Construction of an evolutionary deglaciation model for the Irish midlands based on the integration of morphostratigraphic and geophysical data analyses

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ABSTRACT: Alternative, established models for the deglaciation of the midlands of Ireland are tested against an interpretation of a suite of deglacial sediments covering an area of 600 km^2 in the east central midland area. Interpretation of the sediments is based on geomorphological mapping, lithostratigraphic characterization of exposures and geotechnical data supported by electrical resistivity tomography (ERT) and ground penetrating radar (GPR). GPR depicted small-scale sedimentological and deformational structures within low-conductivity soft sediments, such as cross-bedding, planar bedding, channel-like features and faulting planes, and revealed the internal architecture of eskers, glaciodeltas, subaqueous fans and raised bogs. ERT data permitted the detection of depth to bedrock and the lithological characterization of unconsolidated sediments. The ten sites presented were surveyed by traditional mapping methods and/or geophysical techniques. This allowed the construction of a local model of the deglaciation of the area which recognized five main stages. An ice sheet covering most of Ireland withdrew as a single body as far as the midlands. At this stage, two main directions of ice retreat are identified from the spatial distribution of meltwater/overflow channels, esker and morainic ridges, and ice-marginal glaciolacustrine deposits. A pattern of deglacial sedimentation into an expanding ice-marginal glacial lake is depicted. The glacial lake was dammed to the west by two ice dome fronts, one decaying to the north-west and another to the south-west, and by the Shannon Basin watershed to the east. Glacial lake outlets identified along the watershed and the altitude of the topset/foreset interface zone depicted in glaciodeltaic deposits allowed the identification of three lake water levels. The highest level is at 87–89 m Ordnance Datum (OD), the second lake level at 84 m OD and the third at 78 m OD. Copyright © 2012 John Wiley & Sons, Ltd.

KEYWORDS: esker; geophysics; glacial; glaciolacustrine; Ireland.

Introduction

Ireland was covered by ice for long periods during the Quaternary and the last glaciation effectively shaped the current landscape and defined the underlying geology of the country (Edwards and Warren, 1985). Research on the distribution of glacial landforms in Ireland at a regional scale (e.g. Close, 1867; Charlesworth, 1928; Synge, 1979; Hoare, 1991; Warren, 1991a; Knight et al., 2004) and on the description of glacigenic deposits at a local scale (e.g. Farrington and Synge, 1970; Warren, 1979; Glanville, 1997; Ó Cofaigh and Evans, 2007) over a period of almost 150 years did not produce consensus on either the position of the last glacial maximum (LGM) or the spatial distribution, shape and dynamics of the ice sheets covering Ireland (contrast Warren, 1992, Figs. 3, 7; Eyles and McCabe, 1989, Figs. 3, 30). More recent research using remote sensing, regional glacial landform mapping (Greenwood and Clark, 2009a,b) and dating methods (e.g. Ballantyne et al., 2006, 2007, 2008; Ó Cofaigh et al., 2010a,b; McCarroll et al., 2010) has moved a consensus towards the sort of model outlined by Warren, (1991a,b, 1992): a dynamic model of glaciation and deglaciation consisting of an ice sheet characterized by convergent ice domes and with significant fluctuations at its margins and centres of mass, and extending offshore during the LGM (Ballantyne, 2010; Clark et al., 2010; Chiverrell and Thomas, 2010), and tends to confirm subsequent withdrawal towards discrete ice domes (Warren, 1991b, 1992). The Irish midland region is a key area to

understanding the deglaciation processes within the Irish context. Warren (1991a) proposed a model of deglaciation in the Irish midlands consisting of an expanding glacial lake delimited by the eastern Shannon Basin watershed and by two ice masses gradually retreating north-westwards and south-westwards. Further evidence for the existence of such a lake was compiled from the analysis of glaciolacustrine clays (van der Meer and Warren, 1997; Delaney, 2008) and the identification of subaqueous fans and glaciodeltaic deposits (Warren and Ashley, 1994; Delaney, 2002). A 600-km^2 area in the vicinity of Tullamore, where extensive glaciofluvial and glaciolacustrine deposits are found, was investigated to test the glacial lake model and, by extension, the alternative contradictory models.

With the objective of examining the salient glacigenic depositional features of the area and with the view to testing the results against existing models, a multidisciplinary approach, combining morphostratigraphic analyses supported by geophysical methods, was taken to construct an evolutionary model of deglaciation for the area. The widespread occurrence of glaciolacustrine sediments and landforms and the proposed ice-marginal lake of Warren (1991b) prompted an examination of the watershed areas to assess potential lake overflow channels. Geophysical techniques are fast and inexpensive data collection methods which complement traditional field methods and are increasingly used for the study of the subsurface sedimentology of glacial and postglacial sediments (e.g. Bakker and van der Meer, 2003; Gutsell et al., 2004; Burke et al., 2008; Trafford, 2009; McClymont et al., 2011) particularly where exposed sediments for lithostratigraphic analysis might be limited.

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Geographical and geological settings

The study area covering 600 km² is located in the east-central midlands of Ireland. The distribution of main geographical features is presented in Fig. 1.

Field mapping by one of the authors (X.M.P.) shows that bedrock close to the surface represents less than 2% of the study area. Carboniferous limestone is the dominant petrology underlying about 96% of the area, Carboniferous volcanic rocks, around Croghan Hill, represent about 3% of the bedrock and Devonian sandstone is present south-west of Ballycumber (Hitzman, 1992). Bedrock is generally overlain by glacial and postglacial sediments. These range from 0 to 49 m in thickness with a mean minimum sediment thickness of 7.38 m (Pellicer, 2010). Subglacial sediments are mostly composed of diamicton derived from limestone and form a gently undulating landscape with some poorly developed drumlinoids. Esker ridges and associated glaciofluvial/glaciolacustrine landforms, generally composed of sand and gravel, are the dominant geomorphological features. Seven main eskers, named from north to south Moate, Horseleap, Split Hill, Clara, Ballyduff, Geashill and Kilcormac eskers, form an esker system converging eastwards in a large fan shape (Fig. 1). Other significant glacigenic features are meltwater channels draining from the Shannon Basin to the west into the Boyne/Barrow Basins to the east, and ice-marginal ridges (Fig. 2).

The esker system cuts across the midlands in a west to east direction. The eskers have, over time, been subjected to a variety of interpretations: marine banks formed in a postglacial 'Esker Sea' (Kinahan, 1878); subglacial tunnel-fill deposits (Sollas, 1896); Salpausselkä-type moraines deposited in a shallow glacial sea (Gregory, 1920); recessional moraine ridges associated with terrestrial ice retreating northwards (Charlesworth, 1928); flooded, open-channel and crevasse fillings in a huge expanse of stagnant ice (Flint, 1930); and a combination of tunnel-fill and glaciolacustrine fans deposited in association with a westerly and south-westerly retreating ice margin in the southern part of our area of study (Farrington and Synge, 1970). Warren (1991a, 1992) envisaged two main ice centres acting in the midlands, later named the Northern Dome and the Central Dome (Warren, 1993). These coalesced close to Tullamore Town and advanced as a single ice sheet in an east-southeast direction. During deglaciation the ice of the two domes decoupled in this area opening a glacial lake confined by two separate ice sheets retreating towards the north-west and south-west, respectively (Warren, 1991b; Warren and Ashley, 1994). Greenwood and Clark (2009b) inferred, mainly from remote sensed data, that ice advanced in this region south-eastwards as a single ice sheet and postulate two possible scenarios for ice sheet withdrawal as: (i) two retreating and separating ice domes and (ii) a single ice sheet retreat with a subsequent readvance.

