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1 **Reconstructed centennial variability of Late Holocene storminess from Cors**
2 **Fochno, Wales, UK**

3

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9 **Abstract**

10 Future anthropogenic climate forcing is forecast to increase storm intensity and
11 frequency over Northern Europe, due to a northward shift of the storm tracks, and a
12 positive North Atlantic Oscillation (NAO). However understanding the significance of
13 such a change is difficult since the natural variability of storminess beyond the range
14 of instrumental data is poorly known. Here we present a decadal resolution record of
15 storminess covering the Late Holocene, based on a 4 m long core taken from the
16 peat bog of Cors Fochno in mid-Wales, UK. Storminess is indicated by variations in
17 the minerogenic content as well as bromine deposited from sea spray. Twelve
18 episodes of enhanced storm activity are identified during the last 4.5 cal ka BP.
19 Although the age model gives some uncertainty in the timings, it appears that
20 storminess increased at the onset and close of North Atlantic cold events associated
21 with oceanic changes, with reduced storm activity at their peak. Cors Fochno is
22 strongly influenced by westerly moving storms, so it is suggested that the patterns
23 were due to variations in the intensity of westerly airflow and atmospheric circulation
24 during times when the latitudinal temperature gradient was steepened.

25

26

27 **Keywords:** storminess, Holocene, North Atlantic Oscillation, UK, storm track

28

29

30 Introduction

31 The most intense and damaging storms affecting Europe originate in the
32 Atlantic and impact upon the western seaboard. During the period December 2013 to
33 February 2014 the United Kingdom was affected by frequent, intense storms that
34 caused extensive flooding and infrastructural damage (Kendon and McCarthy,
35 2015). These storms were caused by a number of factors, including a persistent
36 southward perturbation of the jet stream over North America and a strong polar
37 vortex (Slingo *et al.*, 2014). Similar winter conditions are forecasted in response to
38 global warming; predictions suggest that over the next century the storm track will
39 shift northwards and storm frequency will increase in the British Isles, due to an
40 intensified jet stream (Pinto *et al.*, 2009; Stocker *et al.*, 2013). However making such
41 predictions is problematic, as there is poor understanding of natural variability and
42 relatively short instrumental storminess records (Allan *et al.*, 2009; von Storch and
43 Weisse, 2008). Thus there is a requirement for continuous and high resolution
44 reconstructions spanning the Late Holocene that capture changes in the frequency
45 and intensity of storms, termed 'storminess'. The western coast of Wales is a key
46 location for reconstructing these past changes, as recent storms demonstrate that
47 intensification of the jet stream leads to enhanced storminess in this region (Slingo *et*
48 *al.*, 2014).

49 The North Atlantic Oscillation (NAO) is a measure of the pressure difference
50 between the Azores High pressure and the Icelandic Low pressure (Hurrell, 1995). A
51 greater pressure difference during positive NAO anomalies results in the storm track
52 crossing northern Europe and an increased storm intensity, while negative NAO
53 anomalies (with reduced pressure gradients) cause the storm track to cross southern
54 Europe (Hurrell, 1995). During the instrumental period, the NAO is a dominant
55 control on storminess in Europe particularly during the winter months (Allan *et al.*,
56 2009), however research has shown a more complex relationship between
57 storminess and the NAO during the Late Holocene (Dawson *et al.*, 2002; Trouet *et*
58 *al.*, 2012). The Little Ice Age (LIA, c. 0.55-0.15 cal ka BP; 1400-1800 C.E.) is thought
59 to have had more negative NAO conditions (e.g. Trouet *et al.*, 2009, 2012). However
60 records indicate storminess was high across Europe during this time, rather than
61 only in southern Europe as would typically be expected from negative NAO
62 conditions (e.g. Sorrel *et al.*, 2012). It has been hypothesised that during the LIA a

63 steepened temperature gradient caused high intensity but low frequency storms
64 (Lamb, 1995), consistent with a dominant negative NAO (Trouet *et al.*, 2012).

65 The causes of centennial scale storminess variability through the Late Holocene
66 are debated, with oceanic and solar forcings frequently suggested. Reconstructions
67 of oceanic circulation from the North Atlantic region have shown LIA-type events
68 occur with a periodicity of $c.1470 \pm 500$ years; polar waters spread to more southern
69 latitudes, with a weaker Atlantic Meridional Overturning Circulation, North Atlantic
70 Current and subpolar gyre (Bianchi and McCave, 1999; Bond *et al.*, 1997; Thornalley
71 *et al.*, 2009). These episodes have been linked with increased storminess in Europe
72 and southward storm track shifts, suggested as being the result of cold ocean
73 temperatures at high latitudes causing a steepened temperature gradient (Sorrel *et al.*
74 *et al.*, 2012; Fletcher *et al.*, 2012; Sabatier *et al.*, 2012). However other studies have
75 emphasised the importance of solar minima as a main or additional cause of high
76 storminess (Martin-Puertas *et al.*, 2012; Sabatier *et al.*, 2012; Mellström *et al.*, 2015).
77 Further storm reconstructions from Europe are needed to improve the spatial and
78 temporal understanding of storminess. For example, opposite patterns between
79 northern and southern Europe could indicate storm track shifts associated with NAO
80 variability. Furthermore, improved understanding should help to untangle the key
81 drivers of storminess in northwest Europe and their relationship to Late Holocene
82 cold events.

83 A number of methods have been used to obtain extended records of storminess
84 proxies. In Greenland, sea-source sodium concentrations in ice cores provide a
85 proxy of sea-spray (Meeker and Mayewski, 2002). In Europe the deposition of
86 coastal dunes are a proxy for increased wind strength and hence storminess (Clarke
87 *et al.*, 2002; Clarke and Rendell, 2006; Clemmensen *et al.*, 2009). Other methods
88 include over-wash deposits in coastal lagoons (Sabatier *et al.*, 2012), cliff-top storm
89 deposits (CTSDs) left by extreme waves (e.g. Hansom and Hall, 2009) and marine
90 records reflecting wind-blown current strength and storm deposits (e.g. Andresen *et al.*
91 *et al.*, 2005; Billeaud *et al.*, 2009; Hass, 1996; Sorrel *et al.*, 2009). However, all of these
92 methods have potential problems. Dunes are prone to reworking, which bias the
93 record towards more recent events, and it is not clear whether the date of deposition
94 of coastal dunes records the period of most intense aeolian flux, or the waning stage
95 of a period of enhanced activity. Other records such as CTSDs and sand layers

96 within coastal lagoons show only the most extreme events, with erosion potentially
97 causing a bias towards recent events (Haslett and Bryant, 2007), although
98 preservation is good at some sites (Dezileau *et al.*, 2011; Sabatier *et al.*, 2008).

