The age and palaeoclimatic significance of ice-dammed lake deposits, Western Cairngorm Mountains, Scotland.

Final report of contract GA00E19 to the University Collaboration Advisory Committee (UCAC).

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The age and palaeoclimatic significance of ice-dammed lake deposits, Western Cairngorm Mountains, Scotland.

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Front cover

The former glaciolacustrine delta in Gleann Einich, looking south. BGS registered photos P521657 & P521658.

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Foreword

This report is the published product of a study by the British Geological Survey (BGS) in collaboration with the University of Aberdeen and Scottish Natural Heritage. The work presented here is the culmination of an 18 month contract funded by the University Collaboration Advisory Committee (UCAC) employing relatively new laboratory techniques to determine the date of deposition and burial of glaciolacustrine sediments in the Cairngorm Mountains, Scotland. The project was supported within BGS by the Central Highlands strategic mapping project.

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Contents

Foreword i

Acknowledgements i

Contents ii

Figures iii

Tables iii

Summary iv

1 Introduction 5
- 1.1 Aims and justification 5
- 1.2 Palaeoenvironment and previous investigations 5
  - 1.2.1 Study area 5
  - 1.2.2 North Atlantic palaeoclimate 7
  - 1.2.3 Scotland during the last glacial cycle 7
- 1.3 Luminescence dating 8
  - 1.3.1 The luminescence technique 8
  - 1.3.2 Limitations and problems 10

2 Sample sites and sedimentology 12
- 2.1 Gleann Einich 13
- 2.2 Lairig Ghru 14
- 2.3 Summary of facies sampled 15
- 2.4 Section logs 16

3 Methodology 16
- 3.1 Field sampling 16
- 3.2 Summary of sample data 17
- 3.3 Laboratory preparation 18

4 Results 18
- 4.1 Dosimetry 18
- 4.2 Luminescence analysis 19

5 Discussion 22
- 5.1 'Quality control' of data 22
- 5.2 Significance of results 23
  - 5.2.1 Overview 23
  - 5.2.2 Data distribution and weighted means 25
  - 5.2.3 Implications for the chronology of 'local' ice movements 26
    - 5.2.3.1 Gleann Einich 26
    - 5.2.3.2 The Lairig Ghru 27
  - 5.2.4 Regional significance 27

6 Conclusions 29

References **30**

Figures

Figure 1: Location map of field area 6
Figure 2: Northern Hemisphere climate curve 8
Figure 3: Energy-level diagram of the thermoluminescence processes 9
Figure 4: Zeroing of the luminescence signal 10
Figure 5: Impact of water depth on the spectrum of light 11
Figure 6: Fluctuations in luminescence signal intensity 12
Figure 7: Location and geomorphological context of the sample sites 13
Figure 8: The distal glacio-deltaic sequence in Gleann Einich at [NH 923 044] 14
Figure 9: The sediment sequence exposed in the Lairig Ghru at [NH 949 054] 15
Figure 10: Vertical lithological sections showing sample horizons at the two localities 16
Figure 11: Obtaining *in situ* dosimetry data from the Gleann Einich sample site 17
Figure 12: Decay of natural luminescence during infra-red or blue-light stimulation. 19
Figure 13: Luminescence growth curve for a Single Aliquot Regeneration analysis 20
Figure 14: Preheat plateau test for blue-stimulated luminescence for sample ABU_203a 20
Figure 15: Preheat plateau test for IR-stimulated luminescence for sample ABU_203b. 21
Figure 16: OSL - derived ages for glaciolacustrine sediments 24
Figure 17: OSL – derived ages for the most reliable of the samples analysed 24
Figure 18: Frequency histograms of sample ages 25

Tables

Table 1: Sedimentology of the two facies types sampled 15
Table 2: Relative sample locations and lithological characterisitics 18
Table 3: Dosimetry data 19
Table 4: Luminescence data for each of the nine samples. 21
Table 5: Quality assessment of samples and derived data. 23

Summary

In this report we present the findings of a collaborative project run between the British Geological Survey, Aberdeen University and Scottish Natural Heritage to apply the relatively new technique of luminescence dating to glaciolacustrine sediments deposited in two valleys during final deglaciation of the Cairngorms in the Late Devensian. The technique employed, Optically –Stimulated Luminescence (OSL), uses either infra-red or blue light to stimulate either feldspar or quartz grains respectively. On stimulation, 'trapped' electrons are released from locations within the mineral grains, stimulating the generation of photons. The photons can then be counted to give a measure of the intensity of luminescence, from which an age for the deposition of the sediment may be calculated.

Comparisons are made between the data presented here, and that presented by other authors using different techniques in the same area, and an assessment of the palaeoclimatic implications of these data is made. All dates quoted are calendar years unless otherwise stated. Calendar years quoted in addition to radiocarbon years have been calculated using the Internet-served calibration tool 'Calib v. 4.3' at http://radiocarbon.pa.qub.ac.uk/calib/calib.html, 29th October 2002.

1 Introduction

Advances in our understanding of global ice age history have given renewed impetus for the systematic reappraisal of the terrestrial and offshore record of deglaciation of the last British Ice Sheet (BIS). This report examines evidence for the timing of a period of ice dammed lake formation during deglaciation, marking the interaction between a large glacier system lapping around the northern flanks of the Cairngorm Mountain massif, and outlet glaciers from Cairngorm plateau icefields.

1.1 AIMS AND JUSTIFICATION

The project described in this report focussed on four key objectives, set out below. These aims have been amended slightly from those initially defined in the project proposal to accommodate unforeseeable problems, such as the 2001 Foot and Mouth outbreak and restricted land access.

i) To determine, through Optically-Stimulated Luminescence (OSL) dating techniques, the age of glaciolacustrine sediments as identified at two sites on the northern flanks of the Cairngorms; Gleann Einich, and the Lairig Ghru.

ii) To use this information a) to identify within-site variation in age of the sediments, b) to help determine sedimentation rates within the lakes, c) to constrain spatial variations in the timing of local valley glacier and ice lobe fluctuations at those sites and d) to assess how this impacts on other reported ice movement in the Cairngorms.

iii) To relate the age and manner of formation of those glacial lakes and their associated deltas to climatically-driven events elsewhere in the British Isles, and in the north Atlantic region at that time (eg Bond *et al.*, 1993; Dansgaard *et al.*, 1993).

iv) To evaluate the relative merits of luminescence techniques in dating glacial deposits in the Scottish Highlands, and to highlight any limitations of this approach.

1.2 PALAEOENVIRONMENT AND PREVIOUS INVESTIGATIONS

1.2.1 Study area

The Cairngorm Mountains constitute approximately 800 km^2 of the Central Highlands of Scotland and form the largest area of British upland terrain above 1000 m O.D. The massif is a late-Silurian to early-Devonian granite pluton intruded into Neoproterozoic metasediments. Repeated glaciation during the last 2.4 Ma, the Quaternary Period, has selectively exploited pre-existing structural weaknesses in the granite (Sugden, 1968; Rea, 1998) to form a landscape of significantly higher relief than that of the earlier Tertiary surface (Thomas *et al.*, in press). Landforms indicative of extensive glacial erosion, such as overdeepened troughs, corries, and scoured plateaux surfaces are common in the area, and are most likely the product of repeated cycles of ice sheet growth and decay. The numerous depositional landforms seen in the Cairngorms, such as moraines, outwash terraces and former lake deltas mainly reflect the last stage of glaciation in the area, which is thought to have taken place between approximately 22ka BP – 15ka BP.

Two principal models have been proposed to explain the pattern of deglaciation of the Cairngorm Mountains. Early geological mapping in the area identified that during the Main Late Devensian (MLD) glaciation, the ice sheet that developed on Rannoch Moor, 75 km to the south-west of the Cairngorms, spread east and northward following the valley of Strath Spey (Barrow *et al.*, 1913). A lobe of this ice stream became embayed in the topographic depression of Glenmore Basin, below the northern flanks of the Cairngorms (Hinxman and Anderson 1915). A local ice cap, possibly composed of several independent plateau icefields, developed on the Cairngorm Mountains and eventually coalesced with the lobe of ice in Glenmore, most probably around the time of the Late Devensian glacial maximum, c. 22 ka BP.

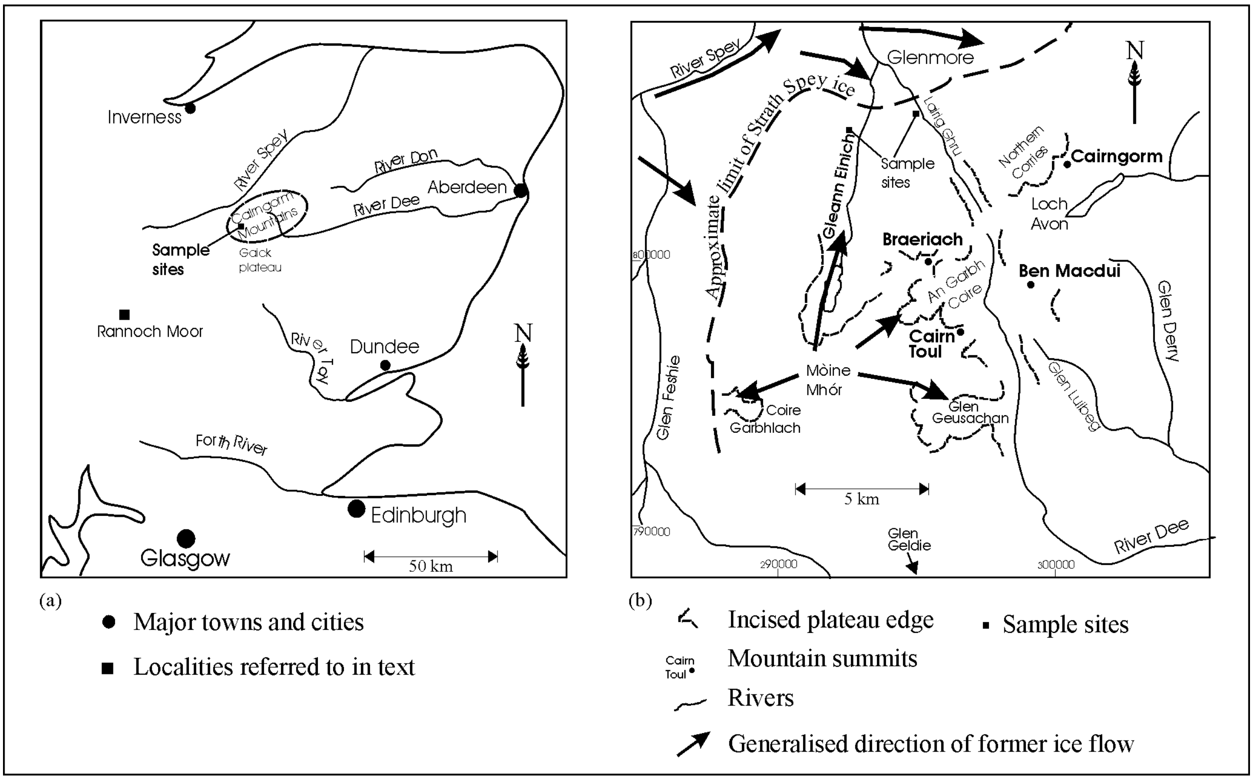


Figure 1: Location map of field area showing a) relation to wider area, and b) generalised ice movements during the Main Late Devensian. Redrawn from: Golledge (2002)

Subsequent climatic amelioration and / or precipitation starvation led to the separation of the two ice masses as the Rannoch Moor ice sheet retreated and the Cairngorm ice cap was reduced to plateau icefields feeding active valley glaciers (Brazier *et al.*, 1996a, b; Golledge, 2002). Loch Etteridge, c. 20 km south-west of the Cairngorms in Glen Truim, is thought to have been ice-free from at least 13 ka ^{14}C BP (c. 15.6 ka cal. BP) (Sissons and Walker, 1974), which implies that complete deglaciation of Strath Spey took place within about 5000 years.

