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Aeolian sediment reconstructions from the Scottish Outer Hebrides: Late Holocene storminess and the role of the North Atlantic Oscillation

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Abstract

Northern Europe can be strongly influenced by winter storms driven by the North Atlantic Oscillation (NAO), with a positive NAO index associated with greater storminess in northern Europe. However, palaeoclimate reconstructions have suggested that the NAO-storminess relationship observed during the instrumental period is not consistent with the relationship over the last millennium, especially during the Little Ice Age (LIA), when it has been suggested that enhanced storminess occurred during a phase of persistent negative NAO. To assess this relationship over a longer time period, a storminess reconstruction from an NAO-sensitive area (the Outer Hebrides) is compared with Late Holocene NAO reconstructions. The patterns of storminess are inferred from aeolian sand deposits within two ombrotrophic peat bogs, with multiple cores and two locations used to distinguish the storminess signal from intra-site variability and local factors. The results suggest storminess increased after 1000 cal yrs BP, with higher storminess during the Medieval Climate Anomaly (MCA) than the LIA, supporting the hypothesis that the NAO-storminess relationship was consistent with the instrumental period. However the shift from a predominantly negative to positive NAO at c.2000 cal yrs BP preceded the increased storminess by 1000 years. We suggest that the long-term trends in storminess were caused by insolation changes, while oceanic forcing may have influenced millennial variability.

Key words: storminess, Holocene, North Atlantic Oscillation, storm track, Outer Hebrides, aeolian sand

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1. Introduction

During winter, the behaviour of the jet stream and atmospheric pressure centres over the North Atlantic, often expressed using the North Atlantic Oscillation (NAO) index, have a strong influence on European storminess, a term which encompasses both the frequency and intensity of storms. When there is a large pressure difference between the Azores High pressure and Icelandic Low pressure the NAO is positive and the storm track crosses northern Europe, whereas when the pressure difference is reduced the storm track is situated across southern Europe (Hurrell, 1995; Van Loon and Rogers, 1978). The northwest British Isles and western Norway are strongly influenced by increased storm frequency during winters when the NAO is in a positive state, whereas there is a much weaker relationship in the south of the UK (Andrade et al., 2008; Figure 1). There is some evidence that the influence from the NAO on storminess in Europe can vary over decadal timescales (Allan et al*.,* 2009; Hanna et al*.,* 2008), so there is reason to believe that over longer timescales and under different climates this relationship may have varied. We can investigate this possibility through the generation and analysis of palaeoclimate reconstructions. Greater understanding of the natural patterns and drivers of storminess is critical in informing future projections that suggest major changes in storminess but with low confidence (Stocker et al., 2013).

To determine the relationship between the NAO and storminess prior to the period with instrumental weather measurements (c.1870 A.D.), reconstructions have been made that span the Late Holocene. The past NAO has been inferred using climate proxy reconstructions from NAO-influenced regions; for example using opposing precipitation and drought patterns in Scotland and Morocco (Trouet et al., 2009). Long-term NAO changes have been inferred from a reconstruction of lake hypolimnic anoxia from Greenland, which is related to NAO-driven winter temperature and precipitation (Olsen et al., 2012) and may also be captured by an aridity index reconstruction from the Azores (Björck et al., 2006). Furthermore, marine source seasalt sodium (ssNa) concentrations measured in the

Greenland Ice Sheet Project Two (GISP2) core have been linked with instrumental pressure records in order to infer past variability of the Icelandic Low pressure (Meeker and Mayewski, 2002). Storminess reconstructions have been generated by dating sand dune activity and buried sand layers to identify periods of enhanced aeolian sand transport (Dawson et al., 2004; Gilbertson et al., 1999; Sommerville et al., 2003; 2007; Tisdall et al., 2013; Wilson, 2002; Wilson et al., 2004;), dating of cliff-top storm deposits (CTSDs) to show when severe storms deposited boulders on cliff tops (Hansom and Hall, 2009; Hall et al., 2006) and through the analysis of sand content within coastal peatbogs (Björck and Clemmensen, 2004; De Jong et al., 2006; Sjögren, 2009). As high precipitation often accompanies storms, it is also possible that patterns of storminess may be reflected by material washed into lakes (Oldfield et al., 2010).

The available Late Holocene reconstructions indicate that there is a changing NAO-storminess relationship through time. The reconstructed negative NAO during the Little Ice Age (LIA; 530-50 cal yrs BP; Lamb, 1995), which would direct the storm track across southern Europe, conflicts with evidence of high storminess in northern Europe at this time (Lamb, 1995; Raible et al., 2007; Sorrel et al., 2012; Trouet et al., 2009, 2012). The recent, long NAO reconstruction (Olsen et al., 2012) makes it possible to investigate the changing NAO-storminess relationship for the Late Holocene. This is a difficult task, however, as there may be differences between reconstructions resulting from the methodologies used and site-specific factors, as well as spatial variations in storminess. This NAO reconstruction spans the last 5000 cal yrs and shows a negative NAO between 4300-2000 cal yrs BP and persistently positive NAO between 2000 and 600 cal yrs BP (Olsen et al., 2012). This trend is corroborated by a transition to more arid conditions after 2000 cal yrs BP in the Azores aridity index reconstruction (Björck et al., 2006). A consistent NAO-storm relationship is supported by reconstructions showing a northwards shift in the storm track over the last 2000 cal yrs BP (Bakke et al., 2008; Giraudeau et al., 2010) but contrasts with those showing millennial variability in Europe, with a cycle of 1500-1700 years identified in some reconstructions spanning the Holocene (e.g. Debret et al., 2007; Fletcher et al., 2012; Sorrel et al., 2012).

We present a Late Holocene storminess reconstruction from the Outer Hebrides, northwest Scotland, selected as a region highly sensitive to NAO-driven variability, as a positive NAO index results in the storm track crossing the region and greater storm intensities (Andrade et al., 2008; Hurrell, 1995; Hurrell and Van Loon, 1997; Figure 1). In order to identify decadal-resolution and continuous changes, we generated aeolian sand records from two ombrotrophic peat bogs (using methods adapted from Björck and Clemmensen, 2004 and De Jong et al, 2006). The sites are close to beaches, sand dunes and machair (grasslands with sandy soils), which provide abundant sources of sand for aeolian entrainment during storms (Arens, 1996; Hotta et al., 1984; Kok et al., 2012). As with many terrestrial climate reconstructions spanning millennial timescales, local changes may have altered the depositional environments being analysed. In particular, there is a long history of human occupation of the Outer Hebrides (Ashmore et al., 2000; Bennet et al., 1990; Fossit, 1996; Garrow and Sturt, 2011; Henley, 2003; Sharples and Pearson, 1999), which may have caused disturbances resulting in enhanced aeolian sand erosion. In order to limit the influence of local factors, two sites have been sampled, compared to each other and to other reconstructions from the region. Furthermore, it has been shown that palaeo-environmental reconstructions from bogs can be influenced by micro-topography (which affects vegetation species, surface wetness variations and accumulation rate) and marginal effects (Bindler et al., 2004; Coggins et al., 2006; Hendon et al., 2001; Mauquoy et al., 2002; Robinson and Moore, 1999). Therefore, replicate cores in addition to a main long core from each bog are used to assess intra-site variability.

