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1 **An 8000-year multi-proxy peat-based palaeoclimate record from Newfoundland:**
2 **Evidence of coherent changes in bog surface wetness and ocean circulation**
3

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17

18 **Abstract**

19 Energy carried by warm tropical water, transported via the Atlantic Meridional
20 Overturning Circulation (AMOC), plays a vital role in regulating the climate of regions
21 bordering the North Atlantic Ocean. Previous phases of elevated freshwater input to
22 areas of North Atlantic Deep Water (NADW) production in the early to mid-Holocene
23 have been linked with slow-downs in the AMOC and changes in regional climate.
24 Newfoundland's proximity in the North Atlantic region to the confluence of the Gulf
25 Stream and the Labrador Current and to an area of NADW production in the
26 Labrador Sea makes it an ideal testing ground to investigate the influence of past
27 fluctuations in ocean circulation on terrestrial ecosystems. We use multi-proxy peat-
28 based records from the east coast of Newfoundland to derive a proxy-climate signal
29 for the last 8000 years, which we have compared with changes in ocean circulation.
30 Prominent shifts towards near-surface bog water table levels, reflecting
31 cooler/wetter climatic conditions, are evident in the early-mid Holocene *c.* 7830,
32 7500, 7220 and 6600 *cal.* BP with minor changes occurring *c.* 6340, and 6110 *cal.* BP.
33 These events are coherent with evidence of meltwater injections into the N. Atlantic
34 and of reduced NADW production. More recent increases in bog surface wetness in
35 the mid-late Holocene *c.* 4290 and *c.* 2610 *cal.* BP are also consistent with reported
36 periods of reduced NADW production. Coherence between the bog-derived

37 palaeoclimate record developed from Newfoundland and evidence of fluctuations in
38 ocean current strength is apparent in the early mid-Holocene.

39

40 **Keywords**

41 Peatland, Newfoundland, palaeoclimate, Holocene, testate amoebae, hydrology,
42 macrofossils

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54 **Introduction**

55 Energy carried by warm tropical water, transported via the Atlantic Meridional
56 Overturning Circulation (AMOC), plays a vital role in regulating the climate of areas
57 bordering the North Atlantic Ocean (Marshall et al., 2001; Rahmstorf et al., 2015;
58 Swingedouw, 2015). The strength of the AMOC, and in particular the production of
59 North Atlantic Deep Water (NADW), has been found to vary through the Holocene at
60 centennial-millennial timescales (Bianchi and McCave, 1999; Oppo et al., 2003;
61 Hoogakker et al., 2011). Reductions in the AMOC strength are often concurrent with
62 increased Ice Rafted Debris, evidencing freshwater export from the Arctic to
63 Labrador and Nordic Seas (Bond et al., 2001).

64 Atmosphere-Ocean General Circulation modelling (Stouffer et al., 2006; Weaver et
65 al., 2012) and ocean sediment records have demonstrated that freshwater input in
66 to areas of NADW production in the Nordic Seas (Elmore et al., 2015) and the
67 Labrador Sea (Carlson et al., 2008; Wagner et al., 2013) can cause potential slow-
68 downs in the AMOC and knock-on effects to regional climate. This is typified by the
69 so-called '8.2 ka climatic event' (Barber et al., 1999; Daley et al., 2009, 2011; Wagner
70 et al., 2013). Meltwater released from glacial lakes, originally dammed by the
71 Laurentide ice sheet, entered the Labrador Sea causing reduced NADW and lower
72 temperatures in the North Atlantic area (Barber et al., 1999). Thereafter, from 7.5 to

6.4 ka, lower magnitude (>0.0015 Sv) pulsed injections of freshwater (up to 30 events) from glacial lakes in the Labrador region have been identified and noted as potentially cooling Labrador Sea Water (Jansson and Kleman, 2004).

Increased freshwater inputs to the ocean from global warming over the 20-21st century that may lead to an on-land cooling effect caused by a slow-down in the AMOC in the North Atlantic region would be nullified by the effect of the global warming trend itself Kuhlbrodt et al. (2009) suggest. However, underestimation of the sensitivity of the AMOC in models, to potential 21st century Greenland ice melt, means that this conclusion is not robust (Swingedouw, 2015). Changes in the strength of the AMOC may be critical for future climate and terrestrial environmental changes in the North Atlantic region (Rahmstorf et al., 2015). Evidence of past variability in terrestrial systems and possible influence from climate forcing mechanisms such as the AMOC are therefore crucial for the development and testing of models determining future climate projections and their influence on terrestrial ecosystems.

Newfoundland is located south of the Labrador Sea and west of the Grand Banks, a crucial junction between subtropical and sub-polar gyres of the North Atlantic where

92 the warm salty North Atlantic Current (NAC) and cool fresher water in the Labrador
93 Current (LC) meet and mix (Rossby 1999; Frantantoni et al., 2010). The Labrador Sea
94 is one of the few areas of open ocean deep convection and contributes to NADW
95 production (Marshall and Schott, 1999). Newfoundland's eastern coast is skirted by
96 the cold LC that exists as an inner and outer branch and transports up to two thirds
97 of the freshwater from the Arctic Ocean (Aksenov et al., 2010), potentially
98 influencing the strength of the NAC (Jones and Anderson, 2008) and therefore
99 climate in Western Europe. The present-day climate of eastern Newfoundland is
100 heavily influenced by these currents, especially the cold LC (Banfield, 1983), the
101 inner branch of which encircles the island and is comprised of water from the Baffin
102 and West Greenland Currents and Irminger Sea. The strength of the LC has been
103 linked to changes in atmospheric circulation, namely the Northern Annular Mode
104 (NAM), with positive phases associated with stronger north westerly winds, sea ice
105 formation, cooler SSTs and iceberg transport (Drinkwater, 1996; Sicre et al., 2014). A
106 stronger LC might be expected to reduce the influence of the Gulf Stream on the
107 climate of Newfoundland and move its position south of Newfoundland. The location
108 of Newfoundland in relation to these ocean currents makes it an ideal testing ground
109 to determine the influence of past fluctuations in their behaviour on terrestrial
110 ecosystems.

111

112 Ombrotrophic peatlands offer sources of decadal to millennial-scale climate records
113 for the Holocene (Aaby, 1976; Barber, 1981; Chambers et al., 2012; Galka et al.,
114 2015; Lacourse et al., 2015; Roland et al., 2014; Blundell et al., 2016) because their
115 surface hydrological conditions react to the balance between precipitation and
116 evapotranspiration, thus reflecting climate. These climate-driven changes in bog
117 water table levels are recorded when the macrofossil remains of plant communities,
118 and associated testate amoebae microfossils from pool, lawn and hummock
119 microforms are preserved in accumulating peat. Water table shifts can be inferred by
120 reconstructing the sequence of changes in past bog surface microforms.

121

122 Although most early peatland palaeoclimate research was carried out in Northwest
123 Europe, studies are now global (Nichols and Huang 2012; Daley et al., 2012; Novenko
124 et al., 2015). However, few investigations on the North Western Atlantic seaboard
125 exist (Hughes et al., 2006; Mackay, 2016; Daley et al., 2016; Peros et al., 2016)
126 despite the widespread occurrence of near-pristine peatlands. Peatlands are
127 abundant in Newfoundland covering 18% (20,000 km²) of the land area (Wells and
128 Pollet 1983), the oldest forming after the Newfoundland Ice Sheet retreated c. 9-10
129 ka (Dyke and Prest, 1987). The abundance of peatlands and its location (Figure 1)

make this area a valuable resource for exploring links between changes in ocean circulation, climate and terrestrial response. Initial work by Hughes et al. (2006) and subsequently by Daley et al. (2009) gave encouraging signs to suggest a link between terrestrial peatland ecosystems and changes in climate, regulated by changes in ocean circulation. Here we present a multi-proxy palaeo-record from Pound Cove Bog (PNDC) in Newfoundland and aim to further examine the link between past variability in terrestrial peatland ecosystems and evidence of variations in North Atlantic Ocean currents. During phases of early-mid Holocene meltwater discharges in to the Labrador Sea we aim to address the following. (1) Do palaeoecological records of water table variability in ombrotrophic bogs represent responses to changes in regional atmospheric moisture balance? (2) Are these responses coherent with the early-mid Holocene fluctuations in the AMOC variability, as seen in time series such as the IRD record (Bond et al., 2001)?

Site description

Pound Cove Bog (PNDC) is a slope bog located (Figure 1) on the western coast of Newfoundland 3 km northwest of Wesleyville on the Bonavista North Peninsula (53° 35' 44" W 49° 9' 59" N). The geological setting is granitic/gabbroic from the Ordovician to Carboniferous Periods and although much of Newfoundland was

covered in ice in the last glaciation, this area remained ice free (Rogerson, 1983). Slope bogs in eastern Newfoundland develop because of frequent fogs, due to the influence of the LC, and high summer precipitation levels (Damman, 1980) and, where the topography is water-shedding in all directions or where upslope snow melt is minimal, are ombrotrophic. The study site supports expansive carpets of oligotrophic Sphagna. *Sphagnum fuscum* occurs widely on hummocks, together with *Rubus chamaemorus*, *Kalmia* spp., *Chamaedaphne calyculata* and several *Cladonia* species. *Picea mariana*, *Pinus strobus* and *Larix laricina* are evident on drier bog hummocks, whilst wetter mud-bottomed hollows and pool edges are inhabited by *Rhynchospora alba* and *Eriophorum angustifolium*. *Sphagnum* section *Cuspidata* occurs in wet hollows and pools.

Field methods

A core was taken distant from expansive pools or hummocks that may prove insensitive to water table changes, in an area of deepest peat (c. 600 cm), from a lawn microform with extensive *Sphagnum*. Peat was recovered using a monolith tin (10 x 10 x 40 cm) from 0-40 cm and a wide-bore (9 cm) 30 cm long Russian corer (Barber, 1984) thereafter with overlaps of 5 cm. All core samples were placed in sealed plastic bags and refrigerated.

168

169 **Recovery, sampling and laboratory analyses**

170 *Recovery and sampling*

171 The core was sub-sampled for humification (2 cm³), macrofossil (4 cm³) and testate
172 amoebae (2 cm³) analyses. Sampling resolution was every 4 cm for testate amoeba,
173 every cm for humification and between 2 and 8 cm for macrofossils. Sampling
174 resolution for macrofossils was variable due to time constraints; lower resolutions
175 were employed at periods of *Sphagnum fuscum* domination.

