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The $\delta^{18}O$ stratigraphy of the Hoxnian lacustrine sequence at Marks Tey, Essex, UK: implications for the climatic structure of MIS 11 in Britain

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ABSTRACT: Marine Isotope Stage 11 (MIS 11) is considered one of the best analogues for the Holocene. In the UK the long lacustrine sequence at Marks Tey, Essex, spans the entirety of the Hoxnian interglacial, the British correlative of MIS 11c. We present multiproxy evidence from a new 18.5-m core from this sequence. Lithostratigraphy, pollen stratigraphy and biomarker evidence indicate that these sediments span the pre-, early and late temperate intervals of this interglacial as well as cold climate sediments that post-date the Hoxnian. The $\delta^{18}O$ signal of endogenic carbonate from this sequence produces several clear patterns that are interpreted as reflecting the climatic structure of the interglacial. As well as providing evidence for long-term climate stability during the interglacial and a major post-Hoxnian stadial/interstadial oscillation the $\delta^{18}O$ signal provides strong evidence for abrupt cooling events during the interglacial itself. One of these isotopic events occurs in association with a short-lived increase in non-arboreal pollen (the NAP phase). The results presented here are discussed in the context of other MIS 11 records from Europe and the North Atlantic, particularly with respect to our understanding of the occurrence of abrupt climatic events in pre-Holocene interglacials. Copyright © 2016 The Authors. *Journal of Quaternary Science* Published by John Wiley & Sons Ltd.

KEYWORDS: abrupt event; Hoxnian; Interglacial; Marine Isotope Stage 11; $\delta^{18}O$ and $\delta^{13}C$.

Introduction

Marine Oxygen Isotope Stage 11 (MIS 11, ca. 425–360 ka; MIS 11c, 425–395 ka) is often suggested to be the most appropriate analogue for the Holocene interglacial (Droxler and Farrell, 2000; Berger and Loutre, 2002; Loutre and Berger, 2003; Candy *et al*., 2014). This suggestion is based on the observation that the pattern of orbital forcing, and consequent insolation variability, that occurred during the Holocene matches that which occurred during MIS 11 more closely than in any other interglacial of the past 500 ka (Berger and Loutre, 2002, 2003; Loutre and Berger, 2003). As temporal variations in insolation patterns will partially control the duration and climatic structure of interglacials, climate records from MIS 11 allow the study of how the climate of a Holocene-like interglacial might evolve in the absence of anthropogenic forcing (Candy *et al*., 2014). Long climate records that provide data on the structure of MIS 11, including the EPICA Dome C ice core (EPICA Community Members, 2004; Jouzel *et al*., 2007), marine cores (McManus *et al*., 1999; Kandiano *et al*., 2012) and long lake records (Prokopenko *et al*., 2002, 2006, 2010; Tzedakis, 2010), are now providing high-resolution evidence for the extended duration, complex structure and potential instability of this climatic episode.

Despite the importance of marine and ice core archives there is an increasing need to understand the expression of MIS 11 in the terrestrial realm. Western and central Europe are ideal locations for such studies as (i) they contain several long lacustrine records that have been correlated with MIS 11 (Nitychoruk *et al*., 2005; Koutsodendris *et al*., 2010, 2011, 2012, 2013), and (ii) they occur in close proximity to the large number of high-resolution palaeoclimatic records that are found in the North Atlantic. Despite the often-fragmentary nature of such records they have the clear advantage over ice and marine core archives in that they can be annually laminated, or varved, and therefore they allow environmental change to be reconstructed at a much higher temporal resolution (Turner, 1970; Meyer, 1974; Müller, 1974; Nitychoruk *et al*., 2005; Mangili *et al*., 2007, 2010a, b; Brauer *et al*., 2008; Koutsodendris *et al*., 2010, 2011, 2012, 2013). In the UK, however, many of these terrestrial records have not been restudied for several decades; consequently, much of the palaeoenvironmental information that has been derived from them is based upon pollen analysis with only limited application of other proxy techniques. Marks Tey (Fig. 1), in southern Britain (Turner, 1970), is a good example of this. The lacustrine sequence at Marks Tey contains the most complete record of the Hoxnian interglacial, the British correlative of MIS 11, more specifically the interglacial interval MIS 11c (Shackleton and Turner, 1967; Shackleton, 1987). Much of this sequence has been proposed to be varved (Shackleton and Turner, 1967), with each varve containing a distinct summer lamination of endogenic carbonate, suitable for $\delta^{18}O$ and $\delta^{13}C$ isotopic analysis. Consequently, Marks Tey holds great potential for increasing our understanding of the expression of the climates and environments of MIS 11 in Western Europe.

This paper presents a re-investigation of the Hoxnian lake sediment sequence preserved at Marks Tey, through the analysis of a new sediment core that was recovered in 2010. The paper describes the lithostratigraphy of the new core and the micromorphological characteristics of the different lithofacies that are present. A pollen diagram for the sequence is also presented, which allows both for the correlation of the new sequence with that of Turner's (1970) original borehole, and for a reconstruction of the vegetation succession preserved within. This record of palaeoecological change is supported by the analysis of lipid biomarkers from the sediments, which allow for long-term patterns of environmental/biological change to be discussed. The $\delta^{18}O$ and $\delta^{13}C$ records of this sequence are also presented. Some of the

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Figure 1. Location of key MIS 11 records discussed in the text: (a) European and Atlantic records, (b) British records.

variability within the isotopic signal can be explained by changes in the style of sedimentation. However, a number of patterns exist that can be interpreted as reflecting climatic, primarily temperature, shifts. These include two episodes of abrupt change: (i) a period of climatic instability that occurs during full interglacial conditions, and (ii) an abrupt warming event that occurs in the cold, post-Hoxnian (post-MIS 11c) sediments. The paper concludes by discussing the significance of the climatic stratigraphy of this record particularly in the context of other records of MIS 11 from Britain and the North Atlantic.

MIS 11 in the British terrestrial record

Marks Tey (Fig. 1a,b) is one of several sites in central and southern Britain that preserve lacustrine sediments that span all, or a significant part, of the Hoxnian interglacial (West, 1956; Kelly, 1964; Turner, 1970; Coxon, 1985; Coope, 1993; Coope and Kenward, 2007; Preece *et al.*, 2007; Ashton *et al.*, 2008). Most of these sequences have accumulated in sedimentary basins that formed as kettle holes or sub-glacial scour features during the preceding Anglian glaciation and became lacustrine systems during the immediate post-glacial period and the subsequent interglacial. An abundance of litho-, bio- and morpho-stratigraphic evidence, coupled with a range of geochronological data, has been used to correlate robustly the Anglian glaciation with MIS 12 and the Hoxnian interglacial with MIS 11, or more specifically with MIS 11c, the interglacial interval of this warm isotopic stage (see Candy *et al.*, 2014 for a recent review of these arguments). Marks Tey is the para-stratotype for the Hoxnian–Interglacial, a temperate episode defined by West (1956) on the basis of pollen stratigraphy from the type sequence at Hoxne in Suffolk (Mitchell *et al.*, 1973). Marks Tey is unique as, unlike most other Hoxnian sequences, it preserves the full interglacial vegetation succession from the end of the Anglian through the entire Hoxnian–Interglacial into the subsequent cold interval.

The pollen stratigraphy of the Hoxnian (Fig. 2) is subdivided into four pollen zones (West, 1956; Turner and West, 1968; Turner, 1970). The pre-temperate zone (Hoxnian I or HoI), where total arboreal pollen (AP) first exceeds total non-arboreal pollen (NAP), is characterized by the presence of closed boreal forest species *Betula* and *Pinus*. The early-temperate zone (HoII) sees the expansion of mixed *Quercus* forest that undergoes a succession of changing species dominance from *Quercus* (HoIIa) to *Alnus* (HoIIb) to *Corylus* (HoIIc). The late-temperate zone (HoIII) is characterized by the progressive decline of mixed *Quercus* forest and an increase in late-migrating temperate trees such as *Carpinus* (HoIIIa) and *Abies* (HoIIIb). Finally, the post-temperate zone (HoIV) is characterized by a return to boreal and heathland species such as *Pinus* and *Empetrum* (HoIVa) as well as *Betula* and Poaceae (HoIVb). At Marks Tey, the Hoxnian sediments are underlain and overlain by beds that contain a high NAP/AP ratio, which have been ascribed to the colder climates of the preceding late Anglian and to the post-Hoxnian cold stage, respectively.

A key characteristic of the pollen record of many Hoxnian sequences is the occurrence of a short-lived increase in NAP relative to AP (Fig. 2) during the early-Temperate zone (HoIIc), commonly referred to as the non-arboreal pollen

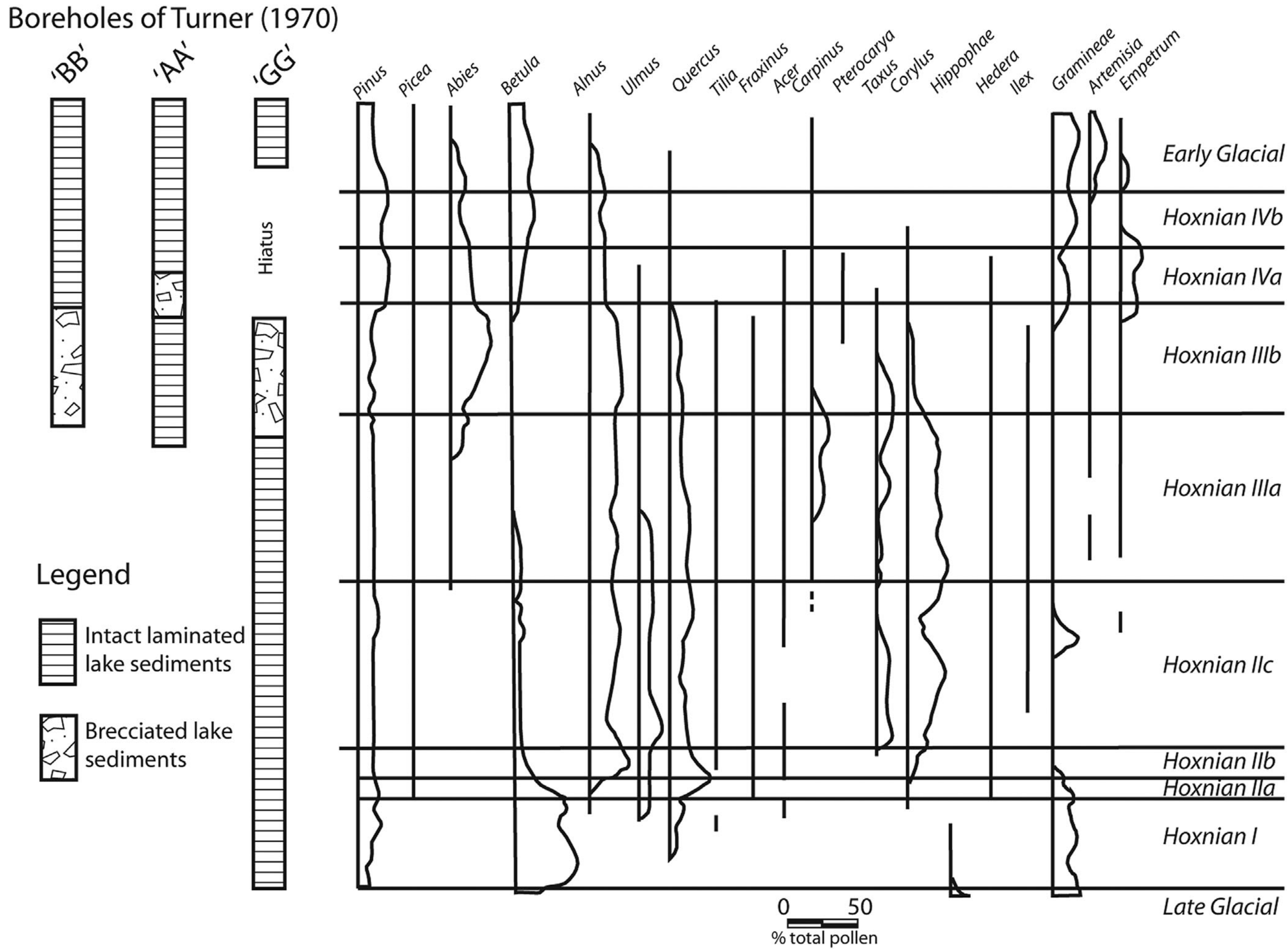

Figure 2. Composite pollen diagram for the Hoxnian interglacial as derived from the sedimentary succession at Marks Tey (Turner, 1970). The pollen diagram is constructed from three main boreholes 'AA', 'BB' and 'GG' (shown on the left-hand side of diagram); the pollen zones that these boreholes record and the nature of the sediments that they contained are shown for illustrative purposes.

