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Glacial Lake Pickering: stratigraphy and chronology of a proglacial lake dammed by the North Sea Lobe of the British–Irish Ice Sheet

DAVID J. A. EVANS,[1]* MARK D. BATEMAN,[2] DAVID H. ROBERTS,[1] ALICIA MEDIALDEA,[2] LAURA HAYES,[2] GEOFF A. T. DULLER,[3] DEREK FABEL[4] and CHRIS D. CLARK[2]

[1]Department of Geography, Durham University, South Road, Durham DH1 3LE, UK
[2]Department of Geography, University of Sheffield, Sheffield, UK
[3]Institute of Geography and Earth Sciences, Aberystwyth University, Penglais Campus, Aberystwyth, UK
[4]School of Geographical and Earth Sciences, University of Glasgow, Glasgow, UK

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ABSTRACT: We report the first chronology, using four new optically stimulated luminescence dates, on the sedimentary record of Glacial Lake Pickering, dammed by the North Sea Lobe of the British–Irish Ice Sheet during the Dimlington Stadial (24–11 ka cal BP). Dates range from 17.6 ± 1.0 to 15.8 ± 0.9 ka for the sedimentation of the Sherburn Sands at East Heslerton, which were formed by multiple coalescing alluvial fans prograding into the falling water levels of the lake and fed by progressively larger volumes of debris from the Wolds. Fan formation ceased ~15.8 ka, at a time when permafrost was degrading and nival-fed streams were no longer capable of supplying sediment to the fans. A further age of 10.1 ± 0.7 ka dates the reworking of coversand into the early part of the Holocene, immediately post-dating Younger Dryas periglacial structures. A 45-m lake level dates to ~17.6 ka, when the North Sea Lobe was already in retreat, having moved eastward of the Wykeham Moraine; it stood further east at the Flamborough Moraine by ~17.3 ka. The highest (70 m) lake level and the occupation of the Wykeham Moraine date to an earlier phase of the North Sea Lobe occupation of the Vale of Pickering.

KEYWORDS: British–Irish Ice Sheet; Glacial Lake Pickering; North Sea Lobe; OSL dating; Sherburn Sands.

Introduction to Glacial Lake Pickering

The sedimentary and stratigraphic record of the recession of the North Sea Lobe of the British–Irish Ice Sheet (BIIS) is best documented along the coast of eastern England, specifically in the tills and associated glacilacustrine deposits of Holderness (Catt, 1991, 2007; Evans *et al.*, 1995; Boston *et al.*, 2010; Evans and Thomson, 2010) and the Humber Estuary and North Lincolnshire (Straw, 1961, 1979; Gaunt, 1981; Bateman *et al.*, 2008, 2015). The style of North Sea Lobe recession, based largely upon these onshore landform–sediment assemblages, has been hypothesized by Clark *et al.* (2012) based upon a restricted number of chronostratigraphic control points. From this, a palaeoglaciology has been proposed that acknowledges the damming of regional drainage to produce glacial lakes along the Yorkshire and Durham coastlines. Significant advances have recently been made in securing a chronology for ice recession from the sites on Holderness and the inner Humber Estuary, the former relating to the Dimlington Stadial type site (cf. Penny *et al.*, 1969; Rose, 1985; Bateman *et al.*, 2015) and the latter relating more specifically to Glacial Lake Humber (Bateman *et al.*, 2008; Fairburn and Bateman, 2015). Glacial lakes further north, such as lakes Pickering, Eskdale, Tees and Wear (Kendall, 1902; Agar, 1954; Smith, 1981; Plater *et al.*, 2000) are relatively poorly constrained chronologically. We report here on attempts to refine the chronology of glacial lake development in the region in relation to the North Sea Lobe, concentrating specifically on Glacial Lake Pickering.

The Vale of Pickering today is an east–west-orientated low-lying plain bounded on three sides by the Howardian Hills (west), the chalky North Yorkshire Wolds (south-east) and the limestone of the North Yorkshire Moors (north; Fig. 1). The damming of pre-glacial easterly river drainage from the North Yorkshire Moors and Yorkshire Wolds by the onshore advance of the North Sea Lobe of the BIIS has long been acknowledged and depicted on palaeoglaciological maps in the form of Glacial Lake Pickering (70–30-m OD shoreline altitude range) and its eastern moraine dam, referred to as the Flamborough Moraine (Fig. 1; Farrington and Mitchell, 1951; Kendall, 1902). This event has had a lasting impact on the regional drainage network, in that the River Derwent, after rising <5 km from the North Sea on the dip slope of the North Yorkshire Moors, no longer flows directly east to its nearest coastline, as it likely did, and is presumed to have done (cf. King, 1965), before the last glaciation, but instead flows 160 km westward up the Vale of Pickering and down the Kirkham Priory gorge (presently at 20 m OD) towards the Humber Estuary. This drainage pattern was dictated by a combination of ice and then moraine damming of the east end of the Vale of Pickering and Vale of Scalby as well as the deepening of the Forge Valley and Kirkham Priory gorge as glacial lake spillways (Fig. 1).

Kendall (1902) was the first to identify shorelines at 68–70 and 45 m OD. For the lake to reach this level, the western end of the vale had to have been blocked by the margin of the Vale of York ice lobe at the Coxwold–Gilling Gap, cutting off that potential drainage route at 68 m OD. Because the pre-spillway incision altitude of the Kirkham Priory gap was at around 60 m (Fig. 1e; Edwards, 1978), the presence of Vale of York ice would also have been required in the Kirkham Priory gap to block or restrict outflow and thereby produce the 70-m OD level of Glacial Lake Pickering. Kendall (1902) associated the 70-m shoreline with a kame terrace/ice-contact delta and hummocky moraine belt between West Ayton and Wykeham (King, 1965; Edwards, 1978). This landform assemblage was termed the Wykeham moraine and used to define the BIIS North Sea Lobe 'Wykeham Stage' of Penny and Rawson (1969), when ice sheet marginal meltwater was forced to flow into the lake along the Forge Valley. With its highest altitude

*Correspondence to: D. J. A. Evans, as above.
E-mail: d.j.a.evans@durham.ac.uk

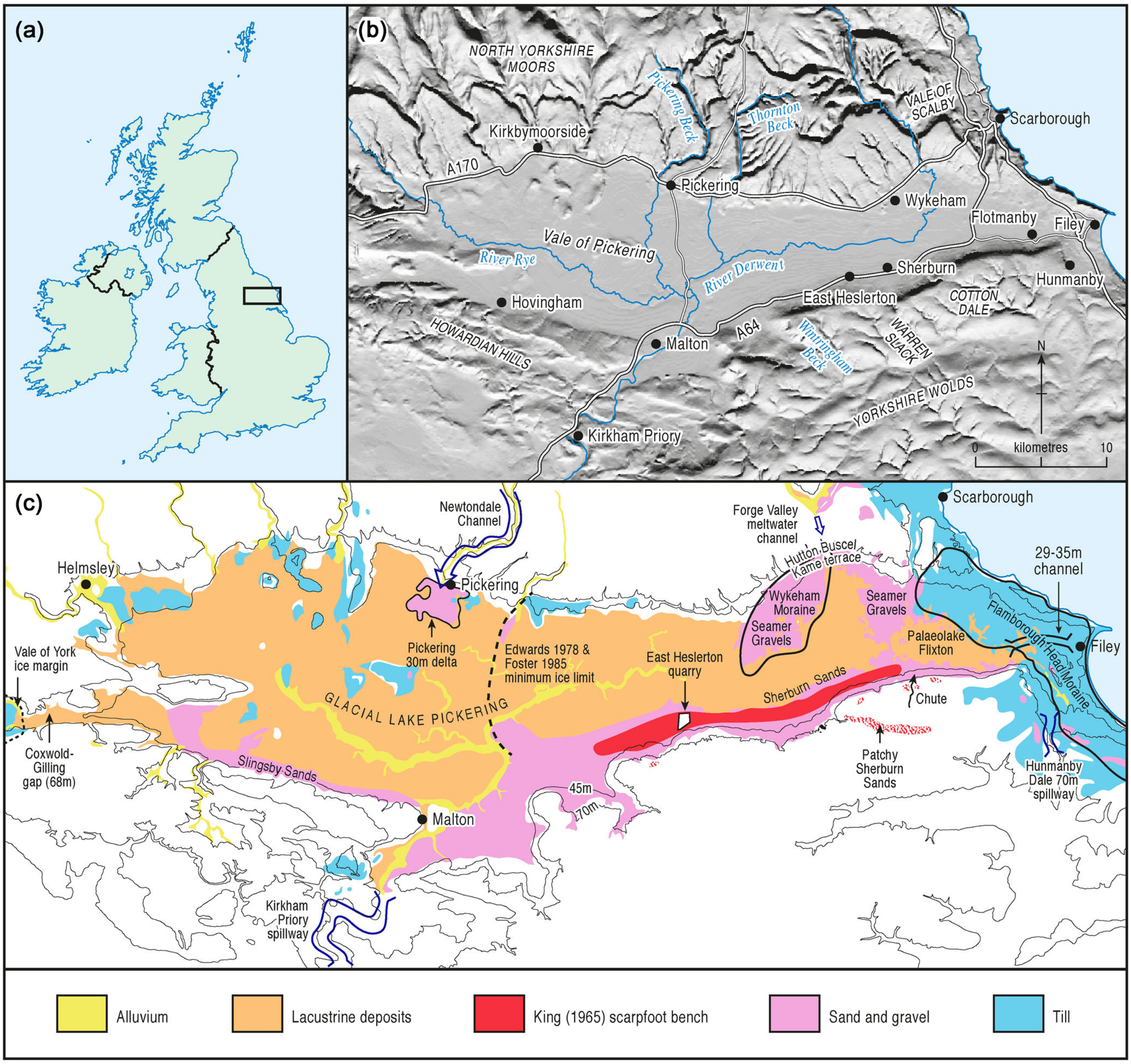

Figure 1. Maps of the study area and the glacial geomorphology pertinent to Glacial Lake Pickering. (a) Location map for the Vale of Pickering; (b) digital elevation model of the Vale of Pickering and surrounding terrain; (c) detailed map of the surficial geology (after BGS Digimap sources) and key landforms and sediments from previous research and this study; (d) traditional palaeoglaciological map of the ice sheet margins and associated ice-dammed lakes around the North Yorkshire Moors (after Kendall, 1902); (e) map of the Kirkham Priory gorge, showing present-day contours and reconstructed pre-incision contours used by Edwards (1978) to propose a 'passive' overflow route at 60–61 m OD. This demonstrates that the Vale of York ice is required as a dam for the 70-m level of Glacial Lake Pickering.