176

177 *Chronology*

178 Depths for Accelerator Mass Spectrometry (AMS) radiocarbon dating were
179 determined after multi-proxy analyses were completed. Major changes in the
180 macrofossil stratigraphy were dated using 13 AMS radiocarbon dates
181 (Supplementary information (SI) Figure 1 and Table 1)). These stratigraphic
182 boundaries represent shifts in the average position of the bog water table; however,
183 they may also indicate points at which the peat accumulation rate changed. Even
184 within *Sphagnum*-dominated peats, accumulation rates can be variable (Aaby and
185 Tauber, 1974) because of differential growth and decay rates between species
186 (Johnson and Damman, 1991). In the phase dominated by *S. fuscum* evenly spaced

187 levels were dated. Sub-sampled 1-cm³ blocks of peat were washed with deionized
188 water in a 125 µm sieve and *Sphagnum* leaves, branches or stems were selected.
189 Non-contemporaneous material was removed to prevent possible reservoir effects
190 (Kilian et al. 1995). Samples were analysed by the NERC Radiocarbon Laboratory and
191 Beta Analytic Inc. (Table 1).

192
193 An age-depth model was produced using the ‘Bacon’ accumulation model (Blaauw
194 and Christen, 2011) in ‘R’ (R Core Team, 2012). Bacon uses Bayesian statistics to
195 determine Bayesian accumulation histories using radiocarbon dates and prior
196 information. Prior information regarding peat accumulation rate and its potential to
197 vary (SI Figure 1) are accounted for in the computations, potentially providing a more
198 realistic environmentally dependent age-depth model. Bacon output provides
199 estimates for every 1 cm interval of total chronological error with maximum and
200 minimum ages within the ‘modelled age range’ (MAR), together with ‘maximum age
201 probabilities’ (MAP), to provide the most likely date. Following Turner et al. (2014)
202 and Blundell et al. (2016), dates quoted in the text are MAP and are followed by
203 MAR values in subscript, where required. All radiocarbon dates quoted unless
204 otherwise stated are calibrated.

206 *Proxy analyses*

207 Macrofossil analyses were carried out using the Quadrat and Leaf Count technique
208 (QLC; Barber et al., 1994), aided by a reference collection of Newfoundland bog flora
209 held at the Palaeoecological Laboratory, University of Southampton. *Sphagna* were
210 identified using Daniels and Eddy (1990) and Bastien and Garneau (1997), whereas
211 for vascular plants and non-*Sphagnum* bryophytes, Robertson (2000) and Smith
212 (2004) were employed. A Hydroclimatic Index (HCI) was calculated by weighted
213 averaging ordination (Dupont, 1986; Daley and Barber, 2012), with each taxon
214 weighted based on their relative positions along the bog water-table gradient.
215 Previous ecological observations (Wells, 1981; Wells and Pollett, 1983) were used to
216 help define weightings, *Picea mariana* = 8, Erica wood and roots = 8, Unidentified
217 organic matter = 8, Monocotyledons (undifferentiated) = 7, Dicotyledons
218 (undifferentiated) = 7, *Eriophorum spissum* = 6, *Trichophorum cespitosum* = 5,
219 *Sphagnum fuscum* = 5, *S. capillifolium* = 5, *Dicranum undulatum* = 5, *Ledum*
220 *groenlandicum* = 5, *S. magellanicum* = 4, *S. flavicomans* = 4, *S. tenellum* = 3,
221 *Rhynchospora alba* = 3, *Drepanocladus fluitans* = 3, *S. pulchrum* = 2, *S. section*
222 *Cuspidata* = 1. Small macrofossils of fruits and seeds were excluded and also
223 macrofossils with low abundance (<10% cumulatively across all levels) together with
224 shrub leaves.

225

226 Analyses to determine the humification of the peat were carried out in accordance
227 with Blackford and Chambers (1993). This involved chemical extraction of humic
228 matter, which was measured for light absorbance in a spectrophotometer at a
229 wavelength of 550 nm. Humification records can be affected by differential decay of
230 vegetation type (Caseldine et al., 2000; Yelhoff and Mauquoy, 2006; Hughes et al.,
231 2012) and may exhibit signal overwriting and temporal lags as a result of secondary
232 decomposition (Morris et al., 2015). This is partly mitigated here by producing a
233 detailed macrofossil diagram to carefully interpret results. The commonly used
234 extraction/measuring technique for peat humification has been questioned
235 (Caseldine et al., 2000; Biester et al., 2014). Biester et al. (2014) found poor
236 correlation between UV-absorption of alkali extractions and other techniques used
237 to examine peat decomposition, such as Pyrolysis GC-MS, C/N ratio, FTIR band
238 intensities and $\delta^{13}\text{C}$ and proposed that changes in the humic acids determined via UV
239 –ABS analysis of alkali extracts may reflect changes in vegetation taxa.

240

241 Samples for testate amoebae analysis were prepared and counted in accordance
242 with Charman et al. (2000). The taxonomy used here is the same as that used in the
243 transfer function applied to PNDC data and therefore taxonomic inconsistencies are

not present (Payne et al. 2010). The transfer function used was developed from North American peatlands (Booth, 2008). Counts of 100 tests are considered sufficient to be representative (Payne and Mitchell, 2009) and this was achieved for all samples. Testate amoeba records can be subject to differential preservation (Wilmhurst et al., 2003, Swindles and Roe, 2007; Mitchell et al., 2008) and some concerns exist regarding associated transfer functions employed, including spatial autocorrelation (Telford and Birks, 2009), uneven sampling of the environmental gradient (Telford and Birks, 2011) and the use of clustered datasets (Payne et al., 2012). Swindles et al. (2015) suggests that most available testate transfer functions are poor at reconstructing absolute values of mean depths to water tables but are reliable in terms of shifts in direction to wetter or drier conditions. Swindles et al. (2015) recommend reporting standardised values to avoid confusion with contemporary water-table data that report reliable magnitudes. We report absolute and standardised water-table values to aid comparison with other studies. Testate amoebae have a short life cycle and limited mobility and therefore may react to hydrological change more rapidly than plant macrofossils or humification leading to potential temporal mismatches (Väliranta et al., 2012).

Results

Plant Macrofossils and chronology

From c. 8070 BP MAR 7936-8207 BP until c. 6660 BP MAR 6534-6816 BP (PNDC a - e) accumulation is relatively low at 14.9 yrs cm⁻¹ as *Picea mariana*, *Carex*, Ericaceae and *Sphagnum fuscum* are the dominant macrofossils, suggesting dry conditions (Figure 2). However, within this period there are two brief phases when *S. magellanicum* is abundant c. 7830 BP MAR 7626-8014 BP and c. 7220 MAR 7052-7365 BP, reflecting changes to wetter conditions. *Drepanocladus fluitans* is abundant c. 6600 BP MAR 6472-6760 BP before a major shift to *Sphagnum* dominance c. 6500 BP MAR 6360-6667 BP up to the surface. The switch from a system that supports *Picea* to one dominated by *Sphagnum* represents a major shift in the development of the site, reflecting progression to more stable water table conditions and to full ombrotrophy as taxa such as *S. fuscum* (Wells and Pollett, 1983) proliferate.

From c. 6500 BP MAR 6360-6667 BP to c. 5530 BP MAR 5378-5634 BP a series of successions between *Sphagnum* taxa exist with an associated increase in mean accumulation to c. 8.3 yrs cm⁻¹. Initial succession from *S. pulcrum* to *S. magellanicum* to *S. capilifolium* is indicative of a shift from a near-surface to a deeper water table.

Thereafter, c. 6340 BP MAR 6198-6497 BP *S. fuscum* peaks and is followed by increases in *R. alba*, *S. magellanicum* and *S. flavicomans*, representing a return to wetter conditions. Subsequent drier conditions, as evidenced by *Trichophorum cespitosum*, are succeeded by a return to *S. magellanicum* and *S. flavicomans* c. 6110 BP MAR 6004-6237 BP, the latter being dominant in zone PNDC-j and indicative of wetter conditions (Wells and Pollett, 1983).

From the mid-late Holocene, c. 5400 BP MAR 5216-5549 BP - 1060 BP MAR 953-1199 BP, the core is dominated by *S. fuscum*, its abundance rarely below 70% (PNDC-k – m). However, notable occurrences of other taxa include an increase in *Dicranum undulatum* (*D. bergerii*), a species that forms tufts in relatively dry conditions (Robertson, 2000), c. 2900 BP MAR 2657-3144 BP and peaks in *S. flavicomans* c. 2460 BP MAR 2303 - 2660 BP, 1920 BP MAR 1790 - 2048 BP and 1580 BP MAR 1307 - 1837 BP, indicating increased wetness. After c. 1060 BP MAR 953 - 1199 BP *S. fuscum* abundance decreases as *S. pulchrum* increases c. 930 BP MAR 658 - 1110 BP – 630 BP MAR 372 - 899 BP suggesting a return to a near surface water table level. Post c. 630 BP MAR 372 - 899 BP *S. pulchrum* is succeeded by *S. magellanicum* and continued *R. alba* presence until c. 230 BP MAR 117 - 399 BP. Thereafter, *S. fuscum* dominates to the present day.

Sphagnum domination c. 6480 MAR 6334 - 6476 BP – 3420 BP MAR 3258 – 3538 BP results in a mean accumulation rate of c. 10.1 yrs cm⁻¹. However, a subsequent change in the bog's development to c. 230 BP MAR 117 - 399 BP leads to a decline in the mean accumulation rate to c. 23.4 yrs cm⁻¹ despite a continued dominance of *Sphagnum*. Between 48- 24 cm (c. 1060 BP MAR 979-1170 BP – 250 BP MAR 128-415 BP) accumulation is particularly slow at 38.6 yrs cm⁻¹. A reduced accumulation rate here may be partly explained by episodes of suboptimal conditions for peat accumulation, such as the development of a pool as evidenced by *S. pulchrum* and *R. alba* in Zone PNDC-n. Taxa associated with pools have been recorded as exhibiting high rates of decay (Johnsen and Damman, 1991; Belyea, 1996; Limpens and Berendse, 2003). From c. 630 BP MAR 372-899 BP (36 cm) *S. magellanicum* and subsequently *S. fuscum* dominate to the surface.

Humification

Periods dominated by *Sphagnum* display low levels of decay, whereas assemblages containing more degradable plant matter, with a lower nitrogen concentrations (Coulson and Butterfield, 1978), such as monocotyledon or Ericaceae remains, register higher levels of humification. Phases are apparent in the record where complex fluctuations occur around a broadly stationary average humification value

(Figure 3). For example, Zones PNDC a-e, containing *Picea*, monocots and UOM remains, show high residual absorbance values compared with zones f-h, which are lower as *Sphagnum* moss becomes abundant. From PNDC-i to midway into PNDC-k a lower average absorbance residual is evident, again commensurate with greater abundance of *Sphagnum*. From 250 cm depth to the start of PNDC-l, average absorbance residuals increase and further increase in PNDC l-m. Major fluctuations in humification in these zones are often associated with increases in monocotyledon and Ericaceae remains that have a disproportionate effect on humification values, representing evidence of ‘species-dependent effects’ (Yeloff and Mauquoy, 2006). Major changes to low absorbance that cannot be accounted for by shifts in the colouration of the contributing plants and potentially reflect water table variability caused by climate change are shown in Supplementary Information (SI) Table 1.