Morphostratigraphic analysis of the main glaciofluvial features in the Irish midlands (Warren and Ashley, 1994) showed that some of the larger eskers were polygenetic, being composed of tunnel-fill sediments partially buried by ice-marginal fans and fan moraines. Delaney (2001, 2002)
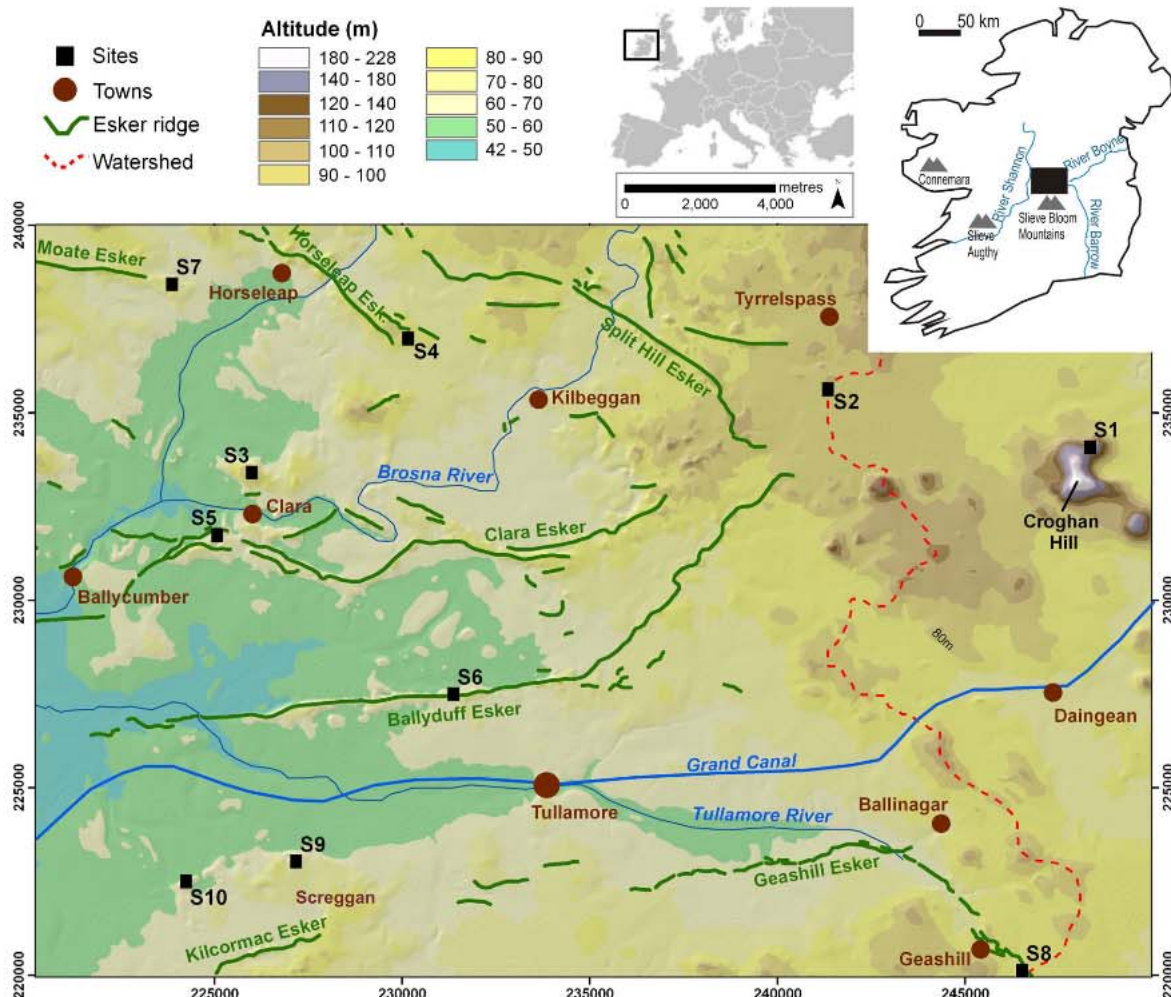

Figure 1. Topographic map of the study area showing the location of the main geographical features, esker ridges and sites S1–S10 discussed. Coordinates shown refer to the Irish National Grid. This figure is available in colour online at wileyonlinelibrary.com/journal/jqs.
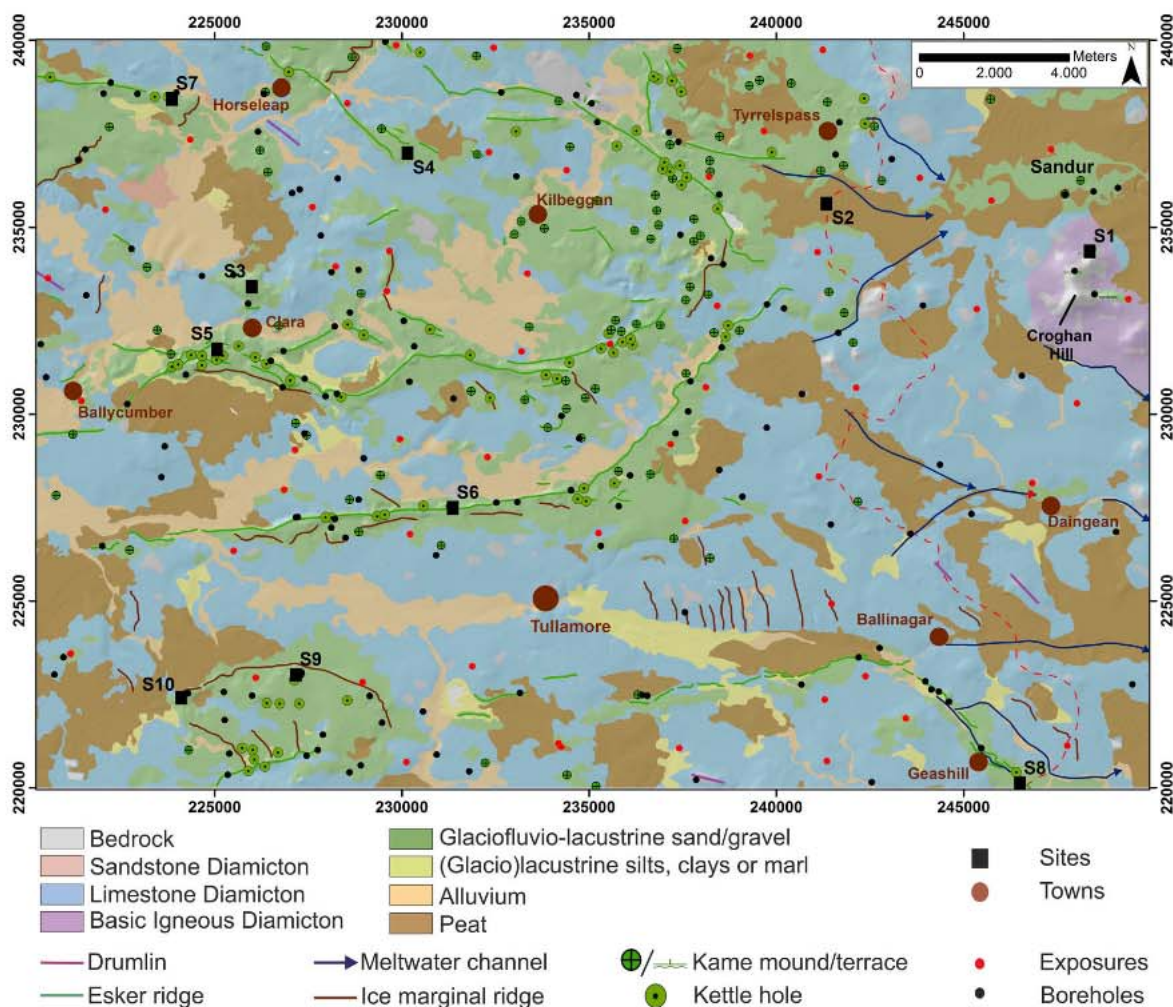
Figure 2. Quaternary Geology map for the study area derived from: (i) desk compilation of previous literature, mapping projects and geotechnical/geophysical surveys; (ii) geomorphological mapping comprising aerial photography interpretation (B&W 1965, 1995 and Colour 2000), analysis of the Digital Elevation Model (DEM) with 10-m pixel size of the Ordnance Survey of Ireland (OSI) and fieldwork; and (iii) field descriptions comprising sample collection and description of 133 exposures, 57 boreholes and geophysical surveys carried out at 17 sites. Coordinates shown refer to the Irish National Grid. This figure is available in colour online at wileyonlinelibrary.com/journal/jqs.

envisaged subaqueous sedimentation in the region west of the study area in a glacial lake with water level at 92 m Ordnance Datum (OD) for at least part of its existence. Sedimentological work on late glacial clays in lacustrine basins identified rhythmites, clays and drop-stones indicative of a glacial lake environment (van der Meer and Warren, 1997). Delaney (2008) identified glaciolacustrine varved sediments deposited during Late Weichselian times (22 10 ka) in a proglacial lake. The integration of geophysical techniques with traditional field mapping methods offers a greatly extended view of the sediments recorded and aids in a three-dimensional recon-struction of the depositional environment.

Methods

The Quaternary geology map of the study area (Fig. 2) is based on geomorphological mapping, lithostratigraphic analysis of exposures, borehole data, previous research (e.g. Farrington and Synge, 1970; Smyth, 1994), mapping projects carried out in the area (Warren and Hammond, 1999; Meehan et al., 2006) and geophysical data (Pellicer, 2010). Field sheets at 1:20 000 derived from the combination of colour aerial photography and a hill-shaded Digital Elevation Model were used for field mapping which involved geomorphological mapping, sedi-mentological description and sampling of exposed sediments.

Exposures showing evidence of glaciolacustrine and glacio-fluvial deposition were identified during the field mapping exercise. Lithostratigraphic characterization is based on the lithofacies coding scheme presented in Supporting information, Table S1.

