Late Devensian deglaciation of the Tyne Gap Palaeo-Ice Stream, northern England

STEPHEN J. LIVINGSTONE,1* DAVID H. ROBERTS,2 BETHAN J. DAVIES,3 DAVID J. A. EVANS,2 COLM O COFAIGH4 and DELIA M. GHEORGHIU4
1Department of Geography, University of Sheffield, Sheffield S10 2TN, UK
2Department of Geography, Durham University, Durham, UK
3Department of Geography, Royal Holloway, University of London, Egham, Surrey, UK
4NERC Cosmogenic Isotope Facility, Scottish Enterprise Technology Park, Glasgow, UK

ABSTRACT: The deglacial history of the central sector of the last British–Irish Ice Sheet is poorly constrained, particularly along major ice-stream flow paths. The Tyne Gap Palaeo-Ice Stream (TGIS) was a major fast-flow conduit of the British–Irish Ice Sheet during the last glaciation. We reconstruct the pattern and constrain the timing of retreat of this ice stream using cosmogenic radionuclide (10Be) dating of exposed bedrock surfaces, radiocarbon dating of lake cores and geomorphological mapping of deglacial features. Four of the five 10Be samples produced minimum ages between 17.8 and 16.5 ka. These were supplemented by a basal radiocarbon date of 15.7 ± 0.1 cal ka BP, in a core recovered from Talkin Tarn in the Brampton Kame Belt. Our new geochronology indicates progressive retreat of the TGIS from 18.7 to 17.1 ka, and becoming ice free before 16.4–15.7 ka. Initial retreat and decoupling of the TGIS from the North Sea Lobe is recorded by a prominent moraine 10–15 km inland of the present-day coast. This constrains the damming of Glacial Lake Wear to a period before ~18.7–17.1 ka in the area deglaciated by the contraction of the TGIS. We suggest that retreat of the TGIS was part of a regional collapse of ice-dispersal centres between 18 and 16 ka. Copyright © 2015 The Authors. Journal of Quaternary Science Published by John Wiley & Sons Ltd.

KEYWORDS: British–Irish Ice Sheet; cosmogenic surface exposure dating; palaeo-ice stream; radiocarbon dating; Tyne Gap.

Introduction

The investigation of palaeo-ice sheet beds is critical to the assessment of recent and future changes in contemporary ice sheets and for understanding the controls that influence their behaviour. However, despite the wealth of geomorphological and sedimentological evidence on the pattern of ice flow in the central sector of the last British–Irish Ice Sheet (BIIS) (e.g. Livingstone et al., 2008, 2010a, 2012; Evans et al., 2009; Davies et al., 2009, 2012), the timing and rate of its retreat are poorly constrained (Chiverrell and Thomas, 2010; Hughes et al., 2011; Clark et al., 2012). This is significant as the development of a robust geochronology is essential for correlating complex ice-flow phasing across different sectors of the last BIIS and for relating ice-dynamic behaviour to internal and external forcing mechanisms.

The mountainous spine of the central sector of the BIIS (northern Pennines and Southern Uplands) was characterized by high-elevation cold-based plateau ice caps bisected by eastward-draining terrestrial ice streams such as the Forth, Tweed, Tyne and Stainmore (Fig. 1) (e.g. Everet et al., 2005; Davies et al., 2009, 2011; Livingstone et al., 2010a). Upland ice-dispersal centres such as the Cheviots, Southern Uplands, Lake District and Pennines played an important role in seeding and modulating ice flow through these fast-flowing corridors, and thus in controlling the relative dominance of the Irish Sea Ice Stream (ISIS) and North Sea Ice Lobe (NSL), both of which have their onset zones in this region (e.g. Livingstone et al., 2012). This is reflected in the geological record, which indicates complex, multi-phase ice-flow behaviour and repeated marginal fluctuations (Livingstone et al., 2008, 2012; Evans et al., 2009).

Of the four palaeo-ice stream corridors, only the Forth is associated with any direct age control. This is from shells (Nuculana pernula) in a marine unit in the Firth of Forth, which provides a minimum age of 16.2 ± 0.1 cal ka BP for ice stream recession (Hedges et al., 1988). Elsewhere, 10Be surface exposure ages on four Shap granite erratics in the Vale of Eden, immediately to the west of the Stainmore Ice Stream, indicate that the Vale of Eden was deglaciated by ~17 ka BP (Wilson et al., 2013a,b) (Fig. 2). This is coincident with widespread thinning and retreat of ice in the Lake District (Hughes et al., 2011; Wilson and Lord, 2014). In the Solway Lowlands, west of the Tyne Gap, an age of 14.4 ± 0.4 ka BP from a bulk sample in a basal peat unit (Bishop and Coope, 1977) provides a minimum age for ice-free conditions. With the exception of these examples there is limited dating control constraining deglaciation of the central sector of the last BIIS and almost no chronology for the four key east-draining palaeo-ice stream corridors.

This paper focuses on the Tyne Gap Ice Stream (TGIS), which flowed eastwards towards the mouth of the River Tyne through a predominantly bedrock-floored, 15–30-km-wide mountainous pass located between the Cheviot Hills and English Pennines (Figs 1 and 2). The TGIS was fed by ice from dispersal centres in the Lake District and Southern Uplands and was a major tributary of the NSL, which flowed southwards down the east coast of England (Eyles et al., 1994; Davies et al., 2011). The aim of this paper is to reconstruct the pattern and constrain the timing and rate of retreat of the TGIS, using cosmogenic radionuclide dating of exposed bedrock surfaces and radiocarbon dating of lake cores, combined with geomorphological mapping of deglacial features. Furthermore, the paper explores the implications of TGIS retreat for changing BIIS dynamics during deglaciation.

Glacial geomorphological context

The Tyne Gap corridor is heavily streamlined below 400 m asl, with lineations typically <1.5 km long, although forms up to 4 km have been observed (Figs 1, 3 and 4). Lineations are composed of both till and bedrock, including visually striking streamlined bedrock ridges controlled by the geological
structure (Livingstone et al., 2010a; Krabbendam and Bradwell, 2011), particularly evident in the western Tyne Gap region (Fig. 3). This has resulted in obliquely angled lineations, which may have formed during a single ice-flow phase (Livingstone et al., 2010a). The ridges often display a distinctive asymmetry, with steep north and west faces and gently dipping southern and eastern flanks (Krabbendam and Bradwell, 2011) (Fig. 3). This is explained by the distinctive geology, which comprises Lower Carboniferous sedimentary rocks composed of rhythmic sequences of limestone, shale, sandstone, greywacke and coal (units 5–20 m thick) intruded in places by the resistant strata of the Whin Sill dolerite that dip 10–20˚ SSE towards the edge of the northern Pennines (Krabbendam and Bradwell, 2011). Further to the north-east the strike of the strata swings around to the NNE, with a dip of 5–10˚ to the east. This was transverse to ice flow, resulting in crag-and-tail features where till has been smeared across the ridges (Krabbendam and Bradwell, 2011). Till is more extensive in this region, with more subdued lineations orientated east to ENE.

