Lateglacial (Younger Dryas) glaciers and ice-sheet deglaciation in the Cairngorm Mountains, Scotland: glacier reconstructions and their palaeoclimatic implications

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Lateglacial (Younger Dryas) Glaciers and Ice-Sheet Deglaciation in the Cairngorm Mountains, Scotland: Glacier Reconstructions and their Palaeoclimatic Implications

by

Matthew Richard Standell

A Doctoral Thesis

Submitted in partial fulfilment of the requirements for the award of Doctor of Philosophy of Loughborough University

May 2014

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Abstract

The Cairngorm Mountains contain an outstanding assemblage of glacial landforms from both the deglaciation of the last British–Irish Ice Sheet and the Younger Dryas readvance. Glaciers are recognised as sensitive indicators of past and present climate change and, thus, these landforms provide information about past climate and glacier-climate interaction that can be used to contextualise the present climate change. Previous interpretations have left doubt over the extent and style of the Younger Dryas readvance. In addition, the pattern and timing of deglaciation in the southern Cairngorms and, particularly, how local and external ice masses interacted is unclear.

New geomorphological mapping from aerial images and fieldwork has been compiled in a GIS for a 600km² area of the Cairngorm Mountains. This has allowed a complex pattern of ice-dammed lakes and local and regionally sourced ice margins to be reconstructed during the retreat of the last British–Irish Ice Sheet. The mapping has been combined with new cosmogenic surface exposure ages taken from areas of ‘hummocky moraine’ previously subject to differing age interpretations. The effect of moraine denudation on apparent ¹⁰Be ages has been checked by inverse modelling of the ¹⁰Be concentration vs. boulder height. The results indicate more extensive Younger Dryas glaciation, with glacier reconstructions and equilibrium-line altitudes (ELAs) comparable with the surrounding areas. Reconstruction of both valley and plateau-fed glaciers are presented, with modelling of local topoclimatic factors, such as radiation, avalanche and snow drifting, combined with precipitation gradients, explaining most of the variation within the glacier ELAs. The geomorphological evidence and palaeoclimatic inferences are important, alongside a growing number of palaeoglaciological studies, in acting as evaluation areas for current numerical models of ice-sheet growth and decay.

Keywords: glacial geomorphology; palaeoclimate; Cairngorms, Scotland; Younger Dryas; glacier reconstruction; equilibrium-line altitude; cosmogenic ¹⁰Be surface exposure dating; topoclimatic factor; snow blow.
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Chapter 1: Introduction

1 Introduction

1.1 Introduction

Present warming of climate is unequivocal; it is causing serious changes to the Earth system and poses a real threat to society (IPCC Summary for Policymakers, 2013; IPCC WGII, 2014). The potential for palaeo-studies to contextualise the present climate change and allow us to better understand past changes is becoming increasingly important (IPCC Chapter 6, 2007). In addition, there is a growing drive to use palaeo-data to test current models used in the prediction of climate change e.g. the Paleoclimate Modelling Intercomparison Project 3 (PMIP3) (Braconnot et al., 2012). Glaciers are recognised as being important sensitive indicators of present and past changes in climate (Nesje and Dahl, 2000). The recent decrease in glacier and ice-sheet mass has important implications, both at a local scale for water resources e.g. in the Andes (Bradley et al., 2006) and Asia (Immerzeel et al., 2010), and for global sea-level rise (IPCC Summary for Policymakers, 2013). Although ice sheets have greater potential to increase sea level over long timescales, glaciers and ice caps will continue to make a substantial contribution to sea-level rise in the 21st century (Radić and Hock, 2011). Glaciers and ice caps are projected to contribute 15–35% of total sea-level rise by the end of this century, second only to thermal expansion 30–55% (IPCC Summary for Policymakers, 2013). Former glaciers often leave a geomorphological record of former climate fluctuations which can be used to assess past shifts in the climate (Nesje and Dahl, 2000). These records provide important information about former climate-glacier interactions and can be used to evaluate current models of climate-glacier growth, demise and dynamics e.g. comparing modelling and field data in Sweden (Napieralski et al., 2007), ice stream location in the palaeo- and modelling environment (Stokes and Clark, 2001) and the extent and dynamics of the British–Irish Ice Sheet and Younger Dryas glaciers (Boulton and Hagdorn, 2006; Hubbard et al., 2009; Golledge et al., 2008, 2009).

This study focuses on the final phases of deglaciation of the last British–Irish Ice Sheet and readvances in the Cairngorm Mountains, Scotland. The aim is to provide a detailed account of the pattern, timing and style of glaciers and their relation to
past climatic changes. This study will feature alongside existing studies to provide a
detailed account of the deglaciation and readvance sequences within the British
Isles; these provide a valuable evaluation area for time-variable ice-sheet and
 glacier modelling.

The purpose of this chapter is to introduce the background and rationale for the
study and give an overview of the aims and structure of the thesis.

1.2 Rationale

1.2.1 Palaeoclimate

Palaeo-studies can be used to further our understanding of Earth systems, the
sensitivity of the systems to past changes, the drivers and feedbacks within these
systems and to provide evaluation information for numerical modelling studies
(Harrison et al., 2013). Ice-core and marine records of climate fluctuation are useful
for large regional- and global-scale patterns of climate change (Blockley et al.,
2012). However, local studies are required to understand the local expression of
these changes in smaller regions (Isarin et al., 1998; Marshall et al., 2001). The
spatial and timing variations between locations, including any lag times, can assist
in understanding climate change and the mechanisms that bring about change
(Lane et al., 2013).

This study focuses on the final stages of the Dimlington Stade (c. 31–14.6 ka BP),
the transition into the warmer Lateglacial Interstade (14.6–12.9 ka BP), the abrupt
Loch Lomond (Younger Dryas) (12.9–11.6 ka BP) cooling event and finally the
rapid warming into the Holocene. This is a period of great importance due to the
abrupt climatic changes taking place over decades to centuries, and the availability
of good terrestrial records. The cool Dimlington Stade contained abrupt climate
changes such as Dansgaard-Oeschger oscillations and Heinrich events (Bond and
Lotti, 1995). Greenland’s ice-core records show Dansgaard-Oeschger oscillations
with spacing of approximately 1500 years and often very sudden 6–10°C warming
lasting a few centuries or maybe even decades, followed by slow cooling in the
subsequent millennia (Alley and Clark, 1999; Rial, 2004). The cause of Dansgaard-
Oeschger cycles is thought to be related to ocean convection but the exact
mechanisms are unclear (Alley and Clark, 1999). Less frequent are the abrupt
Heinrich events which are thought to be associated with the discharge of large icebergs into the North Atlantic by the Laurentide ice sheet; however, there are indications that the Greenland, Fennoscandian and British ice sheets may have also been involved (Grousset *et al*., 2000; Stanford *et al*., 2011 and references therein).

At c.19 ka BP the thermohaline circulation within the Atlantic increased to near interglacial levels (Clark *et al*., 2001). However, as indicated by the Greenland ice-core records, the climatic conditions in the North Atlantic region remained cool. The warming is thought to have been delayed owing to the cooling presence of large ice sheets, low concentrations of greenhouse gases and the influx of cold meltwater from the ice sheets impacting on thermohaline circulation (Clark *et al*., 2001; Nesje *et al*., 2004). The abrupt warming associated with the start of the Lateglacial Interstade at 14.6 ka BP is thought to coincide with the end of a meltwater rerouting event through the Hudson River and the increase of thermohaline circulation to interglacial levels (Clark *et al*., 2001; Nesje *et al*., 2004). Another hypothesis is that the warming was associated with the end of a particularly long 4 ka Heinrich event (Stanford *et al*., 2011).

Within the Lateglacial period abrupt climate-cooling events, most notably the short (0.1 ka) Older Dryas and the longer (1.2 ka) Younger Dryas events, are thought to have been caused by releases of meltwater into the North Atlantic (Nesje *et al*., 2004). This meltwater is thought to have been released from the Laurentide ice sheet but the location is uncertain (Clark *et al*., 2001, Nesje *et al*., 2004; Tarasov and Peltier, 2005; Condron and Winsor, 2012). There was also an increase in fresh water from icebergs released through the Hudson Strait and potentially other locations in the North Atlantic (Andrews *et al*., 1995). The termination of the Younger Dryas event is thought to have been ocean led with the return of the increased levels of North Atlantic Ocean circulation (Pearce *et al*., 2013). As shown by the uncertainties above the mechanisms and thresholds concerned with these abrupt climate changes, such as ocean circulation, are not currently well enough understood (IPCC Chapter 6, 2007).

The last phase of major deglaciation in the North Atlantic region was a complex period of abrupt climate changes and feedbacks between ice sheets and ocean
circulation. Understanding this phase is important for contextualising current and future climate shifts and the vulnerability of the Earth system to internal feedbacks. This being the most recent phase of deglaciation, the period possesses a good terrestrial record of glacier-climate interaction, from which geomorphological studies can better understand regional and local climate changes and potential lags between specific sites.

1.2.2 Palaeoglaciology and evaluation of numerical modelling

The study of former ice sheets and glaciers facilitates a better understanding of how glaciers have previously reacted to climate changes. Information on large-scale ice-margin oscillations and the formation and demise of ice sheets only exist within the palaeo-record and have not been recorded in the short period of direct observation. The terrestrial record of former glaciers can allow us to better understand the dynamics of present-day glaciers, such as the spatial variation in thermal regime (Hall and Glasser, 2003; Fabel et al., 2002), location of ice streaming (Stokes and Clark, 2001) and the development of subglacial drainage networks that have important controls on ice-sheet dynamics (Storrar et al., 2014). Such information is often difficult to extract from beneath current glaciers and ice sheets, and thus the terrestrial record provides useful information on previous glacier and ice-sheet behaviour. The BRITICE project (Clark et al., 2004; Evans et al., 2005) and more recently the BRITICE-CHRONO project are working to better constrain the pattern and timing of the retreat of the last British–Irish Ice Sheet; the aim being to better understand the dynamics of the ice sheet’s retreat, such as whether it stabilised once onshore and whether it reacted to climatic changes synchronously across its surface (BRITICE-CHRONO). Given the high-resolution palaeotemperature record from ice cores, the dated retreat margins provide a unique opportunity, only available through the study of the palaeo-record, to evaluate current models of climate-glacier interaction, and consequently lead to better predictions of sea-level rise (BRITICE-CHRONO).

Ice-sheet models have been previously applied to formerly glaciated areas with good terrestrial records. This includes comparing Last Glacial Maximum (LGM) and Younger Dryas end moraines and glacier lineations with modelled outputs for the Fennoscandian ice sheet (Napieralski et al., 2007). The British geomorphological
record of palaeo-ice-streams has been used to infer the properties of the former ice by utilising ice-sheet models (Boulton and Hagdorn, 2006). The extent and dynamics of the British–Irish Ice Sheet have also been modelled from 38–10 ka BP (Hubbard et al., 2009). A high-resolution model of glaciation during the Lateglacial Interstade and Younger Dryas Stade has been generated for Scotland (Golledge et al., 2008, 2009). These models include both time-dependent variations in glacier margins and also details of ice-flow directions, velocity and the basal thermal regime which have been compared with the geomorphological record (Golledge et al., 2009; Golledge, 2010; Boston et al., 2013). This indicates the potential for well time-constrained landform records to be utilised as evaluation areas for climate-driven ice-sheet and glacier models.

1.2.3 Cairngorm Mountains

The landform record within the Cairngorm Mountains possesses a diverse wealth of information about past climate change (Kirkbride and Gordon, 2010) and, consequently, there is a substantial body of work on the palaeoglaciological record. The Cairngorm Mountains are a landscape of selective linear erosion comprising the highest plateaus within the British Isles and deeply incised glacial troughs (Rea, 1998). Situated on the eastern side of the Western Highlands, the Cairngorm Mountains, despite their altitude, receive less precipitation than the mountains to the west. This pattern has most likely been important throughout the evolution of the Cairngorm landscape, impacting on the basal thermal regime and mass balance of glaciers (Sissons, 1979a; Rea, 1998; Hall and Glasser, 2003).

Although there are important uncertainties, the Cairngorms supported relatively restricted Younger Dryas glaciers compared to the Western Highlands. Sugden (1970) favoured only limited corrie glaciation, whereas Sissons (1979a) favoured additional synchronous glaciation of the upper valleys. There has also been uncertainty surrounding whether the glaciers were plateau fed during the Younger Dryas (Bennett, 1996). While there has been a recent increase in palaeoglaciological studies in upland areas of the British Isles, and many developments in our understanding of glacier style, dynamics and climate records – the Cairngorm Mountains have remained a site of great uncertainty. It is important that accurate glacier margins are established and the style of glaciation understood.
to make robust palaeoclimatic inferences and to evaluate ice-sheet models such as that by Golledge et al. (2008, 2009). The distinctive landscape of the Cairngorms and its location in a precipitation shadow means it is a valuable site for understanding the factors that are important to glaciation in marginal areas and plateau-dominated landscapes.

Owing to the restricted Younger Dryas glacier limits, the Cairngorms possess an important record of pre-Younger Dryas glacier-climate interaction that is not preserved in upland areas that experienced more extensive Younger Dryas glaciation further west. Previous work has focused on the landforms and interaction of local and regionally sourced ice in the northern Cairngorms (Brazier et al., 1998; Golledge, 2002), while the southern Cairngorms have received relatively little attention. A full synthesis of existing research and outstanding uncertainties is given in Chapter 2.

The work to date has made a significant contribution to the understanding of glaciation within the Cairngorms; however, it has often been piecemeal in approach. Our understanding of palaeoglaciology and palaeoclimate has evolved, and previous work has left contentious issues outstanding. There is great uncertainty surrounding the extent and style of Younger Dryas glaciers, and our understanding of ice-sheet deglaciation within the southern and central Cairngorms is limited. The better control of glacier limits and style during this period is important in providing a time-constrained record of glacier fluctuations. This will give location-specific information on former climate change, enabling the relative timing and size of glacier fluctuations to be compared regionally, and explanatory climate mechanisms to be discussed (Kirkbride and Winkler, 2012). High-resolution temperature records can often be reconstructed from biological proxies (e.g. Brooks et al., 2012); however, former glaciers provide location-specific information on former precipitation levels. This can be used to understand past variations in climate and weather systems that can assist in contextualising present climate change. The time-constrained landform record will also enable the evaluation of numerical models of glacier-climate response. Through the better understanding of past mechanisms of climate change, glacier retreat dynamics, and the development of numerical models, it is anticipated our understanding of future climate change and the associated glacier and sea-level predictions can be improved.
1.3 Thesis Aims and Structure

1.3.1 The aims and research objectives

Despite the wealth of research that has been undertaken within the Cairngorm Mountains, it has often been contentious and there remain important unresolved questions.

This project aims to resolve uncertainty around the nature and timing of the deglaciation of the Cairngorm Mountains. Using a multi-disciplinary approach, including geomorphological mapping, surface exposure dating, glacier reconstruction and modelling of topoclimatic factors, the project aims to reconstruct the pattern of glacial retreat prior to the Lateglacial Interstadial, and the style and extent of the Younger Dryas readvance (12.9–11.6 ka BP). This will inform palaeoclimatic inferences such former mechanisms of climate change, and spatial and temporal variations in precipitation, both during ice-sheet retreat and the Younger Dryas readvance.

Research objectives

1. Determine the pattern, style and dynamics of glacier retreat within the Cairngorm Mountains during the retreat of the British–Irish Ice Sheet; including how locally and regionally sourced ice masses interacted.

2. Determine the extent, style (plateau- versus valley-sourced) and retreat dynamics of glaciation during the Younger Dryas readvance (12.9–11.6 ka BP). Reconstruct the glacier equilibrium-line altitudes (ELAs) and determine the topoclimatic factors that were important in supporting glacier formation.

3. Draw inferences on the palaeoclimatic conditions prior to the Lateglacial Interstadial and reconstruct the Younger Dryas palaeoclimatic conditions using innovative topoclimatic analysis; particularly focusing on Younger Dryas precipitation gradients.
1.3.2 Thesis structure

This chapter has introduced the rationale and aims of the project. Chapter 2 provides a detailed account of the research to date within the Cairngorm Mountains and the work’s context in relation to wider issues. The methods used to address the research objectives are set out in Chapter 3. These include remote and fieldwork mapping, cosmogenic surface exposure dating, glacier reconstruction and modelling of topoclimatic factors. Chapter 4 presents the pattern and relative timing of ice retreat, including how local and external ice sources interacted and the location of potential readvances. Chapter 5 presents the new geochronological control for the final phase of valley glaciation and discusses the results in relation to climate data and sources of uncertainty. A synthesis of the mapping, dating and landsystems approach facilitates the proposal of Younger Dryas glacier margins presented in Chapter 6. The chapter also presents the topoclimatic factors that explain some of the variation in Younger Dryas glacier ELA. Chapter 7 includes a synthesis of the ice-sheet retreat margins and how locally and regionally sourced ice interacted. It also includes a synthesis of Younger Dryas glaciation and the implications of the research are discussed in relation to the wider geographical region and palaeoclimatic data. A summary of the research findings and the concluding remarks are made within Chapter 8.
2 The Lateglacial History of the Cairngorm Mountains, Scotland: a Synthesis and Review of Outstanding Research Questions

The chapter aims to synthesise the previous geomorphological and geochronological research within the Cairngorm Mountains, but it also draws on the developments in palaeoclimatic research and numerical modelling to identify areas of uncertainty that require further work.

2.1 Introduction

The landscape of the Cairngorm Mountains is ‘exceptional in Western Europe’ for its outstanding assemblage of glacial geomorphological features (Scottish Natural Heritage and Cairngorms National Park Authority, 2010 p.15) that record a wealth of information about past environmental change and landscape evolution (Kirkbride and Gordon, 2010). As a consequence, the palaeoglaciology of the region has been the subject of extensive research since the glacial theory became widely accepted in the mid-19th Century.

Almost 150 years ago, Jamieson (1865) accepted that landforms in Glen Derry and along the northern flanks of the Cairngorms were most likely of glacial origin. Early work focused on the location of erratics, which were used to infer ice-flow directions (Jamieson, 1908; Bremner, 1929). By the mid-20th century, glacial landforms had been used to identify stages of confluent ice-sheet glaciation and the subsequent separation into valley glaciers, along with a later corrie-glacier stage (Hinxman and Anderson, 1915; Bremner, 1929). More recently, the focus has turned to determining the extent of Younger Dryas glaciation and the significance of ‘hummocky moraine’ to this debate. Albeit a much smaller Younger Dryas extent than further west; Sugden (1970) favoured glaciation in only the highest corries, whereas Sissons reconstructed both corrie and valley glaciers (Sissons, 1979a).

Although the general pattern of glaciation had been established by the mid-20th century, it was not until the 1970s that radiocarbon dating provided the first constraints on the timing of events (Sissons and Walker, 1974; Clapperton et al.,
1975). A good review of geomorphological work in the Cairngorms is provided by Gordon (1993); however significant progress has been made during the last 15 years with the successful use of both cosmogenic nuclide surface exposure dating (Phillips et al., 2006; Everest and Kubik, 2006; Ballantyne et al., 2009a; Kirkbride et al., 2014) and optically stimulated luminescence (OSL) dating (Everest and Golledge, 2004) in the Cairngorms. These techniques have been used to provide considerable detail on particular aspects of the glacial history of the Cairngorm Mountains, but there has not been any systematic review combining the geomorphological and geochronological evidence across the region as a whole.

The aim of this chapter, therefore, is to review the state of current knowledge about the pattern, timing and dynamics of glacier recession and readvance in the Cairngorm Mountains during that period for which direct geomorphological evidence is available (that is, since the Last Glacial Maximum: LGM). The emphasis is on geomorphological and geochronological work, but we also draw upon developments in palaeoclimatic records and numerical modelling. The review concludes with an assessment of current areas of uncertainty requiring further work.

2.2 Geology and Landscape of the Cairngorm Mountains

Figure 2.1 Relief of the Cairngorm Mountains and their location compared to adjacent mountainous regions and within the British Isles. Rectangular box represents the extent of Figures 2.3–5
Chapter 2: The Late-Glacial History of the Cairngorm Mountains, Scotland

The Cairngorm Mountains (Figure 2.1) comprise the most extensive area of land above 1000m in the British Isles (Countryside Commission for Scotland, 1978), including five of their six highest peaks. They are composed almost entirely of Cairngorm Granite, formed as a large pluton into the metasedimentary rocks of the Dalradian Supergroup in the late stages of the Caledonian Orogeny (around 427 million years ago; Thomas et al. 2004). Their morphology is characterised by an extensive montane plateau with a gently rolling morphology of fluvioglacial origin with deeply weathered surfaces on which are found numerous tors, the presence of which has been used to argue that the depth of Quaternary glacial erosion was very limited (Glasser, 1995; Hall and Phillips, 2006). Cut into the plateau surface are deep glacial troughs with depths of up to 500 m, resulting in an impressive landscape of selective linear erosion (Rea, 1998), although their origins have been shown to predate Quaternary glaciation (Thomas et al., 2004, Hall and Glasser, 2003).

The national and international importance of this landscape has been recognised by its designation as a Site of Special Scientific Interest for its Quaternary geology and geomorphology, and as a National Scenic Area for its “outstanding scenic value” (Planning etc. (Scotland) Act 2006). The Cairngorm Mountains form the heart of the Cairngorms National Park, established in 2003.

2.3 Lateglacial Glaciation in Scotland

2.3.1 Late Devensian (Dimlington Stade)

Sea-level records derived using dated coral from marine cores indicate that the global LGM occurred at 26 ka BP, somewhat prior to the conventional age of 21 ka BP (Peltier and Fairbanks, 2006). However, the response of individual ice bodies was not necessarily synchronous, even within individual ice sheets. Indeed, while the British–Irish Ice Sheet reached its northern and western maximum extent (and maximum areal extent) at approximately 27 ka BP (reaching the edge of the continental shelf), its southern maximum extent was not reached until 23 ka BP (Clark et al., 2012).

At the LGM, an ice dome sat over the Cairngorm Mountains. Numerical ice-sheet models indicate that this would have been in a cyclic state of being drawn down and
recovering from multiple competing ice streams (Hubbard et al., 2009). Early in deglaciation (19.4–17.4 ka BP), this modelling suggests that all areas of the British–Irish Ice Sheet underwent active, dynamic and rapid thinning and retreat caused by warmer climatic conditions, followed by a period of stabilisation at round 17 ka BP (Hubbard et al., 2009). Geomorphological evidence indicates that Irish and Scottish ice had separated by 16 ka BP, and the central valley of Scotland was ice free by 15 ka BP (Clark et al., 2012). By this stage, Cairngorm ice had started to dynamically separate from that to the west, and large parts of eastern Scotland were ice free (although ice still reached the coast in the west).

### 2.3.2 Lateglacial Interstade and the Younger Dryas (Loch Lomond) Stade

Greenland ice-core records indicate that the warmer Lateglacial Interstade (GI-1) began shortly after 15 ka BP, but was penetrated by brief colder periods and terminated with the onset of the Younger Dryas (Loch Lomond) Stade at 12.9 ka BP (GS-1) (Lowe et al., 2008). Chironomid- and Coleoptera-inferred mean July air temperatures from lake records at various sites in Scotland broadly correlate with the Greenland ice-core records (Brooks and Birks, 2000; Marshall et al., 2002; Brooks et al., 2012). At Abernethy Forest (located to the north of the Cairngorms at a height of 230m asl) a maximum temperature of 13.6°C was achieved at the start of the Interstade, falling to 6.8°C by the beginning of the Younger Dryas (Brooks et al., 2012). These temperatures are estimated using a modern Norwegian chironomid-based temperature calibration dataset consisting of 157 lakes (Brooks et al., 2012). The temperature model applied has a root mean squared error of prediction of c.1°C (cf. Brooks et al., 2012). Some caution should be acknowledged when using these temperatures due to the assumptions and uncertainties associated with using biological proxies for temperature reconstruction (see Brooks and Birks, 2001; Juggins, 2013).

Recent numerical glacier modelling has indicated that small, isolated, ice masses may have survived the Lateglacial Interstade in parts of western Scotland (Hubbard et al., 2009), or that small glaciers may have formed during colder parts of the Interstade (Golledge et al. 2008). Surface exposure dating has suggested that large remnants of the Devensian ice sheet may even have survived (Bradwell et al., 2008; Ballantyne et al., 2009b), although new recalibration of these ages calls this
interpretation into question (Ballantyne and Stone, 2012). Whether or not ice survived the Interstadial, the cold conditions of the Younger Dryas (12.9–11.6 ka BP) led to the growth of numerous small glaciers in the Scottish uplands (including the Cairngorm Mountains), and the formation of an extensive icefield over the Western Highlands (Golledge, 2010).

Figure 2.2 GRIP ice-core records plotted using the GICC05 time scale for years before 2000 AD (Rasmussen et al., 2008). Ice-core stages and climatic events as proposed by Lowe et al. (2008). Data available at www.icecores.dk

2.4 Glacial Geomorphology in the Cairngorm Mountains

The following discussion of existing research in the Cairngorms has been organised primarily by the stage of deglaciation and secondarily by location.

2.4.1 Last Glacial Maximum (LGM) and early deglaciation (c. 20–19 ka BP)

During the LGM the Cairngorm Mountains were beneath the British–Irish Ice Sheet. The presence of numerous tors on the central and eastern summits, although showing some evidence of glacial modification, indicates that the depth of Quaternary glacial erosion on the plateau surfaces was very limited (Glasser, 1995; Hall and Phillips, 2006), and surface exposure dating confirms the tors have survived multiple glacial cycles (Phillips et al., 2006). The preservation of these tors
provides evidence that the plateau and summit surfaces were predominantly exposed to cold-based glacial conditions, a conclusion supported by numerical modelling of the basal thermal regime (Glasser, 1995; Hall and Glasser, 2003).

The absence of ‘foreign’ erratics on the highest mountains of Ben Macdui and Beinn a’Bhuird (in the central and eastern Cairngorms respectively) was noted by Jamieson (1908) and provides evidence that the mountains acted as a centre of ice dispersal (Bremner, 1929; Hubbard et al., 2009). However, early work by Bremner (1912) identified moraines near Loch nan Cnapan on the Mòine Mhòr plateau that consist of schist originating to the south. On the basis of meltwater channels and schist erratics found on the Meirleach col and in upper Glen Luibeg, Sugden (1970) tentatively reconstructed the flow of ice from the west, crossing the Mòine Mhòr plateau before diverging around the radial drainage of the Cairngorm dome (located over Ben Macdui), via Gleann Einich and Glen Geusachan (Figure 2.3). The absence of schist erratics within Glen Geusachan can be explained by the flushing of material by a locally sourced glacier at a later stage of deglaciation (Sugden, 1970; Bennett and Glasser, 1991). After the LGM, early in deglaciation, it is thought the Cairngorm glaciers would have contributed to the Strathspey (to the north) and Deeside (to the south) outlet glaciers, which would have discharged large volumes of ice (Bremner, 1929).
2.4.2 Dimlington Stade glaciers: the northern Cairngorms and Glenmore lobe (c.18–15 ka BP)

The mountains of the eastern Cairngorms were the first summits to become deglaciated (Brazier et al., 1996b). The best estimate for the timing of this deglaciation is 15.6 ka BP based on surface exposure dates from a single erratic boulder on the summit of Ben Avon, along with two other inheritance corrected dates (Phillips et al., 2006). This date provides an indication of when summits may have become ice free, but local icefields and glaciers are likely to have persisted after this (Brazier et al., 1996b).

Bremner (1929) argued that, following the recession of ice from the eastern summits, the recession of the Glenmore lobe of the Strathspey outlet glacier, and its interaction with the Cairngorm glaciers, is one of the most interesting episodes of
the country's glacial history (see Figure 2.4 and Table 2.1). Early accounts described the presence of the Glenmore lobe leaving lateral moraines high on the northern flanks of the Cairngorm Mountains, and deltaic deposits associated with ice-dammed lakes in Gleann Einich and the Lairig Ghru (Hinxman and Anderson, 1915; Bremner, 1929).

The highest, and apparently earliest, landforms left by the Strathspey glacier on the northern flanks of the Cairngorms are an 800m-long moraine at 940m asl in Coire an t-Sneachda (Brazier et al., 1996b; Brazier et al., 1998), and meltwater channels starting in the west on the col north-west of Sgòran Dubh Mòr at 885m and decreasing in height to the east (Sugden, 1970). Brazier et al. (1998) suggested that each Cairngorm glacier on the northern margin retreated actively, starting in the east, forming ice-dammed lakes in the Lairig Ghru and Gleann Einich but not in Strath Nethy and Glen Feshie. Early in deglaciation, Spey ice entered Strath Nethy leaving a lateral moraine high on the east side of the valley (Brazier et al., 1996b) and eskers containing schist erratics (Sugden and Clapperton, 1975). In addition, a flight of kame terraces mark the thinning of the Nethy ice, while ice and meltwater from the Glenmore lobe continued to overtop the ridge to the west (Sugden and Clapperton, 1975; Brazier et al., 1996b). Surface exposure dating of a rock slope failure (previously interpreted as a rock glacier) suggests lower Strath Nethy was deglaciated by 16.2±1.0 ka based on the uncertainty-weighted mean of two internally consistent ages (Ballantyne et al., 2009a). Recalibration of these ages using the North-West Highlands (NWH) 11.6 local production rate, Lm scaling and 1 mm/ka erosion rate yield ages of 16.7±1.3 ka and 17.2±1.2 ka. The third recalibrated age at the sample site is 20.0±1.3 ka; this may indicate earlier deglaciation of the valley or, as favoured by Ballantyne et al., (2009a), inheritance from prior exposure. These dates also provides a constraint on the recession of ice into Glen Avon, which by this time was no longer overtopping the saddle and feeding ice down Strath Nethy.

Further west, two ice-dammed lakes formed between the Glenmore lobe of the Strathspey glacier and the glaciers occupying Lairig Ghru and Gleann Einich (Brazier et al., 1998). Evidence of the sequential east-to-west break-up of local and external ice is provided by meltwater channels sourced on the western col, with Gleann Einich linking to lacustrine sediment in the Lairig Ghru ice-dammed lake
The Gleann Einich glacier must therefore have been large enough, at least initially, to supply the material and was most likely confluent with the Glenmore lobe. A similar situation occurred later between ice in Glen Feshie and the Gleann Einich ice-dammed lake (Brazier et al., 1998). That these lakes are of different ages implies there must have been at least two readvances or major stillstands in the retreat of the Glenmore lobe (Brazier et al., 1998; Everest and Golledge, 2004; Everest and Kubik, 2006).

Bremner (1912) and Brazier et al. (1996b) noted that schist boulders occur throughout deposits in Gleann Einich, indicating the Einich glacier was an outlet lobe from the ice sheet spilling over the Mòine Mhòr plateau from the south-west. However, after examination of the proglacial foreset beds in both the Lairig Ghru and Gleann Einich deltas no schist was found, although this may need further confirmation (Brazier et al., 1998). Thus the Cairngorm glaciers that terminated in the ice-dammed lakes were locally sourced, received no external ice from the ice sheet to the south (Brazier et al., 1998), and must have looked similar to the reconstruction of a western Cairngorm ice cap presented by Everest and Kubik (2006). This is important, as understanding glacier-source areas and potential connections to larger ice masses is essential if glacier ELAs are to be used for climate reconstructions.
Figure 2.4 Dimlington Stade glacier margins and reconstructions. Radiocarbon age from Clapperton et al., (1975) and cosmogenic surface exposure ages recalibrated from Everest and Kubik (2006) and Ballantyne et al. (2009a). Everest and Kubik (2006) ice margin redrawn with permission of Wiley.
Table 2.1 Authors for ice margins and flow directions contained within Figure 2.4 and 2.5

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2.4.2.1 Timing and uncertainty associated with the separation of Cairngorm glaciers and the Glenmore lobe

Within Gleann Einich itself, Golledge (2002) suggested there were advances of the Gleann Einich glacier as implied from deformation recorded in sediment exposures. Golledge (2002) concluded that an early lake formation occurred, followed by an overriding readvance of the Gleann Einich glacier and the subsequent formation of a second lake. Golledge (2002) pointed out that, if exposure ages (Everest and Kubik, 2006) correctly constrain the later lake formation, the complete retreat-readvance-retreat oscillation must predate this event. Everest and Kubik (2006) reported exposure weighted mean ages from granite boulders of 14.4±0.4 ka for the Glenmore moraine (sourced from regional ice), and 13.1±0.8 ka for the Gleann Einich moraine (sourced from local Cairngorm ice; see Figure 2.4). Climate records show rapid warming at the onset of the Lateglacial Interstade (14.6 ka BP), and the original radiocarbon date 15.5±1.1 cal ka BP (Sissons and Walker, 1974) and a new radiocarbon date 15.2±0.7 cal ka BP (Everest and Golledge, 2004) from the kettle infill at Loch Etteridge (26 km west of the Cairngorms up Strathspey) make some of the younger boulder ages climatically and glaciologically unlikely (Everest and Kubik, 2006). OSL ages from ice-dammed lake sediments in the Lairig Ghru 16.7±5.2 ka BP and Gleann Einich 16.7±0.54 ka BP also suggest the cosmogenic surface exposure ages are too young (Everest and Golledge, 2004). Thus Everest

Recent recalibration of exposure ages in north-west Scotland, using locally derived $^{10}$Be production rates, has shown previous exposure ages to require recalibration, the equivalent of adding in the order of 6.5–12% to previously calculated ages (Ballantyne and Stone, 2012). The Cairngorm cosmogenic surface exposure ages have been recalibrated with the use of the NWH 11.6 local production rate, Lm scaling and 1mm/ka erosion rate to yield ages from 16.3±1.3 ka to 14.6±1.1 ka for the Gleann Einich moraine and 16.0±1.5 ka to 14.2±1.2 ka for the Glenmore moraine (Figure 2.4). In the lower section of the Lairig Ghru, a rock slope failure (RSF) up-valley of the Lairig Ghru moraine limits has been dated using cosmogenic surface exposure dating. The landform should yield ages consistent with the earlier retreat of the Lairig Ghru glacier after lake drainage, as dictated by geomorphological evidence (Brazier et al., 1998). The results yield an age of 15.4±0.8 ka (recalibrated as above to 17.7±1.6 ka to 15.0±1.0 ka); the younger cosmogenic surface exposure ages in this range are in agreement with OSL lake-sediment ages (Ballantyne et al., 2009a), suggesting the earlier formation of an ice-dammed lake 16–17 ka BP, followed by glacier retreat and a subsequent RSF. It has been highlighted that the difference in age between the older Lairig Ghru RSF exposure ages and younger moraine exposure ages in Gleann Einich are difficult to explain (Ballantyne et al., 2009a; Ballantyne, 2010). The local production rate has caused all the surface exposure ages to become older and thus shifted the boulder ages, particularly the older boulders within each dataset, to become more in line with the radiocarbon date for Loch Etteridge. This highlights the advantages of using multiple dating techniques to constrain an event.

2.4.3 Dimlington Stade glaciers: the southern Cairngorms and the Deeside ice mass (c.18–15 ka BP)

The landforms on the southern margin of the Cairngorms have not received the recent focus of work seen to the north, but it has been suggested they may record a similar set of events. This includes the complex relationship between the presence of a large ice mass in Deeside sourced from the Geldie-Feshie watershed, Gaick and Glen Shee hills, locally sourced Cairngorm glaciers and the formation of ice-
dammed lakes (Kirkbride and Gordon, 2010). There is widespread evidence of the Deeside glacier on the southern flanks of the Cairngorms consisting of moraines and meltwater channels; Sugden (1970) described channels at 1000m on Carn a’ Mhaim, 1100m on Monadh Mòr and across the spur to the south of Derry Cairngorm as indicating the height reached by the southern ice masses. An established sequence of events is not clear for the southern margin, thus evidence has been subdivided below, starting in the eastern Cairngorms.

2.4.3.1 Glen Quoich and Dubh-Ghleann

In the south-east Cairngorms, landforms have been used to suggest a complex glacial reconstruction whereby a locally sourced glacier draining the corries of Beinn a’ Bhuirid terminated in a lake dammed by ice in Deeside (Bremner, 1912; Kirkbride and Gordon, 2010). Bremner (1912) noted that moraines at the separation of Glen Quoich and Glen an t-Slugain marked the presence of a glacier sourced in the corries of Beinn a’ Bhuirid (Figure 2.4). Later, Kirkbride and Gordon (2010) recognised the presence of deltaic-like terraces in the area and suggested that geomorphological evidence in the centre of Glen Quoich may indicate that a glacier first calved into an ice-dammed lake and later became grounded in the area.

Approximately 5 km west, but within the same valley system, a rocky gorge passes through a col from Glen Derry to a shallow lake named Poll Bhàt situated on a flat triangular deposit (Figure 2.4), the apex of which terminates in a scree-like face to Dubh-Ghleann beneath (Cunningham-Craig, 1898). Initially it was believed the deposit could only be accounted for by the presence of a glacier in Dubh-Ghleann, allowing deposition to accumulate either against it or in a small lake between the glacier and valley side (Cunningham-Craig, 1898). In addition, Bremner (1912) suggested material may have been partly deposited by the Dubh-Ghleann glacier. An alternative hypothesis is that the delta formed in the same lake as discussed above, which was dammed by ice in Deeside (Kirkbride and Gordon, 2010). The channel that supplied sediment to the delta has no local source and begins at the col with Glen Derry, thus it was identified as a meltwater channel from the Derry glacier (Bremner, 1912). Recently, Kirkbride and Gordon (2010) suggested that the landforms, formerly interpreted as lateral meltwater channels (Sugden, 1970) that fed the overflow channel, are lake shorelines; thus, proposing the hypothesis that
an ice-dammed lake existed in Glen Derry which overflowed into another lower ice-dammed lake occupying part of Dubh-Ghleann and Glen Quoich (Kirkbride and Gordon, 2010).

Three kilometres to the south, the Clais Fhearnaig channel joins Glen Lui and Glen Quoich. Above the main Clais Fhearnaig channel on the northern side there are a series of meltwater inlets either joining up with the main Clais Fhearnaig channel or making their own way across the col from Glen Lui to Glen Quoich. On the Lui side, Bremner (1912, 1929) described the flat intake, low moraines and depicted the start of lateral meltwater channels that flow down towards Glen Quoich where fluvio-glacial deposits are shown. Thus the Lui glacier most probably flowed east into Glen Quoich and the lowering of channels may mark the lowering of the glacier surface. An additional hypothesis is that the water carried in the channels may be associated with the recently proposed ice-dammed lake in Glen Derry.

### 2.4.3.2 Coire Etchachan and Glen Derry

In 1912, Bremner noted the moraines in Glen Derry and Lui for their conspicuous size and freshness. Bremer (1929) later used them to illustrate four long and three short pauses in the retreat of Cairngorm glaciers. Near Derry Lodge, it has been suggested that the large ridges mark a retreat position of a glacier from Glen Derry (Bremner 1929); this would imply that both external ice from Deeside and ice sourced over the Preas nam Meirleach col must have been absent during formation and since (Figure 2.4). Two kilometres up-valley, Derry Dam is a large moraine 500m wide; this moraine spans the valley and would have dammed ‘Loch Derry’ which shrank in size as the stream cut down through the eastern end of Derry Dam (Bremner, 1929). This moraine is thought to mark another pause or readvance of the Derry glacier during deglaciation, prior to the Lateglacial Interstade (Bremner, 1929; Sugden, 1970). The ‘hummocky moraine’ near the head of Glen Derry was suggested by Sugden (1970) to be formed by the ice sheet down-wasting in situ; although Midgley (2001) suggested it was formed by active retreat and of Younger Dryas origin. Up-valley of the ‘hummocky moraine’, moraines are present between Glen Derry and Loch Etchachan which suggest the retreat of a glacier up into the corrie that holds Loch Etchachan (Bremner, 1929). In addition, Bremner (1929) noted the presence of a lateral moraine below Creagan a’ Choire Etchachan, and
Midgley (2001) discussed separate ridges in upper Glen Derry which most likely mark the last stages of glaciation sourced in Glen Derry.

### 2.4.3.3 Glen Allt Cumh na Còinnich and Slochd Mòr

The eastern plateau of Mòine Bhealaidh is thought to have sourced glaciers draining north into the Avon basin (Kirkbride and Gordon, 2010). Mapping of meltwater channels suggests the presence of a plateau ice source on Mòine Bhealaidh with an outlet glacier to the north moving down Allt Cumh na Còinnich (Kirkbride and Gordon, 2010). Despite less geomorphological evidence, it is likely the same plateau glacier fed a glacier south into Dubh-Ghleann. How such a glacier relates to the ice-dammed lake evidence described above is unclear. There is also evidence for a glacier existing in Slochd Mòr and large terraces at the confluence with Glen Avon (Kirkbride and Gordon, 2010). Kirkbride and Gordon (2010) suggested the terraces may have been formed as deltas within an ice-dammed lake blocked by the Builg glacier to the east, but highlighted the need for further work.

Glaciation in the eastern Cairngorms is not well constrained; a radiocarbon date from bulked gyttja cored from an infilled kettle hole near Loch Builg dated at 14,050–13,450 cal yr BP, combined with the stratigraphy and pollen analysis (Clapperton et al., 1975), was used to imply that lower valleys in the eastern Cairngorms were deglaciated prior to approximately 14 ka BP (Phillips et al., 2006). Local glaciers sourced in the mountains of Beinn a’ Bhuird and Ben Avon had retreated by this time, and no externally sourced ice from the main Deeside glacier was present.

### 2.4.3.4 Luibeg

Glen Luibeg lies 3 km west of Glen Derry (Figure 2.4); at its entrance, a large moraine with a steep down-valley face is interpreted to mark the position of a glacier in Glen Dee flowing across the Preas nam Meirleach col, damming Glen Luibeg (Golledge, 2003). A delta surface at 545m exists behind the moraine which is interpreted to mark meltwater flowing northwards into the lake over the moraine. Work on an exposure revealed laminated deposits at 530m which were used to imply a lake depth of 15m (Golledge, 2003). Further up Glen Luibeg, moraines with ice-contact faces on their northern sides and a 250m esker ridge mark the retreat of
a smaller valley glacier; however, it is not known if these ice margins are contemporaneous with the damming of lower Glen Luibeg (Golledge, 2003).

### 2.4.3.5 Glen Dee

The age of glacial landforms in Glen Dee and the valleys and corries that join it have been the subject of extensive debate. Here the landforms that have solely been attributed to deglaciation are discussed; the remainder are discussed in Section 2.4.5. Lake shoreline fragments have been noted in Glen Dee to the south of the confluence between Glen Geusachan and Glen Dee (Kirkbride and Gordon, 2010). These are not to be confused with the separate damming of a lake further north within Glen Dee by a glacier exiting Glen Geusachan (Sissons, 1979a; Bennett and Glasser, 1991).

### 2.4.3.6 Synthesis

The local glaciers of the southern Cairngorms retreated while regional ice, sourced to the west, blocked the lower valleys, forming large ice-dammed lakes. Numerous large moraine assemblages exist in the southern Cairngorms attributable to periods of stillstand or readvance prior to the Lateglacial Interstadial. However, the sequence of events forming these margins, the configuration of ice masses responsible for the formation of ice-dammed lakes and their timing are uncertain. Together the evidence suggests an early, active retreat of Cairngorm glaciers, most likely caused by precipitation starvation, punctuated by possible stillstands and readvances.

### 2.4.4 Final deglaciation and Younger Dryas glaciation (16–11 ka BP)

With the exception of a very restricted phase of glaciation during the Little Ice Age (Harrison et al., 2014; Kirkbride et al., 2014), the last glaciers in the Cairngorm Mountains are thought to have existed during the Younger Dryas Stade. However, the extent of Younger Dryas glaciation is uncertain, and has been the subject of prolonged debate (Sugden, 1970; Sugden and Clapperton, 1975; Sissons, 1979a; Bennett and Glasser, 1991; Bennett, 1996; Purves et al., 1999; Midgley, 2001; Everest and Kubik, 2006; Golledge et al., 2009). The key issue is whether ice was limited to a few of the high corries (as proposed by Sugden, 1970 and Sugden and
Clapperton, 1975), or reached the floor of some of the major glens (as argued by Sissons, 1979a).
Figure 2.5 Margins and reconstructions discussed in relation to Younger Dryas glaciation. Cosmogenic surface exposure ages recalibrated from Everest and Kubik (2006); Ballantyne et al. (2009a)

Everest and Kubik (2006) ice margin redrawn with permission of Wiley

Sissons (1979) ice margins redrawn with permission of Taylor and Francis (www.tandfonline.com)
2.4.5 Western Cairngorms

Sissons (1979a) and Sugden (1970) agreed on the location of three Younger Dryas Stade corrie glaciers in the western Cairngorms (Figure 2.5). In addition, Sissons (1979a) reconstructed three large Younger Dryas Stade valley glaciers in the area, and considerable discussion has occurred regarding the extent, style and timing of glaciers in the area since (Bennett and Glasser, 1991; Bennett, 1996; Brazier et al., 1996b; Midgley, 2001). On the western Cairngorm plateau, Sugden (1970) described the presence of granite moraines on schist bedrock. These were interpreted as marking the westward movement of a plateau glacier towards Glen Eidart, after the Cairngorms had become an independent ice cap during deglaciation, but prior to the Lateglacial Interstade (Sugden, 1970). Based on the moraines in upper Glen Geusachan consisting entirely of granite, Sugden (1970) believed a contemporaneous glacier existed in Glen Geusachan. Bennett and Glasser (1991) suggested the absence of schist erratics stretches as far as the moraines on the western side of Glen Dee north of Allt Garbh. This limit is several hundred metres south of the Younger Dryas ‘hummocky moraine’ mapped by Sissons (1979a). After the existence of a plateau glacier (Sugden, 1970), marked by lake shorelines and overflow channels to the west, an ice-dammed lake formed adjacent to the head of Glen Geusachan (Figure 2.5) (Bennett and Glasser, 1991). Based on the clarity of the lake shoreline cut into otherwise soliflucted material, and the depth and frequency of overflow channels, the ice-dammed lake was thought to be a prolonged event, and Bennett and Glasser (1991) favoured formation during the Younger Dryas Stade. It was suggested that the anomalous size of the Glen Geusachan Younger Dryas Stade glacier, originally reconstructed by Sissons (1979a), may be attributable to ice having survived the Lateglacial Interstade due to favourable valley depth and aspect (Bennett and Glasser, 1991; Bennett, 1996).

The second largest Younger Dryas Stade glacier reconstructed by Sissons (1979a) was in Garbh Choire; geomorphological evidence suggested it terminated in an ice-dammed lake formed by the presence of a glacier exiting Glen Geusachan and damming Glen Dee (Sissons, 1979a; Bennett and Glasser 1991; Midgley, 2001). There is a distinct set of meltwater channels cut around the former terminus of the Glen Geusachan glacier marking the overflow of lake water around the glacier (Bennett and Glasser, 1991; Sissons, 1979a). Exposure ages from boulders
between channels range from 16.3±1.8 ka to 12.0±1.5 ka with a mean age of 13.7±1.6 ka (recalibrated as above to 17.4±1.1 ka to 12.8±1.1 ka, recalculated arithmetic mean 14.6 ka), suggesting deposition from a glacier prior to the Younger Dryas readvance (Everest and Kubik, 2006). The link between the geomorphological evidence of Glen Geusachan and Garbh Choire has always been used to imply synchrony, whether they favoured the Younger Dryas Stade (Sissons 1979a; Bennett and Glasser 1991; Bennett, 1996; Midgley, 2001) or formation prior to the Lateglacial Interstade (Everest and Kubik, 2006). Sugden (1970) favoured separate Younger Dryas Stade corrie glaciers in upper Glen Dee, which did not converge to form a larger valley glacier. On the southern edge of the plateau, Sissons (1979a) reconstructed a Younger Dryas Stade valley glacier in Glen Eidart sourced from Coire Mharconaich. However, Bennett (1996) discussed the geometry of moraines as being inconsistent with the glacier reconstruction by Sissons (1979a) and that ice in Glen Eidart may have drained from the upper reaches of the River Eidart and from the Mòine Mhòr plateau. Bennett (1996) clearly favoured the Younger Dryas Stade for the glacier, but Everest and Kubik (2006) reconstructed a western Cairngorm ice cap supplying ice down Glen Eidart prior to the Lateglacial Interstade. However, based on the moraine limits described by Sugden (1970) above, the readvance of a plateau glacier did not join up with the Glen Eidart glacier. This area of the western Cairngorm plateau requires more work regarding the timing and style of glaciation.

On the north-east side of the western plateau, the topography rises to the summit of Braeriach. Sissons (1979a) used moraine and the glacially transported boulders to reconstruct a corrie glacier in Coire an Lochain with an ELA of 1062m. The neighbouring corries were not thought to have been glaciated during the Younger Dryas Stade and, instead, evidence was found of periglacial activity, such as rock glacier deposits (Sissons, 1979a and Ballantyne, 1996). Ballantyne et al. (2009a) used surface exposure dating to date a possible rock glacier in Coire Beanaidh, yielding an uncertainty-weighted mean age of 11.3±0.6 ka (recalibrated as above to yield ages from 10.1±0.7 ka to 14.1±0.9 ka), suggesting formation during the Younger Dryas Stade.
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2.4.6 Central and eastern Cairngorms

Sissons (1979a) and Sugden (1970) agreed on the existence of three Younger Dryas Stade corrie glaciers in the central Cairngorms (Figure 2.5). However, Sissons (1979a) also reconstructed a valley glacier at the head of Glen Avon. Glen Avon is surrounded by extensive plateau surfaces (1050–1300m a.s.l.) consisting of the slopes of Ben Macdui and Cairn Gorm. Below the loch, Hinxman and Anderson (1915) described the Glen Avon glacier being joined by ice from the col of Glen Derry, and the moraines at the valley intersection mark the retreat of such a glacier. Just beyond Loch Avon, Midgley (2001) noted moraines damming the loch and subdued morainic ridges on the valley sides along the loch. There are also moraines at the head of Strath Nethy which mark the passage of ice down Strath Nethy; this continued until the Glen Avon glacier retreated below the level of the saddle between Glen Avon and Strath Nethy (Hinxman and Anderson, 1915). These lower moraines are assumed to mark the position of glaciers prior to the Lateglacial Interstade; however, at the head of the lake, closely spaced steep-sided moraines covered with boulders can be seen. Sissons (1979a) attributed such moraines to a Younger Dryas Stade glacier sourced at the head of the valley. Other sources for such a glacier may include the unnamed corrie that holds Loch Etchachan or a plateau-sourced glacier. A rock glacier on the eastern side of the unnamed corrie that holds Loch Etchachan (Figure 2.5), reinterpreted to be a rock slope failure and dated by Ballantyne et al. (2009a), yielded a weighted mean age of 12.1±0.6 ka (recalibrated as above using NWH 11.6, Lm scaling and 1 mm/ka erosion, the ages range from 12.9±0.9–12.3±1.1 ka) indicating formation during the Younger Dryas. Thus it cannot be ruled out that a glacier may have existed in Loch Etchachan at this time, particularly as a recent reinterpretation of the dated deposit by Jarman et al. (2013) indicated the landform may have glacial origins (this is discussed further in Chapter 4 Section 4.3.5.1 and Chapter 5 Section 5.6.1).

In the eastern Cairngorms, Sissons (1979a) reconstructed five Younger Dryas Stade glaciers based on the presence of lateral moraines and distinct boulder limits (Figure 2.5). The corrie glaciers on the slopes of Beinn a’ Bhuird have ELAs of 1032m and 1011m, and to the north, the glacier in the head of Slochd Mòr has an ELA of 886m. On the slopes of Ben Avon, Sissons reconstructed two small corrie glaciers with lower ELAs of 831m and 754m. However, Sissons (1979a) noted the
large plateau area to the south-west available for snow blow as being key to the glaciers’ existence. The large plateau area may also have been expected to accumulate ice and thus the area requires further work.

2.4.7 Synthesis

There is consensus that Younger Dryas glaciers were clearly much more restricted in size compared to the Western Highlands, and limited to corrie glaciers and possibly small valley glaciers in favourable topoclimatic locations. A great deal of uncertainty exists regarding which corries and valleys may have been glaciated during the Younger Dryas Stade and whether these were plateau-fed or sourced at the valley head. This uncertainty stems from the overlap of possible sites for Younger Dryas Stade glaciers to have existed with the sites of glaciation just prior to the Lateglacial Interstad. Clearly distinguishing between such events on the basis of geomorphology has been problematic. Some of the key issues, such as plateau glaciation and the variation in size and ELA of previously reconstructed glaciers, are discussed below.

2.5 Discussion/Outstanding Issues

2.5.1 Extent of Younger Dryas ice masses

After the withdrawal of external ice from the British–Irish Ice Sheet, local glaciers formed in the valleys and corries of the Cairngorms. However, there are competing hypotheses on whether all the glaciers were contemporaneous, or whether some pre-dated the Lateglacial Interstad and others existed during the Younger Dryas. While Younger Dryas glaciation of the highest corries is agreed on (Sugden, 1970; Sissons, 1979a; Purves et al., 1999), perhaps the most pertinent question relates to whether valley glaciers co-existed at this time. Brazier et al. (1996b) also mentioned the presence of multiple terminal moraines in several corries and suggested there was no reason for them all to be attributable to the same event; some support for this idea is discussed by Kirkbride (2005) for nearby Lochnager, South-East Grampians. The question does not only concern geomorphology; the Glen Geusachan glacier (0.777 km$^3$) alone contained 46% of the ice mass accumulated in the Cairngorms during the Younger Dryas Stade; Garbh Choire (0.423 km$^3$) and Glen Eidart (0.105 km$^3$), also in the western Cairngorms, were the next largest and
had ELAs of 730m, 897m and 798m respectively (Sissons 1979a). However, the three corrie glaciers reconstructed on the northern flanks of the Cairngorms had ELAs 300m higher than the Glen Geusachan glacier. Given the minimum Younger Dryas chironomid-inferred mean July temperature was 6.8°C at Abernethy Forest (230m asl) (Brooks \textit{et al.}, 2012), the original ELAs can be used to derive the precipitation values that include input from avalanches and wind-blown snow (Ohmura \textit{et al.}, 1992): the 730m Glen Geusachan ELA requires 1720 mm a\(^{-1}\) and the 1000m corrie ELAs 1125 mm a\(^{-1}\) (1141 mm a\(^{-1}\) compared to just 641 mm a\(^{-1}\) respectively when converted to sea-level values (cf. Ballantyne, 2002; Benn and Ballantyne, 2005)). If such a configuration of glaciers existed, this requires explanation through the understanding of local topoclimatic factors and precipitation gradients. There is also growing appreciation of the presence and role of plateau-fed glacier systems within the British Isles during the Younger Dryas and the importance of such ice in increasing ELAs (McDougall, 2001; Benn and Ballantyne, 2005; Finlayson, 2006; Brown \textit{et al.}, 2011; Boston \textit{et al.}, 2013).

There are several key sites within the Cairngorms where geomorphological evidence has been inconclusive or selective evidence has been presented to support reconstructions, and a full reconstruction of events using all the available evidence to test competing theories has been absent. Sissons (1979a) reconstructed the large valley glaciers in Glen Geusachan and Garbh Choire. Sugden’s (1980) reply questioned the use of hummocky moraine to delimit the Younger Dryas Stade glaciers and favoured more limited corrie glaciation; however, there is no evidence known to the author to support the separate corrie glaciers within Garbh Choire depicted by Sugden within Purves \textit{et al.} (1999).

Also within the western Cairngorms, Glen Geusachan remains a valley of much uncertainty, particularly when evidence from multiple authors is combined. The earliest reconstruction is that of an independent western plateau ice cap prior to the Lateglacial Interstade which fed ice into Glen Geusachan and Gleann Einich (Everest and Kubik, 2006). Geomorphological evidence would dictate that later a glacier pushed west across the Mòine Mhòr plateau leaving the granite moraines on schist bedrock, as noted by Sugden (1970). Bennett and Glasser (1991) suggest these ridges formed prior to the ice-dammed lake adjacent to the head of Glen Geusachan, as the lake drained west through the area occupied by the plateau.
glacier. This indicates upper Glen Geusachan was one of the last places to become ice free. It also suggests a change in glacier-source area, either during ice-sheet deglaciation or between deglaciation and renewed Younger Dryas glaciation. The timing of both the plateau glacier and the Geusachan glacier damming the lake in upper Glen Geusachan remain uncertain. Questions still remain over the existence of a Younger Dryas glacier in Glen Geusachan; it seems unlikely to have surpassed the older outer limits based on exposure dating by Everest and Kubik (2006), but did it approach these limits forming the meltwater channels and damming a lake in upper Glen Dee, or was it more limited, forming the hummocky moraine south of Devil’s Point? Did a glacier only exist in the upper reaches of Glen Geusachan where a higher set of moraines mark glacier retreat, or did Glen Geusachan fail to nourish a Younger Dryas Stade glacier? Whichever option is favoured, it has implications for the Garbh Choire glacier further up the Dee, as the lake-terminating outer moraine links it with a large glacier within Glen Geusachan.

Other important sites include Glen Eidart and Glen Avon where Sissons reconstructed Younger Dryas valley glaciers, and upper Glen Derry which was noted by Sugden (1970; 1980) and Midgley (2001) for its hummocky moraine, but no Younger Dryas glacier has been reconstructed. These all have large high plateau areas on their southern and/or western sides, linking them with the need to better understand whether plateaus were important.

Given the uncertainty surrounding the geomorphological evidence, the discussion over the extent of the Younger Dryas glaciers turned to dating and modelling. Most surface exposure dates within the Cairngorms have focused on lower glacier breaches and moraines associated with ice-sheet deglaciation. These only provide down-valley constraints on the location of Younger Dryas glaciers, especially when separate glacier margins are present up-valley. Given the Younger Dryas ages from the discrete debris accumulation (DDA) in Coire Beanaidh and below Carn Etchachan, they provide no unequivocal constraint on Younger Dryas glacier extent, particularly given the uncertainty over the deposits’ origin and whether glacial activity was involved, particularly at the Carn Etchachan site (Jarman et al., 2013).
The use of modelling and topoclimatic factors such as radiation, snow blow and avalanche areas have the potential to resolve the outstanding issues. Sissons (1979a) discussed insolation and snow blow as factors responsible for the variation in ELA; however, the only climatic factor capable of explaining the large differences was accumulation. Thus Sissons (1979a) inferred snow-bearing winds came from southerly directions with snow blow and insolation explaining smaller variations in ELA between adjacent glaciers. A similar pattern of rising ELAs to the north was reconstructed in the nearby South-East Grampians (Sissons and Sutherland, 1976) and this gradient is largely a continuation. More recent work using a GIS-based snow-blow modelling approach concluded that the valley glaciers reconstructed by Sissons (1979a) were unlikely to have formed under the same climatic mechanisms as the glaciers in the higher corries (Purves et al., 1999). The locations of valley glaciers suggested by Sissons (1979a) have some of the lowest snow depths predicted by the model. While useful, the model does have limitations as explained by Purves et al. (1999): it does not account for precipitation gradients both laterally and altitudinally; nor for the acceleration of wind over summits or channelling through valleys; and the defining of accumulation areas is problematic, particularly for larger catchments.

Based on the glacier reconstructions having high variations in ELA and the large plateau surfaces that can act as snow-blow sources, it may be useful to adopt the idea of a regional temperature-precipitation ELA and a more variable temperature-precipitation-wind ELA (Dahl and Nesje, 1992; Lie et al., 2003a, 2003b). Working with such a concept helps to explain high ELAs due to snow deflation and low ELAs owing to leeward accumulation of snow. The Cairngorms' plateaus are known to experience high wind speeds today, and the contrast between rolling plateau and deep well-developed corries and glacial troughs lend themselves to snow redistribution and accumulation. Given our approximate knowledge of likely precipitation gradients to test and basic snow-blow modelling techniques, it should be possible to test whether the configuration of glaciers is climatologically plausible. However, this relies on the correct glacier limits and source areas being identified first.

Snow blow might be less important if evidence of plateau-sourced glaciation is identified, as this would increase the glacier ELAs. Many recent British
palaeoglaciological studies have reinterpreted previous mapping, and there has been a growth in recognition of plateau surfaces as accumulation areas for previously reconstructed Younger Dryas Stade glaciers (McDougall, 2001; Benn and Ballantyne, 2005; Finlayson, 2006; Brown et al., 2011; Hughes, 2012). These have been identified by tracing moraines back onto the plateau surfaces and detailed analysis of the position and geometry of meltwater channels and moraines (McDougall, 2001). Recent modelling of the Younger Dryas Stade extent in Scotland predicted larger, more extensive glaciation in the Cairngorms than currently recognised (Golledge et al., 2008). Although later, Golledge et al. (2009) and Golledge (2010) expanded the description, stating that modelling experiments only predict thin, low velocity ice cover over much of the Cairngorm plateau with no dynamic glaciers and limited infilling of valleys. This continental style cold-based ice, which may have accumulated on plateau surfaces during the Younger Dryas Stade, may have left only very subtle evidence as seen by work on present-day analogues, such as northern Norway (Rea et al., 1998).

Based on Manley’s equation (adapted by Rea et al., 1998), plateaus such as those between Ben Macdui and Cairn Gorm, with altitudes 1100–1309m asl and summit breadths of approximately 2 km, may be glaciated if they occur c.100m above the regional firn line. However, reconstructing a regional ELA in an area of great uncertainty is problematic, thus the ELAs within the Cairngorms are ignored for the purpose of this discussion. ELA reconstructions of 738–816m from eastern Monadhliath Mountains (Boston et al., 2013) and 815m from northern Gaick (Sissons, 1980a), although derived by different ELA calculation methods, both represent plateau-fed glacier systems that give an approximate indication of a regional ELA (Nesje and Dahl, 2000). Thus, depending on the steepness of the rising regional ELA towards the Cairngorms, it might be expected that the highest plateaus could support ice. The western Cairngorm plateau of Mòine Mhòr is lower but wider, thus Manley’s theory (adapted by Rea et al., 1998) suggests glaciation would have occurred when the firn line reached the plateau surface which is at c.900m, making Younger Dryas glaciation of the plateau worthy of further investigation. The inclusion of plateau areas within the calculation of the valley glaciers’ ELA would increase the ELAs affected and potentially decrease the variation amongst currently reconstructed ELAs within the Cairngorms. This
highlights the need for more work to better understand the style of glaciers within the Cairngorms and the precipitation gradient within the region.

A combination of detailed mapping, targeted dating, exploration of plateau-sourced glaciation and modelling of all the topoclimatic factors will provide answers to the extent of Younger Dryas glaciation in the Cairngorms and the factors that controlled glacier size and ELAs in an area of marginal glaciation. Despite the importance of modelling, it should be highlighted that detailed mapping and dating should drive the topoclimatic modelling, otherwise modelling studies can suffer from incorrect glacier configurations. The strength of consistency and agreement between the two approaches will be the ultimate test.

### 2.5.2 Southern margin/ice-sheet deglaciation

The southern margin of the Cairngorms has not received as much recent research as the northern margin, but potentially holds a wealth of information about local glacier, ice sheet and climate interaction prior to the Lateglacial Interstade. The suggestion of ice-dammed lakes occupying Glen Quoich, Glen Derry and Glen Dee indicates the size and relatively longer survival of the Deeside ice mass during deglaciation (Kirkbride and Gordon, 2010). The height of such lakes would suggest they occurred relatively early in deglaciation when ice was thick enough to block cols that otherwise would have acted as overflow points, restricting the heights of the lakes. It has been suggested the Glen Derry and Glen Quoich lakes occurred simultaneously. Questions remain over the extent of the lakes and the location of both Deeside and local glacier ice margins when they were formed. One particular difficulty is how the lakes and the position of the Deeside ice mass which dammed them relate to the formation of large moraines, thought to be locally sourced at Derry Lodge and Derry Dam. Given current geomorphological interpretations, seemingly two hypotheses exist (Figure 2.6): first, ice blocked lower Glen Lui and the Meirleach Col, and later the Glen Derry glacier readvanced forming the Derry Lodge and Derry Dam moraines (Figure 2.6, A). The second formation hypothesis is that the Derry glacier formed the Derry Lodge moraines, and subsequently Glen Lui and Meirleach Col were blocked by ice forming the ice-dammed lakes (Figure 2.6, B). However, the Deeside ice margins must have been approaching the Derry Lodge moraines in order to block off sufficient cols to form the ice-dammed lake. No
lake deposits have been reported in the literature, thus one or more of the lakes may have been a surface lake forming on a depression between the local and external ice sources. It can be assumed that the Glen Dee ice-dammed lake formed later, as the damming ice that blocked Glen Derry and Glen Quoich is likely to have been sourced from the west. This would suggest that the southern Cairngorms went through a similar process of east-to-west deglaciation to the northern Cairngorms. However, on the southern flank of the western plateau, meltwater channels are mapped dipping to the west (Kirkbride and Gordon, 2010). This would suggest the ice was being fed from the south of the Cairngorms and flowing west into Glen Feshie, at least during the later stages of deglaciation.

Many moraines and ice margins exist in Glen Derry, Glen Dee, Glen Quoich and Glen Luibeg, outside of what are thought to be plausible Younger Dryas Stade limits. Of particular interest are the Derry Lodge and Derry Dam moraines believed to be formed by the Derry glacier. The timing of these moraine formations could help improve our understanding of ice-sheet demise and the response of local Cairngorm glaciers to Dimlington Stade climate change. The evidence suggests active moraine formation during ice-sheet glaciation with the possibility of readvance/stillstand events occurring, similar to those described in the northern Cairngorms.
2.5.3 Ice survival

Despite the survival of ice into or throughout the Lateglacial Interstade being a long-term proposition, it has remained difficult to find evidence either in favour of or against such a concept. The amalgamation of mapping and dating in the BRITICE project has shown the similarity between the retreating ice mass prior to the Lateglacial Interstade and the accepted reconstructed ice limits occurring during the Younger Dryas Stade (Clark et al., 2012). This does not imply survival but poses an interesting hypothesis. Unfortunately, most upland areas that would have experienced the last stages of ice-sheet retreat have also been the first to have landform records erased by the readvance or regrowth of glaciers during the Younger Dryas readvance. If the Cairngorms did, as some authors suggest, only experience corrie glaciation during the Younger Dryas Stade (Sugden, 1970; Purves et al., 1999; Everest and Kubik, 2006), then the benefits of carrying out surface exposure dating of moraines in valley heads may be twofold; first confirming the absence or presence of Younger Dryas Stade valley glaciers and second, if ages predate the Younger Dryas Stade, providing an answer regarding
the survival or reformation of glaciers during the Lateglacial Interstade. Given the proximity of the Cairngorms to glaciation today (Sugden and Clapperton, 1975; Harrison et al., 2014), it would require substantial warming to remove ice from the Cairngorms (Sugden, 1974). However, precipitation may have been reduced when compared to levels today, making deglaciation more likely. Subsequent climate cooling during the Lateglacial Interstade may have led to ice accumulation during colder periods of the Lateglacial Interstade, such as the Older Dryas. It may appear illogical to suggest ice accumulation during the Lateglacial Interstade, but with chironomid-inferred mean July temperatures at Abernethy Forest (230m asl) falling to 8°C during the Older Dryas, not dissimilar to the temperatures during the Younger Dryas Stade, it seems realistic to expect ice accumulation. With the known importance of precipitation in restricting glaciation in the Cairngorms, it seems likely this would have been an important factor in determining whether any readvance or reformation of ice during the Lateglacial Interstade occurred.

