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Evidence for the Little Ice Age in upland northwestern Europe: Multiproxy climate data from three blanket mires in northern England

Julia C Webb,[1] Julia McCarroll,[1] Frank M Chambers[1] and Tim Thom[2]

[1] Centre for Environmental Change and Quaternary Research, School of Natural and Social Sciences, University of Gloucestershire, UK

[2]Yorkshire Peat Partnership, Yorkshire Wildlife Trust, UK

Corresponding author:

Frank M Chambers, Centre for Environmental Change and Quaternary Research, School of Natural and Social Sciences, University of Gloucestershire, Francis Close Hall Campus, Swindon Road, Cheltenham GL50 4AZ, UK.

Email: fchambers@glos.ac.uk

ORCID iD

Frank M Chambers https://orcid.org/0000-0002-0998-2093

Abstract

The Little Ice Age (LIA) is a well-recognised palaeoclimatic phenomenon, although its causes, duration and severity have been matters of debate and dispute. Data from a wide range of archives have been used to infer climate variability before, during and after the LIA. Some published proxy-climate data from peatlands imply that two particularly severe episodes within the LIA may be contemporaneous between hemispheres; these echo a previous climatic downturn ca. 2800 cal BP of similar severity but lesser duration. Here, we present palaeoclimate data from the mid- to late-Holocene, reconstructed from three blanket peats in Yorkshire: Mossdale Moor, Oxenhope Moor and West Arkengarthdale. Multiproxy techniques used for palaeoclimatic reconstruction were plant macrofossil, pollen and humification analyses. Dating was provided by a radiocarbon-based chronology, aided by spheroidal carbonaceous particles (SCPs) for all sites, and ^{210}Pb dates for one. The LIA presents as a distinct climatic event within each palaeoenvironmental record at the three sites. These indications are compared with terrestrial datasets from northwest Europe and elsewhere. A broad degree of synchronicity is evident, signifying that the LIA is one of the most pronounced downturns in global climate in the last ca. 6000 years, and arguably the most routinely recorded within the Holocene.

Keywords
blanket peat, Holocene, Little Ice Age, palaeoclimate, palaeoecology, peatlands

Introduction

The Little Ice Age (LIA) is one of the most renowned climate instabilities of the Holocene, with causes being attributed variously to reduced solar activity (Blackford and Chambers, 1995; Grove, 2001; Mauquoy et al., 2002, 2004), volcanic eruptions (Miller et al., 2012) and even to human impact (Ruddiman, 2003). However, Turner et al. (2016) note that solar-type signals in peatlands can be the product of random variations alone, and caution that a more critical approach is required for the vigorous interpretation of such signals. The coldest temperatures of the LIA have been inferred for the interval AD 1400–AD 1700 (Mann et al., 2009), although there is disagreement on the start and end dates for this climatic phenomenon (Fagan, 2001). Matthews and Briffa (2005) suggest that LIA ‘climate’ (as opposed to glacier advance) is defined as a time interval of about 330 years (ca. AD 1570–1900) when Northern Hemisphere summer temperatures (land areas north of 20°N) fell significantly below the AD 1961–1990 mean, whereas in the North Atlantic, the LIA onset is said to be as early as before the early 14th century (Grove, 2001). However, some regional heterogeneity is apparent for northwestern European peatlands, with Mauquoy et al. (2002) identifying the LIA from Lille Vildmose, Denmark and Walton Moss, Cumbria between AD 1449–1464 and AD 1601–1604; De Vleeschouwer et al. (2009) from Slowinskie Blota, Poland AD 1200–1800; the beginning of the LIA AD 1250–1350 as identified by Barber et al. (2004) from Dosenmoor, Germany and Svanemose, Denmark; AD 1410–1540 and AD 1660–1720 from Letterfrack, Ireland (Blackford and Chambers, 1995); and the ‘last pulse’ of the LIA as AD 1650–1850 from Fallahogy, Northern Ireland and Moine Mhor, Scotland (Barber et al., 2000). The geographical variation in these timings is evidence for the multiphase characteristic of the LIA, with possible attribution to the Wolf, Spörer, Maunder and Dalton Minima (Blackford and Chambers, 1995): periods of reduced sunspot activity that occurred at AD 1320 (Wolf), AD 1400–1540 (Spörer), AD 1645–1715 (Maunder) and 1790–1830 (Dalton) (McCracken and Beer, 2014). However, cold temperatures have been recorded outside of these timings, yet still within the LIA, with the early Maunder Minimum (ca. AD 1600), being described as one of the coldest phases of the LIA, with temperatures considered as extreme (Mauquoy et al., 2004). Overall, temperature estimates for the LIA suggest that this event must be considered a time of modest cooling of the Northern Hemisphere, with temperatures dropping by approximately 0.6°C during the 15th–19th centuries (Mann, 2002), which provides motivation behind much academic interest and the focus of the present paper.

The above records all derive from the Northern Hemisphere, but whether the LIA can be regarded a global event has been disputed, largely owing to a relative lack of terrestrial archives in the Southern Hemisphere (Chambers, 2016). A recent paper provides new evidence, supporting the claim that the LIA may be a global phenomenon: using identical methods on a peatland in Tierra del Fuego as have been used for peatlands in northwest and central Europe, the most pronounced episodes of the LIA (at ca. cal AD 1460 and from ca. cal AD 1675 to cal AD 1770 in the Tierra del Fuego mire) were shown to be contemporaneous between hemispheres (Chambers et al., 2014). The bog response in the Southern Hemisphere site (to surface dryness) was opposite to site responses in the Northern Hemisphere (increased surface wetness), which the authors interpreted as indicating globally contemporaneous equatorward shifts of moisture-bearing airmasses during the severest decades.

Proxy-climate data can and have long been inferred from late-Holocene peat stratigraphy (Amesbury et al., 2012a, 2012b; Anderson et al., 1998; Barber, 1981; Barber et al., 1994, 2000, 2003, 2013; Bingham et al., 2010; Blackford, 2000; Blackford and Chambers, 1991, 1995; Blundell and Barber, 2005; Chambers et al., 1997; Charman, 2010; Chiverrell, 2001; Daley and Barber, 2012; Hughes et al., 2000; Sillasoo et al., 2007), with many studies suggesting an association between the visible stratigraphy of peatlands and past climatic change (Barber et al., 2003; Blackford and Chambers, 1991; Stoneman, 1993; Xie et al., 2004). Peatlands contain within them a detailed archive of local and regional vegetation history, making them ideal for researching Holocene environmental and climatic changes (Blackford, 2000).

In the case of blanket mires, such as those in the present study, peat can appear homogenous with little or no stratigraphy; this deterred proxy-climate research in previous decades, as did the perception that some blanket mires contain few identifiable macrofossils (Chambers et al., 1997). This apparent unsuitability was falsified by Chambers (1984) and Blackford and Chambers (1991) using humification data and Tallis (1995) and Mauquoy and Barber (1999) using plant macrofossil data, with blanket peat being used to reconstruct Holocene palaeoclimate in a number of recent studies (Blundell and Holden, 2015; Castro et al., 2015; Swindles et al., 2015).

A strategic way of indicating a regional forcing factor, such as climate change, is to be able to demonstrate a synchronous response in separate mires (Hughes et al., 2000). However, the dating of the LIA can be problematic in peatlands, given that the most widely used dating method is conventional radiocarbon dating, which may have limitations in providing sufficient precision and accuracy for the last few hundred years. Conversely, using techniques such as wiggle-matching and bomb-spike dating at a high enough resolution, it is possible to use radiocarbon alone to produce chronologies to date events such as the LIA (Chambers et al., 2014). Other dating methods are available for peatlands covering the later part of the LIA, for example, tephra, ^{210}Pb, ^{137}Cs and relative dating using Spheroidal Carbonaceous Particles (SCPs) (van der Plicht et al., 2013), but these methods are not without their flaws. van der Plicht et al. (2013) highlight that ^{210}Pb age estimates can be too old, and as such, age estimates of peat samples based only on ^{210}Pb should be used with caution. SCPs can only provide relative age estimates, which are subject to geographical variation (Swindles, 2010).

The LIA has previously been identified in palaeoclimatic reconstructions from peatlands in continental Europe, including Denmark (Aaby, 1976; Barber et al., 2004) and Germany (Barber et al., 2004), and in Britain, including Temple Hill Moss, Scotland (Langdon et al., 2003), Talkin Tarn, Cumbria (Barber and Langdon, 2007), the North York Moors, northern England (Chiverrell, 2001), Greater Manchester and Cheshire, northwest England (Davis and Wilkinson, 2004), Northumberland, northern England (Charman et al., 1999), Cumbria and the Scottish Borders, northern Britain (Mauquoy and Barber, 1999), northern England and Scotland (Stoneman, 1993), the northern Pennines (Swindles et al., 2015) and Yorkshire (Turner et al., 2014). This paper contributes to the record of the LIA in the uplands of northern England with the addition of palaeoclimate data from three Yorkshire upland blanket bogs.

The present study aims to examine the LIA in northern England, based on palaeoclimatic reconstructions using pollen, plant macrofossil and humification data from three Yorkshire blanket bogs: Mossdale Moor (MDM), Oxenhope Moor (OXM) and West Arkengarthdale (ARK). The long-term ecological (palaeoecological) records of these sites were published elsewhere (McCarroll et al., 2016a, 2016b, 2017); in the present paper palaeoclimate data for the LIA are compared with the overall palaeorecords from the three sites, and with proxy-climate data from sites in northwest Europe and elsewhere, to evaluate the LIA compared with previous climatic downturns in the Holocene.

Site descriptions

Site descriptions of MDM, OXM and ARK have previously been provided by McCarroll et al. (2016a, 2016b, 2017) in studies using palaeoecology to advise conservation. For the purpose of this study, only a summary of each site is provided here.

Mossdale Moor

Mossdale Moor is a degraded blanket mire at 550 m altitude located in Upper Wensleydale, North Yorkshire, UK (Figure 1). The modern-day peat supports species characteristic of National Vegetation Classification (NVC) M19 (*Calluna vulgaris-Eriophorum vaginatum* blanket mire) (Rodwell, 1998). Various management practices have altered the vegetation characteristics of the moor from typical blanket mire towards heathland communities.

Oxenhope Moor

Oxenhope Moor is a degraded blanket mire at 430 m altitude located north of Hebden Bridge in West Yorkshire (Figure 1). The modern-day peat supports species characteristic of NVC types M20 (*Eriophorum vaginatum* blanket and raised mire) and M25 (*Molinia caerulea-Potentilla erecta* mire as surveyed by Natural England in 2008) (Rodwell, 1998).

West Arkengarthdale

West Arkengarthdale is a blanket mire at 380 m altitude located north-west of Reeth in North Yorkshire (Figure 1). A vegetation survey conducted during fieldwork identified that the modern-day peat supports species characteristic of NVC type M20 (*Eriophorum vaginatum* blanket and raised mire) (Rodwell, 1998).

