A re-interpretation of the physiographic evolution of the southern end of the Vale of York from the mid-Pleistocene to Early Holocene

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Abstract

The recognition and mapping of planar terraces on the York Moraine led to the belief that these were shorelines of the Late Devensian proglacial Lake Humber and the hypothesis that the progressive demise of the lake was recognisable from stillstands. To test this, landform mapping was initiated across the Vale of York and the flanks of the Wolds between Pocklington and Hessle to identify and record planar land surfaces, which had distinct topographic boundaries resulting from erosional and depositional processes.

The results of this study confirmed the early shoreline mapping and identified strandlines, at lower elevations, down to a terminal lake level of 5.0 m OD. Erosional and deposition effects associated with both stands and retreat stages of Lake Humber have deposited a sand mantle up to 2.0 m thick on the southern face of the York moraine from transgressive and regressive shorelines. In addition, two sets of alluvial fans, originating from dry valleys in the Wolds from frost-fractured Chalk formation were recognised. The older set were terraced by shorelines of Lake Humber, in contrast to the younger set, which clearly post-dated Lake Humber. Corroborative evidence for the existence of the shorelines has been provided by photography and LIDAR imagery. An additional objective was to establish a chronology for key mapped landforms based on luminescence dating of sand samples from shoreline deposits and younger fluvial events. To achieve this 18 sand samples were collected and dated.

The main conclusions of the research are that the older periglacial alluvial fans are from an earlier cold glacial period (possibly MIS 8) and that the younger Late Devensian (MIS 2) glaciation retreated north of the York Moraine about 17 ka BP prior to the main phase of impounding Lake Humber. The dating of this event conforms with OSL dating of *c*. 15.9 ka and *c*. 15.2 ka obtained from sand samples on the York Moraine. The existing two stage model of proglacial Lake Humber has been revised and mapped shorelines, of this lake, now define an 8-stage regressive decline model for Lake Humber, from a high level Stage 1 at 42 m OD down to a low level (partly fluvial) Stage 8 at 5.0 m OD. Mapping has also revealed that the decline

sequence appears to have been punctuated by short-lived, modest rises in lake level from oscillations of the blocking North Sea Ice Lobe. It has been established that the southward slope of the Vale of York is testimony to Holocene flooding and not to isostasy.

The landform mapping also indicates that the gravels forming the Crockey Hill 'esker' probably originated as a fan delta from a drainage gap in the York Moraine, south of York, and that terraces in the Aire and Calder valleys are coeval with stands of Lake Humber.

Acknowledgements

I wish to thank Sheffield University and in particular Professor Mark Bateman for accepting me as an older aged Ph.D. student so that I could complete a research project on post-glacial landforms in the Vale of York. The guidance and help provided by Professor Bateman, in allowing me to generate a chronology for my mapped landforms by Optically Stimulated Luminescence, is greatly appreciated, as it has added significantly to my research and produced some totally unexpected results of events not previously identified in the Vale of York. I would also like to thank Rob Ashurst for his good humour while supplying me with endless support in the preparation of my OSL samples in the laboratory. The patience of other post-graduates in providing me with technical assistance is also appreciated. I would finally like to thank Alison Rothwell for her expertise in typing and collating my Thesis.

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PART I

1. INTRODUCTION

Historically the Vale of York, and in particular the York Moraine, first came into prominence when in 69AD Petilius Cerialis, the commander of the Ninth Roman Legion '*Hispana*', marched northwards from Lincoln along a route which is now the A15 (Ottaway, 2001). The Legion crossed the Humber in the region of Brough and eventually, about 71AD, reached a slightly elevated plateau of land (the York Moraine) between the Fosse and Ouse rivers where a fortress was established (Ottaway, 2001). The fortress named *Eboracum* became the City of York (Ottaway, 2001). The chosen site had a number of advantages: it provided a good vantage point looking south down the Vale of York and the steep slopes, on the southwestern side of the fortress, leading down to the River Ouse, could be easily defended. The navigable River Ouse also provided good communication to the Humber Estuary and the North Sea. The Romans had some awareness of geology and even if they were unaware that the hills around *Eboracum* had no natural source for building stone this fact would have soon become apparent to them. Initially this would have been of little consequence, as the defence of the fortress would have consisted of a ditch and rampart topped by a timber palisade. It is probable that the only building inside the fortress to be built of stone, apart from the bath house, would have been the basilica – a great aisled hall with arcades supported by stone columns. One of these columns, found where it fell, was discovered in excavations below York Minster and has since been re-erected near the south door of the Minister (Ottaway, 2001; Plate 1).

Plate 1. A column from the Roman basilica re-erected near York Minster. The red tiles used to repair the column were probably made from laminated clays worked from pits in the Vale of York. (Photograph. W.A.Fairburn.)

The stone quarried to build this column (and presumably the first building stones to ever be used in York) is Millstone Grit, which may have been quarried from outcrops west of Wetherby (Ottaway, 2001) or even possibly from the Skell Valley near the site for Fountains Abbey. When the wall around the fortress was built about the middle of the 2nd century stone was brought to York from quarries in the Permian Magnesian Limestone near Tadcaster: the stone being transported by barge down the River Wharfe and on the rising tide up the River Ouse to York (Knight, *c*.1950). Other natural resources worked by the Romans included gravel and cobbles from the York Moraine for use in road construction (Ottaway, 2001), and laminated clays, possibly excavated from clay pits near Hoggs Pond [SE 581 492] or elsewhere on Hob Moor [SE 584 505] (north of Askam Bog, Fig. 14) at elevations below 10 m above OD, used in the manufacture of tiles for building and decorative purposes.

Petilius Cerialis, at *Eboracum*, would not have been aware that he was looking down a glacial valley the framework of which had been established between the Late Devensian and the early part of the Holocene, or that much of the Vale of York would be below sea level, but for the covering of till and glacial sediments over glacially-eroded bedrock. As pointed out by Parsons (1887), the Vale represents the most recent page in the evolution of the region 'where geology merges into history'.

My personal interest in the York Moraine and the Vale of York commenced during the summer of 2006 while traversing over the York Moraine near York University. Observations at the time included two facts which seemed to be at variance with or were not included in the published literature of the region (e.g. Kendall & Wroot, 1924; British Geological Survey, 1983b). These were:

- the flanks of the moraine were sand draped and essentially stone free below about 30 m above OD. Above this level, gravels overlying till could be observed in crestal locations partly facilitated by road works and other excavations. In places, concentrations of glacial cobbles suggested reworking of the moraine and the accumulation of lag deposits;
- the moraine also appeared to be a tiered feature with laterally continuous horizontal terracing conspicuous at 20 m and 25 m above OD. Crestal regions of the moraine were also planar, in part, between 30 m – 35 m above OD.

In addition to the above features, the base of the moraine provides a distinct and mappable boundary, at about 15 m above OD, against the level surface of the plain of the Vale of York.

These observations, while forming only an early part of an on-going investigation, led to the development of a hypothesis that reworking had taken place on the moraine along shorelines of a regressive lake margin (i.e. pro-glacial Lake Humber). Preliminary mapping of these features resulted in the preparation of a landform based map of part of the Vale of York covered by the North Sheet of the 1:25 000 Ordnance Survey map (Sheet 290) of York, Selby and Tadcaster. This area of about 500 km^2 straddles a central part of the Vale of York and lies within parts of the British Geological Survey

1:50 000 series sheets of Harrogate (1987), York (1983b), Leeds (2003) and Selby (2008) as shown on Figure 1.

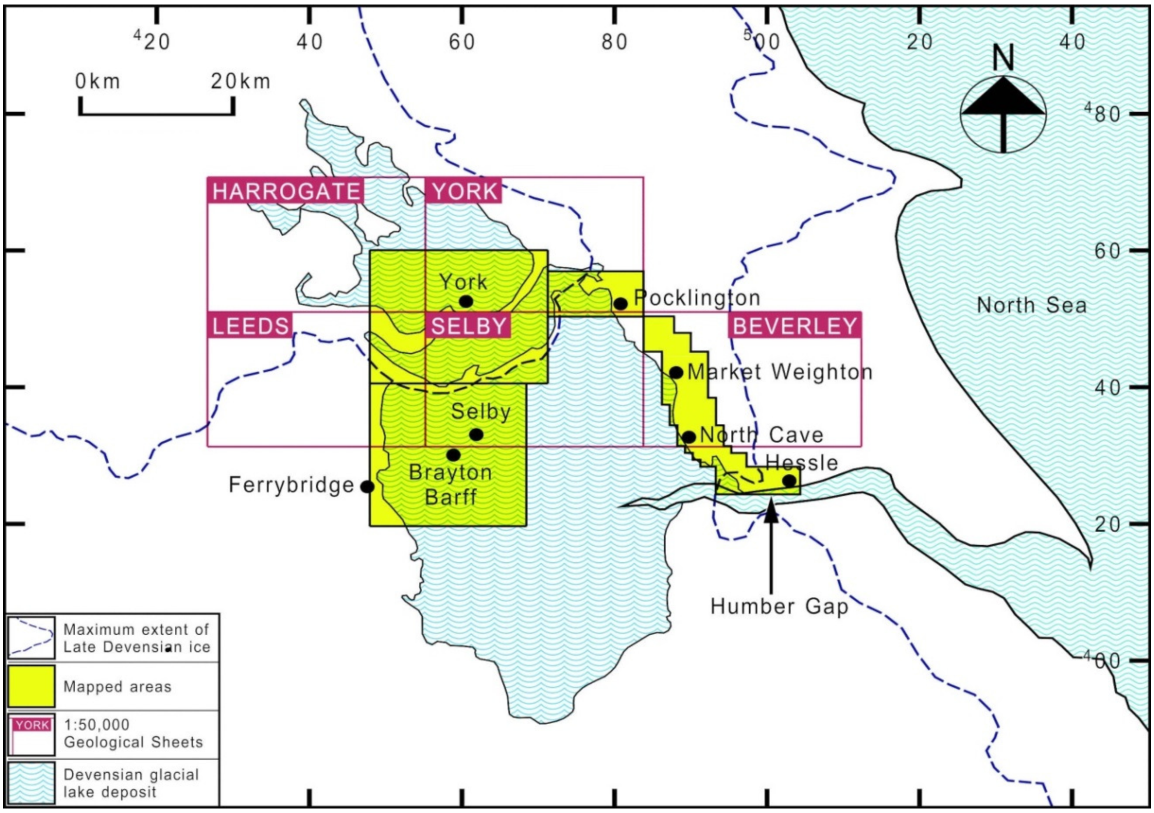


Figure 1. Location plan of landform mapping. The extent of the glacial lake deposit and the limit of Late Devensian ice are adapted from Ford *et al.* (2008).

Publication of the mapping was contained in an article for the Yorkshire Geological Society (Fairburn, 2009). For the purposes of the current text, this early mapping has been modified slightly to conform with more recent mapping (Fig. 14). This landform mapping was later extended to cover an additional 400 km^2 over the South Sheet of the 1:25 000 Ordnance Survey map Sheet 290, an area that covers the southwestern corner of Selby (British Geological Survey, 2008), the southeastern corner of Leeds (British Geological Survey, 2003) and two ridges of Triassic sandstone that extend across the Vale of York at Brayton Barff and west of Snaith (Fig.17).

Additional mapping, also part of the original study of the northern part of the Vale of York, was an investigation of the geometry of the Older Littoral Sand and Gravel described by Gaunt (1981) in the region of Pocklington and Barmby Moor and depicted on the older Selby 1:50 000 geological sheet (British Geological Survey, 1973b). The results of this work, covering just over 100 km^2, were published by the Quaternary

Research Association (Fairburn, 2011), and indicated that the Older Littoral Sand and Gravel was more likely to be of fluvioglacial origin, with the sediment forming distinct fans, as indicated by Ford *et al*. (2008). An updated copy of the mapping is included here in Figure 15. This mapping was later extended to include an area of nearly 150 km^2 along the western slope of the Yorkshire Wolds between Market Weighton and Hessle (Figs 1 and 16). Altogether this work has involved hundreds of hours in the field and hundreds of miles of traversing over every accessible road, track or right-of-way in the mapped area.

The aim of the mapping and the principal purpose of this thesis is to present a re-interpretation of the Late Devensian to early Holocene physiographic evolution of the Vale of York, based on a multi-stage decline system for Lake Humber punctuated by modest rises in lake level. Such a model would have to include an explanation for the planar landforms or terracing on the York Moraine, referred to above, as well as fluvioglacial deposits on the Wolds. This contrasts with the two-stage lake withdrawal model of Gaunt (1976b), which does not recognise any terracing in the Vale of York between his two lake stages (i.e. between high-level and low-level Lake Humber, Fig. 11). Based on the results of research by Hughes (2008), which details the retreat of the Last British-Irish Ice Sheet, an explanation for fluctuations in the level of Lake Humber can most likely be provided by oscillations of the North Sea ice lobe during its northward retreat and not by moraine in the Humber Gap.

2. HISTORICAL REVIEW OF THE QUATERNARY OF THE VALE OF YORK

A brief historical review of research in the Vale of York and related areas, including that preceding acceptance of a glacial origin for the Vale of York, has been included to highlight the work of the early pioneers. Much of the detailed work that was accomplished is still relevant at the present day.

For more that 180 years the nature, origin and depositional history of superficial deposits masking bedrock in the Vale of York and in the Yorkshire Dales has been a matter of debate. Even at the present day, despite increased knowledge gained from detailed mapping by the British Geological Survey (BGS) and a wider based chronology gained by new dating methods, controversy still exists on the recognition, extent, dating and correlation of these deposits.

Prior to 1840, there was a general belief that these deposits (known as diluvium), some of which contained exotic rocks that could not have been locally derived, were the result of the Biblical deluge (Noah's Flood). John Phillips, for instance, who was to become the first Keeper of the Yorkshire Museum, read a paper to the Yorkshire Philosophical Society in 1826 based on his detailed observations on the distribution of Shap Granite (Phillips, 1827). Phillips was able to trace boulders of the granite from Shap Fell across the Pennines by way of Brough and Stainmore through Teesdale to Barnard Castle and Darlington. Boulders were also located in the Vale of York between Thirsk and Pocklington, in gravel pits around York and along the Yorkshire coast between Redcar, Scarborough and Flanborough Head. Boulders of Shap Granite, which can now be seen in the gardens of the Yorkshire Museum, were actually extracted from excavations during the construction of the York Railway Station in the 1870's (Plate 2).

Plate 2. A boulder of Shap Granite in the Museum Gardens at York extracted from excavations at the York Railway Station in the 1870's. (Photograph. W.A.Fairbutn).

Based on his observations, Phillips deduced that the diluvial deluge had moved in an easterly or southeasterly direction. Such a direction differed from the opinions of William Smith who favoured a westerly direction (a personal communication to Phillips) and William Buckland (1823) who preferred a southerly direction.

Buckland, who was also a strong advocate of the diluvial theory, believed that the bones found in the Kirkdale Cave, while being the remains of the food of a pack of hyenas, were contained in mud derived from the great flood (Boylan, 1977): a conclusion he believed was 'in full agreement with the words of the Book of Genesis' (Osborne, 1999).

The glacial theory, that superseded the diluvial hypothesis, was largely due to Agassiz and Charpentier and their work in tracing the ancient limits of Alpine glaciers (Melmore, 1935). Buckland, who was perhaps the first geologist to use the term diluvium, came to accept there had been land-based glaciers in Scotland following a visit to Switzerland with Agassiz in 1838. Agassiz then came to Scotland in 1840, with Buckland, and during a tour of the highlands, which included Glen Roy, recognised abundant evidence of glaciation. In Glen Roy, Agassiz explained the Parallel Roads as the imprint of a former ice-dammed lake (Lowe *et al*., 2008). Armed with this knowledge Buckland was able to present his new beliefs to the Geological Society at three meetings of the Society in late 1840. He was supported at these meetings with presentations by Agassiz and Lyell (Boylan, 1981).

Despite the pioneer work of Buckland and Agassiz, the starting point of modern glacial geology in England has been attributed, by Kendall & Wroot (1924), to Henry Carvill Lewis and the publication of his text 'The Glacial Geology of Great Britain and Ireland' in 1894. In this volume Lewis (1894, Map II) shows lines of ice flow from the Yorkshire Dales (apart from Swaledale) extending into the Vale of York and along the coast of Holderness: the moraine boundary of his ice sheet closely resembling present-day mapping. Lewis (1894, Map IV) also envisaged a great North Sea glacier blocking the mouth of the Humber, The Wash and other drainage exits from Norfolk to trap a fresh-water extra-morainic lake referred to by Lewis (1894, p. 62) as Lake Humber (Fig. 2).

Figure 2. Map showing the extra-morainic or pro-glacial lake mapped by Henry Carvill Lewis (1894) that lay trapped between the great wall of the North Sea glacier and the southern edge of an onshore ice sheet that stretched across northern England into Wales. This lake was considered to have risen to an elevation of 400 ft (122 m) above OD before overflowing into the Bristol Channel and the River Thames.

He estimated that the water level in this lake could have risen to the 400 foot contour (122 m), an elevation that would have included the Vale of Pickering and caused the great lake to overflow into the Vale of Evesham, the Severn Valley and along the Thames Valley to London. There is, however, little evidence available for the extent of such a lake as the highest possible lake level in the Vale of York (without allowing for isostatic recovery) would be the 52 m strandline at the western exit of the Market Weighton spillway recognised by Penny (1974). In the Lower Tees Basin (Murton &

Murton, 2011) a massive clay unit cropping out at *c*. 90 m above OD (Radge, 1939) has not been established as glaciolacustrine in origin and hence indicative of a high-level phase of Lake Tees (Murton & Murton, 2011). More recent mapping of the extent of Lake Humber and its continuation to the south as Lake Fenland, by Straw (1979) and Clark *et al.* (2004) after Gaunt (1976a), shows more restricted bodies of water trapped by the wall of the North Sea ice.

3. THE QUATERNARY GLACIAL RECORD

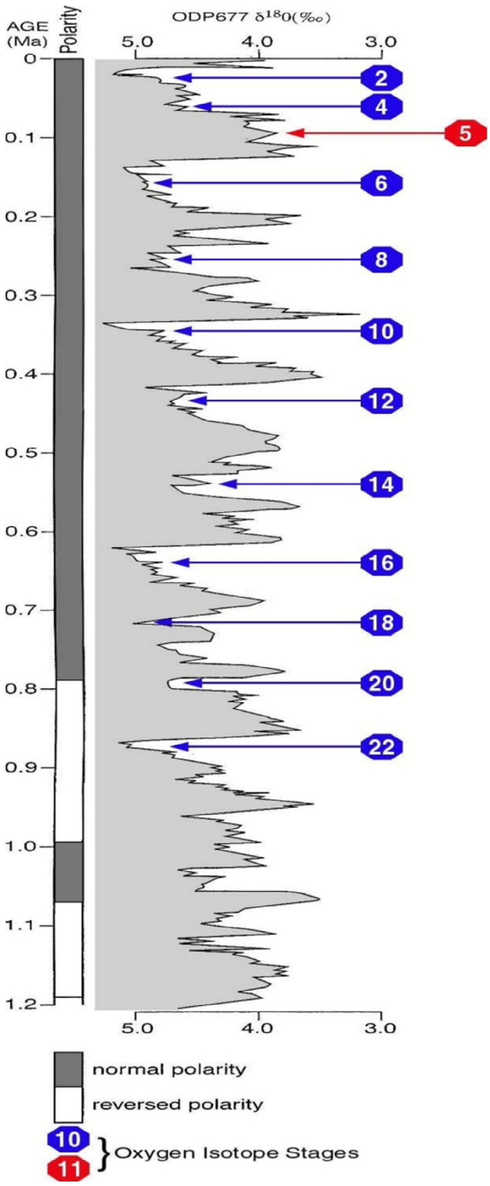
3.1. Summary

It is now known that the 'Great Ice Age', now commonly referred to as the Quaternary Ice Age, commenced about 2.58 Ma BP (Gibbard & Head, 2010). Within this period marked climatic changes, recorded from the $^{18}0/^{16}0$ ratios found in benthic foraminifera, appear to have been caused by long-term cyclical variations in solar energy associated with the Earth's orbital periodicities (Lowe & Walker, 1997; Merritt *et al*., 2003). Based on these cycles, at roughly 100 ka intervals, there could have been 8 to 10 glacial events (or stadials) separated by interglacials (or interstadials) as illustrated in Figure 3 between marine isotope stage 22 (MIS 22) and MIS 2 (Funnell, 1997; Merritt *et al*., 2003, fig. 29). Terrestrial equivalents in the Quaternary succession of the UK, for most of these glacials, is however lacking and reliable evidence appears to be restricted to the existence of only three glaciations. These are the Anglian glaciation (MIS 12), the Saalian (or Wolstonian) glaciation (MIS 6-10) and the Late Devensian glaciation (MIS 2).

The Anglian glaciation was the most extensive of the Pleistocene (Ehlers & Gibbard, 1991) and during its culmination a continental glacier covered much of the British Isles. It is known to have extended as far south as the outskirts of the future site of London and established the present lower course of the River Thames (Funnell, 1997). This glaciation is thought to have occurred between 480 ka – 430 ka BP (Catt, 2007).

The concept of a Middle Pleistocene, pre-Ipswichian/post-Anglian glaciation, given the name Wolstonian, was established from a stratotype in north Warwickshire, even though the glacial sequence at this site has not revealed any interglacial Hoxnian deposits beneath it (Shotton, 1973). This glaciation was thought to cover a cycle from cold protoglacial, through full glacial to a period of ice retreat (Shotton, 1973) but could have occurred at any time between MIS 10 and MIS 6 (350 ka – 128 ka BP, Catt, 2007). The status of the Wolstonian glaciation, in the Midlands, remains controversial as outlined in a review by Rice & Douglas (1991) who suggest that glacial beds, at the Wolston type site, cannot be related directly to dated interglacial deposits and that fluvial sands and gravels, underlying the Wolstonian glacial deposits, can be traced to East Anglia where they underlie deposits of Anglian age.

In a study of late Middle Pleistocene fluvial terrace sequences of the lower Trent system in Lincolnshire, White *et al*. (2010) identified MIS 7 interglacial deposits overlying till, which was consequently assigned to a glacial period no later than MIS 8. This glacial episode, which was mapped down eastern England as far south as Peterborough into The Wash basin was correlated by White *et al*. (2010) with widespread glacial deposits in Europe classified as Saalian. Support for an MIS 8 or Saalian ice sheet, extending into central Lincolnshire and northeast Norfolk, is presented by Straw (2011), who regards usage of the term Wolstonian for these deposits as anomalous, and by Westaway (2010) and Langford (2012).



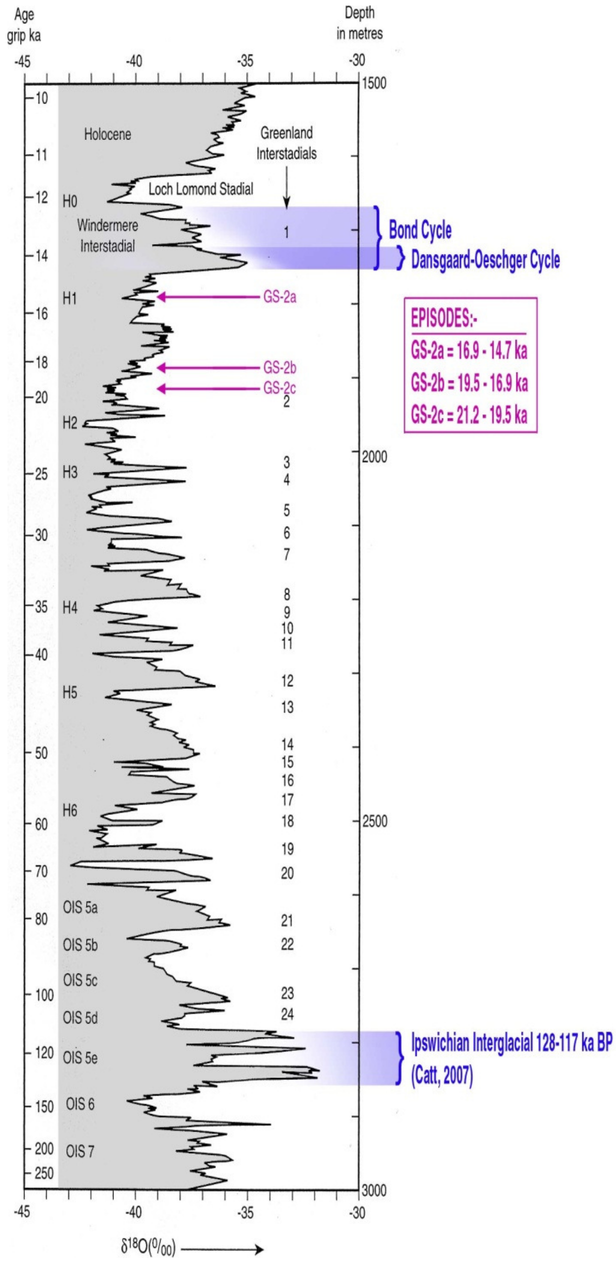


Figure 3. Oxygen isotope and geomagnetic polarity record (ODP 677) after Shackleton *et al*. (1990) and Bowen *et al*. (1999). The glacial events (stadials) are in blue with the Ipswichian interglacial MIS 5 in red.

Figure 4. Greenland (GRIP Summit) oxygen isotope record after Dansgaard *et al*. (1993). Stadial events GS-2a, GS-2b and GS-2c from Björck *et al*. (1998), Heinrich events H0 – H6 after Bond *et al*. (1993), Dansgaard-Oeschger and Bond cycles from Merritt *et al*., 2003 (fig. 32).

The post-Anglian period includes six interglacials, the foremost among these was the Ipswichian interglacial, which peaked about 120 ka BP during MIS stage 5e (Catt, 2007). The Ipswichian is chiefly remarkable for the occurrence of hippopotamus and other tropical or subtropical mammals in Britain, some of which have been identified in the raised beach at Sewerby (Lamplugh, 1891; Catt & Penny, 1966). These remains suggest that temperatures during the Ipswichian were significantly higher than the present day with summer temperatures at least 2˚C higher (Funnell, 1997). Sea levels were some 7m higher than at present (Plimer, 2009).

For the climatic record since the Ipswichian, new high resolution evidence derived from cores through the Greenland ice sheet has shown that dramatic climatic changes have occurred both on the millennial scale and over just a few years during the Devensian Stage. During this period 24 Greenland interstadials have been identified (Fig. 4), which commenced with abrupt warming and typically cooled over only a few thousand years to end with a cold event (Broecker, 1994). These periods, known as Dansgaard-Oeschger (D-O) cycles are sometimes grouped into longer Bond cycles (Fig. 4; Bond *et al*., 1993). Frequently the Bond cycles culminate in Heinrich events, which represent the discharge of ice-rafted debris into ocean sediments from the North American Laurentide ice sheet. Such mass wastage, by iceberg calving of the Laurentide ice sheet could have played a pivotal role in driving world climate variability (Merritt *et al*., 2003; Broecker, 1994). All these events illustrate the dynamic nature of the ice sheets, which fluctuated with time perhaps due to cyclical variations in solar energy or even glacially imposed climate changes.

According to Chiverrell & Thomas (2010), global cooling after MIS 5 (*c*. 75 ka, Fig. 3), marks the beginning of the MIS 4-2 glaciation with cold stage climatic conditions mainly persisting from 60 ka until the termination of MIS 2 (*c*. 15 ka, Fig. 3). A similar climatic break was also noted by Funnell (1997, fig. 4.5) at *c*. 75 ka from a pollen record in southeast France. As a result of this climatic cooling, during the Late Devensian in MIS 2 (from 26 ka to 10 ka BP, Bowen, 1999) ice sheets developed world wide e.g. the Laurentide Ice Sheet of North America (Mayewski *et al*., 1981), the Fennoscandian Ice Sheet (Sejrup *et al*., 1994) and the last British-Irish Ice Sheet (Clark *et al*., 2012).

3.2. The Last British-Irish Ice Sheet

It has been estimated by Clark *et al*. (2012), that the Last British Ice Sheet (BIIS) had an areal extent of *c*. 840 000 km^2 with a probable volume of just below 800 000 km^3. This mass of ice completely covered Ireland and Scotland, extensive areas on the British and Irish continental shelves and North Sea, and at some time was confluent with the Fennoscandian Ice Sheet (Fig. 5; Sejrup *et al*., 1994). The maximum extent and dynamics of this ice sheet in the Vale of York and related areas of Holderness and the Humber Gap provide the basis for the currently accepted model for the evolution and topographic expression of the Vale of York. Much of this is derived from published and unpublished work of G.D.Gaunt (e.g. Gaunt, 1976b). They also provide the basis for a different evolutionary model, particularly if glacial events can be related to the climatic record provided by the GRIP Greenland ice-core (Björck *et al*., 1998; Fig. 4), as indicated by Bateman *et al*. (2011).

Clark *et al.* (2012) have explained that the maximum extent of the BIIS was not synchronous in all areas: to the northeast of Scotland and Ireland the ice reached its maximum extent at the shelf margin around 27 ka BP, while the southern Welsh limit and the Scilly Isles were not reached until 23 ka BP and after *c*. 25 ka BP respectively. Ice advance into the Cheshire Plain has been dated at 25 ka – 21 ka BP (Clark *et al*., 2012, fig. 15). This chronology for the LGM is comparable to dating by Chiverrell & Thomas (2010) who assigned the build-up of the last BIIS to after 35 ka – 32 ka BP with a glacial maximum between 27 ka and 21 ka BP (Fig. 4).

Dating of the Last Glacial Maximum at the south eastern margin of the BIIS is still debated: Clark *et al*. (2012) have suggested that ice did not reach the east coast of England until after 22 ka BP (based on the age of the Dimlington Silts, Penny *et al*., 1969). On the Yorkshire Wolds, dating of loess at the base of the Skipsea Till implies that ice did not reach its limit until after 17.5 ± 1.33 ka BP (Wintle & Catt, 1985), a date that agrees with an OSL date of 17.01 ± 1.33 ka from sand infilling a subglacial stream within the Skipsea Till at Barmston (Hartman, 2011). In the Vale of York Clark *et al*. (2012) have suggested an ice advance after 23.3 ka BP (range 24.8 ka – 21.8 ka BP) for a transient ice stream that formed part of a late surge down the east coast of England but had retreated to the north by 20.5 ka BP (range 21.7 ka – 19.3 ka BP). This age for ice in the Vale of York is based on the dating of loess beneath a diamicton near Ferrybridge (Bateman *et al*., 2008). (It should be noted, however, that there is no other reliable evidence for Late Devensian tills south of the Escrick Moraine.) These dates for a glacial event in the Vale of York are in general agreement with Livingstone *et al.* (2012) who suggested a Stage I Main Glaciation through the Stainmore Gap dated to *c*. 29 ka – 23 ka BP. Cessation of ice flow in the Stainmore Gap and the shutdown of the Vale of York ice lobe occurred during Stage II of their Main Glaciation, an event dated before their Stage IV Blackhall Wood Re-advance (*c*. 21 ka – 19.5 ka BP). Teasdale & Hughes (1999) have also suggested a short-lived southward ice surge, between 22 ka and 18 ka BP, down the east coast from the Firth of Forth and the Cheviot Hills to northern Norfolk.

A further indication of the age of the Vale of York glacier could be provided if it advanced into the Vale of York during a comparable time frame to that of the Blackhall

Till in the Tyne Valley. On the Durham coast, the Blackhall Till, capped in places by an erosional lag-derived boulder pavement, is overlain by the Horden Till (Davies *et al*., 2013a), which has been equated with the Skipsea Till by Davies & Roberts (2013).

Clark *et al*. (2012) have explained that the maximum extent attained by ice in north and west England was curtailed at the edge of the continental shelf, where water was too deep for grounded ice to advance, at a time when global sea levels were still lowering and there was ice expansion in the southern part of the North Sea basin.

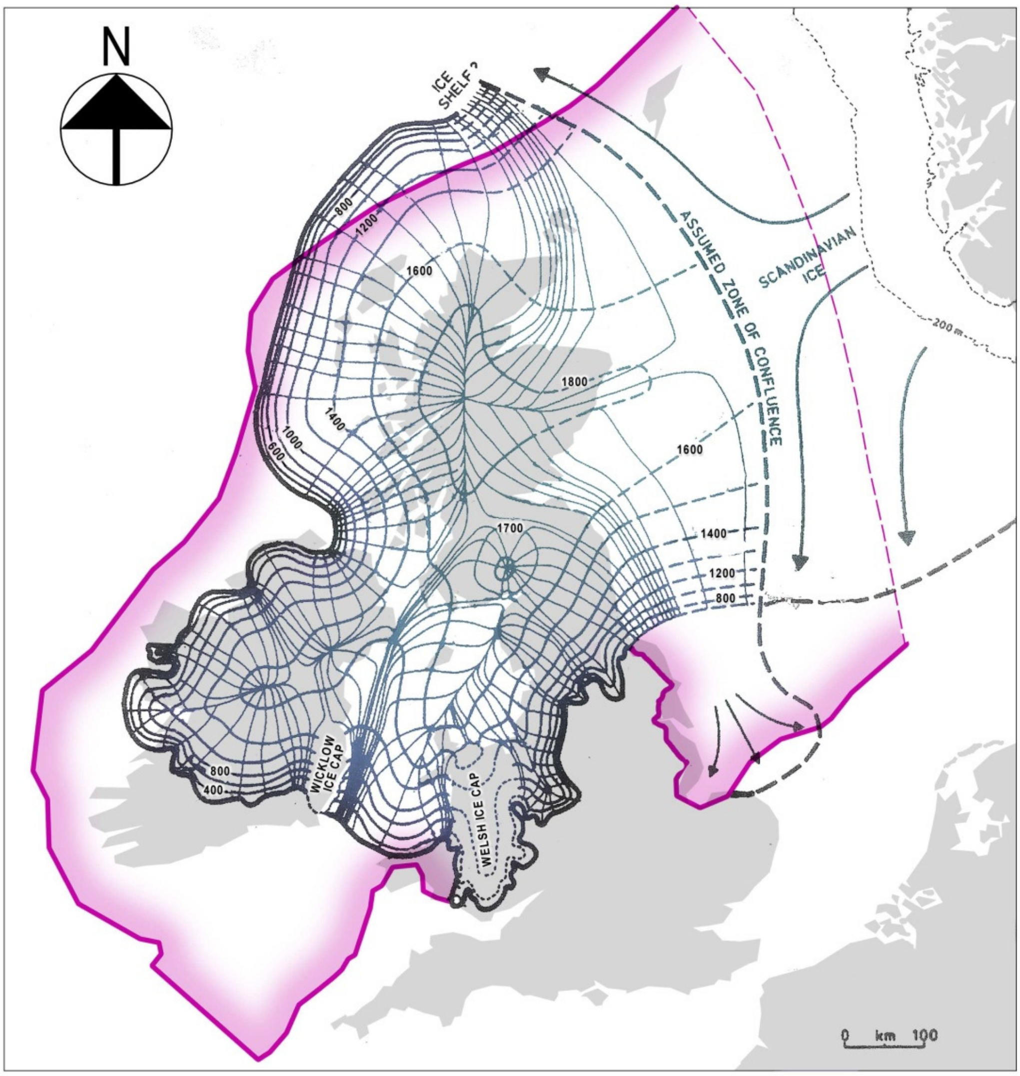


Figure 5. Modelled British-Irish Ice Sheet after Funnell (1997) showing surface topography (in metres) and flow lines. The maximum extent of the BIIS (red line) at the LGM is from Clark *et al*. (2012). Alternative models, with different ice thicknesses have been illustrated by Hughes (2008).

The term Dimlington Stadial was proposed by Rose (1985) as a climatostratigraphic unit to cover the phase of maximum expansion of the BIIS during the Late Devensian between 26 ka and 13 ka ^{14}C yrs BP. The British type site for glacial deposits within this episode is at Dimlington in Holderness. In terms of calendar years the ice advance into Holderness was between *c*. 23 ka – 15 ka BP (Catt, 2007), which is equivalent to GS-5 to GS-2a in the Greenland ice core chronology (Fig. 4; McMillan *et al*., 2011).

The pattern of retreat of the BIIS between 23 ka BP and 15 ka BP has been modelled in detail by Hughes (2008) using remote sensing to record glacial landforms that include moraines, meltwater channels, drumlins, eskers and ice-dammed lakes (Fig 6).

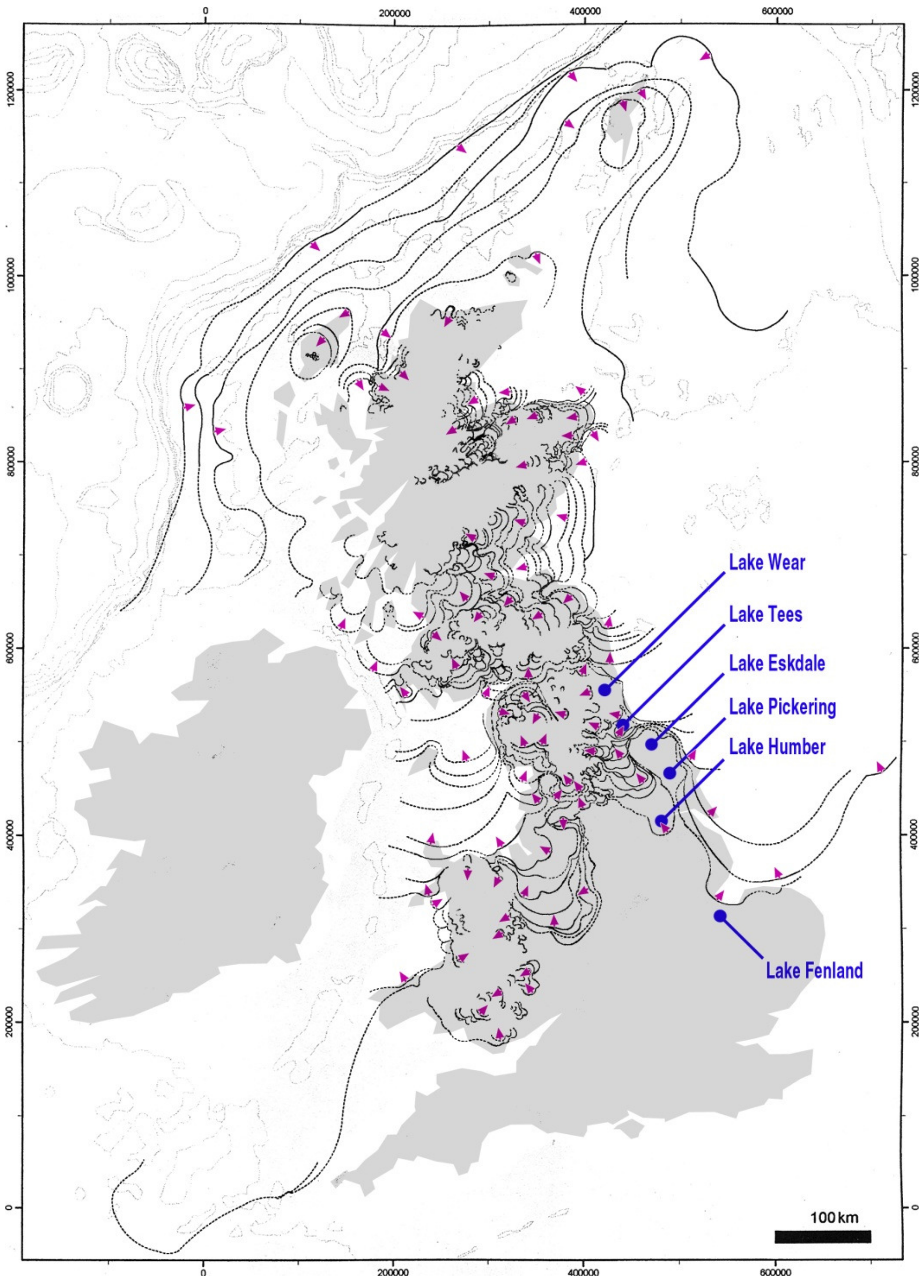


Figure 6 Reconstructed pattern of ice retreat based on synthesis of five independent lines of evidence (moraines, eskers, meltwater channels, flowset retreats and lake dam positions). Arrows show direction of retreat towards ice centres. Approximate locations of glacial lakes Fenland, Humber, Tees and Wear are shown between retreating onshore ice and the North Sea ice lobe (from Hughes, 2008).

All five landform sets largely describe the same pattern of retreat, which Hughes (2008) describes as robust, as the ice retreats back to multiple regional ice centres. The most prominent of these retreat patterns, shown on Figure 6, include:

- migration of Vale of York ice and some east coast ice back to the Lake District and the Cheviots;
- contraction of ice sheets to central areas in Wales and Southern Uplands of Scotland;
- the northerly retreat of ice from the Cheshire Plain into the Irish Sea;
- a marked westerly movement of ice back to a north-south central divide in the Highlands of Scotland;
- the localisation of ice to high ground in the Hebrides and Shetlands.

Initial ice retreat was not always to high ground but did ultimately retreat to higher elevations.

Substantial downwastage of the Last British-Irish Ice Sheet and the emergence of higher summits in NW Scotland has been dated between *c*. 16 ka and *c*. 15 ka BP by Fabel *et al*. (2012) from ^{10}Be dating prior to rapid warming at the termination of the Dimlington Stadial (*c*. 14.7 ka).

Of importance, in the present project, is the interaction during the retreat phase between the North Sea ice lobe, onshore ice and the formation of ice-dammed lakes along the east coast of England. The location of several of these lakes e.g. Lake Wear (Smith, 1994), Lake Tees (Agar, 1954) and Lake Eskdale (Kendall, 1902) are testament to the persistent influence of offshore ice after deglaciation of inland areas (Hughes, 2008). The effect is illustrated in Figure 6 by the retreat of the Vale of York glacier into Wensleydale accompanied by the uncoupling of offshore ice along the eastern coast of England north of The Wash which resulted in deglaciation of the Tees estuary before *c*. 18.3 ka BP[1] (Plater *et al*., 2000; see also 15.19 ± 1.87 cal. ^{14}C ka BP, Jones, 1976). In a study of Late Pleistocene lakes, Murton & Murton (2011) describe the impounding by North Sea Basin ice of pro-glacial lakes in the lower Wear and Tees valleys, North York Moors (Lake Eskdale), Vale of Pickering, Vale of York and the Fens (Fig. 9). The damming effect of North Sea ice was also recorded by Hughes & Teasdale (1999, after

[1] Note that the error factor associated with this age is in excess of 10 ka (see also section 4.2.).

Smith, 1994) in the formation of glacial Lake Wear between the River Wear and the River Tyne.

The important implications from this summary are:

- the glacial lakes often resulted from the westward retreat of inland ice before significant retreat of offshore ice in the North Sea basin;
- the Tees estuary was deglaciated before *c*. 18.3 ka BP prior to the deposition of the Withernsea Till (Livingstone *et al*., 2012).

3.3. The Vale of York Glacial Record

Glacial deposits of both primary and secondary origin, along with their resultant landforms, provide abundant evidence of two terrestrial glaciations in the Vale of York. That evidence for a previous glacial episode still exists has resulted from the deeper penetration of that glaciation into the Vale of York than the more recent glaciation, which has not totally obliterated residuals of the former event. Evidence for the younger glaciation is of course extensive and includes the York and Escrick terminal moraines and the terraced till that lies between these two moraines.

3.3.1. Pre-Ipswichian

The older glaciation has been mainly recognised from the high-altitude drifts including boulder clay, mapped by Raistrick (1926) in the Pennines adjoining Swaledale and Wensleydale. These can be correlated with the 'Older Drift' shown on the one-inch geological map of Leeds (Edwards *et al*., 1950, plate III). Till from this 'Older Drift' has been termed the 'Pennine Boulder Clay' by Gaunt (1976b) based on the pebbles in the tills, which are dominantly of Carboniferous sandstone. Gaunt (1981, fig. 2) has inferred a southerly or south-easterly ice movement to deposit these tills. Whilst no dating has been undertaken on the 'Pennine Boulder Clay', Gaunt (1976b) has suggested a Wolstonian or Anglian age for these deposits. Remnants of an older till, mapped on the Wolds scarp by the British Geological Survey (2008) on flat-lying benches above about 80 m, are dated as possibly of Anglian age by Ford *et al*. (2008). These deposits, near Pocklington, are not included in mapping shown on Figure 15. Within the Vale of York the only occurrences of 'Pennine Boulder Clay' are present on the crestal regions of the Triassic sandstone ridges that extend through Brayton Barff and on the Snaith Ridge between Kellington and Thornfield House (Fig. 8).

Associated with the older till at both Brayton Barff and the Snaith Ridge are gravel deposits, with similar pebble lithologies, exhibiting both erosional and gradational contacts to the till (Figs 7 and 8). These gravels appear to represent fluvial events synchronous with retreat of the ice sheet as they indicate marginal, sub-glacial and supra-glacial drainage. On the Snaith Ridge, Gaunt 1976b recognised a younger set of gravels with depositional features suggesting a vigorous fluvial regime different from the older gravels and Late Devensian shoreline gravels in the Vale of York. These gravels, termed the 'Glacio-Fluvial Sand and Gravel' by Gaunt (1976b), have an erosional basal contact with the older gravels and tills and a different pebble lithology. These two older fluvial deposits differ both in terms of their sedimentary fabric and their lithology: the older set has an 'east Pennine suite' of pebbles (Gaunt, 1976b) while the other shows strong deformational characteristics and is composed of material almost entirely derived from the Coal Measures. This differentiation has allowed correlations to be made with more isolated gravel outcrops, which appear anomalous to fluvial or lacustrine events in the Vale of York (e.g. the cross-bedded gravels in Prescott's Pit, Fig 22, Plate 25).

A further feature, on the Snaith Ridge, is the deeply incised channel deposits recognised by Gaunt (1976b). Having a sub-glacial origin their eastern limit may define a margin to the penultimate ice sheet.

3.3.1.1. Brayton Barff

On Brayton Barff (Fig. 7) two layers of till are present: a lower rather sandy till that has a pronounced bedded appearance including pebbles concentrated into distinct layers and an upper clay till with a less pronounced bedded appearance. These tills overlie laminated clay with a gradational lower contact, marked by a progressive loss of laminae, over fine grained pale red-brown sand on an undulating surface of Sherwood Sandstone (Gaunt, 1976b). The most abundant pebbles, in both tills, are of Carboniferous sandstone with subsidiary Carboniferous and Permian limestones and cherts (Gaunt, 1976b). This assemblage is quite distinct from that in the York and Escrick moraines, which have frequent occurrences of Shap Granite, Brockram and other Eden Valley and southern Lake District erratics as well as Carboniferous sandstones and limestones and Whin Sill (Catt, 2007).

At the summit of Brayton Barff (Fig. 7), the till is overlain by unbedded and poorly sorted gravel referred to by Gaunt (1976b) as 'Older Glacial Sand and Gravel' or 'Glacial Sand and Gravel' (Gaunt, 1994). These gravels have a gradational contact with the underlying till except where it occupies a channel, some 5.5 m deep, incised into the till. This channel, which has a northwest – southeast orientation, extends into the sands below the laminated clay (Fig. 7).

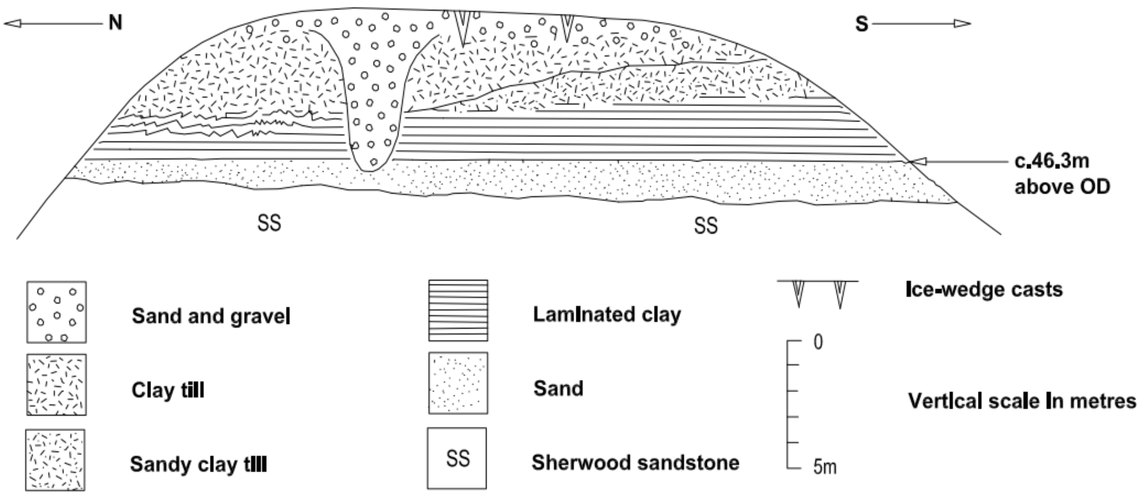


Figure 7. Section through the glacial deposits on Brayton Barff. From Gaunt, 1994, fig. 37.

The pebble content of these gravels is identical with the 'east Pennine suite' of erratics in the till. Similar gravels, visible as float, have also been recorded near the summit of Hambleton Hough (Gaunt, 1976b). As with the till or 'Pennine Boulder Clay' the gravels are considered by Gaunt (1994) to be pre-Ipswichian in age.

In contrast to the Snaith Ridge, the till and gravels on Brayton Barff are not concealed by reworked younger gravels, as they lie above the Lake Humber 40 m Devensian Lake Humber shoreline. However erosion of both the till and gravel at the 40 m above OD level have contributed to the 'east Pennine suite' of gravels on the 15 terrace surrounding Brayton Barff and Hambleton Hough (see section 7.9.4.1.) as well as localised accumulations of gravel on the sides of Brayton Barff at both the 33 Metre and 20 Metre Surfaces (see sections 7.4.3. and 7.8.3.1.).

3.3.1.2. The Snaith Ridge

Till and Glacial Sand and Gravel - On the Snaith Ridge up to 2.5 m of sandy till was exposed in a pipeline trench between Kellington and Thornfield House (Fig 8; see also

Gaunt, 1994, fig. 38). The exposure overlies Sherwood Sandstone except near the southern end of the temporary exposure where the till has a gravelly base. Towards Thornfield House the till has a lateral gradational contact with a sequence of clayey gravel and sandy gravel that comprises (as on Brayton Barff) the 'Older Glacial Sand and Gravel' (Gaunt, 1976b). Both the till and associated gravels contain sub-angular to sub-rounded pebbles and cobbles of mainly Carboniferous sandstones with a few of Carboniferous and Permian limestones. This assemblage, like that from Baryton Barff, is identified as the 'east Pennine suite' (Gaunt, 1976b). Due to the absence of permanent outcrops, erratics from this gravel suite are restricted to gravels that have been reworked into the 15 Metre and 20 Metre Surfaces (see Sections 7.9.4.2. and 7.8.3.2.) near Thornfield House which are referred to as 'Glacial Sand and Gravel' (British Geological Survey, 1971) or as 'Lacustrine Shoreface and Beach Deposits' (British Geological Surevy, 1998). It is also likely that gravel derived from the 'Older Glacial Sand and Gravel' and associated till have been incorporated into the 'Younger Pennine Glacial Sand and Gravel' overlying Sherwood Sandstone in the sand pits at Hensall, Pollington and Great Heck (Plates 38 and 39, Fig. 23).

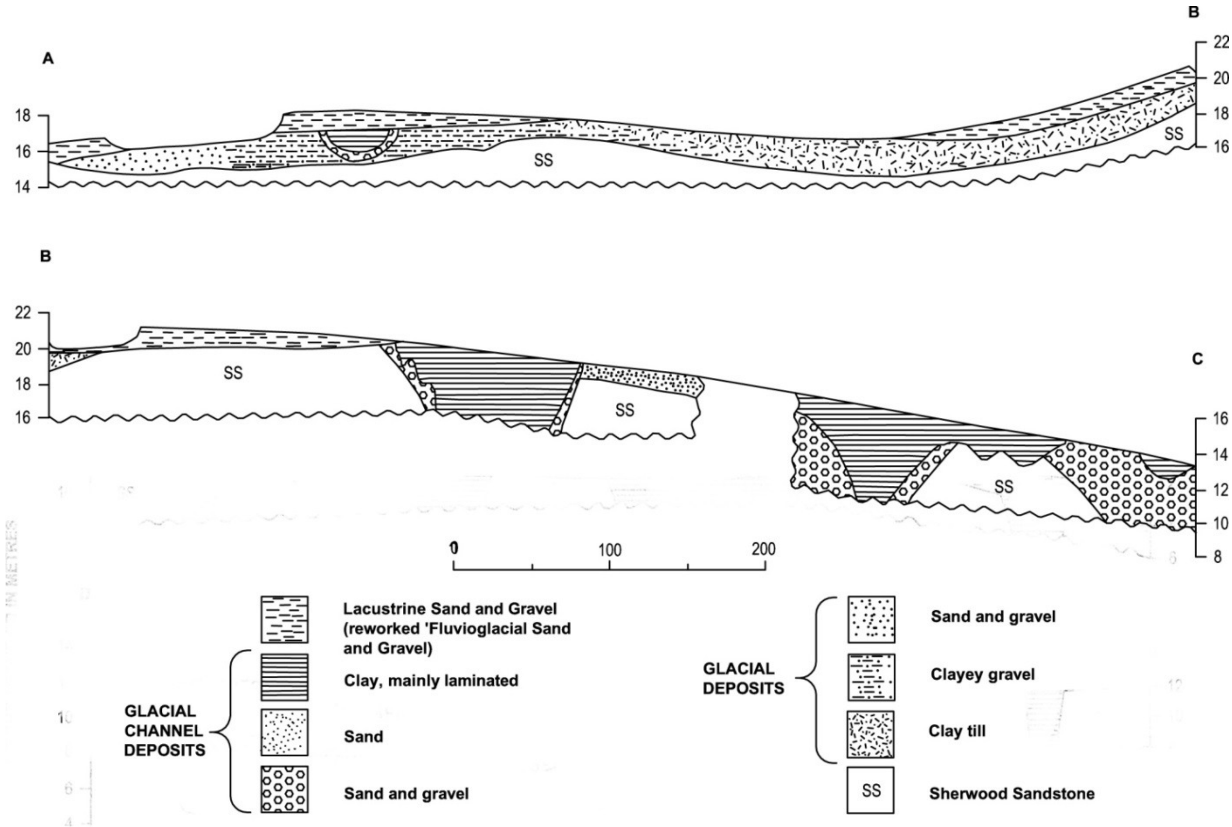


Figure 8. Glacial deposits exposed in a pipeline trench between Kellington and Thornfield House on the Snaith Ridge. From Gaunt, 1994, fig. 38.

Glacial Channel Deposits - Closely adjacent to the till in the northern part of the Kellington trench (Fig. 8) are a number of glacial channels incised into the Sherwood

Sandstone or into gravels associated with the till. These channels are lined by sand and gravel containing sporadic pebbles and a few cobbles and boulders of Carboniferous sandstone and Permian limestone and are infilled mainly by laminated clay (Gaunt, 1976b). In a discussion on the origin of the channels, Gaunt (1976b) has concluded that they originated from sub-glacial drainage, below pre-Devensian ice (i.e. Anglian or Wolstonian), with sufficient hydrostatic pressure to erode into the soft Sherwood Sandstone.

It is now known that the Kellington channel belt forms part of a much more extensive series of closed channels in the Doncaster region that are orientated northwest – southeast or west-east and reached depths of up to 58m below OD (Gaunt, 1976b and Gaunt, 1994). Some of these channels may have originated from flow through gaps in the Permian escarpment as did the Kellington channels from the Aire gap. If these channels are sub-glacial features then their eastern terminations could define a pre-Devensian ice margin east of Doncaster and south of Thorne.

Glacio-fluvial Sand and Gravel - The 'Glacio-fluvial Sand and Gravel' on the Snaith Ridge, was believed by Gaunt (1976b) to possess characteristics, which suggest transport and deposition in a glacial environment. It was mapped as 'Glacial Sand and Gravel' on the Goole 1: 50 000 sheet (British Geological Survey, 1971) and as 'Glaciofluvial Deposits' on the Wakefield 1: 50 000 sheet (British Geological Survey, 1998). It has also been reworked into the 'Lacustrine Sand and Gravel' shown on the Kellington – Thornhill House trench section (Gaunt, 1994; Figs 8 and 38). My landform mapping (Fig. 17) includes the 'Lacustrine Sand and Gravel' as part of the shoreline deposits on the 15 Metre and 20 Metre Surfaces. The deposit, although poorly exposed, apart from outcrops in some old quarries (not now accessible), is mainly underlain by Sherwood Sandstone except where it transgresses, with a marked erosional unconformity, over the 'Pennine Boulder Clay', the 'Older Glacial Sand and Gravel' or the 'Glacial Channel Deposits' (Gaunt, 1976b, 1994). Based on rare exposures the 'Glacio-fluvial Sand and Gravel' is well-bedded or cross-bedded and exhibits isoclinal convolutions, large-scale cut and fill channel structures as well as an appreciable number of cobbles and some boulders (Gaunt, 1976b, 1994). All these features imply torrential fluvial deposition with some post-depositional slumping.

Gaunt (1976b) considers that the Glacio-fluvial Sand and Gravel lies under the lower periglacial surface, which implies it pre-dates the Late Devensian maximum glacial phase. However it is younger than any pre-Devensian glaciation as it is underlain by the 'Pennine Boulder Clay'. Also the pebble suite in the 'Glacio-fluvial Sand and Gravel' precludes derivation from the ice that deposited the 'Pennine Boulder Clay', as the pebbles are predominantly of Carboniferous sandstone, so a derivation from Upper Carboniferous outcrops, in the Pennines, by eastward flowing meltwater through the Aire gap in the Permian escarpment is likely.

The important conclusion that can be reached, from the above account, is that gravel deposits with steeply dipping foresets as at Great Heck (Fig. 23) and Prescott's Pit (Plate 25), which are anomalous in terms of a multi-phase withdrawal model for Lake Humber, resulted from a pre-Late Devensian torrential flooding event possibly equivalent to flooding in the Wolds that deposited the Older Alluvial Fans. These gravels are quite distinct from the mounds of sand and gravel that form the 15 Metre Surface at Pollington and Hensall (Plates 28 and 29).

3.3.2. Late Devensian

It has now been accepted that during the Late Devensian (Dimlington Stadial) ice sourced from the Lake District and Southern Scotland flowed up the Vale of Eden and into eastern England by way of the Stainmore Gap and the Tees Valley (Lewis, 1894; Raistrick, 1934; Catt, 1991) where it was joined by local ice from upper Teesdale derived from the Cross Fell region (Mitchell, 2007). A path for this ice flow has been well established by the distribution of distinctive erratics such as Shap Granite, Whin Sill, and Permian Brockram from the Vale of Eden. The location of these distinctive erratics have been described by several geologists since the time of John Phillips (1827), including Goodchild (1875) in the Vale of Eden and the western parts of the Stainmore, and Raistrick (1926) in the lower Tees valley and the Vale of York and Kendall & Wroot (1924) on the east coast.

Raistrick (1926) considered that the Teesdale glacier initially escaped to the North Sea basin but was later deflected along the coast, east of the Cleveland Hills, by North Sea ice. Eventually the pressure of North Sea ice forced the Stainmore ice into the Vale of York. In contrast Madgett & Catt (1978) suggested that the Stainmore ice (Withernsea Till) and the east coast ice (Skipsea Till) could have been contemporaneous, or nearly so, and that the Stainmore glacier could have over-ridden the east coast glacier. This conclusion is doubted by Bateman *et al*. (2011) who suggest that the Withernsea and Skipsea Tills represent two separate ice advances.

As the Teesdale ice pushed south into the Vale of York it was joined by ice streams from Swaledale and Wensleydale (Raistrick, 1926), Nidderdale (Tillotson, 1934) and Wharfedale (Raistrick, 1931). Based on a recognised western limit for Shap Granite erratics, Raistrick (1926) commented that Wensleydale ice likely formed a buffer zone between the Teesdale – Vale of York ice stream and the Pennine escarpment.

The southern extent for the Vale of York glacier has been a contentious issue. Kendall & Wroot (1924) and Palmer (1966) refer to an early glacial phase, when ice pushed south towards Doncaster, and also to a late or retreat stage when a stable ice front deposited the York and Escrick terminal moraines. A similar southern limit for the Vale of York glacier, as far as Wroot, has also been suggested by Gaunt (1976a), Cooper & Burgess (1993) and Westaway (2010). The expansion of the York glacier south of the Escrick Moraine is largely based on observations by Gaunt (1976a and 1976b) that a series of gravel ridges, between Thorne and Wroot, which contain an appreciable proportion of Permian limestone pebbles, originated as ice-marginal deposits along the western edge of a tongue of ice that had advanced into the southwestern part of the Vale of York. This maximum ice advance was considered to be transient during a short-lived, high-level stage of Lake Humber that followed blocking of the Humber mouth. It was also considered that the southerly surge was probably assisted by the Vale of York ice being buoyant and probably floating over appreciable areas (Gaunt, 1976b).

In contrast to the above, Parsons (1887) has stated that he knows of no 'boulder clay' exposed in the Ouse valley lower than Escrick, while Ford *et al.* (2008) describe and illustrate a maximum limit for Late Devensian ice in the Vale of York at the Escrick Moraine. A small area of till mapped on the Wakefield 1: 50 000 geological sheet (British Geological Survey 1998) located below the Permian escarpment near Monk Fryston [SE 501 293] has been reinterpreted as a littoral deposit on Figure 17.

Mapping by the British Geological Survey (2008), on the 1: 50 000 Selby sheet, has linked the area between the York Moraine and Escrick Moraine by a series of flat-topped gravel ridges, that roughly follow the A19 between Fulford and Escrick (see

Fig.15). These ridges, which include the Crockey Hill 'Esker', together form the Crockey Hill Esker Member of the Vale of York Formation (Ford *et al.,* 2008). Fairburn (2009), however, considers these ridges are eroded remnants of a fan delta deposit south of the York gap at levels of 20 m and 15 m above OD.

3.3.2.1. York Moraine

The York Moraine, which is the inner moraine of Kendall & Wroot (1924), is a twin lobate structure that was established following a retreat phase of the Vale of York glacier (Cooper & Burgess, 1993) during the York – Escrick stage of the Main Dales Glaciation (Edwards *et al*., 1950; Raistrick, 1934). Kendall & Wroot (1924) have however argued that the York Moraine is older than the Escrick Moraine and was overridden during the recession. This view, expressed by Kendall in an address to the Yorkshire Geological Society in 1893 (although not published) was based on a section near Holtby [SE 675 541] that seemed to show that the York Moraine originated as a gravel ridge on to which 'boulder clay' was plastered. The moraine has a western lobe extending from Long Marston to York and an eastern lobe extending from York to north of Sand Hutton and forms a distinct physical feature rising above the plain of the Vale of York (Fig. 14). Along its southern edge the mappable boundary of the moraine is at about 15 m above OD while to the north its boundary against the plain of the Vale of York is about 5 m higher at 20 m above OD. Topographically the moraine reaches a maximum elevation of just over 45 m above OD at Bilbrough and at several locations west of Long Marston and to the north of Sand Hutton (Fig. 14). The width of the moraine varies considerably from about 5 km at Bilbrough, where a ridge of till extends southwards towards the Escrick Moraine, to less than 1.0 km between Dunnington and Heslington. Conspicuous drainage gaps are present in the moraine along Healaugh Beck, at the southern end of the Askham Bog and along the Ouse valley in the York Gap. These gaps were important meltwater channels through the moraine during the northward retreat of the Vale of York glacier and provide physical evidence to illustrate the lowering of depositional surfaces from the north to the south of the moraine (see section 7.16.3).

Along much of its length, between Askham Bog and Long Marston, the moraine straddles an eroded ridge of Triassic sandstone (Melmore, 1935, fig. 5), which is well exposed on its southern flank where it forms a low scarp extending westwards from a small quarry [SE 532 459] south of Bilbrough to a small valley bounded by Ainsty Cliff (Plate 3).

Plate 3. The eroded contact between till and Triassic sandstone south of Bilbrough at Ainsty Cliff [SE 529 461]. The till, which is about 2.0 m thick is overlain by a lag deposit of boulders and cobbles forming a level surface above the 40 m terrace. Note the 30 cm scale. (Photograph M.D. Bateman)

To the east of the quarry, the Triassic outcrop may well continue for some distance, masked by solifluction deposits, towards Redhill Field Road [SE 534 464]. To the south of the Long Marston Obelisk and west of Hutton Wandesley (Fig. 14), the Triassic sandstone has been proven along a depression, known as 'the Glen', on an erosional terrace and in the rising hillside above it by shallow auger drilling on the Harrogate and Leeds 1: 50 000 sheets (British Geological Survey, 1987, 2003) and from sand excavated from animal burrows. Ventifacts were reported, at or near the Triassic sandstone surface, beneath glacial deposits in a borehole drilled on the western side of York [SE 577 528] by Gaunt (1974). These are indicative of arid periglacial conditions prior to deposition of the York Moraine.

The most distinctive topographic feature of the York Moraine is a tiered surface with laterally extensive horizontal terracing at 20 m, 25 m and 30 m - 35 m above OD, and where the elevation of the moraine permits, at 40 m above OD (Fig. 14). An additional formline at 30 m above OD, noted at some locations, may represent a short-lived lake level or a lithological boundary (see section 7.6.).

Superficial deposits – One might expect that the till forming the moraine should be composed of poorly sorted and unstratified sediment ranging from clay and silt up to cobbles and boulders, but based on shallow sections in roadcuts the till seems to have been segregated, to some degree, into fractions of differing grain or clast size. This could have resulted from transgressive shoreline washing (i.e. the development of progressive beaches across the surface of the moraine then at or near lake level). Shallow auger drilling for OSL sand samples indicates that these segregated layers extend for more than 2.0 m below the surface on the flanks of the moraine. This segregation was also noted by Clark (1881) who observed that in places the topmost layers of the moraine had been washed and re-arranged as gravels and brick-earths (laminated clays). It is not thought likely that the segregation could be caused by backwasting of an ice-cored moraine, as can occur in some dead-ice moraines (Schomacker, 2008) as it is more likely that segregation occurred by shoreline washing well after retreat of the Vale of York glacier.

Prior to building construction at York University, on the York Moraine near Heslington, site investigations were undertaken by civil and structural engineers. This mainly involved shallow auger drilling to determine the appropriate building design that could be supported by the subsurface. Whilst the data do indicate some layering in the moraine along crestal locations of sand and gravel over ‘boulder clay’ caution should be exercised as the information obtained was poorly recorded. (Sections and maps from which the above information was obtained were provided by the Borthwick Institute for Archives, University of York, Heslington.) Despite lack of good outcrop three classes of material have been noted from superficial deposits:

Till - Apart from the Ainsty Cliff section (Plate 3) till is generally only exposed on theYork Moraine in crestal locations above the 45 m contour at places such as Bilbrough, northeast of Sand Hutton and west of Hutton Wandesley (Fig. 14). Evidence for this is however meagre and is mainly based on the assumption that the planar surface on the moraine at *c*. 42 m above OD is the maximum elevation of Lake Humber, with gravels lying on this surface between 42 m and 45 m above OD. Temporary sections in the till have also been noted on construction sites at York University and in roadcuts that are now overgrown. Several gravel-covered mounds of till have also been mapped along the crest of the moraine: the most prominent of these being at Severs Howe (Plate 4), a location long supposed to be the burial mound of the Roman Emporer

Septimus Severus, until excavations proved otherwise (Kendall & Wroot, 1924, p. 529). Although access to Severs Howe is now restricted, evidence from nearby building sites indicates that the mound is mainly formed of gravel and not till, as suggested by Kendall & Wroot (1924).

Plate 4. Severs Howe – gravel covered till on the York Moraine forming a mound between 30 m - 35 m on the 25 m Surface [SE 581 519]. The site has restricted access and no till is visible on the photograph. (Photograph W.A. Fairburn)

Although not *in situ*, till, or gravel with cobbles and boulders derived from till, can be seen on the slopes of the moraine between the 25 m and 30 m contours below a levelled area that was cleared to provide space for a car park (Plate 5). The site [SE 623 509] lies immediately southeast of Mill Mound (Fig. 14) where sand and gravel are exposed in a roadcut between *c*. 30 m – 33 m above OD (Plate 16).

Plate 5. Excavated till composed mainly of Carboniferous sandstone and limestone (pebble sizes up to 10 cm), cleared from below the 33 m terrace and now lying on the slope of the moraine between 25 m – 30 m above OD [SE 624 508]. (Photograph W.A.Fairburn)

Lag deposits - Accumulations of boulders, mainly devoid of interstitial clay and sand are believed to have originated as lag deposits at several erosion levels on the York Moraine in a similar fashion to marine shoreline back-beach gravels. Such shoreline washing and segregation on the moraine would result in sand being deposited down the face of the moraine leaving the coarser fractions to form residual deposits. Silt and clay fractions in the till would be swept away into deeper parts of the lake, as suggested by Clark (1881). Such segregation of the morainic deposits was a consequence of the Lake Humber being impounded following the northerly retreat of the glacier and the deposition of the moraine.

