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Late Holocene ecohydrological and carbon dynamics of a UK raised bog: impact of human activity and climate change

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Abstract

Understanding the ecohydrological responses of peatlands to climate change is particularly challenging over the late Holocene owing to the confounding influence of anthropogenic activity. To address this, a core spanning the last ~2400 years from a raised bog in northern England was analysed using a comprehensive suite of proxy methods in an attempt to elucidate the drivers of change. A testate amoebae-based transfer function was used to quantitatively reconstruct changes in water table depth, supported by humification analysis and a plant macrofossil-derived hydroclimatic index. Pollen and plant macrofossil data were used to examine regional and local vegetation change, and human impacts were inferred from charcoal and geochemistry. Chronological control was achieved through a Bayesian age-depth model based on AMS radiocarbon dates and spheroidal carbonaceous particles, from which peat and carbon accumulation rates were calculated. Phases of both increased and decreased bog surface wetness (inferred effective precipitation) are present, with dry phases at *c.* AD 320-830, AD 920-1190 and AD 1850-present, and a marked period of increased effective precipitation at *c.* AD 1460-1850. Coherence with other records from across Northern Europe suggests that these episodes are primarily driven by allogenic climatic change. Periods of high bog surface wetness correspond to the Wolf, Spörer and Maunder sunspot activity minima, suggesting solar forcing was a significant driver of climate change over the last ~1000 years. Following the intensification of agriculture and industry over the last two centuries, the combined climatic and anthropogenic forcing effects become increasingly difficult to separate due to increases in atmospheric deposition of anthropogenically derived pollutants, fertilising compounds, and additions of wind-blown soil dust. We illustrate the need for multiproxy approaches based on high-resolution palaeoecology and geochemistry to examine the recent trajectories of peatlands.

Keywords: Holocene, climatic change, human influence, multiproxy, peatland, UK.

1. Introduction

Peat-based hydrological reconstructions provide an important insight into climatic changes during the Holocene (Aaby, 1976; Blackford, 2000; Chambers and Charman, 2004; Barber, 2007; Chambers et al., 2012). Ombrotrophic peatlands represent very useful ecosystems for the investigation of past climate as precipitation is the sole source of water, thus surface hydrology is a function of effective precipitation (precipitation – evapotranspiration). There has been debate over whether the primary driver of bog surface wetness (BSW) is temperature (Schoning et al., 2005; Barber and Langdon, 2007) or summer water deficit (Charman et al., 2004; Charman, 2007). However, it has been suggested recently that climatic context has a major influence on drivers of BSW, temperature having more influence in a continental setting whereas summer water deficit has the strongest relationship with water table depth in oceanic regions (Lamentowicz et al., 2008). Charman et al. (2009) propose that UK peatland BSW is driven by variations in precipitation, relating primarily to the warm-season water deficit period.

Surface vegetation and microbiota are responsive to changes in effective precipitation, and a wide range of biological remains and geochemical signals are often preserved *in situ* within accumulating peat. Radiocarbon dating of above-ground plant remains in peat (e.g. *Sphagnum*) is also comparatively straightforward, allowing well-constrained reconstructions of Holocene palaeoenvironmental change (Aaby, 1976; Mauquoy et al., 2002). Good chronological control is an essential pre-requisite for meaningful reconstructions, allowing comparison of peat-based records with lacustrine, speleothem and ice core records (Barber, 2007; Swindles et al., 2010). However, separating climatic signals from the effects of human influence in ombrotrophic peatlands is a persistent problem (Chambers, 1993; Rydin and Jeglum, 2006). Direct anthropogenic impacts such as drainage, peat cutting, grazing, burning, or indirect influences such as surrounding land-use changes and atmospheric deposition can heavily influence the ecohydrology of peatland ecosystems (Blackford, 2000; Charman, 2002). For example, the landscape of the UK has been subject to increasing levels of anthropogenic influence since

Mesolithic hunter-gatherers initiated landscape change as early as *c.* 7000 BC (Bush, 1988), as evidenced by charcoal layers found in early Holocene peats and lake deposits throughout the British uplands (Simmons and Innes, 1996). Thus, many peatlands have been influenced by climate change and human impacts for much of the Holocene. Established methods for palaeohydrological reconstruction include testate amoebae-based transfer functions, plant macrofossil and humification analysis. Each proxy has specific strengths and weaknesses; for example, testate amoebae are particularly effective where peat stratigraphy is dominated by a single eurytypic *Sphagnum* species rendering the interpretation of humification and macrofossil analysis unreliable, and humification is useful where peat is highly decomposed (Blackford and Chambers, 1993; Caseldine et al., 2000; Barber et al., 2003; Langdon and Barber, 2005). Combining several proxies potentially gives more robust reconstructions, particularly when combined with proxies of human activities such as geochemical data and pollen analysis (e.g. Lomas-Clarke and Barber, 2007; Coombes et al., 2009; Lamentowicz et al., 2009; Schofield et al., 2010). However, studies of recent environmental change that utilise a comprehensive array of biological, geochemical and physical proxies in ombrotrophic peatlands are relatively uncommon (De Vleeschouwer et al., 2009; van der Knaap et al., 2011; De Vleeschouwer et al., 2012)

The drivers of ecohydrological change in peatlands during the late Holocene become increasingly convoluted as abrupt climatic changes occur contemporaneously with elevated levels of human impact, resulting in complex palaeoenvironmental signals. Therefore a need for a fully integrated approach to palaeoenvironmental reconstructions using multiple proxy methods and empirical evidence is apparent. This study presents a ~2400 year multiproxy surface wetness reconstruction from an ombrotrophic raised bog in northern England, UK. The palaeohydrological data is supplemented by a record of past regional and local vegetation changes, geochemical and physical/sedimentological data. The aim of this study is to present comprehensive multiproxy data from a north-western European raised bog so that drivers of change can be elucidated. The specific aims are to 1) reconstruct local and regional

palaeoenvironmental changes for the last c. 2000 years; 2) identify major changes in bog surface wetness and compare the records inferred from three proxies; 3) determine the source and level of anthropogenic influence; 4) determine whether shifts in bog surface wetness and vegetation correlate with anthropogenic factors or climatic events.

2. Study site

Malham Tarn Moss (MTM) (Fig.1) is a coalescence of three distinct ombrotrophic raised bogs (covering around 40 ha) that forms the major part of a diverse mire complex adjacent to Malham Tarn, a shallow marl lake in the Yorkshire Dales National Park, England (54°05'47"N, 002°10'30"W, 380 m asl.). MTM represents a suitable site for this investigation, as detailed historical records and archaeological data provide evidence of human activity around the area from the Mesolithic period onwards. High-quality climate data from local instrumentation and historical records of land management and agrarian activities are also available. In addition, there are several other palaeoenvironmental reconstructions from raised bogs in the north of England (and therefore within a similar climatic context) with which comparisons can be made.

The underlying geology is impermeable Silurian slates mostly covered by undulating glacio-fluvial deposits, with Carboniferous limestone surrounding the area. It is bordered by a broad belt of calcareous fen to the north/north-west and a narrower 'lagg' to the south/south-west that separate MTM from the surrounding agricultural rough pasture. The fens are fed by strongly calcareous waters draining from the surrounding limestone rocks. To the east, MTM borders directly onto the tarn. The bog stratigraphy indicates that marl sediments, deposited in open water, were succeeded by fen vegetation, evident in highly humified peat containing *Phragmites* and woody remains above the mineral substrate (Pigott and Pigott, 1959). The switch to bog vegetation occurred around 8000 years ago, as shown by subsequent *Sphagnum* peat. Ombrotrophic *Sphagnum austinii*-rich peat then accumulated up to a depth of 5 – 6 m over much of the bog (Pigott and Pigott, 1959; Piggott and Piggott, 1963).

In terms of human impact on Tarn Moss, archaeological artefacts attest to the regular presence of Mesolithic hunter-gathers around Malham Tarn, and settlement during the Bronze Age (Raistrick and Holmes, 1962). The pollen record of Piggot and Piggot (1959) indicates regional woodland clearance from the late Neolithic/early Bronze Age period onwards. It is likely that animals were grazed on and around MTM throughout most of the historical period, and an estate book of the 1780s shows the fen areas divided into agricultural compartments (Cooper and Proctor, 1998). The level of the tarn was raised in 1791 by around 1.3 m (Holmes, 1965) causing rapid erosion of the peat bank, resulting in a vertical exposed peat face several metres high. Peat bank erosion rates have been estimated at $\sim 0.07 \text{ m y}^{-1}$ (Pentecost, 2000). Surface patterns on MTM suggest part of the mire has been drained and cut for peat, though no written records have yet been found (Pentecost, 2000).

3. Methods

3.1. Field sampling

A 100 cm-long peat sequence was extracted from Malham Tarn Moss using a Russian-type D section corer (Jowsey, 1966) with a 50 x 5 cm chamber, using the parallel hole method and 20 cm overlap. Cores were extracted from a lawn microform (De Vleeschouwer et al., 2010) on the eastern cupola. Cores were placed in plastic guttering of a similar diameter to prevent movement in transit, wrapped in cling film, and stored at 4°C prior to laboratory analysis. A description of the core lithostratigraphy was carried out before subsampling following Tröels-Smith (1955).

3.2. Dating and age-depth model

Spheroidal Carbonaceous Particles (SCPs) are a component of fly-ash generated during high temperature combustion of fossil fuels, and therefore represent a stratigraphic age-equivalent marker in recent (post industrial revolution) peat profiles (Rose, 1990). Contiguous 1 cm subsamples were taken from the top 20 cm of the profile for SCP analysis and prepared using acid

digestion (Swindles, 2010). A known mass of material was slide mounted in Naphrax and the number of SCPs counted at 400x magnification. The resulting count is expressed as the number of SCPs per gram of dry peat. Calendar ages were assigned to the start of SCPs in the profile, the rapid rise, and peak concentration using established comparative data from UK lake profiles dated by radiometric methods (Rose et al., 1995; Rose and Appleby, 2005).

Twelve samples from the rest of the peat profile were selected for Accelerator Mass Spectrometry ^{14}C dating at the Queen's University Belfast $^{14}\text{CHRONO}$ laboratory. Sample preparation followed the recommendations of Piotrowska et al. (2011). Where present, identifiable *Sphagnum* remains were handpicked from samples and washed in high-grade deionised water before being stored in vials with deionised water and a few drops of 7% HCl. Where *Sphagnum* was not present, Ericaceous woody material, macro-charcoal, and other above-ground plant remains including leaves and *Eriophorum* spindles were selected. Care was taken to remove Ericaceous roots to prevent contamination with younger carbon as well as possible reservoir effects (cf. Kilian et al., 1995). In highly humified samples, the material was sieved through 300 μm and 125 μm sieves and the remaining humin fraction washed and retained for analysis. The > 300 μm material was examined for datable above-ground remains which were removed, washed and included with the humin fraction.

The age-depth model was constructed using the 'Bacon' piece-wise linear accumulation model of Blaauw and Christen (2011) in R (R Core Team, 2012). Bacon produces potentially more realistic age-depth models than direct linear interpolation methods by using *a priori* knowledge of accumulation rates to account for the variability of environmentally-dependant peat accumulation. The procedure is an improved version of that proposed by Blaauw and Christen (2005) employing a greater number of shorter sections resulting in more robust and flexible chronologies (Swindles et al., 2012a). The model allows quantification of the total chronological error, generating a maximum age probability (MAP) at 1 cm intervals together with maximum and minimum ages within the modelled age range (MAR). Ages discussed in this study are

prefixed *circa* (*c.*) which signifies the MAP. Where ages are followed by subscript text, this signifies ages within the MAR (e.g. *c. AD 1500*_{MAR AD 1445-1580}). The age-depth model output was used to extrapolate peat accumulation rates in 1 cm increments.

3.3. Testate amoebae

Sub-sampling of the core for testate amoebae analysis was carried out at 1 cm contiguous resolution from the surface to 18 cm, then at 2 cm resolution down to 100 cm. Samples were prepared using a modified version of Charman et al. (2000). 2 cm³ peat samples were placed in 50 ml centrifuge tubes, 40 ml of hot water added, and agitated for 2 minutes using a 'whirly-mixer'. The contents were sieved through 300 µm and back sieved through 15 µm meshes, and samples stored in deionised water. Testate amoebae were identified at 200 – 400x magnification under light microscopy using standard taxonomic guides (Corbet, 1973; Ogden and Hedley, 1980; Charman et al., 2000). The taxonomic approach predominantly follows that of Charman et al. (2000), which groups some ecologically similar morphospecies with intraspecific variability into 'types'. Exceptions to this were the splitting out of *Assulina scandinavica* from the *Assulina seminulum* group (cf. Turner et al., 2013). Cladoceran remains of the genus *Alona*, and the Rotifer *Habrotrocha angusticollis* were also counted.

