The evolution of the terrestrial-terminating Irish Sea glacier during the last glaciation
Chiverrell, Richard Christopher; Thomas, Geoff Stephen Powell; Burke, Matthew; Medialdea, Alicia; Smedley, Rachel; Bateman, Mark; Clark, Chris; Duller, Geoffrey A.T.; Fabel, Derek; Jenkins, Geraint; Ou, Xianjiao; Roberts, Helen Marie; Scourse, James

Published in:
Journal of Quaternary Science
DOI:
10.1002/jqs.3229
Publication date:
2021

Citation for published version (APA):
Chiverrell, R. C., Thomas, G. S. P., Burke, M., Medialdea, A., Smedley, R., Bateman, M., Clark, C., Duller, G. A. T., Fabel, D., Jenkins, G., Ou, X., Roberts, H. M., & Scourse, J. (2021). The evolution of the terrestrial-terminating Irish Sea glacier during the last glaciation. Journal of Quaternary Science, 36(5), 752-779. https://doi.org/10.1002/jqs.3229
The evolution of the terrestrial-terminating Irish Sea glacier during the last glaciation

RICHARD CHRISTOPHER CHIVERRELL,1*, GEOFF STEPHEN POWELL THOMAS,1 MATTHEW BURKE,1 ALICIA MEDIALDEA,2 RACHEL SMEDLEY,1 MARK BATEMAN,3 CHRIS CLARK,3 GEOFFREY A. T. DULLER,4 DEREK FABEL,5 GERAINT JENKINS,4 XIANGJIAO OU,4 HELEN MARIE ROBERTS4 and JAMES SCOURSE6

1Geography and Planning, University of Liverpool, Liverpool, Merseyside, UK
2University of Cologne, Institute of Geography, Köln, Nordrhein-Westfalen, Germany
3Department of Geography, University of Sheffield, Sheffield, UK
4Department of Geography and Earth Sciences, Aberystwyth University, Aberystwyth, Ceredigion, UK
5Scottish Universities Environmental Research Centre, SUERC AMS Laboratory, East Kilbride, South Lanarkshire, UK
6Centre for Geography and Environmental Science, University of Exeter, Penryn, Cornwall, UK

ABSTRACT: Here we reconstruct the last advance to maximum limits and retreat of the Irish Sea Glacier (ISG), the only land-terminating ice lobe of the western British Irish Ice Sheet. A series of reverse bedrock slopes rendered proglacial lakes endemic, forming time-transgressive moraine- and bedrock-dammed basins that evolved with ice marginal retreat. Combining, for the first time on glacial sediments, optically stimulated luminescence (OSL) bleaching profiles for cobbles with single grain and small aliquot OSL measurements on sands, has produced a coherent chronology from these heterogeneously bleached samples. This chronology constrains what is globally an early build-up of ice during late Marine Isotope Stage 3 and Greenland Stadial (GS) 5, with ice margins reaching south Lancashire by 30 ± 1.2 ka, followed by a 120-km advance at 28.3 ± 1.4 ka reaching its 26.5 ± 1.1 ka maximum extent during GS-3. Early retreat during GS-3 reflects piracy of ice sources shared with the Irish-Sea Ice Stream (ISIS), starving the ISG. With ISG retreat, an opportunistic readvance of Welsh ice during GS-2 rode over the ISG moraines occupying the space vacated, with ice margins oscillating within a substantial glacial over-deepening. Our geomorphological chronosequence shows a glacial system forced by climate but mediated by piracy of ice sources shared with the ISIS, changing flow regimes and fronting environments.

© 2020 The Authors. Journal of Quaternary Science Published by John Wiley & Sons Ltd.

KEYWORDS: British-Irish ice sheet; deglaciation; geomorphology; glacial lakes; luminescence dating

Introduction

The eastern sector of ice masses located in the Irish Sea Basin (ISB) is unusual in the former British-Irish Ice Sheet (BIIS) in that it comprises an ice mass generated on land, passing offshore, and then flowing back onto land to maximum limits in the English Midlands (Fig. 1) (Thomas, 1989, 2005; Chiverrell and Thomas, 2010). This Irish Sea Glacier (ISG) is the only land-based terminus of the western outlets of the former BIIS, and unambiguous evidence (sensu Stokes and Clark, 1999; Stokes, 2018) for ice streaming is lacking. Repeated and complex ice-marginal oscillations during deglaciation characterize the retreat dynamics of the adjacent and linked marine-terminating western extension of this ice mass called the Irish Sea Ice Stream (McCabe et al., 1998; Thomas et al., 2004; Thomas and Chiverrell, 2007; McCabe, 2008; Livingstone et al., 2010, 2012; Smedley et al., 2017a, 2017b; Chiverrell et al., 2018). The timing in which many glaciers and ice sheets reached their maximum extent in the last glacial cycle was out of phase, but broadly form a time window 27.5–19 ka termed the ‘Last Glacial Maximum’ (LGM) (Clark et al., 2009; Hughes et al., 2013; Hughes and Gibbard, 2015; Palacios et al., 2020) that coincides with the time of minimum sea levels (Lambeck and Purcell, 2001; Lambeck et al., 2014).

At the LGM in north-west England, the ISG flowed into the Cheshire-Shropshire lowlands and 125 km south to maximum limits in the English Midlands (Thomas, 2005; Worsley, 2005). The sediment landform assemblages in Cheshire-Shropshire have received less attention (Johnson, 1971; Morgan, 1973; Thomas, 1985b, 1989, 2005; Worsley, 2005; Parkes et al., 2009) than other sectors of the ice sheet. Further north, recent comprehensive mapping (Chiverrell et al., 2016) in lowland Lancashire shows that the landform signature reflects the uncoupling and realignment of adjacent and increasingly separate ice lobes that developed during deglaciation between 21 and 17.3 ka, showing the interaction of a dominant ISG with smaller ice masses draining eastern Cumbria and the Pennines. The geochronological data published to date suggest that rates of ice-marginal retreat in the eastern ISB were slower than the Irish Sea Ice Stream (ISIS) to the west (Chiverrell et al., 2013, 2016, 2018; Smedley et al., 2017a; Small et al., 2018; Scourse et al., 2019). These differences in the pace of retreat have been attributed to the absence of ice streaming and the terrestrial nature of the eastern terminating ice lobe (Chiverrell et al., 2016).

Here we:

1. Reconstruct the patterns of ice-marginal retreat using new comprehensive ice-marginal landform and sediment-landform
data, including a reassessment of the distribution and nature of extensive proglacial lakes.

2. Define the primary glacial depositional environments and interpret the time–space sequence of ice dynamics using the sediment–landform assemblages.

3. Constrain, using a new optically stimulated luminescence (OSL) chronology, both the timing and the pace of deglaciation from LGM extent for this non‐marine terminating margin and discuss its significance.

4. Evaluate the controls over ice dynamics during >200 km of ice‐marginal retreat and the interaction with ubiquitous proglacial palaeo‐lakes during that retreat.

Geological setting and topographical context

Glaciers have repeatedly invaded the lowlands of north-west England during the Pleistocene and at the LGM they extended south into the English Midlands (Fig. 1B). The region has been divided into nine zones to describe the glacial sediment landform assemblages and interpret the pattern and timing of deglaciation from LGM extent for this non-marine terminating margin and discuss its significance.

Materials and methods

Geomorphology

The glacial geomorphology of the Cheshire–Shropshire–Staffordshire lowlands (12,000 km²) (Fig. 1C) was mapped (Supporting Information Data 1) using the NEXTMap Great Britain digital terrain model following approaches identical to Chiverrell et al. (2016). These new data significantly extend the 4300-km² glacial geomorphological database for Lancashire (NW England) presented by Chiverrell et al. (2016). British Geological Survey
(BGS) digital solid and drift mapping was used to delimit bedrock at the surface. Geomorphological mapping followed the following process steps: (i) identifying the primary breaks in slope together with symbols for largely non-genetic landform geometries such as flats, ridges, mounds and basins; and (ii) interpretation of these forms using sediment–landform associations to produce a layered geomorphological ARCGIS geodatabase. Ground-truthing of the mapping was carried out in selected areas by conventional field investigation. Interpretation of the geomorphology follows conventional typologies for glacial landforms (e.g. Evans, 2003), and was underpinned by both the landform morphology and associated sediments.

**Stratigraphy and sediment landform assemblages**

Exposures of the glacial stratigraphy are limited across the region with sand and gravel quarries providing the main means of accessing sedimentary data. Sediment sequences were described at Bridgwater, Bridgnorth, Seisdon (Wolverhampton), Condover (Shrewsbury), Wood Lane (Evesham), Borras (Wrexham), Cherry Orchard (Delamere), Lydiate Lane (Chorley) and Bradley’s (Preston) quarries (Fig. 1B). Exposures were logged using field sketches, vertical lithofacies logs and photo-mentals following standard procedures (Evans and Benn, 2004; Thomas et al., 2004). Other characteristics recorded included textural classifications, sorting and grain size, palaeocurrents or till fabric indicators, sedimentary structures, nature of contacts and the lithofacies. The exposures at each of these locations were sampled for materials for OSL dating. Borehole data from over 89 000 locations were assessed using the BGS geodatabase, and this was reduced to 56 000 sites by excluding confidential, shallow (<3 m) or uncertain location boreholes. In general, these borehole records were used to corroborate the sediment–landform interpretations from the geomorphological mapping. For boreholes distributed across important geomorphological features, the records were classified from the descriptions into gravel, sand and gravel, sand, silt/clay, laminated silt/clay, and diamictics, and used to construct stratigraphical cross-sections. Ground elevations of the sections were interrogated from the NEXTMap digital elevation model and cross-checked against the borehole reports (Thomas, 1985a, 1989; Chiverrell et al., 2008, 2016). Uncertainties in the interpretation of cross-sections arise from the inherent vertical and lateral variability in glacial sequences, the limited continuity of lithological markers between boreholes and uncertain origin for breaks in the borehole records (e.g. tectonic or cut and fill). However, combining the regional geomorphology with the borehole cross-sections gives us some confidence in these interpretations.

**Dating the retreat of the Irish Sea lobe**

**Geochronological strategy and field sampling**

Samples for OSL dating were collected from eight sites targeting glacioluvial and deltaic outwash sands and gravels, which were selected based on their ice-proximal context and potential to date well-defined ice margins (Fig. 1B; Table 1). The principle underpinning OSL dating is that exposure to sunlight zeros or bleaches an OSL signal that develops within mineral grains underpinning OSL dating is that exposure to sunlight zeros or bleaches an OSL signal that develops within mineral grains. For OSL dating of sand-sized grains the measurement of multiple replicates (typically here ~50) can also identify those grains in a heterogeneously bleached sample that were exposed to sunlight most recently (typically referred to as being well bleached), where statistical models are required to determine an accurate age, e.g. the Minimum Age Model (MAM) (Galbraith et al., 1999). All these luminescence methods date the last exposure of the minerals to daylight, and it is assumed that this relates to the last depositional cycle. For the sand samples from Seisdon, Bridgwater, Condover, Wood Lane, Borras, Cherry Orchard, Lydiate Lane and Bradley’s quarries (Fig. 1B), opaque tubes were hammered into sedimentary sections to prevent exposure to sunlight during sampling. At Bridgwater Quarry and Lydiate Lane, granite cobbles were obtained from gravel-dominated lithofacies. Cobbles were sampled into light-tight bags under dark conditions using an opaque black sheet, targeting cobbles with b-axes varying from ~10 to 3 cm and recording the clast orientation by marking the upper and lower surfaces (Jenkins et al., 2018). The sediment matrix around the cobbles was collected for dosimetry.

