The 5.2 ka climate event: evidence from stable isotope and multi-proxy palaeoecological peatland records in Ireland

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Highlights:

- The 5.2 ka event has been identified globally as a period of abrupt climate change;
- Multiple stable isotope records from Ireland show clear evidence for 5.2 ka event;
- Sluggan Moss possesses a strong palaeoecological evidence of 5.2 ka event;
- 5.2 ka event caused by prolonged positive North Atlantic Oscillation conditions;
- The value of research into peat-based stable isotopes is highlighted.

Keywords: Peatlands; multi-proxy; stable isotopes; testate amoebae; plant macrofossils; Ireland; 5.2 ka event; mid-Holocene; North Atlantic Oscillation (NAO).

Abstract

Evidence for a major climate event at 5.2 ka has been reported globally and is associated with considerable societal disruption, but is poorly characterised in northwest Europe. This event forms part of a broader period of re-organisation in the Earth's ocean-atmosphere circulation system between 6 – 5 ka. This study tests the nature and timing of the event in northwest Europe, a region highly sensitive to change in meridional overturning circulation and mid-latitude westerly airflow. Here we report three high-resolution Irish multi-proxy records obtained from ombrotrophic peatlands that have robust chronological frameworks. We identify the 5.2 ka event by a sustained decrease in $\delta^{18}O_{cellulose}$ at all three sites, with additional and parallel changes in $\delta^{13}C_{cellulose}$ and palaeoecological (testate amoebae, plant macrofossil and humification) data from two sites in northern Ireland. Data from Sluggan Moss demonstrate a particularly coherent shift towards wetter conditions. These data support the hypothesis that the event was caused by a prolonged period of positive North Atlantic Oscillation conditions, resulting in pervasive cyclonic weather patterns across northwest Europe, increasing precipitation over Ireland.

1. Introduction

1.1 A mid-Holocene climatic transition

The occurrence of a substantial transition in the global climate system during the period 6 – 5 ka is widely acknowledged (Steig, 1999; Mayewski et al., 2004; Wanner et al., 2008; Brooks, 2012). This transition marked the termination of the Holocene thermal maximum (HTM), a relatively warm period with temperatures markedly higher than those of the pre-industrial era, as recorded in a range of palaeoclimate archives (e.g. Davis et al., 2003; Kaufman et al., 2004; Jansen et al., 2009; Seppä et al., 2009; Bartlein et al., 2011). Forcing of the HTM is commonly attributed to the orbitally-driven summer insolation maximum in the Northern Hemisphere (NH) (Wanner et al., 2008; Bartlein et al., 2011), with its complex spatio-temporal structure explained by the influence of additional forcing mechanisms and feedbacks, including the decay of the remnant Laurentide ice sheet (LIS) (Renssen et al., 2009, 2012).

Whilst NH summer insolation decreased gradually from the early Holocene onwards, the steepest decline occurred c. 6 ka (Steig, 1999) associated with a decrease in ^{14}C and ^{10}Be residuals, indicating reduced solar activity, which continued until c. 5.1 ka (Finkel and Nishiizumi, 1997; Stuiver et al., 1998). These changes coincided with a global trend of glacial advance (Denton and Karlén, 1973; Hodell et al., 2001; Nesje et al., 2001; Mayewski et al., 2004; Kilian and Lamy, 2012), a major increase in ice-rafted debris in the North Atlantic (Bond event 4) (Bond et al., 2001; Oppo et al., 2003) and South Atlantic (Hodell et al., 2001) and register a strong signal in the glaciochemical proxies of the GISP2 ice core (Mayewski et al., 1997), all of which have been linked with a more positive North Atlantic Oscillation (NAO), and enhanced westerlies across the North Atlantic (Mayewski et al., 2004).

An extensive review of potentially correlative short-lived, multi-centennial climatic event signals recorded in palaeoclimate records from both hemispheres found that the majority occurred 5.6 – 5 ka (Magny et al., 2006 and references therein). Thirty-four of the records examined by Magny et al. (2006) have ages defining the onset of the event, of which the average was 5.23 ka and so the event will be referred henceforth as the '5.2 ka event'. Many of these signals are consistent with the 'cool poles, dry tropics' pattern typical of a number of climate episodes which punctuate the mid- to late-Holocene, including the 4.2 ka, 2.8 ka and Little Ice Age events (Mayewski et al., 2004). During the 5.2 ka event, widespread cooling was accompanied by drier conditions in central and eastern Asia, Africa, the Mediterranean and parts of North America, with wetter conditions in northern Europe and southern South America (Magny et al., 2006), demonstrating that the dynamic processes associated with the event extend beyond the influence of the NAO.

The abrupt termination of the African humid period c. 5.5 ka, following a weakening of the African monsoonal system, was rapid, occurring within several decades to centuries and provides a striking example of a non-linear response to gradual insolation forcing (Demenocal et al., 2000; Kröpelin et al., 2008). A trend towards drier conditions in South America, as recorded in the Cariaco Basin marine sediments, also began c. 5.4 ka (Haug et al., 2001), consistent with numerous other low-latitude records which show a similar drying trend at this time (Magny et al., 2006). This trend suggests a southward migration of the Intertropical Convergence Zone (ITCZ) and is consistent with many other low-latitude records which show a drying trend at this time (Magny et al., 2006).

The period 6 – 5 ka also witnessed the onset of the 'modern' El Niño Southern Oscillation (ENSO) (Sandweiss et al., 1996, 2001, 2007; Moy et al., 2002) and, importantly, the emergence of environmental boundary conditions similar to those of the present day following the final deglaciation of the LIS (Renssen et al., 2012), which played a significant role in early-Holocene climatic events, particularly in the North Atlantic region (Clark et al., 2001). Whilst the precise nature of the mechanisms that drove this ocean-atmosphere variability remains uncertain, it was likely to have been a complex response to variations in solar activity and orbitally-driven insolation changes (Hodell et al., 2001; Magny and Haas, 2004; Mayewski et al., 2004; Magny et al., 2006; Wanner et al., 2008), further complicated by non-linear feedback processes, teleconnections between and thresholds within the climate system components (e.g. NAO, ITCZ, ENSO) discussed here (e.g. Schneider, 2004; Broecker, 2006; Wunsch, 2006; Holmes et al., 2011). As a result, the 5.2 ka and subsequent late-Holocene events have considerable potential for providing 'process analogues' to understand future change (Alley et al., 2003; Broecker, 2006).

Wide scale mid-Holocene aridity across the (sub)tropics, particularly during the 5.2 ka event, has been linked with the abandonment of nomadic lifestyles and the rapid development of the world's first civilizations of large, complex, highly-urbanised, hierarchical and organised societies forming in response to drought and over-population in Egypt, north-central China, northern coastal Peru, the Indus valley, Mesopotamia and more broadly across western Asia (Sirocko et al., 1993; Sandweiss et al., 2001; Brooks, 2006, 2012; Staubwasser and Weiss, 2006). In Europe, cultural development and changes in settlement patterns have also been linked to the event (Berglund, 2003; Magny, 2004; Arbogast et al., 2006). In particular, considerable disruption to marginal Neolithic communities is recorded across Ireland, potentially owing to climatic deterioration, increased storm frequency and a subsequent abandonment of agricultural land (O'Connell and Molloy, 2001; Baillie and Brown, 2002; Caseldine et al., 2005; Turney et al., 2006; Verrill and Tipping, 2010; Ghilardi and O'Connell, 2013). However, despite being an historical focus for palaeoclimatic research, the 5.2 ka event is poorly characterised in northwest Europe. Regional climatic evidence for the Little Ice Age (Mauquoy et al., 2002), 2.8 – 2.6 ka event (Plunkett and Swindles, 2008) and 4.2 ka (Roland et al., 2014) events has been evaluated but an equivalent study for the 5.2 ka event has not been undertaken.

1.2 Stable isotope analysis in peatlands

Stable isotopic analysis of Holocene peat sequences provide a technique for palaeoclimatic and palaeohydrological reconstruction (e.g. Daley et al., 2009, 2010; Loisel et al., 2010). Peatland vascular and non-vascular plants possess significantly different isotopic ratios (Ménot and Burns, 2001; Ménot-Combes et al., 2002; Pancost et al., 2003; Loader et al., 2007; Moschen et al., 2009; Nichols et al., 2010; Stebich et al., 2011) and stable isotopic analysis of bulk peat (e.g. Cristea et al., 2014; Jones et al., 2014) and cellulose extracted from bulk peat (e.g. Aucour et al., 1996; El Bilali and Patterson, 2012; Hong et al., 2000; Jędrysek and Skrzypek, 2005) therefore risk being affected by botanical variation. *Sphagnum* mosses are more suited to stable isotopic analysis as they have relatively simple biomechanical pathways leading to cellulose synthesis, compared to vascular plants (Menot-Combes et al., 2002; Zanazzi and Mora, 2005; Loader et al., 2007; Daley et al., 2010).