phase, or NAP phase (West, 1956; Kelly, 1964; Turner, 1970). This episode is well expressed at Marks Tey, where the presence of assumed varved sediments allowed its duration to be reconstructed at ca. 300 years (Turner, 1970). A comparable event is seen in deposits of the Holsteinian interglacial, the continental correlative of the Hoxnian, in Germany (e.g. Dethlingen and Döttingen) and Poland (Ossówka). In the continental records this event has been referred to as the Older Holsteinian Oscillation, or OHO (Nitychoruk *et al*., 2005; Diehl and Sirocko, 2007; Koutsodendris *et al*., 2011, 2012, 2013). The forcing mechanism for this event is unclear as the occurrence of charcoal fragments at the start of the NAP phase at Marks Tey led Turner (1970) to suggest that the reduction in tree pollen was caused by wildfire, while other researchers have suggested that it was climatically driven (e.g. Kelly, 1964; Turner, 1970; Koutsodendris *et al*., 2011, 2012).

It is important to highlight the fact that the widely cited Marks Tey summary pollen diagram (Fig. 2) is a composite one. This is constructed from pollen records derived from several cores that were extracted from different parts of the Marks Tey basin (Turner, 1970; Rowe *et al*., 1999). The longest of these records, from borehole 'GG', comes from the deepest part of the basin and contains sediments, mostly fine silts and clays, which cover the interval from the late Anglian through pollen zones HoI–HoIIIb inclusive. A hiatus then occurs between the sediments of HoIIIb and those of the subsequent cold climate interval. The later parts of HoIIIb and units corresponding to HoIV were found in pollen records from boreholes 'AA' and 'BB', which were recovered from the lake margins.

In the sediments of 'GG', deposits of the late Anglian and HoI–HoII and much of HoIIIa are intact and finely laminated, while the sediments that correspond to HoIIIb are finely laminated but intensely brecciated, possibly due to the lowering of lake waters and sediment desiccation at this time (Turner, 1970; Gibbard and Aalto, 1977; Gibbard *et al*., 1986). The laminations within the sediments of HoI–HoIIa are irregular and highly variable with respect to their thickness (Turner, 1970). In contrast, from the approximate onset of HoIIb to HoIIIa and within the brecciated fragments of HoIIIb, the sediments have a more regular structure, comprising lamination sets. These sets have three distinct laminations consisting of: (i) diatom laminations, (ii) calcite laminations and (iii) detrital laminations. It is these lamination 'triplets' that have been suggested to be varved (Shackleton and Turner, 1967; Turner, 1970; Candy, 2009), although no quantitative work to validate (or otherwise) the varved nature of these laminations has been published.

It is the endogenic calcite laminations within the Marks Tey deposits that have the greatest potential for investigating British palaeoenvironments during MIS 11c (Candy, 2009; Candy *et al*., 2014). The analysis of the $\delta^{18}O$ and $\delta^{13}C$ values of lacustrine carbonates in British Quaternary sequences has become increasingly widespread over the past 20 years (Whittington *et al*., 1996, 2015; Marshall *et al*., 2002, 2007;

van Asch *et al*., 2012; Candy *et al*., 2015). The $\delta^{18}O$ value of lacustrine carbonates receives the most attention because, within open-lake systems under mid-latitude temperate climates, temperature is suggested to be the primary driver of this proxy (Leng and Marshall, 2004; Candy *et al*., 2015). This is because the $\delta^{18}O$ value of lacustrine carbonate is strongly related to the $\delta^{18}O$ of the lake water, which is in turn a function of the $\delta^{18}O$ of rainfall (Andrews, 2006; Candy *et al*., 2011). In mid-latitude regions, there is a positive linear relationship between prevailing air temperature and the $\delta^{18}O$ of rainfall (Dansgaard, 1964; Rozanski *et al*., 1992, 1993). In general this means that the $\delta^{18}O$ of lacustrine carbonates should increase under warmer climates and decrease under cooler climates. This temperature relationship is clearly seen in numerous lacustrine carbonate records of the Last Glacial to Interglacial Transition (LGIT), which provide the most convincing evidence in the British Isles for abrupt climatic shifts comparable to those preserved in the Greenland ice core records (Marshall *et al*., 2002; van Asch *et al*., 2012; Whittington *et al*., 2015).

Methodology

Core recovery and sedimentology

A sediment sequence (MT-2010) comprising two overlapping boreholes (TL 91081, 24431 and TL91082, 24432 drilling from a surface elevation of 15.80 m OD) was obtained in 2010, within 10 m of the original 'GG' borehole of Turner (1970). The cores were drilled using a wet rotary drilling rig, recovering cores in 3-m lengths, which were then cut in half for ease of transport. The two boreholes were correlated by the identification of key marker beds that were present in both sequences. This produced a composite sequence that was 18.5 m in length. This paper quotes depth as metres below surface (mbs) but depths are also shown in Figs 3–5 as metres below ordnance datum (OD). In the laboratory, macroscopic sediment description of the cores was undertaken to determine facies changes throughout the sequence and to identify the main sedimentary units, or lithofacies. Samples of 1 cm^3 were taken from throughout the sequence to determine the organic carbon content by titration (Walkley

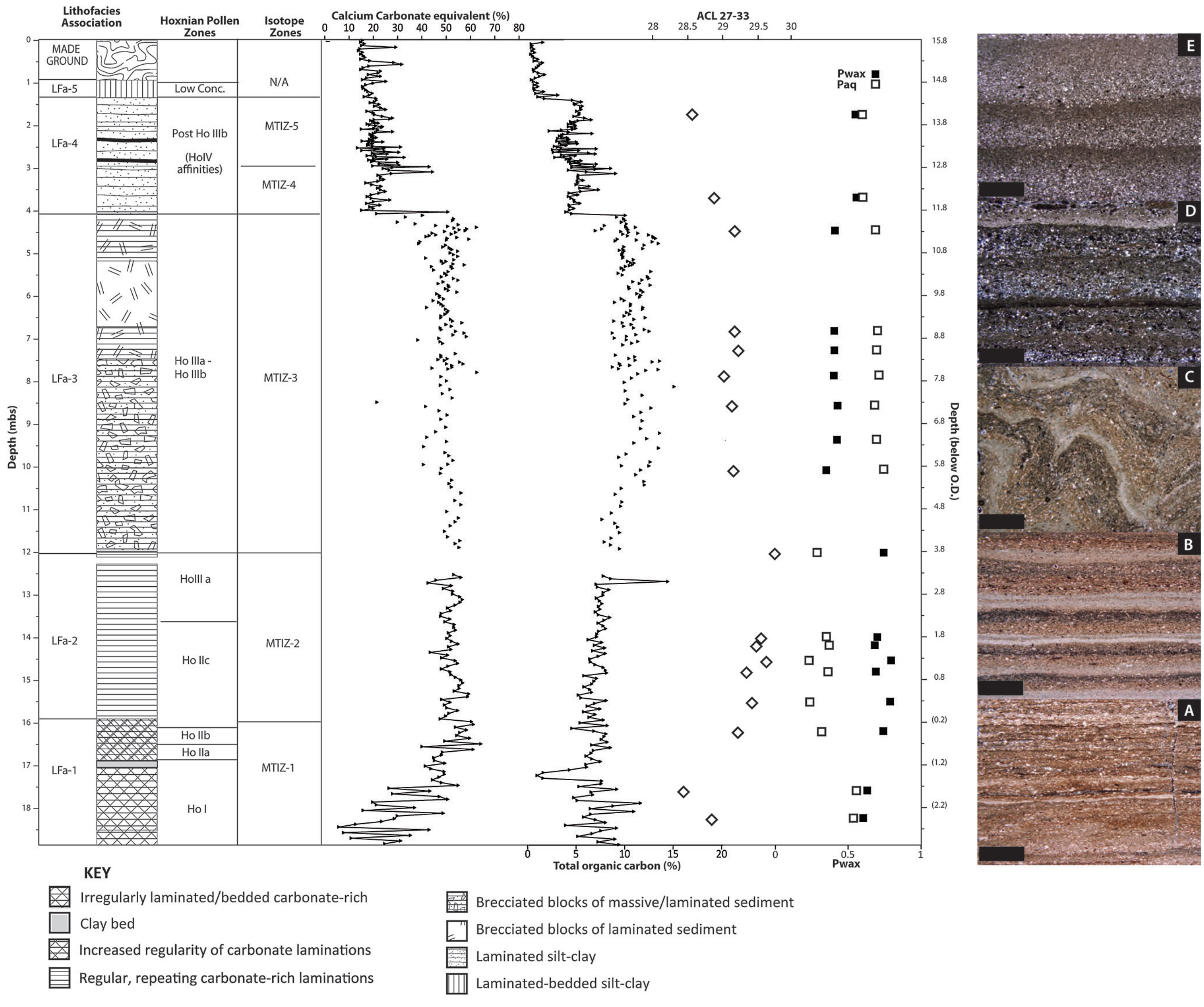

Figure 3. Lithostratigraphy, bulk sedimentology (% carbonate and % organic carbon) and biomarker records for the Marks Tey 2010 (MT-2010) borehole. Lithofacies associations are shown (Table 1, in Supplementary information Table S1, for detailed descriptions) as are the pollen zones recorded in this record (see Fig. 4 for pollen diagram). Photomicrographs (labelled A–E) highlight changes in microfacies through the sequence. LFa-1 consists of finely but irregularly laminated sediments (A), LFa-2 consists of finely but regularly laminated sediments (B) and LFa-3 consists of finely, regularly laminated sediments that are either brecciated or deformed (C). In all of these micrographs the light coloured laminae are calcitic, while the darker sediments are organic. LFa-4 and -5 are characterized by a significant proportion of minerogenic material with LFa-4 (D) containing occasional calcite lamination (top of photomicrograph) but LFa-5 (E) being dominated by fine sand, silts and clays. All photomicrographs were taken using cross-polarized light; scale bar at bottom left is 500 μm in all cases.