Testate amoebae

Dry-indicating testate amoebae including *Hyalsophenia subflava*, *Nebela militaris/minor* and *Trigonopyxis arcula* (Booth, 2008) are prevalent (Figure 4) in Zones PNDC-a – PNDC-e (c. 8020 BP_{MAR} 7865-8163 BP – 6600 BP_{MAR} 6472-6759 BP). However, increases in hygrophilous taxa *Archerella flavum* occur, leading to relatively wetter conditions and lower reconstructed depth to water tables (DTWTs) c. 7830 BP_{MAR}

338 7626-8014 BP, c. 7500 BP MAR 7336 – 7694 BP, and c. 7220 BP MAR 7053-7366 BP. PNDC-e is
 339 dominated by *N. militaris/minor* reflecting drier conditions (Amesbury et al., 2013).
 340 This supersedes a substantial change to wetter conditions as both *Archerella flavum*
 341 and *Amphitrema wrightianum* increase c. 6600 MAR 6472-6759 BP, the latter peaking c.
 342 6480 BP MAR 6334-6643 BP. Further switches from *N. militaris/minor* and *H. subflava* to
 343 the hygrophilous *Archerella flavum* and *Amphitrema wrightianum*, occur at the
 344 lower boundary of PNDC-h c. 6340 MAR 6198-6497 BP, and 6110 MAR 6004-6237 BP. PNDC-I and
 345 PNDC-j are dominated by *Archerella flavum* and only c. 5370 BP MAR 5183 -5531 BP do
 346 more xeric taxa such as *N. militaris/minor* and *T. arcuata* return. In PNDC-k four
 347 substantial increases in *Archerella flavum* occur c. 5010 MAR 4815 -5198 BP, 4500 MAR 4311 -
 348 4669 BP, 4290 MAR 4117-4477 BP, 4010 MAR 3889 -4145 BP and 2950 MAR 2680 -3187 BP and these
 349 events are associated with declines in the xeric taxa *H. subflava*, *N. militaris/minor*,
 350 *T. arcuata* and *Diffugia pulex*, suggesting changes to wetter conditions. Increases in
 351 *Archerella flavum* and *Amphitrema wrightianum* are also evident in PNDC-I c. 2610
 352 BP MAR 2428-2830 BP and c. 2110 BP MAR 1974-2319 BP. These changes are followed by elevated
 353 levels of the xeric *H. subflava*, commencing c. 1920 MAR 1790 -2048 BP. Subsequent
 354 increases in *Archerella flavum*, and *Amphitrema wrightianum* and decline in *N.*
 355 *militaris/minor* and *T. arcuata* are evident c. 1480 MAR 1219-1755 BP, 1250 MAR 1060 -1490 BP
 356 and c. 290 MAR 161 -467 BP, indicating a return to wetter conditions. Substantial changes

to wetter conditions (from high reconstructed DTWTs to low values) are displayed in SI Table 1.

Discussion

Comparisons of proxy palaeoclimate signals

Proxy techniques used in this study for palaeo-hydrological reconstruction require careful cross-examination since ‘palaeoclimate’ signals derived can be influenced by other non-climatic factors (Blundell and Barber, 2005; Swindles et al., 2012).

Reconstruction of bog surface wetness (BSW) should be informed by all available sources of evidence where possible (Blundell and Barber, 2005). Changes to wetter conditions evident in each proxy and when at least two of the three proxies are in agreement (Figure 5) is stated in SI Table 1.

Standardised values for outputs from three testate amoebae transfer functions, humification and macrofossil HCl are compared to highlight similarities and differences between the palaeoclimate proxies at PNDC (SI Figure 2). Transfer functions from Booth (2008) and Amesbury et al. (2013) show good correspondence ($r^2 = 0.696$, $p < 0.0001$), but data from using the Charman and Warner (1997) function do not show good correspondence with the other two functions of Booth

(2008) ($r^2 = 0.058$, $p = 0.004$) and Amesbury et al. (2013) ($r^2 = 0.180$, $p < 0.0001$). At Nordan's Pond Bog (Newfoundland) Hughes et al. (2006) used the Newfoundland-specific transfer function (Charman and Warner, 1997) to reconstruct water table depth; however, its use highlighted concerns related to the representation of specific taxa such as *Hyalosphenia papilio*, and *H. elegans*. In the training set these were modelled (Charman and Warner, 1997) as having relatively high DTWT optima, and resultant DTWT reconstructions displayed 'conflicts' with the other proxies. Optima were substantially higher than those subsequently derived from other European (Charman et al., 2006; Amesbury et al., 2016), British (Woodland et al., 1998), North American (Booth, 2008) and north-eastern Canada and Maine (Amesbury et al., 2013) based functions. This, in part, led to the poor correlations with the other proxy indicators (humification $r^2 = 0.007$, $p = 0.321$; HCl $r^2 = 0.03$, $p = 0.041$) and therefore the Charman and Warner function is not employed.

The macrofossil HCl data exhibit low variability (SI Figure 2a-d) owing to the dominance of *S. fuscum*. Variability that is present is concentrated at the high (dry) and low (wet) extremes and it is broadly in agreement with the extremes of both humification and testate amoebae derived BSW data (SI Figure 2 and SI Table 2). The highest HCl correlation (SI Table 2) is with the testate amoebae transfer function

from Booth (2008) ($r^2 = 0.401$, $p < 0.0001$). Even when the record is dominated by *Sphagnum fuscum*, peat humification values vary considerably, suggesting that these are not species-dependent changes but more likely reflect changes in bog water table. Some peaks in humification are concurrent with small amounts of low-C/N plant material (Coulson and Butterfield, 1978), which can deliver a disproportionate contribution to the peat decay signal (Yeloff and Mauquoy, 2006) owing to their ease of decomposition. The early macrofossil record before c. 5400 BP MAR 5216-5549 BP, which is not dominated by *S. fuscum*, is in close agreement with both testate and humification signals. Thereafter, the macrofossil record is hydrologically insensitive.

The humification record, once reduced to 4 cm resolution by removing values to align it with that of the testate amoebae DTWT reconstruction, displays the closest correspondence ($r^2 = 0.257$, $p = <0.0001$) with the DTWT reconstruction based on the Booth (2008) transfer function, followed by that from Amesbury et al. (2013) ($r^2 = 0.143$, $p < 0.0001$) and a poor correspondence with the function devised by Charman and Warner (1997). The testate transfer functions from Booth (2008) and Amesbury et al. (2013) give similar outputs but the former displayed most correspondence with other proxies and is therefore used here.

Particular differences between the humification record and the DTWT reconstructed from testate amoebae (derived from Booth, 2008) are evident c. 1060 BP MAR 953-1199 BP – 1190 BP MAR 1039-1378 BP, c. 2330 BP MAR 2128-2562 BP, c. 2910 BP MAR 2657-3145 BP, c. 3880 BP MAR 3721-4029 BP, c. 3950 BP MAR 3817-4089 BP, c. 4150 BP MAR 3997-4317 BP and require interpretation. In these instances humification values are relatively high, whereas reconstructed DTWTs from testate amoebae data are relatively low (high water table). Examination of the macrofossil record reveals that these are periods of pool mud deposition or there is evidence of a small percentage of more degradable material such as *Rhynchospira alba*, both of which are indicative of wet conditions. From c. 4890 MAR 4713-5063 BP – 5280 MAR 5073-5458 BP BP and c. 7860 BP MAR 7653-8037 BP the converse applies as *Sphagnum* dominates but high percentages of the testates *H. subflava* and *T. arcuata* point to a low water table. A potential lag between humification and DTWT reconstructed from testate amoebae is evident c. 2980 BP MAR 2720-3212 BP and c. 4520 BP MAR 4329-4684 BP, as the latter appears to react a sample before the humification record. Changes to wetter conditions evident in each proxy and when at least two of the three proxies are in agreement (Figure 5) is stated in SI Table 1.

In the last few decades, changes in vegetation, reconstructed DTWTs, and humification have largely been interpreted as the result of allogenic factors, predominantly changing climate. However, anthropogenic disturbances can affect the bog hydrological conditions, through burning, artificial drainage, peat cutting and atmospheric pollution. The role of autogenic factors has been revisited (Swindles et al., 2012; Morris et al., 2015; Waddington et al., 2015) and highlighted through the differences displayed between multiple cores from the same site (Blaauw and Mauquoy, 2012; Mathijssen et al., 2016), suggesting that internal ‘noise’ exists within the fossil proxy records together with a climate signal. Infilling of pools and increasing DTWTs can, for example, occur as a result of peat accumulation under stable climate conditions (Aaby, 1976). Although peatland water table levels are sensitive to climate change, the peatland archive can be affected by non-linear complex internal responses (Swindles et al., 2012). Determining climate signals can be achieved by examining multiple cores at a site or from adjacent sites, although this approach is costly and not always practical. However, similarities between other well-dated records can provide increased confidence that changes in peat stratigraphy are externally driven.

Comparison of PNDC with Nordans Pond Bog (NDN)

451 The peat-based palaeoclimate record at NDN (Hughes et al., 2006) is located within
452 two kilometres of PNDC. Coherent changes should reflect climate as opposed to
453 autogenic influences. Since publication (Hughes et al., 2006) the age–depth model
454 has evolved (Daley et al., 2009; 2011) and tephra layers have been discovered (Pyne-
455 O'Donnell et al., 2012) at NDN providing additional dating. Five of the tephras found
456 have a reliable date assigned to them and have been used here to improve the
457 original age–depth model (SI Figure 3). To permit comparison with PNDC, testate
458 amoebae-derived DTWTs from NDN have been recalculated using the transfer
459 function of Booth (2008). Here we have compared the two testate amoebae records,
460 as these are the most robust of the three proxies; the macrofossil record post c.
461 5400 BP MAR 5216-5549 BP at PNDC is hydrologically insensitive and the humification data
462 can be affected by site specific changes in vegetation requiring more protracted
463 consideration. Differences in sampling resolution, allied with potential chronological
464 error, make comparison challenging. However, from the early Holocene to c. 4000 BP
465 nine of the 10 most prominent changes in the PNDC testate amoebae record can be
466 associated with similar events at NDN, well within dating errors (Figure 6 and SI
467 Table 1). Thereafter, although variation in the NDN record is relatively low some
468 prominent changes to wetter conditions at PNDC can be linked to changes at NDN
469 within dating errors. The substantial change at PNDC c. 1480 BP MAR 1219-1755 BP is just

within dating errors of a similar change at NDN (c. 1780 BP MAR 1646-1905 BP), but matching these is less robust. Although there are similarities in hydrological changes between the two study sites suggesting a climate-forcing mechanism was acting on these two adjacent but hydrologically-separated peatland systems, the chronological control is insufficient to conclude definitively.