Geomorphological mapping has revealed three distinct flow sets (Livingstone et al., 2008, 2010a): (1) a central trunk zone of W–E to SW–NE lineated terrain; (2) lineations stretching SE down the North Tyne Valley; and (3) a small set of N–S orientated lineations trending down the NE coast (Fig. 4A). Flow set 1 records convergent ice flow sourced from the Lake District and Scottish Southern Uplands, supplemented by locally sourced ice flowing into the main artery from the northern Pennines. During this time, ice overtopped the Tyne Gap and flowed towards the east coast as a topographically controlled ice stream (Beaumont, 1971; Bouledrou et al., 1988; Livingstone et al., 2010a). Subtle differences in lineation orientation (Fig. 4) reflect the sensitivity of the ice stream to western ice-dispersal centres (Livingstone et al., 2008). Towards the east coast lineations become more subdued and eventually are overprinted by flow set 3. The second flow set, trending SE down the North Tyne Valley, documents the subsequent weakening of the Tyne Gap as an easterly draining ice stream and increased dominance of Southern Upland ice (Livingstone et al., 2010a). This probably coincided with inland extension of the southward-flowing NSL (flow set 3) (Davies et al., 2009, 2012).

Meltwater channels and a prominent moraine (‘Crowden Hill Moraine’; Figs 4 and 6) at the eastern end of the main W–E orientated flow set record nascent retreat of the TGIS into the Tyne and Wear Lowlands or an ice-marginal position of the NSL (Smythe, 1912; Livingstone et al., 2010a; Teasdale, 2013). Along the shallower western flank of the moraine, triangular-shaped deposits of glacial sand and gravel have been observed within present-day river valleys (Smythe, 1912). Westward retreat of the TGIS resulted in an extensive proglacial drainage system in the ice-free area between the NSL and TGIS (Davies et al., 2012; Yorke et al., 2012), terminating to the east in a major proglacial lake, Glacial-Lake Wear (Smith, 1994; Teasdale and Hughes, 1999).
Ice-marginal and ice-contact deposits in the lower Tyne Valley recorded periods of stagnation and in situ downwasting (Yorke et al., 2012). This proglacial system can be traced along the Tyne Valley to the west as far as the Brampton Kame Belt, a major glaciofluvial complex formed where ice stagnated in the lee of the Pennines (Trotter, 1929; Livingstone et al., 2010b).

At some point during deglaciation of the TGIS a switch in ice flow resulted in drawdown into the Solway Lowlands and Irish Sea Basin out of Bewcastle Fells and the Vale of Eden (Livingstone et al., 2008, 2010c, 2012).

Field and laboratory methods

Geomorphic mapping

Glacial geomorphological mapping of palaeo-ice flow from glacial lineations, and the derivation of flow sets and relative timings based on cross-cutting relationships has been presented by Livingstone et al. (2008, 2010a) and is summarized above and on Fig. 4A. This is supplemented here by detailed mapping of deglacial features, including: moraines, glaciofluvial deposits and meltwater channels, from NEXTMap (5-m resolution) and LiDAR data (1-m resolution) to reconstruct glacier geometry and ice-marginal positions during retreat.

**Cosmogenic nuclide (10Be) surface exposure dating**

**Sampling strategy**

Single samples for cosmogenic nuclide (10Be) surface exposure dating were taken from five separate glaciated bedrock surfaces at locations along the length of the Tyne Gap to constrain palaeo-ice stream retreat (Figs 4B and 5). The relatively small number of single samples from separate...
glacial erosion features limits our geochronological reconstruction. In particular, a single age is less robust than one derived from multiple samples on the same feature. Moreover, as variance in the exposure age range also increases with number of samples (e.g. Balco, 2011; Applegate et al., 2012) an age derived from one sample may give a false sense of chronological precision. All samples were from quartz-rich Carboniferous sandstones from the Stainmore Formation, Fell Sandstone Group and Border Group. Sites were chosen that showed evidence of glacial erosion, such as abraded and striated surfaces or roches moutonnées, to minimize the possibility of bedrock inheritance. Sampling sites that have had, or continue to harbour, snow, vegetation or sediment cover were avoided. Samples were collected using a rock saw, which allowed the removal of intact surface blocks (10 \times 10 \times 4 \text{ cm}; ca. 3 \text{ kg}) at least 30 \text{ cm} from all outcrop edges. Sampled surfaces were roughened after sampling or filled to avoid leaving unsightly unnatural cuts in the bedrock. All samples had their position, altitude, topographic shielding, surface dip/direction, dimension and surface characteristics (pitted/spalled) recorded.

Sample preparation

Samples were prepared at the Cosmogenic Isotope Analysis Facility in the Scottish Universities Environmental Research Centre. Quartz was isolated from other minerals by mechanical (crushing, grinding, magnetic separation) and chemical (hexafluorosilicic acid treatment and hydrofluoric acid leaching) procedures. In this study, 220 \mu g \textsuperscript{9}Be was added as a carrier to each sample. \textsuperscript{10}Be extraction and target preparation followed procedures modified from Kohl and Nishiizumi (1992) and Child et al. (2000). The \textsuperscript{10}Be/\textsuperscript{9}Be ratios of mixed BeO-Nb targets were measured with the 5-MV accelerator mass spectrometer at Scottish Universities Environmental Research Centre (Xu et al., 2010). The Be ratios were converted to nuclide concentration in quartz. The processed blank \textsuperscript{10}Be/\textsuperscript{9}Be ratio was between 2 and 6\% of the sample \textsuperscript{10}Be/\textsuperscript{9}Be ratios and was subtracted from the measured ratios. Details of sample locations and relevant analytical data are given in Table 1.

Exposure age calculation

The exposure ages were calculated using the Cronus-Earth online calculator v.2.2 (hess.ess.washington.edu/) and are presented in Table 1. \textsuperscript{10}Be ages are calculated using a \textsuperscript{10}Be half-life of 1.36 \text{ Myr} and the sea level and high latitude (SLHL) production rate of 4.39 \pm 0.37 \text{ atoms g}^{-1} \text{ a}^{-1} (Lm scaling) obtained from age-constrained calibration measurements (Balco et al., 2008). To decrease the scaling uncertainties, we calculated the exposure ages using a locally derived production rate (LPR) of 3.99 \pm 0.13 \text{ atoms g}^{-1} \text{ a}^{-1} (Lm scaling; SWH LPR) based on a deglaciation age of 12.2 \text{ ka} from the NW Scottish Highlands (NWH) (Ballantyne, 2012; Ballantyne and Stone, 2012). Thus, the ages calculated using the local production rate become older and more precise than \textsuperscript{10}Be ages calculated using the global production rate. Although the NWH LPR is not

Figure 3. Examples of two bedrock ridges at the western end of the Tyne Gap palaeo-ice stream corridor (ice flow from right to left in both instances). Note the steeper stoss side and shallow lee side in A.
Figure 4. (A) NEXTMap image showing the geomorphology of the study area. The black dotted lines demarcate the edge of the lineated terrain associated with the palaeo-ice stream (see Livingstone et al., 2010a). The three main ice flows are depicted by the white arrows. (B) The mapped deglacial features (meltwater channels, moraine) and interpreted ice-marginal positions (black-dotted lines). Moraine 1 is the Crowden Hill Moraine, moraine 2 is the Thorneyford Moraine and moraine 3 is the North Tyne Moraine. 