3.4. Plant macrofossils

The core was sub-sampled at 2 cm intervals for plant macrofossil analysis. Samples of 4 cm³ were washed over a 125 µm sieve to disaggregate the material and remove fine detritus, and identified under a low power stereomicroscope at 10 – 50x magnification using the quadrat leaf-count (QLC) method of Barber et al. (1994). Bryophyte leaves and epidermal tissue were mounted on slides and examined at 200 – 400x magnification for identification to lowest taxonomic level possible. The proportions of individual *Sphagna* were estimated by randomly selecting > 100 leaves for identification to species or section level under light microscopy at 100 – 200x magnification. Seeds, ericaceous leaves, macro-charcoal (> 125 µm), oribatid mites and *Cenococcum* spp. sclerotia were quantified by totalling the number identified within quadrat

squares. Macrofossils were identified with the aid of Gross-Brauckmann (1972), Smith (2004), Watson (1968), Daniels and Eddy (1990), and Mauquoy et al. (2007). Nomenclature of *Sphagna* follows Daniels and Eddy (1990), other bryophytes Smith (2004), and vascular plants Stace (2010).

A modified Dupont Hydroclimatic Index (DHI) was calculated for the plant macrofossil data by weighted-averaging (cf. Dupont, 1986). The DHI has been found to be more reliable than detrended correspondence analysis (DCA) or non-metric multidimensional scaling (NMDS) ordination for indicating BSW, and can provide a good indicator of BSW variation in peat records where the profile incorporates both *Sphagnum*-dominated and *Eriophorum*/ericaceous-dominated sections (Daley and Barber, 2012). Each peat component was scored relative to its position along the hydrological gradient from driest to wettest; Unidentified organic matter (UOM) = 0, *Racomitrium lanuginosum* = 1, Ericaceae = 2, monocot undiff. = 3, *Eriophorum vaginatum* = 4, *Scirpus cespitosus* = 4, *Sphagnum sec. Acutifolia* = 5, *S. austinii* = 6, *S. papillosum* = 7, *S. sec. Cuspidata* = 8.

3.5. Pollen

Pollen samples were prepared using a modified version of Moore et al. (1991). Sub-samples of 1 cm³ were taken at 4 cm intervals through the peat core. One *Lycopodium* cf. *clavatum* tablet was added as an exotic marker following Stockmarr (1971), and samples were treated with NaOH before acetolysis. Material was stained with 0.2% aqueous safranin and stored in silicon oil. Samples were identified under light microscopy at 400 - 1000x magnification reaching a minimum count of 300 grains. Identification was aided by the taxonomic guides of Moore et al. (1991), Reille (1992), Beug (2004) and by reference to type slides within the School of Geography, University of Leeds. Plant taxonomy follows Stace (2010) and pollen nomenclature follows Bennett (1994), with the exception of 'Rumex-type' which groups all *R. acetosa*, *R. acetosella* and 'Oxyria-type' (Moore et al. 1991) pollen grains. Pteridophyte and Bryophyte spores were counted but were excluded from the total pollen sum, and Ericaceae were

identified to genus level where possible. The category of 'Ericaceae undiff.' includes highly degraded tetrad pollen grains where identification to genus was not possible; however, these pollen grains have been included in 'Total Ericaceae'.

3.6. Bulk density, loss-on-ignition, C/N ratio, C accumulation

Bulk density (BD), loss-on-ignition (LOI) and C/N ratio analysis was carried out at contiguous 1 cm resolution. Samples of 4 cm³ were weighed and oven dried at 105°C to calculate BD. Organic matter content was determined through loss-on-ignition by incinerating samples at 550°C for 4 hours (Chambers et al., 2011). Dried, fine-ground (< 250 µm) homogenised peat sub-samples were analysed for C and N content using a EuroEA 3000 Series Elemental Analyser. The recent (apparent) rate of carbon accumulation (RERCA) and long-term carbon accumulation rate (LORCA) expressed as g C m⁻² y⁻¹ were calculated from the age-depth model peat accumulation rates, bulk density and C content (cf. Tolonen and Turunen, 1996).

3.7. Humification

The colorimetric procedure follows that of Roos-Barraclough et al. (2004), a modified version of Blackford and Chambers (1993) which reduces reagent volumes and improves efficiency. Contiguous 1 cm peat samples were oven dried at 105°C and ground to a homogenous fine powder (< 250 µm) using a high-frequency ball mill. Sub-samples of 0.02 g were placed in test tubes and 10 ml of 8% NaOH added. The tubes were then shaken and placed in a water bath at 95 ±5°C for 60 minutes. Samples were made up to 20 ml with deionised water, shaken and left to stand for 60 minutes before being re-shaken and filtered through Whatman No.1 filter papers. Samples were diluted 1:1 with deionised water before colorimetric analysis at 550 nm wavelength using a Hach DR/2010 spectrophotometer. To reduce the effects of temporal colour-loss, samples were prepared in batches of 10 and the first analysis was repeated at the end of each batch to ensure no colour loss had occurred.

3.8. Geochemical analysis

Geochemical analysis for 23 trace elements (Ar, Ba, Br, Ce, Cl, Cs, Cu, Fe, K, Ni, P, Pb, Rb, S, Se, Si, Sr, Ta, Ti, W, Y, Zn, Zr) of which 12 key markers were selected for interpretation (Ti, Si, P, K, Pb, Fe, Cu, Zn, Br, Rb, Sr, Y), was carried out using X-ray fluorescence (XRF) core scanning, which allows for non-destructive, *in situ* XRF analysis of sediment cores from sub-millimetre resolution upwards. Core sections were scanned using a Cox Analytical Systems ITRAX XRF Core Scanner located at the Institute of Geography and Earth Sciences, Aberystwyth University. Cores were scanned at 500 μm step intervals with a count-time of 30 seconds per point using a Mo tube operating at 30 kV and 30 mA. The ITRAX geochemical data are output as counts and therefore should be considered semi-quantitative in nature (Croudace et al., 2006; Weltje and Tjallingii, 2008).

3.9. Palaeohydrological reconstruction

Quantitative palaeohydrology is reconstructed through a testate amoebae-based water table depth transfer function approach. The transfer function assumes a relationship between water table depth (WTD) and fossil testate amoebae assemblage based on modern species-environment data. The transfer function of Turner et al. (2013), based on a training set of 116 samples from northern England (including 25 from MTM) is used here. The weighted averaging (tolerance downweighted with inverse deshrinking) model has been rigorously tested through spatially-independent cross-validation (SICV) with species-environment data from Northern Ireland (Swindles et al., 2009) and Europe (Charman et al., 2007).

The reconstructed WTD, plant macrofossil DHI and detrended humification data were normalised to z-score (1 standard deviation) to allow direct comparison of proxies.

3.10. Time series analysis

The palaeohydrology z-scores for testate amoebae-based WTD reconstruction, humification and macrofossil DHI were subject to spectral analysis to detect periodicities. Many methods of time

series analysis require that the data are evenly spaced temporally, meaning unevenly spaced data must be interpolated before analysis leading to significant bias (Schulz and Mudelsee, 2002). Analysis was carried out using the REDFIT program in PAST (Hammer et al., 2001) (rectangular window (Schulz and Mudelsee, 2002)) which estimates the first-order autoregression (AR1) process parameter directly from unevenly spaced time series which is then transformed from the time domain into the frequency domain. Statistical significance of spectral peaks was tested using a parametric approach (90%, 95%, and 99% false-alarm levels).

4. Results

4.1. Lithostratigraphy, peat accumulation and age-depth model

The Tröels-Smith description of the core is presented in Table 1 and shown in Fig.6. Fibrous *Eriophorum* and monocotyledonous material dominates most of the core, with *Sphagnum* evident at 100-90 cm and *c.* 35-25 cm. The core undergoes a distinct colour change to a much darker brown/black from *c.* 12 cm to the surface.

Fig.2 shows the age-depth model for Malham Tarn Moss, with details of the individual AMS radiocarbon dates given in Table 2. The Bacon output indicates three outlier radiocarbon dates, UBA-19012 and UBA-19013 appearing older than expected and UBA-17214 much younger. Two of these samples were comprised primarily of the humin fraction with very small amounts of woody material, which can give less reliable results than above-ground plant material (Shore et al., 1995). It is possible that UBA-19012 was contaminated to some degree during the preparation stage; however, the MAR of this sample does marginally fall within the age-depth model. Three SCP dates were assigned to the model following Rose et al. (1995); start AD 1850 \pm 5 at 15.5 cm, rapid rise AD 1950 \pm 5 at 8.5 cm, peak AD 1977 \pm 2 at 5.5 cm, comparable to the Malham Tarn lake SCP profile of Rose and Appleby (2005). A well-defined subsurface peak in SCP profiles, such as the profile at MTM (supplementary file), has also been shown to

correspond with the Hekla AD 1947 tephra layer in Ireland (Swindles and Roe, 2006), and the Askja AD 1875 tephra layer by Amesbury et al. (2012).

The mean accumulation rate throughout the profile is 0.41 mm y⁻¹, and is < 0.5 mm y⁻¹ over much of the profile. Accumulation rates increase significantly from 60 cm to 40 cm (c. AD 910-1070) reaching peak accumulation of 2.7 mm y⁻¹ between 55-51 cm. A second phase of rapid peat accumulation is evident from 15 cm to surface; however, this should be interpreted with caution as the peat at this depth is unlikely to have been fully incorporated into the catotelm and thus subject to further decomposition and carbon loss. The LORCA over the full MTM core is 23.11 g C m⁻² y⁻¹ comparable with that of raised bogs reported by Turunen et al. (2001; 2002).

4.2. Testate amoebae water table reconstruction

The testate amoebae record and water table reconstruction is shown in Fig.3. Taxa with hydrological preferences consistent throughout studies from many regions of the world are represented in the MTM profile: reliable dry indicators *Hyalosphenia subflava*, *Trigonopyxis arcuata* type and *Assulina muscorum*, and wet indicator taxa *Amphitrema wrightianum*, *Archerella flavum* and *Arcella discoidea* type (Tolonen, 1986b; Charman and Warner, 1992; Tolonen et al., 1992; Charman, 1997; Woodland et al., 1998; Mitchell et al., 2001; Lamentowicz and Mitchell, 2005; Charman et al., 2007; Booth, 2008; Swindles et al., 2009). Abrupt wet shifts are apparent at 78, 60 and 12 cm (c. AD 280, AD 910 and c. AD 1920) inferred by the increase of *A. wrightianum* in the profile. A largely stable period of inferred drier conditions is apparent from 54-36 cm (c. AD 960-1200), where *Hyalosphenia subflava*, *Cyclopyxis arcelloidea* type, *A. muscorum* and *Diffugia pulex* are abundant. The WTD reconstruction shows a stable period of increased BSW from 28-16 cm (c. AD 1460-1850) where both *A. wrightianum* and *A. flavum* dominate the record. From around 12 cm to surface the profile is again dominated by *H. subflava* and indicates a rapid return to drier conditions and deeper water table.

4.3. Biostratigraphy

Stratigraphic percentage diagrams for testate amoebae, plant macrofossils and pollen are presented in Figs. 3 – 5. Biostratigraphic zones were assigned to each proxy through constrained Euclidian distance cluster analysis, with fine-tuning by eye based on assemblage changes. The testate amoebae and plant macrofossil diagrams show several distinct changes in sub-fossil assemblages through the profile. In the pollen record these changes are less marked, showing a more gradual pattern in terms of regional vegetation change. Although the zonation of the biological proxies does not coincide, a number of features emerge from the combined data (Fig. 6). Detailed zone descriptions are presented in Table 3.

4.4. BD, LOI, C/N and humification

The results of the bulk density, loss-on-ignition, C/N ratio and degree of humification are shown in Fig.7. LOI is relatively stable ($\pm 0.91\%$) from 100 cm to 8 cm (*c.* AD 1960), after which there is a rapid decline to the surface. From 100-95 cm both C/N ratio and BD decline, and this is the lowest BD figure in the profile. BD then gradually increases to 61 cm (*c.* AD 880) where after a slight reduction remains relatively stable before a gradual trend toward higher BD from around 30 cm (*c.* AD 1420) to the surface. The C/N ratio peaks between 81-57 cm (*c.* AD 190-940), and 56-20 cm (*c.* AD 950-1740). Although the humification data display a greater level of noise distinct patterns are evident. There is a rapid decrease in humification from 89-81 cm (*c.* 30 BC-AD 190), before increased decomposition between 81-70 cm (to *c.* AD 580). The rapid wet shift present in the testate amoebae record at 60 cm (*c.* AD 910) is mirrored in the humification profile, with a peak in transmission values. Light transmission then returns to much lower values indicating increased humification from 58-36 cm (*c.* AD 920-1190), supporting the indication of drier conditions in the biological proxy records. From 28 cm to 12 cm (*c.* AD 1460-1920) there is a very rapid decrease in humification from the highest degree of decomposition to the lowest indicated in the range (27.7% to 45.1% transmission) consistent with the wet

phase indicated in the biostratigraphy. In terms of humification and bulk density, samples from the acrotelm are not directly comparable with those in the catotelm.