The geophysical surveys were carried out in areas lacking in exposed sediments, regions buried by peat and/or other postglacial material, and sites that could potentially aid in understanding deglacial processes in the region. Electrical resistivity tomography (ERT) and ground penetrating radar (GPR) were used for geophysical data collection, a customized coding system for classification of radar surfaces and radar facies allowed the production of a relative chronology within each radargram as well as with adjacent ones (Table 1). A full account of the procedures used and the methodology for characterization of soft sediments in the Irish midlands is available in Pellicer and Gibson (2011). Interpretation and characterization of sediment was supported by geomorpholo-gical, geological and geotechnical data when available (Fig. 2).

Results and Interpretation

Ten sites were surveyed in the study area using a range of methods: lithostratigraphic analysis (S1, S5, S6 and S9), ERT (S2 and S8) and GPR (S2, S3, S4, S7 and S10).

Table 1. Coding scheme for interpretation of radar facies and radar surfaces, modified from Neal (2004).

	Description
Radar facies	
Und	Undifferentiated
Gm	Glacial massive diamicton
Gs	Glacial stratified diamicton
Gfm	Glaciofluvial/lacustrine massive sediments
Gfp	Glaciofluvial/lacustrine plane-stratified sediments
Gfc	Glaciofluvial/lacustrine cross-stratified sediments
Gfch	Glaciofluvial/lacustrine channel fill
Gts	Glaciolacustrine topsets
Gfs	Glaciolacustrine foresets
Gbs	Glaciolacustrine bottomsets
Lh	(Glacio)lacustrine fine sediments horizontally stratified
Pt	Peat
Radar surfaces	
bs	Sedimentological, undifferentiated bounding surface
be	Sedimentological, erosional bounding surface
bol	Sedimentological, onlap
sf	Structural, undifferentiated fault
sfn	Structural, normal fault
wt	Water level/change in moisture content
(d)	deformation

Codes are composed of three parts (e.g. L1f1-Gm) where L1 indicates the profile number, f1 the chronological order, f1, f2, f3, and Gm the interpretation (Gm = Glacial massive).

Site S1

Kame terraces, composed of sand and gravel with interstratified diamicton, perched along the NE-facing slopes of Croghan Hill, were identified at Site S1. These deposits consist of interstratified sand and gravel with bedding planes dipping NNE at angles of 20–30° (supporting Fig. S1). Gravel beds are massive and matrix-supported, presenting crude imbrication towards the slope; sandy beds are dominated by massive, fine to coarse sand, but cross-bedded in places. The sediments are frequently deformed by monoclinal folds and cut by normal faults. These kame terraces are encountered at altitudes of 182, 131 and 98 m OD. A large body of glaciofluvial sediments with poor topographic expression, overlain by peat in places, is located along the north foot of Croghan Hill. The sediments are 1–7 m thick and fine gradually eastwards from coarse gravel to fine sand. These deposits are linked to the esker system to the west and are interpreted as sandur deposits fed by large meltwater channels flowing eastwards located along the Shannon Basin watershed and infilling former shallow glacial depressions (Fig. 2).

Site S2

This site is on a peat-covered low point of the watershed between the Shannon and the Boyne Basins immediately east of the area of convergence of the main esker ridges (Fig. 2) and grading gently eastwards (1:330) towards a spread of sandur deposits. ERT and GPR profiles were collected along a line parallel to the watershed to measure the peat thickness and depth to bedrock. ERT profile RS2-10m (Fig. 3a) shows low
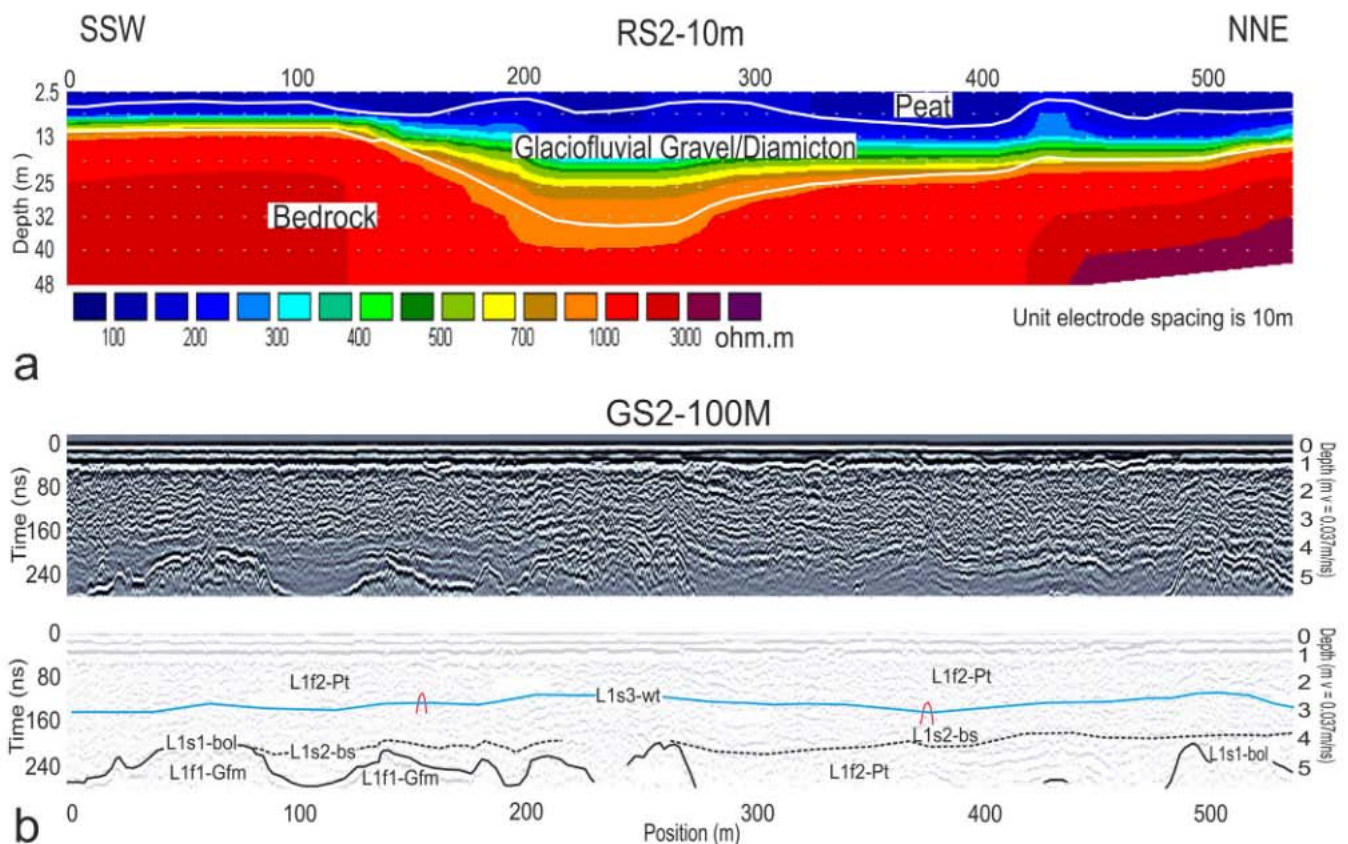

Figure 3. Geophysical data recorded at Site S2. (a) Inverse model of the pseudosection for ERT profile RS2-10m; interpretation of peat thickness is based on the resistivity data, field observations and the GPR radargram. (b) GPR profile GS2-100M: radargram with high pass filter (dewow) and automatic gain control (AGC) (top) and interpretation of the migrated radargram (bottom). Velocity of electromagnetic waves in peat was calibrated at 0.037 m ns[-1] using hyperbolae reflections. Hyperbolae marked in red show the position of draining pipes, which control the water table level in the peat bog. Radar facies codes are described in Table 1.

resistivities ($<150\,\Omega m$) along the surface interpreted as peat, reaching a maximum thickness of over 5.5 m at x-positions 280 430 m. This is underlain by a continuous layer with resistivities of 150 600 Ωm, varying in thickness from 3 to 15 m and corresponding in signature to saturated sand and gravel/diamicton. High resistivities ($>1000\,\Omega m$) underlying these are interpreted as bedrock. A channel-like feature incised in the bedrock is depicted between x-positions 180 and 290 m. The GPR radargram (Fig. 3b) shows a long discontinuous sinuous reflector (L1s1-bol), which is interpreted as the boundary between the peat and the underlying sand and gravel. The discontinuous lines (L1s2-bs) probably relate to changes in the degree of humification. A large channel feature was determined from ERT data, whereas GPR data allowed an estimate of the peat thickness. Maximum peat thickness is offset from the large channel feature incised in the bedrock. The surface level at this location is 95 m OD. This indicates that a potential glacial lake overflow channel draining eastwards across the watershed at this point was at about 89 m OD.