Sphagna have no stomata and are unable to physiologically regulate uptake of atmospheric CO_2 with varying saturation of the hyaline cells providing the only barrier to CO_2 assimilation (Ménot and Burns, 2001). Consequently, stable carbon isotope fractionation in *Sphagnum*, and therefore the ratio of cellulose stable carbon isotopes ($\delta^{13}C_{cellulose}$), is heavily dependent on water availability with lower $\delta^{13}C_{cellulose}$ values associated with drier conditions and *vice versa*. Correlations between *Sphagnum* $\delta^{13}C_{cellulose}$ and modern surface moisture gradients (Price et al., 1997; Ménot and Burns, 2001; Ménot-Combes et al., 2004; Loisel et al., 2009) and independent palaeohydrological proxy records (Lamentowicz et al., 2008; Loisel et al., 2010; van der Knaap et al., 2011) support this, although discrepancies exist (e.g. Markel et al., 2010) and a small number of studies have found a relationship between *Sphagnum* $\delta^{13}C_{cellulose}$ and temperature (Skrzypek et al., 2007; Kaislahti Tillman et al., 2010; Holzkämper et al., 2012).

In the absence of stomata and vascular tissue Sphagna also possess a comparatively simple water use strategy. Although recently challenged (see Sternberg and Ellsworth, 2011), it is generally accepted that a temperature-insensitive, constant enrichment factor between source water and *Sphagnum* cellulose of 27 ± 3‰ for oxygen isotopes exists (Zanazzi and Mora, 2005), meaning that $\delta^{18}O_{cellulose}$ in *Sphagnum* should accurately reflect changes in the source water oxygen isotopic composition (Daley et al., 2010), which is entirely meteoric in an ombrotrophic context.

Isotopic offsets between the different *Sphagnum* components (e.g. leaves, stems, branches) can lead to systematic errors during analysis and so the isolation of stem material is considered preferable (Loader et al., 2007; Moschen et al., 2009; Kaislahti Tillman et al., 2010, 2013). It is also important to isolate a single chemical compound, with α-cellulose favoured owing to the greater level of homogeneity achievable during the purification process (McCarroll and Loader, 2004; Loader et al., 2007). Advances in the extraction and purification of α-cellulose (Loader et al., 1997; Rinne et al., 2005; Daley et al., 2010), developments in stable isotope ratio mass spectrometry (IRMS) (McCarroll and Loader, 2004; Filot and Leuenberger, 2006; Loader et al., 2007; Young et al., 2011; Woodley et al., 2012), including the simultaneous measurement of stable carbon and oxygen isotopes (Woodley et al., 2012; Loader et al., 2015), have also significantly increased the efficiency of the technique.

Studies have also shown that isotopic signals can vary with *Sphagnum* species (Ménot and Burns, 2001; Ménot-Combes et al., 2002) but in sub-fossil samples, particularly when picking stems for analysis, identification to species level is often difficult (Loader et al., 2007; Moschen et al., 2009) However, analysis of the isotopic offsets between *Sphagnum* species have shown them to be of lesser magnitude than the effect of moisture changes (Rice, 2000; Loisel et al., 2009; Daley et al., 2010) and that, therefore, species do not require identification prior to analysis.

Daley et al. (2010) found that *Sphagnum* $\delta^{18}O_{cellulose}$ in a sub-fossil record exhibited far greater variability than could be explained by temperature variation alone and, instead, strong correlation with palaeohydrological proxies suggested a common climatic driver over centennial timescales, indicating that $\delta^{18}O_{cellulose}$ likely reflected either the cooling of or changes in the behaviour of prevailing air masses, as the isotopic composition of precipitation is known to reflect initial moisture source and rainout history along given air mass trajectories (Jouzel et al., 1997, 2000; Cole et al., 1999; Araguás-Araguás et al., 2000).

Stable isotope analysis ($\delta^{18}O$, $\delta^{13}C$) of *Sphagnum* in ombrotrophic raised bogs therefore possesses the potential for cross-proxy and cross-site validation with a range of palaeoecological proxy reconstructions (e.g. testate amoebae, plant macrofossils, humification) linked to warm season water deficit (Charman et al., 2009; Booth, 2010; Amesbury et al., 2012), to identify and verify the influence of a regional climate driver. Here we present multi-proxy palaeoecological (testate amoebae, plant macrofossils, humification) and stable isotope ($\delta^{18}O$, $\delta^{13}C$) records from three ombrotrophic peatlands to provide a temporally-focused examination of the 5.2 ka event in Ireland and its likely climatic causes in the North Atlantic region.

Here we test the timing and impact of the 5.2 ka event by generating three new stable isotopic records as part of a broader multi-proxy study of Irish ombrotrophic peatlands and assess the likely relationship with wider climate dynamics in the North Atlantic and globally.

2. Regional setting

Ireland's maritime climate is heavily influenced by prevailing westerly airflow and is therefore sensitive to ocean-atmosphere circulation changes in the North Atlantic region, with summer precipitation strongly governed by westerly wind strength (McDermott et al., 2001; Anderson et al., 2004; Turney et al., 2005; Swindles et al., 2010, 2013).

Palaeohydrological reconstructions derived from ombrotrophic peat sequences in Ireland have therefore frequently been interpreted as proxies for the strength and position of the westerlies (e.g. Blundell et al., 2008; Swindles et al., 2010; Langdon et al., 2012; Roland et al., 2014). A generalised model of warm season water deficit as the primary control on peatland water tables has been proposed (Charman, 2007; Booth, 2010; Charman et al., 2009). The relative importance of the temperature and precipitation components may vary geographically (Charman et al., 2004) and between methods (Amesbury et al., 2012), however. Comparisons between peat-based palaeohydrological reconstructions and meteorological records in Ireland confirm that summer precipitation is likely to be more influential on peatland water table depth (WTD) in oceanic areas (Charman et al., 2012). Although autogenic ecohydrological processes are likely to influence peatland water tables (Blaauw and Mauquoy, 2012; Swindles et al., 2012; Waddington et al., 2015), palaeohydrological reconstructions have been plausibly replicated both spatially and temporally (e.g. Charman et al., 2006; Swindles et al., 2013) supporting the presence of a common climate signal in these instances and demonstrating the potential for palaeoclimate reconstructions in these archives.

Furthermore, the stable oxygen isotopic composition of precipitation ($\delta^{18}O_{precip}$) in Ireland is closely linked to changes in atmospheric circulation, with the NAO found to account for a significant proportion of the variability on longer timescales, although the precise nature of these relationships is complex (Baldini et al., 2010). A previous study in northern England produced a $\delta^{18}O_{cellulose}$ record spanning the last 4300 years which was strongly correlated with corresponding palaeohydrological reconstructions (Daley et al., 2010), further highlighting the regional potential for using this novel technique to reconstruct past changes in atmospheric circulation.

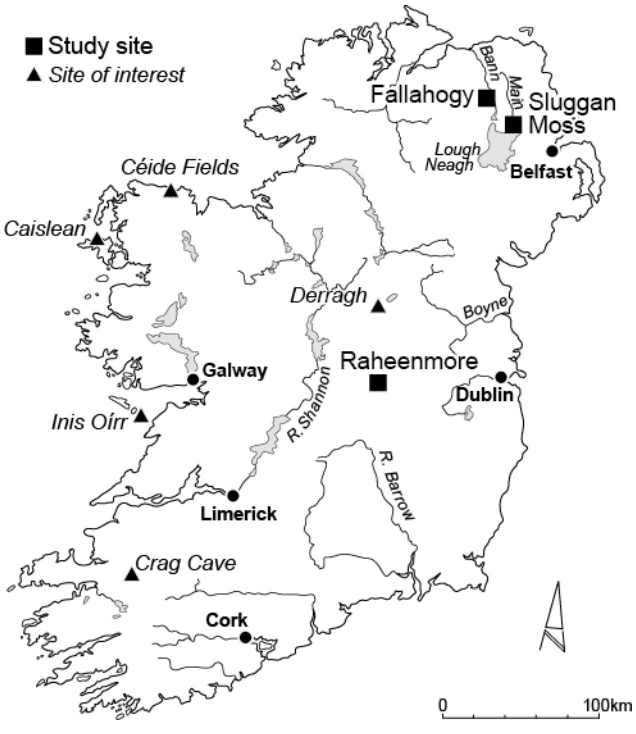

Fig. 1. Location of study sites, Sluggan Moss, Fallahogy Bog and Raheenmore, in Ireland together with other sites discussed in the text.

Three new high-resolution isotopic records from Ireland are reported here (Figure 1). Sluggan Moss (Co. Antrim; 54.766°N, 6.294°W) and Fallahogy Bog (Co. Derry; 54.911°N, 6.561°W) lie c. 22 km apart in the lower valleys of the Rivers Main and Bann respectively, in an extensive lowland area between the Antrim Hills and Sperrin Mountains of northern Ireland. These two sites provide the localised replication of any palaeohydrological signal. A third site, Raheenmore (Co. Offaly; 53.336°N, 7.343°W), located c. 175 km to the south in midland Ireland in a small basin within the catchments of the Rivers Brosna and Boyne, is included as a test of a regional $\delta^{18}O$ signal.

Plant macrofossil and bog stratigraphy data presented in Roland et al. (2014) and Daley (2007) demonstrate the relative dominance of *Sphagnum,* particularly *Sphagnum austinii*, confirming the onset of ombrotrophy prior to this study's temporal focus at all sites and providing abundant material for stable isotope analysis. Whilst recent peat cutting has taken place at all sites, the palaeohydrological record stored in the mid-Holocene portion of these sequences has not been affected by these activities.