99 The aeolian flux of minerogenic material onto peat bogs can be used as a more
100 direct measure of wind strength and hence storminess. Björck and Clemmensen
101 (2004) counted the number of minerogenic grains above 0.2 mm diameter delivered
102 to the surface of two bogs in southwest Sweden per unit time. They termed this the
103 Aeolian Sand Influx (ASI) and although the method of analysis was extremely labour-
104 intensive they preferred ASI to a simple measure of the inorganic residue because
105 their bogs also contained significant silt sized minerogenic component from far-
106 travelled dust. The same approach has subsequently been applied to a number of
107 sites in Europe (e.g. De Jong *et al.*, 2006, 2009; Sjögren, 2009) and in South
108 America (Björck *et al.*, 2012). Reconstructions using this method are often high
109 resolution and continuous, however natural or human alterations to the landscapes
110 surrounding the bogs may influence the amount of sand deposition at times, so must
111 be considered in the interpretation of these records.

112 This study targets a unique site located in the extreme west of the United
113 Kingdom, which provides an ideal location to apply the aeolian flux methods
114 pioneered by Björck and Clemmensen (2004). The key to this approach is the
115 juxtaposition of a westerly facing seashore, with an abundant supply of sand sized
116 material that can be deflated, in close proximity to a continuously aggrading
117 ombrotrophic bog, which can act as a trap for the minerogenic material. The main
118 aim of this study is to produce a high resolution record of storminess for the western
119 margin of the United Kingdom through the Late Holocene using the influx of aeolian
120 sand as a proxy. A secondary aim is to test the effectiveness of bromine (Br)
121 measurements from micro X-ray fluorescence (μ XRF) core scanning as an indicator
122 of storminess. As a marine aerosol, Br records have been interpreted as
123 representing past storm activity (e.g. Unkel *et al.*, 2010; Turner *et al.*, 2014; Schofield
124 *et al.*, 2010). However, Br is also known to accumulate in organic matter and be
125 influenced by humification (e.g. Biester *et al.*, 2004). Comparison of Br profiles with
126 independent records of storminess is needed to validate the potential of this
127 technique. As rapid, high-resolution datasets can be obtained with μ XRF-scanning

128 (Croudace *et al.*, 2006), it is potentially a valuable addition to the range of methods
129 available for reconstructing past storm activity from peat bog deposits.

130

131 Study Area

132 Cors Fochno is a 650 ha ombrotrophic raised peat bog situated in Cardigan
133 Bay, mid-Wales, lying to the east of the 3-4 km long Borth Beach and Ynyslas sand
134 dunes, which act as a sediment source during storms (Figure 1). Station
135 measurements (1981-2010 A.D.) from Llanbedr, located on the coast 30 km north of
136 Cors Fochno, show an annual mean wind speed of 9.4 knots (0.51 m s^{-1}) (Met
137 Office, 2015). The bog is composed predominantly of sphagnum peat and has
138 developed over a mid-Holocene forest bed since c. 4.7 cal ka BP (Shi and Lamb,
139 1991; Wilks, 1979), with the central dome reaching a thickness of 5 m (Hughes and
140 Schulz, 2001). The bog is situated to the south of the village of Borth, The margins
141 have been modified by peat cutting in recent centuries and an artificial channel has
142 been created for the Afon Leri; however the central dome is unaffected by these
143 changes and it forms the largest area of primary raised bog in lowland Britain
144 (Poucher, 2009). The ecological importance and unique oceanic setting of the bog
145 have contributed to its Ramsar status and its inclusion in the UNESCO Dyfi
146 Biosphere. Palaeoenvironmental research has previously been carried out on Cors
147 Fochno to investigate local pollution, vegetation change and coastal and sea level
148 changes (Hughes and Schulz, 2001; Mighall *et al.*, 2009; Moore, 1968).

149

150 Methods

151 Two cores were taken from Cors Fochno using a Russian corer: a short core
152 (*core 1*) from the northern edge of the bog at $52^{\circ}30'28''\text{N}$, $4^{\circ}1'17''\text{W}$, and a main long
153 core (*core 2*) from a central site at $52^{\circ}30'9''\text{N}$, $4^{\circ}0'39''\text{W}$ (Figure 1). Cores were
154 wrapped securely in the field and subsequently stored at 4°C . Core 1 was to a depth
155 of 1 m, although the active peat in the acrotelm (upper 18 cm) was not preserved
156 during coring so has not been analysed. At the central site, a 4 m sequence was
157 obtained; again the acrotelm (upper 14 cm) was not preserved during coring.

158 Cores were subsampled into u-channels to ensure a consistent sample
159 surface and scanned using an ITRAX μ XRF core scanner at a resolution of 200 μ m
160 (30kV, 30mA, 12 second count). The μ XRF bromine (Br) results were normalised
161 using the incoherent + coherent peaks, which represent Compton and Rayleigh
162 scattering and provide an estimation of the organic and water content of the
163 sediments.

164 The sediment within the u-channels was then sliced into 1 cm sections, so
165 that known sample volumes were used (each 2.3 cm³). Loss-on-ignition was used to
166 separate the minerogenic material from the peat. Samples were dried at 105°C
167 overnight, ignited in a furnace at 550°C for 4 hours and weighed between every
168 stage (Dean, 1974; Heiri *et al.*, 2001). This allowed the Ignition Residue (IR) to be
169 calculated, which was the ignited weight as a percentage of the dried weight, and
170 shows the quantity of inorganic material in the sample. This also allowed the Organic
171 Bulk Density (OBD) of each sample to be calculated, which can be a measure of
172 peat humification (Björck and Clemmensen, 2004):

173
$$\text{OBD (g cm}^{-3}\text{)} = (\text{dried weight} - \text{ignited weight}) / \text{sample volume}$$

174 To assess the relationship between the IR measurement and sand content
175 within the peat, we determined the Aeolian Sand Influx (ASI) over two sections using
176 the approach of Björck and Clemmensen (2004). The IR samples collected from
177 depths of 14-110 cm and 150-200 cm in Core 2 were treated twice with 10%
178 hydrochloric acid to remove any carbonates, then with 30% hydrogen peroxide to
179 remove any remnant organics that may have survived ignition, before being mounted
180 on glycerine jelly slides and analysed under a microscope. The number of grains
181 with a diameter over 200 μ m was counted, and the diameter of the largest grain in
182 each sample was recorded. We used the same lower threshold for counting grains
183 as the original research by Björck and Clemmensen (2004). The ASI was then
184 calculated:

185
$$\text{ASI} = (\text{number of grains} > 200\mu\text{m} / \text{sample volume}) / \text{number of years in 1 cm of peat}$$

186 The age-depth model for the main core (core 2) was developed from AMS
187 radiocarbon dates from five bulk peat samples. The calibration and age model was
188 constructed using Bayesian analysis by OxCal version 4.2.2, which used the Intcal13

189 calibration curve (Ramsey, 2009; Reimer *et al.*, 2013). Spectral analysis was carried
190 out on the IR results using a normalised Lomb-Scargle fourier transform (Shoelson,
191 2001), which is a method of spectral analysis that can identify frequency signals in
192 unevenly spaced data (Lomb, 1976; Press and Rybicki, 1989; Scargle, 1982).

193

194 Results

195 The cores consist of dark brown sphagnum peat throughout, but with
196 variations in the degree of humification, as shown by the OBD results (Figure 2).
197 Below 2 m depth there is dense, humified peat while above 2 m there are greater
198 variations in the density and degree of humification.

