The Late Quaternary Palaeoenvironment of the Vale of Pickering

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Abstract

During the Quaternary, repeated glacial cycles left widespread deposits across Britain. These deposits hold an archive of terrestrial responses to changes in climate over the last 2.6 Ma. One such archive is the Vale of Pickering in North Yorkshire: A low-lying depression bounded on all sides save the east end by large hills comprised of Jurassic and Cretaceous bedrock. During the Late Quaternary, this natural basin was blocked by ice sheets forming large proglacial lakes. To understand the advance and retreat of the surrounding ice lobes in the Vale of York to the west and the North Sea Lobe to the east – the deposits of the Vale of Pickering are crucial; however, limited work in the area has failed to ascertain an accurate history of Lake Pickering. Using newly available high-resolution LiDAR data, field observations, historic borehole records and optically stimulated luminescence (OSL) dating, a new chronological model for Lake Pickering has been established. This shows that 1) previously estimated lake levels are too high and that during the LGM, Lake Pickering was no more than 33 m O.D. 2) Ice invaded the eastern coast of the Vale of Pickering on more than one occasion, potentially earlier than the LGM. 3) Several iterations of Lake Pickering exist with a lake during the LGM, but at least one older than 30 ka. 4) The drainage of Lake Pickering is very complex and seaward drainage likely prevailed until the eastern end became blocked by continued deposition of glacial material. This reversed the drainage through the Kirkham Gorge. 5) The use of newer geoscientific techniques like OSL and LiDAR mapping are crucial to the understanding of the palaeoenvironment of the Vale of Pickering and the continued development of these techniques are vital to further work.

Throughout this momentous age in the history of Yorkshire, as far as we can tell, the flaming sunsets that dyed the ice and snow with crimson were reflected in no human eyes. In those far-off times, when the sun was younger and his majesty more imposing than at the present day, we may imagine a herd of reindeer or a solitary bear standing upon some ice-covered height and staring wonderingly at the blood-red globe as it neared the horizon. The tremendous silence that brooded over the face of the land was seldom broken save by the roar of the torrents, the reverberating boom of splitting ice, or the whistling and shrieking of the wind.

The evidences in favour of this glacial period are too apparent to allow of any contradiction; but although geologists agree as to its existence, they do not find it easy to absolutely determine its date or its causes.

Home, 1905

This thesis is dedicated to Carrie Ellen Jennings.

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1. Introduction

1.1 Introduction

During the Quaternary, repeated glacial cycles left widespread deposits across Britain (Gibbard *et al.*, 2005). These deposits hold an archive of terrestrial responses to changes in climate over the last 2.6 Ma. One such archive is the Vale of Pickering in North Yorkshire: A low-lying depression bounded on all sides save the east end by large hills comprised of Jurassic and Cretaceous bedrock. During the Late Devensian period (26 ka – 10 ka) this natural basin was the site of a large proglacial lake (Kendall, 1902; Straw, 1979). This has created an environment in which sediment eroded and transported by Quaternary glacial ice has been deposited and stored. To understand the advance and retreat of the surrounding ice lobes – the Vale of York Lobe (VoYL) to the west and the North Sea Lobe (NSL) to the east – the deposits of the Vale of Pickering are crucial. However, limited work in the area has failed to ascertain an accurate and precise history of Lake Pickering and its relationship to the British Irish Ice Sheet (BIIS). Previous research though field observations (Fox-Strangways, 1880; Reed, 1901; Kendall, 1902; Melmore, 1935; Straw, 1979; Edwards, 1978; Foster, 1985; Franks, 1987) established a series of hypothesised lake levels and mechanisms for controlling these levels but have been unable to provide a robust chronology.

This thesis uses a number of methodologies to reconstruct the palaeoenvironment of the Vale of Pickering during the Late Quaternary period. This is done by using newly available high-resolution LiDAR data, shallow auguring, historic borehole records, and optically stimulated luminescence (OSL) dating to produce a new chronological model for Lake Pickering that offers new insight into the development of the lake and lake-ice sheet interactions. This chapter provides the background information on the physiography of the Vale of Pickering, how changes during the Late Quaternary period affected the landscape, and a history of previous research.

1.2 The Quaternary Period

The Quaternary period is designated as the last 2.588 million years (Pillans and Naish, 2004; Gibbard et al., 2005; Zalasiewicz et al., 2006; Cohen et al., 2013). The Early and Middle Quaternary saw a decline in global temperatures and an increase in periods of intense climate fluctuations resulting in large, land-based ice sheets. The frequency of ice sheets increased, especially after the Mid-Pleistocene Transition (MPT) at 1.2 million years ago (Lee *et al*., 2011). The MPT is marked by a shift from a dominant obliquity (41 ka) cycle to eccentricity (100 ka) cycle that resulted in a pronounced increase in the frequency of glacial cycles and a decrease in sea surface temperature (Clark *et al.*, 2006). However, despite the number of cycles, there is not much landform evidence for early Quaternary glaciations due to erosion from more recent glaciations.

Once important source of evidence for the paleoclimate of the Quaternary comes from oxygen isotope data. Ratios between two types of oxygen isotope (the 'light' $\delta\ ^{16}O$ with eight neutrons and the 'heavy' $\delta\ ^{18}O$ containing ten neutrons) indicate paleoclimate temperature. Cold glacial stages are marked by an increase in $\delta\ ^{18}O$-rich sea water due to the heavier isotope condensing more easily in precipitation events before reaching higher latitudes (Alley, 2000). During colder periods, more of the precipitated lighter isotope becomes locked in glacial ice at higher latitudes. Once climate warming occurs, high concentrations of $\delta\ ^{16}O$ is released as glacial meltwater enters the ocean circulation. Records of these isotopic ratios are found in the calcium carbonate ($CaCO_2$) or silicon dioxide (SiO_2) shells and skeletons of tiny marine animals, corals, and plankton that use $\delta\ ^{18}O$ to build their shells, showing higher concentrations of $\delta\ ^{18}O$ during colder periods than during warmer ones (Alley, 2000; Lisiecki and Raymo, 2005).

These fluctuations in $\delta^{18}O$ and $\delta^{16}O$ ratios are divided into cold and warm stages called Marine Isotope Stages (MIS). Odd numbered stages (MIS1 being the present) represent periods of warmer climate whereas even number stages (MIS2, which is the LGM) represent cooler, glacial periods (Liesiecki and Raymo, 2005).

Offshore stratigraphic records show glaciation by the Scottish Ice Sheet (SIS) at 1.1 ma (MIS 34) (Sejrup et al., 1987), but very little terrestrial evidence of these glaciations is left. The oldest pre-Anglican Quaternary sediments in Northern England are found on the County Durham coast. Clay fills found in some of the subvertical karstic fissures in the Zechstein Group (Mid to Late Permian limestones and dolostones) contain pyritised flora and fauna known as the 'Castle Eden Flora', most of which are no longer endemic to the United Kingdom or are now extinct. This includes the *Mammuthus meridionalis,* a species of elephant found in Britain up until the Cromerian Interglacial (MIS 22 – 13: 866 ka – 478 ka) (Figure 1.2.1) as well as Scandinavian erratics (Stone *et al*., 2010). Lee *et al*. (2004a) suggest that the first major lowland glaciation of Great Britain was during MIS 16. The main evidence comes from the stratigraphic relationship of the Happisburg Formation to the underlying magnetised early Mid-Pleistocene organic sediment layer (MIS 19) and the unconformably overlying Lowestoft Formation (MIS 12). These observations are supported by the application of a climate forcing age model to terrace aggradation of the Bytham River in East Anglia (Lee *et al*., 2004a). However, Amino Acid Racemisation (AAR) dating and biostratigraphy favour a MIS 12 age for the deposits (Preece *et al*., 2009; Rose, 2009).

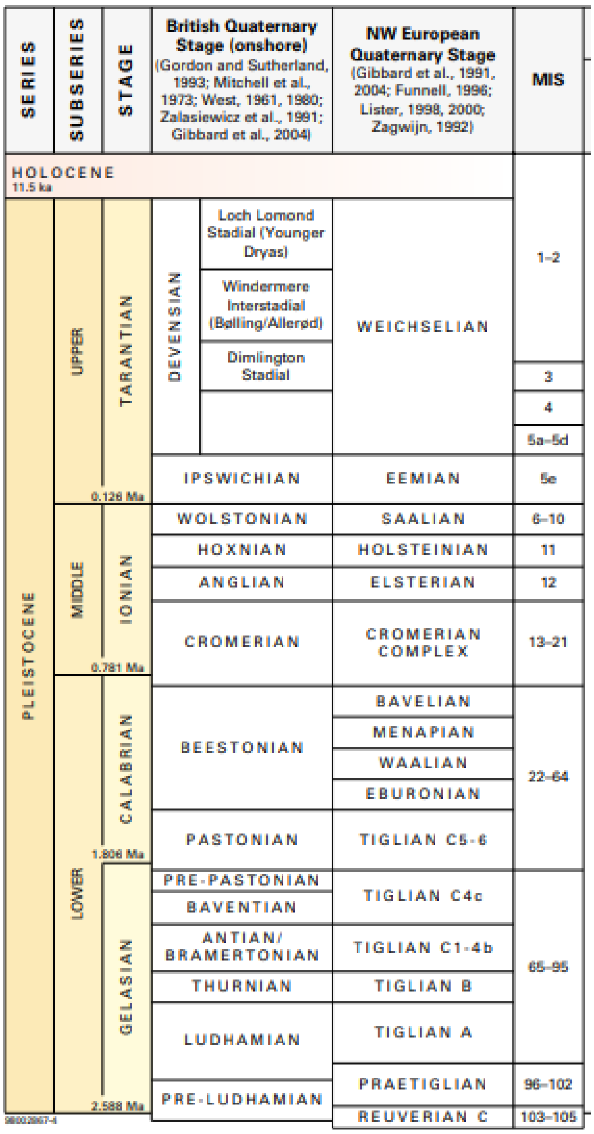


Figure 1.2.1: The UK Quaternary period with European equivalents and associated MIS stages (from Stoker et al, 2011)

In the UK, the glacial periods are divided into the Anglican (MIS 12/478 ka), the Wolstonian (MIS 10 – 6/~300 – 200 ka) and the Devensian (MIS 2/29 ka) (Figure 1.2.1). The Anglican glaciation is thought to be the most extensive Quaternary glaciation in the UK (Ehlers and Gibbard, 1991) reaching as far as southern England and the Isles of Scilly (Figure 1.2.2). Evidence for the Anglican glacial stage is well represented in North Sea seismic records that show a widespread unconformity interpreted as a glacial erosion surface (Lee *et al*., 2011) Deposits of this age include the oldest known deposit in Northern England: The Thornsgill Till Formation, a weathered diamicton found in the Lake District (Stone *et al*., 2010) as well in the Midlands and southeast England, but most especially in East Anglia (e.g.: Bowen *et al*., 1986; Erhlers and Gibbard, 1991; Gibbard *et al*., 1992; Bowen 1999; Gibbard and Clark, 2004; Lee *et al*., 2004b; Hamblin *et al*., 2005 and Rose, 2009).

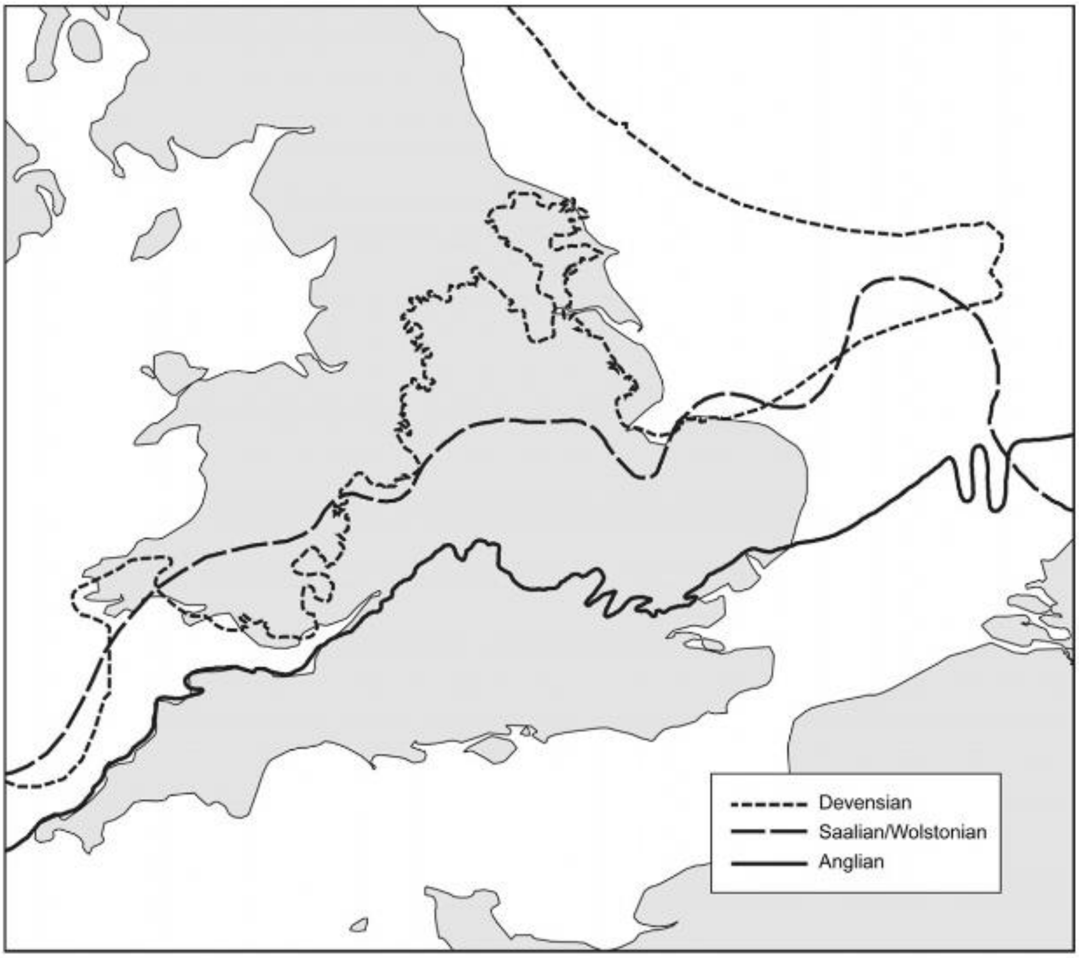


Figure 1.2.2: Limits of the Anglican, Wolstonian and Devensian glaciation. The solid line represents the extent of the Anglican Glaciation; The long-dashed line represents the Wolstonian glaciation, and the short-dashed line represents the Devensian ice limit (Gibbard and Clark, 2011)

Following the Anglican stage is the Hoxnian Interglacial (MIS 11: 424 ka – 374 ka) (Figure 1.21), which is considered to be the best analogue to the current Holocene (Candy, 2009) as climate was controlled by similar orbital forcing as the present day (Loutre and Berger, 2003).

This is followed by the controversial Wolstonian glaciation (MIS 10 – 6: 352 ka – 130 ka) (West and Donner, 1956; Gibbard and Turner 1988; 1990; Gibbard and Clark, 2004; Cohen and Gibbard, 2016) was less extensive than the Anglican, reaching as far as the English Midlands (Figure 1.2). However, investigation into the Wolstonian-age tills such as the Oadby Formation, Sheringham Cliffs Formation, and Britons Lane Formation (Lee *et al*., 2011) reveal that these tills are interpreted by some (Perrin *et al*., 1979; Bowen *et al*., 1986; Banham *et al*., 2001) as being Anglican-aged (MIS 12) while others (Rowe, et al., 1997; Scourse, *et al*., 1999; Rose, 2004; Hamblin *et al*., 2005; Rose, 2009) favour a Wolstonian age.

Following is the Ipswichian interglacial (MIS 5e: 124 ka – 115 ka) and the last significant warm period prior to the present day (Figure 1.2.2). Sea level in the Ipswichian interglacial is thought to be 6 – 9 m higher than current sea levels (Dutton and Lambeck, 2012). Unlike the Hoxnian, the orbital configuration was not the same as the current Holocene period showing that orbital forcing is not the only factor in the changing climates of the Quaternary (Yin and Berger, 2012).

Temperatures declined once more at the end of the Ipswichian marking the beginning of the Devensian (MIS 5d – 2: 117 ka – 10 ka) with a slight increase of temperature during MIS 3 (60 ka – 35 ka). The Main Late Devensian (MLD) occurred between 28 ka and 14.4 ka (Figure 1.3.1). During this time, the BIIS reached its maximum extent, the Last Glacial Maximum (LGM), between 27 ka – 22 ka (Stone *et al*., 2010) (Figure 1.3.1). Post-LGM warming saw the retreat of the BIIS and the transition into the current Holocene (10 ka to present day) began with the Late Glacial-Interglacial Transition (LGIT) between 13 ka - 10 ka (Walker, 1995).

1.3 The Last British-Irish Ice Sheet

The last ice sheet to cover Great Britain is known as the BIIS (Figure 1.3.1). The BIIS has recently been extensively studied under the BRITICE and BRITICE CHRONO projects resulting in an improved understanding of the extent of glaciation and pattern and pace of retreat (e.g. Clark *et al*., 2004; 2012; 2017). Clark *et al*. (2012) estimated the BIIS to have been *c*. 840,000 km^2 at its maximum extent, and covered Scotland and Ireland completely, despite up to two thirds of the ice sheet being marine based (Figure 1.31). Clark *et al*. (2012) argue that today's West Antarctic Ice Sheet (WAIS) is comparable to the conditions of the BIIS during the LGM, which has made the BRITICE project paramount in understanding palaeo-ice sheet modelling. The BIIS of the LGM is currently one of the most extensively studied palaeo-ice sheets in the world.

The build-up of the BIIS is assigned to after 35-32 ka although it did not reach maximum extent or volume simultaneously (Chiverell and Thomas, 2010) (Figure 1.31). Between 27 – 21 ka, the BIIS reached its maximum extent covering Scotland, Ireland, most of Wales, and Northern England (Figure 1.31). The western side of the BIIS was more extensive than the eastern side, reaching as far south as the Isles of Scilly at 26 ka (Smedley *et al*., 2017) earlier than originally estimated by Chiverell *et al.* (2013). Retreat of the western side of ice sheet at 23 ka (Smedley *et al*., 2017) thought to be in response to a climatic forcing known as Heinrich Event (Heinrich, 1987). This was followed by a recession of the Irish Sea Ice Stream (ISIS) that coincided with the advance of the eastern side of the BIIS (Smedley *et al*., 2017; Bateman *et al*., 2017; Roberts *et al*., 2018) (Figure 1.3.1).

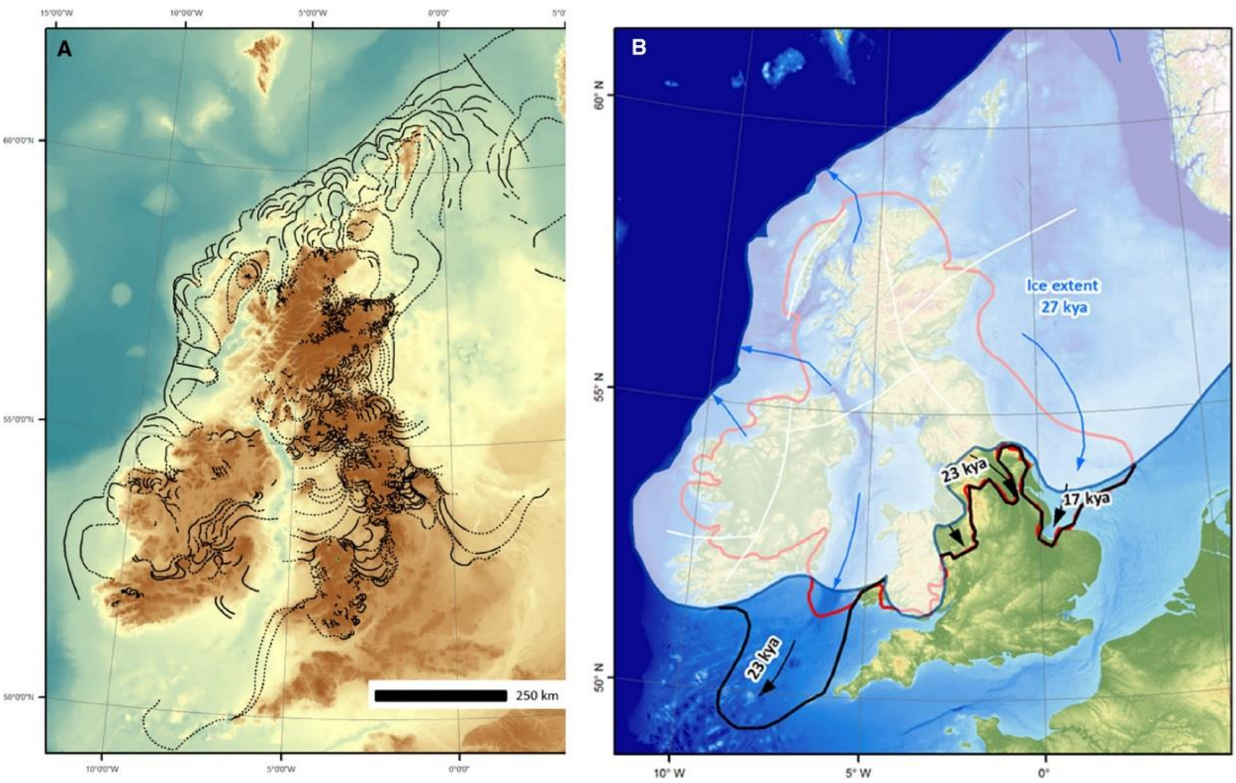


Figure 1.3.1: The extent of the British-Irish Ice Sheet as modelled by the BRITICE project (Clark et al, 2017). The western side of the British Isles and Ireland were glaciated more extensively reaching as far south as the Isles of Scilly at around 24 ka. The eastern side saw an early advance at 27 ka. Ice retreated and then readvanced at 23 ka with a small readvance at 19 ka. Eventually rapid deglaciation of the complete BIIS occurred around 17 ka.

The limits of Late Devensian glaciation (MIS 2) in North East England have been reconstructed by combining data from stratigraphy, geomorphic mapping, and field observations (e.g.: Kendall and Wroot, 1924; Penny and Rawson, 1969; Gaunt, 1974; Straw, 1979; Catt, 2007;

Evans and Thompson, 2010; Clark *et al.*, 2012; Livingstone *et al.*, 2012; Roberts *et al.*, 2013; Bateman *et al.*, 2018) (Figure 1.31). These limits are essential in understanding the regional dynamics and extent of the British-Irish Ice Sheet (BIIS) within north-eastern England. In North Yorkshire, the Vale of Pickering was surrounded by two ice lobes that occupied the low-lying land of Yorkshire and Lincolnshire (Figure 1.3.2): one extended down the west side from the Tees Valley, passing through the Vale of Mowbray, and terminating in the Vale of York (Gaunt, 1976; Straw, 1979; Bateman *et al.*, 2013; Fairburn and Bateman, 2015). The other, extending from Scotland, down the east coast of England into the North Sea and into the North Norfolk (Chiverell and Thomas, 2010; Evans and Thompson, 2010; Clark *et al*., 2012; Busfield *et al*., 2015) (Figure 1.32). These lobes are named the Vale of York Lobe (VoYL) and the North Sea Lobe (NSL), respectively (Figure 1.3.2). The ingress of ice into North Yorkshire by these two ice lobes culminated in the development of large proglacial lakes, most notably Lake Pickering and Lake Humber (Kendall, 1902; Straw, 1979; Murton and Murton, 2012).

Till deposits on the north-eastern coast of England and within the North Sea Basin are thought to have been the result of several Devensian glacial advances into the North Sea Basin. The ice of the NSL originated from East Scotland (Clark *et al*., 2012; Roberts *et al*., 2013) and flowed southwards down the east coast of England, extending as far as the current north Norfolk coast (Clark *et al*., 2012; Dove *et al*., 2017). Erratics from the tills on the North Yorkshire coast (Busfield *et al*., 2015) suggest that the NSL was also sourced from the Lake District via the Stainmore Gap (Livingstone *et al*., 2012). Early in the journey southwards, the North Sea lobe merged with the Fennoscandian Ice Sheet (FIS) within the Central and Northern North Sea Basin (Carr *et al*., 2006; Sejrup *et al*., 2016). The exact configuration and timing of this merger remains somewhat inconclusive, although the two had most likely separated by 18.5 ka (Sejrup *et al*., 2016) creating an extensive proglacial lake in the central North Sea Basin (Hjelstuen *et al*., 2017). The NSL repeatedly oscillated against the east coast of England and, as a result, the majority of the till typesites for the Devensian glaciation (e.g.: Skipsea, Withernsea, Hessle) are located on the East Lincolnshire coast (Evans and Thompson, 2010) in the region of Holderness (Figure 1.3.2).

The tills of north east England have been the focus of study since the 19th century when Wood and Rome (1868), Lamplaugh (1879; 1891), and Reid (1885) first described glacial sediments along the North Yorkshire and East Lincolnshire coast. In the late 1930s, Bisat (1932; 1939;

1940) first developed a systematic nomenclature for the till sequences at Holderness (Table 1.3.1), but this was reworked in the 1970s and the current nomenclature is based on and developed from the work of Madgett and Catt (1978) (Table 1.3.1). The till sequences are highly complex and have been suggested by Eyles *et al.* (1994) to represent a surging NSL although this is refuted by Evans and Thompson (2010) who instead suggest subglacial deformation and onshore accretion of subglacial materials by the NSL for the complexity of the till stratigraphy. This was supported by geochemical analysis of tills by Boston *et al.* (2010), who identified a number of overlapping ice oscillating impinging on Holderness.

At the base of the till sequence is the Basement Till, which is overlain by the Skipsea (formerly 'Drab') and Withernsea (formerly 'Purple') tills (Table 1.3.1). Beneath the Skipsea till lie a suite of glacial silts, sands, and gravels known as the Dimilington Silts that make up the typesite for the Late Devensian in Britain (Rose, 1985; Bateman *et al*., 2011). Penny *et al.* (1969) dated

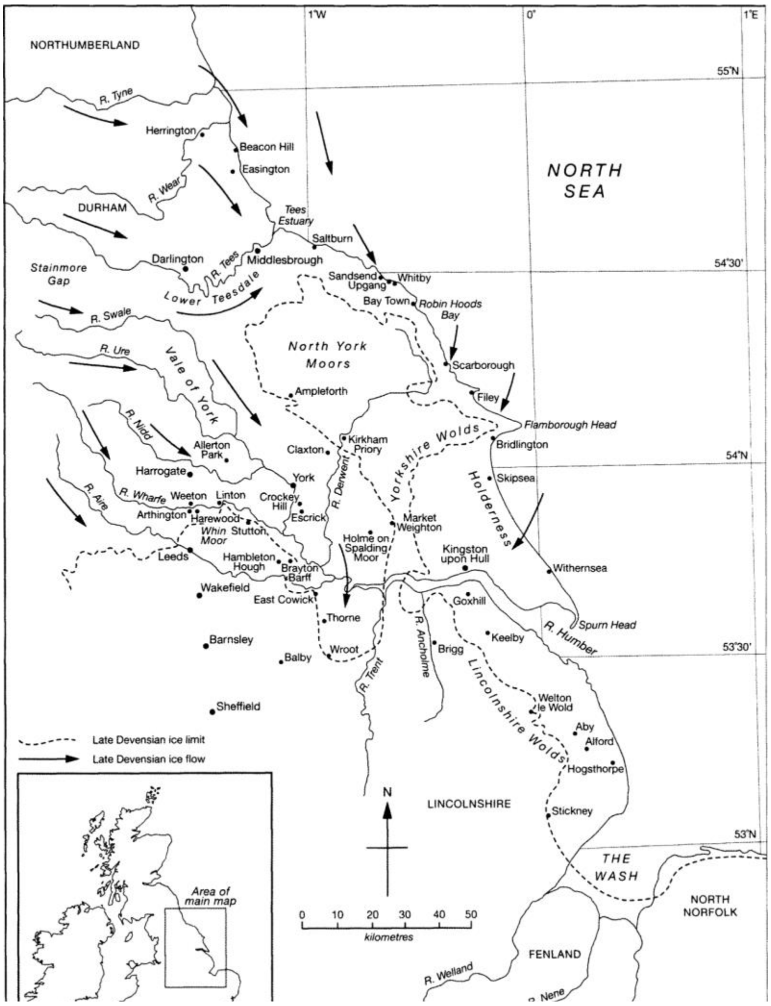


Figure 1.3.2: Devensian Ice limits in North Yorkshire with inferred ice flow direction (Catt, 2007). Note the ice limit in the Vale of York extends beyond Escrick to Wroot, although this limit is controversial. See text for details.

moss samples contained within the silts using radiocarbon dating, obtaining dates of 23.3-21.2 cal. 14C ka BP (18.5 ± 4.0 14C years) and 22.4-21.3 cal. 14C ka BP (18.2 ± 2.5 14C years). From these dates, Clark *et al*. (2012) postulate that ice reached the Holderness area around 22 ka. Catt and Penny (1966) propose that the Basement Till predates the Ipswichian Interglacial, but this has been contested by Eyles *et al*. (1994) and Evans and Thompson (2010), who believe it could belong to an earlier advance of Late Devensian glacial ice; this appears to be supported by recent correlations between marine-based and terrestrial glacial deposits in the central sector of the North Sea (Dove *et al*., 2017).

Table 1.3.1: Comparison of Nomenclature for Coastal Exposures at Filey Bay and Holderness, after Edwards (1987) and Catt and Madgett (1987)

<table>
<tr><th>Lamplugh (1879)
Filey Bay</th><th>Bisat (1940)
Filey Bay</th><th>Catt and Penny (1966)
Holderness</th><th>Madgett and Catt (1978)

Holderness</th><th>Edwards (1987)

Filey Bay</th></tr>
<tr><td rowspan="2">Hessle Till</td><td>Sandy Till

Upper Purple Till</td><td rowspan="2">Hessle Till</td><td rowspan="3">Withernsea Till</td><td rowspan="2">Upper Till</td></tr>
<tr><td>Lower Purple Till</td></tr>
<tr><td>Brown Till

Gravel</td><td>Sand, Silt, and Gravel</td><td>Purple Till</td><td>Gravel</td></tr>
<tr><td>Greenish-Purple Till</td><td>Grey and Stony Drab Tills</td><td>Drab Till

Dimlington Silts</td><td>Skipsea Till

Dimilington Silts</td><td>Lower Till

Chalk Rubble</td></tr>
<tr><td>Basement Till</td><td></td><td>Basement Till</td><td>Basement Till</td><td>Basement Till</td></tr>
<tr><td>Chalk Rubble</td><td></td><td></td><td></td><td>Speeton Shell Bed</td></tr>
</table>

At the eastern end of the Vale of Pickering (Figure 1.3.4), the till complexes at Filey have been investigated by various workers for over a hundred years (Lamplaugh, 1881, Penny and Catt, 1967; Madgett and Catt, 1978; Edwards, 1978; 1981; 1987). The thickest deposits lie between Flamborough Head and Filey Brigg (Edwards, 1978). Between Filey Brigg and Cayton Bay, the till overlies a steep limestone scarp before thinning-out towards Scarborough. Edwards (1987) describes the correlation between the tills of Filey Bay and the tills of Holderness as laterally consistent but divides the tills into Upper and Lower Till Series (Table 1.3.1) due to what he describes as 'greater variation' within the till sections when compared to the tills of Holderness. He argues that the gentle dip of the underlying chalk bedrock of the Holderness area allowed for a more consistent till deposition. Conversely, north of Flamborough Head, at the eastern end of the Vale of Pickering, the bedrock consists of soft Mesozoic clay with the North Yorkshire coast providing many topographic barriers preventing ice advance, which caused more incorporation of the underlying diamicton to be worked and stacked into later ice deposits (Edwards, 1978, 1987).

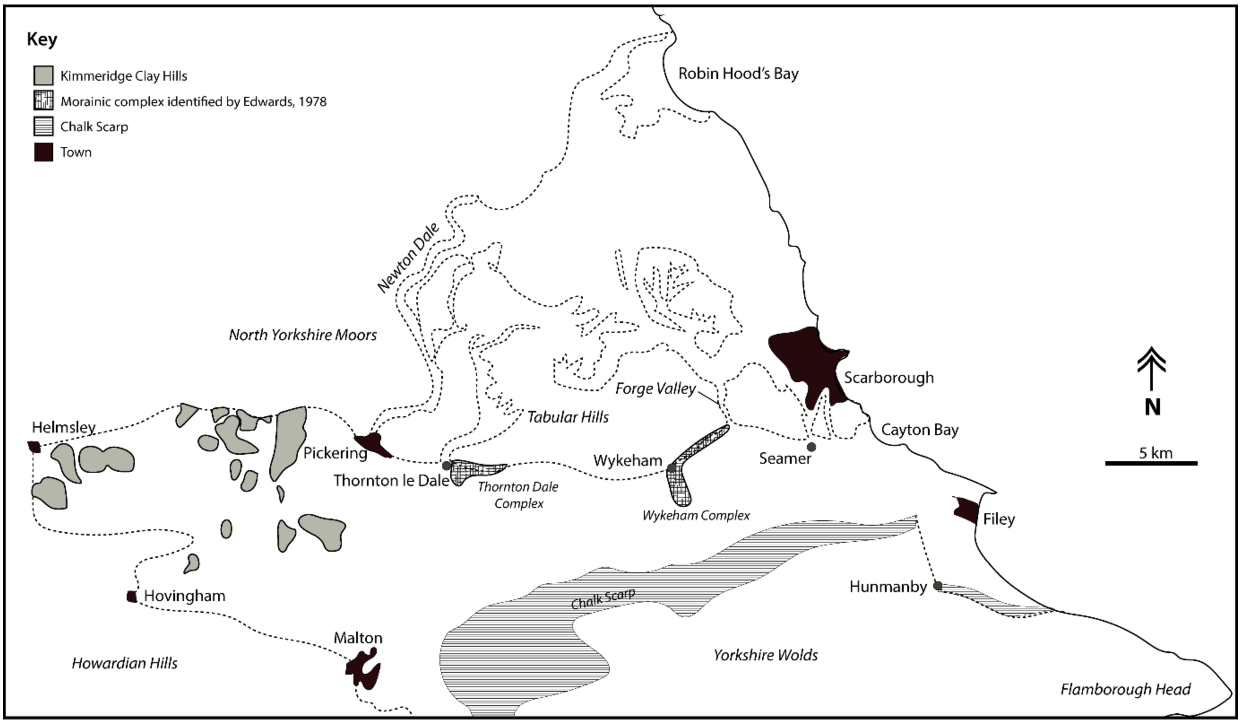


Figure 1.3.3: Features and locations in the Vale of Pickering redrawn from Edwards, 1978

The limit of North Sea ice advance into the Vale of Pickering has been reported to be at Wykeham (Kendall, 1902; Straw, 1979 Evans *et al*., 2017) (Figure 1.3.3). Here, a long kame terrace (Kendall, 1902) known as the Wykeham moraine extends southwestwards across the northern portion of the Vale with the southern half of the moraine speculated to have been eroded by meltwater following the opening of the Forge Valley and redirection of the River

Derwent through the Kirkham Gorge into the Vale of York (Foster, 1985). Straw (1979, 2016) argues for two phases of ice advance at the eastern end of the Vale of Pickering, however, no definitive mapping or interpretation of the moraines and associated glacial features has been completed. Environmental reconstruction from the Late Glacial-Interglacial Transition (LGIT) sediments at Wykeham Quarry, Post-Glacial Lake Flixton, and the early Mesolithic archaeological site of Star Carr (Palmer *et al*., 2015; Lincoln, 2017) provide some evidence of recessional moraines potentially buried beneath the lake deposits of Post-Glacial Lake Flixton.

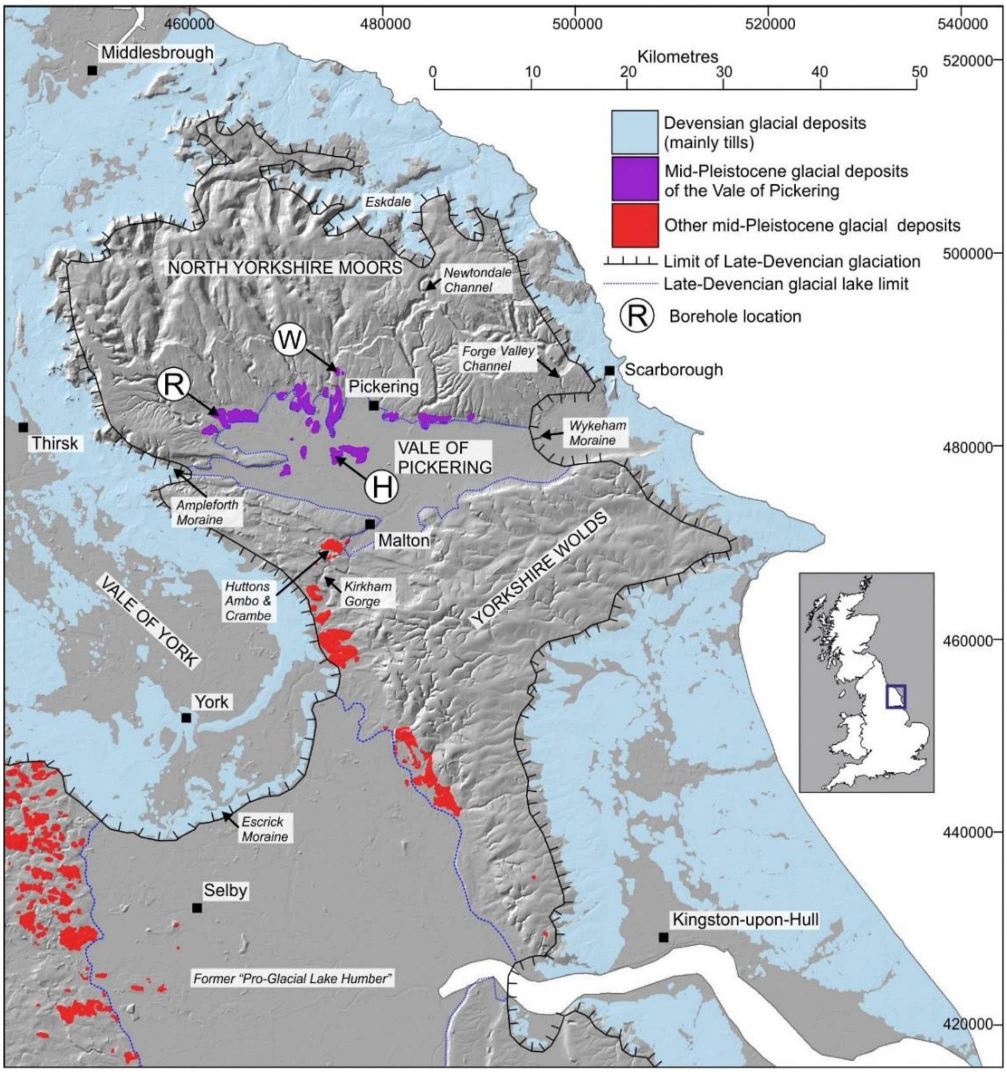


Figure 1.3.4: The Vale of Pickering and Vale of York with associated glacial limits and deposits (Powell et al., 2016). Note the position of the ice limit at the Escrick moraine. The limits for ice within the Northeast Yorkshire and Humber region. Maximum ice limits show that the VoYL and the NSL almost coalesced. Due to the blockage of the main drainage route by ice in the Humber Gap, large proglacial lakes formed in the Vale of Pickering and southern vale of York known as Lake Pickering and Lake Humber, respectively.

To the west of the Vale of Pickering, Pennine ice extended from Stainmore, Swaledale, Wensleydale and Teesdale into the Vale of York, leaving moraines at York and Escrick (Catt, 2007; Livingstone *et al*., 2010) (Figure 1.3.4). Ice from the VoYL blocked the Vale of Pickering at the western end of the Coxwold-Gilling Gap, effectively sealing off the basin on the western side and contributing meltwater to Lake Pickering. This western moraine is known as the Ampeforth moraine (Figure 1.3.3; Figure 1.3.4) (Fox-Strangways, 1892; Kendall, 1902; Harrison, 1935) and is comprised of a large upthrown block of Kimmeridge Clay Formation (Fox-Strangways, 1892). Harrison (1935), however, found little evidence to support this interpretation and believed that ice had not passed into the western end of the Coxwold-Gilling Gap or into the Vale of Pickering. Conversely, Foster (1985) believed the VofYL entered the Vale of Pickering as far as Wath, near Hovingham. The chronological history of the VoYL is somewhat poorly understood and there are three major uncertainties: the extent of the VoYL, its interaction with proglacial Lake Humber, and the resulting chronology (Bateman *et al*., 2017). Work by Bateman *et al*. (2015, 2017) suggest that the NSL and the VoYL almost coalesced with south flowing ice from the NSL intruding into the Humber Gap as far as Winteringham (Figure 1.3.7), which led to the development of a large proglacial Lake Humber. Ice from the NSL is postulated to have blocked the Humber Gap between 22-21 ka (Figure 1.3.5), possibly as part of a readvance of the NSL in the area (Bateman *et al*., 2018). Ice in the Vale of York also restricted water outflow through the Kirkham Gorge from the Vale of Pickering likely contributing to a higher lake level.

Loess under till at Ferrybridge (Bateman *et al*., 2008) suggests that the VoYL advanced into the region around 23.3 ± 1.5 ka with the formation of the York and Escrick Moraines around 22.2 ± 0.5 ka (Murton *et al*., 2009). However, the maximum extent of the VoYL during the LGM has been rather controversial with some researchers suggesting that the lobe extended as far as Wroot in the southern part of the Vale of York (Figure 1.3.5), which has been named by Friend *et al*. (2016) as the Lindholme Advance. Friend suggests this deposit was formed by part of an extended ice shelf advancing into proglacial Lake Humber (Gaunt, 1992; Gaunt, 1994; Friend, 2011; Friend *et al*. 2016; Bateman *et al*., 2017). Evidence for this potential southerly advance comes from gravel deposits that contain large amounts of magnesium limestone found across the Vale of York and into North Lincolnshire (Bateman *et al*., 2015). These are interpreted as ice marginal sediments (Gaunt *et al*., 1992; Bateman *et al*., 2015; 2017). Straw (2002) argues there is little evidence to suggest that the Vale of York lobe extended southwards of Escrick during the Late Devensian (Figure 1.3.5). He argues that the deposits at Wroot, known as the Thorne-Wroot gravels (Friend, 2011), are residual landforms formed during an earlier Devensian glaciation. He argues that the southern limit at Wroot (Figure 1.3.4) is speculative as the Isle of Axeholme shows no glacial bed footprint, a lack of evidence for river diversions around the proposed lobe, and there is an absence of any Late Devensian till or deglacial landforms in the area. Others have argued that the VoYL did not pass the Escrick moraine during the LGM (Kendall and Wroot, 1924; Melmore, 1935; Ford *et al*., 2008; Murton *et al*., 2009; Straw, 2016; Murton, 2018) (Figure 1.3.4). Despite OSL dates produced by Bateman *et al.* (2015) averaging 18.7 ka. Straw (2016) argues that the OSL ages obtained by Bateman *et al.* (2015) are ambiguous as there is no new field evidence to show ice marginal deposits in the area.

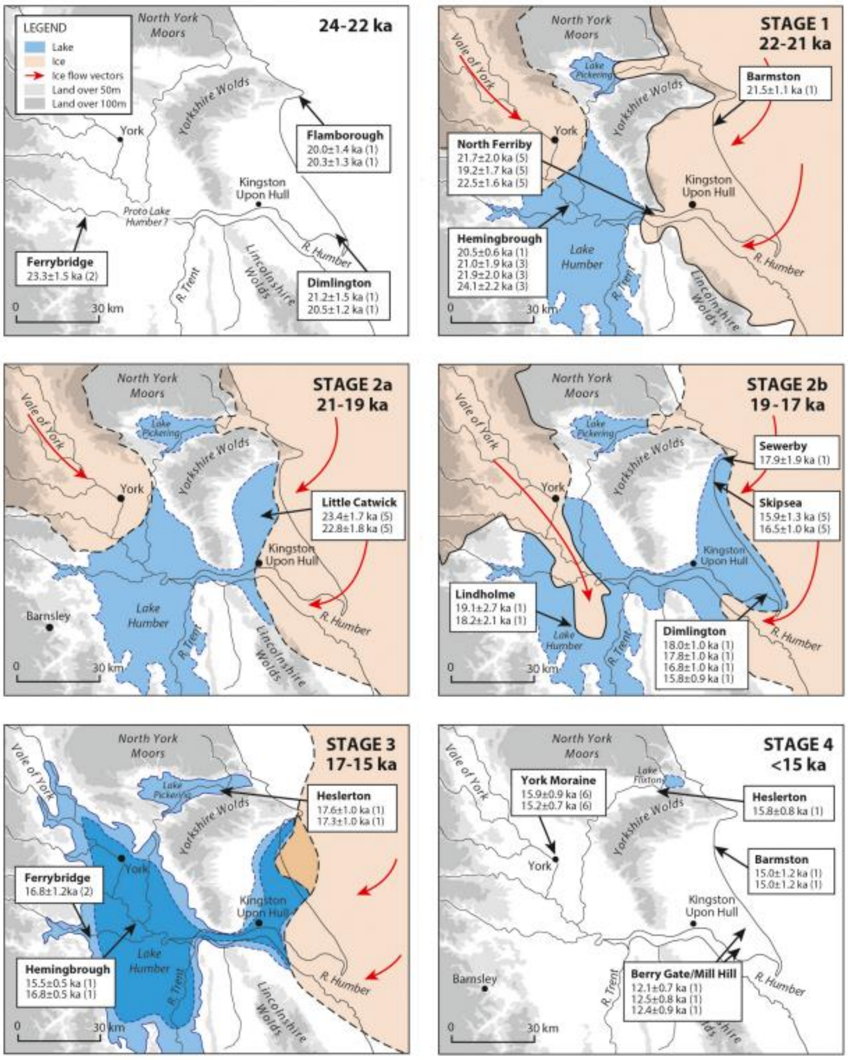


Figure 1.3.5: Chronology of the North Sea Lobe, Vale of York Lobe and stages of Lake Humber. (Bateman et al., 2018)

Undeniably, the regional glacial history is complex and not without controversy with often conflicting interpretations for the mechanisms of sediment deposition (e.g: Eyles *et al*, 1994; Evans and Thompson, 2010), or the extent of ice incursion into an area (e.g.: Straw, 2017; Bateman *et al.*, 2018; Murton, 2018). These disputes indicate the intricacy of the Late Devensian stratigraphy in the area and only through more investigation can these debates become settled.

1.4 Proglacial Lakes

Proglacial lakes are masses of glacial meltwater that has been impounded by sediment, bedrock, or ice. They are well represented throughout the Quaternary and, since lakes are natural sediment traps, proglacial lakes can provide excellent palaeoenvironmental records and landscape evolution. There are many well-documented and researched large palaeo-proglacial lakes throughout the world. For example, work by Yang and Teller (2004) recorded ongoing changes in lake bathymetry as a result of isostatic rebound using changes in paleoshorelines from Lake Agassiz (Teller and Thorliefson, 1983).

Glacial lakes are important in Quaternary studies as their stability and behaviour are linked to ice sheet energy balances that include changes in meltwater and sediment fluxes demonstrated in recent work by Thorndycraft *et al*. (2018) in Patagonia. Furthermore, proglacial lakes also return a stabilizing or moderation effect to nearby ice margins and can help slow summer ice ablation (Carrivick and Tweed, 2013). The changes in these conditions are recorded in the sedimentary archive of the lakes, which are often used to reconstruct proxy records from palaeoenvironmental conditions (e.g. Palmer *et al*., 2015) to glacier mass balance (Larsen *et al*., 2011). By understanding these records of change, comparison to current variations in climate, especially in areas affected by rapid deglaciation, could improve climate modelling and forecasting.

1.5 Proglacial lakes of the Last British Irish Ice Sheet

Along the margin of the BIIS, proglacial lakes formed and fluctuated with the oscillations of the BIIS (e.g. Clark *et al*. 2004; Murton and Murton, 2012; Livingstone *et al*. 2012) (Figure 1.5.1). Evidence of proglacial lakes from the BIIS is scant and aside from Murton and Murton (2012) no complete inventory of British and Irish proglacial lakes exist. This is partly to due to the wide range of lake sizes and durations, ranging from small, ephemeral lakelets lasting a few years to inland seas lasting thousands of years, but also from the lack of lacustrine deposits in some areas (e.g.: Lakes Eskdale, Lapworth, and Fenland) (Murton and Murton, 2012).

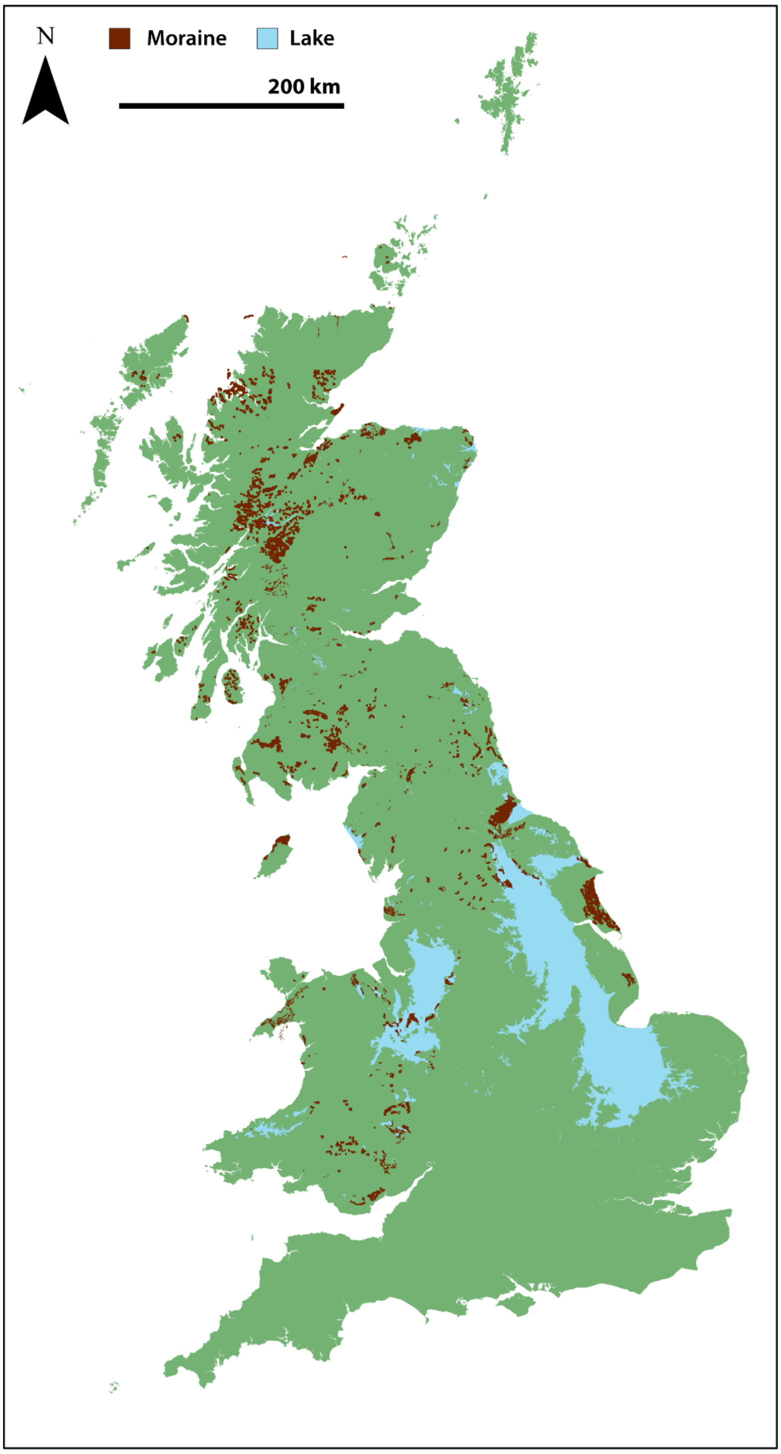


Figure 1.5.1: Proglacial lakes of the last BIIS. Moraines in brown. Data from BRITICE (2018)

One of the largest proglacial lakes to form was Glacial Lake Humber caused by the blockage of the Humber Gap (Lewis, 1894; Melmore, 1940; Gaunt, 1976; Straw, 1979; Murton et al., 2009; Murton and Murton, 2011; Fairburn and Bateman, 2015, Straw, 2017; Bateman *et al*, 2017, Murton, 2018; Evans *et al*., 2018). Currently, ~25% of the drainage of Eastern England exits through the Humber Gap (Versey, 1940; Rees, 2006) and at its maximum extent Proglacial Lake Humber (Figure 1.5.2) covered an area of approximately 4500 km^2 (Clark *et al*., 2004; Bateman *et al*., 2008) encompassing the Humberhead Levels, the Vales of Trent and York, and northwards towards the Vale of Pickering (Fairburn and Bateman, 2016).

Gaunt (1976) proposed two phases of Lake Humber: a high stage at 33 m O.D and a low stage at between 9 and 12 m O.D. Straw (1979) argues for the two-stage model, but with a high stage earlier in the Devensian and a low stage at the LGM, with the lake completely emptying between phases. Bateman *et al*. (2008) dated the higher stage lake using OSL to 16.6 ± 1.2 ka from a shoreline deposit at Ferrybridge postulating that ice occupied the Humber Gap during this late stage of the LGM and contrary to the findings of Murton *et al*. (2009) who report a date of 22.2 ± 0.5. Subsequently, Fairburn and Bateman (2016) added an eight-stage recessional model to the chronology of Lake Humber reporting a pre-LGM high-stage lake level of 52 m O.D. and arguing for an LGM high-stage elevation of 42 m with recessional stands at 40 m, 33 m, 30 m, 20-25 m, 15 m, 10 m, and 5 m. Recent work by Murton (2018) proposes that frequent ice marginal oscillating oscillations of the NSL ice into the Humber Gap led to frequent cyclic drainage of Lake Humber, which is recorded in the sedimentary sequence, thereby creating a sedimentary archive of NSL oscillations in the lower basin of the Humber. Murton (2018) also challenges Fairburn and Bateman (2016) denudation chronology citing misidentification of colluvial deposits sourced by periglacial weathering at Shiptonthorpe as lake shorelines and the coincidence of Fairburn and Bateman (2016) geomorphological mapping and bedrock features such as the Jurassic-Cretaceous unconformity near Market Weighton. Straw (2016) and Murton (2018) both argue against the model proposed by Fairburn and Bateman (2016), the former arguing the high-stage lake was early Devensian and the latter arguing Lake Humber did not exceed 10 m O.D. during the LGM.

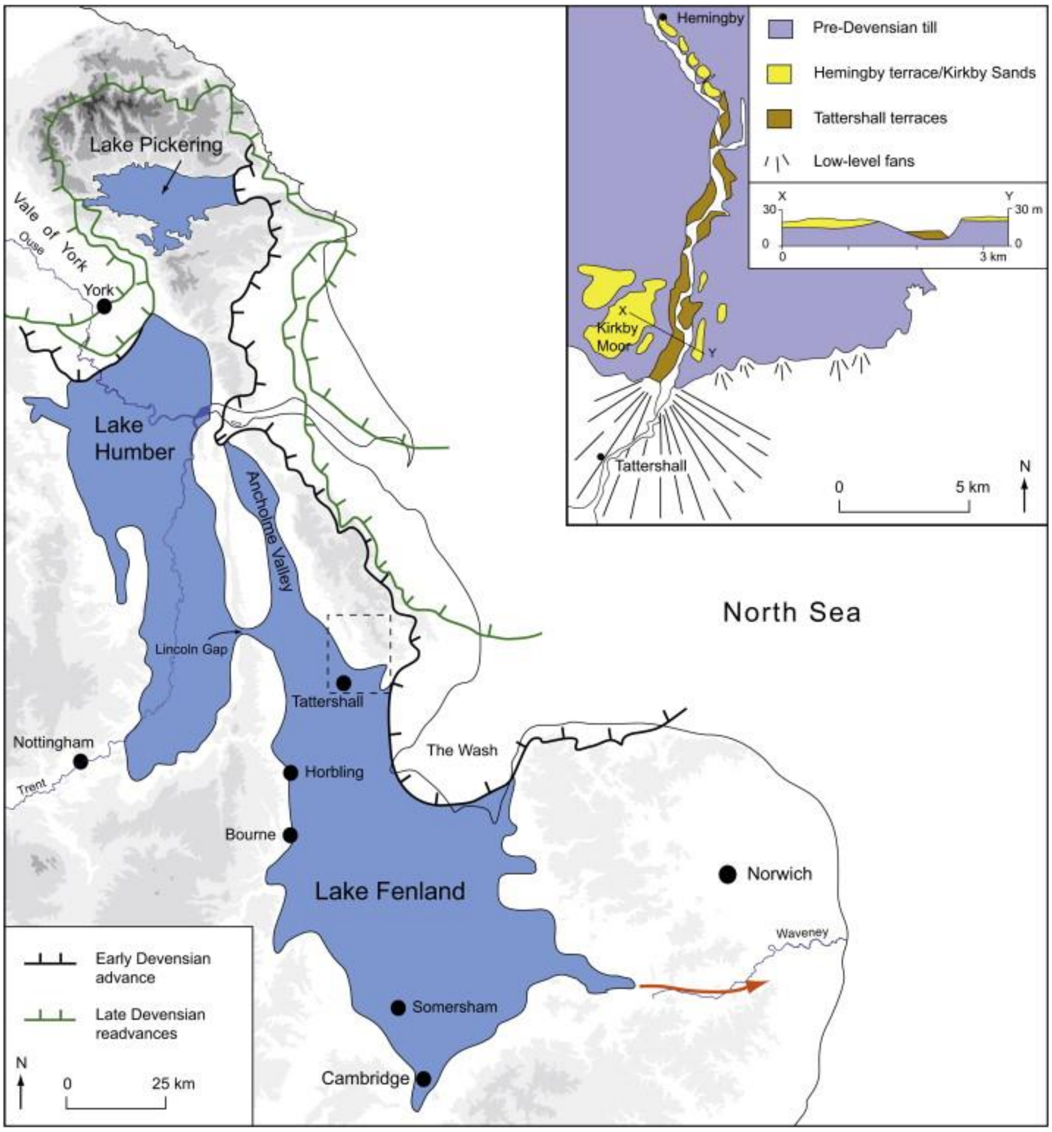


Figure 1.5.2: Extent of Lake Humber in the Vale of York (from Murton and Murton, 2012)

Grey laminated clays cover an extensive area in the Vale of York, reaching a thickness of up to 20 m (Catt, 2007). These clays extend north of the Escrick moraine, suggesting that Lake Humber was in existence at the time of VoYL retreat sometime between 19 and 17 ka (Catt, 2007). Gaunt (1970, 1974) reports lacustrine clay of the 8-9 m Lake Humber overlying the southern flank of the Escrick Moraine. A maximum age for high-stage Lake Humber is suggested by Gaunt (1974) from a 14C date of 24,194 – 28,154 cal. yrs BP from a bone fragment found at the base of strandline gravels. Work by Bateman *et al*. (2008) shows this to be a pre-Lake Humber age and that dating the timing of the blockage of the Humber Gap by the NSL on the Holderness Coast is a more accurate way to date Lake Humber (Figure 1.5.2).

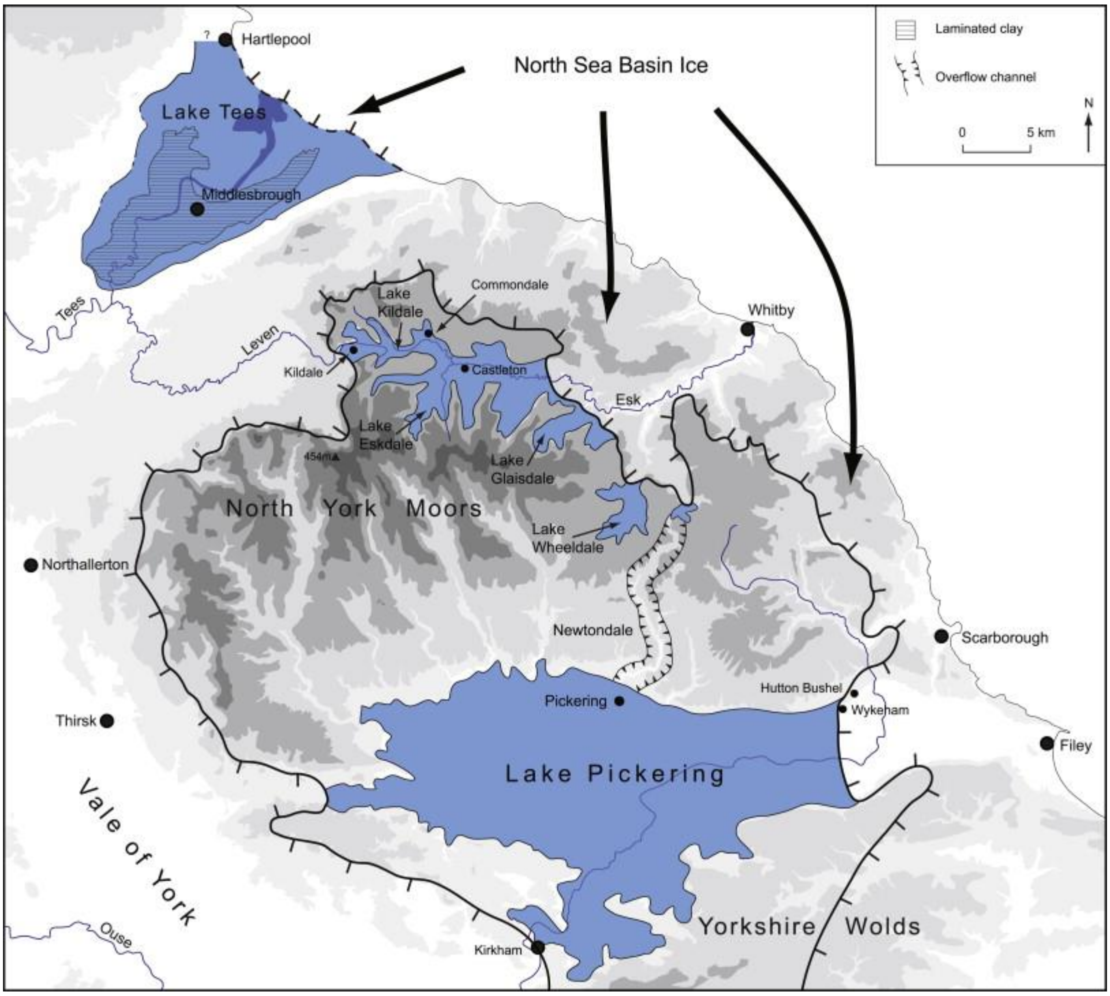


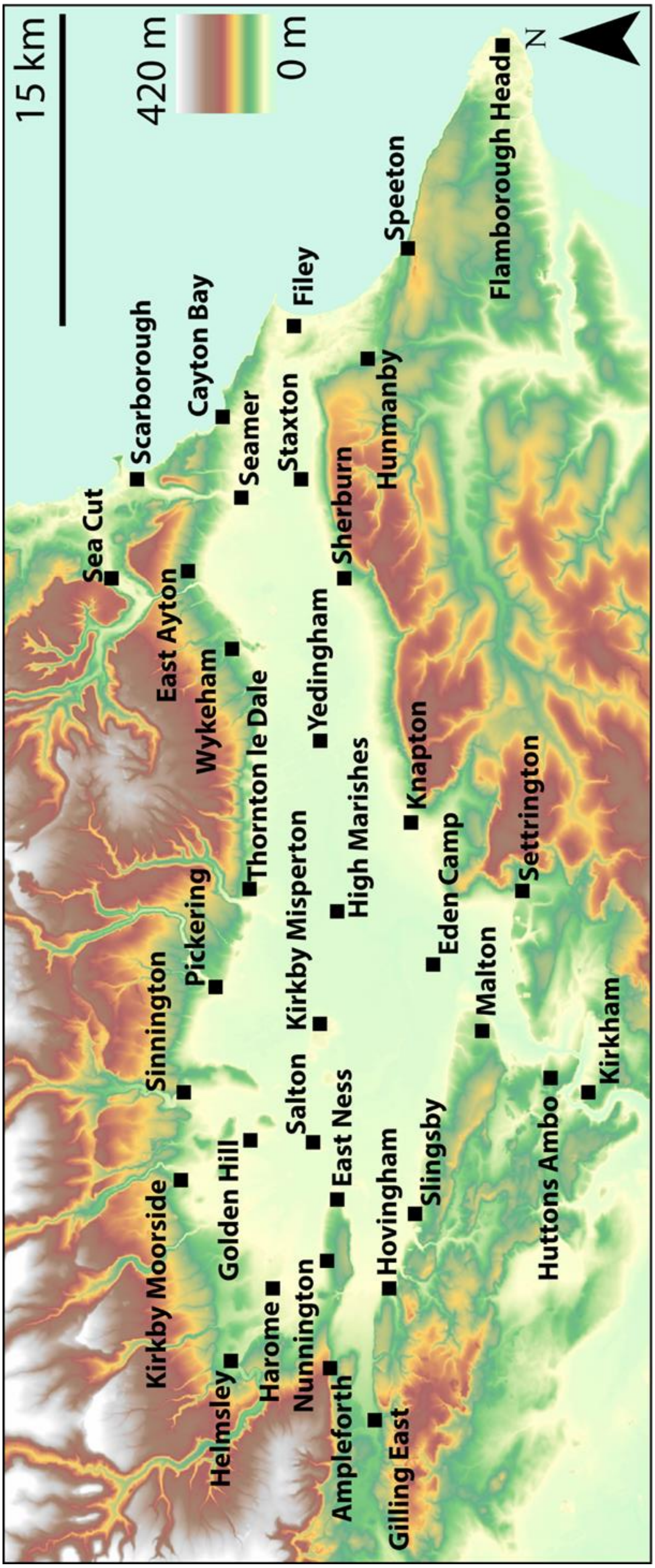
Figure 1.5.3: Extent of Lake Pickering (after Kendall, 1902) and the relationship with the lakes of the North Yorkshire Moors (Murton and Murton, 2012)

North of Lake Humber was Glacial Lake Pickering (Figure 1.5.3), an isolated basin surrounded by limestone and chalk hills, save at the eastern end. The Jurassic clay bedrock is draped by fluvial and lacustrine superficial sediment, except towards the western end of the basin where the Quaternary sediments become thin enough to expose the underlying clays as small hills (See Physiography below). Repeated glaciations have periodically created a barrier at the coast, impounding the eastward drainage of the River Rye, Riccal, and Derwent with moraines and glacial outwash sediment. For this reason, the Vale of Pickering is important because it likely holds a longer record of east coast glaciations due to its unique morphology as a flooded river valley (See Physiography below). With each ice lobe blockage, the drainage of the rivers was impounded, and the basin filled creating a large proglacial ice-dammed lake. Once the ice had retreated, the morainic debris slowed drainage eastwards, leaving the basin cover with slow drying, shallow sandy lakelets. During the Last Glaciation, the remaining moraines and glacial outwash debris, along with a slight rebound from glacial isostatic adjustment (GIA), was large enough to alter the slope of the landscape and redirected lake waters westwards through a col at Kirkham Gorge. Lake levels have been hypothesised to have reached a maximum of 102 m O.D (Harrison and Tweed, 1936) by examining contour heights between the Vale of Pickering and the Vale of York. Kendall, (1902) suggested that the height of Lake Pickering reached 69 m O.D based on the similar height of the top of the kame terrace at Hutton Buscel and the altitude of the col near Ampleforth in the Coxwold-Gilling Gap (Figure 1.5.3). In between these two points, Kendall envisioned a large proglacial lake fed from water draining out of proglacial lakes on the North Yorkshire Moors, Lakes Eskdale, Wheeldale and Glasidale (Figure 1.53). Smaller lake levels of 45-42 m (Kendall, 1902; Straw, 1979), 30-33 m (Kendall, 1902) and 25 m (Clark, 1954) are also often cited

It was Kendall's paper on the glacial lakes of the Cleveland Hills that established criteria for recognising former lakes: 1) evidence of deltaic deposits, 2) beaches (shorelines), 3) floor deposits, and 4) overflow channels. Kendall was able to recognise all but beach deposits in the Vale of Pickering (Kendall, 1902), but the criteria is still used.

1.6 Physiography of the Vale of Pickering

The Vale of Pickering is a 500 km^2 large, moderately-flat, east-west trending enclosed basin located in North Yorkshire, North East England (Figure 1.5.3). The south west margin is marked by the Howardian Hills, a double cuesta of Middle Jurassic limestone and Corallian limestone that is frequently offset by numerous east-west trending faults that reveal occasional exposures of Jurassic-aged Oxford, Ampthill, and Kimmeridge clays (Figure 1.61). A downthrown fault extends along the base of the Howardian Hills from Coxwold to the eastern end at Malton (Figure 1.6.1). This fault makes up the south wall of the Asenby-Coxwold Graben that runs south-eastwards into the Vale of Pickering. On the north wall of the graben is the Kilburn fault (Figure 1.6.1), which is responsible for a 1.5 km east-west trending limestone spur known as Caulkley's Bank (Powell *et al.*, 1992; Wright and Cox, 2001; Powell, 2017) (Figure 1.6.1).



15 km
420 m
0 m
N
Sea Cut
Scarborough
Cayton Bay
Seamer
Filey
Staxton
Speeton
Flamborough Head
Hunmanby
Sherburn
East Ayton
Wykeham
Yedingham
Thornton le Dale
Knapton
High Marishes
Kirkby Misperton
Pickering
Sinnington
Eden Camp
Settrington
Malton
Kirkham
Huttons Ambo
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East Ness
Golden Hill
Kirkby Moorside
Slingsby
Hovingham
Nunnington
Harome
Helmsley
Ampleforth
Gilling East

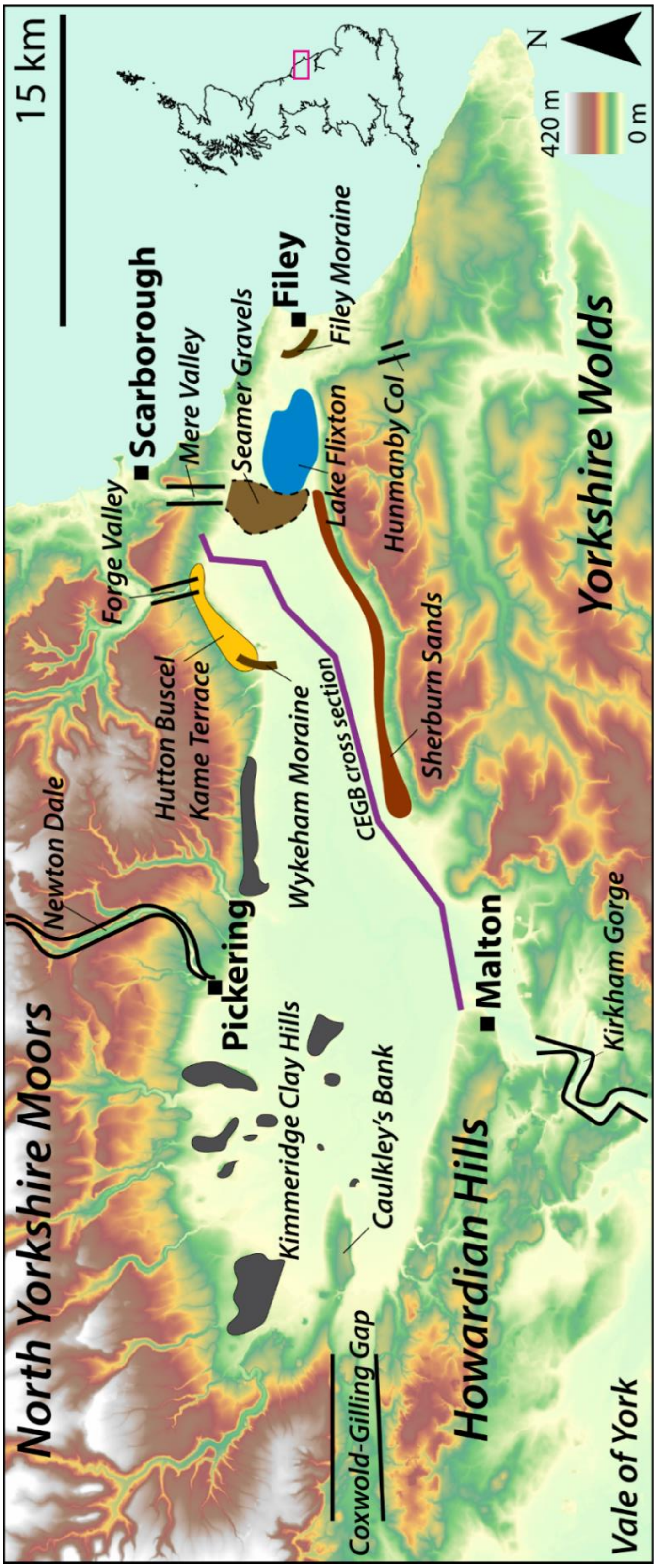


Figure 1.6.1: (Above) Location of places in the Vale of Pickering Mention in the text. (Below) The Vale of Pickering showing the location of the inlets (Forge Valley, Newton Dale, and Mere Valley) and outlet points (Kirkham Gorge, Coxwold-Gilling Gap, Hunmanby Col) as well as the established geomorphic features: Caulkley's Bank (a Jurassic limestone spur); The Wykeham and Filey Moraines; The Hutton Buscel Kame Terrace; The Seamer Gravels; Lake Flixton Basin and the Jurassic-Aged Kimmeridge Clay Hills.