Boulder beds have been noted on the York Moraine below the 30 m contour in a sunken pathway west of Mill Mound [SE 614 510], in a construction pit on the 25 m terrace [SE 618 508], forming a level surface at about 42 m above OD at Bilbrough (Plate 6), on a ridge capped by gravel and boulders 1.5 km to the northeast of Bilbrough (Fig. 14), and at several crestal locations on the moraine between Gate Helmsley and northeast of Sand Hutton (Fig. 14; see also the York 1: 50 000 sheet, British Geological Survey, 1983b).

Plate 6. Stacked boulders collected from gravels above the 40 m terrace at Bilbrough near Ainsty Cliff [SE 530 461]. Most of the boulders, which are of Carboniferous sandstone, are faceted or partly faceted and range in size from 15 cm to 20 cm. (Photograph W.A. Fairburn)

Sand and gravel - Rather chaotic banks of sand and gravel, lacking well-defined bedforms, that have only been identified in crestal regions of the moraine, as at Mill Mound and Severs Howe, appear to have formed as littoral deposits (see also Plates 28 and 29). At Mill Mound a poorly sorted deposit composed mainly of sub-rounded to faceted pebbles, with dimensions generally less than 5.0 cm, occur in a matrix of reddish-brown sand (Plate 16). Many of the pebbles show random orientation, although in places there is some crude alignment along with lenticular streaks of sand. Occasional fragments of exotic material (e.g. pieces of red tile) imply that part of the deposit could have originated from excavations around a nearby water tower. Surface auger holes indicate that the gravels are overlain by up to 0.5 m of pebble-free sand. At Mill Mound the gravels form the sub-surface of the 33 m terrace and are underlain at *c*. 30 m above OD by till or boulder beds on till (see Plate 5).

Sand mantle - Perhaps the most unusual feature of the York Moraine is the lack of surface till on the moraine due to the accumulation of later superficial deposits. In most areas, the flanks of the moraine are mantled by well-sorted sand which is essentially stone free. This sand mantle can be up to 2.0 m thick and liable to landslipage if construction excavations, as at York University, produce unstable slopes. That there is abundant sand in the moraine is not surprising as glacial erosion of the Triassic sandstone bedrock would have produced a very sandy till. The best sections through the sand mantle have been examined at the following locations:

- Below Severs Howe [SE 582 522] several rows of houses terminate against a sand bank forming the hill slope. Here, at least 2.0 m of sand was originally retained showing crude layering produced by occasional boulders. At present, the extent of the visible outcrop has been considerably reduced by further development. A small pit excavated near this location, shown in Plate 7, exposes 1.3 m of stratified sand in a superficial slope deposit that comprises three layers with erosional contacts dipping in a westerly direction at about 16°. The exposure may well represent three episodes of sand washing on the flank of the York Moraine.

Plate 7. Superficial sand on the 20 m terrace of the York Moraine below Severs Howe [SE 582 522]. This OSL sample location exposes 1.3 m of stratified sand. The scale is 30 cm. (Photograph M.D. Bateman)

- East of Bilbrough [SE 538 468] a roadcut through the 40 m surface has again exposed about 2.0 m of sand with a boulder layer that is not always visible: cobbles and boulders are widespread on the surface above. Part of the section was excavated for an OSL sample point (Plate 8).

Plate 8. Superficial sand below the 40 m terrace on the York Moraine in a roadcut east of Bilbrough near Highfield Farm [SE 538 468]. This OSL sample point exposes a 65 cm section of fine-grained, well-sorted sand comprised of angular to sub-rounded sand grains (now thought to have been excavated from the roadcut). The scale is 30 cm. (Photograph W.A. Fairburn)

- At York University an access road behind some of the University buildings [SE 624 508] left an unstable slope on the sand mantle at an elevation of 25 m - 30 m above OD. Part of this slope subsequently slipped on to the road below revealing that the sand cover could be at least 1.0 m thick. The slope has now been stabilised.

The first two of these locations were sampled for OSL dating: slippage and rabbit burrows making the third location unsuitable (Plate 9).

Plate 9. Rabbit burrows in sand scree above stabilised slope between 25 m – 30 m above OD at York University [SE 624 508]. The sand mound from the burrows is over 1.0 m wide. (Photograph W.A. Fairburn)

3.3.2.2. Escrick Moraine

The Escrick terminal moraine, or the outer moraine of Kendall & Wroot (1924), like the York Moraine forms a prominent crescent-shaped ridge that extends north-easterly from Stillingfleet to Wheldrake, where the River Derwent breaks through a gap in the moraine, and then northwards following the eastern bank of the Derwent to High Catton. West of Stillingfleet the moraine is represented by a series of gravel terraces, at about 15 m above OD, that follow the eastern bank of the River Wharfe to north of Tadcaster (Fig. 14). The moraine has also been regarded as part of the York-Escrick stage of the Main Dales Glaciation by Raistrick (1934) or the Escrick phase of a retreating Vale of York glacier (Cooper & Burgess, 1993).

Like the York Moraine there are no natural sections through the Escrick Moraine that are suitable for descriptive purposes. Consequently the most detailed lithological information has been gained from excavations (some temporary) and from published accounts of gravel pits. Kendall & Wroot (1924) refer to sections at High Catton consisting mainly of 'boulder clay' and describe a railway cutting through the moraine, west of Escrick (Fig. 14) as being mainly of 'boulder clay' with some wisps of sand and gravel (see also a description by Parsons, 1887, from the same location). The same authors also examined gravels along the eastern side of the moraine near High Catton, which contained Shap Granite and a suite of rocks from the Eden Valley, the Pennines and eastern Scotland – some of these having a 'desert varnish' (i.e. ventifacts). Melmore (1935) examined a gravel pit at Burtonfields [SE 733 556] where he recorded 'boulder clay' intercalated with bedded sand and gravel containing pebbles of sandstone, limestone, chalk and flint. The deposit was underlain by 13.4 m of clay and gravel (possibly till) over Triassic marl. Stather (1913) described a section in the moraine at High Catton, possibly near Gravel Pit Farm [SE 717 535], exposing cross-bedded sand and well-rounded gravel. Erratics here included Carboniferous sandstone and limestone as well as Shap Granite and Brockram (these gravels are possibly littoral gravels derived from a 33 m shoreline as the moraine near this locality rises to an elevation of 35 m above OD). At Sutton upon Derwent [SE 707 468] Harcourt (1829) found 20 m of diluvium (till) on chalk and flint gravel (these gravels could have originated from the older alluvial fans to be described later in this text (section 7.12.1.). More recent observations include Ford *et al*. (2008) who described the Escrick Moraine as a gravelly, sandy, clay till. A temporary section through the Escrick Moraine examined by Gaunt (1970) exposed 'boulder clay' on the summit of the ridge overlain by sand and gravel on its northern slope below 15 m (Fig. 31).

For the purposes of this present research, the most significant feature of the Escrick Moraine, between Stillingfleet and Wheldrake, is that the ridge rises above the plain of the Vale of York from about 10 m above OD to elevations that seldom exceed more than 2.0 m – 3.0 m above a terrace mapped at 15 m above OD (Fig. 14). In this region, the till forming the moraine is not accompanied by any extensive gravel deposits as is indicated on the Selby 1:50 000 geological map (British Geological Survey, 2008). Gravel was recorded by Gaunt (1970, fig. 2; Fig. 31 this text) on the northern side of the moraine only. Between Sutton upon Derwent and High Catton (Fig. 15), where the base of the moraine is at 15 m above OD, its elevation reaches 35 m above OD. As shown on Figure 15, the moraine in this region has a conspicuous terrace at 25 m above OD. Unlike the Stillingfleet – Wheldrake section of the moraine, however, extensive gravel deposits were worked at High Catton (Stather, 1913) to the west of the summit, suggesting there was more significant erosion and washing of the moraine at this elevation (i.e. above 33 m), as on the York Moraine. This preferential accumulation of gravel on the northern and western face of the moraine could be attributed to a damming effect of the moraine and the movement of lake meltwater to gaps in the moraine cut by the Derwent and Ouse rivers. That accumulation of gravel and sandy gravel can occur on the north side of the moraine, by the lateral movement of meltwater to a Derwent outlet, has been noted to the north of the Escrick Moraine west of Wheldrake (Fig. 31).

3.3.2.3. Vale of York Till

The British Geological Survey (2008) have mapped extensive areas of till, between the Escrick and York Moraines, extending from the west of Stillingfleet and north-eastwards towards Dunnington. These tills have been included in the Vale of York Formation (Ford *et al*., 2008). The deposit has been described as a unit of variable composition ranging from a sandy, gravelly clay with common cobbles and boulders to a slightly clayey sand and gravel: clasts include Carboniferous limestone and sandstone with some volcanic material (Ford *et al*., 2008). Much of the outcrop underlies the 15 Metre Surface (described later) between Stillingfleet and Bridge Dike (Fig. 14). Although the till is mainly seen in cultivated fields, a section in the till is present through the 15 Metre Surface along the disused York to Selby rail line (now part of the Trans Pennine Trail) as shown on Plate 10 [SE 615 436]. In places washing has produced a superficial lag deposit of sand and boulders over the till similar to the boulder pavement above the Blackhall Till at Whitburn Bay described by Davies *et al*. (2013a). The till has also been exposed in the drainage ditch adjacent to the disused line and in Wood Dike east of Bell Hall (Fig. 37) below sections in the Naburn Sand Member (Fig. 25). The till was also extensively exposed in excavations during the construction of the York Railway Station where it is associated with highly involuted laminated clays (Fox–Strangways, 1884, fig.3). These excavations yielded the boulders of Shap Granite that can be seen in the Museum Gardens in York (Plate 2).

Plate 10. Vale of York Formation till exposed beneath the 15 Metre Surface on the Trans Pennine Trail [SE 615 436]. The largest pebble is about 8.0 cm (Photograph W.A. Fairburn)

Parsons (1887) has stated, that to his knowledge, ‘boulder clay’ is not exposed in the Ouse valley south of Escrick: this opinion has been confirmed during the recent mapping (for deposits of Late Devensian age) in the preparation of Figures 14 and 17. Outcrops of till (of Anglian age?), near Monk Fryston [SE 501 293], mapped by the British Geological Survey on the Wakefield 1:50 000 sheet (British Geological Survey, 1998) are regarded in this text as littoral deposits on the 15 Metre Surface. As already stated earlier, occurrences of till in excavations near Ferrybridge (Bateman *et al*., 2008) could be solifluction deposits derived from Pennine Boulder Clay.

3.4. Holderness

The cliff section at Dimlington on the Yorkshire Coast is the type site for the Last Glacial Maximum in Britain (Rose, 1985). Here, the Basement Till is overlain by the Skipsea Till and the Withernsea Till, which are separated by up to 4.0 m of stratified sands, silty clays and fine chalk gravel (Madgett & Catt, 1978; Catt, 2007; Evans & Thomson, 2010). The upper part of this was thought to have been deposited by a two-tiered ice sheet; the Withernsea Till originating from northwest England via the Stainmore Gap and the Skipsea Till from east coast ice originating from southern Scotland, the Cheviots and northern England (Catt, 2007). The intervening sediments were considered to have been deposited by englacial streams from a single ice sheet (Evans *et al.,* 1995; Catt, 2007) or as deglacial lake sediment and subaqueous ice-contact fans (Evans & Thomson, 2010). A more recent study of these sediments by Bateman *et al.* (2011) has suggested a subaerial origin, in a sandur type pond environment, based on sedimentary and fossil insect evidence. Optically stimulated luminescence (OSL) dating of four samples taken from a freshly exposed section in these deposits has given a depositional age of 16.2 ± 0.4 ka (Bateman *et al.*, 2011), which dates glacial retreat between the two tills.

Lithologically similar tills to the Skipsea and Withernsea have been reported from Upgang near Whitby on the North Yorkshire coast by Roberts *et al.* (2013). These authors have described a distinct tripartite sequence composed of two tills (lithofacies associations LFA 2 and LFA 4) separated by a sequence of stratified clay, silt, sand and gravels (LFA 3). Till fabric data and clast lithologies, support a common provenance from the north of England and southern Scotland without an easily discernible westerly imput of ice. Although not explicitly stated by the authors (Roberts *et al*. 2013) the sequence is broadly similar to sections at Dimlington described by Catt & Penny (1966), Madgett & Catt (1978) Evans & Thomson (2010, fig. 11) and Bateman *et al*. (2011, fig.4) that contain the Skipsea and Withernsea tills (LFA 1 and LFA 4).

At Dimlington a section from Catt & Penny (1966) shows that the Skipsea Till is underlain by the Basement Till, which as far as is known is the oldest till in Holderness (Lamplugh, 1889; Catt, 2007). On the surface of the Basement Till lacustrine silts occur in basins 1.0 m – 4.0 m deep and from 5.0 m to over 50 m wide (Catt & Penny, 1966; Catt, 2007). Strands of fossil moss, from these Dimlington Silts were dated by Penny *et al.* (1969) to give ^{14}C ages of 18.5 ± 0.4 ka BP and 18.25 ± 0.25 ka BP (23.0 – 20.8 and 22.3 – 20.9 cal. ka BP). These dates, along with dating of organic matter from a kettle hole in the Withernsea Till at The Bog near Roos (Beckett, 1981; Catt, 2007, fig.1), that gave a ^{14}C age of 13.0 ± 0.27 ka BP (16.3 – 14.5 cal. ka BP) and the OSL dating by Bateman *et al*. (2011) effectively dates the Skipsea Till to the period 20.9 ka – 16.2 ka BP and the Withernsea Till to the period 16.2 ka – 15.5 ka BP (Bateman *et al*., 2011). These dates allow the Skipsea and Withernsea Tills to be equated with episodes GS-2b and GS-2a from the Greenland (GRIP Summit) record (Björck *et al*., 1998; Bateman *et al*., 2011; Fig. 4). It should be accepted though that both these ^{14}C dates above (i.e. Dimlington and the Roos Bog) are from samples that could have been contaminated by hard-water effects, which could have added appreciably to their age. An additional date, obtained by thermoluminescence techniques (Wintle & Catt, 1985) from loess beneath the Skipsea Till at Eppleworth (30 km WNW of Dimlington, [TA 014 320]) could further restrict the age of the inland base of this formation to 17.5 ± 1.6 ka BP in this region.

The age of the Basement Till, which may be relevant in this text, like many other events in the Quaternary, has been a contentious issue. This has largely arisen because of differences in the accepted position of the Basement Till within the section against the buried cliff at Sewerby. Following excavations at the site in 1887 and 1888, funded from a grant by the Yorkshire Geological Society, Lamplugh (1888) considered that in the section above the raised beach, which contains of vertebrate remains resting on a chalk base, the only boulder clay present was the 'Purple Clay' (equivalent to the Drab or Skipsea Till in more modern terminology). He also considered that the 'Basement Clay' (Basement Till) was absent and possibly represented by chalk rubble. Lamplugh, however, later changed his mind and decided that the 'Purple Clay' in the section was the 'Basement Clay' (Lamplugh, 1890 and 1891). This later interpretation has been illustrated in a section by Wilson (1948).

Following a low spring tide in 1963 exposures of the Basement Till, covered with patches of calcreted conglomerate, were seen within 70 m of the buried cliff section (Catt & Penny, 1966). These authors concluded that the conglomerate represented the cemented seaward continuation of Lamplugh's beach shingle and therefore the raised beach is younger than the Basement Till. Confidence was also expressed in their belief that the 'Purple Clay' of Lamplugh was not the Basement Till. Consequently as the Sewerby fauna in the raised beach is of Ipswichian age (Catt & Penny, 1966; Catt, 2001) the older Basement Till must have been deposited in MIS 6 or earlier. Dating of wind-blown sand, above the raised beach, which gave a weighted mean of 120.84 ± 11.8 ka (Bateman & Catt, 1996) confirms the Ipswichian age for the raised beach.

There also remains the interesting possibility that the Basement Till containing Scandanavian erratics (Catt, 2007) could be coeval with the basal till on the Durham coast, the Warren House Till described by Davies *et al*. (2013b). The Warren House Till also with Scandanavian affinities, could be dated to earlier than MIS 7 as it may be the source of erratics in the Easington raised beach (Davies & Bridgland, 2013; Davies *et al*., 2009; Davies *et al*., 2013b).

Further discord is provided by Eyles *et al*. (1994) who dated Pleistocene Arctic marine molluscs, mainly *Macoma balthica* and *Arctica islandica* from the Basement Till by amino acid epimerisation. The results of the dating showed a range of ages from pre-Devensian to Late Devensian. Based on the accepted principal that the youngest molluscs indicate the maximum age for the depositional event of the till, Eyles *et al*. (1994) assigned a Late Devensian age (*c*. 20 ka BP) to the Basement Till. Catt (2001), who contends that the Basement Till underlies the Ipswichian beach at Sewerby, explains the apparent Late Devensian age for the till from incorporation of bivalves into the upper layers of the till when it was remobilised during the advance of the ice that deposited the Skipsea Till. Walker (2005) has also warned that amino acid geochronology can be unreliable as the rate of epimerisation is partly temperature dependent: the techniques should therefore be considered as a guide to relative dating of samples with similar thermal history, rather than as a definitive dating tool.

Livingstone *et al*. (2012) have proposed a six-stage glacial model of Devensian ice-flow history in the central sector of the last British-Irish ice sheet (BIIS) including three major phases of glacial advance. Stage I is correlated with maximum ice sheet expansion, between *c*. 29 ka – 23 ka BP, and consideration is given to including the Basement Till of Dimlington in this interval based on the Late Devensian age of *c*. 20.0 ka BP given by Eyles *et al*. (1994). Two major oscillations of the North Sea ice lobe flowing southwards into Yorkshire during Stages IV and VI are thought to have deposited the Skipsea Till between 21.7 ka – 16.2 ka BP and the Withernsea Till between 16.2 ka – 15.5 ka BP (chronology from Bateman *et al.,* 2011).

Further constraints on the age of the Skipsea Till have been provided by OSL dating from Barmston and Dane's Dyke, on the Holderness Coast, by Hartmann (2011). At Barmston, a sand sample from within the Skipsea Till (subglacial stream deposit) and one from planar cross-bedded sands, above the till, gave dates of 17.1 ± 1.33 ka and 16.51 ± 1.04 ka respectively. Below the Skipsea Till, at Dane's Dyke, an erosional valley in the Chalk contains brecciated chalk rubble, solifluction deposits and lacustrine sediments grading upwards from laminated silts to sands and gravels. A sand sample from the sequence was dated at 24.14 ± 1.66 ka. These results therefore date the advance of the Skipsea ice at before *c*. 16.51 ka BP within the range of *c*. 24.1 ka – 17.1 ka BP. As Hartmann (2011) implies that the lacustrine beds, at Dane's Dyke, could have been deposited in a lake impounded by North Sea ice, then these beds could be coeval with the lacustrine sediments described by Murton *et al*. (2009) at Hemingbrough, which are similarly dated.

3.4.1. Humber Till

Till in the Humber channel exposed at North and South Ferriby, referred to by Gaunt (1976b) as the 'Humber Boulder Clay' has been identified as far west as Elloughton Beck [SE 948 268] (Gaunt, 1976b), where pebble orientations in the till show an ESE peak. The till, while not presenting an impressive landform on the current landscape, must at some stage in the Late Devensian, along with North Sea ice, have formed an impressive barrier in the Humber Gap. Both de Boer *et al*. (1958) and Gaunt (1981) support a view that the glacial deposits survived as a plug in the Humber Gap after the North Sea ice had retreated eastwards. Gaunt (1981) also implies that although high-level Lake Humber may have been short-lived, with drainage through the Lincoln Gap, blockage of the Humber by the till must have preserved a low-level lake, at between 8.0 m – 9.0 m above OD, long enough to allow the accumulation of a thick sequence of laminated clays.

There now seems little doubt that blockage of the Humber mouth did occur (at least once) in the Late Devensian and that this event has been the most important factor in controlling post-glacial deposition in the Vale of York. It is therefore important to establish the extent and age of the till, by reference to the Dimlington sequence, but also to determine, if possible, the precise geometry of the deposit with respect to any post-depositional erosion caused by changing levels of Lake Humber.

At Red Cliff, southwest of Hessle (Fig. 16), two tills were identified on the Humber foreshore and in the low cliff above by Stather (1897) and de Boer *et al*. (1958): a lower leaden coloured till overlying laminated clays, separated from an upper reddish brown clay till by impersistent thin laminated clay (Gaunt *et a*l. 1992). The upper till has been referred to as the Hessle Clay by Bisat (1932) and de Boer *et al.* (1958). De Boer *et al.* (1958) also describes sections in the Hessle 'Boulder Clay' in Little Switzerland, a now disused chalk pit, that has been developed into the Humber Bridge Country Park, and in the Hesslewood quarries [TA 012 263] now used as a rubbish dump. The subdivision of the ‘Humber Boulder Clay’ into two layers is perhaps an oversimplification of the geology, as a line of sections through the till along the edge of the Humber (Gaunt *et al*., 1992, fig. 41) indicates multi-layering and poor correlation of the lithology between sections. This suggests oscillations of the ice-front in the Humber Gap resulting in alternations of till and lacustrine sediments. During a recent inspection of the Red Cliff section in 2011, the highly contorted grey till in the lower part of the cliff (*c*. 5.0 m above OD) was seen to be overlain by laminated silts that also filled steep-sided erosion channels in the till (Plate 11). Distortions in the till are also likely to result from an erosional event during retreat of the ice-front. At a higher part of the cliff (*c*. 10 m above OD) the grey till is overlain by cross-bedded sand and gravel (Plate 27) below a decalcified top (see also plate 5, Gaunt *et al*., 1992).

Plate 11. Laminated silts overlying contorted grey till in the lower part of the cliff in the Red Cliff section. The silts also fill steep-sided erosion channels in the till. (Photograph W.A. Fairburn)

In a review of the Pleistocene deposits of Holderness, Catt and Penny (1966) regarded Red Cliff as a location where the Drab Till (now Skipsea Till) underlies the Hessle Till (see also Stather, 1897, Section A). They also considered that the Hessle Till overlaps the Drab Till near Hessle and rests directly on the Chalk. This view has now been revised by Madgett & Catt (1978) who regard the Hessle Till as a weathered profile that can occur either above the Skipsea Till or the Withernsea Till. Consequently only a single till is now recognised at Red Cliff and that is the Skipsea Till (see Catt, 2007, fig. 1; Fig. 16), which has been given an age of 21.7 – 16.2 ka BP (Bateman *et al*., 2011).

In general, the Skipsea Till does not present a distinct landform and is poorly exposed in the Hessle – North Ferriby region. Apart from temporary building excavations e.g. in extensions to a Motel midway between North Ferriby and Hessle [TA 009 260], the only outcrops occur at Red Cliff and in weathered form around the rim of the Humber Bridge Country Park and in other minor quarry excavations. Consequently, mapping of the Skipsea Till shown on Figure 16 has relied heavily on mapping shown on the Kingston upon Hull geological sheet (British Geological Survey, 1983a). It was however noticed during the recent mapping (2010-2011) that the till has been conspicuously terraced at 33 m, 25 m, 20 m and 15 m above OD (Fig. 16). If these terraces are shorelines of Lake Humber (as the author believes), then much of the Skipsea Till outcrop must have been reworked in the manner of the York Moraine, although perhaps to a lesser degree. This naturally presents a mapping problem and the BGS mapping has been modified to exclude *in situ* till below the 15 m terrace because of suspected extensive reworking of the till at this prominent feature. That significant erosion of the till may have occurred is indicated by extensive shallow excavations on the 33 m terrace, just north of the Humber Bridge County Park [TA 016 265], which may have been developed to exploit gravel lag deposits for road works. It was also noticed that till is absent around the rim of a small chalk pit, located near Hessle, on the 25 m terrace [TA 026 263].

3.5. Lake Humber

The concept of pro-glacial Lake Humber is as valid today as when it was first proposed by Carvill Lewis in 1894. Gaunt (1994) has proposed that Lake Humber originated about 18 C^{14} ka BP (i.e. post-deposition of the Dimlington Silts) by glacial blockage of the Humber Gap – an event linked to the westernmost penetration of the Skipsea Till (Horkstow Moraine) up the Humber estuary (Evans *et al*., 2005; Frederick *et al*., 2001). Gaunt (1974, 1994) considered that the lake rose initially to about 33 m above OD to a high-level stage when a tongue of ice surged southward down the Vale of York. Discontinuous littoral or shoreline deposits were formed around the lake margin at this time. The transient high-level phase was followed by a fall in lake level to below 4.0 m below OD; it then rose to a more prolonged level of about 8.0 m above OD (Gaunt *et al*., 2006; Fig. 11). The low level of the lake, at 4.0 m below OD, is based on a borehole near Carlton, which intersected what has been described as a desiccated layer in laminated clay at this level (Gaunt, 1994).

Evidence for the high-level phase of the lake is based on the recognition by Edwards (1937) of strand-line deposits in the Vale of York close to the 100ft contour (30.5 m) on the Permian escarpment between Tadcaster and Doncaster. Edwards (1937) also refers to the sands, silts and clays forming the wide plain of the Vale of York, at about 25 ft (7.6 m) above OD, as the 25 Foot Drifts. This was perhaps not a wise choice for the topographic level of the Vale of York, as much of this surface to the east of the Permian escarpment and the 100ft strandline lies at about 10 m (32.8 ft) above OD. Another event that can be attributed to high-level Lake Humber is the reversal of drainage up the Trent Valley and through the Lincoln Gap, indicating that ice blocked both the Humber and The Wash (Gaunt, 1981; Straw, 1979; Evans *et al*., 2005). The extent of the impounded glacial Lakes is shown on the Glacial Map of Britain (Clark *et al*., 2004) and by Murton & Murton (2011).

The retreat phase of the Vale of York glacier and the deposition of the York and Escrick moraines is not well documented, although Gaunt (1981) does suggest that as the lower part of the laminated clays continue northwards in places under the morainic deposits then the glacial episode associated with the deposition of the moraine occurred entirely within low-level Lake Humber. Geological mapping by the British Geological Survey for the 1:50 000 Selby sheet (British Geological Survey, 2008), does not concur with this sequence of events as they map two distinct phases of laminated clays with the Escrick Moraine as an intermediate event. This timing is also contrary to the conclusions of Fairburn (2009) who explains terracing on the York Moraine, at levels up to 40 m above OD (Fig. 14), as resulting from high-level Lake Humber, with the Vale of York glacier well to the north of the York Moraine.

Equally poorly known are the age restraints for the advance and retreat of the Vale of York glacier, as well as the rise and fall of Lake Humber. A maximum age for Lake Humber was established by Gaunt (1974) at 21.8 ± 1.6 ka ^{14}C yrs BP (26.2 ± 2.0 cal. ka BP, Murton *et al*., 2009) from an unidentifiable bone fragment found in sand and gravel, near Brantingham, at an elevation of *c*. 25 m above OD. This date is much older than the age of the Dimlington Silts. The age of the bone sample could however be younger than this date suggests if there was carbon exchange between the bone and its depositional environment (Walker, 2005). An OSL age of 16.6 ± 1.2 ka was determined by Bateman *et al*. (2008), at Ferrybridge, from distal sands interpreted as originating from high-level Lake Humber (the Older Littoral Sands and Gravels of Gaunt, 1976b) at an elevation of about 27.5 m above OD. These sands are underlain by periglacial slope deposits dated at 20.5 ± 1.2 ka and loess at 23.3 ± 1.5 ka (Bateman *et al*., 2008). A diamicton, in the sequence, has an age between the latter two dates. The major conclusion from this work is that ice blocked the Humber Gap prior to 16.6 ± 1.2 ka BP. Conflicting evidence, on the age of Lake Humber, has been provided by Murton *et al*. (2009) who dated two wave-rippled sands, from within a massive silt unit assigned to the upper part of the Park Farm Clay Member of the Hemingbrough Glaciolacustrine Formation (Ford *et al*., 2008), in the central part of the Vale of York south of the Escrick Moraine. These OSL dates of 21.0 ± 1.9, 21.9 ± 2.0 and 24.1 ± 2.2 ka were presented as providing the first chronological evidence for low-level Lake Humber.

Clearly an older age for low-level Lake Humber, compared with high-level Lake Humber, is not acceptable based on current understanding of the history of the pro-glacial lake in the Vale of York. Murton *et al*. (2009) overcame this problem by suggesting that the distal beach sands, sampled by Bateman *et al*. (2008), may have been reworked by colluvial processes, but supplied no supporting evidence for this claim. The status of the Hemingbrough Glaciolacustrine Formation with respect to glaciation in the Vale of York will be discussed later in this text (section 10.1.).

Additional dating of sands, associated with an extension of Lake Humber, is provided by a sand sequence near Caistor on the side of the Ancholme Valley in Lincolnshire. While this region is not contained within the mapped area, the data is relevant to the present study. At Caistor, Bateman *et al*. (2000) were able to date four discrete sand units, separated by palaeosols or peat layers, by thermoluminescence dating. The basal lacustrine unit (22.67 ± 1.4 ka) is overlain by fluvial low-angle fans (18.61 ± 2.0 ka – 14.14 ± 1.2 ka) and capped by either aeolian coversand (12.78 ± 1.2 ka – 11.85 ± 4.6 ka) or reworked coversand (6.46 ± 0.6 ka). Radiocarbon dating, of the peat layers, was in accord with the luminescence dating. Bateman *et al*. (2000) suggested that the dating provides a chronological framework for the demise of Lake Humber, from a high-level stand before *c*. 18 ka (i.e. the onset of fluvial deposition). Compared with current dating at Dimlington, it is evident that the basal sand unit at Caistor (range 24.07 ka – 21.27 ka) was deposited in a lacustrine event that could pre-date the Skipsea Till, which overlies the Dimlington Silts dated between 20.8 – 23.0 cal. ^{14}C ka BP (or even younger if hard-water effects are applicable). It therefore might indicate an earlier blocking of the Humber mouth or possibly a minor glacial event preceding the Skipsea Till (see

Summary, *Lake Humber*). The younger low-level fluvial fans gave ages not inconsistent with events in the Vale of York (five out of six samples were dated between 14.14 ± 1.2 ka and 17.67 ± 1.2 ka), as they could post-date the age for high-level Lake Humber determined at Ferrybridge (16.6 ± 1.2 ka, Bateman *et al*., 2008).

A summary of the chronological data, discussed above, is shown in Table 1. The same data are shown in Table 2 with vertical and horizontal lines that separate events that post-date high-level Lake Humber and the emplacement of the Skipsea Till respectively.

TABLE 1
Summary of Chronological Data

Location	Source	Age (ka)	Environment
Barmston	Hartmann, 2011	12.31 ± 0.60 – 12.02 ± 0.72 (3)	post-glacial lacustrine and deltaic sands
Caistor	Bateman *et al*., 2000	12.78 ± 1.2 to 11.9 ± 4.6 (1)	coversand
The Bog, Roos	Beckett, 1981	16.3 – 14.5 (2)	kettle hole in Withernsea Till
Tees estuary	Jones, 1976	15.2 ± 1.9 (2)	organic / carbonaceous
Caistor	Bateman *et al*., 2000	18.6 ± 2.0 to 14.1 ± 1.2 (1)	fluvial sands
Dimlington	Bateman *et al*., 2011	16.2 ± 0.4 (3)	sub-Withernsea Till
New Close (Yorkshire Dales)	Telfer *et al*., 2009	16.5 ± 1.7 (3)	loess
Barmston	Hartmann, 2011	16.51± 1.04 (3)	cross-bedded sands above Skipsea Till
Ferrybridge	Bateman *et al*., 2008	16.6 ± 1.2 (3)	littoral sands
Barmston	Hartmann, 2011	17.01 ± 1.33 (3)	sub-glacial sands within Skipsea Till
Eppleworth (40km WNW of Dimlington)	Wintle & Catt, 1985	17.5 ± 1.6 (1)	solifluction/loess sub-Skipsea Till
Dimlington	Eyles *et al*., 1994	*c*. 20.0 (4)	marine mollusc valves in the Basement Till
Ferrybridge	Bateman *et al*., 2008	20.5 ± 1.2 (3)	periglacial slope
Dimlington	Penny *et al*., 1969	23.0 – 20.8 (2) 22.3 – 20.9 (2)	Dimlington Silts
Hemingbrough	Murton *et al*., 2009	21.0 ± 1.9 21.9 ± 2.0 24.1 ± 2.2 (3)	lacustrine sands
Caistor	Bateman *et al*., 2000	22.7 ± 1.4 (1)	lacustrine sands
Ferrybridge	Bateman *et al*., 2008	23.3 ± 1.5 (3)	loess
Dane's Dyke	Hartmann, 2011	24.14 ± 1.6 (3)	lacustrine sands below Skipsea Till
Brantingham	Gaunt, 1974	26.2 ± 2.0 (2)	bone sample from reworked gravels forming alluvial fans
Dimlington	Bateman *et al*., 2011	16.2 (3) – 15.5 (2)	age range for Withernsea Till
Dimlington	Bateman *et al*., 2011	21.7 (2) – 16.2 (3)	age range for Skipsea Till

(1) thermoluminescence (TL)
(2) radiocarbon (^{14}C) – cal. Ka BP
(3) optically stimulated luminescence (OSL)
(4) amino acid dating – ka BP

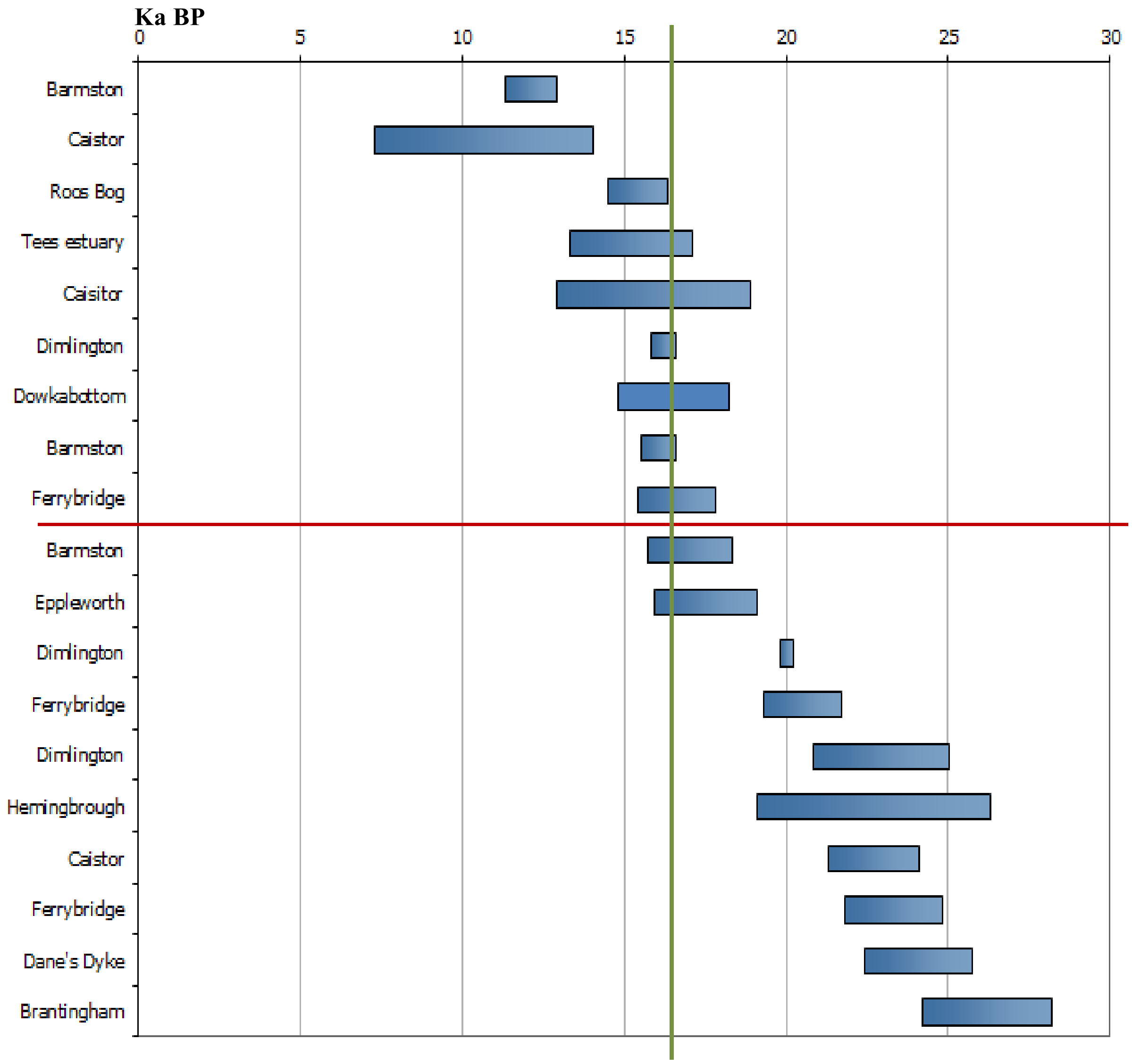


Table 2. Dates relevant to Lake Humber from Table 1. The vertical line is the date of high-level Lake Humber (c. 16.6 ka BP) based on littoral sands from Ferrybridge (Bateman *et al*., 2008). Sites above the horizontal line post-date the Skipsea Till (c. 17.01 ka BP) based on sub-glacial sands within the till (Hartmann, 2011). The above locations are listed in the same chronological order as those in Table 1.

4. PHYSIOGRAPHICAL EVOLUTION OF THE LATE DEVENSIAN EAST COAST PRO-GLACIAL LAKES

It has now become firmly established (Agar, 1954; Smith, 1994; Hughes & Teasdale, 1999; Hughes, 2008; Clark *et al*., 2012 and Murton & Murton, 2011) that the genesis of the Late Devensian pro-glacial lakes along the east coast of England (Fig. 9) commenced mainly with the retreat of inland ice prior to any significant retreat of North Sea ice.

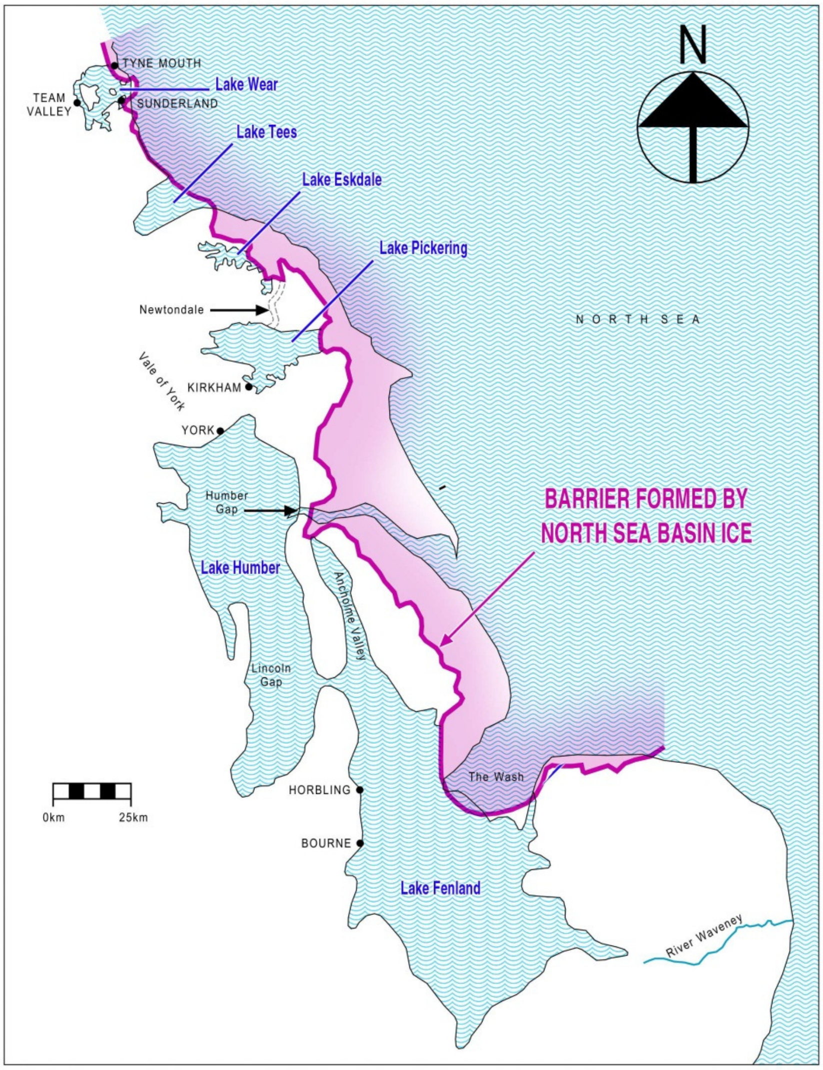


Figure 9. Plan of east coast pro-glacial lakes impounded by the edge of the North Sea basin ice barrier. Glacial lakes Tees (Agar, 1954), Eskdale and Pickering (Kendall, 1902) and Humber (Straw, 1979) probably all exited to the southern North Sea by way of Lake Fenland (Straw, 1979) and the River Waveney at some stage during the retreat of onshore ice. Glacial Lake Wear (Smith, 1994) which discharged to the North Sea through a gorge at Sunderland was probably divided from an older channel in the Team Valley and a connection with the River Tyne.

The ice margin, of the North Sea ice lobe, therefore remained as the principal damming agent that impounded glacial meltwater. The retreat pattern, that resulted in the uncoupling of the inland and offshore ice, is illustrated in Figure 6 (from Hughes, 2008).

Whilst the genesis of the pro-glacial lakes may seem to be a logical result of a glacial retreat pattern, the subsequent drainage history of these lakes is less well understood and only in the Vale of York, occupied by Lake Humber, has a detailed evolutionary model been proposed (Gaunt, 1976b). Factors, some unexplained, that must have been controlled by the dynamics of North Sea ice, particularly during retreat include:

- the maximum elevation attained by the lakes;
- the maintenance of lake levels at certain precise elevations;
- possible multi-staged retreat of the lakes;
- some minor fluctuations in lake levels between the main retreat stages,
- the development of overflow channels.

In Lake Humber, for instance, it has been argued by Straw (1979) and Clark *et al*. (2004) that this lake, along with its contemporary Lake Fenland, could only have reached its maximum level of *c*. 30 m above OD if there was equilibrium between the two lakes through the Lincoln Gap along with closure of the Humber Gap and The Wash by North Sea ice. Any overflow from the system could have been controlled by an outlet through the Waveney Valley gap (Straw, 1979, fig. 3.1), although such an outlet remains unproven. Fluctuations, above or below *c*. 30 m above OD, might therefore be only short-lived events, with the lake level returning to visit its old shoreline.

In the following account of the physiographic evolution of the east coast pro-glacial lakes the role of North Sea ice will be discussed particularly in regard to its control over the re-shaping of some aspects of east coast drainage by the forcing of overflow channels. This factor has been of importance in deflecting drainage of meltwater from Lake Eskdale into Lake Pickering and eventually to the southern North Sea by way of Lake Humber, Lake Fenland and the Waveney Valley (Murton & Murton, 2011, fig. 16). Later drainage, from the system, may well have been through the Humber Gap following the northerly retreat of ice from Holderness. There seems little evidence that Glacial Lake Wear could have formed part of this system, as it developed a different escape route to the North Sea (Smith, 1994). However Lake Tees may well have drained into the Vale of York at some stage (Radge, 1939).

4.1. Glacial Lake Wear

As might be expected, the glacial and post-glacial history of Glacial Lake Wear (Fig. 9), as described by Smith (1994), has many general similarities to events in the Vale of York. A basal ground moraine (Durham Lower Boulder Clay), which rests mainly on eroded bedrock, is overlain by laminated clays and silts (Tyne-Wear Complex) that pass upwards into lacustrine and fluvial sands with some degree of intercalation between the two lithologies. It is evident that Glacial Lake Wear may have stood at several elevations below a maximum level attained by lacustrine sediments of the Tyne-Wear Complex at 132 m above OD as these sediments are also widespread at 70 m above OD (Smith, 1994). Temporary stands have also been recorded by the cutting of benches into the Durham Lower Boulder Clay at 45 m and 43 m above OD. The map of Glacial Lake Wear (Fig. 9), from Smith (1994) has been drawn at this lowest level (i.e. 43 m above OD).

Glacial Lake Wear, denied an exit to the North Sea Basin from the former course of the proto-Wear and an older channel by way of the Team Valley to Tynemouth, initially incised an overflow channel at 90 m above OD to the southeast but later escaped to the sea by cutting a gorge through the Magnesian Limestone at Sunderland (Fig. 9).

4.2. Glacial Lake Tees

Radge (1939) described two lake levels in Lake Tees: a high-level phase at 90 m above OD (established by possible lacustrine clays) and a low-level phase at 25 m above OD. The high-level lake was considered by Radge (1939) to merge with Lake Humber and possibly Lake Wear, whilst the eastward retreat of North Sea Basin ice and the accompanying drainage of Lake Humber enabled the formation of low-level Lake Tees at *c*. 25 m OD (Murton & Murton, 2011). Such an association of Lake Tees with low-level Lake Humber is not unrealistic considering the retreat pattern modelled by Hughes (2008) on Figure 6, the deglaciation of the Tees estuary by *c*. 18.3 ka BP (Plater *et al*., 2000) and the suggestion by Hughes (2008) that retreat of ice into Wenslydale from the Vale of York was in synchrony with the retreat of the Vale of York ice lobe.

Terracing was also noted by Agar (1954), in Lake Tees, where irregular patches and strips of sand occur at intervals along the edges of the lacustrine laminated clays (Agar, 1954, plate 17). These marginal sands mainly occur at elevations between 55 ft - 87 ft (16.8 m – 26.5 m) above OD. At one location Agar (1954) describes a shelf 600 ft -

900 ft (183 m – 274 m) wide, rising from 70 ft – 75 ft (21.3 m – 22.9 m) above OD, backed by a 'cliff' up to 82 ft (25.0 m) above OD.

4.3. Glacial Lakes Eskdale and Pickering

Access to the North Sea for drainage from Lakes Eskdale and Pickering, to maintain or adjust lake levels is even more tenuous. There seems little doubt that Lake Eskdale (eventually composed of several smaller lakes) overflowed into Lake Pickering by way of Newtondale (Kendall, 1902; Fig. 9) from a possible maximum elevation of *c*. 225 m above OD (Murton & Murton, 2011). It is also possible that Lake Pickering discharged into Lake Humber through Kirkham Gorge into the Vale of York (Clark *et al*., 2004; Murton & Murton, 2011) with drainage and lake levels controlled through Lake Fenland or perhaps later through the Humber Gap.

Straw (1979) has proposed two phases for Lake Pickering: a high-level lake at 70 m above OD and a later lower stand at about 45 m above OD.

4.4. Yorkshire Wolds

The earliest reference to ice-dammed lakes in the valleys of the Yorkshire Wolds facing the Holderness coast were made by Carvill Lewis (Lamplugh, 1891) who reportedly pointed out 'there must have been a great accumulation of freshwater in the valleys (of the Wolds) whose mouths were blocked by the North Sea ice. Most of the east-draining Yorkshire valleys would be in this condition'. That damming of the valleys has occurred is detailed by the mapping of de Boer (1994) who recognised a number of channels and lake sites along the lower part of the dip slope of the Wolds where rounded projecting spurs separate valleys. These lake sites and channels mainly extend from Humanby, near the Vale of Pickering to as far south as Walkington at or close to the 200 ft (60 m) contour (Fig. 10), roughly corresponding to the first stage of the ice front during its retreat from the Wolds. De Boer (1944) also linked the main overflow channel formation to the 'Lower Drab clay' (part of the Skipsea Till of Madgett & Catt, 1978, the base of which was assigned an age of 17.5 ± 1.6 ka by Wintle & Catt, 1985) although Lamplugh (1891) has suggested the 'Basement Clay was spread out' at this time.

Of importance to the evolution of the Vale of York is the proposition by de Boer (1944) that the primary lake system, on the east face of the Wolds overflowed through the

Market Weighton Spillway (Fig. 10) into Lake Humber from an intake at between 150 ft – 175 ft (45.7 m – 53.3 m) above OD). It is also of interest to note, that while de Boer (1944) recognised features related to glacial lakes on the eastern side of the Wolds, he did not appear to have recognised similar features on the western side of the Wolds during a later geological investigation between Market Weighton and the Humber (de Boer *et al.*, 1958).

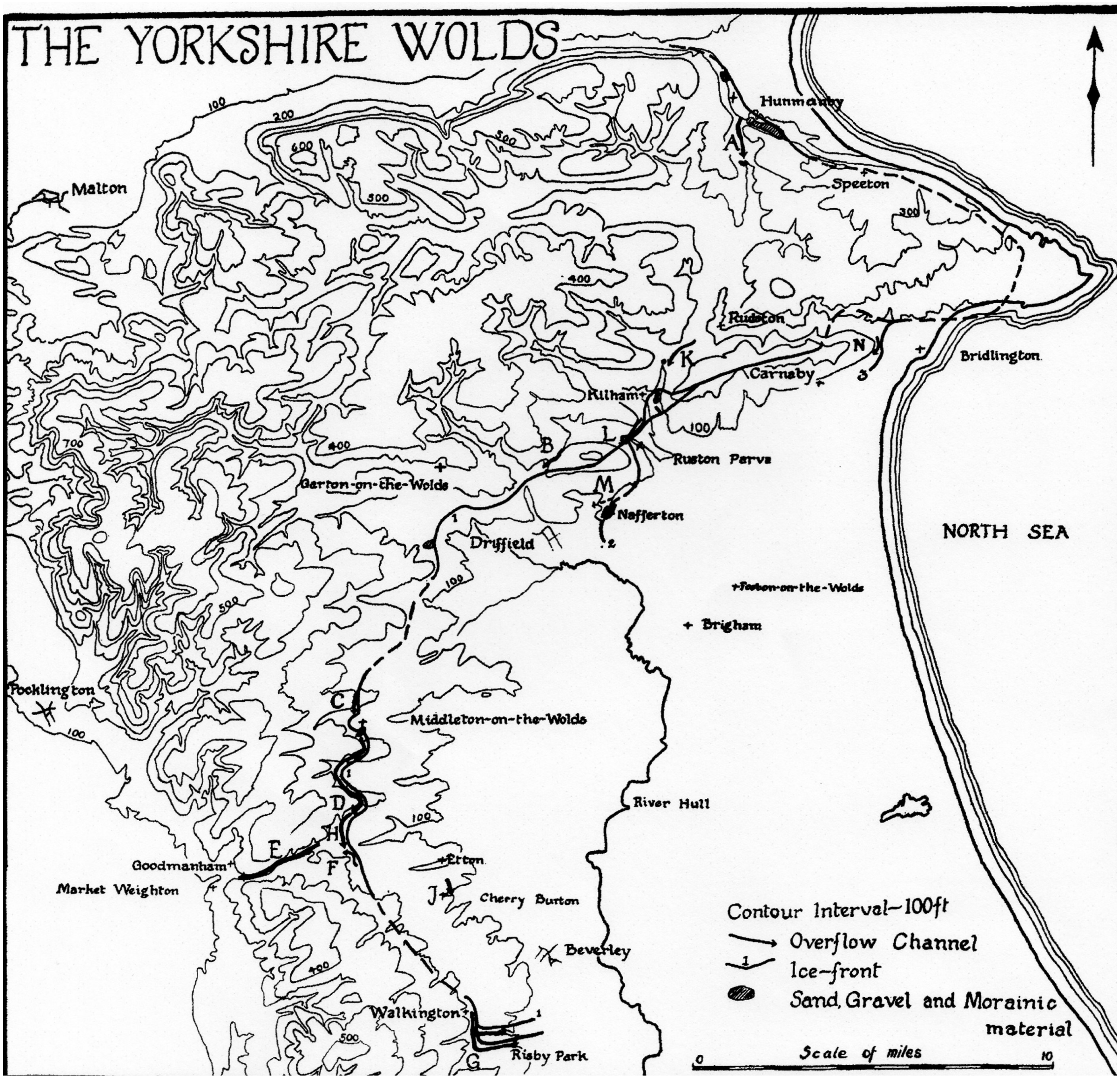


Figure 10. Ice margin features on the eastern flanks of the Yorkshire Wolds from G. de Boer (1944), The Market Weighton spillway is shown to the east of Goodmanham.

4.5. Lake Fenland

Lake Fenland (Straw 1979; Fig. 9), named Lake Ouse by Raistrick (1934), with an extension northwards from the Fenlands into the Ancholme Valley, must have received most of its drainage from Lake Humber through the Lincoln Gap. Evidence for the existence of Lake Fenland is extremely limited as its spatial extent is based on its equalisation through the Lincoln Gap with Lake Humber at *c*. 30 m above OD (Straw, 1979). Strandline features are not widely recognised in Lake Fenland and have only been identified from a narrow bench, at 25 m – 32 m above OD, cut into the western margin of the Fens near Horbling and Bourne (Harrod, 1972; Fig. 9). As referred to earlier, lacustrine sands from the Kelsey pit near Caistor on the eastern side of the Ancholm Valley, at *c*. 30 m above OD, provide a possible date for high-level Lake Fenland. As these sands are overlain by both alluvial fans and aeolian coversand there is little chance of shorelines being preserved. The site does however provide a possible link with high-level lacustrine events in the Vale of York.

4.6. Lake Humber

In his summary of the environmental history and physiographical evolution of the southern part of the Vale of York Gaunt (1976b) recognised thirteen environmental phases between the Holocene and a pre-Ipswichian glaciation. Not all of these phases are relevant to the present text.

Gaunt (1976b) identified two glaciations in the Vale of York, with the older being attributed to the Anglian (MIS 12) or the Wolstonian (MIS 10-6, Catt, 2007). The existence of a Wolstonian glaciation is however controversial and as discussed by Bowen (1999) there appears to be no evidence in East Anglia for glacial deposits intermediate in age between the Hoxnian and the Ipswichian. Merritt *et al*. (2003) have also stated that no Wolstonian terrestrial glacial deposits have been shown unequivocally to occur in Scotland. In contrast, White *et al*. (2010) and Langford (2012) have proposed a Wolstonian (Saalian) MIS 8 glaciation in eastern England with an ice limit that extended as far south as Peterborough. Straw (2011) while supporting an eastern England glaciation in MIS 8 provides evidence for a Saalian ice limit that extends into Norfolk. Irrespective of its precise age the older glaciation in the Vale of York, titled 'Pennine Boulder Clay' by Gaunt (1976b), is presumably equivalent to the older or high altitude gravelly drift and true boulder clay described by Raistrick (1926) from Wensleydale and Swaledale, the first glaciation of Nidderdale (Tillotson, 1934), the older drift of Edwards *et al*. (1950) and the older tills of pre-Devensian age mapped on the British Geological Survey (2008) Selby sheet near Pocklington, which are possibly Anglian in age (Ford *et al*., 2008). In the Vale of York, tills from this older glacial event have been identified on the summit of Brayton Barff (Fig. 7) and in a temporary excavation on the Snaith Ridge near Kellington (Fig. 8). Both these deposits are associated with westerly derived glacial and fluvioglacial sand and gravel (Gaunt, 1994). As with the Late Devensian glaciation, ice for the older glaciation must have originated in northwest England, utilised the Stainmore gap to cross the Pennines, before turning south down the Vale of York. Flowlines for the ice movement are illustrated by Gaunt (1981, fig. 2 and 1994, fig. 37). Although assumptions have been made, the precise age of this older glaciation has not been established. In the Pennines, for instance, Raistrick (1934) has drawn attention to the 'well weathered' appearance of the 'older boulder clay' in comparison to the 'newer clay' as an aspect of age. The only possible age that could be attributed to these older tills is from Ferrybridge where Bateman *et al*. (2008) have dated a diamicton to after 23.3 ± 1.5 ka BP and before 20.5 ± 1.2 ka BP. This dating does not however preclude it from originating during the last glaciation in the Vale of York.

The next most significant feature in the evolution of the Vale of York commenced with the eustatic lowering of sea level by global cooling and ice formation after MIS 5 (*c*. 75 ka BP) and the beginning of the MIS 4-2 glacial (Chiverrell & Thomas, 2010). During this period and prior to glaciation in the Vale of York deep fluvial incision with wide denudation of river valleys took place in response to relative sea level falling to perhaps lower than 19 m below OD (Gaunt, 1976b). Mapping of the base of Devensian deposits by Gaunt (1981, fig. 4 and 1994, fig. 42) has shown that the Ouse valley, which had a more northerly course than at present, was incised to lower than 15 m below OD between Brough and Howden (Gaunt 1981, fig. 4). There was also significant lowering of the Trent and Aire valleys. Infilling of the incised valleys by fluvial and lacustrine processes (e.g. the Hemingbrough Glaciolacustrine Formation, British Geological Survey, 2008) could have occurred either before or following blockage of the Humber mouth to impound Lake Humber (see section 10.1).

The Late Devensian glaciation and high-level lacustrine phase in the Vale of York at *c*. 33 m above OD, according to Gaunt (1976b), commenced with the blocking of the Humber mouth by North Sea ice and morainic material between Brough and Winterton. As explained earlier in the text, the advance of the Lake District ice into high-level Lake Humber was short-lived with the level of Lake Humber soon becoming stabilised at between 9 m and 12 m above OD (Fig. 11): the difference in level being attributed to isostatic depression in the north. This low-level lake is believed to have persisted long after stands of the retreating ice front at the York and Escrick moraines allowing the accumulation of laminated clays and other sediments before the final disappearance of the lake to leave a featureless clay plain. The extent of low-level Lake Humber and the glaciolacustrine deposits have been mapped by Clark *et al*. (2004).

Gaunt (1976b) believed Lake Humber disappeared quickly as a catastrophic or jokulhlaup event (Bateman *et al*., 2007) without leaving regressive shorelines and as Pennine drainage became established there was little incision, with rivers constantly changing their courses and building sandy levees. The importance of alluvium and fluvial sand deposited by these early drainage systems (e.g. Breighton Sand Formation) is considered in this text to be as fundamental in the shaping of the Vale of York as the older lacustrine beds.

The two-stage decline model for Lake Humber (Fig. 11), proposed by Gaunt (1976b), without regressive shorelines, must be considered too simplistic, as he has overlooked or disregarded several recorded examples of such features. Melmore (1940) has described terracing in the Derwent Valley at 15 m above OD and a shoreline at *c*. 25 m above OD near Healaugh, while de Boer *et al*. (1958) and Penny (1974) have mapped terracing (or a strandline) at Goodmanham near Market Weighton.

Friend (2011) in proposing a revised two-stage model for the decline of Lake Humber has reviewed the validity of littoral terraces on the south of the York Moraine (Fairburn, 2009) and has suggested that high-level Lake Humber was coeval with a regressive ice front of the Vale of York glacier north of Escrick.

4.7. Summary

It is apparent that in the evolution of glacial Lakes Humber, Wear and Tees that important terracing has been noted, but not awarded the regional significance it requires to record a likely multi-staged decline in all three lakes. Recording of such events could well influence the status and relative age of other glacial features, associated with the lakes, such as moraines and meltwater channels. It is therefore perhaps regrettable, in the context of this work, that mapping by the British Geological Survey and others has largely concentrated on mapping bedrock geology and the more conspicuous glacial units while minimising the importance of superficial deposits.

5. VALE OF YORK OBJECTIVES

In the Vale of York events that preceded and followed the Late Devensian glaciation, and summarised in earlier sections of this text, have mainly been derived from the many publications of G.D. Gaunt e.g. 1976b and 1994. Shorter summaries by Catt (2007), Evans *et al*. (2005) and Murton *et al*. (2009) have also used data from the same source. The model that has been accepted or partially accepted, which is based largely on lithological mapping, is however flawed, as it cannot explain, has failed to identify or has possibly mis-interpreted many of the erosional and depositional landforms in the Vale of York that are now being recognised (Fairburn, 2009, 2011). The most important of these include:

- terracing on the York and Escrick Moraines, on the Triassic ridges and on the western flanks of the Wolds;
- eskers in the Vale of York, notably the Crockey Hill 'Esker', which may not be sub-glacial deposits but could have originated as alluvial fan deltas;
- the alluvial fans or glaciofluvial fans mapped in the region of Pocklington (Fairburn, 2011; Ford *et al*., 2008) that were previously interpreted as 'Older Littoral Sand and Gravel' (Gaunt, 1981).

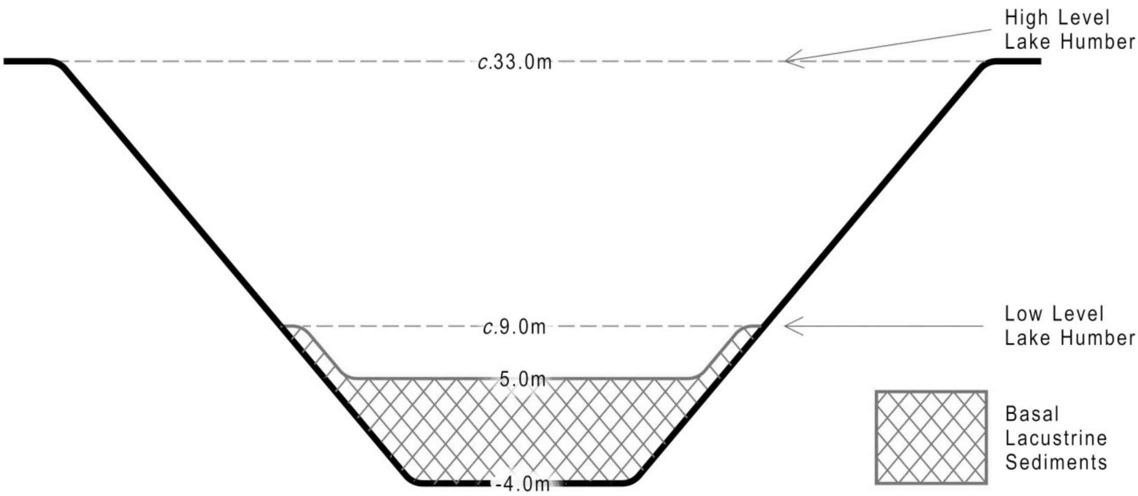


Figure 11. Schematic diagram to illustrate the 2-Stage model for Lake Humber after Gaunt, 1976b, 1994. Lake Humber initially rose to a maximum elevation of *c*. 33 m above OD before falling transitionally to as low as -4.0 m below OD. The lake finally stabilised at a low-level of *c*. 9 m above OD.

The main, primary objective of this research is therefore to present in this Thesis a detailed description of the landforms (including terracing and alluvial fans) in the Vale of York on the York and Escrick moraines, the Triassic ridges and on the flanks of the Wolds that have been mapped in Figures 14 - 17 which will contribute to a revision of the current 2-stage Lake Humber model (Fig. 11) and to include in this model regressive stages.

The description of the new model must attempt to explain, or satisfy, most if not all the controversial issues in the Vale of York that have been discussed over the past 150 years, or since the glacial theory was put on a sound footing, such as the genesis of several gravel deposits that form distinctive morphological features e.g. the Thorne – Wroot Gravels. To achieve this it will be necessary to present and substantiate the completed landform mapping, which covers an area of approximately 1500 km^2 in the Vale of York and on the flanks of the Wolds. For this presentation several approaches are considered necessary to address the issues involved. These are:

- to clearly illustrate, by photography, that the landforms, particularly the terracing, are distinctive features of the landscape;

- describe analogies from elsewhere in the world, as well as the UK, which illustrate that terracing, bounding former glacial lakes, is not unique to the Vale of York;

- use Airborne Light Detection and Ranging (LIDAR) to demonstrate that the mapped landforms can be detected by means other than ground observations.

Such a model will help resolve other well-documented issues that include:-

- the validity of the Crockey Hill 'esker';
- the significance or validity of isostatic depression in the Vale of York;
- the influence exercised by the Skipsea Till (Hessle Till) in the Humber Gap in controlling lake levels;
- the Aire and Calder River terraces;
- the erosional effects of post-glacial, pre-Holocene drainage.

Major secondary objectives (the significance of which are outlined in Part 3) resulting from the revised evolutionary model that need to be addressed for the Vale of York include:

- the implications for Late Devensian and Early Holocene lacustrine and fluvial sequences including the Hemingbrough Glaciolacustrine Formation and the Breighton Sand Formation;
- mammalian chronology;
- a review of the 'east Pennine Suite' of gravels including the Thorne-Wroot Gravels and the Linton-Stutton Gravels.

Whilst landform mapping could prove fundamental in providing the key to the glacial and post-glacial history of the Vale of York, the importance of some older texts containing sections of quarry faces or other excavations that are no longer accessible, cannot be minimised. As stated by Carruthers (1953), 'with relief one remembers that after all the facts gathered with such infinite care, over so many years, are in no wise affected: their permanency is untouched and their value as high as ever'.

Although completion of the landform mapping has been the main objective towards understanding the Late Devensian glacial and post-glacial history in the Vale of York, this work has been augmented by sand sampling from sections in both the York Moraine and the alluvial fans for optically stimulated luminescence (OSL) dating. It is hoped that such dating will place the most recent glacial advance in the Vale of York into context with the Basement, Skipsea and Withernsea Tills at the Dimlington type site; assign the older alluvial fans, if they exist, to an earlier glaciation and provide a more reliable chronology for the glaciofluvial and glaciolacustrine sediments that are either overlain by till between the York and Escrick Moraines or exposed in clay pits south of Escrick. The chronology might also provide a reliable date for the final retreat of the North sea ice lobe from the Holderness coast.

PART 2

6. LANDFORM MAPPING

6.1. Introduction

The Vale of York has evolved as a strike-orientated glacial valley eroded into soft Triassic rocks between the easterly dipping Permian/Carboniferous Pennine escarpment to the west and the easterly dipping Jurassic/Cretaceous escarpment of the Wolds to the east. At least two glacial episodes have been recognised: a pre-Ipswichian glaciation of Wolstonian or Anglian age (Gaunt, 1976b) now thought to be Anglian (British Geological Survey, 2003) and a Late Devensian glaciation referred to as the Dimlington Stadial (Rose, 1985). Before the close of the Late Devensian this deglaciated valley had been transformed into a mostly flat featureless plain, some 30 km wide between Tadcaster and Pocklington, by infilling glaciofluvial and glaciolacustrine sediments. Rising above this plain, which is inclined southerly from about 15 m above OD near York to below 5 m above OD towards the Humber outlet, are ridges of glacial moraine and eroded inliers of Triassic sandstone. On the eastern side of the Vale of York, the Jurassic and Cretaceous rocks are blanketed, between elevations of 15 m – 50 m above OD, by coalescing, terraced, glaciofluvial fans composed mainly of chalk and flint gravel (Fig. 16). These fans may have originated from frost fracturing of Chalk formation (Fairburn, 2011). Because of the different geology similar gravels have not evolved on the western side of the Vale of York although linear mounds of gravel do occur between the villages of Linton and Stutton. These deposits, termed the Linton-Stutton kame belt by Edwards *et al*. (1950) are largely composed of Carboniferous and Permian erratics. Also by the close of the Late Devensian modern drainage was becoming incised into the Pleistocene sediments that underlie the Vale.

The present-day landscape of the Vale of York, the mapping of which has been a primary objective for this research, is therefore the result of at least two glacial events, with their associated tills, and features produced by the development and duration of pro-glacial Lake Humber which led to the deposition of laminated clays and littoral sands and gravels. Marginal to the lake, periglacial effects resulted in the discharge of alluvial fans and gravels. Meltwater channels involved in lacustrine build-up and drainage have also been recorded (section 7.16.3.). Superimposed on this landscape are a series of terraces which it is hypothesised relate to the progressive retreat of Lake

Humber. The results of mapping these terraces, which have been recorded on the moraines and alluvial fans as well as on Permian and Triassic bedrock, along with their likely origin, forms the basis of the next chapter.

It is considered that mapping of the superficial deposits and erosional surfaces, to obtain an understanding of the sequence of events that produced this landscape, cannot be achieved from conventional lithological mapping by auger drilling to shallow bedrock, as the origin and significance of these deposits will either not be recognised or ignored. Because both non-planar and planar landforms may be present, the required mapping should recognise and record all topographic changes in the landscape, so that the geometry of all landforms can be identified and placed into chronological order. In summary, these should include any level or gently inclined planar land surface, which has distinct topographic boundaries resulting from erosional or depositional processes, that can be plotted in the field on the 1: 25 000 Ordnance Survey base map. The most important of these are:

- marginal erosion boundaries along the edge (or edges) of major depositional surfaces, such as the plain of the Vale of York;
- formlines recorded at the point of inflection of erosional terraces with the rising surface of a hillside;
- formlines marking a change of slope forming a visible break on the side of a hill;
- depositional edges produced by sediment terminations, or a down-slope change of gradient resulting from lithological changes, such as sediment fining;
- boundaries of erosion surfaces, particularly those in crestal or near crestal locations on hill tops or ridge lines.