**Optically stimulated luminescence dating**

The sample preparation and analysis methods used were identical to existing studies for SG-OSL dating (Smalley et al., 2017a, 2017b), SA-OSL dating (Evans et al., 2017) and cobble dating (Jenkins et al., 2018). For all samples, the external gamma dose rates were determined using in situ gamma spectrometry. For sand-sized samples, external beta dose rates were calculated from U, Th, K and Rb concentrations that were determined using inductively coupled plasma mass spectrometry (ICP-MS) and inductively coupled plasma atomic emission spectroscopy (ICP-AES). For the cobbles, thick source beta counting using a Rise GM-25-5 instrument (Better-Jensen and Mejdahl, 1988) was used to determine the external beta dose. The dose rate to individual rock slices originates additionally from radionuclides within the cobble. To measure this, a sample of each cobble and a sample of the surrounding matrix were milled to a fine powder for dosimetry measurements (Jenkins et al., 2018). The dose rate to feldspar grains in the cobble also included an internal beta dose contribution arising from the K-content of the grains (assumed to be 12.5 ± 0.5% as per Huntley and Baril, 1997). Appropriate conversion factors (Guérin et al., 2011, 2012) were applied to calculate the final total dose rate (Table 1) including grain size. In terms of palaeoamplitude attenuation, all the sampled sites have water tables that are presently artificially low, owing to sand and gravel extraction and all have water management strategies (e.g. settling lagoons) indicative of higher natural water tables. Maximum pore spaces in 180–250-µm sand are in the range between rhombohedral (26%) and random (40%) packing, which for moderately sorted rounded to sub-rounded sands equates to saturated water contents of around 30%. The inter-annual pattern in the water table levels probably shows a summer draw-down of 2–3 m and in winter tables varying within 8–4 m of the ground surface for three sites in NW England with Permo-Triassic sandstone bedrock overlain by glacial sediments. This is probably representative of post-glacial water tables at these sites (BGS: http://www.bgs.ac.uk/research/groundwater/datainfo/NWRA.html). On this basis, for sediment samples moisture contents of 23 ± 5% were used for shallow and drier samples and 27 ± 5% for deeper saturated samples. In this study, for rock slices more than 1.5 mm below cobble
| Site          | Sample | Depth (m) | Water (%) | U (p.p.m.)* | Th (p.p.m.)* | K (%)* | Rb (p.p.m.)* | Beta dose rate (Gy ka$^{-1}$) | Gamma dose rate (Gy ka$^{-1}$) | Cosmic dose rate (Gy ka$^{-1}$) | Total dose rate (Gy ka$^{-1}$) |
|--------------|--------|-----------|-----------|-------------|-------------|--------|-------------|-----------------------------|-------------------------------|---------------------------------|-------------------------------|
| Seisdon      | T3SEIS01 | 20        | 23        | 1.10        | 3.8         | 1.6    | 56.0        | 1.02 ± 0.12                 | 0.56 ± 0.04                   | 0.06 ± 0.01                    | 1.66 ± 0.12                   |
|              | T3SEIS02 | 20        | 23        | 1.08        | 3.3         | 1.8    | 55.1        | 1.15 ± 0.10                 | 0.43 ± 0.03                   | 0.03 ± 0.00                    | 1.64 ± 0.10                   |
| Bridgwalton  | T3BRID02 | 6         | 23        | 0.70        | 2.7         | 1.1    | 43.6        | 0.70 ± 0.06                 | 0.46 ± 0.03                   | 0.10 ± 0.01                    | 1.24 ± 0.07                   |
|              | T3BRID03 | 40        | 23        | 0.70        | 2.7         | 1.1    | 43.4        | 0.70 ± 0.06                 | 0.42 ± 0.03                   | 0.07 ± 0.00                    | 1.20 ± 0.07                   |
|              | T3BRID04 | 3         | 23        | 0.89        | 3.7         | 1.1    | 46.9        | 0.73 ± 0.06                 | 0.50 ± 0.03                   | 0.14 ± 0.01                    | 1.39 ± 0.07                   |
|              | T3BRID05 | 3.5       | 23        | 0.73        | 2.7         | 1.0    | 41.3        | 0.65 ± 0.06                 | 0.40 ± 0.03                   | 0.13 ± 0.01                    | 1.19 ± 0.06                   |
|              | T3BRID06 | 5         | n/a       | 2.23        | 6.8         | 3.9    | –           | 4.58 ± 0.88                 | 0.70 ± 0.03                   | 0.12 ± 0.06                    | 5.44 ± 0.88†                  |
|              | T3BRID07 | 3         | 27        | 0.94        | 3.6         | 1.0    | 42.2        | 0.68 ± 0.04                 | 0.58 ± 0.04                   | 0.11 ± 0.01                    | 1.38 ± 0.06                   |
| Condover     | T3COND02 | 6.5       | 23        | 0.90        | 3.9         | 1.0    | 43.1        | 0.68 ± 0.06                 | 0.40 ± 0.03                   | 0.09 ± 0.01                    | 1.20 ± 0.06                   |
|              | T3COND04 | 1.5       | 23        | 0.70        | 2.7         | 1.0    | 42.8        | 0.64 ± 0.06                 | 0.37 ± 0.02                   | 0.17 ± 0.01                    | 1.21 ± 0.06                   |
|              | T3COND05 | 2.8       | 23        | 0.66        | 2.3         | 1.0    | 38.8        | 0.63 ± 0.06                 | 0.50 ± 0.03                   | 0.13 ± 0.01                    | 1.14 ± 0.06                   |
| Wood Lane    | T3WOOD04 | 14        | 23        | 0.92        | 3.0         | 1.2    | 46.7        | 0.77 ± 0.07                 | 0.42 ± 0.02                   | 0.05 ± 0.02                    | 1.26 ± 0.07                   |
|              | T3WOOD05 | 15        | 23        | 0.80        | 2.7         | 1.2    | 46.9        | 0.76 ± 0.07                 | 0.47 ± 0.03                   | 0.03 ± 0.00                    | 1.29 ± 0.07                   |
| Cherry Orchard Farm | T3COF03 | 4         | 23        | 0.71        | 2.6         | 1.0    | 39.8        | 0.68 ± 0.07                 | 0.40 ± 0.03                   | 0.12 ± 0.01                    | 1.22 ± 0.07                   |
|              | T3COF04 | 3         | 23        | 0.60        | 2.1         | 1.0    | 37.9        | 0.67 ± 0.07                 | 0.31 ± 0.02                   | 0.14 ± 0.01                    | 1.13 ± 0.07                   |
| Borras       | T3BORR01 | 11        | 27        | 1.15        | 3.4         | 1.0    | 43.2        | 0.65 ± 0.05                 | 0.38 ± 0.03                   | 0.06 ± 0.00                    | 1.11 ± 0.06                   |
|              | T3BORR02 | 5         | 27        | 0.93        | 3.0         | 1.3    | 39.3        | 0.73 ± 0.06                 | 0.39 ± 0.03                   | 0.11 ± 0.01                    | 1.25 ± 0.07                   |
| Dingle       | T3DING01 | 1.5       | 15        | 0.59        | 1.7         | 0.9    | 35.5        | 0.63 ± 0.06                 | 0.44 ± 0.02                   | 0.17 ± 0.01                    | 1.26 ± 0.06                   |
|              | T3DING02 | 1.5       | 15        | 0.66        | 2.5         | 1.3    | 51.3        | 0.90 ± 0.08                 | 0.41 ± 0.02                   | 0.17 ± 0.01                    | 1.49 ± 0.08                   |
| Lydiate Lane | T3LYL04 | 5         | 23        | 0.68        | 2.4         | 1.4    | 50.7        | 0.85 ± 0.08                 | 0.43 ± 0.03                   | 0.11 ± 0.01                    | 1.41 ± 0.08                   |
|              | T3LYL05 | 5.3       | 23        | 0.68        | 2.5         | 1.5    | 58.5        | 0.91 ± 0.09                 | 0.42 ± 0.03                   | 0.11 ± 0.01                    | 1.45 ± 0.09                   |
|              | T3LYL06 | 4.0       | 23        | 1.22        | 2.5         | 0.9    | –           | 0.66 ± 0.03                 | 0.38 ± 0.03                   | 0.12 ± 0.01                    | 1.18 ± 0.04                   |
|              | T3LLD-04† | 2.5 | n/a | 3.92 | 7.2 | 4.5 | – | 4.75 ± 1.01 | 0.84 ± 0.03 | 0.15 ± 0.04 | 5.80 ± 1.02 |
|              | T3LLD-09† | 2.5 | n/a | 3.39 | 6.7 | 4.5 | – | 5.47 ± 0.73 | 0.76 ± 0.04 | 0.15 ± 0.04 | 6.43 ± 0.74 |
| Bradleys     | T3BRAD03 | 5         | 23        | 0.72        | 2.5         | 1.6    | 58.2        | 0.96 ± 0.09                 | 0.48 ± 0.03                   | 0.11 ± 0.01                    | 1.56 ± 0.09                   |
|              | T3BRAD04 | 5         | 23        | 0.70        | 2.5         | 1.6    | 60.1        | 0.97 ± 0.09                 | 0.47 ± 0.03                   | 0.11 ± 0.01                    | 1.56 ± 0.09                   |
|              | T3BRAD06 | 4.5       | 23        | 1.23        | 1.9         | 1.1    | –           | 0.76 ± 0.04                 | 0.40 ± 0.02                   | 0.12 ± 0.01                    | 1.29 ± 0.05                   |

*The analytical chemistry laboratory did not provide uncertainties on individual U, Th, K or Rb concentrations. Based on replicate analyses, uncertainties of 10% were assumed for U, Th and Rb, and 5% for K, and these uncertainties were propagated through the dose rate calculations.

†T3BRID06, T3LLD-04 and T3LLD-09 were cobbles. Geochemical data shown here are for the cobbles themselves, which dominates the dose rate beyond a depth of ~1.5 mm into the cobbles. The dose rate varies slightly with depth. The dose rate values given here are the doses delivered to grains of K-feldspar averaged over the section of the cobbles thought to have been bleached at deposition. The depth to which cobbles were bleached is given in Table 2.
surfaces, ~90% of the dose rate comes from the cobble itself and not from the surrounding sediment matrix, making the dose rate effectively independent of the water content of the surrounding sediment; this is one of the benefits of deriving OSL ages from cobbles.

OSL analyses of all sand quartz samples were performed on grain sizes within the range of 180–250 μm (Table 2), with SA-OSL using aliquots each containing ~20 grains. Given that the proportion of quartz grains emitting an OSL signal for the samples in this study was small, it was thought likely that the SA-OSL signal was dominated by few grains. This has been shown to be the case elsewhere (e.g. Evans et al., 2017). The equivalent dose ($D_e$) distributions determined using SG-OSL and SA-OSL are shown in Fig. 2. All SG-OSL and SA-OSL $D_e$ distributions were asymmetrically distributed with high overdispersion (OD; Table 2). As a result, it would appear likely that all samples had been heterogeneously bleached before burial. MAMs were therefore used to identify the well-bleached component of the heterogeneously bleached $D_e$ distribution to determine the $D_e$ values used for age calculation (Table 2). For SG-OSL the MAM of Galbraith and Laslett (1993) was used with $\sigma_a$ determined for each sample by combining in quadrature the intrinsic OD determined from dose recovery experiments with the extrinsic scatter arising from external microdosimetry (~20%) (see Smedley et al., 2017b, 2020). Due to averaging effects of potentially measuring 2–3 bright grains per aliquot using SA-OSL, it was more difficult to characterize the well-bleached component of the heterogeneously bleached $D_e$ distribution. Final $D_e$ values for age calculation using the SA-OSL were calculated using the internal-external uncertainty (IEU) model (Thomsen et al., 2007) with the parameters $a$ and $b$ used in the model determined from dose recovery tests (calculating the OD of the dose distribution at multiple given doses) for each site. Such an approach has been applied successfully to glacial sediments elsewhere in the BIIS (e.g. Bateman et al., 2018). Inter-comparison of SG-OSL with SA-OSL methods (Fig. 2) when applied to the same samples show similar ages for samples T3COND04 (SG-OSL 35.1 ± 6.8 ka, SA-OSL 34.3 ± 4.3 ka), T3BORR01 (SG-OSL 23.7 ± 3.2 ka, SA-OSL 22.9 ± 2.3 ka) and T3COFO4 (SG-OSL 27.6 ± 4.2 ka, SA-OSL 29.1 ± 4.3 ka). However, the results for the two methods diverge for T3SEIS02 (SG-OSL 48.8 ± 6.4 ka, SA-OSL 22.9 ± 3.4 ka) and T3WOOD05 (SG-OSL 29.2 ± 4.3 ka, SA-OSL 19.8 ± 2.6 ka), which is potentially due to a lack of data characterizing the $D_e$ distribution using SG-OSL ($n$ = 44 grains for T3WOOD05 and $n$ = 41 grains for T3SEIS02).

When using cobbles for luminescence dating, rock cores (either 7 or 8 mm in diameter and between 10 and 20 mm deep) were obtained from the upper face of cobbles under subdued red light. The cores were then cut, producing rock slices of 0.7 mm (Bridgwater) or 0.4 mm (Lydiate Lane) thickness using a water-cooled diamond-edged wafering blade. The clean and dry rock slices were placed into steel planchettes for IRSL measurements. To screen rapidly whether the cobbles had been exposed to light before deposition, a single surface slice was taken and the $L_0/T_0$ ratio was measured, the ratio of the natural ($L_0$) and an applied test ($T_0$) dose (~34 Gy at Bridgwater and ~42 Gy at Lydiate Lane), for both the IRSL$_{50}$ and the post-IR IRSL$_{225}$ signals using a single cycle of the measurement procedure (Jenkins et al., 2018). At Bridgwater, out of 21 cobbles sampled, only one granite cobble yielded evidence of bleaching at deposition and subsequent measurements focused on that cobble, while at Lydiate Lane two of the 12 cobbles measured appeared to have been bleached. For the cobble from Bridgwater, dose recovery measurements were performed on rock slices (Jenkins et al., 2018) using a SOL-2 solar simulator for 7 days to reset any remaining residual signal. The six rock slices used for dose recovery measurements (three used for residual dose measurements and three given a known dose of 85.6 Gy) were taken from the same cobble as those rock slices that were used to determine an age. Measurements show that the IRSL$_{50}$ signal gives a residual—subtracted dose recovery ratio of 0.90 ± 0.02 and the post-IR IRSL$_{225}$ signal a ratio of 0.97 ± 0.03, confirming the suitability of this measurement protocol. Anomalous fading can be a significant issue when dating using feldspars (e.g. Huntley and Lamotte, 2001). Measurements of fading were undertaken on four or five different rock slices for each cobble. Rock slices were irradiated (43 or 150 Gy, Table 3) and preheated before storage for periods up to 1 month to allow calculation of $g$-values (Auclair et al., 2003). The average $g$-values obtained were typically 1.4% per decade for the IRSL$_{50}$ signal, and 0.7% per decade for the post-IR IRSL$_{225}$ signal. Given the similarity of the fading rate measured for the IRSL$_{50}$ signal and the apparent quartz OSL fading rates (1.3 ± 0.3% per decade) (Thiel et al., 2011), no correction has been made here for fading of either the IRSL$_{50}$ or the post-IR IRSL$_{225}$ signals.

A single aliquot regenerative dose (SAR) protocol modified for use on rock slices (Jenkins et al., 2018) was used to generate $D_e$ values using both the IRSL$_{50}$ signal and the post-IR IRSL$_{225}$ signal. The changes in $L_0/T_0$ and $D_e$ with depth were measured (Fig. 3). For three cobbles (T3BRD06, T3LL1D-04 and T3LL1D-09) the data confirm that the outermost part was well-bleached, with low IRSL$_{50}$ and post-IR IRSL$_{225}$ $L_0/T_0$ ratios and $D_e$ values at the surface gradually increasing at depth. The $D_e$ values from slices that had been reset at deposition were then divided by the dose rate to give the age (Fig. 3; Table 2). For T3BRD06 the average age of the upper seven rock slices for the IRSL$_{50}$ signal is 25.3 ± 1.6 ka. The post-IR IRSL$_{225}$ ages (29.3 ± 2.4 ka, Fig. 3a) are consistently greater than the IRSL$_{50}$ ages and this has also been observed in other cobble-dating studies (Freiesleben et al., 2015; Jenkins et al., 2018). It is unclear, at present, why there is a difference between the two signals, but the IRSL$_{50}$ ages are used here given that they are thought to be more reliable based on the agreement of IRSL$_{50}$ ages from cobbles and independent age control at Orrisdale Head, Isle of Man (Jenkins et al., 2018). At Lydiate Lane (Fig. 3b,c) the post-IR IRSL$_{225}$ signal does not appear to have been bleached at the surface for either of the cobbles. The post-IR IRSL$_{225}$ signal is known to bleach much more slowly in daylight than the IRSL$_{50}$ signal (Colarossi et al., 2015; Ou et al., 2018). For two of the cobbles (T3LL1D-04 and T3LL1D-09) the IRSL$_{50}$ signal does appear to have been reset in the uppermost three slices and seven slices, giving ages of 30.2 ± 1.1 and 30.3 ± 1.4 ka, respectively (Fig. 3b,c).

