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The impact of North Atlantic storminess on western European coasts: A review

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Abstract

Instrumental and documentary records of storminess along the Atlantic coast of western Europe show that storm activity exhibits strong spatial and temporal variability at annual and decadal scales. There is evidence of periods of increased storminess during the Little Ice Age (LIA) (AD 1570–1990), and archival records show that these periods are also associated with sand movement in coastal areas. Independent evidence of sand movement during the LIA is derived from dating the coastal sand deposits, using luminescence or radiocarbon methods. The Holocene record of sand drift in western Europe includes episodes corresponding to periods of Northern Hemisphere cooling, particularly at 8.2 ka, and provides the additional evidence that these periods, like the LIA, were also stormy.

1. Introduction

There is growing evidence that periods of sand drift and dune development provide proxy records of the impacts of storms in coastal areas (Wilson and Braley, 1997; Wintle et al., 1998; Clemmensen et al., 2001a, b; Clarke et al., 2002; Dawson et al., 2004; Clemmensen and Murray, 2005). Sand drift and dune building require periods of strong, persistent winds, a supply of sand and sparse vegetation cover (Short and Hesp, 1982; Carter et al., 1990). Strong winds are associated with the passage of synoptic scale storms (Carnell et al., 1996) during which wind velocities exceed the critical threshold for entrainment and transport of the sand-sized grains which form coastal dunes (Carter et al., 1990). Storms are extreme events and the term ‘storminess’ has been used to encompass both the frequency and the intensity of storms (Carnell et al., 1996).

An understanding of the patterns of past storminess is particularly important in the context of future anthropogenically driven climate change (Carnell et al., 1996), with predictions of increased storm frequency and sea level rise by the end of the current century (Lozano et al., 2004). Although the combination of high winds, increased wave activity and storm surges will have impacts on vulnerable coastal areas (Lozano et al., 2004), the empirical basis for these predictions remains problematic (WASA Group, 1998; Osborn, 2004; von Storch and Weisse, 2007). A long-term proxy-based record of storminess, extending back into the Holocene would provide both a basis for the evaluation of trends in past storminess and a firmer foundation for future predictions.

This paper reviews the evidence for storm activity across the North Atlantic region derived from instrumental records together with archival evidence of storm impacts. This information is compared with sedimentological and chronological evidence of sand movement and dune building. The focus on western Europe reflects the availability of historical records of both storms and sand movement (Lamb and Frydendahl, 1991), and the numerous studies which have applied absolute dating techniques, particularly luminescence, to coastal dunefields to develop independent chronologies of sand deposition which is a proxy for storminess (Wintle et al., 1998; Clarke et al., 1999, 2002; Clemmensen et al., 2001a, b; Murray and Clemmensen, 2001; Wilson et al., 2004; Clemmensen and Murray, 2005). This information is then used to evaluate

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the extent to which evidence of sand drift and dune building in western Europe has potential as a storminess proxy for the North Atlantic.

2. Historical records of Atlantic storminess

Records of storm activity are reviewed in three sections: continuous instrumental records that extend back up to 200 years, archival records that cover the last 800 years and proxy records covering the last 2000 years.