Site S3

A flat-topped isolated hill reaching a maximum altitude of 95 m OD occurs north of Clara Town (Fig. 2). The hill is a slightly elongated feature running west-northwest to east-southeast for 1.5 km, reaching a maximum width of 600 m at its central part and protruding about 40 m from the surrounding landscape. Three GPR radargrams were collected at the south central sector of the hill using 100-MHz antennae (Fig. 4). Three main radar facies were recognized: the lower parts of the radargrams are dominated by subhorizontal, moderately continuous to discontinuous reflectors (f1-Gfp/Gfm); these are overlain by a continuous layer 2 3 m in thickness showing moderately continuous, oblique non-parallel reflectors and a number of curved concave-up reflectors (f2-Gfc), which are interpreted as cross-bedded glaciofluvial sediments and channel-like features, respectively. These are overlain by chaotic discontinuous reflectors (f3-Gfm) with a contact consisting of a continuous reflector (s2-bs) subparallel to the surface. The cross-bedded

sediments and channel-like features of radar facies f2 suggest a subaerial depositional environment whereas the others are less diagnostic.

Site S4

An ice-marginal fan associated with the Horseleap Esker (Fig. 1) reaches a width of more than 500 m and a length over 700 m, and rises 10 15 m above the surrounding landscape, with a surface level 88 93 m OD. The GPR radargrams run parallel (L1) and orthogonal (L2) to the esker ridge (Fig. 5). L1f1-Und, showing poor signal, is overlain in L1 from 15 to 50 m by an area dominated by hyperbolic and discontinuous chaotic reflectors (L1f2-Gfm). This is overlain at 15 20 m by a set of moderately continuous reflectors dipping ESE (apparent dip along the survey line) at angles of 5 20° (L1f3-Gfs) with an onlap contact. The latter are overlain at 50 120 m, with a downlap contact, by a set of continuous sinuous reflectors dipping 10 15° ESE at 50 60 m and 90 120 m (L1f3-Gfs); these reflectors show an apparent dip of similar order to the SSW in L2, indicating a true dip SSE. A radar facies 4 7 m thick composed of moderately continuous reflectors dipping 5 10° WNW at x-position 110 160 m (L1f4-Gfs) overlies f3 with an onlap contact; these evolve into discontinuous oblique non-parallel reflectors (L1f4-Gfc) with hyperbolae from x-position 160 to 218 m, indicating the presence of boulders. Radar facies f1 to f4 are overlain, with an erosive contact, by a 1 2-m-thick layer dominated by discontinuous chaotic reflectors (L1f5-Gfm). The contact between f3/f4 and f5 is interpreted as a topset/foreset interface of a glaciodeltaic system at 89 91 m OD, indicating the glacial lake level at the time this landform formed. Radar facies detected in a flat-topped hill 3 km east of S4 are interpreted as a topset/foreset contact at 88 m OD (Pellicer, 2010). Its position, slightly offset from the esker ridge, and the segmented nature of the esker ridges here, suggest it is a fan, associated with a long beaded esker (cf. Warren and Ashley, 1994), which accumulated sufficient sediment to reach the water surface. And 3 km further east a gravel pit in a low hill (86 m OD) on the south side of the Split Hill esker showed it to
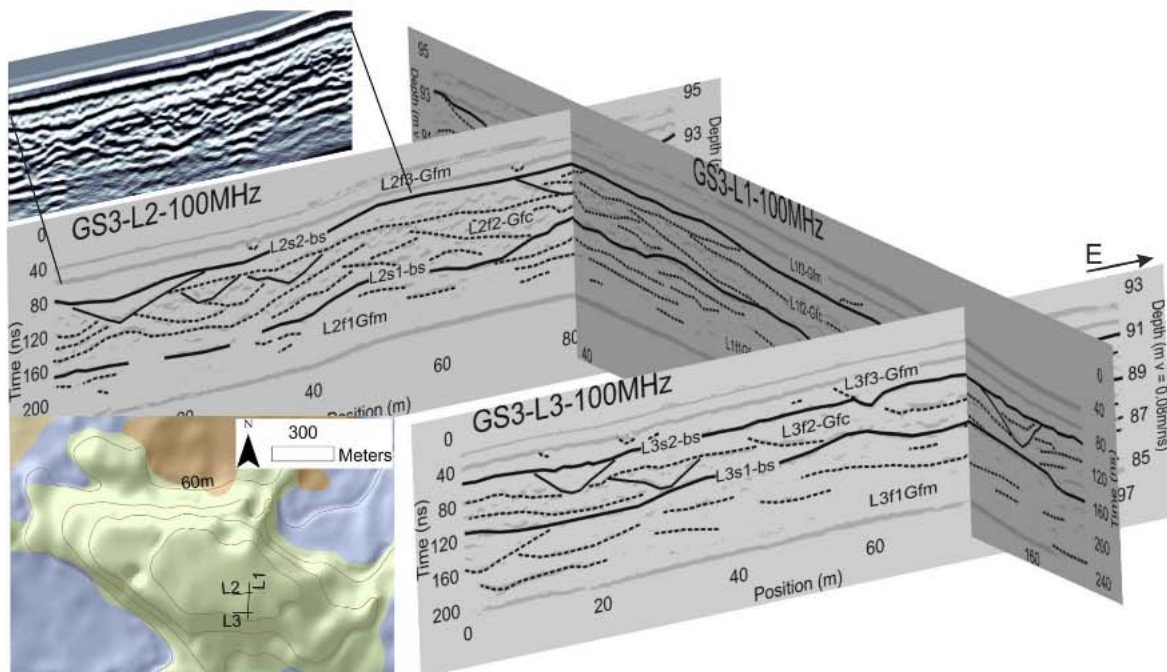

Figure 4. Fence diagram for interpreted GPR profiles collected at Site S3; radargrams are topographically corrected using a velocity of 0.08 m ns [1]. The profiles show main radar facies (continuous black line) and minor sedimentological structures (dashed black line). Location inset from Fig. 2 (10-m contour interval). This figure is available in colour online at wileyonlinelibrary.com/journal/jqs.
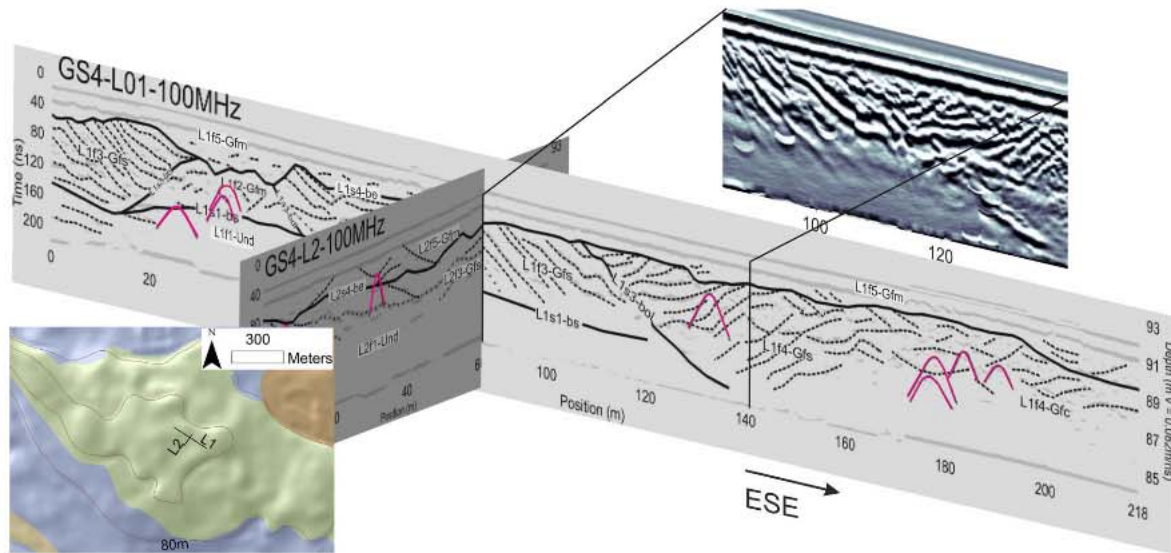
Figure 5. Fence diagrams for interpreted GPR profiles collected at Site S4; radargrams are topographically corrected using a velocity of 0.092 m ns^{-1}. The profiles show main radar facies (continuous black line), minor sedimentological structures (dashed black line) and hyperbolae reflections. Location inset from Fig. 2 (10-m contour interval). This figure is available in colour online at wileyonlinelibrary.com/journal/jqs.

be composed of foreset packages of sand and fine gravel up to 6 m thick and dipping generally to the south-east. This was interpreted as a subaqueous fan associated with a long beaded esker (Ashley and Warren, 1995).