Bayesian age modelling

The deglaciation sequence interpreted from the geomorphology provides here a prior model (i.e. the hypothetical ‘relative-order’ of events) for Bayesian modelling of the geochronological determinations (e.g. Chiverrell et al., 2013). This Bayesian prior model was developed independently of the age information and included all geochronological samples (Brick Ramsey, 2008, 2009a, 2009b; Brick Ramsey and Lee, 2013; Buck et al., 1996). The prior model here covers the ice-marginal retreat from maximum limits in the English Midlands through well-defined ice margin configurations in lowland Shropshire, Cheshire and Lancashire (Chiverrell et al., 2016). The Bayesian modelling used a uniform phase sequence model punctuated by boundaries and was coded using OxCal 4.3 (Bronk Ramsey and Lee, 2013). The approach uses Markov chain Monte Carlo (MCMC) sampling to build up a distribution of possible solutions, generating a probability
Table 2. Luminescence equivalent dose and age data.

| Site              | Sample          | Labcode | Analysis | Grain size (μm) | DR OD (%) | Total analysed* | n* | OD (%) | Age model | a value or Sigma b | Dₑ (Gy) | Age (ka) |
|-------------------|-----------------|---------|----------|-----------------|------------|-----------------|----|---------|------------|-----------------|----------|----------|
| Seisdon           | T3SBS01         | Aber-SBS01 | SG       | 180–212         | 6          | 3600            | 61 | 86     | MAM       | 0.20            | 97.4 ± 17.6 | 58.5 ± 11.5 |
|                   | T3SBS02         | Shfd12096  | SA       | 125–180         | 27         | 77              | 36 | 90     | IEU       | 0.30            | 37.5 ± 5.0 | 22.9 ± 4.4  |
| Bridgwater        | T3BRID02        | Shfd16123  | SA       | 212–250         | 80         | 4300            | 41 | 55     | MAM       | 0.30            | 79.8 ± 12.8 | 48.0 ± 8.4  |
|                   | T3BRID03        | Shfd16124  | SA       | 212–250         | 80         | 4300            | 41 | 55     | IEU       | 0.25            | 63.8 ± 3.5 | 50.3 ± 3.9  |
|                   | T3BRID04        | Shfd14029  | SA       | 212–250         | 23         | 121             | 51 | 59     | IEU       | 0.25            | 35.0 ± 5.4 | 25.2 ± 4.1  |
|                   | T3BRID05        | Shfd14030  | SA       | 212–250         | 58         | 40              | 90 | 92     | IEU       | 0.25            | 34.5 ± 3.8 | 29.0 ± 3.5  |
| Bridgwater        | T3BRID06        | Aber-BW103-1 | Cobble | 360–820         | 5.5*       | 7               | n/a | WM     | n/a       | 1.37 ± 0.88 | 25.3 ± 1.68 |
| Condover          | T3COND02        | Shfd16125  | SA       | 212–250         | 9           | 5700            | 54 | 79     | MAM       | 0.20            | 38.2 ± 5.0 | 27.6 ± 3.8  |
|                   | T3COND04        | Shfd14031  | SA       | 212–250         | 14          | 66              | 38 | 56     | IEU       | 0.15            | 23.6 ± 2.7 | 19.7 ± 2.5  |
|                   | T3COND05        | Shfd14032  | SA       | 212–250         | 66          | 44              | 68 | 80     | IEU       | 0.15            | 41.4 ± 4.7 | 34.3 ± 4.3  |
| Wood Lane         | T3WOOD04        | Shfd14033  | SA       | 212–250         | 4           | 50              | 47 | 65     | IEU       | 0.10            | 65.9 ± 7.8 | 57.8 ± 7.8  |
|                   | T3WOOD05        | Shfd14034  | SA       | 212–250         | 8           | 150             | 66 | 66     | IEU       | 0.10            | 25.5 ± 3.1 | 19.8 ± 2.6  |
| Cherry Orchard    | T3COF03         | Aber-T3COF3 | SG       | 180–212         | 6           | 5500            | 72 | 84     | MAM       | 0.20            | 34.1 ± 4.7 | 27.9 ± 4.2  |
|                   | T3COF04         | Aber-T3COF4 | SG       | 180–212         | 6           | 7100            | 67 | 69     | MAM       | 0.20            | 31.1 ± 4.3 | 27.6 ± 4.2  |
|                   | T3COF04         | Shfd-T3COF4 | SG       | 180–212         | 25          | 104             | 64 | 76     | IEU       | 0.20            | 32.8 ± 4.4 | 29.1 ± 4.3  |
| Dingle Bank       | T3DING01        | Shfd14037  | SA       | 212–250         | 4           | 102             | 39 | 36     | FMM       | 0.20            | 15.9 ± 0.9 | 12.6 ± 1.0  |
|                   | T3DING02        | Shfd14038  | SA       | 212–250         | 4           | 103             | 52 | 36     | FMM       | 0.20            | 16.5 ± 0.7 | 11.1 ± 0.8  |
| Lydiate Lane      | T3LYDL04        | Shfd14039  | SA       | 212–250         | 26          | 107             | 56 | 83     | IEU       | 0.25            | 42.8 ± 3.2 | 30.9 ± 5.2  |
|                   | T3LYDL05        | Shfd14040  | SA       | 212–250         | 3           | 305             | 56 | 83     | IEU       | 0.25            | 42.8 ± 3.2 | 30.9 ± 5.2  |
|                   | T3LYDL06        | Aber-T3LYDL06 | SG       | 212–250         | 24          | 6000            | 47 | 65     | MAM       | 0.45            | 67.2 ± 15.1 | 57.2 ± 13.1 |
|                   | T3LLID04        | Aber-L11-D-04 | Cobble | 260–820         | 2.1*        | 6               | n/a | WM     | n/a       | 176 ± 13 | 30.2 ± 11.1 |
|                   | T3LLID09        | Aber-L11-D-09 | Cobble | 430–1150        | 5.0*        | 7               | n/a | WM     | n/a       | 196 ± 11 | 30.3 ± 1.4  |
| Bradleys          | T3BRAD03        | Shfd14041  | SA       | 212–250         | 17          | 112             | 68 | 71     | IEU       | 0.20            | 44.8 ± 2.8 | 28.7 ± 2.5  |
|                   | T3BRAD04        | Shfd14042  | SA       | 212–250         | 9           | 93              | 51 | 86     | IEU       | 0.20            | 39.2 ± 2.7 | 25.2 ± 2.3  |
|                   | T3BRAD06        | Aber-T3BRAD06 | SG       | 212–250         | 19          | 7900            | 125 | 53    | IEU       | 0.30            | 59.0 ± 6.4 | 45.7 ± 5.2  |

*Total analysed is the number of small aliquots or single grains measured for a sample, while the column headed ‘n’ is the number of small aliquots of single grains accepted for Dₑ modelling. For T3BRID06, L11-D-04, -09, which were cobbles, the ‘Total analysed’ is the depth to which the cobble is thought to be bleached in millimetres, and ‘n’ is the number of rock slices measured in this part of the cobble.

†For single grain measurements the minimum age model (MAM) was used, while the internal-external uncertainty (IEU) model was used for single aliquot data. For the cobbles (T3BRID06, L11-D-04 and -09) the weighted mean of the bleached slices was used.

§Where the IEU model was used, the first parameter ‘a’ is given in this column. The second parameter ‘b’ is 1.5 for all samples. For samples analysed using the MAM, the value given here is that for sigma b.

||
Figure 2. Optically stimulated luminescence data. Abanico plots (Dietze et al., 2016) of the $D_e$ values determined for OSL dating applied at (A) Bridgwalton, (B) Seisdon, (C) Condover, (D) Wood Lane, (E) Cherry Orchard Farm, (F) Borras, (G) Lydiate Lane and (H) Bradley’s Quarries. The plots present the $D_e$ distributions in two plots that share a common $z$-axis of $D_e$ values: (i) a bivariate plot where each $D_e$ value is presented in relation to its precision (shown on the $x$-axis, where those more precisely known are plotted to the right) – this is similar to the radial plot commonly used in luminescence dating; and (ii) a univariate plot showing the age frequency distribution of $D_e$ values, which does not give any presentation of the precision of individual $D_e$ values. The grey shading across both plots shows the $D_e$ used in age calculation for each distribution (2σ shown on the $y$-axis). The combination of these two plots aids interpretation of the scatter in the $D_e$ distributions, where samples with a greater range of $D_e$ values on the $z$-axis have larger amounts of scatter in the $D_e$ distribution.
Table 3. Fading measurements for the three cobbles. The number of slices measured for each cobble is given in the column headed ‘n’.

| Site            | Sample         | Given dose (Gy) | n | IRSL_{50} fading rate (%/decade) | Post-IR IRSL_{225} fading rate (%/decade) |
|-----------------|----------------|-----------------|---|----------------------------------|------------------------------------------|
| Bridgwalton     | Aber-T3BRID06  | 43              | 5 | 1.41 ± 0.63                      | 0.67 ± 0.62                              |
| Lydiate Lane    | Aber-T3LL1D-04 | 150             | 4 | 1.48 ± 0.014                     | 0.81 ± 0.26                              |
|                 | Aber-T3LL1D-09 | 150             | 4 | 1.39 ± 0.75                      | 0.73 ± 0.30                              |

called a posterior density estimate, which is the product of both the prior model and the likelihood (measured ages) probabilities for each sample. This approach generates modelled ages for boundaries for a series of ice margin limit positions between the major zones (L1 to L9: Fig. 1B). Each retreat Zone was coded as a Phase and contained grouped dating information for sites that shared a common relationship with the adjacent zones. Phases were delimited by the Boundary command and generated a modelled age probability distribution output for each ice limit (L1, L2, L4, L5, L6, L8 and L9).

The age model uses the new geochronological data alongside three other clusters of geochronological information. Pre-dating the advance of ice in the region, organic sediments at Four Ashes (Staffordshire) have previously been ^14C dated to around 35–34 cal ka BP (Morgan, 1973) and Telfer et al. (2009) used OSL to date loessal deposits in north Lancashire to 27.2 ± 2.6 ka. In the Bayesian model, these ages pre-dating the ice advance are grouped as a Phase at the beginning of the Prior model. Later in the Prior model the ice margins pass north from the region to limits identified and dated in north Lancashire (Telfer et al., 2009), on the Isle of Man and coastal Cumbria (Chiverrell et al., 2018). Deglacial loessal deposits at Warton Crag (north Lancashire) have been OSL dated to 19.3 ± 2.6 ka (Telfer et al., 2009), and ice-marginal outwash deposits at Turfingland, SE Isle of Man, and Gutterby, SW Cumbria, have been SG-OSL dated to 19.2 ± 2 and 21.7 ± 2.6 ka, respectively (Chiverrell et al., 2018). Together these three age estimates form a grouping immediately post-dating the passage northwards of ice margins from the Lancashire and Cheshire region. The ultimate decline of ice from the region is recorded by a series of cosmogenic isotope surface exposure ages from glacialic features and materials, and these date a series of erratic boulders south of Shap in east Cumbria to 17.35 ± 0.5 ka (Wilson et al., 2013), the erratic boulders at Norber in the Pennines to 17.9 ± 1.0 ka (Wilson et al., 2012a; Wilson and Lord, 2014), and boulders from a series of moraines in the Duddon Valley, SW Cumbria, dated to 16.5 ± 0.8 ka (Wilson et al., 2018). The Sequence model was run to assess outliers in time using a Student’s t-distribution (p < 0.05) to describe the outlier distribution and a scaling of 10^3–10^4 years (Bronk Ramsey, 2009b). Obvious outliers were given a probability scaling of p = 1, with ages that fit poorly within the model allocated probabilities of p < 0.25, p < 0.5, p < 0.75 and p < 0.95 on a scale of increasing outlier severity.

Results

Sediments and landforms of the Shropshire–Lancashire lowlands

At the LGM, ice sourced dominantly in southern Scotland and Cumbria converged as a coherent ice mass in the eastern Irish Sea and expanded reaching maximum limits in the English Midlands (Fig. 1B) (Chiverrell and Thomas, 2010; Clark et al., 2012). In the Lancashire–Cheshire-Shropshire lowlands this ISG interacted with glaciers issuing from the Welsh Borderlands to the west (Thomas, 2005) and Pennines to the east (Chiverrell et al., 2016). Ice contact marginal and proglacial landforms dominate the geomorphological record of these lowlands (Supporting Information Data 1), with streamlined glacial bedforms lacking, which reflects either erosion, sediment burial or perhaps a lack of ice-streaming in this sector. The sediment landform assemblages are dominated by moraines, sandur flats and former lake basins (Fig. 1C; Supporting Information Data 1), and these reflect the oscillations and retreat dynamics of the ice margin. The geometry of former ice margins (Fig. 1B) reconstructed from the distribution of ice-marginal landforms shows the progressive unzipping of the ISG from tributary ice masses.

During the extension to LGM limits ice descended over the Wenlock Edge into the Severn, Worfe and Stour Valleys (Zone 1). The terrain is gently undulating and covered by thin and localized glacial sediments. Moraine ridges are scarce, which is interpreted as reflecting that ice-marginal retreat was relatively rapid. The few moraines form a broad arcuate lobe at the maximum extent (Fig. 1C), but are dissected by deeply

![Figure 3 Luminescence ages for three cobbles as a function of depth. The depth to which cobbles were bleached at deposition is shown with vertical solid lines. The same age scale has been used for all three cobbles to aid comparison, and the scale was chosen to show clearly the data which record the most recent exposure to daylight.](https://wileyonlinelibrary.com)
entrenched valleys floored by glacial outwash deposits. Zone 2 is dominated by an extensive basin to the west, with moraines displaying an ENE to WSW orientation (Fig. 1C). The low-lying terrain, palaeo-lake Newport/Buildwas, is floored by laminated to massive muds, sands and gravels, with southward-draining extensive former sandur fans to the north (Zone 4) (Whitehead et al., 1928; Wills, 1948). The most extensive moraines and kame terraces occur south of the former lake, to the north of the Wenlock Edge–Wrekin escarpment (Fig. 1C). Palaeo-lake Newport drained southwards via lake overflows towards the Severn Gorge, the headwaters of the Stour and feeding the Trent in eastern England (Fig. 4) (Wills and Dixon, 1924; Wills, 1948). Zone 3 is an extensive low-lying 6–7-km-wide belt of lobate moraines that reflects the dynamics of Welsh ice. The lobate form suggests relatively unconfined flows of ice most likely after the retreat of the ISG (Wills, 1948; Shaw, 1972a, 1972b), but the cross-cutting axes to some of the moraines show some interaction of Irish Sea and Welsh-sourced ice during this deglaciation (Fig. 1C). Between the Welsh readvance moraines and retreating ice margin lies former Glacial Lake Severn, which received outwash from Welsh ice and Irish Sea ice through a fan-delta on its northern shores (Fig. 5) (Wills, 1948).

There are few depositional glacial landforms on the bedrock high (Zone 4), where Carboniferous strata in the east and Perm-Oriassic sandstones in the west crop out. Through-valleys dissect these uplands, each former conduits for subglacial and proglacial meltwaters feeding rivers south and eastwards. These over-deepened valleys are the only substantial glacigenic landforms, and are floored with sands and gravels deposited as former glacial outwash sandur that drained southwards to Glacial Lake Prees (Fig. 6) (Thomas, 1985). The ISG climbed an 80–100-m high reverse slope in Zone 5, which preserves an extensive lobate belt of moraines that is widest (14–21 km) in the west and narrower (3–5 km) in the east where the bedrock rises towards the Peak District (Peake, 1961; Poole and Whiteman, 1961; Thomas, 1989; Parkes et al., 2009). Three major ice lobes are reflected in the moraine geometry, and show a planform heavily affected by the bedrock topography. Kettlehole lakes and basins are ubiquitous and form a landform signature of ice stagnation in the west reflecting the widespread abandonment of dead-ice within the moraines during deglaciation (Thomas, 1985). The kettleholes and associated dead-ice are a less abundant component of the land-system in the east. In Zone 6 low-lying lobate arcuate moraines record the retreat of four major glacial lobes (Fig. 1C), with the mid-Cheshire bedrock ridge a significant regulator of the
ice-marginal landform geometry. Proglacial lakes formed between the retreating ice margins and the higher terrain to the south, including Glacial Lake Bangor (Thomas, 1985a, 1989, 2005) and others further east: Glacial Lakes Delamere, Wynburnbury and Congleton (Fig. 1C). Former deltas fronting the ISG margins have left extensive raised terraces formed by glacifluvial waters discharging into the lakes around Wrexham and Delamere (Thomas, 1985a, 1989).

South Lancashire (Zone 7) is dominated by an E–W bedrock high reaching <175 m above the surrounding topography and dissected by NW–SE aligned through-channels (Chiverrell et al., 2016). The largest of these channels are occupied by the Dee and Mersey Estuaries (Longworth, 1985), and they were all incised by subglacial and latterly proglacial meltwaters. Low-amplitude moraines in Zone 8 form two major belts that developed fronting the ISG and eastern Cumbrian ice, with further moraines in the lower Ribble Valley fronting a Ribble Valley Glacier (RVG) (Mitchell, 1991; Mitchell and Hughes, 2012; Chiverrell et al., 2016). The low, sub-parallel ridges of the ‘Kirkham Moraine’ (Gresswell, 1967; Chiverrell et al., 2016) in Zone 9 are dissected by N–S proglacial channel systems (Fig. 1B), with further arcuate sub-parallel moraine ridges stepping back northwards recording the decline of ice across north Lancashire (Chiverrell et al., 2016).