$^{10}$Be ages are presented as ka BP with 1σ error (Table 1). Black text refers to ages calculated with a zero erosion rate and red text refers to ages calculated with an erosion rate of 1 mm ka$^{-1}$. Radiocarbon ages are presented as calibrated years before present (cal ka BP) with 1σ error (Table 2). The black box refers to Fig. 6.
independently constrained, it matches the LLPR (Loch Lomond production rate) of 3.92 ± 0.18 atoms g\(^{-1}\) a\(^{-1}\) (Lm scaling; Fabel et al., 2012, unpublished data). The LLPR was calculated using the \(^{10}\)Be concentration in boulders located on the Loch Lomond terminal moraine and is independently constrained by radiocarbon dating (MacLeod et al., 2011).

The calculated age uncertainties are expressed as 1σ error (Table 1). Corrections for sample thickness (4–5 cm) assume an exponential depth decrease of in situ production rate and an attenuation length of 160 g cm\(^{-2}\). Carboniferous sandstone surfaces show evidence of granular disintegration. We have therefore calculated minimum exposure ages based on both zero erosion of sampling surfaces and for an erosion rate of 1 mm ka\(^{-1}\) to investigate the effect of bedrock erosion on sample age. Higher erosion rates are not considered because they result in unfeasibly old ages that pre-date deglaciation.

Figure 5. Photographs illustrating the five bedrock exposures sampled along the Tyne Gap Palaeo-Ice Stream corridor: (A) TY1: upstanding (1–2 m) bedrock promontory on a Carboniferous Sandstone escarpment; (B) TY2: ice-moulded bedrock (roche moutonée) at the western edge of a Carboniferous Sandstone ridge; (C) TY3: upstanding (1–3 m) bedrock promontory on a glacially abraded Carboniferous Sandstone crag; (D) TY4: large upstanding tor standing up to 5 m above the surrounding crags of the Carboniferous Sandstone ridge; (E) TY5: upstanding (0.5 m) bedrock surface on the lee side of a bedrock-moulded, W–E trending drumlin composed of Carboniferous sandstone.
Table 1.  

| SampleAMS ID | Latitude (˚N) | Longitude (˚W) | Altitude (m) | Thickness (cm) | Density (g cm⁻³) | 10Be Exposure age (a) Internal | 10Be Exposure age (a) External |
|--------------|---------------|----------------|--------------|----------------|------------------|-------------------------------|-------------------------------|
| TY1 b6981    | 55.13         | 1.92            | 202          | 4              | 2.42             | 0.9988                        | 0.0988                        |
| TY2 b6982    | 55.19         | -1.93           | 230          | 4.5            | 2.03             | 0.9999                        | 0.0999                        |
| TY3 b6986    | 55.14         | -2.36           | 231          | 4              | 1.94             | 0.9998                        | 0.0998                        |
| TY4 b6988    | 55.20         | -1.96           | 110          | 5              | 2.36             | 0.9995                        | 0.0995                        |

The 10Be concentrations are corrected for the procedural blank value of (5.83 ± 1.19) × 10⁴ atoms g⁻¹. Shielding factor 10Be/9Be blank-corrected ratios and 10Be concentrations are referenced to NIST SRM 4325 (2.79 ± 0.13 × 10¹¹; Nishiizumi et al., 2007). Uncertainties (¹σ) include all known sources of analytical error. Corrections for sample thickness assume an exponential depth decrease of the in situ production rate and an attenuation length of 160 g cm⁻³. Exposure ages were calculated using a production rate of 3.99, based on a deglaciation age of 12.2 ka in Scotland (Ballantyne and Stone, 2012). The calculated age uncertainties are expressed as 1 1σ. The external uncertainties include the total (systematical) and the internal (analytical) and geometric shielding were taken into account when calculating the exposure ages (Table 1). No correction was made for snow shielding because of the difficulty in constraining long-term snow cover. Any correction for snow shielding and bedrock erosion would increase the nuclide concentrations (and implicitly the ages), and thus all reported 10Be ages calculated with zero erosion in this study are regarded as minimum ages.

Radiocarbon dating

Lake coring, sample preparation and age calculation

Core samples from two small lakes were taken using a Russian-pattern corer (50 cm in length) from a small inflatable boat. Material for radiocarbon dating was sampled at Talkin Tarn, within the Brampton Kame Belt, and Crag Lough (Fig. 2; Table 2). Pre-treatment and accelerator mass spectrometry radiocarbon analyses of the samples (Beta-342469, 342470, 342471) was undertaken at the BETA Analytic Laboratories, following their standard procedures (www.radiocarbon.com/pre-treatment-carbon-dating.htm). The peat sample (L2) from Crag Lough was subjected to an acid–alkali–acid pre-treatment process to eliminate carbonates and secondary organic acids. Due to the absence of macrofossils, samples L1a and L1b from Talkin Tarn were both from bulk organic fractions. These samples underwent repeat acid washes to ensure removal of carbonates. It is acknowledged that bulk organic samples may include some minor components of old carbon, resulting in age overestimations, typically of ~1000 years (e.g. Hägvar and Ohlson, 2013). Thus, although we are dating deglacial lake sediments that provide a minimum age for the onset of ice-free conditions, the bulk organic sediments are maximum age measurements. To ease comparison of radiocarbon and cosmogenic nuclide chronologies, all radiocarbon ages are presented in calibrated form using Calib 7.0.4, the IntCal13 curve (Reimer et al., 2013), a marine ΔR of ~525 years where appropriate (see also Chiverrell et al., 2013) and giving the 1 sigma error range in cal ka BP.

Bayesian analysis of deglacial chronology

To provide further constraints on the deglaciation of the TGIS we constructed a ‘prior model’ of the expected chronological order of ice retreat reasoned solely from glacial geomorphological evidence (after Chiverrell et al., 2013). Bayesian modelling was implemented using a simple Sequence model in Oxcal (c14.arch.ox.ac.uk; Ramsey, 2009), which requires the dated events to be in chronological order, and uses Markov Chain Monte Carlo sampling to form a probability distribution of dates through the sequence. The result is a posterior density, which is a product of the prior model and likelihood probabilities. To run the model we divided the TGIS into three sub-regions (‘before inner TGIS’, ‘inner TGIS’ and ‘after inner TGIS’), which share common relationships and are separated by Boundary commands. We used a Phase command for the inner TGIS sub-region as relative ages are more difficult to define because of the additional complication of ice flow down the North Tyne Valley (Livingstone et al., 2010a). Ages that did not fit the sequence were declared as outliers and excluded from the model.

Results

General geomorphology

Glacial geomorphological mapping of meltwater channels, glacioluvial deposits and moraine hummocks and ridges has
revealed three significant ice-marginal positions within the TGIS corridor (Fig. 4B, labelled 1–3). The easternmost ice-margin position is delineated by a prominent moraine ~60 m asl, which stretches over 30 km S/SE from the River Croquet to the River Tyne. This is the extension of the ‘Crowden Hill Moraine’, previously identified by Smythe (1912) and Teasdale (2013).