Site S5

Clara Esker is a broad, undulating, east–west aligned feature, 100–1400 m wide, in places composed of several esker ridge tunnel-fills (up to 100 m in width) separated by large kettle holes and locally blanketed by bedded sand and gravel that has been interpreted as deltaic (Warren, 1991b) or subaqueous fan sediments deposited from the south (Warren *et al.*, 2002). Glaciolacustrine deposits overtopping boulder gravel forming the core of Clara Esker were identified in a gravel pit south of Clara. The exposure (supporting Fig. S2) shows foresets composed of interstratified sand and gravel with bedding planes dipping north at an angle of 5–10° in the lower parts (Gfo, Sfo). Palaeocurrents indicating deposition northwards are inferred from the dip of the foresets. These beds are overlain by bedded fine-to-coarse sand interstratified with laminated sand beds with occasional outsized clasts interpreted as dropstones (indicating subaqueous deposition) and massive to bedded matrix-supported gravel. The occurrence of numerous kettle holes in the feature suggests deposition over buried ice (Fig. 2). The exposure is interpreted as a subaqueous fan deposited from south to north in a glaciolacustrine ice-marginal environment banked up against and overtopping an esker tunnel-fill ridge composed of coarse poorly stratified sand, gravel and boulders.

Site S6

Glaciolacustrine deposits overlying a subglacial tunnel-fill are encountered in a large exposure recorded in Ballyduff Esker which can be traced as a generally SW–NE-trending chevron for almost 20 km (Fig. 1). North of Tullamore, the W–E-trending ridge varies in width from 400 to 800 m for a distance of more than 8 km and rises up to 30 m above the surrounding ground, reaching a maximum altitude of 80 m OD. In this area, the ridge is expressed as a depositional system composed of esker tunnel-fill gravel in part overlain by glaciolacustrine sediments (Warren and Ashley, 1994). The tunnel-fill is mainly composed of crudely bedded boulder gravel; boulders are generally very well rounded; boulders up to 2 m in diameter are not

uncommon. At Site S6, and over a distance of 5 km to the west and 4 km to the east, overlapping subaqueous fan lobes composed of interstratified sand and fine to medium pebble gravel with bedding planes dipping generally northwards at an angle of about 20° (supporting Fig. S3a) are built up against the south side of the part-buried tunnel-fill deposits, overtopping them in places; the tunnel fill deposits, whether as freestanding ridge or a buried unit covered by glaciolacustrine sediments, vary in width from 75 to 100 m (see Warren and Ashley, 1994). The fans deposited from the south, orthogonal to the tunnel-fill deposits, indicate a minimum lake level of 80 m OD. The tunnel-fill deposits emerge from the glaciolacustrine sediments as a steep-sided, narrow, continuous ridge of over 5 km length north-east from Tullamore. The exposure presented in Fig. S3b shows its internal structure consisting of crudely bedded cobble/boulder gravel (Bms) dipping up-flow, interpreted as back-set beds overlain by sub-horizontally stratified coarse gravel with boulders.

Site S7

A fan-shaped feature 4 km west of Horseleap near the eastern margin of the Moate Esker is 1300 m long and widens in an east-southeast direction to a maximum width of 800 m (Fig. 2). This landform is flat-topped with kettle holes and reaches an altitude of 85 m OD, protruding 15–20 m above the surrounding ground. Three radargrams are presented in Fig. 6. Three main radar facies are depicted: f1-Gfm consists of low-reflectivity, moderately continuous to discontinuous reflectors. These are overlain by a layer 3–4 m thick expressed as continuous reflectors, subhorizontal and moderately sinuous reflectors in L1 (f2-Gfp) and as oblique tangential, continuous reflectors with an apparent dip ESE at angles of 5–15° evolving eastwards into horizontal sinuous reflectors and interpreted as foresets (f2-Gfs) in L2 and L3. This facies is distorted in L3 from 30 to 70 m by a number of reflectors dipping west interpreted as normal faults associated with the kettle hole formation. Radar facies f2 is overlain with an erosive contact (s2-be) by a layer 1–3 m thick (f3-Gts) composed of discontinuous chaotic and hyperbolic reflectors interpreted as topsets. The contact between f2 and f3 at 82–84 m OD is interpreted as the topset/foreset interface of a glaciodelta indicating the lake water level at the time of deposition.
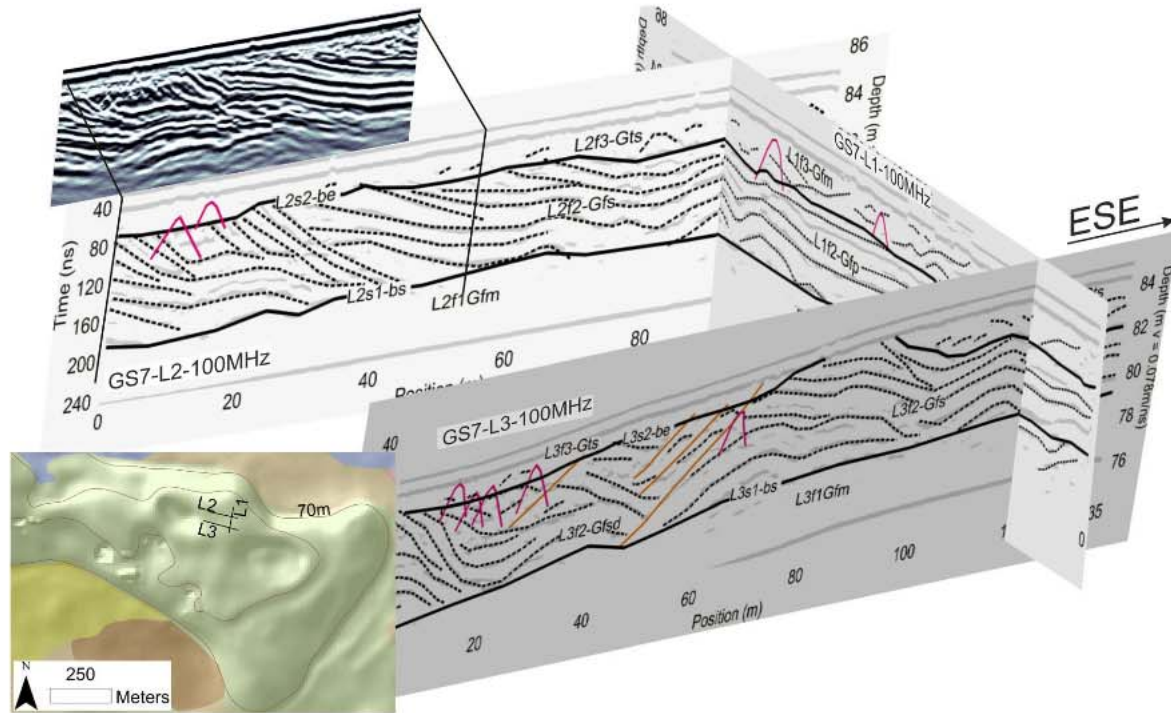
Figure 6. Fence diagram of the interpreted GPR radargrams collected at Site S7; radargrams are topographically corrected using a velocity of 0.078 m ns $^{-1}$. The profiles show main radar facies (continuous black line), minor sedimentological structures (dashed black line) and hyperbolae reflections. Location inset from Fig. 2 (10-m contour interval). This figure is available in colour online at wileyonlinelibrary.com/journal/jqs.

Site S8

South-east of Geashill (Fig. 1), the low point of the watershed between the Shannon and the Barrow Basins runs adjacent to the southern margin of Geashill Esker (Fig. 2). If it operated as a glacial lake outlet it would indicate a former lake level of approximately 84 m OD. Two ERT profiles were recorded along the feature to determine the surficial sediment thickness and their lithological composition (supporting Fig. S4). The profiles were collected inline at 100 m distance from each other and show analogous results. The top region of the profiles is dominated by medium to high resistivity values (500–1500 Ωm) corresponding to glaciofluvial sand and gravel with a varying thickness from x-position 2 to 19 m. These glaciofluvial sediments show continuity southwards where they develop into a large fan-shaped sand and gravel body depicted south of the study area (Warren and Hammond, 1999). A unit with low to medium resistivity values (<500 Ωm) underlying these deposits is interpreted as diamicton. This is underlain by a layer presenting high resistivities (>1000 Ωm) interpreted as limestone bedrock.