3. Materials and methods

3.1 Field and laboratory sampling

Cores were extracted from lawn microforms with a wide-bore Russian peat corer using a parallel-borehole method (De Vleeschouwer et al., 2010) and stored at c. 4°C to minimise biological activity. Sub-sampling procedures followed conventional methods (De Vleeschouwer et al., 2010) with contiguous (humification analysis) and non-contiguous (plant macrofossil, testate amoebae and stable isotope analyses) samples of 1 cm stratigraphic depth. Sampling resolution for palaeoecological analyses (plant macrofossil and testate amoebae) was between 2 and 4 cm for both profiles. Sampling resolution for stable isotope analysis was partially dependent on the availability of sample material. Sampling for tephra analysis followed standard techniques (Pilcher and Hall, 1992; Roland et al., 2015).

3.2 Palaeoecological analysis

Preparation for testate amoebae analysis followed standard techniques (Booth et al., 2010). Taxonomy followed Charman et al. (2000) with the addition of *Centropyxis ecornis* (Booth, 2008) and the reclassification of *Archerella flavum* (Gomaa et al., 2013). At least 100 individual tests were counted for most levels, with counts of 50 only accepted for statistical analysis when testate amoebae concentration and/or preservation was exceptionally poor (Swindles et al., 2007b; Payne and Mitchell, 2009). Testate amoebae-derived water table reconstructions were produced using the pan-European ACCROTELM (Charman et al., 2007) and regionally-specific northern Irish (Swindles et al., 2009) transfer functions, based on weighted averaging tolerance-downweighted regression with inverse deshrinking. Sample-specific errors for the reconstruction were calculated using 1000 boot- strap cycles (Birks et al., 1990; Line et al., 1994).

Preparation for plant macrofossil analysis followed standard techniques (Mauquoy et al., 2010) and identifications were made using a range of type specimens and texts (Grosse-Brauckmann, 1974; Daniels and Eddy, 1990; Smith, 2004; Mauquoy and van Geel, 2006). Plant macrofossil data were transformed into univariate bog surface wetness (BSW) indices for ease of interpretation using detrended correspondence analysis (DCA) and the Dupont Hydroclimatic Index (DHI) methods (Daley and Barber, 2012). DHI estimates relative hydrological conditions based on the weighted averaging of plant macrofossil data (Dupont, 1986). Modified weightings, revised and expanded to include species commonly found in ombrotrophic bogs, were used (Mauquoy, 1997; Daley and Barber, 2012). Palaeoecological diagrams were produced using TILIA and TGView (Grimm, 1991, 2004) and DCA was undertaken in the *Vegan* package (Oksanen et al., 2013) in R (R Core Team, 2014)

Humification analysis followed the standard colorimetric procedure (Blackford and Chambers, 1993). Percentage light transmission data are presented as detrended residuals, following linear regression of the raw data to remove the down-core tendency towards a higher degree of humification (Borgmark and Wastegård, 2008). Palaeoecological and humification data for Raheenmore were not available.

3.3 $\delta^{13}C$ and $\delta^{18}O$ analysis

Sphagnum stem material was isolated through sieving (1000 μm) with distilled water and cleaning under 10x magnification to remove fine detritus and other plant remains, before being prepared to α-cellulose using standard methodology, which included ultrasonic homogenisation and freeze drying (Loader et al., 1997; Daley et al., 2010). Sample sizes of 0.3 – 0.35 mg of prepared α-cellulose were used, in triplicate where possible. $\delta^{18}O$ and $\delta^{13}C$ data from Sluggan Moss and Fallahogy Bog were produced following high-temperature pyrolysis (Young et al., 2011; Woodley et al., 2012), whereas $\delta^{18}O$ data from Raheenmore were produced separately following low-temperature (1090°C) pyrolysis described by Daley et al. (2010). Stable isotope ratios were reported as per mil (‰) deviations from the VSMOW standard for oxygen ($\delta^{18}O$) and the VPDB standard for carbon ($\delta^{13}C$) and where:

$$\delta^{18}\mathrm{O} \text{ or } \delta^{13}\mathrm{C} = \left[\left(\frac{\mathrm{R_{sample}}}{\mathrm{R_{standard}}}\right) - 1\right] \times 1000 \qquad \text{(Eq. 1)}$$

where R is the ratio of $^{18}O/^{16}O$ or $^{13}C/^{12}C$ in the sample and standard. A mean of three isotope measurements was calculated for each sample where possible.

3.4 Chronology

Full details of the chronological methodologies and data used (i.e. AMS ^{14}C, tephra) in the development of age-depth models for all sites can be found in Roland et al. (2014) and Daley (2007). The presence of rhyolitic tephra shards from the Hekla 4 eruption (4.345-4.229 ka; Pilcher et al., 1996) was confirmed in sequences from Sluggan Moss and Fallahogy Bog by Roland et al. (2014), and although smaller concentrations of shards were found elsewhere in these sequences they yielded no distinct horizons after further investigation. An absence of Hekla 4 shards in central western Ireland has been noted (Chambers et al., 2004; Schettler et al., 2006) and so its absence from Raheenmore is not unexpected. Age-models (Fig. 2) were produced for all sites using the Bayesian age-depth modelling package, *Bacon* (Blaauw and Christen, 2011) in R (R Core Team, 2014). Default priors were accepted for accumulation rate unless the software suggested otherwise; faster accumulation rates of 10 yr/cm were therefore accepted for Raheenmore and Fallahogy Bog. ^{14}C dates were calibrated using the IntCal13 calibration curve (Reimer et al., 2013) and assume a Student's-t distribution with wide tails instead of the usual Gaussian distribution (Christen and Pérez, 2009). Ages quoted hereafter are based on weighted average means of each age-model (Telford et al., 2004).

4. Results

Summaries of the palaeoecological data (plant macrofossils, testate amoebae) for Sluggan Moss and Fallahogy Bog are presented in Figures 3 and 4, respectively. Both peat sequences are characterised by a dominance of *Sphagnum austinii*, indicating ombrotrophic conditions, with periodic incursions of other *Sphagnum* species of the sections *Acutifolia* and, in the case of Sluggan Moss, *Cuspidata*. A stratigraphical survey at Raheenmore confirmed the presence of *Sphagnum*-rich peat, indicating ombrotrophic conditions, at depths beyond this study's period of focus (Daley, 2007). A previous DCA of plant macrofossil results (Roland et al., 2014), with no transformation or downweighting of rare species produced eigenvalues of 0.6145 at Sluggan Moss and 0.4738 at Fallahogy Bog, therefore approaching or exceeding the desired value of >0.5 (ter Braak, 1995). At both sites the distribution of plant species along the first axes indicates the presence of a latent hydrological gradient within the data, with species indicative of drier (e.g. Ericaceae) and wetter (e.g. *S*. s. *Cuspidata*) conditions positioned at opposite ends of the axes. Distribution of species scores at both sites approached 5 standard deviations along this first axis, suggesting a lack of overlap between species of different hydrological preferences, further strengthening subsequent interpretation (Daley and Barber, 2012). An absence of *S*. s. *Cuspidata*, together with an increased abundance of unidentified organic matter (UOM) compared with levels seen at Sluggan Moss, indicates that Fallahogy Bog was historically the drier of the two northern sites. Summaries of the fossil testate amoebae data (Figs. 3 and 4) confirm this assertion with species indicative of moderate to dry conditions (e.g. *Difflugia pulex* and *Trigonopyxis arcula*) broadly more prevalent than at Sluggan Moss.

Figure 5 presents normalised contiguous detrended percentage light transmission data alongside univariate palaeoecological reconstructions and $\delta^{13}C$ data for ease of comparison from Sluggan Moss and Fallahogy Bog. A number of shifts towards higher transmission values and lower degrees of humification are suggested, potentially indicative of wetter and/or cooler climatic conditions (Blackford and Chambers, 1993). A marked wet shift is suggested in the proxy data from Sluggan Moss c. 5.5 ka, but a relatively complacent trend is recorded in Fallahogy Bog, where hydrological conditions appear to be more stable. The relatively large uncertainty in reconstructed changes in surface wetness precludes an unambiguous identification of a significant moisture shift across this period.