### Marginal limits and dynamics

Subglacial bedforms are scarce across the region and so ice flow patterns have been interpreted from the very well-developed moraine record (Fig. 1B). These former ice margins have been subdivided based on the ice source areas, distinguishing the ISG from tributary ice masses sourced from Wales, eastern Cumbria and the Pennines. Major and minor ice-marginal positions were discerned from interpretation of moraine ridge continuity and geometry, with larger and more continuous ridges interpreted as major ice limits and smaller, less continuous ridges forming minor ice limits (Fig. 1B). Major suites of ridges have been identified as ice-marginal limits (L1–9) and interpreted in terms of readvance, oscillation or stand-still of the ISG margins. These limit positions (L1–9: Fig. 1B) have guided our selection of sites for geochronological analysis.

Maximum ice limits are reflected in sporadic moraines, and their small scale and the limited accumulation of proglacial sediments suggests that this advance (L1: Fig. 1B) was either short-lived or not particularly erosive and with no significant standstill or readvance during subsequent retreat (Fig. 1C). Sediment transport pathways to Zone 1 had to overcome the Wenlock Edge – Wrekin ridge with the main outwash conduit through the Severn Gorge. Ice-marginal moraines and kame terraces with a strong E–W continuity occur north of the Wenlock Edge – Wrekin ridge (L2: Fig. 1C), alongside other minor limits also reflecting an E–W configuration. Stepping northwards towards Zone 4 indications of former ice limits are well defined (L3: Fig. 1C), particularly along the Zone 2–3 boundary where Welsh ice moraines are prevalent. These moraines record the step back of the Severn Valley Glacier (SVG) into Wales with a series of arcuate major limits (Fig. 1C). To the north the SVG was constrained by the presence of the ISG, but further south Welsh

Figure 5. (A) Glacial geomorphology of Glacial Lake Severn (Wills, 1948), borehole cross-section lines, former lake shorelines S1 to S3 and location of Condover Quarry. (B) Borehole cross-section extending across former Glacial Lake Severn. [Color figure can be viewed at wileyonlinelibrary.com]
ice was unimpeded and extended as a lobate form (Shaw, 1972a, 1972b). Former Glacial Lake Severn is largely devoid of moraines (Fig. 5), but this may in part reflect burial beneath thick accumulations of lake sediment.

Fronting at least four ISG lobes, the largest moraine complex in the region (L4–L5: Fig. 1C) is compound and comprises multiple ridges. Some of the morphological complexity reflects ice stagnation and disintegration, with ice compressional stresses against a reverse slope encouraging potentially more supraglacial debris to be deposited and then ice stagnation leaving extensive dead-ice buried within the deposits. Further east, the moraines contain fewer kettleholes and geophysical investigations have shown glacitectonism and a more active ice margin in the glacial deposits (Parkes et al., 2009). Thus, variations in the ice-marginal dynamics occurred along the 100-km ice front and between the ice lobes (L4–L5: Fig. 1C). Progressing northwards from L4 to L5 the moraines are typically subdued and often buried by substantial accumulations of outwash and glacialustrine sediment. In places the moraines are more prominent, for example the well-developed ice-marginal moraine system bordering Wales (Thomas, 1985a). The most substantial depositional landforms are the delta terraces around Wrexham in the west (Thomas, 1985a) and Delamere east of the mid-Cheshire ridge. The ice contact slopes to the rear of both deltas were major still-stands of the ice margin (L6: Fig. 1C) during which outwash sands and gravels entered former glacier lake basins.

A series of ice-marginal still-stands has formed a belt of moraines and kames (L7: Fig. 1B) along the break in gradient as ice pulled back into south Lancashire (Chiverrell et al., 2016). Climbing bedrock slopes northwards, these moraines reduce in size and are widely spaced perhaps reflecting more rapid ice retreat (Chiverrell et al., 2016). L8 (Fig. 1B) describes a belt of ice-marginal kames and moraines and reflect the pull-back of ice generating a major limit as ISG and Cumbrian ice margins retreated and the ice masses decoupled, including the Tarleton and Chorley moraines (Price, 1963; Chiverrell et al., 2016). The moraine geometry reflects this decoupling, with moving northwards the ISG moraines increasingly either subdued or now offshore (Chiverrell et al., 2016). The ice-marginal geometries show the interaction of ISG and Cumbrian ice with a confluent RVG. North of the Ribble Valley there is the second largest moraine system (L9: Fig. 1B) in the region, the Kirkham Moraine, a belt of ridges separated by marginal

Figure 6. (A) Glacial geomorphology and borehole cross-section lines of former Glacial Lake Prees, showing the location of probable former ice margins and Wood Lane Quarry in the Oswestry–Whitchurch–Congleton moraine. (B) Borehole cross-section fence sections extending into Glacial Lake Prees. [Color figure can be viewed at wileyonlinelibrary.com]
Glacial lakes have long been postulated in the Cheshire–Shropshire lowlands (Wills and Dixon, 1924; Wills, 1948; Worsley, 1975), and debates about their existence and extent are bound to the concept of the classical ‘Lake Lapworth’ (Wills and Dixon, 1924). Refining understanding of glacial lakes in Cheshire–Shropshire has been through several iterations (Worsley, 1975; Thomas, 1985a, 1989; Worsley, 2005; Murton and Murton, 2012), and includes the interpretation of former glacial lake environments at Preses, in the Dee and Severn Valleys, and in the major L4–L5 and L9 moraine complexes (Fig. 1B). The lack of continuity to their sedimentary record led to them being regarded as small, irregular and disconnected pro- and sub-glacial lakes rather than major regional lakes (Cannell and Harries, 1981; Cannell, 1982; Wilson et al., 1982). The topography provides more restricted opportunity for the development of proglacial lakes in the south, for example Glacial Lake Morville (Wills and Dixon, 1924) forming between ISG ice margins and the Wenlock Edge bedrock escarpment (Fig. 7). This small and temporary proglacial lake was ponded between a N–S aligned ice margin and bedrock slopes to the west. The sediment–landform signature and geochronology of Glacial Lake Morville has been developed using exposures at Bridgwater Quarry (Fig. 7).

Upstream of Ironbridge Gorge (Fig. 4A), more widespread flat ground floored by glacilacustrine sediments evidence an extensive glacial lake, including deliсa ice-proximal deposits that have been attributed to Lake Buildwas and Lake Newport (Wills and Dixon, 1924; Shaw, 1972a, 1972b). Lake Buildwas drained south to the Severn through Ironbridge Gorge and Lake Newport eastwards to the Trent at Gnosall (Fig. 4A), but with ice-marginal retreat these lakes merged and meltwater discharge was solely focused through Ironbridge Gorge and the River Severn (Wills and Dixon, 1924; Wills, 1948; Worsley, 1975). This lake is heavily associated with both the ‘Lake Lapworth’ concept (Wills and Dixon, 1924; Poole and Whiteman, 1961; Worsley, 1975; Thomas, 1989) and the dynamic evolution of drainage thresholds associated with the incision or re-excavation of Ironbridge Gorge. The sediment–landform record is equally complicated, showing a mixture of glacilacustrine silts/clays, outwash sands and gravels, and glacial diamictic formations. Former quarry exposures at Buildwas, near Ironbridge (Fig. 4A) (Wills and Dixon, 1924), showed 30 m of laminated glacilacustrine muds, including diamict layers with intercalated afloat icebergs, over lain by diamict, outwash gravels and sands near the entry to Ironbridge Gorge. Borehole transects (Fig. 4B, from the western edge of the Lake Newport basin northwards, show <20 m thickness of spatially variable glaciogenic sediments overlying bedrock (Hamblin, 1986), with diamictic layers interspersed with sporadic glacilacustrine muds and a dominance of glaci fluvial sands and gravels to the north prograding into the lake basin (Fig. 4). Glacial Lake Newport developed as a relatively shallow basin (30–20 m water depth) and evolved with the retreat of the ice margins. Supraglacial debris and widespread abandonment of dead-ice with retreat of the ice margins is reflected in the kettleholes ubiquitous across the lake floor. The sediment–landform assemblages present a model of relatively shallow and time transgressive lakes, with shallowing driven by a combination of base level incision at Ironbridge and basin infill (Fig. 4B). Further west, the borehole stratigraphy underlying Glacial Lake Severn (Figs 1C and 5) shows thick accumulations of glacilacustrine sediment that evidence a substantial over-deepening terminating at the confluence of Welsh and ISG ice. The basin is filled predominantly by glacilacustrine muds (Fig. 5B), descends >75 m below present sea level and is bounded in the east by the rising rock-head and Welsh readvance moraines (Fig. 5A). The scale of over-deepening (>170 m depth and dimensions of 19 × 6 km) is intriguing and not a feature of the regions’ other former glacial lakes. The sediment fill and process of over-deepening may span more than one glacial cycle, but compressional stresses generated as Welsh ice collided with the ISG compounded by the rising rock-head offer a mechanism for generating intense basal scour and over-deepening. Condover Quarry lies between Glacial Lakes Newport and Severn (Figs 4A and 5A), and the exposures were sampled to date the ice margins associated with these proglacial lake basins and the readvance of Welsh ice.

Glacial Lake Prees (Thomas, 1989) was a lengthy and narrow (20 × 5 km) ice-marginal lake that formed between the L4–L5 ice margin feeding ISG outwash and the Triassic bedrock escarpment to the south (Zone 4: Fig. 6A). Borehole transects show that the lake was divided into separate basins by diamict ridges (Fig. 6B), and these lakes were probably in existence at different times. Lake sediment thicknesses reach 25 m in the deeper (40–30 m water depths) basin and are commonly intercalated with diamict and outwash sands and gravels. Patches of diamictic, overlain by outwash sands and gravels, dominate the northern margins of the lake l ain down initially by ice-contact lacustrine processes and latterly by proglacial outwash prograding into the basin (Fig. 6B). Changes in sediment compositions between adjacent boreholes occur over short distances and reflect the dynamic nature of the glacilacustrine sub-environment. Glacial Lake Prees developed as a long-term system but became proglacial as the ice margins withdrew towards the L4–L5 moraine (Fig. 6). As the ice margins retreated, prograding deltaic and sandur sediments filled the basin. Ultimately, Glacial Lake Prees decoupled from the ISG with ice-marginal retreat northwards (Thomas, 1985a; Thomas, 1989), and set inside the L4–L5 moraines further extensive proglacial lakes developed in the lower Dee and Weaver basins (Fig. 1C). Wood Lane Quarry, near Ellesmere, lies immediately north of Glacial Lake Prees (Fig. 6A), and the exposures were sampled to constrain both retreat from the lake basin and the development of the L4–L5 moraine complex (see ‘Zone 5’ below).

Glacial Lakes Bangor (Thomas, 1989) and Delamere formed as proglacial basins as the ice margins retreated further north developing a series of ice-marginal moraines (Fig. 1C). Extensive deltas form the western margins of Glacial Lakes Bangor (Thomas, 1989) and Delamere. Based on the height of delta top-sets exposed in the Wrexham Delta at Borras Quarry (Thomas et al., 1985), Glacial Lake Bangor was at its maximum 10 km wide and 30 km long, had a maximum lake level of 70 m OD and waters <70–80 m deep were ponded against the L4–L5 moraines and bedrock slope. Glacial Lake Delamere was one of a series of smaller lakes that developed in the Weaver basin with north-westward retreat of the ice margins. The delta complex on the western margins around Delamere had a surface elevation of 85–75 m and water depths probably of 40–60 m. The basin sediments comprise a thick sequence of
lacustrine muds, sands and gravels, which are well exposed owing to several decades of sand and gravel extraction at multiple quarries in the area. Sediment exposures at Borras Quarry (Lake Bangor) and Cherry Orchard Farm (Lake Delamere) were sampled to constrain the L6 ice margins associated with the deltas prograding into these lake basins (see ‘Zone 6’ below).

The topography fronting the retreat of ice margins northwards through Zone 7 is less conducive to the development of proglacial lakes, but between ice limits L8 and L9 temporary lakes developed in front of the retreating ice margins and in the lower Ribble Valley (Fig. 1B) (Earp et al., 1961; Aitkenhead et al., 1992; Chiverrell et al., 2016). The retreat of Cumbrian and RVG ice produced a proglacial lake ponded to the west by the ISG (Chiverrell et al., 2016), and this expanded eastwards with retreat of RVG ice margins up the Ribble Valley (Earp et al., 1961; Aitkenhead et al., 1992). Sediment exposures at Lydiate Lane Quarry (L8 ice margin) and Bradley’s Sandpit (L9 ice margin) (Fig. 1B) were sampled principally to constrain these former ice marginal configurations and bracket the development of proglacial lakes in the lower Ribble Valley (see ‘Zones 8 and 9’ below).
Sedimentology and geochronology

Zone 1: maximum ice limits in the English Midlands

The maximum extent achieved by the ISG in the English Midlands impinged on a small tributary (Mor Brook) of the River Severn, south of Ironbridge Gorge near Bridgnorth (Fig. 1C). East of Mor Brook (Fig. 7) there are low-relief ridges composed of glacial diamict that mark the maximum limit of ice advance. Thick accumulations of outwash sands and gravels have accumulated in the valley in front of this ice limit as ISG ice blocked the valley creating a small ice-dammed lake, Glacial Lake Morville (Fig. 7A). The sand and gravel deposits associated with Glacial Lake Morville have been exploited as mineral aggregate for >50 years. Decades of sand and gravel extraction at Bridgwater Quarry have shown a >15-m sequence of sands and gravels overlying bedrock (Fig. 7B), in which low-angle stacked packages of planar and trough cross-stratified gravels become more sand dominated towards the surface (Fig. 7C). Development of the quarry has progressed from ice-proximal exposures in 2013 (Fig. 7C) more or less contiguously to ice-distal exposures in 2014–2017 (Fig. 7D). Together the stratigraphical sections are dominated by low-angle (~20°) delta-fore-sets sands and gravels dipping and thickening towards the south-east and these are overlain by <2 m of planar massive stratified delta-top set sands. Throughout the sequence, palaeocurrent evidence indicates flows to the south. The stratified medium-coarse sandier units contain granule to pebble gravels and frequent out-sized clasts (Fig. 7C) and are interpreted as typical of delta-fore-set beds deposited as cohesionless debris-flows down the delta front into a shallow lake (Smith and Ashley, 1985; Nemec et al., 1999; Nemec, 2009). The outsized clasts may either be drop-stones or gravity-fed debris rolling down the delta front. The upper 3–2 m (Fig. 7C) comprises planar massive stratified gravels (Gms) with occasional planar cross-stratified gravels (Gp) and pockets of sands (Sh and Sp) becoming gravel dominated with proximity to the former ice margin. Locally, a thin (~2 m) diamict caps the more ice-proximal sections (Fig. 7C), whereas the more ice-distal exposures showed no evidence of this diamictic; it probably reflects a small-scale advance of ice onto the outwash surface. To the south-east, the uppermost planar geometry sheet of laminated gravel and medium-coarse sand (Sh) with some fine laminations (Fl) becomes thicker and is interpreted as a delta top-set (Fig. 7D).