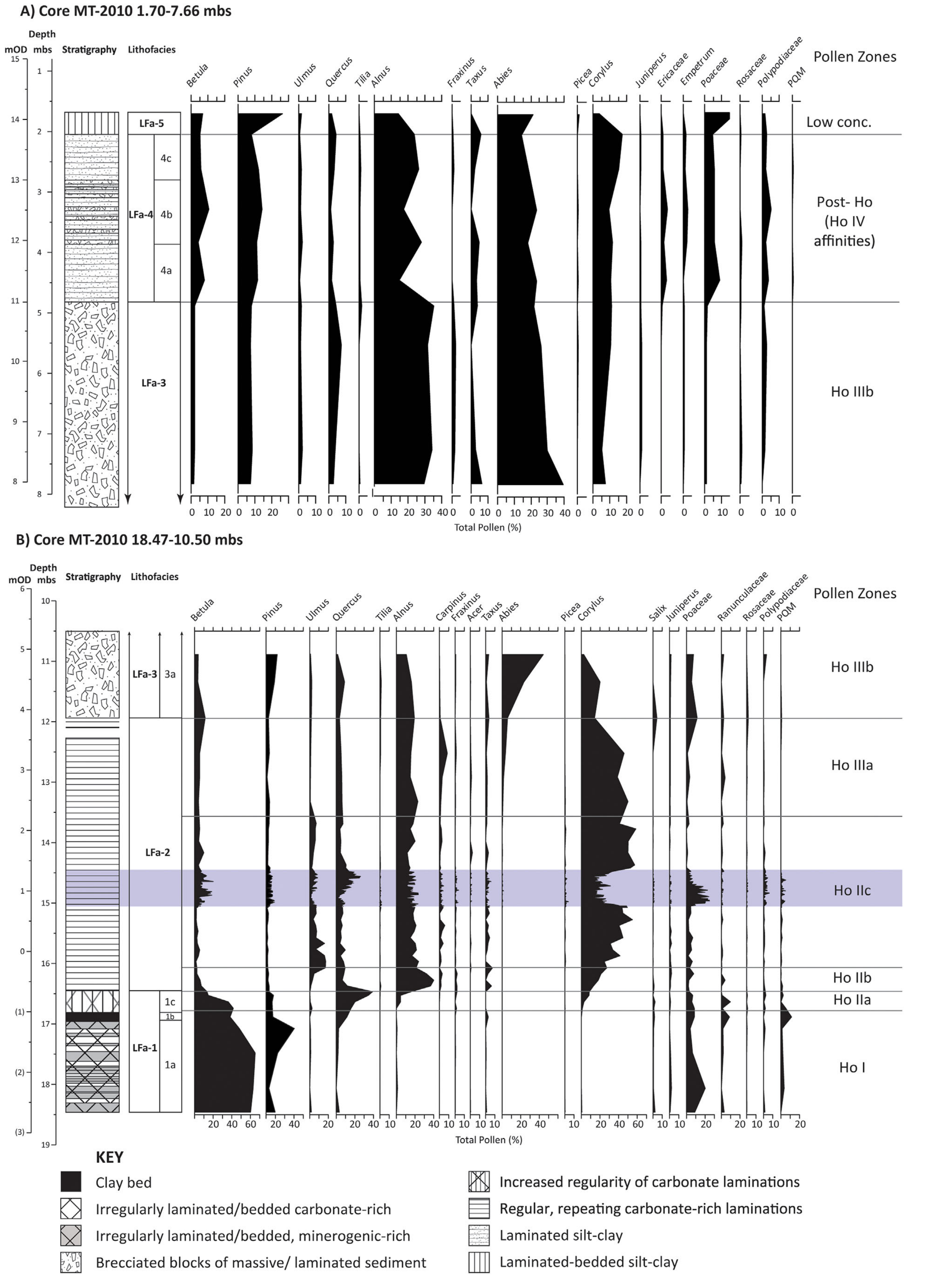

Figure 4. Pollen diagrams for the Marks Tey 2010 (MT-2010) record. No pollen analysis was carried out on the middle 2 m of the brecciated sediments of LFa-3. The key characteristics of MT-2010 replicate that seen in the 'GG' borehole of Turner (1970) including the NAP phase (shaded).

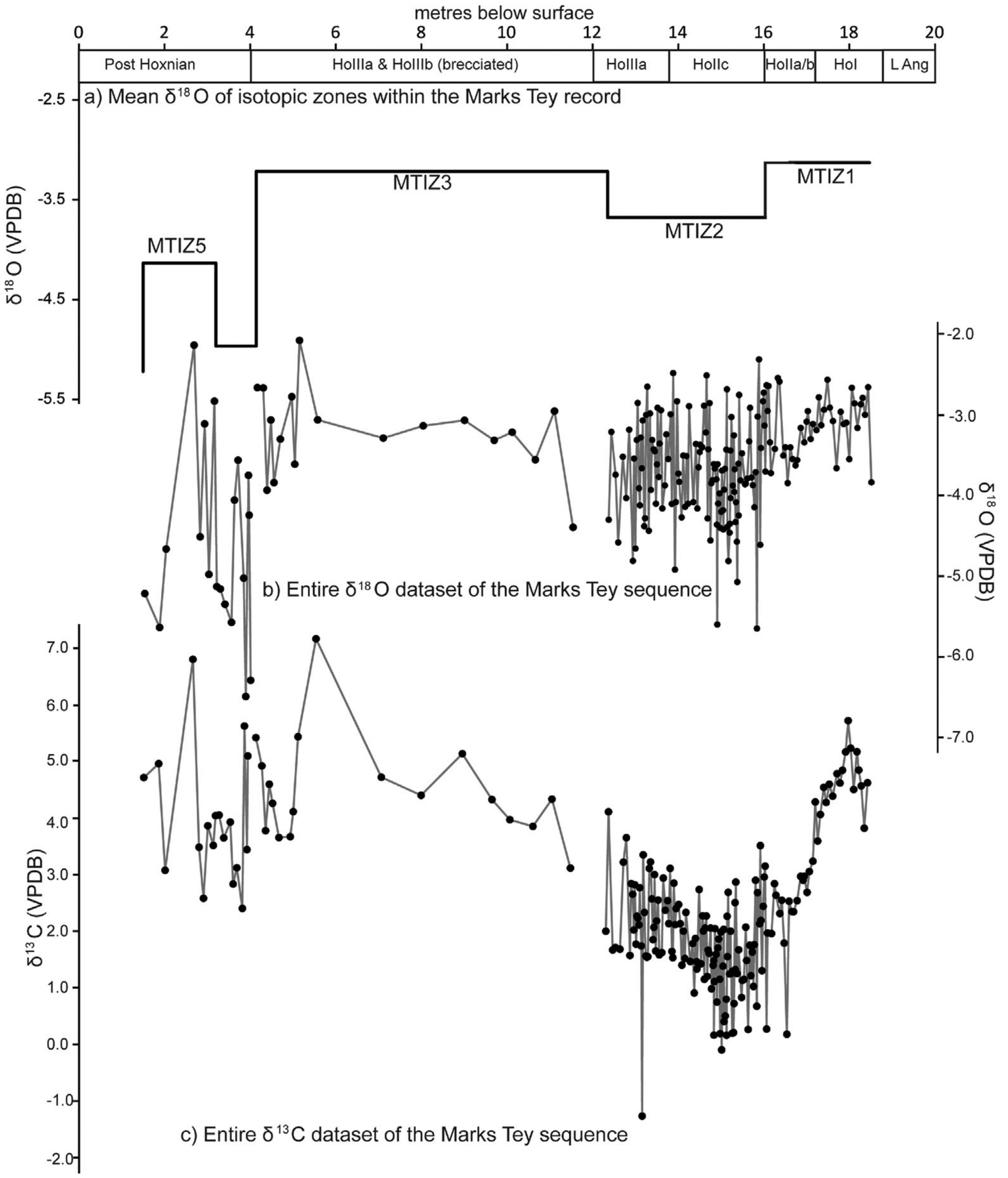

Figure 5. $\delta^{18}O$ and $\delta^{13}C$ record from the MT-2010 borehole. (a) Mean $\delta^{18}O$ values for each isotopic zone (see Table 1 for descriptive statistics), (b) $\delta^{18}O$ data plotted against depth (nd (c) $\delta^{13}C$ data plotted against depth. Note the increase in scatter within the $\delta^{18}O$ data in the varved section (equivalent to MTIZ-2). $\delta^{13}C$ values are for much of the record consistent with those recorded in open system lakes where dissolved carbon has equilibrated with atmospheric CO_2, although in the later part of the record more enriched values occur, this may reflect greater anoxic decay in the water body or lower plant activity in the surrounding catchment. Note that there is no evidence in the sediments for detrital contamination, a suggestion supported by the lack of co-variance within the $\delta^{18}O$ and $\delta^{13}C$ dataset. That is, where $\delta^{13}C$ values become more enriched there is no concomitant change in $\delta^{18}O$ values (see Supplementary information Table S2).

and Black, 1934), and calcium carbonate content using a Bascomb Calcimeter (Gale and Hoare, 1991).

Micromorphology

Samples of undisturbed sediment were taken from each lithofacies for the production of thin sections. The aim of this was to support the macroscopic description of the different lithofacies with observations made at the microscale. Furthermore, microscopic analysis of the sediments can provide information on detrital inwashing, which may act as a source of detrital contamination during the isotopic analysis of the endogenic carbonate. Thin sections were prepared from fresh sediment blocks ($100 \times 30 \times 20$ mm) using standard impregnation techniques involving a slow-curing crystic resin developed in the Centre for Micromorphology at Royal Holloway, University of London (Palmer *et al.*, 2008). Thin sections were analysed using an Olympus BX-50 microscope with magnifications from 20× to 200× and photomicrographs were captured with a Pixera Penguin 600es camera.

Pollen

Sediment samples of 1 cm^3 were taken for pollen analysis. In total, 88 samples were prepared from the lowermost 8 m of the core (Fig. 4b) and ten samples were prepared from the uppermost 6.5 m of the core (Fig. 4a). Samples were prepared following standard techniques, including sample weighing, treatment with sodium pyrophosphate ($Na_4P_2O_7$), hydrochloric acid (HCL, 10%), hydrofluoric acid (HF, 40%), heavy liquid separation with sodium polytungstate, acetolysis ($C_2H_4O_2$), and slide preparation using glycerine jelly. A tablet with a known number of *Lycopodium* spores (Stockmarr, 1971) was added to the samples before preparation to enable concentrations (grains g^{-1}) to be calculated (lowermost 88 samples prepared with *Lycopodium* batch 177745, uppermost ten samples prepared with *Lycopodium* batch 1031).

Biomarkers

Lipid biomarkers were extracted from 18 freeze-dried and homogenized subsamples of ca. 1 cm^3 following the microwave-assisted extraction methodology of Kornilova and Rosell-Melé (2003). Known concentrations of 2-nonadecanone, 5α-cholestane and hexatriacontane (all Sigma-Aldrich) were added as internal standards. An aliquot of each lipid extract was separated into apolar, ketone and polar fractions using silica column chromatography (5% H_2O) using *n*-hexane, dichloromethane and methanol, respectively. The apolar fractions were analysed by gas chromatography-mass spectrometry (GC-MS), using a 30-m HP-5MS fused silica column (0.25 mm i.d. 0.25 μm of 5% phenyl methyl siloxane). The carrier gas was He, and the oven temperature was programmed as follows: 60–200 °C at 20 °C min^{-1}, then to 320 °C (held 35 min) at 6 °C min^{-1}. The mass spectrometer was operated in full-scan mode (50–650 amu s^{-1}, electron voltage 70 eV, source temperature 230 °C). Quantification was achieved through comparison of integrated peak areas in the total ion chromatograms and those of the internal standards.

Stable isotopes

Two hundred samples for stable isotope analysis were taken from individual carbonate laminations using a fine bladed scalpel and needle under a magnifier stand. It is common procedure to sieve bulk lacustrine sediment samples, before isotopic analysis, to remove the coarser fraction, typically greater than either 125 or 63 μm (Marshall *et al*., 2002; Leng *et al*., 2010; Candy *et al*., 2015), which is more likely to contain detrital material or shell fragments. This was not undertaken in this study because the sediments are silt grade and the process of sieving removes no further material from the sample. All samples were then left to dry, powdered and then weighed using a Mettler Toledo XP6 microbalance. The $\delta^{18}O$ and $\delta^{13}C$ values of each sample were determined by analysing CO_2 liberated from the reaction of the sample with phosphoric acid at 90 °C using a VG PRISM series 2 mass spectrometer in the Earth Sciences Department at Royal Holloway. Internal (RHBNC) and external (NBS19, LSVEC) standards were run every four and 18 samples, respectively. The 1σ uncertainties are 0.04‰ ($\delta^{18}O$) and 0.02‰ ($\delta^{13}C$). All isotope data presented in this study are quoted against VPDB.