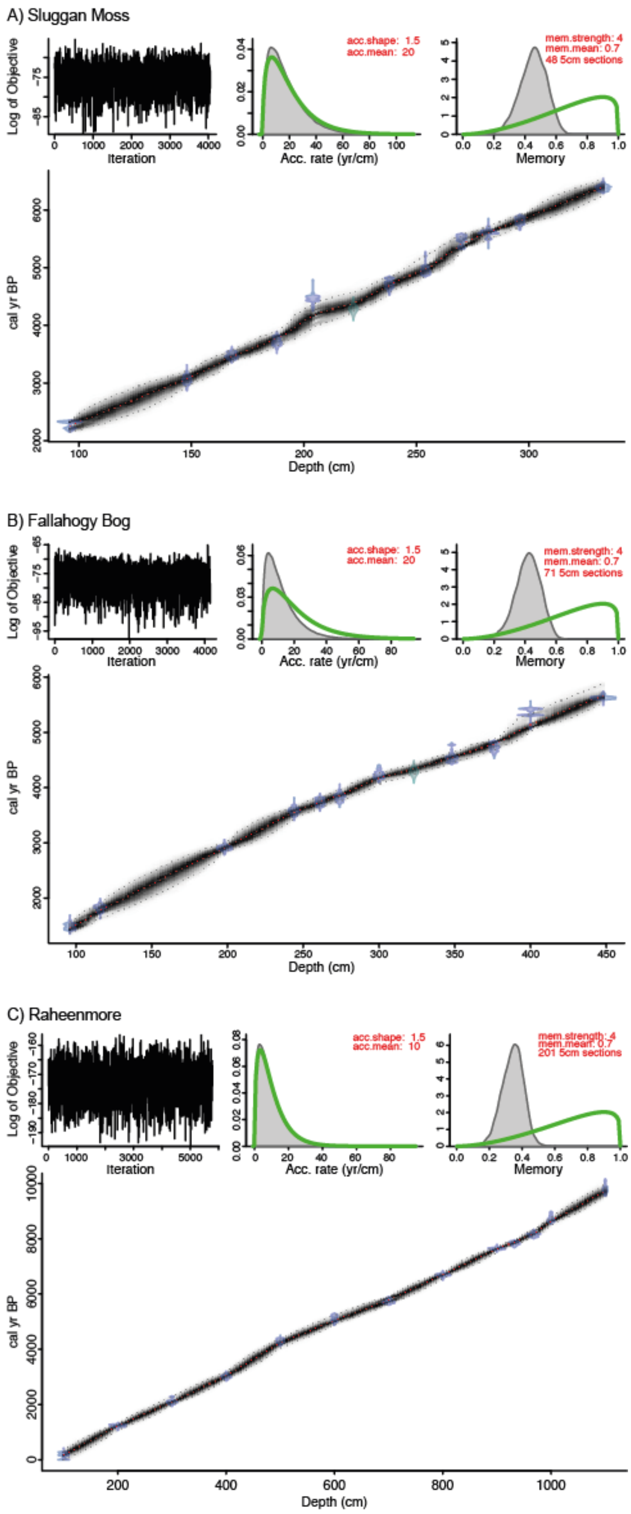

Fig. 2. *Bacon* age-depth models from Sluggan Moss, Fallahogy Bog and Raheenmore using radiocarbon ages (blue distributions) and Hekla 4 tephra (turquoise distributions; Sluggan Moss and Fallahogy only). The upper panels in each model represent *Bacon* settings and results for (left to right) number of iterations used, accumulation rate and model memory (see Blaauw and Christen (2011) for more details).

Importantly, $\delta^{18}O_{cellulose}$ demonstrates a robust and sustained shift across the period of interest (Figure 6 and Table 1). Error bars present 2σ error ranges based on replicate measurements of the same sample where available. Triplicate measurements were taken for the majority of samples but there were a small number of instances where only duplicate or single measurements were possible as a result of small sample size (Table 2). An apparent drift correction in the standards was evident in replicate $\delta^{13}C_{cellulose}$ but not $\delta^{18}O_{cellulose}$ data for samples dating from 4.2 – 3.25 ka at Fallahogy Bog and so these samples do not possess error bars.

Linear regression of $\delta^{18}O_{cellulose}$ and $\delta^{13}C_{cellulose}$ data at Sluggan Moss resulted in a moderate but statistically significant correlation ($r^2 = 0.33$, $p<0.001$), potentially suggesting a common forcing mechanism. Conversely, no statistically significant relationship between the two stable isotopes was found at Fallahogy Bog ($r^2 = 0.02$, $p = 0.32$).

Tab. 1. A summary of the isotopic data from Sluggan Moss, Fallahogy Bog, Raheenmore and Lough Corrib, 2 – 6.5 ka.

Parameter	Isotope	Sluggan Moss (this study)	Fallahogy Bog (this study)	Raheenmore (Daley, 2007)
Highest value	$\delta^{18}O$	25.18‰	26.02‰	22.65‰
	$\delta^{13}C$	-27.25‰	-26.25‰	-
Lowest value	$\delta^{18}O$	20.90‰	21.77‰	19.05‰
	$\delta^{13}C$	-21.46‰	-28.89‰	-
Range	$\delta^{18}O$	4.28‰	4.25‰	3.60‰
	$\delta^{13}C$	5.79‰	2.64‰	-
Mean	$\delta^{18}O$	23.74‰	23.42‰	20.75‰
	$\delta^{13}C$	-25.63‰	-27.31‰	-
Standard deviation	$\delta^{18}O$	0.96‰	0.72‰	0.77‰
	$\delta^{13}C$	1.30‰	0.57‰	-
Variance	$\delta^{18}O$	0.92‰	0.52‰	0.59‰
	$\delta^{13}C$	1.68‰	0.33‰	-

Tab. 2. Problematic isotopic measurements at Fallahogy Bog. Potential 'outliers' are italicised.

Depth (cm)	Approximate age (ka)	$\delta^{18}O$ measurements (‰)			Average (‰ ± 1σ)	
		1	*2*	*3*	*With 'outlier'*	*Without 'outlier'*
306	4.225	27.165	27.362	*23.540*	26.02 ± 2.15	27.26 ± 0.14
324	4.35	23.627	23.803	*26.749*	24.73 ± 1.75	23.72 ± 0.12
		$\delta^{13}C$ measurements (‰)			**Average (‰ ± 1σ)**	

344	4.525	-25.017	-26.368	-27.350	-26.25 ± 1.17

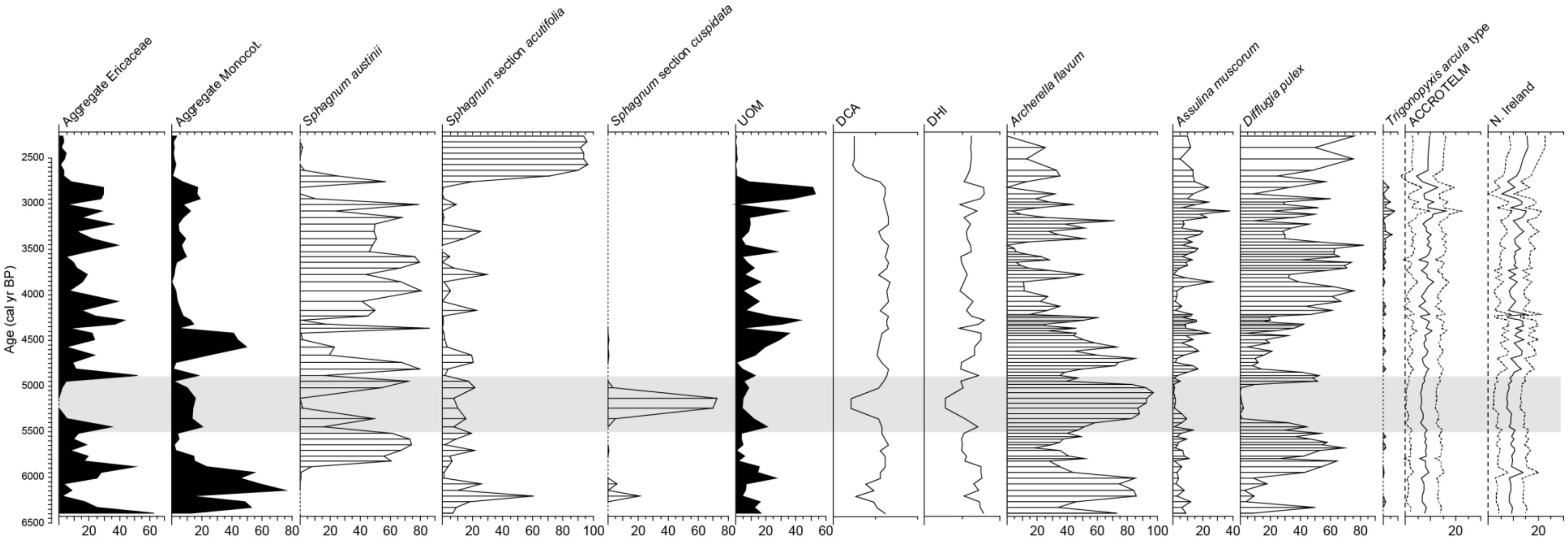

Fig. 3. Summary palaeoecological diagram showing selected plant macrofossil, and testate amoebae data from Sluggan Moss. Plant macrofossil and testate amoebae counts are displayed as percentages; plant macrofossil DCA axis one scores and DHI scores have been normalised for comparison. Inferred WTD reconstructions based on the ACCROTELM pan-European (Charman et al., 2007) and northern Ireland (Swindles et al., 2009) transfer functions (black curve) with errors derived from bootstrapping (grey curves). Increasing bog surface wetness conditions are indicated by shifts to the left in all curves. The periods 5.5 – 4.95 and 3.65 – 3.85 ka are highlighted in grey.