The palaeoenvironmental record at PNDC and comparisons with proxies of ocean circulation and terrestrial paleoclimate records.

Mid Holocene (8000 – 5000 BP) The PNDC record displays a series of prominent increases in bog water table levels to wetter conditions in the mid-Holocene c. 7830 BP MAR 7626-8014 BP and 7500 BP MAR 7336 – 7694 BP, followed by further episodes c. 7050 MAR 6847-7247 BP and 6480 MAR 6341-6653 BP. Comparisons with wider oceanographic evidence from the eastern seaboard region, discussed below, suggest that these changes in BSW on the Bonavista North Peninsula are most likely the result of reduced evapotranspiration from cooler climatic conditions (Figure 5 and 7) and/or elevated levels of precipitation. Evidence presented by Jansson and Kleman (2004) suggests that the eighth millennium BP was a period of heightened sensitivity of the AMOC as numerous meltwater injections (up to 30 over 0.015Sv), from remnant glacial lakes

489 on the North American continental margin, into Ungava Bay and other points in the
490 Labrador Sea took place. Six meltwater events (>0.015 Sv, based on 30-day duration)
491 c. 7500 to 7000 BP are reported to have occurred. Despite relatively low outflow
492 volumes, the pulsed nature of these freshwater discharges could have led to
493 increased sensitivity of the AMOC, thus altering the degree of NADW production in
494 the Labrador Sea and the degree of heat transport north. Global ocean circulation
495 models suggest that small volumes (<0.06 Sv) of freshwater input are sufficient to
496 cause a shutdown of convection in the Labrador Sea (Rahmstorf, 1995). The
497 existence of substantial meltwater inputs to the Labrador Sea in this period is
498 supported by evidence of abundant *Neogloboquadrin pachyderma* from ocean
499 sediments off Nova Scotia c. 7.1 ka BP (Keigwin and Jones, 1995) and a $\delta^{18}\text{O}$
500 minimum in the carbonate shells of the foraminifera from these sediments. Both of
501 these features of the ocean record have been interpreted as a consequence of the
502 input of meltwater from either the Great Lakes or Labrador-Ungava region. Evidence
503 of reduced ocean circulation in the North Atlantic Ocean between c. 7600-7000 BP
504 can also be inferred from decreasing mean sortable silt sizes from cores in areas of
505 North West Atlantic Bottom Water circulation east of Newfoundland (Hoogakker et
506 al., 2011), in Iceland Scotland Overflow Water (ISOW), in the north eastern Atlantic
507 (Bianchi and McCave, 1999; Figure 7d) and from declining sea surface temperatures

(SSTs) in the Icelandic Sea (Bendle and Rosell Mele, 2007). Sheldon et al. (2015) concluded from dinoflagellate records in Eastern Newfoundland that between c. 7.2–5.5 ka, SSTs were dominated by cold water sourced from the Arctic via the LC and greater sea ice and iceberg transport resulting from the increased strength and influence of the LC. Dominance of colder waters off the coast must have affected the local climate near the PNDC site, most likely lowering air temperatures and decreasing evapotranspiration, leading to increases in bog water table levels.

The most substantial and abrupt increase in BSW at PNDC occurs c. 6600 BP MAR 6472–6759 BP, peaking c. 6480 BP MAR 6334–6643 BP with two further minor increases c. 6340 BP MAR 6198–6497 BP and 6110 BP MAR 6004–6237 BP (Figure 5). Thereafter, until c. 5200 BP the record implies continuity of wet bog surface conditions. There is evidence to suggest that this was also an important period of climate change globally and one of changing AMOC and solar activity (Wanner et al., 2011). Before and during this period further meltwater discharge events from eastern North America (between 7.0 – 6.4 ka) directly into the Labrador Sea and/or Ungava Bay region are reported (Jansson and Kleman 2004). These events may have been large enough to reduce NADW production. Increased advection of cooler surface waters, driven by northerly winds from the Nordic and Labrador Seas, further south into the Atlantic is

suggested by a fluctuating but overall increasing level of IRD (Figure 7a) from 7200–5500 BP in the Atlantic Ocean (Bond et al., 2001). In the eastern North Atlantic the importance of this period is clear; from the record of $\delta^{13}\text{C}_{\text{Calcite}}$ in benthic foraminifera (Oppo et al., 2003) a major reduction in NADW contribution between c. 6.5 ka and 5.0 ka is inferred and this feature, producing cooler climatic conditions, is considered to be the most pronounced isotopic event in the Holocene section of the record (Figure 7c). SSTs around Iceland are also reported to have reduced substantially at this time (Anderson et al., 2004; Bendle and Rosell Mele, 2007), together with increasing sea ice (Cabedo-Sanz et al., 2016) and strength of nearby ocean currents, such as the ISOW (Bianchi and McCave, 1999) c. 6.4 ka (Figure 7d). West of Greenland, periods of cooler sea surface temperatures c. 6500 BP have been demonstrated (Hald et al., 2007) and attributed to southward displacement of Arctic waters. The period c. 6600–5200 BP was clearly one of abrupt changes in ocean circulation in the N. Atlantic often providing periods of cooler climatic conditions, and these are reflected in the record at PNDC.

Although evidence of climate change in the neighbouring NDN bog surface wetness record is less clear than that at PNDC, much evidence exists to suggest that this period of changing ocean circulation between c. 6700–5200 BP was paralleled by a

significant period of change on the Eastern Canadian/American continent. Wetter conditions are evident c. 6800 BP to c. 6500 BP and c. 5800 to 5500 BP in a record derived by Nichols and Huang (2012) of the *Sphagnum*/vascular plant ratio derived from the abundance of biomarker *n* – alkanes from a raised coastal bog to the south of PNDC in Maine, closely reflecting changes seen in the testate amoebae record from PNDC. Sediments of Eastern Lake Ontario in the continental interior also exhibit distinct palaeoenvironmental changes between 6.3 and 6.0 ka ending at 5.0 ka with decreases in % Organic Carbon and % Total Nitrogen and $\delta^{15}\text{N}$ (Mc Fadden et al., 2005), reflecting lower lake productivity related to increasing lake levels at this time – an event known as the ‘Nipissing rise’. Although often interpreted as a consequence of water moving from the Upper Great Lakes to the Lower Great lakes following isostatic uplift, Booth et al. (2002) noted that a wetter climate prevailed at the time of Nipissing Rise and that increased atmospheric moisture availability was an important factor. Elevated precipitation between c. 6600 BP and 5000 BP has also been reconstructed by lake level and pollen records to the east in the St Lawrence lowlands (Muller et al., 2003).

Mid - Late Holocene (5000 – 3000 BP) The PNDC record displays long-term (millennial) lowering of BSW (drier surface conditions) c. 5000 – 3000 BP. However,

565 numerous centennial-scale variations occur, some of which can be tied within dating
566 errors to changes in the NDN testate record (e.g. c. 5010 MAR 4815-5198 BP, 4500 BP MAR
567 4311 -4669 BP, 4290 BP MAR 4117-4477 BP) and disruption in ocean circulation. Although
568 evident, these changes appear to be of smaller scale compared with those observed
569 earlier in the Holocene c. 8000–5200 BP (Figure 5 and 7). Increased NADW
570 production, as interpreted by Oppo et al. (2003) from benthic $\delta^{13}\text{C}$, is evident
571 between c. 4800–3000 BP, and coincides with the general trend towards lower BSW
572 at the PNDC site (Figure 7c). Decreasing mean sortable silt sizes in sediments,
573 reflecting reduced North West Atlantic Bottom Water circulation east of
574 Newfoundland (Hoogakker et al., 2011), are evident c. 4800 BP and c. 4500–3900 BP.
575 Less vigorous ocean circulation is also inferred in the eastern North Atlantic since
576 reductions in ISOW (Figure 7d) are recorded c. 4800 BP and c. 4400 with a low c.
577 4200 BP (Bianchi and McCave, 1999) and evidence of increasing IRD c. 4800 BP to
578 highs c. 4600 BP and c. 4200 BP exists (Bond et al., 2001). Increases in IRD,
579 suggesting cooler SSTs are further supported by Andersen et al. (2004) further south
580 in the N. Atlantic and possible slowdowns of NADW. Reports of climate change c.
581 4200 BP in the palaeoenvironmental literature are frequent (Mayewski et al., 2004;
582 Booth et al., 2005), with drier conditions evident at mid-lower latitudes and some
583 evidence of increased wetness in more northerly higher latitudes, although the latter

appears to be less coherent (Roland et al., 2014). The period around 4200 BP is also often reported as the onset of neoglaciation in the Northern Atlantic region (Larsen et al., 2012; Balascio et al., 2015) as ice caps and glaciers re-advance. A change to cooler conditions (c. 4600-4200 BP) has also been recorded in $\delta^{18}\text{O}$ from cellulose from *Sphagnum* from Mer Bleu bog in Ontario (Bilali et al., 2013) and $\delta^{18}\text{O}$ from lake sediments in Newfoundland (Finkenbinder et al., 2016) point to the onset of wetter conditions c. 4300 BP, which is in agreement (Figure 6) with elevated BSW at PNDC and NDN.

Late Holocene (3000 BP – present) Although numerous changes in the PNDC testate amoebae record exist in the Late Holocene, those that can both be replicated by at least one other proxy at PNDC and can be well matched, within dating errors, at NDN, are limited to one event c. 2610 MAR 2428-2830 BP (SI Table 1). The substantial decline in DTWT and the other proxy values c. 1480 BP MAR 1219-1755 BP in PNDC can be replicated within dating errors at NDN but it is at the extreme of the error range, reducing confidence in the coherence of this second event between the two bogs. If changes registered in the testate amoebae record from PNDC are used alone without corroboration with other proxies from PNDC then the reduction in DTWT c. 2080 MAR 1852-2260 BP is also evident within dating errors at NDN.