2.5.4 Palaeoclimatic inferences and wider glaciation

![Figure 2.7 The Younger Dryas ice limits and generalised ELA contours (in metres). Note the large West Highland ice cap and the smaller glaciers of the Cairngorms and South-East Grampians. Adapted from Sissons (1979c) with permission of Nature Publishing Group and Benn (1997) with permission of Elsevier. (Note: many ice limits and ELAs may have been reinterpreted since but the overall pattern remains largely correct)](image-url)
Past glacial environments are often studied to increase our understanding of palaeoclimate and past climate-glacier interactions. It is important to consider that glaciers respond to both temperature and precipitation variations, and that glacier response also varies according to other climatic factors such as continentality (see Nesje and Dahl, 2003; Chinn et al., 2005). It appears that during ice-sheet deglaciation, climate factors – most likely a lack of precipitation – caused local Cairngorm glaciers to reduce in size and split from externally sourced ice masses. However, moraines, sedimentological evidence and ice-dammed lakes show that the local glaciers, both on the northern and southern margin of the Cairngorms, remained active and readvanced or stood still multiple times prior to the Lateglacial Interstade. The timing of these events remains unclear, but taken together the evidence would suggest 15–17 ka BP. Wider questions also exist concerning how these readvances and stillstands relate to readvances proposed elsewhere in the British Isles, such as the Wester Ross Readvance. Recently, the recalibration of exposure ages on the Wester Ross Readvance moraines suggested a readvance occurred approximately 14.7 ka BP (Ballantyne and Stone, 2012). Ballantyne and Stone (2012) expanded an idea initially put forward by Sissons (1981) that surviving ice masses may have readvanced as the atmospheric polar front moved northwards across the Highlands. The enhanced cyclonic activity, when warm water replaced cold water, would have increased snowfall prior to the rapid warming at the onset of the Lateglacial Interstade (Ballantyne and Stone, 2012). This would have also had implications for seasonality that may have been important to glacier retreat and readvance. It is possible these changes in precipitation and seasonality also affected glaciers in the Cairngorms. Also, there may be interest in dating studies along past ice streams (Ballantyne, 2010) to enable the linking of knowledge in the Cairngorms with earlier stages of deglaciation in the coastal lowlands of north-east Scotland (Phillips et al., 2008). This may involve building on the work of Brown (1993, 1994) who reconstructed ice-dammed lakes and retreating ice-marginal positions along Deeside from Aberdeen towards Braemar.

The substantial difference in Younger Dryas glacier size between the Cairngorms and Western Highlands is well documented (Sissons, 1980a; Golledge, 2010), and strong precipitation gradients are thought to be responsible (Sissons, 1979b, 1979c; Benn and Ballantyne, 2005). Deriving more accurate precipitation values and
gradients between western Scotland and the Cairngorms is currently hindered by the uncertainty over glacier extent and the differences in glacier style between the regions.

Often studies have focused on limited geographical areas without making important comparisons with work in nearby mountain ranges. Original ELAs, reconstructed by Sissons and Sutherland (1976) for the wider area, indicate ELAs of approximately 500m in the south-east corner of the South-East Grampians, rising to 760m for valley glaciers and 880m for corrie glaciers in the north-west corner (770m based on newly reconstructed plateau icefield (Hughes, 2012)), nearer the Cairngorms (Figure 2.7). The glaciers of the Gaick Plateau, reconstructed by Sissons (1980a), possess ELAs in the south-west corner of 678m, rising to 815m in the north-east, nearer the Cairngorms. Recent work in the Monadhliath Mountains suggests plateau ELAs rising from 560–816m from west to east (Boston et al., 2013). If hypothetically the Younger Dryas Stade glaciers are limited to the high corries of the Cairngorms, this reduces the spatial variation of ELAs within the Cairngorms. However, it increases the precipitation gradient required between the Cairngorms and the larger valley and corrie glaciers of the South-East Grampians (Sissons and Sutherland, 1976) and the plateau-fed reconstructions of the Gaick (Sisson, 1974) and Monadhliath Mountains (Boston et al., 2013). For this reason, a full reinvestigation of the extent and style of glaciers within the Cairngorms is needed.

However, the surrounding areas are not without their own issues. Merritt et al. (2004) suggested a Cairngorm-type model for the Gaick Plateau with ice caps existing prior to the Lateglacial Interstade and a more restricted Younger Dryas readvance glaciation. Benn and Ballantyne (2005) refuted the proposal, reconstructing and providing evidence for a Younger Dryas readvance icefield in the West Drumochter Hills, but the Gaick region requires dating to clarify Younger Dryas readvance limits (Merritt et al., 2004). Dating will also be required to clarify whether large valley/plateau-fed outlet glaciers existed in the South-East Grampians during the Younger Dryas Stade. It is clear that making concrete Younger Dryas Stade palaeoclimatic inferences in the region is limited by uncertainties. As progress is made in adjacent regions, the regional consequences for palaeoclimate, particularly regarding precipitation gradients, will require revision.
2.6 Conclusion and Chapter Summary

The polygenetic and palimpsest landscape of landforms from both locally and externally sourced glaciers, from both prior to the Late-glacial Interstade and during the Younger Dryas Stade, has provided researchers with an ongoing challenge. Almost 150 years have passed since it was recognised that glaciers had formed many of the landforms found within the Cairngorms. Originally landforms and erratic rocks were used to reconstruct past glaciers and the direction of ice flow; later the extent of Younger Dryas readvance glaciers and their palaeoclimatic inferences became the focus. More recently, the advance in dating techniques has allowed individual landforms to be attributed to climatic events; thus providing date-supported insights into glacier-climate response.

Reviewing the existing work has tested the compatibility of studies both within the Cairngorms, between neighbouring geographical areas and between connected disciplines. Through the synthesis of the literature, it would appear that, early in deglaciation, external ice from the west divided around the central Cairngorms ice cap with erratics being left high on the flanks of the Cairngorms. Later, locally sourced Cairngorm glaciers continued to contribute to the main glaciers of Strathspey and Deeside until they broke up, forming ice-dammed lakes approximately 16–17 ka BP. Moraines mark several stillstand positions of both local and external ice masses at this time, suggesting possible climate-driven events. Glaciers are thought then to have retreated at the onset of the Late-glacial Interstade with no current evidence of ice survival or reformation during the short cold oscillations of the Late-glacial Interstade. The next glaciers formed during the Younger Dryas Stade; while some corries were definitely occupied, the presence or absence of valley or plateau glaciers remains uncertain. The only glaciers shown by cosmogenic surface exposure ages (Kirkbride et al., 2014) and modelling (Harrison et al., 2014) to have existed during the Holocene have been restricted to the confines of the corrie backwalls during the Little Ice Age.

While substantial progress has been made in the Cairngorms, the work has been highly focused. Some areas of the southern and eastern Cairngorms have received little attention or the original accounts from over 100 years ago remain our only insight into deglaciation. While these remain valuable, further research is needed.
and important research questions remain: how did local and Deeside ice masses interact during deglaciation; were ice-dammed lakes formed and did Deeside ice invade the valleys of the southern Cairngorms; what was the timing of the local ice readvances in Glen Derry prior to the Lateglacial Interstade and how do these compare with events in the northern Cairngorms; did ice survive into or reform during the Lateglacial Interstade; were glaciers valley- or plateau-sourced or did this vary spatially and temporally; what was the extent of the Younger Dryas readvance glaciation in the Cairngorms; how did ELAs vary during deglaciation and the Younger Dryas readvance, and what were the topoclimatic and palaeoclimatic factors responsible for this? Solving these issues will be important in combining the growing number of palaeoglacial reconstructions in Britain to provide key evaluation areas for numerical models of ice-sheet growth and demise.

Given the limited Younger Dryas readvance, the Cairngorms provide a unique record of ice-sheet deglaciation. The Younger Dryas Stade glaciers provide an insight into an area of marginal glaciation where understanding topoclimatic factors, such as the plateau surfaces and snow drift, are essential to unlocking the glacial history. It is expected a combination of new mapping, targeted surface exposure dating and modelling will all contribute to our understanding of glaciation in the Cairngorm Mountains.
3 Methodology

3.1 Introduction

Numerous methods have been used in this study to address the research objectives set out in Chapter 1. These have included remote and fieldwork mapping combined in a geographical information system (GIS) to produce a map of glacial landforms within the Cairngorms. This has been temporally constrained in key locations by new cosmogenic surface exposure ages. Glacier surfaces have been reconstructed based on the geomorphology and flow-line modelling, enabling the glacier ELAs to be calculated. Finally, topoclimatic factors such as snow redistribution, avalanche inputs and solar radiation have been modelled to assist with explaining variations in ELA and drawing palaeoclimatic inferences.

3.2 Geomorphological Mapping

3.2.1 Introduction and approach

The Cairngorm Mountains are located within north-east Scotland; the mapping presented within this project covers an area of c.600 km² that is recognised for containing a wealth of information about palaeoclimatic changes (Kirkbride and Gordon, 2010). These landforms represent a complex combination of locally and externally sourced glaciers during the retreat of the last British-Irish Ice Sheet at the end of the Dimlington Stade, and the growth of smaller local glaciers during the Younger Dryas. The area has attracted considerable palaeoglaciological work; however, many interpretations have been conflicting (Sugden, 1970; Sissons, 1979a) and many studies have been piecemeal in approach, often focusing on specific sites (e.g. Bennett and Glasser, 1991; Midgley, 2001; Everest and Kubik, 2006). While much of the landform interpretation remains valuable, our understanding of glaciology and Younger Dryas glaciation has developed considerably since much of this work was undertaken. Recently, the Scottish Natural Heritage (SNH) carried out extensive mapping of the Cairngorms (Kirkbride and Gordon, 2010); however, because of the nature of the project it lacked detailed interpretation of the landforms and discussion of their significance for glacier reconstruction. Much of the existing work was consulted before the new mapping of landforms within this study was undertaken.
The mapping presented within this project consists of a new seamless map of the geomorphology within the Cairngorms (Figure 3.1). This approach had numerous benefits; it minimised the possibility of landforms being overlooked, allowed existing detailed sedimentological, geochronological ages and mapping to be contextualised within the wider geomorphology, and provided a basis for site selection for new targeted surface exposure ages. An inherent benefit of this comprehensive approach was the inclusion of both landforms associated with the Younger Dryas and Dimlington Stade, facilitating a full reanalysis of evidence without bias from site selection or existing work. The aim of the mapping is to provide a detailed relative retreat pattern of glaciers within the Cairngorm Mountains, to understand the source locations of these glaciers and whether the overall retreat was punctuated by one or more stages of stillstand or readvance. This will include the evaluation of the pattern and distribution of landforms within the Cairngorm Mountains to establish whether there are landsystems that represent synchronous stages of readvance.

![Figure 3.1 Area covered by new mapping](© Crown Copyright/database right 2014. An Ordnance Survey/EDINA supplied service)
3.2.2 Methods

3.2.2.1 Imagery analysis

The mapping of landforms has been carried out in Esri GIS software using ArcMap 10 and ArcScene 10. Digital orthorectified and georeferenced 25cm resolution aerial images were obtained from the Ordnance Survey. The NEXTMap digital surface model (DSM), digital terrain model (DTM) and orthorectified radar image (ORI) provided 5m resolution elevation data with a vertical accuracy of 0.5–1m for the area (Smith et al., 2006). It is collected using synthetic aperture radar (SAR) which is less detailed than LiDAR, but it can be flown at higher altitudes and thus reduces costs (Smith et al., 2006). Mapping was undertaken onscreen using ArcMap 10; although, ArcScene 10 was used on an adjacent monitor to view the aerial imagery draped over the DSM, which thus facilitated the viewing of landforms either in perspective or stereo.

The imagery was merged and organised as raster datasets within a file geodatabase, ensuring the imagery retained its quality and display speeds. The aerial imagery was predominantly viewed using its natural colour. However, stretching the values by ‘standard deviations’ for the ‘current display extent’ was particularly useful for enhancing areas of shadow. This stretched the pixel values for the displayed extent, such that landforms in shadow beneath steep cliffs could be mapped, this technique also assisted in identifying subtle changes in the DSM. Larger scale landforms were often digitised using a hill-shaded DSM, which was particularly useful for linear landforms (Figure 3.2). This was carried out by displaying two hill-shaded layers at 90° intervals simultaneously, with the sun’s angle at 45° from the horizon and ensuring mapping was replicated from multiple shading azimuths to reduce bias (Smith and Clark, 2005). Other ArcGIS tools, such as slope, were used to identify changes in gradient; these were useful in identifying linear landforms such as lake shorelines. In addition, the contour and profile tools within 3D Analyst were used to assist in interpreting whether landforms, such as lake shorelines, were continuous or to determine the profile of landforms.
3.2.2.2 Fieldwork

The ground truthing of interpretations from aerial photos was an important process, particularly for distinguishing between landforms that can appear similar on aerial images, e.g. lateral meltwater channels, terraces and lake shorelines. It also permitted the distinction between depositional and erosional landforms, and the identification of smaller landforms that can only be identified within the field. A Trimble Juno handheld GPS tablet with ArcPad mapping software was used to enable the mapping to be viewed in the field. This also facilitated the addition of point and line features electronically in the field. In addition, printed versions of the remote mapping were taken into the field to permit annotations and corrections. Sketches, photos and notes were also taken and, where applicable, sediment analysis was carried out to assist with landform interpretation and to distinguish between formation hypotheses.

3.2.2.3 Ground-based photogrammetry

At sites where a higher resolution digital elevation model (DEM) was deemed useful to fulfilling the mapping objectives, such as Glen Eidart where the orientation of the moraines is key to identifying the glacier-source area (Bennett, 1996), photos were collected to undertake ground-based photogrammetry. Photographs were collected
using a standard point and shoot Cannon IXUS 105 camera. Photos were taken from multiple vantage points in the landscape such that all landforms of interest appeared in multiple sets of photographs from different angles (see Westoby et al., 2012). The locations of the photograph sites were recorded using a GPS, in case these were required for camera positioning at the DEM generation stage.

The Agisoft Photoscan software package was first used to spatially align the camera positions and to generate a point cloud; once this was corrected manually the software was used to build a DEM (Figure 3.3). Although the mesh resolution was generally good enough for the project’s purpose, the impact of the limited topographic vantage points was noticeable in the resulting mesh. Areas of low point cloud density/resolution were found to be where the angles of photographs were not sufficient to capture all the sides of the landform. The models were later georeferenced to match the aerial imagery and NEXTMap DEM, with control point coordinates generated using GIS. This method was sufficient for the mapping purposes here, and by referencing it to the existing aerial images it ensured the seamless mapping of landforms. The DEM was then used in ArcGIS to assist in mapping the head of Glen Eidart.

![Figure 3.3 Camera positions around the head of Glen Eidart for DEM generation using Agisoft Photoscan. Flags mark the locations used for georeferencing](image-url)
3.2.2.4 **Software and map design**

The digitisation of landforms and map design was carried out in Esri ArcMap 10, aided by 3D visualisation of imagery in ArcScene 10. Mapping was transferred to a Trimble Juno Handheld GPS with ArcPad 8.0 to enable viewing in the field. The map consists of vector layers for each landform/superficial deposit, with each being attributed the appropriate feature type, e.g. polygon, line or point. The feature datasets were all mapped on the British National Grid coordinate system using OSGB36 datum, and stored within a file geodatabase. Past mapping styles and symbology were analysed from existing maps (e.g. Sahlin and Glasser, 2008; Brown *et al.*, 2011), and a suitable structure and approach devised.

3.2.2.5 **Geomorphological mapping**

The pattern of glacial retreat can most easily be established through the morphology of moraines and orientation of meltwater channels, particularly when these features are formed at the ice margin and record the former ice retreat positions. However, to assign landforms in separate valleys to a contemporaneous event, multiple lines of evidence and a repeated spatial relationship between an assemblage of individual landforms is required (Lukas, 2006). This approach also allows an independently dated margin in one location to be extrapolated to undated sites of the same distinguished geomorphology (Lowe and Walker, 1997; Benn and Ballantyne, 2005).

Lukas (2006) reviewed landforms in upland Britain that had been used in determining the sites and extent of Younger Dryas glaciers, and used the concept of morphostratigraphy in the North-West Highlands to show that there were key differences between multiple landforms inside and outside dated Younger Dryas limits. The key features suggested as being used to identify Younger Dryas limits are: the presence of ‘hummocky moraine’, and clear end moraines or boulder moraines inside Younger Dryas limits, which differ from subdued and isolated moraine ridges outside Younger Dryas limits (Lukas, 2006). It should be possible to test this principle with the use of new cosmogenic surface exposure ages presented in Chapter 5. Differences also occur within glaciofluvial landforms: outwash fans, small terraces and beaded eskers are often present inside Younger Dryas limits, whereas large kames, eskers and kettled outwash are more typical outside...
Younger Dryas limits (Lukas, 2006). Additional features include gullied slopes, immature talus slopes, lateral termination of drift limits, glacially transported boulders, meltwater channels and lateral moraines inside Younger Dryas limits compared to mature talus, protalus ramparts, large debris fans, blockfields and large solifluction lobes outside Younger Dryas limits (Lukas, 2006). These features, such as mature well-developed talus slopes and large debris fans, would have been highly active during the Lateglacial Interstadial and cold Younger Dryas Stade, in addition to any further development during the Holocene; whereas inside Younger Dryas glacier limits, the talus/debris fans would have been removed by ice flow and only subsequent immature features have developed during the Holocene. The landform evidence described above is summarised from over 30 years' work in the British uplands and provided a good basis for distinguishing areas of potential Younger Dryas glaciation; however, local patterns in landform assemblages have also become apparent and these are discussed in Chapter 6.

Given the plateau landscape of the Cairngorm Mountains, more specific plateau landsystems were also considered. In Ellesmere, Canada the retreat of plateau-ice margins is recorded by incision into bedrock or drift by lateral meltwater channels (Rea et al., 1998). Moraines can also be generated on plateaus where erosion has taken place underneath warm-based ice and, where the topography is suitable, periglacial trimlines can assist in discerning if plateau ice existed (Rea et al., 1998). The Cairngorms' position on the eastern side of the British Isles, in more continental conditions than the Western Highlands, would favour the presence of cold-based ice; longer-term evidence for this is given by the survival of tors and chemically weathered surfaces (Hall and Glasser, 2003). Cold-based plateau ice has often been shown to leave only a faint signature on the landscape, with present-day analogues creating only subtle lateral meltwater channels or moraines (Rea et al., 1998). In Lyngen, northern Norway, the presence of cold-based ice and the absence of supraglacial debris sources meant there was no potential for moraine formation (Rea et al., 1998) and the plateau ice is retreating to reveal undisturbed blockfields and patterned ground (Gellatly et al., 1988). To add to the complexity of potential cold-based ice leaving little geomorphological signature, the Cairngorms’ characteristics as a landscape of selective linear erosion restrict the preservation of evidence on the steep valley sides and back walls.
3.2.3 Geomorphological landforms

It was decided, based on the strength of the morphestratigraphic approach and the need to understand the spatial pattern of landforms, including locations where glacial evidence has been lost to postglacial activity, that all landforms and surfaces within the Cairngorms would be mapped. However, because of the nature of the project, there is a bias in detail and interpretation towards glacial and periglacial landforms. The landforms and surfaces have been divided into the following groups: glacial deposition, glacial erosion, glaciofluvial, periglacial and miscellaneous. The landforms within each group, their mapping criteria and their use in glacier reconstruction can be seen in Table 3.1.
<table>
<thead>
<tr>
<th>Landform</th>
<th>GIS Feature</th>
<th>Mapping Criteria and Physical Description</th>
<th>Use in Glacier Reconstruction</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glacial Erosion</td>
<td>Polygon</td>
<td>Steeply cut, glacially eroded rock. Common at the corrie back wall or truncated spurs associated with erosion from corrie and valley glaciers.</td>
<td>Indicates ice erosion and presence of warm-based ice.</td>
</tr>
<tr>
<td>Ice-Moulded and Scoured Bedrock</td>
<td>Polygon</td>
<td>Bedrock surfaces shaped by glacier activity. Often smooth bedrock on the upper surface and plucked surfaces on the leeward side. May show signs of ice scouring or striations.</td>
<td>Indicates warm-based ice and the direction of ice flow. This flow direction does not necessarily represent the most recent phase of glaciation if the ice was cold based or not sufficiently erosive.</td>
</tr>
<tr>
<td>Roche Moutonée</td>
<td>Polygon</td>
<td>Distinctive shape with smooth bedrock surface on up-ice side and a steeper plucked surface on the leeward side.</td>
<td>Indicates presence of warm-based ice and the direction of ice flow, but not necessarily the most recent phase of glaciation.</td>
</tr>
<tr>
<td>Glacial Deposition</td>
<td>Polygon</td>
<td>Ridges or mounds of glacigenic material. Often linear or arcuate ridges on valley floors and sides. Composed of angular, sub-angular unsorted material. This includes hummocky moraine which can appear chaotic and irregular; however, often shows order when viewed from aerial photography.</td>
<td>Depending on type and formation processes, it can be used to infer the pattern, style and speed of glacier retreat. More subdued moraines may exist outside the Younger Dryas limits.</td>
</tr>
<tr>
<td>Crest</td>
<td>Line</td>
<td>Orientation of the crest line of the moraine. Note this is not always the same as the larger moraine footprint.</td>
<td>Useful for identifying the pattern of glacier retreat.</td>
</tr>
<tr>
<td>Fluted Moraines</td>
<td>Line</td>
<td>Linear ridges up to 0.5m in height which can run over moraine/drift surfaces. Typically 10–100m in length and 1m wide. Often begin at a large embedded boulder.</td>
<td>Indicate direction of ice flow and can be used to identify the extent and direction of glacier readvances. Indicate subglacial modification of sediment.</td>
</tr>
<tr>
<td>Glacial Drift</td>
<td>Polygon</td>
<td>Thick glacigenic deposit which does not have the topographic expression of moraine ridges. Can be identified by a change in vegetation or a change in the density of glacially transported boulders. Can occur in association with a trimline which separates areas of solifluxion from areas of glacial drift, or areas of mountain top detritus from glacial drift.</td>
<td>Used to indicate horizontal and vertical extent of glaciers.</td>
</tr>
<tr>
<td>Ice-Marginal Deposits</td>
<td>Polygon</td>
<td>Drift that has been heavily shaped by meltwater at the glacier margin. Includes moundsy deposits with no clear orientation and which are often associated with the presence of meltwater channels.</td>
<td>Indicate the former position of the glacier margin.</td>
</tr>
<tr>
<td>Glaciofluvial Landforms</td>
<td>Polygon</td>
<td>Drift that has been heavily shaped by meltwater at the glacier margin. Includes moundsy deposits with no clear orientation and which are often associated with the presence of meltwater channels.</td>
<td>Indicate the former position of the glacier margin.</td>
</tr>
</tbody>
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</thead>
<tbody>
<tr>
<td>Glacial Meltwater Channel (Rock)</td>
<td>Line</td>
<td>Often steeply sided channels cut into bedrock. The large size of the channel often does not fit with the small stream it now contains. The channel may run obliquely across the slope, whereas postglacial drainage channels will follow gravity. Meltwater channels often have abrupt inceptions or terminations.</td>
<td>Indicates the position of a glacier or ice-dammed lake drainage route. Where channel cross sections on valley sides are only benches not full channels, it suggests the glacier supported the inner side of the channel. This indicates formation as a lateral meltwater channel which can be used to infer the vertical position of the glacier and the ice gradient. If the channels have up and down long profiles this suggests subglacial formation.</td>
</tr>
<tr>
<td>Glacial Meltwater Channel (Drift)</td>
<td>Line</td>
<td>As above but incised into superficial drift deposits.</td>
<td>As above.</td>
</tr>
<tr>
<td>Delta Deposits</td>
<td>Polygon</td>
<td>Gently sloping upper surface which descends towards a steeper outer apex on the lake-side of the deposit. If the material was sourced from the nearby topography, the delta may have a meltwater channel or stream at its upper end or the delta may have an associated glacier margin or ice-contact slope if formed by glacial drainage.</td>
<td>Can be used to reconstruct former glacier margins and the height of former lake surfaces.</td>
</tr>
<tr>
<td>Feature</td>
<td>Type</td>
<td>Description</td>
<td>Significance</td>
</tr>
<tr>
<td>------------------------------</td>
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<td>-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------</td>
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</tr>
<tr>
<td>Kettle Hole</td>
<td>Polygon</td>
<td>A depression within another glacial deposit such as outwash. May now be filled with water, peat or fluvial deposits. Often approximately circular in shape and found in clusters.</td>
<td>Can infer the timing of the deposit in which the kettle hole has formed was simultaneous with the period of glaciation. Can be used to identify if a terrace or surface is of glacial origin/timing or post-glacially formed.</td>
</tr>
<tr>
<td>Esker</td>
<td>Line</td>
<td>Narrow, sinuous ridge composed of sand and gravel deposited on valley floors. Some eskers are single crested, others are braided and they can have up and down long profiles.</td>
<td>Eskers mark the former location of subglacial meltwater channels. These are parallel or subparallel to the former ice-surface slope, allowing the reconstruction of the direction of ice flow.</td>
</tr>
<tr>
<td>Kame Terraces</td>
<td>Polygon</td>
<td>Flat-topped terrace features that occur on the valley sides. Typically 10m in height and composed of sand and gravel with steep outer slopes marking the former ice-contact slope. Upper surface may be marked by kettle holes.</td>
<td>Indicate the position of the glacier margin and can be used to infer the style and pattern of retreat. Formed by sediment deposition from meltwater flowing laterally along the ice margin.</td>
</tr>
<tr>
<td>Undifferentiated</td>
<td>Polygon</td>
<td>Generic grouping of surfaces that exhibit some or all of the features above, or have other evidence of being shaped by glaciofluvial action. These surfaces may include kettle surfaces, meltwater channels and fragments of eskers.</td>
<td>Require further investigation to discern individual landforms.</td>
</tr>
<tr>
<td>Glaciofluvial Deposits</td>
<td>Polygon</td>
<td>Infill of sediment by accumulation of fluvial deposition on the valley floor; the surface gently slopes away from the former glacier margin. If buried ice was present at the time of formation, the surface may be kettled or have developed into a series of kames. If there was relatively little buried ice, the surface may still exhibit palaeochannels and the fluvial processes that formed it.</td>
<td>Indicates the position of the glacier margin. Outwash surfaces of different altitudes, origins or morphology can be used to identify different glacial events.</td>
</tr>
<tr>
<td>Outwash</td>
<td>Polygon</td>
<td>Ridge deposit found at the base of a steep slope. Surface is composed of unsorted local material and angular boulders transported via a snow slope to the ridge.</td>
<td>Infers a long-lived snow slope existed in a former period of climate deterioration. The presence of a protalus rampart indicates the absence of a glacier; however, more complex interactions between glaciers and protalus ramparts have been suggested.</td>
</tr>
<tr>
<td>Protalus Rampart or Pronival</td>
<td>Polygon</td>
<td>Arced alignments of boulders with a steep riser on their downhill side. Formed by the downhill movement of boulders.</td>
<td>Developed in a former period of climate deterioration. The last period of development is thought to be the Younger Dryas, and thus they can be used to infer the absence of warm-based glaciers at this time.</td>
</tr>
<tr>
<td>Boulder Lobes</td>
<td></td>
<td>Arced alignments of boulders with a steep riser on their downhill side. Formed by the downhill movement of boulders.</td>
<td>Developed in a former period of climate deterioration. The last period of development is thought to be the Younger Dryas, and thus they can be used to infer the absence of warm-based glaciers at this time.</td>
</tr>
<tr>
<td>Blockfield</td>
<td>Polygon</td>
<td>Angular blocks of rock that have broken up in situ during prolonged exposure to periods of cold weather and frost action. Common on plateau surfaces.</td>
<td>Blockfield development has occurred during former periods of climate deterioration. Their limits cannot always be used to infer the absence of ice because of their preservation under cold-based ice. In such situations their presence can be useful to understanding basal thermal regime.</td>
</tr>
<tr>
<td>Soliflucted Surfaces (Sheets</td>
<td>Polygon</td>
<td>Surfaces and formations that have formed through the downhill movement of sediment in cold environments. The better developed soliflucted surfaces form well-defined arcuate lobes with prominent risers, and often appear in a series.</td>
<td>The last phase of major solifluction development in Britain was during the Younger Dryas. Thus better developed solifluction formations can be used to indicate the absence of warm-based Younger Dryas glaciers.</td>
</tr>
<tr>
<td>and Lobes)</td>
<td></td>
<td></td>
<td>Tors have been shown to survive multiple glacial cycles under cold-based ice.</td>
</tr>
<tr>
<td>Tors</td>
<td>Point</td>
<td>Prominent rock outcrops that have persisted in situ after the weathering of surrounding rock.</td>
<td>Tors have been shown to survive multiple glacial cycles under cold-based ice.</td>
</tr>
<tr>
<td>Rock Glacier Deposits</td>
<td>Polygon</td>
<td>Ridge deposits comprised of coarse boulders with relatively well-defined margins consisting of lateral and terminal ridges, and often numerous ridges and depressions within this limit.</td>
<td>Formed during previous phases of climate deterioration. This would suggest the absence of a Younger Dryas glacier. Their presence has been associated with relatively cold and arid climatic conditions.</td>
</tr>
</tbody>
</table>
## Miscellaneous Landforms

<table>
<thead>
<tr>
<th><strong>Bedrock Outcrops</strong></th>
<th>Bedrock surfaces that have been subject to weathering including frost action.</th>
<th>May indicate a nunatak or the preservation of weathered features under cold-based ice.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Discrete Debris Accumulation</strong></td>
<td>Polygon Descriptive term used to describe a deposit without implying a formation process. Often comprises a ridge or mound of sediment and boulders beneath an overlooking cliff. Multiple formation methods are possible, but include rock glaciers, snow slopes, mass movement and glacier deposition.</td>
<td>Detailed work considering the processes that formed them can assist with establishing their palaeoclimatic significance.</td>
</tr>
<tr>
<td><strong>Lake Shoreline</strong></td>
<td>Line Bench-like change in profile of the slope, either through erosion or deposition of the bedrock or superficial deposits. The shoreline must be horizontal after taking into account isostatic adjustment and solifluction activity.</td>
<td>Used to reconstruct the dimensions of former ice-dammed lakes and contemporaneous positions of former glacier margins.</td>
</tr>
<tr>
<td><strong>Talus Deposits</strong></td>
<td>Polygon Accumulation of angular boulder deposits below overlooking crags, typically concave in form and sorted with larger material at the base. Slopes may be modified by debris flows and avalanche activity.</td>
<td>Rates of formation are thought to have been increased during periods of climatic deterioration. Differences in talus development/maturity and vegetation cover are thought to occur inside and outside Younger Dryas glacier limits.</td>
</tr>
<tr>
<td><strong>Regolith</strong></td>
<td>Polygon Unconsolidated rock and its components that have been broken up in situ by weathering processes.</td>
<td>It has been suggested that it is likely to have survived multiple glaciations and thus its location may infer cold-based ice.</td>
</tr>
<tr>
<td><strong>Terrace Undefined</strong></td>
<td>Generic term for terrace features that may have been formed as outwash, kame terraces or river terraces.</td>
<td>Additional work is required to understand their importance to glacier reconstruction.</td>
</tr>
<tr>
<td><strong>Alluvial Fan</strong></td>
<td>Polygon Cone-shaped deposit that typically originates at the base of a gully or chute. Formed by the deposition of fluvially transported material.</td>
<td>Deposits can mask important glacial and periglacial landforms. Larger alluvial fans, it has been suggested, form outside Younger Dryas limits.</td>
</tr>
<tr>
<td><strong>Gully</strong></td>
<td>Line A steep drainage channel that has been incised into sediment or bedrock.</td>
<td>Often modifies the shape and appearance of glacially deposited landforms.</td>
</tr>
<tr>
<td><strong>River Terrace</strong></td>
<td>Polygon Gently sloping terrace that occurs above the present river floodplain.</td>
<td>It has been suggested that one river terrace may exist above the floodplain inside Younger Dryas limits, whereas multiple river terraces may exist outside. In addition, the features may mask glacial landforms or be misinterpreted as outwash or delta deposits.</td>
</tr>
<tr>
<td><strong>Fluvial Surfaces/Deposits</strong></td>
<td>Polygon Postglacial surfaces reworked by fluvial activity in the immediate vicinity of the current river channel. This includes the present-day floodplain.</td>
<td>Used to account for the absence of glacial landforms and the reshaping or infilling of original glacial landscape.</td>
</tr>
<tr>
<td><strong>Undifferentiated Drift</strong></td>
<td>Polygon Superficial deposit with little morphological expression. These surfaces are likely to have had glacial origins; however, either they never had a strong morphological expression or it has been reworked by periglacial and postglacial processes.</td>
<td>No distinguishable importance to glacier reconstruction.</td>
</tr>
<tr>
<td><strong>Palaeochannels</strong></td>
<td>Polygon Sinuous network of former meanders and channels no longer in use by the present river.</td>
<td>May explain the absence or distribution of glacial landforms on the valley floor.</td>
</tr>
<tr>
<td><strong>Miscellaneous</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Stream, River</strong></td>
<td>Main drainage network used at the present time.</td>
<td>May explain the absence or distribution of glacial landforms on the valley floor.</td>
</tr>
<tr>
<td><strong>Lake</strong></td>
<td>Body of standing water.</td>
<td>Lakes often represent valley over deepening by glacial activity or the lake may be dammed by glacial deposits.</td>
</tr>
</tbody>
</table>

Sources: Ballantyne, 1996; Hubbard and Glasser, 2005; Lukas, 2006; Phillips et al., 2006; Harrison et al., 2008; Ballantyne et al., 2009a; Bennett and Glasser, 2009; Fabel et al., 2012.
3.3 Cosmogenic Surface Exposure Dating

3.3.1 Introduction and approach

There has been a recent focus on geochronological studies within the Cairngorms (Everest and Golledge, 2004; Everest and Kubik, 2006; Ballantyne et al., 2009a). The pattern of Dimlington Stade deglaciation in the northern Cairngorms is now established (Brazier et al., 1996b; Everest and Golledge, 2004; Everest and Kubik, 2006; Kirkbride and Gordon, 2010). In contrast, the extent of Younger Dryas glaciation has remained controversial; Sugden (1970) argued that glaciation was limited to the high corries, whereas Sissons (1979a) argued that it included the upper section of selected valleys. Recent surface exposure dating has established there was a small corrie glacier in Coire an Lochain (near Cairn Gorm) during the Younger Dryas (Kirkbride et al., 2014); however, there is no dating control elsewhere with which to resolve this controversy. The dating in this project focuses on determining whether Younger Dryas valley glaciers existed in the Cairngorms.

Cosmogenic nuclide surface exposure dating has been used to provide dating control in two valleys where the timing of glaciation is controversial and previous interpretations have been differing. The quartz-rich granite landscape of the Cairngorms lends itself to the measurement of the cosmogenic isotope beryllium 10 ($^{10}\text{Be}$). By accurately measuring the $^{10}\text{Be}$ concentration per gram of quartz, and knowing the $^{10}\text{Be}$ half-life and production rate at which $^{10}\text{Be}$ is produced, the exposure age of a boulder’s surface on a glacially deposited moraine can be calculated.
There are some important assumptions associated with the surface exposure dating of moraine boulders:

- Boulders do not possess inheritance from prior exposure on the valley sides or valley floor (to minimise this rounded boulders were selected that indicated erosion during transportation).

- Boulders have not moved/rotated since deposition by the glacier (boulders were sampled that were embedded within the moraine matrix to minimise the risk of sampling a rotated boulder).

- There has been no significant covering or shielding of the boulder surface by sediment or snow (depending on how well it is constrained, the effect of sediment and snow cover may be modelled and, to a degree, taken into account).

The accuracy of the exposure ages largely depends on how well these criteria are met (Phillips, 2001).

### 3.3.2 Production of cosmogenic nuclides

Cosmic rays are produced outside our solar system from supernova explosions (Diehl et al., 2006). On entering the atmosphere the primary cosmic rays collide with atomic nuclei producing a secondary cosmic radiation shower; the make-up of this shower turns progressively from proton to neutron dominated prior to reaching sea level (Dunai, 2010). Cosmogenic nuclides are the product of these cosmic rays’ interaction with atomic nuclei within the rock’s surface (Dunai, 2010). The cosmogenic nuclides can be produced through different production pathways, but the dominant pathway for $^{10}$Be is spallation reactions, which occur within the top 3m of rock (Phillips, 2001; Dunai, 2010). Muons penetrate much deeper into the rock; however, they have much lower nuclide production rates (Phillips, 2001). At sea level high latitude (SLHL) the production of $^{10}$Be is 96% by spallation and 4% by muons (Dunai, 2010). Combined, in quartz these production pathways produce c.4–5 atoms per gram per year; and despite the small quantities, these can be measured using an accelerator mass spectrometer (AMS).
3.3.3 Production rates

The best estimate for the production rate for $^{10}$Be has changed several times in the last few decades and will continue to be improved (Stone, 2000; Phillips, 2001). Recent updates have occurred in the light of the new $^{10}$Be half-life and independent calculation of AMS standard isotope ratios (Nishiizumi et al., 2007; Korschinek et al., 2010; Chmeleff et al., 2010; Balco 2009 and 2010 update documentation version 2.2 and 2.2.1). The default CRONUS-Earth online calculator uses a worldwide calibration dataset to derive reference production rates that vary with each scaling scheme (Balco et al., 2008).

Recent research suggests there may be reason to advocate the use of locally calibrated production rates (Ballantyne and Stone, 2012; Ballantyne, 2012). In Scotland, this was suggested after boulder samples in north-west Scotland yielded anomalously young ages, thought to be as a result of inaccuracies in globally calibrated production rates (Ballantyne and Stone, 2012). The majority of Scottish ages have been calculated using the default CRONUS-Earth online calculator; this bases production rates on a globally distributed calibration dataset where independent landform ages are known. Ballantyne (2012) noted the calibration sites as being mainly from mountainous environments between 30°N and 45°N, and that the ages suffered from large uncertainties and poor independent dating control. Instead, local calibration data from north-west Scotland, assumed to represent Younger Dryas deglaciation, has been used to derive a local production rate (Ballantyne and Stone, 2012; Ballantyne, 2012). These local production rates were used to recalculate ages of the nearby Lateglacial moraines which were found to be 6.5–12% older than when calculated using the globally calibrated production rate (Ballantyne and Stone, 2012).

The CRONUS-Earth website provides a separate calculator (developmental version) to use local samples of a known age to calculate a locally derived production rate; this can then be used to calculate sample ages of unknown ages. A further local production rate has been derived from Younger Dryas moraines, associated with a lake deposit independently dated using radiocarbon dating (D. Fabel, pers. comm.). This production rate has been used by Fabel et al., (2012) and Gheorghiu and Fabel (2013) and produces similar ages to the NWH production rate calibrated to 12.2 ka. This is a rapidly developing area and as better calibration
sites become available the accuracy of the production rates will improve. The ages in this study are presented both using the global and local production rates available. In addition, the raw data required for the future recalibration of the ages is provided.

3.3.3.1 Elevation and latitude

Production rates vary worldwide based on regional atmospheric pressure anomalies and variations in geomagnetic fields (Lal, 1991; Dunai, 2000; Stone, 2000). Scotland does not suffer from long-term atmospheric anomalies such as those experienced in Antarctica (Stone, 2000; Phillips, 2001), or large geomagnetic variations which become more important at latitudes less than 40° (Dunai, 2000; Phillips, 2001). However, scaling factors are important to account for the elevation and latitude of the sample site (Lal, 1991). For this reason, reference production rates are often presented for sea level high latitude (SLHL) areas (≥60°) and then scaled to the specific sample site (Phillips, 2001). Production rates rise by approximately 1% with every 10m increase in elevation (Stone, 2000); thus making it important to record the position and, in particular, the altitude of the sample site. The CRONUS calculator then scales the production rate to the specific sample site location.

3.3.3.2 Sample shielding, sample thickness, erosion and density

The sample-specific production rates have also been corrected for topographic shielding, sample thickness and surface erosion rates. Despite the samples being taken from a mountainous area, the corrections are minor because of the cosmic radiation being strongly focused from the vertical (Gosse and Phillips, 2001). The method for the shielding calculation is given later. As the production rates attenuate with depth, corrections have been applied for the sample thickness within the CRONUS calculators. This form of shielding was corrected for by measuring the thickness of the sample and its density. The density of the granite samples is assumed from the formerly used density of c.2.7 g cm$^{-3}$ for the Cairngorms (Phillips et al., 2006). However, slight variations of (in the order of 0.1 g cm$^{-3}$) have negligible impact on the ages.

Additional corrections have been applied for the erosion of the boulder surface due to weathering. When using a single isotope such as $^{10}$Be, a rate can be estimated.
from the height of quartz veins or known rates derived from rocks of similar lithology and climatic history. Granite is composed predominantly of feldspar, quartz and mica. Feldspars weather rapidly by chemical reaction with water; however, the quartz is very resistant (Thomas et al., 2004), thus if quartz veins are found on boulder surfaces this can be used to identify minimum postglacial erosion rates. This faster weathering of the feldspar component of granite, combined with other methods such as different thermal expansion and contraction rates of the granite components, can cause the break-up of granite through granular disintegration (Gómez-Heras et al., 2006). Postglacial weathering rates from Lapland for granite and metaphoric rocks ranged from 0.2 to 1.2mm ka (André, 2002). This is likely to be representative of the erosion rate, and any additional errors from boulder erosion were circumvented by avoiding sampling areas that have visible signs of erosion. Burial in peat and waterlogged ground can increase rates of weathering (Curran et al., 2002), but this is unlikely to be a problem because of sampling from moraine crests.

3.3.3.3 Snow and sediment cover

Cover by snow, peat, soil, volcanic ash and vegetation all have the capacity to decrease the boulder surface production rate and thus reduce the apparent boulder age (Dunai, 2010). An example is that if snow cover persisted for 4 months of the year with a depth 75–150 cm it would reduce the production rate by 5% (Dunai, 2010). Modelling work by Schildgen et al. (2005) has estimated the impact of snow cover on cosmogenic ages within the Cairngorms. This work, alongside further analysis of the vulnerability of the new ages to snow shielding, is discussed in Chapter 5. To minimise the new ages’ vulnerability to snow shielding, the samples were taken from boulders on the crests of moraines, sites expected to experience snow deflation by wind.

Sediment shielding has the potential to have a significant impact on the apparent age of boulder samples (Putkonen and Swanson, 2003; Putkonen and O’Neal, 2006; Heyman et al., 2011). Often it is incorrectly assumed that the age derived from a boulder surface is representative of the moraine formation age. While some boulders will remain on the surface of the moraine since the time of deposition, others will be exhumed later, as the fine-grained moraine matrix is eroded (Putkonen and Swanson, 2003). The more recently exhumed boulders will have
been subjected to a reduced production rate while covered, and thus yield younger apparent boulder ages. Fortunately it is thought that most moraine-crest lowering takes place soon after moraine formation, therefore most boulders will be exhumed within the first few thousand years (Putkonen and Swanson, 2003). However, after this rapid period of degradation, a reduced rate of moraine matrix erosion will continue. Modelling analysis has suggested that the moraine size applicable to this study, less than 20m in height, suffered less from exhumation complications and suggested fewer samples are required to yield an age >90 % of the true moraine formation age (Putkonen and Swanson, 2003). The sampling and modelling strategy described below is aimed to identify boulders that have suffered from sediment cover, and endeavours to calculate and correct the moraine formation age for the matrix erosion.

3.3.4 Sampling strategy

3.3.4.1 General precautions

There are numerous considerations that must be taken into account when selecting a boulder for sampling. Firstly the rock type: for $^{10}$Be cosmogenic dating a source of quartz is required; this can either be quartz veins or a high percentage of quartz grains, such as in Cairngorm Granite. Selecting a boulder that has been glacially transported is essential, thus only boulders rounded from glacial transportation were selected. More angular boulders and boulders with a close proximity to rock cliff faces or steep slopes were avoided. Boulders were selected from relatively flat areas of the moraine crest, not on or at the bottom of slopes where the boulder may have suffered postglacial movement or rotated. For the same reason, only boulders embedded within the moraine matrix were selected; this reduces the chance of selecting a rotated boulder either from erosion of the supporting matrix or through tree throw (Á. Rodés, pers. comm.) (Figure 3.4). Postglacial areas of erosion and accumulation of sediment were avoided when selecting sample sites. Moraines often dam large volumes of water as the glacier recedes; this can cause sediment infilling behind moraines and channels to be eroded over the moraine crest. Therefore any postglacial channel surfaces were avoided and samples were taken from the higher original moraine surface. These general criteria above formed the basis for the paired sampling strategy outlined below.
Figure 3.4 Boulders included within the moraine matrix are less likely to have rotated through matrix erosion or tree throw (figure courtesy of Á. Rodés).

3.3.4.2 Paired sampling and constant matrix erosion modelling

Reconnaissance fieldwork of the proposed sampling sites found most boulders on the moraine crests to have relatively low heights, approximately 1m or less above the surrounding terrain. Collaboration with Á. Rodés of NERC-CIAF led to the derivation of a sampling and modelling strategy to calculate both matrix erosion and the exposure age of the moraine. Á. Rodés wrote the programme using Mathematica; a brief explanation of the model is given below and further details can be found within the Appendix.

The present-day moraine surface is composed of boulders exposed at the time of moraine formation, but also boulders that have been exhumed more recently. Moraine matrix material can be eroded by water erosion such as splash erosion, sheet-wash, rill erosion, or by wind erosion such as surface creep, saltation and suspension. Erosion is likely to be variable with climate such as aridity, and the development of vegetation cover is likely to reduce erosion. In addition, erosion is likely to reduce with time due to changes in moraine shape from sharp-crested to more rounded shallower profiles (Putkonen and O’Neal, 2006). Furthermore, it is likely that armouring of the moraine surface with larger boulders and coarser matrix will occur progressively as finer material is eroded first (Poesen et al., 1994; Granger et al., 2001). However, such time-variable influences are difficult to constrain; thus, given the importance of moraine degradation to cosmogenic ages
and the assumption of no moraine matrix erosion being incorrect, a constant moraine matrix model should be considered; albeit with the acknowledgment that complex time-variable components do exist. The approach below describes a data driven method that assumes the moraine surface has been eroding at a constant rate.

Given that some moraine surface erosion is likely, nearby boulders of different heights above the moraine surface at present may have been exhumed from the moraine surface at different stages in the moraine’s history. Thus using a paired sampling technique, of two or more boulders of different heights above the moraine surface, should yield two different ages if the moraine surface has undergone lowering, or the same age if the lowering has been insignificant (Figure 3.5). This is owing to the buried boulder/s receiving a reduced production rate during burial.

Numerical modelling has been used to estimate the moraine surface erosion rate and moraine age. The modelling assumes a moraine matrix density of 2g/cm³, an attenuation length of 160g/cm² for spallation reactions, and no muon attenuation over the small distance concerned. The sampling technique provided a set of data with different boulder heights and $^{10}$Be concentrations. Numerical modelling provided the moraine age and sediment erosion rate values that best fit this dataset; further details are given in Chapter 5. In order to do this, the height of the boulder above the surrounding moraine matrix was recorded.
Figure 3.5 Theoretical $^{10}$Be and $^{26}$Al concentrations on the surface of paired boulders of 20cm and 100cm above the terrain, for moraines formed at 12, 15 and 18 ka BP with differing erosion rates (0,50 and 100mm ka$^{-1}$). The difference in concentration between the boulders is used to calculate the erosion rate at which the surface has been eroded and to deduce the likely moraine age (figure courtesy of Á. Rodés)

### 3.3.4.3 Sample extraction

The selected boulders were sampled using a chisel and hammer (Figure 3.6 and Figure 3.7). The sample was taken from the uppermost surface of the boulder, avoiding shielding by the boulder’s own surface or larger neighbouring boulders. A sample of 0.5–1.2 kg was taken and weighed using portable digital scales. The average sample thickness was recorded in the field, as were the GPS coordinates and GPS altitudes (m a.s.l. Newlyn) (Table 3.2). The GPS data was collected with a handheld GPS Garmin Etrex and checked later using NEXTMap and Ordnance Survey Maps. GPS errors in the sample site altitudes were identified and have been corrected for prior to age calculation (Table 3.2). The height of the boulder above the moraine surface and the proximity to other sampled boulders was sketched and recorded in the field (Table 3.2).
Figure 3.6 Glen Geusachan (LGG1,3,5) sampled boulders with sample bags indicating sampled boulder.
Figure 3.7 Glen Derry (HH1,2,3) sampled boulders with sample bags indicating sampled boulder.
Table 3.2 Summary table of sample field data

<table>
<thead>
<tr>
<th>Sample</th>
<th>Lithology</th>
<th>Lat (°N)</th>
<th>Long (°W)</th>
<th>Altitude GPS Corrected* (m a.s.l.)</th>
<th>Weight (kg)</th>
<th>Thickness (cm)</th>
<th>Density (g cm⁻³)</th>
<th>Shielding</th>
<th>Boulder Height (cm)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glen Derry</td>
<td>Granite Boulder</td>
<td>57.07729</td>
<td>-3.59593</td>
<td>647</td>
<td>642</td>
<td>0.52</td>
<td>2.0</td>
<td>2.7</td>
<td>0.99064</td>
<td>80</td>
</tr>
<tr>
<td>HH2 Granite Boulder</td>
<td>57.07729</td>
<td>-3.59593</td>
<td>647</td>
<td>642</td>
<td>0.56</td>
<td>2.0</td>
<td>2.7</td>
<td>0.99066</td>
<td>30</td>
<td>65</td>
</tr>
<tr>
<td>HH3 Granite Boulder</td>
<td>57.07464</td>
<td>-3.59549</td>
<td>612</td>
<td>612</td>
<td>0.92</td>
<td>2.0</td>
<td>2.7</td>
<td>0.98928</td>
<td>140</td>
<td>140</td>
</tr>
<tr>
<td>Glen Geusachan</td>
<td>Granite Boulder</td>
<td>57.02818</td>
<td>-3.68160</td>
<td>517</td>
<td>517</td>
<td>1.23</td>
<td>2.5</td>
<td>2.7</td>
<td>0.99078</td>
<td>n/a</td>
</tr>
<tr>
<td>LGG3 Granite Boulder</td>
<td>57.02754</td>
<td>-3.68021</td>
<td>518</td>
<td>518</td>
<td>1.08</td>
<td>2.0</td>
<td>2.7</td>
<td>0.99159</td>
<td>n/a</td>
<td>n/a</td>
</tr>
<tr>
<td>LGG5 Granite Boulder</td>
<td>57.02802</td>
<td>-3.68162</td>
<td>527</td>
<td>518</td>
<td>0.95</td>
<td>1.5</td>
<td>2.7</td>
<td>0.99019</td>
<td>n/a</td>
<td>n/a</td>
</tr>
</tbody>
</table>
As described above, the shielding of the boulder surface from the surrounding mountains can be important. The skyline was recorded in the field using a compass and clinometer (Table 3.3). The inclination was recorded at horizon intervals that best approximated the topography, rather than using predefined intervals. These were later converted into a shielding factor using the CRONUS online Geometric Shielding calculator (version 1.1. written by Greg Balco
http://hess.ess.washington.edu/math/general/skyline_input.php). All the samples were from relatively open sites, thus the effect of shielding is negligible (Table 3.2).

Table 3.3 Shielding azimuths and elevations recorded for each sample site

<table>
<thead>
<tr>
<th>Sample</th>
<th>Azimuths: 0 20 40 90 125 150 170 185 190 200 225 240 260 265 270 290 295 300 325 335 355</th>
<th>Elevations: 7 10 17 18 7 6 0 1 2 6 17 21 18 11 12 8 8 11 19 15 12</th>
</tr>
</thead>
<tbody>
<tr>
<td>HH1-2</td>
<td>Azimuths: 0 20 65 105 120 130 170 175 180 190 195 215 240 260 280 290 295 300 310 330 340 355</td>
<td>Elevations: 7 11 18 8 9 8 2 0 2 3 5 10 22 25 20 12 9 9 12 17 13 11</td>
</tr>
<tr>
<td>HH3</td>
<td>Azimuths: 0 10 20 60 80 105 130 145 165 180 190 210 240 260 290 295 340 355 360</td>
<td>Elevations: 9 7 10 17 13 3 2 3 2 2 9 15 17 9 7 11 27</td>
</tr>
<tr>
<td>LGG1</td>
<td>Azimuths: 0 5 10 20 60 75 105 135 145 160 185 190 210 240 250 280 295 305 335 345 355</td>
<td>Elevations: 6 5 9 9 17 13 4 3 4 2 3 10 16 17 8 12 10 18 24</td>
</tr>
<tr>
<td>LGG3</td>
<td>Azimuths: 0 10 15 20 30 70 85 105 130 135 150 160 175 180 185 210 235 245 285 290 310 345 355</td>
<td>Elevations: 9 17 9 10 16 12 3 2 4 1 1 3 10 17 18 8 7 10 18 27</td>
</tr>
</tbody>
</table>

Figure 3.8 Skyline plots from the online calculator for HH1 (Glen Derry) and LGG1 (Glen Geusachan), with the corresponding shielding factor

HH1 = 0.990638111

LGG1 = 0.99078283
3.3.5 Sample preparation

All stages of sample preparation and AMS measurement were completed at Natural Environment Research Council-Cosmogenic Isotope Analysis Facility (NERC-CIAF) within the Scottish Universities Environmental Research Centre (SUERC) East Kilbride, Glasgow. The author was trained and carried out all aspects of the physical preparation and most stages of the chemical preparation. Sample preparation is carried out to purify the measured target material for AMS analysis; this can be subdivided into physical and chemical preparation.

3.3.5.1 Physical

First the 0.5–1.2kg granite samples were crushed and sieved to 0–250 microns, 250–500 microns and >500 microns fractions. A mass of 250g of the 250–500 micron fraction was required; for the smaller samples this involved additional crushing or use of a disk mill to reduce the grain size of the >500 micron fraction. The samples were then bagged according to their respective grain size. After each sample was crushed and sieved, the equipment was cleaned to reduce contamination. The 250–500 micron fraction was then washed to remove lichen, dust and organics and placed in a heated cupboard to dry the samples.

A Frantz isodynamic magnetic mineral separator was used to separate the magnetic and non-magnetic minerals within the sample; the non-magnetic fraction was then bagged and labelled ready for the chemical preparation. As the samples were granite, this stage removed the darker coloured micas and magnetic material, leaving the lighter coloured feldspar and quartz grains. Again, between each sample run, the machinery was cleaned with compressed air to avoid contamination.

3.3.5.2 Chemical

Where possible a mass of c.200g per sample was etched to ensure there was a sufficient purified quartz mass to allow for good resolution and distinction between concentrations of Younger Dryas and Dimlington Stade age. Each sample was weighed (samples ranged from 123–206g) before adding 200ml of hexafluorosilicic acid within a fume cupboard. The sample was then secured to a shake table for 24 hours. The acid was then poured into a tank for neutralisation and another 200ml of
hexafluorosilicic acid added; this process was repeated multiple times. This selective etching process of removing impurities and meteoric $^{10}$Be can also dissolve up to 30% of the target mineral, in this case quartz (Kohl and Nishiizumi, 1992). The etching was followed by analysis of the purity of the quartz under a microscope, and a small test sample was dissolved in 40% hydrofluoric acid (HF) to allow the quartz purity to be measured. Results of an Al concentration less than 200 ppm indicate pure quartz with the meteoric $^{10}$Be removed (Wilson et al., 2008). The samples measured between 48 and 60 ppm, substantially below the threshold required for both Be and Al analysis.

The purified quartz was then dried, added to a beaker and weighed. Next the quartz was dissolved in 40% HF acid and heated to evaporate the HF. This process was repeated until all the quartz was dissolved. Two samples were not fully dissolved (HH3 and LGG3); the mass that did not dissolve was subtracted from the initial mass of quartz.

After quartz purification, the isotope was extracted. The aim was to extract as much cosmogenic nuclide and remove unwanted components (Dunai, 2010). First, a known mass of $^9$Be carrier with a known concentration (Uhann – 405.4ppm) was added to the quartz samples, and the samples were then dissolved in HF. The addition of a known quantity of $^9$Be allows the AMS to determine the ratio of $^{10}$Be/$^9$Be. The carrier has not been exposed to cosmogenic radiation; however, any $^{10}$Be added during the chemical preparation was corrected for with the use of a blank. Next, selective precipitations were conducted to remove unwanted elements. Ion chromatography was used to separate the solutions, anion exchange removed Fe, and the samples were then converted to a sulphate, prior to cation exchange to remove Ti and separate the Be and Al fractions.

The Be was then brought to a pH of 9 by adding Ammonia Na$_4$OH and centrifuged to precipitate beryllium hydroxide Be(OH)$_2$. The samples were then washed, transferred to quartz crucibles and dried on a hotplate. The samples were then heated in a furnace for conversion to beryllium oxide BeO.
3.3.6 Accelerator mass spectrometry (AMS)

The $^{10}\text{Be}/^{9}\text{Be}$ ratio measurement took place using the 5MV NEC Pelletron accelerator mass spectrometer at the SUERC (Freeman et al., 2004). The AMS process separates the required isotopes before counting the isotopes and recording the $^{10}\text{Be}/^{9}\text{Be}$ ratio (Dunai, 2010); further details on the procedure can be found within Wilson et al. (2008).

3.3.7 Exposure age calculation: CRONUS-Earth online calculators

3.3.7.1 Inputs

The CRONUS-Earth online calculators were used with the field, laboratory and AMS data to calculate the samples’ exposure ages. The calculators were designed to enable the direct comparison of multiple exposure dating studies using common methods to calculate the ages. This is important as it allows the synthesising of results and recalculation of ages in the future as long as the necessary information is published (Balco et al., 2008). The majority of the inputs (Table 3.4) are self-explanatory, have been described in the relevant sections above and are given again in Chapter 5.

Table 3.4 Required information for input into CRONUS online calculators (adapted from Balco et al., 2008)

<table>
<thead>
<tr>
<th>Field</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample Name</td>
<td>Text</td>
</tr>
<tr>
<td>Latitude</td>
<td>Decimal Degrees</td>
</tr>
<tr>
<td>Longitude</td>
<td>Decimal Degrees</td>
</tr>
<tr>
<td>Elevation</td>
<td>m a.s.l.</td>
</tr>
<tr>
<td>Sample Thickness</td>
<td>cm</td>
</tr>
<tr>
<td>Sample Density</td>
<td>g cm$^{-3}$</td>
</tr>
<tr>
<td>Shielding Correction</td>
<td>Nondimensional between 0 and 1</td>
</tr>
<tr>
<td>Erosion Rate</td>
<td>cm yr$^{-1}$</td>
</tr>
<tr>
<td>Nuclide Concentrations</td>
<td>atoms g$^{-1}$</td>
</tr>
<tr>
<td>Uncertainties in Nuclide Concentrations</td>
<td>atoms g$^{-1}$</td>
</tr>
<tr>
<td>Name of $^{10}\text{Be}$ Standardisation</td>
<td>Text</td>
</tr>
</tbody>
</table>

The only calculator inputs that require further explanation are the nuclide concentrations and standardisation used. The AMS measures the $^{10}\text{Be}/^{9}\text{Be}$ ratio of the sample compared to the isotope ratio of a standard material whose absolute isotope ratio is known independently (Balco et al., 2008). There have been several
recent advances in this area. Nishiizumi et al. (2007) calculated the $^{10}\text{Be}/^{9}\text{Be}$ ratio of the standards independent of the $^{10}\text{Be}$ half-life. The paper also compared the isotope ratio between regularly used AMS standards, thus allowing the conversion between different standards, which is now included in the online calculators (Balco, 2009 online update 2.2). Recently the accuracy of the half-life of $^{10}\text{Be}$ has been improved by two studies whose experimental methods were different, giving a weighted mean half-life of 1.387±0.012Ma (Korschinek et al., 2010; Chmeleff et al., 2010). This has been included within the updated version of the online calculator (2.2.1). Much of the complexity is circumvented by the use of the online calculator and it allows easy recalculation of the ages as technical advances are made; however, it is essential the laboratory data is converted and input into the calculator correctly.

The AMS measures the $^{10}\text{Be}/^{9}\text{Be}$ ratio; SUERC report the ratio and uncertainty referenced to the standard for three different assumed $^{10}\text{Be}$ half-lives: 1.53Ma, 1.34Ma and 1.36Ma. The report also includes the standard used and the ratio recorded for the standard given the three different assumed half-lives. The standard material has a known $^{10}\text{Be}/^{9}\text{Be}$ ratio which the sample measurements are referenced against. The results have been calculated using the 1.36Ma half-life ratio; however, this is irrelevant as all the half-life ratios yield the same exposure ages when input with the corresponding nominal ratio. The nominal ratio for the 1.36Ma half-life is $2.79 \times 10^{-11}$ thus the standardisation code used to input the data into CRONUS online calculators was NIST_27900. All nuclide concentrations are now renormalised to the 07KNSTD standard internally within the calculator; thus sample concentrations referenced to different standards and with different assumed half-lives can be normalised prior to age or production rate calculation (Balco, 2009 updated version 2.2).

The sample and blank $^{10}\text{Be}/^{9}\text{Be}$ ratios and 1 sigma uncertainty ratios from the AMS were converted to atoms per g of quartz prior to input into the CRONUS online calculators. The mass of the carrier added (g) and carrier concentration were used to calculate the Be added. This was converted to the number of atoms of $^{9}\text{Be}$ added, using Avogadro’s number (6.02214129 x $10^{23}$ atoms mol$^{-1}$) and the molar weight of Be (9.012182g mol$^{-1}$). The known number of $^{9}\text{Be}$ atoms added was then multiplied by the $^{10}\text{Be}/^{9}\text{Be}$ ratio to calculate the unknown number of $^{19}\text{Be}$ atoms.
The number of \(^{10}\)Be atoms within the blank was subtracted from each sample to calculate the number of \(^{10}\)Be atoms within each sample. Finally, the number of \(^{10}\)Be atoms was divided by the mass of the quartz to calculate the number of \(^{10}\)Be atoms per gram of dissolved quartz. This method is explained by Balco (2006) in the CRONUS-Earth online documentation.

The one-sigma uncertainty was calculated using error propagation. This includes the chemistry errors associated with carrier addition and a balance error of 0.0005, measured AMS error, and AMS and chemistry errors associated with the blank. The final stage is to convert the error in atoms to a concentration of atoms per gram of quartz including the incorporation of the necessary balance errors. This concentration was then included within the online calculations to generate one-sigma age uncertainties.

### 3.4 Glacier Reconstruction and ELA Calculation

#### 3.4.1 Introduction and approach

Using the mapped geomorphological evidence presented in Chapter 4, interpretations were made of the former glacier margins. Glacier reconstruction was then undertaken; where possible this used vertical geomorphological constraints such as lateral moraines and boundaries between glacial drift and soliflucuted surfaces. However, suitable geomorphological evidence to constrain the vertical limit of the glaciers was often absent. In addition, evidence for or against the simultaneous glaciation of the plateau surfaces was often subtle or absent. Thus flow-line modelling was used to extrapolate between areas of known vertical control. Where the contribution of plateau ice was uncertain both valley- and plateau-sourced glaciers were reconstructed. These reconstructions were then used to calculate the former glacier ELAs.

#### 3.4.2 Ice-surface profiler

The ice-surface profile of the glaciers has been reconstructed through a combination of geomorphological evidence and a modified version of the ice-surface profile model created by Benn and Hulton (2010). The model (Profiler v.2) allows the former ice-surface profile to be calculated from a known glacier terminus, based on the assumed basal shear stress and measured valley shape factor. The
present-day valley-floor profile extracted from the NEXTMap DEM was input from the glacier terminus to the head of the back wall or ice divide of the plateau, as appropriate. The shape factor was calculated using cross sections of the valley at intervals that were deemed to best characterise the changes in the valley geometry (Rea and Evans, 2007). Once the cross sections were extracted from the NEXTMap DEM at key locations along the valley (Figure 3.9), the glacier's cross-sectional area was calculated and divided by the ice thickness multiplied by the glacierised perimeter to generate the shape factor (Benn and Hulton, 2010). This factor adjusts the shear stress based on the proportion of driving stress that is supported by the bed versus drag from the valley sides (Benn and Hulton, 2010). At this stage of DEM data extraction, the presence of lakes and postglacial changes in valley shape are a potential source of error. The shape factors for the Cairngorm valleys and corries were typically between 0.55 and 0.75. The lower values are locations where a greater proportion of ice is supported by drag from the valley sides and values closer to 1 where the ice is mainly supported by the bed. On the plateau surfaces a shape factor of 1 was applied to represent the lack of any supporting valley slides (Rea and Evans, 2007).