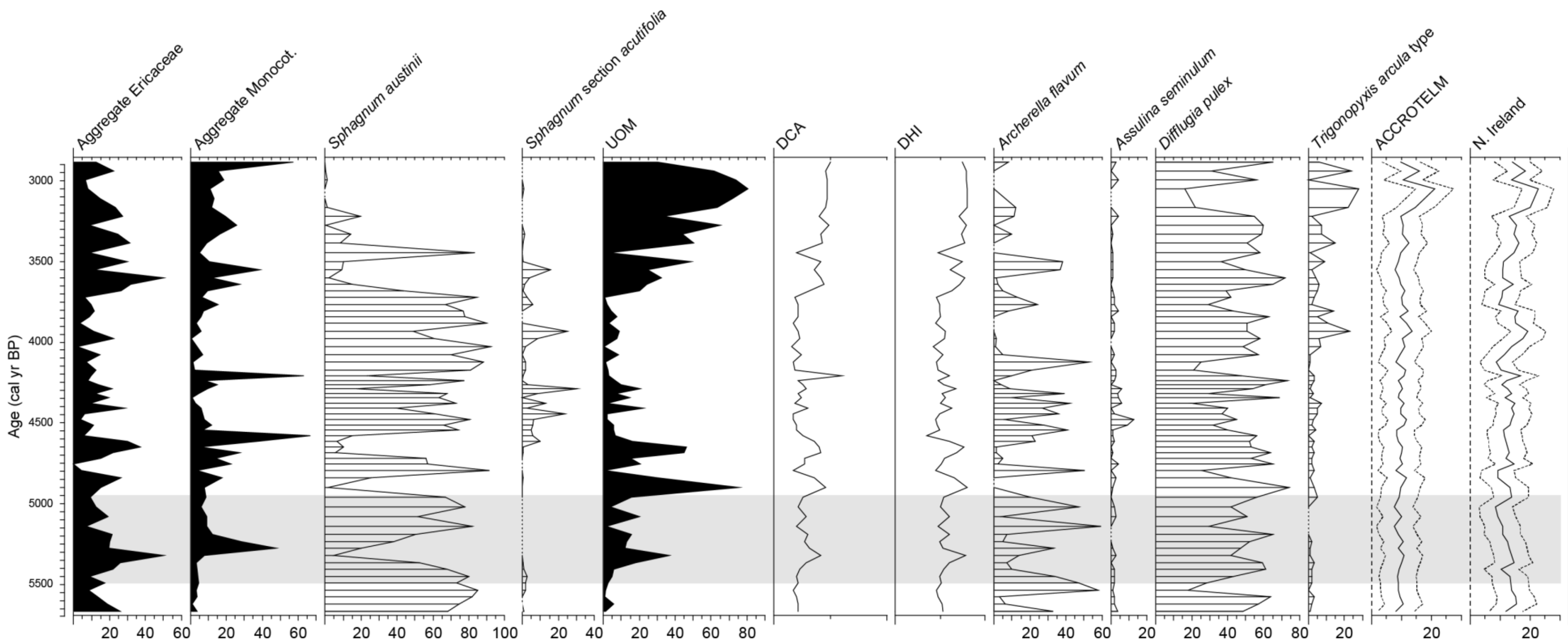

Fig. 4. Summary palaeoecological diagram showing selected plant macrofossil, and testate amoebae data from Fallahogy Bog. Plant macrofossil and testate amoebae counts are displayed as percentages; plant macrofossil DCA axis one scores and DHI scores have been normalised for comparison. Inferred WTD reconstructions based on the ACCROTELM pan-European (Charman et al., 2007) and northern Ireland (Swindles et al., 2009) transfer functions (black curve) with errors derived from bootstrapping (grey curves). Increasing bog surface wetness conditions are indicated by shifts to the left in all curves. The periods 5. 5 – 4.95 and 3.65 – 3.85 ka are highlighted in grey.

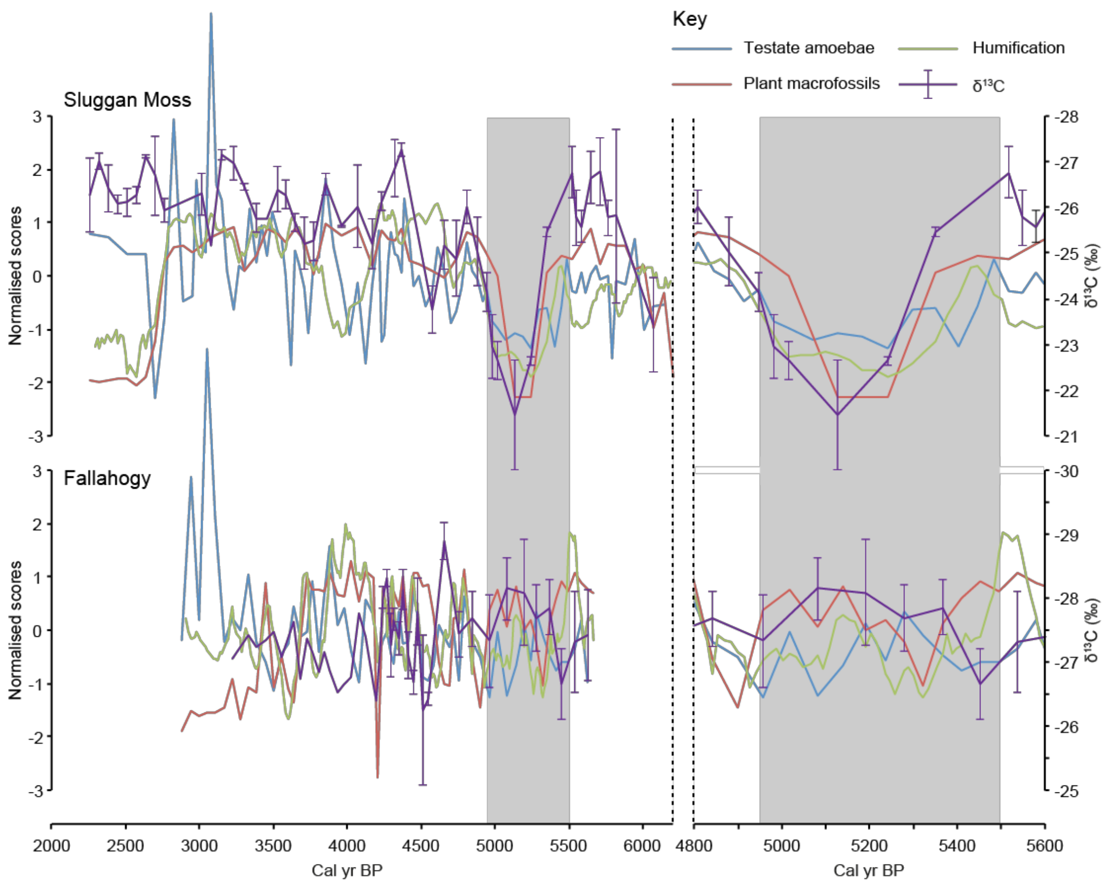

Fig. 5. Comparison of normalised data for testate amoebae-based reconstructed water table (ACCROTELM transfer function; blue), humification (green) and plant macrofossil (red) DCA normalised palaeoecological data with $\delta^{13}C_{cellulose}$ data (purple) from Sluggan Moss and Fallahogy Bog. Error bars associated with $\delta^{13}C_{cellulose}$ data present 2σ error ranges based on replicate measurements of the same sample where available. Equivalent data were not available for Raheenmore. The period 5.5-4.95 ka is highlighted in grey.

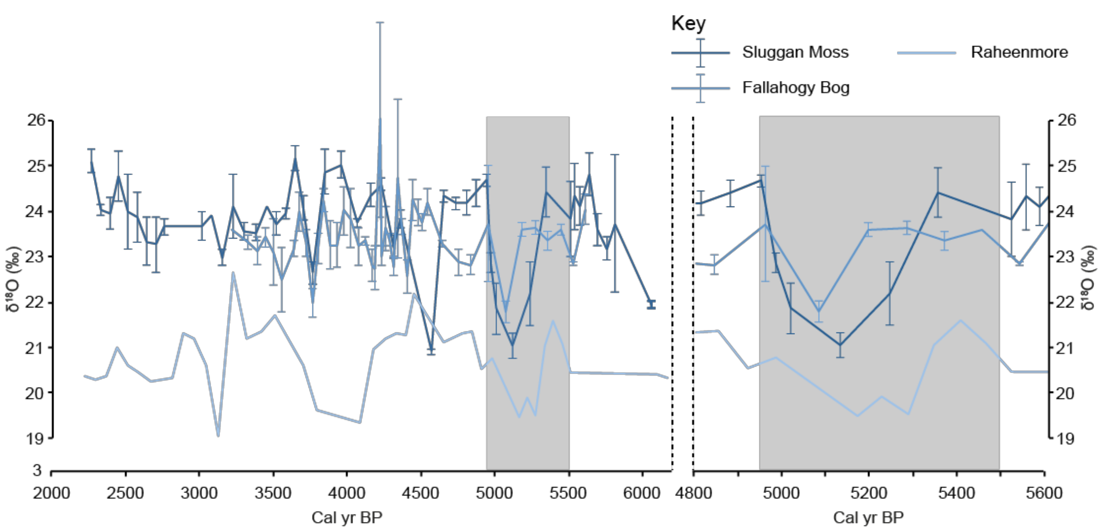

Fig. 6. $\delta^{18}O_{cellulose}$ data from Sluggan Moss (dark blue), Fallahogy Bog (mid-blue) and Raheenmore (light blue). Error bars present 2σ error ranges based on replicate measurements of the same sample where available. The period 5.5-4.95 ka is highlighted in grey.

Overall, the $\delta^{18}O_{cellulose}$ records at Sluggan Moss and Fallahogy Bog demonstrate strong agreement, although there is some slight disparity in the magnitude of $\delta^{18}O$ changes. While there are a small number of isotopic excursions that are present in single records (possibly indicating the presence of locally-specific factors and the isotopic variation of meteoric water) the use of multiple sites allows us to confidently identify common periods of change.

