The evolution of Palaeolake Flixton and the environmental context of Star Carr: II An oxygen and carbon isotopic record of environmental change for the early Holocene

Candy, I[1], Farry, A.[1], Darvill, C.M.[1,2], Palmer, A.[1], Blockley, S. P. E.[1], Matthews, I. P.[1], MacLeod, A., Deeprose, L.[1], Farley, N.[1], Kearney, R., Conneller, C.[3], Taylor, B.[4], Milner, N.[4]*

[1]Department of Geography, Royal Holloway, University of London, Egham, Surrey, TW20 0EX,UK

[2]Department of Geography, Durham University, Science Laboratories, South Road, Durham, DH1 3LE, UK

[3]Archaeology, School of Arts, Languages and Cultures, University of Manchester, Manchester, M13 9PL

[4]Department of Archaeology, University of York, The Kings Manor, York, YO1 7EP

[*]corresponding author: Ian Candy (ian.candy@rhul.ac.uk), +44 (0)1784 443649

Abstract

This paper presents $\delta^{18}O$ and $\delta^{13}C$ values from three early Holocene lacustrine carbonate sequences from Palaeolake Flixton in northeastern England. The $\delta^{13}C$ values are typical of carbonates precipitating in a large open lake system with the exception of samples from the very uppermost parts of these sequences which have values more typical of palustrine or tufaceous carbonates and, therefore, indicate the progressive contraction and shrinkage of the lake system overtime. The $\delta^{18}O$ values record an initial increase to an early peak at the onset of the Holocene but a subsequent decline in values of such a magnitude that by ca 8,000 yrs B.P. (the approximate age for the end of marl accumulation) the $\delta^{18}O$ value of the precipitated carbonate is consistent with carbonates that precipitated at the end of the Lateglacial stadial. The early increase and peak in $\delta^{18}O$ values is suggested to reflect the climatic amelioration at the onset of the Holocene, as temperatures progressively rose. The decline is anomalous and cannot be explained by any known temperature shifts, however, this “depletion trend” is seen in several other early Holocene lacustrine records from across the British Isles. The study concludes by suggesting that this decline reflects a shift in the seasonality of precipitation during the early Holocene from a more seasonal precipitation regime typical of a “continental” (strong seasonal contrasts in both temperature and rainfall) climate to a more “maritime” (reduced seasonality in terms of both temperature and rainfall) climate which was characterised by consistent levels of rainfall occurring across the year. The relative timing of the archaeological site of Star Carr suggests that human occupation occurred after this isotopic decline and, therefore, under maritime rather than under continental climate.

Key words

Oxygen and carbon isotopes, lacustrine carbonate, Holocene, Star Carr

Introduction

Lacustrine systems that occur in catchments that are dominated by limestone-rich bedrock are frequently characterised by the precipitation and accumulation of authigenic carbonates (Platt and Wright, 1991; Talbot and Allen, 1996; Verrechia, 2007). In such systems, carbonate precipitation from the water column is driven by biological controls, primarily photosynthesis, on the acidity of the lake water and, consequently, the solubility of carbonate (Verrechia, 2007). During periods of minimal allogenic input into the lake basin, carbonate accumulation can dominate lacustrine sedimentation leading to the deposition of relatively thick beds of authigenic carbonate, or marl (Platt and Wright, 1991; Verrechia, 2007). In recent years the analysis of the oxygen and carbon isotopic values of marl sequences has provided important records of abrupt climate change in the British Isles during the Lateglacial and the early Holocene (Marshall et al., 2002; 2007; Leng and Marshall, 2004; Leng et al., 2005; van Asch et al., 2012). In such studies it is variations in the $\delta^{18}O$ value of the marl sediment that provides the most important palaeoclimatic indicator acting, as it does, as a record of changes in the $\delta^{18}O$ value of meteoric water which is primarily controlled by prevailing air temperature (see Leng and Marshall, 2004; Andrews, 2006; Candy et al., 2011).

Although the $\delta^{18}O$ value of lacustrine carbonates can potentially provide important palaeoclimate data it has, in fact, been applied to comparatively few records in Britain, despite the abundance of early Holocene and Lateglacial marl sequences in this region. A good example of this would be the early Holocene carbonate records from Palaeolake Flixton in the Vale of Pickering, northeast England (Cloutman, 1988a; 1998b; Cloutman and Smith, 1988, Day, 1996; Dark, 1998; Palmer et al., submitted). This lake system, formed during the Lateglacial after the retreat of the Last Glacial Maximum (L.G.M.) ice from this locality

(Figure 1a), accumulated sediments across the early Holocene until ca 8,000 yrs B.P (Catt, 1987; Cloutman, 1988a; Palmer et al., submitted). The importance of studying the isotopic record of the early Holocene marl at this site is twofold. Firstly, the early Holocene archaeological site of Star Carr, one of the most important Mesolithic sites in Europe, occurs in association with the shoreline deposits of Palaeolake Flixton (Mellars and Dark, 1998; Conneller and Schadla-Hall, 2003; Conneller et al., 2012). An oxygen and carbon isotopic record from this sequence would, consequently, provide a palaeoenvironmental context to this important archaeological sequence. Secondly, the two key Lateglacial/early Holocene lacustrine isotope records in the British Isles occur on the west coast of Britain (Marshall et al., 2002; 2007) and Ireland (van Asch et al., 2012). An early Holocene isotope record from Palaeolake Flixton, located on the eastern side of Britain, would increase the spatial coverage of such records and begin to allow regional variations in isotopic response during this time interval to be understood.

In this paper we present the first oxygen and carbon isotopic records from the early Holocene lacustrine carbonates of Palaeolake Flixton. The aim of this study is not to generate a high-resolution isotope record but to use isotopic variations to look at major trends and patterns in environmental and hydrological change across the early Holocene. As well as discussing the significance of this record for the evolution of the landscape and environment of the palaeolake, the paper concludes by comparing this record with other oxygen isotope records from early Holocene lacustrine carbonate sequences in Britain. The main patterns of oxygen isotopic variability seen in the Flixton sequences are also seen in these other records and are interpreted in the context of early Holocene climatic change.

Holocene marl records from Palaeolake Flixton

During the L.G.M. the Vale of Pickering contained an extensive ice-dammed lake, constrained by high-ground to the west and the ice front to the east (Catt, 1987; Cloutman, 1998a; Clark et al., 2004; Palmer et al., submitted)). During deglaciation, ice-retreat exposed a series of topographic lows which, during the Lateglacial and early Holocene, became a more restricted lake basin, Palaeolake Flixton (Figure 1a & b), within which sediment accumulated. Surveys of the geometry of the sediment fill of the basin indicates that the bathymetry of this lake system was complex (Cloutman, 1988a; Cloutman and Smith, 1988) and was characterised by deep localised sub-basins separated by topographic highs (Figure 1c, after Palmer et al., submitted). This bathymetry reflects a typical glacial "Kettle and Kame" landscape that would have formed during deglaciation after the L.G.M (Palmer et al., submitted).

The complex bathymetry of this lake basin meant that changing water levels across the Lateglacial and early Holocene would have had significant impacts on the landscape and the sediment record at different locations within the palaeolake (Palmer et al., submitted). High lake levels (ca 24 metres O.D.) during the early Holocene would have resulted in the entire basin being submerged and flooded, resulting in early Holocene aged sediments accumulating across the entire area of the former lake basin. During low lake levels, i.e. the Lateglacial stadial, only the deepest part of the lake basin would have remained lacustrine with the topographic highs becoming emergent and exposed. Consequently, the deepest parts of the basin would preserve sediments deposited during the entire Lateglacial/early Holocene interval, see Figure 2 (Day, 1996; Dark, 1998Lowe et al., 1999). The topographic high-ground within the basin would, however, only have accumulated sediments during the early Holocene when lake levels were at their highest, Figure 2 (Cloutman, 1988a; Cloutman and Smith, 1988).

Periods of climatic amelioration, the Lateglacial Interstadial and the early Holocene, are characterised by the accumulation of carbonate-rich lacustrine marl (Day, 1996; Dark, 1998; Palmer et al., submitted). This reflects the relatively high vegetation cover in the area around Palaeolake Flixton at this time, which stabilises the landscape and reduces allogenic input, and the high bio-productivity of the lake waters, which cause regular changes in the pH of the lake waters and triggers carbonate precipitation (Palmer et al., submitted). The intervening cold interval, the Lateglacial Stadial, is dominated by allogenic inputs, reflecting the reduction in vegetation cover in the area surrounding Palaeolake Flixton and the subsequent increase in soil/sediment erosion (Palmer et al., submitted). During the early Holocene the progressive sediment infilling of the basin caused sufficient shallowing to propogate a shift from a lacustrine to a mire environment, as reflected by a shift from carbonate marl to peat accumulation, as part of a hydroseral succession (Day, 1996; Dark, 1998). A radiocarbon date at the base of the peat generated an age of 7640 14C yrs +/-85 (OxA-4042, Day, 1996) that is calibrated, using Intcal 13 (Reimer et al., 2013) to 8415-8050 cal yrs BP, implying that marl accumulation had ceased by this age.

Work by Palmer et al. (submitted) has shown that the longest (ca 3 metres) Holocene marl record in the basin is core B which occurs on the flanks of a deep sub-basin approximately 50 metres east of the site of Star Carr (Figure 1 and 2). The Holocene sequence from core D (ca 1.40 metres) comes from one of the deepest and widest sub-basins (Figure 2). The early Holocene sequence in both of these cores shows a progressive increase in % calcium carbonate composition in the lower 20 cm, suggested to reflect the climatic amelioration during the Lateglacial to Holocene transition, and then high and relatively stable % calcium carbonate values across the remainder of the early Holocene until the onset of peat accumulation. Core F comes from a topographic high (Figure 1c), approximately equidistant from core B and D and contains a relatively short (ca 0.5 metres) marl sequence. At the base of core F there is no progressive increase in % calcium carbonate values (Figure 2), the sequence, underlain by glacial gravels, shows high and stable % calcium carbonate values from the base of this sequence until the onset of peat accumulation.