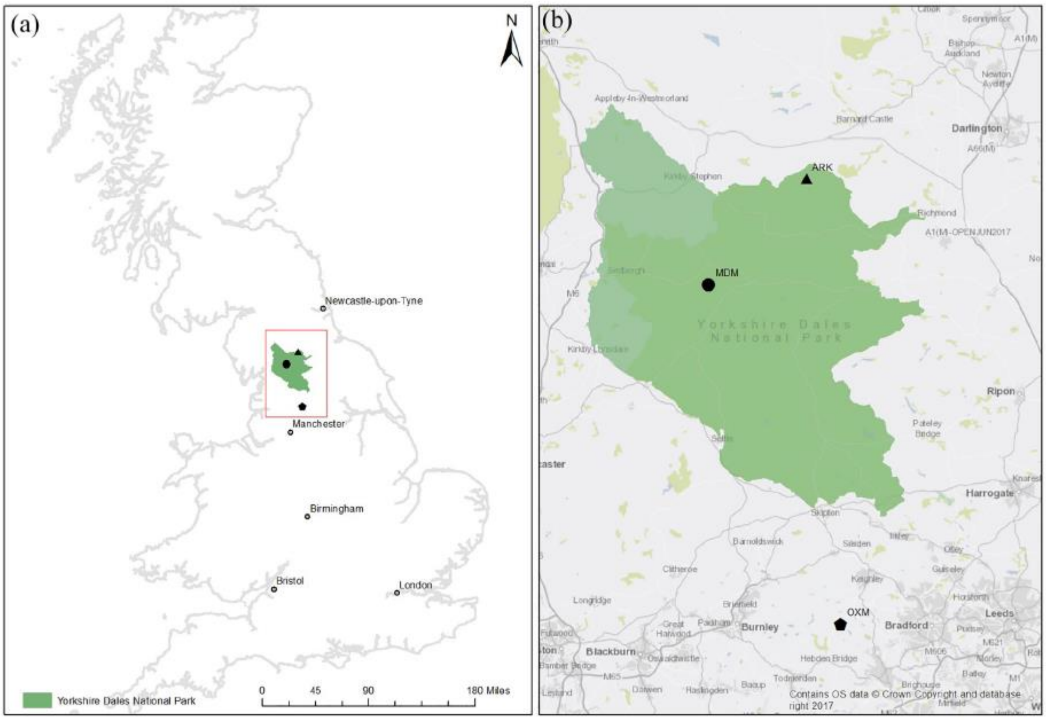

Figure 1 (a) Map of United Kingdom showing the site locations of Mossdale Moor, Oxenhope Moor and West Arkengarthdale. (b) Map of Yorkshire Dales National Park showing site locations of Mossdale Moor (MDM: circle), exact coring location at Latitude: 54.300292°N, Longitude: 2.315507°W; Oxenhope Moor (OXM: pentagon), exact coring location at Latitude: 53.793759°N, Longitude: 1.977952°W; and West Arkengarthdale (ARK: triangle), exact coring location at 54.458815°N, Longitude: 2.067067°W

Methods

Field sampling strategy

The field sampling strategy was a modified version of the method used in the ACCROTELM Research Project (Chambers, 2006; De Vleeschouwer et al., 2010). First, site morphology was established by measuring peat depths across each site to allow the deepest ombrotrophic zone closest to the highest point of the bog to be identified. By describing multiple cores using the Troels-Smith (1955) method, the gross site stratigraphy was determined, in turn allowing the identification and selection of the master core, taken in a zone between a hummock and a hollow, as recommended by De Vleeschouwer et al. (2010). Using a 5 cm diameter Russian corer, overlapping adjacent cores were extracted, described, photographed for the full length of the profile before being placed in labelled plastic guttering, wrapped in airtight carbon-stable bags and transported to the laboratories at the University of Gloucestershire where they were stored at 4°C.

Laboratory methods

Laboratory methods of analysis were the same at all three sites, with the exception of Mossdale Moor, where ^{210}Pb dating was also undertaken. Radiocarbon dates were obtained from plant macrofossils where possible (11 of 27 samples – see Table 1). Where this was not possible, bulk peat samples were sent to Beta Analytic Miami for analysis, where following pre-treatment, the plant fraction (as opposed to bulk organic carbon) of the sample was selected for dating. The depths selected for radiocarbon dating were chosen following pollen analysis and their positioning at particular points of interest or on boundaries selected by CONISS (an agglomerative cluster analysis technique that compares the total pollen assemblage of each sample with that of its stratigraphic neighbours) and by eye. The pollen preparation method used in this study does not degrade SCPs (Chambers et al., 2011b) and therefore SCPs were counted alongside pollen. 210Pb samples from Mossdale Moor were analysed every 1 cm for the first 32 cm depth at the University of Exeter using alpha spectrometry (rather than gamma spectrometry; see Zaborska et al. (2007) for comparison), which measures the decay of ^{210}Po, a daughter product of ^{210}Pb. As a chemical yield tracer, a Polonium spike (^{209}Po) was added before Po was extracted from the sample using acid digestion and electroplated onto a silver disc. Ages were calculated using the Constant Rate of Supply (CRS) model (Appleby and Oldfield, 1978).

Peat humification analysis has been subject to criticism in recent decades (Hughes et al., 2012; Yeloff and Mauquoy, 2006). Studies show that changes in humification may not be influenced solely by climate and that plant species composition can also be a contributor to observed changes (Hughes et al., 2012; Yeloff and Mauquoy, 2006). Yeloff and Mauquoy (2006) suggest that the assumption that the major influence on peat humification is the surface wetness of the bog (and therefore climate) should be tested. Hughes et al. (2012) found that, overall, humification data are validated, but that the species signal is sometimes sufficient to change the timing and number of events and trends recorded in the data. Furthermore, the role of secondary decomposition during dry periods may also influence the humification signal (Borgmark and Schoning, 2006). Provided the shortfalls are considered, humification analysis can aid palaeoclimatic reconstructions when incorporated as part of a multiproxy palaeoclimate study; this method remains a straightforward and fast way to produce high-resolution, contiguous datasets.

The preparation and quantification of humification samples follows a modified methodology based on that described by Chambers et al. (2011a). The protocol used differs where 0.1 g of sediment is used as opposed to 0.2 g. Humification was analysed contiguously at every centimetre for each site, which equates to a 10-year sampling resolution. 1 cm^3 of peat was sampled and oven-dried for 48 h at 60°C before being ground and returned to the weighing boat. 0.1 g of sample was accurately weighed, recorded and then transferred into a 150 mL beaker where 100 mL of 8% NaOH solution was added to each beaker before the samples were simmered on a hotplate for 1 h. The contents of each beaker were then poured into a 200 mL labelled volumetric flask before being filtered into separate 50 mL labelled volumetric flasks. Only 50 mL of filtrate was transferred. The samples were then measured in a spectrophotometer (set at a wavelength of 540 nm). Measurements were repeated three times and an average recorded.

The preparation and quantification of pollen samples follows a modified methodology based on that of van Geel (1978) in Chambers et al. (2011b). The sampling resolution was a minimum of 8 cm at Oxenhope Moor, and 4 cm at Mossdale Moor and Arkengarthdale, with a maximum resolution of 2 cm at points of interest, equating to a minimum of a 40-year sampling resolution over the LIA. *Lycopodium clavatum* tablets were added to the samples in order to calculate microfossil concentrations. Pollen grains were identified using Moore et al. (1991) and a reference collection of type slides at the University of Gloucestershire. The pollen sum (500 grains) is Total Land Pollen, which includes the total number of pollen from trees, shrubs and terrestrial herbaceous plants.

For the analysis of plant macrofossils, a minimum sampling resolution of 8 cm at OXM and 4 cm at MDM and ARK was employed. Sub-samples measuring 4 cm3 were taken using a scalpel and sieved through a 125 μm mesh with a standard 5 L volume of tap water. The samples were transferred to three glass petri dishes and spread out to form a monolayer, before quantification using the quadrat and leaf count (QLC) method of macrofossil analysis described by Barber et al. (1994). Plant macrofossils were identified using type collections and with reference to modern plant material sampled from the study sites. In addition, Daniels and Eddy (1985) was used to identify *Sphagna* and Smith (2004) was used for non-*Sphagnum* bryophytes.

An adapted version of the weighted-average Dupont Hydroclimatic Index (DHI) (Dupont, 1986) was applied to the macrofossil data. Weights were assigned to species based on those used in Daley and Barber (2012) and Mauquoy et al. (2008) and the DHI scores were calculated in Excel using plant macrofossil percentages.

Results and interpretation

The data presented in this paper have previously been published elsewhere (McCarroll et al., 2016a, 2016b, 2017) as individual sites with a focus on using palaeoecology to inform conservation. These records have now been brought together, with the addition of seven radiocarbon dates (four for MDM; three for ARK) to revise chronologies. The data are now used to tackle a different palaeoenvironmental research question, providing insight into palaeoclimatic changes in upland Yorkshire blanket bogs over the span of the LIA.

Basis of chronology

Four of the 27 radiocarbon dates obtained did not show increasing age with depth (three from MDM, one from ARK), suggesting that the accumulation of peat may have ceased or that some material may have been eroded (Table 1). At both sites, these anomalies occur approximately at the mid depths of the profiles, at a time predating the LIA. Ages were modelled in OxCal version 4.2 (Ramsey, 2009), to include dates from Spheroidal Carbonaceous Particles (SCPs) and ^{210}Pb (site MDM) (Figures 2 and 3); those that show increasing age with depth were accepted. Calibration of the radiocarbon ages to calendar years BP was conducted using INTCAL13 (Reimer et al., 2013) in OxCal version 4.2.

^{210}Pb dating

Depth profiles for ^{210}Pb typically show exponential decline with depth throughout the core (Figure 3). The Constant Rate of Supply (CRS) model was used to develop an age-depth model from 210Pb data using the method described by Appleby (2001). The CRS model is thought to be the most appropriate in ombrotrophic peats as ^{210}Pb inputs are dominated by atmospheric inputs (Appleby, 2008). Uncertainty was calculated by colleagues at Northeast Normal University, China using analytical uncertainty and the error propagated using the CRS model. The level of activity decreases below 17 cm depth (see Supplemental Material) and at this point, the error margin becomes unreliably larger, and the dates calculated exceed beyond the 130-year range of ^{210}Pb dating (Appleby, 2008) and so were excluded from the age-depth model.

Table 1 Dates for MDM, OXM and ARK including ^{210}Pb (MDM only), SCPs and radiocarbon[1].

Depth (cm)	**Radiocarbon Date (yr BP); SCP/^{210}Pb date (AD)**	**Lab number**	**Calibrated Age (95.4%) (cal yr BP) from-to**	**Material/method**
MDM				
1	2008 ± 2			^{210}Pb
1.5	1980 ± 10			SCPs
2	2001 ± 2			^{210}Pb
3	1991 ± 3			^{210}Pb
3.5	1950 ± 10			SCPs
4	1982 ± 4			^{210}Pb
5	1975 ± 4			^{210}Pb
6	1968 ± 5			^{210}Pb
7	1959 ± 6			^{210}Pb
8	1949 ± 8			^{210}Pb
9	1941 ± 8			^{210}Pb
10	AD 1933 ± 10			^{210}Pb
11	AD 1925 ± 12			^{210}Pb
12	AD 1919 ± 14			^{210}Pb
13	AD 1906 ± 17			^{210}Pb
13.5	AD 1850 ± 10			SCPs
14	AD 1891 ± 22			^{210}Pb
15	AD 1878 ± 26			^{210}Pb
34.5	530 ± 30	BETA-327196	645-502	*Polytrichum commune*
51.5	770 ± 30	BETA-426568	734-668	Bulk plant material
58.5	480 ± 30*	BETA-327197	625-469	Monocot leaves
65.5	1070 ± 30	BETA-385275	1055-929	Bulk plant material
69.5	930 ± 30*	BETA-426569	925-785	Bulk plant material
74.5	350 ± 30*	BETA-327198	503-306	Monocot leaves
94.5	1340 ± 30	BETA-327199	1334-1180	Monocot leaves
118.5	1520 ± 30	BETA-327200	1528-1314	Monocot leaves
135.5	2610 ± 30	BETA-385276	2775-2720	Bulk plant material
150	5310 ± 30	BETA-364579	6267-5944	Charred material
OXM				
0.5	1990 ± 10			SCPs
8.5	1945 ± 10			SCPs
24.5	1850 ± 10			SCPs
28.5	350 ± 30	BETA-382650	503-306	Bulk plant material
108.5	1300 ± 30	BETA-382651	1304-1088	Bulk plant material
212.5	2270 ± 30	BETA-382658	2354-2152	Bulk plant material
310.5	3170 ± 30	BETA-382653	3480-3253	Bulk plant material
380.5	3910 ± 30	BETA-382654	4512-4160	Plant macrofossils
436.5	4730 ± 30	BETA-382655	5588-5322	Bulk plant material
479.5	6090 ± 30	BETA-382656	7159-6799	Plant macrofossils
ARK				
16.5	1850 ± 10			SCPs
12.5	1950 ± 10			SCPs
8.5	1970 ± 10			SCPs
4.5	1990 ± 10			SCPs
39.5	300 ± 30	BETA-381604	485-156	Plant macrofossils
100.5	1780 ± 30	BETA-385277	1822-1570	Plant macrofossils
141.5	2710 ± 30	BETA-426566	2859-2758	Bulk plant material
147.5	3410 ± 30	BETA-379805	3828-3563	Wood
155.5	3430 ± 30	BETA-444008	3830-3578	Bulk plant material
159.5	3640 ± 30	BETA-385278	4089-3841	Plant macrofossils
182.5	3320 ± 30*	BETA-379806	3685-3450	Bulk plant material
199.5	3970 ± 30	BETA-426567	4524-4299	Bulk plant material
260.5	4820 ± 30	BETA-379807	5645-5468	Bulk plant material
289.5	5610 ± 30	BETA-379808	6478-6300	Bulk plant material

[1]Initial radiocarbon chronologies of MDM and ARK are discussed in McCarroll et al. (2016b, 2017). However, further radiocarbon dates have beenobtained for the purpose of the present paper for these sites, so interpretation is briefly revised.
*Date rejected as an outlier.