603

604 Substantial evidence of global climate change exists between c.2.8–2.6 ka (Wanner
605 et al., 2011) and this has been linked to both ocean circulation and solar activity (van
606 Geel et al., 1999; Chambers et al., 2007). This was a period of reduced NADW
607 production (Oppo, et al., 2003; Bianchi and McCave, 1999), greater IRD (Bond et al.,
608 2001) (Figure 7a) and evidence of migration of arctic waters south (Dourain et al.,
609 2015) and changing atmospheric circulation (O’Brien et al., 1995). Ice caps and
610 glaciers to the north, around Baffin Island (Miller et al., 2005) and Greenland,
611 advanced (Levy et al., 2014; Balasico et al., 2015) and peatland records in
612 Newfoundland (Hughes et al., 2006) and mainland Canada/North America (Nichols
613 and Huang, 2012; Bilali et al., 2013) point towards cooling and/or increased wetness.
614 The PNDC record therefore helps to build the evidence base that suggests there was
615 an increase in atmospheric moisture availability c. 2.8 – 2.6 ka on the eastern
616 seaboard of North America at the time of a major global palaeoenvironmental
617 change.

618

619 Conclusions

620 1) Multi-proxy methods have been used to derive a detailed BSW record for the
621 last 8000 years for PNDC in Newfoundland. Comparisons with a neighbouring

622 record at NDN suggests similar timings of water table variability in the period
623 pre-4000 BP, implying a common climate forcing mechanism was a
624 prominent driver of water table level at both sites. However, a common
625 water table response is less clear after 4000 BP, possibly because of
626 insufficient chronological control; however, the possibility that local or
627 autogenic processes (*sensu* Swindles *et al.* 2012) masked low magnitude
628 climate signals in this period cannot be discounted.

629 2) Pronounced evidence of increased BSW in the PNDC record is coherent with
630 meltwater pulses and evidence of reduced NADW activity in the early to mid-
631 Holocene, especially c. 7500-6500 BP, suggesting terrestrial responses to
632 substantial ocean circulation changes at this time.

633 3) In the mid-late Holocene the peatland palaeoclimate records appear to be
634 noisier and a distinctive pattern of BSW variability that links to ocean
635 circulation is not clear. There is some apparent coherence between increased
636 BSW c. 2610 and 4210 BP in PNDC and evidence of reduced AMOC at these
637 times. These events are two of the high-magnitude palaeoclimate anomalies
638 of the mid to late Holocene. To better recognise the remainder of late
639 Holocene regional peat-based palaeoclimate signal of Eastern Newfoundland
640 improved chronologies pinned with tephras, where possible, and suites of

641 water table constructions from hydrologically-separated sites located within
642 distinct, well-defined spatial climate zones are required.
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Figure Captions

Figure 1. Location map of Pound Cove Bog and Nordan's Pond Bog.

Figure 2. Macrofossil diagram for PNDC. Peat Components are derived from averaged quadrat counts (15) under low power magnification (x10). The Leaf Counts are a breakdown of % Identifiable *Sphagnum* and consist of proportions based on a random selection of leaves (>100 per sample interval) identified at high magnification (x400). HCI indices are displayed.

Figure 3. a) Raw and b) detrended humification data for PNDC at 1 cm sampling resolution. Macrofossil zones are superimposed to aid comparison.

Figure 4. Testate amoebae diagram for PNDC. All data are percentages of the total number counted per level. Additional non-testate data such as abundance of *Alona rustica* and *Copepod* spermatophores are absolute counts. Transfer function (Booth, 2008) depth to water table level reconstructions are displayed.

Figure 5. Comparison of a) macrofossil derived dupont index, b) humification (grey line 1 cm resolution, black line 4cm resolution), c) DTWT from testate data. Grey shading denotes ‘wetter periods’ based on at least two proxies.

Figure 6. Comparison of PNDC and NDN testate amoebae reconstructed DTWTs. Grey shading represents 95% probability intervals from the age depth model. Black and grey dashed lines between sites records represent events at PNDC that are coherent and well within dating errors at NDN and those that are less robust respectively. SI Table 1 lists details of the dates for each event and associated errors.

Figure 7. Comparison of a) IRD combined record (Bond et al., 2001), b) PNDC DTWT record, c) $\delta^{13}\text{C}_{\text{Calcite}}$ of benthic foraminifera (Oppo et al., 2003) and d) ISOW record (McCave and Bianchi 1999).

Table captions

Table 1. AMS radiocarbon dates from Pound Cove Bog (PNDC) core. Dates were calibrated using IntCal13 (Reimer et al., 2013) within the ‘Bacon’ computer program (Blaauw and Christen, 2011).

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Figure 1.