Methodology

In order to produce a continuous record for sites B, D and F overlapping cores were extracted using 1 metre long, 50 mm wide Russian cores. Coring reached depths of 4 metres (B), 5.6 metres (D, the lowermost 2 metres being Lateglacial in age) and 1.4 metres (F) at which point the presence of either consolidated glacial clay (B and D) or glacial gravel (F) made further penetration impossible. The lithostratigraphy of each core was described using Troel-Smith and the % calcium carbonate content of the sediments were analysed using a Bascomb calcimeter (Gale and Hoare, 1991). As core B preserved the longest marl sequence this record was studied in the greatest detail, consequently, a continuous thin section record was produced through the sequence based on extracting material using a 3cm wide U-channel. A continuous sequence of 35 overlapping thin sections was then generated from depths 1.00 to 4.40 meters in this core. Sediment samples for thin section were air-dried and then impregnated using crystic resin (Palmer et al., 2008). The microfacies of the sediments were described along measured transects with the abundance of different features being quantified.

The sampling and preparation of carbonates for oxygen and carbon isotopic analysis is designed to reduce, or remove, the risk of contamination from allogenic detrital limestone and/or the mixing of different authigenic carbonate fabrics, i.e. marl, ostracod, chara, mollusc shells etc (Leng et al., 2010). In this study two approaches were utilised in order to reduce this risk. Firstly, microfacies analysis was used to identify areas where the sediment was comprised entirely of fine-grained authigenic marl with minimal evidence for in washing of sediments and/or the presence of mollusc shell or chara fragments (Figure 3). Samples were then drilled from these areas using a fine (1 mm wide) diamond tipped drill. Secondly, unimpregnated bulk sediment samples were taken from the same sampling points in B but at a lower resolution. These bulk samples were sieved over a 63µm mesh with the <63µm fraction being collected and treated with hydrogen peroxide to remove organic material. Both sets of samples were analysed for oxygen and carbon isotopic values and the values produced by the different sampling techniques were statistically identical. In order to account for the effect of resin impregnation samples of pure resin were mixed with standards of known $\delta^{18}O$ and $\delta^{13}C$ values and the presence of resin was found to have no effect of the derived isotopic values. For core D and F unimpregnated bulk sediment samples were extracted, sieved and treated with hydrogen peroxide and then analysed for stable isotopic values.

Samples for isotopic analysis were weighed using a Cahn C-31 Microbalance. The stable $\delta^{18}O$ and $\delta^{13}C$ values were measured by analysing CO_2 liberated from sample reaction with phosphoric acid at 90^{o}C. Internal (RHBNC-PRISM) and external (NBS-19, LSVEC) standards were run every 10 samples. The carbonate stable isotopes were analysed using a VG PRISM series 2 mass spectrometer. All stable isotopic values are quoted with reference to VPDB. Internal precision produces analytical uncertainties of +/-0.07 ($\delta^{18}O$) and +/-0.04 ($\delta^{13}C$).

It is important to acknowledge that the correlation of the marl in these three sequences is based on; 1) existing ^{14}C ages that correlate the onset of peat accumulation to ca 8,000 yrs B.P. (Cloutman, 1998a; Cloutman and Smith, 1988; Day, 1996; Dark, 1998) and, 2) the stratigraphy of the basin fill, i.e. the relative height of the units and the superimposition of these sequences, at various locations, above deposits with attributions to the Lateglacial stadial (Palmer et al., submitted). Consequently, the marl sequences analysed in this study are described as “early Holocene” and this is used simply to imply that the stratigraphic sequences discussed here must have accumulated between the end of the Lategalcial stadial/Holocene and ca 8,000 yrs B.P. No attempt is made to attribute an absolute chronology to any of the isotopic shifts seen in these sequences. When the isotopic patterns are discussed it is, therefore, with the aim of describing the isotopic shifts that occur across a broad and loosely defined time slice of the early Holocene without implying the attribution of any of these shifts to events that are known to affect the wider region.

Results

$\delta^{18}O$ and $\delta^{13}C$ values of core B

The record of δ^{18}O values in core B exhibit an increase, from -8.2 to -5.8‰, to an early maximum between 19.50 to 19.75 metres O.D. From 19.75 to 22.60 metres there is an overall trend of decreasing δ^{18}O values from the peak of -5.8‰ at 19.75 metres to -9.0‰ at 22.60 metres (Figure 4a). Although the overall trend is of decreasing values there is isotopic variability around this trend generating a series of peaks and troughs with an amplitude of ca 1.5 to 2.0‰. The δ^{13}C values show a similar trend to the δ^{18}O values with values increasing, from -1.2 to +2.8‰ between 19.50 to 19.75 metres O.D., and then showing an overall trend of decreasing δ^{13}C values, from +2.8 to -5.2‰, between 19.75 and 22.60 metres O.D. (Figure 4a). Although both δ^{18}O and δ^{13}C values show a similar trend (R^2 statistic 0.62) the magnitude of the isotopic shifts is much greater in the δ^{13}C values than in the δ^{18}O values.

$\delta^{18}O$ and $\delta^{13}C$ values of core D

The record of δ^{18}O values in core D show a trend consistent with that seen in core B (Figure 4b). An increase occurs, from -7.5 to -5.6‰, to an early maximum between 20.50 to 20.75 metres O.D. From 20.75 to 21.80 metres O.D. there is an overall trend of decreasing δ^{18}O values from the peak of -5.6‰ at 20.75 metres to -7.4‰ at 21.80 metres O.D (Figure 4b). As is the case for core B there is variability in $\delta^{18}O$ values around this trend. The $\delta^{13}C$ values show a similar trend to the $\delta^{18}O$ values with values increasing, from -0.9 to +1.8‰ between 20.50 to 20.75 metres O.D., and then showing an overall trend of decreasing $\delta^{13}C$ values, from +1.8 to -1.9‰, between 20.75 and 21.80 metres (Figure 4b). Although both $\delta^{18}O$ and $\delta^{13}C$ values show a similar trend (R^2 statistic 0.55) the magnitude of the isotopic shifts is much greater in the $\delta^{13}C$ values than in the $\delta^{18}O$ values.

$\delta^{18}O$ and $\delta^{13}C$ values of core F

The $\delta^{18}O$ and $\delta^{13}C$ values show no clear trend across core F (Figure 4c). Although there is variability in both the $\delta^{18}O$ (-7.4 to -6.4‰) and $\delta^{13}C$ (-1.6 to -0.7) values (R^2 statistic 0.52) there is no clear upward trend through the sequence.

Evidence for detrital contamination

Care has been taken, through microfacies analysis and sieving/bleaching procedures, to ensure that detrital contamination has not affected the isotopic results. To highlight the fact that detrital contamination is not an issue it is worth comparing the results produced from the authigenc marl with those of the regional bedrock geology, from which there is a relatively extensive isotopic dataset (Irwin et al., 1977; Hudson et al., 1978; Astin and Scotchman, 1988; Price, 1988; Kantotowicz, 1990; Price et al., 2000). This is done in Figure 5, which shows that the marl dataset has minimal overlap with any bedrock source with the exception of the Oxford clay, the nearest outcrop of which is more than 3miles away from the site. There is either no or minimal overlap with the local rock types (the Kimmeridge and Speeton Clays). Consequently it is argued that detrital contmaintaion is not an issue in the isotopic dataset presented here.

Core correlation using the $\delta^{18}O$ values of lacustrine carbonate

The $\delta^{18}O$ signal of each of the cores can be to align the three sequences. This is suggested because, in a lake system, variations in $\delta^{18}O$ values should be a function of changes in the $\delta^{18}O$ value of the lake waters, which are controlled by the $\delta^{18}O$ value of the meteoric waters, and the temperature of carbonate mineralisation (Darling et al., 2003; Darling, 2004; Leng and Marshall, 2004). Over a short distance within a single lake there is no reason to assume major differences in the $\delta^{18}O$ value of the lake water. This is not true for $\delta^{13}C$ values which, due to local differences in biological or physical processes, may show local gradients (Stuiver, 1975).

Core B and D both show an increase in $\delta^{18}O$ values to an early maximum and, in both cases, the magnitude and absolute values that occur in this shift are consistent. After this early peak both records show a progressive trend of decreasing $\delta^{18}O$ values. It is, therefore, considered reasonable to assume that the early increase and early peak in $\delta^{18}O$ values seen in cores B and D can be directly correlated. The magnitude of the decline in $\delta^{18}O$ values is much greater in core B, $\delta^{18}O$ values of -9.0‰ just prior to the onset of peat formation, than in core D, where values are ca -7.4‰ directly before the onset of peat formation. The simplest explanation to account for this difference is that the record of B is longer and that lacustrine sedimentation had persisted at the site of core B after peat accumulation had begun at the location of core D.

F does not show the early rise and early peak in $\delta^{18}O$ values seen in B and D. Equally, the $\delta^{18}O$ values generated from core F are not as high as those seen in the early peak of B and D but are consistent with those seen in depths 20.25 to 22.25 metres O.D. in Core B and depths 21.00 to 21.80 metres O.D. in core D. The marl sequence in F is relatively short and is located on a topographic high in the Palaeolake Flixton basin. The difference in the $\delta^{18}O$

record between core F and cores B and D could, therefore, reflect a progressive rise in water levels across the early Holocene.

