Irish Ice Sheet dynamics during deglaciation of the central Irish Midlands: Evidence of ice streaming and surging from airborne LiDAR

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Abstract

High resolution digital terrain models (DTMs) generated from airborne LiDAR data and supplemented by field evidence are used to map glacial landform assemblages dating from the last glaciation (Midlandian glaciation; OI stages 2-3) in the central Irish Midlands. The DTMs reveal previously unrecognised low-amplitude landforms, including crevasse-squeeze ridges and mega-scale glacial lineations overprinted by conduit fills leading to ice-marginal subaqueous deposits. We interpret this landform assemblage as evidence for surging behaviour during ice recession. The data indicate that two separate phases of accelerated ice flow were followed by ice sheet stagnation during overall deglaciation. The second surge event was followed by a subglacial outburst flood, forming an intricate esker and crevasse-fill network. The data provide the first clear evidence that ice flow direction was eastward along the eastern watershed of the Shannon River basin, at odds with previous models, and raise the possibility that an ice stream existed in this area. Our work

demonstrates the potential for airborne LiDAR surveys to produce detailed paleoglaciological reconstructions and to enhance our understanding of complex palaeo-ice sheet dynamics.

Keywords: LiDAR; Ireland; crevasse-squeeze ridges; ice surging

1. Introduction

Airborne LiDAR (Light Detection and Ranging) surveying allows the remote sensing of earth surface topography at unprecedented high spatial resolutions (<1 m horizontal resolutions and vertical resolutions of <0.25 m). This allows the relatively swift identification and precise measurement of archaeological and geomorphological features at a higher level of detail and accuracy than that achieved by traditional field study methodologies (Slatton et al., 2007; Roering et al., 2013).

In paleoglaciology, digital terrain models (DTMs) of >20 m horizontal resolution generated from satellite radar surveying have been central in the mapping of ice sheet subglacial bedforms and reconstructing the changing dynamics of ice sheets(e.g., Clark, 1993; McCabe et al., 1998; Greenwood and Clark, 2009a; Clark et al., 2012; Hughes et al., 2016). In Ireland, DTMs generated from radar, aerial photos, and satellite images have been used to map subglacial bedforms from the last Irish Ice Sheet (IIS) (Midlandian; OI stage 2-3; McCabe et al., 1998; Knight et al., 1999; Clark and Meehan, 2001; Greenwood and Clark, 2008). This approach has been used to infer bedform 'flowsets' (Clark and Meehan, 2001; Greenwood and Clark, 2008, 2009a) related to ice sheet flow phases and thus reconstruct ice dynamics where bedforms (e.g., drumlins) and contemporaneous ice marginal positions (e.g., terminal moraines) are well preserved.

Very low-amplitude features (<1 m high) are unlikely to be recognised on radar data or photographs, especially if they have poorly defined slope breaks (Smith et al., 2006). This category of topography includes gently undulating and low-amplitude hummocky terrain, which commonly occurs in marginal zones of former ice sheets and is widespread across the Irish Midlands (Synge and Stephens, 1960; Synge, 1979; Warren, 1992; Meehan, 1999; Delaney, 2001). However, the evidence is increasing that higher resolution spatial data can significantly improve the detection of large- and small-scale glacial features, improving the accuracy of landform interpretation (e.g., Cline et al., 2015; Dowling et al., 2015). Small-scale features — including controlled and uncontrolled hummocky terrain, ice-flow transverse ridges, and crevasse squeeze ridges — often record critical changes in ice sheet dynamics at the time of formation, including in basal thermal regime, ice flow velocity and direction, and ice marginal readvances (e.g., Eyles et al., 1999; Evans et al., 2008; Ottesen et al., 2008; Evans, 2009; Andreasson et al., 2014). Recent evaluations of high-resolution multibeam acoustic surveys in offshore areas and airborne LiDAR onshore indicate that such features are critical to understanding evolving ice dynamics during overall ice sheet recession during glacial terminations (e.g., Andreasson et al., 2014; Bjarnadóttir et al., 2014; Cline et al., 2015; Möller and Dowling, 2015).

In this study we use two high-resolution DTMs generated from airborne LiDAR data to map the glacial geomorphology of two nearby areas in the central Irish Midlands. The mapping is supplemented by field investigations of sediment exposures, producing morphosedimentary evidence of glacial events in the region during the last glaciation. The evidence is used to generate new palaeoglaciological models of ice sheet dynamics in a region where previous interpretations of ice sheet dynamics have been based almost entirely on the large-scale pattern of glaciofluvial evidence.

2. Regional setting

2.1. Regional geology and topography

The central Irish Midlands consists of a low-relief plain (~40-70 MOD, meters above ordnance datum), drained by the River Shannon and its tributaries (Fig. 1). The plain is underlain by Carboniferous limestones, with inliers of Paleozoic sandstones and siltstones (Sevastopulo and Wyse-Jackson, 2009). The area is poorly drained but tilts generally westward from a watershed at

~80 MOD, ~12 km east of Tullamore, County Offaly, to ~37 MOD at the River Shannon. The plain is defined to the south by the rising topography of the Slieve Bloom Mountains (480-514 MOD) and to the north and east by gently rising, NE-SW striking (bedrock controlled) topography.

2.2. Regional glacial landform distributions

The distribution of larger glacial landforms dating from the last (Midlandian) glaciation across the Irish Midlands is well documented (e.g., Sollas, 1896; Synge and Stephens, 1960; Synge, 1979). The dominant glacigenic landforms in the study area are well developed (≤60 m high and up to 40 km in length) west- to east-trending eskers (Fig. 1; Sollas, 1896; Farrington with Synge, 1970; Warren and Ashley, 1994; Pellicer et al., 2012). Smaller SE-trending eskers occur along the northern margin of the basin (Delaney, 2001a,b, 2002), and short SE-aligned segments also occur within the west- to east-trending esker group (Gallagher et al., 1996). Moraines are rare in the area, and reconstructions of ice marginal positions through time are based primarily on the occurrence of ice-contact deltas and subaqueous outwash fans at the downstream ends of esker conduit deposits, together with some minor, short moraines (Delaney, 2001a,b, 2002; Pellicer et al., 2012). Drumlins and mega-scale glacial lineations (MSGLs) occur on the higher ground around the basin margins but are rare on lower ground (Greenwood and Clark, 2008, 2009a; Fig. 2A). Inter-esker areas are partly covered by glaciolacustrine deposits, overlain by Holocene peat and alluvium, thought to have been deposited in a topographically controlled, proglacial, ice-contact lake (Paleolake Riada). This lake formed as ice receded westward and downslope during deglaciation and was dammed by ice to the west and northwest and by rising topography to the north, east, and south. At its maximum extent the lake drained through a col at 82 MOD, 10 km east of Tullamore. This is also the height of many ice-contact deltas around the lake basin, indicating a relatively stable lake surface level (Delaney, 2002, 2007; Pellicer et al., 2012). Further recession of ice westward is presumed to have allowed drainage of the lake southward around the western margin of the Slieve Bloom uplands.

Between the glaciolacustrine basins, flat, undulating, and hummocky terrain is underlain by a mix of glaciofluvial sand and gravel deposits and sandy-silt and silty-clay diamicton, both rich in limestone (Pellicer et al., 2012). Diamicton thicknesses, where known, are generally 1-10 m; and hummocky terrain is particularly concentrated around the northern and western margins of the basin but also occurs in other areas around the eskers (Fig. 1; GSI, 2016).

(FIGURE 1 HERE)

2.3 Regional palaeoglaciological models

The absence of widespread indicators of active ice flow, together with the dominance of glaciofluvial sediments, means that the glacial morphosedimentary sequence in this area has been interpreted as a deglaciation landscape, formed by ice recession and sedimentation into an expanding ice marginal lake (Pellicer et al., 2012). This absence of evidence for active ice flow has led to contestation over the interpretation of ice recession direction (Fig. 2). Warren and others (Warren, 1992; Warren and Ashley, 1994; Pellicer et al., 2012) have used water palaeoflow directions, constructed from subglacial esker and ice-marginal kame sediments, to suggest that the esker Riada and the Ballyduff esker formed contemporaneously in an interdomal position. This model requires an ice dispersal centre to the southwest feeding ice northeastward, and a second dispersal centre to the northwest feeding ice flow southeastward (Fig. 2B). An alternative multiphase model, involving ice flow eastward and initial recession westward, followed by readvance from the north has also been proposed (Synge, 1952; Delaney, 2002; McCabe, 2007). Mapping of subglacial lineations in the area by Greenwood et al. (2009a, b) has failed to resolve this issue (Fig. 2A), which is of critical importance in the construction of regional palaeoglacological models (cf Warren et al., 1992).

(FIGURE 2 HERE)

3. Methods

LiDAR data acquired during 2011 for flood risk assessment and infrastructure planning were obtained for three areas in the Irish Midlands. For areas 1 and 3 (see below), the data were processed to remove survey artefacts, vegetation cover (using the last return of the laser survey signal), and anthropogenic structures; and the resulting point cloud data was used to interpolate DTM rasters for each area of a nominal 0.5-1 m horizontal and 0.12-0.25 m vertical resolution. For area 2, the data had been previously processed in a similar manner and was already in raster format.

The DTM rasters were relief-shaded using the Relief Visualisation Toolbox application, a programme originally developed for archaeological surveys (Kokalj et al., 2011; Zakšek et al., 2011). Relief-shaded terrain models were generated using the multidirectional shading tool, which composites relief-shade models lit from 16 different azimuths. Incident light elevation angle was set between 20 and 25°, with a vertical exaggeration of up to x4 applied to exaggerate low-amplitude features. In addition, where landforms were particularly faint, a principal component analysis (PCA) of the original hillshaded image was undertaken. This procedure shows the three main components of the image only and minimises noise (Kokalj et al., 2011). Landforms in areas immediately outside the DTM footprints were mapped where appropriate, using a combination of stereo pair and single vertical aerial photos.

Mapping of candidate glacial landforms from DTMs was field checked and supplemented by field surveying. During field surveying, available sedimentary exposures were photographed and logged; and fabric data on clast orientations in diamictons were obtained where appropriate.

4. Results

4.1 Area 1: Tullamore

The LiDAR DTM covers a low-relief (54-82 MOD), trapezoid area of 13 km by 5 km around the Tullamore River, centred on Tullamore town (Figs. 1, 3). About 9 km^2 of the central part of the area is urbanised, and construction has resulted in some landscape alteration and the removal of small-scale landforms. At its margins, the data patch clips the southern flank of the Ballyduff esker to the north and the Blackwood ice-contact delta-kame (Pellicer et al., 2012) to the southwest.

Warren and Ashley (1994) suggested that the area was covered by ice feeding from an ice dome to the southwest and interpreted ice flow direction as northeast. However, Farrington (with Synge, 1970) and Warren (1987) identified small moraines in the area, which indicate ice recession was west to southwest, implying ice flow was roughly eastward, but deflected northward around the Slieve Bloom mountains. More recently, the GSI have mapped drumlins trending southeast across the area (GSI, 2016; Fig. 1).

(FIGURE 3 HERE)

The following glacial features were identified on the LiDAR DTM.

4.1.1. Lineations

Multiple (<10) low-amplitude, highly elongate, streamlined ridges with subparallel long axes and associated grooves trend eastward (average bearing 083°) across Area 1 (Figs. 3, 4, 5). Smooth ridges crests are between 0.3 and 4 m higher (average height 0.98 m) than surrounding terrain, with ridge widths between 50 and 116 m and measured lengths of 1.9-4.7 km. Some ridges show abrupt truncation or extend beyond the DTM coverage, so maximum ridge length is greater. Minimum elongation ratios are >1:30. Five ridges with well-defined, wider, rounded western ends are visible in the eastern part of area 1, and additional traces of lineations in the form of grooves are also visible in places (Figs. 3, 4, 5). Two ridges in the north of the study area extend eastward from a protuberance (Fig. 3A); however, the two most southerly ridges are unrelated to any prominent bedrock irregularities and have subtle seeding points, widening over 500 m to their maximum width and height before tapering out gradually eastward (Fig. 5). Judging by the occurrence of faint linear traces on aerial photos, the ridges and grooves appear to form part of a larger group of these features that continues eastward.

(FIGURES 4 AND 5 HERE)

A roadside exposure measuring ~1 m high by 10 m long crosses one well-developed lineation orthogonally (X on Fig. 3A) and shows one lithofacies, a limestone clast-rich diamicton with a pale yellow, silty matrix (Fig. 6A). The diamicton contains fractured clasts with angular clast slabs off-lapping each other on centimeter-scales progressively upward to ~85° (Fig. 6B).

(FIGURE 6 HERE)

Interpretation: The consistent ridge long-axis azimuth indicates formation under a persistent and strong controlling mechanism. The low-amplitude, highly streamlined elongate morphology and the occurrence of a tectonically deformed diamicton, with sheared clasts indicating a high normal confining stress and shear stress applied in the direction of ridge long axes, both indicate modification of a diamicton in a highly stressed, confined environment. The ridges and accompanying grooves are interpreted to form part of west- to east-trending mega-scale glacial lineations (MSGLs). The precise mechanism of formation of MSGLs is debated; however, the landforms are thought to form from a combination of erosion, transfer, and deposition of diamicton in a deforming subglacial bed, in association with accelerated ice flow (Ó Cofaigh et al., 2013). These features are discussed further below.

4.1.2. Cross-cutting ridge sets

Overlapping sets of low-amplitude ridges occur over large parts of the study area. Preservation of the ridges is partial, but where best preserved they form a rectilinear pattern (Figs. 4, 5).

Three ridge orientations can be identified on the LiDAR DTMs. The most prominent ridge set (ridge group A) are aligned N-S to NNW-SSE (mean alignment 343-163°) and were partly mapped by Farrington with Synge (1970) and Warren (1987). The ridges are arranged in an east-west oriented belt, with further ridges detectable on aerial photos up to 4 km to the east, but not north or south of the DTM area. Group A ridges have rounded or slightly flattened crests, with sideslope angles of <25°, are between 300 and900 m long (average length 558 m; aligned segments traceable for up to

2500 m), up to 8 m high, and between 40 and 100 m wide. Ridge crestlines are increasingly broken westward. Spacing between ridges is between 53 and 157 m, with an average spacing of 112 m (Fig. 3C). Ridges are generally straight, becoming slightly arcuate (convex to the east) along the southern margin of the group. Minor exposures indicate that the ridges are composed of diamicton. A 4-m-long trench cut in undulating terrain at the western limit of the ridge distribution (Y on Fig. 3A (Irish National Grid (ING) 229813, 225513; UTM 596368m E 904157m N) is shown in Fig. 7. A basal matrix-poor boulder breccia, containing angular and often fractured limestone clasts (facies A) and presumed close to bedrock, is overlain by a crudely stratified, highly consolidated, matrix-rich, grey-pale brown diamicton (facies B). Clasts are pebble to boulder-sized, subangular to subrounded, occasionally striated and include pale and dark blue limestones (>90%) and yellow sandstones (<10%). Clasts exhibit a well-developed, bimodal orientation. The uppermost facies (C) consists of massive, yellow sandy silts with sparse, generally boulder-sized clasts of angular pale limestones.