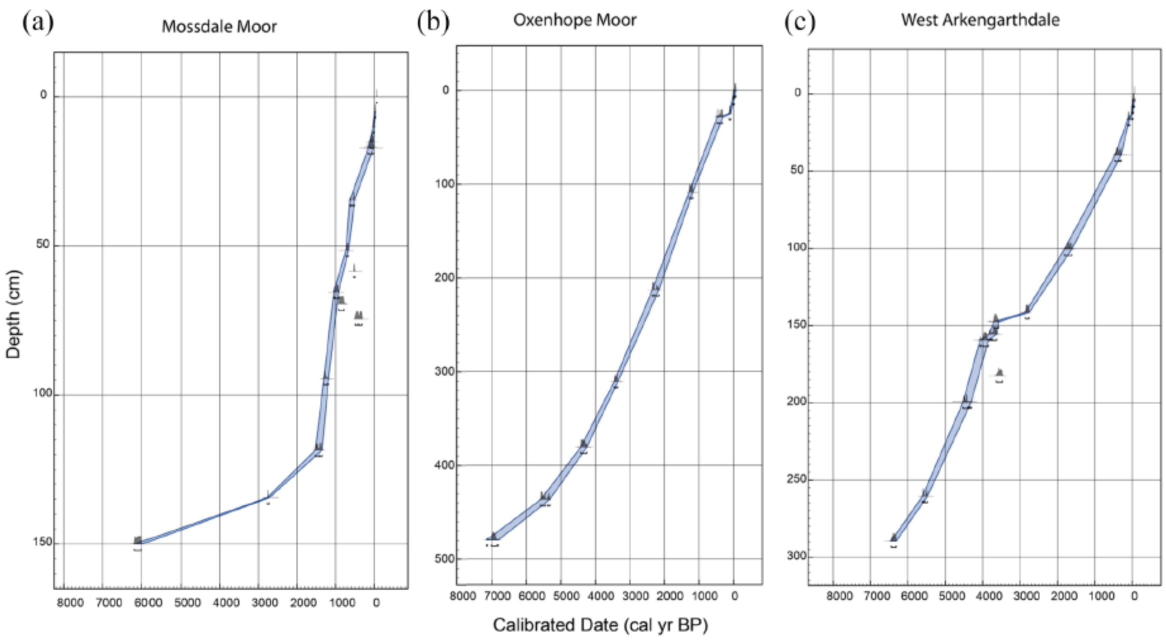

Figure 2 Full sequence age-depth models for (a) MDM, (b) OXM and (c) ARK using radiocarbon, SCP and ^{210}Pb dates (MDM only).

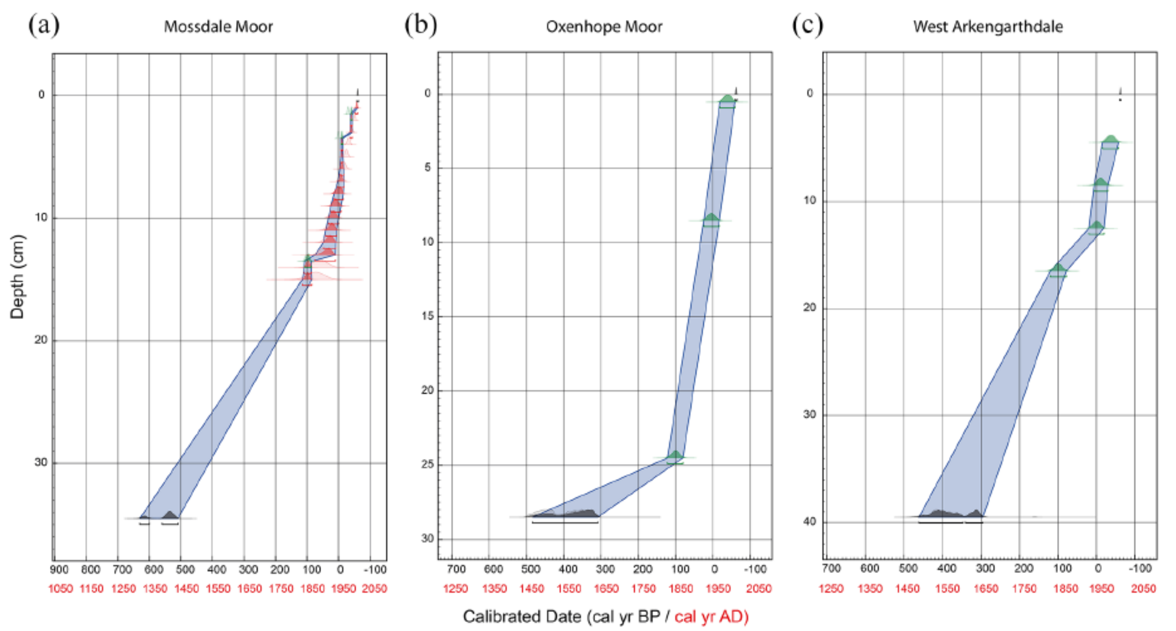

Figure 3 LIA age-depth models for (a) MDM, (b) OXM and (c) ARK using radiocarbon (grey), SCP (green) and ^{210}Pb dates (red – MDM only).

Humification

Mossdale Moor.

At the base of the profile, high T values (detrended % transmitted light) suggest low humification (ca. 6000–5000 cal yr BP) (Figure 4a). Following this, humification is higher between 5000 and 1600 cal yr BP, as indicated by low T values. T values then fluctuate, suggesting alternations between more and less humification occurred, but the peat appears more humified between ca. 1400 cal yr BP and 700 cal yr BP (cal yr AD 1250), which overlaps with and could be interpreted as the climatic amelioration known as the Mediaeval Warm Period (MWP).

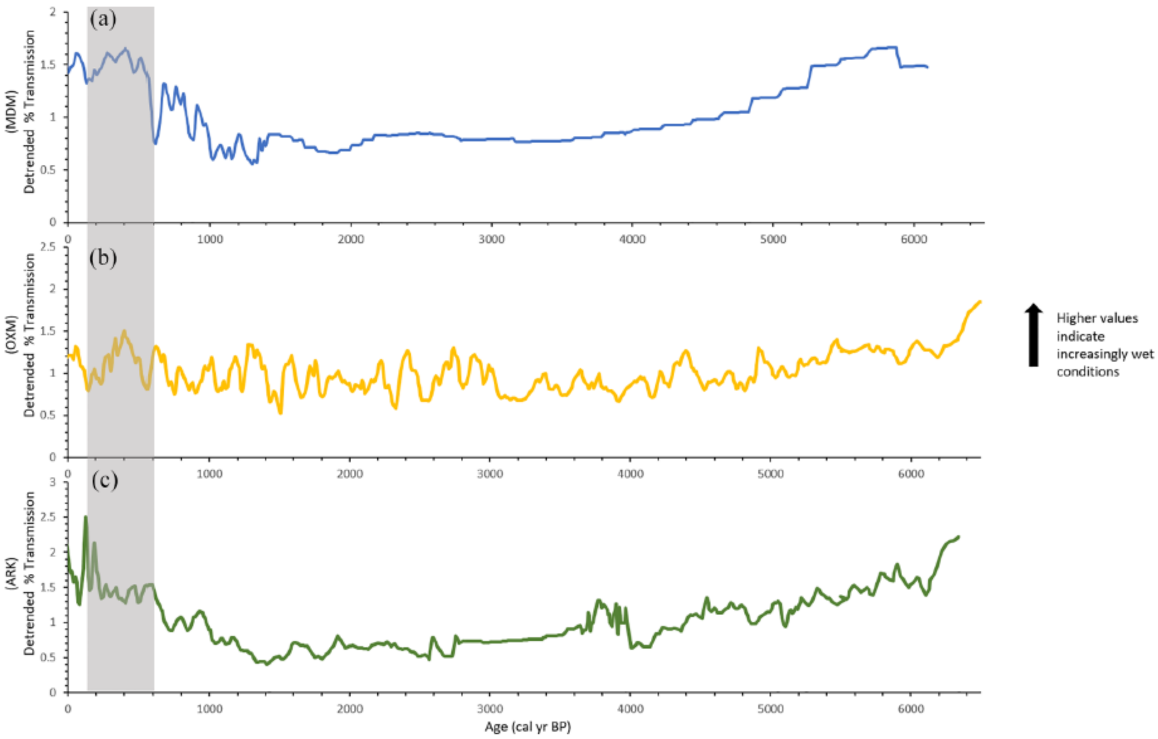

Figure 4 Humification data plotted against age for (a) MDM, (b) OXM and (c) ARK. %T values have been detrended and are therefore displayed as residual values from a linear trendline. The shaded bar approximates the LIA, for visual guidance only.

Between ca. 600 and 140 cal yr BP, generally high T values suggest low humification. However, the earlier date may reflect the low decomposition rates of the acrotelm (5–10 cm depth) and that this plant material has not had as long to decompose as the lower layers of peat in the catotelm. The T values are particularly high between ca. 500 and 300 cal yr BP (beneath the acrotelm), which is interpreted likely as the cooling associated with the Little Ice Age (LIA).

Oxenhope Moor.

The T curve is highly fluctuating and could be evidence of cyclicity. There is less humification taking place between 7000 and 5000 cal yr BP when compared with the rest of the profile, with notably high

T values, the highest in the bottom half of the profile, between 7000 and 6000 cal yr BP. More humification occurred between 5000 and 3000 cal yr BP (Figure 4b), followed by lower humification identified by higher T values between 3000–2600 cal yr BP and 1300–1100 cal yr BP. Furthermore, there are particularly high T values towards the surface of the profile particularly between 500 and 300 cal yr BP consistent with the Little Ice Age. The values here are higher than the rest of the profile (with the exception of high T values between 7000 and 6000 cal yr BP). T then briefly drops before rising again towards the surface, interpreted as fresh peat in the acrotelm that has comparatively less time to decompose than the peat below.

West Arkengarthdale.

Towards the base of the profile, between 6200 and 4200 cal yr BP, T values increase with age suggesting low humification, reaching the highest T values at ca. 6200 cal yr BP (Figure 4c). T values are low suggesting more humification at 4000 cal yr BP but less between 3900 and 3600 cal yr BP. Lower T values indicate more humification taking place between 3500 and 1100 cal yr BP, gradually becoming less towards the surface of the profile. T values are highest between 600 and 150 cal yr BP (being particularly high between 300 and 150 cal yr BP) and is interpreted as the Little Ice Age (LIA). T values are then briefly lower subsequently up until approximately cal yr AD 1960 with a return to less humified conditions. Despite this, peat at this depth has had much less time to decompose than the layers of peat below and therefore the high values here are likely a reflection of this.

Pollen and charcoal

Each of the sites was likely surrounded by mixed woodland as indicated by high percentages of tree and shrub pollen: at MDM, between ca. 6100 and 5000 cal yr BP; at OXM, between ca. 7000 and 5500 cal yr BP; and at ARK, between ca. 5600 and 4400 cal yr BP (Figure 5 and Supplemental Material). Mixed mire communities then succeed at each of the sites, evidenced by *Sphagnum* spores and Ericales species at MDM, between ca. 5000 and 1500 cal yr BP; OXM, between ca. 5500 and 3400 cal yr BP; and ARK, between ca. 4400 and 3700 cal yr BP.