The absolute magnitude of $\delta^{18}O_{cellulose}$ variability from Raheenmore, developed using the low-temperature pyrolysis technique cannot be straightforwardly compared with data from Fallahogy Bog and Sluggan Moss, where data was produced using the high-temperature approach, as there is no way of verifying the degree to which the $\delta^{18}O_{cellulose}$ signal reflects local fractionation, rainout or other hydrological processes. Variation in pyrolysis temperature can explain differences in variance in the stable isotope record, but there is no evidence that a systematic offset, which would require consistent selective pyrolysis of heavy and/or light oxygen atoms within a cellulose unit, exists (Woodley et al., 2012). However, examining the relative magnitude and timing of changes between these records can still provide useful comparisons.

Crucially, a high degree of coherence exists between $\delta^{18}O_{cellulose}$ in the three sequences during the period 5.5 – 4.95 ka. Here, large shifts (i.e. >1‰) towards more depleted values occur, apparently contemporaneously (Figure 6). An additional shift towards more depleted values occurs at Sluggan Moss c. 4.575 ka, but is not shown at the other sites. At Fallahogy Bog, two excursions towards enriched $\delta^{18}O$ values occur at 4.35 (306 cm) and 4.225 (324 cm) ka possess large associated errors, potentially indicating significant variation in intra-sample $\delta^{18}O$ measurement. Both samples were subject to measurements in triplicate and have one outlying measurement, most likely caused by incomplete homogenisation of the α-cellulose fraction during laboratory preparation. Table 2 highlights these potential outliers and demonstrates that upon their exclusion, the average $\delta^{18}O$ value for the sample at 324 cm falls within the range of normal variability within the core in terms of both variance and one standard deviation (23.72 ± 0.12‰), thus suggesting preparation error. Removal of this outlier for the sample at 306 cm (27.26 ± 0.14‰), however, appears to confirm the presence of a genuine enrichment event.

The $\delta^{18}O_{cellulose}$ excursion associated with the 5.2 ka event at Sluggan Moss is larger (4.78‰) than that of Fallahogy Bog (1.84‰) and is also coincident with a change in the dominant *Sphagnum* species from *S. austinii* to *S.* s. *Cuspidata*. At least part of this difference in magnitude could be attributed to a species effect with two of the lowest values (22.18‰, 21.03‰) occurring under *S.* s. *Cuspidata* dominance as modern annual seasonal variation in precipitation ^{18}O of northwest Europe is just ~2-4‰ (Rozanski et al., 1993). The excursion persists beyond this, however, and into a period of *S. austinii* dominance producing a comparable isotopic value of -21.88‰ supporting the existence of a genuine isotopic shift associated with the 5.2 ka event. A concurrent stable isotopic shift also occurs at Fallahogy Bog, where no such shift in *Sphagnum* species occurs.

5. Discussion

5.1 Evidence for the 5.2 ka event

The major shifts between *Archerella flavum* and *Difflugia pulex* in the Sluggan Moss record particularly around the 5.2 ka event (c. 5.5 – 4.95 ka), are not well reflected in the transfer function reconstructions, but are clearly indicative of considerable changes in peatland surface environmental conditions (Fig. 3). *A. flavum* and *D. pulex* are both considered intermediate species in terms of their hydrological tolerances within the transfer function models employed here (Charman et al., 2007; Swindles et al., 2009). The latter species is, however, poorly represented in both models owing to a relative lack of modern analogues. *A. flavum* is typically associated with moderate to wet conditions, and sometimes standing water, whereas *D. pulex* is considered a relatively dry indicator (Charman et al., 2000) but the two species may also occupy opposite ends of an environmental variability index, with *D. pulex* indicative of highly variable conditions and *A. flavum* of environmental stability at the peatland surface (Sullivan and Booth, 2011). The abundance of *D. pulex* has been shown to exhibit vertical zonation in favour of subsurface samples, which may also explain its underrepresentation in transfer function training sets (van Bellen et al., 2014). The broadly concurrent emergence of *Sphagnum* section *Cuspidata*, coupled with a shift towards lower levels of peat humification at Sluggan Moss emphasise a period of prolonged wet conditions at this site, with particularly strong coherence between all proxies at this point (Fig. 5).

Palaeoecological data from Fallahogy Bog provide less convincing evidence for a 5.2 ka event but as the drier of the two sites, based on palaeoecological evidence, this is not unexpected. A notable shift back to a dominance of *Sphagnum austinii* at c. 5.3 ka following the establishment of ericaceous and monocotyledonous plants indicates wetter conditions at the bog surface, but this change is not reflected in the testate record which remains dominated by *D. pulex* for the majority of the sequence (Fig. 4). The broadly coincident disappearance of two dry indicators, *Assulina seminulum* and *Trigonopyxis arcula* type (Charman et al., 2000), could indicate a shift to wetter bog surface conditions c. 5.2 ka and is consistent with a trend towards reduced peat humification, indicating wetter conditions (Fig. 5).

At Sluggan Moss, $\delta^{13}C_{cellulose}$ values respond consistently with the palaeoecological data during the 5.2 ka event, supporting the suggestion that $\delta^{13}C$ in *Sphagnum* is strongly influenced by bog surface wetness (Loisel et al., 2010). At Sluggan Moss, a large shift of 5.3‰ occurs between c. 5525 BP (-26.8‰) and 5125 BP (-21.5‰), before returning to previous levels by c. 4950 BP (-24.2‰), indicative of a multi-centennial wet event. $\delta^{13}C_{cellulose}$ values at Fallahogy Bog remain relatively stable throughout this period, suggests relatively stable water table levels (Fig. 5).

Strong coherence exists between the $\delta^{18}O_{cellulose}$ records at Sluggan Moss, Fallahogy Bog and Raheenmore, suggesting that $\delta^{18}O_{cellulose}$ in *Sphagnum* has been caused by simultaneous variation in the oxygen isotopic composition of the source water available at all three sites (Daley et al., 2010). This is likely to be the result of significant change in the mode of atmospheric circulation, altering air mass histories and sources of precipitation, with a persistent shift in the NAO and the associated westerlies a likely mechanism. Following a two-year monitoring study, Baldini et al. (2010) found that on monthly timescales, $\delta^{18}O_{precip}$ was heavily influenced by NAO with little effect from temperature. Back trajectory analysis demonstrated that amount-weighted mean $\delta^{18}O_{precip}$ of rain events possessing southerly and northerly trajectories can be depleted in ^{18}O by c. 2.0‰ relative to those with westerly trajectories. The most $\delta^{18}O$-depleted rain events were also associated with cyclonic weather conditions during which continuous moisture recycling might explain low $\delta^{18}O_{precip}$ values.

The NAO represents the dominant mode of atmospheric circulation variability in the mid-latitude North Atlantic, exerting major influence on the temperature and precipitation patterns in the region. When its index is positive (NAO^{+}), mid-latitude westerlies are intensified and enhanced zonal flow occurs across the North Atlantic, bringing warm and wet conditions to western Europe (Timm, 2008; Olsen et al., 2012). It therefore seems possible that the $\delta^{18}O_{cellulose}$ depletion episode (5.5 – 4.95 ka) identified here was a result of a multi-decadal to centennial increase in the frequency of cyclonic weather systems, associated with NAO^{+} conditions that significantly changed the dominant air mass trajectory over Ireland. This would also be consistent with palaeoecological and $\delta^{13}C_{cellulose}$ data, particularly from Sluggan Moss, which indicate a notable and coincident shift to wetter conditions.

5.2 Northwest European context

Few Irish palaeoclimate records extend beyond 5 ka (Swindles et al., 2013). Figure 7 summarises the small number of records available that can provide context for the 5.2 ka event in Ireland. Blanket peat records from Achill Island (Fig. 7a; Caseldine et al., 2005) provide evidence for an extreme inwash event associated with increased storminess between 5.3 and 5.05 ka, following a period of relative dryness since c. 5.8 ka as interpreted from corresponding humification and palynological data. A number of peat sequences in Scandinavia record increased aeolian input c. 5.2 – 4.8 ka suggesting this period of increased storminess was experienced regionally (Björck and Clemmensen, 2004; Clemmensen et al., 2006; Jong et al., 2006). Similarly, a major phase of aeolian activity was dated to c. 5.6 ka in Portuguese coastal dunefield (Costas et al., 2012). The consistency in these records along the Atlantic European seaboard provides strong support for a period of enhanced westerly winds and increased storminess.

A testate amoebae record from Derragh bog, midland Ireland (Fig. 7b; Langdon et al., 2012), demonstrates a period of relatively high BSW 6.0 – 5.0 ka, with the wettest conditions occurring between 5.5 and 5.2 ka. This wet period is also characterised by greater WTD fluctuation than elsewhere in the record, potentially indicating increased climatic variability. In the absence of palaeoecological data from Raheenmore, the record from Derragh confirms that midland Ireland is likely to have experienced wetter conditions during the 5.2 ka event.