An alternative model proposed by Sugden (1970) was influenced by the then current Scandinavian model of rapid, large-scale stagnation of the ice sheet, largely downwasting in situ and with increasing topographic control exerted on meltwater discharge. Sugden reinterpreted moraine ridges identified by earlier workers, as glaciofluvial landforms, formed subglacially. Detailed mapping of the extensive suites of moundy glaciofluvial deposits and landforms found on the valley floors of Abernethy forest, Glenmore and Glen Feshie lent further support to the ice stagnation hypothesis (Young, 1974, 1975a, 1975b). Recent research in various localities in the Scottish Highlands, however, reveals a greater complexity to ice sheet deglaciation history than accommodated by this hypothesis (Robinson and Ballantyne, 1979; Sutherland, 1984; Brown, 1993; Merritt *et al.*, 1995; Brazier *et al.*, 1996a, b; Golledge, 2002).

Whilst this 'mass-wasting' hypothesis has been superceded in recent years, the relative chronology of the 'active retreat' model is now widely accepted, although little evidence exists to constrain the various stages of deglaciation that are reflected in the landform assemblages. Until recently, no landforms within the Cairngorm Mountains had been dated at all, due to the relative paucity of material suitable for established techniques such as ^{14}C or tephrochronology. Recent advances in the application of laboratory techniques to glacial deposits have enabled some initial dates to be presented for deglaciation of the area (Everest and Kubik, 2001).

1.2.2 North Atlantic palaeoclimate

The air masses whose influence prevails in Scotland are significantly affected by the warm ocean water of the Gulf Stream, part of the North Atlantic Current circulating from the tropics to the Arctic Ocean. The link between ocean and atmosphere is well understood, and it is becoming increasingly apparent that abrupt changes in circulatory patterns of the Atlantic Ocean can promote dramatic climatic fluctuations affecting not only Scotland but the whole of the northern hemisphere. It is thought that numerous rapid shifts in ocean circulation occur throughout much of a glacial cycle (Clapperton 1997), a pattern that is reflected in oxygen isotopes derived from both seabed sediments and ice cores.

Trigger mechanisms for ocean circulation changes and consequent climate transitions are numerous and inter-connected, or 'coupled', where each individual factor promotes feedback from others. For example, the periodic release of iceberg swarms from the North American Laurentide Ice Sheet (LIS) into the North Atlantic during the last glacial period, termed Heinrich Events (HE), occurred on a c. 7 – 10 ka timescale. They are postulated to be the cause of dramatic cooling in the Late-glacial and during the onset of the Loch Lomond (or Younger Dryas) Stadial (Figure 2). The feedback mechanisms involved relied on a coupled ocean – atmosphere system, which translated the massive iceberg discharge into cooler ocean surface water and suppressed the North Atlantic Current. In turn this prevented the warm water of the Gulf Stream from reaching as far north, which consequently allowed the polar front to migrate south. However, it is now thought that the LIS was *itself* responding to a climate forcing (perhaps reduced solar radiation), and simply propagated the effect throughout the North Atlantic. Glaciers in Scotland are likely to have advanced during such Heinrich Events (Clapperton 1997).

Other climate fluctuations are known to exist, such as Dansgaard-Oeschger (D-O) Cycles (2 –3 ka periodicity), and Bond Cycles (grouped D-O cycles resulting in a HE). These patterns can be identified in the high-resolution climate signals derived from ice cores and seabed sediments, and suggest an overall pattern of a highly variable and unstable climate. Identifying these fluctuations in the Scottish landform assemblage is an ongoing challenge, particularly as a lag time may exist between initial climate change and the transmission of this into glacier oscillations and landform genesis.

1.2.3 Scotland during the last glacial cycle

During the last glacial cycle, the Devensian, ice growth in Scotland led to the formation of an ice sheet that extended well into the North Sea, and may have become confluent with the ice sheet covering Scandinavia at around 40 ka BP (Heinrich Event 4). The maximum extent of the British Ice Sheet (BIS) was achieved during a period known as the Last Glacial Maximum (LGM), and is thought to have occurred during the MLD c. 22 ka BP (Heinrich Event 2). This period of time also coincided with lower levels of solar insolation in mid to high latitudes in the northern hemisphere, due to the earth's tilt and precession (Clapperton, 1997). Partial deglaciation may then have taken place, around 21 ka BP, before the ice again expanded during a cold reversal c. 19.5 BP (Bowen *et al.*, 2002). It is suggested by Clapperton (1997) that the submerged Wee Bankie - Bosies Bank moraines found c.35 km off the east coast of Scotland, and submerged moraines found offshore of the Outer Hebrides date from this last major expansion of the British ice sheet, c.18 ^{14}C ka BP (Hall and Bent, 1991; Sutherland and Gordon, 1993).

Interpretations made on the basis of ice sheet modelling have also refined earlier assumptions of a single ice dome covering Scotland. The reality was probably far more complex, with undulating ice surfaces, fast-flowing ice streams and a number of exposed nunataks away from the main centres of ice accumulation in the west (Boulton *et al.*, 1991). Indeed, recent research on ice sheet trimline evidence in western Scotland has identified a number of nunataks rising above the ice sheet surface (Ballantyne, 1990, 1997, Ballantyne *et al.*, 1998; Ballantyne and McCarroll, 1995; McCarroll *et al.*, 1995). The evidence from these studies strongly suggests that both the timing of ice sheet maxima (extent and / or thickness) varied across the British Isles, and that the balance in the interaction between locally generated ice masses and regional ice masses also changed throughout the duration of the last British Ice Sheet. Following these maxima, climate cooling appears to have proceeded until Heinrich Event 1 c. 16 - 17 ka BP, which promoted a final expansion of the BIS prior to its complete collapse. Scotland is likely to have been entirely ice-free by c.13.5 ka BP, the Windermere Interstadial (WI), although this is still widely debated.

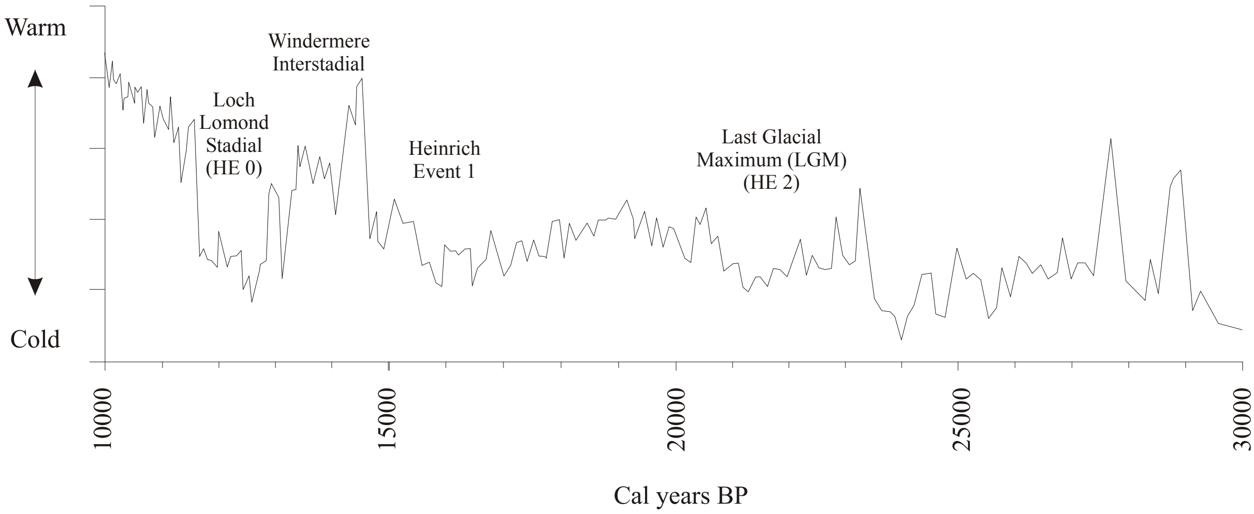


Figure 2: Northern Hemisphere climate curve for the Late Devensian, redrawn from the GISP2 ice core record.

The final stage of the last glacial cycle was a brief return to cold conditions following the W.I., a period known as the Younger Dryas, or more specifically in Scotland as the Loch Lomond Stadial. During this period, c. 12.5 – 11.5 ka BP, an ice cap grew in the Western Highlands, centred around Rannoch Moor, and renewed glacial cover spread across the surrounding high ground. In the Cairngorm Mountains and Central Highlands, a 'rain-shadow' effect prevented the growth of significant ice bodies, and glaciation was restricted to corrie glacier growth on the higher slopes. Given our present climate, it is estimated that a drop of only 2.5 – 3 ºC in our current mean domestic summer temperature might precipitate this kind of glacier re-growth in the west of Scotland (Clapperton 1997).

1.3 LUMINESCENCE DATING

1.3.1 The luminescence technique

Luminescence dating uses the impact of natural ionizing radiation on the atomic structure of mineral grains to determine the date of deposition of those grains as part of a sedimentary accumulation. This is possible because the luminescence signal from the grains will gradually increase over time following the deposition of the mineral particles, the rate of increase of the signal being a function of the ambient flux of ionizing radiation the grains are exposed to, and the inherent characteristics of the minerals of which the grains are composed. If the total amount of stored luminescence (the so-called Environmental Dose) can be measured and the rate of increase of that store (the Dose Rate) can be determined, then the length of time it took the stored luminescence to accumulate in the mineral grains can be determined by simple division. That length of time can be considered the 'age' of the sedimentary deposit. This simple relationship is predicated upon the mineral grains having been 'zeroed' (i.e. any stored luminescence having been removed prior to deposition so that the luminescence 'clock' is set to zero at that moment).

In order to understand the nature of the luminescence 'clock', it is necessary first of all to consider the basis of the actual luminescence signal. Natural ionizing radiation emitted by isotopes of uranium, thorium and potassium in any soil or sediment, together with cosmic radiation, causes electrons within the crystal lattice of a mineral to become detached from the parent nuclei (Figure 3(i)). While the great majority of these electrons recombine almost instantaneously with the parent, a certain proportion of them become held at other locations in the mineral lattice (Figure 3(ii)), usually associated with defects in the structure such as impurities or damage resulting from nuclear radiation. These defects, particularly those exhibiting a negative ion vacancy, attract ionized electrons, and may effectively trap them, preventing their immediate recombination with parent nuclei. Once 'trapped' in this fashion, electrons remain in the trap until the mineral grain is subject to sufficient energy for them to be released - Aitken (1985) refers to the process as vibrations in the crystal lattice causing the trapped electrons to be 'shaken out' of the traps (Figure 3(iii)). In the absence of such energy levels affecting the mineral grains, increasing numbers of electrons become trapped over time in the crystal lattice. The rate of accumulation of trapped electrons is dependent upon a number of factors, but one of the most important is the flux of ionizing radiation affecting the grains (the Dose Rate). The greater the flux, the more rapidly the number of trapped electrons builds up.

The actual luminescence is produced by the release of 'trapped' electrons from the defects where they were held. As mentioned above, this usually occurs when the mineral grains are subject to an increase in energy levels. Such increases usually result from application of heat energy or light energy to the mineral grains. Released electrons may be retrapped, or they may recombine with ions which had previously lost an electron. A recombination may result in the emission of a photon, and it is these photons that are the luminescence, Figure 3(iii)). The number of photons can be counted, and that count is used as the measure of the intensity of the luminescence. In simple terms the more trapped electrons released, the more intense the associated luminescence. Luminescence consequent upon the application of heat to the mineral lattice is known as thermoluminescence (TL), while that resulting from the exposure of mineral grains to light is referred to as either optically-stimulated luminescence (OSL) or photo-stimulated luminescence (PSL).