The main variable that controls the ice-surface profile is the basal shear stress. The key controls on basal shear stress are ice thickness and surface slope (Bennett and Glasser, 2009); however local fluctuations are caused by other factors such as bedrock protrusions and cavities. The spatial difference between accumulation and ablation causes the ice surface to increase until a shear stress is reached that will cause the ice to deform and flow (Bennett and Glasser, 2009). Basal shear stress values for flow over a rigid substrate in modern glaciers are consistently 50–100 kPa (Bennett and Glasser, 2009). Lower basal shear stresses and therefore shallower ice-surface profiles can occur where water or deformable sediments are present at the glacier base (Nesje and Dahl, 2000). Previous glacier reconstruction studies have used values of 50–100 kPa (Ballantyne et al., 2011; Hughes et al., 2011) with a constant value of 100 kPa used by Rea and Evans (2007) and an approximate value of 50 kPa inferred from geomorphological evidence by Finlayson et al. (2011). In locations of well-preserved vertical geomorphological control, the basal shear stress value can be tuned to match the geomorphology (Schilling and Hollin, 1981; Murray and Locke, 1989; Benn and Hulton, 2010). The approach taken within this study was to use basal shear stress values of 50–100 kPa to tune
the ice surface to the geomorphological evidence. Where no vertical control existed, a shear stress value of 75 kPa was applied.

The profiler v.2 model was modified to allow the incorporation of a higher resolution step length, calculation of larger glaciers and easier output of the ice-surface data for GIS-based glacier reconstruction at 50m altitude intervals; the equations remained the same as in Benn and Hulton (2010). The model visually displayed the ice-surface profile (Figure 3.9). Where the ice thickness attenuated towards 0m at the back wall of the valley/corrie, this indicated the upper end of the glacier (Benn and Hulton, 2010). This does not, however, indicate the absence of ice on the plateau above – this must be considered independently. Where the plateau may have contributed ice, both valley- and plateau-sourced reconstructions were undertaken. On the plateau surfaces, a decrease in shear stress towards zero was required to simulate the approach of the ice divide (Benn and Hulton 2010; Finlayson et al., 2011; Trelea-Newton and Golledge, 2012). This approach provided the centreline ice-surface profile for the former glaciers; multiple flowlines were used when the glacier had numerous accumulation areas (Figure 3.9). The glacier contours were then digitised at 50m intervals to reconstruct the glacier shape; this was completed with reference to modern glaciers using concave contours above the ELA and convex contours below the ELA. This generated 50m altitudinal bands that were utilised for ELA calculation.
Figure 3.9 Example glacier reconstruction. a) the extraction of the valley long (black) and cross (red) profiles. b) example valley cross section and corresponding shape factor, $F$. c) example long profile bed elevation and calculated ice surface from the ice-profiler model. d) valley glacier reconstruction using output from ice-profiler model

### 3.4.3 ELA calculation

The ELA of a glacier is the average elevation at which the net annual accumulation and ablation are equal; this line varies across a glacier surface according to spatial variations in accumulation and ablation (Benn and Lehmkuhl, 2000). When a glacier is in equilibrium with the climate, i.e. the mass balance of the whole glacier is zero, it can be described as having a steady-state ELA (Benn and Evans, 2010). This is unusual and is largely a theoretical concept; however, it is what is meant when describing palaeoglacier ELAs (Benn and Evans, 2010).

Numerous methods are available for ELA calculation. The Area-Altitude Balance Ratio (AABR) method is commonly thought of as the most favourable method and is used here because it takes into account glacier mass balance and glacier hypsometry – the variation of glacier area with altitude (Furbish and Andrews, 1984; Osmaston, 2005). This is important as glacier accumulation or ablation typically becomes more pronounced with altitudinal distance from the ELA; therefore the
AABR weights areas further above or below the ELA more than those closer to it (Osmaston, 2005). The method therefore assumes that the ablation gradient and accumulation gradient are approximately linear, commonly with steeper ablation gradients; and the ratio between the two is known or can be accurately estimated (Benn and Lehmkuhl, 2000). However, it is important to recognise that mass-balance gradients are influenced to varying degrees by localised factors such as snow redistribution, radiation shading, avalanching and debris cover (Benn and Lehmkuhl, 2000). Despite this, the AABR approach is thought to be suitable for most glaciers, with the exception of debris-covered glaciers and glaciers that receive high inputs from avalanches (Benn and Lehmkuhl, 2000).

The glacier areas within each 50m contour band were extracted from GIS and the ELA calculated using the spreadsheet provided by Osmaston (2005). First the areas within each contour band were weighted according to their mean altitudinal distance above or below the trial ELA. The correct ELA elevation was then identified where the ablation and accumulation were equal for the specified balance ratio (Osmaston, 2005). The ELA was calculated using multiple balance ratios; this is the ratio between the ablation and accumulation gradients and is thought to vary with climatic conditions (Furbish and Andrews, 1984; Rea, 2009). The ELAs have been calculated using recommended balance ratios based on modern glaciers in mid-latitude maritime 1.9 ± 0.81 and Western Norway 1.5 ± 0.4 climates (Rea, 2009). The ELAs have also been calculated using balance ratios of 1.0, 1.67, 1.8 and 2.0 to allow comparison with other studies within the British Isles. A balance ratio of 1.67 is equivalent to using the accumulation-area ratio (AAR) method with a ratio of 0.6 for a planar glacier (Ballantyne, 2002).

3.5 Palaeoprecipitation: Temperature–Precipitation Relations

3.5.1 Introduction and approach

It is recognised that a relation between temperature and precipitation exists at the ELA of modern glaciers (Ohmura et al., 1992). Given the available palaeotemperature records derived from proxies, such as chironomids, the former precipitation values at the ELA can be estimated.

The chironomid-derived palaeotemperature record from the Abernethy Forest, at 230m a.s.l. on the northern margin of the Cairngorms, indicated the minimum
Younger Dryas mean July temperature was 6.8°C (Brooks et al., 2012). The temperature was estimated using a model derived from a modern Norwegian chironomid-based temperature calibration dataset consisting of 157 lakes and has a root mean squared error of prediction of c.1°C (cf. Brooks et al., 2012). Therefore a temperature of 6.8±1°C has been used to derive the palaeoprecipitation values. Some additional caution should be acknowledged given the assumptions and uncertainties associated with using biological proxies for temperature reconstruction (see Brooks and Birks, 2001; Juggins, 2013).

This July temperature was then converted to a 3-month summer temperature by multiplying by 0.97 based on an empirical relationship derived from meteorological stations in Scotland and Scandinavia (cf. Benn and Ballantyne, 2005; as used by Ballantyne, 2006; Finlayson, 2006; Ballantyne, 2007a, 2007b; Carr and Coleman, 2007; Bendle and Glasser, 2012). Some caution should be acknowledged regarding the strength of such a relationship and whether the present-day relationship can represent Younger Dryas climate. The temperature was scaled to the ELA of the glacier using a lapse rate of 0.0065±0.0005°C per m. This temperature was then used within the temperature–precipitation equations to derive a precipitation value; these equations are described in Section 3.5.2 and 3.5.3 below.

This provides a former precipitation value for the altitude of the ELA. In turn, this can be scaled to other altitudes or sea level using a non-linear relationship between precipitation and elevation. Ballantyne (2002) used data originally collected between 1884 and 1903 from the foot and summit of Ben Nevis to calculate that precipitation increased by 5.8% per 100m rise in elevation. An alternative value of 8–9% per 100m rise in elevation has been suggested for Norway (Haakensen, 1989; Dahl and Nesje, 1992).

**3.5.2 Global relations (Ohmura et al., 1992)**

Ohmura et al. (1992) used a database of 70 globally distributed glaciers to determine the relation between temperature and precipitation at the ELA. Using local palaeotemperature this dataset has been widely used to reconstruct former precipitation values (Benn and Ballantyne, 2005; Finlayson, 2006; Ballantyne, 2007a; Bendle and Glasser, 2012). The Ohmura et al. (1992) equation is based on
a globally derived dataset and thus smoothes differences between climatically different regions (Benn and Ballantyne, 2005). However, this means important local climate factors are ignored and it has been suggested to overestimate precipitation values in Scotland during the Younger Dryas (Golledge et al., 2010). The equation for the best-fit polynomial regression curve by Ohmura et al. (1992) is:

\[ P_a = 645 + 296T_3 + 9T_3^2 \]

**Equation 3.1**

Where \( P_a \) is precipitation in mm per annum (winter balance plus summer precipitation) and \( T_3 \) the mean 3-month summer temperature in °C, the standard error of the relationship is 200mm per annum (Ohmura et al., 1992). Note the winter balance includes additional snow redistributed by wind or avalanches onto the glacier surface.

### 3.5.3 Scottish Younger Dryas relation (Golledge et al., 2010)

The use of a globally derived temperature–precipitation relation has its limitations and is thought to overestimate precipitation values in Scotland, particularly in colder, more arid regions (Golledge et al., 2010). Based on numerical modelling of the Younger Dryas glaciers in Scotland (Golledge et al., 2008, 2009), Golledge et al. (2010) designed a method to calculate precipitation values that takes into account the seasonality of the climate. It uses an annual climate variation of 30°C based on biological proxy-based records, approximately three times that of the present climate (Golledge et al., 2010). The model can also take into account the seasonality of precipitation; this includes a winter-dominated, neutral and summer-dominated scenario. Golledge et al. (2010) favour the summer-dominated scenario, as winter precipitation is thought to have been reduced by the presence of sea ice. The equation for precipitation is given below:

\[ P = S(14.2T_3^2 + 248.2T_3 + 213.5) \]

**Equation 3.2**

Where S is a scaling coefficient for precipitation under different seasonal scenarios; 1 for neutral, 1.4 for summer-dominated and 0.8 for winter-dominated and \( T_3 \) is the 3-month mean summer temperature at the ELA.
3.6 Topoclimatic Modelling

3.6.1 Introduction and approach

Local variations in glacier ELA are known to be caused by topoclimatic factors such as differences in contribution from snow blow, avalanching and solar radiation. Better understanding of these factors may explain any variations in glacier ELAs, facilitate the better reconstruction of former climatic conditions and enable the derivation of a more accurate climatic ELA. Below are the methods used to calculate the topoclimatic factors presented within Chapter 6.

3.6.2 Avalanche areas

Avalanches provide additional accumulation to the surface of glaciers. The size and importance of this input varies depending on the catchment topography of individual glaciers. Glaciers in the Himalayas are known to gain considerable amounts of their accumulation through avalanching (Inoue, 1977; Benn and Lehmkuhl, 2000). Khumbu glacier, Nepal is thought to receive c.75% of its accumulation from avalanching and c.25% from snowfall (Inoue, 1977). The Ama Dablam Glacier, Nepal also receives a large proportion of its accumulation through avalanches, and the ELA is situated at the base of the avalanche cones. This glacier's ELA is much lower than the climatic ELA, and the neighbouring Chhukung glacier has a snout altitude c.200m above the ELA of the Ama Dablam glacier (Benn and Lehmkuhl, 2000). The Cairngorm Mountains are considerably smaller; however, it is worth considering that avalanching, particularly from leeward slopes heavily loaded by snow redistribution, may have been important to the mass balance of the glaciers and could account for some of the ELA variation. Where the snow accumulation is heavily focused within avalanche deposits and in the lee of mountain ridges, this could reduce the size of the required accumulation area, thus effectively increasing the altitude of the glacier ELA. In such scenarios, using the AABR method to reconstruct the former glacier ELA would be incorrect due to the assumption of linear ablation and accumulation gradients being violated. In this case the ABBR method would generate a lower ELA than formerly existed. However, the main consideration within the topography of the Cairngorm Mountains is that the focusing of avalanche deposits could have lowered the glacier ELA below the climatic ELA.
In order to assess the relative potential importance of avalanches on the reconstructed glaciers, the potential avalanche areas were calculated. It was decided to include areas that sloped towards the glacier surface with a slope of 20° or greater (as used by Sissons and Sutherland, 1976; Sutherland, 1984; Coleman et al., 2009). The 20° slope threshold used is shallower than the slopes on which most avalanches start (Schweizer et al., 2003). However, importantly, the inclusion of these slopes allows avalanches that began on higher steeper terrain to continue their track and runout over shallower terrain prior to reaching the glacier surface. Previous studies limited the avalanche area to include only the potential avalanche area above the reconstructed ELA (Sissons and Sutherland, 1976; Sutherland, 1984; Coleman et al., 2009). In reality, the accumulation on the glacier surface may be much more spatially variable, and avalanche accumulation may be important within the lower part of the glacier. Mass-balance studies indicate that often patterns of accumulation and ablation are spatially variable, and locations near the back and sidewalls of valley glaciers have greater accumulation (Rogerson, 1986; Hock and Holmgren, 2005), possibly from shading and/or avalanches. For this reason, the potential avalanche area both above and below the reconstructed ELA were digitised (Figure 3.10) and the areas are presented in Chapter 6. These areas were divided by the glacier size to generate avalanche ratios, which can be used to compare the relative importance of potential avalanche inputs.

Figure 3.10 Digitised avalanche areas both above and below the ELA
Another approach is to include the potential avalanche area within the calculation of the ELA. The inclusion of south-west snow-blown areas in the calculation of Drumochter glacier ELAs was undertaken by Benn and Ballantyne (2005), and the inclusion of the glacier headwall areas in Himalayan glacier ELAs was undertaken by Owen and Benn (2005). Given the avalanche areas are likely to contribute to the accumulation of the glacier, it seems appropriate to incorporate them into the ELA calculation. This only includes the potential avalanche areas above the originally reconstructed ELA. While slopes below the ELA adjacent to the glacier may provide accumulation through avalanching, the ELA calculation method would treat them as ablation areas, thus they have not been included. The ELAs are presented in Chapter 6, as are the differences between the glacier-only ELA and the glacier-plus-potential-avalanche-area ELA.

### 3.6.3 Solar radiation

Solar radiation is an important component of the energy balance of most glaciers (Hock, 2005) and local variations in incoming solar radiation can have important impacts on the ELAs of glaciers in close proximity (Rogerson, 1986; Holmlund and Jansson, 1999; Cossart, 2011). The incoming components of radiation can be subdivided into shortwave (direct, diffuse and reflected from the terrain) and longwave radiation (sky and emitted from the surrounding terrain) (Hock, 2005). The use of GIS allows the computation of the relative incoming direct, diffuse and total radiation (direct plus diffuse) values for a given DEM. Note the commonly used term, global radiation, includes reflected radiation which is not included within this modelling, thus the term total radiation is used. The calculated incoming radiation values vary spatially based on the shading by local topography and also the incidence angle of the surface’s slope and aspect with the incoming radiation. The solar radiation modelling has been undertaken using a 10m resolution present-day DEM and the 10m resolution DEMs, including the valley- and plateau-sourced reconstructed glacier surfaces. To generate the later DEMs, the glacier surface was created by interpolation between the glacier contours and margins; this was then combined with the present-day DEM. The modelling utilised the Area Solar Radiation Toolbox within ArcGIS based on the methods of Rich et al. (1994), Rich and Fu (2000) and Fu and Rich (2000, 2002), which generates raster files of the
direct, diffuse and total radiation in watt hours per square metre. Examples of the solar radiation output can be found in Chapter 6.

Modern-day orbits and solar activity were used, whereas Milankovitch variation in obliquity and eccentricity would have been different during the Younger Dryas (Berger, 1978; Berger et al., 1981). Based on the use of only relative comparisons between former glaciers this is assumed to be unimportant. The modelling was run from the 1st May to 30th September to identify important spatial variations in radiation during the ablation season (Sutherland, 1984). The modelling was set up using a sky size of 1000 to balance accuracy with computational resources, a seven-day interval to account for the variation in the sun's track through the sky during the ablation season, and a 0.5-hour interval for the sun's track position within each of the days. The calculation of the slope and aspect of the surface was derived from the specified DEM. A horizon value of 32 angles was used for the viewed calculation of the sunmap; this is suitable for complex topography (Fu and Rich, 1999), with eight zenith and azimuth divisions for the skymap, which is suitable for most terrains (Fu and Rich, 1999). The diffuse radiation was modelled using the default uniform sky model, in which incoming diffuse radiation is the same from all sky directions (Fu and Rich, 1999). A generally clear-sky scenario was run with the default values of 0.3 for the diffuse proportion and 0.5 for the transmissivity. A second partially cloudy scenario was generated to represent Aviemore’s present-day climate with a diffuse to global radiation ratio of 0.55–0.65 from May to October 1995–2000 (Aviemore Weather Station – available through MIDAS). To simulate these conditions a diffuse portion of 0.6 and a transmissivity of 0.3 was applied. The transmissivity parameter has an inverse relation with the diffuse proportion (Fu and Rich, 1999). The nature of the GIS toolbox means both the transmissivity and diffuse proportion are changed to simulate differences in cloud cover. Thus it is not possible to isolate the diffuse proportion in order to facilitate direct comparisons between the clear-sky and partially cloudy scenarios. The untested accuracy of the diffuse and transmissivity values applied means that the datasets are not suitable for comparison with other modelled or measured datasets. This is not an issue for the purpose of comparing relative radiation values between different glaciers within a modelled scenario.
The raster outputs were then used to extract mean radiation values for the ablation areas of the reconstructed glaciers. The ablation area was chosen because it is the area of mass loss and the lower albedos of the exposed ice cause a greater percentage of the radiation to be absorbed compared to the accumulation area where higher albedo snow exists throughout the summer (Paterson, 1994; Benn and Evans, 2010). Further details of the method and analysis can be found in Chapter 6: Section 6.3.2.2.

3.6.4 Snow redistribution

3.6.4.1 Background and previous approaches

There have been numerous different approaches to the issue of snow redistribution and its impact on former glacier ELAs and the derived palaeoclimatic information (Sissons and Sutherland, 1976; Mitchell, 1996; Dahl et al., 1997; Purves et al., 1999; Benn and Ballantyne, 2005; Coleman et al., 2009; Bendle and Glasser, 2012). Sissons and Sutherland (1976) defined the snow-blown area as ground lying above the accumulation area that was sloping down towards the glacier surface or potential avalanche area. Dahl et al. (1997) suggested that, as they were working in steep mountainous areas, the upward transport of snow was minimal. Thus the ratio between the present drainage area above the reconstructed ELA and the reconstructed glacier accumulation area was used; this is effectively a similar approach to that of Sissons and Sutherland (1976). However, this approach is most likely unsuitable for the Cairngorm Mountains which are characterised by their large rounded plateaus.

The approach of Sissons and Sutherland (1976) was modified to include uphill movement of snow up to a maximum of 10° on ground lying above the ELA and continuous to the glacier surface (Robertson 1988; Mitchell, 1996; Coleman et al., 2009). Typically these areas have then been divided into 16 segments centred on the intersection of the ELA and glacier centreline (Coleman et al., 2009); these can then be combined to generate areas for different snow-blown directions. This approach is suitable for small circular corrie glaciers; however, it becomes problematic for larger and complex shaped glaciers, as snow can often reach the glacier surface from multiple wind directions.
Benn and Ballantyne (2005), and later Ballantyne (2007a), and Bendle and Glasser (2012) generated south-west (180–270) snow-blow quadrants. These quadrants included areas that were upslope and upwind of the glacier surface, including plateau, all glacier-facing slopes, and all other plateau-edge slopes with gradients of less than 5° irrespective of orientation. They suggested the inclusion of slopes less than 5° was conservative. Using a specific south-west quadrant based on the prevailing wind direction is beneficial when working with a mixture of different sized and shaped glaciers, as any snow sources that could reach the glacier or avalanche area through a south or west wind direction are included.

Many authors have generated ratios and factors to normalise the snow-blow areas to their respective glaciers (Sissons, 1980a, 1980b; Mitchell, 1996; Coleman et al., 2009; Bendle and Glasser, 2012). Thus dividing the snow-blow area by the glacier area generates a snow-blow ratio. Sissons (1980b) noted that because of the elongated shape of most snow-blow areas, the area further away from the glacier had less potential to supply snow to the glacier, thus the ratio was square rooted to create a snow-blow factor. This has been adopted by other authors since, often with both ratios and factors being presented (Mitchell, 1996; Coleman et al., 2009). The snow-blow ratios have previously been combined with avalanche ratios and plotted against the glacier ELA to identify if any of the variation in ELA can be attributed to the ratios (Bendle and Glasser, 2012).

A different approach for analysing snow redistribution has emerged through the use of cell-based models within GIS. This allows an initial snow depth to be redistributed based on the speed and direction of wind over a DEM (Purves et al., 1998, 1999). The models are based on the transportation of snow between cells when the wind speed exceeds the snow-transport threshold, thus eroding snow from plateaus and ridges and depositing in sheltered locations. The modelling is suitable for relative measures and distribution patterns, but often uses arbitrary values and thus is not fully quantitative (Purves et al., 1998, 1999). There are numerous complicating factors that impact on the amount of snow and the threshold at which snow is transported: these include the availability of snow for transportation, temperature and humidity at which the snow was deposited, particle size and whether melt–freeze cycles have occurred (Purves et al., 1998). This modelling approach has been previously applied to the Cairngorm topography for
south, south-west and south-easterly wind directions and a ‘cookie-cutting’ technique used to extract values for the accumulation zones of the previously proposed Younger Dryas glaciers (Purves et al., 1999). The redistributed snow depths were used to make relative comparisons between Younger Dryas glaciated and unglaciated sites (Purves et al., 1999). Similar modelling of the spatial redistribution of snow within the Cairngorms has been undertaken by Treglia (2011) and more recently in relation to possible Little Ice Age glaciers (Harrison et al., 2014).

The approach was taken to use both the manual digitisation of snow-blow areas and a cell-based GIS model to investigate the redistribution of snow within the Cairngorm Mountains. The methods used are similar to those that have been undertaken previously but have some important differences; these are described below.

3.6.4.2 Manual digitisation method

A conventional approach was used for the calculation of snow-blow areas by manual digitisation. The glacier area above the ELA and the corresponding avalanche area (i.e. the adjacent terrain with slopes greater than 20°, as described in section 3.6.2) were used as the target area for potential snow to reach. The terrain within the south-west (180–270) quadrant was included, as long as it had slopes of no more than 15° with aspects facing away from the glacier. This allowed the inclusion of slopes facing towards the glacier and plateaus, slopes adjacent to the corrie or valley, and slopes that faced away from the glacier as long as they were not steeper than 15° (Figure 3.11). This style of the approach was employed by Benn and Ballantyne (2005) and Bendle and Glasser (2012). It was decided to increase the conservative 5° slope threshold used by Benn and Ballantyne (2005) and the 10° slope used by Coleman et al. (2009) to 15°; however, this figure remains arbitrary and the use of such a threshold is simplistic. This increase is supported by observations from Greenland and Svalbard that, provided there is sufficient wind speed, snow transport uphill is efficient, even on steep slopes, and the notion that uphill snow transport has often been underestimated (Humlum, 2002).
Chapter 3: Methodology

3.6.4.3 GIS cell-based snow-redistribution model

This cell-based GIS approach is modified from the work of Purves et al. (1998, 1999) and Treglia (2011). The modelling can be broken into three stages: wind flow over the terrain, snow redistribution, and extraction of snow depths.

For the first stage, the freely available Wind Ninja 2.2.0 software was utilised. It is a mass-consistent model and is designed for modelling forest fires over complex terrain (Forthofer, 2007; Forthofer and Butler, 2007). It was used to modify the wind speed and direction over the Cairngorm Mountains; the surrounding areas were also included to avoid the boundary of the DEM impacting on the results. The model has been validated against measured wind speeds over the Askervein Hill, Scotland; this included sensitivity analysis, such as the effect of different resolutions between 23.5 and 120m (Forthofer, 2007). The model generated wind speeds within 15% of the measured values, except in some leeside hill positions where errors up to 150% were found, the worst being a measured value of 3.2m/s and a simulated value of 8m/s (Forthofer, 2007). The wind speeds mimic the trend of the measured values, although some leeside speeds may be overestimated. Visual inspection of the modelled wind-speed outputs in GIS showed that wind speeds were reduced within the corries and valleys, particularly on the leeside of sharp ridges. Thus while it may be overestimating the leeside values, the pattern mimics the expected higher wind speeds on plateaus and ridges and lower speeds within corries and valleys, thus the model results are likely to be sufficient for our purpose. Computational fluid dynamics (CFD) models are better at producing more accurate
wind speeds but they are more computationally expensive and are not freely available (Forthofer, 2007).

The Wind Ninja modelling was performed at a 30m resolution on a high-powered computer. The average wind speed for the domain was set at 15m/s at 5m above grassy terrain, generating wind speeds of over 20m/s on some of the higher summits and ridges (Figure 3.12). Although realistic, these are arbitrary values and are only important relative to the wind-speed threshold at which snow is deemed to be transported. The wind directions were varied from 135° to 270° at 45° degree intervals to simulate likely prevailing wind directions from the south-east, south, south-west and west. The model was run for present-day terrain, reconstructed valley glacier and reconstructed plateau-glacier terrain, generating three sets of point data for the wind speeds and directions.

The second stage of the model uses python script within an ArcGIS toolbox to redistribute the snow. The python script is based on that of Treglia (2011) but has been heavily modified. First the wind-speed point data was interpolated to create a 30m resolution wind-speed raster; the wind direction was kept uniform to save computational requirements. The toolbox uses the wind-speed raster directly and no further alterations for shelter indexes were applied. The python-scripted toolbox redistributed the snow according to the specified wind direction when the wind speed within a cell was above the arbitrary wind-speed snow-transport threshold. When the threshold was exceeded, all the snow within the cell was moved to the adjacent cell in the specified wind direction. This is simplistic; however, it avoids complications of how much snow to move, which will depend on multiple factors including the length of exposure, snow wetness, and melt-freeze cycles. When all the cells above the snow-transport threshold (source cells) have been redistributed into cells below the threshold (deposition cells) the model is complete. This created an output of snow redistribution, with high values representing snow deposition and zero values in areas where snow has been removed. This approach is sensitive to the snow-transport threshold, thus multiple thresholds have been used: 10, 12 and 14m/s. These were not chosen to represent the threshold at which snow is entrained or deposited by wind; instead they are arbitrary values that for the input wind speed generated a high, medium and low snow-transport scenario (Figure 3.13).
The output of the model was analysed visually to identify areas of accumulation and deflation, and was also used to generate snow-blow ratios. To do this, snow values that were on the full reconstructed glacier surface (accumulation and ablation areas) or within the adjacent avalanche area were included. The avalanche area was automatically calculated with the use of toolboxes within ArcGIS. It included areas within the watershed of the glacier, with slopes greater than 20° that directly abut the glacier surface. The sum of the snow values within the full glacier and avalanche area were then divided by the glacier area to create a combined snow-blow and avalanche ratio.

Despite modelling snow towards the glacier, instead of working from the glacier backwards, this approach has many similarities with the conventional method of calculating the snow-blow area by digitisation. One of the important differences is the lack of an arbitrary slope-angle threshold. This allows larger areas to be included that have greater slope angles, as long as the wind speed exceeds the snow-transport threshold. Often the wind speeds were reduced near steep slopes, so the unrealistic transportation of snow up very steep slopes is avoided. No limit has been imposed on the distance snow can be transported, unlike other studies that have included a maximum transport distance (e.g. Harrison *et al*., 2014). Thus a large c.3250km² DEM was used for the modelling to ensure snow from the terrain to the south and west of the Cairngorms could be redistributed to the glaciers, provided the balance between the wind-speed and snow-transport threshold permitted. This was important to ensure the size of the DEM did not restrict the transportation of snow to the depositional areas; instead locations where the snow-transport threshold was not exceeded acted as natural boundaries.
Figure 3.12 Modelled wind speed at 5m above the present-day DEM using a south-west wind direction
Figure 3.13 Low (top), medium (middle) and high (bottom) snow-transport scenarios for the south-west wind direction
3.7 Chapter Summary

This chapter has presented the methods applied within the thesis. For clarity, some of the specific details are included or described further within the appropriate analysis chapters. The methods have evolved throughout the project to provide meaningful analysis with which to answer the research objectives. The holistic mapping approach will ensure all evidence is considered and utilised in reconstructing retreat patterns and identifying locations of glacier readvance through the landsystems approach. It will also provide robust landform interpretations for the new cosmogenic surface exposure ages and existing geochronology studies. The new cosmogenic surface exposure samples have been taken from contentious sites in order to understand whether Younger Dryas valley glaciers existed. Samples have been collected from embedded boulders of differing heights above the moraine surface in order to use inverse modelling techniques to better understand the impact of moraine degradation on the boulder ages. The reconstruction of the former glacier surfaces, ELAs and precipitation values use well-established techniques. However, innovative GIS analysis of topoclimatic factors such as avalanches, wind redistribution of snow and solar radiation has been undertaken. These factors are combined using statistical analysis to better understand precipitation gradients and climatic ELAs in Chapter 6. The methods used build on each other throughout the thesis to develop our understanding of the geomorphology, geochronology and palaeoclimatic inferences from the reconstructed glaciers. At all stages, areas of uncertainty are acknowledged and, where possible, explored with further research. Given the nature of the work, uncertainties are often cumulative from one stage to the next; where these are quantifiable they have been included, such as when deriving palaeoprecipitation values. The analysis of methods used in previous studies has been crucial in order to endeavour to improve the methods described within this chapter, and thus the quality of the data that is presented in the next three chapters.
4 The Pattern and Style of Retreat and Readvances

4.1 Introduction

This chapter presents the mapping results that concern the pattern of ice-sheet deglaciation, the interaction of local and external ice masses, and the extent and style of any later readvances. Where evidence permits, a relative chronology is stated but attribution to specific climatic events is presented in Chapter 6. The pattern and style of glacial retreat has been assessed through the mapping of glacial landforms and principles of morphostratigraphy described in the methodology (Chapter 3: Section 3.2). Previous work and interpretations made important contributions to the new mapping presented below (and are reviewed in Chapter 2) but the work has been omitted here for clarity, except where more detailed work assists in explaining the geomorphology. The findings have been arranged primarily by their geographical location, and secondly by the relative stage of deglaciation. The landform morphology and any sediment analysis are stated first; this is followed by interpretation or, if insufficient evidence is available to support one interpretation, multiple alternative hypotheses are discussed. The final presentation and discussion of contemporaneous margins and attribution to climatic events are presented in Chapter 6. The location of the map extracts presented throughout this chapter can be seen on Figure 4.1.1. The maps follow the symbology presented in Figure 4.1.2 and the landforms and features defined in Chapter 3 (Table 3.1). Scale details and north arrows can be found on the individual maps.
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Figure 4.1.1 Location map for geomorphological mapping figures for eastern (red), central (blue) and western Cairngorms (black)
4.2 Eastern Cairngorms

4.2.1 Deeside, River Gairn and Glen Builg

On the southernmost slopes of the Cairngorms that bound the main Dee valley, large meltwater channels and moraines modified by meltwater have been identified. In the south-east Cairngorms, both within the Gairn valley and on the flanks of the Carn Liath and Culardoch mountains, moraines and lateral meltwater channels cover many of the lower cols and valley sides. The cols contain glacial drift and moraines heavily shaped by meltwater channels; these have no local ice source and represent northwards ice flow from the main Dee valley to the south (Figure 4.2.1). At the col between the Carn Liath and Culardoch mountains, the highest
meltwater channels in the main Dee valley flow north over the col towards the River Gairn, whereas below the col the channels continue down-valley within the main Dee valley.

**Figure 4.2.1 Parallel meltwater channels cut into glacial drift – upper Gairn valley**

This evidence is interpreted to mark the presence of the Dee glacier sourced from the south and west, which breached cols and flowed north into valleys such as Glen Gairn, continuing northwards towards Loch Builg (Figure 4.1.1). Dee ice was able to overtop this lower eastern topography, both because it is a lower physical barrier and the topography is not high enough to generate substantial counteractive local ice flow. The deglaciation of this south-eastern corner of the Cairngorms and cessation of ice from Deeside overtopping into Gleann an t-Slugain, Gairn and towards Loch Builg is likely to have occurred relatively early in deglaciation, as later Gleann an t-Slugain would have been ice free when acting as an overflow channel for the ice-dammed lake in Glen Quoich, discussed in Section 4.2.4.

**4.2.2 Slochd Mòr and Ben Avon valleys and corries**

There are several valleys and corries which drain the eastern plateau slopes of Ben Avon; although, in general, the topography does not possess that same glacial trough erosion and modification as seen further west in the Cairngorms.
4.2.2.1 Slochd Mòr and Glen Avon

Figure 4.2.2 Slochd Mòr geomorphological map.
Slochd Mòr has a large diversity of glacial landforms (Figure 4.2.2). At the lower northern end, an area of ridges and large flat-topped terraces has been identified (Figure 4.2.3). The flat-topped terraces gently slope down-valley from south to north and stand in excess of 20m above the incised channels, some of which support streams today. The terraces on the east side of the valley are at a slightly higher altitude than those on the west side of the central channel.

The formation of terraces across the central valley floor suggests they are not kame terraces. Another formation hypothesis is that the terraces may be kames: these form when sediment accumulates on or within the ice and inverts when the supporting ice melts out (Johnson and Menzies, 2002). However, in Slochd Mòr the flat-topped surfaces are separated by deeply eroded channels, which suggests the surface may have been more continuous prior to channel incision. Such a surface may have been formed by glacial outwash. While kettle holes can often be found within outwash surfaces, these have not been identified in Slochd Mòr. The final
formation method is that the terraces were formed as a delta with sediment deposition in a lake environment. However, these last two hypotheses cannot easily explain the variation in height of landforms, unless there were different sediment sources, glacier margins or lake levels involved. The favoured formation method is the terraces formed either as an outwash terrace or a delta surface. At a later stage fluvial incision dissected the surfaces leaving the landscape present today.

There has been the suggestion that the Loch Builg glacier may have dammed lower Glen Avon forming an ice-dammed lake (Kirkbride and Gordon, 2010). Mapping from aerial imagery supports this idea. Meltwater channels within Glen Builg record ice flow from the south, northwards towards the exit of Glen Avon. To the west, 2.5-3.5km up-valley into lower Glen Avon, there are numerous large ridges interpreted to be moraines, their orientation suggests they are from the invading Builg glacier. There are numerous flat-topped, candidate delta formations within the adjacent valley to the south, further up Glen Avon and most noticeably the Slochd Mòr terraces described above. This remote mapping work provides limited support for the notion of an ice-dammed lake in this area; however, detailed fieldwork including sedimentology is required.

Up-valley in Slochd Mòr, to the south of the terraces, more irregular glacial drift and moraines exist; these are dissected by the same channels as the terraces down-valley. This suggests the channels were formed at a later time or were formed contemporaneously with the glacier that formed the area of irregular moraines. These deposits continue up-valley with no major well-defined former ice margins. The deposits stop 2km short of the valley head where the valley sides are heavily dissected and alluvial-fan formation may mask glacial evidence.

From the valley head, a distinctive spread of highly concentrated glacially transported boulders exists that terminate just over 1km down-valley. The frontal limit continues as a lateral limit on the northern side and joins with a lateral moraine higher up on the northern side of the valley. The boulder limits cannot be traced on the southern side as a result of erosion and deposition from postglacial streams and alluvial fans. The moraine, boulder ridges and concentration of boulders are morphologically separate to the irregular drift higher on the valley side and further down the valley. No soliflucted surfaces occur within this upper glacier limit, yet with the exception of small areas, solifluction does not feature heavily within the lower
valley, instead the valley sides are dominated by gullied drift. Above the head of Slochd Mòr is the Cnap a’ Chlèirich plateau; this is covered in thin regolith, wind-scoured surfaces, limited solifluction and isolated candidate meltwater channels. Thus contributing ice from the plateau either in earlier or later stages of glaciation is plausible.

The probable deglaciation sequence in Slochd Mòr is that the lowermost moraines and meltwater channels formed first during ice retreat. Later outwash or delta formation created the large terrace and sediment accumulations that occupy the valley floor, prior to incision of the terraces by meltwater most probably when a glacier occupied the area of unorganised glacial drift. More detailed fieldwork of the lower Slochd Mòr deposits and landforms within Glen Avon would be beneficial to understand this phase of ice-sheet deglaciation. There was later a readvance of a smaller glacier in the Slochd Mòr valley head.
4.2.2.2 Ben Avon northern valleys

The corrie that holds Lochan nan Gabhar has distinctive glacial drift, boulder accumulations and moraines that extend c.700m from the backwall; these are
morphologically separate from the more widespread drift down-valley (Figure 4.2.4 and Figure 4.2.5). Within this area, arcuate moraine ridges have been mapped and a higher concentration of glacially transported boulders is present. The crests of the moraines indicate ice flow from the southern side sourced within the corrie. The plateau slopes above the corrie are covered in well-developed solifluction lobes. A plateau area drains into the corrie on the south-east side and aerial images indicate channels of potential meltwater origin in this area, thus some ice may have been sourced from this area but evidence to support or refute this is limited. This evidence suggests a later locally sourced glacier within the corrie.

Figure 4.2.5 Lochan nan Gabhar – note the moraine ridges and glacially transported boulders extending from the corrie

To the immediate west just outside the corrie moraine limits, are a pair of subtle ridges on the eastern side of the neighbouring valley. These may indicate a retreat stage of a valley- or plateau-fed glacier in the valley immediately to the west of the corrie. The lack of other evidence in the valley hinders further interpretation; however, the presence of solifluction on the surrounding plateau-source area and on the neighbouring glacial drift suggests glaciation of this valley predates the corrie glaciation to the east.

To the east is the valley head that contains the Meur Gorm stream; no clear evidence exists here for a period of local glaciation. This may be due to its lower corrie-floor altitude.
4.2.2.3 Eastern Ben Avon

The corrie floor below Sgor Riabhach Crag has a thick accumulation of glacial deposits within it (Figure 4.2.4). There are subtle arcuate moraine ridges within the deposits and faint meltwater channels are located amongst the moraines on the northern side (Figure 4.2.6). The moraines are studded with glacially transported boulders. The higher plateau surfaces surrounding the corrie are covered in extensive solifluction, and the valley side adjacent to the corrie moraines shows subtle solifluction which is not present inside the moraine margins. This indicates a later glacier existed within the corrie, most likely sourced from within the corrie.

To the north, the corrie/valley head to the east of East Meur Gorm Craig has a thick accumulation of sediment within it. However, other glacial evidence such as moraine ridges, meltwater channels and glacially transported boulders are not present and there is no obvious source area for a later local glacier.

4.2.2.4 Southern Ben Avon

To the south the Allt an Eas Mhòir and Allt an Eas Bhig valleys have glacial drift and subtle ridges that imply ice flow from the up-valley direction (Figure 4.2.4). These compare with the larger mounds of drift and meltwater channels that are associated with externally sourced ice-sheet retreat in the Gairn valley below. Up-valley of these deposits on the valley sides and plateau are extensive areas of soliflucted surfaces. Whether these landforms reflect local or externally sourced ice
is unclear; however, given the presence of solifluction, these margins are unlikely to be from a later phase of glaciation.

4.2.2.5 Plateau of Ben Avon

The Ben Avon plateau surface is dominated by periglacial activity (Figure 4.2.4). Much of the summit surface and upper plateau are covered in tors and blockfields and most of the slopes are covered in well-developed solifluction surfaces and lobes. However, candidate meltwater channels have been identified on cols and other areas of relatively flat topography. In addition, small areas of drift exist where the irregular nature of the surfaces and the oblique channels indicate a glacial origin. However, these features are isolated and not conclusive, thus gathering evidence for glacial reconstructions is problematic. The prevailing conclusion is that the eastern plateau became deglaciated early and is unlikely to have been covered by a later readvance of ice. However, thin areas of isolated ice or cold-based ice cannot be eliminated.
4.2.3 Corries of Beinn a’ Bhuird

Figure 4.2.7 Geomorphological map of the corries of Beinn a’ Bhuird

Coire na Ciche on the east side of the South Top of Beinn a’ Bhuird contains a well-defined arcuate moraine ridge covered with large boulders that spans the width of the corrie (Figure 4.2.7). Immediately behind the outer ridge, a second less well-defined ridge exists with infilling by fluvial deposition behind the ridges. Outside the moraine limits, morphologically separate gullied glacial drift exists with small areas of solifluction. This suggests the distinctive corrie moraines represent a separate later readvance of a glacier.

Further north, on the east side of Beinn a’ Bhuird, the two well-developed corries of Coire an Dubh-lochan and Coire nan Clach converge and descend before the valley turns south towards Glen Quoich (Figure 4.2.7). Both corrie floors are covered by a high concentration of large glacially transported granite boulders.
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(Figure 4.2.8); these stop abruptly at the point where the valley begins to turn south. This coincides with a small subtle moraine ridge, where the corrie streams are forced to converge (Figure 4.2.7). Within this limit there is a high concentration of boulders, some forming lines parallel to ice flow. It is possible the central curving line of boulders marks a medial moraine formed where the two ice sources converged. Most boulders are relatively angular; this may be due to being transported supraglacially or due to the relatively short transportation distances. In the area of the central and southern lake small moraines can be found that mark stages of retreat. Outside the boulder limit there is a distinct change to older drift that has suffered solifluction processes. This is strong evidence that the glacier that occupied the corries was formed later during a separate readvance.

Figure 4.2.8 View looking towards Coire an Dubh-lochain. Note the high concentration of glacially transported boulders

To the north a small shallow unnamed corrie exists on the eastern side of Cnap a’ Chlèirich (Figure 4.2.7). At the lip of the shallow corrie drift deposits occur which have been incised by channel erosion. The deposits are tentatively interpreted here as moraines trend down-valley originating from a glacier sourced to the west. Whether the glacier was sourced from the corrie or from the plateau above is uncertain and the age of these deposits is also unclear.
4.2.4 Quoich and Dubh-Ghleann

4.2.4.1 Central Glen Quoich and Gleann an t-Slugain

At the separation of Gleann an t-Slugain and Glen Quoich a large assemblage of glacial, glaciofluvial and glaciolacustrine landforms exist (Figure 4.2.9). On the eastern side of Glen Quoich, moraine ridges and meltwater channels descend into Glen Gairn and Gleann an t-Slugain. These mark the retreat of the Quoich glacier from a time when the Quoich glacier flowed east into Glen Gairn, where it merged with ice overtopping the southerly col from the main Dee valley.

On the western side of Glen Quoich, a rocky ridge with a low gradient gently slopes down-valley. This is interpreted to be a lateral moraine and its shallow gradient is supportive of the interpretation below of a water-terminating glacier margin. At the southern outermost limit of the moraine area, the moraines curve down-valley suggesting ice flow from the north. The area is covered in large granite boulders suggesting locally sourced glaciation. Just inside the outermost moraines a large flat-topped deposit can be found on the western side of the valley (Figure 4.2.10). The surface of this feature is c.600m but consists of two levels: the higher approximately 3-4m higher than the lower (Figure 4.2.11). The eastern side slopes at c.22 degrees and is a continuous curved face with no split level, whereas the southern side slopes at c.20 degrees and possesses both the upper and lower more gently sloping terrace surfaces. Based on the position of the feature at an altitude of 600m, with evidence of a lake at this height elsewhere in the valley, and the double terrace flat upper surfaces being typical of a change in lake height, this feature has been interpreted as a delta. On the eastern side, the delta side has a continuous curved face with no indication of lake level change and thus is interpreted as being the ice-contact slope. There are also moraine banks that continue towards the valley centre from the ice-contact slope, indicating one continuous curved margin (Figure 4.2.10 and Figure 4.2.12). The glacier margin has been cut through by a meltwater channel between the delta and the moraine banks. The size and clarity of the delta would tend to indicate it may have been fed by a single stable meltwater stream (Ashley, 2002). Given the double level of the southern face, this is interpreted to be the lake-terminating side. Based on its steepness, this delta is likely to have formed in a high energy environment with coarse grained material in its make-up, whereas, in a lower energy environment it
would be expected to observe a steep side on the ice-contact face and a shallower side on the lake side (Ashley, 2002). The double terrace surface on the southern side indicates a rise in the lake level during delta formation (Bennett and Glasser, 2009) (Figure 4.2.11 and Figure 4.2.13). The formation of an ice-contact delta and the associated moraines suggest the glacier that formed these was stable and not a highly dynamic calving margin, consistent with a shallow lake depth in this area. Altogether this margin, based on a large assemblage of meltwater channels, moraines and an ice-contact delta, suggests the presence of a glacier fed from the corries/plateau of Beinn a’ Bhuird to the north.
Figure 4.2.9 Glen Quoich geomorphological map

(© Crown Copyright/database right 2014. An Ordnance Survey/EDINA supplied service)
Figure 4.2.10 Glen Quoich glacier margin. Note flat-topped delta levels, moraines and meltwater channel

Figure 4.2.11 Profile of the lakeside delta face. Note the two terrace surfaces marking a change in lake level. In background overflow channel over col into Gleann an t-Slugain
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Figure 4.2.12 Close-up of moraine profile adjacent to delta margin, up-valley and ice flow from the left, Glen Quoich

Figure 4.2.13 Schematic delta formation responses from rising and lowering lake levels (adapted from Bennett and Glasser (2009) with permission of Wiley)

Approximately 300-400m up-valley, a 200m broad shallow ridge crosses some 900m across the valley floor (Figure 4.2.9 and Figure 4.2.14). The ridge has been dissected on the eastern side by Quoich Water. Palaeochannels in this area suggest Quoich Water took multiple routes across the moraine prior to incising the present-day river channel. A continuous ridge, like that seen, can be formed by moraine delta-building, where deltas and fans coalesce if the margin is stationary for long enough (Ashley, 2002). This deposit is thought to mark a later stage of glacial retreat and fits with the notion of the glacier retreating to a grounding line (Kirkbride and Gordon, 2010). Despite its position up-valley, the moraine is just below the altitude of the delta surface and is consistent with a lake level of c.600m. Up-valley from this location there are no major ice margins until those associated
with the readvance of ice from the Beinn a’ Bhuidr corries, described in Section 4.2.3.

At the head of Gleann an t-Slugain, where it divides from Glen Quoich, a steep-sided channel exists with no present-day water source (Figure 4.2.15); although partially formed by lateral meltwater channels from the Glen Quoich glacier, it is suggested here it was deepened when acting as a spillway for the adjacent lake in Glen Quoich. It is suggested the flat ground, now covered by peat, acted as a spillway and the lake water drained through the channel down Gleen an t-Slugain.
4.2.4.2 Poll Bhàt

A large triangular deposit that supports the Poll Bhàt lake is formed on the western side of Dubh-Ghleann at c.600m (Figure 4.2.9 and Figure 4.2.16). The deposit surface dips gently east towards Dubh-Ghleann before it steeply descends into the valley (Figure 4.2.17). Linked to the deposit is a large meltwater channel that begins at the col with Glen Derry to the west and abuts the upper end of the deposit (Figure 4.2.9). Sediment logging of a naturally occurring stream cut exposure in the central section of the deposit revealed a clast-supported sediment comprising of angular clasts with a-axis up to 45cm in length, which were found dipping at 18–30 degrees towards the north-eastern deposit edge (Figure 4.2.18). A second pit closer to the deposit edge revealed a similar clast-supported sediment with angular clasts of up to 25cm, with unconsolidated finer material infilling the inter-clast spaces. Some similar dipping of clasts towards the deposit edge was recorded here. The morphology of the deposit and the presence of the feeder meltwater channel suggest the landform is of deltaic origin. The sediment logging, although not conclusive, is consistent with dipping foresets formed in a lake environment when sediment avalanches down the face of the delta (Bennett and Glasser, 2009). The delta surface exists at two levels, the outer edges being c.4–5m higher than the central section, indicating the lake level dropped during formation (Figure 4.2.16 and Figure 4.2.19) (Bennett and Glasser, 2009). The angular nature, lithology and dipping recorded in the exposures are consistent with the formation of a delta protruding into a lake, with the majority of the material being transported only a short distance by meltwater erosion of the gorge above (Figure 4.2.20). The steep outer edge of the delta and steep angles of sediment within the exposure indicate the delta was formed in a relatively high energy environment and is consistent with relatively coarse material being eroded by meltwater travelling down the from the col with Glen Derry. The source of this water in Glen Derry is discussed in Section 4.3.1.
Figure 4.2.16 Poll Bhàt delta deposit and Dubh-Ghleann and Glen Quoich in the background. Note the central section of the delta is lower, due to the down cutting from a lowering lake level.

Figure 4.2.17 Poll Bhàt delta deposit – flat top with steep apex to Dubh-Ghleann below.
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Figure 4.2.18 Poll Bhàt delta sediment section. Delta edge is to the left of the photo

Figure 4.2.19 Generalised diagram of the Poll Bhàt deposit and the likely lowering lake level that formed it
4.2.4.3 Lower Glen Quoich and Clais Fhearnaig

Multiple shallow bench-like fragments have been identified between 570–600m in the central area of Glen Quoich (Figure 4.2.9): these are most prominent on the southern slopes of Carn Allt nà Beinne (Figure 4.2.21) and the northern slopes of Carn Elrig Mòr (Figure 4.2.22). The features appear as faint horizontal lines on the valley side and are best illuminated by shallow sunlight; closer investigation showed they are shallow changes in the slope profile. The discontinuity of the fragments is likely to be due to postglacial slope processes including solifluction. The morphology of these benches, continuity on opposing sides of the valley and their altitudinal link with the deltas described above suggests they are lake shorelines. Together this evidence indicates the presence of a large ice-dammed lake within Glen Quoich.
Down-valley in Glen Quoich, both valley sides are covered with ice-marginal deposits heavily shaped by meltwater channels (Figure 4.2.9). On the western side, north of the main Clais Fhearnaig channel, multiple meltwater channels flow obliquely across the col from Glen Derry dipping towards Glen Quoich (Figure 4.2.23). The channels begin as shallow benches but develop into full channels with banks on both sides. The nature, particularly the one-sided inlets/steps to the channels, suggest these are lateral meltwater channels which existed when ice flowed east through the Clais Fhearnaig passage from Glen Lui to Glen Quoich. These channels start at approximately 670m, noticeably just a few metres lower than the col that feeds Poll Bhàt located 2km further north. This starting altitude, the channels one-sided cross-sectional profile at their inception and their link with lake shorelines further north (described in Section 4.3.1) suggest these channels may have become the new outflow from the Derry lake as the damming ice lowered during deglaciation. Draining east, at their lower end they end abruptly as they meet the valley side of Glen Quoich; here their termination is associated with large,
rounded deposits which also end abruptly as the hillside meets Glen Quoich. The abrupt nature of their termination and the associated large deposits may suggest they terminated in a lake environment or against a glacier.

Figure 4.2.23 Meltwater channel dipping towards Glen Quoich

On the opposite side of Glen Quoich, the slopes of Carn na Criche and Carn Elrig possess two distinguishable sets of glacial deposits and meltwater channels. On the slopes of Carn na Criche above c.630m, meltwater channels can be seen dipping northwards in towards the Cairngorms (Figure 4.2.9 and Figure 4.2.24). These channels start high on the valley sides of the main Dee valley and continue for 2.5 km towards the intersection of Glen Quoich and Gleann an t-Slugain. The channels end as they descend down the valley side into Glen Quoich. Assuming these channels are lateral to the former ice mass, they mark the ice gradient at the time of formation. Thus early in deglaciation external Deeside ice sloped towards the central area of Glen Quoich and may have joined with locally sourced ice from Beinn a’ Bhuird. Meltwater channels from the same high ice mass in the main Dee valley can be seen overtopping the Carn na Criche ridge at c.750m and flowing north downslope into Gleann an t-Slugain. This would suggest Gleann an t-Slugain was at least part deglaciated at this time.
Conversely, on the lower slopes of Carn na Criche and Carn Elrig Beag, meltwater channels dip to the south out from the Cairngorms (Figure 4.2.9). One of the main channels starts on the slopes of Carn Elrig Beag at 568m; it then cuts through the col between Carn Elrig Mòr and Carn Elrig, and then through the large shallow deposits to the east as it descends towards the floor of Glen Quoich. Further channels start at a similar altitude and location on the slopes of Carn Mòr, but descend more steeply on the western slopes of Carn Elrig Beag. Three formation hypotheses exist: (i) the channels are lateral in origin and formed by a locally sourced Cairngorm glacier; (ii) the channels formed subglacially underneath an ice mass from the main Dee valley that invaded lower Glen Quoich; (iii) the channels were subglacial or lateral draining around ice sourced from the west through the Clais Fhearnaig and the lower parts of the neighbouring ridge. The first hypothesis effectively explains the slope of the channels, but is at odds with the evidence for an ice-dammed lake within Glen Quoich, and would require a large readvance of local ice given the evidence of external ice sloping towards the Cairngorms higher up on the valley sides. The second hypothesis fits with the evidence for the Glen Quoich ice-dammed lake, and the channels can be explained as acting as subglacial drainage from the lake through the invading Deeside ice as it retreated. The third hypothesis is favoured and has some additional evidence to support it. The lateral meltwater channels north of Clais Fhearnaig, described above, indicate ice from the west joined Glen Quoich at c.550–600m and it is likely ice was higher than this previously. This ice would have filled the floor of Glen Quoich and dammed water further up Glen Quoich. First, water flowed over the Gleann an t-
Slugain col at 600m and later, when the damming ice lowered, it drained around the eastern margin of the damming ice mass, forming the lower channels on the eastern side of Glen Quoich. It is possible this western damming ice was also fed to an extent by ice from the main Dee valley to the south. Some of the terraces and kettle holes in lower Glen Quoich may also be linked to this regional ice from the west/south.

4.2.4.4 Dubh-Ghleann

Joining Glen Quoich from the north-west is Dubh-Ghleann, which drains the plateau of Mòine Bhealaidh. A large tree-covered ridge has been identified in the lower area of Dubh-Ghleann opposite Poll Bhàt. The broad ridge is c.20m above the valley floor and has a gently arcuate planform which connects to the northern valley side. It is tentatively suggested it may be a moraine sourced from up-valley. On the opposite side of the valley, immediately north-east of Poll Bhàt, many ridges can be seen. These appear to be sourced from the col with Glen Derry and slowly decline in altitude before turning back on themselves and declining towards the Dubh-Ghleann valley floor (Figure 4.2.25). The highest feature turns back on itself, suggesting the presence of a glacier in Dubh-Ghleann sourced from the north. These are interpreted to be landforms that mark the passage of ice over the col from Glen Derry into Dubh-Ghleann and possibly the contemporaneous presence of a glacier in Dubh-Ghleann.

Figure 4.2.25 Moraines and glaciofluvial landforms associated with ice from Glen Derry and Dubh-Ghleann. Poll Bhàt to the left of the photo
4.2.5 Mòine Bhealaidh

Figure 4.2.26 Geomorphological map of the northern side of Mòine Bhealaidh plateau
Mòine Bhealaidh is a large plateau area c.850–900m in altitude, located between Glen Derry and the higher slopes of Beinn a’ Bhuird; it drains north into Glen Avon and south into Glen Derry, and via Dubh-Ghleann into Glen Quoich. Much of the inner plateau surface is now covered by peat.

On the northern side, dry meltwater channels are present draining from the plateau surfaces obliquely down the shallow valley that holds Allt Cumh na Còinnich (Figure 4.2.26). These are particularly prominent on the western side where they run obliquely from the plateau and slopes of Beinn a’ Chaorainn Bheag towards the valley floor for up to 2km (Figure 4.2.27 and Figure 4.2.28). The highest channels begin at the summit of Beinn a’ Chaorainn Bheag and drain directly into Glen Avon, whereas the lower channels descend into Allt Cumh na Còinnich before joining Glen Avon. At the upper end of the channels they are clearly directed by ice and are most likely ice marginal, however lower in their profile this becomes less certain. Between the lower channels, rounded ridges can be seen trending obliquely down-valley on either side of the valley. Some conflict in orientation exists between the channels which are more oblique than the ridges that descend more quickly to the valley floor. At the valley head the drift deposits rapidly merge with moulded bedrock (Figure 4.2.29 and Figure 4.2.30). Solifluction exists on many of the higher outer channels but to a lesser extent on the central valley area. This change coincides with the most prominent meltwater channel on the eastern side, and the segmented ridges also only occur within this limit. However, given the morphologically similar channels throughout the valley it is likely the landforms represent the retreat of one event. In addition, the ridges are heavily shaped by meltwater and much of their relief is associated with meltwater erosion, unlike the sharp-crested moraine formations associated with the readvance of ice elsewhere. This evidence indicates the retreat of an ice mass sourced from the southerly direction, either from the plateau or including external ice from the south or west given the altitude of the highest channels.
Figure 4.2.27 Channels and ridges on west side of Allt Cumh na Côinnich

Figure 4.2.28 Channels and ridges on east side of Allt Cumh na Côinnich
Within the shallow Coire Ruairidh, a ridge has dammed a series of small lakes on the up-valley side (Figure 4.2.31). Down-valley multiple large mounds and smaller ridges have been mapped. From the orientation of the ridges it is difficult to identify the geometry of the glacier that formed them. The moraine surfaces and the surrounding slopes have suffered from solifluction, thus formation during ice-sheet deglaciation is favoured.
Figure 4.2.31 Ridge in Coire Ruairidh

On the southern side of Mòine Bhealaidh, several tributaries drain into the deeply eroded head of Dubh-Ghleann. The lowermost section of Dubh-Ghleann is described in Section 4.2.4.4. The middle section is predominantly covered in soliflucted surfaces and mature talus. At the head of the valley, on the western side, a long subtle ridge descends from the valley side to the north-east of Allt Clais nam Balgair and can be matched with subtle ridges descending from the opposite side of the stream (Figure 4.2.32). The ridges, particularly on the southern side, coincide with a change in solifluction development. However, much of the surrounding higher plateau above the ridges is heavily soliflucted and some of the differences in solifluction may be coincidental, due to more recent postglacial processes inside the limit. Given the subtle ridge and solifluction on the likely plateau-source areas, this margin is favoured to be from ice-sheet deglaciation. On the opposite side of Dubh-Ghleann, the valley side is heavily soliflucted with multiple sub-horizontal features at 800–850m on the valley side, which may mark the position of older meltwater channels. Lower in the valley floor, deposits are present but whether these are fluvial or glacial in origin is not clear.

On the western side of the plateau, Glas Allt Mòr drains into Glen Derry. This is a narrow V-shaped valley with heavily dissected drift on the valley sides. No evidence for a glacier readvance has been preserved here.
On the northern side of the plateau is the high ground of Beinn a’ Chaorainn. The flanks of this mountain are heavily soliflucted. On the north-east side of the mountain, the upper section of the shallow Coire nan Clach is soliflucted, but lower down broad ridges can be seen, particularly to the east of the stream. Their location within an area of solifluction and the morphology of these ridges suggests they are features from ice-sheet deglaciation.

Figure 4.2.32 Ridges in upper Dubh-Ghleann
4.2.6 Summary of interpretation: eastern Cairngorms

A summary of the geomorphology and relative retreat patterns is presented below. Further discussion of the wider retreat patterns and schematic diagrams of the retreat margins can be found in Chapters 6 and 7.

- Early in deglaciation the lower eastern valleys and hills were overtopped by ice from the main Dee valley as it flowed north-east around the south-east flanks of the Cairngorms through valleys such as the Gairn valley and Glen Builg. This also included ice flow into Glen Quoich and Gleann an t-Slugain.

- This was followed by the retreat of ice from the Gairn valley and Gleann an t-Slugain. However, regional ice remained in lower Glen Quoich and dammed a 600m a.s.l. lake in Glen Quoich and the lower part of Dubh-Ghleann. At this time a glacier sourced from the corries of Beinn a’ Bhuidr terminated within the lake. The altitude of the lake suggests drainage via the spillway into the now deglaciated Gleann an t-Slugain. At this time the Derry glacier must have ceased overtopping the col into Dubh-Ghleann, as meltwater from a contemporaneous ice-dammed lake in Glen Derry drained over the col to form the Poll Bhàt delta deposit. The presence of a contemporaneous c.675m a.s.l. lake in Glen Derry implies that Glen Lui and lower Glen Quoich remained full of ice which joined through Clais Fhearnaig, most likely with ice being sourced from the west. It is likely that as the damming ice lowered, the channels north of Clais Fhearnaig became the new point of Derry lake drainage, still east into the Quoich lake. The position of the Dubh-Ghleann glacier at this stage is unclear but may have been near the moraine in lower Dubh-Ghleann.

- The relative timing of the deglaciation of Slochd Mòr is unclear. If the terraces were formed as a delta in an ice-dammed lake, then they must predate the sequence above as the damming ice would have been sourced via the Gairn valley and Glen Builg. If the terraces are otherwise formed then no inferences about the relative chronology can be drawn. Further detailed work, including sedimentological work, could resolve this uncertainty and provide a relative chronology.
• The valley of Allt Cumh na Còinnich records the retreat of ice from the south, most likely sourced from the plateau of Mòine Bhealaidh. Similar retreat evidence is not as clear within Dubh-Ghleann to the south. More limited moraines in the valley head of Dubh-Ghleann indicate more restricted glaciation but the timing of this and the last ice to flow north from the plateau remains unclear, although ice-sheet deglaciation is favoured.

• The high eastern plateau surfaces of Ben Avon contain no clear evidence of glacier margins and extensive solifluction surfaces indicate it has most likely been ice free since ice-sheet deglaciation, unless limited cold-based ice existed.

• There is good evidence for a later readvance of small corrie glaciers. This is suggested by boulder-rich arcuate moraine deposits extending from some of the highest corries Coir an Dubh-Lochain and Coire na Ciche of Beinn a’ Bhuird, upper Slochd Mòr, Lochan nan Gabhar, and Sgor Riabhach.
4.3 Central Cairngorms

4.3.1 Glen Derry and Glen Lui

4.3.1.1 Glen Lui

It has long been recognised that Glen Derry and Glen Lui possess an outstanding group of landforms that mark several phases of glacial retreat (Bremner, 1929). Where Glen Lui joins the main Dee valley, glacial drift and meltwater channels have been mapped (Figure 4.3.1). The channels originate in the main Dee valley and descend towards the exit of Glen Lui. Above these lower channels, on the col between Carn an ‘Ic Dubh and Sgòr Dubh, meltwater channels are present orientated south-west to north-east from the main Dee valley across the col into Glen Lui. These channels are interpreted to be associated with the retreat of the main Dee glacier. Within the valley on the northern side of Lui Water, south-west of the Clais Fhearnaig channel, subtle moraine ridges slope down-valley out of the Cairngorms (Figure 4.3.1). Their orientation is not conclusive but combined with the moraines consisting predominantly of granite it suggests they mark ice retreat from within the Cairngorms. Above these deposits, c.500–700m on the northern side of Glen Lui are glacial deposits and meltwater channels that descend towards and into Glen Quoich (described in Section 4.2.4.3).

At the exit of Glen Derry into Glen Lui, near Derry Lodge, there are a large number of high-relief ridges (Figure 4.3.1). On the western side, ridges descend from Creag Bad an t-Seabhaig south towards the valley floor. In this area the morphology is the most complex, with ridge crests cross-cutting each other (Figure 4.3.2 and Figure 4.3.3). The ridges continue on the southern side of Lui Water, where they curve to the east (Figure 4.3.4). The ridges then stop/cannot be followed under the trees and steeper ground. These ridges, both north and south of the river, are covered in granite boulders, suggesting they were associated with locally sourced ice. Outside the most prominent ridge, more discontinuous and less prominent ridge and mound segments mimic the inner ridges’ shape. On the eastern side of the Derry/Lui confluence, multiple lower relief ridges can be found near the Bob Scott bothy, some dissected by Lui Water. Forestry plantations make following the ridges to the north of the river difficult. However, higher at 500–550m on the northern side of the
planted, a large ridge can be found below the Glen Derry lake shorelines and deposits that lead towards Glen Quoich. The orientation of the ridge crest is obliquely down-valley towards the ridges described immediately above. Together these ridges are interpreted as an arcuate moraine system, forming the lateral and frontal position of the Glen Derry glacier extending into Glen Lui. Within these limits, immediately behind Derry Lodge, a large curved sharp-crested ridge exists. The sharp nature of this feature at its down-valley end is likely to be due to fluvial modification, as the Derry Burn runs nearby and the crest can be traced up-valley into an area of more gently undulating moraine morphology.
Figure 4.3.1 Glen Lui and Derry geomorphological map
Figure 4.3.2 North-west moraines near Derry Lodge (below Creag Bad an t-Seabhaig) – note complex cross-cutting crests

Figure 4.3.3 North-west moraines below Creag Bad an t-Seabhaig – note complex cross-cutting crests. Rectangle contains the linear bench features in Coire Craobh an Oir
Within Glen Derry, 1km to the north of Derry Lodge, a large ridge exists on the western side of the Derry Burn (Figure 4.3.1). The feature has a steep east face and a palaeochannel on the northern side; these faces have been incised and modified by fluvial activity. The original ridge crest is arcuate and can be followed back to an area of mounedy glacial drift on the valley side. The origin of the ridge is not clear, but here it is thought to be a recessional moraine segment from the retreat of the Glen Derry glacier.

A further kilometre up-valley, a large broad deposit, c.400m from front to back, spans the valley floor (Figure 4.3.1). Several shallow palaeochannels run over the deposit surface from north to south. These may have been formed at the time of deposition or at a later time when acting as the spillways from the dammed lake up-valley. Direct evidence of the palaeolake can be seen across the northern side of the deposit, where a lake shoreline has formed (Figure 4.3.5). Today the lake is fully drained by a channel cut on the eastern side of the deposit; however, the flat lake infill can be seen extending up-valley behind the moraine (Figure 4.3.6).

The deposit has sporadic large granite boulders on its surface and is interpreted to be a moraine. The planform shape of the shallow moraine crest does not immediately indicate an ice-flow direction; however, the up-valley side is steeper suggesting it is the proximal ice-contact face. In addition, the granite boulders and
some smaller crests on the deposit suggest formation by a glacier up-valley. This would also be consistent with the southerly flow of ice through Glen Derry to form the Derry Lodge moraines. There are no lateral moraines associated with this feature, but the altitude of the gullied drift limits on the valley sides to the north match with the notion of glacier flow from up Glen Derry. Additional hypotheses for the formation of the relatively broad flat-natured moraine include formation at a water-terminating glacier margin or streamlining of the feature by glacial readvance. The idea of this large deposit being a palimpsest landform cannot be ruled out.

