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Middle and Late Pleistocene glacial lakes of lowland Britain and the southern North Sea Basin

Della K. Murton*, Julian B. Murton

Permafrost Laboratory, Department of Geography, University of Sussex, Brighton BN1 9QJ, UK

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ABSTRACT

Focusing on lowland Britain and the southern North Sea Basin, this article reviews the sedimentary and geomorphic evidence for the main glacial lakes inferred during the Middle and Late Pleistocene and evaluates their impacts on drainage-basin development. Glacial lakes are best known from glaciations during Marine Isotope Stages (MIS) 12, 6 and 2, although glacial lakes have also been inferred during MIS 10 and 4. Some lakes – for example, Bosworth, low-level Humber and the lakes of the eastern Fenland margin – are reconstructed from unequivocal sedimentary evidence, including rhythmites and subaqueous outwash, whereas others lakes – for example, Lapworth and Fenland – are inferred mostly from erosional features and remain to be substantiated. The largest known glacial lake developed in the southern North Sea Basin between an ice sheet to the north and a chalk bedrock ridge in the Strait of Dover area, first during the Anglian/Elsterian glaciation of MIS 12 and again during the late Wolstonian/late Saalian Drenthe glaciation of MIS 6. The palaeohydrological impacts of lake drainage are thought to include cutting of the Strait of Dover as a result of catastrophic drainage from the North Sea Lake during MIS 12, incision of a number of gorges and river valleys in England, and diversion or even reversal of major rivers such as the Thames and the proto-Soar/Avon system. Recently, varve chronologies have been correlated with the Greenland ice-core record, although caution is needed to discriminate between varves and non-annual rhythmites. Future work on Pleistocene glacial lakes needs to test chronologies of lake development by luminescence dating of glaciolacustrine sediments deposited in non-ice-proximal locations – (1) fine-grained rainout deposits and (2) wave-rippled sands deposited in shallow water – and to model the impacts of glacial isostasy in order to reconstruct lake extents. All of this work should be based on the rigorous application of sedimentology to interpret sedimentary facies and depositional environments.

1. Introduction

The Pleistocene glacial lakes of Britain and the southern North Sea Basin constitute an important archive of cold-climate environmental conditions and a driver of landscape evolution. They have a long history of investigation, beginning with pioneering studies like those of Lewis (1894) and Harrison (1898). Using modern examples as analogues, Kendall (1902) developed a set of criteria to identify former glacial lakes in the British landscape: beaches, deltas, floor deposits and overflow channels. Importantly, he observed that glaciolacustrine sediments differ from alluvial deposits in that the former exhibit regular lamination and, being deposited parallel to the subjacent surface, could be highly inclined. The first summary of glacial-lake systems in Britain was undertaken by Charlesworth (1957), more recently tabulated by Palmer (2005). However, given the variable quality of evidence used to delimit Pleistocene glacial lakes in Britain and the often generalised sedimentological descriptions of their deposits, synthesis and evaluation of more than a century of glacial lake studies in Britain is timely.

This paper therefore provides the first systematic review of Middle and Late Pleistocene glaciolacustrine deposits in lowland Britain and the southern North Sea Basin. It aims to provide a balanced view of the evidence used to identify the former glacial lakes and, where possible, to place it within a stratigraphical framework. It is not intended to be exhaustive, but to focus on the more significant lakes that developed in lowland England or those which have prompted most discussion in the literature (Fig. 1). Its objectives are to (1) describe the historical development of ideas about glacial lakes and their regional context; (2) summarise and

* Corresponding author.
E-mail address: d.murton@sussex.ac.uk (D.K. Murton).

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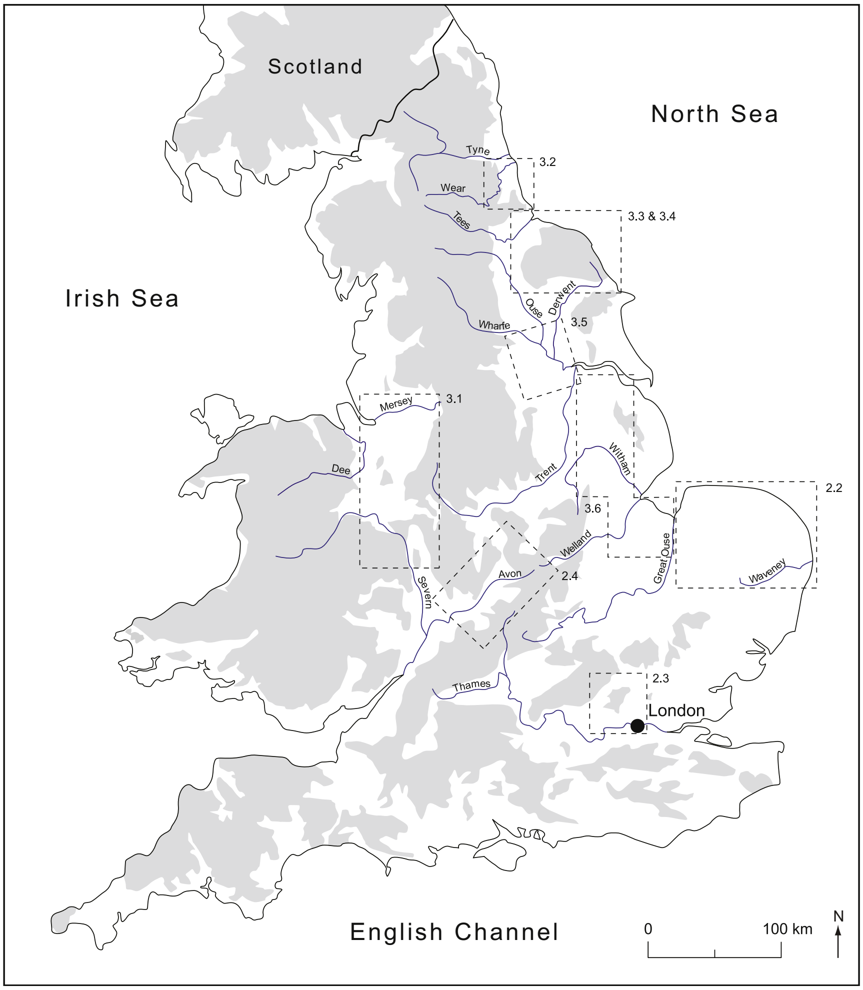

Fig. 1. Physiographic map of England and Wales showing the principal rivers of the contemporary drainage network and rivers mentioned in the text. Dashed boxes indicate terrestrial regions discussed in the text, with numbers beside them denoting the section headings. Shaded areas are >100 m OD.

evaluate the sedimentary and geomorphic evidence used to reconstruct them; (3) assess their influence on Quaternary landscape evolution – principally on drainage networks, both onshore and offshore; and (4) evaluate the methodologies used to devise an annually resolved varve chronology for Britain. We group the lakes into two main time periods (pre-Devensian and Devensian) and examine them regionally within each period, before considering the varve chronology and making recommendations for future research. Diagrams of the lakes are derived from the original publications and are adapted to clarify the Pleistocene and physiographic context of the lakes.

2. Glacial lakes of the pre-Devensian (MIS 12–6) cold stages

The principal periods of glacial lake formation inferred in Britain relate to glaciations during Marine Isotope Stages (MIS) 12, 6 and 2, although some evidence has been presented for glacial lakes during MIS 10 and 4. In Britain, MIS 12 is conventionally assigned to the Anglian glaciation, MIS 4–2 to the Devensian glaciation, and the intervening glacial stages (MIS 10, 8 and 6) to the Wolstonian Complex (Catt et al., 2006). Glacigenic landforms and glaciolacustrine deposits attributed to MIS 2 are the most extensive, as mapped by the BRITICE project (Clark et al., 2004). Such an authoritative database as BRITICE does not yet exist for older glaciolacustrine deposits, which are preserved beyond the MIS 2 ice margin in central and eastern England, and in the southern North Sea Basin.

2.1. The southern North Sea Basin and the Strait of Dover

The Strait of Dover, underlain by chalk of the Weald-Artois anticline, forms a shallow sill that separates the English Channel and southern North Sea Basin (Fig. 2). While the Channel floor is dominated by a system of narrow, anastomosing and often overdeepened valleys, the southern North Sea Basin is a sedimentary sink composed of fluvial, glaciolacustrine and marine deposits (reviewed in Gibbard, 1995). Significantly, the main valley apex of the Channel River ('Fleuve Manche') system – the Fosse Dangeard (Destombes et al., 1975), an infilled palaeovalley – and the adjacent Lobourg Channel originate from the western edge of the chalk sill. This section examines the relationship between sediments in the southern North Sea and the geomorphology of the English Channel seabed.

2.1.1. Sedimentary evidence from the southern North Sea

Detailed descriptions of the southern North Sea Basin sediments are sparse and, in the British sector at least, restricted to seismic profiles supplemented by occasional shallow borehole records (Cameron et al., 1987, 1992). Of relevance to this review are the sediments assigned to the Middle Pleistocene (ca. 780–128 ka; i.e.,

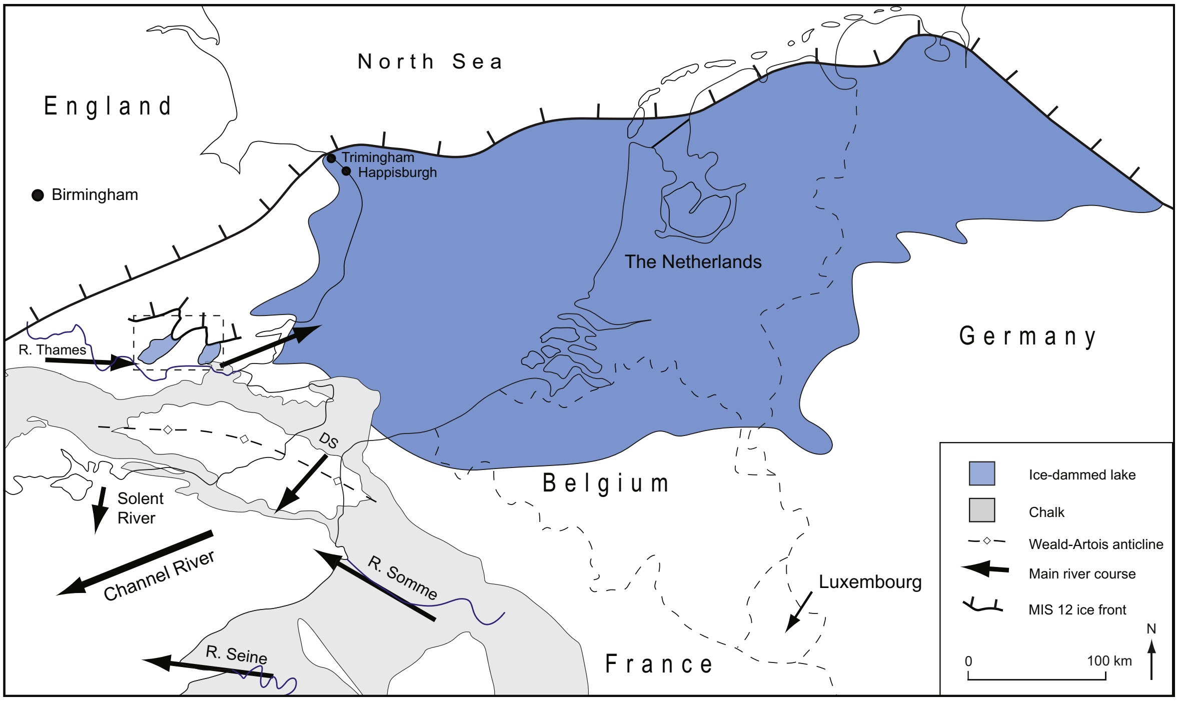

Fig. 2. North Sea Lake during MIS 12, and its relationship with the chalk bedrock ridge underlying south-east England and north-west France. DS: Dover Strait. The dashed box refers to the region discussed in section 2.3; the ice limit shown is the MIS 12 maximum extent of the Anglian/Elsterian Ice Sheet. Adapted from Gibbard (1995, 2007) and Cohen et al. (2005).

between MIS 19 and the MIS 6–5e boundary; Catt et al., 2006, p. 441), when the influence of glaciation on the lowland landscape of southern Britain became more prevalent. Within the southern North Sea Basin, the Middle Pleistocene lithostratigraphy comprises the Yarmouth Roads Formation unconformably overlain by the tripartite Swarte Bank Formation (SBF) (Long et al., 1988). The former is composed of fluvial sediments derived from the prograding deltas of major north European rivers such as the proto-Thames, Rhine and Maas/Meuse (Cameron et al., 1987) which flowed across the exposed sea floor. During this time the Dover Strait area was still a bedrock interfluve, and so North Sea ventilation was restricted to the Norwegian Sea. West of the Dover Strait, the proto-Somme, Seine and Solent rivers were tributaries of the Channel River system (Fig. 2), incising deep palaeovalleys (Antoine et al., 2003; Lericolais et al., 2003; Toucanne et al., 2009a).

In Britain, the most extensive Pleistocene ice sheet identified formed during MIS 12 (Gibbard and Clark, 2011), which offshore in the North Sea Basin is represented by the SBF (Long et al., 1988). During this glacial period, NNW–SSE-trending valleys were incised into the underlying sediments and subsequently infilled by the SBF (Cameron et al., 1987). As the British and Fennoscandinavian ice sheets coalesced for the first time in the northern North Sea Basin ca. 450 ka ago (Toucanne et al., 2009b), glacial meltwater and outflow from the northward-draining rivers became impounded in a massive proglacial lake in the southern North Sea Basin (Gibbard, 1988, 1995) (Fig. 2). It is now thought that the North Sea Lake was considerably larger than originally suggested in Gibbard's (1988) reconstruction and extended from eastern East Anglia across the southern North Sea Basin and into northern Belgium, the Netherlands and Germany, fed by drainage from much of western Europe (Cohen et al., 2005; Gibbard, 2007). This lake is thought to have measured ca. 550 km east–west and ca. 250 km north–south, covering a total area of nearly 140,000 km^2 (Fig. 2). The middle member of the SBF is composed of glaciolacustrine clay with beds of silty clay and fine-grained sand (Long et al., 1988) deposited during this period.

2.1.2. Sedimentary and geomorphic evidence from the English Channel

The similarity of the Channel River valley system to the Channel Scabland in the northwest USA prompted suggestions that incision of the Dover Strait was initiated by catastrophic drainage of ice-dammed water in the southern North Sea (Smith, 1985). This corroborated sedimentary evidence from Wissant, near Calais, France (Roep et al., 1975), where two units of sand and gravel, interbedded with sands and clays, were interpreted as fluviatile. Critically, the lowermost unit occupied a fossil bench incised in the underlying bedrock, intraclasts of which were incorporated in the gravels. Structurally, the cross-cutting channels ca. 10 m wide were infilled with sands showing climbing ripples. Elsewhere in the section, 1.5-m-thick intercalated sand beds exhibited mega-cross-bedding. Similar structures were observed in the upper sand and gravel unit. Palaeocurrent measurements from these coarse-grained units indicated a southerly flow direction. It was inferred that overflow from an ice-dammed lake in the southern North Sea Basin occurred twice.

Using high-resolution sonar data from the floor of the English Channel, Gupta et al. (2007) identified three lines of evidence consistent with a high-magnitude flood hypothesis and rapid lowering of base level: (1) streamlined islands of bedrock and longitudinal erosional grooves; (2) knickpoints at the confluence of the palaeo-Solent and the main Channel valley; and (3) a sub-horizontal bench incised in the chalk bedrock of the northern valley flank.

The bench morphology in particular suggests two separate episodes of flooding, while small truncated and beheaded channels superimposed in the streamlined islands suggest subaerial exposure of the valley floor, with normal fluvial processes dominant between these floods (Gibbard, 2007). This is consistent with the

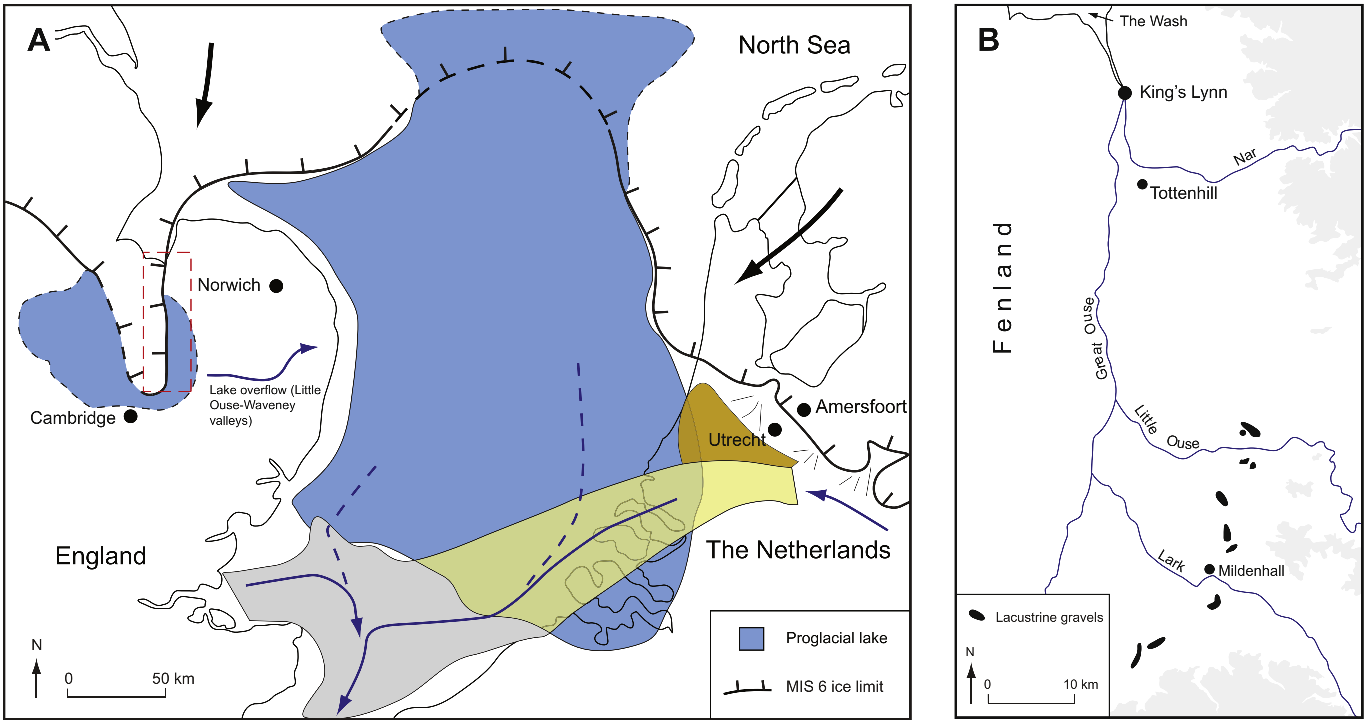

Fig. 3. (A) North Sea Lake and lake extent in the Fenland region of eastern England during MIS 6. The grey area denotes the Southern Bight of the southern North Sea Basin, which formed part of a Southern Bight/Dover Strait land bridge to the south of the North Sea Lake. The brown area denotes glaciofluvial gravelly sands deposited on the proglacial braidplain of the Rhine-Meuse river system at the time of maximum ice-sheet extent during the late Saalian Drenthe glaciation (unit S4 in Busschers et al., 2008, fig. 7C). The braidplain terminated in a delta extending into the North Sea Lake, the water level of which was around mean interglacial sea level. The yellow area indicates gravelly sands deposited in the area of the braidplain dissected by the River Meuse during and after lake drainage in the Drenthe deglaciation (unit S5 in Busschers et al., 2008, fig. 7D). The River Thames did not enter the lake itself in the Saalian Drenthe stage, but joined the spillway river downstream of it. The red dashed box highlights the part of the eastern Fenland shown in (B). Adapted from Busschers et al. (2008), Hijma et al. (in review) and K. Cohen (pers. comm. 2011). (B) Distribution of glaciolacustrine gravels, deposited in MIS 6, along the eastern margin of the Fens. Contour at 50 m OD. Adapted from Gibbard et al. (2009). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

fluvial evidence at Wissant. But precise dating of these events remains problematic, although the lithostratigraphy of the southern North Sea Basin strongly supports a MIS 12 age for at least one of the floods, and a major peak in X-ray fluorescence-determined Ti/Ca ratios – a proxy for fluvially derived terrigenous input – from an ocean sediment core in the Bay of Biscay is consistent with catastrophic drainage causing initial breaching of the Weald-Artois structural barrier ca. 455 ka ago (Toucanne et al., 2009b).