The southern edge of the Vale is lined by the Cretaceous chalk scarp of the North Yorkshire Wolds. This is the northern most remnant of chalk in England and lies unconformably over the Jurassic Clays (Figure 1.6.2). Beneath the North Yorkshire Wolds lies a series of faults known as the Howardian-Flamborough Fault Bed (Powell, 2010). Reactivation of this fault group resulted in a crush of the chalk scarp, presenting a noticeable fault ridge running from Hunmanby to the Flamborourgh Head scarp (Figure 1.6.2). At the coast, the Speeton Clay (Rawson and Wright, 2000) outcrops from beneath Quaternary deposits and underlies much of the eastern Vale of Pickering (Figure 1.6.2).

Bedrock Geology of the Vale of Pickering

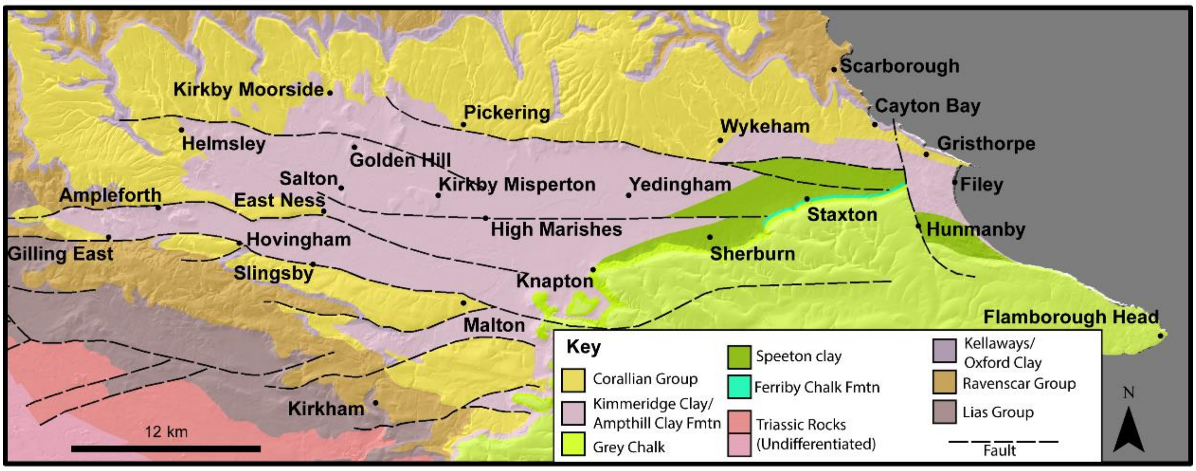


Figure 1.6.2: The bedrock geology and major faults of the Vale of Pickering. (Created using BGS 625k data)

The northern margin is dominated by the Tabular Hills of the North Yorkshire Moors that mark the boundary between the Middle and Upper Jurassic limestones and calcareous gritstone and sandstones. The sequence of Jurassic and Cretaceous rocks of the Howardian Hills and North Yorkshire Wolds had once extended over to the North Yorkshire moors, but due to tectonic activity during the Tertiary (Alpine), the rocks were inverted, faulted, and folded resulting in a reactivation of the Vale of Pickering faults (Hemmingway and Riddler, 1982; Kirkby and Swallow, 1987; Powell, 2017). Subsequently, the Jurassic and Cretaceous rocks were downthrown, resulting in a southwards dip, and were subsequently eroded to expose the lower Jurassic clays forming the basin of Vale of Pickering. Continued erosion by a pre-Quaternary proto-River Ure that flowed through the Coxwold-Gilling Gap to Filey (Figure 1.6.3), created a deep river valley now infilled with Quaternary sediments (Reed, 1901; Versey, 1948) currently with an altitude between 21 m O.D. and 24 m O.D.

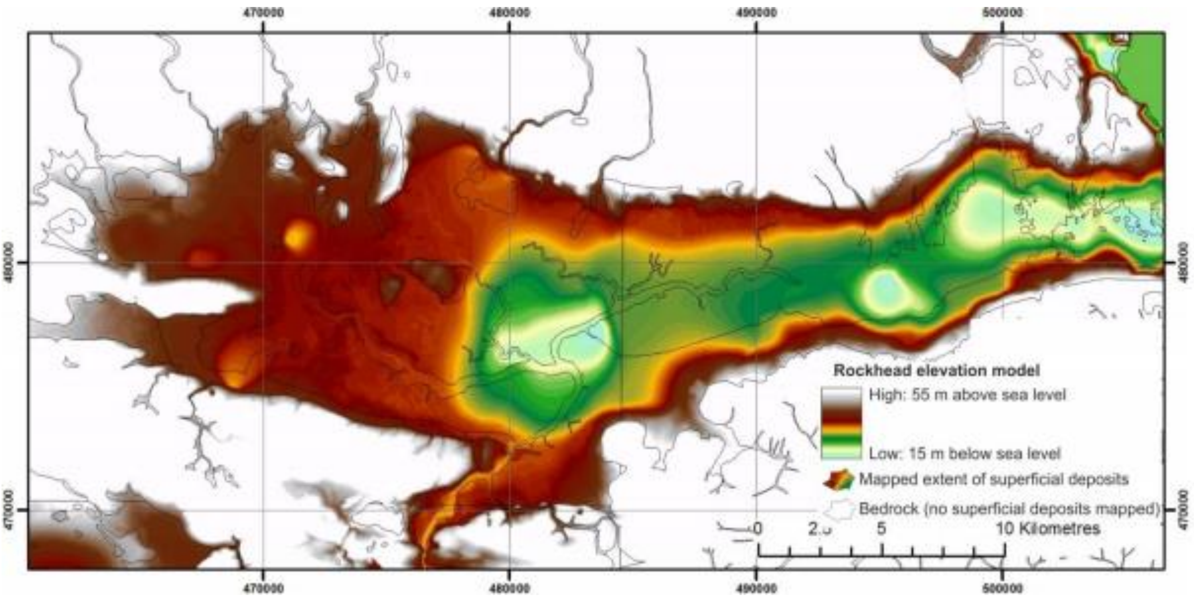


Figure 1.6.3: Rockhead Elevation Map from Ford et al. (2015). The change in deposit thickness is clearly different between the west and eastern halves of the Vale of Pickering. The buried valley is the green section running from the middle of the page to the east side. It must also be noted that robust data from boreholes is greater for the eastern side of the Vale of Pickering and this map reflects that.

The underlying clays, the Upper Jurassic-aged Kimmeridge Clay and Ampthill Clay formations, are up to 297 m thick (Powell, 2017) and appear to dip south-eastwards. Shown in a depth to rockhead elevation model by Ford *et al*. (2015), data from historic boreholes model a palaeovalley depression with bedrock start heights reaching up to 15 m below the overburden of Quaternary deposits in the central and eastern Vale of Pickering (Figure 1.6.3). This is referred to as a buried valley by Edwards (1978) and Foster (1985). In the west, the bedrock sits higher, and the Quaternary deposits are shallow enough to allow lumps of upthrown Kimmeridge clay to break the surface (e.g at: Golden Hill; Kirkby Misperton) in a series of small, flat-topped hills (Table 1.6.1; Figure 1.6.2). The Jurassic clay also crops out along the northern margin of the North Yorkshire Moors in two places: from Helmsley to Sinnington culminating with a 1 km south trending spur known as Riseborough Bank, and again further eastwards starting from a 0.5 km southwards trending spur at Thornton le Dale and ending near Snainton. Below the Jurassic Clay lies organic-rich mudstones and Carboniferous mudrocks, which are potential sources of shale gas (Smedley *et al*., 2016; Newell *et al*., 2016; Powell, 2017) and fracking work is currently being carried out at Kirkby Misperton in the central Vale of Pickering (Figure 1.6.2).

The eastern part of the Vale of Pickering is noticeably more recently glaciated, with large till deposits exposed along the coast from Scarborough to Flamborough Head particularly between

Filey and Bempton (Fox-Strangways, 1880; Edwards, 1978; Evans and Thompson, 2010). Till deposits between Filey and Scarborough are thinner due to a southerly dipping Jurassic limestone scarp (Gristhorpe Cliffs) that creates the tip of Filey Brigg and runs north-westwards towards Cayton Bay (Table 1.6.1). Between Seamer and Filey there is evidence of kame and kettle topography, although much of it buried by the sediments of Post-Glacial Lake Flixton (Palmer *et al*., 2015; Table 1.6.1). An extensive gravel deposit called the Seamer Gravels (Figure 1.6.1) runs south-westwards from the Mere Valley into the Vale of Pickering. It merges with another gravel deposit that exits the mouth of the Forge Valley consisting of the reworked Hutton Buscel sands and gravels (Franks, 1987; Lincoln, 2017). The culmination of coastal moraines and alluvial deposits (Table 1.6.1) create a rise in the altitude of the ground with the moraine at Filey reaching 40 m O.D. before descending to sea level at Filey Bay.

At Wykeham a large curved feature extends for 5 km from the scarp of the North Yorkshire Moors in a southwest trending arc (Table 1.6.1; Figure 1.6.1). It is thought to be the north half of a terminal moraine (Straw, 1979) possibly as young as the LGM (Kendall, 1902; Edwards, 1978; Foster, 1985). Lining the north scarp between the NYM and the Wykeham moraine is a long, stepped kame terrace that runs from Wykeham to East Ayton (Kendall, 1902). The Hutton Buscel kame terrace (Table 1.6.1) is comprised of outwash sands and gravels overlying kame and kettle topography (Franks, 1987; Eddey *et al*., 2017; Lincoln, 2017). Also, at East Ayton is the mouth of the Forge Valley where the River Derwent now runs southwards in to the Vale of Pickering (Figure 1.6.1).

Along the base of the chalk scarp opposite the Wykeham moraine is a long stretch of thick late glacial aged (17 ka) lacustrine sand interspersed with or overlain by LGIT-aged (15 ka – 11 ka) intermittent chalk gravel fans (Evans *et al*. 2017). The sand deposit is known as the Sherburn Sands (King, 1965; Foster, 1985; Table 1.6.1) and extends westwards along the base of the chalk scarp from Staxton towards Knapton (Figure 1.6.2) and is referred to by Foster (1985) as a southern outwash train.

Much of the central portion of the Vale of Pickering is underlain by deep lacustrine clays proved in numerous BGS boreholes and in the field logs of researchers like Fox-Strangways (1881; 1904) and Kendall (1902). A long borehole section from Malton to Scarborough (Figure 1.6.5) was recorded by the Central Electric Generating Board (CEGB)

Table 1.6.1: List of geomorphic features in the Vale of Pickering (modified from Eddey *et al.*, 2017)

Landform	**Classification**	**Interpretation**
Kimmeridge clay hills	Isolated hills in the centre of the western Vale, comprised of Kimmeridge clay and till	Pre-Late Devensian tills representing expansion of a plateau icefield in the Tabular Hills (Powell et al. 2016), or landforms derived from fault-based uplift of Kimmeridge clay (Edwards, 1978).
Thornton-le-Dale till	Undulating topography consisting of a diamicton, containing various lithologies on the northern side of the vale.	Formed either via an initial Late-Devensian North Sea Ice Lobe advance (Edwards, 1987), or by glacial activity prior to the Late Devensian (see Kimmeridge clay hills)
Ampleforth Moraine	Elevated ridge composed of diamicton, crossing the Coxwold-Gilling Gap at the western terminus of the Vale of Pickering	Ice terminus of the Vale of York Ice Lobe (Kendall 1902)
Hutton Buscel terrace	Flat topped, horizontal-gently graded landform (60-70m OD), perched upon the valley side in the eastern Vale of Pickering	Kame terrace formed during the 'Wykeham Stage' when ice stood at the Wykeham, used to support a 70 mOD lake in the valley (Kendall 1902; Straw 1979)
Hunmanby Gap	Deeply incised valley, through the watershed of the Yorkshire Wolds scarp, in the eastern Vale of Pickering.	Submarginal spillway, draining a 70m OD Glacial Lake Pickering (Evans *et al*. 2017)
Subglacial, Marginal/Submarginal meltwater channels	Channel forms running outside contemporary drainage, obliquely to the valley side, or as misfit features in contemporary systems	Meltwater channels formed underneath, or at the margin of a glacier (Kendall 1902; Edwards 1978; Straw 1979; Foster 1985; Franks 1987)
Hutton Buscel sands anc gravels	Irregular collection of mounds and ridges, with frequent flat based depressions in the eastern Vale	Either kettle-kame topography from the Late Devensian recession of the North Sea ice lobe (Franks 1987), or deltaic outwash from the Forge Valley (Fraser *et al*. 2009).
Thorn Park sands and gravels	Narrow ridge of sand and gravel, oriented parallel to the a-axis of the Seacut Valley	Glaciolacustrine and glacifluvial sediments, eroded by subglacial meltwater
Kirkham Gorge	Deeply incised, sinuous valley, incised through the Howardian Hills watershed in the southwestern Vale	Spillway for the Glacial Lake Pickering, during the Late Devensian (Kendall 1902; Evans et al. 2016), Or a subglacial meltwater channel formed prior to the Late Devensian (Edwards 1978; Franks 1987)
Mere Valley	Deep and wide valley, incised from the VoP, to the Scarborough Bay.	Pre-existing valley utilised as a meltwater spillway through the Late Devensian (Penny and Rawson 1969; Straw 1979).

Forge Valley	Deeply incised valley through the Corallian Dipslope watershed in the eastern Vale of Pickering	Spillway for a Late Devensian Glacial Lake Hackness into the eastern Vale (Kendall 1902; Penny and Rawson 1969), or a pre-existing valley, utilised as a meltwater conduit through the Corallian Dipslope during the Late Devensian (Foster 1985; Franks 1987)
Glacial Lake Pickering	Clays, in some instances laminated, spread across the lowlands of the VoP	Ice/moraine dammed, proglacial lake occupying a 70-60, 45 and 30m OD lake levels (Kendall 1902; Penny and Rawson 1969; Straw 1979)
Sherburn Sands	Undulating, benched deposits, spread across the northern scarp slope of the Yorkshire Wolds	Scarp fed fan deltas, fed by nival runoff from the Yorkshire Wolds (Evans *et al*. 2016), or a marginal glacial outwash train (Foster 1985).
Seamer sands and gravels	Hummocky, low angled-graded sands and gravels, interbedded with diamicton in the eastern Vale of Pickering	Ice proximal outwash fan from the Mere Valley, when ice stood in the Seamer vicinity (Kendall 1902; Penny and Rawson 1969), or ice-proximal glacifluvial outwash deposits from ice in the eastern Vale of Pickering (Foster 1985; Franks 1987).
Flamborough Moraine/ Cayton-Speeton Moraine	Undulating topography consisting of a two-tiered diamicton, correlated to the Skipsea till on Holderness.	Ice contact landform corresponding to North Sea ice positioned close to the contemporary Yorkshire coast (Farrington and Mitchell, 1951; Penny and Rawson, 1969; Straw 1979)
Palaeolake Flixton	Well sorted, fine grained sediments infilling a topographic depression in the eastern Vale, north of Flixton village.	Palaeolake formed within kettle-kame topography in the eastern Vale of Pickering, after Late Devensian ice recession (Palmer *et al.* 2015).

in 1968. Over the course of a few years, the CEGB recorded at least 74 boreholes, 400 meters apart that ran across the south side of the Vale of Pickering across the moraine at Wykeham and up into the Hutton Buscel kame complex. Using data from the log book, Edwards (1978) and Foster (1985) provide a cross section of the Vale of Pickering that reveals multiple lake stages (Figure 1.6.5). The cross section provided by Edwards (1978) shows that the clay has layers of sand intermittently dispersed. These sand layers range from a few cm thick to several meters. Over the top of the lacustrine clay are outwash sands and gravels capped by a post-glacial alluvium of varying thickness.

In the north-western Vale, the landscape is more dissected. The deposits here lack the deep clays of the central vale (Figure 1.6.3) and a series of boreholes (Reeves *et al*., 1978) describe a layer of gravels underlying post-glacial alluvium, although no dating evidence is provided. Work by Powell *et al*. (2016) record weathered till deposits capping the Kimmeridge clay hills

(Table 1.6.1; Figure 1.6.4). Edwards (1978) argues that till caps the Jurassic clay spur at Thornton le Dale (Table 1.6.1).

In the middle of the basin north of Malton, the River Rye meets the River Derwent. The Rye flows north eastwards before being captured by the reversed River Derwent. Initially, the Derwent had flowed eastwards meeting the sea near Filey, but morainic debris left during the LGM reversed the Derwent. The canalised outlet at Sea Cut into Scarborough was also restricted to force the river to flow southwards through the Forge Valley (Table 1.6.1) and alleviate flooding of the Derwent in times of heavy rain. The Derwent flows south-westwards across the Vale of Pickering through Yedingham and Rillington before turning southwards east of Malton. The river channel is large and well established. The Derwent flows east of Malton into the Malton Embayment where the bedrock becomes dominant and the glacial sands and gravels thin. The river flows along a series of old fault scarps to make an angular steep-sided gorge known as the Kirkham Gorge (Table 1.6.1). The Kirkham Gorge is thought to have been created after the waters of Lake Pickering (Table 1.6.1) overtopped the scarp at nearly 70 m O.D. (Kendall, 1902). There is evidence of terracing (Eddey and Lincoln, 2017) along the sides of the gorge, but the density of the undergrowth makes it difficult to discern whether these are geological or fluvial.

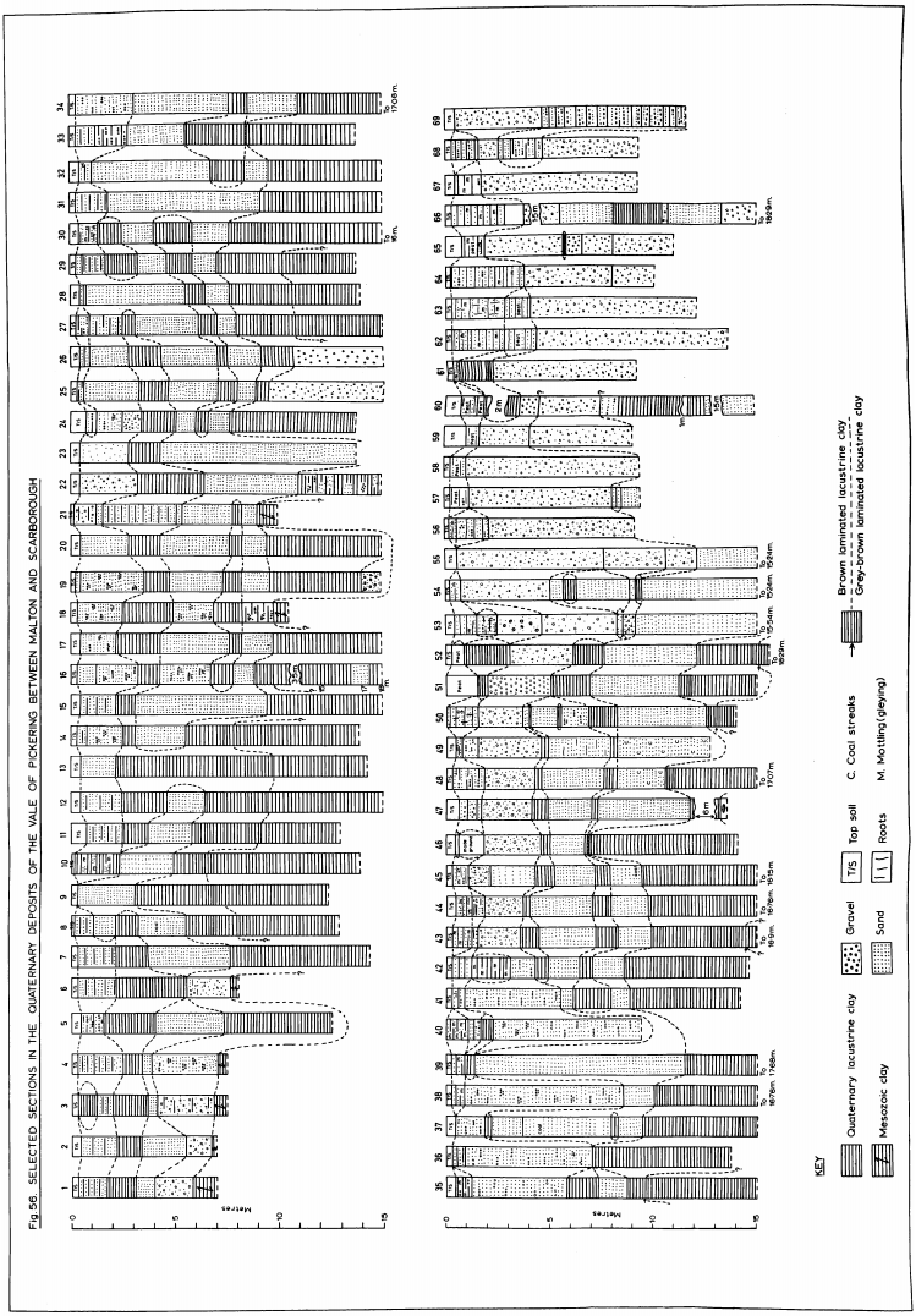


Figure 1.6.5: The Central Electricity Generating Board (CEGB) cross section (Edwards, 1978). Several alternating layers of sand and lacustrine clay can be seen suggesting that there were multiple iterations of Lake Pickering

1.7 History of research in the Vale of Pickering

Percy Fry Kendall (1902) must be credited with the first attempt to establish evidence for the existence of a former proglacial lake in what is now the Vale of Pickering (Figure 1.7.1). He envisioned the flat basin filled with glacial meltwater to a height of 70 m as it flowed from the North Yorkshire Moors into the Vale of York though the Vale of Pickering. Following on from work by Fox-Strangways (1892) and Lamplugh (1879), Kendall (1902) had also established mapping evidence for a moraine complex along the east coast, leading the idea of a former glacial ice dam. Evidence for a large glacier in the Vale of York (Dakyns *et al*., 1886; Lewis, 1887, 1894; Kendall and Wroot, 1924; Edwards *et al*., 1950) provided a western ice barrier at Ampleforth (Kendall and Wroot, 1924). Kendall (1902) also noted the Kirkham Gorge as the lake's exit point from the Vale of Pickering. He suggested that lake waters reached 70 m O.D. and overtopped the Howardian Hills, carving the Kirkham Gorge and emptying Lake Pickering in a singular catastrophic event (Figure 1.7.1).

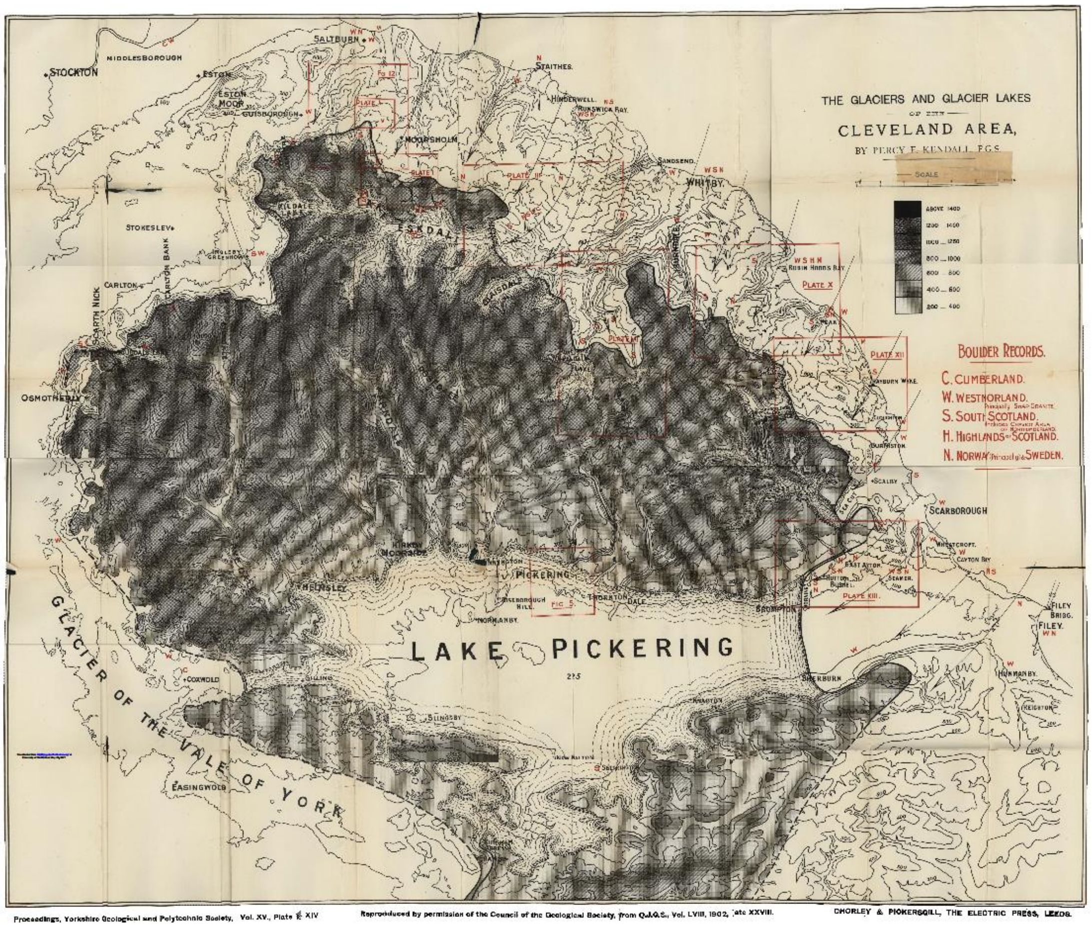


Figure 1.7.1: Kendall (1902) map of Lake Pickering and associated NSL and VofY ice based on the altitude of the kame terrace at Hutton Buscel and the outlet at Ampleforth of 69 m.

For many years, Kendall's interpretation was unmatched. However, in 1979 Straw suggested a two-stage model with the 70 m Lake Pickering coinciding with an earlier glacial maximum at Wykeham and a later 45 m O.D. once the NSL had retreated to Cayton Bay. He suggests that the lake emptied during the retreat phase through an outlet at Filey and into the North Sea, not through the Kirkham Gorge (Straw, 1979; 2016).

Subsequent unpublished PhD theses of Edwards (1978), Foster (1985), and Franks (1987) did not identify any of Kendall's (1902) evidence for a 70 m lake. Although Edwards (1978) describes laminated clay at an elevation of 76 m O.D near Thornton-le -Dale, outcrops of Ampthill and Oxford clay are frequently laminated, suggesting a misidentification by Edwards (Franks, 1987). The lack of shoreline evidence for high lake stages (e.g.70 m), however, is problematic. Gregory (1965) suggesting many reasons for this: 1) fluctuating water levels, 2) a soft, easily eroded bed rock, 3) an unfavourable direction of the prevailing winds, or 4) subsequent periglacial erosion.

Edwards (1978) believed that the presence of the channel at the base of Golden Hill (Figure 1.6.1) indicated that during the LGM the summit was sub-aerial and that lake levels did not rise above the height of the hill (54 m O.D.). He proposed that during the LGM, hill tops in the western portion of the Vale of Pickering were not submerged beneath lake waters, but instead accumulated firn. As the firn melted, channels carrying local gravels infilled and smaller lakelets emerged where water ponded (Edwards, 1978). He suggests the lakelets had oscillating water levels due to seasonal climatic variations and localised melting from the distribution of examining clay deposits. He also notes the lack of geomorphological evidence for a recent extensive lake in the western part of the Vale of Pickering (Edwards, 1978) such as a lack of shorelines or related features. The presence of birch twigs in much of his sediment descriptions lead Edwards (1978) to suggest the lake still existed at 16,713 ± 340 BP (Jones, 1977; Edwards, 1978).

Foster (1985) estimates Lake Pickering between 40 m O.D. and 27 m O.D. Foster (1985) suggests Lake Pickering is partially a subglacial lake with a floating shelf ice shelf extending from the terminal moraine at Wykeham to match a spur of Kimmeridge clay at Thornton-le-Dale and later advancing as far as Malton. His reasoning is not entirely clear, but he cites evidence of deep lacustrine clays buried under thick deposits of outwash sands and gravels (Figure 1.7.2) (Foster, 1985) and a small erratic lump of clay found at Wath Quarry. He concludes that the end of Lake Pickering resulted from the silting up of the marshy depressions left after the glaciers had retreated.

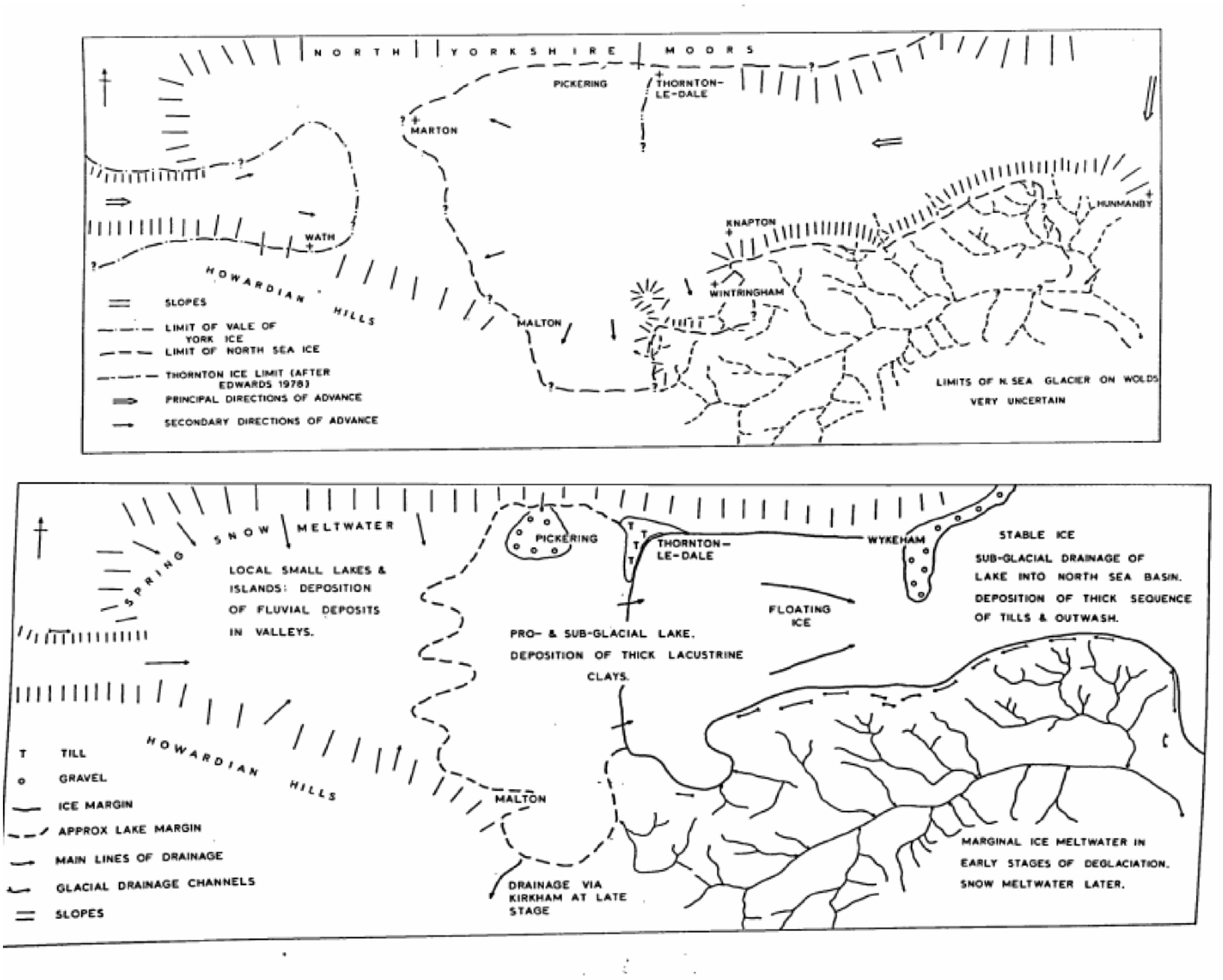


Figure 1.7.2: Foster (1985) hypothesised maximum Devensian ice extent in the Vale of Pickering. Note Edwards (1978) maximum limit at Thornton le Dale (top of map, middle)

Recent work by Evans *et al.* (2017) investigated sands (Sherburn Sands Figure 1.6.1) deposited at East Heslerton, exposed though excavations at Cooks Quarry. They interpreted from this a series of prograding alluvial fans overlying lacustrine sand deposits. The lake deposits were dated by OSL to 17.6 ± 1.0 ka, which Evans *et al*. (2017) attribute to a 45 m lake stage while the ice had retreated east of Wykeham. They describe deep water rhythmites in their LF1 overlain by pore-water escape structures indicating shallow lake or marshy conditions by 17.3 ± 1.0 ka (Evans *et al*., 2017) The post-glacial sedimentation evidence from Cook's Quarry reveal the lake had disappeared by 15.8 ± 0.9 ka supporting Livingstone *et al*. (2012) model of establishing LGM Lake Pickering at 22 ka that lasted until 16 ka.

The origin of the Kirkham Gorge as a Late Glacial spillway cut by the waters of Lake Pickering is often cited (e.g: Marshall, 1820; Fox-Strangways, 1880; 1881; Carvill Lewis, 1887; Reed, 1901; Kendall, 1902; 1903; Kendall and Wroot, 1924; Versey, 1929; Penny and Rawson, 1969; Edwards, 1978; Straw and Clayton, 1979; Foster, 1985; Franks, 1987; White, 1989; Murton and Murton, 2012; Palmer *et al*., 2014; Evans *et al*. 2015), but very little research has been focused on how the gorge was formed. It is generally accepted that prior to the blocking of the eastern end of the Vale of Pickering by glacial ice, the course of the River Derwent ran towards the coast and met the sea at Filey (Reed, 1901). The subsequent advance of ice and formation of a proglacial lake during the Devensian reversed the course once the lake had started to fill the old river channel running along the foot of the North Yorkshire Wolds scarp. The main question seems to be whether the Kirkham Gorge was cut during the Devensian or was already in existence before the onset of glaciation. Melmore (1935) suggests that the lake did not empty through the Kirkham Gorge, but rather through an older stream course between Westow and Gally Gap (at 61 m O.D.).

Reed (1901) wrote extensively on the river evolution of the Vale of Pickering and its relationship to the North Yorkshire Moors, Howardian Hills, North Yorkshire Wolds and Vale of York. In his work, he suggests that the col at Kirkham was cut post-glacially and that lake waters had originally escaped between Cram Beck bridge and Crambe (Reed, 1901). Versey (1929) was critical of this analysis. He notes the disparity between the steepness of the Kirkham Gorge and the shape of the Malton embayment with the sudden opening of the river channel north of Huttons Ambo. Versey suggests that prior to the LGM, the River Derwent had flowed northwards, capturing many of the smaller northward flowing streams and that the apparent disparate shape (from wide and flat to steep and narrow) of the Kirkham Gorge to where the Derwent enters the Howardian Hills at the Malton embayment suggests that the channel of the Kirkham Gorge existed in some form prior to the Devensian. Versey (1929) also suggests that the surface of the Howardian Hills has been considerably eroded during the Quaternary and that, prior to the onset of glaciation, the chalk of the North Yorkshire Wolds had once extended over the Howardian Hills. If this is the case, then it is possible that the Pleistocene glaciations have eroded away the chalk and the channel that is now the Kirkham Gorge is some remnant of that process (Versey, 1929).

The Kirkham Gorge is also an important feature in constraining Glacial Lake Pickering lake levels. In the Vale of Pickering, currently the lowest outlet is the Malton embayment at around 15 m O.D., but prior to the Holocene this was likely higher. Outlets in the Howardian Hills at Eddlethorpe Hall (40 m O.D.), Wath (50 m O.D.), Gally Gap (61 m O.D.), and Bulmer (65 m O.D.) suggest that the highest lake level proposed by Kendall (1902) is not possible without some sort of blockage or obstruction at these points (Figure 1.7.3). However, it is possible that ice or glacial deposits within the Vale of York could create such a barrier.

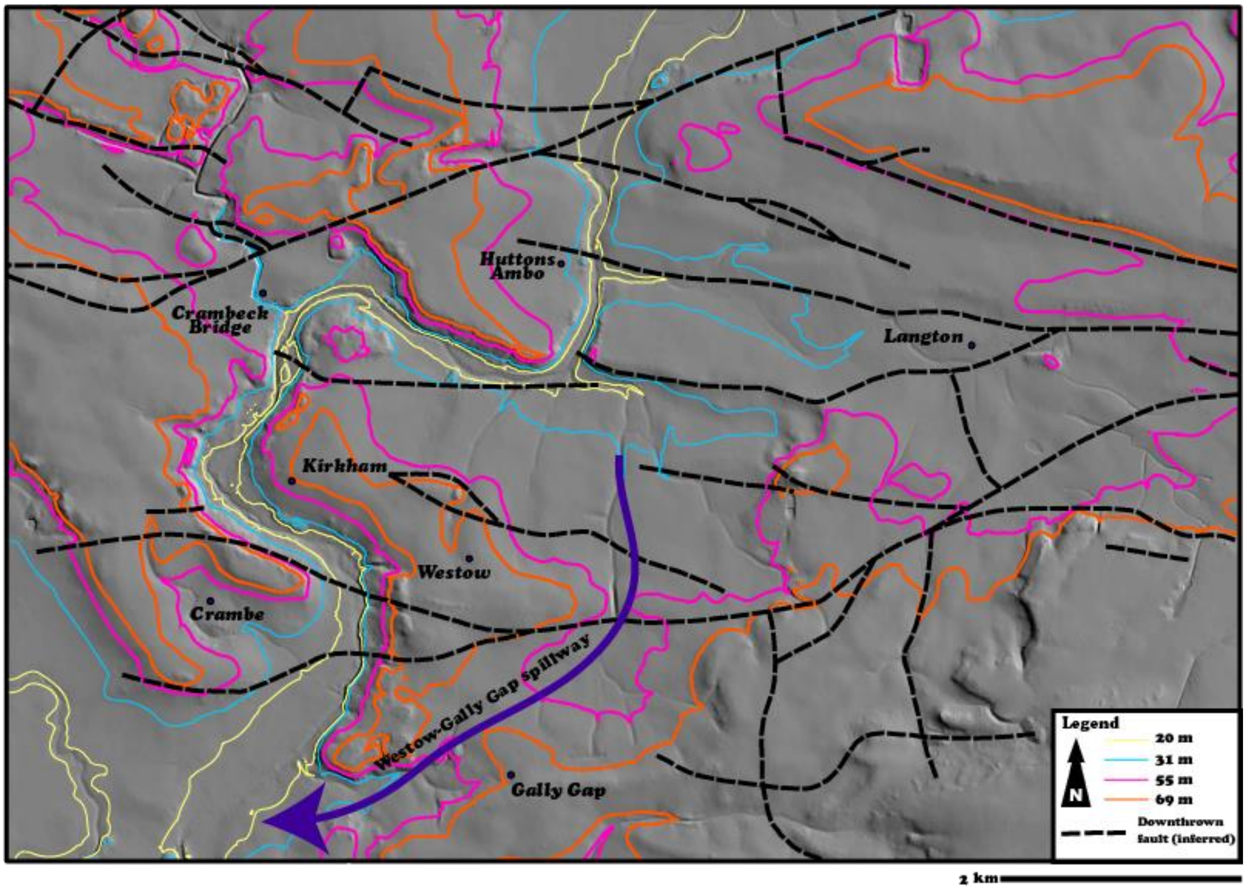


Figure 1.7.3: The Derwent Gorge and heights therein. Note the spillway between Westow and Gally Gap, preventing a 70 m lake. The barrier of the Howardian Hills is not enough to create a high stage lake, although ice in the Vale of York possibly could.

Edwards argued that the Kirkham Priory channel pre-existed the Late Devensian and was likely formed as a subglacial channel during the Anglian glaciation. He cited the general lack of available water and time to cut the channel from 75 m O.D down to its present level of 15 m O.D. as well as the presence of a lower outlet at 61 m O.D to the south and east of Westow (Edwards, 1978).

Although Kendall's research primarily focused on the formation of Glacial Lake Eskdale in the North Yorkshire Moors, Kendall stated the lake overflowed cutting the Newton Dale Gorge and Forge Valley Gorge emptying into the Vale of Pickering. He suggested that other, smaller proglacial lakes on the north-western side of the Cleveland Hills also drained into Lake Pickering cutting a gorge north of Helmsley and through the Coxwold-Gilling Gap (Kendall, 1902; Figure 1.7.3). Kendall asserted that Glacial Lake Pickering had reached a maximum height of 225 feet (~69 metres) and eventually overflowed through the Kirkham Gorge, draining into Lake Humber (Kendall, 1902).

Kendall's (1902; 1903) analysis of the role of Newton Dale has been extensively reviewed by Gregory (1962) and others (Peel, 1949; 1956 and Sissons, 1958). In short, Gregory argued that the Newton Dale channel was not supplied with the amount of water Kendall suggested and instead was likely a pre-existing channel. Gregory also asserted that the other overflow channels described by Kendall (1902) were probably sub-glacial channels in origin and represent stagnating ice rather than an active ice retreat front (Gregory, 1965).

1.8 Aims of the thesis

This thesis aimed to reconstruct the Late Quaternary landscape evolution of the Vale of Pickering

To achieve this, several research questions were considered, and four objectives defined;

1. What is the maximum extent of Glacial Lake Pickering?
2. What is the extent of ice and pattern of retreat within the Vale of Pickering and how is it related to the dynamics of the British Irish Ice Sheet (BIIS)?
3. What are the mechanisms behind the filling and emptying of Glacial Lake Pickering?
4. What is the duration of Glacial Lake Pickering and did it exist for one or several stage(s)?

Objective 1: Assessment of lake extent involves geomorphological analysis using high resolution LiDAR (0.5 m and 2 m) and NextMap (5 m) imagery and mapping of shoreline and lacustrine related features. Relationships between sediment deposits, analysed through sediment geochemistry and particle size analysis, were assessed by correlating new and historical borehole data. Groundtruthing was used to establish positive identification of landforms and ascertain their relationships to one another.

Objective 2: Using high resolution (0.5 m and 2 m) LiDAR imagery and NextMap (5 m), glacial landforms were mapped. Glacial limits were recognised through stratigraphic relationships and identifying geomorphological expression within the landscape.

Objective 3: To understand the hydrological dynamics within the Vale of Pickering, a number of floodfills were made using high resolution, Glacial Isostatic Adjustment (GIA) corrected DEM models using Arc GIS annotated with different ice limit scenarios from objective 2.

Objective 4: Results from objectives 1-3, when combined with new optically stimulated luminescence (OSL) from sediments within the Lake basin is used to create a full reconstruction of Lake Pickering and its relationship to the broader dynamics of the British Irish Ice Sheet (BIIS).

2. Methodology

2.1 Introduction

To understand the palaeoenvironmental history, many analytical methods were implemented. Geomorphic mapping from high resolution imagery was undertaken to establish lake levels and extent. Isostatically corrected elevation models were constructed to see how these lakes might have fitted into the landscape and their relationship to ice sheets blocking topographic lows. Existing and new boreholes were analysed in terms of their stratigraphy and sedimentological characteristics using Loss on Ignition (LOI), particle size analysis, and magnetic susceptibility. This was used to establish a stratigraphic framework for the deposits found within the Vale of Pickering. Concurrently, to establish an age chronology, optically stimulated luminescence (OSL) dating was applied to a range of sediments at sites across the Vale of Pickering.

2.2 Geomorphic mapping

The most recent mapping of the area was presented as part of the BRITICE v.2 project (Clark *et al.*, 2018) and mapped using NEXTMap Britain™ data from Intermap Technologies. Since that time, the Environment Agency (EA) has released data acquired through an airborne mapping system known as Light Detection and Ranging or LiDAR. LiDAR data is available for almost all UK locations and is available at 2 m, 1 m, 50 cm, and 25 cm resolutions. The mapping presented herein attempts to improve the quality and accuracy of the geomorphic mapping for the Vale of Pickering by using a combination of primary and secondary remote sensing sources, as well as fieldwork and ground truthing.

Data for the Vale of Pickering mapping was based on Kendall, (1902); Penny and Rawson, (1969); Straw, (1979); Foster, (1985; 1987), Powell *et al*. (2016) and BRITICE V.2 data (Clark *et al*. 2018). Aside from Foster (1985), the interpretation for geomorphological features is from a few references to field evidence and very little has been mapped from aerial imagery. Most of the data presented for the Vale of Pickering area in the BRITICE mapping (Figure 2.2.1) was from an archive of scarce academic literature and there is a real need for the landforms within the area to be examined. BRITICE v.2 has not, as a result, been able to provide any new data for the area.

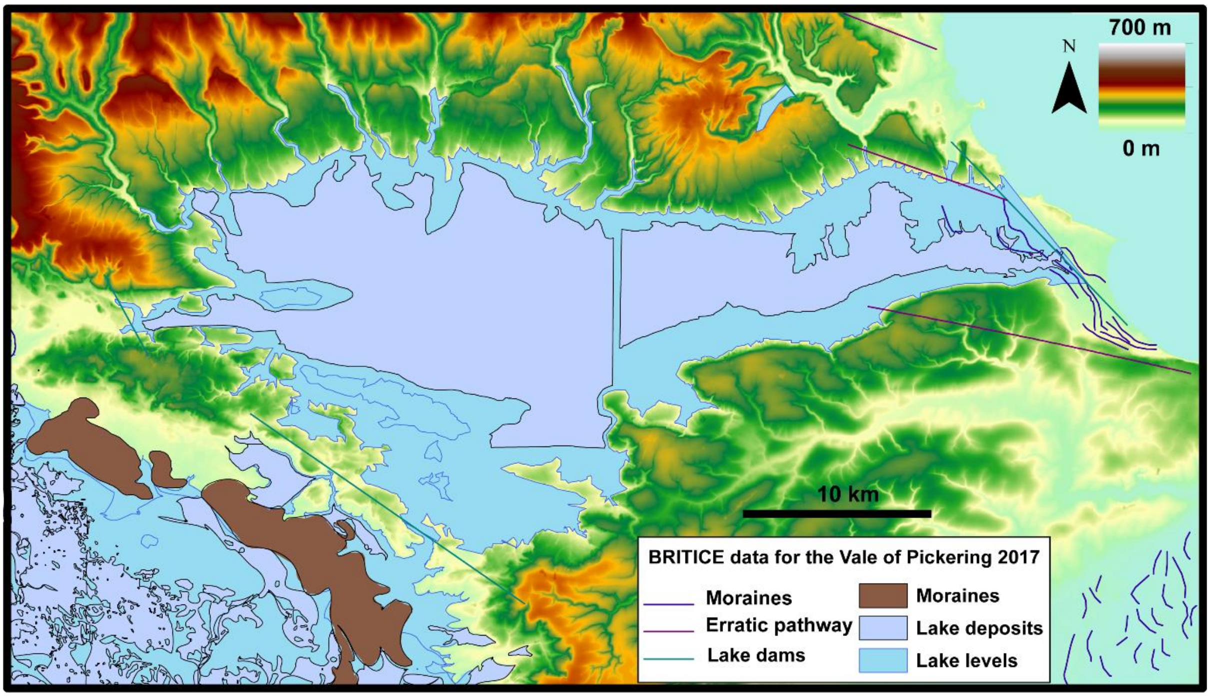


Figure 2.2.1: Data from the BRITICE v.2 map by Clark et al. 2018. Most of the data is from the surficial geology map produced by the BGS. The map has not been adequately updated and many incorrect designations still exist. Academic literature was used for the lake heights shown of 70 m (Kendall, 1902) and 45 m (Straw, 1979), although there are several other lake levels not represented.

The Vale of Pickering was mapped from both digital (NEXTMap, LiDAR) and print data (OS maps, BGS maps, academic literature). For both the NEXTMap and LiDAR data a Digital Elevation Model (DEM) was used rather than a Digital Surface Model (DSM). This is because DSMs contain all surface features, including buildings and trees, whereas a DEM does not.

Mapping was done using three different sets of DEM data. The first DEM was compiled from NEXTMap data. The data is collected through interferometric synthetic aperture radar (IfSAR). NEXTMap data is displayed at a resolution of 5 m with a vertical accuracy of 1 m. The other two were from LiDAR data at resolutions of 2 m, 1 m, and 50 cm, with a reported vertical accuracy between ±15 cm and ±40 cm, depending on resolution, location and elevation. For this study, the NEXTMap data had the advantage of covering the entire Vale of Pickering, while the LiDAR data was patchier (Figure 2.2.2). The LiDAR data, however, provided much clearer resolution of smaller landforms and the 2 m resolution was the preferred data set for mapping. The 1 m data had the same coverage as the 50 cm data. The 50 cm suffered from the vertical exaggeration needed and had a much smaller area of coverage.

NEXTMap data tiles come divided by British National Grid (BNG) reference. These two data sets were mosaicked in Arc Map and several hillshades were then applied.

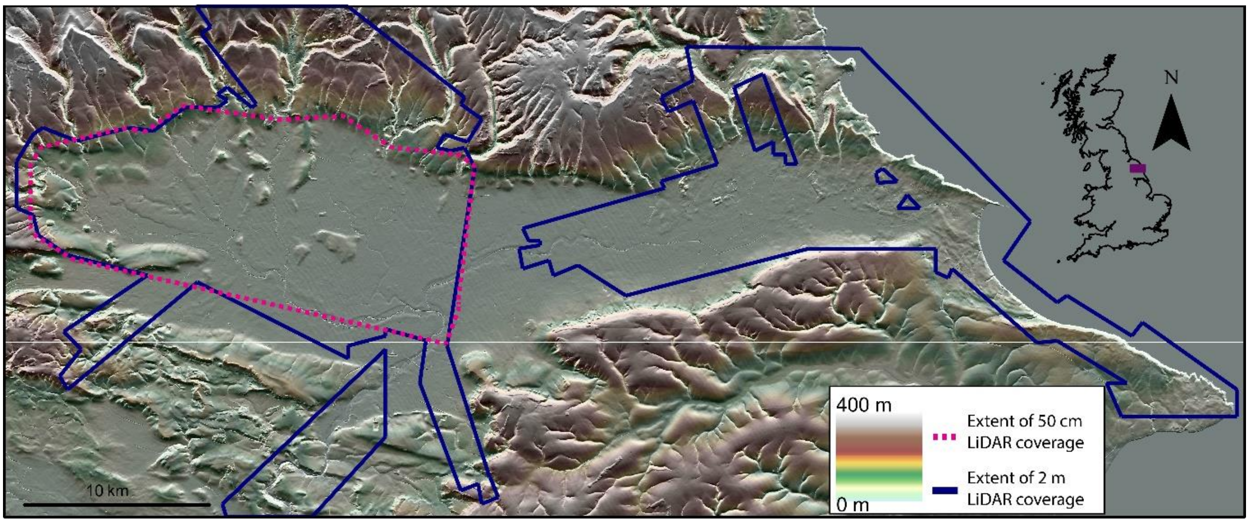


Figure 2.22: Map showing the coverage of IfSAR and LiDAR data for the Vale of Pickering. The NEXTMap 5 m DEM has full coverage. The blue outline shows the extent of the 2 m LiDAR data currently available and the dashed red line shows the 50 cm LiDAR data currently available. The 1 m data is comparable to the extent of the 50 cm data and was not used. The 25 cm data is negligible.

LiDAR data sets of 2 m, 1 m, and 50 cm were obtained from Open Survey online store (data.gov.uk) provided by the EA and processed from ACII format to raster following methods outlined in Davis (2012) and mosaicked into one raster data set. The data was then hillshaded.

In terms of processing the data for visualisation, several factors were considered. In the Vale of Pickering, many landforms align north-south or east-west, so the data needs to be illuminated from at least two perpendicular angles so that bias from shadowing is reduced (Smith & Clark, 2005). The data also needs to be presented with both a low sun altitude angle, as longer shadows help to distinguish individual features, and a high sun angle, which highlights large features, but can leave more delicate features flattened or hidden (Figure 2.2.3). Finally, the *Z* factor needs to suit the terrain and the scale of the data as to help pick out individual features easily, but not as to create pseudo-features (e.g.: Figure 2.2.6). A visual comparison of the two data sets (5 m and 2 m) is shown in Figure 2.2.4 (Panel A and B). While the resolution is clearer, the Vale of Pickering landforms are very muted due to lacustrine deposits draping the Vale. A vertical exaggeration is needed to elevate the appearance of features. Panel B shows a *Z* factor of 2, while for the two lower panels (C and D), the vertical exaggeration is set to *Z*=10. In comparison with panel B, the features become much clearer (Figure 2.24).

For a hillshade to be most effective, the lighting angle (azimuth) needs to be perpendicular to the features displayed (Clark and Meehan, 2001; Smith and Clark, 2005). For the Vale of Pickering, hillshades for each resolution were made with four different azimuths at 90°, 180°, 270° and 360° and at an oblique angle of 45° due to the east-west trend. In Figure 2.2.4, panel C shows a hillshade with an azimuth of 180° and panel D has an azimuth of 90°. Not the change in the appearance of the hill (centre-left) through all four panels.

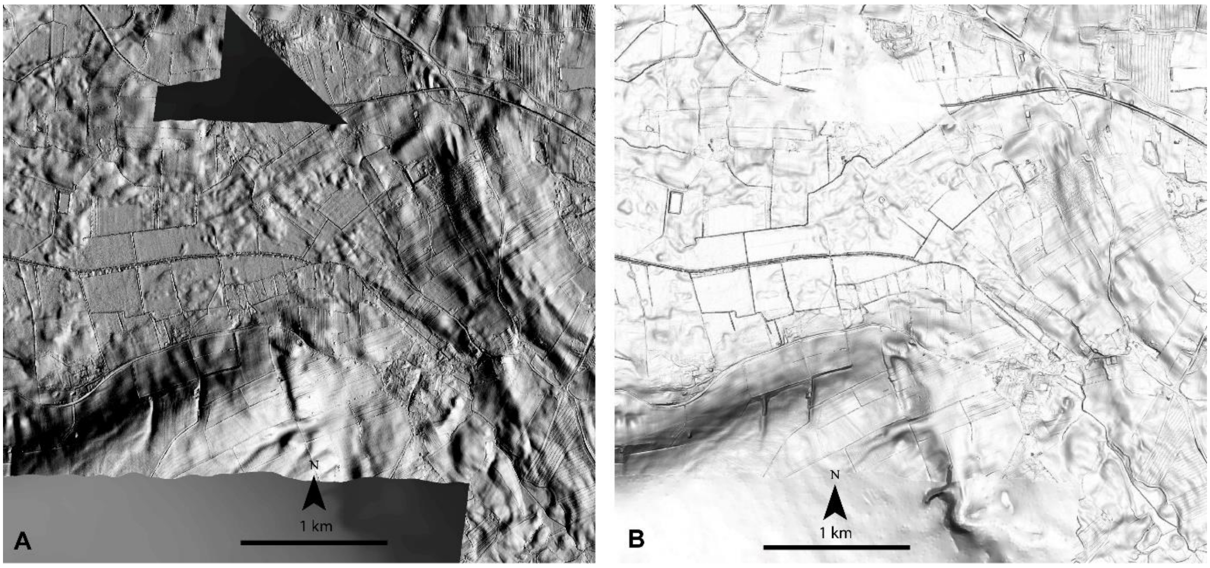


Figure 2.2.3: Difference between sun angle. Panel A shows an image at 90˚ azimuth and 35˚ sun angle. Panel B shows 90˚ azimuth with 85˚-degree sun angle. Both are useful for mapping landforms as the features become more muted at a higher sun angle allowing focus to the larger landforms. The 35˚ sun angle allows for more delicate features to be recognised. Z factor is 10 in both images

The 50 cm LiDAR data covered a smaller area of the Vale of Pickering. Its advantages are that it has very high resolution, which is helpful for examining features at a small scale and smoothing transitions between landforms where jumps in data in 2 m resolution may be mistaken for terraces or ridges. As shown in Figure 2.2.5, the contrast between the 2 m and 50 cm shows how much smoother the data is displayed, however, the 50 cm was not utilised in mapping as much because overall it did not provide much more evidence than could be garnered from the 2 m imagery. The 50 cm data was processed in the same way as the other data with multiple hillshades at the same azimuth and sun angle. The *Z* factor was too noisy at 10, so it was halved to a value of 5, which was about the limit for displaying the data.

Although the 5 m NEXTMap is advantageous in its resolution considering its UK-wide availability, it does suffer from false-positives. This is a data processing error common with

IfSAR due to the data collection method and the conversion of NEXTMap DSM to DTM. These false-positives primarily occur where large clumps of woodland or trees occur and show up as hummocks.

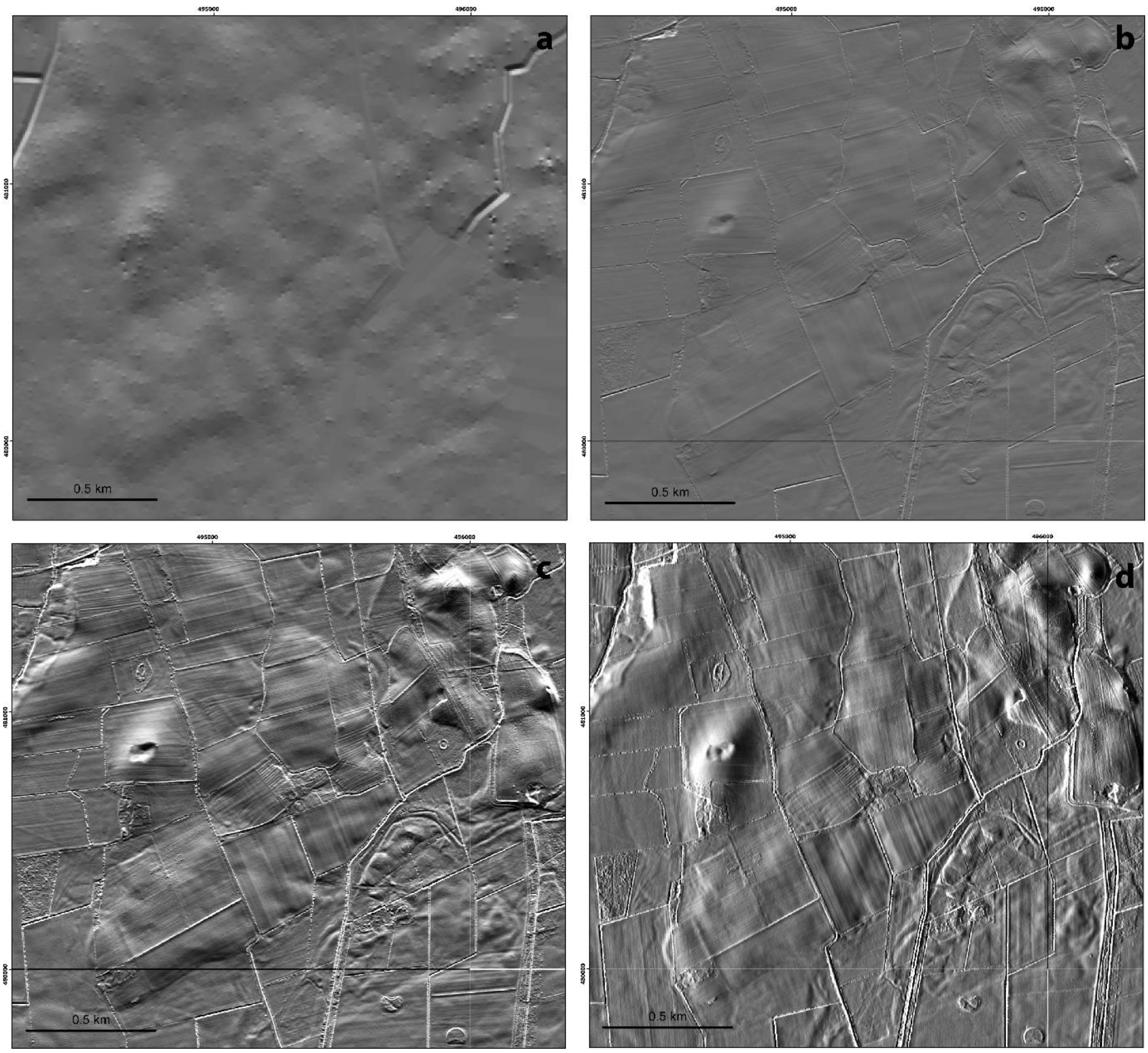


*Figure 2.2.4: The presentation of data. Panel **a** show the 5 m NEXTMap DEM compared with the 2 m LiDAR DEM in panel **b**. In panel b, Z=2 which shows how muted the landforms in the Vale of Pickering are. Panels **c** and **d** show the same area at 2m, but with a Z factor of 10. Finally, the data in panel **c** has an azimuth of 180° while panel **d** has an azimuth of 90°. All are produced with a sun angel of 35°.*

In order to correct for this, unexplained hummocks on the 5 m NEXTMap were compared with either the LiDAR data or, if LiDAR was not available, through secondary data like the Ordinance Survey map (Figure 2.2.6, panel B and D) or Google Earth. As in Figure 2.2.6, in two place there are spurs of raised ground. This feature is absent from all resolutions of LiDAR data and the OS map confirms there to be a small crop of trees at both points (Figure 2.2.6 red arrow).

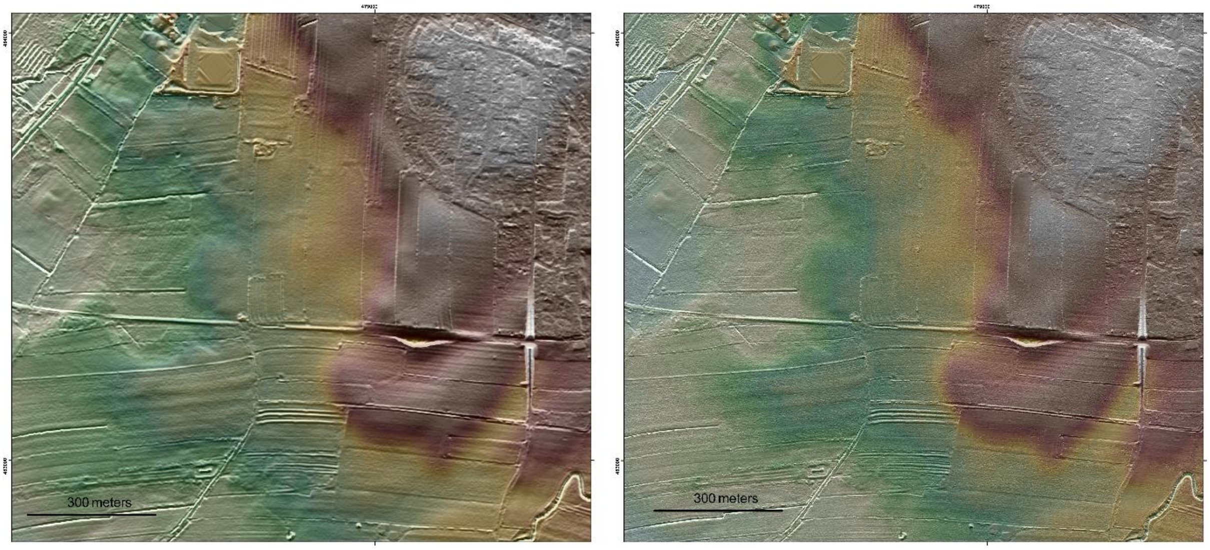


Figure 2.2.5: The difference between resolution of 2m LiDAR (on the left) and 50 cm LiDAR presented (on the right) at a scale of 1:4000. Although the features are sharper, the 50 cm data does not provide much more information than the 2 m. Therefore, it was primarily utilised for further investigating any questionable landforms.

Mapping with accuracy is dependent on three factors a) the type of landform; b) the data and c) the experience and confidence of the mapper with the data set. In order to establish a relationship with the data, a repeat-pass procedure is best utilised. This procedure involves repeatedly analysing the data from different scales and resolutions to be confident in the features that are present. In an area, like the Vale of Pickering, where little to no prior mapping has taken place, confidence in the type of landforms present is achieved through this repeated procedure.

When mapping the Vale of Pickering, the first pass was on the 5 m NEXTMap DEM, followed by the 2 m and then 50 cm LiDAR DEMs. Each type of landform was mapped completely before moving onto the next type. Any miscellaneous features were noted and added on a separate shapefile for further investigation later. The mapping scale varied from full Vale coverage at 1:100 000 on the 5 m DEM down to a scale of 1:4 000 for the 50 cm DEM. This variability in scale helped to view landforms as both a collective set and individually, which helps to visualise the relationships between landforms more effectively.

All landforms were mapped using polygons and/or polylines, depending on the nature of the feature. Polygons were mapped hollow and then coloured once all mapping had finished (Figures 17 - 20). A table of features in shown in Table 2.2.1.

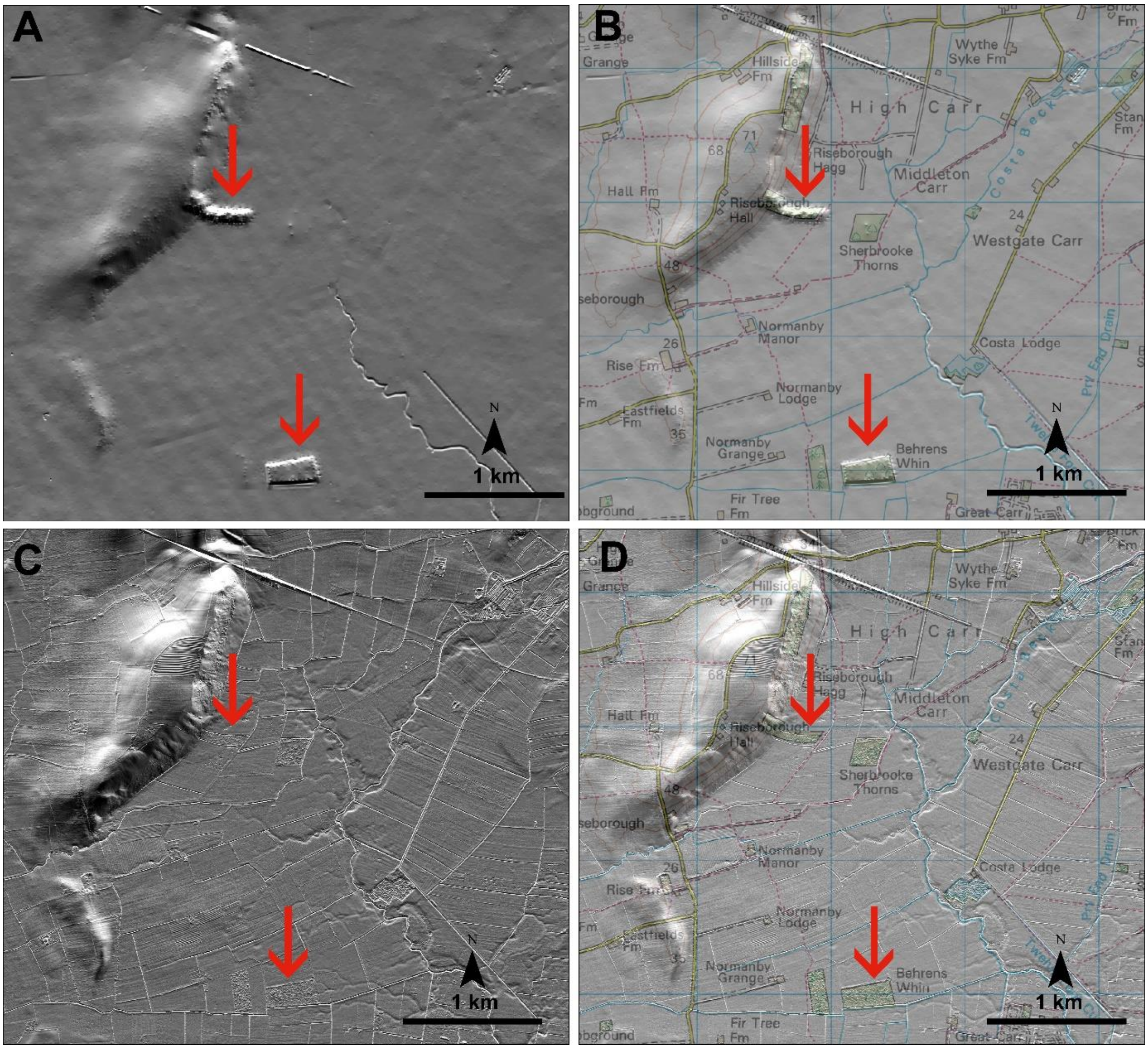


Figure 2.26: The false-positives of the 5 m NEXTMap. In panel A, two hummocks are seen, although when further investigated using the OS map, the hummocks are revealed to be clumps of woodland. Not all woodland is displayed as hummocks as post-processing of the NEXTMap data has removed some of the undesired surficial features as shown in the woodlands adjacent to the ones under the lower red arrow. The absence of these hummocks as landforms is confirmed in the 2 m LiDAR data shown in the panels C and D.

Once all landforms were mapped, quality control was established by checking each area in all available data sets, as well as considering other sources like BGS geological maps, OS maps, and the academic literature. Several trips were undertaken to the Vale of Pickering over the mapping period to ascertain the correct interpretation of features for mapping. Some were confirmed or re-classed as different features. Occasionally, a landform was reclassified or discovered. A more in-depth discussion of these features is found in the following results portion of this chapter.

Table 2.2.1: Features mapped within the Vale of Pickering

Feature	**Evidence**
Jurassic and Older diamicton	BGS maps, published and unpublished literature
Alluvial Fans and Deltas	DEMs and published and unpublished literature
Terraces and Benches	DEMs
Hummocks	DEMs
Moraines	DEMS, published and unpublished literature
Basins	DEMs, Lincoln, 2017
Palaeochannels	DEMs, OS map
Shorelines	DEMs.

As previously mentioned, secondary data sources for the mapping came from OS maps, BGS maps, academic literature and Google Earth. The 1:625 000 (Geology of Britain Viewer) and 1:50 000 BGS geological map were used to locate edges of bedrock and for information regarding fault lines. The Ordnance Survey 1:25 000 maps were used for extra information regarding features (Figure 2.2.6). Both maps were helpful in deciding how features related to the landscape. Widespread auguring in the western Vale of Pickering by Edwards (1978) was added to the map, as well as the approximate locations of the boreholes used in the Malton to Scarborough Transmission Line (CEGB, 1968).

2.3 Sedimentology

During September 2015, a shallow borehole campaign was launched to extract six cores from three sites within the Vale of Pickering in collaboration with the DANDO drilling team of the British Geological Survey (BGS). The three sites were located at Yedingham, High Marishes, and Salton (Figure 2.3.1). Locations were chosen based on historic borehole records (Section 2.4) and preliminary geomorphic mapping (Section 2.2). Landowners for the three sites were contacted by the BGS and a survey was completed for any obstructions from infrastructure such as power lines, buried cables, or gas pipelines.