4.5. Geochemistry

The elemental analysis from the XRF core scan is presented in Fig.8. Owing to the very high resolution analysis of this method (500 μm increments) the output displays a high degree of noise; however, patterns are evident in the data and a LOWESS smoothing function (Cleveland, 1979) (degree of smoothing = 0.01) was applied to aid data interpretation. The soil erosion indicator Ti (Holzer and Holzer, 1998) is relatively constant throughout the profile, then increases from around 15 cm (*c.* AD 1880). Yttrium (Y), a rare earth element of mineral origin (Murad, 1978), and Rubidium (Rb) have been associated with lithogenic deposition (Schofield et al., 2010). They are thus potential indicators of sporadic soil inputs, and both display a marked peak around 3 cm (*c.* AD 1990) to the present day. Several elements indicative of industrial pollutant deposition, such as Fe, Cu, Zn and Pb, increase significantly from around 10 cm to surface (*c.* AD 1950 onwards). Post-depositional mobilisation of Fe is improbable as there is no evidence of Fe increase with depth. Pb also registers a smaller peak at around 31 cm (*c.* AD 1380) and this may be indicative of lead smelting. Elements associated with agricultural fertilisers include P and K. P remains relatively constant through the profile displaying no trend; however, K has an overall increasing trend from 100 cm ($r^2= 0.22$), with a noticeable peak near the surface. The halogen bromine (Br), which is retained stratigraphically in accumulating peat (Roos-Barraclough et al., 2002), peaks between 26-14 cm (*c.* AD 1500-1890), coeval with the increased BSW phase apparent in the biological proxies and humification record.

4.6. Palaeohydrology: comparison of the proxies

The testate amoebae-based water table depth reconstruction, humification data and plant macrofossil DHI were normalised and plotted on the same axis to allow direct comparison of the three proxies (Fig.9). WTD depth errors have been removed for diagrammatic clarity. Although

there are clearly some anomalies, the three proxies show broadly similar changes in direction throughout the record. The wetter conditions are reflected in all three records at the base of the profile, and all register a drying trend from 100-89 cm (*c.* 410-30 BC). The rapid return to increased BSW registered in the testate amoebae record at 78 cm (*c.* AD 280) appears earlier in the humification record by approximately 65 years, and is not registered in the DHI at all. The proxies complement each other particularly well over the prolonged dry phase in the medieval period (*c.* AD 920-1190); however, the preceding rapid wet shift at around 60 cm (*c.* AD 890) is apparent in the testate amoebae and humification record only. Again, all proxies register the prolonged period of increased BSW between 28-16 cm (*c.* AD 1460-1850), although the humification record exhibits a large degree of noise, possibly an artefact of the higher sampling resolution. The rapid shift to drier conditions *c.* AD 1920 is evident in the testate amoebae and DHI profiles; however, the humification results are rendered unreliable in the acrotelm. Although magnitudes of change should be viewed with caution when employing a transfer function approach to palaeohydrological reconstruction (Turner et al., 2013), this rapid deepening of the water table represents the largest shift within the last 2,000 years, from a reconstructed WTD of ~3 cm to ~29 cm.

4.7. Periodicities

Data generated from REDFIT analysis for the palaeohydrological proxies are presented in Fig.10. Multiple periodicities are evident in each proxy, and REDFIT analysis confirms that these cannot be explained by the first-order autoregressive process, and are therefore not the product of 'red noise' (a continuous decrease of spectral amplitude with increasing frequency). Statistically significant periodicities of *c.* 240, 176, 150, 130, 80 and 60 years are present in at least one proxy (>90% level). The periodicity at *c.* 130 occurs in all three proxies. A periodicity of *c.* 1230 years is also present in the humification record, although this cannot be deemed a reliable signal due to the temporal length of the data

5. Discussion

5.1. Climate events and periodicities – the case for climate forcing

Distinct BSW phases are evident in the multiproxy reconstruction from Malham Tarn Moss. Peatlands are complex systems in which autogenic processes may result in disproportionate responses to external forcing mechanisms (Morris et al., 2011). However, where distinct phases of dry or wetter conditions are broadly synchronous with similar phases apparent in numerous other palaeohydrological studies, teleconnections may be implied provided that the dating methods used are suitably robust and of sufficient resolution (Barber et al., 2000). Historical accounts of notable events may also provide an invaluable chronologically precise means of ‘calibration’ (Jones and Mann, 2004).

Several studies have reported a ‘Dark Ages’ climate deterioration around AD 550 (e.g. Blackford and Chambers, 1991; Hughes et al., 2000; Langdon et al., 2003). A European cool period from AD 300-800 is also described in van der Linden et al. (2008), with several glacial advances recorded by Holzhauser et al. (2005) during this period. The MTM record indicates drier conditions across this period *c.* AD 340 MAR AD 219-549 to AD 840 MAR AD 709-889, with a shift to wetter surface conditions at *c.* AD 910 MAR AD 741-931, apparent in all three palaeohydrological proxies. The signal at MTM is coherent with a period of higher lake levels between *c.* AD 650-850 shown in a comprehensive review of Holocene palaeohydrological data in Swiss-French alpine lakes by Magny (2004), and a precisely dated wet shift at AD 770 in Ireland (Kerr et al., 2009). The level of human impact on the bog at this time is unclear as there appears to be no significant increase in burning regimes or other evidence of intensive land use in the geochemical record. The apparent delay in hydrological change at MTM may be the result of chronological error, unquantifiable human impact such as WTD recovery following peripheral peat cutting, or autogenic processes (cf. Barber et al., 1998). However, taking the MAR into account, a tentative comparison to the early Medieval climatic deterioration in Ireland (Kerr et al., 2009) is plausible.

The dry phase suggested in all three proxies between *c.* AD 920-1190, supported by an abundant count of *Cenococcum* spp. sclerotia (associated with aerated peat (Hughes and Barber, 2004)) at *c.* AD 950, is coherent with the 'Medieval Warm Period' (MWP), defined by Lamb (1977) as a period of climate amelioration in Europe between AD 1150-1300. Although the chronology for this phase at MTM does not appear to be entirely coeval with Lamb's definition, evidence from several later studies suggests that this climatic episode was variable globally (e.g. AD 650-1250 (Moberg et al., 2005), AD 1050-1400 (Wilson et al., 1979), AD 830-1150 (Moschen et al., 2011)) and even regionally in Britain e.g. AD 1100-1200 (Chambers et al., 1997), AD 800-900 (Hendon et al., 2001), AD 920-1160 (Langdon et al., 2003), AD 820-1070 (Blundell and Barber, 2005). In terms of human impact indicators during this period, the plant macrofossil profile records a more regular occurrence of macro-charcoal, most likely of local origin (cf. Tolonen, 1986a). Climatically, this period was favourable for agriculture in North-West Europe (Berghlund, 2003) and therefore increased agrarian activity may be expected around MTM at this time. Given the regularity of charcoal in the record, it is unlikely that these fire events can be solely related to natural wildfire events (Innes and Simmons, 2000) and therefore some level of human impact is assumed. It is possible that the drier climate rendered the surface of MTM more conducive to rough grazing, and burning was carried out to encourage grass and low shrub vegetation during a period where land demand was especially high. A rise in *Calluna vulgaris* and Poaceae pollen is evident during this phase; however, assigning this to a causal factor would be conjecture. The very slight increases in both Ti and Y, visible in the latter half of the apparent drought phase, could reflect soil destabilisation through increased agricultural activity in the region (Lomas-Clarke and Barber, 2004). This apparent dry phase at MTM is in agreement with data reported by Langdon et al. (2003) from south-east Scotland.

Hey (1986, p.86) describes a period of "harvest failure and livestock disasters almost without parallel in the history of English agriculture" between AD 1315-1322, a well-documented period of famine in England (Kershaw, 1973). A poor summer recorded in AD 1314 was followed by an exceptionally wet summer the year after: harvests failed, and a weakened population

succumbed to disease. Taking chronological error into consideration, this historical description corresponds with the water table depth reconstruction from MTM, where a rapid shift from dry to wet is apparent *c.* AD 1280-1350. The population suffered further soon after, as the Black Death reached Yorkshire in AD 1349, with death rates in nearby Nidderdale reported at > 40% (Hey, 1986). It is therefore logical to suggest that human impact on MTM around this period would have been low, and a clear reduction of charcoal in the profile from *c.* AD 1240 MAR AD 1112-1362 onwards may be indicative of this. Magny (2004) also found a period of increased lake levels between AD 1200-1300, with teleconnections to the GISP2-Polar Circulation Index record, the North Atlantic ice-rafting debris (IRD) events and the $\Delta^{14}\text{C}$ record, although the approach used by Magny (2004) has been recently debated (Bleicher, 2013a; Bleicher, 2013b; Magny, 2013).

The prolonged phase of increased BSW apparent in the palaeohydrological proxies at MTM between *c.* AD 1460-1850 is synchronous with the so-called 'Little Ice Age' (LIA) climate deterioration, amply supported by a large number of peatland studies across northern Europe (e.g. Mauquoy and Barber, 1999b; Mauquoy et al., 2002; Blundell and Barber, 2005; van der Linden et al., 2008; De Vleeschouwer et al., 2009). The LIA presents an ideal period to examine the impact of solar forcing as empirical records are available for many areas, and the impact of freshwater inputs to the North Atlantic are minimal (Plunkett and Swindles, 2008). Peaks in BSW apparent in the testate amoebae WTD reconstruction occur at *c.* AD 1280 MAR AD 1163-1378, *c.* AD 1570 MAR AD 1507-1642 and *c.* AD 1740 MAR AD 1616-1816. These points correspond well with periods of lower sunspot activity: the Wolf Minimum (AD 1281-1342), Spörer Minimum (AD 1416-1534) and Maunder Minimum (AD 1645-1715) (Bell and Walker, 2005), adding support to the theory that solar forcing was a major driver of climate change over the LIA period (Blackford and Chambers, 1995; Plunkett and Swindles, 2008; van der Linden et al., 2008; Gray et al., 2010).

A local weather station was installed on the Malham Tarn Estate in the mid-19th Century, and a decadal running mean for precipitation between AD 1860-2000 (Burt and Horton, 2003) provides a means to examine the palaeohydrological reconstruction over this period. Peak decadal average maximum of 1645 mm is centred at AD 1927 and minimum at AD 1973 around the well documented British drought phase of AD 1975-76 (Ratcliffe, 1978). The testate amoebae-based BSW reconstruction shows a rapid but brief return to wetter conditions at *c.* AD 1920 MAR AD 1892-1942, and deep water tables at *c.* AD 1970 MAR AD 1949-1974. Taking the total chronological error into account, the timing of these apparent phases in the WTD reconstructed values are not dissimilar, suggesting that testate amoebae respond rapidly to changes in surface hydrology and may indicate decadal-level effective-precipitation changes where the sampling resolution is high enough.

In the geochemical record, a distinct rise in bromine is evident over the LIA period *c.* AD 1500-1920. Br is a common element in sea water and sea salt inputs were found to account for effectively all the Br recorded at two oceanic peatland sites in Scotland by Shotyk (1997). Mass balance calculations to quantify Br retention in peat is somewhat contradictory; Shotyk (1997) concludes that Br is not retained and therefore cannot be used as a quantitative measure of sea salt inputs, whereas Roos-Barraclough et al. (2002) assert that Br is retained and can be at least semi-quantitative. Elevated Br levels found in minerotrophic peats in Greenland have shown comparability to increased Na⁺ ions in the GISP2 core, linked to increased North Atlantic storminess from AD 1400 (Schofield et al., 2010). Strong evidence of increased winter storminess is reflected in historical records and aeolian sand drift activity on western coasts of Europe (Dawson et al., 2004; de Jong et al., 2007; Clarke and Rendell, 2009). Although there are other possible terrestrial sources of Br, such as the release of methyl bromide during plant litter decay (Lee-Taylor and Holland, 2000) or abiotic oxidation of organic matter (Keppler et al., 2000), the timing and nature of the rise of Br at MTM suggests that 1) it is, at least to a large degree, retained in the peat profile, and 2) the predominant source is marine in origin and thus Br may be an indicator of increased North Atlantic storminess. The findings here contradict the

theory of Biester et al. (2005) that increased halogens (including Br) are correlated with dry periods and high peat decomposition, as the Br signal is clearly associated with the increased BSW over the LIA.