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2. Regional setting

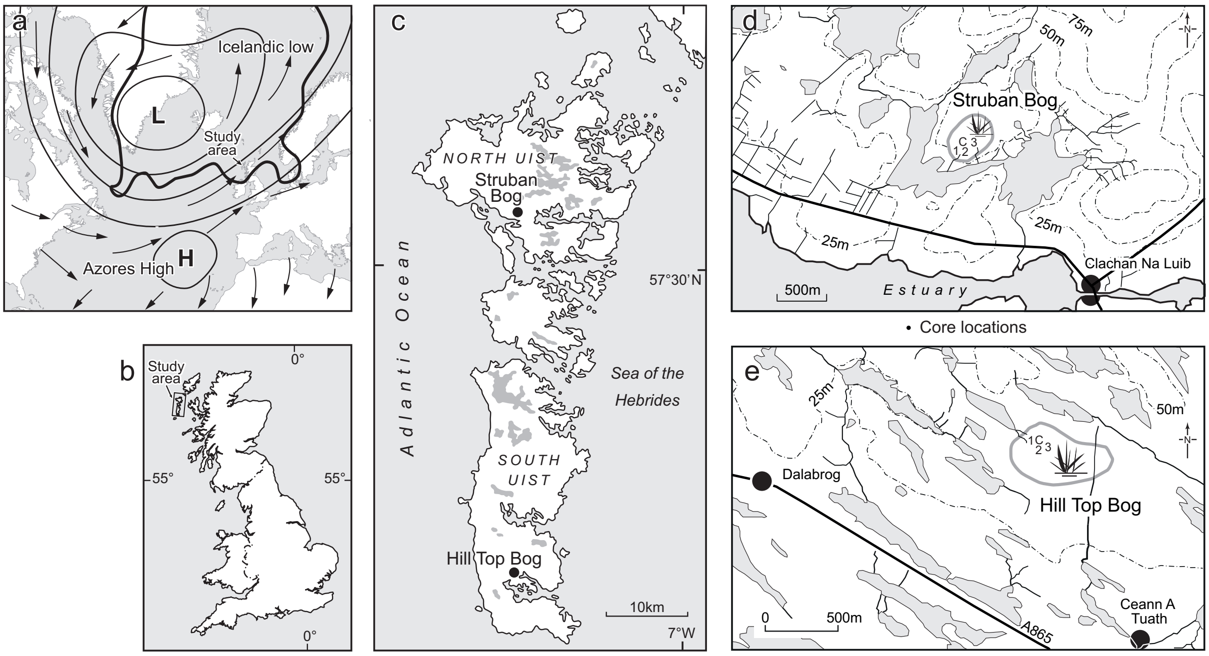

Figure 1: Location maps of Struban Bog (North Uist) and Hill Top Bog (South Uist) in the Outer Hebrides, UK. Inset a: map of the North Atlantic region with North Atlantic Oscillation pressure centres (Hurrell and Deser, 2010); the region enclosed by the thick line is the area with the strongest correlation between the NAO and cyclone frequency (Andrade et al., 2008). Inset b: location of the Outer Hebrides in the northwest of the UK. Inset c: locations of Struban Bog and Hill Top Bog in the Outer Hebrides. Inset d: site map of Struban Bog, with the locations of the main core (labelled C) and replicate cores (labelled: 1,2,3). Inset e: site map of Hill Top Bog, with the locations of the main core (labelled C) and replicate cores (labelled: 1,2,3).

The location of the Outer Hebrides in the eastern Atlantic Ocean provides the islands with a hyperoceanic climate, with little variation between winter and summer temperatures (Birse, 1971; Langdon et al., 2005). There are mild temperatures (annual averages between 7-11ºC), strong winds (annual averages of 8 m/s) and high precipitation (annual average totals of 1200 mm), particularly in winter (Benbecula Airport records, 1981-2010; Met Office, 2014).

Extensive peatlands and peaty soils occur across the islands (Angus, 1997; Hudson, 1991). These lie in the lee of the unique machair ecosystem lining the western coastline of the islands, which comprise beaches, sand dunes and grassland (Boorman, 1993) and provide the main potential source of windblown sand to the peatlands.

Two ombrotrophic peat bogs were sampled; Struban Bog on North Uist and Hill Top Bog on South Uist (Figure 1). Struban Bog is located at 57°33'35"N, 7°20'45"W, around 1 km from the tidal sands that separate North Uist from the island of Baleshare. The peat bog is on an outcrop of bedrock surrounded by four lochs, which are connected by streams. Towards the western, seaward end of this outcrop Struban Bog has formed in a bedrock depression. The peat bog extends over an area of approximately 0.05 km^2 and has a maximum depth of 6 m. Hill Top Bog on South Uist at 57°10'5"N, 7°20'52"W, is a peat bog around 3 km east of the coastline. The edges have been cut for fuel in the past, but the central peat bog is untouched and accumulating peat. The area of the peat bog is approximately 0.03 km^2 and it reaches a maximum depth of 6.1 m.

3. Material and methods

A core was sampled to 3 m depth from the centre of each bog in September 2011; two adjacent holes were used, with overlapping 10 cm segments, to limit disturbance caused by coring of the above sediment. To assess spatial variability in sand deposits within each bog, in June 2012 three replicate cores were sampled to depths of 1.5 m from across each peat bog. These were sampled using two coring holes, but without overlapping sections. Coring was carried out using a Russian Corer with a barrel length of 50 cm, with cores transferred to plastic tubing and wrapped securely in plastic film. These were transported horizontally to limit water movement, and stored in a cold store following fieldwork.

The 3 m main core chronologies were constructed using radiocarbon dating (AMS ^{14}C) of above-ground plant material (such as leaves, twigs and seeds) to prevent younger roots being dated (Kilian et al., 1995; Piotrowska et al., 2011). Radiocarbon dating on thirteen of the samples was carried out by the East Kilbride node of the NERC Radiocarbon Facility, with the ^{14}C analysed at the SUERC AMS Laboratory. A further sample was dated by the radiocarbon facility (^{14}CHRONO) at Queens University, Belfast. Age-depth models were produced using Bayesian analysis by OxCal version 4.2.3, using the IntCal13 calibration curve (Bronk Ramsey, 2009; Bronk Ramsey and Lee, 2013; Reimer et al., 2009, 2013). The median of the modelled 2-sigma age range was used to estimate the age for individual samples down the core.