However only those features that can be mapped with certainty and have regional continuity were used to delineate the major landforms shown on Figures 14-17.

Mapping of mainly non-planar glacial landforms was widely used by Clark *et al*. (2012) to model the retreat of the last British-Irish Ice Sheet. These landforms included features of both ice-marginal origin, such as moraines and lateral meltwater channels and those of sub-glacial origin, particularly drumlins and eskers. While remote sensing has greatly facilitated the recognition of sub-glacial bedforms using a dataset that includes a Digital Surface Model (DSM) with a ground resolution (grid size) of 5.0 m and a vertical accuracy 0.5 m to identify minor changes in elevation (e.g. Hughes, 2008), plotting of drumlins and meltwater channels can still be undertaken as a field-mapping procedure (e.g. Mitchell, 2007). Recognition of distinctive landforms in sequences of poorly exposed strata has also been used as an aid to compile regional geological maps when conventional mapping has proved difficult. Fairburn (2004), for example, was able to differentiate poorly exposed Tertiary and Quaternary sequences from the underlying Neoproterozoic by mapping characteristic landforms and identifying subcrop from boreholes, roadcuts and building excavations.

Recognition of planar landforms has also been a part of geological mapping: the most commonly mapped surfaces being the floodplains of major rivers, as shown on Figure 14 flanking the Wharfe, Ouse and Derwent rivers. These surfaces are inclined downstream and rise gently towards the containing hill slopes. Lacustrine terracing, which will be a major topic of the next chapter, differs in being mainly laterally horizontal unless affected by differential isostatic adjustment and will be underlain by mostly coarse littoral sediments rather than river alluvium. Floodplains, which are mainly depositional in origin, are generally more laterally extensive than erosional lacustrine terracing.

6.2. Terracing and Planar Landforms

This section of the text essentially provides the justification for the research that was provided by the regional recognition of planar landforms and terracing in the Vale of York (Fairburn 2009 and 2011) and includes a description of the innovative technique used to illustrate the mapping. Background support for this work is provided by some previous localised accounts of planar landforms in the Vale of York but more importantly by analogues of lacustrine shorelines of pro-glacial lakes elsewhere in the world.

6.2.1. Previous Recognition

Landform mapping and the recognition of terracing on the York and Escrick Moraines led to conclusions that these planar surfaces could be shorelines resulting from stands of Lake Humber in the Late Devensian (Fairburn, 2009). This hypothesis also implied that the Vale of York glacier had deposited the York Moraine and retreated back towards the Pennines before the impounding of Lake Humber; a sequence of events not envisaged by Gaunt (1976b). This mapping, with some important amendments to better depict the full extent of terracing at and above 40 m above OD, has been included in this text as Figure 14. A follow-up program of mapping in the Pocklington area (Fairburn, 2011) showed that two sets of alluvial fans were present flanking the western side of the Wolds. These two sets of fans were differentiated in the field because one set of fans, at a slightly higher elevation, were distinctly terraced and had erosional contacts with the lower level fans, which were not terraced. As the terracing was at comparable elevations to that on the York and Escrick moraines a logical deduction was made that terracing on the older fans could also be shorelines of Lake Humber, with the younger fans forming from fluvial events post-Lake Humber. To confirm that this early date of fluvial events bordering the Wolds, shown on Figure 15, had regional significance, landform mapping was extended south of Pocklington, along the Wolds, towards the Humber (Fig. 16).

While the main objective of this mapping was to confirm that two sets of alluvial fans could be recognised in a wider context, priority was also given to examining the effects of terracing, if any, on the outcropping Skipsea Till near Hessle. As a further priority mapping was also extended south of the Escrick Moraine to determine whether ridges of Triassic sandstone (Sherwood Sandstone Group) that form inliers through the Quaternary cover at Brayton Barff, Hambleton Hough and between Kellingley and Snaith were also flanked by gravel deposits that had been terraced by lacustrine action (Fig. 17).

A further objective of the landform mapping was to determine an understanding of the origin of the plain of the Vale of York. This planar, southerly tilted surface is far from simple in origin and must have evolved from both lacustrine and fluvial events. Continued acceptance of the term 25 Foot Drifts (Edwards, 1937) to describe the sediments underlying the plain (of which this author is also guilty) has perhaps been detrimental in understanding its true origin.

That planar landforms can be recognised within or around the margins of the Vale of York is not a new concept, as several authors have published articles dealing with the identification of these features. All this mapping has however been localised in its extent and there seems to have been little attempt to pursue these objectives on a regional basis. It is perhaps pertinent to point out, that any erosional or depositional features, related to shorelines of Lake Humber, should have regional continuity (provided these are preserved), and irrespective of the timing for the impounding of the glacial lake, significant events imposed on one bank of the lake will be mirrored on the other bank. The significance of this has been overlooked, by the British Geological Survey.

The most important of the early work to be published on shorelines in the Vale of York was the mapping by Edwards (1937) of gravels on a Pleistocene strandline (the 100 Foot Strandline) at or close to the 100 foot contour (30.5 m) on the dip slope of the Permian rocks along the western edge of the Vale of York between Tadcaster and Doncaster (Fig. 20). Edwards considered that the upward limit of the gravels was clearly defined between 100 ft and 110 ft above OD (30.5 m – 33.5 m above OD). The location of these gravels is now referred to as the 33 metre strandline. Melmore (1940) also, in a study of terraces flanking the Ouse and Derwent rivers, recognised terraces above modern river alluvium. The higher 25 foot terrace, (i.e. 25 ft above river alluvium or a little over 15 m above OD), was identified on the right bank of the River Derwent near the Stamford Bridge Viaduct [SE 708 554], while the lower 10 foot terrace (10 ft above river alluvium), at just over 10 m above OD, was recognised on the right bank of the River Ouse between Beningbrough [SE 530 577] and the railway yards at York [SE 581 526]. These two surfaces are shown on Figure 14 as the 20 Metre Surface and the Older River Terrace respectively.

In addition, Melmore (1941) described an 85 ft (25.9 m) surface near Healaugh [SE 493 479], which he regarded as a shoreline at 'an arrest in the draining-off of Lake Humber'. The same beach was also reported by Edwards *et al*. (1950). This shoreline has been mapped as the 25 m strandline at this location (Fig. 14) and as the 25 m terrace on the Escrick Moraine near High Catton (Fig. 15). Both de Boer *et al*. (1958) and Gaunt *et al*. (1992) have described *Gryphaea* gravels from pits [SE 904 511] west of South Cave associated with cross-bedded sand and gravel. These gravels lie on a surface between 15 m and 20 m above OD (20 Metre Surface, Fig. 16) and must form part of a littoral deposit over shallow sub-cropping Liassic strata. De Boer *et al*. (1958, Pl. 11) also describe cross-bedded gravels forming a 'storm beach' at 50 ft above OD (15 Metre Surface) on Cave Oolite at Elloughton Quarry South [SE 932 286].

Well-established shorelines that have been mapped in other glaciated regions of the world where pro-glacial lakes have been impounded include the 'Parallel Roads' of

Glen Roy and the strandlines of Lake Agassiz, which formed during the retreat of the Laurentide Ice Sheet in North America.

6.2.1.1. *Glen Roy*

That the 'Roads' at Glen Roy are in fact lacustrine shorelines resulting from the impounding of a glacial lake in Glen Roy by an ice advance from Ben Nevis during the Loch Lomond Stadial (Fig. 4), was first suggested by Agassiz during a tour of Scotland with William Buckland in 1840. This assumption, by Agassiz, was later elaborated on by Jamieson (1863) who identified the overspill cols in Glen Roy that controlled the ice-dammed lake levels. Detailed investigations of the Late Quaternary history of Glen Roy and its vicinity by Sissons (e.g. Sissons, 1978) has been summarised by Lowe *et al.* (2008). These authors explain how rising lake levels, controlled by spillways to connecting valleys, produced visible shorelines at 260 m, 325 m and 350 m in Glen Roy and a fourth shoreline at 355 m in the adjacent Glen Gloy (Glen Gluoy). This 'rising sequence' was followed by a 'falling sequence' as the shorelines were revisited by falling lake levels during termination of the Loch Lomond Advance. The imprint of the 'Parallel Roads' is clearly shown on Plate 12.

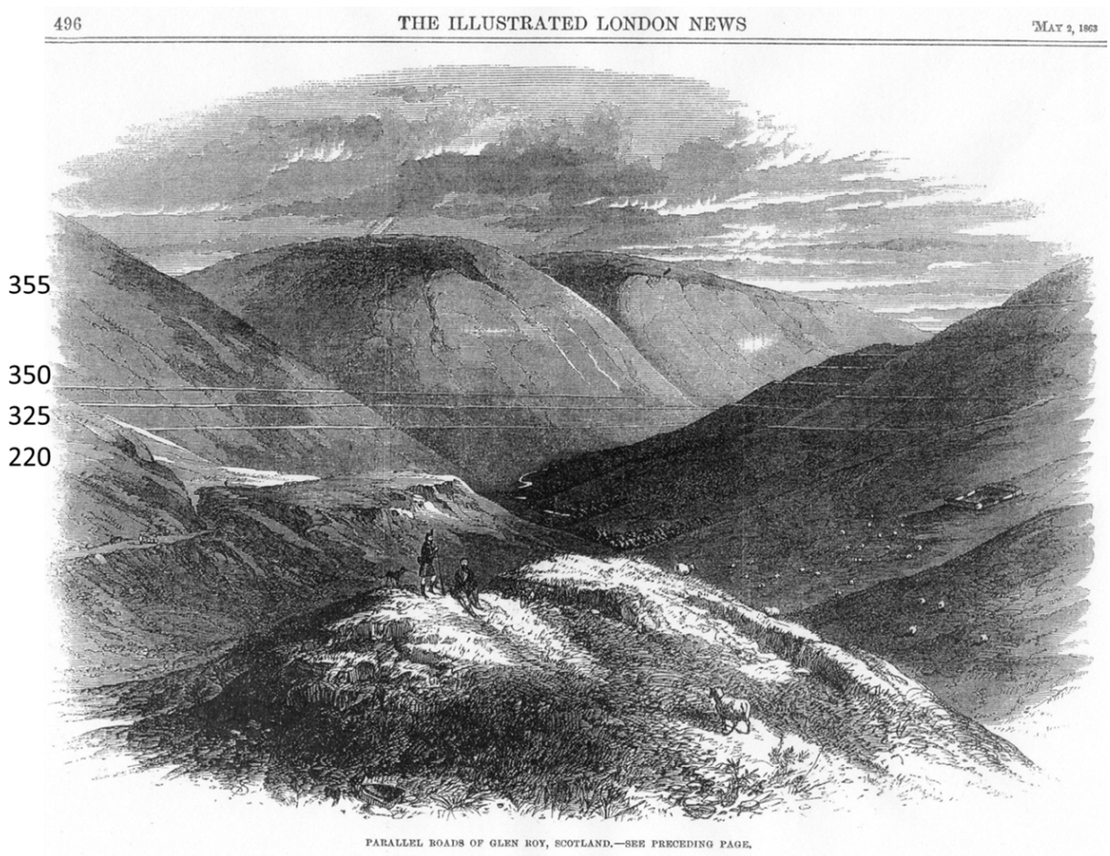


Plate 12. Engraving of the parallel roads of Glen Roy from the Illustrated London News, 1863.

Mapping of the 'Roads', shown by Lowe *et al.* (2008, fig. 2) is based on the width of the shoreline (*c*. 10 m) between the eroded valley back-wall and the depositional front edge. Sissons (1978) has attributed the formation of the 'Roads' to both frost and wave action, with the profile of the 'Road' being concave at the back-wall and convex at the front edge. Minor eastward tilting of the 'Roads', that was recognised by Jamieson (1863) at about 0.19 m per km, has been explained in terms of a glacio-isostatic event by Dawson & Hampton (2008) who estimated a tilt of between 0.11 and 0.14 m per km.

6.2.1.2. *Lake Agassiz*

Lake Agassiz, North America's largest pro-glacial lake, reached an areal extent of 260 000 km^2 about 9.4 ka ^{14}C yr BP during the retreat of the Laurentide Ice Sheet (Teller, 2001). Beaches and wave-trimmed cliffs formed around Lake Agassiz and more than 50 strandlines have been identified (Elson, 1967) all of which rise in elevation towards the northeast as a result of isostatic rebound (Fig. 12). Both transgressive (the most widespread) and regressive beaches have been recognised due to the changes in lake geometry caused by changing outlets of the lake (Teller, 2001). Over 4 000 years of lake history were recorded by optically stimulated luminescence dating of strandlines (Lepper *et al*., 2011) originating between 14 ka - 10 ka BP.

In Lake Humber, the 100 Foot strandline of Edwards (1937), on the flanks of the Permian escarpment, appears to be regressive and erosional in nature, as a result of a receding lake level: in this environment a transgressive shoreline could only form if the culmination of topographic features were at similar elevations to lake levels.

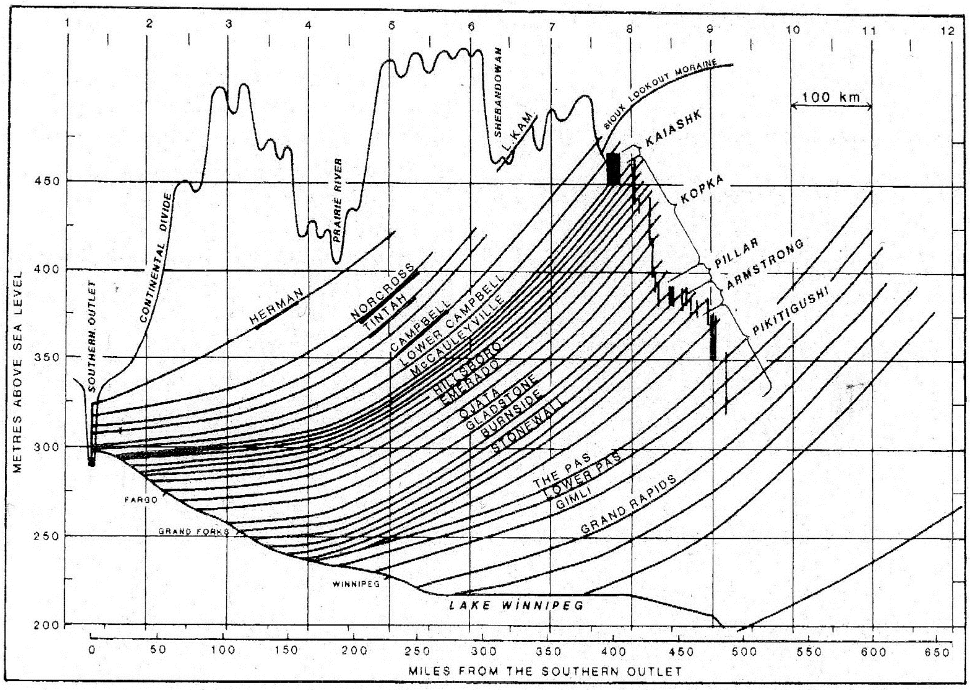


Figure 12. Regressive and transgressive shorelines developed around Lake Agassiz during the retreat of the Laurentide Ice Sheet. The shorelines rise from a south-western outlet towards the northeast due to isostatic rebound. From Teller, 2001.

6.2.2. Mapping Methods

Terracing, mapped on the York and Escrick Moraines and along the eastern edge of the Vale of York, interpreted as shorelines by Fairburn (2009 and 2011), can be described in terms of three components. These are an erosional concave back-wall at the strandline and a depositional convex front edge, separated by a mainly horizontal terrace (Fig. 13). In contrast to Glen Roy (Plate 12), the terraces of the Vale of York are not such conspicuous features, and because they are mainly imprinted in unconsolidated sand or gravel, their permanency has been less secure. Because of this, accurate mapping to define a particular terrace is not possible as the precise location of the depositional front edge is either difficult or impossible to locate due to the convex curvature of the land surface, as shown on Figure 13. In contrast, the back-wall, or presumed strandline, of the terrace can usually be accurately mapped on the 1: 25 000 Ordnance Survey maps because the point of inflection between the level terrace and the rising back-wall is a visible feature and for this reason is often marked by a bottom-slope drainage ditch, or field-boundary hedgerow. Consequently as the back-wall, or strandline, can be plotted on the field map with a vertical accuracy of *c.* 1.0 m in most cases, it was decided to use, as a mapping unit, the interval between two successive strandlines i.e. the interval between ‘C’ and ‘F’ on Figure 13. This interval has been referred to as a ‘Surface’ with a metric numerical prefix e.g. the C-F interval, on Figure

13, is the 33 Metre Surface. If the Glen Roy 'Roads' had been mapped in the same manner then the 350 Metre Surface would cover that part of the hill slope between the 350 m shoreline and the 325 m shoreline (see Plate 12). Whilst it will be expedient to map the 'Surfaces' as the intervals between recognisable strandlines or shorelines, for reasons stated above, it is nevertheless more appropriate to describe the landforms in terms of their main components, that is terracing and strandlines and formlines. Whilst this terminology may be fully self-explanatory some amplification is required for the text as follows:

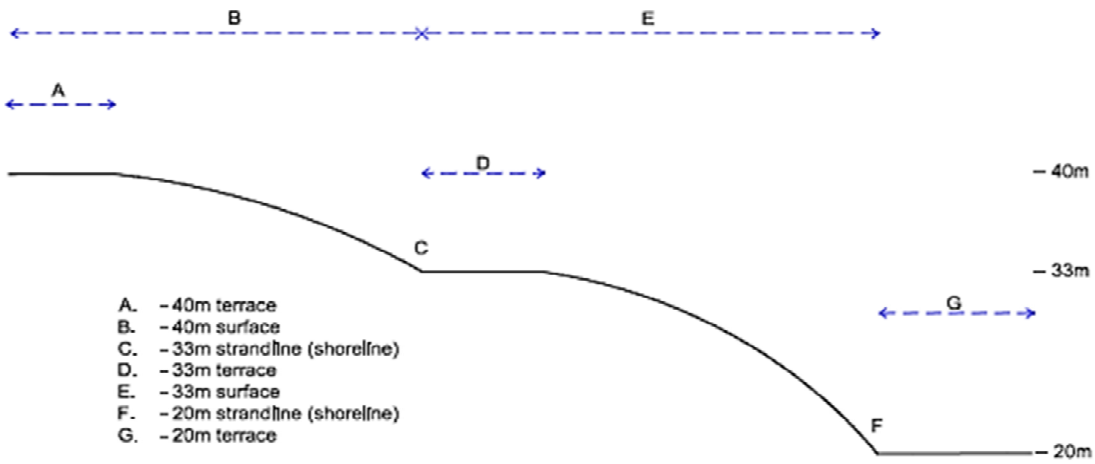


Figure 13. Diagram to illustrate the components of the landforms used in mapping the regressive shorelines of Lake Humber.

Terracing Terracing is applied here to generally laterally horizontal planar landforms, which may be gently inclined into or away from the hill slope. Some terraces are only tens of metres wide but in places they can by up to a kilometre in width. Generally the surface of the terrace is not laterally inclined.

Strandlines These have been selected at the point of inflection where the flat erosional terrace forms a marked boundary against the rising surface of a hillside. Some sections (included later in this text) suggest that the mapped strandline has been slightly elevated by later depositional events on the terrace.

Formlines Formlines were mapped to mark any change of slope forming a visible break on the side of a hill. These may be erosional, such as minor strandlines, or may be caused by down-slope gradient variations resulting from lithological changes, such as sediment fining.

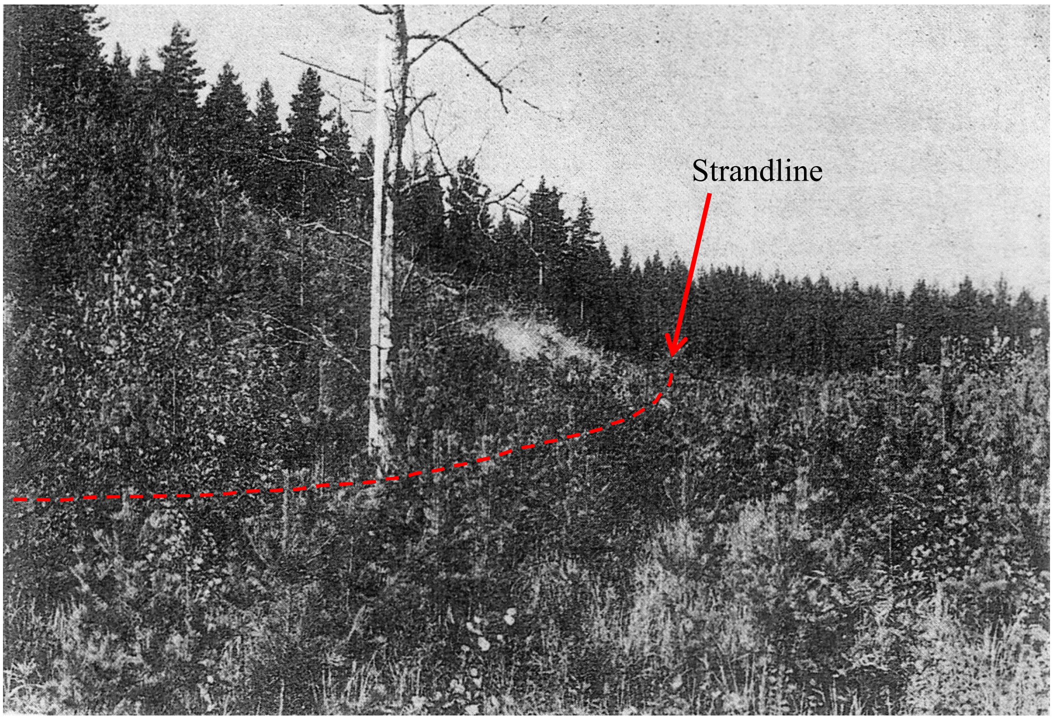
Surfaces The term surface, with a metric numerical prefix, is given to that part of a hillside between two strandlines. For instance the 33 Metre Surface is the interval between the 33 m strandline and the 20 m strandline (Fig. 13). The other mapped surfaces are indicated in the legends of the main maps (Figs. 14 – 17). Because the 25 m strandline is not everywhere a mappable event it has been incorporated into a single surface between the 20 m strandline and the 33 m strandline.

6.2.3. Analogues

While good analogues to the terracing around Lake Humber, which is depicted on Figures 14 – 17 in this text, is provided by the ‘Parallel Roads’ of Glen Roy, and the strandlines of Lake Agassiz, the best analogue of terracing with distinct erosional back-wall scarps is provided by the ancient shorelines of the Saimaa Lake complex in Finland.

6.2.3.1. *Saimaa Lake*

In southern Finland, the Late Weichselian (Late Devensian) and Flandrian history of the Saimaa Lake complex has been studied by Saarnisto (1970). Here ancient shorelines resulting from the retreat of the Baltic Ice Lake were elucidated by determining lake elevations from erosional shores, backed by shore cliffs (Plates 13a and 13b), and lake margin sandur-deltas.



Plates 13a.

Plates 13a and 13b Erosional shorelines or beach terraces at Lake Saimaa terminating against a strandline at the base of a scarp forming the back-wall of the terrace. From Saanisto, 1970

In most instances, the shorelines were eroded into glaciofluvial sediments with the highest beach being referred to as the washing limit; above the washing limit bedrock is usually overlain by till. It was noted, in some cases, that the level of the marginal delta had been lowered by later abrasive action. Although the overall genesis of the numerous lakes in the Saimaa Lake complex differs from Lake Humber certain similarities do occur and comparisons can be made between beach levels at Lake Saimaa and terracing around Lake Humber, as both are backed by scarps or cliffs as illustrated on Plates 13a and 13b.

A major difference between Lake Saimaa and Lake Humber is that the Lake Humber terraces are horizontal, while the Lake Saimaa shorelines are tilted due to isostatic re-emergence. The land uplift at Lake Saimaa has been estimated at between 3 – 9 mm/yr (Saarnisto, 1970, fig. 3).

7. LANDFORM MAPPING RESULTS

This section of the text mainly describes the fundamental landform mapping that was acquired over an area of nearly 1200 km^2, in the Vale of York, that has provided the basis for a re-interpretation of its physiographic evolution since the Last Glacial Maximum. The proposed model, based on shoreline recognition, requires that Lake Humber was only impounded after substantial retreat of the Vale of York Glacier back to the Pennines and that it subsequently declined from 42 m above OD, down to 5.0 m above OD. During this decline, existing glacial landforms, and older alluvial fans below a 52 m strandline, were either modified or overprinted by younger landforms. The section also discusses some of the major implications, for other aspects of Vale of York physiography, that have evolved from this new interpretation such as isostasy and a re-interpretation of terracing in the Aire and Calder river valleys.

7.1. 52 Metre Strandline

De Boer *et al.* (1958) recorded a 200 ft (61 m) above OD terrace at Goodmanham, northeast of Market Weighton, which they were able to trace along the escarpment as far as South Cave. It is believed however that a distinctive terrace between about 50 m and 55 m above OD, at Goodmanham, recognised during the landform mapping, which extends between the outcropping Chalk and the surface of the flanking alluvial fans, is equivalent to the de Boer *et al*. (1958) terrace. In a later publication Penny (1974) changed the description of this terrace to that of a 52 m strandline; a level recorded as a shoreline of Lake Humber by Thomas (1999). This boundary, shown on Figure 16, was also mapped on the Beverley 1:50 000 sheet (British Geological Survey, 1995) as the boundary between Chalk formation and the underlying Jurassic Redcar Mudstone Group. A significant feature of this terrace is that it forms the upper level of fluvial discharge from Middlethorpe Dale and Goodmanham Dale; the latter being the Market Weighton Spillway (Fig. 10) considered by de Boer *et al*. (1958) and de Boer (1944) to be an overflow channel from the eastern side of the Wolds into the Vale of York.

Although de Boer *et al*. (1958) did recognise a southerly extension of their 200 ft terrace to South Cave and to the dale northwest of Elloughton, the feature was not shown on their geological map of the area between Market Weighton and the Humber (de Boer *et al*., P1. 12), which shows only the Jurassic and Cretaceous solid geology at heights above alluvium of the Vale of York. In contrast, the more recent landform mapping, of a similar area (Fig. 16), has concentrated on the importance of this feature as part of the glacial geology of the Vale of York.

During the new work, done as part of this research, planar features such as terraces, were not mapped as separate landforms. This was based on the realisation that if the terrace represents a lacustrine shoreline then the only distinct boundary that can be recorded, that may have lateral continuity, is the eroded back-wall of the terrace against the rising hillside. For this reason the 200 ft terrace of de Boer *et al*. (1958) has been mapped only as a strandline between outcropping Jurassic or Cretaceous strata and the down-slope alluvial fans. The name 52 m strandline has been applied to the terrace as it occurs at this elevation near Goodmanham where it was first recorded by Penny (1974). It has been accepted as a strandline based on its lateral continuity, its extension from Goodmanham across various Mesozoic stratigraphic levels (so that it is not a bedrock lithological boundary) and that it lies on the upper surface of a sequence of alluvial fans.

South of Goodmanham, the 52 m strandline stays at this elevation as far south as South Newbald before gradually falling to 45 m above OD east of Everthorpe Hill and then 40 m above OD between South Cave and Brantingham (Fig. 16). From Brantingham to just west of Hessle the strandline remains at the 40 m above OD level where it merges with the 40 Metre Surface of the Vale of York. As the mapping characteristics that distinguish the strandline remain constant between Goodmanham and Brantingham, and the feature is not controlled by a regional bedrock dip, it is concluded that the inclination of the strandline is an isostatic effect. That the strandline falls to about 40 m above OD to merge with the proposed maximum level of Lake Humber adds weight to an isostatic influence, as this may be the similar level of an older lake.

Not unexpectedly there may be some apparent minor deviations from the general southerly inclination of the strandline, particularly in areas where the outcrop boundary is poorly defined or where there is a lack of distinctive ‘float’. As the mapping was however based on the best field interpretation, any reinterpretation of the mapping has been avoided as much as possible, despite a now presumed isostatic influence. The only major problem encountered during mapping of the 52 m strandline occurred in the region of Everthorpe Quarry, where a spur of Cave Oolite projects westwards towards the gravel pits at Everthorpe (Fig. 16). Here, quarry sections, illustrated by Stather (1922), show a mass of Cave Oolite, up to 1.2 m thick and occupying an area of some

150 m by 50 m, which has been displaced. This is underlain by a thin layer of till and younger Upper Estuarine sands (Gaunt 1976b) and overlain by a thin superficial cover of chalk and flint gravel. Prior to a review of the Stather article, it was considered incorrectly, from field mapping, that the superficial colluvium on the crest of the spur, was part of the older alluvial fan sequence. The final mapping was later modified to place the upper boundary of the alluvial fan sequence to the west of the Everthorpe quarry, at *c*. 45m above OD, with the Cave Oolite erratic raft included as part of the pre-Devensian outcrop (Fig. 16).

[3]

Gaunt (1976b) concluded that the displaced mass of Cave Oolite could only have reached its present location from ice moving over the surface or by rafting on floating ice. The former is the most likely explanation, as the location is above the 'washing line' of the 52 m strandline. Whatever the cause, the event would have to be pre-Devensian in age.

7.2. 42 Metre Terrace (Stage 1)

Evidence for a 42 m terrace, in the Vale of York, marking a Stage 1 maximum elevation for Lake Humber, is rather scant and is largely based on a cobble-strewn level surface, above the 40 m terrace on the York Moraine around Bilbrough (Fig. 14). It is also represented by gravel deposits mapped along the crest of the moraine between Gate Helmsley and Sand Hutton, above the 40 Metre Surface (Fig. 14; see also mapping on the York 1: 50 000 sheet by the British Geological Survey, 1983b). That shoreline washing may have occurred, above the 40 Metre Surface on the York Moraine, is indicated by sand and boulder beds, on the southern flanks of the moraine, in a road-cut east of Bilbrough [SE 538 468], unfortunately no longer visible.

Similarly in the Wolds, evidence for a 42 m terrace is also meagre and is restricted to only a few occurrences near Mask Hall [SE 893 402] (Fig. 40), where terracing, at this elevation, has been imprinted on to sand deposits of the Older Alluvial Fans below the 52 m strandline (see section 7.3.3.). Elsewhere in the Wolds, topographic detail has not been precise enough to differentiate a 42 m terrace from the more obvious 40 m terrace, for example near Hessle.

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[3] Figure 14 (next page).
Geomorphological map of the York and Escrick Moraines showing landforms that developed between shorelines during the retreat of Lake Humber.

It has been concluded that there is sound evidence for a 42 m above OD lake level for Lake Humber and where this coincides with the crest of the York Moraine, as at Bilbrough, a transgressive shoreline was produced.

7.3. 40 Metre Terrace (Stage 2)

Following the mapping by Edwards (1937) of the 100 Foot Strandline, on the western side of the Vale of York, all subsequent accounts of Lake Humber have presumed a maximum elevation for the lake of *c*. 30 m above OD (e.g. Clark *et al*., 2004; Gaunt, 1994). An alternative lake elevation of *c*. 40 m above OD has however been suggested by Fairburn (2009), based on landform mapping near Bilbrough and Sand Hutton on the York Moraine, some 7.0 m higher than the level proposed by Gaunt (1974). More recent landform mapping, over the York Moraine and the flanks of the Wolds south of Market Weighton, has shown there is now abundant evidence to define a 40 m strandline and more localised evidence for a 42 m terrace wherever ground elevations rise to over 40 m and a 40 Metre Surface (defined between the 40 m and 33 m strandlines) has been mapped in both these regions (Figs 14 and 16). At present there is no evidence for a 40 m terrace on the western side of the Vale of York; a situation either due to inadequate observation or failure of only a short-lived shoreline to imprint a strandline on the more resistant Magnesian Limestone. There is however evidence for a 40 m strandline on Brayton Barff (Fig. 17).

7.3.1. York Moraine

The 40 m terrace at the top of the 40 Metre Surface (Fig. 14), forms a most conspicuous feature on the York Moraine at Bilbrough and an almost continuous terrace extends around the summit of Bilbrough top. A small scarp rising above the terrace to a level surface at 42 m above OD is most clearly seen on Redhill Field Road and to the west of the road, where the scarp is underlain by Triassic sandstone (Plate 14), and along most of the northern side of the hill (Plate 15).

Plate 14. The scarp above the 40 m strandline [SE 532 462] trending south-westerly from Red Hill Field Road southeast of Bilbrough. At this location the 40 m strandline has been eroded into Triassic sandstone. (Photograph. M. D. Bateman)

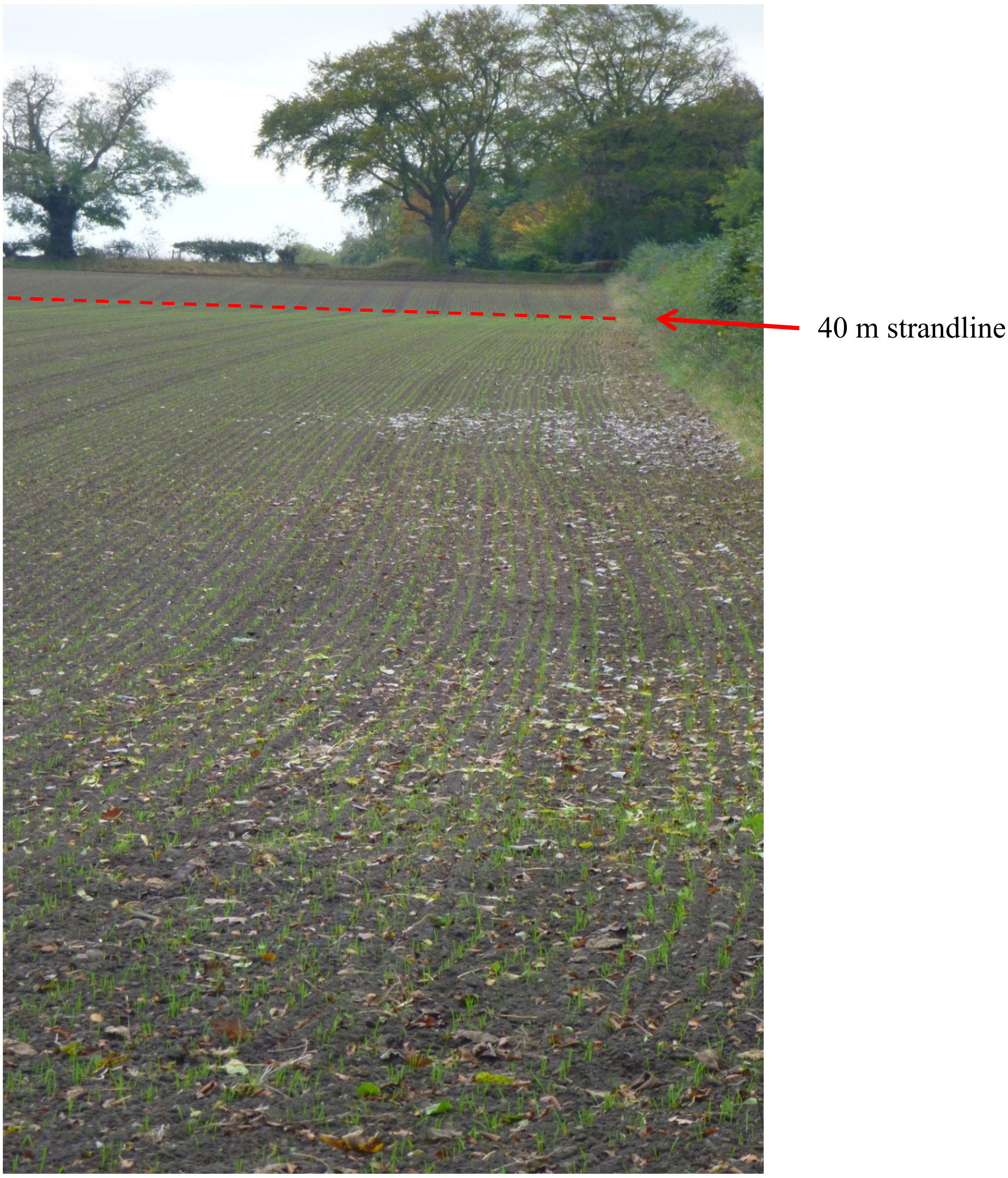


Plate 15. The 40 m strandline north of Bilbrough [SE 533 470], the low scarp above the strandline rises to a level surface on the York Moraine at *c*. 42m. (Photograph. M. D. Bateman)

Much of the hilltop above the 40 m terrace at Bilbrough, is a planar surface strewn with abundant boulders and cobbles, which must have resulted from washing by a transgressive shoreline near the crest of the moraine at *c*. 42 m and segregation of the till to leave a layer of residual gravel and boulders (Plate 6) and sand wash on the southern slope of the moraine (Plate 8). This lag deposit is evidently quite thin, as it is not conspicuous in a 2.0 m section of till overlying Triassic sandstone, above the 40 m terrace, along a valley south of Bilbrough (Plate 3).

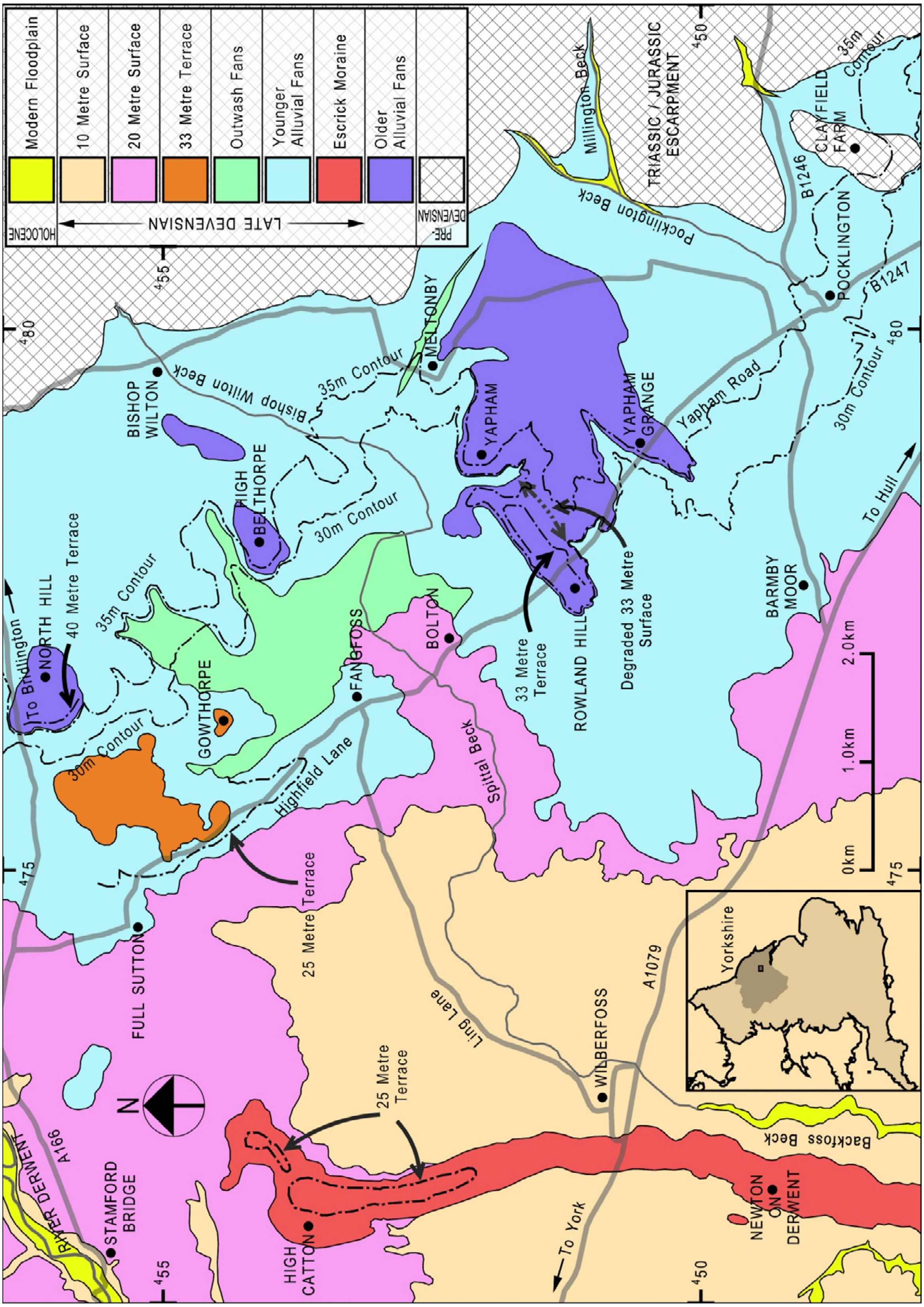


Figure 15. Landforms in the region of Pocklington resulting from the interaction of prograding alluvial fans with the receding shorelines of Lake Humber.

The 40 m terrace is also present along the crest of the York Moraine between Holtby and Sand Hutton, where it is again overlain in places by deposits of gravel (Fig. 14; British Geological Survey, 1983b). Near Long Marston, where the York Moraine rises to over 45 m above OD (Fig. 14), a 40 m terrace is not apparent. Because of the lack of a 40 m terrace, the 40 Metre Surface has here been extended from *c*. 30m to about *c*. 45 m above OD, a situation that does not conform with mapping elsewhere on Figure 14.

7.3.2. The Wolds

Near Pocklington (Fig. 15) the 40 metre terrace is not a distinctive landform and it is only indicated by localised terracing near the 40 m contour at North Hill [SE 764 560] and above Yapham Grange [SE 791 505], and by an erosional surface underlain by Triassic marl near Clayfield Farm [SE 816 485].

Further south, the 40 m terrace forms a conspicuous level surface composed of beach sand in the region of Houghton Hall [SE 886 392] and Houghton Moor, west and southwest of Sancton (Fig. 16). The surface of the beach which rises gently to the east from 35 m – 40 m above OD, forms a border up to 750 m wide fringing the rising surface of the older alluvial fan deposits to the east and the Jurassic ridge east of North Cliffe and South Cliffe to the west (the Cliffe ridge). Characterised by well-sorted, medium to fine-grained sand composed of sub-rounded to rounded grains, typical of a beach environment (Steele & Baird, 1968; Imhansoloeva *et al*., 2011), with only occasional small nodules of flint and chalk, the beach appears to have been reworked from the older alluvial fans along a shoreline of high-level Lake Humber at *c*. 40 m above OD. As described earlier, the fluvial sands forming the fans, from which the beach sand was derived, probably originated from Middlethorpe and Goodmanham Dales. The 40 m terrace also forms a small feature on the road between South Newbald and Hotham on the south side of Hotham Creek. Here a small scarp is present above the 40 m strandline (Plate 19) rising to about 42 m above OD. A 42 m level was also noted above the 40 m terrace near Mask Hall [SE 893 402].

To the east of South Cave the 40 m terrace appears to merge with the 52 m strandline (now at a lower level) suggesting that at least two glacial lakes may have occupied the Vale of York, both with maximum shoreline elevations reaching *c*. 40 m above OD (discounting any isostatic effects).

7.3.3. Hotham Valley

The Hotham Valley, while being a part of the Vale of York and hence very much a part of Lake Humber, forms a distinct open-ended, sediment-filled, flat-bottomed valley between the Cliffe ridge and the Cretaceous chalk escarpment east of Sancton and North Newbald (Fig. 16). It extends southward from Houghton Moor, through Hotham to North Cave, for a distance of 5.0 km. Understanding the erosional and depositional history of the valley which can be considered as a mini Vale of York, can perhaps provide the key to understanding much of the Late Devensian (and some pre-Late Devensian) physiographic evolution of the Vale of York. Events leading to the present geomorphology of the valley, although partly speculative, can be accounted for by two glacial cycles. These two cycles include the last glaciation in the Vale of York and an undated previous glacial cycle probably between MIS 12 (Anglian) and the very early part of MIS 2 (Episode GS – 2c in the Greenland ice core, Fig. 4). Unlike the more recent glaciation, the earlier and much larger ice sheet resulted in differential isostatic depression as evidenced by the changing elevation of the 52 m strandline which reached a maximum elevation of 52 m above OD from its point of greatest depression. Release of sediment into an older proglacial lake from Middlethorpe Dale, the Market Weighton spillway and the now dry valleys east of Sancton and North Newbald produced extensive alluvial fans on the flanks of the Wolds, prior to any isostatic change, that were either initially sub-aqueous or became submerged by rising lake levels to a bounding shoreline at an elevation equivalent to the Market Weighton spillway i.e. the 52 m strandline following later isostatic rebound. These fans may well have been terraced at lower levels during retreat stages of the lake.

[4]

Following the last glaciation, based on evidence principally from the York Moraine at Bilbrough (Fig. 14), the more recent Lake Humber rose to at least 42 m above OD, a process that would have obliterated any terracing on the alluvial fans up to this level and produced the 'beach' levels near Houghton Moor. During retreat of the lake the alluvial fans would have been resculptured to their current geomorphology, which includes a stepping-down of planar surfaces from 42 m above OD at Houghton Moor to below 15 m above OD near North Cave. The erosion in this arm of the Vale of York could well have been assisted by fluvial drainage from the Wolds. There is also evidence from

[4] Figure 16 (next page)
Geomorphological map along the western face of the Wolds between Market Weighton and Hessle showing the imprint of Lake Humber shorelines on the Skipsea Till and the Older Alluvial Fans.

OSL dating (section 9) of renewed fluvial activity in the Hotham Valley during the Holocene.

In contrast to the above, mapping by the British Geological Survey on the Beverley 1:50 000 sheet (British Geological Survey, 1995) indicates that the sand deposits in the Hotham Valley originated as 'blown sand'. While aeolian processes have deposited very fine sand and silt over the Cliffe ridge and parts of the Chalk escarpment that had been winnowed from the coarser fluvial sands within the valley, these deposits are quite distinctive because of their fine particle size and their location as a thin veneer (*c*. 5.0 cm) on the valley sides above any influence from fluvial processes. Some of the thinking by the British Geological Survey, on the nature of the Hotham valley sand, may have been influenced by a reference to 'blown sand' at Houghton Moor by de Boer *et al*. (1958). These authors probably sourced their data from Dakyns *et al*. (1886) who refer to a 'vast deposit of sand' that was blown from the plain to the west. The deposit was also referred to by Melmore (1934) who did not mention blown sand. Two sand samples collected for OSL dating in the Houghton Moor area of the Hotham Valley (Shfd11112 and Shfd11113) were also subjected to particle size analysis (Appendix). The results indicate a unimodal well-sorted distribution for both samples suggesting derivation from a beach deposit (cf. Imhansoloeva *et al*., 2011).

7.3.4. Brayton Barff

Brayton Barff, which is a multi-tiered, isolated hill of Triassic sandstone, located southwest of Selby (Fig. 17), is capped by laminated clay and till (Fig. 7) equated by Gaunt (1976b) with the 'Pennine Boulder Clay', which is presumed by Gaunt (1976b), to be of pre-Devensian age. The mapped boundary of the till with the underlying Triassic sandstone is marked by a steepening of the hill slope between 40 and 45 m above OD. Evidence that this boundary could be a shoreline of Lake Humber is provided by a gently inclined surface, on the southwest side of the hill that extends from the topographic break at 40 m above OD down to nearly 30 m above OD. This surface, composed of reworked Triassic sandstone strewn with rounded quartz pebbles may be a littoral apron preserved on a shoulder of the hill, but without a back-wall for the 33 m terrace.

7.4. 33 Metre Terrace (Stage 3)

Since the recognition and mapping of the 100 Foot Strandline by Edwards (1937), along the western side of the Vale of York from Tadcaster to south of Ferrybridge (Fig. 20) mapping of Lake Humber, at its maximum extent, has been based on a shoreline of *c*. 30 m (Straw, 1979; Clark *et al*., 2004). Despite this, recognition of a *c*. 30 m shoreline (with a high level up to 33 m above OD, Gaunt, 1974) elsewhere in the Vale of York is lacking, even though gravel deposits composed largely of worn pebbles of chalk and flint, at about this elevation, have been well documented on the eastern side of the Vale of York at Everthorpe and Mill Hill near Elloughton (Fig. 16). The Everthorpe gravels, which occur at elevations between 30 m and 35 m above OD have been described by Dakyns *et al*. (1886), Edwards (1937) and de Boer *et al*. (1958). Gaunt (1976b), however, while considering that this deposit could be 'older littoral sand and gravel' (i.e. gravels on the 100 Foot Strandline), thought that they were more likely to be of pre-Devensian, fluvio-glacial origin. This assumption was partly based on the absence of locally derived pebbles of Liassic rocks. At Mill Hill, near Elloughton, similar gravels forming the upper part of a sequence, that rises to an elevation of 33 m above OD, were described by Lamplugh (1887), Sheppard (1897) and de Boer *et al*. (1958). The significance of this deposit, as a beach of high-level Lake Humber, was dismissed by Gaunt (1976b) as an 'altimetric coincidence'. Both the Everthorpe and Mill Hill gravels will be described more fully later in this text (section 7.4.2.).

[5]

In contrast, Fairburn (2009, 2011) reported a shoreline at *c*. 30 m above OD in the Vale of York on the York Moraine (Fig. 14) and near Pocklington (Fig. 15). Follow-up mapping south of Pocklington towards Hessle and on Brayton Barff (Figs 16 and 17) confirmed the persistence of this shoreline over an extensive part of the Vale of York. The shoreline (the 33 m terrace), for this text, includes the gravel deposits at Everthorpe and Mill Hill.

7.4.1. York Moraine

That the 33 m shoreline, or Stage 3 of the regressive decline model for Lake Humber (Fig. 52) was a prolonged event is well displayed by extensive planar surfaces on the York Moraine, near York University and at Bilbrough (Fig. 14). Near York University, at Mill Mound, the 33 m terrace is a gravel and sand covered surface, underlain by till,

[5] Figure 17 (next page).
Geomorphological map in the region of Selby showing Lake Humber terraces developed on the eastern edge of the Permian escarpment and on the Brayton Barff and Snaith Triassic ridges.

that extends down-slope to *c*. 30 m above OD where the surface is replaced by the generally steeper sand-draped slope of the moraine (Plate 20). The gravels are not well exposed and are mainly recognised in roadcuts through the 33 m terrace that provide access to the University (Plate 16), while the sand drape has been extensively burrowed, in place, by rabbits (Plate 9).

Plate 16. Sandy gravel in a roadcut below Mill Mound at York University near 30 m contour [SE 621 508]. The largest pebbles are between 6 cm – 8 cm: most are less than 5 cm. (Photograph. W.A. Fairburn)

A 33m strandline and terrace can be identified at the eastern end of Mill Mound (Plate 17), but its location is poorly defined possibly by the dumping of excavated sand and gravel.

33m strandline

Plate 17. The 33 m terrace on Mill Mound that rises from a poorly-defined strandline by 2.0 m to Siward's How at 35 m above OD [SE 622 500]. (Photograph. W.A. Fairburn)

The mapped 33 Metre Surface of the moraine, that extends down to the 20 m terrace is mainly sand covered except where some lag boulders are present at or near the 25 m terrace (section 7.7.).

On the western side of York, in the region of Severs Howe (Plate 4), several mounds of gravel covered till rise quite steeply above a 30 m above OD base level towards 33 m above OD (Fig. 14). These mounds, as with Mill Mound, are essentially remnants of the 33 m terrace standing above the sand draped slope of the moraine (Plate 7).

At Bilbrough, where the moraine rises to over 45 m above OD, the 33 m strandline forms a very conspicuous feature to the south of Bilbrough, where the 33 m terrace has been eroded into a back-wall of Triassic sandstone (Plate 18).

Plate 18. The 33 m terrace south of Bilbrough that terminates against a well-defined strandline [SE 531 459]. The erosional back-wall of the terrace, underlain by Triassic sandstone, rises 7.0 m to the 40 m terrace. (Photograph. M.D. Bateman)

Owing to the low-angle slope of the moraine south of Bilbrough (and south of the A64) the front edge of the 33 m terrace is not well defined, or mappable, as it grades into the 33 Metre Surface, that slopes gently to the south, for over 1.0 km, to the 20 m strandline (Fig. 14). On this surface, which is strewn with erratics, the 25 m strandline is poorly developed as a minor feature.

Immediately south of the A64, several small hills rising to a little over 30 m above OD, or about 5.0 m above local ground level, are probably remnants of the 33 m terrace, which has been downgraded by erosion from a retreating shoreline. At Street Houses, a section through the largest hill, provided by a roadcut at the side of the A64 (Fig. 18), shows an eroded and channelled surface on till overlain by sand and laminated clays that drape the underlying irregular surface.

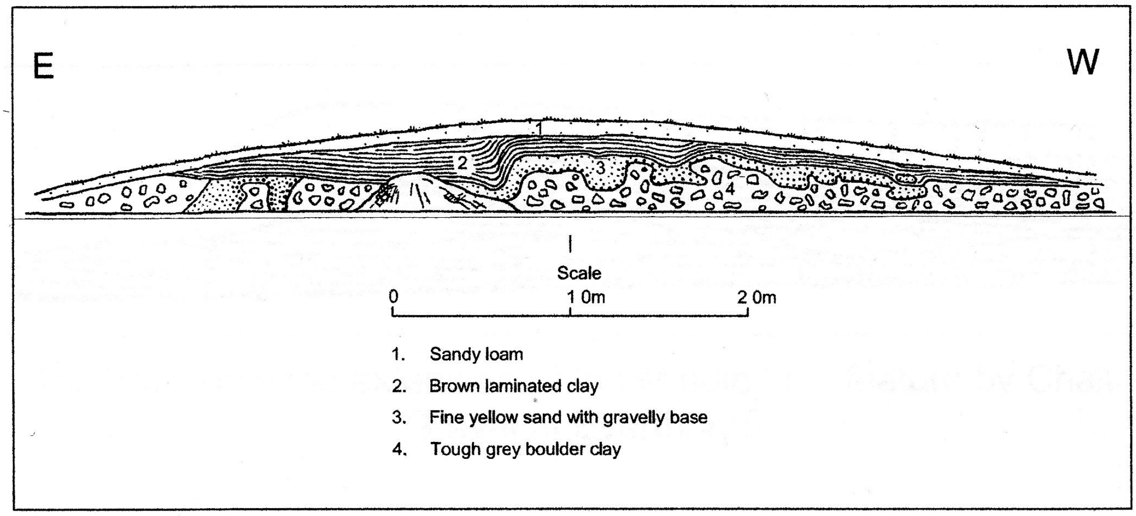


Figure 18. Erosional surface of the 33 m terrace on channelled till in a road cutting at Street Houses [SE 531 456]. The till is overlain by sand and laminated clay. From Edwards *et al*., 1950.

The sequence appears to show that there was a rise in lake level, or local ponding, following erosion of the till, resulting in the deposition of sand and laminated clays. For clarity, on Figure 14, the deposits forming these mounds on the erosional terrace have been mapped as part of the 100 Foot Strandline, though they were not defined in this way by Edwards *et al.* (1950).

Further west, on the northern side of the York Moraine near Long Marston (Fig. 14), the 33 m terrace has again been eroded into a back-wall of Triassic sandstone. At this location, in contrast to Bilbrough, the 33 m terrace is clearly defined by a surface that is inclined into the hillside, where it forms an erosional hollow subject to flooding. This hollow or shallow valley, known as 'the Glen', was believed to have been used by the Royalist cavalry, as a hidden refuge, following the Battle of Marston Moor in 1644 (Tincrey, 2003).

7.4.2. The Wolds

Near Pocklington (Fig.15) the 33 metre terrace has a special significance as its imprint (along with the imprint of the 40 metre terrace) differentiates the older alluvial fans, which are terraced, from the younger alluvial fans, which are not terraced. The 33 m terrace has been mapped on Rowland Hill and to the east of Rowland Hill, where a planar surface representing a degraded part of the 33 Metre Surface forms the Yapham cricket field. Planar-topped erosional mounds of Triassic marl, above the 30 m contour at Gowthorpe and southwest of North Hill, which were not covered by later fan deposits, exist as remnants of a once more extensive terrace. Other erosional features that could mark the 33 m shoreline are the distal terminations of the older alluvial fans at North Hill and High Belthorpe.

In contrast to the Pocklington area (Fig. 15), where the older alluvial fans have been mapped as discrete lithological units, the remainder of the mapping along the edge of the Wolds (Fig. 16) has differentiated the older alluvial fans, below the 40 m terrace, into separate depositional surfaces separated by strandlines as shown on Figure 14.

Generally along the face of the Wolds, between Market Weighton and Hessle (Fig. 16) the 33 metre strandline is not everywhere a distinct mappable feature. This is due to the unconsolidated nature of the older fan alluvium that has not been stable enough to preserve the 33 m terrace as a stable landform. Often the hillside falls away on a convex slope either down to an intermediate strandline at 25 m or to the top of the 20 Metre Surface. The 33 m terrace is however clearly defined on the south bank of Hotham Creek near South Newbald (Plate 19) and on surfaces that must originally have been headlands projecting westwards into high-level Lake Humber. These headlands, now indicated by the 30 m contour, occur at Everthorpe and Elloughton, where gravel deposits have been preserved, and near South Cave where the headland is formed by an outcrop of Cave Oolite.

Plate 19. The 33 m terrace and strandline in Hotham Creek near South Newbald [SE 963 353]. The back-wall above the 33 m terrace is about 2.0 m high. The ground rises to the 40 m strandline just below the tree line. (Photograph. W.A. Fairburn)

The mound containing the Everthorpe gravels [SE 905 320] lies between 30 m and 35 m above OD near Everthorpe village (Fig. 16). A good section through the deposit, in a nearby railway cut, was described by Dakyns *et al.* (1886) who observed that the gravel had been deposited against an eroded bank of Middle Lias, which extended the cut feature a further 200 yards (183 m) to the west (Fig. 19). The extent of the deposit, when examined by de Boer *et al*. (1958), was restricted to a few small exposures north of the railway line.

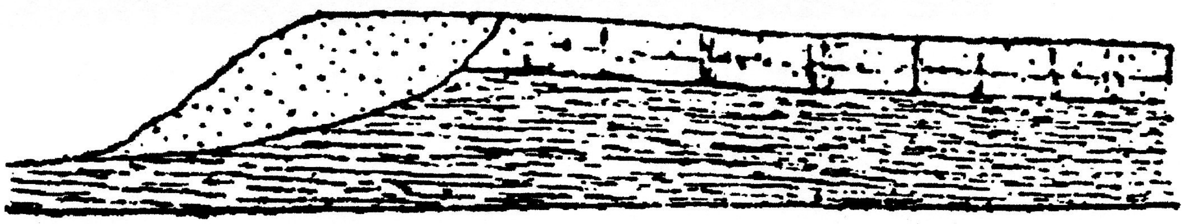


Figure 19. Sand and gravel banked against an eroded spur of Liassic rocks at a gravel pit near Everthorpe [SE 905 320]. From Dakyns *et al*., 1886

At the present day, traces of the former gravel pits, which had been worked to a depth of 4.0 m – 5.0 m, can still be seen on both sides of the railway. From surface occurrences the deposit consists mainly of abundant sub-rounded to rounded chalk and sub-angular flint pebbles, with a few pebbles of Cave Oolite and sandstone (Gaunt, 1976b). While Gaunt (1976b) preferred a pre-Devensian fluvio-glacial origin for the deposit the current mapping (Fig. 16) suggests a Late Devensian age, with accumulation of the deposit during a high-level phase of Lake Humber. That the gravels extend to an elevation of over 35 m above OD need not be a major objection to the deposit being 'Older Littoral Sand and Gravel', as reworked alluvial fan gravels, which were the source of the deposit, occur along the spur to the east of the gravel pit, toward Everthorpe Quarry, up to an elevation of over 40 m above OD (i.e. a higher level of Lake Humber).

The Mill Hill [SE 942 278] gravel deposit, near Elloughton, caps a small isolated hill, now taken over by housing, that rises to nearly 35 m above OD from a base at about 30 m above OD. Because mammalian bones were found during quarrying of the deposit it has attracted considerable interest and several descriptions of the gravels or of the mammalian remains have been made. Lamplugh (1887) originally described a 5.0 m succession composed of a lower gravel bed, some 1.5 m thick, composed of cross-bedded yellow sand with stony layers, that contained the mammalian remains, lying on hard grey clay (Jurassic Ancholme Clay), overlain by a cross-bedded rough, stony upper gravel bed about 2.7 m thick. Based on a description by de Boer *et al*. (1958), the whole sequence is current bedded (generally to the southeast, Sheppard, 1897) with the two gravel beds separated by an erosional unconformity (Fig. 48). The lower gravel, according to de Boer *et al*. (1958), contains large blocks of Coal Measure sandstone and Millstone Grit while the upper gravel is composed of locally derived flint, chalk and oolite limestone with only a few foreign pebbles. Sheppard (1897) also commented on large boulders of Kellaways Rock and one large boulder of Whin Sill in the floor of the pit. A precise age for the mammalian remains has not been confirmed (see Section 11.2.2.).

Near Hessle, in the vicinity of the A1105 (Fig. 16), the 33 m terrace forms a series of features that have been eroded into the Skipsea Till (formerly Hessle Till). Three locations were mapped; the most conspicuous being north of the Humber Bridge Country Park where a planar surface over 200 m wide has been excavated to a depth of about 1.0 m to provide construction material for embankments leading up to the Humber Bridge. This development of the 33 m terrace, near Hessle, therefore has special significance as Stage 3 in the regressive decline model for Lake Humber, as it is younger than the Skipsea Till. Dating of sediments, below the Skipsea Till, includes 17.5 ka (range 19.1 ka – 15.9 ka) for loess in solifluction deposits (Wintle & Catt, 1985, see section 3.4.) and 16.2 ka (range 16.6 ka – 15.8 ka) for cross-bedded sands deposited in ponds or a sandur environment (Bateman *et al*., 2011). A sub-glacial stream deposit in the Skipsea Till has been dated by Hartmann (2011) at 17.01 ka (range 18.34 ka – 15.68 ka). Dates from pro-glacial fluvial deposits cut into the Skipsea Till extend from 15.07 ka (range 16.03 ka – 14.11 ka) to 16.51 ka (range 17.55 ka – 15.47 ka) at Barmston (Hartmann, 2011). These data, while providing only a proxy for dating Stage 3 or high-level Lake Humber of Gaunt (1976b), nevertheless support a date of 16.6 ka (range 17.8 ka – 15.61 ka) for distal sands from the 'Older Littoral Sands and Gravels' (33 m terrace) at Ferrybridge (Bateman *et al*., 2008).

Terracing of the Skipsea Till in the Humber Gap, also indicates that the till did not provide a seal at any level of Lake Humber: this must have been provided by a lobe of North Sea ice. That the till was not an effective seal has been suggested by Straw (a written contribution in Madgett & Catt, 1978) who doubts if a morainic ridge ever crossed the Gap.

7.4.3. Brayton Barff

Edwards (1937) recognised a well-defined beach formed of gravel at about 70 ft – 85 ft (21.3 m – 25.9 m) above OD on Hambleton Hough and Brayton Barff. This probably corresponds to mapping, on Figure 17, of a terrace between 20 m and 30 m above OD that is best developed along the eastern and western sides of Brayton Barff. Whilst the back-wall of the terrace falls short of 33 m it must be equivalent to part of the 33 Metre Surface that has been recognised on the York Moraine (Fig. 14) and on the Older Alluvial Fans flanking the western side of the Wolds south of Market Weighton (Fig. 16).

7.5. The 100 Foot Strandline (Older Littoral Sand and Gravel - Stage 3)

Gravel deposits mapped by Edwards (1937) along the Permian escarpment on the western side of the Vale of York, at a constant elevation close to or above the 100 ft contour (30.05 m), were considered by Edwards (1937) to be remnants of a beach deposit laid down on the shoreline of glacial Lake Humber: they were later referred to by Gaunt (1976b) as the 'Older Littoral Sand and Gravel'. Edwards (1937) regarded these gravels as pre-Hessle in age i.e. older than what was considered, at the time, as the final glaciation in East Yorkshire. Mapping by Edwards shows that the scattered patches of gravel sloped gently eastwards from a maximum elevation of 100 ft – 110 ft above OD (30.05 m – 33.5 m) down to about 50 ft above OD (15.2 m), but did not extend into the 25 Foot Drifts (7.6 m) of the Vale of York. The deposits extend from Tadcaster to near Doncaster, a distance of over 20 km (see Edwards, 1937, fig. 1). The relationship between the outcrop of the shoreline gravels between Tadcaster and Ferrybridge and the littoral deposits on the 15 and 20 Metre Surfaces (mapped on Fig. 17) is shown on Figure 20.

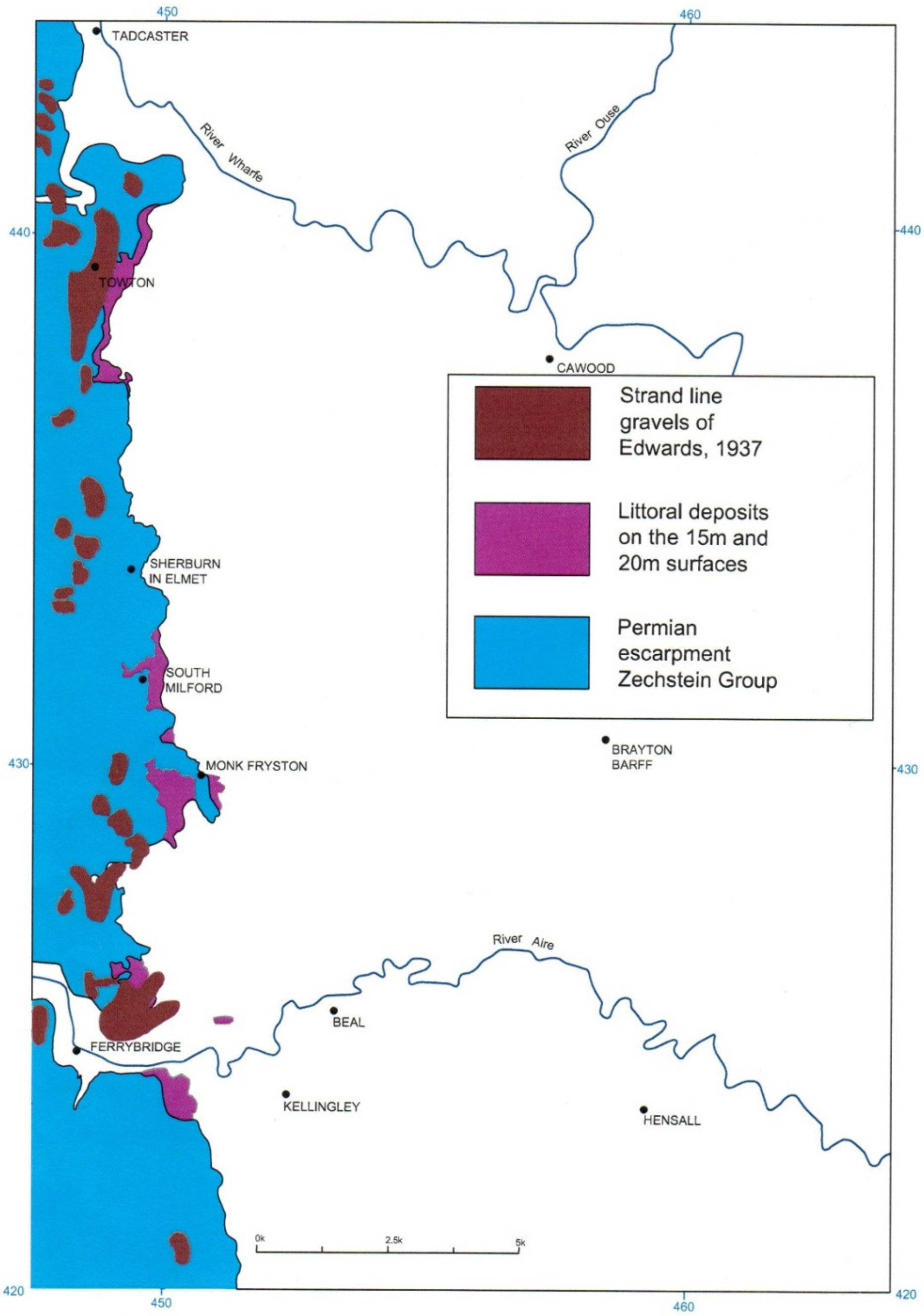


Figure 20. Location plan of the 100 Foot Strandline gravels reproduced from Edwards, 1937, fig. 1. The 15 m and 20 m littoral deposits are taken from landform mapping shown on Fig. 17 (this text). (Note that the register between the two sets of data is not precise).