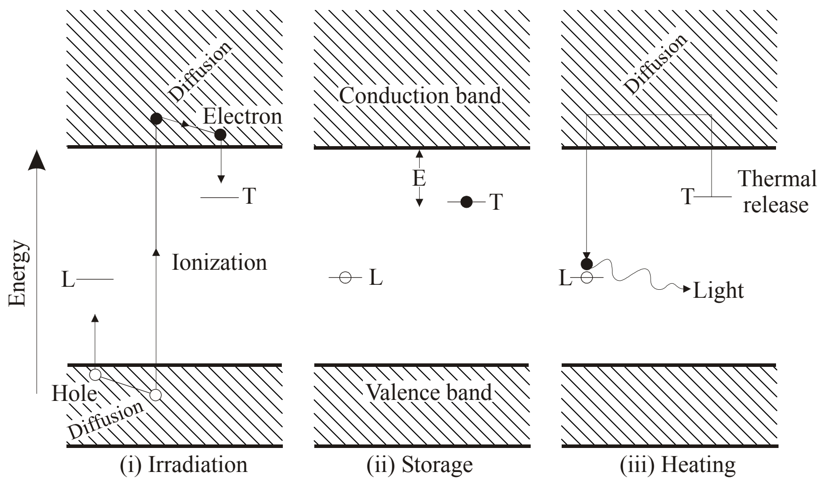


Figure 3: Energy-level diagram representing the processes associated with thermoluminescence. In (i) the action of ionizing radiation to detach an electron from the parent nucleus is represented, with the subsequent trapping of the electron in a trap (T). In (ii) the trapped electron is stored in the trap. E represents the depth of the trap below the conduction band, that depth effectively representing the amount of energy required to free the electron from that trap. In (iii) the input of energy (in this case thermal, but it could equally be light energy) releases the electron from

the trap, It then travels to a luminescent centre (L) where it recombines with a nucleus, in the process emitting a photon (luminescence). Redrawn from Aitken, (1985).

There is usually a range of 'traps' present in the crystal lattice of a mineral grain. These traps can be described as being of differing depths, i.e. different intensities of energy are required to 'shake' trapped electrons out of them. Some traps may be emptied by thermal energy but not by light, while others may be emptied only by light energy and some are emptied by both of them. The efficiency with which the different stimulating energies empty traps also varies. In general traps associated with TL are less easily emptied by exposure to sunlight than are traps associated with OSL. This makes OSL the preferred option for dating many sedimentary accumulations, as it is more likely that the traps associated with OSL will have been completely emptied of electrons by the relatively brief exposure to sunlight that some grains will experience during transport prior to deposition. In contrast, TL is often associated with traps that are hard to empty, and it is more likely that there is residual TL still present in the mineral lattice at the time of deposition. In other words it is easier to 'zero' the OSL clock prior to deposition of sediment than it is to 'zero' the TL clock (Aitken, 1998). If the clock has not been reset prior to deposition, then the number of trapped electrons will start to build up following deposition from a non-zero base, and the Environmental Dose determined at some later date will be greater than the age of the deposit would warrant (Figure 4). Under such circumstances, luminescence techniques will give erroneous dates (exaggerating the age of the deposit).

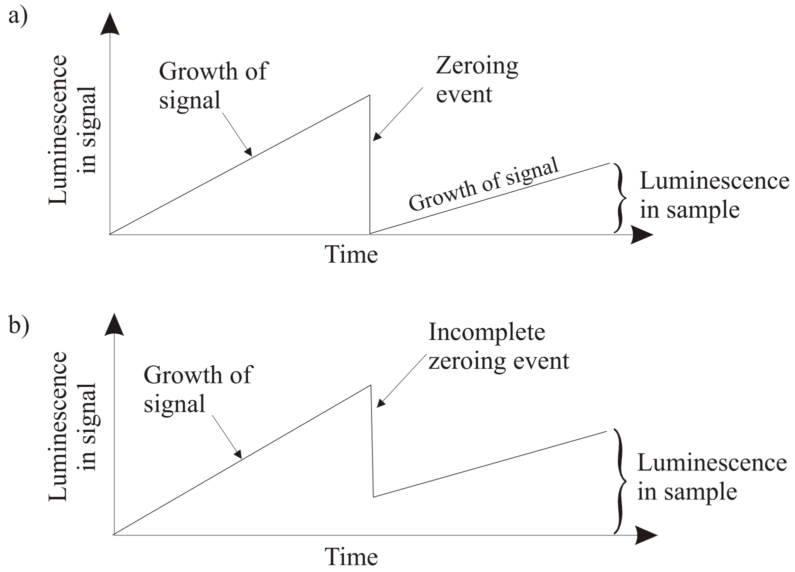


Figure 4: a) The growth of the luminescence signal over time may be interrupted by a zeroing event (exposure to heat or light) which is sufficiently intense to empty all the electron traps and reset the luminescence clock. Subsequent to that luminescence again grows as electrons become trapped. A sample collected now will contain a luminescence signal, the size of which when divided by the rate of growth of that signal will give the time that has elapsed since the zeroing event.

b) The same scenario as in a), save that the zeroing event is incomplete. The rate of signal growth subsequent to this event is the same as in the event portrayed in a), but the luminescence signal now is greater than is the case in a). In consequence dividing the size of signal by the rate of growth will give an inaccurate age for the partial zeroing event.

1.3.2 Limitations and problems

Although luminescence dating techniques have been successfully applied to sediments from a wide range of depositional environments, the use of the technique with glacial and glacially-derived sediments has met with rather more mixed fortunes. This is largely a consequence of the fact that ice is an effective screen preventing sunlight from reaching sub-glacial sediments with the consequence that many of these, particularly lodgement tills, have not been zeroed and still retain the geological luminescence signal on deposition (Huntley *et al*., 1983; Berger, 1984; Gemmell, 1988; Forman & Ennis, 1992). Such sediment is also a major component of the sediment load of glacial meltwater streams, with substantial quantities of fine-grained sediment of this type finding their way into glacial lakes or into fiords into which glaciers are calving or which have glacial termini close to sea level. Although suspended sediment in such water bodies is exposed to sunlight, the degree of zeroing experienced prior to deposition is reduced compared to that experienced by sediment lying on the ground surface. Even in clear water, the spectrum and intensity of light rapidly attenuate with depth (Jerlov, 1970), though Rendell *et al*., (1994) found that quartz grains could still be zeroed in terms of OSL at depths of 12-14m by as little as 3 hours daylight exposure, though the TL signal only reduced by <50% under the same circumstances. The presence of suspended sediment in the water is a further factor influencing the rate of zeroing of both the TL and the OSL signal (Ditlefson, 1992), there being a clear negative correlation between the concentration of suspended sediment and the speed with which the luminescence clock in the sediment is reset (Gemmell, 1985). Berger (1990) demonstrated that for suspended sediment concentrations in the range 15-21 mg l^{-1} the light penetrating to 7m below the surface of a lake was of an intensity 5 orders of magnitude less than that at the lake surface (Figure 5). Ditlefson, (1992) estimated that OSL signals were zeroed only very slowly at high concentrations of suspended sediment, while Duller *et al*., (1995) concluded that '…luminescence dating of……fluvial deposits within a few kilometres of the front of an ice sheet is unlikely to be successful'. Gemmell, (1997, 1999) estimated that complete zeroing of suspended sediment could occur in Alpine environments within 3 km of the ice front, but only for sediment transported during daylight hours. In the case of sediments transported during the hours of darkness, there was no zeroing detectable (Figure 6). This suggests strongly that glaciofluvial and glaciolacustrine sediments are likely to consist of a mixture of grains only some of which are fully zeroed at the time of deposition.

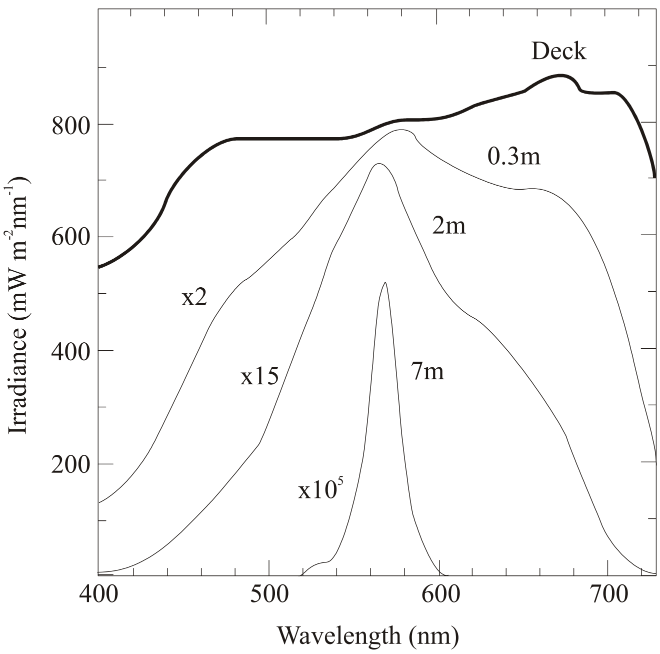


Figure 5: Impact of water depth on the spectrum of light in water containing a suspended sediment concentration of 15-21 mg l-1. Redrawn from Berger, (1990).

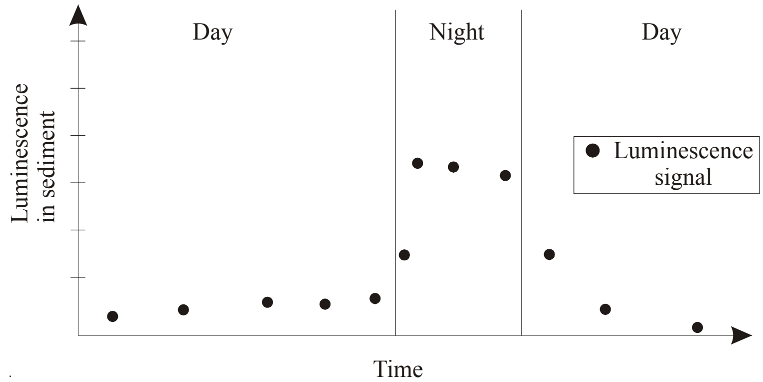


Figure 6: Fluctuations in luminescence signal intensity in suspended sediment being transported by a glacial meltwater stream in the Italian Alps. Note the sediment sampled during daylight hours is well zeroed, but samples collected during the hours of darkness contain substantial amounts of residual luminescence. Redrawn from Gemmell, (1999).

Recent developments in luminescence techniques have enabled the zeroing problem discussed above to be at least partially overcome. In particular, the use of inter-aliquot scatter analysis (Li, 1994; Clarke, 1996; Clarke et al., 1999) enables estimates of the degree of bleaching (zeroing) of the sediment prior to deposition to be made, so that age estimates derived from un-zeroed or only partially zeroed sediments can be identified and rejected as true dates.

2 Sample sites and sedimentology

The two sites chosen for application of the luminescence dating technique are Gleann Einich and the Lairig Ghru, both north-trending valleys on the northern side of the western Cairngorm Mountains (Figure 7). These valleys both exhibit well-formed and accessible former glaciolacustrine delta terraces, with natural sections exposing the sediments, and have been the focus of earlier work by Brazier, *et al.*, (1998), and Golledge, (2002). The sites are important because the former deltas are located between moraines of the Glenmore Lobe at the northern end of the valleys, and moraines of valley glaciers sourced from Cairngorm plateau ice fields in the south, and can therefore be used to constrain the timing of the retreat of the two different ice masses.