Prior to this first flood, in MIS 12, the lowest col in the Chalk barrier was ca. 30 m OD (Smith, 1985). Thus, the water level of the North Sea Lake at its maximum in the Anglian/Elsterian must have been at least ca. 25 m, and possibly 30 m, above present sea level (Cohen et al., 2005), although the level declined through time as the spillway was deepened (a common phenomenon in glacial lakes ponded by a bedrock or morainic ridge). At the end of the Anglian, the water level probably stood at ca. 12–14 m asl (Gibbard et al., 1996). Gupta et al. (2007) calculated maximum peak discharges for the flood events of between ca. 0.2 and $1.0 \times 10^6\ m^3\ s^{-1}$, discharges of sufficient magnitude to initiate erosion of the Dover Strait. As chalk readily fractures by ice segregation (Murton et al., 2006) and given that the Dover Strait was likely underlain by permafrost during its incision, as advocated by Smith (1985), it is suggested that overflowing floodwaters would have rapidly eroded the brecciated and ice-rich upper layer of permafrost developed in chalk.

A second period of impoundment of a major proglacial lake in the southern North Sea Basin – with levels comparable to interglacial sea level stands there – occurred during MIS 6, to the south of the late Wolstonian/late Saalian Drenthe ice margin (Fig. 3A). This lake is inferred from proglacial fluvial sediments of the Rhine-Meuse river system that fed into a delta lying at an enigmatically high elevation in The Netherlands at the time of maximum ice-sheet extent (Busschers et al., 2008). The high elevation of the delta – at the same level as MIS 7 and MIS 5 coastal deposits – indicates that the Strait of Dover and Southern Bight land bridge to the southwest of the lake had not fully opened at this time (Cohen, pers. comm. 2011). These Rhine-Meuse glaciofluvial sediments were later dissected by the incising River Meuse during and after deglaciation. The incision phase was associated with a drop in lake level to below present-day mean sea level as the land bridge was eroded during the later part of MIS 6. The late Saalian Drenthe age of the North Sea Lake is inferred from optically stimulated luminescence (OSL) ages from the fluvial sediments. Although uncertain whether lake drainage and resulting spillway incision were catastrophic or gradual, Busschers et al. (2008) argued that this breach of the Dover Strait/Southern Bight land bridge to a level below present-day mean sea level resulted in the separation of Britain from mainland Europe during subsequent interglacials.

2.2. *Terrestrial glacial lakes: East Anglia*

2.2.1. *Stratigraphic context*

In East Anglia the Middle Pleistocene lithostratigraphy is complex. Originally described by Reid (1882), the stacked sequences of tills interbedded with separate units of gravels, sands and clays have prompted contrasting litho- and chronostratigraphical models (reviewed in Preece et al., 2009; Lee et al., 2010) (Table 1), with markedly different palaeoenvironmental

Table 1
Lithostratigraphic units and chronostratigraphical models proposed for the Middle Pleistocene deposits of Norfolk, eastern England, showing the stratigraphical context of the sedimentary units referred to in the text. The full lithostratigraphy and palaeoenvironmental reconstruction is presented in Lunkka (1994) and Lee et al. (2004a).

Bowen (1999)	Lunkka (1994)	Proposed MIS stage	Lee et al. (2004a,b); Hamblin et al. (2005)		Proposed MIS stage
			Briton's Lane Formation		6
			Sheringham Cliffs Formation	Trimingham Sand Member Trimingham Clay Member Ivy Farm Laminated Silt Member	10
Lowestoft Formation	Lowestoft Till/Marly Drift	12	Lowestoft Formation	Walcott Till Member/Lowestoft Till Member (lateral equivalents)	12
North Sea Drift Formation	Trimingham Sands Trimingham Clay Cromer Diamicton Mundesley Diamicton Mundesley Upper Sands Walcott Diamicton Lower Mundesley Sands Happisburgh Clays Happisburgh Diamicton	12	Happisburgh Formation	Corton Sand Member Happisburgh Sand Member Ostend Clay Member Happisburgh Till Member	16

reconstructions. As it is beyond the scope of this review to discuss these in detail, the focus is on key sedimentary units indicative of waterlain deposition exposed at Trimingham, Happisburgh and Tottenhill (Figs. 2 and 3B).

2.2.2. Sedimentary evidence

2.2.2.1. Northeast Norfolk. In outline, the waterlain sediments in northeast Norfolk were either deposited in an extensive glacial lake in the southern North Sea Basin, during MIS 12 (Lunkka, 1994), or principally reflect subaerial outwash and localised glaciolacustrine deposition (Hart, 1992) associated with different glacial stages (Lee et al., 2004a, 2008). In the former model the associated diamictons were deposited predominately by Scandinavian ice, while in the latter the diamictons are attributed to repeated incursions of the British Ice Sheet, which flowed southeastwards adjacent to the present-day coastline of eastern Britain.

Detailed descriptions of the sediments at Trimingham are presented in Hart (1992), Lunkka (1994) and Lee (2003). Hart (1992) identified five units (in parentheses) constituting – as it was then termed – the Trimingham Member. The lowermost unit (unit 1) comprises laminated sediments. At the base these are interbedded

Fig 4. Symmetrical ripple form sets (arrows) with peaked crests (wave ripples) draped by clayey silt, Ostend Clay Member at Happisburgh, Norfolk. Wave ripples, wavy bedding and flaser bedding are common throughout this vertical section through glaciolacustrine deposits. The wave ripples are probably of MIS 12 age. The trowel handle is 11 cm long.

with a massive silty clay diamicton, but couplets of silt or silty sand grading into clay gradually become more dominant. The upper contact of each clay lamina with the succeeding silt lamina is sharp. The silt laminae frequently exhibit micro-laminae of sand or silt, which are prevalent near the base of the sequence. Higher up the sequence, the thickness of the clay laminae increases, grading first into massive clay (unit 2) with indistinct laminae before becoming increasingly calcareous (unit 3). This unit is overlain by a fine-grained, pale yellow sand that contains a few small ripple structures (unit 4). The uppermost unit (5) is a clay bed that sharply overlies the fine-grained sand. It becomes increasingly laminated and deformed towards the top. It is overlain by diamicton.

Hart (1992) interpreted this sequence as glaciolacustrine, deposited in glacial Lake Trimingham. The thin silty sand layers in the lowermost laminated sediments are consistent with a proximal or intermediate proximal position within the lake (Ashley, 1975), suggesting deposition close to a delta front. The interbedded diamicton reflects reworking of the underlying till by debris flows. To determine whether the laminated sediments were varves or turbidites, Hart (1992) devised the rhythmite index (see Section 4.2), which compared the ratio between the thicknesses of the clay laminae with those of silt/clay laminae. From this she concluded that these laminated sediments were varves, which, if the homogenous clay bed is also included, recorded deposition over a period of 1000–2000 years. Lee (2003), however, attributed the graded contacts between the silt and clay laminae to deposition by turbidity currents. With the exception of the upper unit of laminated sediment, which suggested a resumption of meltwater influx, the remaining sedimentary sequence indicated that Lake Trimingham became progressively shallower as the sediment source diminished, possibly associated with climatic amelioration.

More systematic investigations, however, determined that the lower and upper laminated sediments exposed at Trimingham were distinct units, termed, respectively, the Happisburgh Clays and the Trimingham Sand and Clay Member (Lunkka, 1994), or the Ostend Clay Member and the Ivy Farm Laminated Silt Member (Lee et al., 2004a). This latter unit is at least 21 m thick, and comprises wave and current-bedded sand, with two beds of light grey silt and dark grey clay couplets, usually with sharp contacts between each lamina. An interbedded sand bed with climbing ripples and massive fines with diamicton lenses separates the finer-grained sediments.

Farther southeast at Happisburgh, the stratigraphy differs from that exposed at Trimingham. Here, the Happisburgh Clays/Ostend Clay Member typically comprises, from the base up: (1) intercalated

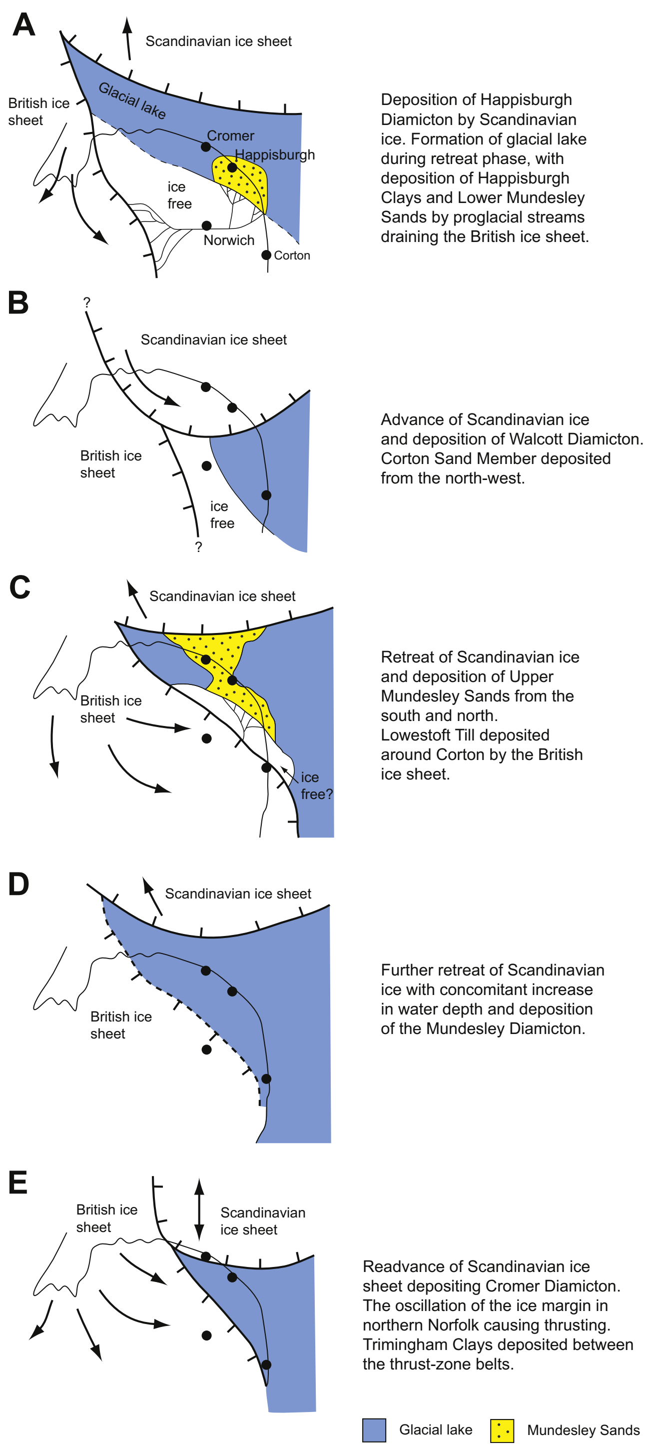

Fig 5. Schematic representation of palaeoenvironmental change in north-east Norfolk during MIS 12. Adapted from Lunkka (1994).

beds of sandy diamicton and coarse silt; this unit also contains ripple-bedded sand interbedded with clay and silt laminae (Fig. 4), with both symmetric and asymmetric ripple structures; (2) massive silt beds separated by silt and clay couplets, with occasional lenticular and convolute beds; and (3) massive silt with occasional discrete laminae and ripples (Lunkka, 1994; Lee et al., 2008).

The Ostend Clay Member was deposited in a proglacial environment associated with retreat of the Happisburgh Till ice and subsequent advance by Walcott Till ice (Hart, 1992; Lunkka, 1994; Roberts and Hart, 2000) or Corton Till ice (Lee et al., 2008). Although Hart (1992) envisaged deposition in an isolated glacial lake (Lake Trimingham – not to be confused with the Trimingham lake bed of Preece et al. (2009), which is a localised bed incised into the top of the Briton's Lane Formation), Lee (2003) recognised a series of coalescing proglacial lake basins that had formed on the ridged surface of the Happisburgh Till.

As the Ostend Clay Member is coarser at Happisburgh than at Trimingham, these sediments may reflect a continuum of deposition in a single body of water, deepening towards the northwest. Lunkka (1994) interpreted the Happisburgh Clays (and the overlying Lower Mundesley Sands) as prograding deltaic bottomsets, deposited in an extensive glacial lake impounded by Scandinavian ice in the southern North Sea Basin (Fig. 5A). The northern extent of this lake was determined by episodic retreat and readvance of the Scandinavian ice, while its southern limit was restricted by the bedrock col of the Dover Strait, as discussed above. Other overlying sedimentary units, in particular the Corton Sand Member and Mundesley Diamicton, have also been attributed to deposition in this lake (Fig. 5B–D), implying that it persisted throughout MIS 12.

The base of the Corton Sand Member is defined by a thin gravel lag deposit. The remainder of the deposit is a well-sorted, stratified sand containing detrital chalk grains (Bridge, 1988). It exhibits a variety of structures from planar cross-bedding, type A and type B climbing ripples to horizontal bedding (Lee et al., 2006, 2008). Although the consensus interpretation is that these deposits are deltaic, Lee et al. (2004b, 2008) considered these sands to be distal glaciofluvial outwash derived from meltwater draining into the Bytham River during the Happisburgh Glaciation (MIS 16). This MIS 16 age, however, is inconsistent with biostratigraphic and amino-acid data, which strongly support the conventional view that the Happisburgh Till is of Anglian (MIS 12) age (Preece et al., 2009) and represents the first diamicton member of the North Sea Drift Formation.

Mundesley Diamicton overlies the Mundesley Sands, with a predominately gradational contact (Lunkka, 1994). It comprises sandy diamicton lenses interbedded with horizontally bedded sands, and fine sand and silt/clay couplets. The lower and upper contacts of these couplets are sharp. Dropstones and shelly fragments are present throughout the unit.

In conclusion, the northeast Norfolk sequences can be interpreted as basically the product of glaciolacustrine sedimentation formed where the Scandinavian Ice Sheet impinged on the British landmass in the North Sea basinal area. The evidence includes the laminated sequence attributed by Hart (1992) to deposition in 'Lake Trimingham', which in fact represents the North Sea Lake (Fig. 5E) that filled the basin through most of Anglian/Elsterian Stage time (MIS 12; Fig. 2). However, the glacial lake sediment is not restricted to bottomset fines, and some of the diamictons are waterlain, especially the Mundesley and Cromer Diamicton Members (Ehlers and Gibbard, 1997). The sand units are glaciodeltaic (including, for example, the Corton Sands Member), and the overall sequence represents ice-front oscillation in the relatively deep-water lake (Clark et al. 2004; Gibbard and Clark, 2011). If an analogue, or at least a parallel series of deposits exists for the NE East Anglia sequences, it is those that occur on the margins of the great glacial lakes, such as the Weichselian of the southern coast of the Baltic Sea, or the Great Lakes' Wisconsinan sequences, such as those in southern Ontario on Lakes Erie and Ontario.

2.2.2.2. Western Norfolk and eastern Cambridgeshire. In western Norfolk and eastern Cambridgeshire, some distinctive coarse-grained glacial lake sediments flank the eastern margin of the Fenland (Fig. 3B; Gibbard et al., 1992, 2009). Here, a line of sands and gravel accumulations – assigned to the Tottenhill Member –

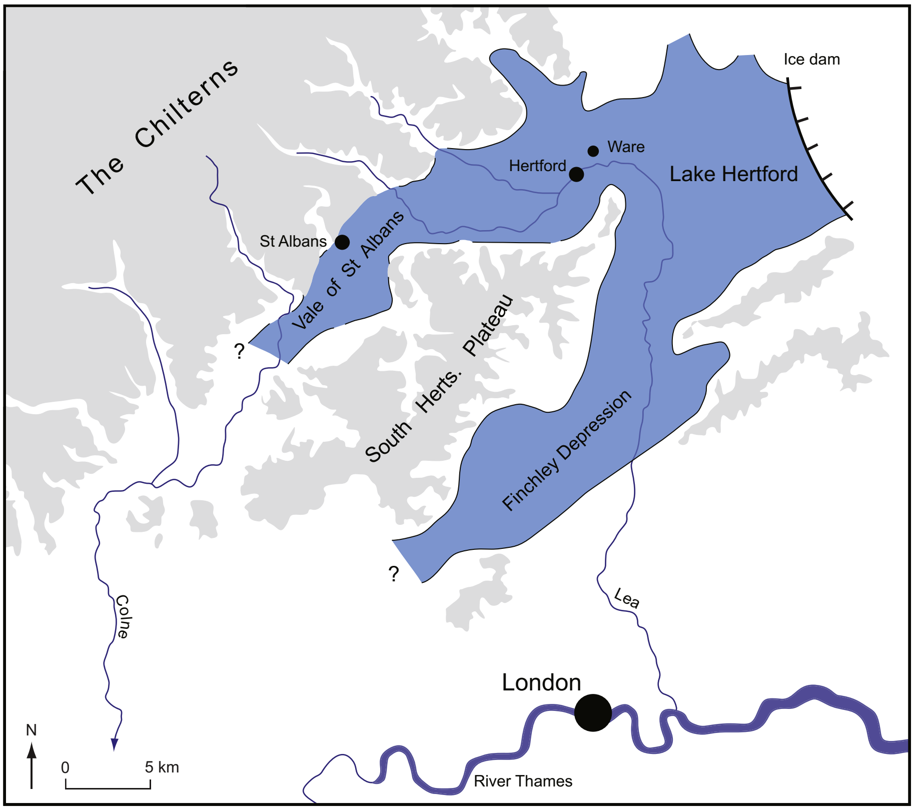

Fig. 6. The maximum extent of glacial Lake Hertford and its postulated ice dam during MIS 12. The rivers are shown in their contemporary courses. Contour at 90 m OD. Adapted from Clayton and Brown (1958).

occurs on hills or ridges near the rivers Little Ouse and Lark, close to the foot of the Chalk escarpment. The ridges are fan-shaped and have a typical ice-contact morphology, with the steeper (stoss) sides facing west or northwest, and the gentler (lee) sides facing east or southeast. The dominant sediments within them are matrix-supported gravels rich in chalk clasts. The gravels commonly contain steeply dipping foresets up to at least 2 m high. Subsidiary sands – some with climbing ripple cross-lamination and draped lamination – and rare silts and clays are also present.