5.1.1. Periodicities

A number of periodicities in testate amoebae based WTD reconstruction and humification data were found in two ombrotrophic peatland records in Northern Ireland: *c.* 130, 180, 260, 540 and 1160 years (Swindles et al., 2012c). These periodicities were present in at least one proxy at each site, and both records have tightly constrained Bayesian piece-wise linear accumulation age-depth models, with several tephra layers adding precision to them. A periodicity of *c.* 130 years is present in all three palaeohydrology proxies at MTM, which has been associated with sub-harmonics of the 22-year Hale cycle (Attolini et al., 1990) and also found in several other records (e.g. Patterson et al., 2007; Nichols and Huang, 2012). A *c.* 176-year cycle is present in the DHI record and is similar to the *c.* 180-year periodicity of Swindles et al. and also found in numerous other records (see Swindles et al., 2012b for a full review) and is coherent with the Suess solar cycle. A *c.* 240-year periodicity apparent in the humification record could be correlative with the *c.* 260-year described by Swindles et al. and found in many other records from Europe (Aaby, 1976; Borgmark, 2005; Swindles et al., 2007). Chambers and Blackford (2001) detected periodicities of *c.* 80 and 60 years, and both these frequencies are present in the humification record at MTM. The *c.* 80-year periodicity is coherent with the Gleissberg cycle, and the *c.* 60-year frequency with the 52 year fourth harmonic of the Suess cycle (Damon et al., 1998). The *c.* 1230-year periodicity present in the DHI record cannot be interpreted with any confidence owing to the limited duration of the proxy records (~2400 years); however, this does fall within the range of 'Bond event' cycles of 1373 ± 502 years over the Holocene (Bond et al., 1997; 2001), and similar periodicities (1100-1160 years) have been reported in other northern European peatlands (Hughes et al., 2000; Langdon et al., 2003; Swindles et al., 2012c). As previously discussed, each palaeohydrological proxy has its own limitations, and the

relationship between climate and humification is particularly complex. However, given the presence of the periodicities encountered within the MTM proxies, particularly the *c.* 130-year periodicity found in all three, and their coherence with numerous other records, the evidence for common climate forcing across northern hemisphere peat hydrology is convincing, despite possible underlying autogenic processes (cf. Swindles et al., 2012b).

5.2. Human impact at MTM – indicators of anthropogenically driven change

Human activity can influence vegetation change and WTD via several mechanisms: grazing pressure or stock type (cattle or sheep), burning regime, drainage or cutting, nutrient input via grazing animal faecal enrichment or atmospheric deposition (e.g. atmospheric N, aeolian mineral soil dust), and other atmospheric input (Chambers et al., 2007). These factors can alter moisture availability and community composition, the primary drivers of peat production and decay rates (Clymo, 1984).

The disappearance of *Sphagnum austinii* (Sull. Ex Aust.) registered in numerous British peatland records is also apparent at MTM. There has been wide debate on the causal factors of this, including grazing, drainage and burning (Pearsall, 1956; Piggott and Piggott, 1963), autogenic processes (Barber, 1981), changes in bog hydrology (Mauquoy and Barber, 1999a) or nutrient enrichment (van Geel and Middelorp, 1988). Hughes et al. (2007; 2008) temporally linked agricultural land-use intensity with the decline of *S. austinii*, inferred through the decline of arboreal pollen, increase in anthropochore and apophyte pollen, and increased Ti loading. They therefore suggest that enrichment through soil erosion may be a causal factor, and that *S. austinii* is sensitive to even comparatively minor land-use impacts. This theory is also in part supported by McClymont et al. (2008), who suggest a combination of nutrient loading together with rapid increases in BSW drove the decline of this species. At MTM *S. austinii* has virtually disappeared by *c.* AD 30 with only a slight incursion at *c.* AD 970, which coincides with a reducing C/N ratio over this period. Unlike other plant macrofossil records (Mauquoy and Barber, 1999a; Langdon and Barber, 2005; Hughes et al., 2007), *S. austinii* is not replaced by

other *Sphagna* but by increased UOM and monocotyledonous material, implying that the *S. austinii* decline was not driven by inter-species competition (e.g. Barber, 1981). There is evidence of higher Ti, Y and Rb in the geochemical data suggesting soil dust loading at the bog surface. The palaeohydrological reconstructions all indicate a gradual lowering of water table from c. 410 BC – AD 230; therefore the evidence from MTM appears to support the work of Hughes et al. (2007; 2008) that agrarian activity was a primary driver for the loss of *S. austinii* in British peatlands.

Although lead mining and smelting in the Yorkshire Dales was in existence in Roman times, the industry became much more widespread and organised after monastic involvement in the 12th Century. Early smelting techniques consisted of open bonfire-like furnaces known as ‘bales’. The XRF data from MTM show a very low background level of Pb which increases c. AD 1280-1380 before returning to low levels. This may be related to localised atmospheric Pb deposition from these medieval bales, as later smelt mills incorporated flues and chimney systems that probably increased the distribution range of smelting fumes. The lead mining and smelting industry in the Yorkshire Dales rapidly declined after the 1880s as more economically viable sources became available (White, 1997). Atmospheric deposition of lead increased markedly following the start of the industrial revolution (c. 1850s onwards), peaking in the mid to late-20th Century, related in part to the rapid expansion in vehicular use and leaded petrol emissions. This is followed by a subsequent decrease in the late 20th century owing to the decline of heavy industry, improved emissions control and the ban on leaded petrol (Lee and Tallis, 1973; Clymo et al., 1990; Shotyk, 1996; MacKenzie et al., 1998). This trend is clearly replicated in the MTM profile where Pb deposition begins to increase from the 1870s (as does Cu, Zn and Fe) and then quickly increases from the 1960s. A positive correlation between increased Pb pollution and a reduction in testate diversity and density living within *Sphagnum fallax* has been demonstrated by Nguyen-Viet et al. (2008), with *Euglypha strigosa*, *Heleopera sphagni* and *Nebela carinata* described as the least tolerant taxa to Pb pollution. It is difficult to detect similar responses in the testate amoebae assemblage at MTM over the Pb influx increase from 10-2 cm, and thus

there is the possibility that Pb has affected the community and subsequent WTD reconstruction. Humification as a palaeohydrology proxy at this depth is unreliable as the plant material is not yet fully decomposed; therefore the reconstruction over this core section is potentially less secure.

The premise that *Hyalosphenia subflava* is an indicator of major disturbance, and in particular fire events (cf. Turner and Swindles, 2012), is not apparent in the palaeoecological data from MTM. A Spearman's rank correlation test revealed no significant correlation between charcoal concentration and the relative abundance of this taxon within the profile. However, a significant correlation ($r_s = 0.285$, $p < 0.05$) between charcoal presence and *Trigonopyxis arcula* type was found, supporting the suggestion of Warner (1990) that *T. arcula* is sometimes an indicator of fire events. There is clearly a complex association between both these taxa and burning that requires further exploration, as other factors may also be influencing community dynamics.

5.3. The intensification of human impact

On the balance of the evidence presented so far, the ecohydrology of MTM from *c.* 410 BC to the late medieval period has primarily been driven by allogenic climate forcing. Human impact may have amplified the effects of climate amelioration in terms of vegetation change when the surface of the mire was rendered more exploitable. Prior to recent centuries, enrichment of MTM was more likely to be from the presence of grazing animals rather than aeolian inputs of mineral soils, and thus vegetation impact would have been localised. However, over the last two centuries the level of human impact has increased to such a level that drivers of change become increasingly difficult to separate.

An extremely rapid drying phase from *c.* AD 1850 MAR AD 1819-1884 is evident in the WTD reconstruction, indicated by the increased presence of testate amoebae *Assulina muscorum* and *Cyclopyxis arcelloides* type. In addition, *Sphagnum* disappears from the plant macrofossil record, replaced by *Racomitrium lanuginosum*, which suggests a prolonged period of drier climate (Tallis, 1995; Blackford, 2000). The raising of the tarn level in AD 1791 would undoubtedly

have had an effect on WTD as wave action reduced the lateral extent and therefore the hydraulic head (cf. Morris et al., 2011) of the adjacent mire by 15-20 m (Pentecost, 2000; Proctor, 2003). The spatial and temporal extent of this drainage is unknown. However, the current strong dominance of *Calluna vulgaris* toward the tarn, over-steepening of the peat rand and the presence of pipes and gullies indicate that water is lost from this eastern edge. Therefore the raising of the tarn level, with subsequent erosion of the eastern side of the bog and loss of hydraulic head, could be a contributory factor to the apparent dry shift as the effects would likely involve substantial time-lag. Aerial photographs of MTM clearly show several man-made linear drains (installation date unknown); however, the two main soakways running north and west from Spiggot Hill (Fig.1.) are thought to be natural features (Meade and Jones, 2010). MTM was acquired by the UK National Trust (www.nationaltrust.org.uk) in 1947 at which time very little *Sphagnum* was present on the bog surface except around deeper pools, and records would suggest the site had been heavily grazed, cut and burnt over the previous century (Cooper and Proctor, 1998), although this appears unsupported in the palaeoecological record. Drain blocking has been underway since the 1970s, including damming of the natural soakways to instigate higher WTD. The testate amoebae-based WTD reconstruction clearly shows an increase in BSW at c. AD 1960 MAR AD 1949-1974 potentially coeval with the onset of grip blocking, which would suggest that this management has been partly successful in raising water tables (a further deepening of WTD is apparent from c. AD 1990). However, Spiggot Hill (Fig.1), a wooded glacio-fluvial mound of approximately 1.5 hectares located to the southern section of MTM, is elevated above the peat surface and precipitation run-off is minerogenic in nature, with additional nutrients deposited via the faeces of roosting bird species. Blocking of the natural soakways may therefore be a source of eutrophication to the ombrotrophic mire. The XRF geochemical profile shows greatly increased inputs of minerogenic enrichment from the 1900s indicated by increases in Ti, Rb and Y, and increases in agricultural fertiliser (K). Thus, the damming of the minerotrophic soakways could exacerbate the potential eutrophication of the bog.

Atmospheric nitrogen deposition increased considerably from the industrial revolution, peaking in the 1970/80s although has declined over the last three decades in the UK (Fowler et al., 2005). Primary sources include NO_x from industrial processes and vehicular emissions, and NH₃ from coal burning and agriculture (Fowler et al., 2004). Excess N has been linked to negative ecosystem changes through substrate eutrophication and acidification (Woodin and Farmer, 1993; Hogg et al., 1995; Fowler et al., 2004; Edmondson et al., 2010). Many *Sphagnum* species are highly adapted to oligotrophic conditions and are nutrient-efficient (Rydin and Jeglum, 2006). Increased N availability has been found to have a deleterious effect on *Sphagnum* biomass favouring dominance of vascular plants, which could lead to changes in the microtopography of peatlands, particularly hummock formation and growth (Gunnarsson and Rydin, 2000). Hummocks are often formed from the *Sphagnum* Section *Acutifolia* with structural support from ericaceous plants. As they form thicker acrotelmic peat than lawn and hollow microforms, they have a greater aerobic zone and are thus greater oxidisers of methane (Nedwell and Watson, 1995; Sundh et al., 1995). The potential for reduced hummock formation or even collapse (Gunnarsson and Rydin, 2000) therefore has serious implications on greenhouse gas (GHG) cycling in peatlands.

The apparent rapid shift to drier bog surface conditions over the last few decades is interesting in terms of carbon storage and sequestration. The testate amoebae-based reconstruction indicates the deepest water table depths over the whole peat section occur since the 1960s. Lowered WTD increases peat decomposition (Armstrong et al., 2009) and carbon loss in the form of CO₂ efflux and dissolved and particulate organic carbon run-off (Laiho, 2006; Worrall et al., 2006; Ellis et al., 2009; Holden et al., 2011). The highest carbon accumulation values at MTM between c. AD 920-1060 occurred during the MWP climate amelioration. While other studies have noted a reduced C accumulation rate over periods of increased BSW, particularly during the LIA (Oldfield et al., 1997; Mauquoy et al., 2002) the analysis from MTM suggests a relatively constant accumulation rate over the last ~2400 years *except* the MWP. Lower accumulation over the wetter periods could be related to reduced net primary productivity (NPP) during

colder conditions (Mauquoy et al., 2002; Charman et al., 2013) and/or differential decay of more aquatic species such as *Sphagnum cuspidatum* (Johnson and Damman, 1991). The high C accumulation over the MWP shown here is most likely a result of increased NPP and species composition/litter type (Borren et al., 2004; Laiho, 2006; Beilman et al., 2009) as higher LORCA rates have been associated with woody and ericaceous peats (Loisel and Garneau, 2010). This raises some interesting questions about long-term carbon sequestration given the vast amount of resources currently being expended on peatland restoration schemes across Europe and North America, predominantly to raise water tables and restore *Sphagnum* dominated flora communities (Komulainen et al., 1999; Waddington et al., 2010; Holden et al., 2011).

In the review by Laiho (2006) consensus suggests high WTD in peatlands results in C sequestration; however, the relationships between water level, decomposition and C sequestration are complex (Waddington and Roulet, 2000; Bubier et al., 2003). The key factor in C sequestration is rather obvious, however: accumulation must exceed decomposition. Therefore regardless of the evident C accumulation over the MWP where WTD appears lower than at other phases over the last ~2400 years, the apparent deepening of WTD over recent decades, coupled with increased eutrophication should be viewed as deleterious to the ecohydrology and GHG exchange capacity of Malham Tarn Moss.