Both sites expose sediments of five main facies types, labelled A – E by Brazier *et al.*, (1998). The nomenclature is maintained here to aid correlation of results.

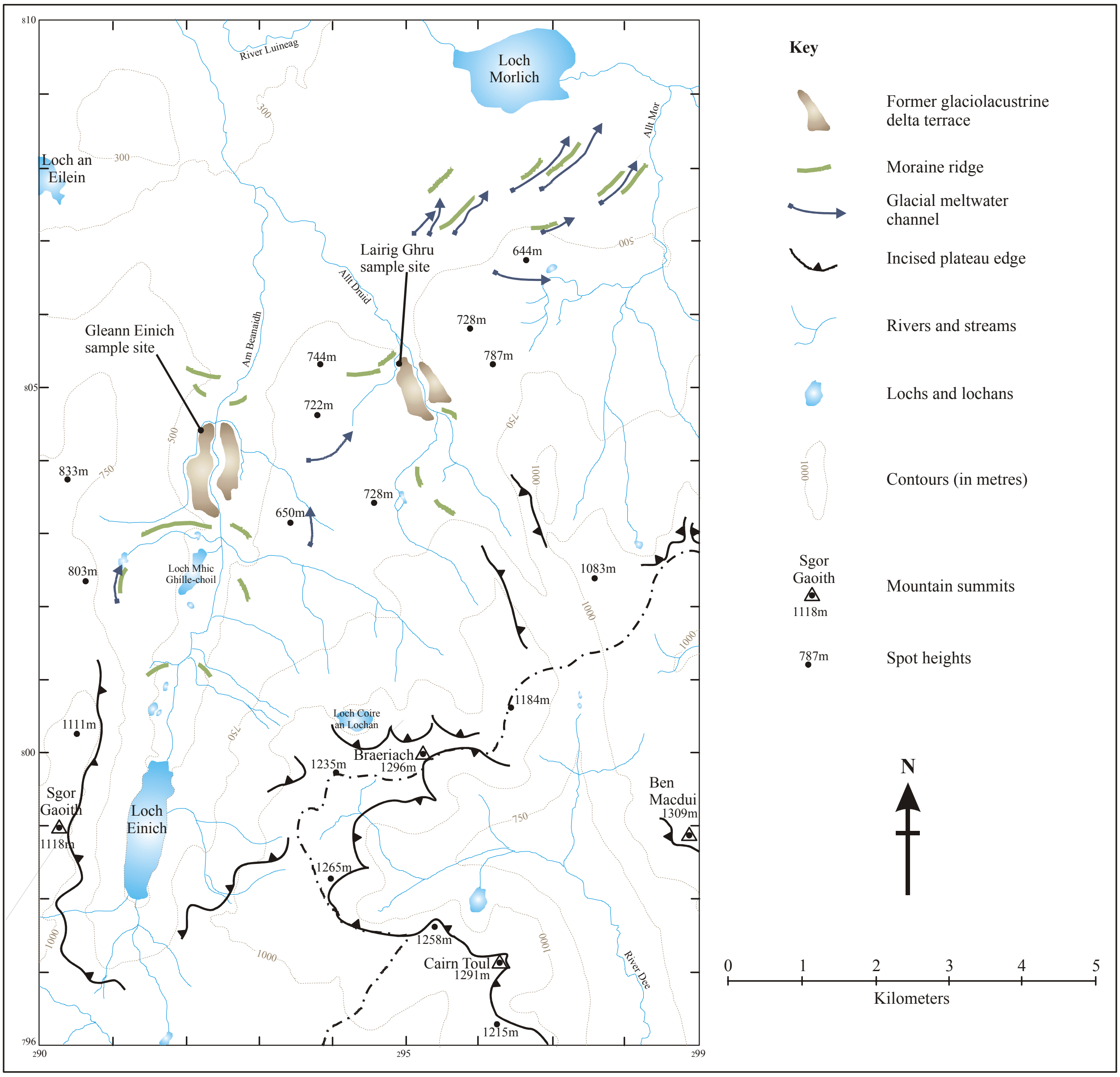


Figure 7: Location and geomorphological context of the sample sites

2.1 GLEANN EINICH

The 25 m high section sampled in Gleann Einich at [NH 923 044] is cut through a sediment sequence (Figure 8) deposited at a distal location from the former ice front, which was located some 1 – 1.5 km to the south. The sequence grades from laminated silt and fine sand (Facies B) upwards into coarser, more thickly laminated and bedded sand (Facies C), and then into cross-bedded sand and fine gravel (Facies D). The top of the section is not well exposed but appears to consist of a poorly sorted cobble gravel unit (Facies E) (Table 1; Figure 10).

Sampling at this locality was restricted to Facies B and C, as these were considered to comprise the most suitable grain sizes for luminescence dating. Two small sections to the north of the main section were used for sampling of sediments lowest in the sequence, as these were poorly exposed at the main section (Figure 10).

Figure 8: The distal glacio-deltaic sequence in Gleann Einich at [NH 923 044].

2.2 LAIRIG GHRU

The 20 m high section at [NH 949 054] in the Lairig Ghru (Figure 9) is also cut through a distal sediment sequence with respect to the former ice front to the south. However, in contrast to the situation in Gleann Einich, input of meltwater to the Lairig Ghru lake occurred from the valley glacier, the margin of the Glenmore ice, and also from a meltwater stream flowing over the col to the west, from the ice occupying Gleann Einich.

The section exposes four of the five facies types identified by Brazier *et al.*, (1998). The lowest unit is a poorly stratified matrix-supported diamicton (Facies A) on top of which are the laminated fine sediments of Facies B, and the coarser material of Facies C. Facies D is not present, though the cobble gravel of Facies E is (Figure 10).

Figure 9: The sediment sequence exposed in the Lairig Ghru at [NH 949 054]

2.3 SUMMARY OF FACIES SAMPLED

Table 1 (below) describes the sedimentology of the sampled Facies B and C, and provides a generalised interpretation of their origins and depositional environment.

FACIES	MIALL FACIES CODE	SEDIMENTOLOGY	INTERPRETATION
B	Fl, Flv	Horizontally-laminated, normally graded couplets of silt and fine sand with graded silt and fine sand microlaminae and rare, persistent, thin (1-3mm) clay layers. Laminae become coarser and thicker up-section, with increase in cm-scale normal faulting and current action. Type B climbing ripples (Ashley *et al* ., 1982) occur infrequently towards the top. Facies B grades upwards into Facies C. Occasional dropstones are present.	Lake floor turbidity currents. Decreasing flow velocity results in laminated silty, fine sand grading into fine silt. Lacks thin and laterally persistent clay layers. Rapid sedimentation by underflows from melt streams. Bed deposition taking hours only. Only very low in the facies, where laminae drape over underlying coarse sediment is there any case to be made for annual sedimentation. (See **sample 7 LG).**
C	Suc, Sl, Sr, Fl	Bedded fine sand coarsening-up to medium and coarse sand, though laminated lower in the unit, with type B climbing ripples (stoss side preservation). Upward thickening of beds with planar cross-bedded sands, and fine gravel up to 0.5m thick becoming more abundant, normally lying above sharp, convoluted loaded contacts. Thin, but internally massive granule beds also increase in frequency up section. Ripples occur more commonly up section and parallel laminae are rare.	Lower delta slope, increasing proximity to delta front. Type B sand ripples grading into draped silt laminae indicate waning flow deposition following passage of turbid underflows. Appearance of subsidiary planar cross-bedded sands indicates migration of small dunes and trend to more energetic flow and sediment delivery. Thin beds of coarse sand/granules due to greater mobility of the granule fraction of a poorly sorted underflow. Loaded contacts show rapid deposition prior to dewatering.

Table 1: Sedimentology of the two facies types sampled in both Gleann Einich and the Lairig Ghru, modified from Brazier *et al.*, (1998).

2.4 SECTION LOGS

Figure 10 (below) shows vertical lithological logs for the sections in Gleann Einich and the Lairig Ghru, including the two small exposures north of the main section in Gleann Einich. Sample numbers indicate the approximate level at which the samples were taken, and from which facies type.

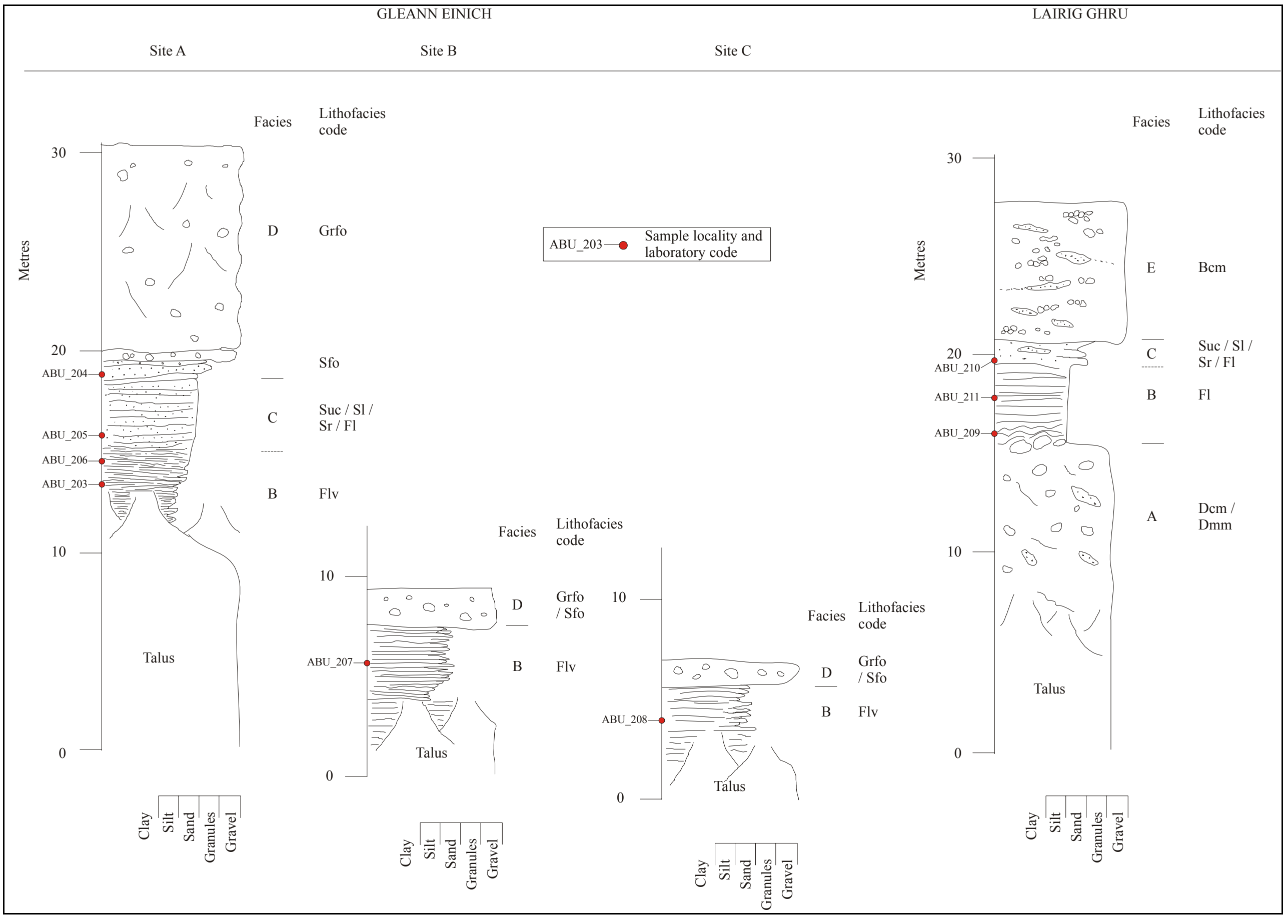


Figure 10: Vertical lithological sections showing sample horizons at the two localities

3 Methodology

3.1 FIELD SAMPLING

A two-day visit was made to the field area between the 6th and 7th August 2001, in order to sample the glaciolacustrine sediments and to take *in situ* gamma ray measurements. A total of nine samples were collected, six from Gleann Einich (ABU_203 – ABU_208) and three from the Lairig Ghru (ABU_209 – ABU_211). The samples were obtained by hammering 400mm lengths of 65mm diameter black plastic tubing horizontally into the exposed and 'cleaned' sediments. Each tube was then sealed with duct tape at the exposed end and labelled before the sediment was dug from around the tube to allow its removal from the face. The other end of the tube was sealed in the same way. A MicroNomad portable gamma spectrometer was then placed into the hole left by the tube to obtain *in situ* environmental dosimetry data (Figure 11).

