



The timing and consequences of the blockage of the Humber Gap by the last British–Irish Ice Sheet

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The eastern England terrestrial glacial sequences are critical to the spatial and temporal reconstruction of the last British–Irish Ice sheet (BIIS). Understanding glacial behaviour in the area of the Humber Gap is key as its blockage by ice created extensive proglacial lakes. This paper maps the glacial geomorphology of the Humber Gap region to establish for the first time the extent and thickness of the North Sea Lobe (NSL) of the BIIS. Findings establish the westerly maximal limit of the NSL. Ten new luminescence ages from across the region show the initial Skipsea Till advance to the maximal limits occurred regionally at *c.* 21.6 ka (Stage 1) and retreated off-shore *c.* 18 ka (Stage 2). Punctuated retreat is evident in the south of the region whilst to the immediate north retreat was initially rapid before a series of near synchronous ice advances (including the Withernsea Till advance) occurred at *c.* 16.8 ka (Stage 3). Full withdrawal of BIIS ice occurred prior to *c.* 15 ka (Stage 4). Geomorphic mapping and stratigraphy confirms the existence of a proto Lake Humber prior to Stage 1, which persisted to Stage 3 expanding eastward as the NSL ice retreated. It appears that proglacial lakes formed wherever the NSL encountered low topography and reverse gradients during both phases of both advance and retreat. These lakes may in part help explain the dynamism of parts of the NSL, as they initiated ice draw down and associated streaming/surging. The above record of ice-dammed lakes provides an analogue for now off-shore parts of the BIIS where it advanced as a number of asynchronous lowland lobes.

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Eastern England and the North Sea basin remains one of the key regions of uncertainty in understanding the advance and retreat patterns of the last British–Irish Ice Sheet (BIIS). Without this understanding, the nature of the BIIS's interaction with the Fennoscandian Ice Sheet, of drainage and/or lakes in the exposed North Sea basin during the sea-level lowstands, and of ice loading and glacio-isostatic rebound patterns all remain contested. Bateman *et al.* (2015) highlighted the critical nature of the Humber Gap region because of its location in the landscape where the Vale of York lobe and the North Sea Lobe (NSL) almost coalesced (Fig. 1). The NSL is thought to have emanated from southern Scotland and advanced rapidly southwards to form at its maximum a 400-km-long lobe reaching the present north Norfolk coastline (e.g. Moorlock *et al.* 2008; Evans & Thomson 2010; Busfield *et al.* 2015). Whilst primarily flowing south, localized radial flow caused the NSL to impinge upon the Humber gap area (Evans & Thomson 2010; Bateman *et al.* 2015). By advancing into the Humber Gap the NSL is thought to have dammed the drainage through it, leading to the development of the extensive proglacial Lake Humber (Lewis 1894; Edwards 1937; Gaunt *et al.* 1992; Bateman *et al.* 2015; Fairburn & Bateman 2016).

Despite being a critical region for understanding the advance and recession of the BIIS (including as it does the Last Glacial Maximum typesite at Dimlington; Rose 1985), the ice limits discussed above are only broadly

known and the underpinning geomorphic and geological evidence for events in the Humber Gap region have yet to be mapped systematically. As a result, the evidence for palaeoglaciological reconstructions from the Humber Gap has yet to be included on deglacial maps of the BIIS (Clark *et al.* 2004) and BIIS reconstructions (Clark *et al.* 2012; Hughes *et al.* 2014, 2016). Additionally, Bateman *et al.* (2015) encountered difficulties in reconciling the timing of the NSL blockage of the Humber Gap with the wider regional advance and retreat chronology of the BIIS.

This paper has three main aims. Firstly, to map the glacial geomorphology on either side of the Humber Gap as it pertains to NSL dynamics in the region. Secondly, to use both mapped till extents and glacial geomorphological features to establish the extent and first-order thickness of the NSL in this area. Finally, using both sediment stratigraphy and luminescence dating, to establish a chronological framework for ice advance into and recession from the Humber Gap. Our findings are then used to provide a critical regional chronological control on the NSL, and hence the eastern sector, of the BIIS.

Background

In eastern England, Cretaceous Chalk forms the north–south aligned Lincolnshire and Yorkshire Wolds, which respectively extend up to 168 m above sea level (a.s.l.;

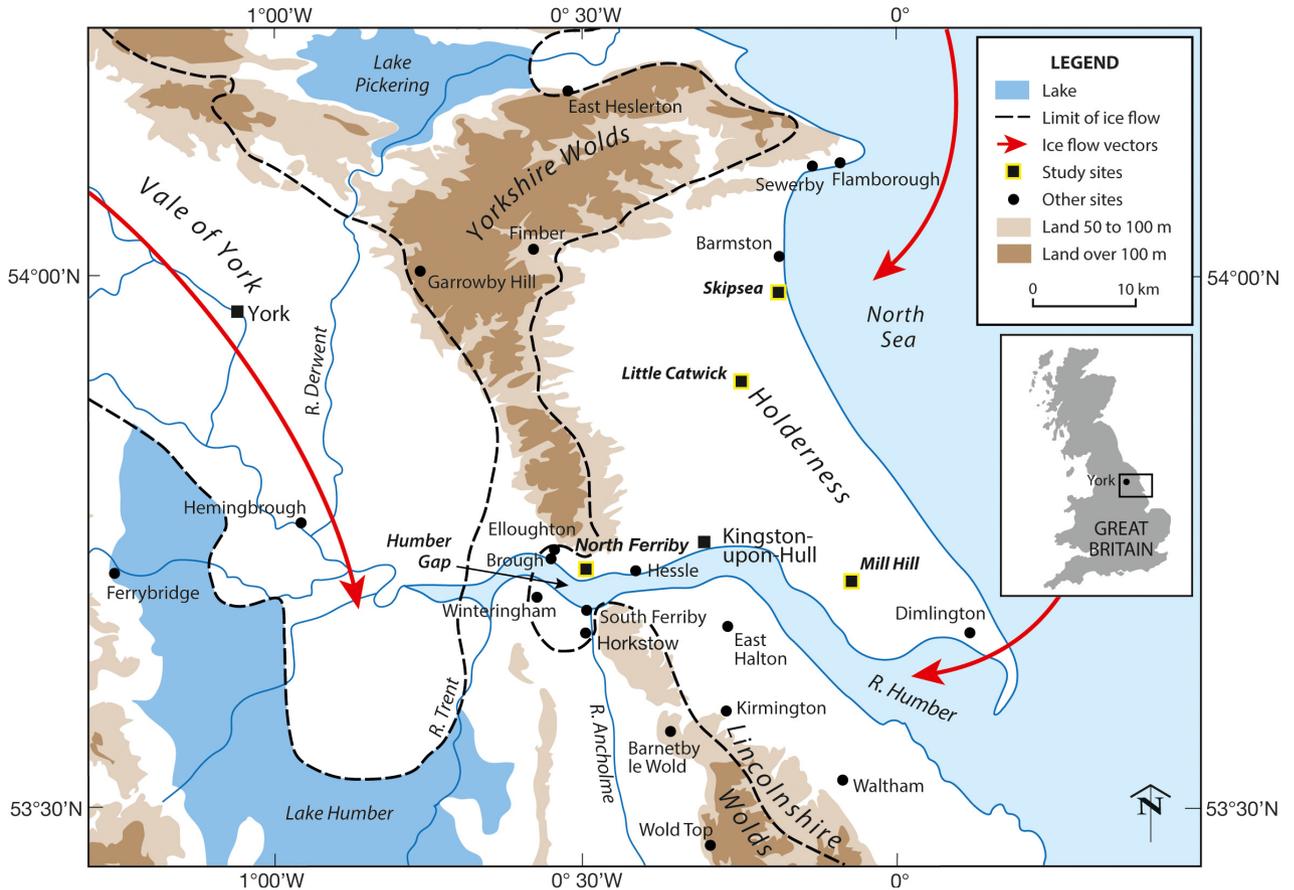


Fig. 1. Regional map of Humber Gap and East Yorkshire including study sites and other sites mentioned in the text. Ice limits based on those of Clark *et al.* (2004).

Wold Top) and 246 m a.s.l. (Garrowby Hill; Fig. 1). These hills are heavily dissected by periglacial dry valleys (e.g. Fimber; Fig. 1; Buckland 2001) and some glacial melt-water drainage valleys (e.g. Kermington; Gale & Hoare 2007) as well as the polygenetic Lincoln and Humber Gaps (Gaunt *et al.* 1992). Of these, arguably the Humber Gap is of greatest significance having been initiated by the pre-Devensian fluvial exploitation of WNW faults located in the Brough-North Ferriby area (Fig. 1; Gaunt *et al.* 1992). This eroded a 2-km-wide gap (at its narrowest) in the Chalk down to at least -30 m a.s.l. (Fig. 1; Gaunt *et al.* 1992). To the east of the Humber Gap extends a low-lying till plain punctuated with low and fragmented ridges interpreted as transverse marginal ridges (Evans *et al.* 2001; Evans & Thomson 2010). Currently, water flowing through the Humber Gap drains a fifth of the area of England ($24\,000$ km²; Rees *et al.* 2000). During the last glaciation the Humber Gap would also have been affected by increased volumes of water melted from the BIIS (including the Vale of York lobe), from associated proglacial lakes (Lake Humber; Fairburn & Bateman 2016) and overland flow due to the presence of impermeable permafrost as well as melting of ice-rich permafrost. Isostatic depression caused by the presence of ice in the

Vale of York and Lake Humber outflowing south has been cited as a possible cause of the diversion of the River Trent from the Lincoln to Humber Gap during the Late Devensian (Bridgland *et al.* 2015).

The last glacial history of the Humber Gap is principally inferred from sediment outcrops either side of the Humber Estuary, at North and South Ferriby (Fig. 1). At the South Ferriby cliffs, up to 7 m of clay matrix dominated diamicton is exposed (Stather 1896; Penny *et al.* 1972; Frederick *et al.* 2001) with a clast macrofabric showing a preferred orientation from NE to SW (Gaunt *et al.* 1992). The diamicton contains *in situ* striated boulders of carboniferous limestone, Shap granite, Millstone grit, dolerite (Whin Sill) and sandstone (Frederick *et al.* 2001). Unaltered miospores (*Rotaspora fracta* and *Tripartites vetustus*) found in coal from the diamicton, whilst not diagnostic, are thought to be sourced from the Midland Valley of Scotland, Tyne Valley and the eastern-most parts of Yorkshire (Frederick *et al.* 2001). At Eastfield farm, Winterringham, ~ 5 km to the west (Fig. 1), interbedded glacial fluvial sands, gravelly sands and thin clayey silts are overlain by a thin (20 cm) diamicton and then further laminated sands and muds (Fig. 1; Frederick *et al.* 2001). Between these two

exposures the River Ancholme flows in a north–south aligned valley between the Wolds and the Lincolnshire Edge. Within this valley, a borehole at Low Farm recorded a thin glacial diamicton at ~ -6 m a.s.l. within 5 m of laminated clays (Gaunt *et al.* 1992). Augering across the Ancholme valley around Horkstow (Fig. 1) established the distribution of diamicton and outwash gravels, which reach up to 20 m a.s.l. near Winterton and have been interpreted as the vestiges of a moraine ridge (Straw 1957, 1958). Whilst glacial diamicton at this point is recorded up to 12.5 m thick (Gaunt *et al.* 1992), the ridge has little observable surface expression, as this has largely been muted by subsequent on-lapping of fluvial sediments.