Site S9

Kilcormac Esker originates outside the study area. It runs SW–NE for 23 km and 400 m within the area turns sharply to run a further 3 km WSW–ENE. The ridge is 15–20 m high and 50–100 m wide. Exposures recorded along the esker are dominated by poorly stratified well-rounded boulder/cobble gravel with sandy matrix. Kilcormac Esker consists of a continuous subglacial tunnel-fill deposited simultaneously with ice retreat (Farrington and Synge, 1970). A broad spread of glaciodeltaic sediments are exposed immediately to the north along the final 3-km stretch. Several exposures recorded within the glaciodeltaic complex show that the deposits are dominated by well-sorted glaciofluvial sand and gravel. Topsets, foresets and bottomsets were recognized in a face orientated north–south

(Fig. 7); bottomsets are composed of horizontally laminated muddy sand with dropstones. These are overlain by foresets composed of interstratified silt, sand and granules with bedding planes dipping NNE at an angle of 20–30°. Topsets overlying these are composed of interstratified horizontal and cross-bedded gravels. Kame and kettle topography recognized on the glaciodelta indicates the presence of stagnant ice blocks during deposition. The exposures recorded do not show deformation structures, and thus the topset/foreset transition zone recorded at 75–78 m OD is inferred to reflect the glacial lake water level during the deposition of these sediments. Such a level does not agree with any of the potential overflow channels identified in the study area. A glacial lake outlet located along the watershed between the Shannon and the Barrow Basins at 81 m OD, 9 km to the south, may be associated with such a lake level, although the outlet is blanketed by peat of unknown thickness, masking its palaeo-topography during deglaciation times.

Site S10

A 400-m-long isolated mound composed of sand and gravel on the north-west edge of the Kilcormac glaciodeltaic system 3 km west-southwest of S9 attains a maximum altitude of 73 m OD, protruding 10–12 m above the surrounding landscape. Borehole BHS10 shows interstratified sand and gravel dominating the top 16 m (Fig. 8). Three GPR radargrams were recorded on the site (Fig. 8). Profile L1 runs parallel to the feature's long axis; it is dominated by sinuous, subparallel to oblique, discontinuous reflectors (L1f2-Gfc) at x-positions 0–50 and 100–150 m and by parallel, continuous strong reflectors at 50–100 m (L1f2-Gfpd). Lines L2 and L3 run west-northwest to east-southeast across the long axis of the mound, showing analogous radargrams. Moderately continuous subparallel reflectors identified in L3 are depicted at 6 m depth from x-position 0 to 90 m and are interpreted as bottomsets (L3f1-Gbs). These are overlain with a downlap contact by various sigmoidal,
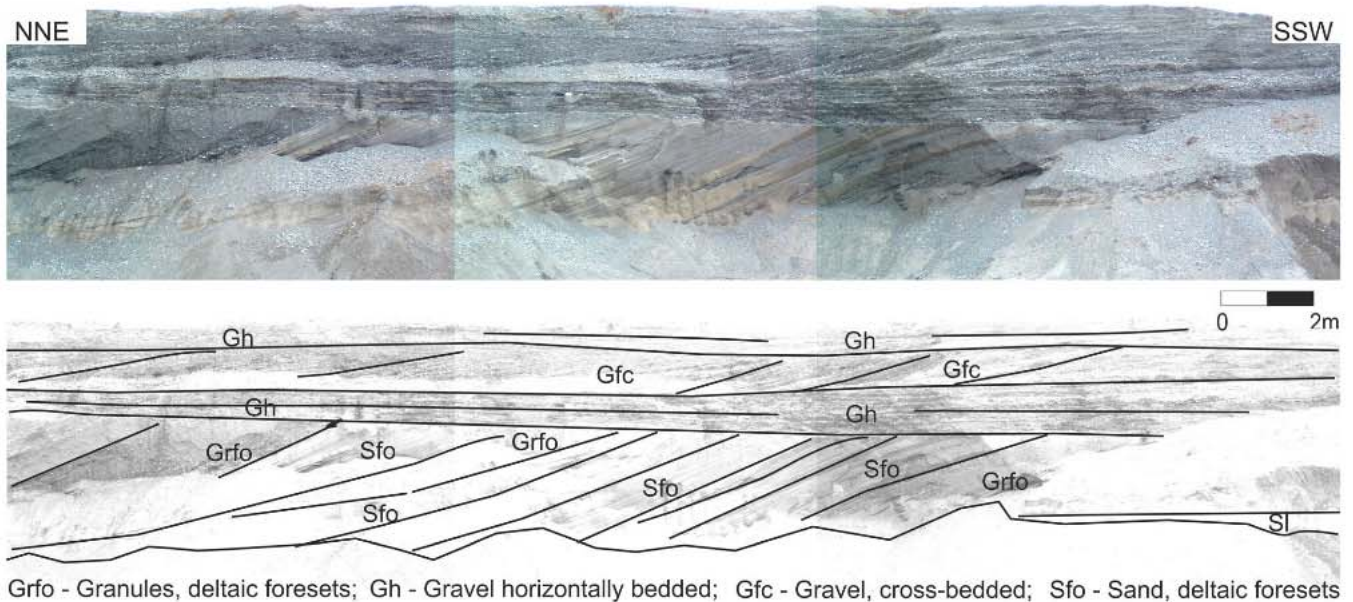
Grfo - Granules, deltaic foresets; Gh - Gravel horizontally bedded; Gfc - Gravel, cross-bedded; Sfo - Sand, deltaic foresets

Figure 7. Exposure illustrating very well-developed topsets and foresets overlying bottomsets recorded on gravel pit (Site S9) excavated on glaciodeltaic sediments deposited north of Kilcormac Esker. This figure is available in colour online at wileyonlinelibrary.com/journal/jqs.

tangential, continuous reflectors dipping WNW (apparent dip) at angles of 10 20° depicted in L2 and L3 and interpreted as foresets (f2-Gfsd). The foresets occur continuously along the profile as far as a convex change in slope at x-position 130 m in L2 and at 90 m in L3. Several sinuous, subparallel, moderately

continuous reflectors dipping SE at angles of 20 30° are depicted along the SE-facing slope. These are interpreted as foresets partly offset by normal faults caused by the collapse of the sediments subsequent to ice withdrawal or by dead ice melt (L2f2-Gfsd).
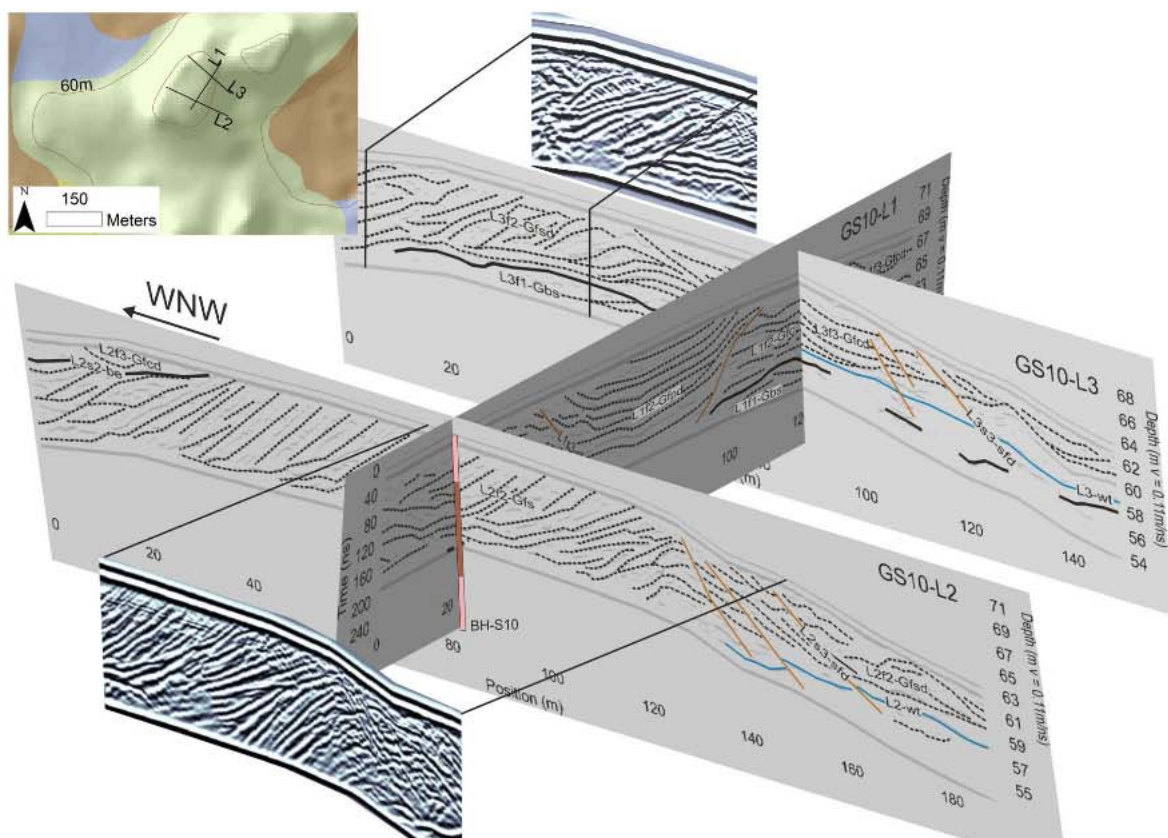

Figure 8. Interpreted radargrams collected at Site S10 presented as a fence diagram. The profiles show main radar facies (continuous black line), minor sedimentological structures (dashed black line), fault planes and water level. Radargrams are topographically corrected using a velocity of 0.11 m ns⁻¹. BH-S10 is composed of fine to coarse sand with well-rounded limestone pebbles, coarsening downwards at 0–4 m, medium coarse gravelly sand with rounded pebbles at 4–12.2 m and soft to medium, dense, fining-downwards sand, with pebbles at 12.2–16.8 m depth. Location inset from Fig. 2 (10-m contour interval). This figure is available in colour online at wileyonlinelibrary.com/journal/jqs.