6. Limitations & Future work

There is an inherent risk in comparing data and making assumptions on coevality between sites where a level of chronological and reconstruction uncertainty exists (cf. Swindles et al., 2012a). Reliable time-stratigraphic markers such as tephra layers incorporated into Bayesian age-depth models go some way to quantify chronological error, although studies with this level of confidence remain in the minority. Also, the moderating effects of local environmental controls on climatically driven change may play a central role in the variation in palaeoclimate histories across Europe (Chiverrell, 2001). However, whilst avoiding 'force-fitting' of changes in the palaeoenvironmental record at MTM, the synchronicity of these phases with numerous other

peat and non-peat based records is compelling. Chronological confidence could be improved for the MTM record with the addition of reliable time-stratigraphic markers, such as the Hekla AD 1947 (AD 1947) or Glen Garry (260-12 BC) tephra layers (Lawson et al., 2012; Rea et al., 2012); however, the Bayesian piece-wise linear accumulation model combined with historical records that have been employed in this study provide a large degree of certainty. Further refinement of the potential effects of eutrophication at the bog surface due to animal grazing could be aided by the addition of coprophilous fungal spore analysis (van Geel, 1978; Davis, 1987; van Geel and Aptroot, 2006).

7. Conclusions

1. This study presents well-dated palaeoenvironmental records of bog surface hydrology and vegetation change at Malham Tarn Moss (MTM), northern England. Allogenic climate and human influence signals have been identified from *c.* 413 BC to AD 2009.
2. Several phases of both increased and decreased bog surface wetness have been detected in the peat profile. Dry phases are apparent at *c.* AD 320-830, AD 920-1190 (Medieval Warm Period; MWP), AD 1850 and AD 1920-present. A period of increased effective precipitation is detected at *c.* AD 1460-1850 (Little Ice Age; LIA).
3. The synchronicity of the MWP and LIA climatic phases inferred from the MTM multiproxy data with other records from across northern Europe implies that these episodes are primarily driven by allogenic climatic changes.
4. Distinct wet phases at *c.* AD 1280 MAR AD 1163-1378, *c.* AD 1570 MAR AD 1507-1642 and *c.* AD 1740 MAR AD 1616-1816 correspond to the Wolf, Spörer and Maunder sunspot activity minima, supporting existing evidence that solar forcing was a significant driver of climate change over the last ~1000 years.
5. Ecohydrological changes at MTM were primarily driven by allogenic climate forcing up to approximately the early 19th century; however, following the intensification of

industry and agriculture over the last two centuries, the combined forcing effects become increasingly difficult to separate.

6. The recent drying phase apparent in the testate amoebae-based palaeohydrological reconstruction since *c.* AD 1850 has the potential to exert serious negative feedbacks on carbon storage, carbon sequestration and GHG cycling. The causal factors of this apparent drying phase warrant further research.

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Figures and tables

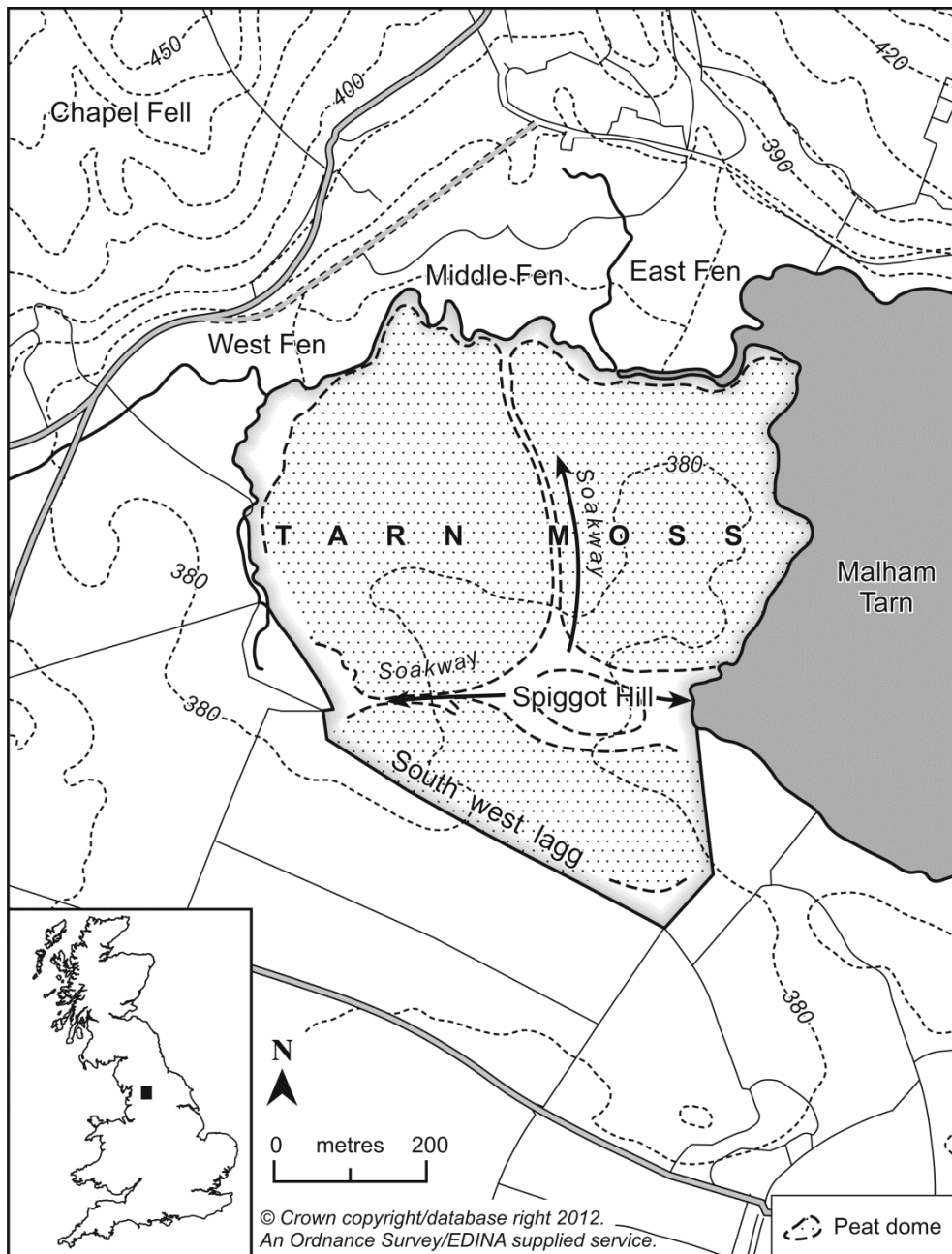


Fig. 1. Location map of Malham Tarn Moss.

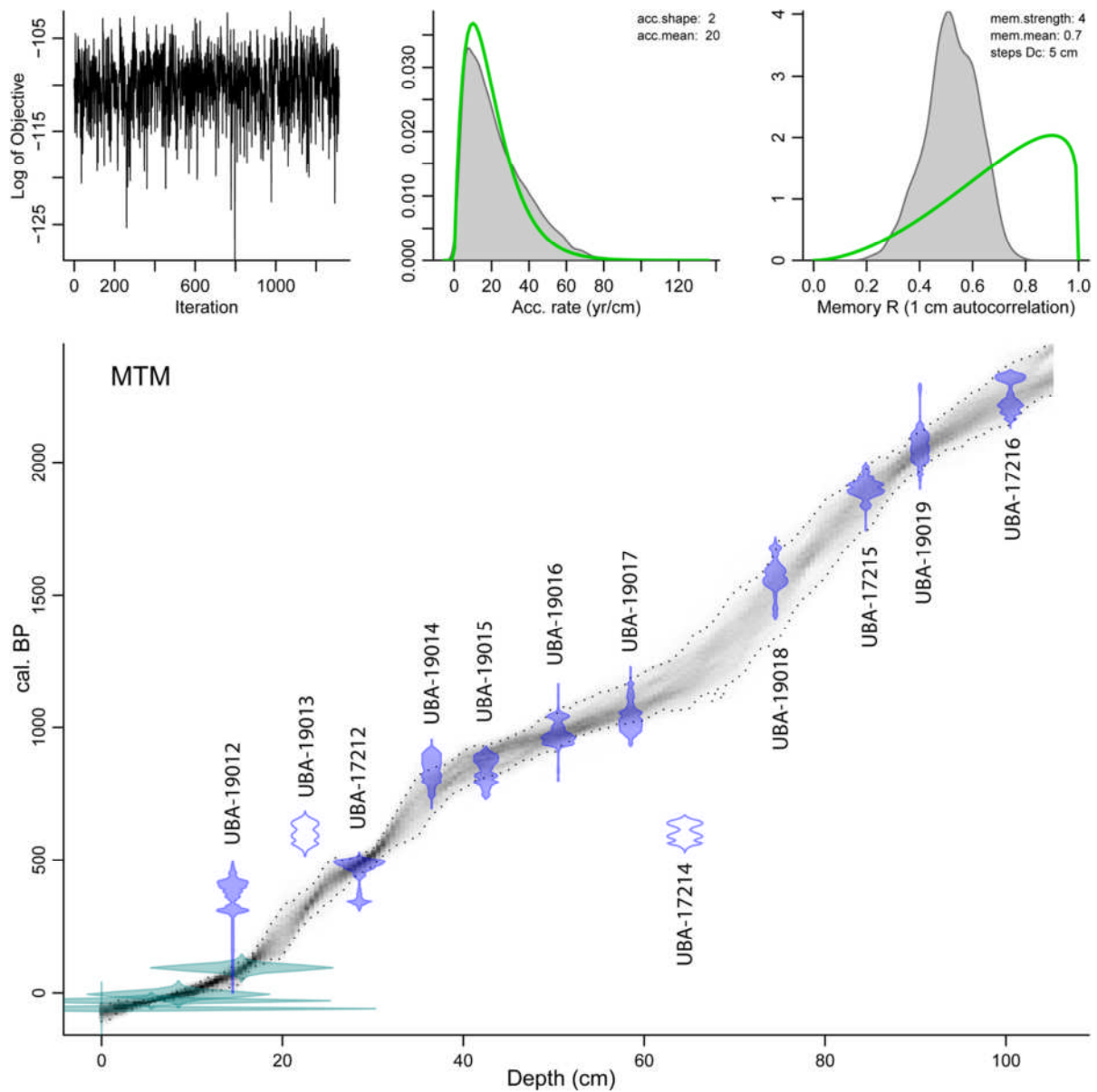


Fig. 2. Bayesian age-depth model for MTM. Major plot shows the age distributions of calibrated ¹⁴C dates (blue) and SCP age-equivalent time markers (pale green). Calibrated dates that lie outside the model are shown unfilled. The upper left plot shows the stability of the Markov Chain Monte Carlo iterations (>1000 iterations), the middle plot shows the prior (green curve) and posterior (grey fill) distribution for accumulation rate (yr cm⁻¹), and the upper right plot the prior (green) and posterior (grey fill) dependence of accumulation rates between piece-wise sections.