Figure 4.3.5 Glen Derry moraine and the lake shoreline cut into its northern side
Figure 4.3.6 Flat lake fill looking north up Glen Derry from moraine

Above the moraine, on the eastern side of Glen Derry, from 575–675m a.s.l., widespread horizontal bench-like features can be seen (Figure 4.3.1 and Figure 4.3.7). The benches are fragmented in places by solifluction; nevertheless, they can be followed as continuous features for over 2km. The benches are c.10–25m wide and their flatter surfaces dip at c.10° compared to the inter-bench slope dip of c.20° (Figure 4.3.8). A smaller but distinctive fragment can be seen on the opposite western side of the valley above Derry Dam moraine (Figure 4.3.9, Figure 4.3.10 and Figure 4.3.11). This feature at c.675m matches both the altitude of the most prominent bench on the eastern side and the spillway channel that descends east into Glen Quoich (Figure 4.3.1). This is strong evidence that the benches formed as lake shorelines in a large ice-dammed lake that occupied at least this section of Glen Derry.

Some 20m above the main spillway into Glen Quoich, a large inlet channel into the main channel exists, as do other channels that descend obliquely into Glen Quoich. These features above the col must have formed while ice was still overtopping the col from Glen Derry into Glen Quoich. The spillway at the lowest point of the col is of the same altitude as the uppermost and most prominent lake shoreline, indicating this was the overflow point of the ice-dammed lake. Thus it is proposed that as the
Derry glacier lowered and no longer fed ice over the col, water became dammed in Glen Derry and overflowed towards Glen Quoich/Dubh-Ghlean, creating the incised spillway and meltwater channel towards the Poll Bhàt delta (described in Section 4.2.4.2.).

The lake shorelines continue south and link with the meltwater channels above Clais Fhearnaig. This suggests that as the damming ice lowered or retreated, the channels immediately north of Clais Fhearnaig became the new lower overflow point (Figure 4.3.12). Multiple lake shorelines and the corresponding meltwater channels have been identified, suggesting this continued for some time while the damming ice mass gradually lowered. There are also candidate lake shorelines in Coire Craobh an Oir (Figure 4.3.1 and Figure 4.3.3). In the northerly direction, the clarity of the shorelines fades at the Allt Coire an Fhir Bhogha. This is likely to be due to the change in slope aspect and increased shelter from wave action at this point, rather than being indicative of the lake’s extent. The slopes below Craig Derry become steeper and heavily soliflucted; although horizontal soliflucted vegetated surfaces can be seen in the otherwise boulder-dominated soliflucted slope (Figure 4.3.13). These lines do occur at broadly the same heights as the lake shorelines further south, but they may also represent soliflucted lateral moraines or otherwise generated patterns within the soliflucted surfaces. Thus determining the northern, up-valley extent of the lake is difficult to resolve.

Described above is strong evidence for ice-dammed lakes at c.675m within Glen Derry and at c.600m in Glen Quoich. They are thought to have occurred simultaneously as meltwater from the Glen Derry lake spillway drained into Glen Quoich to form the Poll Bhàt delta. Consequently, the minimum damming heights required are: Glen Quoich to 600m, the exit of Glen Derry to 675m, and the Clais Fhearnaig col which links the two valleys must also have been dammed to 675m to enable the existence of two lakes of differing heights. Thus lower Glen Lui, lower Glen Quoich and Clais Fhearnaig were blocked, either by ice from the south or by ice sourced from the western Cairngorms across the Meirleach col; the Glen Quoich evidence is discussed further in Section 4.2.4.3.
Figure 4.3.7 Lake shorelines east Glen Derry

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<td>12°</td>
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Figure 4.3.8 Field measurements of shoreline profiles for the shoreline level with the spillway channel on the eastern side of Glen Derry
Figure 4.3.9 Glen Derry shoreline west side

Figure 4.3.10 Profile of western shoreline, Glen Derry
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Figure 4.3.11 Field-measured shoreline profiles and width for western side of Glen Derry

Figure 4.3.12 Flat benched landforms that link into the meltwater channels above Clais Fhearnaig (right).
Figure 4.3.13 Eastern side of Glen Derry below Craig Derry. Linear patterns within the soliflucted talus are candidate continuations of the lake shorelines

Discussion

Separately, the evidence of a lake in Glen Derry and readvance of ice from Glen Derry into Glen Lui (near Derry Lodge) are strong. However, combining the events into a relative chronology is more problematic. The first hypothesis is that the Derry Lodge moraines were formed, followed by the ice-dammed lake. This would imply the full or partial deglaciation of Glen Lui to enable the Glen Derry glacier to form the moraines near Derry Lodge, followed by the readvance of a c.300m-thick ice mass, from the south into Glen Lui. This must have advanced to within hundreds of metres of the Derry Lodge moraines to block Clais Fhearnaig at 675m. The second hypothesis is that the ice-dammed lake existed first, followed by the readvance of a glacier from Glen Derry into Glen Lui. This would imply that during retreat the Derry glacier lowered or retreated, while regional Glen Lui ice, sourced from the west or south, remained higher. This formed a lake either on the glacier surface or between the separated glaciers. Later the glacier margins near Derry Lodge and Derry Dam were formed by the readvance of ice from Glen Derry, possibly as a response to the drainage and change to a land-terminating margin. The second hypothesis is more favourable from a geomorphological and glaciological viewpoint, as it puts fewer constraints on the margin of the damming ice. However, the second hypothesis requires glaciological testing to identify whether the ice surface of the Derry glacier that formed the Derry Lodge moraines can have existed below the preserved Glen Derry lake shorelines. Alternatively, the landforms at Derry Lodge have been misinterpreted.
4.3.1.3 Upper Glen Derry

Figure 4.3.14 Overview geomorphological map of upper Glen Derry, Glen Avon, Loch Etchachan and the Ben Macdui plateau

At its head, Glen Derry joins the glacial breach Lairig an Laoigh, while the main valley turns west into the valley head Coire Etchachan (Figure 4.3.14). At this point, glacial drift and small individual moraine ridges cover much of the valley floor, and in certain areas the individual moraines can be joined to indicate former glacier margins. Several spatial groupings of landforms have been distinguished; these are described below.

To the north, moraines are found at the col with Lairig an Laoigh. The geometry of the moraines suggests ice flowed from upper Glen Derry northwards towards Glen
Avon. These continue north down-valley; however, the clarity of the former glacier margins quickly reduces and the deposits merge into an area of eskers and kettle holes, similar to the ice stagnation lakes seen in Ashley (2002). The moraines at the col are comparable in appearance to those within upper Glen Derry, which could be argued to indicate contemporaneity. However, there are some important differences: the moraines within upper Glen Derry are fresher in appearance and are more consistent in alignment; those at the col are significantly higher in altitude than the main area of moraines within Glen Derry and the two areas are not linked by continuous ice-marginal retreat moraines (Figure 4.3.15).

Figure 4.3.15 Glen Derry ice-marginal moraines (in valley centre) and moraines at the col with Lairig an Laoigh (on the left-hand side)
Figure 4.3.16 Close-up geomorphological map of upper Glen Derry and Coire Etchachan. Arrow marks down-valley limit to closely spaced high-relief moraines

Up-valley of the alluvial fan descending from Glas Allt Mòr, shallow subtle moraine ridges can be seen on the valley floor (Figure 4.3.14). The moraines become more concentrated and more defined into steeper-sided prominent moraine ridges at (57.073380° N, 3.594753° W) (Figure 4.3.16). These moraines in the head of Glen Derry and Coire Etchachan are interpreted as lateral and frontal moraines that indicate the retreat of a glacier sourced to the west, not the neighbouring lower eastern Mòine Bhealaidh plateau (Figure 4.3.17). The individual recessional moraine ridges stop 300m short of the Hutchison Memorial Hut, where a ridge consisting of large boulders crosses the width of the valley (Figure 4.3.18). Some 130m up-valley, another ridge joins the central spur of the valley and the southern valley side, with the Hutchison Memorial Hut located upon it. A further 250m up-valley, another ridge joins the central spur to the southern side. These uppermost ridges have significant amounts of fluvial infilling on their up-valley sides. The origin and internal composition of the large central spur is not immediately apparent – it may be a combination of a glacial deposit and underlying bedrock feature. At the up-valley end, the crest contours beneath the Creagan a’ Choire Etchachan rock slope above, but the feature’s size and its spatial separation from the rock slope at its lower end is not consistent with a protalus rampart interpretation, thus it is
thought to be at least partially a glacial deposit. Indeed the surface of the spur is covered with subtle moraine ridges that trend obliquely down-valley. On the opposite side, north of the incised stream sourced from Loch Etchachan, multiple corresponding small moraine ridges are present (Figure 4.3.16). Together these moraines mark the retreat margins of a glacier sourced from the corrie that holds Loch Etchachan. No solifluction exists within the moraine limits of the glacier in upper Glen Derry, whereas it can be found outside the southern moraine limits and on the valley sides north of the Lairig an Laoigh col. This suggests a later stage of glaciation may have occurred in Coire Etchachan and upper Glen Derry.

Figure 4.3.17 Glen Derry ice-marginal moraines. Col to Loch Etchachan in the centre
Figure 4.3.18 Moraine ridges in upper Glen Derry/Coire Etchachan. Blue = boulder ridge, red = Hutchinson Hut Ridge, black = uppermost ridge, orange = ridges from Loch Etchachan (Adapted from Midgley, 2001)

The described landforms at the head of Glen Derry and Lairig an Laoigh are interpreted to be from the sequence of glacial events below. The eskers and kettled topography in lower Lairig an Laoigh indicate local stagnation of ice, potentially as the glacier became cut off from the Derry source areas. Up-valley at the Lairig a Laoigh/Glen Derry col and within Glen Derry, moraines mark the retreat of a glacier in Glen Derry towards the west. The southernmost, more subtle moraines in upper Glen Derry and the moraines at the Lairig a Laoigh col are interpreted here to predate the main area of sharp-crested ice-marginal moraines located on the Glen Derry valley floor. Within the head of Glen Derry and the surrounding plateau surfaces, no evidence for plateau-sourced glaciation has been mapped; whether the linking Loch Etchachan glacier was plateau-fed is discussed in Section 4.3.5.1 and 4.3.5.2. During the Glen Derry glacier’s final retreat it divided into two smaller glaciers. Two ridges mark the final stages of glaciation in the head of Glen Derry/Coire Etchachan, while small moraine ridges mark the retreat of a glacier into the corrie that holds Loch Etchachan. The absence of solifluction within upper Glen Derry indicates that a later glacier readvance in this area cannot be eliminated and will be discussed further in Chapter 5.
4.3.1.4 Corries of Glen Derry

On the western side of Glen Derry is the shallow corrie of Coire na Saobhaidh (Figure 4.3.1). Only a few hundred metres from the rock backwall, an area of large boulders and a small but well-defined arcuate moraine ridge marks a former corrie-glacier margin (Figure 4.3.19). Within this outer ridge a faint inner moraine can be seen. At both ends of the moraine, channels descend to the valley below. These may be solely postglacial features, but they may have initially carried meltwater from the glacier. Little solifluction can be seen within the glacier limits, but no distinct change in solifluction level is present outside the glacier margin, thus inferring timing is problematic.

Figure 4.3.19 Coire na Saobhaidh. Note boulder area and small arcuate ridge

Approximately 3km to the north, Coire an Lochain Uaine is eroded into the eastern side of Derry Cairngorm. The corrie floor is covered by large ridge deposits (Figure 4.3.20). The outermost is a moundy semi-circular ridge covered with embedded transported boulders (Figure 4.3.21). Within this there are two large deposits, again covered with large boulders. On the gentle south-west corrie backwall directly behind these deposits, the corrie is covered in boulders from rockfall or mass movement, with many perched boulders present. The semi-circular outer ridge and inner deposits appear to be sourced from the highest south-west side of the corrie. Interpretation of these features is complex. The slope side of the features has been modified by talus formation from the crags and slopes above. The ridge landforms
are not typical of readvance corrie glaciation in the Cairngorms, but whether all the ridges formed through mass movement events or whether the ridges mark glacial activity is uncertain.

Figure 4.3.20 Coire an Lochain Uaine in Glen Derry

Figure 4.3.21 Northern ridges in Coire an Lochain Uaine, Glen Derry (looking east). Black line indicates the position of the outermost ridge
4.3.2 Glen Luibeg, Meirleach Col and Coire an Lochain (Ben Macdui)

Figure 4.3.22 Glen Luibeg geomorphological map
Glen Luibeg runs parallel to Glen Derry on the western side, and exits into the same east-to-west orientated valley that joins the Meirleach col and Glen Lui/Derry Lodge (Figure 4.3.22). The area south of the river, between the Derry Lodge moraine system and the Luibeg Bridge, possesses large rounded deposits and relict channel systems. These deposits are relatively close to the valley floor and have an abrupt upper limit above which they do not exist and the drift becomes smoother. On the western side of the confluence of the Luibeg Burn and Allt Preas nam Meirleach, the channels are close to the present-day stream Allt Preas nam Meirleach. The geometry and elevated level of the channels above the valley floor suggests they were redirected by the presence of ice. The channels may have been incised by the locally sourced stream, but their size may be better explained by the former meltwater that was supplied over the Meirleach col. Therefore a tentative explanation of these landforms is that as the Dee glacier lowered, it stopped supplying ice over the Meirleach col; down-wasting ice was left near the exit of Glen Luibeg, then while meltwater was still being fed over the col the channels were formed around the down-wasting ice. The meltwater presumably would have been supplied, at least at first, directly from the retreating ice margin, but later, meltwater from the ice-dammed lake in lower Glen Dee (Section 4.4.3.1) and the ice-dammed lake within upper Glen Dee (Section 4.4.3.2 and 4.4.3.3) are thought to have overflowed across the Meirleach col. The landforms 500m to the east are similar in appearance and at their eastern limit have an abrupt termination with the Derry Lodge moraine system. Although being more difficult to interpret, they too are likely to represent the combination of down-wasting ice and meltwater or stream activity.

Given the knowledge of the surrounding valleys, this area would have undergone a complex deglaciation history and some alternative hypotheses include: the landforms marking the retreat or down-wasting of ice sourced from Glen Derry or external ice to the south; or the area may have been dammed by the Derry glacier’s outermost position near Derry Lodge, forming a possible ice-dammed or moraine-dammed lake. Thus the history of the area south of Glen Luibeg is difficult to interpret and may be a composite landscape.

Just within the exit of Glen Luibeg a large moraine ridge can be seen rising above the valley floor (Figure 4.3.22). Meltwater channels are present on the south-west side of the moraine draining to the south (Figure 4.3.23). On the northern side of
the moraine gently sloping delta surfaces are present (Figure 4.3.24). Previous mapping and sediment work indicated the large moraine and delta surfaces were formed by ice on the southern side blocking lower Glen Luibeg, forming the moraine, and then meltwater from the southern margin building the delta surface in a lake behind the moraine (Golledge, 2003).

Figure 4.3.23 Glen Luibeg moraine – channels on the south-west side can be seen draining to the south (picture left)
To the north smaller moraines and meltwater channels are present; their orientation down-valley indicates the ice-contact slope was on the northern side (Golledge, 2003). These moraines continue up-valley to the divide of Glen Luibeg. The western branch possesses further moraines orientated down-valley, and the moraines and glacial drift continue up-valley to the saddle that divides Glen Luibeg from the neighbouring Glen Dee (Figure 4.3.22). At the saddle, ridges and meltwater channels are present; their position and oblique orientation marks glaciation from the west, indicating ice flow in this valley was sourced over the col from Glen Dee. The retreat of the eastern branch of Glen Luibeg is less well defined but areas of glacial drift and moraines mark the separate retreat of a glacier up-valley. Evidence is sparse and it is not clear whether the glacier was valley- or plateau-sourced. There are no well-defined sharp-crested moraines typical of a later readvance and the valley sides are heavily soliflucted.

The unnamed corrie that holds Lochain Uaine is situated on the western side of the Luibeg valley and is eroded into the high plateau slopes of Ben Macdui (Figure 4.3.22). At the corrie lip a distinct moraine ridge has been mapped, below which soliflucted surfaces continue to the valley floor. Inside this limit there is a well-defined second moraine ridge that dams Lochain Uaine (Figure 4.3.25). No
soliflucted surfaces are found within either moraine limit, and there is a reduced talus development inside the glacier limits. These glacier margins are interpreted to mark a later stillstand or readvance, but whether they both mark the same event is unclear.

![Figure 4.3.25 Unnamed corrie that holds Lochain Uaine, Ben Macdui. Note the inner and outer moraine ridges beyond the lake](image)

### 4.3.3 Glen Avon, Strath Nethy and Loch Etchachan

At the intersection of Glen Avon and Lairig an Laoigh, two distinct sets of landforms have been identified (Figure 4.3.14). On the southern side of the intersection, in Lairig an Laoigh, fluviglacial deposits including eskers and kettle holes are present. The most prominent esker lies to the west of the northern lake and its sinuous ridge can be followed for 500m (Figure 4.3.26). At the intersection of Lairig an Laoigh and Glen Avon, on the south-west side, a higher surface of glacial deposits is present. The surfaces consist of several broad parallel ridges orientated obliquely down-valley, separated by shallow channels. The orientation of these ridges and channels, combined with the continuation of the same surfaces up-valley into Glen Avon, suggests these are deposits from a glacier sourced further up Glen
Avon. At the south-east limit of these deposits, the topography descends to smaller ridges which are seemingly associated with the same surface as the esker and glaciofluvial deposits within Lairig and Laoigh to the south. These lower landforms are interpreted to mark the deglaciation of a glacier sourced from Glen Derry to the south. The deposits appear to be overprinted by the higher and morphologically separate Glen Avon deposits above.

In front of these deposits, in the central area of the valley intersection, palaeochannels orientated north-west to south-east may mark braided outwash or postglacial migration of the River Avon. To the north of the river, further glacial deposits and meltwater channels are present; these descend towards the River Avon. The orientation of the deposits and channels suggest they are either lateral to the easterly flowing Avon glacier, frontal to the northerly flowing glacier from Glen Derry, or formed when the glaciers were still confluent. The overprinting of the Derry deposits by the Glen Avon deposits, described above, suggests the Glen Avon glacier survived or readvanced into the area at a later time.

Figure 4.3.26 Esker in Lairig an Laoigh. Blue = esker crest.

The deposit surfaces from the Avon glacier can be traced up-valley to the eastern end of Loch Avon. Moraine ridges and meltwater channels are particularly clear immediately down-valley of Loch Avon, on the northern side of the river. These mark the retreat of the Avon glacier and the final valley-wide moraine-ridge dams Loch Avon (Figure 4.3.27). These moraines mark the presence of a locally sourced glacier, most likely sourced from the high plateaus of Ben Macdui and Cairn Gorm. The moraine surfaces continue on either side of the lake, with two particularly
prominent deposits opposite the saddle with Strath Nethy; these reduce both the width and depth of the lake.

![Glen Avon moraine dam and Loch Avon outflow](image)

Figure 4.3.27 Glen Avon moraine dam and Loch Avon outflow

The glacial drift continues up-valley on both sides, though a distinct change in the thickness of drift and onset of closely spaced, individual moraine ridges occurs nearer the valley head (Figure 4.3.14 and Figure 4.3.28). Within this area, on the north (Figure 4.3.29) and south sides of the lake (Figure 4.3.30), lateral moraine ridges mark glacier margin retreat up-valley. To the west of the lake interpreting the deposits is more problematic. The area is a complex interplay between large sediment accumulations and smaller individual ridges and channels. This is most prevalent below the Shelter Stone Crag, where three large deposits are found (Figure 4.3.31). Instead of the deposits being abutted to the valley side as typical for other moraines within the Cairngorms, the deposits stand several tens of metres from the headwall. Below these deposits, another large deposit forms a second level, on which small individual ridge crests can be seen. Large channels can be found running west to east, dissecting the large accumulations and continuing into the lower-level moraine areas (Figure 4.3.32). Midgley (2001) argued the smaller
ridges appear to show imbricated stacking, and thus suggested formation through proglacial or englacial thrusting.

A composite explanation is favoured for the complex landforms beneath the Shelter Stone Crag. It is interpreted here that the large accumulation deposits were formed prior to the last glacial activity in the area, either by mass movement or earlier glacial deposition. Later they were modified and overprinted by a glacier, forming the smaller moraine ridges and meltwater channels. On the northern side of the river a large moraine is present; it is of similar size and height to the lower deposit group on the southern side.

Figure 4.3.28 Geomorphology at the head of Glen Avon

To the west, more subtle longer ridges exist that trend obliquely down-valley from the bedrock above. These mark the retreat of the glacier into the valley head or up onto the plateau (Figure 4.3.33). The head of the valley consists of glacially smoothed bedrock surfaces. No moraines have been traced on the plateau surface and no other direct evidence linking this glacier with plateau glaciation has been found. The plateau surrounding Glen Avon is described in Section 4.3.5.

On the southern side of Glen Avon, ridges can be seen running up the valley side towards the col with Loch Etchachan. The outer and most prominent ridge can be seen below the cliffs and talus in Figure 4.3.31. This is interpreted to be a lateral
moraine of contemporaneous age to the moraines below at the up-valley end of Loch Avon. The ridge does not rise onto the col; however, its steep nature and position at the exit of the col is indicative of ice flow from Loch Etchachan into Glen Avon. The floor of the col is covered with glacially transported boulders and irregular drift (Figure 4.3.34). On the eastern side of the col, small meltwater channels can be seen dipping towards Glen Avon. This evidence suggests a glacier flowed north across the col into Glen Avon. The Loch Etchachan geomorphology is described in more detail in Section 4.3.5.1.

Together, the change in thickness of drift and the clarity of individual moraine crests at the head of Glen Avon suggest a later glacier may have existed in the area. The glacier was sourced from the valley head to the west, with ice also most likely being contributed from the Loch Etchachan corrie. No solifluction surfaces have been identified within this moraine limit, neither are many soliflucted surfaces seen further down the valley. Instead the soliflucted surfaces occur on the higher flanks, which does not assist in constraining the timing of glaciation.

Figure 4.3.29 Lateral moraines on the northern side of upper Glen Avon
Figure 4.3.30 Lateral moraines on the southern side of upper Glen Avon

Figure 4.3.31 Upper Glen Avon. Blue polygon indicates the position of the three large deposits. Clear polygon indicates the lower level covered by smaller ridges. Red polygon shows the location of the lateral moraine ridge – note it rises towards the col with Loch Etchachan. The back line indicates the outer position of the closely spaced moraine ridges
Figure 4.3.32 Channel between the large accumulations of sediment at the head of Glen Avon

Figure 4.3.33 Glen Avon – moraine ridges marking the retreat into the valley head/plateau
Figure 4.3.34 Glacially transported boulders and irregular glacial drift on the col between Loch Etchachan and Glen Avon. Looking north towards the deeply incised Glen Avon and Cairn Gorm in the distance.

4.3.3.1 Strath Nethy

No additional fieldwork has been undertaken in Strath Nethy and interpretation from aerial images is consistent with existing literature (Brazier et al., 1996b). On the western side, kame terraces dip northwards up-valley; these mark the lowering of the external Speyside lobe (Figure 4.3.35). These are associated with the large meltwater channels that cross the spur from the west, on the western side of Strath Nethy. Up-valley of the terraces, irregular deposits fill the valley floor and are draped in large boulders, particularly on the western side (Figure 4.3.35). These continue largely uninterrupted for 2km. Interpretations differ as to whether this represents a rockslide, or rock avalanche in situ over pre-existing till (Ballantyne et al., 2009a); or rock emplacement while ice was decaying, either stagnantly or advancing slowly (cf. Jarman et al., 2013).

4.3.4 Speyside and Northern Corries

The landforms on the northern flanks have been well documented elsewhere (Hinxman and Anderson, 1915; Sugden, 1970; Sissons 1979a; Brazier et al., 1996a, 1996b, 1998); these include those from both ice-sheet deglaciation and later corrie glaciation. On the lower mountain flanks, extensive thick drift, terraces,
large meltwater channels cut into both drift and bedrock, and lateral moraines are present. The highest channels occur at 780m, above the large rock-cut channel of Chalamain Gap; channels of a similar altitude are common to most spurs east towards Strath Nethy. Between the spurs, in the lower sections of the valleys that extend from the northern corries, lateral moraines from the Speyside glacier are present; some of these are described later within this section. Lower on the mountain, the features become more continuous and larger in size, the most obvious being the large meltwater channel and deposits that can be seen to the west of the base-station car park at 500–600m (Figure 4.3.36). Other areas of particular interest are the meltwater channels and terraces towards and inside Strath Nethy; these are discussed in Section 4.3.3.1.
Figure 4.3.35 Northern corries and Strath Nethy geomorphological map

Figure 4.3.36 Northern Cairngorms, view from base station, looking west
Starting at the eastern end of the northerly facing corries, the shallow corrie of Coire Laogh Mòr has glacial-drift ridges within it (Figure 4.3.35). The deposits do not show the extensive solifluction of the plateau flanks, but they are of similar morphology to the wider glacial drift and are located at a similar altitude to the meltwater channel on the Creagan Dubh spur to the west. The evidence is not indicative of a locally sourced readvance; instead, given the morphological and altitudinal similarity with the surrounding ice-sheet deglaciation landforms, it suggests earlier deposition.

Côire na Ciste has extensive soliflucted surfaces in the upper reaches, and lower down-valley glacial drift has been identified. Lateral meltwater channels orientated parallel to the main Spey valley can be seen on the spurs either side of the corrie, the highest at c.780m. These link with the uppermost drift deposits in the corrie, indicating they were formed by the Speyside glacier. The presence of solifluction and the lack of evidence for separate corrie glaciation indicate there was no later readvance of a glacier at this site.

To the west, Coire Cas has a more developed corrie shape. Unfortunately some of the landforms within this corrie may have been affected by the building of ski facilities. The lower section of Coire Cas contains the wider glacial drift associated with the deglaciation of the Spey glacier, and the flanks of the surrounding spurs are covered in solifluction. Above the White Lady Shieling, soliflucted surfaces are present across the central mouth of the corrie, signifying no later readvance of ice through this area. On the upper corrie floor, just below where the gradient becomes flatter, the concentration of boulders increases and they become clast supported (Figure 4.3.37 and Figure 4.3.38), whereas the corrie sides remain soliflucted, except for the main backwall which is covered by talus, and the eastern side, which has been affected by postglacial fluvial processes. No arcuate ridges or crests exist within the boulder deposits and thus emplacement by rockfall and avalanche activity is favoured.
Located to the west is the well-developed Coire an t-Sneachda; the lower section below 770m consists of moraine ridges predominantly orientated with the main Spey valley and found on the neighbouring mountain flanks. These are interpreted to be lateral features from the deglaciation of the main Spey valley. Higher up the corrie, further deposits are found, although their orientation is less clear, and much of the central area of the corrie is heavily soliflucted. It is not easy to discern whether the deposits represent local or Spey glaciation but, based on the solifluction, it is clear these were formed during ice-sheet deglaciation.

Within 500m of the corrie backwall, an abrupt change occurs in the concentration of transported boulders. Inside this limit, arcuate ridges and boulder lines can been seen, with particularly clear examples on the eastern side (Figure 4.3.39). This is
strong evidence for a small corrie-glacier readvance. On the western side, the lateral limit of the glacier can be identified by the boulder ridge on the corrie side. The backwall and overlooking corrie sides are covered by immature talus and alluvial fans, and only contain limited soliflucted surfaces; unlike the extensive well-developed soliflucted surfaces on the outer flanks. These moraine crests and the abrupt change in solifluction development are strong evidence for a later stage of corrie glaciation.

On the eastern side of the corrie, a prominent ridge runs for almost 700 metres below the corrie sidewall. Several formation hypotheses exist: including a lateral moraine of a large corrie or plateau-fed glacier, a protalus rampart, or a bedrock-cored feature. Recent detailed work by Jarman et al. (2013) suggested the feature has 2–3 components and the authors provided an additional hypothesis that it may have been formed by a complex interplay between a snow bank and a pre-Younger Dryas corrie glacier (cf. Jarman et al., 2013).

Figure 4.3.39 Coire an t-Sneachda – the prominent ridge in the corrie foreground, with boulder limit and arcuate moraine ridges on the corrie floor

The similarly well-developed Coire an Lochain possesses a comparable set of glacial features. The lower section is covered by lateral moraines from Spey valley deglaciation, with a particularly prominent ridge occurring at 760m. The
neighbouring valley to the east, along with the northern and eastern slopes of Lurcher’s Crag, possesses well-developed solifluction lobes. It is assumed these slopes deglaciated early and thus, to an extent, can be used as a reference to compare solifluction development within the corrie. The walls of the lower corrie section are extensively soliflucted; however, solifluction on the corrie floor is less well developed. Indeed, boulder-covered arcuate ridges have been identified on the outer corrie floor. The orientation of the 300m long lateral segments and the curved frontal segments of these ridges indicate they may be lateral and frontal moraine ridges from a glacier sourced within the corrie (Figure 4.3.35, Figure 4.3.40 and Figure 4.3.41). There are small areas of immature solifluction on the ridges and the associated glacial drift up-valley. At the head of the corrie, just 500m from the backwall, there is a distinct arcuate limit to the higher concentration of transported boulders (Figure 4.3.42). A few hundred metres inside this outer limit, a 250m long boulder frontal-lateral moraine can be seen on the eastern side. No solifluction occurs within these boulder limits or on the surrounding corrie walls, thus this upper stage of corrie glaciation is interpreted to be from a later readvance.

The evidence within Coire an Lochain suggests there was possibly a large glacier that extended c.1200m from the backwall; given the subdued ridges and solifluction this is likely to have been during ice-sheet deglaciation. Later a glacier readvanced to the well-defined boulder limit, just 500m from the backwall (Figure 4.3.35). A further limit or retreat stage may be associated with the well-defined moraine ridge within the boulder limit. The abrupt boulder limit and absence of solifluction indicate these upper margins are from a later readvance stage. Inside these limits at the corrie backwall, recent research, including new surface exposure ages, has indicated a Little Ice Age glacier may have formed small moraine ridges (Kirkbride et al., 2014).
Figure 4.3.40 Coire an Lochain – lower section showing outer candidate moraine ridges sourced from the corrie to the right of the picture. Note drift and solifluction in background.

Figure 4.3.41 Coire an Lochain – ridge crest of lower candidate lateral moraine sourced from the corrie.
4.3.5 Ben Macdui and Cairn Gorm Plateau

The slopes of Ben Macdui and Cairn Gorm form a continuous plateau area over 1100m in altitude. It is bound by Glen Dee and the Lairig Ghru to the west, the northern corries to the north and the valleys of Glen Derry and Glen Avon to the east. It is these eastern valleys that predominantly drain the plateau surface and towards which most of the topography slopes. In general, the summits of Ben Macdui and Cairn Gorm are covered in blockfields, with the steeper slopes having developed into boulder lobes. Solifluction is not widespread in the Ben Macdui and Loch Etchachan areas of the plateau; it is more prevalent on the slopes of Cairn Lochan, Stob Coire an t-Sneachda and Cairn Gorm. The individual corries and valleys that drain the plateau are described below.

4.3.5.1 Loch Etchachan

Within the unnamed corrie that contains Loch Etchachan, on the corrie floor to the north of the lake, there are moundy deposits and, particularly on the eastern side, channels exist (Figure 4.3.43). Large transported boulders can also be found within this area between Loch Etchachan and Glen Avon, and on the southern side of Loch Etchachan (Figure 4.3.14 and Figure 4.3.44). The glacially transported...
boulders characterised by their high concentration, random deposition and rounded edges contrast with the weathered granite located higher on the slopes which is characterised by the in situ break-up of granite into blocks. This evidence, combined with the down-valley moraines in Glen Derry and Glen Avon, is interpreted to mark the presence of a glacier within the Loch Etchachan corrie, feeding ice into Glen Avon and Glen Derry. The slopes below the crags of the corrie are covered in immature talus, an observation consistent with the suggestion of upper Glen Derry and Glen Avon representing a later readvance. On the slopes to the south of Loch Etchachan and above the meltwater channels on the eastern side towards Glen Avon, some limited solifluction exists. This may suggest the glacier was thin, which is consistent with the low height of transported boulders and channels on the slope sides. Alternatively, it may suggest older glaciation, early retreat within a period of climatic deterioration or continued solifluction during the Holocene.

Below Carn Etchachan, the origin of a boulder deposit has been debated (Sissons, 1979a; Ballantyne et al., 2009a and Jarman et al., 2013). Originally interpreted as a talus rock glacier (Sissons, 1979a), it was recently reinterpreted as an RSF (Ballantyne et al., 2009a) and, subsequently, partly due to the lack of a cavity source for an RSF, it has been interpreted as a moraine covering over a rock boss, with only minor local rockfall evidence (cf. Jarman et al., 2013). Viewing aerial imagery and a fieldwork survey indicated that much of the conical feature away from the backwall is comprised of subtle mounds and ridges, with both surface and embedded boulders on its surface (Figure 4.3.43 and Figure 4.3.45). Importantly, this matches the surrounding pattern of mounds and depressions on the wider corrie floor. Only when approaching the steep corrie side does the deposit change to being more boulder dominated and clast supported, more reminiscent of rockfall activity. The underlying conical shape of the deposit may have its origins as a bedrock feature or an older pre-existing rock slope failure. However, the mounds and ridges, particularly nearer the corrie floor, suggest later overriding by the same glacier that left the moraine deposits on the corrie floor. This interpretation largely supports that of Jarman et al. (2013), that the underlying shape and characteristics of the deposit are controlled by bedrock and moraine formation, with limited rockfall activity closer to the valley side. Ballantyne et al. (2009a) dated three boulders from
the feature. Given the reinterpretation, the dates now represent either moraine formation or a rockfall – both provide a minimum age for the last glaciation of the site.

Figure 4.3.43 Floor of unnamed corrie that holds Loch Etchachan (left), view looking west. Note the mounds, channels and glacially transported boulders on the corrie floor. On the far side the Carn Etchachan deposit can be seen (outlined by dotted line); note the continuation of the same moudny morphology onto the deposit

Figure 4.3.44 Note glacially transported boulders to the south of Loch Etchachan and, in the distance, one of the Ben Macdui plateau valleys that feeds into Loch Etchachan
4.3.5.2 Ben Macdui

The corrie that holds Loch Etchachan is relatively shallow, and a large area of the Ben Macdui plateau is on its southern and eastern sides (Figure 4.3.14). To the south of Loch Etchachan, above the boulders, the valley consists of boulder-covered ridges that in places appear to be bedrock cored (Figure 4.3.46). To the west of the stream, smoothed and glacially moulded bedrock exists; the smoothed up-valley and plucked down-valley sides of these features indicate ice flow from the up-valley direction (Figure 4.3.47).

The parallel valley situated to the north-west that contains Garbh Uisge Mòr is likewise covered in ice-moulded and scoured bedrock with areas of rock weathering (Figure 4.3.14 and Figure 4.3.48). Weathered bedrock is common, with smooth up-valley and upper surfaces, and leeside removal of blocks by ice indicating the former direction of ice flow from upslope (Figure 4.3.49) (Hall and Glasser, 2003). The present-day drainage from this valley is north towards Glen Avon; however, given its shallow form, under glacial conditions ice would have flowed both north towards Glen Avon and east towards Loch Etchachan (Figure 4.3.50). The valley that descends to Loch Etchachan possesses similar characteristics before its steep descent over bedrock to meet Loch Etchachan. These valleys are over 1050m,
sheltered to the north-east of the Ben Macdui plateau, with many snow patches still present in the summer months today. Although clearly glacially modified, probably through multiple glacial cycles, no evidence has been found to directly link the plateau valley to the last glaciers to have existed in Loch Etchachan, Glen Derry and Glen Avon below. However, it remains likely that thin cold-based ice would have existed in these valleys during a later readvance.

Figure 4.3.46 Ridged topography to the south of Loch Etchachan

Figure 4.3.47 Moulded bedrock with smooth upper sides and plucked down-valley sides (left), looking south towards Loch Etchachan (picture centre)
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Figure 4.3.48 Shallow valley of Garbh Uisge Mòr looking north towards Glen Avon

Figure 4.3.49 Weathered bedrock with leeside block removal and disruption by ice flow from the up-valley side (picture left)
Figure 4.3.50 Looking towards Loch Etchachan from saddle with Garbh Uisge Mòr

To the north-west, the valley that contains Garbh Uisge Beag drains into Glen Avon (Figure 4.3.14); this too contains outcrops and ridges of ice-moulded bedrock and weathered bedrock with leeside block removal. On the flatter plateau above this valley, less disturbed weathered bedrock and fewer blockfields exist with small weathering pits on the rock surfaces (Figure 4.3.51). This indicates reduced glacial erosion compared with the plateau valleys below and is consistent with the work of Hall and Glasser (2003).

Directly to the west of Glen Avon is the plateau valley that contains Fèith Buidhe (Figure 4.3.52). It is a good example of a preglacial valley, which drains over steep ice-moulded bedrock into the glacially eroded head of Glen Avon. The shallow valley does not contain evidence of glacial activity and is dominated by fluvial activity and snow slopes. There is strong evidence that ice from these plateau valleys has fed into Glen Avon; however, no evidence was found to support or disprove whether this occurred during the formation of the most recent sequence of moraines at the head of Glen Avon. The only tentative evidence may be the eroded clean slabs leading from the plateau into the valley that exhibit little rockfall debris and thus may suggest contemporaneous recent glaciation (Figure 4.3.53).
Figure 4.3.51 Weathered bedrock with small weathering pits on slabs in foreground. See insert for close-up of weathered pits (boot for scale)

Figure 4.3.52 Preglacial valley of Féith Buidhe, above Glen Avon. Note late-lying snow patches
4.3.5.3 Cairn Gorm

On the southern side of the plateau between Cairn Gorm and Cairn Lochan there are a series of shallow corries. The westernmost is Coire Domhain; the corrie floor is dominated by fluvially-modified surfaces and the western side by late-lying snow slopes (Figure 4.3.14). Within the floor of Coire Raibeirt there is an area of subtle ridges; however, these have been formed by postglacial channel incision. Up-valley peat and further watercourses are present. Shallow corrie sides rise above this and are covered in soliflucted surfaces extending up onto Cairn Gorm.

On the north-east side of Cairn Gorm, within Ciste Mhearad, there are subtle ridges and channels (Figure 4.3.35). Fieldwork showed the majority of these channels originate at the back of the corrie and mark the outlet of springs (Figure 4.3.54). At the time of fieldwork, a large bank of snow was retained in the hollow on the north-facing side; however, the corrie is shallow and has limited space for glaciation.
In summary, the slopes of the Cairn Gorm–Cairn Lochan section of the plateau are soliflucted and the shallow south-facing corries exhibit no evidence of renewed glaciation. Thus, unless thin cold-based ice existed, this section of the plateau does not seem to have undergone a later stage of glaciation.

The plateau valleys on the north and east side of Ben Macdui exhibit ice-moulded bedrock; however, the timing of this activity is unknown and may be from multiple glacial cycles. Evidence for late or renewed glaciation in the corrie that holds Loch Etchachan is stronger, with links to the upper glacial landforms in Glen Derry and Glen Avon. Whether ice in the Etchachan-Avon-Derry system is linked to wider cold-based glaciation of the plateau valleys and plateau summits is unclear. Given the shallow nature of the Loch Etchachan backwall, and the altitude and sheltered aspect of the plateau valleys north and east of Ben Macdui, it is probable ice was sourced from at least parts of the plateau. However, there is no direct evidence to suggest or dispute this.
4.3.6 Summary of interpretation: central Cairngorms

Apart from where described, no timing or chronology can be inferred from the order of the bullet points below; this is only a summary of the geomorphology and interpretation. Further discussion of the wider retreat patterns and schematic diagrams of the retreat margins can be found in Chapters 6 and 7.

- Spey ice lowered on the northern side of Cairngorms: marked by channels, moraines and kame terraces. A particularly noteworthy sequence of kame terraces associated with Spey ice occurs within lower Strath Nethy. There would also have been a corresponding stage of corrie glaciation marked by the outer moraines in Coire an Lochain (Cairn Gorm).

- In the southern Cairngorms, Derry ice initially flowed east over the col into Glen Quoich. The external ice from Deeside or ice sourced further west within the Cairngorms then became higher than the lowering Derry ice and dammed a high-level lake at 675m in Glen Derry, which drained over the Poll Bhàt col into a contemporaneous 600m lake in Glen Quoich. Almost certainly after the lake formation, the Derry glacier advanced into Glen Lui forming the moraines near Derry Lodge.

- The same damming ice sourced from the west is likely to have later formed the large moraine at the exit of lower Glen Luibeg. It is assumed the local deglaciation of Glen Luibeg, with its limited source area and link to the overtopping ice from Glen Dee, would also have occurred at a broadly similar time.

- The position of the Derry glacier’s northern outlet down Lairig an Laoigh is unclear at this time. It is thought the glacier would have been predominantly sourced from the corries and plateau of Ben Macdui, but it may also have included ice from the eastern Mòine Bhealaidh plateau. The overprinting of landforms at the Fords of Avon indicates the Derry glacier vacated this area first, followed by the Avon glacier, which retreated while actively forming moraines up Glen Avon. During this retreat, ice supply into Strath Nethy would have stopped.
A later stage of deglaciation or a readvance then occurred at the head of Glen Derry and Glen Avon, marked by the morphologically similar moraine systems. These glaciers were linked via the Loch Etchachan source area, and potentially by ice on part or all of the wider Ben Macdui plateau. The upper sections of both valleys possess closely spaced ice-marginal moraines, and the linking corrie floor has a high boulder concentration, typical of valley and corrie readvance elsewhere in the Cairngorms.

A later readvance occurred in Lochain Uaine (Ben Macdui), Coire an Lochain and Coire an t-Sneachda (Cairn Gorm) – all marked by high boulder concentrations and arcuate moraines. More subtle evidence for a corrie readvance exists in Coire na Saobhaidh (Glen Derry).

In the northern Cairngorms, there is evidence that there was an early deglaciation of Cairngorm valleys, while ice remained within the Spey valley. It is possible the outermost candidate moraines within Coire an Lochain (Cairn Gorm) may be from a similar time during ice-sheet deglaciation. On the southern side there is similar evidence that local glaciers retreated early. However, the source of the damming ice is less clear: it could have been external Deeside ice or locally sourced ice from further west in the Cairngorms. There is also a common pattern of deglaciation appearing from east to west on both the northern and southern sides of the Cairngorms.
4.4 Western Cairngorms

4.4.1 Lairig Ghru, Gleann Einich, Glen Feshie and Glenmore

The pattern of deglaciation in the lower part of the Lairig Ghru and Gleann Einich is well documented; no new fieldwork has been carried out and mapping from aerial imagery is consistent with previous studies (Brazier et al., 1998; Golledge, 2002). There is a large group of well-defined curved moraine ridges at the exit of the Lairig Ghru (Figure 4.4.1); these mark the former margin of the Glenmore lobe occupying the lower section of the Lairig Ghru. Up-valley, multiple terraces of different heights exist, with their upper surfaces gently sloping down-valley; these are delta deposits from a southerly input. Up-valley, particularly on the western side, large moraines and channel systems can be seen; these predate the lower lake deposits. However, the lower moraines, in addition to an ice-contact slope, on the southern side of the terraces (Brazier et al. 1998) suggest the Lairig Ghru glacier terminated in the lake.

On the western side, a meltwater channel originating from the col with Gleann Einich descends towards the lake deposits (Figure 4.4.1) (Brazier et al., 1998). Part-way down the channel at c.570m there is a small deposit, potentially with deltaic origins, and to the north, a candidate lake shoreline fragment (Brazier et al., 1998). This indicates that either the Einich glacier was at the col or an ice-dammed lake was present in Gleann Einich to supply water over the col.

On the eastern side of the Lairig Ghru, up-valley of the ice-dammed lake, there is a 200–300m wide boulder-covered deposit (Figure 4.4.1). Former interpretations include: snowbed activity (Sissons, 1979a), talus rock glacier (Maclean, 1991), and rock slope failure (Ballantyne et al., 2009a). Recently Ballantyne et al. (2009a: p. 23) recognised that the ‘boulder cover is draped over thick pre-existing valley fill (probably of till)’ that has been incised by the river. Aerial mapping here concurs that the underlying drift topography matches the wider undulating glacial drift in the valley and that the boulders are overlaid. One explanation for this is through postglacial RSF and rockfall events (Ballantyne et al., 2009a); the other through rockfall and supraglacial transport during the deglaciation of the Lairig Ghru (Jarman et al., 2013). Segments of the glacial drift can be traced up-valley; however, they are increasingly overlaid or obscured by more recent periglacial and
postglacial activity from the steep valley sides. This continues through the col to upper Glen Dee, where glacial drift and moraines become more prominent; these are discussed in Section 4.4.3.3.

The mapping in Gleann Einich is consistent with previous interpretations (Brazier et al. 1998; Golledge, 2002). At the exit of the valley, large moraine ridges mark the position of the Glenmore lobe invading the valley (Figure 4.4.1). On the southern side, multiple terrace levels gently slope to the north; these are deltas formed by meltwater from the Einich glacier. The Einich glacier terminated immediately north of Loch Mhic Ghille-chaoil, where terminal moraines can be followed into lateral moraines, particularly on the western side (Brazier et al., 1998). Here, multiple lateral moraines and meltwater channels mark the retreat of the Einich glacier. These continue up-valley until the valley floor becomes dominated by peat deposits north of Loch Einich. Up-valley, the steepening valley sides and lake-filled valley floor show little evidence of preserved moraines and are instead covered by gullied drift and alluvial fans. The valley head is predominantly covered in dissected drift, alluvial fans and fluvial surfaces, although none of the surfaces within upper Gleann Einich possess the well-developed solifluction surfaces of the neighbouring higher flanks of Braeriach. However, given the absence of moraines within the upper section of the valley, there is no evidence for a later readvance of ice.

The long profile of Gleann Einich has a very low gradient with the floor at c.500m for most of the valley, until it rises steeply onto the Mòine Mhòr plateau above (Figure 4.4.2). This would have implications for its glaciation, as the valley would require substantial lowering of the ELA before a glacier could exist. In addition, ice fed from the plateau would steeply descend/calve from the plateau edge down to 500m. The evidence on the surrounding plateau is described in Section 4.4.4.
Figure 4.4.1 Gleann Einich and Lairig Ghru geomorphological map
4.4.2 Northern and western corries of Braeriach

The Braeriach summit supports several corries of varying development; the geomorphological evidence for glaciation in each will be discussed from east to west. Common across the mountain flank are the well-developed soliflucted surfaces that occur on the spurs between the corries (Figure 4.4.3). Below c.700–800m the glacial landforms from the Gleann Einich glacier become dominant.
Within the lower section of Coire Beanaidh, on the eastern side, there is a limit to the altitude of the thicker drift (Figure 4.4.4). Both the drift and the valley side are heavily soliflucted and thus are associated with ice-sheet deglaciation. Immediately up-valley, several ridges occur to the east of the stream on the corrie floor; it could be tentatively suggested that one ridge is an arcuate lateral-frontal moraine from a glacier sourced within the corrie. However, these features are again soliflucted and most likely represent evidence of ice-sheet deglaciation. Opposite, below the western corrie wall, a 700m long boulder-covered ridge is present, which curves into the corrie side at its northern end (Figure 4.4.4). Within this curve, multiple inner ridges have been identified from aerial imagery. This feature was first interpreted as a talus rock glacier (Sissons, 1979a); Ballantyne et al. (2009a) indicated it may be an RSF but could not eliminate a rock glacier origin. Recent work and detailed discussion by Jarman et al. (2013) was not conclusive. Less clear deposits, that may be a continuation of the feature, occur across the south-western side of the corrie head. The surrounding corrie walls are covered by talus and alluvial fans. Although the possibility of a thin extensive Younger Dryas glacier is discussed by Jarman et al. (2013), the position of the controversial deposit, rock glacier or otherwise-interpreted deposit, particularly the upper segment, is not obviously conducive to formation by a corrie glacier. Thus a later readvance of ice is not thought to have occurred at this site, with only the controversial deposit potentially forming as a periglacial rock glacier and/or protalus rampart combination at this time. The palaeoclimatic significance of these different interpretations is discussed in Chapter 7: Section 7.2.3.
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Figure 4.4.4 Geomorphological map of the northern corries of Braeriach

Coire Ruadh contains soliflucted drift within its lower section, with no apparent moraine ridges (Figure 4.4.4). Higher, at an altitude of c.1000m on the eastern side of the corrie floor, a bench-like feature of boulder-covered mounds is present. Recently, the feature has been described as declining parallel with the corrie floor and having a sharp edge, with talus accumulation against the ice margin of a thin corrie glacier being proposed as a potential formation mechanism (Jarman et al., 2013). Formerly it has been interpreted as and remains a candidate rock glacier site (Ballantyne, 1996; Jarman et al., 2013). Within the upper corrie, immature talus and alluvial fans cover much of the walls and floor, potentially masking important glacial and periglacial evidence. Talus formation against a corrie glacier cannot be ruled out for the western deposit; however, other evidence to support or refute a corrie glacier and its timing is limited. Thus no or a limited readvance of ice is favoured at this site.

In contrast, the neighbouring Coire an Lochain contains extensive evidence for corrie glaciation (Figure 4.4.4). The outermost ridges associated with the corrie occur on both sides and descend to c.800m a.s.l., towards Gleann Einich below. These are surrounded by soliflucted surfaces and are most likely associated with
ice-sheet deglaciation. At c.900m a.s.l., lobate boulder and moraine ridges contrast with steep soliflucted surfaces that descend into the valley below. A lateral moraine and a sharp contrast with soliflucted surfaces mark the glacier’s extent on the east and west respectively. Many arcuate moraine crests can be traced across the corrie floor within this limit and no changes in solifluction development can be identified within the moraine area. The corrie walls are covered in immature talus and some solifluction can be seen on the plateau surface of Braeriach above. This evidence indicates a later corrie-glacier readvance, most likely sourced within the corrie. Although the moraines may represent multiple stages of glaciation, given their morphological similarity and the absence of different solifluction levels, one readvance event is favoured here (cf. Sissons, 1979a).

Within Coire Bogha-cloiche, large deposits exist on both sides of the corrie (Figure 4.4.5). Behind these, alluvial fans and fluvial activity cover the corrie floor. On the northern side, a 450m long deposit slopes down-valley, with an undulating upper surface that appears to be locally infilled by talus from above (Figure 4.4.6). It begins as a narrow feature at its uphill end, growing in prominence and relief towards its lower end. This feature has similarities with the Coire Ruadh deposit and may have similar talus-glacier origins (Jarman et al., 2013). Deposits also occur on the southern side but these may represent postglacial slumping. The shallow corrie headwall is covered in immature talus and fluvial activity. With the exception of the candidate talus-glacier deposit, no evidence for a readvance of a corrie glacier has been preserved, thus the presence of a small glacier during a later readvance cannot be ruled out.
Figure 4.4.5 Geomorphological map of the western corries of Braeriach

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The floor of Coire nan Clach is dominated by an alluvial fan descending from an area of gullied cliff erosion on the eastern side (Figure 4.4.5). To the north, the corrie side is covered by soliflucted surfaces, with the exception of a subtle ridge near the northern stream that drains the corrie. Below the alluvial fan, an area of boulders exists above more soliflucted surfaces. No large readvance occurred within this corrie; however, a small readvance cannot be eliminated as evidence may have been masked by alluvial-fan development.

The lower section of Coire Dhondail is dominated by soliflucted drift. Above on the western side, glacial drift, boulders and small subtle ridges are present, whereas the more central and eastern part of the corrie is dominated by alluvial fans and gullying activity (Figure 4.4.5). Within the glacial deposits, the most prominent ridge is an arcuate ridge that originates on the western side, curves to the centre of the corrie at 800m a.s.l., and ends with the large boulder seen in Figure 4.4.7. This ridge is interpreted to mark the margin of a corrie glacier. Up-valley of this ridge a higher concentration of boulders exists, and closer to the corrie backwall further arcuate ridges and boulder lines exist. Determining the eastern limit of this stage of
corrie glaciation is significantly hindered by the infilling from fluvial activity. It is possible that the ridge followed by the footpath may be the corresponding eastern limit (Figure 4.4.8). The corrie backwall is covered in immature talus, and limited solifluxion occurs within the main ridge limit. This is new evidence for a corrie glacier within Coire Dhonail; although more subtle, the evidence is morphologically similar to other sites of corrie readvance. The corrie also highlights the potential for postglacial slope processes to mask glacial evidence, particularly where the evidence is close to corrie walls.

Figure 4.4.7 Coire Dhondail – arcuate outer ridge curving to large boulder at end
4.4.3 Glen Dee, Glen Geusachan and Garbh Coire

The Glen Dee valley and its tributaries can be split into three distinct areas: lower Glen Dee, Glen Geusachan and Garbh Coire (upper Glen Dee).

4.4.3.1 Lower Glen Dee

At the lower end of Glen Dee, multiple ridges with shallow crests trend obliquely up-valley on the western side (Figure 4.4.9 and Figure 4.4.10). Opposite, on the eastern side, a glacial-drift limit obliquely descends northwards; the lower part of this drift is now heavily dissected by gullying. This evidence is interpreted to mark ice flow from the south, from the main Dee valley pushing north into the exit of Glen Dee.

Adjacent to this glacier margin and to the north, horizontal lines have been identified on both the eastern and western sides of Glen Dee. The lines are most prominent at 576m on the western side of the valley and can be followed more or less continuously for 2.5km (Figure 4.4.9 and Figure 4.4.11). The line is less continuous on the eastern side and is largely confined to segments adjacent to the southern moraines described above (Figure 4.4.12), and a 300m segment just
beyond the lowermost moraines and meltwater channels at the exit of Glen Geusachan (Figure 4.4.9). The features’ bench profile, with matching altitudes on opposing sides of the valley and for their full length, indicate these are lake shorelines from a former ice-dammed lake. A potential explanation for the absence of the lake shoreline for much of the eastern valley side is that the steeper topography may be less conducive to lake shoreline formation and survival. The clarity of the most prominent lake shoreline at 576m and the altitudinal link with the Meirleach col indicate the lake continued north to the Meirleach col where it drained east towards Derry Lodge.

However, preserved lake shoreline evidence stops on both sides of the valley at the onset of the moraines and meltwater channels south of the exit of Glen Geusachan. This indicates a glacier existed in this part of Glen Dee, either at the time of lake formation or at some point after the lake drained; this is further discussed in Section 4.4.3.2. Based on the assumption that ice readvance was limited to corries, plateaus and the upper valleys in the Eastern Highlands, it is assumed the lake shoreline was formed by ice damming lower Glen Dee during ice-sheet retreat, prior to the Lateglacial Interstadial. Thus the upper end of the lake shoreline can act as a down-valley limit for any potential later readvance of ice.
Figure 4.4.9 Glen Dee geomorphological map
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Figure 4.4.10 Lower Glen Dee – subtle moraine ridges obliquely descending up-valley, marking the position of glacier flow from the south (left) northwards (right)

Figure 4.4.11 West side of Glen Dee. Lake shoreline cutting beneath large pre-existing moraine deposit
Near Alt Garbh, a large ridge deposit can be seen at c.600m on the western valley side (Figure 4.4.11 and Figure 4.4.13). On its eastern side, shallow lakes and peat exist between the ridge and the valley side, and multiple channels run between the deposit and valley side. On its western side, the ridge falls steeply to the valley below and the lower Glen Dee lake shoreline can be seen cutting across the lower slopes of the ridge. The ridge becomes shallower at its southern end, where it starts to descend towards the valley floor, but becomes dissected and modified by fluvial activity. The granite boulders along the ridge’s full length, the crest geometry and meltwater channels on the western side indicate this feature is likely to have been formed by a glacier flowing south. However, no further evidence of this margin has been found on the valley floor or the opposing valley side. The lake shoreline described above cuts across the lower part of the moraine and continues a further 1km towards the exit of Glen Geusachan, thus this feature must predate the ice-dammed lake (Figure 4.4.11).
4.4.3.2 Glen Geusachan

Extending from approximately 400m south of the confluence of the River Dee and Geusachan Burn, an area of outwash, palaeochannels and kettle holes continues for almost 3km towards the exit of Glen Dee (Figure 4.4.9). However, this area consists of multiple different outwash areas potentially from different glacial events (W. Mitchell, pers. comm.). In the lower section, a terrace with extensive modification by palaeochannels is present on the eastern side of the River Dee, while on the western side a terrace with numerous kettle holes is present (Figure 4.4.14). To the north of the confluence of the Caochan Roibidh stream and the River Dee, a lower terrace covered in large granite boulders and kettle holes continues north, mainly on the eastern side of the River Dee, until it meets an area of highly concentrated moraines on the valley floor at (57.018323° N, 3.674755° W) (Figure 4.4.15, Figure 4.4.16 and Figure 4.4.17). The large nature of the boulders on the lowermost terrace and their continuation down-valley suggests they were released from high energy events, from a contemporaneous glacier margin (Mitchell and Tipping, unpublished report, National Trust for Scotland). Up-valley, the valley floor is covered by the present-day floodplain; there is no continuation of a terrace with kettle holes or a high concentration of boulders. The difference in terrace height and distinctive change in boulder concentration between the northern and
southern terraces indicate there are at least two different outwash-terrace formation events. The linking with the glacier moraine evidence and further interpretation is provided later within this section.

Figure 4.4.14 Lower Glen Dee – southerly outwash terraces on either side of the River Dee

Figure 4.4.15 Glen Dee – boulder-covered lower terrace (looking south)
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Figure 4.4.16 Glen Dee – in foreground the boulder-covered terrace with kettle holes on left-hand side, behind this the terrace ends with the area of moraine ridges on the valley floor

Figure 4.4.17 Glen Dee – panoramic photo of the boulder-covered terrace on the nearside of the River Dee and the predating moraine ridges and meltwater channels on far side
The moraines and glacial landforms within Glen Dee, at the exit of Glen Geusachan, can be divided into multiple groups. At the southern end, at the northern limit of the lake shoreline, a boulder-covered moraine ridge starts on the western valley side and descends towards the upper end of the southern higher outwash terrace, indeed, only being separated by the postglacial incision of the Caochan Roibidh stream. On the opposite side of the valley, moraine ridges and meltwater channels descend obliquely across the valley side towards the valley floor (Figure 4.4.17).

These continue up-valley on the eastern side until opposite Glen Geusachan, where the glacial evidence is dominated by large meltwater channels cut into drift. The longest continuous channel runs obliquely across the valley side for 1500m. The most prominent are full double-sided channels, whereas many are single-sided features, suggesting these were ice-marginal channels with ice supporting the western side. The channels appear to be cut into a pre-existing surface, rather than being associated with moraine building as previously interpreted (Bennett and Glasser, 1991; Everest and Kubik, 2006). No individual moraine crests are seen in this area; instead the meltwater channels dominate an otherwise flatter area of glacial drift and glacially transported boulders (Figure 4.4.18). A single particularly large channel crosses the Meirleach col, continuing east towards Glen Luibeg. Channels and glacially modified surfaces can be seen on the northern side of this channel for the first 600m. Above the main channels, higher on the Carn a’ Mhaim mountainside, the landforms are soliflucted but channels may have continued up onto this slope. Inside the most prominent channels, fainter channels and linear lines within the vegetation mark the former position of meltwater channels; these continue towards the Glen Dee valley floor. Combined with the evidence for a water-terminating glacier margin further up Glen Dee (Midgley, 2001) and some small candidate lake shorelines, these channels are thought to mark the drainage of a lake in upper Glen Dee, around the damming Glen Geusachan ice (Bennett and Glasser, 1991).
Figure 4.4.18 Glen Dee opposite Glen Geusachan, looking south. Large meltwater channels incised into an area of glacial drift. Note pre-existing flat inter-channel topography

Just north of the Geusachan and Dee intersection, the drift limit descends out from Glen Geusachan north into Glen Dee. Moraines within this area also have subtle arcuate form and their orientation suggests former ice flow from Glen Geusachan north up Glen Dee. This evidence stops c.700m short of Corrour Bothy, indicating the limit of the former glacier.

Inside these outer limits, an inner area of more prominent closely spaced ridges has been identified. Their down-valley southern limit occurs c.600m south of the confluence of the River Dee and Geusachan Burn. This coincides with the upper limit of the boulder-covered outwash terrace. A short distance up-valley, at the confluence of the River Dee and Geusachan Burn, lateral moraines on the eastern valley side are particularly well preserved.

Immediately to the east of this moraine area, a wide shallow channel runs north to south circumventing the moraines, ending near the present River Dee channel. This is interpreted to mark the deflection of the River Dee around the margin of the Glen Geusachan glacier that formed the inner moraines described immediately above. The north-eastern limit of this inner moraine area is less clear, but is interpreted to be marked by the edge of the prominent ice-marginal moraine ridges to the south of Devil's Point (Figure 4.4.19). Deposits of similar appearance do exist outside this
limit; however, their relief is often enhanced by gullying and they are more sporadic rather than marking closely spaced glacier margins.

Figure 4.4.19 Exit of Glen Geusachan. Note change between small moraine-ridge-dominated system in foreground, and the meltwater-channel incision behind

The glacial landforms at the exit of Glen Geusachan all seem to have been formed by a glacier sourced within Glen Geusachan, rather than a glacier sourced in upper Glen Dee. The absence of evidence for a glacier margin terminating in the southern Glen Dee lake, and the presence of meltwater channels descending to the valley floor rather than terminating as deltas at the lake level, indicate that the outer and inner glacier margins were both formed by later readvances of ice, subsequent to the drainage of the southern lake. The outer margin, dominated by meltwater incision on the eastern side, and moraine and meltwater-channel formation on the southern side, is seemingly linked to the northern end of the higher outwash terrace. The inner margin, within which closely spaced moraine crests mark ice-margin retreat, terminates at the start of the inner, lower, boulder-covered outwash terrace. These two margins may represent two different stages of glaciation or the retreat of one glacier. The interpretation as two different glaciers can explain the
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change to more prominent moraine building and a second inner outwash surface. However, the landform evidence on the eastern margin may have been heavily impacted by the drainage of the upper Glen Dee ice-dammed lake around the glacier margin. Thus some internal change in the landsystem would be expected to have occurred when the Geusachan glacier no longer dammed an ice-dammed lake which drained around its eastern margin. The closely spaced ice-marginal moraine system, lack of soliflucted surfaces and inner boulder-covered terrace indicate the inner margin may represent a later readvance. Whether the outer margin also represents the same readvance or earlier ice-sheet deglaciation is uncertain. The discussion of these sites in relation to new and existing cosmogenic surface exposure ages continues in Chapter 5.

Within the exit of Glen Geusachan, sharp-crested moraine ridges continue up-valley. These are more sporadic, but this may be due to the narrower valley floor making fluvial erosion and alluvial-fan formation less favourable for moraine preservation. The north-facing Coire Cath nam Fionn contains heavily dissected drift; this hinders identification of individual moraine crests and their orientation, so no evidence for ice retreat into the corrie or separate corrie glaciation exists. However, the drift thickness and appearance matches that of the surrounding Glen Geusachan valley sides and it is likely the corrie would have contributed ice to the Geusachan glacier.

At the head of Glen Geusachan a higher concentration of oblique moraine ridges and glacial features can be identified (Figure 4.4.20 and Figure 4.4.21). It must be noted that the moraines and glacial drift continue outside this area, but a change in their clarity makes them more easily identifiable. This may be due partly to their original form and partly to extensive gullying and the steeper slopes outside this area. The lower end of the moraine group is marked by a broad deposit which has been incised by the Geusachan Burn. Behind this deposit, on both sides of the valley, closely spaced oblique moraine ridges are present, which have been crossed vertically by postglacial gullying (Bennett and Glasser, 1991). The crests of these moraines mark ice-marginal retreat to the north into the head of Glen Geusachan; these become less organised in the upper reaches of Allt Clais an t-Sabhail. Above the moraines, on the eastern side, soliflucted surfaces are present, suggesting the south-west-facing slope above the glacier was ice free. On the
western side, between the head of Glen Geusachan and Loch nan Stuirteag, a large 450m long north-to-south orientated ridge is found (Figure 4.4.20 and Figure 4.4.22). At its higher northern end, it abuts the spur, and at its southern end, it curves into Glen Geusachan. On its western side, smaller moraine mounds with less obvious orientation cover the ground towards Loch nan Stuirteag. On the northern side of Loch nan Stuirteag, at a height of 895m, a horizontal bench in the drift is interpreted to be a lake shoreline based on its continuity, horizontal form and cross section (Figure 4.4.23) (Bennett and Glasser, 1991). The height of the lake shoreline is consistent with the height of the col through which it drained to the west. Thus the lake, dammed by ice in upper Glen Geusachan, must post-date deglaciation of the plateau to the west. The lake shoreline is not seen on the southern side; this may be due to the prevailing wind favouring lake shoreline formation on the northern side or the presence of ice or snow on this more sheltered north-east-facing side. The lake shoreline is continuous and has not suffered the solifluction experienced by lake shorelines elsewhere in the Cairngorms. This suggests more recent formation, and the local relative chronology concurs that Glen Geusachan was one of the last places to support a glacier. The lack of solifluction, both within the glacier margin and to the lake shoreline adjacent to upper Glen Geusachan, suggests the site may represent a glacial readvance. It is important to note here that to dam the lake adjacent to upper Glen Geusachan requires a substantial amount of ice in Glen Geusachan, not just ice within the valley head. An area of interest is how the evidence in Glen Geusachan relates to meltwater-channel and moraine evidence in the shallow plateau valley to the west; this is discussed in Section 4.4.4.
Figure 4.4.20 Upper Glen Geusachan geomorphological map
Figure 4.4.21 Head of Glen Geusachan. Oblique moraine crests and vertical cross-cutting by gullies. Solifluction surfaces can be seen above the moraines. Photo taken from Monadh Mòr

Figure 4.4.22 Upper Glen Geusachan. Large ridge on the western side of Upper Glen Geusachan. Smaller moraines can be seen in Glen Geusachan below and towards Loch nan Stuirteag above
Glen Geusachan is surrounded by high plateau above 1000m, with Monadh Mòr to the west and Beinn Bhrotain to the south. Monadh Mòr is covered by blockfields and thin regolith on the top, and ice-moulded bedrock, immature talus and snow patches as the slopes steepen and descend into Glen Geusachan (Figure 4.4.20). It is also noticeable in the field and by looking at the contours that the eastern side of Monadh Mòr has a sheltered corrie named Creagach, which has late-lying snow patches and may have been important for snow accumulation.