The sand content of all the cores was initially analysed using loss-on-ignition to determine the organic and inorganic content (also called the Ignition Residue, IR). The cores were sampled at 1 cm resolution, at each depth using 5 cm^3 of wet peat. Loss-on-ignition was done by drying the samples in an oven overnight at 102 ºC and then igniting at 550 ºC for 4 hours, with weighing before and after each stage (Dean, 1974; Heiri et al., 2001). The Organic Bulk Density (OBD) was calculated (organic weight/wet volume), as some authors have shown that low values are associated with poorly humified peat, suggesting high moss production and/or a high water table (Björck and Clemmensen, 2004; Chambers et al., 2011; Yu et al., 2003).

To confirm that the IR reflects the sand content, the inorganic content of the peat was also analysed using a similar technique to the Aeolian Sediment Influx (ASI) method used previously (Björck and Clemmensen, 2004). The ASI method involves a count of the number of sand grains above a size threshold contained in a fixed-volume of peat (Björck and Clemmensen, 2004; De Jong et al., 2006). However, instead of counting sand grains, we wet-sieved the IR samples into two size fractions of medium sand (120-180 µm) and coarse sand (>180 µm) and weighed these. The lower boundary was selected to remove the influence of long-range dust and tephra, which is <100 µm in size (Bücher and Lucas, 1984;

Hall and Pilcher, 2002), with the >180 μm size selected to distinguish intense storms capable of transporting coarse sand, from less intense storms transporting finer particles. Before being sieved the IR samples were left overnight in 10% hydrochloric acid, before 10 ml of 30% hydrogen peroxide was added, at which point they were heated gently for 4 hours to stimulate a reaction. This process removed any remnant organics as well as carbonate. Shell ($CaCO_3$) is a constituent of sand in the Outer Hebrides (Ritchie, 1972) and this, once deposited in acidic bogs, would be expected to dissolve over time. Therefore we removed any remaining carbonate so that the upper peat, potentially with $CaCO_3$ remaining, would not be biased towards higher values. After sieving the samples were washed with distilled water into beakers and dried in an oven, before being cooled in desiccators. The beaker and sand fractions were weighed, the samples removed and the empty beakers reweighed, with the difference used to calculate the sand weight. This method minimised the loss of sand grains during the sieving and weighing process.

Changes in peat accumulation rate can potentially cause enhanced or reduced sand accumulation in a given layer. For example, 1 cm^3 of peat might accumulate over 10 or 50 years, which would most likely result in higher sand content in the latter sample. To account for this the sand influx was calculated on the dated main cores:

Sand Influx (g cm^{-3} yr^{-1}) = sand weight / wet volume / age span of 1 cm sample.

Finally, to assess the variability of the reconstructions, cycles in the ignition residue results of the main cores were analysed using the Lomb-Scargle spectral analysis method for unevenly spaced data (Lomb, 1976; Press and Rybicki, 1989; Scargle, 1982; calculated using Shoelson, 2001).

4. Results

The Hill Top Bog record spans 5-330 cm depth and consists of humified sedge dominated peat with small amounts of *Sphagnum*. The OBD values of <0.1 g cm^{-3} are typical of ombrotrophic bogs (De Jong et al*.,* 2009) and the results indicate no long term trends through the core (Figure 2). The 3.3 m main core was dated using eight radiocarbon dates (Table 1). The modelled age-depth curve for the main core, along with 2-sigma errors, are shown in Figure 2. The core spans the period from 4000 to -13 cal yrs BP and suggests there has been a fairly constant peat accumulation rate, apart from before 3500 cal yrs BP when the accumulation rate was slightly higher.

Table 1: Results of radiocarbon dated peat samples from the main core of Struban Bog and Hill Top Bog.

Sample Depth (cm)	Laboratory Code	$\delta^{13}C$ (‰)	Radiocarbon Age (^{14}C yr BP ± 1σ)	Median calibrated age (cal yrs BP) and 2σ range
Struban Bog				
37-38	SUERC-51107	-29.4	417 ± 35	492 (340-530)
62-63	SUERC-51108	-29.4	1024 ± 35	938 (801-987)
93-94	SUERC-51109	-30.0	1447 ± 35	1364 (1301-1474)
147-148	SUERC-41801	-26.8	2201 ± 37	2225 (2129-2314)
167-168	SUERC-41802	-29.4	2488 ± 37	2512 (2369-2639)
247-248	SUERC-41803	-29.4	3249 ± 37	3483 (3398-3568)
Hill Top Bog				
55-56	SUERC-51102	-22.8	449 ± 37	508 (446-540)
78-79	SUERC-51103	-25.2	845 ± 37	768 (693-903)

120-121	SUERC-51106	-26.8	1585 ± 35	1442 (1386-1530)
159-160	SUERC-41797	-27.7	1913 ± 35	1881 (1817-1970)
189-190	SUERC-41798	-28.6	2439 ± 37	2421 (2350-2606)
242-244	SUERC-41804	-25.8	3063 ± 37	3239 (3157-3326)
263-264	SUERC-43068	-26.0	3243 ± 35	3434 (3376-3507)
329-330	UB-No 19468	-24.7	3579 ± 37	3954 (3848-4083)

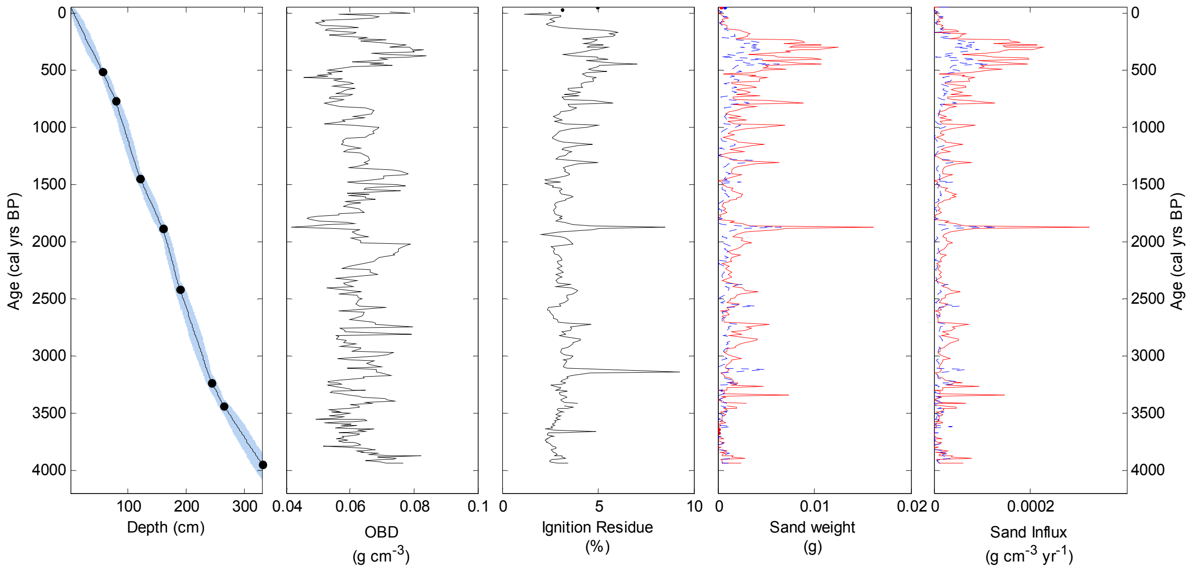

Figure 2: Hill Top Bog results. From left: age-depth model including dated points with shaded 2-sigma errors; Organic Bulk Density results; Ignition residue results; sand weight results for 120-180 μm sand fraction (red, continuous line) and >180 μm sand fraction (blue, dashed line); sand influx results to account for changes in the accumulation rate (line colours of sand fractions as previous).