Discussion

It is clear from the data collected across the area that a substantial portion of the deglacial sediments were deposited as deltas and subaqueous fans into a body or bodies of standing water. We have attempted, by correlating water level indicators, ice marginal ridges and classical tunnel-fill esker ridges, to reconstruct the conditions that would have allowed for this. The topset/foreset interface in deltaic deposits serves as a proxy for a former lake level (see Burbank and Anderson, 2011, Fig. 2.9). Delta topset/foreset interface and corresponding potential outlet channels along with ice contact features that mark points along the former ice margin can be used to define the shape of the retreating ice front.

The kame terraces identified along the N- and NE-facing slopes of Croghan Hill at 182, 131 and 98 m OD (Site S1) indicate a gradual ice thinning along the slope at the onset of the deglaciation of this area.

The topset/foreset interface at S4 is at 89 91 m OD and matches the outlet on the watershed at S2 (89 m OD). The topsetset/foreset interface at site S7 is at 82 m and could relate to either the col west of Daingean (83 m OD) or the meltwater channel interpreted at S8 (84 m).

The interface at S9, which is the most southerly occurrence of a clear topset/foreset contact, suggests a lake level that was controlled by one of the lower watershed cols south of the area of study. As none of the other sets of lacustrine deposits studied shows any clear topsets they must remain less definitive. The lacustrine foresets at Ballyduff (S6) reach 80 m and thus might relate to any of the three potential lake outlets on the eastern watershed. Likewise, those exposed at Clara (S5) at 73 m OD cannot be linked to a definitive outlet. On the other hand, the flat-topped feature at S3, which reaches 94 m OD, shows cross-bedding and probable channel structures at 87 m OD and suggests that any lake in the area at the time of deposition must have had a level lower than that.

We therefore have clear evidence that during the deglaciation of this area an ice-marginal lake formed which was controlled, at least for a time, by the cols on the Shannon/Boyne and Shannon/Barrow watershed, and for a time by outlets to the south of the area. A consideration of the palaeo-flow directions inherent in the bedding structures of the stratified sediment and the disposition of ice-contact landforms reveals a clear pattern in ice retreat across the area. Two very distinct palaeo-flow directions emerge, one indicating flow to the east and south-east (sites S3, S4 and S7) and a second indicating flow between the north-east and north-west (S6, S9 and S10), and point to the disintegration of the ice sheet along two separating ice fronts, one receding to the south-west and the other to the north-northeast (Fig. 9). When the disposition of ice contact ridges (Fig. 2) is taken into account, a more specific pattern becomes apparent, or the interpretation of a more specific pattern is possible, one showing the gradual mutual distancing of the ice fronts one from the other while sustaining an interlobate lake held up by the watershed to the east. A model of deglaciation encompassing five main stages of ice retreat is proposed (Fig. 9).

Stage I of deglaciation is illustrated by the suite of kame terraces built up against the north and east side of Croghan Hill (234 m). The hill is composed of volcaniclastic agglomerate, basalt and Carboniferous limestone which protrude through a general cover of glacial diamicton (till) which contains clasts of limestone, chert, sandstone and volcanics. The kame terraces indicate a gradual thinning of the ice sheet during the early stages of deglaciation of the area. At this stage of deglaciation, with the exception of the emergent Croghan Hill, the study area is covered by the ice sheet.

Diamicton in the watershed area is generally thin; however, it is expressed as three drumlinoids indicating NW SE ice movement in the vicinity of Ballinagar. Meltwater channels dominate the relatively depressed areas (Fig. 2). A large body of glaciofluvial sand and gravel identified north of Croghan Hill (Fig. 9b) composed of coarse sandy gravel, gradually fining eastwards into fine sand (partially covered by peat in Fig. 2), is interpreted as a sandur. These deposits were fed by meltwater channels flowing east and cutting across the watershed between the Shannon and Boyne Basins (Fig. 2). These were large glaciofluvial meltwater channels (Site S2, Fig. 3), connected with the subglacial tunnels in which Split Hill, Clara and Ballyduff eskers formed, and feeding large quantities of sediment towards the sandur deposits at the time the ice margin was located approximately along the watershed; this complex illustrates Stage II of deglaciation (Fig. 9b).

There are three points on the watershed between 87 and 89 m OD through which the ever-widening, insipient lake drained during Stage III. Two main directions of ice retreat (SW and NW) are inferred from the alignment of the eskers and ice-marginal ridges, and the meltwater flow directions reflected in sedimentary structures in the glaciolacustrine sands and gravels. Two ice masses with a suture zone slightly north of Clara Esker began to separate, opening an ever-widening inlet. As the ice receded from the watershed the core esker ridges of cobble to large boulder gravels with muddy matrix at Split Hill, Clara and Ballyduff started to form. These ridges do not extend beyond the Shannon/East Basin watershed. They formed as continuous subglacial conduit infill, time-transgressive and younging generally westward (Warren and Ashley, 1994; cf. Makinen, 2003).

The disposition of the ice-marginal sediments suggests that the most northerly outlet was the first to open and as the lake expanded those further south came into play, slightly lowering the lake level. As the ice fronts withdrew, the glacial lake continued expanding probably as far west as S3. Several glaciolacustrine sediments agree with such lake water level and their disposition enables the definition of the ice margin: (i) the topset/foreset interface at S4 (89 91 m OD); (ii) the topset/foreset interface east of site S4 (88 m OD); (iii) the subaqueous fan associated with Split Hill esker (86 m OD) indicating SE-flowing palaeocurrents (Ashley and Warren, 1995); (iv) cross-stratified sediments and small channel features characteristic of a subaerial depositional environment (Site S3, Fig. 4), although unclear in the radargram, are probably topset beds implying a lake water table at less than 87 m OD.

The glacial lake continued expanding during Stage IV and, as the ice fronts retreated from the watershed, lower lake outlets became available, one west of Daingean at 83 m OD and a second one east of Geashill at 84 m OD (Site S8). The altitude of the former should be treated with caution as its original topography may have been lowered during the construction of the Grand Canal which passes through it (Fig. 1). The sediments deposited at this stage are often associated with kame and kettle topography, indicating fairly rapid collapse of the ice sheets, leaving large masses of dead ice. A topset/foreset interface at S7 (82 84 m OD) with palaeocurrents running east-southeastwards is consistent with these lake outlets. Ice-contact faces rising abruptly 30 m above peat- and lacustrine clay-covered till (Warren et al., 2002) at Clara (S5) and almost 20 m above the till plain to the south of Ballyduff (S6) show subaqueous sediments built up against and overtopping tunnel infill deposits. They demonstrate an ice margin retreating south-west and a lake level of over 80 m OD.