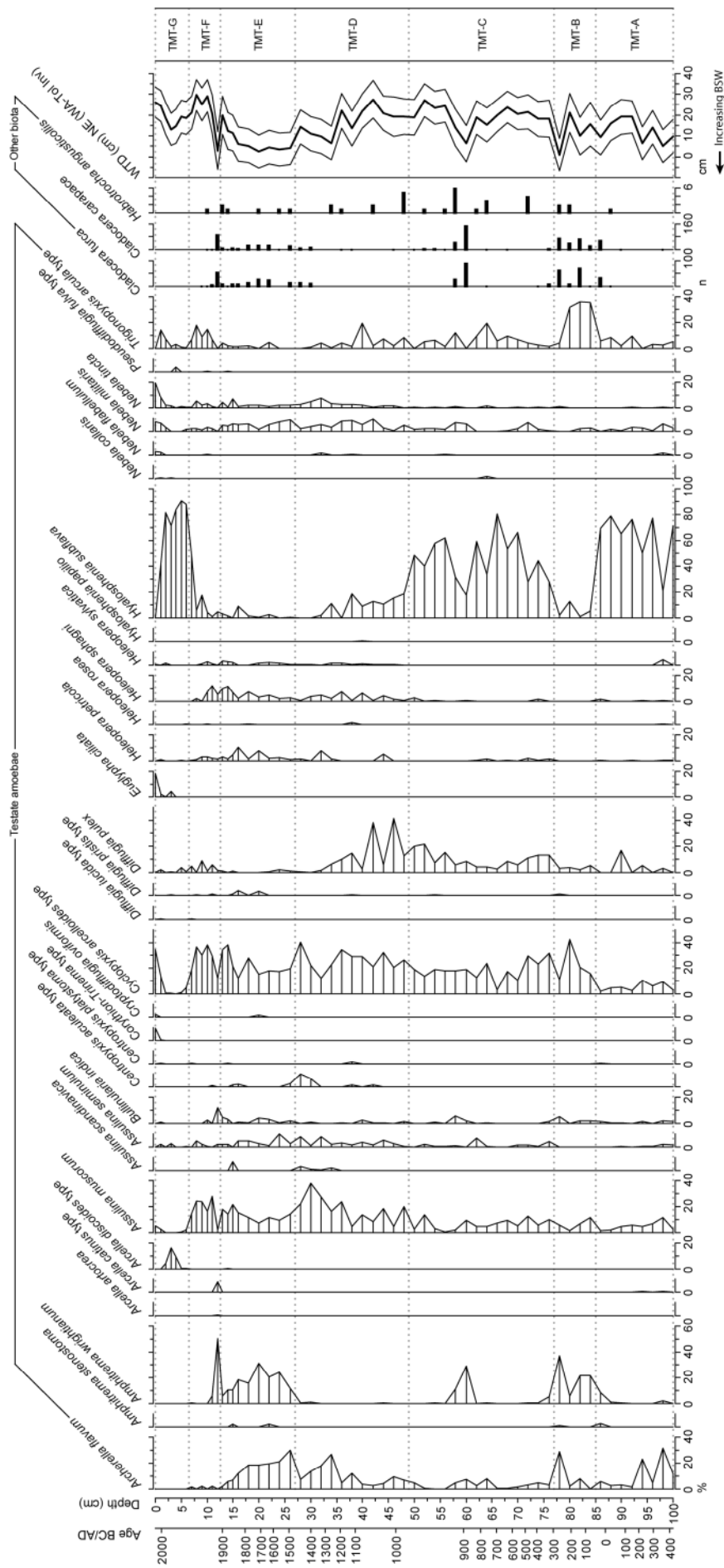


Fig. 3. Percentage testate amoebae diagram from MTM. A water table reconstruction based on the Northern England transfer function (Turner et al., 2013) is shown with maximum sample-specific error ranges based on 1,000 bootstrap cycles. The diagram has been zoned using constrained Euclidian distance cluster analysis, with fine-tuning by eye based on assemblage changes.

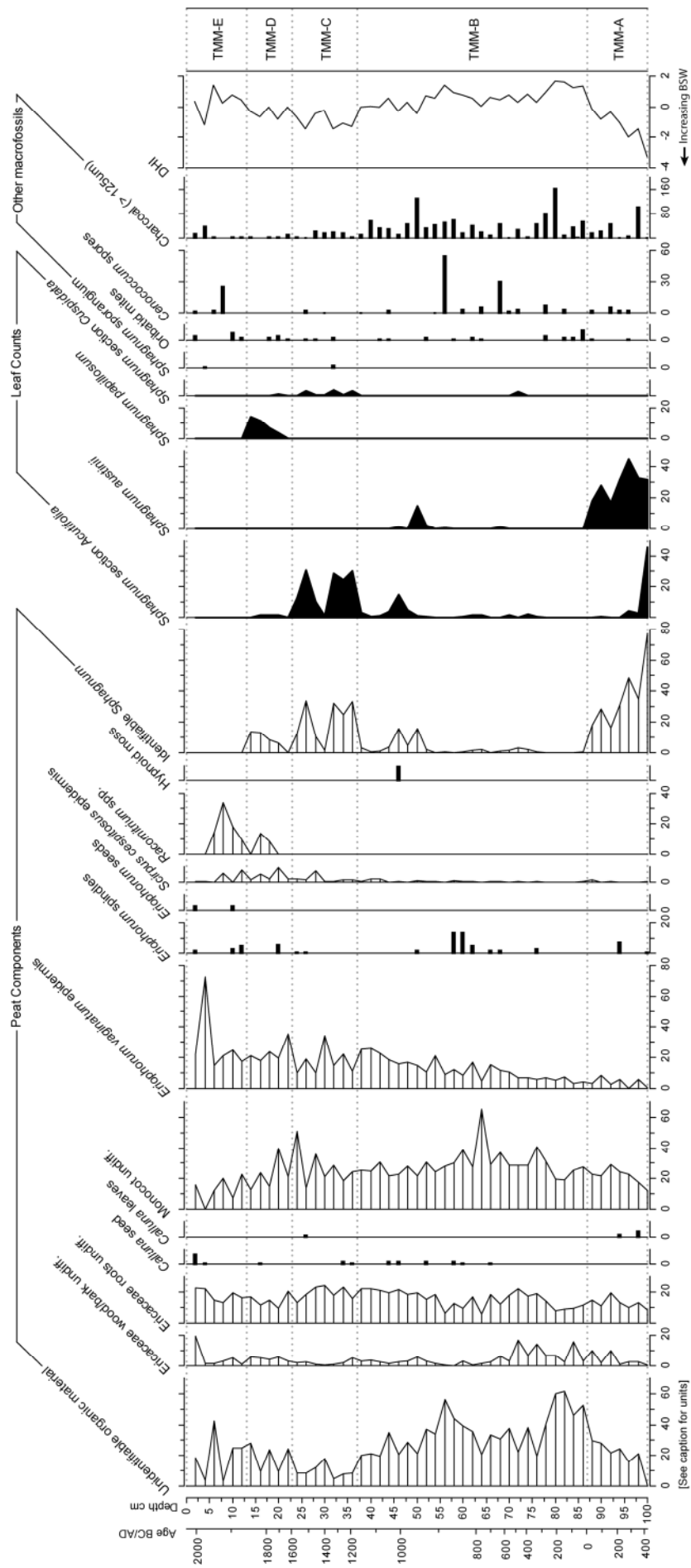


Fig. 4. Plant macrofossil diagram from MTM. Peat components are derived from averaged quadrat counts (15), and leaf counts the percentage representation of >100 randomly selected identifiable *Sphagnum* leaves. Bar graphs are used to express absolute counts. The normalised DHI score is shown; negative values represent increased BSW. The diagram has been zoned using constrained Euclidian distance cluster analysis, with fine-tuning by eye based on assemblage changes.

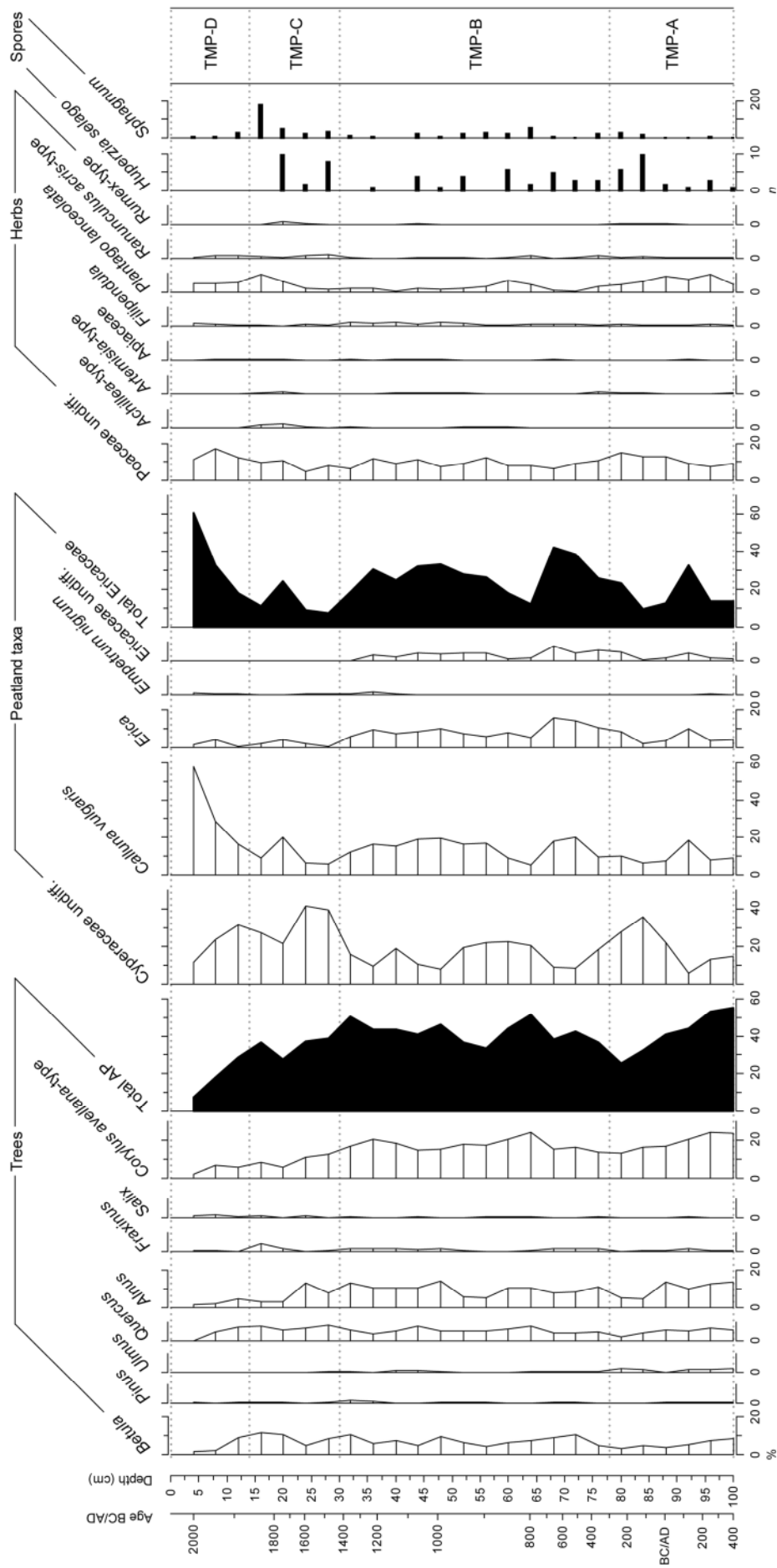


Fig. 5. Percentage pollen diagram from MTM (selected taxa only - occasional occurrences have not been included). Solid-fill graphs are shown for total arboreal pollen and total Ericaceae. Bar graphs are absolute counts. The diagram has been zoned using constrained Euclidian distance cluster analysis, with fine-tuning by eye based on assemblage changes.

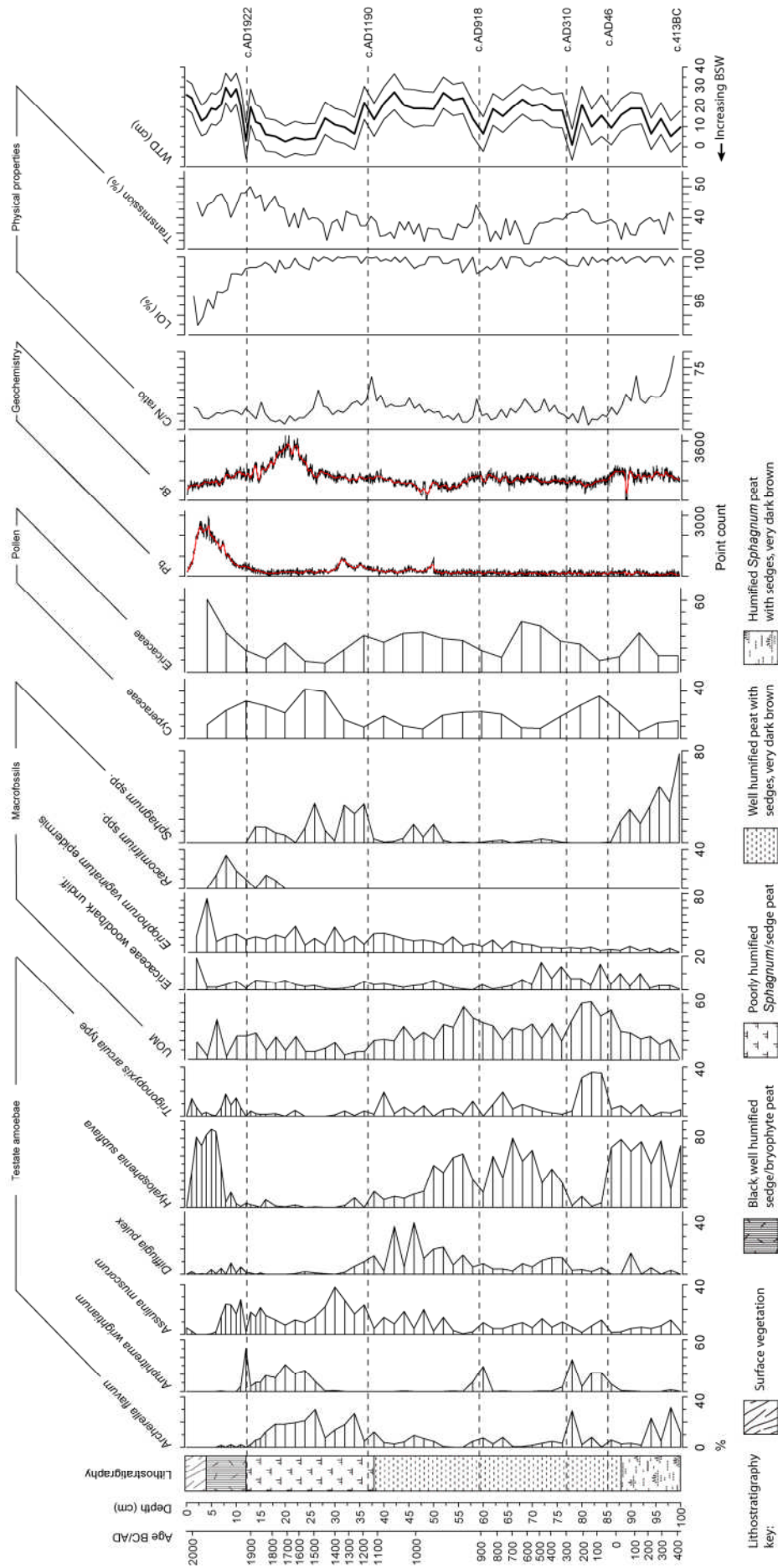


Fig. 6. Key indicator proxies and stratigraphy for MTM. Units for each variable are shown on the graph.