The summit area of Beinn Bhrotain is covered in blockfields, sandy regolith and wind-scoured surfaces (Figure 4.4.20). The northern slopes of Beinn Bhrotain have bedrock-cored ridges and areas of bedrock that have suffered glacial erosion and modification. There is no evidence of solifluction on either the plateau to the west or south of Glen Geusachan, unlike the soliflucted surfaces on the opposite south-west-facing slopes of Stob Coire an Saighdeir. This may indicate the later presence of ice cover on these plateaus, or at least on the sheltered northern and eastern sides. These large plateaus would have been important for either snow blow or plateau-ice accumulation, and also shading the deeply incised Glen Geusachan below.

In summary, the continuous glacial drift and the sporadic preserved moraine ridges indicate that the evidence in upper Glen Geusachan and Coire Cath nam Fionn
marks the retreat of the same glacier that formed the inner closely spaced moraine ridges south of Devil’s Point. This fits with the lake shoreline adjacent to the valley head, which requires a substantial ice mass to be present within Glen Geusachan. This glacier would have been sourced from upper Glen Geusachan, Coire Cath nam Fionn, Coire Creagach and would most likely have received plateau ice from Monadh Mòr and Beinn Bhrotaín. In addition, the plateaus and sheltered northern and eastern slopes may have been important for snow redistribution.

4.4.3.3 Upper Glen Dee

Within Glen Dee, to the north of the lobate moraines left by Glen Geusachan ice, the valley floor is covered by flat areas of peat, consistent with this once being the bottom of an ice-dammed lake. Glacial drift can be seen above the peat in isolated areas, such as 500m north of Corrour Bothy, but this section of the valley is dominated by gullying and alluvial fans which encroach on the valley floor.

On the western side, several corries have been eroded into the plateau between Devil’s Point and Cairn Toul (Figure 4.4.24). The southernmost, Coire Odhar, does not contain moraine or meltwater-channel evidence; instead drift containing little morphological expression, but most likely of glacial origin, is well dissected by gullying and locally covered by fluvial deposition.

To the north, the higher and better developed Coire an t-Saighdeir contains a well-defined area of glacial deposits, the most prevalent being a 500m long continuous arcuate ridge that starts on the northern side and curves round to the mouth of the corrie at 850m a.s.l. (Figure 4.4.24). This limit continues on the southern side as a limit of highly concentrated boulders and ridges, which also continue further back within the corrie. Talus accumulations mask much of the upper sides of the corrie, but based on the moraine geometry the glacier seems to have been sourced from the south-west side. Below the corrie mouth, boulders are less common and the drift is gullied with only subtle signs of solifluction. However, given the abrupt boulder limit and continuous arcuate moraine ridge, this is interpreted to mark a later glacier readvance.

Immediately north, the small but high Coire an t-Sabhail is less well developed, with a corrie floor above 1100m. No clear moraines are mapped here. However, the
upper part of the corrie is covered in a moundy irregular deposit, covered in boulders. This could have many origins, but given the altitude and aspect of the corrie, glacial or periglacial activity is probable during former periods of climate cooling.

Figure 4.4.24 Geomorphological map of the corries on the western side of Glen Dee
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Figure 4.4.25 Upper Glen Dee geomorphological map

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Towards the upper end of Glen Dee, a straight 600m wide, 200m broad, shallow ridge fills the valley floor (Figure 4.4.25 and Figure 4.4.26). Behind the ridge a flat area of fluvial surfaces exists, likely to have been formed by sedimentation in a lake, while the River Dee incised through the moraines’ eastern side. The ridge has previously been interpreted to be the grounding line of a glacier sourced from Garbh Choire, terminating in a lake, dammed by ice from Glen Geusachan (Midgley, 2001). The size and clarity of the feature suggests the glacier margin was stable for some time. A few moraines exist just 100–200m beyond the main margin, and up-valley they quickly turn to being smaller individual moraine ridges.

![Linear moraine ridge in upper Glen Dee (up-valley side on the right)](image)

**Figure 4.4.26 Linear moraine ridge in upper Glen Dee (up-valley side on the right)**

Above this margin on the eastern valley side, there is a distinct limit between the soliflucted drift and talus surfaces, and the lower glacial drift and moraines that descend down-valley towards the linear moraine. On this eastern valley side, moraine crests descend obliquely towards the valley floor at orientations of c.180 to 200 degrees. The orientation of these moraines, particularly higher on the valley side, is more consistent with ice flow from the north, rather than ice sourced only from Garbh Choire to the north-west. Some of the steepest descending moraines, which indicate ice flow from the north, are just south of the alluvial fan from Allt a’ Choire Mhòir. However, to the north of the fan, moraines are not present on the eastern side, with the exception of subtle features and a candidate kettle-hole feature. Many of the lower moraines on the eastern side towards the stream Allt na
Lairig Ghru and the River Dee indicate ice flow from Garbh Choire to the west, and in places their orientation is less clear.

To the west of the Allt na Lairig Ghru, near the Lairig Ghru-Dee col, ridges are subtle and postglacial processes dominate. Down-valley as the valley widens, on the central spur of the valley, moraine ridges and particularly meltwater channels can be seen curving round from the west towards the south (Figure 4.4.27 and Figure 4.4.28). This section is interpreted to be good evidence for the retreat of a glacier sourced in Garbh Choire to the west. On the west side of the spur, there are channels that drain south obliquely across the side of the spur; these are interpreted to mark a slightly later stage of retreat when the glacier no longer fed ice across the spur. Below these a curved band of granite boulders is present, the origin of which is up-valley where it merges with the drift limit extending from Garbh Choire Dhàidh (Figure 4.4.25). The boulders’ down-valley termination is near the confluence of the Dee and Lairig Ghru streams. For the central part of the band’s length, the boulders are associated with a large ridge located between the central spur and the River Dee (Figure 4.4.29); this is interpreted to be a lateral moraine from a glacier sourced in Garbh Choire Dhàidh. This ridge terminates above the confluence of the River Dee and Allt na Lairig Ghru streams, with a small continuation of the ridge on the south-east side of Allt na Lairig Ghru. This last segment of the ridge is eroded on the north-west side by the stream and on the north-east side by palaeochannels. South-east of the palaeochannels, a shallow slope covered in similar large boulders exists. Its proximity to the boulder-covered lateral moraine indicates these too may have a glacial origin; it is tentatively suggested they could mark a small outwash area from the same glacier margin.

Up-valley, inside the lateral band of boulders, sharp-crested closely spaced moraine ridges occur, with a similar set of moraines present to the south of the River Dee. Together these moraines mark the retreat of a glacier into the Garbh Choire corries. Bennett (1996) tentatively suggested the straighter moraine crests may not record the former ice margins, instead they may mark thrusts when the glacier compressed against the valley side. Down-valley moraines on the southern side are best preserved on higher areas of the valley floor; this may mark the overprinting of moraines on larger underlying deposits.
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Figure 4.4.27 Upper Glen Dee. Moraines and meltwater channels curve from west to south (see arrow). Ice sourced from Garbh Coire behind.

Figure 4.4.28 Central spur of upper Glen Dee. Foreground – meltwater channel cut into moraines on the spur. In the distance, the steeper moraines on the western side of Glen Dee can be seen.
Figure 4.4.29 Boulder-covered lateral moraine to the north-east of the River Dee, looking towards the linear moraine in the distance

On the northern side of Garbh Choire is the well-developed Coire Bhrochain, which descends via a steep corrie lip into Glen Dee below. At the mouth of the corrie, at 925m a.s.l., a large arcuate moraine ridge is present marking a corrie-glacier limit, behind which several other moraine ridges can be identified, although none as prominent as the outer ridge (Figure 4.4.25). The corrie-glacier moraine and boulder area forms a distinct contrast with the talus and soliflucted surfaces surrounding the outer limit. Below the corrie-glacier moraines, the area of soliflucted drift contains subtle oblique moraine ridges that descend towards Glen Dee below. Particularly, the upper moraine ridges indicate an earlier more extensive period of glaciation, with ice flow from the corrie into Glen Dee. The geometry of the moraines below, within Glen Dee, does not indicate contributing ice from Coire Bhrochain. Indeed the band of boulders and associated ridge, described above, could tentatively be suggested to continue up-valley below Coire Bhrochain. Along with the solifluction evidence, this indicates the contribution of ice from Coire Bhrochain into the Dee valley predates the formation of at least the boulder-covered lateral moraine and the moraines inside it. The solifluction development indicates the more extensive Coire Bhrochain glaciation was prior to the most recent readvance when the Bhrochain glacier only reached the corrie lip.
To the west lies the slightly lower Garbh Choire Dhàidh, with a corrie-floor altitude of 900–950m. The distinctive features on the corrie floor are the parallel flutes and boulder lines that record the former ice-flow direction towards the exit of the corrie (Figure 4.4.30). The edge of the corrie is undulating with moraine mounds covered in boulders; they are particularly prevalent on the eastern side, where they continue into Glen Dee. At the back of the corrie a shallow ridge crosses the corrie floor (Figure 4.4.31). The origin of this is unclear; however, a line of large boulders crosses the ridge running from the backwall towards the mouth of the corrie. This would suggest the ridge is an older feature and has been modified by later ice flow. The evidence within this corrie suggests ice flow into Glen Dee, where it joined the probably more dominant flow of ice from Garbh Choire Mòr to the west.

Figure 4.4.30 Flutes in Garbh Choire Dhàidh, looking towards Glen Dee
Situated at approximately the same altitude, just to the south-west is Garbh Choire Mòr. Several moraine-ridge fragments exist at its exit; these mark the retreat of a glacier that once created the moraines further down Glen Dee. In the highest, north-west corner of the corrie, a boulder ridge can be seen below the backwall (Figure 4.4.32). It consists of a mixed size of angular boulders, with an open framework and limited finer material. The landform appears relatively fresh in appearance and is unstable to walk on (Figure 4.4.33). It has been suggested elsewhere this moraine may have been formed during the Little Ice Age (Rapson, 1990; Gordon, 1993; Harrison et al., 2014). While the feature’s morphology indicates it is a moraine, the fresh appearance may be in part due to the supply of rockfall and avalanche debris to the upper side of the feature. This poses the question as to whether the underlying ridge has closer relations to the ridge at the back of the neighbouring Garbh Choire Dhàidh (see Figure 4.4.31). Detailed geomorphological work will be required in this remote location to identify the process and timing of its formation.
The final corrie within upper Glen Dee is to the south, Coire an Lochain Uaine, with the corrie lake at an altitude of 910m. Much of the corrie floor is masked by the lake; however, at the corrie mouth, glacial drift with glacially transported boulders is present until the corrie steeply drops over bedrock into Glen Dee. The corrie walls...
are covered in immature talus, and soliflucted deposits are not found within the corrie. It is most probable that ice from this north-facing corrie would have contributed to the Glen Dee glaciers.

In summary, upper Glen Dee is likely to have initially joined with the wider Cairngorm system of glaciers, including supplying ice to the Lairig Ghru glacier to the north. Later the linear margin, combined with the highest glacial-drift limit, marks a readvance or period of stability. During this time the glacier was most likely sourced from all the corries to the west and possibly also from the northerly direction, given the orientation of some of the higher moraines on the eastern side: this marks phase 1. Within these limits, the moraines and meltwater channels sourced from only the west have an outer limit; this is partially across the central spur and onto the lower western valley side: this marks phase 2. Inside this, the lateral moraine and band of glacially transported boulders that post-date ice contribution from Coire Bhrochain represent phase 3. The moraines within each limit are not dissimilar; however, inside phase 3 the moraines show greater alignment into continuous ridges, whereas the relief of the phase 2 landforms is more dominated by meltwater-channel incision. Whether the phases mark a single retreat or multiple readvances is unclear, but here it is favoured that a later readvance glacier did exist within upper Glen Dee and reached at least phase 3, while a simultaneous corrie glacier existed within Coire Bhrochain.

Given the ice-dammed lake link between the Geusachan and Upper Dee glaciers, the evidence needs to be reviewed together. It is important to note that to create even a shallow water-terminating glacier in upper Glen Dee, the lake would need to be c.600m a.s.l. This is above the Meirleach col and thus requires not only the Geusachan glacier to be near the large meltwater channels at c.575m but at a higher more extensive position opposite Glen Geusachan. There is agreement that the outermost limits associated with the ice-dammed lake in upper Glen Dee and Glen Geusachan may not be the final glacier margins to have existed in the valleys. Instead a more restricted readvance phase may have occurred in both Glen Geusachan and upper Glen Dee; this is investigated further with reference to new and existing surface exposure ages within Chapter 5.

4.4.4 Mòine Mhòr
Figure 4.4.34 Geomorphological map of the Mòine Mhòr plateau

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The western plateau possesses the most evidence of deglaciation of the four main Cairngorm plateaus (Figure 4.4.34). The landforms mainly consist of subtle moraines, ice-moulded bedrock and more distinctive meltwater channels. Firstly, there is an area of ice-moulded bedrock, meltwater channels and glacial drift on the plateau above the head of Gleann Einich and to the north of Loch nan Cnapan. These features start c.1500m south of the plateau edge and converge towards the valley head, before disappearing as the plateau descends steeply into Gleann Einich. To the east, a large area of peat covers any evidence, but nearer the plateau edge before it descends to Coire Odhar nan Each and A’ Phòcaid, meltwater channels have been mapped. To the south of Loch nan Cnapan, the channel orientation is north to south. The channels’ position relatively far south on the plateau suggests these may be of subglacial origin when external ice flowed over the plateau towards Gleann Einich from the south.

Channels to the east of Loch nan Cnapan curve around the lower slopes of Carn na Criche from the east, towards the head of Gleann Einich (Figure 4.4.34). The favoured interpretation of these channels is that they are ice marginal and mark the thinning and retreat of a glacier towards the north-east section of the plateau. Linear features also exist on the northern flank of Monadh Mòr; these slope from south-east to north-west obliquely across the hillside. These features, assumed to be meltwater channels, may be subglacial, but if lateral, they indicate ice flow from the head of Glen Geusachan westwards onto the plateau.

Within the plateau valley that contains Allt Luineag, an area of moraines and drift, heavily shaped and dissected by well-defined meltwater channels, is present. The channels start on the northern side of the col to the east of Loch nan Stuirteag (Figure 4.4.35 and Figure 4.4.36) and continue west, gradually descending while following the gentle contours of the topography, before joining the stream on the valley floor (Figure 4.4.34 and Figure 4.4.37). Some of the highest channels may continue to link with the channels above Gleann Einich. It should be noted that these channels are some 150m lower than the previously described channels on the northern flank of Monadh Mòr. It is also important to clarify that these channels start up to 15m above the col that separates Glen Geusachan and the plateau – it is
only the lowest channel that is associated with the lake shoreline above Loch nan Stuirteag. Instead, the higher channels must predate the lake, unless the lake was more variable in height and was dammed by ice both on the Glen Geusachan and the plateau side, thus draining around or underneath the plateau glacier. However, there are no shorelines at higher altitudes to support this.

North, above this area of meltwater activity on the plateau valley, heavily soliflucted lobate boulder lines can be seen; at their lower end they are more continuous bands of small boulders (Figure 4.4.35 and Figure 4.4.38). Some of these features may have formed through solifluction, but their position may mark the earlier lateral limits of a lowering plateau glacier. Solifluction can also be seen on the west-facing side of Monadh Mor.

Figure 4.4.35 Plateau valley that contains Allt Luineag. Note meltwater channels on right that begin on col with Glen Geusachan just out of the photo; these curve to join channels in the left of picture. Higher up, see the soliflucted boulder lobes that become straighter at their flatter down-valley end
Figure 4.4.36 Meltwater channel on the col to west of Loch nan Stuirteag, looking west into plateau valley

Figure 4.4.37 Meltwater channel north-east of Allt Luineag, looking south
In summary, the glacial evidence above Gleann Einich represents large plateau glaciation with ice flow from the southerly direction, either sourced locally or including external ice. The landforms within the plateau valley that contains Allt Luineag indicate local ice flowed south and west towards Loch nan Cnapan. The higher boulder bands and lower meltwater channels mark the retreat of a glacier, while water flowed around the glacier margin sourced from the col with Glen Geusachan. The ice-dammed lake associated with the lake shoreline adjacent to upper Glen Geusachan must post-date the earlier plateau landforms. The solifluction of the slopes surrounding the plateau suggests widespread large plateau glaciation did not occur during the latest period of climatic deterioration; while the evidence to the west of the col with Glen Geusachan may mark a continuation of the Geusachan glacier onto the plateau. The seamless continuation of meltwater channels into the higher soliflucted slopes and towards Gleann Einich suggests this marked the retreat of ice during ice-sheet deglaciation as previously interpreted by Bennett and Glasser (1991). Thus the lake shoreline and distinctive change to closely spaced moraine mounds near Loch nan Stuirteag are interpreted to mark a potential readvance limit.
4.4.5 Glen Eidart and south-west flanks of the Cairngorm Mountains

The southern flanks, particularly to the west of the River Eidart, are covered in large ice-sheet deposits and meltwater channels. These meltwater channels descend obliquely to the west, with some of the highest being above 800m immediately to the west of Glen Eidart (Figure 4.4.39). Numerous similarly orientated channels below record the lowering of the ice during deglaciation. Similar channels can be seen on the western side of the plateau, descending north towards and beyond the exit of Coire Garbhlagh.

Glen Eidart is covered in small closely spaced moraine mounds and ridges that extend c.2.5km from the valley head. Outside this, the moraines become less well defined and the surrounding mountain flanks are covered by the ice-sheet landforms described above. Detailed analysis by linking the individual moraine fragments within Glen Eidart has been used to create palaeo-ice margins of retreat (Figure 4.4.40 and Figure 4.4.41). This indicates the retreat of a glacier up-valley, and then the retreat into Coire Mharconaich, marked by arcuate lateral-frontal
moraines that extend out from the corrie sides. Oblique lateral moraine ridges are also found on the northern and southern sides of the north-west valley head (Figure 4.4.42). The orientation of the lateral moraines indicates that part of the Eidart glacier retreated into the north-west valley head. The plateau channels to the west may be linked to such an ice mass or may represent earlier glaciation or postglacial activity (Figure 4.4.41).

The moraines do not continue northwards onto the plateau or north-east up the narrow valley towards Tom Dubh, suggesting the glacier in Glen Eidart was sourced within the corrie and the north-west valley head. To the north of the valley on the plateau edge, dry meltwater channels exist (Figure 4.4.43). The narrow channels are deep, steep-sided, and are orientated east to west; however, they are unlikely to represent the same stage of glaciation as the landforms c.150m below within the valley.

The landforms within Glen Eidart mark the retreat of a glacier predominantly sourced from Coire Mharconaich, but also from the north-east valley head and potentially the plateau behind. Solifluction is absent within this area and the moraines appear particularly well preserved, thus a later readvance at this site is favoured.
Figure 4.4.40 Eastern side of Glen Eidart, looking north up-valley. Hill-shaded DEM self-generated from ground-based photographs. Note the gently curved former ice-marginal positions recorded by moraine ridges and meltwater channels.
Figure 4.4.1 Geomorphological map and palaeo-ice-margin positions. Note glacier retreat both into the corrie and the north-west valley head
Figure 4.4.42 Moraines on northern valley side, within the north-west tributary valley, at the head of Glen Eidart

Figure 4.4.43 Meltwater channel at the head of Glen Eidart, on the corner with the north-east tributary valley. Looking west towards Coire Mharconaich and the north-west tributary
4.4.6 Coire Garbhach and western flanks of the Cairngorm Mountains

Coire Garbhach descends from the western Mòine Mhòr plateau to join Glen Feshie. At its termination in Glen Feshie, the Feshie valley side is covered in thick glacial drift and meltwater channels that descend northwards, marking the southerly retreat of ice during ice-sheet deglaciation. Coire Garbhach is a steeply sided valley with talus, gully ing and alluvial fans covering the slopes, many still active today. Within the upper part of the valley, on the north-east-facing slope, two limits of thick deposits have been identified, below which the river has incised to create a further terrace surface (Figure 4.4.44 and Figure 4.4.45). Above these, talus slopes are present and have covered part of the uppermost deposit limit for a short section. The upper limit consists of a ridge that descends down-valley, behind which rockfall debris has collected. Towards its lower, northern end, a semi-circular section of the crest has collected debris inside it (Figure 4.4.46). On the opposite side of the valley, small areas of glacial drift are present; however, these are limited because of the substantial slope processes that have occurred on the valley side. While the deposit on the north-east-facing side may too have glacial origins as a lateral moraine/drift limit from a glacier margin within Coire Garbhach, there is insufficient evidence to conclude on the presence or absence of a later readvance of ice.

![Figure 4.4.44 Moraine and drift limits in Coire Garbhach. The deposit can be seen below the talus on the north-east-facing slope, to the bottom left-hand side of the photo](image-url)
Figure 4.4.45 Geomorphological map of Coire Garbhlaruch

Figure 4.4.46 Coire Garbhlaruch – stood on the deposit ridge looking up-valley, rockfall debris infilling the upper side. The ridge continues up-valley, depicted by the change in vegetation.
4.4.7 **Summary of interpretation: western Cairngorms**

This is a summary of the geomorphology and interpretation: except where described, no timing or chronology can be inferred from the order of the bullet points below. Further discussion of the wider retreat patterns and schematic diagrams of the retreat margins can be found in Chapters 6 and 7.

- Cairngorm ice within the Lairig Ghru broke from the Glenmore ice lobe, forming an ice-dammed lake between the local and regional ice masses. Later a similar pattern of events occurred in Gleann Einich. At this time, ice still filled neighbouring Glen Feshie to the west.

- The relative timing compared to the northern margin is unclear, but early within the deglaciation history of Glen Dee, the large moraine near Allt Garbh was formed, probably by the still-confluent Geusachan/Dee glacier. After this, an ice-dammed lake existed within at least lower Glen Dee and overflowed to the east over the Meirlea col. This requires the more eastern valleys of Glen Luibeg, Glen Lui and Glen Derry and further east down Deeside to have deglaciated, at least partially, indicating an east-to-west deglaciation of the southern Cairngorms. The favoured source for the damming ice is from the south, as there is meltwater-channel evidence on the southern flanks of Mòine Mhòr for ice retreat towards the east, and ice sourced directly from the east is unlikely, given the lake drained to the east. There is no evidence for the position of the Geusachan or Dee glaciers while this lake existed; presumably it has been destroyed by a readvance.

- Later, several phases of readvance or pauses in retreat have been identified at the exit of Glen Geusachan and upper Glen Dee. The lowermost upper Glen Dee linear moraine marks a glacier terminating in a lake. This is assumed to be dammed by Geusachan ice protruding into Glen Dee. Note: the Geusachan glacier must have been above the altitude of the Meirleach col to generate a water-terminating margin in upper Glen Dee; the main meltwater channels at the col are too low to be the corresponding overflow channels. Additional candidate readvance limits occur up-valley of these
outer margins: south of Devil’s Point in Glen Geusachan and associated with a boulder-covered lateral moraine in upper Glen Dee.

- The western plateau possesses evidence from the northerly flow of ice into Gleann Einich and the retreat of a glacier into the north-east plateau valley/head of Glen Geusachan. These are both interpreted to be from ice-sheet deglaciation, implying any later readvance of plateau ice was limited to localised areas and possibly the higher western mountain summits of Monadh Mòr, Beinn Bhrotaín and Braeriach.

- Glen Eidart contains evidence for a later readvance, as do many of the corries: Coire an Lochain (Braeriach), Coire Bhrochain, Coire an t-Saighdeir and Coire Dhondail. Some suggestions of possible corrie glaciation are evident in Coire Ruadh and Coire Bogha-cloiche with curious talus-glacier deposits. The corrie with the most substantial evidence for periglacial activity is Coire Beanaidh.

- Gleann Einich contains no evidence for a readvance of ice within the valley head; however, this may be expected given the steep slopes and low 500m valley-floor altitude. Evidence in Coire Garbhlach was limited and did not resemble the typical readvance geomorphology seen elsewhere in the Cairngorms.
4.5 Chapter Summary

This chapter has seen the presentation of new mapping and geomorphological evidence from both the analysis of aerial imagery and fieldwork. Lower glacier margins and ice-dammed lakes have been identified and interpreted to be from the interaction of regionally and locally sourced glaciers during ice-sheet retreat. The local relative chronologies have been established through direct use of the geomorphological evidence. The wider scale relative patterns, chronology and palaeoclimatic implications of this ice-sheet deglaciation phase are explored and discussed in Chapter 7: Section 7.1.

Sites of potential later readvance have been identified; these concern favourable corries and upper valley sections. These sites tended to be characterised by a high concentration of glacially transported boulders and arcuate moraine ridges within corries, and closely spaced sharp-crested moraine ridges within the valleys. These patterns are further discussed in Chapter 6. The more contentious readvance sites of Glen Geusachan and upper Glen Derry have been selected for new surface exposure dating (Chapter 5). The work presented within this chapter forms the foundation for much of the dating, glacier reconstruction and palaeoclimate inferences presented later in the thesis.
5 The Extent of Younger Dryas Glaciation

5.1 Introduction

In Chapter 4 multiple phases of glaciation were identified, including a more recent stage of corrie and valley glaciation. With the exception of the development of small glaciers during the Little Ice Age, there is a consensus that the last phase of major corrie glaciation occurred during the Younger Dryas (Sugden, 1970; Sissons, 1979a; Kirkbride et al., 2014). However, the age of the most recent phase of valley glaciation is unclear; this may also have been during the Younger Dryas (Sissons, 1979a; Bennett and Glasser, 1991; Midgley, 2001), or may have occurred during a prior period of climate deterioration most likely during ice-sheet retreat (Sugden, 1970; Purves et al., 1999; Everest and Kubik, 2006). This chapter presents results and discussion of new geochronological control for the final phase of valley glaciation in the Cairngorms. New cosmogenic surface exposure ages taken from boulders on moraines crests were processed with the objective of determining the extent of Younger Dryas glaciation. The work focuses on two specific geographical areas, the Lateglacial history of which is contested. The new ages provide minimum ages for the deglaciation of the last valley glaciers to exist in both Glen Geusachan and Glen Derry. The ages obtained from these sites – combined with the geomorphological mapping and landsystems analysis – will be used to draw wider inferences about the extent of other potential Younger Dryas glaciers in the Cairngorms.

5.2 Sample Sites

Following a review of the literature, new geomorphological mapping and reconnaissance fieldwork, two valleys were selected for dating. These were based on their importance to clarifying whether Younger Dryas valley glaciation occurred and an assessment of their suitability for surface exposure dating.

5.2.1 Glen Geusachan

Evidence of a substantial valley glacier (8.89 sq. km) was mapped by Sissons (1979a) in Glen Geusachan, and attributed to the Younger Dryas. This interpretation was largely supported by Bennett and Glasser (1991) who suggested
ice retreated actively and may have survived the Lateglacial Interstadial. Everest and Kubik (2006) obtained six surface exposure ages from boulders on what they interpreted as the glacier's lateral moraines. Although the samples yielded a large range, the ages were predominantly pre-Younger Dryas (recalculated NWH 11.6, 1mm/ka ages, Lm scaling: 17.4±1.1 ka to 12.8±1.1 ka, recalculated arithmetic mean 14.6 ka). However, it is recognised that meltwater channels in this area carried water from a lake in upper Glen Dee being dammed by the Geusachan ice, past the former glacier margin and into lower Glen Dee (Sisson, 1979a; Bennett and Glasser, 2001; Everest and Kubik, 2006). New geomorphological mapping suggests that the sampled boulders lie on a predominantly pre-existing surface into which meltwater channels were cut, rather than moraines of the Geusachan glacier, so the existing ages may not accurately constrain the last phase of glaciation (Figure 5.1). In addition, new mapping has identified a probable inner glacier margin within the previously dated channels to the south of Devils Point (Chapter 4: Section 4.4.3). This consisted of a high concentration of prominent closely spaced moraine ridges, the start of a boulder-covered outwash terrace and well-preserved lateral moraines. Glen Geusachan has important implications for our understanding of Younger Dryas glaciation in the Cairngorms. If associated with the Younger Dryas, the terminus of the glacier would have been at the lowest altitude of any glacier from this period and would be one of the largest ice masses within the Cairngorms. This would have important implications for understanding palaeoprecipitation gradients and local topoclimatic factors that supported such an ice mass. Thus new boulders from unambiguous moraine surfaces within the inner Geusachan glacier limits, south of Devils Point have been dated (Figure 5.1).
Chapter 5: The Extent of Younger Dryas Glaciation

5.2.2 Glen Derry

There is similar controversy over the timing of deglaciation at the head of Glen Derry, a valley unique in the Cairngorms for its distinct moraines marking clear stages of glacial recession. Sugden (1970) argued that the moraine assemblage was deposited during ice-sheet deglaciation, whereas Midgley (2001) favoured a Younger Dryas origin; Brazier et al. (1996b) noted that systematic mapping and dating was needed in this area. Although Sissons did not recognise a Younger Dryas glacier at this site, Midgley (2001) noted it would have been of a similar size to others reconstructed by Sissons (1979a). New geomorphological mapping found numerous recessional moraines within the head of Glen Derry/Coire Etchachan, with ice retreat both into the head of Coire Etchachan and also towards the higher unnamed corrie that holds Loch Etchachan (Chapter 4: Section 4.3.1). Closely spaced, consistently aligned, prominent moraine ridges on the valley floor at the head of Glen Derry were interpreted as marking the outermost limit of glacial readvance. Outside this limit (down-valley of N 57.07380, W 3.594753) moraines were shallower and less continuous and were interrupted to be from ice-sheet deglaciation. Three new ages from boulders within the prominent moraine-mound...
complex in upper Glen Derry have been dated and are discussed in relation to whether Younger Dryas valley glaciation occurred at this site (Figure 5.2).

5.3 Hypotheses

The competing hypotheses tested using surface exposure dating were:

1. Valley glaciers existed in Glen Geusachan and/or upper Glen Derry during the Younger Dryas (i.e. all samples yield Younger Dryas ages, c. 12.9–11.6 ka).

2. No glaciers existed in Glen Geusachan and/or upper Glen Derry during the Younger Dryas (i.e. all samples yield ages prior to the Lateglacial Interstadial >14.6 ka from a phase ice-sheet deglaciation).
5.4 Results

Further field data and details of the techniques used in the field, laboratory and the calculation of the ages can be found within the methodology (Chapter 3).

Using the field, laboratory and AMS data (Table 5.1), surface exposure ages have been calculated using the CRONUS-Earth online calculator v.2.2. (Balco, 2008 and subsequent updates). Erosion rates of 0mm/ka and 1mm/ka have been employed to permit easier comparison with other studies (which often quote ages assuming these erosion values). The erosion rate of 1mm/ka is based on measured erosion rates of postglacial granite and metamorphic rocks in Scandinavia which ranged from 0.2 to 1.2mm/ka (André, 2002). A 1mm/ka erosion rate increases the ages by 77–156 years depending on the sample and production rate when using the Lm scaling scheme. All internal uncertainty values are presented at the 68.27% (1σ) uncertainty level for AMS, blank, carrier and chemistry errors. The internal uncertainty can be used to compare samples within the same site as long as the same scaling scheme has been used (Balco et al., 2008). The external uncertainty includes additional uncertainties associated with the production rate and scaling schemes and must be used when comparing the results with samples in other geographical locations, other dating techniques or climate records. The ages according to the different scaling schemes (Table 5.2) are presented below; none of the schemes is currently thought to be more accurate than another and their distribution is unknown, thus averaging the ages from the schemes is not thought to provide more accurate results (Balco et al., 2008). The Lm and De schemes are quoted below as they have been adopted most commonly within the British Isles and they yield the oldest and youngest ages for the time-varying scaling schemes; thus it is assumed they bound the true age (Ballantyne, 2012). The variation in ages is largely a result of the altitudinal scaling of the schemes. Sites above 270m yield older ages with the Du scheme (Ballantyne, 2012). Due to the fast moving nature of cosmogenic surface exposure dating, and particularly the production rates used to calculate the ages, the new ages have been calculated with all the applicable production rates. The ages are given for the global production rate (GPR), North-West Highlands (NWH) and the Loch Lomond (LL) production rate.
Table 5.1 Sample details and required information for future recalibration of the exposure ages

<table>
<thead>
<tr>
<th>Sample</th>
<th>Lat</th>
<th>Long</th>
<th>Elevation</th>
<th>Shielding</th>
<th>Thickness</th>
<th>Density</th>
<th>Quartz</th>
<th>Carrier Added</th>
<th>Be Added</th>
<th>10Be/9Be</th>
<th>10Be/9Be Error</th>
<th>10Be Conc</th>
<th>10Be Error Conc</th>
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<tbody>
<tr>
<td></td>
<td>(ºN)</td>
<td>(ºW)</td>
<td>(m a.s.l.)</td>
<td>(factor)</td>
<td>(cm)</td>
<td>(g cm⁻³)</td>
<td>(g)</td>
<td>(µg)</td>
<td></td>
<td>(atoms g⁻¹)</td>
<td>(atoms g⁻¹)</td>
<td></td>
<td></td>
</tr>
<tr>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>HH1</td>
<td>57.07729</td>
<td>-3.59593</td>
<td>642</td>
<td>0.9906</td>
<td>2.0</td>
<td>2.7</td>
<td>28.7667</td>
<td>0.5325</td>
<td>215.87550</td>
<td>1.7940x10⁻¹³</td>
<td>5.9427x10⁻¹⁵</td>
<td>87721</td>
<td>3012</td>
</tr>
<tr>
<td>HH2</td>
<td>57.07729</td>
<td>-3.59593</td>
<td>642</td>
<td>0.9906</td>
<td>2.0</td>
<td>2.7</td>
<td>21.3884</td>
<td>0.5317</td>
<td>215.55118</td>
<td>1.2890x10⁻¹³</td>
<td>5.0778x10⁻¹⁵</td>
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<td>0.5298</td>
<td>214.78092</td>
<td>2.0507x10⁻¹³</td>
<td>6.9471x10⁻¹⁵</td>
<td>99690</td>
<td>3481</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
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</tr>
<tr>
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<td>517</td>
<td>0.9908</td>
<td>2.5</td>
<td>2.7</td>
<td>28.7323</td>
<td>0.5324</td>
<td>215.83496</td>
<td>1.6182x10⁻¹³</td>
<td>4.0176x10⁻¹⁵</td>
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<td>2064</td>
</tr>
<tr>
<td>LGG3</td>
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<td>-3.68021</td>
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<td>0.9916</td>
<td>2.0</td>
<td>2.7</td>
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<td>0.5328</td>
<td>215.99712</td>
<td>1.4899x10⁻¹³</td>
<td>4.4919x10⁻¹⁵</td>
<td>72358</td>
<td>2291</td>
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<tr>
<td>LGG5</td>
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<td>0.9902</td>
<td>1.5</td>
<td>2.7</td>
<td>28.3001</td>
<td>0.5174</td>
<td>209.75396</td>
<td>1.6880x10⁻¹³</td>
<td>4.9662x10⁻¹⁵</td>
<td>81324</td>
<td>2500</td>
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<tr>
<td>Blank</td>
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<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>0.5325</td>
<td>215.87550</td>
<td>4.4640x10⁻¹⁵</td>
<td>8.5932x10⁻¹⁶</td>
<td>n/a</td>
</tr>
</tbody>
</table>

Errors are reported at the ±1σ confidence level and 10Be concentration error propagation includes AMS, blank and chemistry error (including carrier addition and balance errors). Isotope ratios are normalised to NIST SRM4325 standard with an assumed isotope ratio of 2.79 x 10⁻¹¹ (10Be half-life value of 1.36 x 10⁶); for input into the CRONUS-Earth Calculators using NIST_27900. A blank value of 64,394 ±12,396 10Be atoms was subtracted from the samples. The correction for the blank <3.5% of the total 10Be in the samples.
Table 5.2 Scaling schemes used in the calculation of surface exposure ages (adapted from Balco \textit{et al.}, 2008 with permission of Elsevier)

<table>
<thead>
<tr>
<th>Constant production rate model</th>
<th>Time-varying production rate models</th>
</tr>
</thead>
<tbody>
<tr>
<td>Based on the latitude–altitude scaling factors of Lal (1991), as recast as functions of latitude and atmospheric pressure by Stone (2000). The scaling factor is a function of geographic latitude and atmospheric pressure. Does not take account of magnetic field variations—the nuclide production rate is constant over time</td>
<td>The scaling factor is a function of cut-off rigidity and atmospheric pressure. Production rates vary with time according to magnetic field changes</td>
</tr>
<tr>
<td><strong>Du</strong> Dunai (2001)</td>
<td><strong>Li</strong> Lifton \textit{et al.} (2005)</td>
</tr>
<tr>
<td>The scaling factor is a function of cut-off rigidity and atmospheric pressure. Production rates vary with time according to magnetic field changes</td>
<td>The scaling factor is a function of cut-off rigidity, atmospheric pressure, and a solar modulation parameter. Production rates vary with time according to changes in solar output as well as changes in the Earth’s magnetic field</td>
</tr>
<tr>
<td>An adaptation of the Lal (1991) scaling scheme that accommodates palaeomagnetic corrections. Production rates vary with time according to magnetic field changes. Based on the palaeomagnetic correction described in Nishiizumi \textit{et al.} (1989)</td>
<td></td>
</tr>
</tbody>
</table>
Chapter 5: The Extent of Younger Dryas Glaciation

Table 5.3 Surface exposure ages calculated using default production rate and 0mm/ka erosion rate

<table>
<thead>
<tr>
<th>Sample</th>
<th>Int Unc</th>
<th>Production Rate (atoms g⁻¹ a⁻¹)</th>
<th>Time Constant Production Rate</th>
<th>Time-varying Production Rate</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>St</td>
<td>Muons</td>
<td>Spallation</td>
</tr>
<tr>
<td>Glen Derry</td>
<td>HH1</td>
<td>352</td>
<td>0.227</td>
<td>8.37</td>
</tr>
<tr>
<td></td>
<td>HH2</td>
<td>406</td>
<td>0.227</td>
<td>8.37</td>
</tr>
<tr>
<td></td>
<td>HH3</td>
<td>419</td>
<td>0.225</td>
<td>8.13</td>
</tr>
<tr>
<td>Glen Geusachan</td>
<td>LGG1</td>
<td>271</td>
<td>0.217</td>
<td>7.44</td>
</tr>
<tr>
<td></td>
<td>LGG3</td>
<td>299</td>
<td>0.217</td>
<td>7.48</td>
</tr>
<tr>
<td></td>
<td>LGG5</td>
<td>326</td>
<td>0.218</td>
<td>7.50</td>
</tr>
</tbody>
</table>

Table 5.4 Surface exposure ages calculated using default production rate and 1mm/ka erosion rate.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Int Unc</th>
<th>Production Rate (atoms g⁻¹ a⁻¹)</th>
<th>Time Constant Production Rate</th>
<th>Time-varying Production Rate</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>St</td>
<td>Muons</td>
<td>Spallation</td>
</tr>
<tr>
<td>Glen Derry</td>
<td>HH1</td>
<td>358</td>
<td>0.227</td>
<td>8.37</td>
</tr>
<tr>
<td></td>
<td>HH2</td>
<td>412</td>
<td>0.227</td>
<td>8.37</td>
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<tr>
<td></td>
<td>HH3</td>
<td>428</td>
<td>0.225</td>
<td>8.13</td>
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<tr>
<td>Glen Geusachan</td>
<td>LGG1</td>
<td>276</td>
<td>0.217</td>
<td>7.44</td>
</tr>
<tr>
<td></td>
<td>LGG3</td>
<td>304</td>
<td>0.217</td>
<td>7.48</td>
</tr>
<tr>
<td></td>
<td>LGG5</td>
<td>331</td>
<td>0.218</td>
<td>7.50</td>
</tr>
</tbody>
</table>

Internal and external uncertainty is reported at the ±1σ confidence level and calculated using the CRONUS online calculator. Internal uncertainty error propagation includes AMS, carrier addition, blank and balance errors. External uncertainty includes uncertainty associated with the production rate and scaling scheme in addition to the internal uncertainty. The production rates and surface exposure ages were calculated using the CRONUS-Earth online calculator (Version: Wrapper script 2.2, Main Calculator 2.1, Constants 2.2.1 and Muons 1.1 with the default calibration dataset) (see Balco et al., 2008 and updates available online http://hess.ess.washington.edu/math/docs/al_be_v22/al_be_docs.html). Sample-specific production rates were calculated using the time constant production rate based on Lal (1991) and Stone (2000). A sea level high latitude spallation Lm reference production rate of 4.39±0.37 atoms g⁻¹ yr⁻¹ was used. The muon production rate is a constant production rate based on Heisinger et al., (2002a,b). The spallation production rates for the time-varying schemes cannot be calculated due to their inherent variation with time.
5.4.1 Surface exposure ages (global production rate)

The ages above have been calculated using the default CRONUS worldwide calibration dataset with no erosion rate applied (Table 5.3) and 1mm/ka erosion rate applied (Table 5.4). The samples for the Glen Derry (HH1–3) moraine site range from 10.0±0.9 ka to 12.2±1.1 ka using the Lm scaling scheme (0mm/ka). HH1 and HH2 fall within the early Holocene, whereas HH3 lies within the Younger Dryas Stade (12.9–11.6 ka BP); some potential reasons for the variability in age are discussed in Section 5.5. The ages for the Glen Geusachan boulders (LGG) range from 9.6±0.9 ka to 10.8±1.0 ka; all within the 2 ka subsequent to the termination of the Younger Dryas.

5.4.2 Surface exposure ages (local production rate: NWH 11.6)

A local production rate is available through the CRONUS-Earth project for the North-West Highlands, Scotland; the dataset is based on 18 boulders and 2 bedrock samples from Torridon, Applecross Peninsula and Skye. Most of the boulders are large and angular, thought to be inheritance-free rockfall deposits that fell onto Younger Dryas glaciers (Ballantyne and Stone, 2012). The dataset is available at http://depts.washington.edu/cosmolab/cronus/UW_CRONUS_data_June09.html and site descriptions at http://depts.washington.edu/cosmolab/cronus/cronus_cal.html. The NWH dataset can be calibrated to 11.6±0.3, 11.9±0.3 and 12.2±0.3 depending on when the calibration sites deglaciated (Ballantyne and Stone, 2012); here it is calibrated to 11.6±0.3 (referenced production rate for Lm of 4.20±0.15 atoms g⁻¹a⁻¹ and Du of 4.39±0.15 atoms g⁻¹a⁻¹) as used by Ballantyne (2012). As 11.6±0.3 ka is seen as the minimum deglaciation age for the calibration dataset it too provides minimum ages for the new samples. If the calibration sites were deglaciated earlier, then the new ages would also be proportionally older (Ballantyne, 2012). Using this local production rate, the ages become older shifting towards the Younger Dryas; with the HH3 sample yielding an age at the onset of the Younger Dryas (Table 5.5).
Table 5.5 Surface exposure ages calculated using the NWH 11.6 and 1mm/ka

<table>
<thead>
<tr>
<th>Sample</th>
<th>Int</th>
<th>Unc&lt;sup&gt;a&lt;/sup&gt;</th>
<th>Production Rate (atoms g&lt;sup&gt;-1&lt;/sup&gt; a&lt;sup&gt;-1&lt;/sup&gt;)</th>
<th>St</th>
<th>Unc&lt;sup&gt;b&lt;/sup&gt;</th>
<th>Age</th>
<th>Unc&lt;sup&gt;b&lt;/sup&gt;</th>
<th>Age</th>
<th>Unc&lt;sup&gt;b&lt;/sup&gt;</th>
<th>Age</th>
<th>Unc&lt;sup&gt;b&lt;/sup&gt;</th>
<th>Age</th>
<th>Lm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glenn Derry</td>
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<td></td>
<td></td>
<td></td>
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<tr>
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</tbody>
</table>

Ages calculated using the CRONUS online calculator (Developmental version: 10<sup>Be</sup> production rate calibration: Wrapper script 2.2, Main calculator 2.1, Constants 2.2.1 and Muons 1.1). The NWH 11.6±0.3 calibration dataset with a 1mm/ka erosion rate provided reference production rate for Lm scaling scheme of 4.20±0.15 atoms g<sup>-1</sup> a<sup>-1</sup> and Du of 4.39±0.15 atoms g<sup>-1</sup> a<sup>-1</sup>. NWH dataset available at: http://depts.washington.edu/cosmolab/cronus/UW_CRONUS_data_June09.html. <sup>a</sup>Internal and <sup>b</sup>external uncertainty is reported at the ±1<sup>ϭ</sup> confidence level.

5.4.3 Surface exposure ages (local production rate: Loch Lomond)

An additional local production rate has recently been used in Scotland (Fabel et al., 2012; Gheorghiu and Fabel, 2013); however the calibration details at the time of writing remain unpublished. It is derived from erratic boulders from the terminal moraine of the Loch Lomond glacier advance (Fabel et al., 2012), and an independent control based on radiocarbon ages of macrofossils associated with a varve sequence from a contemporaneous lake deposit (MacLeod et al., 2011). A boulder erosion rate estimate of 3mm/ka based on the relief of quartz veins was applied, resulting in a reference 10<sup>Be</sup> production rate of 3.92±0.18 atoms g<sup>-1</sup> a<sup>-1</sup> for the Lm scheme. This production rate is in approximate agreement with the NWH 12.2 local production rate. This production rate produces the oldest ages for the new samples (Table 5.6.).
### Table 5.6 Surface exposure ages calculated using the LL production rate and 1mm/ka

<table>
<thead>
<tr>
<th>Sample</th>
<th>Int Unc</th>
<th>Production Rate (atoms g⁻¹ a⁻¹)</th>
<th>St</th>
<th>De</th>
<th>Du</th>
<th>Li</th>
<th>Lm</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Muons Spallation Age Unc²</td>
<td>Age Unc²</td>
<td>Age Unc²</td>
<td>Age Unc²</td>
<td>Age Unc²</td>
<td>Age Unc²</td>
</tr>
<tr>
<td>Glen Derry</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>HH1</td>
<td>410</td>
<td>0.227</td>
<td>7.3</td>
<td>11796</td>
<td>680</td>
<td>11980</td>
<td>691</td>
</tr>
<tr>
<td>HH2</td>
<td>472</td>
<td>0.227</td>
<td>7.3</td>
<td>11262</td>
<td>700</td>
<td>11437</td>
<td>711</td>
</tr>
<tr>
<td>HH3</td>
<td>490</td>
<td>0.225</td>
<td>7.1</td>
<td>13814</td>
<td>803</td>
<td>14028</td>
<td>816</td>
</tr>
<tr>
<td>Glen Geusachan</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>LGG1</td>
<td>316</td>
<td>0.217</td>
<td>6.49</td>
<td>11925</td>
<td>633</td>
<td>12101</td>
<td>642</td>
</tr>
<tr>
<td>LGG3</td>
<td>348</td>
<td>0.217</td>
<td>6.53</td>
<td>10849</td>
<td>607</td>
<td>11009</td>
<td>616</td>
</tr>
<tr>
<td>LGG5</td>
<td>379</td>
<td>0.218</td>
<td>6.55</td>
<td>12177</td>
<td>676</td>
<td>12357</td>
<td>686</td>
</tr>
</tbody>
</table>

Reference ¹⁰Be production rate of 3.92±0.18 atoms g⁻¹ a⁻¹ for the Lm scheme. *Internal and *external uncertainty is reported at the ±1σ confidence level.

## 5.5 Age Robustness to Geomorphic Effects

The published NWH 11.6 local production rate has been used for the analysis below unless otherwise stated. The ages have been investigated graphically, statistically and through modelling to better understand the variability of ages and evaluate whether geomorphic effects such as shielding or inheritance may be impacting on the ages. If the spread of ages is a function of the analytical uncertainty alone, the ages will overlap at the generous 95% (1.96σ) internal uncertainty level (Figure 5.3). The Glen Geusachan (LGG) samples all show overlap which indicates no outliers are present and an average age for the sample site can be calculated (Dunai, 2010). The Glen Derry (HH) samples do not show the same overlap, with sample HH3 being older than HH1 and HH2. This suggests the difference in ages in Glen Derry is not caused by analytical uncertainty alone and that geomorphic effects are influencing the spread of ages. Thus an error-weighted mean value may not be the best representation of the true moraine formation age at the Glen Derry site or such a value may need to be viewed more critically.
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Figure 5.3 Surface exposure ages calculated using the NWH 11.6 production rate, Lm scaling scheme and 1mm/ka erosion rate, presented with internal 95%(±1.96) uncertainty error bars

Table 5.7 \( \chi^2 \) values for the Glen Derry and Glen Geusachan sites.

<table>
<thead>
<tr>
<th>Sample Site</th>
<th>( \chi^2 ) ( (1 \sigma) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glen Derry</td>
<td>7.94</td>
</tr>
<tr>
<td>Glen Geusachan</td>
<td>3.99</td>
</tr>
</tbody>
</table>

Similarly, the calculation of the sample sites’ reduced chi-square value can be used to identify whether the variation in observed ages at one site is caused by the expected internal/analytical uncertainty (uncertainty associated with the laboratory and AMS procedures) (Applegate et al., 2012). The reduced chi-square test compares the variation amongst the apparent ages at the sample site with the variation expected from the stated analytical uncertainties. To do this, the apparent ages and corresponding rigorous 1\( \sigma \) (68%) internal uncertainty levels from the CRONUS online calculator are used (Applegate et al., 2012). A \( \chi^2_R \) value close to 1 suggests the spread of ages is caused by analytical uncertainty alone, whereas higher values indicate the spread of ages is likely to be due to geomorphic effects such as inheritance or moraine degradation (Applegate et al., 2012). Table 5.7 indicates the variation in Glen Derry exposure ages is not from analytical uncertainty alone. In Glen Geusachan the \( \chi^2_R \) value is smaller suggesting more but not all of the scatter may be analytical uncertainty. Based on this approach, mean ages should be avoided as geomorphic effects are likely to have impacted on the ages.

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To better understand whether the source of boulder age variation Applegate et al., (2012) advocated the use of skewness as a tool to identify whether apparent boulder ages are impacted by inheritance or moraine degradation. If the skewness is less than ±0.5 it reflects a normal distribution resulting from analytical uncertainty. Greater positive skewness values indicate boulder inheritance and negative skewness indicates moraine degradation (Applegate et al., 2010; Applegate et al., 2012). However, with just three samples at each site, whether these samples reflect the true parent population is uncertain. In addition, by sampling boulders embedded within the moraine matrix we have changed the assumption of randomly sampled boulders used by Applegate et al. (2010). Thus our approach may have preferentially sampled homogenously young apparent ages that yielded young ages with a deceptively small scatter (Applegate and Alley, 2011). The measure of skewness may be a popular strategy in future studies, but in this case, due to the sample size and sampling strategy, it is not possible to proceed with this approach.

With the exception of HH3, the samples using both the default calibration dataset and the NWH 11.6 do not fall within the expected Younger Dryas or the Dimlington Stade. Instead the samples are grouped within the subsequent 2 ka after the termination of the Younger Dryas. The grouping of the samples at both sites within this time period does not suggest postglacial rock falls, boulder rotation or misinterpretation of the geomorphology. Thus assuming glaciation ended with the rapid increase of temperatures at the end of the Younger Dryas (11.6 ka BP), the ages must represent glacial deposition during the Younger Dryas or a prior cold period; this is thought to be the termination of the Dimlington Stade at 14.7 ka BP. Given this assumption, the boulders must have experienced shielding from overlying sediment and/or winter snow accumulations. Some limited support for this is given by the relation between the boulder height and sample age as seen in Figure 5.4 and Figure 5.5. This would suggest the higher boulders experienced less snow shielding or sediment cover; this is evaluated below, as are the implications of sediment and snow cover on the interpreted moraine formation ages.
Figure 5.4 Glen Derry apparent ages (NWH 11.6, Lm and 1mm/ka) plotted against boulder height above the surrounding moraine surface. Error bars for the ages are presented at the 68% (1σ) internal uncertainty level and boulder heights at 1 S. D.

Figure 5.5 Glen Geusachan apparent ages (NWH 11.6, Lm and 1mm/ka) plotted against boulder height above the surrounding moraine surface. Error bars for the ages are presented at the 68% (1σ) internal uncertainty level and boulder heights at 1 S. D.

Snow and sediment shielding act on the apparent ages of the samples by reducing the production rate of $^{10}$Be at the sampled surface. The production pathways for $^{10}$Be in quartz consist of spallation reactions, muon capture and fast muons (see Dunai, 2010 and references therein). Spallation reactions control the production
rate near the surface, with muons becoming the dominant production pathway at depths greater than 3m in rock (Figure 5.6). It is the effect of snow and sediment on the relatively short attenuation length of the spallation production pathway that primarily affects the production rate at the boulder surface, whereas the lower production rates from muons remains more constant with depth.

Figure 5.6 Production pathways for $^{10}$Be in quartz. from Dunai (2010) with permission of Cambridge University Press. Calculated using a rock density of 2.7g cm$^{-3}$. Depth dependency and surface production rates from Heisinger et al., (2002a, 2002b)
5.5.1 Snow shielding

A simple exponential attenuation equation can be used to calculate the impact of snow shielding on the production pathways and surface exposure ages. The annual impact of snow shielding on production rate can be calculated from the equation below:

\[
S_{\text{snow}} = \frac{1}{12} \sum_{i}^{12} e^{-((Z_{\text{snow}},i-Z_{\text{sample}})\rho_{\text{snow}},i/A)}
\]

Equation 5.1

Where \( Z_{\text{snow}},i \) is the monthly average snow height (cm), \( Z_{\text{sample}} \) is the boulder height, \( \rho_{\text{snow}},i \) is the monthly snow density. \( A \) is the attenuation length varying with latitude, 165 g cm\(^{-2}\) is used here for 57°N (Schildgen et al., 2005). In this case monthly intervals were used for summation. The snow density will vary according to climate and tend to increase through the winter. In northern Eurasia a snow density range of 0.16–0.48 g cm\(^{-3}\) has been measured with an average range of 0.16–0.33 g cm\(^{-3}\) (Onuchin and Burenina, 1996). The impact of different depths and densities on the apparent ages calculated is seen in Figure 5.7.

Based on the difference between the boulder age and the assumed true moraine formation age, it is possible to calculate the snow depth required to provide this reduction in production rate. LGG3 is c. 1.5 ka younger than the youngest assumed age of moraine formation at 11.6 ka, thus requires an addition of c. 15% from shielding. Consequently assuming snow survives for 4 months per year and there is a constant snow density of 0.3g cm\(^{-1}\), c.280cm of snow is required on top of the boulder to provide the equivalent of c. 1.5 ka of shielding. While these values are simplified and snow depth and density will change throughout the winter, it allows the impact of snow shielding to be assessed. Such depths are too large for the Holocene climate at the Glen Geusachan and Glen Derry sample sites, c.520m and c.610–640m respectively. In addition the true snow depth required would be greater given the boulder height and likely snow deflation from the moraine crest. Even greater depths would be needed to force the apparent ages back to the older assumed formation age during ice-sheet deglaciation (14.6 ka).
The Cairngorms have been the site of a study to determine the effect of snow shielding on cosmogenic surface exposure ages. A simulation of snow shielding carried out in the Cairngorms at 900 m found that, under present-day conditions, snow led to the reduction of production rates by 6% (constant-plus-exponential) or 9% (exponential profile) (Schildgen et al., 2005). However, at 600 m a.s.l., more representative of the sample sites within this study, the effect on production rate is considerably smaller, from 15.5 ka to present the effect is 2 (CPE) or 4% (EP), or for present-day conditions it is 0% (CPE) or 2% (EP). The explanation for the two values is based on the uncertainty surrounding how production rates decrease with depth at the surface interface. The exponential decrease has been used in most reports but the work of Masarik and Reedy (1995) suggested production rates stayed constant within the air-surface interface (12 g cm\(^{-2}\)); for snow of 0.3 g cm\(^{-3}\) density this constant depth is 40 cm (Schildgen et al., 2005). For our interpretation purposes here, this is not important.
Table 5.8 Snow-shielding effect on attenuation of production rate and age (data taken from Schildgen et al., 2005)

<table>
<thead>
<tr>
<th>Altitude (m a.s.l.)</th>
<th>Models run through to present day (Constant-Plus-Exponential or Exponential Profile) (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Present day</td>
</tr>
<tr>
<td>600</td>
<td>0 or 2</td>
</tr>
<tr>
<td>900</td>
<td>6 or 9</td>
</tr>
</tbody>
</table>

The Glen Derry and Glen Geusachan samples in this study are both taken from the central areas of the valley floor at altitudes below 650m. Thus the samples will not have been affected by the build-up of semi-permanent snow patches on sheltered north-east facing slopes. Based on the modelling of Schildgen et al. (2005), the altitude of the samples suggests that even if the samples were deposited prior to the Lateglacial Interstadte and suffered increased shielding during the Younger Dryas, the ages would only be affected by 2% (CPE) or 4% (EP) (up to c. 500 years). Applying such a correction would not shift all the ages from the Holocene/Younger Dryas to the Dimlington Stade. Thus based on this snow-shielding modelling, the ages are likely to represent Younger Dryas deposition, thus applying the 11 ka – present-day correction of 0% (CPE) or 2% (EP) may be necessary. For an age of 10 ka this would increase the exposure age by up to 200 years.

The modelling does not taken into account wind redistribution or slope aspect that have major effects on snow depth (Schildgen et al., 2005). Local factors such as the boulder’s position on the moraine and the height of the boulder will increase the boulder’s susceptibility to snow deflation by wind, thus reducing the chances of snow shielding. Although many of the boulders sampled were relatively low in height above the surrounding moraine matrix, the samples were taken from moraine crests. This exposure on the moraine crest would have a similar effect to sampling a larger boulder, increasing the exposure to wind; thus reducing the susceptibility to snow shielding.

Overall the effect of snow shielding is thought to be minimal in determining whether moraine formation took place during the Younger Dryas or Dimlington Stade. While the lower boulders may have suffered shielding up to the equivalent of a few hundred years, the larger boulders, such as HH3, are unlikely to have been
impacted. The limited impact of snow shielding is partly due to its low density and the limited length of the winter months, thus the samples still receive the full production rate for the majority of the year.

5.5.2 Constant matrix erosion modelling

Commonly it is assumed that the exposure age of a sample dates the formation of the moraine. However, no such straightforward relationship exists; while some boulders will remain on the moraine surface from the time of deposition, others will be exhumed from the surface as the fine-grained matrix is eroded by wind, water and creep (Putkonen and Swanson, 2003; see Chapter 3: Section 3.3.4.2). These more recently exhumed boulders will yield apparent exposure ages less than the true moraine formation age.

As we cannot assume constant exposure since boulder deposition, and sediment cover has an important impact on the exposure age, a model that considers constant erosion of the moraine matrix should be considered (Á. Rodés, pers. comm.). There is a positive relationship between the boulder height above the surrounding terrain and the apparent boulder age at both the Glen Derry site and more tentatively at the Glen Geusachan site (Figure 5.4 and Figure 5.5). This relationship is investigated further, below, by a constant rate moraine matrix erosion model created by Á. Rodés (see Appendix for details). Using the ages, boulder heights and associated uncertainties, it is possible to infer both the erosion rate and exposure age of the moraine.

Ideally, this modelling would have been carried out using multiple paired samples. Paired samples consisting of two neighbouring boulders of different heights, increases the chance the boulders have experienced the same surface lowering history. Unfortunately, due to the lack of suitable boulders and the expense of processing samples, this was not possible. The Glen Derry site contains one paired sample (HH1 and HH2) and a single boulder (HH3) and the Glen Geusachan site contains three single samples from different but morphologically similar moraine crests. This limitation must be kept in mind when considering the results below.
Table 5.9 Inputs for constant rate matrix erosion modelling (NWH 11.6, 0mm/ka and St scaling scheme)

<table>
<thead>
<tr>
<th>Samples</th>
<th>Age (yrs)</th>
<th>Uncertainty (yrs)</th>
<th>Height (cm)</th>
<th>Uncertainty (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>HH1</td>
<td>10917</td>
<td>375</td>
<td>90</td>
<td>10</td>
</tr>
<tr>
<td>HH2</td>
<td>10427</td>
<td>432</td>
<td>48</td>
<td>18</td>
</tr>
<tr>
<td>HH3</td>
<td>12742</td>
<td>445</td>
<td>140</td>
<td>0</td>
</tr>
</tbody>
</table>

Figure 5.8 shows the increase in exposure age with height above the surrounding moraine surface in Glen Derry. A steeper line of best fit would indicate greater sediment impact and higher moraine degradation rates, whereas a shallow or flat line would suggest no moraine degradation and all the ages representing the true moraine formation age. The line of best fit plateaus at 12.3 ka, within the uncertainty of the largest boulder, suggesting the oldest boulder accurately constrains the moraine formation age and the moraine matrix lowering has not adversely affected the exposure ages. This indicates that the surface lowering is less than the highest boulder (HH3).

Sampling provided a set of data with different ages and boulder heights. Numerical modelling was performed to calculate the time (t) and lowering since moraine formation (which is related to the time averaged erosion rate c) values that best fit
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the data obtained from the boulders. To fit the model to the data, an inverse method must be used. The same modelling of the $x^2$ fit-based inverse method as used by Siame et al. (2004), Braucher et al. (2009) and Rodés et al. (2011) was applied to define the solution to the age ($t$) and erosion ($\varepsilon$) space by minimising the $x^2$ value:

$$x^2 = \sum_{i=0}^{N} \left( \frac{C_i - C(h_i, \varepsilon, t)}{\sigma_i} \right)^2$$

Equation 5.2

where $C_i$ are the measured concentrations from the $N$ samples at $h_i$ heights, $C(h_i, \varepsilon, t)$ is the concentration predicted by the model for each $h_i$, and $\sigma_i$ are the uncertainties of the measured concentrations (Å. Rodés, pers. comm.).

As the data have inherent measurement errors, and erosion processes are more complex than the model assumes, the data never fully fit the model but there is only one solution that minimises the $x^2$ value (Rodés et al., 2011). The change in $x^2$ defines a constant confidence contour region around the modelled best-fit solution (Siame et al., 2004). The $\varepsilon$-$t$ values that fit the data within the $1\sigma$ and $2\sigma$ confidence levels (68.27% and 95.45%) can be calculated using the $x^2_{\text{min}}$ value and the chi-square distribution based on the number of degrees of freedom (cf. Rodés et al., 2011).

The erosion and age values that best fit the field data are shown in Figure 5.9. The $x^2$ values are associated with the confidence regions surrounding the minimum best-fit value. The inner two contours mark the $1\sigma$ and $2\sigma$ uncertainty levels. The geometry of the tilted ellipses shows the degree of correlation between the two variables’ age and erosion (Bevington and Robinson, 1992). Note the ellipses become shallower towards the lower erosion rates; this is expected as it indicates the lower erosion rates having less impact on the ages. Figure 5.9 indicates a moraine formation age between 11.4 and 13.1 ka at the $1\sigma$ uncertainty level and moraine surface lowering values between 80cm and 150cm. The erosion modelling was calculated using the St scaling scheme and the NWH 11.6 production rate. Except for the variations in the moraine formation ages, the modelling is not thought to be sensitive to different scaling schemes and production rates. Thus the results suggest the Glen Derry ages have not been adversely affected by sediment cover.
and the moraine formation age is likely to be accurately constrained by the oldest sample age given by the chosen production rate.

![Figure 5.9 Glen Derry moraine age and total surface lowering at the 1σ(solid red) and 2σ(dotted) confidence levels according to chi-square inverse modelling. Chi-square values are also shown](image)

The modelling for the Glen Geusachan moraines did not assist in constraining the upper age of moraine formation. The line of best fit indicated a formation age over 20 ka and a surface lowering above 2m; such an age is not realistic given Scotland was beneath a large ice sheet at this time (Clark et al., 2012). The 1σ uncertainty modelling shows that the moraine age may vary between 11.4 and 180 ka and surface lowering between 0.5 and 35m. This indicates the moraine was formed before 11.4 ka but does not assist in giving a realistic upper age constraint. It is thought the modelling was less successful due to the sample dataset characteristics; the samples were from different moraines, therefore may have undergone different erosion histories and in addition the small variations in boulder heights may have caused the large uncertainties within the modelling. Unfortunately larger boulders were unavailable at the sample site.
5.5.3 Moraine degradation modelling

As it is not possible to eliminate the possibility that the ages from the Glen Geusachan moraine samples (LGG) have been affected by sediment cover using the modelling above, an evaluation of the potential effect of sediment cover on the apparent ages has been undertaken. Existing modelling of fine-grained matrix erosion attempts to model processes that act on moraines such as soil creep; the modelling suggests most boulders will be exhumed within the first few thousand years after moraine formation (Putkonen and Swanson, 2003). Putkonen and Swanson also suggested smaller moraines (<20m in initial height) generally suffer less from degradation, and thus exhumation complications may be less of a problem than for taller and older moraines (Figure 5.10). Table 5.10 indicates that the ages of 3–4 boulders (>1m), from young small moraines similar to those in this study, should yield an age >90% of the true moraine formation age. Some support for this is given by the modelling above, where in Glen Derry sediment cover is not thought to have adversely affected the ages, particularly of the largest boulder.