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The reconstruction of sand deposition onto Hill Top Bog is based on the IR, sand weight and sand influx measurements (Figure 2). The constant rate of peat accumulation means that the sand influx results resemble the sand weight measurements, so peak concentrations of sand do not appear to have been caused by slower peat formation. Sand deposition on the bog was low but variable between 4000 and 1500 cal yrs BP, aside from small peaks at 3350, 3150, 2850, 2730, 2440 and 1580 cal yrs BP and a large peak at 1870 cal yrs BP. After 1500 cal yrs BP sand deposition gradually increased, with four peaks at c.1300, 1150, 980 and 780 cal yrs BP. Maximum sand deposition occurred between c.500 and 200 cal yrs BP.

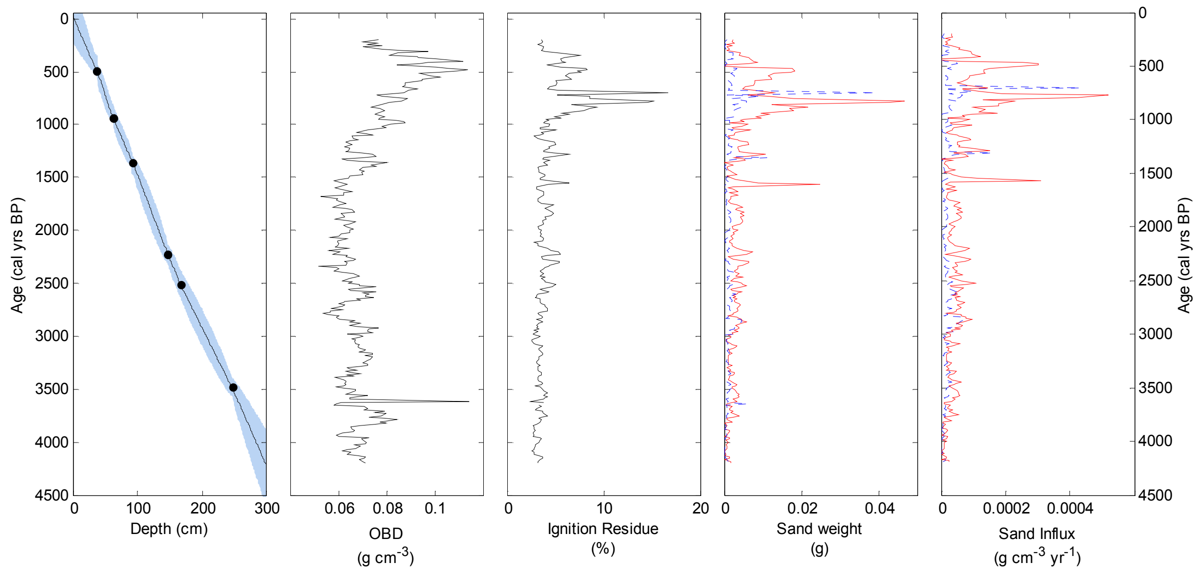

Figure 3: Struban Bog results. From left: age-depth model including dated points with shaded 2-sigma errors; Organic Bulk Density results; Ignition residue results; sand weight results for 120-180 µm sand fraction (red, continuous line) and >180 µm sand fraction (blue, dashed line); sand influx results to account for changes in the accumulation rate (line colours of sand fractions as previous).

The Struban Bog core spans from 16-300 cm depth; the peat is humified and sedge dominated with little *Sphagnum* content. The OBD results show small variations through most of the core before a gradual increase between 100 and 40 cm (Figure 3). As on Hill Top Bog, the OBD values are less than 0.1 g cm^{-3}. The 3 m main core was dated using six radiocarbon dates, which are summarised in Table 1. The modelled age-depth curve for the 3 m core, along with the 2-sigma errors, are shown in Figure 3. The core spans the period from 4200 to 200 cal yrs BP and appears to have had a steady peat accumulation rate through this time.

As on Hill Top Bog, the constant peat accumulation rate means the sand influx results on Struban Bog reflect the ignition residue and sand weight results (Figure 3). Sand deposition onto the bog was relatively low between 4200-1000 cal yrs BP, although small sand peaks occurred at c.3550, 2870, 2530, 2300, 2200, 1860 and 1190 cal yrs BP, with larger peaks at 1570 and 1280 cal yrs BP. After 1000 cal yrs BP sand deposition increased significantly, with three peaks at c.850-750, 550-470 and 400 cal yrs BP.

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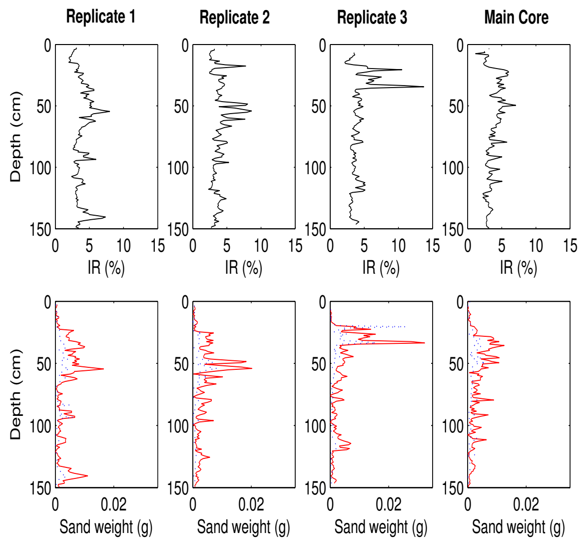

Figure 4: Hill Top Bog replicate core results. Ignition residue results (top row) and weights of the 120-180 µm (red, continuous lines) and >180 µm (blue, continuous lines) sand fractions (bottom row) from the replicate cores 1 to 3 and the main core (left to right).

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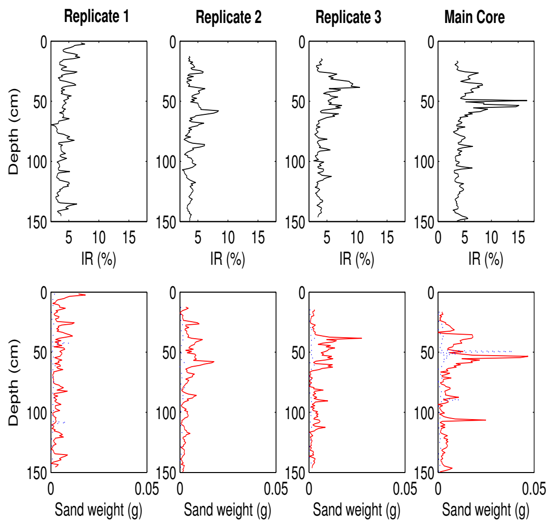

Figure 5: Struban Bog replicate core results. Ignition residue results (top row) and weights of the 120-180 μm (red, continuous lines) and >180 μm (blue, dashed lines) sand fractions (bottom row) from the replicate cores 1 to 3 and the main core (left to right).