It should be noted in Figure 21, that in a location north of Ferrybridge, the 100 foot shoreline deposits appear to coalesce with, or overlie, the lower level littoral deposits on the 20 Metre Surface. This is probably a registration error.

Edwards (1937) has described the gravels as consisting mainly of thoroughly water-worn coarse gravels and boulders (up to 0.5 m in diameter) of locally derived Magnesian Limestone, with a small percentage of foreign rocks that include cobbles of Carboniferous sandstone and chert plus occasional quartz pebbles probably derived from 'Older Drift'. During an earlier investigation of these deposits, by Melmore & Harrison (1934), a 12 ft (3.7 m) section of gravel from a pit near Garnet Lane in the western part of Tadcaster was described as consisting of pebbles of Magnesian Limestone (up to 15 cm in size) in a purplish sandy matrix capped by sand, blue clay and alluvium (Fig. 21).

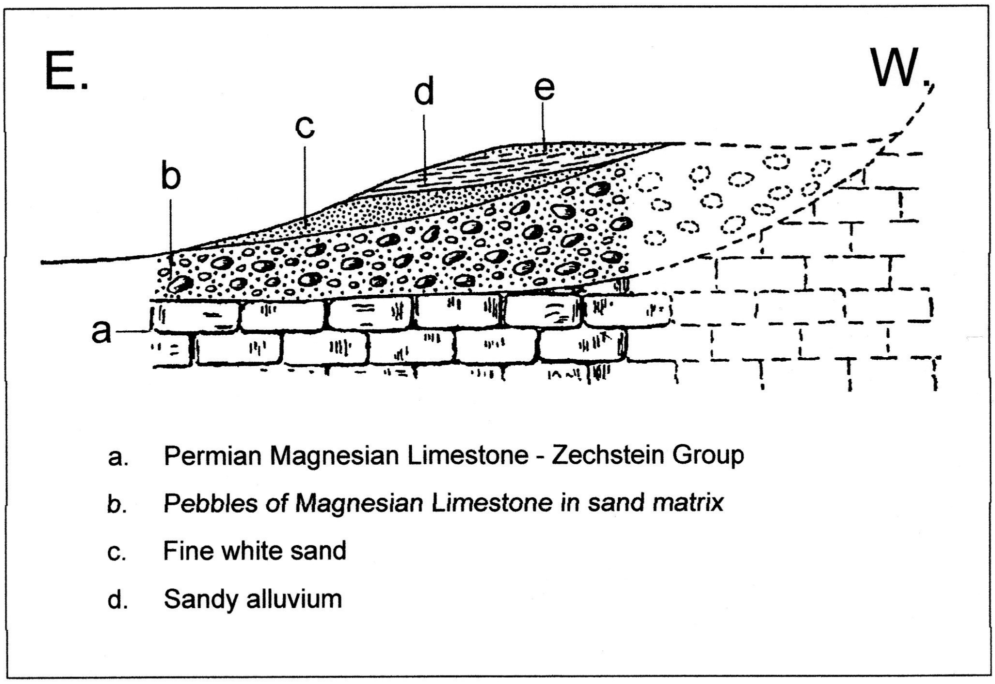


Figure 21. Section of pit in the 'Strand Line Gravels' of Edwards, 1937, located near Garnet Lane west of Tadcaster (Melmore & Harrison, 1934). The site of the pit now forms part of a housing estate and cannot be located.

That the 100 Foot Strandline gravels do outcrop at a fairly constant elevation of *c*. 30 m above OD is clearly illustrated by Murton & Murton (2011, fig. 15B). As with the gravel deposit north of Ferrybridge (referred to above), some of the gravel also extends below the 30 m contour at Askam, just south of the River Went (Murton & Murton, 2011, fig. 15B). Here again the gravel deposit appears to be located on a 20 m terrace. Both the deviant Ferrybridge and Askam gravel deposits are younger and could be related to a later littoral event. It is also implicit in this constant level for the 100 Foot

Strandline gravels that no post-depositional isostatic adjustment has occurred. A similar conclusion was reached by Fairburn (2011) following mapping of shorelines in the Pocklington area.

In contrast to the conclusions of Edwards (1937), that the gravels are beach deposits laid down on a shoreline at 100 ft above OD, during an event described by Gaunt (1976b) as the high-level lacustrine phase of Lake Humber, Carruthers (1948) has asserted that the 'strandlines' gravels represent hill-wash poured on to the margin of a residual ice-tongue. Part of Carruthers' assertion, is based on the descriptions of Edwards (1937 and 1940) that the shoreline is intermittent and there is no true terrace or old cliff line. His views were however influenced by his belief that there are no lacustrine clays in the Vale of York and therefore no Lake Humber: an opinion strongly opposed by Gaunt (1976b). Strong evidence supporting the 100 Foot Strandline as a Lake Humber shoreline is provided by the landform mapping that has recognised terracing, at this elevation, on the eastern side of the Vale of York (Fig. 16) and on the York Moraine (Fig. 14).

The term 'Older Littoral Sand and Gravel' was introduced by the British Geological Survey to describe sequences of sand and gravel 'whose pebble composition is virtually identical lithologically to adjacent rocks or pre-existing Quaternary deposits, and which have clearly originated by localised re-working of these older formations' (Gaunt, 1976b). The stratigraphic context of the Older Littoral Sand and Gravel was restricted by Gaunt (1976b) to a horizon between the lower and upper periglacial surfaces; events believed by Gaunt to be of Devensian age. A further distinctive feature of the 'Older Littoral Sand and Gravel' recognised by Gaunt (1976b) is that it does not extend higher than 33 m above OD and that the upper limit of the deposits rises gradually northwards because of isostatic uplift. The term appears to have been first used on the Selby 1:50 000 sheet (British Geological Survey, 1973b) and replaced usage of the term 'Glacial Sand and Gravel' used on the Goole 1:50 000 sheet (British Geological Survey, 1971). On the more recent Selby 1:50 000 sheet (British Geological Survey, 2008), the 'Older Littoral Sands and Gravels' have been re-named as the Pocklington Gravel Formation, described by Ford *et al*. (2008) as a glaciofluvial fan deposit and in comparable terms by Fairburn (2011) as an alluvial fan deposit (see Fig. 16).

Gravel deposits included by Gaunt (1976b) as 'Older Littoral Sand and Gravel' include:

- the 100 Foot Strandline gravels of Edwards (1937);
- the now renamed 'Pocklington Gravel Formation';
- the Gryphite (*Gryphaea incurva*) gravels of Tate and Blake (1876) occurring in gravel pits west of South Cave at elevations between 15 and 20 m (see also Dakyns *et al*., 1886). These deposits are now referred to as 'Marginal Sand and Gravel' by Gaunt (1992) and are mapped on Figure 16 as part of the 20 Metre Surface;
- numerous occurrences of gravel commonly overlying cryoturbated Triassic sandstone in sand pits on the Snaith Ridge e.g. near Hensall and Pollington. A good example of this was described by Parsons (1887) from a section through the gravels at Great Heck railway station (Fig. 23).

While the term 'Older Littoral Sand and Gravel' was used by Gaunt (1976b) to encompass widely distributed gravel deposits in differing geological situations and elevations, the term appears to have a more restricted usage in later publications and was not used for instance in the Kingston Upon Hull and Brigg Memoir (Gaunt *et al*., 1992) or on the Kingston Upon Hull 1:50 000 sheet (British Geological Survey, 1983a). In these two publications the terms 'Marginal Sand and Gravel' and 'Sand and Gravel' were preferred. In this text it is considered that the term 'Older Littoral Sand and Gravel' should strictly only apply to the 100 Foot Strandline gravels of Edwards (1937) but can be equated with the gravel deposits at Everthorpe and Mill Hill near Elloughton (Fig. 16), sand and laminated clay overlying an erosional surface on the 33 m terrace at Street Houses (Fig. 18) and gravel deposits overlying till on the York Moraine at Mill Mound and near Severs Howe (Fig. 14). More distal deposits on the 33 m terrace that could also be included as 'Older Littoral Sand and Gravel' include pebbly sand mapped on the Wakefield 1:50 000 sheet (British Geological Survey, 1998) above the 25 m contour as a 'lacustrine shoreface and beach deposit'.

Dating of the 'Older Littoral Sand and Gravel' is restricted to a single OSL determination made by Bateman *et al*. (2008) who sampled distal sands from deposits on the 100 Foot Strandline in excavations near Ferrybridge (see section 3.5.). An age of 16.6 ± 1.2 ka given for these sands is however disputed by Murton *et al*. (2009) who have suggested that the tested Ferrybridge sands may have been reworked from older deposits by colluvial processes or else represent a younger high-level phase for Lake Humber (Murton & Murton, 2011). There is however no evidence for this assertion.

7.6. 30 Metre Formline/Strandline

This formline is a distinctive low-amplitude ridge, at 29 m - 30 m, above OD, that marks the crestal region of Mill Mound above York University. As such, the feature provides a boundary between the fairly level 33 m terrace at the summit of Mill Mound and the steeper slope of the York Moraine that descends towards the 25 m terrace. The feature is most conspicuous to the west of the water tower on Mill Mound (Plate 20) but is also recognisable northeast of the water tower on the LIDAR (airborne light detection and ranging) digital elevation maps (Figs 33b and 33c).

30 m
formline

Plate 20. The 30 m formline below the water tower at Mill Mound on the York Moraine [SE 621 509]. Note the level surface of the 33 m terrace (here at about 32 m) forming the top of the moraine just below the tree line and some 2.0 m above the formline. (Photograph. W.A. Fairburn)

The 30 m formline can be best explained as the distal or depositional front edge of the gravels that underlie the 33 m terrace on Mill Mound (shown on Figure 14 as equivalent to the 100 Foot Strandline gravels of Edwards, 1937). These gravels are exposed in a roadcut below the water tower (Plate 16), and can be penetrated by hand auger drilling at depths of about 0.5 m on the 33 m terrace. Down-slope of the formline auger holes, to a depth of 1.0 m, encountered only gravel-free sand.

Near Acomb and Severs Howe (Fig. 14) several dissected mounds of gravel, similar to Mill Mound, rise above a base level of 30 m above OD to levels close to 35 m above

OD. Till has not been proved at these locations. A 30 m formline has also been recognised on the York Moraine northeast of Heslington and Holtby, whilst at Sand Hutton a planar surface, below the 30 m contour, was also mapped on the 33 Metre Surface (Fig. 14). Near Pocklington, the 30 m formline is not a conspicuous feature but it does seem to form an erosional lower edge to the Older Alluvial Fans below High Belthorpe and between Rowland Hill and Yapham Grange (Fig. 15). Further south, on the Wolds, the 30 m formline was mapped near Market Weighton and Hotham and at intervals for several kilometres southwest of Drewton Dale (Fig. 16).

It seems likely that as the 30 m formline is only intermittent and poorly defined, it probably marks either a temporary shoreline of Lake Humber as the lake receded below 33 m above OD, the lowest level of a shoreline oscillating above and below the 33 m terrace, or a later temporary rise in lake level to 30 m above OD that allowed a period of lacustrine deposition on eroded remnants of the 33 m terrace as at Street Houses (Fig. 18).

7.7. 25 Metre Terrace (Stage 4)

Although forming a distinctly visible terrace around Mill Mound at York University (Fig. 14) and near High Catton where the Escrick Moraine rises to 35 m above OD (Fig. 15), the 25 m terrace is not a conspicuous feature elsewhere in the Vale of York and appears to be only preserved in certain small areas. As an example of this, the terrace has been recorded on Hambleton Hough but not on Brayton Barff (Fig. 17). Because of its localised occurrence, it was considered misleading to extrapolate a major boundary for this terrace around the length of the York Moraine and along the Wolds despite the fact that the 25 m terrace may have been a significant shoreline of Lake Humber. Consequently, the 25 m shoreline is only recorded as a strandline on the 33 Metre Surface between 20 m and 33 m above OD (Figs 14, 15 and 16). It was therefore considered pragmatic not to elaborate the landform mapping with another defined surface.

7.7.1. York Moraine

The 25 m terrace forms a continuous surface traceable around the summit of the York Moraine below Mill Mound (Fig. 14): it can be up to 50 m wide but is generally less than 20 m. At one location (Plate 21) the width of the planar surface forming the terrace has been extended by mechanical levelling to provide playing fields.

Plate 21. The eroded back-wall of the 25 m terrace near York University [SE 623 510]. To the right of the back-wall the 25 m terrace has been modified by mechanical levelling. The fence posts are 1.2 m. (Photograph W.A. Fairburn)

In many places the level surface of the terrace has been as asset for road construction and the alignment of University buildings and nearby housing. Although there is only minimal sub-surface data available above and below the 25 m terrace, a small construction pit, near the University [SE 618 508], does reveal that the terrace is underlain by boulder beds with about 0.5 m of sand cover, while boulder beds above the terrace are exposed in a sunken pathway [SE 614 509] between 25 m and 30 m above OD (Plate 22).

Plate 22. The 25 m terrace and strandline (centre of photo) rising to over 30 m above OD at the crest of the ridge [SE 614 509]. Boulder beds are exposed in the sunken pathway that cuts the ridge. The brick wall to the right of the photograph is 3.0 m high. (Photograph W.A. Fairburn)

On the western side of York the 25 m strandline is developed below Severs Howe and provides the mappable boundary to a number of prominent mounds of sand and gravel along the crest of the York Moraine southwest of Severs Howe (Fig. 14). To the northeast of York University the 25 m terrace was only noted near Dunnington, just below the crest of the moraine (Fig. 14). In the Bilbrough / Healaugh region, on the southern side of the York Moraine, the 25 m terrace is again only a localised feature, being mapped at a few locations south of the A64 and at one location northwest of Healaugh, where it had been reported by Melmore (1940).

On the northern side of the York Moraine, near Bilbrough and Healaugh, defining an equivalent to the 25 m terrace, as recognised at Mill Mound, is somewhat enigmatic as the damming effect of the York Moraine appears to have resulted in higher elevations for Lake Humber (referred to as Glacial Lake Alne by Ford *et al.*, 2008) and consequently somewhat higher elevations for shoreline terracing. As an example of this, the back-wall of the 20 Metre Surface rises to nearly 25 m above OD compared with only 20 m above OD south of the moraine. The 'two' terraces do however join through the Healaugh gap leading to Catterton Beck (Fig. 14). It is therefore likely that an equivalent of the 25 m terrace is only recorded by a surface approaching 30 m above OD that outlines two small mounds north and northeast of Bilbrough [SE 530 479 and SE 549 483]. It is however possible that these mounds represent eroded remnants of the 33 Metre Surface as shown on Figure 18.

7.7.2. Escrick Moraine

Along much of its length between Stillingfleet and Wheldrake the crest of the Escrick Moraine lies below 20 m above OD and it is only in the region of High Catton, where gravel beds have been worked (Kendall & Wroot, 1924; Stather, 1913), that the crest of the moraine rises to over 30 m above OD. In this elevated part of the moraine, a conspicuous terrace at 25 m above OD can be traced from the northern tip of the moraine to nearly as far south as Wilberfoss (Fig. 15).

7.7.3. The Wolds

Northwest of Pocklington the 25 m terrace is an erosional shoreline and a depositional boundary of the prograding Younger Alluvial Fans. The erosional nature of the shoreline is evident adjacent to Highfield Lane, northwest of Fangfoss (Fig. 15), where the terrace forms the boundary between shallow subcropping Triassic marl (partly covered by fine gravel or grit) and the Younger Alluvial Fans. It also forms part of the erosional edge of the Older Alluvial Fans to the east of Bolton [SE 781 523]. The depositional origin of the 25 m terrace, as a shoreline, is marked southwest of High Belthorpe [SE 776 538] and west of Gowthorpe [SE 761 544] by the termination of the Younger Alluvial Fans, where they grade into outwash fans of clayey sediment.

In contrast to the Pocklington area, where the two ages of alluvial fans are clearly defined by the mapping (Fig. 15), the status of gravels below a well-defined 25 m terrace to the northwest and south of Market Weighton (Fig. 16) is less certain. These gravels, as the mapping indicates (Fig. 16) could represent the Younger Alluvial Fans deposited following earlier erosion of the Older Alluvial Fans along a 25 m shoreline or they could in fact represent part of the Older Alluvial Fans which have been terraced by the 25 m shoreline. Further south between Hotham and Malton, where the 25 m terrace is again a mappable feature, the Younger Alluvial Fans are not present so the problem does not exist and the terrace has been imprinted onto the Older Alluvial Fans (the relationship of the alluvial fans to the shoreline of Lake Humber will be discussed later in the text).

7.8. 20 Metre Terrace (Stage 5)

Following a retreat of lake level, from a short-lived stand at *c*. 25 m above OD down to a more prolonged Stage 5 stand at *c*. 20 m above OD, wide depositional tracts, mapped as the 20 Metre Surface, became established around the margins of Lake Humber and in Lake Alne to the north of the York Moraine. In Lake Humber, this surface or terrace is mainly restricted to the shallower parts of the lake bordering the York Moraine and adjacent to eroded remnants of the York Moraine east of Tadcaster and south of Bilbrough near Colton (Fig. 14). The 20 m terrace also forms a level platform between Barmby Moor and Stamford Bridge (Fig. 15), interpreted here as a fan delta, but further south beyond Market Weighton it is only a minor beach on the western scarp of the Wolds (Fig. 16). A 20 m terrace has also developed along the crestal regions of the Snaith Ridge near Kellington (Fig. 17) but elsewhere south of the Escrick Moraine the 20 m terrace is only represented by a localised strandline. Much of the Escrick Moraine remained submerged at the 20 m lake level apart from the higher summit of the moraine around High Catton (Fig. 15). North of the York Moraine in the area of Lake Alne, the 20 Metre Surface forms a widespread terrace, in places detached from the York Moraine, between 20 m and 25 m above OD some 5.0 m higher than south of the moraine (Fig. 14). This extensive development therefore not only reflects a degree of damming by the York Moraine to produce higher lake levels but also a new phase in the evolution of Lake Humber, or Lake Alne, with the first signs that the pro-glacial lake was beginning to silt-up.

The 20 m terrace has not been recognised by the British Geological Survey on the 1: 50 000 York and older Selby sheets (British Geological Survey, 1983b and 1973b) where the littoral sediments have been incorporated into the gravel, sand, silt and lacustrine clay of the 25 Foot Drifts, which underlie the Vale of York. On the newer Selby sheet (British Geological Survey, 2008) the terraced sediments are included as either part of the Naburn Sand Member or the Bielby Sand Member of the post-Lake Humber Breighton Sand Formation. Earlier recognition of the 20 m terrace was however made by Melmore (1940) who identified a 25 foot terrace (i.e. 25 ft above river alluvium) at a level a little over 15 m above OD on the right bank of the River Derwent near Stamford Bridge. In this region, near the Stamford Bridge Viaduct, the 20 Metre Surface rises from a frontal edge at 15 m above OD to a back-wall at 20 m above OD.

Gaunt (1976b), in his account of the Quaternary geology of the southern part of the Vale of York, describes a littoral facies around the flanks of Lake Humber as marginal sand, which is either overlain by clay of the 25 Foot Drifts, or passes laterally in to it. The marginal sand was therefore accepted as a littoral facies of the low-level lacustrine phase of Lake Humber. Clearly Gaunt (1976b) did not recognise any erosional boundaries for his littoral zones. In the southern part of the Vale of York, they probably included both the 15 m and 20 m terraces as defined in this text. The fact that the lacustrine clays may be overlain by the littoral sands is an expected situation in an offlapping or regressive sequence resulting from a retreating shoreline. Similarly to Gaunt (1976b), Edwards *et al*. (1940) stated ‘there is no well-defined strandline’ at the edge of the 25 Foot Drifts ‘although in one or two places uncertain evidence of marginal beaches can be found above the 25 ft (7.6m) level’ (this level is closer to 10 m above OD in the area referred to by Edwards *et al*., 1940).

7.8.1. York Moraine

The 20 m terrace forms a well-defined littoral zone, between 15 m and 20 m above OD, along the southern edge of the York Moraine between the River Derwent and the River Wharfe (Fig. 14). Extensions of the terrace project southwards, from the York Moraine, adjacent to the Derwent and Ouse rivers, where glacial drainage has broken through the moraine into Lake Humber. The terrace also projects southwards, as a littoral deposit, from Bilbrough through Colton towards Bolton Percy (Fig. 14) inline with a southerly bulge in the Moraine. Between this southerly extension and the River Wharfe the wide expanse of the 20 m terrace is indicative of early silting-up of Lake Humber south of Catterton Beck and north of Bolton Percy (Fig. 14). Sections through the terrace were not recorded during the mapping, so sub-surface data on the sedimentology and structure of the terrace is mainly lacking. Surface evidence, from cultivated fields, does indicate that near-surface sediment, on the terrace, is dominantly sandy alluvium with some pebbles and rare cobbles. This conforms with observations by Melmore (1940) who reported fine loamy sand in an ‘old gravel pit’ near Stamford Bridge and in nearby clay pits.

Plate 23a. The 20 Metre Surface in the foreground (location SE 618 507), to the west of York University, is inclined into the hillside to form a shallow depression that marks the boundary of slumping in the scree-apron sand mantle below the 20 m strandline. A spoon-shaped slump scar at the left-hand edge of the strandline is not clearly visible on the photograph but can be seen on the LIDAR image Figure 33b and on Plate 23b. (Photograph W.A. Fairburn)

Plate 23b. Enlargement of part of Plate 23a showing back-wall of the 20 m terrace rising to the 25 m terrace. Note OSL sample point for Shfd12068 and Shfd13034. (Photograph M.D. Bateman)

The 20 m terrace as a distinctive landform, on the southern side of the York Moraine, is best seen in a region to the west of York University and south of Mill Mound near 'The

Retreat' (Fig. 14). Here two landscapes across the terrace including a backwall rising up to the 25 m terrace are illustrated on Plates 23a, 23b and 24. The first profile (Plates 23a and 23b) shows the level surface of the terrace, in the foreground, dotted with sandy mole hills, that has become detached from its back-wall by slumping. The line of slumping is indicated by the inclination of the terrace into the hillside to form a shallow depression complete with ponding not visible in the photograph, just behind the sawn tree trunk. The actual 20 m strandline occurs several metres above this depression, as shown on Plate 23a, with the back-wall rising quite steeply to the 25 m terrace near the top of the photograph.

The second profile (Plate 24 and Fig. 35), located some 250 m to the west, illustrates a more dramatic rise of the 20 m terrace back-wall which has been emphasised by landscaping. The building in the background (The Retreat) is located on the 25 m terrace.

Plate 24. The back-wall of the 20 m terrace rising 4.0 m to 'The Retreat' located on the 25 m terrace. Note that both the terrace, in the foreground, and the back-wall have been emphasised by landscaping. The approximate location is SE 616 509. (Photograph W.A. Fairburn)

To the north of the York Moraine, the 20 m terrace is again widely developed as a narrow fringing littoral deposit bordering the moraine, but also in broader depositional tracts that are somewhat detached from the moraine. The main difference between the terracing north and south of the moraine is that those to the north are about 5.0 m higher in elevation at levels of 20 m – 25 m above OD. This difference does not appear to be an isostatic effect as the change in elevation from north to south seems to occur in the meltwater channels through the moraine (see section 7.16.3.). It is considered that the difference in elevation is a consequence of damming by the York Moraine allowing water levels to stand some 5.0 m higher north of the moraine. It is also noticeable both to the east of the River Ouse and to the west of Healaugh Beck that the bounding edges of the 20 m terrace are slightly lower in elevation than elsewhere along the northern edge of the York Moraine. To the west of Healaugh Beck, in the region of Long Marston and Hutton Wandesley (Fig. 14), it is quite clear that this effect is erosional and is caused by movement of meltwater towards Healaugh Beck. A similar erosional effect is also suggested for the lower levels of the 20 m terrace east of the River Ouse (the York Gap, section 7.16.3.1.).

Whilst the flanking littoral deposits on both sides of the York Moraine were probably derived by erosion and wash from the moraine, the origin of the larger more detached areas of alluvium and sand forming the 20 m surface north of the moraine are probably different. Two of these larger deposits have been mapped: one to the north of Warthill and west of Sand Hutton and the other surrounding Knapton west of Severs Howe (Fig. 14). As other smaller detached areas of the 20 m surface occur elsewhere north of the moraine, notably at Shipton and Newton-on-Ouse, it would seem likely that the Lake Alne extension of Lake Humber may well have partially silted-up, albeit temporarily, when the level of Lake Humber, to the south, stood at 20 m above OD. That there was later erosion of the 20 m terrace, north of the York Moraine, is evident west of Severs Howe where a meltwater channel has incised down below 15 m above OD to detach the Knapston terrace from the edge of the moraine (Fig. 30). More recent erosion, subsequent to deposition of a very localised 15 m terrace adjacent to the Nidd and Ouse rivers, is also indicated by the isolation of remnants of this terrace at Skelton, Overton, Beningbrough and at two locations southwest of Beningbrough (Fig. 14).

7.8.2. The Wolds

In the region of Stamford Bridge (Fig. 15), the 20 m terrace forms an extensive level surface that extends for about 5.5 km from Gate Helmsley (Fig. 14) to Full Sutton (Fig. 15), between the 15 m and 20 m contours. To the northeast of Stamford Bridge the width of the terrace decreases markedly as it becomes more confined in the Derwent valley leading up to the glacial overflow channel at Kirkham Priory, that connected

Lake Pickering to Lake Humber (Fig. 9). This wedge of sediment also extends south of Gate Helmsley, on the west bank of the Derwent River, to below Low Catton on the east bank and to the south of Full Sutton in the broad valley between High Catton on the Escrick Moraine and the edge of the alluvial fans west of Fangfoss (Figs 14 and 15). The expanse of sediment, that could have partly originated by over-flow from Lake Pickering or even Lake Eskdale via the River Derwent, does therefore not take the form of a fringing littoral deposit but more of the form of a fan delta (see Tucker, 1991, p.76) that was deposited in a discharge area from an early Derwent River into Lake Humber when it stood at 20 m above OD. This landform, deposited by the River Derwent, is more extensive than a similar but more degraded feature that extends along the banks of the River Ouse, south of York, to Middlethorpe and Fulford (Fig. 14).

Elsewhere in the region, near Pocklington (Fig. 15), the 20 m terrace appears to form a littoral deposit, standing above the Vale of York stretching from west of Barmby Moor towards Full Sutton. To the north of Spittal Beck, and towards the Escrick Moraine, the western and southern edge of the 20 m terrace forms a distinctive topographic feature as it rises about a metre above the lacustrine and fluvial sediments of the Vale of York. Its eastern edge is also a distinct mappable feature as it passes into the rising terminal margin of the younger alluvial fans. South of Spittal Beck (Fig. 15) where the terrace has been equated with the Bielby Sand Member of the Breighton Sand Formation on the Selby 1: 50 000 sheet (British Geological Survey, 2008), the western edge of the terrace is often marked by extensive sand deposits, forming dunes and bars running parallel to a possible palaeo-shoreline that has provided an environment for rabbit warrens. The eastern edge, which may also be marked by sand ridges, particularly on a line southwest of Bolton towards the A1079, the boundary is more frequently only indicated by a barely perceptible change of slope or by mounds of wind-blown sand developed on the younger alluvial fans. Southeast of Barmby Moor the boundary is partly obscured by earthworks along the A1079.

From Barmby Moor, south-eastwards towards Market Weighton, the 20 m terrace continues to flank the western edge of the younger alluvial fans (Figs 15 and 16), but south of North Cave to as far as North Ferriby, the terrace is eroded into the older alluvial fans and is bounded to the southwest by the younger 15 m terrace. Evidence that the 20 m terrace could be littoral in origin is provided by gravel pits near South Cave and northwest of Elloughton (Fig. 16). The South Cave deposits [SE 904 311] termed the ‘Gryphite gravel’ by Tate and Blake (1876), because of the abundance in them of rolled *Gryphaea incurva*, has been described by de Boer *et al.* (1958) as a gravel deposit derived very largely from the Lias but containing (besides *Gryphaea incurva*) chalk, flint, Carboniferous sandstone and quartz pebbles. The beds are now poorly exposed, but based on the de Boer *et al.* (1958) account, were mainly horizontally bedded but in places dipping to the northwest or northeast, directions not inconsistent with beach deposition. The deposit does reflect a shoreline environment, with *Gryphaea* forming an erosional lag deposit, associated with locally derived chalk and flint from the alluvial fans.

The gravel deposit located northeast of Elloughton (Fig. 16) referred to by Sheppard (1897) as Prescott’s Pit and by de Boer *et al.* (1958) as the Elloughton Quarry South, are more difficult to explain as a littoral deposit. The section (Fig 22), from de Boer *et al.* (1958), shows an erosional surface on the Cave Oolite at *c.* 50 ft (15 m) above OD covered by easterly dipping cross-bedded sand and gravel up to 1.0 m thick.

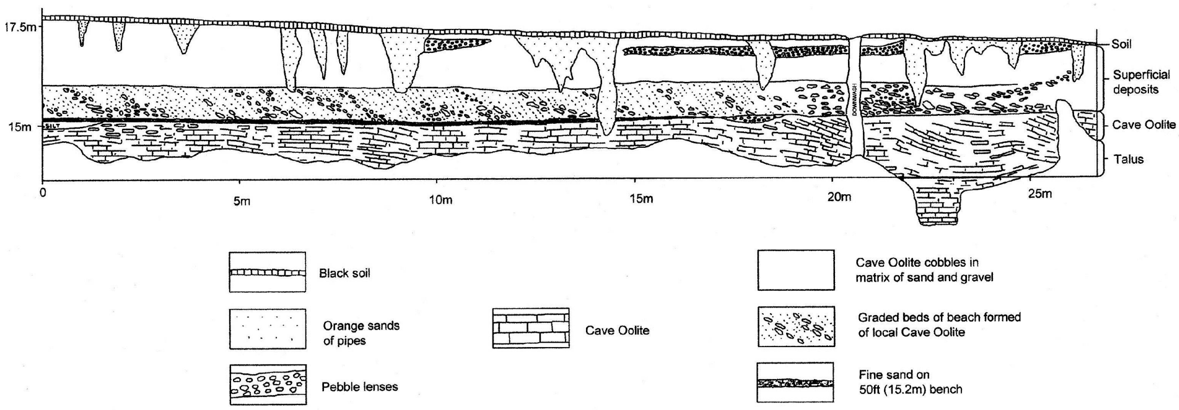


Figure 22. Easterly dipping cross-bedded gravels overlying an erosion surface on the Cave Oolite at *c.* 50 ft above OD (15 m) on the northern face of the Elloughton Quarry South (Prescott’s Pit) considered by de Boer *et al.* (1958) to be a storm beach. The quarry section is no longer visible.

It was considered by de Boer *et al.* that the deposit represents a storm beach formed with a water surface at about 50 ft (15 m) above OD. In Plate 25 (from Sheppard (1897)), the cross-bedded sand and gravel, dipping to the southeast at 45°, is shown to be capped by more chaotic horizontally bedded sand and gravel, of similar composition, on an erosional contact. Sheppard (1897) describes the gravels as consisting largely of Millepore Oolite (Cave Oolite) with some fossils from the Lias and Chalk.

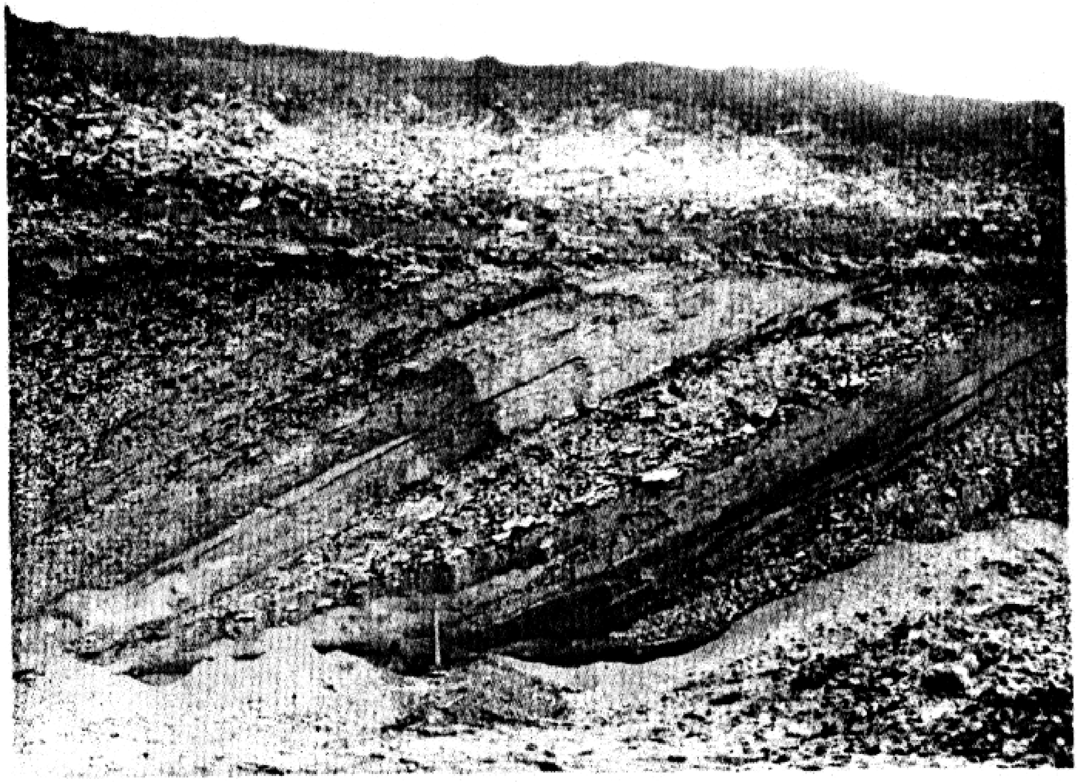


Plate 25. Prescott's Pit (Elloughton Quarry South) on the 20 m terrace showing cross-bedded gravels dipping towards the south east at about 45° overlain by horizontally bedded gravels. Note the truncated top to the cross-bedding. The section, which is about 4.5 m high, is located at SE 932 288 (Fig. 17). Photograph from Sheppard (1897).

Whilst the 15 m erosion level, in the quarry, could be a shoreline covered by fine sand (Fig 22), the overlying sand and gravel would appear to be more fluvial in origin rather than littoral: the capping of near horizontally bedded gravel could however be a back-beach deposit. The supposed fluvial interval therefore marks an event not compatible with a multiple-stage decline model, proposed in this text, for the drainage of Lake Humber, and could be equivalent to fluvial events associated with the Older Alluvial Fans (*cf.* Fig. 23).

7.8.3. Triassic Ridges

Terraced sand and gravel deposits, present on both the Brayton Barff and Snaith Triassic ridges, have evolved along shorelines of Lake Humber by reworking of both the Sherwood Sandstone and gravel accumulations that probably originated during pre-Late Devensian glacial and fluvioglacial events. Of these two regions the Snaith Ridge has the most complex geological history as Gaunt (1976b and 1994) has described four generations of sand and gravel deposition (see Section 3.3.1.2.). Reworking of this older sand and gravel has principally occurred to form terraces associated with the 15m lake level around Brayton Barff and with the 15 m and 20 m lake levels along the Snaith Ridge.

7.8.3.1. Brayton Barff

Around Brayton Barff and Hambleton Hough the 20 Metre Surface, between 15 m and 20 m above OD, forms a continuous or near continuous terrace (Fig. 17). The terrace is everywhere underlain by Triassic sandstone with only a thin covering of pebbles and cobbles of mainly Carboniferous sandstone. At both these locations, the landform mapping indicates that the lower boundary of the Triassic outcrop extends only as far as the 15 m contour, in contrast to mapping on the Goole 1: 50 000 sheet (British Geological Survey, 1971), which shows a much more extensive Triassic outcrop or subcrop. The erratics on the 20 m terrace were probably derived from sand and gravel deposits (and in the case of Brayton Barff from till, see Gaunt, 1994, fig. 37; Fig. 7) at or near the summit of each hill. Gaunt (1976b) has described the small outcrop of unbedded and poorly sorted sand and gravel that rests on 'Pennine boulder clay' at the summit of Brayton Barff as consisting principally of Carboniferous sandstone, with subsidiary Carboniferous and Permian limestone, and rare siltstone, chert, quartz, quartzite, and igneous rocks. He concluded that the assemblage is consistent with the suite of erratics found in the underlying 'Pennine Boulder Clay' and it has been classified by Gaunt (1976b) as 'Older Glacial Sand and Gravel'.

7.8.3.2. Snaith Ridge

The Snaith Ridge, which extends eastwards from Ferrybridge towards Snaith, is also an elongate ridge of Triassic sandstone, as many sand pits in the region of Hensall and Great Heck testify (Fig. 17). As with Brayton Barff, the Snaith Ridge is also a terraced landform with the 20 Metre Surface forming the crestal region south of Kellington, except for a small area of sand and gravel (mapped as a distal part of the 33 Metre Surface), referred to on the Wakefield 1: 50 000 sheet (British Geological Survey, 1998) as a lacustrine shoreface and beach deposit. Smaller areas of the 20 Metre Surface occur near Thornfield House, Whitley and Great Heck (Fig. 17). All these areas of the 20 Metre Surface are surrounded by the younger and more widespread 15 Metre Surface (Fig. 17). The gravel deposits terraced by the 15 Metre and 20 Metre Surfaces on the Snaith Ridge are equated with 'Glacial Sand and Gravel' on the Goole 1: 50 000 sheet (British Geological Survey, 1971) and 'Glaciofluvial Deposits' on the Wakefield 1: 50 000 sheet (British Geological Survey, 1998). Gaunt (1976b) has described gravels on the Snaith Ridge, present at heights between 7.0 m and 17 m above OD (i.e. generally corresponding to the 15 Metre and 20 Metre Surfaces), as being composed mainly of rounded to sub-rounded pebbles of Carboniferous rocks. It was concluded by Gaunt (1976b) that these gravels, which are almost entirely derived from the Coal Measures, do not correspond to the 'east Pennine suite' of erratics found in the 'Pennine Boulder Clay' based on the absence of pebbles of Carboniferous and Permian limestone and of Lake District rocks.

At Great Heck (Fig. 17) sand and gravel, underlying the 20 Metre Surface, were formerly exposed in a sand pit over Triassic sandstone. Here, Parsons (1887) has described a two-fold sequence of gravels overlying cryoturbated sandstone (Fig. 23).

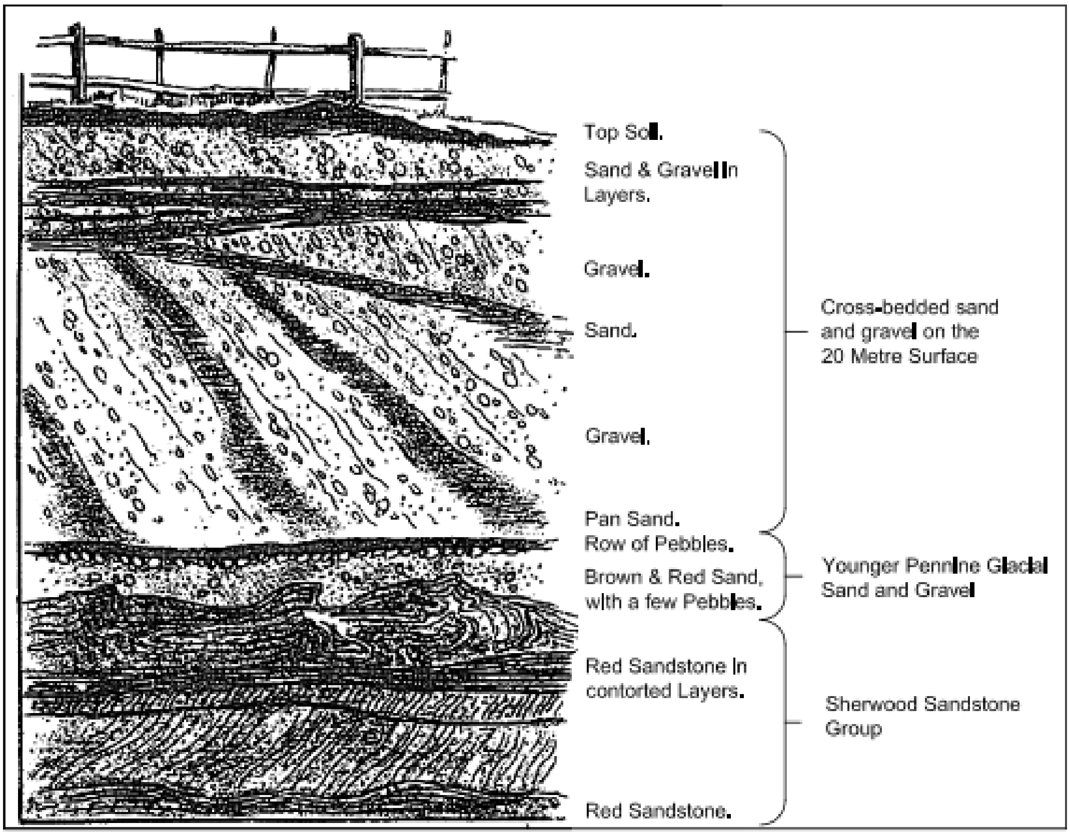


Figure 23. Section near Great Heck Station (dismantled) SSE of Hensall (Fig. 17) showing cross-bedded sand and gravel (*cf.* Glacio-fluvial Sand and Gravel of Gaunt, 1976b) overlying Younger Pennine Glacial Sand and Gravel (Gaunt, 1976b) on cryoturbated Triassic sandstone. (The Pan Sand row of pebbles is about 2.5 cm thick and contains fist-sized pebbles.) From Parsons (1887) – no scale given.

The basal gravel, or 'pan sand row of pebbles', comprises a single layer of fist-sized pebbles, covered by clay-cemented sand, and rests on cryoturbated sandstone containing cavities filled with sand and gravel. This sequence is overlain by steeply, southerly dipping, cross-bedded sand and gravel, which forms the 20 Metre Surface. Parsons (1887) describes these and similar gravels at Whitley and Pollington (it is presumed he refers to the upper sequence) as consisting of rounded boulders of Coal Measure sandstone and Millstone Grit up to about six inches in diameter: he noted that pebbles of Carboniferous and Permian limestone were entirely absent. Parsons (1887) did however record flat fragments of Magnesian Limestone from gravels, at a lower level (probably at or below 10 m) from a quarry near Hensall Station [SE 585 230]; a similar observation was made by the writer from a quarry to the east of Hensall [SE 598 235]. It is however likely that the gravels east of Hensall represent the basal gravels at Great Heck and not the upper sequence. Parsons (1887) has therefore, like Gaunt (1976b; see section 3.3.1.2.), recognised two gravel sequences: an older set containing pebbles of Magnesian Limestone and a younger set mainly of Carboniferous sandstone and devoid of Magnesian Limestone.

An important observation, that results from the recognition of the 20 Metre and 15 Metre Surfaces, is the similarity between the sections at Great Heck (Fig. 23) and the section presented by de Boer *et al*. (1958) from Prescott's Pit (Fig. 22). At both locations an erosion surface, at *c*. 15 m, is capped by a thin compact sand layer, which may represent a residual deposit or a paleosol on a periglacial land surfaces overlain by steeply dipping cross-bedded sand and gravel over 1.0 m thick. The foresets of these cross-bedded deposits, are, as was observed earlier, anomalous in terms of the proposed multi-stage decline model for Lake Humber, as 45° dips are too steep for natural deposition in shoreline deposits but are likely in a vigorous fluvial regime (e.g. Hodgson, 1978). This anomaly could be resolved if these deposits are equivalent to the 'Glacio-fluvial Sand and Gravel', described by Gaunt (1976b), from now inaccessible quarries on the Snaith Ridge, near Eggborough (Fig. 17), which he considers could be considerably older than the Late Devensian glacial maximum. An account of these pre-Devensian glacial deposits and older 'Pennine Boulder Clay' is included in this text (see section 3.3.1.).

7.8.4. Eskers

That linear banks of gravel in the Ouse valley, south of York, originated as eskers has long been accepted with no firm alternative views being suggested for their genesis. The earliest use of the term esker can probably be attributed to Clark (1881) who referred to the ridge of the York Moraine that extends from York to beyond Bilbrough for a distance of 14 miles as a 'Scotch Esker'. Whilst this assertion has never been accepted, Clark (1881) did make an important historical contribution in his reference to workings in gravel beds stretching down both sides of the River Ouse from York to

Fulford and on the Bishopthorpe road: ridges that form part of the 20 Metre Surface (Fig. 14). Clark (1881) also reported the discovery of a well-preserved bone recovered from a pit near the Bishopthorpe road that was later identified as the metatarsal of a bear (*Ursus spelaeus*). In a note, also of some historical interest, Kendall & Wroot (1924) refer to gravel ridges south of York, adjacent to Knavesmire (Fig. 14), as esker ridges and their significance in providing two of the mediaeval roads to the south, one or both of which may have been used by Harold in his march to Stamford Bridge in 1066. Kendall & Wroot (1924) also describes a 'curious ridge' of gravel that connects the York and Escrick moraines (through Crockey Hill) as an esker or elongated delta in which the 'bones of the great mammalia have been found'. This landform was referred to as a 'well-developed esker-like ridge' by Cooper & Burgess (1993) who attributed its origin to stagnation of the Vale of York ice sheet as the ice front retreated. Palmer (1966) has similarly suggested that the gravel ridge between York and Escrick is a clear example of esker formation from decay of ice by stagnation. It was formalised as the Crockey Hill 'Esker' Member of the Vale of York Formation on the Selby 1: 50 000 geological map (British Geological Survey, 2008).

Despite its location south of the York Moraine, the Crockey Hill 'esker' was included as part of an esker belt extending for about 50 km to include the Helperby-Aldwark esker system on the Harrogate 1: 50 000 geological map (British Geological Survey, 1987). Evidence for this tenuous extrapolation is mainly based on its alignment with sand and gravel deposits at Linton-on-Ouse and beneath later glaciolacustrine deposits (proven in boreholes) as far as Beningbrough (Cooper & Burgess, 1993, fig.29). Cooper & Burgess (1993) also include in this alignment spreads of sand and gravel at York (presumably south of York) but with no precise location given.

On the Selby 1:50 000 geological map (British Geological Survey, 2008) the Crockey Hill 'esker' has been shown as a linear outcrop of sand and gravel extending from Escrick to Crockey Hill essentially following the alignment of the A19 (see also Fig. 14). North of Crockey Hill the esker (or esker belt) expands to a width of nearly 2.0 km towards Fulford where it extends along the east bank of the River Ouse. Gravels along the west bank of the river adjacent to the Bishopthorpe road, referred to by Clark (1881), are not included as part of the Crockey Hill 'esker', but have been included in the York Moraine Member, a unit composed of sand and gravel (British Geological Survey, 2008). In general, the mapped Crockey Hill 'esker' changes its morphology from a flat-topped narrow ridge at about 15 m above OD, south of Crockey Hill, to a broader level land surface at about 20m above OD in the region of Fulford.

In the various accounts of eskers, in the Vale of York, detailed descriptions of their morphology is lacking. However it can be presumed that reference to stagnant ice implies an origin as ice-contact ridges and probably to classical tunnel-fill deposits (Warren & Ashley, 1994).

Landform mapping by Fairburn (2009, included in Fig. 14), indicates that the 20 Metre Surface south of York is represented by three gravel ridges separated by the River Ouse and probable earlier channels of the Ouse that now form Knavesmire and Askam Bog/Hob Moor. It therefore appears likely that the 20 Metre Surface could have originally been a much more extensive terrace, with eroded remnants now in the subsurface overlain by younger deposits including laminated clays. One such buried eroded remnant may have been observed by Kendall & Wroot (1924) who recorded laminated clays draped over ‘a great hemispherical mound of gravel’ in a clay pit near Hogs Pond, which lies in the drainage channel through Hob Moor and Askam Bog (Fig. 14). South of Crockey Hill the gravel ridge persists as a flat-topped landform to Gravel Pit Farm (Fig. 14) where it is part of the 15 m terrace. The lower surface probably represents a lower lake level, but loss of gravel towards Escrick, where the 15 m terrace can be underlain by till or by residual sand and gravel on till, could be a progradational effect resulting from a diminishing supply of coarse sediment.

Despite the prominence of these gravel ridges, the lithology of the sediment forming them is only known from surface ‘float’ and from a few overgrown or partly infilled gravel pits. These pits include several grass covered hollows at Crockey Hill [SE 626 425] and Gravel Pit Farm [SE 626 452], and a pit north of Middlethorpe near the Bishopthorpe road [SE 601 496], where exposures in the gravel are largely obscured by infilling waste. At this last locality, which lies at the top of a low scarp above river alluvium, the previously worked deposit of poorly sorted gravels contains mainly sandstone clasts ranging from faceted cobbles to well rounded small pebbles (Plate 26): several sand layers in the deposit have been picked-out by rabbits. This pit may well be one of many referred to by Clark (1881) in the vicinity of Knavesmire and the Bishopthorpe Road.

Plate 26. Poorly sorted gravels on the slope of a gravel pit in the River Ouse fan delta north of Middlethorpe. (Photograph W.A. Fairburn)

Based on the results of the landform mapping, it is concluded that the Crockey Hill 'esker' is a distal remnant of a once more extensive 20 m terrace now preserved as ridges between glacial or post-glacial drainage channels. This conclusion is based on the belief that horizontal terracing in the Vale of York originated as shorelines of Lake Humber long after the Vale of York glacier had retreated, so that the Crockey Hill 'esker' could not have originated from stagnant ice. It would also seem unlikely that any eskers to the north of York on the Harrogate sheet (British Geological Survey, 1987), developed in till, could form part of an esker system extrapolated through the York Moraine. The ridge forming the Crockey Hill 'esker' also lacks sinuosity, while its defining western edge, flanked by the 10 Metre Surface, between Crockey Hill and Gravel Pit Farm (Fig. 14), is more consistent with an erosional event rather than a depositional one.

As with the expanse of sediment where the River Derwent discharges into Lake Humber, it is considered here that Crockey Hill is a fan delta as defined by Tucker (1991) and described below.

7.8.5. Fan Deltas

Nemec (1990a), in a review of the terminology and classification of deltas, has suggested a first stage broad subdivision of deltas into alluvial and non-alluvial deltas:

two specific examples of his alluvial deltas, which are applicable to this text, are fan deltas and scree-cone or scree-apron deltas (Fig. 24). The second type of delta, described by Nemec (1990b) as steep-face, coarse-grained deltas are akin to the sand screes described in section 7.16.1.1.

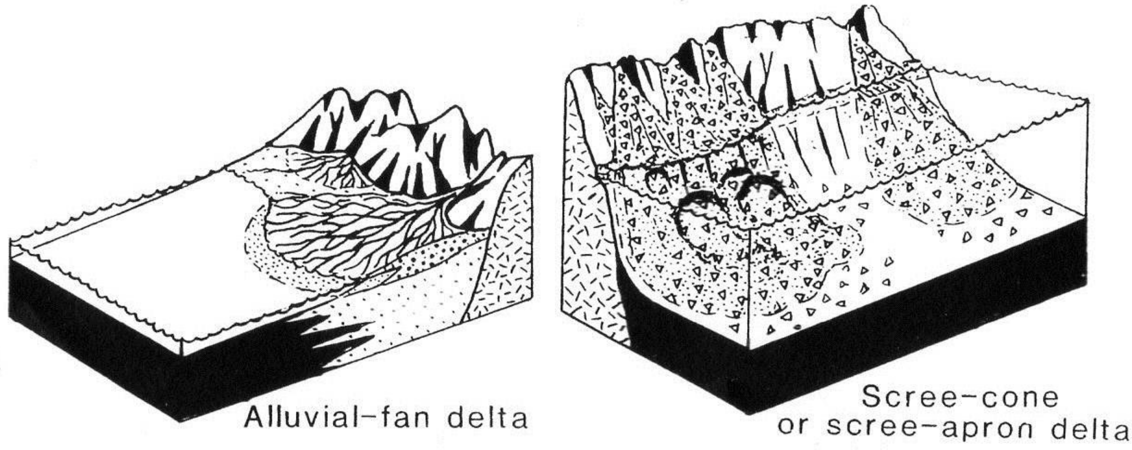


Figure 24. Two types of fan deltas, associated with the York Moraine, that have been identified in the Vale of York. Note the spoon-shaped slump scars on the steep-faced scree-cone delta. Illustrations from Nemec (1990a).

As already stated (sections 7.8.2. and 7.8.4.) it is believed that the wedges of sediment deposited into Lake Humber at the 20 m above OD level south of the York Moraine at York and Stamford Bridge originated as fan deltas. Support for this conclusion is contained in a description by Tucker (1991) who has defined fan deltas as 'cones of sediment shed directly from a source area ……. through canyons to alluvial fans which pass straight into the sea or lake as the fan delta'. Internally fan deltas mostly consist of steeply dipping foresets of coarse sand and gravel topped by flat-bedded or smaller-scale cross-bedded sands with a bottom-set of fine sediment (Tucker, 1991, fig. 2.86). In an account of Pleistocene fan deltas in south Ontario, Martini (1990) has described various settings developed under the influence of Pleistocene continental glaciers. These settings include fan deltas in contact with the glacier; those fed by glacial streams; or those at the end of a narrow bedrock valley, which channelled glacial meltwater. In an example of the last type, which most resembles the environment of the York Moraine fan deltas, Martini (1990) has described a fluvial-dominated fan delta at Joe Lake, in southern Ontario, which formed at the end of a torrential valley. This fan delta contains large foresets up to 20 m thick, with dips up to 25° - 30°, composed of poorly-sorted sandy gravel, which may contain some large boulders.

In contrast with the Canadian example, the internal fabric of the postulated York Moraine fan deltas is virtually unknown due to lack of sections, with the only outcrop of note being the Middlethorpe gravel pit (Plate 26) where only degraded surfaces exist. Despite the lack of detailed sedimentology, the fluvioglacial setting of these gravel deposits at meltwater channel exits through the York Moraine into Lake Humber as at Stamford Bridge, coupled with their spatial extent does suggest these deposits are fan deltas rather than eskers. It is also likely that the meltwater channels were torrential, as the proto-Derwent at Stamford Bridge would have had a head of water back to Lake Eskdale (Fig. 9), and Lake Humber (Lake Alne) to the north of York would have had a lake level some 5.0 m higher than the lake to the south (see section 7.7.1.).

7.9. 15 Metre Terrace (Stage 6)

A distinct 15 m terrace, as part of the 15 Metre Surface, defined by a 15 m above OD back-wall and a 10 m above OD depositional front edge, is not represented everywhere in the mapped area. It resulted from the progressive shallowing of Lake Humber, with a southerly retreat of the northern shoreline towards the Escrick Moraine and a contraction of the lake westwards from the Wolds and eastwards from the Permian escarpment. It is mainly restricted to the south of a line curving north-westwards from North Cliffe (Fig. 16) towards Elvington on the River Derwent, and then west-southwest to include Crockey Hill, Naburn and Appleton Roebuck (Fig. 14). To the north of this line, Lake Humber had silted or sanded-up, and any 15 m terrace is mantled by an influx of fluvial sand (the Breighton Sand Formation) on to the Vale of York in the Early Holocene (see OSL dating, section 9). The 15 m terrace and the 15 Metre Surface has therefore only been mapped on the sides of the Escrick Moraine, north and northeast of the moraine towards Crockey Hill, Naburn and Appleton Roebuck (Fig. 14), around the Brayton Barff and Snaith Triassic ridges (Fig. 17), at several locations along the edge of the Permian escarpment, south of Tadcaster (Fig. 17) and on the flanks of the Wolds between North Cliffe and Hessle (Fig. 16).

Similar observations were made by Palmer (1966) who recorded beaches at a 50 ft (15.2 m) lake level and suggested that Lake Humber may have drained from the 100 ft lake level and reformed at 50 ft.

7.9.1. York Moraine

From the River Wharfe, in the west, to the River Derwent, in the east, there is a distinct topographic boundary, at just below 15 m above OD, between the level plain of the Vale of York and the rising ground formed by the depositional front edge of the 20 m terrace. This boundary is not only a lacustrine shoreline marking the maximum northerly extent of lacustrine laminated clays south of the York Moraine, but also the northern boundary to the flooding events that brought extensive fluvial deposits (Breighton Sand Formation) into this part of Lake Humber (see British Geological Survey, 2008). It also defines an edge for re-worked sand scree delta sands that had been deposited into an earlier sub-aqueous environment south of the York Moraine. The 15 m strandline is also recognisable as the boundary around projections of the York Moraine that extend south of Bilbrough towards Appleton Roebuck and Bolton Percy (Fig. 14). In this region, the moraine, based on mapping by the British Geological Survey (2008), is bounded either by laminated clays, of the Elvington Glaciolacustrine Formation, or till underlying the plain of the Vale of York.

Between the York and the Escrick Moraines, the 15 m terrace forms a widespread planar landform, that is underlain by fan-delta gravels from Crockey Hill to Gravel Pit Farm and more extensively by till in areas between Stillingfleet and Escrick, east of Deighton and southeast of Naburn (Fig. 14; British Geological Survey, 2008).

7.9.2. Escrick Moraine

Apart from an area near High Catton, where a 25 m terrace has been mapped (Fig. 15), much of the Escrick Moraine remained submerged below the surface of Lake Humber until the lake had receded to 15 m above OD. At this lake elevation, most of the crest of the moraine was exposed between Stillingfleet and Wheldrake (Fig. 14) and north of the River Derwent towards Wilberfoss (Fig. 15). Erosion of a strandline at this level is evident on the moraine from Stillingfleet to Wheldrake, while the 15 m contour, east of the River Derwent, coincides with the formline that separates the rising edge of the moraine from sands and clays underlying the plain of the Vale of York, here mapped by the British Geological Survey (2008), as the Bielby Sand Member of the Breighton Sand Formation and the Thorganby Clay Member of the Hemingbrough Glaciolacustrine Formation. In the region of High Catton (Fig. 15) the boundary edge of the moraine has risen further in elevation to form the back-wall of the 20 m terrace (Fig. 15).

7.9.3. The Wolds

The 15 m strandline between Stamford Bridge and Full Sutton (Fig. 15), and possibly as far south as Ling Lane, while marking the southern and western frontal edge of the 20

Metre Surface, probably also marks an erosional termination to the River Derwent fan delta system, which further south probably extends beneath the Bielby Sand Member of the Breighton Sand Formation. On the Selby 1:50 000 geological map (British Geological Survey, 2008) the 15 m strandline has not been recognised as much of the 20 Metre Surface is covered by the Bielby Sand Member. South of Ling Lane, to beyond Market Weighton, the 15 m strandline continues as a mappable feature just below the 15 m contour. To the west the 15 m terrace, as in the area south of the York Moraine, merges into the plain of the Vale of York, which generally lies at an elevation of *c*. 10 m: the plain being mapped as either part of the Breighton Sand Formation (British Geological Survey, 2008) or as undifferentiated sand and gravel of the 25 Foot Drift (British Geological Survey, 1995).

South of North Cliffe, the 15 m terrace becomes a recognisable landform as part of the 15 Metre Surface, which can be traced along the base of the Wolds to beyond Hessle (Fig. 16). The feature is significant as it contains the lowest shorelines of Lake Humber etched into the alluvial fan gravels that cover the western slope of the Wolds as far as the Humber Bridge Country Park (Fig. 16). From here to north of Hessle the 15 m terrace has been eroded into the Skipsea (Hessle) Till, again confirming that the Skipsea Till did not provide a lake-impounding barrier across the Humber Gap. The decision to map the 15 Metre Surface at Hessle as a landform and not part of the Skipsea Till is based on the strong imprint of the terrace near Hessle that suggests substantial reworking of the till at the 15 m above OD level.

Gaunt (1976b) in an account of gravel deposits between North Cave and Brough refers to gravels at elevations between 11 m and 23 m above OD near South Cave and between 11 m and 33 m above OD in the region of Ellerker and Brough. These deposits are included by Gaunt (1976b) as 'Older Littoral Sand and Gravel' and therefore akin to similar gravel deposits mapped previously on the older Selby 1: 50 000 sheet (British Geological Survey, 1973b). It is worth noting however that widespread gravel deposits in the region, referred to in the Ellerker-Brough region, are not shown on the Kingston upon Hull 1: 50 000 sheet (British Geological Survey, 1983a). Gaunt (1976b) concluded from surface evidence that the gravels between North and South Cave consist mainly of brown fine-grained sand and silty sand with locally abundant pebbles of sub-angular to sub-rounded flint and sub-rounded chalk. During fieldwork sand with a few flints was also noted in a ditch north of Everthorpe Hall [SE 897 313] at a location on the 15 m terrace. The distribution and lithology of these gravels clearly indicates that they are equivalent to the alluvial fan deposits described in this text that have been terraced at levels between 10 m and 40 m above OD.

The only gravel in the Ellerker – Elloughton region that may not have been subjected to Late Devensian lacustrine terracing are the enigmatic deposits at Prescott's Pit referred to on the Kingston upon Hull 1: 50 000 sheet (British Geological Survey, 1983a) as the 'Cockle Pits'. These gravel deposits, described earlier in the text (section 7.8.2., Fig. 22, Plate 25), are considered to be fluvial in origin and therefore unrelated to the history of Lake Humber. Easterly dipping, cross-bedded fluvial sands and gravel underlying the 15 m terrace and deposited on an erosional surface above the Skipsea Till (Plate 27), do however occur at Red Cliff [SE 980 250] on the north bank of the River Humber. At this location, much closer to the Humber outlet, fluvial deposition is more likely and would be expected where Lake Humber drains through the Humber Gap.

Plate 27. Easterly dipping cross-bedded sands and gravels underlying the 15 Metre Surface exposed at Red Cliff on the bank of the River Humber. The gravels were deposited on an erosional surface above the Skipsea Till and have been decalcified near the top of the 4.0 m section. (Photograph W.A. Fairburn)

7.9.4. Triassic Ridges

The 15 m terrace is a more widespread feature than the 20 m terrace as it forms a low platform, particularly around Brayton Barff and Hambleton Hough, that is barely perceptible in places above the plain of the Vale of York. In the central part of the Snaith Ridge the 15 m terrace has reworked gravels from the 20 Metre Surface but to the east of Eggborough the 15 Metre Surface covers degraded parts of the Snaith Ridge represented by low mounds of Sherwood Sandstone from which these littoral deposits have been eroded.

7.9.4.1. Brayton Barff

The 15 m terrace forms a raised platform around Brayton Barff and Hambleton Hough (Fig. 17) that has a back-wall at nearly 15 m above OD eroded into the Sherwood Sandstone forming both these hills. A frontal edge of the platform, at 9 m – 10 m above OD, rises above the plain of the Vale of York, which is underlain by silt and clay of the 10 Metre Surface (the 25-Foot Drift of the Vale of York). In places the 10 Metre Surface can be overlain by up to 1.0 m of cross-bedded red sand derived from the Sherwood Sandstone (noted in a pipeline trench near Burn [SE 588 280]). On the Goole 1: 50 000 sheet (British Geological Survey, 1971) much of this platform has been mapped as Bunter Sandstone (Sherwood Sandstone) with only patches of glacial sand and gravel. Scattered pebbles and cobbles on the 15 m terrace principally consist of Carboniferous sandstone derived by progressive reworking from poorly sorted gravel beds that occur near the summit of both Brayton Barff and Hambleton Hough (Gaunt, 1994, fig. 37). These gravels are associated with clay till and conform to an 'east Pennine suite' of erratics (Gaunt, 1994).

7.9.4.2. Snaith Ridge

The 15 m terrace on the Snaith Ridge extends from Kellingley in the west to Pollington and West Cowick in the east (Fig. 17) and is more extensive than the 20 m terrace as it flanks this terrace around the crest of the ridge near Kellingley and at several locations near Great Heck and Pollington. Unlike the 20 m terrace, exposed sections through the sand and gravel of the 15 m terrace are still accessible in some disused sand pits though many of these abandoned pits are overgrown or have been landscaped for other purposes and informative sections have been lost. Areas of the 15 m terrace are broadly similar to those of 'Glacial Sand and Gravel' mapped on the Goole 1: 50 000 sheet (British Geological Survey, 1971) and Glaciofluvial Deposits' mapped on the Wakefield 1: 50 000 sheet (British Geological Survey, 1998).

The 15 m terrace along the Snaith Ridge is a somewhat unusual feature, in the Vale of York, in that sections through the sand and gravel deposits underlying the terrace are exposed in a number of disused sand pits. While clean sections are either not available or totally accessible, the lithology of these deposits is sometimes clearly visible. For example in the north face of a sand pit, near Pollington [SE 612 201], poorly bedded yellowish sand containing some gravel layers covers a mound of planar-bedded and cross-bedded Sherwood Sandstone, from which the sand was derived (Plate 28).

Plate 28. Section in the sand pit at Pollington composed of roughly bedded littoral sand and gravel, containing the sampled sand lens above the 30 cm scale, forms the 15 Metre Surface over horizontally-bedded weathered sand on cross-bedded Sherwood Sandstone. The gravel, formed mainly from clasts of Carboniferous sandstone, is overlain by a thin dark soil layer below a gravel 'pavement' forming the base for quarry waste. (Photograph M.D. Bateman)

A similar association, although more poorly exposed, can also be seen in a sand pit near Hensall railway station [SE 585 229]. East of Hensall a 2.0 m section through the 15 m terrace is exposed along the western face of a sand pit [SE 597 235], where post-operational landscaping has now restricted access to some of the faces. Here deposits on the eroded surface of the Sherwood Sandstone consist of a rather chaotic assemblage of sand and quartz gravel, mainly derived from the Sherwood Sandstone (Plate 29).

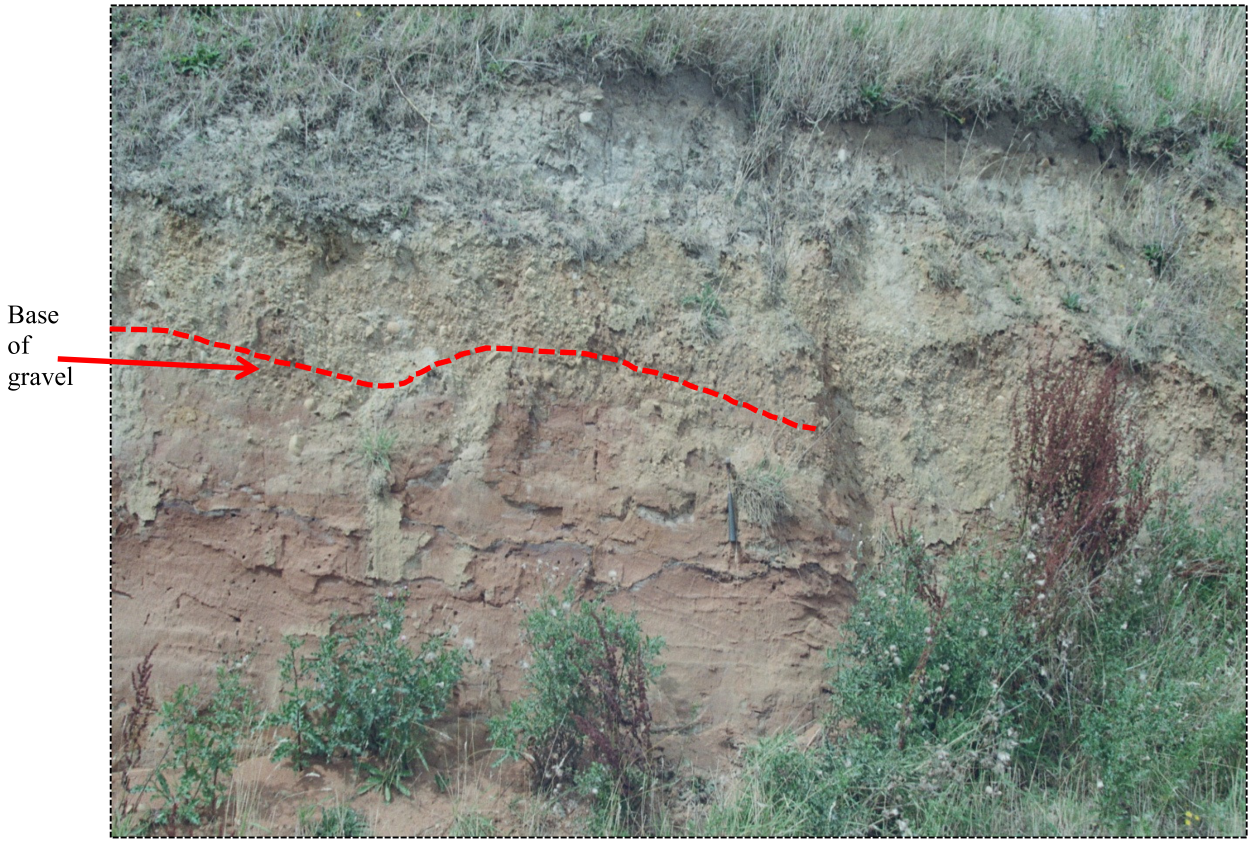


Plate 29. A 2.0 m section of sand and quartz gravel, bleached near the surface, overlying an erosional surface on Sherwood Sandstone. The section is exposed in the wall of a sand pit [SE 597 235] east of Hensall.Note the rather obscure contact between the gravel and the Sherwood Sandstone. (Photograph W.A. Fairburn)

In all the above sections, the nature of the sand and gravel deposits implies a littoral origin as there is no evidence for fluvial deposition. This environment contrasts with the cross-bedded fluvial deposits described earlier (section 7.8.2. and 7.8.3.) from Prescott's Pit (Fig. 22) and Great Heck (Fig. 23) that form part of the 20 m terrace.