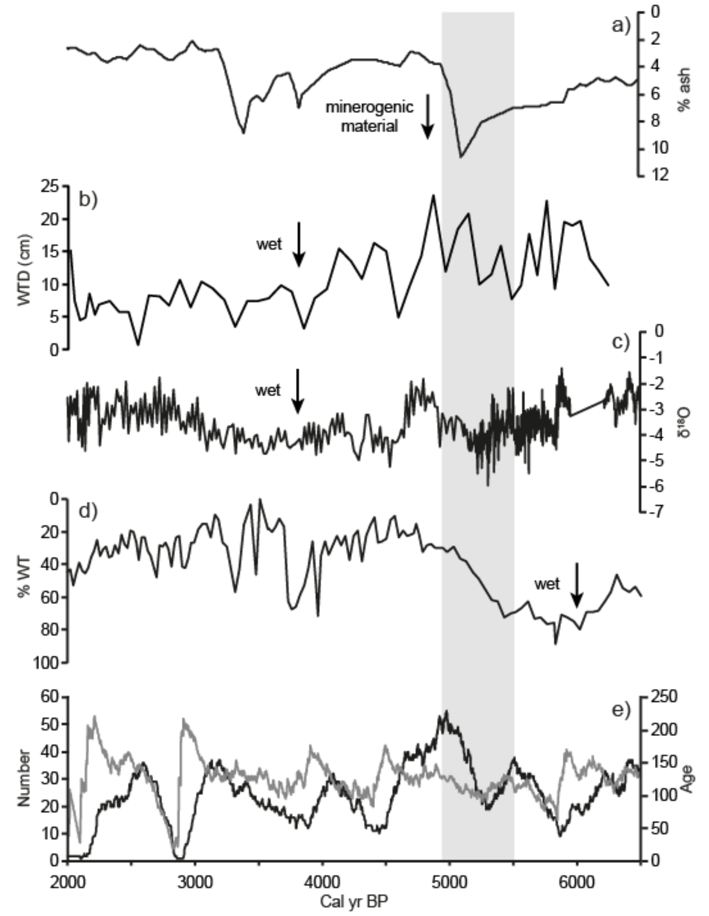

Figure 7 Palaeoclimatic records from Ireland, 6.5-2 ka. a) Loss-on-ignition records Caislean peat cores, Achill Island, western Ireland (Caseldine et al., 2005); b) Testate amoebae-based water table reconstruction (ACCROTELM transfer function) from Derragh Bog, central Ireland (Langdon et al., 2012); c) $\delta^{18}O$ from speleothem CC3, Crag Cave, southwest Ireland (McDermott et al., 2001); d) CaCO3 record, expressed as a weight percentage, from Inis Oírr, western Ireland (Schettler et al., 2006); e) Northern Irish dendro-dated bog oak population (black) and average age (grey) (Turney et al., 2005). The period 5.5-4.95 ka is highlighted in grey.

$\delta^{18}O$ variations in a speleothem record from Crag Cave, southwest Ireland (Fig. 7c; McDermott et al., 2001), have been shown to correlate with those in the GISP2 ice core and are thus postulated to reflect Holocene climate signals. After a period of relative stability in the Crag Cave record from c. 5.8 ka, $\delta^{18}O$ declines from 5.5 to 5.2 ka indicating an increased prevalence of wetter and possibly cooler conditions.

In a brackish karst lake record from An Loch Mór, western Ireland (Fig. 7d; Schettler et al., 2006), higher proportions of $CaCO_3$ are interpreted as corresponding with periods of increased freshwater discharge from the catchment, as a result of increased precipitation. Conversely, low

$CaCO_3$ relates to increased seawater infiltration, enhancing lake productivity. However, the onset of diurnal seawater infiltration at c. 5.1 ka is further complicated by the hydrological effects of anthropogenic changes in catchment vegetation dynamics. This followed an increase in farming intensity from c. 5.5 ka with woodland regeneration evident c. 5 ka onwards as permanent settlement on the island appears to cease, possibly linked to the hypothesised freshwater shortage. In the same record, varve formation ceased c. 5.2 ka (Holmes et al., 2007) which, together with increased seawater infiltration into the lake system could be the result of increased storminess and increased prevalence of wetter conditions during the 5.2 ka event.

Population dynamics of Irish bog oaks (Fig. 8d) have been explained as a function of BSW, with population peaks said to represent colonisation during dry periods and troughs indicating population reductions owing to waterlogging (Turney et al., 2005), although comparison with independent climate proxies found no consistent relationship (Swindles and Plunkett, 2010).

Instead, Charman (2010) suggests that rather than interpreting peaks and troughs, it is the rising and falling limbs of the population curve that are significant, where a declining population indicates a reduction in tree recruitment as a result of wet conditions and *vice versa*. Based on this interpretation, correspondence between the bog oak record and regional BSW records improves. It follows that a declining population should also correspond with an increase in its mean age, reflecting this reduction in recruitment, a pattern that can clearly be seen in most of the record. Whilst a sharp reduction in oak population occurs at 5.5ka, the corresponding peak in average age does not appear and instead remains fairly constant through the period associated with the 5.2 ka event, potentially indicating a disruption in population dynamics of these trees.

The difficulty in interpreting many of the records presented here highlights the importance and potential of longer peatland records for early- and mid-Holocene palaeoclimate reconstruction in Ireland. A number of other peatlands across northwest Europe demonstrate shifts towards wetter conditions c. 5.3 – 5.2 ka but chronological resolution and precision varies dramatically between records (e.g. Aaby, 1976; Hughes et al., 2000; Barber et al., 2003; Langdon et al., 2003; Barber, 2007; Borgmark and Wastegård, 2008; Kylander et al., 2013). Wetter conditions are also recorded 5.3 – 4.85 ka in peatland records from northwest Spain (Castro et al., 2014). Pollen-based mean annual temperature reconstructions from northern Europe also indicate colder conditions c. 5.3 ka (Seppä et al., 2009).

The data presented in this study also lend tentative support to the hypothesis that the economic and agricultural stability of Neolithic farming communities inhabiting marginal regions of western Ireland was detrimentally affected by a prolonged period of climatic deterioration c. 5.5 – 5 ka. Evidence from Achill Island (Caseldine et al., 2005), the Céide Fields (O'Connell and Molloy, 2001) and Belderrig (Verrill and Tipping, 2010) in Co. Mayo, and further south in Galway Bay (Schettler et al., 2006) suggests a abandonment of agricultural sites and eventual regional depopulation occurred at this time. More broadly in Ireland, archaeological evidence of human activity is sparse c. 5.4 – 5.1 ka and is coupled with a reduction in cereal remains, an increase in wild resources and palynological evidence for re-afforestation (Whitehouse et al., 2014). Similarly, a multi-centennial reduction in Neolithic landscape impact, beginning c. 5.3 ka, can also be inferred from archaeological and palynological records from across Great Britain (Woodbridge et al., 2012) though recent studies have suggested deterministic links between climate and prehistoric societies should be approached with caution (e.g. Armit et al., 2014) and although this period appears to be one of considerable environmental, landscape, settlement and economic change, any causal relationships between these factors is likely to be complex (Whitehouse et al., 2014).

5.3 A global context

The period 6 – 5 ka is characterised by a series of negative total solar irradiance (TSI) anomalies centered around 5.6, 5.45 and 5.3 ka (Fig. 8b, Steinhilber et al., 2012). A modelling study suggests that this reduction in solar activity was a major driver of the 5.2 ka event through an expansion of sea ice that led to cooling across the wider North Atlantic region (Renssen et al., 2006). Figure 8c demonstrates that the 5.2 ka event occurred at the height of ice-rafted debris (IRD) or Bond event 4 (Bond et al., 2001). Interestingly, this modelling also suggested the influence of negative TSI anomalies on other periods of cooling in the North Atlantic including the 2.8 – 2.6 ka event (Plunkett and Swindles, 2008) and Little Ice Age (Mauquoy et al., 2002) but not for the 4.2 ka event (Roland et al., 2014), suggestive of an alternative driver.

Major ion concentrations in the GISP2 ice core also demonstrate considerable atmospheric circulation change during the period 6 – 5 ka. During this time, increased levels of Na^+ and K^+ are interpreted as proxies for the expansion of the northern polar vortex (Fig. 8e; O'Brien et al., 1995) and strengthening of the Siberian high (Fig. 8d; Mayewski et al., 1997) respectively, caused by increased atmospheric loading of aerosols over the ice sheet. Such changes are consistent with the evidence presented in this study suggesting increased westerly intensity and prevalence of cyclonic weather patterns as drivers of the 5.2 ka event. There is also evidence for cooler temperatures over Greenland during this time from the GISP2 $\delta^{18}O$ record (Stuiver et al., 1995; Alley, 2000).

A brief interval of warmer sea surface conditions centring on c. 6 ka, as indicated by a Holocene maximum in a North Atlantic Current indicator diatom species, ended by 5.3 ka in the subpolar North Atlantic (Miller and Chapman, 2013) and marine records from the southern Labrador Sea indicate a substantial reduction in deep ocean flow speeds 5.6 – 4.8 with a pronounced minimum c. 5.1 ka (Hoogakker et al., 2011). The proposed occurrence of a weakening Atlantic meridional overturning circulation (AMOC) during what other records suggest is a NAO^+ period, however, conflicts with a recent modelling study that found, by examining an NAO index based on instrumental and documentary proxy data covering 1659 – 2000, that a NAO+ leads to a strengthening AMOC and warmer sea surface temperatures (Sun et al., 2015).