(FIGURE 7 HERE)

A second set of ridges aligned at 017-197° (ridge group B) overlies the group A ridges, forming a partially preserved rectilinear grid in this area (Figs. 4, 5). Group B ridges are shorter than group A types (mean length 227 m), lower (heights <5 m) and narrower (20-30 m width), and have less clearly defined basal slope breaks. A third group of ridges aligned at 127-307° (group C) occur throughout the area. These have a mean length of 189 m, are under 5 m high and are 30 m wide.

The ridges do not appear to have been deposited in a distinct chronological sequence. Where ridges intersect, a groove is often visible in the underlying ridge, indicating reworking of material from the lower ridge (arrows on Fig. 4, 5). No consistent pattern of reworking occurs — for example, in Fig. 4, group A ridges overlie (X on Fig. 4) and are overlain by Group B ridges (Y on Fig. 4).

Interpretation: Small transverse ridges are very common in glaciated terrain and form in a number of ways, including annual and subannual, ice-marginal push and squeeze ridges (Price, 1970; Sharp, 1984; Kruger, 1994; Evans and Hiemstra, 2005); controlled moraine formed by supraglacial meltout of debris-rich bands within the ice margin (e.g., Evans, 2009); and subglacially, either as crevasse-squeeze ridges (CSRs) formed in flow-transverse crevasses (including washboard moraine, e.g., Cline et al., 2015) or as ribbed moraine formed of meltout till and linked to a thermal transition from warm- to cold-based ice within the ice sheet (e.g., Möller and Dowling, 2015).

Any interpretation must explain two features: firstly no evidence of reshaping of the ridges by active ice flow is visible, as expected if overriding had occurred; and secondly group A ridges appear to overlie and underlie group B and C ridges. One possibility is that the lower transverse ridges were formed ice-marginally as push and squeeze ridges (Price, 1970; Sharp, 1984; Kruger, 1994; Evans and Hiemstra, 2005). However, if this was the case, a readvance of ice across the ridges would be necessary to form the overlying oblique ridges, a process that normally involves fluting and reworking of material to form asymmetric ridge profiles and arcuate planforms, features not seen in the flow-transverse ridges (Sharp, 1984; Benn and Evans, 2012). Formation as controlled moraine by supraglacial meltout of debris-rich bands within the ice margin (Evans, 2009) is also discounted, as again, reworking into upper ridges would not be possible without readvance and reshaping.

Instead, formation of the entire ridge network subglacially is preferred and is supported by the diamictic composition of the ridges and well-developed fabric. Two possibilities exist. The first possibility, formation as ribbed moraine formed of meltout till and linked to a thermal transition from warm- to cold-based ice within the ice sheet (e.g., Möller and Dowling, 2015) is again unlikely, as such ridges are usually strongly asymmetric and laterally variable in profile because of fluting of the surface.

Instead, we consider formation as crevasse-squeeze ridges (CSRs; Sharp, 1985) to be the most likely mechanism for all the ridge sets, with formation of the three different ridge orientations taking place near-concurrently but after formation of the underlying MSGLs. Crevasse-squeeze ridges forming geometrical networks were originally identified in front of modern surging glaciers and are considered a diagnostic feature of ice stagnation after a period of accelerated, extensional ice flow (Sharp, 1985; Evans and Rea, 1999, 2003; Evans et al., 2007). The CSRs are thought to form by the injection of wet basal sediments upward into extensional crevasses, under high basal water pressures (Sharp, 1985; Rea and Evans, 2011). Similar rectilinear ridge networks have been observed on the beds of modern and Quaternary temperate and polythermal glaciers and are associated with surging glaciers and ice streams in onshore and offshore situations (Boulton et al., 1996; Ottesen et al., 2008; Ó Cofaigh et al., 2010; Andreasson et al., 2014; Jónsson et al., 2014; Evans et al., 2014, 2016; Cline et al., 2015; Flink et al., 2017). Ice-flow transverse (set 1) ridges in area 1 are somewhat larger than usual; however, similar larger transverse ridges have been observed in front of modern surging glaciers in Svalbard (Boulton et al., 1996; Flink et al., 2017) and on the bed of former ice streams (Evans et al., 2016).

4.2. Areas 2 and 3: Birr

These areas are located in the southwest of the study area around Birr, in a low-relief area around the Little Brosna and Ballinurig rivers, which drain WNW toward the Shannon River. The available high resolution LiDAR data cover an irregularly shaped polygon (~59 km^2) north of Birr (area 2; Figs. 1, 8, 9) and a 2 km^2 square west of Birr (area 3; Figs. 1, 10). Both areas have a low relief range (between 40-60 MOD, excluding prominent esker ridges). Surfaces in this area below 50 MOD are characterised by raised bog, reclaimed bog, and alluvial floodplain, while glacial features form the areas above 50 MOD.

(FIGURE 8 HERE)

(FIGURE 9 HERE)

Previous studies have identified a complex of glaciofluvial ridges oriented west to east and southwest to northeast (Gallagher et al., 1996; Greenwood and Clark, 2009a; GSI, 2016; Fig. 1), with a set of NNW-SSE–oriented ridges interpreted as recessional moraines by Gallagher et al. (1996). The glaciofluvial ridges have been interpreted as conduit fills (eskers) leading to subaqueous ice-marginal deposits formed in a proglacial, ice contact lake, with changing esker orientation reflecting deflection of an eastward-flowing ice sheet around the Slieve Bloom massif to the southeast (Gallagher et al., 1996). Hummocky terrain is present around the eskers and has been mapped as either hummocky sand and gravel or as underlain by limestone till (GSI, 2016). Drumlins and MSGLs have been mapped to the west and northwest of both areas (Fig. 1).

(FIGURE 10 HERE)

The following glacial features were identified on the LiDAR DTMS.

4.2.1. Lineations

Elongate, closely spaced ridges and grooves trending at 307-127° similar to those seen in area 1 are visible in the northern part of area 2 (Figs. 8, 9, 11) and are interpreted as MSGLs. Ridges have similar heights and widths to those in area 1 but are shorter, between 700 and 1600 m (average 1055 m) in length, with an elongation ratio of >1:13. However, they are partly truncated and overlain by hummocky terrain to the south, so original lengths are likely to have been greater. No good exposures were found, but drilling by the Geological Survey of Ireland has enabled mapping of the area as 'limestone till'.

(FIGURE 11 HERE)

Two drumlins were also identified (Figs. 8, 9, 11). One immediately NW of the MSGLs has been mapped by the GSI and Greenwood et al. (2008) within an area of limestone till; a second along the southern margin of the DTM was mapped by the GSI (2016) as a sand and gravel deposit but has a fluted surface.

4.2.2. Hummocky terrain

Hummocky terrain (HT) occurs as 1-2 km^2 patches in area 2 and covers all of area 3 (Figs. 8, 10). These areas have been mapped previously by the Geological Survey of Ireland (GSI 2016). We classify HT into three main types on the basis of morphological characteristics visible on the LiDAR DTMs.

Hummock terrain type 1 (HT1): HT1 is present in the northern part of area 2 (Figs. 8, 9, 12A,B), immediately downstream of the MSGLs, and throughout area 3 (Fig. 10). It is confined to higher ground and appears to have been partly eroded by post-glacial fluvial action. The HT1 consists of ridges and mounds <5 m high, with rounded or flattened crests, and with intervening hollows and grooves. Mounds and ridges and grooves exhibit preferred orientations.

Where HT1 occurs in area 2 two prominent ridge and groove orientations are visible (320-140° and 235-055°) with two minor orientations at 275-095° and 350-170° (Figs. 12A,B; 13A,B). Truncation of ridges by grooves indicates that the oldest orientation is at 320-140°, parallel to MSGLs to the north (shown in Fig. 11), and these features are interpreted as remnants of MSGLs; they are round crested and relatively wide (>50 m), with height differences between crest and trough of between 0.8 and 1.5 m. Of the remaining ridges and grooves, the largest group is aligned transverse to the MSGLs, at 235-055°. Larger ridges in this group (height 2-4 m; width 40-120 m) have flattened tops, well-defined basal slopes, and are steeper on the distal (downice) side; smaller ridges are narrower, lower and near-symmetrical. Minor exposures indicate that the ridges are underlain by diamicton containing fractured clasts, but with no evidence of directional shearing. All ridges vary in width and height along their length. A third ridge alignment at 350-170° appears in places to have formed from reshaping of the flow-transverse ridges into flutes. The final orientation is dominated by grooves oriented at 275-095° that truncate other ridge and groove orientations.

Area 3 has a similar pattern, with MSGL remnants aligned at 305-125° overlain by a set of closely spaced ridges aligned at right angles (215-035°; Figs. 10A, B). As in area 2, morphology varies with size, and some fluting has occurred. Two further ridges aligned at 0-180° in the northern part of the DTM are 70-120 m long, up to 6 m high, with well-defined basal slope breaks and sharp-crested and flattened tops.

(FIGURE 12 HERE)

Hummocky terrain type 2 (HT2): This type of hummocky terrain is visible in area 2 (Figs. 12C,D; 13C,D). It contains features in common with HT1 topography, i.e., is primarily formed from MSGL remnants overlain by transverse ridges at similar orientations to those seen in HT1 topography in area 2. In addition, higher, sharp-crested ridges with slightly sinuous crests also occur and are interpreted as remnants of conduit fills (eskers). The principal difference from HT1 is the presence of discontinuous, anastomosing channels cut into the hummocky surface (Figs. 12C,D). These are commonly flat-bottomed with steep sides and undulating long profiles that rises southeastward. They are interpreted as subglacial meltwater channels.

(FIGURE 13 HERE)

Hummocky terrain type 3 (HT3): This type of hummocky terrain is present in the southern part of area 2 (Figs. 12E,F) and is mapped as hummocky sand and gravel by the GSI (2016). It is bounded by eskers to the west, south, and east (GSI, 2016). The HT3 consists of multiple, short, sinuous ridges forming an interconnected network. These ridges have been partly anthropogenically modified, but the overall pattern of ridges is preserved. The dominant ridge orientation is 335-155°, parallel to the eskers to the east and west, and subparallel to the MSGLs to the north. Shorter, sharp-crested, 1-7 m high discontinuous sections with well-defined basal slopes and exhibiting low sinuosity oriented at 235-055° are also present. Traces of a third, NE-SW–oriented set of straight ridge segments can be detected underlying the sinuous ridge set. The ridges are similar in appearance to conduit fill eskers.

Interpretation: We consider the varieties of hummocky terrain seen here to be multigenetic in origin, involving subglacial squeezing of wet sediments and subsequent glaciofluvial erosion and sedimentation. The HT1 and HT2 are interpreted as a continuum of subglacial landforms, formed under soft bed conditions but under varying ice flow regimes. The MSGLs indicate accelerated ice flow southeastward across the area. They have been partly eroded during emplacement of the overlying ridges. The overlying flow-transverse ridges have an asymmetrical cross-profile and may have originated as ice-marginal features. However, the presence of fluting, grooves, and oblique superimposition of further ridges supports a subglacial origin; and we consider HT1 and HT2 to have formed as CSRs, i.e., by the subglacial squeezing of wet basal sediment into tensional crevasses, in a similar manner to the ridges in area 1. In areas 2 and 3 the relatively poor preservation of MSGLs is considered the result of a more aggressive reworking of sediment into closely spaced crevasses, and the further overprinting of ridge sets because of to the continuation of active ice flow during formation of the overlying ridges and grooves. This characteristic is shared with ribbed moraine described by Möller and Dowling (2015). In addition, HT2 appears to have been eroded subsequently by channelized subglacial meltwater flow in Nye-channels, with channels cut generally subparallel to local esker azimuths (see below).

The HT3 differs in form to HT1 and HT2 but also contains evidence for channelized meltwater flow within the ice sheet. The area is underlain by sand and gravel (GSI, 2016), and the sinuosity and composition of the larger ridges in this area, together with links to adjacent eskers, indicates that these are also ice-walled conduit or channel fills. We consider the alignment of the sinuous ridges and the occurrence of linear ridges and grooves within HT3 to indicate formation within crevasses, either supraglacially or subglacially. Similar ridge networks have been described at modern glaciers where subglacial water has been diverted into crevasses within a fractured snout (Roberts et al., 2000; Russell et al., 2006; Bennett et al., 2000; Evans et al., 2012) and in Quaternary eskers in Finland (Mäkinen and Palmu, 2008). The modern analogues formed during glacial outburst floods as discharge increased rapidly, beyond preexisting conduit capacity, and are often associated with hydrofracturing of adjacent crevasses. This is discussed further below.

4.2.3. Eskers and kames

Eskers and associated kames are the dominant landform in area 2 (Figs. 1, 8, 9). Previous mapping in the area southeast of the study area (Gallagher et al., 1996; GSI, 2016) has identified a major esker, the Kilcormac esker (A on Fig. 8). The esker undergoes a number of minor high angle changes in direction before changing orientation sharply from WNW-ESE to SW-NE (Figs. 8, 9). Geological Survey of Ireland mapping (2016) also identified a short tributary (A1) that feeds in to its northern flank just before a sharp (60°) reorientation. A distributary (A2) is also visible, leading eastward from the main ridge. A second esker (B on Fig. 8) runs southeastward across the centre of the study area, parallel with tributary A1, terminating close to the point where A1 joins A. Esker B also appears to have a tributary, B1, which joins the main ridge close to the terminus. Further esker segments lie immediately south of esker A.