Deposition of these gravelly sediments most likely occurred within a series of subaquatic fans prograding towards the east or south-southeast into one or more shallow, ice-marginal lakes. The fans developed where high-energy and sediment-charged water emerged, probably from esker-like tunnel mouths, along an ice margin immediately west or northwest of the fans. The parent ice lobe is thought to have advanced east- to southeastward into the Fenland Basin (Fig. 3A), damming meltwater between the ice margin and the Chalk escarpment to the east and south. Eventual merging of a number of small lakes may have culminated in a single proglacial lake that drained eastward through the Little Ouse-Waveney valleys and onwards to the large proglacial lake in the southern North Sea Basin (Section 2.1). However, detailed reconstructions of the development of the southern lake complex near the Little Ouse Valley relative to the Nar Valley lake near Tottenhill (Fig. 3B) remain to be established.

The age of these glacial lake sediments is assigned to the late Wolstonian and corresponds to the late Saalian Drenthe substage in The Netherlands (Gibbard et al., 2009), that is ca. 175–160 ka (Toucanne et al., 2009a). This is based on regional correlation, supported by OSL dating of sands, which places the so-called 'Tottenhill glaciation' at ca. 160 ka (MIS 6; Gibbard and Clark, 2011).

Additional sedimentary evidence for a proglacial ice-dammed lake is found in the Nar Valley, in the form of laminated clays and subaquatic till. This lake, however, is thought to have developed during deglaciation of the Anglian Ice Sheet in MIS 12 (Gibbard and Clark, 2011).

2.3. Terrestrial glacial lakes: Vale of St Albans

Concomitant with a glacial lake in the southern North Sea Basin during MIS 12 were smaller ice-dammed lakes in the northern tributary valleys of the contemporary river Thames (Fig. 2). The most significant of these lakes developed in the Vale of St Albans (Figs. 6 and 7) and contributed, in conjunction with at least two ice advances into the region, to diversion of the Thames from its pre-MIS 12 course through the Vale of St Albans to its current route through London (Gibbard, 1977; Bridgland, 1994 and references therein).

2.3.1. Sedimentary evidence

Significant glaciolacustrine deposits in this region were originally recognised by Sherlock and Pocock (1924), although the presence of a lake in the Vale of St. Albans was first proposed by Clayton and Brown (1958). However, systematic description of the evidence for a lake was first presented by Gibbard (1974, 1977), and included an attempt at varve correlation.

Clayton and Brown (1958) identified a 6-m-thick bed of laminated sediment at Ware, although generally these sediments were just a few metres thick. In addition, the laminated sediments were often locally interdigitated with diamicton, interpreted as till, and overlain by current-bedded sands and gravels. Extrapolating from known observations of glaciolacustrine sediments, up to ca. 73 m OD, they defined a proglacial lake – Lake Hertford – which occupied both the Finchley Depression and the Vale of St Albans (Fig. 6).

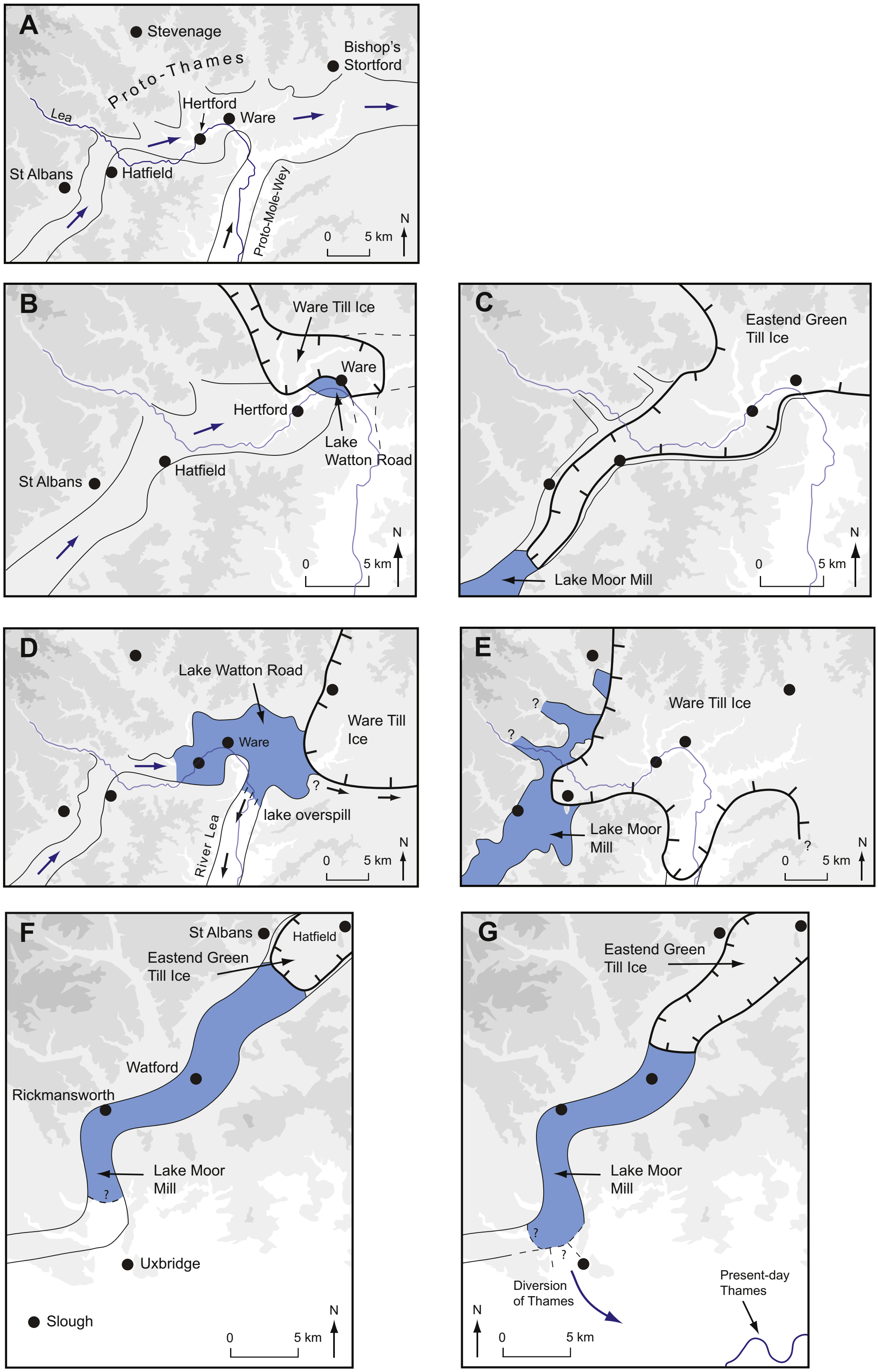

Fig. 7. Models of glacial lake development and palaeohydrological changes in the Vale of St Albans region during MIS 12. Blue arrows denote the course of the proto-Thames. Contours are at 50 m intervals, the lightest shading corresponding to 50 m OD. (A) Palaeogeography of the Vale of St Albans, with the proto-Mole-Wey as a right-bank tributary to the proto-Thames. The course of the contemporary river Lea is shown for orientation. (B and C) Formation of glacial Lake Watton Road, and glacial Lake Moor Mill, attributed to two separate ice advances. Adapted from Gibbard (1977). (D and E) Formation of glacial Lake Watton Road, and glacial Lake Moor Mill, attributed to a single ice advance that deposited the Ware Till. Overspill from glacial Lake Watton Road initiated the southwards-flowing river Lea. Adapted from Cheshire (1986). (F and G) Diversion of the river Thames to its present-day course as a result of overspill from Lake Moor Mill, which extended southwestwards to Uxbridge concomitant with advance of Eastend Green Till ice. Adapted from Gibbard (1977). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

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Focusing on the Vale of St Albans, Gibbard (1977) identified two separate glaciolacustrine units towards its eastern end, which he termed the Watton Road Laminated Silts and the Moor Mill Laminated Clays. These were deposited in glacial Lake Watton Road and glacial Lake Moor Mill, respectively, with both units underlain by fluvial sands and gravels. The Watton Road Laminated Silts are identified by 2.5 m of brown laminated clayey silts composed of silt and clay couplets up to 10 mm and 2–3 mm thick, respectively. From this exposure 485 varve-like couplets were counted, suggesting lake duration was at least 485 years. The couplets are overlain by massive grey brown silty clay which grades into the overlying Ware Till.

The Moor Mill Laminated Clays are up to 5 m thick and cover an area of at least 4 km by 1.2 km. They comprise silt and clay couplets up to 2 mm thick, of which 342 varve-like couplets were counted. Interbedded within the laminated sediments are massive horizons of silty clay with slump bedding structures. The uppermost metre of these clays exhibits soft-sediment deformation structures, with flame structures extending into the overlying Eastend Green Till.

2.3.2. Palaeohydological implications

The relationship between the course of the proto-Thames (Fig. 7A) and inferred glacial lakes is controversial. Clayton and Brown (1958) argued that Lake Hertford formed by an ice advance from the north east, which dammed drainage of the proto-Thames, then flowing – according to these authors – in the Finchley Depression. This palaeoenvironmental reconstruction, however, is unlikely because subsequent work has failed to substantiate a single, large glacial lake connecting the Finchley Depression and the Vale of St Albans. Moreover, there is now general consensus that the proto-Thames, prior to its diversion during the Anglian Stage, did not flow along the Finchley Depression but instead flowed along the Vale of St Albans, before being diverted directly into its modern valley (reviewed in Bridgland, 1994). Today, the proto-Thames valley in Vale of St Albans is occupied by the southwestward flowing river Colne.

Gibbard (1977) attributed these laminated sediments in the Vale of St Albans to two periods of localised glacial lake development associated with separate ice advances southwestwards

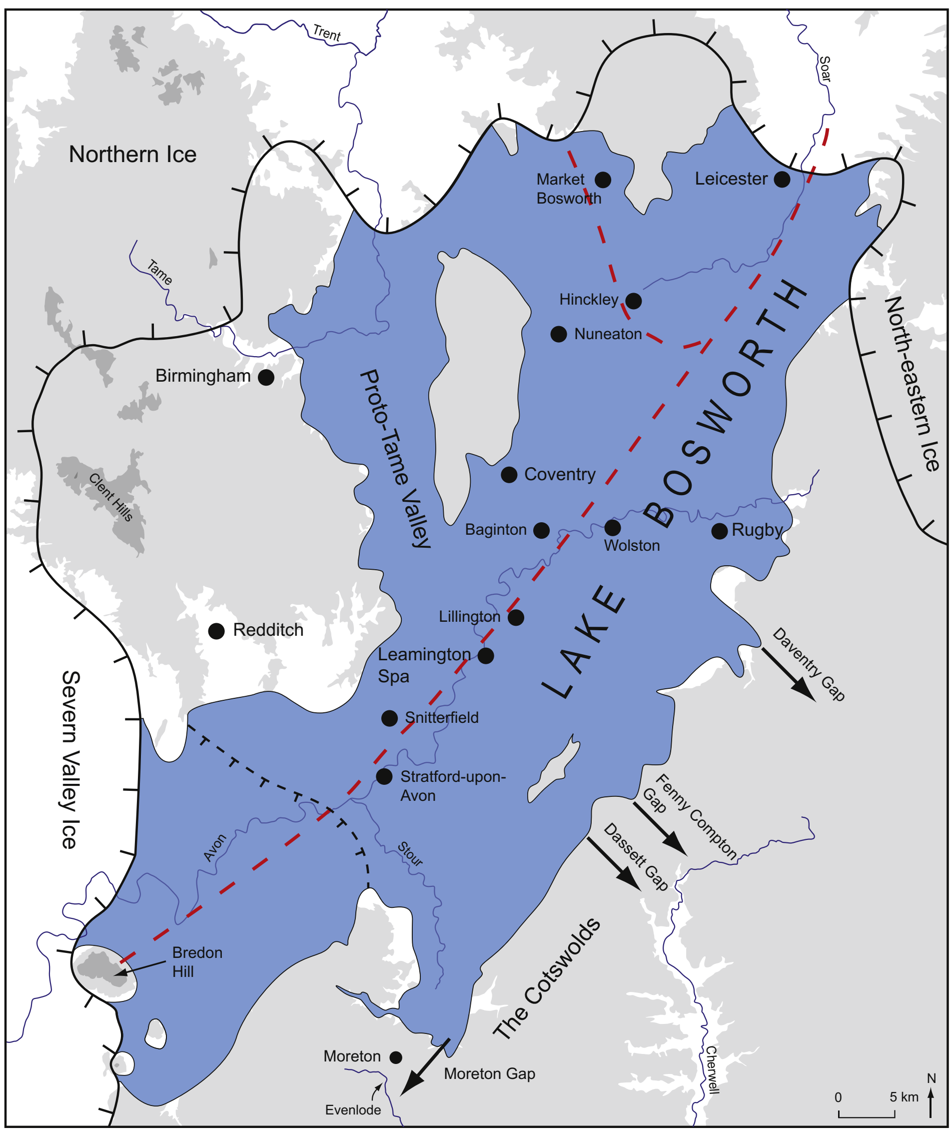

Fig. 8. Generalised extent of glacial Lake Bosworth at its maximum lake level of 125 m OD during the Wolstonian glaciation (?MIS 6 or 10) or the Anglian glaciation (MIS 12). The dashed line denotes the original lake margin depicted by Shotton (1953) for the south-western extent of Lake Bosworth, as defined by the margin of Welsh ice in the Severn Valley. Red dashed line denotes the axis of the proto-Soar valley (from Douglas, 1974). The contemporary drainage network and places mentioned in the text are shown. Contours are at 100 m intervals, the lightest shading corresponding to 100 m OD. Adapted from Shotton (1983). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

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Table 2
Nomenclature, sedimentary properties and palaeoenvironmental interpretation of the Pleistocene sequence between Leicester, Market Bosworth, Redditch and Moreton-in-Marsh, preceding and during the period of deposition within Lake Bosworth. Additional sedimentary descriptions are derived from Pickering (1957); Douglas (1980); Sumbler (1983); and Old et al. (1987).

Stratigraphical unit		Sedimentary properties	Palaeoenvironmental interpretation
Shotton (1953)	Shotton (1976); Rice (1981)		
Dunsmore gravel	Dunsmore gravel	Brown, poorly sorted sandy and clayey gravel typically 3–4 m thick. Gravel is subangular and interbedded with subordinate beds of sand. Pebbles mainly flint, with some Bunter pebbles and ironstone.	Outwash gravels from retreating ice sheet of north-eastern origin.
Upper Wolston clay	Upper Oadby Till Lower Oadby Till	Highly calcareous and pinkish grey, weathering to brown. Significant proportion of chalk, flint, and Jurassic limestone and sandstone pebbles; Triassic lithologies are much reduced.	Advance of ice from the east and NNE, with transient ice-marginal lakes ponding in front of advancing ice lobe.
Wolston sand	Wolston Sand and Gravel	Basal contact is characterised by a gravel lag of Triassic lithologies, often intercalated with underlying silts and clays. Upper contact is sometimes similarly intercalated. The deposit thickens (up to 13 m) and becomes more gravelly towards the north. Sands are red, and fine- to medium-grained. Trough-cross-bedded sand in Hinckley valley.	Braided deposits of an outwash floodplain associated with glacial retreat.
Lower Wolston clay	Bosworth Clays and Silts	Approximately 16–18 m thick near Coventry and Leamington Spa, ≤48 m thick near Market Bosworth. Mainly massive, calcareous and grey brown to chocolate brown, weathering to reddish brown. Frequently contains pebbles of Bunter and Triassic sandstone in lower few metres, with flint, chalk and Jurassic limestone prominent in upper part. Contains fine couplets of red-brown silt and clay; couplet thicknesses exhibit a similar spatial distribution, from ~1.6 to 3 mm. In Hinckley valley the clay lamina often fines upwards, and the silt lamina occasionally exhibits small-scale cross-lamination. Clays are intercalated with sands or sands and gravel 1.8–2.3 m thick.	Glaciolacustrine, with isolated pebbles deposited as dropstones by icebergs. Lenses of intercalated till suggest ice remained nearby throughout the lake's duration, with clast lithological differences corresponding with the Thrussington (Trias-dominated pebbles) and Oadby (chalk and Jurassic pebbles) ice advances, respectively. Normal grading and cross-lamination within couplets in Hinckley valley suggest non-annual deposition by turbidity currents.
	Thrussington Till	Three to five meters thick, red-brown clay. Contains Triassic sandstone and siltstone, Bunter pebbles and coal fragments. Sharp basal contact; gradational upper contact over ~1 m with Bosworth Clays and Silts.	Southerly ice advance up proto-Soar valley.
Baginton sand	Baginton sand	Mainly cross-bedded with scour-and-fill channels, becoming level-bedded towards the top and often intercalated with beds of silt and clayey sand.	Fluvial sedimentation of reduced competence. Intercalation of fine-grained sediment reflects initiation of Lake Bosworth. Period of ice advance.
Baginton–Lillington gravel	Baginton–Lillington gravel	Matrix-supported gravel of rounded pebbles of quartz and quartzite at Baginton, and Jurassic limestone and ironstone at Lillington.	Fluvial sedimentation. Lithological differences reflect tributaries to main pre-glacial valley of proto-Soar.

across the Vale (Fig. 7B and C), whereas Cheshire (1986) considered both lakes to be the product of a single ice advance that deposited the Ware Till (Fig. 7D and E). In both palaeoenvironmental reconstructions the proto-Thames flowed through the Vale of St Albans. Of the localised glacial lakes, Lake Watton Road formed first. Although the varve record indicated that this lake persisted for at least 485 years, Gibbard (1977) observed little evidence for lake overflow during this time and concluded that the course of the proto-Thames after the lake's demise remained unaltered. With additional sedimentary evidence derived from borehole records and previously published observations he inferred instead that overspill from the more extensive Lake Moor Mill in the western part of the Vale initiated the diversion of the Thames southwards at Uxbridge (Fig. 5F and G). A similar, separate glacial lake formed in the Finchley valley, and possibly other valleys of south bank tributaries (Gibbard 1979), but according to this author these lakes never merged with those in the Vale of St Albans, and certainly the Thames never flowed northwards through the Finchley (Mole-Wey) valley. The river's diversion arose from overspill from valley to valley.

Cheshire (1981, 1986), by contrast, argued for an earlier diversion of the Thames, via the river Lea, as the Ware ice advance impounded drainage in Lake Watton Road (Fig. 5D).

2.4. Central England

2.4.1. Stratigraphic context

In the English Midlands, the existence of a former glacial lake was first postulated by Harrison (1898), who recognised that the configuration of ice masses and relief in northwest Leicestershire would have impounded a lake. He named this body of water Lake Bosworth, after Bosworth Field, situated 5 km south of Market Bosworth. Shotton (1953), however, renamed it Lake Harrison, arguing that a specific place name was inappropriate for a postulated lake that straddled the watershed of the contemporary river network (Fig. 8). But given Harrison's precedence in identifying the former lake, in addition to the substantial thickness of glaciolacustrine deposits around Market Bosworth (≤48 m thick; Douglas, 1980) and their occurrence in a regionally mappable unit – the Bosworth Clays and Silts (Douglas, 1974; Shotton, 1976), here it is considered appropriate to revert to the original name: Lake Bosworth.