In the central part of the Snaith Ridge, adjacent to the 20 Metre Surface near Kellington (Fig. 17), deposits on the 15 m terrace would have been derived by shoreline reworking of gravels from the 20 Metre Surface and not from the Sherwood Sandstone. Pebble assemblages noted here in cultivated fields, on this surface, conform to a conclusion by Gaunt (1976b) that gravels on the Snaith Ridge, between 7 m and 17 m, are almost entirely from the Coal Measures, with an absence of pebbles of Carboniferous and Permian limestone and Lake District rocks (i.e. the 'east Pennine suite', see section 3.3.1.).

7.9.5. The Permian Escarpment

Fragments of the 15 m terrace occur along the eastern edge of the Permian escarpment at Towton, South Milford, Monk Fryston and Hallam (Fig. 17). In all cases the terrace lies between the 10 m and 15 m contours with clasts on the terrace being mainly locally derived angular fragments and slabs of Magnesian Limestone set in a matrix of yellow clay. The deposits can also contain abundant sub-rounded or faceted pebbles of Carboniferous sandstone derived from nearby exposures of till; a circumstance that has led to one occurrence, near Monk Fryston [SE 501 292], on the Wakefield 1: 50 000 sheet (British Geological Survey, 1998) to be incorrectly identified as till.

7.10. 10 Metre Terrace (Stage 7)

In practical terms of mapping the 10 Metre Surface occupies parts of the Vale of York shown on Figures 14-17, that are lower in elevation than the 15 m terrace. This wide, mainly featureless, plain, which extends in places from the Permian escarpment in the west to the foothills of the Wolds in the east, is underlain by a vast mass of generally laminated clay described by Gaunt (1976b) as the largest Quaternary deposit in the southern part of the Vale of York. The clay, which is virtually stoneless and over 20 m thick in some localities, plus associated sands and silts, was named the '25 Foot Drifts' by Edwards (1937) as much of the landform it occupies lies below 25 ft (7.62 m). Edwards (1937) considered the upper beds of this unit belonged to a late glacial stage (presumably Late Devensian). This nomenclature, proposed by Edwards (1937) was later incorporated into the mapping of the British Geological Survey e.g. the Goole and older Selby 1:50 000 sheets (British Geological Survey, 1971, 1973b). In later mapping, the term '25 Foot Drifts' was replaced by 'Glaciolacustrine Deposits' of the Vale of York e.g. the Wakefield and Leeds 1:50 000 sheets (British Geological Survey, 1998 and 2003).

In terms of the evaluation of Lake Humber, which is the main concern of this current research, the 10 Metre Surface marks the final full phase of lacustrine and shoreline deposition in the Vale of York: a landform that is mainly underlain by laminated clays and to a lesser extent by erosional surfaces on Triassic sandstone or marl with some minor aggradation. Landform mapping (Figs. 14 – 17) confirms that the 10 m terrace, south of the York Moraine, lies almost everywhere below 10 m above OD while north of the moraine contemporary lacustrine deposition, for instance in the Hob Moor region, occurred at higher elevations. Erosion of this surface by river incision, fluvial flooding and emplacement of a 5 m erosional terrace extended into the Holocene.

Recent mapping by the British Geological Survey on the newer Selby 1:50 000 sheet (British Geological Survey, 2008) has now segregated the '25 Foot Drifts' into two components: the lacustrine clays, and the younger Breighton Sand Formation, which has an erosional contact with the underlying clay formation (Ford *et al*., 2008). A geographical separation of the laminated clays was also proposed by Ford *et al*. (2008) with the Thorganby Clay Member of the Hemingbrough Glaciolacustrine Formation mapped to the south of the Escrick Moraine; the Elvington Glaciolacustrine Formation to the north of the Escrick Moraine and the Alne Glaciolacustrine Formation, in the Hob Moor region to the north of the York Moraine (Ford *et al*., fig. 6; Fig. 45). The landform mapping suggests this regional nomenclature is over simplified, as deposition of the laminated clays probably commenced soon after the initial regressional stages of Lake Humber in a number of discrete sub-basins within the Vale of York. These sub-basins, probably evolved with time, separated initially by the York and Escrick Moraines, by extensions of the York Moraine south of Bilbrough, the east-west Triassic ridges and later by the coarse-grained deltas discharging from the River Derwent and through Crockey Hill. The Hobb Moor deposit, for instance post-dates erosion of the 20 m terrace south of York in the Ouse valley, whereas other basins formed later.

Gaunt (1974) has stated that the '25 Foot Drifts' were deposited during a prolonged stand of low-level Lake Humber at 10 m to 14 m above OD with the northerly rise in this depositional surface attributed by Gaunt (1981) to isostatic depression. This conclusion cannot now be considered tenable as the northerly rise in the plain of the Vale of York can be attributed to fluvial and deltaic sand deposition south of the York Moraine. However an observation by Gaunt (1981) that Lake Humber finally disappeared because it silted-up conforms with the landform mapping.

The most informative data on the nature of the laminated clays beneath the 10 Metre Surface has been provided by sections in excavations that are now infilled or inaccessible. In Dakyns *et al*. (1886), a section drawn by J.C. Ward of a drainage ditch 4.0 ft – 5.0 ft (1.2 m – 1.5 m) deep extending north of Escrick Grange Farm [SE 619 419], beside a disused railway line, showed about 1.0 m of laminated clay draped over an irregularly eroded surface of till (Fig. 25). These laminated clays, referred to by Dakyns *et al*. (1886) as 'Warp Clay', are of lacustrine origin and were not deposited as tidal alluvium or from artificial flooding.

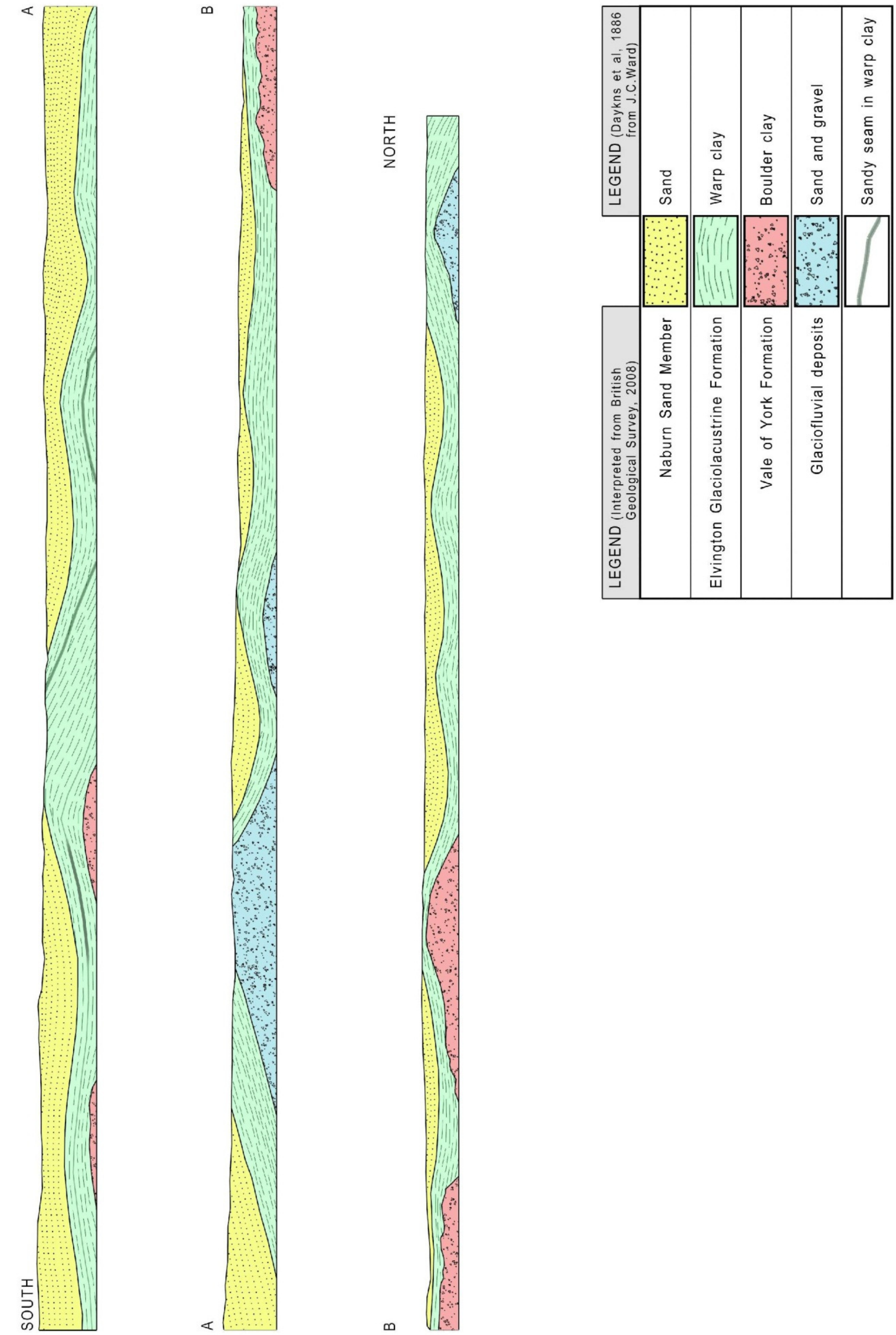


Figure 25. Section exposed in a ditch along the now disused railway line between Selby and York; beginning close to Escrick Grange Farm [SE 619 419] and extending northwards towards Naburn Wood (Figs 14 and 37). From a drawing by J.C. Ward in Dakyns, 1886, scale not given.

A similar draping effect was noted by Kendall & Wroot,1924, in a clay pit near Hogs Pond (see section 7.8.4.). For most of the section (Fig. 25), the clay appears to be conformably overlain by fluvial sand of the Naburn Sand Member of the Breighton Sand Formation. This relationship is unlikely as the Naburn Sand Member is defined as having an erosional lower contact (Ford *et al*., 2008). A more plausible erosional contact between the clay and the fluvial sand is shown in a section across the Escrick Moraine near Wheldrake (Fig. 31, from Gaunt, 1970) where the sands appear to 'feather' into the bounding laminated clays. A similar section to the one in the railway drainage ditch can be partially seen in the recently (2007) re-excavated Wood Dike which runs from north of Naburn Wood into the River Ouse (Fig. 37) near Bell Hall [SE 596 436].

In another early account of the laminated clay Parsons (1887) describes the bed as very constant in character over a large area with a thickness of 48 ft (14.6 m) at Selby and 57 ft (17.4 m) at Cawood (Fig. 17).

An account of the 'Silt and Clay of the 25 Foot Drift', by Gaunt (1976b), describes numerous sections and boreholes in the deposit. His description of two lengthy pipeline trenches extending from Elvington, on the River Derwent, to southwest of Brayton Barff shows that the clay is monotonously laminated below depths of 1.0 m to 1.5 m, with laminae of cross-bedded silt or very fine-grained sand traceable for distances perhaps greater than half a kilometre. Most of these sand beds were noticed towards the base of the clay (Gaunt, 1976b). An extension of the trench southwest of Brayton Barff towards the River Aire, west of the Selby Canal (Fig. 17), exposed laminated clay to depths of up to 3.4 m. Some of the sandy laminae here contain finely divided detrital coal and exhibit southerly dipping cross-bedding and small-scale ripples up to 20 mm – 30 mm wide and approximately 2.0 mm thick.

Despite widespread exploitation for brick clays in clay pits and lengthy pipeline and drainage trenching, sections through the laminated clays are not now common, although field evidence supported by shallow drilling has enabled Ford *et al*. (2008) to give maximum thicknesses of 1.0 m – 4.0 m for the Thorganby Clay Member, 3.0 m – 6.0 m for the Elvington Glaciolacustrine Formation and 2.5 m for the Alne Glaciolacustrine Formation. Sections that are still accessible through the laminated clays are probably now restricted to clay pits at Hemingbrough [SE 674 317] logged by Murton *et al*.

(2009) and Newton upon Derwent [SE 727 503] referred to by Ford *et al*. (2008) as supplying evidence for the Escrick Moraine being pushed into the lacustrine deposits. Both these clay pits expose the Thorganby Clay member as the uppermost unit. Of particular importance at the Newton upon Derwent pit are the discontinuous sand lenses, up to 15 cm thick (Plate 30), one of which was sampled for OSL dating.

In areas where the laminated clays are absent below the 10 Metre Surface, such as near Brayton Barff and the Snaith Ridge, where the Sherwood Sandstone subcrops at shallow depths, the 10 Metre Surface is often underlain by a thin sequence of cross-bedded or rippled red sand. These sequences have been noted in a pipeline trench at Burn [SE 588 280] and in sand pits at Hensall (Plate 38) and Pollington (Plate 39). Although generally less than 1.0 m thick, the sands at Pollington (Plate 39) form the basal part of a succession that contains gently, southerly dipping gravel bedforms up to 3.0 m thick. Such southerly dipping beds could have originated from Lake Humber drainage towards the Humber Gap outlet.

Plate 30. Smeared sand lens, approximately 10 cm thick, within dark blue-grey laminated clay of the Thorganby Clay Member, in the Newton upon Derwent clay pit [SE727 503]. (Photograph W.A. Fairburn)

7.11. 5 Metre Terrace (Stage 8)

The final phase in the regression of Lake Humber (Stage 8) is represented by a shallow lake with a 5.0 m above OD shoreline (an area still subject to flooding at the present day) and river terracing intermediate between the 10 Metre Surface and modern river floodplains.

A terrace, on the River Ouse, some 10 ft (3.0 m) above river alluvium was recognised by Melmore (1940) extending from Beningbrough to Poppleton along the right bank of the river. This terrace, which Melmore (1940) named the 10 ft terrace, was also identified at Holgate Beck and forms part of the playing fields at St Peter's School (Fig. 14). As a significant landform this feature was mapped as the 'Older River Terraces' by Fairburn (2009) and is included with this title on Figure 14. The mapping on Figure 14 shows the extent of the terrace between Beningbrough and Holgate Beck, which extends into a meltwater channel between Severs Howe and Knapton, north of Kexby Bridge on the River Derwent, and part of the Bridge Dike – Dunning Dike – Horsecourse Dike drainage system. The terrace also occupies the Knavesmire – Middlethorpe area, although its elevation here is possibly only slightly above alluvium of the nearby River Ouse. On the Selby 1: 50 000 sheet, the British Geological Survey (2008) shows this area to be underlain by laminated clay of the Elvington Glaciolacustrine Formation.

Later mapping (Fig. 17) has shown that the 'Older River Terraces' expand south of Wistow into an extensive floodplain, flanking the River Ouse, that extends through Selby and beyond the River Aire to at least as far south as Pollington and West Cowick. In places where the floodplain surrounds the villages of Camblesforth, Carlton and Drax (Fig. 17) it is over 6.0 km wide. On the north bank of the River Aire, above Gowdall, a ridge of higher ground, possibly a natural levee, separates the now named 5 Metre Terrace from alluvium bordering the Aire River.

The edge of the 5 Metre Terrace, where it abuts against the older 10 Metre Surface is generally a recognisable feature throughout the mapped area. South of Snaith, for instance, just north of the M62 [SE 648 210] the edge of the 5 Metre Terrace is quite conspicuous (Plate 31).

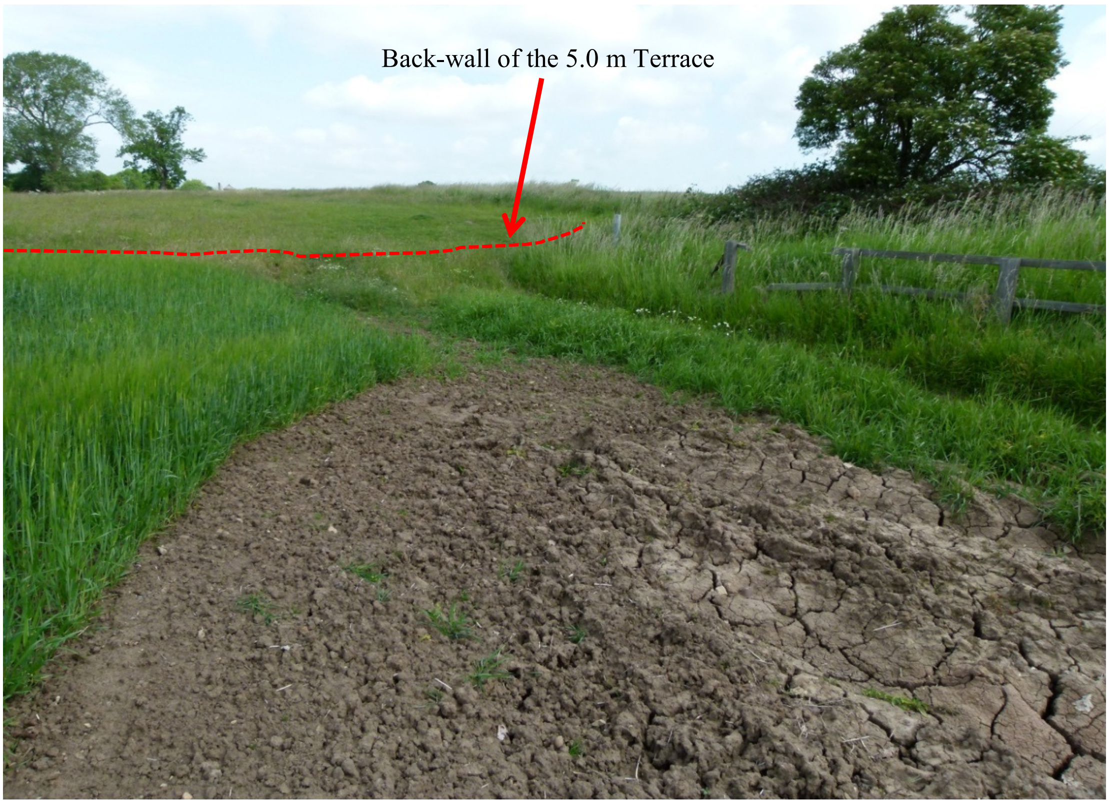


Plate 31. The edge of the 5 Metre Terrace rising about 2.0 m to the 10 Metre Surface south of Snaith [SE 648 210]. (Photograph W.A.Fairburn)

Elsewhere to the north of the River Aire the edge of the floodplain has been mapped just below the 5.0 m contour until the feature is lost among housing east of Selby near the River Ouse.

Isolated remnants of the 10 Metre Surface, surrounded by the 5 Metre Surface, include ‘islands’ of Triassic sandstone occupied by the villages of Camblesforth and Carlton. Historically the older houses, in these villages, were not constructed below the distinct boundary that marks the change in elevation above the 5.0 m contour. That the lower surface was prone to flooding is still apparent at the present day by recent erosion of a face above the 5.0 m contour along part of the northern side of Camblesforth (Plate 32).

Plate 32. An erosional scarp of about 3.0 m above the 5 Metre Surface near Camela Lane on the northern side of Camblesforth [SE 648 263]. The steepness of the slope suggests recent erosion. (Photograph W.A. Fairburn)

Based on lithological mapping shown on the Selby 1: 50 000 sheet (British Geological Survey, 2008), the 5 Metre Surface between Wistow and Selby is underlain by either the Thorganby Clay Member of the Hemingbrough Glaciolacustrine Formation or the Breighton Sand Formation. South of Selby, on the Goole 1: 50 000 sheet (British Geological Survey, 1971), these formations are probably still present but are not distinguished from the clays, silts and sands of the '25 Foot Drifts'. That there is a lateral transition from the Breighton Sand Formation to the Thorganby Clay Member, near Wistow, on the 5 Metre Surface, reflects the unconformity between the two units and incision of the younger Breighton Sand Formation into the laminated clays.

7.12. Alluvial Fans

The alluvial fans shown on Figure 15 and part of the northwest segment of Figure 16 were formerly mapped on the older Selby 1: 50 000 sheet (British Geological Survey, 1973b) as a combination of 'Older Littoral Sand and Gravel' (Gaunt, 1981) and 'Head'. 'Head' has been defined by French (2007), as a product of mass-wastage or solifluction: Reid (1887) has also described 'coombe rock' or chalk rubble of dry Chalkland valleys as a form of Head. Updated mapping by the British Geological Survey (2008) on the

Selby 1: 50 000 sheet has replaced Older Littoral Sand and Gravel and part of the Head outcrop by the Pocklington Gravel Formation – a glaciofluvial fan deposit. This mapping however has not differentiated between the low-level and high-level fans (that comprise the Pocklington Gravel Formation) mapped by Fairburn (2009) in the Pocklington area (terms that have now been replaced by younger and older alluvial fans on Figure 15). Ford *et al*. (2008) suggest that part of the Pocklington Gravel Formation may have been derived from older glaciofluvial deposits. Also at deviance with mapping by Fairburn (2009) and on Figure 15 is the mapping by the British Geological Survey (1983b) on the York 1: 50 000 sheet. This shows extensive areas of 'boulder clay' between Yapham and Bishop Wilton and not two ages of alluvial fans[6].

The recognition of two ages of fluvial deposition that comprise the Pocklington Gravel Formation is based not only on the separation of the two units by lateral erosional contacts but by their geographical distribution and the differences produced by imprinted shorelines resulting from two recognisable and temporally separated glacial episodes that impounded glacial lakes.

7.12.1. Older Alluvial Fans

The superficial deposits that form the Older Alluvial Fans occur in two distinct regions controlled by the intensities of erosion that occurred during later fluvial events that deposited the younger fans. To the north of Market Weighton, in an area with more extensive Younger Alluvial Fans, the older fans have suffered extensive dissection and are now typically represented by the erosional remnants of a once extensive, planar sloping surface now separated by younger drainage channels. Invariably these residuals form discrete landforms, now detached from source valleys in the Wolds and are each bounded by lateral erosion contacts with the younger fans. They also have distal depositional terminals near the 30 m or 35 m contours; locations that approximate to the 33 m terrace mapped on Figure 15. The largest of these remnants extends from a ridge, at an elevation of about 55 m above OD southeast of Meltonby, down to terminal points between Rowland Hill and Yapham Grange (Fig. 15) on a slope of about 10 m / km. This slope, to the southwest, has been modified northeast of Rowland Hill by strandlines and erosion surfaces, which are included in this text as part of the 33 Metre

6 Mapping on the York 1: 50 000 sheet, which is of some antiquity, has failed to recognise these gravel deposits as alluvial fans with a pebble assemblage derived solely from valleys cut into the Jurassic and Chalk formation above Bishop Wilton. Benches of till, on the higher ground in the Wolds, is more clayey and contains exotic faceted erratics of Carboniferous sandstone and igneous rocks.

Surface. Other remnants of the older fans occur at High Belthorpe, North Hill and a small ridge southeast of Bishop Wilton (Fig. 15). In places, such as southeast of Yapham Grange, the terminal slope of the older fans may extend further as a less-prominent low-profile mound below the 35 m contour. Such extensions could result from sub-lacustrine deposition below the 33 m above OD lake level.

Although the Older Alluvial Fans north of Market Weighton appear to form low-angle sheets of colluvium on ridges with post-depositional erosional edges, underlain and bordered by near-surface Triassic marl, it is likely that these features originated as alluvial fans, following an older glacial period, that were later terraced by the most recent impounding of Lake Humber following the Last Glacial Maximum (LGM). This conclusion is based mainly on a correlation with the better defined fans to the south.

To the south and southeast of Market Weighton, the Older Alluvial Fans form a near continuous mantle on the western slope of the Wolds to beyond North Ferriby (Fig. 16). Here they have an upper boundary, or washing limit, controlled by the 52 Metre Strandline, which descends from Goodmanham to about 40 m above OD (isostatic adjustment) just south of Brantingham, to apparently coincide with high-level Lake Humber formed during the LGM. The fans are clearly terraced at intervals between 40 m and 10 m above OD. Unlike the fans north of Market Weighton, the terracing has been depicted, on Figure 16, as erosional surfaces similar to those on the York Moraine (Fig. 14). As with the fans north of Market Weighton, the fans south of Market Weighton are detached from source areas, in the Wolds, by later valley erosion that was not accompanied by significant younger alluvial fan formation.

To the east of North Ferriby, the Older Alluvial Fans are laterally replaced by Skipsea Till (Fig. 16) which is itself terraced by shorelines of Lake Humber. An important conclusion from this part of the mapping is that the Older Alluvial Fans preceded the Skipsea Till and that the Skipsea Till is older than the Late Devensian lacustrine terracing.

Another factor resulting from the recognition of the Older Alluvial Fans, on the western slope of the Wolds, is that similar flint-rich gravels also occur below the 10 Metre Surface above the mammalian deposits at Beilsbeck Farm, near North Cliffe (Fig. 47), and in the gravel pits below the 5 Metre Surface that now form the wetlands to the east of North Cave (Plate 33). These data suggest the possibility that all these gravel deposits originated during the same time interval, with the deposits on the wetlands being reworked at the 10 m level of Lake Humber and redeposited by fluvial events prior to the MIS 2 glaciation (see sections 9 and 13).

Plate 33. Interbedded sand and flint gravel, with numerous ice wedges, exposed in a gravel pit at the North Cave Wetlands Nature Reserve. The horizontal or gently inclined bedding may represent a reworked progradational sequence originally deposited during MIS 8 deglaciation. (Photograph M Bateman)

Although bedrock is not well exposed, the Older Alluvial Fans were deposited over a sequence of Mesozoic strata ranging upwards from Triassic marl near Pocklington to Cretaceous Chalk at Hessle. Immediately south of Goodmanham Dale however and possibly extending as far south as Sancton, in a region where the fans display a dominantly sandy facies, they are underlain by, or marginal to, up to 5.0 m of pale green clay. The clay which forms a terrace at over 45 m above OD near Market Weighton School [SE 884 422], where the clay is laminated, was also encountered by shallow auger drilling at Mask Hall [SE 893 402] at about 44 m above OD under 0.85 m of alluvial sand. This clay terrace probably represents a marginal remnant of a once more extensive deposit of lacustrine clays, which may have extended across the Vale of York to include laminated clays, at a similar elevation, on Brayton Barff (Fig. 7). Later events, including the discharge of sediment from Goodmanham Dale and elsewhere in the Wolds, would have either obliterated or buried this terrace, leaving a few localised remnants. Although firm evidence is lacking, the laminated clays, where preserved, are probably everywhere underlain by pre-Quaternary deposits.

7.12.2. Younger Alluvial Fans

The Younger Alluvial Fans occupy most of the mapped area between North Hill and Market Weighton, from the edge of the Triassic – Jurassic escarpment down towards the 20 Metre Surface (Figs 15 and 16). Generally the fans form a coalescing surface that slopes southwesterly (or more rarely southerly) from about 30 m / km down to less than 5 m / km. In many places, such as Bishop Wilton Beck and Millington Beck, the alluvial source of the fans can be traced back to dry valleys in the Wolds.

Compared with the older fans, the younger fans are not terraced at *c*. 33 m and generally terminate at or just below the 20 m contour on the 20 Metre Surface: an exception is in the region of Gowthorpe where terminations are closer to 25 m. Terminations at the *c*. 20 m shoreline generally form distinct margins (i.e. changes in elevation and/or sediment) in most places, such as southeast of Full Sutton or southwest of Bolton (Fig. 15). In other locations, the boundary between the 20 Metre Surface and the younger fans is almost imperceptible and the contact has been extrapolated between known points. This is particularly evident southeast of Barmby Moor, where the boundary is adjacent to earth works marginal to the A74. Such fluctuations in the nature of this boundary must reflect the balance between sediment supply and the dynamics of the lacustrine shoreline.

The 25 m above OD terminals are most conspicuous between Gowthorpe, High Belthorpe and Bolton (Fig. 15). Here the clastic sediments of the fans are replaced quite sharply by clayey alluvium, devoid of clastic sediment, which forms outwash fans that grade down towards the 20 m contour. Here again, the contact with the 20 Metre Surface may be rather elusive. The clayey outwash fans seem to form shallow drainage channels between discrete lobes of the younger fans, as can be seen between Gowthorpe and Highfield Lane, and between Gowthorpe and High Belthorpe (Fig. 15). It seems likely that the outwash fans originated by cessation of clastic sediment supply to the younger fans at the 25 m above OD shoreline of Lake Humber, but with some drainage or dewatering from the fans continuing towards a lower lake level.

The lithology of the younger fans is comparable to the older fans with most information coming from temporary building excavations, because of the scarcity of natural sections through the deposit. Ford *et al*. (2008) defined a type-section for the Pocklington Gravel Formation in a temporary exposure at Pocklington [SE 8042 4801], where 1.0 m of light-coloured, matrix-supported, sandy gravel, dominated by sub-rounded chalk and sub-angular flint fragments, but also with abundant clasts of ironstone and Jurassic limestone (Plate 34) showed a gradational lower contact with weathered Triassic sandstone. A similar lithological sequence was noted in gravel pits, during the landform mapping, to the west of Market Weighton [SE 865 418]. At this location, the 2.0 m – 3.0 m section in cross-bedded gravels (Plate 35), as explained earlier in section 7.7.3., is either in the Younger Alluvial Fans or reworked Older Alluvial Fans on the 25 m terrace.

Plate 34. Pocklington Gravel Formation (Younger Alluvial Fans) exposed in house foundation trenches at Pocklington. The gravel comprises mainly chalk and flint with abundant ironstone and Jurassic limestone. The dark brown soil is partly decalcified with mainly flint fragments. From Ford *et al*., 2008.

Plate 35. Section in gravel pit, some 2.0 m – 3.0 m high, west of Market Weighton [SE 865 418]. The gravel which is mainly composed of chalk and flint exhibited crude cross-bedding (now only visible in the top right-hand corner of the photograph) prior to collapse of the face. (Photograph W.A. Fairburn)

7.12.3. Discussion

The Pocklington Gravel Formation, comprising both the Older Alluvial Fans and the Younger Alluvial fans that extend westwards from incised valleys in the Wolds, must have originated mainly by erosion and fluvial transport of frost-shattered Chalk locked in place by permafrost (French, 2007; see also Hitchens, 2009). The distinctive shape of many of the Chalk clasts has resulted from the platy brecciation of bedded Chalk bedrock, that can occur at depths up to 5.0 m – 10.0 m, under periglacial conditions (French, 2007, figure 13.2). Degradation of the permafrost by fluvial erosion would have been caused, under ameliorating weather conditions, by meltwater from local ice or snow accumulations (Price *et al*., 2008 from Waltham *et al*., 1997; Bateman *et al*., 2000).

An interesting outcome, derived from the mapping of the alluvial fans, is the inference that the comparative areal extent of the two sets of fans can be taken as a proxy for the comparative magnitude of the two cold periods that produced them. The younger fans, which are not terraced and do not appear to extend further south than Market Weighton, contrast with the older set of fans, which reach at least as far south as Hessle. This difference in distribution may therefore reflect the extent of periglacial or permafrost conditions that caused fracturing of Chalk formation that produced the gravels. Along with the contrast in isostatic response between the southerly tilted 52 Metre Strandline and the more recent horizontal terracing of Lake Humber, this suggests that the last glaciation, in the Vale of York was not so prolonged or intense as the penultimate or pre-Devensian glaciation.

7.13. Isostasy

Teller (2001) has suggested that most large pro-glacial lakes, in glacial regions, have a number of beaches around their margins, each of which reflects the outline of a former stand of the lake. If differential isostatic rebound occurs after formation of these beaches, then their elevations will rise towards the area of maximum glacio-isostatic depression. In the case of Lake Agassiz, North America's largest pro-glacial lake that developed on the southwestern margin of the Laurentide Ice Sheet, differential isostatic rebound has resulted in over 50 mapped shorelines, to the lake, rising in elevations to the northeast where the Laurentide ice was thickest (Teller, 2001; fig. 11).

Gaunt (1981) invoked differential isostatic rebound to account for (a) the northerly increase in elevation of high-level sand and gravel (i.e. the 100 Foot Strandline deposits of Edwards, 1937) up to 33 m above OD and (b) the northerly increase in elevation of low-level laminated clay and sand, underlying the surface of the plain of the Vale of York from 10 m above OD at Doncaster to 14 m above OD at York (Gaunt, 1974). Such a model for isostatic control of shorelines in the lacustrine basin of the Vale of York with a southerly outlet and southerly retreating, regressive shorelines, should result in a schematic model for shoreline development similar to that proposed by Teller (2001; Fig. 26) for a lacustrine basin undergoing differential isostatic rebound. The Teller (2001) model indicates convergence of the shorelines to the point of outlet, a system that cannot be demonstrated in the Vale of York.

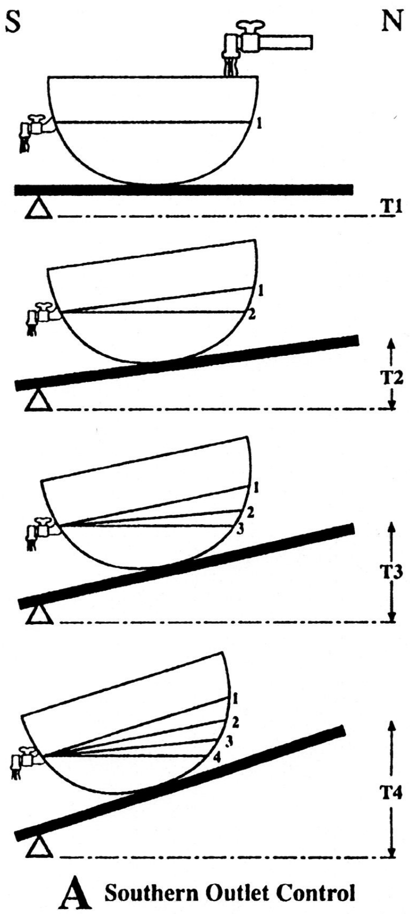


Figure 26. Schematic model of shorelines developing in a basin undergoing differential isostatic rebound with a southern outlet and shorelines everywhere regressing across the basin floor (from Teller, 2001).

Much of the evidence that has been used to support isostatic rebound in the Vale of York can now be regarded as unreliable. Murton & Murton (2011), in a review of Late Pleistocene glacial lakes, show that most of the 100 Foot Strandline gravels mapped by Edwards (1937), lie on or about the 30 m contour (Murton & Murton, 2011, fig. 15B) with no obvious southerly decline in elevation of this deposit. Some gravels at a lower level (i.e. below 30 m above OD), mapped during this research, have been included in this text as part of either the 20 m or 15 m terrace (Figs 17 and 20). Gravel deposits mapped at Mill Mound and Severs Howe, on the York Moraine (Fig. 14), and at Everthorpe (Fig. 19) and Mill Hill near Elloughton (Fig. 48), on the western flanks of the Wolds (Fig. 16), can be correlated with the 100 Foot Strandline deposits, as indicated on Figures 14 and 16, provide further evidence that the 100 Foot Strandline deposits occur at a fairly constant elevation a little over 30 m above OD.

Publication of the newer Selby 1: 50 000 sheet (British Geological Survey, 2008) and its brief explanation (Ford *et al*., 2008) now very clearly illustrate that the sand component of the '25 Foot Drifts' (now designated as the Breighton Sand Formation) is a widespread fluvial sand formation, up to 4.0 m in thickness, with an erosional lower contact (Ford *et al*. 2008). This indicates that much of the surface of the Vale of York, covered by the Selby 1: 50 000 geological sheet (British Geological Survey, 2008), can be accepted as a flooding surface and not as a lacustrine surface. Thus the northerly rise of the plain of the Vale of York can now be attributed to fluvial deposition and not to isostatic adjustment. A more detailed account of the Holocene-dated Breighton Sand Formation (given later in this text – section 10.2.) indicates that it was deposited in southerly sloping bedforms at angles comparable to modern drainage.

Credence to the effect of isostasy on landform mapping between York and the Escrick Moraine was given by Fairburn (2009) because he detected gradual southerly lowering in elevation of depositional surfaces in this region of about 5.0 m. A similar rise of these depositional surfaces to the north of the York Moraine was however attributed to differences in lake levels; a view that is maintained. The elevation change south of the York Moraine was based on the following:

- the strong persuasion, from earlier literature (e.g. Gaunt, 1981), that elevation changes in the plain of the Vale of York were caused by isostatic rebound were correct;
- lack of knowledge of the extent and relationship between the fluvial sands (now the Breighton Sand Formation) and the laminated clays, which then comprised the '25 Foot Drifts' until this was clarified by the explanatory text for the Selby (2008) 1: 50 000 sheet (Ford *et al*., 2008);
- failure to recognise a distinct 15 m terrace below the 20 m terrace, which led to the erroneous suggestion of a southerly tilt to the 20 m terrace;
- failure to recognise a 25 m terrace, which suggested that the 33 m shoreline, in places, formed a wide littoral zone extending down to *c*. 20 m above OD, e.g. between Bilbrough and Appleton Roebuck (Fairburn, 2009).

It was not until a later article, based on the relationship of alluvial fans forming terminations against lacustrine terracing near Pocklington was published (Fairburn, 2011; Fig. 15) that conclusions were reached, by the author, that the lacustrine shorelines of Lake Humber do not provide any evidence for isostatic change and that if any change occurred it must have preceded the impounding of Lake Humber. Later mapping to the south of Pocklington, between North Cliffe and Hessle (Fig. 16), not only confirmed the presence of a 15 m above OD shoreline for Lake Humber but that all the terracing was horizontal or as near horizontal as could be judged from on the 1: 25 000 mapping.

The only terrace on the western side of the Wolds that appears to have been influenced by isostatic rebound is the 52 m strandline of Penny (1974). As described earlier (section 7.1.), the 52 m strandline, which marks the 'washing limit' of the Older Alluvial Fans, gradually falls in elevation from 52 m above OD at Goodmanham to 40 m above OD at Brantingham (Fig. 16), where it eventually merges with the 40 m terrace that has been proposed as the maximum level of Lake Humber. North of Market Weighton, the 52 m strandline remains at this elevation but its location becomes increasingly obscure near Pocklington where the Older Alluvial Fans have been extensively eroded by the younger fans.

Similarly to using the smaller areal extent of the Younger Alluvial Fans compared to that of the Older Alluvial Fans as a proxy to compare the relative magnitude of the MIS 2 glaciation with that of an older glaciation so can differences in the glacio-isostatic response of the two events be used in the same manner. Thus the lack of an isostatic response for the MIS 2 glaciation, based on constant elevations for the Lake Humber shorelines, suggest that this younger glaciation in the Vale of York was either of shorter duration or of less magnitude (or both) than the glaciation during which the 52 m strandline and the Older Alluvial Fans originated.

The current elevation of the Sewerby raised beach over a wave-cut platform at approximately 2.0 m above OD (Catt, 2007) and for a possible raised wave-cut platform further north, at approximately 5.0 m above OD, on headlands of Magnesian Limestone below Late Devensian till, at Whitburn Bay, County Durham (see fig. 21, Davies *et al*., 2013a) is also suggestive that there is a lack of significant or recognisable glacio-isostatic response since the Ipswichian.

7.14. Aire and Calder River Terraces

The only large river to enter directly into the Vale of York from a Pennine valley, within the mapped area, is the River Aire, which discharges into the Vale near Ferrybridge from a gorge eroded into the Permian Magnesian Limestone. While it might be tempting to assume that glacial deposits, temporarily exposed in a pipeline trench crossing the Snaith Ridge south of Kellington and to the east of the Aire gap

(Fig. 8), may have resulted from glaciation in the Aire Valley, it was concluded by Gaunt (1976b) that these deposits of 'Pennine Boulder Clay' probably originated from a land-based glacier that crossed the Pennines through the Stainmore Gap, as did ice in the younger MIS 2 glaciation, recorded by till at Ferrybridge (Bateman *et al*., 2008). Gaunt (1976b) based this conclusion on the lithology of erratics found in the till, though he did consider that glacial channel deposits on the Snaith Ridge (see section 3.3.1.2.) could be related to the Aire gap.

In a description of the 'Geology of the Country around Wakefield' Edwards *et al*. (1940) subdivided the Pleistocene deposits into two categories: the 'earlier drifts' comprising boulder clay and gravels that cap many of the hilltops in the region and the 'later drifts' that include river terraces, in valley bottoms, and strandline gravels (i.e. the 100 Foot Strandline gravels) on the western side of the Vale of York (Edwards *et al*., 1940, plate VIII). Generally two sets of river terraces were recognised, a higher (Second) terrace whose surface lies about 20 ft (6.1 m) above the river floodplain and a lower (First) terrace between 5.0 ft – 10 ft (1.5 m – 3.0 m) above the river floodplain. While both levels resemble river terraces topographically, they were considered to show 'evidence of deposition in part under deltaic conditions'. The mapping of Edwards *et al*. (1940, plate VIII) also depicts additional river terraces between the higher and lower terraces. In a later description of the Aire and Calder terraces, Lake (1999), confirms the First and Second terracing at 3.0 m and 6.0 m above river alluvium, both of which are shown on the Wakefield 1:50 000 geological sheet (British Geological Survey, 1998) along with river terraces not assigned to these two levels (i.e. undifferentiated). In a schematic cross-section of the river terraces, Lake (1999, fig. 28) indicates the First Terrace at about 10 m above OD and the Second Terrace at nearly 15 m above OD. The section also shows river alluvium incised in a channel below OD with the base of the First Terrace extending below the surface of river alluvium. Lake (1999) reported that some of the undifferentiated terracing occurs at *c*. 25 m above OD, while higher terracing, at or above 30 m above OD, is shown on the Wakefield geological sheet (British Geological Survey, 1998), in the Aire valley near Hunslet.

Provided the Aire valley was in existence, prior to the LGM, as suggested by Gaunt (1976b), the river terracing must have been influenced by the water levels of Lake Humber. Because of this, a more realistic assessment can be made of terracing in the Aire and Calder valleys based on their elevation above OD (rather than height above river alluvium) taken in relationship with shoreline terracing of Lake Humber. Such a relationship is outlined below:

10 Metre Lake Terrace This terrace, which would be the lowest terrace in the Aire valley above modern alluvium, probably pinches-out against the Aire floodplain near Ferrybridge, so its relationship to the Vale of York 10 m terrace is uncertain (Fig. 17).

15 Metre Lake Terrace The 15 metre lake terrace is well-defined on both banks of the River Aire at Ferrybridge and appears to extend up the valley (beyond the western edge of Figure 17) as the First Terrace of the River Aire to beyond the Calder valley confluence. It therefore lies between the modern floodplain and about 15 m above OD (i.e. the 50 ft contour shown by Edwards *et al*., 1940, on plate VIII). Any mapping of the First Terrace above 15 m above OD would not be realistic. This suggests that the First Terrace originated as depositional fill in the Aire valley when the level of Lake Humber stood at *c*. 15 m above OD. Any subsequent lowering of erosional base level in the Aire valley, in response to a fall in Lake Humber, would produce an erosional, but offlapping contact, between the First Terrace and the newly evolved floodplain (see Lake, 1999, fig. 28).

20 Metre Lake Terrace As this marks a prominent stand of Lake Humber it should be represented by a significant terrace in the Aire valley. This is not clearly apparent, either on the mapping of Edwards *et al*. (1940, plate VIII) or on the Wakefield 1:50 000 geological sheet (British Geological Survey, 1998), largely due to poor topographic detail on both these maps. Nevertheless, it must be represented by terracing between the 50 ft and 100 ft contours (15.2 m and 30.5 m) shown by Edwards *et al*. (1940, plate VIII) that is either incorrectly labelled, as the First or Second Terrace, or is an undifferentiated terrace. Deposition of the 20 m river terrace, which may well be represented by the second Terrace near Methley and Mickletown (Edwards *et al*., 1940, plate VIII), as with other terracing in the Aire valley, would be a prograding surface (note the earlier comment from Edwards, 1940, on deltaic deposition) that could have remained as a discrete landform, above modern alluvium, or it could have coalesced down valley with the First Terrace at 15 m above OD.

33 Metre Lake Terrace Mapping of the Second Terrace in the Aire valley is represented by terracing both above and below the 100 ft (30.5 m) contour (Edwards *et al*., 1940), notably near Hunslet. Again, topographic control is inadequate to assign this terracing either to 'high-level' Lake Humber at *c.* 33 m above OD or to the 100 Foot Strandline gravels mapped near Ferrybridge by Edwards (1937) and Bateman *et al*. (2008).

7.15. Dry River Valleys

Dry river valleys of the Yorkshire Wolds form a dendritic pattern of steep-sided, deeply incised valleys on the easterly dip-slope of the Chalk formations with wider valleys on the steeper westerly facing escarpment overlooking the Vale of York (Price *et al*., 2008). Generally the dry valleys have a flat bottom filled with gravelly clay or silty clay, which is a solifluction deposit referred to as 'Head' by the British Geological Survey e.g. on the Selby 1:50 000 geological sheet (British Geological Survey, 2008) or as 'coombe rock' (Reid, 1887; Kerney, 1963). Thin sandy clay or silt is also present on the flanks of the valleys (Price *et al*., 2008). As already explained (section 7.12.), the dry valleys evolved by rapid fluvial incision, under ameliorating weather conditions, of frost-shattered Chalk that had developed platy brecciation under periglacial conditions (French, 2007; Hitchens, 2009; Waltham *et al*., 1997) or as explained by Dakyns *et al*. (1886) as the disintegration of the beds *in situ*. Typically, the valleys of the Yorkshire Wolds are now dry following loss of permafrost and the chalk regaining a permeable and porous character (French, 2007). Similar descriptions for the origin of steep-sided dry valleys, or coombes, and the deposition of fans from the chalk escarpment of Kent have been made by Kerney (1963) and Kerney *et al*. (1965) who radio-carbon dated marsh deposits below the fans to 12.0 ka – 10.8 ka BP. Chalk and flint gravel, eroded from these now dry valleys of the Wolds by fluvial processes, has produced the alluvial fans described by Fairburn (2011, fig. 2 and on Figs 15 and 16), also referred to as glaciofluvial fan deposits by Ford *et al*. (2008).

Of major significance, for this text, has been the recognition, during mapping, of two sets of temporally distinct alluvial fans that have originated from the dry valleys: an older set of fans mapped from North Hill (Fig. 15) to east of North Ferriby (Fig. 16) that include the well-documented gravel deposits at Everthorpe (e.g. Dakyns *et al*., 1886) and Elloughton (e.g. Lamplugh, 1887) and a younger set of fans restricted to the Wolds north of Market Weighton (Figs 15 and 16). These alluvial fans probably represent periglacial cold climate conditions associated with two glacial episodes: the MIS 2 glaciation in the Vale of York and the older tills on the Pennines and Wolds equivalent to the 'Pennine Boulder Clay' of Gaunt (1976b).

The only other dry valley of significance, noted during the recent mapping, which has eroded the post-glacial sediments in the Vale of York, is a sand-filled channel that has detached the 20 m terrace surrounding Knapton from the western edge of the York Moraine near Severs Howe (Figs 14 and 30). Meltwater from this channel may have drained into Hob Moor through a small dry valley in the York Moraine south of Acomb [SE 565 501].

7.16. Pre-Holocene Drainage

While regional factors, such as the dynamics of the North Sea ice lobe, may have controlled the level of Lake Humber, the deposition and erosion of fluvioglacial and lacustrine deposits in the Vale of York were controlled by a number of local factors. These include:

- transgressive shoreline erosion on the York Moraine;
- the damming effect of the York and Escrick moraines;
- constricted meltwater channels through the York Moraine and Lincoln Gap;
- lateral erosion along the frontal edges of the York and Escrick moraines;
- drainage in the underfit valleys.

7.16.1. Transgressive Shoreline Erosion

It is proposed in this text (*Sand screes*) that segregation of the till on the York Moraine into three principal components (i.e. sand, sand and gravel, and boulders) could only have occurred during high stands of Lake Humber, when lake levels coincided with crestal regions of the moraine, as at Severs Howe and Mill Mound during the 33 m above OD lake level. Such a situation would allow transgressive beach development across the top of the moraine with separated sand being washed down the southern face of the moraine to produce a steep-face, sand scree or scree-apron delta in a sub-aqueous environment. This down-slope gravity driven system would not preclude the transfer of gravel or even boulders to the delta slope, although gravel does seem to be mainly confined to crestal regions. As shown on Figure 27 (from Nemec, 1990b) the steep-face delta would be a layered structure, a feature that has been noted, on the York Moraine, in a small excavation under Severs Howe (Plate 7). Subsequent reworking and terracing of the delta, at lower lake levels, could well produce instability and slumping on the delta slope as indicated on Figure 27.

7.16.1.1. Sand Screes

Sand screes (or the sand mantle) that have been generated on the slopes of the York Moraine are similar in appearance to the scree-cone or steep-face coarse-grained deltas of Nemec (1990a and 1990b). These screes, which are restricted to the York Moraine, while not producing distinctive landforms nevertheless have had a profound effect on the landscape of the moraine. The screes originated on the York Moraine as a washing effect or degradation wherever the crest of the moraine coincided with a prolonged stand of Lake Humber (e.g. the 33 m strandline). As a consequence of this, the major development of the sand screes are probably restricted to the slopes of the moraine below Mill Mound, Severs Howe and Bilbrough where thick sand deposits (perhaps 2.0 m plus) have been identified (Plates 7, 8 and 9). At these locations, the development and progradation of sand screes would have little or no subaerial exposure such as a distributary plain, and groups of screes could coalesce, like scree-cones, to form slope aprons (Nemec, 1990b). In a schematic diagram to illustrate the development of the underwater deltas, Nemec (1990b, fig.2) indicates that the slope of the delta can be quite steep (exceeding 27°) but as the delta aggrades the concave slope decreases. Ravinement at the spillpoint on the crest of the moraine would be minimal, as meltwater levels on both sides of the moraine would have equalised, and may only be present between the highest points at Severs Howe.

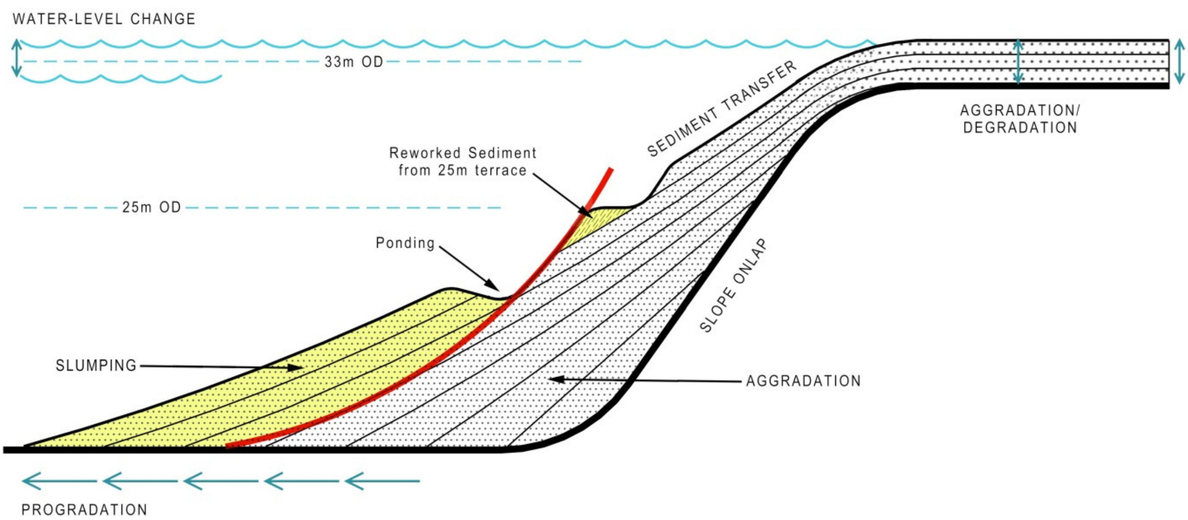


Figure 27. Schematic section (not to scale) showing the development of the sand scree on the southern slopes of the York Moraine below Mill Mound. Note slumping of the scree below 20 m OD and reworking of the primary sand scree at the 25 m OD shoreline. The section is adapted from Nemec (1990b, fig. 2).

Figure 27 also illustrates a multi-layered on-lapping fabric for the delta that would be similar to a sand scree. On the York Moraine sections through the sand apron are restricted to a roadcut east of Bilbrough [SE 538 467] and an excavation on the slope below Severs Howe [SE 583 521]. At both these locations (not now visible) low-angle crude bedding in the sand was noted by the alignment of intercalated boulders, roughly parallel to the hill slope. No dip slopes were recorded however. Near the Severs Howe section a bedding dip of 16° was measured from sand layers exposed in a shallow pit excavated for an OSL sand sample (Plate 7). Instability of the sand slope below Mill Mound, to the west of York University (Plates 23a and 23b) is particularly evident on LIDAR imagery (Fig. 33b) where slumping has occurred below the 20 m terrace and below an associated slump scar. An example of a slump scar on a scree-cone delta is shown on Figure 24 (see also Nemec 1990b, fig. 1A).

An additional effect on the York Moraine sand scree could be the redistribution of sand from the primary apron (i.e. from the 33 m terrace) by lower lake levels and hence adding an offlapping component to the illustration in Figure 27.

It should also be noted that sand screes are not a part of the Escrick Moraine, except perhaps for a small area near High Catton, where gravel deposits have been worked (see Section 3.3.2.2.), as the rest of the moraine lies below 33 m above OD.

7.16.2. Damming Effects

It has been suggested by Russell *et al*. (2006) that retreat of a glacier from a prominent moraine may produce a trench or topographic low between the glacier and the moraine. This topographic low could be occupied by a pro-glacial lake, or a fluvial system that would flow parallel to the glacier margin before breaking through the containing moraine. Such a fluvial system could well have evolved following retreat of the Vale of York glacier from the Escrick Moraine. In this case, break-through would have occurred at locations now occupied by the Ouse and Derwent rivers. While firm evidence for this sequence of events may be lacking, there is good erosional evidence for lateral movement of water along the northerly face of the moraine during later events related to drainage of Lake Humber as shown in Figure 31. The moraine–confined backwater would act as a major control on sedimentation with much finer grained deposits downstream of the backwater zone (Russell *et al*., 2006, fig.10). However if cataclysmic outwash flooding (jökulhlaups) occurred in the Vale of York, then this could perhaps have only taken place during primary break-through of the moraines.

Events similar to these could also have occurred following retreat of the glacier to the north of the York Moraine: though this was a more formidable obstacle. Break-through of this moraine would have occurred through gaps at York and along Healaugh Beck (Fig. 14). These gaps do not appear to be random, as they are located above buried valleys eroded into the underlying Triassic Sherwood Sandstone. At York, this buried valley is some 35 ft (10.7 m) below OD (Melmore, 1935, fig. 5), while Healaugh Beck seems to lie above a buried valley at a low point in the Sherwood Sandstone escarpment between Hutton Wandesley and Bilbrough, where the Sherwood Sandstone rises to above 30 m above OD. A valley eroded to a depth of 80 ft (24.4 m) below OD, into the Trias, was also recorded by Melmore (1935, fig. 5) to the east of Wheldrake, where the River Derwent has broken-through the Escrick Moraine. However there remains uncertainty that the underlying valleys in the Trias were totally excavated at this time, or whether they were initiated as glacial channels.

Another factor of some relevance, to be considered in the break-through of the moraines, particularly the York Moraine, is whether the fan deltas (discussed in section 7.17.1.) could have formed at this time.

The York Moraine also seems to have acted as a barrier to the flow of water and sediment from the northern part of Lake Humber (i.e. Lake Alne) to the southern part of the lake, south of the York Moraine (see section 7.16.3.1.). This has resulted in depositional terraces forming in Lake Alne some 5.0 m higher than the rest of Lake Humber and the virtual silting-up of that part of Lake Humber following the fall in lake level below 20 m (see section 7.8.1.). Such a damming effect does not appear to have been caused by the Escrick Moraine. Any consideration for differences of lake levels in Lake Alne, as being an isostatic effect, have been discussed in section 7.13.

7.16.3. Meltwater Channels

As already discussed (sections 7.7., 7.8. and 7.9.) the York Moraine appears to have acted as a restrictive dam to the flow of meltwater, into the southern part of Lake Humber, and as a sediment trap; factors that have allowed the build-up of contemporaneous lacustrine depositional terraces to higher elevations north of the moraine than to the south. The elevation difference is most apparent for the 20 m, 15 m and 10 m terraces. That this has occurred can be partly attributed to inadequate drainage through the two lobes of the York Moraine, which was restricted to only two major and one minor outlet.

A further control on Lake levels in Lake Humber may have been exercised by meltwater escaping to Lake Fenland and possibly the River Waveney by way of the Lincoln Gap, Straw (1979, fig. 3.1) and Murton & Murton (2011, fig. 16), prior to retreat stages of the ice barrier in the Humber Gap (Figs 9 and 28) or elsewhere. Release of meltwater from Lake Fenland by the River Waveney is however contentious, as Clark *et al*. (2004, fig.1) consider this unlikely, based on a maximum level for Lake Humber at 30 m above OD. But their conclusion does not totally invalidate a discharge into the River Waveney for lake levels at 40 m or 42 m above OD. Such a high level for Lake Humber has been noted on the York Moraine, near Bilbrough, with a possible lower stand at *c*. 33 m above OD (Plate 18). This lower stand can therefore be considered as the first sign of shrinkage for Lake Humber resulting from a retreat of the North Sea ice barrier.

7.16.3.1. York Gap

Because of its location, the York Gap is of particular interest both geographically and historically and consists of two gorges cut through the York Moraine. One gorge is occupied by the present valley of the River Ouse and the other is an older abandoned channel of the River Ouse, possibly entering the Hob Moor/Askham Bog embayment by way of Holgate Beck before mainly flowing into Knavesmire (Fig. 14). Both Hob Moor and Knavesmire are underlain by laminated clays belonging to the Alne Glaciolacustrine Formation and the Elvington Glaciolacustrine Formation respectively (British Geological Survey, 2008) indicating erosion had occurred prior to formation of the 10 m terrace. Erosion through the gap, probably to enlarge an existing channel, must have commenced shortly after the level of Lake Humber had fallen below 30 m above OD as there is a conspicuous development of the 25 m terrace below Severs Howe (Figs 14 and 30). There is no noticeable fall in the elevation of the lacustrine terraces through the York Gap as erosional effects to accommodate this change have occurred by lateral erosion along the northern face of the moraine both to the east and west of the gap (section 7.16.4., Figs 29 and 30).

A subsidiary third channel, at 15 m above OD, may also have existed temporarily near Micklegate Bar, in York, as evidenced by a small drainage channel just south of Micklegate Bar, leading to the River Ouse (Fig. 14).

7.16.2.3. Healaugh Gap

Mapping adjacent to this channel, which cuts through the York Moraine at a low point in the Sherwood Sandstone escarpment between Healaugh and Bilbrough, clearly demonstrates the stepping down of depositional surfaces between the northern and southern parts of Lake Humber. It is most noticeable with respect to the boundaries of the 20 Metre Surface: the front edge of which steps down from 20 m to 15 m, while the elevation of the back-wall lowers from 25 m to 20 m (Fig. 14). Healaugh Beck passes southwards to Catterton Beck and into the lacustrine sub-basin drained by the Great Marsh (an underfit valley) near Bolton Percy (Fig. 14).

7.16.3.3. Askam Bog and Knavesmire

All the major roads, south of York, lie along narrow ridges that appear to be eroded remnants of a once more extensive 20 m terrace. This 20 m platform would have extended from Fulford across the Knavesmire embayment to the western side of the Hob Moor/Askham Bog embayment (Fig. 14). Following a fall in lake level to 15 m above OD and erosion of the 20 m terrace, the two embayments must have evolved as distinct outlets separated by a narrow ridge of gravel (one of the esker-ridges of Kendall & Wroot, 1924) that underlies the old Roman road from Copmanthorpe to York.

The Askham Bog embayment remained as a drainage outlet during deposition of the Alne Glaciolacustrine Formation, under the 10 m terrace, with an exit northeast of Copmanthorpe (Fig. 14) now used by the main rail line to the south. In contrast, the Knavesmire embayment persisted into the Holocene with the erosion of the 10 m terrace, now preserved as a narrow ledge around Knavesmire, to form a part of the Older River Terrace (5 m terrace). During this later phase, Knavesmire must have drained into the River Ouse between Middlethorpe and Bishopthorpe, with possible spillage southwards into the underfit valley of The Fleet river (Fig. 14).

7.16.3.4. Lincoln Gap

The validity of the Lincoln Gap, as a glacial overflow channel connecting Lake Humber to Lake Fenland is essentially based on the belief that stability of high-level Lake Humber, at *c*. 30 m, could only be achieved when it became equalized through the Lincoln Gap with Lake Fenland. Impounding of such a relatively stable system could only have occurred if North Sea ice blocked both the Humber Gap and the Wash Basin at the same time allowing southwest-flowing water into the Fen Basin (Clark *et al*., 2004). For Lake Fenland not to exist would require blockage of the Humber Gap and also the Lincoln Gap by ice or by some other means (Clark *et al*., 2004) and an overflow outlet from Lake Humber other than the Waveney Valley. The full extent of the combined ice-dammed lakes Pickering, Humber and Fenland is shown by Straw (1979, fig. 3.1) and Clark *et al*. (2004, fig.7). This combined body of water has an estimated volume of 257 km^3, a maximum depth of 31 m and an average depth of 20 m (Clark *et al*., 2004). Land form mapping, while indicating a possible maximum elevation of Lake Humber at *c*. 42 m above OD, as evidenced by terracing near Houghton Hall (section 7.3.3.) and by gravel deposits above the 40 Metre Surface on the York Moraine (Fig. 14), does confirm a more stable lake level at *c*. 33 m above OD, for a period that allowed erosion of the 100 Foot Strandline and therefore supports the suggestion of Lake Humber connecting through the Lincoln Gap with Lake Fenland.

Unlike Lake Humber, good evidence for shorelines in Lake Fenland to support its existence is lacking except for a narrow bench at 25 m – 32 m near Horbling and Bourne recorded by Harrod (1972; Fig. 9). Even in Lake Humber, shorelines and shoreline deposits, in the past, have either not been recognised or have been mis-interpreted and evidence for its existence is largely based on the recognition of the 100 Foot Strandline (Edwards, 1937) and the presence of laminated clays. Lack of evidence, for shorelines of Lake Fenland can partly be attributed to an extensive blanket of aeolian sand, which has been identified at many locations in Lincolnshire to the west of the Wolds, including the eastern side of the Ancholme valley in the southern part of Lake Humber (Bateman *et al*., 2000). Similarly laminated clays are not known to be widely distributed and have only been reported from near Somersham in the southern part of Lake Fenland (West, 1993; Murton & Murton, 2011) in an area referred to by West (1993) as Lake Sparks. West (1993) attributes the impounding of Lake Sparks to entry of ice into The Wash with sediment for laminated clays in the lake originating locally from the southern margin of the Fen. West's conclusions are however not a rebuttal for a Lake Fenland even though the laminated clays may not have been sourced through the Lincoln Gap.

In the Lincoln Gap, where former shorelines of Lake Fenland might be anticipated, mapping on the Lincoln 1:50 000 sheet (British Geological Survey, 1973a) has recognised 'undifferentiated river sand and gravel' above modern alluvium in the Witham valley. This deposit does however appear to be poorly controlled topographically as a cursory field examination by the writer revealed that above modern alluvium, at about 5.0 m above OD, there are two river terraces: a first terrace at 5.0 m - 10 m above OD and a second at 10 m – 15 m above OD (Fig. 28).

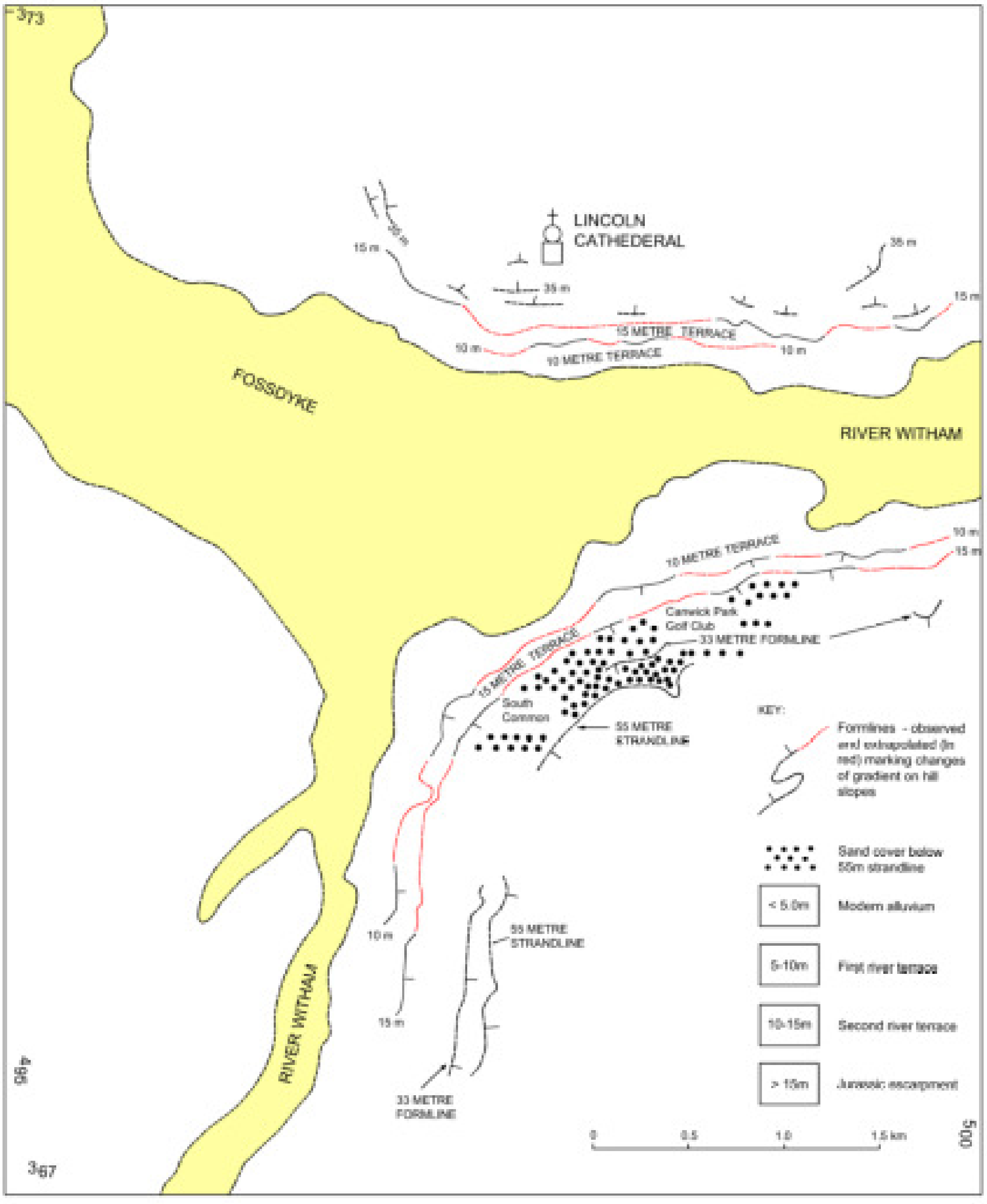


Figure 28. The Lincoln Gap showing river terracing and other linear landforms above the River Witham. Note the extensive sand deposits below a postulated 55 m strandline, near the Canwick Park Golf Club (mapping by W.A. Fairburn).

Elsewhere on both sides of the Witham valley near Lincoln there are abundant formlines, with minor terracing, up to elevations of 55 m above OD. Despite the significance that some of these features lie at or about 35 m above OD, they cannot be interpreted with confidence as shorelines, as lithological changes in the near horizontal underlying Jurassic sequences could produce similar geomorphological features.

7.16.4 Lateral Erosion Channels

Landform mapping, based on terracing, completed for this research project, has demonstrated that Lake Humber after being impounded by North Sea ice, following a retreat of the Vale of York glacier, drained in stages from a Stage 1 lake high of *c*. 42 m above OD to a Stage 8 lake low of 5.0 m above OD. This multi-stage decline model for Lake Humber is illustrated in Figure 52, which shows the main proposed stands for the lake. The conflicting two-stage decline model (Fig. 11) proposed by Gaunt (1976b) is included in Figure 52 for comparison.