In this model water levels would have risen during the early Holocene but, during the time interval represented by the increase and early peak in $\delta^{18}O$ values at B and D, not to a sufficient level to inundate the site of F (Figure 6a). A progressive rise in lake levels, or a discrete second phase of water level increase, then occurred after this early peak and after the subsequent drop in $\delta^{18}O$ values witnessed in B and D (Figure 6b). This later water level rise was sufficient to allow the site of core F to be inundated and for marl accumulation to begin at this location (Figure 6b). This model of water level rise is supported by the sedimentary record of the three cores. In B and D the % calcium carbonate record shows a progressive increase in values that would be expected in a record which records the stabilisation of the landscape and the climatic amelioration at the very onset of the Holocene. At F the onset of marl production is sharp with high (ca 85%) calcium carbonate values at the marl/gravel contact with no evidence for a progressive increase in marl production. This is more consistent with a surface that is inundated later in the early Holocene after landscape stabilisation and climatic amelioration has occurred. As progressive sedimentation caused the water body to shallow, parts of the basin, i.e. at D and F, saw lacustrine environments replaced by mire environments, consequently, at these locations marl accumulation ceased (Figure 6c). Marl accumulation persisted at the site of core B, where a more spatially restricted palustrine environment existed until mire accumulation began there also (Figure 6c).

Discussion

Interpretation of the Palaeolake Flixton $\delta^{18}O$ and $\delta^{13}C$ records

The $\delta^{13}C$ values of the Palaeolake Flixton marl are consistent with those of lacustrine carbonates (Stuiver, 1970; 1975; Talbot, 1990; Leng and Marshall, 2004; Candy et al., 2011). The $\delta^{13}C$ of lake carbonates is controlled by the dissolved inorganic carbon (DIC) of the lake waters. Lake water DIC is a function of that of the water that recharges the lake and uptake/degassing processes within the lake. Although ground and surface waters are typically depleted with respect to $\delta^{13}C$, due to the uptake of plant respired CO_2 in the soil zone during groundwater recharge, the equilibration of lake water CO_2 with atmospheric CO_2 that typically occurs after the water has been resident in the basin for a short time interval leads to an increase in DIC $\delta^{13}C$ values.

The samples in the uppermost part, 1.24 to 1.75 metres, of B have lower values -3.5 to -5.5‰ that are more typical of tufa/palustrine carbonates (Alonso-Zarza, 2003; Andrews, 2006; Candy et al., 2011). In such water bodies the high rate of recharge means that the equilibration with atmospheric CO_2 is not complete and, consequently, the resulting authigenic carbonate partially inherits the more negative isotopic signal of soil zone carbon (Alonso-Zarza, 2003; Andrews, 2006; Candy et al., 2011). If this is correct then during the accumulation of most of the early Holocene marl, Palaeolake Flixton was a relatively extensive lake body and this surface water body persisted until the onset of mire development. The isotopic record of B suggests, however, that during the final phases of basin in filling, localised, shallow water bodies, probably with significant throughflow, persisted in the landscape prior to peat development.

The $\delta^{18}O$ value of lacustrine carbonate is primarily controlled by; 1) the temperature at which carbonate mineralisation occurs and 2) the $\delta^{18}O$ value of the lake water. The second factor is controlled by the $\delta^{18}O$ of rainfall which is controlled by a range of factors, including; air temperature, amount of rainfall, seasonality of rainfall and distance from the moisture source. (Rozanski et al., 1992; 1993; Darling, 2004). A large number of studies have shown that, in mid-latitude temperate regions, the $\delta^{18}O$ value of rainfall is strongly controlled by air temperature at a relationship of approximately +0.58‰ per +1°C (Rozanski et al., 1993; Andrews, 2006; Candy et al., 2011). Increasing temperatures will, therefore, lead to rainfall, and consequently meteoric and lake waters, with higher $\delta^{18}O$ values. The control of air temperature on the $\delta^{18}O$ value of lake waters is consequently recorded in the $\delta^{18}O$ value of the lacustrine carbonates that precipitate from these lake waters.

If this relationship is accepted then the increase in $\delta^{18}O$ values that is present at the base of cores B and D can be interpreted as representing the increase in temperature that occurs during the transition from the Lateglacial stadial to the early Holocene. This suggestion is supported by the fact that the magnitude of the increase in $\delta^{18}O$ values that occurs at the base of the sequence is of a comparable magnitude, ca -2.5‰, to the Lateglacial to Holocene transition in lacustrine carbonate sequences at sites such as Haweswater, the English Lake District (Marshall et al., 2002)) and Fiddaun, the west coast of Ireland (van Asch et al., 2012).

Although the positive shift that occurs during the deposition of the lowermost part of B and D is consistent with the onset of the Holocene, the subsequent pattern of declining $\delta^{18}O$ values is not and is anomalous. Most records of Holocene climate from western Europe and the north Atlantic region indicate that, after the climatic amelioration at the onset of the Holocene, a number of short-lived climatic oscillations (e.g. Thomas et al., 2007) occurred but temperatures across the remainder of the Holocene remained; 1) relatively high, and 2) relatively stable (e.g. Davis et al., 2004; Lowe et al., 2008). If temperature is considered the primary control on the $\delta^{18}O$ value of the Palaeolake Flixton marl sequence this record is, therefore, anomalous because it implies that after the initial warming that occurred at the onset of the Holocene the climate must have cooled to such a degree that by ca 8,000 yrs B.P. (the timing of the onset of peat development) the $\delta^{18}O$ value of the marl, and, therefore, by inference the $\delta^{18}O$ value of rainfall and hence air temperature, was comparable to that which occurred during the Lateglacial stadial.

To place this into context the decrease in $\delta^{18}O$ values that occurs in this sequence after the early Holocene peak is of a magnitude consistent with that which would occur across a glacial/interglacial transition (see Marshall et al., 2002; Candy et al., 2012; Gallant et al., 2014). As no temperature decline of such a magnitude occurs during any part of the Holocene it is difficult to justify proposing temperature as the main control of the isotopic signal. Although events such as the pre-Boreal oscillation and the 8.2ka event (Thomas et al., 2007; Marshall et al., 2007; Lowe et al., 2008) do have an impact on temperatures these are relatively short-lived (ca 100-200 years) and cause only 1-2°C worth of cooling (Marshall et al., 2007).

It is considered unlikely that the decline in $\delta^{18}O$ values is down to changes in the hydrology of the lake basin. As mentioned above the progressive in filling by sediments of the Palaeolake Flixton basin would mean that, over time, the water body would contract and become smaller. As such it is possible that hydrological processes which may modify the isotopic signature of the water body, such as evaporation, may become more effective overtime (Stuiver, 1970; Leng and Marshall, 2004; Candy et al., 2012). This is considered unlikely because increased evaporation will typically lead to an increase in the $\delta^{18}O$ value of the lake water, and consequently the authigenic carbonate, not a decrease, which is the pattern that is observed here.

An early Holocene Oxygen isotope depletion trend in British Lacustrine carbonates?

Although it has received very little attention a trend towards decreasing $\delta^{18}O$ values across the early Holocene appears to occur in all available lacustrine $\delta^{18}O$ records in Britain and other parts of western Europe (Figure 7). A review of the literature indicates that the pattern of $\delta^{18}O$ values seen at Flixton, is also seen at Loch Inchquin (west coast of Ireland (Diefendorf et al., 2006; 2008)), Haweswater (the English Lake District (Marshall et al., 2002; 2007)), Lundin Tower (southeast Scotland (Whittington et al., 1996)) and Lake Tibetanus (northern Sweden ((Hammarlund et al., 2002)). All of these records show the same basic pattern of; 1) an increase in $\delta^{18}O$ values at the onset of the Holocene to an early peak, 2) a decrease in $\delta^{18}O$ values after that peak and 3) a stratigraphy/chronology that implies that this decline had occurred prior to ca 8,000 yrs B.P. In all cases the magnitude of the decrease in isotopic values is such that by at least 8,000 yrs B.P., $\delta^{18}O$ values are comparable with those of carbonates precipitated during the Lateglacial stadial (Figure 7). It should be highlighted that the chronology of many of these sites is of low resolution, there is, therefore, no implication that the timing of these peaks are synchronous or that the onset of the depletion trend is uniform across the British Isles. Figure 7 simply shows that during the first 3,500 yrs of the Holocene, effectively the pre-Boreal and early part of the Boreal (the Blytt-Sernander) or the Early Holocene (Walker et al., 2012), these sites contain a consistent $\delta^{18}O$ isotopic pattern.

The absolute value of authigenic carbonates precipitated during this interval shows a spatial variation which is consistent with known variations in the $\delta^{18}O$ value of modern day precipitation, i.e. the early Holocene peak in $\delta^{18}O$ values is higher in western sites such as Loch Inchquin and Haweswater, ca -4.0 to -4.5‰ (Marshall et al., 2002; Diefendorf et al., 2006; 2008)) than at eastern sites such as Lundin Tower and Palaeolake Flixton, ca -6‰ (Whittington et al., 1996). However, at all sites the magnitude of the decrease is relatively consistent, between 2 and 3‰. As this depletion event is observed in carbonate sequences from lake basins of a range of sizes and from still extent basins as well as palaeo-lake basins it is difficult to explain this decline in $\delta^{18}O$ values in terms of changing lake basin hydrology.

As it is unlikely that temperature variability, and its effect on the $\delta^{18}O$ value of rainfall, can explain the isotopic pattern observed in the early Holocene, it is here proposed that the most probable explanation is that the isotopic signal is driven by variations in the precipitation regime (see Hammarlund et al., 2002; Diefendorf et al., 2006; 2008). It is unlikely that, in a maritime mid-latitude location such as the British Isles, a change in the amount of annual precipitation would generate such a major shift in $\delta^{18}O$ values. The amount effect is strongest in tropical regions where high levels of precipitation (>2000mm) can lead to the modification of the $\delta^{18}O$ value of precipitation (Rozanski et al., 1992; 1993). The amount of rainfall does affect the $\delta^{18}O$ value of rainfall in Britain (Jones et al., 2007), however, it is unclear whether a shift of 2-3‰ could be explained solely by changes in the amount of rainfall at this latitude. Although it is currently difficult to prove an exact cause for this depletion event it is considered more likely that a change in the seasonality of the precipitation regime of western Europe could account for the isotopic patterns reported here (see Diefendorf et al., 2006).