199 The results from the five radiocarbon dates are shown in Table 1, including the
200 median probability of the 2 σ range and the errors representing the analytical error
201 propagated through the calibration software. The 4 m core spans the last 4.5 cal ka
202 BP. The upper radiocarbon date at 50 cm depth had a modern age, thought to be
203 due to contamination, so this age was not included in the age-depth model. The four
204 other ages and known age of the modern bog surface at 0 cm depth were used to
205 create the age-depth model (Figure 2). This shows a consistent peat accumulation
206 rate of ~0.8-1.2 mm/year, similar to the rate of growth calculated by Hughes and
207 Schulz (2001) for the bog during the period from 7.04 to 3.27 ^{14}C ka BP. Mighall *et*
208 *al.* (2009) obtained a chronology for a peat core from the centre of Cors Fochno
209 covering the period from 3.63 ^{14}C ka BP to the present day. The ^{210}Pb dates for the
210 upper 17 cm, and a series of radiocarbon dates from 53 cm to 325 cm depth,
211 confirmed an essentially constant accumulation rate at this site, and is very similar to
212 our age model. The lack of age-control in our study between 130 cm depth and the
213 surface means that there is additional chronological uncertainty in this interval for our
214 core, but comparison with these others studies implies that any discrepancies should
215 be small.

216 The comparison between the ASI, IR and maximum grain sizes over two
217 sections showed that there was less similarity between the ASI and IR results with
218 depth. Maximum grain size also decreased downcore (Figure 3). In the main core
219 (core 2) the average IR is 2.2%, and throughout much of the core the IR results vary

220 from this by less than 1% (Figure 4), However, in some parts of the core there are
221 higher IR values, reaching a maximum of 6.8%, with some peaks spanning depths of
222 >10 cm. The IR results revealed twelve peaks at: 4.46-4.44, 3.98-3.92, 3.77-3.61, 3-
223 2.97, 2.84-2.8, 2.31-2.24, 2.09-1.97, 1.56-1.55, 1.4-1.35, 1.09-1.05, 0.58-0.47 and
224 0.21-0.12 cal ka BP (Figure 4). The bromine record shows peaks at c. 3.9, 3.4, 3.3,
225 2.9, 2.8, 2.3, 1.9, 1.7, 1.4, 0.9, 0.5 and 0.2 cal ka BP. The Lomb-Scargle spectral
226 analysis shows the Cors Fochno IR record has cycles that are significant at the 95%
227 confidence limit with periodicities of 1740, 870, 445, 395, 315 and 290 years (Figure
228 5).

229

230 Proxy Interpretation

231 By comparing the ASI, IR and maximum grain size results in two sections
232 (Figure 3) it was clear that with depth in the core the ASI no longer captured changes
233 in sand content due to smaller grain sizes. Sites in Scandinavia where the ASI
234 method has been used have a significant fine dust input originating from long
235 distance transport of grains (Björck and Clemmensen, 2004), so it was necessary for
236 the ASI proxy to be used to measure the coarse sand and exclude the fine sand and
237 silt fractions, which may not have been indicative of storms. Cors Fochno however
238 can be expected to have had minimal fine dust input because it is on the Atlantic
239 seaboard, meaning the IR results here give an unambiguous storminess signal. The
240 IR is a preferable proxy because the ASI relies on correctly defining the minimum
241 threshold for counting grains. It is suggested that a gradual sea level rise during the
242 Mid-Late Holocene at this site resulted in a transgression (Kidson and Heyworth,
243 1978); the change in distance between the coring site and the beach appears to
244 have altered the size of sand grain reaching the core site over time (Figure 3).
245 Therefore at Cors Fochno the use of the IR was considered the most appropriate
246 proxy for detecting storminess when considering the location, environmental
247 evolution of the surrounding area and the limitations of the ASI method.

248 Archaeological evidence suggests a human presence in the area throughout the
249 Late Holocene but particularly since the early 19th century, when there was land
250 drainage and diversion of the River Leri, greater agriculture, the building of a railway
251 track and expansion of Borth (Poucher, 2009). It is noteworthy that many of the

252 archaeological sites are situated to the east and south-east of Cors Fochno, so the
253 prevailing westerly winds would not transport sediment onto the bog from these. As
254 the beach and dunes are in close proximity to the bog and in the path of the
255 prevailing winds, we consider that these have been the dominant source of sand
256 delivered to Cors Fochno, with the human activities resulting in a negligible level of
257 disturbance prior to the 19th century.

258 The similarity between the independent IR and Br records suggest that both are
259 capturing storminess signals rather than human activity as Br (a marine aerosol) is
260 less likely to be affected by anthropogenic disturbance. The position of the site on
261 the western seaboard of Wales mean that these most likely represent periods of
262 increased westerly storm activity. However there is a discrepancy between the two
263 proxies at 3.7 cal ka BP, with very high IR values but low Br. This may have been
264 caused by local factors, such as human disturbance, though there is no independent
265 evidence for this. Alternatively, as Br will be more easily transported than sand, the
266 concentrations in the bog may reflect storm frequency rather than intensity, so this
267 may have been a time of generally low storminess punctuated with a number of
268 intense storms capable of sand transport. Phases of enhanced storminess (shown
269 by the sand influx) last for between 10 and 160 years. The peaks have an average
270 age error of ± 160 years but with greater uncertainty before 3.8 cal ka BP and after
271 1.34 cal ka BP.

272

273 Discussion

274 *European Spatial Patterns of Storminess*

275 Records of storminess from Europe have been compiled to show temporal as
276 well as spatial changes over the Late Holocene (Figure 6). Analysing multiple
277 reconstructions within a region reduces the impact of local factors and
278 methodological limitations. Comparing sites from northern and southern Europe may
279 also allow changes related to storm track shifts and the NAO to be detected.

280 During the last 2000 years the Cors Fochno reconstruction (Figure 6F) shows
281 four peaks in storminess which, despite large age errors in both reconstructions,
282 appear to be in phase with strengthened bottom water currents in the Skagerrak

283 Sea, thought to be driven by westerly storms (Hass, 1996; Figure 6G). Storm
284 reconstructions from around northern Europe, including Scotland and Northern
285 Ireland as well as Scandinavia and France, show some but not all of these events
286 (see references within Figure 6). Together the results support that the Cors Fochno
287 storminess record is capturing a regional climate signal, and that northern and
288 central Europe have experienced greater storminess at c. 1.5, 1.05, 0.5 and 0.1 cal
289 ka BP. A similar comparison of European reconstructions by Sorrel *et al.* (2012)
290 suggested that during the last 2000 years storminess was higher across Europe at
291 1.9-1.05 and 0.6-0.25 cal ka BP. Our comparison agrees with these broad periods
292 but indicates that within these times storminess was more variable, with periods of
293 increased storminess lasting around 100-200 years.