To the north of the River Humber, the British Geological Survey mapped a discrete area of glacial diamicton extending west from the Humber Gap to ESE of Brough (Fig. 1) and up to ~ 20 m a.s.l. on the Yorkshire Wolds to the north (Gaunt *et al.* 1992). On the Humber foreshore ~ 6 km WSW of North Ferriby, up to 7 m of sediment is exposed (Fig. 1; Stather 1896; Bisat 1932). At this locality (referred to elsewhere as Red Cliff), two clay diamictons were originally recognized (Lower and Upper; Stather 1896) with occasional intercalated laminated clays. The diamictons contain erratics, including local chalk, basalt, dolerite, gneiss, gritstones, sandstones and flint (Stather 1896). Laminated lacustrine clays were described as on-lapping with the diamicton to the east and overlain by a disturbed ‘intermediate bed’. This is described as comprising ‘one bed embedded in a matrix of another...in the form of cups and hollows’ (Stather 1896: p. 213). Both diamictons have been correlated with the Skipsea Till (formerly Drab till), with the upper part (Hessle till) being a weathering horizon of the lower (Madgett & Catt 1978). Clast macrofabrics from the diamicton at North Ferriby reveal NNW–SSE and WSW–ESE orientations (Gaunt *et al.* 1992). Nearby, within the boundaries of the mapped glacial diamicton at Hessle nearby Elloughton (Fig. 1), the predominant clast orientation is W–E. A borehole just to the west of the North Ferriby exposures reached ~ 10 m below sea level (Gaunt *et al.* 1992). This showed that the above described diamictons rest on ~ 2 m of silty gravel, a lower diamicton (~ 2.5 m thick), ~ 2.5 m of laminated silts, a further 2 m of sand and a gravel lag (Gaunt *et al.* 1992: fig. 41). A lower, below sea level diamicton, was also recorded in a borehole nearby (Gaunt *et al.* 1992: fig. 41). Despite the detailed work over many years at North Ferriby, South Ferriby and in the Ancholme valley no ages have been secured for any of the sedimentary units described above.

To advance to and retreat from the Humber Gap, the NSL had to cross the low-lying region to the east of the Humber Gap (Fig. 1). It is thought that the advance associated with the Skipsea Till (LFA 1 of Evans & Thomson 2010) was over pre-existing patchy Basement Till and the Dimlington Silts. This advance, based on

work at the Last Glacial stratotype at Dimlington (~ 42 km to the east), occurred after *c.* 21 ka (Bateman *et al.* 2015). Re-advances are indicated across East Yorkshire by the glacial deformation of lake sediments and subaqueous ice-contact fans into a hummock terrain of glaciotectonically folded ridges (LFA 2; Evans & Thomson 2010). Re-advance of the NSL associated with the Withernsea Till accentuated the sediment thickening and imparted gentle folding (LFA 4; Evans & Thomson 2010).

In summary, sedimentological evidence suggests that at some point after *c.* 21 ka, ice from the NSL of the BIIS flowed westwards through the Humber Gap, reaching a maximal extent just immediately west of this topographical depression. The duration of NSL occupation of this landscape, whether or not it was subject to multiple advances, and the timing of its final retreat are not well constrained and are addressed below.

Material and methods

Geomorphic mapping

Glacial landforms were identified and mapped on digital elevation models (DEMs) in consultation with 1:10 000 surficial geology maps from the British Geological Survey (Geological Map Data BGS © NERC 2016). Where available, mapping was conducted upon Environment Agency 2-m LiDAR DEMs (<http://environment.data.gov.uk/ds/survey/>) with the 5 m NextMap Great Britain™ used elsewhere. To avoid azimuth biasing, DEMs were hill-shaded from multiple directions, including 315° , 45° and directly overhead (Smith & Clark 2005). Moraine ridges/complexes were recognized as mounds, often semi-arcuate in form, which were revealed by the British Geological Survey surficial geology maps to contain Devensian diamicton (till). Large moraine ridges were digitized around their break of slope. Where this was not possible, a separate class of minor moraine ridges was mapped along their crestlines. Glacifluvial meltwater channels were also mapped along their thalweg and distinguished due to their discordance with contemporary fluvial drainage patterns (e.g. Greenwood *et al.* 2007).

Fieldwork and sampling

New sedimentological analysis and stratigraphical logging of the exposure just west of North Ferriby, where sediments have been eroded by the River Humber, were carried out in 2013. Vertical profile logs were compiled from cleared sections following the procedures set out in Evans & Benn (2004) and employing the lithofacies description and coding approach of Eyles *et al.* (1983). In order to convey the lateral changes in architecture and localized complexity in structural features, section sketches and annotated photographs were also compiled. Where

possible, palaeocurrent measurements were taken on stratified deposits. In addition to North Ferriby within the Humber Gap, three previously studied sites (Evans & Thomson 2010) across adjacent East Yorkshire (Mill Hill/Berrygate Hill, Little Catwick and Skipsea) were selected to provide an optically stimulated luminescence (OSL) chronology on the depositional events associated with the advance and retreat of the NSL. At each site sandy sediment at key points in the stratigraphy was selected for dating and collected in opaque plastic tubes to avoid exposure of the sediment to sunlight. A total of 10 samples were collected; three from North Ferriby, three from Mill Hill/Berrygate Hill, two from Little Catwick and two from Skipsea.

Luminescence dating

All luminescence measurements were carried out at the University of Sheffield luminescence laboratory. Samples were prepared by wet sieving and chemically treated following standard procedures to isolate and clean quartz for luminescence measurements (Bateman & Catt 1996; Porat *et al.* 2015). Beta dose rates are based on the concentration of U, Th and K measured using inductively coupled plasma mass and atomic emission spectroscopy at SGS laboratories, Ontario, Canada. Gamma dose rates are based on activities measured at sampled locations using an EG&G MicroNomad gamma field spectrometer. Appropriate conversion factors (Guérin *et al.* 2011) and cosmic radiation contributions (based on Prescott & Hutton 1994) were then used to calculate the dose rate. The total dose rates were calculated according to attenuation caused by moisture and grain size. Uncertainty of 5% on all water contents has been adopted. Assumed water contents, beta, gamma and cosmic dose rates and derived total dose rates to an infinite matrix are summarized in Table 1. For estimation of the palaeodose (D_e), OSL measurements are based on small multigrain aliquots (SA, containing ~20 grains each). This aliquot size has shown to be appropriate to measure samples potentially affected by incomplete bleaching in this region as it provides similar resolution to single grain measurement but better signal to noise ratios (Medialdea *et al.* 2014; Evans *et al.* 2017). All luminescence measurements were carried out using an automated Risø TL/OSL DA-15 luminescence reader (Bøtter-Jensen *et al.* 2010). Blue (470±30 nm) stimulation was used and the corresponding emitted luminescence was detected through a 7.5 mm Hoya U-340 filter. Grains were mounted as a monolayer on stainless steel discs using silicone oil and between 40 and 100 SA replicates were measured for each sample. All samples were measured using the SAR protocol (Murray & Wintle 2000, 2003) with three to four regeneration doses, a zero dose point to check recuperation of the signal, a recycling of the first regeneration dose and a second recycling including IR stimulation prior to OSL mea-

surement. The latter allowed calculation of the IR depletion ratio (i.e. the ratio between the post-IR blue OSL signal response to a specific dose and the blue OSL signal without previous IR stimulation) to test for feldspar contamination. For each site a dose recovery preheat test with temperatures ranging from 160 to 260 °C was performed. This resulted in selection of a preheat of 200 °C for 10 s for the North Ferriby, Mill Hill/Berrygate Hill and Skipsea sites and of 220 °C for 10 s for the Little Catwick site. Derived D_e estimates were accepted if the relative uncertainty on the natural test-dose response was less than 20%, the recycling and the IR depletion ratio including uncertainties were within 20% of unity and the recuperation, i.e. measurement at zero dose, was less than 5%.

Ice-surface reconstruction

In order to ascertain the elevation of the Devensian (Dimlington Stadial) ice sheet, a regular grid of points, 50 m apart, was constructed over the area covered by Devensian diamicton. The elevation value of each point was then measured from the Nextmap DEM. This was then compared to the overall hypsometry of the region in order to establish any topographical control upon the distribution of glacial sediments. The former ice-sheet surface profile was reconstructed using the estimation procedures of Ng *et al.* (2010) based on surface profiles of modern-day ice masses. The maximum extent of ice was taken either to be the most westerly moraine ridge position or the limit of Devensian diamicton. A flowline was then constructed away from this margin towards the ENE. The surface elevation along this profile was calculated by constructing a parabola using the equation:

$$H_{ICE} = H_{START} (C^* \sqrt{x}) \quad (1)$$

where H_{ICE} is the ice-surface elevation, H_{START} is the elevation at the start of the profile, C^* is a shape parameter and x is the distance along the flowline. A fairly conservative value of 1.9 was chosen for C^* , the minimum measured value of Ng *et al.* (2010) based on measurements of contemporary ice sheets and glaciers. This produces an ice elevation that is the lowest that is consistent with modern steady state ice-surface profiles, although shallower slopes have been used to reconstruct lobes of former ice sheets elsewhere (e.g. Clark 1992). This was then compared to surfaces derived using the ArcGIS-based tool of Pellitero *et al.* (2016), which uses the assumption of perfectly plastic ice (see Nye 1952; Paterson 1994). This requires an input of basal shear stress, which here was kept constant at 20 kPa (a value consistent with those derived for Antarctic ice streams situated over weak sediments; e.g. MacAyeal 1992; Joughin *et al.* 2004). The low assumed values used in both methods produce gradual ice-surface slopes to a maximum thickness of 400 m. However, short-lived

Table 1. OSL dates and associated information from North Ferriby, Berrygate Hill/Mill Hill, Little Catwick and Skipsea.

Site	Lab. code	Depth (m)	W (%)	β dose rate (Gy ka ⁻¹)	γ dose rate (Gy ka ⁻¹)	Cosmic dose rate (Gy ka ⁻¹)	Total dose rate (Gy ka ⁻¹)	Aliquots accepted (measured)	OD (%)	D _e (Gy)	Age (ka)
Ferriby (53°42'44"N, 0°31'1"W)	Shfdl13073	1.9	15	0.92±0.07	0.71±0.04	0.16±0.008	1.80±0.08	40 (96)	63	34.8±2.7	19.4±1.8
	Shfdl13071	2.4	15	0.70±0.06	0.58±0.03	0.15±0.008	1.45±0.07	48 (80)	58	32.6±1.7	22.5±1.6
	Shfdl13072	3.4	15	1.15±0.09	0.72±0.03	0.13±0.007	2.05±0.10	48 (96)	41	44.4±3.5	21.7±2.0
Berrygate Hill/Mill Hill (53°42'33"N, 0°8'33"W)	Shfdl13063	1.2	15	0.53±0.04	0.52±0.03	0.18±0.009	1.24±0.05	28 (72)	11	15.0±0.6	12.1±0.7
	Shfdl13062	2.8	15	0.50±0.04	0.49±0.03	0.15±0.007	1.15±0.05	38 (72)	14	14.4±0.7	12.5±0.8
	Shfdl13061	4.8	15	0.42±0.04	0.34±0.02	0.11±0.006	0.77±0.04	37 (70)	15	11.0±0.6	12.4±0.9
Little Catwick (53°53'3"N, 0°17'48"W)	Shfdl13065	2.0	15	0.57±0.05	0.46±0.03	0.16±0.008	1.20±0.06	57 (71)	64	28.0±1.5	23.4±1.7
	Shfdl13064	6.3	15	0.59±0.05	0.38±0.02	0.10±0.005	1.08±0.05	53 (62)	63	24.6±1.5	22.8±1.8
Skipsea (53°58'45"N, 0°11'58"W)	Shfdl13070	6.0	15	0.58±0.05	0.46±0.03	0.10±0.005	1.15±0.05	26 (43)	40	18.3±1.2	15.9±1.3
	Shfdl13069	7.0	15	0.65±0.05	0.55±0.03	0.09±0.004	1.31±0.06	51 (72)	38	21.6±0.9	16.5±1.0

events such as surging or the presence of a weaker bed (as discussed later) could have further reduced the ice-surface slope/ice thickness. The reconstructed elevations should therefore be considered as low estimates if the NSL was in steady state but high estimates if it was surging.