Results

Core stratigraphy and sedimentology

The composite stratigraphy and sedimentology of the MT-2010 core is presented in Fig. 3. Based on changes in macroscale and microscale sedimentology throughout the sequence, the core can be divided into five main lithofacies associations (LFa-1 to 5). The main characteristics of these associations are summarized in Supplementary Table S1. The lowermost two units, LFa-1 (18.47–16.47 mbs or −2.67 tp −0.67 m OD) and LFa-2 (16.47–12.00 mbs or −0.67 to +3.8 m OD), comprise intact and *in situ* laminated deposits that are rich in organic and endogenic sediments. The laminations that comprise LFa-1 are irregular with carbonate laminations becoming more frequent upwards through this unit. LFa-2 comprises regular millimetre-scale laminations composed of the repeating organic/diatom/calcite triplets described earlier (Turner, 1970). Following the interpretation of Turner (1970) the sediments that comprise LFa-2 are considered varved, although this cannot be definitively proven. However, the irregular pattern of sedimentation seen in the sediments from LFa-1 is untypical of seasonal/annual laminations, questioning whether they are truly varved, as proposed by Turner (1970). Further detailed work on the pollen and diatom composition of the lamination subsets of both LFa-1 and -2 is required to definitively address this question and either prove/disprove their seasonal origin; this work is currently being undertaken but is beyond the scope of this paper. LFa-3 (12.00–4.10 mbs or +3.8 to +11.7 m OD) consists of blocks of brecciated and deformed sediment ranging in size from millimetre to centimetre scale. The blocks comprise sediments with the same regular lamination structure as LFa-2 but these frequently show evidence for folding and faulting. LFa-4 (4.10–1.35 mbs or +11.7 to +14.45 m OD) and LFa-5 (1.35–0 mbs or +14.45 to +14.80 m OD) are distinct from LFa-1-3 in that they are dominated by minerogenic material and represent a return to undisturbed and intact sediments after the brecciated beds of LFa-3. Both units are characterized by graded beds of silt, frequently of centimetre scale, with LFa-4 containing a higher $CaCO_3$ and total organic carbon (TOC) content than LFa-5. A noticeable increase in $CaCO_3$, from 20 to 40%, occurs mid-way through LFa-4.

Pollen

Two pollen diagrams spanning LFa-1, -2 and the lowermost part of LFa-3 (18.50–10.50 mbs) and the uppermost part of LFa-3 and LFa-4 and 5 (7.00–0.00 mbs) are presented in Fig. 4. The lower pollen diagram records a characteristic interglacial vegetation succession from a pre-temperate *Pinus–Betula*–Poaceae assemblage (18.50–16.75 mbs) through an early temperate assemblage of deciduous woodland taxa, primarily *Ulmus–Quercus–Alnus–Corylus* in varying proportions (16.75–13.60 mbs), to a post-temperate assemblage characterized by a rise in *Abies* (13.50 mbs onwards and into the brecciated sediments of LFa-3). A short-lived expansion of grass pollen at the expense of deciduous taxa (particularly *Corylus*) occurs between 15.00 and 14.50 mbs, producing a pronounced NAP phase. The brecciated sediments of LFa-3 in the upper pollen diagram also record a post-temperate vegetation assemblage dominated by *Abies*. From 4.00 mbs upwards, temperate tree pollen is still abundant but grass and other open ground taxa become increasingly significant.

The pollen sequence presented here is characteristic of the Hoxnian interglacial and replicates almost exactly the vegetation succession record in the 'GG' borehole of Turner (1970). The lowermost pollen diagram records pollen zones HoI, HoIIa–IIc and HoIIIa, with the onset of HoIIIb being broadly consistent with the transition to the brecciated sediments of LFa-3. The position of the NAP phase, midway through HoIIc, is identical to that described by Turner (1970). Between 7.00 and 4.00 mbs in the upper pollen diagram, the dominance of *Abies* highlights the continuation of HoIII. However, the overlying sediments (4.00–0.00 mbs) are marked by an expansion of non-arboreal taxa although temperate pollen is still present. Turner (1970) argued that a sedimentary hiatus occurred at this level, with the uppermost sediments in 'GG', the equivalent of LFa-4 and -5, accumulating during a cold-climate interval immediately after the Hoxnian. In Turner's model, sediments of HoIV are absent from the sequence and the temperate taxa represented in the pollen spectrum reflect reworking of material from interglacial sediments exposed at the lake margin, a common taphonomic issue in Hoxnian lake sequences (West, 1956; Coope and Kenward, 2007). This proposal is also accepted in this study to explain the pollen record of LFa-4 and -5 and these sediments are suggested to reflect a post-Hoxnian cold interval.

n-Alkane biomarkers

This study focuses on the distribution of the biomarkers found within the apolar fraction at Marks Tey, which is dominated by mid-chain (C_{20}–C_{26}) and long-chain (C_{27}–C_{33}) *n*-alkanes, with summed concentrations ranging from 5.4 to 26.3 μg g^{-1} and 2.4 to 18.6 μg g^{-1}, respectively. The dominant *n*-alkane varies between C_{27} (5.0–12.4 μg g^{-1}; 1773–1838 cm), C_{29} (1.8–3.4 μg g^{-1}; 1633–1211 cm) and C_{23} (4.5–10.2 μg g^{-1}; above 1016 cm). Minor contributions are also recorded from taraxast-20-ene (0–0.6 $\mu g\,g^{-1}$), which has been linked to Ericaceae (Pancost *et al.*, 2002) and the C_{31} methylhopanes sourced from aerobic bacteria (0–1.1 $\mu g\,g^{-1}$; Sinninghe Damsté *et al.*, 2004).

The relative distributions of the *n*-alkanes provide useful indicators of the dominant contributions of different higher plants to sediment sequences, alongside indications of the relative importance of aquatic algal inputs (Castañeda and Schouten, 2011). For example, short- and mid-chain *n*-alkanes dominate aquatic algae (C_{17}–C_{21}, Cranwell *et al.*, 1987) and submerged macrophytes (C_{23}–C_{25}, Ficken *et al.*, 2000), whereas long chain length *n*-alkanes (C_{27}–C_{33}) are important components of the epicuticular waxes of higher plants (Eglinton and Hamilton, 1967). The contribution of submerged vegetation is calculated using the P_{aq} ratio (Ficken *et al.*, 2000). Furthermore, the relative importance of different long-chain *n*-alkanes, described by the 'average chain length (ACL)' has been linked to changes to the higher plant assemblage and/or shifts in temperature and humidity (Gagosian and Peltzer, 1986; Hinrichs *et al.*, 1997; Rinna *et al.*, 1999; Pancost *et al.*, 2003; McClymont *et al.*, 2008). In contrast, the P_{wax} ratio is believed to distinguish between contributions from plant roots and the above-ground parts, albeit based on evidence from peatland environments (Zheng *et al.*, 2007; Ronkainen *et al.*, 2013).

Values of the ACL, P_{aq} and P_{wax} indices all vary across the MT-2010 sequence (Fig. 3). The lowest ACL values (28.5–29) occur within sediments of LFa-1 (HoI) and LFa-4 (the post-Hoxnian sediments). ACL values peak in LFa-2, HoIIc–HoIIIa (29.5–29.8), and then decline in the later part of the interglacial, LFa-3 or HoIIIb (29–29.5). Note that the samples analysed from HoIIIb are from the brecciated zone, and the extracted material includes both brecciated fragments and the intra-fragment clay, which were mixed after freeze-drying (it was not possible to separate these components before extraction). These lower values may therefore represent an 'averaging' of the ACL values of HoIIIb sediments and the clays that were deposited post-brecciation. P_{wax} values show a similar pattern to ACL values in that they peak in LFa-2 (0.7–0.8 in HoIIc and HoIIa) but are low in the early (ca. 0.6 in LFa-1/HoI) and late (ca. 0.4 in LFa-3/HoIIIb) interglacial. There is a slight increase in P_{wax} values from HoIIIb into the post-Hoxnian sediments (0.5–0.6 in LFa-4). In contrast, P_{aq} values show the reverse trend to P_{wax} (reflecting the different emphasis on long- versus mid-chain *n*-alkanes between the two indices). P_{aq} values decline from 0.55 in HoI, are low (0.25–0.38) through HoIIa, IIb, IIc and IIIa, before increasing to 0.68–0.73 through HoIIIa–IIIb. A slight decline to values of 0.60 occurs in HoIIIb.

$\delta^{18}O$ and $\delta^{13}C$ values of the lacustrine carbonates

The Marks Tey sequence is divided into five Isotopic Zones, MTIZ (Fig. 5, Table 1 and Supplementary information Table S2). These zones are delimited on the basis of variations and patterns with the $\delta^{18}O$ signal, rather than the $\delta^{13}C$ signal. This is because in all isotopic studies of lacustrine marl sequences in the British Quaternary, it is the $\delta^{18}O$ values that provide the most useful climatic/environmental data because of their relationship to prevailing temperature (as outlined above). $\delta^{13}C$ values record more localized information on hydrology, biological activity, carbon sources and organic decay (Leng and Marshall, 2004; Candy *et al.*, 2015).

Table 1. Descriptive statistics of the MT-2010 isotopic dataset presented as summary data for the entire dataset (top row) as summary data for the individual isotopic zones.

	Covariance *P*	r^2	Mean $\delta^{18}O$ (‰)	1σ	Mean $\delta^{13}C$ (‰)	1σ
Dataset	0.23	0.05	−3.64	0.76	1.60	1.52
MTIZ 1	0.18	0.03	−3.13	0.37	2.32	1.31
MTIZ 2	0.31	0.20	−3.75	0.62	0.77	0.84
MTIZ 3	0.26	0.31	−3.22	0.53	3.49	0.92
MTIZ 4	−1.19	0.36	−4.97	0.97	3.43	2.23
MTIZ 5	0.26	0.03	−4.13	1.27	3.12	1.34

MTIZ-1 (18.47–15.90 mbs, LFa-1, pollen zone HoI–HoIIb)

The $\delta^{18}O$ (mean = −3.13‰) and $\delta^{13}C$ (mean = 2.32‰) values in MTIZ-1 are high with relatively low standard deviations ($\delta^{18}O$ 1σ = 0.4; $\delta^{13}C$ 1σ = 1.3). $\delta^{18}O$ values decrease from 18.28 mbs (−2.65‰) to 16.48 mbs (−3.40‰). This declining trend is also seen in the $\delta^{13}C$ signal but the scale of the decline is much greater; from 3.62‰ at 18.37 mbs to −0.82‰ at 16.48 mbs. There is no co-variance between $\delta^{18}O$ and $\delta^{13}C$ ($r^2 = 0.03/P = 0.18$).