294 In southern Europe the identification of storm events during the last 2000 years
295 is less certain as a result of fewer available reconstructions. In Portugal, sand dune
296 development at 1.5 and 0.1 cal ka BP (Clarke and Rendell, 2006; Figure 6M) occurs
297 at similar times to the periods of high storminess seen in northern Europe, so may
298 indicate increases in storminess across mainland Europe. Similarly a Mediterranean
299 storm reconstruction, measuring wave overtopping of a lagoon barrier, shows high
300 storminess during the periods 1.95-1.4 and 0.4-0.05 cal ka BP (Sabatier *et al.*, 2012;
301 Figure 6N), although these are longer periods of increased storm activity than those
302 identified in northern Europe.

303 The reconstructions from the high latitudes (Figure 6 A-C) show conflicting
304 patterns of storminess. A proxy for bottom current strength on the Icelandic shelf
305 indicates that storminess increased between 1.1-0.7 cal ka BP (Andresen *et al.*,
306 2005, Figure 6B). However a terrestrial proxy of loess grain size in Iceland suggests
307 increased storminess between c.2-1 cal ka BP and to a lesser degree after 0.5 cal ka
308 BP (Jackson *et al.*, 2005, Figure 6C), while the GISP2 sea spray proxy supports the
309 finding that there was increased storminess after 0.5 cal ka BP (Mayewski *et al.*,
310 1997; Figure 6A).

311 The Cors Fochno record suggests that the period between 4.5 and 2 cal ka BP
312 had as frequent storms as the time since 2 cal ka BP. The large peak in the Cors
313 Fochno reconstruction at c.3.7 cal ka BP is not clearly shown in the other
314 reconstructions, however the enhanced storminess between around 2.3-2 cal ka BP

315 is in agreement with a single reconstruction in Scandinavia and another from
316 Portugal (Figure 6 I and M). There are also indications that storminess increased
317 widely in Europe c. 2.9 cal ka BP, with peaks in Wales, Scandinavia, France, Ireland
318 and Greenland at this time (Figure 6; Mellström *et al.*, 2015), as well as in coastal
319 deposits in northwest Spain (Gonzalez-Álvarez *et al.*, 2005).

320 The compiled storm reconstructions indicate that during the last 2000 years
321 periods of enhanced storminess were simultaneous (within dating error) across
322 northern Europe, during the LIA (0.55 and 0.1 cal ka BP) and around 1.1 and 1.5 cal
323 ka BP. The pattern is less clear before 2 cal ka BP, most likely due to fewer records,
324 although there is some suggestion that widespread storminess increases occurred
325 c.2.9 cal ka BP and there was potentially a period of enhanced storminess at 3.7 cal
326 ka BP.

327

328 *Storm Track Shifts*

329 The LIA appears to have had high storminess particularly over the transitions
330 (c.0.6-0.4 and c.0.2-0.05 cal ka BP) from the Medieval Climate Anomaly and post-
331 AD 1900, although the uncertainty in the age of the peak at 0.6-0.4 cal ka BP is 200
332 years. Nevertheless fluvial flooding reconstructions from northern, western and
333 central Europe, also show increases at 0.7-0.4 and 0.2-0.05 cal ka BP (Rumsby and
334 Macklin, 1996). Between these periods, during the mid-LIA, reconstructions have
335 contradictory findings for the magnitude of storminess. Some indicate intense storms
336 in northern Europe at this time (Lamb, 1995; Wheeler *et al.*, 2010) and in spring and
337 autumn a more southerly storm track is believed to have crossed the British Isles
338 (Luterbacher *et al.*, 2001). However the Cors Fochno reconstruction suggests
339 reduced storminess during the mid-LIA and there is thought to have been reduced
340 flooding in northern and central Europe (Rumsby and Macklin, 1996). Furthermore a
341 proxy reconstruction of wind-driven Atlantic Water Inflow into the Norwegian Sea
342 suggests that the storm track was not in a northerly position during the LIA
343 (Giraudeau *et al.*, 2010). This reduced storminess in the Cors Fochno reconstruction
344 may in part result from the bogs location making it sensitive to westerly tracking
345 storms. Westerly airflow was suggested as the cause of the increased flooding at the
346 LIA transitions (Rumsby and Macklin, 1996), while at the peak of the LIA (the

347 Maunder Minimum, 1645-1715 A.D.) documentary and modelling evidence has
348 indicated that there were more meridional circulation patterns, blocking high
349 pressures across northern Europe, a southerly storm track in winter and lower
350 precipitation particularly on Britain's west coast (Jacobeit *et al.*, 2003; Lamb, 1966;
351 Luterbacher *et al.*, 2001; Raible *et al.*, 2007). In support of this, reconstructions from
352 across southern Europe suggest precipitation and flooding increased between 0.45-
353 0.25/0.15 cal ka BP (Benito *et al.*, 1996; Magny *et al.*, 2008; Pfister, 1984). Therefore
354 these findings support the idea that storminess increased during the LIA, potentially
355 due to a steepened temperature gradient (Trouet *et al.*, 2012), however they also
356 support that circulation patterns and storm track shifts were important.

357 A similar pattern can be observed during the period 1.55-1.05 cal ka BP. As
358 during the LIA, temperatures in the extra-tropical northern hemisphere were lower
359 between c.1.65-1.15 cal ka BP, during what is often termed the Dark Ages cold
360 period (Ljungqvist, 2010). The reconstructions presented imply that northern Europe
361 experienced increased storminess at c.1.56-1.35 and c.1.09-1.05 cal ka BP, and this
362 is supported by documentary evidence pieced together by Lamb (1995), showing
363 increased storminess c.1.4 and 1.1 cal ka BP. In the intervening period, as during
364 the LIA, southerly storm tracks are indicated by the north-south index based on
365 Norwegian glacier reconstructions, with a maximum southerly extent at 1.2 cal ka BP
366 (Bakke *et al.*, 2008), and by reconstructions suggesting increased precipitation and
367 flooding in southern Europe at this time (Arnaud *et al.*, 2005; Magny *et al.*, 2007).
368 However as the wind-driven Atlantic water inflow into the Norwegian Sea remained
369 fairly high (Giraudeau *et al.*, 2010), it is possible that the storm track shift was not as
370 persistent or as far south as during the LIA.

371

372 *Oceanic and Atmospheric Circulation Changes*

373 During the instrumental period the NAO is the dominant control on storminess.
374 We compare the Cors Fochno storm reconstruction with an NAO reconstruction
375 based on weather-driven changes in hypolimnic anoxia from a lake from south west
376 Greenland (Olsen *et al.*, 2012; Figure 7E). This indicates that storm events coincide
377 with some negative NAO periods, particularly since 2.2 cal ka BP (c.1.1, 1.4, 1.6 and
378 2.1 cal ka BP), however the earlier section of the record before 2.2 cal ka BP does

379 not show increased storminess at times of negative NAO. As the locations of the
380 NAO pressure centres are non-stationary (Schmutz *et al.*, 2000), it is possible that
381 the NAO influence on regional climate has changed over the Late Holocene, which
382 may explain the lower correspondence in the earlier part of the record. As
383 hypothesised for the LIA (Trouet *et al.*, 2012) a steepened temperature gradient may
384 have caused intensified storms despite frequent negative NAO patterns at the times
385 of high storminess since 2.2 cal ka BP. These may have been climate transitions
386 associated with both high, westerly storminess in northern Europe as well as
387 negative NAO conditions. Although contradictory, as weather patterns vary on the
388 timescales of days-weeks it is likely that periods with frequent negative NAO
389 anomalies could also have had strong westerly airflow across northern Europe.
390 Overall the cold events of the Late Holocene may have had a steepened meridional
391 temperature gradient, like the LIA, which resulted in periods characterised by both
392 strengthened westerly airflow as well as NAO negative events.