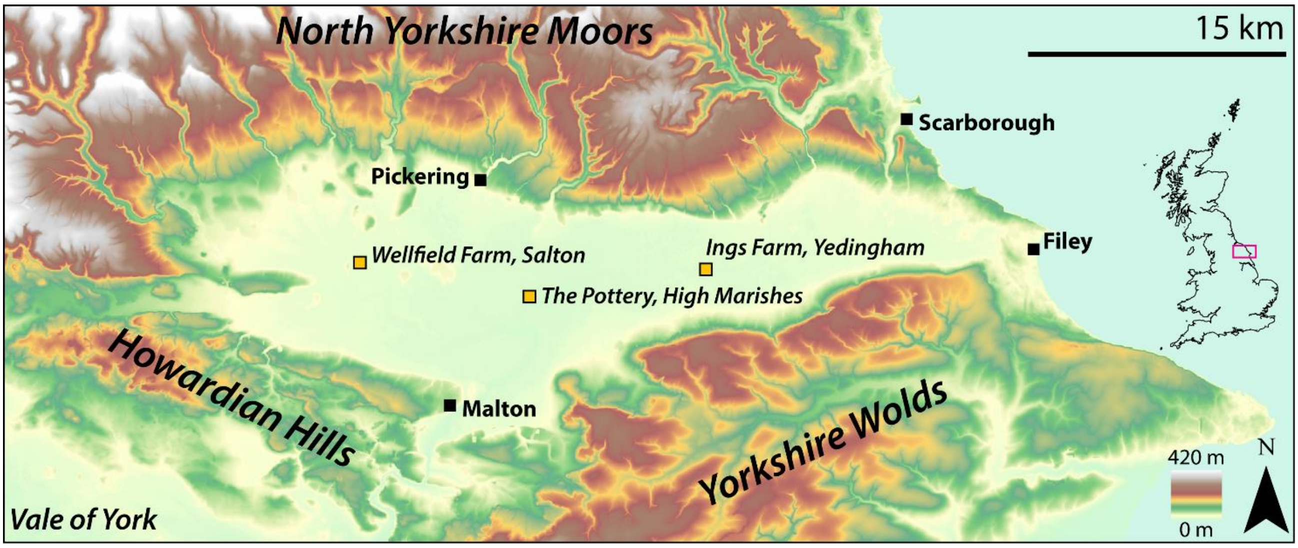


Figure 2.3.1: Location of sites (yellow squares) where DANDO shallow borehole cores were extracted. Two cores were extracted at each site: one in a clear liner and one in an opaque liner to protect sediments from light contamination for OSL dating.

In the field, preparation for drilling at each site included another subsurface sweep to check for any unknown service cables or pipes. At each site, two cores were completed. One in a clear liner and one in an opaque liner to protect sediments from light for OSL dating. The boreholes were logged according to BS5930.

Sedimentology techniques

Characterising the depositional and environmental history of unconsolidated sediment can be done by applying several laboratory techniques, like loss on ignition (LOI); particle size analysis (PSA); and magnetic susceptibility. The procedures for measuring these factors followed those of Gale and Hoare (1991).

Sediment samples were collected in the field from exposures, hand auguring, and through percussion coring for geochemical and particle size analysis. Each exposure and core were visually logged, noting changes in stratigraphy including Munsell colour, approximate grain size, degree of sorting, and sedimentary structures. Lithofacies codes are based upon those of Evans and Benn (2004) with sedimentary interpretations based on Lewis and Maddy (1999) in Jones *et al*., (1999). Samples were collected from exposures (the quarry faces at Ings Farm, Yedingham), DANDO shallow borehole cores (The Pottery, High Marishes; Wellfield Farm, Salton; and Ings Farm, Yedingham) and by hand augur (Slingsby and Caulkley's Bank) at intervals between 5 cm and 10 cm (depending on the size of the section and the availability of sediment) and placed in sealed sample bags or 5 cm cubes. In the laboratory, samples were analysed for low frequency and high frequency magnetic susceptibility, LOI, and particle size analysis.

For PSA, LOI, and magnetic susceptibility, all cores were sampled at 5 cm, except where tough clay meant a 10 cm interval had to be used. For the outcrops at Yedingham Quarry 1 samples were collected initially at 10 cm due to coarse gravel and then at 5 cm when the size fraction fined to sands and silts. North of Quarry 1 was Quarry 2. This site was sampled at 10 cm intervals and by sediment package, which ranged in size from 5 cm to 25 cm in thickness.

Particle size analysis

The behaviour of sediment in different depositional environments is reflected in the distribution of their particle size. For this reason, PSA is an important and sensitive parameter that can reveal much of the depositional history of unconsolidated material. Because the size of material transported can vary over eight orders of magnitude, a logarithmic scale developed by Udden (1898) and later expanded by Wentworth (1922) is frequently used to represent the spread of data. Krumbien (1934) developed phi the units (ϕ) the origin of which is set to 0 ϕ (equivalent to 1 mm). For each phi-unit increase in diameter, the corresponding size in millimetres is halved (Gale and Hoare, 1991).

In all, 294 samples from the Vale of Pickering were processed for PSA. The samples were dry sieved through a 2 mm mesh to separate sand, silt and clay from coarser fractions. Each sample was soaked in 10% sodium hexametaphosphate solution and dried. The coarse fraction was bagged and weighed. Particle size was then measured using a Horiba LA-920 laser diffraction particle size distribution analyser.

A number of statistical methods were used to further describe the distribution of particle size with a view of characterising the sediments and any changes in them. For each sample, the mean (M_{z}), sorting (σ_{I}), skewedness (Sk_{I}), and Kurtosis (K_{G}) were calculated. From these the descriptive terms of Folk and Ward (in Gale and Hoare, 1991) could be applied as well as the percent sand, silt/clay determined.

Magnetic Susceptibility

Magnetic susceptibility measures the induced magnetisation of a sample when placed in a weak magnetic field (Gale and Hoare, 1991). Magnetic minerals are abundant in most sedimentary environments and can reveal information about the history of change in environmental conditions. Since magnetic minerals are usually fine-grained and difficult to separate from a non-magnetic matrix, samples are measured in bulk (Gale and Hoare, 1991). Much of the magnetic material brought into lake environments is allogenic (from outside) rather than authigenic (created *in situ*) (Engstrom and Wright, 1984). This means that the surrounding bedrock can be highly influential in the magnetic susceptibility of a deposit and pulses of sediment entering into a lake system can be noticeable. The rock sources which make up the deposits in the Vale of Pickering are varied because of the mechanisms of transport – fluvial, glacial – that brought them into the area. Therefore, the magnetic susceptibility of the following sites can be used to estimate whether far-travelled rocks, such as granite or quartzite that are not local to the area, are contained within the deposits. A higher value for magnetic susceptibility indicates source material, generally as

Basic extrusive rocks > basic intrusive rocks > acid igneous rocks > metamorphic rocks > sedimentary rocks (Gale and Hoare, 1991).

Other sources of magnetic minerals may come from organic matter, the atmosphere, groundwater, lake waters, and rivers. Since the surficial deposits in the Vale of Pickering are predominantly glacially-derived, there is generally very little organic matter below the top meter of the deposits. For the detection of organic material, these results are combined with LOI (see LOI section below). Atmospheric sources of magnetic minerals are from dust or fires, which may have some input into the glacial lake system but would likely not be a dominant source. In lake waters, deposition of iron occurs near the sediment-water interface in the form of coatings, surface crusts, concretions/nodules, or through reduction by bacteria living just below the sediment-water interface (Gale and Hoare, 1991).

To measure for magnetic susceptibility, samples from all cores were dried in 5 cm cubes and measured using a Bartington MS2 magnetic susceptibility system. To ensure quality of data, blanks were used before and during measurement. Each sample was measured three times to allow for instrument fluctuations, and an average value used. Values for high frequency (χ_{HF}) and low frequency (χ_{LF}) were taken but provided similar values so only χ_{LF} was plotted as a function of depth.

Loss on Ignition

LOI is a technique used to establish the amount of plant organic carbon content in a sample. It is a simple and quick method established by Davies (1974) and recommended as the preferred method for measuring plant organic matter in samples by Gale and Hoare (1991). Samples were dried at 105˚C for 24 hours to remove moisture. Once dried in a desiccator and weighed, samples were then furnaced at $430\pm^{20}_{55}$ ºC for 24 hours and then reweighed. To calculate the percentage by loss on ignition, the following equation is used

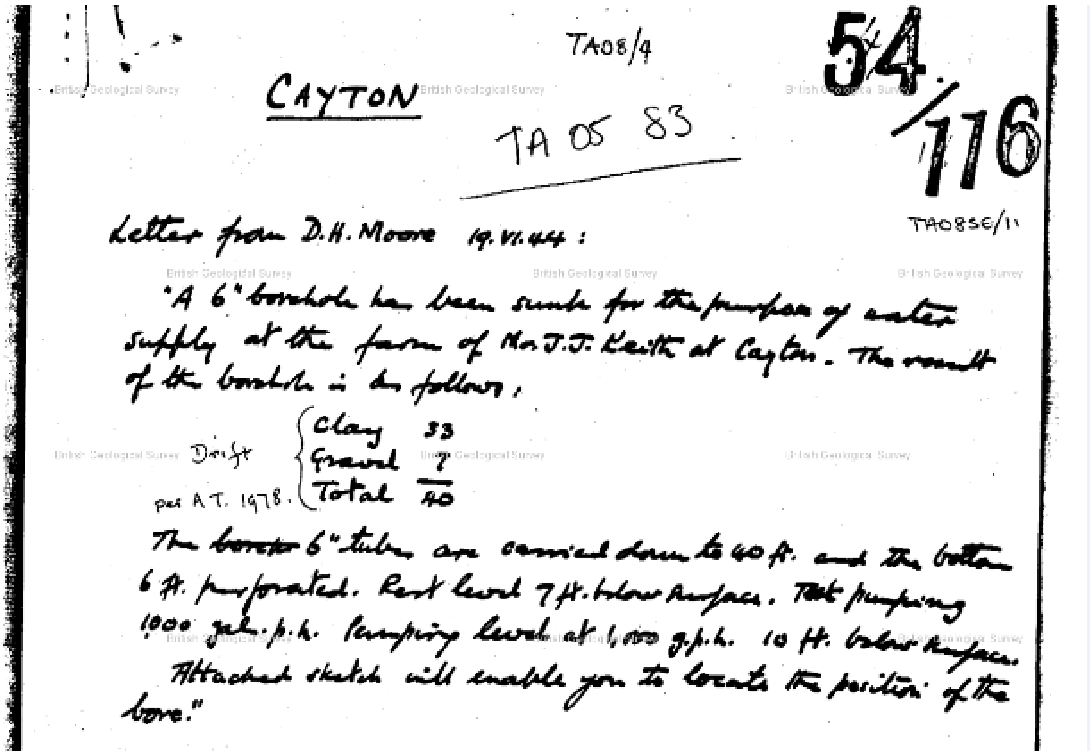
Equation 2.31*:*

$$\text{LOI (\%)} = 100\,\frac{(M_2 - M_1) - (M_3 - M_1)}{(M_2 - M_1)}$$

if M_1 = crucible weight, M_2 = dry (105ºC) weight, and M_3 = furnace (430ºC) weight (Gale and Hoare, 1991).

2.4 Historic borehole data

There are over 1000 historic borehole logs for the Vale of Pickering. However, many contain little information or are derived from old field slips and personal observations that can often be illegible (Figure 2.4.1).



TA08/4

CAYTON

TA 05 83

54/116

TA08SE/11

Letter from D.H. Moore 19.VI.44:

"A 6" borehole has been sunk for the purpose of water supply at the farm of Mr J.J. Keith at Cayton. The result of the borehole is as follows,

Drift	Clay	33
	Gravel	7
per A.T. 1978.	Total	40

The borehole 6" tubes are carried down to 40 ft. and the bottom 6 ft. perforated. Rest level 7 ft. below surface. Test pumping 1000 gal.p.h. Pumping level at 1,000 g.p.h. 10 ft. below surface.

Attached sketch will enable you to locate the position of the bore."

Figure 2.4.1: A historic borehole record from Cayton in the Vale of Pickering. Many contain little to no useable information.

The historic borehole records are most often located along main roads, such as the A64, or are related to works on infrastructure. As a result, there are more boreholes in towns around the fringe of the basin rather than in the lake basin itself (Figure 2.4.2).

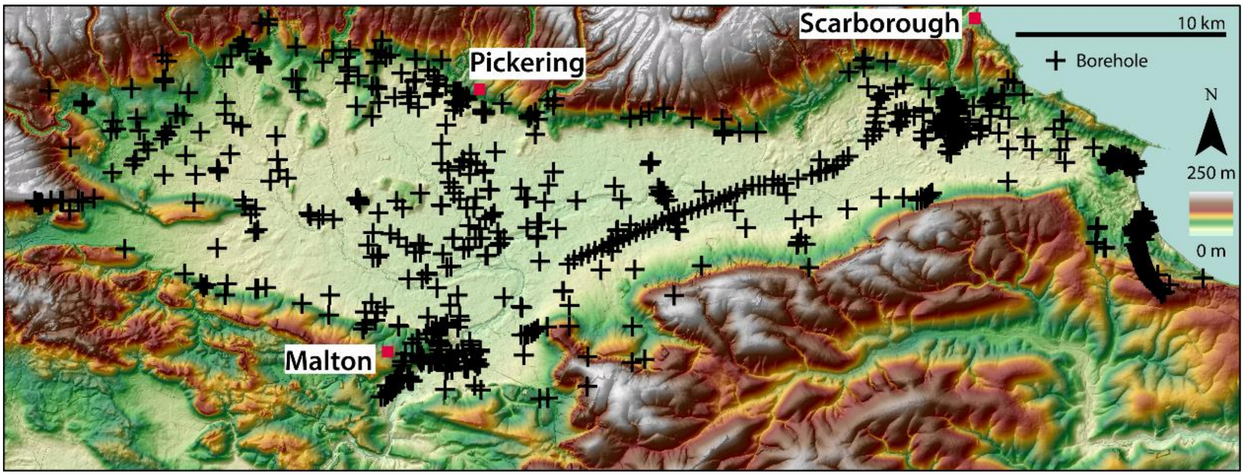


Figure 2.4.2: Boreholes in the Vale of Pickering. Most contain little information on subsurface deposits or are locations of disused wells

In 1967, the Central Electricity Generating Board (CEGB) ran a series of boreholes across the Vale of Pickering from Malton to Irton Moor just south of Scarborough. Initially the boreholes were drilled every forth tower (2 km), but due to the unpredictable nature of the undulating lacustrine sand and clay subsurface in the Vale of Pickering borehole were taken every 400 m. The result was a suite of 76 boreholes that stretch across the lake basin (Figure 3.1.31).

To create a 2D cross-section from the borehole logs, the measurements of each record were first converted from feet and inches to metres and then divided into 10 cm intervals. In Excel, a table was created the sedimentary description per 10 cm of each record was logged and assigned a sedimentary code based on the sediment type or the logger's description (Figure 2.4.3; Table 2.4.1). The descriptions for sedimentary units ranged from noting the sediment type (ie: "Sand") to a more detailed record (ie: "Firm brown laminated clay with occasional silty fine partings").

Additionally, the CEGB technical report also included data on moisture content and PSA for individual sediment packages within the cores (Figure 2.4.3). These were incorporated in the results section for comparison with the sediment packages from The Pottery, High Marishes and Ings Farm, Yedingham DANDO cores.

RECORD OF BOREHOLE 47 Malton SE 97 NW | 21

BORING DIA. TO DEPTH 6" to 46'0"

BORING: STARTED 14.6.68 FINISHED 16.6.68 93737969

GROUND LEVEL Ft. O.D.

SAMPLE OR TEST DEPTH	TYPE	LEGEND	DEPTH	O.D. LEVEL	CHANGE OF STRATA DESCRIPTION
1'0"	D1		0·46 1'6"		Firm brown topsoil.
1'9"	D2		2'0" 0·61		Firm dark brown organic clay with pockets of peat.
3'0" 3'0"	B3 W1		1·83 6'0"		Soft clay, ashes and gravel (made ground).
7'0" - 8'0" 10'0" 12'0" - 13'0"	C(25) B4 C(20)		4·57 15'0"		Medium dense gravel and sand.
15'6" 15'6" - 17'0"	D5 U(4)1		16'6"		Firm brown laminated clay.
17'6" - 18'6" 22'0" - 23'6"	S(11) U(4)2		5·03 6·71 22'0"		Medium dense grey silty fine sand.
			22'6"		Firm brown laminated clay.
23'0" - 24'0" 23'6" 25'0" 26'6"	S(54) D6 B7 D8		6·86 7·92 26'0"		Dense brown medium sand.
27'0" - 28'6" 28'6" 32'0" - 33'6" 33'6" 38'0" - 39'6" 39'6" 43'0" - 44'6" 44'6"	U(4)3 D9 U(4)4 D10 U(4)5 D11 U(4)6 D12		14·02 46'0"		Firm brown laminated clay with occasional silty fine partings.

REMARKS Water seeping in at 17' to 18', standing at 5 ft. on 15 6.68, and 16 6.68 and at 3 ft. on completion.

KEY: D – DISTURBED SAMPLE B – BULK DISTURBED SAMPLE W – WATER SAMPLE S ( ) STANDARD PENETRATION TEST C ( ) CONE PENETRATION TEST NO. OF BLOWS FOR 12 INCH PENETRATION IN BRACKETS U(4) – 4 INCH DIA. UNDISTURBED SAMPLE

PRE-PILING SURVEYS LTD., PETERBOROUGH

Figure 2.4.3: Example borehole record (#47) from the Central Electricity Generating Board (CEGB). Measurements in feet have been converted to metres. Sediments description ranges from a simple sediment type (i.e.: sand) to more descriptive instances where observations on particle size and sedimentary structures are included. In the second column marked Type, codes beginning with an S indicate where PSA samples were taken by the loggers.

Table 2.4.2: Lithofacies codes devised and attributed to the CEGB borehole logs

Code	Sediment	Description
MADE	Man-made	Disturbed ground, concrete, asphalt, made ground
6	Topsoil	Topsoil of any colour, texture, friability incl. agricultural
1ORG	Organics	Silty organic clay, any colour
1PEAT	Organics	Peaty clay, any colour, texture
1ORGP	Organics	Organic clay with peat any colour, texture
PEAT	Organics	Peat, any colour, texture
PC	Organics	Clayey peat
12P	Clay/Organics	Clay, sandy, with pockets of peat
12P4	Clay/Organics	Clay, sandy, with pockets of peat and gravel
123ORG	Clay/Organics	Organic silty sandy clay may have root/rootlets
1	Clay	Clay
1A	Clay	Firm to stiff brown mostly laminated including silty fine partings
1A.1	Clay	Grey/brown laminated clay
1A/1K	Clay	A mix of clay between 1A and 1K type
1AC	Clay	Clay with no silt partings
1G	Clay	Clay with gravel, any colour
1K	Clay	Deep lacustrine clay. Not Kimmeridge, but overlies and/or is derived from Kimmeridge clay
12	Clay	Clay with sand
12.1	Clay	Clay with sand, part of a section
12.2	Clay	Clay with sand, part of a section above 12.1
124	Clay	Clay, gravel and sand, any colour
12R	Clay	Clay with sands and roots, any colour
13	Clay	Clay with silt, any colour
13R	Clay	Clay with silt and roots, any colour
14R	Clay	Clay with gravel and roots, any colour
14G	Clay	Clay with a lot of gravel, any colour
K	Clay	Kimmeridge Clay, Jurassic
GK	Clay	Green clay, Jurassic
RED	Clay	Red clay, Jurassic
2	Sand	Sand
2A	Sand	Medium dense, medium to coarse sand
2A.1	Sand	Medium dense, medium coarse sand, appearing above 2A
2A/2C	Sand	A mix of 2A and 2C
2B	Sand	Loose grey silty fine sand
2C	Sand	Medium dense, medium brown sand with occasional gravel
2G	Sand	Sand with gravel, any colour
2COAL	Sand	Sand with coal fragments
21	Sand	Sand with clay, any colour
213	Sand	Sand with clay and silt, any colour
214	Sand	Sand with clay and gravel, any colour
23	Sand	Sand with silt, any colour
231	Sand	Sand with silt and clay, any colour
24	Sand	Sand with gravel, any colour
24A	Sand	Loose to medium dense, grey silty, fine to medium sand with occasional fine chalk and flint gravel
24B	Sand	Medium dense, brown, medium to coarse sand with occasional fine gravel
241	Sand	Sand with gravel and clay, any colour
234	Sand	Sand with silt and gravel, any colour
3	Silt	Silt, any colour
312	Silt	Silt with clay and sand, any colour
4	Gravel	Gravel
4CH	Gravel	Gravel with/made with chalk
41	Gravel	Gravel with clay, any colour

42	Gravel	Gravel with sand, any colour
42A	Gravel	Medium dense, brown fine to coarse gravel with a little sand
42B	Gravel	Dense gravel with some sand
42C	Gravel	Loose gravel and sand
42D	Gravel	Medium dense gravel and some sand
42R	Gravel	Gravel and sand with roots, any colour
COBBLES	Cobbles	Cobbles
COAL	Coal	Coal dust, deposit, gravel

In other parts of the Vale of Pickering, historic boreholes can reveal much about the subsurface and were either used as supporting evidence for subsurface interpretation or to locate potential areas for shallow auguring. These records, plus the CEGB borehole record was used to inform the DANDO shallow borehole drilling campaign.

2.5 Optically Stimulated Luminescence Dating

Luminescence dating is a relatively new chronological technique. It is the only dating method that extends from the present day into the Mid-Pleistocene (125,000 – 780,000 years) (Lian, 2007). It is a technique that is growing in application, not only in earth science, but also in other areas like archaeology where it is used in ceramic dating, fire history and mapping, and artefact authenticity (Lian, 2007). OSL can be applied to glaciogenic sediments.

The main advantages of luminescence dating over other techniques are four-fold. 1) Luminescence dating does not require the presence of organic material; 2) it is able to date material that is older than the maximum range for radiocarbon dating; 3) it is relatively inexpensive, and 4) it provides a useful chronological tool when combined with other methods of dating that can support interpretations of depositional history e.g: the degree of partial bleaching or the presence of bioturbation (Lian, 2007).

Principles of luminescence dating

Luminescence dating is a radiometric dating technique that measures the accumulation of unbound electrons in shielded or buried mineral grains such as feldspar and quartz (Murray and Wintle, 2000; Lian, 2007). Unbound electrons are created from the decay of unstable isotopes, which then act as electron donors. Some are captured and released unnoticed, but others are encounter defects within the lattice of the grain and become trapped (Duller, 2008). Over time, the number of trapped electrons increases (Lian, 2007). Since there are a finite number of traps, continued exposure to radiation means that once the traps are filled, any additional electrons cannot be stored. This is known as grain saturation. The time it takes for a grain to reach saturation depends on the number of traps in a grain, amount of radiation and the rate at which it is received. If a grain is exhumed and exposed to light or becomes heated, the trapped electrons are released and recombine producing light or heat (Figure 2.5.1). This results in ‘bleaching,’ ‘zeroing,’ or ‘resetting’ of the grain until it is reburied whereby the accumulation of free electrons starts again (Lian, 2007).

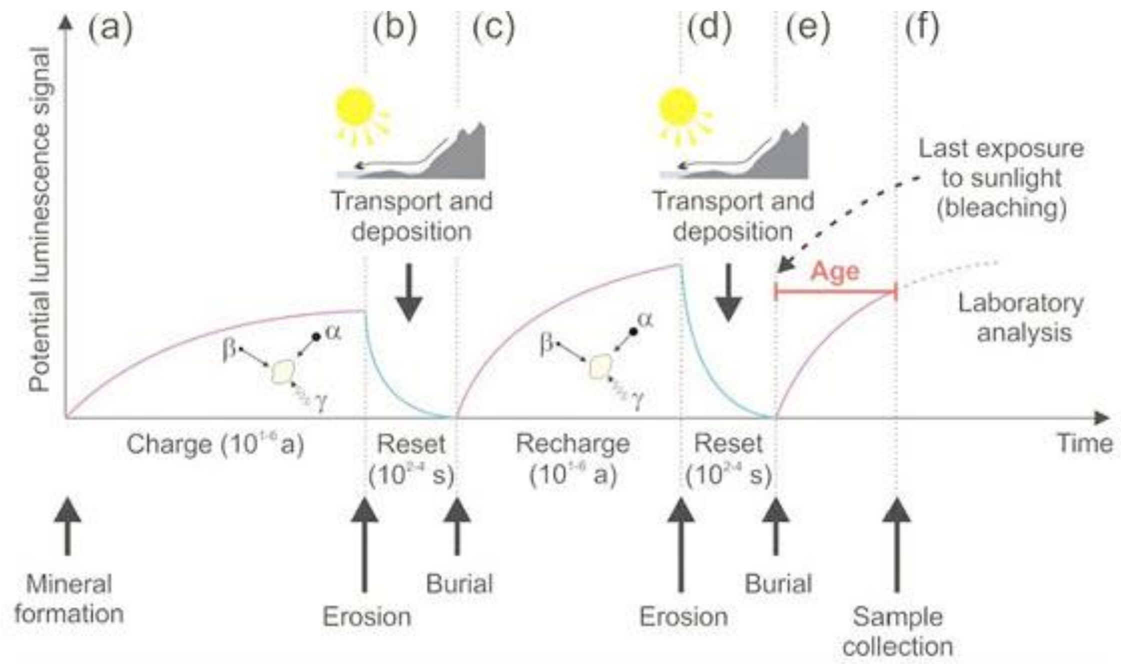


Figure 2.5.1: The luminescence cycle (Mellet, 2013). A) Signal builds in the sediment over time as the sediment is exposed to background alpha, beta and gamma radiation. B) The sediment is disturbed through erosion and/or transport and is exposed to sunlight whereby the electron traps holding the store radiation are emptied. C) The sediment is redeposited and the accumulation of background radiation is restarted. D) The sediment is disturbed and becomes zeroed again. E) The sample is reburied and the signal builds up once again. E) The sediment is sampled without exposure to light. F) The sample is prepared and analysed in darkroom conditions for age estimation.

To recreate this process for dating, samples are stimulated in a controlled environment either through heat (thermoluminescence or TL) or light (optically stimulated luminescence or OSL, infrared stimulated luminescence or IRSL). The techniques of OSL and IRSL are more common, with quartz grains used for OSL and feldspar grains used for IRSL. The age of a sample is calculated using the following equation:

Equation 2.5.1

$$Age\ (years) = \frac{Equivalent\ dose\ (Gy)}{Dose\ rate\ \left(\frac{Gy}{year}\right)} = \frac{D_e}{D_r}$$

The equivalent dose, or D_e, is a laboratory dose created to replicate the natural dose of the grain since the initial bleaching event. The dose rate (D_r) is calculated by sampling the surrounding sediment and calculating the concentration of radioisotopes present or by direct measurement while in the field using a gamma spectrometer. The resultant calculated age has an associated uncertainty due to systematic and random errors in the D_e, as well as differences in water content and mineral composition (Fuchs and Owen, 2008).

The mechanisms behind the lattice traps in quartz grains are complex. Bailey *et al.* (1997) identified three first-order decay components referred to as "fast," "medium," and "slow." The difference between these components is the rate at which the traps are emptied, with the "fast" component being the easiest to bleach under optical stimulation. This is the component most suitable for luminescence dating as, while fast, it also bleaches well, leaving little residual signal (Murray and Wintle, 2000). Murray and Wintle's (2000) Single Aliquot Regeneration (SAR) protocol assumes most quartz signals are dominated by the fast component. Where this is not the case, the SAR method may not be as effective for measurement (e.g. Thrasher *et al.*, 2009).

Application in glaciogenic sediments

The application of luminescence dating to glaciogenic sediments has been substantially reviewed by Lian and Roberts (2006), Lian (2007), Duller (2008), Fuchs and Owen (2008), Thrasher *et al.* (2009). The glacial environment is complex due to the number of sedimentary processes involving erosion, redeposition, transportation, and deformation, and many units of glacial sediment are highly variable in clast size, shape, and mineralogy.

The introduction of the luminescence dating and optically stimulated luminescence technique to glacial sediment has revolutionised Quaternary dating methods. Until recently, there were few techniques that allowed for the direct dating of glacial deposits and researchers were often limited to searching for entrenched organic matter suitable for radiocarbon dating. Radiocarbon dating organic materials within glacial sediment, however, is difficult due to the nature of deposition – ice-contact and proglacial environments do not preserve organic matter well – and potential for age-overestimation from older organic material. Furthermore, the short half-life (5730 ± 40 yr) of ^{14}C limits the dating range to Late Pleistocene and Holocene deposits, and requires complex calibration based on carbon fluctuations over time (Fuchs and Owen, 2008).

Other numerical dating methods, such as magnostratigraphy, tephrachronology, dendrochronology, varve chronology, potassium-argon, argon-argon, uranium-series, and fission track dating have limited use in glacial sediments and often require complex calibrations, involve rare materials, or must be matched with known stratigraphic markers (Walker, 2005). Even the increased popularity of terrestrial cosmogenic nuclide (TCN) dating has not been able to contribute to glacial sediment chronology with as much precision and accuracy as luminescence dating due to the underlying assumption that a surface has been stable with little to no weathering or shielding since exposure, something very hard to achieve within a glacial environment. Moreover, difficulties with age-models often produce age underestimations where many glacial environments occur such as high-altitude environments (Gosse and Phillips, 2001; Putkonen and Swanson, 2003; Balco *et al*. 2008).

Luminescence dating, conversely, can be applied to a plethora of terrestrial sediments, ranging in ages from 10^1 to 10^5 years (Fuchs and Owen, 2008). The direct dating of sediment is particularly useful to glacial environments where many of the sedimentary processes drive landform formation. Furthermore, the nature of luminescence dating allows for researchers not only to date the time of deposition, but also the duration of events, which can provide valuable information into the nature and scope of glacial sedimentary processes.

The nature of glacial sediment is diverse. Grain sizes range from boulders tens of meters in diameter to fine silt and clay particles. These grains are transported in many ways and through several processes that can affect the resulting light exposure to reset grains. Subglacial landforms, such as eskers and drumlins would not have been aerially exposed at the time of deposition and therefore have a much lower chance of containing clasts that may have been reset by light. Glacial moraines, deformation tills, and subglacial meltwater channels, too, are unlikely to produce adequately bleached sediment since the probability of light exposure is extremely low (Figure 2.5.2). Conversely, sediment that is deposited atop glaciers, supraglacial, does have high potential to be reset by sunlight, but the transport and subsequent deposition of the sediment makes dating more complicated. Some sediment may have been deposited supraglacially, yet there is still a chance that this sediment is not reset due to deposition during winter months or at night; immediate cover by other sediment, ice or snow, or by mixing with subglacial sediment making identification of the supraglacial grains more difficult. Along with these complications, any mass movement of supra or subglacial sediment reduces the potential for bleaching (Figure 2.5.2; Lian, 2007; Rhodes, 2011).

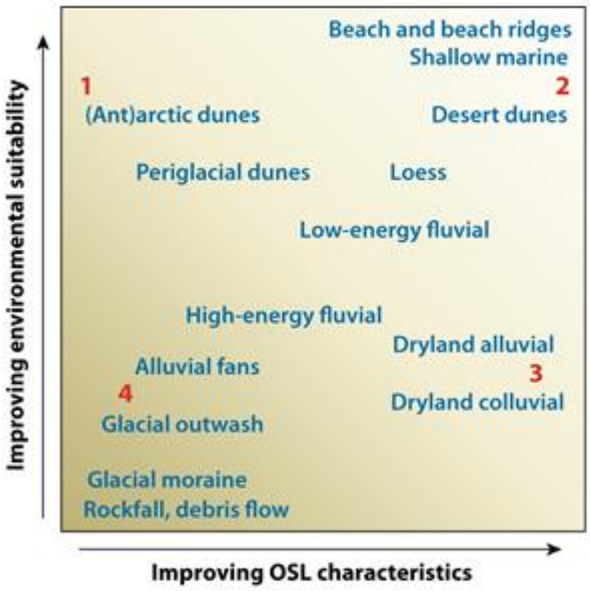


Figure 2.5.2: Potential bleachability for sediments depending on environment (Rhodes, 2011). Dryland sediments, such as dunes, have a higher chance of bleachability due to aeolian erosion increasing potential light exposure. Sediments contained within ice-proximal sediments are least likely to be reset and more likely to be affected by partial bleaching.

Proglacial sediment, or sediment that is deposited or carried from the front of the glacier, is far more conducive to OSL dating (Thrasher *et al*., 2009; Livingstone *et al*., 2012). This is due to the transport mechanisms washing or transporting the sediment away from the glacier margin, which increases potential for exposure to sunlight (Lian, 2007; Thrasher *et al.*, 2009; King *et al*., 2014). Many studies (Duller, 2006; Klasen *et al*., 2006, Alexanderson and Murray, 2007, Bøe *et al*., 2007; Prusser *et al*., 2007; Thrasher *et al*., 2009; Alexanderson and Murray, 2012; Carson *et al*., 2012; Gaar *et al*., 2013; King *et al*., 2014; Wyshnytzky *et al*., 2014) have shown that bleaching potential increases with distance travelled from the ice margin. Furthermore, sediment deposited and reworked within shallow aqueous environments, such as braided rivers and sandurs (Thrasher *et al*., 2009) have a much higher potential for bleaching than sediment carried in temperate deep river channels due to the difference in depth and turbidity of the water column. Shallower and slower fluvial environments, like a braided stream, allow more opportunity for the reworking and exposure of grains to sunlight, whereas a turbid water column is more likely to prevent attenuation from sunlight due to the high sediment load. Grain sizes suitable for OSL dating are also more likely to be transported as bedload than suspended load, further reducing the potential for bleaching (Thrasher *et al*., 2009). Bedload sediments are also more likely to be coated with fine grained clays or amalgamated into clusters of sediment, making light penetration difficult.

Luminescence Methodology

Luminescence measurements are usually taken at the aliquot, small aliquot, or single grain level. At aliquot level, there are approximately 2000 grains per disc; small aliquots hold between 200 and 20 grains, and single grain is just one grain (Duller, 2008). For environments where incomplete bleaching is suspected, small aliquot or single grain measurements are preferred as the variation that occurs with these environments can be muted (due to averaging) in measurements with a large number of grains (Figure 2.5.3). Typically, 95% of the signal observed in a sample comes from 5% of the grains (Duller, 2008), which means that single grain measurements are typically best for environments where partial bleaching is a problem.

Following work by Evans *et al*. (2017) at East Heslerton, a site situated within the Vale of Pickering that has comparable sediment to the sites used in this study, a comparison between single grain (SG) dating and single aliquot (SA) dating showed marginal differences between the estimated D_e. Evans *et al*. (2017) reported each sample the D_e replicates were normally distributed and showed an over-dispersion (OD) of 32-44% for SG and 22-38% for SA. Dose recovery experiments on sample Shfd13055 showed an OD of 29% for SG and 11% for SA suggesting that the OD over 20% was not from partial bleaching but other intrinsic factors (Roberts *et al*., 2005; Thomsen *et al.*, 2005; Jacobs *et al*., 2006). Since the grains used in Evans *et al*. (2017) study were considered well bleached, despite the OD over 20% (Olley *et al*., 2004), the decision was made to use small single aliquot over single grain for the present research due to the lower uncertainties and better signal-to-noise ratio in single aliquot data. The exception to this is where very restricted datable material was available.

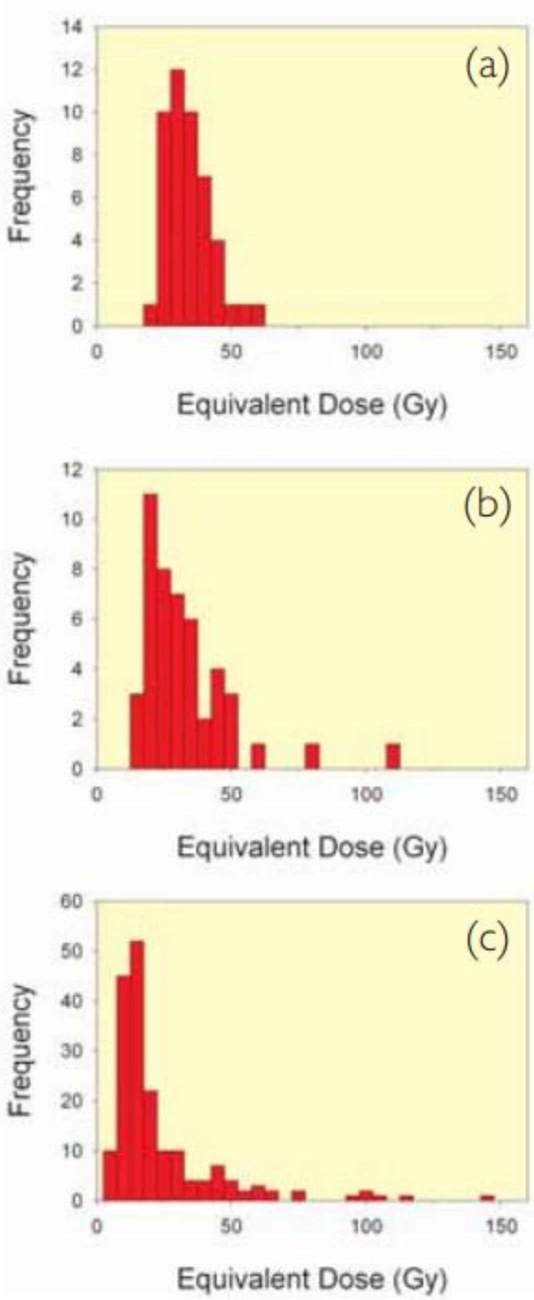


Figure 2.5.3: Distribution of equivalent dose (D_e) from A) aliquot B) small aliquot and C) single grain measurements. Note the channges in the distribution of of D_e. In environments where partial bleaching is a conern, aliquot size measurements may mute the extent of partial bleaching. Small aliquot (B) or single grain (C) show variation in D_e much more clearly (Duller, 2008)

Feldspar vs quartz

While both quartz and feldspars are abundant in glaciolacustrine and glaciofluvial deposits, the decision to use quartz related to two issues: 1) the rate at which feldspar bleaches is slower than quartz (Figure 2.54; Rhodes, 2011) and 2) feldspar has issues with anomalous fading, which has been shown to under-estimate age (Wintle, 1997; Huntley and Lamothe, 2001; Thomas *et* *al*. 2003). Since most of the sand-sized sediment within the Vale of Pickering comes from glaciofluvial deposits, quartz was preferred due to the faster rate of bleaching since sediment in the Vale of Pickering may have been exposed to light for only a brief period. Secondly, since quartz is abundant within the Vale of Pickering and the grains at East Heslerton (Evans *et al*., 2017) behaved well, quartz OSL was considered appropriate.

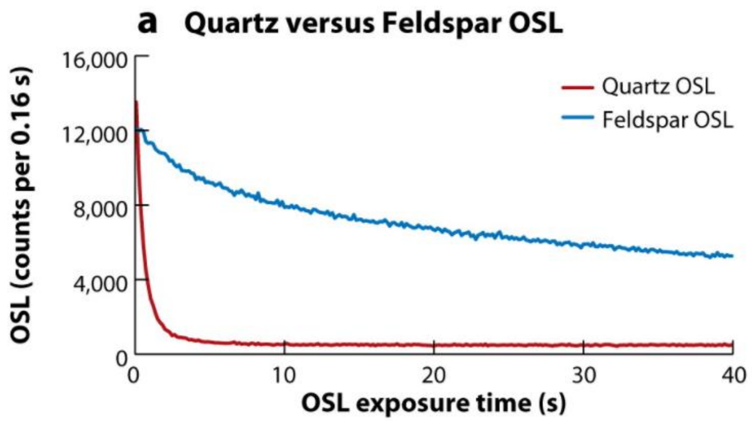


Figure 2.5.4: Quartz vs feldspar OSL decay (Rhodes, 2011)

Table 2.5.1: Comparison of advantages and disadvantages between quartz and feldspar

Quartz		**Feldspar**	
Advantage	**Disadvantage**	**Advantage**	**Disadvantage**
Highly resistant to weathering	Relatively low OSL intensity; some quartz samples do not emit measurable OSL; signal may consist of thermally unstable components	OSL saturates at higher radiation doses than does that from quartz	Weathers more readily from the environment than does quartz
OSL signal bleaches more rapidly in sunlight than that from feldspar	OSL saturates at lower radiation doses compared to that emitted from feldspar	OSL intensity may be orders of magnitude higher than that emitted from quartz	Suffers from anomalous fading
Does not appear to suffer from anomalous fading	Thermal transfer can be higher in quartz than in feldspar	IRSL can be stimulated preferentially in unseparated mineral mixtures (polymineral samples)	Difficult or impossible to correct for sensitivity change in regenerative-dose data when using SAR

Equivalent dose determination

Equivalent dose determination assumes that the changes in sensitivity during the main measurement of the OSL signal can be observed by repeated measurements, which are in turn displayed in a sensitivity-corrected dose-response curve (Murray and Wintle, 2000; 2003). The D_e can then be interpolated between the dose-response curve and the natural OSL signal (Figure 2.5.5).

The determination of equivalent dose is arguably the most important and time-consuming aspect of OSL measurements. The D_e is an equivalent to the natural environmental dose rate because it is measured in the laboratory using only beta radiation (a calibrated $^{90}Sr/^{90}Y$ beta source), whereas in the natural environment the dose rate is comprised from a mixture of alpha and beta particles and gamma and cosmic rays (Lian, 2007). Techniques for determining D_e have changed over the years from the less reliable additive dose method to the regenerative dose method (Duller, 1995). To combat inaccuracies due to grain sensitivity from repeated stimulations in the regenerative dose method, the Single Aliquot Regeneration (SAR) protocol (Table 2.5.2), developed by Murray and Wintle (2000; 2003), has been widely adopted and is considered the preferred method for establishing D_e. It follows the same procedure as the regenerative dose protocol but incorporates an additional standardised dose (test dose) and measurement after each cycle allowing for close monitoring of sensitivity changes (Murray and Wintle, 2000). To establish whether sensitivity was adequately adjusted, Murray and Wintle (2000) applied regenerative doses non-sequentially. They observed that the order in which doses were administered had a negligible effect on the results, meaning that residue from prior treatments was essentially eliminated. A study by Murray and Clemmensen (2001) confirmed this to be the case.

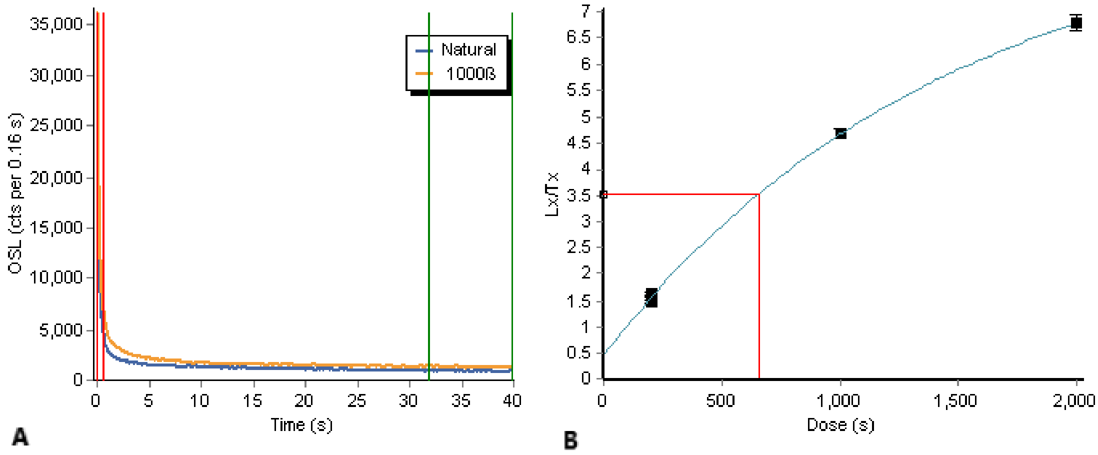


Figure 2.5.5: A shows a shinedown curve, which can be an indication of how quickly the quartz traps empty. B is a SAR growth curve of interpolated doses (here 1000s and 2000s)

The SAR technique is used to obtain a D_e value in the following way (Table 2.5.2): In the first instance, a measurement of the 'natural' signal is taken, and the grain is then subsequently bleached. This is followed by repeated cycles of increased dosing and measurement with the sample preheated between measurements to empty the thermally unstable traps. When complete, a dose response is interpolated on to a growth curve (Figure 2.5.5, panel B) using a series of regeneration points that bracket the natural dose to give a D_e.

Using the SAR protocol

The SAR protocol is given in Table 2.5.2. Before using the SAR protocol, two factors must be determined: preheat temperature and regeneration dose point values. Both factors vary between samples and need to be established for accuracy in measurement.

The need for preheating comes from the presence of a small amount of charge held in the shallow, thermally-unstable traps that must be removed as they are generally empty during burial but are filled once sediment is dosed in the laboratory as part of the SAR protocol so can contribute erroneously to the OSL measurement (Lian, 2007). To remove this extra charge, samples are preheated, but the correct preheat must be established or further errors in establishing D_e (i.e: age over estimation) can occur. Optimal preheat temperatures for sites were derived from a dose recovery preheat test (Table 2.5.2). Eighteen aliquots of each sample were made, which were divided into five groups of three aliquots. Each of the aliquot groups were bleached at room temperature using green LED light for 100s and given a known lab dose.

Table 2.5.2: Standard single aliquot sequence for the SAR protocol (after Murray and Wintle, 2000)

Step	Treatment	Observed
1	Give dose, D_x*	
2	Preheat (180-220°C for 10 s, heating rate of 2°C s^{-1})	
3	Stimulate at 125 °C for 60 s	L_x†
4	Give test dose, D_t‡	
5	Cut-heat (160°C for 0 s, heating rate of 2°C s^{-1})	
6	Stimulate at 125 °C for 60 s	T_x†
7	Return to step 1	

* For the natural sample, applied dose = 0 Gy. The final dose will be equal to the first regeneration dose, allowing calculation of a recycling ratio. The fourth regeneration point is always 0 Gy allowing recuperation to be observed

† Based on the first 0.8 s of stimulation with a background (based on last 3 s of stimulation) subtracted. The first part of the signal shows increased sensitivity to light and a better signal to noise ratio (Duller and Augustinus, 1997)

‡ A standard test dose of 50 s was selected as appropriate for all samples, but it should be noted that this value is equivalent to a value of between approximately 3 and 6 Gy, depending on the dose rate of the machine used. Test dose size has been found to have no significant effect on D_e or level of recuperation (Murray and Wintle, 2000)

Each group was then assigned one of five temperature doses, between 160°C to 260°C, and measured using the SAR protocol. The most suitable preheat temperature was chosen where the recycling ratio was closest to unity and the resultant recovered dose from the SAR measurements was the nearest to the given laboratory dose (Figure 2.5.6).

Table 2.5.3: Preheat temperature used for each sample

Lab code	**Site**	**Time (s)**	**Preheat (˚C)**
Shfd16018	Caulkley’s Bank	100	220
Shfd16019	Slingsby	100	180
Shfd16020	Ings Farm, Yedingham Quarry 1 OSL2	100	220
Shfd16021	Ings Farm, Yedingham Core 4	100	260
Shfd16022	Wellfield Farm, Salton Core 3	100	220
Shfd16023	Ings Farm, Yedingham Core 1	100	260

Shfd16024	The Pottery, High Marishes Core 3	100	220
Shfd16025	Ings Farm, Yedingham, Quarry 1 OSL 1	100	260
Shfd16026	Ings Farm, Yedingham, Quarry 2 OSL 1	100	260
Shfd16154	Ings Farm, Yedingham Quarry 2 OSL	100	180
Shfd16157	Hunmanby Gap, OSL 1	100	180
Shfd16158	Hunmanby Gap OSL 4	100	180
Shfd16159	Cayton Bay	100	180
Shfd16161	Eden Camp	100	180
Shfd16162	The Pottery, High Marishes Core 1	100	180
Shfd16163	Ings Farm, Yedingham Core 2	100	180
Shfd16164	Wellfield Farm, Salton Core 1	100	180
Shfd16165	Wellfield Farm, Salton Core 2	100	180
Shfd16174	The Pottery, High Marishes Core 2	100	180

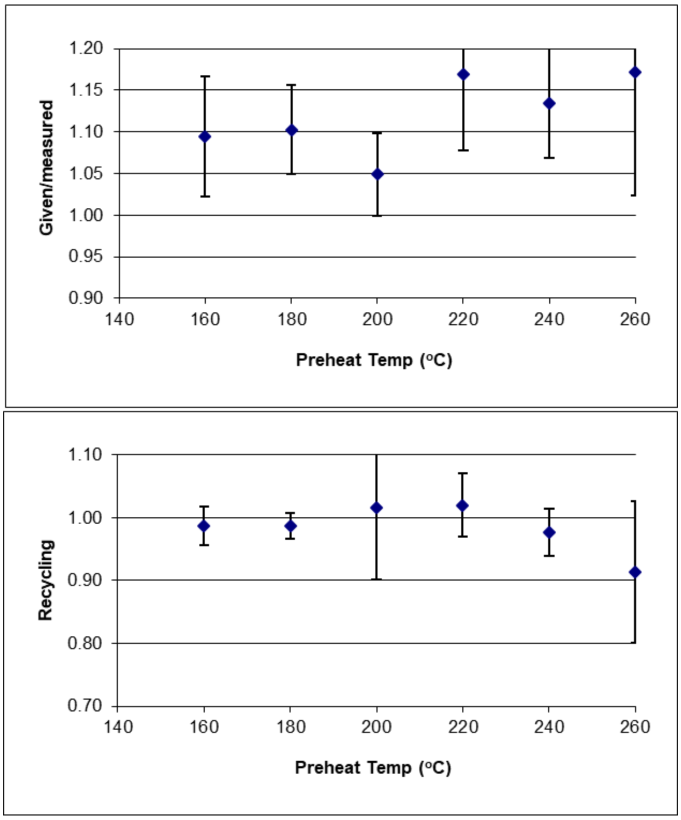


Figure 2.5.6: Shfd16157 Preheat test with recycling ratios. A preheat of 180˚ was used (as opposed to 200˚) because although the measured dose is higher, the error on recycling is very small.

The SAR protocol contracts a growth curve based on a range of regeneration values upon which the natural signal is interpolated (Figure 2.5.5 panel B). The regeneration points need to be within range of the natural dose. The rangerfinder test uses three aliquots per sample and an abbreviated version of SAR (three regeneration points only and no recycling) to establish roughly where the palaeodose will intercept the SAR growth curve so regeneration points can be optimised.

Instrument specifications

All measurements were made on a TL-DA-15/18 automated Risø TL/OSL reader with a Hoya U-340 filtered photomultiplier tube and blue LEDs for stimulation. Laboratory doses were applied using a calibrated $Strontium^{90}$ beta source.

Sample collection and preparation

In the field, samples were collected either by driving lightproof PVC tubes of 5 cm diameter and 12 or 22 cm in length, depending on the availability of sediment, into sediment exposures. At two sites, a Dormer engineering drill was used to drill into sediment where a light-proof PVC tube was inserted into a light proof, metal sediment extractor. Opaque-lined cores were also extracted at DANDO sites taken concurrently with clear liners (Section 2.3).

Table 2.5.4: The effect of different moisture values on estimated OSL ages for three samples within the Vale of Pickering with percent change shown on age estimations. *Ages calculated using CAM

Sample Name	**Water (%)**	**Age (ka)***	**% change**
Shfd16023	10	16.6 ± 1.0	-
Shfd16023	15	17.7 ± 1.1	6.6
Shfd16023	20	19.0 ± 1.1	7.3
Shfd16026	10	39.7 ± 4.4	-
Shfd16026	15	42.0 ± 4.6	5.8
Shfd16026	20	44.6 ± 4.9	6.2
Shfd16159	10	38.0 ± 3.0	-
Shfd16159	15	40.5 ± 3.2	6.6
Shfd16159	20	43.3 ± 3.5	6.9

The OSL samples were extracted in a dark room designed for OSL sample preparation following the methods outlined in Bateman and Catt (1997). For tube samples, 2.5 cm from each end was removed and bagged for the calculation of moisture content. Samples were then dried in an oven at 30°C overnight (or until completely dry) and treated with 10% H_2O_2 and 10% HCl to remove organic matter and carbonates, respectively. Each sample was wet sieved through a nest of sieves ranging in mesh size from 90µm - 250µm to attain a narrow size fraction. To separate the quartz-rich fraction from heavy minerals, each sample was placed in sodium polytungstate with a specific gravity of 2.7g $^{cm-2}$ Following removal of heavy minerals, the samples were etched in HF for one hour to removed feldspars and remove the alpha-irradiated skin (~20 µm) from each quartz grain. Samples were then given a final dry sieve at the smallest fraction kept at each sample to remove any small fractions and any residual feldspars.

Samples from the DANDO cores were sampled by examining the stratigraphy of the open cores that were taken concurrently. Sites were chosen with a preference for massive sand deposits. Cores were sawn with a handsaw in the dark room and processed as above.

Aliquots of each sample were made under darkroom conditions. Up to 48 aliquots were made per sample. Quartz grains were mounted as a mono-layer using Silkospray silicon glue spray sprayed through a 2 mm mask on to 9.6 mm diameter stainless steel disks. Each aliquot contained approximately 100 grains of quartz.

Dose Rate determination

For each sample total dose rates were based on the concentrations of U, Th, K and Rb. To determine these, samples were milled and sent to SGS laboratory, Canada to be measured using inductively couple plasma mass spectroscopy (ICP-MS). Resultant concentrations were converted to dose rates attenuating appropriately for density and grain size. Where exposures allowed, the *in-situ* gamma dose rate was measured with an EG & G MicroNomad gamma spectrometer and this data used for the gamma dose rate. The contribution from cosmic rays was determined by formulae presented in Prescott and Hutton (1994). Where heterogeneous sediment was sampled, and gamma spectrometry was not possible additional samples from sediments adjacent to those samples were analysed using ICP and gamma dose rate contributions from these units to the sample modelled using the data presented in Aitken (1985).

The calculation of moisture values for OSL samples is one of the largest causes of uncertainty in age estimation (Aitken, 1998; Pressuer *et al*., 2009). Areas that have complex environmental histories, where groundwater levels have changed dramatically, present an especially unique challenge in calculating moisture for OSL dating as current moisture values do not reflect past hydrological change (Table 2.5.4; Figure 2.5.7). In an effort to compensate for these changes, OSL ages need to be calculated with a moisture value that reflects these changes and not purely based on present day values. Table 2.5.4 shows three samples from the Vale of Pickering showing the effect on age calculations due to variations in moisture content. In the samples shown, an over estimation of >7% is possible if moisture values are incorrect.

The degree of saturation within a deposit depends on three factors 1). The size of the pore space, 2) the ratio of air to water in the pore space and 3) the type of sediment (Juschus *et al*., 2007; Presusser *et al*., 2009). The size of the pore space is related to the sediment type and to the ratio of air to water within that pore space. For example, sand (above water table) will have a larger and drier pore space than a similarly located clay, indicating that, since water also absorbs ionising radiation, drier samples will have a higher effective dose (Pressuer *et al*., 2009). Furthermore, consolidation of lacustrine clays reduces pore space and due to this determining D_e in lacustrine sediments often leads to age overestimations (Juschus *et al*., 2007).

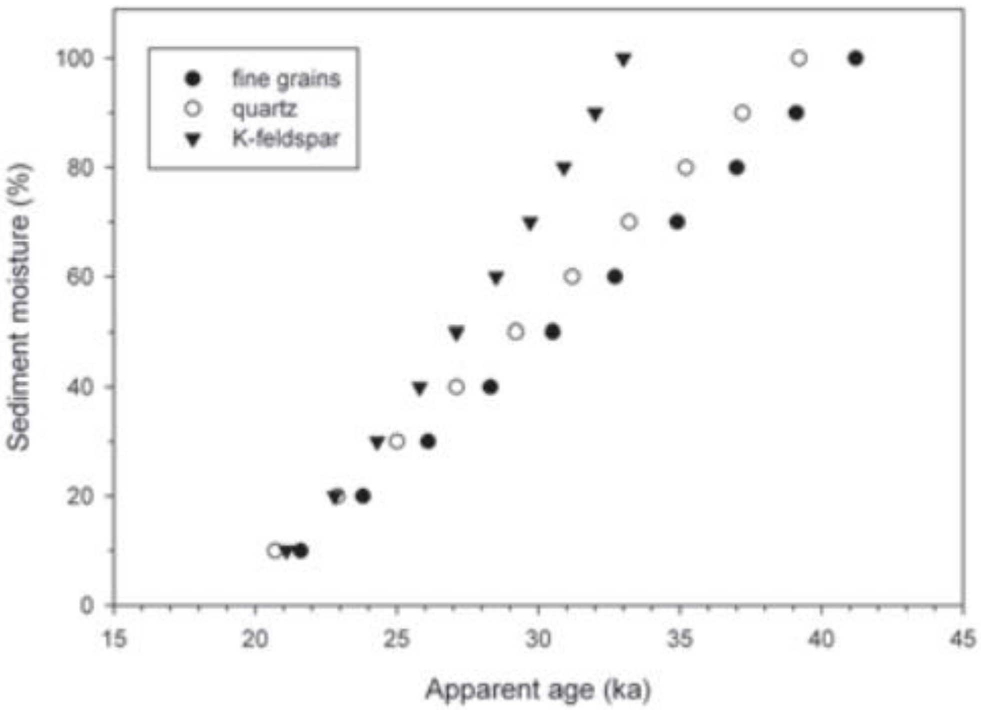


Figure 2.5.7: Impact of sediment moisture on calculated age (from Preusser et al., 2009)

Climatic changes in the Vale of Pickering since the LGM have reduced water table levels from potentially as high as 70 m OD (Kendall, 1902), to 45 m OD (Straw, 1979) as the glacial ice retreated, and further lowering during the onset of the Holocene to 24 m OD (Palmer *et al*., 2015). Figure 2.5.8 approximates these changes from a predominantly saturated environment (30 ka – 16 ka), during which the majority of samples were deposited, to an intermittent period (15 ka – 5 ka) where lake water, meltwater, and seasonal waters drained slowly out of the Vale. Current conditions in the Vale of Pickering are still predominantly marshy and wet (5 ka – present) although drier than the onset of the Holocene (Bearcock *et al*., 2016). This is evidenced by the towns of the Vale lining the dry 30 – 40 m lip where the Kimmeridge clay outcrops from under the superficial lacustrine deposits. Very few towns lie in the marshier central flat lands, but since Roman settlement, the marshes and meres have been drained for agricultural use (Sheppard, 1948). Sheppard (1948) estimates a post-glacial Lake Pickering around 20 m OD at 2,000 BC, with Seamer Carr described as an open body of water during the 15th century (Lincoln *et al.*, 2017). The current water table lies between 17 m OD in the east and 20 m OD in the west (Bearcock *et al*., 2016)

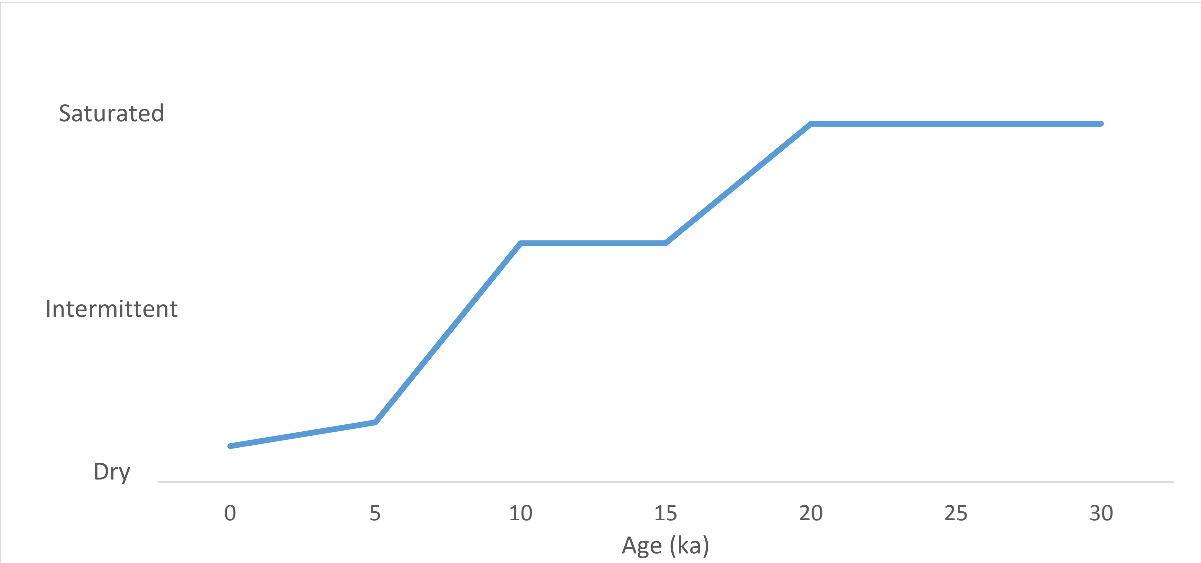


Figure 2.5.8: Approximation of water saturation in the Vale of Pickering. During the LGM, the Vale of Pickering was a large proglacial lake that slowly drained after the glacial ice retreated. Current land use has decreased the height of the water table, but it still lies between 1-2 m below the land surface in some areas, especially the eastern half of the Vale.

Other work involving OSL dating in the region (Bateman *et al*., 2015; Evans *et al*., 2017; Bateman *et al*., 2017) have assumed moisture values for sandy samples based on the environmental history of the region. Evans *et al.* (2017) assumed 20 ± 5% for samples below the water table and 10 ± 5% for those close to but above the water table. This approach has been used in this study but with a third option of 15 ± 5% added for those sandy samples that are currently at the transition zone but which for part of their burial history were definitely saturated under Lake Pickering for long periods of time. For the majority of samples collected in the Vale of Pickering estimated moisture content is similar to present day moisture content, especially in samples that are located in the intermittent zone; however, there are still some anomalies that need to be addressed. Three samples (Shfd16154, 16157, and 16159) are now very dry due to their exposure on sea cliffs; however, during deposition, these samples would have been fully saturated and likely remained so for some time before eventually draining as the ice retreated and so the current moisture value does not reflect the samples full moisture history (Table 2.5.5).

Table 2.5.5: Samples taken from the Vale of Pickering and their associated sedimentology and moisture history.

Sample	**Sediment Type**	**Present moisture (%)**	**Relationship to present water table**	**Relationship to past water table**	**Moisture used (%)**
Shfd16018	Clayey sand	11.7	Dry	Intermittent	10
Shfd16019	Sand	16.4	Intermittent	within	15
Shfd16020	Sand	20.0	Intermittent	Saturated	20
Shfd16021	Sand	15.6	Saturated	Saturated	20
Shfd16022	Sand	13.0	Intermittent	Intermittent	15
Shfd16023	Sand in clay	24.5	Saturated	Saturated	26
Shfd16024	Sand	16.6	Intermittent	Saturated	20
Shfd16025	Sand	13.6	Intermittent	Saturated	20
Shfd16026	Sand	5.5	Intermittent	Saturated	20
Shfd16154	Sand	7.0	Intermittent	Saturated	20
Shfd16157	Sand	3.5	Dry	Saturated	15
Shfd16158	Sand	3.7	Dry	Saturated	15
Shfd16159	Sand	4.6	Dry	Saturated	15
Shfd16161	Sand	16.1	Dry	Saturated	15
Shfd16162	Sand	8.5	Dry	Intermittent	10
Shfd16163	Sand	24.8	Saturated	Saturated	20
Shfd16164	Clayey sand	19.1	Intermittent	Intermittent	15

Shfd16165	Sand	20.5	Intermittent	Saturated	20
Shfd16174	Sand	21.3	Intermittent	Saturated	20

Shfd16023 is a sand sample washed out from a clay deposit that lies below the water table and has likely always been saturated, but the saturation value for sand would underestimate moisture content for a clay. Therefore, for clay-rich samples under the water-table, an experiment was constructed to determine this more precisely.

For this, an intact clay sample was removed from below the water table *in situ*, weighed, and dried in a low temperature oven (36°C) to preserve the clay's structural integrity. The sample provided a moisture content of 25%, which is thought to be close to the true saturation value for this sample. However, to check this the dried clay was coated in wax except for a small hole at the top of the sample. This was then set under a burette of slow dripping water to gradually replace all air-filled pores with water and attain a saturation value (Figures 2.5.9 and 2.5.10).

Figure 2.5.9: (left) Apparatus set up and (right) wax-coated clay sample

Table 2.5.6: Moisture values for clay saturation from 24 to195 hours

Time (hours)	Wax weight (g)	Wax+clay weight (g)	Wet weight (g)	Moisture (%)
24	7.01	45.6	52.57	18.06
48	7.01	45.6	55.13	24.70
120	7.01	45.6	55.27	25.06
144	7.01	45.6	55.72	26.22
168	7.01	45.6	55.49	25.63
192	7.01	45.6	55.53	25.73
195 (Final)	7.01	45.6	55.43	25.47

Figure 2.5.10: After 195 hours of wetting, clay sample is saturated

The outer wax surface was dried, and the sample was then weighed every 24 hours until the moisture content reached plateaux. The results (Table 2.5.6) show that the value of 20 ± 5% for sand would likely underestimate the age of the sample and a moisture value of 26 ± 5% for the clay is more accurate.

Statistical age models

Bailey and Arnold (2006) found that no single method of analysis is applicable to all samples, but it is possible to choose the most appropriate model based on the shape and degree of scatter of the D_e distribution. Since partial bleaching potentially could have affected most of the samples, three models were considered for determining ages: The Central Age Model (CAM), the Minimum Age Model (MAM), and the Finite Mixture Model (FMM).

The CAM (Galbraith *et al.*, 1999) uses a weighted mean based on the log of each D_e and is designed for well bleached samples with a normal distribution (Figure 2.5.11), although it does assume some variation in D_e (i.e.: overdispersion is >0) in D_e from natural variations in water content and microdosimetry.

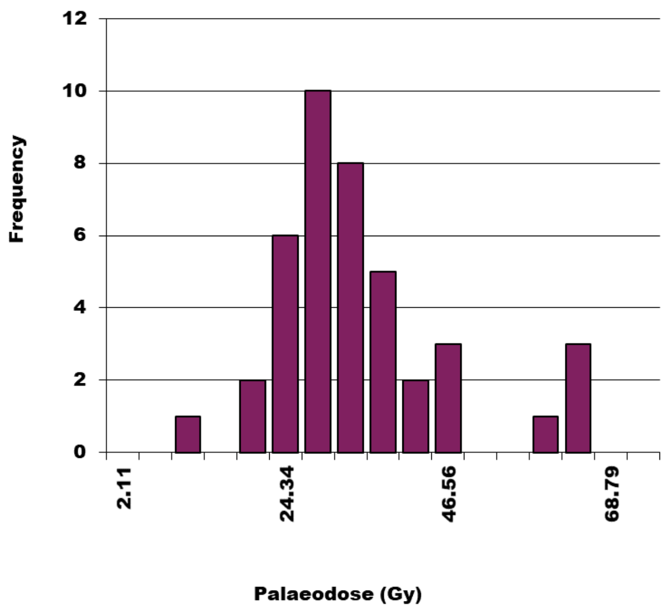


Figure 2.5.11: Distribution for a sample where CAM was applied. Once the two high De outliers are removed, the distribution is normalised and the CAM is the best statistical model for this data.

In glaciofluvial environments, however, partial bleaching can be present. This means that the distribution for a sample is non-normal and so statistical methods for recognising this have led to several models, including the 'lowest 5% '(Olley *et al.*, 1998), and 'leading edge' method

(Lepper *et al*., 2000; Lepper and McKeever, 2002), both of which have become uncommon due to the more commonly applied Minimum Age Model (Galbraith *et al*., 1999).

The minimum age model (MAM) (Galbraith *et al.*, 1999) attempts to define the minimum D_e acquired by the fully bleached proportion of grains within a heterogeneously-bleached D_e distribution (Figure 2.5.12). The distribution of D_e values is approximated by a truncated log-normal distribution, with the truncation point giving the estimate of D_e. Olley *et al*. (2004) found that the MAM provided a more accurate estimate of D_e than the CAM for sediments from aeolian, fluvial, and marine environments.

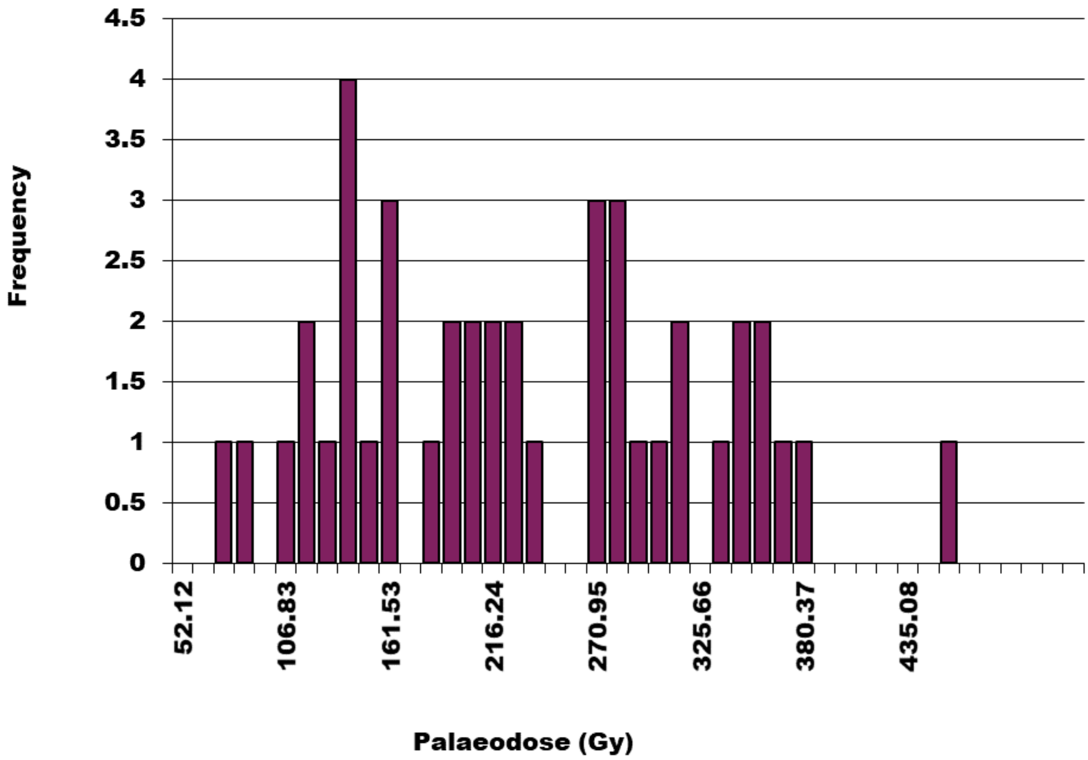


Figure 2.5.12: Distribution of a partially bleached sample. Note that the spread of data covers values from 50 Gys to 435Gys. There are many peaks indicating a multimodal distribution which indicates partial bleaching. The FMM or MAM model would be more appropriate for this sample.

The Finite Mixture Model (FMM) of Galbraith and Green (1990) identifies 'populations' (or clusters) of aliquots/single grains within the entire heterogeneously-bleached D_e distribution. Each component population of aliquots/single grains has its own D_e and error associated with it. Additionally, the Bayesian Information Criterion (BIC) is given, which is used to assess the 'best fit' number of populations to fit a dataset. BIC reduces to a minimum at the optimum number of component populations (Rodnight, 2006). The lowest D_e population identified through this process is thought to reflect the most well-bleached proportion of the sample.