Figure 11: Obtaining *in situ* dosimetry data from the Gleann Einich sample site, using the MicroNomad portable gamma spectrometer.

3.2 SUMMARY OF SAMPLE DATA

Table 2 (below) summarises the location, facies type and local lithological characteristics of each of the nine samples taken.

Field sample number	Laboratory Code	Height (m) above base of section	Facies	Sampling notes
1 GE-A	**ABU_203**	10.3	B	Predominantly fine silt - silt laminae, with very thin sandy partings. Sample site deformed and dewatered during extraction.
2 GE-A	**ABU_204**	16.2	C	Sample taken from upper end of facies C, in transitional deposits adjacent to Facies D. Sample taken from unconsolidated sand bed.
3 GE-A	**ABU_205**	13.2	C	Medium to coarse grained sands, near dispersed boundary with underlying Facies B.
4 GE-A	**ABU_206**	11.3	B	Horizontal inter-bedded silts and sands. Silts less abundant than in sample 1.
5 GE-B	**ABU_207**	5.8	B	Sample taken from equivalent height of 1.2m in adjacent section A. Fine silt laminae with significant clay content. Silts draped over dropstones (granite). Sample site deformed and dewatered during sample extraction.
6 GE-C	**ABU_208**	4.4	B	Sample taken from silt bed in repeated fining upward sequences of silts and fine sands. Pipe hit an obstruction (possibly a dropstone), so that sampling was shallow in the section face at this site.
7 LG	**ABU_209**	20	B	Sample taken from base of Facies B, just above contact with underlying facies A. Silts draped over underlying cobbles, which may indicate more gradual less energetic rain out of fines, than found typically in Facies B. Deformation and dewatering of the sample site during extraction, including bulging. Outer end of sample liquefied and came out of tube.
8 LG	**ABU_210**	24.5	C	Sample taken in a sandy layer. Adjacent deposits show signs of faulting, deformation (eg of parasitic folding). 1m above site overlain by Facies E. The sample site is interpreted as high energy, but probably in shallow water. Deformation occurring with burial.

9 LG	**ABU_211**	22.5	B	Sampled from predominantly silt rich layer. Site below transitional boundary with sandier Facies C. Other structures indicate this was deposited in a higher energy environment than sample 7, but quieter than 8.

Table 2: Relative sample locations and lithological characterisitics

3.3 LABORATORY PREPARATION

In the laboratory, samples were initially weighed and dried to determine the field moisture content. Subsequent to this the samples were treated with 10% HCl to eliminate any carbonates present, and then with 30% H_2O_2 to remove organic matter. The remaining sediment was then sieved, and appropriate fractions reserved for luminescence analysis. Where suitable material was present, 90µ-125µ, and 125µ-180µ quartz grains are extracted from the sample by flotation in sodium polytungstate at a density of 2.62gm cm-3, followed by etching in 10% HF for 40 minutes to dissolve any remaining feldspar grains and remove the surface layers of the quartz grains. A further flotation at a density of 2.74 gm cm-3 was used to separate any residual heavy minerals from the quartz grains. The quartz grains were then washed and dried before being mounted onto stainless steel discs using silicone oil.

In samples where quartz grains were in short supply, feldspar grains were separated by flotation in sodium polytungstate at a density of 2.58 gm cm-3. These grains were not etched in HF.

If a sample is taken from silt or clay horizons, the initial procedures are the same as detailed above for coarser sediments. Once the material has been sieved, however, the 4-11µ fraction is isolated using Stokes' settling in acetone, and is precipitated onto discs for luminescence analysis.

4 Results

4.1 DOSIMETRY

The Environmental Dose Rate for each of the samples analysed has been calculated using a combination of thick source alpha counting (TSAC), together with flame photometry, to determine the uranium, thorium and potassium content of each sample. The results have been combined with those obtained by portable gamma spectrometry to calculate the Environmental Dose Rate for each sample. The cosmic radiation component of the Environmental Dose Rate has been calculated following the method of Prescott and Hutton (1994). Details of the dose rates are presented in Table 3.

Sample	**K(%)**[a]	**Dosimetric Components** *(mGy a^{-1})*				**Water content (% dry wt.)**	**Dose rate (mGy a^{-1})**
		α-dose[b]	**β-dose**[b]	**γ-dose**[c]	**Cosmic dose**[d]		
ABU_203	3.13	5.0±0.7	3.52±0.19	2.7±0.01	0.038±0.002	3.13±0.31	11.81±0.73
ABU_204	2.96	4.3±0.81	5.56±0.29	2.0±0.01	0.064±0.003	11.24±3.0	12.53±0.88
ABU_205	2.96	5.6±1.37	5.74±0.29	2.6±0.01	0.049±0.002	19.35±5.0	14.53±1.41
ABU_206	3.33	6.5±0.91	4.58±0.17	4.1±0.01	0.041±0.002	19.02±3.0	15.13±0.93
ABU_207	3.4	6.6±1.48	6.12±0.26	3.2±0.02	0.033±0.002	29.28±5.0	16.62±1.52
ABU_208	3.23	5.1±0.75	3.65±0.19	2.6±0.01	0.16±0.008	27.5±5.0	12.09±0.78

Sample	K(%)[a]	Dosimetric Components *(mGy a^{-1})*				Water content (% dry wt.)	Dose rate (mGy a^{-1})
		α-dose[b]	β-dose[b]	γ-dose[c]	Cosmic dose[d]		
ABU_209	3.09	4.9±1.22	3.46±0.19	2.5±0.01	0.063±0.03	28.74±5.0	11.54±1.24
ABU_210	3.35	3.7±0.59	5.74±0.26	2.4±0.01	0.078±0.004	12.24±3.0	12.56±0.67
ABU_211	3.25	3.7±0.61	4.97±0.26	2.4±0.1	0.093±0.005	25.07±5.0	11.69±0.69

[a] Determined by flame photometry. Errors assumed to be ± 10% of K value

[b] Determined by thick-source alpha counting

[c] In-situ measurement using portable gamma spectrometer

[d] Calculated following Prescott and Hutton (1994)

Table 3: Dosimetry data

4.2 LUMINESCENCE ANALYSIS

Samples were analysed using a Risø luminescence reader (model TL/OSL-DA-15). This apparatus has the facility to stimulate luminescence by use of an infra-red laser at 830nm, or by blue light-emitting diodes, which stimulate the sample at 470nm. Infra-red stimulation is used to generate a signal from feldspar grains, but not from quartz. Indeed it can be used as a quick test of the purity of a quartz separation. The blue light is used primarily in quartz dating, though it does produce a response from feldspar as well. The luminescence emitted from the samples upon stimulation was passed through two Hoya U-340 filters, which screen out unwanted components of the light spectrum. All measurements were made at a temperature of 125^{o}C.

The principle of SAR protocols is that the natural luminescence signal of a series of aliquots is first measured, by infra-red or blue light stimulation. This stimulation is maintained for a period of time (usually 100 seconds) long enough to release all the trapped electrons in the mineral grains in each sample (Figure 12).

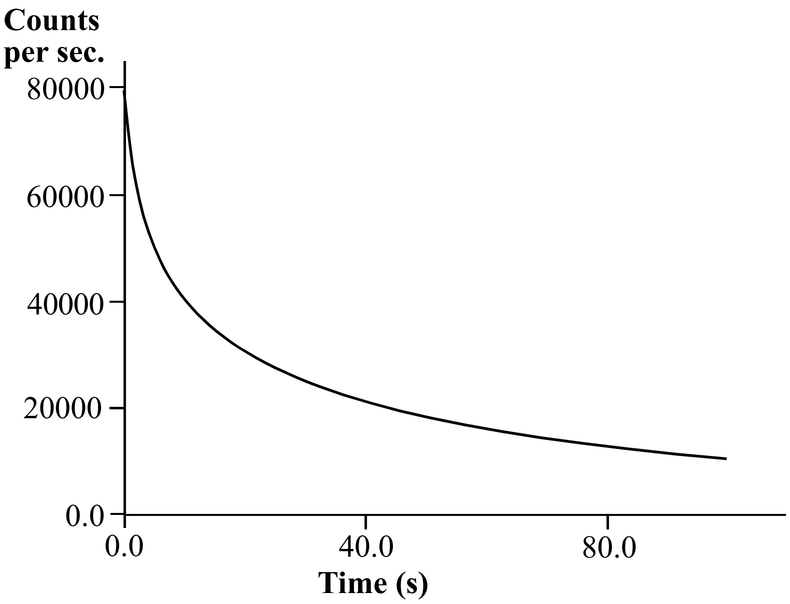


Figure 12: Decay of natural luminescence during infra-red or blue-light stimulation.

Signals are then regenerated by irradiation using a beta source present within the apparatus and which delivers a dose of approximately 0.04 Gy sec^{-1} to the samples. Following irradiation, the luminescence is again measured by stimulation in the same way as for the natural signal. This process is repeated for each aliquot using a range of irradiation doses designed to bracket the natural dose (Environmental Dose), and the Environmental Dose is then determined by interpolation (see Figure 13). Sensitivity changes between measurements were measured by assessing the luminescence reaction to a standard test dose, this reaction being used to normalise the regenerated luminescence signals.

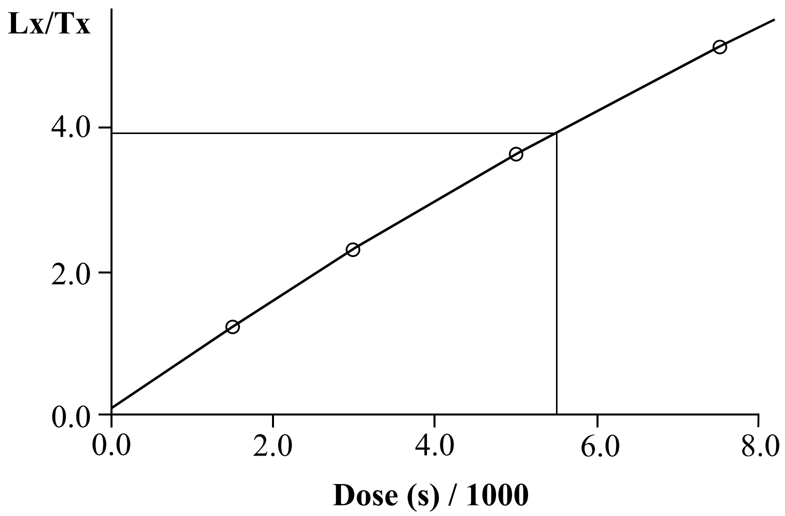


Figure 13: Luminescence growth curve for a Single Aliquot Regeneration analysis. The horizontal line marks the luminescence intensity of the natural sediment, while the points on the growth curve have been generated by successive laboratory irradiations. The point where the growth curve intersects the natural luminescence intensity allows the amount of irradiation needed to produce the natural signal (the Environmental Dose) to be determined.