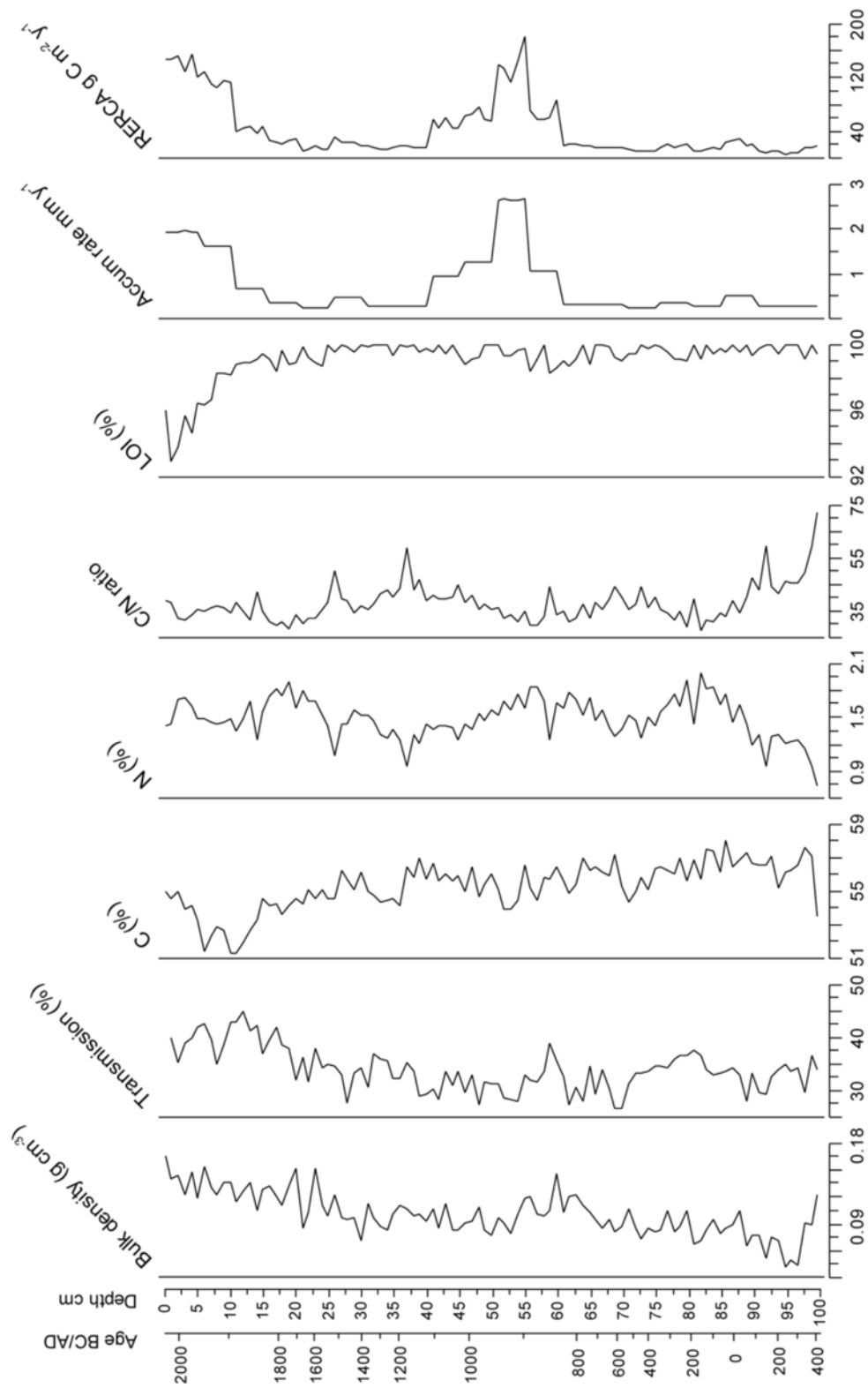


Fig. 7. Results of bulk density, degree of humification (% transmission), C/N ratio, loss-on-ignition, peat accumulation rate and apparent rate of C accumulation (RERCA). Units for each variable are shown on the top of each graph.

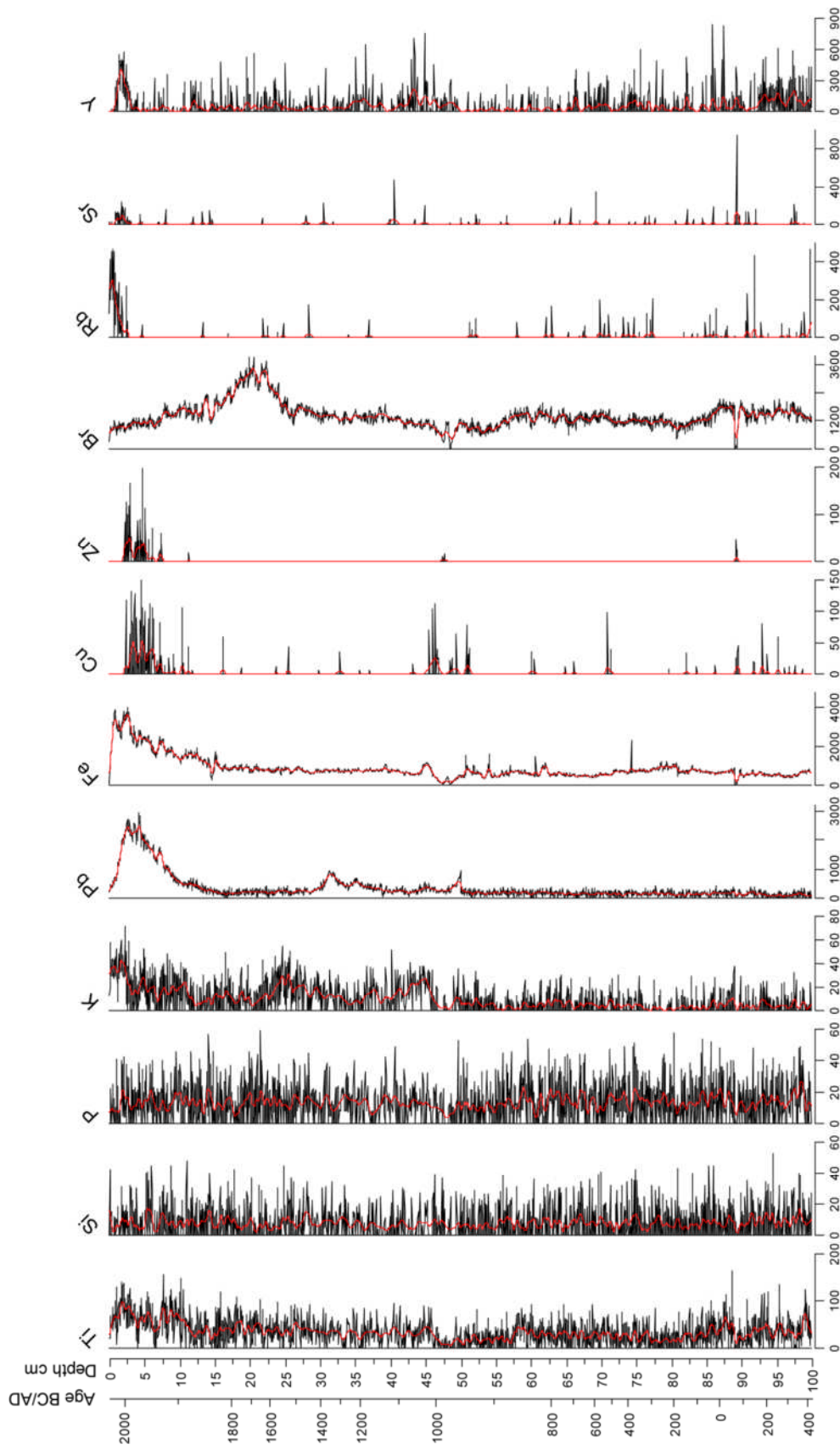


Fig. 8. XRF trace element analysis results. Values represent a time-based point count (30 seconds per 500 μm increment). A LOWESS smooth (0.01 span) is shown (red line) to aid interpretation.

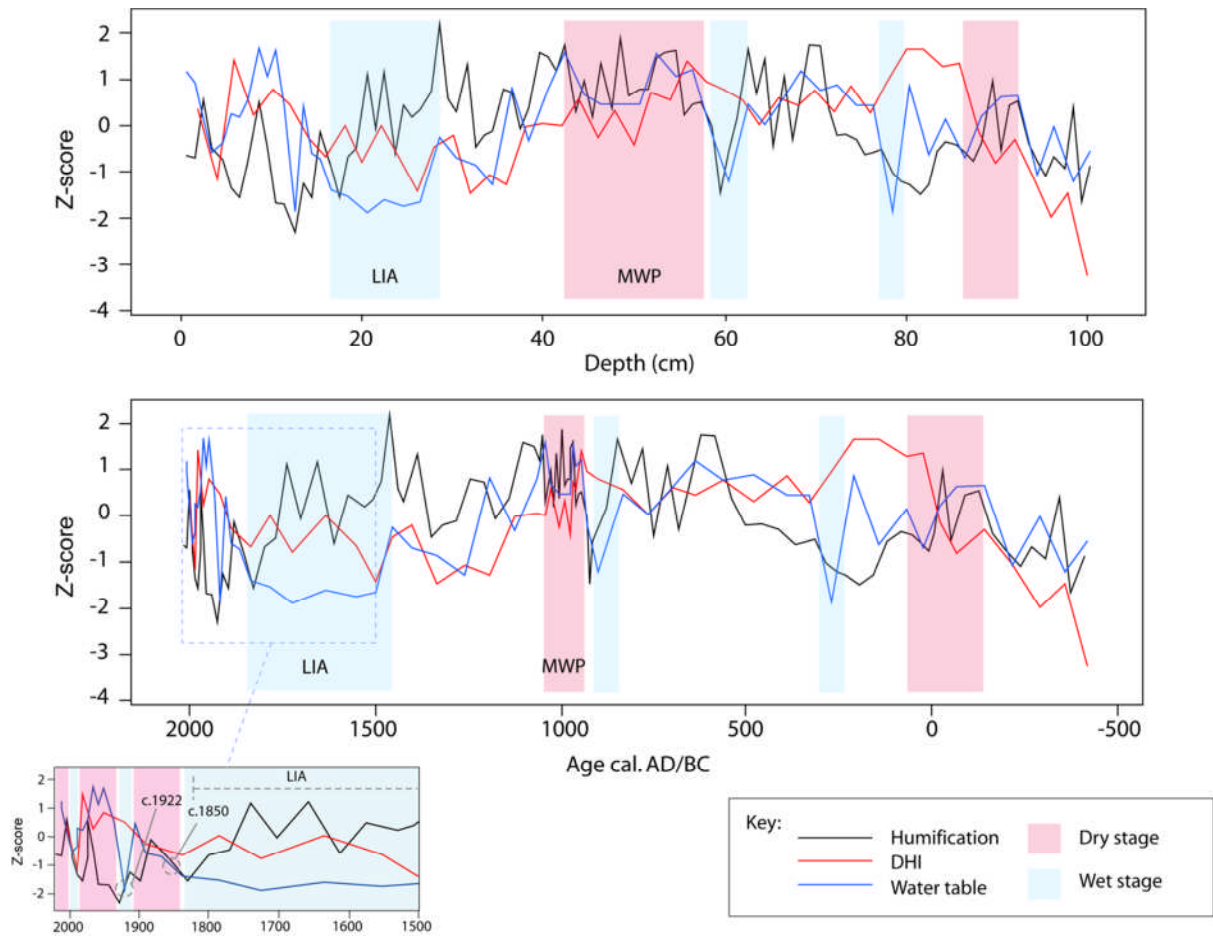


Fig. 9. Comparison of the three hydrological proxies (testate amoebae WTD, humification, plant macrofossil DHI) shown against depth and age. All values have been normalised to z-scores. Negative values indicate increased BSW. Inset shows an expanded view of the last 500 years.