A second ice-margin position (Thorneyford Moraine), 10–15 km further inland, is marked by a discontinuous series of ridges arranged in a lobate configuration, bisected at the northern end by flights of short SW–NE orientated meltwater channels (Fig. 6). To the east of the Thorneyford Moraine (Fig. 6) there are two flat-bottomed basins and several straight to sinuous channels that cut across higher ground on the eastern flank of the basins. The terrain to the west of the Thorneyford moraine comprises W–E orientated lineations associated with the main ice-flow phase. A third, less distinct ice-marginal position (North Tyne Moraine) is identified by a series of small moraine ridges, hummocky moraine and/or glaciofluvial deposits within the North Tyne Valley. This still-stand position is associated with the NW–SE flow-set.

**Overview of individual sites**

Most of the sites sampled for cosmogenic surface exposure ages occur within the central corridor of the palaeo-ice stream with the exception of sample TY4. TY1–3 and TY5 are in areas of glacially abraded terrain situated on bedrock highs controlled by the predominant strike of the local Carboniferous Sandstone. TY4 is situated on the northern Table 2. Radiocarbon ages.

| Lab code     | Site           | Sample ID (depth cm) | Material     | $^{13}$C/$^{12}$C | Measured radiocarbon age ($^{14}$C a BP) | Calibrated $^{14}$C age (cal a BP)* |
|--------------|---------------|----------------------|--------------|-------------------|------------------------------------------|-------------------------------------|
| BETA-342471 | Talkin Tarn   | L1a (971–972)        | Organic sediment | $-22.2$          | $13 080 \pm 60$                          | $15 690 \pm 130$                     |
| BETA-342470 | Talkin Tarn   | L1b (948–949)        | Organic sediment | $-24.7$          | $11 760 \pm 50$                          | $13 550 \pm 70$                      |
| BETA-342469 | Lough Crag    | L2 (320–321)         | Peat          | $-26$             | $11 190 \pm 50$                          | $13 060 \pm 40$                      |

*Accelerator mass spectrometry radiocarbon ages are $^{14}$C a BP $\pm 1\sigma$. Calibrated ages are in calendar years before present (BP) $\pm 1\sigma$. 

Figure 6. The Crowden Hill and Thorneyford moraine ridges (white dotted lines) (Fig. 4B for locations). The Thorneyford Moraine has a distinctive lobate planform. The terrain to the west of the moraine is lineated, while to the east there are two small basins, with overspill channels at their eastern margin. The digital elevation model is a combination of NEXTMap and LiDAR data.
edge of the ice stream corridor. With respect to the lake samples, Crag Lough is also situated within the main ice stream corridor, at the foot of a bedrock escarpment controlled by a NNW–SSW striking Whin Sill intrusion. In contrast, Talkin Tarn is located to the west of the main flow corridor in an area of hummocky ice stagnation terrain related to a later deglacial phase of the ice stream (Livingstone et al., 2010b).

Sample TY1 was taken from the upper flat surface of Pipers Rock (302 m asl), which is part of Shattock Crags, a large Carboniferous escarpment located in the central corridor of the main W–E flow set of the ice stream. The escarpment comprises interbedded argillaceous rocks and sandstone from the Stainmore Formation. The surface is covered with lichens and has been heavily weathered, with signs of granular disintegration, large pits and numerous runnels, up to 30 cm deep in places. The bedrock promontory is 2–3 m above local ground level and has clearly not been vegetated for some time.

Sample TY2 is from Rothley Crags (230 m asl), a Carboniferous ridge trending roughly N–S. The ridge comprises interbedded argillaceous rocks and sandstone from the Stainmore Formation. The western leading edge has roches moutonnées indicating moulding during easterly ice flow. The sample was taken on the upper surface of a roche moutonnée from a stoss/mid location about 1–2 m above ground level. There are fractures and joints and some surface weathering, with large pits up to 20 cm deep and evidence of granular disintegration. However, weathering appears limited at the sample site. The rock surface is lichen covered and may have been vegetated in the past.

TY3 is from Snabdaugh Crags (221 m asl), a Carboniferous outcrop trending roughly N–S, with a partially glacially abraded upper surface. The outcrop comprises interbedded argillaceous rocks and limestone from the Border Group. This region is associated with the later NW–SE flow phase down the North Tyne Valley. The sample was taken from the upper surface of a bedrock promontory 1–3 m above local ground level. There is some evidence of surface weathering, including spalling, rills, small pits (up to 5 cm) and granular disintegration.

TY4 was sampled at Simmonside (431 m asl), a W–E trending Carboniferous sandstone ridge of the Fell Sandstone Group, upstanding above the surrounding landscape. This site has been mapped as the northern edge of the ice stream (Livingstone et al., 2010a), and is altitudinally distinct from the other samples which occupy the floor of the Tyne Gap corridor. The sample was taken from the upper surface of a large pedestal rising up to 5 m above the surrounding terrain. The structure is part of a large crag comprising independent (~2 × 2 m) pedestals. There is some lichen cover and evidence of granular disintegration, spallation and large (~5 cm) weathering pits.

Sample TY5 was from Brierton Hill (110 m asl), a W–E trending whaleback towards the southern margin of the ice stream corridor (Livingstone et al., 2010a). The whaleback comprises interbedded argillaceous rocks and sandstone from the Stainmore Formation. The sample was taken from a bedrock surface up to 0.5 m above the local ground level. The surface does not show significant weathering, although some granular disintegration and spalling is evident. The rock is also lichen-covered.

Two lake sites were identified as coring targets to supplement the cosmogenic nuclide samples. This included Talkin Tarn (L1), an enclosed kettle hole in the Brampton Kame Belt, and Crag Lough (L2), a natural lake formed in the northern lee of the Whin Sill, Tyne Gap (Fig. 4B).

### Chronology

Cosmogenic 10Be from Carboniferous Sandstone bedrock and basal radiocarbon ages from lake cores in the palaeo-ice stream corridor reveal the timing and duration of ice stream deglaciation (Figs 4B, 5 and 7). Sample TY4 (431 m asl) yielded an age of 47 ± 1.9 10Be ka BP, which is anomalously old compared with the other samples. Given this, and its ice-stream marginal position, we suggest the surface suffered minimal erosion during the Late Devensian and can therefore be excluded from our geochronological model of ice retreat due to incomplete re-setting.

The remaining four cosmogenic samples were taken between 100 and 220 m asl, within the main trunk of the TGIS. The two most easterly samples, TY1 and TY2, returned 10Be ages of 17.8 ± 0.9 and 17.6 ± 1.1 ka BP, respectively. TY5, from the southerly edge of the palaeo-ice stream, provided a 10Be age of 17.2 ± 0.9 ka BP, while further west sample TY3 was dated to 16.5 ± 0.8 ka BP. With an erosion rate of 1 mm ka−1 the 10Be ages become 242–260 years older.