MTIZ-2 (16.48–12.00 mbs, LFa-2, pollen zones HoIIc–IIIa)

MTIZ-2 is characterized by average $\delta^{18}O$ and $\delta^{13}C$ values that are lower than MTIZ-1 ($\delta^{18}O$ mean = −3.75‰ and $\delta^{13}C$ mean = 0.77‰). The standard deviation in $\delta^{18}O$ increases from that of MTIZ-1 ($\delta^{18}O$ 1σ = 0.6) while that of $\delta^{13}C$ decreases but is still high ($\delta^{13}C$ 1 σ = 0.8). $\delta^{13}C$ values continue to decrease across the boundary between MTIZ-1 and -2 to a low of −1.10‰ at 15.06 mbs, after which they increase across the rest of this zone. Within MITZ-2 there is no clear trend in $\delta^{18}O$, although the lowermost part of this zone (15.88–14.71 mbs) has a lower mean $\delta^{18}O$ value (−3.92‰) than the uppermost part (14.71–12.00 mbs) which has a mean $\delta^{18}O$ value of −3.61‰. The lower mean $\delta^{18}O$ value for 15.88–14.71 mbs is a function of two factors: (i) the lowest individual $\delta^{18}O$ values occur between these depths, and (ii) the occurrence of a 30-cm section of sediment (15.02–14.71 mbs) where $\delta^{18}O$ values are persistently low (mean $\delta^{18}O$ value = −4.06‰). These low values occur at the same depths within the sequence as the NAP phase. There is no co-variance between $\delta^{18}O$ and $\delta^{13}C$ values ($r^2 = 0.20/P = 0.31$).

MTIZ-3 (12.00–4.00 mbs, LFa-3, HoIIIa–IIIb)

MTIZ-3 covers the 'brecciated' zone. Samples for isotopic analysis from this lithofacies were taken from carbonate laminations from within the brecciated fragments at regular depths across this interval. This lithofacies does not have stratigraphic integrity due to the brecciation and disturbance. Thus, although on the basis of pollen content the fragments of lake sediments that make up this zone (HoIIIb) are younger than the sediments of LFa-2 and older than the sediments of LFa-4, within LFa-3 depth does not equate to age. The isotopic value of the samples taken from across this lithofacies therefore does not provide information on how the environment evolved across this interval but provides an indication of the isotopic characteristics of lacustrine carbonates that precipitated during this time. The mean $\delta^{18}O$ (−3.22‰, 1σ = 0.53) and $\delta^{13}C$ (3.49‰, 1σ = 0.91) values both show an increase from those of MTIZ-2. MTIZ-3 contains some of the highest $\delta^{18}O$ values of the whole record. There is no evidence of co-variance between $\delta^{18}O$ and $\delta^{13}C$ values ($r^2 = 0.31$).

MTIZ-4 (4.11–3.19 mbs) and MTIZ-5 (3.19–1.51 mbs)

The sediments of lithofacies 4 are characterized by major oscillations in $\delta^{18}O$ values, with a zone (MTIZ-4) of relatively low values (mean $\delta^{18}O = -4.97‰$, $1\sigma = 0.97$) between 4.11 and 3.19 mbs and a zone of relatively high values (mean $\delta^{18}O = -4.13‰$, $1\sigma = 1.27$) between 3.19 and 1.51 mbs (MTIZ-5). It is these characteristics that are used to define MTIZ-4 and -5. As indicated by the high standard deviation values of MTIZ-4 and-5, there is significant scatter within the data of both zones; however, the $\delta^{18}O$ values effectively show a decrease, in MTIZ-4, to the lowest values of the whole dataset and then an increase, in MTIZ-5, to values as high as those seen in MTIZ-1, -2 and -3. At the end of MTIZ-5, $\delta^{18}O$ values decrease again to values consistent with those seen in MTIZ-4. $\delta^{13}C$ shows no trend, with mean values (MTIZ-4 = 3.43‰, $1\sigma = 2.23$; MTIZ-5 = 3.12‰, $1\sigma = 1.34$) being relatively consistent across the two zones. In both of these zones r^2 values are low, indicating little co-variance between $\delta^{18}O$ and $\delta^{13}C$ (MTIZ-4 $r^2 = 0.36$; MTIZ-5 $r^2 = 0.03$).

General characteristics of the Marks Tey isotopic record

The isotopic dataset presented above is characteristic of an open-system lake basin in a temperate climate (Fig. 5). In general the $\delta^{13}C$ values (mean = 1.6‰, $1\sigma = 1.52$) are typical of lake waters that have equilibrated with atmospheric CO_2 (Talbot, 1990; Leng and Marshall, 2004), although in places (early MTIZ-1 and late MTIZ-3) carbon values are significantly higher. It is likely that this reflects the presence of anoxic conditions at the lake bed, resulting in an increase in methanogenesis leading to a consequent rise in $\delta^{13}C$ values (Talbot, 1990). The absence of any significant relationship between $\delta^{18}O$ and $\delta^{13}C$ values (the r^2 of the whole data set = 0.05, never exceeding 0.36 for any isotopic zone) suggests that evaporation was not a major environmental control on lake basin hydrology (Leng and Marshall, 2004).

It is also proposed that the dataset is unaffected by detrital contamination. Isotopic values of samples of the carbonate-rich Lowestoft Till, which underlies the Marks Tey basin and is the most likely source of detrital contamination, are plotted against the lacustrine carbonate dataset (Fig. 6). The detrital values and the lacustrine samples overlap, which is not unusual as marine limestone, the source of the carbonate in the Lowestoft Till, and lake carbonates have similar $\delta^{13}C$ values (Candy *et al.*, 2015). However, the overlap occurs between Lowestoft Till samples and samples from LFa-2, the endogenic-rich sediments that show the least evidence for detrital contamination. There is no overlap between the isotopic values of the minerogenic rich sediments of LFa-4 and -5, the units that have the greatest potential for detrital contamination, and those of the Lowestoft Till samples. It is therefore suggested that the impact of detrital contamination on the Marks Tey isotopic record is negligible, with the overlap observed between the isotopic values of bedrock samples and the samples from LFa-2 being coincidental.

It is also important to note that some shifts within the isotopic dataset may be a function of changes in sedimentary style. For example, there is a large increase in the standard

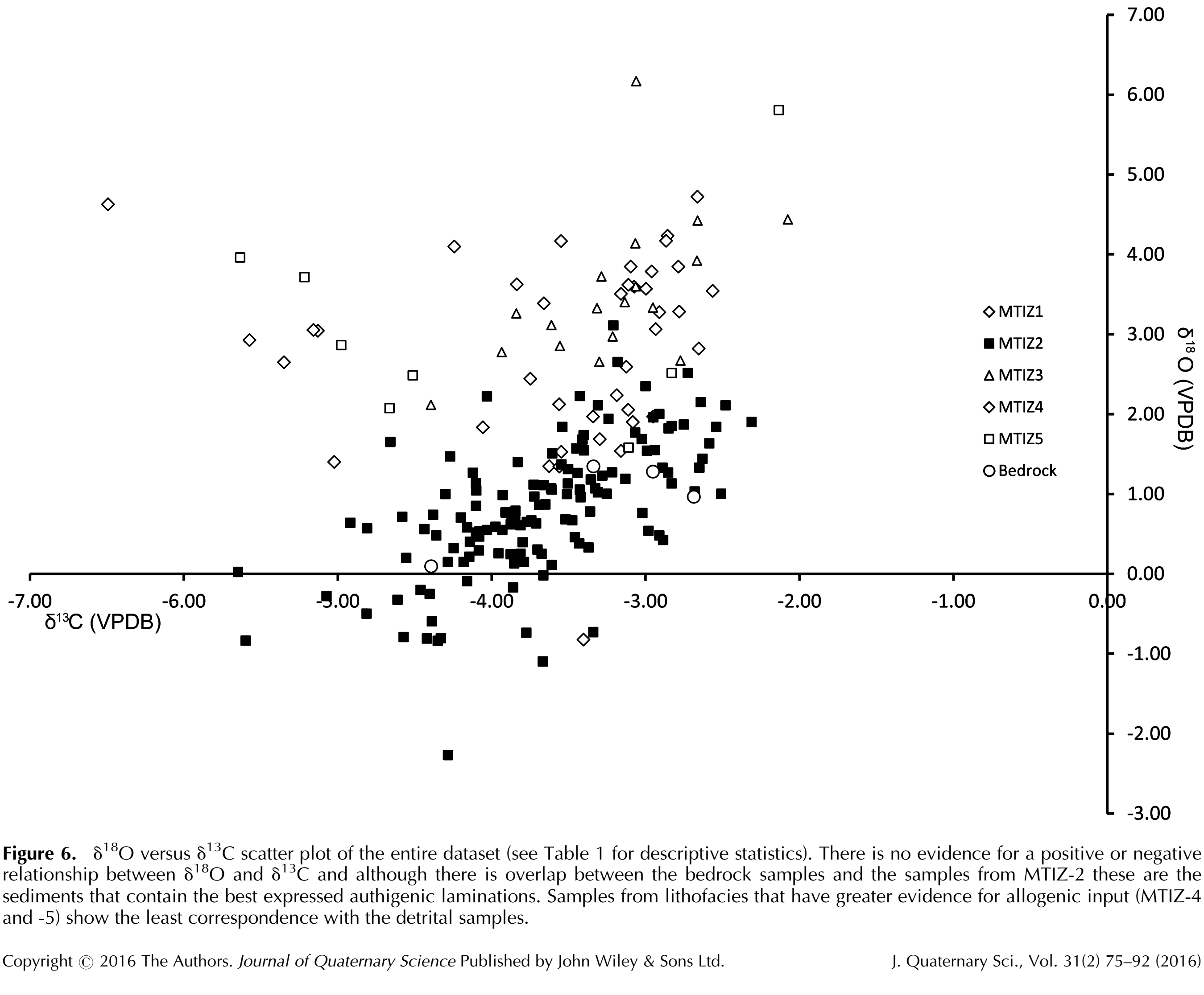

Figure 6. $\delta^{18}O$ versus $\delta^{13}C$ scatter plot of the entire dataset (see Table 1 for descriptive statistics). There is no evidence for a positive or negative relationship between $\delta^{18}O$ and $\delta^{13}C$ and although there is overlap between the bedrock samples and the samples from MTIZ-2 these are the sediments that contain the best expressed authigenic laminations. Samples from lithofacies that have greater evidence for allogenic input (MTIZ-4 and -5) show the least correspondence with the detrital samples.

deviation of $\delta^{18}O$ values from MTIZ-1 to MTIZ-2. If the sediments within LFa-2 are interpreted as varves, but those within LFa-1 are non-annual in nature (see above for discussion and associated caveats), the shift in standard deviation could simply reflect the resolution of the archive: in this instance, a shift from a sediment where each carbonate bed/lamination may reflect many years or even decades of accumulation (LFa-1) to a unit where each lamination represents carbonate that has accumulated in a single summer (LFa-2). This increase in resolution, because of a change in sedimentology, will automatically produce an increase in scatter within the dataset. In such contexts, this change in scatter has no environmental/climatic significance, but averaging the $\delta^{18}O$ datasets from MTIZ-1 and -2 removes the effect of changes in sample resolution and allows the $\delta^{18}O$ values of these two datasets to be compared. Consequently, the decrease in mean $\delta^{18}O$ values that occurs from MTIZ-1 to -2 is considered to have an environmental significance.

Discussion

Lithostratigraphy, pollen stratigraphy and n-alkane biomarkers of the Marks Tey sequence

The lithostratigraphy and the pollen stratigraphy of MT-2010 replicate the sequence recovered from borehole GG by Turner (1970). MT-2010 records intact sediments preserving continuous sedimentation across the early to middle part of the Hoxnian interglacial (HoI–HoIIIa, in LFa-1 and -2). The later part of the interglacial (HoIIIa and b) is preserved in the brecciated and deformed sediments of LFa-3. As with the GG borehole sequence, there is no evidence for intact sediments of HoIV, and consequently the very end of the Hoxnian interglacial is absent from this sequence. The minerogenic sediments of LFa-4 and -5 directly overlie the brecciated sediments of HoIIIb (LFa-3). The accumulation of deposits dominated by allogenic sediments, accompanied by a decrease in $CaCO_3$ concentrations to 15–20%, suggests that these lithofacies represent a cold climate interval. Reduced vegetation cover during such an interval would increase allogenic sediment supply and dilute the input of endogenic material (Palmer *et al.*, 2015).