A robust stratigraphy for central England – encompassing the counties of Warwickshire, Leicestershire and Gloucestershire – was established by Shotton (1953). He identified six units (Table 2), of which the Wolston Series (now termed 'Wolston Formation') – comprising the Lower Wolston clay, Wolston sand and Upper Wolston clay – is directly attributed to deposition in Lake Bosworth. From

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auger data, Shotton (1953) deduced that these deposits straddle the current watershed divide in England and concluded that a different drainage network configuration existed prior to the development of Lake Bosworth. This pre-glacial valley, then envisaged as part of the River Trent system, extended northeastward from Bredon Hill towards Leicester (Fig. 8), and within it accumulated the Baginton-Lillington sands and gravel. The younger part of the Wolston Formation extends beyond the confines of the proto-Soar valley system.

Subsequent revisions of stratigraphic nomenclature (Shotton, 1976; Rice, 1981) recognised a clay-rich diamicton – the Thrussington Till – within the Lower Wolston clay and re-assigned the Upper Wolston clay as Oadby Till, with a Lower and Upper facies. Glaciolacustrine sediments were incorporated within the Lower Oadby Till. The Thrussington Till extends to Leamington Spa (Old et al., 1987), whereas the Oadby Till terminates farther south at Moreton-in-Marsh (Bishop, 1958) (Fig. 8). The overall lithostratigraphy remains robust, and more recent correlations and nomenclature are summarised in Bowen (1999). The chronostratigraphy, however, remains controversial and it has yet to be established for certain whether the sediments represent one or more glacial periods during the Wolstonian or Anglian glaciations (MIS 12, 10 or 6). Nonetheless, with growing recognition of the widespread late Wolstonian glaciation in MIS 6 (ca. 160 ka) in both continental Europe and Britain, Gibbard and Clark (2011) have suggested that there is a strong likelihood that the Wolston Formation is indeed of this age, as originally envisaged by Shotton. For a more detailed discussion see Catt et al. (2006) and Gibbard and Clark (2011) and references therein.

2.4.2. *Evidence for Lake Bosworth*

Although Harrison (1898) conceptualised Lake Bosworth, the lithostratigraphical model presented in Table 2 provided the contextual understanding of how such a lake could develop and be sustained. In this model, Shotton (1953, 1983) envisaged Lake Bosworth to have existed between the Thrussington Till and Oadby Till ice advances, because in the upper reaches of the proto-Soar, at Snitterfield (Fig. 8), laminated sediments directly overlie Baginton sand (Tomlinson, 1935), suggesting initial lake impoundment was most probably concomitant with the Thrussington Till ice advance from the north. Based on observations of lake clay elevations in the Leamington Spa-Rugby-Coventry region, Shotton (1953, 1976) reconstructed the maximum lake level at 125 m OD. At this elevation it was impounded by the Cotswolds to the southeast, with Welsh ice occupying the Severn valley, northern ice in the Tame valley, and northeastern ice in the proto-Soar (Fig. 8). Only during recession of the Thrussington Till ice lobe did Lake Bosworth extend northwards into south-western Leicestershire (Douglas, 1980), reaching 90 km long at its maximum extent (West, 1968). Calculations from thicknesses of silt and clay couplets near Nuneaton suggested this phase of Lake Bosworth lasted at least 9600 years (Shotton, 1976).

Independent mapping by Dury (1951) in southeast Warwickshire identified a dissected bench at ca. 122 m OD cut into the Lower Lias rocks along the Cotswolds escarpment, which he interpreted as a wave-cut platform, or former shoreline of Lake Bosworth. Additional geomorphological mapping by Bishop (1958) confirmed the continuation of this bench through the Fenny Compton Gap. This was deemed significant, since gaps or cols in the Jurassic escarpment – at Fenny Compton, Daventry, Dassett and Moreton (Fig. 8A) – provided overflow outlets for Lake Bosworth (Shotton, 1953). Of these, Fenny Compton was inferred as the most probable (Dury, 1951; Bishop, 1958).

Bishop's (1958) mapping, however, also recorded lake clays up to 130 m OD, higher than Shotton's postulated maximum lake level. To account for these elevation differences Bishop (1958) invoked a two-stage model: (1) extra-morainic Lake Bosworth, and (2) inter-morainic Lake Bosworth (Fig. 9). In phase one, the Oadby Till ice advance initially impounded Lake Bosworth up to 133 m OD (Fig. 9A), with a lower level at ca. 126 m OD (Fig. 9B) and overflow through the Moreton gap. Phase two of Lake Bosworth was associated with Oadby Till ice retreat, with lake levels first at 131 m OD (Fig. 9C) and finally at 125 m OD (Fig. 9D). In this reconstruction the Moreton gap was the initial overflow route, replaced by Fenny Compton at a later stage. The palaeohydrological implications of this are discussed in Section 2.4.3.

It is not clear whether Bishop's (1958) reconstruction of Lake Bosworth depicted in Fig. 9D was comparable to its original reconstruction as envisaged by Shotton (1953). As a result it is possible that the glaciolacustrine deposits are not contemporaneous, but may instead represent transient glacial lakes developed in front of, upon or within the oscillating ice lobes (Sumbler, 1983; Old et al., 1987; Harwood, 1988).

Two tributary valleys of the proto-Soar, the Tame and Hinckley, formed embayments of Lake Bosworth (Fig. 8). In the Tame Valley, three units of laminated sediments, interbedded with sands and gravels, or diamicton and gravels are recorded (Pickering, 1957). The lowest, at ca. 130 m OD, has similar sedimentary properties to the Lower Wolston clay and is included within the Lake Bosworth depicted in Fig. 8. The remaining units are associated with the Oadby Till ice advance. The two laminated sediment units, at maximum elevations of 143 m and 166 m OD, respectively, indicate progressive ice advance, with impoundment at increasingly higher elevations.

In the Moreton-in-Marsh area – located on a col separating the headwaters of the Warwickshire Stour and River Evenlode – Sumbler (2001) has inferred that ice advanced southward up the Stour valley, depositing Trias-rich tills and ponding water in front of it in the head of the valley. The sedimentary evidence is contained within the Moreton Member – a variable assemblage of clays, silts, sand and Trias tills, which underlies the Oadby Till. Sumbler correlated the Moreton Member with the lithologically similar Wolston Clay and Thrussington Till in their type area in Warwickshire. The waterlain clays and silts occur to an elevation of 137 m OD, presumed to be the initial height of a col where lake waters were able to overflow into and incise the Evenlode valley of the Upper Thames Basin (Figs. 1 and 8).

2.4.3. *Palaeohydological implications of Lake Bosworth*

Within the main valley of the proto-Soar the latter stages of Lake Bosworth, subsequent to the period of Thrussington Till deposition, remain uncertain and speculative. Prior to Lake Bosworth, palaeodrainage in the region was dominated by the River Trent, there was no Avon tributary to the River Severn, and the Cotswolds escarpment formed the watershed with the River Thames (Shotton, 1953). Today, the River Avon, which flows southwestwards towards the Bristol Channel, occupies the upper part of the proto-Soar (Fig. 8). A key question is therefore: *what role, if any, did Lake Bosworth have in initiating this river?*

Shotton (1953) argued that the Oadby Till ice advance – which extended to Moreton-in-Marsh and is evidenced by glaciofluvial sediments (Wolston sand) overlain by Oadby Till – obliterated Lake Bosworth. At this time the upper reaches of the proto-Soar were sufficiently shallow that glacial meltwaters were not impounded but instead flowed through the Moreton gap into the Upper Thames Basin (Bishop, 1958). Critically, it was withdrawal of Severn/Welsh ice from the vicinity of Bredon Hill that facilitated the initiation of an incipient river Avon: meltwaters from a receding Oadby Till ice lobe rapidly incised into the newly exposed superficial deposits infilling the proto-Soar valley (Shotton, 1953).

An alternative hypothesis infers that a diminished Lake Bosworth developed during retreat of the Oadby Till ice lobe (Bishop,

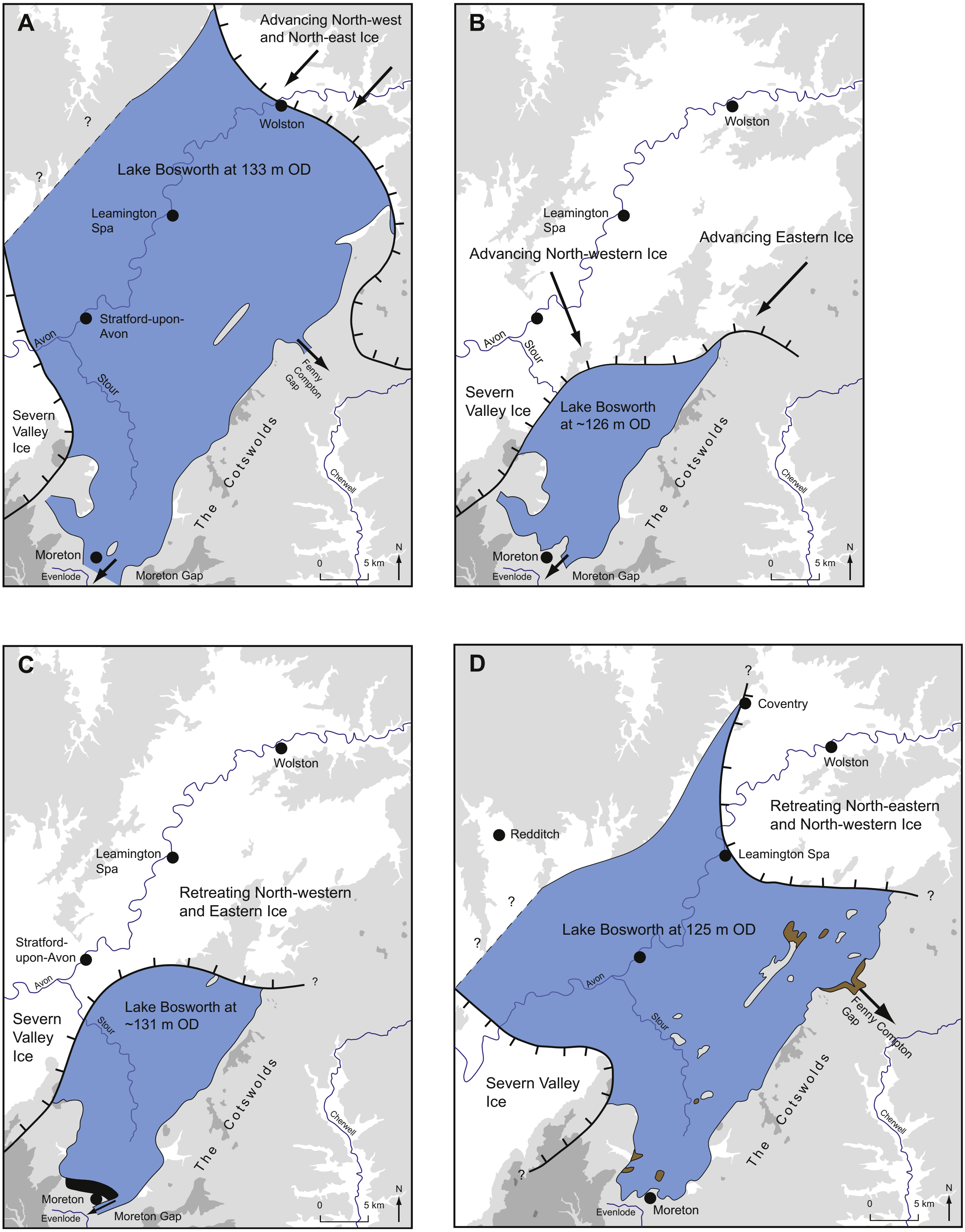

Fig. 9. A two-stage model for glacial Lake Bosworth associated with advance and retreat of the Oadby Till ice. (A–B) show an 'extra-morainic' stage, and (C–D) show an 'inter-morainic' stage. In A–C overflow from Lake Bosworth was through the Moreton Gap, and in D overflow was through the Fenny Compton Gap. In (C) the black area denotes the Moreton moraine ridge. In (D) the brown areas denote a rock bench. Adapted from Bishop (1958). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

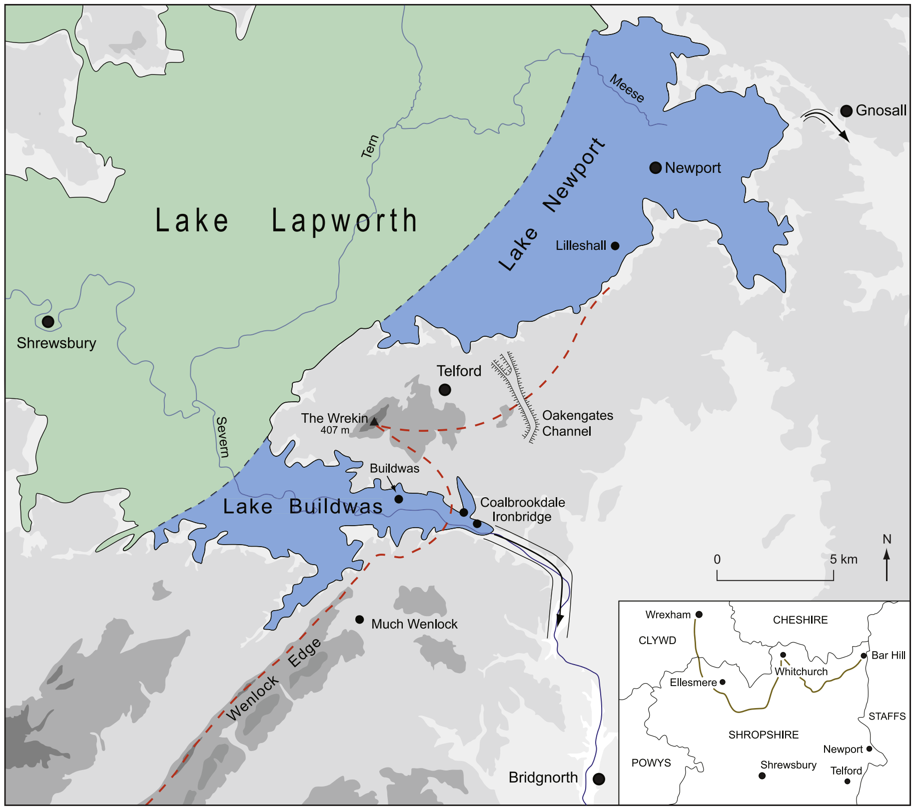

Fig. 10. Generalised extent of glacial Lakes Buildwas and Newport, and part of glacial Lake Lapworth during MIS 2. The dashed black line shows a hypothetical ice margin needed to pond Lakes Buildwas and Newport, with respective overflow routeways at Ironbridge and Gnosall. Dashed red line denotes the pre-glacial Dee-Severn watershed, which was breached by the Oakengates meltwater channel. Contours are at 50 m, 100 m, 200 m, 250 m, 300 m and 400 m OD. Inset map shows the location of the Wrexham–Ellesmere–Whitchurch–Bar Hill moraine complex (thick brown line). Adapted from Worsley (1975) and Hamblin (1986). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

1958; Shotton, 1977) (Fig. 9C and D). Two geomorphic features supported this inference: (1) a flat-bottomed valley at ca. 128 m OD eroded within the terminal moraine ridge at Moreton, interpreted as an overflow channel; and (2) the rock bench traversing the Fenny Compton gap at ca. 122 m OD. In this model, headwater erosion of the river Avon was initiated by the withdrawal of Severn/Welsh ice, releasing the impounded waters of Lake Bosworth. Overflow from Lake Bosworth through the Fenny Compton Gap drained into the River Cherwell, in the upper Thames Basin, causing incision.

Several lines of evidence are inconsistent with this latter palaeoenvironmental reconstruction. First, no glaciolacustrine sediments overlie the Dunsmore gravel. Second, geophysical markers from borehole records in the Warwick district suggest that the rock bench traversing the Fenny Compton gap and originally mapped by Dury (1951), is not a wave-cut platform but is structural and more commonly reflects cemented calcareous layers within the Lower Lias (Ambrose and Brewster, 1982). Furthermore, the formation of a wave-cut platform implies stable water levels with incremental lateral erosion. This seems incompatible with overflowing lake water funnelled through a rock col, which is more likely to induce vertical incision.

3. Glacial lakes of the Devensian (MIS 4–2) cold stage

During the Devensian cold stage, glacial lakes formed in northeastern and eastern England as ice masses in the North Sea Basin (Tweed-Cheviot) impinged upon the terrestrial landmass, often coalescing with ice flowing through the Tyne and Stainmore Gaps (Catt et al., 2006; Livingstone et al., 2010a). In particular, the decoupling of these ice lobes formed embayments into which drainage became impounded. A similar decoupling of Irish Sea ice and Welsh ice in the Cheshire–Shropshire lowlands of northwestern England also produced glacial lakes (Thomas, 1989). This section discusses the resultant lakes in the Cheshire–Shropshire lowlands, lower Wear and Tees valleys, North York Moors and Vale of Pickering, Vale of York, and the Fens (Fig. 1).

A large proglacial lake may also have developed for a third time in the North Sea Basin – dammed again by coalescing British and Scandinavian ice sheets to the north and the Strait of Dover to the south – during the period between ca. 30 and 25 ka, and possibly to ca. 20 ka (see Toucanne et al., 2009a, 2010). However, the existence of this lake is debated, and, if it existed, it does not appear to have been a significant sediment trap (Toucanne et al., 2010); thus it is not discussed further.

3.1. The Cheshire–Shropshire Lowlands

3.1.1. Physiographic and stratigraphic context

During MIS 2, ice from the Irish Sea Basin advanced southwards across the Cheshire–Shropshire Lowlands, depositing the Stockport Formation (Bowen, 1999), which broadly comprises two reddish brown tills separated by sands. In Shropshire, its maximum extent is at Bridgnorth, although farther north, the Wrexham–Ellesmere–Whitchurch–Bar Hill (WEWB) moraine complex likely denotes a later readvance (Fig. 10; Chiverrell and Thomas, 2010). Confluent or subsequent, eastward-flowing Welsh ice deposited till and glaciofluvial sediments assigned to the

Shrewsbury Formation, its limit at Shrewsbury. More detailed reviews are presented in Worsley (2005), and Chiverrell and Thomas (2010).

The principal rivers in the region are the Dee and the Severn, which flow into the Irish Sea. The pre-glacial watershed of the region, shown in red on Fig. 10, extends along the Wenlock Edge to the Wrekin, traversing the Coal Measures escarpment to Lilleshall (Hamblin and Coppack, 1995). Then, the proto-upper Severn drained into the Dee to the west, while the proto-lower Severn flowed east. Valleys such as the Oakengates channel, now buried by glacigenic sediments, cut across this watershed. The exception is the Ironbridge Gorge, a 10-km-long overdeepened valley section of the River Severn between Ironbridge and Bridgnorth (Fig. 10) which is devoid of surficial deposits. Lapworth (in Watts, 1898) suggested that diversion of the upper Dee drainage during glaciation initiated deepening of the River Severn to the south of Ironbridge. However, the concept of a glacial lake occupying the Cheshire–Shropshire lowlands was first proposed by Harmer (1907), who instead inferred that the Ironbridge Gorge was incised by its overflow. Wills (1924) mapped this region, aiming to reconstruct former ice margins and to present the evidence of this former glacial lake. Utilising data from E.E.L. Dixon, he identified two smaller glacial lakes, Lake Buildwas and Lake Newport, which he concluded amalgamated to form Lake Lapworth. Despite this, sedimentary and geomorphic evidence for Lake Lapworth, and its relationship to the Ironbridge Gorge, remain equivocal.