2.1. Instrumental records

Atlantic storminess has been characterised in different ways using extreme wind velocities, storm surge heights on land-fall and by sea level pressure fields (von Storch and Weisse, 2007). Storminess is normally defined in terms of the number of days with strong winds in the range 7 (13.9–$17.1\,\mathrm{m\,s^{-1}}$) to 11 ($>28.4\,\mathrm{m\,s^{-1}}$) on the Beaufort wind scale for one or more measurements during a 6-h period over the course of any single day (Qian and Saunders, 2003). Storm tracks have been identified and analysed using synoptic charts to track individual weather systems over time periods of hours to days, or by employing statistical methods and gridded datasets (Hoskins and Hodges, 2002). Although wind speed is the most obvious measure of storminess, little use is made of long-term observational wind speed data owing to inhomogeneities in such records (Schmith et al., 1998; Smits et al., 2005; von Storch and Weisse, 2007). Instead storminess tends to be characterised using a range of instrumental proxies for near surface wind speeds, e.g. using surface air pressure readings to calculate the geostrophic wind speed (Alexandersson et al., 1998), using pressure tendencies (Schmith et al., 1998), or making use of the gridded wind speed reanalysis data sets such as those available from the National Center for Environmental Prediction, the National Center for Atmospheric Research (NCEP-NCAR) (Hoskins and Hodges, 2002) and the European Centre for Medium-range Weather Forecasting (ECMWF) (Bartholy et al., 2006). Coastal storminess has also been reconstructed using hourly tide-gauge data (Orford, 2001; Betts et al., 2004). Instrumental records of storminess have been taken on land, near-shore on lighthouses and out at sea on individual vessels and recorded in ships' log-books (García Herrera et al., 2003). Ship-borne observations provide useful additional information allowing reconstruction of sea level pressure changes across the ocean although the spatial coverage within the North Atlantic is still limited.

Spatial and temporal variability in Atlantic storminess based on continuous instrumental records using hourly or daily data are summarised in Table 1. From this, we can see that storminess during the 19th and 20th centuries exhibits no clear trends. Storminess shows significant temporal and spatial variability (Lozano et al., 2004) that may or may not be related to the status of the North Atlantic Oscillation (NAO) (Dawson et al., 2002; Betts et al., 2004; Matulla et al., 2008). The difficulty of correlating storms with the NAO is partly one of temporal scale, i.e. there are inherent difficulties in relating transient synoptic scale storms, with durations of 2–3 days, with monthly averaged NAO indices.

2.2. Archival records

Archival records provide a rich source of evidence for Atlantic storm activity. In his description of the county of Cornwall, in southwest England, Sir Richard Carew writing at the start of the 17th century notes that:

> The Countrie is much subject to stormes, which fetching a large course in the open Sea, doe from thence violently assault the dwellers at land, and leave them uncovered houses, pared hedges and dwarfe-growne trees, as witnesses of their force and furie; yea, even the hard stones, and yron barres of the windowes, doe fret to be so continually grated. One kind of these stormes they call a flaw, or flaugh, which is a mightie gale of wind, passing suddainley to the shore. And working strong effects, upon whatsoever it encountreth in his way (Carew, 1602, p. r6).

The range of information available from archival sources includes lighthouse keepers' records (Dawson et al., 2002), weather diaries (Golinski, 2001; Naylor, 2006), event accounts (Defoe, 1704; Shields and Fitzgerald, 1989; Brayne, 2002), shipwreck registers (e.g. Lloyd's List 1741–1914; Lloyd's Register of Shipping 1764–1995), regional descriptions (e.g. Carew, 1602), travellers' tales and discontinuous instrumental records (Zuidervaart, 2005, 2006; Naylor, 2006). Lamb and Frydendahl (1991) used many of these sources to reconstruct a series of individual storms affecting the northwest Atlantic and the North Sea over the period AD 1570–1990. These storms are classified using a storm severity index (SSI) that is defined as: $V_{\max}^3 \times A_{\max} \times D$ where: $V_{\max}$ is the maximum surface windspeed in knots (wind power is expressed as the cube of windspeed), $A_{\max}$ is the maximum area affected by damaging winds in units of $10^5\,\mathrm{km}^2$ and D is the duration of the storm in hours (Lamb and Frydendahl, 1991); thus, the storm of AD 1703 (Defoe, 1704), that affected much of England, has an SSI value of 9000. The SSI approach allows the impacts of storms to be compared directly so that the October AD 1987 storm, the most severe storm in the last 50 years affecting an area covering both southern England and northern France, has an SSI of 8000. Six of the most severe storms within this record (SSI > 6000) occurred in the periods AD 1690–1720 and AD 1790–1840 (Lamb and Frydendahl, 1991). A mechanism linking increased winter storminess to Little Ice Age (LIA) cooling (AD 1570–1900) was suggested by Lamb (1995), who hypothesised that a southward spread of sea ice and polar water would create an increased thermal gradient between 50°N and 65°N, intensifying storm activity in the North Atlantic.