That the sediments of MT-2010 record the major part of an interglacial and the subsequent return to cold climate conditions is supported by the biomarker studies. This pattern contrasts with the biomarker-derived indicators of a more stable interglacial climate, interrupted by two short-lived climate oscillations that occurred within the Holsteinian interglacial at Dethlingen lake, Germany (Regnery *et al.*, 2013), although this is likely to be a function of two factors. Firstly, the biomarker analysis at Marks Tey has been carried out at a lower resolution that at Dethlingen, with the aim of characterizing the broad structure of the interglacial rather than looking for evidence for abrupt and short-term change. Secondly, the later abrupt event described by Regnery *et al.* (2013), if present in the Marks Tey sequence, would lie within the brecciated sediments of LFa-3, making it unlikely that it would be identified in this study. It is important to highlight, however, that in both studies the patterns and interpretations of the *n*-alkane ratios are supported by multiple lines of evidence. *n*-Alkane chain length (ACL) is generally used as a broad indictor of climatic conditions, with longer chain lengths occurring under warmer/drier conditions (Gagosian and Peltzer, 1986; Hinrichs *et al.*, 1997; Rinna *et al.*, 1999; Pancost *et al.*, 2003; McClymont *et al.*, 2008), although the strength of the relationships can vary between plant genera and species (e.g. Sachse *et al.*, 2006; Hoffmann *et al.*, 2013; Ronkainen *et al.*, 2013).

In the Marks Tey record, ACL increases across HoI and is highest during HoII and HoIIIa, suggesting that peak interglacial conditions may have occurred during these pollen zones. ACL then declines within the brecciated sediments of LFa-3, which may be a function of the later part of the interglacial being cooler, although as mentioned earlier this may also be an artefact of sediment mixing at this level, between the brecciated fragments and the clay matrix that they occur within. ACL values of sediments from lithofacies 4 and 5 are low, supporting the idea that these represent a cold climate interval. P_{aq} provides an indication of the proportion of *n*-alkanes within each sample coming from higher (terrestrial plants) relative to aquatic plants. The low P_{aq} values from HoI to HoIIIa indicate a dominance of input from terrestrial plants (Ficken *et al.*, 2000). From HoIIIb onwards, there is a shift to a greater contribution from submerged/floating macrophytes (Ficken *et al.*, 2000) given the increasing P_{aq} signal. The mirror-image pattern of the P_{wax} ratio (high in HoI–HoIIIa, low from HoIIIb) may also indicate a shift towards a reduced contribution from higher plants and emergent macrophytes through time (Zheng *et al.*, 2007). Although this increase in aquatic plant markers could be related to sediment mixing as described above, alternative environmental controls on the observed trends include either: (i) long-term changes in nutrient availability in the basin as the lake system evolves over time, or (ii) an environmentally controlled reduction in vegetation cover around the lake basin.

$\delta^{18}O$ stratigraphy of the Marks Tey sequence

The $\delta^{18}O$ stratigraphy of the Marks Tey sequence (Figs 5 and 7) shows two zones, MTIZ-1 (HoI–HoIIb) and MTIZ-3 (HoIIIa and b), with high mean $\delta^{18}O$ values (−3.13 and −3.22‰, respectively), separated by a zone, MTIZ-2 (HoIIc and HoIIIa), with a lower mean $\delta^{18}O$ value (−3.75‰). In the interpretation of this signal it must be remembered that the samples of MTIZ-3 come from brecciated fragments of lake deposits. The magnitude of the isotopic decline between MTIZ-1 and MTIZ-2 and the recovery into MTIZ-3 is relatively small (ca. 0.6–0.7 ‰). Across a large part of the Hoxnian interglacial (HoI–HoIIIb) there is therefore minimal evidence for major, long-term shifts in mean $\delta^{18}O$ values. This is not the case for the post-Hoxnian sediments of LFa-4 and -5 where values decline to −6.49‰ early in MTIZ-4 (LFa-4), increase to −2.83‰ early in MTIZ-5 and then decline again to −5.63 at the end of MTIZ-5 (Fig. 5).

It is common, in most lacustrine carbonate $\delta^{18}O$ studies in Western Europe, to interpret isotopic shifts in the context of temperature changes (Marshall *et al.*, 2002, 2007; Leng and Marshall, 2004; Candy, 2009; Candy *et al.*, 2011, 2015; van Asch *et al.*, 2012). Such interpretations are based on the well-recorded control that air temperature exerts on the $\delta^{18}O$ of rainfall (Dansgaard, 1964; Rozanski *et al.*, 1992, 1993; Darling, 2004), which is, in turn, the main control on the $\delta^{18}O$ of meteoric waters and, through groundwater recharge, the main control on the $\delta^{18}O$ of lake waters. Following this approach, the $\delta^{18}O$ value of the Marks Tey carbonate record can be interpreted in terms of long-term temperature variations if the following two assumptions are accepted. Firstly, that the $\delta^{18}O$ value of the Marks Tey lacustrine carbonates is primarily controlled by the $\delta^{18}O$ value of the lake water from which it precipitates. Secondly, that the modification of the $\delta^{18}O$ value of lake water by processes such as evaporation is minimal. If this is accepted then the following observations can be made. That the Hoxnian

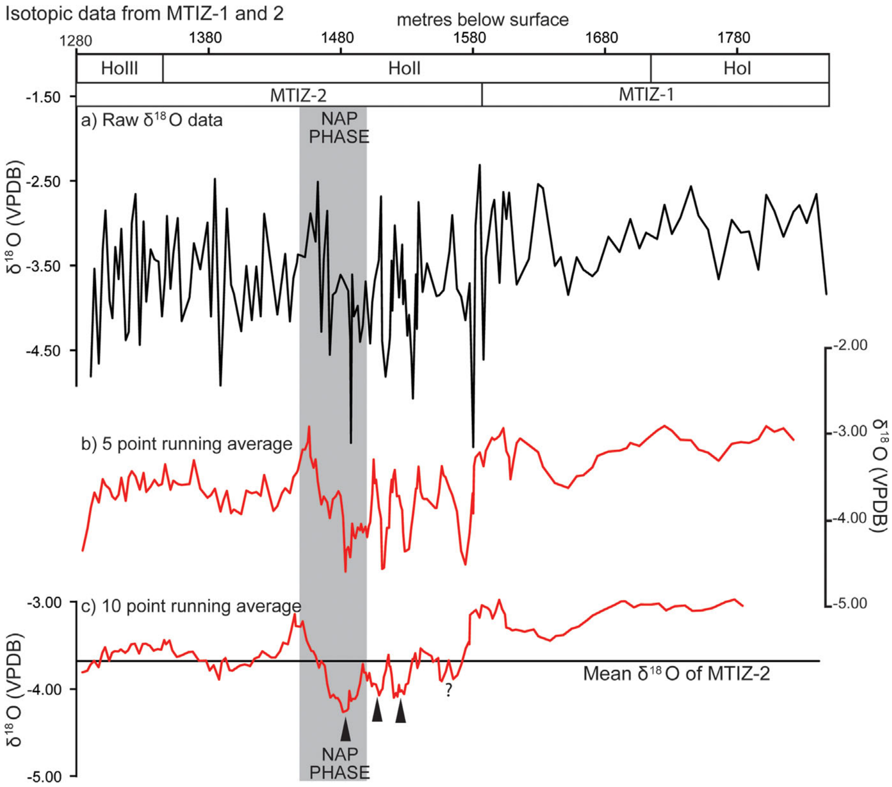

Figure 7. Raw $\delta^{18}O$ data for MTIZ-1 and -2 (a) and that data converted into five- (b) and ten- (c) point running averages. Much of the scatter within the $\delta^{18}O$ data from the varved sediments is smoothed out when averaged because the distribution of high and low values are part of no consistent pattern or trend. This is not the case within HoIIc where intervals, such as that associated with the NAP phase, contain persistent low $\delta^{18}O$ values that produce clear isotopic excursions/events within the running averages. Three noticeable isotopic 'depletion' events occur during HoIIc, the last of which occurs in association with the NAP phase.

interglacial was characterized by long-term climate stability, with minimal changes in mean $\delta^{18}O$ value from HoI to HoIII. The higher mean $\delta^{18}O$ values in MTIZ-1 and MTIZ-3, relative to MTIZ-2, suggest that two temperature peaks may have occurred during the Hoxnian, one early in the interglacial and one in the mid-/late part of the interglacial. It must be stressed, however, that the changes in mean $\delta^{18}O$ values across the Hoxnian sediments preserved at Marks Tey are small, and consequently if there are two periods of higher temperatures then the degree of cooling between them must have been relatively minor (as witnessed by the persistence of deciduous woodland across this interval).

The pronounced variation in $\delta^{18}O$ values that occurs within LFa-4 is suggested to represent a more pronounced climatic oscillation for two main reasons. Firstly, the scale of the $\delta^{18}O$ variations (ca. 3‰) represents a large shift and, secondly, the oscillation occurs in association with significant changes in allogenic versus endogenic sediment input. In LFa-4 (MTIZ-4), the low $\delta^{18}O$ values occur in association with reduced $CaCO_3$ concentrations (20% or less), implying increased allogenic input. The increase in $\delta^{18}O$ values in the later part of LFa-4 (MTIZ-5) occurs in association with a return to higher $CaCO_3$ concentrations (25–30%), implying an increase in endogenic carbonate precipitation in the lake basin or a reduction in allogenic inputs from the catchment. The reduction in $\delta^{18}O$ values in the later parts of MTIZ-5 is associated with a reduction in $CaCO_3$ (20% or less), again implying that allogenic inwashing has increased and erosion is dominating the catchment. The magnitude of the isotopic shifts, combined with the evidence for changing sedimentology, is therefore used to suggest that during LFa-4, a significant post-Hoxnian cold interlude occurred, followed by a short-lived return to interglacial-like temperatures, i.e. the $\delta^{18}O$ values at this point are consistent with those of the underlying Hoxnian beds. This warm event is then followed by a return to lower $\delta^{18}O$ values and, consequently, lower temperatures.

The above discussion assumes that temperature is the driving control on $\delta^{18}O$. However, it is worth considering two other environmental factors that could have influenced the $\delta^{18}O$ signal presented here. Firstly, several recent studies have argued that under interglacial climates, rainfall, both amount and seasonality, may have a major influence on the $\delta^{18}O$ value of lacustrine carbonates in western and central Europe (Hammarlund *et al.*, 2002; Nitychoruk *et al.*, 2005; Diefendorf *et al.*, 2006; Candy *et al.*, 2015). These studies have proposed that the occurrence of more continental-style climates at the onset of interglacials may produce, because of high levels of summer rainfall (enriched in $\delta^{18}O$) relative to winter rainfall (depleted in $\delta^{18}O$), high $\delta^{18}O$ values in recharging surface and groundwaters. As the interglacials progress, the onset of more maritime climates, characterized by an increase in winter rainfall (and therefore an increase in the contribution of relatively depleted $\delta^{18}O$ precipitation), will result in a decline in the $\delta^{18}O$ value of recharging waters. This has been used to explain an early interglacial peak in the $\delta^{18}O$ signal of both Holocene and Hoxnian/Holsteinian lacustrine carbonate records (Nitychoruk *et al.*, 2005; Diefendorf *et al.*, 2006; Candy *et al.*, 2015). It is therefore possible that the decline in $\delta^{18}O$ values that occurs between MTIZ-1 and -2 could reflect a shift in the seasonality of rainfall regime in the early part of the Hoxnian rather than, or as well as, a decrease in temperature.