At OXM, between ca. 3400 and 2300 cal yr BP, anthropogenic influence is suggested by an introduction of arable and ruderal species but percentages of tree and shrub pollen are still high at this stage. At ARK, between ca. 3700 and 600 cal yr BP, a mixed mire community (indicated by Ericales pollen and *Sphagnum* spores) prevails on site, likely surrounded by mixed woodland as tree and shrub pollen are still high. Charcoal is high, indicative of burning, possibly deliberately by human activity.

At MDM, between ca. 1500 and 1300 cal yr BP, pollen of trees and shrubs is lower than preceding sections, as are *Sphagnum* spores. A higher percentage of Ericales suggests drier conditions, perhaps

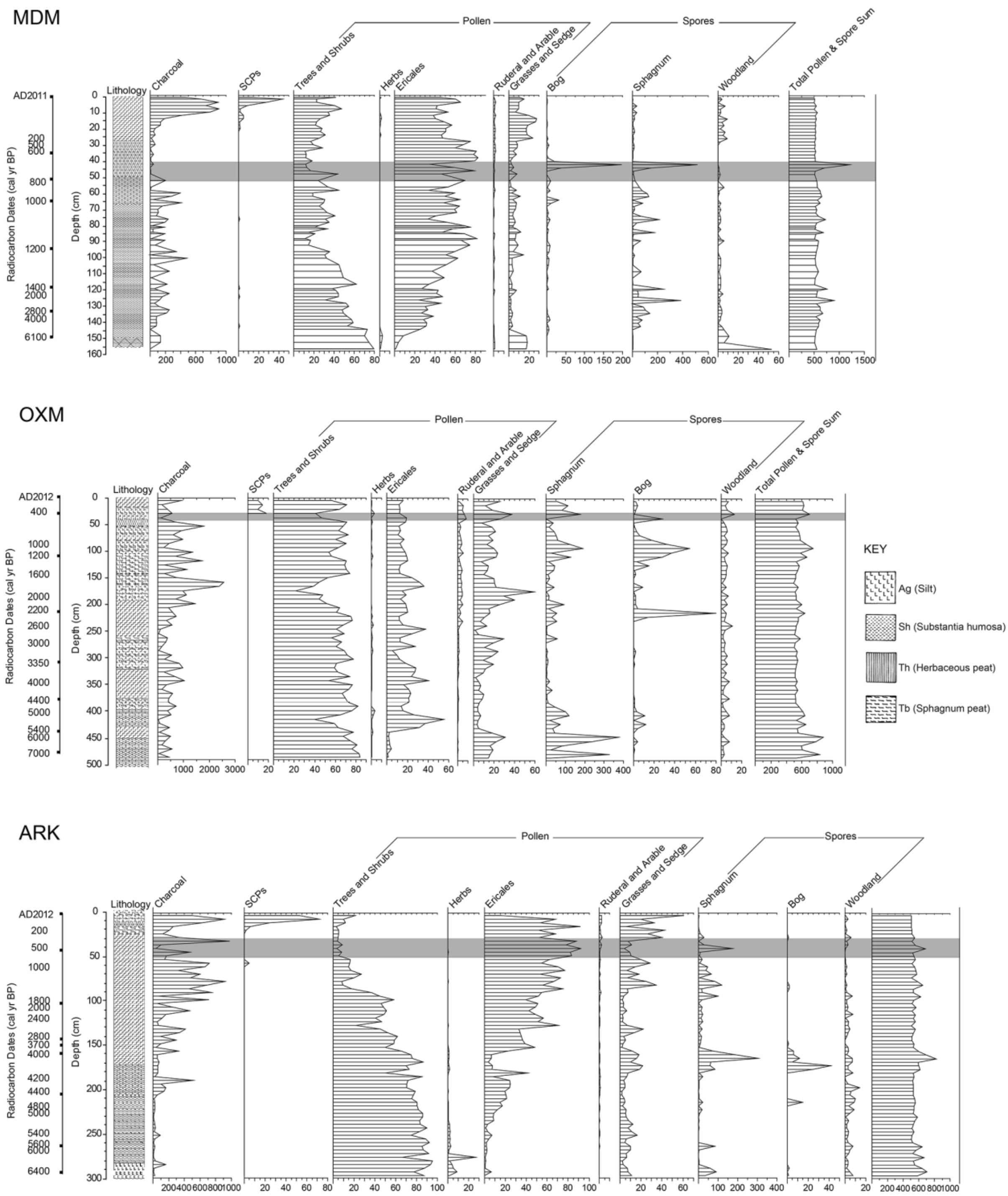

Figure 5 Summary pollen diagram for MDM, OXM and ARK. Shaded bar represents LIA. Full pollen diagrams available from McCarroll et al. (2016a, 2016b, 2017).

attributable to the Mediaeval Warm Period. At OXM, between ca. 2300 and 1200 cal yr BP, evidence suggests either a climatically dry environment or clearance by fire (high percentage of grasses and sedge pollen, high charcoal and fluctuating tree and shrub pollen, low *Sphagnum* spores). A brief move to wetter conditions is suggested by a peak in bog spores at ca. 2250 cal yr BP.

At MDM, between 1300 and 600 cal yr BP, wet conditions are suggested by low charcoal, and high percentages of *Sphagnum* and bog spores. At OXM, between ca. 1200 and 350 cal yr BP, wet conditions are also identified by a high number of *Sphagnum* and bog spores. Anthropogenic activity is also evidenced by a high percentage of ruderal and arable pollen.

Between ca. 600 and −60 cal yr BP (cal yr AD 1350–2010) at MDM and ARK, percentages of trees and shrub pollen are low in comparison to preceding sections, consistent with high charcoal numbers, indicative of clearance by man. However, at OXM, between ca. 350 and −60 cal yr BP (cal yr AD 1600–2100), charcoal fragments are low and *Sphagnum* spores are high, indicative of wet conditions.

Plant macrofossils

At OXM between ca. 7000 and 5200 cal yr BP, the presence of *Sphagnum* along with monocots and Cyperaceae (including *Eriophorum vaginatum*) suggests a mixed *Sphagnum*, sedge and graminoid mire (Figure 6 and Supplemental Material). At ARK, between ca. 6400 and 5600 cal yr BP, standing water is suggested by the presence of aquatic species (including *Nymphaea*), before the formation of peat. Once peat is established, Ericaceae, monocot roots and Cyperaceae species suggest the presence of a mixed ericaceous, sedge and graminoid community.

At MDM, between ca. 6100 and 2400 cal yr BP, UOM reaches 90%, owing to the very degraded nature of the blanket peat at this depth. At OXM between 5200 and 3200 cal yr BP, the percentage of UOM is high, Ericaceae (*Calluna vulgaris*) are present and the percentage of identified *Sphagnum* is lower than above, suggestive of drier conditions. At ARK, between ca. 5600 and 4400 cal yr BP, shallow water is indicated by the presence of aquatic species (including *Equisetum fluviatile*). However, a predominantly sedge community exists (consisting of *Trichophorum cespitosum, Eriophorum vaginatum* and *E. angustifolium*). Following on from this, between ca. 4400 and 3700 cal yr BP, wet conditions are indicated by the presence of *Sphagnum cuspidatum* and low UOM.

At MDM between ca. 2400 and 1300 cal yr BP, UOM is still high. Aquatic species (including *Equisetum fluviatile* and *Scheuchzeria palustris*) suggest shallow water and the presence of charred remains signifies local fire, before giving way to a sedge-graminoid community between ca. 1300 and 1100 cal yr BP, indicated by Cyperaceae and monocot species. At OXM, between ca. 3200 and 1000 cal yr BP, wet conditions are indicated by low UOM and a high percentage of *Sphagnum* Section A*cutifolia*, mainly *S. austinii*. A *Sphagnum*-sedge-Ericales-graminoid community is present, identified by the presence of Ericales rootlets, monocot roots, and Cyperaceae species. At ARK, between ca. 3700 and 500 cal yr BP, shallow water is indicated by the presence of aquatic species (*Scheuchzeria palustris* and *Equisetum fluviatile*). The community consisted of Cyperaceae

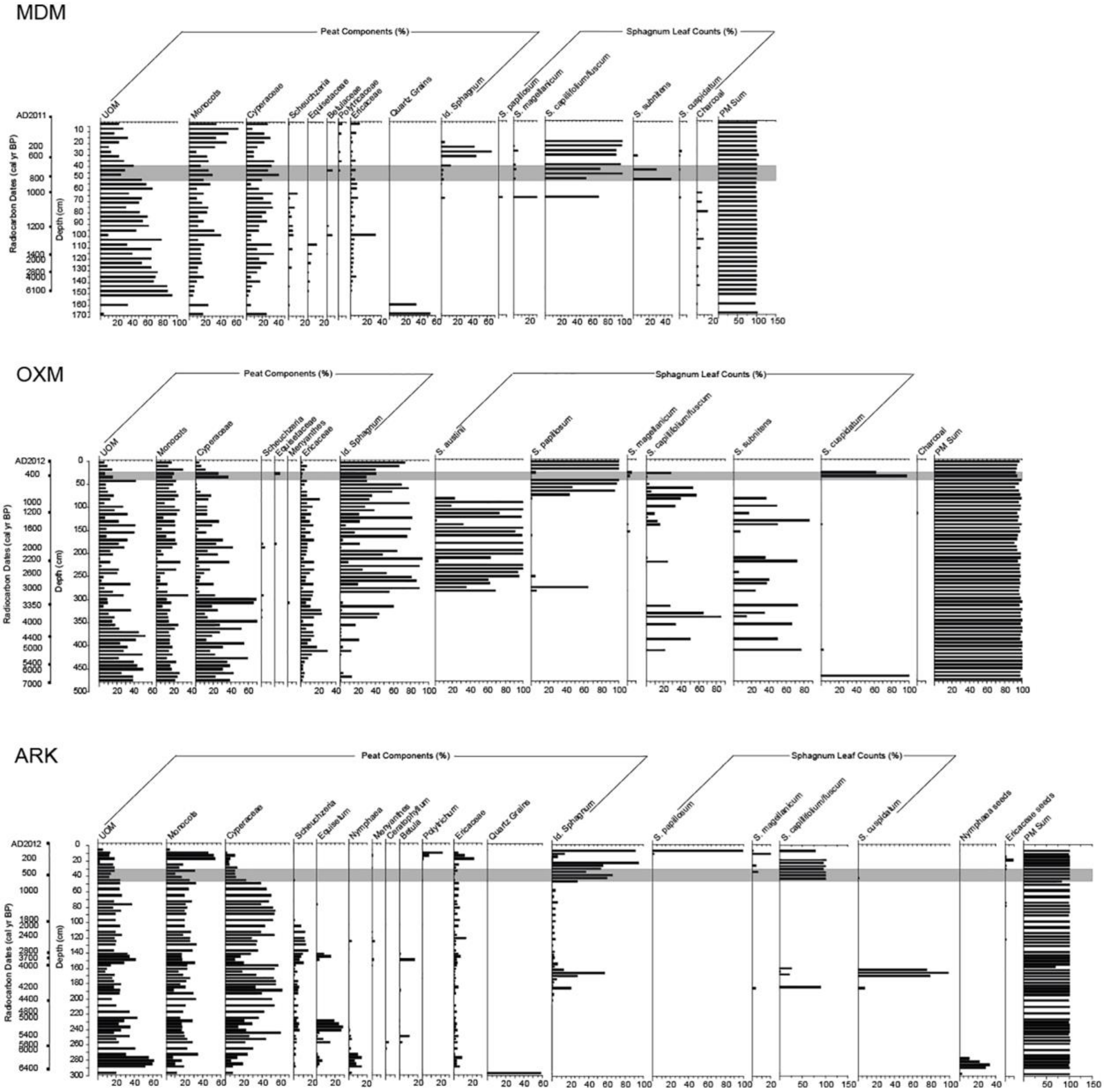

Figure 6 Summary plant macrofossil diagram for MDM, OXM and ARK. Shaded bar represents LIA. Full plant macrofossil diagrams indicating species within the summary groups Cyperaceae (*Eriophorum vaginatum*, *E. angustifolium* and *Trichophorum cespitosum*) and Ericaceae (*Calluna vulgaris*, *Erica tetralix* and *Vaccinium oxycoccus*) available from McCarroll et al. (2016a, 2016b, 2017).