3.1.2. Evidence for Lake Lapworth and its subsidiary lakes

Just west of Ironbridge are hummocks of sand and gravel at ca. 90 m OD, which structurally are a rather chaotic mix of unbedded sands and gravels, bedded gravels with a prevailing eastward dip, a bed of fining-upwards coarse gravel to silty sand, and clasts of grey brown till. Locally, cross-stratified units with cut-and-fill structures have been observed (Hamblin, 1986). Laminated sediments are sparse, the only occurrence being a 3.8-m-thick isolated patch at ca. 135–138 m OD on the valley side at Lincoln Hill, Ironbridge (Wills, 1924).

Between Newport and the Wrekin, up to 12.8 m of laminated silts and clays with scattered pebbles and gravelly sand underlie, and occasionally interdigitate with till. In the vicinity of Lilleshall, laminated clays form terraces with sandy bluffs at 114 m, 110 m, 99 m, 71 m and 69 m OD. These deposits are interpreted as glaciolacustrine sediments of Lake Newport (Hamblin, 1986).

Northwestwards of the Wrekin, between Shrewsbury and Wrexham, fence diagrams derived from borehole records do not

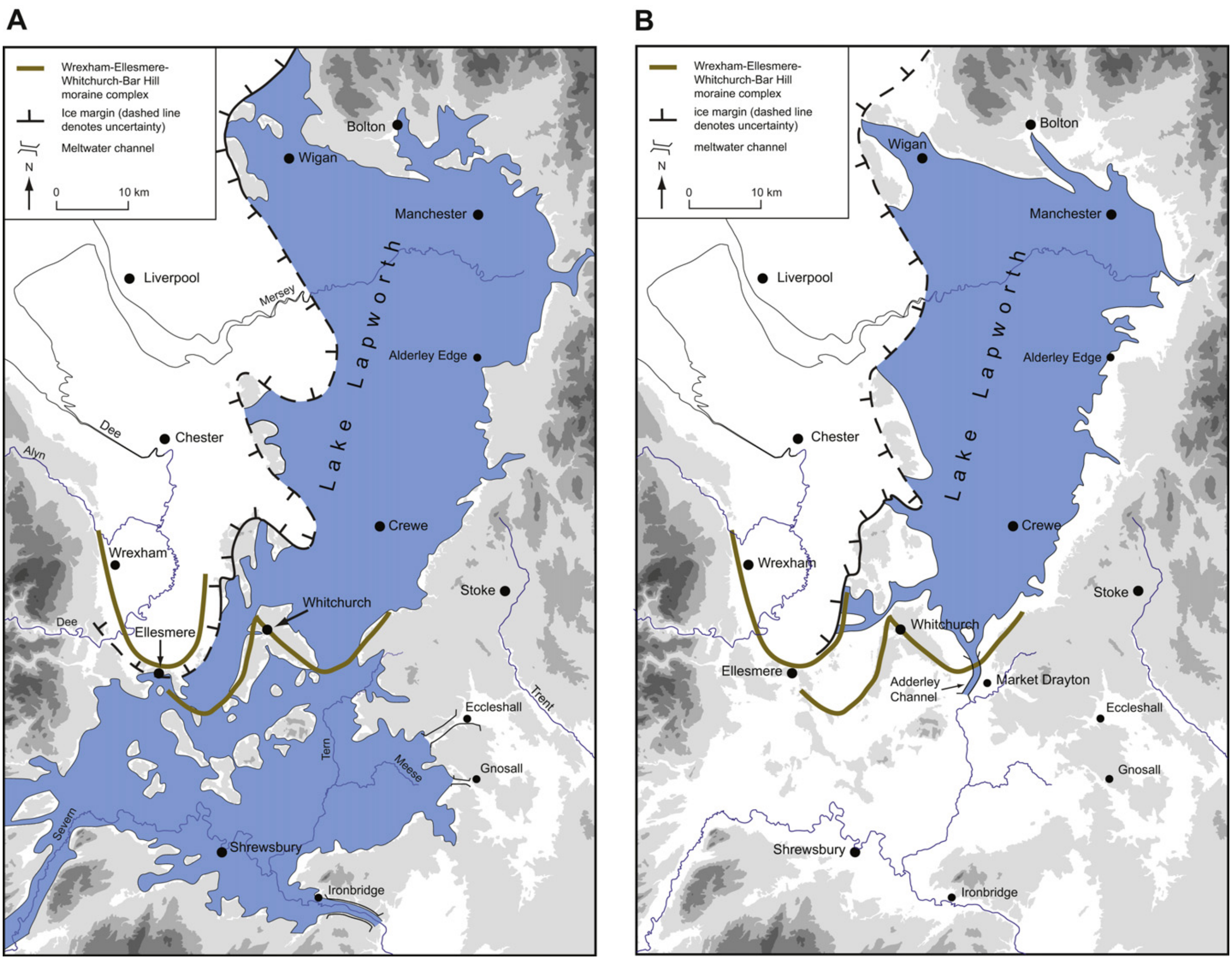

Fig. 11. Generalised extent of glacial Lake Lapworth with overflow channels at its 101 m OD level (A) and 76 m OD level (B). Contours are at 100 m intervals, the lightest shading corresponding to 100 m OD. Adapted from Poole and Whiteman (1961).

depict a regionally mappable glaciolacustrine deposit indicative of an extensive body of water (Lake Lapworth). Instead, the lithostratigraphy comprises laminated reddish brown, brown and greyish brown silts and clays interdigitating with till, sand and gravel (Thomas, 1989). There is much local variability. Thicknesses of laminated sediments range from >23.1 m in the bedrock depression south of Prees, to just 0.1 m within overlying glaciofluvial sands and gravels (James, 1983). In the Wem district the laminated sediments are calcareous (Cannell and Harries, 1981), while in the vicinity of Whitchurch they are overconsolidated and highly contorted (Jackson et al., 1983). Although sand lenses within the laminated silts and clays were observed across the surveyed region, rhythmically alternating silts and fine sand were most common north of the Triassic escarpment around Wem and Prees (Cannell and Harries, 1981; James, 1983), where they were interpreted as shallow-water glacial lake sediments.

In the vicinity of Borras, ~2 km north-east of Wrexham, open-face exposures in isolated quarries record a progressive southeastward fining sequence (Thomas, 1985). The sedimentary architecture reflects this, with the coarsest sediments composed of gravels intercalated with stony diamicton, while at the south-easternmost extent sands exhibiting parallel lamination and ripple-drift cross-lamination alternate with discontinuous layers of silt and clay. In between, sands show fining-upward tabular bedsets with a maximum foreset thickness of 8 m and sequences of parallel-laminated and cross-laminated sand separated by thick sets of trough or planar cross-bedded sands. The sedimentary sequences exposed around Borras are interpreted as a series of prograding fan deltas.

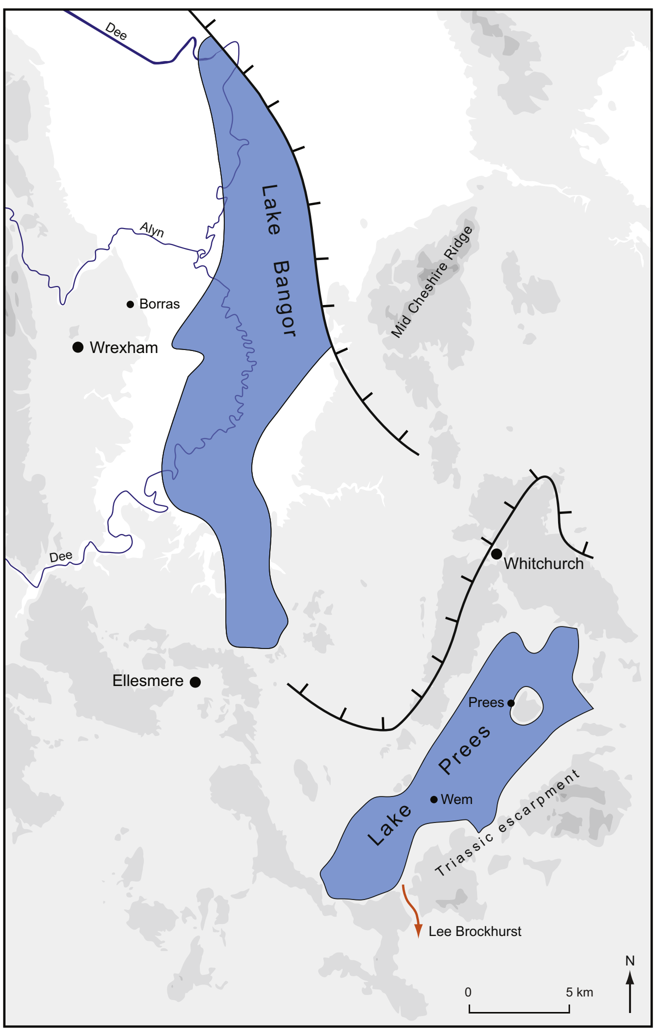

Fig. 12. Generalised extent of glacial Lakes Prees and Bangor, with the approximate locations of the ice margins at the time of their formation during MIS 2. Orange arrow shows overflow through a col in the Triassic escarpment at Lee Brockhurst. Contours are at 50 m intervals, the lightest shading corresponding to 50 m OD. Adapted from Thomas (1989). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

3.1.3. Palaeoenvironmental reconstructions

Palaeoenvironmental reconstructions of glacial lake development in this region vary considerably, with only the earlier models supporting the Lake Lapworth concept. Wills (1924) inferred that Irish Sea ice situated just west of the pre-glacial watershed impounded meltwater in the vicinity of Coalbrookdale (Fig. 10). He attributed the laminated sediments near Ironbridge to deposition in this incipient lake, up to a level of ca. 141 m OD. Continued withdrawal of Irish Sea ice westwards enlarged this lake (Lake Buildwas), eventually amalgamating with Lake Newport after ice had receded from the base of the Wrekin to form Lake Lapworth. Concomitant was a fall in lake level to ca. 90 m OD (Fig. 10). In this model, the northern limit of Lake Lapworth is defined by the WEWB moraine complex. At Shrewsbury, outwash terraces interpreted as relict deltas at ca. 66 m OD (Worsley, 1975) and attributed to the Shrewsbury Formation suggest progressive lake lowering prior to the advance of Welsh ice.

Poole and Whiteman (1961) proposed a more regionally extensive Lake Lapworth, with four distinct stages associated with retreat of Irish Sea ice (Fig. 11). The highest lake level (101 m OD) is delimited from an unspecified shoreline feature, traceable across the region (Fig. 11A). Overflow during this time occurred at Eccleshall, as well as Gnosall and Ironbridge. The second stage, at ca. 93 m OD is also delimited by shoreline features, albeit less well developed than that of the previous stage. The third and fourth stages (Fig. 11B) are confined to the region north of the WEWB moraine complex, with drainage through the Adderley channel. Given the presence of the WEWB moraine complex, however, it is questionable whether this should all be regarded as Lake Lapworth, or as two separate lakes.

Recent palaeoenvironmental reconstructions based on more substantive sedimentary evidence (Thomas, 1985, 1989) are inconsistent with the geomorphic evidence of earlier models promulgating Lake Lapworth. Instead these reconstructions suggested that decoupling of the Irish Sea ice and Welsh ice, initially in the vicinity of Shrewsbury, then northwestwards towards Wrexham and the Alyn valley, impounded a series of transient and smaller ice-marginal lakes. The largest identified were glacial Lake Prees and glacial Lake Bangor (Fig. 12), the inferred length by width dimensions of which were 20 km by 5 km, and 30 km by 10 km, respectively (Thomas, 1989). The former developed as meltwater emanating from Irish Sea ice situated at the WEWB moraine complex became impounded between it and the Triassic escarpment. Overflow was south through the col at Lee Brockhurst. As ice progressively receded north it infilled Lake Prees with prograding outwash sands and gravels whilst subsequently impounding glacial Lake Bangor, as meltwater and drainage from the Alyn and Dee rivers became impeded by the WEWB moraine complex. Delta foresets on the western margin at Borras infer a lake height of 70 m OD. This lake, too, was infilled by prograding sandur deposits.

3.1.4. Palaeohydrological implications of Lake Lapworth

Wills (1924) argued that overflow from Lake Buildwas initiated incision of a col in the pre-glacial watershed at Ironbridge,

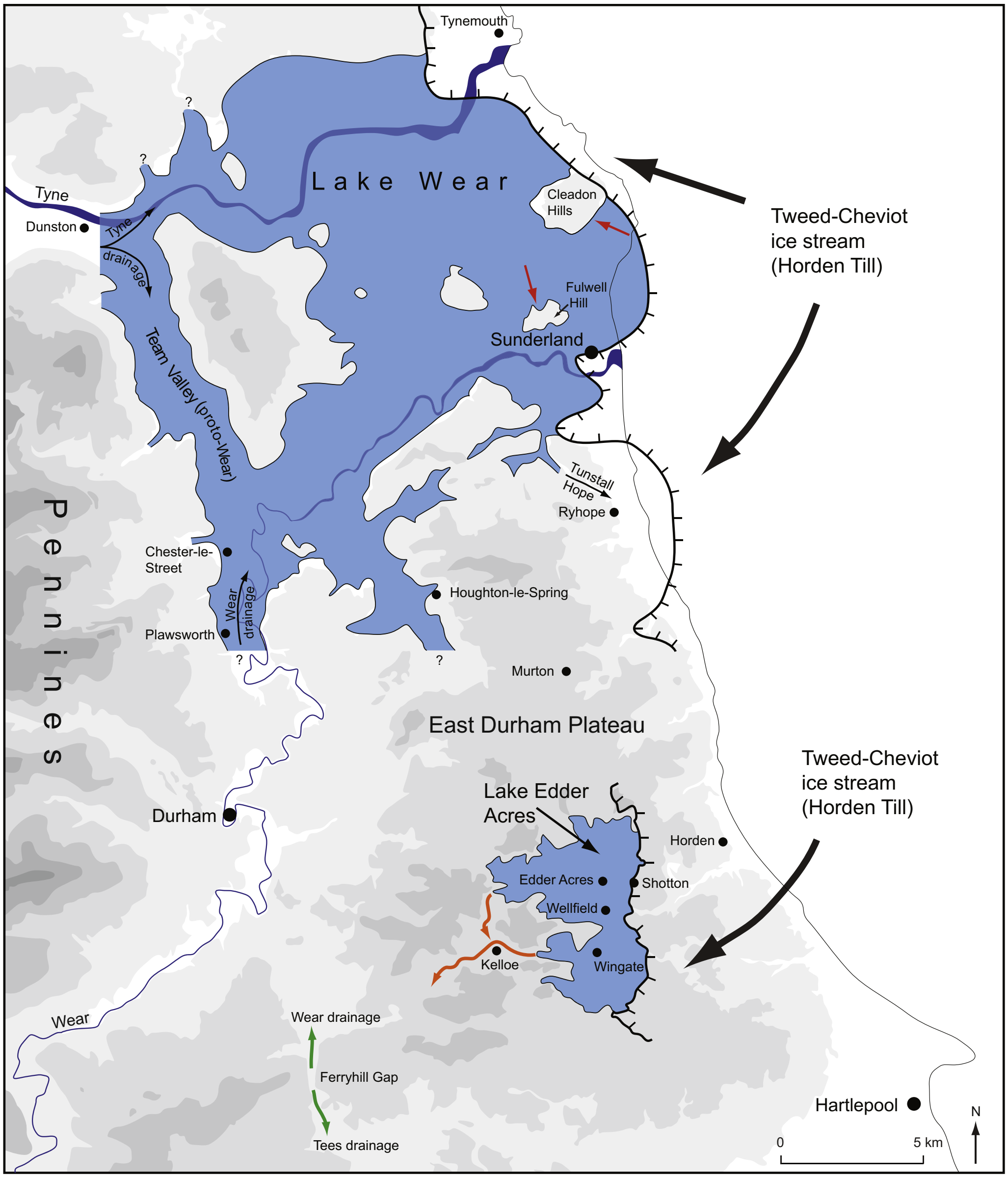

Fig. 13. Generalised extent of glacial Lake Wear at its 43 m OD level and glacial Lake Edder Acres during MIS 2. For both lakes the exact position of the ice margin is unknown, although for Lake Edder Acres it is inferred from the mapped western extent of Horden Till. Red arrows indicate the general location of isolated patches of Tyne–Wear Complex sediments; green arrows show the contemporary watershed – Ferryhill Gap – between the rivers Tees and Wear; orange arrows are inferred overflows. Place names and additional landforms mentioned in the text are shown. Contours are at 50 m intervals, the lightest shading corresponding to 50 m OD. Adapted from Smith (1981), and Smith and Francis (1967). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

downcutting of which continued throughout the retreat of Irish Sea ice, to 90 m OD (Fig. 10). During this time, overflow from Lake Newport was initially through the Oakengates channel (Wills, 1924) at 134 m OD, and subsequently at Gnosall, based on the regressive levels inferred from the terraced sands and clays (Hamblin, 1986). As the Gnosall outlet was slightly higher – at 92 m OD – than the Ironbridge col, so increased drainage resulting from an enlarging Lake Lapworth inevitably exploited the lower outlet, thereby forming the valley in which the contemporary river Severn flows. However, incision of the Ironbridge Gorge can also be explained without invoking a regionally extensive glacial lake. Hamblin (1986) proposed that the Ironbridge Gorge was initiated as an ice-marginal channel when the Irish Sea ice margin was immediately north of the present course of the Severn which subsequently became an overflow channel for Lake Buildwas.

3.2. The Wear Lowlands

3.2.1. Physiographic and stratigraphic context

The [Tyne and] Wear Lowlands encompass the lower valleys of the rivers Tyne and Wear, bounded by Newcastle-upon-Tyne to the north, the Pennines in the west and the East Durham Plateau in the east (Figs. 1 and 13). The region has a distinctive tripartite Late Pleistocene sequence that comprises, from the base upwards: Blackhall Till (formerly Lower Boulder Clay); sands, gravels and finer-grained sediments; and Horden Till (formerly Upper Boulder Clay) (Davies et al., 2009). Locally, the sands, gravels and finer-grained sediments are termed the Tyne–Wear Complex (Smith, 1981), Peterlee Sands (Davies et al., 2009), or Middle Sand (Smith and Francis, 1967). It is these sediments which form the focus of the following two sub-sections.

3.2.2. Evidence for Lake Wear

The Tyne–Wear Complex sediments vary in both their extent and composition. Although they predominately infill the pre-glacial buried valleys of the Sunderland district, they are also recorded up to 70 m OD in the Wear Lowlands, with isolated pockets of potentially similar sediments up to 132 m OD. They exhibit a sharp contact with the underlying Blackhall Till and show an upward transition from laminated silty clays and clayey silts to bedded sands. Locally, they are interbedded with lenses or beds of stony clay (Smith, 1981, 1994).

In detail, the laminated clays are dark brown with discontinuous to continuous films, laminae or thin beds of fine-grained pale brown or grey sand. Commonly their surface is current rippled. The transition from laminated sediments to bedded sand is gradational, interbedded over a vertical distance of 1–2 m. The sands are medium to fine-grained and contain small-scale cross-lamination and current ripples, the latter indicating flow towards the north and east (Smith, 1994). These sedimentary properties suggest that deposition occurred initially in deep water which became progressively shallower (Smith, 1994). The stony clay contains abundant subangular to subrounded pebbles, cobbles and boulders of similar lithologies to the Blackhall Till. It is interpreted as periglacial mudflows deposits derived from Blackhall Till situated above the elevation of the laminated sediments (Smith, 1981).