Secondly, it should also be highlighted that any isotopic record of the early part of MIS 11 may be influenced by the uniquely long and protracted deglaciation that occurred across Termination V (Rohling *et al.*, 2010). Termination V, as well as being characterized by the most extreme deglaciation of the past 500 000 years (Droxler *et al.*, 2003), also lasted for approximately twice the length of most other terminations (Rohling *et al.*, 2010; Vázquez Riveiros *et al.*, 2013). This would have meant that during the early part of MIS 11, when interglacial conditions were already established in much of Europe, Atlantic waters (the source of British precipitation) would have remained relatively enriched with respect to $\delta^{18}O$, because a significant proportion of $H_2{}^{16}O$ was still held within the residual ice masses. This may have elevated the $\delta^{18}O$ value of rainfall during the early part of the Hoxnian until significant ice melt had occurred. The decline in $\delta^{18}O$ values across pollen zones HoI–HoII could therefore represent a complex environmental signal of temperature-, precipitation- and source water-controlled change.

Significant oscillations in $\delta^{18}O$ values occurred during MTIZ-2 in association with the NAP phase of HoIIc. The scatter in the $\delta^{18}O$ dataset of the varved sediments of LFA-2 obscures this pattern, but the depth at which the NAP phase occurs is relatively unusual within MTIZ-2 in that it correlates with a zone of consistently low $\delta^{18}O$ values. The mean $\delta^{18}O$ value of MTIZ-2 is $-3.75‰$ and between 15.10 and 14.70 mbs, the depth of the NAP phase, all the individual $\delta^{18}O$ values (mean $= -4.08‰$) are either below the mean value or within uncertainties of the mean value. This situation is unique within MTIZ-2 where low $\delta^{18}O$ values do occur but rarely as part of a consistent pattern of low values. The occurrence of a $\delta^{18}O$ excursion or event in association with the NAP phase can clearly be seen if a five- or ten-point running average is plotted through the dataset (Fig. 7b,c) as this removes scatter and highlights areas of persistently low or high values.

The NAP phase is therefore characterized by a pronounced low in $\delta^{18}O$ values and is followed by a return to higher mean $\delta^{18}O$ values, although the pattern of $\delta^{18}O$ variations before the NAP phase is complex. This section of the sequence is characterized by three zones of low $\delta^{18}O$ values, separated by a short-lived return to higher $\delta^{18}O$ values. It has long been debated whether the NAP phase of HoIIc is a climatic event, possibly analogous to the 8.2-ka event of the early Holocene, or a regional event driven by wildfire or volcanic eruptions (Kelly, 1964; Turner, 1970; Kukla, 2003; Koutsodendris *et al.*, 2012). If it is assumed that changing $\delta^{18}O$ values in the Marks Tey record reflect temperature variability then the NAP phase does occur in association with a significant climatic oscillation. This oscillation may not, however, be a single cold event but part of a more complex series of cold/warm oscillations that occurred in the early part of this interglacial.

The Marks Tey sequence in the context of MIS 11 in Britain, Europe and the North Atlantic

The application of stable isotopic analysis to the Marks Tey sequence has identified three key characteristics of the Hoxnian (MIS 11c) interglacial in Britain. Firstly, over the scale of the entire interglacial, the climate is one of relative stability. Secondly, at least one post-Hoxnian (MIS 11c) climatic oscillation occurs. Finally, the NAP phase is associated with a significant decline in $\delta^{18}O$ values and therefore is likely to have occurred in association with an abrupt cooling event. These three points will be discussed here with reference to other sites in Britain and the North Atlantic. Although several stable isotopic records have been produced from MIS 11 sequences in Europe, comparing them with the Marks Tey record is problematic. The main stratigraphic proxy for the La Celle (northern France) tufa sequence (Dabkowski *et al.*, 2012) is a molluscan assemblage that makes a direct correlation with the pollen zones of Marks Tey difficult. Equally, comparing the Pianico pollen record of northern Italy (Mangili *et al.*, 2007, 2010a, b) with that of Britain is problematic, as it is unclear how compatible the vegetation history of northern Europe is with that of southern Europe during MIS 11c (see Koutsodendris *et al.*, 2011, 2012). Koutsodendris *et al.* (2012) have presented $\delta^{18}O$ values for the Holsteinian lake sediments at Dethlingen but detailed analysis is restricted to the interval spanning the OHO, with little data from the rest of the interglacial. The complexity of understanding $\delta^{18}O$ shifts in the early part of interglacials has been discussed above with reference to the Polish site of Ossówka (Nitychoruk *et al.*, 2005). However, it is the variability in $\delta^{18}O$ values that occurs at Dethlingen in association with the OHO that is significant for the discussion here.

While absolute alignment of British and European MIS 11c records is problematic, there is also much debate about how the pollen history of this interglacial should be aligned with marine records of the North Atlantic. In southern Europe, it would appear that woodland conditions persisted for most of MIS 11c, albeit with a noticeable decline ca. 415 ka (Desprat *et al.*, 2005; Tzedakis, 2010). This has led researchers such as Ashton *et al.* (2008) to assume that the Hoxnian interglacial spans most of MIS 11c (ca. 30 ka), a suggestion supported by estimated varve counts at Marks Tey (Shackleton and Turner, 1967). In contrast, varve chronologies from lacustrine records of the Holsteinian indicate that the woodland phase of this interglacial persisted for only ca. 15 ka (Müller, 1974; Kukla, 2003; Koutsodendris *et al.*, 2011, 2012), i.e. only half of MIS 11c. Koutsodendris *et al.* (2012) have suggested that the Hoxnian/Holsteinian should therefore be correlated with the second half of MIS 11c, which in North Atlantic sea surface temperature (SST) records (Fig. 8) contains the thermal maximum, and in Mediterranean pollen records contains the maximum expansion of woodland (Tzedakis, 2010).

In the following section we follow the approach of Ashton *et al.* (2008) in assuming that the Hoxnian equates to the entirety of MIS 11c, because: (i) the extrapolation of annual-layer counting (Shackleton and Turner, 1967) remains the only age model for this period in Britain, and (ii) the model of Koutsodendris *et al.* (2012) requires Britain to remain treeless for 15 ka, despite the fact that North Atlantic SST records indicate that during this 'treeless' interval, this region had achieved Holocene levels of warmth. Consequently, the isotopic variability recorded in Marks Tey is compared, but not directly correlated to, the climatic complexity seen across the entirety of MIS 11c in North Atlantic SST records. The varve chronology for a 'short' Holsteinian (Koutsodendris *et al.*, 2012) is, however, compelling and it should be considered that the Marks Tey isotopic record may actually represent climatic variability in the second half of MIS 11c.

The Hoxnian as an interglacial of prolonged climatic stability

The suggestion that the Hoxnian interglacial is a time of prolonged climatic stability is consistent with records from the North Atlantic (McManus *et al.*, 1999, 2003; Martrat *et al.*, 2007; Stein *et al.*, 2009). In particular, the work of McManus *et al.* (2003) on ODP 980 has highlighted the occurrence of relatively stable SSTs at the same approximate latitude as Britain. This is consistent with the $\delta^{18}O$ signal of the Marks Tey sequence, which shows remarkably little variation in mean $\delta^{18}O$ value across the major part of the interglacial (HoI–HoIIIb). The apparent occurrence in the Marks Tey sequence of an early and a late peak in $\delta^{18}O$ values is, if these are interpreted as being temperature maxima, consistent with many of the North Atlantic records of MIS 11 (Fig. 8), e.g. MD01-2443 (Martrat *et al.*, 2007), ODP 982 (Lawrence *et al.*, 2009), U1313 (Stein *et al.*, 2009) and MD03-2699 (Voelker *et al.*, 2010; Rodrigues *et al.*, 2011). All of these workers have shown that North Atlantic SSTs during MIS 11c were characterized by an early (centred on ca. 420 ka) and a late (centred on 405 ka) temperature peak, separated by a short-lived and relatively minor cooling interval. It is important to highlight that the difference in the

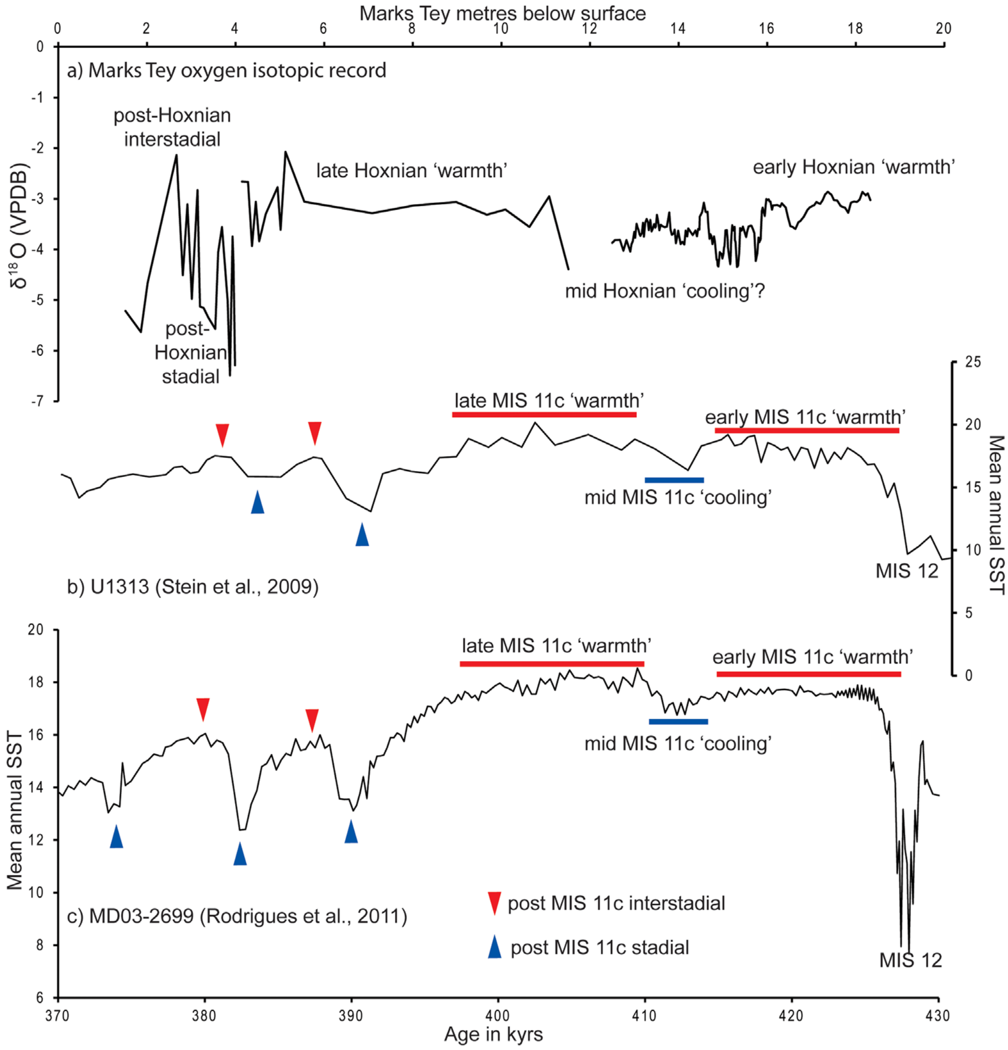

Figure 8. Comparison between the MT-2010 δ^{18}O record, plotted against depth (data from MTIZ-1 and -2 shown as a five-point running average), and sea surface temperature records from two high-resolution North Atlantic cores (U1313 from Stein *et al.*, 2009; MD03-2699 from Rodrigues *et al.*, 2011). Note that there is no absolute chronology associated with the Marks Tey record so this figure *does not* represent a direct correlation but is used to highlight the fact that some of the patterns seen in the MT-2010 isotopic records are consistent with those seen in SST records from the North Atlantic. This comparison uses the approach taken by Ashton *et al.* (2008) that the Hoxnian interglacial corresponds to all of MIS 11c, as opposed to that of Koutsodendris *et al.* (2010, 2012), which assumes that the Hoxnian/Holsteinian correlates with only the later warm peak of MIS 11c (see Candy *et al.*, 2014, for discussion).

magnitude of the early and late Hoxnian δ^{18}O peaks is small: 0.1‰ difference in the average of the two isotopic zones in question (MTIZ-1 and 3) and 0.49‰ difference in the peak value from each zone. This is, however, consistent with SST estimates from U1313 (Stein *et al.*, 2009) and MD03-2699 (Rodrigues *et al.*, 2011), which show that the temperature difference between the late and the early peaks is in the order of ca 1 °C, even without considering the uncertainties associated with these temperature estimates.