The replicate core results of Hill Top Bog (Figure 4) and Struban Bog (Figure 5) both show that the long-term trends in sand content (IR and sand fraction weights) are similar across the bogs. There are some dissimilarities in the depths and in the smaller sand peaks, although this comparison is limited by a lack of age-control. On both sites there is a trend towards higher sand content around 50 cm depth. This peak is absent in Struban Bog replicate core 1 (Figure 5) and is at 30 cm depth rather than 50 cm in Hill Top Bog replicate core 3 (Figure 4). A small peak is also apparent in the Hill Top Bog replicate cores between depths of 100-150 cm, which is not present in the Struban Bog cores. The results indicate that small variations in sand concentrations differ across the bogs but the broad trends are similar between cores.

Finally, spectral analysis showed that in the Hill Top Bog reconstruction there are significant cycles of 2290 (range of 2670-2040), 1330 (1460-1180), 890 (1000-870), 730 (750-700), 330 and 290 (295-280) years (Figure 6). Significant cycles identified in the Struban Bog reconstruction are 1450 (range of 1610-1230) and 940 (1000-910) years (Figure 6).

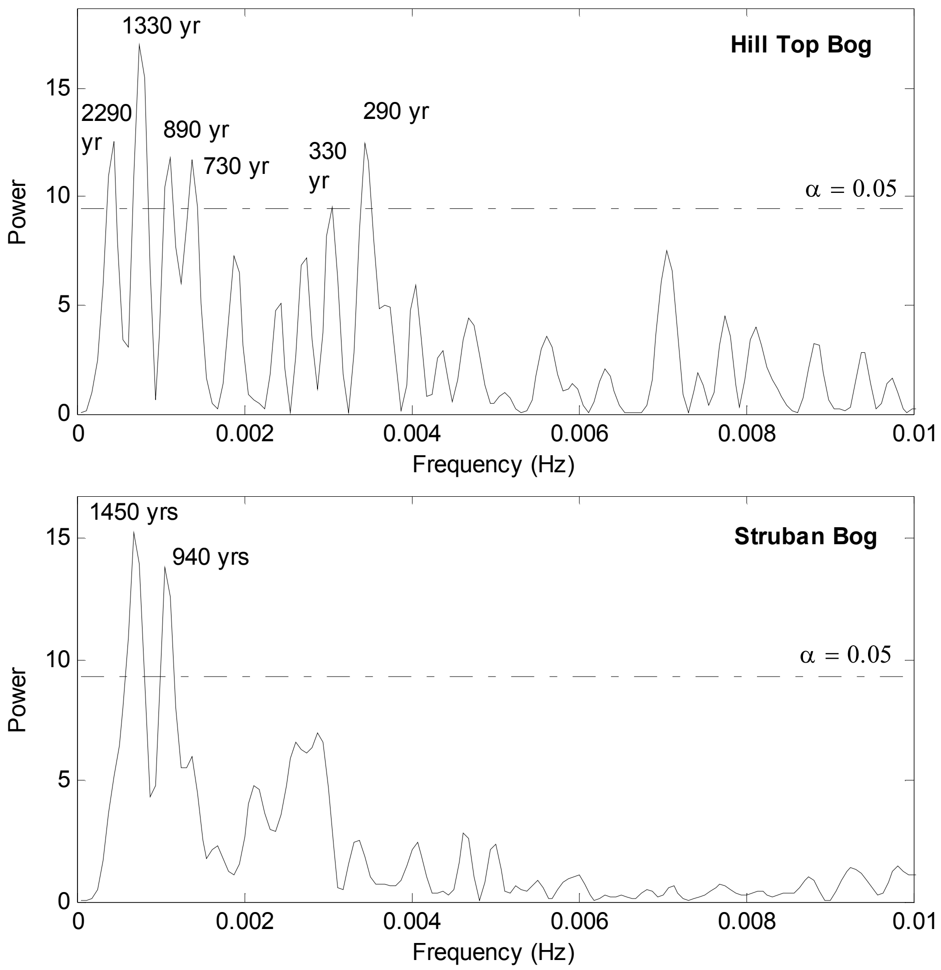

Figure 6: Lomb-Scargle spectral analysis of the ignition residue results of Hill Top Bog (upper figure) and Struban Bog (lower figure). Dashed line indicates the 95% significance level.

5. Interpretation of sand content

Replicate cores were sampled to assess the variability in sand content within each bog, although the cores have no chronological control. The results suggest that the large magnitude trends are similar across each bog and between each site, but that short-term, low magnitude peaks in sand content are less comparable, which may be the result of real differences in sand deposition or differences in accumulation rate. Previous palaeoclimatic reconstructions comparing multiple cores and other proxies such as heavy metals have shown similar results, with the same general trends but differences in the smaller variations (e.g. Bindler et al., 2004; Coggins et al., 2006). Intra-site differences have been suggested as being due to microform variability (such as hummocks/hollows/lawns), as the level of water logging or the vegetation type of these features may influence the trapping of heavy metals, or in this case sand (Bindler et al., 2004; Coggins et al., 2006). Furthermore, accumulation rates are likely to be locally variable (e.g. Robinson and Moore, 1999), so this may have influenced the depths at which the replicate cores have increased sand concentrations. Finally, the core margins have been found to produce more variable reconstructions than central bog cores (Hendon et al., 2001), which could explain why the Struban replicate core 1 is different. The similar large magnitude variations in the replicate cores support the contention that sand deposition is recorded across the peat bogs, giving greater confidence that the main cores are recording real changes in sand influx at each site.

As the two peat bog sites are 40 km apart they are not expected to share variability produced by local factors, such as sediment disturbance from human activities, but both should include a regional storminess signal. Small peaks in sand content occurred in both of the reconstructions at approximately 2850, 1860, 1580, 1300 and 1150-90 cal yrs BP, with larger peaks at c. 780 and just after 500 cal yrs BP. The sand peaks identified support previous evidence of aeolian activity on the Outer Hebrides; sand wedges formed locally during the LIA (Gilbertson et al*.,* 1999), particularly at the LIA onset c.500 cal yrs BP (Dawson et al*.,* 2004). Prior to the LIA a peak in sand influx during the Medieval Climate

Anomaly (MCA; 1050-600 cal yrs BP; Graham et al*.,* 2011) at c.800 cal yrs BP is supported by sand deposition phases identified on South Uist and Barra, as dating at the base of sand wedges gives ages of 940 ± 40 and 670 ± 130 cal yrs BP respectively for the onset of aeolian activity (Dawson et al., 2004).