(*Eriophorum vaginatum*), monocots and Ericaceae (*Erica tetralix*), with the presence of birch shrubs growing on site, indicated by *Betula* bark.

At MDM, between 1100 and 50 cal yr BP, wet conditions are indicated by the presence of *Sphagnum* Section *Acutifolia* and *Polytrichum commune*. A *Sphagnum*-sedge-graminoid community prevails, evidenced by identified Cyperaceae (*Eriophorum vaginatum*) and Monocot leaves. Following on from this, between ca. 50 and −60 cal yr BP (cal yr AD 1910–2100), a low percentage of UOM, high percentage of Ericaceae (*Vaccinium oxycoccus* stems) and the presence of *Polytrichum commune* are suggestive of wet conditions. However, the percentage of Monocot roots is high, suggesting dry or unfavourable conditions. At OXM, between ca. 1000 and −60 cal yr BP, a low percentage of UOM and the presence of *Sphagnum* (including *S. papillosum* and *S. cuspidatum*) are suggestive of wet conditions. Monocots are also present, as is Cyperaceae (*Trichophorum cespitosum*), reaching its highest in the profile, indicating a mixed *Sphagnum* and sedge mire. At ARK, between ca. 500 and −60 cal yr BP, a mixed mire community is present including *Polytrichum commune*, some Cyperaceae (*Eriophorum vaginatum* and *E. angustifolium*), and Ericaceae (*Calluna vulgaris*). There is a high percentage of identified *Sphagnum*, including *S. papillosum* towards the surface, indicative of the presence of hummocks.

DHI

DHI (Dupont, 1986) has been applied to the plant macrofossil data to provide a qualitative indication of changes in water table (Figure 7). DCA was applied to the plant macrofossil data but was unsuccessful, perhaps owing to the very infrequent presence of certain species, or rare species, such as *Nymphaea*, occurring only once in a profile. Despite a recent favoured use of Detrended Correspondence Analysis (DCA) in studies on late-Holocene *Sphagnum*-dominated peats, DHI has some benefits over DCA and other ordination options (Daley and Barber, 2012). Unlike DCA, DHI does not require any portion of the data to be removed for it to function properly. One of the criticisms of DCA is that samples with rare species, such as is the case with the present datasets, can have a large effect on the resulting ordination (Daley and Barber, 2012). When applying DHI, the full range of data can be used. Furthermore, in DCA, the resulting axis may not be influenced by a single environmental variable. In DHI, weightings are assigned to species based on ecological knowledge of response to specific environmental variables and so despite the apparent subjectivity of the method, it is probable that DHI provides important results (Daley and Barber, 2012).

The indices used were Unidentified Organic Matter (UOM) 8, *Calluna vulgaris* 8, Ericaceae undifferentiated 8, *Betula* 7, *Polytrichum commune* 7, *Eriophorum vaginatum* 6, monocots undifferentiated 6, *Trichophorum cespitosum* 6, *Vaccinium oxycoccus* 5, *Sphagnum papillosum* 4, *Sphagnum* Section *Acutifolia* 4, *Sphagnum magellanicum* 3, *Menyanthes* 3, *Eriophorum angustifolium* 2, *Scheuchzeria palustris* 2, *Ceratophyllum* 1, *Nymphaea* 1, *Sphagnum cuspidatum* 1, based upon the weights used in Daley and Barber (2012) and Mauquoy et al. (2008) and knowledge of ecological tolerances and habitat preferences of each species.

Mossdale Moor.

Low values indicate wet conditions, implying that the mire was wet at 1300, 950 and 600–350 cal yr BP (LIA) (Figure 7), although single-sample examples must be treated with caution. The DHI curve largely agrees with the humification curve for the LIA. However, the reliability of the DHI curve is

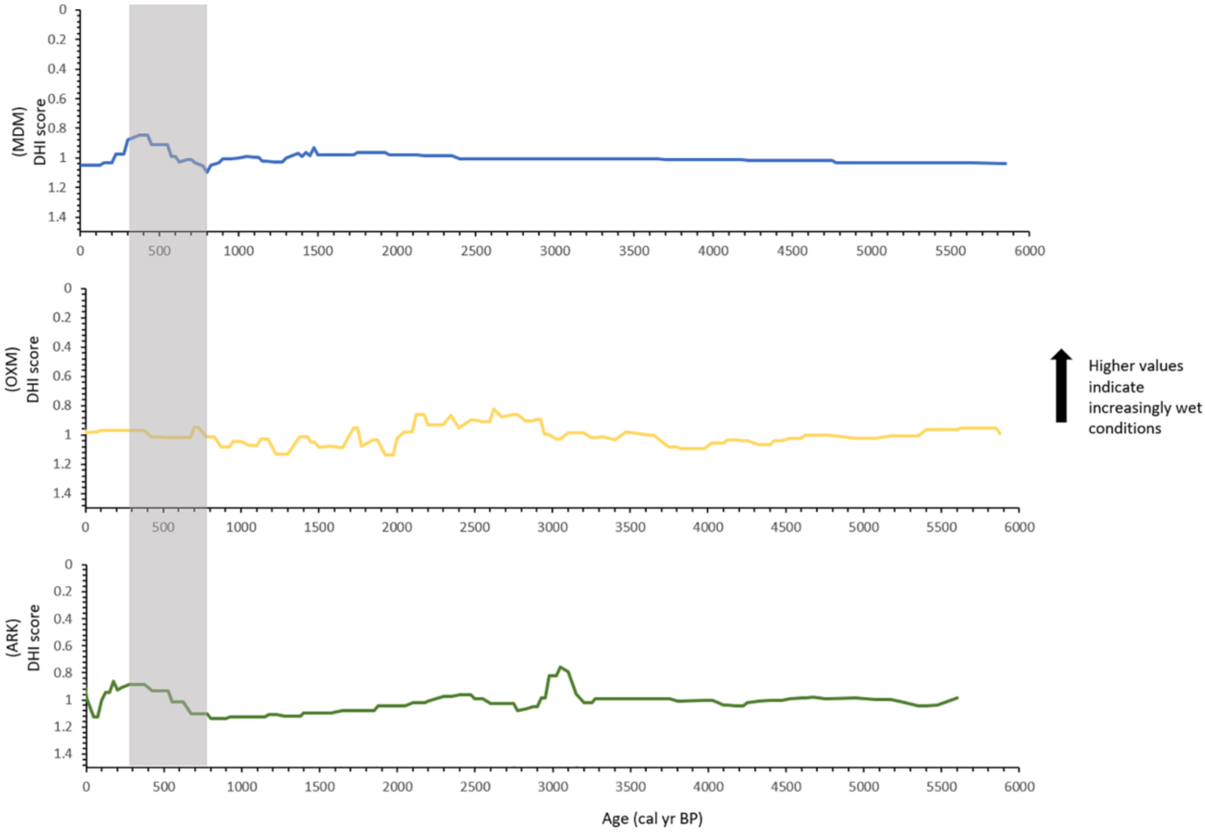

Figure 7 DHI scores plotted against age for MDM, OXM and ARK. The shaded bar represents the LIA, for visual guidance only.

problematic, given that there seems to be a disappearance of *Sphagnum* in the plant macrofossils below 65 cm depth, perhaps owing to decomposition. *Sphagnum* may have been present at the site despite the absence in the plant macrofossils, given the presence of *Sphagnum* spores counted in pollen analysis.

Oxenhope Moor.

The DHI results are plotted against age rather than depth (cp. McCarroll et al., 2016a). At 6400 cal yr BP, UOM is high, owing to higher decomposition levels at this depth in the catotelm. The high score attributed from a high percentage of UOM dilutes the wetness signal indicated by the presence of *Sphagnum cuspidatum* at this depth. Low values indicate wet conditions, suggesting that the mire was wet at 2800 cal yr BP, 1700 cal yr BP, 1350 cal yr BP, 1000 cal yr BP and between 750–300 cal yr BP (LIA).

West Arkengarthdale.

A summary plotting the DHI data against age as opposed to depth is provided here. Low values indicate wet conditions, and these suggest that the mire was wet at 3900 cal yr BP, 2600–2100 cal yr BP, 1800 cal yr BP and 450–200 cal yr BP (LIA).

Palaeoenvironmental evidence for the LIA from Yorkshire blanket peat

At Mossdale Moor, regional wet conditions are identified after 660 cal yr BP, indicated by rising T values and large numbers of *Sphagnum* and associated *Tilletia sphagni* spores, and by low levels of charcoal and *Calluna* pollen, which would be indicative of dry conditions. Upon re-examination of the data published by McCarroll et al. (2017), including DHI scores using a revised chronology, we identify a reduction in charcoal fragments and a reduction in tree and shrub pollen from pollen and spore data between 700 and 600 cal yr BP (Figure 5). A sharp increase in bog and *Sphagnum* spores is present between 650 and 600 cal yr BP, suggestive of wet conditions. The plant macrofossil data show remains of *Sphagnum*, mainly *S. capillifolium* and *S. subnitens* and low levels of UOM, suggesting wet conditions locally. The lowest degree of peat humification in the profile is shown where % T reaches its highest peak at ca. 340 cal yr BP (ca. cal yr AD 1610), suggesting the most marked of climatic deteriorations (to wet and/or cold) observed throughout the profile at MDM, concluding by 135 cal yr BP (ca. cal yr AD 1815). From ca. 700 to 250 cal yr BP (cal yr AD 1250–1700), DHI scores are lower, with the lowest DHI score throughout the profile observed at ca. 430 cal yr BP/cal yr AD 1520).

The combined data suggest a climatic deterioration between 600 and 300 cal yr BP (cal yr AD 1350–1650), consistent with the LIA, defined as occurring between cal yr AD 1200 and 1850 from the peatland palaeoenvironmental evidence from northwestern Europe.

At Oxenhope Moor, evidence for the LIA is apparent at 400 cal yr BP/cal yr AD 1550 from humification data (high T values) and at ca. 500 cal yr BP from the identification of *Sphagnum cuspidatum* from plant macrofossils. Between 45 and 40 cm depth, 100% of 35% identified *Sphagnum* is attributable to *S. cuspidatum* and pollen and spore data show low charcoal fragments, comparatively lower tree and shrub pollen than the rest of the profile, increasing bog species and *Sphagnum* spores. Low DHI values are observed and the highest T values throughout the profile are evident between 600 and 200 cal yr BP (cal yr AD 1350–1750), consistent with the LIA as defined within this paper.

Similarly, at West Arkengarthdale, ca. 450 years ago, the LIA is evidenced by low humification, low UOM and low DHI values (from 600 to 200 cal yr BP). The pollen and spore data show low tree and shrub and grass and sedge pollen and a high number of *Sphagnum* spores. From the plant macrofossils, a return of *Sphagnum* is evident at 45 cm depth (450 cal yr BP) consisting of *S. capillifolium* and *S. fuscum*. T values are at their highest between 300 and 130 cal yr BP. DHI values reduce from ca. 500 cal yr BP and are at their lowest in the profile at ca. 150 cal yr BP. All proxies agree on a climatic deterioration between 700 and 250 cal yr BP (cal yr AD 1250–1700), consistent with the LIA as defined within this paper.

Discussion

Differences in the onset and duration of the LIA in Yorkshire

The multiproxy evidence (from pollen, plant macrofossil and humification analysis) suggests that the LIA has been identified at each of the sites but with slightly differing ages (Mossdale Moor: 600– 300 cal yr BP (cal yr AD 1350–1650); Oxenhope Moor: 600–200 cal yr BP (cal yr AD 1350–1750); West Arkengarthdale: 700–250 cal yr BP (cal yr AD 1250–1700). This may be owing to differing response times of the vegetation communities at each of the sites: for instance, there is a high percentage of *Sphagnum papillosum* at Oxenhope Moor compared to *Sphagnum* Section *Acutifolia* at Mossdale Moor and West Arkengarthdale, although both have the same DHI scores and can form lawns, so it is not immediately apparent how this should delay response. Monocots and *Eriophorum vaginatum* are present at all three sites in varying amounts. *E. vaginatum* is ecologically tolerant in a range of environments (Atherton et al., 2010) and therefore perhaps has varying response times at different locations. An additional contributing factor may be variations in geographical location. For instance, West Arkengarthdale sees the earliest onset of the LIA and is located furthest north; however, it is situated at the lowest altitude of the three sites. Dating uncertainties may also play a role in the differences seen between sites, given that ^{210}Pb dates were obtained for MDM but not for OXM and ARK. A possible way to resolve this would be to obtain ^{210}Pb dates from OXM and ARK.