The $\delta^{18}O$ value of groundwater is effectively an "average" of the $\delta^{18}O$ value of annual rainfall (Darling et al., 2003; Darling, 2004). In Britain this has been shown to vary across the year in response to changing air temperature (Darling, 2004, although see Jones et al., 2007). That is to say that the $\delta^{18}O$ value of summer rainfall in Britain typically has a higher value than the $\delta^{18}O$ value of winter rainfall (Darling, 2004). This seasonal variability is "homogenised" during groundwater recharge, resulting in meteoric waters with relatively constant $\delta^{18}O$ values throughout the year (Darling et al., 2003; Candy et al., 2011; Waghorne et al., 2012). In eastern England there is minimal seasonal variability in the distribution of rainfall across the year, with each month containing comparable levels of precipitation (see Candy et al., 2006 for a discussion of modern seasonality in rainfall in eastern England). This means that isotopically "depleted" winter rainfall contributes as much to the $\delta^{18}O$ value of groundwater as isotopically "enriched" summer rainfall. In such a region it is possible to generate large shifts in the $\delta^{18}O$ value of groundwater, and, therefore, surface water, if the seasonality of rainfall changes. A reduction in the amount of summer rainfall (reducing the contribution of isotopically "enriched" rainfall) or an increase in the amount of winter rainfall (increasing the amount of isotopically "depleted" rainfall) would result in a decrease in the $\delta^{18}O$ value of lake waters without the necessity of a concomitant shift in either temperature or the actual amount of mean annual rainfall.

In this model the early Holocene increase and peak in $\delta^{18}O$ values would occur as a result of increasing temperatures under a strongly seasonal rainfall regime, most probably characterised by high levels of summer rainfall relative to low levels of winter rainfall. This is suggested because compilations of Holocene temperature variations in Europe suggest that, in western Europe, "continental" style climates (i.e. summers and winters that were warmer and colder than the present day respectively) occurred during the very earliest Holocene (Zagwijn, 1994; Davis et al., 2004). Continental style climates are characterised by high summer rainfall relative to low winter rainfall, partly because of the fact that cooler air masses can carry less water vapour. After the early Holocene peak in $\delta^{18}O$ values a reduction in the seasonality of the rainfall regime occurred, either through a reduction in summer rainfall and/or an increase in winter rainfall. This resulted in a progressive decrease in the $\delta^{18}O$ value of groundwaters/lake waters and consequently the $\delta^{18}O$ value of authigenic carbonate. This decrease in rainfall seasonality may have resulted from a reduction in the "continentality" of the climate in association with the progressive rise in sea level which occurred across this time interval, generating a more "maritime" precipitation regime, characterised by more consistent rainfall across the year. Such a model would be consistent with palaeoecological evidence for an influx of warm water into the Irish Sea during the early Holocene that elevated winter temperatures (Zagwijn, 1994).

At the moment this explanation of early Holocene isotopic must be speculative. It is, however, consistent with; 1) the isotopic evidence, 2) our understanding of patterns of early Holocene climate change, and 3) the need for an explanation that excludes a major early Holocene temperature decrease. Currently, precipitation levels and seasonality are difficult to quantify in temperate regions, where the environmental tolerance of many paleocological proxies are used to reconstruct temperature (Candy et al., 2010). Furthermore, as most palaeoecological proxies in Britain reconstruct summer temperatures more precisely than winter temperatures (Candy et al., 2010), testing models of changing seasonality is highly problematic. If, however, this model is correct then it has two implications. Firstly, the isotopic record of Holocene lacustrine carbonates may allow a detailed understanding of changes in precipitation dynamics across this interval. Secondly, that the climate of the early Holocene was dynamic not just in terms of the occurrence of abrupt, short-lived cooling events (e.g. the 8.2 ka event) but in respect to the precipitation regime of this time interval.

With respect to the Mesolithic site of Star Carr, which occurs on the shorelines of Palaeolake Flixton, early human occupation was set against a backdrop of climatic change, not necessarily with regard to temperature variations, but in terms of precipitation regime as the climate system of Britain shifted from a more “continental” to a more “maritime” climate. Occupation at Star Carr appears to occur in association with the highest water levels attained by the palaeolake in the early Holocene. The correlation of cores on the basis of $\delta^{18}O$ values (Figure 6) would imply that this occurred “late” in the early Holocene with low water levels being associated with “continental” climates and high water levels being associated with “maritime” climates. A preliminary conclusion to this work would, therefore, be that the occupation of Starr Carr did not occur until after the onset of “maritime” climates, a conclusion that requires further testing.

Conclusions

- A sequence of lacustrine marls has been recovered from Palaeolake Flixton, northern England close to the Early Mesolithic site of Star Carr.
- A $\delta^{13}C$ and $\delta^{18}O$ record is produced for this sequence and a number of major isotopic shifts are observable.
- A significant depletion in $\delta^{13}C$ values occurs across the early Holocene from +3‰ to -5‰. This is interpreted as reflecting hydrological shifts associated with the evolution and contraction of Palaeolake Flixton from a deep extensive lake basin to a paludal environment characterised by shallow water bodies with significant throughflow of water.
- The $\delta^{18}O$ values show an initial enrichment in line with the temperature rise at the beginning of the Holocene. After the early Holocene peak in $\delta^{18}O$ values, however, the isotopic values decrease by ca 3‰ and these cannot be explained by temperature variations.
- An early Holocene $\delta^{18}O$ depletion trend appears to be common to several carbonate lake records in Britain and Ireland, and possibly in Mainland Europe. Its cause is suggested to reflect a shift in the seasonality of rainfall that is proposed to occur during the early Holocene.
- This proposed shift in rainfall regime is tentative but, if correct, it would mean that human occupation in the region occurred after this shift from “continental” to “maritime” climates had occurred.

Acknowledgements

This project was funded as part of a NERC grant (NE/1015191/1) fur further investigations at Star Carr in 2010. We would like to thank Scarborough Council and Tim Burkinshaw

(Cayton and Flixton Wetland and Carrs Project) and Margaret Nieke and Jackie Roberts of Natural England for granting permission to core in this field. Dave Lowry (Department of Earth Sciences, Royal Holloway, University of London) is thanked for help in the analysis of the samples. This manuscript was greatly improved by the comments and advice of two anonymous reviewers.

References

Alonso-Zarza AM, 2003. Palaeoenvironmental significance of palustrine carbonates and calcretes in the geological record. *Earth Science Review,* **60**; 261-298.

Andrews JE, 2006. Palaeoclimatic records from stable isotopes in riverine tufas: Synthesis and review. *Earth-Science Reviews*, **75**; 85– 104.

Astin TR, Scotchman IC, 1988. The diagenetic history of some spetarian concretions from the Kimmeridge Clay, England. *Sedimentology* **35**; 349-368

Candy, I., Rose, J., Lee, J.R., 2006. A seasonally 'dry' interglacial climate in eastern England during the early Middle Pleistocene: palaeopedological and stable isotopic evidence from Pakefield, UK. Boreas, 35, 255-265.

Candy, I., Rose, J., Coope, G.R., Lee, J.R,., Parfitt, S.P., Preece, R.C., Schreve, D.C., 2010. Pronounced climate warming during early Middle Pleistocene interglacials: investigating the mid-Brunhes event in the British terrestrial sequence. Earth Science Reviews 103, 183-196.

Candy I, Stephens M, Hancock JDR, Waghorn RS, 2011. Palaeoenvironments of Ancient Human Occupation: The application of oxygen and carbon isotopes to the reconstruction of Pleistocene environments. In Ashton, N., Lewis, S.G. and Stringer, C. (eds). *The Ancient Human Occupation of Britain Project. Developments in Quaternary Science*, Elsevier, 23-37.

Candy, I., Adamson, K., Gallant, C., Maher, E., Pope, R., *2012*. Oxygen and carbon isotopic composition of Quaternary meteoric carbonates from western and southern Europe: their role in palaoenvironmental reconstruction. *Palaeoclimatology, Palaeogeography, Palaeoecology*. 326-328, 1-11.

Catt JA, 1987. The Quaternary of East Yorkshire and adjacent areas. In Ellis (ed) *East Yorkshire, Field Guide*, Quaternary Research Association, Cambridge, 1-14.

Cerling TE, Quade J, 1993. Stable carbon and oxygen isotopes in soil carbonates. In: Swart, P.K., Lohmann, K.C., Mckenzie, J., Savin, S., (eds). *Climate Change in Continental Isotopic Records*. Geophysical Monograph 78, 217-231. American Geophysical Union, Washington.

Clark CD, Evans DJA, Khatwa A, Bradwell T, Jordan CJ, Marsh SH, Mitchell WA, Bateman, MD, 2004. Map and GIS database of landforms and features related to the last British Ice Sheet. *Boreas* 33; 359-375.

Cloutman EW, 1988a. Palaeoenvironments in the Vale of Pickering, part 1: stratigraphy and palaeogeography of Seamer Carr, Star Carr and Flixton Carr. *Proceedings of the Prehistoric Society* **54**; 1-19.

Cloutman EW, 1988b. Palaeoenvironments in the Vale of Pickering, part 2: environmental history at Seamer Carr. *Proceedings of the Prehistoric Society* **54**; 21-36.

Cloutman EW, Smith AG, 1988. Palaeoenvironments in the Vale of Pickering, part 3: environmental history at Star Carr. *Proceedings of the Prehistoric Society* **54**; 37-58.