The amplification of the sand peaks in the last millennium may have been the result of human disturbance, with erosion increasing the amount of sand available for aeolian transport. People have lived on the Outer Hebrides throughout the Late Holocene (Garrow and Sturt, 2011; Henley, 2003), with settlements situated on the machair between the middle of the first millennium B.C. and the 14^{th} century (Sharples and Pearson, 1999). For example, the settlement of Cladh Hallan was situated between Hill Top Bog and the coastline (Pearson et al., 2005). Since the 14^{th} century, settlements moved inland onto the peatlands (Sharples and Pearson, 1999), which coincides with the time when sand content increased in Hill Top Bog in particular, so this increase may reflect more proximal human activities. However, the cause and effect is uncertain, as it has been speculated that the LIA climate deterioration destabilized the sand dunes and was the reason for the settlement move (Sharples and Pearson, 1999). Furthermore, the climate change led to an enhanced use of marginal lands, with harvesting of seaweed and marram grass, which may have led to further destabilisation of the beaches and sand dunes (Angus and Elliot, 1992; Sommerville et al*.,* 2007). Pollen reconstructions from South Uist indicate that there was gradual deforestation between 4000 and 2500 cal yrs BP after which blanket peat dominated, which it is thought may have been caused by slight but sustained human activity in combination with climatic changes (Bennet et al., 1990; Fossit, 1996), The pollen records also show cereal cultivation after c.1700 cal yrs BP and in Loch Lang there was a sharp increase in eroded sediment after 480 cal yrs BP (Bennet et al., 1990; Fossit, 1996). This evidence therefore suggests that there was a recent intensification in human activity, which may have enhanced the sand deposition to the bogs during recent centuries through disturbance of sand sources and a more open landscape.

A different explanation for the trend towards higher sand content since 1500 cal yrs BP is that a transgression may have reduced the distance between the sand sources and the bogs. We can discount this, however, as a relative sea level reconstruction from the nearby island of Harris indicates that present sea levels were reached after 3100-2100 cal yrs BP (Jordan et al., 2010), approximately 1500 to 500 years before the sand content increased in the bogs.

There is a difference in the timing of the maximum sand content at each site. In the Hill Top Bog reconstruction the sand influx reached a maximum after 500 cal yrs BP, while in the Struban Bog reconstruction it was c.800 cal yrs BP. An explanation for this may be the topographic position of each bog in relation to the coastline and sand sources, as well as obstacles such as lakes. The amount of sand transport may have been influenced by the greater distance of Hill Top Bog than Struban Bog from the coastline, the aspects of the coastlines and barriers in the form of surrounding lochs (see Figure 1). Struban Bog is currently surrounded by lochs to the south, east and west, and as sand 70-500 μm in size travels by saltation (Bagnold et al., 1937; Kok et al., 2012), which cannot occur on water, it is possible that the reconstruction is biased towards storms with northerly winds. Alternatively, warmer periods may have resulted in lower lake-levels, thus reducing the size of the water barrier, or colder periods may have had more frequent snow and ice cover, enabling greater sand transport through niveo-aeolian processes (Björck and Clemmensen, 2004).

The climatic conditions that caused the sand deposits in the bogs is uncertain, as they may have occurred in severe storms but may also reflect periods of drier, anticyclonic weather. Moisture creates inter-particle forces so wet sand requires higher wind speeds to initiate movement than dry sand, however once initiated wet and dry sand are transported equally (Kok et al., 2012; Chepil, 1956; McKenna-Neuman and Nickling, 1989; Hotta et al., 1984). Research into sand transport on beaches in the Netherlands found that moderate precipitation, such as showers, can result in greater sand transport than dry conditions, as the raindrops can act to dislodge grains, however once saturated sand transport is negligible

(Arens, 1996). This research concluded that moderate storm events, with less rainfall, are more important for sand transport than extreme events with more rainfall (Arens, 1996), which may be significant for the interpretation of the Outer Hebrides reconstructions. On the other hand, the amount of sand within each centimetre layer of the core is small (on average 0.018 g or 3.7% of the dry weight), which may show that sand transport onto the bog sites is uncommon, perhaps only occurring during storms with extreme winds but unsaturated ground. We suggest that the bogs locations far inland, and with loch barriers, may mean that strong winds were required for sand to reach the sites.

As highlighted by the above discussion, separating the climatic, human and local environmental causes of aeolian sand deposition remains a challenge. To address the issues connected with local changes (particularly human activity) we have detrended the two records before the following discussion, as it appears possible that the long term increase in sand deposition may have resulted from greater human activities, with increasing exposure of the sand sources and reduced forests. We have also combined the two records, which we did by normalising and downsampling to the same resolution, and then calculating their mean values at each 20 year interval (Figure 7A). As both sites should be influenced by the same storms, combining the records will remove or reduce peaks seen in single records. In addition, the error associated with the combined proxy record was calculated (see method in supplementary information) to show the range of possible ages for the peaks. The resulting reconstruction (Figure 7A) indicates that there was greater sand deposition at c.3150, 1900, 800 and 500 cal yrs BP.

6. Discussion

6.1. Late Holocene storminess in northwest Europe

The interpretation of the combined sand influx reconstruction as a storminess proxy is supported by comparison with other published records from the Outer Hebrides. The uppermost peaks at c.500-200 cal yrs BP were deposited during a period of historically documented severe storms that buried in sand the Outer Hebridean settlements of Udal in 1697 A.D. and Baleshare in 1756 A.D. (Gilbertson et al., 1999; Lamb, 1984; 1991). Furthermore, an increase in maritime plant communities on St Kilda during the LIA suggests more saline conditions caused by sea-spray, probably linked to high storminess (Walker, 1984).

To investigate the patterns of storminess in northwest Europe, we have compiled the results of reconstructions from elsewhere in the region (Figure 7). Those that are based on blown sand (such as dated sand wedges and sand dune formation) from Scotland and northern Ireland have been combined (Dawson et al*.,* 2004; Gilbertson et al*.,* 1999; Sommerville et al*.,* 2003; Tisdall et al., 2013; Wilson et al*.,* 2004) and for each 100 year time slice the number of these studies that suggest high storminess was calculated (Figure 7C). The results, as well as those of CTSD's from Shetland (Figure 7D; Hansom and Hall, 2009), support the common assertion that storminess was high during the LIA, as well as c.1500, 2800-2400 and 3300 cal yrs BP. A precipitation proxy from lake inwash also shows similar increases at c.600, 2700-2200 and 3300 cal yrs BP (Figure 7E; Oldfield et al., 2010). The timings of these increases in storminess differ from those shown by the Outer Hebrides reconstruction, aside from the peak during the LIA. This is possibly because the reconstructions have been formed using a range of different methods that may be influenced differently by storm frequency and intensity. For example, CTSDs reflect severe storm events (Hansom and Hall, 2009), while phases of sand dune development are more likely to reflect changes in the frequency of moderate intensity storms, as dry sand can be transported by relatively weak winds (Kok et al., 2012). The reconstructions may also have been influenced by temperature changes, for example the Cairngorms lake inwash reconstruction is thought to be connected to enhanced freeze-thaw weathering, as well as wind erosion (Oldfield et al., 2010). Similarly, lower precipitation may have influenced sand dune based reconstructions, as water deficits can cause reduced vegetation cover, leading to greater sand erosion (Lancaster, 1989; Wasson and Nanninga, 1986). Therefore many of the reconstructions available from the region may not solely reflect storminess, with the exception of the CTSD reconstruction.