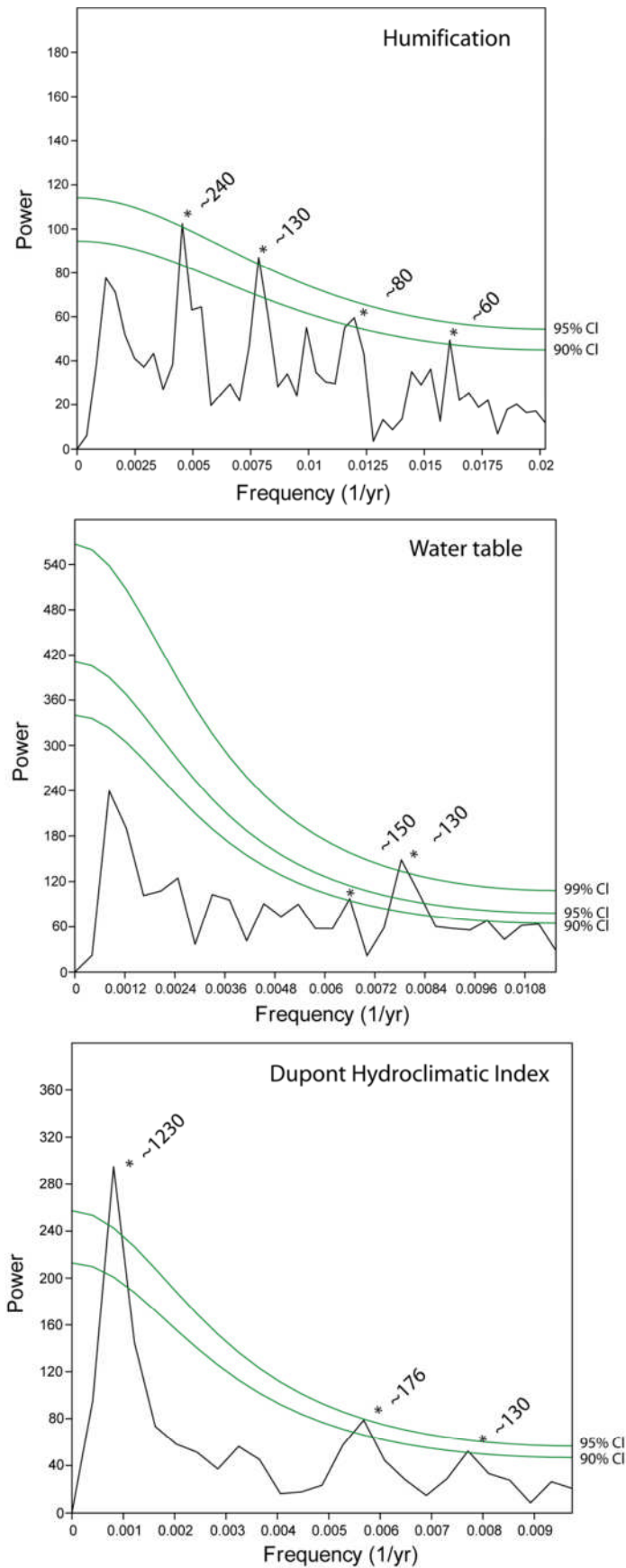


Fig. 10. Spectral (REDFIT) analysis of humification, testate amoebae-derived water table reconstruction and plant macrofossil DHI z-scores. (99% significance omitted where not intersected by peaks).

Table.1 Lithological description of MTM core.

Depth (cm)	Troels-Smith	Description
0 – 4	Acrotelm	Hypnoid mosses
4 – 12	Tb2Th2Sh+	Black, well-humified peat with sedges/bryophytes, boundary less gradational
12 – 38	Th2Tb2 (Sphag)	Poorly-humified <i>Sphagnum</i> /sedge peat, boundary gradational
38 – 88	Th2Sh2	Well-humified sedge peat, very dark brown, <i>Eriophorum</i> remains, <i>Calluna</i> is absent
88 – 100	Tb3(Sphag)Th1	Well-humified <i>Sphagnum</i> and sedge peat, very dark brown, <i>Eriophorum</i> remains, <i>Calluna</i> is absent

Table.2 Results of AMS ¹⁴C dating of the MTM core.

Code	Depth (cm)	¹⁴ C age (BP)	±	δ ¹³ C (‰)	Calibrated age (2σ)	Material
UBA-19012*	14	295	24	-33.0	AD 1523 – 1654	<i>Sphagnum</i> stems/leaves, <i>Racomitrium</i> stems/leaves, Ericaceous woody material.
UBA-19013*	22	617	36	-29.1	AD 1298 – 1404	Humin fraction (300µm - 125µm), Ericaceous woody material.
UBA-17212	28	401	22	-25.6	AD 1446 - 1616	<i>Sphagnum</i> leaves/stems.
UBA-19014	36	922	32	-29.5	AD 1044 - 1185	<i>Sphagnum</i> leaves.
UBA-19015	42	909	23	-31.5	AD 1046 - 1205	Humin fraction, Ericaceous woody material.
UBA-19016	50	1061	31	-31.1	AD 905 - 1023	<i>Sphagnum</i> leaves, charcoal fragments, Ericaceous woody material.
UBA-19017	58	1127	32	-26.0	AD 889 – 991	Humin fraction, <i>Eriophorum</i> spindles.
UBA-17214*	64	621	22	-24.5	AD 1299 – 1397	Humin fraction, Ericaceous woody material.
UBA-19018	74	1668	33	-28.3	AD 343 – 526	Ericaceous woody material.
UBA-17215	84	1952	23	-23.8	AD 25 – 122	Humin fraction, Ericaceous woody material.
UBA-19019	90	2087	33	-29.4	162 – 3 BC	Humin fraction, Ericaceous leaves and woody material.
UBA-17216	100	2259	25	-30.9	389 – 210 BC	<i>Sphagnum</i> leaves/stems.

*denotes erroneous date excluded from age-depth model

Table.3 Description of biostratigraphic zones by individual proxy.

Zone	Age	Description
<i>Testate amoebae</i>		
TMT-G (0 – 6 cm)	c. AD 2009 - 1970	Zone dominated by <i>H. subflava</i> > 90% before a sharp decline from 2 cm to surface, replaced by a diverse assemblage including <i>T. arcula</i> type, <i>Nebela tinctoria</i> , <i>N. militaris</i> , and a sharp increase of <i>C. arcelloides</i> type. A brief peak of <i>Arcella discoidea</i> type appears between 4 and 2 cm, and <i>Corythion</i> – <i>Trinema</i> type and <i>Euglypha ciliata</i> feature at low abundances, the only zone where these taxa are present.
TMT-F (6 – 12 cm)	c. AD 1970 - 1920	Highest presence of <i>Amphitrema wrightianum</i> in the entire profile at 12 cm before rapid decline to absence. Return to co-dominance of <i>Assulina muscorum</i> , <i>C. arcelloides</i> type and <i>T. arcula</i> type. Highest presence of <i>Bullinularia indica</i> and <i>Heleopera sphagni</i>
TMT-E (12 – 27 cm)	c. AD 1920 - 1480	Rapid appearance of <i>A. wrightianum</i> to co-dominate the zone with <i>A. flavum</i> . Decrease of <i>A. muscorum</i> and virtual absence of <i>H. subflava</i> .
TMT-D (27 – 49 cm)	c. AD 1480 - 980	Decline of <i>H. subflava</i> until total absence at 32 cm. Sharp peaks of <i>D. pulex</i> at 46 and 42 cm, and steady increases of <i>C. arcelloides</i> type and <i>A. muscorum</i> . Rise then slight decline of <i>Archerella flavum</i> peaking at 34 cm. <i>N. militaris</i> consistent but sparse throughout the zone with more

TMT-C (49 – 77 cm)	c. AD 980 - 310	frequent appearances of <i>Heleopera petricola</i> , <i>Heleopera sphagni</i> , <i>Assulina seminulum</i> and <i>Assulina scandinavica</i> . Zone dominated by <i>H. subflava</i> . Disappearance of <i>A. flavum</i> before a brief peak between 60 and 58 cm, and very low abundance of <i>A. wrightianum</i> . <i>A. muscorum</i> , <i>D. pulex</i> , <i>C. arcelloides</i> type and <i>T. arcuata</i> type feature at low abundance throughout the zone.
TMT-B (77 – 85 cm)	c. AD 310 - 50	Significant reduction in <i>H. subflava</i> , replaced by <i>Amphitrema wrightianum</i> and <i>A. flavum</i> and increased presence of <i>T. arcuata</i> type and <i>C. arcelloides</i> type.
TMT-A (85 – 100 cm)	c. AD 50 – 410 BC	Generally dominated by a very high percentage of <i>Hyalosphenia subflava</i> , though some significant presence of <i>Archerella flavum</i> with peaks at 98 and 96 cm. <i>Assulina muscorum</i> <i>Diffugia pulex</i> , <i>Nebela militaris</i> , <i>Cyclopyxis arcelloides</i> type and <i>Trigonopyxis arcuata</i> type occur throughout the zone at low abundance.
<i>Plant macrofossils</i>		
TMM-E (0 – 13 cm)	c. AD 2009 - 1910	<i>Sphagnum</i> disappears totally from the profile, replaced by <i>R. lanuginosum</i> and <i>E. vaginatum</i> which peaks at 4 cm.
TMM-D (13 – 23 cm)	c. AD 1910 - 1610	Only appearance of <i>Sphagnum papillosum</i> in the profile, with some <i>S. s. Acutifolia</i> also present. <i>Racomitrium lanuginosum</i> and <i>S. cespitosus</i> also present and undifferentiated monocotyledon remains reduce. Macro-charcoal remains virtually absent.
TMM-C (23 – 37 cm)	c. AD 1610 - 1170	Reappearance of <i>Sphagnum</i> to dominate the zone – mainly <i>S. s. Acutifolia</i> with very low presence of <i>S. s. Cuspidata</i> . Reduced UOM. Slight increase in <i>Scirpus cespitosus</i> . Reduction in charcoal to total absence.
TMM-B (37 – 87 cm)	c. AD 1170 - 10	Dominance of UOM and monocot remains. Increasing <i>E. vaginatum</i> . Very low presence of <i>Sphagnum</i> , though a small peak of <i>S. austinii</i> at 50 cm and a similar peak of <i>S. section Acutifolia</i> at 46 cm. Charcoal present throughout the zone with peaks at 80 cm and 50 cm. <i>Cenococcum</i> spores in high numbers at 68 cm and 56 cm.
TMM-A (87 – 100 cm)	c. AD 10 – 410 BC	Zone dominated by <i>Sphagnum</i> at the base, declining steadily to absence at 87 cm. Leaf count predominantly <i>S. austinii</i> which then virtually disappears throughout the rest of the profile. Increasing UOM and monocot as <i>Sphagnum</i> declines. Low levels of Ericaceous remains and <i>Eriophorum vaginatum</i> . Significant charcoal count at 98 cm declining with <i>S. austinii</i> .
<i>Pollen</i>		
TMP-D (0 – 14 cm)	c. AD 2009 - 1890	Rapid decline of arboreal pollen to the lowest abundance in the profile. Rapid increase in <i>Calluna</i> toward dominance (60%) at 4 cm. Increase in Poaceae and reduced Cyperaceae toward the surface.
TMP-C (14 – 30 cm)	c. AD 1890 - 1420	Decline of both arboreal pollen and Ericaceae, replaced by Cyperaceae. <i>Plantago</i> becomes more prominent from 24 cm. Higher numbers of <i>Sphagnum</i> spores encountered in the count.
TMP-B (30 – 78 cm)	c. AD 1420 - 280	Arboreal pollen is dominant and stays relatively stable at 42% ± 9. Cyperaceae, <i>Calluna</i> , <i>Erica</i> and Poaceae all feature throughout the zone fairly consistently, though a sharp trough in the total Ericaceae pollen sum is evident at 64 cm. A very slight peak of <i>Plantago</i> features at 60 cm.
TMP-A (78 – 100 cm)	c. AD 280 BC – 410 BC	Reduction in all arboreal pollen from 55% to 25% of the pollen sum. Slight increase in Poaceae. <i>Plantago</i> is present throughout, but declining through time. There is a sharp increase in Cyperaceae from 92 cm to 84 cm

Supplementary data

Spheroidal carbonaceous particle profile at MTM