Conneller C, Milner N, Taylor B, Milner N, in press. Structured settlement in the European early Mesolithic: New research at Star Carr. *Antiquity*.

Conneller C, Schadla-Hall T, 2003. Beyond Star Carr: the Vale of Pickering in the 10th millennium BP. *Proceedings of the Prehistoric Society* **69**; 85-106.

Dark P, 1998. Palaeoecological investigations. In Mellars, P. And Dark, P. (eds) *Star Carr in context*. Chapters 9-15, 111-181. McDonald Institute Monographs, Oxbow Books, Oxford.

Darling WG, 2004. Hydrological factors in the interpretation of stable isotopic proxy data present and past: a European perspective. *Quaternary Science Reviews* **23**; 743-770.

Darling WG, Talbot JC, 2003. The O and H stable isotopic composition of fresh waters in the British Isles 1. Rainfall. *Hydrology and Earth System Sciences* **7**; 163–181.

Davis, B.A.S., Brewer, S., Stevenson, A.C., Guiot, J., Data Contributors, 2003. The temperature of Europe during the Holocene reconstructed from pollen data. Quaternary Science Reviews, 22, 1701-1716.

Day P, 1996. Devensian Late-glacial and early Flandrian environmental historv of the Vale of Pickering, Yorkshire, England. *Journal of Quaternary Science* **11**; 9-24.

Diefendorf AF, Patterson MP, Mullins HT, Tibert N, Martini A, 2006. Evidence for high-frequency late Glacial to mid-Holocene (16,800 to 5500 cal yr B.P.) climate variability from oxygen isotope values of Lough Inchiquin, Ireland. *Quaternary Research* **65**; 78 – 86

Diefendorf, A., Patterson, W.P., Holmden, C.. Mullins, H.T. (2008) Carbon isotopes of marl and lake sediment organic matter reflect terrestrial landscape change during the late Glacial and early Holocene (16,800 to 5,540 cal yr B.P.): a multiproxy study of lacustrine sediments at Lough Inchiquin, western Ireland‘, *Journal of Paleolimnology*, **39**, 101-115

Gallant, C.E., Candy, I., van der Bogaard, Turner, E., Silva, B., in press. Stable isotopic evidence for Middle Pleistocene environmental change from a loess paleosol sequence: Kärlich, Germany. Boreas

Gale, S., Hoare, P., 1991. *Quaternary Sediments: Petrographic methods for the study of unlithified rocks*, Belhaven and Halsted Press, New York

Hammarlund D, Barnekow L, Birks HJB, Buchardt B, Edwards TWD, 2002. Holocene changes in atmospheric circulation recorded in the oxygen-isotope stratigraphy of lacustrine carbonates from northern Sweden. *The Holocene* **12**; 339–351.

Leng MJ, Marshall, JD, 2004. Palaeoclimate interpretation of stable isotope data from lake sediment archives. *Quaternary Science Reviews* **23**; 811–831.

Leng MJ, Jones MD, Frogley MR, Eastwood WJ, Kendrick CP, Roberts C, 2010. Detrital carbonate influences on bulk oxygen and carbon isotope composition of lacustrine sediments from the Mediterranean. *Global and Planetary Change* **71**; 175–182

Leng MJ, Lamb A., Heaton THE, Marshall JD, Wolfe BB, Jones MD, Holmes JA, Arrowsmith C, 2005. Isotopes in lake sediments. In: Leng M. J. (ed.). *Isotopes in Palaeoenvironmental Research*. Springer, Dordrecht, The Netherlands. 148-176 pp.

Lowe, J.J., Birks, H.J.B., Brooks, S.J., Coope, G.R., Harkness, D.D., Mayle, F.E., Sheldrick, C., Turney, C.S.M., Walker, M.J.C. 1999. The chronology of palaeoenvironmental changes during the Last Glacial-Holocene transition towards an event stratigraphy for the British Isles‘, Jo*urnal of the Geological Society*, London, **156**, 397-410

Lowe, J.J., Rasmussen, S.O. Björck, S., Hoek, W.Z., Steffensen, J.P., Walker, M.J.C., Yu, Z.C. & the INTIMATE group, 2008. Synchronisation of palaeoenvironmental events in the North Atlantic region during the Last Termination: a revised protocol recommended by the INTIMATE group‘, *Quaternary Science Reviews*, **27**, 6-17

Marshall JD, Jones RT, Crowley SF, Oldfield F, Nash S, Bedford A, 2002. A high resolution late glacial isotopic record from Hawes Water, Northwest England. Climatic oscillations: calibration and comparison of palaeotemperature proxies. *Palaeogeography, Palaeoclimatology, Palaeoecology* **185**; 25–40.

Marshall JD, Lang B, Crowley SF, Weedon GP, van Calsteren P, Fisher EH, Holme R, Holmes J., Jones, RT, Bedford A, Brooks SJ, Bloemendal J, Kiriakoulakis K, Ball JD, 2007. Terrestrial impact of abrupt changes in the North Atlantic thermohaline circulation: early Holocene, UK. *Geology* 35; 639-642

Mellars P, Dark P, 1998. Summary and conclusions. In *'Star Carr in context: new archaeological and palaeoecological investigations at the Early Mesolithic site of Star Carr, North Yorkshire'*, Mellars, P. and Dark, P. (ed's). McDonald Institute Monographs.

Palmer AP, Lee JA, Kemp RA, Carr SJ, 2008. *Revised laboratory procedures for the preparation of thin sections from unconsolidated sediments*. Centre for micromorphology publication.

Palmer,A.P., Blockley, S.P.E.M., Candy, I., Darvill, C., MacLeod, A., Matthews, I.P., Milner, N., Taylor, B., Farry, A., Flowers, K., submitted. Landscape evolution of the Star Carr Area of North-east Yorkshire during the Last Glacial Interglacial Transition. Proceedings of the Geologists' Association.

Platt, N.H., Wright, V.P. 1991. Lacustrine carbonates: facies models, facies distributions and hydrocarbon aspects', *Special Publication of the International Association of Sedimentologists*, **13**, 57-74

Price GD, 1998. Isotopic variation in fossils and matrix of the Cretaceous Red Chalk at Speeton and South Ferriby, Yorkshire, England. *Proceedings of the Yorkshire Geological Society* **52**; 107-112.

Price GD, Ruffell AH, Jones CE, Kalin RM, Mutterlose J, 2000. Isotopic evidence for temperature variation during the early Cretaceous (late Ryazanian-mid Hauterivian). *Journal of the Geological Society, London* **157**; 335-343.

Rose J, 1985. The Dimlington Stadial/Dimlington Chronozone: proposal for naming the main glacial episode of the Late Devensian in Britain. *Boreas* **14**; 225-230.

Rozanski K, Araguas-Araguas L, Gonfiantini R, 1993. Isotopic Patterns in Modern Global Precipitation. In: Swart PK, Lohmann KC, McKenzie J, Savin S. (eds), *Climate Change in Continental Isotopic Records*. Geophysical Monograph **78**, American Geophysical Union, 1 - 36.

Stuiver, M. 1970. Oxygen and carbon isotope ratios of freshwater carbonates as climatic indicators', *Journal of Geophysical Research*, **75**, 5247-5257

Stuiver, M. 1975. Climate versus changes in 13C content of the organic component of lake sediments during the Late Quaternary', *Quaternary Research*, **5**, 251-262

Talbot, M.R. 1990. A review of the palaeohydrological interpretation of carbon and oxygen isotopic ratios in primary lacustrine carbonates', *Chemical Geology*, **80**, 261-279

Talbot, M.R. & Allen, P.A. 1996. Lakes' in Reading, H.G. (ed.'s) *Sedimentary Environments: processes, facies and stratigraphy*. Blackwell Science, Oxford

Thomas E., Wolff EW, Mulvaney R, Steffensen JP, Johnsen SJ, Arrowsmith C, White JWC, Vaughn B, 2007. The 8.2 ka event from Greenland ice cores. *Quaternary Science Reviews* **26**; 70-81.

Verrecchia EP, 2007. Lacustrine and Palustrine Geochemical Sediments. In Nash DJ, McLaren SJ, (eds) *Geochemical Sediments and Landscapes*. Blackwell Publishing Ltd, Malden, 298-330.

Waghorne RS, Hancock JDR, Candy I, 2012. Environmental controls on the $\delta^{18}O$ composition of freshwater calcite and aragonite in a temperate, lowland river system: Significance for palaeoclimatic studies. *Proceedings of the Geologists' Association* **123**; 576-583.

Whittington G, Fallick AE, Edwards KJ, 1996. Stable oxygen isotope and pollen records from eastern Scotland and a consideration of Late-glacial and early Holocene climate change for Europe. *Journal of Quaternary Science* **11**; 327-340

Figure captions

Figure 1 – The location of the study site. 1A – the position of Palaeolake Flixton in the context of the margins of the Last Glacial Maximum (L.G.M.), or *Dimlington Stadial* (Rose, 1985), ice sheet. 1B – The approximate extent of Palaeolake Flixton during the Holocene, black circles show the location of key archaeological sites. 1C – The bathymetry of Palaeolake Flixton in the area between Star Carr and No Name Hill (shown in the dashed box on 1B). This figure shows depth (in metres O.D.) to glacial sediments (diamicton, glaciolacustrine clays or outwash gravels). The study area consists of deep/isolated sub-basins separated by topographic highs reflecting a typical "Kettle and Kame" topography (Palmer et al., submitted). The arrows show the location of cores described in Palmer et al. (submitted) but only cores B, D and F are described here.

Figure 2 – Lithostratigraphy, calcium carbonate values and chronostratigraphy of cores B, D and F from Palaeolake Flixton. The stratigraphic correlation of the different units in each core follows the correlation of Palmer et al. (submitted).