Radiocarbon ages were taken from two lake cores at Talkin Tarn (L1) and Crag Lough (L2). The Crag Lough sediment core was 2.0 m long, consisting of two bio/lithofacies, an upper brown–black well-humified peat with some sand and silt stringers, and a lower grey silty clay (Fig. 8). The radiocarbon sample was taken directly above the lower contact at the base of the humified peat (300–301 cm) and provided an age of 13.1 ± 0.04 cal ka BP. The 0.5-m-long Talkin Tarn sediment core comprises a lower light-brown to grey partially laminated silt with some sand and an increasing organic content up core (Fig. 9). A bulk sample taken from the base of this facies (L1a) at 971–972 cm depth provided an age of 15.7 ± 0.1 cal ka BP. A further bulk sample was taken at the top of the lower facies (948–949 cm depth), 1 cm below a prominent 5 cm sand layer (L1b). This provided an age of 13.6 ± 0.07 cal ka BP. Above the sand facies is a 6-cm grey silt facies followed by 46 cm of gyttja, which comprises the bulk of the sequence.

Bayesian modelling of this geochronology used a relative age order (cf. Chiverrell et al., 2013) characterized by westward retreat of the TGIS into the Solway Lowlands (Fig. 7). The geochronological results are compatible with this model, with samples to the west across a distance of ~55 km yielding progressively younger ages. The exception is the 14C date at Lough Crag, which is too young, and is therefore considered an outlier. The model constrains the retreat of ice from the eastern edge of the TGIS to before 18.7–17.1 ka BP (Boundary Basal, Fig. 7), from the western edge of the TGIS to before 16.4–15.7 ka BP (Boundary after Inner TGIS, Fig. 7) and from Talkin Tarn, at the confluence between the TGIS and ice in the Vale of Eden, to before 15.7 ± 0.1 cal ka BP (Fig. 9).

### Discussion

Decoupling of the TGIS from the NSL

Previous mapping has established the ice-flow history of the TGIS (Livingstone et al., 2008, 2010a). This demonstrates three clear flow phases: (1) a main W–E ice flow through the Tyne Gap discharging into the North Sea at Newcastle; (2) a later NW–SE flow down the North Tyne Valley, which infers a switch to a more northerly Scottish ice source into the Tyne corridor; and (3) encroachment of the N–S flowing NSL into the coastal lowlands of Northumbria and Durham (Fig. 4). However, the deglacial dynamics and timing of these competing ice flows are poorly understood.
Our new mapping and geochronology help to constrain the decoupling of the TGIS and NSL, and the formation of Glacial Lake Wear. Of the three still-stand positions mapped, the Crowden Hill Moraine (No. 1 on Fig. 4) is geomorphologically distinct and has a NNW–SSE orientation. Its westward-facing convexity in planform suggests it does not relate to the TGIS as it began to retreat westwards; more probably it relates to the onshore movement of the NSL. This is supported by the work of Smythe (1912) and Teasdale (2013), who link this ice-margin position to the formation of regional ice-dammed lakes, such as Glacial Lake Wear (Fig. 4).

The Thorneyford Moraine (No. 2 on Fig. 4) has a nested, arcuate planform that indicates ice downwasting in a westwards direction. It is the first definitive evidence for the uncoupling of the TGIS as it began to retreat westwards; more probably it relates to the onshore movement of the NSL. This is supported by the work of Smythe (1912) and Teasdale (2013), who link this ice-margin position to the formation of regional ice-dammed lakes, such as Glacial Lake Wear (Fig. 4).

The North Tyne Moraine (No. 3 on Fig. 4) is the least distinctive, but may mark the continued retreat of ice into the North Tyne system.

There is additional evidence for local damming of glacial lakes other than Lake Wear. To the west of the Thorneyford Moraine the terrain is clearly glacially streamlined, whereas to the east the terrain is composed of channels and two flat-bottomed basins previously reported as ice-marginal lakes impounded by the NSL near Ponteland (Teasdale, 2013). An alternative hypothesis is that these lakes were dammed by local topography in front of the westerly retreating TGIS. Overspill channels decanting across higher ground to the east of the basins suggests this terrain was at least partially deglaciated at the time of their formation. The North Tyne Moraine (No. 3 on Fig. 4) is the least distinctive, but may mark the continued retreat of ice into the North Tyne system.

To dam Glacial Lake Wear (Smith, 1994; Teasdale and Hughes, 1999) the NSL must have decoupled from the TGIS. At this time, the western edge of the NSL must have been from the Tyne, southwards into County Durham and westwards into the Tyne Valley (Smith, 1994).

Figure 7. Bayesian sequence model of the dating control for the palaeo-TGIS, showing the OxCal keywords that describe the model (Ramsey, 2009). Each distribution (light grey) represents the relative probability of each age estimate with the posterior density estimate (black) generated by the modelling.
located subparallel to the present east coast, damming the Tyne Valley. It is probable that this margin was coincident with the Crowden Hill Moraine, which extends south of the Tyne into County Durham where it is also associated with the formation of Glacial Lake Wear (Davies et al., 2009; Teasdale, 2013). TY1 and TY2 provide maximum limiting dates for moraine 2, which constrains the formation of Glacial Lake Wear to a period before 18.7–17.1 ka BP (Fig. 7, Boundary Basal), in the area deglaciated following contraction of the TGIS. This broadly agrees with dating control at Dimlington along the east coast of Yorkshire, which suggests the Skipsea Till, associated with the NSL, was deposited between 21.7 and 16.2 ka BP (Bateman et al., 2011), while Glacial Lake Humber was still in existence until 16.6 ka BP (Bateman et al., 2008, 2011; Murton et al., 2009).

Retreat of the TGIS

The $^{10}$Be surface exposure ages and $^{14}$C radiocarbon ages recovered from bedrock and lake cores respectively indicate deglaciation of the TGIS from 18.7 to 17.1 ka BP, with progressive retreat from the east coast of England into the Solway Lowlands over a distance of ~55 km by 16.4–15.7 ka BP at the latest (Fig. 7). In particular, Bayesian modelling indicates that the basal $^{14}$C radiocarbon age from Talkin Tarn is consistent with the $^{10}$Be surface exposure age history (Fig. 7). Together these dates provide the first constraints on the deglacial history of the Tyne Gap region. At Crag Lough it appears there was a lag between deglaciation and the $^{14}$C age that sampled a horizon stratigraphically higher in the deglacial sequence compared with the Talkin Tarn measurements.

Our new geochronology was unable to differentiate between the main eastwards ice flow across the Tyne Gap and later SE ice flow down the North Tyne. This favours ice-flow reorganization rather than a two-phase model of TGIS retreat into the Solway Lowlands followed by re-advance of ice down the North Tyne Valley. This may be a product of eastward expansion of the Southern Upland–Scottish Highlands ice divide (Finlayson et al., 2010; Livingstone et al.,...
LATE DEVENSIAN DEGLACIATION OF THE TYNE GAP PALAEO-ICE STREAM

Recent studies in the Lake District have suggested increased topographic control during deglaciation (Hughes et al., 2010a) and increased topographic control during deglaciation (Hughes et al., 2014). The Thorneyford and North Tyne moraines constrain the pattern of rapid retreat through the ice stream corridor as ice downwasted. At the eastern end of the corridor, Yorke et al. (2012) demonstrate that retreat was accompanied by the widespread deposition of ice-marginal glaciolacustrine and glaciofluvial sediments, which may have fed down valley into Glacial Lake Wear during early deglaciation.