Table 1
Studies of trends in Atlantic storminess derived from instrumental records

Time period of observations	Location	Conclusions of study	Reference
1796–1999	N Ireland	Significant variations in storminess. High numbers of storms 1796–1825, 1835–1844. Decline in storms 1983–2001	Hickey (2003)
~1800–2000	Scandinavia	No robust signs of any long-term trend in storminess indices	Bärring and von Storch (2004)
1823–2005	Iceland	No increased overall storminess in the study period	Jónsson and Hanna (2007)
1875–1995	NE Atlantic	No common trend with some stations showing decreasing or increasing storminess for part of the length of record, but with some evidence of increasing trends over the last two to three decades of record in the most northeasterly part of the study area	Schmith et al. (1998)
1875–1992	SW approaches of eastern Atlantic	Decadal scale changes in storm surge data with maximum storminess detected for early 1880s and mid-1900s	Orford (2001)
1876–1990	German Bight	No significant trends in geostrophic winds	Schmidt and von Storch (1993)
1876–1996	N Scotland, NW Ireland	Strong interannual variability in gale days. Station data show both increases (Lerwick) and decreases (Stornoway) over study period	Dawson et al. (2002)
1881–1960	NW Europe	Decrease in storm frequency between 1881 and 1960	Alexandersson et al. (1998)
1881–1998	NW Europe	Decreasing trend in storminess but with an increase again in the 1980s and peaking in the early 1990s followed by a reduction in storminess in the late 1990s	Alexandersson et al. (2000)
1881–2003	N Atlantic	A significant decrease in cyclone density over the North Atlantic over the study period	Bhend (2005)
1903–1987	N Atlantic and northwest Europe	Variability of cyclonic activity on decadal timescales but no significant overall trends	Kaas et al. (1996)
1949–2004	N Atlantic (Iceland and UK)	Decrease in severity of severe storms in Iceland after 1980; increase in severity of storms over UK since 1950s	Alexander et al. (2005)
1950–1992	Eastern N Atlantic	No trends in cyclone frequency over the study period	Betts et al. (2004)
1955–1995	Iceland	Intensification of cyclone activity over the study period	Bartholy et al. (2006)
1957–2002	North Atlantic	A decrease in the frequency but increase in the intensity of mid-latitude (30–60°N) cyclones and an increase in both the frequency and intensity of high-latitude (60–90°N) cyclones	McCabe et al. (2001)
1962–2002	Netherlands	Decrease in storminess based on the analysis of near-surface wind data	Smits et al. (2005)
1965–1995	N Atlantic	Decline in the number of storms but strong interannual variability	Lozano et al. (2004)

Other independent evidence of storminess comes from storm surges recorded in the Netherlands (Jelgersma et al., 1995) in AD 1421 (Elizabeth Flood), AD 1530 (Zealand flood), AD 1570 (All Saints flood), AD 1682, AD 1775, AD 1825 and AD 1953. Increased storm surge frequencies are also recorded in Denmark (Aagaard et al., 2007) for the periods AD 1600–1630, 1690–1720 and 1820–1850.

2.3. Proxy records of storminess

Beyond the instrumental and documentary records, the changing concentration of Na^+ ions in Greenland ice cores has been used as a proxy for storminess (Meeker and Mayewski, 2002; Dawson et al., 2003; Fischer and Miedling, 2005) for the period covering the last 1000–1500 years. In contrast to the synoptic (<10 day) resolution of the instrumental records, the ice core data have a seasonal resolution (Meeker and Mayewski, 2002) and appear to show a dominance of high Na^+ values for the period AD 1400 to 2000 compared with lower values for the previous 500–1000 years (Meeker and Mayewski, 2002; Dawson et al., 2003).

3. Timings of periods of sand movement and deposition along western European coasts

Instrumental records of storminess demonstrate interannual and decadal scale variability, while archival records identify impacts of single or multiple storm events on agriculture, infrastructure and livelihoods. The evidence for the impact of changing patterns of storminess on coastal sediments is examined below.