A silty clay forms the uppermost unit in the Wear lowlands. Locally termed the Pelaw Clay (Smith, 1981) or Upper Wear Clay (Beaumont, 1971), it is differentiated on the basis of colour, and is at least 0.5 m thick but often 1–2 m. Its lower contact is generally sharp, but often gradational with either the underlying laminated sediments or Blackhall Till. It is generally unstratified, but shows occasional weak stratification, with lenses or thin beds of sand. Pebbles and small cobbles of similar lithologies to those in the Blackhall Till are scattered throughout the clay (Smith, 1994). X-ray diffraction analysis demonstrates that the clay mineralogy of the Blackhall Till and Pelaw Clay are similar, with the exception of montmorillonite, which only occurs in the latter (Beaumont, 1971).

Woolacott (1913, 1921) first alluded to an ice-dammed lake in the lower Wear, resulting from retreat of the Tweed-Cheviot ice stream. Detailed geomorphological mapping by Raistrick (1931) elaborated this model to depict this lake – which he named Lake Wear – as having formed by the impoundment of water draining from Lake District-Tyne ice in the north and north-west, Tweed-Cheviot ice in the east, and Tees-Stainmore ice in the south. He inferred an initial lake level at 120 m OD, with a lower phase at 90 m OD. This correlates with known elevations of the Tyne–Wear Complex (Smith, 1994), and a delta at Plawsworth formed as the Wear debouched into the lake (Raistrick, 1931). In addition, a lower level of ca. 43 m OD (Fig. 13) is inferred from isolated patches of Tyne–Wear Complex sediments overlying brecciated Magnesian Limestone bedrock of the Fulwell and Cleadon Hills (Smith, 1981). The origin of the Pelaw Clay is uncertain, but its stratigraphic position suggests that it was deposited after the drainage of Lake Wear, most probably derived from periglacial modification and reworking of the underlying laminated sediments (Smith, 1981).

3.2.3. Palaeohydological implications of Lake Wear

The most significant palaeohydrological roles attributed to Lake Wear are the incision of the Wear Gorge at Sunderland and the dry valley of Tunstall Hope (Fig. 13). The sub-drift topography shows the pre-glacial drainage of the region. In particular, a deeply incised channel underlies the present Wear valley up to Chester-le-Street, where it continues northward to Dunston to merge with the river Tyne (Raistrick, 1931). It is uncertain how the Magnesian Limestone gorge and the concomitant diversion of the Wear to Sunderland were initiated. One possibility is meltwater emanating from the retreating Blackhall Till ice (Trechmann, 1952), or more probably drainage from Lake Wear (Raistrick, 1931).

The dry valley of Tunstall Hope was incised during overflow of Lake Wear (Woolacott, 1921; Smith, 1981) at its 90 m OD level. Evidence for this is derived from the relationship between the valley fill, which comprises laminated sediments in its north-western end and gravels at its southern margin, and sandy gravels at Rythorpe. The latter gravels sharply overlie Blackhall Till, have a similar lithological composition to it, and exhibit imbrication and palaeocurrent orientations that indicate first eastwards and then southwards flow (Smith, 1994). Smith (1981, 1994) interpreted these gravels as remnants of a delta and outwash fan radiating from the mouth of the Hope. But it is difficult to reconcile the juxtaposition of the laminated sediment and gravel within Tunstall Hope and beyond to Rythorpe, because laminated sediments are indicative of quiescent conditions with suspension settling of fine sediment, while coarser grained sediment reflects higher-energy transportation. An alternative interpretation is that ice segregation fractured the frost-susceptible Magnesian Limestone of Tunstall Hope such that seasonal nival floods, enhanced by overflow waters from Lake Wear, thermally eroded the gravels and transported them southwards.

3.2.4. Lake Edder Acres

Farther south, Smith and Francis (1967) mapped significant deposits of laminated sediments in the vicinity of Edder Acres and Wingate, which they interpreted as glaciolacustrine deposits of glacial Lake Edder Acres (Fig. 13). As with Lake Wear, these sediments overlie the Blackhall Till. The contact is sharp, but undulating, so while the general thickness of glaciolacustrine sediment ranges from 7.0 to 12.5 m, a maximum thickness of 29.5 m is recorded at Wellfield.

In detail, the sediments occur no higher than 130 m OD and are typified by exposures at Shotton (Smith and Francis, 1967). The sediments here show a progressive eastward increase in texture from thinly laminated to interbedded sands and clays. In western exposures, 7.0 m of laminated brown and dark brown clays with thin lenses and beds of rippled, brown fine sand are overlain by 0.6 m of dark brown laminated clay, and 0.75 m of homogeneous dark brown clay. In the eastern exposures, the lowermost laminated unit is thicker (12.5 m) and sandier. It is overlain by 4.4 m of interbedded sand and clay, the sand beds of which show current ripples and foreset bedding, indicating a westwards flow. The laminated sediments are generally overlain by Horden Till. It is likely that Lake Edder Acres had a negligible palaeohydrological influence, although its drainage through the Magnesian Limestone escarpment at Kelloe resulted in a tributary to the contemporary Wear.

3.3. The Lower Tees Basin

In the Lower Tees Basin, near Middlesbrough, clay has been observed at two elevations. The higher clay – termed gutta-percha by Gayner and Melmore (1936) – is massive and crops out at ca. 90 m OD (Radge, 1939). It has yet to be firmly established if this clay is glaciolacustrine and indicative of a high-level phase of Lake Tees, or represents remnants of till from an earlier glacial stage. The lower clay is more extensive, occurs up to 16.5 m OD (Fig. 14), and sharply overlies a red till with a thin gravel lag at its contact. Typically, its thickness ranges from 4 to 8 m to the south of the River Tees, up to 15 m to the north of it. Its lamination is locally variable, with the coarsest lamination, sometimes wavy or interbedded with sand, occurring in the middle of the deposit, or towards its margins. The deposit commonly fines upwards, becoming massive (Radge, 1939; Agar, 1954). The lake margins are characterised by a bench, 180–270 m wide, incised into the red till

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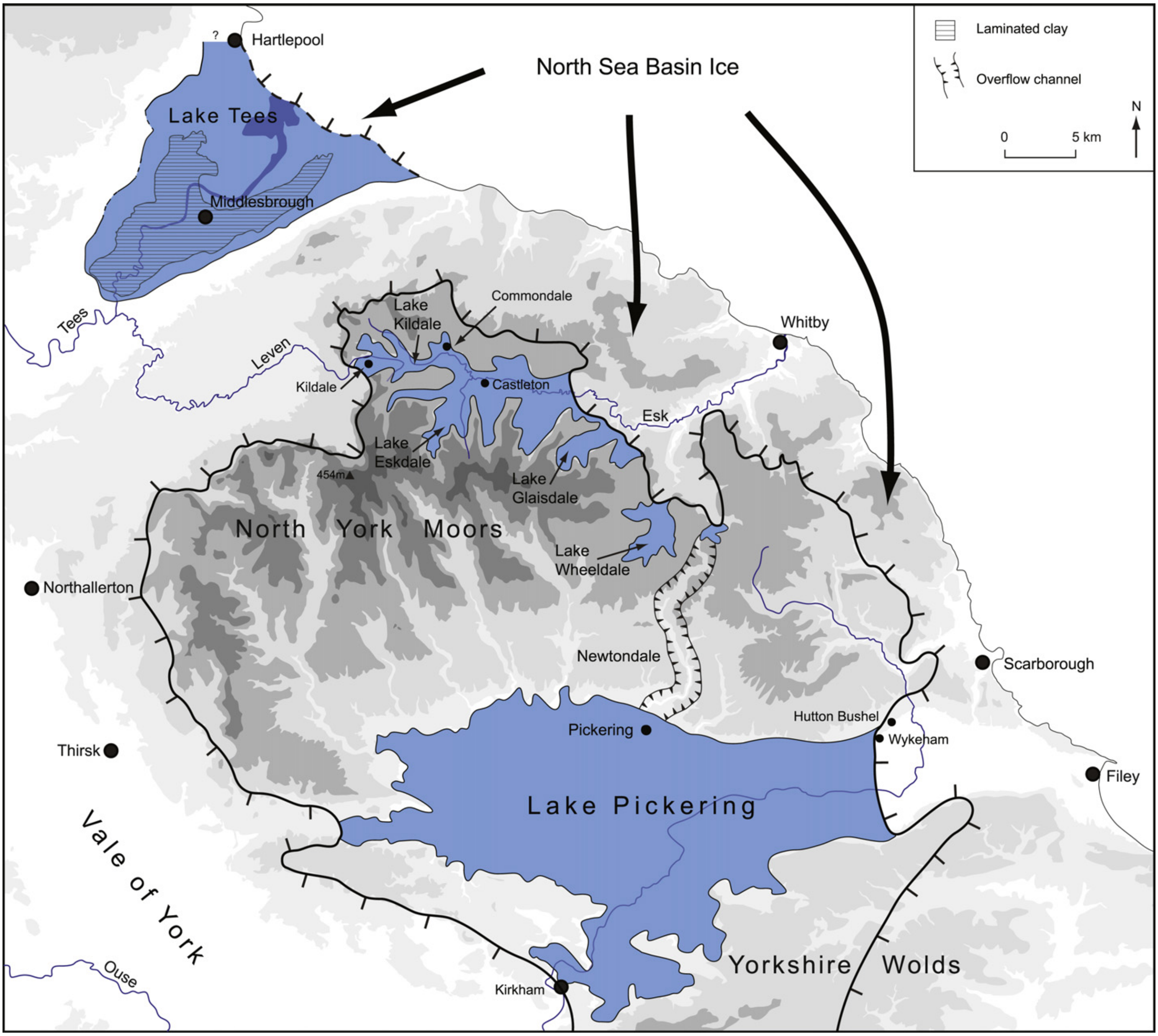

Fig. 14. Generalised extent of glacial Lake Tees, glacial Lake Pickering, and glacial lakes in the North York Moors during MIS 2. The ice margin impounding lakes in the Vale of Pickering and within the North York Moors is the maximum extent of the Late Devensian ice sheet. The configuration of the British Ice Sheet during emplacement of the laminated sediments in the lower Tees Basin is uncertain, and is tentatively placed adjacent to the present-day coastline. Glacial Lake Tees is delimited from a shoreline mapped at 25 m OD, dashed lines denoting uncertainty. Contours are shown at 50 m, 100 m, and at 100 m intervals thereafter. Adapted from Kendall (1902), and Agar (1954).

at ca. 25 m OD. It is filled with up to 3 m of yellow sand, often interdigitating with the laminated sediments.

In detail, the sand–clay couplets of the laminated deposits are usually very thin and frequently indistinct or contorted. Three types of laminae are identified: basal light brown silty sand, reddish brown silty clay and black to dark grey clay (Agar, 1954; Plater et al., 2000). The sand laminae have a rippled surface and the contact between the upper clay lamina and the succeeding silty sand lamina is sharp (Agar, 1954).

Radge (1939) hypothesized that both levels of Lake Tees developed during the Late Devensian deglaciation, following more rapid recession of Pennine ice whilst North Sea Basin ice impeded drainage. At its higher level (ca. 90 m OD) he envisaged Lake Tees merging with Lake Humber and possibly Lake Wear also, with drainage into the latter through the Ferryhill Gap in the Permian escarpment (Trechmann, 1920). Continued eastwards retreat of North Sea Basin ice and concomitant drainage of Lake Humber enabled the formation of low-level (ca. 25 m OD) Lake Tees (Fig. 14), the thickness of deposits indicating that this was a prolonged phase (Radge, 1939). However, the description of wavy or indistinct lamination is similar to that observed in the Vale of York (Murton et al., 2009) (Section 3.5), and it can be speculated that Lake Tees may also have been characterised by fluctuating lake levels rather than a relatively stable one.

3.4. The North York Moors and the Vale of Pickering

In the North York Moors, the most extensive glacial lake was Lake Eskdale (Kendall, 1902; Kendall and Wroot, 1924), formed as the rivers Esk and Leven were impounded by North Sea Basin ice to the east, and Pennine ice occupying the Vale of Mowbray to the west (Fig. 14). This lake is unusual in that evidence used to infer its former existence is predominately derived from channels incised into the underlying bedrock rather than the distribution of glaciolacustrine sediments. In a seminal paper, Kendall (1902) interpreted these features as overflow channels and used their elevations to reconstruct ice-margin recession and hence proglacial lake development. From this, the maximum reconstructed level of Lake Eskdale was 225 m OD, corresponding with the lowest point in the Cleveland anticline at Newtondale, through which the lake overflowed into the Vale of Pickering. With continued incision of Newtondale to 218 m OD (height of the current sill in the valley) Lake Eskdale fragmented into a series of smaller lakes (Kildale, Wheeldale and Glaisdale) occupying the tributary valleys (Fig. 14). Gregory (1965), however, re-interpreted these channels as subglacial rather than subaerial lake overflows.

In addition to overflow channels, Kendall (1902) also identified sedimentary evidence to support his glacial lake hypothesis. In particular, he interpreted relatively flat-topped gravel accumulations at confluences of tributary valleys with the Esk as deltaic. One such example occurs at Commondale, where a gravel plateau at ca. 165–173 m OD composed of locally derived Jurassic sandstone and erratics such as Shap Granite reflects debouchment from Sleddale Beck. At the western end of Lake Eskdale, at Castleton, a similar feature is identified from the debouchment of Ewe Crag Slack (Kendall and Wroot, 1924). At the confluence of Ewe Crag Slack and Black Beck two gravel terraces occur at 180 and 173 m OD, which are interpreted as representing two levels of Lake Eskdale (Kendall, 1902). Critically, no laminated, fine-grained sediments have been observed for Lake Eskdale, or its subsidiary lakes, despite the formation of deltas and proximity of ice. Without chronological control it is not yet possible to constrain the age of deposition and so these terrace features may be pre-MIS 2.

Farther south, between the North York Moors and Yorkshire Wolds, is the Vale of Pickering. Phillips (in Gayner and Melmore, 1936) alluded to the existence of a glacial lake here, inferring that morainic debris blocking the eastern end of the vale impounded drainage of the proto-Derwent (Fig. 14). A detailed review of the conceptual development of Lake Pickering is presented in Foster (1985).

Geomorphic evidence delimiting Lake Pickering is sparse. On the northern margin at Hutton Buschel (Fig. 14), a prominent terrace of sand and gravel at 60 m OD has been interpreted as littoral (Fox-Strangways, 1892) or deltaic (Kendall, 1902) in origin. Farther east at Pickering, a single fan-shaped gravel deposit of rounded, locally derived lithologies is also considered deltaic, with sediment derived from waters draining from Lake Eskdale southwards through Newtondale (Kendall, 1902). Fox-Strangways (1892) noted less prominent gravel terraces at 42 and 30 m OD, which he attributed to reduced incision of an overflow channel at Kirkham Abbey through which Lake Pickering drained into the Vale of York. No comparable deposits were observed on the southern margin, which Kendall (1902) attributed to streams of low competence draining the chalk of the Yorkshire Wolds.

Within the inter-connecting glacial lake system proposed by Kendall (1902) only Lake Pickering retains sedimentary evidence, derived principally from borehole records. These show laminated or massive silts and clays overlain by interbedded sands and laminated sediments (Kendall and Wroot, 1924; Foster, 1985). The

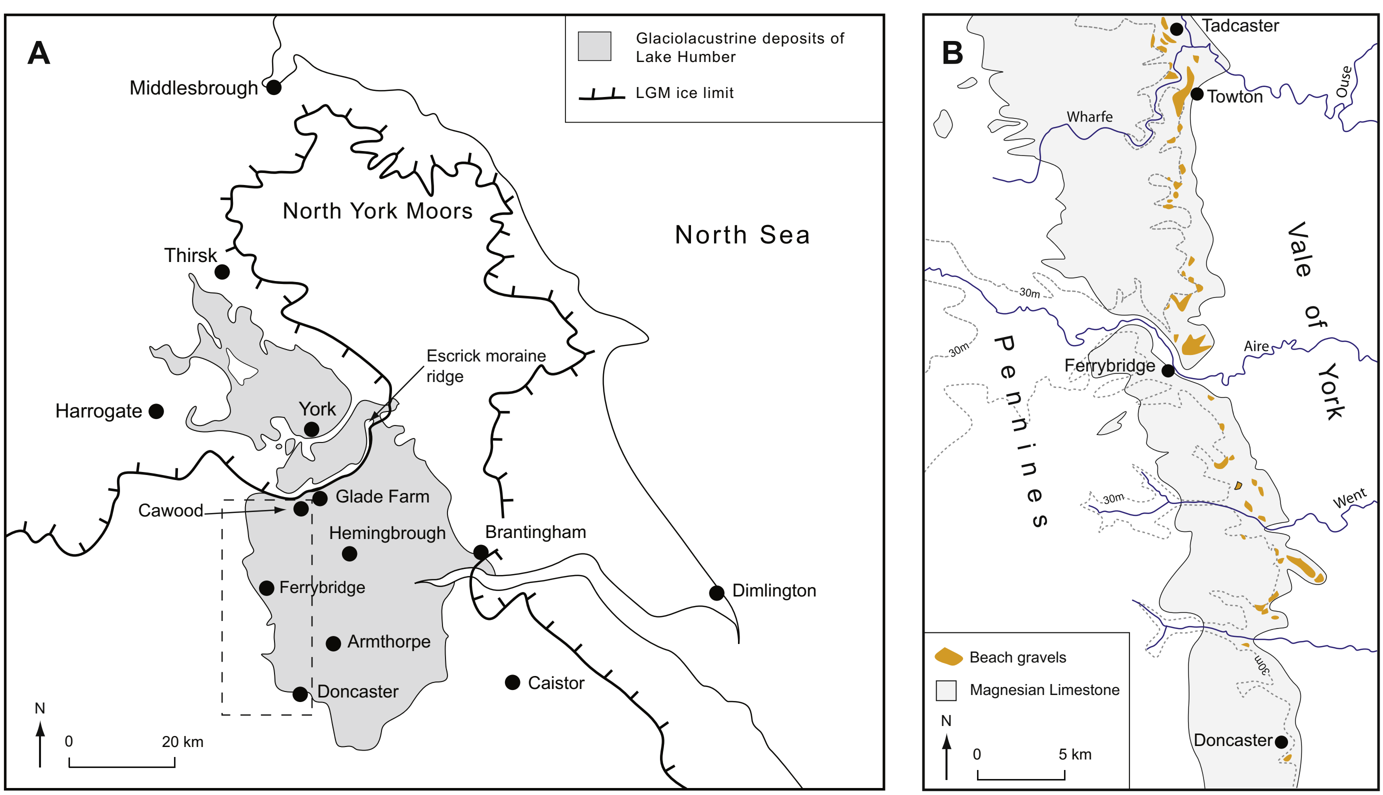

Fig. 15. (A) Distribution of glaciolacustrine sediments deposited in glacial Lake Humber. The upper part of the glaciolacustrine sequence is of MIS 2 age and accumulated when the Vale of York ice was at its southern limit at the Escrick moraine ridge. Dashed box denotes location of (B). Topographic setting of Lake Humber is shown in Fig. 16. Adapted from Murton et al. (2009). (B) Distribution of discontinuous patches of gravels at ca. 30 m OD on the western fringes of the Vale of York. Adapted from Edwards (1937).