The Brampton Kame Belt records downwasting of ice at the confluence between the TGIS and ice in the Vale of Eden. The basal radiocarbon age of 15.7 ± 0.1 cal ka BP, from Talkin Tarn, provides a minimum age for downwasting of the two ice streams and potentially limits the deglaciation of the Solway Lowlands. The timing of retreat out of the Tyne Gap raises questions over the timing of ice-drawdown and flow-phase switching in the Solway Lowlands (Livingstone et al., 2010c, 2012). 26Be surface exposure ages on four Shap granite erratics at the upstream head of the Vale of Eden (275–325 m asl), immediately to the west of the Stainmore Gap (Fig. 2), suggest ice-free conditions prevailed by ~17 ka BP (Wilson et al., 2013a), and is consistent with widespread emergence of the Lake District during this period (cf. Wilson and Lord, 2014). Combined with our new geochronology, this evidence limits the window for complex flow-phase switching westwards into the Solway Firth (Blackhall Wood–Gosforth Readvance) to before 17–16 ka BP. Indeed, to account for the continued activity of the TGIS during this flow switch we need to invoke an ice divide across the Tyne Gap, allowing drainage of ice westwards into the Solway Firth and eastwards into the North Sea.

Wider implications for the British Ice Sheet

Livingstone et al. (2012) produced a six-stage model of ice-flow history and behaviour in the central sector of the last BIIS (Fig. 10A). Our new mapping and geochronology has enabled us to further constrain and finesse this model (Fig. 10B).

Given that complex flow-phase switching westwards into the Solway Firth (Blackhall Wood–Gosforth Readvance – Stage IV) must have occurred before 17–16 ka BP, it is unlikely to be associated with the Killard Point Stadial, which reached its maximum ~16.5 ka (McCabe et al., 2007). This supports our previous reconstruction, although the timing may have been slightly later (fig. 10A, Livingstone et al., 2012). The influence of this switch on the ice-flow history of the eastern Irish Sea Basin is poorly constrained. The ISIS is thought to have retreated back into the northern Irish Sea Basin by 22.6–20.9 ka, with relatively rapid retreat of the western Irish Sea Basin by 20–19.1 ka BP (Chiverrell et al., 2013). There is therefore no indication that flow-phase switching in the central sector of the BIIS was part of a wider pan-Irish Sea reorganization.

The retreat of the TGIS from 18.7 to 17.1 ka, deglaciation of the Brampton Kame Belt by 15.7 ± 0.1 cal ka BP (Fig. 7) and the southern edge of the Vale of Eden by ~17 ka BP (Wilson et al., 2013a), coupled with widespread thinning of Lake District ice between 17 and 15 ka BP (e.g. Ballantyne et al., 2009; McCarroll et al., 2010; Wilson et al., 2013b) suggests widespread collapse of the southern and central source areas between about 18 and 16 ka BP (Fig. 10B). This occurred during a period when both the NSL and the ISIS were active and extended considerable distances southwards (e.g. Bateman et al., 2011; Clark et al., 2012). In the northern Irish Sea Basin, ice re-advanced at Killard Point, north-east Ireland, sometime after 16.9 ± 0.2 ka BP, reaching a maximum extent at ~16.5 ka BP (Killard Point Stadial; McCabe et al., 2007). Optically stimulated luminescence ages ranging from 16.4 to 14.1 ka on Isle of Man outwash deposits provide further constraints (Thrasher et al., 2009), while this event may also have been coincident with the Scottish Re-advance in the Solway Lowlands, although there is currently no age control. In eastern England the NSL, fed by Scottish ice emanating from the Forth and Tweed (Davies et al., 2011), readvanced southwards to the Yorkshire coast, depositing the Withernsea Till within the period 16.2–15.5 ka (Bateman et al., 2011; Roberts et al., 2013). Together, the geochronology implies that the smaller and lower latitude dispersal centres had already crossed a threshold for collapse and were no longer sustainable by 18–16 ka. Conversely, northern dispersal centres remained healthy and were able to respond to renewed climatic cooling possibly associated with Heinrich Event 1 at ~17–16 ka (e.g. McCabe and Clark, 1998; McCabe et al., 1998, 2005, 2007; Bateman et al., 2011; Chiverrell et al., 2013; Roberts et al., 2013).

Conclusions

There is considerable uncertainty over the timing and rate of retreat and complex flow-phase switching in the central

Figure 9. Log and photograph of the 0.5-m Talkin Tarn core.
sector of the last BIIS during deglaciation. This makes it difficult to correlate ice-flow phases across the ice sheet and ultimately to identify and investigate controls (internal and external) governing ice-sheet behaviour. In this paper we present a simple chronological model based on single $^{10}$Be surface exposure and $^{14}$C radiocarbon ages, which provide the first constraints on deglaciation of the TGIS corridor. This is supplemented by high-resolution mapping of deglacial features, including meltwater channels and moraine ridges and hummocks.

We demonstrate that westward retreat of the TGIS had begun by 18.7–17.1 ka BP and reached the Solway Lowlands by 16.4–15.7 ka BP. The first definitive evidence for retreat and uncoupling from the NSL is a prominent lobate moraine, 10–15 km inland of the present-day coast. Two $^{10}$Be ages of 17.8–17.6 ka BP provide maximum bounds on the moraine's formation, and constrain the formation of Glacial Lake Wear to a period before $\sim$18.7–17.1 ka BP in the area deglaciated by the retreating TGIS. Several smaller lakes also became dammed and overspilled across higher ground to the east of the TGIS during this still-stand. Our new geochronology is not of high enough resolution for us to differentiate between the W–E and later NW–SE ice flow phases. This short time window favours a reorganization of ice flow rather than a

Figure 10. (A) Reconstruction of ice dynamics in the central sector (stages IV–VI) from Livingstone et al. (2012). (B) Updated reconstruction based on our new geochronological and geomorphological constraints. Red lines numbered 1–3 refer to the three moraine systems identified in Fig. 4B. In particular, note that the central sector underwent widespread and collapse between 18 and 16 ka BP, while regions sourced from Scottish ice continued to stream a considerable distance south during this time.
separate late-phase re-advance of ice down the North Tyne Valley.

An age of 15.7 ± 0.1 cal ka BP from Talkin Tarn provides a minimum age for deglaciation of the Brampton Kame Belt, which records downwasting of ice at the confluence of the TGIS and ice in the Vale of Eden. This indicates continued retreat of the TGIS into the Solway Lowlands and limits the timing of complex flow-phase switching westwards into the Solway Firth to before 17–16 ka BP.

Together with other published ages (cf. Wilson and Lord, 2014) our data indicate that southern ice dispersal centres (e.g. Lake District, Howgill Fells and Pennines) and their drainage outlets (e.g. Tyne and Vale of Eden) collapsed between 18 and 16 ka BP. This response is in stark contrast to more northerly ice dispersal centres, which remained active, feeding the re-advance of the NSL into Yorkshire (~16–15 ka BP) and the Killard Point Stadial in the northern Irish Sea Basin (~16.5 ka BP).