A number of records provide evidence for a shift in global climatic between 6 – 5 ka, including demonstration of the development of the mid-Holocene neoglacial period of glacier expansion in Scandinavia (Fig. 8g; Nesje, 2009), the abrupt onset of the African humid period (Fig. 8i; deMenocal et al., 2000) and the initiation of a shift towards colder temperatures in the Southern Hemisphere as evidenced by $\delta^{18}O$ values in the Taylor Dome (Steig et al., 2000) and Huascaran (Thompson et al., 1995) ice core records. It is therefore likely that the 5.2 ka event, as seen in the $\delta^{18}O_{cellulose}$ data presented in this study (Fig. 8h), was the manifestation of a non-linear response to broader climate forcing and reorganization taking place 6 – 5 ka. 'Event' type signals can also be seen in range of palaeoclimate archives from the mid-latitudes, including the $\delta^{18}O$ record from Kilimanjaro (Fig. 8l; Thompson et al., 2002), marine sediments in the Gulf of Oman (Fig. 8m; Cullen et al., 2000) and $\delta^{18}O$ and $\delta^{13}C$ records from a speleothem in Soreq Cave, Israel (Fig. 8n; Bar-Matthews et al., 2003), all of which support a southward migration of the ITCZ. A broad scale shift to wetter conditions, potentially linked to enhanced southern westerlies has also been observed in a range of palaeoclimate archives in southern South America 5.5 – 4.9 ka (Hermanns and Biester, 2013), further demonstrating a major climatic reorganisation took place at that time.

6. Concluding remarks

Evidence for a pronounced climatic event c. 5.2 ka has been reported globally (Magny et al., 2006) and resulted in wetter and/or cooler conditions in central and northern Europe, where the event is hypothesised to have resulted in considerable societal disruption (Magny and Haas, 2004; Caseldine et al., 2005). The event is also associated with a broader period of re-organisation in the global ocean-atmosphere circulation system between 6 – 5 ka (Mayewski et al., 2004; Wanner et al., 2008).

This study represents the first temporally focused examination of the 5.2 ka event in the ombrotrophic peatlands of northwest Europe. The climatic sensitivity and chronological potential possessed by these archives creates considerable potential for the reconstruction of past climate change through multi-proxy palaeoecological and novel stable isotopic analyses ($\delta^{18}O$ and $\delta^{13}C$). Despite this, few such studies exist beyond 4.5 to 5 ka in Ireland or neighbouring Great Britain (Charman et al., 2006; Swindles et al., 2010, 2013).

The 5.2 ka event is apparent in three peat-based $\delta^{18}O_{cellulose}$ records from across Ireland and its occurrence is supported by selected proxy data demonstrating a coherent shift towards wet conditions. It is suggested that this event was caused by a prolonged period of NAO^{+} conditions, resulting increased prevalence of cyclonic weather patterns over Ireland and an associated increase in precipitation.

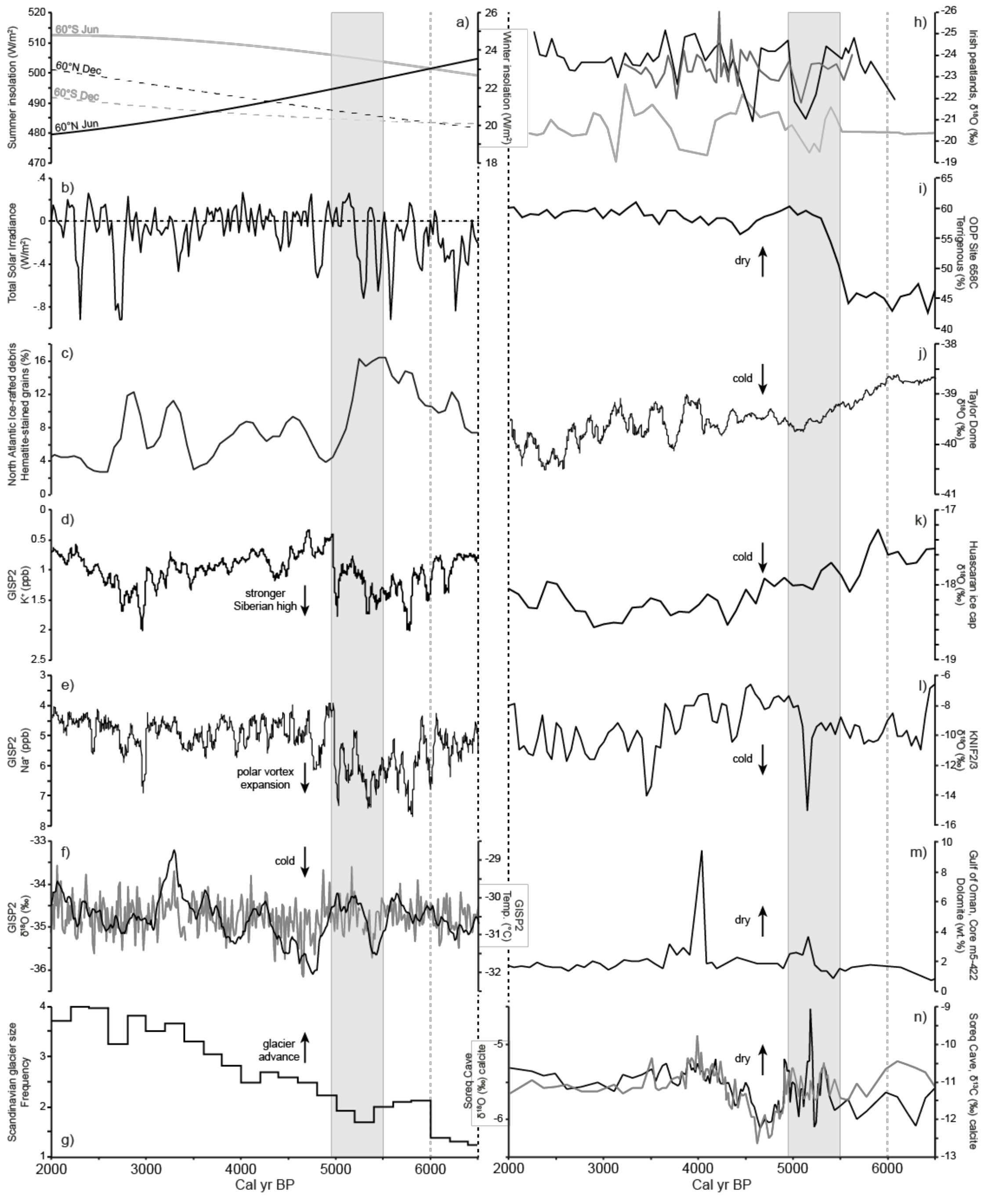

Figure 8. Global palaeoclimate records, 6.5 - 2 ka. a) Summer (solid) and winter (dashed) insolation from 60°N (black) and 60°S (grey) (Berger and Loutre, 1991); b) Total solar irradiance based on ^{10}Be ^{14}C in tree rings and polar ice cores (Steinhilber et al., 2012); c) North Atlantic hematite-stained grains record (stacked cores: MC52, V29191, MC21 and GGC22) (Bond et al., 2001); d) Potassium ion (K^+) content of GISP2 ice core (Mayewski et al., 1997); e) Sodium ion (Na^+) content of GISP2 ice core (O'Brien et al., 1995); f) Temperature over Greenland (GISP2) (Alley, 2000) derived from $\delta^{18}O$ (Stuiver et al., 1995); g) Frequency-distribution histogram (100-yr interval) of Scandinavian glacier-size variations (Nesje, 2009); h) $\delta^{18}O_{cellulose}$ data from Sluggan Moss (black), Fallahogy Bog (mid-grey) and Raheenmore (light grey) (this study); i) Terrigenous (aeolian) sediment percentage record from ODP Site 658C, West Africa (Demenocal et al., 2000); j) $\delta^{18}O$ record from Taylor Dome ice core, Antarctica (Steig et al., 2000); k) $\delta^{18}O$ record for Huascaran ice-cap, Peru (Thompson et al., 1995); l) $\delta^{18}O$ record for Kilimanjaro ice-core, Tanzania (Thompson et al., 2002); m) Percentage by weight of dolomite of Mesopotamian origin in marine sediments from the Gulf of Oman (Cullen et al., 2000); n) $\delta^{18}O$ (black) and $\delta^{13}C$ (grey) from a speleothem in Soreq Cave, Israel (Bar-Matthews et al., 2003).

The NAO has exhibited enhanced multi-decadal scale variability during the twentieth century (Goodkin et al., 2008) and whilst it is suggested that NAO conditions are likely to shift to a more positive mean state under anthropogenic warming scenarios (IPCC, 2013), the instrumental record is too short to assess this possibility fully (Timm, 2008). Proxy data that record the NAO either directly (e.g. $\delta^{18}O_{cellulose}$, this study) or indirectly (e.g. $\delta^{13}C_{cellulose}$ and other BSW proxies, this study) are therefore even more valuable to improve understanding of this complex process and its effects.

This study demonstrates the considerable palaeoclimatic value of stable isotopic analysis in ombrotrophic peat. This value could be further enhanced by the application of an optimised and standardised protocol, tackling methodological inconsistencies relating to sample selection, preparation and analysis (e.g. Daley et al., 2010; Loisel et al., 2010; Jones et al., 2014) and further research to address uncertainties as to the nature of biochemical pathways and isotopic signal preservation in *Sphagnum* (Kaislahti Tillman et al., 2010).

An examination of a number of global palaeoclimate records in addition to those presented in this study provides evidence of considerable variability (e.g. NAO, ITCZ, AMOC) and, in the case of ENSO, a shift to a new systematic regime, in ocean-atmosphere circulation systems between 6 – 5 ka, culminating in the termination of the HTM. The 5.2 ka event appears to be a component of this wide scale reorganisation – a non-linear and spatially complex response to broader forcing mechanisms. It is clear that further research is required to elucidate the precise drivers and effects of this climatically complex and archaeologically important period.

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