at 70 m OD, Hunmanby Dale is a likely candidate for a lake spillway for the Wykeham Stage; it was interpreted by King (1965) as a spillway for a small lake located north of Hunmanby but its large size is inconsistent with drainage of such a small volume of water. It could have served as a spillway for the 70-m Lake Pickering stand by one of two mechanisms: (i) the southern lateral margin of the North Sea lobe receded a short distance to expose the lower scarp slope of the Yorkshire Wolds above Flotmanby; or (ii) the ice lobe allowed lake water to escape via a sub-marginal tunnel due to partial snout flotation, akin to the cyclical emptying of ice-dammed lakes along modern ice cap margins (e.g. Björnsson, 1976; Clarke, 1982, 2003; Tweed and Russell, 1999; Roberts *et al.*, 2005). The lower 45-m shoreline, purportedly controlled by down-cutting of the Kirkham Priory gorge (Edwards, 1978), was associated with an outwash fan fed by the Mere Valley at Seamer and used to define the BIIS North Sea lobe at the 'Cayton-Speeton' Stage of Penny and Rawson (1969). At this time the Flamborough Moraine (Farrington and Mitchell, 1951) was thought to have been constructed, as it forms the topography against which the Seamer Gravels (Mere Valley fan delta) were deposited (Fig. 1c). A gravel delta prograded from the Newtondale channel (spillway?) at Pickering documents a former lake level as low as 30 m OD (Kendall, 1902). Later lowstands such as this are presumed to have been controlled by further downcutting of the Kirkham Priory gorge (spillway), which presently is as low as 20 m OD at its intake, but there is no morphochronological control on such downcutting. The only other exit from the Vale of Pickering that could have controlled a 30-m lake stand is a 29–35-m OD channel through the Flamborough Moraine between Flotmanby and Filey (Fig. 1c).

Borehole records (Fig. 2) reveal that parts of the former lake floor contain up to 12 m of fine-grained laminations lying

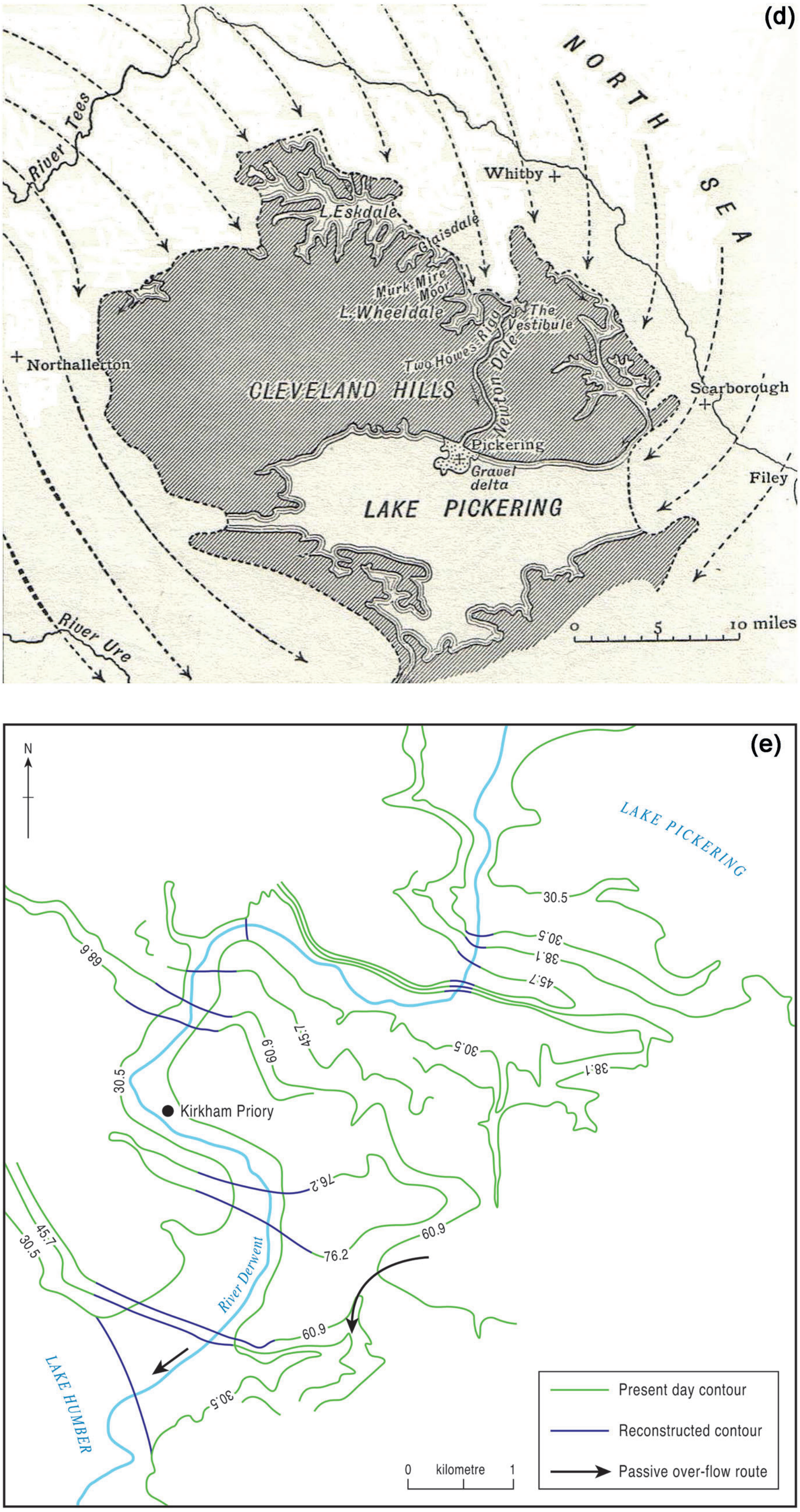

Figure 1. Continued

over bedrock and capped by up to 12 m of interbedded sequences of sands and gravels, with gravels becoming more dominant towards the Wykeham and Flamborough moraines (Edwards, 1978). The westward change along the centre of the vale from sand- and gravel-dominated sequences at around borehole 52 to sand and clay sequences (Fig. 2) reflects the increasingly ice-distal depositional environment. Borehole SE97NW31 at the Wykeham Moraine is representative of subaqueous ice-contact sedimentation and borehole SE97NE11 at Sherburn is representative of the Sherburn Sands of Fox-Strangways (1880, 1881) and the deposits central to the findings of this paper.

Although the Wykeham Moraine has been widely regarded as the limit of the North Sea Lobe in the Vale of Pickering

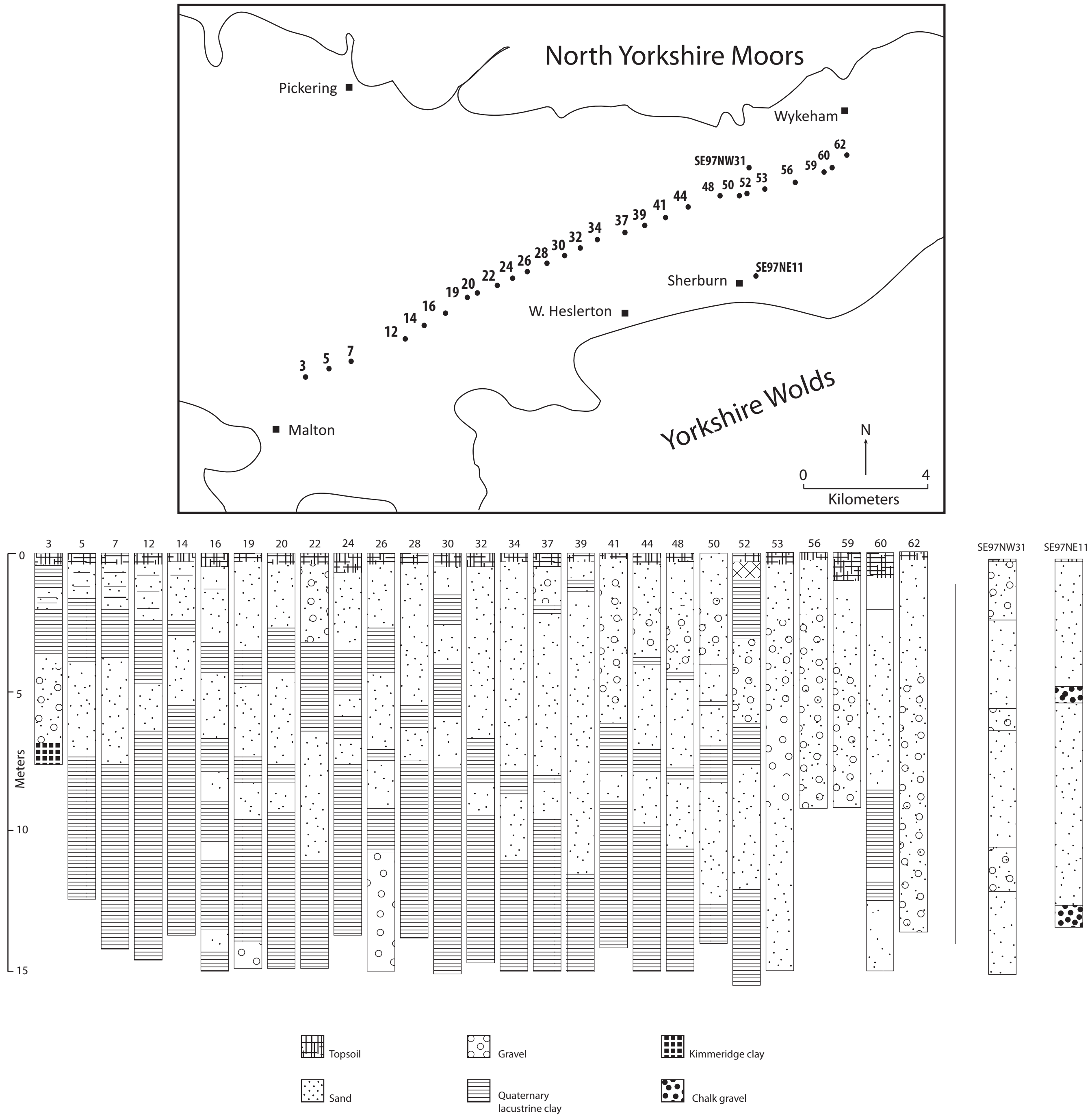

Figure 2. Compilation of selected borehole logs from the Vale of Pickering (after Edwards, 1978) and including BGS boreholes at Sherburn (SE97NE11) and at the outer edge of the Wykeham Moraine (SE97NW31).

during the Dimlington Stadial (Fig. 1a), Edwards (1978) and Foster (1985) have proposed that the lobe penetrated further west up the Vale of Pickering; Edwards (1978) used an apparent ice-marginal assemblage of till and glacifluvial ridges to propose Thornton-le-Dale as the limit, whereas Foster (1985) suggested the limit was around the town of Pickering (Fig. 1). Foster's (1985, 1987a,b) ice margin was reconstructed using glacial meltwater channels on the Wolds escarpment and the occurrence and distribution of the Sherburn/Slingsby Sands (*sensu* Fox-Strangways, 1880, 1881), which he interpreted as the deposits of an outwash train that stretches from Flotmanby to Hovingham, the full length of the former Glacial Lake Pickering (Fig. 1a). However, any variability in the sedimentary architecture and landform manifestation of the Sherburn/Slingsby Sands that must have been created by sequential ice-marginal recession has never been documented.