At Malham Tarn Moss, Yorkshire, a marked period of increased effective precipitation and a prolonged phase of increased Bog Surface Wetness (BSW) is apparent in palaeohydrological proxies (testate amoebae, plant macrofossil and humification analysis) between ca. AD 1460 and 1850 (490–100 cal yr BP) (Figure 8 and Table 2) and is synchronous with the LIA climate deterioration (Turner et al., 2014). The LIA starts earlier at the sites investigated in the present study, where the onset is identified as AD 1350 at MDM and OXM and AD 1250 at ARK. Furthermore, the duration of the LIA at Malham Tarn Moss is longer, ending AD 1850, compared to AD 1650 at MDM, AD 1750 at OXM and AD 1700 at ARK.

A testate amoebae profile and water-table reconstruction from Moor House blanket peat in Northern England shows a transition from near surface to deeper water tables at the boundary of the LIA (Swindles et al., 2015). However, Swindles et al. (2015) suggest that the magnitude of the reconstruction is too dry and therefore, these data have been excluded from Figure 8 and Table 2.

Figure 8 Geographical and temporal patterns of the LIA across northwestern Europe. See Table 2 for sites included and age ranges.

Table 2 Geographical and temporal patterns of the LIA across northwestern Europe including site names, start and end dates for the LIA, country and coordinates and authors.

Site name	LIA start date (AD)	LIA end date (AD)	Country	Latitude (ºN)	Longitude (ºE; -ve: ºW)	Authors
Bolton Fell Moss	1160	1400	England	55.012925	-2.7990421	Mauquoy and Barber (1999)
Walton Moss	1160	1400	England	54.992401	-2.7767524	Mauquoy and Barber (1999)
Coom Rigg Moss	1160	1400	England	55.108665	-2.48507	Mauquoy and Barber (1999)
Raeburn Flow	1160	1400	Scotland	55.933171	-3.1300138	Mauquoy and Barber (1999)
Felecia Moss	1160	1400	England	55.092938	-2.4386886	Mauquoy and Barber (1999)
Bells Flow	1160	1400	Scotland	55.073781	-3.0665036	Mauquoy and Barber (1999)
Slowinske Blota	1200	1800	Poland	54.399925	16.483501	De Vleeschouwer et al. (2009)
Dosenmoor	1250	1350	Germany	54.136281	10.019157	Barber et al. (2004)
Svanemorse	1250	1350	Denmark	55.22	9.3	Barber et al. (2004)
West Arkengarthdale	1250	1700	England	54.458815	-2.067067	Present paper
Walton Moss	1335	1615	England	55.012925	-2.7990421	Stoneman (1993)
Bolton Fell Moss	1335	1615	England	55.012925	-2.7990421	Stoneman (1993)
Glasson Moss	1335	1615	England	54.932438	-3.189822	Stoneman (1993)
Carsegowan Moss	1335	1615	Scotland	54.900419	-4.455654	Stoneman (1993)
Ellergower Moss	1335	1615	Scotland	55.087103	-4.384487	Stoneman (1993)
Cranley Moss	1335	1615	Scotland	55.710083	-3.699848	Stoneman (1993)
Drone Moss	1335	1615	Scotland	55.895281	-2.249653	Stoneman (1993)
Dogden Moss	1335	1615	Scotland	55.73942	-2.506115	Stoneman (1993)
Letham Moss	1335	1615	Scotland	56.0545	-3.7993760	Stoneman (1993)
Blairbech Moss	1335	1615	Scotland	56.018719	-4.5155371	Stoneman (1993)
Talkin Tarn	1350	1350	England	54.922164	-2.711438	Barber and Langdon (2007)
Tore Hill Moss	1350	1350	Scotland	53.70463	-2.378667	Charman et al. (2006)
Mallachie Moss	1350	1350	Scotland	57.234743	-3.717148	Charman et al. (2006)
Shirgarton Moss	1350	1350	Scotland	56.1392	-4.179245	Charman et al. (2006)
Killorn Moss	1350	1350	Scotland	56.137636	-4.220133	Charman et al. (2006)
Temple Hill Moss	1350	1350	Scotland	55.837639	-3.4178588	Charman et al. (2006)
Langlands Moss	1350	1350	Scotland	55.734948	-4.176072	Charman et al. (2006)
Longridge Moss	1350	1350	Scotland	55.839884	-3.6686636	Charman et al. (2006)
Butterburn Flow	1350	1350	England	55.056049	-2.509428	Charman et al. (2006)
Coom Rigg Moss	1350	1350	England	55.108665	-2.48507	Charman et al. (2006)
May Moss	1350	1350	England	54.3532	-0.6531684	Charman et al. (2006)
Traligill Basin	1350	1450	Scotland	58.148555	-4.977197	Baker et al. (1999)
May Moss	1350	1450	England	54.3532	-0.6531684	Chiverrell (2001)
Mossdale Moor	1350	1650	England	54.300292	-2.315507	Present paper
Oxenhope Moor	1350	1750	England	53.793759	-1.977952	Present paper
Coom Rigg Moss	1400	1500	England	55.108665	-2.48507	Charman et al. (1999)
Letterfrack	1410	1540	Ireland	53.550965	-9.958596	Blackford and Chambers (1995)
Talla Moss	1410	1410	Scotland	55.463916	-3.34104	Chambers et al. (1997)
Lille Vildmose	1449	1464	Denmark	56.889176	10.211426	Mauquoy et al. (2002)
Malham Tarn	1460	1850	England	54.097024	-2.168611	Turner et al. (2014)
Astley Moss	1488	1796	England	53.476002	-2.433262	Davis and Wilkinson (2004)
Danes Moss	1488	1796	England	53.237879	-2.143975	Davis and Wilkinson (2004)
Traligill Basin	1550	1800	Scotland	58.148555	-4.977197	Baker et al. (1999)
Walton Moss	1601	1604	England	54.987636	-2.778144	Mauquoy et al. (2002)
Fallahogy	1650	1850	Northern Ireland	54.345	-6.336	Barber et al. (2000)
Moine Mhor	1650	1850	Scotland	56.103303	-5.501073	Barber et al. (2000)
Coom Rigg Moss	1650	1900	England	55.108665	-2.48507	Charman et al. (1999)
Letterfrack	1660	1720	Ireland	53.550965	-9.958596	Blackford and Chambers (1995)
Temple Hill Moss	1700	1800	Scotland	55.883372	-3.1025082	Langdon et al. (2003)

In the North York Moors in northern England, Chiverrell (2001) identifies a shift to wetter or cooler climatic conditions at May Moss after cal yr AD 1400–1620, which is contemporaneous with the onset of the LIA (Figure 8, Table 2). At May Moss, this deterioration lasts from ca. AD 1350 to 1450 to ca. AD 1900. The LIA is more long-lived at May Moss, ending AD 1900, compared to AD 1650 at MDM, AD 1750 at OXM and AD 1700 at ARK. This is interesting given that May Moss and the sites in the present study are all located in Yorkshire. The differences may be owing to varying sensitivity of the sites, perhaps as May Moss is located closer to the coast, therefore making the site susceptible to cold easterly winds from the North Sea.

The LIA in northwestern Europe

At Talkin Tarn, Cumbria, wetter conditions have been identified at 600 BP by low chironomid-inferred July temperatures and wetshifts in plant macrofossils and testate amoebae (Barber and Langdon, 2007) (Figure 8, Table 2). Charman et al. (2006) also recognise 600 BP as being a period of higher water tables in northern Britain. The same deterioration is identified by Baker et al. (1999) at 600–500 BP and is claimed to correlate with Maunder and Spörer sunspot minima. The LIA is also identified by Langdon et al. (2003) from Temple Hill Moss, Pentland Hills, southeast Scotland and in southern Scotland from Talla Moss, where Chambers et al. (1997) recognise a cool and wet episode commencing at ca. 540 BP from pollen and humification analysis. Peat sequences in Denmark (Aaby, 1976; Barber et al., 2004) show wet shifts of an equivalent age and Barber et al. (2004) recognise the LIA from northern Germany and Denmark from peat macrofossil investigations. These data overlap with the age ranges proposed for the LIA in Yorkshire in the present study, with the onset identified as 600 cal yr BP for MDM and OXM although slightly earlier at ARK at 700 cal yr BP.

Davis and Wilkinson (2004) identify *Amphitrema* spp. From testate amoebae analysis at ca. 300 BP (cal yr AD 1488–1796) at Astley Moss, Greater Manchester and Danes Moss, Cheshire, northwest England, which the authors suggest likely corresponds to the LIA. Again, each of these suggested date ranges overlaps with the LIA as identified at MDM, OXM and ARK.

Charman et al. (1999) recognise prominent wet peaks at Coom Rigg Moss, Northumberland, northern England from the reconstructed water-table record using testate amoebae at cal yr AD 1400–1500 and cal yr AD 1650–1900. These wet periods are also well replicated in the plant macrofossil record at the same site. In the present study, one wet period is identified at each site for the LIA, as opposed to two as identified by Charman et al. (1999). This is perhaps owing to differences in the location of the sites, or perhaps the difference in methods used to reconstruct palaeoclimatic changes. However, Mauquoy and Barber (1999) also studied Coom Rigg Moss and only report one wet shift from this site, associated with the decline and local extinction of *Sphagnum austinii* between cal yr AD 1395 and 1485. This does overlap with the dates for the LIA from the sites in the present study. Mauquoy and Barber (1999) provide further evidence supporting the occurrence of climatic deteriorations synchronous with the LIA in northern Britain. There are wet shifts associated with the decline and local extinction of *S. austinii*, dated to cal yr AD 1160–1400 from Raeburn Flow and Bell's Flow (Scotland) and cal yr AD 1030–1400 from Bolton Fell Moss and Walton Moss (England), which incorporates the onset of the LIA. As well as the aforementioned Coom Rigg Moss, they also studied Felecia Moss, where wet shifts are associated with the decline and local extinction of *S. austinii* between cal yr AD 1395–1485, again synchronous with the earlier dates for the LIA.

Stoneman (1993) found evidence from northern England and southern Scotland from 10 raised peat bogs (Walton, Bolton Fell, Glasson, Carsegowan, Ellergower, Cranley, Dogden, Drone, Letham and Blairbech Mosses) for a climatic deterioration at ca. cal AD 1335–1615 from plant macrofossil and humification data. These dates closely overlap with the dates proposed in the present study, with the LIA beginning at ca. AD 1350 for MDM and OXM and AD 1250 for ARK, ending AD 1650 at MDM, AD 1750 at OXM and AD 1700 at ARK. It would be useful to revisit each of the sites studied by Stoneman (1993) and the sites in the present study to understand whether testate amoebae water table depth reconstructions would confirm or deny these propositions, particularly given the age ranges proposed by Charman et al. (1999) for Coom Rigg Moss, where testate amoebae were used and how these differed from the ages proposed by Mauquoy and Barber (1999) for the same site, without testate amoebae analysis. Furthermore, many of these studies were conducted in the 1990s and early 2000s; revisiting these datasets with advances in dating may refine results. It may also be the case that sites at higher altitude respond earlier, owing to decreasing temperature with height.

The global record of the LIA

Increasing evidence allows the LIA to be regarded as a global event, with differing responses dependant on location. For example, Chambers et al. (2014) argue that the most severe episodes of the LIA are contemporaneous between hemispheres but with opposite site responses at a mire in Tierra del Fuego in the Southern Hemisphere when compared to those in a mire in north-central Europe in the Northern Hemisphere.

The LIA is also identified in Australasia; for example, Winkler (2000) identifies a LIA glacier maximum at Mueller Glacier and Yan et al. (2015) document a retreat of the East Asian Summer Monsoon and the Australian Summer Monsoon into the tropics. In south central Alaska, Barlow et al. (2012) identify glacial advance; yet, Fritz et al. (1994) provide evidence for aridity and drought during the LIA from Devils Lake, North Dakota, USA. Dry intervals have also been suggested by Platt

Bradbury (1988) from Elk Lake, Minnesota, again suggesting that for this part of the globe, an ‘ice age’ does not appropriately describe the climatic event.