3.1. Historical records of sand movement

Historical accounts from western European show that coastal communities have been particularly vulnerable to sand invasion and the burial of settlements and agricultural land (Lamb and Frydendahl, 1991). Records of sand movement in Denmark date back to AD 1427 (Lamb and Frydendahl, 1991). Sand drift in the Skagen area of northern Jutland was particularly severe between AD 1700 and 1800, forcing the abandonment of churches and settlements (Clemmensen and Murray, 2005). The Skagen dunes were still active in the 1830s and 1840s (Clemmensen

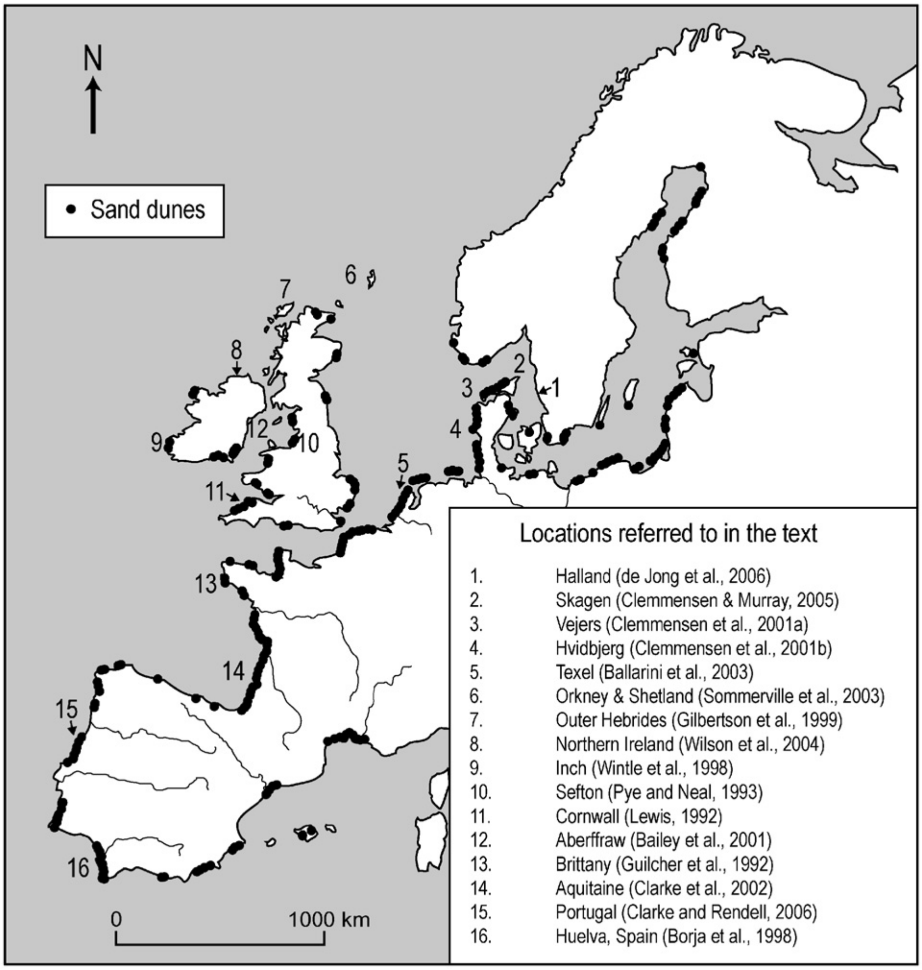

Fig. 1. Distribution of coastal dunefields in western Europe (after Klijn, 1990) with location of particular locations and studies referred to in the text.

and Murray, 2005). Further south along the Jutland coast the dunefields at Hvidbjerg and Vejers (Fig. 1) were active from AD 1550 until 1883, and from AD 1660 until 1900, respectively (Clemmensen et al., 1996, 2001a, b; Clemmensen and Murray, 2005).