Results

Geomorphological mapping

The glacial geomorphology and distribution of Devensian diamicton are shown in Fig. 2, with examples of mapped moraine ridges shown in Fig. 3. A series of subtle moraine complexes was mapped on the north bank of the River Humber and to the west of the Humber gap (Fig. 2). The largest of these, the Ferriby moraine ridge (Fig. 3A), is ~5 m in amplitude at its highest point. South of the Humber and to the east of Winterton, a series of moraine ridges was found in an arcuate arrangement (Fig. 2). The largest of these ridges, shown in Fig. 3B, has a maximum relief of ~4 m. The Horkstow moraine ridge and the moraine ridge complex to the west of North Ferriby coincide with the westerly limit of mapped Devensian diamicton (Fig. 2). Diamicton encroaches on the eastern side of the Wolds, but no clear morainic features were identified on the high ground (Fig. 2). Although diamicton is most commonly found on the valley floors and surrounding lowlands, its occurrences have been mapped at elevations of up to ~98 m a.s.l. (Fig. 4). However, glacial sediment is lacking on the highest regions of the Wolds (Fig. 2). Minor moraine ridges and a series of meltwater channels were observed to the east of the Wolds, on ground lower than 40 m a.s.l. (Fig. 2). Below this elevation, a large moraine ridge complex stretching for ~25 km and trending in a NNW to SSE direction, occurs south of the Humber (Fig. 2). In places this large moraine ridge complex is superimposed by minor moraine ridges and is dissected by meltwater channels (e.g. Fig. 3C). The appearance and scale of this moraine ridge complex suggest that this is a southerly extension of the westernmost of the hummocky moraine belts in the Holderness area of East Yorkshire described by Evans & Thomson (2010).

Site stratigraphy

Results from four sites are included in this paper but of these three (Skipsea, Little Catwick and Mill Hill/Berrygate Hill) have already been described in detail so readers are directed to Evans & Thomson (2010) for more detailed descriptions.

The stratigraphical architecture of the exposure at North Ferriby is summarized in Fig. 5. Vertical profile logs 1–3 (Fig. 6) and the architecture of the undeformed stratigraphical sequence of the eastern part of the site (Fig. 7) reveal four major sedimentary units, which are

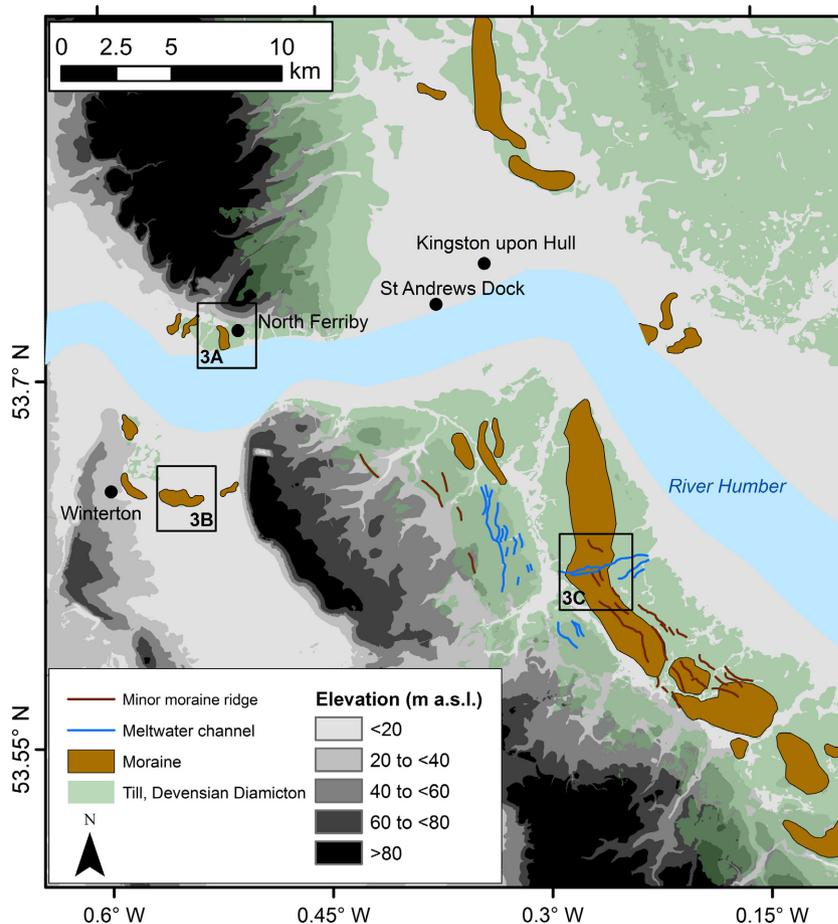


Fig. 2. Glacial geomorphology and distribution of Devensian diamicton across the Humber Gap region. Boxes show detailed images shown in Fig. 3.

here described, interpreted and then lithostratigraphically assigned, based upon their characteristics and likely genesis, to the lithofacies associations (LFAs) recognized more widely in the East Yorkshire Devensian glacial deposits (cf. Evans & Thomson 2010).

The basal lithofacies exposed at North Ferriby is a grey-coloured, clay matrix-supported diamicton that varies spatially from massive to pseudo-laminated in appearance but can contain stratified sediment intraclasts (rafts) of varying dimensions, displaying varying intensities of internal deformation (Fig. 8); sandy bedforms such as ripple-drift cross-lamination are contorted from near horizontal to vertical in individual rafts (Fig. 8F). The pseudo-lamination and textural heterogeneity created by the rafts allow the identification of strain markers within the diamicton, such as intense folding and multiple thrust faults (Fig. 6A), all of which indicate a shearing direction from the ENE. Macrofabrics previously reported by Gaunt *et al.* (1992) indicate that this diamicton contains a strong ENE–WSW clast alignment, which together with its well-documented northerly erratic assemblage, suggests that it is an equivalent to the Skipsea Till or LFA 1 of Evans & Thomson (2010). The large size

of some stratified sediment intraclasts (rafts) allows an assessment of their source materials. They predominantly comprise massive to micro-laminated silty clays, silty sands with minor fine gravel lags and displaying climbing ripple drift, and rhythmically laminated silts and clays. The sands themselves also contain numerous tiny clay intraclasts and associated vein-like dykes that have been injected postdepositionally into the host ripple forms as vein-like or tentacle networks during the deformation of the larger enclosing sand rafts (Fig. 8F). The sedimentology of the stratified rafts is similar to those previously identified and classified as LFA 2b by Evans & Thomson (2010). The complex folding and thrust stacking of LFA 1 and LFA 2b along with the laminated intraclasts in the western part of the exposure are typical of many similar exposures through the Dimlington Stadial glacial deposits in East Yorkshire and have been equated to the oscillation of the North Sea Lobe into a proglacial lake environment and the concomitant production of glacio-tectonized subaqueous fans.

The capping brown diamicton has various characteristics indicating that it is a distinct (tectonostratigraphical) unit even though it derives from the same parent

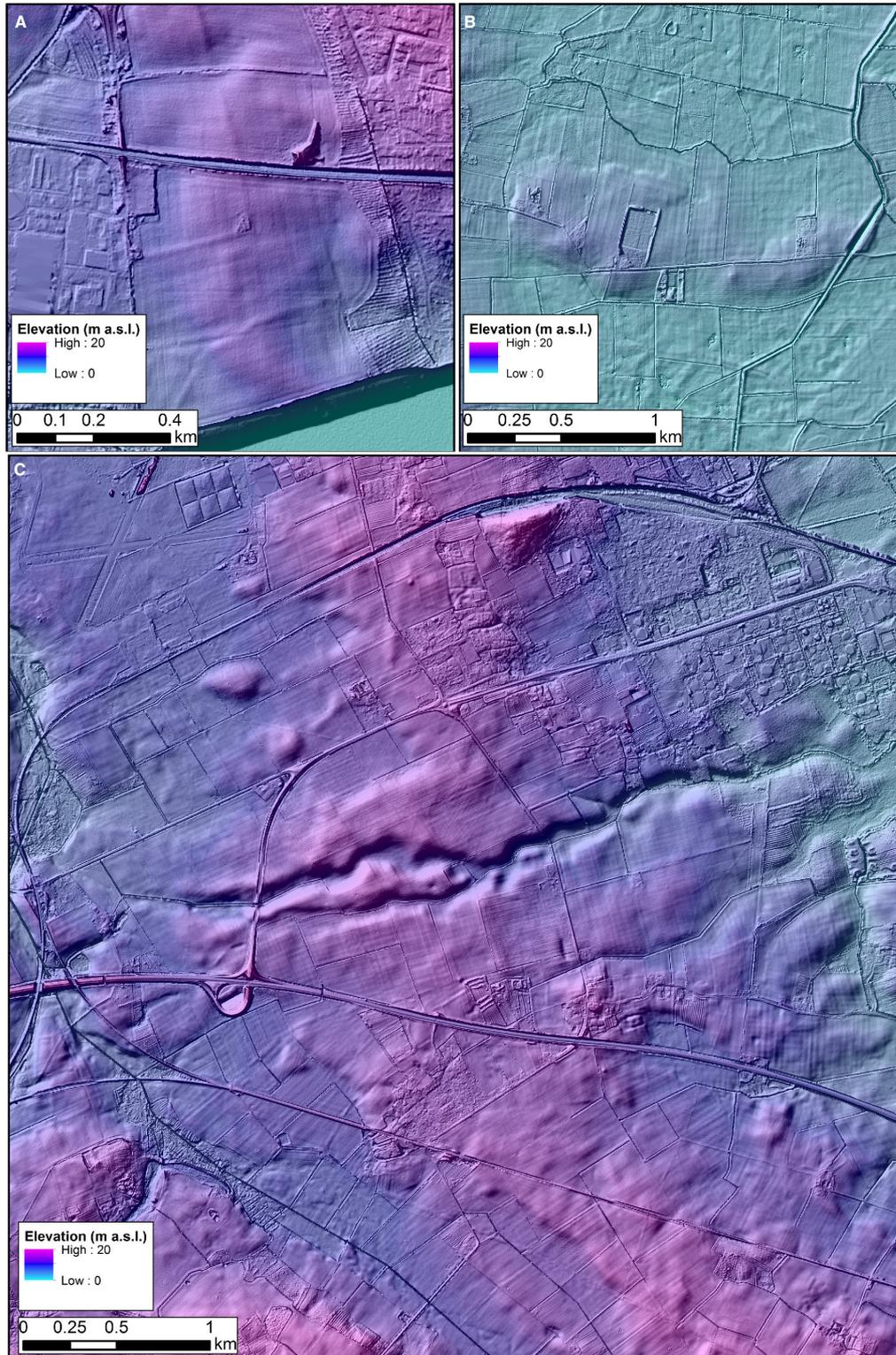


Fig. 3. Hill-shaded LiDAR images of moraine ridges/complexes, illuminated from the northwest. LiDAR data shown © Environmental Agency, 2016. Ferriby moraine ridge (A), Horkstow moraine ridge (B) and Halton to Waltham moraine complex (C). Note that elevation and spatial scale change between images.

material. Firstly, it contains numerous attenuated rafts of highly disturbed, red-coloured stratified sandy material and appears as a melange in places where the ingestion of

the rafts into the diamicton matrix is at a more advanced stage (Fig. 8D, E). This melange cross-cuts the deformation structures formed by the underlying rafts and

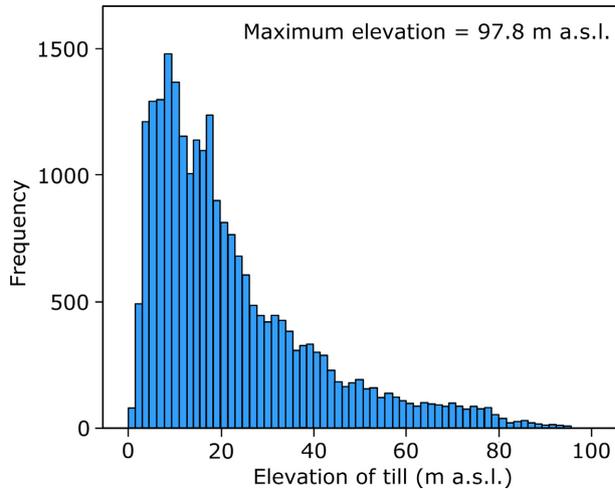


Fig. 4. Distribution of maximum altitude of mapped diamict (till) within the region showing limited material above 80 m a.s.l. Elevation measurements were taken from a regular grid of points, spaced 50 m apart, covering regions where Devensian diamict has been mapped (see Fig. 2). Maximum elevation of terrain is around 130 m a.s.l.

intervening folded and sheared grey diamict, thereby constituting an erosional contact (Figs 5, 6A, 8E). It therefore appears that the upper brown diamict has been created by shear deformation and partial homogenization of the upper surface of the deformed grey diamict and its incorporated rafts. When cleaned, the brown diamict reveals colour banding, including grey, grey-brown and red brown pseudo-lamination, and is therefore only brown in appearance following surface weathering. This brown colouring is probably related to the greater groundwater mobility created in the diamict by its higher sand content compared with the lower

clay-rich grey diamict. This is a characteristic of a glacitectorite (*sensu* Benn & Evans 1996; Evans *et al.* 2006; Evans 2017), in this location created by deformation and ingestion of LFA 2b within a clay matrix and hence strictly a further layer of glacial diamict emplaced as a distinct tectonostratigraphical unit.