The work of Candy *et al*. (2015) and Palmer *et al*. (2015) has led to significant improvements in our knowledge of the postglacial evolution and archaeology of the east end of the Vale of Pickering. In the complex depression between the Wykeham and Flamborough Head moraines a vestige of Glacial Lake Pickering persisted as Palaeolake Flixton (Fig. 1a) the shoreline of which saw human occupation at Star Carr. While this research reports the oldest sedimentation records for Glacial Lake Pickering, it relates to the base of the sedimentary sequence that accumulated in the later stages of lake damming at the Last Glacial–Interglacial Transition.

Although the traditional palaeoglaciological reconstructions of Glacial Lake Pickering (Fig. 1b) have endured (King, 1965; Catt, 1991; Clark *et al.*, 2004; Evans *et al.*, 2005), the landform and sedimentary evidence proposed to support both minimum and maximum western full glacial limits for North Sea Lobe advances (Fig. 1a) has not been fully scrutinized and never dated, especially in the centre and at the western end of the vale. This paper reports on the first attempt to provide a chronology on the sedimentary record pertaining to the operation of Glacial Lake Pickering during the Dimlington Stadial (24–11 ka cal BP), specifically on the sedimentation recorded on the south side of the vale in the Sherburn Sands at East Heslerton (Fig. 1). The chronology presented for the East Heslerton deposits forms part of a wider dating programme (BRITICE-CHRONO) aiming to constrain the rate and timing of recession of the last BIIS.

Study site and methods

During excavations at the East Heslerton quarry of R. Cook and Son, extensive exposures have been created through a large valley-side depositional sequence, locally known as the Sherburn Sands (Fig. 3; Fox-Strangways, 1880, 1881). These deposits interdigitate with the Seamer Gravels at the eastern end of the vale (Foster, 1987a), a body of ice-proximal outwash associated with the Wykeham and Flamborough moraines (Fig. 1a). To the west of Malton, Fox-Strangways (1880, 1881) proposed a change in nomenclature to the Slingsby Sands in recognition of their finer grain size distribution. The Sherburn Sands occur predominantly as an elongate strip lying below the Wolds escarpment and skirting the south edge of the Vale of Pickering between the altitudes of ca. 60 and 27 m OD and mapped by the British Geological Survey (BGS) as a mixture of glacifluvial sands and gravels and sands and gravels of unknown age and origin (Fig. 1a). Small pockets of Sherburn Sands are reported by Foster (1987a) to lie on the Wolds scarp and in the south-easterly draining valleys of Warren Slack and Cotton Dale on the dip slope. The area depicted by Foster (1987a) and the BGS as the Sherburn Sands on the south side of the Vale of Pickering coincides with King's (1965) 'scarp-foot bench' or terrace of chalky gravel, which is depicted by King (1965) as lying below the uppermost Glacial Lake Pickering shoreline at 225 ft (68.5 m OD) and interpreted as the product of a postglacial solifluction terrace. East of the Wykeham moraine, Foster (1987a) and the BGS depict the Sherburn Sands 'strip' as more indented and narrow (Fig. 1a).

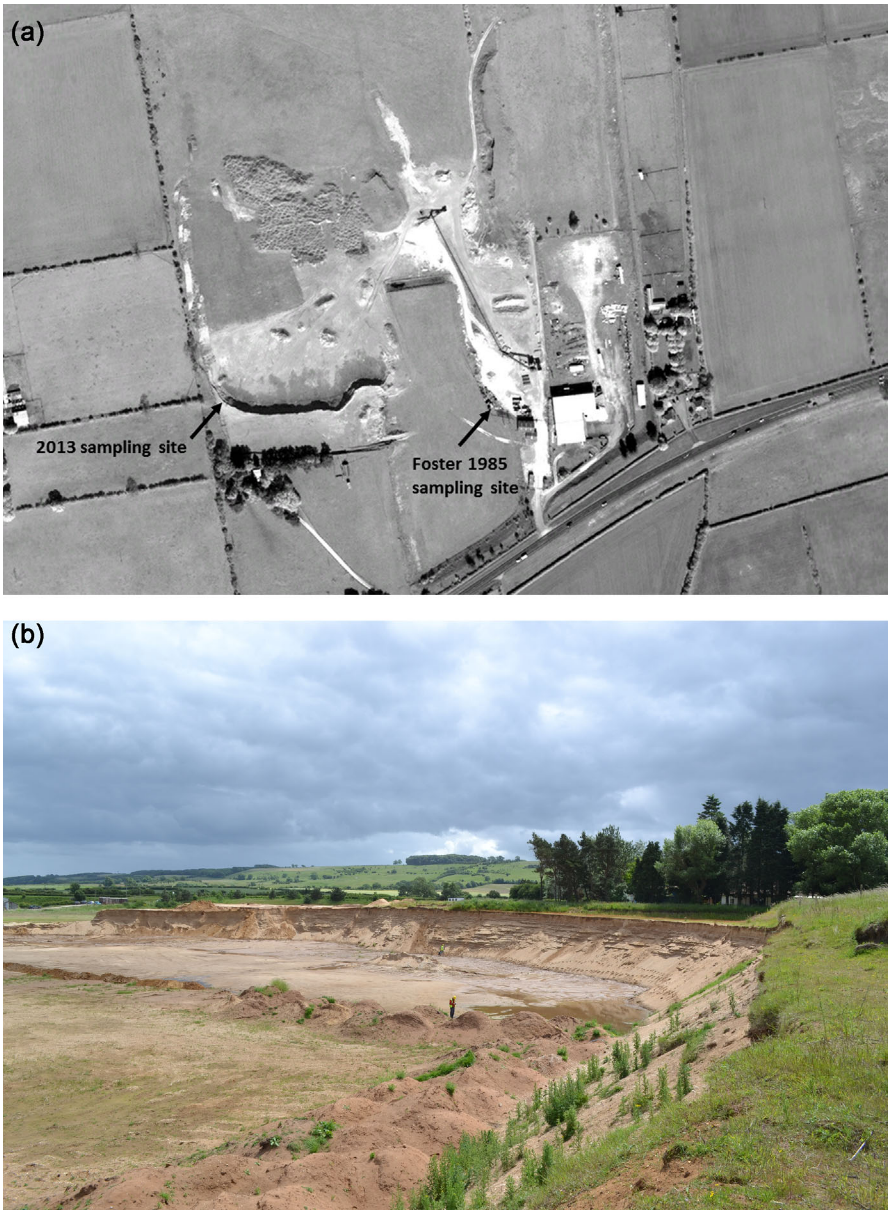

Figure 3. Locational details of the East Heslerton exposures: (a) Google Earth image of the quarry in 2013, showing the location of our sampling and logging in 2013 and by Foster in the 1980s; (b) overview of the quarry face in 2013, looking south towards the Wolds escarpment. Note the prominent tabular architecture of the sands and gravels.

At East Heslerton the bench or terrace of Sherburn Sands appears to dip northwards towards the centre of the Vale of Pickering from 50 to 30 m OD, although the northern edge is not particularly pronounced; hence it more resembles a dipping shelf than a bench. The Sherburn Sands that have been exposed occur in the middle to outer edge of the shelf (Fig. 3a). Exposures through the stratigraphic sequence at East Heslerton have been logged previously by Foster (1985, 1987a), who proposed a glacial outwash origin, with the deposits grading distally into the Slingsby Sands at the west end of the Vale of Pickering and interdigitating in the east with the glacier-proximal Seamer Gravels at the Wykeham and Flamborough moraines. Earlier notions that the deposits were lacustrine in origin and associated with Glacial Lake Pickering sedimentation (Kendall, 1902; Clark, 1954; Sheppard, 1956) were dismissed by Foster (1987a), who explained the shelf-like, valley-side distribution of the deposits as indicative of sedimentation along the left lateral margin of the glacier lobe that penetrated the east end of the vale, presumably as a feature similar to a kame terrace but grading to the falling base level of the proglacial lake to the west. Pockets of the Sherburn Sands on the Wolds scarp and dip slope valleys document the earliest stages of such sedimentation, when the ice margin overtopped the eastern escarpment summit and fed meltwater down valleys such as Cotton Dale. Foster (1987b) also identifies glacier sub-marginal chutes along the escarpment which contain pockets of Sherburn Sands. Because the Sherburn Sands shelf-like deposit continues westwards and backfills small escarpment valleys such as the Wintringham Beck valley, it has been used by Foster (1987a) to justify the more westerly or maximum glacier margin reconstruction.

Two outstanding problems arise from this reconstruction of the Sherburn Sands depositional environment: first, as acknowledged by Foster (1987a), the source of the Sherburn Sands is difficult to reconcile with direct glacial meltwater drainage, as the materials are predominantly fine grained and contain few far-travelled erratics typical of those contained within the east coast tills (Madgett and Catt, 1978); second, there are no indicators of ice-contact sedimentation, which would be abundant if the positioning of the Sherburn Sands bench was conditioned by the margin of a glacier lobe extending as far west as Thornton-le-Dale, as proposed by Edwards (1978). However, the upper shelf altitude of 40–60 m is compatible with Kendall's (1902) lower shoreline of 45 m OD and the altitude of the outwash fan delta at Seamer used to define the Cayton–Speeton Stage (Penny and Rawson, 1969). At this stage the North Sea ice lobe constructed the Flamborough Moraine (Farrington and Mitchell, 1951) and formed the topographic barrier against which the Seamer Gravels were deposited.