Flow chart for OSL model decision making based on data distribution

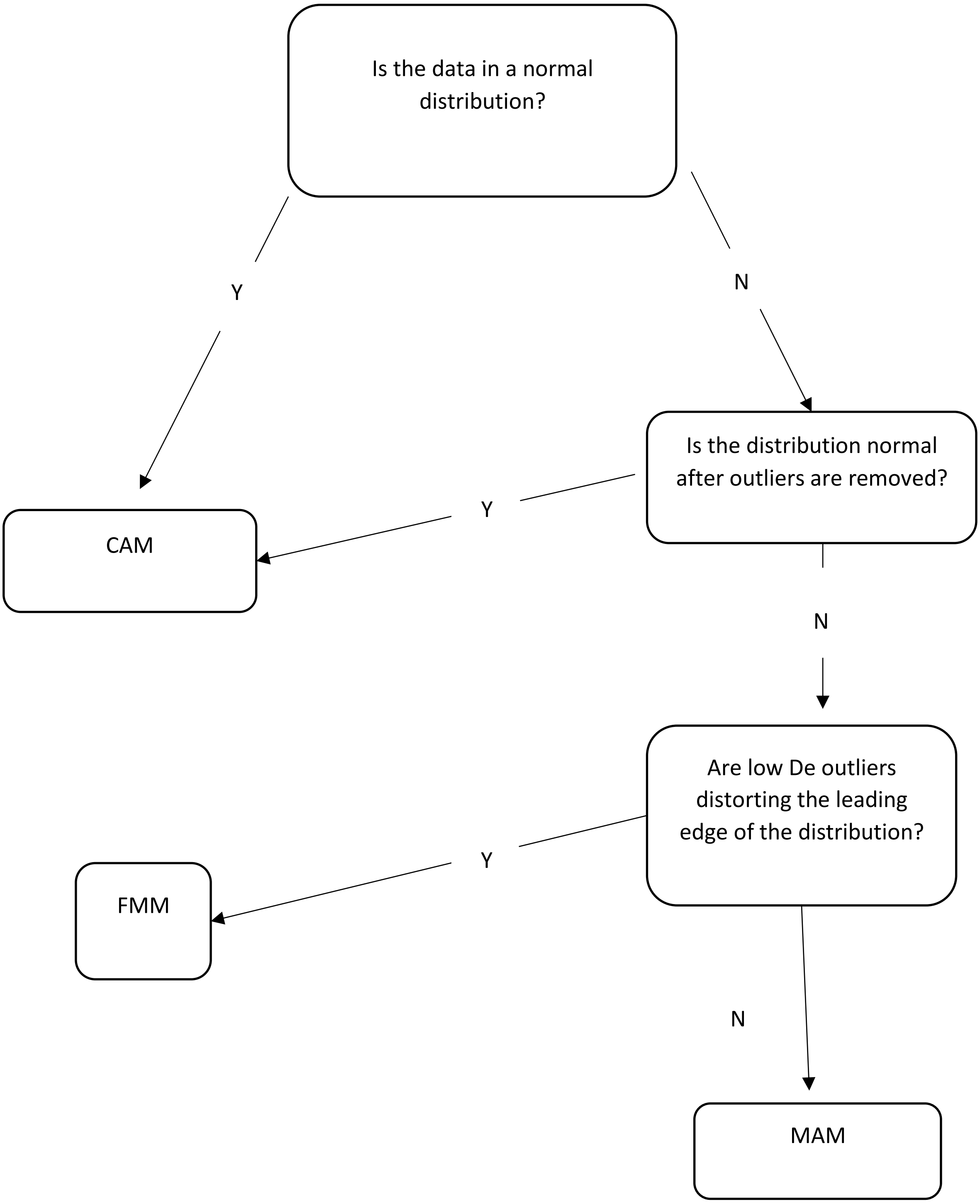


Figure 2.5.13: Model choosing flow chart. Samples with a normal distribution were analysed with the Central Age Model (CAM). Samples with a normal distribution after outliers were removed were also analysed under CAM. If low De outliers were not present and not the possible cause of the skewed distribution, samples were analysed using the Minimum Age Model (MAM). Samples with low De outliers that may indicate reworking of sediment were analysed using FMM and FMM components <10% of the total proportion were not considered.

In determining how to apply the models to the samples collected from the Vale of Pickering, observations on the spread of the palaeodose data was needed (Figure 2.5.13). Each of the models (CAM, FMM, and MAM) was applied to each sample. The CAM was not considered appropriate for samples with an OD (minus outliers) >25% as this often indicates samples with an incomplete bleaching problem. It was noted that often, when the CAM was the best model for the data, the FMM and the MAM would have provided a similar palaeodose. The MAM was most chosen for samples with multimodal distributions and high OD (Figure 2.5.12; Figure 2.5.14) The FMM was applied to samples (using a sigmab of 0.2) where partial bleaching was a possibility based on the high OD of D_e values, skew in the D_e data (most often a positive skew or a bimodal distribution where the frequency of higher palaeodose values were lower). It was particularly used where a few low D_e values would have caused problems for minimum age models. In such cases the lowest FMM De component with >10% of the data was used for age calculation purposes.

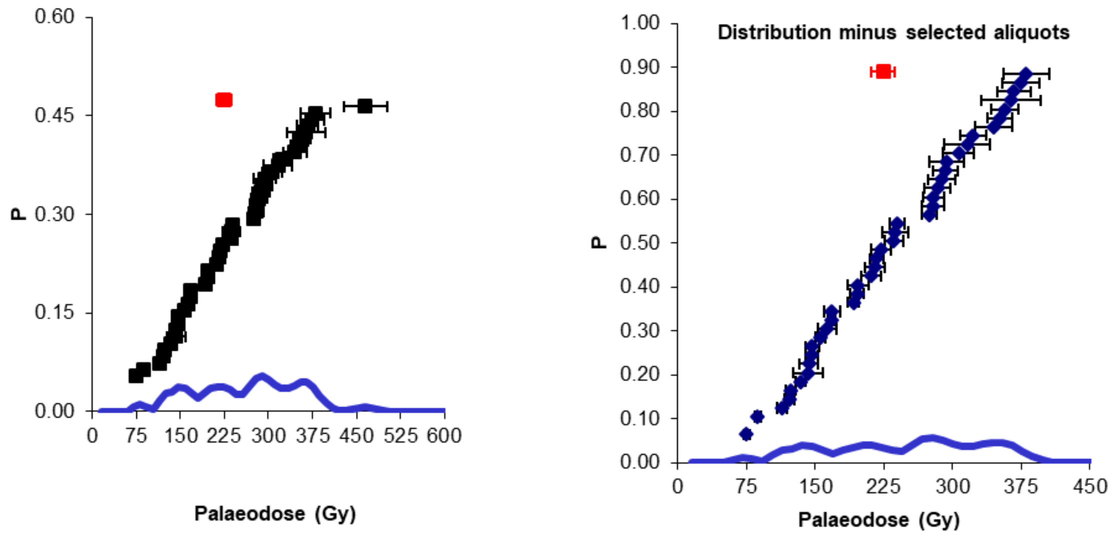


Figure 2.5.14: A sample where MAM is the best model. The distribution is multimodal, and the OD is 42.4% (41.3% minus outliers)

Statistical models for each sample were chosen based on the best fit of the model to the data presented (Section 3.1). Resultant ages using the selected D_e and the calculated sample total dose rate are reported in thousand years before date of sampling. These ages are burial ages and are reported with 1σ uncertainties.

One single grain sample was processed from one of the recovered DANDO cores, 6.2 meters deep as the opaque core was unable to reach the same depth. Sand from within a clay matrix core (with the outer 2 cm of core sediment discarded to limit contamination by partial bleaching) was washed out and prepared as above. The grains were mounted on a grain holder and stimulated individually using a green laser rather than LED. Measurements were made using SAR as outlined above. Results were also run through all three statistical models and analysed for the best fit for the data.

2.6 Lake level models

A series of flood fills were produced to model the effect of Lake Humber, the VoYL, and NSL on the extent of Lake Pickering and its drainage through the Kirkham Gorge. The link between the lake water height of Lake Humber and of Lake Pickering is crucial to understanding 1) The deglaciation of the VoYL and NSL due regional drainage of both lakes through the Humber Gap; 2) The creation of the Kirkham Gorge, through which Lake Pickering is thought to have emptied and the current course of the south-flowing River Derwent; 3) The timing of the reversal of the River Derwent, and 4) The maximum level reached by the waters of Lake Pickering before the Kirkham Gorge spillway was created (See section 1.4).

Modelling was completed in ArcGIS using the method outlined below (Figure 2.6.1). Lake Humber levels modelled were based on shapefiles available from BRITICE v2 (Clark *et al*., 2018). GIA data was provided by Dr. Sarah Bradley in the form of Kuchar plots (Kuchar *et al*., 2012) available as .XYZ files (Figure 2.6.1). Kuchar plots show the predicted solid earth depth at each time slice (Kuchar *et al*., 2012) and so it was necessary to add the lake height by assigning a value to each of the pixels inline with the predicted height for each Lake Humber shapefile. One of the limitations of this model is it does not incorporate overlying unconsolidated deposits or their geomorphology, which may differ to bedrock geomorphology.

Table 2.6.1: Lake Humber height (in m) associated with each Kuchar plot time slice (ka) (Figure 2.6.2).

Lake Level (m)	GIA time period (ka)
10	16
10	17
15	18
15	19
20	20
25	21
25	21.5
33	22
33	24

The shapefile of Lake Humber at various heights was added depending on time period (Table 2.6.1) and ice position based on Bateman *et al*. 2018 (Figure 1.3.6). The GIA adjusted DEM was then flood filled. This was repeated for each lake height and time period (Table 2.6.1).

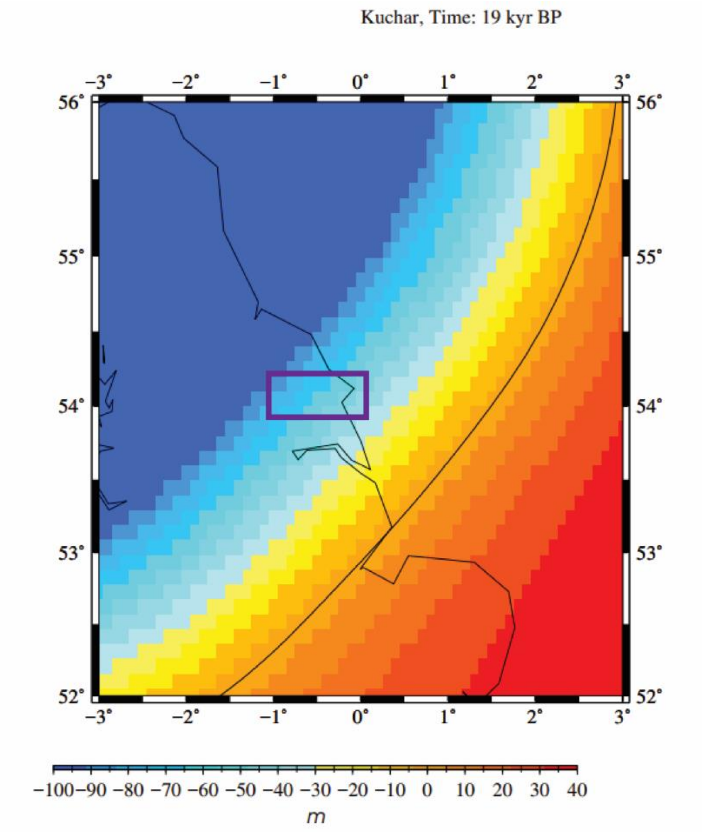
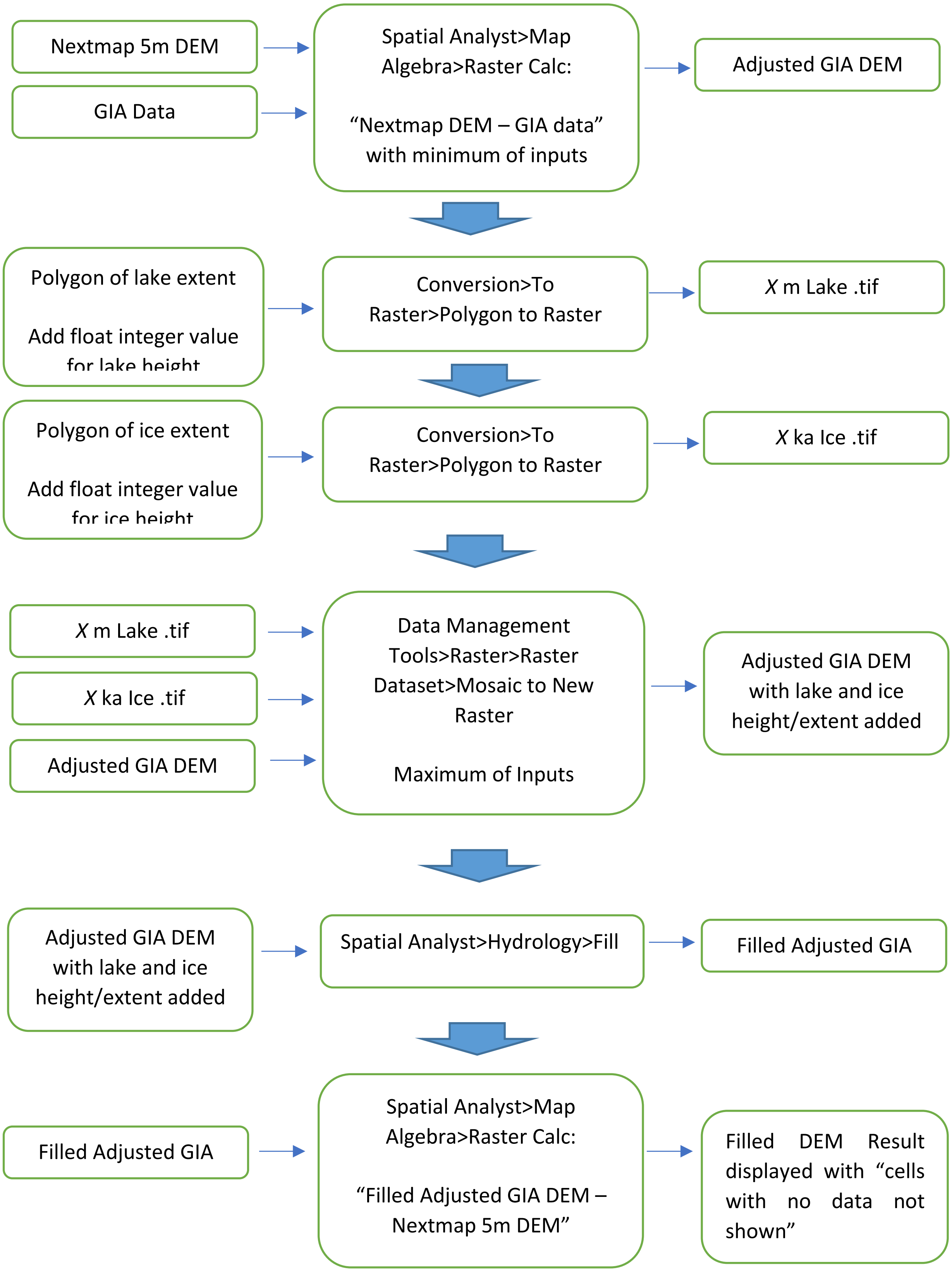


Figure 2.6.1: Example of one of the Kuchar plots of solid earth deformation based on models of ice thickness (Kuchar et al., 2012). Blue colours represent an increase in deformation, whereas red and oranges show uplift. The black line marks the point of equilibrium.

Figure 2.6.2: Flow chart for flood fill procedure. First step is to adjust current NextMap DEMs with GIA data. Polygons of ice position and stage of Lake Humber are added. The resulting TIFF is flooded and then a calculation applied to create a model of flood extent.



2.7 Summary

By using all the techniques listed above (OSL, geomorphic mapping, sediment characterisation, and modelling) to analyse and describe the environmental conditions in the Vale of Pickering, new data has been generated that can provide insight into the changing conditions since (at least) the LGM. These techniques cover environmental evolution from a very large scale, such as the entire Vale of Pickering, down to the crystals of the clay minerals in individual sediment packages.

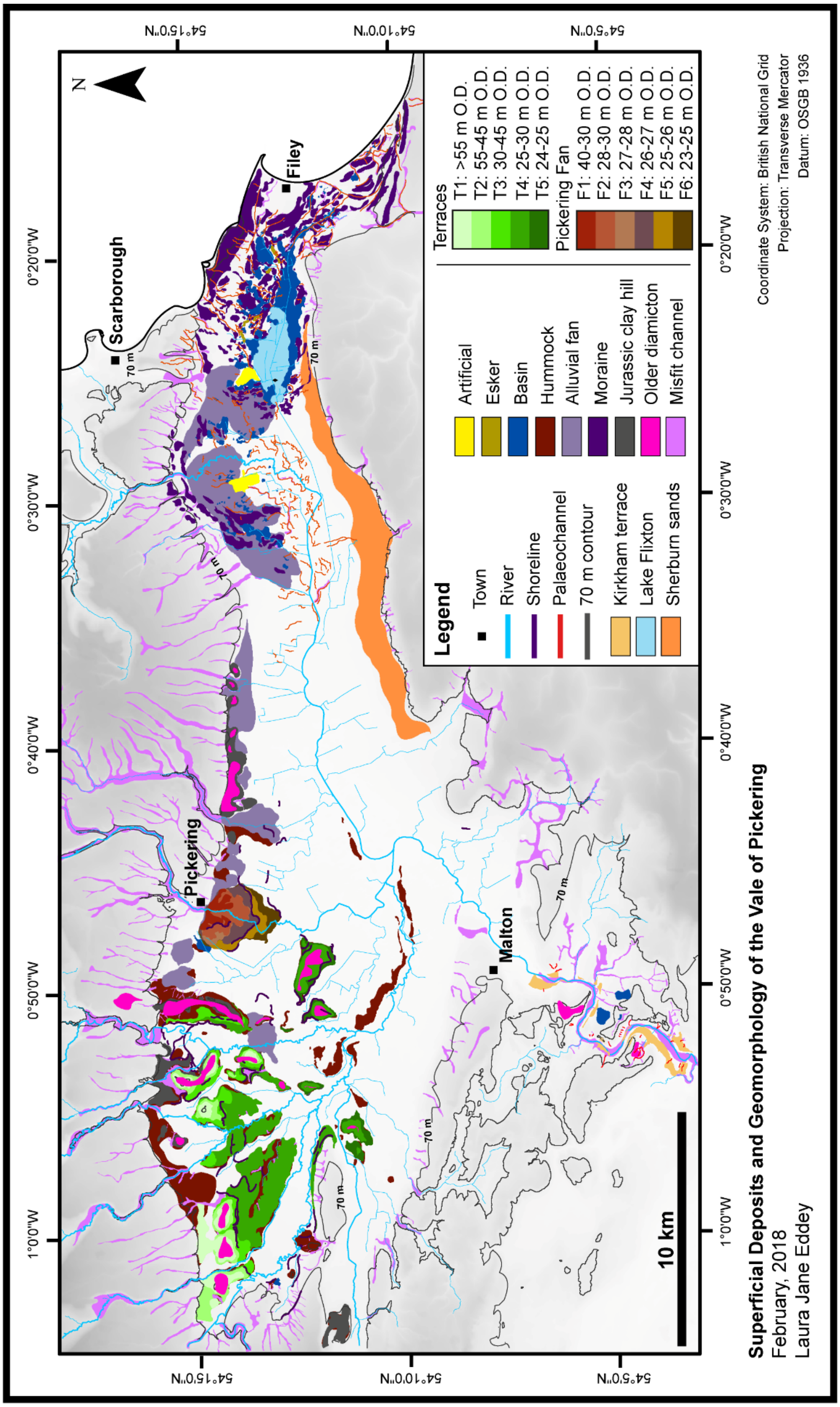


Figure 3.1: Superficial deposits of the Vale of Pickering. Presented is the complete geomorphic mapping. For explanations on how each feature was mapped see section 3.1

3.1 Geomorphic Mapping

Geomorphic mapping was completed as outlined in section 2.2. Landforms were examined using desktop mapping techniques, site visits, and from existing literature. Setting and relationship with other landforms was also considered in interpretation. There is always a degree of uncertainty in mapping and further investigation of some sites are necessary to establish their morphology. This is discussed in the following section where applicable.

Pre-Devensian landforms: Jurassic clay hills and older diamicton deposits

In several areas, most noticeably the western Vale of Pickering, spurs of Jurassic Clay break the surface of the Quaternary-aged sediments to form small, rounded hills (Figure 3.1.1). The Jurassic-aged marine clays (the Kimmeridge, Ampthill and Oxford) range from very thick, dark grey, shaley deposits to brittle, light grey, finely laminated clays, which, when wet, are very tenacious and sticky.

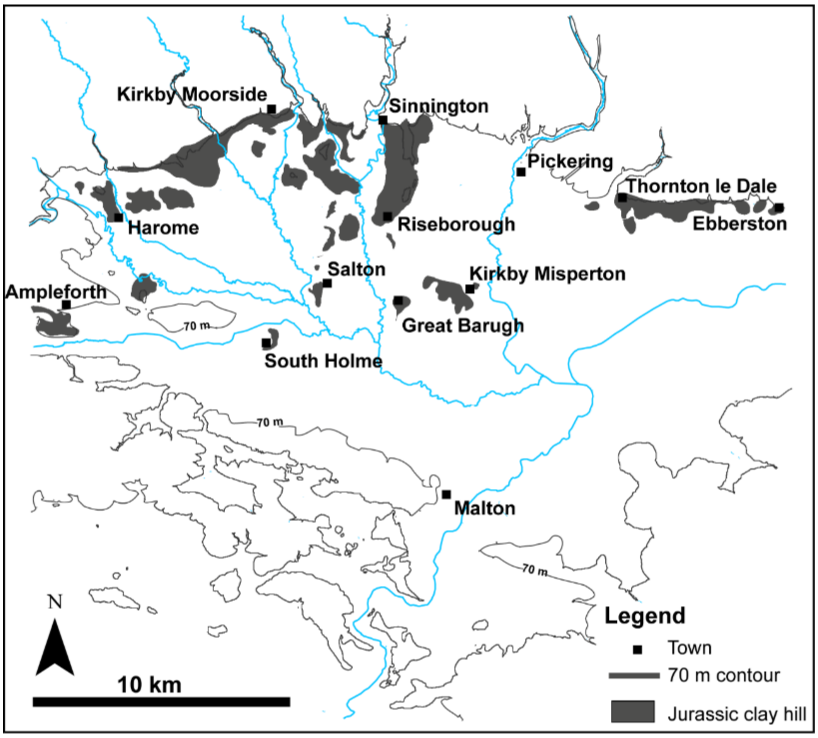


Figure 3.1.1: The Jurassic-aged clay hills in the north central and western Vale of Pickering. The hills are upthrown blocks of Kimmeridge, Ampthill or Oxford clay that break the surface of the thinner Quaternary-aged deposits.

They occasionally contain the remains of ammonites allowing for the finely laminated clay to be distinguished from the Quaternary-aged laminated lacustrine clay, however, this is not always the case and the clay can appear similar to the lacustrine deposits. These hills often show evidence of erosion by fluvial processes, like small channels and terraces. Some have a veneer of weathered, older diamicton recorded in BGS geological maps, historic boreholes, shallow auguring by Edwards (1978), and in cores analysed Powell *et al.* (2016) (Figure 3.1.5).

The distribution of Jurassic-aged clay deposits is confined to the western Vale of Pickering and in places where the clay outcrops against the edge of the Corallian dip slope of the North Yorkshire Moors (Figure 3.1.2). One section, running in a long bank between Thornton le Dale and Ebberston, is downfaulted against oolitic limestones and sandstones in a series of fluvially dissected spurs that run from 72 m O.D. to 40 m O.D at Harrow Hill (Figure 3.1.2). At the mouth of Thornton Dale, the bank of clay is dissected by the misfit channel of Thornton Beck. Extending to the southwest is the High Riggs gravel complex, a fluvial deposit emanating from the mouth of Thornton Dale (see alluvial fans). The outcrop of Kimmeridge clay is capped by a gelifluctate underlying a middle layer of silts and a topped by another gelifluctate (Franks, 1987).

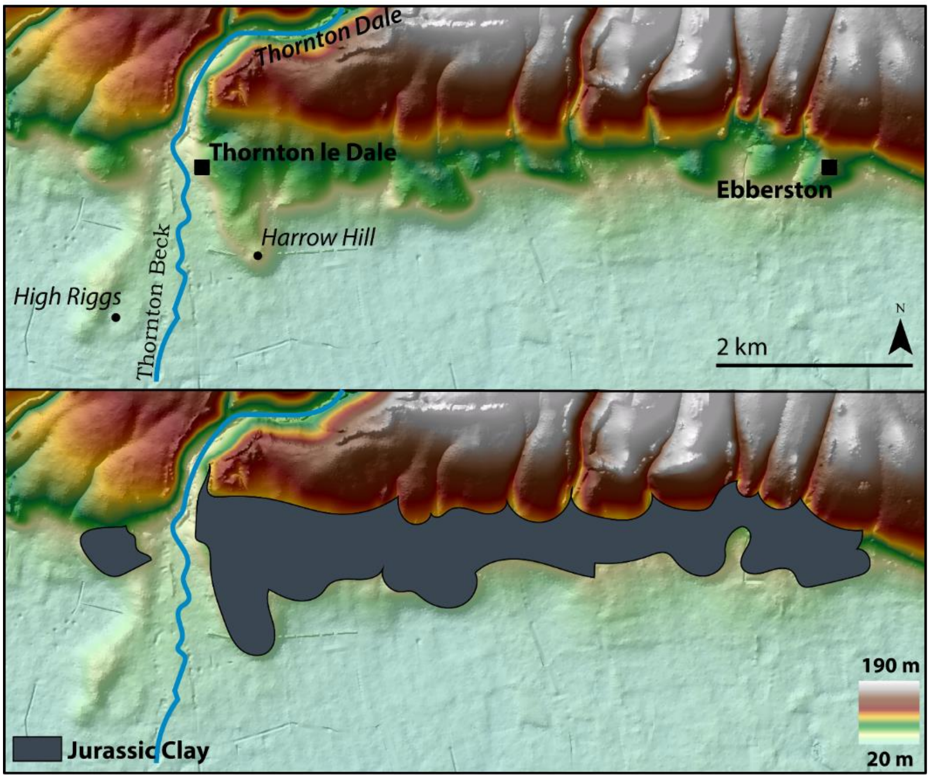


Figure 3.1.2: The Jurassic Clay spur that outcrops from under the Tabular Hills at Thornton le Dale east to Ebberston. Note where Thornton Beck incised the clay bank at Thornton le Dale. The High Riggs gravel complex is a n alluvial fan that extends south west of Thornton le Dale.

West of Pickering, the long north-south trending spur at Riseborough (Figure 3.1.3) splits the Vale between the low, flat, lacustrine landscape and one of more relief. The Quaternary deposits are shallowest here, and altogether absent or just a thin post-glacial alluvium covers the bedrock north of Harome. The extent of the Kimmeridge clay and the depth of Quaternary diamicton deposit is important as in some places older lake shorelines are preserved as planation surfaces eroded onto the surface of the clay bedrock (see shorelines; terraces).

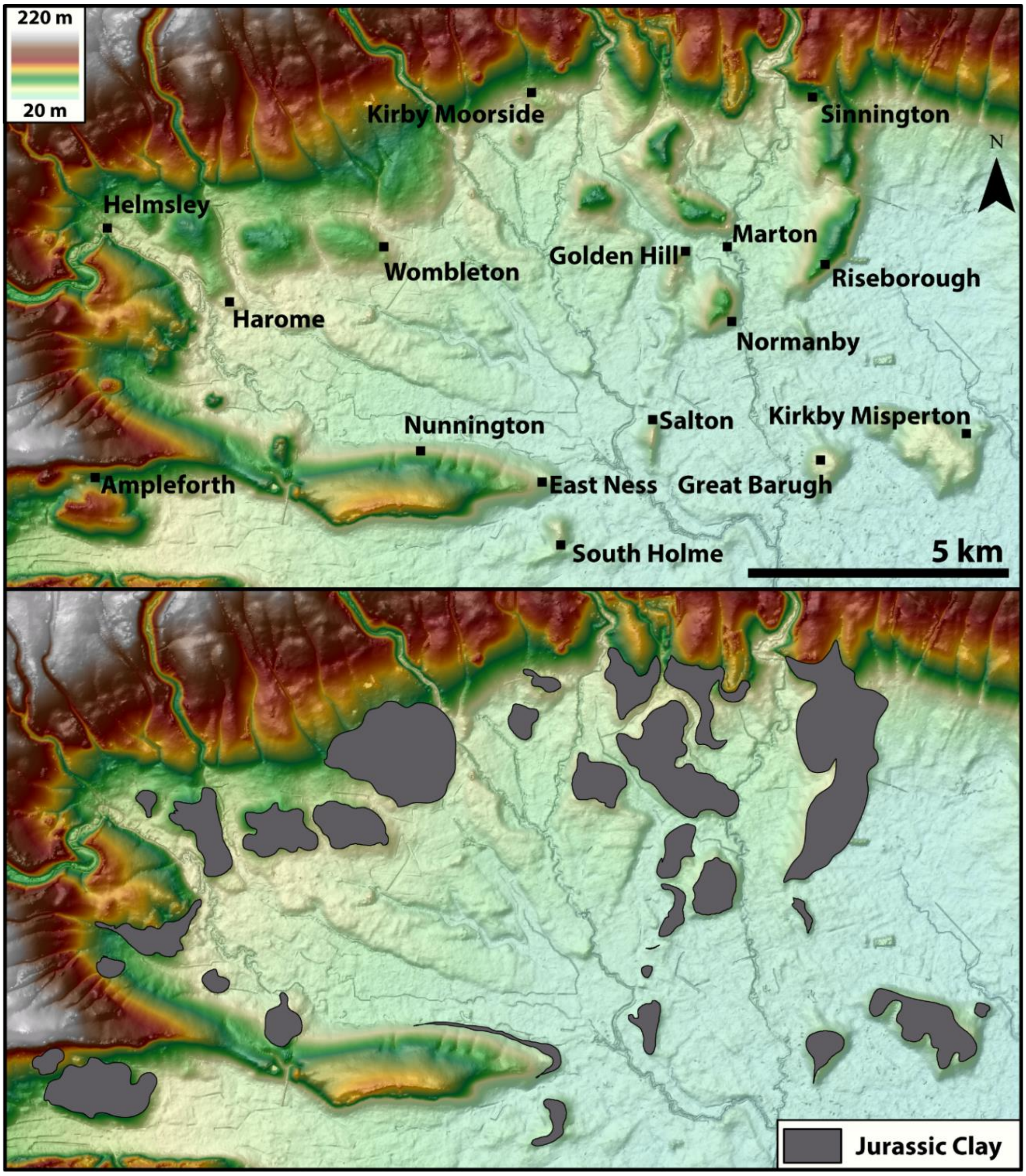


Figure 3.1.3: The western end of the Vale of Pickering showing the location of Jurassic clay hills. The Quaternary sediments are thinner here than in the central and eastern Vale of Pickering, which allow for upthrown blocks of clay to break the surface.

These hills are well rounded and smoothed. Occasionally, the hills have a thin veneer of weathered diamicton attributed to MIS 8 by Powell et al, 2016.

At Ampleforth (Figure 3.1.3), a large hill of Kimmeridge clay has been often cited as a terminal moraine from Vale of York ice intruding into the Coxwold-Gilling Gap (Fox-Strangways, 1881; Kendal, 1902). Although when investigated by Reed (1935) it was found to be a spur of Kimmeridge clay not quite dissected by the tectonic processes that formed Coxwold-Gilling Gap with no evidence of glacial diamicton.

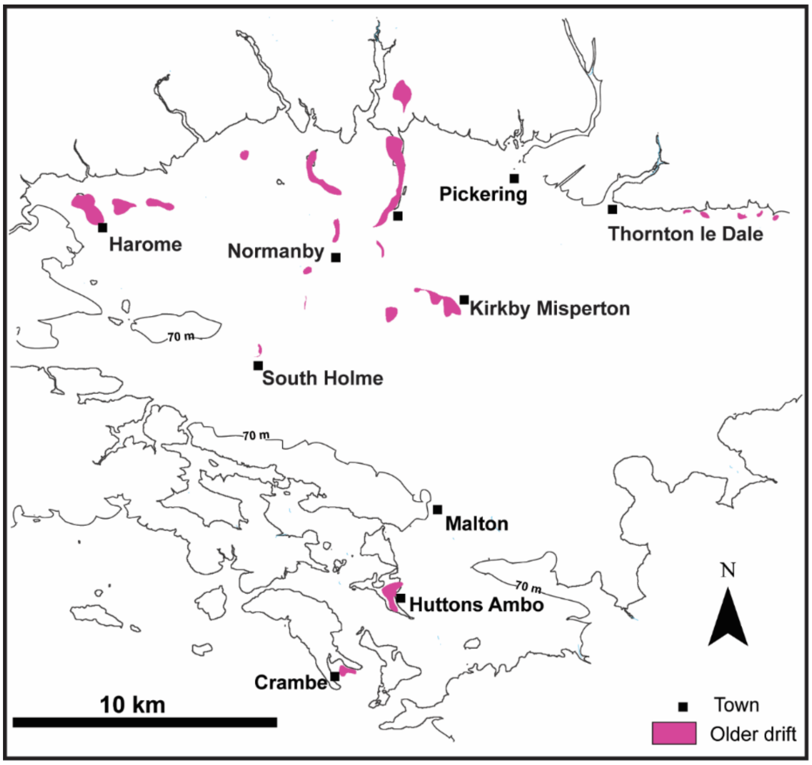


Figure 3.1.4: Older diamicton deposits of the Vale of Pickering. The adherent nature of the underlying Jurassic clay has helped preserve the diamicton on the crests of some of the hills. Deposits in the Howardian Hills (at Crambe and Huttons Ambo) are undated and considered by Powell et al (2016) as a separate mid Pleistocene deposit to the diamicton found on top of the hills within the Vale of Pickering basin.

Older diamicton deposits

The older diamicton deposits that are occasionally found on the top of the Jurassic hills is highly weathered, containing clasts dominated by local lithologies including sandstones, limestones (quite often the Malton oolite) and chalk (Figure 3.1.4). At Crambe and Huttons Ambo, the diamicton is attributed to a separate, undated, mid Pleistocene glaciation (Powell *et al*., 2016).

Interpretation

The Jurassic clay hills are important as they provide a long record of the geomorphic processes that have shaped the geomorphology of the Vale of Pickering. However, the clay is grey and laminated and has been reworked in places into Quaternary lake deposits so, occasionally, the two clay units can appear very similar. This has led to some misidentification in places. For example, Edwards (1978) interpreted an exposure of laminated clay near Thornton le Dale at 76 m O.D as evidence of a high-stage Lake Pickering (Figure 3.1.2), but after comparison with the BGS 625k geological map and independent verification by Franks (1987), it is apparent that Edwards encountered the Kimmeridge clay and not Quaternary clay.

Furthermore, it is likely that the tenacious nature of the Jurassic clay has allowed for the preservation of older diamicton deposits (Figure 3.1.5) and could reveal more about the glacial history of the Vale of Pickering. Likewise, the clay is somewhat easily eroded and has been reworked and incorporated into the lacustrine deposits in the Vale of Pickering basin. In places, like at East Ness and Kirby Misperton (Figure 3.1.2), the clay has been eroded into planation surfaces and small embayments. Fluvial channels also preserve a record of lake height (see channels) terminating at lake base levels.

Alluvial fans and deltas

There are several large alluvial fans and deltas in the Vale of Pickering and it may be that some deltas continued as alluvial fans as the lake drained (Ford *et al*., 2015). There are also some sites where smaller alluvial fans coalesce to become one unit, such as the Sherburn Sands (Foster, 1985; Evans *et al*., 2017) that lie at the foot of the Yorkshire Wolds scarp (Figure 3.1.6). Many of the fans and deltas are undated and stratigraphic relationships with other deposits in the Vale of Pickering have been used to estimate ages (e.g.: Edwards, 1978; Foster, 1985; Franks, 1987; Lincoln, 2017). However, it is sometimes difficult to establish whether a landform was deposited as an alluvial fan or a delta (Ford *et al*., 2015) as many deposits have been reworked or significantly eroded (e.g.: The High Riggs gravel complex south of Thornton le Dale) or are not covered by LiDAR imagery and the NextMap resolution is too coarse. For this reason, both were mapped together.

The largest delta is found at Pickering and merges with a series of smaller alluvial fans to the east that extend to Thornton le Dale (Figure 3.1.6). The largest alluvial fan in the Vale of Pickering forms part of the Hutton Buscel kame complex and merges with another large delta found at the mouth of the Mere Valley known as the Seamer Gravels (Figure 3.1.6).

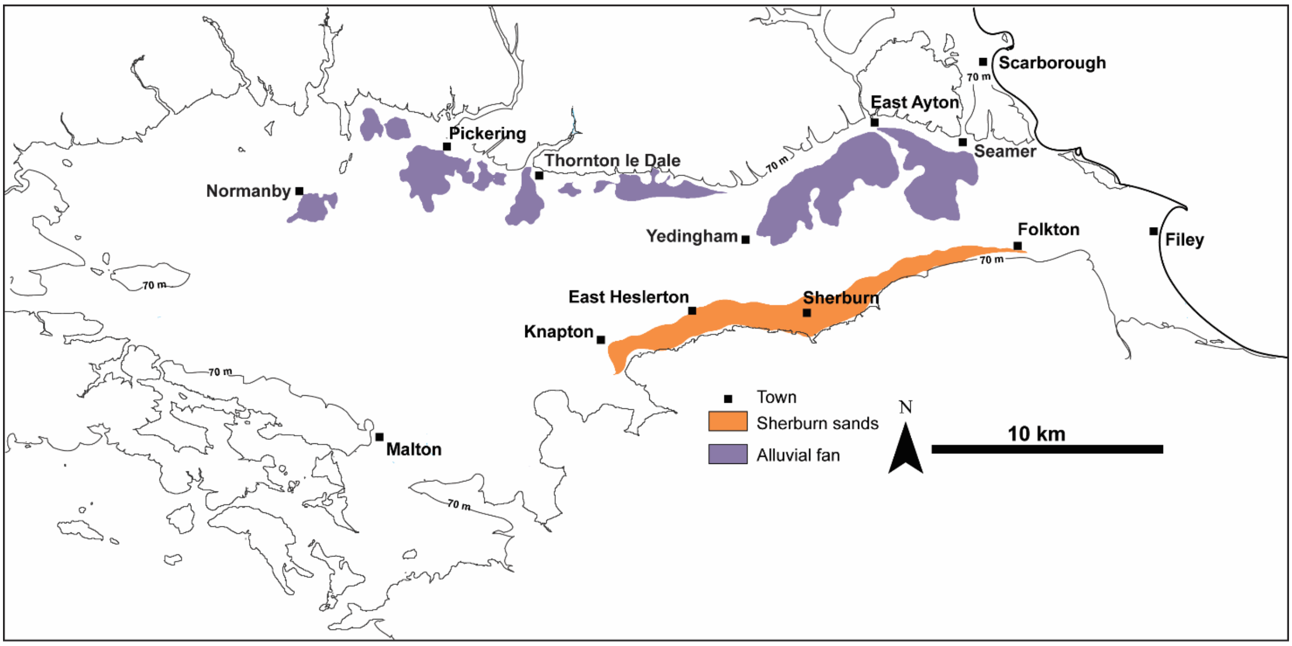


Figure 3.1.6: The alluvial fans and deltas in the Vale of Pickering shown in purple with the partially alluvial Sherburn sands shown in orange.

The Seamer Gravels extend southwards from the mouth of the Mere Valley for 3 km towards Staxton. The eastern edge of the deposit forms a crescent shaped moraine and the western rim of the Lake Flixton basin (Figure 3.1.7). The shows terracing from 33-35 m O.D and 28-32 m O.D (Lincoln, 2017) and merges with the Hutton Buscel sands and gravels on its western edge.

The Hutton Buscel Sands and Gravels (Franks, 1987) extend for 6.5 km in a south-westerly direction from the mouth of the Forge Valley at East Ayton towards Wykeham (Figure 3.1.8) starting at a height of 78 m O.D. with a falling surface gradient of 1:345 from east to west, reaching as far as Rye Topping 2 km southwest of the Wykeham moraine (Figure 3.1.8). The Hutton Buscel gravels are comprised of four distinct units described by Franks (1987) (Section 1.6). It is not possible to identify the individual units from LiDAR especially where the deposit merges with the Seamer gravels on the eastern side. West of Wykeham, a secondary delta emanating from north of Ruston is dissected by a number of channels draining from the North Yorkshire Moors and from the long channel that sweeps eastwards across the Corallian dip slope above the Hutton Buscel kame (Figure 3.1.8.). A small rounded hill known as Rye Topping is situated at the west-central point of this deposit and may be a kame or portion of the Wykeham moraine (see Morianes).

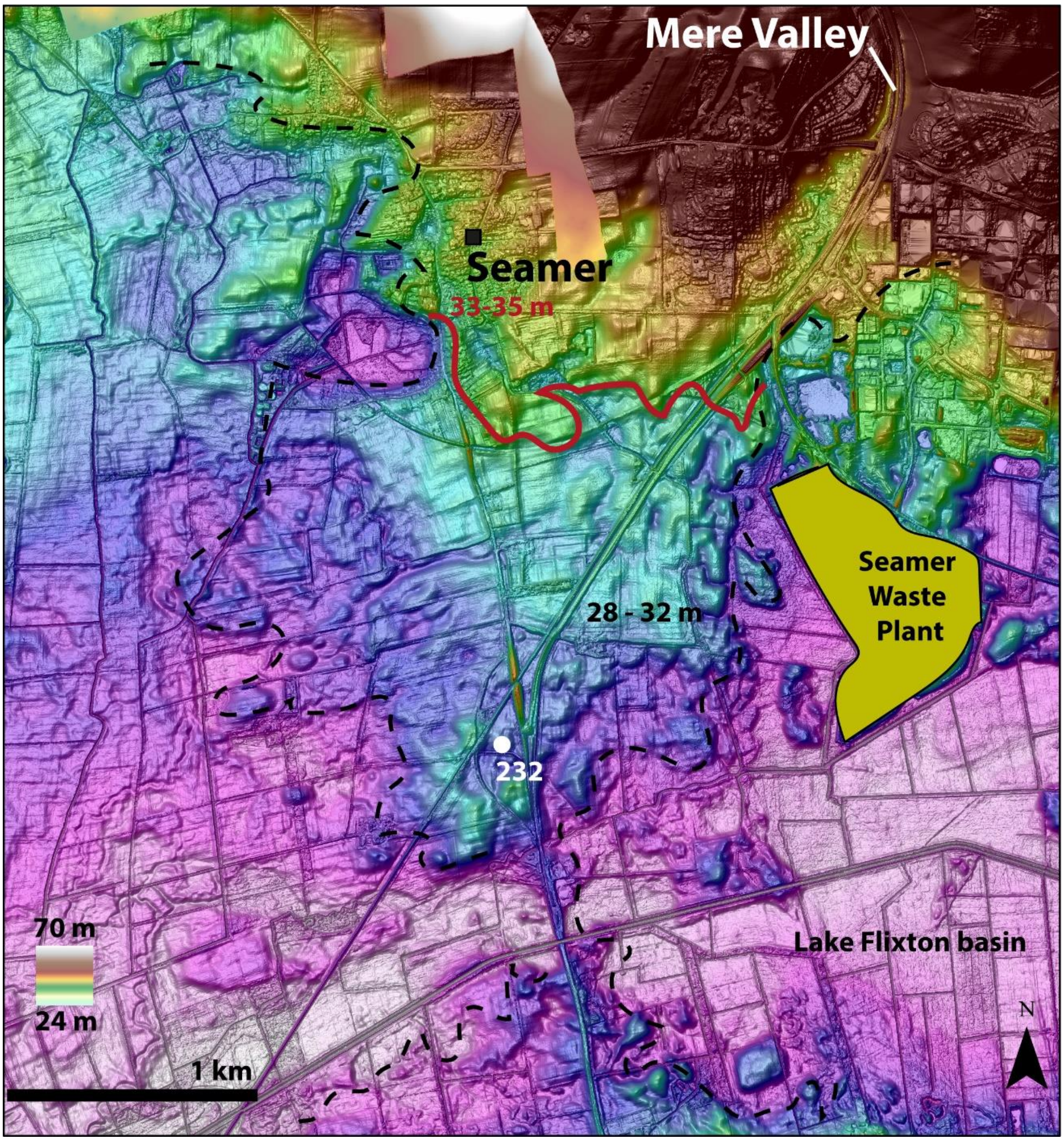


Figure 3.1.7: The Seamer gravels extending from the mouth of the Mere Valley. Note the flat area of the Lake Flixton Basin. The Seamer gravels likely delineate a former ice margin.

West of the Hutton Buscel Sands and Gravels there are a series of shallow alluvial fans emanating from the mouths of a line of dry valleys and coalesce at the base of the Jurassic clay outcrop. LiDAR imagery is absent in this area and the NextMap 5 m resolution is too coarse for a detailed interpretation. At Thornton le Dale, the remains of what is thought to be a large delta extend for 2 km southwards into the Vale of Pickering basin from the mouth of Thornton Dale. The delta is known as the High Riggs Gravel Complex (Edwards, 1978; Franks, 1987) and has been incised by the once larger Thornton Beck that occupies Thornton Dale.

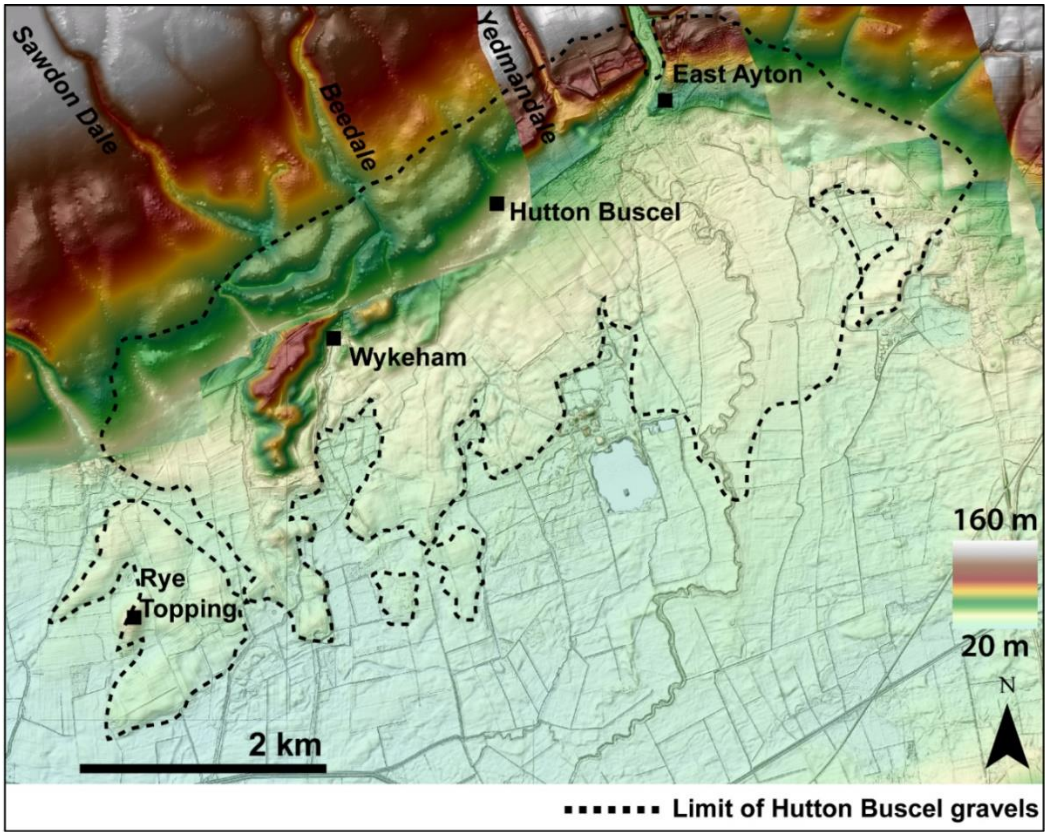


Figure 3.1.8: The dashed line delineates the extent of the Hutton Buscel sand and gravel complex. The raised ground at Wykeham is the Wykeham moraine. The Rushton delta lies east of the Wykeham moraine and includes Rye Topping, a small conical hill that is potentially a small kame or eroded extension of the Wykeham limit.

A few miles west, the extensive Pickering delta that emanates from the mouth of Newtondale meltwater channel (Kendall, 1902; Gregory, 1965). The town of Pickering is situated directly on the delta deposits and changes from human settlement has reorganised some of the deposits with some of the terraces now landscaped. The delta extends south for 3.5 km into Pickering Low Carr, an area of marshy land and is 3 km at its widest point. The delta starts at 35 m O.D falling to 23 m O.D. in a series of terraces. West of the Pickering delta in Ailsby Carr are two previously identified uncategorised hummocks both approximately 500 meters in length and 30 m high. A historic borehole (BGS ID: 126135) records “boulder clay” overlying Kimmeridge clay. Field investigation of these hummocks prove them to be a stoney, structureless, mottle grey and orange clay. Further west, at Normanby, a low angle fan extends 2 km southeast from between two outcrops of Kimmeridge clay into the Vale of Pickering.

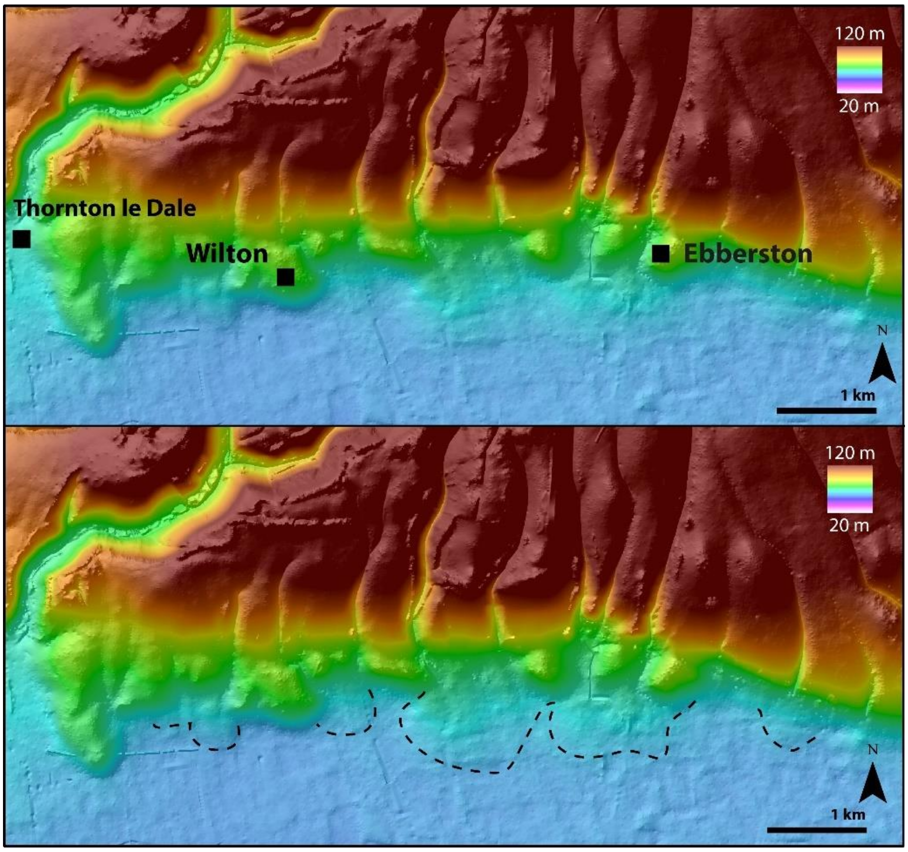


Figure 3.1.9: Alluvial fans from the dry valleys between Thornton le Dale and Ebberston. The lack of fine resolution DEMs makes interpretation difficult.

In the south, along the chalk scarp, a line of coalescing alluvial fans run 18 km from Knapton to Folkton. These fans known as the Sherburn Sands (Foster, 1985) and extend from a starting elevation at the base of the chalk scarp of approximately 50 m OD into the Vale of Pickering for up to a kilometre. King (1965) noted that the width of the fans correlates with the steepness of slope and that they extend in to the dry valley near Sherburn as an aeolian deposit. Shorelines are recorded at the base of the fans by the reworking of alluvial sediment (See shorelines). These sands have been reworked by wind and as a result have an aeolian cap of between 10 cm and 50 cm thick

Interpretation

The Seamer Gravels have been interpreted as an ice marginal fan delta (Franks, 1987; Evans *et al.*, 2017; Lincoln, 2017) deposited into a 33-35 m O.D lake with a lower terrace between

32 -28 m O.D. that was worked as the lake level dropped. Franks (1987) interprets ice-contact features within the Seamer gravels as indicative of concurrent deposition with a stagnant ice front sitting in the Lake Flixton basin. Historic boreholes drilled by the North Yorkshire County Council (N.Y.C.C.) record “boulder clay” in borehole 232 south of Seamer station that Franks (1987) matches to the Lower Till Series (Skipsea Till) (Table 1.3.1) analysed by Edwards (1978). The Lower till is noted by Franks to extend along the north eastern edge of the Seamer gravels. If this comparison is correct, the Lower Till Series can be correlated with the extension of ice into the Vale of Pickering at least as far as Seamer. However, a core north of Seamer to limestone bedrock at 27 m depth suggests that ice may have advanced as far as Seamer more than once. This borehole stratigraphy shows Seamer gravels capped by a “boulder-type clay”. Edwards (1978) considered this deposit to be the western most limit of the Upper Till Series (see Moraines).

The Hutton Buscel complex in more difficult to interpret. The Hutton Buscel sands and gravels have been interpreted as a kame terrace deposited as a terminal moraine as ice reached Wykeham (see Moraines). Its most famous interpretation is from Kendall (1902) as evidence of a large proglacial lake of 70 m O.D. Franks (1987) argues that the terraces falling from 68 m to 60 m O.D. is the result of localised ponding between the Corallian dip slope and the ice front. The Hutton Buscel gravels also predate the incision of the Forge Valley (Franks, 1987; Lincoln, 2017).

The Hutton Buscel gravels are comprised of four units (Franks, 1987). A basal unit that sits unconformably on the Jurassic bedrock is known as the Cooks Quarry Gravels (although to avoid confusion with Cooks Quarry at East Heslerton, the unit is now the Ayton Quarry Gravels). They are not as laterally consistent as the overlying units and are comprised of horizontally stratified gravels that thicken in the south portion of the Hutton Buscel unit. They include a high concentration of exotic lithologies (jasper and agates) not seen elsewhere in the Vale of Pickering and attributed by Franks (1987) to an earlier pre-Devensian ice advance. Above this unit is the Toft Hill silts. These are described as an aeolian deposit similar to the loess on the Wolds (Catt *et al*., 1973; Franks, 1987) and are lithological similar to the Harrow Hill silts (see below). The third unit is the Canborough Gravels, which are comprised of angular local limestones with rip up erratics of the underlying Toft Hill silts suggesting reworking by ice (Franks, 1987). Sporadic instances of delta forests are recorded throughout the unit but lack consistency throughout, which Franks (1987) attributes to ephemeral lakes forming between the ice margin and the Corallian dip slope. These lakes may have drained frequently along the long east-west trending channel resulting in the fan near Ruston.

The delta at Ruston is fluvially eroded and likely represents the lowering of the lake from ~35 m O.D to 23 m O.D. The mouth of the east-west channel enters the Vale of Pickering at approximately 34 m O.D. just above the 33 m contour A series of shallow terraces cut between 30 m and 33 m suggest lowering of lake levels while the delta is fluvially incised down to 23 m O.D. A long channel running southeast between Brompton by Sawdon and the southern point of the Wykeham moraine suggest the continuation of the Hutton Buscel channel once the lake lowered from 27 m to 23 m. Smaller fluvial channels that show abandonment between 30 m and 25 m O.D suggest that the form of the Hutton Buscel sands and gravels and the Wykeham moraine may have been reworked by lowering lake waters.

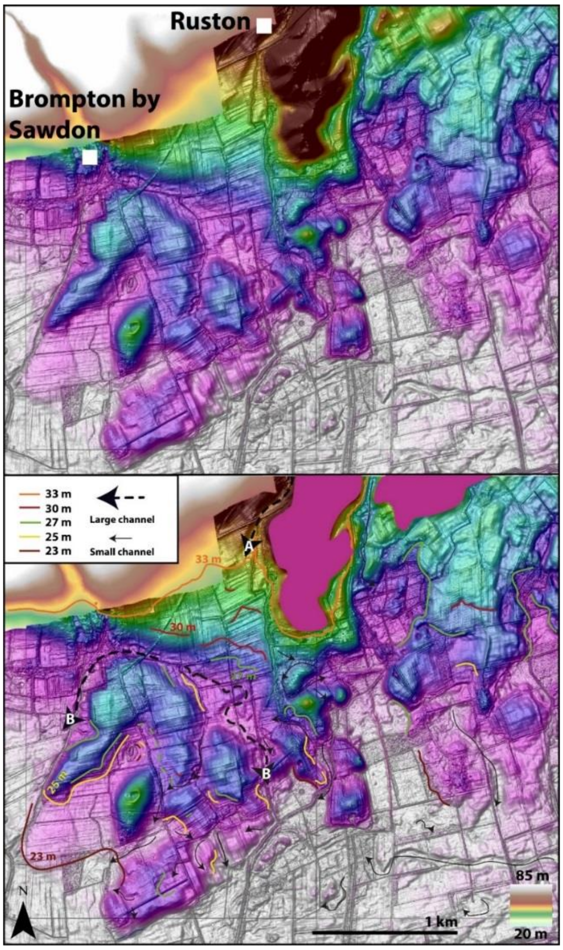


Figure 3.1.10: The dissected delta at Ruston. Meltwater circumventing the Hutton Buscel Kame Terrcae entered the Vale of Pickering near Ruston. Lowering lake levels and increased meltwater during deglaciation eroded the delta into its current configuration.

Figure 3.1.11: The delta at Pickering is divided into six terraces between 40 m O.D. at the mouth of Newton Dale to 23 m O.D. at the basin floor in Pickering Low Carr

The deposits west of Ruston to the High Riggs Gravel complex at Thornton le Dale are low angle alluvial fans that were likely formed as nival meltwater entered the Vale of Pickering via the numerous north-south trending dales (see Channels) of the Corallian dipslope. Lack of high resolution DEM makes further interpretation difficult.

The delta at Pickering has not been extensively studied aside from a few notes made by Sewell (1905) on what he considered to be glacial deposits. He noted through a number of well borings and gas main excavation pits that the delta is comprised of a series of waterworn gravel and boulders sourced from the Jurassic rocks of the North Yorkshire Moors. Like the delta at Ruston, the Pickering fan has been fluvially eroded and reworked by lowering lake waters and a diminishing source of meltwater through the Newton Dale channel that is now occupied by the Pickering Beck. A historic borehole from next to the A619 reveals clay overlying bedrock and underlying delta gravels. The delta deposits are capped with a sandy yellow clay suggesting that the lake existed both prior to and after the deposition of the Pickering delta. Whether the two lakes were part of the same lacustrine phase is unknown.

Finally, the shallow but extensive fan at Normanby (Figure 3.1.12) is a result of a reduction in fluvial energy as water from the North Yorkshire Moors enters the wider part of the basin. It is likely related to post-glacial fluvial activity revealed by the dates and deposits analysed at Wellfield Farm, Salton (Section 3.2) But does suggest that the Kimmeridge Clay hills were significant in maintaining the 33 m O.D lake level.

In the south, the source of the Sherburn Sands is unknown, but it is likely that they are derived from the Hutton Buscel and Seamer complexes as the Forge Valley and Mere Valley opened and the River Derwent reversed its course to the south (Lincoln, 2017). These sands have been reworked in the LGIT, with work by Evans *et al*. (2017) interpreting these deposits as late glacial (18 -15 ka) coalescing alluvial fans with an aeolian cap. Shorelines have been terraced into the sands as the lake reduced from 30 m to 24 m O.D (See shorelines).

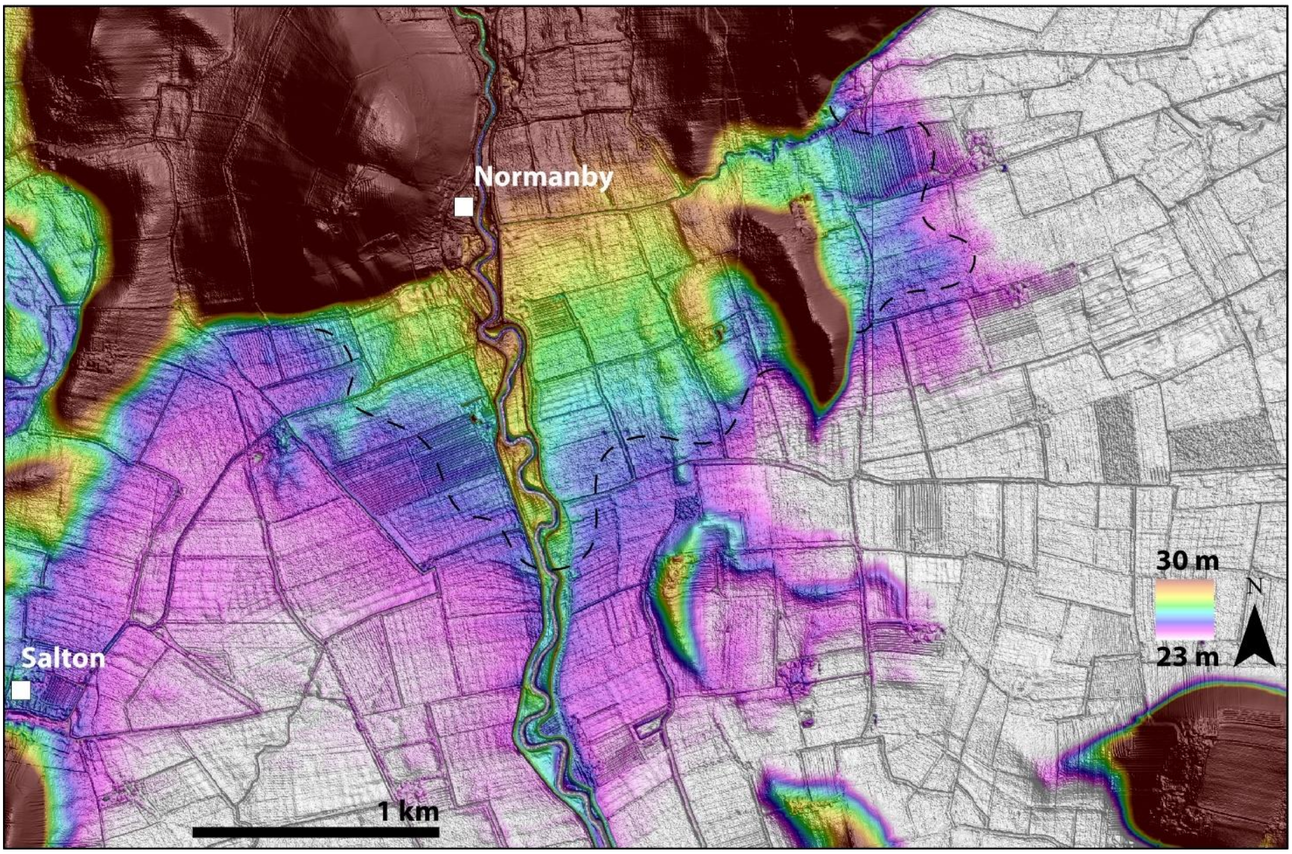


Figure 3.1.12: A low angle alluvial fan enters the basin near Normanby from between two of the Jurassic Hills. The Jurassic hills likely helped for a barrier for lake extension past 33 m O.D.

Terraces and benches

Terraces are present in the western portion of the Vale of Pickering, at the Malton Embayment and in the Kirkham Gorge (Figure). In the western Vale of Pickering, the dissected remains of the former Vale floor are divided into five long terraces with two terrace stubs. The terraces all start either in the north or northwest at 45 - 48 m O.D and grade downwards to 27 m O.D at the confluence of the Rivers Riccal, Rye and Dove where they merge into the current valley floor. The longest of these terraces runs approximately 6.5 km from Harome to Muscoates. The widest terrace sits below Great Edstone and is 1.8 km wide.

Planation surfaces are evident on the limestone bank known as Caulkley's Bank that runs east west from Oswaldkirk to East Ness (Figure X). Terraces west of Nunnington are subtle and originate around 45 m O.D. The road that runs east-west across the bank uses this terrace and so it is almost completely obscured. To the north of the road, a manmade bank runs west of Nunnington 200 m for 1.2 km, but due to the unusual shape, it unclear if this bank is a reworked natural feature. East of Nunnington at West Ness, the terrace becomes more distinguishable at 40 m O.D (Figure X). At the very tip of Caulkley's Bank, at East Ness, the planation surface becomes notably longer and flatter (Figure X) with a terrace dipping from 39 m O.D. to 37 m O.D.

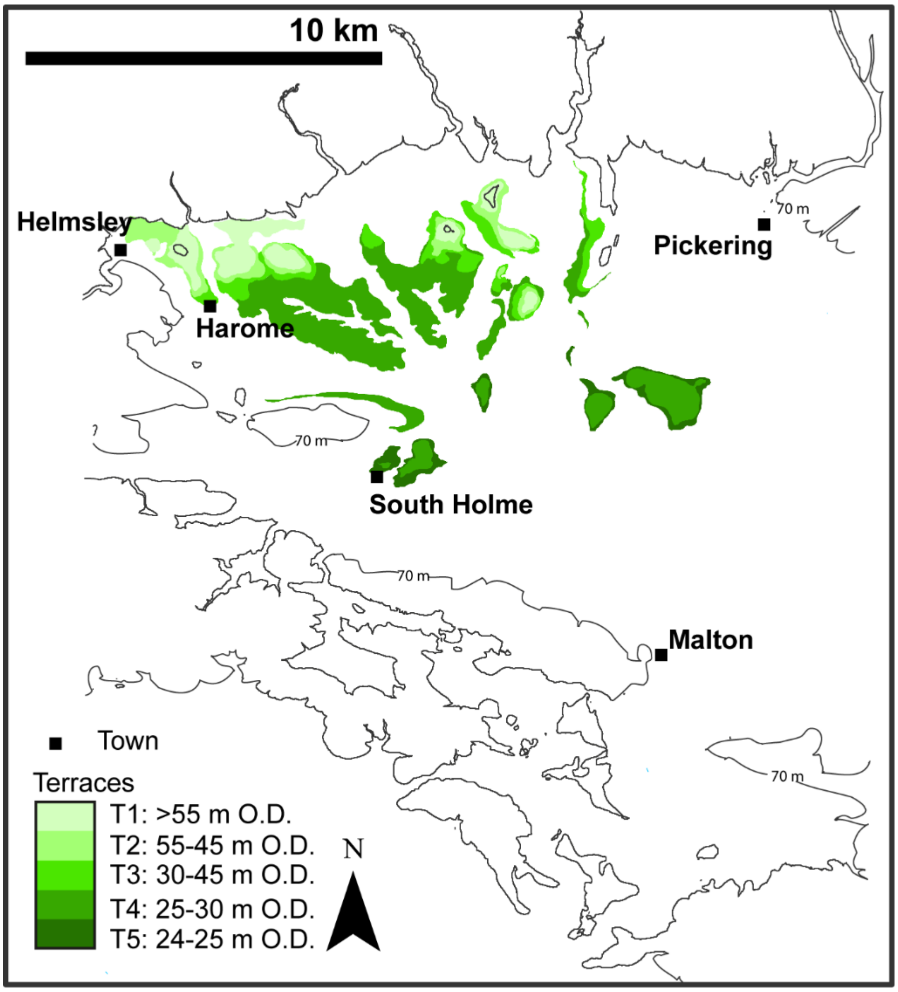


Figure 3.1.13: The terraces of the Vale of Pickering are primarily comprised of the former lake bed and planation and erosional surfaces thought to have been carved by lake waters.

The surrounding Kimmeridge Clay hills also have some evidence of planation potentially by lake waters. At Great Barugh and Kirkby Misperton planation surfaces suggest on lapping by lake waters at heights from 33 m O.D to 25 m O.D. with small channels terminating at 30 m (Figure 3.1.15).

Around South Holme, there are a series of north-south trending hummocks with an overlapping east-west trending feature. The largest of the hummocks is 1.2 km long and reaches 40 m O.D at its highest point. The east-west trending feature is 2 km long rising from 28 m O.D at the ends to 31 m in the centre. A historic borehole record (BGS ID: 612932) at South Holme Farm records 1.1 m of boulder clay overlying grey shale and clay.

Interpretation

The terraces in the western vale of Pickering are likely to be older surfaces from a pre-LGM 35 m lake, although this lake is undated. The flat surface that stretches from Nunnington to East Ness is likely the result of lake waters reworking a section where the Kimmeridge Clay outcrops from under the overlying limestone creating a planation surface that drops from 45 m O.D. at Nunnington to 40 -38 m O.D. at East Ness (Figure 3.1.14). Historic boreholes at East Ness sewage processing plant record laminated clay present in 6 boreholes up to a depth of 10 m, although it is uncertain whether this is the Kimmeridge clay or the Quaternary lacustrine clay.

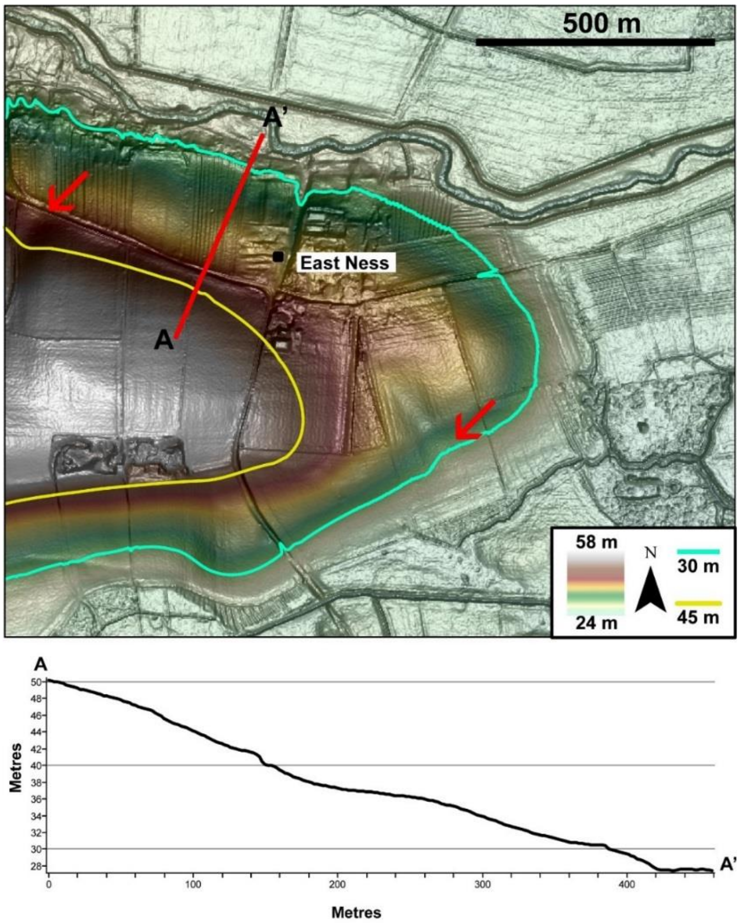


Figure 3.1.14: Benches and possible planation surface at East Ness. The shoreline drops from 45 m near Nunnington to below 40 m at East Ness. The red arrow marks a short channel that terminates at 33 m O.D.