Figure 5.10 Moraine-crest lowering through time for six moraines of different initial heights. The topographic diffusivity (i.e. soil erosion rate) is based on survey results from moraines in Bloody Canyon, Sierra Nevada. Note for small moraines the initial crest lowering is relatively large followed by a long period of relatively little change (reproduced from Putkonen and Swanson (2003) with permission of Elsevier)
Table 5.10 The number of boulders (diameter > 1 m) that need to be randomly sampled from a moraine crest to find a boulder whose age is > 90% of the moraine age with 95% probability, for a given initial moraine height and age (adapted from Putkonen and Swanson (2003) with permission of Elsevier)

<table>
<thead>
<tr>
<th>Moraine age (ka)</th>
<th>100</th>
<th>90</th>
<th>80</th>
<th>70</th>
<th>60</th>
<th>50</th>
<th>40</th>
<th>30</th>
<th>20</th>
<th>10</th>
</tr>
</thead>
<tbody>
<tr>
<td>100</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>6</td>
<td>5</td>
<td>5</td>
<td>3</td>
<td>1</td>
</tr>
<tr>
<td>80</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>6</td>
<td>5</td>
<td>5</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>60</td>
<td>7</td>
<td>7</td>
<td>7</td>
<td>7</td>
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<td>6</td>
<td>6</td>
<td>5</td>
<td>4</td>
<td>2</td>
</tr>
<tr>
<td>40</td>
<td>6</td>
<td>6</td>
<td>6</td>
<td>6</td>
<td>6</td>
<td>6</td>
<td>5</td>
<td>5</td>
<td>4</td>
<td>2</td>
</tr>
<tr>
<td>20</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>4</td>
<td>3</td>
<td></td>
</tr>
</tbody>
</table>

Using the same principles as for snow shielding, it is possible to calculate shielding from sediment cover. Peat and soil do not currently cover the samples, however it is possible that fine glacially transported moraine material covered the samples and has since been eroded. The following equation can be used to calculate the production rate at the shielding depth $P_s$

$$P_s = P_0 \cdot e^{-\left(z \cdot \rho_{\text{sediment}}\right)/A}$$

**Equation 5.3**

where $z$ is depth of sediment, $\rho_{\text{sediment}}$ is density of sediment, $A$ is attenuation length (160 g cm$^{-2}$), and $P_0$ is cosmogenic production rate at Earth’s surface. The values for sediment density vary according to its make-up and water content; it is thought the values used in Figure 5.11 bound most natural occurring sediment densities. Variable sediment densities of 1.8 g cm$^{-3}$ (uncemented gravel) – 2.5 g cm$^{-3}$ (gravels with well-developed calcrete crust) were found in outcrops from alluvial fans (Rodés et al., 2011). Most studies have used values of 2.0g cm$^{-3}$ to represent moraines (Applegate et al., 2010; Heyman et al., 2011). The implications of different sediment densities and depths on the exposure ages can be seen in Figure 5.11; where depths of just 20cm can cause the true formation age to be c.25% older than the apparent exposure age. The shielding acts through the dominant spallation production pathway and although production by muons has been included, the attenuation has been ignored due to its relatively small impact.
Figure 5.11 Effect of sediment shielding on exposure ages. Includes fixed production by muons but ignores muon attenuation. Calculated for spallation production using an attenuation length of 165m g cm\(^{-2}\). Based on equations from Gosse and Phillips (2001)

However, the samples have not been covered for their entire exposure history. Thus there has been a recent focus of research in understanding and modelling moraine degradation and exhumation of fresh boulders (Putkonen and Swanson 2003; Applegate et al., 2010; Applegate and Alley, 2011; Applegate et al., 2012).

The modelling can be divided into two stages: the modelling of moraine degradation/lowering, and then the impact of the sediment cover on the eventually exhumed apparent boulder ages. The degradation modelling is controlled by the initial surface slope and a topographic diffusivity coefficient. Some uncertainty is introduced by assuming a sharp-crested moraine ridge at the model onset. The moraine slope is assumed to be 34° based on field observations from recently deglaciated moraines (Putkonen and Swanson, 2003; Putkonen et al., 2008). The slopes become shallower with time as shown in the valleys of North America, the oldest Pleistocene moraines had slope angles of 15° and the up-valley younger Pleistocene moraines 23° (Putkonen and O’Neal, 2006). While 34° seems a reasonable assumption for lateral moraines, end moraines tend to have more varied cross-profiles, particularly if ice-cored, feature multiple crests or have been streamlined by a readvance (Putkonen et al., 2008). To alleviate this argument that moraine lowering from sharp crests is rapid or that moraines start as smoother shaped landforms, Putkonen and O’Neal (2006) pre-aged moraine crests to smoother profiles, thus making the subsequent lowering a minimum estimate. This
has not been carried out here as the initial sharp-crested ridge rapidly erodes (Applegate et al., 2010).

A second source of uncertainty is introduced by the topographic diffusivity coefficient which represents the erodibility of the slope sediments. The value is the volume of sediment which moves downslope per unit time and per unit length transverse to the slope divided by the slope inclination (cf. Hallet and Putkonen, 1994; Applegate et al., 2010). Proposed values for this coefficient range over several orders of magnitude from $10^{-1}$ to $10^{-4}$ m$^2$/yr (Putkonen et al., 2008 and references therein) and possibly as low as $10^{-5}$ m$^2$/yr in Antarctica (Putkonen et al., 2007). This value varies according to climatic factors and the alluvial substrate (Putkonen and Swanson, 2003). The values are generally poorly constrained, but some analysis based on a range of plausible values may be useful. The final uncertainty is the initial moraine height, which must be estimated. While this is less important for larger moraines where deposition on the lower flank will not reduce the slope angle at the eroding moraine crest; on smaller moraines, such as those within the study area, deposition of the eroded material will immediately reduce the moraine-crest slope angle, thus reducing the rate of degradation.

The second stage of the model is based on boulders occurring throughout the moraine matrix at different initial depths. Effectively the moraine degradation model described above then lowers the moraine surface causing these boulders to become closer to the moraine surface and eventually be exhumed onto the moraine surface. During their submerged period they receive a reduced production rate based on the matrix depth above them, and after exhumation they receive the full production rate. This approach results in a range of apparent boulder ages on the final moraine surface. Further information on the model used can be found in Applegate et al. (2012) and references therein. The published MATLAB code from Applegate et al. (2010) and Applegate et al. (2012) was used for the analysis below.

For a prescribed time period, the model outputs the initial moraine surface, end moraine surface, lowering rate and a distribution graph of the boulder ages on the moraine surface. The final output equates to the probability of randomly sampling a boulder from the different age intervals given a random sampling strategy. However, the sampling strategy employed in this study, of sampling boulders still
partially embedded within the moraine matrix, makes the use of this approach invalid. Thus the full approach outlined in Applegate et al. (2012) is not followed. However, of interest, is that the modelling can be used to calculate whether a given set of moraine parameters can cause a moraine produced during the Dimlington Stade (>14.6 ka BP) to yield an apparent Younger Dryas age (<11.6 ka).

The moraines within the sample area have heights today of c.2–5m and slopes predominately less than 15° (based on data extracted from a 5m resolution DEM). Thus the maximum value within this range was taken to be representative of the start height; this was to avoid starting with erroneously large moraines. The suggested initial slope value of 34° was used for the modelling (Putkonen and Swanson, 2003; Putkonen et al., 2008). The model was highly sensitive to the k diffusivity coefficient (Figure 5.12). k values of \(10^{-1} - 10^{-1.5}\) cause moraines to flatten quickly (Figure 5.12) and, irrespective of the initial depth, most boulders yielded apparent ages close to the true moraine formation age due to rapid exhumation (Figure 5.12). Conversely k values of \(10^{-4}\) result in little rounding of the moraine crest with limited lowering and yielded ages close to the moraine formation age (Figure 5.12). The intermediate k values yielded younger minimum apparent ages and are more problematic for interpretation of the moraine formation age (Figure 5.12).
Modelling $k$ diffusivity values of $10^{-2}$ to $10^{-3.5}$ on a 5m moraine for 14.6 ka caused crest lowering between c. 1.6–4m. Sampling boulders 35cm in height, such as in Glen Geusachan, would yield apparent ages within or younger than the Younger Dryas termination. This would suggest in the order of 1.25–3.65m of sediment on top of the sampled boulder surface at the time of moraine formation (Figure 5.12). Although this sediment shielding is shown to be possible, the assumption that the moraines were sharp crested and had initial slope angles of 34º cannot easily be tested and the correct $k$ diffusivity value is poorly constrained and varies with climate, substrate and vegetation. However, this modelling indicates the vulnerability of surface exposure ages to moraine surface lowering and the
capability to yield younger apparent ages. The larger boulders such as those in Glen Derry are less susceptible to shielding from the modelled crest lowering.

5.5.4 Constant matrix erosion and moraine degradation discussion

The modelling and discussion demonstrates that larger boulders are less susceptible to snow and sediment shielding. The use of multiple samples of different heights above the moraine surface proved to be a promising technique which requires further validation. The constant matrix erosion modelling indicated the Glen Derry moraine formation age was between 11.4 and 13.1 ka at the 1σ uncertainty level and moraine-surface lowering values were between 80cm and 150cm. This equates to erosion rates in the region of 0.07mm a⁻¹ to 0.11mm a⁻¹. These denudation rates are larger than values derived from the depth profile dating of Pinedale (0.02mm a⁻¹) and Bull lake (0.01mm a⁻¹) moraines in America (Schaller et al., 2009). However, the Glen Derry values are less than denudation rates of c.0.5mm a⁻¹ from degradation modelling of 10–20m high moraines over 10–20ka (see Figure 5.10; Putkonen and Swanson, 2003). A major strength of the constant matrix erosion model is that it suggests the oldest apparent boulder age in Glen Derry is representative of the true moraine formation age. The constant matrix erosion model was not able to eliminate the possibility that sediment cover adversely affected the apparent ages in Glen Geusachan; this was most likely due to restricted boulder size variability at the Glen Geusachan site.

The sampling of boulders still embedded within the moraine matrix meant it was not possible to use the Applegate et al. (2012) model as it was originally intended; however this additional control on the position of the boulders caused the minimum apparent age from the modelling to be relevant. Modelling potentially representative parameters for the Glen Geusachan site showed the susceptibility of apparent ages to moraine-crest lowering. However, the high moraine surface lowering depths required seem improbable, although there is no clear evidence to support this interpretation. Perhaps more importantly, the relatively tightly grouped ages yielded from Glen Geusachan from different moraine crests and positions on the crests make it unlikely that the Younger Dryas aged boulders are from an older moraine. Rather it seems more probable the ages represent Younger Dryas formation with small amounts of surface lowering causing only small variations in the exposure ages. Thus we conclude it more probable the Glen Geusachan samples represent
Younger Dryas glaciation, although greater impacts by sediment cover cannot be eliminated.

The potential impact of sediment shielding on the use of surface exposure ages is considerable. The constant moraine matrix model is useful as it uses the data derived from the field to calculate the best-fit erosion rate and moraine formation age. The moraine degradation model is advantageous when paired samples are unavailable; however it is very sensitive to the input parameters that are not well controlled. It is also time-varying so responds to the likely rapid lowering of the sharp crest within in the first 1 ka before slowing as the moraine shape evolves. However, neither model fully represents reality. A potential further weakness may be the change in erosion rates with climate and vegetation cover, for example, if the moraines pre-dated the Younger Dryas they may have experienced increased erosion during the Younger Dryas due to a combination of the severe climate and reduced vegetation cover.

Despite the high uncertainties related to these modelling approaches, this study has attempted to access the robustness of the apparent ages to moraine degradation. The exploration of approaches has been worthwhile and a quantitative understanding of the surface exposure ages vulnerability to moraine degradation has been gained. Future studies should provide a more detailed account of the potential shielding from sediment as it can impact on the interpretation of the ages and on a wider scale if such boulders are used to derive production rates.

5.6 Discussion: Wider Geochronology and Palaeoclimate

5.6.1 Glen Derry and the Carn Etchachan RSF/moraine

The new Glen Derry ages can be compared with the surface exposure dating of a rock slope failure (RSF) deposit at Carn Etchachan (Ballantyne et al., 2009a); formerly interpreted as a rock glacier (Sissons, 1979a). A recent reinterpretation suggested this feature is a moraine, with only small localised areas of rock fall (Jarman et al., 2013). This reinterpretation is supported by the continuation of the moraine topography within the corrie floor onto the conical deposit, suggesting that the feature may in part have glacial origins, with moraine deposits overlying the underlying bedrock/pre-existing rock slope failures (see Chapter 4: Section 4.3.5.1). The published ages of this deposit range from 12.4±1.1–11.8±1.2 ka (recalibrated
with NWH 11.6: 12.9±0.9–12.3±1.1 ka) suggesting formation early within the Younger Dryas Stade. The age of this feature is important as it falls within the likely source area for the glacier that deposited the newly dated moraine surfaces in Glen Derry (Figure 5.13). Geomorphological evidence indicates that ice sourced in the unnamed corrie that holds the dated RSF/moraine flowed east through the col and contributed to the Derry glacier, as later recorded by recessional moraines (Chapter 4: Section 4.3.1.3). If interpreted as a RSF the deposit can be used to infer a minimum age for the glacier. Similarly, if the deposit is interpreted as a moraine the ages still strictly indicate a minimum age for the glacier; however, the ages become more relevant to the timing of the final deglaciation of the site. Using the NWH 11.6 local production rate, the Carn Etchachan RSF/moraine ages occur within the Younger Dryas with the oldest age being at its onset. Using the same production rate and scaling scheme, the ages from the Carn Etchachan RSF/moraine are similar to the oldest sample in Glen Derry (HH3) (Figure 5.14). This suggests these ages may represent the retreat of a glacier from upper Glen Derry and the linked unnamed corrie that holds Loch Etchachan. This would suggest the younger apparent ages in Glen Derry (HH1 and HH2) may, as the modelling suggests, have suffered from sediment cover/moraine lowering. Such issues with the Carn Etchachan RSF/moraine deposit are not as likely due to its rocky nature and absence of finer material. This would indicate a glacier existed within the Loch Etchachan and Glen Derry area after the retreat of ice during the Dimlington Stade. The exact timing of this glacier largely depends on the production rate applied. The default global production rate suggests deglaciation during the Younger Dryas, the NWH 11.6 local production rate at the onset of the Younger Dryas and the NWH 12.2 and LL local production rates suggest deglaciation during the Lateglacial Interstade. This makes linking the glacier to climatic events problematic and this is discussed in Section 5.6.3.
Figure 5.13 New Glen Derry sample sites and existing Carn Etchachan RSF/moraine sample site. Glacier outline shows the Glen Derry and Glen Avon glacier with a valley-style source area (full key can be found in Chapter 4: Figure 4.1.2)

Figure 5.14 Glen Derry and Carn Etchachan exposure ages recalibrated from Ballantyne et al. (2009a) (NWH 11.6 and 1mm/ka: Lm (black rectangles) and Du (blue circles))

Climate Data NGRIP on GICO5 age (years before 2000 AD) (Andersen et al., 2006; Rasmussen et al., 2006; Svensson et al., 2006) (Downloaded from www.icecores.dk). Ice-core age for climatic events as proposed by Lowe et al. (2008)
5.6.2 Glen Geusachan

The new Glen Geusachan samples in this study yielded ages younger than the existing boulder ages near the meltwater channels opposite the exit of Glen Geusachan (Everest and Kubik, 2006). Everest and Kubik (2006) dated 6 boulders from 'lateral moraine crests' associated with the meltwater channels opposite the exit of Glen Geusachan (Figure 5.1). The published ages show a large range from 16.5±1.0 to 12.2±0.7 ka (recalculated NWH 11.6, 1mm/ka ages, Lm scaling: 17.4±1.1 ka to 12.8±1.1 ka) which includes the Younger Dryas, Lateglacial Interstadial and the later part of the Dimlington Stade. This compares to the younger ages in this study which range from 10.1±0.5–11.4±0.5 ka using the same NWH 11.6 local production rate. With the exception of the sampling sites' location and the landform type, the only notable difference between the sampled boulders is their size and height above the surrounding terrain. Everest and Kubik (2006) sampled surfaces are in excess of 1m above the terrain whereas the new samples at this site are less than 0.5m. This difference may alter the boulders susceptibility to snow and sediment cover. The competing explanations for the differences in yielded ages between the new and existing sample areas are outlined below:

1. The ages represent glacier margins associated with separate readvances. The older samples near the meltwater channels represent an older glacier margin prior to the Lateglacial Interstadial and the younger samples from this study represent a smaller later Younger Dryas glacier that terminated up-valley of the channels.

2. The area marks the retreat of one older Dimlington Stade glacier and no Younger Dryas readvance. The older boulder ages on the interfluves between the channels represent the true deglaciation age, based on the oldest ages being the most representative of deglaciation. The new samples with younger ages have suffered shielding by sediment cover due to their lower boulder height above the terrain. Thus all the samples represent deglaciation prior to the Lateglacial Interstade and the exhumation of fresh boulders explains both the younger ages and the variation in ages.
3. The third explanation is that a Younger Dryas glacier readvanced to the channels of the outer existing sample site. The existing older ages yielded a large range of ages which can be explained by:
   - The incorporation of inherited boulders from a pre-existing glacial foreland/valley side.
   - The dated boulders are not from an area of new Younger Dryas moraine formation; instead they are on interfluves of a pre-existing surface through which Younger Dryas aged meltwater channels have been eroded.

Thus the existing boulder ages have a complex exposure history and do not accurately represent one glacier margin; whereas the new younger ages from the unambiguous moraine crests indicate ages closer to the true moraine formation age during the Younger Dryas.

![Figure 5.15 Area of previously sampled boulders. Note continuous flat-topped pre-existing surfaces through which channels have been eroded](image_url)

The viability of these options is discussed in turn. The first explanation of two glacier margins can be discussed in relation to the geomorphological evidence. Mapping evidence suggests differences within moraine morphology are present at the exit of Glen Geusachan, and different sections are discussed in Chapter 4 Section 4.4.3. The most prominent margin is the change to sharp-crested, closely spaced moraines, south of Devils Point and the inner lower boulder-covered
outwash terrace that is associated with this margin. A Younger Dryas readvance to this inner margin would explain the younger ages at the new sample site, and the older ages outside the inner glacier limit as originally interpreted by Everest and Kubik (2006) would mark a stage of ice-sheet deglaciation.

The second explanation has some support based on the importance of sediment cover on apparent exposure ages, as seen by the modelling and discussion above. This would suggest that, all things being equal, the larger boulders near the channels would yield ages closer to the true moraine formation age than the newly sampled smaller boulders. However, the ages of the six larger boulders are highly variable (recalculated NWH 11.6, 1mm/ka ages, Lm scaling: 17.4±1.1 ka to 12.8±1.1 ka). This large variation cannot be explained by sediment cover alone as all the samples were taken from boulders larger than 1m and from relatively flat surfaces that are unlikely to have suffered significant surface lowering. In addition, two of the ages are within the Younger Dryas which cannot easily be explained by sediment cover, as similar sized boulders yielded ages 2 ka older. Thus this variation in the existing samples’ ages may represent the presence of another geomorphic factor. In addition, the newly sampled surfaces would require substantial sediment depth and for a sustained period to shift the apparent sample ages back to prior to the Lateglacial Interstade. While this cannot be eliminated, it is deemed unlikely that three samples from different moraines all underwent very similar moraine surface lowering histories to enable them all to yield apparent ages in the Younger Dryas/early Holocene, despite being from ice-sheet deglaciation.

The third explanation provides two plausible reasons for the existing older samples and the variability in their ages. The incorporation of boulders with inheritance from prior exposure on the valley floor or valley side is not unexpected as the boulders are near the maximum extent of the glacier, and it is likely boulders from a prior glacial foreland would have been carried to the maximum extent of the glacier early in the readvance. Some limited support for this is provided by the skew of the existing sample dataset which is 0.75, indicating the possible presence of inheritance (Applegate et al., 2012). The two younger ages of 12.2 ka and 12.4 ka (recalculated NWH 11.6: 12.8±1.1 ka and 13.0±0.9 ka) may represent direct Younger Dryas glacier deposition and the older ages may have prior exposure either on cliffs or on the valley floor. However, it is unlikely that four of the six
existing samples contained inheritance. Inheritance will be focused on a single side/corner of the boulder, thus unless the boulder is not rotated during transportation there is a good chance the boulder will be deposited with an inheritance-low side facing up. Rudimentary calculations indicate the bottom of a 1m diameter boulder would yield exposure ages c. 22% and a 2m boulder c. 7% of the full boulder top ages.

The second explanation for the older samples and variation in age is a misinterpretation of the geomorphology. The boulders, instead of being incorporated in moraine ridges, seem to be lying on a pre-existing surface that has subsequently been modified by meltwater activity and eroded to form the channels (Figure 5.15). Dating boulders from such a surface would yield older and more variable ages, as boulders may have been exposed to cosmogenic rays either on the terrain surface or beneath sediment prior to Younger Dryas glacier and meltwater modification. The new lower younger boulder ages from the moraine crests are therefore likely to represent Younger Dryas moraine formation with limited sediment cover.

Here the first explanation is favoured as geomorphological evidence indicates a potential inner readvance position, down-valley of the new younger ages in this study and short of the older existing ages. The third hypothesis cannot be ruled out, as the readvancing glacier may have approached/produced the channels opposite the exit of Glen Geusachan where the existing older ages were taken. Such a situation, with the mixing of new and old boulders and meltwater cut surfaces may assist in explaining the large variation of ages at this site. The sediment cover hypothesis is unlikely to be able to explain all the variation in boulder ages both between the two sites and the variability of ages within their respective sample sites. While sediment cover cannot be ruled out and the new ages strictly provide minimum ages for deglaciation, the presence of a Younger Dryas glacier in Glen Geusachan is worthy of further climatological investigation.

5.6.3 Production rate influence on ages and climate interpretation

Using the full uncertainty values, the exposure ages can be compared to climate data from the Greenland ice core records or chironomid-inferred temperature records from sites within the British Isles. These are thought to correlate well
throughout the period of interest – Lateglacial Interstade, Younger Dryas and the onset of the Holocene (Brooks and Birks, 2000; Brooks et al., 2012). Figure 5.16 and Figure 5.17 show the surface exposure ages and their associated uncertainty plotted against dissolved oxygen concentrations from NGRIP, a proxy for temperature. As the work above suggests, unless a boulder suffers from inheritance it is likely the oldest sample from each site best represents the true deglaciation age and younger ages are likely to be due to shielding from the moraine matrix.

The globally derived production rate produces ages indicating deposition during the Younger Dryas and Holocene (Figure 5.16). For reasons aforementioned the local production rate ages are favoured; the NWH requires assumptions to be made about the timing of local Younger Dryas deglaciation 11.6, 11.9 or 12.2 ka. The NWH 11.6 and NWH 12.2 have been employed here to provide minimum and maximum ages (Figure 5.17). These local production rates produce older ages shifting most of the ages towards and into the Younger Dryas. However, HH3 falls at the onset of the Younger Dryas when the NWH dataset is calibrated to 11.6 ka and HH3 falls within the Lateglacial Interstade when NWH is calibrated to 12.2 ka, although the uncertainty error bars still include the onset of the Younger Dryas. The Loch Lomond local production rate has independent radiocarbon control and the ages according to this can be seen in Table 5.6 (D. Fabel, pers. comm.), the details of this production rate remain unpublished but yield similar ages to the NWH 12.2.

In light of the current uncertainty surrounding production rates, it is useful to compare the new ages to other Scottish sites where samples are thought to represent Younger Dryas glaciation (Figure 5.18). The figure shows the new ages to be grouped on the younger side of the dataset with the exception of HH3 which is comparable with the older ages. The ages are not dissimilar to the other sample sites which have been interpreted to represent Younger Dryas glaciation. Note in Table 5.11 the new Cairngorm ages are from a relatively high altitude compared to the other sample sites; thus the age comparisons change slightly with the use of the Lm and Du scaling schemes.
Chapter 5: The Extent of Younger Dryas Glaciation

Figure 5.16 Surface exposure ages plotted with NGRIP climate record. Default calibration and no erosion rate

Climate Data NGRIP on GICO5 age (years before 2000 AD) (Andersen et al., 2006; Rasmussen et al., 2006; Svensson et al., 2006) (Downloaded from www.icecores.dk). Ice-core age for climatic events as proposed by Lowe et al. (2008). The age and full uncertainty error bars are shown for both the Lm (square) and Du (circle) scaling schemes calculated using the default calibration dataset and no erosion rate.

Figure 5.17 Surface exposure ages plotted with NGRIP climate record. NWH 11.6 (Black rectangles) and NWH 12.2 (Blue circles) 1mm/ka erosion rate and Lm scaling scheme

Climate Data NGRIP on GICO5 (years before 2000 AD) (Andersen et al., 2006; Rasmussen et al., 2006; Svensson et al., 2006) (Downloaded from www.icecores.dk). Ice-core age for climatic events as proposed by Lowe et al. (2008). The age and full uncertainty error bars are shown for both the Lm scaling scheme calculated using the NWH 11.6 and NWH 12.2 calibration dataset (as used by Ballantyne (2012) with a 1mm/ka erosion rate).
Figure 5.18 a–c Scottish sample sites interpreted to be of Younger Dryas age (data from Ballantyne (2012) with permission from Wiley) Grey area indicates Younger Dryas event based on ages from Lowe et al. (2008)
Table 5.11 Source and sample details for Figure 5.18 (adapted from Ballantyne (2012) with permission from Wiley)

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<th>Sample code</th>
<th>Site</th>
<th>No. of samples</th>
<th>Altitude (m a.s.l.)</th>
<th>Sample type</th>
<th>Source</th>
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<td></td>
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<td></td>
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<td>595–609</td>
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<td>325</td>
<td>Bedrock on shoreline</td>
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<td>Boulder moraine</td>
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5.6.3.1 Glen Derry summary

The constant moraine modelling suggests the oldest boulder in Glen Derry best represents the moraine formation age and has not been adversely impacted by sediment cover. The age is also unlikely to have suffered from snow shielding. This age is consistent with the RSF/moraine ages from the nearby corrie that may have acted as a source area, providing additional evidence that the oldest boulder is the most representative of deglaciation. This age is contemporaneous with those from many other sites in Scotland interpreted to mark Younger Dryas glaciation.

However, the timing of deglaciation varies with the different production rates; during the Younger Dryas (GPR), at the onset of the Younger Dryas (NWH 11.6) and in the Lateglacial Interstade (LL; NWH 12.2). Including uncertainty the local production rates place the oldest age early within the Younger Dryas or late within the Lateglacial Interstade. Note that using the NWH 12.2 and LL production rates would force many other sample sites currently thought to be Younger Dryas in age into the Lateglacial Interstade. While the ages are not dissimilar to the other sites in Scotland interpreted to be Younger Dryas (Figure 5.18), the oldest age is clearly
older than the ages used to derive the NWH and LL production rates. This may mark Younger Dryas glaciers reaching their maximum position earlier within the Cairngorms, a possibility that is discussed further in Chapter 7: Section 7.2.4.3. Alternatively, the ages may represent the growth or persistence of an ice mass during the Lateglacial Interstadte and the area becoming ice free during the Younger Dryas. This is thought to be unlikely given the Younger Dryas was the most severe and prolonged period of climate deterioration in the Lateglacial period (Brooks et al., 2012). The conclusion from Glen Derry is that a Younger Dryas readvance glacier may have existed early within the Younger Dryas Stade. This is supported by the geomorphological evidence (Chapter 4) and requires further investigation with reference to the palaeoclimate (Chapter 6).

5.6.3.2 Glen Geusachan summary

The constant moraine matrix modelling could not be used to rule out sediment cover at the Glen Geusachan site. However, it is deemed unlikely that sustained sediment cover shielded all three boulders to greatly reduce their ages, and the effect of snow shielding at the site is thought to be minimal on the age interpretation. This suggests, with the use of all the available production rates that a Younger Dryas glacier existed in Glen Geusachan. The most likely reason for the difference in age between the Younger Dryas moraine ages and the published older and more varied ages near the meltwater channels is thought to be due to a more limited Younger Dryas glacier readvance position, which is in keeping with new geomorphological mapping of the area (Chapter 4: Section 4.4.3). The new dating thus facilitates the prospect of a valley glacier in Glen Geusachan during the Younger Dryas to be evaluated with respect to the palaeoclimate.

A matter of further debate is the difference in age between the oldest age in Glen Derry and the oldest age in Glen Geusachan (Figure 5.17). This suggests Glen Geusachan deglaciated later which is unlikely given their close proximity. The most likely suggestion is that the glaciers were contemporaneous and the younger ages in Glen Geusachan reflect a short period of moraine stabilisation or sediment cover, a similar pattern to the younger ages in Glen Derry. Despite small uncertainties, the ages clearly suggest a separate period of glaciation to the older RSF/moraine ages from Strath Nethy (recalculated NWH 11.6: 20.0±1.3 ka to 16.7±1.3 ka) and Lairig Ghru (recalculated NWH 11.6: 17.7±1.6 ka to 15.0±1.0 ka) which put deglaciation of
the lower glacial breaches within the Dimlington Stade (Ballantyne et al., 2009a). The ages in this study are closer to the Coire an Lochain (Cairn Gorm) ages attributed to a Younger Dryas glacier (Kirkbride et al., 2014).

5.7 Chapter Summary

Six new $^{10}$Be samples from Glen Geusachan and Glen Derry yielded ages that put deglaciation of the upper valleys of the Cairngorms in the context of the Lateglacial Interstade and the Younger Dryas. The robustness of the ages to sediment cover and snow shielding has been modelled and discussed. Results suggest snow shielding is unlikely to be a major influencing factor in the interpretation of the ages. However, sediment cover has greater potential to adversely interfere with apparent exposure ages, causing them to no longer reflect the true age of moraine formation. In Glen Derry, modelling of moraine matrix erosion has been successful and suggests the ages are not adversely affected by moraine matrix erosion and the exhumation of fresh boulders. Thus the true age of the Glen Derry moraine formation is likely to be represented by the oldest boulder and this suggests moraine formation at the onset of the Younger Dryas. Modelling of the Glen Geusachan samples did not assist in constraining the upper moraine formation age. Thus earlier deposition and subsequent shielding prior to exhumation cannot be ruled out, but moraine formation during the Younger Dryas is favoured. While the impacts of geomorphic effects, particularly sediment cover, are commonly overlooked, and the use of statistical techniques requires a high number of samples, the use of paired sampling deserves further exploration. The ages at both sites fall within the range of ages given for other sites in Scotland interpreted to represent Younger Dryas glaciers. This suggests that while the production rate may vary, the ages represent a contemporaneous readvance event. The new ages indicate deglaciation occurred late in the Lateglacial Interstade/early in the Younger Dryas Stade, thus the hypothesis of Younger Dryas valley glaciation cannot be rejected, indeed it agrees with the geomorphological evidence and requires further investigation with reference to palaeoclimate.
6 Glacier Reconstruction, Topoclimatic Factors and Palaeoclimate

This chapter builds on the new geomorphological mapping and surface exposure dating from Chapters 4 and 5. The new and existing dating evidence, combined with morphostratigraphy and landsystems derived from the geomorphological mapping, are used to suggest contemporaneous margins of glacier readvance. The reconstruction of the glaciers and their ELAs are presented, and the style of glaciation is evaluated. Any variation within the ELAs is evaluated with the modelling of topoclimatic factors such as variation in solar radiation, avalanches and snow redistribution by wind. This facilitates a better understanding of how the glaciers relate to the climatic ELA and the regional precipitation gradient. A synthesis of the pattern of ice-sheet deglaciation will be presented in Chapter 7: Section 7.1, along with the palaeoclimate implications.

6.1 Establishing a Chronology of Glacial Events

A landsystem has been commonly defined as an area with ‘common terrain attributes, different to those of adjacent areas’ (Evans, 2005 and references therein: p.1). Multiple different landsystems have been identified within the Cairngorm Mountains and are discussed below. The structure for the identification of landsystems associated with Younger Dryas glaciation was in part taken from previous work (Lukas, 2006; see Chapter 3: Section 3.2.2.5). It was important to remain objective while grouping the sites into the different landsystems; consequently some sites have not been assigned a landsystem and have thus been discussed individually in Section 6.1.4. The approach has been useful for distinguishing synchronous events and attributing specific glacier limits to climatic events. Together the combination of mapping, dating and landsystems has been used to generate a summary map of readvance margins, candidate readvance margins and ice-sheet deglaciation margins.
6.1.1 Corrie landsystem

A robust corrie landsystem exists that can include all or some of the following: a well-defined limit to glacially transported boulders, with arcuate and often bouldery moraines within this boundary (Figure 6.1). Often a particularly large or continuous arcuate ridge is at or near to the boulder limit, and the lateral/vertical limits of the glaciers are often distinguishable by boulder ridges. Within the corrie, the walls are covered by immature talus and alluvial fans. Outside the glacier limits, particularly where the corrie mouth descends steeply, there is often a sharp change to soliflucted surfaces.

This landsystem is characteristic of the highest corries of the Cairngorms, but also includes the lower valley heads and corries in the eastern Cairngorms, such as upper Slochd Mòr. The best examples of this landsystem are Coire an t-Sneachda (with a corrie-floor altitude of approximately 925m), Coire an Lochain (Cairn Gorm 900m, Braeriach 1000m and Ben Macdui 950m), Coire an t-Saighdeir (950m), Coire na Ciche (900m), Coire nan Clach (Beinn a’ Bhuird 900m) and Coire Bhrochain (950m) which all have corrie floors above 900m. Coire Dhondail and the head of Slochd Mòr are also included within this landsystem, although they have slightly lower corrie floors at c.830m. Coire na Saobhaidh is a very shallow corrie on the west side of lower Glen Derry. The small amount of evidence available here fits with the Younger Dryas corrie landsystem; however, it is lower than most corries with a headwall under 900m and moraine ridges at 650m. In the east, the corries of Lochan nan Gabhar and Sgor Riabhach contain less distinguished geomorphological evidence than the higher corries mentioned above; however, the geomorphology fits the same landsystem description. The broadly similar altitudinal range of these corries and the similarity in geomorphological evidence indicate their last major glaciation was most probably from one concurrent readvance event. The absence of solifluction activity within the glacier limits suggests this was during the last period of climatic deterioration – the Younger Dryas (Sissons, 1972; Benn and Ballantyne, 2005; Ballantyne, 2007a). This has been supported by surface exposure dating within Coire an Lochain (Cairn Gorm) (Kirkbride et al., 2014).

Some important variation in corrie altitude is present within this group; this is evaluated with the use of topoclimatic factors later within the chapter. The locations of these Younger Dryas readvance margins are shown in Figure 6.3.
Figure 6.1 Younger Dryas corrie landsystem
6.1.2 Valley landsystem

A separate landsystem occurs within many of the valley heads. It comprises sharp-crested, fragmented moraine crests, often aligned obliquely across the valley sides, with the central frontal section typically eroded by postglacial fluvial activity (Figure 6.2). Local areas where the orientations of the features are not discernible do occur; however, more often the direction of ice-margin retreat can be distinguished. Meltwater channels are sometimes identifiable between the ridges, and postglacial gullies commonly dissect the ridges. Relatively small boulders are often found on top and embedded within the ridges; larger boulders are rare or have local explanations for their presence, e.g. rockfall sources. These landforms can be distinguished from other features based on their closely spaced nature, well-preserved positive sharp-crested moraine form, and strong alignment; whereas moraines outside these limits tend to be broader, shallower, less concentrated and their profile can often be enhanced by meltwater erosion or postglacial gullying. In addition, outside of the limits, preserved lateral moraines on the valley sides are less common and solifluction is often present; whereas solifluction is limited within the readvance limits.

This landsystem is found in the upper parts of Glen Derry, Glen Dee, Glen Geusachan, Glen Avon and Glen Eidart. The altitudes of the valley floors where this landsystem is found range from 500–900m. The landforms often retreat to the valley head; however, they have not been identified retreating onto the plateaus above. These valleys all have their head or side walls on the eastern and/or northern side of high plateau areas, consistent with sites favourable for readvance at the onset of climate deterioration (Figure 6.3). This landsystem is not found east of Glen Derry. The new cosmogenic surface exposure ages presented within Chapter 5 are taken from moraine ridges within this landsystem in Glen Geusachan and Glen Derry. The new minimum ages indicated Younger Dryas glaciation of these sites was probable. Thus, given the spatial pattern of the landsystem within the Cairngorms, the recognition of this and similar landsystems in the British Isles as being characteristic of the Younger Dryas (Lukas, 2006) and the new surface exposure ages within this landsystem indicating a later readvance, it is thought this landsystem represents the readvance of Cairngorm glaciers during the Younger
Dryas. The location of these Younger Dryas readvance limits can be seen in Figure 6.3.

Importantly, never does a valley with the Younger Dryas valley landsystem contain the Younger Dryas corrie landsystem at the valley head or in contributing corries. Thus the landsystems are mutually exclusive, suggesting they formed simultaneously, otherwise it would be reasonable to expect some cross-cutting of the landsystems. One location that supports this is upper Glen Dee; Coire Bhrochain contains the Younger Dryas corrie landsystem, while the rest of the Garbh Coire corries do not contain the Younger Dryas corrie landsystem as they fed ice into upper Glen Dee, which contains the Younger Dryas valley landsystem. The only location at which both landsystems merge is through the Derry-Loch Etchachan-Avon valley system. Here the glacially transported boulders dominated the corrie and col floor near Loch Etchachan, whereas the larger sediment-rich moraine crests typical of the Younger Dryas valley landsystem dominate the lower valley floors of Glen Avon and Glen Derry. This may be explained by thinner cold-based ice on the corrie floor above c.900m and thicker more dynamic ice within the lower (c.700m) valleys; this is discussed further in Chapter 7: Section 7.2.2.4.
Figure 6.2 Younger Dryas valley landsystem
6.1.3 Dimlington Stade deglaciation

The landforms outside these Younger Dryas readvance landsystems are interpreted to represent ice-sheet deglaciation. These landforms tend to be larger and more rounded in appearance; however, there are many localised factors that impact on the landforms, making the application of a single landsystem model inappropriate. Nevertheless, some key similarities are discussed below. The major ice-sheet deglaciation margins presented in Chapter 4 are included on the summary map (Figure 6.3).

The landforms associated with external ice on the outer Cairngorm Mountain flanks are distinctive for their size, thick drift and shaping by meltwater channels. These lateral meltwater channels record the surface gradient of the ice and suggest the ice was cold based, forcing water to flow at the ice margin (Ó Cofaigh et al., 2005; Bennett and Glasser, 2009). Ice-dammed lakes are a common feature on both the northern and southern side of the Cairngorms. The shallower northern lakes of Lairig Ghru and Einich have large delta surfaces and well-defined glacier margins. The landforms associated with the deeper southern lakes such as in lower Glen Dee, Glen Derry and Glen Quoich are soliflucted lake shorelines and high perched deltas. The ice-marginal positions, with the exception of near the Glen Quoich and Glean an t-Slugain intersection, are less obvious.

The stagnation of ice appears to have been limited within the Cairngorms, with only localised areas at the exit of Glen Luibeg and near Dubh Lochan in Lairig an Laoigh. These were most probably where cols cut off ice supply, leading to ice stagnation. Thus evidence suggests deglaciation was largely active and in synchrony with changes in the climate. Large broad moraines with rounded crests occur in both the northern (e.g. Gleann Einich) and southern (e.g. Glen Dee and Glen Derry) Cairngorms that mark stages of glacier retreat during the Dimlington Stade.

The plateaus can be grouped, based partially on their altitude, topography and geomorphology. The higher rounded summits of Ben Avon (1171m) and Cairn Gorm–Cairn Lochan (1245m and 1215m respectively) are heavily soliflucted. The broad low 850–950m plateaus of Mòine Mhòr and Mòine Bhealaidh have meltwater
and ice-moulded bedrock evidence of ice flow northwards during ice-sheet glaciation. The sheltered north-east side of the Ben Macdui (1309m) plateau contains evidence of ice erosion, and to a lesser extent so do Beinn Bhrotaín (1108m) and Monadh Mòr (1113m). Braeriach (1296m) is one of the highest and broadest plateaus within the Cairngorms; the edges possess particularly well-soliflucted areas but the timing of the last phase of glaciation in unclear. No evidence was found to indicate the most recent phase of glaciation on the high Beinn a’ Bhuird (1197m) plateau or the narrower summit of Cairn Toul (1291m). These higher plateaus, particularly the broader Ben Macdui plateau, may have supported cold-based ice during the Younger Dryas.

Many solifluction sheets and lobes in Britain are relics of a colder climate (Ballantyne, 2008); however, some smaller solifluction forms have been active during the Holocene (Sugden, 1971; Mottershead, 1978; Ballantyne, 1986; Ballantyne, 2013). The extent and development of solifluction varies throughout the Cairngorms. Some of the most well-developed solifluction occurs on the northern flanks of Braeriach, on the spurs between the northern corries and the slopes of Ben Avon. Solifluction is absent or very limited within the Younger Dryas land system limits. This indicates that the more developed solifluction must have occurred during the Younger Dryas or prior periods of climate deterioration. This corresponds to other sites in Scotland where the solifluction of valley sides stops abruptly at the Younger Dryas glacier limits (e.g. Sissons, 1972; Benn and Ballantyne, 2005; Ballantyne, 2007a). Despite local factors, including elevation, slope gradient, moisture content and snow cover, influencing solifluction development (Matsuoka, 2001; Harris et al., 2008), the location of the more developed solifluction is in accordance with sites that are thought to have deglaciated early, based on the relative chronology of ice-sheet retreat. Thus it is likely that the most well-developed solifluction sites began development once deglaciated during ice-sheet retreat, and have been further developed during cold periods since.
6.1.4 Sites of uncertainty

Within the Cairngorms there are corries and valleys that contain a unique set of landforms specific to the site, and therefore cannot be included in the landsystems above. However, the sites may have supported Younger Dryas glaciers, so where evidence is sufficient to derive a glacier margin and a glacier reconstruction is possible, a candidate margin/glacier has been generated.

Coire Bogha-cloiche and Coire Ruadh contain deposits of potential talus-glacier origin. Their geomorphological evidence does not fall within the Younger Dryas corrie landsystem, yet the neighbouring Coire an Lochain is included and thought to have supported a Younger Dryas glacier. In Coire Bogha-cloiche the deposit on the northern side may be indicative of a thin ice mass during the Younger Dryas. Alternatively, the shallow nature and west-facing aspect of the corrie may have prevented a Younger Dryas glacier forming within this corrie. Coire Ruadh is of similar altitude and orientation to the adjacent Coire an Lochain, with only slight differences in how their headwalls align with the Braeriach plateau. The absence of a Younger Dryas glacier in Coire Ruadh is difficult to explain, as seen in previous attempts to model snow blow by Purves et al., (1999). Both the Coire Bogha-cloiche and Coire Ruadh sites are marked as candidate readvance sites in Figure 6.3.

Similarly, Coire Garbhlich in the western Cairngorms does not contain geomorphological evidence typical of the Younger Dryas landsystems. However, lateral ridges and deposits likely to be from valley glaciation are present, but the timing is unclear. Thus a candidate readvance margin has been included on Figure 6.3. A small shallow corrie exists on the eastern side of Cnap a Chlèirich. Deposits and channel erosion occur beneath the crag, but the landforms are not typical of the Younger Dryas landsystems. However, the site’s altitude and position on the east side of the Beinn a’ Bhuird plateau make it favourable for readvance glaciation and thus a candidate margin has been included on Figure 6.3.

The north-facing Coire Beanaidh has a corrie floor above 900m, with a headwall at c.1250m. The lack of strong evidence for a readvance within this corrie is at odds with its altitude and orientation. The only characteristic it lacks is a sizable plateau surface behind the headwall. Either the geomorphology has been misunderstood or the lack of plateau-fed ice or snow blow from the plateau was a contributing factor
in restricting Younger Dryas glacier formation. Within Glen Derry, Coire an Lochain Uaine is a corrie with large deposits which are not typical of the Younger Dryas corrie landsystem and the positioning of the ridges is not consistent with a local corrie glacier; this was supported by the failure to reconcile the ridges with a glacier reconstruction. This site is therefore favoured to mark an earlier period of glaciation or a mass movement event (Kirkbride and Gordon, 2010). An additional hypothesis is that the corrie deposits may be partially formed as a cirque infill deposit from the Derry glacier during ice-sheet deglaciation, such as those in the Khibiny Mountains, Kola Peninsula, Russia (Hättestrand et al., 2008).

The north-facing Coire Cas (near Cairn Gorm) has a corrie floor above 1000m yet does not possess moraine crests from a glacier readvance. This may be because the relatively shallow rounded backwall of c.100m in height stopped a glacier forming. Alternatively, a small thin cold-based Younger Dryas glacier may have existed, which either failed to create a clear geomorphological signature or has been destroyed by postglacial activity. These last three sites of uncertainty have not been included as candidate readvance margins or glaciers because of the lack of geomorphological evidence to constrain a margin, or difficulty in reconciling the geomorphology with a glacier reconstruction.
Figure 6.3 Summary of the glacier margins within the Cairngorm Mountains

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6.2 Younger Dryas Glacier Reconstruction

Glacier reconstructions have been undertaken based on the geomorphological evidence from Chapter 4 and, where altitudinal constraints are absent, guided by modelling of the former ice surface (Benn and Hulton, 2010). The modelling was based on modern-day shear stress values (50–100kPa) and calculated shape factors; further details can be found within the methodology (Chapter 3: Section 3.4). All the palaeoglaciologists that exhibited the Younger Dryas corrie or valley landsystems were reconstructed; where uncertainties over the glacier style exist, both valley- and plateau-style reconstructions are presented. In addition, glaciers were also reconstructed for the Coire Bogha-cloiche, Coire Ruadh, Coire Garbhlagh and Cnap a Chlèirich candidate sites where limited evidence of former glacier positions exists, but the timing of glaciation based on geomorphological evidence is uncertain. These are referred to as candidate glaciers or candidate reconstructions below. Glaciers were not reconstructed for the Coire Beanaidh, Coire Cas and Coire an Lochain Uaine (Derry Cairngorm) sites because no or insufficient evidence was available to reconstruct a glacier. The reconstruction of the additional candidate glaciers allows the likelihood of their existence, based on ELAs and topoclimatic factors, to be assessed.

6.3 Younger Dryas Glacier ELAs and Palaeoclimate

The ELAs calculated for the glacier reconstructions are presented in Figure 6.4, Figure 6.5 and Table 6.1. These have been calculated using the Area-Altitude Balance Ratio (AABR) technique with a balance ratio of 1.9 ± 0.81 for mid-latitude maritime glaciers (Rea, 2009) and 1.5 ± 0.4 for Western Norway glaciers (Rea, 2009); the latter suggested for the British Isles during Younger Dryas by Carr et al. (2010). Other balance ratios are included to allow comparison with existing work elsewhere within the British Isles. The ratio 1.9 ± 0.81 has been used within the following analysis as this is thought to encompass all possible likely balance ratios. This generates corrie-style ELAs between 734m (727–745m) and 1050m (1041–1062m) and plateau-style ELAs of between 734m (727–745m) and 1106m (1091–1131m). The ELAs for the glaciers including their avalanche area above the ELA are also presented – these are discussed in Section 6.3.2.1.
The ELAs have been given in metres above present-day sea level. Their original position during the Younger Dryas would have been somewhat different due to glacio-isostatic uplift (cf. Firth and Stewart, 2000; Smith et al., 2006). There have been substantial changes in sea level since the Younger Dryas as a result of both isostatic changes (change in height of the land) and eustatic changes (change in the volume of water in the oceans); these have both increased since the former glaciers existed (Shennan and Horton, 2002; Peltier et al., 2002).

The precipitation values (Table 6.2) have been calculated using the Ohmura et al. (1992) equation based on the global relationship between temperature and precipitation at the ELA. The values have been calculated using a balance ratio of 1.9, lapse rate of 0.0065°C per m, and a July temperature of 6.8°C at Abernethy Forest (Brooks et al., 2012). An upper and lower precipitation estimate has been calculated using the cumulative errors from the balance ratio (±0.81), lapse rate (±0.0005°C per m), temperature (±1°C) and a standard error for the equation of ±200mm a\(^{-1}\) (Ohmura et al., 1992). Combining these errors produces a large range of precipitation values (Table 6.2). The precipitation values are the required water equivalent at the ELA based on summer precipitation and winter balance, the latter including wind-blown snow and avalanche inputs. The winter balance and ablation gradients, based on the work of Schytt (1967), Carr and Coleman (2007) and Carr et al. (2010), have been calculated. The winter balance was also calculated as it is thought to give a better indication of the ablation gradient (Carr et al., 2010). The ablation gradients are relatively shallow < 5 a\(_z\) mm m\(^{-1}\), indicating relatively low change in accumulation and ablation over the glacier profile, typical of glaciers in cold polar environments (Schytt, 1967; Andrews, 1972; Carr and Coleman, 2007; Carr et al., 2010). The precipitation values have also been calculated using a climate-specific model for Scotland during the Younger Dryas (Golledge et al., 2010). The neutral, winter and focused precipitation scenarios are presented (Table 6.3) with the cumulative errors (including balance ratio, temperature, lapse rate). The precipitation values are not discussed further at this stage because of their variability and the influence of topoclimatic factors on the glacier ELAs, and thus on the derived precipitation values.
Figure 6.4 Valley-style reconstructed glaciers and candidate glaciers

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Figure 6.5 Plateau-style reconstructed glaciers and candidate glaciers

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Table 6.1 Glacier ELAs

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<tr>
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<td>1057</td>
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<tr>
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<td>956</td>
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<tr>
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<td>1078</td>
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</tr>
<tr>
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<td>1075</td>
<td>1090</td>
</tr>
<tr>
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<td>33 p</td>
<td>0.31</td>
<td>1139</td>
<td>1163</td>
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</table>

Suggested balance ratios for mid-latitude maritime glaciers and Western Norway from Rea (2009). Other balance ratios given to facilitate comparison with studies elsewhere within the British Isles.* Refers to glacier style v = valley-sourced and p = plateau-sourced.
Table 6.2 Precipitation values and ablation gradients (precipitation values based on equation from Ohmura et al., 1992)

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Number</th>
<th>Style</th>
<th>ELA (ABBR)</th>
<th>ELA Temp °C</th>
<th>Pa mm a</th>
<th>Winter Balance bw mm a</th>
<th>Ablation Gradient (Pa) az mm m</th>
<th>Ablation Gradient (WB) az mm m</th>
<th>Sea level Pa mm a</th>
<th>Sea level Pa mm a</th>
<th>Upper Precipitation El (°C)</th>
<th>Lower Precipitation El (°C)</th>
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<tbody>
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<td>v</td>
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<td>922</td>
<td>1.617</td>
<td>1090</td>
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<td>869</td>
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<td>1909</td>
</tr>
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<td>v</td>
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<td>999</td>
<td>960</td>
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<td>893</td>
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<td>1094</td>
<td>4.45</td>
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</tbody>
</table>

Based on temperature of 6.8 ± 1.0°C at 230m, lapse rate of 0.0065 ± 0.0005°C per m and AABR ELAs BR=1.9 ± 0.81. The upper and lower precipitation values are based on the cumulative errors of temperature, lapse rate and ELA. The sea-level equivalent precipitation values are converted using 5.8% per 100m (Ballantyne, 2002). * Refers to glacier style v = valley-sourced and p = plateau-sourced.
### Table 6.3 Precipitation values (based on equation from Gollledge et al., 2010)

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Number</th>
<th>Style</th>
<th>At ELA</th>
<th>Neutral (1.0)</th>
<th>Winter focused (0.8)</th>
<th>Summer focused (1.4)</th>
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<td></td>
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<td>mm a</td>
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The sea-level equivalent precipitation values are converted using 5.8% per 100m (Ballantyne, 2002).
6.3.1 Qualitative analysis of Younger Dryas glacier spatial distribution, retreat dynamics and palaeoclimate

The spatial pattern of glaciers and their ELAs, and the retreat patterns of individual glaciers can inform our knowledge of palaeoclimate (e.g. Mitchell, 1996; Lukas and Benn, 2006; Coleman et al., 2009; Bendle and Glasser, 2012). Previous studies within the Cairngorms suggested limited sites where the moraines may represent thrusting and non-ice-marginal moraine-building processes (Bennett, 1996; Midgley, 2001); however, the moraines are largely thought to be ice-marginal recessional moraines (Bennett and Glasser, 1991; Bennett, 1996; and Midgley, 2001). Thus most of the moraines can be joined to form palaeo-ice fronts of glacier retreat (Lukas and Benn, 2006), implying the glaciers within these valleys retreated actively, and the moraine spacing and pattern of retreat can be used to identify important palaeoclimatic factors. A valuable example is the joining of ice-marginal moraines within the Eidart valley; this showed retreat into Coire Mharconaich and the north-west valley head, rather than onto the plateau to the north or the north-east valley head (Chapter 4: Section 4.4.5). This indicates valley-style glaciation, at least during glacier retreat, and indicates snow accumulation from the west. Similarly, the moraine geometry suggests the readvances in Slochd Mòr, Coire an t-Sneachda, Coire an Lochain (Cairn Gorm) and Coire an t-Saighdeir were sourced from the south-west sections of their respective corries. This greater accumulation in the south-west parts of the corrie is likely to be due to snow redistribution by winds from the south-west.

Many of the glaciers, particularly those with lower ELAs, have large plateaus on their southern and western sides. These may have been important for snow redistribution and possibly plateau-fed ice, as seen in the plateau-style reconstructions (Figure 6.5). Examples of corries that are likely to have benefitted from plateau-blown snow are Slochd Mòr, Lochan nan Gabhar and Sgor Riabhach in the eastern Cairngorms. Also Glen Eidart in the western Cairngorms has a large plateau area surrounding it; although, as discussed above, the western side was probably the most important for snow accumulation. The larger glaciers such as those in Glen Geusachan, upper Glen Dee, Glen Derry and Glen Avon have some important common attributes. They all have large plateau areas surrounding them,
particularly on their south and western sides. They have steep valley sides which would have assisted in shading the glacier surface and reducing wind speeds within the valleys, thus creating favourable locations for snow accumulation from south-westerly winds. In addition, the majority of the perimeter of these former glaciers would have been surrounded by the steep valley sides which would have provided regular avalanches onto the glacier surface. Importantly, the avalanches would have occurred not only within the accumulation zone of the glacier but also along its full length.

These properties are different from some of the characteristics of the higher corries. The corrie glaciers are often shaded, offered wind protection, and subject to regular avalanches in the accumulation zone of the corrie where the back and side walls surround the glacier. However, the lower section of the glacier can advance beyond this zone of favourable topoclimatic conditions and be exposed to increased radiation, higher wind speeds and less avalanche accumulation. The glacier within Coire an t-Saighdeir readvanced out of this zone, with its lower part beyond the confines of the corrie back and side walls. This may be particularly common where the corrie has not developed a full protective semi-circular shape or if the corrie is too shallow or open to offer favourable topoclimatic conditions. It is also a feature of the Cairngorms that few of the corries gradually merge into the glacial troughs; instead, partially due to the Cairngorms being a landscape of selective linear erosion (Rea, 1998), many corries have a hanging relation to the much lower valley floors. For example, the northern and western Braeriach corries descend rapidly into Gleann Einich, Coire an t-Saighdeir into Glen Dee, and Coire an Lochain (Ben Macdui) into Glen Luibeg. Noticeably, where the corries do gently converge into sheltered glacial troughs, e.g. An Garbh Choire/upper Glen Dee, a larger readvance glacier has been recognised in the geomorphology. Thus it is probable that valleys such as upper Glen Dee, Glen Avon, Glen Derry and Glen Geusachan provided a continued zone of favourable topoclimatic conditions that allowed glaciers to readvance further down-valley.

Some evidence for these concepts has been recognised in previous studies. In the lower part of Glen Geusachan, the alignment and spacing of the ice-marginal retreat moraines indicate differential retreat between the northern and southern valley sides (Bennett and Glasser, 1991). The curved ice fronts are further down...
the valley on the southern side; Bennett and Glasser (1991) suggest this may be due to differences in solar radiation impacting on ablation. It may also include greater wind-blown snow and avalanche accumulation on the southern side. Within upper Glen Derry, part of the glacier retreated into the south-west corner of Coire Etchachan (Midgley, 2001). This again indicates the favourable conditions on the southern side, most likely a combination of radiation shading and greater snow accumulation from the overlooking plateau.

Multiple corries have two or more ice-margin positions within them and the valleys have numerous moraines, most of which are interpreted to be ice marginal. This indicates the active retreat of glaciers during the Younger Dryas, and that they were dynamically responding to climate fluctuations during retreat. No particularly large terminal moraines are found, which is consistent with the relatively short-lived climatic event. In addition, it is likely there were low debris fluxes due to the lack of extra glacial source areas and low basal debris loads due to restricted basal erosion. Some corries do have a large arcuate moraine at the limit of the glacier; this may represent a stable position or the transportation of existing material within the corrie to the glacier limit early on. This pattern, of the valleys containing numerous former ice-margin positions and the corries relatively few, was also found in north-west Scotland (Lukas and Benn, 2006) and on Skye (Benn, 1990; Benn et al., 1992). This has been hypothesised to be due to the different ways smaller and larger glaciers respond to climate fluctuations (cf. Lukas and Benn, 2006). The flutes in Garbh Choire Dhàidh, upper Glen Dee suggest the last phase of retreat in this valley may have been relatively rapid as they have not been extensively modified by moraine building during glacier retreat; similar observations were found on Skye and in the Northern Highlands (Benn, 1990). This differs from Glen Geusachan and Glen Eidart where recessional moraine building continued into the valley heads. This may represent retreat at slightly different times, or a change in the thermal regime or sediment supply of the Glen Dee glacier as it thinned and retreated into the higher corries.

It is notable that there are variations in the ELAs within the Cairngorms. The valley-style reconstructions have ELAs between 734m (727–745m) and 1050m (1041–1062m) and plateau-style ELAs of between 734m (727–745m) and 1106m (1091–1131m). Variations in palaeo-ELAs are seen in Snowdonia between 380m and
837m (Bendle and Glasser, 2012), Mull between 211m and 551m (Ballantyne, 2002) and Creag Meagaidh Massif between 601m and 703m (Finlayson, 2006). Some variation in the ELAs is to be expected due to topoclimatic factors such as snow redistribution, avalanche accumulation, radiation inputs and precipitation gradients. For this reason, it is best to accept models such as that of Dahl and Nesje (1992), where there is a regional temperature-precipitation ELA and a more variable ELA associated with individual glaciers that reflects the combination of topoclimatic factors. Consequently, combined ELAs from icecaps and icefields that include all the outlet glaciers have ELAs close to the regional ELA; whereas corrie, valley and outlet glaciers with favourable conditions have lower ELAs, and glaciers with less favourable factors have higher ELAs. To gain a better understanding of the regional ELA and with the aim of explaining some of the variation in ELAs within the Cairngorms modelling of the topoclimatic factors has been undertaken.

6.3.2 Quantitative analysis of topoclimatic factors

The modelling of topoclimatic factors is presented within this section and extra details of the methods used can be found within Chapter 3 (Section 3.6). The main topoclimatic factors thought to impact on glaciers within the Cairngorms are snow blow, avalanche area, radiation and precipitation: these are outlined in Figure 6.6, although, note that this is not an exhaustive list. The modelling has been carried out for the corrie- and plateau-style reconstructions of the glaciers, and also the additional candidate reconstructions. This allows the topoclimatic factors of the different glaciers to be compared, and also comparisons to be made between the glaciers and the candidate glaciers. Where appropriate, the modelling has been undertaken for the glaciated terrain, defined here as the present-day DEM including the ice-surface topography of the Younger Dryas glacier reconstructions, and also for the unglaciated terrain which only utilises the present-day DEM. The results are presented below, first for the individual factors, and then combined using multiple regression analysis in Section 6.3.3.
Figure 6.6 Diagram to illustrate the topoclimatic factors and the contributing factors. Note precipitation is usually similar between neighbouring glaciers. This is not an exhaustive list. Other factors can impact on the ELA, such as wind speeds impacting on glacier melt, calving margins and debris cover.

6.3.2.1 Avalanche areas

Avalanches are a well-documented source of additional snow accumulation for glaciers from the surrounding topography, particularly in high mountainous environments such as the Himalayas (Benn and Lehmkuhl, 2000). The potential avalanche area above and below the ELA has been digitised in GIS; details can be found in Chapter 3: Section 3.6.2. These have then been normalised to the glacier area to create avalanche ratios (above the ELA, below the ELA and total) for each glacier (Table 6.4 and Figure 6.7). This allows the relative potential impact of avalanches on the glacier ELA and mass balance to be analysed. The largest (>0.50) avalanche ratios for above the ELA are Coire na Saobhaidh (Glen Derry), Coire na Ciche (Beinn a’ Bhuiird), Lochan nan Gabhar and Coire Bhogha-cloiche. The particularly high value for Coire na Saobhaidh (Glen Derry) may have aided its existence at this altitude and explain its lower ELA. The lower ratios are on the plateau-fed glaciers and also Eidart, Glen Derry and Glen Avon. This is expected as there are relatively small areas of overlooking slopes in the accumulation zones of these glaciers.
The avalanche areas above the ELAs have been incorporated into the ELA calculations (Table 6.1), as in Benn and Ballantyne (2005). This causes the Glen Dee glacier ELA to rise by 58m if reconstructed as a valley glacier, or 8m if reconstructed as a plateau glacier. The Geusachan glacier ELA increases by 27m or 12m for valley and plateau reconstructions respectively. The ELAs of Slochd Mòr, Coire na Ciche (Beinn a’ Bhuird), Lochan nan Gabhar, Coire Bhrochain, Coire an t-Saighdeir, Coire Bogha-cloiche and Coire Ruadh all increase by more than 30m for their valley-style reconstructions when the avalanche area above the ELA is included.

The additional snow accumulation by avalanches will not only be advantageous above the theoretical altitudinally uniform reconstructed ELA; instead, the ELA will vary locally on the glacier and some important accumulation from avalanches may occur below the reconstructed ELA. Thus the avalanche areas and ratios for below the ELA and full glacier perimeter can be analysed. In most cases the pattern is similar, but Slochd Mòr, Coire an t-Sneachda, Glen Derry and Avon, Geusachan, Eidart and Glen Dee, amongst others, benefit from including overlooking slopes, regardless of whether they are above or below the ELA. These glaciers all have valley or corrie sides along the lower perimeter of the glacier, unlike some corries such as Coire na Saobhaidh that only have overlooking slopes at the backwall. The inclusion of both above- and below-the-ELA avalanche areas causes the lower valley glaciers such as Glen Derry and Glen Avon, Glen Geusachan and Glen Eidart to have total avalanche ratios that are more in line with the corrie glaciers (Figure 6.7).
Table 6.4 Glacier avalanche source areas and ratios to glacier size

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Number</th>
<th>*Style</th>
<th>ELA AABR BR=1.9 (m)</th>
<th>Glacier Area (km²)</th>
<th>Avalanche Above ELA (km²)</th>
<th>Avalanche Below ELA (km²)</th>
<th>Avalanche Whole Glacier (km²)</th>
<th>Ratio</th>
<th>Ratio</th>
<th>Ratio</th>
</tr>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td></td>
<td>0.68</td>
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<td>1.15</td>
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<td></td>
</tr>
<tr>
<td>Cnap a’ Chlèirich</td>
<td>6 v</td>
<td></td>
<td></td>
<td>1,009</td>
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<td>0.02</td>
<td>0.19</td>
</tr>
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<td>0.00</td>
</tr>
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<td></td>
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<td>0.12</td>
<td>0.50</td>
</tr>
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<td>0.03</td>
</tr>
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<td>0.01</td>
<td>0.03</td>
</tr>
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<td>0.14</td>
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</table>
Figure 6.7 Whole glacier avalanche ratios, subdivided into avalanche ratios for accumulation and ablation zones
6.3.2.2 Solar radiation and shading

Weather-dependent, net radiation (shortwave and longwave) is the most important component of the energy balance of most glaciers (Hock, 2005). The incoming part can be subdivided into shortwave (direct, diffuse and reflected from the terrain) and longwave radiation (sky and surrounding terrain) (Hock 2005). Only part of the incoming shortwave radiation is absorbed: the albedo of a surface controls the proportion of radiation that is reflected. Values for dry snow can be as high as 0.80–0.97 and as low as 0.15–0.25 for dirty ice (Paterson, 1994). This causes greater absorption of energy later in the ablation season as older ice is exposed. It also causes greater absorption in the lower ablation section of the glacier compared to the higher altitudes where clean snow survives throughout the ablation season (Benn and Evans, 2010).

Solar radiation separates into direct and diffuse when entering the atmosphere: diffuse being radiation that is scattered by particles in the atmosphere and is largely influenced by cloud cover (Hock, 2005). Slope angle, aspect and shading by the surrounding topography are important in calculating site-specific radiation values (Hock, 2005; Hock and Holmgren, 2005). Modelling of solar radiation has been carried out for the ablation season (1st May–30th September) in ArcGIS, for predominantly clear-sky and partially cloudy conditions: the latter representative of Aviemore’s modern climate. The partially cloudy scenario was based on data from May to October 1995–2000 which showed diffuse to global radiation ratios of 0.55–0.65 (Aviemore Weather Station – available through Met Office integrated data archive system – MIDAS). Modelling of the direct and diffuse parts of the incoming solar radiation for both the unglaciated terrain and the terrain including the reconstructed glaciers has been undertaken; further details can be found in Chapter 3: Section 3.6.3. Based on the assumption that there was relatively little modification of the terrain by Younger Dryas glaciers or postglacial activity, the unglaciated terrain can be used to represent the likely conditions that would have been present at the onset of Younger Dryas glaciation. The modelling utilised the Area Solar Radiation Toolbox in ArcGIS, based on the methods of Rich et al. (1994), Rich and Fu (2000), Fu and Rich (2000; 2002), which generates raster files of the total direct and diffuse radiation for the chosen terrain (Figure 6.8 and Figure...
A mean was then generated for the ablation area of each glacier. It should be noted the outputs are suitable for relative comparisons of solar radiation between each glacier, but are not appropriate for comparison between clear-sky and partially cloudy scenarios, or with other modelled or measured datasets; this is explained in Chapter 3: Section 3.6.3.

The spatial pattern of areas of high and low total radiation are broadly similar between the clear-sky and partially cloudy models; the latter having a greater proportion of diffuse radiation, thus causing deeper corries and valleys to exhibit lower relative radiation values. The direct radiation values are lowest on the north-facing surfaces and on flatter terrain immediately in front of steep cliffs, and highest on the south-facing slopes and plateau. The diffuse radiation values are lowest within the deepest corries and valleys: while the incidence angle is still important, diffuse radiation is less impacted by the aspect. The spatial pattern of the combined total radiation (diffuse and direct) for the partially cloudy scenario can be seen for the unglaciated terrain (Figure 6.8) and the valley-sourced glaciated terrain (Figure 6.9). A similar output, not shown, was generated for the plateau-sourced glaciers and clear-sky conditions. The inclusion of the glacier reconstructions within the DEM causes the shading immediately in front of the cliffs to shorten, as the glacier fills up the valley. This occurs to varying extents in all the corries and the valleys such as Glen Geusachan and Glen Dee. The corries with shallow headwalls or relatively high glacier surfaces within them provide little shading from radiation, e.g. Coire Mharconaich (head of Glen Eidart) (Figure 6.9). Comparison of the mean total radiation values of the glacier ablation areas between the unglaciated and glaciated terrain shows that the mean radiation generally increased when the glaciers were included within the DEM (Figure 6.10 and Figure 6.11). This is expected as the ice fills up the corries and valleys, and reduces the potential for topographic shading. It is also interesting that the inclusion of the glaciers in the glaciated terrain increases the inter-site variability.