Figure 3 - Photomicrographs of the three main carbonate dominated lithofacies found in core B (all taken undercross-polarised light). 3a) Lithofacies 1, the light band in the centre of the field of view represents bands of increased authigenic calcite whilst the darker bands above and below contain predominantly silt-sized minerogenic material. 3b) Lithofacies 2, the dark bands represent calcite microspar interspersed with lighter bands of calcite spar, minimal minerogenic material is present. 3c) Lithofacies 3, dominated by coarse spar with mollusc shell fragments and chara.

Figure 4 - $\delta^{18}O$ and $\delta^{13}C$ values of carbonates from cores B, D and F plotted against height within core in metres O.D. In Cores D and F all the samples have been prepared for analysis through sieving through 63µm sieves and treatment with H_2O_2. In core B the open symbols indicate data from samples that were drilled from impregnated blocks, whereas the closed symbols represented samples sieved and treated with H_2O_2 prior to analysis. There is no difference between the data generated through drilling and the data generated through sieving, both sets of samples show comparable values and trends. The data produced by the two preparation methods is, therefore, directly comparable.

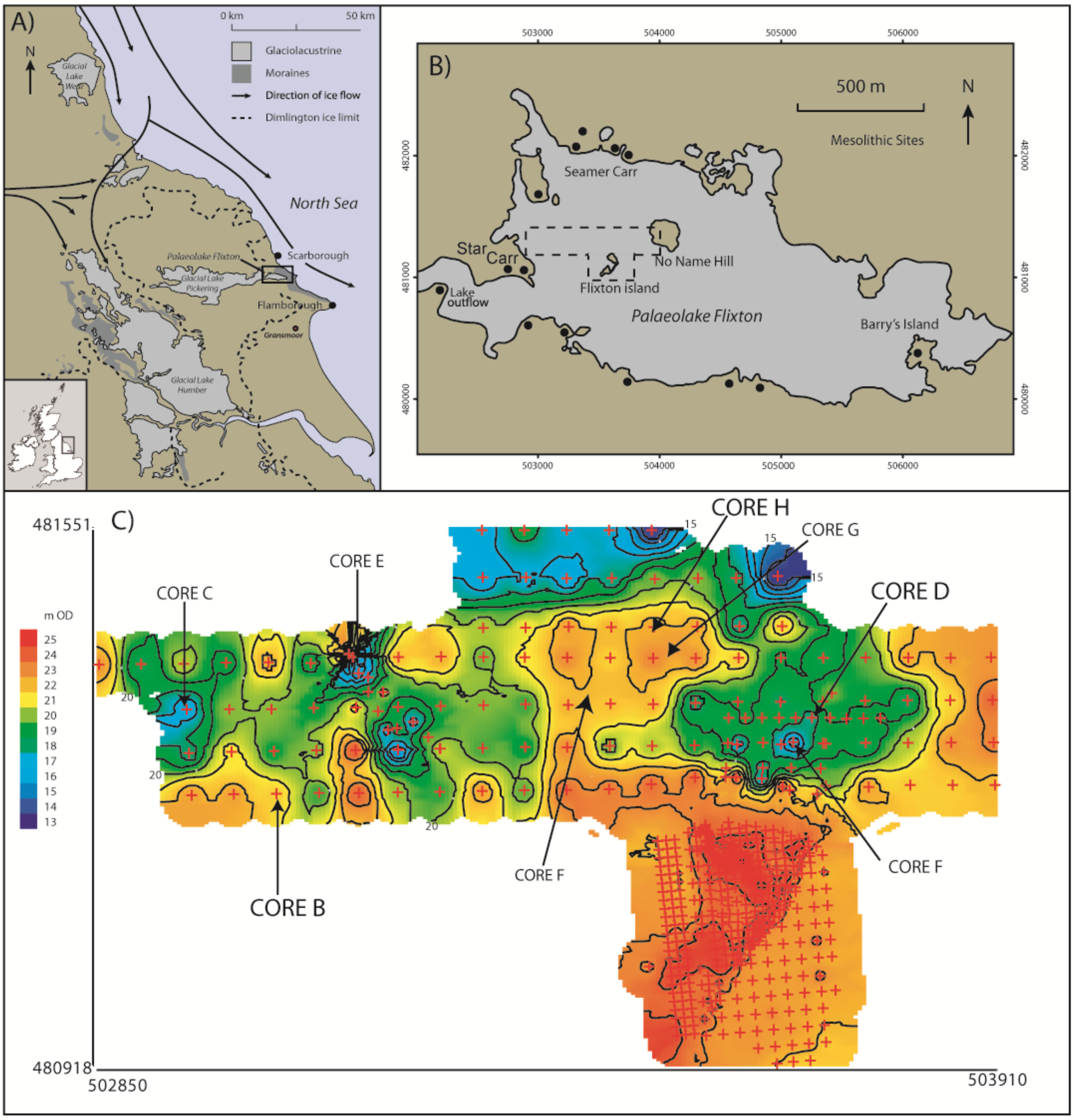
Figure 5 - Comparison of the Star Carr marl dataset (red squares) with previously published data on bedrock limestone of different sources in the Star Carr region. Star Carr is underlain by Kimmeridge Clay, which contain Septarian nodules (the primary source of carbonate) and the Speeton clays. There is very little affinity between the marls and any of the major geological groups, overlapping with only a few outliers. Data from Price (1998), Price et al. (2000), Irwin et al. (1977), Astin and Scotchman (1988), Hudson et al. (1978) and Kantotowicz (1990).

Figure 6 – Schematic evolution of the Palaeolake Flixton sequence that accounts for the variations in; 1) stratigraphy, and 2) isotopic record that are found in each core. During the "earliest" Holocene (6A) lake levels have risen to a level that allowed marl accumulation at the location of cores B and D but not at the location of core F. Consequently, marl accumulates at the former sites and the isotopic record of each core records the early Holocene increase, peak and decline in $\delta^{18}O$ values . "Later" in the early Holocene (6B) water levels have risen to such a level that all three coring sites are submerged and marl accumulation occurs at all three sites. During the "latest" early Holocene (ca 8,000 yrs B.P.)

sedimentation has resulted in large parts of the basin being so shallow that mire environments have succeeded lacustrine deposition (6C at the sites of cores D and F). Only at the location of site B does freshwater sedimentation persist in a palustrine ("paludal") environment.

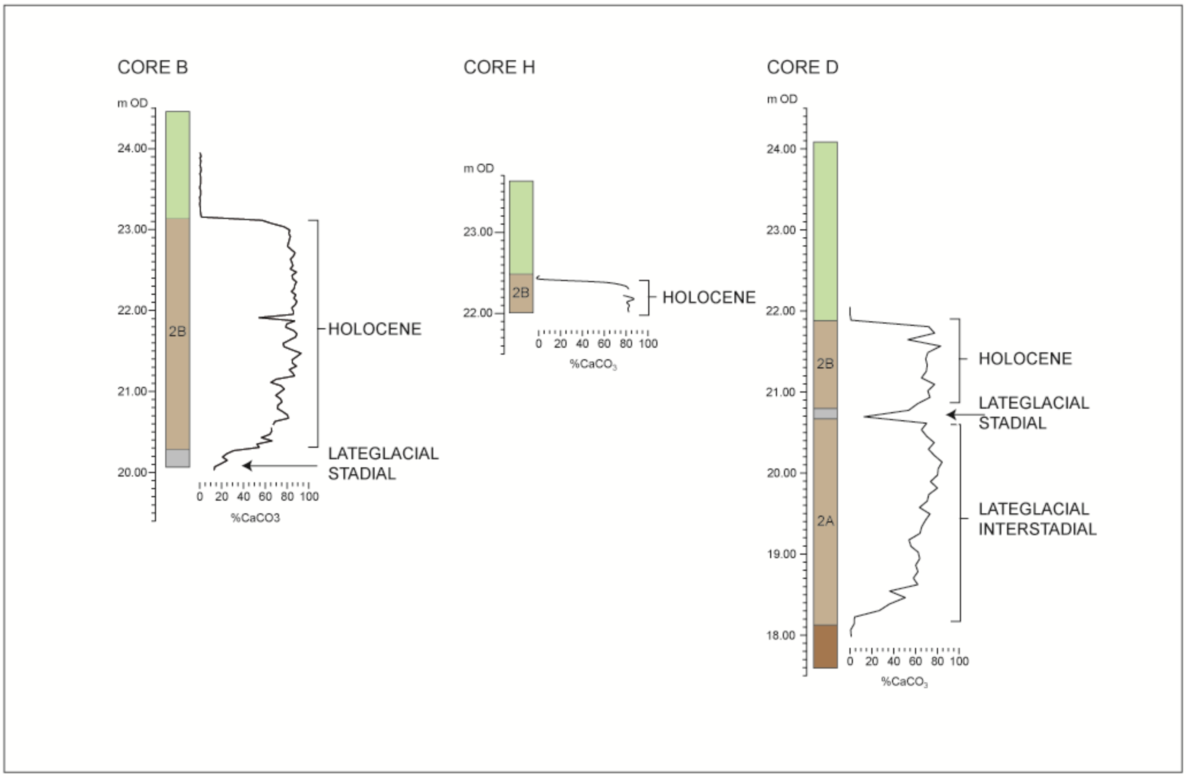
Figure 7 - Figure 6 – Comparison between the isotopic records of Loch Inchquin (Diefendorf et al., 2006; 2008), Haweswater (Marshall et al., 2002; 2007), Star Carr (this study), Lundin Tower (Whittington et al., 1996), Lake Tibetanus (Hammarlund et al., 2002). The chronological precision on these records is highly variable. At sites like Loch Inchquin and Lake Tibetanus there are a high concentration of radiocarbon dates. At Haweswater there are only two radiocarbon dates for the Lateglacial and a number of U-series dates for the Holocene. At Lundin Tower and Star Carr there is minimal radiometric dating control but the existing pollen work, aligned to a small number of radiocarbon dates, allows these sediments to be robustly correlated with the early Holocene. In general the uppermost part of each record, as shown here, correlates with ca 8,000 years BP. Consequently the alignment here is to show broad patterns in $\delta^{18}O$ values at key stratigraphic intervals and is not an attempt at high-resolution correlation. Essentially all records have a stratigraphy that is robust enough to identify the onset of the Holocene, at Loch Inchquin, Haweswater and Star Carr it is possible to say that the youngest end of the curve presented here is ca 8,000 yrs BP. The pollen stratigraphy at Lundin Tower would imply a similar time interval. The Haweswater record is patched together from two isotopic record a Lateglacial (Marshall et al., 2002) and a (smoothed) Holocene (Marshall et al., 2007) sequence. This is not an attempt to produce a continuous record but merely to illustrate that the depletion event seen in Star Carr, Lundin Tower and Loch Inchquin is replicated here. The magnitude of depletion from the early Holocene peak varies from site to site but is typically between -2.0 and -3.0‰ (PDB).