393 Oceanic forcing of storminess is suggested by the dominant cycle of 1740 years
394 in the Cors Fochno record. Similar length cycles have been found in other
395 reconstructions of cyclonic activity (precipitation and wind) from the Mediterranean,
396 Iceland and Greenland (Debret *et al.*, 2007; Fletcher *et al.*, 2013; Giraudeau *et al.*,
397 2000; Jackson *et al.*, 2005; O'Brien *et al.*, 1995). It has been suggested that the
398 1700 year cycle is the result of internal oceanic forcing, which imprints on cyclonic
399 activity in the North Atlantic, as the 1700 year cycle has been identified in North
400 Atlantic marine cores (Bianchi and McCave, 1999; Debret *et al.*, 2007; Fletcher *et al.*,
401 2013; Giraudeau *et al.*, 2000).

402 This ocean-atmosphere relationship is potentially supported by comparing the
403 Cors Fochno storminess reconstruction with those reflecting the strength of the N.
404 Atlantic thermohaline circulation (Figure 7 and references therein), although dating
405 errors in parts of our reconstruction and the marine reconstructions make such
406 comparisons difficult. There appear to be increases in storminess at the transitions of
407 periods with weak thermohaline circulation, as shown by high ice-rafting debris (IRD)
408 concentrations in North Atlantic cores (showing when polar waters moved south;
409 Figure 7B), and reduced strength of the Iceland-Scotland Overflow Water (ISOW;
410 Figure 7C) and sub-polar gyre (SPG) circulation (Figure 7D). These changes
411 occurred during the above described LIA (c.0.6-0.05 cal ka BP) and Dark Ages

412 (c.1.6-1.05 cal ka BP), although the SPG strength proxy shows some difference in
413 the timing of the weakening during the Dark Ages. Earlier in the Late Holocene the
414 ISOW speed was reduced between 4.2-3.5 and 3-2.2 cal ka BP, the SPG circulation
415 weakened c.4.5-3.8 and 3-2 cal ka BP and the IRD increased at c.4 cal ka BP
416 (Figure 7 B-D). At the transitions of these periods the Cors Fochno period has single
417 or pairs of periods with enhanced storm activity (Figure 7A), although the increases
418 in the IRD at c.2.8 cal ka BP occur simultaneously with high storminess in records
419 from Europe. The Cors Fochno reconstruction indicates that at the transitions of LIA-
420 type events during the Late Holocene storminess (or westerly airflow) may have
421 increased in northern Europe.

422 Solar maxima and minima have both been suggested as causes of variations in
423 storminess (Gleisner and Thejll, 2003; Huth *et al.*, 2006; Lamb, 1991; Mayewski *et*
424 *al.*, 2005; Poore *et al.*, 2003; Wheeler *et al.*, 2010; Mellström *et al.*, 2015). This may
425 be supported by cycles of 445 and 320 years in the Cors Fochno record, which are
426 similar to 420 and 315 year solar cycles (Stuiver and Braziunas, 1989; Poore *et al.*,
427 2003), although the centennial-length cycles may have been distorted by the age
428 errors. The comparison between the Cors Fochno reconstruction and the total solar
429 irradiance (TSI) reconstruction (Figure 7F; Steinhilber *et al.*, 2008) does not indicate
430 that storminess increased at either solar maxima or minima, therefore it is not
431 possible to ascertain a solar influence on storminess.

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441 Conclusion

442 Aeolian-transport of sand onto the coastal ombrotrophic peat bog of Cors
443 Fochno, situated on the west coast of Wales, has allowed a storminess
444 reconstruction to be made spanning 4500 years. Twelve peaks in storminess have
445 been identified at 4.46-4.44, 3.98-3.92, 3.77-3.61, 3-2.97, 2.84-2.8, 2.31-2.24, 2.09-
446 1.97, 1.56-1.55, 1.4-1.35, 1.09-1.05, 0.58-0.47 and 0.21-0.12 cal ka BP. Comparison
447 between sand content and normalised μ XRF bromine measurements support the
448 use of bromine as a proxy for sea spray and therefore storminess in peat bogs.

449 By comparison with other European Late Holocene storminess
450 reconstructions it is possible to identify synchronous increases in storminess across
451 northern Europe. The Cors Fochno reconstruction and others in northern Europe
452 show enhanced storminess at the transitions of times that were calm and cold, at
453 1.6-1 cal ka BP and 0.55-0.05 cal ka BP (LIA). Evidence indicates that the ocean
454 circulation in the North Atlantic during these periods, and similar events of the Late
455 Holocene, was weakened, and the identified storm events occur at the transitions of
456 these times. The 1740 year cycle found in the Cors Fochno reconstruction has also
457 been found in westerly wind and oceanic proxies elsewhere also suggesting a link
458 with oceanic variability. The findings support the hypothesis that a steeper
459 temperature gradient increased storm intensity across Europe during the LIA (Trouet
460 *et al.*, 2012; Lamb, 1995) and similar events earlier in the Late Holocene; however
461 these times were temporally variable, possibly as the result of circulation variability
462 and episodes of strengthened westerly airflow.

463

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- 727

728 Table 1: Radiocarbon dates and calibrated ages from core 2, Cors Fochno.

Sample Depth (cm)	Laboratory Code	$\delta^{13}\text{C}$ (‰)	Radiocarbon Age (^{14}C yr BP $\pm 1\sigma$)	Calibrated Age (2σ interval)	Calibrated Median age (cal yrs BP)
50-51	Beta-289917	-27.2	post-bomb	NA	NA
130-131	Beta-281364	-22.4	1700 ± 40	1705 - 1535	1610
185-186	Beta-289918	-25.6	2060 ± 30	2119 - 1946	2030
260-261	Beta-289919	-26.8	2760 ± 30	2941 - 2779	2850
330-331	Beta-281365	-26.2	3430 ± 40	3828 - 3586	3690

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732 Figure Captions:

733 Figure 1: *left*: Location of Cors Fochno in the United Kingdom (inset) and map of the
734 Dyfi estuary and Cors Fochno bog with the two coring locations. *Right*: Map of
735 Europe showing the approximate locations of the NAO pressure centres (Hurrell and
736 Deser, 2010) and sites of storminess reconstructions discussed in this research
737 (letters corresponding to those in Figure 6).

738 Figure 2: *from left*: core 1 ignition residue results and core 2 ignition residue results,
739 organic bulk density and age-depth model.

740 Figure 3: Comparison of the proxies for sand content in core 2: The Aeolian
741 Sediment Influx, ignition residue and maximum grain size.