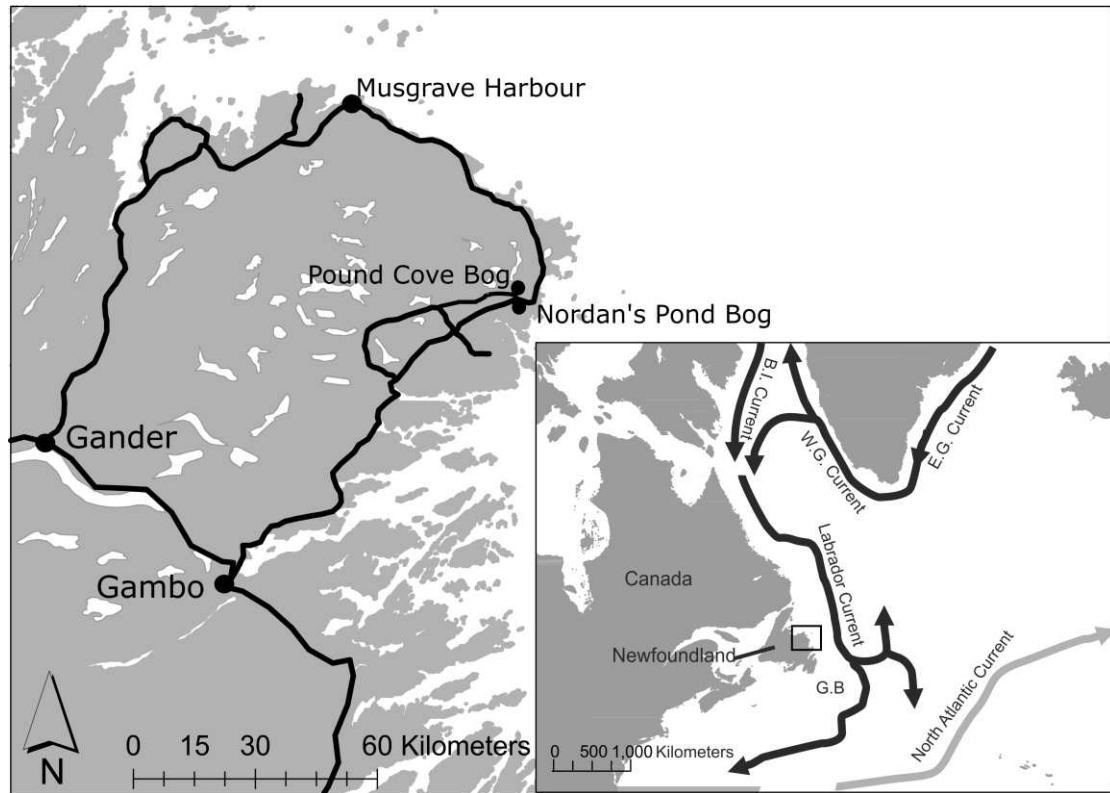


Figure 2

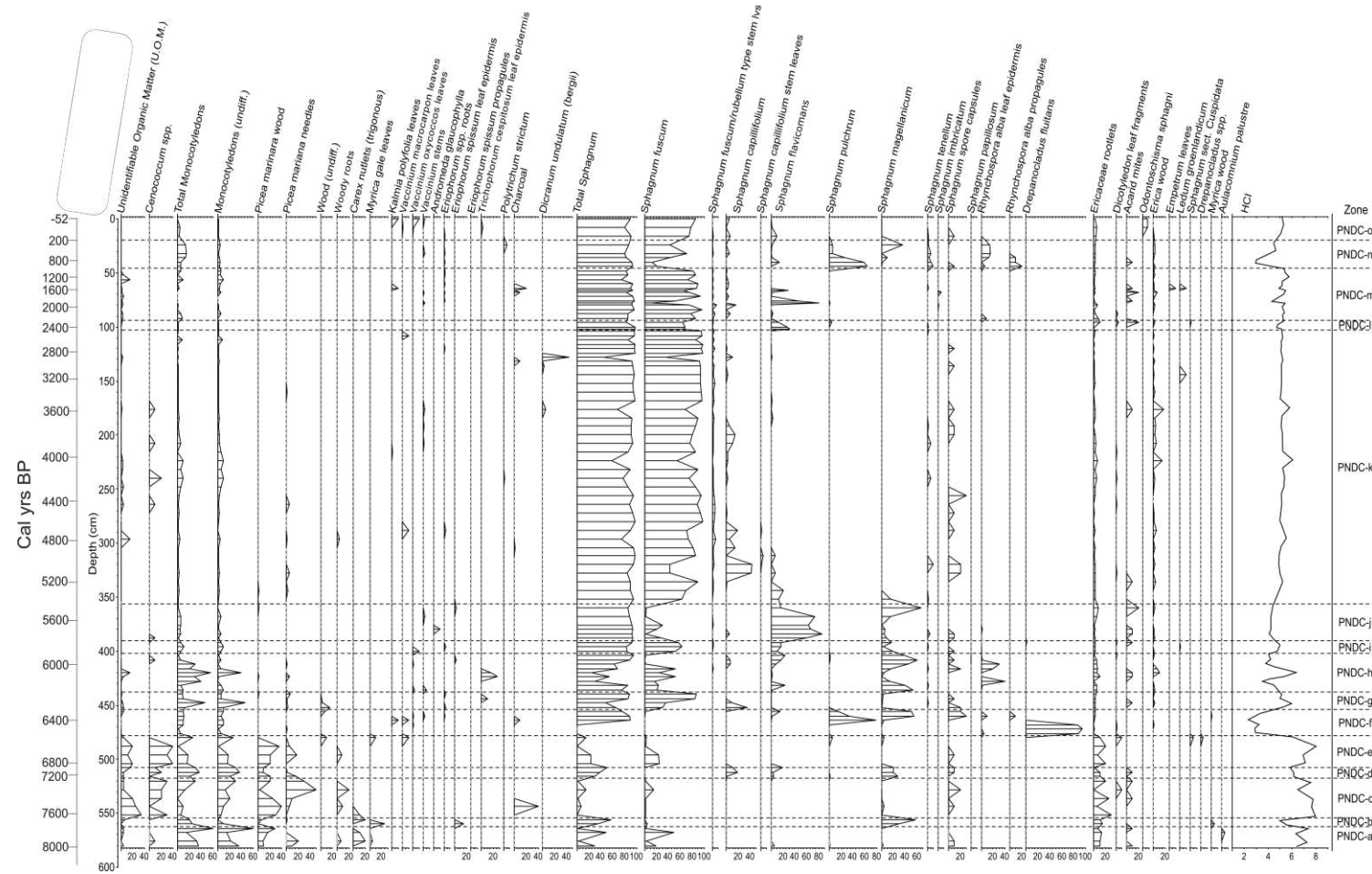


Figure 3

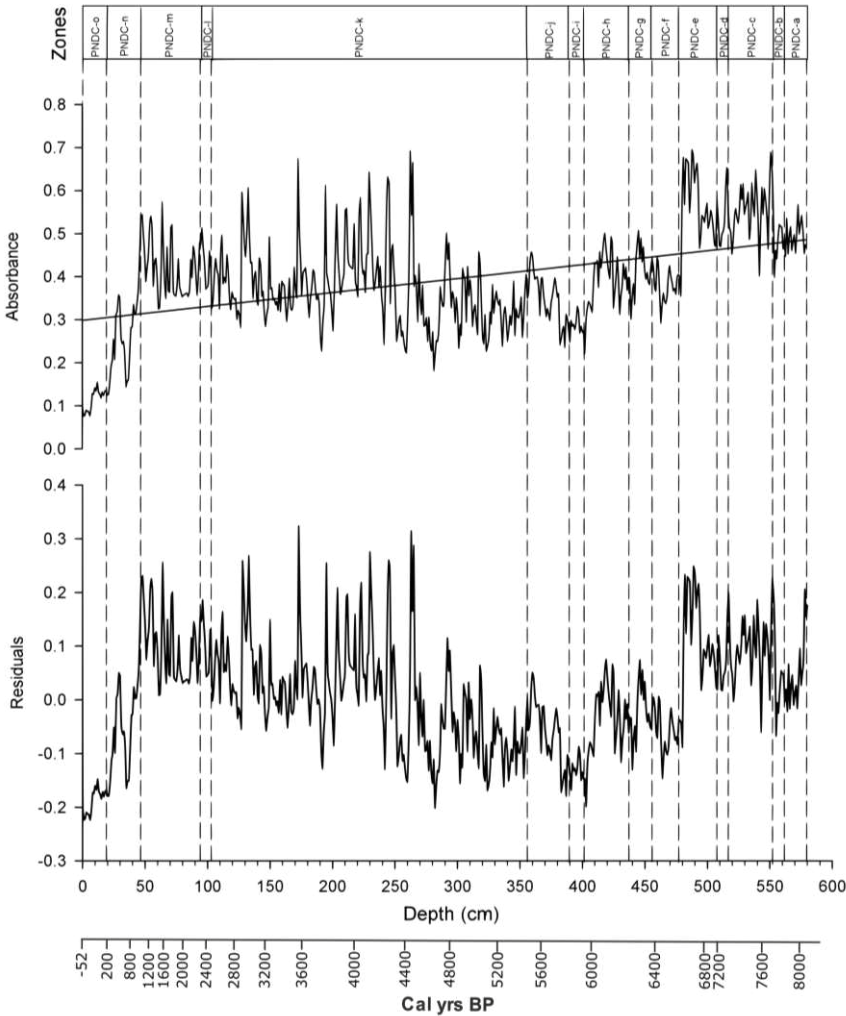


Figure 4

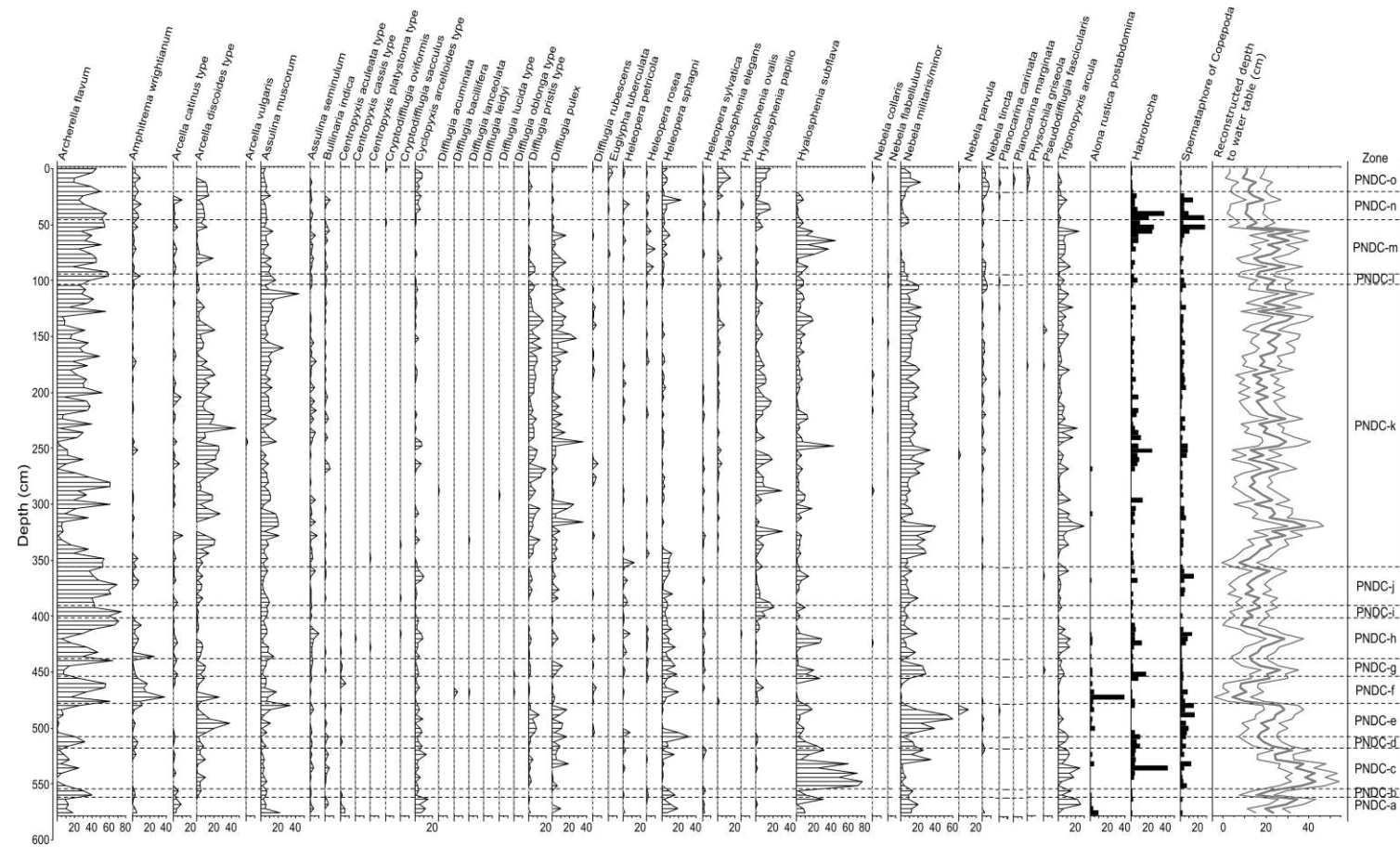


Figure 5

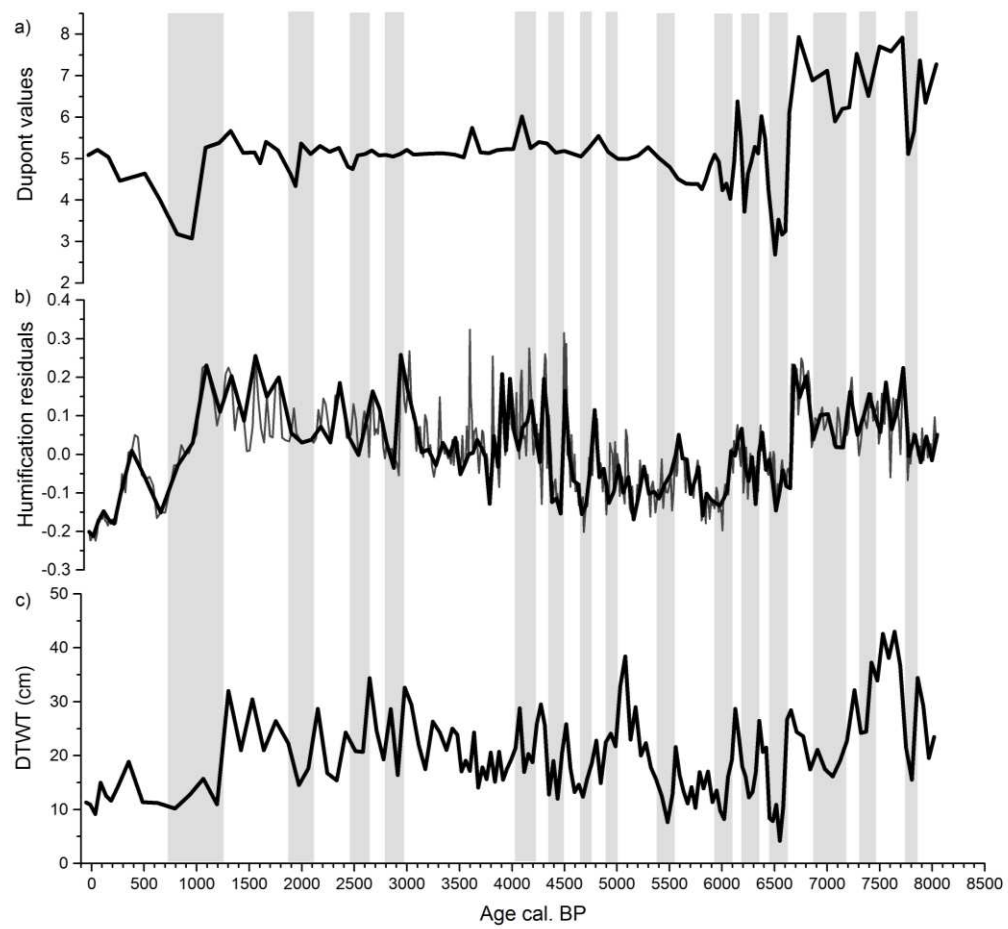


Figure 6

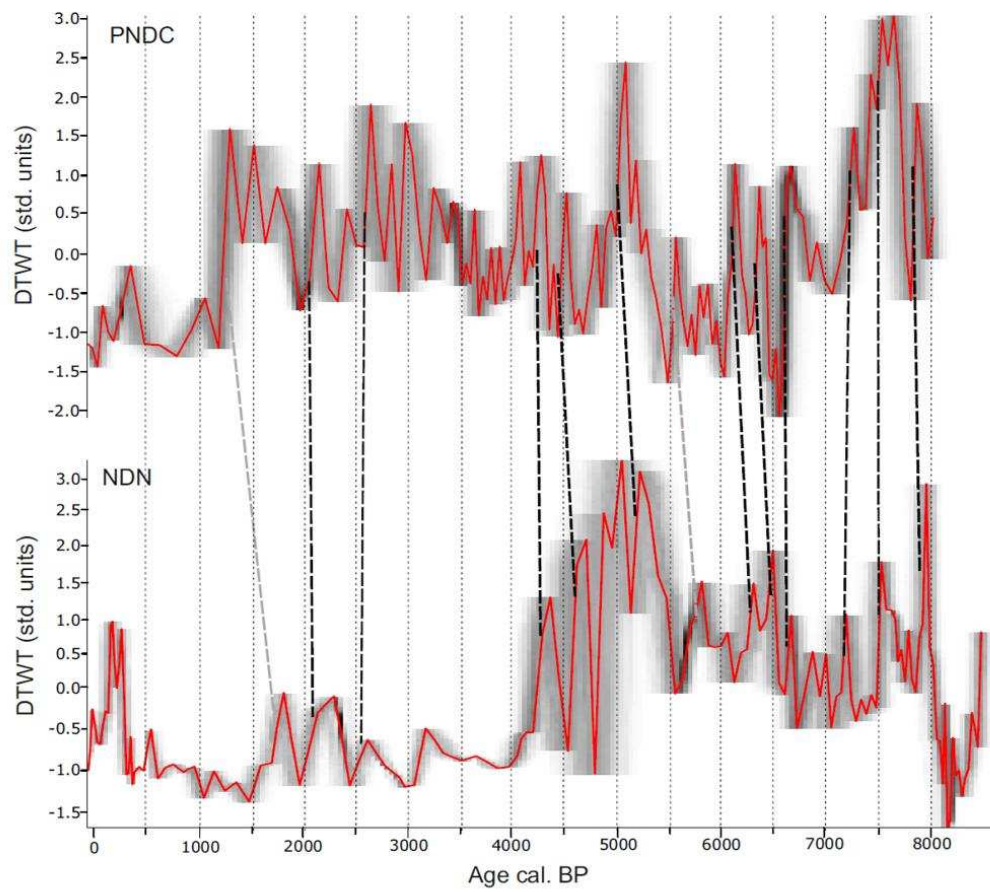


Figure 7

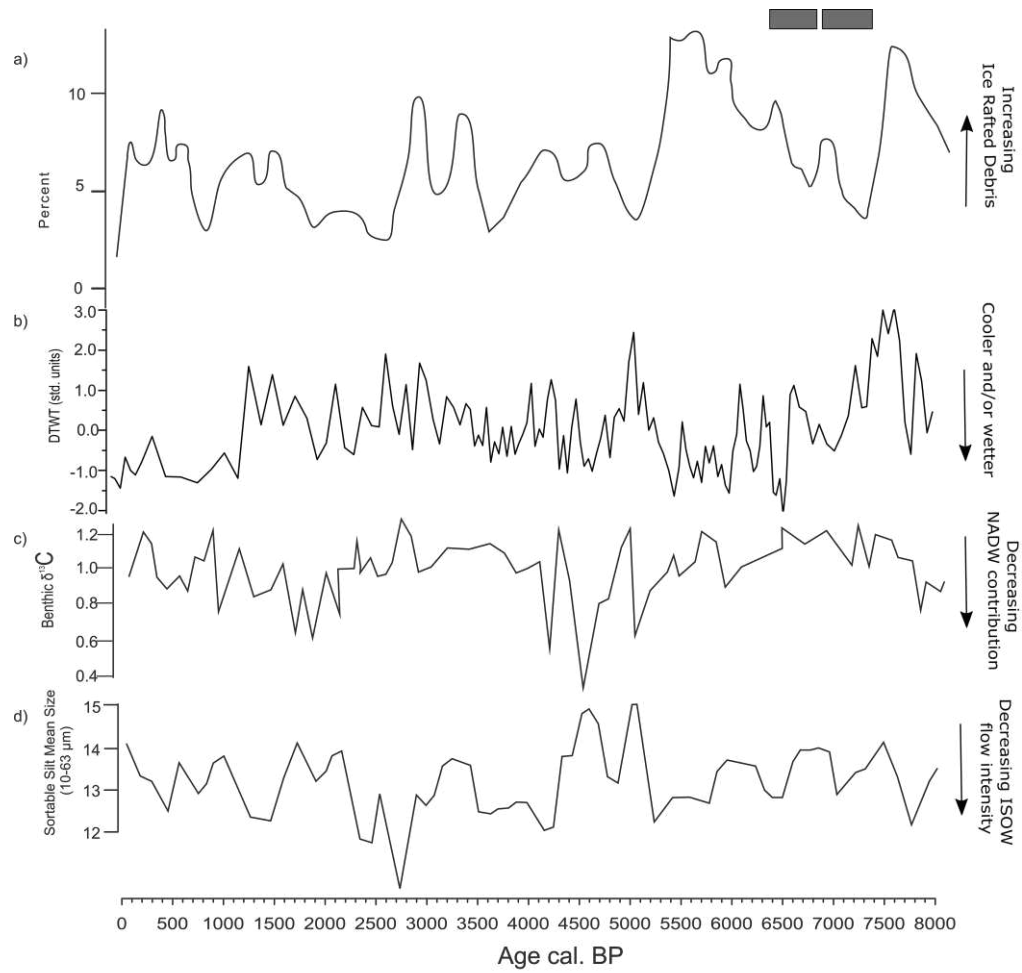


Table 1

Lab no.	Depth (cm)	Material	AMS RC date BP (uncal.)	1 σ error	Cal. range (2 σ) BP
SUERC-11763	24	<i>Sphagnum</i>	144	35	2 - 282
SUERC-11764	48	<i>Sphagnum</i>	1157	35	979 - 1172
SUERC-542	80	<i>Sphagnum</i>	2027	26	1899 - 2056
SUERC-11765	104	<i>Sphagnum</i>	2479	35	2366 - 2717
SUERC-11767	160	<i>Sphagnum</i>	3235	35	3383 - 3556
SUERC-543	216	<i>Sphagnum</i>	3660	41	3874 - 4139
BETA-195385	291	<i>Sphagnum</i>	4140	40	4535 - 4824
SUERC-11768	352	<i>Sphagnum</i>	4753	35	5330 - 5587
SUERC-11769	392	<i>Sphagnum</i>	5113	35	5749 - 5926