In Britain, documentary records exist of sand invasions at Aberffraw, Anglesey, in AD 1331 (Owen, 1953), in South Wales in AD 1316 (Higgins, 1933; Lees, 1982), and North Uist, Outer Hebrides in AD 1697 (Lamb and Frydendahl, 1991). There are also records of repeated sand invasion in Orkney and Shetland in AD 1492 and AD 1500 (Sommerville et al., 2003) and along the north Cornwall coast between AD 1100 and AD 1200 (Polsue, 1868) and during the period AD 1600–1900 (Lamb and Frydendahl, 1991; Lewis, 1992). The low dunes at Sefton in northwest England were apparently stable until the period AD 1200–1400 when a major phase of coastal erosion and sand drift caused substantial dune mobility leading to the abandonment of several settlements and farmsteads which were either lost to the sea or overwhelmed by blown sand (Pye and Neal, 1993). Blowing sand was also a nuisance in this area in the period AD 1700–1850 (Atkins, 2004). Documentary records of sand drift in Ireland, in the period AD 1600–1800, note the impact on settlements and agricultural land in Donegal, Sligo and Mayo (Wilson, 1990). In Brittany, northern France, the scattered coastal dunefields have migrated inland several times, particularly during the Middle Ages and the period AD 1600–1800, often requiring stabilisation works to protect villages and farmlands (Meur et al., 1992) while in southwest France, repeated sand invasion forced the relocation or abandonment of churches and villages from AD 1480 until 1750 (Clarke et al., 2002). In Portugal sand invasion of medieval cemeteries is recorded at Fão and Chafé between AD 1300 and 1700 (Granja and Soares de Carvalho, 1992) and along the Atlantic coast of southern Spain, watch towers built between AD 1600 and 1700 were buried by sand movement (Borja et al., 1999).

Historic accounts reveal numerous instances of sand movement over the past 1000 years. In some cases these involve specific events, in others periods of years or decades during which sand movement was documented. In Cornwall, sand invasion caused damage and provoked countermeasures, as illustrated by the following quotation:

> The Barton of Upton, one of the principal farms in the Parish, was overwhelmed and now lies buried in the sands. This was so suddenly done during one tempestuous night that members of the family were obliged to

make their escape through the chamber windows; shortly after the house disappeared .... The (Gwithian) Churchtown would have shared the same fate had it not been for the parish officers who promptly resorted to an expedient, which simple as it may seem, has everywhere proved to be the most efficacious in arresting the progress of this gigantic evil, that of planting rushes; these completely stay the sand, and greatly facilitate the growth of other vegetation on the surface, so as to create a thin turf (Polsue, 1868, p.161)

Since the 14th century many European coastal dunes have been artificially stabilised by the planting of marram grass (*Ammophila arenaria*, also known as *Arundo arenaria, Calamagrostis arenaria*, *Psamma arenaria* and by the common names: sea bent, sea rush, starr grass and beach grass) or by afforestation (Carter et al., 1990; Barrère, 1992). Other strategies for stabilising bare areas have included using cattle dung, the dumping of rubble (Angus and Elliott, 1992), heather, hay and straw laid out in layers and weighed down by sand (Feilberg and Jensen, 1992). About half the coastal dunes in west Denmark were forested between 1850 and 1930 (Feilberg and Jensen, 1992). Afforestation of coastal dunes in both Portugal and France was enforced by law (Clarke et al., 2002) and in Britain Acts of Parliament were passed in Scotland, in 1695 (*Act for Preservation of Meadowes, Lands and Pasturages Lying Adjacent to Sand Hills*) and England, in 1741 (*The Starr and Bent Act*), to make the cutting of marram grass illegal (Polsue, 1868). In this context of successful stabilisation by vegetation, it is therefore interesting to note that modern dune management strategies are increasingly focusing on 'naturalness' (rather than re-vegetation). Where coastal sand dunes have been artificially disturbed, as part of a 'landscape rejuvenation' project (Arens and Geelen, 2006), vegetated surfaces were buried in winter by sand sheets and dune forms (shadow and coppice dunes). The dune system in this example was supply limited and thus some restabilisation occurred where burial was curtailed by invasion of pioneer species and re-growth of roots (Arens and Geelen, 2006). However many coastal dunes are not supply limited and, if storminess has remained more or less unchanged over the last 150 years (as indicated in Table 1) then, without these stabilisation measures, the dunefields of western Europe would still be actively developing.