Samples T3BRID02 and T3BRID03 were taken in 2013 from back bar sands in an ice-proximal delta top-set and sampled rippled medium sands (Sr) within planar cross-stratified sands (Sp). In April 2014, new exposures of the more ice-distal delta top-set allowed for sampling of planar geometry sheet laminated medium–coarse sand (Sh) with some fine laminations (Fl) (T3BRID04) and sheet laminated medium sand (Sh) at the fore-set to top-set transition (T3BRID05). These distal exposures were revisited in August 2017 showing a gravel-dominated upper 2–3 m of the delta top-set, which were sampled collecting ~20 cobble-sized gravels and an adjacent sand sample (T3BRID07) was taken from a channel fill in the top-set gravels. SA-OSL analyses focused on four samples (T3BRID02–05). The asymmetrical \( D_0 \) distributions determined for T3BRID02 and T3BRID03 (Fig. 2A) suggest that these samples were bleached heterogeneously before burial, where a small proportion of the grains characterize the minimum dose population due to limited exposure to sunlight before burial. The \( D_0 \) distributions determined for T3BRID04 and T3BRID05 (Fig. 2A) were also asymmetrical and therefore heterogeneously bleached before burial. However, the minimum dose populations in the \( D_0 \) distributions for T3BRID04 and T3BRID05 contained a greater proportion of the grains than those determined for sample T3BRID02 and T3BRID03 (Fig. 2A). T3BRID02 and T3BRID03 were taken from ice-proximal settings with diamictons bedded within and capping the sequence, which suggests relatively short sediment transport distances, with T3BRID04 and T3BRID05 from more ice-distal settings within the delta (Fig. 6D). Samples T3BRID06–07 were taken from the same section as T3BRID04–05, but from more ice-proximal and coarser grained lithofacies. The single-grain \( D_0 \) distributions determined from quartz grains of sample T3BRID07 were asymmetrical and therefore the grains were also heterogeneously bleached before burial. Age–depth profiles determined for the cobble (BWT03–1: T3BRID06) confirm that the outermost parts were well-bleached, with low IRSL50 and post-IR IRSL225 \( L_T/T_0 \) ratios at the surface gradually increasing at a depth of ~6 mm. IRSL50 ages are consistent to a depth of ~7 mm and this supports the interpretation that the cobble had been well-bleached at deposition (Fig. 3).

Seisdon Quarry (Fig. 1C) is located 16 km further east of Bridgwater and is part of a series of quarries that have exploited the Trewyll sands and gravels for >70 years (Morgan, 1973). The bulk of the stratigraphy at the Trewyll–Seisdon quarries spans several marine isotope stages (MIS) including interglacial units, but the uppermost glaci-fluvial outwash sand and gravels containing Irish Sea erratics are unformable on a cryoturbated palaeo-lake surface and potentially date to the last glaciation (Morgan, 1973). The exposures in November 2012 revealed ~10 m of these uppermost orange stratified sands, and two adjacent samples were taken for OSL dating (T3SEIS01 and T3SEIS02) to constrain the timing of the ISG impinging on the River Stour basin within 3 km of the LGM limit south of Wolverhampton (Fig. 1C). Both samples were horizontally laminated sand (Sh) lithofacies within a channel fill. Asymmetrical \( D_0 \) distributions (Fig. 2B) determined by parallel SA-OSL and SG-OSL suggest that both samples were bleached heterogeneously before burial, and only a small proportion of the grains characterize the minimum dose population probably because of limited exposure to sunlight before burial.

The ice-marginal position at Bridgwater constrains the maximum extent of ice in the English Midlands (L1: Fig. 1C). The SA-OSL age determined for T3BRID02 of 50.3 ± 3.9 ka is considered an inaccurate estimate of the time of deposition reflecting poor bleaching of the OSL signal. Therefore, the maximum ice limit at Bridgwater was constrained by the SA-OSL ages of 25.2 ± 4.1 ka (T3BRID04) and 29.0 ± 3.5 ka (T3BRID05), alongside SG-OSL ages of 30.1 ± 8 ka (T3BRID03) and 27.6 ± 3.8 ka (T3BRID07). The cobble (T3BRID06) yielded an age of 25.3 ± 1.6 ka. The geochronology from Seisdon is equivocal, with T3SEIS02 yielding an SG-OSL age of 48.8 ± 8.4 ka, different to the SA-OSL age 22.9 ± 3.4 ka for the same sample. T3SEIS01 also measured for SG-OSL yielded an age of 58 ± 8.3 ka. The two SG-OSL ages overlap within uncertainties and pre-date MIS 2, whereas the SA-OSL age for T3SEIS01 is ~26 ka younger than the paired single grain measurement. It is unclear why the SA-OSL approach should recover a younger minimum dose population than equivalent single grain measurements for the same materials.

Zone 2: Glacial Lakes Newport and Severn

After the retreat of the ISG margins northwards, a readvance by Welsh ice produced a series of lobate moraines that extend as far east as Shrewsbury. These moraines form the boundary between former Glacial Lakes Newport and Severn (Figs 1C, 4 and 5). The sand and gravel workings at Condover have...
exploited a sandur fan issuing from gaps in a Precambrian bedrock ridge into a basin south of Bayston Hill (Fig. 8). A series of E–W aligned moraine ridges show Irish Sea ice rode over this ridge, but SW–NE orientated ridges show a Welsh readvance extending to the ridge crest (Figs 5 and 8A). The quarry workings were limited to steps of 2–3 m in height, and so the vertical sequence was compiled from multiple short-lived exposures (2014–2016). The generalized vertical succession recorded for the quarry revealed repeating cycles of barform massive (Gm) or trough cross-stratified channel (Gt) gravels fining upwards into back-bar parallel-laminated sands (Sh) and planar cross-bedded units (Gp and Sp) (Fig. 8A). The sequence is typical of variable high-discharge and high-energy units, which were probably adjacent to the ice margins given the occasional diamictons elsewhere in the quarry and the extensive post-depositional faulting and numerous kettle basins (Scourse et al., 2009) caused by the melt-out of dead-ice. OSL dating of the sequence has focused on the uppermost 5–8 m and the most recent phase of outwash deposition related probably to the readvance of Welsh ice. T3COND04 sampled cross-stratified back-bar or channel fill sands, whereas T3COND04 and T3COND05 sampled horizontal sheet laminated medium sands between back-bar planar cross-stratified sands (Sp). The \( D_b \) distributions determined for all the samples were asymmetrically distributed (Fig. 2C), which indicates that the sediment was heterogeneously bleached before burial. The geochronology from the uppermost units is equivocal, with T3COND04 yielding paired SG-OSL and SA-OSL ages for the same sample of \( 35.1 \pm 6.8 \) and \( 34.3 \pm 4.3 \) ka, respectively, which overlap within uncertainties, but they are younger than the SA-OSL determination of \( 57.8 \pm 7.8 \) ka measured for the adjacent T3COND05 sample. T3COND02 ~3–4 m lower in the sequence was also measured for SA-OSL yielding an age of \( 19.7 \pm 2.5 \) ka. The older age determinations (T3COND04 and 05) are geologically implausible, pre-dating the advance to maximum limits evidenced further south at Bridgwater. The age of \( 19.7 \pm 2.5 \) ka (T3COND02), although overlapping within the uncertainties, is slightly younger than the cluster of ages from Wood Lane Quarry documenting the retreat of the ISG northwards to L4 (Fig. 1B). However, the T3COND02 result provides potentially the first age estimation for the readvance of Welsh ice late during deglaciation and in that context the deposits at Condover and Wood Lane plausibly could be of similar age. T3COND04 and T3COND05 appear to be clear outliers, but T3COND02 is a good fit constraining the readvance of Welsh ice to \( 19.7 \pm 2.5 \) ka (Fig. 7B) and the establishment of Glacial Lake Severn in the period after \( 19.7 \pm 2.5 \) ka (L3–4: Fig. 1B).

**Zone 5: the Oswestry–Whitchurch–Congleton moraine**

The largest ice-marginal glaciogenic feature in the region is the Oswestry–Whitchurch–Congleton moraine complex (Thomas, 1985a, 1989, 2005), which is classified here as the L4 ISG marginal position (Fig. 1B). The L4 margin was reached after the ISG pulled back from Glacial Lake Prees, and subsequent retreat northwards led to the establishment of large proglacial lakes in the lower River Dee and Weaver basins (Fig. 1C). The geomorphology of the Oswestry–Whitchurch–Congleton moraine is complex and riddled with hollows and lakes formed by the melt-out of dead-ice leaving a chaotic and gently undulating ice disintegration terrain (Fig. 9A) (Shaw, 1972a; Thomas, 1985a, 1989, 2005). The only substantial exposures in the moraine are at Wood Lane Quarry (Fig. 9), with the stratigraphy first described in the 1960s encountering two glacial diamictons that interdigitated with outwash sands and gravels (Shaw, 1972a). The recent exposures...
suggest that aggregate extraction may have removed the upper till as overburden and the vertical sequence 2014–2018 was compiled from multiple sections with aggregate extraction limited to steps of 2–3 m in height through the deposit. A generalized vertical succession (Fig. 9B) shows a laterally discontinuous 2–3 m basal reddish sandy lower diamicton of Irish Sea origin that is interspersed with equivalently red outwash sands and gravels. A 3–4 m thick sequence of trough cross-stratified and planar outwash gravels overlies the lower diamicton, and is interpreted as probable high-energy channel-fill and associated bar-forms that have a more Welsh lithological origin while still relatively rich with Irish Sea erratic clasts. Overlying the gravels is 4.5 m of medium sands including horizontally stratified (Sh), planar cross-stratified (Sp) and rippled (Sr) facies typical of ice-distal proglacial outwash (Fig. 9B). From this unit T3WOOD04 sampled horizontally stratified medium sands and T3WOOD05 more reddish rippled medium sands. The only differences in material were the reddish character to the T3WOOD05 sands, even though the samples were within 0.5 m of each other. The colour reflects dilution of the typically reddish ISG sediments with grey materials of Welsh provenance, and in the sections, there are locally units with a more red and grey character. The sands are capped variably by thin gravels and a 1 m thick weakly consolidated sandy diamicton of Irish Sea Ice origin interpreted as an ice-proximal debris flow into a proglacial lake basin. The diamicot is overlain by <9 m of fining-upwards cycles (cycles 4–11: Fig. 9B) of massive stratified or cross-stratified gravels (Gt and Gms) grading to horizontally bedded, planar cross-stratified (Sp) and rippled sands (Sr) reflecting migration of channels and bars within an active sandur. Faulting and tectonic structures related to collapse and subsidence support the geomorphological evidence for endemic persistence and then melt-out of dead-ice buried within the deposits. T3WOOD04 and T3WOOD05 both display heterogeneously bleached D_e distributions (Fig. 2D), which reflect poor resetting of the OSL signal. The stratigraphy suggests this was an ice-proximal high-energy outwash sequence interspersed with overbank or back bar deposits, providing sediment transport distances limiting opportunity for resetting of the OSL signal. Abanico plots of the SA-OSL D_e distributions show a minimum age population is present comprising better bleached quartz and was better resolved for T3WOOD04 with an age of 21.2 ± 2.9 ka (Fig. 2D). T3WOOD05 underwent parallel SA-OSL and SG-OSL analysis. The SA-OSL analysis returned a similar minimum D_e population, and age 19.8 ± 2.6 ka, as T3WOOD04. The SG-OSL measurements of T3WOOD05 gave an older age (29.2 ± 5.4 ka), although this overlaps at two standard deviations (Fig. 2D; Table 2). The OSL ages for samples T3WOOD04
and T3WOOD05 overlap within uncertainties and date the Oswestry–Whitchurch–Congleton (L4) configuration of the ice margin (Fig. 1C) and the abandonment of Glacial Lake Prees (Figs 6 and 9).

Zone 6: Glacial Lakes Bangor and Delamere in lowland Cheshire

Running parallel to the western margin of the ISG were a series of NE–SW orientated moraines that confine the Mold –Caergwyle ice-marginal sandur (Thomas, 1985a, 1989, 2005); this was the main outwash feeder for the >50-km² Wrexham delta (Fig. 10A) (Thomas, 2005). The eastern frontal slopes to the delta terrace are raised <30 m overlooking the Dee Valley, and since first description (Wedd et al., 1927) the terrace has been identified variably as a delta, alluvial fan and outwash sandur (Peake, 1961, 1979, 1981; Poole and Whiteman, 1961; Wilson et al., 1982; Worsley, 1985). Comprehensive morphological and sedimentological analysis (Thomas, 1985a, 1989, 2005) has revealed areas of ice-disintegration terrain (melt-out basin or pitted sandur), ice front fan, ice-marginal sandur, and a proglacial to ice-contact delta, all of which fed into a large former lake basin called Glacial Lake Bangor (Fig. 10) (Thomas, 1985a, 1989). The serial sections have been devised from exposures and borehole data (Fig. 10A,C). For the delta to function ice is required to be in contact with both the Mold–Caergwyle ice-marginal sandur and the delta slopes immediately to the north of the terrace. Borras Quarry has provided access to exposures towards the distal end of the Wrexham delta (Fig. 10C). A composite >15-m vertical succession was developed in 2014–2016 using a series of quarry faces that were limited in height to 2–3 m (Fig. 10B). T3BORR01 sampled horizontally stratified medium to coarse sands deep within the sequence that appear to be low-angle delta foresets given adjacent units of trough cross-stratified gravel, granule and sand. Within 2–4 m of the delta surface (~70 m O.D.), T3BORR02 sampled horizontally stratified fine to medium sands and T3BORR03 rippled fine to medium sands, both of which were from delta top-set outwash deposits that showed a planar geometry. All three SA-OSL Dₜ distributions were asymmetrical and were therefore probably heterogeneously bleached before deposition (Fig. 2F). T3BORR01 underwent parallel SA-OSL and SG-OSL analyses returning similar minimum Dₜ populations and ages. OSL ages of 22.9 ± 2.3, 23.7 ± 3.2 and 22.9 ± 2.3 ka are tightly clustered, whereas sample T3BORR02 appears older at 29.2 ± 2.8 ka, indicating that this SA-OSL age is an overestimate including Dₜ values potentially from partially bleached grains.

Discharge from the ISG margin between the ice and the mid-Cheshire sandstone ridge has formed an extensive wedge-shaped terrace of ice-contact kame and pitted outwash delta surfaces. This flat terrace extends 10 km as a wedge southward and has a pitted surface with kettlehole basins (e.g. Hatchmere). Bordering in the east by a probable ice-contact slope, the feature resembles the Wrexham delta (Thomas, 1985a) in its morphology (Fig. 1C). Sediments exposed in the former...
Cherry Orchard Farm Quarry (Fig. 11A) have been OSL dated to constrain an ice-marginal configuration straddling the terrace (Fig. 11). The 2013 exposures showed a 10-m-thick sequence of predominantly sand with occasional sand and gravel overlying at depth a diamicton (Fig. 11C). The upper 8 m was dominated by low-angle trough and planar cross-beds of stratified medium–coarse sand with granule to pebble gravels dipping south-east interpreted as delta fore-set beds and deposited as cohesionless debris-flows into a <20–30-m deep proglacial lake (Smith and Ashley, 1985; Nemec et al., 1999; Nemec, 2009). The cross-cutting nature of the cross-sets probably reflects scour and variations in flow direction on the low-angle delta front (Fig. 11C). The depositional environments graded from delta-proximal...
toe-set or bottom-set sands, through stacked units of trough cross-stratified granular sands laid down as cohesionless debris flows in subaqueous channels on the delta front. The units contain occasional out-sized clasts that are probably dropstones. T3COF03 sampled one of the uppermost trough cross-stratified channel fills, whereas T3COF04 sampled horizontally stratified sands probably laid down in relatively shallow water on a low-angle delta fore-set slope near the top of the sequence (Fig. 11D). The three $D_\alpha$ distributions (Fig. 2E) determined for samples from Cherry Orchard farm were asymmetrical and therefore heterogeneously bleached before burial. The $D_\alpha$ distributions characterize the minimum dose population well and so the OSL ages appear to reflect the true burial age. The SG-OSL ages of 27.9 ± 4.2 ka (T4COF3) and 27.6 ± 4.2 ka (T4COF4) and SA-OSL age of 29.1 ± 4.3 ka (T4COF4) date the L6 configuration of the ice margin and timing of Glacial Lake Delamere (Fig. 11B).