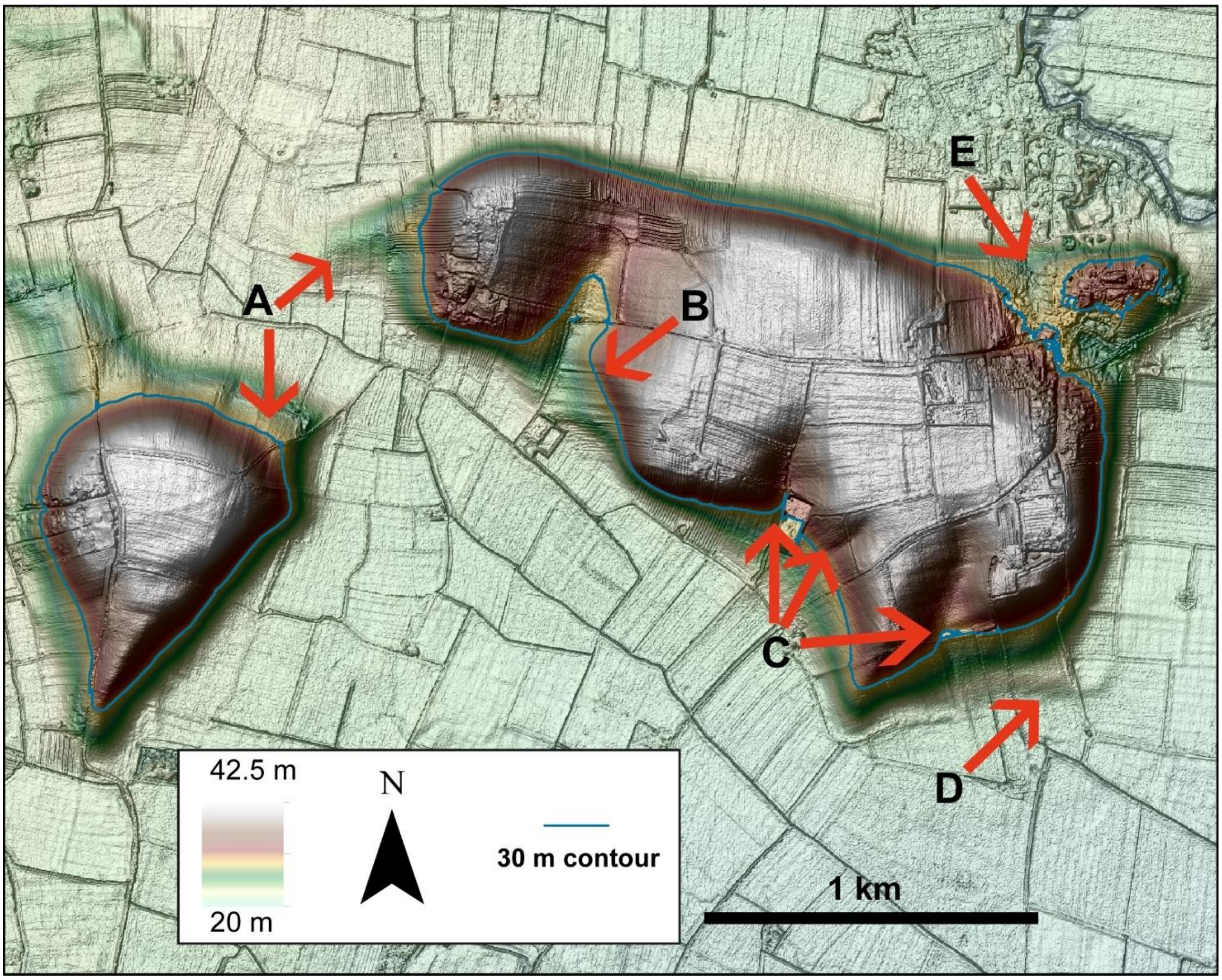


Figure 3.1.15: Planation surfaces, inlets and abandoned channels eroded into the Jurassic clay hills of Great Barugh (Left) and Kirkby Misperton (right). Powell et al., 2016 prove weather MIS 6 till on the tops of the Jurassic hills.

Terraces in the Kirkham Gorge (see Kirkham Gorge below) and along the edge of the Pickering delta support the existence of a 33 m O.D. lake that receded in several stages to a height of 25 m O.D. A small terrace between 40 and 45 m O.D may be related to lake height, but due to the nature of the bedrock, the terrace could equally be related to the complex bedrock faulting of the Howardian Hills.

Hummocks

The majority of uncategorised relief are simply interpreted as hummocks. Along the northern bank of the River Rye from southeast of Salton to Ryeton (approximately 5 km) there is a long embankment of sand, roughly 50 cm high and 600 m at the widest point, which continues along the southern bank of the river. Similar, but less extensive, deposits are located along the River Derwent and at Yedingham. North of Great Barugh towards Normanby

Around South Holme, there are a series of north-south trending hummocks with an overlapping east-west trending feature. The largest of the hummocks is 1.2 km long and reaches 40 m O.D at its highest point. The east-west trending feature is 2 km long rising from 28 m O.D at the ends to 31 m in the centre. A historic borehole record (BGS ID: 612932) at South Holme Farm records 1.1 m of boulder clay overlying grey shale and clay.

Interpretation

The hummocks mapped are generally fluvial sand deposits that are a result of post-glacial reworking. The hummocks along the River Rye and Derwent are likely overbank flood deposits or small natural levees.

Moraines

The eastern end of the Vale of Pickering is dominated by glacial topography and a series of moraines can be identified (Figure 3.1.16). The most westerly extent of the morainic deposits is 18 km inland, near Wykeham, which is regarded as the terminal moraine. A series of recessional moraines run predominately north-south from Wykeham to Cayton Bay. Between Cayton Bay and Filey, the moraines predominant direction changes to overlapping ridges running both NE-SW and NW-SE. Previous workers (Edwards, 1978; Foster, 1985) have suggested ice entered the Vale of Pickering as far as Thornton le Dale (Edwards, 1978) and Marton (Foster, 1985). Currently, no conclusive evidence has been provided to show that ice from the east passed the limit at Wykeham. By revisiting and revaluating Edwards' (1978) sampling sites, Franks (1987) concluded that the landform was incorrectly identified by Edwards and that the till is a misinterpretation of a gelifluctate deposit. Foster's (1985) suggestion of a floating ice shelf covering the central portion of the Vale of Pickering (section 1.6) is unproven.

The Wykeham moraine is a long ridge approximately 5 km long trending south-eastwards from East Ayton towards the village of Wykeham following the shape of the Corallian scarp (Figure 3.1.17). Franks (1987) suggests that the western portion of the moraine is a continuation of the

Corallian bedrock where ice was restricted from entering the Vale any further to the west. He suggests that the high point of the moraine at Wykeham is a comprised of a south trending spur of the Coral Rag that has been covered with a veneer of Devensian till. No boreholes exist from the Wykeham moraine, so it is unconfirmed, but given the trend of the underlying bedrock, it is likely to be the case. The deposit generally lies between 65 m and 55 m O.D with its highest point reaching 70 m at Wykeham, where it turns directly southwards for 2 km into the Vale of Pickering, terminating at Ruston Carr. As it turns southwards, the deposit lowers to 45 m O.D at Wykeham Abbey before becoming noticeably fragmented (Figure 3.1.17) and reaching only 26 m O.D. at Ruston Carr.

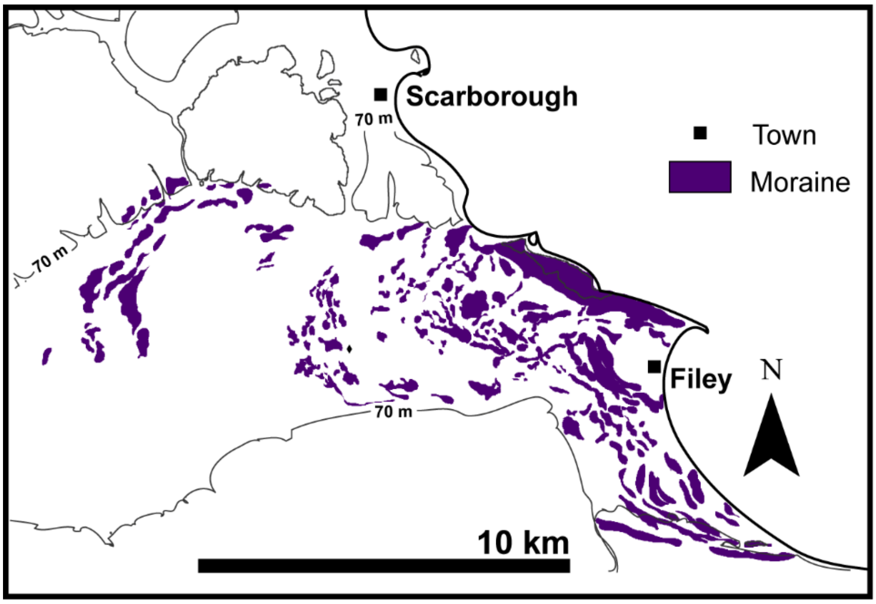


Figure 3.1.16: Moraines of the Eastern end of the Vale of Pickering (purple). A waste treatment plant near Seamer creates a topographic high but is unrelated to the glacial topography. The moraine at Wykeham represents the most westerly extent while the recessional moraines between Seamer and Staxton divide the Vale of Pickering from the Post Glacial Lake Flixton basin.

West of the Wykeham limit is a small hill known as Rye Topping (Figure 3.1.10; Figure 3.1.17). It is potentially the remains of the terminal moraine as it follows the line and shape of the Wykeham deposit. Its crest reaches 31 m OD the same as a portion of the Wykeham moraine 1.4 km to the northeast.

East of the Wykeham moraine, the glacial deposits are dissected by two meltwater outlets: The Forge Valley near East Ayton and the Mere near Seamer. Both have extensive alluvial fans that cover the area between Wykeham and Seamer (see Alluvial Fans). At the mouth of the Mere Valley, the Seamer gravels spill in a southwestwards direction, although the eastern and southeastern side of the fan are reworked into part of a series of recessional moraines (Figure 3.1.18). East of Seamer, the topography is dominated by recessional moraines and the basin of Post-Glacial Lake Flixton. The Flixton basin ends at Filey where a series of moraines overtop one another to create the highest feature in the area at 58 m O.D (Figure 3.1.18; Figure 3.1.19)

At the western end of the Lake Flixton basin are two recessional moraines that show an interesting relationship with the paleochannel development. There appear to be breaks in the moraine with an overdeepened channel running west-east, along the course of the (now canalised) River Hereford. The height of either side of the dissected moraine is the same at 27 m O.D.

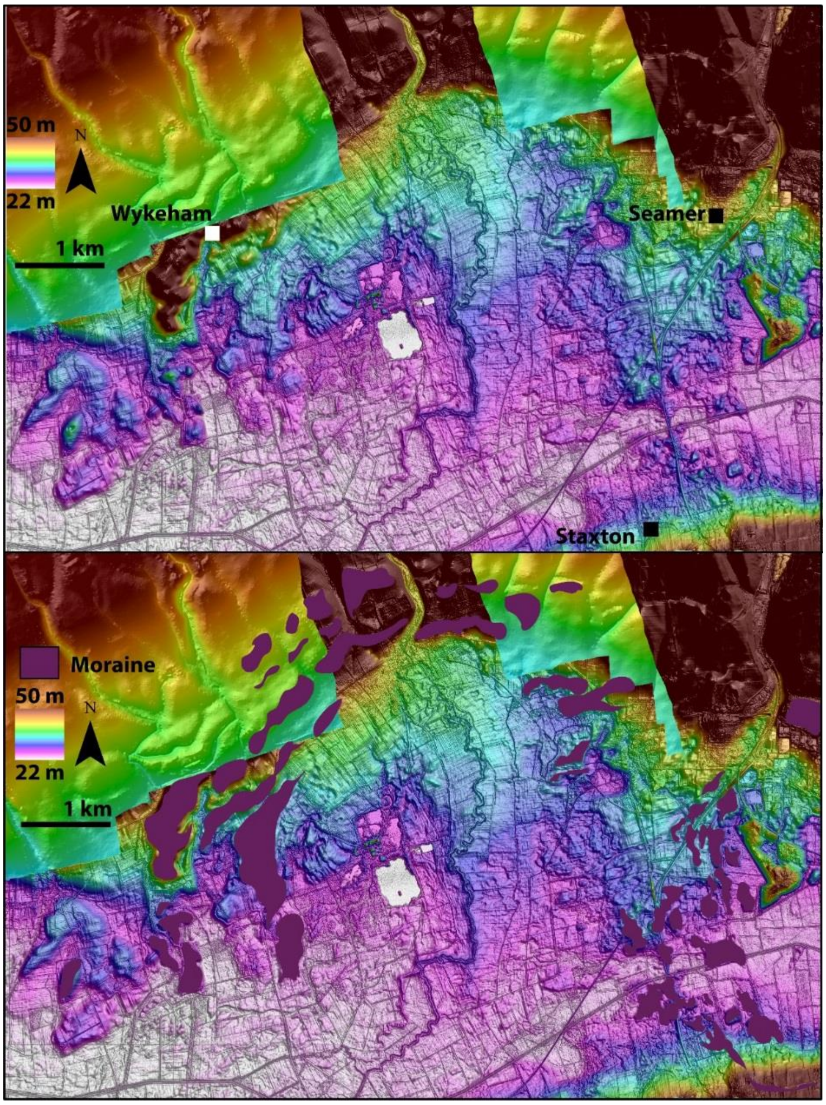


Figure 3.1.17: The Wykeham moraine is thought to be the most western limit of ice during the Devensian glaciation of the eastern end of the Vale of Pickering. It is suggested that that limit should be extended to incorporate the deposit at Rye Topping.

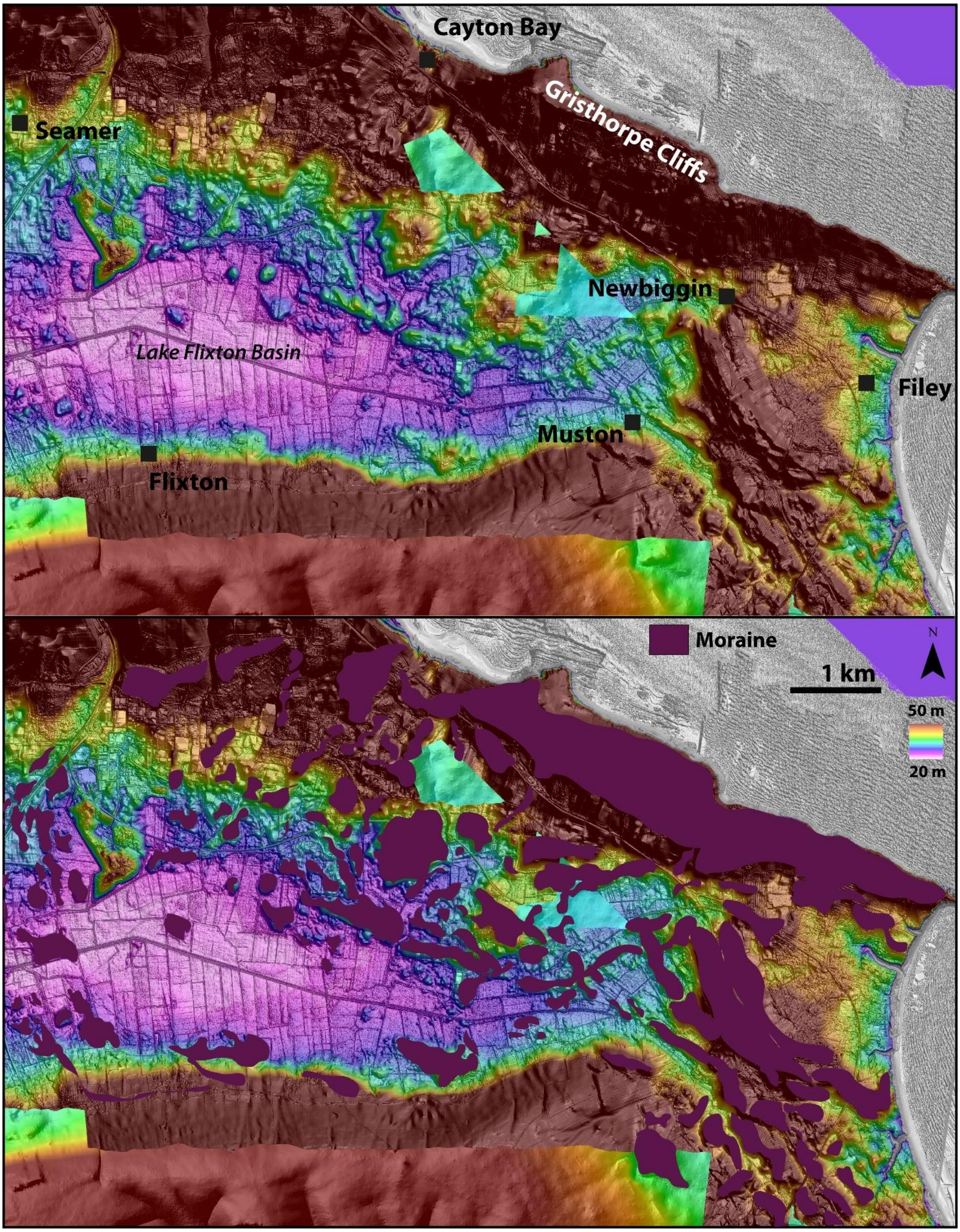


Figure 3.1.18 Moraines from Seamer to Filey, some of which are covered by the post-glacial peat deposits of Lake Flixton

The Lake Flixton basin is comprised of filled-in hollows (Palmer *et al*., 2015) that hide a series of rescessional moraines beneath a post-glacial infill of peat. Evidence for the ice recession across the Lake Flixton basin is found, however, at the margins; in the south, moraine ridges line the chalk scarp of the North Yorkshire Wolds. To the northeast, along the coast between Cayton Bay and Filey, a limestone ridge called Gristhorpe Cliffs is covered in till in places reaching over 20 m thick. At Cayton Bay there is a topographic low of 38 m OD between Osgodby and Gristhorpe Cliffs that is overlain by a moranic ridge. This ridge continues along the line of the Gristhorpe Cliffs.

The largest moraine lies in the east end of the Vale of Pickering near west Filey and is known as the Filey Moraine. The Filey moraine is comprised of a series of stacked north-south trending ridges that run between Newbiggin to 2 km northeast of Hunmanby. With an average height of 40 m O.D, it is the tallest feature separating the Lake Flixton Basin from the coast with the tallest point reaching 58 m O.D at Beacon Hill. The south end of the moraine curves eastwards towards the sea. South of the Filey moraine, the topography appears more fluvially dissected.

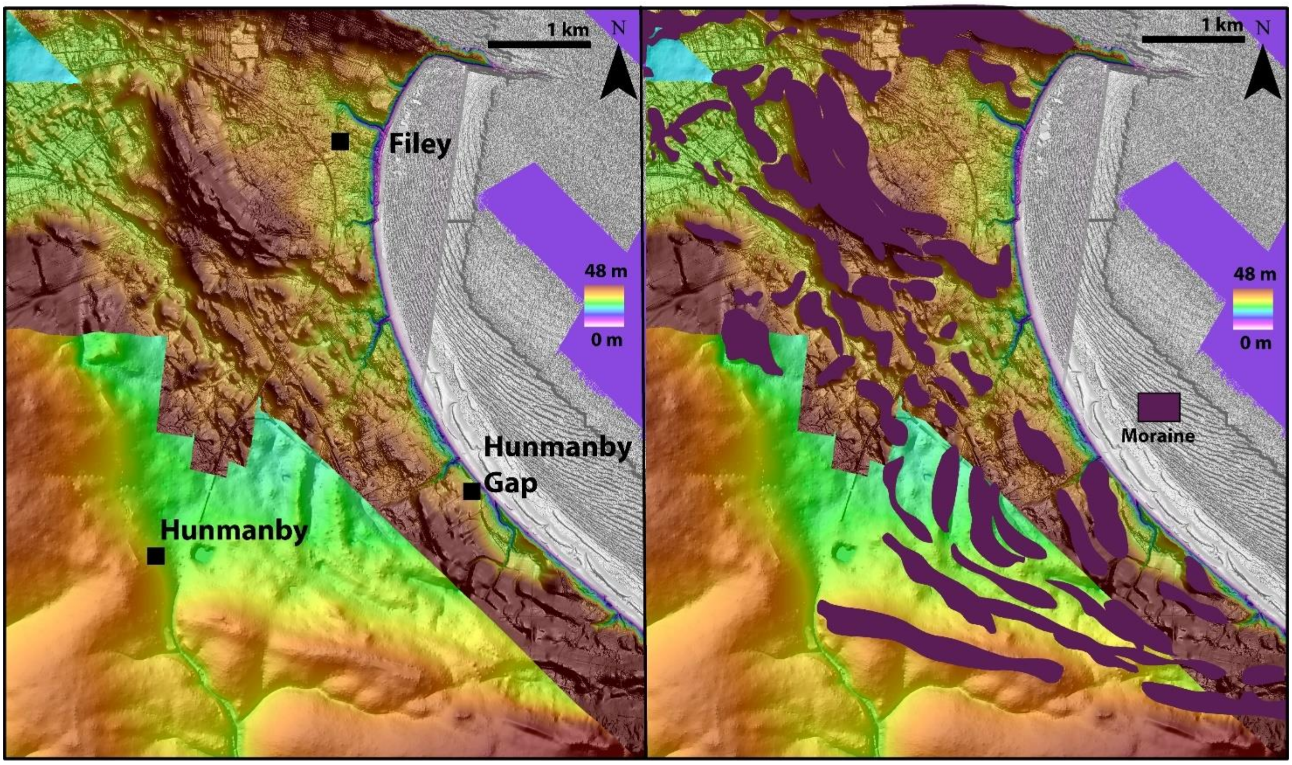


Figure 3.1.19: The Filey moraine is comprised of a series of stacked moraines due to NE ice overlapping moraines from a recessional SE ice. Note the relational position of the moraine to the sharp corner of the chalk bedrock due to the Hunmanby Fault at Muston. The highest point is at 58 m OD at Beacon Hill.

South of Filey, the morainic ridges predominantly run north south until they meet the scarp of the Yorkshire Wolds at Reighton. The moraine is wavy suggesting some compression due to the shape of the bay and the sharp right angle of the Hunmanby fault. There are also two east-west trending ridge that appear to lie beneath the north-south moraine.

Interpretation

If Ice did terminate at Wykeham due to a bedrock barrier, it explains why there is no southern arm to the Wykeham moraine. Instead, the ice front south of the moraine would terminate directly into the lake. Any evidence of ice contact between the floor of the Vale of Pickering and the ice margin south of Wykeham has been buried or eroded by post-glacial alluvium and reworking, so the limit of this southern portion is limited to conjecture. There is little evidence along the scarp of the Wolds for ice along the southern margin as much of the area is now covered by the Sherburn Sands. There are two small sandy hummocks at Ganton and at Sherburn, which are possibly an alluvial fan or delta that further investigation may reveal any evidence of ice contact, but the southern limits of the Wykeham moraine are still unknown.

At Staxton, the moraines delineated by the edge of the Seamer Gravels show evidence of possible moraine damming in the Flixton Basin. A wide paleochannel connects the Lake Flixton basin with the main basin in the Vale of Pickering. If such an event occurred where the moraine dam burst, it could account for some of the source of the Sherburn sands as well as the reorganisation of the River Derwent.

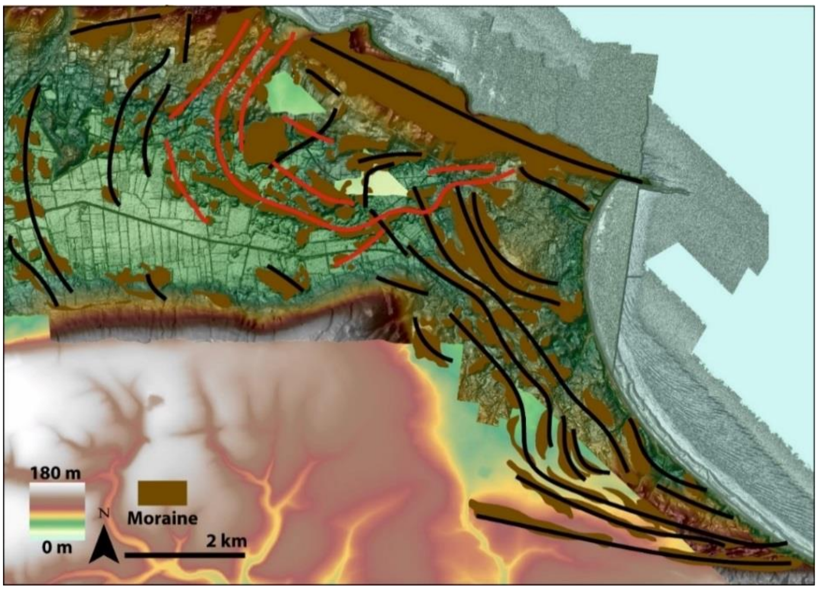


Figure 3.1.20: The moraines at the east coast overlap each other suggesting more than one advance during the LGM.

The moraines mapped at the eastern end along the Gristhorpe Cliffs from Cayton Bay to Filey show several areas of overlapping. This suggests that the ice ingressed on to the east coast more than once. It is difficult to discern which advance came first due to the fragmented nature of the moraines (See section 4.2 for further discussion)

Basins

There are several small basins within the Vale of Pickering (Figure 3.1.21) attributed to a kame and kettle topography associated with a stagnant ice margin (Franks, 1987; Lincoln, 2017). Evidence of these landforms is especially prevalent in the eastern Vale of Pickering between Wykeham and Filey. Lincoln (2017) mapped 43 small basins interpreted as kettle holes in this area.

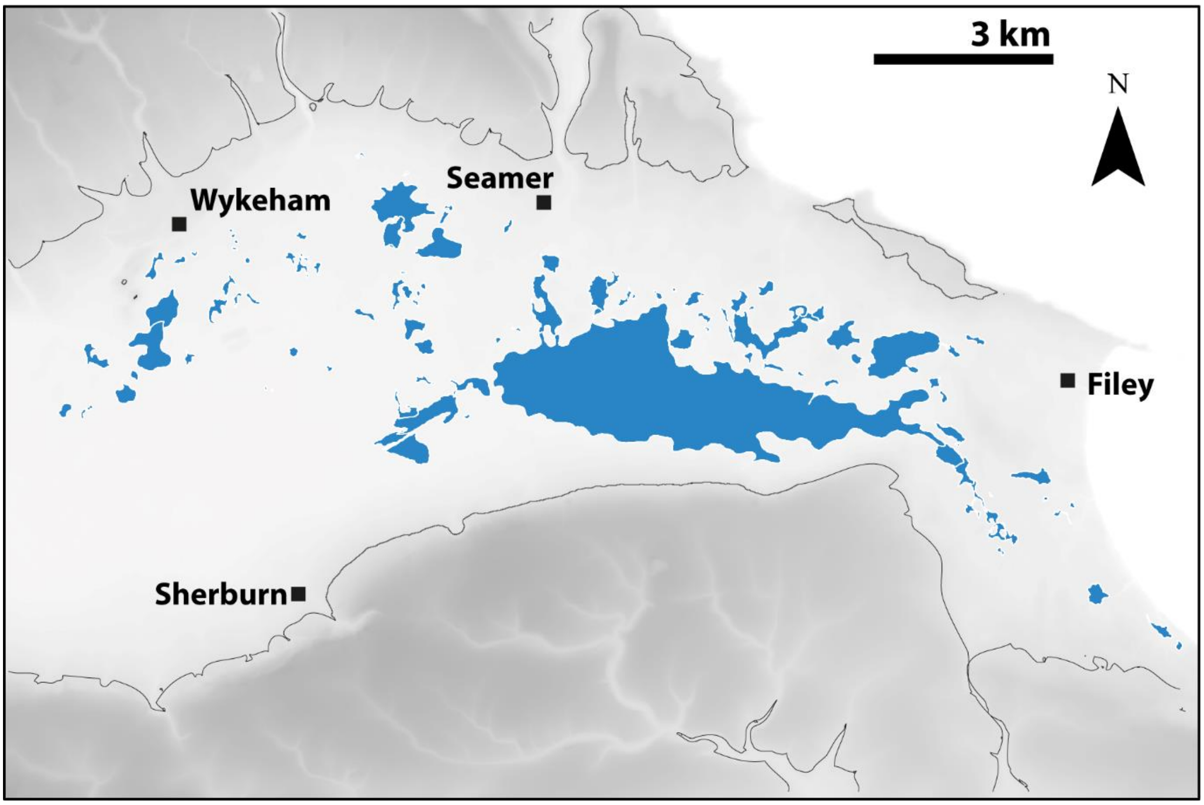


Figure 3.1.21: Basins within the Vale of Pickering. The largest is the Lake Flixton basin with sever small kettle basins present.

The largest basin is the site of post-glacial Lake Flixton, which existed until about 10 ka (Clark, 1954; Humphries, 1994; Palmer *et al*., 2015; Lincoln, 2017). The majority of these basins are comprised of peat and marly clay that are associated with climate warming from the end of the LGM to the beginning of the Holocene. Deposits within the Flixton basin show evidence of climate change from the Loch Lomond stadial and from the Windermere interstadial (Palmer *et al*., 2015; Lincoln *et al*., 2017). These basins were mapped but were not extensively scrutinised so as to not overlap with more detailed work by Lincoln (2017).

Interpretation

The basins are interpreted as post-glacial kettle holes and are typical of the kame and kettle type topography resulting from a stagnating ice front (Franks, 1987; Lincoln, 2017). Lake Flixton is likely the result of a number of small kettle lakes infilling and merging to create one larger water body that slowly silted up, as many of these smaller lakes did (Lincoln, 2017).

Palaeochannels

The Vale of Pickering is surrounded by dry palaeochannels. The majority of channels trend north to south from the North Yorkshire Moors with the channels on the southern escarpment trending south to north. The main exception to this is at Hutton Buscel where a long channel trends northeast-southwest following the northern edge of the kame deposit.

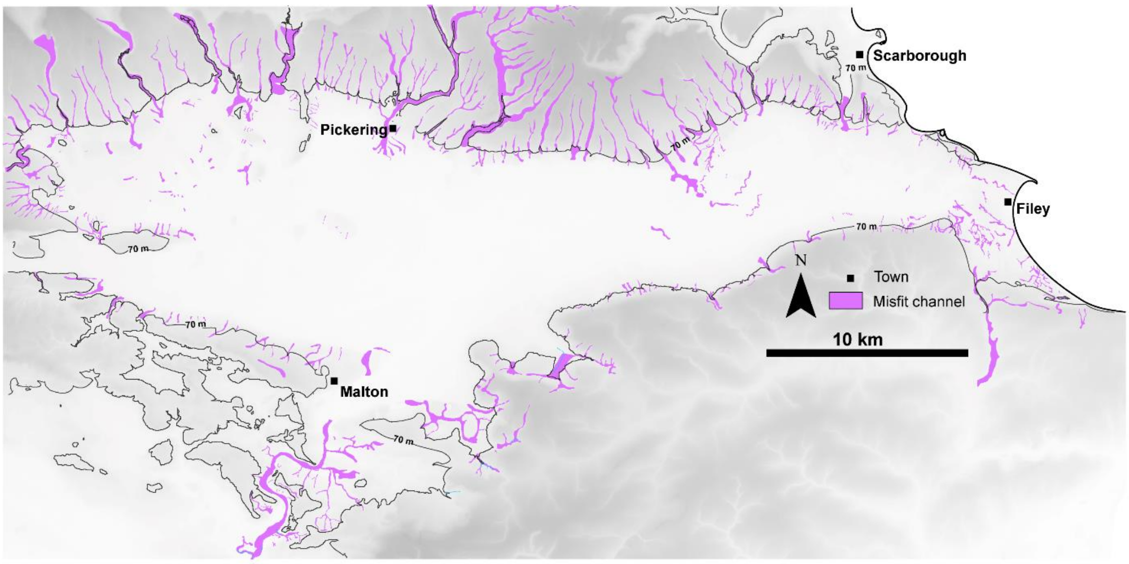


Figure 3.1.22.: Paleochannels of the Vale of Pickering. The majority of channels are situated on the Corallian displope of the North Yorkshire Moors, are long and sinuous, and drain directly into the Vale of Pickering. The channels on the southern side along the Chalk escarpment of the North Yorkshire Wolds are shorter and less incised.

There are still many small active streams in the region that are misfit in much larger channels. Pickering Beck, which flows through Newton Dale, and the Derwent, which flows through the

Forge Valley and across the Vale of Pickering to the outlet at the Kirkham Gorge are two examples. A few small channels from the northern Howardian Hills also still actively drain northwards. However, there are notably less channels along the chalk scarp of the North Yorkshire Wolds.

A total of 384 channels considered to be misfit or ephemeral palaeochannels were mapped (Figure 3.1.22). The average length of the channels is 1.5 km with the longest channel (not including Newton Dale) measuring 13.7 km long and the shortest only 28 m. Channels originating in the North Yorkshire Moors are deeply incised ranging from five to over 90 meters deep with an average depth of 19.5 meters. Along the scarp of the Yorkshire Wolds the channels are less deeply incised with depths ranging from less than a meter to 50 meters. The average channel depth is 10 m but is skewed due to two deeply incised channels dropping to 7 m with these channels omitted.

Channel lengths are noticeably shorter and shallower along the southern margin of the Vale of Pickering than in the north. This is due to the steeper slopes of the North Yorkshire Wolds chalk escarpment and the comparatively short distance to the floor of the lake basin. The channels on the north side on the Corallian dipslope are much more frequent, tend to have many tributaries and are much longer and straighter. The low relief of the channels along the chalk escarpment make them more difficult to map on the 5 m NextMap DEM because they do not show as clearly as the deeper cut northern channels as many of them are shallower than 5 m.

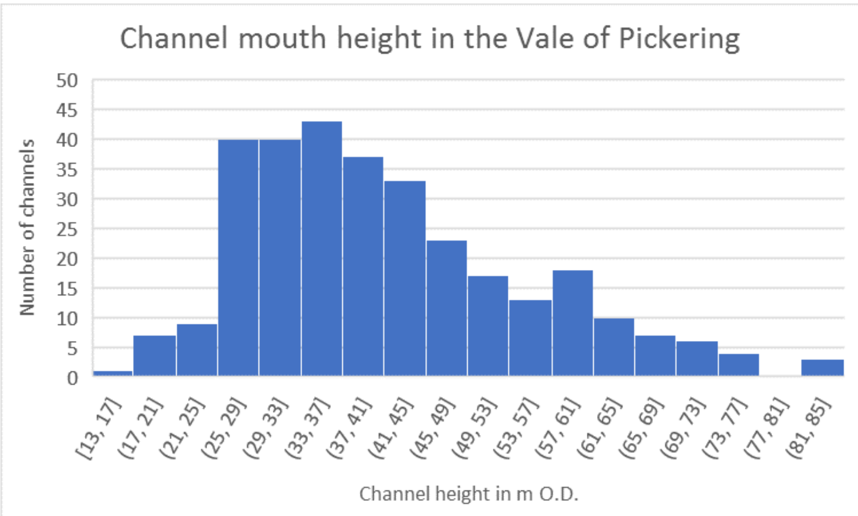


Figure 3.1.23.: Elevation of palaeochannels and dry valleys in the Vale of Pickering. The majority of values are at elevations between 25 m and 49 m O.D with a few channels terminating at higher elevations.

A total of 311 channels that decanted into the Vale of Pickering basin were analysed for the elevation of the river mouth as it enters the Vale. The mean elevation for a channel mouth was 42 m, the lowest was 13 m and the highest 82 m with a standard deviation of 13 m (Figure 3.1.23).

Interpretation

These channels are primarily meltwater channels that were incised and expanded during times of ice advance and retreat (Gregory, 1965). Some of these channels have large deltas at their mouth, like at Pickering, suggesting that they were once very active, but meltwater availability was reduced as ice retreated or water found alternative routes (Gregory, 1965; Edwards, 1978). Some of the channels were formed through spring sapping, such as the channels on the dipslope of the North Yorkshire Moors. This is evidence by the presence of the Hole of Horkham, north of Thornton le Dale. However, shallow laterally incised channels suggest that these channels were utilised by overland flow of meltwater.

Shorelines

Shorelines are the most difficult landform to distinguish in the Vale of Pickering with Kendall (1902) concluding that there were none. Shorelines are best preserved in environments that are relative stable following the recession of the lake body (Carrivick and Tweed, 2012). The Vale of Pickering basin, however, has many paleao- and active channels entering and so it is not surprising that much of the shoreline deposits have been eroded or incorporated elsewhere. However, since the sediment and underlying bedrock are both soft, there are two areas where lake shorelines may be inferred: eroded/reworked alluvial fans and planation surfaces.

The base of the 45 m lake has been discussed in the section on terraces. Figure 3.1.24 shows where the former lake bed has been eroded by a subsequent lake or fluctuating lake and inferred shorelines from 45 m to 28 m, down to the surface height of 25 m are shown. Three small basins on the western side (labelled 1, 2, and 3 in Figure 3.1.24) suggest a higher lake level of 45 m O.D.

West of Pickering at Aislaby Carr (Figure 3.1.25) there are a series of subtle channels running from the North Yorkshire Moors. The shorelines are very feint and difficult to distinguish due to the flat nature of the Vale of Pickering basin, but are seen at 34 m OD, 33 m OD, 32 m OD, 30 m OD, 27 m OD and 25 m OD. They are evident at these elevations from Aislaby to Thornton le Dale and cut into the delta at Pickering.

The extensive alluvial fan at Pickering has evidence of shorelines. At 25 m OD a shoreline extends from the west of Pickering eastwards, turning south as it reaches the alluvial fan it rises to 25.5 m Other shorelines are seen at 24.5 m and at 27 m OD and 30 m OD.

Along the southeast border of the Vale of Pickering, where the basin meets the chalk scarp of the Yorkshire Wolds, there are shorelines cut into the alluvial fans that extend from the base of the scarp at 50 m OD to the floor of the Vale. These shorelines are found at 30 m, 25 m, 25 m, 23 m, and 22 m.

Overall, evidence for shorelines is very sparse. The most convincing evidence is for a lake of ~33 m, 28 m O.D., and 25 m O.D.

Interpretation

Two small hummocks in the middle of the Vale of Pickering, where the towns of Great Baraugh and Kirkby Misperton are located also show some geomorphic evidence of a 33 m lake. Near Kirkby Misperton, there a are a number of channels and nickpoints that lake waters reached 33 m O.D, with lower terraces at 27 m O.D., 26 m O.D and 23 m O.D. (Figure X). At Great Baraugh, a bench at progrades from 29 m O.D. to 26 m O.D is comprised of sand and gravels suggest reworking of sediment by lake waters at this elevation. Opposite, at the western end of the Kirby Misperton hummock, a similar deposit sits at 26 m O.D. These have likely been subsequently reworked by draining lake waters which gives them an unusual shape. Another unusual feature is the small embayment in the south west section of the Kirby Misperton hummock. This feature reaches 32 m OD. Grading down to 27 m O.D and was likely eroded by lake waters. Similar to this feature is an abandoned channel that dissects the north eastern corner of the Kirby Misperton hummock, again showing terracing at 32 m O.D and 30 m O.D.

In the north central portion of the Vale of Pickering between the long north-south trending spur of Riseborough to the west and the extensive delta at Pickering to the east, there is evidence of shorelines from 33 m O.D to 26 m O.D. There are two hummocks of gravel overlying mottled orange and brown lacustrine clay. These deposits have been recorded in historic boreholes as "boulder clay" but they are more likely to be dissected and reworked alluvial deposits from the dales that have been eroded post-glacially with the eastern hummock likely to be a remnant of the Pickering delta.

The Pickering delta is terraced by lake waters although it is unclear exactly when the delta was formed. One historic borehole record from the centre of the delta (Fox-Strangways field notes, 1881) records 1.8 m of clay overlying 5.4 m of gravel. On the western flank of the delta, a suite of boreholes drilled in 1987 show silty clays, gravelly clays, and gravel ranging in thickness between 2.2 m and 6.1 m overlying Jurassic mudstones. From the boreholes, however, it would appear that the gravels of the Pickering delta sit on top of bedrock Jurassic clay, but it is uncertain when these gravels were deposited, and the lack of boreholes make it difficult to ascertain the relationship between the delta and the lake shorelines.

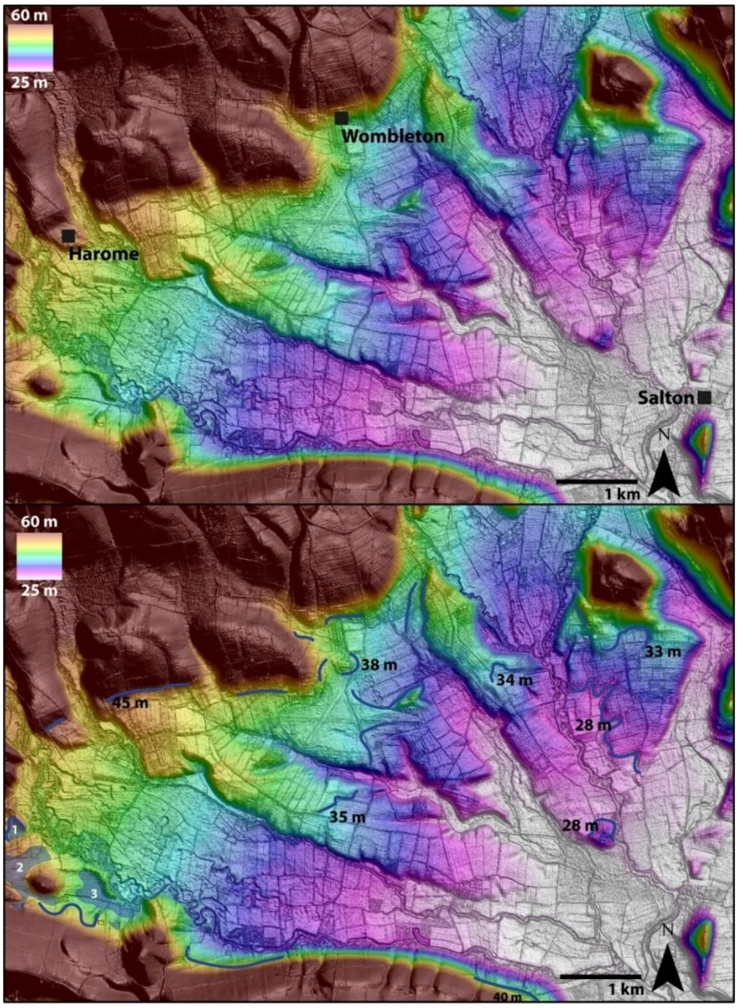


Figure 3.1.24: The western Vale of Pickering is dominated by the fluvially eroded former lake bed. It has been cut into terraces with shorelines evident between 45 m O.D and 28 m OD.

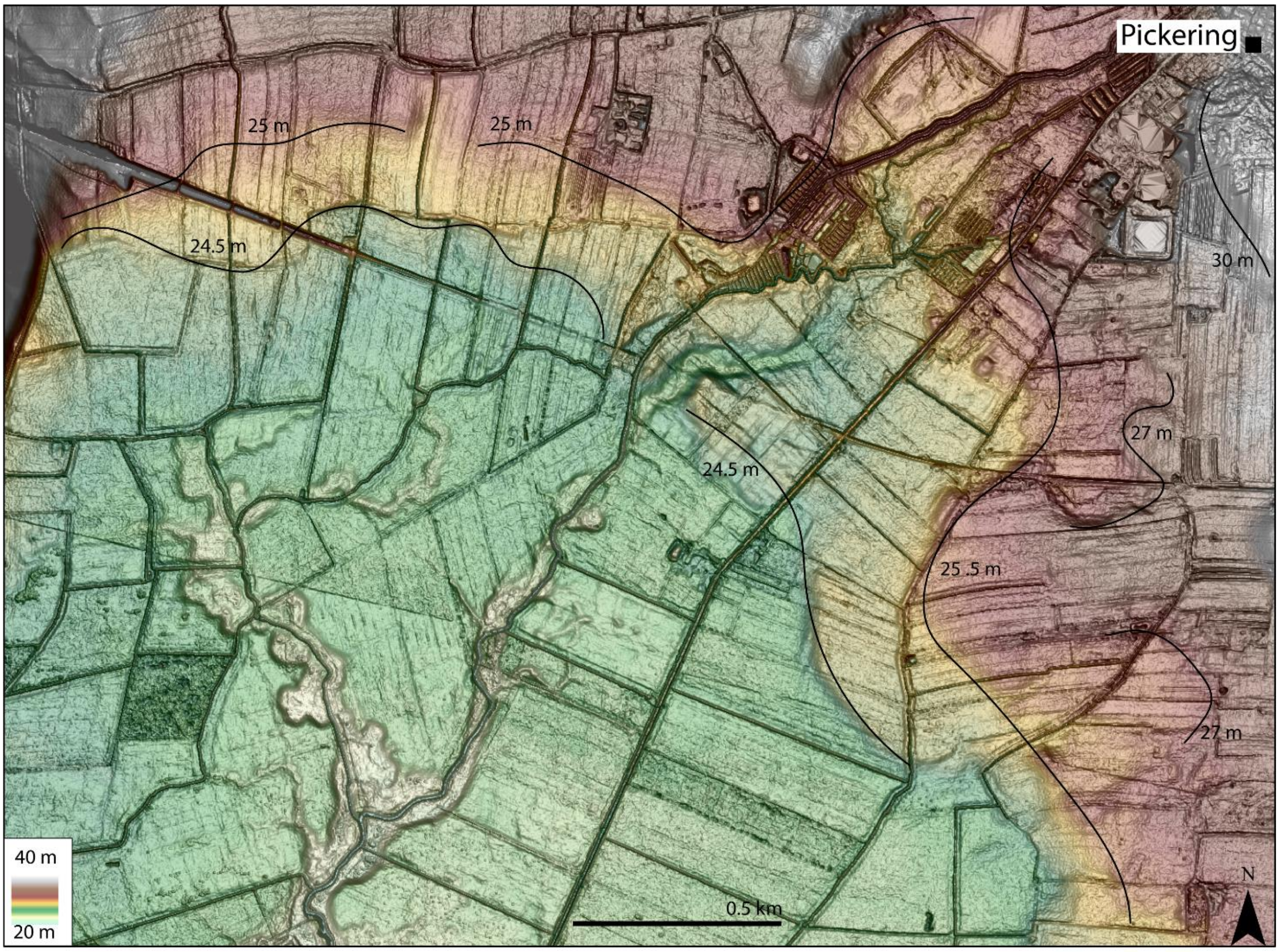


Figure 3.1.25: The shorelines surrounding Pickering cut into the extensive alluvial fan. These shorelines extend eastwards towards Thornton le Dale and westwards towards the clay scarp of Riseborough. Shorelines are seen at 25.5 m, 25 m, 25.5 m, 27 m, and 30 m.

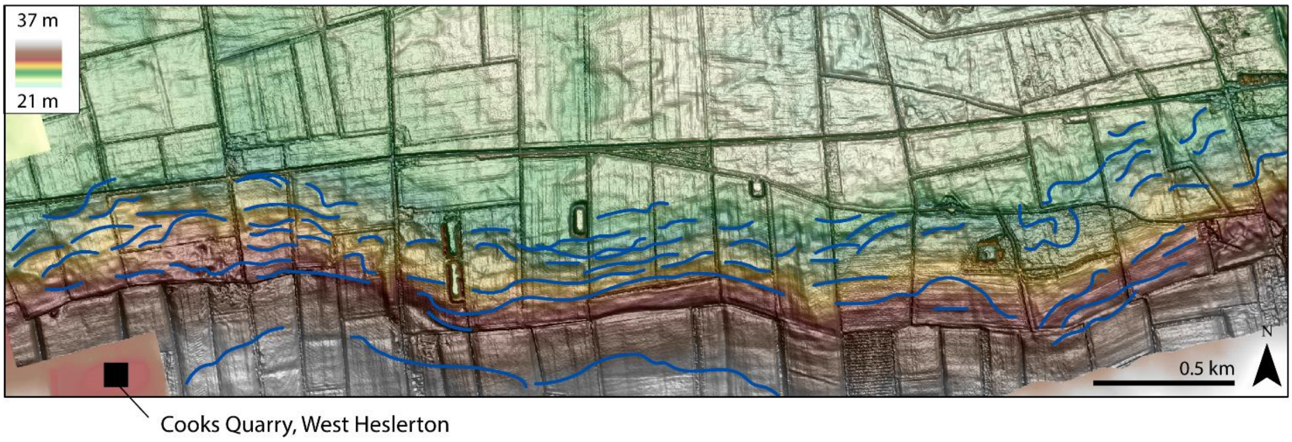


Figure 3.1.26: Shorelines running along the southern Vale of Pickering against the scarp of the Yorkshire Wolds. Shorelines are seen at 30 m, 25 m, 24 m, 23 m and 22 m

East of the Pickering delta some lake shoreline evidence is interpreted towards Thornton Le Dale. Thornton le Dale sits between two large outcrops of Kimmeridge clay with a diamict cap that has been suggested by Edwards (1978) as evidence of ice advance further into the Vale of Pickering. On the western hummock, an alluvial fan spreads westward that shows a planation surface at 28 m O.D before dropping down to 25 m O.D. At the base of the western hummock, a north-south trending deposit of gravels interpreted as an older, incised delta or alluvial fan by Edwards (1978), Foster (1985), and Franks (1987) shows some evidence of terracing at 27 m O.D on the south eastern side. A topographic low of 31 m O.D suggests erosion of material by lake waters.

On the southern side of the Vale of Pickering, shorelines for 33 m lake are hard to locate. This is partly due to the deposit of the Sherburn Sands between Knapton and Ganton. Along the scarp of the Howardian Hills between Hovingham and Malton, LiDAR imagery is unavailable, which makes the identification of the subtle shorelines difficult. These deposits are also covered by post-glacial fluvial deposits and are difficult to identify. The topography around Hovingham is more hummocky than in the central Vale of Pickering, although there is some evidence of planation at 29 m O.D.

Kirkham Gorge

The outlet for Lake Pickering is thought to have been through the Kirkham Gorge and into the Vale of York. Prior to utilizing the Kirkham Gorge, Lake Pickering drained eastwards into the North Sea. Previous workers (Reed, 1901; Kendall, 1902; Straw, 1979; Edwards, 1978; Foster, 1985; Franks, 1987) have all noted that the build-up of moraines at the eastern end of the Vale of Pickering, as well as ice in the Scarborough, area prevented drainage to the North Sea and instead the River Derwent reversed course to Malton and the Kirkham Gorge. Furthermore, any connection between Lake Pickering and Lake Humber should be reflected in the geomorphology of the Kirkham Gorge.

Prior to the LGM, it is unclear if previous iterations of Lake Pickering drained to the east or through the Kirkham Gorge. While the Forge Valley was carved during the LGM (Franks, 1987), it is uncertain if all or some of the Kirkham Gorge was created in the same manor. Mapping of watersheds and tributary streams in the Howardian Hills where the Kirkham Gorge cuts through many faulted limestones, shows that the river's course was initially into the Vale of Pickering through the Malton Embayment. Reed (1901) established that this was a small tributary channel of a larger proto-River Ure whose course ran through the Coxwold-Gilling Gap and along the scarp of the North Yorkshire Wolds where the Sherburn sands are found today.

Mapping in the Kirkham Gorge reveals a series of river terraces between 26 m O.D and 21 m O.D (Figure 3.1.27). These terraces are undated. They are important as the height of the base level of the Kirkham Gorge has been shown to influence the drainage of the lake (Section 3.5).

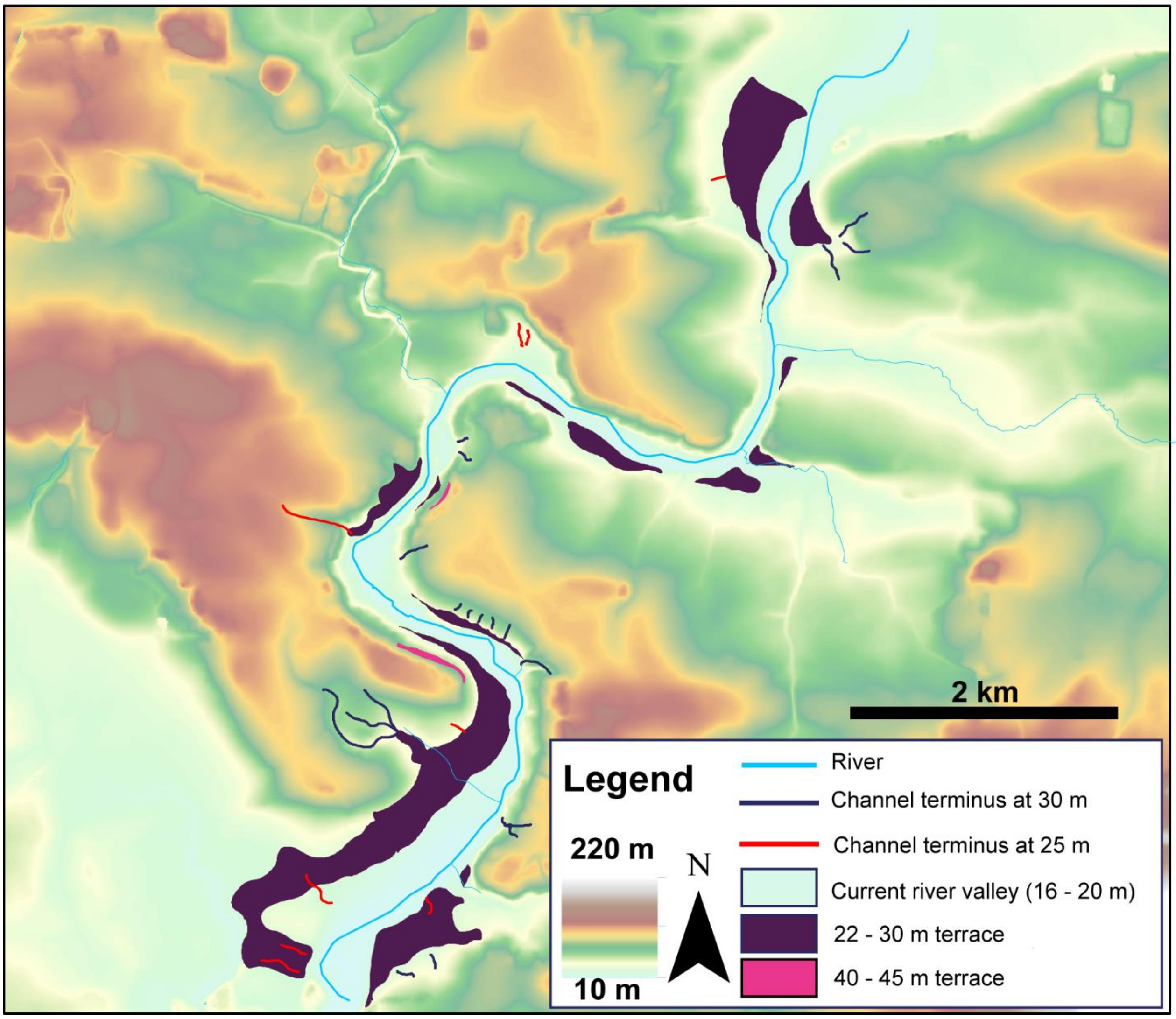


Figure 3.1.27: terraces in the Kirkham Gorge lie between 30 m and 22 m O.D with one potential terrace as high as 45 m and 40 m O.D. Channels terminate at similar levels of the terraces at 25 m and 30 m O.D.

Interpretation

It is likely that parts of the Kirkham Gorge were cut prior to the LGM and ran in a north-easterly course, debouching into the Vale of Pickering at Malton. The shape of the palaeovalleys cut along the many east-west trending faults suggest that the course of the gorge was already established (Figure 3.1.28) and that the river has been downcutting along these fault planes. However, there has been no investigation into the timing of the river valley development in the Kirkham Gorge and so dating the reversal of the River Derwent falls to dating the blockage of the eastern end of the Vale of Pickering during the LGM as a minimum age. There is some erosion at Eddlethorpe Hall through a col at 40 m O.D. that suggest that the base of the river was much higher prior to LGM. This is likely because the river through the Kirkham Gorge would have been much smaller, possibly several different river networks draining along the many faults of the Howardian Hills and it is possible that the reversal of drainage created a larger capacity river leading to increased downcutting of the Gorge. Terraces at the mouth of the Malton Embayment suggest downcutting started at a lake level height of between 30 m and 25 m O.D and may have been impounded by the limestone ridges at High Hutton and Low Hutton (Figure 3.1.29).

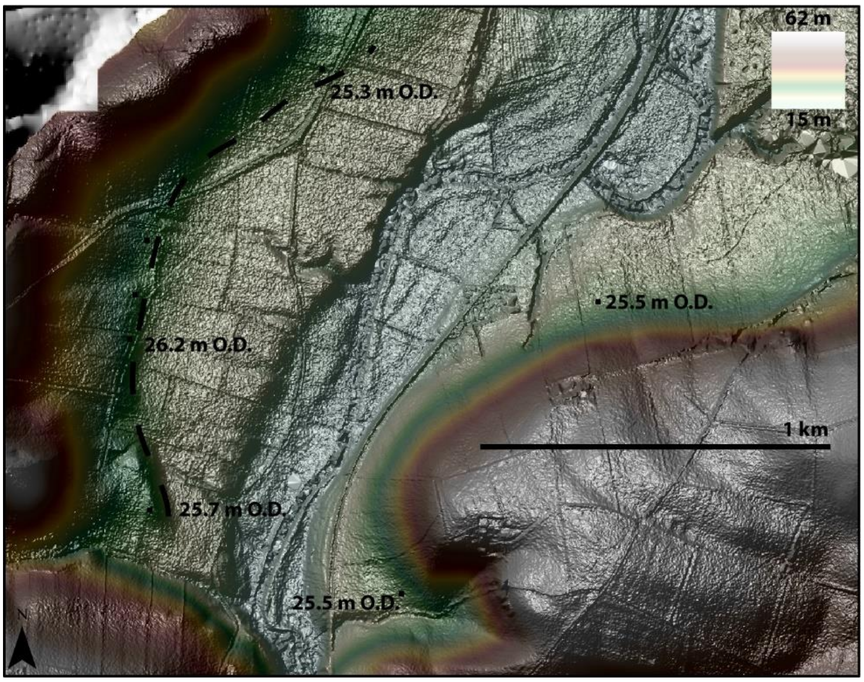


Figure 3.1.28: Terraces from the River Derwent in the Malton Embayment grade from 26.2 m down to 25 m and the floor of the current river bed is 21 m O.D lowering to 15 m O.D as the river courses through the Kirkham Gorge

It is difficult to map terraces in the Kirkham Gorge due to the southerly dip of the limestone and the There is little diamicton or lacustrine deposits in the Kirkham Gorge so any connection between Lake Humber and Lake Pickering was likely short-lived as no significant terracing can be traced.

Borehole logs (#958; #968) from the A64 along the terrace into the Vale of Pickering from High Hutton to Malton show orange mottled sands and clays, while a borehole (#906) from Low Hutton shows the older diamicton noted by Powel *et al*. (2016) overlying laminated Jurassic clays (Kimmeridge clay) (Figure 3.1.29). This suggests that the terraces cut could either be very old as little diamicton remains or very young with the diamicton having been eroded. The Kirkham Gorge is undoubtable a site that with further investigation could lend a lot of information about the history of lakes in the Vale of Pickering.

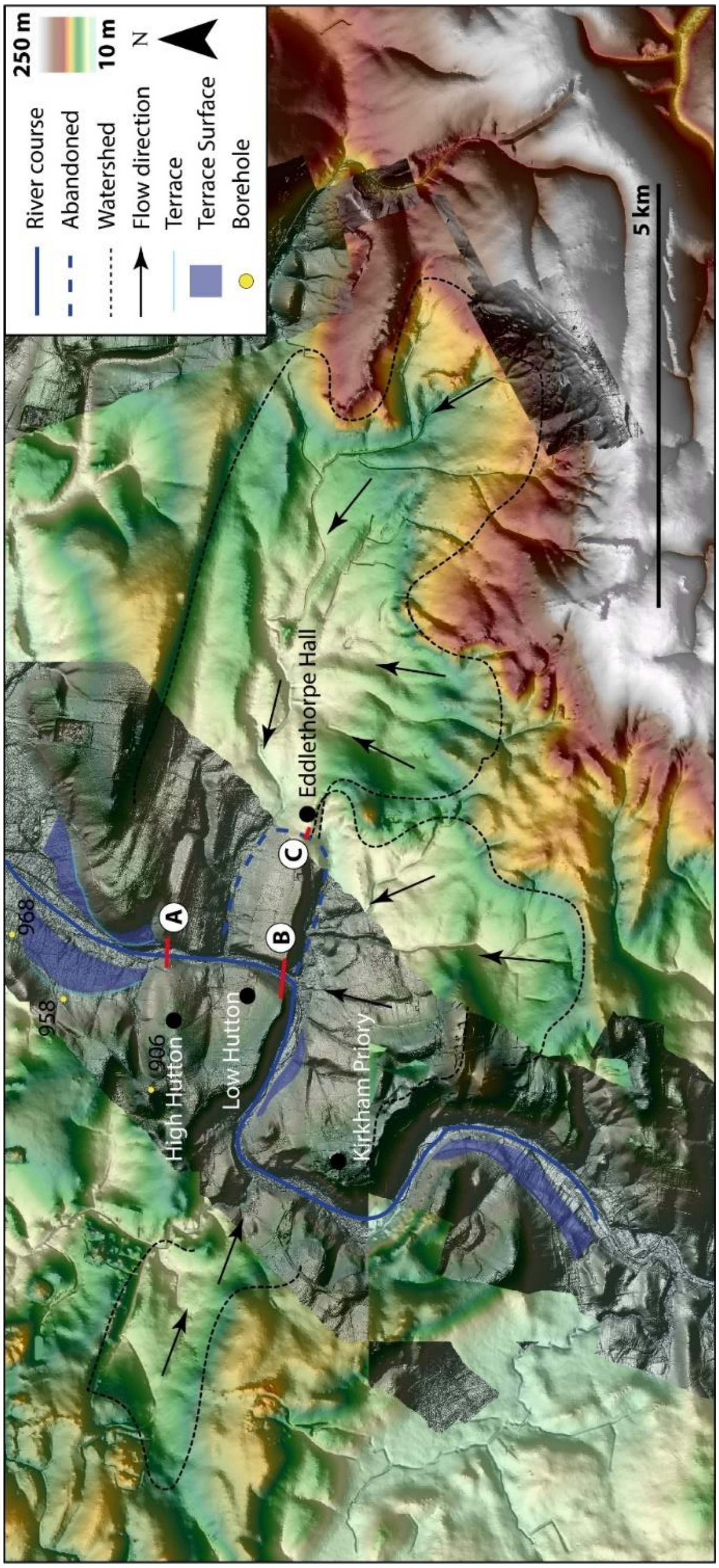


Figure 3.1.29: The geomorphology of the Kirkham Gorge is highly complex, and it is difficult to ascertain when the waters of Lake Pickering may have breached the limestone scarp. It is likely that it occurred near High Hutton (point A) where the scarp is only about 30 m O.D. and the velocity helped to erode the scarp at Low Hutton (B). Prior to the LGM, rivers in the Kirkham area of the Howardian Hills flowed northwards into the Vale of Pickering Basin.

3.1a Historic borehole data

To help establish geomorphic relationships in mapping, historic borehole data was used to understand the subsurface deposits. The most significant borehole data set is comprised from a 25 km line of boreholes running across the Vale of Pickering from Malton to Irton Moor, Scarborough (Figure 3.1.30) from work completed between 1964 and 1968 by the Central Electricity Generating Board (CEGB). These boreholes were collated into a 2D cross section by Edwards, 1978 (see section 1.6). The data set also included particle size analysis from units within the boreholes, hydraulic conductivity, and stress from overburden. The PSA data was used to ascertain whether there was any relationship between the historic borehole data and the core extracted at High Marishes and Yedingham (See Section 3.2).

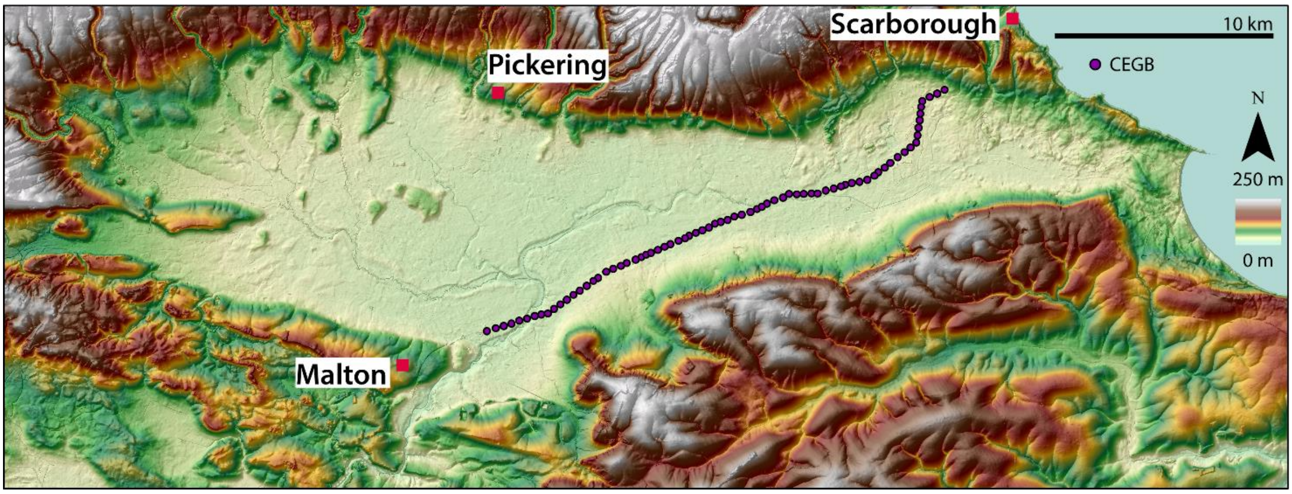
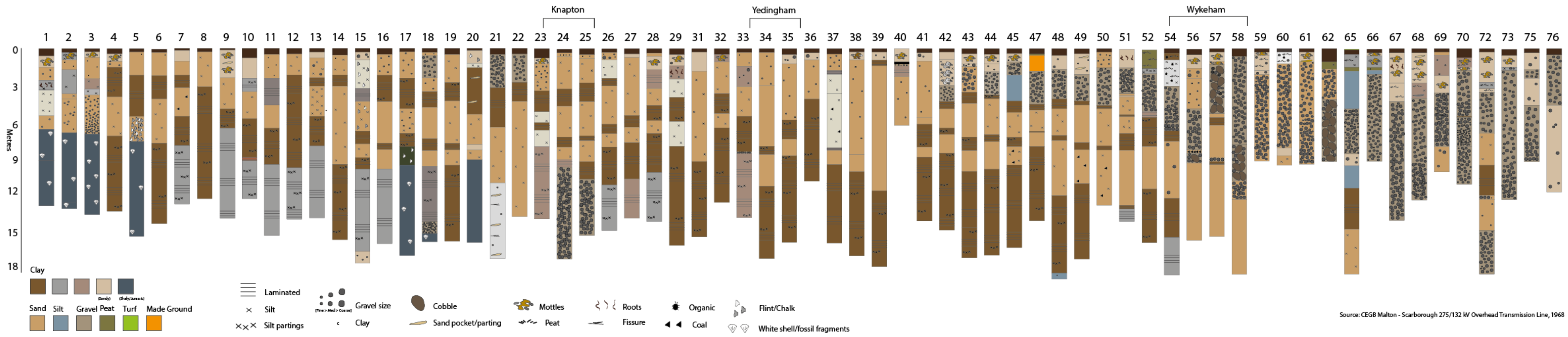


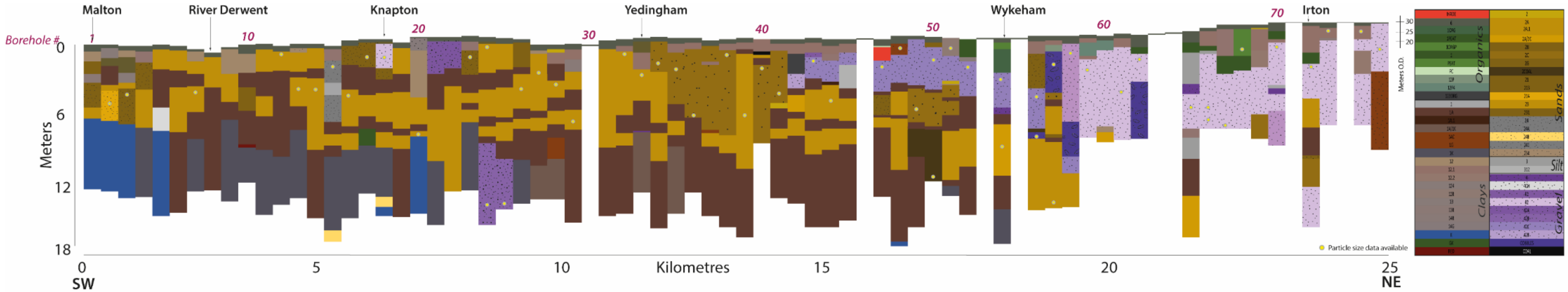
Figure 3.1.30: The line of boreholes stretching from Malton to Irton Moor, Scarborough. The logs come from work carried out by the Central Electricity Generating Board (CEGB) between 1964 and 1968.

The logs were drawn and are presented in Figure 3.1.31. The log shows a coarsening of sediments towards the east with glacial sands and gravels overlying lacustrine deposits in the eastern section. These are the Yedingham gravels, the Hutton Buscel sands and gravels some of which are capped by Holocene alluvium and peat in the east (Edwards, 1978; Franks, 1987; Lincoln, 2017). A second cross section was drawn to more accurately represent the 2D architecture and was adjusted to the boreholes

Next Page: Figure 3.1.31: (Left) The CEGB borehole line drawn per log. (Right) The CEGB borehole line drawn with 400 m increments and presented with borehole elevation start heights

Borehole cross section. Malton - Scarborough (CEGB)





Borehole cross section of the Malton to Scarborough 275/132kV Overhead Transmission Line
Central Electricity Generating Board, 1968.

height in meters O.D and spaced at 400 m intervals (Figure 3.1.31). The boreholes reached a depth between 10 and 18 m.

The section shows the bedrock of Jurassic clays at the base of the boreholes 1, 2, 3, 5, 18, 20 and 48 (Shown in blue in Figure 3.1.31). These instances are more common near Malton where the depth to bedrock is shallower than the central basin. An outcrop at the base of boreholes 18 and 20 suggests a palaeovalley heading into the Malton embayment. Past borehole 20, the Jurassic clay is not encountered again until borehole 48, due it being one of the deepest boreholes drilled to a depth of 18 m. This illustrates the depth of the central palaeovalley identified by Ford *et al*. (2015) (section 1.3). Logs describing the units of Jurassic clay show that it occasional contains fragments of shells and chalk gravel (CEGB, 1968).

The CEGB cross section shows lacustrine clay lying above the Jurassic bedrock. These clays are almost continuous and are located at a depth between 6 m and 10 m. They are primarily described as grey-brown and are occasionally laminated. There is no record in these deposits of organic material or shells.

Above the lower clay is a discontinuous deposit of medium brown sand, occasionally with inclusions of coal. In borehole 15, the deposit is described as grey coarse sand with chalk gravel. This is the only unit of this description in the whole cross section. These sand deposits are sometimes capped by a second layer of lacustrine clay described as a brown clay with occasional laminations. In boreholes 15-20, 23-26, 28, 29, 33, 43-46, and 48-53 there are three or four alternating layers of sand and clay.

From Yedingham to Irton Moor, the boreholes reveal a coarse gravel overlying the lacustrine deposits. These gravels increase in size towards the east with deposits of diamicton in boreholes 56-58 and 63 shown in dark purple (Figure 3.1.31). The gravels are likely glaciofluvial from the Hutton Buscel and Forge Valley deposits (Franks. 1987) that have been reworked as the lake levels dropped and the rivers redistributed the overlying glacial sediment during the LGIT (Lincoln, 2017). The change in height eastwards is also important as it reveals the extent to which the landscape has been altered by the presence of moraines and glacial gravels. The borehole cross section turns north-eastwards at borehole 65, but the elevation caused by the coarse sediment is apparent as far west as Yedingham. It is this topographic high that contributed to the reversal of the River Derwent towards the south to the outlet at the Kirkham Gorge.

Overall, the CEGB borehole cross section reveals many iterations of Lake Pickering. The sequence of lacustrine episodes is likely related to the position of the NSL. As the NSL advances and retreats off the Yorkshire coast (Catt, 2007; Roberts *et al*., 2018), the lake outlet at the sea is opened allowing the lake to drain. However, as glacial sediment is built up by moraine deposits at the eastern end of the Vale of Pickering, these sediment barriers create dams, slowing the drainages of the lake and allowing the sands to remain waterlogged. As the NSL readvances onto the Yorkshire coast, the drainage is halted, and conditions return to low-energy allowing the deposition of lacustrine clay.

In the western Vale of Pickering, historic boreholes reveal an older lacustrine floor overlying Jurassic clay. These boreholes reveal that downcutting by the River Rye from Helmsley to Salton has eroded and reworked the clays.