Consequently, some outstanding but not unrelated questions need to be addressed using the sedimentology and geomorphology of the Vale of Pickering. First, why do the Sherburn Sands and equivalent deposits coincide with and outcrop below the Glacial Lake Pickering shorelines (45–70 m OD) if they are not lacustrine? Second, if they record glacier-marginal recession, why are they devoid of ice-contact characteristics and why is their depositional shelf at a consistent altitudinal range inside and outside the Wykeham Moraine? Third, if they represent glacial outwash, why are they concentrated in a valley-side shelf which has no kame terrace characteristics and why do palaeocurrents record former water flow in all directions except south? Fourth, as the East Heslerton deposits in particular lie below 45 m and beyond the Wykeham Moraine (i.e. in an area formerly submerged by the 70- and 45-m lake stands) why are they not deltaic?

Further to the work of Foster (1987a) we now present new logging, together with sampling for optically stimulated luminescence (OSL) dating, which was undertaken in 2013 in the exposures through the thickest part of the deposit, below the bench surface at 40 m OD, at the south end of the quarry (Fig. 3b). Vertical profile logs were compiled following the procedures set out by Evans and Benn (2004) and employing the lithofacies description and coding approach of Eyles *et al.* (1983).

Luminescence dating

Samples for OSL dating were taken using opaque plastic tubes. In the laboratory, sediment preparation followed standard procedures to isolate and clean the quartz fraction including wet sieving to separate out the dominant 180–250-μm fraction (Bateman and Catt, 1996; Porat *et al.*, 2015). Beta dose rates are based on the concentration of U, Th and K measured using inductively couple plasma mass spectroscopy (ICP-MS). Gamma dose rates are based on site measurement of radionuclide activities carried out with an EG & G MicroNomad gamma spectrometer. Cosmic radiation contributions were calculated based on average burial depths through time (Prescott and Hutton, 1994). Appropriate conversion factors (Guérin *et al.*, 2011) including attenuation by moisture and grain size were used to calculate the final total dose rate (Table 1). Moisture values were assumed at $20 \pm 5\%$ for samples well below the current water table (Shfd13054) and $10 \pm 5\%$ for those currently close to but above the current water table.

Burial doses (D_e) were measured at both the single grain (SG) and ultra-small multigrain aliquot (SA, containing ~20 grains each) levels. All luminescence measurements were carried out on automated Risø readers with blue (470 ± 30 nm) LED and green (532 nm) Nd:YVO_4 laser

Table 1. OSL dates and associated information from East Heslerton.

Field code	Lab code	Depth (m)	w (%)	β dose rate (Gy/ka)	γ dose rate (Gy/ka)	Cosmic dose rate (Gy/ka)	Total dose rate (Gy/ka)	Aliquots accepted (measured)	Aliquot size	OD (%)	Equivalent dose (Gy)	Age (ka)
HES13/1/4	Shfd13057	1	10 ± 5	0.58 ± 0.04	0.43 ± 0.03	0.18 ± 0.01	1.22 ± 0.05	72 (150)	SA	38	12.3 ± 0.6	10.1 ± 0.7
								37 (4500)	SG	32	13.8 ± 0.9	
HES13/1/3	Shfd13056	2	10 ± 5	0.61 ± 0.05	0.32 ± 0.02	0.17 ± 0.01	1.11 ± 0.05	83 (140)	SA	26	17.5 ± 0.6	15.8 ± 0.9
								40 (4500)	SG	39	19.4 ± 1.3	
HES13/1/2	Shfd13055	5.5	10 ± 5	0.53 ± 0.04	0.18 ± 0.01	0.12 ± 0.01	0.85 ± 0.04	86 (140)	SA	22	14.7 ± 0.4	17.3 ± 1.0
								37 (4500)	SG	40	16.4 ± 1.3	
HES13/1/1	Shfd13054	8.6	20 ± 5	0.65 ± 0.05	0.34 ± 0.02	0.08 ± 0.01	1.09 ± 0.05	81 (140)	SA	22	19.2 ± 0.6	17.6 ± 1.0
								51 (5000)	SG	44	21.3 ± 1.5	

stimulation for measurement of SA and SG, respectively. OSL was detected through a Hoya U-340 filter. All samples were measured using the SAR protocol (Murray and Wintle, 2003) including an IR depletion ratio step to test for feldspar contamination and a preheat of 200 °C for 10 s. The latter was derived experimentally from a dose recovery preheat test. For each sample, 100–120 SA replicates and 4500–5000 grains were measured. Derived D_e estimates were accepted if the relative uncertainty on the natural test-dose response was <20%, the recycling and the IR depletion ratio (including uncertainties) were within 20% of unity and recuperation <5%. The resulting data show that for each sample the measured D_e replicates were normally distributed with over-dispersion (values of 22–38% for SA and 32–44% for SG. While these are above the 20% normally applied to differentiate between well-bleached and incompletely bleached samples (Olley *et al.*, 2004), this may be due to extra over-dispersion associated with intrinsic factors (Roberts *et al.*, 2005; Thomsen *et al.*, 2005; Jacobs *et al.*, 2006). Dose recovery experiments carried out on artificially bleached and irradiated material from sample Shfd13055, thereby only affected by intrinsic factors, returned over-dispersion values of 11% for SA and 29% for SG. Therefore, the over-dispersion values observed on the natural dose distributions, above the 20% threshold, are not necessarily derived from incomplete bleaching. In addition, the characteristic lack of asymmetry observed in these natural distributions, similar to those reported in previous studies (Alexanderson and Murray, 2007; Rowan *et al.*, 2012), leads us to the conclusion that the four samples studied here are well bleached. Therefore, the application of Minimum Age or Internal External Consistency Criterion (IEU) models as per Medialdea *et al.* (2014) was not required. Final D_e values for age calculation purposes are therefore based on the Central Age Model (CAM, Galbraith *et al.*, 1999). An average of 9% of the accepted D_e values have been identified as outliers following the criterion of those out of $1.5 \times$ interquartile range (Tukey, 1977) and excluded from the calculations. Results also show that estimated D_e values from SG and SA are consistent within 1σ and have similar distributions (Fig. 4) indicating that no resolution is lost when using the very small (SA) multigrain aliquots. The SA data with better signal-to-noise ratio and lower uncertainties are therefore reported for age calculation purposes (Table 1).

Sedimentology and stratigraphy of East Heslerton

The earliest stratigraphic exposures at East Heslerton were logged by Foster (1985, 1987a), whose summary vertical profile log is reproduced here as Fig. 5. This exposure was located beneath what was the ca. 40-m shelf surface at SE918767 in 1985, in the east side of the pit. Foster (1985, 1987a) identified three main sedimentary units which we here classify as lithofacies (LF) 1–3. At the base of the sequence, LF1 comprised around 3.5 m of interbedded and commonly internally upward-fining tabular units of horizontal and trough cross-bedded, coarse to fine sands, also containing erratic coal and shale clast lags, angular flint and chalk fragments and rare gritstone pebbles. Localized ripple drift laminations recorded palaeocurrents towards the west and north-west. Internal structures included small (<0.50 m deep) ice wedge pseudomorphs but deeper structures extended through the sands from the overlying deposits. The sands of LF1 were overlain by 2–3 m of sandy gravel to gravelly sand (LF2), largely devoid of internal structures with the exception of distorted sand lenses. The gravel clasts in LF2 were angular to sub-angular chalk (≤90%), flint and rare gritstone. The contact between LF1 and LF2 was sharp and locally loaded. Importantly, towards the north of the pit, the gravels of LF2 became sub-ordinate to the sand and pinched out into interdigitating lenses. Internal structures included narrow (<0.20 m) but deeply penetrating sand-filled dykes, interpreted as ice wedge pseudomorphs, which extended from the top of LF2 through underlying LF1. Wider ice wedge pseudomorphs (<1.0 m deep) were evident, together with cryoturbation structures, at the upper contact of LF2. The sequence was capped by 2 m of LF3, a massive sand unit with scattered, pebble-sized clasts, which infilled cryoturbation structures and ice wedge pseudomorphs in underlying LF2 and contained localized pockets of sand 'reworked by wind action' (i.e. presumably horizontally cross-laminated).

Exposures available in 2013 were also located directly beneath the ca. 40 m OD mid-shelf area but 250 m west of Foster's (1987a) sampling sites (Fig. 3b). The sequence (Fig. 6) comprised three sedimentary units, similar to those of Foster (1987a) and hence classified similarly here as LF1–3, although the sedimentological details allowed a refinement of the lithofacies into sub-units. A further upper LF4 was also recognized.

At the base of the sequence in 2013, the basal fine to medium sands identified by Foster (1987a) and classified here as LF1, displayed a significant proportion of rhythmically bedded sand and silt laminations in its lower 1.5 m and is therefore classified LF1a. A sharp contact then separated the rhythmites from an overlying 0.75 m of scour and fill features and climbing ripple drift (LF1b; Fig. 7a). Bedforms were locally disturbed above the rhythmites by the development of small-scale water escape necks and associated angular autochthonous intraclasts, although the climbing ripples recorded a palaeocurrent towards east-north-east and east.

A scoured and erosional contact separates LF1 from overlying LF2, which comprises three sub-units. First, LF2a comprises planar-bedded sands with rare isolated clasts, clast lags and scour fills (Fig. 7b). It occurs both at the base and at the top of LF2. At the base it is 0.60–0.75 m thick and internally displays two fining-upwards sequences wherein planar-bedded sand grades upwards into horizontally bedded to laminated or rippled sands. Palaeocurrents measured from this basal unit of LF2a record a range of flow directions towards the south-west through to east. Second, LF2b comprises 2.6 m of tabular units of planar-bedded sands and sandy granule gravels separated by thin beds of horizontally bedded to laminated sand, locally draping undulating surfaces in the coarser sands and gravels (Fig. 7c). Third, LF2c comprises 2.10 m of stacked tabular units of planar-bedded sands with scour fills and clast lags, separated by horizontally bedded to laminated sand and climbing ripple drift (Fig. 7d). The top of LF2 is characterized by 0.50 m of LF2a, which has been penetrated by secondary, vertical wedge infills descending from overlying deposits (Fig. 7d). Clast lithologies in LF2 were consistent with those identified by Foster (1987a), and included a majority of angular to sub-angular chalk, with flint and gritstone.