Abbreviations. BIIS, British–Irish Ice Sheet; ISIS, Irish Sea Ice Stream; LLPR, Loch Lomond production rate; NSL, North Sea Ice Lobe; NWH LPR, NW Scottish Highlands locally derived production rate; TGIS, Tyne Gap Palaeo-ice Stream

Acknowledgments. This work was funded by a NERC Cosmogenic Isotope Analysis Facility grant (9125/1012). We thank the various land owners and farmers who gave us permission to work on their land. Richard Chiverrell and an anonymous reviewer are thanked for their constructive reviews.

References

Applegate PJ, Urban NM, Keller K et al. 2012. Improved moraine age interpretations through explicit matching of geomorphic process models to cosmogenic nuclide measurements from single landforms. Quaternary Research 77: 293–304 [DOI: 10.1016/j. yqres.2012.11.002].

Balco G. 2011. Contributions and unrealized potential contributions of cosmogenic-nuclide exposure dating to glacier chronology, 1990–2010. Quaternary Science Reviews 30: 5–27 [DOI: 10.1016/ j.quascirev.2010.11.003].

Balco G, Stone JO, Lithon NA et al. 2008. A complete and easily accessible means of calculating surface exposure ages or erosion rates from 10Be and 26Al measurements. Quaternary Geochronology 3: 174–195 [DOI: 10.1016/j.quageo.2007.12.001].

Ballantyne CK. 2012. Chronology of glaciation and deglaciation during the Loch Lomond (Younger Dryas) Stage in the Scottish Highlands: implications of recalibrated 14C exposure ages. Boreas 41: 513–526 [DOI: 10.1111/j.1502-3885.2012.00251.x].

Ballantyne CK, Stone JO. 2012. Did large ice caps persist on low ground in north-west Scotland during the Lateglacial Interstadial? Journal of Quaternary Science 27: 297–306 [DOI: 10.1002/ jqs.1544].

Ballantyne CK, Stone JO, Fifield LK. 2009. Glaciation and deglaciation of the SW Lake District, England: implications of cosmogenic 36Cl exposure dating. Proceedings of the Geologists’ Association 120: 139–144 [DOI: 10.1111/j.1525-1385.2009.00803].

Bateman MD, Buckland PC, Chalmers B et al. 2008. The Late-Devensian proglacial Lake Humber: new evidence from littoral deposits at Ferrybridge, Yorkshire, England. Boreas 37: 195–210 [DOI: 10.1111/j.1502-3885.2007.00013.x].

Bateman MD, Buckland PC, Whyte MA et al. 2011. Re-evaluation of the Last Glacial Maximum typesite at Dilimmingt, UK. Boreas 40: 573–584 [DOI: 10.1111/j.1502-3885.2011.00204.x].

Beaumont P. 1971. Stone orientation and stone count data from the lower till sheet, eastern Durham. Proceedings of the Yorkshire Geological Society 38: 343–360 [DOI: 10.1144/ pysgs.38.3.343].

Bishop WW, Coope GR. 1977. Stratigraphical and faunal evidence for Lateglacial and early Flandrian environments in south-west Scotland. In Studies in the Scottish Late Glacial Environment, Gray JM, Lowe JJ (eds). Pergamon Press: Oxford; 61–88.

Bouleodrou A, Tarling DH, Lunn AG. 1988. Glacial drift thickness in the Tyne Gap, Northumbria. Transactions of the Natural History of Northumbria 55: 20–27.

Child D, Elliott G, Mifsud C et al. 2000. Sample processing for earth science studies at ANTEARS. Nuclear Instruments and Methods in Physics Research B 172: 856–860.

Chiverrell RC, Thomas GSP. 2010. Extent and timing of the Last Glacial Maximum (LGM) in Britain and Ireland: a review. Journal of Quaternary Science 25: 535–549 [DOI: 10.1002/jqs.1404].

Chiverrell RC, Thrasher IM, Thomas GSP et al. 2013. Bayesian modelling the retreat of the Irish Sea Ice Stream. Journal of Quaternary Science 28: 200–209 [DOI: 10.1002/jqs.2616].

Clark CD, Hughes ALC, Greenwood SL et al. 2012. Pattern and timing of retreat of the last British–Irish Ice Sheet. Quaternary Science Reviews 44: 112–146 [DOI: 10.1016/j.quascirev.2010.07.019].

Dasgupta AJ, Roberts DH, Bridgland DR et al. 2011. Provenance and depositional environments of Quaternary sediments from the western North Sea Basin. Journal of Quaternary Science 26: 59–75 [DOI: 10.1002/jqs.1426].

Davies BJ, Roberts DH, Bridgland DR et al. 2012. Dynamic Devensian ice flow in NE England: a sedimentological reconstruction. Boreas 41: 337–336 [DOI: 10.1111/j.1502-3885.2011.00237.x].

Davies BJ, Roberts DH, Ø Coifaigh C et al. 2009. Interlobate ice-sheet dynamics during the Last Glacial Maximum at Whithorn Bay, County Durham, England. Boreas 38: 555–578 [DOI: 10.1111/j.1502-3885.2008.00083.x].

Evans DJA, Livingstone SJ, Vieli A et al. 2009. The palaeoglaciology of the central sector of the British and Irish Ice Sheet: reconciling glacial geomorphology and preliminary ice sheet modelling. Quaternary Science Reviews 28: 739–757 [DOI: 10.1016/j.quascirev.2008.05.011].

Everest J, Bradwell T, Galloway N. 2005. Subglacial landforms of the Tweed palaeo-ice stream. Scottish Geographical Journal 121: 163–173 [DOI: 10.1080/03091530518732729].

Eyles N, McCabe AM, Bowen DQ. 1994. The stratigraphic and sedimentological significance of Late Devensian ice sheet surging in Holderness, Yorkshire, U.K. Quaternary Science Reviews 13: 727–759 [DOI: 10.1016/0277-3791(94)90102-3].

Fabel D, Ballantyne CK, Xu S. 2012. Trimlines, blockfields, mountain-top erratics and the vertical dimensions of the last British–Irish Ice Sheet in NW Scotland. Quaternary Science Reviews 55: 91–102 [DOI: 10.1016/j.quascirev.2012.09.002].

Finlayson A, Merritt J, Browne M et al. 2010. Ice sheet advance, dynamics, and decay configurations: evidence from west central Scotland. Quaternary Science Reviews 29: 969–988 [DOI: 10.1016/j.quascirev.2009.12.016].

Hedges REM, Housley RA, Law IA et al. 1988. Radiocarbon dates from the Oxford AMS system archaeometry datafile 8. Archaeometry 30: 291–305 [DOI: 10.1111/j.1475-4754.1988.tb00454.x].

Hughes ALC, Clarck CD, Jordan CJ. 2014. Flow-pattern evolution of the last British Ice Sheet. Quaternary Science Reviews 89: 148–168 [DOI: 10.1016/j.quascirev.2014.02.002].

Hughes ALC, Greenwood SL, Clark CD. 2011. Dating constraints on the last British–Irish Ice Sheet: a map and database. Journal of Maps 7: 156–184 [DOI: 10.4111/jom.2011.1145].

Hägvar S, Ohlson M. 2013. Ancient carbon from a melting glacier gives high 14C in living pioneer invertebrates. Scientific Reports 3: 2820 [DOI: 10.1038/srep02820].