Zones 8 and 9: deglaciation of lowland Lancashire

Retreat of the ISG margins northwards (Fig. 1B) into Lancashire led to an increasing influence of differing ice sources (Irish Sea, Cumbria and Ribble Valley) over the configuration of the ice-marginal geomorphology (Fig. 12A). Exposure of the stratigraphy in the region has been scarce since the coastal retreatment of the former cliff sections at Blackpool (De Rance, 1877), and limited to Lydiate Lane and Bradley’s Quarries (Fig. 12A) providing the only substantial glaciogenous exposure as described by Chiverrell et al. (2016). Retreat of Cumbrian ice margins north of the Ribble and the RVG margins eastwards while the ISG was still in a more advanced position led to the establishment of an ice-dammed proglacial lake, with the large debris fan west of Lydiate Lane Quarry (Fig. 12A) formed by flows into that lake basin. The Kirkham Moraine (BL6) is a substantial moraine complex that extends ~30 km E-W across central Lancashire (Fig. 1) (Gresswell, 1967; Chiverrell et al., 2016). Lydiate Lane exploits a bench interpreted as an ice-contact kame terrace, ice limit L8 (Fig. 1B), between the eastern margin of Cumbrian ice and the West Pennines Moors, and comprises a thick sequence of outwash sands and gravels (Chiverrell et al., 2016). Further north, exposures at Bradley’s Quarry exploit an E-W aligned trough sandur, bounded by moraine ridges to the north and south (Fig. 12A). The ridges of the Kirkham Moraine trend west before curving north-west towards the coast and form ice limit L9 (Fig. 1B). The ridges are separated by sandur flats and punctuated by numerous kettle basins showing a significant presence of former dead-ice within the moraine complex.

Lydiate Lane Quarry has evolved rapidly from March 2013 to the present (Fig. 12B), and the working faces have revealed a generalized vertical succession of <15 m of outwash sands and gravels overlain by >2 m of basal diamicton (see Chiverrell et al., 2016). The reddish sand and gravels reflect their derivation from eroded Permo-Triassic sandstones and contain numerous Cumbrian erratic clasts (e.g. Shap granite). The deposits reflect an evolution from an ice-proximal to more distal sandur lithofacies (Thomas et al., 1985), with the upper of these cycles produced by prograding bar-forms and channel migration (Miall, 1977; Thomas et al., 1985). T3LYDL04-6 (Fig. 12C-D) were taken from rippled medium to coarse sand (Sr) between thin units of planar cross-set coarse sand (Sp), with all three samples lain down in relatively low-energy settings, in theory with reasonable opportunities for bleaching. However, the SA-OSL $D_\alpha$ distributions for T3LYDL04-05 and SG-OSL distributions for T3LYDL06 suggest that the samples were heterogeneously bleached (Fig. 2G). The SA-OSL ages of 30.9 ± 5.2 ka (T3LYDL04) and 44.5 ± 3.9 ka (T3LYDL05) and the SG-OSL age of 57.2 ± 13.1 ka (T3LYDL05) in theory constrain the L8 configuration of the ice margin (Fig. 1B). A clearer picture emerges from analysis of two cobbles at Lydiate Lane that yielded ages of 30.2 ± 1.1 ka (T3LL1D-04) and 30.3 ± 1.4 ka (T3LL1D-09). The pattern of age with depth into these cobbles, especially T3LL1D-09 that has seven slices to a depth of 5.0 mm that give consistent ages (Fig. 3c), is strong evidence that the last exposure to daylight for these samples was at that time. However, these ages pre-date the timing of advance to maximum limits at Bridgwalton, which if taken at face value suggest the deposits at Lydiate Lane relate to an earlier glacial episode, with the cluster of ages around 30 ka constraining the advance of the ice sheet during the build up towards the LGM. The setting at Lydiate Lane is intriguing, with the deposits forming a lateral bench or series of benches at 50–80 m O.D. wedged against the 200–300 m bedrock rise to the West Pennines Moors. The deposits display no evidence for over-consolidation and structural deformation that might be expected had there been later over-ride by ice, particularly so given the numerous ice-marginal ridges in the immediate vicinity (Fig. 12A). An alternative explanation is that the Lydiate Lane deposits are younger and relate to the most recent deglaciation, but the OSL signals for these samples had not been reset. The quantity of analysis undertaken here is substantial, assessing >17 000 grains and two cobbles; perhaps the transport distances from the ice margin were very short and the kame environment may have been a turbid subaqueous shallow lacustrine setting rather than outwash sandur. The chronology and context derived at Lydiate Lane remains equivocal.

The sections at Bradley’s Quarry have also evolved with time (Fig. 13A) and description of the exposures from 2009–2016 has shown >25 m vertical succession (Fig. 13B) (see Chiverrell et al., 2016), with the upper units of distal outwash deposits (Thomas et al., 1985; Sambrook Smith et al., 2005, 2006) sampled for OSL dating. Samples T3BRAD03-05 (Fig. 13C-D) were taken from adjacent rippled medium to coarse sands (Sr) from relatively low-energy flows with reasonable opportunities for bleaching in this upper and distal sandur succession (Thomas et al., 1985). SA-OSL (T3BRAD04-05) and SG-OSL (T3BRAD06) $D_\alpha$ distributions suggest that all samples were heterogeneously bleached (Fig. 2H), but they contain a minimum dose population that may reflect the burial age. The SA-OSL ages were 28.7 ± 2.5 ka (T3BRAD03) and 25.2 ± 2.3 ka (T3BRAD04), and the SG-OSL age was 45.7 ± 5.2 ka (T3BRAD06). The SA-OSL ages appear a little old and the SG-OSL age substantially too old in constraining the timing of this ice margin configuration (L9: Fig. 1). The rationale for poor resetting of the OSL signal is less clear at Bradley’s Sandpit given the low-energy sandur environment and again the quantity of analysis was substantial, assessing >20 000 grains.

Bayesian age modelling of the deglaciation

Bayesian age modelling (Brong Ramsey, 2009a) of the dating control calculates the timing for the advance and retreat northwards of the ISG across the Shropshire, Staffordshire, Cheshire and Lancashire lowlands. Ultimately, Bayesian analysis informed a conformable age model (Fig. 14; Table 3) with an overall agreement index of 109%, thus exceeding the >60% threshold advocated by Bronk Ramsey (2009a). Italicics denote the posterior density estimates or modelled ages derived from the Bayesian modelling throughout to distinguish them from the unmodelled individual ages of directly dated samples. Pre-dating ice advance to LGM limits were radiocarbon ages from Four Ashes (Morgan, 1973) and the OSL ages.
from Dowkabottom (Telfer et al., 2009), and these were supported here by the cluster of OSL ages around 29.9 ± 1.2 ka from Lydiate Lane Quarry in constraining the advance of ice into the region. In Zone 1, six OSL ages from Bridgwalton and Seisdon produced good agreement with the model, but the SG-OSL ages for T3SEIS01 and T3SEIS02 were treated as outliers. Thus, ice advance to maximum limits occurred after 28.2 ± 1.4 ka and before 25.5 ± 1.2 ka, with an ice margin.
discharging into Glacial Lake Morville $26.5 \pm 1.1$ ka. The inner shoreline of Glacial Lake Newport (Zone 2) was more challenging to date, with three ages being clear outliers (T3COND03-‐SG and –SA, T3COND05), but T2COND02 gives a better but low (42.5%) agreement with the age range refining to $25.5 \pm 1.2$ ka. However, given the potential for readvance of Welsh ice influencing the site, the original age of $19.7 \pm 2.5$ ka for T3COND02 is regarded as a better fit. Given that the Welsh ice advance was in part contemporary with the ISG ice margins at or near to the Wood Lane site, T3COND02 arguably would be better positioned in the Bayesian sequence model as part of the Phase Zone 5 (Fig. 14). The Oswestry—Whitchurch—Congleton moraine OSL ages produce a good agreement in the model, with T3WOOD05 a minor outlier (48%) and slightly too young (Fig. 14). Ice margin retreat to the L4 limits occurred after $24.3 \pm 1.1$ ka, with an ice margin discharging into sandur around Wood Lane $23.9 \pm 1.0$ ka. Two sites, Borras and Cherry Orchard Farm Quarries, constrain the dynamics of proglacial lakes and ice margins in the Dee and Weaver basins. Dating the materials at Cherry Orchard Farm was challenging, but the OSL ages overlap within uncertainties; they are conformable within the model producing an average modelled age of $23.3 \pm 1.0$ ka which is towards the young end of the measured $2\sigma$ uncertainties. The OSL ages from Borras, with the exception of the complete outlier T3BORR02-‐SA, form a tight cluster centred on a modelled age of $22.5 \pm 1.0$ ka. Together these sites constrain the retreat of ice margins into Glacial Lakes Bangor and Delamere to after $23.6 \pm 1.0$ ka (L4), with ice margin retreat northwards into Lancashire by $22.1 \pm 1.0$ ka (L5 to L6). The chronology for ice margins in Zones 8—9 was challenging, with only one out of eight ages producing a weak agreement with this phase in the age model. The Lydiate Lane chronology is a better fit and has been included earlier in the model documenting the advance stage build up to the LGM advance. The OSL measurement for T3BRAD04-‐SA is a minor outlier (52.8%) and has produced a modelled age of $21.5 \pm 1.1$ ka, but much of the constraint on this comes from well-‐dated sites further south and ages from further north documenting the retreat of ice margins into north Lancashire, the Isle of Man and SW Cumbria by $20.5 \pm 1.3$ ka (L9: Fig. 14), and then ultimately into upland Cumbria and the Pennines by $17.5 \pm 0.8$ ka (L10: Fig. 14). In summary, the Bayesian modelling has calculated modelled age probability distributions for ice limits of $28.2 \pm 1.3$ to $25.4 \pm 1.2$ ka (L1), $23.6 \pm 1.0$ ka (L4), $22.1 \pm 1.0$ ka (L5 to L6) and $20.5 \pm 1.3$ ka (L9).

**Discussion**

**Influence of lake-terminating ice margins**

The Irish Sea sector was unusual in the former BIIS in feeding ice towards two disparate termini: in the east the lake/land-terminating ISG and in the west the marine-terminating ISIS. Terrain in the lowlands of northwest England and Midlands lacks convincing evidence for higher ice velocities in the form of elongated bed forms unlike the ISIS to the west.
(Van Landeghem et al., 2009). It also terminated on land at maximum limits extending into the English Midlands (Fig. 1). Extensive proglacial lakes formed during the LGM advance and retreat stages between the ISG and the reverse bedrock slopes (Fig. 15) that generally fronted these former ice margins (Wills and Dixon, 1924; Wills, 1948; Shaw, 1972a, 1972b; Worsley, 1975, 2005; Thomas, 1989, 2005). This complex bedrock topography provided numerous niche opportunities for establishing localized ice-marginal lakes (Wills and Dixon, 1924). The lake extents presented here are not compatible with the more extensive definitions of ‘Lake Lapworth’ (see discussions in Maw, 1864; Watts, 1898; Harmer, 1907; Wills and Dixon, 1924; Poole and White, 1961; Worsley, 1975; Thomas, 1989; Murton and Murton, 2012), but show instead a time transgressive sequence of smaller lakes that developed and evolved with retreat of the ice margins (Thomas, 1989). The geochronology presented here is the first time that dating of the evolution of these lakes has been attempted. The styles of lake vary from bedrock-confined ice contact lakes (e.g. Glacial Lake Morville; Fig. 7), ubiquitous ice disintegration hollows and kettle basins concentrated in the Oswestry–Whitchurch–Congleton moraine complex (Fig. 6), to proglacial lakes confined by the fronting topography (e.g. Glacial Lake Bangor; Fig. 10).

Downstream of Ironbridge Gorge (Zone 1), slopes normal to the ice flow direction drained south and limited glacial lakes to localized ponding between the ice margins and bedrock terrain (e.g. Glacial Lake Morville and others as described by Wills and Dixon, 1924). North of the Wenlock Edge – Wrekin escarpment (Zone 1/2 boundary), slopes reverse to the ice flow direction ponded initially small and latterly more substantial compound lakes. Glacial Lakes Newport/Buildwas expanded and combined with the north-westerly retreating ice margin. The lake spillways drained east feeding the River Trent (Gnosall) and increasingly south to the River Severn with (re-)incision at Ironbridge Gorge. Continued ice margin retreat led to the development of proglacial lakes north of the moraine and bedrock highs at limits L4 (Glacial Lake Prees).
Figure 15. (A) The boundary ages (circle ± 1 sigma whisker plots) from the Bayesian model plotted against net axial retreat distance and summer insolation (pecked) for 60°N (Berger and Loutre, 1991). (B) The dolomitic carbon (DC – solid orange) and total ice-rafter debris (IRD – grey outline) flux records from the OME2X2 marine core from the Goban Spur continental slope SW of Ireland (Hapaniemi et al., 2010) and the North Atlantic core SO82-2 from the Reykjanes Ridge at 59°N (Morton et al., 2002; Rasmussen et al., 2016; Waebroeck et al., 2019). Heinrich Events H2 and H1 are highlighted (Bond et al., 1992). (C) δ18O concentrations, Greenland Stadials (GS) and Interstadials (GI) from the GS2P and GRIP Greenland ice cores (Rasmussen et al., 2014). (D) Sea surface temperature records determined for the North Atlantic using SST (°C) calculated using planktonic foraminifera for core SO82-2 at 59°N, 31°W (red line) (Van Kreveld et al., 2000; Rasmussen et al., 2016) plotted using an updated age model (Waebroeck et al., 2019) and the MD01-2461 site from the Porcupine Seabight at 51.7°N, 12.9°W (blue line) (Peck et al., 2006, 2007). (E) Ice volume equivalent sea level (Lambeck et al., 2014) and modelled relative sea level for Anglesey (blue dots) derived from a glacial isostatic adjustment (GIA) model (Bradley et al., 2011). (F) Mean and 95% trough elevations estimated from the NEXTMap elevation and EMODnet bathymetry (http://www.emodnet-hydrography.eu/) datasets plotted against the modelled boundary ages (Fig. 14). [Color figure can be viewed at wileyonlinelibrary.com]

Advances in geochronological methods

This study is the first to apply a combination of SA-OSL, SG-OSL and cobble-based OSL methods to date glacial sediments. The distributions of $D_e$ values obtained for all the sediment samples (SA-OSL and SG-OSL) are more scattered than can be explained by the measured uncertainties on individual aliquots, and this scatter is likely to result from incomplete resetting of the OSL signal in some grains at the time of deposition. Unlike some other studies of glacial sediments of similar age, no samples are seen in this study where all grains have had their OSL reset at deposition (cf. sites in the Isles of Scilly; Smedley et al., 2017b). In this study, models designed to extract the lowest dose population have been used to calculate the $D_e$ for use in age calculation in Table 2. The proportion of the grains in any given sample that were exposed to enough daylight at deposition to reset the OSL signal will vary, for example T3BRID07 and T3WOOD05 provide an interesting contrast. T3BRID07 (Fig. 2A) has a clear population of grains with a $D_e$ of ~38 Gy that is picked out by the MAM, and these are apparent in the kernel density estimate as a distinct peak. A range of grains with higher $D_e$ values are also seen, with values appearing to cluster between 150 and 250 Gy, but these are likely to be close to saturation. A clear differentiation of grains in this way was seen in similar samples from glacial sediments in eastern Ireland (Small et al., 2018, figure R3). In contrast, T3WOOD05 appears to have a very small proportion of grains that had their signal reset, whether measured on small aliquots or on single grains (Fig. 2D). This latter type of sample is common in this transect, and OSL dating of such samples, whether using single grains or small aliquots, is challenging. It is also plausible that some of the samples either have no grains
that had their OSL signal completely reset at the most recent cycle of erosion and deposition, or that the proportion of grains that were reset was so small that it has not been possible to identify them using the statistical approaches used here. A critical part of the analysis has been to collect multiple samples from sites, use multiple scales of analysis, use cobbles for dating at selected sites and use a Bayesian framework to identify outliers.