Towards the top of the 2013 sequence was predominantly ≤1.0 m of crudely horizontally bedded but heavily disturbed (convoluted) sandy gravel with localized surface pockets of horizontally cross-laminated sand with isolated clasts, together regarded as the equivalent of Foster's (1987a) upper unit (LF3). Where less disturbed, LF3 appears as shallow-dipping interbeds of gravel clinoforms (small-scale foresets) and planar to horizontally bedded openwork gravel and

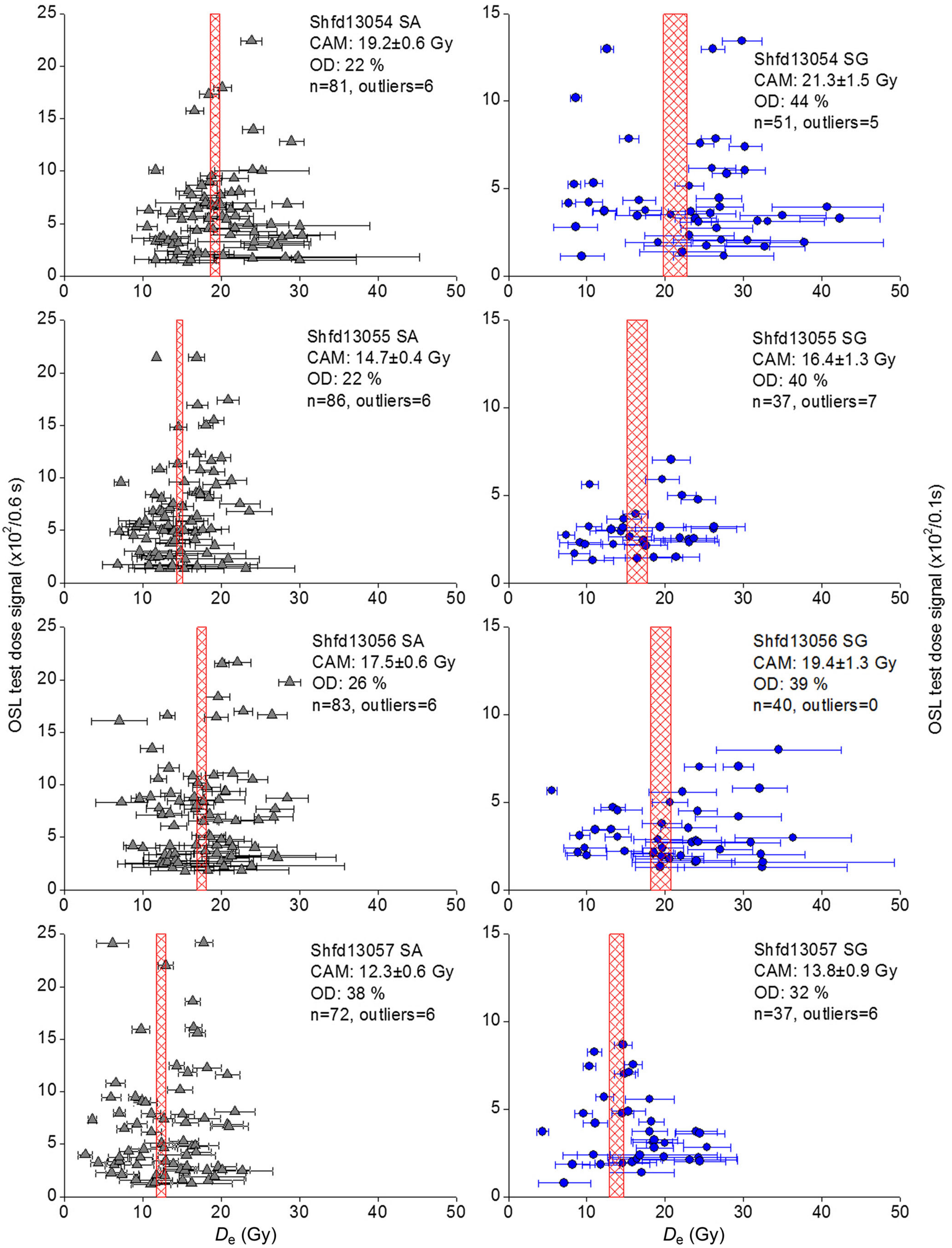

Figure 4. Dose distributions from four samples measured using the SAR protocol on small multi-grain aliquots (left column) and single grains (right column). We show the signal of the natural test dose as a function of measured dose (D_e). The estimated doses derived from CAM calculation are indicated by the vertical red bar. Over-dispersion values (OD), number of accepted aliquots (n) and the number of identified outliers are included in the plots. Outliers have been excluded from the plotted distributions and from the corresponding CAM and over-dispersion values reported.

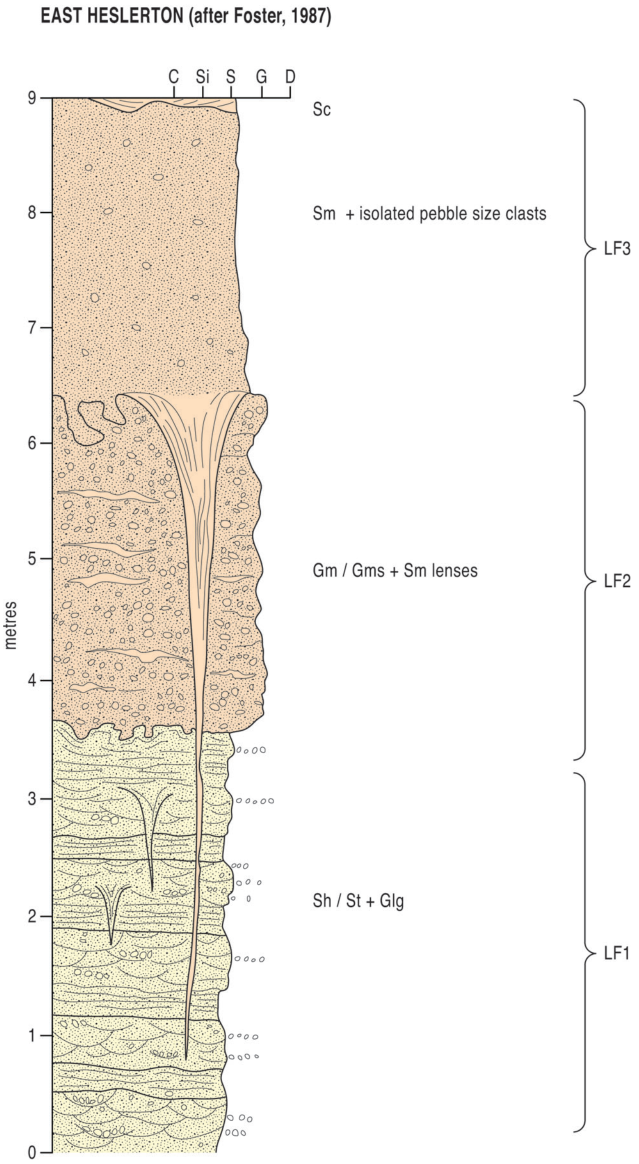

Figure 5. Redrawn vertical profile log compiled by Foster (1987a) from East Heslerton.

sandy and matrix-supported gravel (Fig. 7e). Palaeocurrents indicated by the prominent dips of the gravel clinoforms indicate flow towards the north and north-north-west. The sediments in the top 1–3 m of the sequence (LF2a, b and 3) are crosscut by vertical wedges filled with a mix of sands and gravels, arranged in sub-vertical beds that are aligned parallel to the wedge margins and overturned at the wedge top (Fig. 7d). We concur with Foster's (1987a) conclusion that these characteristics are diagnostic of ice wedge pseudomorphs and that together with the convolutions, which have characteristics indicative of cryoturbation structures (e.g. Murton and Bateman, 2007), they record the development of permafrost conditions sometime after the deposition of the sands and gravels.

Capping the 2013 sequence was a deposit not recognized previously by Foster (1987a) and classified here as LF4. It is a horizontally bedded massive brown sand with rare isolated clasts and a sharp, erosional basal contact. It lies stratigraphically above the ice wedge pseudomorphs of LF3 and represents the cold climate aeolian coversands as mapped by the BGS and found elsewhere in Vale of York and North Lincolnshire (Bateman, 1998).

Depositional environment at East Heslerton and implications for Glacial Lake Pickering

The sedimentology documented at East Heslerton is critical to our understanding of the relationship between Glacial Lake Pickering and the North Sea Lobe of the BIIS. The earliest deposits recorded in the exposures in LF1a are fine-grained rhythmites and hence document subaqueous suspension sedimentation when the water level was above 34 m OD. As the deposits lie at the margins of former Glacial Lake Pickering, where at least 12 m of laminated lake sediments have been reported (Edwards, 1978), and immediately proximal to the Wykeham Moraine, which is associated with the upper Glacial Lake Pickering shoreline at 70 m OD, the simplest interpretation of the LF1a rhythmites is that they represent deep-water glacilacustrine sedimentation.

Rhythmite deposition was terminated and sedimentation changed abruptly in LF1b, which contains stacked, locally scoured and filled, sequences of sandy fluvial bedforms such as horizontal, planar and trough cross-beds and climbing ripple drift separated by fine-grained laminations or waning discharge deposits, all indicative of shallow water. Some minor clast lags or isolated clasts indicate coarse-grained sediment starvation or a sediment source that was predominantly sandy but the grain size range indicates pulsed sedimentation by a highly variable current. Foster (1987a) proposed that the sand source for the Sherburn Sands in their entirety could have been wind-blown sand sheets from the slopes of the Wolds. Indeed, the palaeocurrents derived from bedforms in LF1b and 2a indicate a radial pattern of water flow from the base of the Wolds escarpment as a low-angled subaerial fan. A local source for materials within the Sherburn Sands would help to explain the paucity of far-travelled erratics and coarse gravels typical of proximal glacial outwash. However, the wide range of grain size from fine to coarse sand precludes an entirely wind-blown origin. The small ice wedge pseudomorphs in LF1 recorded by Foster (1987a) are syngenetic, indicating that sediment progradation was on a subaerial surface in a periglacial climate, and hence post-dating the lacustrine sedimentation recorded in LF1a. The localized disturbance of bedforms by water escape necks and brittle failed blocks is most likely indicative of elevated porewater pressures brought about by rapid fan sedimentation over freshly exposed lake bed deposits.