Kohl CP, Nishizumi K. 1992. Chemical isolation of quartz for measurement of in-situ-produced cosmogenic nuclides. Geochimica et Cosmochimica Acta 56: 3583–3587 [DOI: 10.1016/0016-7037(92)00401-4].

Krabbendam M, Bradwell T. 2011. Lateral plucking as a mechanism for elongate erosional glacial bedforms: explaining megagrooves in Britain and Canada. Earth Surface Processes and Landforms 36: 1335–1349 [DOI: 10.1002/esp.2157].

Livingstone SJ, Evans DJA, Ø Coifaigh C et al. 2012. Glaciodynamics of the central sector of the last British–Irish Ice Sheet in Northern England. Earth-Science Reviews 111: 25–55 [DOI: 10.1016/j.earscirev.2011.12.006].
Livingstone SJ, Evans DJA, Ó Cofaigh C et al. 2010b. The Brampton kame belt and Pennine Escarpment meltwater channel system (Cumbria, UK): morphology, sedimentology and formation. Proceedings of the Geologists’ Association 121: 423–443 [DOI: 10.1016/j.pgeola.2009.10.005].

Livingstone SJ, Ó Cofaigh C, Evans DJA et al. 2010c. Glaciolacustrine sedimentation in the Solway Lowlands (Cumbria, UK): evidence for a major glacial oscillation during Late Devensian deglaciation. Boreas 39: 505–527.

Livingstone SJ, Ó Cofaigh C, Evans DJA. 2008. Glacial geomorphology of the central sector of the last British-Irish Ice Sheet. Journal of Maps 4: 358–377 [DOI: 10.4113/jom.2008.1032].

Livingstone SJ, Ó Cofaigh C, Evans DJA. 2010a. A major ice drainage pathway of the last British-Irish Ice Sheet: the Tyne Gap, northern England. Journal of Quaternary Science 25: 354–370 [DOI: 10.1002/jqs.1341].

MacLeod A, Palmer A, Lowe J et al. 2011. Timing of glacier response to Younger Dryas climatic cooling in Scotland. Global and Planetary Change 79: 264–274 [DOI: 10.1016/j.gloplacha.2010.07.006].

McCabe AM, Clark PU. 1998. Ice-sheet variability around the North Atlantic Ocean during the last deglaciation. Nature 392: 373–377 [DOI: 10.1038/32866].

McCabe AM, Clark PU, Clark J. 2005. AMS 14C dating of deglacial events in the Irish Sea Basin and other sectors of the British-Irish Ice Sheet. Quaternary Science Reviews 24: 1673–1690 [DOI: 10.1016/j.quascirev.2004.06.019].

McCabe AM, Clark PU, Clark J et al. 2007. Radiocarbon constraints on readvances of the British-Irish Ice Sheet in the northern Irish Sea Basin during the last deglaciation. Quaternary Science Reviews 26: 1204–1211 [DOI: 10.1016/j.quascirev.2007.01.010].

McCabe AM, Knight J, McCarron S. 1998. Evidence for Heinrich Event 1 in the British Isles. Journal of Quaternary Science 13: 549–568 [DOI: 10.1002/(SICI)1099-1417(19981101)13:6<549::AID-JQS394>3.0.CO;2-A].

McCarroll D, Stone JO, Ballantyne CK et al. 2010. Exposure-age constraints on the extent, timing and rate of retreat of the last Irish Sea Ice Stream. Quaternary Science Reviews 29: 1844–1852 [DOI: 10.1016/j.quascirev.2010.04.002].

Murton DK, Pawley SM, Murton JB. 2009. Sedimentology and luminescence ages of Glacial Lake Humber deposits in the central Vale of York. Proceedings of the Geologists’ Association 120: 209–222 [DOI: 10.1016/j.pgeola.2009.09.001].

Nishizumi K, Imamura M, Caffee MW et al. 2007. Absolute calibration of 10Be AMS standards. Nuclear Instruments and Methods in Physics Research, Section B: Beam Interactions with Materials and Atoms 258: 403–413 [DOI: 10.1016/j.nimb.2007.01.297].

Ramsey CB. 2009. Bayesian analysis of radiocarbon dates. Radiocarbon 51: 337–360.

Reimer PJ, Bard E, Bayliss A et al. 2013. IntCal13 and MARINE12 radiocarbon age calibration curves, 0–50,000 years cal BP. Radiocarbon 55 [DOI: 10.2458/azu_js_rc.55.16947].

Roberts DH, Evans DJA, Lodwick J et al. 2013. The subglacial and ice-marginal signature of the North Sea Lobe of the British-Irish Ice Sheet during the Last Glacial Maximum at Upgang, North Yorkshire, UK. Proceedings of the Geologists’ Association 124: 503–519 [DOI: 10.1016/j.pgeola.2012.08.009].

Smith DB. 1994. Geology of the Country around Sunderland. Memoir of the British Geological Survey (sheet 21). HMSO: London.

Smythe JA. 1912. The glacial geology of Northumberland. Transactions of the Natural History Society of Northumberland, Durham and Newcastle upon Tyne 4: 86–116.

Teasdale D. 2013. Evidence for the western limits of the North Sea Lobe of the BIS in North East England. In QRA Field Guide: The Quaternary of Northumberland, Durham and Yorkshire, Davies BJ, Yorke L, Bridgland DR, Roberts DH (eds). Quaternary Research Association: Cambridge; 106–121.

Teasdale D, Hughes D. 1999. The glacial history of north-east England. In Quaternary of North East England, Bridgland DR, Horton BP, Innes JB (eds). Quaternary Research Association: London; 10–17.

Thrasher IM, Mauz B, Chiverrell RC, Lang A, Thomas GPS. 2009. Testing an approach to OSL dating of Late Devensian sediments of the British Isles. Journal of Quaternary Science 24: 785–801.

Trotter FM. 1929. The glaciation of Eastern Edenside, the Alston Block, and the Carlisle Plain. Quarterly Journal of the Geological Society 85: 549–612 [DOI: 10.1144/QJGJS.1929.085.01-04.17].

Wilson P, Lord T. 2014. Towards a robust deglacial chronology for the northwest England sector of the last British-Irish Ice Sheet. North West Geography 14: 1–11.

Wilson P, Lord T, Rodes A. 2013a. Deglaciation of the eastern Cumbria glacikarst, northwest England, as determined by cosmogenic nuclide (10Be) surface exposure dating, and the pattern and significance of subsequent environmental changes. Cave and Karst Science 40: 22–27.

Wilson P, Schnabel C, Wilcken KM et al. 2013b. Surface exposure dating (39Cl and 10Be) of post-Last Glacial Maximum valley moraines, Lake District, northwest England: some issues and implications. Journal of Quaternary Science 28: 379–390 [DOI: 10.1002/jqs.2628].

Xu S, Dougars AB, Freeman SPHT et al. 2010. Improved Be-10 and Al-26 AMS with a 5 MV spectrometer. Nuclear Instruments and Methods B 268: 736–738.

Yorke L, Rumsby BT, Chiverrell RC. 2012. Depositional history of the Tyne valley associated with retreat and stagnation of Late Devensian Ice Streams. Proceedings of the Geologists’ Association 123: 608–625 [DOI: 10.1016/j.pgeola.2012.03.005].