The sequence displayed at Log 1 is off-lapped towards the east by undisturbed stratified deposits, as depicted in Logs 2 and 3 (Figs 5, 6B, 7). At these locations, the lower grey-coloured diamict of LFA 1 is conformably overlain by massive to laminated clays and silty clays with limestones/dropstones and with occasional laminated and massive fine sands. Lamination frequency increases up sequence and soft-sediment deformation in the form of ball and pillow structures and contorted bedding occurs near the upper contact with overlying sands and gravels. These massive to laminated fine-grained sediments have the typical characteristics of LFA 2b reported more widely in association with LFA 1 (Evans & Thomson 2010). An erosional contact separates LFA 2b from overlying gravels and sands, which are arranged in vertically stacked off-lapping sequences of clinofolds (Fig. 7). These comprise tabular sets of largely openwork sandy and gravelly foreset bedding, including coarsening-up sequences of openwork to matrix-supported gravel typical of subaqueous debrisflows. The clinofolds are conformably overlain in Log 3 by a coarsening-up sequence of poorly sorted, horizontally bedded sands and gravels. A dominant palaeocurrent is immediately apparent in the easterly to east-northeasterly dips of the clinofolds. Ice-wedge pseudomorphs cross-cut these sediments and clearly post-date their deposition. These sands and gravels have all the characteristics of glacial outwash deposits (LFA 3; Evans & Thomson 2010). The occurrence of openwork gravels and shallow debris-

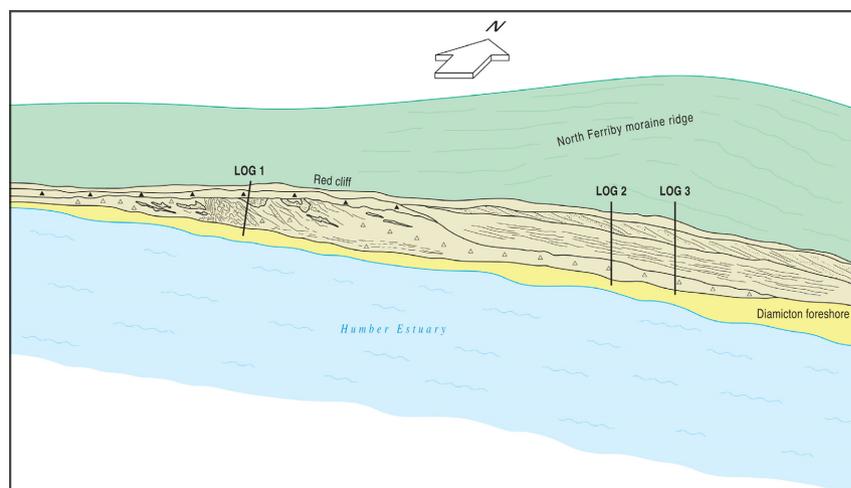


Fig. 5. Schematic sketch of the North Ferriby moraine ridge, viewed obliquely from the air from the southeast. Total length shown is 200 m. Four major units are identified, including a wedge-shaped basal grey diamict (containing some complex folds, faults and deformed stratified sediment rafts), capped by a brown diamict in the west and by fine-grained laminated deposits in the east; the latter are then unconformably overlain by gravel and sand clinofolds dipping eastwards.

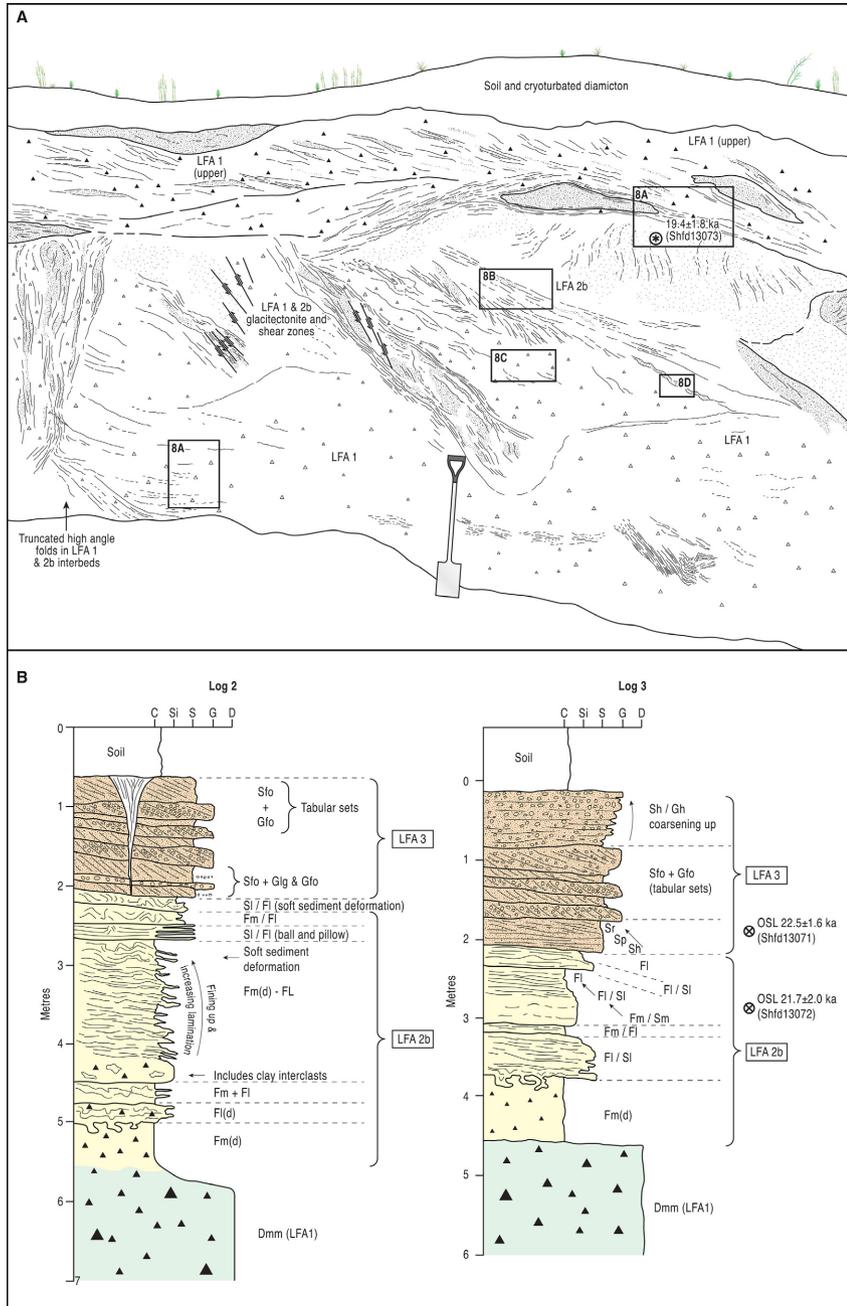


Fig. 6. A. Annotated photograph log (Log 1) of the complex deformation structures in the western half of the exposure at North Ferriby. Boxes refer to photographs in Fig. 8A. B. Vertical profile logs 2 and 3 from the undeformed stratigraphical sequence in the eastern half of the exposure at North Ferriby.

flow deposits in prograding foresets with strong unidirectional palaeocurrents are indicative of rapid sedimentation in fast-flowing, highly turbid flows, typical of an environment characterized by rapidly aggrading, transverse fluvial bars.

In summary, the basal grey diamicton (LFA 1) is classified as Skipsea Till displaying sedimentary structures indicative of subaqueous deposition followed by subglacial deformation. Overlying this on the west is a

zone of heavily folded, thrust and attenuated stratified deposits (LFA 2b) stacked between and separated from LFA 1 by an amalgamation of shear zones. This unit is derived from glaciotectionic deformation and thrust stacking of submarginal till and distal subaqueous fan and lake-floor deposits (cf. Evans & Thomson 2010). Overall this vertical sequence is typical of the exposures through LFA 1 (Skipsea Till) and its associated ice-dammed lake deposits (LFA 2; Boston *et al.* 2010; Evans

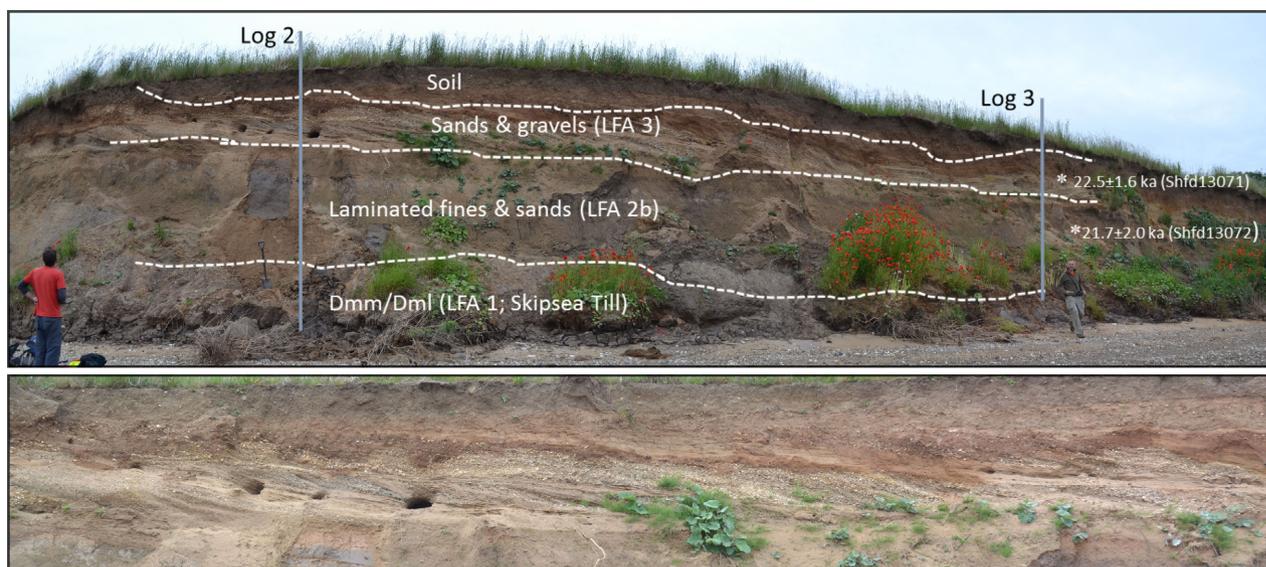


Fig. 7. Architecture of the undeformed stratigraphical sequence in the eastern half of the North Ferriby exposure, showing the details of the gravel and sand clinoforms and their unconformable (erosional) basal contact with underlying laminated sediments.

& Thomson 2010). The compressional glacioteconic structures developed within the North Ferriby exposure (Fig. 7) are indicative of a composite thrust moraine (Aber *et al.* 1989) that was overridden by westward-flowing glacier ice. The off-lapping nature of LFA 2b over the glacioteconically deformed sequence exposed in the western part of the cliff demonstrates that proglacial lake conditions replaced subglacial conditions after the construction of the North Ferriby thrust moraine. The drainage of this lake eastwards once the ice dam was no longer competent is recorded by the emplacement of LFA 3 on the eastern slopes of the moraine. It is likely that the lake drained within a channel that was directed through a low point in the moraine crest somewhere in the present location of the Humber Estuary.