Figure 1



A)
N
Glacial Lake Wear
0 km
50 km
Glaciolacustrine
Moraines
Direction of ice flow
Dimlington ice limit
North Sea
Palaeolake Flixton
Glacial Lake Pickering
Scarborough
Flamborough
Gransmoor
Glacial Lake Humber
B)
503000
504000
505000
506000
482000
481000
480000
500 m
N
Mesolithic Sites
Seamer Carr
Star Carr
No Name Hill
Flixton island
Lake outflow
Palaeolake Flixton
Barry's Island
C)
481551
480918
502850
503910
m OD
25
24
23
22
21
20
19
18
17
16
15
14
13
CORE B
CORE C
CORE D
CORE E
CORE F
CORE F
CORE G
CORE H

Figure 2



CORE B
m OD
24.00
23.00
22.00
21.00
20.00
2B
HOLOCENE
LATEGLACIAL
STADIAL
0 20 40 60 80 100
%CaCO3
CORE H
m OD
23.00
22.00
2B
HOLOCENE
0 20 40 60 80 100
$\%CaCO_3$
CORE D
m OD
24.00
23.00
22.00
21.00
20.00
19.00
18.00
2B
2A
HOLOCENE
LATEGLACIAL
STADIAL
LATEGLACIAL
INTERSTADIAL
0 20 40 60 80 100
$\%CaCO_3$

Figure 3

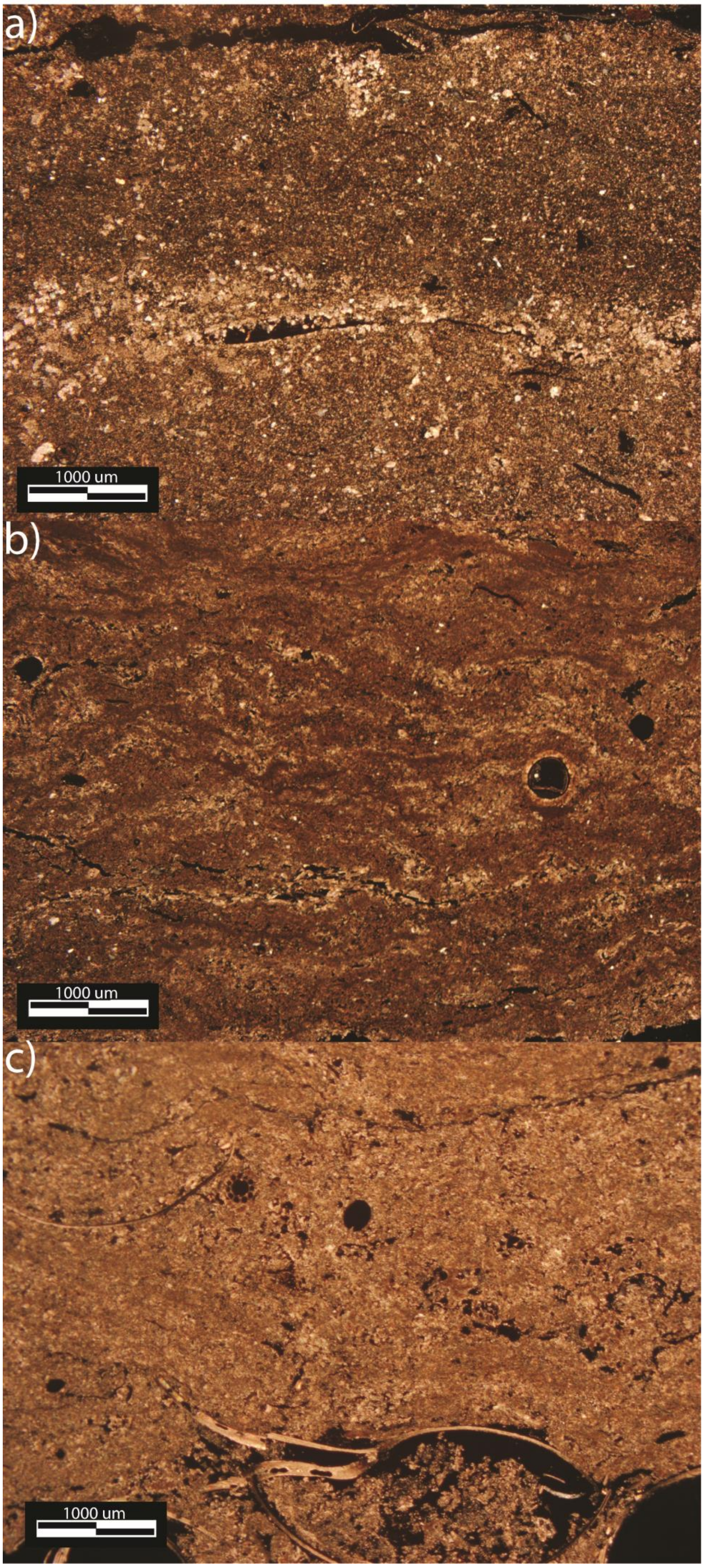
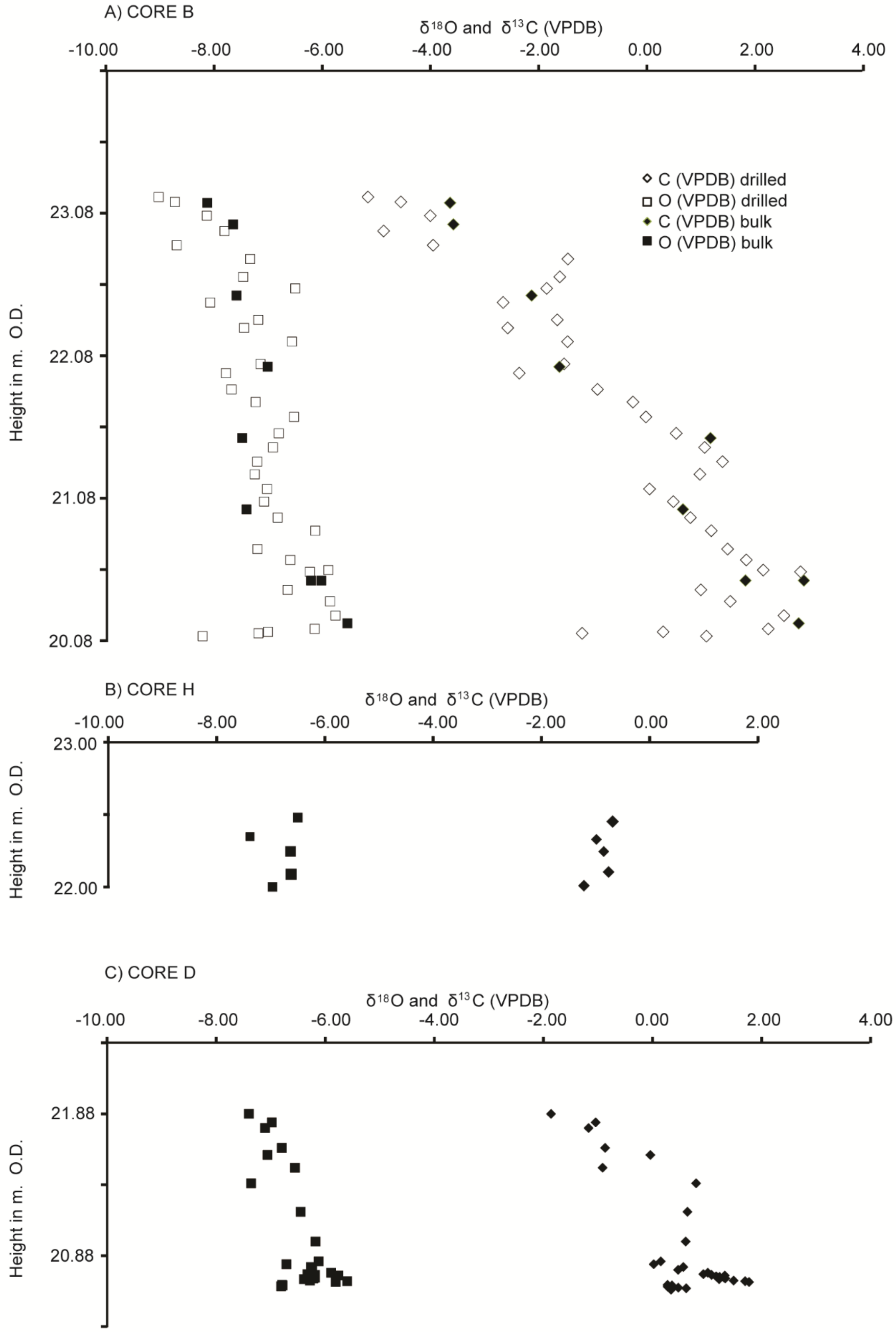
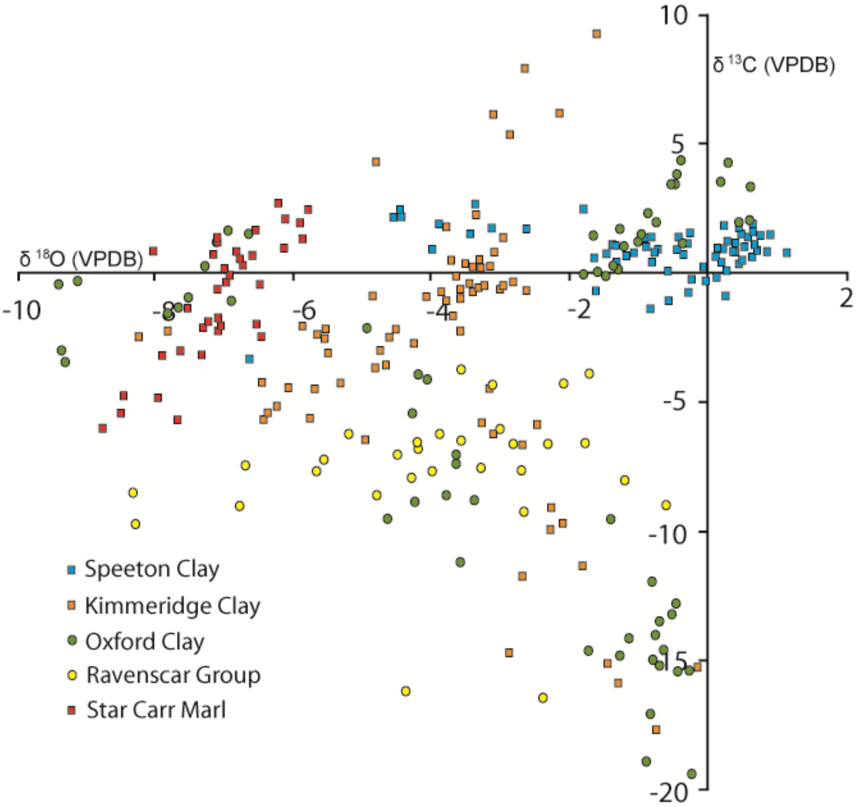