Initial drainage of the high-level lake, to perhaps outlets into Lake Fenland, would have been fairly passive with erosional effects restricted to transgressive shorelines on crestal regions of the York Moraine at Bilbrough and Mill Hill / Severs Howe. As already outlined (section 3.3.2.1.), this shoreline washing segregated the till forming the moraine into boulder lags, banks of sand and gravel and a sand mantle formed by sand screes. At lower lake levels, with the progressive silting-up of the lake, there was lateral drainage to the north of both the York and Escrick moraines which resulted in channelling towards drainage outlets with significant fluvial erosion. As outlets through both moraines had probably been made by break-through, prior to the build-up of Lake Humber, further break-through of either moraine probably did not occur at this late stage so there was little likelihood of any outburst flooding (jökulhaups).

7.16.4.1. York Moraine

In contrast to much of the Vale of York, where shoreline terraces are essentially horizontal due to lack of any isostatic rebound, the 20 Metre Surface north of the York Moraine, which stands at 20 m – 25 m above OD near Holtby, declines westwards by about 5.0 m to a level between 15 m and 20 m above OD just east of York University (Figs 14 and 29). At this location, the elevation of the 20 Metre Surface north of the moraine is equal to that of the 20 Metre Surface south of the moraine (Fig. 29). Not surprisingly, evidence of gentle spillage marked by ponding and hummocky sand mounds (possibly exploited) are present [SE 629 508]. There is also evidence of a small drainage channel in this region [SE 637 509], which has largely been infilled for the construction of roads and buildings, with only a small area of remnant marsh remaining. West of the University, towards the York Gap, the front edge of the 20 Metre Surface remains at about 15 m above OD while the backwall rises slightly to above 20 m above OD.

The most plausible explanation for the changes in the elevation of the 20 Metre Surface and the offlapping 10 Metre Surface, is that both surfaces were subject to erosion by lateral drainage of meltwater towards spillpoints east of the University and through the York Gap, as lake levels fell below 20 m above OD.

Unlike the inclination of the 20 Metre Surface north of York University, that can only be demonstrated by carefully mapping the bounding formlines, an erosional channel near the back-wall of the 33 m terrace, near Long Marston (Fig. 14), can easily be observed visually. This channel, known south of the Long Marston Obelisk as the 'Glen', can be traced south-eastwards past Hutton Wandesley into the Healaugh Gap. Following heavy rain, the channel can take a moat-like form, due to linear ponding along part of its length. The channel probably developed by the flow of meltwater from the northwest into the Healaugh Gap against a scarp of Sherwood Sandstone, which here forms the back-wall of the 33 m terrace.

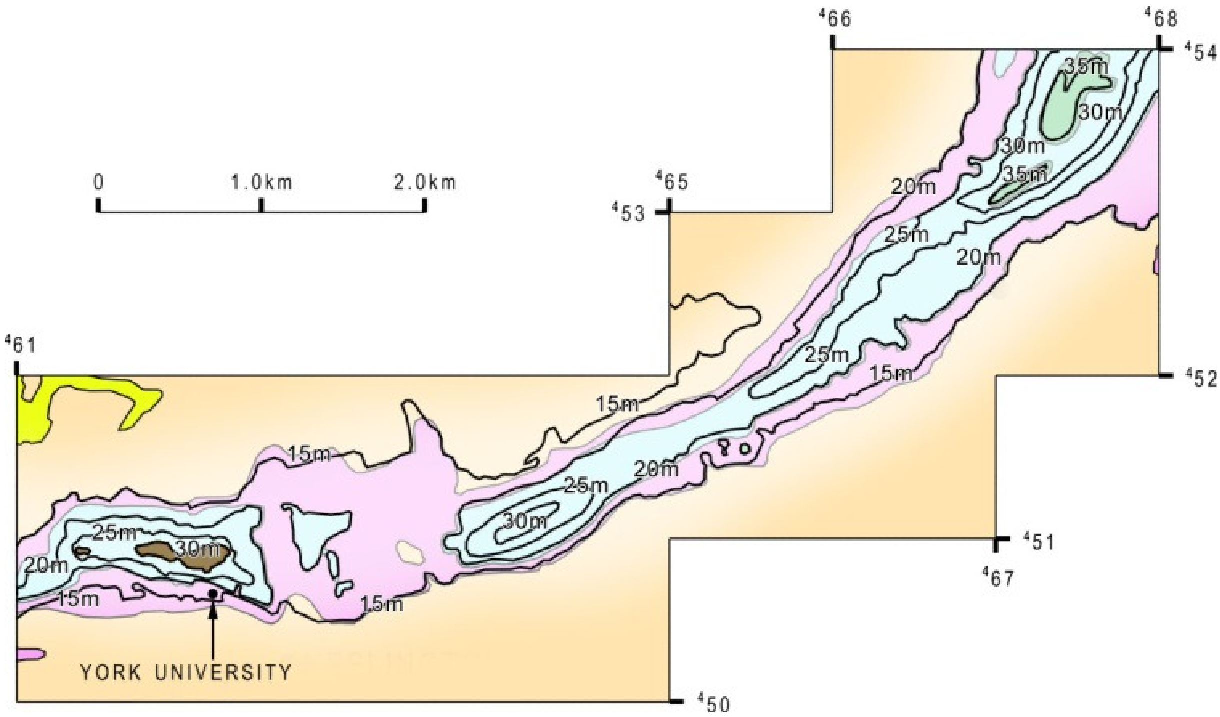


Figure 29. This diagram is part of Figure 14 and shows elevation changes of the 20 Metre Surface to the northeast of the York Gap. To the north of the York Moraine the level of the 20 Metre Surface falls from between 20 m to 25 m in the northeast down to between 15 m to 20 m near York University. South of the York Moraine the 20 Metre Surface is fairly constant between 15 m to 20 m and is at a level equal to the Surface north of the moraine just east of the University. The elevation changes are attributed to the erosional effects caused by lateral movement of meltwater to spillways through the York Gap and across the moraine east of the University. (Legend Fig. 30)

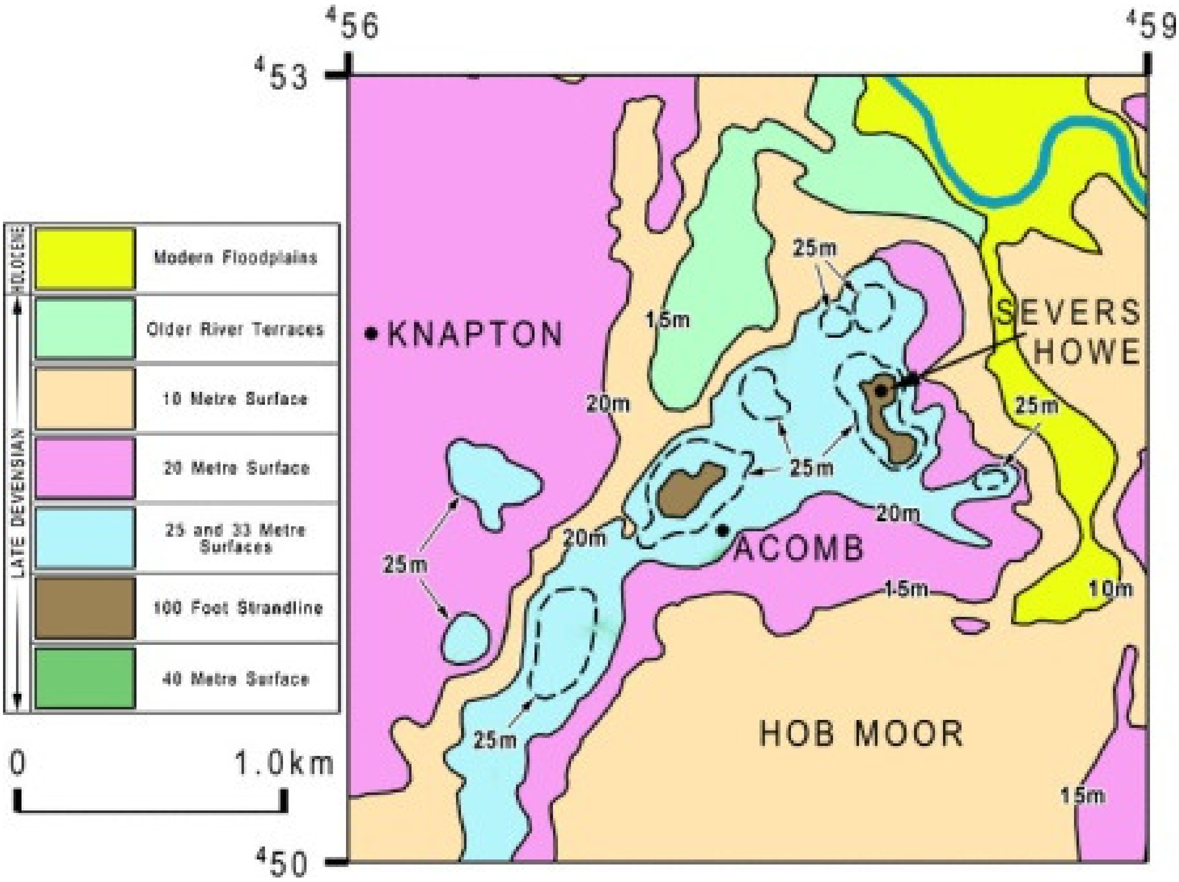


Figure 30. Drainage channel incised to below 15 m above OD west of Severs Howe that has detached part of the 20 Metre Surface around Knapton from the edge of the York Moraine. The channel was probably initiated following a fall in lake level below 20 m above OD to the north of the Moraine. Meltwater may have spilled from this channel into Hob Moor through a dry valley south of Acomb [SE 565 502]. The northern part of the channel was re-entered during incision of the Older River Terrace of the River Ouse.

A further visible channel, indicated by a shallow depression in the 20 Metre and 25/33 Metre Surfaces by the westerly projection of strandlines, was also mapped on the northern side of the York Moraine near Copmanthorpe, where meltwater drainage has flowed eastwards into Askham Bog (Fig. 14, SE 556 479).

A fourth lateral erosion channel, recognised along the edge of the York Moraine is the dry sand-filled valley that has detached part of the 20 Metre Surface around Knapton from the edge of the moraine (Fig. 30). This channel, which seems to have developed from an embayment between the York Moraine and the 20 Metre Surface west of Severs Howe, either originated from a fluvial channel late in the drainage of that part of the Vale of York or from a crevasse splay of the Ouse valley during incision of the Older River Terrace.

7.16.4.2. Escrick Moraine

The best example of a lateral erosional channel or trench in both plan and sectional format lies along the northern edge of the Escrick Moraine. This channel extends from the River Ouse in the west to the River Derwent in the east and is currently occupied by Stillingfleet Beck, Horsecourse Dike and extensions of Bridge Dike from Dunning Dike (Fig. 14). Mapping (Fig. 14) suggests that the trench evolved over at least two erosional stages following shrinkage of Lake Humber to below 15 m above OD and the emergence of the Escrick Moraine as a drainage barrier. These stages commenced with the separation of the 15 Metre Surface, a landform largely underlain by till, from the northern edge of the Escrick Moraine, to an area north of Stillingfleet and Escrick. This produced a shallow valley later infilled with sediments, including laminated clays, below the 10 Metre Surface. Later Holocene erosion of the valley established the equivalent of the 5 Metre Surface along Bridge Dike and Horsecourse Dike and eventual incision of Stillingfleet Beck.

The above account for the genesis of the Escrick channel, based on two-dimensional mapping, is however too simplistic when compared with sub-surface evidence obtained from a pipeline trench between Elvington and Wheldrake (Fig. 31). In a section, along the trench, described by Gaunt (1970, fig.2), there is clear evidence of two periods of channel incision into the till north of the Escrick Moraine (Fig. 31). Early formed channels containing gravel, presumably derived by erosion of till forming the 15 Metre Surface, were later excavated and refilled with cross-bedded and parallel (horizontally) bedded sands that comprise part of the 10 Metre Surface. The section (Fig. 31) also shows the feathered erosional contact between the laminated clays of the Elvington Glaciolacustrine Formation and the fluvial sands of the Naburn Sand Member of the Breighton Sand Formation.

It is considered that, once the level of Lake Humber had fallen below 15 m above OD, the Escrick Moraine emerged as a dam across the Vale of York between the Ouse and Derwent river outlets. Meltwater was largely diverted south-westwards towards the River Ouse outlet with perhaps only a lesser diversion towards the Derwent River outlet. Both systems then had the potential to incise fluvial channels across the northern face of the moraine. This effect could well have produced steepening on the northern face, of the moraine, as noted by Melmore (1935, fig.7) at Stillingfleet.

Gaunt (1970) has attributed the deposits exposed in the trench north of the Escrick Moraine to 'the denuded remnants of a succession of meltwater deposits and flow tills derived from the retreating ice front' with 'a subsequent phase of down-slope sludging or solifluxion' infilling the 'closed hollows'. Such a sequence of events is however not in accordance with the proposition advanced, in this research, that Lake Humber originated well after the retreat of the Vale of York glacier.

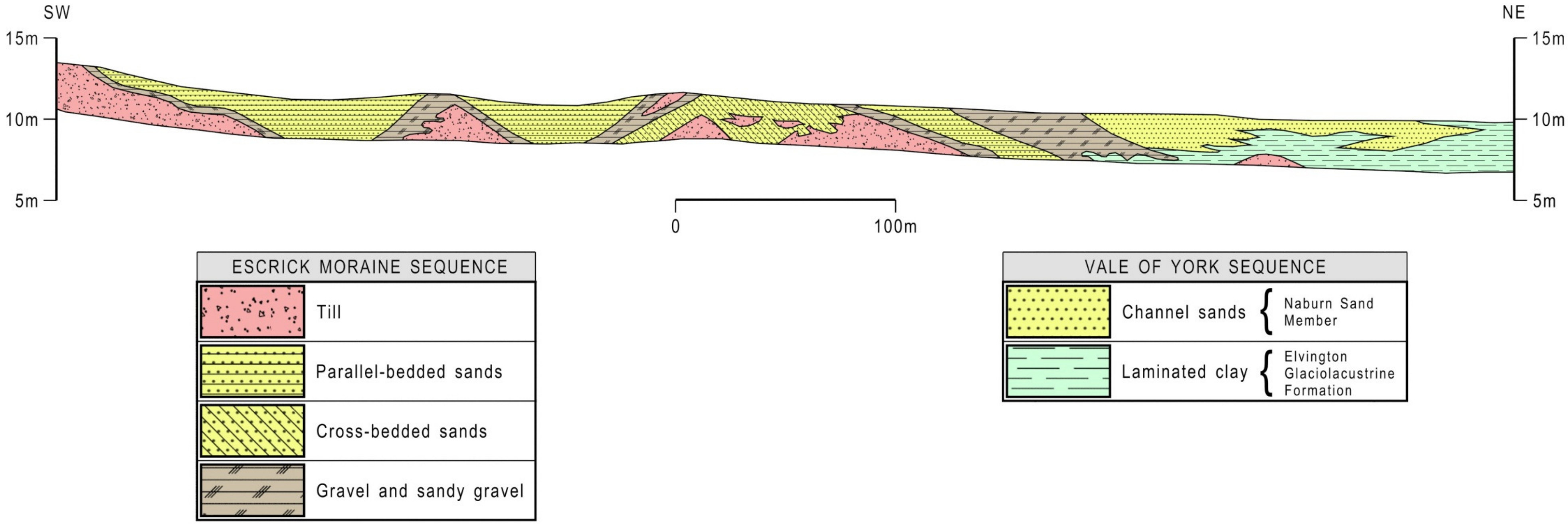


Figure 31. Lateral erosion channels along the northern face of the Escrick Moraine, which have detached the 15 Metre Surface and in places the 10 Metre Surface from the edge of the moraine (see Fig. 14). The channels indicate two phases of incision into the till below the 10 Metre Surface north of the moraine. The initial incision, containing a gravel and sandy gravel infilling, was followed by a second incision and the deposition of parallel-bedded and cross-bedded sands. Modified from Gaunt, 1970.

7.16.5. Underfit Valleys

Drainage systems that were more active, in the Vale of York, during the closing stages of sedimentation in Lake Humber and in post-lacustrine times, have often left their imprint as underfit valleys.

7.16.5.1. Great Marsh

This small river drains a lacustrine sub-basin of Lake Humber that originally must have received major flows of meltwater through Healaugh Gap and Catterton Beck. The sub-basin, which forms part of the 10 Metre Surface fringed by littoral deposits of the 10 m and 15 m terraces (Fig. 14), is mainly underlain by till with localised areas of glacio-lacustrine and glacio-fluvial deposition (British Geological Survey, 2003).

7.16.5.2. The Fleet

In contrast to the Great Marsh drainage sub-basin, The Fleet drains an area of the 10 Metre Surface to the west of the River Ouse, mainly underlain by laminated clays of the Elvington Glaciolacustrine Formation. It was extensively eroded before or during deposition of the Naburn Sand Member of the Breighton Sand Formation (see section 10.2.3. and British Geological Survey, 2008). Whilst the sub-basin may have received Late Devensian drainage from Askham Bog and Knavesmire, it must also have been an outlet for drainage during the deposition of the Naburn Sand Member in the Early Holocene.

7.16.5.3. Stillingfleet Beck

The most spectacular of the underfit valleys is Stillingfleet Beck, with flooding nowadays by flood water backing up from the River Ouse towards Stillingfleet (Plate 36). The geological history of the system must, however, extend back to the Late Devensian, with first erosion of till forming the 15 Metre Surface followed by erosion of the later established 10 Metre Surface (Fig. 14). These two phases of incision are illustrated in Figure 31. In the early Holocene the system enlarged even further with the incision of the older river terraces that are still seen at Dunning Dike, Bridge Dike and Horsecourse Dike (Fig. 14).

Plate 36. Flooding in Stillingfleet Beck (Photograph A.Rothwell)

This system may also have drained an extensive area between Crockey Hill and the River Derwent during deposition of the Naburn Sand Member across an eroded surface underlain by the Elvington Glaciolacustrine Formation and till of the Vale of York Formation (British Geological Survey, 2008). The final event along the Escrick trench would have been the incision of Stillingfleet Beck either through the older river terrace or the 10 Metre Surface (Fig. 14).

8. LIDAR IMAGERY

8.1. Introduction

The following account of LIDAR remote sensing is based on technical descriptions by Oliver Davis for the Royal Commission on the Ancient and Historical Monuments in Wales titled 'Processing and Working with LIDAR Data: in ArcGIS: A Practical Guide for Archaeologists' (2012) and from an article (after Holden *et al*., 2002) explaining the principles of LIDAR collection.

Airborne Light Detection and Ranging (LIDAR) is a method of remote sensing using a laser mounted on an aircraft that provides the ability to acquire detailed three-dimensional data over a large area in a relatively short period of time. In order to collect data the LIDAR sensor, mounted below the aircraft, emits short infrared pulses towards the earth's surface, which are fan-shaped across the flight path (Fig. 32) with each pulse giving multiple echoes or 'returns'. Typically the first return to be received is from the tops of trees and vegetation while the late return is received from the ground surface. Digital Elevation Models (DEM) generated from the data can therefore include a Digital Surface Model (DSM) from first return points, which include features such as vegetation and buildings, and a Digital Terraine Model (DTM) or 'bare-earth' model generated from last return points, which strips out vegetation even in forested country. The clarity and detail of ground features depends upon the spatial resolution or the density of measured points per square metre and it is suggested that for many features the basic minimum requirement is 1.0 m resolution. It is however recommended that more detailed information can only be obtained from a resolution of two hits per square metre gridded to 0.5m.

It is claimed that high-resolution digital elevation maps have led to significant advances in geomorphology with LIDAR able to detect subtle topographic features such as river terraces and river channel banks.

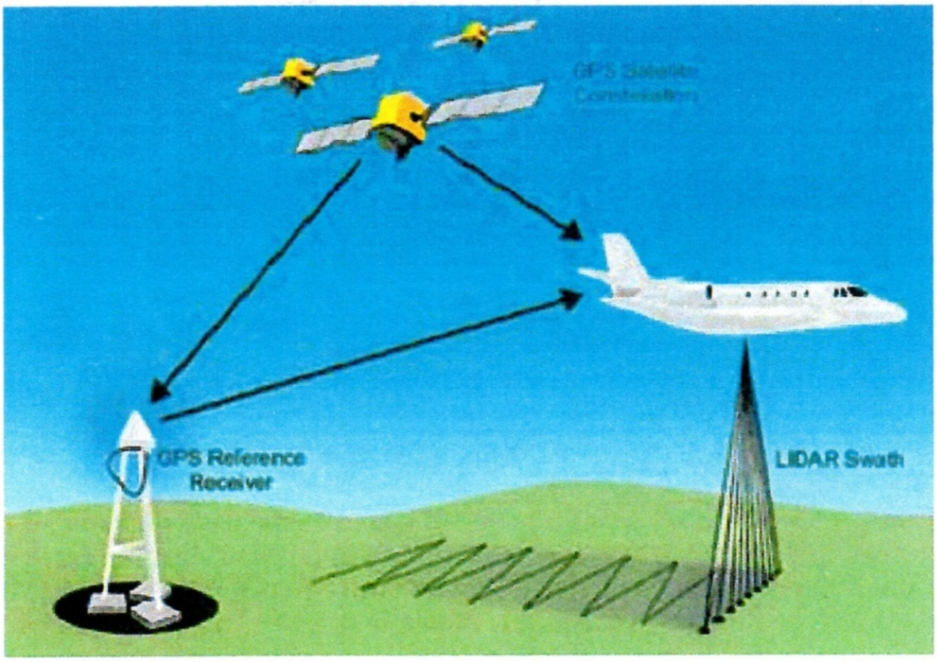


Figure 32. Principles of LIDAR collection (after Holden *et al*. 2002).

High-resolution LIDAR DEM data have previously been used for geomorphological mapping, for example, by Yang & Teller (2012). They used LIDAR to create DEM mosaics and hillshade maps over a large region in North Dakota to better define and measure elevations of strandlines associated with glacial Lake Agassiz. They reported that the LIDAR images clearly revealed the complexity of the strandlines and identified additional small and low relief ridges not recognised and mapped before, either from aerial photographs or topographic maps.

LIDAR topographic mapping has been therefore applied as part of this research to corroborate the field-based mapping undertaken.

8.2. Landform Mapping

In order to evaluate the field mapping, LIDAR imagery was obtained for two areas: one on the York Moraine below Mill Mound and the other on the western face of the Wolds between Everthorpe and South Cave. In both these areas the results of the LIDAR DEM were compared to field mapping with particular emphasis on identifying terraces and their back-walls. In addition, profiles were plotted from the southern side and along the crest of the York Moraine, from LIDAR topographic data, to illustrate the terracing and its elevation.

8.2.1. York Moraine

On the York Moraine, LIDAR Digital Surface Model (DSM) mapping with a surface spatial resolution of 50 cm was acquired over an area of 1.0 km^2 on the crestal part of the moraine that included much of Mill Mound and most of York University (Fig. 33b, 33c and 33d). Similar colour coding was used on both the field-prepared geomorphological map (Fig. 33a) and the LIDAR maps (Figs 33b and 33c) for comparison of back-wall mapping of the principal terracing at approximately 15 m, 20 m, 25 m, 30 m and 33 m above OD.

While there is good agreement in recording landforms by both forms of mapping including the distinctive alignment of the University buildings along both the 20 m and 25 m terraces, as already noted for the geomorphological landform description of the 25 m terrace (section 7.7.1.), there are significant differences in the quality of the mapping. These include:

- LIDAR mapping is most detailed where there are indentations or recesses in the back-wall of the terracing resulting from earthworks or landscaping particularly between the University buildings (e.g. the 15 m and 20 m terraces), or where there are distinctive features incorporated into the back-wall (e.g. the brick steps on the 20 m terrace back-wall below the Retreat Fig. 33b);
- the field geomorphological mapping is more suitable than the LIDAR mapping in open country where the transition from terrace to back-wall is a concave surface as shown on Figure 13;
- because of restricted access, the field geomorphological mapping through the University is more generalised than the LIDAR mapping;
- the LIDAR mapping clearly illustrates a spoon-shaped slump structure below the back-wall of the 20 m terrace (Fig. 33b), which was noticed in the field.

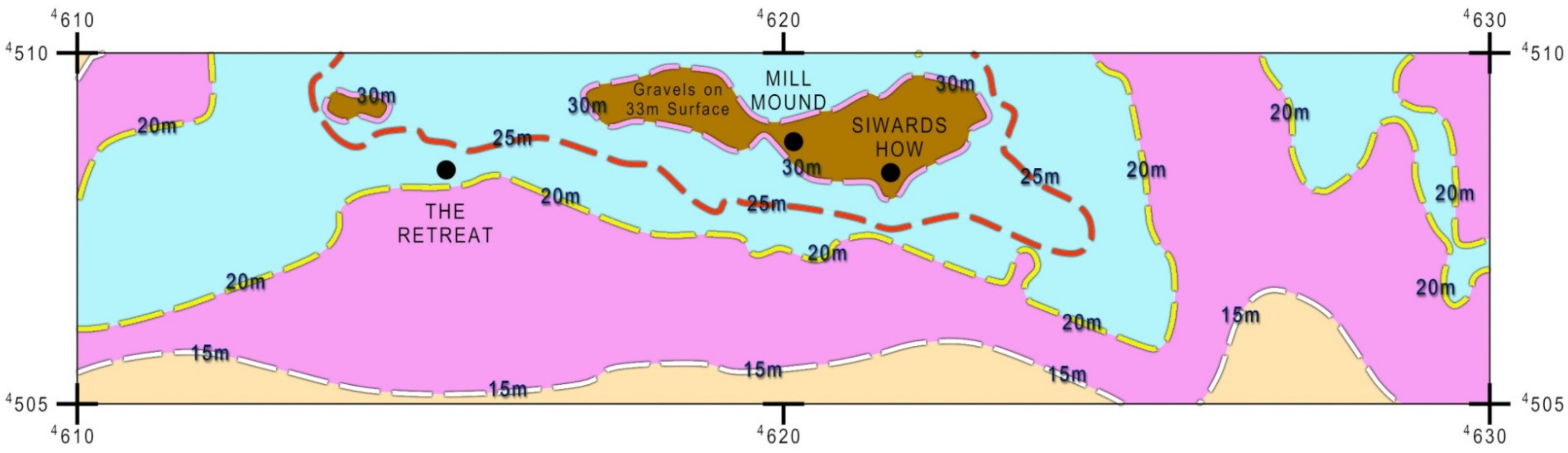


Figure 33a. Geomorphological map of part of the York Moraine covering an area of 1.0 km x 0.5 km around Mill Mound above York University. The mapping, taken from Figure 14, includes terracing at 15 m, 20 m, 25 m, 30 m and 33 m above OD. For the legend see Figure 14.

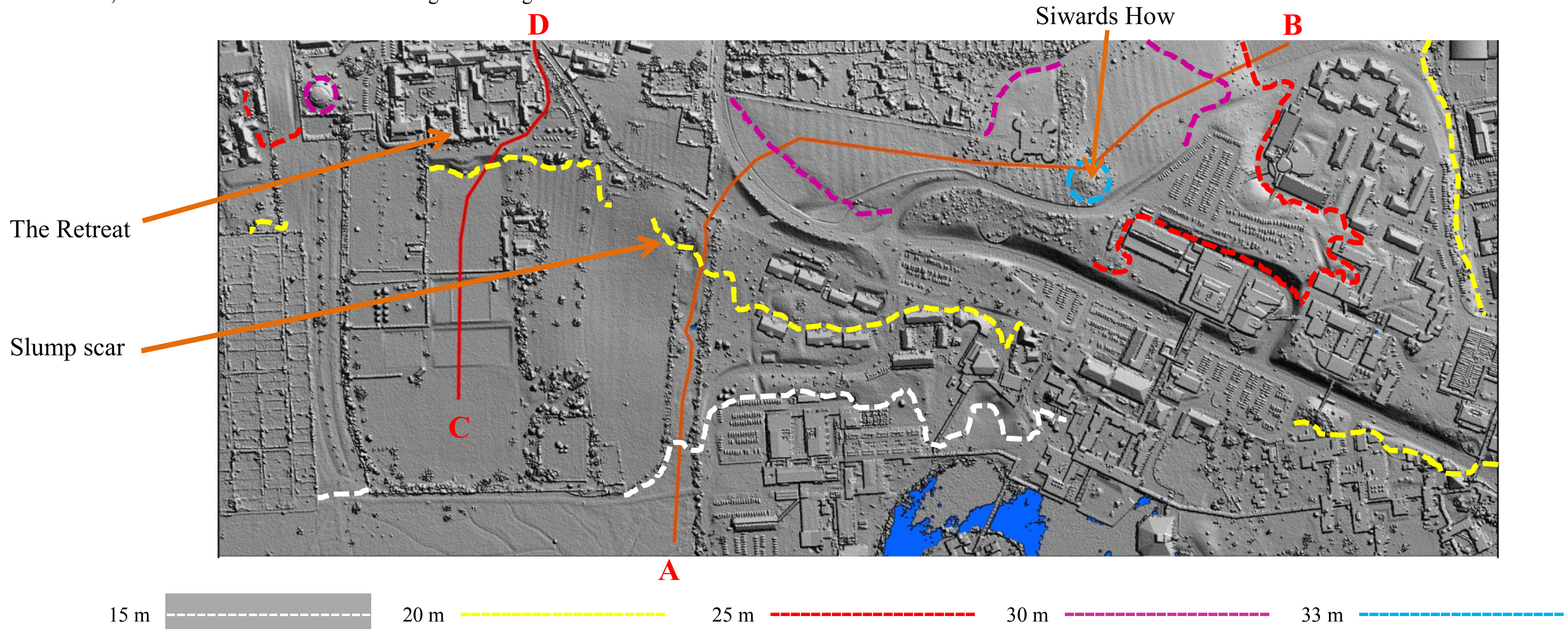
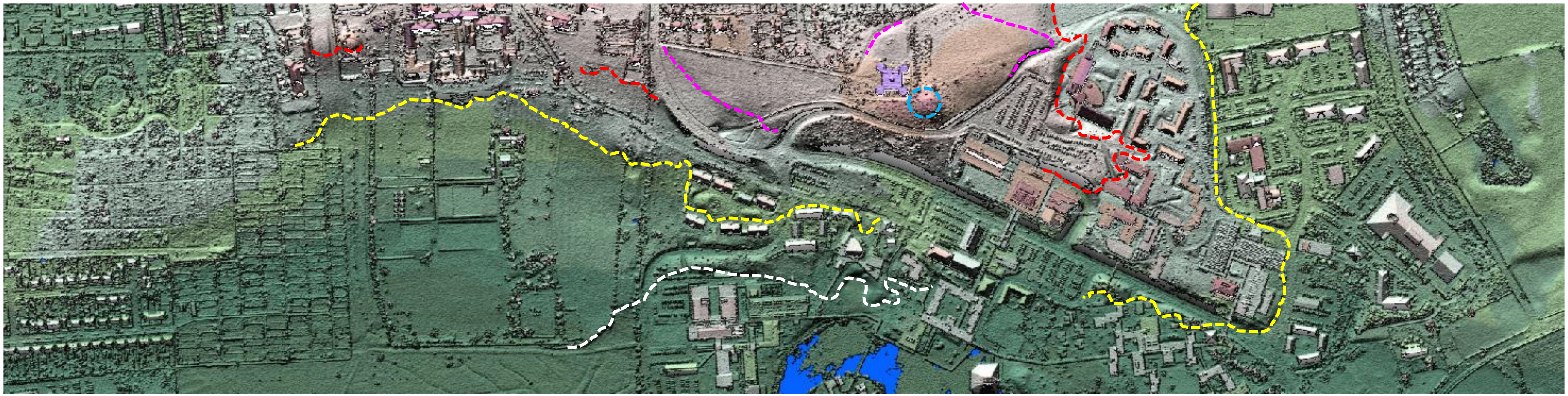


Figure 33b. LIDAR DSM hill-shade map of part of the area shown on Figure 33a. Note that the back-wall of the terracing, where recognised has been colour-coded to match terracing on Figure 33a. (Scale 0.5 km x 1.24 km)

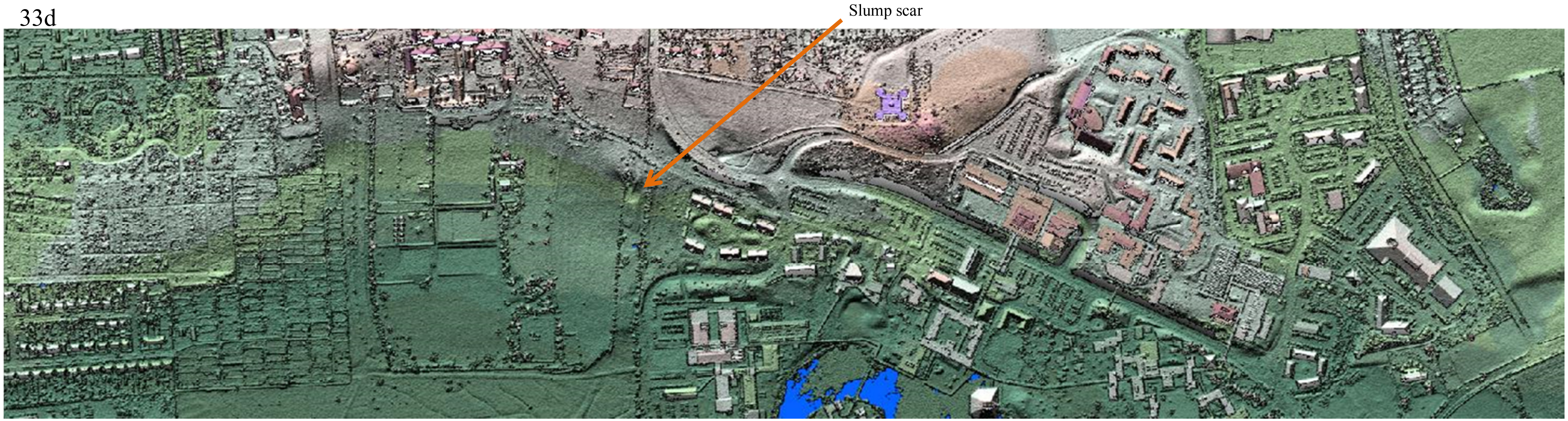
33c



15 m ------------------------ 20 m ------------------------ 25 m ------------------------ 30 m ------------------------ 33m ------------------------

Figures 33c and 33d. Alternative LIDAR DSM hill-shade maps of the area shown on Figure 33a. Note that the back-wall strandlines shown on Figure 33c have been excluded on Figure 33d. (Scale 0.5 km x 2.0 km)

33d



- the back-wall of the 33 m terrace, illustrated on Plate 17 below Siwards How, is a conspicuous feature on the LIDAR map (Fig. 33b). Although it is a mappable landform it has not been differentiated from the gravels shown on the 33 Metre Surface of Figure 33a.

To demonstrate that terracing on the York Moraine, recorded visually in the field, can be replicated by LIDAR DSM mapping, two profile graphs were plotted across the moraine. One profile (Fig. 34) extended from the southern flank of the moraine on to the crestal region of Mill Mound, as far as Siwards How, and then down to 25 m above OD on the northern side of the moraine (A-B, Fig. 33b). The other profile (Fig. 35) was specifically selected to cross the 20 m and 25 m terraces east of the Retreat (C-D, Fig. 33b).

Profile A-B

This profile (Fig. 34) illustrates the 25 m terrace on both the southern and northern slopes of the moraine, the back-wall of the 30 m formline and the 33 m terrace including the distal parts of this terrace at just over 30 m above OD.

Due to the scale of the profile, the 20 m terrace, shown on Plate 23a, is not visible on the section line. Of some interest on the profile is the sand mound near the base of the moraine below 15 m above OD. This sand may have accumulated as part of a scree-apron delta (Fig. 27) below a 15 m above OD lake level either by slumping from the 20 m terrace as illustrated in Figure 27 or by normal down-slope aggradion from a scree-cone delta. That slumping has occurred is indicated by the spoon-shaped slump scar shown on Figures 33b, 33c and 33d just below the 20 m strandline immediately west of York University (see also an example of slump scars in Fig. 24, from Nemec, 1999a).

Profile C-D

This profile (Fig. 35), near The Retreat, has been included to show that the 20 m terrace and back-wall, with the 25 m terrace above, is a robust feature on the York Moraine below Mill Mound. Its abbreviated presentation on Profile A-B (Fig. 34, Plate 23a) has been attributed to slumping. Man-made terracing on the C - D profile is shown below 16 m and 14 m above OD.

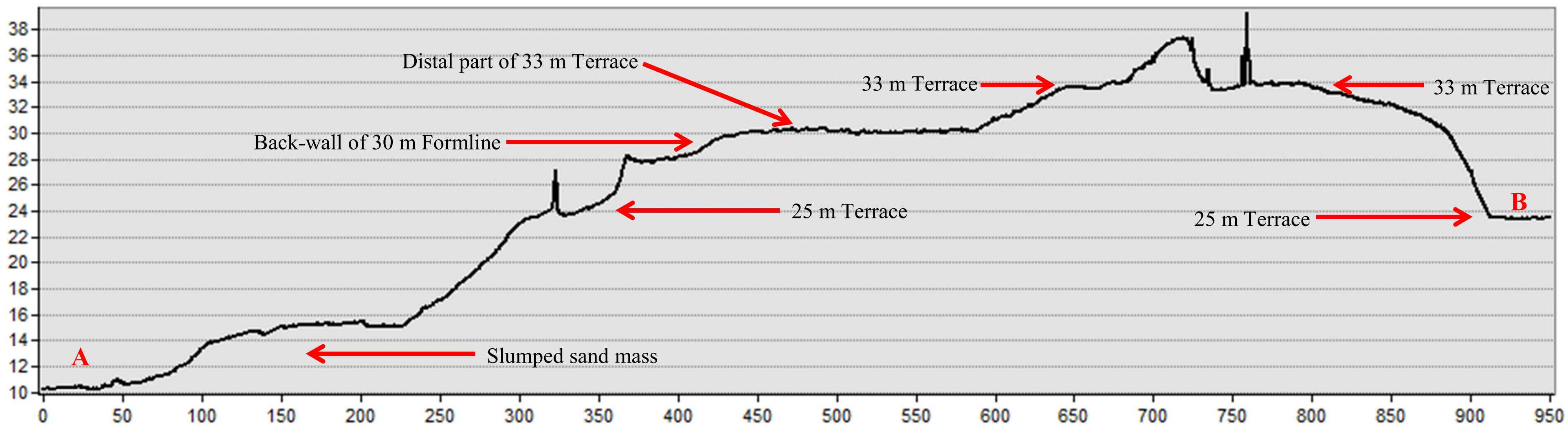


Figure 34. Profile graph A-B west of York University. The profile illustrates the loss of the 20 m terrace by slumping, the extensive 33 m terrace and the well-defined eastern part of the 25 m terrace, which marks a drainage channel through the moraine.

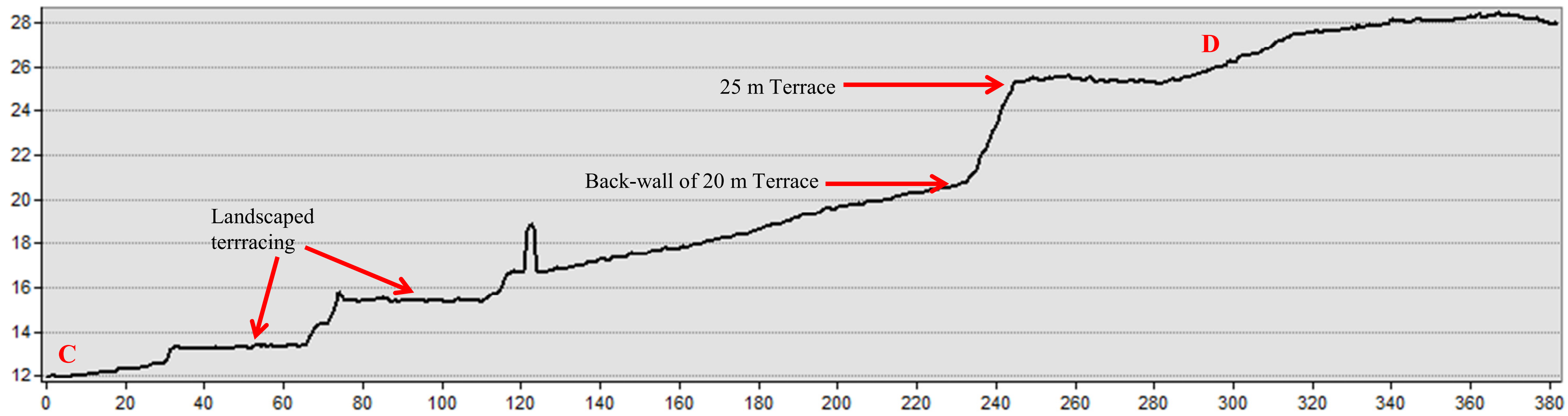


Figure 35. Profile graph C-D near The Retreat, showing landscaped terracing at 15 m above OD and the conspicuous back-wall of the 20 m terrace below The Retreat.

8.2.2. Everthorpe – South Cave

LIDAR DSM mapping was also acquired over an area of 1.0 km^2 extending between the gravel deposits on Everthorpe Hill and the Castle golf course at South Cave. In this area the acquired DSM mapping had a spatial surface resolution of 25 cm (Fig. 36b).

In general there is good back-up for the landform mapping (Fig. 36a) by the LIDAR hill-shade imagery (Fig. 36b) although the latter does appear mainly as 'ghost-like' images. This is best illustrated along the back-wall of the 20 m and 25 m terracing and parts of the 30 m and 33 m terracing. The most conspicuous agreement occurs where sections of the 25 m terrace mark the boundary between agricultural land and the rising, often uneven slope, towards farm buildings and housing in Everthorpe village. In places, however, the LIDAR mapping is not as robust as the elevation changes would suggest, for instance at one location east of Everthorpe village (Fig. 36b), where the 25 m terrace becomes less distinct in spite of its line being marked by ditching and hedgerows.

In contrast to the shadowy LIDAR presentation of the geomorphology, the detail on the golf course (southeastern part of Fig. 36b, 36c and 36d) is quite sharp with shot-hole bunkers, tee boxes and elevated greens being conspicuous: ploughed fields are also distinctive.

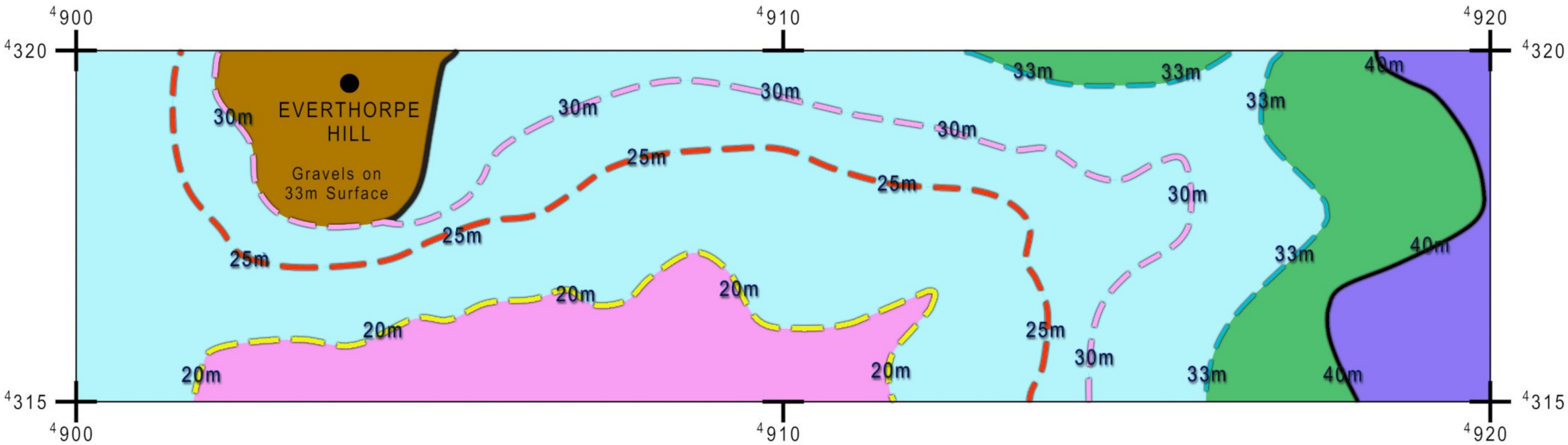


Figure 36a. Geomorphological map on the western slope of the Wolds between Everthorpe Hill and the Castle golf course at South Cave covering an area of 1.0 km x 0.5 km. The map, taken from Figure 16, includes terracing between the back-wall of the 20 m terrace to the back-wall of the 40 m terrace. For the legend see Figure 16.

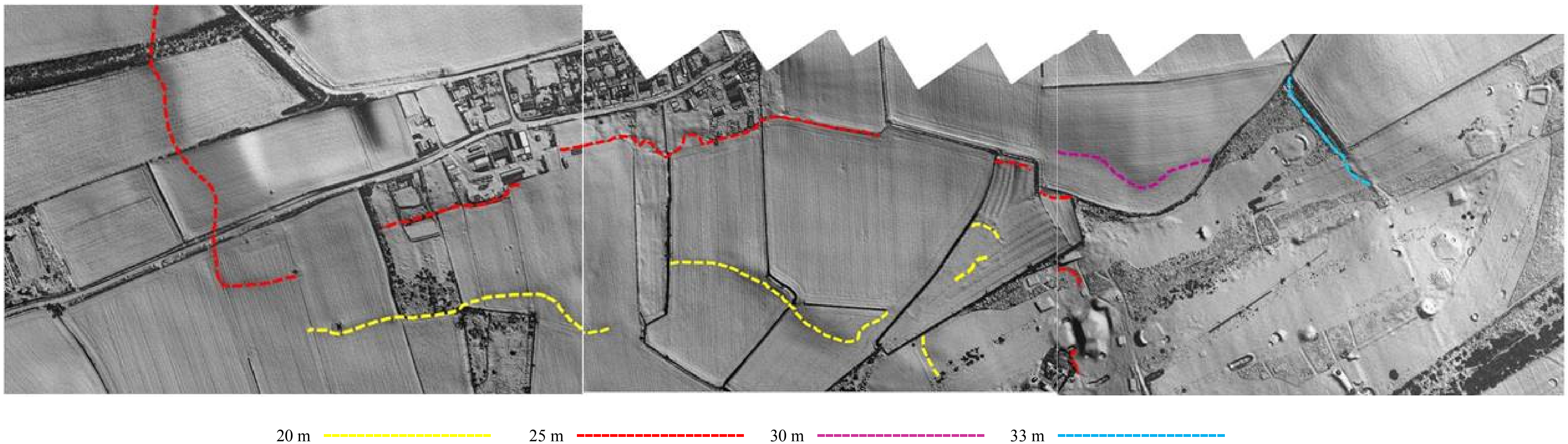


Figure 36b. LIDAR DSM hill-shade map of the area shown on Figure 36a. Surface resolution is at 0.25 cm. (Scale 1.0 km x 0.5 km). Note the old gravel pit near the bend in the road west of Everthorpe.

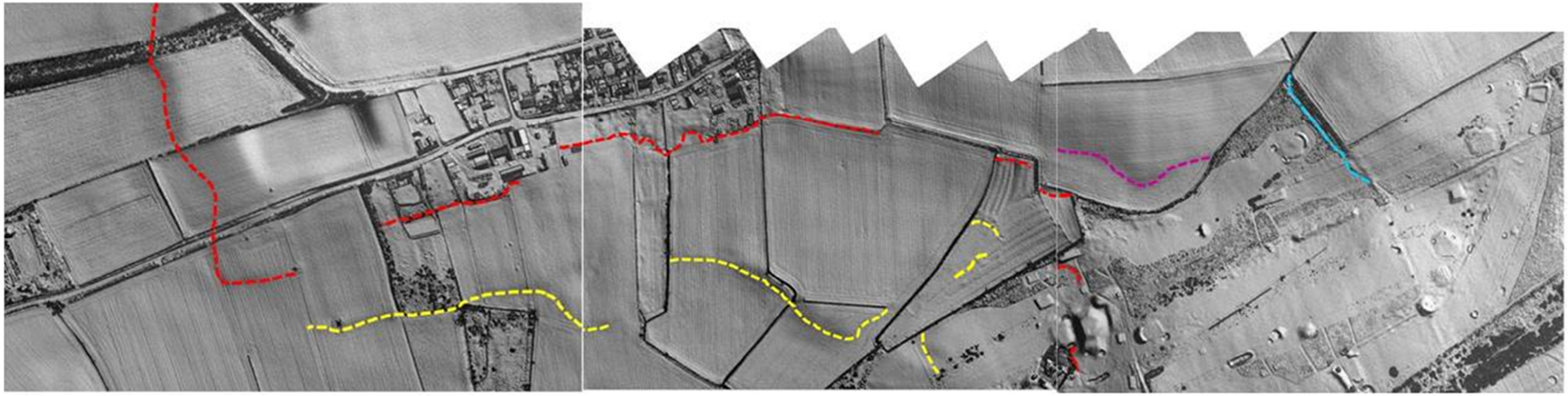


Figure 36c. LIDAR DSM hill-shade map as in Figure 36b.

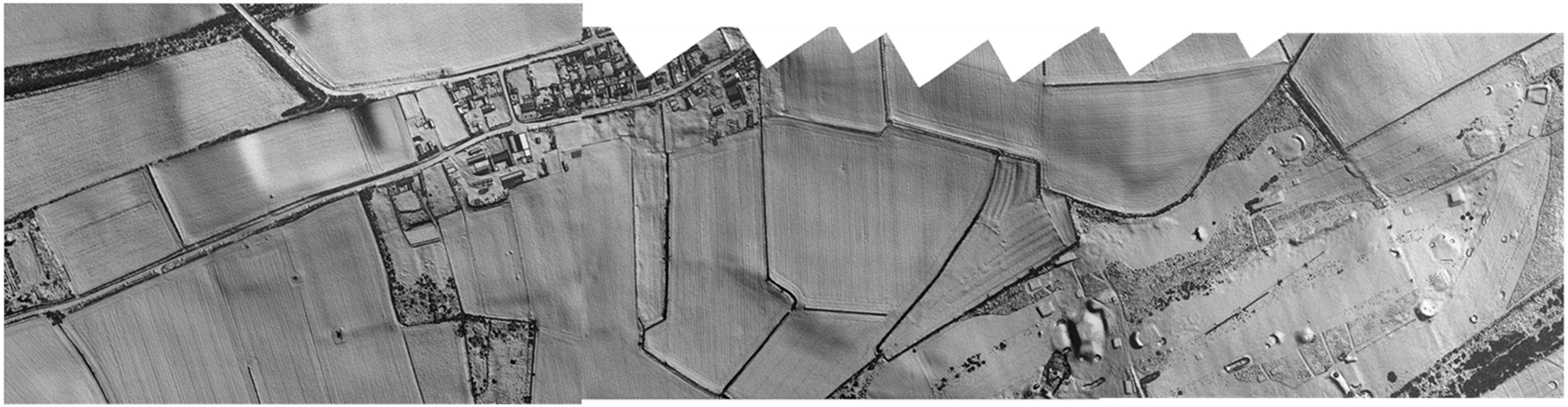


Figure 36d. LIDAR hill-shade map as in Figure 36c but without the colour-coded back-wall terracing.

8.2.3. Conclusions

It can be concluded that over these two limited areas LIDAR imagery does corroborate the regional landform mapping, which has been the major objective of this research, particularly detailing usually small, discrete, well-defined minor landforms. Larger landforms such as terracing with concave back-walls, are also apparent on the imagery, even though the back-wall definition is poor. LIDAR can therefore be used in a supportive role for field observations plotted on quality topographic maps such as those issued by the Ordnance Survey.

9. OPTICALLY STIMULATED LUMINESCENCE (OSL)

9.1. OSL Dating

To obtain OSL ages from quartz-rich sediments two measurements are required. Firstly, the palaeodose (De), or the weak flux of ionising radiation absorbed and stored by the sediment, in traps formed by defects and impurities in crystals, which is sourced from naturally-occurring radionuclides, predominantly uranium, thorium and potassium (the environmental dose rate) and from cosmic radiation (the cosmic dose rate), since burial. Secondly the annual dose rate derived from the same sources. To obtain an age the palaeodose (De in Grays) is divided by the annual dose rate (Grays/yr). For further details of these measurements see Aitken (1998). It is assumed that an accurate age can only be achieved if the sediment was bleached or its OSL signal fully reset by sunlight during transport prior to burial. Although the range for OSL dating exceeds that of ^{14}C dating and is therefore suitable for dating fluvial and lacustrine sediments during the Late Quaternary, eventually over long periods of burial the electron traps within the quartz grains will become saturated i.e. they cannot absorb any more radiation and can therefore only preserve a minimum age.

The suitability of using OSL dating for fluvial and lacustrine shoreline deposits has been documented by Bateman *et al*. (2008 and 2011) and Hartman (2011) who have provided a consistent chronology for the emplacement of the Skipsea Till and subsequent high level phase of Lake Humber. Earlier luminescence work by Bateman *et al*. (2000), from Caistor, had shown good agreement between TL dating of fluvial, lacustrine-shoreline and aeolian sands with ^{14}C dating of interbedded peat deposits. Also relevant to this text has been the dating of the Easington raised beach, or shoreline sand, by Davies and Bridgland (2013), which has been given a mean OSL age of 201 ± 28 ka considered reliable for its geological context.

9.2. Sampling Strategy

The main objective of the OSL dating was to provide a chronology, since the Last Glacial Maximum in the Vale of York, that includes the subsequent fluvial and lacustrine events, following the impounding of Lake Humber. A further objective was to determine the age, if possible, of the older periglacial deposits represented by the Older Alluvial Fans, which terracing indicates are older than Lake Humber. The Late Devensian part of this chronology should fit into an existing, partly open-ended time-framework provided by Penny *et al*., (1969); Bateman & Catt, (1996); Bateman *et al*., (2008); Bateman *et al*., (2011); and Hartmann, (2011).

An important part of the chronology, missing from the above framework, is an age for the final decline of Lake Humber, an event marked by the cessation of laminated clay deposition (i.e. the Thorganby Clay Member of the Hemingbrough Formation) followed by fluvial and erosive flooding of the Vale of York (Breighton Sand Formation) and initial incision and drainage by the major rivers into a residual or remnant 5.0 m above OD Lake (Stages 7 and 8, Fig. 52).

It was expected that the chronology could be provided by sand sampling from the York Moraine and reworked sand derived from the Older Alluvial Fans on lacustrine terraces in the Vale of York. Dating of the latter required locating sands which had not been reworked by Lake Humber i.e. sediments between the 52 m strandline and the 42 m terrace of Lake Humber.

Several problems associated with such a sampling program have been revealed by the results of the OSL dating. These include extensive flooding events prior to the MIS4 – MIS2 glaciation (see section 13 – MIS 4 stadial) that resulted in a redistribution of older fluvial deposits (e.g. the Older Alluvial Fans) and in the early Holocene with incision of the laminated clays and widespread flooding in the Vale of York (Breighton Sand Formation). Another problem associated with the York Moraine was the shoreline reworking of the sand screes (see section 7.16.1.): here age differences have been recorded from partially bleached sand deposited on the sand screes derived by flooding across the top of the moraine at elevations between 33 m and 42 m above OD that was later reworked, principally at 25 m and 20 m above OD lake levels. Other problems include human disturbance, bioturbation and aeolian reworking.

9.3. OSL Sample Collection and Preparation

To achieve the proposed chronology 18 sand samples, listed in Table 3, were collected from 15 locations shown on Figures 37 – 40. All samples were either extracted by hammering opaque PVC tubes (4.0 cm in diameter and 25 cm long) into cleaned outcrop sections or by hammering a steel sampling tube into the bottom of a hole drilled using a manually operated Dormer Engineering sand auger to obtain a 30 cm long core of sediment, which was immediately pressed into one of the PVC tubes. All tube ends were tightly capped to prevent exposure to sunlight.

In the laboratory all sample preparation was conducted under red-light conditions. Initially, potentially light-exposed material was removed from the ends of each sample tube to calculate moisture content. Part of the remainder of each sample was then treated chemically following the procedure outlined by Bateman & Catt (1996) to separate clean quartz grains from any contaminants. This involved first treating the samples with 10% hydrochloric acid to remove carbonates and then 10% hydrogen peroxide to remove organic matter. The samples were then dry-sieved into four size fractions between 90 - 125, 125 - 180, 180 - 212 and 212 - 250 µm. The heavy mineral content of one of these fractions (commonly 125 – 180 µm – providing quantity was sufficient) was then removed using sodium polytungstate ($Na_6\,O_{39}\,W_{12}.\,H_2O$) with a specific gravity of 2.7 g/cm^3 and the remaining quartz-rich material was etched in 40% hydrofluoric acid for 60 minutes, to corrode and reduce in size any feldspar present and to remove the outer alpha-irradiated skin (~10µm) of each quartz grain. The samples were then washed in hydrochloric acid to remove fluorites and re-sieved to remove any sub-size feldspar grains. Material from each sample for single aliquot analysis was then mounted, as a mono-layer, on to 9.6 mm diameter stainless steel discs with silicon oil spray.

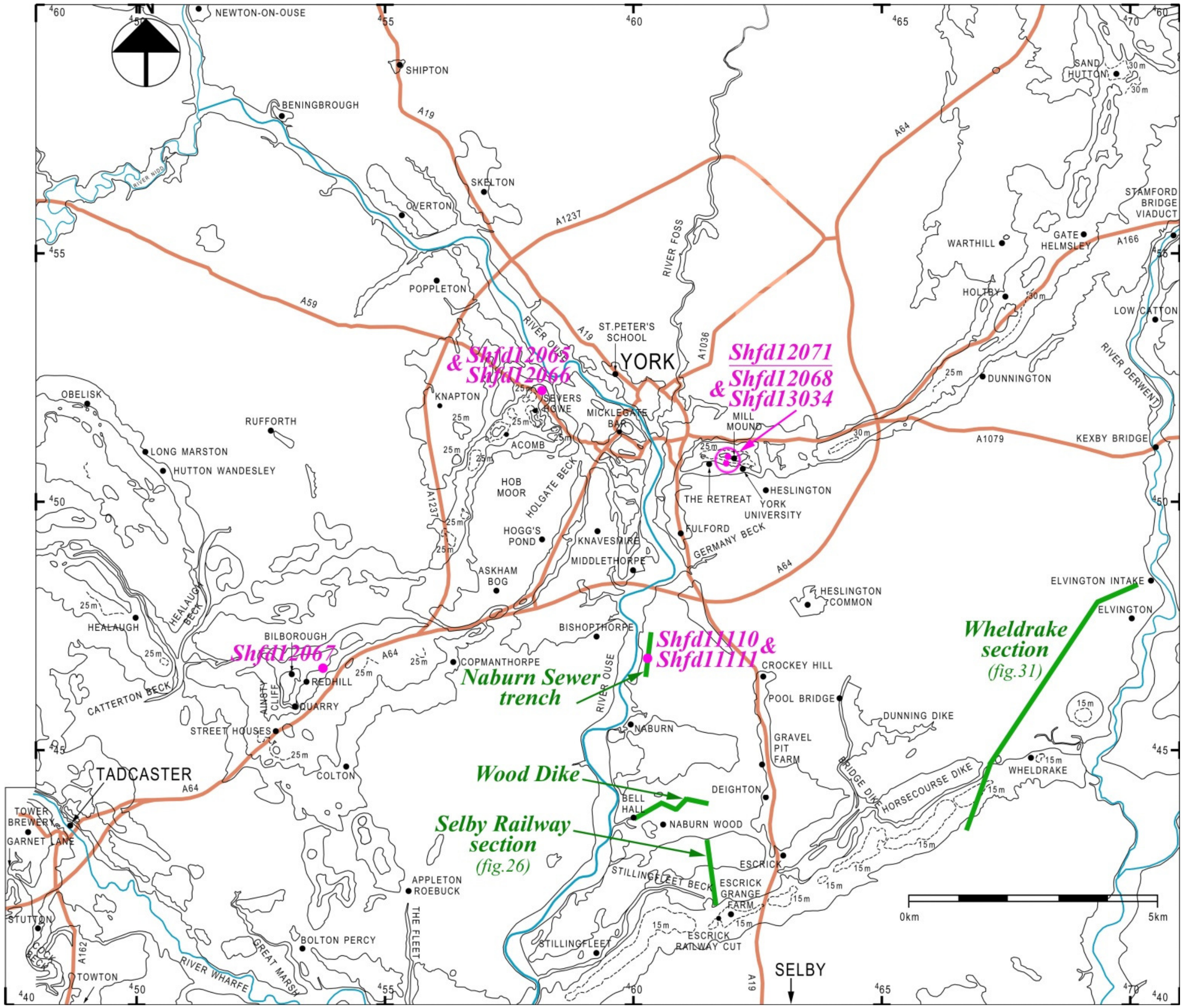


Figure 37. Location plan for (1) Shfd OSL sample nos 11110, 11111, 12065, 12066, 12067, 12068, 12071 and 13034 (2) Trench or ditch sections at Naburn sewer, Wood Dike, Selby railway (disused) and Wheldrake.

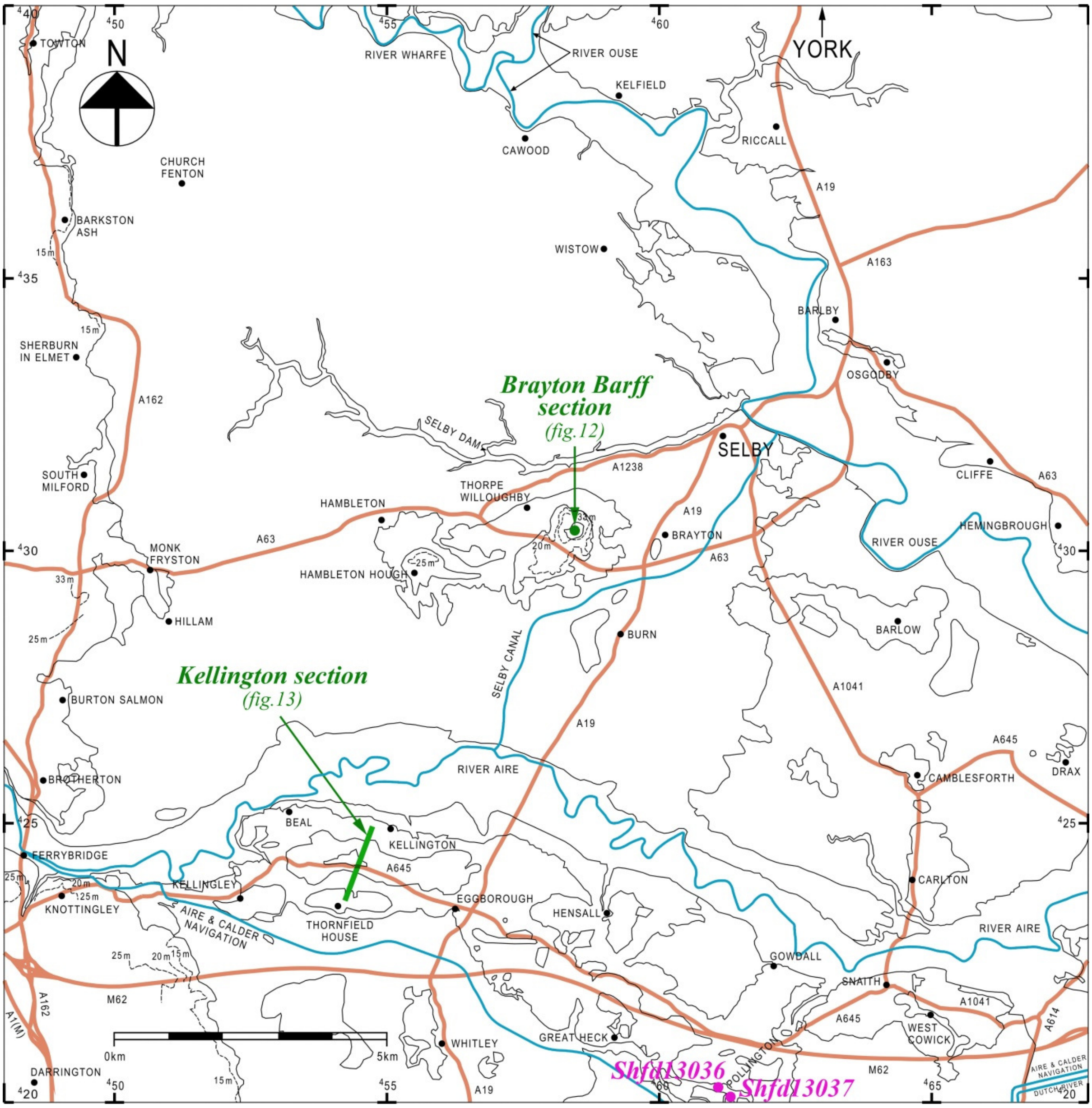


Figure 38. Location plan for (1) Shfd OSL sample nos 13036 and 13037 (2) Trench sections at Brayton Barff and Kellington.

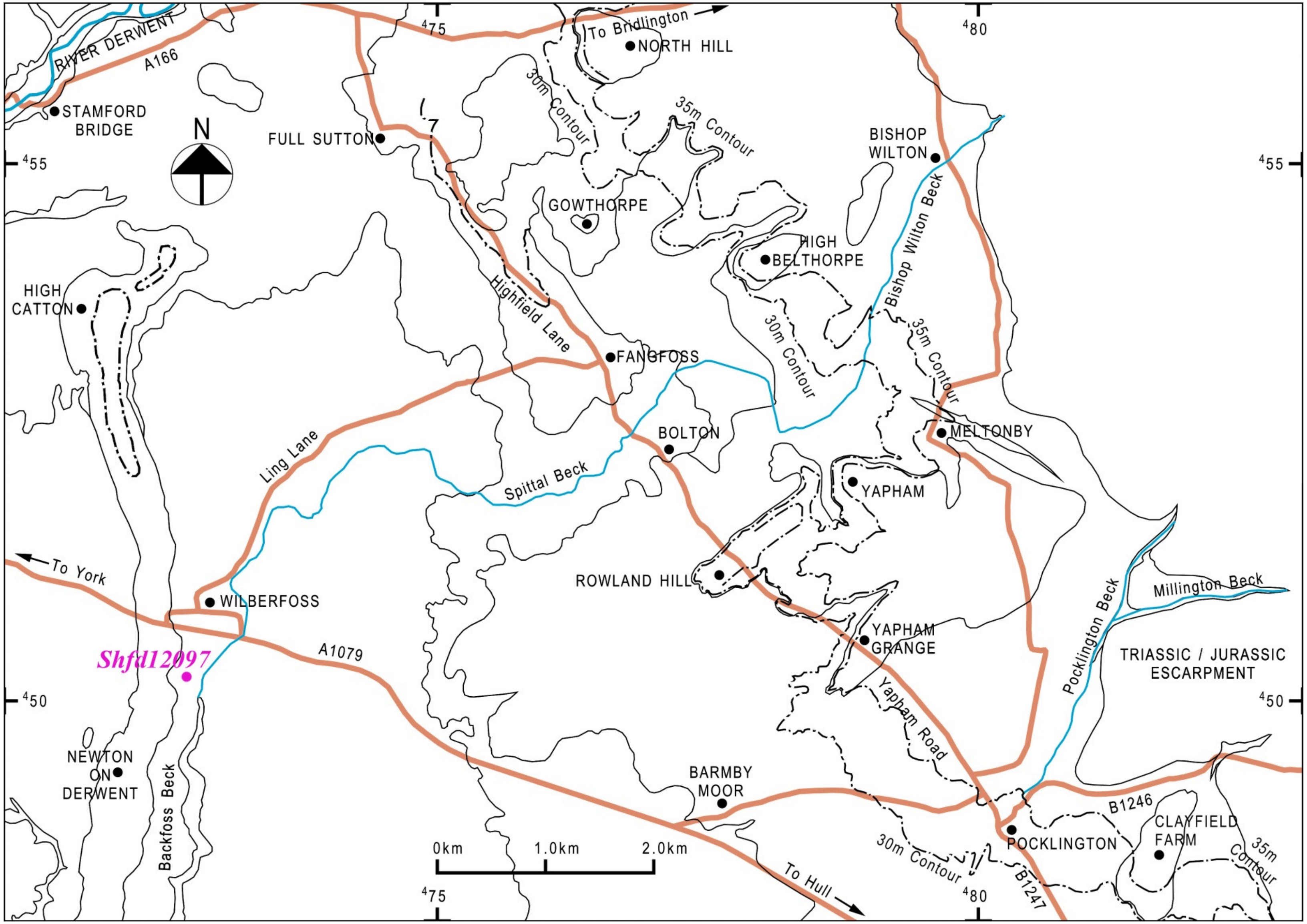


Figure 39. Location plan for OSL sample Shfd 12097.