Analysis of the glaciated terrain (Figure 6.9) shows that some of the glaciers with low-ablation area mean radiation values (Figure 6.10 and Figure 6.11) benefit from their north-facing aspect and low incidence angle to the sun’s radiation, such as Coire Dhondail, the northern corries and Lochan nan Gabhar. This indicates that it is not only backwall shading that is important, but also aspect and incident angle to
the incoming radiation. Unsurprisingly, the south-facing Coire Bhrochain (Braeriach) had a high radiation value, as did the south-facing Lochain Uaine (Ben Macdui) and Eidart glaciers. The glaciers within Coire Saighdeir (Glen Dee) and East Beinn a’ Bhuird also had relatively high radiation values: these both face east but the ablation zones are beyond the corrie walls and thus are no longer provided with shading. The larger glaciers, such as those in Glen Geusachan, Glen Dee, Glen Derry and Glen Avon, all have ablation areas facing east within relatively deep valleys, and have intermediate radiation values compared with the other glaciers.

The mean total radiation values for the ablation area of each glacier are included within the multiple regression analysis in Section 6.3.3. Interestingly, the diffuse portion alone has a stronger relationship with the glacier ELAs. This may suggest the climate was cloudier and direct radiation was less important. However, it was not used within the regression analysis because there is no justification to include only diffuse radiation.
Figure 6.8 Ablation season total radiation (direct and diffuse) for unglaciated terrain and partially cloudy scenario
Figure 6.9 Ablation season total radiation (direct and diffuse) for glaciated terrain (valley-sourced) and partially cloudy scenario
Figure 6.10 Mean ablation area total radiation values for partially cloudy conditions. Top graph is for unglaciated terrain and bottom for glaciated terrain (Note: absolute values are only valid for relative comparison and not for comparison with the clear-sky model or other datasets)
Figure 6.11 Mean ablation area total radiation values for clear-sky conditions. Top graph is for unglaciated terrain and bottom for glaciated terrain (Note: absolute values are only valid for relative comparison and not for comparison with the partially cloudy model or other datasets)
6.3.2.3 Snow-blow areas (manual digitisation)

The redistribution of snow by wind is documented in relation to modern glaciers in Svalbard (Hodgkins et al., 2006), Switzerland (Machguth et al., 2006), Greenland (Mernild et al., 2006), alpine terrain (Elder et al., 1991; Purves et al., 1998; Winstral et al., 2002; Schirmer et al., 2011) and palaeoglaciers (Mitchell, 1996; Dahl et al., 1997; Coleman et al., 2009; Bendle and Glasser, 2012). Within the British Isles during the Younger Dryas the inferred prevailing wind direction is from the south-west (Mitchell, 1996; Sutherland, 1984; Benn and Ballantyne, 2005; Ballantyne, 2007a; Coleman et al., 2009; Bendle and Glasser, 2012). Two different approaches have been used to generate snow-blow ratios and factors for each glacier; firstly, within this section, the results of the digitisation of the potential snow-blow areas are presented. Later the results of a GIS-based model of snow redistribution are presented. The details of the techniques used can be found within the methodology (Section 3.6.4).

The potential south-west snow-blow quadrant areas have been digitised for each glacier, as used by Benn and Ballantyne (2005). Potential snow-blow areas were deemed to be continuous areas of plateau or neighbouring slopes that had gradients of no more than 15° and that could reach the accumulation area of the glacier, either directly or through the glacier’s avalanche area (Section 3.6.4.2). Note the avalanche area is not included within the snow-blow area to avoid double-counting the avalanche source area. The area was then normalised to the glacier size as a ratio and converted to a factor by taking the square root (Sissons and Sutherland 1976; Sissons, 1980a; Mitchell, 1996; Coleman et al., 2009). The largest factors are for the corries of the eastern Cairngorms, such as Sgor Riabhach and Lochan nan Gabhar (Table 6.5 and Figure 6.12). The highest value in the central Cairngorms is the Coire na Saobhaidh (Glen Derry) which, although not a large absolute area, normalised to the glacier size it becomes a favourable snow-blow factor. As expected, the snow-blow factors of Coire an Sneachda and Coire an Lochain Uaine on the northern side of the Cairn Gorm plateau are larger than Lochain Uaine on the southern side of Ben Macdui. In the western Cairngorms, Coire Dhondail has a large snow-blow factor due to the extensive Mòine Mhòr plateau to the south. Coire an Lochain (Braeriach) and Coire Bhrochain are larger than the nearby candidate sites of Coire Bhogha-cloche and Coire
Ruadh. Note that Coire Ruadh would have a much larger snow-blown area if the entire Braeriach plateau was included; it has not been included here because of being separated by a slope greater than 15°. This highlights a weakness of this approach as it is likely that some snow from the larger plateau would reach the corrie. The Glen Eidart glacier snow-blown area factor is larger than the Glen Geusachan glacier which, in turn, is favourable to the Glen Dee glacier; however, they all have factors greater than the nearby Coire an t-Saighdeir, which may in part explain the higher ELA of this corrie glacier. The plateau-style reconstructions tend to have much smaller snow-blown area factors; this is due to much of the potential snow-blown area being incorporated within the glacier, and also the division of the snow-blown area by the larger plateau-glacier area reducing the ratio and factors.

### Table 6.5 Digitised snow-blown areas, ratios and factors for the south-west quadrant

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Number</th>
<th>*Style</th>
<th>ELA AABR</th>
<th>Glacier Area</th>
<th>Source Area</th>
<th>Ratio</th>
<th>Factor</th>
</tr>
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<tbody>
<tr>
<td><strong>Eastern</strong></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Slochd Mòr</td>
<td>1</td>
<td>v</td>
<td>871</td>
<td>0.68</td>
<td>3.02</td>
<td>4.47</td>
<td>2.11</td>
</tr>
<tr>
<td>Slochd Mòr Plateau</td>
<td>24</td>
<td>p</td>
<td>969</td>
<td>1.15</td>
<td>2.50</td>
<td>2.18</td>
<td>1.48</td>
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<tr>
<td>East Beinn a’ Bhuird</td>
<td>2</td>
<td>v</td>
<td>977</td>
<td>2.04</td>
<td>1.30</td>
<td>0.64</td>
<td>0.80</td>
</tr>
<tr>
<td>East Beinn a’ Bhuird Plateau</td>
<td>25</td>
<td>p</td>
<td>1,020</td>
<td>3.14</td>
<td>1.59</td>
<td>0.50</td>
<td>0.71</td>
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<tr>
<td>Coire na Ciche (Beinn a’ Bhuird)</td>
<td>3</td>
<td>v</td>
<td>983</td>
<td>0.11</td>
<td>0.61</td>
<td>5.64</td>
<td>2.37</td>
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<td>4</td>
<td>v</td>
<td>826</td>
<td>0.31</td>
<td>5.54</td>
<td>18.06</td>
<td>4.25</td>
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<tr>
<td>Sgor Riabhach</td>
<td>5</td>
<td>v</td>
<td>757</td>
<td>0.18</td>
<td>4.19</td>
<td>23.39</td>
<td>4.84</td>
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<tr>
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<tr>
<td>Lochan Uaine (Ben Macdui)</td>
<td>7</td>
<td>v</td>
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<td>0.31</td>
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<td>1.11</td>
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<tr>
<td>Coire an t-Sneachda</td>
<td>8</td>
<td>v</td>
<td>1,010</td>
<td>0.28</td>
<td>1.47</td>
<td>5.16</td>
<td>2.27</td>
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<tr>
<td>Coire an Lochain (Cair Gorm)</td>
<td>9</td>
<td>v</td>
<td>998</td>
<td>0.32</td>
<td>2.28</td>
<td>7.22</td>
<td>2.69</td>
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<tr>
<td>Coire na Sgobhraidh (Glen Derry)</td>
<td>10</td>
<td>v</td>
<td>734</td>
<td>0.09</td>
<td>0.71</td>
<td>7.71</td>
<td>2.78</td>
</tr>
<tr>
<td>Deny and Aven Valley</td>
<td>13</td>
<td>v</td>
<td>932</td>
<td>5.92</td>
<td>5.72</td>
<td>0.97</td>
<td>0.98</td>
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<tr>
<td>Deny and Aven Plateau</td>
<td>27</td>
<td>p</td>
<td>992</td>
<td>8.38</td>
<td>2.65</td>
<td>0.32</td>
<td>0.56</td>
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<tr>
<td><strong>Western</strong></td>
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<td></td>
<td></td>
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<tr>
<td>Coire an Lochain (Braeriach)</td>
<td>14</td>
<td>v</td>
<td>1,029</td>
<td>0.74</td>
<td>2.21</td>
<td>2.97</td>
<td>1.72</td>
</tr>
<tr>
<td>Coire an Lochain Plateau (Braeriach)</td>
<td>28</td>
<td>p</td>
<td>1,092</td>
<td>1.20</td>
<td>1.94</td>
<td>1.61</td>
<td>1.27</td>
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<tr>
<td>Coire Dhondail</td>
<td>16</td>
<td>v</td>
<td>891</td>
<td>0.18</td>
<td>1.51</td>
<td>8.17</td>
<td>2.86</td>
</tr>
<tr>
<td>Coire Bhrochain (Braeriach)</td>
<td>17</td>
<td>v</td>
<td>1,050</td>
<td>0.37</td>
<td>1.13</td>
<td>3.07</td>
<td>1.75</td>
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<tr>
<td>Coire Bhrochain Plateau (Braeriach)</td>
<td>30</td>
<td>p</td>
<td>1,106</td>
<td>0.56</td>
<td>1.06</td>
<td>1.91</td>
<td>1.38</td>
</tr>
<tr>
<td>Coire an t-Saighdeir (Glen Dee)</td>
<td>18</td>
<td>v</td>
<td>1,028</td>
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<td>0.31</td>
<td>0.85</td>
<td>0.92</td>
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<tr>
<td>Eidart</td>
<td>19</td>
<td>v</td>
<td>812</td>
<td>1.97</td>
<td>5.73</td>
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<td>1.71</td>
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<td>Glen Geusachan Valley</td>
<td>20</td>
<td>v</td>
<td>791</td>
<td>8.20</td>
<td>14.33</td>
<td>1.75</td>
<td>1.32</td>
</tr>
<tr>
<td>Glen Geusachan Plateau</td>
<td>31</td>
<td>p</td>
<td>831</td>
<td>10.41</td>
<td>13.12</td>
<td>1.26</td>
<td>1.12</td>
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<tr>
<td>Glen Dee Valley</td>
<td>21</td>
<td>v</td>
<td>869</td>
<td>2.78</td>
<td>1.98</td>
<td>0.71</td>
<td>0.84</td>
</tr>
<tr>
<td>Glen Dee Plateau</td>
<td>32</td>
<td>p</td>
<td>947</td>
<td>3.86</td>
<td>1.66</td>
<td>0.43</td>
<td>0.66</td>
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<tr>
<td><strong>Candidate Readvance Sites</strong></td>
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<tr>
<td>Cnap a’ Chèirich</td>
<td>6</td>
<td>v</td>
<td>1,009</td>
<td>0.10</td>
<td>1.47</td>
<td>14.93</td>
<td>3.86</td>
</tr>
<tr>
<td>Cnap a’ Chèirich Plateau</td>
<td>26</td>
<td>p</td>
<td>1,057</td>
<td>0.25</td>
<td>1.52</td>
<td>6.19</td>
<td>2.49</td>
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<tr>
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<td>15</td>
<td>v</td>
<td>936</td>
<td>0.24</td>
<td>0.59</td>
<td>2.44</td>
<td>1.56</td>
</tr>
<tr>
<td>Coire Bogha-clichte Plateau</td>
<td>29</td>
<td>p</td>
<td>1,078</td>
<td>0.65</td>
<td>1.27</td>
<td>1.94</td>
<td>1.39</td>
</tr>
<tr>
<td>Coire Ruadh</td>
<td>23</td>
<td>v</td>
<td>1,075</td>
<td>0.18</td>
<td>0.31</td>
<td>1.74</td>
<td>1.32</td>
</tr>
<tr>
<td>Coire Ruadh Plateau</td>
<td>33</td>
<td>p</td>
<td>1,139</td>
<td>0.31</td>
<td>0.25</td>
<td>0.82</td>
<td>0.91</td>
</tr>
<tr>
<td>Coire Garbhlaich</td>
<td>22</td>
<td>v</td>
<td>781</td>
<td>0.39</td>
<td>1.57</td>
<td>4.05</td>
<td>2.01</td>
</tr>
</tbody>
</table>

* Refers to glacier style v = valley-sourced and p = plateau-sourced.
6.3.2.4 Snow-redistribution modelling

A GIS-based snow-blown model has been used to redistribute snow within the Cairngorms; further methodology details can be found within Chapter 3 (Section 3.6.4.3). The model was run for wind directions from the south-east, south, south-west and west. These directions were chosen based on the inferred prevailing wind direction from nearby mountains being from the south-west (Sutherland, 1984; Purves et al., 1999; Benn and Ballantyne, 2005; Ballantyne, 2007a), a possible south-east influence in the South-East Grampians (Sissons and Sutherland, 1976; Sissons, 1979) and modelling of Younger Dryas climate (Isarin et al., 1997). Low, medium and high snow-transport scenarios were modelled for each direction. After redistribution, the ‘volume’ of snow lying within the glacier area and surrounding avalanche area was calculated. This was divided by the glacier area to generate a snow-blown ratio and the square root taken to derive a snow-blown factor, as preferred by some authors (e.g. Sissons, 1980b). Here the ratios are described and
used within the regression analysis; however, the factors show very similar relationships. This modelling was carried out for the unglaciated and glaciated terrain, including both valley- and plateau-sourced glaciation styles. Figure 6.13 shows the redistributed snow from the south-west medium snow-transport model. Note the higher snow values on the lee side of ridges where the snow has been deposited immediately once the wind transport threshold is no longer met. Note also that the glacier extraction areas include the avalanche area and thus include snow which is assumed to reach the glacier surface through avalanching. The highest snow values occur in the sheltered heads of corries and in valleys that are orientated perpendicular to the wind direction.

Snow-blow areas, ratios and factors for the plateau unglaciated (Table 6.6) and glaciated terrains (Table 6.7), and valley unglaciated (Table 6.8) and glaciated terrains (Table 6.9) are presented below. Some glaciers with unfavourable wind directions and speeds have values of zero; this represents the removal of snow from the combined glacier and avalanche area. The general patterns are consistent throughout the different scenarios; the plateau glaciers tend to have lower snow-blow ratios and factors than their valley counterparts due to a similar snow-blow area being normalised to a larger glacier area. The glaciated terrain scenarios tend to have lower snow-blow factors than the unglaciated terrain due to the infilling of favourable deposition locations by the glacier. The unglaciated medium transport threshold snow-blow ratios show particularly high values for Coire Dhondail from the south (Figure 6.14 and Figure 6.15), due to being on the northern side of the large Mòine Mhòr plateau. However, this is much reduced in the glaciated terrain, caused by the presence of the glacier reducing the sheltered area with low wind speeds (Figure 6.14 and Figure 6.15). The eastern corries of Coire na Ciche, Lochan na Gabhar and Sgor Riabhach have high snow-blow ratios especially from the south-west. In addition, Coire na Saobhaidh (Glen Derry), despite not being surrounded by a large plateau, has a large snow-blow ratio due to its small size. Coire an Lochain (Braeriach) has small snow-blow ratios particularly when glaciated. Low snow-blow ratios for corries such as Coire an Lochain (Braeriach) may be due to the overestimation of the glacier size. Small errors in corrie-glacier size may generate large differences in snow-blow ratio when the snow-blow area is divided by glacier size, thus creating sensitivity to any errors in glacier extent,
particularly of small corrie glaciers. The large glaciers such as Derry, Geusachan and Derry/Avon have relatively small snow-blow ratios; however, they are consistent from multiple directions (Figure 6.14 and Figure 6.15). The ratios show that these larger valleys are well situated to gain snow from multiple directions, and that they are not vulnerable to snow deflation from any of the modelled directions.

Using the different snow-blow directions and transport scenarios, it was possible to establish their relationships with the glaciers’ ELAs (Table 6.6, Table 6.7, Table 6.8 and Table 6.9). The $R^2$ and coefficients were calculated for the reconstructions based on geomorphological and dating evidence; the candidate sites were not included. The strongest $R^2$ values were consistently found for the south-west wind direction and middle transport scenario; the plots can be seen in Figure 6.16. They show an expected negative relationship between glacier ELA and south-west snow-blow ratio, with more variation being explained in the unglaciated terrain scenarios. The unexplained scatter within the plots may be due to the other topoclimatic factors or precipitation gradients, which are not taken into account at this stage.

This method has independently selected the south-west medium transport scenario as explaining the most variation in the glacier ELAs. A south-west wind direction has also been inferred from other palaeoglaciological studies elsewhere in the British Isles (Mitchell, 1996; Ballantyne, 2007a; Bendle and Glasser, 2012). Mean ratios were calculated to combine the low, medium and high transport scenarios for each wind direction and, in addition, multiple wind directions were combined; however, this did not increase the explained variation in ELA. Again the mean south-west snow-blow ratios explained the most variation in ELA, but it was no stronger than the medium transport scenario.
Figure 6.13 Redistributed snow from the south-west wind direction medium transport scenario

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Table 6.6 Plateau (unglaciated) scenario snow-blow areas, ratios and factors with coefficients and R² values for relationships with glacier ELA
Glacier ELA
ELA Av
Glacier
No. Style (km²) BR=1.9 BR=1.9
Slochd Mòr Plateau
24
p
1.15
969
970
East Beinn a’ Bhuird Plateau
25
p
3.14 1020
1026
Coire na Ciche (Beinn a’ Bhuird) 3
v
0.11
983
1014
Lochan nan Gabhar
4
v
0.31
826
860
Sgor Riabhach
5
v
0.18
757
781
Lochan Uaine (Ben Macdui)
7
v
0.25 1029
1052
Coire an t-Sneachda
8
v
0.28 1010
1026
Coire an Lochain (Cairn Gorm) 9
v
0.32
998
1019
Coire an Lochain Plateau (Braeriach)
28
p
1.20 1092
1093
Coire Dhondail
16
v
0.18
891
899
Coire Bhrochain Plateau (Braeriach)
30
p
0.56 1106
1109
Coire an t-Saighdeir (Glen Dee) 18
v
0.37 1028
1058
Coire na Saobhaidh (Glen Derry)10
v
0.09
734
760
Derry and Avon Plateau
27
p
8.38
992
992
Eidart
19
v
1.97
812
817
Glen Geusachan Plateau
31
p
10.41
831
843
Glen Dee Plateau
32
p
3.86
947
955

(km²)
3.89
10.69
1.04
2.00
0.20
2.45
1.13
1.17
1.01
0.69
1.41
0.74
0.00
14.09
9.31
28.29
19.22

South-East 135
Low
Medium
Ratio Factor (km²) Ratio Factor (km²)
3.39
1.84 3.21 2.80 1.67 2.26
3.40
1.84 3.82 1.22 1.10 3.56
9.61
3.10 0.39 3.57 1.89 0.21
6.52
2.55 1.48 4.82 2.19 1.45
1.12
1.06 0.23 1.30 1.14 0.25
9.74
3.12 0.45 1.81 1.34 0.42
3.99
2.00 1.20 4.21 2.05 1.15
3.70
1.92 1.24 3.92 1.98 0.91
0.84
0.92 1.02 0.85 0.92 1.32
3.72
1.93 1.41 7.65 2.77 0.88
2.53
1.59 0.49 0.89 0.94 0.68
2.03
1.43 0.78 2.13 1.46 0.82
0.00
0.00 0.41 4.50 2.12 0.11
1.68
1.30 10.40 1.24 1.11 10.21
4.73
2.17 5.74 2.91 1.71 3.38
2.72
1.65 16.51 1.59 1.26 12.43
4.98
2.23 7.94 2.06 1.43 5.69

High
Ratio Factor (km²)
1.97
1.40 6.59
1.13
1.06 7.88
1.91
1.38 1.09
4.73
2.17 1.31
1.41
1.19 1.83
1.67
1.29 0.53
4.05
2.01 4.21
2.86
1.69 1.70
1.10
1.05 1.19
4.79
2.19 3.65
1.23
1.11 0.59
2.25
1.50 1.48
1.23
1.11 0.58
1.22
1.10 17.18
1.71
1.31 13.08
1.19
1.09 39.49
1.48
1.22 13.12

South 180
Low
Medium
Ratio Factor (km²) Ratio Factor (km²)
5.74
2.40 3.95 3.44 1.86 3.15
2.51
1.58 6.90 2.20 1.48 4.32
10.01
3.16 0.92 8.45 2.91 0.39
4.26
2.06 0.94 3.06 1.75 0.93
10.21
3.20 0.76 4.25 2.06 0.34
2.10
1.45 0.76 3.02 1.74 0.69
14.80
3.85 2.06 7.23 2.69 1.24
5.38
2.32 1.89 5.96 2.44 1.46
0.99
1.00 1.51 1.26 1.12 1.67
19.81
4.45 2.98 16.16 4.02 0.78
1.05
1.03 0.70 1.26 1.12 0.73
4.04
2.01 1.42 3.88 1.97 1.05
6.34
2.52 0.53 5.75 2.40 0.50
2.05
1.43 15.08 1.80 1.34 12.12
6.64
2.58 7.66 3.89 1.97 4.21
3.80
1.95 17.04 1.64 1.28 15.60
3.40
1.84 8.05 2.09 1.45 6.31

South-West 225
High
Low
Medium
Ratio Factor (km²) Ratio Factor (km²) Ratio Factor (km²)
2.75 1.66 2.62 2.28 1.51 2.72 2.37
1.54 2.84
1.37 1.17 7.55 2.40 1.55 6.78 2.16
1.47 4.90
3.57 1.89 1.06 9.73 3.12 0.56 5.15
2.27 0.29
3.04 1.74 3.28 10.69 3.27 1.89 6.16
2.48 0.93
1.89 1.37 1.45 8.07 2.84 1.50 8.36
2.89 0.63
2.73 1.65 0.48 1.92 1.39 0.57 2.25
1.50 0.62
4.37 2.09 1.38 4.87 2.21 1.20 4.22
2.05 1.01
4.61 2.15 1.11 3.51 1.87 1.15 3.63
1.90 1.18
1.39 1.18 1.38 1.15 1.07 1.25 1.04
1.02 2.33
4.24 2.06 0.29 1.58 1.26 0.36 1.95
1.40 0.38
1.31 1.14 1.31 2.36 1.54 1.23 2.21
1.49 1.06
2.86 1.69 1.25 3.42 1.85 1.37 3.74
1.93 1.41
5.39 2.32 0.59 6.36 2.52 0.71 7.75
2.78 0.75
1.45 1.20 16.32 1.95 1.40 14.35 1.71
1.31 13.57
2.14 1.46 12.64 6.42 2.53 7.43 3.78
1.94 4.27
1.50 1.22 31.07 2.99 1.73 19.62 1.89
1.37 14.62
1.64 1.28 12.76 3.31 1.82 11.56 3.00
1.73 7.35

High
Ratio Factor
2.48
1.57
1.56
1.25
2.71
1.65
3.03
1.74
3.49
1.87
2.48
1.58
3.55
1.89
3.73
1.93
1.94
1.39
2.06
1.44
1.90
1.38
3.85
1.96
8.14
2.85
1.62
1.27
2.17
1.47
1.40
1.19
1.90
1.38

(km²)
5.61
6.94
0.34
0.62
0.20
0.82
0.63
1.02
0.98
0.28
2.13
0.26
0.00
10.02
13.04
19.32
7.42

West 270
Low
Medium
Ratio Factor (km²) Ratio Factor
4.89 2.21 2.70 2.36 1.54
2.21 1.49 7.06 2.25 1.50
3.17 1.78 0.67 6.17 2.48
2.04 1.43 0.51 1.67 1.29
1.14 1.07 1.34 7.46 2.73
3.26 1.81 0.93 3.69 1.92
2.21 1.49 0.55 1.95 1.40
3.22 1.79 1.00 3.15 1.78
0.82 0.90 1.88 1.56 1.25
1.54 1.24 0.29 1.58 1.26
3.83 1.96 2.05 3.68 1.92
0.72 0.85 0.49 1.33 1.15
0.00 0.00 0.34 3.65 1.91
1.20 1.09 12.54 1.50 1.22
6.62 2.57 9.07 4.61 2.15
1.86 1.36 19.96 1.92 1.39
1.92 1.39 7.89 2.04 1.43

(km²)
2.27
4.58
0.68
0.44
0.96
0.72
0.49
0.79
2.47
0.29
0.89
0.57
0.49
12.63
4.42
13.89
7.35

High
Ratio Factor
1.98
1.41
1.46
1.21
6.25
2.50
1.43
1.20
5.34
2.31
2.85
1.69
1.72
1.31
2.50
1.58
2.06
1.43
1.56
1.25
1.59
1.26
1.57
1.25
5.30
2.30
1.51
1.23
2.24
1.50
1.34
1.16
1.91
1.38

Relationship with ELA

Coefficient
R²

0.16
0.03

0.26
0.07

-0.37
0.13

-0.40
0.16

-0.12
0.01

-0.10
0.01

-0.33
0.11

-0.43
0.19

-0.21
0.04

-0.26
0.07

-0.27
0.07

-0.27
0.07

-0.57
0.33

-0.60
0.35

-0.68
0.47

-0.66
0.43

-0.44
0.20

-0.41
0.17

0.10
0.01

0.25
0.06

-0.35
0.12

-0.33
0.11

-0.38
0.15

-0.37
0.14

Relationship with ELA Av

Coefficient
R²

0.20
0.04

0.28
0.08

-0.34
0.12

-0.37
0.13

-0.08
0.01

-0.07
0.00

-0.32
0.10

-0.42
0.18

-0.19
0.04

-0.24
0.06

-0.22
0.05

-0.22
0.05

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0.26

-0.54
0.29

-0.63
0.39

-0.60
0.36

-0.40
0.16

-0.36
0.13

0.08
0.01

0.23
0.05

-0.32
0.10

-0.31
0.09

-0.34
0.12

-0.33
0.11

0.22 0.89
2.97 9.72
0.71 1.09
8.96 23.13

0.94
3.12
1.04
4.81

0.12 0.47
0.55 1.81
0.95 1.45
4.32 11.14

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3.34

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4.56

0.64
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1.57
3.62

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2.22
7.74

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1.98

2.24
0.97
0.66
1.41

2.41
1.18
0.73
1.81

1.55
1.09
0.86
1.35

Candidate Readvance Sites
Cnap a’ Chlèirich Plateau
Coire Ruadh Plateau
Coire Bogha-cloiche Plateau
Coire Garbhlach

26
33
29
22

p
p
p
c

0.25
0.31
0.65
0.39

1057
1139
1078
781

1057
1140
1080
804

0.10
0.47
0.89
1.77

0.51
0.74
1.45
3.00

0.19
0.52
0.94
2.09

0.16
2.53
0.71
1.38

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0.27
0.48
0.81
1.13

1.64
0.50
0.00
1.20

1.23
0.29
0.29
0.77

0.59
0.36
0.48
0.70

310


<p>| Table 6.7 Plateau (glaciated) scenario snow-blow areas, ratios and factors with coefficients and R² values for relationships with glacier ELA |</p>
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<th>ELA Av</th>
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## Table 6.8 Valley (unlagated) scenario snow-blow ratios and factors with coefficients and R² values for relationships with glacier ELA

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<td>Medium</td>
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### Chapter 6: Glacier Reconstruction, Topoclimatic Factors and Palaeoclimate
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<th>Coefficient</th>
<th>R²</th>
<th>Relationship with ELA</th>
<th>Coefficient</th>
<th>R²</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coire an Chleiren</td>
<td>0.10</td>
<td>-0.18</td>
<td>0.03</td>
<td>0.02</td>
<td>0.04</td>
</tr>
<tr>
<td>Coire Buchaille</td>
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<td>0.03</td>
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<tr>
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<td>-0.18</td>
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<td>Coire Ghearch</td>
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<td>0.03</td>
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<td>0.02</td>
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</table>

Candidate Reassurance Sites

<table>
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<tr>
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<th>Style</th>
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<th>BR=1.9</th>
<th>BR=1.9 (km²)</th>
<th>Ratio</th>
<th>Factor (km²)</th>
<th>Ratio</th>
<th>Factor (km²)</th>
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<td>1091</td>
<td>1015</td>
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<td>0.00</td>
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<td>Coire Bheag-theoch</td>
<td>15 v</td>
<td>Medium Low</td>
<td>0.24</td>
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<td>857</td>
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<td>0.18</td>
<td>1075</td>
<td>1195</td>
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</table>

Table 6.9 Valley (glaciated) scenario snow-blow ratios and factors with coefficients and relationships with glacier ELA
Figure 6.14 Plateau scenario snow-blown ratios for unglaciated (top) and glaciated (bottom) terrain
Figure 6.15 Valley scenario snow-blow ratios for unglaciated (top) and glaciated (bottom) terrain
Figure 6.16 ELA versus south-west medium transport snow-blow ratio. Plateau unglaciated (top left) and glaciated (top right), valley unglaciated (bottom left) and glaciated (bottom right)
6.3.3 Topoclimatic regression analysis

6.3.3.1 Introduction

Multiple regression analysis has been used to establish whether the variation in the dependent glacier ELAs can be explained by the independent topoclimatic factors; alternatively, one or more of the glacier ELAs may not be explained. In such cases, this would indicate that the glacier or glaciers may be from a separate climatic event, that there were factors present that have not been included in the regression analysis, that the topoclimatic factors have not been modelled sufficiently to represent the processes, or that there were errors in the glacier reconstructions. The approach taken was to generate the regression models using the reconstructed glaciers of probable Younger Dryas age, based on the geomorphological mapping, landsystems and dating described previously. The candidate glaciers have been omitted at the model-building stage, so as not to build the models using uncertain cases. However, their topoclimatic factors and ELAs have been compared with the multiple regression equations in Section 6.3.3.6. Numerous regression models have been generated using both the valley-sourced scenario and the plateau-scoured scenario, and the topoclimatic factors calculated from the unglaciated terrain (present-day DEM) and the glaciated terrain (DEM including glacier reconstructions). The Derry and Avon glaciers have been combined as one case for both the valley- and plateau-sourced scenarios to avoid uncertainties regarding the position of the ice divide and the influence of this on ELA and calculation of individual topoclimatic factors. The AABR (BR = 1.9) ELAs have been used for the regression analysis; given the relatively small variations in ELA caused by the different balance ratios applied (Table 6.1) the relationships are likely to be consistent regardless of the balance ratio used.

Multiple linear regression has some important assumptions including: linear relationships, independent errors, homogeneity of variance, normally distributed errors, no perfect multicollinearity and absence of outliers (Miles and Shevlin, 2001; Field, 2013). The analysis has been carried out in the statistical package SPSS, and these assumptions have been checked visually with scatter plots, histograms and P-P plots, or with statistical test scores such as variance inflation factors (VIF) < 10 for multicollinearity, Durbin-Watson close to 2 for independent errors and
Cook’s distance < 1 for influential cases (Field, 2013). Given the number of glaciers is 17, the number of possible predictors is limited to ensure over-fitting does not become an issue.

The diagnostic statistics used to evaluate which explanatory variables were included within the regression models are explained below. The standardised coefficients are useful to compare the relative importance of each explanatory variable within the model. The $R^2$ value reports the amount of variance in the glacier ELAs that is accounted for by the regression model, and the adjusted $R^2$ the amount of variance accounted for by the model if the whole population was used (Field, 2013). The latter penalises the inclusion of surplus variables that are not helpful in explaining the variation in ELA and is useful for comparing different models, particularly when sample size is low (Montgomery et al., 2012). The $F$-ratio and associated $p$-value indicate the models’ overall significance relative to using the mean ELA to predict the individual glaciers ELA. The coefficients indicate the relation between the individual explanatory variables and the ELA, and whether the relation is significant (less than 0.05).

### 6.3.3.2 Approach

All the topoclimatic factors described above including avalanche ratios (Table 6.4), radiation values (Figure 6.10 and Figure 6.11), manually digitised snow-blow ratios (Table 6.5) and snow-redistribution modelled snow-blow ratios (Table 6.6, Table 6.7, Table 6.8 and Table 6.9) were input into the regression analysis. The topoclimatic factors (explanatory variables) which explained most of the variation in the glacier ELA (independent variable) were selected for inclusion in the final models. The modelled snow-blow ratios were found to explain a greater proportion of the variation in glacier ELA than the manually digitised snow-blow areas. The south-west medium transport snow-blow ratio was the best snow-blow predictor for both the unglaciated and glaciated models. Note that the snow-blow factor was also used and gave similar results. The ablation area total radiation value for the partially cloudy conditions explained more of the variation in ELA than the clear-sky conditions for both the glaciated and unglaciated models. The clear-sky conditions explained slightly less variation in ELA; this is thought to be due to the partially cloudy conditions having an increased diffuse to total radiation ratio, such that the
deeper valleys received relatively less radiation. The repeated selection of these two topoclimatic factors gave some encouragement that these are genuine explanatory variables. An important question is whether the explanatory variables have scientific justification for their inclusion, and whether there is a causal relationship with glacier ELA. These variables both have good scientific reasons for their inclusion, with south-westerly snow-blow factors having been found to be important for supporting Younger Dryas glaciers throughout the British Isles (Mitchell, 1996; Sutherland, 1984; Ballantyne, 2007a; Bendle and Glasser, 2012) and ablation area radiation being important to glacier melt (Hock, 2005). The avalanche ratios were not found to have a significant relationship with the glacier ELAs, but may be helpful in explaining some individual glaciers.

6.3.3.3 Results

The results (Table 6.10 and Table 6.11) include the coefficients for the raw data and standardised data values. The adjusted R² values show that up to 63% of the variation in glacier ELA can be explained by the topoclimatic factors using the unglaciated terrain factors (Table 6.10), compared to up to 26% for the glaciated terrain factors (Table 6.11). The significance levels of the unglaciated models are significant and are superior to the glaciated terrain models. The unglaciated terrain models meet the assumptions of regression and have significant explanatory variables, whereas the glaciated terrain models have predictors that are not significant and low Durbin-Watson values, suggesting the residuals may not be independent. The adjusted R² and significance values are similar between the plateau- and valley-sourced scenarios in the unglaciated terrain, and superior for the plateau-sourced scenario in the glaciated terrain. The ELA residuals from the regression analysis were plotted spatially across the Cairngorm Mountains (Figure 6.17 and Figure 6.18). The spatial distribution revealed a negative tendency in the southern and western parts of the Cairngorms and positive trend further north. It is proposed that this spatial trend is due to a precipitation gradient from higher accumulation in the south-western corner of the Cairngorms decreasing further north. This is explored further in Section 6.3.3.4.
Table 6.10 Multiple regression models (unglaciated terrain)

<table>
<thead>
<tr>
<th>Model Scenario</th>
<th>Explanatory Variables</th>
<th>R²</th>
<th>Adjusted R²</th>
<th>Standard Error</th>
<th>Model F-Ratio</th>
<th>Model Sig.</th>
<th>Coefficients Unstandardised</th>
<th>95% Confidence Interval</th>
<th>Coefficients Standardised</th>
<th>Sig.</th>
<th>Partial Correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plateau-sourced</td>
<td>Constant SW Snow-Blow</td>
<td>0.68</td>
<td>0.63</td>
<td>68.96</td>
<td>14.65</td>
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Table 6.11 Multiple regression models (glaciated terrain)

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<th>Model Scenario</th>
<th>Explanatory Variables</th>
<th>R²</th>
<th>Adjusted R²</th>
<th>Standard Error</th>
<th>Model F-Ratio</th>
<th>Model Sig.</th>
<th>Coefficients Unstandardised</th>
<th>95% Confidence Interval</th>
<th>Coefficients Standardised</th>
<th>Sig.</th>
<th>Partial Correlation</th>
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<td>0.36</td>
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<td>97.39</td>
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<td>1434.2409</td>
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<td>SW Snow-Blow Radiation</td>
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Figure 6.17 ELA residuals from the plateau regression models (ELA bar chart residuals given in metres, scale is consistent with figures below)
Figure 6.18 ELA residuals from the valley regression models (ELA residuals given in metres)
6.3.3.4 Precipitation

Given the spatial distribution of ELA residuals in Figure 6.17 and Figure 6.18, it is hypothesised that this pattern results from a precipitation gradient across the Cairngorm Mountains. Although assuming a single linear precipitation gradient across the Cairngorm Mountains is simplistic, it may assist in understanding whether a precipitation gradient was accountable for the spatial trend. The perpendicular distance of each glacier to arbitrary lines was calculated using GIS (Figure 6.19).

![Figure 6.19 Method used for deriving distances which act as a proxy for precipitation gradients (dotted and solid lines represent the lines used for deriving the perpendicular precipitation distances)](image)

These distances represent precipitation gradients from the south-south-east, south, south-south-west, south-west, west-south-west and west. When included in the regression analysis, the precipitation gradient that generated the greatest increase in adjusted $R^2$ value was the south-to-north gradient. The inclusion of the south-to-north precipitation gradient improves the adjusted $R^2$ and significance values of all the models, with the three explanatory variables (south-west snow-blow ratio, partially cloudy radiation and south-to-north precipitation) explaining up to 80% of the variation in ELA in the unglaciated models and up to 61% in the glaciated models (Table 6.12 and Table 6.13). In addition, the explanatory variables and the models are all now significant and the models meet the assumptions of multiple
linear regression. The partial plots of the plateau- and valley-sourced unglaciated regression models can be seen in (Figure 6.20). The standardised coefficients indicate that the precipitation gradient is less important than the topoclimatic factors in the unglaciated terrain models and equal to the snow-blow ratio in the glaciated terrain models (Table 6.12 and Table 6.13). The spatial distribution of the ELA residuals with the inclusion of the precipitation gradient can be seen in Figure 6.21 and Figure 6.22. Note the overall reduction in residual size compared to Figure 6.17 and Figure 6.18; the outstanding glaciers with large positive or negative ELA residuals and some possible explanations are discussed in Section 6.3.5.
### Table 6.12 Multiple regression models including precipitation (unglaciated terrain)

<table>
<thead>
<tr>
<th>Model Scenario</th>
<th>Explanatory Variables</th>
<th>R²</th>
<th>Adjusted R²</th>
<th>Standard Error</th>
<th>Model F-Ratio</th>
<th>Model Sig.</th>
<th>Coefficients Unstandardised</th>
<th>95% Confidence Interval</th>
<th>Coefficients Standardised</th>
<th>Sig.</th>
<th>Partial Correlation</th>
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<td><strong>Plateau-sourced including precipitation</strong></td>
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<td>0.80</td>
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<td>-216.9423 278.7538</td>
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### Table 6.13 Multiple regression models including precipitation (glaciated terrain)

<table>
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<th>Model Scenario</th>
<th>Explanatory Variables</th>
<th>R²</th>
<th>Adjusted R²</th>
<th>Standard Error</th>
<th>Model F-Ratio</th>
<th>Model Sig.</th>
<th>Coefficients Unstandardised</th>
<th>95% Confidence Interval</th>
<th>Coefficients Standardised</th>
<th>Sig.</th>
<th>Partial Correlation</th>
</tr>
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<tbody>
<tr>
<td><strong>Plateau-sourced including precipitation</strong></td>
<td>Constant SW Snow-Blow Radiation S to N Precipitation</td>
<td>0.68</td>
<td>0.61</td>
<td>71.34</td>
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<td>126.6184 313.9877</td>
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<td>0.0001 0.0027</td>
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<td>0.007 0.546</td>
<td>0.003 0.708</td>
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<th>Model Scenario</th>
<th>Explanatory Variables</th>
<th>R²</th>
<th>Adjusted R²</th>
<th>Standard Error</th>
<th>Model F-Ratio</th>
<th>Model Sig.</th>
<th>Coefficients Unstandardised</th>
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<th>Coefficients Standardised</th>
<th>Sig.</th>
<th>Partial Correlation</th>
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<td><strong>Valley-sourced including precipitation</strong></td>
<td>Constant SW Snow-Blow Radiation S to N Precipitation</td>
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<td>0.0001 0.0027</td>
<td>0.0023 0.0083</td>
<td>0.007 0.546</td>
<td>0.003 0.708</td>
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</table>
Figure 6.20 Partial plots from the plateau-sourced and valley-sourced unglaciated models
Figure 6.21 ELA residuals from plateau regression models including a south-to-north precipitation gradient (ELA residuals given in metres)
Figure 6.22 ELA residuals from valley regression models including a south-to-north precipitation gradient (ELA residuals given in metres)
6.3.3.5 Discussion

The cause of the unglaciated terrain factors explaining a greater proportion of the variation in glacier ELA may be due to some important conditions for glacier inception and growth being reduced when the glacier’s maximum readvance position was reached. Lower $R^2$ values were apparent for the glaciated ratios when directly comparing the snow-blow ratios and glacier ELAs in Table 6.6, Table 6.7, Table 6.8 and Table 6.9. Some of the glaciers may have lost their favourable topoclimatic conditions when they reached their maximum readvance position. It seems probable that the glaciers readvanced making use of the topoclimatic factors until a state of equilibrium was reached. Therefore modelling using the glaciated terrain does not reveal the key topoclimatic factors that allowed the glacier to readvance.

The unstandardised regression coefficient for the south-to-north precipitation gradient is approximately 0.014 for the unglaciated models, equating to an increase in ELA of 14m per km from south to north. The glaciated models indicate a higher increase in ELA of c.21m per km. These gradients are comparable to those from other currently and formerly glaciated regions. ELAs rise across Ellesmere Island by 400m in less than 10km (40m per km) and 15m per km along the north-western margin of the Canadian archipelago (Miller et al., 1975). The Southern Alps, New Zealand, are recognised today for their steep ELA gradients (Rother and Shulmeister, 2006), with palaeo-reconstructed ELA gradients of between 19 and 23m per km (Porter, 1975). In Scandinavia the ELAs on a cross section from Memurubreene through Jostendalsbreen to Alfotbreen change by 900m in 150km (6m per km) (Ohmura et al., 1992). Whether such a gradient may have existed regionally or for a short distance across the topography of the Cairngorms is discussed further in Chapter 7: Section 7.2.2 and 7.2.4. A south-to-north increase in ELA was recognised by Sissons (1979a) as largely a continuation of that from the South-East Grampians and Gaick Plateau (Sissons and Sutherland, 1976; Sissons, 1979a, 1979c, Sissons, 1980a). A more north-easterly and easterly rise in ELA has been found closer to the west coast and across the Grampian Highlands (Sissons, 1979b, 1979c; Benn and Ballantyne, 2005; Finlayson, 2006; Gheorghiu et al., 2012) and in Snowdonia (Bendle and Glasser, 2012). Sissons suggested the pattern of glaciers within Britain could be explained by southerly snow-bearing winds, eastward moving fronts and winds from
the south-west redistributing snow (Sissons, 1979b, 1979c). This has been supported by work in Skye (Ballantyne, 1989) and Mull (Ballantyne, 2002). Further discussion of whether the local precipitation gradient within the Cairngorms fits with the regional trend can be found in Chapter 7: Section 7.2.4.

6.3.3.6 Climatic ELA

The regression equations can also be used to suggest a ‘climatic’ ELA, free from favourable topoclimatic factors. Using a snow-blow ratio of one, implying no snow-blow accumulation or deflation, a radiation value for the ablation area derived from the flat plateau (566074 WH/m²) and a location on the central Ben Macdui Summit (11,750m north of the arbitrary precipitation line) gives an ELA of c.1275m using the unglaciated terrain and c.1135m using the glaciated terrain. These values are above all but the highest summits and plateaus within the Cairngorms. This demonstrates the importance of the reduced radiation within the ablation zones of the corries and valleys, and also the additional accumulation from snow redistribution in allowing glacier formation below the climatic ELA. However, care must be taken because the radiation value is extrapolated beyond the higher range of the input values that were used to build the regression models.

The regression equations can also be used to infer whether the candidate glaciers were contemporaneous with the glaciers used to construct the models. Where the predicted ELAs are similar to the reconstructed ELAs, this would suggest the glaciers could have existed in the same climatic environment. Table 6.14 shows that the candidate glaciers all have predicted ELAs similar to or lower than their reconstructed ELAs, suggesting these glaciers may have existed under the same Younger Dryas climatic conditions. The Coire Ruadh candidate glacier has particularly favourable topoclimatic factors and a high reconstructed ELA. Further work is required to investigate whether the candidate glaciers in Table 6.14 were contemporaneous with the Younger Dryas glaciers or whether there are factors not accounted for that stopped the glaciers forming; this is discussed further in Chapter 7: Section 7.2.3.
In summary, the topoclimatic factors – specifically the south-west snow-blow ratio and the ablation area total radiation values – had negative and positive relations with the glacier ELAs respectively, and together explained up to c.63% of the variation in glacier ELA. The residuals from this regression analysis had a spatial pattern of negative values in the south-west and south, and higher positive values on the northern side of the Cairngorm Mountains. Up to 80% of the variation in glacier ELA could be explained when a south-to-north precipitation gradient was included within the regression models. The regression equations indicated a climatic ELA of c.1135m or c.1275m, and the candidate glaciers had predicted ELA values similar to or lower than their reconstructed ELAs, indicating they may have existed during the same climatic event.

### 6.3.4 Glaciation style and palaeoprecipitation values

It has not been possible to reliably discern the style of glaciation of all the glaciers during the Younger Dryas within the Cairngorm Mountains. The regression analysis of the valley- and plateau-style options yielded the same explanatory variables and very similar regression statistics. The adjusted $R^2$ values are 0.80 for both the valley- and plateau-style reconstructions when using the unglaciated terrain, and 0.50 and 0.61 respectively for the glaciated terrain. Thus it is not possible to distinguish the glaciation style from the analysis, and it is possible a mixture of plateau- and valley-sourced glaciers existed or that both existed at different times during the Younger Dryas Stade. However, the impact on the climatic inferences that can be made may be insubstantial. The modelled ‘climatic’ ELAs from both the valley- and plateau-sourced scenarios generated similar climatic ELAs at c.1275m for the unglaciated terrain and c.1135m for the glaciated terrain. Using the same methods as in Section

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Plateau-sourced</th>
<th>Reconstructed ELA (m)</th>
<th>Predicted ELA (m)</th>
<th>Difference (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Cnap a' Chléirich Plateau</td>
<td>1057</td>
<td>1096</td>
<td>39</td>
</tr>
<tr>
<td></td>
<td>Coire Bogha-cloiche Plateau</td>
<td>1078</td>
<td>994</td>
<td>-84</td>
</tr>
<tr>
<td></td>
<td>Coire Ruadh Plateau</td>
<td>1139</td>
<td>899</td>
<td>-240</td>
</tr>
<tr>
<td></td>
<td>Garbhlaich</td>
<td>781</td>
<td>749</td>
<td>-32</td>
</tr>
<tr>
<td>Valley-sourced</td>
<td>Cnap a' Chléirich</td>
<td>1009</td>
<td>1052</td>
<td>43</td>
</tr>
<tr>
<td></td>
<td>Coire Bogha-cloiche</td>
<td>936</td>
<td>922</td>
<td>-14</td>
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<tr>
<td></td>
<td>Coire Ruadh</td>
<td>1075</td>
<td>779</td>
<td>-296</td>
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<tr>
<td></td>
<td>Garbhlaich</td>
<td>781</td>
<td>710</td>
<td>-71</td>
</tr>
</tbody>
</table>
6.3 and described in Chapter 3: Section 3.5, the ELAs would require precipitation values of 590mm (0–1240mm) and 860mm (240–1520mm) per annum respectively based on the Ohmura et al. (1992) equation. These values are much lower than the precipitation values of up to 1700mm per annum required for the lower glacier ELAs (Table 6.2). However, as shown by the regression analysis, the snow-blow accumulation and reduced radiation can explain the lower glacier ELAs. The climatic ELAs of c.1275m and c.1135m would require just 165mm (0–560mm) and 400mm (50–810mm) of precipitation respectively based on the neutral seasonal equation from Golledge et al. (2010). Some caution must be applied when using the climatic ELA and the associated derived precipitation values because of extrapolation beyond the radiation values used to calculate the regression equations, and some uncertainty as to whether the original dataset of 70 glaciers used in the Ohmura et al. (1992) equation can represent such a climatic ELA. Indeed, Ohmura et al. (1992) explained that some of the scatter in the temperature-precipitation relationship can most likely be explained by variations in global radiation and longwave net radiation.

6.3.5 Discussion of Younger Dryas glaciers, topoclimatic factors and palaeoclimate

The reconstruction of the Younger Dryas glaciers and their ELAs within the Cairngorm Mountains showed a variation of valley-sourced ELAs between 734m (727–745m) and 1050m (1041–1062m) and plateau-style ELAs of between 734m (727–745m) and 1106m (1091–1131m). Modern-day glacier ELAs are known to vary within close proximity based on avalanche accumulation e.g. Himalayas (Benn and Lehmkuhl, 2000), snow redistribution e.g. Norway (Aa, 1996) and topographic shading and aspect e.g. Pyrenees, Spain (Chueca and Julián, 2004). Some important qualitative topoclimatic differences between the glaciation sites had been identified from fieldwork and observations from DEMs. The quantitative topoclimatic factors encouragingly revealed similar patterns to those that had been identified qualitatively. The strength of the quantitative approach was demonstrated when the factors were combined using multiple regression analysis and explained up to 63% of the variability in ELA, and 80% when a precipitation gradient was included. The partial plots and residuals showed that there were no significant outliers from the regression
models, suggesting the reconstructed glaciers could have all been from a contemporaneous climatic event, most likely the Younger Dryas Stade.

The calculation of a climatic ELA showed that all of the reconstructed glaciers benefitted from favourable topoclimatic conditions, either from snow redistribution or reduced radiation. This is to be expected as glaciers, particularly in more marginal areas, will utilise favourable topoclimatic locations. The results highlight the importance of the concept of variable glacier ELAs (Dahl et al., 1997; Nesje and Dahl, 2000). The Cairngorm landscape of selective linear erosion – consisting of large plateaus, deep valleys and perched corries – may accentuate the variability in glacier ELA compared to gentler terrain with less extensive plateaus.

Although the inclusion of the precipitation gradient greatly reduces the residuals, there are a few corrie glaciers, such as Coire na Ciche and Sgor Riabhach, where the topoclimatic factors and precipitation gradient cannot explain the ELA (Figure 6.21 and Figure 6.22). This may be due to topoclimatic factors that have not been modelled or included, such as the avalanche factors, or generalisations within the models that are important at a local scale, such as wind deflection. It may also reflect the errors associated with glacier reconstruction and ELA calculation. Questions have also arisen at the candidate glacier sites where topoclimatic factors suggest they could have supported glaciers. This is an area for additional work.

6.4 Chapter Summary

The mapping and dating evidence has been used to establish readvance landsystems of Younger Dryas age. The geomorphological evidence, assisted by flow-line modelling, has been used to reconstruct the former glacier geometry, both as valley- and plateau-sourced glaciers, where geomorphological evidence left the style of glaciation uncertain. The calculation of the glacier ELAs revealed local and regional variability, but this was partially explained by regression analysis including snow-blown ratios, radiation values and precipitation gradients. This indicates the glaciers could have existed simultaneously during the Younger Dryas, despite the variation in ELA, particularly when a precipitation gradient was included. It is thought all the glaciers benefitted from favourable topoclimatic conditions, with the climatic ELA being similar in altitude to the highest plateaus within the Cairngorms.
Chapter 7: Discussion

This chapter reports the findings of the study with respect to the objectives set out in Chapter 1 (included below) and discusses them in relation to our wider understanding of glaciology and palaeoclimate.

Research objectives

1. Determine the pattern, style and dynamics of glacier retreat within the Cairngorm Mountains during the retreat of the British–Irish Ice Sheet; including how locally and regionally sourced ice masses interacted.

2. Determine the extent, style (plateau- versus valley-sourced) and retreat dynamics of glaciation during the Younger Dryas readvance (12.9–11.6 ka BP). Reconstruct the glacier equilibrium-line altitudes (ELAs) and determine the topoclimatic factors that were important in supporting glacier formation.

3. Draw inferences on the palaeoclimatic conditions prior to the Lateglacial Interstadial and reconstruct the Younger Dryas palaeoclimatic conditions using innovative topoclimatic analysis; particularly focusing on Younger Dryas precipitation gradients.

The detailed analysis presented within Chapters 4, 5 and 6 progresses within each chapter to increase our understanding of the overall deglaciation of the Cairngorm Mountains. A synthesis of the analysis presented in Chapters 4, 5 and 6 is provided at the start of Section 7.1 for the Dimlington Stade deglaciation and Section 7.2 for the Younger Dryas readvance. These findings are then compared with modern analogues, previous studies and modelling studies, both within the Cairngorm Mountains and the wider British Isles. This enables further palaeoclimatic implications to be discussed again, both within the Cairngorm Mountains and regionally. Later in the chapter, the transferable findings of the research and the methods utilised are discussed, in addition to discussion of areas for future research.
Chapter 7: Discussion

7.1 Dimlington Stade and Transition into Lateglacial Interstade

7.1.1 Pattern of deglaciation

The mapping presented within Chapter 4 provided a relative chronology of deglaciation within individual valleys or areas. Within Chapter 5 and 6 the Younger Dryas glacier limits have been established through the identification of landsystems, dating evidence and modelling. This permits the pattern of ice-sheet deglaciation outside these readvance limits, combined with existing dating evidence, to be explored further. These findings, outlined below, show the progress towards objective 1 and 3 in Chapter 1, including addressing the style, dynamics and pattern of retreat; how local and external ice masses interacted; and the implications for palaeoclimate.

The pattern of deglaciation of the northern margin is well established (Brazier et al., 1996b), with the east-to-west deglaciation of both Cairngorm and Spey ice marked by the system of ice-dammed lakes and meltwater channels. However, a detailed sequence of deglaciation for the southern Cairngorms has not been presented before. There is strong evidence for a similar east-to-west deglaciation of the southern Cairngorms marked by numerous ice-dammed lakes set out in Section 7.1.2. Integrating the models of retreat for the northern and southern Cairngorms is more complex; this discussed in Section 7.1.3. The relations between landforms described in Chapter 4 have been used to infer retreat patterns (Figure 7.1). The geomorphological detail has been omitted here for clarity; please refer back to Chapter 4 for additional details.
Chapter 7: Discussion

Figure 7.1 Summary of glacier margins and existing chronological ages associated with ice-sheet deglaciation (recalculated using NWH 11.6, Lm scaling scheme and 1 mm/ka). Ages recalculated from Everest and Kubik (2006) and Ballantyne et al., (2009a)
Chapter 7: Discussion

7.1.2 Model of deglaciation for the southern Cairngorm Mountains

The ice within Deeside has formed meltwater channels high on the southern flanks of the Cairngorms; in the south-east section of the Cairngorms these indicate ice flow within the Dee to the east. There was also ice flow from Glen Derry into Glen Quoich, and from Glen Dee into Glen Luibeg, suggesting a west-to-east overflow of ice between neighbouring valleys. Early in deglaciation, Deeside ice overtopped the southern flanks of the south-east Cairngorms flowing north towards Loch Builg. This ice may have dammed a lake within lower Glen Avon as it retreated (Figure 7.2: Stage A). As the Deeside ice lowered, this ice supply ceased and ice remained within Deeside, lowering as shown by meltwater channels near the Carn Liath and Culardoch mountains. The Cairngorm ice from Beinn a’ Bhuidh would once have been confluent with external Deeside ice, and can also be seen to retreat from the Glen Quoich-Gleann an t-Slugain intersect. Here an ice-dammed lake formed in lower Glen Quoich at c.600m, blocked by ice flowing through the Clais Fhearnaig and possibly directly from Deeside (Figure 7.2: Stage B). At this time the locally sourced Beinn a’ Bhuidh glacier terminated within the ice-dammed lake. A simultaneous c.675m lake occurred in neighbouring Glen Derry as marked by lake shorelines and a shoreline-channel-delta system at Poll Bhàt (Figure 7.2: Stage B (1)). At this stage the c.600m Quoich ice-dammed lake overflowed east via Gleann an t-Slugain (Figure 7.2: Stage B (a)). Later, as the damming ice and consequent lake levels lowered, the Derry ice-dammed lake is thought to have drained via the channels on the northern side of Clais Fhearnaig, and the Quoich ice-dammed lake drained around the edge of the damming ice (Figure 7.2: Stage B (2 and b)). The damming ice is thought to have been sourced from the west and possibly included ice from Deeside. However, the final retreat of ice from Glen Lui seems to have been sourced from further west within the Cairngorms, most likely via the Meirleach col, not directly from Deeside. Later the same damming ice is likely to have formed the large moraine and ice-dammed lake at the exit of Glen Luibeg, and at a similar time the glacier is likely to have formed the large moraine near Allt Garbh in lower Glen Dee (Figure 7.2: Stage C). Later, as the ice lowered over the Meirleach col, stagnant ice and meltwater formed the landforms near Luibeg Bridge (Figure 7.2: Stage D). Once the ice had retreated sufficiently, a readvance of the locally sourced
Derry glacier out to Derry Lodge (Figure 7.2: Stage C) and later to Derry Dam (Figure 7.2: Stage E) occurred, but the exact timing is unclear.

After the Meirleach col became ice free, an ice-dammed lake formed within lower Glen Dee, and drained across the Meirleach col (c.575m) east towards Luibeg Bridge, possibly draining around the previously mentioned stagnant ice (Figure 7.2: Stage D). This lake was dammed by ice to the south of Glen Dee, which created shallow moraines in lower Glen Dee. The source of this ice is unknown but, given the lake drained east into valleys that link with Deeside, ice must have come from the south or west. The channels to the south of the Mòine Mhòr plateau indicate ice flow west around the south-west corner of the Cairngorms into Glen Feshie; thus the last stages of ice flow are most likely to have been from the south, and thus the damming ice may too have been from the south. Further work on the area to the south and west of the Cairngorms may provide clarity on its source. The positions of the locally sourced Dee and Geusachan glaciers at the time of lake formation are unclear; however, after lake drainage, ice is thought to have readvanced to create a distinct northern limit to the lake shoreline in lower Glen Dee (Figure 7.2: Stage E). It is likely that at this time the Geusachan glacier dammed a lake within upper Glen Dee in which the Glen Dee glacier terminated; note that to create a lake-terminating glacier within upper Glen Dee, the Geusachan glacier must have been above the height of the Meirleach col (Figure 7.2: Stage E). The Geusachan glacier later retreated to allow meltwater flow around its margin within Glen Dee as seen in Figure 7.2 Stage F.

These final stages may have been in the form of a western ice cap similar to that presented by Everest and Kubik (2006), with ice flow into Glen Geusachan, Gleann Einich and upper Glen Dee. The Lairig Ghru and Einich outlets of this ice cap were contemporaneous with the presence of the Glenmore ice lobe, as marked by the ice-dammed lakes (Brazier et al., 1998; Golledge, 2002; Everest and Kubik, 2006). If such a contemporaneous set of margins existed, this western ice cap provides a key link between the externally sourced Glenmore lobe and the local glaciers of both the northern and southern Cairngorms. As dictated by meltwater and moraine evidence, first the Lairig Ghru ice-dammed lake was formed (Figure 7.2: Stage E), followed by the Glen Einich ice-dammed lake (Figure 7.2: Stage F). The retreat of the Einich glacier is marked by numerous moraines lower in the valley but relatively
few nearer the valley head, suggesting rapid retreat or low debris transport during the last phase of deglaciation, whereas Glen Geusachan and the upper part of Glen Dee contain evidence for a later readvance and thus it is not possible to make further ice-sheet deglaciation inferences. However, the evidence on the Mòine Mhòr plateau indicates the retreat of ice was towards Allt Luineag and upper Glen Geusachan.

The relative timing of the final phase of deglaciation of the central Ben Macdui plateau is assumed to have been at a similar time, based on its close proximity. The large Ben Macdui plateau is thought to have supported plateau ice which contributed to the locally sourced Derry, Avon, Dee and Lairig Ghru glacier systems. It is possible the moraines in lower Glen Derry and near the Fords of Avon formed during similar climatic events to those in the Lairig Ghru, Gleann Einich and Glen Dee (Figure 7.2: Stages C–F). The Avon glacier was the last to retreat from the Fords of Avon, most likely as it is closer to its large source area which included much of the Ben Macdui and Cairn Gorm plateaus.

7.1.3 Linking the northern and southern models of deglaciation and outstanding issues

Difficulties do exist with the model of deglaciation. The northern and southern deglaciation models both suggest an east-to-west deglaciation. However, linking the models into one model is challenging due to them being, for the most part, topographically separate. The linking of the northern and southern models of deglaciation in the tentative schematic reconstruction diagrams is based on the key links described within this section.

The best links may be found through glacial breaches, such as the Lairig Ghru, and around the edges of the Cairngorm Mountains. The Lairig Ghru glacial breach does not have its own large source area, instead it would have benefitted from ice sourced from the plateaus of Ben Macdui and Braeriach (Figure 7.2), and potentially at times from ice overflow from upper Glen Dee. Thus it would be expected that the Lairig Ghru moraines formed when there was still considerable ice on the plateaus, and within the corries and valleys systems. The angle of the highest moraines on the east side of upper Glen Dee suggests ice flow from the
Lairig Ghru col to the north (Chapter 4: Section 4.4.3.3). This indicates that ice from the Lairig Ghru col flowed both north into the Lairig Ghru ice-dammed lake dammed by the Glenmore ice lobe and south to join the ice from the upper Glen Dee corries (Figure 7.2: Stage E).

The central and northern Cairngorms can be linked through the Strath Nethy glacial breach from Glen Avon to the Spey valley. The mapping suggests ice within the Spey valley, sourced from the west, entered lower Strath Nethy. This indicates the locally sourced Glen Avon glacier had retreated relatively early as it was no longer supplying ice into the lower part of Strath Nethy; at most, the large Ben Macdui plateau was only feeding ice into Glen Avon. This illustrates the contrast between the Spey ice lobe which was still of substantial size, when one of the highest plateau areas in Scotland only supported a limited outlet glacier at most (Figure 7.2: Stages B–F).

A further link is via Glen Feshie around the western side of the Cairngorm Mountains. The simultaneous presence of Feshie and Einich ice when the Gleann Einich ice-dammed lake was formed (Brazier et al., 1998) indicates substantial ice thicknesses to the west and most probably south of the Cairngorms at this time (Figure 7.2: Stage F). The direction of ice flow near the Feshie/Geldie watershed during the final phase of deglaciation is west and then north around the western side of the Cairngorms (Figure 7.2: Stage C–F). This is different from the general east-to-west retreat of ice identified within the southern Cairngorms. This suggests that while the overall pattern along the southern margin of the Cairngorms was from east to west, the deglaciation of the south-western corner near the Feshie/Geldie watershed was more complex. It is not thought such ice was sourced from within the Cairngorms, given the evidence of ice-dammed lakes along the southern margin. Thus it most likely marks the retreat of ice from the south, although further work is needed here, particularly in understanding the upper Tay and Gaick areas. Similarly, further work on the relative and absolute timing of the deglaciation of the north-east Cairngorms, particularly lower Slochd Mòr and the potential ice-dammed lake in Glen Avon, could provide information on the early deglaciation of the eastern Cairngorms.
The geometry of ice around the Cairngorms must have been complex. The Spey ice is thought to have been predominantly sourced from the west. Some ice would have contributed to this around the western side of the Cairngorms from Glen Feshie. The southern ice would seemingly have deglaciated from east to west along the southern side of the eastern and central Cairngorms, and then in the opposing direction around the south-west corner. While the northern Cairngorms have a relatively abrupt topographic transition with the Spey valley, the southern Cairngorms have a more gradual transition from Deeside to lower hills and then to the higher mountain flanks of the plateau. Thus it is more complex to understand the dominance and maximum position of external ice along and through the valleys of the southern Cairngorms. This is partially shown by previous mapping of erratics found near the Meirleach col and in Glen Luibeg (Sugden, 1970). However, given the granite boulders on moraines near the Glen Quoich-Glean an t-Slugain junction, Derry Lodge, Derry Dam, and the Allt Garbh moraine in lower Glen Dee, the locally sourced Cairngorm ice formed many of the final moraine positions within the lower southern valleys (Figure 7.2: Stage B−F).
Note: in Stage B the ice-dammed lake in Glen Derry first overflowed into Glen Quoich via Poll Bhàt (1), while the Glen Quoich ice-dammed lake overflowed via Gleann an t-Slugain (a). After the damming ice lowered, the ice-dammed lake in Glen Derry overflowed near Clais Fhearnaig (2) and the Quoich ice-dammed lake drained around the edge of the damming ice at (b).
Chapter 7: Discussion

Figure 7.2 Tentative schematic reconstruction of phases of ice-sheet deglaciation

Note: it is hoped these tentative deglaciation phases will provide a basis for discussion and further work. There is strong evidence for the east-to-west pattern of retreat and sequence of ice-dammed lakes along the southern margin of the Cairngorms. Note: some uncertainties have been introduced by matching this pattern with the established retreat of the northern Cairngorm margin – it is hoped this will provide a basis for future work.
7.1.4 Palaeoclimatic inferences

Several important palaeoclimatic implications have been established from the deglaciation sequence within the Cairngorms. There is evidence of predominantly active retreat with only limited areas of ice stagnation; these are located where the source area of the glacier concerned has been separated by the local topography. On both the northern and southern margin the retreat has been punctuated by stillstands and readvances; these are likely to represent the response to climatic fluctuations, although internal factors – such as changes in the type of glacier margin from lake to terrestrial – may have also been significant. Importantly, the geomorphology and derived deglaciation sequence indicate the formation of locally sourced glaciers during ice-sheet deglaciation prior to the Lateglacial Interstadial. These have interacted through numerous ice-dammed lakes on both the southern and northern sides of the Cairngorms, and suggest multiple periods of stillstand and readvance. The exact relative and absolute timing of the locally sourced glacier retreat and survival of ice within high corries and on plateaus is complex to establish; however, a tentative model of the main phases of deglaciation has been provided above to stimulate future discussion and further work.