BETA-195386	412	<i>Sphagnum</i>	5320	40	5991 - 6263
SUERC-544	480	<i>Sphagnum</i>	5727	32	6442 - 6633
SUERC-11770	520	<i>Sphagnum</i>	6345	36	7172 - 7413
SUERC-545	576	<i>Sphagnum</i>	7222	39	7964 - 8159

Supplementary Figure and Table captions

Supplementary Figure 1. 'Bacon' based Bayesian age-depth model (Blaauw and Christen, 2011) for the core from PNDC. The three upper charts from left to right denote the stability of the Markov Chain Monte Carlo iterations (>1000 iterations), and the prior (green line) and posterior (grey shading) for accumulation rate and memory employed. For the lower chart: blue shading shows age distributions of calibrated AMS ^{14}C dates and grey shading denotes the posterior age-depth model bounded by grey dots showing the 95% probability intervals of the model.

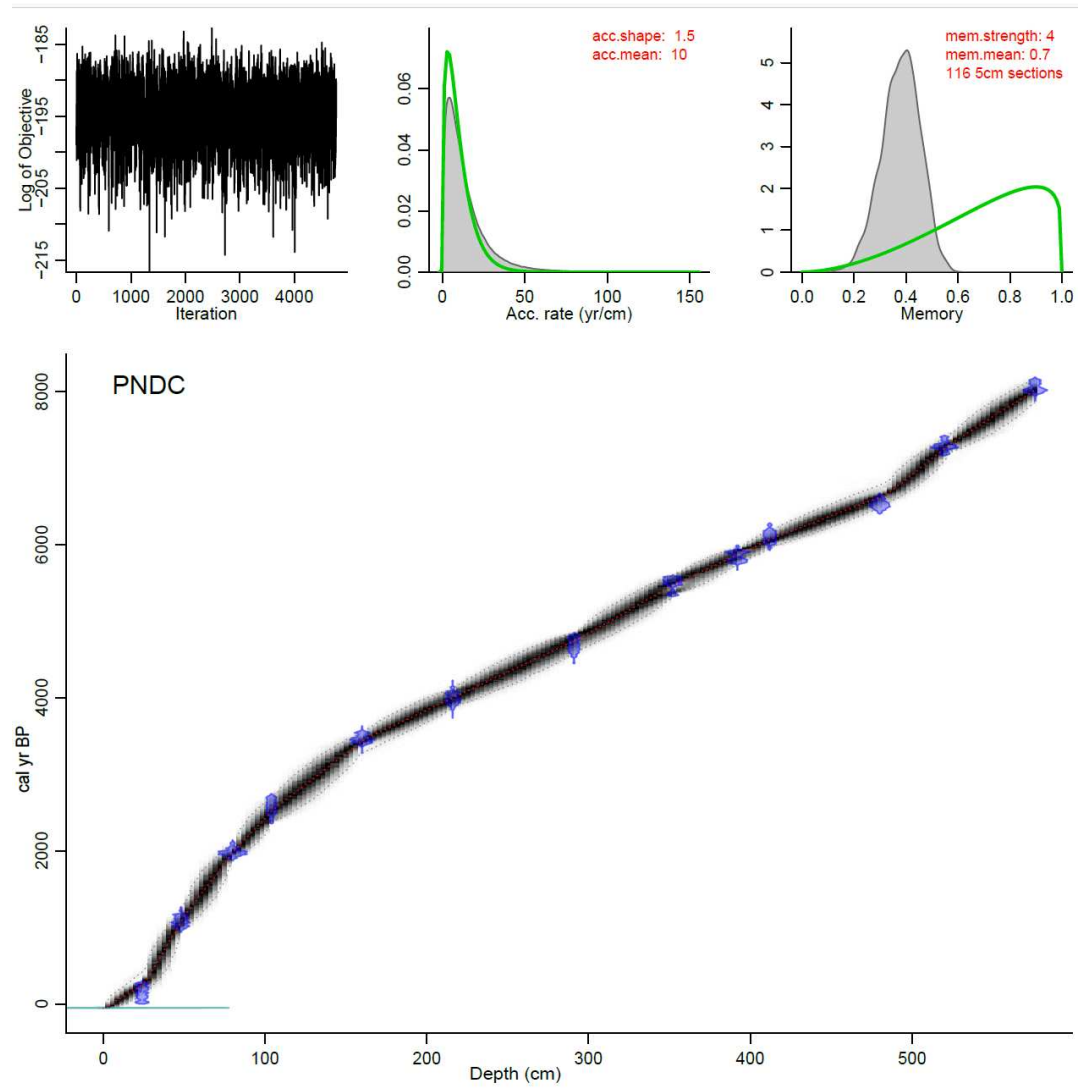
Supplementary Figure 2. Scatterplots showing comparisons between standardised proxy values from PNDC.