The sediments contained within LF2 record increased sediment discharges. This is manifest in the scoured and loaded contact at the LF1b/2a boundary, the predominantly coarser, gravelly grain sizes, greater number of partially gravel filled scours, and interbeds of planar-bedded sands and granule gravels, as well as Foster's (1987a) massive sandy gravel (matrix-supported) facies. Pulsed flows are locally recorded in fining-upward sequences in LF2a, but LF2 predominantly represents the deposits of sandy to granule gravel bedforms in braided channels typical of the fluctuating discharges of intermediate sandur systems (Miall, 1985) but with clear evidence of matrix-supported gravel deposition. Rapid distal fining of LF2 appears to be recorded by Foster (1987a) in his interdigitating lenses of sand and increasingly subordinate gravel towards the north of the pit, suggesting that sedimentation was on a shallow fan that rapidly dissipated water flow energy after the feeder streams emerged from the Wolds scarp channels. The three sub-facies visible in LF2 in 2013 record a vertical sequence of first increasingly high discharges through LF2a to LF2b and then falling discharges through LF2c to LF2a, the latter represented by a decrease in scour fills and clast lags.

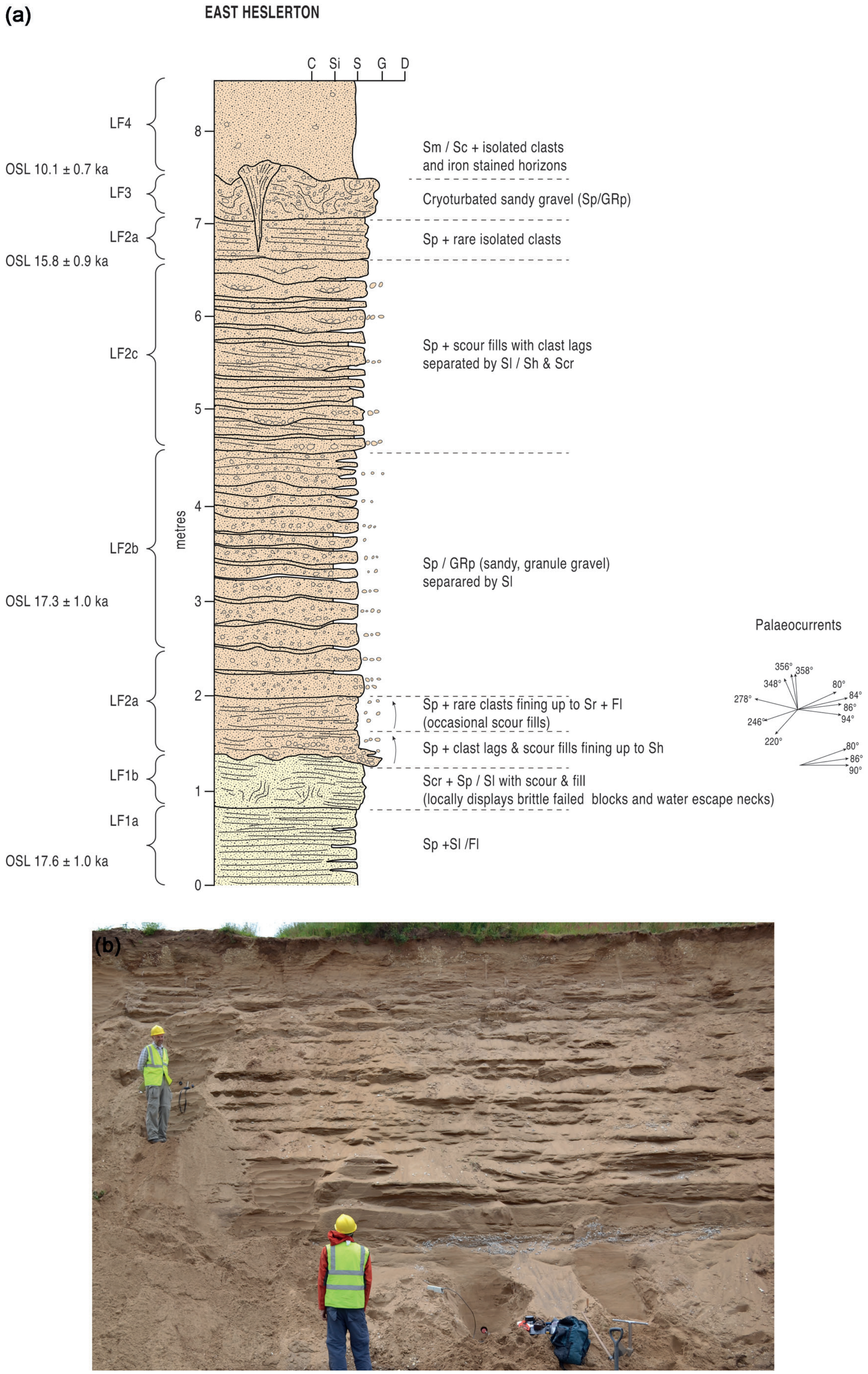

Figure 6. Details of section sampled at East Heslerton in 2013: (a) vertical profile log with OSL dates presented to the left and palaeocurrent measurements in rose plots on the right; (b) photograph of section represented in the vertical profile log, showing horizontal and tabular sedimentary architecture.

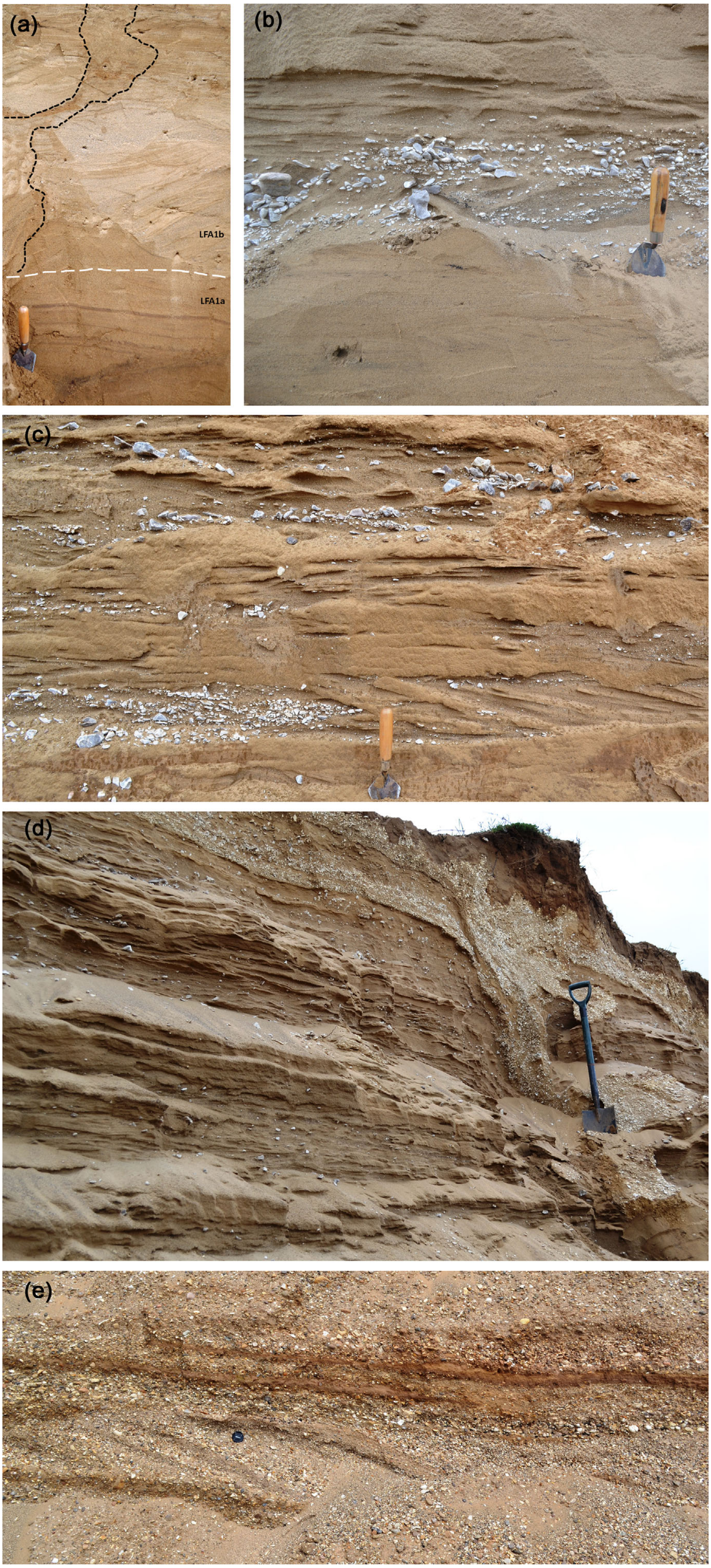

Figure 7. Photographs of the main lithofacies exposed in 2013: (a) rhythmically bedded sand and silt laminations of LF1a overlain by scour and fill features and climbing ripple drift of LF1b – areas of local disturbance of bedforms in LF1b by water escape necks and associated brittle failed blocks are highlighted; (b) planar-bedded sands with rare isolated clasts, clast lags and scour fills of LF2a; (c) planar-bedded sands and sandy granule gravels separated by thin beds of horizontally bedded to laminated sand of LF2b; (d) details of LF2c, showing planar-bedded sands with scour fills and clast lags, separated by horizontally bedded to laminated sand and climbing ripple drift and crosscut by vertical wedge infills (ice-wedge pseudomorphs); (e) undisturbed exposure through LF3 showing shallow-dipping interbeds of gravel clinoforms (small-scale foresets) and planar to horizontally bedded openwork gravel and sandy and matrix-supported gravel.

The sands and gravels that comprise LF3 at the top of the sequence have been largely post-depositionally modified by cryoturbation and ice wedge development (Foster, 1987a). Localized reworking by wind was also proposed by Foster (1987a). Some exposures in 2013 revealed that the deposits were originally interbeds of gravel clinoforms and planar to horizontally bedded openwork gravel and sandy, matrix-supported gravel and hence record a return to high discharges in a braided stream network similar to that reflected in LF2 but with gravel bedforms accumulating in the style of transverse bars (Miall, 1992).