The externally sourced ice impinging on the lower valleys of the Cairngorms illustrates the relative size of the ice sheet. Given the high topography of the Cairngorms, the relatively small locally sourced valley and outlet glaciers can only be explained by differences in precipitation. This suggests that, as at present (Met Office) and inferred during the Younger Dryas (Sissons, 1979b, 1979c; Benn and Ballantyne, 2005), there was a strong precipitation gradient across Scotland during ice-sheet deglaciation. The best estimates for the timing of the deglaciation of the Cairngorms come from surface exposure ages from moraines associated with the Gleann Einich glacier (16.3±1.3 ka to 14.6±1.1 ka), the simultaneous Glen More Lobe position (16.0±1.5 ka to 14.2±1.2 ka) (NWH 11.6, Lm and 1 mm/ka recalibrated ages from Everest and Kubik, 2006) and RSF deposits (reinterpreted as candidate moraines by Jarman et al., 2013) within the Lairig Ghru (17.7±1.6 ka to 15.0±1.0 ka) and Strath Nethy (20.0±1.3 ka to 16.7±1.3 ka - upper age has suffered from inheritance) (NWH 11.6, Lm and 1 mm/ka recalibrated ages from Ballantyne et al., 2009a). The aforementioned moraine ages are younger than the RSF/moraine ages, but overall the ages indicate deglaciation of the lower valleys by
15–17 ka BP and this is likely to be closer to 17 ka BP given that exposure ages provide minimum ages. This is before the rapid climate warming from c.6°C to c.12°C at the onset of the Late Glacial Interstade (c.14.6 ka BP) as recorded at Whitrig Bog, south-east Scotland (Brooks and Birks, 2000), which matches with the NGRIP record (Rasmussen et al., 2006).

7.1.5 Wider comparisons and implications

The record of both the regional retreat of ice and local glaciers is particularly useful: this section discusses how the Cairngorm deglaciation fits into the wider context of the retreat of the British–Irish Ice Sheet. Summary time-slice maps within Clark et al. (2012) show the north-east of Scotland to be one of the first locations for ice to retreat onshore at 19 ka BP; this compares to the later separation of the British and Irish ice sheets at c.16 ka BP, this is likely to be due to precipitation starvation of the eastern side of the ice sheet (Sutherland, 1984). The early deglaciation of the north-east coast near Buchan allowed the Moray Firth and Strathmore ice streams to come onshore after 18 ka BP (Peacock and Merritt, 2000; Merritt et al., 2003; Clark et al., 2012). Surface exposure dating of hills near Buchan (215–527m a.s.l.) yielded ages c.20 ka; however, lower altitude sites near Caithness and Orkney yielded younger ages indicating deglaciation between 16.0–14.8 ka (see Phillips et al., 2008; note these ages were calculated using the worldwide calibration dataset). Unfortunately no lower sites were dated near Buchan, but a similar pattern of higher summits being deglaciated early while ice persisted on the lower ground may have existed (Phillips et al., 2008). This would imply retreat of ice from the coast to the Cairngorms, where RSF/moraine ages suggest some valleys were ice free by c.16 ka BP (using default global calibration) (Ballantyne et al., 2009a) or proportionally earlier using the locally derived production rates. Dating of the mapped ice margins between Aberdeen and Ballater (Brown, 1993) or the Deeside roches moutonées described by Glasser (2002) would assist in understanding the timing and rate of deglaciation in the Dee valley. There is also a requirement to better understand how stillstand and readvance margins within the Cairngorms relate to other margins within east Scotland, such as the Perth Readvance (See McCabe et al., 2007), Ardersier, near Inverness (Gordon and Merritt, 1993) and glacier margins in the Moray Firth (Graham et al., 2009). How the Cairngorm deglaciation events compare with readvances elsewhere in the British Isles, such as the Killard Point Stade
Readvance in Ireland (McCabe et al., 1998), the Wester Ross Readvance (WRR) in north-west Scotland (Everest et al., 2006; Ballantyne et al., 2009b; Ballantyne and Stone, 2012) also needs further investigation. Despite such margins having been discussed in relation to the Heinrich 1 event (McCabe et al., 1998; Everest and Kubik, 2006) and also increased precipitation due to enhanced cyclonic activity as the atmospheric polar front moved northwards across the British Isles prior to the Lateglacial Interstadte (Sissons, 1981; Ballantyne and Stone, 2012), establishing a concrete link is difficult.

The geomorphological and dating evidence from the Cairngorms can be compared with numerical ice-sheet modelling outputs by Hubbard et al. (2009). Despite the detailed resolution of the modelling, Hubbard et al. (2009) highlight the broad methodological assumptions applied. The description within Hubbard et al. (2009) suggested that by 15.7 ka BP the ice sheet had reduced to ice caps over the Cairngorms and Western Highlands; however, the time-slice figures for the optimal model show extensive ice on the lower ground in both west and east Scotland at 15.7 and 15.6 ka BP. The description is consistent with the geomorphological evidence for both the local retreat of Cairngorm glaciers and the retreat of regional ice sourced from the west and possibly south of the Cairngorms. The RSF/moraine ages by Ballantyne et al. (2009a) indicate the glacial breaches within the Cairngorms were deglaciated by c.17 ka, and thus the ice extent is likely to have been reduced compared to the modelled time-slices for the period 17–15.6 ka BP. These differences between the model time-slices and terrestrial evidence are likely to be due to local reductions in precipitation. The model uses modern precipitation values that are linearly reduced in cold climates. However, there is likely to have been time-variable spatial variations in precipitation values that caused the Cairngorms to become deglaciated relatively early. The precipitation within the model reacts to the changes in geometry; however, Hubbard et al. (2009) recognise the method is simplistic and the climate forcing is not locally accurate. These local effects are likely to be balanced out when ice is transferred 100 km (Hubbard et al., 2009); however, it may be of greater importance during the build-up and retreat of ice masses.
7.2 Younger Dryas Glaciation

7.2.1 Synthesis

The geomorphological evidence presented within Chapter 4 found evidence to suggest a later period of glaciation within some of the corries and upper valleys. The moraines within Coire an Lochain have been attributed to the Younger Dryas Stade with the use of surface exposure dating (Kirkbride et al., 2014). The analysis of six new cosmogenic ages within Glen Geusachan and Glen Derry indicates the upper valleys may also have experienced limited glaciation at this time (Chapter 5). The combination of geomorphological and dating evidence, with the morphostratigraphy and landsystem approach, allowed the extrapolation of the dated landforms to sites of similar characteristic geomorphology (Chapter 6). The reconstruction of the glaciers showed some variations in glacier ELA (734–1050m or 734–1106m valley- and plateau-style respectively); qualitative observations suggested the high ELAs of some corrie glaciers may have been due to them extending out of the corries’ favourable topoclimatic zone; whereas selected valleys had a more extensive zone of favourable topoclimatic conditions. Quantitative analysis showed up to 80% of the ELA variation was explained with the use of modelled topoclimatic factors and inferred precipitation gradients, indicating the glaciers could have been from a contemporaneous event (Chapter 6). A contemporaneous readvance during the Younger Dryas is favoured for the reconstructed glaciers based on the geomorphological, geochronological and topoclimatic evidence summarised above. However, other options should not be eliminated from future work and discussion. In an attempt to establish a climatic ELA from the regression analysis, a neutral snow-redistribution scenario, plateau radiation value and central location were used to generate climatic ELAs of c.1135m and c.1275m. Derived annual precipitation values for the glaciers ranged from 1030 mm to 1730 mm for valley-style or 920 mm to 1730 mm for plateau-style reconstructions using the Ohmura et al. (1992) relationship. The use of the Scottish- and Younger Dryas-specific relationship by Golledge et al. (2010) reduces these precipitation values to between 550 mm and 1190 mm for valley-style or 450 mm to 1190 mm for plateau-style reconstructions. Given the favourable topoclimatic
conditions experienced by these glaciers, the precipitation values are likely to have been reduced (see Chapter 6: Section 6.3.4).

This analysis has made a valuable contribution towards the second and third research objectives in Chapter 1. The analysis has shown that larger valley glaciers may have existed during the Younger Dryas, benefitting from topoclimatic conditions and higher precipitation values in the southern and western Cairngorms. Little conclusive evidence was found to discern the style of glaciation; however, both valley- and plateau-sourced glaciers have been considered and this shortcoming does not alter the palaeoprecipitation inferences, given the detailed attention to topoclimatic factors (Chapter 6). No valley glaciers were reconstructed draining the easternmost plateaus of the Cairngorms near Ben Avon, but cold-based plateau ice cannot be eliminated. This is most likely due to the narrower and more rounded domed-shaped topography being less favourable for snow accumulation and also reduced precipitation further east.

7.2.2 Modern analogues and comparisons with modern glaciated environments

It is useful to compare the reconstructed glaciers within the Cairngorms with modern analogues. This can aid our understanding of the likely style of glaciation and whether the variation in glacier ELAs is seen within modern glaciated environments.

7.2.2.1 Variation in ELA and mass balance

The lowest mid-latitude ELAs in the northern hemisphere are often in the north-east facing valleys owing to them receiving less solar radiation and being on the leeward side of the prevailing south-westerly winds (Dahl and Nesje, 1992; Benn and Evans, 2010). However, even in regions of extensive glaciation, the summits can remain ice free owing to their steep relief, shape and area, which can lead to snow removal by wind and avalanches (Benn and Evans, 2010).

In the southern Coast Mountains of British Columbia, Canada, the all-sided glaciation level was 300m above the local glaciation level; commonly the lower more favourable sites had northern aspects (Benn and Evans, 2010). In the
Canadian Arctic, work has shown that the ELA of local ice caps is c.100m above the ELA of adjacent corrie glaciers (Miller et al., 1975). This is to be expected as corrie sites typically benefit to varying degrees from avalanches, snow redistribution and radiation shading. This is consistent with the Younger Dryas reconstructions within the Cairngorms, where only the highest plateaus are thought to have possibly been glaciated. The lower plateaus are thought to have been ice free, which matches the notion that glaciated plateaus will be above the ELA of most corrie glaciers. Some corrie glaciers do have particularly high ELAs; however, these have been shown to have high radiation values, such as Coire Bhrochain. In Canada, glaciation levels were found to increase inland, but the pattern was complicated by the variability of the topography; particularly steep gradients of 15m per km occurred along the north-west margin of the Queen Elizabeth Islands (Miller et al., 1975). Four corrie glaciers in the Torngat Mountains, northern Quebec and Labrador, Canada, had a local ELA difference that varied annually but also spatially by an average of c.340m (Rogerson, 1986). The highest ELA was thought to be closest to the regional ELA, while the three more similar ELAs, despite varying internally by 120m, were thought to represent shading effects (Rogerson, 1986).

In south-west Norway, Grovabreen is a plateau glacier with short outlets in all directions. The ELA is 1400m on the southern side and 1315m on the eastern side, with the difference resulting from leeward accumulation of snow (Aa, 1996). Variability also exists between the 20 outlet glaciers of Jostedalsbreen, Norway, where ELAs calculated using the AAR method vary between 1420m and 1735m (Torsnes et al., 1993). The most important factors controlling the ELA here were thought to be topography, glacier hypsometry, climate conditions and aspect (Torsnes et al., 1993). The precipitation associated with south-westerly winds and the precipitation tracks through large fjords, such as Sognefjorden, cause greater accumulation of snow and thus lower the ELAs on the south-west side (Østrem et al., 1988; Torsnes et al., 1993). Local variations in ELAs exist between the western and eastern sides of the ice cap and between nearby outlets (see Torsnes et al., 1993). The Tarfala mass-balance monitoring programme includes seven glaciers in two east-west transects predominantly within Sweden but with one Norwegian glacier (Holmlund and Jansson, 1999). Storglaciären has the largest winter balance of the northern transect due to being on the leeward side of Kebnekaise and the
increased precipitation from the heaving of air masses as they pass over the mountains (Holmlund and Jansson, 1999). The Riukojietna ice cap is closer to the coast and does not rely on leeward accumulation, although local maximum accumulation is on the leeward side of the glacier plateau summit. Similarly if Younger Dryas plateau glaciation occurred within the Cairngorms, it is likely that the highest accumulation depths would have been in sheltered locations on the leeward side of the plateau. Måmglikiären to the north receives less accumulation as it is protected by several large mountains to the west. The summer balances of the glaciers are less variable; however, shading such as on Rabots glacier and differences in cloud cover do lead to variations in the radiation balance (Holmlund and Jansson, 1999). Such differences in radiation from topographic shading and aspect have been identified in Chapter 6.

In Chapter 6 the topoclimatic factors for each glacier were modelled and the best explanatory variables (snow redistribution, solar radiation and south-to-north precipitation gradient) were included within regression analysis which explained up to 80% of the variation in ELA. Similar studies within currently glaciated environments have encouraging similarities. A study within Glacier National Park, Montana, showed that local topoclimatic factors (headwall height, area and slope, and wind exposure/shelter) had significant relationships with the ELA; however, in this case it was not possible to remove all the sub-regional effects with regression analysis to generate a regional ELA for the study area (Allen, 1998). This was thought to be due to missing variables and local meteorological processes, such as local variations in wind and precipitation (Allen, 1998); similar factors are likely to be contributing to the unexplained variation in ELA within the Cairngorms. A study in the French Alps, in an area c.20 km north to south and c.27 km east to west, had a mean present-day ELA of 3024m and a large standard deviation of 175m (Cossart, 2011). Modelling of numerous topoclimatic factors and their inclusion within a regression model showed that solar radiation, longitude, wind-effect and slope were significant and explained approximately half of the variation in ELA, with residuals of 55m (Cossart, 2011). Solar radiation explained some of the spatially abrupt 300–350m variations in ELA; wind explained abrupt 150m variations in ELA from exposed to sheltered slopes; and longitude explained 150m of variation in ELA across the study area (Cossart, 2011). The studies within modern glaciated
environments (Allen, 1998; Cossart, 2011) have similar significant topoclimatic explanatory variables to those used within Chapter 6. The French Alps study explains a similar amount of the variation in modern ELAs to the use of the glaciated topoclimatic factors in the regression models within Chapter 6, whereas the unglaciated topoclimatic factors in Chapter 6 explain a greater proportion of the variation in ELA in the Cairngorms. In summary, modern glaciated environments possess a great deal of variation in glacier ELA and the regression analysis in Chapter 6 compares well with studies of topoclimatic factors in currently glaciated areas.

### 7.2.2.2 Style of glaciation

The Cairngorm Mountains supported relatively small glaciers during the Younger Dryas, with both plateau- and valley-sourced glaciers being reconstructed within Chapter 6 (Figure 6.4 and Figure 6.5). Modern glaciated environments with similar topography and glacier extents can be used as analogues to identify the likelihood of plateau ice. It is understood that there is a non-linear relationship between the breadth of a plateau/summit and the height it must be above the ‘firn line’ (ELA) of adjacent glaciers to support ice (Manley, 1955; 1959). Using data from northern Norway, Rea et al. (1998) showed that one unglaciated plateau existed above the curve and numerous glaciated plateaus existed below the curve, this indicated that the originally recommended relationship may not hold (Figure 7.3). Manley (1955) described the reference ‘firn line’ (ELA) as that of the prevailing glaciers surrounding the summits/plateau; however, owing to topoclimatic factors this can be highly variable. In the Cairngorms only the highest plateaus are considered candidates for plateau Younger Dryas plateau glaciation. It is thought unlikely that the lower 850–900m plateaus, such as Mòine Mhòr, supported plateau ice, except in localised areas, as this would mean plateau-ice accumulation below the ELA of many of the corrie glaciers, and corries and valleys are typically favourable sites for snow accumulation and reduced radiation (Chapter 6). The higher plateaus and summits, such as Beinn Bhrotaín, Braeriach and particularly Ben Macdui, owing to its width, may have been glaciated.
Maps of modern glaciated environments can be used to identify the patterns mentioned in the literature. To the south of Longyearbyen, Svalbard, valley glaciers are present on the northern side of the plateaus and ridges. Ice is not found on the broad lower plateau surfaces such as Platåberget at c.480m, nor the high narrow plateaus such as Foxtoppen at 955m; however, the lower c.800m, broader, nearby Foxfonna plateau supports ice (Figure 7.4). It may have been a similar situation within the Cairngorms with the broad lower plateaus (e.g. Mòine Mhòr and Mòine Bhealaidh) and higher narrow summits (e.g. Cairn Gorm) unglaciated, but where the altitude was high enough and breadth wide enough the plateaus may have been glaciated (e.g. Ben Macdui). Indeed, many of the higher narrow peaks are not glaciated even within the larger Svalbard ice caps. In the more marginal areas of glaciation on Svalbard, the glaciers are confined to sheltered north-facing valleys and corries.
The change in style of glaciation from ice cap, icefields and their associated outlet glaciers in western Scotland to more restricted corrie, valley and limited plateau ice in the Cairngorms is very apparent (see Figure 7.5). An appropriate modern analogue is the west-to-east change in glaciation extent and style in Norway. In western Norway lies the large ice cap of Jostendalsbreen; however, approximately 100–150 km to the east, despite being on the highest mountains in Norway at Jotunheimen, the glacier extents are much more restricted because of a more continental regime (Winkler et al., 2009). In the east, the glaciers are more topographically confined, with ridges and small plateau areas being free of ice and greater preferential glaciation of the north-eastern mountain sides. The mean regional ELAs range from 1200m near the west coast to 2200m near Jotunheimen further east (Liestøl, 1967; Lie et al., 2003b), with increasing ELA gradients of up to 10m per km near Jotunheimen (Figure 7.6). The variation in ELA is caused partly by summer temperature and partly by precipitation (Lie et al., 2003b). The gradients inferred from the regression analysis for the Cairngorms were slightly higher but, given the gradients in Norway (and other locations discussed in Section 6.3.3.5), they do not seem unrealistic. An interesting similarity is that the steepening of ELA gradients towards Jotunheimen is similar to the steepening of gradients towards the Cairngorms, seen in the somewhat outdated regional ELA map by Sissons (1979c).
Figure 7.5 The Younger Dryas ice limits and generalised ELA contours (in metres). Note the large West Highland ice cap and the smaller glaciers of the Cairngorms and South-East Grampians. Redrawn from Sissons (1979c) with permission of Nature Publishing Group and Benn (1997) with permission of Elsevier. Note: many ice limits and ELAs have been reinterpreted since but the overall pattern remains largely correct.

Figure 7.6 Norwegian mean observed regional ELAs (reproduced from Lie et al., 2003b with permission of SAGE Publications, based on Østrem et al., 1988 with permission of NVE Library and Liestøl, 1967)
7.2.2.3 Spatial distribution of modern snow patches

Leonard (1984) compared modern snow accumulation patterns and palaeo-ELAs in the San Juan Mountains of Colorado. When the effect of elevation was removed, there was a resemblance to the lowest ELAs in areas of highest modern snow accumulation (Leonard, 1984). The snow patches that have lasted more than one summer from 1971–2000 in north-east Scotland have been compiled by Watson et al. (2002). The sites within the Cairngorms that have supported snow patches for at least 2 years during that period are: Garbh Chòire Mòr and Coire an Lochain (Braeriach); Garbh Uisge Beag, Garbh Uisge Mòr, low Fèith Buidhe and high Fèith Buidhe (Ben Macdui plateau); Coire an Lochain Uaine (Ben Macdui); Coire Domhain, Ciste Mhearad and Coire an Lochain (Cairn Gorm); Coire an t-Sneachda (Beinn Bhrotain); Coire Creagach (Monadh Mòr); and Ear-choire Sneachdach (Beinn a’ Bhuid). It is not surprising that this list includes some of the highest corrie locations; however, it also includes numerous sites on the sheltered parts of the Ben Macdui and Cairn Gorm plateau. The aspect of these snow patches is predominantly north-east with no patches facing south or south-west (Watson et al., 2002). Given that similar prevailing wind directions are inferred, this indicates that at least the sheltered parts of the Ben Macdui plateau and other high plateaus are likely to have supported Younger Dryas ice. It is also noticeable that there are no snow patches that have survived two years in the Ben Avon range to the east; similarly no Younger Dryas plateau ice is thought to have existed here.

7.2.2.4 Analogues for readvance landsystems

The Younger Dryas glacial landforms within the Cairngorms can be compared both with other sites elsewhere within the British Isles and also with currently glaciated areas to better understand the dynamics of the former glaciers. The boulder-rich, matrix-poor, relatively low-relief moraines and boulder limits associated with the corrie landsystem may be indicative of cold-based ice with low erosion rates (Fitzsimons, 2005), whereas larger moraines composed of finer material, such as those found within valleys such as Glen Geusachan and Glen Derry, are more commonly associated with warm-based ice and glacial erosion (Ó Cofaigh et al., 2005). These moraines suggest larger, thicker, more dynamic glaciers than the Younger Dryas corrie landforms. Studies elsewhere in the British Isles indicate that
valley moraines of similar morphology are thought to have formed at oscillatory ice margins from the dumping and pushing of supraglacial fluvial and mass flow deposits (Lukas and Benn, 2006; Benn and Lukas, 2006). It has been suggested that the material required to produce such moraines may have been supplied from existing glacial and paraglacial deposits, with large moraines being a feature of high sediment availability at the onset of renewed glaciation (Ballantyne, 2002). The largest moraines within the corries were most likely formed when the glacier margin was stationary for a period of time (Bennett and Glasser, 2009), or where their position coincides with the readvance limit; this may mark the rapid transportation of pre-existing debris to the glacier margin (Ballantyne, 2002). The difference between the corrie and valley landsystems may be explained by the availability of sediment and the control that ice thickness and thermal regime has on glacier movement, erosion and moraine-building processes (Fitzsimons, 2005).

Given the lack of corresponding evidence on the plateaus for the moraines within the valleys, it is suggested that if simultaneous plateau ice existed it was predominantly cold-based. Moraines and meltwater channels have been found on Younger Dryas glaciated plateaus elsewhere within the British Isles (e.g. McDougall, 2001; Finlayson, 2006; Brown et al., 2011; Boston et al., 2013); however, the lower temperatures on the higher Cairngorm plateau and the arid climate further east may have restricted landform formation. This may have been similar to the cold-based plateau ice which has more recently retreated to reveal undisturbed blockfields within northern Norway (Gellatly et al., 1988). In addition, the existence of tors suggests that the absence of plateau ice or the presence of cold-based plateau ice within the Cairngorm Mountains is likely to have been a longer-term trend (Hall and Glasser, 2003; Phillips et al., 2006).

7.2.3 Comparisons with previous geomorphological and modelling studies within the Cairngorm Mountains

The Younger Dryas limits proposed within this study have many similarities with those of Sissons (1979a). However, there are some important differences. The Glen Geusachan glacier is slightly more limited than the margin proposed by Sissons (1979a) and Bennett and Glasser (1991); this is for important geomorphological and geochronological reasons presented within Chapters 4 and 5
respectively. The more restricted margin suggests the Glen Geusachan glacier did not dam a lake within upper Glen Dee during the Younger Dryas. Thus the corresponding straight, outermost moraine position within upper Glen Dee, thought to mark the position of the lake-terminating glacier (Midgley, 2001), is older and a slightly more restricted Younger Dryas glacier is also favoured within upper Glen Dee. The more restricted glacier limits assist in explaining the orientation of moraines in upper Glen Dee, and the existence of different landsystems, numerous outwash terraces, and the differences in surface exposure ages at the exit of Glen Geusachan (Chapters 4 and 5). A Younger Dryas glacier was not recognised within Glen Derry by Sissons (1979a), but later Midgley (2001) proposed such a glacier may have existed; a glacier has been reconstructed in upper Glen Derry based on the new geomorphological and dating evidence (Chapters 4 and 5). Coire Dhondail has not been recognised before as supporting a Younger Dryas glacier, but contains good moraine evidence for a later readvance of ice (see Chapter 4 and 6).

There are also important differences in the style of reconstructed glaciers. Sissons (1979a) reconstructed valley glaciers; however, there has since been a greater recognition of plateau ice within the British Isles during the Younger Dryas (e.g. McDougall, 2001; Finlayson, 2006; Brown et al., 2011; Boston et al., 2013). There is little direct evidence for or against Younger Dryas plateau glaciation in the Cairngorms, thus this study has reconstructed and considered both plateau- and valley-sourced glaciers. The prevalence of plateau ice elsewhere within the British Isles alone does not suggest its presence within the Cairngorms due to local variations in climate and topography. There are also differences in the dynamics of glacier retreat. Sissons (1979a) suggested the hummocky moraine marked the stagnation of ice; however, this study concurs with more recent work (e.g. Bennett and Glasser, 1991) that most of the Younger Dryas moraines within the Cairngorms mark former ice-marginal positions during active glacier retreat. This suggests the glaciers were in synchrony with the climate during retreat.

It is also of interest to discuss the importance of the candidate rock glaciers within the Cairngorms. Rock glaciers are thought to occur in dry to moderately humid climates with cool summers (Humlum, 1998). They have been associated with mean annual air temperatures of < -2°C and low accumulation of snow, otherwise the sites would promote glacier formation (Haeberli, 1985; Humlum, 1998;
The most probable Younger Dryas rock glacier site within the Cairngorms is within Coire Beanaidh (Ballantyne et al., 2009a). It is located at c.920m on the north-east side of a thin ridge separating Corrie Beanaidh with the neighbouring Coire Ruadh. This is important as it formed within a corrie that lacked additional snow redistribution from plateau sources (and similarly any plateau-fed ice), otherwise this site may have favoured glacier growth. The periglacial landform within Coire Beanaidh has formerly been used to suggest a precipitation figure of 350–550 mm/yr (19–27% of present day) (Ballantyne and Harris, 1994). The glacier-derived precipitation values are approximately double this estimate using the Ohmura et al. (1992) equation, but when using the Golledge et al. (2010) equation the glacier-derived precipitation values are more similar (Chapter 6). Slightly higher derived precipitation values from the adjacent glaciers may be expected as it is thought these glaciers would have benefitted from greater snow redistribution by wind. Rock glaciers typically occur in similar environments to glaciers with slightly less precipitation; often rock glaciers form where talus inputs are high relative to snow, and glaciers form where snow inputs are high relative to talus (Humlum, 1998). Similar marginal sites have been suggested to be in part pronival (protalus) ramparts. These also indicate environments close to glaciation with perennial or semi-permanent snowbeds, typically beneath rock cliffs (Shakesby, 1997).

Additional candidate glacier sites were identified for their altitude and aspect that do not possess landforms typical of Younger Dryas glaciation, such as Coire Bogha-cloiche, Cnap a’ Chlèirich, and Coire Garbhachlach. The topoclimatic regression analysis indicated that these sites may have contained contemporaneous glaciers (Chapter 6: Section 6.3.3). One particularly noteworthy candidate glacier was Coire Ruadh; this site had favourable topoclimatic factors indicating a glacier of the dimensions reconstructed in Chapter 6 or potentially larger may have existed at this site. Further detailed work in Coire Ruadh could be undertaken to better understand whether a Younger Dryas glacier existed leaving little geomorphological signature or whether evidence has been masked by postglacial slope activity. Other high corries, such as Coire Cas, may have been too shallow for glaciation (Chapter 6).

Previous topoclimatic work within the Cairngorms has included investigating snow redistribution (Sissons 1979a; Purves et al., 1999) and radiation (Sissons 1979a).
Sissons (1979a) suggested the eastern corries benefitted from snow blown from the plateau surfaces; a similar finding was found with both the manual digitisation and cell-based approaches within this study (Chapter 6). Sissons’ (1979a) approach generated smaller snow-blow ratios for the glaciers than the ratios calculated within this study; this was most likely due to Sissons’ ratios being restricted by arbitrary rules such as snow having to travel downhill. The radiation modelling by Sissons (1979a) assisted in explaining some of the higher ELAs, such as Coire Bhrochain, which also had one of the highest radiation values within this study; however, it did not take into account topographic shading or cloud conditions (Sissons and Sutherland, 1976; Sissons, 1979a). Sissons recognised that much of the variation in ELAs was attributable to snow redistribution and radiation differences, but did not attempt to show this statistically or generate a generalised ELA surface owing to the sparse and unequally distributed glacier locations. Purves et al. (1999) used a GIS cell-based snow-transport model to suggest that the corrie glaciers and valley glaciers cannot have existed under the same climatic conditions. The use of a cell-based model within this study found that the valley glaciers had comparable snow-blow ratios to many of the corrie glaciers (Chapter 6). These different conclusions may not only be due to differences in the model itself but also to differences in the model resolution, differences in the extraction area used for each glacier, and to differences between the relative methods for comparing between glaciers/sites. For example, Purves et al. (1999) used 50m resolution, ‘half-moon’ shaped accumulation zones between the cliff tops of the corries/valleys and their mouths, and relative snow depths for comparisons. Purves et al. (1999) accepted that their ‘cookie-cutting’ technique may be limited, especially for larger catchments, and was not fully objective. Whereas this study used a 30m resolution DEM, defined the extraction zone as the whole glacier area and overlooking avalanche zones with slopes greater than 20°, and summed the total snow depth for the glacier and avalanche area before dividing by the glacier area to generate a snow-blow ratio for each glacier. While, in this study, the snow-blow ratios for the larger glaciers were more comparable with those of the corrie glaciers, they alone could not fully explain the lower ELAs. This required the inclusion of radiation modelling and a precipitation gradient within the regression modelling (Chapter 6).
The work of Golledge et al. (2008, 2009) modelled the extent and thermal regime of Younger Dryas glaciers in Scotland. This used the UKCIP (5 km resolution) dataset for spatial variations in temperature and precipitation; the temperature was forced by the GRIP record and the precipitation modified by a west-to-east and south-to-north precipitation gradient away from the West Highland ice cap (80% and 60% respectively). This generated cold-based ice formation over the Cairngorm Mountains (Golledge et al., 2009). Detailed comparisons between the modelling and reconstructed glaciers are not applicable due to the 500m resolution of the modelling and the exclusion of local topoclimatic factors such as wind-blown snow. However, the model fits the overall pattern of there being a reduced ice presence within the Cairngorm Mountains compared with the Western Highlands. This required the model to include larger precipitation gradients from east to west and south to north than exist at present, otherwise a large ice mass would be generated in the Eastern Highlands (Hubbard, 1999; Golledge et al., 2008). The same modelling work has also suggested local ice growth within the Lateglacial Interstade, prior to the onset of the Younger Dryas (Golledge et al., 2008). The geomorphological evidence cannot prove or disprove this proposition, but the build-up of ice during the Lateglacial Interstade may contribute in explaining the seemingly early maximum ice positions reached within the Cairngorms during the Younger Dryas; other reasons are discussed in Section 7.2.4.3.

7.2.4 Palaeoglaciological and palaeoclimatic regional implications

7.2.4.1 Geomorphology

This study has shown that the morphostratigraphic approach can be applied within the Cairngorms and is most likely valid for the nearby regions. This approach has been carried out objectively by undertaking new holistic mapping of the Cairngorms, with analysis of multiple landforms indicating the same Younger Dryas limits. Complications have existed whereby moraines of similar appearance have been identified outside Younger Dryas limits but, by using numerous lines of evidence such as moraines, outwash terraces, talus development, drift and solifluction limits, it has been possible to identify Younger Dryas landsystems that have been supported by dating both inside Younger Dryas limits within Glen Geusachan and Glen Derry, and outside the limits within Glen Geusachan (Everest and Kubik,
In addition to the morphostratigraphy and dating, the mutual exclusivity of the corrie Younger Dryas landsystem and valley Younger Dryas landsystem, and the lack of any cross-cutting of the landsystems indicate they were contemporaneous. This suggests that the Younger Dryas glaciers generated variable moraines depending on sediment availability, glacier size and dynamics, not dissimilar to differences in moraines created recently in modern glaciated environments such as Norway (Winkler and Matthews, 2010).

7.2.4.2 Glacier ELAs, glacier style and climate

Given that derived precipitation values have large associated errors, it seems more appropriate to compare the spatial variation in ELAs. The maximum extent of the proposed Younger Dryas glaciers is used for ELA reconstruction; however, the glaciers may not have reached their maximum extent contemporaneously – this is discussed further within the next section. It is also worth noting that the individual glacier ELAs have different relationships to the regional temperature-precipitation ELA depending on glacier style and the topoclimatic factors. Locally, across a mountain range, the lowest ELAs may be on the windward side owing to enhanced precipitation when precipitation tracks meet the mountain range (Østrem et al., 1988; Torsnes et al., 1993), or on the leeward side, if the windward side has suffered from snow deflation and the leeward side has benefitted from snow accumulation (Dahl and Nesje, 1992; Benn and Ballantyne, 2005). Without the removal of these complicating factors, only general patterns can be established.

The South-East Grampians are located to the south-east of the Cairngorm Mountains (Figure 7.7). Sissons (1972) and Sissons and Sutherland (1976) reconstructed numerous corrie, valley- and plateau-sourced glaciers with rising ELAs from c.500m in the south-east to c.850m in the north-west towards the Cairngorms. Recently, Hughes (2012) reconstructed a plateau ice field in the Lochnager, Callater and Muick region of the South-East Grampians; the ice field had an ELA of 777m, yielding a derived precipitation estimate of 1630±360 mm a⁻¹. It is unclear if Younger Dryas glaciers existed to the west of The Cairnwell (A93) towards Glen Tilt, and further work is required here. To the west of Glen Tilt towards the Drumochter Pass, Sissons (1974) reconstructed the Gaick ice cap consisting of plateau ice and valley outlets. The northern outlets had outer positions
that descended towards the Feshie/Geldie watershed in the south-west corner of the Cairngorm Mountains. This ice mass was thought to be Younger Dryas in age (Sissons, 1974) but others have cast doubt on this (Lukas, 2004; Merritt et al., 2004). The ELAs of the Gaick glaciers rise from 741m to 815m towards the Cairngorms (Sissons, 1980a). However, it is likely these north-east outlet glaciers benefitted from additional wind-blown snow and reduced radiation. Immediately to the west of the Gaick Plateau, the Younger Dryas West Drumochter Icefield has been reconstructed (Benn and Ballantyne, 2005). This plateau-fed system has a best estimate climatic ELA of 622m without snow-blow areas, and 656m including snow-blow areas (Benn and Ballantyne, 2005). The Drumochter Icefield lies c.35 km south-west of the south-west corner of the Cairngorm Mountains. Situated to the north of the Drumochter Hills, on the northern side of the Spey valley, are the Monadhliath Mountains. Recent work has suggested a Younger Dryas icefield existed here with an average ELA of 714±25m, but with rising ELAs from 560–646m in the west to 738–816m in the east (Boston et al., 2013). The Cairngorms are situated to the east and slightly south of the Monadhliath Mountains on the opposite side of the Spey valley. The ELAs rise from the South-East Grampians and Gaick northwards and from the Monadhliath Mountains eastwards towards the Cairngorm Mountains. There is a well-documented rise in ELAs eastwards towards this area from the west coast of Scotland (Sissons, 1979b, 1979c, 1980a; Benn and Ballantyne, 2005).

Despite the Cairngorms possessing high topography including large plateaus up to c.1300m a.s.l., the Younger Dryas glaciers reconstructed within this study are more restricted than those in the surrounding mountain ranges. This is likely to be due to the Cairngorm Mountains’ location situated on the eastern/northern side of the mountain ranges described above (Figure 7.7), causing a reduction in precipitation. This effect must have been substantial given the reconstruction of plateau ice on the c.730–900m Gaick Plateau and c.750–945m Monadhliath Mountains, but absence of ice on plateaus of similar altitude (e.g. Môine Mhòr and Môine Bhealaidh) and even within some of the corries within the Cairngorms. This rise in ELA from the surrounding mountain ranges towards the Cairngorm Mountains suggests that snow-bearing winds were indeed from the west and south (Sissons, 1980a). More detailed analysis of the ELA gradients is hindered by the uncertainty
of the glacier extents in some regions and the uncertainty of how the glaciers compare to the temperature-precipitation ELA. Bearing this in mind, plateau-fed glacier ELAs of c.714m for Monadhliath Mountains, c. 622m (c.656m including snow-blow areas) for West Drumochter Icefield and c.787m for Gaick Plateau can be compared with more topographically favourable sites with typical ELAs of > c.800m in the south-west Cairngorms and > c.900m within more central areas. It is encouraging that the residuals, after the removal of topoclimatic factors within Chapter 6, indicated lower ELAs within the southern and western Cairngorms and that the precipitation gradient favoured by the regression was from south to north. This largely matches with both the regional trend across Scotland and the sub-regional pattern of ELAs within the surrounding mountain ranges described above.

As in Norway at the present time, where variations in ELA are attributable to a combination of precipitation (c.70%) and temperature (c.30%) (Lie et al., 2003b), it is likely that the variations in Younger Dryas ELA across the British Isles were caused primarily by spatial variations in precipitation, but temperature may have also been important. Recent work at the Abernethy Forest site (230m, Northern Cairngorms) suggested mean July air temperatures were higher than at the Loch Ashik site (50m a.s.l., Isle of Skye) during the Younger Dryas (Brooks et al., 2012). This could reflect warmer continental conditions during the summer further inland near the Cairngorms; however, the difference may be influenced by the proximity of the Skye site to a Younger Dryas icefield (Brooks et al., 2012).
7.2.4.3 Readvance timing and climate

To further complicate the comparison of glacier ELAs, it is likely that the maximum positions of the Younger Dryas glaciers did not occur all at once (Golledge, 2010) and there are many good reasons discussed in this section to explain this. The surface exposure dating evidence presented within Chapter 5 and also for the Coire an Lochain glacier (Cairn Gorm) (Kirkbride et al., 2014) suggest the outer glacier limits were reached relatively early within the Younger Dryas. Regardless of uncertainties in the absolute ages due to different calibration datasets, there appears to be a relative difference between the Cairngorm ages and the Younger Dryas calibration sites in western Scotland, suggesting that the Cairngorm glaciers reached their maximum position earlier. This coincides with the coldest period of the Younger Dryas at Abernethy Forest, before a gradual increase in the temperature in the later part of the Younger Dryas, prior to rapid warming at the onset of the Holocene (Brooks et al., 2012). However, sites further west are thought to have reached their maximum positions later in the Younger Dryas. The main Younger Dryas West Highland ice cap reached and then maintained its maximum position near Loch Lomond after 12.0 ka BP (MacLeod et al., 2011). Work by Ballantyne (2012) highlights the variability of deglaciation ages interpreted to
represent Younger Dryas glaciation (Chapter 5); however, it is difficult to determine whether this is due to sample-specific geomorphic effects (e.g. snow and sediment shielding) or spatial differences in climate and glacier response to climate.

While chironomid data has shown that temperature varied both temporally and spatially during the Younger Dryas (Brooks et al., 2012), it is likely precipitation would have also varied throughout the Younger Dryas (Isarin et al., 1998). Although the relationship at the ELA is non-linear, 1°C of temperature equates approximately to 350 mm of precipitation (Ohmura et al., 1992), thus variations in both would have been important to glacier fluctuations. The importance of both temperature and precipitation in controlling the extent and timing of glacier readvance has been shown during the Little Ice Age (Nesje and Dahl, 2003). Most glaciers in Scandinavia are thought to have reached their maximum extent during the mid-eighteenth century whereas, in the European Alps, glaciers reached their maximum extent in the mid-nineteenth century (Nesje and Dahl, 2003). This asynchronous readvance is thought to be controlled by differences in precipitation caused by the North Atlantic Oscillation (cf. Nesje and Dahl, 2003). More recent spatial variations in glacier readvance have been recorded within Norway; these have been attributed to differences in precipitation and to the differences in mass-balance regime between maritime and continental glaciers (cf. Chinn et al., 2005)

Within Scotland, using the oscillatory moraine margins and pollen data on Skye, it was suggested that, within the latter part of the Younger Dryas, more arid conditions caused the glaciers to retreat prior to the rapid retreat at the onset of the Holocene warming (Benn, 1997 and references therein). In addition, other factors such as glacier size/response time (MacLeod et al., 2011; Kirkbride and Winkler, 2012), and temperature and precipitation feedbacks from the growth of large ice masses would have impacted on the former glaciers (Benn and Ballantyne, 2005). The response time of glaciers is greater for larger glaciers and glaciers in more continental climates (Kirkbride and Winkler, 2012); thus it is likely that the smaller Cairngorm glaciers were more sensitive to climatic fluctuations than the larger glaciers further west. The growth of the West Highland ice cap would have caused greater atmospheric chilling and thus reduced snow carried by westerly air masses to the Cairngorms (Benn and Ballantyne, 2005). The reduction in snow carried by westerly tracks may have meant southerly tracks became more important in the
provision of precipitation to the Eastern Highlands, explaining the rise in ELA from south to north across from the Gaick, South-East Grampians and Cairngorms (Sissons, 1980a).

The location of the polar front is an important factor in the climate of the British Isles; it currently lies to north of British Isles, with Scotland receiving a mild maritime climate brought from lower latitudes by the North Atlantic Meridional Overturning Circulation (AMOC). During the Younger Dryas the AMOC weakened, the North Atlantic oceanic polar front migrated southwards and extensive sea ice developed in the North Atlantic (Lane et al., 2013). This caused a decrease in temperature and an increase in seasonality (Isarin et al., 1998; Lie and Paasche, 2006). The polar front is thought to have been south of 50°N during the early Younger Dryas, and then an increase of the AMOC caused the polar front to gradually migrate north towards its present-day position c.62°N in the second part of the Younger Dryas (Lane et al., 2013). This shift and the associated shifts in the southern extent of sea ice and storm tracks during winter may explain the often proposed subdivision of the Younger Dryas (Isarin et al., 1998). Thus while it may have got drier and warmer in mainland western Europe (Isarin et al., 1998), a northwards shift would bring increased precipitation to the Britain Isles. Similarly, Bakke et al. (2009) proposed that sea ice was more extensive during the first part of the Younger Dryas, then ‘flickering’ of the temperature occurred during the second half of the Younger Dryas caused by the alteration between sea ice and warm water; during the warm water phase the westerly winds drifted northwards, melting glaciers in Norway.

It has been suggested that the reduced sea ice during the second phase of the Younger Dryas may have increased precipitation and facilitated the Loch Lomond glacier to reach its outer limit late within the Younger Dryas (MacLeod et al., 2011). However, this increase in precipitation may not necessarily have transferred eastwards to the Cairngorm Mountains owing to the increasing size of the West Highland ice cap. This may have been similar to the more recent positive mass balances of glaciers in western Norway due to increased precipitation, with a less obvious change to glaciers further east (Chinn et al., 2005). Instead, the Cairngorm glaciers may have retreated early within the Younger Dryas because of increased summer temperatures (Brooks et al., 2012), and also because of potentially
reduced seasonality in the second phase of the Younger Dryas, due to the northward shift in winter sea-ice cover. Given that increased seasonality reduces the required precipitation at the ELA (Golledge et al., 2010; Finlayson et al., 2011), seemingly it would increase the required precipitation values to sustain the same ELA and thus the Cairngorm glaciers would have retreated. In addition, in climates of reduced seasonality, glaciers are more sensitive to changes in mass balance (De Woul and Hock, 2005). Regardless of any changes in precipitation and seasonality, the accumulation of ice in the Western Highlands is likely to have reduced the transfer of snow towards the Cairngorms, making the early part of the Younger Dryas favourable for glacier growth in the Cairngorms. An alternative explanation for temporal difference in maximum extent may be that the larger western ice mass had a longer response time to the gradual rise in temperature during the latter part of the Younger Dryas (MacLeod et al., 2011). Similarities exist with Norway, where different parts of the ice sheet, cirque glaciers and valley glaciers responded differently during the Younger Dryas (Mangerud, 1980; Mangerud et al., 2010). Mangerud et al. (2010) also noted that the largest positive mass balances may not have occurred during the coldest phase of the Younger Dryas and Lateglacial Interstad; instead they may have coincided with intermediate climates when precipitation was higher. Given the low precipitation values within the Cairngorms, it is likely any changes in precipitation would have been important to glacier growth both within the Younger Dryas and potentially during earlier periods of climate deterioration within the Lateglacial Interstad. In summary, there are numerous mechanisms for the early maximum position of the Cairngorm glaciers. These include: smaller ice masses responding more rapidly to climate forcing; growth of the West Highland ice cap reducing precipitation; and decreased seasonality and increased temperature during the Younger Dryas.

A recent study has suggested that much of our understanding of Younger Dryas glaciation in Scotland could be incorrect. New radiocarbon ages from Rannoch Moor suggested complete deglaciation by the middle of the Younger Dryas (Bromley et al., 2014). The preferred driver for this was summertime warming and snowline rise caused by increased seasonality during the Younger Dryas (Bromley et al., 2014). The presence of winter sea ice would have resulted in continental winters; while, during the summers, a buoyancy-stratified North Atlantic Ocean
would have increased summer sea-surface temperatures and thus downwind air temperatures in Scotland (cf. Bromley et al., 2014). These new findings are at odds with much of the existing dating and geomorphological work in Scotland; the authors suggest that either the Younger Dryas was a small readvance of existing glaciers early within the Younger Dryas Stade or that some of the glacier margins currently attributed to the Younger Dryas Stade predate the Younger Dryas (Bromley et al., 2014). This will be an area of great interest and discussion for the future.

### 7.2.4.4 Comparisons with modern precipitation values

Modern precipitation values for 21 years of overlapping data from 1983 to 2011 were analysed for Braemar (Met Office Historic Station Data) and Aviemore (Tutiempo, http://www.tutiempo.net/en/). The Braemar station had a mean precipitation value of 915 mm per annum at an average altitude of 336m (the station site moved from 339m to 317m a.s.l. in 2005). The Aviemore station had a mean precipitation value of 980 mm at 228m a.s.l.. Using the procedure of Ballantyne (2002) and Benn and Ballantyne (2005), the values can be scaled to the desired altitude: the Ben Nevis rate is employed here at 5.78% per 100m rise in elevation (Ballantyne, 2002). The scaled precipitation values at 800m and 1000m are 1187 mm and 1329 mm for Braemar, and 1352 mm and 1513 mm for Aviemore. A value of 8% per 100m has been suggested for Norway (Lie et al., 2003b). Using such a value increased the precipitation values to 1307 mm and 1525 mm for Braemar, and 1523 mm and 1777 mm for Aviemore at 800m and 1000m respectively. Scaling to a valley altitude of 350m, Aviemore receives 100–150 mm per annum more precipitation than Braemar. Unfortunately no weather station precipitation data exist for high-altitude sites within the Cairngorm Mountains. The Met Office monthly 5 km x 5 km interpolated datasets (Perry and Hollis, 2005) suggest annual values of c.2000 mm in the high central Cairngorms, much less than values in excess of 3500 mm in the high areas of the Western Highlands (Figure 7.8).

The ELA-derived Younger Dryas precipitation values are substantially reduced compared to modern values. This is likely to have been partly due to increased distances from moisture sources, caused by the presence of sea ice within the
North Sea and much of the North Atlantic during winter (Isarin et al., 1998; Lane et al., 2013). The location of the polar front also led to changes in weather patterns and precipitation as it migrated during the Younger Dryas (Isarin et al., 1998). In addition, the cooling effect of the West Highland ice cap has been suggested to reduce the transfer of precipitation eastwards towards the Cairngorms (Benn and Ballantyne, 2005).

Figure 7.8 Annual precipitation values from the Met Office monthly 5 km x 5 km dataset (Perry and Hollis, 2005). Data available at: http://www.metoffice.gov.uk/climatechange/science/monitoring/ukcp09/
7.3 Informing Future Predictions and Further Research

7.3.1 Informing modelling and future predictions

This work has provided a time-constrained landform record, which can be combined with existing and ongoing work within the British Isles, to be used as an evaluation area for numerical models of glacier-climate interaction. It has been shown that local factors, such as precipitation, impact on the spatial synchrony of ice-sheet retreat and should be taken into account when modelling ice sheets. In addition, the effect of ice growth in the Western Highlands reducing the precipitation transfer further east is another time-transgressive feedback that requires consideration when modelling ice masses. It is also likely that the dynamic variation in storm tracks and sea-ice extent is important to delivery of precipitation and seasonality. The feedbacks mentioned above, such as changes in ice geometry, sea-ice extent and weather systems, may cause location-specific changes in mass balance, and their inclusion within modelling past and current ice masses may be beneficial.

At a smaller scale, topoclimatic factors are important to supporting valley and corrie glaciers and may be useful for predicting future retreat of present glaciers. This may assist in identifying glaciers that will continue to be supported by favourable topoclimatic factors in a warming climate where the regional climatic ELA is rising. This may be beneficial to future planning in communities that rely on glacier meltwater for water supplies.

7.3.2 Development of approaches/methods and relevance to wider research

There are many aspects of this research and its methods that may be valuable to future studies. The increasing resolution of surface exposure dating means understanding the geomorphological processes that may influence the boulder ages are becoming more important (Kirkbride and Winkler, 2012). In this study the cosmogenic surface exposure dating showed a positive relationship between boulder size and apparent exposure age. Inverse modelling was used to calculate the probable moraine degradation rate and the probable moraine formation age. The use of embedded paired boulders of differing heights above the moraine surface and inverse modelling deserves further exploration to identify whether it can assist in removing uncertainties associated with the later exhumation of boulders at
other sites. This technique of using embedded boulders has the potential to account for moraine degradation and assist in identifying boulders that suffer from inheritance or rotation.

The use of the topoclimatic factors employed in this study may allow the comparison of ELAs in areas of differing styles of glaciation, or with differing degrees of favourable topoclimatic factors. The approach may provide a method for comparing the ELAs of relatively small valley and corrie glaciers with larger plateau-sourced glaciers. It may also be useful for comparing areas such as the Cairngorms characterised by the deep valleys and large plateaus where topoclimatic factors have proven to be important, with gentler terrain with less expansive plateaus where topoclimatic factors would presumably be less important to supporting glaciers. Further work may benefit from testing the topoclimatic modelling, or aspects of it, in currently glaciated areas. This may involve utilising data collected from modern glaciers, such as Finsterwalderbreen, Svalbard, where accumulation patterns are thought to be impacted by topoclimatic variables (Hodgkins et al., 2006). In this example, the snow-blow model could be used to ascertain whether it can explain the variation in modern snow depths.

The Younger Dryas margins within the Cairngorms could be utilised for comparisons with higher resolution modelling and specifically models that can be modified to include topoclimatic factors (e.g. Plummer and Phillips, 2003; Harrison et al., 2014). The modelling of topoclimatic factors may also assist in identifying modern glaciers that are likely to retreat quickly or persist for longer as the temperature-precipitation ELA rises in a warming climate.

In palaeoglaciological studies, the upwind limit of the snow-blow area is often defined by a 5° slope (e.g. Benn and Ballantyne, 2005). It is of interest that the modelling of snow redistribution based on wind speeds explained a greater proportion of the glacier ELA than the conventional approach using a slope angle to determine the outer limit of the snow-blow area. This supports the opinion of Humlum (2002) that winter observations in modern glaciated environments such as Greenland and Svalbard suggest that the importance of snow-blow upslope has been underestimated. Using the snow-redistribution modelling approach within this study more widely may explain variations in former glacier ELA at other locations.
7.3.3 Cairngorm Mountains and adjacent regions

This study has provided detailed geomorphological mapping, targeted dating and analysis of topoclimatic factors to establish the extent of Younger Dryas glaciers and the pattern of ice-sheet deglaciation. Progress has been made towards answering numerous important questions, but further uncertainties have arisen.

In the eastern Cairngorms there remains uncertainty regarding the terraces within lower Slochd Mòr and the whether an ice-dammed lake occurred in lower Glen Avon, this requires detailed fieldwork and sediment analysis. There are also uncertainties as to whether Younger Dryas glaciers existed within the candidate sites, such as Coire Bogha-cloiche, Cnap a’ Chlèirich, Coire Garbhlaich and other high corries, such as Coire Cas (Chapter 6). Coire Ruadh is a site of particular interest given the favourable modelled topoclimatic factors. Glaciation within Coire Beanaidh is likely to have been restricted due to limited snow accumulation from plateau-blown snow. This may explain the development of periglacial features, including a candidate rock glacier, at the site. Additional dating of the moraines that dam Loch Avon within Glen Avon would be beneficial to understanding the age of the locally sourced glacier to have existed prior to the Lateglacial Interstade and confirm the extent of the Younger Dryas glacier.

Confirming the proposed Younger Dryas glacier extents within adjacent regions, such as the Gaick Plateau, South-East Grampians and the intervening area, is important to making concrete regional inferences regarding spatial variation in Younger Dryas glacier ELA and the derived precipitation gradients. In addition, the linking of the Cairngorm dates for ice-sheet deglaciation with the north-east coast of Scotland via the Dee and Spey valleys would be beneficial to understanding the retreat of ice once onshore. Retreat patterns for the surrounding areas would help identify the sources and retreat dynamics of the regional ice that dammed the lakes on both the northern and southern sides of the Cairngorms.

The recent abundance of palaeoglaciological studies within the Scottish Highlands and greater utilisation of GIS will allow the amalgamation of glacier reconstructions regionally. By systematically using topoclimatic modelling to better understand how the Younger Dryas ELAs relate to the temperature-precipitation ELA, a better
understanding of regional ELA gradients may be developed. Using the relationship between summer temperature and precipitation will enable us to understand what percentage of the ELA gradient is attributable to local changes in summer temperature and precipitation.
7.4 Chapter Summary

This chapter has provided a synthesis of the results and analysis presented within Chapters 4, 5 and 6, and a discussion of the findings with reference to modern glaciated environments, and previous geomorphological and modelling studies.

The establishment of Younger Dryas limits allowed patterns of Dimlington Stade deglaciation to be finalised. The southern Cairngorms underwent a similar east-to-west deglaciation as the northern Cairngorms, with a complex juxtaposition of regional ice around the Cairngorms. The large ice-dammed lakes and the limited extent of the locally sourced Cairngorm glaciers, while Dee and Spey ice still impeded on the edges of the Cairngorms, illustrate the early deglaciation of much of the Cairngorms. The retreat of local glaciers at this time suggests a lack of precipitation in the Cairngorms when the ice sheet remained relatively expansive. The Cairngorm ice remained active and formed moraine sequences in the lower valleys on both the northern and southern sides of the Cairngorms. The final stage of ice-sheet deglaciation and its timing is unclear because of the readvance of ice in the most favourable topoclimatic sites during the Younger Dryas.

The variation within the Younger Dryas ELAs is not unusual compared with variations in ELAs in glaciated environments, and the topoclimatic factors responsible are consistent with the modern glaciated environment. The pattern of ELA residuals, after removal of topoclimatic factors, fits with the regional pattern of ELAs rising north and east towards the Cairngorms. These ELA gradients are not dissimilar to gradients found within modern glaciated environments such as Norway and the Canadian Arctic. The readvance of ice during the Younger Dryas would seemingly have been earlier within the Cairngorms than the Western Highlands. Many mechanisms exist including different glacier response rates and precipitation differences due the presence of the West Highland ice cap. It is clear that the precipitation was much reduced compared to the Western Highlands, a pattern that was most likely seen during ice-sheet retreat and former periods of glaciation. The Younger Dryas values are much reduced compared to modern precipitation values owing to lower sea levels, the presence of sea ice and the cooling effect of the West Highland ice cap. The final section of this chapter has suggested future avenues for research both within Scotland and the broader research area.
8 Conclusion

This thesis has presented new detailed seamless geomorphological mapping for the Cairngorm Mountains which, combined with new geochronological control, has provided the basis for the reconstruction of time-constrained former glacier margins. The evidence suggests active retreat during Dimlington Stade ice-sheet deglaciation, punctuated by multiple stillstands and/or readvance events on both the north and south sides of the Cairngorm Mountains. During ice-sheet deglaciation prior to the Lateglacial Interstadial locally sourced glaciers formed within the Cairngorms. New evidence has been presented for a sequence of ice-dammed lakes that formed from east to west along the southern side of the Cairngorms between locally sourced and regionally sourced ice masses. These ice-dammed lakes, in addition to those on the northern side of the Cairngorms, indicate the relatively early retreat of locally sourced glaciers from the lower valleys. This was most likely instigated by low precipitation values on the eastern side of the British–Irish Ice Sheet.

The new geomorphological evidence, combined with new cosmogenic surface exposure ages from moraines previously subject to differing age interpretations, indicate a later readvance occurred within selected corries and valleys. Analysis of surface exposure ages indicates the readvance within the Cairngorm Mountains may have occurred early relative to the final readvance of ice within the Western Highlands, but both were most likely caused by the climate deterioration during the Younger Dryas event. New modelling of topoclimatic factors showed that the snow-blow ratios and radiation values were comparable between the larger valley glaciers and the more restricted corrie glaciers. Indeed, the majority of the Cairngorm glaciers benefitted from south-westerly winds transporting snow across the extensive high-altitude plateaus and rounded topography, and also the low radiation values associated with the glaciers’ favourable aspects and shading from the steeply incised valley and corrie walls. The use of multiple linear regression to remove the modelled topoclimatic factors generated a spatial variation within the ELA residuals, rising from the south-west corner towards the northern margin of the Cairngorms, with a south-to-north precipitation gradient explaining the greatest proportion of ELA variation. These inferred precipitation gradients match with the
spatial pattern of ELAs within adjacent mountain ranges, providing reassurance to the approach and findings. The reduced precipitation in the Cairngorm Mountains mimics the inferred pattern during the retreat of the British–Irish Ice Sheet and is likely to have been a longer-term feature of Quaternary glaciation.

The work would benefit from comparisons with future studies that progress our understanding of Younger Dryas glaciation within adjacent mountain ranges, and from sophisticated regional analysis of ELAs and precipitation gradients, including the use of topoclimatic factors. Targeted dating within adjacent mountain ranges to establish Younger Dryas limits, and dating of landforms between the Cairngorm Mountains and the north-east coast of Scotland would improve our knowledge of Younger Dryas glacier extent and ice-sheet retreat rates. The use of boulders of differing heights to better understand geomorphic impacts on derived surface exposure ages and detailed analysis of the vulnerability of the ages to geomorphic impacts may prove fruitful in extracting more robust information from surface exposure ages. The glacier margins and palaeoclimatic inferences associated with the demise of the British–Irish Ice Sheet and the Younger Dryas readvance can be included within larger mapping projects such as the BRITICE database and utilised by the modelling community for model evaluation.

In summary, the Cairngorms contain a high density and variety of geomorphological landforms from both the British–Irish ice-sheet deglaciation and the Younger Dryas readvance. Based on the new geomorphological mapping, this thesis provides a model of ice-sheet deglaciation for the Cairngorms, including ice-dammed lakes, numerous former glacier margins and an understanding of how regional and local ice interacted. The thesis has also provided new geomorphological, geochronological and topoclimatic evidence proposing that Younger Dryas glaciation in the Cairngorm Mountains, while restricted in comparison to the Western Highlands, may have been more extensive than the previously accepted consensus and included glaciation of favourable valleys.


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Appendix

Quantifying age and matrix erosion rates from cosmogenic surface concentration in paired moraine boulders.

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1 Introduction

The accumulation of cosmogenic isotopes in a exposed surface depends on the exposure time and the erosion processes that affects the surface. Assessing the exposure age of a given surface requires to quantify the erosion processes or to assume that their effects are negligible. In case of unconsolidated sediments, such as till, erosion processes cannot be ignored. To solve this, I can build a simple model that reproduce the concentration of $^{10}$Be and $^{26}$Al at the top of the boulders in a moraine which matrix is been eroded at a constant rate. This model may be used to infer both the erosion rate and exposure age of a moraine from the cosmogenic concentrations measured at the top of a set of neighbouring boulders.

2 Theory

2.1 Model basis

The cosmogenic concentration ($C$) at the surface of a deposit, which has been eroding at a constant rate \( \varepsilon \) (g cm\(^{-2}\) a\(^{-1}\)) since its deposition (\( t \) years ago), was described by [3] as:

\[
\frac{\delta C}{\delta t} = P + \varepsilon \frac{\delta C}{\delta x} - \lambda C
\]

where \( x \) is depth in g cm\(^{-2}\), \( P \) is the surface production rate in at g\(^{-1}\) a\(^{-1}\), which can be calculated using the CRONUS-Earth online calculator [1], \( \Lambda \) is the attenuation length of the cosmic radiation (g cm\(^{-2}\)), and \( \lambda \) is the $^{10}$Be decay constant (a\(^{-1}\)). Eq. 1 may be solved as:

\[
C(\varepsilon, t) = \frac{P}{\lambda + \varepsilon} \left( 1 - e^{-\frac{t}{\lambda + \frac{\varepsilon}{\lambda}}} \right)
\]  

(2)

Considering that the production of $^{10}$Be and $^{26}$Al are due to spallation, stopping muons and fast muons, which have different attenuation lengths and production rates, the superficial $^{10}$Be or $^{26}$Al concentrations can be expressed as:

\[
C(\varepsilon, t) = \frac{P_{\text{spal.}}}{\lambda_{\text{spal.}} + \varepsilon} \left( 1 - e^{-\frac{t}{\lambda + \frac{\varepsilon}{\lambda_{\text{spal.}}}}} \right) + \frac{P_{\text{stop}}}{\lambda_{\text{stop}} + \varepsilon} \left( 1 - e^{-\frac{t}{\lambda + \frac{\varepsilon}{\lambda_{\text{stop}}}}} \right) + \frac{P_{\text{fast}}}{\lambda_{\text{fast}} + \varepsilon} \left( 1 - e^{-\frac{t}{\lambda + \frac{\varepsilon}{\lambda_{\text{fast}}}}} \right)
\]  

(3)

In the case of the concentration at the top of a boulder that was buried after deposition and exposed later due to a constant erosion rate (Fig. 1), and considering a negligible erosion rate of the boulder surface, the final concentration will be:

\[
C_{(h, \varepsilon, t)} = C_{(\varepsilon, t - \frac{e^h}{\varepsilon})} e^{-\lambda \frac{e^h}{\varepsilon}} + C_{(0, \varepsilon)}
\]  

(4)

where \( h \) is the height of the boulder in cm and \( \rho \) is the density of the erodable material, i.e. the matrix, expressed in g cm\(^{-3}\). If the top of the boulder has been exposed since its deposition, implying \( t < \frac{e^h}{\varepsilon} \), the final concentration will be:

\[
C_{(h, \varepsilon, t)} = C_{(0, t)}
\]  

(5)

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2.2 Paleo-depth-profile model

As we cannot, a priori, assume a constant exposure since boulder deposition, a general model that considers constant erosion of the surrounding matrix should be considered. Eqs. 5 and 4 describe the concentration on the currently exposed top of a boulder that has been exposed during and after its deposition, respectively. Using these equation a continuous model can be constructed as follows:

\[
C_{(h,t)} = \begin{cases} 
C_{(0,t)} & \forall t \leq \frac{\rho h}{\varepsilon} \\
C_{(\varepsilon, t)} & \varepsilon < \frac{\rho h}{\varepsilon} + C_{(0, \varepsilon)} & \forall t > \frac{\rho h}{\varepsilon} 
\end{cases} \tag{6}
\]

The main limitations of this models rely in the assumptions of constant erosion rate and boulder stability. However, if only nearby boulders that are partially buried in the sediment are sampled, these assumptions may be reliable. Moreover, generalized boulder rotation would produce inversions in the concentration dataset and could be detected.

2.3 Model fitting

Moraine sampling may provide a set of data of \( N \) concentrations \((C_i)\) measured in samples obtained from several nearby boulders, which may have different heights \((h_i)\). Numerical modeling can be performed to calculate the time \((t)\) and erosion rate \((\varepsilon)\) values that fit the data obtained from the sampled boulder. To fit the model (Eq. 6) into the data, it is necessary to compute it with an inverse method, especially if uncertainty boundaries need to be demarcated in the \( \varepsilon \)-\( t \) space. We will use the same \( \chi^2 \) fit-based inverse method employed by [2] or [5]. This method defines the solution in the \( \varepsilon \)-\( t \) space by minimizing the \( \chi^2 \) value:

\[
\chi^2 = \sum_{i=0}^{N} \left( \frac{C_i - C(h_i, \varepsilon, t)}{\sigma_i} \right)^2 \tag{7}
\]

where \( C_i \) are the measured concentrations from the \( N \) samples at \( h_i \) heights, \( C(h_i, \varepsilon, t) \) is the concentration predicted by the model for each \( h_i \), and \( \sigma_i \) are the uncertainties of the measured concentrations.

As the data are subject to measurement errors and as the erosion processes are more complex than the model assumptions, data never fully fit the model, but only one solution that minimizes the \( \chi^2 \) value is always observed in the \( \chi^2 \) function. This solution, which is defined by the maximum likelihood \( \varepsilon \)-\( t \) combination, is called \( \chi^2 \) best fit value \((\chi^2_{\text{min}})\). To establish to what degree a model reflects the distribution of the data and the \( \varepsilon \)-\( t \) values that fit the data within a certain confidence level, the quality factor of the \( \chi^2 \) values can be used [4].

3 Cairngorms paired samples

Figure 2 show the theoretical \( ^{10}\text{Be} \) and \( ^{26}\text{Al} \) concentrations we can expect for the different hypothesized ages and different erosion rates. Fitting the paleo-depth-profile model to paired boulder \( ^{10}\text{Be} \) datasets will allow us to discern ages and erosion rates of moraine surfaces being eroded at erosion rates ranging 0-50 mm ka\(^{-1}\). This erosion rate range implies surface lowerings of c. 0-0.5 m since moraine abandonment.

If the \( ^{10}\text{Be} \) data yield (a) unexpected ages or (b) erosion rates > 50 mm ka\(^{-1}\), measurement of \( ^{26}\text{Al}/^{10}\text{Be} \) concentration ratios from the same
Figure 2: Theoretical $^{10}$Be and $^{26}$Al concentrations at the top of paired neighbouring boulders (20 and 100 cm high) exposed for 10, 15 and 18 ka at the top of a moraine that is eroded at different rates (0, 50 and 100 mm ka$^{-1}$). Production rates were calculated according to [1] for the lowest sampling site in the study area. Considering AMS, blank and chemistry error propagations, typical concentration uncertainties are depicted (5% and 7% of $^{10}$Be and $^{26}$Al concentrations, respectively). Paired boulder sampling allows to calculate the erosion rate at which the surface is being eroded and, therefore, to deduce minimum and maximum exposure ages.

samples will allow us (a) to detect inheritance effects ($^{26}$Al/$^{10}$Be$\ll P_{^{26}Al}/P_{^{10}Be}$), or (b) to get a better fitting based in the paleo-depth-profile model and the difference in the $^{26}$Al/$^{10}$Be between the paired samples ($P_{^{26}Al}/P_{^{10}Be}$ varies with depth [1] and, therefore, also with the exposure history)

References