Supplementary Figure 3. 'Bacon' based Bayesian age-depth model (Blaauw and Christen, 2011) for the core from NDN. The three upper charts from left to right denote the stability of the Markov Chain Monte Carlo iterations (>1000 iterations), and the prior (green line) and posterior (grey shading) for accumulation rate and memory employed. For the lower chart: blue shading shows age distributions of calibrated AMS ^{14}C dates and grey shading denotes the posterior age-depth model bounded by grey dots showing the 95% probability intervals of the model.

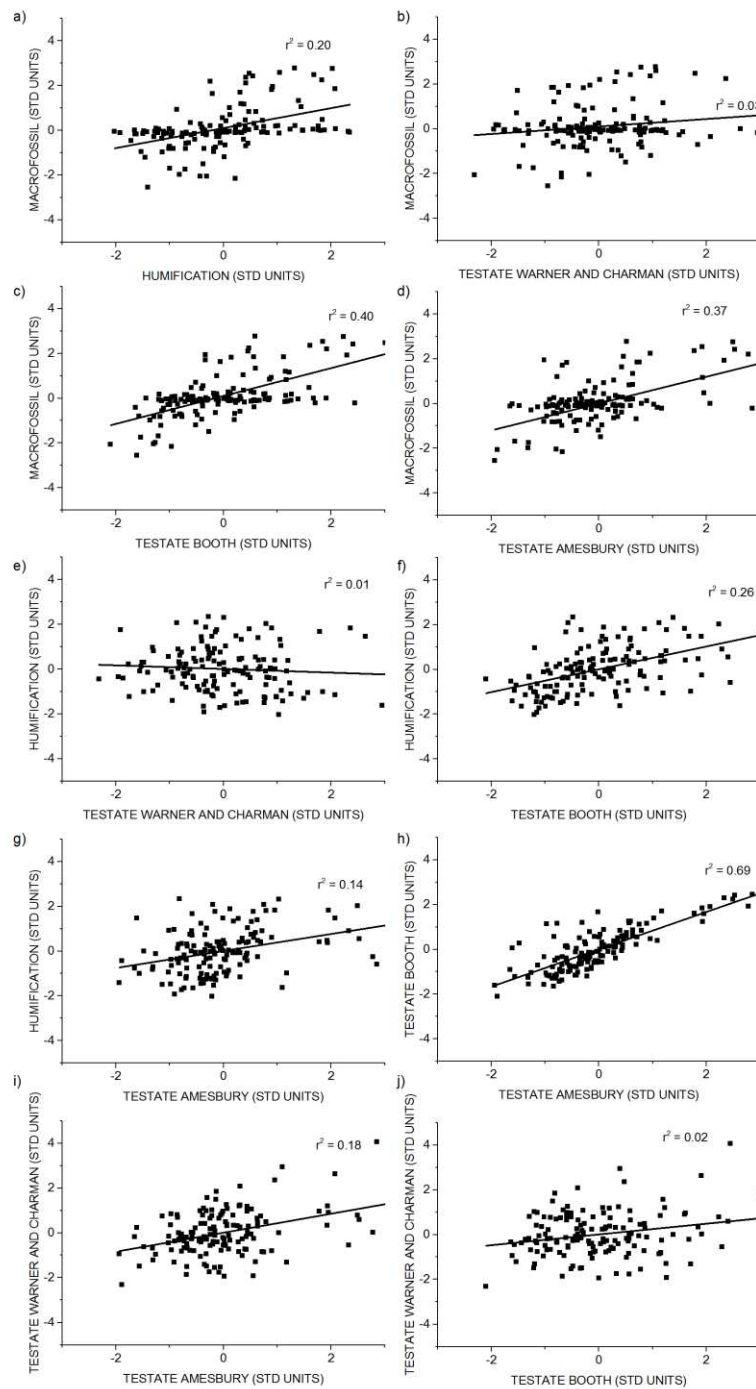
Supplementary Table 1. Dates and error ranges of prominent changes to reduced HCl, reconstructed DTWTs and levels of humification implying elevated water tables at PNDC. Grey shading denotes replication of changes by at least two proxies at PNDC. Changes to elevated water tables at NDN as derived from testate amoebae are listed and are in bold when they are well within the radiocarbon errors of those from the PNDC testate amoebae data.

Supplementary Table 2. Coefficient of determination and associated p values in parentheses for each standardised proxy and testate transfer function.

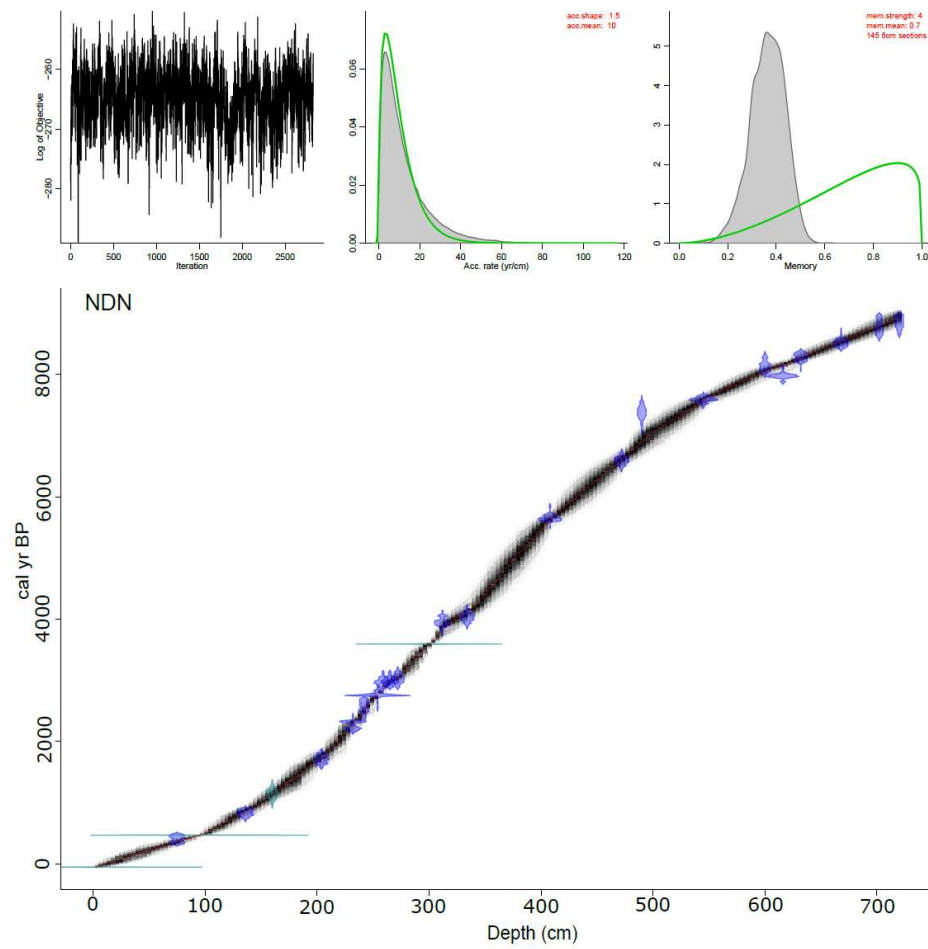
Supplementary Figure 1



Supplementary Figure 2



Supplementary Figure 3



Supplementary Table 1

PNDC			NDN
Macrofossil (HCl)	Humification	Testate amoebae (DTWT)	Testate amoebae (DTWT)
7830 MAR 7626-8014 BP		7830 MAR 7626-8014 BP	7940 MAR 7800-8053 BP
7500 MAR 7336 – 7694 BP	7500 MAR 7336 – 7694 BP	7500 MAR 7336 – 7694 BP	7500 MAR 7330-7627 BP
7220 MAR 7052-7365 BP	7160 MAR 6951-7318 BP	7220 MAR 7053-7366 BP	7170 MAR 6930-7378 BP
6600 MAR 6472-6760 BP	6640 MAR 6517-6797 BP	6600 MAR 6472-6759 BP	6630 MAR 6504-6772 BP
6340 MAR 6198-6497 BP	6340 MAR 6198-6497 BP	6340 MAR 6198-6497 BP	6460 MAR 6236-6641 BP
6110 MAR 6004-6237 BP	6110 MAR 6004-6237 BP	6110 MAR 6004-6237 BP	6280 MAR 6031-6505 BP
	5550 MAR 5389-5655 BP	5550 MAR 5389-5655 BP	5790 MAR 5633-5983 BP
	4750 MAR 4580-4850 BP	5010 MAR 4815-5198 BP	5180 MAR 4862-5461 BP
	4440 MAR 4252-4622 BP	4500 MAR 4311 -4669 BP	4570 MAR 4309-4886 BP
	4290 MAR 4117-4477 BP	4290 MAR 4117-4477 BP	4310 MAR 4129-4584 BP
	3860 MAR 3702 -4011 BP		
	2890 MAR 2646-3131 BP	2950 MAR 2680 -3187 BP	
	2610 MAR 2428-2830 BP	2610 MAR 2428-2830 BP	2600 MAR 2452-2741 BP
	2310 MAR 2092-2548 BP	2110 MAR 1974-2319 BP	2080 MAR 1852-2260 BP
	1480 MAR 1219-1755 BP	1480 MAR 1219-1755 BP	1780 MAR 1640-1897 BP
	1250 MAR 1060 -1490 BP	1250 MAR 1060 -1490 BP	
	1000 MAR 790-1157 BP		
	290 MAR 161 -467 BP	290 MAR 161 -467 BP	

Supplementary Table 2

	HUMIFICATION	HCI	BOOTH TF	AMESBURY TF
HCI	0.455 (0.000)			
BOOTH TF	0.507 (0.000)	0.633 (0.000)		
AMESBURY TF	0.378 (0.000)	0.610 (0.000)	0.834 (0.000)	
CHARMAN TF	-0.081 (0.321)	0.170 (0.041)	0.240 (0.004)	0.424 (0.000)