The largest ice wedge pseudomorphs at East Heslerton penetrate up to 3 m vertically, through LF3 and 2a at the top of the sequence. Together with the cryoturbation structures, the ice wedge pseudomorphs record a phase of permafrost conditions that post-dates the deposition of the Sherburn Sands. These features are typical of many surface sands and gravels in eastern England, with excellent examples in the same stratigraphic position at Sewerby and Barmston (Evans *et al.*, 1995; Evans and Thomson, 2010) and well-developed polygons being visible on upland surfaces in the region (Dimbleby, 1952).

The dominance of angular to sub-angular clast forms within the gravels at East Heslerton indicates the mechanical breakdown (frost shattering) of cold climate conditions but also of short travel distances, the former being compatible with the development of intraformational ice wedges in LF2. They also reflect low-energy fluvial conditions and/or short travel distances and hence are not likely to have been delivered to the site after significant transport through and then along the margin of a glacier snout. This further supports the notion that the depocentre represented at East Heslerton is a scarp base fan fed by runoff from the Wolds. A fan interpretation, however, does not explain the Sherburn Sands 'shelf' located just below the altitude of the lower Glacial Lake Pickering shoreline (45 m OD), unless a series of channels through the escarpment were used by runoff to prograde sediments into the vale in a series of fans that coalesced over time and, at the last stages of sediment production, aggraded to a base level below the 45-m shoreline. If deposited at the margin of Glacial Lake Pickering, the Sherburn Sands were probably deposited in a scarp foot fan delta. The occurrence of a small outcrop of rhythmites in LF1a is probably the stratigraphic equivalent of the lake sediments previously reported from the former lake floor and hence the lake deposits appear to continue under the Sherburn Sands. Additionally, the more gravel-rich sediments of LF2 interdigitate with sands in a distal-fining architecture at East Heslerton (Foster, 1987a). The paucity of lake deposits in areas covered by the Sherburn Sands can be explained by their location within the limits of the former Vale of Pickering ice lobe, so that fan progradation only started once the ice margin began its recession eastwards. Moreover, the subaerial nature of LF2, as indicated by the fluvial bedforms and intraformational ice wedge development at East Heslerton, indicates that lake water levels had fallen to below 34 m OD by the time it was deposited. This must have taken place sometime after the 45-m shoreline phase of the Cayton–Speeton Stage (Penny and Rawson, 1969) and a lower lake stand is recorded by the Pickering delta at 30 m OD (Kendall, 1902), potentially recording a later incision level at the intake of the Kirkham Priory spillway. However, the 45-m lake stand must have impacted upon the Sherburn Sands because they extend up the Wolds to at least 60 m OD. Lake waters probably trimmed fans west of the Flamborough Moraine, East Heslerton being an example, rejuvenating them with new base levels. This enabled these fans to continue carrying sediments from the Wolds. It therefore appears that the distribution of the Sherburn Sands over an altitude range of 60–27 m OD to produce King's (1965) and Foster's (1987a) 'shelf' is the result of various lakes stands (i.e. 70, 45 and 30 m), with the East Heslerton deposits relating to sedimentation patterns in the later, lower lake stands.

OSL dating of the East Heslerton stratigraphy

Four OSL ages have been obtained from the East Heslerton stratigraphy (Table 1) and are located on Fig. 6(a). The oldest age of 17.6 ± 1.0 ka (Shfd15054) comes from the LF1a glacilacustrine deposits and therefore dates the sedimentation in Glacial Lake Pickering at the time the ice margin lay at or immediately east of the Wykeham Moraine. If the Last Glacial Maximum limit lies at the maximum western margin proposed by Edwards (1978) and Foster (1985), this date relates to initial deglaciation of the Vale of Pickering. Alternatively, if the Last Glacial Maximum limit lies at the Wykeham Moraine as proposed by Kendall (1902) the age of 17.6 ± 1.0 ka relates to the development of a full glacial lake, although we do not know the depth of water above the sample and hence it dates lake sedimentation any time after the glacier margin stood at the Wykeham Moraine.

An age of 17.3 ± 1.0 ka (Shfd15055) from LF2b records fan aggradation and permafrost conditions following the lowering of Glacial Lake Pickering, to somewhere below 34 m OD. Together with the 17.6 ± 1.0 ka age on LF1b, this date indicates a drop in lake level sometime around 17.5 ± 1.0 ka and below the 45-m stand. This drop from the 45-m lake stand indicates that a lower altitude spillway outlet than that at the Kirkham Priory gorge was exploited by the lake waters. This could simply relate to incision of the gorge down to a new altitude, controlled by the next level of relatively resistant strata, but the gorge could not have incised down to a base level lower than the altitude of Glacial Lake Humber at that time, which stood at >30 m OD (Fairburn and Bateman, 2015). Alternatively, it was controlled by the unblocking of the lower topography in the area between Flotmanby and Filey where a corridor of low terrain at 29–35 m OD dissects the Flamborough Moraine. If the drainage of lake waters did take place through this spillway to create the 30-m lake level, the age of 17.6 ± 1.0 ka dates the thinning of the North Sea Lobe and its recession eastward from the Flamborough Moraine.

The reduced discharges and sediment grades recorded by LF2c are dated to 15.8 ± 0.9 ka (Shfd15056). As this deposit contains no evidence of obvious lacustrine sedimentation or shoreline reworking despite lying below 45 m OD, it must relate to fan progradation to a lower lake stand and hence post-date the Cayton–Speeton Stage and the production of the Flamborough Moraine, at which time the 45-m lake developed. The general fining-upwards evident in LF2c to LF2a sometime after 15.8 ± 0.9 ka probably records the exhaustion of sediment and meltwater energy in the Wolds drainage channels feeding the lake marginal fans. This cessation might be coincident with the ameliorating climate associated with the Windermere Interstadial (Allerod/Bolling; GI-1 14.7–12.9 ka, Lowe *et al*, 2008). Permafrost degradation as reported elsewhere in lowland UK (e.g. Murton *et al.*, 2003) would have reduced overland flow on the Chalk Wolds and the peak discharge associated with spring nival melting would have been reduced.

An age of 10.1 ± 0.7 ka (Shfd15057) was obtained from the LF4 sand unit above the cryoturbated sandy gravels of LF3. The final phase of significant epigenetic ice wedge development occurred between 15.8 ± 0.9 and 10.1 ± 0.7 ka, indicating probably a Younger Dryas age for the periglacial features. This further supports previous reports of significant lowland permafrost during the stade (e.g. Bateman *et al.*, 2014). Sediment from LFA4, dated to 10.1 ± 0.7 ka, most likely reflects coversand reworking into the early part of the Holocene, before the Sherburn sands were stabilized by the development of vegetation in the ameliorating climate of that time. However, the youngest possible derived age of 9.4 ka, if it is the more accurate, would coincide with the establishment of birch and corylus vegetation, as documented by pollen analysis by Day (1996), indicative of vegetation stabilization. Hence further age determinations on the uppermost facies of the Sherburn Sands are warranted.

Discussion

The extensive outcrop through the Sherburn Sands at East Heslerton clearly records the aggradation of a fan from the base of the Wolds escarpment after a drop in the level of Glacial Lake Pickering sometime around 17.5 ka (Fig. 8). Although Foster (1987a) previously rejected earlier notions that the deposits were lacustrine, their occurrence at altitudes <40 m OD and therefore well below the 70- and 45-m lake stands associated, respectively, with when the BIIS was at the Wykeham and Flamborough Moraines requires explanation. Although the sedimentology presented here based upon new exposures does include evidence of lake sedimentation at ca. 17.6 ka in LF1b, we have no evidence of water depth at that time. Above LF1b, the East Heslerton deposits predominantly record subaerial fan aggradation. Because they lie below 45 m OD, they must have aggraded to a lake level no higher than that altitude and probably to the lowest Glacial Lake Pickering shoreline indicated by the Pickering delta at 30 m OD.

We now attempt to answer the questions we posed above regarding the Sherburn Sands and equivalent deposits. First, they coincide with and outcrop below the Glacial Lake Pickering shorelines (45–70 m OD) because they relate to alluvial fans receiving progressively larger volumes of debris and adjusting to falling shallow lake levels. Hence, although they are not lacustrine in nature, they are related to lake surface base level. Second, they probably do not directly record glacier-marginal recession and hence are devoid of ice-contact characteristics; their depositional shelf at a consistent altitudinal range inside and outside the Wykeham Moraine is dictated by the accommodation space afforded by the shallow lake margins and the dominant 45- and 70-m lake level altitudes at the time of sediment delivery from the Wolds. Third, they only partially represent glacial outwash, with most sediment-laden streams emerging from the Wolds probably being fed by nival melt and hence explaining their grain size, clast forms and lithologies; palaeocurrents are typical of alluvial fans that coalesced and aggraded to falling lake levels between 70 and 30 m OD. Fourth, they are not deltaic because of the lack of vertical accommodation space necessary for the development of foreset beds in a shallow lake margin subject to falling water levels; this would be compounded by localized incision and regrading of alluvial fans adjusting to falling lake levels. Interesting in this respect is the source of the abundant sand and gravel, which must reflect the activity of nival melt on the Wolds escarpment.