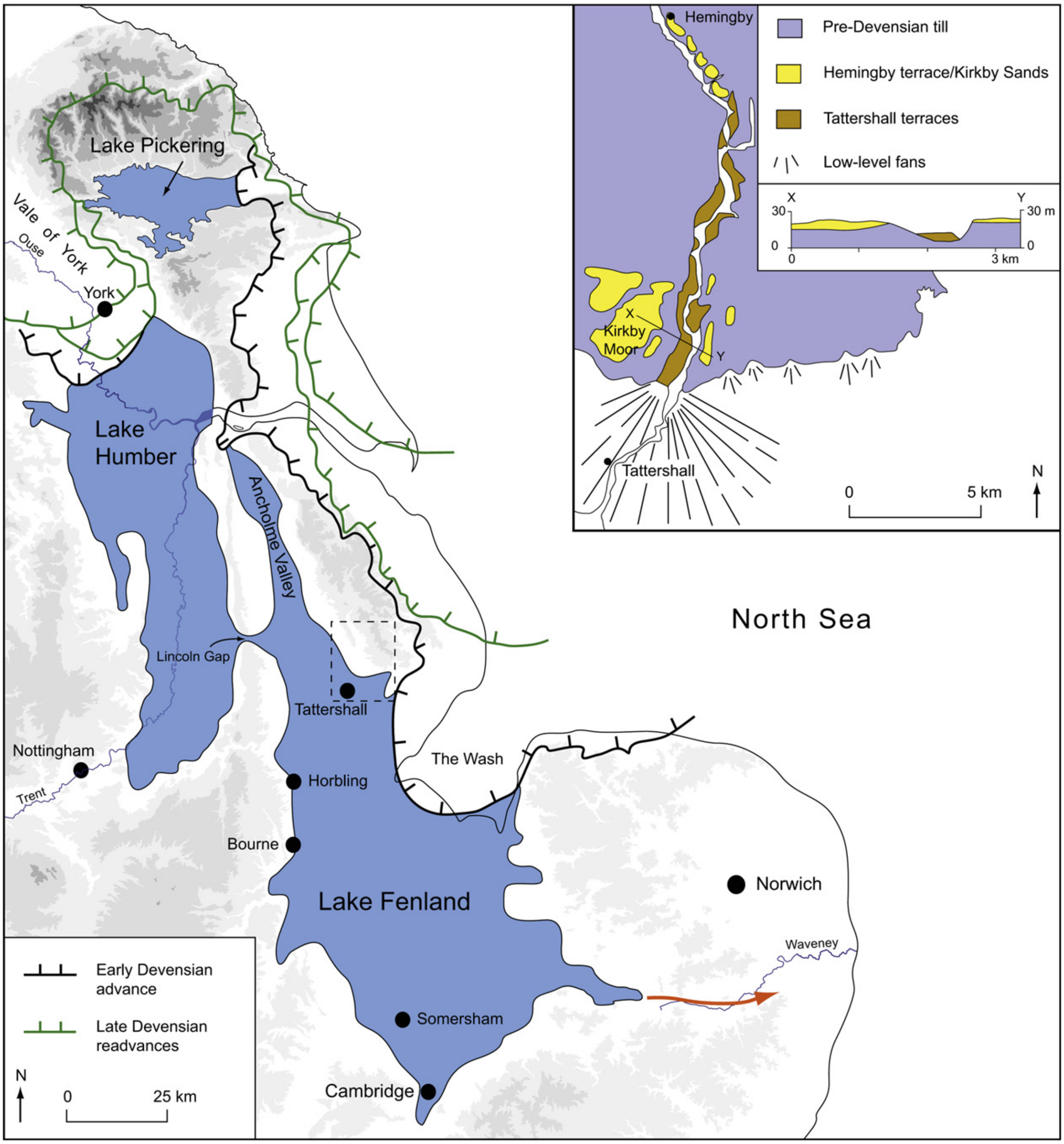

Fig. 16. Potential extent of glacial Lake Humber during its high-level (ca. 30 m OD) phase, and the coeval glacial Lake Fenland during MIS 4. Orange arrow denotes drainage eastwards into the Waveney valley. Inset box shows stratigraphic relationship between inferred MIS 6 deposits and gravels containing organic lenses in the lower Bain valley, Lincolnshire. Contours are shown at 50 m, 100 m, and at 100 m intervals thereafter. Adapted from Straw (1979). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

basal unit ranges in thickness from 33 m to 4 m, its variability in part determined by a buried valley mapped within the vale (Foster, 1985). The upper unit is thinner, often just a few metres, but thickens towards the west where it crops out up to ca. 30 m OD.

Palaeoenvironmental reconstructions of Lake Pickering vary. Gayner and Melmore (1936) considered its lake levels to be up to 120 m OD, merging with Lake Humber across the Howardian Hills to form a single lake. The prevailing interpretation, however, is that Lake Pickering was an isolated feature, possibly part proglacial and part subglacial (Foster, 1985, fig. 89). It may have existed with two separate lake-level phases (Kendall and Wroot, 1924), although the sedimentary evidence is consistent with one at least one at ca. 30 m OD. Furthermore, the laminated clays overlying the gravel fan at Pickering suggest that Lake Pickering continued to receive water after Newtondale ceased to operate as an overflow channel for Lake Eskdale (Kendall and Wroot, 1924).

3.5. *The Vale of York*

To the south of Lake Pickering, Lake Humber (Lewis, 1894) formed in the central and southern parts of the Vale of York, south of the Escrick moraine ridge (Figs. 15 and 16). Lake Humber developed when North Sea Basin ice dammed the Humber Gap, ponding meltwater draining from ice occupying the Vale of York and Pennines (Fig. 15A). The concept of a two-phase Lake Humber was promulgated by Gaunt (1976), who argued for a brief, high-

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level phase at ca. 30 m OD, with a subsequent prolonged, low-level phase at ca. 8 m OD. The two phases were separated by lake drainage and subaerial exposure of the lake floor. Drainage of the lake is considered to have occurred eastward through the Humber Gap and intermittently southeastward through the Lincoln Gap.

The high-level phase is inferred from discontinuous patches of rounded, but predominately locally derived sands and gravels at ca. 30 m OD on the Magnesian Limestone escarpment of the eastern Pennines (Edwards, 1937) (Fig. 15B). As all observations of these localised gravels occur south of the Escrick moraine ridge, the overriding interpretation is that they represent a former beach of Lake Humber (Gaunt, 1976; Bateman et al., 2008). More recently, Murton et al. (2009) suggested that given their proximity to hillslope surfaces, these deposits may have been reworked by periglacial processes. On the eastern margin of the Vale of York, distal terminations of interpreted alluvial fans and planar surfaces at ca. 30 m OD etched into the underlying Triassic bedrock are also inferred to represent a former shoreline of Lake Humber (Fairburn, 2011). These fans, and others emanating from the valleys of the Yorkshire Wolds which terminate at ca. 20 m OD, constitute the Pocklington Gravel Formation (cf. Ford et al., 2008). The low-level phase is recognised by up to 24 m of sands and laminated silts and clays – termed the Hemingbrough Glaciolacustrine Formation (Ford et al., 2008) – which form the basin fill in the Vale of York. Detailed in Murton et al. (2009), the laminated sediments of the uppermost 10 m are characterised by wavy parallel, planar parallel or lenticular lamination. They most commonly form couplets of light-coloured silt or fine sand either grading into, or sharply underlying, dark-coloured clayey silt. The most prominent lenticular lamination is characterised by round-crested to peaked symmetrical ripple form sets of silt or sandy silt interspersed within clayey silt.

In the centre of the basin, at Hemingbrough, the laminated sediments are overlain by dark greyish brown to brown massive silt that is brecciated and contains distinctive sandy interbeds or laminae. These sandy beds increase in thickness up the sedimentary sequence and commonly have well developed ripple form sets with symmetrical, straight to slightly sinuous crests that have peaked or rounded forms. These ripple forms have silt drapes forming flaser bedding. At Glade Farm, ca. 1 km south of the Escrick moraine ridge, the laminated sediments are gradationally overlain by massive clayey silt that contains abundant slickensides. This unit is sharply overlain by massive silt with similar texture and sedimentary structures as at Hemingbrough. It grades upwards into medium to fine-grained sand, interbedded with clayey silt in the lowermost 30 cm. At Hemingbrough the uppermost unit is an overconsolidated laminated to massive clayey silt.

While Carruthers (1947) interpreted the laminated sediments as shear clays deposited sub-glacially by undermelting of a stagnant or decaying ice sheet, the conventional interpretation infers quiescent rainout deposition in a standing body of water (Gaunt, 1976, 1994). Although ripple structures had previously been recorded within the laminated sediments (Gaunt, 1994), their precise sedimentary characteristics and significance remained unexplored. Murton et al. (2009) have clarified that wave ripples are interbedded within both the laminated clayey silts, and overlying silty sand beds. Critically, these structures suggested that Lake Humber, in its latter stages, was characterised by shallow, fluctuating lake levels that were basin-wide, rather than shallow-water conditions being restricted to the basin margins. Furthermore, as the wave-rippled silty sands become more prominent towards the top of the Lake Humber deposits, they probably record the southward advance of Vale of York ice into Lake Humber (Fig. 15A).

3.6. The Fens

The largest putative former glacial lake in England is Lake Fenland, which extended from the Fenlands northwards into the Ancholme Valley (Raistrick, 1934; Straw, 1979), deriving additional

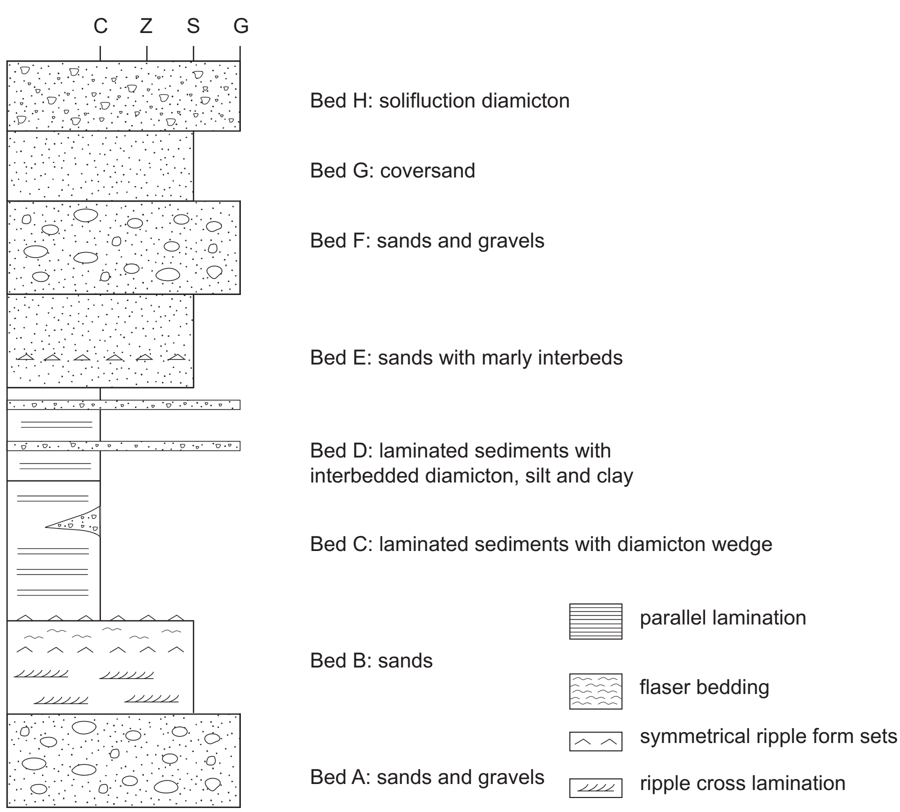

Fig 17. Schematic representation of the sedimentary sequence exposed at Somersham, Cambridgeshire. After West et al. (1999).

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drainage from Lake Humber through the Lincoln Gap (Fig. 16). More recent palaeoenvironmental reconstructions using a digital elevation model of topography also conjoin Lake Pickering via the Kirkham Gorge at Kirkham (Fig. 14). Collectively, this combined body of water is computed to have had a volume of 257 km^3, a maximum depth of 31 m, and an average depth of 20 m (Clark et al., 2004).

Evidence for Lake Fenland is extremely limited. The only directly attributable evidence is a narrow bench at 25–32 m OD observed on the western margin around Horbling and Bourne (Harrod, 1972). More commonly, its spatial extent is derived from extrapolation of the ca. 30 m OD, high-level phase of Lake Humber (Straw, 1979), which necessitates three assumptions (Clark et al., 2004): (1) North Sea Basin ice concurrently blocked the Humber gap and Wash Basin; (2) overflow from the high-level phase of Lake Humber was prevented from draining through the Lincoln gap by the presence of Lake Fenland; and (3) lakes Humber and Fenland existed in equilibrium at ca. 30 m OD.

However, on the southern margin of the Fens, at Somersham, sedimentary evidence for a former glacial lake is more substantive. Here, a channel fill comprising sands overlain by laminated sediments is incised into gravel (West, 1993; West et al., 1999); the full stratigraphic sequence is shown in Fig. 17. While the sands are interpreted as fluvial, the laminated sediments were deposited during the subsequent impoundment of a lake – termed Lake Sparks (West, 1991) – by the advance of North Sea ice into the Wash. Since both Lake Fenland and Lake Sparks formed as North Sea ice in the Wash impounded fluvial drainage it is possible that both names actually refer to the same water body (cf. Clark et al., 2004). But a re-evaluation of the sedimentary evidence for Lake Sparks suggests that this is unlikely, not least because the incised channel at Somersham is just 95 m wide, so hardly indicative of a regionally extensive feature.

Of the eight sedimentary units identified at Somersham (Fig. 17), Beds B–E can be interpreted as glaciolacustrine. The lowermost unit, Bed B, comprises sand, which in its upper part shows ripple structures overlain by silt and clay drapes forming flaser bedding. These ripples have a wavelength of 8–12 cm, a height of 2 cm and are orientated east–west. Significantly, the final ripple drift of this bed, which has peaked symmetrical crests and possible ripple form sets, is sharply overlain by laminated sediments (Bed C) (West, 1993; Fig 5).

The implications for this are twofold. First, the peaked symmetrical ripples characterise very shallow or near-emergent conditions associated with wave action (Collinson et al., 2006), which is inconsistent with the fluvial interpretation postulated by West et al. (1999), where near-bottom currents would prevail. Furthermore, the preservation of ripple sets sharply overlain by laminated sediments implies that deposition was near continuous, otherwise subaerial exposure would have obliterated the ripple structures. Higher up the sequence, the occurrence of ripple-bedded sand (Bed D), and graded sand, often with flaser bedding, interbedded with progressively thinner massive marly beds (Bed E) suggests continuation of shallow-water conditions.

If the glaciolacustrine sediments of Lake Sparks are a remnant from Lake Fenland, then it is difficult to reconcile the sedimentary evidence outlined above with the deep-water (ca. 30 m) environment postulated for Lake Fenland. In the Vale of York, symmetrical ripple form sets and flaser bedding are attributed to the low-level phase of Lake Humber (Murton et al., 2009). An alternative interpretation, proposed here, is that Lake Sparks and the low-level phase of Lake Humber were more or less coeval, albeit as separate features. At present, the co-existence of Lake Fenland with the high-level phase of Lake Humber remains to be substantiated with firm geological evidence.

3.7. Chronology of glacial lake formation

Robust chronological control on Pleistocene glacial lakes in Britain is sparse, more often inferred from stratigraphic relationships with known ice limits or from organic material within under- or overlying units. More recent applications evaluating the potential for developing annually resolved chronologies will be discussed in the next section. Of those lakes discussed above, the chronological framework is strongest for Devensian lakes, particularly for Lake Humber.

Three ages constrain the high-level phase of Lake Humber (Fig. 15A). On the eastern margin at Brantingham, a maximum age of $26,163 \pm 2046$ cal BP is inferred from a bone fragment reworked into or below the base of littoral deposits (Gaunt, 1974), while farther south at Caistor, thermoluminescence dating on beach sand returned an age of 22.7 ± 1.4 ka (Bateman et al., 2000). In contrast, comparable littoral deposits from the western margin at Ferrybridge, returned an OSL age of 16.6 ± 1.2 ka (Bateman et al., 2008). This age though is inconsistent with both the earlier determinations, and with ages of $22,043 \pm 497$ and $21,734 \pm 372$ cal BP from moss within the Dimlington Silts, which underlie the Skipsea Till at Dimlington, in Holderness (Penny et al., 1969). If this 16.6 ± 1.2 ka age is accepted as indicating beach deposition, then it identifies a previously unrecognised secondary high-level phase, with significant implications for ice dynamics in the North Sea and Humber Gap, for Lake Fenland and the subsequent low-level phase of Lake Humber.

Although not directly addressing this issue, an alternative model was proposed by Straw (1979) (Fig. 16), whereby the high-level phase of Lake Humber – and so Lake Fenland – existed during the Early Devensian (MIS 4–3 transition). In the lower Bain valley, around Kirkby Moor, cross-bedded sands with thin beds of flint gravels occur at ca. 30 m OD. Interpreted as deltaic, these sediments are underlain by till of postulated MIS 6 age. Approximately 4 km farther south at Tattershall, gravels situated at a lower elevation than the Kirkby Moor sediments form fan-like features or low terraces grading up the Bain valley. As organic remains from clayey and silty lenses within the Tattershall gravels yielded ages ranging from $36,167 \pm 377$ cal BP to $>46,557 \pm 888$ cal BP, it is inferred that the Kirkby Moor sands and gravels are older. In common with the Kirkby Moor sediments and correlative Hemingby Terrace, littoral deposits on the western margin of the Vale of York are preserved on interfluves, suggesting they too may be older than MIS 2.

Until recently, the low-level phase of Lake Humber was constrained by three minimum ages derived from peat at or near the base of blown sand overlying the glaciolacustrine sequence. These yielded ages of $12,965 \pm 180$ cal BP from Armthorpe (Gaunt et al., 1971), $12,879 \pm 168$ cal BP from near York (Matthews, 1970) and $12,398 \pm 109$ cal BP from Cawood (Jones and Gaunt, 1976), all coincident with, or slightly preceding, Greenland Stadial 1 (the Younger Dryas Stadial). Murton et al. (2009) pioneered the application of OSL dating to shallow-water wave ripples in order to provide the first direct chronological determination for the low-level phase of Lake Humber. Two wave-rippled sand beds at Hemingbrough returned ages of 21.0 ± 1.9, 21.9 ± 2.0, and 24.1 ± 2.2 ka (weighted-mean age 22.2 ± 0.5 ka), firmly constraining the low-level phase within the Last Glacial Maximum (LGM) Chronozone ca. 19.0–23.0 ka (Mix et al., 2001). Given these chronological constraints, it seems improbable that the underlying ca. 20 m of glaciolacustrine sediments were emplaced during the relatively short-lived high-level phase.

It was suggested previously that Lake Sparks and the low-level phase of Lake Humber were coeval. At Somersham, *Salix* leaves from drift mud in an adjacent correlative channel fill to that occupied by sediments of Lake Sparks yielded an age of $21,820 + 2409$ cal BP to $21,800 - 1746$ cal BP (West, 1993). This is consistent with the weighted-mean OSL age 22.2 ± 0.5 ka from Hemingbrough (Murton et al., 2009). It is probable, therefore, that Lake Fenland and Lake Sparks were two temporally and spatially distinct glacial lakes, the former developing during the Early Devensian, and latter during the LGM. However, a critical prerequisite of a high-level Lake Humber is the existence of Lake Fenland to prevent overflow through the Lincoln Gap. Murton et al. (2009) suggested that the discontinuous sands and gravels indicative of this lake phase may reflect deposition or reworking by colluvial rather than littoral processes. This interpretation is compatible with the paucity of laminated sediments within the postulated Lake Fenland limits, implying that Lake Fenland is conceptual.