742 Figure 4: OBD, IR and Br/inc + coh results for core 2. Arrows highlight periods of
743 enhanced storm activity shown by the IR and bromine proxies. The grey line on the
744 bromine plot gives the raw measurements (0.2 mm resolution) and the black line the
745 smoothed measurements (using a 1 cm moving average).

746 Figure 5: Lomb-Scargle Powerspectrum analysis of the IR results of core 2

747 Figure 6: Latitudinal differences of European storminess records

748 Sites (from top): **A** Greenland (Meeker and Mayewski, 2002), **B** Iceland
749 (basalt/plagioclase ratio) (Andresen *et al.*, 2005), **C** Iceland (Jackson *et al.*, 2005), **D**
750 Scotland (Hansom and Hall, 2009), **E** northern Ireland (Wilson *et al.*, 2004) **F** Cors
751 Fochno, Wales (*this study*) **G** Skagerrak Sea (Hass, 1996), **H** Halland Coast,
752 Sweden (De Jong *et al.*, 2006), **I** Denmark (Clemmensen *et al.*, 2009), **J** Seine
753 Estuary, France (Sorrel *et al.*, 2009), **K** Mont-Saint-Michel Bay, France (Billeaud *et*
754 *al.*, 2009) **L** Aquitaine coast, France (Clarke *et al.*, 2002), **M** Portugal (Clarke and
755 Rendell, 2006) **N** French Mediterranean coast (Sabatier *et al.*, 2012), Dotted lines
756 show periods of enhanced storminess identified in the Cors Fochno reconstruction.

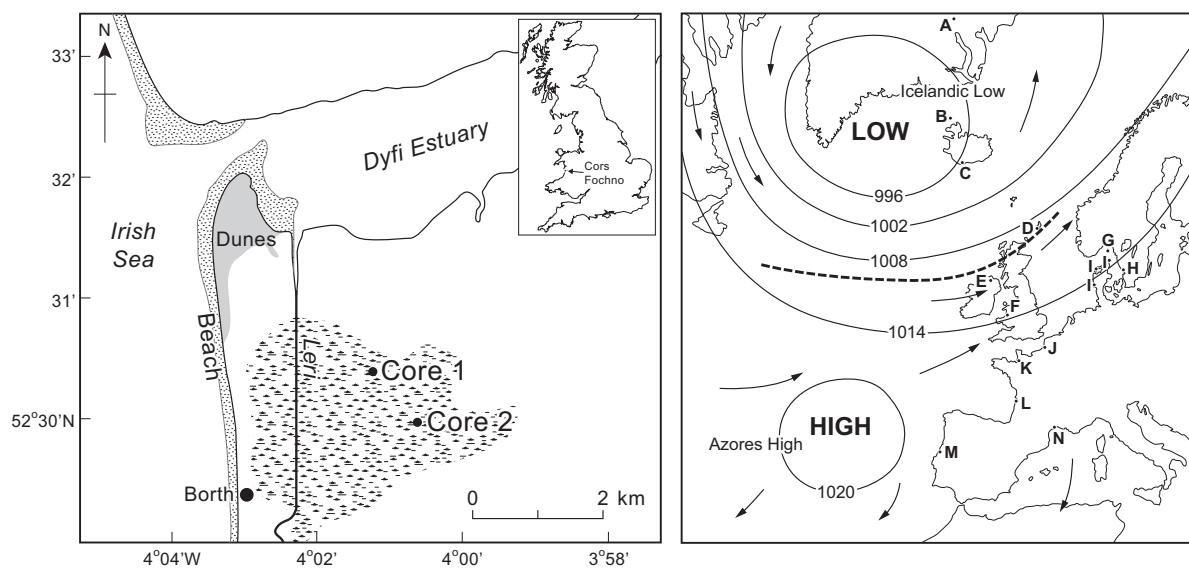
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758 Figure 7: Comparison between the Cors Fochno storm reconstruction and potential
759 forcings. From top: A) Cors Fochno storm reconstruction (*this study*), B) percentage
760 of Haematite Stained Grains (HSG) from 4-stacked records from the North Atlantic
761 as a proxy for IRD (Bond *et al.*, 2001), C) mean sortable silt (10-63 μ m) mean size as
762 a proxy for Iceland-Scotland Overflow Water (ISOW) current strength (Bianchi and

763 McCave, 1999), D) inferred water column stratification (density difference) based on
764 temperature and salinity reconstructions from two planktonic foraminifera
765 (*Globigerina bulloides* and *Globorotalia inflata*), which can be used as a proxy for
766 Sub-Polar Gyre strength (Thornalley *et al.*, 2009), E) reconstruction of the North
767 Atlantic Oscillation index (Olsen *et al.*, 2012), F) reconstructed Total Solar Irradiance
768 (TSI) (Steinhilber *et al.*, 2009). The dashed rectangles highlight the periods
769 discussed in the text that appear to have weaker ocean circulation and peaks in
770 storminess at the transitions."

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772 Figure 1:



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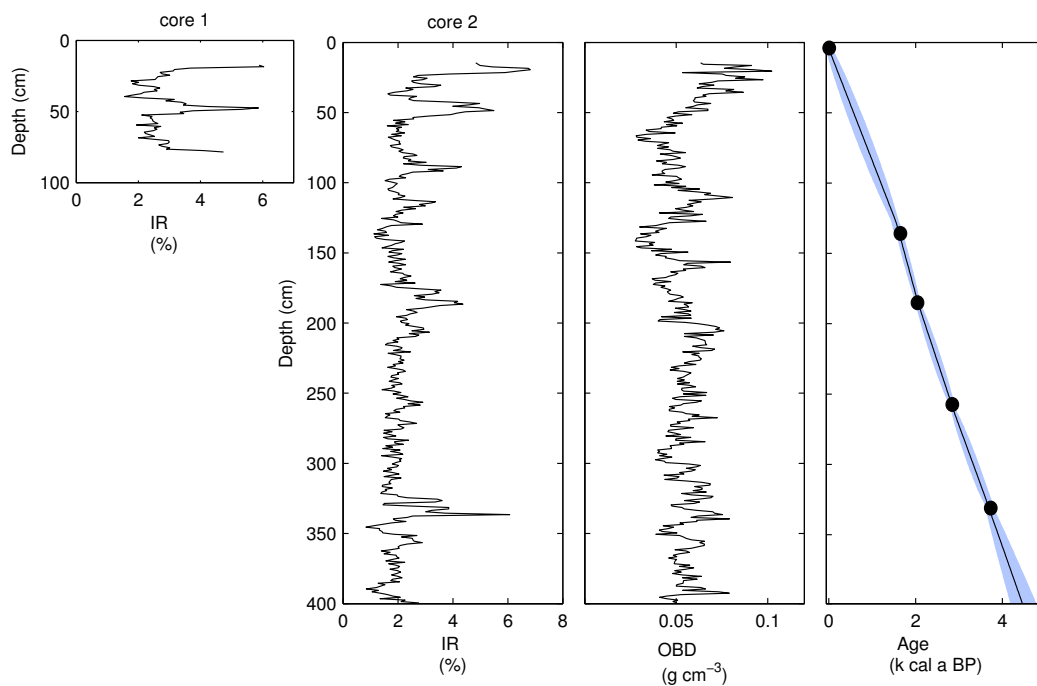
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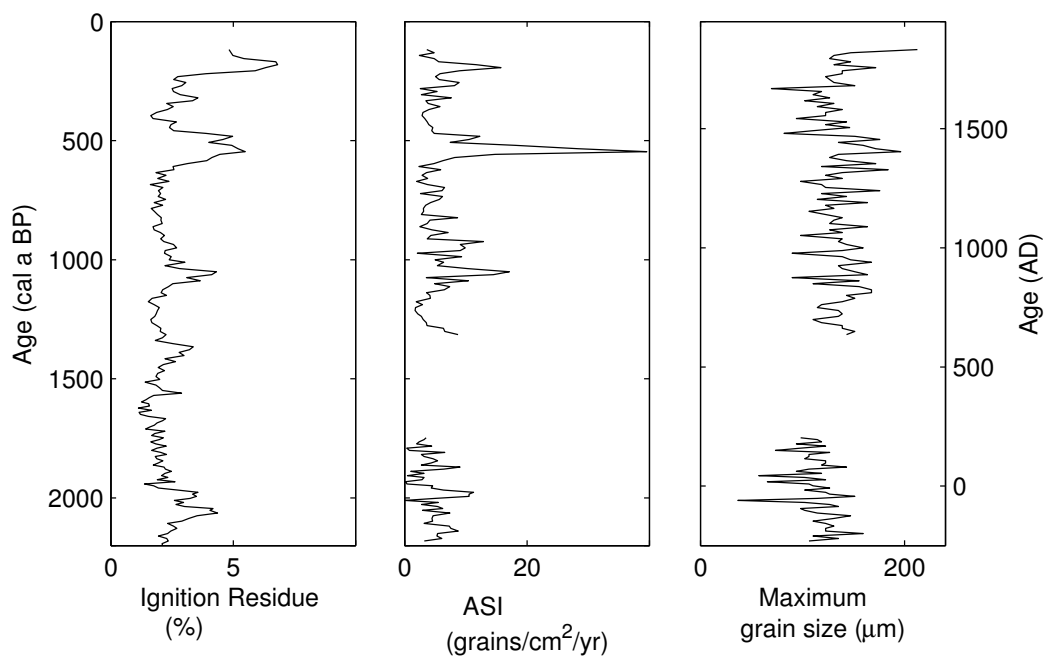
783 Figure 2:



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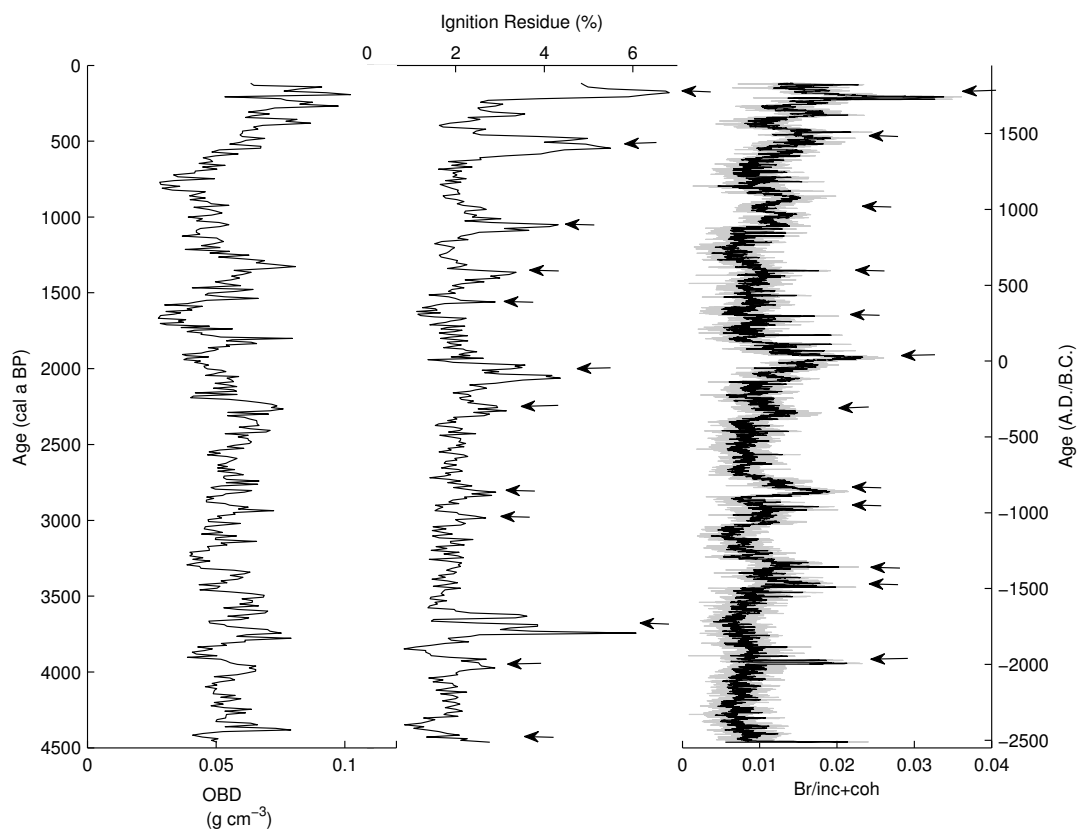
786 Figure 3:



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789 Figure 4:



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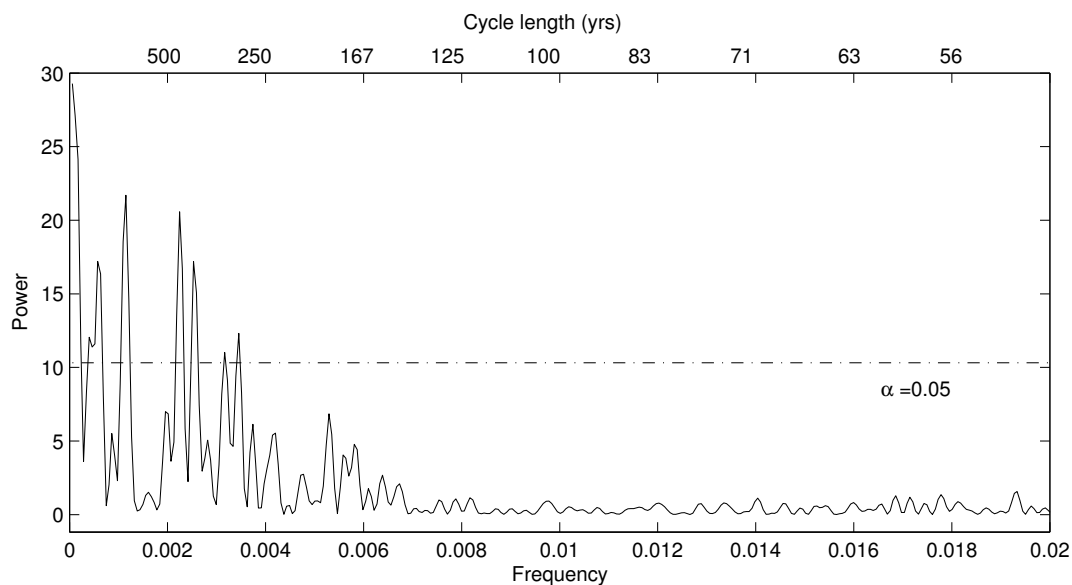
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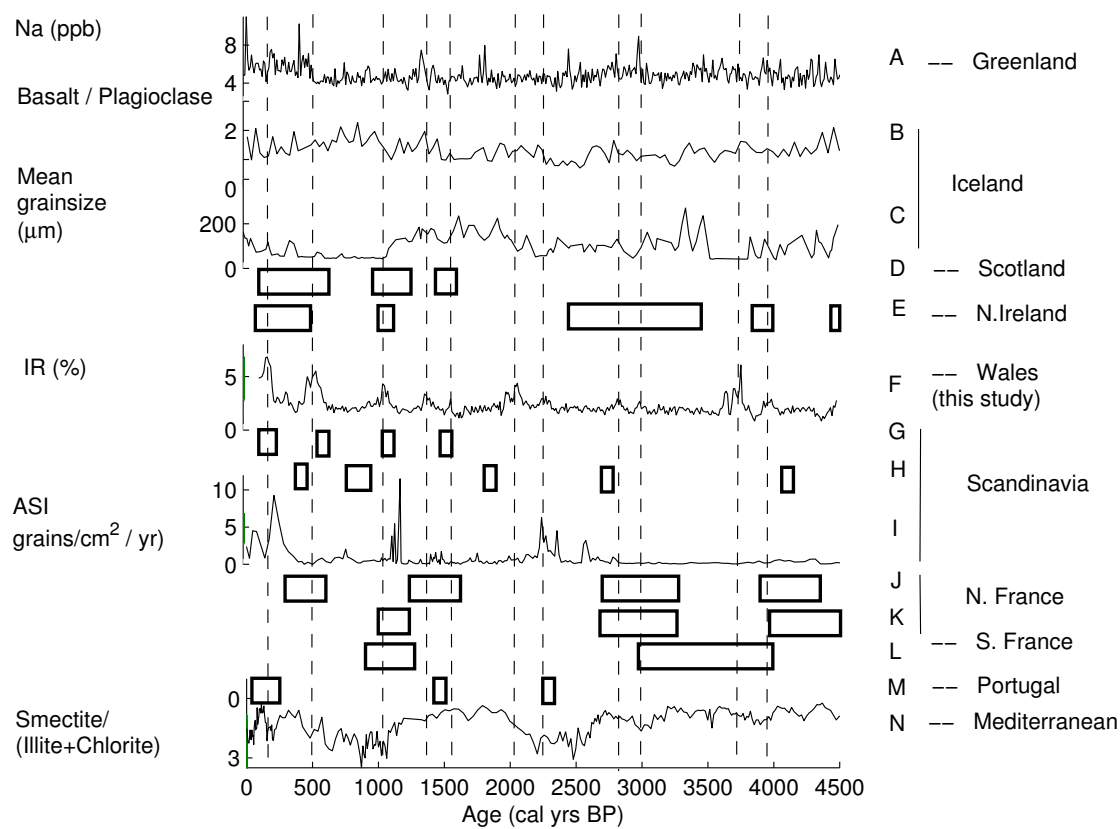
806 Figure 5:



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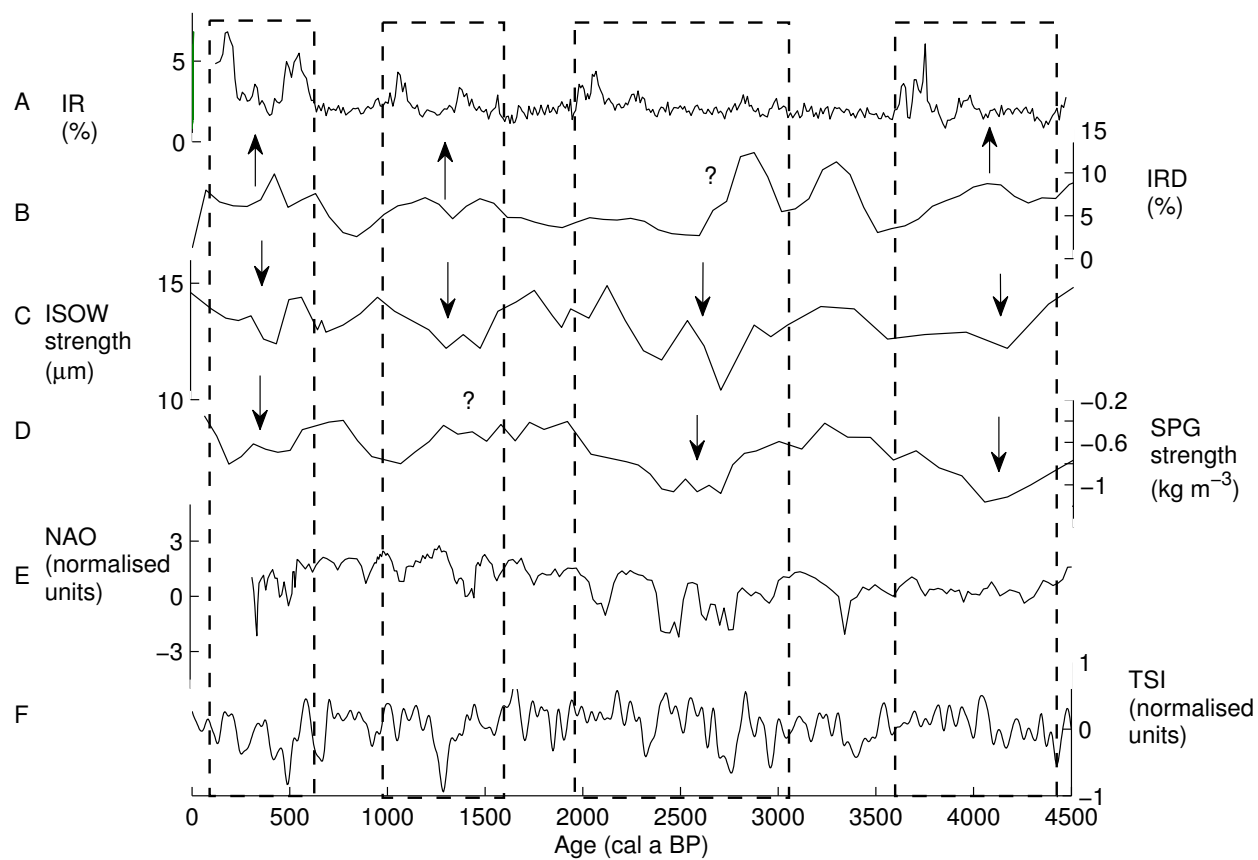
809 Figure 6:



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812 Figure 7:



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