The Skipsea site is dominated by matrix-supported diamicton locally up to 10.5 m thick (LFA 1; Figs 9A, 10C). It ranges from massive to crudely laminated or locally stratified and is also locally complexly folded. Within LFA 1 at Skipsea is a laterally extensive zone of sand and gravel intrabeds (LFA 2b) up to 1.5 m in thickness. Original bedding and cross-stratification within these intrabeds have been heavily distorted (Fig. 9) with complex internal fold structures. Evans & Thomson (2010) interpreted the site as evidence of an ice advance, reflected in the deposition of lower LFA 1 (Skipsea Till), followed by ice thinning and subglacial meltwater canal infilling (LFA 2), which grade laterally inland into ice-marginal subaqueous outwash fans. These fans are thought to be the lateral equivalents of the basin infilling rhythmites found at Barmston (e.g. Bateman *et al.* 2015).

Canal filling and fan progradation was terminated by an ice re-advance as recorded in the capping till/glaciteconite of upper LFA 1. Although they are both attributed to the Skipsea Till Member (Evans *et al.* 1995), differences in geochemistry between upper and lower LFA 1 shown by Boston *et al.* (2010) reveal a change in till provenance but not the equivalent to the Skipsea and Withernsea Till bipartite sequence traditionally related to this coast.

Little Catwick is an exposure through a linear, low-amplitude ridge located inland of Skipsea and represents one of a number of syn-depositionally glacioteconized depo-centres comprising alternating sequences of subaqueous ice-contact fans/deltas (LFA 2) and outwash gravels (LFA 3). The basal unit at Little Catwick is a structureless, massive and clast-poor diamicton with an east-northeasterly macrofabric (LFA 1); Figs 9B, 10B), which is assigned to the Skipsea Till Member. This is overlain by 2 m of cross-cutting planar bedded, well-sorted sands and sandy gravels arranged in cut and fill sequences (LFA 3) interpreted as braided glaci-fluvial outwash deposits laid down by ice-marginal streams flowing towards the south. This coarsens upwards abruptly into 1.1 m of sandy gravel and sandy clinoforms (LFA 2a) interpreted as deltaic foresets prograding towards the west and hence away from the former ice-sheet margin in the area. Deeper water is then recorded in an overlying 1.5-m-thick coarsening upwards sequence of rhythmically laminated to massive fine sands, silts and clays. (LFA 2b) interpreted as resulting from distal subaqueous fan and lake-floor suspension deposits. The uppermost sediments at the site comprise 0.5 m of weakly

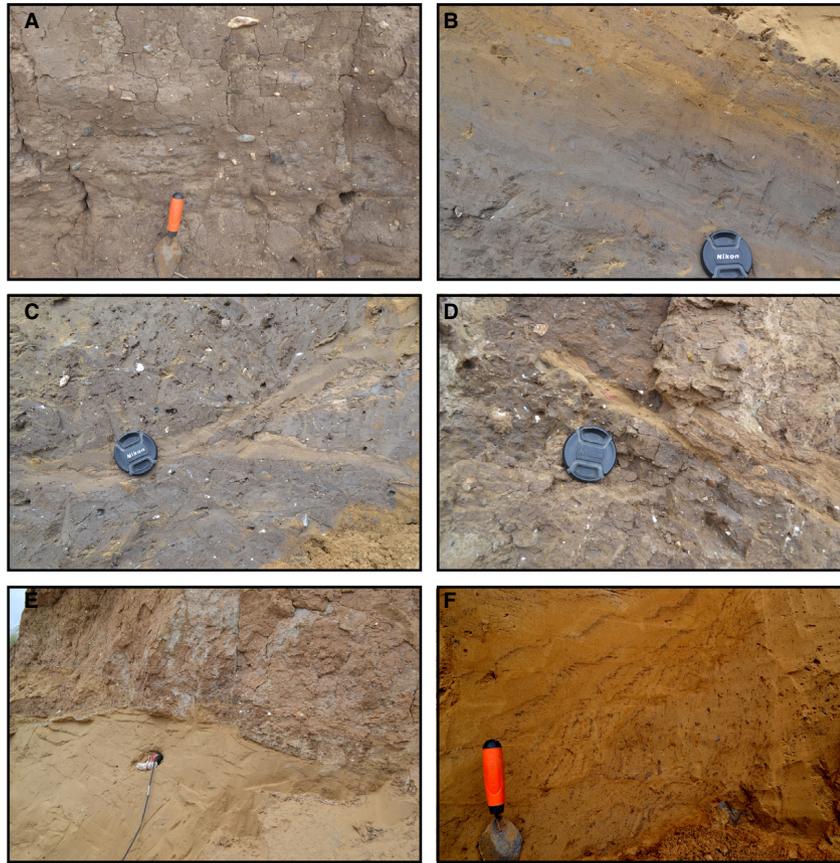


Fig. 8. Typical characteristics of LFA 1 (Skipsea Till) and associated LFA 2b rafts at Log 1 (see Fig. 4A for locations). A. Base of LFA 1 showing vertical change from laminated to massive structure. B. Amalgamation or shear zone at the base of a rippled sand raft and developed in underlying massive silty clays and diamicton. C. Highly attenuated silty sand lenses forming an anastomosing network within LFA 1. D. Attenuated sandy lens showing interdigitated margins with surrounding LFA 1. E. Erosional contact at top of rippled sand raft, showing pseudo-laminated appearance of upper LFA 1 and its highly attenuated sandy inclusions. Location of OSL sample Shfd13073 is marked by spectrometer. F. Steeply dipping beds of climbing ripple drift (small coal fragments pick out bedload lines), which have been folded and thrust from right to left. Injected clay intraclasts are scattered throughout the sands in the lower part of the image.

horizontally bedded matrix-supported gravel (LFA 3) and indicate a resumption of glaci-fluvial outwash deposition. Overall the stratigraphical sequence at Little Catwick records the alternating emplacement of subaqueous fans and deltas, proglacial outwash and finer distal lake sediments in a proglacial lake that was subject to fluctuating water levels immediately after the deglaciation recorded by the Skipsea Till.

The stratigraphy of the Mill Hill/Berrygate Hill area is presented as a composite of various exposures in the undulatory linear hummocks of the area. The stratigraphy presented here pertains to the deposits that overlie the Withernsea Till (LFA 4; Figs 9C, 10C). The Withernsea Till Member records ice re-advance over the preceding lacustrine depo-centre sediments (LFA 2b). Ice recession and continued proglacial lake sedimentation after the deposition of LFA 4 is recorded by a 2-m-thick sequence of rippled to planar bedded sands and massive granule gravels containing shell fragments and arranged in Cross-cutting shallow clinofolds dipping up

to 20° towards the north (LFA 2a). These deposits are equated to lake basin infilling by subaqueous fans that appear to have been prograding predominantly northwards at this location. The sequence is capped by horizontal, trough cross-bedded and massive, medium to fine sands, which record north-flowing palaeocurrents (LFA 3). These deposits are interpreted as postglacial fluvial infills of depressions between fan depo-centres and/or glaciotectonically derived undulations on the former proglacial lake floor.

Chronology

Samples taken for OSL determinations from both the Berrygate Hill/Mill Hill and Skipsea sites show normal D_e distributions (Table 1; Fig. S1). Over-dispersion (OD) values from Skipsea are relatively high (38 and 40%) but similar to what previous studies from the region (Heslerton in the Vale of Pickering; Evans *et al.* 2017) have shown to be well-bleached samples. Samples from both

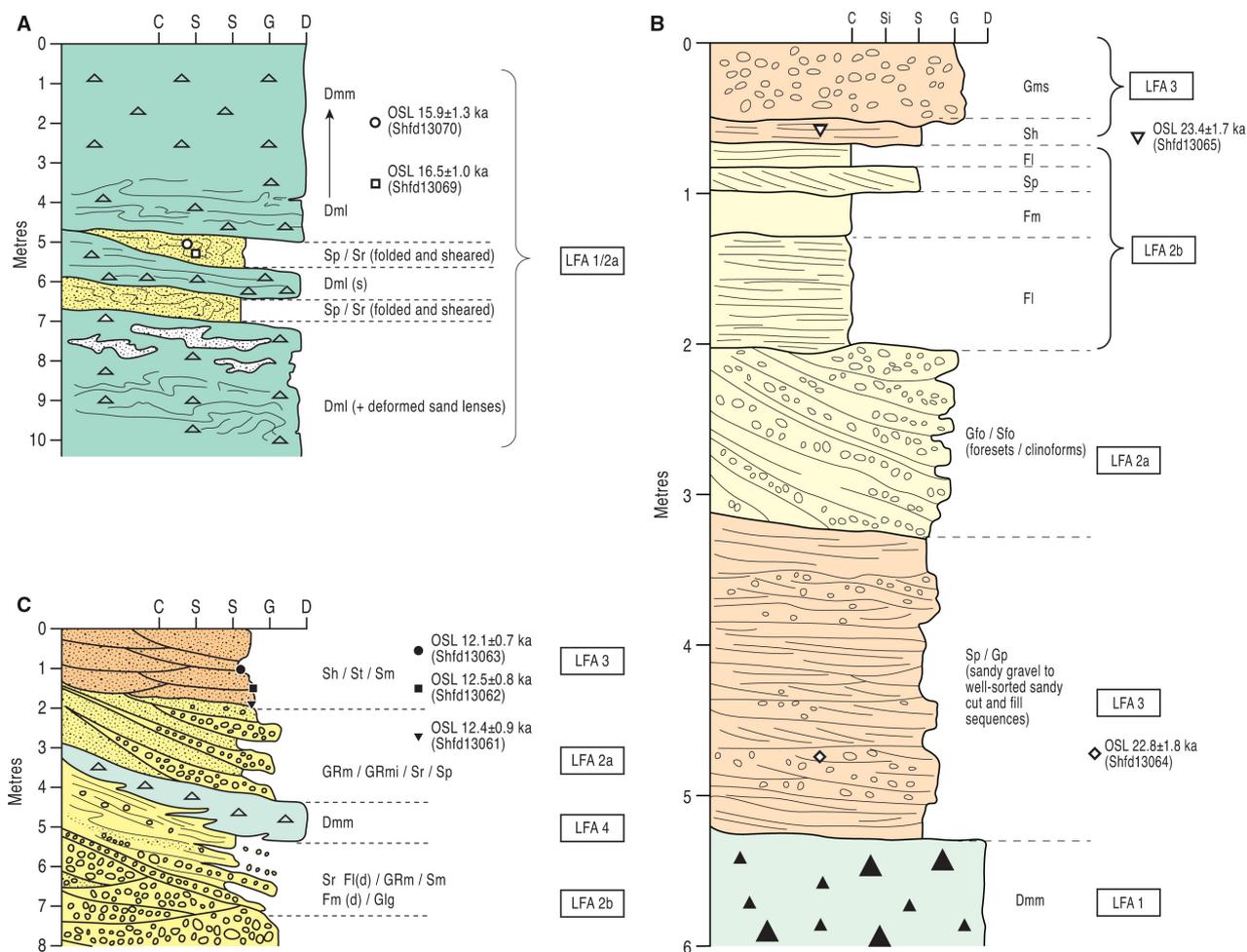


Fig. 9. Composite vertical profile logs and OSL stratigraphical positions for Skipsea site east of the Humber Gap (A), Little Catwick site east of the Humber Gap (B), the Berry Hill/Mill Hill site east of the Humber Gap (C) (adapted from Evans & Thomson 2010).

these sites were therefore assumed to be well bleached with D_e values for age calculation purposes derived using the Central Age Model (Galbraith *et al.* 1999) once outliers were excluded (those out of 1.5 times the interquartile Range; Tukey 1977). In contrast, samples from the North Ferriby and Little Catwick sites showed skewed D_e distributions and high OD (41–64%; Table 1) suggesting that these samples are affected by incomplete bleaching prior to burial. Use of weighted averages of the entire D_e distributions would therefore result in over-estimation of ages. As a result, a minimum age approach has been applied to estimate the true burial D_e using the Internal-External-Consistency Criterion (IEU, Thomsen *et al.* 2003, 2007). This is used to identify the lowest dose population which is presumed to be the population of grains most likely to have been well bleached at deposition. IEU requires the definition of starting parameters based on the characterization of OD of a well-bleached dose distribution (not available at the sites presented here) or by artificially bleaching and irradiating a subsample of the sediment

under study (prone to increased laboratory-induced uncertainty). In this case, it has been preferred to consider three well-bleached sediments from East Heslerton (Evans *et al.* 2017) as the IEU reference. The determined relation from East Heslerton is $OD (D_e) = 0.189 \times D_e + 1.5$. This was used in the calculation of IEU to derive the D_e for the samples summarized in Table 1 along with the associated ages.