Figure 4



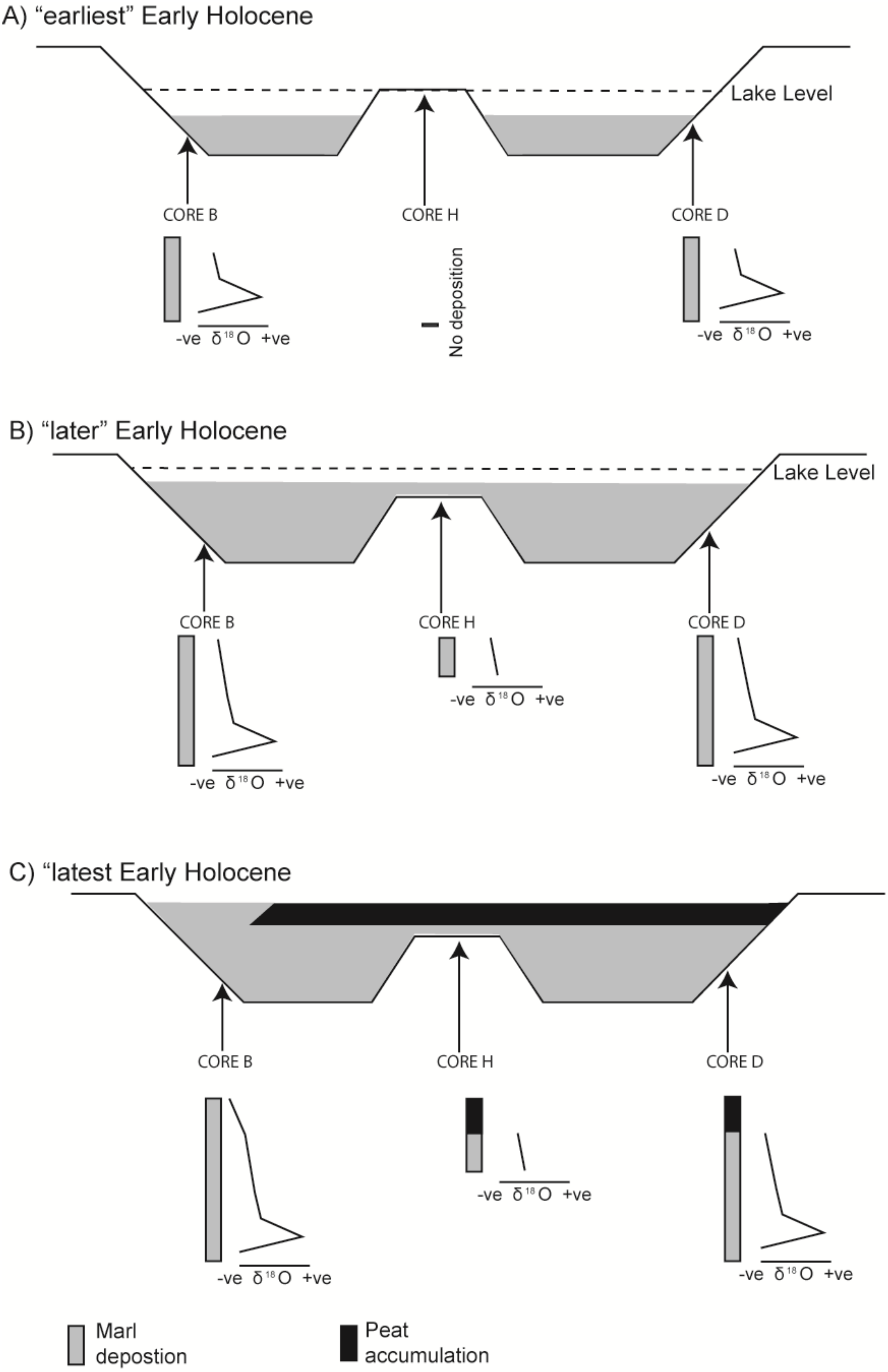
A) CORE B
δ18O and δ13C (VPDB)
-10.00
-8.00
-6.00
-4.00
-2.00
0.00
2.00
4.00
Height in m. O.D.
23.08
22.08
21.08
20.08
C (VPDB) drilled
O (VPDB) drilled
C (VPDB) bulk
O (VPDB) bulk
B) CORE H
δ18O and δ13C (VPDB)
-10.00
-8.00
-6.00
-4.00
-2.00
0.00
2.00
Height in m. O.D.
23.00
22.00
C) CORE D
δ18O and δ13C (VPDB)
-10.00
-8.00
-6.00
-4.00
-2.00
0.00
2.00
4.00
Height in m. O.D.
21.88
20.88

Figure 5



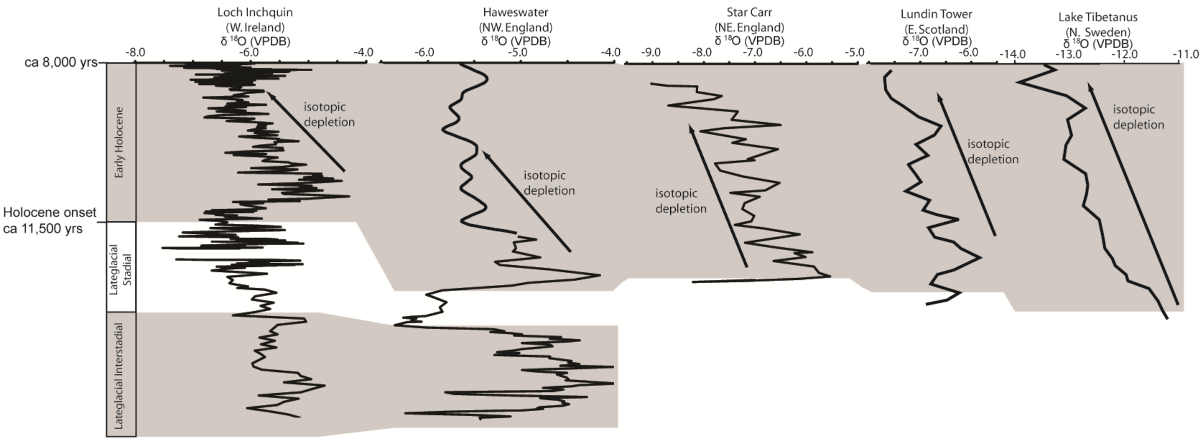
10
δ 13C (VPDB)
5
δ 18O (VPDB)
-10
-8
-6
-4
-2
2
-5
-10
-15
-20
Speeton Clay
Kimmeridge Clay
Oxford Clay
Ravenscar Group
Star Carr Marl

Figure 6



A) "earliest" Early Holocene
Lake Level
CORE B
CORE H
CORE D
-ve δ18O +ve
No deposition
-ve δ18O +ve
B) "later" Early Holocene
Lake Level
CORE B
CORE H
CORE D
-ve δ18O +ve
-ve δ18O +ve
-ve δ18O +ve
C) "latest Early Holocene
CORE B
CORE H
CORE D
-ve δ18O +ve
-ve δ18O +ve
-ve δ18O +ve
Marl depostion
Peat accumulation

Figure 7



Loch Inchquin
(W. Ireland)
δ 18O (VPDB)
Haweswater
(NW. England)
δ 18O (VPDB)
Star Carr
(NE. England)
δ 18O (VPDB)
Lundin Tower
(E. Scotland)
δ 18O (VPDB)
Lake Tibetanus
(N. Sweden)
δ 18O (VPDB)
ca 8,000 yrs
Holocene onset
ca 11,500 yrs
Early Holocene
Lateglacial Stadial
Lateglacial Interstadial
isotopic depletion