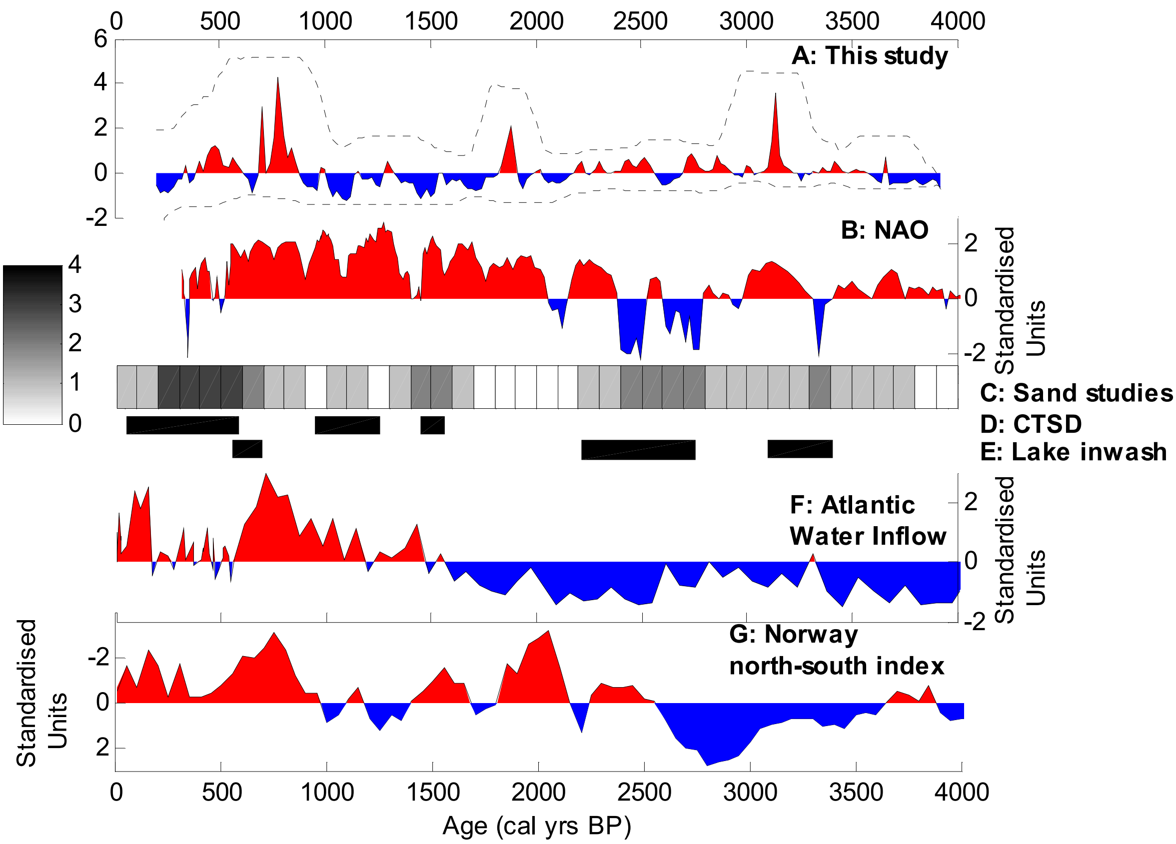

Figure 7: Comparison between regional storminess and precipitatio n reconstructions and the NAO. A: storminess reconstruction from the Outer Hebrides (this study), based on the detrended and standardised mean sand influx values from the Hill Top Bog and Struban Bog reconstructions, with dashed lines reflecting the confidence interval (see supplementary

information). B: NAO reconstruction, based on temperature and precipitation driven hypolimnic anoxia from a lake in Greenland (Olsen et al., 2012). C: numbers of studies from Scotland and northern Ireland showing phases of sand transport (sand dunes or sand wedges) within 100 year time slices (Gilbertson et al., 1999; Dawson et al., 2004; Wilson et al., 2004; Sommerville et al., 2003; Tisdall et al., 2013). D: Cliff Top Storm Deposits from sites on Shetland (Hansom and Hall, 2009). E: minerogenic input into a remote lake in the Cairngorms thought to reflect catchment erosion during climate deteriorations, associated with enhanced debris flows, slope wash and wind action (Oldfield et al., 2010). F: standardised reconstruction of Atlantic Water Inflow through the Iceland-Scotland Ridge in the Norwegian Sea, based on concentrations of the coccolith *Gephyrocapsa muellerae*, thought to be a proxy for the strength and position of the westerlies (Giraudeau et al*.,* 2010). G: north-south index of the position of the westerlies, based on the difference between two reconstructions of winter precipitation from glaciers from northern and southern Norway. Note that for comparison with the other records the y-axis has been inverted: negative values indicate lower precipitation gradients and therefore a northerly position of the westerlies (Bakke et al*.,* 2008).

The Outer Hebrides reconstruction has greater similarity with reconstructions from the Norwegian Sea region that capture changes in the storm track. Atlantic Water Inflow in the Norwegian Sea increases when the storm track is farther north; the increased inflow since around 1500 cal yrs BP (Figure 7F; Giraudeau et al., 2010) is around the time when storminess began to increase in the Outer Hebrides. Similarly, a north-south index of westerly airflow from Norway also suggests a northwards shift in the storm track (Figure 7G; Bakke et al., 2008), which began c.2500 cal yrs BP with a maximum after 1000 cal yrs BP. In addition, a storminess reconstruction from northern Norway, also based on sand content within a coastal peat bog, showed an increase after c.1500 cal yrs BP (Sjögren, 2009). As these studies are not based on aeolian sand transport (and the AWI is uninfluenced by temperature), these support the hypothesis that the Outer Hebrides reconstruction reflects storminess rather than anticyclonic circulation patterns.