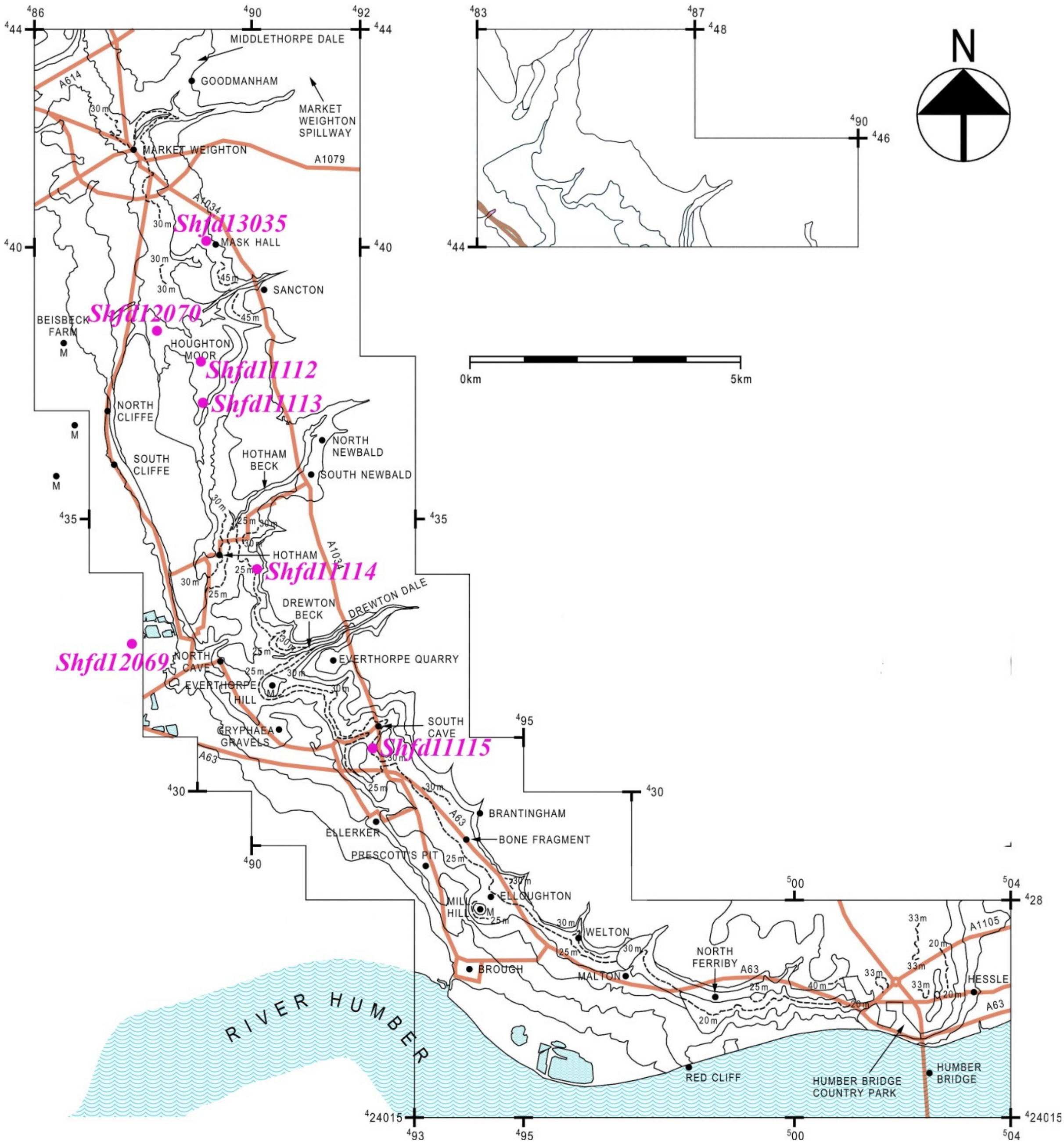


Figure 40. Location plan for Shfd OSL sample nos 11112, 11113, 11114, 11115, 12069, 12070 and 13035.

9.4. Dose Rate Determination

The environmental dose rate from naturally occurring uranium, thorium and potassium, which decay over long time scales thus ensuring a constant irradiation, was normally determined directly in the field using an EG&G Micronomad field gamma-ray spectrometer fitted with a 2 inch sodium iodide crystal. Measurement times amounted to 45 minutes per sample with the instrument probe inserted in the sample hole. In some instances, when the instrument was not available, elemental concentrations were obtained from a small pulverised portion of the sample sent to SGS Laboratories in Canada for inductively coupled mass spectrometry (ICP – MS) for uranium and thorium and inductively coupled plasma spectroscopy (ICP-AES) for potassium. In all cases the elemental concentrations were converted to annual dose rates (Table 3) using data from Adamiec & Aitken (1998), Aitken (1998) and Marsh *et al*. (2002).

TABLE 3

OSL Data for Sample Sites in the Vale of York

Site	Sample No.	Lat. (°N)	Long. (°W)	Alt. (m)	Depth (m)	MMIStur e (%)	Size (µm)	K (%)	U (ppm)	Th (ppm)	Dose Rate		Cosmic		De		Age	
											(µGY/ka)	±	(µGY/ka)	±	(Gy)	±	(ka)	±
Naburn sewer 1	Shfd11110	53.9161	1.0833	10.0	1.0	14.16	125-180	1.2	1.33	4.6	1739	87	184	9	7.51	0.32	4.32	0.28
Naburn sewer 2	Shfd11111	53.9161	1.0833	10.0	1.0	14.81	125-180	1.4	1.38	5.0	1925	98	184	9	17.58	0.76	9.13	0.61
Houghton Moor 1	Shfd11112	53.8289	0.6442	38.8	0.75	15.58	125-180	0.8	0.89	2.3	1166	56	191	10	8.75	0.21	7.51	0.40
Houghton Moor 2	Shfd11113	53.8236	0.6431	38.5	1.30	18.27	125-180	0.7	0.60	1.5	936	46	178	9	7.71	0.27	8.24	0.49
Hotham 1	Shfd11114	53.7958	0.8981	25.0	0.8	24.66	125-180	0.9	1.63	4.5	1363	62	190	9	1.04	0.10	0.76	0.08
South Cave 1	Shfd11115	53.7328	0.6020	35.0	0.85	16.05	125-180	1.3	1.63	4.4	1829	92	189	9	370.8	10.6	202.7	11.7
Severs Howe 1	Shfd12065	53.9622	1.1131	20.0	0.60	5.70	125-180	1.43	1.39	4.79	2162	112	195	10	112.4	3.8	52.0	3.2
Severs Howe 2	Shfd12066	53.9622	1.1131	20.0	0.95	4.60	125-180	1.29	1.23	3.81	1952	101	186	9	123.1	3.9	63.1	3.8
Bilbrough Top 1	Shfd12067	53.9142	1.1803	42.0	0.65	6.50	125-180				204	10	194	10	1.24	0.07	6.07	0.46
Mill Mound 2	Shfd12068	53.9495	1.0581	22.0	0.86	16.10	125-180	1.55	1.55	6.62	2139	109	188	9	33.95	0.73	15.9	0.9
North Cave 1	Shfd12069	53.7839	0.6658	6.0	4.00	9.10	125-180	1.11	0.83	3.75	1535	82	124	6	89.62	1.77	58.4	3.3
Manor Farm 1	Shfd12070	53.8350	0.6575	37.0	0.55	7.20	125-180	0.74	1.56	4.25	1509	68	197	10	2.67	0.11	1.77	0.11
Mill Mound 1	Shfd12071	53.9503	1.0561	29.5	0.55	20.50	125-212	1.17	1.58	4.48	1624	74	196	10	3.41	0.19	2.10	0.16
Newton 1	Shfd12097	53.9442	0.8897	14.0	1.0													
Sand						13.50	90-180	1.9	0.83	3.0								
Clay						20.90		3.0	2.93	12.3								
Combined		-	-	-	-	-	-	-	-	-	2345	119	169	8	10.26	0.63	4.37	0.35
Mill Mound 3	Shfd13034	53.9495	1.0581	22.0	1.76	10.85	125-180	1.06	1.09	4.33	1596	68	166	8	24.27	0.33	15.21	0.68
Mask Hall 1	Shfd13035	53.8500	0.6438	45.0	0.60	7.40	125-180	0.81	1.05	4.60	1476	68	196	10	4.9	0.19	3.32	0.20
Pollington 1	Shfd13036	53.6745	1.0747	11.0	0.55	3.85	125-180	0.83	1.26	4.76	1545	77	127	6	23.95	0.27	15.5	0.8
Pollington 2	Shfd13037	53.6730	1.0722	6.2	3.20	4.58	125-180	1.65	0.89	4.44	2174	126	116	6	103.6	2.6	47.6	3.0

The cosmic dose rate, which depends upon altitude, latitude and burial depth (Prescott & Hutton, 1994), provides a small but not negligible contribution to the total dose rate. This value is calculated using an algorithm given by Prescott & Hutton (1994, fig. 2). It was assumed for all samples that the location details had remained constant since burial.

Final calculation of the dose rate takes into account the attenuation and variability relating to sediment grain size used (Duller, 2008a) and possible fluctuations between palaeomoisture and present-day moisture values (Aitken, 1998). Although the presence of water in inter-grain void space can have a significant effect on the radiation dose from the radioactivity in the sample and a correction must be made for this attenuation (Chauhan, 2013), the correction can only be based on the moisture content of the tube-collected sample, as changes in water saturation for the sample, since burial, are unknown.

9.5. Determination of the Palaeodose or Equivalent Dose (De)

All luminescence measurements were made in the Sheffield University laboratory using a TL-DA-18 (Ermintrude) automated Risø TL/OSL reader equipped with blue LED stimulation. Irradiation of the aliquots of samples mounted on 9.6 mm diameter steel discs was from calibrated sealed ^{90}Sr beta sources with luminescence measurements made through Hoya U-340 filters.

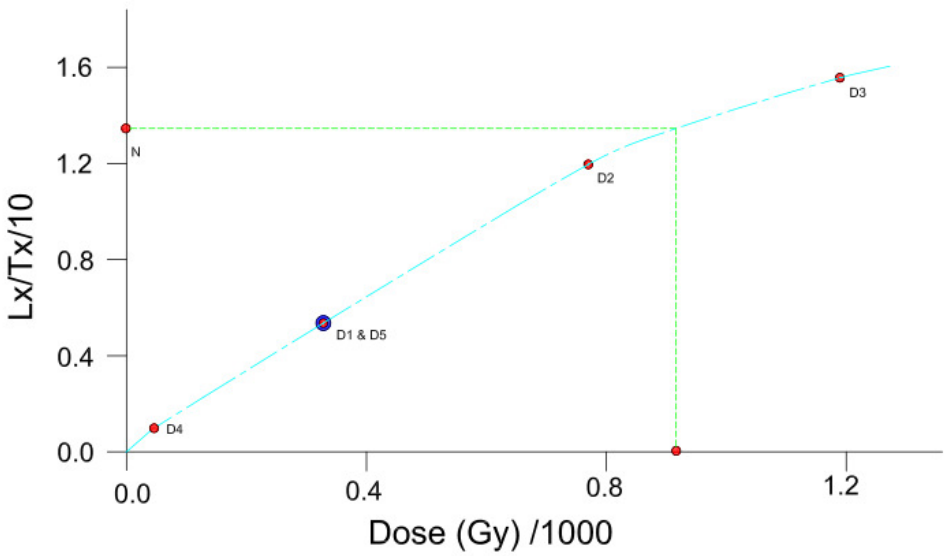


Figure 41a. Growth curve generated using SAR OSL. The luminescence response (Lx) to a series of known doses (D1-5) is normalised by a test dose response (Tx) and plotted against dose. The green line represents the interpolation of the natural dose (De) from the natural luminescence response (N). Sample Shfd12069

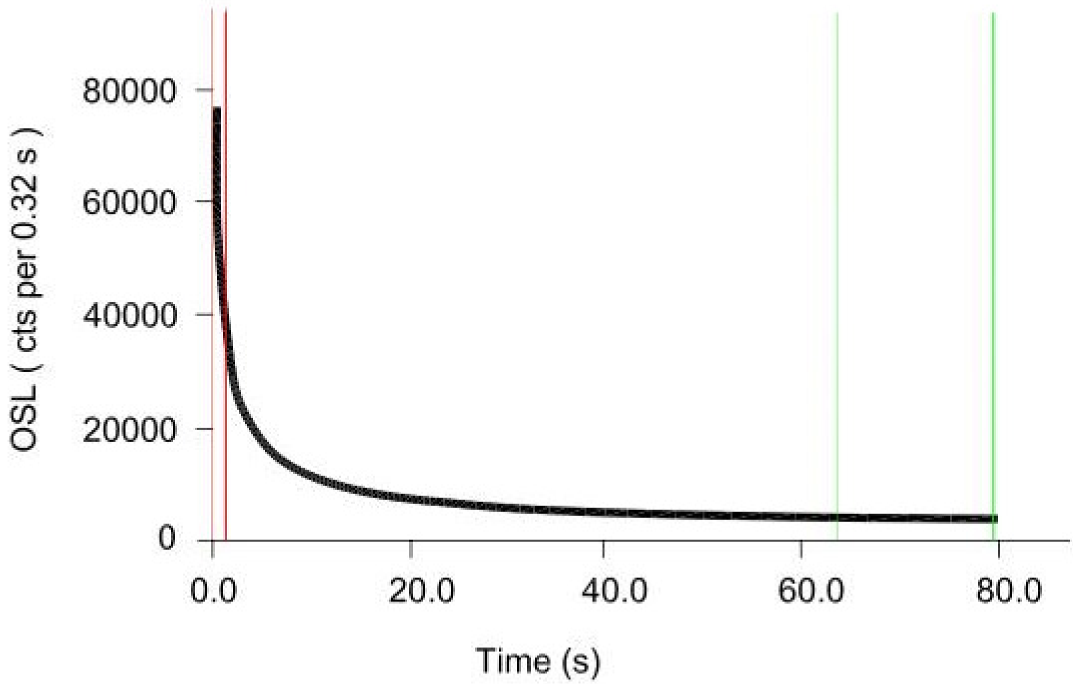


Figure 41b. Decay or shine down curve for sample Shfd12069. The red lines indicate the integration limits for the signal measurement and the green lines the background measurement once the signal has been zeroed.

Samples were analysed using the single aliquot regenerative (SAR) protocol of Murray and Wintle (2000), which involved constructing a growth curve from repeated measurements of an aliquot that had been given various laboratory irradiations, or doses (Fig. 41a). Generally five regenerative points were used to establish the growth curve with the first three (D1, D2 and D3, Fig. 41a) bracketing the natural dose the value of which had been roughly established from an initial range finder test using an abbreviated SAR protocol on three aliquots with only two regenerative points and a preheat of 200°C. The fourth regenerative point (D4) on the growth curve was a zero point with the fifth point (D5) identical to D1. The principal feature of the SAR protocol is the use of a long stimulation period to effectively zero the luminescence signal (Fig. 41b) after each measurement permitting intermediate measurements of a fixed test dose to be made to normalise any sensitivity changes that occur during the sequence (Fig. 41a). The recycling ratio of D1/D5 should be as close to unity as possible and acceptable within 10% of unity, to assess the efficiency of the test dose normalisation. The natural dose (De) of the sample can be interpolated from the growth curve knowing its natural luminescence (N) determined from the integration limits of the 'fast' component on the decay curve (Fig. 41b).

9.6. Recuperation

An important check on the reliability of the corrected growth curve is based on the assumption that the dose-response starts at zero (Murray & Wintle, 2000). In an ideal sample one would not expect any significant OSL response by using O Gy as a regenerative dose (e.g. D4 on Fig. 41a). This principle is therefore used in the recuperative check to determine whether some of the charge carriers from inactive deep traps have transferred to active luminescence traps by preheating. A value for recuperation can be determined by calculating the Lx/Tx ratio of the zero dose point (D4) as a percentage of the Lx/Tx ratio of the natural dose (N). A recuperative value of <5.0% is regarded as acceptable as for sample Shfd12069 (Fig. 41a).

9.7. Preheating

As explained by Aitken (1998) preheating of samples prior to stimulation in the SAR protocol is a necessary requirement to avoid contamination of the artificial OSL by direct contribution of electrons from light sensitive shallow traps, which are not contributed to by natural OSL. It also removes contributions in shallow traps, from the natural OSL that have resulted from thermal transfer during burial. The principal aim therefore is to normalise the artificial OSL with respect to the natural OSL.

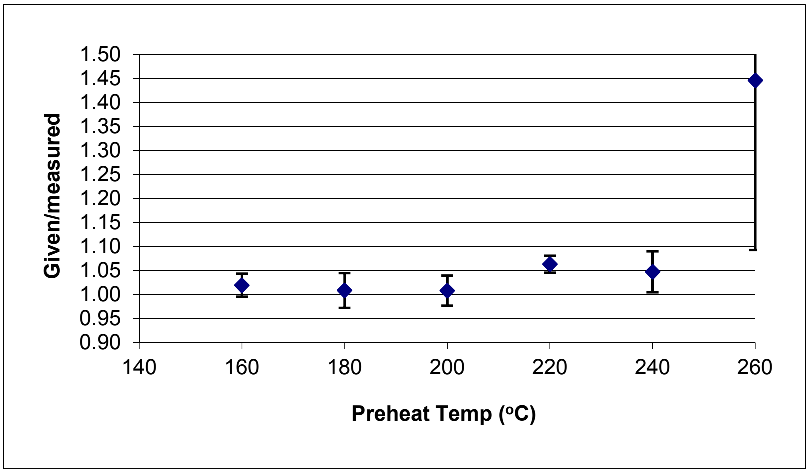


Figure 42a Results of preheat plateau test for sample Shfd12065 using preheat ranging from 160°C to 260°C. A preheat of 160°C was selected.

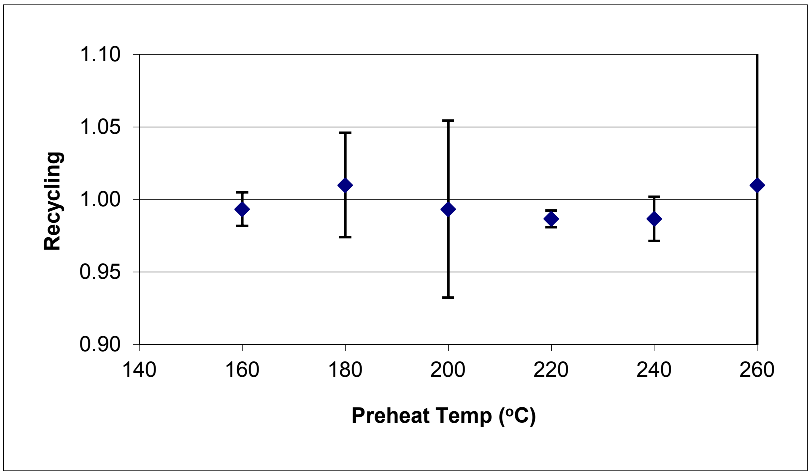


Figure 42b. Results of the recyling ratios D1/D5 for sample Shfd12065 over the same preheats as in Fig. 42a.

To determine the most appropriate preheat temperature, dose recovery preheat tests were undertaken to establish the preheat plateau test of Murray & Wintle (2003). To achieve this, 3 aliquots of each sample were optically bleached at room temperature and then given a known laboratory dose close to the rangefinder De of the sample. The artificial De was then determined using the SAR OSL protocol with five regeneration points. This procedure was completed six times with different preheats ranging from 160°C to 260°C. The results shown in Figure 42a indicate a plateau close to unity (as expected if the procedure is working correctly) between 160°C and 240°C. The recycling ratios D1/D5, over the same temperature range, are also close to unity (Fig. 42b). An optimum preheat for each sample site was chosen in this way in a range from 160°C to 220°C.

9.8 Feldspar Contamination

In contrast to quartz, where electrons exhibit stability in deep traps, laboratory experiments have shown that electron storage in feldspar is less stable than expected, a situation called anomalous fading (Duller, 2008b). This means that the magnitude of the stimulated luminescence signal can decrease during sample storage. To overcome this problem and to obtain aliquots of quartz grains only, the prepared sample for luminescence measurement is etched with hydrofluoric acid (HF) or in the case of silt-sized fractions by hydrofluorosilicic acid ($H_2S_iF_6$) to remove any feldspar contaminants. To ensure that the procedure has been successful, before beginning the OSL measurement, some aliquots of the sample are first subjected to infrared stimulated luminescence (IRSL) measurement. If no IRSL is detectable it can be assumed that feldspar is absent, but if this is not the case, a re-etch may be required or the sample subjected to a modified SAR protocol.

Banerjee et al. (2001), to overcome the problem of feldspar contamination, particularly in polymineralic fine-grained (4-11μm) colluvium, reported the behaviour of three OSL signals, namely blue-stimulated (OSL), infrared-stimulated luminescence (IRSL) and blue-stimulated luminescence following infrared (IR) stimulation [post-IR] OSL. These authors concluded that the [post-IR] OSL provides reliable thermally-stable estimates of De that could have originated from either quartz or feldspar and that the IRSL estimates of De are usually higher that the [post-IR] OSL. This procedure of a [post-IR] OSL stimulation has been referred to by Roberts & Wintle (2001) and Roberts (2007) as the double-SAR protocol or the IRSL 'wash' by Wilson *et al*. (2008). A precise account of sample preparation and measurement is detailed by Roberts (2007). Roberts & Wintle (2001) noted that values of De are greater for IRSL than for [post-IR] OSL (*cf*. Banerjee *et al*., 2001) but there is a close agreement between both sets of data for preheats below 220°C. Roberts (2007) also confirms that where the proportion of quartz to feldspar is large, the OSL signal appeared to be dominated by quartz, so that the double-SAR technique potentially provides an effective means of obtaining a quartz-dominated signal from a polymineralic source.

Of the eighteen samples collected during the research for OSL dating only five (Shfd12097, Shfd13034, Shfd13035, Shfd13036 and Shfd13037) exhibited feldspar contamination after the HF etch and for these the double-SAR protocol was applied.

9.9 OSL Dating Results

The eighteen samples (listed in Table 3) collected to establish a chronology for the events described and mapped in this research mainly originated from five primary sources. These are:

- deposits older than the Older Alluvial Fans;
- sediment deposited as part of the Older Alluvial Fans;
- sand derived from the York and Escrick moraines which most likely originated from the Sherwood Sandstone;
- sand derived directly from the Sherwood Sandstone;
- sediment eroded from the Pennine escarpment by the 100 Foot Strandline.

Erosion, transport and burial could have occurred with all these primary deposits (a situation that may not be obvious from field evidence) and it is therefore important to differentiate between any primary or secondary burial events along with other factors, which may have altered the final OSL signal for the sample. This point is particularly relevant for the York Moraine, where the most important factor involved in the interpretation of the OSL signal recorded for four of the six samples taken, has been the influence of transgressive shorelines across the crest of the moraine to produce sand screes (see section 3.3.2.1.; Figs 24 and 27). It seems very likely that sand deposited as part of the original sand scree, probably under sub-aqueous lacustrine conditions, might not have been as fully bleached as sand reworked from the sand scree along shorelines on the moraine.

Naburn Sewer samples 1 and 2 (Fig. 37)

These duplicate samples from the fluvial, partly aeolian, Naburn Sand Member of the Breighton Sand Formation were taken in a temporary sewer trench section. Dating was expected to give a Late Devensian or Early Holocene age for the sand member and provide a minimum age for the cessation of deposition of the underlying laminated clays.

Shfd11110 and Shfd11111 (Fig 43 (1) and (m)

Both these samples show a wide almost bimodal distribution of De values particularly Shfd11111 (Fig. 43, (m)). This tends to reflect mixing of older and younger end-members (see Bateman *et al*., 2003), as is implied in a description of the sand by Ford *et al*. (2008) who interpret the sand 'as having been deposited by a complex interaction of fluvial and aeolian processes. Because of the bimodal distribution, the mean age of 4.32 ± 0.26 ka for Shfd11110 has two components, a younger age of 3.5 ± 0.29 ka and an older age of 5.0 ± 0.28 ka, while the mean age of 9.13 ± 0.61 ka for Shfd11111 similarly comprises 7.8 ± 0.54 ka and 11.0 ± 0.66 ka.

Houghton Moor 1 and 2 (Fig. 40)

These two samples, either reworked from, or forming remnants of, the 40 m terrace at Houghton Moor, were expected to provide an age for high stands of Lake Humber at either 40 m or 33 m above OD lake level (Stages 2 and 3).

Shfd11112 and Shfd11113 (Fig. 43 (g), (h), (i) and (j))

The two Houghton Moor samples show a unimodal distribution of De values that give OSL ages of 7.51 ± 0.04 ka and 8.24 ± 0.49 ka that dates them to the early Holocene. Outliers in both distributions (Fig. 43 (g) and (i)) may indicate some older sand grains.

The results indicate that the sand has been redeposited from older sources, by fluvial or aeolian activity, in the early Holocene and the OSL signal reset.

Hotham 1 (Fig. 40)
This sample, taken from the side of a ditch on a terrace in the Hotham Valley, was intended to date sand deposited on the 25 m terrace of Lake Humber (Stage 4).
Shfd11114 (Fig. 43 (n))
The sample gave consistently low De values with a well-defined modal distribution of 1.04 Gy that gave an age of 0.76 ± 0.08 ka. Outliers giving some older De are not considered significant based on the overall De distribution. The result suggests that the sample came from infill in an older ditch excavation.

South Cave 1 (Fig. 40)
Originating from the landscaped back-wall of the 33m terrace in the Older Alluvial Fans, this sample was designed to date sand, which had not been subjected to superficial reworking by Lake Humber (cf. the Severs Howe samples Shfd12065 and Shfd12066).
Shfd11115 (Fig. 43 (o))
The De values, with a slightly bimodal distribution, reflecting that some grains have an antecedent De, gave an age of 202.7 ± 11.7 ka. This age provides the only substantive evidence that the Older Alluvial Fans may have been deposited close to the MIS 7 interglacial.

Severs Howe 1 and 2 (Fig. 37)
Two sand samples were taken from a layered sand sequence, with erosional contacts, exposed in a small excavation cut into the back-wall of the 20 m terrace (Stage 5) below Severs Howe (Plate 7). It was expected that these samples could provide an age for episodes of sand washing down the southern side of the moraine during the formation of the primary sand scree, which had not been subsequently reworked on the 20 m terrace.
Shfd12065 and Shfd12066 (Fig. 43 (a) and (b))
Both these samples show a strong, almost bimodal De tail, reflecting partial bleaching of some (perhaps most) grains that are carrying an antecedent De unrelated to the final burial event (see Bateman *et al*., 2003 and 2007). This suggests that the sampled section (Plate 7) was in the primary sand scree, where the OSL signal was not fully reset before burial. Calculated OSL ages of 52.0 ± 3.3 ka (with the minimum De giving an age of 39.07 ± 2.99 ka) and 63.1 ± 3.8 ka (with the minimum De giving an age of 39.56

± 3.62 ka) are anomalous with respect to the age of the moraine, or terracing on the moraine. Note that both these minimum ages are probably still over-estimates of the burial date.

Bilbrough 1 (Fig. 37)

To establish an age for a high-level phase of Lake Humber (Stage 2), on the York Moraine, a sand sample was obtained from a roadcut northeast of Bilbrough (Plate 8). The sand was believed to have been washed from near the crest of the moraine and deposited below the 40 m terrace of Lake Humber.

Shfd12067 (Fig. 43 (e))

The sample gave an anomalous age of 6.07 ± 0.46 ka with respect to its location. The sample probably came from sand excavated from the roadcut that resulted in the OSL signal being largely reset for most of the sand grains.

Mill Mound 1, 2 and 3 (Fig. 37)

These three samples were obtained by auger drilling the sand scree or reworked sand at three sites on the southern slope of the York Moraine below Mill Mound. Mill Mound 1 specifically targeted sand near the crest of the moraine below the 30 m formline (Plate 20). As this sample was taken at a depth of 0.55 m, due to poor penetration of the auger drill, in possibly disturbed ground, dating of the sample could prove unreliable. Both Mill Mound 2 and Mill Mound 3 were taken between the 20 m and 25 m terraces (Plate 23b): the former at a depth of 0.86 m and the latter at a depth of 1.76 m. Mill Mound 2 was expected to test sand reworked from the 25 m terrace on the sand scree, while Mill Mound 3, at a greater depth, was thought might penetrate the primary sand scree (Fig. 27).

Shfd12071 (Fig. 43 (f))

As anticipated this sample from disturbed ground, at the 30 m above OD level, gave an anomalous OSL age of 2.10 ± 0.16 ka.

Shfd12068 and Shfd13034 (Fig. 43 (c) and (d))

The near unimodal distribution of the De values, with only a few aliquots as outliers giving a slightly positively skewed distribution, suggests good bleaching with the OSL burial signal reset for most of the sand grains. It therefore seems likely that the samples, representing a sand thickness of 1.76 m (depth for Shfd13034), were reworked from the earlier deposited sand scree. The two dates of 15.9 ± 0.9 ka and 15.21 ± 0.68

ka for this level of Lake Humber are suitably younger than an OSL age of 16.6 ± 1.2 ka for the 33 m terrace at Ferrybridge (Bateman *et al*., 2008).

North Cave 1 (Fig. 40)

A sand sample, at an elevation of about 1.0 m above OD, was taken from a sequence of sand and flint gravel exposed in the wall of a gravel pit in the North Cave Wetlands close to the pit shown in Plate 33. It was previously proposed (sections 11.2.1. and 7.12.1.) that these deposits originated from the Older Alluvial Fans and were transported to the wetlands by a proto-Foulness River (Halkon, 1999, fig. 8.2). As such, it would be expected that the sample should date approximately to the MIS 7 interglacial.

Shfd12069 (Fig. 43 (q))

Although there is a fairly wide distribution of De values, the sample does display a normal distribution with a mean De of 89.62 Gy once outliers are removed. The outliers, shown on Fig. 43 (s), do not indicate any possible antecedent De or any incomplete bleaching. An age for the sample of 58.4 ± 3.3 ka is however not compatible with the perceived origin of the gravels by run-off from the Older Alluvial Fans, during approximately the MIS 7 interglacial. As with Shfd13037, the OSL signal for the sample could have been partially re-set during final fluvial activity across the plain of the Vale of York once Lake Humber had declined to below 5.0 m above OD (Stage 8).

Manor Farm 1 (Fig. 40)

To obtain a more reliable sample from the 40 m terrace, near Houghton Moor, that would exclude any reworking, as might have occurred for samples Shfd11112 and Shfd11113, an auger hole sample was obtained at a depth of 0.55 m from the level land surface of the terrace near Manor Farm southwest of Mask Hall. It was hoped that this sample would give an age for the 40 m terrace of Lake Humber (Stage 2).

Shfd12070 (Fig. 43 (k))

The sample displayed a fairly tight unimodal distribution, with only two or three outliers, giving a mean De value of 2.84 Gy. An OSL age of 1.77 ± 0.11 ka is highly anomalous for the location and suggests that the sand could have been re-activated by aeolian processes during the Late Holocene.

Newton upon Derwent 1 (Fig. 39)

Because the final stages of Lake Humber were marked by the deposition of laminated clays, finding a satisfactory date for this event has proved elusive. It was thought that the problem could be solved by dating a sample from a 10 cm sand lens near the top of the Thorganby Clay Member of the Hemingbrough Formation, in a clay pit, at Newton upon Derwent (Plate 30).

Shfd12097 (Fig. 43 (r))

As the sample was taken from a thin 10 cm sand lens, within laminated clay, and therefore irradiated from two different sources an age was obtained for the sample using dose rates from the sand and clay members. The date obtained was 4.37 ± 0.35 ka and is much younger than expected. There is also no indication on the De plot, which shows a fairly widespread range of De values, of an older De signal, or in fact whether the De signal has been reset.

Mask Hall 1 (Fig. 40)

An important objective of the OSL sampling program was to establish a date for the age of the Older Alluvial Fans. The location selected for an auger hole sample was from a dominantly sandy part of the alluvial fans, mapped south of the Market Weighton Spillway, near Mask Hall, at an elevation of 45 m above OD (i.e. above any known Lake Humber shorelines).

Shfd13035 (Fig. 43 (p))

The auger hole terminated at a depth of 0.6 m in greenish possibly lacustrine clay. The sample, taken above this depth, gave a fairly wide scatter of low De values with an age of 3.32 ± 0.20 ka that is incompatible with an origin from the Older Alluvial Fans. It is likely, at this site, that the OSL signal has been modified by exhumation caused by burrowing animals (see Bateman *et al*., 2007).

Pollington 1 (Fig. 38)

The sample was taken from a sand lens in a roughly bedded littoral deposit banked over and against a mound of Sherwood Sandstone (Plate 28), below the 15 Metre Surface. It was expected to provide an age for the deposition of the 15 m terrace (Stage 6) in the decline of Lake Humber.

Shfd13036 (Fig. 43 (s))

This sample displays well bleached characteristics as it has a unimodal and tight distribution of De values. This result reflects the depositional model for sand worked into a beach deposit with associated gravel. The OSL age of 15.5 ± 0.8 ka is in the same range as dates for samples Shfd12068 and Shfd13034 taken from below the 25 m terrace on Mill Mound, suggesting no major time gap between the 25 m and 15 m above OD lake levels.

Pollington 2 (Fig. 38)

This sample was from a sequence of cross-bedded, rippled and horizontally bedded sands overlying 15 cm of gravel containing Permian limestone erratics ('east Pennine suite'), referred to by Gaunt (1976b) as the 'Younger Pennine Glacial Sand and Gravel' (Plate 39). The sample, from below 5.0 m above OD, was taken to date a glaciofluvial event believed to represent a likely pre-Devensian deglaciation.

Shfd13037 (Fig. 43 (t))

In contrast to Shfd13036, this sample displays a wide scatter of De values, possibly due to incomplete bleaching, with a positively skewed unimodal distribution. A date of 47.6 ± 3.0 ka, probably a reliable maximum age for burial of the sample, was obtained from normally distributed De values once outliers were removed. This date does not disprove an original possible pre-Devensian burial age for the 'east Pennine suite' of gravels.

9.10 Conclusions

The OSL sampling program gives only an incomplete chronology for distinct glacial, lacustrine and fluvial episodes prior to or since the LGM, but does provided a framework for the inclusion of additional data that is compatible with the Pleistocene history of the Vale of York, based on previous results and the landform mapping. Hopefully the framework can be used in future to improve the existing record. New data should come from: further sampling of the Older Alluvial Fans not reworked by Lake Humber shorelines; single grain OSL dating of the York Moraine sand scree, to obtain the youngest possible date for this event; and single grain dating of the re-distributed gravel deposits at North Cave and Pollington to determine whether these gravels resulted from a pre-Devensian glaciation. Most of the anomalous dates, reflecting failure of the sampling strategy, are attributable to previous human activity, burrowing animals and re-activation by aeolian processes. The only anomalous result that cannot be explained by re-working in this way is the sample from a sand lens in the Thorganby Clay Member at Newton on Derwent. A summary of the recently derived chronology is included in Table 4.

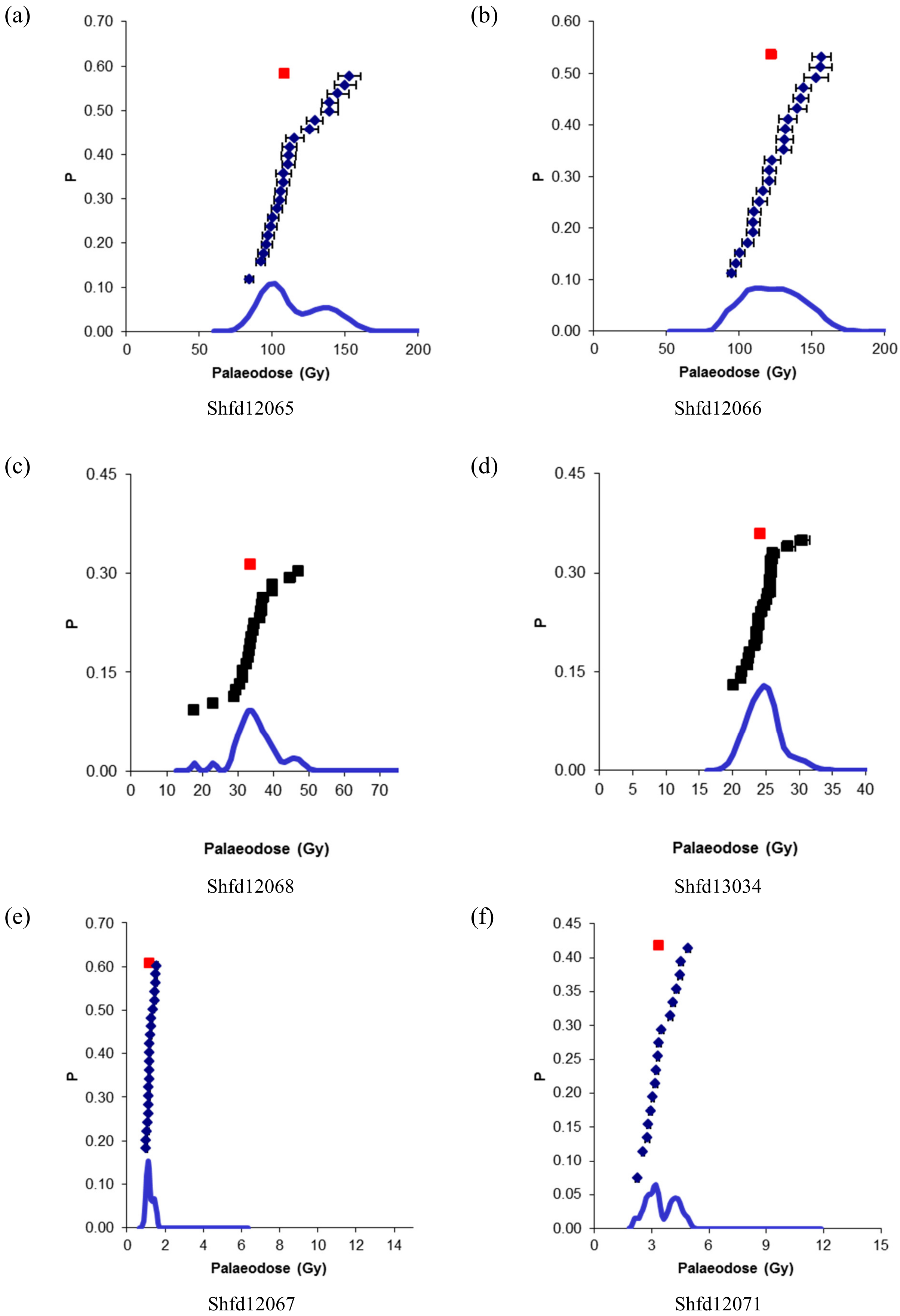


Figure 43. De probability distribution diagrams for OSL samples listed in Table 3. The red square is the mean De value.

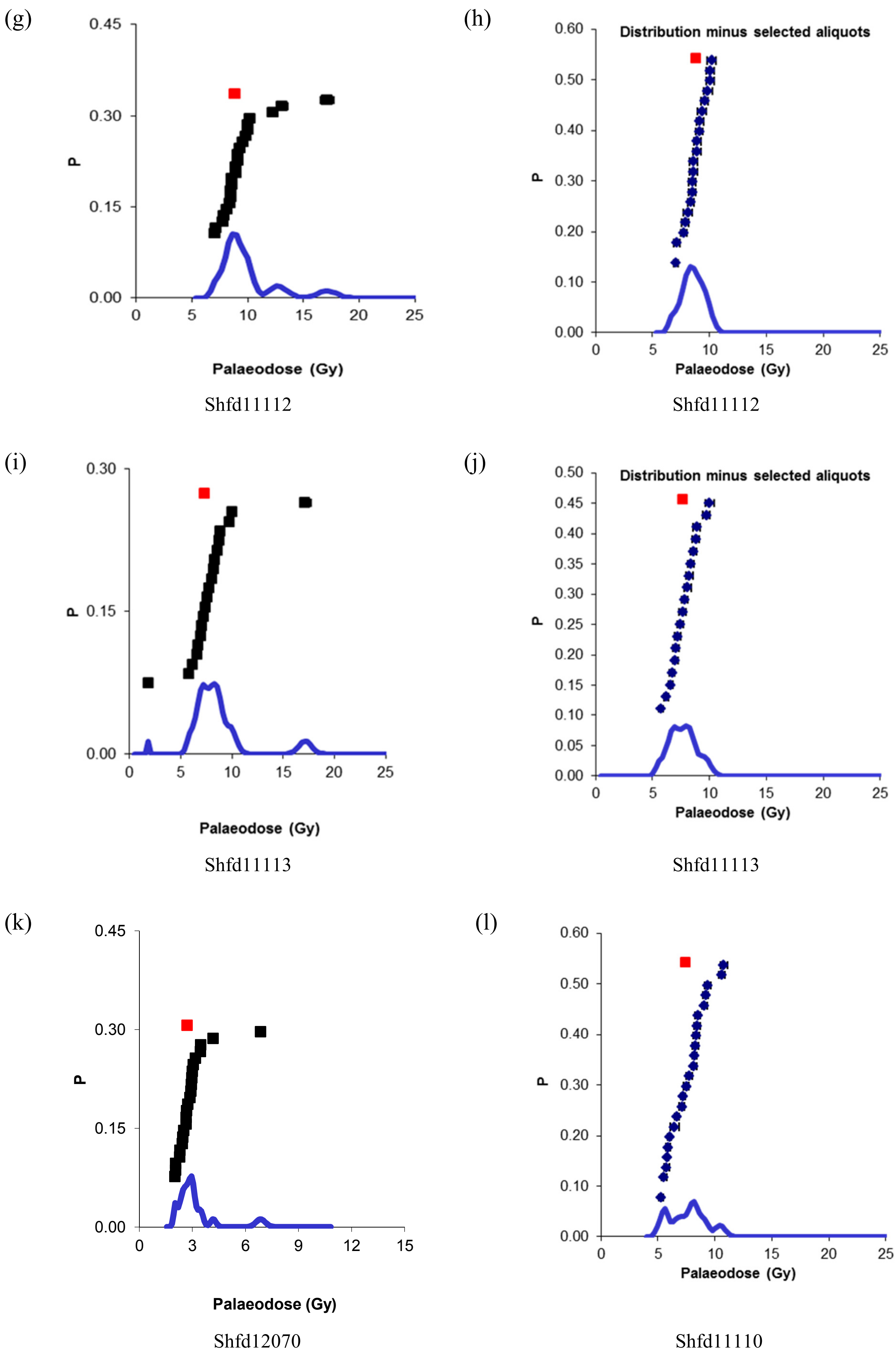
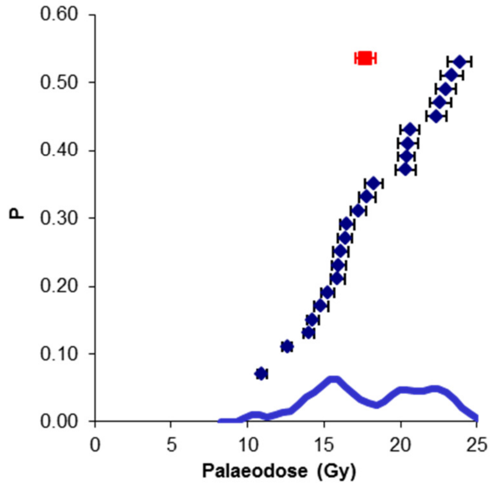


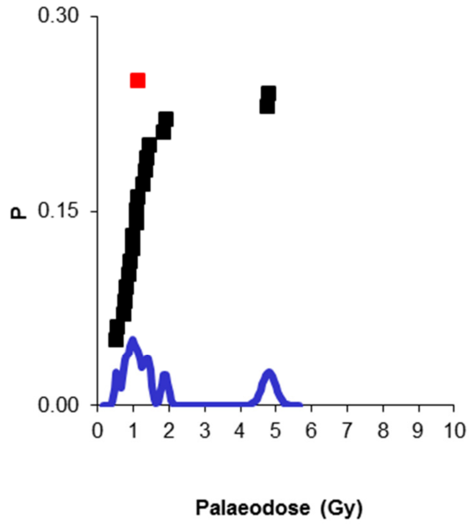
Figure 43. Continued

(m)



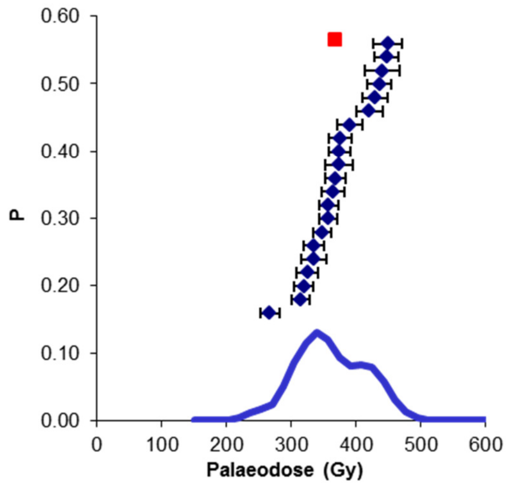
Shfd11111

(n)



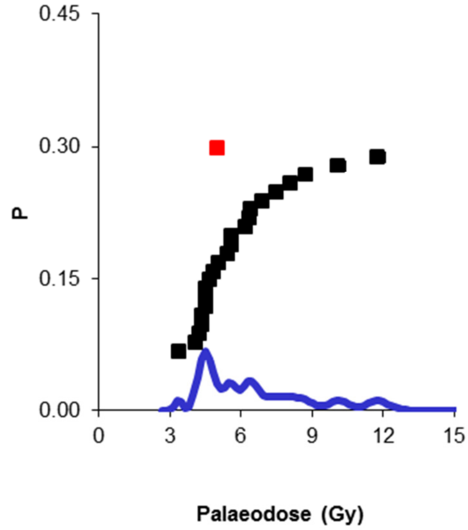
Shfd11114

(o)



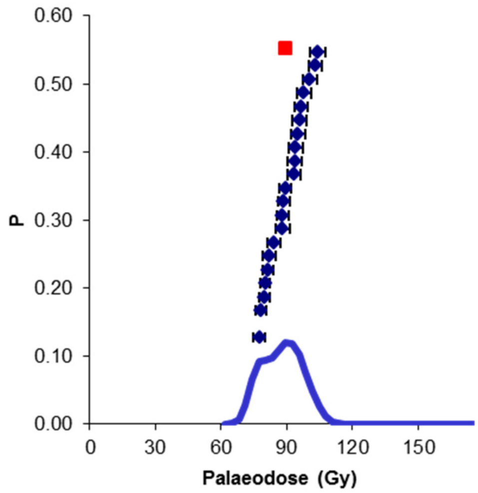
Shfd11115

(p)



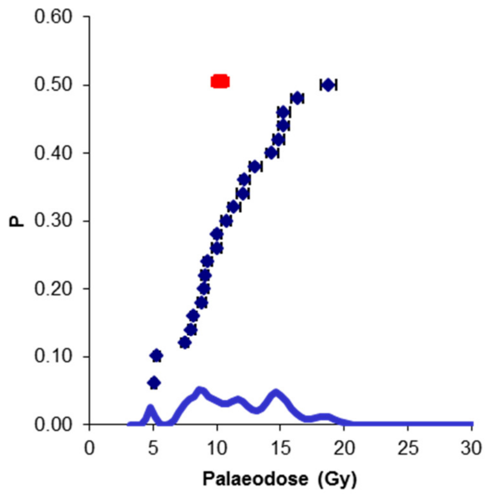
Shfd13035

(q)



Shfd12069

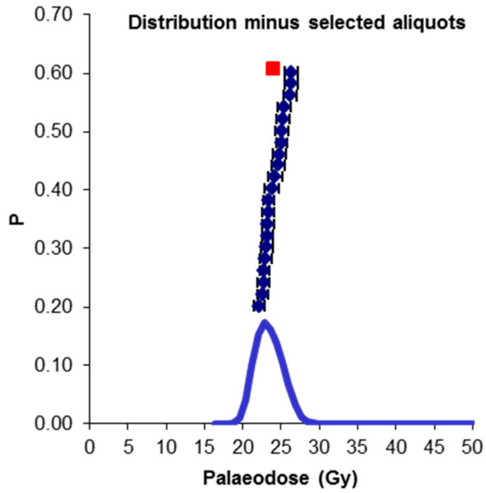
(r)



Shfd12097

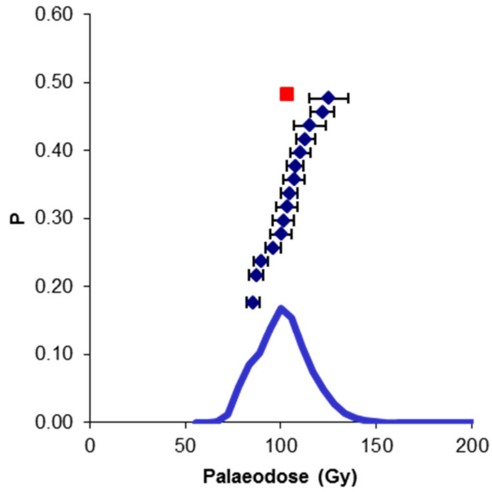
Figure 43. Continued

(s)



Shfd13036

(t)



Shfd13037

Figure 43. Continued

PART 3

Part 3 of this text deals with a re-evaluation, based mainly on implications derived from the landform mapping, of several depositional sequences, related to Vale of York geology and geomorphology, that may range in age from mid-Pleistocene to Early Holocene. These deposits include both fluvial and lacustrine sequences and the enigmatic Thorne-Wroot and Linton-Stutton gravels. Understanding the depositional origin and age of these deposits, while not an integral part of this research, is nevertheless important in any appreciation of the physiographic events that have shaped the Vale of York during the last two glaciations. To exclude them from this research could result in overlooking some significant conclusions that have been derived from the mapping including isostasy, an early low-stage Lake Humber and the possibility of an MIS 8 glaciation.

10. A REVISED STRATIGRAPHIC INTERPRETATION OF LATE DEVENSIAN / EARLY HOLOCENE – LACUSTRINE AND FLUVIAL SEQUENCES

A revised stratigraphic interpretation and provenance for the Hemingbrough Glaciolacustrine Formation has been made possible based on the conclusion that Lake Humber formed after retreat of the Vale of York glacier. This has allowed a re-interpretation to be made for the origin of the fluvial Hemingbrough Formation compared with that of Murton *et al*. (2009). Mapping by the British Geological Survey (2008) has also recognised that sands once thought to be a component of the '25 Foot Drifts' are now a discrete formation, the Breighton Sand Formation, overlying the '25 Foot Drifts'. A similar conclusion is contained in section 10.2. along with the belief that the sand formation is incised into the clays of the '25 Foot Drifts'.

10.1 Hemingbrough Glaciolacustrine Formation

The Hemingbrough Glaciolacustrine Formation was formerly named the Hemingbrough Formation by Thomas (1999) from a type section (in a clay pit?) north of Hemingbrough [SE 675 516] east of Selby. Previously the sediments were included in the 25 Foot Drifts of the Vale of York by Edwards *et al*. (1950) and Gaunt (1976a). Although laterally persistent, in much of the Vale of York, the formation everywhere lies below either the 10 m or the 5 m terrace so its detailed stratigraphy is mainly known from quarry sections and boreholes drilled to the underlying Carboniferous Coal Measures (Section 1, British Geological Survey, 2008; hereafter referred to as BGS

Section 1). As described by Ford *et al*. (2008), the Hemingbrough Formation south of the Escrick Moraine, between the Ouse and Derwent rivers (Domaine B, Fig. 45), consists of up to 24 m of laminated clay and silt split by sand beds into three members resting on Triassic bedrock or basal glaciofluvial deposits (BGS Section 1). Between the Escrick and York moraines, the sequence was deeply eroded by ice that deposited till of the Vale of York Formation. Because the till represents a substantial unconformity in the Hemingbrough Formation, it might well have been prudent to subdivide the formation into two parts i.e. pre-glacial and post-glacial. But as BGS Section 1 was presented, the till could be included as part of the Hemingbrough Formation.

Despite what appears to be a fairly definitive description, by Ford *et al*. (2008), of the three-fold sub-division of the Hemingbrough Formation, there are many contentious issues involving the Hemingbrough Formation. These include the depositional extent of the individual members and their chronological status. The source of the sediment, from which the formation is composed, is also speculative, as some parts, particularly the sands, could have been sourced either as sandurs from the Escrick Moraine, from meltwater channels eroded through the moraines or from lateral erosion along the northern flanks of the moraines (notably the Escrick Moraine). A definition of the Hemingbrough Formation also poses a problem with the Crockey Hill gravel and its precise context in the post-glacial sediments south of the York Moraine. As shown on BGS Section 1, these gravels have been deeply eroded into the till of the Vale of York Formation, a feature that does not preclude their origin either as an esker or as a coarse-grained delta (fan delta) associated with an erosional channel through the York Moraine (see sections 7.8.4. and 7.8.5.).

10.1.1. Park Farm Clay Member

The basal unit of the Hemingbrough Formation, the Park Farm Clay Member, was defined by Ford *et al*. (2008) from a clay pit at Park Farm [SE 622 406] 2.0 km north of Riccall. Northwards from Park Farm, this Member (mainly below OD) underlies till of the Vale of York Formation (BGS Section 1) and extends into the Harrogate district (Cooper & Burgess, 1993). In the region of York, highly contorted and deeply eroded beds of sandy laminated clay, overlain by till (Fig. 44) were exposed in sections during excavations for the York Railway Station (Fox-Strangeways, 1884, from Clark, 1881).

These laminated clays, which overlie fine sand, are probably equivalents of the Park Farm Clay Member.

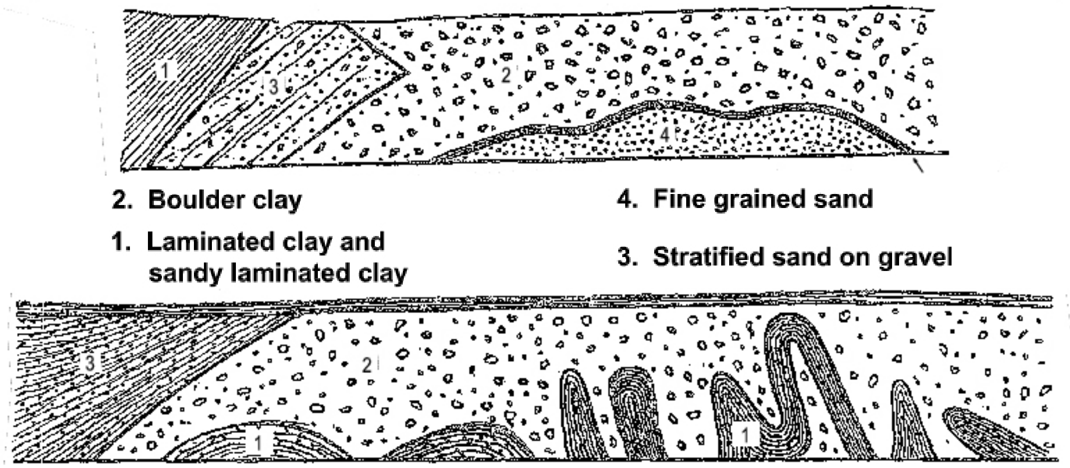


Figure 44. Highly contorted beds of sandy laminated clay overlain by till that were exposed in excavations during construction of the York Railway Station. From Fox-Strangeways, 1884, after Clark, 1881 (no scale shown).

The Park Farm Clay Member was also described by Murton *et al*. (2009) from a 10.1 m section in the clay pit at Hemingbrough [SE 674 317] east of Selby. Here, basal laminated clayey silts with massive to faintly stratified clayey silts (5.6 m thick) are overlain by a 3.3 m massive silt unit containing sandy interbeds. This section is capped by 0.7 m – 1.2 m of laminated to massive clayey silt (Murton *et al*., 2009). It is however apparent, at Hemingbrough, that the two lower units considered by Murton *et al*. (2009) to form part of the Park Farm Clay Member are quite distinct, as they are separated by a planar surface marking an aquiclude between the highly permeable silt unit above and the impermeable laminated clays below (Plates 37a & 37b); a fact clearly visible after heavy rain.

37a

Plates 37a and 37b. The Hemingbrough clay pit showing the tripartite subdivision of the sequence. A distinct planar surface separates the laminated clays below from the massive silts above. The upper bed of laminated clay (Thorganby Clay Member) can be seen around the rim of the quarry. (Photographs W.A Fairburn)

37b

The Hemingbrough section, as illustrated on Plates 37a and 37b, could therefore be re-interpreted as a tripartite sequence with only the lower unit representing the Park Farm Clay Member.

10.1.2. The Lawns House Farm Sand Member

The Lawns House Farm Sand Member overlying the Park Farm Clay Member (Ford *et al*., 2008, plate 1) is considered by Ford *et al*. (2008) to form a lobe extending south of the Escrick Moraine for up to 15 km (i.e. as far as Hemingbrough), thinning gradually from a maximum of 2.5 m. As shown on BGS Section 1, the Lawns House Farm Sand Member does not extend north of the Escrick Moraine and appears to be banked-up against the base of the moraine; a feature regarded by Ford *et al*. (2008) as evidence that ice depositing the Escrick Moraine was pushed into the laminated deposits. Such a feature could also have originated as a sand scree from which the Lawns House Farm Sand Member was derived.

At the Hemingbrough clay pit, Murton *et al*. (2009) placed their middle silty unit into the Park Farm Clay Member, as there are no thick sand layers present comparable with the Lawns House Farm Sand Member. However they did regard several prominent wave-rippled sandy beds as possible distal equivalents of the Lawns House Farm Sand Member. This generalised conclusion is acceptable in terms of a northern limit for the Lawns House Farm Sand Member at the Escrick Moraine, a significant factor for the source of the sand, but does not consider any of the possible erosional events associated with the Escrick Moraine discussed in this text. Erosional events that could have supplied sediment include:

- meltwater draining from the southern face of the Escrick Moraine between Stillingfleet and Wheldrake (Fig. 14), to produce scree-slope fans, as described in section 7.16.1.1. for the York Moraine;
- early breakthrough of the Escrick Moraine to provide meltwater in line with the present Ouse and Derwent valleys;
- lateral erosion channels along the north face of the Escrick Moraine, in to both the early Ouse and Derwent drainage systems during partial retreat of ice from the northern edge of the moraine.

10.1.3. Thorganby Clay Member

The Thorganby Clay Member overlies the Lawns House Farm Sand Member with a gradational or sharp contact (Ford *et al*., 2008, plate 1). The subcrop pattern of the member, on the southern and central parts of the Selby 1: 50 000 geological sheet (British Geological Survey, 2008), shown on Figure 45, where the unit can be up to 4.0 m thick, is largely controlled by its erosional contact with the overlying Bielby and Skipwith Sand Members of the Breighton Sand Formation (Ford *et al*., 2008, fig. 7; Fig. 46). This erosional contact is illustrated most clearly by the Skipwith Sand Member on the east bank of the River Derwent (Fig. 46) being adjacent to the Thorganby Clay Member on the west bank (Fig. 45).

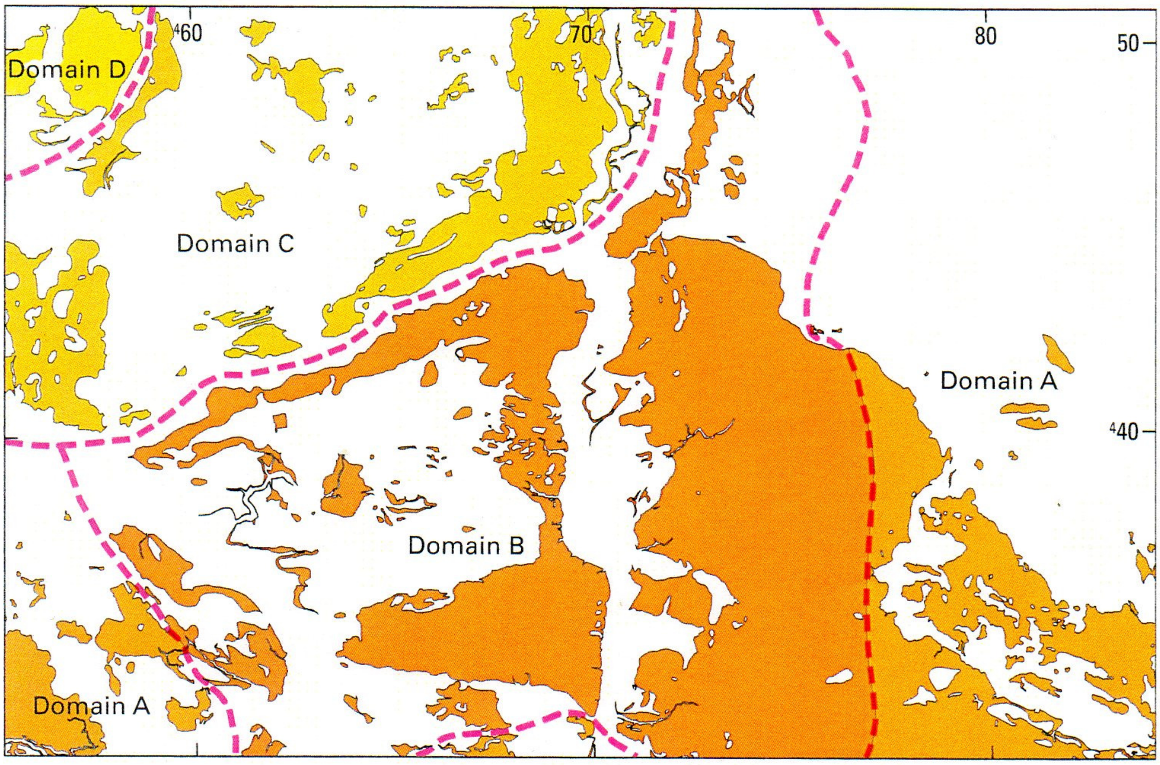


Figure 45. The distribution of laminated clays on the 1: 50 000 geological map of Selby, from British Geological Survey, 2008: Domain A = undivided Hemingbrough Glaciolacustrine Formation; Domain B = Thorganby Clay Member; Domain C = Elvington Glaciolacustrine Formation; Domain D = Alne Glaciolacustrine Formation. River alluvium and the Breighton Sand Formation, not shown on this illustration are included on Figure 46. The Escrick Moraine lies along the boundary between Domains B and C.

Elsewhere in the Vale of York, laminated clays equivalent to or partly equivalent to the Thorganby Clay Member include the Elvington Glaciolacustrine Formation, exposed between the southwestern margin of the Naburn Sand Member and the Escrick Moraine (Figs 45 and 46) and the Alne Glaciolacustrine Formation which underlies Hob Moor (Figs 14 and 45).

Although the site of the Hemingbrough clay pit is located on the Goole 1: 63 360 geological sheet (British Geological Survey, 1971) in an area mapped as silt and clay of the 25 Foot Drifts of the Vale of York, it does lie close to the Selby 1: 50 000 sheet (British Geological Survey, 2008) where the 25 Foot Drifts are now included as part of the Hemingbrough Formation. As the formation in this region is capped by laminated clay of the Thorganby Clay Member, it is highly likely that the laminated to massive clayey silt mapped by Murton *et al*. (2009) at the top of the section in the Hemingbrough pit, is equivalent to the Thorganby Clay Member.

10.1.4. Summary

The sediments exposed in the Hemingbrough pit (Murton *et al*., 2009) span a sequence of events, in the Vale of York, that extends from pre-glacial fluvial deposits (i.e. prior to advance of the Vale of York glacier) to a post-glacial flooding surface represented by the Breighton Sand Formation. Understanding the depositional environment of all the deposits in this sequence and creating a chronology of the events would resolve many of the issues related to MIS 2 glaciation in the Vale of York.

The Basal Glaciofluvial Deposits of Ford *et al*. (2008) shown on BGS Section 1, below the Park Farm Clay member, are concealed deposits, up to 2.0 m thick (Ford *et al*., 2008), that mainly lie below 10 m below OD. Sands equivalent to these deposits were possibly exposed below laminated clay in the Vale of York Railway Station sections (Fig. 44). The age of these sediments has yet to be confirmed.

The Basal Glaciofluvial Deposits are overlain by the Park Farm Clay Member, which underlies the Vale of York till. This member, which extends to at least 5.0 m below OD in the Hemingbrough clay pit (Murton *et al*., 2009), appears to have been deposited as a transgressive sequence into a restricted, shallow lacustrine environment formed by rising but fluctuating lake levels. Evidence of emergence could account for the variability of the sediments, as described by Murton *et al*. (2009). The Park Farm Clay Member, which has not been dated, may have been confined to wide mature valleys, cut down to nearly 20 m below OD, that converge towards the Humber Gap (Gaunt, 1994, fig. 42; Gaunt, 1981, fig. 4). Containment of lacustrine conditions in these valleys could well have been provided by an early MIS 2 glacial blockage of the Humber mouth at the time of GRIP episode GS-2a but not to the extent of creating a younger Lake Humber.

The middle silt unit in the Hemingbrough clay pit (Murton *et al*., 2009) overlies the Park Farm Clay Member with a sharp non-erosional contact (Plates 37a and 37b) and is considered, in this text, to be equivalent to the Lawns House Farm Sand Member beneath the Thorganby Clay Member. This silt unit is characterised by shallow fluctuations of water level, as indicated by abundant wave ripples (Murton *et al*., 2009). In the Hemingbrough sequence, three optically stimulated luminescence (OSL) ages, obtained from prominent sand interbeds, gave a weighted-mean age of 22.2 ± 0.5 ka BP (Murton *et al*., 2009). If these Hemingbrough dates are valid, then several firm conclusions can be reached regarding the status of the sampled sandy interbeds (inferred to be part of the Lawns House Farm Sand Member) both in terms of stratigraphy and chronology. These are:

- as they underlie the Thorganby Clay Member, which is a component of a sequence of laminated clays that filled up the region to about 7 m to 9 m above OD from a low-level lake (Gaunt, 1981 & Gaunt *et al*., 2006), they do not represent low-level Lake Humber, as stated by Murton *et al*. (2009), and must belong to an earlier event;
- the 22.2 ± 0.5 ka BP date does also not relate to low-level Lake Humber, as it is much older than high-level Lake Humber, dated by Bateman *et al*. (2008) at 16.6 ± 1.2 ka BP and older than the Skipsea Till, dated by Hartmann (2011) at *c*. 17.01 ± 1.33 ka BP, which has been terraced by high-level Lake Humber near Hessle (see section 7.4.2.), a discrepancy that has been explained by Murton *et al*. (2009) as being due to a previously unrecognised secondary high-level phase of Lake Humber;
- if the dated sandy interbeds were sourced from the Escrick Moraine then their OSL dating provides an age for the Escrick Moraine (i.e. *c*. 22 ka BP).

10.2 Breighton Sand Formation

The Breighton Sand Formation is widespread in the Vale of York, although it was not clearly recognised as a discrete mappable unit until after publication of the Selby 1: 50 000 geological sheet (British Geological Survey, 2008). Because it has an erosional contact with the underlying laminated clays of the Vale of York, such as the Thorganby Clay Member, and the tills of the Vale of York Formation (see Sections 1 and 2 in British Geological Survey, 2008) it was originally but incorrectly described as 'Sand of the 25-Foot Drift of the Vale of York' (Edwards *et al*., 1950; British Geological Survey, 1971). The Breighton Sand Formation is therefore younger than the final stages of lacustrine sedimentation in Lake Humber but older than a stage of the 5 metre terrace, which forms an erosional surface across the formation north-northwest of Selby (Figs 17 and 46). Based on the geographical distribution of the formation into discrete locations, which have been attributed to sourcing from separate fluvial systems (Ford *et al*., 2008), the formation has been subdivided into three members: the Bielby Sand Member, the Naburn Sand Member and the Skipwith Sand Member (Fig. 46).