For the North Ferriby site, sample Shfd13073, which gave an age of 19.4 ± 1.8 ka, was collected from the top of a rippled sand raft at the core of the thrust moraine (Figs 6A, 8E). This thereby gives an age for the accumulation of glacial lake sediments immediately prior to glacier advance. Sample Shfd13072, collected from the middle of LFA 2b, returned an age of 21.7 ± 2.0 ka. This age is associated with the accumulation of glacial lake sediments immediately after moraine ridge construction. Finally, sample Shfd13071, from the base of LFA 3 (Fig. 6B), and dating to 22.5 ± 1.6 ka denotes the timing of glacial lake drainage and the beginning of easterly draining outwash sedimentation.



Fig. 10. Exposed sedimentary sections at time of sampling showing position of OSL samples from (A) the horizontal, trough cross-bedded and massive, medium to fine sands (LFA 3) exposed at Berrygate Hill/Mill Hill; (B) sands and gravels of LFA 3 (lower sample) and LFA 2a (upper sample) exposed at Little Catwick; and (C) contorted intra-till sands (LFA 2b) in LFA 1 exposed at Skipsea.

The new OSL ages from Skipsea of 15.9 ± 1.3 ka (Shfd3070) and 16.5 ± 1.0 ka (Shfd13069) are from the same unit and within errors of each other. As this unit overlies the lower Skipsea Till (LFA 1) these ages post-date this ice advance. That the sands are contorted and are over-riden by a further unit of diamicton (upper LFA 1) indicates that a second phase of ice thickening and renewed bed deformation took place after these ages. Ages from Little Catwick of 23.4 ± 1.7 ka (Shfd13065) and 22.8 ± 1.8 ka (Shfd13064) are also within errors of each other indicating rapid sedimentation at this site and that the diamicton attributed to the Skipsea Till Member (LFA 1) pre-dates this time. The ages also record the earliest phases of ice recession and lake sediment progradation in central Holderness, over which the later Skipsea Till units (retreat tills, Evans & Thomson 2010), that were emplaced further east at Skipsea, are absent. The Berrygate Hill/Mill Hill samples gave OSL ages of 12.1 ± 0.7 ka (Shfd13063), 12.5 ± 0.8 ka (Shfd13062) and 12.4 ± 0.9 ka (Shfd13061) again internally consistent with one another and again showing rapid sedimentation at the site. This post-dates ice retreat from the area east of the Humber Gap and the demise of extended Lake Humber and gives ages for postglacial fluvial activity infilling the top of the fan depo-centres and/or glaciotectonically derived undulations on the former proglacial lake floor.

Ice-surface reconstruction

A two-stage reconstruction of the NSL ice extent in the Humber Gap area based on the mapping and the

estimated ice thicknesses is shown in Fig. 11. During its maximum extent evidence shows that the NSL predominantly remained to the east of, and did not overtop, the Wolds (Fig. 11A). Within the Humber Gap the ice attained 100–200 m in thickness (Fig. 11A, C). Toward the North Sea, ice elevation increases to a maximum of 400 m in the study area, with the outlet lobe in the Humber gap drawing in ice, and locally deflecting surface contours (Fig. 11A). The reconstruction therefore depicts a small ice lobe spreading out once through the Humber gap, at the Late Devensian Maximum (Fig. 11). This probably dammed an early phase of Lake Humber (cf. Fairburn & Bateman 2016) by stopping the eastward drainage of the lake through the Humber Gap. Ice advancing into a lake might be expected to float but the mapped moraine ridges appear to resemble terrestrial push/thrust moraines. The evidence from the Ferriby Moraine ridge supports this interpretation with its stacked sediments developed as distal facies of an ice grounding line. The maximum significant elevation reported for Lake Humber is 33 m (Fairburn & Bateman 2016) although this elevation may have been attained later than the initial ice advance into the gap and moraine ridge formation. With a lake of this depth ice would need only to be 36 m thick to resist floatation (Friend *et al.* 2016). Surface ice profile reconstructions based on simple physics and vertically constrained by the lack of diamicton found above 100 m a.s.l. in the area indicate that this outlet glacier was probably thin but thick enough not to float (Fig. 11).

Subsequent to the NSL reaching the Humber Gap, it retreated to a second marginal position forming the Halton to Waltham moraine ridge complex (Figs 1, 3C), joining up with the Westerly moraine ridge complex of Evans & Thomson (2010) (Fig. 11B). This indicates the removal of the Ferriby ice dam out of the Humber Gap, and allows water to flow over the smaller gap by Barnet le Wold (Figs 1, 11B). The latter contains glaci-fluvial sands and gravels, which would support this scenario. This second marginal position could have had the following possible consequences for the water contained in Lake Humber: (i) drainage to the north; (ii) drainage to the south; (iii) expansion of the lake to meet the new ice frontal position; (iv) drainage via a subglacial route; (v) drainage via another outlet opened by ice retreat elsewhere. The fate of Lake Humber water in scenarios i–iii depends upon the configuration of the ice sheet outside of the study area, along the east coast of England. Ice-free corridors to the north (i) or the south (ii) of the Wolds would have allowed drainage along either. As the NSL expanded from and retreated to the north an ice-free corridor to the north would appear less likely. If both northerly and southerly routes were blocked by ice, it would have facilitated an expanded Lake Humber impounded by the NSL (Fig. 11B; option iii). Alternatively, water could have drained into the ice sheet, via a subglacial conduit (iv), which would in turn have had an impact on the dynamics of the ice lobe. Finally, drainage may have continued to flow through the Lincoln Gap 50 km to the south of the Humber Gap (see Fairburn & Bateman 2016). The potential for multiple re-advances of ice in the area leaves open the possibility that more than one of these scenarios could have occurred during the glacial occupation.

Discussion

The new data on the glacial geomorphology and chronology of glacial events in the Humber Gap and environs to the east, when used in conjunction with previously published OSL ages and events from the region (Table 2; Fig. 12), provide a number of important details on the operation of the NSL of the BIIS during the last glaciation. Most significantly these details help determine the timing of the maximal westerly advance of the NSL during the last glaciation of the BIIS. They also provide clear evidence of the NSL advance on-shore and its subsequent oscillatory recession in contact with proglacial lakes, the latter persisting much longer and extending over a wider area to the east of the Wolds than previously thought. The new data also facilitate a much better spatial understanding of the pattern and chronology of ice advance and retreat within the region. Additionally, this yields a better connection between the records of off-shore and on-shore ice dynamics of the NSL of the BIIS.

Ages relating to non-glacial sediment deposited just prior to the onset of the final glacial advance in the region of south Holderness and the Humber Estuary (Table 2, Fig. 12) yield an average age of 20.5 ± 0.44 ka. Ages that directly date the Skipsea Till ice advance or require ice-damming in the Humber Gap show a high level of correspondence across the whole region and yield an average age of 21.6 ± 1.3 ka (Stage 1, Table 2, Fig. 12). This is consistent with ice-advance ages further north at Upgang, North Yorkshire (~85 km north of the Humber Gap) where ice moved on-shore after 23.3 ka (Roberts *et al.* 2013). It would appear that the Humber Gap and the whole of the region east of it was ice covered at this time. Based on the moraine ridges at North and South Ferriby, Horkstow and two indistinct moraine ridge complexes located around Brough-Elloughton 3 km to the west of North Ferriby (Fig. 2), this age also relates to the NSL attaining its maximal westerly limit. The occurrence of the lower (below sea level), and as yet still undated, diamicton at North Ferriby may be an oscillation within this glacial advance or relate to an earlier event within the Humber Gap.

Ages relating to the retreat, re-advances and deglaciation pattern appear more regionally variable (Stages 2–4, Table 2, Fig. 12). The upper unit of LFA 1 at North Ferriby indicates that at some point after the Ferriby moraine ridge was formed, ice re-advanced over it. Whether this was a minor ice-marginal oscillation or a post Skipsea Till re-advance of more regional significance is unclear. Based on the geomorphological evidence from east of the Lincolnshire Wolds, the NSL retreated eastward from its maximal limits punctuated by a number of stillstands, resulting in the moraine ridges at Halton to Waltham (and its equivalent north of Hull, Fig. 3) and at St Andrews Docks (Rees *et al.* 2000). The exact timing of this retreat phase is yet to be firmly established. Ages from sediment directly overlying LFA 1 (Skipsea Till) at Dimlington and Skipsea indicate that the NSL retreated off-shore before 17.0 ± 0.9 ka (Stage 2, Table 2) and this may have been as early as *c.* 18.0 ka based on the stratigraphically lowest ages from Dimlington and Sewerby (Bateman *et al.* 2015). This places the Halton-Waltham stillstand after *c.* 21.6 and before 18 ka. The new data also help with the ‘glaciologically’ problematic western extension of the NSL as shown in Bateman *et al.* (2015: see fig. 10, 17–18 ka reconstruction). This NSL extension arose as data available at that time showed Lake Humber was impounded (presumed by ice in the Humber Gap) whilst at the same time the Dimlington site (~45 km to the ESE) was ice free. However such a large and discrete extension of the NSL westward when most ice in the NSL was flowing southward is unlikely unless the rest of the NSL margin across the Holderness region was confined in some way. The new data presented here showing a punctuated withdrawal of ice eastward of the NSL across the south of the region mean any extension