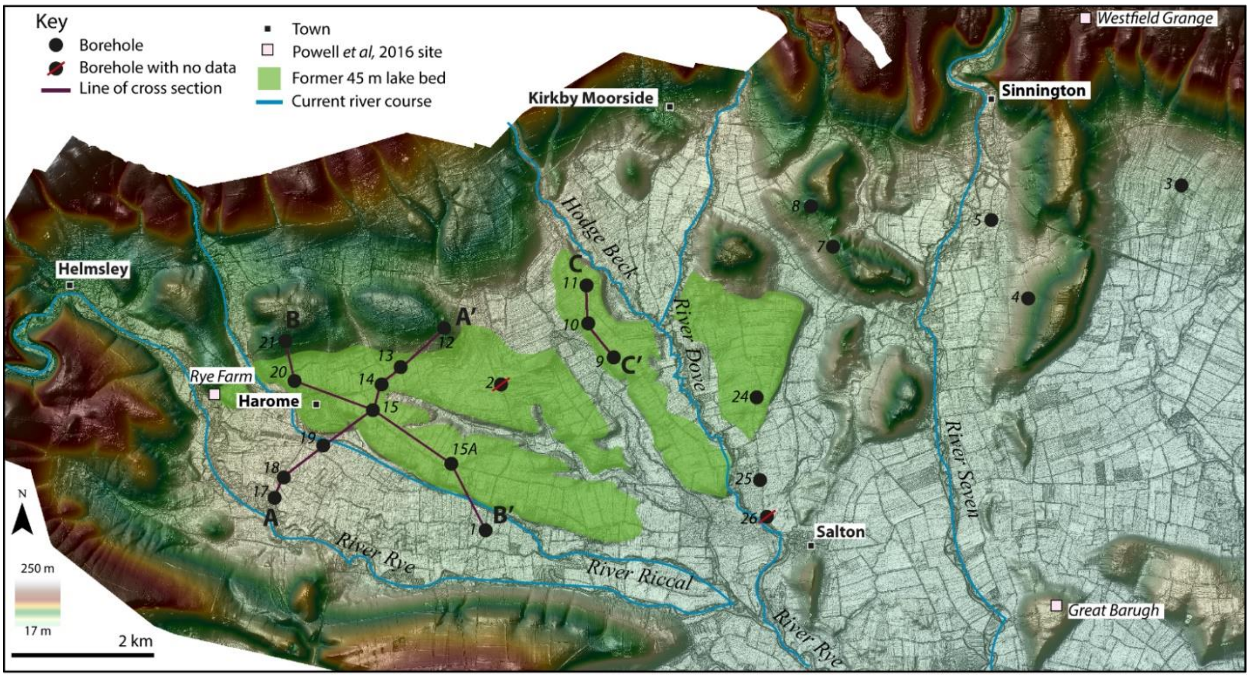


Figure 3.1.32: Cross sections using historic boreholes in the western Vale of Pickering were used to establish the presence of a former lake bed (see accompanying borehole logs)

The stratigraphy of the boreholes here is very different from the CEGB boreholes. Cross section A-A' (Figure 3.1.X; Figure 3.1.X Panel A) shows in the Rye river valley (Boreholes 17, 18, and 19) the bedrock of Kimmeridge clay underlying sands and gravels with a layer of reworked alluvial sands and clay. Once cross section A-A' reaches the hummock of the old lake bed

(Section 3.1) boreholes 15, 14, 13, and 12 show a Kimmeridge clay underlying medium large gravel and sand, with a mottled clay cap thought to be the base of the 45 m lake. This stratigraphy is confirmed in Powell *et al.* (2016) in a borehole taken from Harome. The same stratigraphy can be seen in a borehole cross sections B-B' and C-C' (Figure 3.1.X and 3.1.X). This layer is undated, however. This is an area suitable for further investigation to both constrain the timing of the 45 m lake and could potentially be used to measure the rate of GIA using the relationship between the downcutting of the Rye river valley.

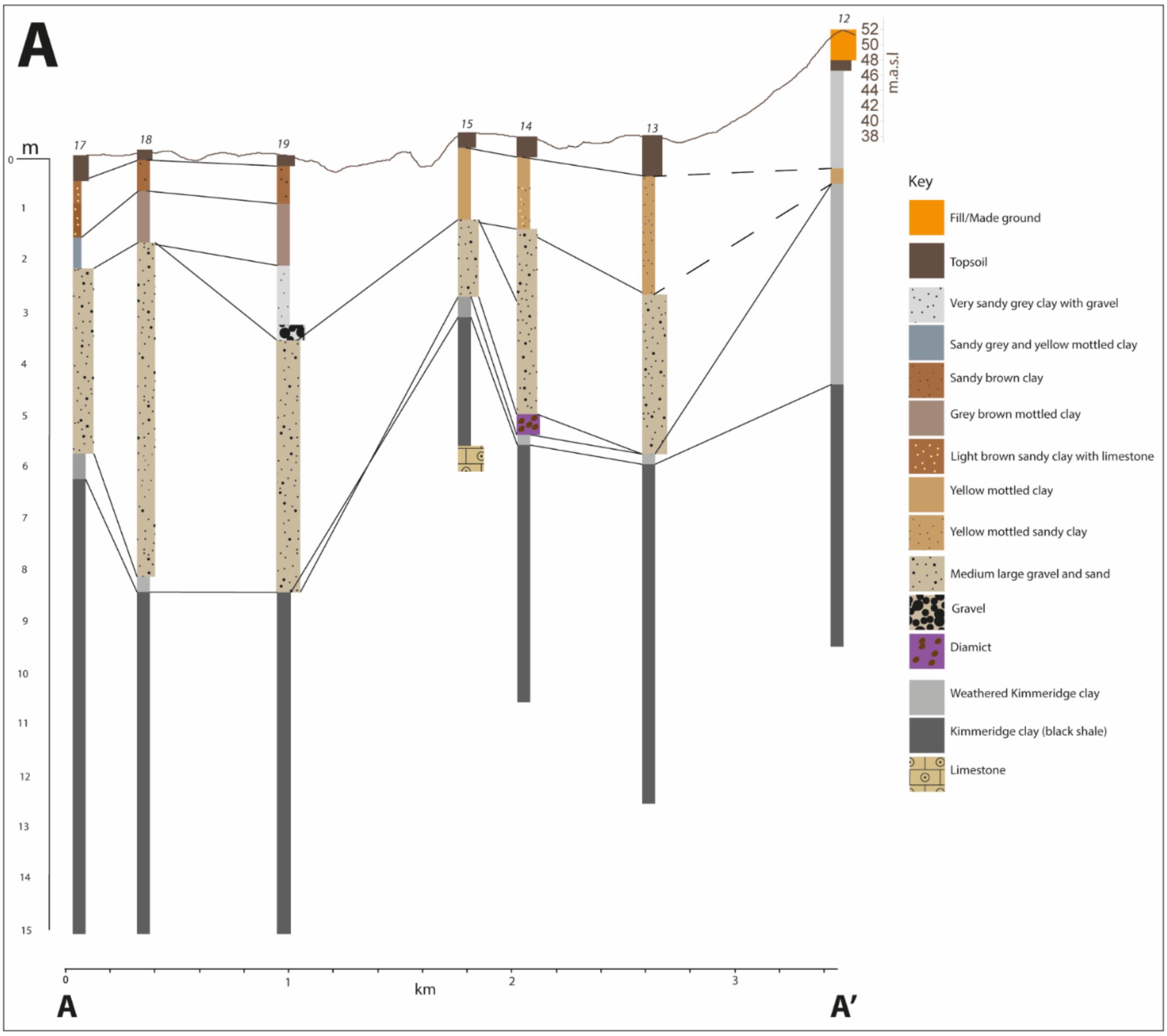


Figure 3.1.33: A-A' from Figure 3.1.32. Boreholes 17-19 show a much sandier clay than boreholes 13-15 which show a weathered mottled clay.

The historic boreholes used to help understand the subsurface deposits are invaluable especially as the Vale of Pickering has very few open exposures and quarries due to the high water table.

These cross sections show a complex depositional history that differ between the eastern and western ends of the basin. The potential for further work is revealed by these records.

Logging descriptions for the Helmsley series are rudimentary, but there are consistencies between the logs and the geomorphology of the area. For example, boreholes (Helmsley 9, 10, 11, 13, 14, 15, 15a, and 24) that were drilled through the raised hummocks are remarkably consistent. They are comprised of deposits of medium to large gravel and sand between 1 m and 6 m thick, underlying a mottled yellow clay (sometimes described as sandy) between 1 m and 1.5 m thick, capped by a topsoil of 50 cm or less. Logging descriptions for the boreholes that are situated in the river valley flood plain (Helmsley 1, 17, 18, 19, and 25) are more varied. Three cross sections (A-A'; B-B'; C-C') highlight the relationship between the topography and borehole stratigraphy.

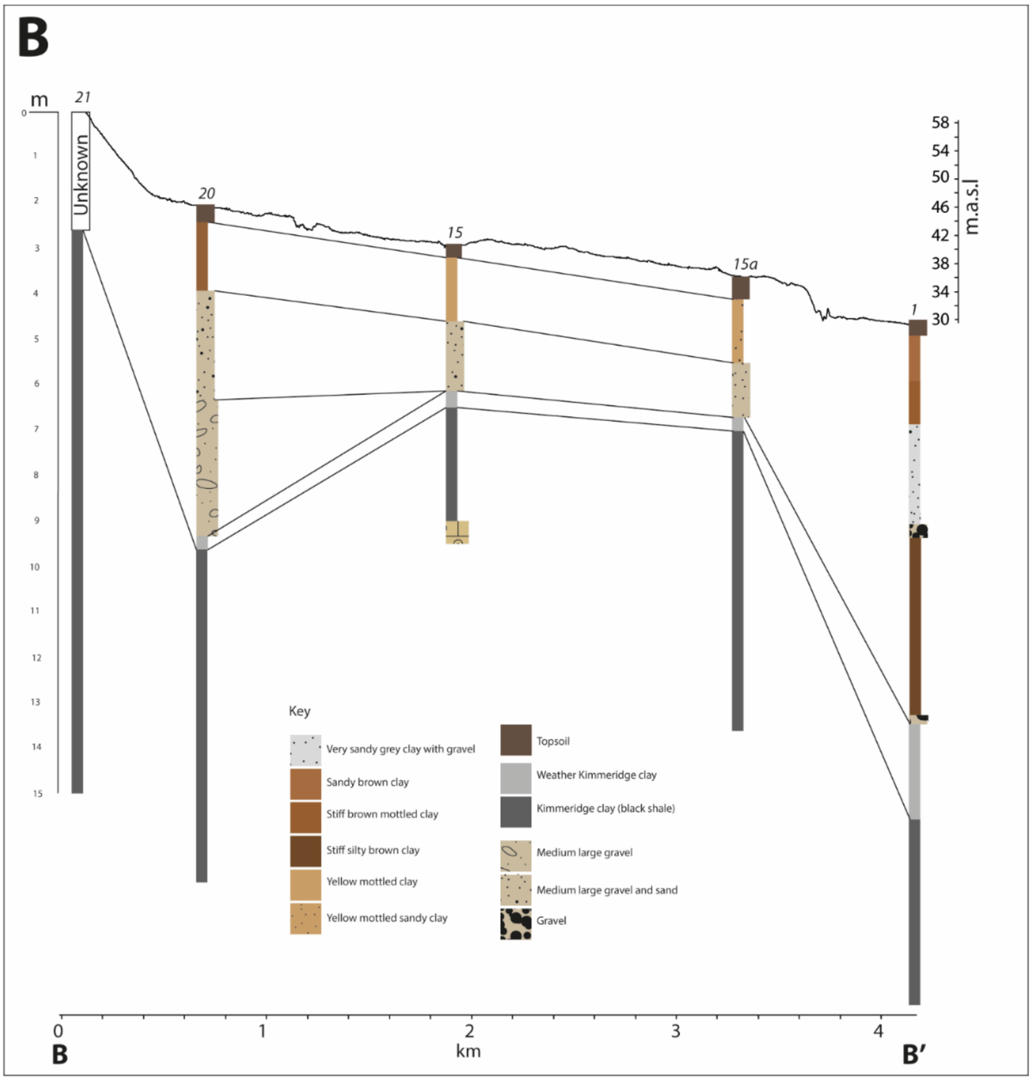


Figure 3.1.34: Section B-B' in the western Vale of Pickering shown in Figure 3.1.32.

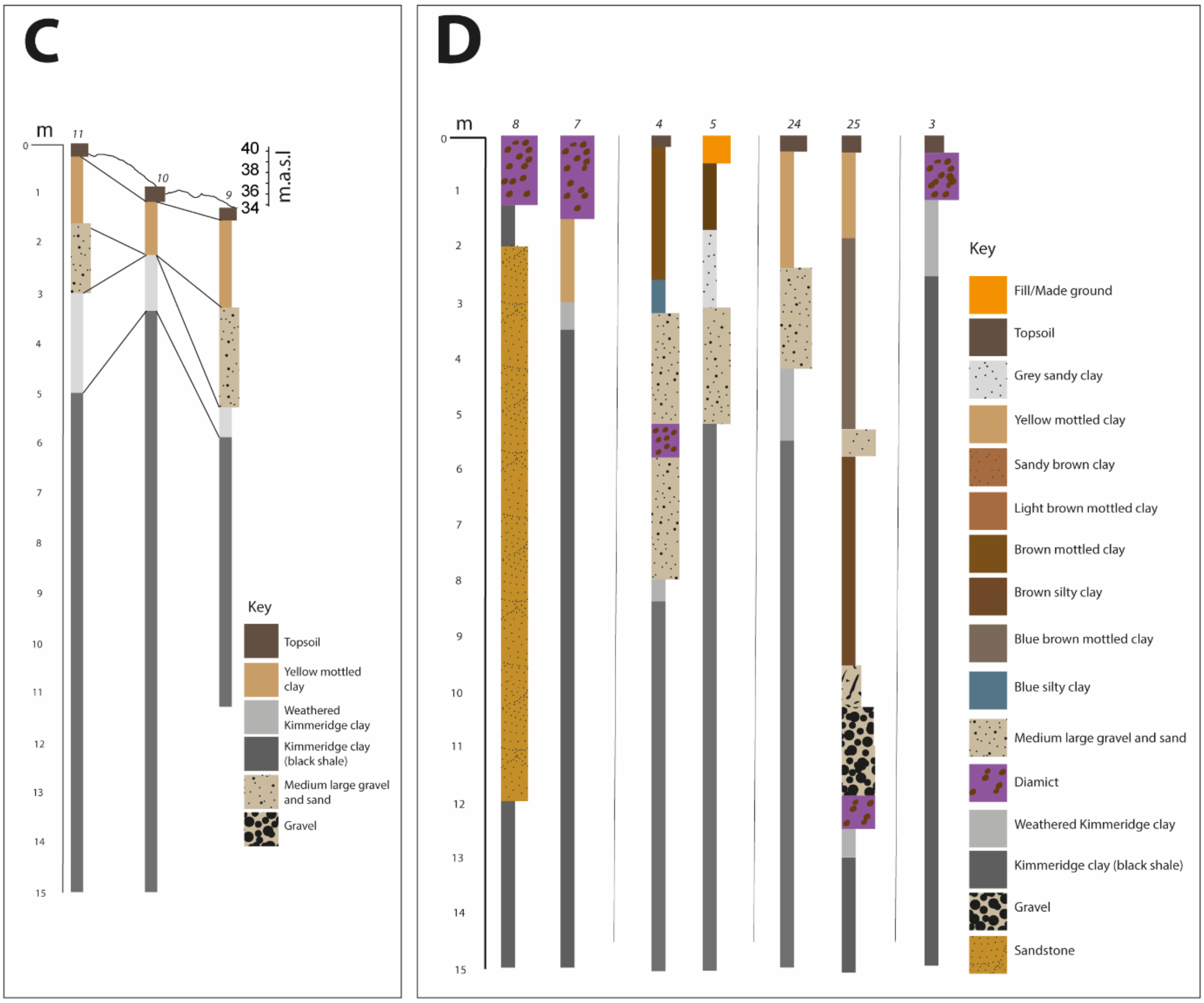


Figure 3.1.35: cross section C-C′ from the western Vale of Pickering shown in Figure 3.1.32. D shows the remaining boreholes without any stratigraphic relationship shown.

There are seven additional boreholes scattered to the east of the River Dove that are not included in the cross sections, but relationships between the topography and the stratigraphic logs remain consistent. Borehole 24 shows the same medium to large gravel and sand underlying a mottled yellow clay capped by topsoil. Boreholes 5 and 25 both show varied deposits of mottle alluvial clay consistent with others in the river valleys. There are extensive meanders of the River Rye and River Riccal and it is likely that due to the erosion of the older lake bed, the clays have been redistributed and reworked into many of the two-toned mottled clays seen in the borehole logs (Helmsley 1, 17, 18, 19, 25). None of the clays are described as laminated.

In boreholes 3,4,7, 8, 14, and 25 the stratigraphic description states “boulder clay.” Since this term is outdated, the deposit was changed to diamicton as in some instances (borehole 3) the deposit is not boulder clay but has been proved in the field and from conversations with landowners to be alluvial gravels washed down through the misfit channels that enter the Vale of Pickering from the North Yorkshire Moors. Powell *et al.* (2016) proved the presence of diamicton (interpreted as MIS 8 till deposits) at Harome, Great Barugh, and Cropton. Without further investigation and dating, the 45 m lake age is only speculated to be pre-Devensian due to the degree of erosion compared with lower elevations and the apparent weathering of deposits described in historic boreholes.

3.2 Sedimentology

Ings Farm Quarry 1, Yedingham

Ings Farm, Yedingham, is located approximately 6 km south west of the Wykeham moraine. The area is low relief at 21 m OD on the Derwent floodplain (Figure 3.2.1). The current course of the River Derwent has been artificially straightened, but the course of the Old Derwent is visible on 2 m LiDAR and OS maps (Figure 3.2.1). There are two quarries located at Ings Farm. Quarry 1 is in the south-east corner. It forms a long rectangle and is the newer of the two quarries to be opened. Quarry 2 is to the north of Quarry 1 and is now a pond (Figure 3.2.1).

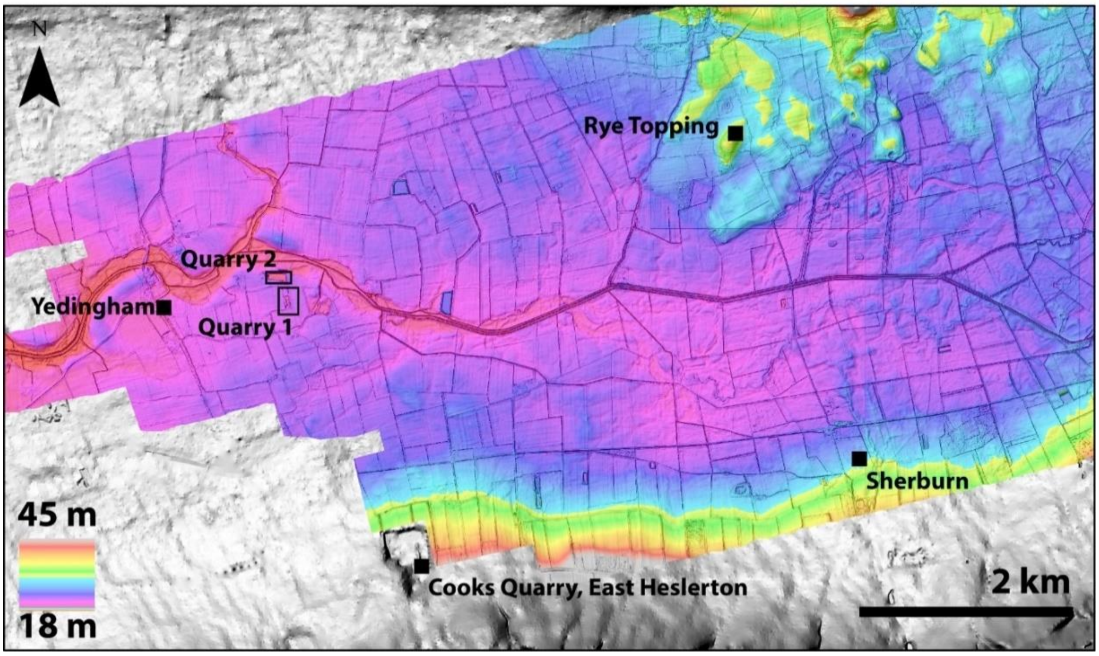


Figure 3.2.1: Location of Quarry 1 and Quarry 2 (outlined in red) at Ings Farm, Yedingham.

In September 2015, 2.5 m of the exposed quarry top were logged and sampled for sedimentological analyses and for OSL dating. In an attempt to extend the section downwards, the British Geological Survey provided a DANDO percussion corer to extract two 2.5 m cores (one in an opaque liner), from the floor of quarry 1 at Yedingham. Coring was suspended at this point due to hitting an impenetrable clay layer (Figure 3.2.2). Samples taken from the opaque borehole and sediment exposures in both quarries were dated using optically stimulated luminescence dating. This dating programme at the age of glaciolacustrine deposits, establish when Glacial Lake Pickering first formed and examine whether there is evidence for lake draining and refilling during the Late Quaternary.

Lithofacies Analysis

LF1 Is a grey-brown clay (Figure 3.2.6). This unit represents the base of the section. Historic boreholes from Yedingham and the CEGB (Figure) prove this clay to extend to at least 10 m depth. Above this clay base, lies LF2 comprising of 2 m of moderately sorted, light brown, medium grained sand (M_z 1.8 ϕ) with occasional weak laminae and coal streaks towards the top of the unit (Figure 3.2.6). The unit fines upwards to a mean particle size of 3.0 ϕ as the silt fraction increases above 3.6 m (Figure 3.2.6). LOI shows between 1 and 3% organic matter (OM) from the base of LF2 to 3.7 m where it spikes to 4.3%. Above this, the OM% drops to <1% for the continuation of the unit.

Above this sand sits LF3 (Figure 3.2.5). This is comprised of 1 m of dark brown clay with light brown silty laminae that are highly distorted. It is difficult to be certain if the distortion is a relic of the DANDO rig drilling process. Grain size varies between 4.7 and 7.0 ϕ as coarse grains are incorporated in to the laminae. LOI varies between 3.8% and 8.9%. There is no apparent relationship between LOI% and grain size except where the smallest grain size (7.0 ϕ) has the lowest LOI% (3.8%) at 3 m (Figure 3.2.5).

Above LF3 is 2.5 m of sand and gravels that make up the open face of the quarry. This section is separated into LF4, LF5, and LF6 (Figure 3.2.4). LF4 is a 50 cm thick deposit of coarse (0.6 -1.4 ϕ) laminated sands with <2% LOI (Figure 3.2.6) but has elevated magnetic susceptibility. LF5 (Figure 3.2.6) is comprised of horizontally-bedded medium sand with occasional bands of small gravel. The sand beds are thicker (up to 20 cm) and stratified to cross-laminated, with occasional coal fragments draped over the cross laminated sands There is evidence of cryoturbation with small frost cracks seen at the top of the unit (Figure 3.2.5). LF6 is comprised of bedded sands and gravels that vary in thickness. The <2 mm fraction fines from medium sand (1.7 ϕ) to fine sand (2.4 ϕ) with LOI at 2-3% for the whole section (Figure 3.2.5). Sections approximately 6-10 cm thick of well-rounded pebbles and granules in a sandy matrix (Figure 3.2.5). Clasts are comprised of both local and exotic lithologies. In the center of the south quarry face (Figure 3.2.5) is a long frost crack extends for nearly a meter from a thicker sandy bed in LF6 and penetrates the underlying LF4. The sequence is capped by 0.5 m of ploughed top soil.

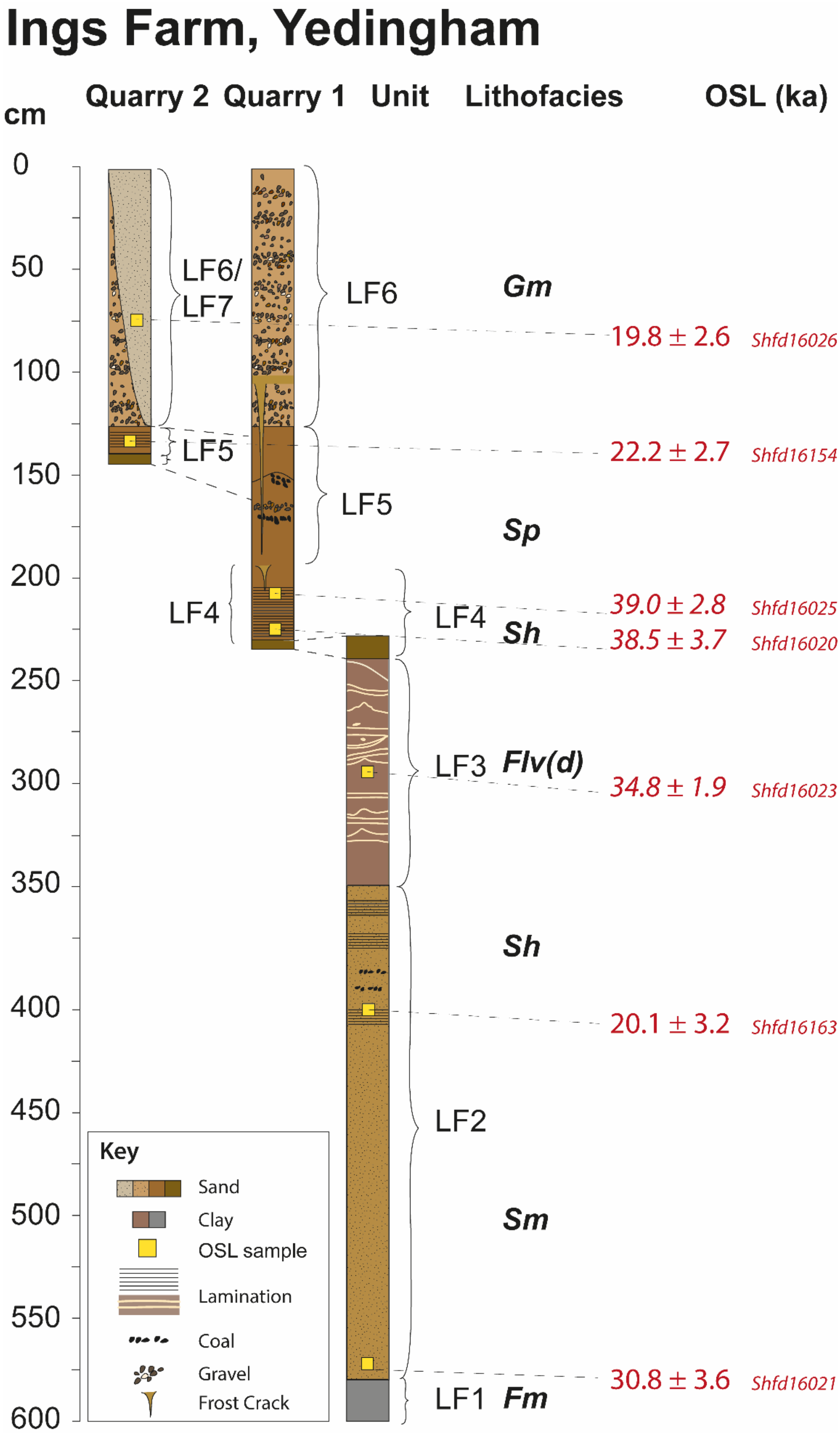


Figure 3.2.2: Stratigraphic relationship between Ings Farm Quarry 1, Quarry 2 and the extracted cores. Ages considered unreliable are given in italics.

At Yedingham Quarry 2, the sequence is the same with LF4, LF5 and LF6 overlying LF1-3. At Quarry 2, LF4-6 is intersected by a massive, structureless, light brown sandy deposit (LF7).

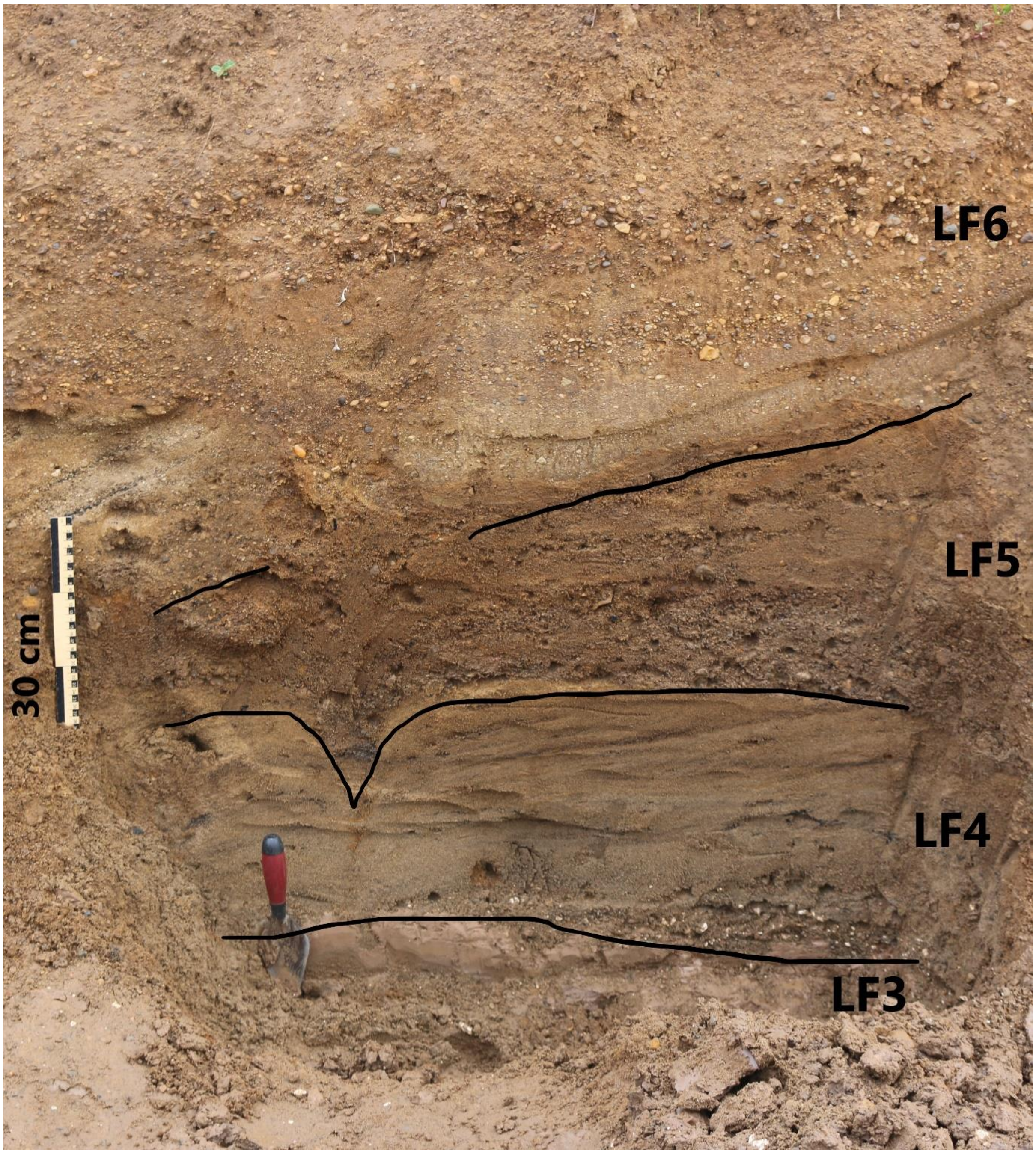


Figure 3.2.4: Quarry 1 at Ings Farm Yedingham. Sands and gravels (LF4-6) overly laminated lacustrine clay (LF3). Note the frost crack extending from LF6 to LF4. Frost cracks appear at the top of LF4, LF5 and LF6 throughout the quarry.

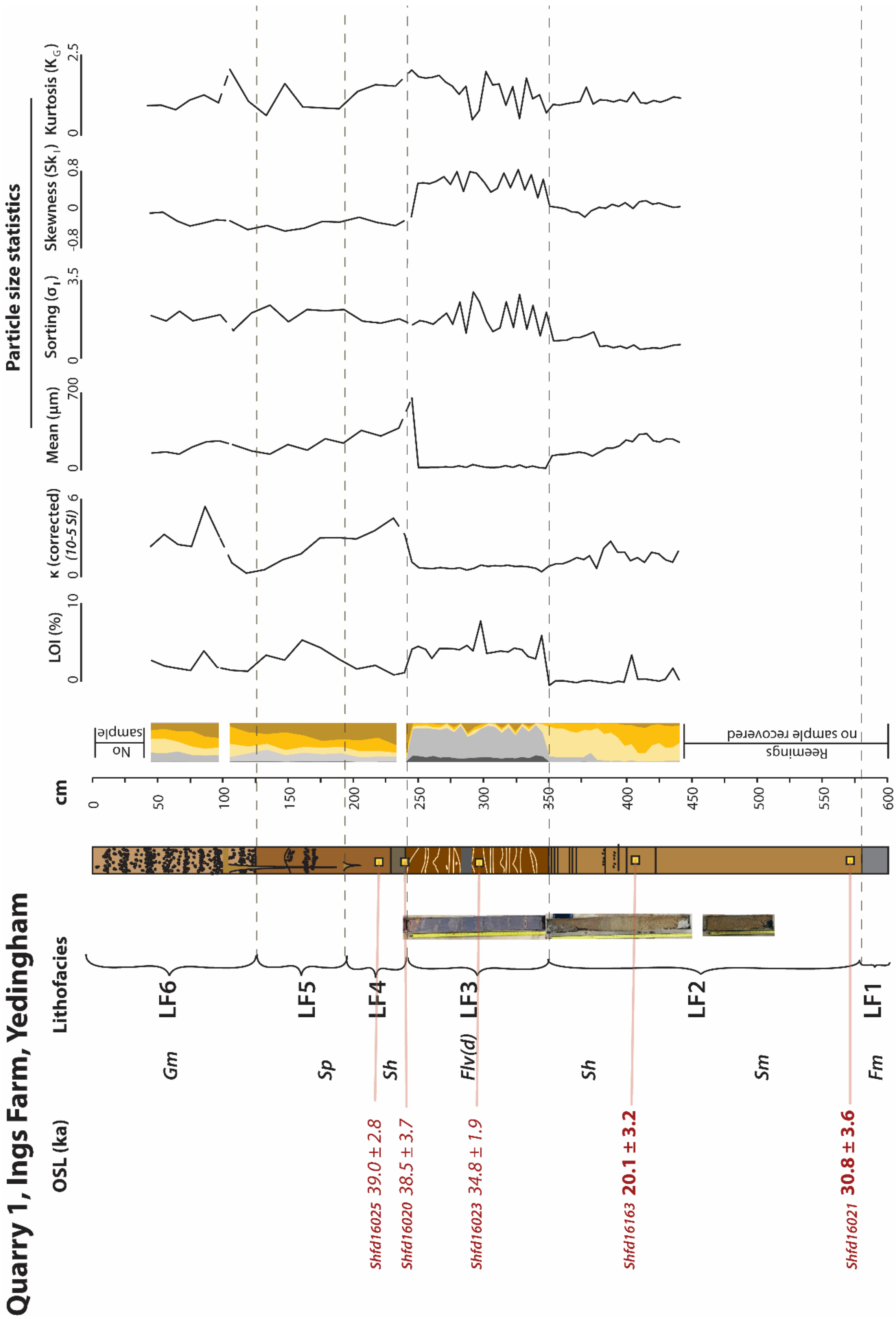


Figure 3.2.5: Core description and data for Ings Farm, Yedingham. OSL and lithofacies analysis are shown on the right. LOI, PSA, and magnetic susceptibility are shown on the left. Figure 3.2.2 shows the full stratigraphic relationship between the core and the quarries. Ages considered unreliable are shown in italics.

Interpretation

LF1 with its fine silts and clays represents deposition in a very low energy environment and is interpreted as being a deep water, low-energy lacustrine unit (Smith and Ashley, 1985). LF2 indicates a change in energy and/or sediment availability with moderately-sorted medium sand suggesting a higher-energy system. This is interpreted as indicating lake levels dropping and/or the site becoming closer to the lake margins. The energy decreases through LF2 as the median grain size drops from medium sand (1.6 ϕ) to very fine sand (3 ϕ) at the top of the section. This is further supported by the deposition above of LF3, which is dominated by fine silt (5-7 ϕ). This is interpreted as indicating a refilling of the lake. Above LF3, energy increases with a layer of coarse sand in LF4 (1.6 ϕ) with higher magnetic susceptibility indicating an influx of different sourced sediment. This is taken to indicate the lake level had dropped by this point and rivers had re-established themselves across the region. Both LF5 and LF6 show a marked change in depositional environment. Gravel-sized clasts with laterally discontinuous beds of sand and gravels, interrupted occasionally by ice wedge casts (Figure 3.2.4) are taken to indicate the development of a cold-climate braided river system with switching channels and periodic high discharge as part of a sandur plain (Miall, 1977). The increase in coarseness from LF5 to LF6 is interpreted as indicating increasing energy levels, which given the low gradient of the area, it taken to indicate increasing proximity to an ice margin.

The Pottery, High Marishes

The Pottery, High Marishes is located in the central Vale of Pickering at 21.5 m O.D. The site was chosen because of its central location and borehole logs from the area showed the potential for a long core of lacustrine deposits. The site is also 1.5 km from the current course of the River Derwent, which seemed sufficient to avoid recent floodplain deposits.

Lithofacies Analysis

In September 2015, a core of 10 m was extracted that showed 5.4 m of lacustrine clay (LF1 (Fl) - 2 (Fm)) underlying 3.5 m of coarse sand with occasional gravel (LF 3-4), comprised of mostly chalk and flint clasts (Figure 3.2.7). LF1 is a grey, laminated clay (6.0 ϕ) (Figure 3.2.) with light brown silt laminations (5.6 ϕ) varying from 3 cm to 0.1 cm in thickness. LOI shows 3-4% OM. The laminations show evidence of deformation and faulting, which is likely a relic of the drilling process. Between 9.75 m and 9.5 m lies a thick, dark, structureless silt (6.3 ϕ). However, magnetic susceptibility and LOI do not differ from LF1 (Figure 3.2.9).

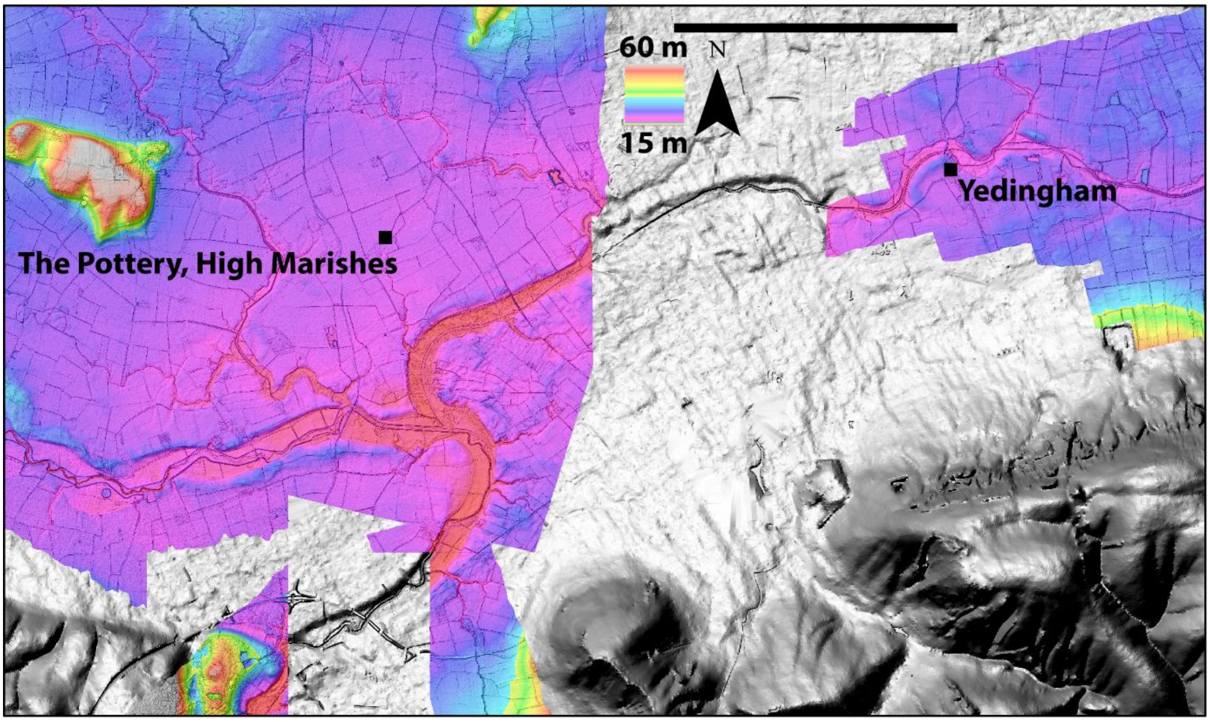


Figure 3.2.6: Location of the Pottery, High Marishes in the central basin of the Vale of Pickering. The Pottery is located 9 km west of Yedingham and 1 km north of the River Derwent.

From 9.5 m to 8.4 m, LF1 returns as a grey laminated clay with distorted laminae ranging from 2 cm to 0.1 cm. From 8.4 m to 7.35 m, LF2 reappears as a dark brown, massive clay (6.0 ϕ) with occasional small outsized chalk clasts. LOI shows an increase to 5-6% along with a slight increase in magnetic susceptibility. From 7.35 m to 5.4 m the deposit returns to LF1 where the clay is brown-grey with silty laminae ranging from 5 cm to 0.1 cm. There are numerous distortions and rip-up clasts. From 4.75 to 5.40 m the clay is firmer, but laminations are still highly distorted (Figure 3.2.8). There are no definite varves in any of the laminated sections, no organics, and very few dropstones.

LF3 (4.75 m to 1.40 m) is a massive, poorly sorted, medium (1.3 ϕ) to coarse (0.7 ϕ) sand deposit with well-rounded clasts of chalk, sandstone, flint, and other locally derived stones. In the top half of LF3 (3 m - 2 m), a small proportion are exotic and identified as quartz, granite, and jet, which is shown by a slight increase in magnetic susceptibility (Figure). Also at the top of LF3, some structure is preserved with faint laminae present. LOI ranges between 3 and 5%, although a large spike at 76.2% is present at very top of LF3 likely from charcoal contamination (Figure 3.2.9).

LFA4 is comprised of bedded sands and gravels with a light brown, fine (3.2 ϕ) to very fine poorly sorted (4 ϕ) sand matrix. The gravels range from -1 ϕ to -4 ϕ and are comprised of both local sedimentary rocks and exotic clasts such as jet, quartz, and granite, although magnetic susceptibility values remain low. A spike in the LOI of 12% is present at 1.25 m, which is likely related to the high number of charcoal fragments in the deposit, but otherwise values remain between 1 and 3%. In both LF3 and LA4 there are clasts of charcoal, sometimes laminate. The deposit is capped by 10 cm of agricultural topsoil, shown in the LOI as 13.5% OM.

Interpretation

The deposit at The Pottery, High Marishes show abyssal lacustrine clays with often distorted laminae (LF1). The majority of the distortions are likely to have been caused by compression of the DANDO rig, although some appear more in situ (Figure 3.2.7) and SEM images also show some faulting and fracturing (Delaney, Pers. Com.). The laminations are rhythmites and no evidence of seasonal varves are present. Above the lacustrine clay, there is a fluvial deposit of grey, medium sand with chalk clasts (LF3) that coarsens towards the middle of the deposit. This suggests it deposited quickly and with increasing energy as the lake level lowered. Above this fine grained sand suggests decrease in energy as meltwater energy decreases with availability with the sites distal location to the ice margin and drainage may have slowed as the lake turned to marshier conditions. This is capped by alluvium and an agriculture topsoil. This sequence represents a transition from a deep proglacial lake to a sandur plain as the ice margin retreated. The top 1 m is comprised of Holocene-aged alluvium and flood deposits.

Figure 3.2.7: The cores taken from The Pottery, High Marishes. Note the change from lacustrine clay to coarse sand suggesting change from a low energy environment to a higher energy one

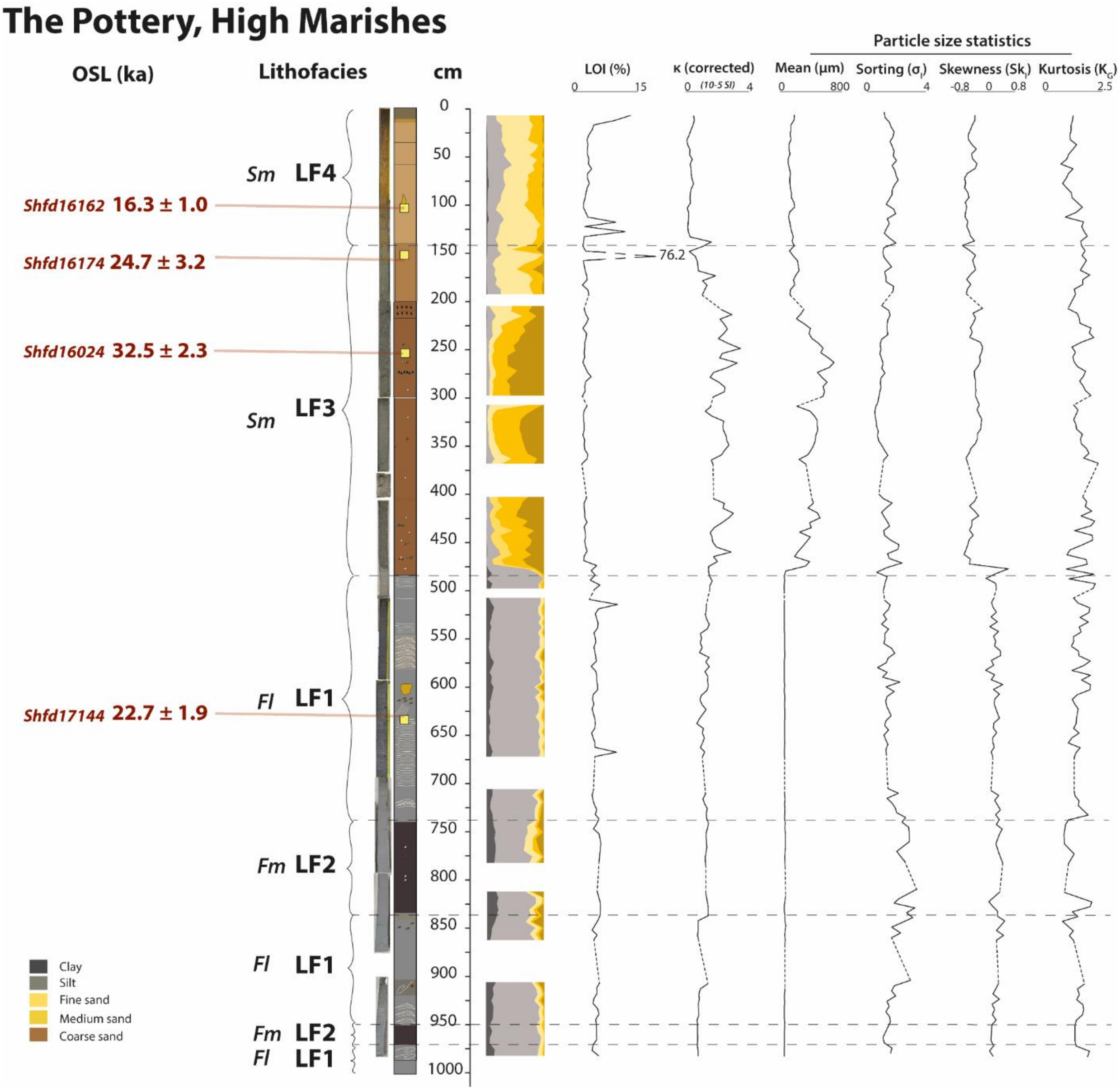


Figure 3.2.8: Core description and data for The Pottery, High Marishes. OSL and lithofacies analysis are shown on the right. LOI, PSA, and magnetic susceptibility are shown on the left

Wellfield Farm, Salton

Wellfield Farm is located in the western floodplain of the Vale of Pickering at approximately 26 m O.D. This site was chosen for its potential of recording lowering lake levels. Three meters of core were extracted from a cornfield ~1 km east of Salton (Figure 3.2.10).

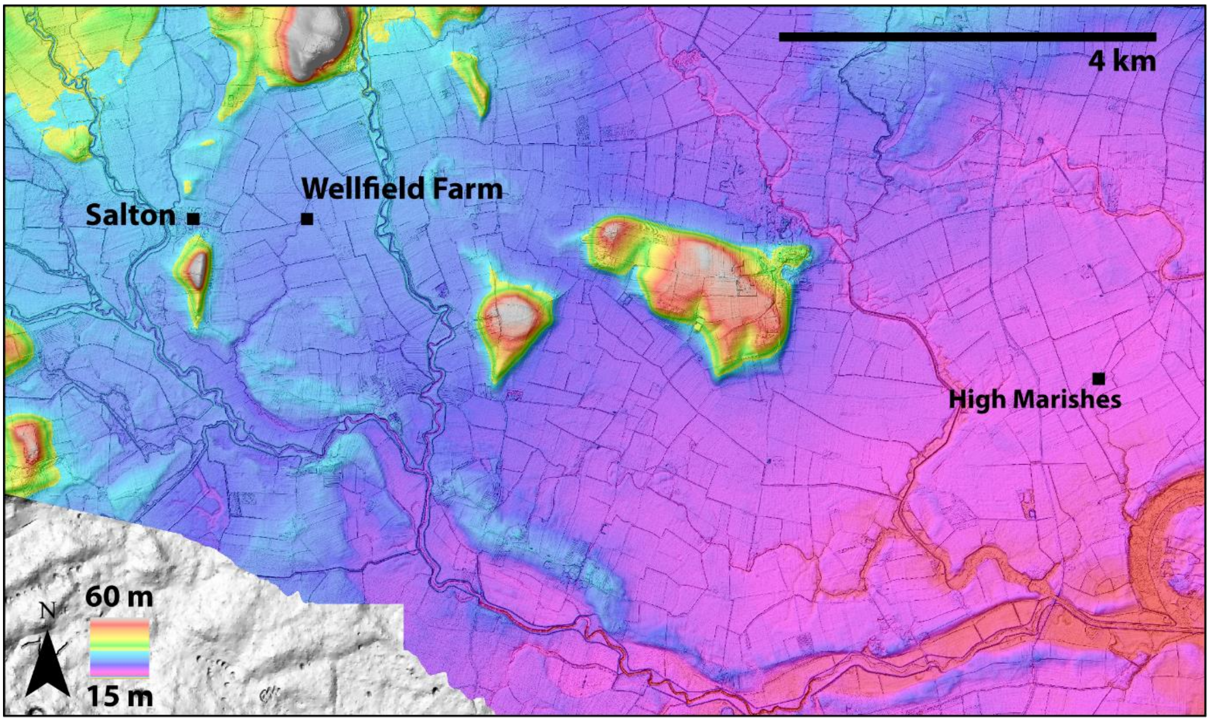


Figure 3.2.10: Location of Wellfield Farm, 1 km east of Salton and 9 km west of High Marishes. The location is a floodplain at 26 m O.D

Lithofacies Analysis

LF1 from 2.9 m to 2.5 m is made up of laminated, poorly to very poorly sorted fine (3.3 ϕ) dark grey sand. At 2.75 m there is possible evidence of a larger scale cross bedding beyond the diameter of the core. There is a spike in LOI to 6% before dropping to almost 0% as the sand changes to a light grey laminated fine sand (3.0 ϕ) at 2.65 m. At 2.40 m there is a spike in magnetic susceptibility, although from 2.40 m to 1.60 m, the deposits are badly distorted with evidence of fluid escape structures most likely a consequence of the percussion corer. At 1.60 m, the laminated grey sand shows an increase in silt content, but the mean grain size remains fine to very fine sand (3.7 ϕ) and shows faint planar cross bedding with some distortion by the drilling. At 1.4 m is a clay drop stone approximately 4 cm in diameter. Above this, most of the structure is obscured by brown-orange mottling.

LF2 (1.05 m to 0.5 m) is comprised of small, soft, well-rounded granules in a sandy silt matrix

(4 – 5 ϕ). The small, well rounded gravels are soft, locally derived chalks and limestones. Again, there appears to be evidence of larger cross bedding stratigraphy beyond the dimension of the core. LF3 (0.55 to 0 m) is a silty sandy clay alluvium that has been plowed in the upper 0.30 m.

Interpretation

The base of LF1 comprises a dark coloured deposit that shows a spike in organic content similar to the current agricultural topsoil. This may be a previously subaerially exposed surface prior to lacustrine conditions. Above this is a section of laminated sands underlying some large flame structures and distorted sands that are likely a relic of the coring process. Above the disturbed sand, the laminations continue, suggesting a low energy fluvial or lake proximal environment (Smith and Ashley, 1985). If the latter is correct, it implies that lake water at Salton was not deep enough or still enough to allow lacustrine clays and silts to fully settle, although silt content does increase above the dark subaerial layer. Above the laminated clay, a gravel deposit in a clay matrix shows evidence of cross bedding showing some evidence of redistribution by water, possibly caused by slumping as the climate warmed. A layer of post-glacial alluvium sits above the gravel layer, with the top 0.2 m plowed for agricultural practices.

Figure 3.2.11: Cores taken from Wellfield Farm, Salton. Disturbed sediment as a result of the coring process is present at the base of core 2 and the top of core 3.

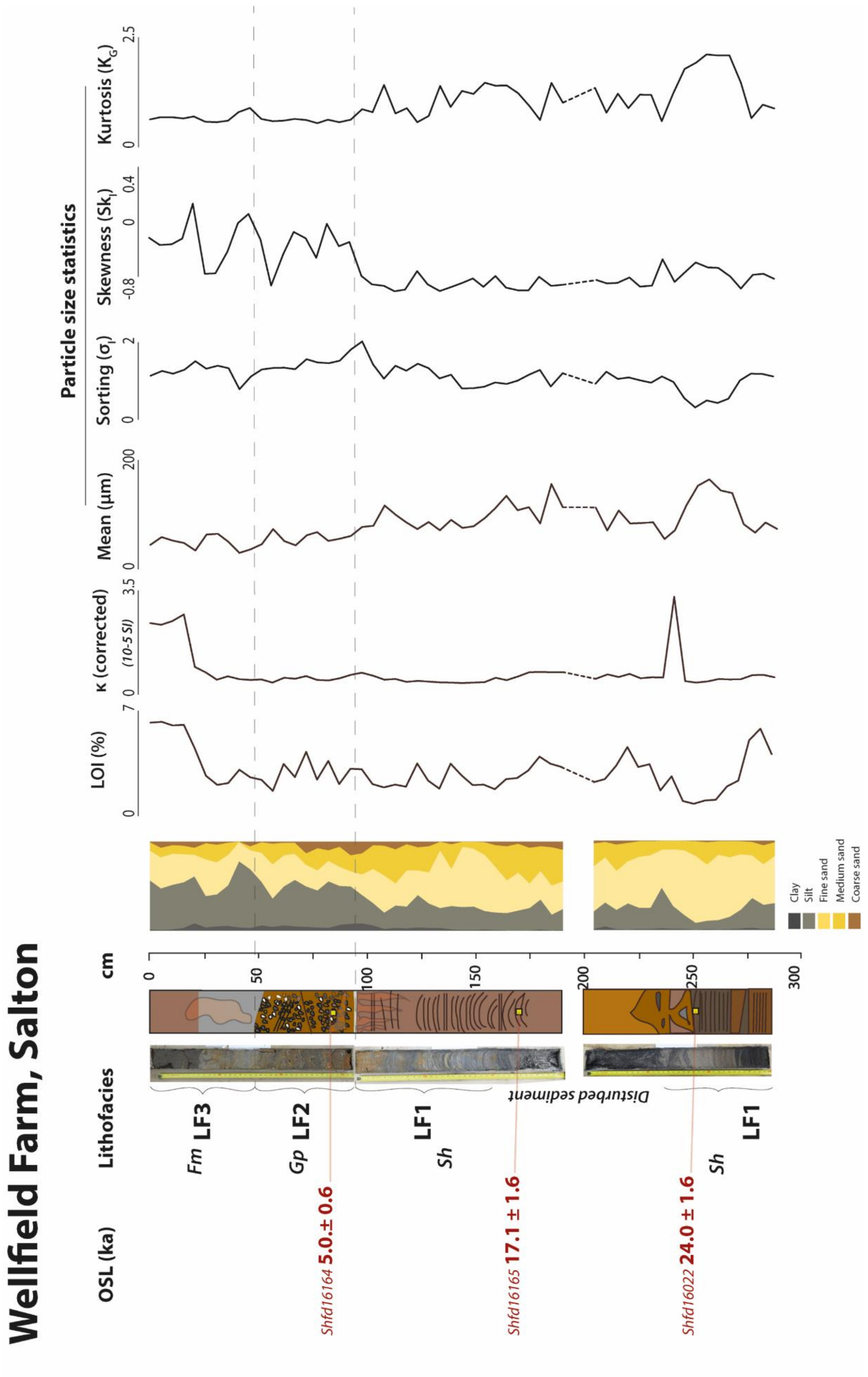


Figure 3.2.12: Core description and data for Wellfield Farm, Salton. OSL and lithofacies analysis are shown on the right. LOI, PSA, and magnetic susceptibility are shown on the left

Comparison with historical particle size data

Four samples of sand, two taken from Yedingham (LF2 and LF6) and two from High Marishes (LF3), were compared with historical PSA data from the Malton to Scarborough Transmission Line Report (CEGB,1968). These were dry sieved in using the same fractions given in the CEGB data. This was undertaken to help verify if sedimentary units identified from disparate historic logs were the same as sedimentary units examined as part of the current research. Table 1 shows the samples:

Table 3.2.1: Comparison between PSA for cores taken in this study and historical data provided by the CEGB.

Sample	Depth (m)	LF	CEGB borehole number	CEGB Stratcode*	Depth (m)	R^2 value
Yedingham Core 2	3.5 - 4.0	LF2	26	2C	4.1	0.3
Yedingham Core 2	3.5 - 4.0	LF2	36	2A/2C	2.1	0.4
High Marishes Core 3	3.0 – 4.0	LF3	14	2A	4.6	0.6
High Marishes Core 4	4.0- 5.0	LF3	15	2B	1.6	0.8

* For full historical borehole data see Methods 2.1

The sample from High Marishes matches very well with the historic data from boreholes 14 and 15 (Table 1). The grey sand with chalk gravel is unique to borehole 15 and is likely to be sourced from the south of the Vale of Pickering due to the chalk exposures in the Wolds and that the grain size of the High Marishes deposit is a little finer, which suggests further travel. The sand from Yedingham is not as similar in PSA, but the description as a brown medium sand with occasional coal laminae is very typical of the Yedingham sands in LF2.

3.4. Optically Stimulated Luminescence Dating

Luminescence ages were obtained from 19 samples from a range of environments across the Vale of Pickering (Figure 3.4.1). Proglacial sedimentary environments provide a wealth of challenges for dating sediments (See section 2.2) and many of these issues were realised in the dating of sediments from the Vale of Pickering, most notably partial bleaching. The moisture values were also difficult to calculate as some sites have experienced fluctuations in the water table since deposition. Each sample was assigned a moisture value by estimating the likely moisture content. These are saturated: for samples below the water table and likely to have always been submerged; an intermediate zone between the top of the current water table and the current land surface where samples may have been submerged for some time (although, this is difficult to estimate accurately); and unsaturated where samples were unlikely to be deposited and fully submerged in water for any extended period of time. Most samples were either saturated or intermediate. Other issues include partial bleaching where grains were not fully bleached or were redeposited without exposure to sunlight. Samples that are considered unreliable are given in italics. Details of dating each of the samples is provided in this section.

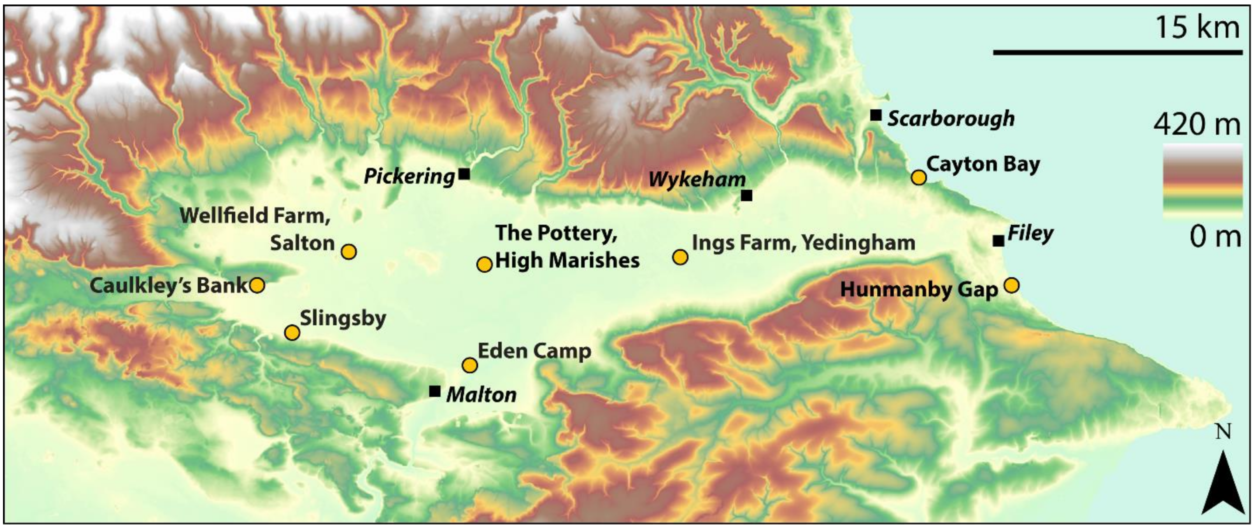


Figure 3.4.1: Sites sampled in the Vale of Pickering for OSL dating. Wellfield Farm at Salton, The Pottery at High Marishes and Ings Farm, Yedingham all were sampled from cores taken via percussion drilling.

Samples and sample sites

Ages below (Table 3.4.2) are presented in stratigraphic order for each site followed by results for the individual sites (see Table 3.4.1 for depth, coordinates, and water content %). Ages reported for CAM are minus outliers, denoted by an *. Over dispersion (OD) values for each sample is given with outliers and then minus outliers in parenthesis. For samples that were analysed using the FMM, the number of components, mean De and proportion are given with the selected data in bold italic. A final map of ages by location is presented at the end of the section.

Table 3.4.1: Location data and site conditions for OSL samples taken from the Vale of Pickering

	Lab code	Sample name	Depth (m)	Coordinates	Water (%)	β (μGy/ka)	γ (μGy/ka)	cosmic (μGy/ka)
Ings Farm Quarry, Yedingham	Shfd16026	Quarry 2 OSL 1	0.6	54.203436 N, -0.613165 E	15	716 ± 55	461 ± 29	150 ± 8
	Shfd16154	Quarry 2 OSL 4	0.9	54.203436 N, -0.613165 E	20	1067 ± 83	670 ± 43	187 ± 9
	Shfd16025	Quarry 1 OSL 1	2.3	54.202162 N, -0.612886 E	20	719 ± 56	457 ± 29	154 ± 8
	Shfd16020	Quarry 1 OSL 2	2.5	54.202162 N, -0.612886 E	20	735 ± 58	469 ± 29	150 ± 8
	Shfd16023	Core 1	3	54.202162 N, -0.612886 E	26	373 ± 31	256 ± 15	132 ± 7
	Shfd16163	Core 2	4.7	54.202162 N, -0.612886 E	20	586 ± 45	433 ± 27	176 ± 9
	Shfd16021	Core 4	6.5	54.202162 N, -0.612886 E	20	527 ± 41	342 ± 22	92 ± 5
The Pottery, High Marishes	Shfd16162	Core 1	0.9	54.197288 N, -0.750721 E	10	897 ± 69	622 ± 39	187 ± 9
	Shfd16174	Core 2	1.5	54.197288 N, -0.750721 E	20	619 ± 48	411 ± 26	172 ± 9
	Shfd16024	Core 3	2.7	54.197288 N, -0.750721 E	20	603 ± 47	378 ± 24	145 ± 7
	Shfd17144	Core 8	6.4	54.197288 N, -0.750721 E	26	1715 ± 134	1180 ± 74	94 ± 5
Wellfield Farm, Salton	Shfd16164	Core 1	0.5	54.213924 N, -0.890281 E	15	937 ± 72	823 ±52	198 ± 10
	Shfd16165	Core 2	1.9	54.213924 N, -0.890281 E	15	1000 ± 79	736 ± 46	163 ± 8
	Shfd16022	Core 3	2.5	54.213924 N, -0.890281 E	15	816 ± 68	377 ± 25	148 ± 7
Other sites	Shfd16019	Slingsby	1.6	54.170392 N, -0.934687 E	15	729 ± 56	504 ± 32	169 ± 8
	Shfd16018	Caulkley's Bank	2.3	54.192058 N, -0.951128 E	10	624 ±48	256 ± 15	154 ± 8
	Shfd16161	Eden Camp	1.3	54.153409 N, -0.778805 E	15	586 ± 45	433 ± 27	176 ± 9
	Shfd16157	Hunmanby Gap OSL 1	15	54.177556 N, -0.266662 E	10	1277 ± 99	600 ± 38	41 ± 2
	Shfd16158	Hunmanby Gap OSL 4	17	54.177556 N, -0.266662 E	10	761 ± 59	577 ± 36	36 ± 2
	Shfd16159	Cayton Bay	10	54.243046 N, -0.357282 E	10	401 ± 30	350 ± 22	184 ± 9

Table 3.4.2: OSL ages for sites within the Vale of Pickering. OD presented for all aliquots and in parenthesis once outliers were removed. FMM % refers to the proportion of data contained in that component

	Lab code	**Sample name**	**OD %**	**D_e (Gy)**	**Dose rate (μGy/a)**	**Model**	**FMM Components*† K #**	**K**	**Mean De (Gy)**	**Error**	**%**	**Age (ka)**
Ings Farm Quarry, Yedingham	Shfd16026	Quarry 2 OSL 1	53 (34)	24.8 ± 3.0	1256 ± 56	**MAM**	Not applicable					**19.8 ± 2.6**
	Shfd16154	Quarry 2 OSL 4	42 (39)	36.8 ± 4.1	1660 ± 79	**MAM**	n/a					**22.2 ± 2.7**
	Shfd16025	Quarry 1 OSL 1	25 (19)	48.9 ± 2.7	1254 ± 59	**FMM**	3	***1***	***48.9***	***2.7***	***77***	***39.0 ± 2.8***
								2	73.6	10.5	20	
	Shfd16020	Quarry 1 OSL 2	29 (24)	53.1 ± 3.9	1380 ± 66	**FMM**	2	***1***	***53.1***	***3.9***	***66***	***38.5 ± 3.7***
								2	80.7	9.5	34	
	Shfd16023	Core 1	37 (19)	67.0 ± 2.1	1923 ± 88	**CAM**	n/a					***34.8 ± 1.9***
	Shfd16163	Core 3	50 (38)	22.2 ± 3.4	1102 ± 52	**MAM**	n/a					**20.1 ± 3.2**
	Shfd16021	Core 4	27 (23)	28.6 ± 3.0	929 ± 44	**MAM**	n/a					**30.8 ± 3.6**
The Pottery, High Marishes	Shfd16162	Core 1	29 (21)	28.3 ± 1.2	1735 ± 80	**CAM**	n/a					**16.3 ± 1.0**
	Shfd16174	Core 2	28 (22)	31.3 ± 3.8	1261 ± 56	**MAM**	n/a					**24.7 ± 3.2**
	Shfd16024	Core 3	39 (21)	35.9 ± 1.8	1102 ± 51	**CAM**	n/a					**32.5 ± 2.3**
	Shfd17144	Core 8	19	69.3 ± 4.2	3047 ± 153	**COM**	n/a					***22.7 ± 1.9***
Wellfield Farm, Salton	Shfd16164	Core 1	55 (48)	10.6 ± 1.2	2120 ± 96	**FMM**	3	***1***	***10.6***	***1.2***	***21***	**5.0 ± 0.6**
								2	24.5	1.9	55	
								3	44.1	4.0	25	
	Shfd16165	Core 2	43 (34)	40.0 ± 2.2	2346 ± 113	**FMM**	3	***1***	***40.0***	***2.2***	***58***	**17.1 ± 1.6**
								2	73.9	52	40	
	Shfd16022	Core 3	29 (16)	31.9 ± 1.2	1330 ± 71	**FMM**	2	***1***	***31.86***	***1.40***	***81***	**24.0 ± 1.6**
								2	54.72	5.45	19	
Other sites	Shfd16019	Slingsby	29 (22)	36.0 ± 2.3	1451 ± 66	**FMM**	2	***1***	***36.0***	***2.3***	***65***	**24.8 ± 2.0**
								2	58.3	6.2	35	
	Shfd16018	Caulkley's Bank	34 (10)	114.3 ± 4.8	1258 ± 57	**CAM**	n/a					***90.9 ± 5.4***
	Shfd16161	Eden Camp	33 (24)	26.4 ± 1.0	1233 ± 54	**FMM**	4	***1***	***26.4***	***1.0***	***87***	**21.7 ± 1.3**
								2	51.5	9.6	13	
	Shfd16157	Hunmanby Gap OSL 1	42 (41)	138.2 ± 9.9	1962 ± 106	**FMM**	2	***1***	***138.2***	***9.9***	***37***	***70.5 ± 6.3***
								2	286.0	14.4	63	
	Shfd16158	Hunmanby Gap OSL 4	63 (56)	49.8 ± 3.1	1402 ± 70	**FMM**	3	***1***	***49.8***	***3.1***	***44***	***35.5 ± 2.8***
								2	101.1	11.4	37	
								3	184.8	0.1	17	
	Shfd16159	Cayton Bay	54 (48)	62.8 ± 4.1	1260 ± 65	**FMM**	3	***1***	***62.8***	***4.1***	***62***	***49.0 ± 4.1***
								2	145.5	11.5	30	

* FMM components are reported with the number of components (K#) chosen based on the lowest BIC and the D_e, error, and proportion (%)† for each sample by component (K). † Only data from proportions >10% are reported. Proportion is rounded to the nearest 1

Ings Farm, Yedingham

Yedingham is in the central part of the Vale of Pickering (Figure 3.4.1). Samples for OSL were taken from three locations: two quarry faces and from cores that were drilled below the quarry floor. Many of the deposits from the quarries (Shfd16026, Shfd16154, Shfd16025 and Shfd16020) are glaciofluvial and there is a strong likelihood that some grains were partially bleached. There is also evidence of post-depositional disturbance (see 3.3 for more detail).

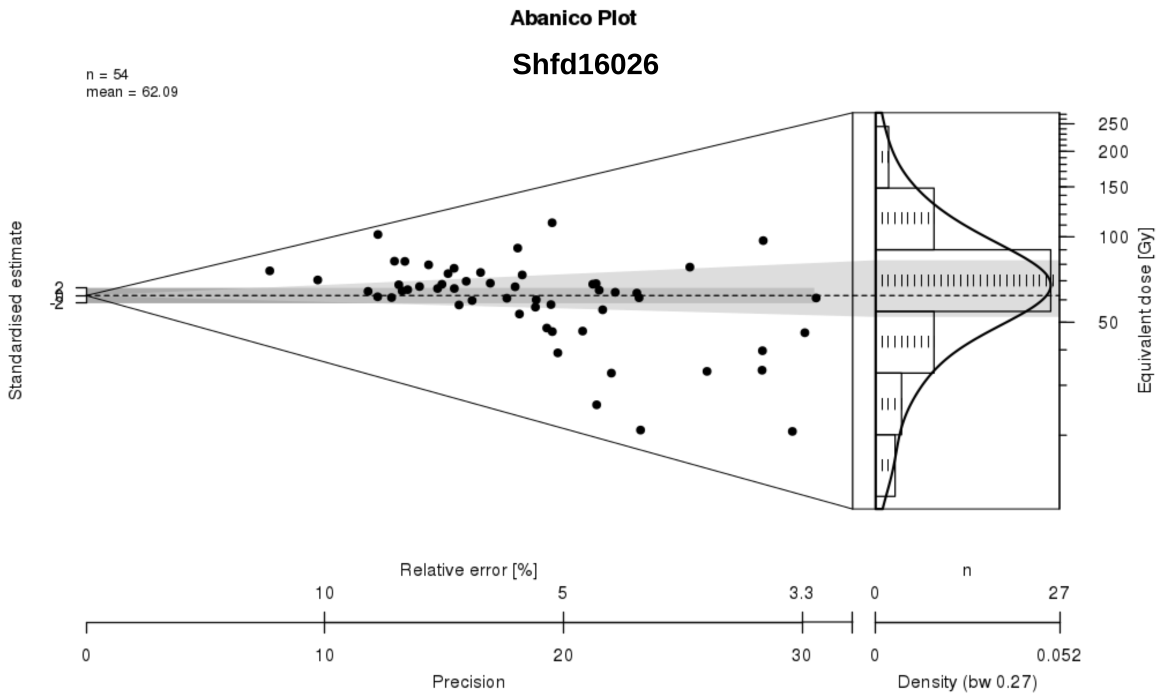


Figure 3.4.2: Abanico plot for Shfd16026 showing the data distribution of D_e values. The scatter of the data in this sample suggests partial bleaching. Using the MAM provides a weighted average of low D_e values (24.8 ± 3.0 Gys) resulting in an age of 19.8 ± 2.6 ka.