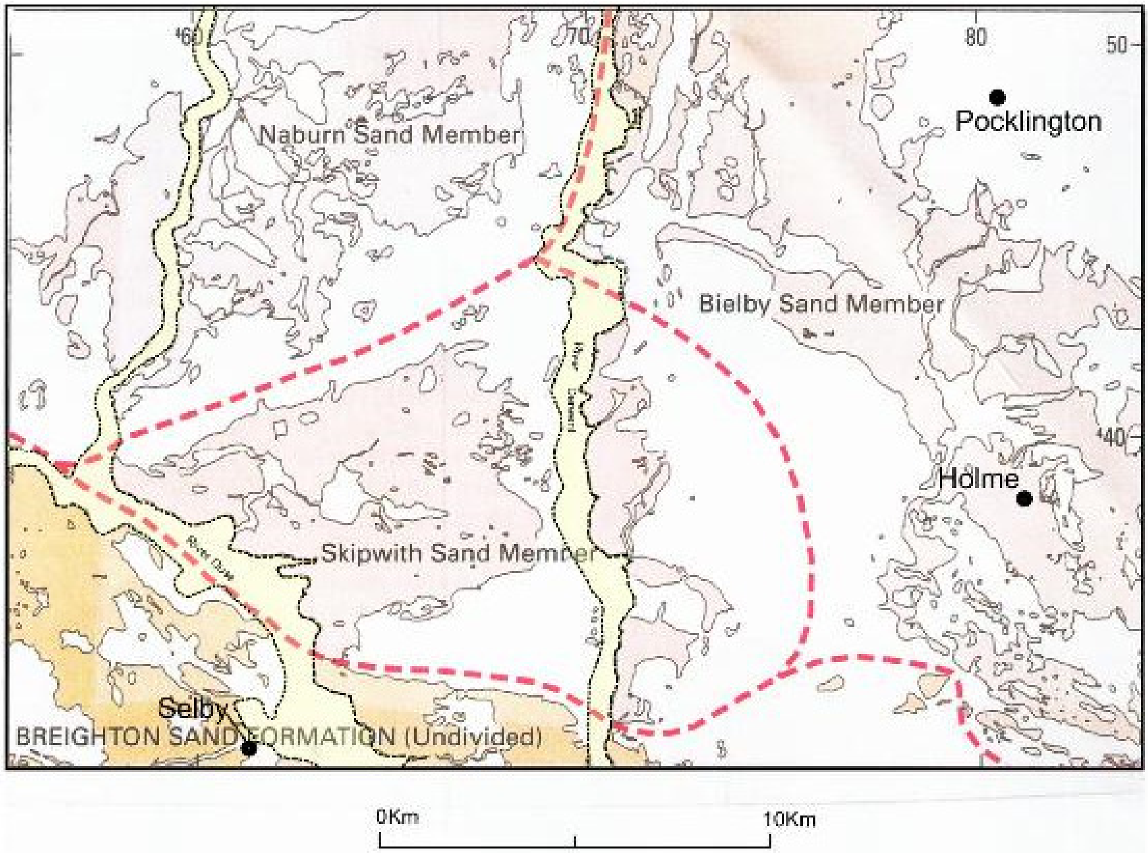


Figure 46. The distribution of the Breighton Sand Formation and its component members in the Selby district. From Ford *et al*., 2008

The Breighton Sand Formation must mark a return of strong fluvial activity in the Vale of York associated with Holocene warming as the Naburn Sand Member is dated by OSL to the early Holocene (section 9.9.). However it predates the incision of modern drainage south of the York Moraine. Such unconfined drainage resulted in a flooding surface across the Vale of York with deposits sourced from channels through and around the Vale of York moraines that were cut by meltwater drainage prior to and during the impounding of Lake Humber. The location of some of these channels were noted by Gaunt (1976b) and Palmer (1966) where they form linear patterns of ridges and discontinuous mounds. Subsequent to their deposition the sediments were probably modified and redistributed by aeolian processes (Ford *et al*., 2008) to produce the 'blow-away sand' and dunes observed by Parsons (1887).

10.2.1. Bielby Sand Member

The Bielby Sand member (Fig. 46), as shown on the Selby 1: 50 000 geological sheet and accompanying Section 2 (British Geological Survey, 2008), appears to occupy a unique, shallow, southeasterly trending erosional valley sloping southwards at about 0.62 m/km between the Pocklington Gravel Formation (alluvial fans) to the east, and the Escrick Moraine and Thorganby Clay Member to the west. The unconformable relationship between the member and the underlying Thorganby Clay Member and the Pocklington Gravel Formation is clearly illustrated in Section 2 (British Geological Survey, 2008).

Although the landform mapping, shown on Figure 15, has not recognised any topographic features associated with the Bielby Sand member, parts of the member may well be embedded into the 20 Metre Surface in the region of Bolton and Barmby Moor. While evidence for this, in the field, is only provided by some scattered conspicuous patches of yellow sand near Bolton and along parts of Spittal Beck, there has to be some degree of overlap between mapping of the Bielby Sand Member, shown on the Selby geological sheet (British Geological Survey, 2008), and the 20 Metre Surface shown on Figure 15.

The Bielby Sand member appears to have originated from sediment supplied to a proto-channel of the River Derwent, eroded east of the Escrick Moraine, with some minor additions from channels originating in the Wolds, such as renewed drainage through the Spittal Beck/Bishop Wilton Beck system. The main problem arising from such a conclusion is that landform mapping has not revealed a defined channel through the 20 Metre Surface, from the Derwent Valley, west of Full Sutton (see Fig. 15).

10.2.2. Skipwith Sand Member

In contrast to the Bielby Sand Member, the Skipwith Sand member originated from a proto-channel of the River Derwent eroded west of the Escrick Moraine that discharged into this part of the Vale of York through the gap in the moraine east of Wheldrake. The outcrop (or subcrop) of the member is mainly confined to the east bank of the River Derwent valleys (Fig. 46), where the unconformity, at the base of the member, has extended down into the Park Farm Clay member east of Riccall (British Geological Survey, 2008). As with the Bielby Sand Member, the base of the Skipwith Sand Member dips gently southwards, similar to modern drainage, as indicated by its extensive lateral boundary against alluvium along the River Derwent.

10.2.3. Naburn Sand Member

As mapped on the Selby geological sheet (British Geological Survey, 2008) this member is confined to an area between the York and Escrick Moraines that extends from the west bank of the River Ouse to the west bank of the River Derwent, where in places it is adjacent to the Bielby Sand Member on the east bank (Fig. 46). The sand mainly overlies either till of the Vale of York Formation or laminated clays of the Elvington Glaciolacustrine Formation, but in places it also rests unconformably on the Crockey Hill gravels and eroded segments of the Escrick Moraine (Section 1, British Geological Survey, 2008; Fig. 31). Unlike the other members of the Breighton Sand Formation, the Naburn Sand Member has been extensively exposed in a number of excavations, notably in the sewerage pipeline trench north of Naburn (Fig. 37), a ditch along the disused railway line near Naburn Wood (Figs 25 and 37), in the re-excavated Dike Wood north of Naburn Wood (Fig. 37) and in the water pipeline trench near Wheldrake (Figs 31 and 37). The unconformable contact with the underlying clays of the Elvington Glaciolacustrine Formation is best illustrated in Figure 31. Dating of sand samples from a 1.0 m section in the Naburn Sand Member in the Naburn Sewerage trench, by optically stimulated luminescence (OSL), gave mean dates of 4.32 ± 0.28 ka and 9.13 ± 0.61 ka. These dates (discussed in section 9.9) do suggest an early Holocene age for the Breighton Sand Formation probably with some younger aeolian reworking.

Widespread wind-blown sand has been mapped northeast of York by Matthews (1970) associated with organic horizons near Sutton on the Forrest. Radiocarbon dating of this material with ages of 10.7 ± 0.19 ka BP and 9.95 ± 0.18 ka BP does however suggest that these sands are older than the aeolian component of the Naburn Sand Member, and therefore represent an older aeolian event.

11. MAMMALIAN DEPOSITS – CHRONOLOGICAL IMPLICATIONS

Since the mid-Pleistocene Anglian glaciation (MIS 12) the glacial record has been punctuated by six interglacials, which are now defined by their oxygen isotope ratios (Fig. 3). In an earlier classification of Quaternary deposits by Shotton (*in* Mitchell *et al*., 1973) temperate interglacial stages were defined on the basis of their vegetational history while cold glacial stages were defined on lithological criteria. The interglacial stages can also be characterised by distinctive mammalian remains and indications of human habitation, which not only assist in defining the stratigraphy but also provide important knowledge on past climates. Within Yorkshire two sets of mammalian deposits have now been recognised: one set belonging to the Ipswichian (MIS 5e) and the other to MIS 7. Besides these two dated intervals, some minor less well-documented mammalian remains have also been discovered in gravel associated with depositional events that must post-date the MIS 2 glaciation in the Vale of York. These minor occurrences (located within the mapped area of Fig. 14), which must originate from unknown Ipswichian deposits, as they are indicative of an interglacial with remains of *Hippopotamus*, include 'bones of the great mammalia' from the Crockey Hill gravels (Kendall & Wroot, 1924), the metatarsal of *Ursus spelaeus* (Grizzly Bear) discovered in coarse gravels near the Bishopthorpe road (Clark, 1881) and deposits near Overton that yielded 'Mammoth, Hippopotamus etc.' (Clark 1881).

11.1. Ipswichian

The best documented of the mammalian remains are from the Ipswichian interglacial (MIS 5e) given a date of 128 ka – 117 ka BP by Catt (2007). Of these, the most important because of its relevance to this text, is the Ipswichian mammalian deposit on the raised beach at Sewerby (Lamplugh, 1891; Catt & Penny, 1966; Boylan, 1967). An Ipswichian age of 120.8 ± 11.8 ka for the raised beach was confirmed from thermoluminescence (TL) dating of blown sand above the raised beach, by Bateman & Catt (1996).

Mammalian deposits, possibly equivalent to those on the Sewerby beach, have been described by Reid (1885) from chalk pits west of Hessle Railway Station. Here the Hessle Gravel, containing the bones of *Elephas primigenius* and *Rhinoceros* (Reid, 1885), lie on chalk and chalk rubble covered by till on a planar surface, at 20ft (6.1 m) above OD (Melmore, 1935). The till, at this location, is the Skipsea Till (see Catt, 2007, fig.1; also referred to as the Hessle Boulder Clay by de Boer *et al*., 1958), as the Basement Till has not been recorded in this region.

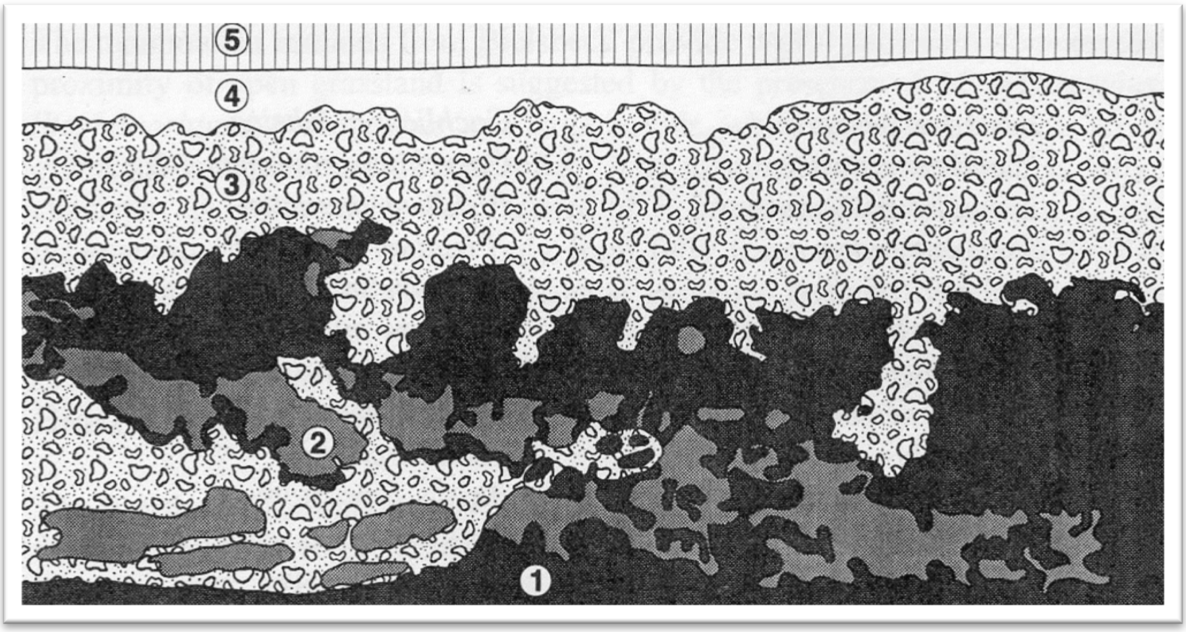
11.2. Marine Isotope Stage 7

While the age of an older glacial period appears uncertain, a possible age may be inferred from the MIS 7 mammalian deposits at Bielsbeck Farm and from deposits of a possibly similar age from Elloughton.

It is also worth noting that the raised beach at Easington, situated some 33 m above OD on the coast of County Durham (Davies & Bridgland, 2013), has been dated to MIS 7 with a mean OSL age of 201 ± 28 ka (Davies *et al*., 2009).

11.2.1. Bielsbeck Farm

The Bielsbeck Farm mammalian deposit (Fig. 16) was discovered in 1829 during extraction of marl from a pit some 12.5 ft deep, believed by Kendall & Wroot (1924) to lie in a hollow of a former land surface underlain by 'Keuper Marl'. The first detailed account of the deposit by Harcourt (1829) described a section some 5.95 m thick composed of basal marl overlain along an irregular erosional contact by chalk/flint gravel and topped by sand (Fig. 47).



5.	**Black sand**	**0.23m**	**VALE OF YORK SEQUENCE**
4.	**Yellow sand**	**0.45m**	
3.	**White gravel consisting of small pebbles of chalk and angular fragments of flint with *Gryphaea Incurva* and sandstone pebbles**	**0.75m**	
2.	**Blue marl containing bones**	**1.5m**	**MERCIA MUDSTONE GROUP**
1.	**Blacker marl containing bones: land marsh and freshwater Mollusca, *Coleoptera* and plant remains**	**>3.0m**	

Figure 47. Section at Bielsbeck Farm from Schreve, 1999, after Harcourt, 1829

In listing the fauna, contained in the deposit, de Boer *et al*. (1958) comment that the bones in most cases were perfectly preserved with some found 'in original relative positions'. The de Boer *et al*. (1958) description also contains an incompletely referenced quote from Phillips (1875) 'that the bones in the black marl were inhumed at only a very short distance from the place where the animal died'. In a more recent account of the fauna Schreve (1999) lists eleven species and also comments that the bones were not transported far from where they were originally deposited. Schreve (1999) has concluded that the mammalian assemblage is fully interglacial, with the presence of *Palaeoloxodon antiquus* (straight-tusked elephant) and a late morphotype of *Mammuthus primigenius* (woolly mammoth) now referred to as *Mammuthus trogontherii* (D. Schreve, *pers. comm.*) diagnostic of MIS 7, possibly substage 7a.

At a further site near Galley Moor [SE 841 398], about 3.2 km northwest of Bielsbeck, remains of *Mammuthus primigenius* were found in two adjacent pits (Halkon, 1999). Radiocarbon dates from one of the mammoth teeth and some associated organic material, reported by Halkon (1999), were 47 ka and 46 ka ± 2.3 ka BP respectively. These dates must however be considered unreliable as they are close to the maximum limit for radiocarbon dating.

The relative abundance of these mammalian deposits, occurring along a northwest – southeast trend, has prompted Halkon (1999) to suggest that the deposits occur along the projected route of a pre-Devensian proto-Foulness River eroded into 'Keuper Marl' (Halkon, 1999, fig. 8.2). If this is the case then the scattered deposit may form only a part of a linear mammalian graveyard.

In terms of the landform mapping, the Bielsbeck Farm deposit lies under the 10 Metre Surface (Fig. 16). The valley eroded during or before the MIS 7 interglacial was later infilled by fluvial and lacustrine events extending into the Late Devensian with gravel derived from the Older Alluvial Fans including the Market Weighton Spillway (de Boer *et al*., 1958).

Another factor of significance, suggested by preservation of the Bielsbeck Farm and other mammalian sites, is that the fragile environment in which they occur, survived MIS 2 flooding, but could probably not have survived being over-ridden by any post-MIS 7 ice sheet, such as an MIS 6 ice advance into the Fen Basin (Westaway *et al*., 2012) if this also affected the Vale of York. This suggests that any mid-Pleistocene (?Saalian) glaciation of the area would have to be older than MIS 7. An MIS 8 Saalian glaciation of Lincolnshire has been proposed by Westaway (2010), White *et al*. (2010) and Straw (2011).

11.2.2. Mill Hill - Elloughton

Mill Hill [SE 942 278], near Elloughton, which rises to just over 33 m above OD is capped by sand and gravel composed mainly of chalk and flint pebbles (Gaunt *et al*., 1992). A section through the gravel layers (Lamplugh, 1887) shown in Figure 48, indicates two sets of gravels separated by an erosional unconformity (blue line in Fig. 48).

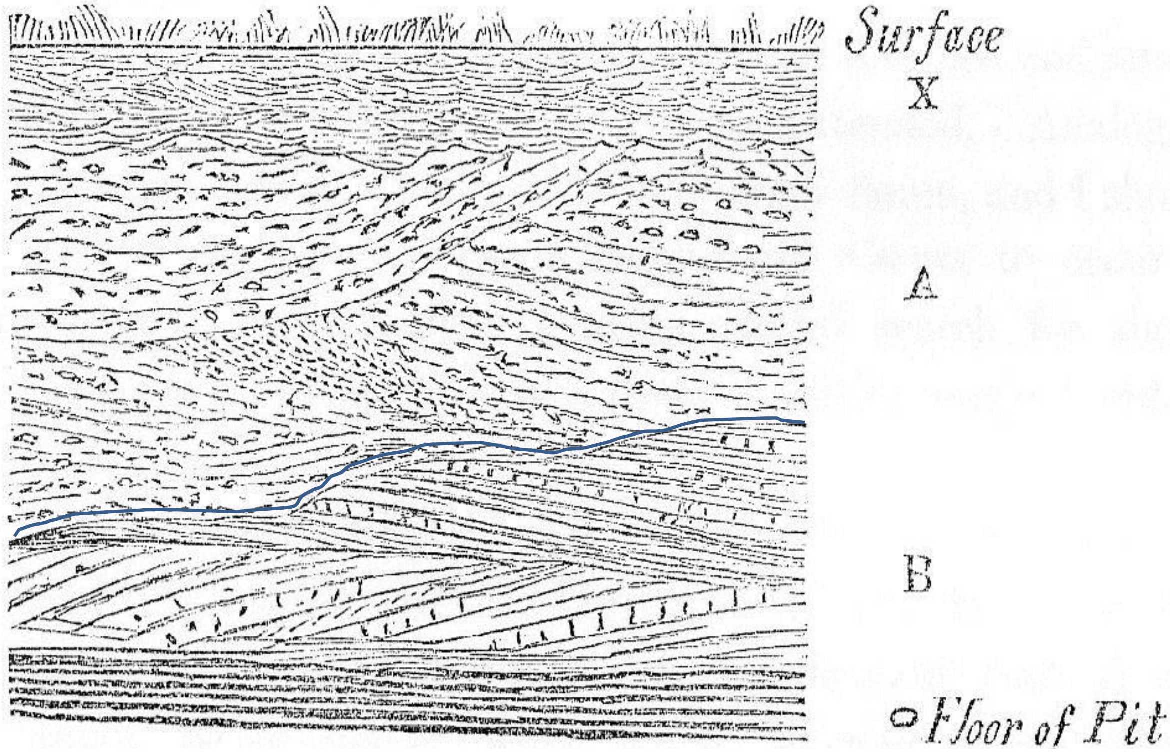


Figure 48. Section through the Mill Hill gravels at Elloughton from Lamplugh, 1887.

X	Top soil – 0.76 m
A	Rough stony gravel with some sand – 2.7 m
B	Yellow sand with stony layers – 1.5 m
O	Floor of pit – Jurassic Ancholme clay

A list of mammalian species recognised from the Mill Hill site, and reported by Sheppard (1897) and de Boer (1958), includes both *Elephas (Mammuthus) primigenius* and *Elephas (Palaeoloxodon) antiquus*. It is not known, however, whether more than a dozen teeth attributed to *Mammuthus primigenius* could be referred to *Mammathus trogontherii*, identified from the Bielsbeck site by Schreve (1999), which is diagnostic of MIS 7, due to loss of these remains in the Hull museum.

It is clear that the two sets of gravels, identified by Lamplugh (1887), had differing sources and can be attributed to two separate events. The lower gravels, with exotic boulders of Whin Sill and Millstone Grit, could only have been sourced from a Pennine glaciation, while the upper gravels, which contain flint, chalk and oolitic limestone could have been sourced from the Older Alluvial Fans mapped on Figure 16. Both sets of gravels, however, would have to be older than MIS 2 (see section 7.12.1.).

11.3. Brantingham Bone Fragment

In 1971, a bone fragment was found southeast of Brantingham [SE 935 292] during construction of the Elloughton by-pass (Gaunt, 1974). This fragment yielded a radiocarbon age of 21.84 ± 1.66 ka (Gaunt, 1974) calibrated to 26.16 ± 2.05 ka BP

(Murton *et al*., 2009). It was presumed by Gaunt (1976) that the bone fragment was contained within the 'Older Littoral Sand and Gravel' thus providing a maximum age for high-level Lake Humber. The landform mapping (Fig. 16) does however suggest that the bone fragment, excavated from an elevation between 25 m and 30 m above OD, was probably re-worked from part of the Older Alluvial Fan sequence along a shoreline of Lake Humber at *c*. 33 m above OD.

12. RE-APPRAISAL OF THE EAST PENNINE SUITE OF GRAVELS

A number of gravel deposits that may have a primary origin before the Devensian have been modified or reworked by lacustrine or fluvial processes, in the late Devensian to produce many of the land forms mapped within or bordering Lake Humber. These include the 'Younger Pennine Glacial Sand and Gravel' (Gaunt, 1976b), the 'Thorne - Wroot Gravels' (Straw, 2002), and the 'Linton-Stutton Kame-belt' (Edwards *et al*., 1950). These deposits have a distinctive pebble composition consisting mainly of Carboniferous sandstone and Permian limestone, but with some Carboniferous limestone, chert and rocks from the Lake District (Gaunt, 1976b, Gaunt *et al***.**, 2006). The proportion of Permian limestone pebbles, which distinguishes the gravels from the other deposits, is commonly as high as 20% – 40% and in some deposits is slightly more than 50% (Gaunt, 1976b, Straw, 2002). This assemblage of pebbles, called the 'east Pennine suite' is also contained in the 'Pennine Boulder Clay' (Gaunt, 1976b).

12.1. Younger Pennine Glacial Sand and Gravel

During field mapping, to compile Figure 17, gravels attributed to the 'Younger Pennine Glacial Sand and Gravel' were extensively recognised below the 10 metre surface, overlying an (in places cryoturbated) erosional surface of Sherwood Sandstone (Plate 38). At all locations, the gravels, which are seldom over 0.25 m thick are invariably covered by up to 1.0 m of cross- bedded red sand derived by erosion from nearby outcropping Sherwood Sandstone. For this reason the gravel is often seen in shallow temporary sections e.g., a pipeline trench near Burn [SE588 280] and quarry sections at Hensall (Plate 38) and Pollington (Plate 39), towards the eastern end of the Snaith Ridge.

Plate 38. A thin band of gravel (10 cm – 20 cm), the 'Younger Pennine Glacial Sand and Gravel',(or the 'east Pennine suite' of gravels, Gaunt, 1976b), exposed above an erosional surface on Sherwood Sandstone in a quarry east of Hensall [SE 598 235]. The gravel is overlain by cross-bedded red sand below the 10 Metre Surface. (Photograph W.A. Fairburn)

It was noticeable in both the Pollington and Hensall sand pits (Plates 38 and 39) that the gravels were restricted to below the 10m surface (as already stated) and were not present below quartz gravels forming the 15 metre surface over 'highs' of Sherwood Sandstone (Plates 28 and 29). In a section at Great Heck Station (Fig. 23) the 'Younger Pennine Glacial sand and Gravel', overlying cryoturbated Sherwood Sandstone was mapped by Parsons (1887) as 'pan sand row of pebbles'.

Plate 39. A 3.5 m section [SE 614 200] in the Pollington sand pit composed of sand and gravel overlying cross-bedded Sherwood Sandstone. The base of the section (near the scale) is formed by 5.0 cm – 15 cm of gravel containing Permian limestone erratics referred to by Gaunt (1976b) as the 'Younger Pennine Glacial Sand and Gravel', or the 'east Pennine suite', overlain by cross-bedded, rippled and horizontally bedded sand at the OSL sample point. The section is capped by a southerly dipping bedform of sandy gravel below the 10 Metre Surface. (Photograph M.D. Bateman)

The distribution of the gravels described above, implies a major fluvioglacial flooding event to account for the widespread occurrence of such coarse deposits. As the gravels probably originated by erosion of 'Pennine Boulder Clay' or the associated 'Older Glacial Sand and Gravel' (section 3.3., Fig. 7), it is probable that the 'Younger Pennine Glacial Sand and Gravel' could also be of pre-Devensian age. While some of the deposits were re-worked into the 15m shoreline, near Brayton Barff (Parsons, 1887; Melmore,1934), other areas of gravel, below 10m above OD to the east of the Burn and the Snaith Ridge were little modified by Lake Humber.

12.2. Thorne-Wroot Gravels

The deposits of 'Younger Glacial Sand and Gravel' that form the discontinuous ridge that trends south-southeast from Thorne through Bradholme Hill, Tudworth Hall and Lindholme Hall to Wroot (Gaunt, 1976b; Fig. 49) are also known as the Thorne-Wroot Gravels (Straw, 2002) or the Tudworth Hall type deposits (Gaunt, *et al.*, 2006).

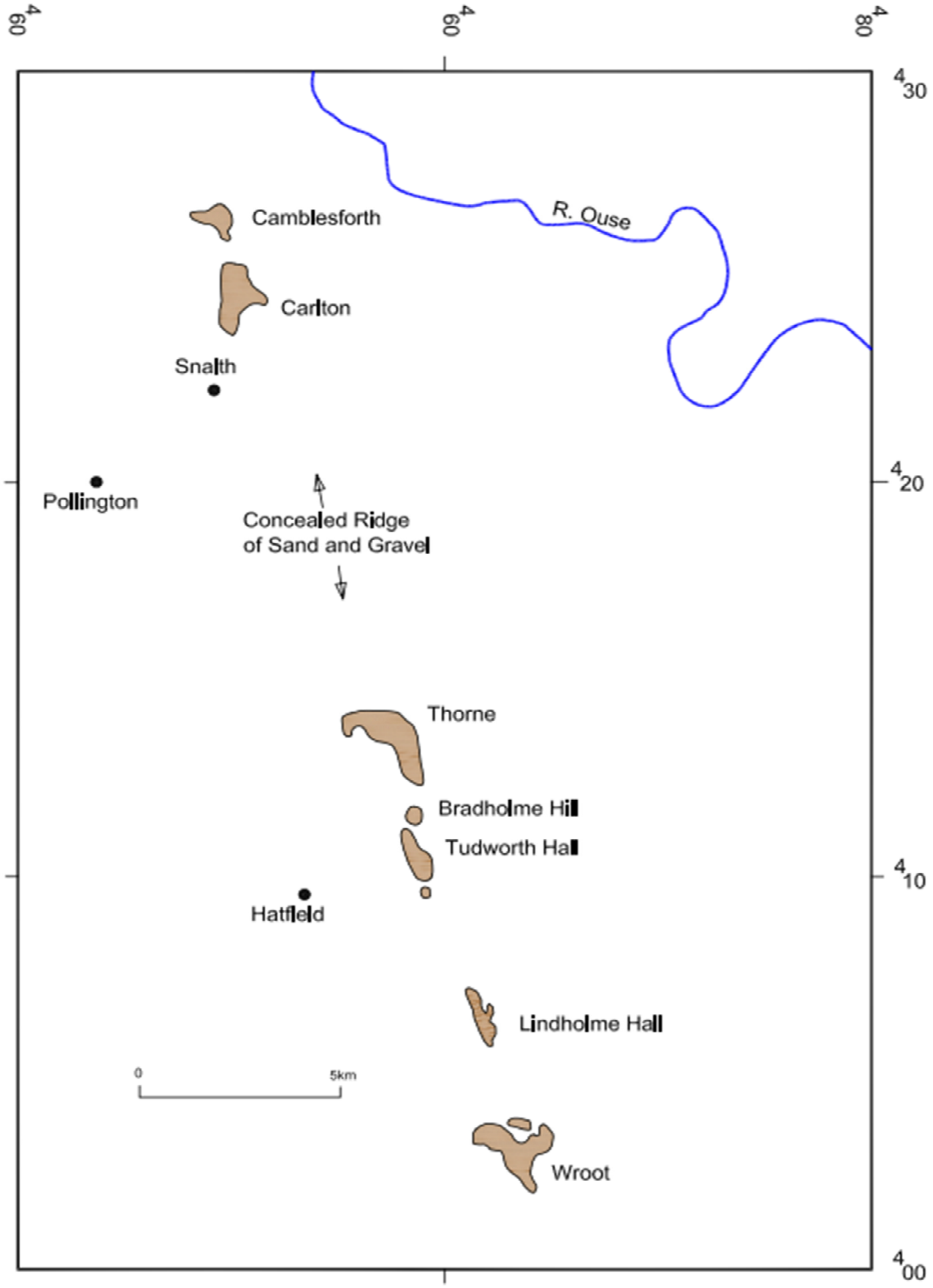


Figure 49. Location diagram of Thorne-Wroot Gravels south-southwest of Snaith (from Gaunt, 1976a). Note alignment with remnant 'islands' of the 10 Metre Surface at Carlton and Camblesforth.

12.2.1. Glacial Association

Gaunt (1976a) proposed that the Thorne-Wroot Gravels were formed along the western margin of a tongue of Lake Devensian ice, which had advanced into the south-eastern part of the Vale of York as far as Wroot south of the Escrick Moraine. Evidence supporting this ice incursion in the form of tills or deglacial features was not recorded during the landform mapping (see also Straw, 2002 and Parsons, 1887). A factor at variance with ice sourced through the Vale of York is the pebble composition of the gravels, which resembles assemblages in pre-Devensian tills (Gaunt, 1994) rather than the Teesdale assemblage of erratics that would be expected in the Late Devensian (Straw, 2002).

12.2.2. Physiography and Stratigraphy

As shown along the eastern edge of Figure 17, small eroded remnants or 'islands' of the lacustrine 10 Metre Surface, surrounded by the fluvial erosional terrace of the 5 Metre Surface, have been preserved around the villages of Carlton, Camblesforth, Barlow and Drax. At both Carlton and Camblesforth, the 10 Metre Surface is underlain by Sherwood Sandstones with perhaps a thin covering of quartz gravel. Similarly the 'Thorne-Wroot Gravels' located southeast of Snaith (Fig.49), lie on 'islands' of Sherwood Sandstone, below the 10 Metre Surface, surrounded by the erosional edge of the 5 Metre Surface. It is highly likely that the general level – bedding, with some southerly or south-easterly dipping cross-bedded units within these gravels, plus delicate ripple bedding indicating deposition in quiescent or gently flowing water, reported by Gaunt *et al*., (2006) and Straw (2002), resulted from a redistribution of sediment during formation of the 10 Metre Surface (*cf.* cross-bedded gravels below the 15 Metre Surface at Red Cliff, Plate 27). Similarly Straw (2011) has suggested that these gravel deposits lie on pedestals of Sherwood Sandstone.

12.2.3 Age

Gaunt (1976a, 1976b,1981 and 1994) has suggested that the Late Devensian glacial, fluvial and lacustrine deposits lie between a lower periglacial surface and an upper periglacial surface. In adhering to this rigid concept, Gaunt (1976a) drew a rather tenuous correlation between the Thorne-Wroot Gravels and the Older Littoral Sand and Gravel. Straw (2002) has pointed out that periglacial conditions affected Britain over many periods in the Early and Middle Devensian and that 'almost any deposit of this time could rest on and lie beneath a surface betraying periglacial activity'. This statement could include periglacial surfaces attributable to pre-Devensian cold stages. At Lindholme Hall (Plate 40) the

Thorne-Wroot Gravels could well have been re-arranged in the Late Devensian but the underlying cryoturbation (Plate 40) could well be contemporaneous with periglacial conditions in the Wolds that produced the Older Alluvial Fans during a pre-Devensian glaciation (see discussion section 7.12.3.). Indeed Gaunt (1981 and 1994) has drawn attention to sub-glacial channels, to the west of Lindholme Hall, to which Gaunt (1976b) previously assigned a pre-Devensian age.

Plate 40. Poorly stratified Thorne-Wroot Gravels overlying cryoturbated Sherwood Sandstne at Lindholme Hall. (Photograph W.A. Fairburn)

The similarity between the pebble lithology in the Thorne-Wroot Gravels with erratics from the Pennine Boulder Clay (the 'east Pennine suite', Gaunt, 1976b) supports a pre-Devensian age for these deposits rather than a Late Devensian age based on derivation from the Vale of York glacier. Re-working of these gravels below the 10 Metre Surface and in places below the 5 Metre Surface is supported by OSL dating of the laminated sands exposed in the Lindholme Hall 'island' pit (Friend, 2011, fig. 4.12). These dates of 32 $\pm$ 2.4 ka and 34 $\pm$ 2.2 ka were considered by Friend (2011) to represent the maximum age of sedimentation.

12.3 Linton – Stutton Gravels

The Linton-Stutton gravels form a linear deposit between Linton and Stutton that were referred to by Edwards *et al***.** (1950) as the Linton – Stutton Kame-belt. Mapping of the gravels by Edwards *et al*. (1950, plate III) has been included on the Leeds 1:50 000 geological sheet (British Geological Survey 2003). A map of the gravels (Fig. 50, from British Geological Survey, 2003) shows that the outcrops mainly lie on interfluves reflecting post-depositional valley erosion of a once more continuous outcrop. The linear nature of the deposit is marked, but as the lower edges of many of the outcrops terminate between 40m – 50m above OD, e.g. the Wingate Hill deposit (Fig. 51), then this linearity can be partly explained as a facet of elevation that could have resulted from shoreline erosion. Many of the deposits do however originate from areas of the Anglian 'Boulder Clay' at elevations over 70m above OD (British Geological Survey, 2003). Lithologically the gravels consist largely of Permian limestone (the largest boulders), Carboniferous sandstone and limestone with fewer pebbles of Whin Sill and Cheviot porphyrite (Edwards *et al.* 1950).

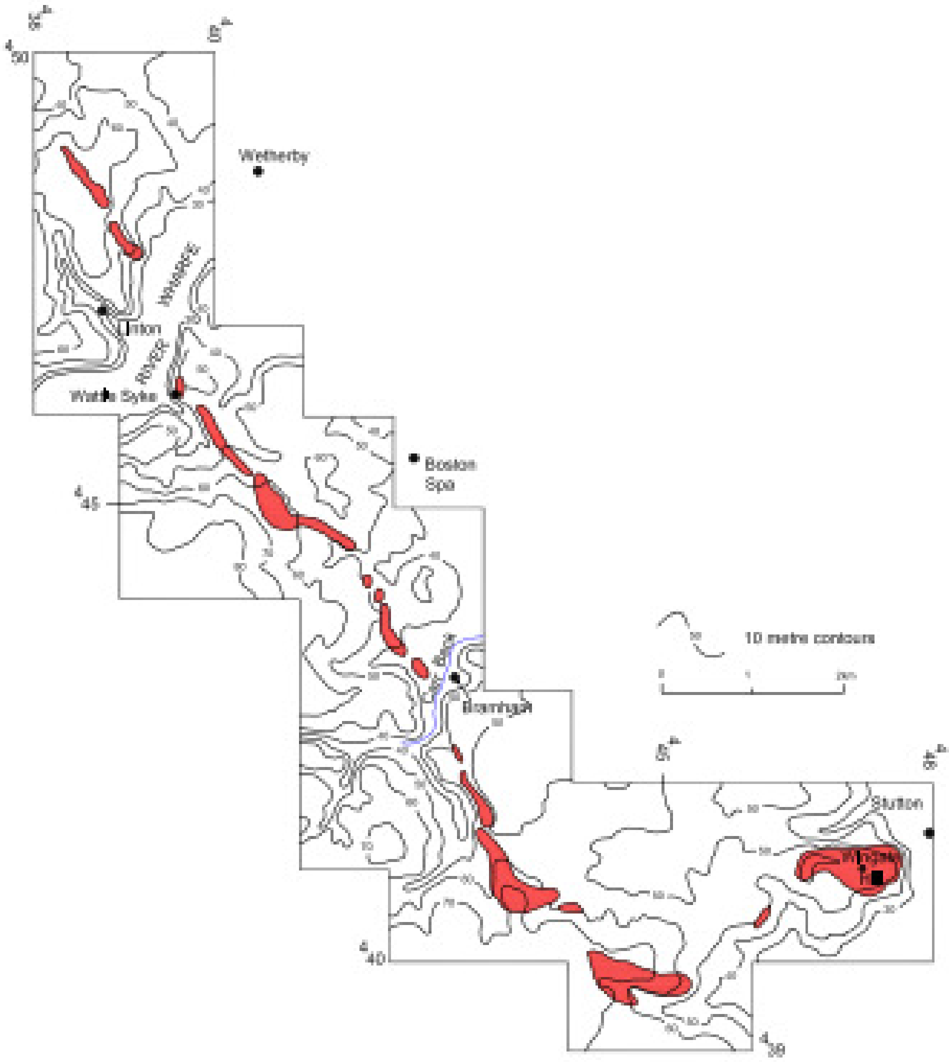


Figure 50. Location map of Linton-Stutton gravels. From Leeds 1: 50 000 Geological Map, British Geological Survey, 2003.

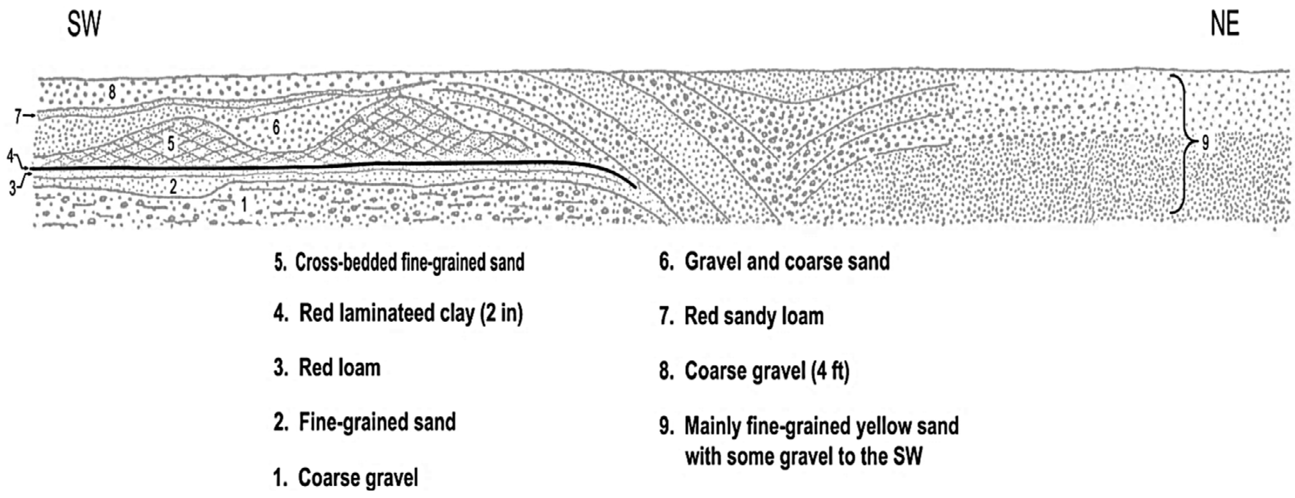


Figure 51. Section through the Linton – Stutton gravels at an elevation of 61.0 m at the top of Wingate Hill [SE 474 410]. From Melmore, 1934, fig. 5 no scale given.

It has generally been concluded that these gravels form a Kame-belt with steeper bedding dips to the east or northeast (Edwards *et al.*, 1950; Palmer, 1966 and Straw, 1979), or possibly a sub-marginal esker (Straw, 1979). The deposit, the outcrop of which appears to have erosional affinities with the associated mapped 'Boulder Clay' (Edwards *et al*., 1950, plate III) or Anglian Till (British Geological Survey, 2003) and has been dated by Edwards *et al.* (1950), to the Early Main Dales Stage of glaciation. It is consequently older than his York – Escrick Moraine-belt and the 100 foot strandline of Edwards (1937).

Based on the recognition of alluvial fans on the western face of the Wolds, probably associated with a pre-Devensian glacial episode, coupled with the termination elevations for the Linton-Stutton Gravels at between 40 m – 50 m above OD, it would seem feasible, that these gravels are the Pennine equivalents of the Older Alluvial Fans.

PART 4

13. SUMMARY

This summary is a catalogue of events that have shaped the Vale of York from a pre-Lake Humber periglacial episode to the imposition of a widespread flooding surface in the Early Holocene that preceded the final disappearance of Lake Humber and the incision of modern drainage. Some of these events are considered well proven, but others, such as the age of the earlier periglacial environment are conjectural. Nevertheless, this overview tries to explain and place into context many issues concerning deposits of uncertain origin that have been the subject of controversy for many years (see Table 4).

1. *The Pre-Devensian*

Evidence for pre-Devensian periglacial environments in the Vale of York, is only provided by the Older Alluvial Fans that were terraced during retreat stages of Lake Humber. These fans, which are underlain by laminated clay near Market Weighton, at 45 m above OD, are believed to have originated from permafrost shattered Chalk formation. The upper demarcation of the fans is marked by the 52 Metre Strandline that rises in elevation northwards probably due to isostatic adjustment. Dating of the alluvial fans is however restricted to a single OSL age of 202 ± 12 ka (Shfd11115) from the 33 m terrace at South Cave, which is indicative of MIS 7. This age, however, has not been substantiated from the Older Alluvial Fans elsewhere in the Wolds, nor in the gravel deposits of the North Cave Wetlands, which may have a similar original depositional age. In the Wolds, an OSL sample (Shfd13035), taken from a shallow depth in a sandy facies of the Older Alluvial Fans at Mask Hall near Market Weighton (Fig. 40), gave an anomalous age of 3.32 ± 0.20 ka that has resulted from surface reactivation. At North Cave, a sand sample (Shfd12069) taken at a depth of 4.0 m, in one of the gravel pits, provided a maximum age of 58.4 ± 3.3 ka, from an OSL signal that was probably re-set during the MIS 4 – MIS 2 glacial period. These gravels may well have been reworked from an older deposit.

Another set of gravels, that could be referred to the pre-Ipswichian, are the 'Younger Pennine Glacial Sand and Gravel' of Gaunt (1976b). These gravels, with a distinctive suite of pebbles (the 'east Pennine Suite'), composed of Permian limestone, Carboniferous sandstone and Lake District volcanics, frequently overlie cryoturbated

Sherwood Sandstone (Fig. 23, Plate 38) or the lower periglacial surface i.e. the base of the Devensian deposits (Gaunt, 1974). Because of this, the gravels are regarded as Late Devensian in age by Gaunt (1976b). As however the pebble content does have affinities with the pre-Devensian Pennine Boulder Clay (Gaunt, 1976b) a pre-Devensian age is more likely. A sand sample (Shfd13037), taken from immediately above these gravels at Pollington (Plate 39), gave an age of 47.6 ± 3.0 ka. This again reflects an OSL signal that was reset during the MIS 4 – MIS 2 glacial period.

2. The LGM (MIS 2)

While there is no new dating relevant for the advance of the British-Irish Ice Sheet into the Vale of York, an approximate age for this event can be derived from other data. Bateman *et al*. (2008) have given an OSL age after 23.3 ± 1.5 ka for the incursion of the Vale of York ice lobe based on dating loess below a diamicton at Ferrybridge. A similar age can also be inferred from the age of the Park Farm Clay Member of the Hemingbrough Formation, which must be older than 22.2 ± 0.5 ka; an age for sand overlying this unit at Hemingbrough (Murton *et al*., 2009). This member is overlain by the Vale of York till between the York and Escrick moraines (British Geological Survey, 2008).

Livingstone *et al*. (2012) have also proposed that the maximum expansion of the BIIS during the LGM occurred between *c*. 29 ka – 23 ka BP with ice draining eastwards through the Tyne and Stainmore gaps. In the later stages of this period Livingstone *et al*. (2012) have also suggested that there was migration of ice back towards uplands resulting in a cessation of the Stainmore ice flow, a switch that may have coincided with a shutdown of the Vale of York ice lobe and a weakening of the Tyne Valley ice stream. Additional constraints, on the onset of glaciation in the Yorkshire Dales to later than *c*. 27 ± 2.0 ka and deglaciation prior to *c*. 17 ± 2.0 ka, have been presented by Telfer *et al*. (2008) based on OSL dating.

While restraints for the onset of glaciation in the Vale of York do suggest a possible age of 23.0 ka BP, or younger, a date for deglaciation is even less precise. An indication for deglaciation can however be inferred from deglaciation data elsewhere in the UK, particularly if these can be related to the Vale of York. Bateman *et al*. (2011) has established an age for the Skipsea Till of between *c*. 21.7 ka BP and 16.2 ka (the age of sands between the Skipsea and Withernsea Tills). This age range is in agreement with a

TL date of 17.5 ± 1.6 ka for loess beneath the Skipsea Till (Wintle & Catt, 1985) and an OSL date of 17.01 ± 1.33 ka from a subglacial stream deposit in the Skipsea Till (Hartmann, 2011). This date of Hartmann (2011), not only provides an approximate upper age limit for the Skipsea Till, but also signifies deglaciation and a recession of the North Sea ice lobe, that allowed the accumulation of fluvial and lacustrine sediments in ice-marginal lakes extending from Dimlington to Upgang north of Whitby (Evans & Thomson, 2010; Bateman *et al*., 2011; Roberts *et al*.; 2013). This fluvial-lacustrine phase, dated at 16.2 ka (Bateman *et al*., 2011) was terminated by a further readvance of North Sea ice and the emplacement of the Withernsea Till (lithofacies association LFA 4, Evans & Thomson, 2010) dated by Bateman *et al*. (2011) to 16.2 ka – 15.5 ka. A realistic date for deglaciation in the Vale of York, prior to 17.01 ± 1.33 ka, would be in agreement with the date for deglaciation in the Yorkshire Dales proposed by Telfer *et al*. (2008).

3. Lake Humber

The main phase in the impounding of Lake Humber and the imprinting of shorelines could not have occurred until after retreat of the Vale of York glacier to north of the York Moraine and after blocking of the Humber mouth by the advance of the North Sea glacier that deposited the Skipsea Till. This latter event has been dated to *c*. 17.01 ± 1.33 ka by Hartmann (2011). Prior to this, there appears to have been an earlier minor blockage of the Humber, presumably caused by North Sea ice that left no recognisable glacial till. This earlier blockage resulted in shallow, but rising ponding in the Vale of York (pre-cursor, Table 4) initially restricted to channels that had been cut up to 20 m below OD (Gaunt, 1994) during earlier falling sea levels. The ponding resulted in the deposition of a transgressive sequence of shallow water silts and laminated clays forming the Park Farm Clay Member of the Hemingbrough Formation (Table 4). This member has been identified in the Hemingbrough clay pit, mainly below OD, by Murton *et al*. (2009), where it is exposed as an impervious sequence below massive silts (Plates 37a and 37b). It also lies beneath Vale of York till, as shown in Section 1, on the Selby geological sheet (British Geological Survey, 2008), again mainly below OD.

The main phase of Lake Humber, initiated by a more extensive and sustained blockage of the Humber mouth by the North Sea glacier that deposited the Skipsea Till resulted in the impounding of a pro-glacial lake that rose to a maximum elevation of 42 m above OD. This larger Lake Humber eventually declined progressively from the high stand of

42 m above OD down to a low stand of 5.0 m above OD. Mapping of shoreline terraces etched into the York and Escrick Moraines, the Older Alluvial Fans on the western face of the Wolds and into ridges of Triassic sandstone during the punctuated decline has been the major focus of this research. This has resulted in the formulation of a revised 8-stage regressive decline model for Lake Humber (Fig. 52) that replaces the older 2-stage high-stand, low-stand Lake Humber model that proved inadequate to explain the mapped geomorphology (Fig. 11). The model contains both transgressive and regressive shorelines affecting the moraines: the former only occurred when the stand of the Lake coincided with the crest of the moraine. To adequately illustrate the extent and attitude of the landforms, defined between the 8 stages in the new decline model for Lake Humber, an innovative mapping procedure was introduced, as explained on Figure 13 and shown on the geomorphological map of the York and Escrick Morines (Fig. 14). Stage 1 marks the maximum elevation of Lake Humber at 42 m and is mainly recognisable on the York Moraine at Bilbrough where a transgressive shoreline has created a planar surface across the crest of the moraine and left a lag of cobbles and boulders (Plate 6) with discrete mounds of sand and gravel (Fig. 14). Sand, segregated from the moraine at this level, has produced localised sand screes on the southern face of the York Moraine east of Bilbrough. A strandline developed by the 42 m lake level has been recognised between Market Weighton and Sancton (not depicted on Figure 16) and elsewhere on the Wolds southwards towards Hessle (note also a stand of about 45 m above OD for Lake Pickering (Straw, 1979).

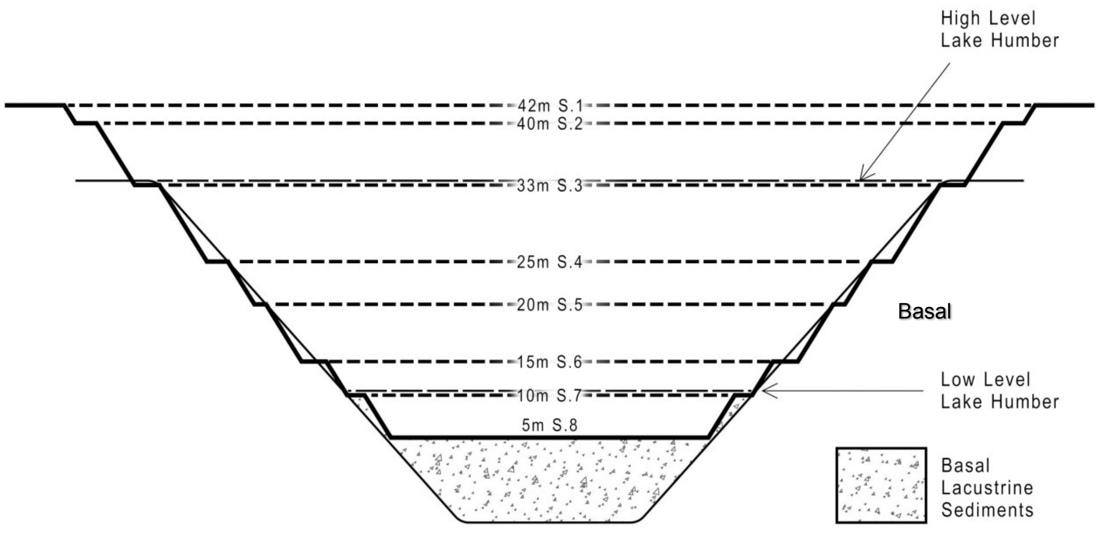


Figure 52. Schematic diagram to illustrate the proposed 8-stage regressive model for Lake Humber (bold line). The diagram includes the 2-stage high-level/low-level model for Lake Humber shown in Fig.11.

A stand at 40 m above OD (Stage 2), mainly identified on the York Moraine near Bilbrough (Plates 14 and 15) and along the Wolds was followed by the long-recognised stand at 33 m above OD (Stage 3) which was the former high level Lake Humber. This stand, which includes the 100 Foot Strandline of Edwards (1937), has particular relevance in both the chronology and physiography of the Vale of York as it has terraced the Skipsea Till near Hessle (and is therefore younger than the till) and formed a transgressive shoreline across the crest of the moraine at Mill Mound and Severs Howe. This latter feature washed sand from the moraine and deposited it down the southern slopes of the moraine to form a sand scree or sand apron (Figs 27 and 24). The 33 m terrace has been dated to 16.6 ± 1.2 ka (Bateman et al., 2008). Below the 33 m terrace less conspicuous stands of the lake have been recorded locally at 30 m and 25 m above OD (Stage 4): the 25 m stand has been dated by Shfd12068 and Shfd13034, from Mill Mound, at 15.9 ± 0.9 ka and 15.21 ± 0.68 ka respectively (Table 3). A fall in lake level to 20 m above OD (Stage 5) south of the York Moraine, and a widespread depositional stand at this level, some 5.0 m lower than to the north of the moraine, permitted deepening and widening of the York Gap and the discharge of a coarse-grained delta towards Crockey Hill. Sand samples taken from the landscaped back-wall of the 20 m terrace below Severs Howe (Shfd12065 and Shfd12066, Plate 7) and believed to be from the original sand scree on the York Moraine that has not been reworked by later Lake Humber shorelines, gave ages of 52.0 ± 3.2 ka and 63.1 ± 3.8 ka but from only partially bleached samples (Fig. 43(*a*) and (*b*)).

Whilst there has been no age established for the 20 m terrace, the lower 15 m (Stage 6) terrace has been dated by Shfd13036, from Pollington, at 15.5 ± 0.8 ka. The duration of the final level of Lake Humber at 5.0 m above OD (Stage 8) is difficult to determine due to lack of dated samples, but it could be expected that Lake Humber reached the 5.0 m level and finally emptied before 15.38 ka; the retreat date of the ice lobe that deposited the Withernsea Till in northeast Holderness. This date also marks the closure of episode GS – 2a in the Greenland oxygen isotope record (Fig. 4). There is however mapping evidence that the 5.0 m terrace, northwest of Selby, is underlain by an eroded surface composed of Thorganby Clay Member incised by the Breighton Sand Formation (British Geological Survey, 2008). This would imply that the 5.0 m terrace was locally revisited in the Holocene following a fall in lake level below OD, as suggested by Gaunt (1994).

As indicated in Table 4 the initial phase of Lake Humber from *c*. 23 ka to after 22.2 ± 0.5 ka was followed by the main phase that declined in at least 8 stages from a highest level of 42 m before 16.6 ± 1.2 ka and after 17.5 ± 1.7 ka to its lowest level of 5.0 m above OD after 15.5 ka.

4. Holocene

The present day southward inclination of the Vale of York is not the result of isostasy but the imprint of an extensive flooding surface that is underlain by the Breighton Sand Formation. This formation overlies and is incised into the lacustrine and laminated clays of the Vale of York. Approximate dating of this event to the Early Holocene is provided by OSL samples from the Naburn Sand Member (Shfd11110 and Shfd11111) that gave ages of 4.32 ± 0.28 ka and 9.13 ± 0.61 ka. (Table 3). These dates are however produced from bimodal distribution of the De data that may reflect a mixed fluvial and aeolian origin of the sand member. This conforms with a description of the sand member by Ford *et al*. (2008). A bimodal distribution of the sand is also indicated by the particle size analysis given in the Appendix. Flooding that reworked older sand deposits in the Houghton Moor region has been dated to the same period by samples Shfd1112 (7.51 ± 0.40 ka) and Shfd11113 (8.24 ± 0.49 ka).

14. CONCLUSIONS

The recognition and interpretation of terracing on the York Moraine and the western face of the Wolds, as shorelines of Lake Humber impounded following the LGM, has led to two major conclusions concerning the physiographic evolution of the Vale of York since a previous glaciation. These two conclusions, supported by photography, OSL dating and LIDAR imagery, can be used to explain much of the glacial history of the Vale of York since the proposed penultimate glaciation.

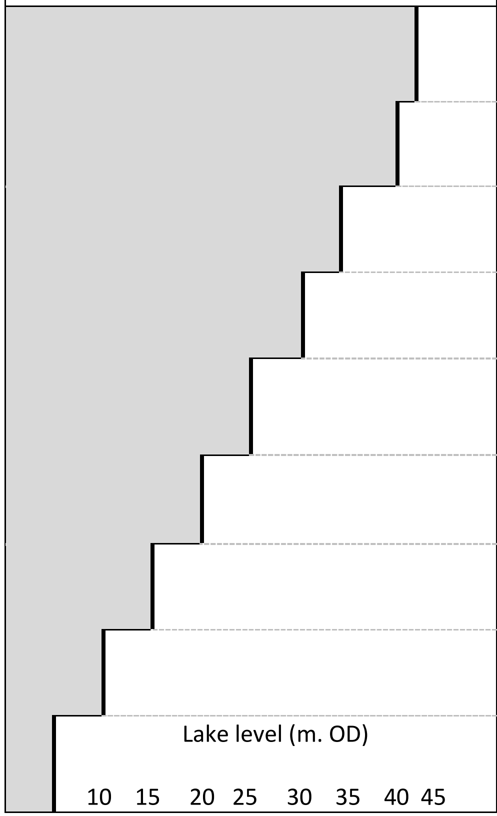
The first major conclusion is that the Vale of York glacier must have retreated to the north of the York Moraine prior to imprinting of shorelines formed by rising and/or falling levels of Lake Humber impounded by glacial blockage of the Humber Gap. Terracing has confirmed that this pro-glacial lake rose to a maximum level of 42 m above OD, before it spilled over the crest of the moraine, prior to a staged withdrawal of the lake from its Stage 1 high stand down to a 5.0 m above OD, Stage 8, low stand. An important stillstand, during this withdrawal, resulted in washing across the York Moraine at the 33 m above OD lake level (Stage 3) with the production of sand screes or sand aprons, down the southern face of the moraine at Severs Howe and Mill Mound near York. A second significant stand, at 20 m above OD (Stage 5), partly produced by the moraine now having a damming effect, resulted in the deeper incision of the York Gap and the discharge at its exit of a coarse-grained delta once interpreted as the Crockey Hill esker. Evidence supporting the mapped terracing has been given by the acquisition of LIDAR imagery near Mill Mound on the York Moraine and at Everthorpe on the Wolds. An important feature of the terracing is that it appears to be horizontal and not affected by isostasy. The results of OSL dating tend to confirm that the interval between deglaciation of the Vale of York and the final withdrawal of Lake Humber can be assigned to a period after 17.0 ± 1.33 ka and before 15.5 ka.

The second major conclusion is that there have been two generations of periglacial alluvial fans discharged from dry valleys in the Wolds by the release of frost-shattered Chalk formation. This conclusion is based on the recognition that the older set of fans has been terraced by recessional stages of Lake Humber in contrast to the younger set. The later set, which only extends as far south as Market Weighton, appears to have been discharged during the waning stages of Lake Humber while the older set seem to have originated during a penultimate glaciation, that could be MIS 8 in age. Such an age, deduced from one OSL sample dated at 202.7 ± 11.7 ka could not be substantiated by later sampling.

It is also important to realise, that while the mapped shorelines in the Vale of York do represent a final imprint for stands of Lake Humber, they do not nescessarily represent a simple punctuated decline system, as oscillations of the retreating North Sea ice could have resulted in a lake decline with many unrecorded fluctuations.

Table 4 is a summary of chronological data for the Late Devensian, including recently acquired OSL data, for a pre-cursor low-level Lake Humber and the demise of the high-level lake.

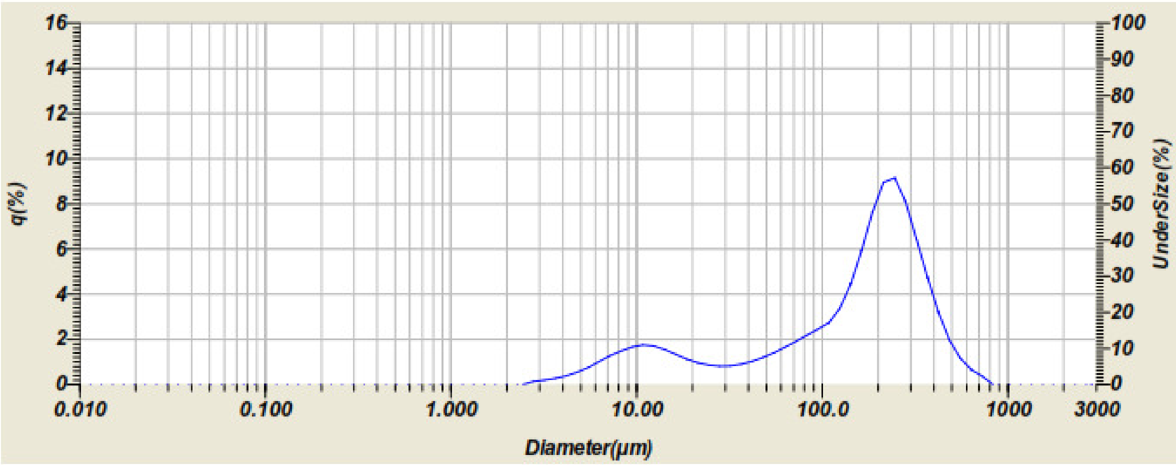
TABLE 4
Chronology of the Multi-stage Demise of High-level Lake Humber

Lake Levels	Time (ka)	Depositional event	BIIS / Vale of York glacier
Pre-cursor Low-Level Lake Humber	<23.3 ± 1.5	Loess under till Ferrybridge (Bateman *et al*., 2008)	Advance Vale of York glacier
	< 22.2 ± 0.5	Park Farm Clay Mbr – lacustrine seds (Murton *et al*., 2009)	Deposition Vale of York till
	c. 22.2 ± 0.5	Lawns House Farm Sand Mbr – lacustrine seds (Murton *et al*., 2009)	Emplacement Escrick Moraine
	<17.5 ± 1.7 17.01 ± 1.33	Loess below Skipsea Till (Wintle & Catt, 1985) Sediments in Skipsea Till (Hartmann, 2011)	BIIS advance Humber Gap blocked
	>16.6 ± 1.2	High-level Lake Humber (Bateman *et al*., 2008)	
	> 16.6 ± 1.2	40 m terrace at Bilbrough, Brayton Barff and the Yorkshire Wolds (This study)	
	16.6 ± 1.2	33 m terrace at Ferrybridge (Bateman *et al*., 2008) Deposition of sand scree on Mill Mound (This study)	Skipsea Till terraced at Hessle, 33 m OD
	16.2 ± 0.4	Sands below Withernsea Till (Bateman *et al*., 2011)	Skipsea ice retreat before 16.2 ka Advance Withernsea ice after 16.2 ka Humber Gap blockage stabilised
	15.9 ± 0.9 (Shfd12068) 15.2 ± 0.68 (Shfd13034)	Reworked sand scree between 20 m – 25 m below Mill Mound (This study)	
		Termination for some Younger Alluvial Fans	
	15.38 ± 0.92 cal. BP	Roos Bog (Bateman *et al*., 2011) develops over Withernsea Till Most Younger Alluvial Fans terminate	Retreat of Withernsea ice before 15.38
	15.5 ± 0.8 (Shfd13036)	Sand below 15 m terrace at Pollington (This study)	
	< 15.5 ± 0.8	Southerly dipping bedforms of sand & gravel below 10 m terrace at Pollington (Plate 39) (This study)	
	< 15.5 ± 0.8	Lowest level of Lake Humber (Plates 31 & 32) (This study)	

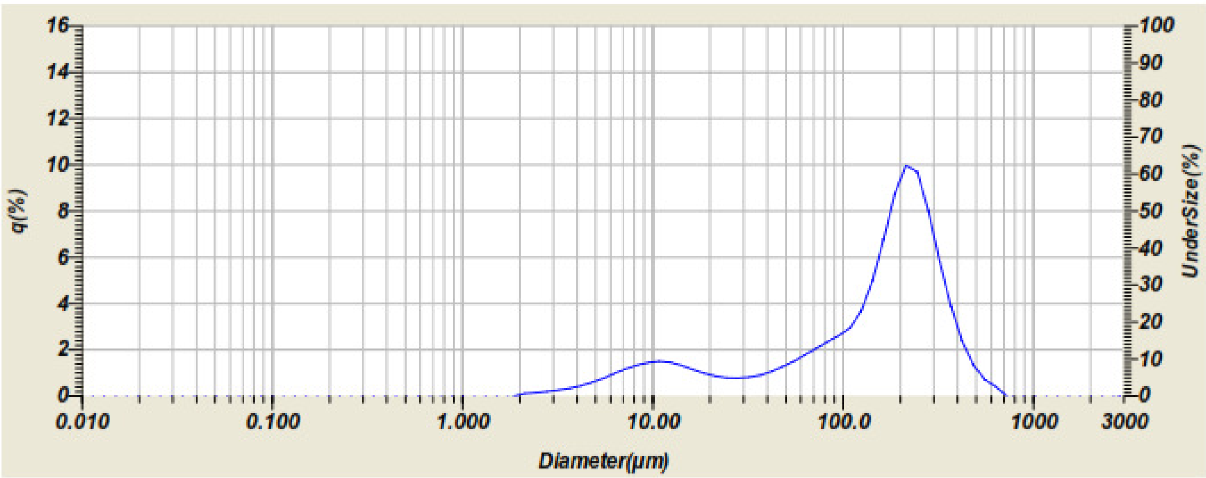
15. APPENDIX

Particle Size Distribution of Sand Samples

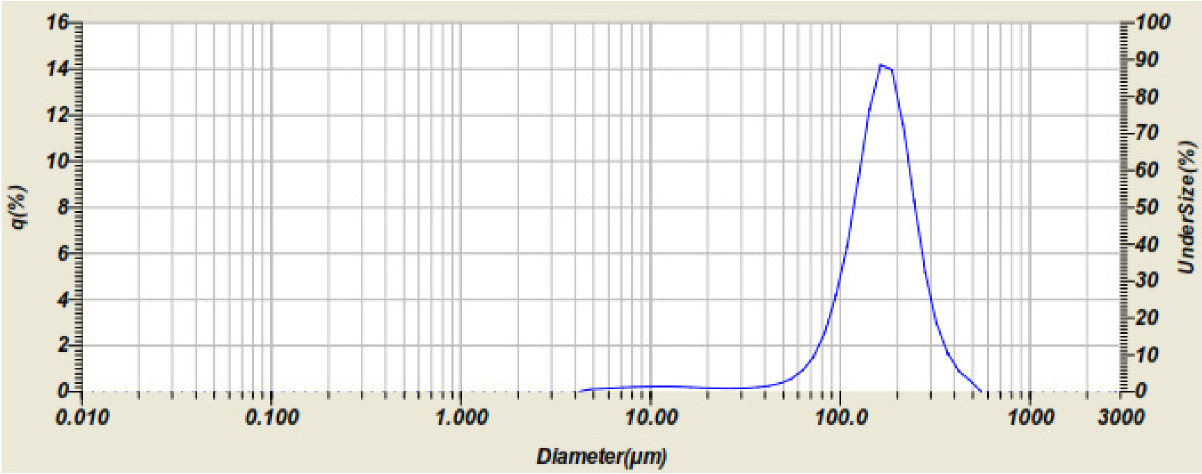
Sand Sample – Naburn Sewer 1 (Naburn Sand Member) – Shfd 11110



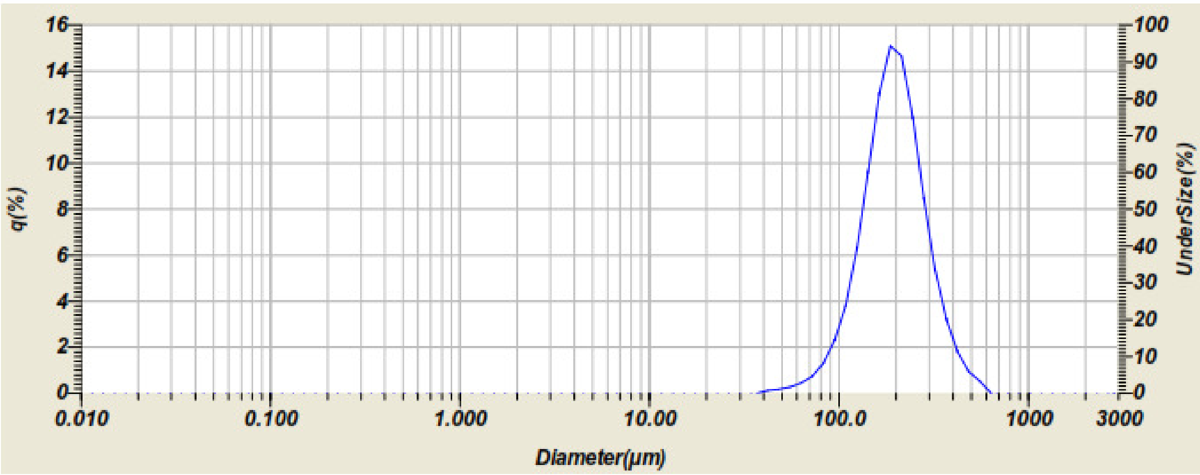
Sand Sample – Naburn Sewer 2 (Naburn Sand Member) – Shfd 11111



Sand Sample – Houghton Moor 1 – Shfd 11112



Sand Sample – Houghton Moor 2 – Shfd 11113



Sample No.	Median Size (µm)	Mean Size (µm)	Mode Size (µm)
Shfd 11110	181.98	187.19	243.65
Shfd 11111	176.91	177.02	215.44
Shfd 11112	166.36	174.93	164.50
Shfd 11113	194.43	207.86	188.25

(The particle size distribution was undertaken using a Horiba 950 Laser Particle Size Analyzer.)

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