SA-OSL and SG-OSL ages are within errors for the deposits at Cherry Orchard Farm, Wood Lane and Borras Quarries. While no direct comparisons were made at Bridgwater, the SG-OSL and cobble ages are similar, as are two of four of the SA-OSL ages. The SA-OSL and SG-OSL measurements do not agree at Seisdon, but these were particularly poorly bleached sediments. On this basis we believe that our methodologies are consistently extracting the most bleached component of the $D_e$ distribution. However, given the complexity of the $D_e$ distributions and the very small component of many of the samples that appear to have been bleached, the extracted final $D_e$ used for age calculation purposes may still include some poorly bleached grains leading to age over-estimation. Given the ice-marginal contexts sampled, the sedimentation rates may have been very high, the sediment transport pathways short and water columns opaque with heavy sediment loads. As a result, it is not a surprise that sunlight exposure was often insufficient to bleach samples before burial. Bayesian analysis of sites and samples within zones supports this by identifying outlier ages which were affected particularly by poor bleaching. Age over-estimation is less of an issue for the cobble approach because poorly bleached cobbles were identified and not analysed. This, along with a lower reliance on understanding palaeo-moisture fluctuations, has led to a lower uncertainty for the cobble ages.

Forcing of the glacial dynamics

The build-up and advance of the BIIS into NW England had not been dated previously (Chiverrell and Thomas, 2010; Hughes et al., 2016a). Ice-free conditions are evidenced in the central Scottish sector of the ice-sheet and radiocarbon dated to 33.1–34.4 ka BP at Sourlie (40 km SW from Glasgow, 55°38′ 18.80′′N, 4°38′31.46′′W) (Jardine et al., 1988; Bos et al., 2004) and 31.5–34.8 ka BP at Balglass Burn (20 km north of Glasgow, 56°02′N, 04°17′W) (Brown et al., 2007). Those radiocarbon ages are similar to the 35–34 cal ka BP range obtained for equivalent ice-free conditions at Four Ashes in the English Midlands (Morgan, 1973). Here, we use the convergence of quartz grain- and granite cobble-based OSL ages from Lydiate Lane to show that by 30.2 ± 1.1 and 30.3 ± 1.4 ka eastern Lake District and Irish Sea (Galloway Hills) ice masses had advanced 65 and ~150 km, respectively, into south Lancashire. The Lydiate Lane kame complex requires an ice margin abutting against the lower slopes (30–90 m O.D.) of the West Pennines (maximum elevation 450 m). The Lydiate Lane deposits contain ISB and Lake District erratic cobbles including Shap granite and volcanioclastic sediments (Borrowdale Volcanic Group). The lack of deformation in the sequence (e.g. consolidation, faulting or folding) point to limited subsequent override by ice on this eastern margin. Later expansions of Irish Sea ice to LGM limits 28.3 ± 1.9–26.5 ± 1.8 ka in the English Midlands and ~26 ± 1.5 ka in the south Celtic Sea (Praeg et al., 2015; Smedley et al., 2017b; Scourse et al., 2019), coupled with faster ice flows down the axis of the Irish Sea (Van Langenhem et al., 2009), may have reduced ice flows towards this less dynamic eastern ice margin. Our discussion of ice margin advance and retreat rates reflects the net movement of ice margins and by necessity integrates smaller episodes of margin advance, slowdown, stillstand, retreat and acceleration. Build-up of ice, from what were probably reduced-ice or ice-free conditions in Britain 34.0–31.5 ka (Fig. 15A), comprised net marginal advance rates of 100–46 m a⁻¹ from the uplands of SW Scotland and NW England into south Lancashire by 30.3 ± 1.4 ka.

The subsequent LGM advance of the ISG occurred from 28.3 ± 1.9 ka (Fig. 15A), reaching and ponding Glacial Lake Morville at the maximum limit at 26.5 ± 1.8 ka. This ice margin advance of 130 km occurred at ~50 m a⁻¹, and was coeval with the shift to colder conditions of Greenland Stadial 3 (GS-3) (Fig. 15C), a colder North Atlantic (Fig. 15D) and the June insolation minimum (Fig. 15A) (Berger and Loutre, 1991; Van Kreveld et al., 2000; Peck et al., 2007; Rasmussen et al., 2014, 2016). The chronology for the ISG maximum is equivalent to dating of maximum limits to the west for the ISG on the Isles of Scilly at 26 ± 1.5 ka (Smedley et al., 2017b) and in the south Celtic Sea at 27–24 ka (Praeg et al., 2015; Scourse et al., 2019). The advance of the ISG was larger, extending 550 km to the south Celtic Sea (Praeg et al., 2015; Lockhart et al., 2018; Scourse et al., 2019) and it was also faster, at ~180 m a⁻¹. Comparison of the timing for the LGM in the English Midlands with other global ice-sheets is better framed in the context of the response of the entire BIIS, although the Irish Sea sector was a large ice mass discharging >17% of the total BIIS. The LGM maxima for the Irish Sea sector displays a strong similarity in timing to the other west-draining and marine-terminating ice streams of the former BIIS (Callard et al., 2018; Ó Cofaigh et al., 2019; Scourse et al., 2019; Callard et al., 2020). The LGM in the English Midlands, expanding after 28.3 ± 1.9 ka and reaching the maximum limit at 26.5 ± 1.8 ka, is early in the context of an LGM during GS-3 at 27.5–23.3 ka (Hughes and Gibbard, 2015) and it is early also in the time window for the eustatic sea-level minimum and maximum global ice volume (Fig. 15) (Lambeck et al., 2014). This relatively early build-up and expansion of ice reflects potentially the wetter oceanic climate on the Atlantic eastern seaboard during the prolonged colder conditions of late MIS 3 and GS-5, and was followed by substantial glacier growth during the cold conditions of GS-3 (Rasmussen et al., 2014, 2016).

Retreat of the ISG was initially slow and relatively even in pace, with ice margins passing through the ~80 km of Shropshire and Cheshire between 25.3 ± 1.6 and 22.5 ± 1.0 ka at net rates of ~30 m a⁻¹ (Fig. 15A). This retreat begins under cooler conditions of GS-3 and with cold surface waters in the North Atlantic (Fig. 15C,D), posing the question: why is the land-terminating ISG retreatting? A possible explanation lies in the ISG sharing the same source regions as the larger ISIS, an ice mass that underwent a ~550-km advance at net advance rates of 180 m a⁻¹. The extension of the ISIS into the Celtic Sea was short-lived (Scourse et al., 1990, 2019; Ó Cofaigh and Evans, 2007; Chiverrell et al., 2013), and was followed by a 500- km retreat coeval with Heinrich Event H2 at net retreat rates of 150 m a⁻¹ (Smedley et al., 2017a). These ice margin retreat rates are five times faster than the coeval slower retreat of the ISG from the English Midlands. The discharge of ice to the Celtic Sea effectively beheaded the ISG in the English Midlands, which would account for the ubiquitous evidence for ice stagnation and disintegration topography in the form of kettle-holes throughout the retreat sequence in the English Midlands (Thomas, 1989, 2005; Chiverrell et al., 2016). Net ice-marginal retreat rates then accelerated with the step back of ice margins into the Irish Sea and Lancashire covering ~130 km at ~70 m a⁻¹, before slowing as the ice margins stepped back into the Lake District (Fig. 15A). The increase in pace of retreat around 22.5 ± 1.0 ka corresponds with a number of factors including...
increasing June insolation (Fig. 15A) and higher temperatures during Greenland Interstadial 2 (GI-2: Fig. 15C,D), wider ice margins and increasing influence of calving margin initially in the broader deeper glacial lakes of the lower Dee and Weaver basins and latterly in the glacimarine eastern Irish Sea (Smedley et al., 2017a; Chiverrell et al., 2018). Our interpretation of this retreat sequence highlights the importance of considering the competition between glaciers and internal dynamics in controlling the behaviour of ice-masses during retreat episodes.

The ISG was the dominant ice mass in the English Midlands, but moraines reflecting ice issuing from the Severn and other valley systems evidence a lobate readvance of Welsh ice taking advantage probably of the accommodation space vacated by Irish Sea ice. Welsh ice moraines appear a corollary of equivalent ISG moraines near Wood Lane dated to 23.9 ± 1.0 ka (Fig. 9) and the outwash deposits of Welsh ice have been dated here for first time at Condover to 19.7 ± 2.5 ka. There are lobate moraines on the sea floor west of Wales, and ice-sheet modelling experiments (Patton et al., 2013) have pointed to a westwards readvance by Welsh Ice after the retreat of the ISIS that is undated but postulated at around 21 ka. There was a readvance by Welsh ice both to the east and to the west after the retreat of Irish Sea ice, and this is constrained here to the period 23.9 ± 1.0 to 19.7 ± 2.5 ka. Hughes et al. (2016b) dated the thinning of the Welsh Ice Cap using 15 Be ages. Those measurements have been recalculated here using Loch Lomond Production Rates (Fabel et al., 2012) and combined with a reduced Chi-square test (χ² = 21; 5% = 23.7) to give an age of 20.7 ± 0.23 ka. A Welsh readvance is more difficult to sustain later during the period 23.9 ± 1.0 to 19.7 ± 2.5 ka with the Welsh Ice Cap thinning by 20.7 ± 0.23 ka. These constraints place the Welsh readvance during the transition to colder conditions from GI-3 to GI-2.

Rasmussen et al., 2014, 2016). The forcing of the readvance could also have included an opportunistic expansion to occupy the accommodation space vacated by the ISG. The lobate Welsh moraines are located on the western shoreline on an over-deepened basin, Glacial Lake Severn (Fig. 7), which raises glacier instability at the transition from land-terminating to a calving glacialaustrine ice margin as a further possible forcing mechanism. This convergence of a cooler climate, accommodation space and dynamic instability owing to ice-bed topography and environments provided conditions conducive to and explaining the readvance and oscillation of the Welsh ice margin in the English Midlands.

Conclusions
We present a novel combination of cobble-; SA- and SG-OSL to constrain the timing of BIIS advance to LGM limits in the English Midlands and the pace of ice margin retreat. New geomorphological mapping and assessment of the glacial stratigraphy at key sites have revealed the evolution of environments fronting the ice margins during deglaciation of the region. Patterns in this geochronology provide the first constraint on what, from a global perspective, is an early build-up and advance of ice into lowland Lancashire during GS-5 at 30 ± 2.0 ka, with the advance to terrestrial LGM limits during GS-3 between 26.3 ± 1.9 ka and a maximum extent at 26.5 ± 1.8 ka. This land-terminating advance slightly pre-dates or is coeval with the large expansion of the ISIS into the south Celtic Sea. Evidence for a reduced ice presence on the ISG eastern ice margins fronting eastern Lake District ice and endemic ice stagnation and disintegration topography accompanied the relatively early retreat of ice from the English Midlands during GS-3. The correspondence of these events point to the ISIS Celtic Sea advance and onset of ice streaming in the central Irish Sea as key internal drivers of the glacial dynamics. The advance of ice to maximum limits occurs during the climate cooling and summer insolation minima of GS-3. Thereafter, still under the cooler conditions of GS-3, piracy of shared ice sources by the ISIS led to a decline in impetus behind the ISG feeding towards the English Midlands. In the English Midlands the ISG retreat was slow and relatively even in pace between 25.3 ± 1.6 and 22.5 ± 1.0 ka. The establishment and abandonment of a series of time transgressive lakes accompanied the retreat of ice margins from the English Midlands. Retreat of ice margins from the larger and deeper of these lakes in the lower River Dee and Weaver basins provides a possible explanation for the acceleration in retreat around 22.5 ± 1.0 ka with the ice front eventually vacating the region by 20.6 ± 2.2 ka, but this also corresponds with increasing June insolation and higher temperatures of GI-2. The geomorphology west of Shrewsbury shows unambiguous evidence for what might be an opportunistic readvance of Welsh ice as the Severn Valley Glacier expanded into space vacated by the ISG between 23.9 ± 1 and 19.7 ± 2.5 ka, but this also corresponds with the GI-3 to GS-2 transition to colder conditions and occurred in a location conducive to ice margin instability on the down-ice shoreline of a substantial glacial over-deepening.

In summary, we have shown a GS-5 build-up and then GS-3 expansion of Irish Sea ice into the English Midlands, which reflects potentially wetter oceanic conditions affecting the BIIS on the Atlantic eastern seaboard during these cold episodes. The timescale and geomorphology of the English Midlands retreat sequence point to a glacial system driven by climate, but heavily mediated by competition between ice masses, internal adjustments in the flow regime and the nature of the environment fronting the ice margins during the advance and retreat from maximum limits.

Supporting information
Additional supporting information may be found in the online version of this article at the publisher’s web-site.

Acknowledgements. This work was supported by a Natural Environment Research Council consortium grant: BRITICE-CHRONO NE/J009768/1. Thanks are due to the technical support staff at the Aberystwyth Luminescence Research Laboratory and Sheffield Luminescence Laboratory. Phil Hughes, an anonymous reviewer and the editorial input of Arjen Stroeven are acknowledged for their detailed constructive comments, which helped to improve the paper.

Data availability statement
The data that support the findings of this study are available from the corresponding author upon reasonable request.

Abbreviations. BGS, British Geological Survey; BIIS, British-Irish Ice Sheet; GS, Greenland Stadial; ICP-AES, inductively coupled plasma atomic emission spectroscopy; ICP-MS, inductively coupled plasma mass spectrometry; IEU, internal–external uncertainty; IRSL, infra-red stimulated luminescence; ISB, Irish Sea Basin; ISG, Irish Sea Glacier; ISIS, Irish-Sea Ice Stream; LGM, Last Glacial Maximum; MAM, Minimum Age Model; MCMC, Markov chain Monte Carlo; MIS, Marine Isotope Stage; OSL, optically stimulated luminescence; RVG, Ribble Valley Glacier; SA, small aliquots; SG, single grains; SVG, Severn Valley Glacier.

References
Aitkenhead N, Bridge D, Riley NI et al. 1992. Geology of the Country Around Garstang, Memoir of the British Geological Survey, Sheet 67. HM Stationery Office: London.
Aucilair M, Lamothé M, Huot S. 2003. Measurement of anomalous fading for feldspar IRSL using SAR. Radiation Measurements 37: 487–492.