The first sample to be analyzed (lab. code ABU_203) yielded predominantly fine-grained sediment. Accordingly the 4-11μ fraction was isolated by suspension in acetone, mounted on discs, and then analyzed following a Single Aliquot Regeneration (SAR) protocol in which polymineralic samples are first exposed to infra-red laser stimulation at 830nm, to remove the signal from feldspar grains prior to stimulation with blue light (470nm) emitting diodes to generate a quartz signal (Banerjee *et al.*, 2001; Roberts & Wintle, 2001). Prior to measurement, aliquots of the sample were preheated at temperatures ranging from 160°C to 300°C.

Results from the blue-stimulated luminescence show only very limited signs of plateau behaviour, with Environmental Dose appearing to be strongly dependent on the preheat temperature (Figure 14). The mean Environmental Dose corresponding to the limited preheat plateau (236.67 ± 4.816 Gy) has been taken as the correct value here and corresponds to an age of 20.04 ± 1.3 ka.

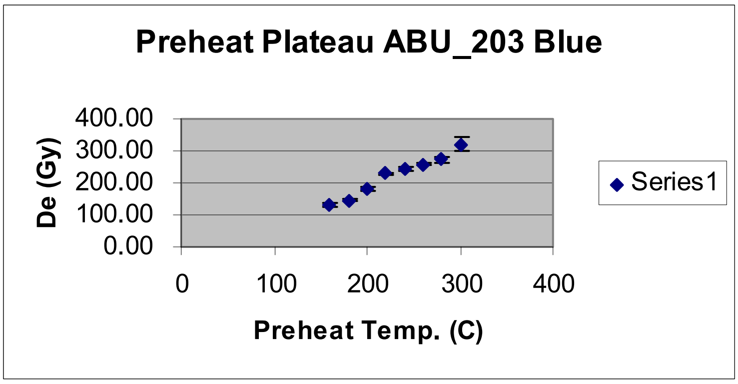


Figure 14: Preheat plateau test for blue-stimulated luminescence for sample ABU_203a.

In contrast, results from the infra-red stimulated luminescence show clear evidence of an extensive plateau (Figure 15) corresponding to an Environmental Dose of 199.13 ± 1.079Gy. This corresponds to an age of 16.86 ± 1.04 ka.

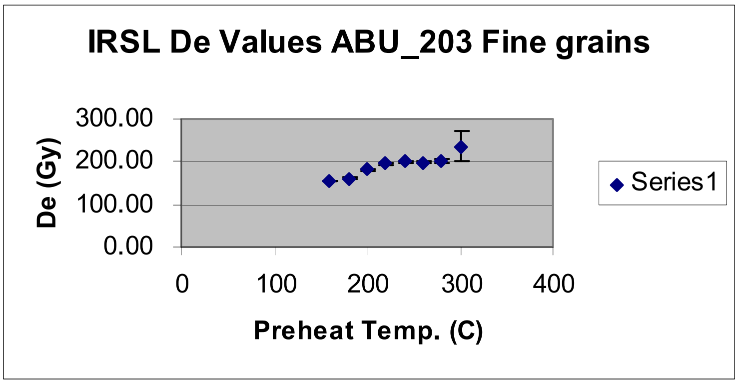


Figure 15: Preheat plateau test for IR-stimulated luminescence for sample ABU_203b.

It should be noted that the IR signal from feldspars is prone to fading – that is, the loss of stored luminescence at ambient temperatures (Banerjee *et al*., 2001). This means that the IRSL ages presented in Table 4 may be based on an underestimate of the Environmental Dose, and hence of the age of the sample. The apparent discrepancy between the dates obtained using the two techniques may be attributable to this mechanism. The dates are greater than might have been expected given present knowledge of the stratigraphic relationships of the sampled horizon (Brazier *et al*., 1998), and it would appear likely that in this case only partial bleaching has been accomplished prior to deposition.

Data for all of the samples analysed are shown below.

Sample	**Particle size**	**Stimulus**	**Plateau (^{o}C)**	**Environmental Dose (Gy)**	**Dose rate (mGy/a)**	**Age (ka)**
ABU_203	4-11µ	Blue diodes	220-240	236.67 ± 4.816	11.81 ± 0.73	**20.04 ± 1.3**
	4-11µ	IR laser	220-300	199.13 ± 1.079	11.81 ± 0.73	**16.86 ± 1.04**
ABU_204	90-125µ	IR laser	160-220	192.75 ± 21.147	12.525 ± 0.88	**15.39 ± 2.00**
	125-180µ	IR laser	180-220	168.65 ± 19.65	12.438 ± 0.88	**13.56 ± 1.85**
ABU_205	125-180µ	Blue diodes	240-260	220.6 ± 25.69	14.53 ± 1.41	**15.18 ± 2.3**
ABU_206	4-11µ	Blue diodes	240-300	354.38 ± 37.25	15.13 ± 0.93	**23.43 ± 2.85**
	4-11µ	IR laser	260-300	313.18 ± 50.13	15.13 ± 0.93	20.7 ± 3.55
ABU_207	125-180µ	Blue diodes	220-280	199.87 ± 45.21	16.619 ± 1.52	12.03 ± 2.93
ABU_208	4-11µ	Blue diodes	240-280	399.14 ± 46.13	12.09 ± 0.78	**33.01 ± 4.37**
ABU_209a	4-11µ	Blue diodes	260-280	299.68 ± 16.4	11.54 ± 1.24	**25.96 ± 3.13**
ABU_209b	4-11µ	IR laser	240-300	251.87 ± 15.27	11.54 ± 1.24	**21.82 ± 2.69**
ABU_210	90-125µ	Blue diodes	No plateau	468.01 ± 139.17	12.56 ± 0.67	37.27 ± 11.26
ABU_211	90-125µ	Blue diodes	240-260	239.59 ± 36.09	11.69 ± 0.69	20.49 ± 3.32

Table 4: Luminescence data for each of the nine samples. Plateau ranges as described above and as shown in Figs. 14 & 15. Emboldened ages are those deemed to be most reliable, according to the Quality Assessment test (Table 5).

5 Discussion

5.1 'QUALITY CONTROL' OF DATA

A number of factors need to be taken into consideration in order to assess the reliability of the data, such as the sediment type, (grain size, mineralogy), the assumed transport path the sediment has followed prior to deposition, and the luminescence technique used to derive a date.

Problems such as fading of feldspar signals (see section 4.2) are but one of the considerations that need to be taken into account when evaluating dates obtained using luminescence analysis. To this end, Clarke (1996) and Clarke *et al.* (1999) compiled a series of tests to which data from samples were subjected with the objective of assessing the likely accuracy of each date. These tests have been applied to the dates obtained in the present study, and the results are presented in Table 5. There are, however, some caveats that should be borne in mind. Clarke *et al.* (1999) performed all their work on samples of sand-sized material, and used infra-red stimulation of the feldspar grains as their dating protocol. The present study uses this approach only for sample ABU_204, as quartz is generally considered to give a more stable signal and one that is less prone to loss of stored signal at ambient temperatures. In the case of ABU_204 difficulty was experienced in obtaining a suitable quantity of quartz grains from the sample, and so feldspar grains were used instead. The test results for ABU_204 show it to have contained unbleached or poorly-bleached grains at the time of deposition. These variations in the residual levels of luminescence at the time of deposition are now equivalent to only a very small percentage of the total luminescence, and will not result in an erroneous determination of the age of deposition.

The fact that ABU_204 is a sample taken from Facies C (Table 5) is of interest. Facies C sediments are deposited quite close to the delta front, and it seems likely that the transport of sediments across the delta would have given them some exposure to sunlight before they travelled down the delta front and became screened from light by the opacity of the lake water. While inspection of Table 5 shows that not every sample of sediment interpreted as being Facies C has been well bleached at deposition, the majority of dates from them are believed to be reliable. Sediments interpreted as being from Facies B, in contrast, show significant problems due to lack of bleaching. The rapid deposition and highly turbid nature of the water in which they were deposited may account for this.

To date, the authors of this report are unaware of any equivalent work on quality assurance that has been done on fine-grained polymineral sediments or on quartz samples, other than the preheat plateau test (see section 4.1). For purposes of comparison, those samples other than ABU_204 have been assessed using the same criteria, and the conclusions included in the table. In all cases the aliquots used for the evaluation are those which plot on the preheat plateau, save for ABU_210 where there was no clear plateau, and the aliquots used are those for which the data points could be fitted with a saturating exponential curve. The conclusions for all the non-feldspar samples with respect to quality of data should be regarded as indicative only.

Table 5. below shows an objective determination of sample quality based on the methods described by Clarke (1996) and Clarke *et al.* (1999), with consideration given to the additional factors described above.

Sample No.	Facies type[a]	De (Gy)[b]	$\sigma_{(n-1)}$[b]	S_N[b]	Age (ka)	Sample Quality[b]
ABU_203a	B	236.67	11.79	0.04	20.04 ± 1.3	**Poorly bleached at deposition, but not a problem for dating**
ABU_203b	B	199.13	4.03	0.02	16.86 ± 1.04	**Well bleached at deposition**
ABU_204a	C	192.78	20.92	0.108	15.39 ± 2.00	**Moderately poorly bleached at deposition, but no longer a significant problem**
ABU_204b	C	168.6	19.46	0.11	13.56 ± 1.85	**Moderately poorly bleached at deposition, but no longer a significant problem**
ABU_205	C	222.06	25.67	0.115	15.18 ± 2.3	**Moderately poorly bleached at deposition, but no longer a significant problem**
ABU_206a	B	354.38	37.24	0.105	23.42 ± 2.85	**Moderately poorly bleached at deposition, but no longer a significant problem**
ABU_206b	B	313.18	49.86	0.159	20.7 ± 3.55	Poorly bleached at deposition - problem for dating
ABU_207	B	199.74	45.01	0.22	12.03 ± 2.93	Poorly bleached at deposition - problem for dating
ABU_208	B	399.14	50.28	0.13	33.01 ± 4.37	**Moderately poorly bleached at deposition - no longer a significant problem for dating**
ABU_209a	B	299.68	15.70	0.05	25.96 ± 3.13	**Poorly bleached at deposition, but no longer a problem for dating**
ABU_209b	B	251.87	14.73	0.06	21.82 ± 2.69	**Poorly bleached at deposition, but no longer a problem for dating**
ABU_210	C	465.01	138.9	0.29	37.27 ± 11.26	Poorly bleached at deposition - problem for dating
ABU_211	B	239.59	37.83	0.16	20.49 ± 3.32	Poorly bleached at deposition - problem for dating

[a] Facies type descriptions as per those detailed in Table 1.