The LiDAR DTM, combined with sedimentological evidence, provides considerable extra information on the nature and distribution of glaciofluvial sediments in the Birr area. Glaciofluvial ridges are more extensive than previously thought and converge in the south of the area covered by the DTM (Fig. 8). The eskers mapped generally are narrow and steep sided with a well-defined basal slope break, but with some variation in width and height occurs along the length of individual ridges (from <10 m high with widths of 25-70 m, to 10-18 m high with widths of 80-120 m). Exposures in the larger ridges indicate that they are underlain by horizontal- and cross-bedded boulder to sand-sized sediments indicating transport parallel to the ridge long axis (Fig. 14A). Sediments become increasingly organised downstream (southeast and eastward), with a transition from matrix-rich to dominantly bimodal sediment distributions with a corresponding transition from internally massive to well-bedded gravels. Sandy, cross- and ripple-laminated beds increase in frequency downstream, indicating a transition from high energy, episodic, sediment-laden, often hyperconcentrated flood flows to fully Newtonian flows with variable current strengths, typical of glaciofluvial environments (e.g., Brennand, 1994; Delaney, 2001a, 2002). This morphology and sedimentology is characteristic of conduit fills (Brennand, 2000; Perkins et al., 2016), and these sediments are interpreted as such. Faulting at the sides of ridges has occurred, but the sediments at the ridge core are undisturbed, indicating a probable subglacial origin for the conduits.

(FIGURE 14 HERE)

Wider, equally elevated areas (>40 m width) have a less well-defined basal slope. Along esker A, a wider area at the western (upstream) end is clearly flat-topped and has multiple adjoining short ridges extending northward (Fig. 15A). Exposures indicate this is a truncated surface, and beach deposits occur along the esker flanks (Fig. 14B). The flat top and short ridges are likely to have originated from erosion and reworking by wave action in Paleolake Riada. Wider areas at the downstream end of eskers B and B1 are underlain by horizontally bedded cobble and pebble gravels and by cross-bedded and ripple-laminated coarse to fine sands, with interbeds of ripple and drape-laminated fine sand, silt, and clays, indicating subaqueous deposition (Figs. 14C,D). These are interpreted as subaqueous outwash fan deposits.

(FIGURE 15 HERE)

Previously unidentified eskers in the area are distinguished using the LiDAR-generated DTMs. The eskers trend southeastward and eastward. Esker B1 is distinguishable as an entirely separate ridge, running parallel to esker B. Newly identified southeastward-trending eskers include esker C to the SW of esker A (previously interpreted as hummocky sandy and gravel and two short ridges (D, E) immediately west of a major NW-SE to SW-NE change in the direction of esker A (Fig. 8). Ridges D and E cross an eastward-trending glaciofluvial ridge (F) that runs subparallel to the minor eastward-aligned section of esker A to the north (Figs. 8, 15B). Ridges D and E clearly overlie F, almost orthogonally, although the path of ridge D deflects along the top of ridge F for a circa 50 m (Fig. 15B). Additional eastward trending ridges can be seen to the east of esker A (A2 on Fig. 8), in the direction of a large isolated kame. Fragments of these ridges are also traceable to the west of esker A, subparallel to shorter ridge segments within the HT3 zone in this area (H on Fig. 8).

Previously unidentified ridges northwest of esker B appear to lead toward the western upstream end of the esker and are interpreted as the infills of tributary conduits (B2 and B3 on Fig. 8).

Interpretation: Eskers in area 2 are interpreted as conduit and channel fills, with wider areas underlain by ice-marginal point discharges from conduits/channels or conduit/channel sediments reworked as pro-glacial lake shoreline deposits. Larger eskers appear to be subglacial, but smaller eskers and sinuous ridges within HT3 are less continuous and may be partly fills of en- or supra-glacial conduits and channels. This network of conduits and channels exhibit two distinctive features. First, channels and conduits have two distinct, near orthogonal alignments: southeastward and east-northeastward. The SE alignment dominates in the northern part of area 2 and is subparallel to subjacent subglacial lineations, and we interpret this as reflecting control by the ice sheet surface slope upon hydraulic potential within the ice mass. To the south, ENE-aligned ridges become more common. In places, these connect directly with southeastward-aligned esker segments, and these eastnortheast-aligned sections are interpreted as a consequence of meltwater routing along a set of eastnortheast-aligned crevasses in the ice sheet. However, one of these eastnortheast-oriented ridges, esker A, continues eastnortheast for a considerable distance outside area 2, paralleling other Midlands eskers, indicating that ice sheet surface slope was also a factor in orientation. In addition, some ENE-aligned ridges underlie southeastward-aligned ridges, suggesting that they partly predate the southeastward-aligned landforms and suggesting a major reorganisation of the subglacial drainage network, from an east-northeastward-draining to a southeastward-draining system. We consider this reorganisation to have accompanied a southeast-directed ice marginal readvance, which also partly removed eastnortheast-trending conduit fills by subglacial erosion. Readvance was followed by the emplacement of sediment along new conduits draining southeastward toward an ice margin striking approximately northeast-southwest, beyond the area of the DTM.

A second significant feature is that the eskers indicate the occurrence of many closely spaced, converging conduits, suggesting concentration of flow in a topographic low. This differs from conduit spacing that develops under normal meltwater fluxes in modern systems, where conduits are usually regularly spaced along the ice margin (Boulton et al., 2007; Storrar et al., 2014). The routing of subglacial meltwater through closely spaced subglacial and englacial conduits and into connecting crevasses as indicated by the adjacent HT3 zone is associated with subglacial outburst floods (Roberts et al., 2000; Russell et al., 2006; Bennett et al., 2000; Evans et al., 2012). This is discussed further below.

5. Discussion

The suite of low-amplitude landforms visible on high-resolution DTMs provides critical new evidence for elucidating ice sheet events during overall deglaciation of the Irish Midlands. When combined with the palaeohydrological information provided by regional glaciofluvial features, changing ice sheet dynamics, ice flow directions, and surface slope directions can be reconstructed to construct a temporal sequence of ice sheet reorganisation events. We identify evidence for five major stages in the development of the late Midlandian ice sheet during deglaciation of the central Midlands (Fig. 16).

(FIGURE 16 HERE)

Stage 1: The oldest glacial landforms identified are the MSGLs in area 1, as these underlie other landforms in the area (Fig. 16). The MSGLs are characteristic of ice streams and surging glacier beds (Clark, 1993; Stokes and Clark, 2001) and provide clear evidence of wet-bedded, accelerated ice flow involving bed deformation across the central Midlands plain trending eastward. Where full lengths are traceable in area 1, MSGLs are close to maximum elongation values measured for these landforms under modern and former ice streams (Spagnolo et al., 2014). A comparison of Irish Midland MSGL parameters to known ice stream examples indicates that the Irish Midland MSGLS are likely to have formed under conditions of low basal shear stress (<10kPa), under relatively thin ice (<650 m) at velocities in excess of 900 m a^{-1} (Jamieson et al., 2016).

The MSGL orientation towards 083°E in area 1 differs significantly from the southeastward drumlin orientation mapped across this area by the GSI (2016). Although topographic highs in the area are mapped as drumlins, we see no indication of subglacial streamlining in a SE direction. Neither is there evidence for ice flow northeastward, as proposed by Warren and others (Warren, 1992; Warren and Ashley, 1994; Pellicer et al., 2012). The evidence in area 1 points to active ice flow eastward, toward an ice limit located beyond the study area.

Stage 2: MSGLs in area 1 are overlain by a geometric network of ridges and are interpreted as crevasse-squeeze ridge (CSR) networks, formed during ice stagnation following accelerated ice flow (Fig. 16). The ridges form by squeezing of highly deformable (wet) subglacial sediments into extensional crevasses formed during accelerated ice flow (Sharp, 1985; Evans and Rea, 1999, 2003; Evans et al., 2007; Rea and Evans, 2011).

Crevasse-squeeze ridges were first observed in association with surging glaciers, where acceleration is relatively shortlived (tens of years or less) and often involve a readvance of the ice margin (e.g., Meier and Post, 1969; Kamb et al., 1985; Ingólfsson et al., 2016). More recently, CSRs have been observed in the footprint of palaeo ice streams (Ó Cofaigh et al., 2010; Evans et al., 2016), where acceleration is more prolonged, and transition to a slower velocity regime occurs over 100s of years (Catania et al., 2012; Evans et al. 2016). The expected ridge network geometry differs between the two situations. TheCSRs formed by surging tend to occur as nets extending laterally transverse to ice flow with a significant arcuate component to the overall ridge network (e.g., Evans and Rea, 1999, 2003; Evans et al., 2016). The CSRs formed under ice streams are thought to form confined corridors along the central part of the ice stream, where tensional crevasses are best developed; ice stream CSRs are also likely to have well-developed transverse ridges reflecting this (Evans et al., 2016).

The CSR network in area 1 is arranged in a relatively narrow corridor that continues to the east of the LiDAR DTM, parallel to ice flow direction, but not to the north or south. The MSGLs also extend eastward, and GSI (2016) and earlier mapping indicated further subglacial lineations and esker fragments aligned parallel to these MSGLs up to 50 km east of area 1. This geometry is consistent with formation along the trunk of a laterally confined glacier or ice stream and resembles MSGL and rectilinear networks identified as the footprint of long-periodicity surging glaciers and surging ice streams elsewhere (Andreasson et al., 2014; Evans et al., 2016; Flink et al., 2017).

Stage 3: CSR formation was followed by a period of relative ice mass stasis, down- and backwasting accompanied by widespread glaciofluvial meltwater production and the consequent formation of eskers and kames to the northwest and south of the MSGLs and CSRs (Fig. 16). These features have been documented elsewhere and consist of subglacial conduit fills terminating in, or overlain by, ice-marginal subaqueous sediments (Farrington with Synge, 1970; Warren and Ashley, 1994; Pellicer et al., 2012). The establishment of subglacial conduits under the ice sheet is consistent with a switch from a high-pressure, distributed subglacial meltwater system to a relatively low-pressure channelized system on the cessation of accelerated flow (Fig. 16; Kamb et al., 1985; Raymond, 1987) and probably happened during, or shortly after, CSR formation, with esker formation occurring towards the end of conduit life. Esker orientation indicates that these conduits drained eastward.

Subsequently, ice recession westward and downslope resulted in the ponding of water along the ice margin (Paleolake Riada), with the deposition of subaqueous outwash sediments at conduit mouths (Fig. 16). These are represented by the Ballyduff esker and Blackwood kame-delta (Fig. 1; Warren and Ashley, 1994; Pellicer et al., 2012) and indicate ice-marginal water flow was northeastward. We consider this difference in esker orientation and ice-marginal water flow direction to reflect a change from subglacial water flow driven by ice surface slope to ice-marginal flow responding to a combination of ice margin geometry and local topography. This combination of ice and water flow directions supports the interpretation of the Ballyduff esker as an interlobate moraine (Warren and Ashley, 1994) but indicates that the dispersal centre for ice feeding the southern lobe lies to the west of the Irish Midlands basin rather than the southwest.

During later westward ice-marginal recession, ENE-ward-trending conduit fills formed at the southern margin of area 2. Their alignment supports models of ice flow deflection from eastsoutheastward to eastnortheastward around the margin of the Slieve Bloom uplands (Fig. 16; Gallagher et al., 1996).

Stage 4: Formation of MSGLs in areas 2 and 3 around Birr indicates a second phase of accelerated, extensional ice flow, toward 127° (southeastward) (Fig. 16). This supports existing interpretations of NW to SE ice flow in this area (Greenwood et al., 2009a,b). This flow direction is at a high angle to preexisting eskers in the area and indicates ice marginal readvance during MSGL formation, a possibility supported by the orientation of esker ridges formed during subsequent ice sheet drainage (stage 5 below).

Stage 5: As in area 1, accelerated ice flow is followed by deceleration and ice stagnation, resulting in the formation of CSRs (Fig. 16). The CSR formation in areas 2 and 3 involved significant reworking of underlying MSGLs, so much so that in places only MSGL grooves remain. The overall geometry of the CSR network also differs: CSRs occur as a ca.2-km-wide zone extending southwestward across areas 2 and 3, interrupted only by major esker ridges occupying lower ground. This geometry is consistent with formation of CSRs by surging of an ice marginal zone rather than in a discrete corridor underlying a linear ice stream (Evans et al., 2016).

Ice stagnation following surging around areas 2 and 3 also involved a switch from a distributed drainage system to drainage through intricate subglacial conduits, represented by the interweaved formation of esker segments and subglacial meltwater channels (Fig. 16). This subglacial drainage network was more complex than that formed in area 1, with closely spaced larger conduits flanked by networks of subglacial Nye channels (in HT2) and by crevasse fills (in HT3) indicating en- and supra-glacial diversion of water from conduits laterally across the sole of, and upward into, the ice sheet. Clustering of these meltwater channels and conduits indicates high meltwater discharges passing rapidly through the system and exceeding the carrying capacity of the preexisting conduit system. The formation of these features points to the occurrence of a subglacial outburst flood type event following CSR formation. The southeastward draining conduits also cross older ENE-trending glaciofluvial ridges, indicating that they formed under a reorganised ice sheet geometry that may reflect a shift in ice surface slope, consistent with an ice-margin readvance model.

The combination of landforms in area 2 is characteristic of surging glacier landsystems described in front of modern, temperate surging glaciers (e.g., Evans and Rea, 1999, 2003). The occurrence of a readvance during surging, followed by a post-surge outburst flood is also a common feature of modern surging glaciers (e.g., Kamb et al., 1985; Bennett et al., 2000; Eisen et al., 2005;

Burke et al., 2010). This surge event in the western part of the Irish Midlands basin post-dated the accelerated flow associated with MSGL formation in the eastern part of the basin (at Area 1), indicating that at least two distinct phases of accelerated flow occurred during overall deglaciation of the region.

Implications for dynamics of the Irish Ice Sheet: Accelerated ice flow has previously been associated with formation of extensive drumlin fields in the northern half of Ireland (Fig. 1) and with formation of MSGLs in the Irish Sea basin associated with the Irish Sea ice stream (Van Langehem et al., 2009). One theory is that the drumlins formed as a result of repeated surging initiated by drawdown of ice in response to the rapid breakup of the Irish Sea Ice Stream by calving along the tidewater margin of an 'Irish Sea Ice Stream' (Eyles and McCabe, 1989; McCabe, 1996; McCabe et al., 1998). However, this model relies on an external trigger, i.e., episodic relative sea-level rise to destabilise ice margins and to generate drumlin formation. It does not account for those drumlins formed behind land-terminating ice margins of the last British Irish Ice Sheet. An alternative cause of surging in these cases may be a disequilbrium within the ice sheet. Sevestre and Benn (2015) showed that surging in modern glaciers is caused by imbalances in mass and enthalpy (defined as the internal energy of glacier system) transfers within the glacier. We suggest that surging in the Irish Midlands reflects a similar internal imbalance in the Irish Ice Sheet during deglaciation rather than oscillations driven by external triggers such as relative sea-level influencing marine-terminating ice sheet sectors.