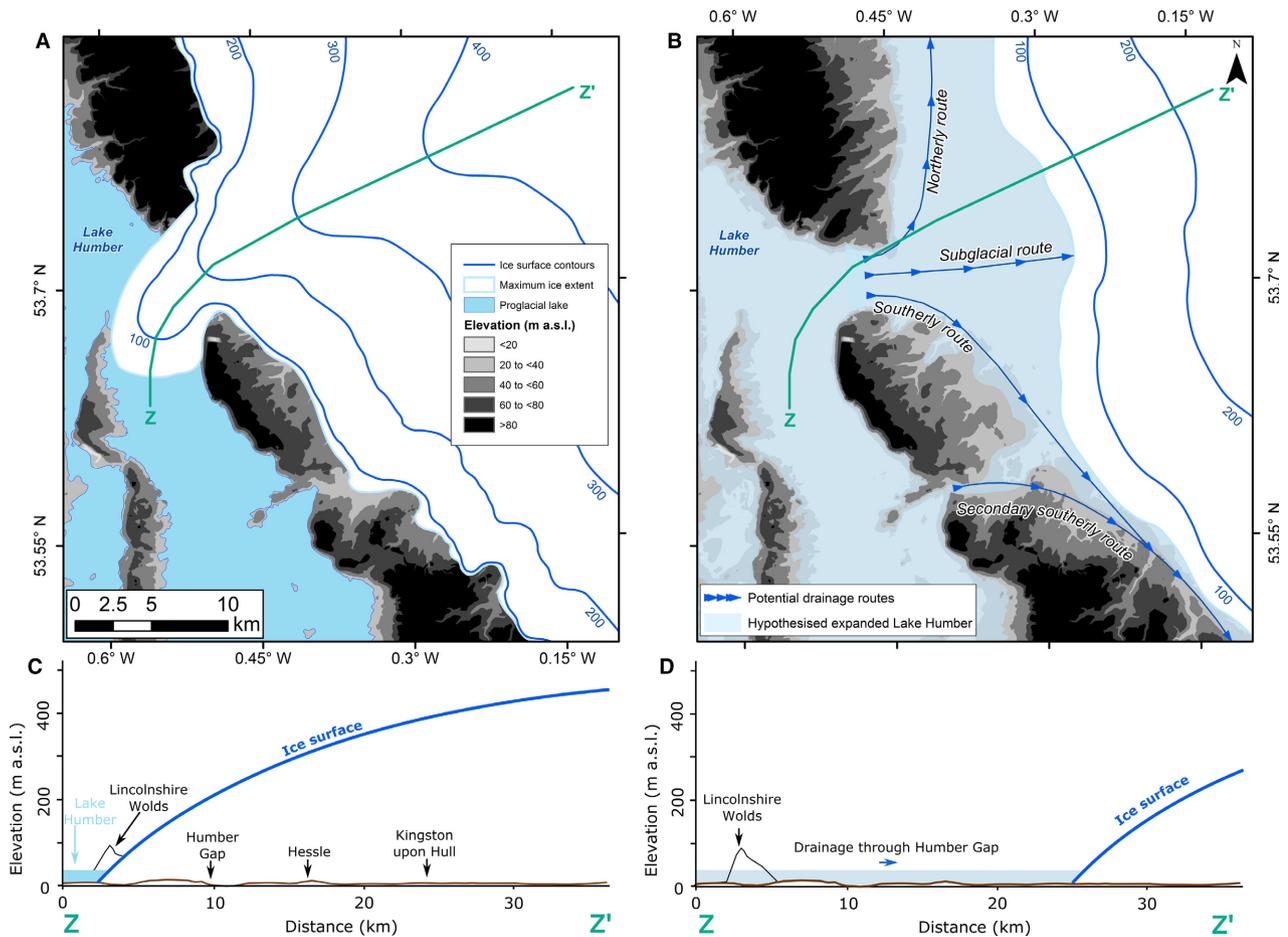


Fig. 11. Estimated NSL ice extent and thickness across the Humber gap region during the Late Devensian (Dimlington Stadial) at two stages. A. Maximum extent is based upon mapped diamicton distribution and geomorphological evidence (Fig. 2). Lake Humber level of 32 m a.s.l. shown from Fairburn & Bateman (2016). B. A second stage is shown whereby ice has retreated to the east. Several possible scenarios for the drainage of Lake Humber are depicted. C. Cross-section profile of topography and reconstructed ice surface at its maximal limit. D. Cross-section profile of topography and reconstructed ice surface at its second marginal position. Both basal topography and ice surfaces are based on transect Z to Z'. Lincolnshire Wold topography shown in Fig. 5C and (D) is for reference only and taken from a transect parallel to Z to Z'.

would have been much smaller and confined to the topographical low associated with the outer Humber estuary. This improves its glaciological plausibility. Perhaps of more significance is the timing and style of re-advances in the region. Re-advances are recorded at Skipsea and Dimlington whilst proximity of ice to the region is reflected in the ice-dammed lake sediments found at Hemingbrough, Heselton and Ferrybridge. When combined these date to around 16.8 ± 0.8 ka. As the upper diamict at Skipsea is geochemically distinct from the Withernsea Till found at Dimlington (Boston *et al.* 2010) but apparently of similar age, this may indicate the occurrence in Stage 3 of near-synchronous re-advances of small lobes of the NSL with different provenances at the eastern side of the study region (Fig. 12).

In terms of final deglaciation of the region (Stage 4, Table 2, Fig. 12), the results from Little Catwick show an

early deglaciation (before 22.8 ± 1.8 ka) at the time that NSL ice in the Humber region was still advancing over Dimlington (after 20.5 ± 1.2 ka) and extending to the North Ferriby site (mean site age of 21.2 ± 1.6 ka). Whilst it is possible that the OSL ages from Little Catwick are over-estimates due to partial bleaching, they could indicate that ice remained only briefly in the north-western part of the region before retreating. Such an early retreat would have been to a position west of Barmston where an age of 21.5 ± 1.1 ka was obtained for the uppermost part of the Skipsea Till (Bateman *et al.* 2015). A short-lived advance and early retreat in the north of the region is supported by the regional geomorphology, because the moraine ridge complex north of Kingston upon Hull does not extend into this area and ice-transverse ridges are only mapped on the eastern side of the region (Evans *et al.* 2001; Fig. 1). Final withdrawal of the ice from the present-day terrestrial landscape

Table 2. Summary events and chronology for the NSL based on new and previously published OSL ages within the Humber Gap and region to the east of it. (1) = Bateman *et al.* (2015); (2) = Bateman *et al.* (2008); (3) = Murton *et al.* (2009); (4) = Evans *et al.* (2017); (5) = this paper; (6) = Fairburn & Bateman (2016). Averages are arithmetic with 1 SD uncertainties.

Event	Site	Relationship to ice	Ages (ka)	Average ages (ka)
Pre-Skipsea Till ice advance	Flamborough	Sand beneath LFA 1 diamicton	20.0±1.4 (1)	20.5±0.44
	54°6'15" N, 0°7'6" W		20.3±1.3 (1)	
	Dimlington	Silts beneath LFA 1 diamicton	21.2±1.5 (1)	
	53°40'11" N, 0°6'26" E		20.5±1.2 (1)	
Stage 1. Skipsea Till ice advance	Hemingbrough	Proto-Lake Humber requiring ice blockage in region	24.1±2.2 (3)	21.6±1.3
	53°46'35" N, 0°58'42" W		21.9±2.0 (3)	
			21.0±1.9 (3)	
			20.5±0.6 (1)	
	North Ferriby	Moraine	21.7±2.0 (5)	
	53°42'44" N, 0°31'1" W		19.2±1.7 (5)	
	Barmston	Sand in LFA 1 diamicton	22.5±1.6 (5)	
	54°0'59" N, 0°12'55" W		21.5±1.1 (1)	
Stage 2. Skipsea Till ice retreat	Dimlington	Sand between two diamictons	18.0±1.0 (1)	Pre 17.0±0.9
	53°40'11" N, 0°6'26" E		17.8±1.0 (1)	
			16.8±1.0 (1)	
			15.8±0.9 (1)	
	Skipsea	Sand between two diamictons	15.9±1.3 (5)	
	53°58'45" N, 0°11'58" W		16.5±1.0 (5)	
	Sewerby	Sand & gravel over LFA 1 diamicton	17.9±1.9 (1)	
	54°5'54" N, 0°10'7" W			
Stage 3. Ice re-advances and proximal ice dam in region	Ferrybridge	Ice-dammed lake	16.8±1.2 (2)	16.8±0.8
	53°42'47" N, 1°17'15" W			
	Heslerton	Fan into ice-dammed lake formed after ice retreat	17.6±1.0 (4)	
	54°10'43" N, 0°35'52" W		17.3±1.0 (4)	
	Hemingbrough	Ice-dammed lake	16.8±0.5 (1)	
	53°46'35" N, 0°58'42" W		15.5±0.5 (1)	
Stage 4. Deglaciation, early in NW, later at coast	Little Catwick	Sand over LFA 1 diamicton (inland and north)	23.4±1.7 (5)	Pre 23.1
	53°53'3" N, 0°17'48" W		22.8±1.8 (5)	
	Barmston	Sand and gravel above LFA 1 diamicton (coast, north)	15.0±1.2 (1)	Pre 14.2±1.5
	54°0'59" N, 0°12'55" W		15.0±1.2 (1)	
	Heslerton	Moraine-dammed lake	15.8±0.8 (4)	
	54°10'43" N, 0°35'52" W			
	York Moraine	Moraine-dammed lake	15.9±0.9 (6)	
	53°56'56" N, 1°3'39" W		15.2±0.7 (6)	
Berrygate Hill/Mill Hill	Sand above LFA 4 diamicton (coast, south)	12.1±0.7 (5)		
53°42'33" N, 0°8'33" W		12.5±0.8 (5)		
			12.4±0.9 (5)	

occurred before 14.2±1.5 ka based on ages from sediment overlying diamicton at Barmston and Berrygate Hill/Mill Hill as well as ages from proglacial lakes no longer ice dammed but retained by moraine ridges at Heslerton and York (Table 2, Fig. 12). Of these ages, those at Barmston, based as they are on sediment laid down directly above diamicton are probably closest to the true deglacial age of the region.

The spatial and temporal complexity of advances, stillstands and re-advances calls into question the current two-stage advance/retreat model for the NSL, wherein the Skipsea advance dates to *c.* 20.9–17.3 ka and the Withernsea advance to 17.1–15.1 ka (Bateman *et al.* 2015). The data presented here strengthen the previously proposed reconstructions implying the NSL associated with the Skipsea Till initially advanced into the region and attained its maximum extent at around 21.6±1.3 ka. In the northwest of the region the NSL did not lie at the

maximal limits for long but instead rapidly retreated to a point just inland of the present-day coastline. Elsewhere ice stayed at the maximal extent before punctuated retreat off-shore by probably *c.* 18 ka. After this, at around 16.8 ka, there occurred in different parts of the region a number of near-synchronous re-advances of small subsidiary lobes of the NSL. For ice to have advanced westwards, the NSL as a whole must have been expanding and fluctuating at this time. All evidence suggests that ice no longer affected the region directly by *c.* 15 ka.

To put this into the wider context, the Eurasian ice sheet as a whole attained its maximum extent at a similar time to the Stage 1 advance of the NSL at 21 ka (Hughes *et al.* 2016). Within this however, an west–east asynchronicity was noted for the Eurasian ice sheet (Hughes *et al.* 2016). To the west, the Irish Sea lobe (ISL) on the western side of the BIIS advanced to its