Figure 9. Proposed evolutionary model of deglaciation for the study area. Dashed grey lines indicate the estimated suture zone between the Central and Northern Domes. Arrows illustrate ice movement (grey), meltwater channels (dark blue), glacial lake outlets (blue) and palaeocurrents of glaciolacustrine sediments inferred from cross-bedding (orange). Dots illustrate the position and altitude (m OD) of glaciodeltaic deposits and lake outlets (black) and sub-aqueous fans (brown). Lines show the position of eskers (green), ice-marginal ridges (brown) and drumlins (purple); red dashed line shows the position of the watershed between the Shannon and the Boyne/Barrow basins. Limits of the glacial lake are derived from the DEM contours.
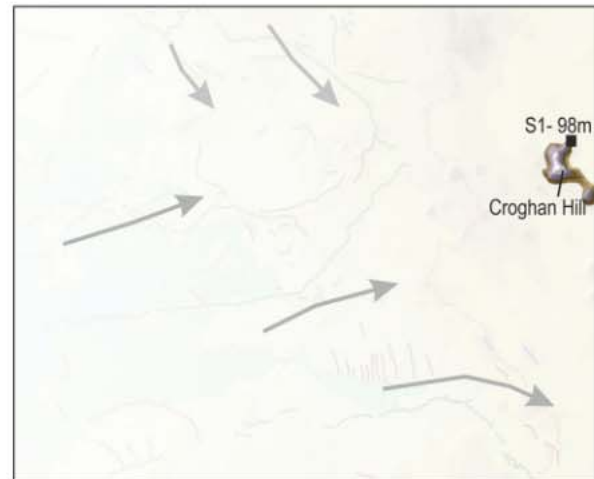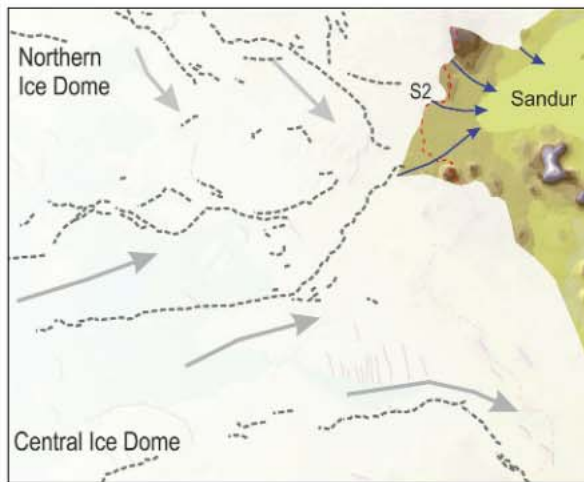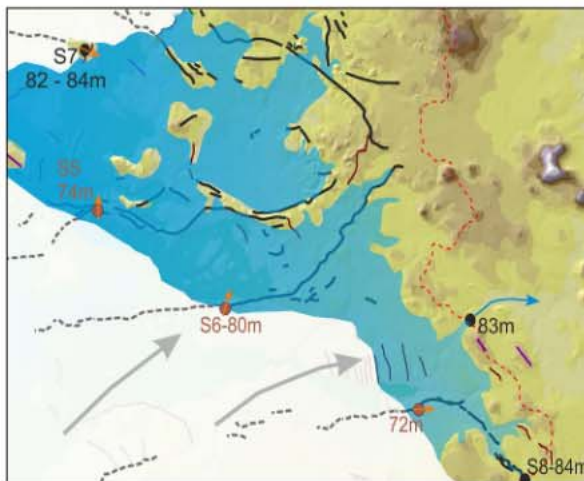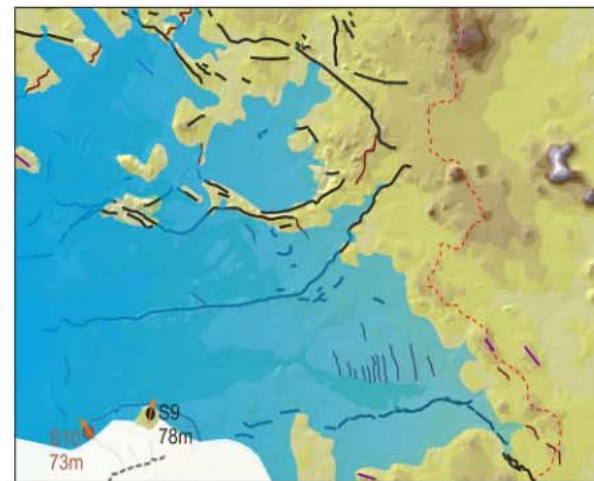
The ice-contact ridge north of the Kilcormac Esker and the associated fan (S10) and delta (S9) deposits show that the ice front retreated southwards from the Ballyduff Esker, while the alignment of the Geashill Esker and the disposition of the moraine ridges east of Tullamore reflect the retreat pattern from Geashill deflected around the northern slopes of the Slieve Bloom Mountains. These show that the ice front retreated first northward as far as Geashill, then west to the Kilcormac Esker and finally south-westward once free of this topographic control.

This ice retreat opened lower-level glacial lake outlets south of the research area; the topset/foreset interface of glaciodeltaic sediments recorded at S9 at an altitude of 78 m OD and subaqueous fans indicate a lake water-level dropping as ice withdrew during Stage V of deglaciation (Fig. 9e). Sites S9 and S10 occur in a large glaciolacustrine sedimentary system that initially formed as a set of subaqueous fans associated with Kilcormac Esker but evolved into glaciodeltaic systems once the sediment reached the lake water level (78 m). Palaeocurrents to the north (S9) and north-west (S10) confirm these

landforms were associated with an ice front located to the south.

The clear evidence across this area of a water-flooded interlobate area between an ice front retreating north-westward on the one hand and one retreating to the south-west on the other renders the simple south north deglacial models of both Gregory (1920) and Charlesworth (1928) redundant. Gregory (1920) recognized the widespread evidence of ice-marginal sedimentation into standing water, but on the assumption of a simple ice retreat to the north-west discounted an ice-marginal lake and postulated a high Lateglacial sea. Flint (1930) also recognized that the sediments were deposited 'chiefly in ponded water' (p. 628), but his proposition that all constructional features were supported on every side by dead ice and that this was part of 'a great mass of stagnant ice standing over the whole of the central lowland' (p. 628), which melted away in place, is contradicted by the consistent northern component in all palaeoflow indicators south of an east west line running just north of Clara Esker and the southern component in those to the north of it, and indeed with the consistent pattern of ice-marginal ridges and contact faces stepping in one case to the south-west and in the other to the north-west. Farrington and Synge (1970) recognized a series of moraines representing halt stages of an ice front retreating to the west and south-west in the area of the Ballyduff, Geashill and Kilcormack Eskers. They seem not to have identified clear evidence of standing water, although they draw attention to similarities with the eskers at Trim (Synge, 1950; Warren and Ashley, 1994), which are characteristically associated with an ice-marginal lake. The most recent attempt to reconstruct the deglacial pattern in the Irish midlands (Greenwood and Clark, 2009b) suggests that an 'unzipping' of two ice lobes, as suggested by Warren (1991b) and Warren and Ashley (1994), and a simple retreat to the west followed by a re-advance from the north-west are equally feasible. Neither of their scenarios (Greenwood and Clark, 2009b, Fig. 9) can accommodate the south to north palaeoflow directions at Clara (S5), Ballyduff (S6) and Skreggan (S9, S10). Nor can either case accommodate the topset/foreset interface levels at S4, the site 3 km east of this or S7. In addition, none of the sediments examined in the area exhibit any evidence of a glacial re-advance or even of ice pushing. The deglacial model outlined by Warren (1991a), Warren and Ashley (1994) and Ashley and Warren (1995) in which ice fronts actively retreated to two major domal centres, a central dome and a northern dome, requires an interdomal lake, but neither of Greenwood and Clark's (2009b, Figs 8, 9) scenarios, although they would result in the emergence of a lake across part of the midlands, fits the data outlined in this paper, particularly the persistent northerly palaeoflows indicated in the sediments south of Clara. The model outlined by Warren (1993) and used in Warren and Ashley (1994) recognized a central dome that was conjoint between Slieve Aughty and Slieve Bernagh to the south-west of the area under study, and Connemara to the west. Ice retreating to such a dome will accommodate all of the features with palaeoflows to the north.

Conclusions

The deglaciation of east central Ireland was characterized by the progressive westerly separation of two formerly conjoint ice lobes, one to the north and the other to the south. These ice bodies supported an ice-marginal lake, the eastern margin of which was the Shannon/Boyne/Barrow watershed on which a number of glacial lake outlets controlled the water level (Fig. 9). Three water levels for this lake are proposed. The highest level is at 88 89 m OD with three possible outlets located along the watershed south of Tyrrelspass. A second lake level related to

outlets occurring at 84 m OD located along Geshill Esker and 83 m OD on the route of the Grand Canal, and a third water level of up to 78 m OD was marked by the topset/foreset interface of glaciodeltaic sediments at Site S9 (Fig. 7), the potential lake outlet of which is located south of the study area. These outlets, the glaciodeltas and the subaqueous fans help delineate the approximate shape of an expanding glacial lake (Fig. 9). Analysis of the collected geophysical and sedimentological data confirms a model of deglaciation involving the gradual separation of two conjoined ice sheets which supported an expanding interdomal lake and has added substantial refinement to it in terms of the relative positions of correlative ice fronts and the water level of the lake as deglaciation progressed. This study confirms in outline the model proposed by Warren (1991b) and Warren and Ashley (1994) and it indicates that there was a much stronger north-eastern component in the flow pattern of the deglaciating ice sheet that lay to the west and south-west than is allowed for in recent models of the last Irish ice sheet (e.g. Greenwood and Clark, 2009b).

Supporting Information

Additional supporting information can be found in the online version of this article:

Figure S1. Exposure in deformed interstratified glaciofluvial sand and gravel, dipping NNE, deposited on a kame terrace positioned along the north-west slope of Croghan Hill.

Figure S2. Cross-section of glaciolacustrine sediments overlying Clara Esker ridge.

Figure S3. Exposure in the Ballyduff Esker complex and longitudinal profile of Ballyduff Esker ridge.

Figure S4. Inverse model of the pseudosection for ERT profiles RS8-L1-10m and RS8-L2-10m and interpretation of the datasets.

Table S1 Lithofacies coding scheme used for interpretation of exposures, modified from Benn and Evans (1998).

Abbreviations. DEM, Digital Elevation Model; ERT, electrical resistivity tomography; GPR, ground penetrating radar; LGM, Last Glacial Maximum; OD, Ordnance Datum.

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