Sample Shfd16026 was taken at a depth of 0.6 m in Quarry 2from a massive, fluvially deposited sand, thought to be a former point bar of the River Derwent (section 3.3). The MAM was chosen for this sample (Figure 3.4.2) as the distribution is not normal and the OD is high (53%), even without outliers (34%), so it is likely that some grains were not, or only partially, reset.

The MAM (Table 3.4.2) gives a weighted average of 24.8 ± 3.0 Gy providing an age of 19.8 ± 2.6 ka.

Sample Shfd16154 (Figure 3.4.3) was taken from the top of a suite of fluvial sands in Quarry 2 (section 3.3). The OD is high at 42% and remains high at 39% with outliers removed. There is one potentially low D_e outlier, which did not affect the MAM, but did skew the FMM and so the MAM was preferred. The MAM (table 3.4.2) gives a weighted average of 36.8 ± 4.1 Gy resulting in an age of 22.2 ± 2.7 ka.

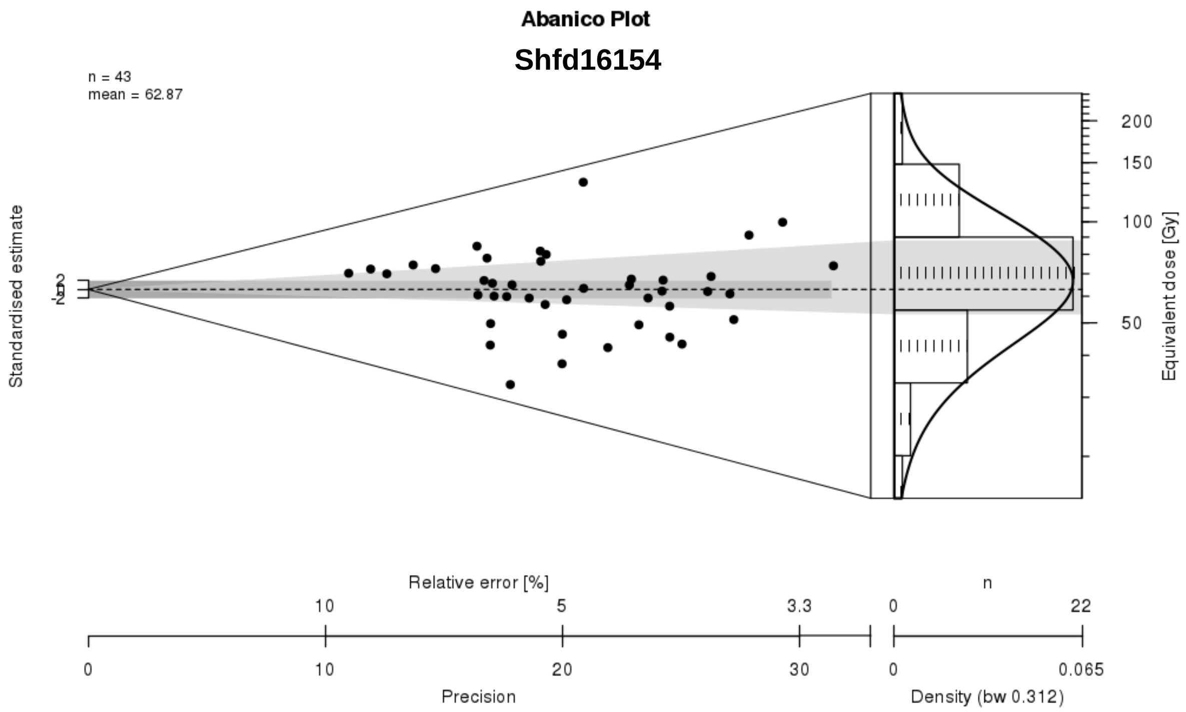


Figure 3.4.3: Shfd16154 has a high OD of 42% (39% minus outliers). The MAM gives an age of 22.2 ± 2.7 ka

Shfd16025 (Figure 3.4.4) and Shfld16020 (Figure 3.4.5) are fluvial sands deposited above the lacustrine clay quarry floor (section 3.3). Shfd16020 does not have a normal distribution (Figure 3.4.4) with an OD of 25% reduced to 19% once outliers are removed. Most of the values are left of the mean skewing the data to the right. Because of this, the FMM was chosen over the CAM. Component 1 (48.9 ± 2.7 Gy) of the FMM (Table 3.4.2) returns an age of 39.0 ± 2.8 ka. However, this result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates. Shfd16020 (Figure 3.4.5) has an OD of 29%

reducing to 24% once outliers are removed. The sample shows a bimodal distribution and the MAM was affected by a low D_e value of ~30 Gy so the FMM was chosen. The first component of the FMM has an average of 53.1 ± 3.0 Gy returning an age of 38.5 ± 3.7 ka. Again, this result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

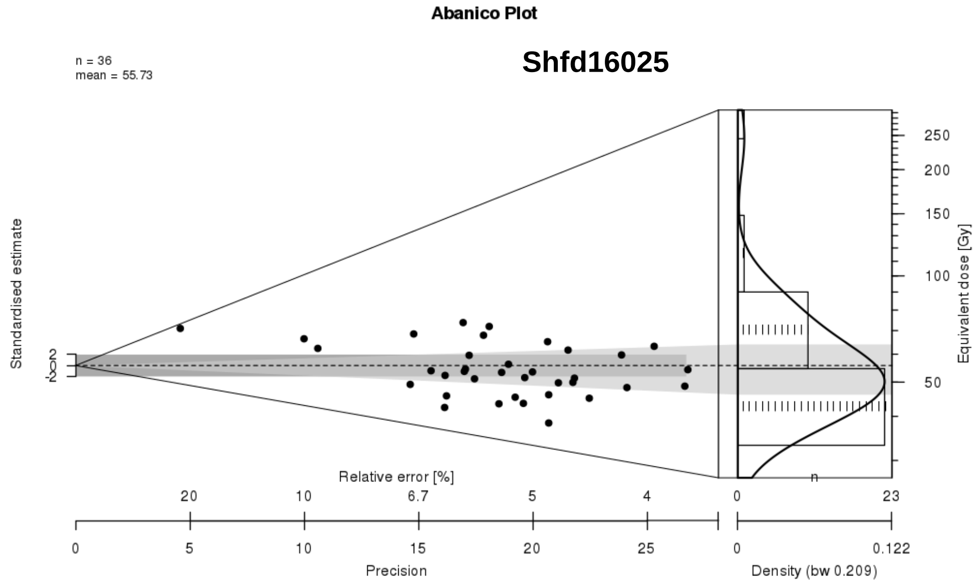


Figure 3.4.4: Even though the data are somewhat normal after removal of outliers, the distribution is skewed to the right. As a result, the first component (48.9 ± 2.7 Gys) of the FMM was chosen over the CAM. This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

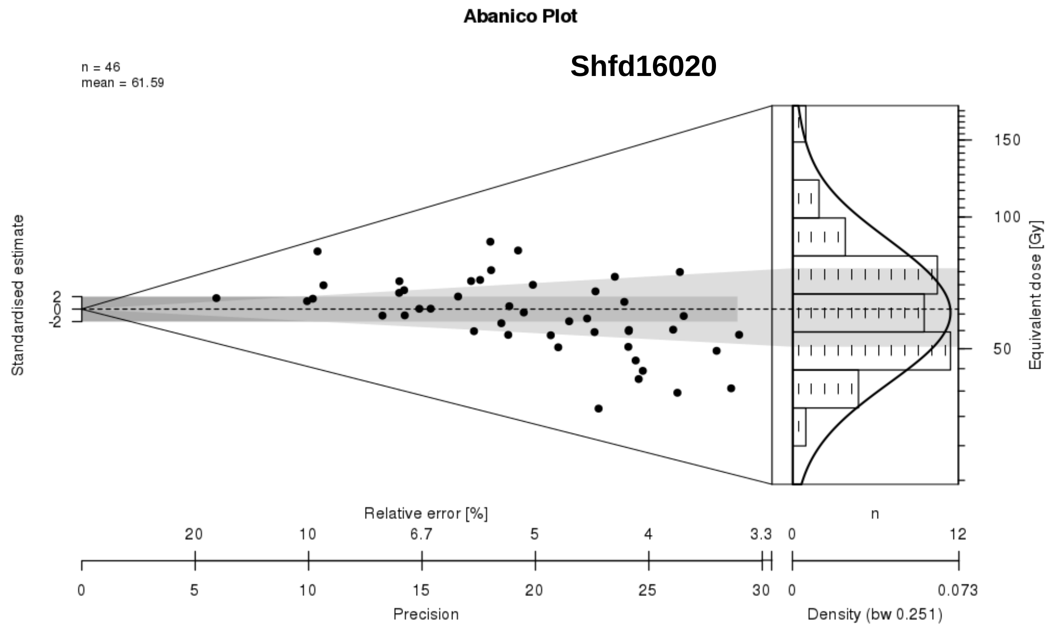


Figure 3.4.5: Shfd16020 has a bimodal distribution and a single lower D_e outlier. Component 1 of the FMM was chosen giving a mean of 53.1 ± 3.0 Gys, which equates to an age of 38.5 ± 3.7 ka.

Samples Shfd16023 (Figure 3.4.6), Shfd16163 (Figure 3.4.7) and Shfd16021 (Figure 3.4.8) were taken from cores extracted from beneath the lacustrine quarry floor (section 3.3). Mixing of sediment due to the nature of coring and the sloughing and mixing of loose sediment in the borehole as the cores were removed were important factors to consider in assigning an age model. Sand grains measured from Shfd16023 (Figure 3.4.6) were washed out of a clay matrix core with the dose rate adjusted to compensate for the decreased grain size of the deposit (section 3.3). The sample has an OD of 37% that reduces to 19% once outliers are removed (Table 3.4.2). The D_e distribution is scattered with a mean of 71.1 ± 3.9 Gys but displays a normal distribution with a mean of 67.0 ± 2.1 Gy once outliers are removed. Due to this, the CAM was used with a mean D_e of 67.0 ± 2.1 Gy providing an age of 34.8 ± 3.7 ka (Table 3.4.2). This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

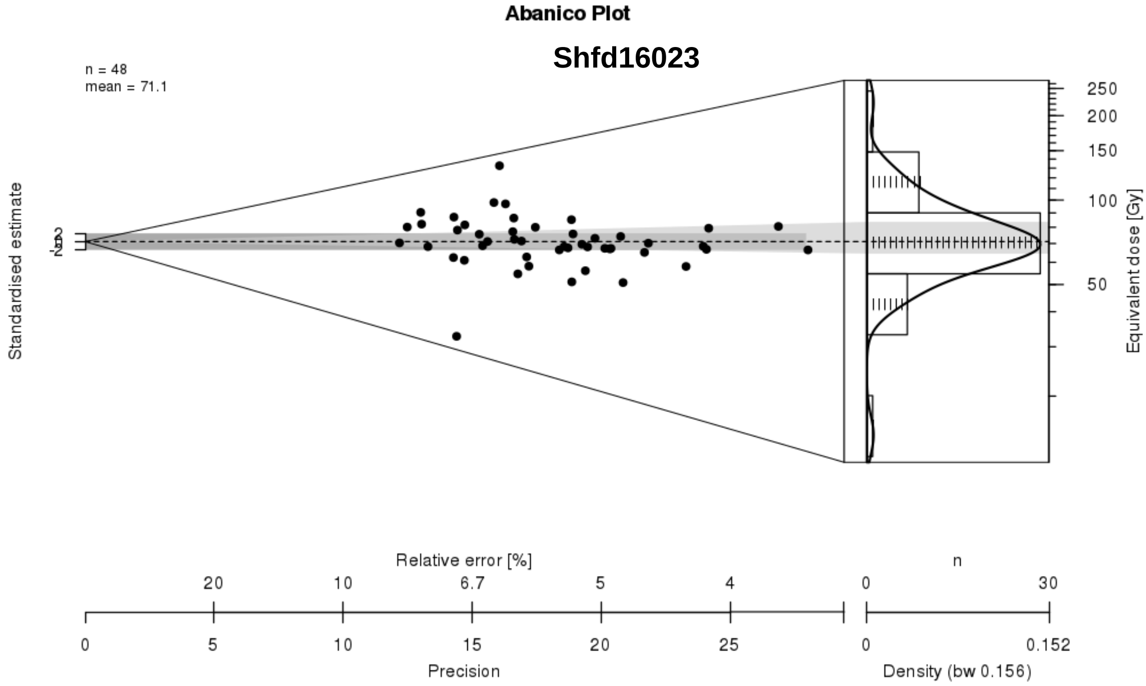


Figure 3.4.6: Grains from Shfd16023 were washed out of a lacustrine clay matrix. The CAM is the best model for this data as the mean D_e is normalised once outliers are removed from 71.1 ± 3.9 Gys to 67.0 ± 2.1 Gys. The sample returned an age of 34.8 ± 1.9 ka. This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

Shfd16163 (Figure 3.4.7) is a sand deposit below lacustrine clay (section 3.3). The D_e distribution has an OD of 50% reducing to 38% once outliers are removed (Table 3.4.2). The distribution is also skewed to the right with a large proportion of the grains to the left of the mean with no low D_e outliers, which makes the MAM a good choice for this data as the OD shows this deposit is partially bleached. The weighted mean from the MAM of 22.2 ± 3.4 Gy results in an age of 20.1 ± 3.2 ka (Table 3.4.2).

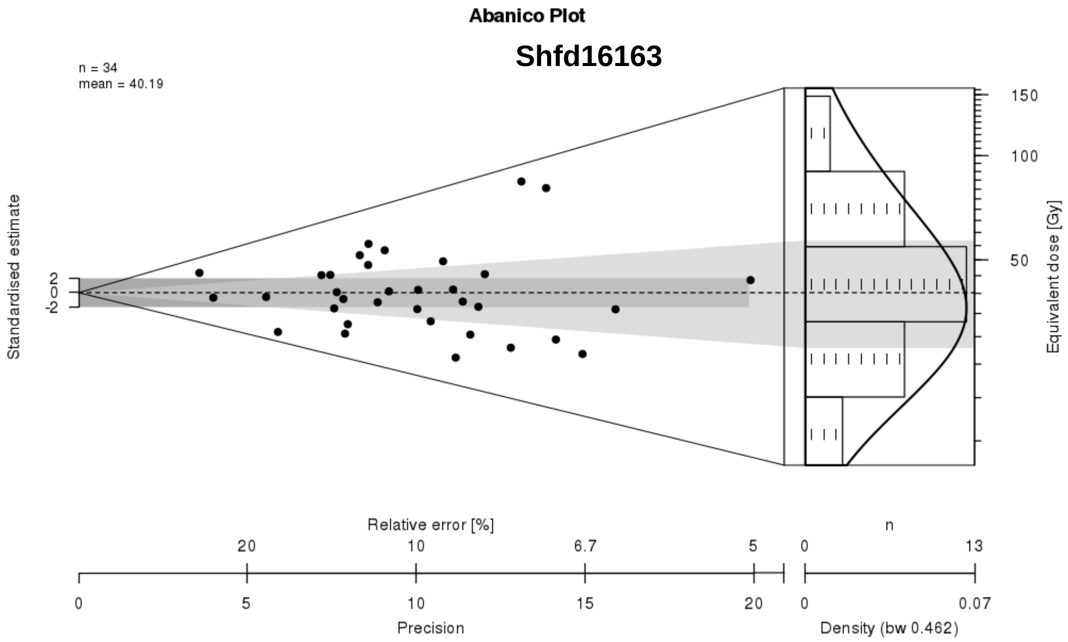


Figure 3.4.7: The D_e distribution of Shfd16163 has an OD of 50% reducing to 38% once outliers are removed. The distribution is skewed to the right and so the MAM was used giving a weight mean D_e of 22.2 ± 3.4 Gys that equates to an age of 20.1 ± 3.2 ka.

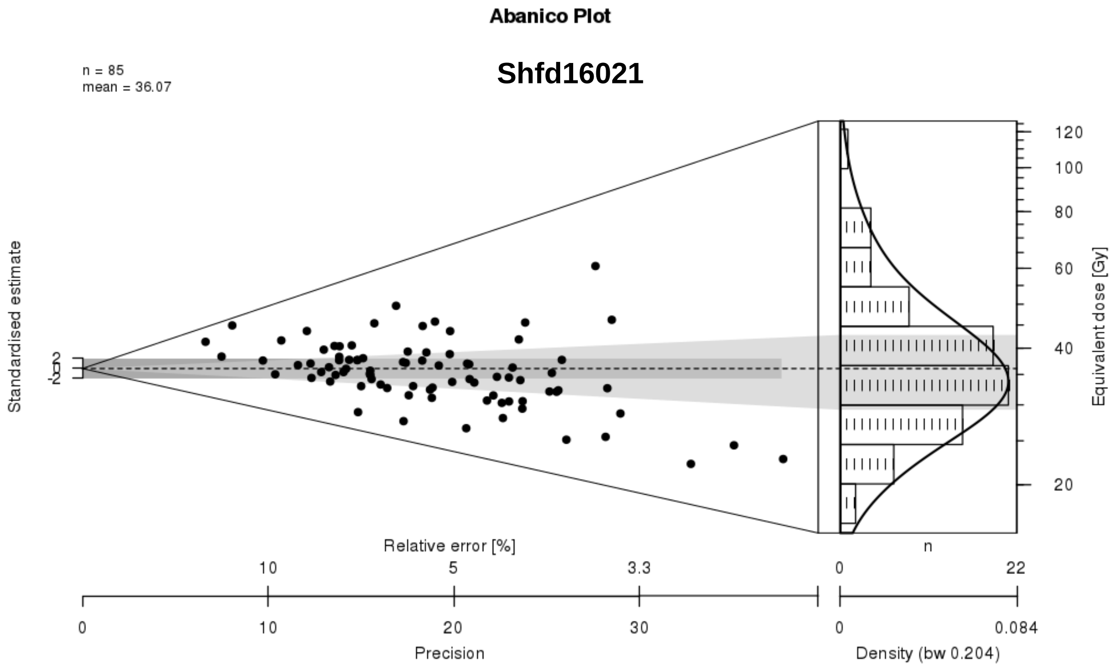


Figure 3.4.8: Shfd16021 has an OD of 27% reducing to 23% once outliers are removed. The distribution is skewed to the right but not affected by low D_e values. The MAM gives a weighted mean of 28.6 ± 3.0 Gys, resulting in an age of 30.8 ± 3.6 ka.

Shfd16021 has an OD of 27% that reduces to 23% once outliers are removed. The distribution is skewed to the right but is not affected by low D_e values, so the MAM is a good choice giving a weighted mean of 28.6 ± 3.0 Gys, resulting in an age of 30.8 ± 3.6 ka.

Interpretation

The deposits at Yedingham show a suite of lacustrine sediments underlying fluvial sands and gravels (Figure 3.2.2; section 3.2). The ages returned show a stratigraphic reversal. At the base of LF2, Shfd16021 returns an age of 30.8 ± 3.6ka. This followed by an age of 20.1 ± 3.2 ka (Shfd16163) from the top of the unit, just below LF3. Shfd16023 from LF3 gives an age of 34.8 ± 1.9 ka (Shfd16023). Above this, samples taken from LF4 give ages of 38.5 ± 3.7 ka (Shfd16020) and 39.0 ± 2.8 ka (Shfd16025). In Quarry 2, a sample from LF5 returns an age of 22.2 ± 2.7 ka (Shfd16154). Sand from LF7 gives an age of 19.8 ± 2.6 ka (Shfd16026).

The stratigraphically deepest sample, Shfd16021 at the base of LF2, suggests that an extensive and deep lake existed prior to the LGM (before 30.8 ± 3.6 ka) as it sits above a deeper layer of lacustrine clay (LF1). The date from the top of LF2, under the second unit of lacustrine clay of LF3, suggests that the environment returned to a second deep and low energy lacustrine phase of the lake after 20.1 ± 3.2 ka (Shfd16163). However, the stratigraphically reversed ages from LF3 and LF4 likely entered the lake system without exposure to light it is likely the age from the top of LF2 (Shfd16163) is giving a minimum age for the formation of the lake.

The age reversal starts with the sand grains taken from within the lacustrine clay. Sample Shfd16023 returns an age of 34.8 ± 1.9 ka (Shfd16023). This is supported by samples taken from LF4 that return ages of 39.0 ± 2.8 ka (Shfd16025) and 38.5 ± 3.7 ka (Shfd16020). This suggests that the fluvial sands above the clay were subaqueously deposited without exposure to light from older deposits reworked into the lake system as the quieter lake phase ended and sediment began to move into the lake, possibly from reactivation of meltwater channels as glacier ice retreated from the east coast. As the lake shallowed, more of the sand grains were light exposed, although sample Shfd16154 returns an age of 22.2 ± 2.7 ka, it is likely that a portion of these grains were also partially bleached or potentially mixed with sediment from LF4 thereby providing an age overestimation. Sample Shfd16026 is also likely overestimated as it is partially bleached, and the age obtained from the MAM.

The Pottery, High Marishes

Samples for OSL dating were collected via DANDO percussion coring (section 3.3) at The Pottery, High Marishes (Figure 3.4.1): a site situated in the central Vale of Pickering where the thickest deposit of lacustrine sands and clays are found. Three 1 m opaque-lined cores were recovered for OSL dating, as well as a sample taken from the interior of core 8 for single grain OSL dating (Shfld17144).

Shfd16162 (Figure 3.4.9) is taken from 0.5 m (Table 3.4.1) the first core (0-1 m) and is composed of sand (section 3.3). It has an OD of 29% reducing to 21% once outliers are removed. Additionally, after outliers are removed, the distribution is normal with CAM providing a mean of 28.3 ± 1.2 Gys and returning an age of 16.3 ± 1.0. Shfld16174 is from a coarse section of grey sand and showed a bimodal distribution, even after removing outliers.

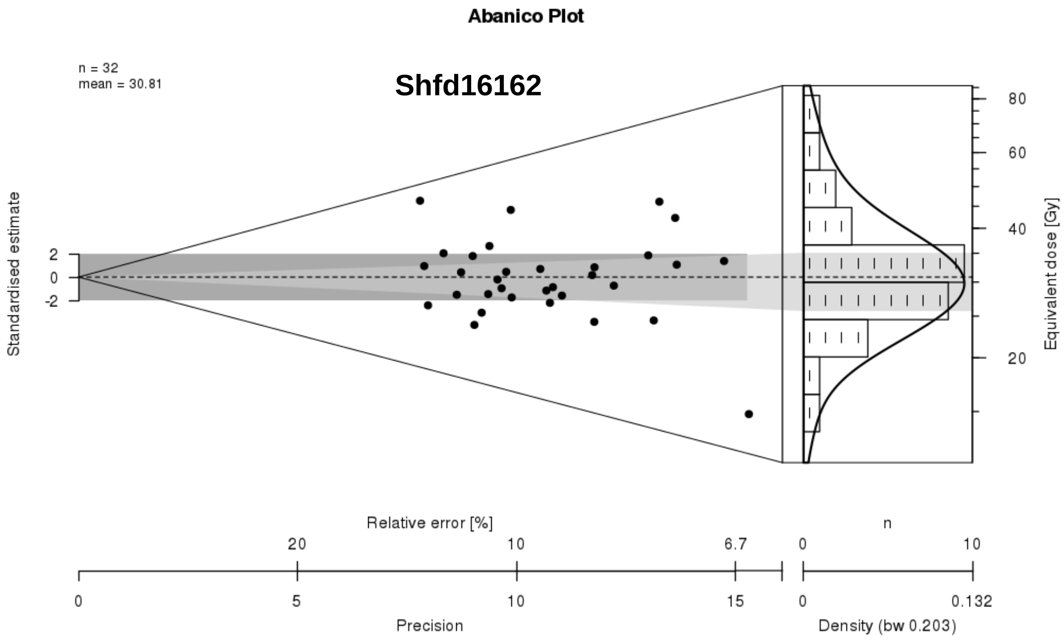


Figure 3.4.9: Shfd16162 with a normal distribution after outliers are removed. The mean of 28.3 ± 1.2 Gys equating to an age of 16.3 ± 1.0 ka.

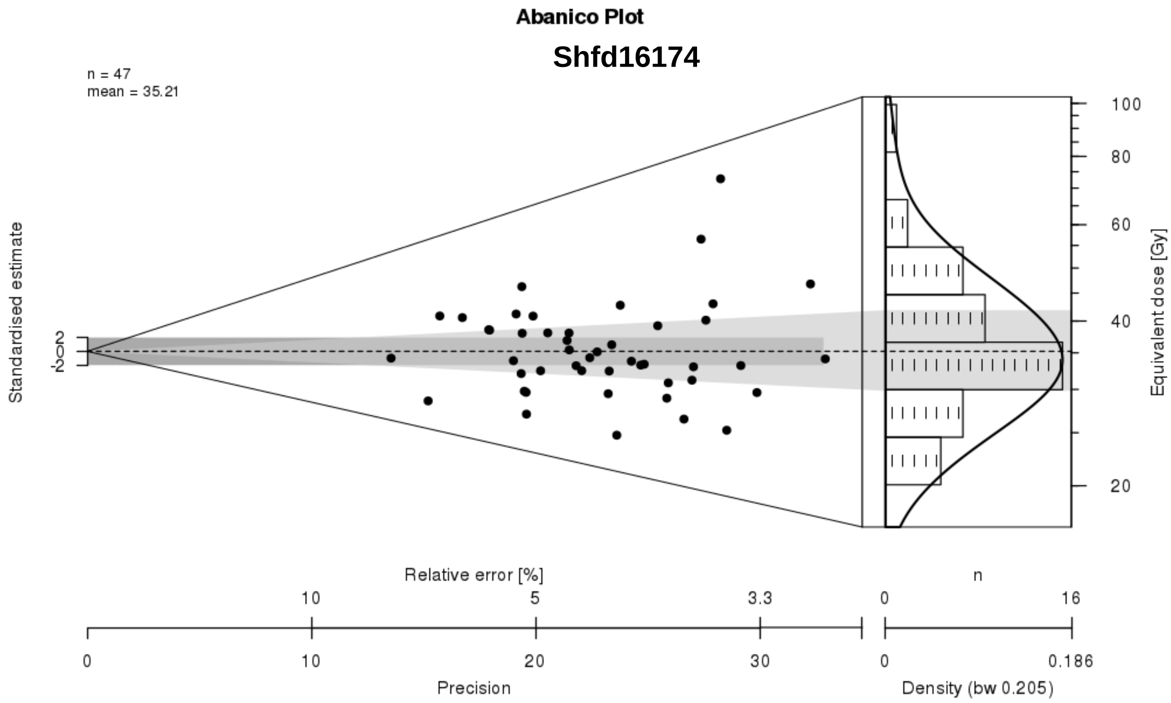


Figure 3.4.10: Shfld16174 has an OD of 28% reduced to 21% after outliers are removed. The MAM gives a weighted mean of 31.3 ± 3.8 Gys resulting in an age of 24.7 ± 3.2 ka.

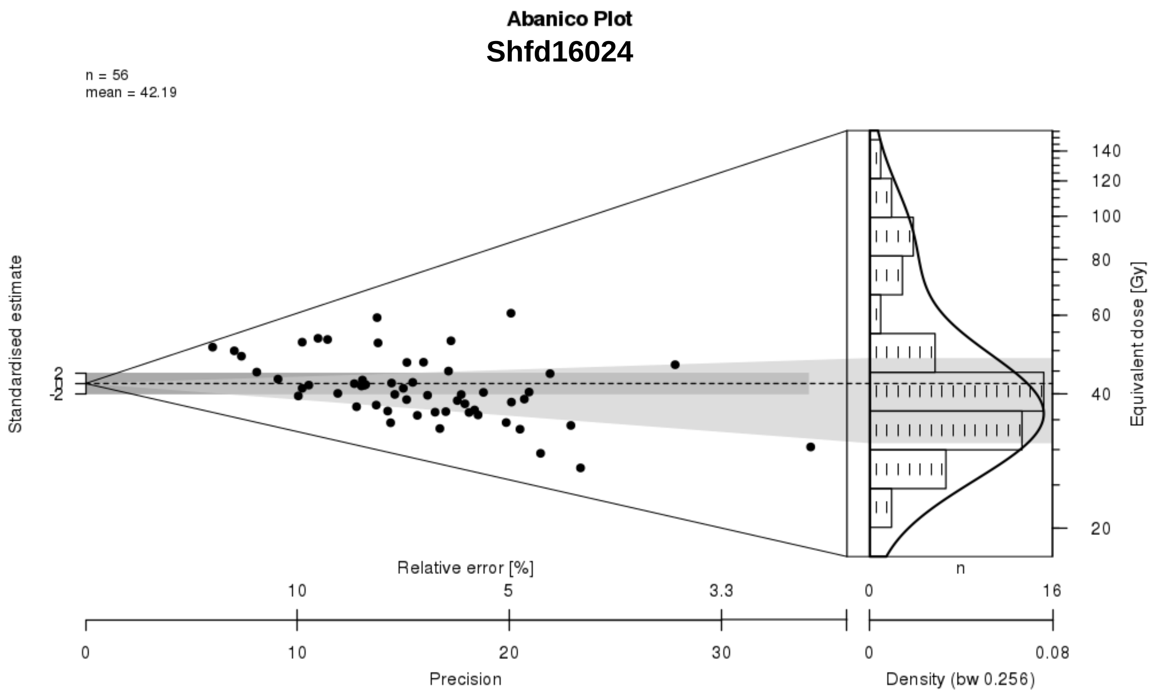


Figure 3.4.11: Shfld16124 has an OD of 39% dropping to 21% once outliers are removed The CAM was used, giving a mean of 35.9 ± 1.8 Gys resulting in an age of 32.5 ± 2.3 ka.

Figure 3.4.11

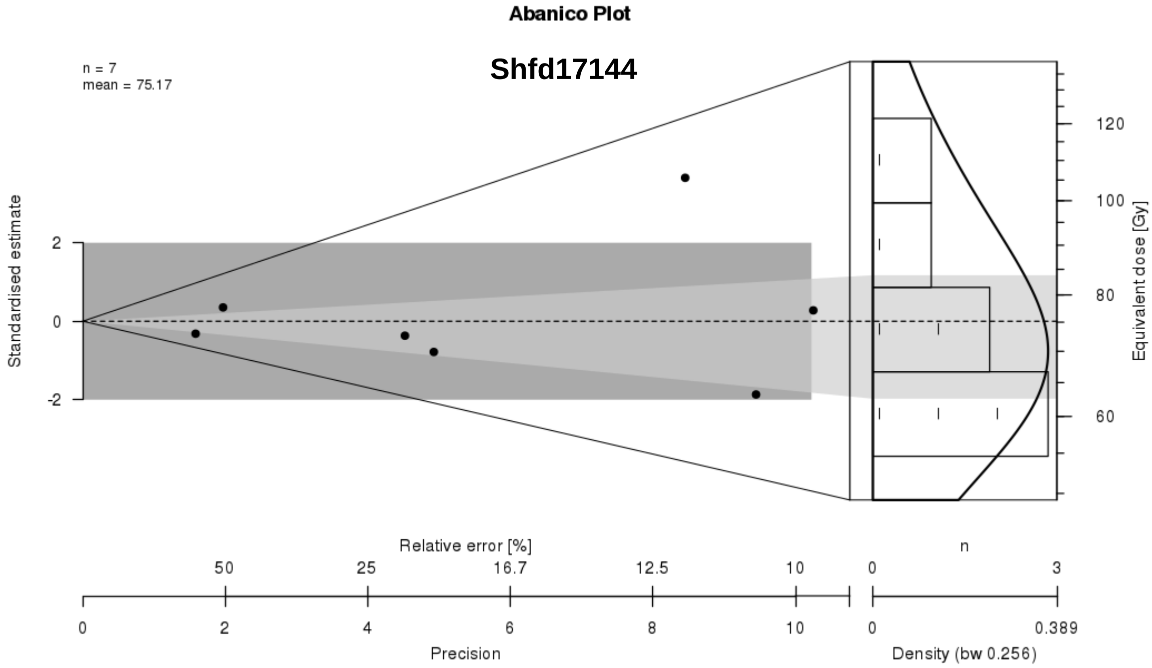


Figure 3.4.12: Single grain data gives a mean of 69.3 ± 4.4 Gys (the CAM provided a mean after outliers were removed of 69.9 Gys), which returns an age of 22.7 ± 1.9 ka. This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

Sample Shfd16174 (Figure 3.4.10) is taken from the second opaque core from 2-3 m (Table 2.3.1). The OD is 28% reducing to 22% once outliers are removed (Table 3.4.2). The distribution is bimodal with a subtle skew to the right (Figure 3.4.10), but there is no low D_e outliers so MAM was used. The weighted mean is 31.1 ± 3.8 Gys, returning an age of 24.7 ± 3.2 ka.

Similarly, sample Shfd16024 shows a bimodal distribution with a skew to the right, only more pronounced than Shfd16174 (Figure 3.4.11). This sample is taken from the third opaque core of 2-3 m (Table 3.4.1). The OD is 39% dropping to 21% once outliers are removed (Table 3.4.2). Likewise, the distribution is normal once outliers are removed so CAM was used with a mean of 35.9 ± 1.8 Gys resulting in an age of 32.5 ± 2.3 ka (Table 3.4.2).

Shfd17144 produced an age based on data from seven single grains (Figure 3.4.12). This was included in an attempted to constrain the age of the lake clays at High Marishes and Yedingham. The data was normally distributed after the removal of one outlier. With such a small dataset the Common Age Model (COM) was used (the CAM was unable to calculate an error once outliers were removed). This gave a mean of 69.3 ± 4.4 Gys (the CAM provided a mean after outliers were removed of 69.9 Gys), which returns an age of 22.7 ± 1.9 ka (Table 3.4.2). This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

Interpretation

The dates provided by the High Marishes cores show good stratigraphic agreement and increasing age with depth. The ages show that the upper sand unit is post-glacial with an age of 16.3 ± 1.0 ka (Shfd16162). Underlying this sand deposit is a coarse, grey sand deposit that potentially represents a pulse of sediment entering into the lake system (section 3.3) providing an age of 24. 24.7 ± 3.2 ka. The final core (Shfd16024) is medium grained sand, similar to the deposit at Yedingham (Shfd16021) and providing a similar age of 32.5 ± 2.3 ka. A sample of sand grains washed out from a core of clay taken from core 8 of the clear lined cores gives an age of 22.7 ± 1.9 ka. The difficulty with this sample is the small amount of grains. They are potentially partially bleached since they do not align in stratigraphic order with the above sediments. This is supported by the distal location of the site at High Marishes where the upper water column may have been clearer allowing for some light to attenuate as icerafted debris rained out, partially resetting the grains. Alternatively, although every precaution was taken when extracting the grains from the clay, they could potentially be contamination from the coring process, which may have allowed fractures to open along the planes of the laminations and move sand grains from the exterior into the interior.

Overall, the OSL ages calculated from the sediments at High Marishes reveal that, in agreement with Yedingham, there is more than one iteration of Lake Pickering with an older lake prior to 32.5 ± 2.3 ka.

Wellfield Farm, Salton

Wellfield Farm is in the western section of the Vale of Pickering (Figure 3.4.1). This site was chosen as it is potentially located at the edge of the Devensian-aged lake between 26 m and 30 m O.D. Three opaque cores were taken for OSL dating (Section 3.3).

Sample Shfd16164 (Figure 3.4.13) was extracted from the middle of core 1 (Table 3.4.1; section 3.3). It shows partial bleaching indicated by the high OD of 55%, which only reduces slightly to 48% when outliers are removed. A low D_e value affected the MAM and so the FMM was used. Component 1 has a mean of 10.6 ± 1.2 Gys, which provides an age of 5.0 ± 0.6 ka (Table 3.4.2).

Sample Shfd16165 (Figure 3.4.14) was taken from the second core (section 3.3) and also shows partial bleaching with a high OD of 43%, reducing to 34% once outliers are removed. A single low D_e value skewed the MAM and so the FMM was used. Component 1 has a mean of 40.0 ± 2.2 Gys that provides an age of 17.1 ± 1.6 ka (Table 3.4.2).

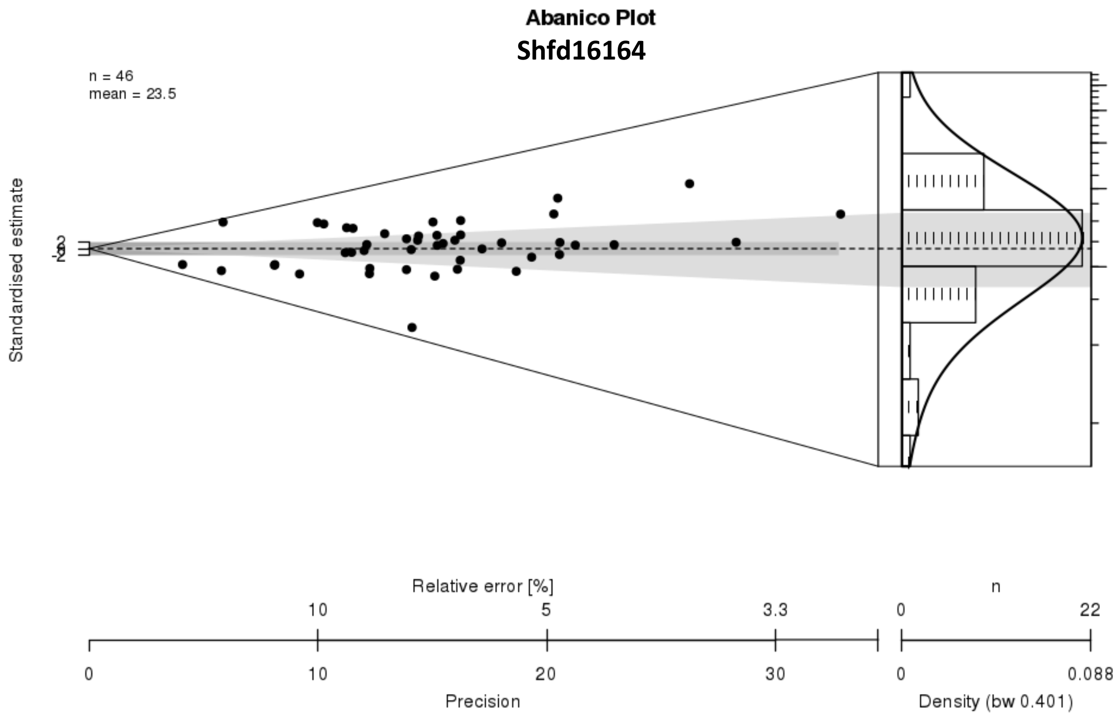


Figure 3.4.13: Shfd16164 has an OD of 55% dropping to 48% with outliers removed. The sample shows partial bleaching. The weighted mean of the MAM is 13.9 ± 3.0 Gys, providing an age of 6.6 ± 1.5 ka.

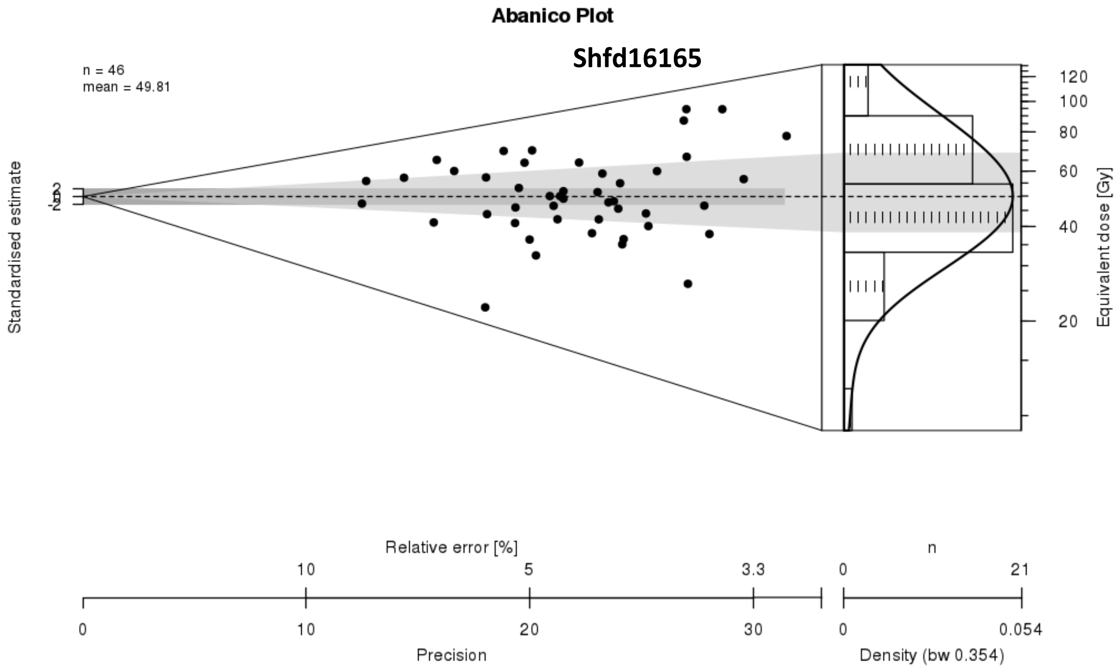


Figure 3.4.14 Shfd16165 shows partial bleaching and has an OD of 42%, reduced to 34 after outliers are removed. The mean of 40.0 ± 2.2 Gys from component 1 gives an age of 17.1 ± 1.6 ka.

Sample 16022 (Figure 3.4.15) was taken from laminated sands in the third core (section 3.3). The data is skewed to the left. The OD is 29% reducing to 16% after outliers are removed. A low D_e value skews the MAM and so component 1 of the FMM was used. This gives a mean of 31.9 ± 1.2 Gys, which equates to an age of 24.0 ± 1.6 ka (Table 3.4.2)

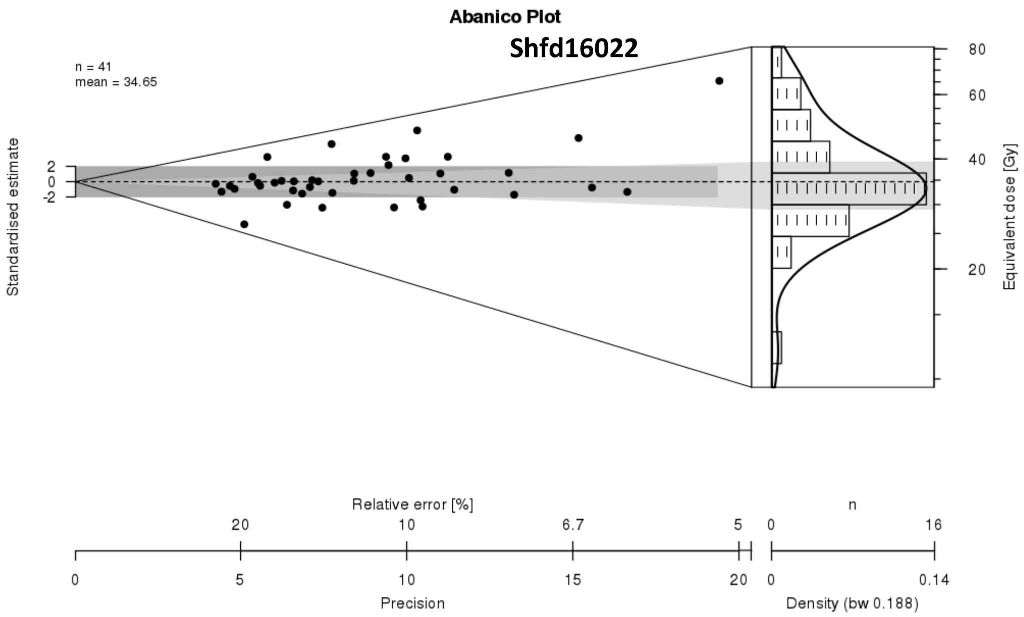


Figure 3.4.15: Sample 16022 has an OD of 29% reducing to 16% once outliers are removed. The data is skewed by a low D_e value so the FMM is applied. A mean of 31.9 ± 1.2 Gys gives an age of 24.0 ± 1.6 ka.

Interpretation

The deposit at Wellfield Farm shows a relatively low energy lacustrine environment that was punctuated by sediment pulses. A large portion of the core is disturbed between where sample Shfd16022 and Shfd16164 making interpretation difficult. The dates show this to be the LGM phase where the site at Salton was on the edge of the 33 m lake, but where reorganisation of sediment through a small outlet west of Salton may have impacted the depositional environment with pulses of sediment movement and reorganisation in a shallow lake. The lack of lacustrine clay at this site, but within the lacustrine phase of the LGM suggests that the sediment here was under a more fluvially active portion of the lake and so finer particle were unable to settle out to form a clay layer. This is likely because the site at Wellfield Farm sits at 26 m O.D and was likely more influenced by lake level fluctuations than the sites at High Marishes and Yedingham. The high OD of Samples Shfd16164 and Shfd16165 suggest that the sediment was partially bleached, most likely due to the extensive subaqueous reworking and redistribution at an active lake margin where high sediment load may have protected grains from exposure or provided very limited exposure.

Slingsby

Slingsby is located in the south west Vale of Pickering (Figure 3.4.1). A sample of sand was taken from below lacustrine clay (section 3.3). Shfd16019 (Figure 3.4.16) has an OD of 29%, reducing to 22% once outliers are removed (Table 3.4.2). The data is skewed to the right and a low D_e value skewed the MAM, so component 1 of the FMM was used. This gave a mean of 36.0 ± 2.3 Gy returning an age of 24.8 ± 2.0 ka.

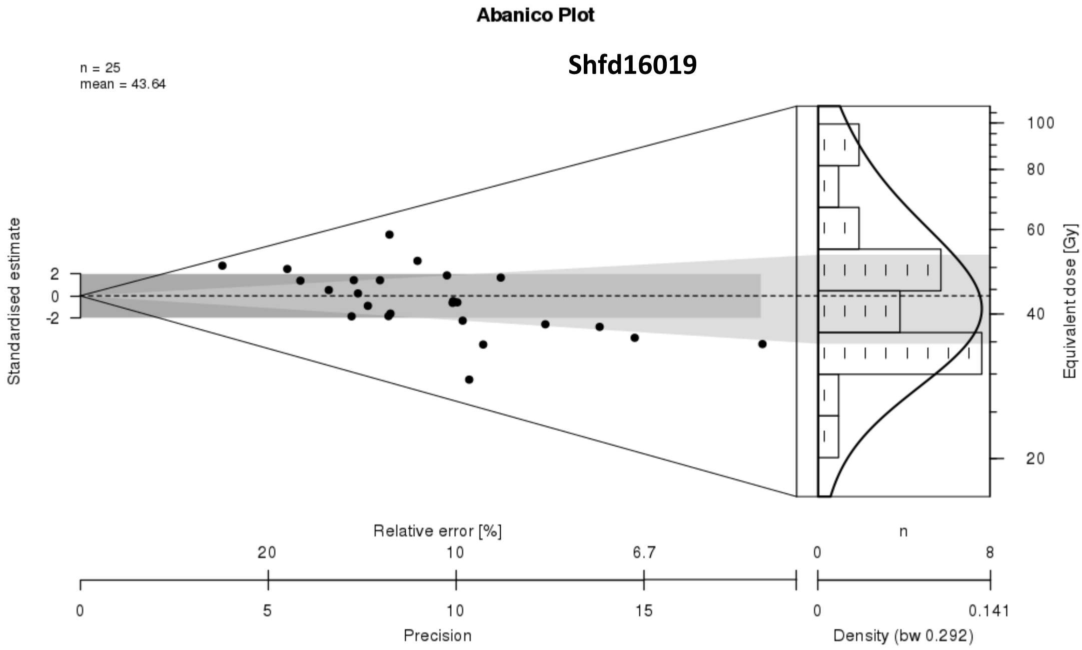


Figure 3.4.16: Shfd16019 has an OD of 29% reducing to 22% with outliers removed. Component 1 of the FMM was used as a low D_e value skewed the MAM. A mean of 36.0 ± 2.3 Gy gave an age of 24.8 ± 2.0 ka.

Interpretation

Sample 16019 was taken from a sand deposit beneath a meter of lacustrine clay (section 3.3). The age returned of 24. 8 ± 2.0 suggests that the LGM lake developed near Slingsby after this time. Slingsby sits at an altitude of 30 m O.D, suggesting that the lake extended to at least this height, but probably represents the maximum extent of LGM Lake Pickering of 33 m since Slingsby is at the western limit of lacustrine clay. Laminated sand from Salton (which lies 6.5 km to the northeast) shows a similar age of 24.0 ± 1.6 ka (Shfd16022).

Caulkley's Bank

The site at Caulkley's Bank lies 2 km north of Slingsby at the footslope of a long east-west trending limestone spur (Figure 3.4.1). This site was selected for dating as it sits on the 33 m O.D. contour (section 3.3), the hypothesised maximum LGM lake extent. Sample Shfd16018 (Figure 3.4.17) has a OD of 34% reducing to 10% with outliers removed. Once outliers are removed, the distribution is normal, so CAM was used. This provided a mean of 114.1 ± 2.7 Gy returning an age of 90.0 ± 5.4 ka (Table 3.4.2). This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

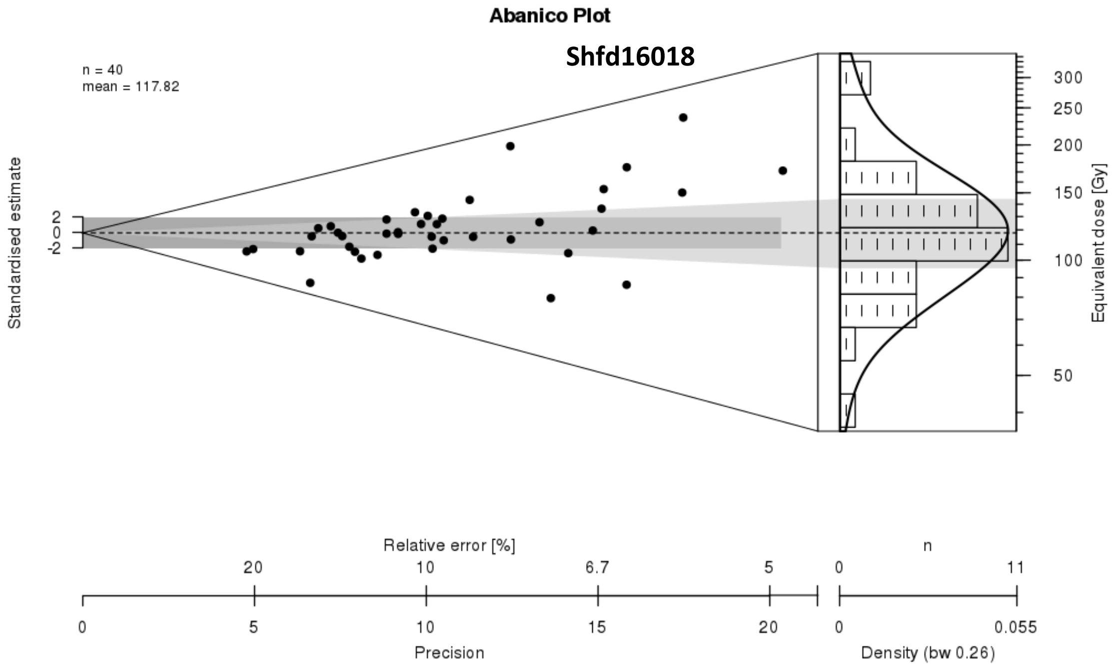


Figure 3.4.17: Sample Shfd16018 has a n OD of 34% reducing to 10% with outliers removed. After outliers are removed, the distribution is normal, so CAM was used. This gave a mean of 114.3 ± 4.8 Gy, resulting in an age of 90.9 ± 5.4 ka. This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

Interpretation

Shfd16018 (Figure 3.4.17) was taken from a hand augured sample (section 3.3) at a depth of 2.3 m (Table 3.4.1). No lacustrine clay was encountered in any of the 2.5 m cored. Mapping in this locality (section 3.2) shows a potential shoreline 400 m to the east at 29.4 m O.D suggesting that LGM lake waters may not have reached 33 m at the base of Caulkley's Bank. However, as the deposit likely represents slumping of material (section 3.3), these features may be covered. The older age returned of 90.9 ± 5.4 ka and the normal distribution of the D_e values (after outliers are removed) with a low OD (after outliers are removed) of 10% suggest the grains have not been reset and may be remnant from the MIS 5 interglacial.

Eden Camp

Eden Camp is located in the south-central portion of the Vale of Pickering (Figure 3.4.1) 2 km north east of Malton. Sample Shfd16161 (Figure 3.4.18) taken from a massive sand deposit (section 3.3) at a depth of 1.3 m (Table 3.4.1) from a pit opened for pipe repair work by Yorkshire Water. Care was taken to extract a sample from below any disturbed sediment. Shfd16161 has a bimodal distribution with an OD of 33%, reducing to 24% once outliers are removed. Component 1 of the FMM was used as two low D_e values skewed the MAM giving a mean of 26.4 ± 1.0 Gy. This provides an age of 21.7 ± 1.3 ka.

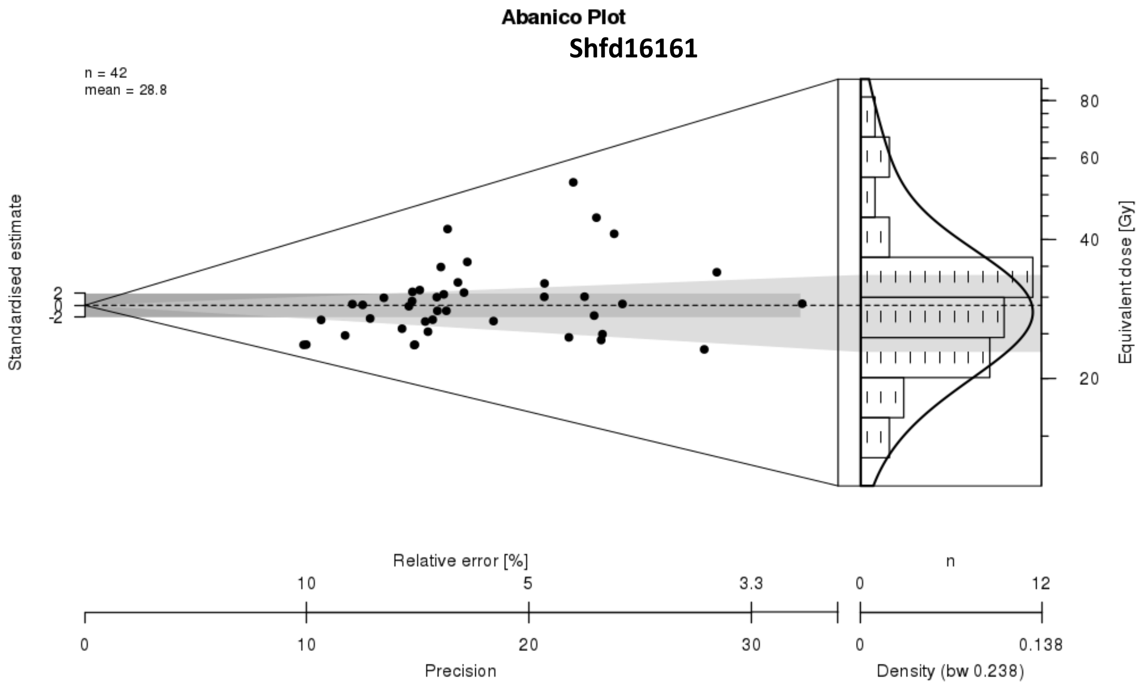


Figure 3.4.18: Shfd16161 has an OD of 33% reducing to 24% once outliers are removed. Component 1 of the FMM was used as two low D_e values skewed the MAM giving a mean of 26.4 ± 1.0 Gy. This provides an age of 21.7 ± 1.3 ka.

Interpretation

The site at Eden Camp is located close to the confluence of the River Rye with the River Derwent (Section 3.3). The River Derwent then heads south to the Kirkham Gorge. The deposit represents an over-estimation (by the use of the MAM). There is no sedimentary structure (Section 3.3) and is possibly sand reorganised and deposited meltwater probably ponding from a reduction in the velocity of water as it waited to exit through the outlet at Malton as lake levels dropped.

Hunmanby Gap

The site at Hunmanby Gap is located on the Yorkshire coast 4.5 km southeast of Filey (Figure 3.4.1). It is an infilled (potentially subglacial) lake basin situated between two till packages (section 3.3). Sample Shfd16157 was taken from a sandy layer at the top of the deposit. The palaeodose distribution is multimodal with a high OD% of 42%, reducing to 41% with outliers removed. Component 1 of the FMM was used because the MAM result would have been skewed by a few low D_e values. This resulted in a mean of 138.2 ± 9.9 Gy and an age of 70.5 ± 6.3 ka (Table 3.4.2). This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

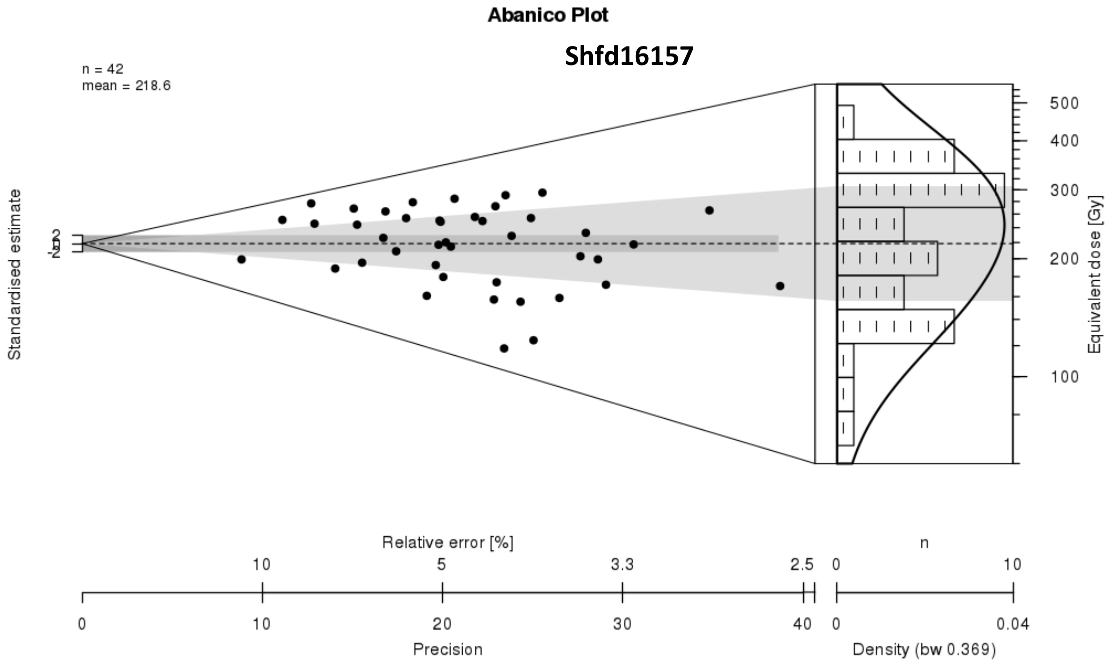


Figure 3.4.19: Shfd16157 has an OD of 42% reducing slightly to 41% with outliers removed. Component 1 of the FMM provides a mean of 138.9 ± 9.9 Gy, returning an age of 70.5 ± 6.3. This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

Shfd16158 (Figure 3.4.20) was taken two metres below Shfd16157 from a sand layer. The sample also shows partial bleaching as indicated by an OD of 63% reducing to 56% after the removal of outliers. It has a multimodal distribution and, as a result, it was thought that the MAM if used would have led to age underestimation due to two low D_e outliers. Therefore, component 1 of the FMM was considered a better representation with a mean of 49.8 ± 3.1 Gy, which resulted in an age of 35.5 ± 2.8 ka. This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

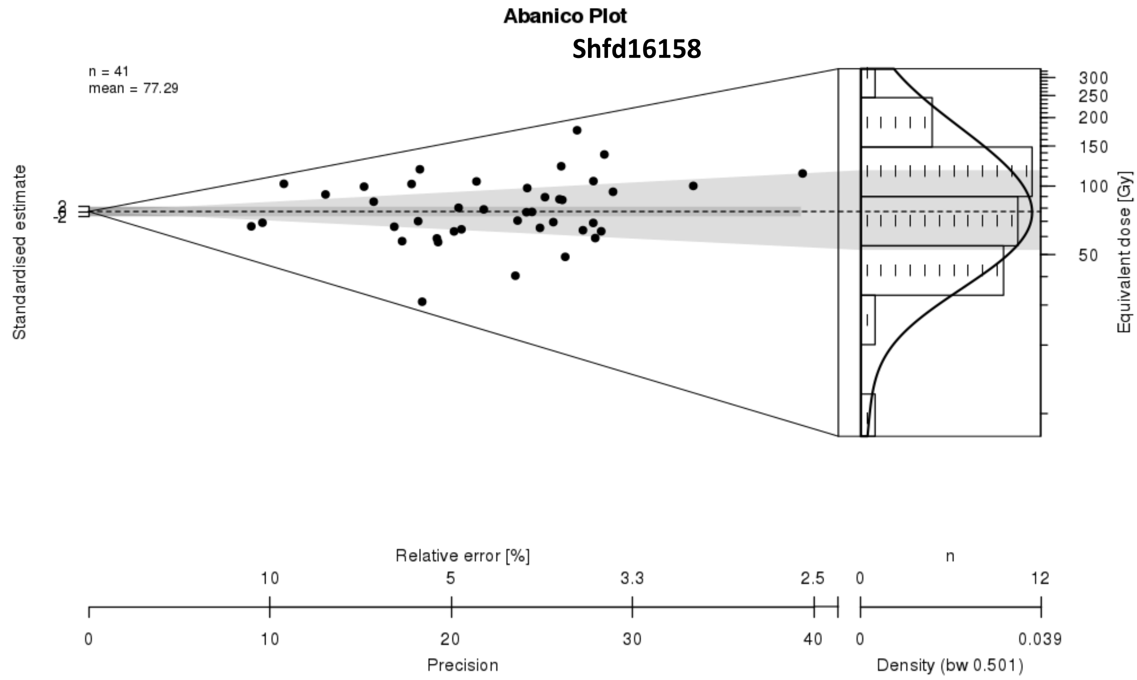


Figure 3.4.20: Sample 16158 has an OD of 63% reducing to 56% after outliers are removed. Component 1 of the FMM gives a mean of 49.8 ± 3.1 Gy returning an age of 35.5 ± 2.8 ka. This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

Interpretation

Looking at the site ages as a whole, it is clear that there is a large age discrepancy between the two samples. Given the stratigraphic context of a discrete sand-filled lake basin, which may have been sub-glacial between two tills (the upper which is thought to be LGM), it is unlikely that both ages are correct. As both samples suffered from partial bleaching and at the single aliquot level FMM will only mitigate this, and it is considered that Shfd16157 is definitely too old and over-estimating the burial age. Shfd16158 may be correct or at least only a small over-estimation. This age gives a maximum age for the overlying till emplacement and potentially indicates an earlier glacial advance within the Late Devensian (i.e: Roberts *et al*., 2018).

Cayton Bay

Cayton Bay is located on the north side of the North Yorkshire coast towards Scarborough. The deposit here is a rotated slump, possibly comprised of sands from between two diamicts like the deposit at Hunmanby Gap (Figure 3.4.1).

The OD values for Shfd16159 (Figure 3.4.21) are high with 54% dropping to 48% once outliers are removed. The OD of 54%, dropping slightly to 48% with the removal of outliers, and multimodal D_e distribution indicates this sample was only partially bleached at deposition. Field evidence of bioturbation from nesting birds led to the decision to avoid an age based on MAM and instead base the age on component 1 of the FMM (Table 3.4.2). This resulted in a mean of 62.8 ± 4.1 Gy, resulting in an age of 49.0 ± 4.1 ka. Whilst it may still be slightly over-estimating true age, is considered to give a maximum age for the overlying till emplacement. This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

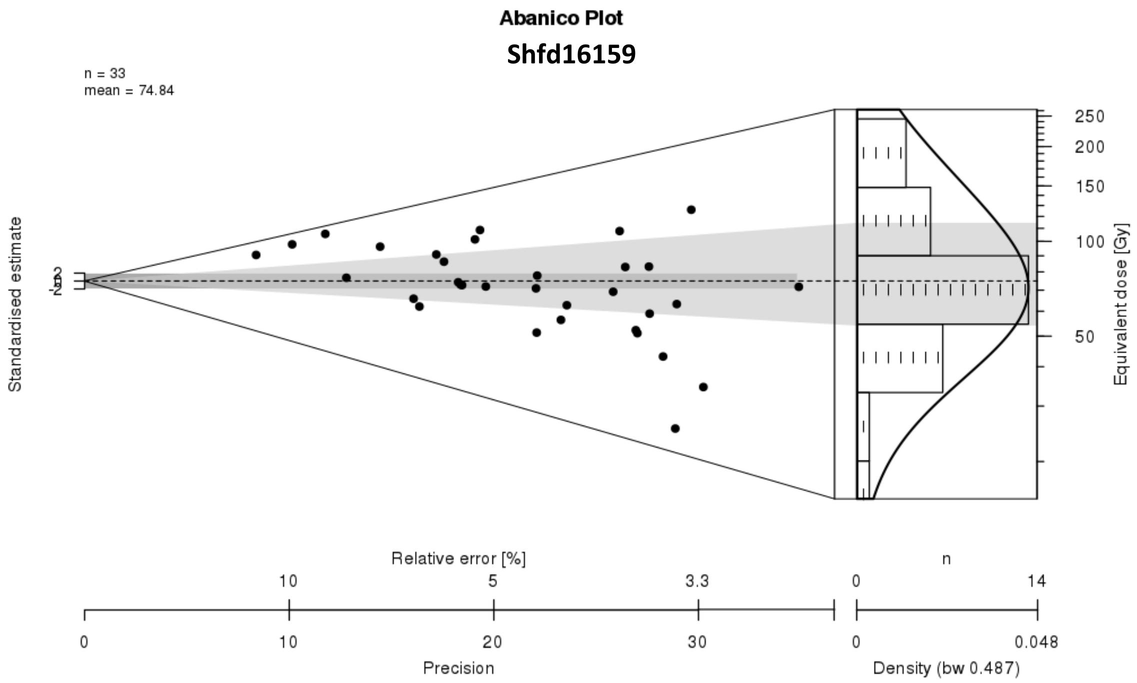


Figure 3.4.21: Shfd16159 has an OD of 54% reducing slightly to 48% with outliers removed. Component 1 of the FMM was used giving a mean of 62.8 ± 4.1 Gy, resulting in an age of 49.0 ± 4.1 ka, This result is considered unreliable as it does not fit the straigraphic constraints provided by other age estimates.

Interpretation

Similarly, to Hunmanby Gap, the deposit at Cayton Bay is comprised of partially bleached glaciogenic sediment. It returns an age of 49.0 ± 4.1 ka that is likely to be an overestimation and the deposit is likely to date to a pre-LGM advance (Roberts *et al*., 2018).

3.5 Lake Level Models

Several flood fills corrected for glacial isostatic adjustment (GIA: see section 2.6) were created to establish the relationship between Lake Humber and Lake Pickering. The first were created following the chronology set out in Bateman *et al*. 2018 (Figure 3.5.1) The results showed that the level of Lake Humber did affect the level of Lake Pickering and that if the Kirkham Gorge was incised to its current base level (10 – 15 m), then the two lakes were interconnected and had a hydraulic stabilizing affect. The model fails to create a 70 m lake, with 45 m reached between 22 -21 ka with a high stage Lake Humber at 33 m and ice is at the terminal position at Wykeham in the Vale of Pickering.

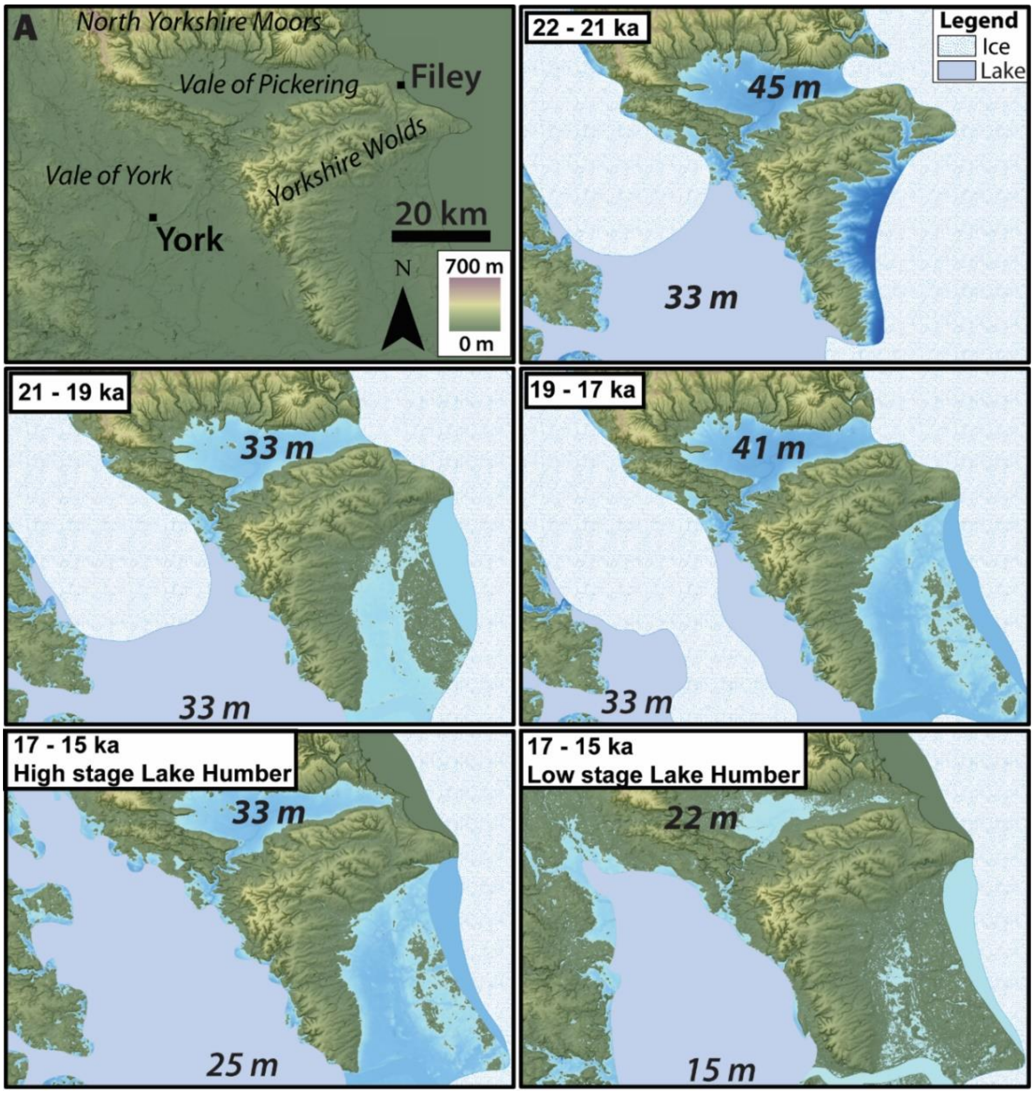


Figure 3.5.1: The relationship between Lake Pickering and Lake Humber based on reconstructions by Bateman et al., 2018. Lake Pickering and lake Humber appear to have a strong influence upon each other.