If the high δ^{18}O values that occur in HoI are not a function of higher temperatures but rather a function of precipitation changes or the prolonged pattern of deglaciation, then the fact that the highest δ^{18}O values occurred in HoIII (the later part of MTIZ-2 and MTIZ-3) might imply a thermal maximum relatively late in the Hoxnian. This would again be consistent with North Atlantic SST records that typically show that the later temperature peak in MIS 11c contained the interglacial thermal maximum (Stein *et al.*, 2009; Rodrigues *et al.*, 2011). It is also consistent with British records of the Hoxnian, which show tentative evidence for climates during HoIII being warmer than during HoI and HoII. In the last interglacial, MIS 5e, fossils of the most thermophilous species are concentrated in pollen zone IpII, i.e. the early temperate phase (Candy *et al.*, 2010). In Hoxnian sequences, there is less convincing evidence for the extreme warmth that is seen in MIS 5e but where evidence for exotic plant species does occur, e.g. seeds of *Trapa natans*, which indicate summer temperatures of >20 °C (Candy *et al.*, 2010), they are found in sediments of HoIII, i.e. the late temperate phase (Gibbard and Aalto, 1977; Coxon, 1985; Gibbard *et al.*, 1986). Although the evidence of warm climate conditions in HoIII at Marks Tey is heavily reliant on isotopic data from brecciated sediments, it is consistent with palaeoclimatic data from other Hoxnian sites.

A post-Hoxnian/MIS 11c stadial/interstadial oscillation

The sedimentological/isotopic oscillations observable in LFa-4 suggest the occurrence of climatic instability directly after the Hoxnian. There is now a growing body of evidence, supported by the new work at Marks Tey, for the existence of short-lived warm climate episodes after the end of the Hoxnian interglacial in Britain. This is seen at both the Hoxne typesite (West, 1956; Ashton *et al.*, 2008) and Quinton (Coope and Kenward, 2007), where a combination of floral and faunal data indicates that post-HoIII, a cold interlude occurred during which summer temperatures decreased by ca. 7 °C and winter temperatures decreased by at least 10 °C from their interglacial peak. This cold interlude was followed by a return to warm, but not fully interglacial conditions (Coope and Kenward, 2007; Ashton *et al.*, 2008; Candy *et al.*, 2014). That this later 'interstadial' occurred directly after the Hoxnian and does not relate to a younger interglacial, MIS 9 for example, is supported by the amino acid racemization values of shells from this bed at Hoxne (Ashton *et al.*, 2008; Penkman *et al.*, 2011) and the mammalian fauna, which has Hoxnian affinities (Schreve, 2000, 2001).

Ashton *et al.* (2008) tentatively correlated the cold interlude with MIS 11b and the later interstadial with MIS 11a. A lack of radiometric dating makes it impossible to provide a direct correlation between this interstadial and the marine record or even to prove that the post-Hoxnian climatic oscillations seen at Quinton, Hoxne and now Marks Tey are even the same event. This situation is complicated by the fact that there is growing evidence from marine (Martrat *et al.*, 2007), ice core (Jouzel *et al.*, 2007) and long lake records (Prokopenko *et al.*, 2006) that the latter part of MIS 11 was punctuated by multiple stadial/interstadial cycles (Candy *et al.*, 2014). The record from Marks Tey does not aid the stratigraphic correlation of post-Hoxnian stadial/interstadial events between sites or provide a greater understanding of the number of stadial/interstadial events that occurred in Britain during the transition from MIS 11 to 10. However, it supplies the most detailed evidence yet for post-Hoxnian climatic complexity, indicating that the British mainland was sensitive and susceptible to such events.

The occurrence of abrupt climatic events in pre-Holocene interglacials

There is now clear evidence for abrupt climatic events in the early part of the current interglacial (Daley *et al.*, 2011). A key research question is whether such events are common in pre-Holocene interglacials (Tzedakis *et al.*, 2009). Within Europe, the most convincing evidence for an abrupt 'event' in any pre-Holocene interglacial is the NAP phase/Older Holsteinian oscillation (OHO) that occurs in multiple Hoxnian/Holsteinian interglacial sequences across Europe (Koutsodendris *et al.*, 2011, 2012; Candy *et al.*, 2014). The duration of this event is approximately 300 years (Turner, 1970; Koutsodendris *et al.*, 2011, 2012) and although it is not (yet) possible to absolutely date this event, it is frequently considered synchronous across Europe because it occurs in the same position in the regional pollen stratigraphy (Koutsodendris *et al.*, 2012).

Although the NAP phase/OHO clearly represents an ecological 'event', it has long been debated whether it is a response to a climatic trigger (Kelly, 1964; Koutsodendris *et al.*, 2011, 2012), a wildfire (Turner, 1970) or a major volcanic eruption (Diehl and Sirocko, 2007). The study of the $\delta^{18}O$ signal of the lake sequences in which the NAP phase/OHO occurs would address this issue. In many of these lacustrine records, however, carbonate precipitates are absent. Koutsodendris *et al.* (2012) attempted to address this by analysing the $\delta^{18}O$ value of diatom silica across the OHO interval at Dethlingen. Although a reduction in $\delta^{18}O$ values was evident in the OHO, the section of the sequence analysed was short, making it difficult to establish whether this shift was significant in the context of long-term changes within the $\delta^{18}O$ signal. Koutsodendris *et al.* (2012) were, however, able to show that the shift in $\delta^{18}O$ values and other palaeoenvironmental proxies, such as diatom assemblages (Koutsodendris *et al.*, 2013), did occur before the shift in vegetation assemblage, implying that the former led the latter. This would also appear to be the case at Marks Tey, as the $\delta^{18}O$ values decline before the vegetation response (Fig. 7). The data presented here allow for the first time: (i) a record of the NAP phase to be directly compared with the $\delta^{18}O$ value of lacustrine carbonate, and (ii) the $\delta^{18}O$ values associated with the NAP phase to be placed within the context of the $\delta^{18}O$ signal of an entire interglacial. As outlined above, the NAP phase is associated with an interval of persistently low $\delta^{18}O$ values and therefore occurs during a period of climatic cooling. Figure 7 clearly shows that three distinct episodes of low $\delta^{18}O$ values occurred during HoIIc, implying both that HoIIc is characterized by numerous abrupt cooling events, and that the NAP phase occurs in association with the last of these.

Koutsodendris *et al.* (2012) have argued that the OHO may be analogous to the 8.2-ka event of the early Holocene in terms of both its relative timing within the stratigraphy of the interglacial and the expression of this event in terrestrial sequences. Several studies in Europe have used the $\delta^{18}O$ of early Holocene lacustrine carbonates in Europe to characterize the 8.2-ka event (Daley *et al.*, 2011); in Britain the best example is from Hawes Water (Marshall *et al.*, 2002, 2007). A comparison between the isotopic record of Marks Tey and that of Hawes Water highlights two similarities between the isotopic characteristics of the abrupt events in these records. Firstly, the 8.2-ka event is just one of several isotopic oscillations, and therefore abrupt cooling events seen in the early Holocene record of Hawes Water (Marshall *et al.*, 2007). This sequence records a number of abrupt events in the early Holocene, including the 8.2-ka event, the 9.3-ka event and other unnamed $\delta^{18}O$ events. The 8.2-ka event is not therefore an isolated cooling event but one of several early Holocene abrupt climatic shifts, a situation comparable to that seen in HoIIc. Secondly, the 8.2- ka event in the Hawes Water sequence is characterized by a negative shift away from the moving average of the $\delta^{18}O$ dataset of ca. 0.8‰. The largest isotopic shift in the ten-point running average in HoIIc is 0.6‰ from the average of the dataset. With respect to the relationship between temperature and the $\delta^{18}O$ of freshwater carbonates (Andrews, 2006; Candy *et al.*, 2011) the difference between these shifts is negligible, implying that both sequences record an isotopic response to a temperature shift of a comparable magnitude.

It is apparent that the NAP phase in Marks Tey occurs in association with a period of climatic instability. It is therefore likely that it represents a vegetation response to an abrupt cooling event and, given its position within the early Hoxnian, this event may be comparable to the 8.2-ka event. The Marks Tey sequence has the potential to address this further as: (i) the $\delta^{18}O$ signal of this time interval can, with further sampling, be constructed in much greater resolution, and (ii) if it can be proved that these laminations are varved then this archive can be used to quantify the absolute duration of the isotopic events, the intervals between them and the lag time between climatic shifts and vegetational response/recovery. The duration of these events can then be compared with the known duration of the 8.2-ka event, as preserved in the Greenland ice cores (Thomas *et al.*, 2007). The data presented in this study provide strong evidence that abrupt events occurred in pre-Holocene interglacials and that these events had significant impacts on the ecosystems of Western Europe.

Conclusions

The recovery and analysis of overlapping boreholes (MT-2010) from the Hoxnian deposits at Marks Tey replicate the previously published record of Turner (1970) and allow, through the application of pollen, biomarker and stable isotopic analysis, the following conclusions to be drawn:

- The sequence preserved in MT-2010 records sediments spanning the pre-, early and late temperate phases of the Hoxnian and cold-climate sediments deposited in the immediate post-Hoxnian period.

- The $\delta^{18}O$ record of the endogenic carbonates from this sequence records the climatic structure of this interglacial which is characterized by: (i) relative long-term climatic stability, (ii) the existence of a possible early and late temperature peak and (iii) a stadial/interstadial oscillation in the immediate post-Hoxnian.
- The $\delta^{18}O$ record from the early temperate phase records several short-term fluctuations that are interpreted as abrupt cooling events.
- The most pronounced of these oscillations occurs in association with the well-documented NAP phase during which grassland expands at the expense of deciduous woodland taxa.
- The implication is that the NAP phase represents a response to a short-lived cooling event (possibly analogous to the 8.2-ka event of the early Holocene) and consequently provides one of the best recorded examples of an abrupt climatic event in a pre-Holocene interglacial.

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Abbreviations. MIS, Marine Isotope Stage; AP, arboreal pollen; NAP, non-arboreal pollen; OHO, Older Holsteinian Oscillation; LGIT, Last Glacial to Interglacial Transition; OD, ordnance datum; TOC, total organic carbon; ACL, average chain length; MTIZ, Marks Tey Isotopic Zone; SST, sea surface temperature

Supplementary information

Additional supporting information can be found in the online version of this article:

Table S1. Descriptions of the sedimentary characteristics of the five main lithofacies that comprise the MT-2010 sequence (taken from Sherriff *et al.*, 2014).

Table S2. Sediment data from the Mrks Tey sequence: % organic carbon, % calcium carbonate and carbon and oxygen isotopic values.

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