6.2. Storminess and the North Atlantic Oscillation

The strongest spatial positive correlation between the NAO and cyclone occurrence during the instrumental period is over an area including western Scotland, in particular the Outer Hebrides, the Norwegian Sea and the west coast of Norway (Andrade et al*.,* 2008; Figure 1), which may explain why the reconstructions of storminess from these regions are similar. If considering the last 1000 years, the transition from high to moderate storminess in the Outer Hebrides at the MCA-LIA transition, when the NAO switched from positive to neutral/negative (Trouet et al., 2009), supports a positive NAO-storminess relationship consistent with the instrumental period. On the other hand the ssNa changes in the GISP2 core indicate deeper Icelandic Low pressures and a positive NAO during the LIA compared to the MCA (Meeker and Mayewski, 2002), which contradicts the reconstructed higher storminess in the North Sea region during the MCA. However the GISP2 reconstruction also indicates that the LIA had intensified circulation (Meeker and Mayewski, 2002), which is supported by evidence of enhanced storminess at this time elsewhere in Scotland and Ireland (Figure 7), as well as Europe (e.g. Sorrel et al., 2012).

When considering the last 4000 years the reconstructions from the Norwegian Sea region (*this study*; Bakke et al., 2008; Giraudeau et al., 2010) differ from the Olsen et al. (2012) NAO reconstruction. The suggested NAO transition from a negative to positive index c.2000 cal yrs BP (Björck et al., 2006; Olsen et al., 2012) does not correspond closely with an increase in storminess in the records from the Norwegian Sea, as the storminess reconstructions show a much later transition to enhanced storminess c.1000 cal yrs BP (Figure 7). This disparity raises a number of questions, including whether the NAO is the major influence on storminess over millennial timescales and other drivers may have caused this delay. Alternatively, this lag may be an artefact of changes in the sensitivity of the NAO reconstructions, given that the NAO can exhibit non-stationarity over time that can alter the influence of the NAO on climate proxies (Schmutz et al., 2000).

The timings of high storminess indicated by the reconstructions from elsewhere in Scotland and Northern Ireland (c.3300, 2800-2400, 1500 and 500 cal yrs BP; see references in Figure 7) all coincide with reconstructed negative NAO events. High storminess during the LIA has been suggested as being the result of intense but infrequent storms caused by steep temperature gradients during periods of negative NAO (Raible et al., 2007; Trouet et al., 2012). When there is a meridional, meandering circulation of the jet stream (such as when the NAO is negative) temperature contrasts both vertically (in response to ocean temperatures) and between air masses can cause increased cyclogenesis (Betts et al., 2004; Van Vliet-Lanoë et al., 2014). Furthermore, cold polar waters further south in the North Atlantic and extensive sea ice may have caused the steepened temperature gradient by cooling the air (Lamb, 1979), leading to high pressure development and negative NAO conditions, which would have shifted the storm track southwards and increased storm intensity (Dawson et al., 2002). Pulses of cold water extending across the North Atlantic earlier in the Holocene at c.2800 and 1400 cal yrs BP have been linked with changes in storminess elsewhere in Europe, with southward shifts in the oceanic fronts and/or sea ice extent again hypothesised as the cause of atmospheric reorganisations (Bond et al., 1997; Pena et al., 2010; Sabatier et al., 2011; Sorrel et al., 2012). The timings of these oceanic changes resemble the increases in storminess at 2800-2400 and 1500 cal yrs BP (Figure 7 C,D,E), supporting the link between ocean and climate variability. Furthermore, the cycles of 1450 (1610-1230) and 1330 (1460-1180) years identified in the Struban Bog and Hill Top Bog reconstructions respectively may reflect this oceanic forcing, which has a cyclicity of c.1450 years (Bond et al*.,* 1997), within the error range of the cycles identified here. Similarly cycles of 940 (1000-910) and 890 (1000-870) years are also present in the Struban Bog and Hill Top Bog reconstructions respectively, and these resemble a 1000 year cycle identified in proxies of North Atlantic Deep Water circulation strength (Chapman and Shackleton, 2000).

The increasing trend in storminess in the Outer Hebrides and reconstructions from the Norwegian Sea region, and suggested northwards storm track shift, may have been caused by increasing winter insolation, as also suggested by other research on mid-latitude climate reconstructions that are related to atmospheric circulation (e.g. Kirby et al., 2007; Moros et al., 2004; Bakke et al., 2008). Brayshaw et al. (2010) suggest, based on theory and model simulations, that changes in winter insolation from the Mid to Late Holocene would have resulted in expansion of the Hadley Cell, a steeper latitudinal temperature gradient and a northwards shift in the North Atlantic storm track. Therefore it is suggested that the long-term shift in storm tracks may have been the result of insolation-driven changes in the temperature gradient, while millennial variability may have resulted from changes in ocean temperature gradients.

7. Conclusion

Storminess reconstructions, based on aeolian sand deposits from two ombrotrophic peat bogs on the Outer Hebrides, allow the patterns of storminess during the Late Holocene to be investigated. Multiple cores from each site show that there is intra-bog variability of the high-frequency, low-magnitude changes in sand content, probably due to bog micro-topography. However, the long-term trends in sand content are replicated. The two reconstructions have been combined to show the regional signal, which indicates that storminess increased after 1000 cal yrs BP.

Over the last millennium reconstructed storminess was higher during the MCA than the LIA, reflecting the positive to negative NAO transition (Trouet et al., 2009) and supporting that there has been a consistent NAO-storminess relationship. In areas such as the Outer Hebrides, the Norwegian Sea and western Norway there appears to have been enhanced storminess in the last 1000 years compared to 4000-1000 cal yrs BP. The reconstructed negative-to-positive NAO shift (Olsen et al., 2012) precedes this increase by 1000 years, raising questions over the NAO influence on storminess over the Late Holocene. However reconstructions from elsewhere in Scotland and Northern Ireland (Gilbertson et al., 1999; Dawson et al., 2004; Wilson et al., 2004; Sommerville et al., 2003; Tisdall et al., 2013; Oldfield et al., 2010; Hansom and Hall, 2009), where the present-day NAO-storm relationship is weaker, have indicated that storminess increased when the NAO was negative. This may have been the result of increased storm intensities caused by steeper latitudinal temperature gradients at these times (e.g. Trouet et al., 2012). We speculate that the inferred northwards storm track shift during the Late Holocene may have been the result of insolation changes, while there is some evidence that oceanic drivers caused the millennial-scale variability in storminess.

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Supplementary Information

Method used for combining two reconstructions and calculating the proxy error

We sought to combine the Hill Top Bog and Struban Bog reconstructions of sand influx and calculate the error in the resulting proxy reconstruction given the age errors. To do this we first calculated the proxy error of each reconstruction separately. For each 1 cm sample we randomly selected 300 possible ages from within the 2 sigma age errors. Using this data we identified the ages that fell within each 10 year interval from 3920-200 cal yrs BP (the period shared by the two records) and compiled the sample depths associated with these. For each 10 year interval we determined the maximum and minimum IR values associated with the compiled depths, which gave us the maximum and minimum proxy error for each age interval. These were smoothed and downsampled to the 20 year resolution. The maximum proxy error reflects the potential range of timings of peaks in the IR. To combine the two reconstructions we smoothed and downsampled the IR results to a 20 year resolution. We then normalised the IR and proxy error records and calculated the averages between the two bogs.

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