The range of OSL dates on the East Heslerton stratigraphy provides chronological control on lake-marginal infilling by aggrading glacial and nival-fed fans on the distal side of the Wykeham Moraine. The basal date of 17.6 ka could conceivably relate to lacustrine sedimentation (LF1a) as early as the 70-m lake stand in front of the Wykeham Moraine. However, work elsewhere has shown that the North Sea Lobe of the BIIS had extended sufficiently south to have blocked the Vale of Pickering to form Glacial Lake Pickering from ~20.5 ka onwards (Bateman *et al.*, 2015). It is more likely therefore that the East Heslerton basal age of 17.6 ± 1.0 ka reflects sedimentation within the later 45-m lake, a period not to too dissimilar to that of the highstand of Lake Humber (16.6 ± 1.2 ka; Bateman *et al.*, 2008) into which Glacial Lake Pickering flowed (Fig. 8). The fluvial rather than deltaic signature of overlying LF1b and LF2 indicates that fan construction at ca. 17.3 ka was grading to lower lake levels and hence must post-date the 45-m level of Glacial Lake Pickering. Therefore, the construction of the Flamborough Moraine, which is associated with the 45-m lake level, must have taken place before 17.3 ka (Fig. 8). Indeed, although the basal age of Shfd13054 is thought to directly date this 45-m lake, as only the upper sediments were observed, this age more likely represents the final phase of the 45-m lake. The duration of the 45-m lake, and therefore the Flamborough Moraine which impounded it, remains undated but with 17.6 ± 1.0 ka being the best minimum age at present. Further adjustment to the 30-m lake level at ~17.3 ka is coincident with the time when a regional-scale BIIS North Sea Lobe is thought to have retreated a short distance eastward (Bateman *et al.*, 2015) and would have held Lakes Pickering and Humber at similar levels while the Vale of York Lobe retreated northward (Fairburn and Bateman, 2015). With the shrinkage and demise of Lake Humber, flow through the Kirkham Gap resumed lowering Glacial Lake Pickering further (down to the 20-m OD level of the gap; Fig. 8). This in combination with silting up led to fragmentation and demise of Glacial Lake Pickering, leaving only Lake Flixton to survive into the Holocene (Palmer *et al.*, 2015).

The first direct chronology that Glacial Lake Pickering existed at least by ~17.6 ka (probably with an earlier higher lake level) until sometime before 15.6 ka provides further information for the dynamics of the BIIS North Sea Lobe. Given ice impoundment for the existence of the lake is required, the North Sea Lobe must have been established and further south of the Vale of Pickering by this time range. It must have also endured close to the present-day coastline for this period. Livingstone *et al.* (2012) showed extension of the North Sea ice lobe to the Vale of Pickering by 25 ka, establishment of Glacial Lake Pickering by 22 ka and with it persisting and blocking the Vale of Pickering until 16 ka. Clark *et al.* (2012) showed ice approaching the Vale of Pickering at 23 ka but not reaching it until 19 ka and gone by 16 ka. Given the ice had to retreat before the lake could exist at East Heslerton, the new data broadly agree with the onset times of both studies but that the North Sea lobe retreated northward later than either suggest. The latter point is borne out by new work by Bateman *et al* (2015) who propose that ice arrived in the Vale of Pickering ~20.5 ka but did not retreat northward until just before 15.1 ka. They also show two ice-marginal advances (within 20.9–17.1 and 17.1–15.1 ka) on the Holderness coastline to the south. The occurrence of a low col at around 29–35 m OD on the Flamborough Moraine between Flotmanby and Filey indicates that the change of level of Glacial Lake Pickering to 30 m at around 17.3 ka was probably initiated by the retreat of ice that was blocking the Vale of Pickering at that time (Fig. 8). Hence the lake level was maintained at ca. 30 m by the moraine dam rather than ice. Although Glacial Lake Humber stood at 33 m OD from 18 to 17 ka (Bateman *et al.*, 2015; Fairburn and

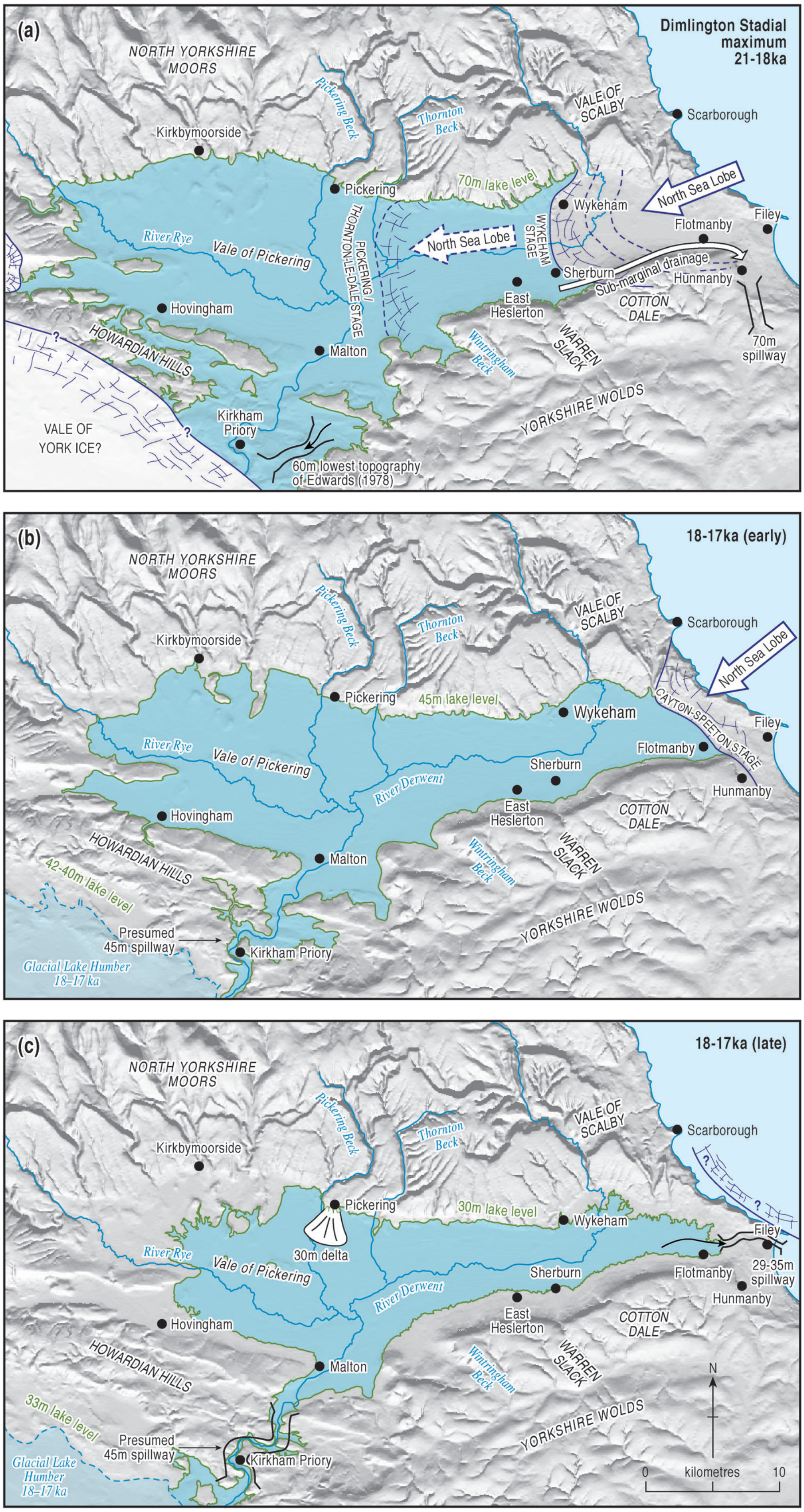

Figure 8. Palaeogeographical reconstructions of glacial events and the development of Glacial Lake Pickering during the last glaciation, incorporating the regional reconstructions of Fairburn and Bateman (2015) and Bateman *et al.* (2015): (a) the alternative potential scenarios for the Dimlington Stadial maximum, bracketed at >17.6 ka to <20.5 ka – drainage of the 70-m lake level is assumed to be sub-ice-marginal through Hunmanby Dale, which would require the North Sea lobe to be located as far east at least as the Wykeham Stage limit; (b) the 45-m OD lake level at ~17.6 ka at the Cayton–Speeton Stage; (c) the 30-m OD lake level at ~17.3 ka, assumed to be controlled by the 29–35-m OD spillway between Flotmanby and Filey and hence dating the recession of the North Sea lobe from the Flamborough Moraine. Note that at this stage it is unknown whether the 33-m level of Lake Humber and 30-m level of Lake Pickering were exactly coeval but Lake Pickering waters could not have decanted through the Kirkham Priory gorge.

Bateman, 2015), it is unknown whether there was a link between it and Glacial Lake Pickering. This is because the altitude of the Kirkham Gap (presently <20 m OD) at that time is unknown even though a newly incised altitude of 45 m OD is assumed as the control on the 45-m Lake Pickering level.

Conclusions

- The 45-m level of Glacial Lake Pickering dates to ~17.6 ka. The lake's initial impoundment to the 70-m level was probably much earlier and was related to Vale of York ice damming both the Coxwold–Gilling gap (68 m OD) and the Kirkham Priory gorge (ca. 60 m OD). The maximum extent of the North Sea lobe is unknown but the Pickering/Thornton-le-Dale limit could be related to the early stages of the 70-m lake level. At the Wykeham Stage, the lake waters probably spilled out along the Hunmanby Gap (70 m OD) but this route could not have operated if the lobe stood further west at the Pickering/Thornton-le-Dale limit. Indeed, initial North Sea lobe thinning and recession probably triggered the sub-marginal drainage.

- The Sherburn Sands and equivalent deposits were formed by multiple coalescing alluvial fans receiving progressively larger volumes of debris from the Wolds and being rejuvenated by the falling water levels of Glacial Lake Pickering.
- Fan formation ceased ~15.8 ka as climate ameliorated and permafrost waned. This enhanced the percolation into the Chalk of the Wolds and reduced the power of the nival-fed streams which had been supplying sediment to the fans.
- During the Loch Lomond Stadial, periglacial conditions resumed with extensive epigenetic ice wedge formation and coversand deposition.
- At the time when the 45-m lake level formed (~17.6 ka) the North Sea lobe of the BIIS was already in retreat, having moved eastward of the Wykham Moraine in the Vale of Pickering to the Flamborough Moraine (Cayton–Speeton Stage).
- The operation of the 29–35 m OD easterly draining spillway through the Flamborough Moraine at Filey, which most likely controlled the 17.3 ka lake level, indicates that the North Sea lobe lay east of the Flamborough Moraine by 17.3 ka.

This paper provides the first chronology for Glacial Lake Pickering during the Dimlington Stadial. In doing so its gives an initial indication of the lake's sensitivity not only to the dynamics and position of North Sea Lobe of the BIIS but also to its relationship with Glacial Lake Humber and climatic controls within its catchment. Future work is required to establish on the exact timing and position of the North Sea Lobe within the Vale of Pickering as well as the initiation and duration of the different lake levels.

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Abbreviations. BIIS, British–Irish Ice Sheet; BGS, British Geological Survey; CAM, Central Age Model; OSL, optically stimulated luminescence; SA, ultra-small multigrain aliquot; SG, single grain.

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