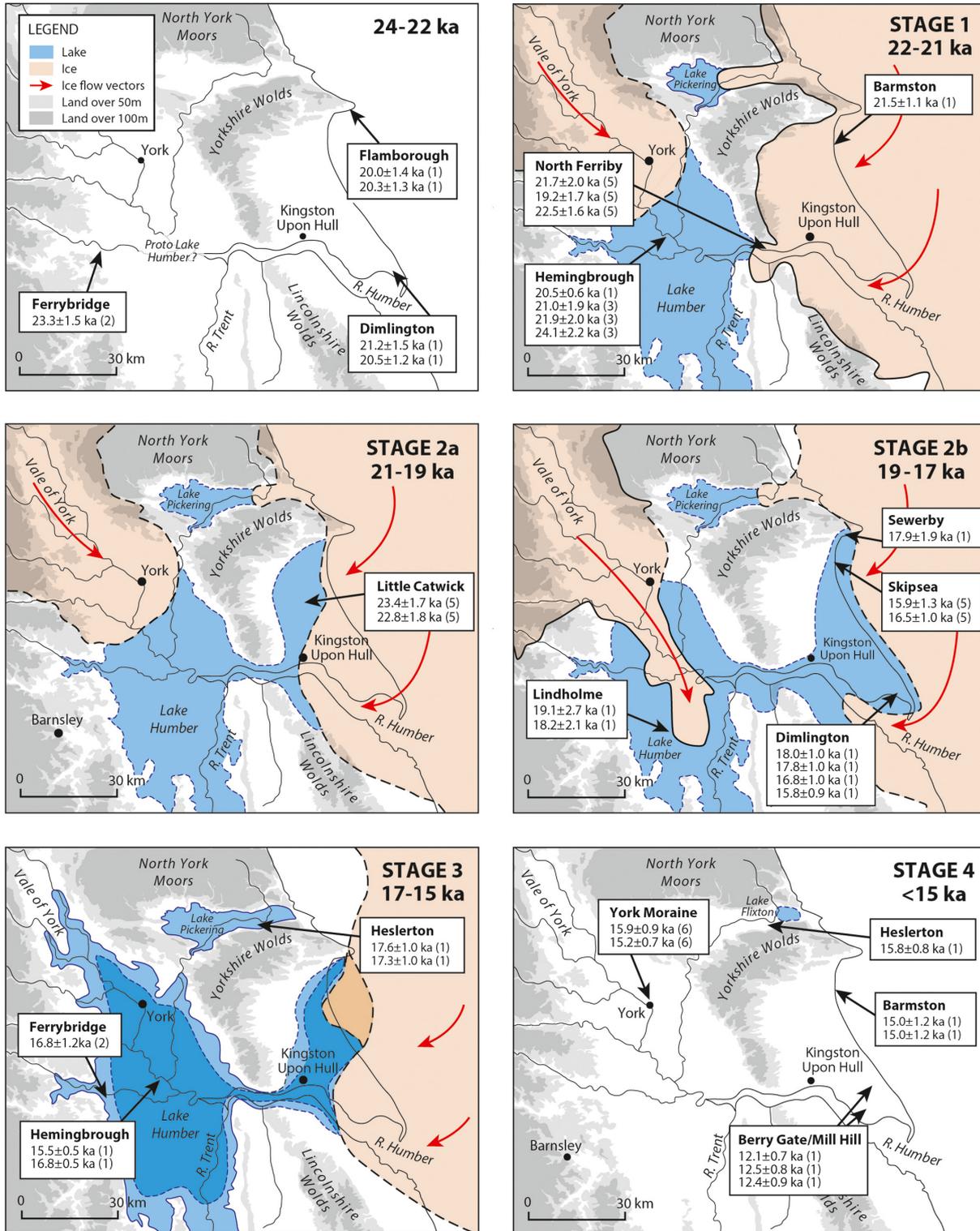


Fig. 12. Schematic reconstruction through time of the North Sea Lobe (NSL) of the BIIS in the Humber Gap region based on geomorphic, sedimentological evidence and luminescence ages (updated from Bateman *et al.* 2015). Ice margins are illustrative only and are based, where available, on dated stratigraphical and geomorphic evidence indicating either ice-free or ice-inundated conditions. Please note, within stages margins may have advanced/retreated considerably from that shown. Vale of York Ice presumes rapid advance to the York-Esrick moraine complex and then the short-lived more southerly extent as outlined by Friend *et al.* (2016). Between 24 and 22 ka, neither proto Lake Humber's extent nor the proximity of NSL to the coast at this time is known and so neither is shown. Lake Humber margins in Stages 1 and 2 are schematic and in Stage 3 are based on those of Fairburn & Bateman (2016). Lake Pickering extent is based on Clark *et al.* (2004) and Evans *et al.* (2017) with the vestige (Lake Flixton) based on Palmer *et al.* (2015). Ages are from the sources outlined in Table 2.

maximum on the Scilly Isles before 24 ka (Smedley *et al.* 2017). Elsewhere, the NW part of the BIIS retreated back from maximal positions on Lewis before *c.* 25 ka (Bradwell & Stoker 2015). These ages are both earlier than the NSL Stage 1 advance. In contrast, maximal limits of the Eurasian ice sheet were only attained to the east on the Russian Plain at *c.* 20–19 ka (Hughes *et al.* 2016). This is a period when the NSL was already starting to retreat and the ISL had retreated back almost 400 km to the Llŷn Peninsula (Chiverrell *et al.* 2013). The NSL Stage 3 advance correlates well with oscillations of the Fennoscandian ice sheet timed at 18–16 ka (e.g. Sejrup *et al.* 2009) and the Fladen 2 northern North Sea advance (e.g. Sejrup *et al.* 2015). However, that the NSL was still as far south as Holderness, Yorkshire, at *c.* 16.8 ka with ice-free conditions only attained by 15 ka remains at odds with other dated BIIS retreat evidence. Livingstone *et al.* (2012) indicate retreat of the Tyne gap ice lobe, ~170 km north of the Humber Gap, before 16.4–15.7 ka. Additionally, Hughes *et al.* (2016) show the BIIS confined to Ireland at Scotland by 16 ka with the moraines at Wester Ross (~480 km north of the Humber Gap) dating to 14 ka (e.g. Ballantyne *et al.* 2009). Thus the final retreat pattern of the NSL requires further work.

The findings presented here also provide insights into the relationship of NSL advances and retreats and proglacial lakes. In the Vale of York, Humber Gap and areas to the east, proglacial lakes appear to have been present (although not necessarily continuous or connected) as NSL ice moved on-shore at around 21.6 ka through to sometime after *c.* 15.5 ka (Fairburn & Bateman 2016). Bateman *et al.* (2015: see fig. 10, 21–19 ka reconstruction) proposed a proto Lake Humber based on laminated clays found at Hemingbrough and dated to around 21.9 ka. This lake could only have formed if eastward discharges of water were blocked by NSL ice advancing towards the Humber Gap, although the position and evidence for the ice dam that caused the drainage impoundment were uncertain (Bateman *et al.* 2015). In addition to the laminated basal (below sea level) laminated sediments from boreholes (Gaunt *et al.* 1992), the new work at North Ferriby in the Humber Gap, with intraclasts in LFA 1 and development of an LFA 1 glaciectonite derived from LFA 2b lake deposits, confirm the existence of proto Lake Humber sediments extending into the Humber Gap. The attribution of LFA 1 at North Ferriby to Skipsea Till and the fact that the new OSL assessments are coeval with the dated lacustrine sediments at Hemingbrough (Murton *et al.* 2009; Bateman *et al.* 2015) prove ice moving into the Humber Gap dammed a proto Lake Humber. Additionally, the laminated sediments of LFA 2b have been cannibalized by northeasterly sourced glacier stress to create the moraine ridge and hence they indicate that the ice dam was to the east of the Humber Gap. Thus as the ice margin

thickened and moved westwards and on-shore, a proglacial lake or a series of proglacial lakes were formed. The lake associated with the Dimlington silts (*sensu* Penny *et al.* 1969; Rose 1985) could be associated with this phase (Bateman *et al.* 2015). The new evidence from North Ferriby (onlapping of LFA 2b onto a moraine) also shows that as ice pulled back, an ice dam was maintained allowing continued impoundment of Lake Humber. Evidence from continued high levels of Lake Humber through Stages 2 and 3 (Table 2) shows that the lake must have expanded eastward as the ice withdrew. Ice must therefore have been able to maintain a dam, perhaps remaining pinned to the south side of Flamborough head and the eastern limb of the Lincolnshire Wolds (Fig. 1).

Others have shown that the NSL caused ice-dammed lakes during similar time frames. Livingstone *et al.* (2015) showed that the formation of Lake Wear was by separation (retreat) of the Tyne Gap ice from the NSL at 18.7–17.1 ka. Evans *et al.* (2017) dated the later stages of Lake Pickering to *c.* 17.5 ka when the NSL was blocking the eastern end of the Vale of Pickering. Davies *et al.* (2012) in their Stage III NSL ice advance (21 ka until 16 ka) showed in addition to Lakes Pickering and Wear, a lake impounded in the Tees basin (Lake Tees). Two new distinctions can be drawn. Firstly, our new evidence shows for the first time that proglacial lakes formed as NSL ice advanced eastward upslope as well as during its retreat. Secondly whilst the number, spatial coverage, duration and depth of proglacial lakes on the western margin of the NSL will have varied through time, proglacial lakes were relatively frequent and persistent. They appear to have been largest where reversed gradients (relative to the ice) were low and the topography had low relief. Both of these points have implications for ice marginal stability (see below).

The dynamic nature of the NSL has been discussed previously, and is thought to have been characterized by a rapid advance from Scotland down the North Sea to reach a maximal limit around Norfolk in a short space of time (e.g. Eyles *et al.* 1994). Modelling suggests that this also retreated quickly (e.g. Boulton & Haggdorn 2006). Such a dynamic NSL is not clearly attributable to climatic changes (Bateman *et al.* 2015). Switches in ice flux from the Tyne, Tweed and Forth ice streams have been suggested to have influenced the flow dynamics of the NSL (Livingstone *et al.* 2012). Boston *et al.* (2010) highlighted this by recognizing at least five flow units at the southern end of the NSL of which three are to be found in the area to the east of the Humber Gap. The NSL appears to have been particularly sensitive to changing ice dynamics and its interactions with Scandinavian ice (e.g. Carr *et al.* 2006; Clark *et al.* 2012). Debutressing of the BIIS from the Fennoscandian ice may have been the main driver (e.g. Sejrup *et al.* 2016). For example, an easterly shift in the main, Scottish-nourished axial trunk of the NSL could have allowed

terrestrially based inland ice (e.g. from the Forth) to flow into the North Sea basin and southward into eastern England. The large longitudinal extent of the NSL has been described as an ice stream or unstable surging lobe (Boston *et al.* 2010; Evans & Thomson 2010) drawing down significant amounts from BIIS ice centres and forming a relatively thin ice lobe. This model of thin ice is supported by the mapped till limits being only up to ~100 m above sea level on the Yorkshire Wolds (Fig. 4). Whilst the reconstructed ice thicknesses of 400 m in our new work would indicate thick ice, these estimated thicknesses would be over-estimates where ice was unstable. Ice draw down and streaming/surging have been demonstrated to be inextricably linked to the development of proglacial, ice-dammed lakes, for example at the receding margins of the Laurentide Ice Sheet (Evans *et al.* 1999, 2006, 2008; Stokes & Clark 2004). The persistence and prevalence of ice-dammed lakes such as shown in the Humber Gap and its vicinity are evident in the glacial stratigraphy of eastern England (e.g. Straw 1979; West 1993; Roberts *et al.* 2013) and may therefore have also been an important contributory factor in the NSL dynamism.

As much of the former footprint of the NSL is now submerged under the North Sea, the Humber Gap and the region to its east serve as an important evidential analogue for what was probably happening elsewhere in NSL when it was active. For example as NSL ice moved southwards to the topographical rise of Dogger Bank a proglacial lake may have formed. Likewise with the confluence of the NSL with Fennoscandian ice North Sea drainage would have been limited to the channel between France and England. As a result much of the NSL margin at its maximal limit may have been in contact with a proglacial lake (Clark *et al.* 2012: fig. 18). In both examples high levels of ice draw down and increased ice velocities could have occurred. As the off-shore NSL ice limits become better understood, so modelling in areas of potential proglacial lakes may be of significance in understanding the causes of both the advance and retreat patterns of this dynamic ice body.

Conclusions

- The identification and dating of the Ferriby moraine ridge in the Humber Gap provide, for the first time, the westerly maximal limit of incursion of the North Sea Lobe of the BIIS.
- New ages further refine the timing of advance and retreat of the NSL, leading to modification of the pre-existing two-stage model. In Stage 1, Skipsea Till advance occurred regionally at *c.* 21.6 ka and retreated off-shore by *c.* 18 ka (Stage 2). Punctuated retreat is evident in the southern part of region whilst in the north retreat was initially rapid towards the present-day coast. In Stage 3, a series of near-

synchronous ice advances (including the Withernsea Till advance) occurred *c.* 16.8 ka with full withdrawal of ice from the region occurring *c.* 15 ka (Stage 4).

- Laminated intraclasts in LFA 1 at North Ferriby confirm the existence of a proglacial lake (proto Lake Humber) impounded by ice advancing westwards across the Humber Gap region.
- Proglacial Lake Humber persisted during Stages 2 and 3 expanding eastward as the NSL ice retreated from the Humber Gap with an ice dam being maintained across the Holderness embayment.
- The formation of proglacial lakes both during the advance and retreat of NSL wherever it encountered low topography and reverse gradients were relatively common and persistent. As such they provide an additional mechanism to explain the dynamism of the NSL.
- Advance and retreat into proglacial lakes is a scenario probably repeated in other off-shore parts of the NSL and elsewhere in the BIIS. Quantifying this would greatly help in the understanding of the ice sheet and its demise.

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Supporting Information

Additional Supporting Information may be found in the online version of this article at <http://www.boreas.dk>.

Fig. S1. Example palaeodose (D_e) distributions as measured from small (2 mm) aliquots from the North Ferriby (left column) and Berrygate Hill/Mill Hill (right column) sites.