Bateman MD et al. 2017. The timing and consequences of the blockage of the Humber Gap by the last British–Irish Ice Sheet. Boreas 47: 41–61.

Berger A, Loutre MF. 1991. Insolation values for the climate of the last 10 million years. Quaternary Science Reviews 10: 297–317.

Bond G, Heinrich H, Broecker WS et al. 1981. Onset and timing of the last glacial maximum. Science 212: 1071–1080.

Bradley SL, Milne GA, Shennan I et al. 2011. An improved glacial isostatic adjustment model for the British Isles. Journal of Quaternary Science 26: 541–552.

Brazier V, Kirkbride MP, Gordon JE. 1998. Active ice-sheet deglaciation and ice-dammed lakes in the northern Cairngorm Mountains, Scotland. Boreas 27: 297–310.

Brunk Ramsey C. 2009a. Bayesian analysis of radiocarbon dates. Quaternary Science Reviews 28 (2020): 289–311.

Buck CE, Cavanagh WG, Litton CD. 1996. Bayesian Approach to Bayesian Approach to single and multiple grains of quartz from Jinmium rock shelter, northern Australia: Part I. Experimental design and statistical models. Archaeometry 41: 319–364.

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 35: 91–102.

Frechen R, Schibati R, Murray A et al. 2015. Mathematical model quantifies multiple daylight exposure and burial events for rock surfaces using luminescence dating. Radiation Measurements 81: 16–22.

Galbraith RF, Laslett GM. 1991. Statistical models for mixed fission track ages. Nuclear Tracks and Radiation Measurements 21: 459–470.

Galbraith RF, Roberts RG, Laslett GM et al. 1999. Optical dating of single and multiple grains of quartz from Jinmium rock shelter, northern Australia: Part II. Experimental design and statistical models. Archaeometry 41: 319–364.

Guérin G, Mercier N, Adamiec G. 2011. Dose-rate conversion factors: update. Ancient TL 29: 5–8.

Hamblin RJ. 1986. The Pleistocene sequence of the Telford district. Proceedings of the Geologists’ Association 97: 363–377.

Hughes PD, Gibbard PL, Plint AC. 2015. A stratigraphical basis for the Last Glacial Maximum (LGM). Quaternary International 383: 174–185.

Hughes PD, Glasser NF, Plint AC. 2016. The abanico plot: visualising chromatic data with individual standard errors. Quaternary Geochronology 31: 12–18.

Dietze M, Kreutzer S, Burow C et al. 2016. The abanico plot: visualising chromatic data with individual standard errors. Quaternary Geochronology 31: 12–18.

Earp JR, Poole EG, Whiteman AJ. 1961. Geology of the Country Around Clitheroe and Nelson. Memoir of the Geological Survey of Great Britain, 68. HMS Stationery Office: London.

Evans DJA. 2003. Glacial Landscapes. Arnold: London.

Evans DJA, Bateman MD, Roberts DH et al. 2017. Glacial Lake Pickering: stratigraphy and chronology of a proglacial lake drowned by the North Sea Lobe of the British–Irish Ice Sheet. Journal of Quaternary Science 32: 295–310.

Evans DJA, Benn DI 2004. A Practical Guide to the Study of Glacial Sediments. Arnold, London. 266pp. ISBN:10: 0340759593.

Galbraith RF, Laslett GM. 1991. Statistical models for mixed fission track ages. Nuclear Tracks and Radiation Measurements 21: 459–470.

Galbraith RF, Roberts RG, Laslett GM et al. 1999. Optical dating of single and multiple grains of quartz from Jinmium rock shelter, northern Australia: Part I. Experimental design and statistical models. Archaeometry 41: 319–364.

Guérin G, Mercier N, Adamiec G. 2011. Dose-rate conversion factors: update. Ancient TL 29: 5–8.

Hamblin RJ. 1986. The Pleistocene sequence of the Telford district. Proceedings of the Geologists’ Association 97: 363–377.

Hammer FW. 1907. On the origin of certain canyons and their associations. Scientific Monthly 24: 321–329.

Hughes PD, Glasser NF, Fink D. 2016b. Rapid thinning of the Welsh Ice Cap at 20°N. Nature 541: 619–622.

Hughes PD, Gibbard PL. 2015. A stratigraphical basis for the Last Glacial Maximum (LGM). Quaternary International 383: 174–185.

Hughes PD, Gibbard PL, Plint AC. 2016. The abanico plot: visualising chromatic data with individual standard errors. Quaternary Geochronology 31: 12–18.

Jardine WG, Dickson JH, Haughton P DW et al. 1988. A late Middle Devensian interstadial site at Sourlie, near Irvine, Strathclyde. Scottish Journal of Geology 24: 288–295.
driven by glacial-isostatic depression and high relative sea level. Quaternary Science Reviews 208: 76–96.

Ou XL, Roberts HM, Duller GAT et al. 2018. Attenuation of light in different rock types and implications for rock surface luminescence dating. Radiation Measurements 120: 305–311.

Palacios D, Stokes CR, Phillips FM et al. 2020. The deglaciation of the Americas during the Last Glacial Termination. Earth-Science Reviews 203.

Parkes AA, Waller RI, Knight PG et al. 2009. A morphological, sedimentological and geophysical investigation of the Moine Moraine, Shropshire, England. Proceedings of the Geologists’ Association 120: 231–244.

Patton H, Hubbard A, Bradwell T et al. 2013. Rapid marine deglaciation: asynchronous retreat dynamics between the Irish Sea Ice Stream and terrestrial outlet glaciers. Earth Surface Dynamics 1: 53–65.

Peake DS. 1961. Glacial changes in the Alyn river system and their significance in the glaciology of the North Welsh border. Quarterly Journal of the Geological Society 117: 335–363.

Peake DS. 1979. The limit of the Devensian Irish Sea ice sheet on the north Welsh border. Quarterly Newsletter 27: 1–4.

Peake DS. 1981. The Devensian glaciation on the North Welsh border. In The Quaternary in Britain 49–59.

Peck VL, Hall IR, Zahn R et al. 2006. High resolution evidence for linkages between NW European ice sheet instability and Atlantic Meridional Overturning Circulation. Earth and Planetary Science Letters 243: 476–488.

Peck VL, Hall IR, Zahn R et al. 2007. The relationship of Heinrich events and their European precursors over the past 60 ka a r: a multi-proxy ice-rafted debris provenance study in the North East Atlantic. Quaternary Science Reviews 26: 862–875.

Pocock RW, Wray DA 1925 The Geology of the Country Around Wem. Geological Survey, Memoirs.

Poole EG. 1966. Geology of the Country Around Nantwich and Whitchurch: (Explanation of One-Inch Geological sheet 122). Her Majesty’s Stationary Office: London.

Poole EG, Whitman AL. 1961. The glacial drifts of the southern part of the Shropshire-Cheshire basin. Quarterly Journal of the Geological Society 117: 91–130.

Praeg D, McCarron S, Dove D et al. 2015. Ice sheet extension to the Celtic Sea shelf edge at the Last Glacial Maximum. Quaternary Science Reviews 111: 107–112.

Price D 1963. Geology of the Country Around Preston. 75. H M Stationary Office: London.

Rasmussen SO, Bigler M, Blockley SP et al. 2014. A stratigraphic framework for abrupt climatic changes during the Last Glacial period based on three synchronized Greenland ice-core records: refining and extending the INTIMATE event stratigraphy. Quaternary Science Reviews 106: 14–28.

Rasmussen TL, Thomsen E, Moros M. 2016. North Atlantic warming during Dansgaard-Oeschger events synchronous with Antarctic warming and out-of-phase with Greenland climate. Scientific Reports 6: 20535.

Sambrook Smith GH, Ashworth PJ, Best JL et al. 2005. The morphology and facies of sandy braided rivers: some considerations of scale invariance. Special Publications of the International Association of Sedimentologists 35: 145–158.

Sambrook Smith GH, Ashworth PJ, Best JL et al. 2006. The sedimentology and alluvial architecture of the sandy braided South Saskatchewan River, Canada. Sedimentology 53: 413–434.

Scourse JD, Austin WEN, Bateman RM et al. 1990. Sedimentology and micropaleoecology of glacimarine sediments from the Central and southwestern Celtic Sea. In Glacimarine Environments: Processes and Sediments, Dowdeswell JA, Scourse JD (eds). Special Publication of the Geological Society. The Geological Society of London: London.

Scourse JD, Coope GR, Allen JRM et al. 2009. Late-glacial remains of woolly mammoth (Mammuthus primigenius) from Shropshire, UK: stratigraphy, sedimentology and geomorphology of the Condover site. Geological Journal 44: 392–413.

Scourse J, Saher M, Van Landeghem KJJ et al. 2019.Advance and retreat of the marine-terminating Irish Sea Ice Stream into the Celtic Sea during the last glacial: timing and maximum extent. Marine Geology 412: 53–68.
Shaw J. 1972a. The Irish Sea glaciation of north Shropshire – some environmental reconstructions. Field Studies 4: 603–631.
Shaw J. 1972b. Sedimentation in the ice-contact environment with examples from Shropshire (England). Sedimentology 18: 23–62.
Small D, Smedley RK, Chiverrell RC et al. 2018. Trough geometry was a greater influence than climate-ocean forcing in regulating retreat of the marine-based Irish-Sea Ice Stream. GSA Bulletin 130: 1981–1999.
Smedley RK, Chiverrell RC, Ballantine CK et al. 2017a. Internal dynamics condition centennial-scale oscillations in marine-based ice-stream retreat. Geology 45: 787–790.
Smedley RK, Duller GAT, Ruter D et al. 2020. Empirical assessment of beta dose heterogeneity in sediments: implications for luminescence dating. Quaternary Geochronology 56.
Smedley RK, Scourse JD, Small D et al. 2017b. New age constraints for the limit of the British–Irish Ice Sheet on the Isles of Scilly. Journal of Quaternary Science 32: 48–62.
Smith ND, Ashley GM. 1985. Proglacial lacustrine environment. In Glacial Sedimentary Environments 135–216.
Soibati R, Murray AS, Porat N et al. 2015. Age of a prehistoric ‘Rodedian’ cult site constrained by sediment and rock surface luminescence dating techniques. Quaternary Geochronology 30: 90–99.
Stokes CR. 2018. Geomorphology under ice streams: moving from form to process. Earth Surface Processes and Landforms 43: 85–123.
Stokes CR, Clark CD. 1999. Geomorphological criteria for identifying Quaternary Geochronology 30: 67–74.
Telfer MW, Wilson P, Lord TC et al. 2009. New constraints on the age of the last ice sheet glaciation in NW England using optically stimulated luminescence dating. Journal of Quaternary Science 24: 906–915.
Thiel C, Buylaert JP, Murray A et al. 2011. Luminescence dating of the Stratzing loess profile (Austria) – Testing the potential of an elevated temperature post-IR IRSL protocol. Quaternary International 234: 23–31.
Thomas GSP. 1985a. The Late Devensian glaciation along the border of northeast Wales. Geological Journal 20: 319–340.
Thomas GSP. 1985b. The Quaternary of the Northern Irish Sea Basin. In: The Geomorphology of North-West England, Johnson RH (ed). Manchester University Press: Manchester; 143–158.
Thomas GSP. 1989. The Late Devensian glaciation along the western margin of the Cheshire–Shropshire lowland. Journal of Quaternary Science 4: 167–181.
Thomas GSP. 2005. North-East Wales. In: The Glaciations of Wales and Adjacent Areas, Lewis CA, Richards AE (eds). Logaston Press: Bristol; 41–58.
Thomas GSP, Chiverrell R, Huddart D. 2004. Ice-marginal depositional responses to readvance episodes in the Late Devensian deglaciation of the Isle of Man. Quaternary Science Reviews 23: 85–106.
Thomas GSP, Chiverrell RC. 2007. Structural and depositional evidence for repeated ice-marginal oscillation along the eastern margin of the Late Devensian Irish Sea Ice Stream. Quaternary Science Reviews 26: 2375–2405.
Thomas GSP, Connaughton M, Dackombe RV. 1985. Facies variation in a Late Pleistocene supraglacial outwash sandur from the Isle of Man. Geological Journal 20: 193–213.
Thomsen KJ, Murray AS, Better-Jensen L et al. 2007. Determination of burial dose in incompletely bleached fluvial samples using single grains of quartz. Radiation Measurements 42: 370–379.
Van Kvevel D, Samthin M, Erlenkeuser H et al. 2000. Potential links between surging ice sheets, circulation changes, and the Dansgaard–Oeschger Cycles in the Irminger Sea, 60–18 Kyr. Paleoceanography 15: 425–442.
Van Landeghem KJ, Wheeler AJ, Mitchell NC. 2009. Seafloor evidence for palaeo-ice-streaming and calving of the grounded Irish Sea Ice Stream: implications for the interpretation of its final deglaciation phase. Boreas 38: 111–118.
Waelbroeck C, Lougheed BC, Vazquez Riveiros N et al. 2019. Consistently dated Atlantic sediment cores over the last 40 thousand years. Scientific Data 6: 165.
Watts WW. 1898. Long Excursion to the Birmingham District. Proceedings of the Geologists’ Association 15: 417–419.
Wedd CB, Smith B, Wills LJ et al. 1927. The Geology of the Country Around Wrexham: Explanation of One-Inch Geological sheet 121, New Series. Memoirs of the Geological Survey of Great Britain, England and Wales (Sheet – New Series). HMSO: London.
Whitehead TH, Robertson T, Pocock RW et al. 1928. The Country Between Wolverhampton and Shrewsbury. HMSO: London.
Wills LJ. 1948. The Palaeogeography of the Midlands, University Press of Liverpool: Liverpool.
Wills LJ, Dixon EEL. 1924. The development of the Severn valley in the neighbourhood of Iron-Bridge and Bridgnorth. Quarterly Journal of the Geological Society 80: 274–308.
Wilson AC, Mathers SJ, Cannell B 1982. The Middle Sands, a prograding sandur succession; its significance in the glacial evolution of the Wrexham–Shrewsbury region. United Kingdom, Institute of Geological Sciences Report 82: 30–35.
Wilson P, Barrows TT, Lord TC 2012a. Cosmogenic isotope analysis and surface exposure dating in the Yorkshire Dales. In Cave Archaeology and Karst Geomorphology of North West England, Regin HJO, Faulkner T, Smith IR (eds). Quaternary Research Association and British Cave Research Association: London; 117–135, ISBN: 090778084925323.
Wilson P, Barrows TT, Lord TC et al. 2012b. Surface lowering of limestone pavement as determined by cosmogenic (10Be) analysis. Earth Surface Processes and Landforms 37: 1518–1526.
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 Á. 2013. Deglaciation of the eastern Cumbria glaciokarst, 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, Rodés Á, Smith A. 2018. Valley glaciers persisted in the Lake District, north-west England, until ~16–15 ka as revealed by terrestrial cosmogenic nuclide (10Be) dating: a response to Heinrich event 11. Journal of Quaternary Science 33: 518–526.
Worsley P 1975. An appraisal of the glacial Lake Lapworth concept. In Environment, Man and Economic Change, Phillips ADM, Turton B (eds). Longman: London; 98–118.
Worsley P 1985. Pleistocene history of the Cheshire–Shropshire Plain. In The Geomorphology of North-West England, Johnson RH (ed). Manchester University Press: Manchester; 201–221.
Worsley P. 2005. The Cheshire–Shropshire plain. In: The Glaciations of Wales and Adjacent Areas, Lewis CA, Richards AE (eds). Logaston Press: Bristol; 59–72.