Elsewhere, OSL dating applied to silty sand laminae from Lake Tees returned an age of 18,365 ± 10,015, but it is unreliable given the significant error, attributed to issues with water-content reconstruction and variability to sensitivity in radiation dose (Plater et al., 2000). From sedimentary analyses Davies et al. (2009) correlated the Holden Till in Durham with the Skipsea Till in Holderness. As the Holden Till commonly overlies the glaciolacustrine deposits of Lake Wear their age is broadly LGM.

4. Establishing a varve chronology from glaciolacustrine sediments in Britain

Glaciolacustrine sediments can potentially provide a high-resolution record of palaeoenvironmental change, and their rhythmicity can sometimes be used as a dating tool (Gilbert, 2003). In Britain, stratigraphic variations in varve thickness have been correlated with the Late Devensian GISP2 $\delta^{18}O$ record (Plater et al., 2000), used to determine rates of ice-margin recession in the Brecon Beacons, Wales (Palmer et al., 2008), the duration of lake levels in Glen Roy, Scotland (Palmer et al., 2010) and, supplemented with radiocarbon ages, to constrain the timing and duration of the maximum extent of ice at Loch Lomond, Scotland (MacLeod et al., 2010).

4.1. General principles

Fundamental to any geochronological or palaeoenvironmental application is determining whether a rhythmic sequence of alternating coarse (sand/coarse silt) and fine (fine silt/clay) layers – together constituting a couplet – records annual or non-annual deposition. Where sediment input into a glacial lake is dominated by inter- and overflows in summer, the couplets tend to exhibit sharp contacts, both at the base and the top of the coarse layer (Smith, 1978; Smith and Ashley, 1985; Leonard, 1997); such couplets are interpreted as clastic varves. The coarse-grained sediment tends to be deposited as a result of large inflows of water and sediment and vigorous circulation in summer, whereas the fine-grained sediment is deposited in low-energy conditions (usually in winter) when inflow to lakes tends to be small (Gilbert, 2003). The discrimination of such classic glaciolacustrine varves, however, can be ambiguous where a graded or diffuse contact is present between the coarse layer and overlying fine layer of a couplet (Desloges, 1994) and the varves are attributed to turbidity currents/underflows (Gilbert et al., 1997).

Turbidity currents are a form of sediment gravity flow in which sediment is held in suspension by fluid turbulence (see reviews by Middleton, 1993; Meiburg and Kneller, 2010). Sediment concentrations are low (<9%), which permits development of graded bedding (Postma, 1986). In summer, turbidity currents commonly transport sediment into lakes and deposit numerous graded laminae (Gilbert, 2003). Turbidity currents encompass a range of flow behaviours from very short duration surge and surge-like turbidity flows, to quasi-steady turbidity currents operating at timescales from hours to months (Mulder and Alexander, 2001). It is argued that collectively their deposits should be termed turbidites (Mulder and Alexander, 2001; Shanmugam, 2002), where each turbidite represents a single depositional event. Turbidites are identified in the sedimentary record by a graded bed associated with waning flow.

Rhythmite type	Varve	Surge deposit
Sediment dispersal mechanisms	Suspension settling in winter Overflows-interflows Underflows in summer Surge currents	Slump-generated surge current
Time for depositing each rhythmite	1 Year	Minutes as deposition occurs as a single event
Silt/clay contact within a couplet	Sharp	Gradational
Thickness of silt and clay layer at a given site	Silt layer varies dependent upon effectiveness of sediment dispersal mechanism. Clay layer is relatively constant.	Both silt and clay thicknesses vary in proportion
Bedding	Silt layer may contain multi-laminated micrograded beds, but as a whole unit not necessarily graded. Clay layers frequently graded	Normal grading through the whole couplet.

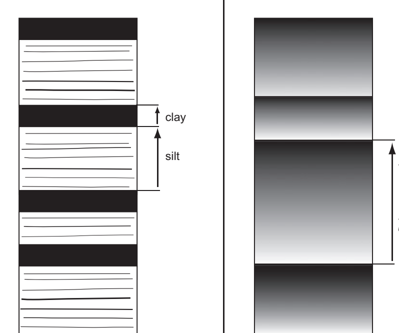

Fig. 18. Sedimentary criteria used to distinguish between turbidites and varves. After Smith and Ashley (1985).

The issue of nomenclature is of greater concern in glacial lakes where density-driven underflows prevail and sediment deposition can occur from traction and suspension settling. The underflows are derived either from quasi-continuous currents originating from sediment-laden river water, or from slump-generated currents triggered by mass-movement processes (Smith and Ashley, 1985). These processes are likely comparable with the respective quasi-steady turbidity currents and surge-like turbidity flows of Mulder and Alexander (2001). As a varve consists of two texturally and genetically distinct units (Ashley, 1975) it cannot represent a single period of sedimentation.

4.2. Applications

To clarify the subjective nature of varve identification, Smith and Ashley (1985) devised empirical criteria by which rhythmites deposited by slump-generated currents could be distinguished from varves (Fig. 18). Importantly, varves have a sharp contact between the silt layer and overlying clay layer (i.e., *within* each couplet), whereas in surge-generated deposits the contact is gradational. In both end members the contact *between* succeeding couplets is sharp.

Adopting a more quantitative approach, Hart (1992) devised the rhythmite index (R) to compare relative thicknesses of silt and clay layers of a couplet:

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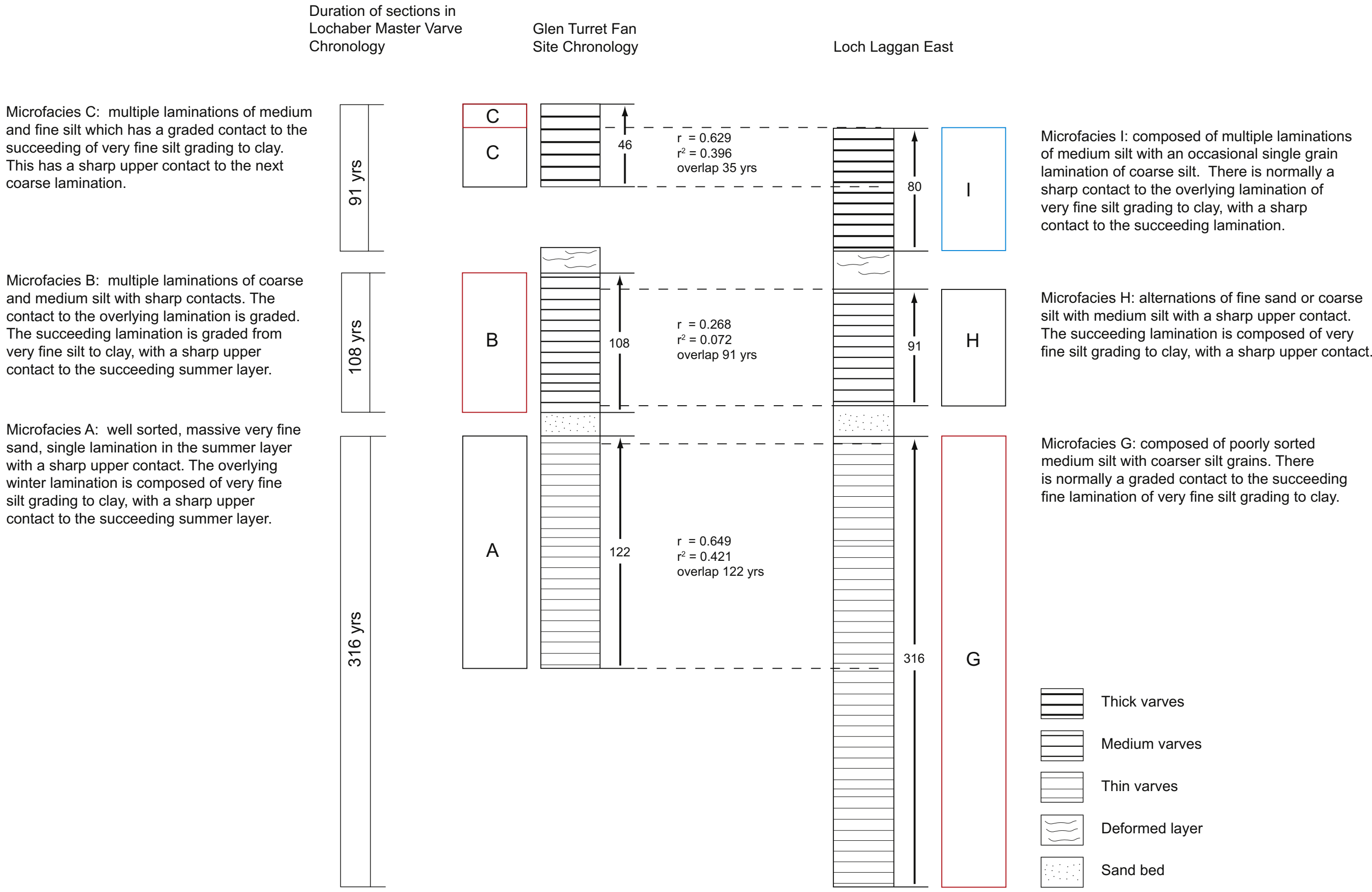

Fig. 19. Master varve chronology derived by wiggle-matching varve thickness records from Glen Roy (Glen Turret Fan data) and Glen Spean (Loch Laggan East data) in Lochaber, western Scotland. Dashed lines show optimal cross-matches based on Pearson's product moment correlation (r), with corresponding r^2 values. Solid black lines show limits of a particular varved section, the numbers representing varve counts. Microfacies descriptions are given for each section, with microfacies G, B, I and C used to produce the master varve chronology. For these four microfacies, the red boxes denote structures with a graded silt-clay contact, whilst the blue box denotes a sharp silt-clay contact. Adapted from Palmer et al. (2010). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

$$R = C_N/(C/Z)_N$$

where C_N = clay-layer thickness, and $(C/Z)_N$ = clay/silt ratio.

$$C_N = \sum \sqrt{(i_N - x_N)^2/n} \times 100$$

where i_N = normalised values of clay thickness, x_N = normalised average value of clay thickness, and n = no of samples.

$$(C/Z)_N = \sum \sqrt{(j_N - y_N)^2/n} \times 100$$

where j_N = normalised values of clay/silt thickness, and y_N = normalised average value of clay/silt thickness.

R values <1 denote a varve, while R values >1 indicate a turbidite. The key determinant is clay-layer thickness, which for turbidites is proportional to the strength of the underflow. It is not apparent whether Hart's (1992) turbidite nomenclature refers specifically to a surge-like turbidity flow or a quasi-steady turbidity current, but is implicit that R values >1 indicate non-annual rhythmites.

In Britain, the first varve thickness record, utilising the R index, was determined from laminated sediments deposited in Lake Tees (Plater et al., 2000). Despite indicating a predominant waning flow or turbidite origin, Plater et al. (2000) noted similarities in sedimentary properties to contemporary laminated sediments deposited near a fluvial inlet, interpreted as varves (Smith, 1978), and so adjusted their interpretation accordingly. From this a tentative correlation to the GISP2 $\delta^{18}O$ record (cf. Grootes et al., 1993) was proposed. However, the facies model proposed by Smith (1978) was derived from Hector Lake, Canada, where documented underflows are rare and sediment is transported in suspension by over- and interflows. As the R indices for Lake Tees are indicative of density-driven underflow currents, it is perhaps inappropriate to interpret the sediments as varves without demonstrating more clearly that the observed cyclicity is annual.

A critical component of the R index is the precise delineation of the silt-clay contact. At the macroscale, as undertaken by Hart (1992) and Plater et al. (2000), this is likely to produce analytical errors, particularly where the contact is gradational, or in relatively thin laminae. Larsen and Stalsberg (2004) demonstrated that selective thin-sections can validate interpretations derived from the R index, highlighting the potential for microscale analysis of glaciolacustrine sediments.

Of wider application is the derivation of a microfacies classification, developed by Ringberg and Erlström (1999) and substantiated by Palmer (2005), to define annually resolved records. In summary, this method uses overlapping thin-sections which are photographed, with lamina thickness and structure determined by digital image-analysis software. Microfacies are distinguished according to grain size and mineralogy, sorting, comparative thickness of silt and clay laminae, occurrence of anomalous grains and presence of plasmic fabric within the clay component (Palmer, 2005). The microfacies identified are currently site-specific, but have been applied to Late Pleistocene glaciolacustrine sediments in the Brecon Beacons, south Wales (Palmer et al., 2008); Lochaber, western Scotland (Palmer et al., 2010); and the Solway Lowlands, northern England (Livingstone et al., 2010b).

The Lochaber study – which produced a composite varve thickness record by wiggle-matching records from two neighbouring sites – shows the need for caution in using this for dating where a significant noise-to-signal record may be present. As illustrated in Fig. 19, the microfacies show distinct between-site differences, attributed to local variations in climate, sediment supply and glacial dynamics. In microfacies A and G, for example, the coarser layer is composed respectively of well-sorted sand with a *sharp* upper contact (varve *sensu* Smith and Ashley, 1985), and poorly sorted medium silt with coarser silt grains and a *graded* upper contact. This suggests that each lamination in microfacies G was the product of continuous sedimentation, without a switch to suspension settling during the winter months. Such differences are borne out by the coefficient of determination (r^2) values calculated in Fig. 19: they indicate that more than 50% of the variation in varve thickness in these three examples is unexplained. This suggests that there is significant 'noise' within the rhythmite record, and that a non-annual, turbidite interpretation might be more probable for some deposits with a graded silt-clay contact.

Geochemical analysis offers an alternate method of analysing glaciolacustrine varves. Variations between the geochemical signatures of silt and clay laminae, in particular Ti, from Bogwood in central Scotland have been used to produce a varve thickness record indicating the advance of ice of Younger Dryas Stadial age into the region (MacLeod et al., 2010). However, a comparable varve thickness record derived from digital image and micromorphological analyses highlights the significant variability, often more than 10 mm, between the different proxy indicators. It is therefore uncertain which record is the most accurate, and correlations between varve records derived from separate sites can, at this stage, only be tentative.

5. Discussion and conclusions

From this review, it is clear that the Middle and Late Pleistocene glacial lakes of lowland Britain and the southern North Sea Basin come in all shapes and sizes and span a wide range of time. Some lakes were the size of small, ephemeral ponds, at the one extreme, and the size of major inland seas, at the other. Some lakes lasted a few years, whilst others lasted several thousand years or more. Some had a major influence on drainage patterns and Britain's link to the European mainland, others had little impact. This makes it very difficult to find a common thread linking all these bodies, except that they were ponded water bodies.

This review highlights a question: *why have most glacial lakes deposited laminated or other glaciolacustrine sediments whereas some – Lakes Eskdale, Lapworth and Fenland – apparently have not?* Although such sediments may later have been removed by erosion, an alternative explanation concerns denudation chronology.

Much of the early work that identified Pleistocene glacial lakes in Britain did so under a paradigm of denudation chronology. This applied the Davisian cycle of erosion to the British landscape (Wooldridge and Linton, 1939; see review in Jones, 1981) and focussed attention on erosional rather than depositional features. Such focus may account for the application of overflow channels, 'gaps', gorges and benches to delineate some glacial lakes. But heavy reliance on such features to map them has subsequently been questioned, as outlined above, not least because of difficulties in interpreting these landforms and dating them to particular glacial stages. Thus, the reconstructions of the glacial lakes discussed above vary in their reliability, and a number of them remain more conceptual than substantial. Interestingly, however, recent techniques such as cosmogenic isotope (^{36}Cl) surface exposure dating now offer potential for dating bedrock features such as overflow channels and gorges, and hence of improving lake reconstructions and relating them to wider landscape evolution. Since denudation chronology fell from favour in the mid 20th Century, studies of British glacial lakes have focussed more on sedimentary deposits. However, the application of mainstream sedimentology methods and interpretations (e.g., Collinson et al., 2006) remains limited to relatively few lake studies, and now needs to be systematically and widely applied in order to interpret depositional conditions – for example, shallow vs. deep water, ice-marginal facies and subaqueous outwash. Glacial lakes are, by themselves, a subsystem of the overall glacial sedimentary system, and as such they cannot exist in a vacuum as isolated basin fills. Therefore in discussing glacial lake genesis, evolution and sedimentation, it is difficult not to involve other, related sedimentary processes and environments. Glacial lake sedimentation, although characterised by laminites, or rhythmites (e.g., clastic varve sequences), also includes deltaic, littoral, gravity mass-flows and in some cases diamicton deposition and tephra might also occur. Integration of depositional and erosional features needs to be coupled with studies of glacial isostasy.

The role of glacial isostasy on glacial lake development in Britain is barely known. To date, reconstructions of the extent of some of the large lakes – Bosworth, Lapworth, Fenland, high-level Humber, for example – have been based on extrapolation of horizontal surfaces across present-day topography, placing lakes between high ground, former ice fronts and 'gaps'. This provides a useful first approximation of former lake extents, but does not take into account subsequent glacio-isostatic movements that must have tilted the landscape. Ideally, what is now needed are palaeotopographic modelling studies, such as those of J.T. Teller and colleagues working on the Canadian glacial lakes such as Agassiz. Such work applies GIS palaeotopographic modelling, based on adjusting modern digital elevation model surface elevations for differential isostatic rebound at a specific time (Yang and Teller, 2005; Murton et al., 2010). Application of such studies to Britain will help to constrain the extent of some of the conceptual lakes like Fenland and identify the sequence of outflow routings from lakes such as Bosworth and its impact of the development of the surrounding drainage networks. To determine the timing of the lake development, however, requires further dating studies.

Dating of Pleistocene glaciolacustrine deposits in Britain is likely to benefit from a combination of luminescence and sedimentological methods, as well as development of the varve chronologies discussed above. In North America, G. Berger and colleagues have successfully applied thermoluminescence (TL) dating to fine silts deposited in Arctic lakes and former glacial lakes (Berger et al., 1987; Berger and Easterbrook, 1993; Berger and Anderson, 2000). This work has highlighted the importance of first identifying lithofacies in which the light-sensitive TL is likely to have been zeroed during subaqueous sediment deposition. Fine silts deposited slowly by rainout from suspension are thought to provide an effective natural zeroing of TL, and hence, the resulting clayey laminae have generally proved suitable for TL dating. By contrast, coarser silty laminae deposited by turbid meltwater, as well as sediments deposited in ice-proximal locations, are generally incompletely zeroed and unsuitable for TL dating. In Britain, Murton et al. (2009) extended the scope of luminescence dating of glaciolacustrine sediments, again based on prior sedimentological analysis. By targeting wave-rippled sand beds deposited in very shallow-water or near-emergent conditions – favourable to zeroing of OSL – they obtained optical dates from sands in Lake Humber. Importantly, these sediments were emplaced 10 km south of the ice margin where the influence of turbulent discharge of meltwater into the

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lake would be significantly less than in an ice-proximal setting. Here, it is highly probable that the deltaic sediments would be incompletely bleached and so are unsuitable for OSL dating. As most other glaciolacustrine deposits themselves remain undated, luminescence dating offers the potential to develop other chronologies and test stratigraphic models and regional correlations, not only in Britain, but elsewhere where wave-rippled sands within glaciolacustrine deposits are identified.

Lastly, there is a need to develop a database with information on varve thickness and the distribution and age of glaciolacustrine deposits throughout Britain in order to elucidate the development of Pleistocene glacial lakes. This could build upon the BRITICE database and mapping of MIS 2 deposits and landforms (Clark et al., 2004).

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