Drought is identified in Québec, Canada by Archambault and Bergeron (1992) but despite this, Bégin and Payette (1988) show that snowfall increased towards the end of the LIA in the same province; Petersen (1994) notes that the LIA is cold and dry in the southern Rocky Mountains, U.S.A., differing from the wet and cold conditions observed in northern England and Europe. Regardless of the climatic variation of the LIA across the globe, the identification of this event at this scale is evidence for its severity.

Evidence for other climatic deteriorations in this study

Other climatic deteriorations identified in the present study and from the wider literature are indicated in Table 3. These events were identified in two of the sites from the present study as well as by other authors. None of these events is present at all sites in the present study, none is found routinely in other studies, and therefore, perhaps none of these events can be regarded as pronounced as the LIA.

Table 3 Evidence for other climatic deteriorations in terms of wetness/coldness.

Climatic deterioration (time period)	**Sites**	**Authors**	**Evidence**
3200-2100 cal yr BP	Mossdale Moor, Oxenhope Moor, Bolton Fell Moss, Butterburn Flow, Walton Moss, Talkin Tarn, Bigland Tarn (northern Britain), Mongan Bog, Abbeyknockmoy Bog (Ireland), Temple Hill Moss (southeast Scotland), North Atlantic Ocean core, ^{14}C calibration curve.	Present paper, Barber et al. (2003, 2013), Langdon et al. (2003), Bond et al. (1997), van Geel et al. (1996), van Geel and Renssen (1998).	MDM: *Sphagnum* spores present, low humification. OXM: *Tilletia sphagni* spores present, low humification, low DHI values, low *Calluna* pollen, low charcoal.
4400 cal yr BP	Oxenhope Moor, West Arkengarthdale, Walton Moss (northern Britain), Temple Hill Moss (southeast Scotland).	Present paper, Hughes et al. (2000), Langdon et al. (2003).	OXM: *Sphagnum* spores present, low *Calluna* pollen, low humification. ARK: low humification, low charcoal, low *Calluna* pollen, present of *Scheuchzeria palustris* in plant macrofossils.
5000 cal yr BP	Oxenhope Moor, West Arkengarthdale, Walton Moss, Talkin Tarn (northern Britain), Eilean Subhainn, Glen Tarridon, Glen Carron (northwest Scotland).	Present paper, Barber and Langdon (2007), Anderson et al. (1998).	OXM: *Sphagnum* spores present, low humification, low charcoal. ARK: Low humification, presence of *Equisetum fluviatile* in plant macrofossils.

The widely discussed and globally identified 4.2 ka event (Bond et al., 2001; Booth et al., 2005; Drysdale et al., 2006; Liu and Feng, 2012; Thompson et al., 2002) has not been identified in the present study, with the possible exception of at ARK, where *S. cuspidatum* is seen to increase around this time; however, this is not enough evidence alone to suggest the presence of this climatic event. The identification of this event is somewhat contested in northwest European peatlands, particularly in the UK (Roland et al., 2014). Roland et al. (2014) attribute this to the possibility that either the dominant forcing mechanisms of this period lie outside of the North Atlantic, or any changes in the region may not have been severe enough to be recorded in peatlands. This is a possible explanation for the absence of this event from the climatic changes identified in the present study.

The significance of the LIA

While there is evidence for other climatic deteriorations recorded in the data from the sites in the present paper, the LIA is the only such event recorded by the proxy-climate data from all three sites. It is the most widely recorded climatic instability in many other peatlands and other records of palaeoclimate across northern England and continental Europe, suggesting that the LIA is the most noticeable palaeoclimate downturn here in the latter half of the postglacial. The humification data from MDM, OXM and ARK show the most pronounced shift towards wetter conditions in the entirety of the ca. 6000-year records at ca. 500 cal yr BP. Similarly, the DHI records for MDM and ARK show that the harshest climatic conditions in terms of wetness/coldness were present during this time. In particular, there was a vegetation shift at OXM from *Sphagnum papillosum* dominated peat before the event to wet-loving *S. cuspidatum* dominance during the LIA. The dominance then reverts to *S. papillosum* following the end of the LIA. A comparable shift in *Sphagnum* dominance is not present in the rest of the profile. A climatic cause for this vegetation shift is also supported by a decrease in humification values and a signal for a wider than local climatic deterioration in the pollen signal. The palaeoclimatic changes observed throughout the rest of the profiles at each site appear more modest in comparison to the apparent wetness of the LIA. This is perhaps the most pronounced downturn in climate observed in the northern England terrestrial record, as potentially in Europe.

Conclusions

The Little Ice Age has been identified in the form of a wet phase from three blanket peats in Yorkshire using pollen, plant macrofossil and humification data, providing new palaeoclimate data for this area of Britain. Although other climatic events are apparent at each site, this is the only event clearly evident in all three in the last ca. 6000 years. The data are insufficiently detailed to identify multiple phases within the LIA clearly. Slightly differing ages are apparent for the LIA between sites, although this is in keeping with differences between other sites in the literature and may be attributed either to the limitations in radiocarbon dating spanning across this time or to the individual sensitivity of each site to changes in climate. The LIA has previously been identified in published literature for northern England and northwest Europe from terrestrial records and from various other sources across the globe, highlighting the severity of this event. The present paper adds further evidence in support of this. It is perhaps the most widespread and routinely recorded climatic event evident in peatlands in northern England over the mid- to late-Holocene.

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Supplemental material

Supplemental material for this article is available online.

References

Aaby B (1976) Cyclic climatic variations in climate over the past 5,500 yr reflected in raised bogs. *Nature* 263: 281–284.

Amesbury MJ, Barber KE and Hughes PDM (2012a) Can rapidly accumulating Holocene peat profiles provide sub-decadal resolution proxy climate data? *Journal of Quaternary Science* 27(8): 757–770.

Amesbury MJ, Barber KE and Hughes PD (2012b) The relationship of fine-resolution, multi-proxy palaeoclimate records to meteorological data at Fågelmossen, Värmland, Sweden and the implications for the debate on climate drivers of the peat-based record. *Quaternary International* 268: 77–86.

Anderson DE, Binney HA and Smith MA (1998) Evidence for abrupt climatic change in northern Scotland between 3900 and 3500 calendar years BP. *The Holocene* 8(1): 97–103.

Appleby PG (2001) Chronostratigraphic techniques in recent sediments. In: Last WM and Smol JP (eds) *Tracking Environmental Change Using Lake Sediments Vol. 1: Basin Analysis, Coring, and Chronological Techniques*. Dordrecht: Kluwer, pp. 171–203.

Appleby PG (2008) Three decades of dating recent sediments by fallout radionuclides: a review. *The Holocene* 18: 83–93.

Appleby PG and Oldfield F (1978) The calculation of lead-210 dates assuming a constant rate of supply of unsupported ^{210}Pb to the sediment. *CATENA* 5(1): 1–8.

Archambault S and Bergeron Y (1992) An 802-year tree-ring chronology from the Quebec boreal forest. *Canadian Journal of Forest Research* 22(5): 674–682.

Atherton I, Bosanquet S and Lawley M (2010) *Mosses and Liverworts of Britain and Ireland: A Field Guide*. London: British Bryological Society.

Baker A, Caseldine CJ, Gilmour MA et al. (1999) Stalagmite luminescence and peat humification records of palaeomoisture for the last 2500 years. *Earth and Planetary Science Letters* 165(1): 157–162.

Barber K, Brown A, Langdon P et al. (2013) Comparing and cross-validating lake and bog palaeoclimatic records: A review and a new 5,000 year chironomid-inferred temperature record from northern England. *Journal of Paleolimnology* 49(3): 497–512.

Barber K, Chambers F and Maddy D (2004) Late Holocene climatic history of northern Germany and Denmark: Peat macrofossil investigations at Dosenmoor, Schleswig-Holstein, and Svanemose, Jutland. *Boreas* 33(2): 132–144.

Barber KE (1981) *Peat Stratigraphy and Climatic Change: A Palaeoecological Test of the Theory of Cyclic Bog Regeneration*. Rotterdam: Balkema.

Barber KE, Chambers FM, Maddy D et al. (1994) A sensitive high-resolution record of late holocene climatic change from a raised bog in northern England. *The Holocene* 4(2): 198–205.

Barber KE, Chambers FM and Maddy D (2003) Holocene palaeoclimates from peat stratigraphy: Macrofossil proxy climate records from three oceanic raised bogs in England and Ireland. *Quaternary Science Reviews* 22(5–7): 521–539.

Barber KE and Langdon PG (2007) What drives the peat-based palaeoclimate record? A critical test using multi-proxy climate records from northern Britain. *Quaternary Science Reviews* 26(25–28): 3318–3327.

Barber KE, Maddy D, Rose N et al. (2000) Replicated proxyclimate signals over the last 2000 yr from two distant UK peat bogs: New evidence for regional palaeoclimate teleconnections. *Quaternary Science Reviews* 19(6): 481–487.

Barlow NLM, Shennan I and Long AJ (2012) Relative sea-level response to Little Ice Age ice mass change in south central Alaska: Reconciling model predictions and geological evidence. *Earth and Planetary Science Letters* 315–316: 62–75.

Bégin Y and Payette S (1988) Dendroecological evidence of lake-level changes during the last three centuries in subarctic Québec. *Quaternary Research* 30(2): 210–220.

Bingham EM, McClymont EL, Väliranta M et al. (2010) Conservative composition of n-alkane biomarkers in Sphagnum species: Implications for palaeoclimate reconstruction in ombrotrophic peat bogs. *Organic Geochemistry* 41(2): 214–220.

Blackford J (2000) Palaeoclimatic records from peat bogs. *Trends in Ecology & Evolution* 15(5): 193–198.

Blackford JJ and Chambers FM (1991) Proxy records of climate from blanket mires: Evidence for a Dark Age (1400 BP) climatic deterioration in the British Isles. *The Holocene* 1(1): 63–67.

Blackford JJ and Chambers FM (1995) Proxy climate record for the last 1000 years from Irish blanket peat and a possible link to solar variability. *Earth and Planetary Science Letters* 133(1–2): 145–150.

Blundell A and Barber K (2005) A 2800-year palaeoclimatic record from Tore Hill Moss, Strathspey, Scotland: The need for a multi-proxy approach to peat-based climate reconstructions. *Quaternary Science Reviews* 24(10–11): 1261–1277.

Blundell A and Holden J (2015) Using palaeoecology to support blanket peatland management. *Ecological Indicators* 49: 110–120.

Bond G, Kromer B, Beer J et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294: 2130–2136.

Bond G, Showers W, Cheseby M et al. (1997) A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates. *Science* 278: 1257–1266.

Booth RK, Jackson ST, Forman SL et al. (2005) A severe centennial-scale drought in midcontinental North America 4200 years ago and apparent global linkages. *The Holocene* 15: 321–328.

Borgmark A and Schoning K (2006) A comparative study of peat proxies from two eastern central Swedish bogs and their relation to meteorological data. *Journal of Quaternary Science* 21(2): 109–114.

Castro D, Souto M, Garcia-Rodeja E et al. (2015) Climate change records between the mid- and late Holocene in a peat bog from Serra do Xistral (SW Europe) using plant macrofossils and peat humification analyses. *Palaeogeography Palaeoclimatology Palaeoecology* 420: 82–95.

Chambers FM (1984) *Studies on the initiation, growth rate and humification of blanket peats in South Wales*. PhD Thesis, University of Keele, Keele.

Chambers FM (2006) *ACCROTELM: Abrupt Climate Changes Recorded Over the European Land Mass. Final Report to European Commission*, Contract no. EVK2-CT-2002-00166. Cheltenham: CECQR, p.117.

Chambers FM, Beilman DW and Yu Z (2011a) Methods for determining peat humification and for quantifying peat bulk density, organic matter and carbon content for palaeostudies of climate and peatland carbon dynamics. *Mires and Peat* 7(7): 1–10.