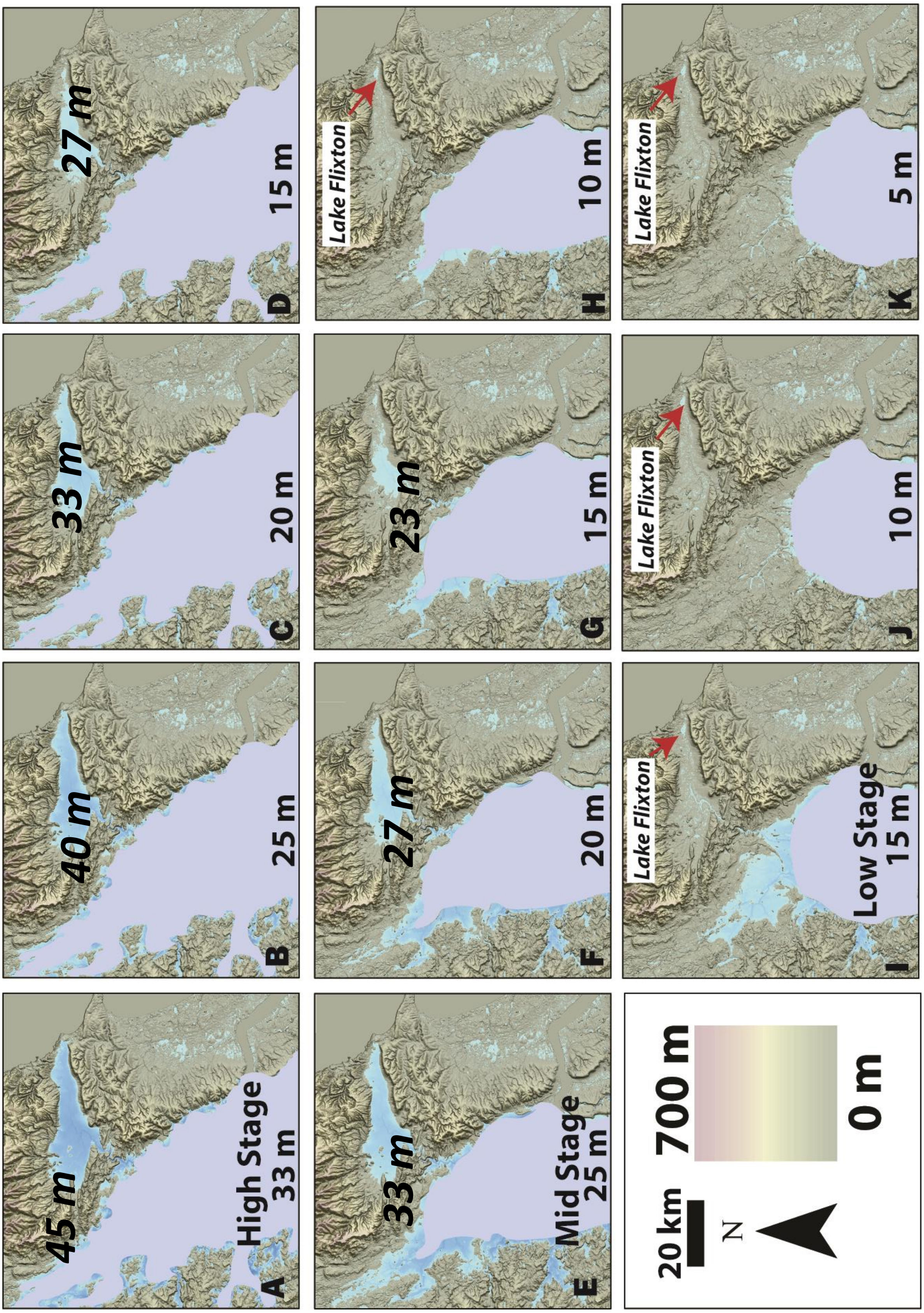


Figure 3.5.2: Experiments to understand the effect of Lake Humber on the height of Lake Pickering based on the height and extent of Lake Humber

A slight retreat of the ice margins between 21 and 19 ka reduces the lake to 33 m. However, when ice readvances in the Vale of York, the model shows an increase from 33 m to 41 m between 19 and 17 ka. As the ice retreats from 17 ka, the lake drops once again to 33 m and down to 22 m by 15 ka when the NSL had left the area completely. After this time, the lake would have drained and only Lake Flixton remained.

A second experiment was conducted based upon the height and extent of Lake Humber (Figure 3.5.2). These results show that a larger and extensive Lake Humber needs to occupy the Vale of York to control Lake Pickering through the Kirkham Gorge. Once the lake drops past a height of 10 m in the mid stage, there is not enough of a hydraulic barrier to constrain the flow of the Derwent through the Kirkham Gorge. With a low stage Lake Humber, only Lake Flixton remains. This suggested that the base of the Kirkham Gorge was responsible for lake height and not the presence of Lake Humber.

The limits of these flood fills are apparent. Despite being adjusted for GIA, the flood fills are based upon a post-glacial landscape and were not altered to reflect a pre-glacial topography. This is evident in the height of the River Derwent through the Kirkham Gorge, which has likely been down cut post-glacially. To rectify this, terraces mapped at the mouth of the Kirkham Gorge were used to model a lake based on a Kirkham Gorge river bed height of 25 m (Figure 3.5.3) in line with the terraces identified. This showed that even a small increase in river base level in the Kirkham Gorge can form a lake without the need for a connection with Lake Humber. This confirmed that it is the height of the river bed for the outflow of the lake at Kirkham Gorge that affects the height of Lake Pickering. This is in agreement with recent work by Murton (2018) who concluded that the LGM height of Lake Humber did not exceed 10 m O.D.

The final flood fill experiment was to create Kendall's (1902) 70 m lake. Two ways of creating a 70 m lake were tested. The first was by plugging the col at Kirkham by filling the Gorge to a height of 70m (Figure 3.5.4). The second was to create an ice dam in the Vale of York, blocking the Kirkham Gorge outlet (Figure 3.5.5).

To model the effect of ice filling the Vale of York, the ice was expanded from the Escrick terminal moraine to the reach the east edge of the Yorkshire Wolds. It is important to note that this limit was not reached during the LGM according to mapping by the BGS and published literature (e.g.: Bateman *et al*., 2018).

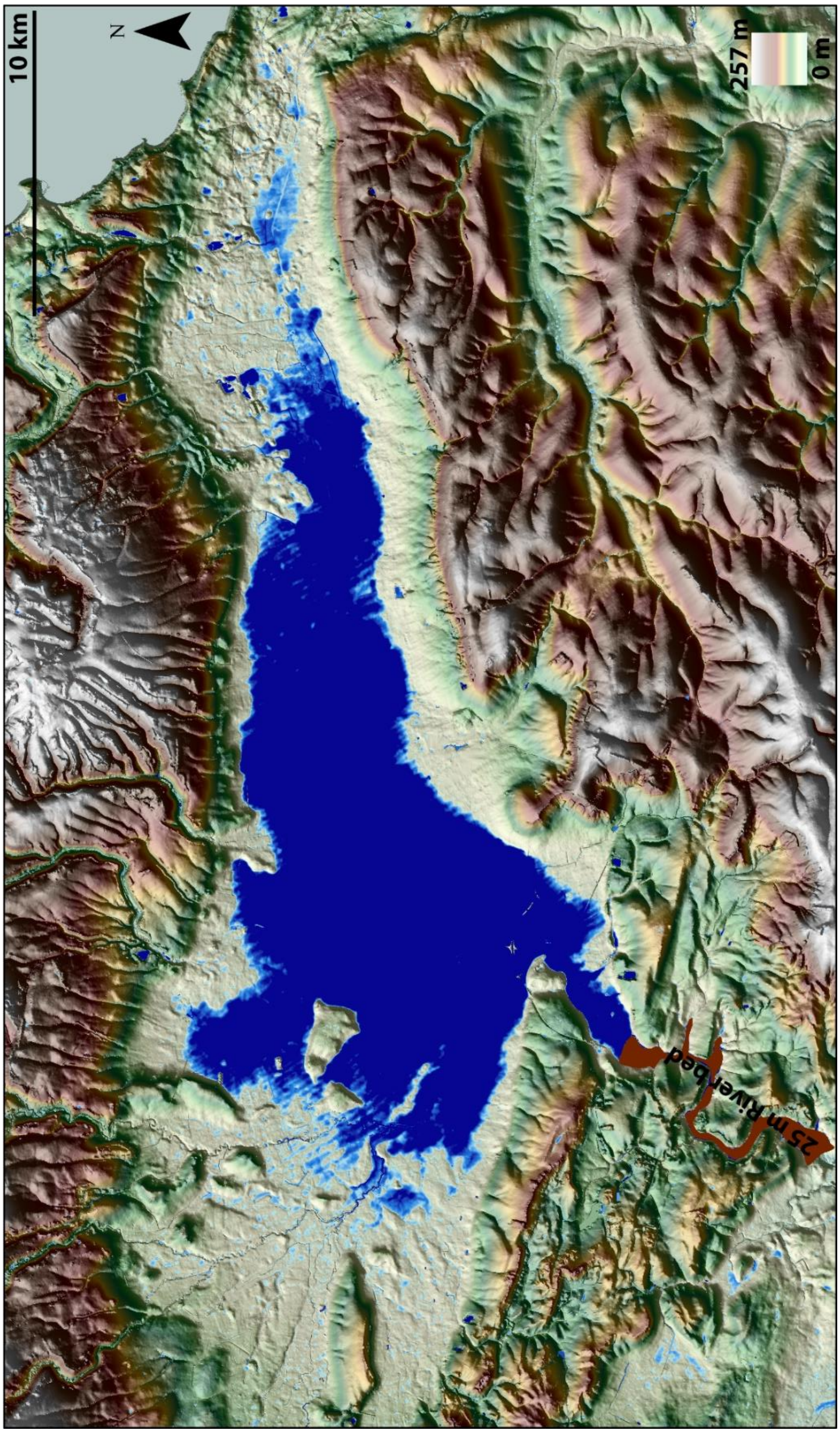


Figure 3.5.3: A Kirkham Gorge river bed height of 25 m creates a 25 m lake in the Vale of Pickering, suggesting that the base level of outlet in Kirkham Gorge has more of an affect that the presence of Lake Humber.

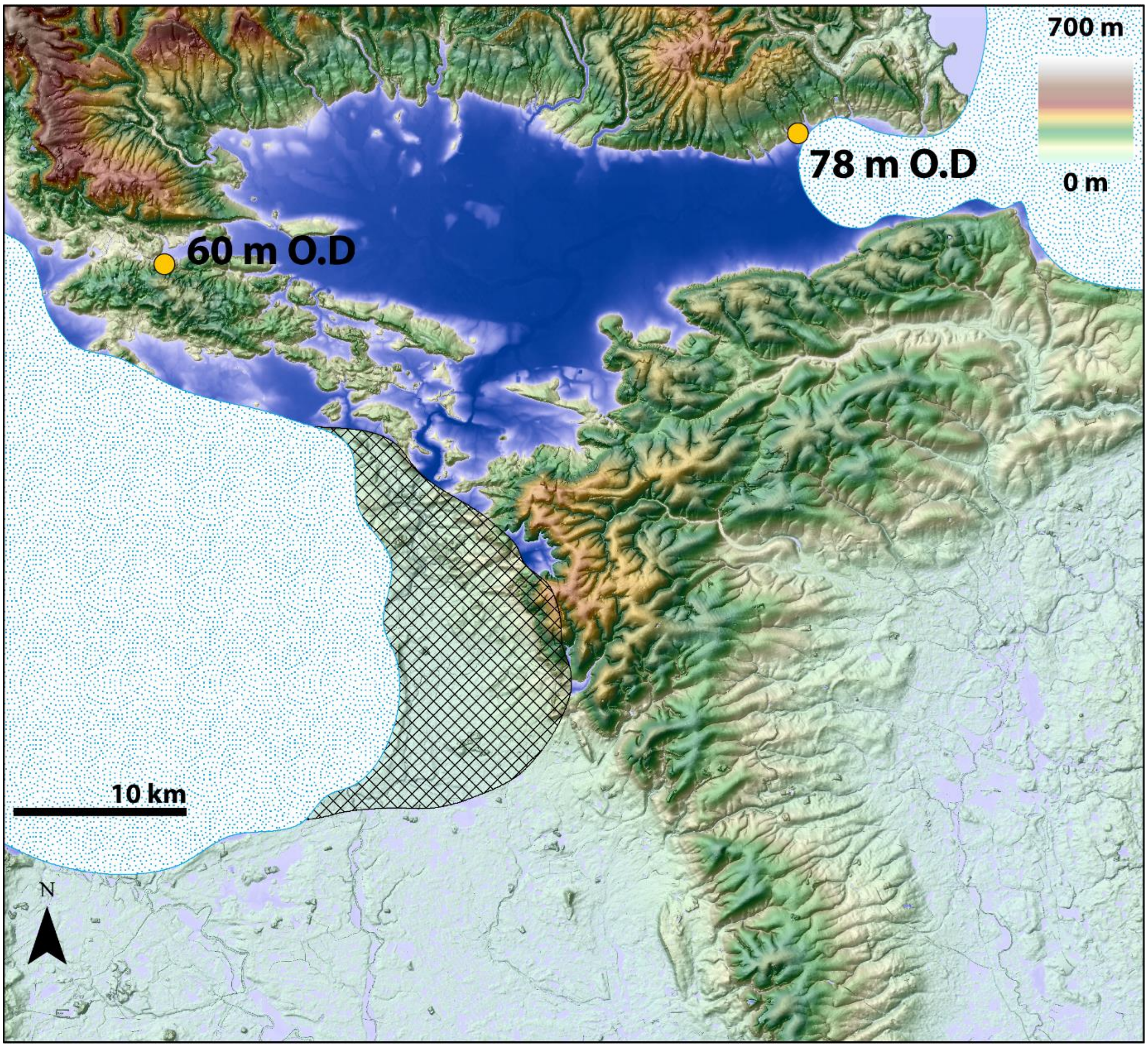


Figure 3.5.5: Ice expanded (shaded area) to fill the Vale of York creates a 60 m O.D lake reaching heights of 78 m O.D next to the Wykeham moraine, although it should be noted that localised affects from glacial ice (e,g: forebulge, sediment blockages, etc, at ice fronts) that may affect drainage cannot be modelled.

In Figure 3.5.5, the expanded ice creates lake with an average extent of 60 m O.D, reaching 78 m O.D near the Wykeham moraine. It is also important to note the overspill of the lake in the Kirkham Gorge supporting Franks (1987) and work by Eddey and Lincoln (2017) showing that a 70 m O.D lake could not be supported due to various spillways that exist in the Howardian Hills. It is also worth noting that the Hunmanby spillway at the east end (Section 3.2) is not utilised due to the presence of moraines. It is likely that this outlet was used. It is also possible drainage of the lake eastwards occurred under the ice sheet.

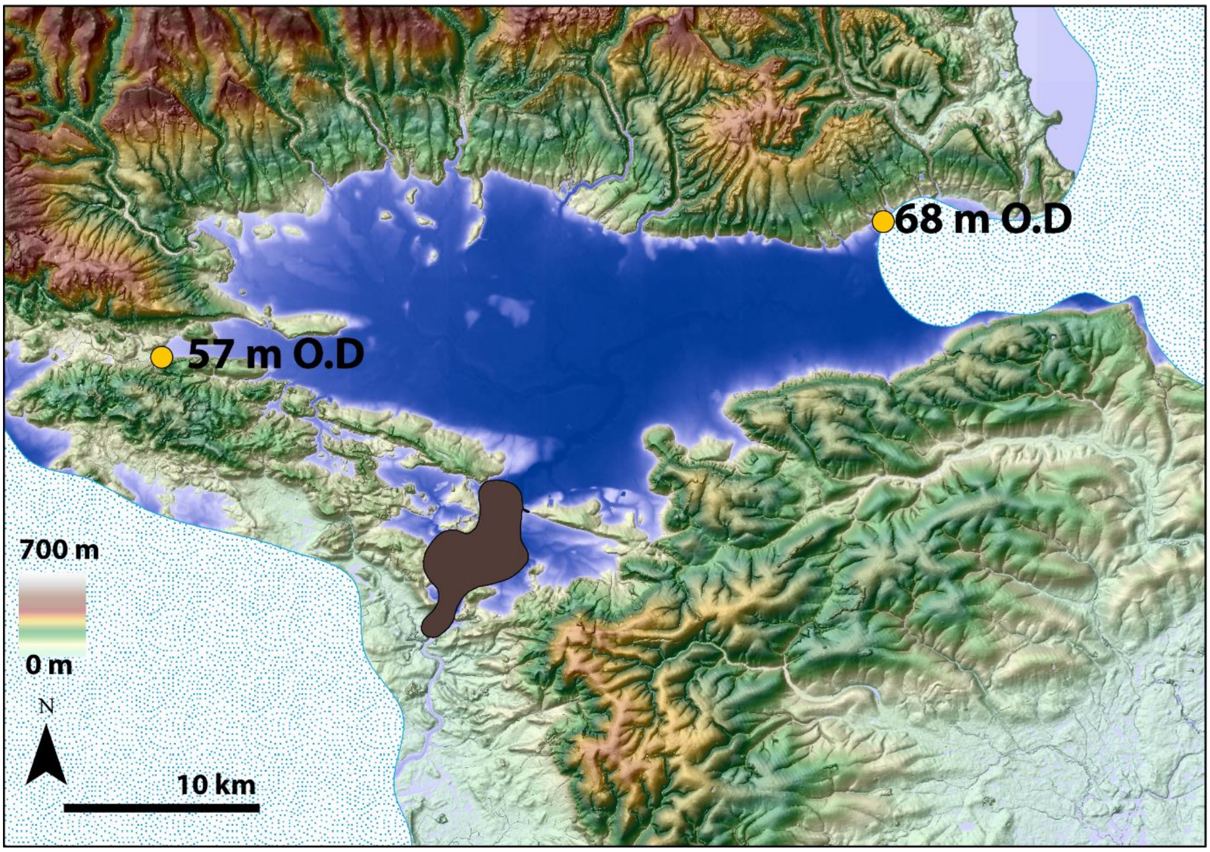


Figure 3.5.6: A blockage in the Kirkham Gorge create a lake with an average extent of 57 m O.D. At the Wykeham moraine, a height of 68 m O.D is reached lending support to Kendall's lake height.

Similarly, to the ice block, restricting the outlet through the Kirkham Gorge with 70 m O.D blockage creates a higher lake level with an average level of 58 m O.D rising to 68 m O.D at the Wykeham moraine. However, lake waters here still utilise other spillways available in the Howardian Hills. From these two experiments, it is likely that in the Howardian Hills the evidence for these higher lake levels is likely to be found.

The final experiment involved taking a cross section of the Vale of Pickering using the available GIA data (Bradley, *pers comm.)* to establish the change in elevation from 24 ka to 16 ka. The results are shown in Figure 3.5.7. The graph shows that the GIA response to the fluctuating margin of the NSL is reflected in the basin response at its lowest at 16 ka lowering by up to 13 m and highest at 20 ka, lowering by 7 m. This could have an impact on the drainage of the channels as the basin is also tilted to become flatter towards the eastern end, which would have an effect on whether the lake was draining eastwards or through the outlet at the Kirkham Gorge.

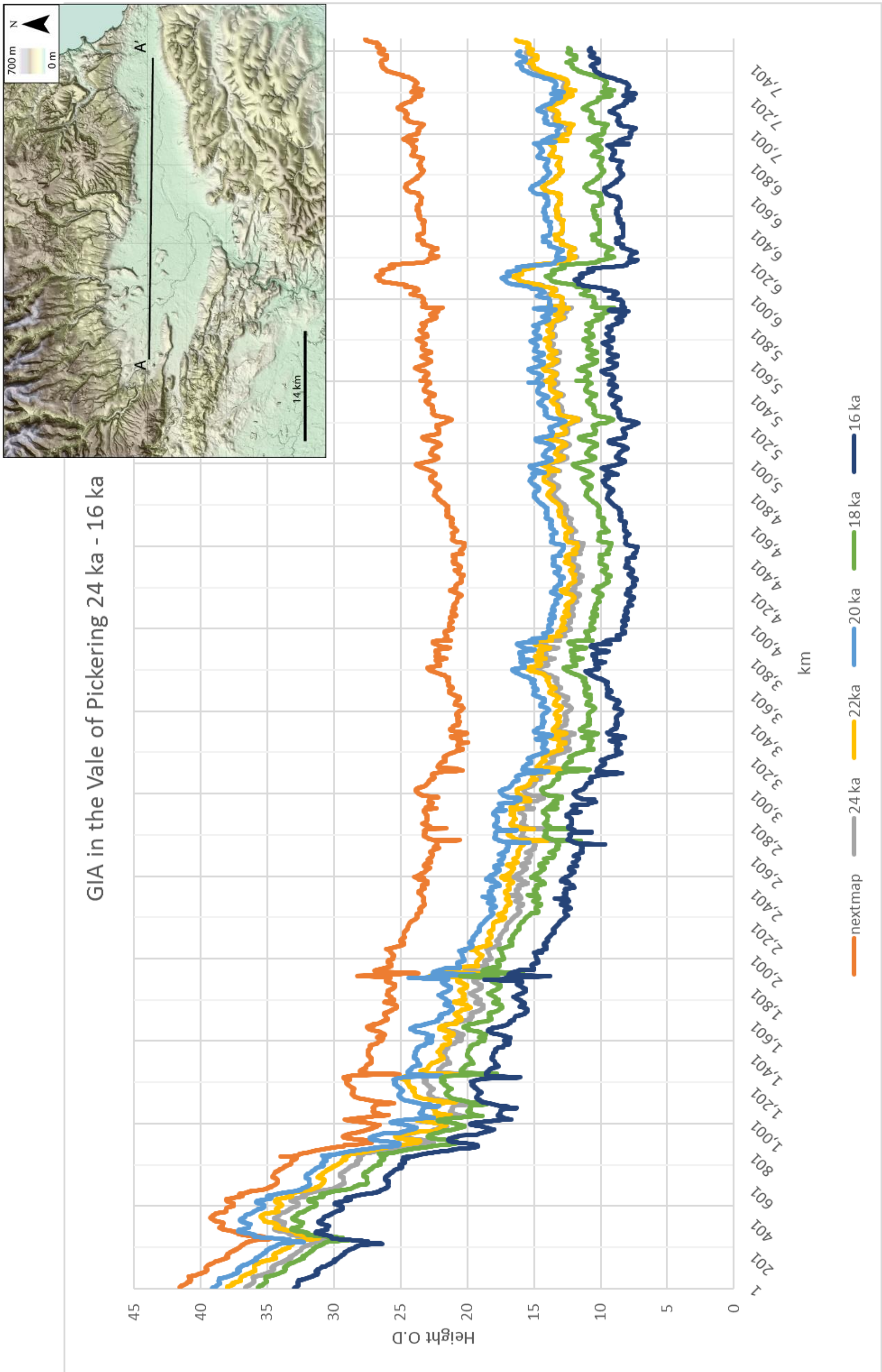


Figure 3.5.7: GIA data for the Vale of Pickering courtesy of Dr. Sarah Bradley. Data shows that the largest change in altitude comes at 16 ka with the least at 20 ka.

4. Interpretation

4.1. What was the maximum extent of Glacial Lake Pickering?

Glacial Lake Pickering has long been associated with a maximum height of 70 m O.D. (Kendall, 1902; Straw, 1979), lowering to 45 m O.D., to 33 m O.D and eventually to 25 m O.D. However, while there is plenty of evidence for lacustrine conditions, geomorphological analysis of the area finds little indication of a 70 m lake. The natural break in bedrock between the sedimentary rocks of the surrounding hills and the softer underlying Jurassic clays create a basin at about 70 m O.D. Previous authors (Kendall, 1902; Straw, 1979) envision this basin filling up with water before overspilling into the adjacent Vale of York cutting the Kirkham Gorge down from 70 m O.D. to its present river base height of 15 m O.D. It may be that the natural shape of the Vale of Pickering basin that lends to the image of a 70 m lake.

The evidence presented by Kendall (1902) for a 70 m lake is the height of terraces of the Hutton Buscel kame terrace (69 m O.D.) and the elevation of the col near Coxwold (68 m O.D.) If ever there was a lake of this height, all evidence has since been eroded. Additionally, there are at least four points, notably at in the Howardian Hills (such as Gally Gap) and south of Hunmanby, that sit lower than 70 m where water can drain from the Vale of Pickering (Section 1.7). Furthermore, Kendall made no allowance for changes in landscape due to GIA, which would have lowered the landscape and tilted it towards the northwest (Bradley *et al*., 2011; Kuchar *et al*., 2012) (Section 2.6). Nor did he consider more localised factors like ponding between the ice margin at Wykeham and the Corallian dipslope. This could have created the 68 m terrace on the Hutton Buscel kame, but would not have extended into the Vale of Pickering basin (Franks, 1987).

Lake height modelling in Section 3.5 reveals that very high lake levels (58-78 m O.D.) are possible but requires either significant blockages in the Kirkham Gorge or ice filling the entirety of the Vale of York. While the ice margins in the Vale of York for the LGM may have extended past the Escrick moraine (Friend *et al*., 2011; Bateman *et al*., 2018), the flood fill modelling shows that this is not enough to create a 70 m lake height (Figure 3.5.3).

This conjecture is supported by deposits at the famous Kirkdale Cave (Buckland, 1823) situated in the Vale of Pickering near Kirby Moorside at 58 m O.D. These were reanalysed by

McFarlane and Ford (1998) and dated to 121.0 ± 4.0 ka from flowstone overlying the cave deposits, placing the occupation of the cave in MIS 5e, the last interglacial. Deposits were not noted to be covered or affected by lacustrine deposits, suggesting that the lake level did not reach this high after the Ipswitchian. This suggests that the 45 m lake is possibly younger than MIS 5e while the 70 m lake cannot be. Furthermore, evidence for lacustrine deposits above ~ 48 m O.D. from both historic boreholes and field walking is non-existent. Despite these limitations, it is not impossible that a lake of this height existed during the Mid Pleistocene during MIS 6 or 12, but that the amount of time passed has removed the lake's signature.

Evidence for the 45 m lake

Mapping shows that there is geomorphic evidence for a 45 m O.D lake. The western Vale of Pickering has a distinctive 45 m surface that runs between the north side of Caulkely's Bank and Wombleton. Evidence of this is also seen in a series of historic boreholes drilled by Holst Soil Engineering, Leeds during the summer of 1973 (Section 3.1). The geomorphology of the area is interpreted as a former raised lake bed where the southern portion has been eroded by the meandering River Rye. Although undated, these sediments are likely to be older than LGM as OSL dates from Slingsby at 30 m O.D. (Shfd16019) and Wellfield Farm (Shfd16164, Shfd16165, Shfd16022) at 26 m O.D. suggest the extent of the LGM lake was around 33 m O.D. Further work is needed in this area.

The 45 m lake is likely to be pre-LGM and probably correlates with LF1 at Yedingham and LF1 and LF2 at High Marishes. OSL dates from both sites suggest a lacustrine phase prior to 30 ka. A sand unit above a lower lacustrine clay (LF1) at Yedingham provides an OSL age of 30.8 ± 3.6 ka at the base of LF2 and a sandy unit above a lower lacustrine (LF1-2) unit at High Marishes gives an OSL age of 32.5 ± 1.9 ka from the middle of LF3. It should be noted that both sites sit within the basin of the buried valley discussed in Section 1.5 and that as a deep channel may have exercised geomorphic control on the extent and depth of the lake. The clays described in the Helmsley series (Holst Soil Engineering boreholes) in the western Vale of Pickering differ in colour and texture and are not noted to be laminated like the deeper clays found in the CEGB section. This could be due to driller logging, but also suggests two different sources for the clays. The mention of mottling also suggests a degree of weathering from subaerial exposure (Franks, 1987). More work would need to be done to investigate the link between the clays in the western Vale of Pickering thought to make up the floor of the 45 m lake and the clays analysed in the 10 m borehole from High Marishes.

Additionally, work by Franks (1987) near Thornton le Dale shows the tops of Kimmeridge clay deposits to be a sequence of aeolian silts underlying a weathered gelifluctate (Section 1.6; section 3.2). There are no lacustrine deposits along the bank between Thornton le Dale and Ebberston over 40 m O.D. The lack of available LiDAR data makes it difficult to discern any shorelines between these two points, but erosion of the High Riggs Gravel complex suggests water levels no higher than 43 m O.D.

Evidence for the 33 m lake

Evidence for the 33 m lake is found at elevations between 35 and 27 m O.D due to changes in topography from post-glacial deposition, erosion of sediment as well as rebound from isostatic adjustment. In the western end of the Vale of Pickering where the rivers Rye and Riccal meet, a number of wide river valleys meet and grade into the flat lying floor at 25 m O.D with small potential shorelines at 35 m O.D, 33 m O.D, 31 m O.D. and 29 m O.D. Section 3.1. shows that many of the dry channels terminate at heights between 35 and 27 m O.D (Figure 3.1.23).

An OSL sample (Shfd16019) taken from sand under lacustrine clay 1.6 m deep at Slingsby, returned an age of 24.8 ± 2.0 ka. The site at Slingsby sits on the 30 m contour and geomorphic mapping and historic borehole records show that the lacustrine clay does not extend far beyond Slingsby. However, on the south side of Caulkley's Bank, Shfd16018 returned an age of 90.9 ± 5.6 ka at a depth of 2.3 m from a sandy deposit at 30 m O.D, although no lacustrine clay was encountered. It is possible that the lake waters were ephemeral but ponded at Slingsby due to lower lying land caused by the old river valley running from the Coxwold-Gilling Gap towards Malton. It is also possible that lacustrine clay exists below the deposit at Caulkely's Bank dated, but no historic boreholes exist in this area with which to compare.

Evidence for a 30 + m lake is seen in the terraces of the Pickering delta and from mapped shorelines in the basin between Riseborough and Pickering. On the 1:25 000 OS maps, the 33 m contour makes a sharp right from following the scarp of Riseborough across to Pickering. LiDAR reveals a series of shorelines showing the demise of the lake between 30 m and 23 m O.D. (Figure 4.1). This area suggests that the limit of the 33 m lake may have been influenced by the presence of the long Jurassic clay spur of Riseborough that has a maximum height of 72 m.

At Wykeham, the 33 m contour curves around the base of the Wykeham moraine and both the Forge Valley delta and Seamer delta enter the Vale of Pickering at 33 m O.D. Overall the erosion of the Vale of Pickering between 33m and the base level of 21 m O.D support the presence of 33 m lake. The lack of till in the upper portion of the CEGB logs (Figure 3.1.31) suggests that the deposits are reorganised sands and gravels from the Hutton Buscel complex reworked by the opening of the Forge Valley during the LGM. The kame kettle topography described by Lincoln (2017) has been estimated to have become subaerially exposed at a minimum of 15.24 ± 0.22 ca; ka BP based on plant microfossils at the bottom of a quarry pit at North Wykeham (Lincoln *et al*., 2017). These dates come from Lincoln's LFA2 a dark grey carbonate rich, coarse, silty fine sand that lies above a glaciolacustrine unit over 8 m deep. Lincoln's LFA1 has been correlated with LF3 from Yedingham, suggesting that the 33 m lake extended as far as Wykeham during the LGM and that the Hutton Buscel sands and gravels were deposited into the lake by opening of the Forge Valley. Dates of 39.0 ± 2.8 ka and 38.5 ± 3.7 ka are taken from sand above lacustrine clay at Yedingham that lie beneath glaciofluvial sands and gravels (Figure 3.2.5) and an age of 34.8 ± 1.9 ka from sand within the lacustrine clay of LF3 at Yedingham suggest that these ages are related to the deposition of the Canbrough gravels (Franks, 1987) and were not exposed to light when deposited at Yedingham. If these gravels were redistributed by the opening of the Forge Valley, it is inferred that Forge Valley was breached and cut sometime during the LGM in a 33 m lake while ice stood at the Cayton-Speeton stage (see 4.2)

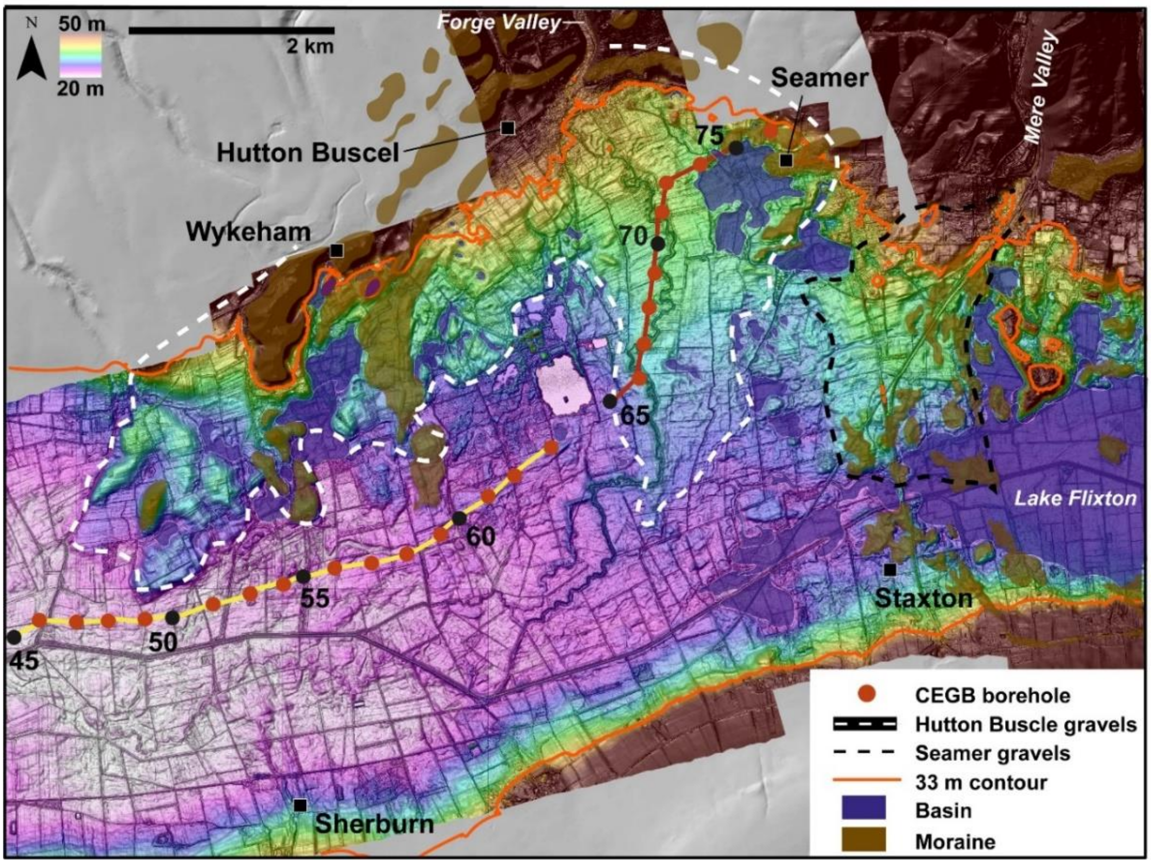


Figure 4.1.1: The Wykeham moraine, Hutton Buscel sands and gravels, Seamer gravel and the western side of Lake Flixton. The 33 m contour is outlined in orange and follows a fluvially eroded surface around to Ayton and the Forge Valley and eastwards towards Seamer. The boreholes of the CEGB from 45 to 76 are shown in their relationship to the surficial deposits and reveal a coarsening of material eastwards. There is little evidence of till deposits in the boreholes suggesting that the 33 m lake existed after the ice had left the Wykeham limit.

There is some evidence along the chalk scarp of the Yorkshire Wolds for the 33 m lake, but most of the evidence for higher lake levels underlies post-glacial slumping and alluvial fans of the Sherburn Sands (Evans *et al*., 2017). The lowest LF1 of Evans *et al*., 2017 gives an OSL age of 17.4 ka. This is likely to be from when the lake levels were dropping at the end of the LGM and may have been influenced by pulse of meltwater entering the system as deglaciation reorganised drainage into the basin (Edwards, 1978; Foster, 1985).

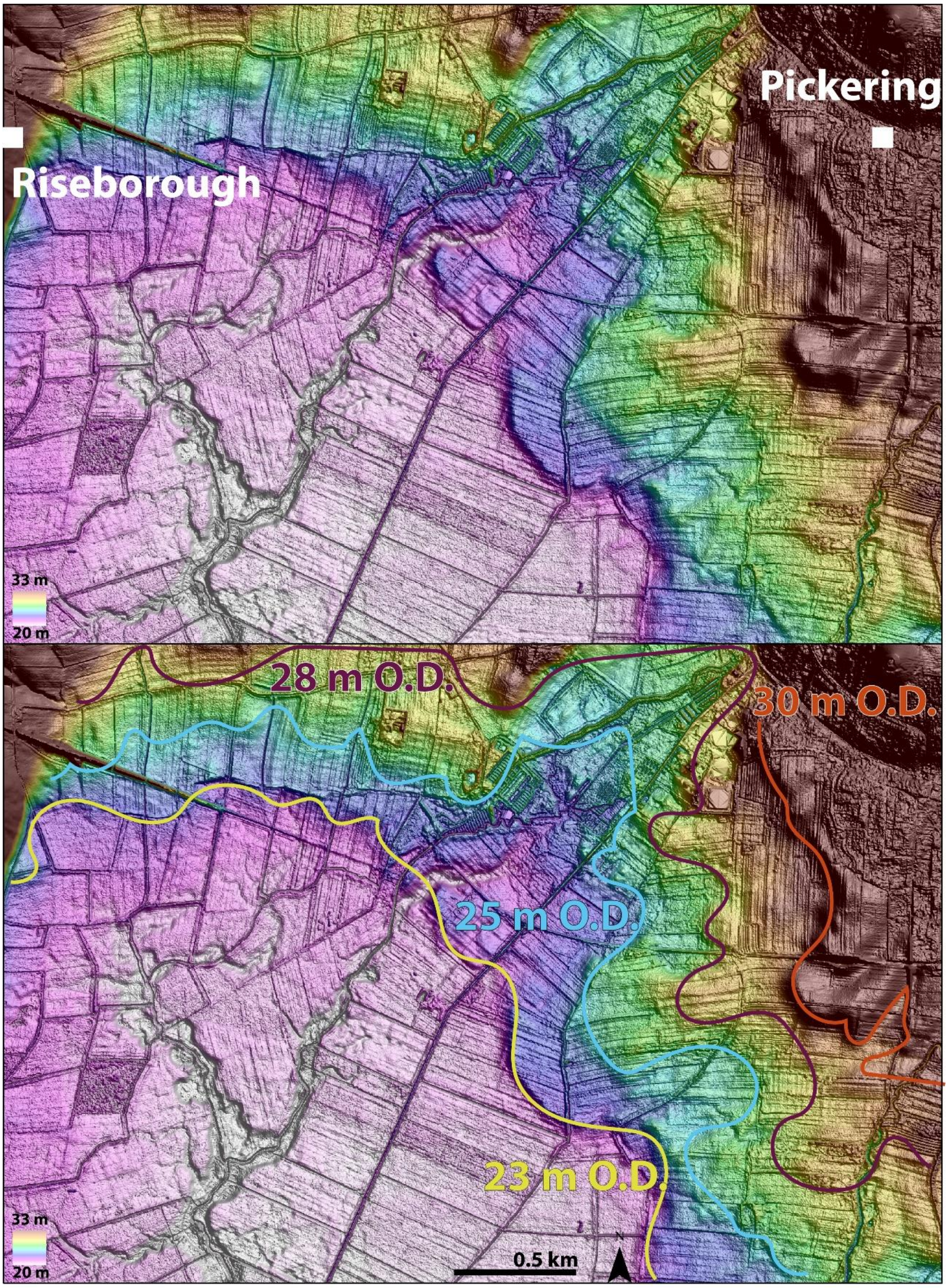


Figure 4.1.2: Shorelines outlining the Lake Pickering basin dropping from 30 m + at the mouth of Newton Dale in Pickering to 23 m in the low carrs between Pickering and Riseborough.

25 m lake and lower

The 25 m shoreline is the easiest to identify in the Vale of Pickering and is seen at elevations from 26 m O.D to 23 m O.D. The most prominent features are a suite of shorelines cut into the Sherburn Sands where alluvial fans fed into a lowering Lake Pickering (Figure 3.1.26). These potential shorelines are seen to run east-west along the foot of the chalk scarp from East Heslerton to Sherburn, although their extent past East Heslerton is unknown as the LiDAR data does not extend beyond this point, but the deposits extent to Knapton (King, 1965; Foster, 1985). Evans *et al*. (2017) describe the lacustrine sands below the alluvial fans as deposited during the 45 m – 33 m lake stage and dated to 17.4 ka, which, due to their sedimentology, are likely to be from a fluctuating shoreline of a shallow lake as it drained at the end of the LGM At this point, it is unlikely the lake was higher than 33 m and had probably dropped to the 25 m stage.

The 25 m terrace at the mouth of Kirkham Gorge suggests that the lake had to be 25 m or higher to cut the terraces so these deposits below the alluvial fans could be related to the drainage of the 33 m - 25 m lake. The 25 m – 23 m O.D. shoreline is also incised into the base of the Pickering delta, the Thornton le Dale alluvial fan and the base of the Kimmeridge clay hills located in the centre of the Vale. Drainage through the Kirkham Gorge may have been slow, and the underlying clays create an aquitard extending the life of Lake Pickering into the LGIT and probably existed for a time concurrently with a pro-Lake Flixton before separating. An OSL date from Eden Camp (Shfd16161) near to the mouth of the Malton Embayment at a depth of 1.3 m returned an age of 21.7 ± 1.3 ka suggesting that sand continued to accumulate in a shallow lacustrine setting. King (1965) suggests the Vale of Pickering existed into the Holocene as a reed swamp. This is supported by the presence of small areas of peat in the lower lying areas of the Vale and in the kettle holes examined by Lincoln (2017).

Overall, evidence for lake levels is present but often difficult to interpret due to subsequent erosion, reorganisation of deposits by post-glacial fluvial reworking, and from changes due to GIA. More work is needed to create a robust chronology for lake levels and extent beyond the LGM.

4.2 What was the extent of ice and pattern of retreat within the Vale of Pickering and how is it related to the dynamics of the British Irish Ice Sheet (BIIS)?

From the mapping completed in section 3.2, it is likely that ice ingress into the Vale of Pickering occurred several times during the Devensian. Contary to Edwards (1978) and Foster (1985) there is no evidence of LGM ice extending into the Vale of Pickering past the moraine at Wykeham (Franks, 1987; Evans *et al*., 2017). It is difficult to asertain the exact timing of the Wykeham advance as the deposits here are undated below the 15.24 ± 0.22 cal ka BP obtained from the base of a kettle hole located in the northern extension of Wykeham Quarry (Lincoln, 2017). Using borehole logs described by Hanson Aggregates (owners of Wykeham Quarry) Lincoln (2017) describes the basal deposits (LFA1) as consisting of glaciolacustrine laminated clays up to 8 m thick interbedded with massive sands containing occasional gravel clasts interpreted as dropstones (Lincoln *et al*., 2017) similar to the deposits found in the CEGB borehole cross section near Wykham (Figure 3.1.31) and LF3 at Ings Farm, Yedingham (Figure 3.2.5). Above this deposit, Lincoln (2017) describes LFA2 as lower and upper gravels with local and erratic lithologies similar to the lithologies found in the Skipsea Till and the Hutton Buscel sands and gravels (Lincoln *et al*., 2017). This description matches with the sands and gravels of LF6 at Ings Farm,Yedingham. On the CEGB cross section (Figure 3.1.31), the extension of the Hutton Buscel sands and gravels is shown from Borehole 65 to 76, although boreholes in the Derwent River valley (45-64) show these gravels extending beyond the mapped limit (Figure 4.1.1) fining as the sands and gravels are deposited westwards. Beneath these deposits, instances of lacustrine clay are seen in boreholes 45-57 and in 65 and 72, providing the most northward instance of evidence for Glacial Lake Pickering. However, which lacustrine layer the lacutrine clay from borehole 72 corrolates with (LF1 or LF3 at Yedingham) is uncertain, however, given the proximity to the Forge Valley, it is likely to be LF3 giving the 33 m lake an LGM age and confirimng the Forge Valley opened during the LGM.

There are some instances of patchy till in the CEGB cross section (Figure 3.1.31: Boreholes 57, 58, and 62) overlying lacustrine sand with a small unit of lacustrine clay in borehole 57. This lacustrine unit appears to laterally corrolate with sand below LF3 and LF1 at Ings Farm, Yedingham, suggesting it may be as old or older than LF2 at Yedingham. Furthermore, dates at Yedingham from LF3 (Shfd16023: 34.8 ± 1.9 ka) and LF4 (Shfd16025: 39.0 ± 2.8 ka and

Shfd16020: 38.5 ± 3.7 ka; Figure 3.2.5) suggest reworking of glaciofluvial sediment that was not exposed to light since last deposition by the opening of the Forge Valley. These sediments may likely be derived from material deposited during the Wykeham advance or retreat stages suggesting a pre-LGM date, possibly corresponding with the advance described in Roberts *et al*. (2018).

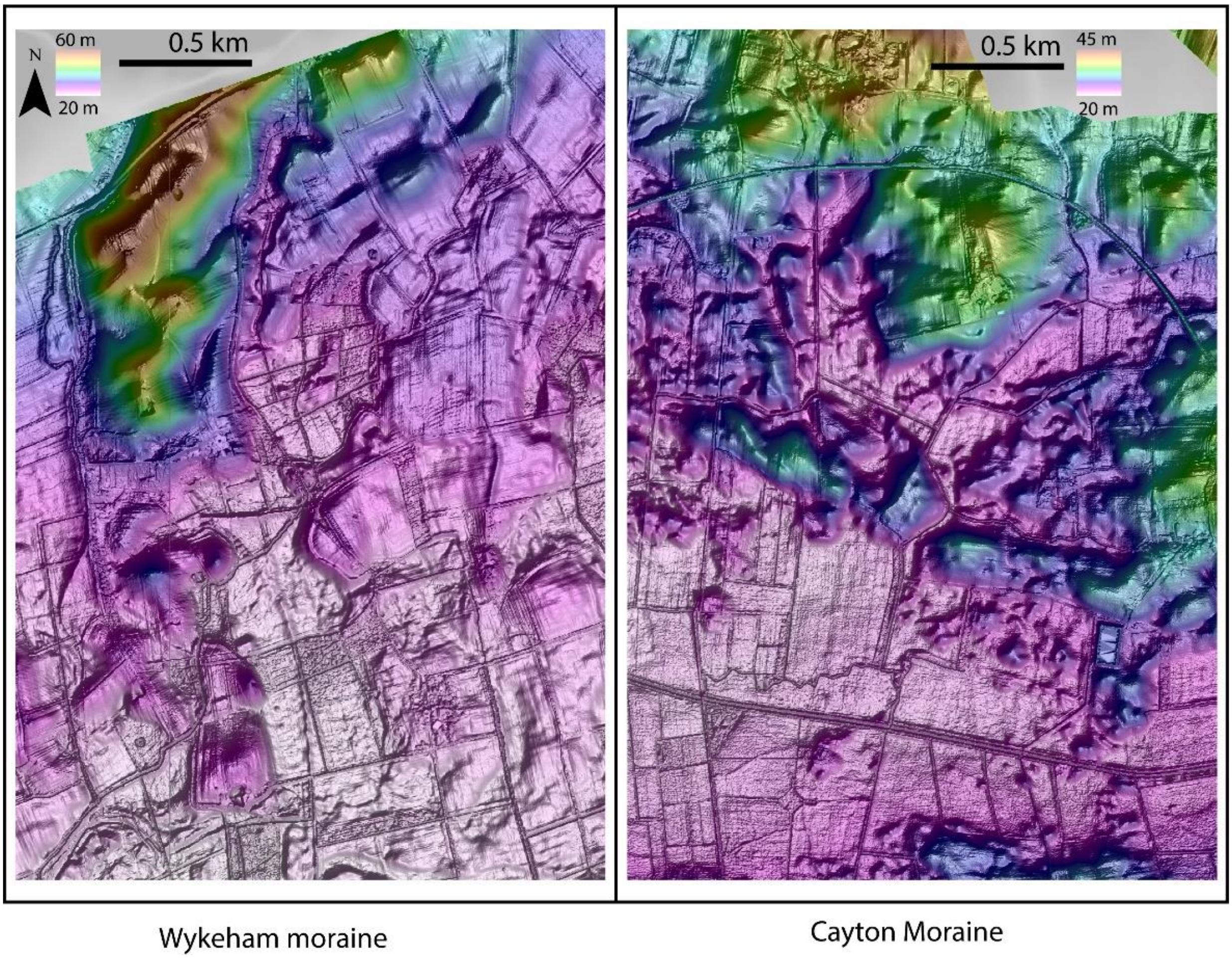


Figure 4.2.1: Comparison between the moraine at Wykeham (left) and the moraine near Cayton (left). The moraine at Wykeham appears less sharp, more rounded, and draped with sediment from interaction with a 33 m lake. The deposit at Cayton is unlikely to have been affected by a lake at 33 m

This is also supported by the work of Franks (1987) who identified a lower layer of horizonally stratified gravels (The Ayton Gravels), which contain ice-contact features (although these features are not fully described by Franks) that he attriburttes to a pre-LGM glacial episode. Above this unit is a layer of aeolian silt (Toft Hill silts) Franks correlated with the Harrow Hill silts at Thornton le Dale and with loess on the Yorkshire Wolds identified by Catt (1973). If Franks is correct, dating of the silts above the Ayton gravels could establish some timing of the extent of the Wykeham limit. More detailed comparison with the deposits at Yedingham and the Hutton Buscel complex could reveal further relationships between the two sites. Furthermore, the 33 m contour winds along the base of the terminal moraine at Wykeham (Figure 4.1.1) and along the lower terrace of the Hutton Buscel gravels. The small amount of till present in the deposits of the CEGB borehole cross section and the presence of dropstone described by Lincoln (2017) suggest that Wykeham ice may have advanced into Glacial Lake Pickering, which likely formed as ice advanced from the east coast blocking the drainage route at Filey. This advance may have resulted in an increased lake height, possibly forcing the opening of the Kirkham Gorge at Huttons Ambo (See section 4.3).

East of Wykeham, the evidence for recessional moraines is sporadic until Staxton where the western end of Paleolake Flixton is bound by a moraine couplet and the Seamer gravels. To the north is the Cayton moraine. The deposits appear slightly fresher here, with sharper ridges having not been drapped lacustrine deposits or softened by lake waters (Figure 4.2.1).

Moraines were mapped in section 3.1 from the limit at Wykeham to the Yorkshire coast between Cayton Bay and Flamborough Head. Figure 4.2.2 shows seven moraine clusters with 1 at Wykeham and 7 at Gristhorpe Cliffs. The relationship between the recessional moraines from Wykeham to Flixton is fairly straight forward as the ice leaves the Vale of Pickering, although the interaction between moranic deposits, meltwater, and lacustrine conditions in the Lake Flixton basin makes the moraines discontinuous. The moraines at Cayton and Filey, however, show a cross-cutting relationship that is difficult to unpick (Figure 4.2.2). At least two flow directions can be deduced. One northeast-southwest and one east-northeast to west-southwest (Figure 4.2.2.) Looking closer at the interaction between the moraines, the Cayton series seems to underlie the Filey series. This is area that would need more investigation to understand the true relationship between the retreat and readvance of ice, although work by Evans and Thompson (2010) show repeated osccilations by the NSL. This is supported by the work of Edwards (1978; 1987) and Boston *et al*., (2010) who show there is great variation between the numerous till packages on the East Yorkshire Coast (Section 1.3).

In the area of Filey Bay, there are several moraines that follow the shape of the chalk scarp of the Yorkshire Wolds as well as a large stacked moraine west of Filey town. This suggests that ice was oscillating on and off the coast, creating the stacked sequence of tills at Filey, but the ice was also likely affected by the high chalk scarp running from Muston to Hunmanby., Moraines at Hunmanby are crumpled following the shape of the angular fault that forms the topographic high of Flamborough Head. Penny and Rawson (1969) and Edwards (1978) both suggest that the landforms east of Hunmanby are drumlins, which seems unlikely as ice would have been thin and stalled at meeting the scarp from Hunmanby to Speeton. It is likely that they are part of the post-glacial fluvially dissected moraine ridges.

The dates for ice advance and retreat in the Vale of Pickering during the LGM can be correlated with work by Bateman *et al*. (2018) on the timing of the blockage of the Humber Gap. Bateman *et al*. (2018) stage 1 advance and deposition of the Skipsea Till is dated to an average age of 21.6 ± 1.3 ka, which is in agreement with the upper date (Shfd16163) from LF2 at Yedingham of 20.1 ± 3.2 ka. Skipsea Till ice retreat is dated by Bateman *et al*. (2018) to pre-17.0 ± 0.9 ka, which correlates (within errors) to sand dates at Yedingham Quarry 2 (Shfd16026: Figure 3.2.2) of 19.8 ± 2.6 ka. Dates at the Pottery, High Marishes (Figure 3.2.8) from the upper portion of LF3 (Shfd16174) of 24.7 ± 3.2 ka and LF4 (Shfd16162) of 16.3 ± 1.0 ka, which is located only 60 cm above Shfd16174 suggest low energy conditions between these two dates. At Salton (Figure 3.2.12), dates between LF1 (Shf16022) of 24.0 ± 1.6 ka and LF2 (Shfd16165) of 17.1 ± 1.6 ka (70 cm above Shfd16022) indicate a similar setting with low sedimentation rates during the LGM.

The timing of an earlier glaciation is supported by work from BRITICE and Roberts *et al*. (2018) who suggest an early ice advance down the Yorkshire coast pre-30 ka with a short-lived retreat at around 23 ka. OSL dates from sands lying between the upper and lower till series of 49.0 ± 4.1 ka at Cayton Bay (Shfd16159) and 35.5 ± 2.8 ka from Hunmanby Gap (Shfd16158) support an earlier advance. Dates from the sand above a LF1 at Yedingham (Shfd16021) of 30.8 ± 3.6 ka and below LF3 at Yedingham (Shfd16163) of 20.1 ± 3.2 ka and from below lacustrine clay at Slingsby (Shfd16019) of 24.8 ± 2.0 ka suggest movement of sediment into the lake between 30 ka and 25 ka. This lake may have shallowed given the presence of coal laminations in core 2 (LF2) at Yedingham similar to the ones seen at East Heslerton (Evans *et al*., 2017) indicating the opening of the east coast from glacial retreat. As a result, frequent oscillations of ice lobes on the east coast had an effect on the drainage of Lake Pickering, with perhaps short-lived openings allowing lake water to drain eastwards and into the North Sea Basin and cause the layers of coarser sediments to accumulate before once again damming the eastern end.

The ice advance and retreat of the NSL is summarised in Figure 1.3.5 by Bateman *et al*. (2018) and the correlation with the Vale of Pickering in Table 4.2.1. The table shows pre-LGM lacustrine

Table 4.2.1: Relationship between Lake Pickering and the North Sea Lobe

Event (After Bateman *et al*. 2018)	**Site**	**Sample**	**Relationship to ice**	**Age (ka)**	**Average age (ka)**
Pre-LGM	Yedingham	Shfd16021	Proximal Glaciolacustrine	30.8 ± 3.6	35.2 ± 2.9
	Yedingham	Shfd16023	Non-light exposed reworked glacialfluvial sediment	34.8 ± 1.9	
	Yedingham	Shfd16020		38.5 ± 3.7	
	Yedingham	Shfd16025		39.0 ± 2.8	
	High Marishes	Shfd16124	Delta sands/glacial sands	32.5 ± 2.3	
	Hunmanby Gap	Shfd16158	Subglacial sands between till packages	35.5 ± 2.8	
Pre-Skipsea	Slingsby	Shfd16019	Sand beneath lacustrine deposits	24.8 ± 2.0	23.1 ± 2.3
	Yedingham	Shf16163		20.1 ± 3.2	
	Salton	Shfd16022	Distal proglacial lake sediments	24.0 ± 1.6	
	High Marishes	Shfd16174		24.7 ± 3.2	
	Eden Camp	Shfd16161		21.7 ± 1.3	
Skipsea retreat	Yedingham	Shf16154	Proglacial sediment with meltwater input	22.2 ± 2.7	18.9 ± 2.0
	Yedingham	Shfd16026		19.8 ± 2.6	
	High Marishes	Shfd16162		16.3 ± 1.0	
	Salton	Shf16165		17.1 ± 1.6	

conditions from sand above clay at Yedingham and High Marishes, as well as beneath clay at Slingsby. These dates are supported by a date from Hunmanby Gap of 35.5 ± 2.8 ka from subglacial sand between two till packages, suggesting that this sand was reworked during the Skipsea advance. Deposits at Yedingham from LF3 and LF4 are interpreted as non-light exposed sediments reworked during the LGM retreat but are recycled from Wykeham advance deposits given their proximity giving an average age of 35.2 ± 2.9 ka (Table 4.2.1). Sand directly below lacustrine clay at Yedingham gives an approximate age for low energy lacustrine conditions suggesting ice blockage at the eastern end by the NSL with an average ag of 23.1 ± 2.3 ka (Table 4.2.1). As deglaciation occurs during the retreat of the NSL, sediment starts moving into Lake Pickering giving an average age of 18.9 ± 2.0 ka (Table 4.2.1).

To put this into wider context, the retreat pattern mapped shows a terminal moraine at Wykeham (Stage 1 in Figure 4.2.2) followed by a series of recessional moraines (2-4), which may include a readvance at Staxton from till overlying alluvial fan deposits (See section 3.1 on alluvial fans). It is interpreted that the ice limit at Cayton Bay (Stage 5) was prior to the ice limit at Filey (Stage 6) with a short-lived limit at Gristhorpe Cliffs (Stage 7) before retreating northwards.

Undoubtedly the moraine sequence on the east coast of the Vale of Pickering is highly complex. A simplified version of events as they related to the deposits studied is presented, but the timing between stages 5 to 7 is difficult to interpret and would benefit from more intensive study.

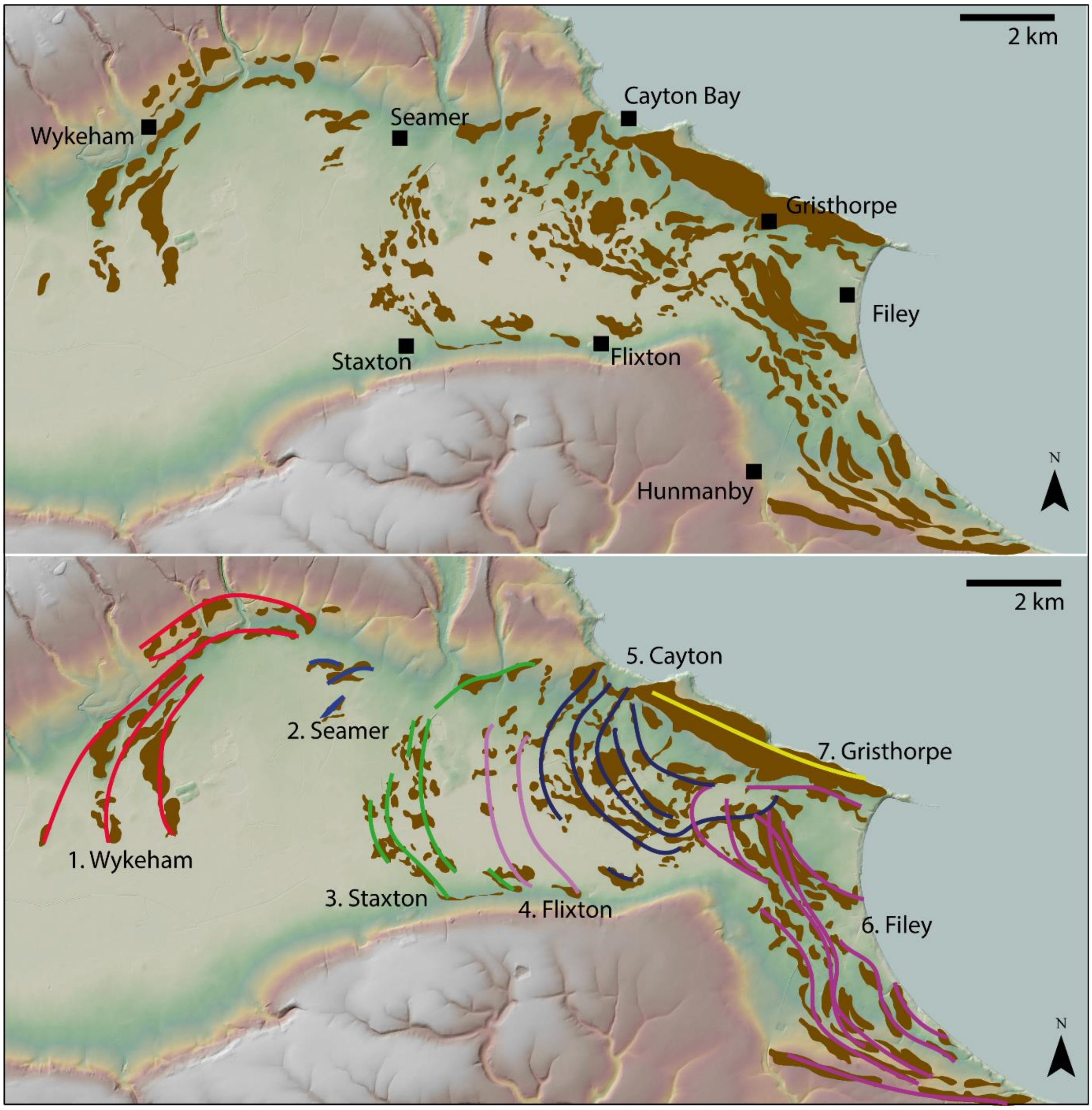


Figure 4.2.2: Retreat pattern in the eastern Vale of Pickering. 1. Wykeham stage: the terminal moraine and earliest advance of ice in to the Vale of Pickering. 2. Seamer: portions of recessional moraines remain interspersed with the Seamer Gravels. 3. Staxton: Seamer gravels frame the outline of the ice sitting in the Lake Flixton basin. Boreholes show till overlying delta gravels. 4. Flixton: moraines in the Flixton basin are partially buried by LGIT and Holocene sands, silts, and peats (e.g.: Palmer et al., 2015). 5. Cayton: or the Cayton-Speeton stage, which may have been two different sets of moraines with ice retreating northwards before readvancing in stage 6. 6. Filey: Ice moves back on to the east coast, crumpled moraines against the Hunmanby fault. 7. Gristhorpe. Final ice margin before complete retreat northwards at the end of the LGM.

4.3. What were the mechanisms behind the filling and emptying of Glacial Lake Pickering?

Establishing the mechanisms behind the filling and emptying of Lake Pickering is difficult to do with any precision. The lake was likely impounded many times during the Late Quaternary, but most likely drained eastwards due to the over deepened channel proved in the CEGB borehole cross section (Figure 3.1.13) created by the proto-River Ure. The reversal of the River Derwent is a response to the build-up of moraines at the eastern end of the Vale of Pickering that created a topographic barrier to the previously established drainage pattern to the east. Only the drainage of the River Derwent reversed direction; all other streams still flow their initial direction. Therefore, it is likely that prior to the LGM, the phases of Lake Pickering drained eastwards towards Filey. When this occurred is unknown, but it is likely a result of LGM glaciations given the large amounts of glacial debris deposited at Filey and along the coast. The reversal of the Derwent is likely to concurrent with the ice limit at Staxton.

Evidence for the reversal in drainage is seen at the embayment near Malton (Figure 3.1.28). The size and extent of the embayment suggests a well-developed river valley, with many smaller streams joining its trend north-eastwards. The general drainage into the Vale of Pickering is through a number of rivers including the River Rye, Riccal, Dove, Severn and several smaller becks: Holbeck, Costa Beck, Thornton Beck, Settrington Beck, Welldale Beck, Brompton Beck, Marrs Beck, and Wath Beck that sit in misfit channels (Section 3.1). Due to the chalk bedrock, no becks or rivers flow from the Yorkshire Wolds into the Vale of Pickering. The most south easterly channel being Scampston Beck, five miles east of Malton, still drains northwards into the Vale of Pickering. There is spring fed channel at Sherburn where two small lakes rest in an old chalk river valley. This over-deepened river channel is likely all that is left of the terminal ice margin on the southern side of the Vale of Pickering (See section 4.2).

The number of north draining streams in the Vale of Pickering is considerably lower than the number of abandoned channels. Currently only Scampston Beck, Marrs Beck, Wath Beck, and Settrington Beck drain northwards into the Vale of Pickering while there are over 40 dry channels with a northerly trend. The channels on the scarp of the Yorkshire Wolds are likely to be periglacial channels when the chalk acted as a frozen aquitard creating temporary meltwater channels. The origin of the river channels etched into the limestone of the north slope of the Howardian Hills may have a similar origin, but some are dendritic, suggesting a degree of river maturity and it is the adjustment due to isostatic depression or uplift that has caused the drainage divide to shift northwards, and result in the abandonment of these channels.

Drainage from the North Yorkshire Moors is generally confined to the large limestone dales and open valleys that run north-south through the Tabular Hills into the Vale of Pickering with the deposits of cobbles in the delta at Newton Dale suggesting that at times there was large amounts of high-energy meltwater entering the Lake Pickering from the North Yorkshire Moors and the Eskdale Lakes (Kendall, 1902).

The drainage of Lake Pickering through the Kirkham Gorge during this stage was likely to have been slow due to ice in the Vale of York, and potentially the presence of Lake Humber, but predominantly the grade of erosion down from around 26 m to 15 m presently. It is likely that the drainage of the River Derwent through the southern half of the terminal and recessional moraines from Wykeham to Sherburn resulted in a large amount of sediment deposited along the southern margin of the Yorkshire Wolds known as the Sherburn Sands.

The Kirkham Gorge is undoubtedly a complex area where underlying geology controls much of the geomorphic expression, but events in the Vale of Pickering have interrupted the natural order. Borehole logs (#958; #968) from the A64 along the terrace into the Vale of Pickering from High Hutton to Malton show orange mottled sands and clays, while a borehole (#906) from Low Hutton shows the older diamicton noted by Powell *et al*. (2016) overlying laminated Jurassic clays (Kimmeridge clay). This suggests that the terraces cut could either be very old as little diamicton remains or very young with the diamicton having been eroded. Terraces at either end of Kirkham Gorge may be able to shed some light on the precise timing of the drainage reversal, as well as further investigation of the meltwater channels at Filey and Hunmanby.

The drainage of Lake Pickering and when the River Derwent reversed course is a very difficult question to answer. Further work is needed in the Kirkham Gorge to understand its role in controlling the lake levels in the Vale of Pickering. The likelihood that a lake larger than 45 m O.D. did not exist prior to MIS 5e means that the Kirkham Gorge no longer needs a 70 m barrier to hold back Lake Pickering. Flood fills in Section 3.5 show that the base level of the River Derwent does not need to increase much to impound a lake. Figure 3.5.3 shows that (by using 25 m terraces in the Malton Embayment) a 25 m lake forms with just the base level of the Kirkham Gorge returning to 25 m O.D. Limestone ridge in the Kirkham Gorge of 32 m O.D near High Hutton is sufficient enough to impound a 33 m lake. The presence of ice and permafrost would have lowered permeability and increased the amount of surface run off, which makes it difficult to assess their degree of influence on lake level height. Furthermore, if Lake Humber occupies the Vale of York at a height of 25 m O.D or more, depending on the thickness and extent of the VoYL, Lake Pickering can reach 45 m without the need to incise the base of the Kirkham Gorge. Overall the series of events leading to the creation of the Kirkham Gorge in its present form are very complex and would need significant further investigation.

4.4 What is the duration of Glacial Lake Pickering and did it occur in one or several stage(s)

From the substrata revealed by the CEGB cross-section (Figure 3.1.13), it is evident that many phases of lacustrine conditions prevailed during the history of the Vale of Pickering. For simplicity, the deposits have been divided into the lower lacustrine clays and the upper lacustrine clays overlying Jurassic Kimmeridge and Ampthill clays with intermittent periods of glaciofluvial sands and gravels worked in phases of meltout or during warmer climate phases.

At sites where samples were taken from below lacustrine clay units (Slingsby, Yedingham) dates obtained suggest movement of sediment into the Vale of Pickering basin prior to the onset of glaciation. These sand and lacustrine units can be correlated at Yedingham with the upper layer of deposits in the CEGB borehole cross section (Figure 3.1.31) between 0 and 6 m depth. These deposits are non-continuous and show potentially erosion by the meandering River Derwent or are interpreted as small, shallow lakelets as the lake drained. Below approximately 6 m depth, the lacustrine clay deposits are thicker and more continuous suggesting a larger, longer-lived lake that pre-dates the LGM from OSL dates at Yedingham and High Marishes above lacustrine clay units that correlate with these deeper clay units shown on the CEGB record. This suggests that Lake Pickering existed in several phases and prior to the LGM.

The relationship of these deposits to lake levels is difficult, however, it is likely that the 33 m lake is associated with the upper clay (LF3 at Yedingham) but has undergone post-glacial erosion in places making it appear patchier. The deeper clays related to a deeper and longer-lived lake given their thickness and higher silt content as shown from PSA at High Marishes (Figure 3.2.6) as compared with the clay from Yedingham (Figure 3.2.5) (Smith and Ashley, 1985). This older lake may be correlated with the 45 m lake, but further work is needed.

Overall, the Vale of Pickering remains an excellent sedimentary archive of proglacial lake deposits that extends beyond the LGM and could potentially reveal much about the conditions for the onset of the Devensian glaciation after the end of the Ipswitchian interglacial.

Conclusion

The results of this investigation provide abundant new data on the Quaternary palaeoenvironmental history in the Vale of Pickering. Prior work was unable to satisfactorily provide evidence of lake age or extent, but due to advancements in chronological techniques and from higher resolution DEM data the foundation for subsequent in-depth studies relating to the history of Lake Pickering and its relationship with the surrounding landscape evolution. From this study, several key points have been identified as well as areas for further investigation.

The following key points can be surmised:

1. The lake level heights attributed to Lake Pickering are more complex than initially thought. The lack of evidence for a 70 m O.D. lake as well as the potentially pre-LGM 45 m lake mean that Lake Pickering has existed for much longer than previously thought.

2. The substrata of the Vale of Pickering revealed by historic boreholes and newly acquired boreholes within the scope of this study show that there was more than one iteration of Lake Pickering and at least one of these lakes was present prior to the LGM in North Yorkshire.

3. The relationship between the drainage of Lake Pickering at the outlets at both Filey and through the Kirkham Gorge are complex, but the lake likely utilised both depending on sediment distribution, ice position, and GIA.

4. The size and extent of Lake Humber does have an effect on the size and extent of Lake Pickering, providing the base of the Kirkham Gorge is below 25 m O.D and Lake Humber is higher than 20 m O.D.

5. Ice advanced more than one on the east coast of the Vale of Pickering, blocking the eastern end and altering the drainage of the lake. This is seen in the stacking of the Filey moraine and the cross-cutting moraines north of Filey.

6. The furthest advance was to Wykeham and began at least >30 ka.

7. The application of optically stimulated luminescence dating to sediments within the Vale of Pickering does work and provide robust dates.

8. New mapping produced from LiDAR imagery shows that there is a wealth of previously undescribed landforms that show evidence of multiple iterations of a proglacial lake and a new sequential moraine retreat pattern.

Further work

The results of this study show that there are multiple new lies of enquiry that can help to constrain further the results of this study. These include:

1. The investigation of older till deposits identified by Edwards (1978) and Powell *et al.* (2016) to establish their age and relationship to the formation of the Vale of Pickering basin and previous iterations of Glacial Lake Pickering.

2. Further investigation into the deposits left by the 45 m O.D. lake and further enquiry into evidence of 70 m O.D lake by investigating the role of the Kirkham Gorge and associated overflow points in the Howardian Hills.

3. Further investigation into the rate of isostatic uplift by investigating the downcutting of the River Rye through the former lake plain of the 45 m O.D. lake.

4. Further constraining the retreat and readvance history of the North Sea Lobe at the eastern end of the Vale of Pickering by further investigation in to the extent, type, and morphology of the tills found from Scarborough to Flamborough Head.

Further investigation into the early advance at Wykeham and its relationship with newly aquired data from the North Sea and Dogger Bank (Roberts *et al*., 2018).

6. Investigation into the surrounding periglacial deposits on the North Yorkshire Moors and Yorkshire Wolds, especially the relationship between loess found on the Yorkshire Wolds (Catt *et a*l., 1973) and the Toft/Harrow Hill silts (Franks, 1987).

7. Further work to constrain the relationship between Lake Humber and Lake Pickering.

8. The extent of the Slingby Sands and their relationship to the Sherburn Sands.

9. Research into the timing and extent of other proglacial lakes in the area: e.g.: Lake Eskdale, Lake Harwood Dale, Lake Hackness, Lake Hunmanby and their relationship to the timing of the North Sea Lobe.

10. The extent and timing of the Pickering delta and its relationship to the formation of Newton Dale and with Lake Pickering.

Ultimately, the palaeoenvironmental history of the Vale of Pickering is incomplete. This investigation has shown that there is a wealth of opportunity to develop many research projects in the future. It is a fascinating area and extremely complex, despite the limitations from the high-water table and lack of available, decent sedimentary exposures. It is the intention of this research to go beyond previous attempts to establish a more robust foundation upon which a high-resolution record of the late Quaternary can be completed.

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