Recent work on the morphometrics of drumlins and MSGLs suggests that they form a continuum, with elongation controlled by a combination of glacier bed sedimentary properties, ice velocity, and time (Barchyn et al., 2016, Jamieson et al., 2016). Greenwood and Clark (2008, 2010) showed that drumlin length increases in a down-ice direction across the Irish Midlands, with MSGLs occurring at the downstream end of some flow sets, and showed that this increase in length does not reflect changes in the subglacial bed. Instead, they suggested that changes in velocity are the primary control on lineation length. The evidence for a temporal variation from ice streaming to a surging ice margin presented here suggests that the length of time over which accelerated flow phases operated may also have been a factor in controlling bedform lineation length. The reduction in lineation length westward across central Ireland therefore may reflect a reduction in ice flow velocity *and* length of time of accelerated flow.

Andreasson et al. (2014) suggested a similar transition in ice dynamics during recession around Svalbard. Surging glaciers with current cycle lengths in the region of 50-500 years (Sevestre and Benn, 2015) are linked offshore and through time to the location of former ice streams. Similar subglacial bedform assemblages to those occurring in central Ireland are preserved downstream from the modern termini of these glaciers (Otteson et al., 2007; Andreasson et al., 2014). Surging behaviour has also been observed in the Kamb ice stream, Antarctica, during overall ice retreat (Engelhardt and Kamb, 2013).

An external control on ice marginal dynamics during recession may have been the presence of a large proglacial lake (Paleolake Riada), as modern and Quaternary proglacial lakes have been linked to enhanced ice flow velocity, changing flow direction and enhancing retreat rates through iceberg calving (e.g., Kirkbride and Warren, 1999; Stokes and Clark, 2003; Walder et al., 2006; Tsutaki et al., 2013). The proglacial lake had not yet formed during MSGL and CSR formation in area 1, as the ice margin extended across the watershed. However, in area 2, the depth of Paleolake Riada was around 40 m at the ice margin during ice recession, possibly sufficient to have influenced ice flow velocity and direction during a local readvance. This may help explain the shift in ice surface slope direction and associated conduit orientation following surging.

The other impact of Paleolake Riada on the glacial landform assemblages seen in the LiDAR DTMs may have been to allow their preservation in the decades after ice sheet recession as subaqueous landforms, preventing proglacial fluvial and periglacial weathering and erosion. The removal of landforms along river floodplains is noticeable in areas 1 and 2, indicating Holocene fluvial erosion. A further factor in landform preservation may be the widespread practice of pastoral farming in the Irish Midlands throughout the twentieth century, so that mechanised ploughing has not occurred over much of the area.

6. Conclusions

- Low-amplitude landforms revealed by high-resolution DTMs constructed from LiDAR provide evidence for two distinct phases of accelerated ice flow and subsequent stagnation in the central Irish Midlands. The first phase was associated with the operation of an eastward-directed ice stream extending beyond the bounds of the study area. Landforms associated with the second phase are typical of modern surging glacier landsystems
- Ice flow directions during both accelerated ice flow events are parallel to post-surge subglacial conduit orientations as indicated by nearby major eskers. These indicate that a switch occurred from a distributed basal meltwater drainage system to a more efficient channelized system at the cessation of accelerated flow. In addition, following the second surge event, a subglacial outburst flood formed tunnel channels that linked to ice-walled conduits down-ice. Discharge was sufficiently high to overwhelm the existing conduit system and expand laterally into crevasses.
- The high-resolution DTMs provide the first clear evidence for ice flow directions around the Tullamore area (area 1). This previously undetected subglacial lineation flowset indicates that ice flow in this area was eastward, during a phase of ice streaming during regional deglaciation.
- Overall, our work demonstrates the potential for airborne LiDAR surveys to enhance existing glacial geomorphological maps and to improve reconstructions of complex ice dynamics, even in areas previously mapped by remote sensing and fieldwork.

Acknowledgements

We thank the Office of Public Works, Ireland, for provision of LiDAR data for areas 1 and 3. LiDAR for area 2 was purchased at a reduced rate from Fugro-BKS Ltd. Purchase of LiDAR data covering area 2 was funded by Manchester Metropolitan University. We are grateful to an anonymous reviewer and Dr. Ian S. Evans for helpful reviews that have considerably improved this paper.

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FIGURE CAPTIONS

Fig. 1.(A) Map of Ireland showing main glacial features and location of (B) (after Synge, 1979; redrawn by P. Coxon). (B) Map of central Midlands basin showing position of LiDAR DTM, main glacial landforms and places mentioned in the text. Land below 82 MOD is shaded grey-blue, indicating areas likely to have been covered by Paleolake Riada when an ice dam was in place west of the River Shannon and south of Birr. Eskers mapped by C. Delaney; drumlins and MSGLs mapped by Geological Survey of Ireland (2016). TWO COLUMN WIDTH

Fig. 2. Published models of ice flow directions in the Irish Midlands. (A) Model of the Irish Ice Sheet by Warren (1992) showing ice flowing northeast- and southeast-ward across the area discussed in this paper, from ice domes to the north and southwest. Grey arrows added to the original diagram show ice flow directions implied by dome contours. (B) Ice flowsets constructed from subglacial lineations (drumlins and MSGLs) in the area around central Irish Midlands esker system, from Greenwood et al. (2009a, b). Multiple flow directions can be inferred from subglacial bedforms around the eskers. The area discussed in this paper is outlined. TWO COLUMN WIDTH

Fig. 3. (A) LiDAR DTM of Tullamore area, central Ireland. MSGLs and rectilinear grids of small ridges are visible. Locations of Figs. 4 and 5 are shown. X – location of exposure in MSGL discussed in text and shown in Fig. 6. Y – position of location illustrated in Fig. 7. (B) Interpretative sketch of glacial features, including MSGL and crevasse-squeeze ridge (CSR) network fragments. (C) Cross section of transverse ridge terrain at Cappancur, east of Tullamore town. TWO-COLUMN WIDTH

Fig. 4. (A) Closeup of hill-shaded DTM shown in Fig. 3A, showing the eastern part of rea 1 (location shown in Fig. 3A). The DTM shows MSGL crests and grooves and a partly preserved grid of cross-cutting ridges, interpreted as crevasse-squeeze ridges (CSRs). Arrows indicate points where an

underlying ridge has been truncated by formation of the overlying ridge. MSGLs are visible towards the bottom of the image. (B) PCA of hillshade shown in Fig. 4A. Arrows indicate points where ridges have been truncated by the overlying ridge formation. (C) Interpretative sketch of glacial features shown in Figs. 4A and 4B. ONE-COLUMN WIDTH

Fig. 5. (A) Closeup of southern part of hill-shaded DTM shown in Fig. 3A, showing part of an esker, MSGL and the CSR network south of Tullamore town. (B) Interpretation sketch of glacial features shown in Fig. 5A. ONE-COLUMN WIDTH

Fig. 6. (A) Exposure in MSGL showing highly compacted, silty diamicton. (B) Sheared clast within diamicton. The location of the exposure is marked X in Fig. 3A. ONE-COLUMN WIDTH

Fig. 7. Photomontage of exposure at Y on Fig. 2A with interpretative sketch. TWO-COLUMN WIDTH

Fig. 8. LiDAR DTM of area 2, Birr, showing glacial landforms and position of figures. MSGL, hummocky terrain, and eskers are visible. Letters refer to landforms mentioned in the text. TWO-COLUMN WIDTH

Fig. 9. Interpretation of landforms seen in area 2 DTMs and in surrounding area (mapped using air photos). TWO COLUMN WIDTH

Fig. 10. (A) LiDAR DTM of area 3, southwest of area 2, showing type 1 hummocky terrain (HT1) and glaciofluvial ridges. (B) Interpretative sketch of landforms seen in (A). ONE-COLUMN WIDTH

Fig. 11. Closeup of MSGLs in area 2. Orientation is 307-127°. An esker is visible at the bottom left of the image. ONE-COLUMN WIDTH

Fig. 12. LiDAR DTM closeups and interpretations of hummocky terrain types. (A) and (B) HT1. (C) and (D) HT2. (E) and (F) HT3. TWO-COLUMN WIDTH

Fig. 13. Rose diagrams of ridge and groove orientations in hummocky terrain (HT). (A) HT1 ridges. (B) HT1 grooves. (C) HT2 ridges. (D) HT2 grooves. (E) HT1 ridges in Area 3. (F) HT1 grooves in area 3.

Fig. 14. (A) Conduit fill sediments, esker B. (B) Beach deposits overlying conduit fill sediments in esker A. (C) Interbedded horizontal and cross-bedded pebble and cobble gravels, horizontally bedded and ripple laminated coarse to fine sands and ripple- and drape-laminated fine sands and silts, subaqueous outwash fan, esker A. (D) Climbing-ripple cross-laminated and drape-laminated fine sands and silts, esker F. TWO-COLUMN WIDTH

Fig. 15. (A) Short ridges extending of flat-topped ridge, interpreted as possible beach ridges reworked from esker sediments. (B) Overlapping eskers. Eskers X, Y, Z override esker A, formed at an earlier point in time. ONE COLUMN WIDTH

Fig. 16. Model showing successive stages in the deglaciation of the central Midlands. Individual stages are explained in the text.

Figure 1:

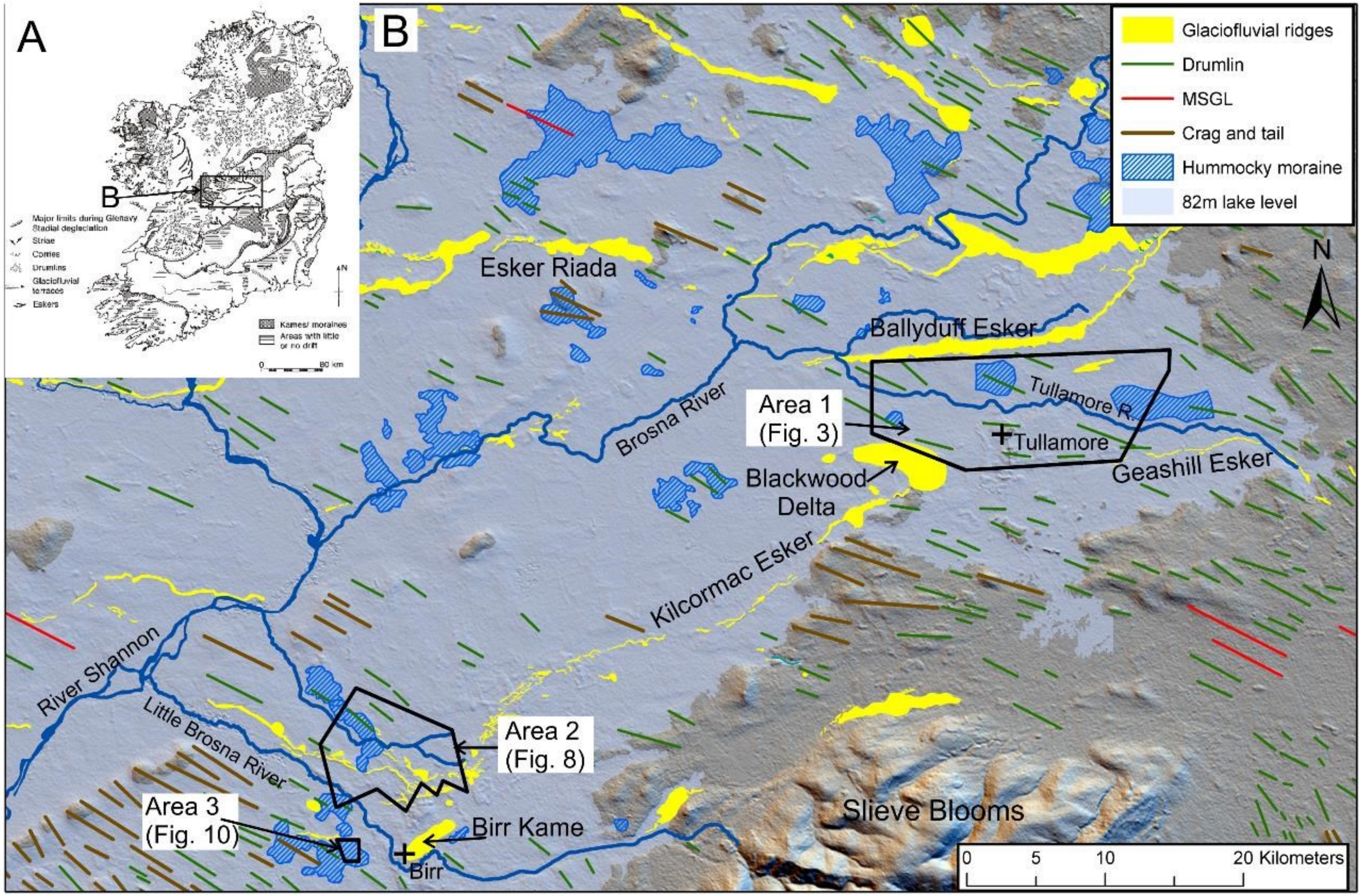

A
B
Major limits during Glenavy Stadial deglaciation
Striae
Corries
Drumlins
Glaciofluvial terraces
Eskers
N
Kames/ moraines
Areas with little or no drift
0
80 km
B
Glaciofluvial ridges
Drumlin
MSGL
Crag and tail
Hummocky moraine
82m lake level
N
Esker Riada
Ballyduff Esker
Brosna River
Tullamore R.
Area 1
(Fig. 3)
Tullamore
Blackwood
Delta
Geashill Esker
Kilcormac Esker
River Shannon
Little Brosna River
Area 2
(Fig. 8)
Area 3
(Fig. 10)
Birr Kame
Birr
Slieve Blooms
0
5
10
20 Kilometers

Figure 2:

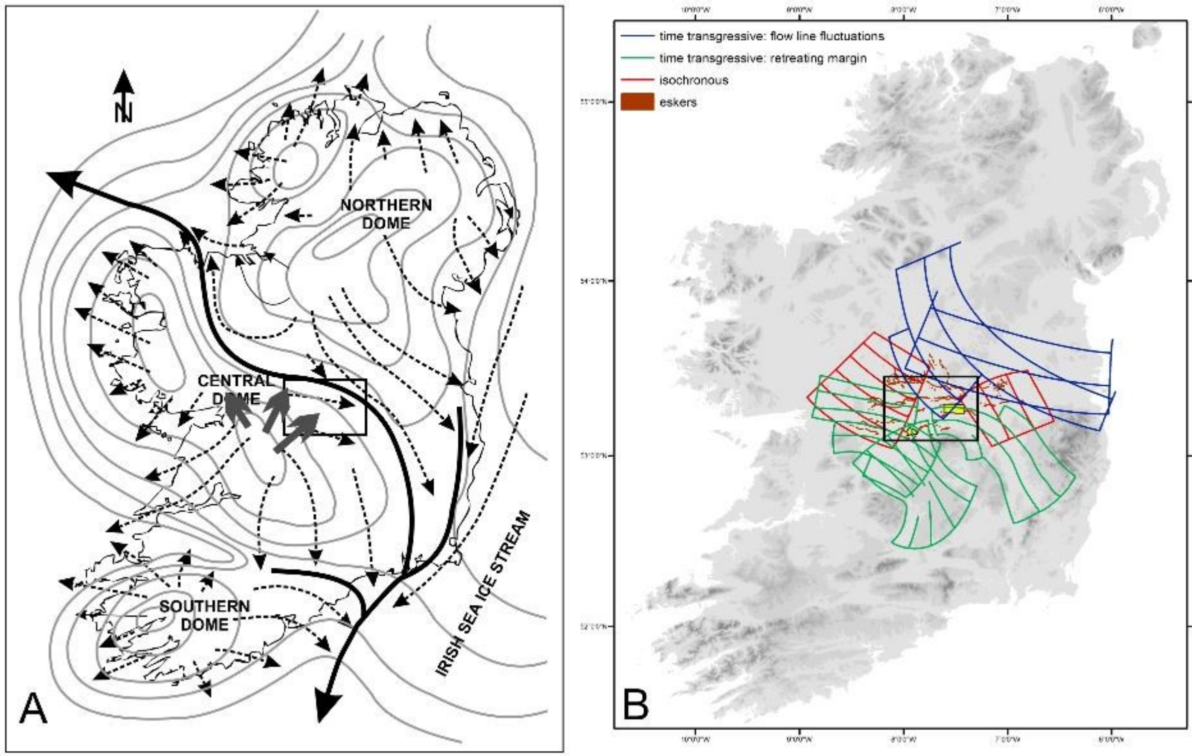

N
NORTHERN DOME
CENTRAL DOME
SOUTHERN DOME
IRISH SEA ICE STREAM
A
time transgressive: flow line fluctuations
time transgressive: retreating margin
isochronous
eskers
B

Figure 3:

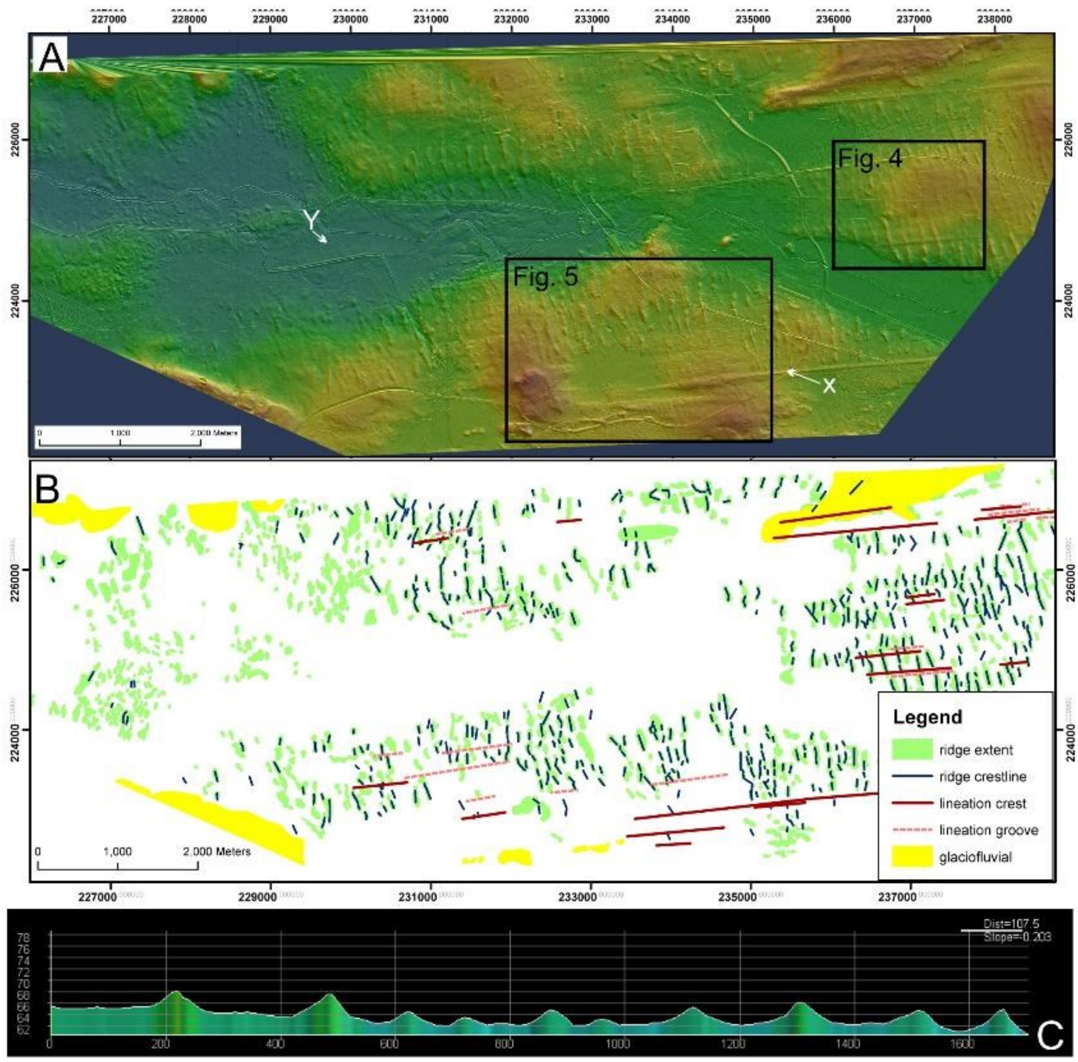
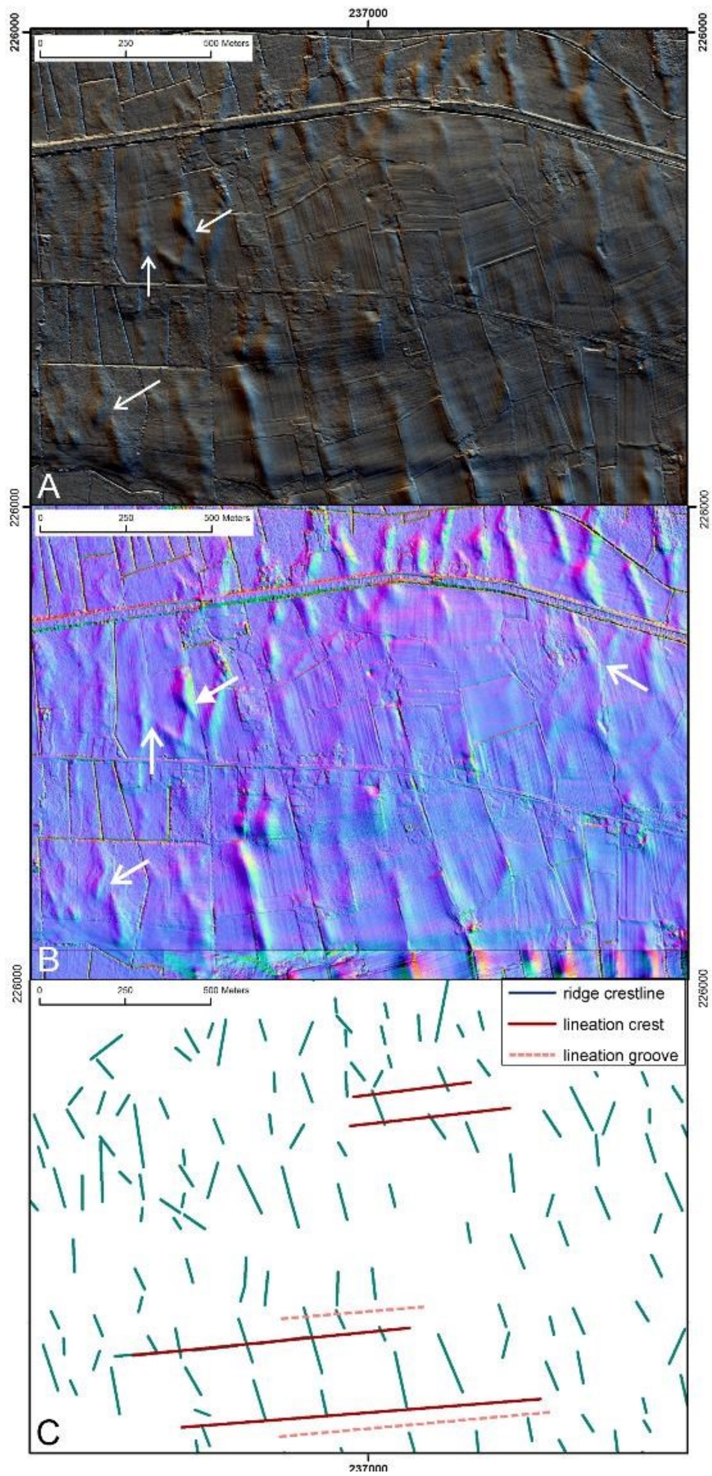

A
Fig. 4
Fig. 5
Y
X
0 1,000 2,000 Meters
B
Legend
ridge extent
ridge crestline
lineation crest
lineation groove
glaciofluvial
0 1,000 2,000 Meters
227000
229000
231000
233000
235000
237000
224000
226000
C
Dist=107.5
Slope=-0.203

Figure 4:



237000
226000
0
250
500 Meters
A
B
C
ridge crestline
lineation crest
lineation groove

Figure 5:

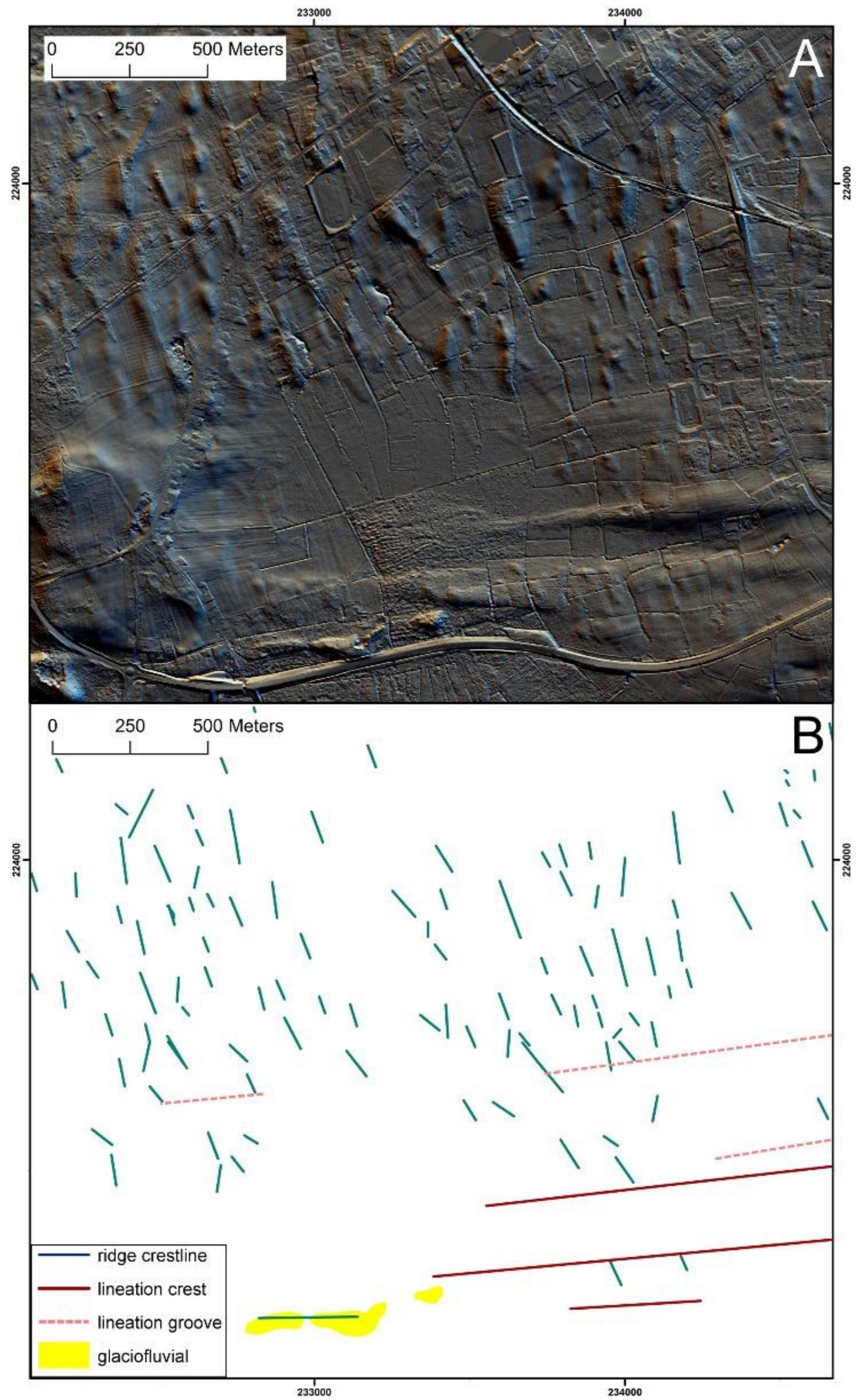
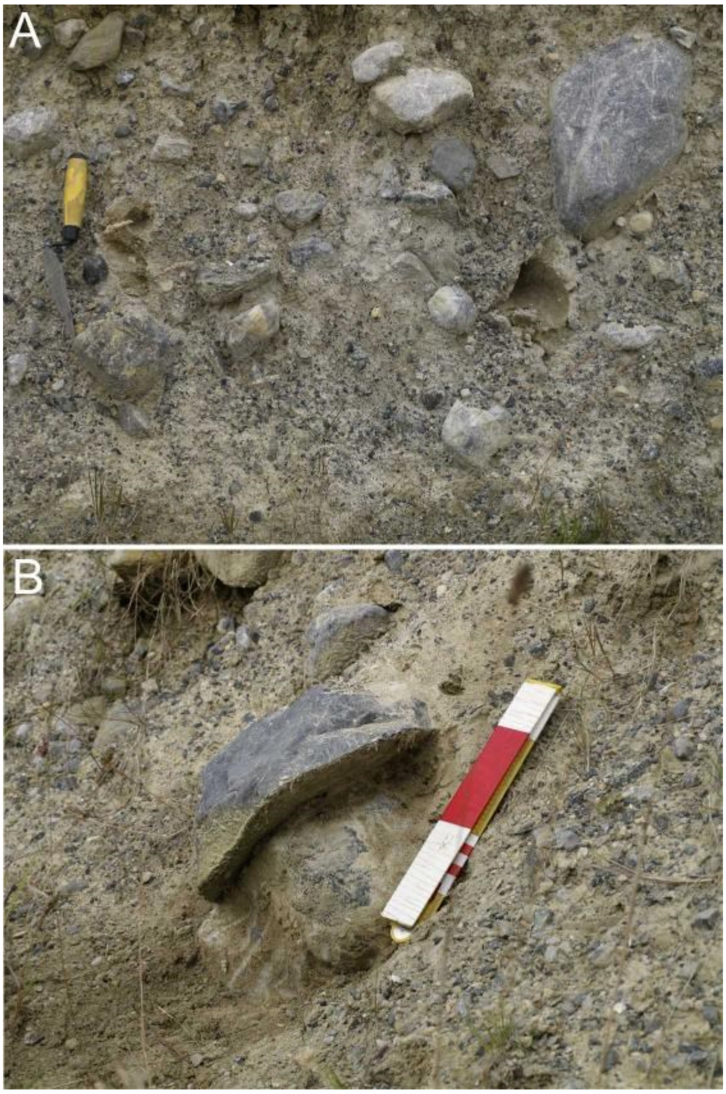
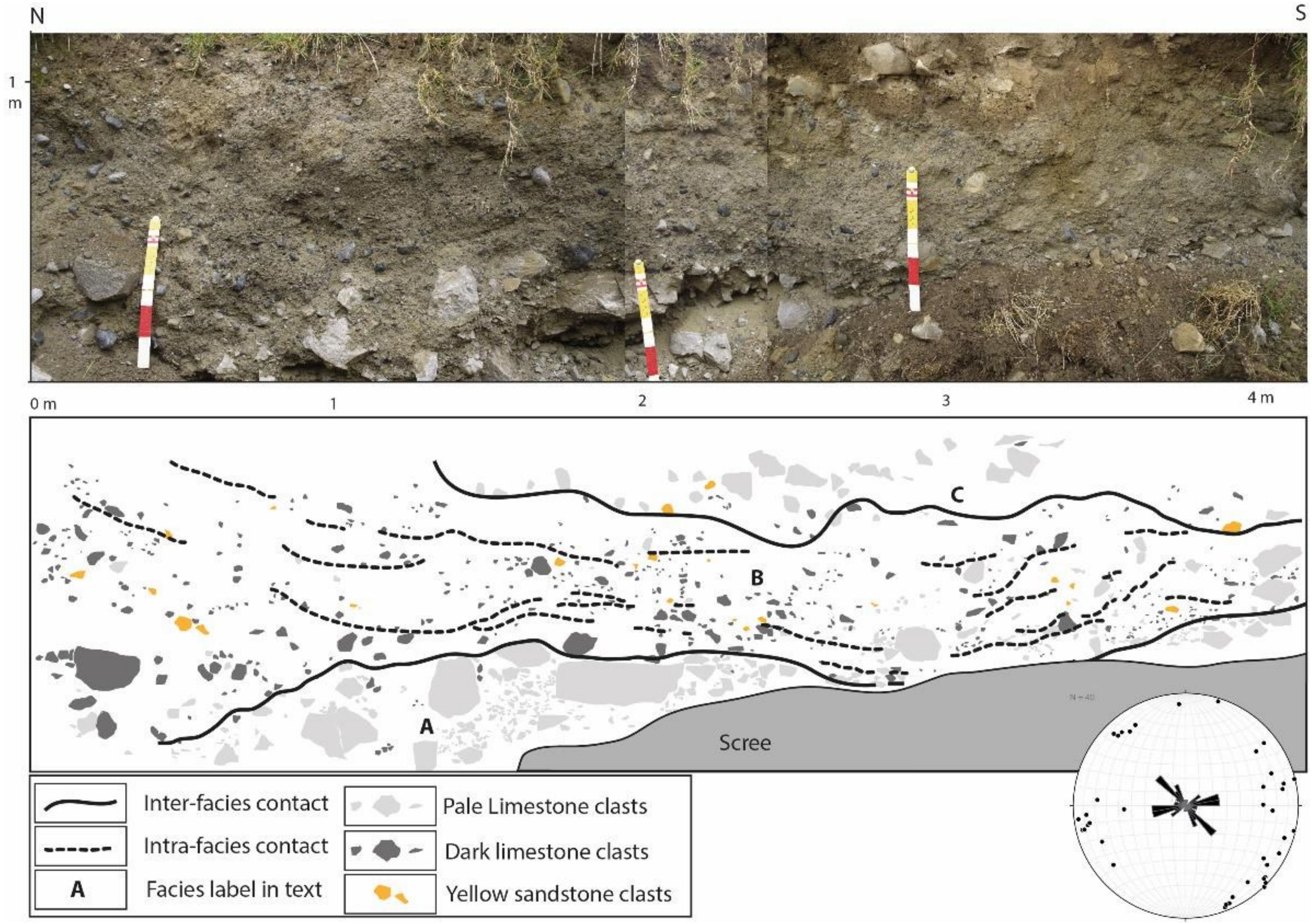

233000
234000
224000
0
250
500 Meters
A
B
ridge crestline
lineation crest
lineation groove
glaciofluvial

Figure 6:

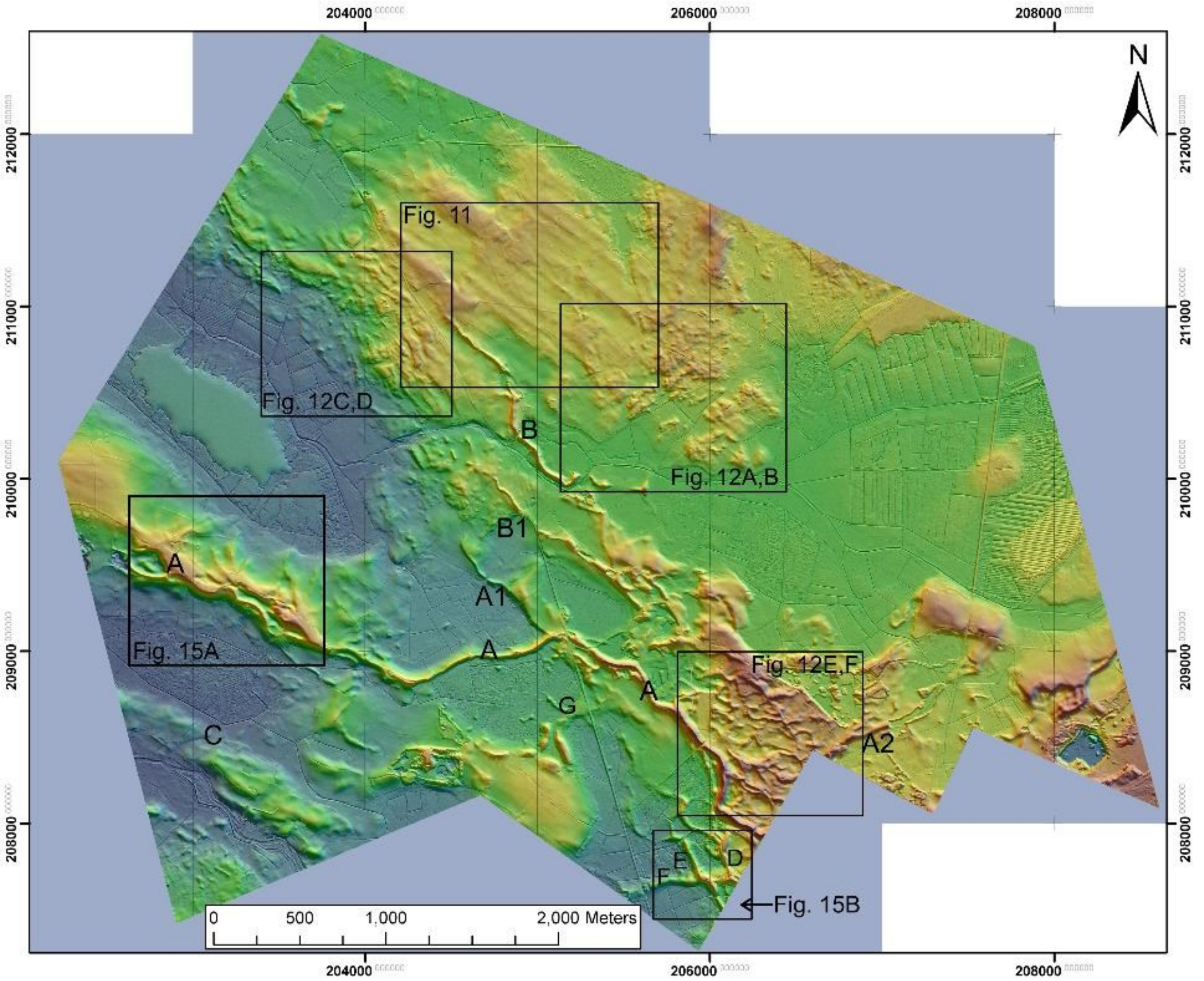


1037 Figure 7:



1038

1039 Figure 8:



1040

Figure 9:

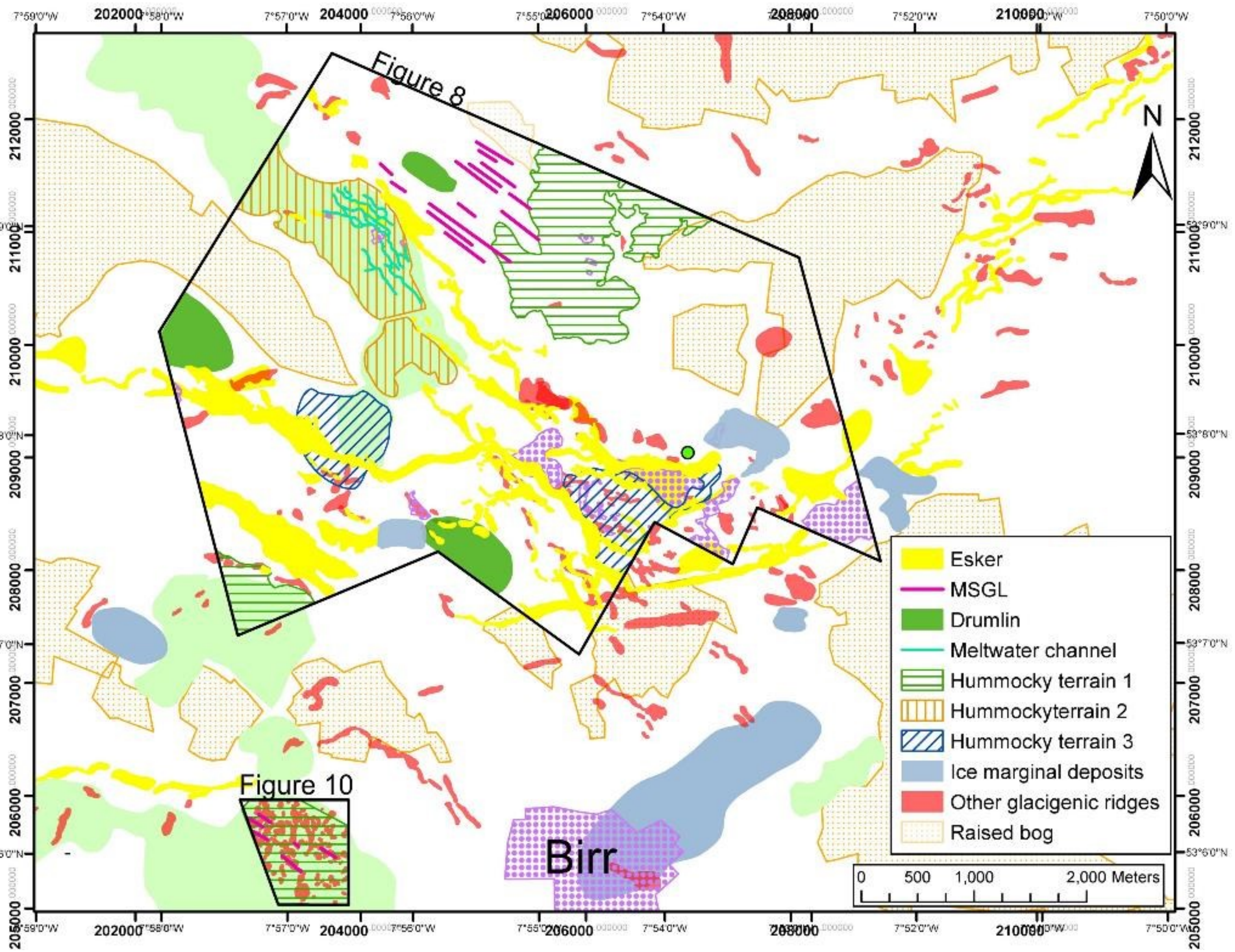
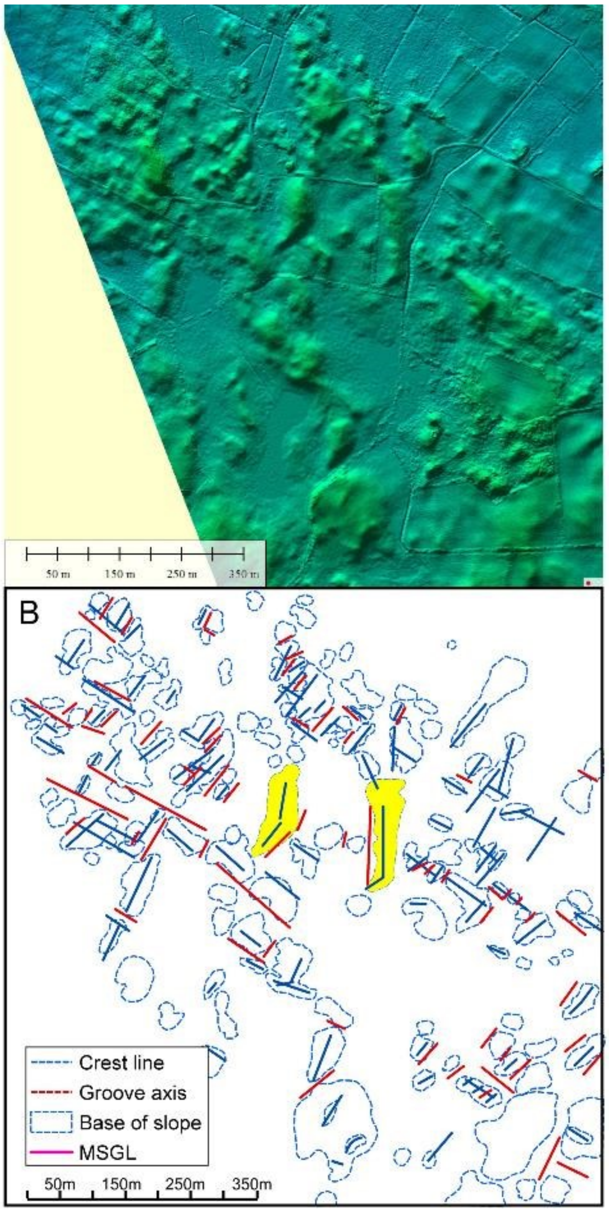
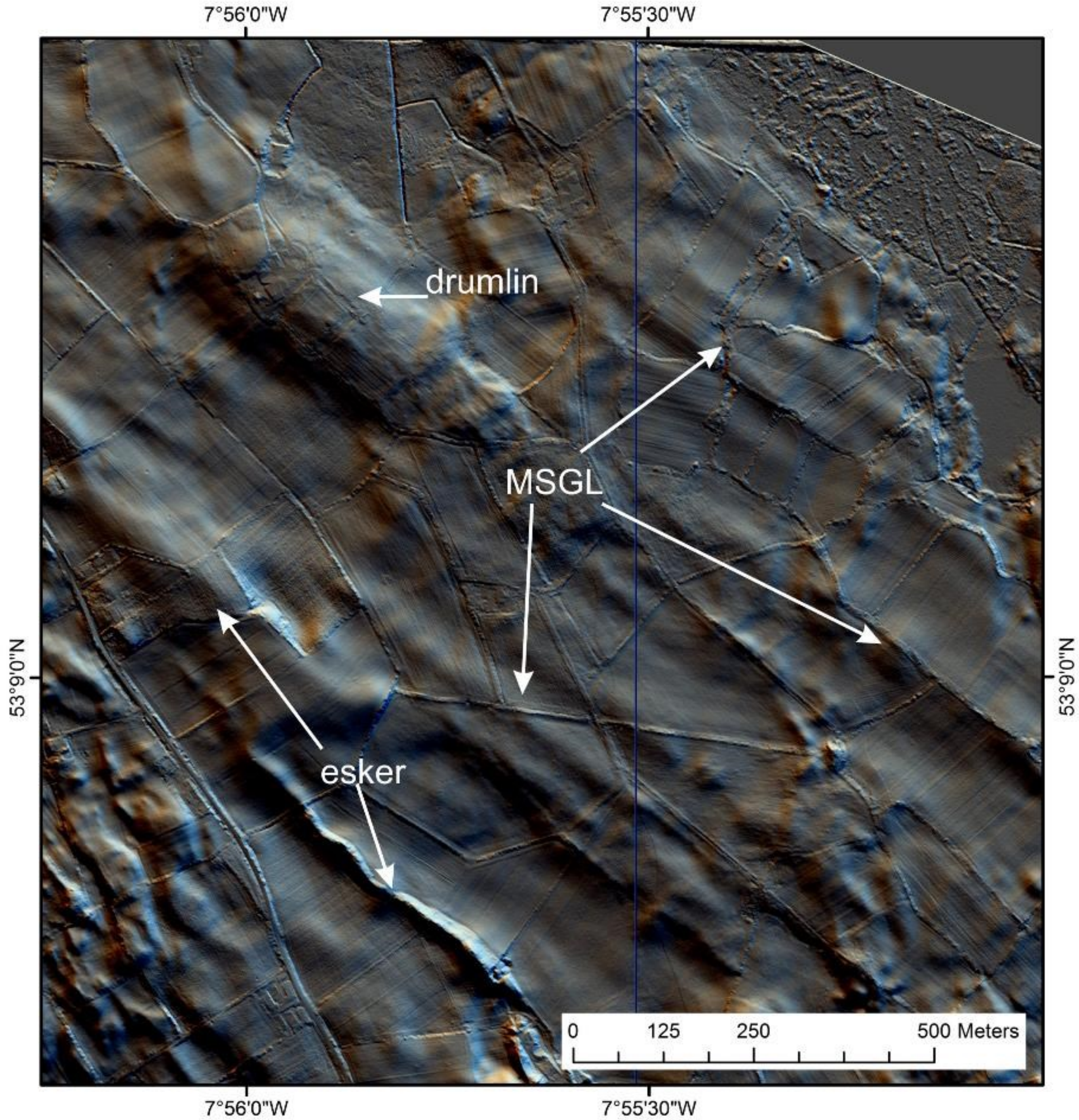


Figure 8
Figure 10
Birr
N
Esker
MSGL
Drumlin
Meltwater channel
Hummocky terrain 1
Hummockyterrain 2
Hummocky terrain 3
Ice marginal deposits
Other glacigenic ridges
Raised bog
0 500 1,000 2,000 Meters
202000
204000
206000
208000
210000
212000
211000
210000
209000
208000
207000
206000
205000
7°59'0"W
7°58'0"W
7°57'0"W
7°56'0"W
7°55'0"W
7°54'0"W
7°52'0"W
7°50'0"W
53°9'0"N
53°8'0"N
53°7'0"N
53°6'0"N

Figure 10:


50 m 150 m 250 m 350 m
B
Crest line
Groove axis
Base of slope
MSGL
50m 150m 250m 350m

1045 Figure 11:



1046

1047 Figure 12:

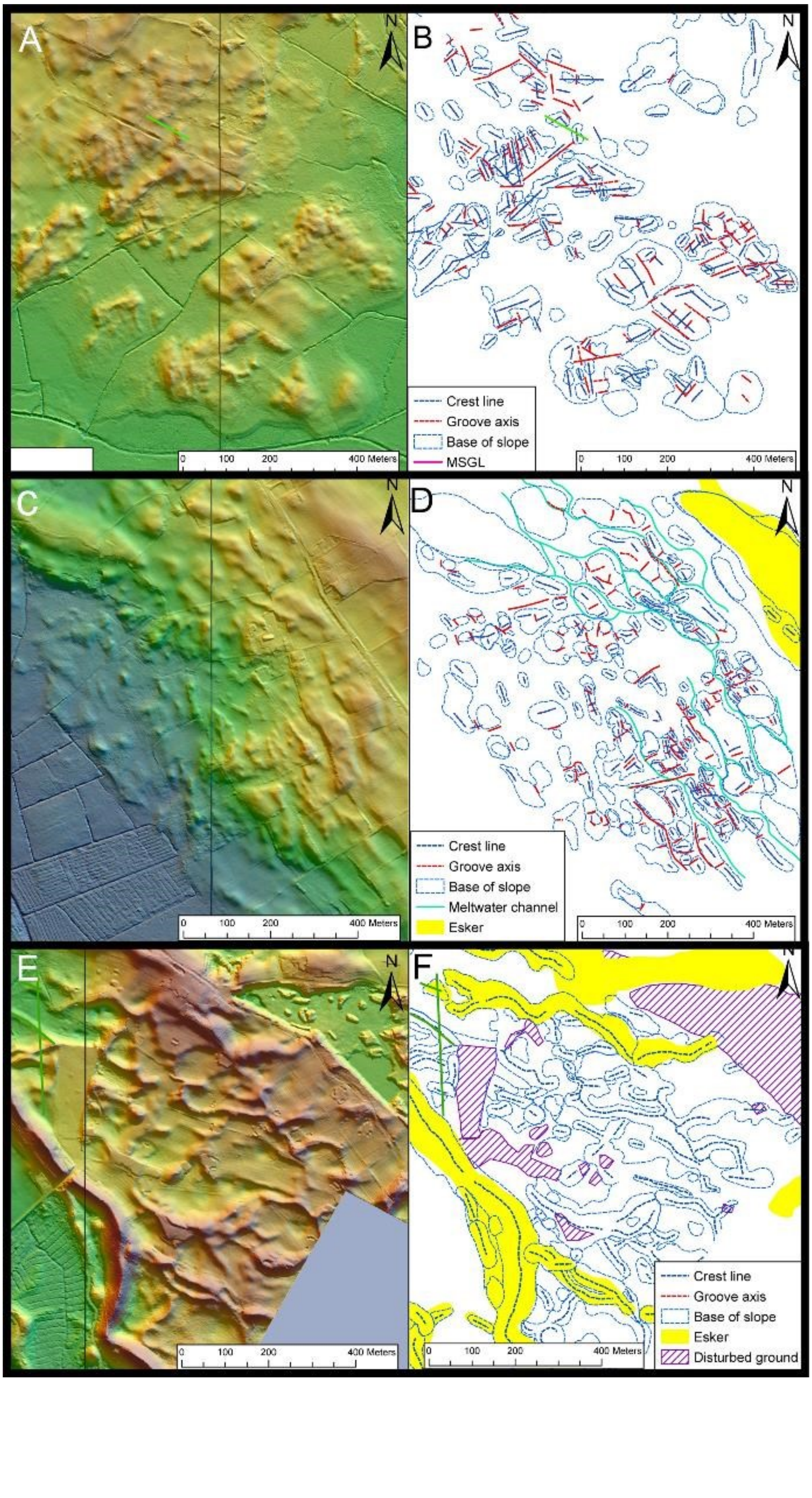
1048 

Figure 13:

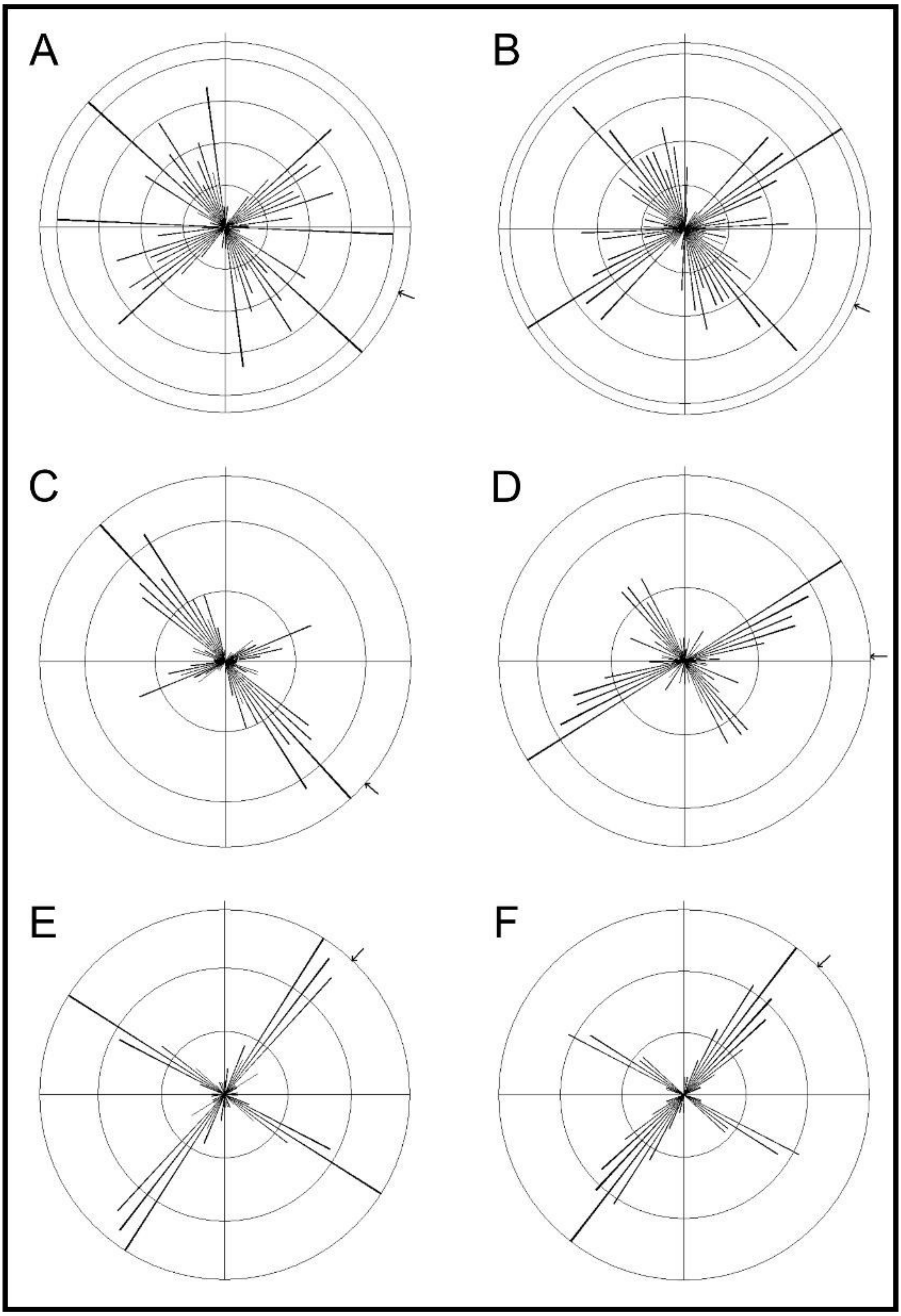
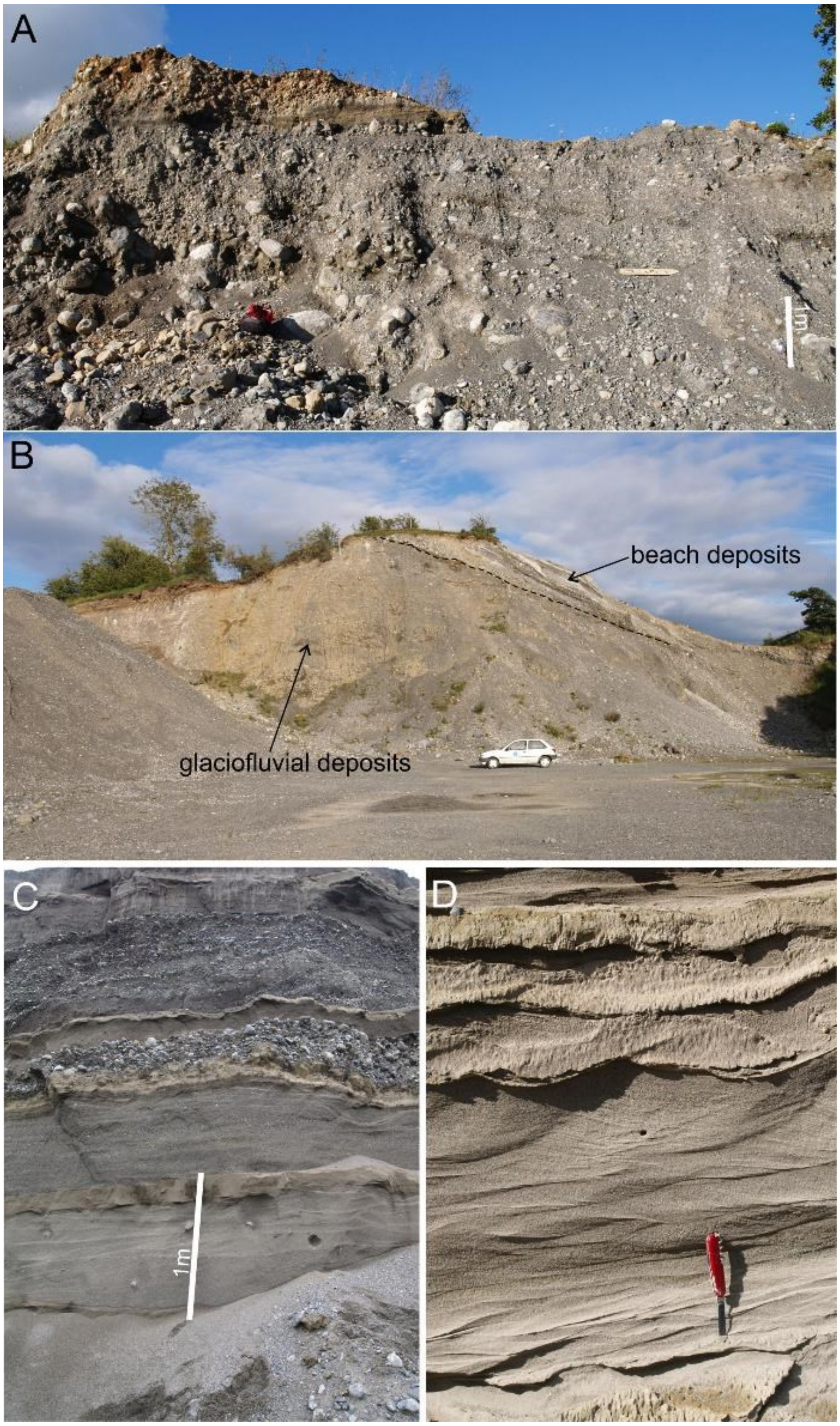
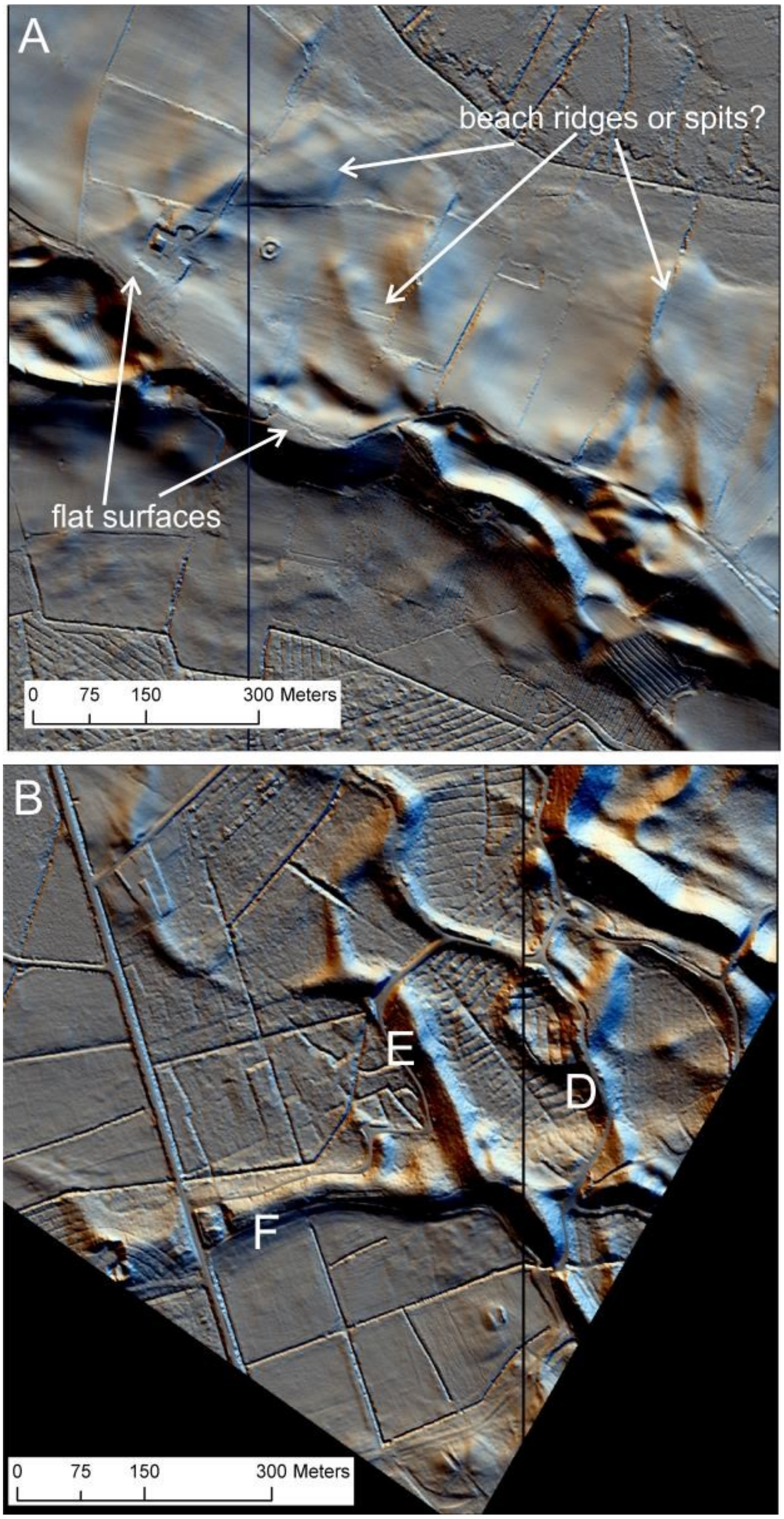


Figure 14:



1053 Figure 15:



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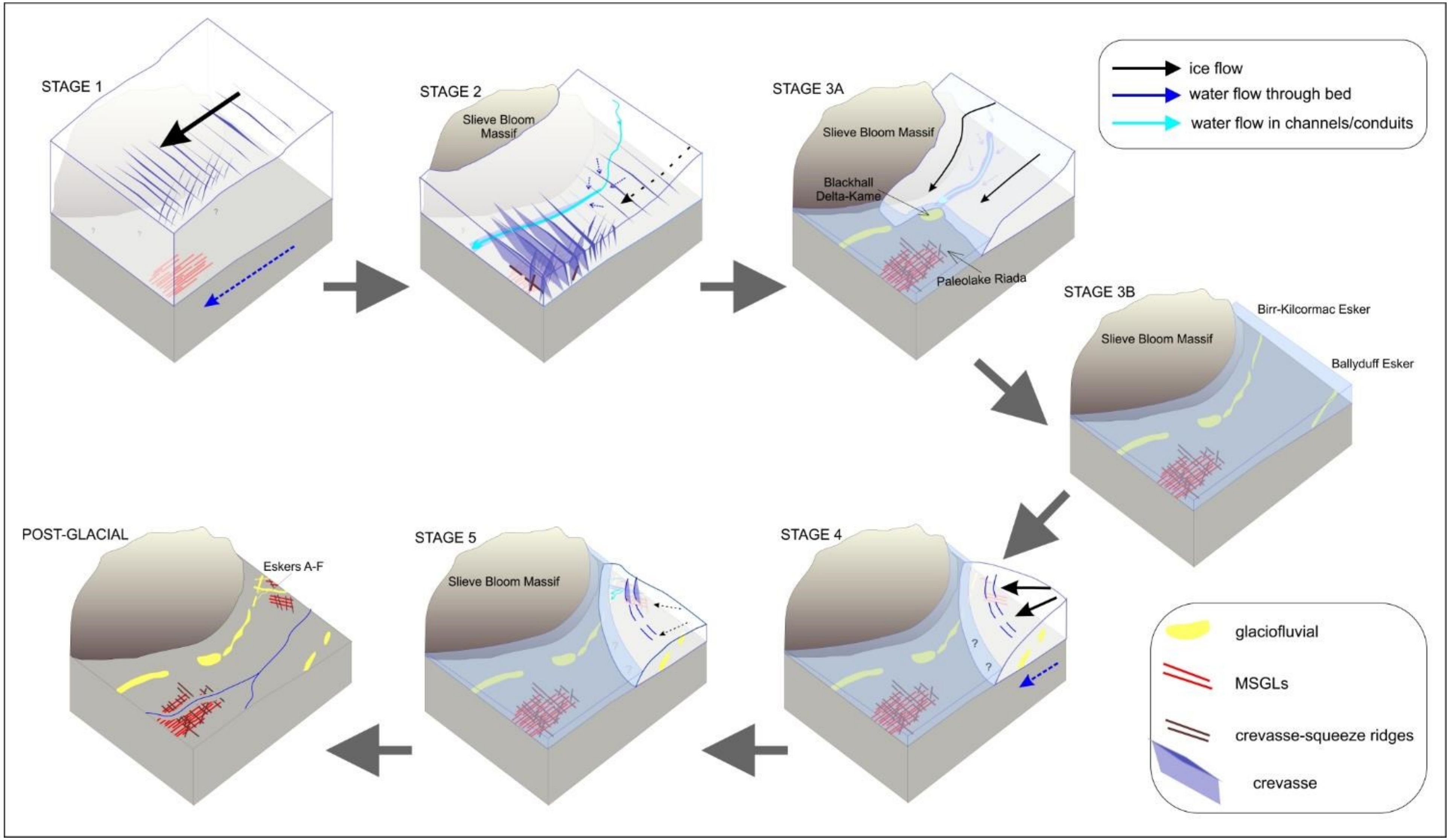
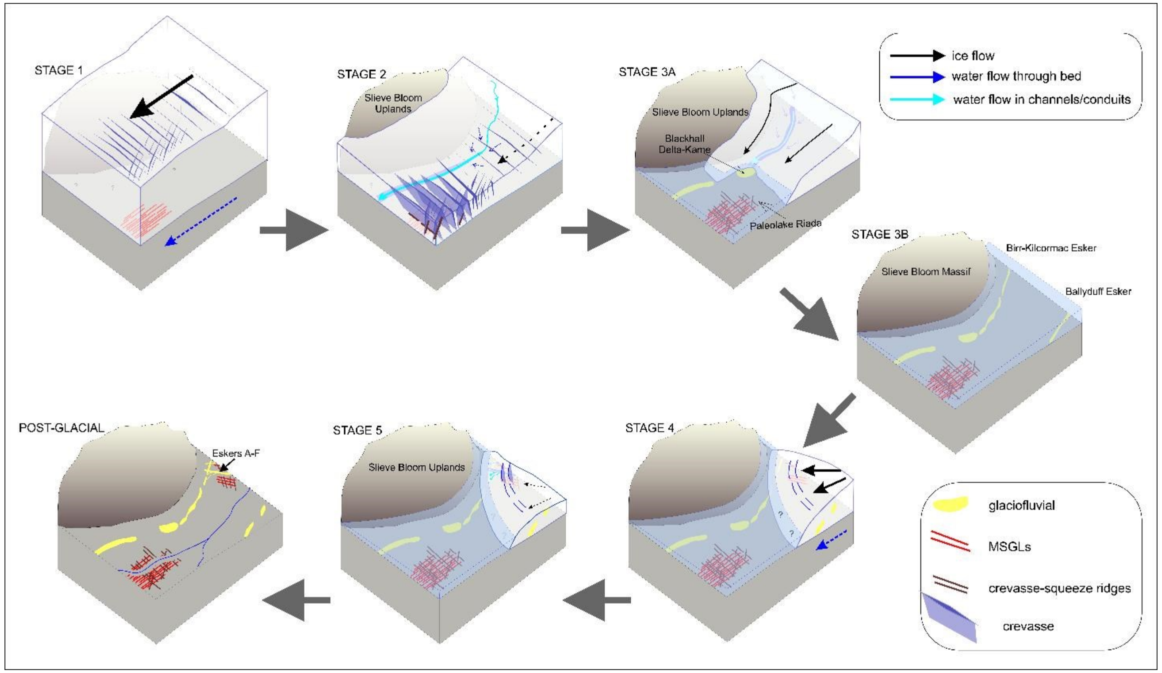
1055 Figure 16:



1056

1057


STAGE 1
STAGE 2
Slieve Bloom Uplands
STAGE 3A
Slieve Bloom Uplands
Blackhall Delta-Kame
Paleolake Riada
STAGE 3B
Slieve Bloom Massif
Birr-Kilcormac Esker
Ballyduff Esker
STAGE 4
STAGE 5
Slieve Bloom Uplands
POST-GLACIAL
Eskers A-F
ice flow
water flow through bed
water flow in channels/conduits
glaciofluvial
MSGLs
crevasse-squeeze ridges
crevasse