Chambers FM, van Geel B and van der Linden M (2011b) Considerations for the preparation of peat samples for palynology, and for the counting of pollen and non-pollen palynomorphs. *Mires and Peat* 7(11): 1–14.

Chambers FM (2016) The ‘Little Ice Age’: The first virtual issue of the Holocene. *The Holocene* 26(3): 335–337.

Chambers FM, Barber KE, Maddy D et al. (1997) A 5500-year proxy-climate and vegetation record from blanket mire at Talla Moss, Borders, Scotland. *The Holocene* 7(4): 391–399.

Chambers FM, Brain SA, Mauquoy D et al. (2014) The ‘Little Ice Age’ in the southern Hemisphere in the context of the last 3000 years: Peat-based proxy-climate data from Tierra del Fuego. *The Holocene* 24(12): 1649–1656.

Charman DJ (2010) Centennial climate variability in the British Isles during the mid–late Holocene. *Quaternary Science Reviews* 29(13–14): 1539–1554.

Charman DJ, Blundell A, Chiverrell RC et al. (2006) Compilation of non-annually resolved Holocene proxy climate records: Stacked Holocene peatland palaeo-water table reconstructions from northern Britain. *Quaternary Science Reviews* 25(3–4): 336–350.

Charman DJ, Hendon D and Packman S (1999) Multiproxy surface wetness records from replicate cores on an ombrotrophic mire: Implications for Holocene palaeoclimate records. *Journal of Quaternary Science* 14(5): 451–463.

Chiverrell RC (2001) A proxy record of late Holocene climate change from May Moss, northeast England. *Journal of Quaternary Science* 16(1): 9–29.

Daley TJ and Barber KE (2012) Multi-proxy Holocene palaeoclimate records from Walton Moss, northern England and Dosenmoor, northern Germany, assessed using three statistical approaches. *Quaternary International* 268: 111–127.

Daniels RE and Eddy A (1985) *Handbook of European Sphagna*. Abbots Ripton, Huntingdon: Institute of Terrestrial Ecology.

Davis SR and Wilkinson DM (2004) The conservation management value of testate amoebae as ‘restoration’ ‘indicators: Speculations based on two damaged raised mires in northwest England. *The Holocene* 14(1): 135–143.

De Vleeschouwer F, Chambers FM and Swindles GT (2010) Coring and sub-sampling of peatlands for palaeoenvironmental research. *Mires and Peat* 7(1): 1–10.

De Vleeschouwer F, Piotrowska N, Sikorski J et al. (2009) Multiproxy evidence of 'Little Ice Age’ palaeoenvironmental changes in a peat bog from northern Poland. *The Holocene* 19(4): 625–637.

Drysdale R, Zanchetta G, Hellstrom J et al. (2006) Late Holocene drought responsible for the collapse of Old World civilizations is recorded in an Italian cave flowstone. *Geology* 34: 101–104.

Dupont LM (1986) Temperature and rainfall variation in the Holocene based on comparative palaeoecology and isotope geology of a hummock and a hollow (Bourtangerveen, The Netherlands). *Review of Palaeobotany and Palynology* 48(1–3): 71–159.

Fagan B (2001) *The Little Ice Age: How Climate Made History 1300–1850*. New York, NY: Basic Books.

Fritz SC, Engstrom DR and Haskell BJ (1994) 'Little ice age' aridity in the North American Great Plains: A high-resolution reconstruction of salinity fluctuations from Devils Lake, North Dakota, USA. *The Holocene* 4(1): 69–73.

Grove JM (2001) The initiation of the "Little Ice Age" in regions round the North Atlantic. In: Ogilvie EJ and Jónsson T (eds) *The Iceberg in the Mist: Northern Research in Pursuit of a "Little Ice Age."* New York, NY: Springer, pp.53–82.

Hughes PD, Mallon G, Essex HJ et al. (2012) The use of k-values to examine plant 'species signals' in a peat humification record from Newfoundland. *Quaternary International* 268: 156–165.

Hughes PDM, Mauquoy D, Barber KE et al. (2000) Mire-development pathways and palaeoclimatic records from a full Holocene peat archive at Walton Moss, Cumbria, England. *The Holocene* 10(4): 465–479.

Langdon PG, Barber KE and Hughes PD (2003) A 7500-year peat-based palaeoclimatic reconstruction and evidence for an 1100-year cyclicity in bog surface wetness from Temple Hill Moss, Pentland Hills, southeast Scotland. *Quaternary Science Reviews* 22(2–4): 259–274.

Liu F and Feng Z (2012) A dramatic climatic transition at 4000 cal. yr BP and its cultural responses in Chinese cultural domains. *The Holocene* 22: 1181–1197.

Mann ME (2002) Little Ice Age. In: MacCracken MC and Perry JS (eds) *Encyclopedia of Global Environmental Change Vol. 1*. Chichester: John Wiley, pp. 504–509.

Mann ME, Zhang Z, Rutherford S et al. (2009) Global signatures and dynamical origins of the little ice age and medieval climate anomaly. *Science* 326(5957): 1256–1260.

Matthews JA and Briffa KR (2005) The 'Little Ice Age': Reevaluation of an evolving concept. *Geografiska Annaler: Series A, Physical Geography* 87(1): 17–36.

Mauquoy D and Barber K (1999) Evidence for climatic deteriorations associated with the decline of *Sphagnum imbricatum* Hornsch. ex Russ. in six ombrotrophic mires from northern England and the Scottish Borders. *The Holocene* 9(4): 423– 437.

Mauquoy D, van Geel B, Blaauw M et al. (2002) Evidence from northwest European bogs shows 'Little Ice Age' climatic changes driven by variations in solar activity. *The Holocene* 12(1): 1–6.

Mauquoy D, van Geel B, Blaauw M et al. (2004) Changes in solar activity and holocene climatic shifts derived from 14C wiggle-match dated peat deposits. *The Holocene* 14(1): 45–52.

Mauquoy D, Yeloff D, van Geel B et al. (2008) Two decadally resolved records from north-west European peat bogs show rapid climate changes associated with solar variability during the mid-late Holocene. *Journal of Quaternary Science* 23(8): 745–763.

McCarroll J, Chambers FM, Webb JC et al. (2016a) Informing innovative peatland conservation in light of palaeoecological evidence for the demise of Sphagnum imbricatum: The case of Oxenhope Moor, Yorkshire, UK. *Mires and Peat* 18(8): 1–24.

McCarroll J, Chambers FM, Webb JC et al. (2016b) Using palaeoecology to advise peatland conservation: An example from West Arkengarthdale, Yorkshire, UK. *Journal for Nature Conservation* 30: 90–102.

McCarroll J, Chambers FM, Webb JC et al. (2017) Application of palaeoecology for peatland conservation at Mossdale Moor, UK. *Quaternary International* 432(Part A): 39–47.

McCracken KG and Beer J (2014) Comparison of the extended solar minimum of 2006-2009 with the Spoerer, Maunder, and Dalton Grand Minima in solar activity in the past. *Journal of Geophysical Research Space Physics* 119(4): 2379–2387.

Miller GH, Geirsdóttir Á, Zhong Y et al. (2012) Abrupt onset of the Little Ice Age triggered by volcanism and sustained by sea-ice/ocean feedbacks. *Geophysical Research Letters* 39(2): n/a–n/a.

Moore PD, Webb JA and Collinson ME (1991) *Pollen Analysis*. Oxford: Blackwell.

Petersen KL (1994) A warm and wet little climatic optimum and a cold and dry little ice age in the southern Rocky Mountains, USA. In: Hughes MK and Diaz HF (eds) *The Medieval Warm Period*. New York, NY: Springer, pp.243–269.

Platt Bradbury J (1988) A climatic-limnologic model of diatom succession for paleolimnological interpretation of varved sediments at Elk Lake, Minnesota. *Journal of Paleolimnology* 1(2): 115–131.

Ramsey CB (2009) Bayesian analysis of radiocarbon dates. *Radiocarbon* 51(1): 337–360.

Reimer PJ, Bard E, Bayliss A et al. (2013) IntCal13 and marine13 radiocarbon age calibration curves 0–50,000 years cal BP. *Radiocarbon* 55(4): 1869–1887.

Rodwell JS (1998) *British Plant Communities, 2*. Cambridge: Cambridge University Press.

Roland TP, Caseldine CJ, Charman DJ et al. (2014) Was there a ‘4.2 ka event ‘in Great Britain and Ireland? Evidence from the peatland record. *Quaternary Science Reviews* 83: 11–27.

Ruddiman WF (2003) The Anthropogenic greenhouse era began thousands of years ago. *Climatic Change* 61(3): 261–293.

Sillasoo Mauquoy D, Blundell A et al. (2007) Peat multi-proxy data from Männikjärve bog as indicators of late Holocene climate changes in Estonia. *Boreas* 36(1): 20–37.

Smith AJE (2004) *The Moss Flora of Britain and Ireland*. Cambridge: Cambridge University Press.

Stoneman R (1993) *Holocene palaeoclimates from peat stratigraphy: extending and refining the model*. PhD Thesis, University of Southampton, Southampton.

Swindles GT (2010) Dating recent peat profiles using spheroidal carbonaceous particles (SCPs). *Mires and Peat* 7(3): 1–5.

Swindles GT, Holden J, Raby CL et al. (2015) Testing peatland water-table depth transfer functions using high-resolution hydrological monitoring data. *Quaternary Science Reviews* 120(lement C): 107–117.

Tallis JH (1995) Climate and erosion signals in British blanket peats: The significance of Racomitrium lanuginosum remains. *Journal of Ecology* 83(6): 1021–1030.

Thompson LG, Mosley-Thompson E, Davis ME et al. (2002) Kilimanjaro ice core records: Evidence of Holocene climate change in tropical Africa. *Science* 298: 589–593.

Troels-Smith J (1955) Characterisation of unconsolidated sediments. *Danmarks Geologiske Undersøgelse Series* 4(3): 39–73.

Turner TE, Swindles GT, Charman DJ et al. (2016) Solar cycles or random processes? Evaluating solar variability in Holocene climate records. *Scientific Reports* 6: 23961.

Turner TE, Swindles GT and Roucoux KH (2014) Late Holocene ecohydrological and carbon dynamics of a UK raised bog: Impact of human activity and climate change. *Quaternary Science Reviews* 84: 65–85.

van der Plicht J, Yeloff D, van der Linden M et al. (2013) Dating recent peat accumulation in European ombrotrophic bogs. *Radiocarbon* 55(3): 1763–1778.

van Geel B (1978) A palaeoecological study of Holocene peat bog sections in Germany and the Netherlands, based on the analysis of pollen, spores and macro- and microscopic remains of fungi, algae, cormophytes and animals. *Review of Palaeobotany and Palynology* 25(1): 1–120.

van Geel B and Renssen H (1998) Abrupt climate change around 2,650 BP in North-West Europe: evidence for climatic teleconnections and a tentative explanation. In: Brown N and Issar AS (eds) *Water, Environment and Society in Times of Climatic Change*. Dordrecht: Kluwer, pp. 21–41.

van Geel B, Buurman J and Waterbolk HT (1996) Archaeological and palaeoecological indications of an abrupt climate change in The Netherlands, and evidence for climatalogical teleconnections around 2650 BP. *Journal of Quaternary Science* 11: 451–460.

Winkler S (2000) The ‘Little Ice Age’ ‘maximum in the southern Alps, New Zealand: Preliminary results at Mueller Glacier. *The Holocene* 10(5): 643–647.

Xie S, Nott CJ, Avsejs LA et al. (2004) Molecular and isotopic stratigraphy in an ombrotrophic mire for paleoclimate reconstruction. *Geochimica et Cosmochimica Acta* 68(13): 2849–2862.

Yan H, Wei W, Soon W et al. (2015) Dynamics of the intertropical convergence zone over the western Pacific during the Little Ice Age. *Nature Geoscience* 8(4): 315–320.

Yeloff D and Mauquoy D (2006) The influence of vegetation composition on peat humification: Implications for palaeoclimatic studies. *Boreas* 35(4): 662–673.

Zaborska A, Carroll J, Papucci C et al. (2007) Intercomparison of alpha and gamma spectroscopy techniques used in ^{210}Pb geochronology. *Journal of Environmental Radioactivity* 93: 38–50.