[b] Sample quality derived as per Clarke *et al.* (1999), save that the critical value of S_N that has been used to classify samples are those indicated by Clarke (1996)

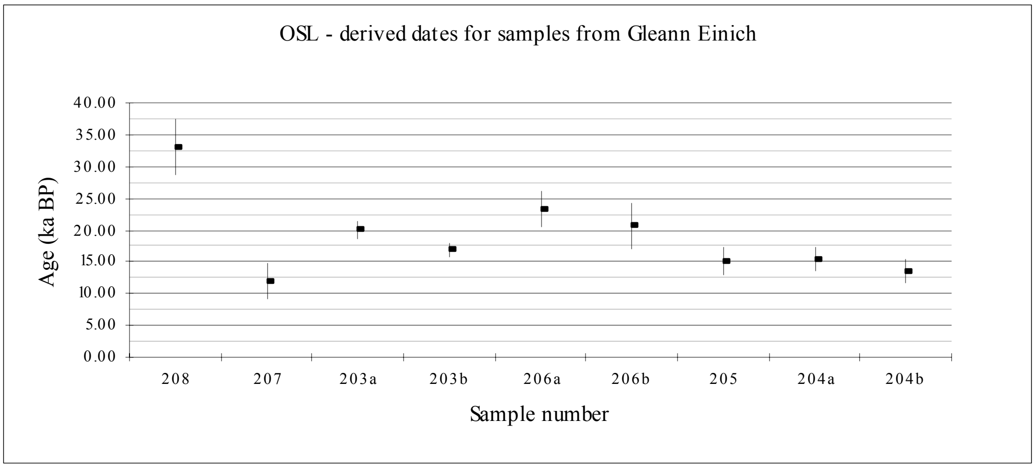
Table 5: Quality assessment of samples and derived data, emboldening as for Table 4.

5.2 SIGNIFICANCE OF RESULTS

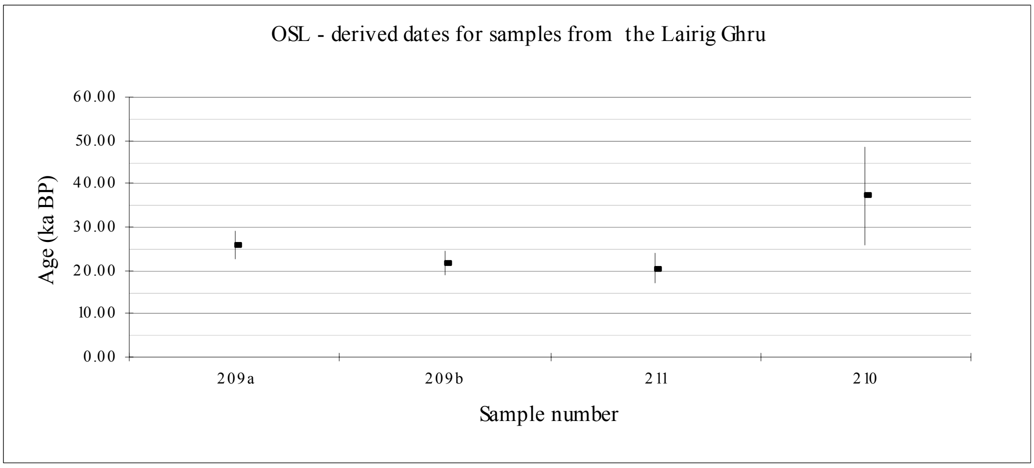
Table 5 illustrates the variability in sample quality, ranging from 'well bleached at deposition' to 'poorly bleached at deposition - problem for dating'. The ages presented and their respective errors reflect the differing degrees of sample quality, those ages with the largest errors and more widely distributed ages appear to be the samples whose reliability is most questionable. Those samples from Facies C are considered the most reliable based on their greater degree of bleaching prior to deposition; the one exception is sample ABU_210 which appears to show no preheat plateau (Table 4), and therefore is unlikely to represent complete zeroing. Where two dates are derived from the same sample using different methods, for example ABU_203 and ABU_204, each case is considered individually. In the case of ABU_203, the larger plateau shown by (b) suggests that this is the more reliable of the two. In ABU_204, it was found that signal fading in the feldspar grains was least in (a), and it is therefore this date that is most reliable.

5.2.1 Overview

The dates derived from the Optically-Stimulated Luminescence experiments on the glaciolacustrine sediments from Gleann Einich and the Lairig Ghru show a total age range from 12.03 (± 2.93) ka to 37.27 (± 11.26) ka (Figure 16a & b).



a)



b)

Figure 16: OSL - derived ages for glaciolacustrine sediments in the Cairngorm Mountains, showing dates and error bars according to Table 5.

In order to enable reasonable analysis of the data, the Quality Control procedures outlined above were employed to determine which samples produced the most reliable dates, and it is these that are most fully discussed below. Given these criteria, (Table 5 and above), it is apparent that the most reliable samples show an age range from 13.56 (± 1.85) ka to 33.01 (± 4.37) ka, Figure 17, (below).

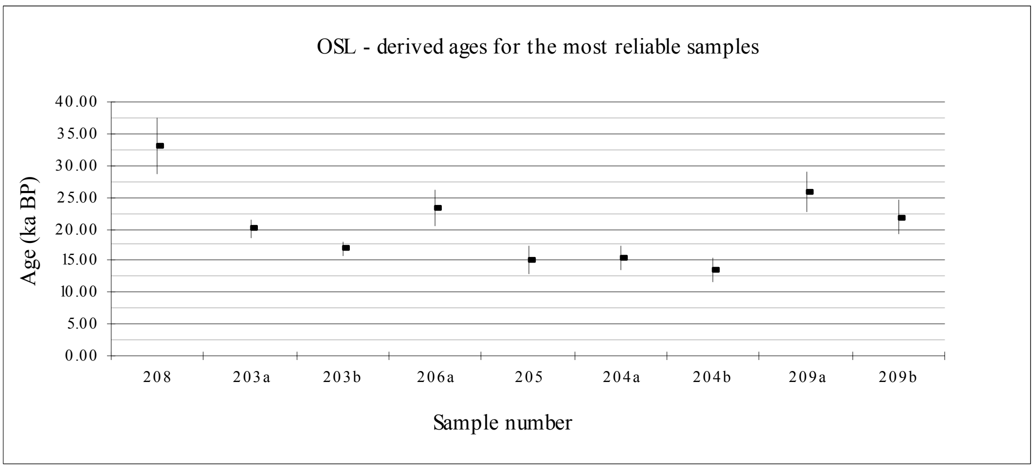
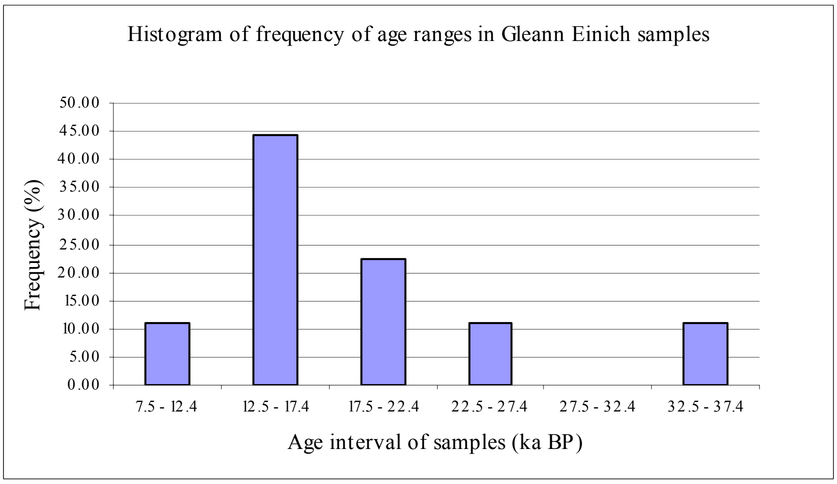


Figure 17: OSL – derived ages for the most reliable of the samples analysed, arranged with the stratigraphically lowest samples in Gleann Einich on the left, and the highest further right. ABU_209a & b are from the Lairig Ghru.

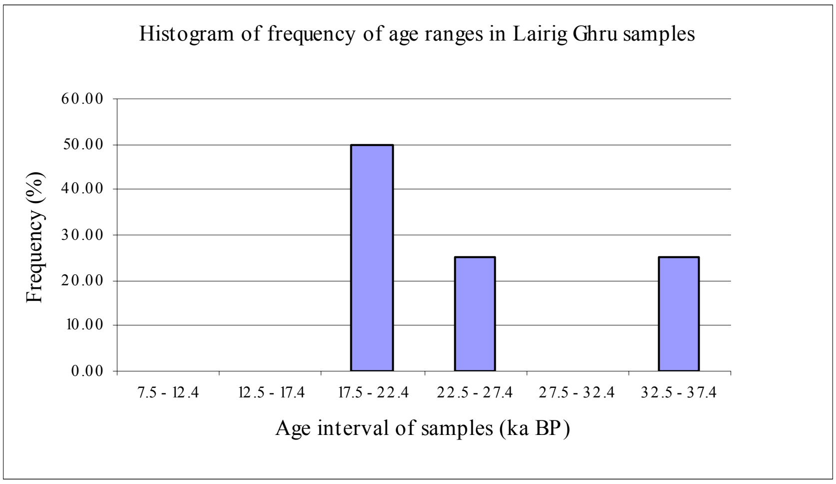
The relationship between the derived dates and their stratigraphic position should theoretically produce a younging-upwards sequence. Figure 17 shows that some of the dates do follow this trend.

5.2.2 Data distribution and weighted means

By plotting histograms of age interval against frequency, the distribution of dates within each dataset can be more clearly seen, (Figure 18a & b).



a)



b)

Figure 18: Frequency histograms showing the percentage distribution of sample ages from a) Gleann Einich and b) the Lairig Ghru.

The large spread of ages within the dataset can be attributed to factors such as the relative degree of bleaching that the material has been subjected too, as discussed above. In order to derive a 'mean' value that reflects these factors accurately, a 'weighted mean' was produced using the method set out by Aitken (1985).

Using Aitken's formula , a weighted mean of 17.48 (± 0.3) ka BP was calculated for the Gleann Einich samples, and 23.07 (± 1.0) ka BP for those from the Lairig Ghru. This value was produced using the entire population of samples, not just those deemed to be significant by the QA test.

5.2.3 Implications for the chronology of 'local' ice movements

5.2.3.1 GLEANN EINICH

The data presented above define the age of the glaciolacustrine sediments in Gleann Einich well, with the most reliable date of 16.86 ± 1.04 ka BP derived from sample ABU_203b. Whilst some of the other data suggest ages significantly older or younger than this, their reliability and overall 'quality' are questionable (Table 5).

The dates obtained for Gleann Einich show within-site differences, and these may be related to their relative time of deposition or to variation in their relative degrees of bleaching. There is a tendency towards younging upward in the samples shown in Figure 17, from the lowest of the five, ABU_208, showing the oldest ages, up to the highest of the five, ABU_204, showing the youngest ages. Although samples ABU_203 and ABU_205 are both stratigraphically and chronologically located between these two, sample 206a appears to deviate substantially from the general trend. Whilst these samples show a total age range of 13.56 (± 1.85) ka to 33.01 (± 4.37) ka, there appears to be a clustering of dates from samples 203b, 204a and 205 between 16.86 (± 1.04) ka and 15.18 (± 2.3) ka BP. The age-interval distribution histogram of Figure 18(a) shows that the greatest proportion of dates from all the samples lie in the 12.5 – 17.4 ka BP age interval, whilst the weighted mean lies around the upper end of this age bracket at around 17.5 ka BP. An age of c. 15 ka – 17 ka BP may therefore best represent the true age of the sediments.

Assuming that the above data are correct, they imply that Gleann Einich was filled with a glacial lake at and around c. 16 ka BP. Preservation of the landform since formation suggests that no later glacier advances occurred in this valley, and that the lake formed during the final stages of Main Late Devensian deglaciation. By implication, the Gleann Einich outlet lobe (fed from the Móine Mhòr plateau icefield), must have retreated from a glacial maximum position of confluence with Strath Spey ice in Glenmore, to the moraines at [NH 923 032] (Figure 7). The size of the moraines and the delta still seen suggest that this 'stillstand' position was occupied for a significant period of time, forming a lake whose lifetime was estimated to be 'several centuries' by Brazier *et al.*, (1998). The data presented here are of insufficient resolution to refine this estimate, or to give any indication of sedimentation rates during the lifetime of the lake.

One further implication of these dates is that the moraine dam at the northern end of the glen must also have been coherent and stable during this time in order to confine the meltwater and form the lake. This may suggest that the ice in Glenmore was still present, as it is assumed that the moraine would have collapsed following removal of its supporting ice, resulting in catastrophic drainage of the lake. In the absence of an assessment of the engineering properties of the moraine material it is impossible to say whether this is the case or not. The 'sediment wedge' on which the moraine is situated is c. 100m thick, and is most likely a composite landform built up by numerous oscillations of the ice margin at this location. The moraines on the top of the landform therefore reflect only the last of the oscillations. This landform evidence lends support to the theory that ice in Strath Spey was active throughout much of its lifetime, and may have responded as part of the readvance of the BIS thought to have taken place coevally or immediately following HE 1 (Bowen *et al.*, 2002) at around 14 ^{14}C ka BP (McCabe *et al.*, 1998) (c. 16.5 – 17 ka BP).

Additionally, recent data derived from cosmogenic isotope analysis by Everest and Kubik (Pers. Comm.), suggest that the confining moraine of the Glenmore lobe has a mean age of c. 16.6 ka BP, whilst the moraine of the Gleann Einich glacier has a mean of c. 15.2 ka BP. These dates show a tantalising degree of overlap with the OSL dates presented here and the radiocarbon dates of other authors above, although further investigation is required before any significant correlations can be made.

5.2.3.2 THE LAIRIG GHRU

Half of the four dates obtained from the Lairig Ghru samples appear to show incomplete zeroing and therefore an over-estimate of their ages. Of the four, ABU_209a and b are the only two to pass the QA test for dating reliability, although even these are considered to be poorly bleached. These samples yield ages of 25.96 (± 3.13) ka BP and 21.82 (± 2.69) ka BP respectively. ABU_211 also shows signs of a preheat plateau and gives an age of 20.49 (± 3.32) ka BP, but is also considered to be poorly bleached, and according to the QA tests may pose a problem for dating. ABU_210 shows no preheat plateau, is poorly bleached and consequently a problem for dating, factors which are reflected in the derived age of 37.27 (± 11.26) ka BP which is clearly anomalous. Calculation of a weighted mean has enabled a date of 23.07 (± 1.0) ka BP to be established.

In light of these considerations, it may be suggested that the lake in the Lairig Ghru formed at or around 22 – 25 ka BP, a period coincident with the accepted Late Devensian glacial maximum (Bowen *et al.*, 2002). This date does, however, lend support to the relative chronology proposed by Brazier *et al.*, (1998), suggesting that the delta in the Lairig Ghru must predate the delta in Gleann Einich. Further work may be required to determine a more precise age for these deposits.

5.2.4 Regional

The geomorphological and sedimentological legacy of the deglaciation of the northern Cairngorms provides evidence of a period of ice dammed lake formation, that occurred as a consequence of both regional (ice sheet) and local (ice field outlet glacier) readvances and or still stands. The distribution of ice marginal landforms found along the northern flank of the Cairngorm Mountains, (Brazier *et al.*, 1996a, b), indicate a widespread series of events during deglaciation. It has already been stated above that two separate and differently-sourced ice masses were occupying different parts of both Gleann Einich and the Lairig Ghru during the lifetime of the lake. The larger Strath Spey ice mass, filling the Glenmore area, was ultimately sourced from the ice accumulation area centred around Rannoch Moor, to the south-west of the study area (Figure 1). As such it would have been largely influenced by the prevailing air masses on the west coast, and perhaps would be expected to show overall mass-balance fluctuations in synchrony with changes in sea-surface temperature and moisture content of the dominant air masses. Through the coupled ocean – atmosphere system described in 1.2.2 above, these changes would have been propagated into terrestrial ice fluctuations almost directly. By contrast, the independent ice masses supported on the Cairngorm plateaux would be distant from the influence of ocean currents. The prevailing winds from the south-west would have brought moisture-bearing air over the western highlands before reaching the Cairngorms, releasing much of their content in the west and producing a 'rain-shadow' effect farther east. Consequently, advances of the western ice masses due to heavy snowfall may not have been matched by ice formation in the more arid, precipitation-starved Cairngorms. An asynchronous response to ocean – atmosphere fluctuations may therefore have occurred across the country, making correlations between individual ice limits in different locations particularly difficult without additional accurate dating.

Elsewhere in Scotland there is evidence for readvances during deglaciation. Clapperton (1997) suggested, for example, that the Cairngorm ice marginal deposits may correspond with the Moray Firth readvance deposits at Lhanbryde. Other examples of readvance features have been found in Deeside, at Dinnet, (older than 12.6 to 11.5 ka ^{14}C BP) and in the lower Dee (Brown, 1993), as well as the Wester Ross moraine (Robinson and Ballantyne, 1979; Ballantyne, 1993) which is thought to have been deposited between 13.5 and 13 ka ^{14}C BP (c. 15.5 – 16 ka BP). Clapperton (1997) speculated that HE 1 caused a final phase of readvances across the last Scottish Ice Sheet, and specifically identified the northern flank of the Cairngorms as one of the localities recording this event. The results of the luminescence dating reported here do not refute this assertion.

In assessing the wider significance of the results, frameworks such as that provided by the GISP2 ice core record (Figure 2) are particularly useful. The pattern of climate change shown by the GISP2 ice core record (Figure 2) indicates that after the Last Glacial Maximum (HE 2) at c. 22 ka BP, gradual warming took place over a period of 2 – 3 ka before a cooling trend was re-established shortly after c. 19 ka BP. This period of cooling lasted a further 3 ka, terminating in Heinrich Event 1 around 16 ka BP. Subsequent rapid warming is thought to have led to the complete collapse of the British Ice Sheet within a few thousand years. Within each period of overall warming or cooling, many smaller oscillations are recorded; some indicating very abrupt changes in temperature over periods of as little as a few hundred years.

This pattern of warming and cooling may be partly reflected in the landform assemblage present in the Cairngorms. Initial warming after the LGM may have led to the retreat of the thin plateau icefields of the Cairngorms, with their outlet valley glaciers decoupling from the larger Strath Spey ice mass in areas such as the Lairig Ghru. Possibly as a result of the precipitation gradient that promoted eastern areas to become ice-free soonest, or maybe due to differing mass balance characteristics, lake formation in the Lairig Ghru occurred whilst the Gleann Einich valley glacier was still confluent with the Strath Spey ice. The Gleann Einich ice decoupled from the Strath Spey ice later and retreated, before readvancing to the moraine-marked terminus at [NH 923 032].

Although the exact chronology is difficult to assess, it is likely that the plateau icefields of the Cairngorms were being starved of precipitation and were unable to respond dynamically throughout much of the period between HE 2 and HE 1. The lifetime of the lakes, estimated at centuries rather than millennia, is too short to fit neatly to any one individual climatic oscillation. However, it seems reasonable to suggest that the lake in the Lairig Ghru may have existed at, or shortly after, the LGM, at around 22 ka BP, whilst the lake in Gleann Einich formed somewhat later, around 16 - 17 ka BP. Significantly, landforms in the Lairig Ghru appear to show evidence of an ice-marginal oscillation subsequent to initial lake formation. This approximate chronology accords with recent data (Bowen *et al.*, 2002), and supports the idea of precipitation starvation leading to glacier recession during a period of wider climatic cooling, followed by a readvance around HE 1 and subsequent deglaciation.

The larger and more stable Strath Spey ice stream may have been less susceptible to short-term climate fluctuations, requiring longer period oscillations to significantly affect its overall mass balance, but is likely to have responded to the readvance at HE 1. Retreat of the Strath Spey ice therefore only took place following this readvance, as a result of rapid climatic amelioration. This supports the scenario presented by Brazier *et al.*, (1998), postulating that the lake in Gleann Einich drained catastrophically when the northern most dam was left unsupported by retreat of the Strath Spey ice, whilst the lake in the Lairig Ghru drained periodically with the Strath Spey ice still present.

Sudden global warming immediately following HE 1 would have forced the polar front to migrate northwards over Scotland, producing increased precipitation that may have augmented the readvance described above. Rapid glacier retreat and widespread collapse of the BIS then ensued. One radiocarbon date in the area, from Loch Etteridge, south of Strath Spey, has been used to suggest that the whole area was ice-free by 13.1 ^{14}C ka BP (Sissons and Walker 1974), equivalent to c. 15.2 – 16.4 ka BP calendar years. Further dated landforms and sediments are required to constrain this chronology more precisely.

6 Conclusions

The sampling of glaciolacustrine sediments from the northern Cairngorms, and their subsequent analysis using Optically-Stimulated Luminescence techniques, has yielded a number of interesting results.

In processing the material obtained from the former glaciolacustrine deltas in Gleann Einich and the Lairig Ghru it has been demonstrated that such sediment is not entirely suitable for the application of OSL techniques, due to the relatively short transport paths along which the sediment travelled. Inadequate bleaching of quartz and feldspar minerals by sunlight during their transport resulted in a number of the samples being of poor reliability, in general, over estimating the age of the deposit. Overall, the samples taken from the Gleann Einich delta proved more reliable than those taken from the Lairig Ghru, most probably because a greater number of samples were taken in Gleann Einich, which allowed a greater chance of finding suitable material.

A sufficient number of the samples were suitable for OSL analysis, however, and these yielded reliable dates in at least four or five cases. Two of the samples, ABU_203 and ABU_204 were subjected to two tests. In each case, the result that showed the greatest reliability was chosen for discussion. Reliability was determined using an established methodology, and enabled the discussion to focus on the samples that were deemed most likely to reflect the true age of the sediments. When plotted on a graph, a number of the reliable dates showed a degree of 'clustering' around c. 16ka BP, which was taken as the best timeframe around which an approximate chronology of deglaciation in the area can be constructed.

The results suggest that lake formation may have begun in the Lairig Ghru soon after the Last Glacial Maximum whilst the lake in Gleann Einich formed later. The former seems to have existed prior to the climax of the cooling trend at around 16.9 ka BP (HE 1) shown by ice-rafted debris in North Atlantic sediment cores (Bowen *et al.* 2002), whilst that in Gleann Einich may have formed coevally with, or directly after this event.

Climatic forcing may have been of key importance in the onset of initial deglaciation in the Cairngorms, whilst the effect of precipitation starvation in the Central Highlands during the Late Devensian prevented significant build-up of ice leading up to HE 1. However, the manner of retreat of the larger Strath Spey ice stream remains poorly understood, though it was probably characterised by numerous oscillations and readvances. Final deglaciation perhaps culminated in rapid and widespread retreat, resulting in the deposition of the large quantities of glaciofluvial sand and gravel identified by earlier workers in the low ground of Strath Spey and the Glenmore basin.

Further work is necessary in order to more precisely constrain the pattern of deglaciation in the Cairngorm Mountains, and to establish whether climatic oscillations defined in ice core records such as GISP2 can be identified in the landform assemblages present today. Luminescence techniques have proved to be highly valuable in providing a time frame for deglaciation in a formerly glaciated Scottish highland terrain. However, it has been found that bleaching of sediments prior to deposition is highly variable, and that numerous samples need to be taken in order to increase the chances of getting a few 'reliable' results. The application of the quality assessment procedures was essential in determining the relative importance to attach to the widely ranging results, and enabled an objective analysis of the data to be made.

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