3.2. Sand movement during the last 1000 years

Independent chronologies for the historic events such as those described above can be provided by luminescence dating of the sand deposition event or, in some cases, by radiocarbon (^{14}C) dating of enclosing organic-rich deposits. Luminescence dates the last exposure to light of sediment grains prior to burial, whereas ^{14}C dating of organic carbon provides an age for the death of the particular organism. Using these absolute dating techniques, multiple phases of dune building have been identified along the western coast of Denmark (Christiansen et al., 1990) and the Netherlands (Jelgersma et al., 1995). The most recent period of dune building in northern Jutland was dated at 0.3–0.1 ka (AD 1700–1900), in agreement with the available historic records (Clemmensen et al., 2001b; Clemmensen and Murray, 2005) while the onset of Medieval sand drift is recorded around 0.9–0.6ka (AD 1100–1400) (Strickertsson and Murray, 1999; Dalsgaard and Odgaard, 2001; Murray and Clemmensen, 2001). Three periods of dune activity have been recognised in the Netherlands: 1.1–0.8 ka (AD 900–1200); 0.5–0.2 ka (AD 1450–1750), when the production of parabolic dunes occurred (Jelgersma et al., 1995), and 0.2–0.1 ka (AD 1800–1850) prior to the formation of the modern littoral foredunes (Jungerius et al., 1991). Dunes formed by sand blown onshore from shoals migrating on to the currently prograding southwest coast of Texel, a barrier island off the north-west coast of the Netherlands, provide ridge ages, based on luminescence, of 0.3 ka (AD 1700) and 0.1 ka (AD 1900) (Ballarini et al., 2003) which show good agreement with historic records.

In Britain, sand invasion at Aberffraw on the south coast of Anglesey, north Wales, occurred around 0.7, 0.5 and 0.3 (AD 1300, 1500 and 1700, respectively), on the basis of luminescence dating, with the earliest dated incursion at 0.7 ka in agreement with documentary evidence of sand inundation of agricultural fields during a major storm in AD 1331 (Bailey et al., 2001). Sand drift has also been identified in the Outer Hebrides of Scotland, around 0.6–0.2 ka (AD 1400–1800) using luminescence dating of both quartz and calcareous machair sands (Gilbertson et al., 1999). A second study in the Outer Hebrides, using ^{14}C dating of enclosing peats (Dawson et al., 2004), demonstrated a concentration of aeolian activity between 0.5 and 0.3 ka BP (AD 1500–1700). Multiple luminescence dates on blown sand in Orkney and Shetland date sand movement to 0.6–0.1 ka (AD 1400–1900) (Sommerville et al., 2003), but with a cluster of dates around 0.5 ka (AD 1500). The most recent phases of dune building in Ireland date to 0.5–0.1 ka (AD 1500–1900) (Wilson et al., 2004) and 0.6–0.1 ka (AD 1400–1900) (Wintle et al., 1998).

Along the Aquitaine coast of France, to the south of the Gironde estuary coastal dunes have been dated at 0.5–0.3 ka (AD 1500–1800), by ^{14}C dating of organic material from palaeosols and luminescence dating of dune sands (Clarke et al., 1999, 2002) while in western Portugal the most recent phase of dune development dates to 0.2–0.1 ka (AD 1800–1900), based on luminescence dating of sands from transverse and parabolic dune ridges (Clarke and Rendell, 2006). Both studies demonstrate good agreement between luminescence ages and documentary records of both sand drift and the timing of large scale afforestation programmes to stabilise the dune fields (Clarke et al., 2002; Clarke and Rendell, 2006). In southern Spain the dune building is associated with the Medieval

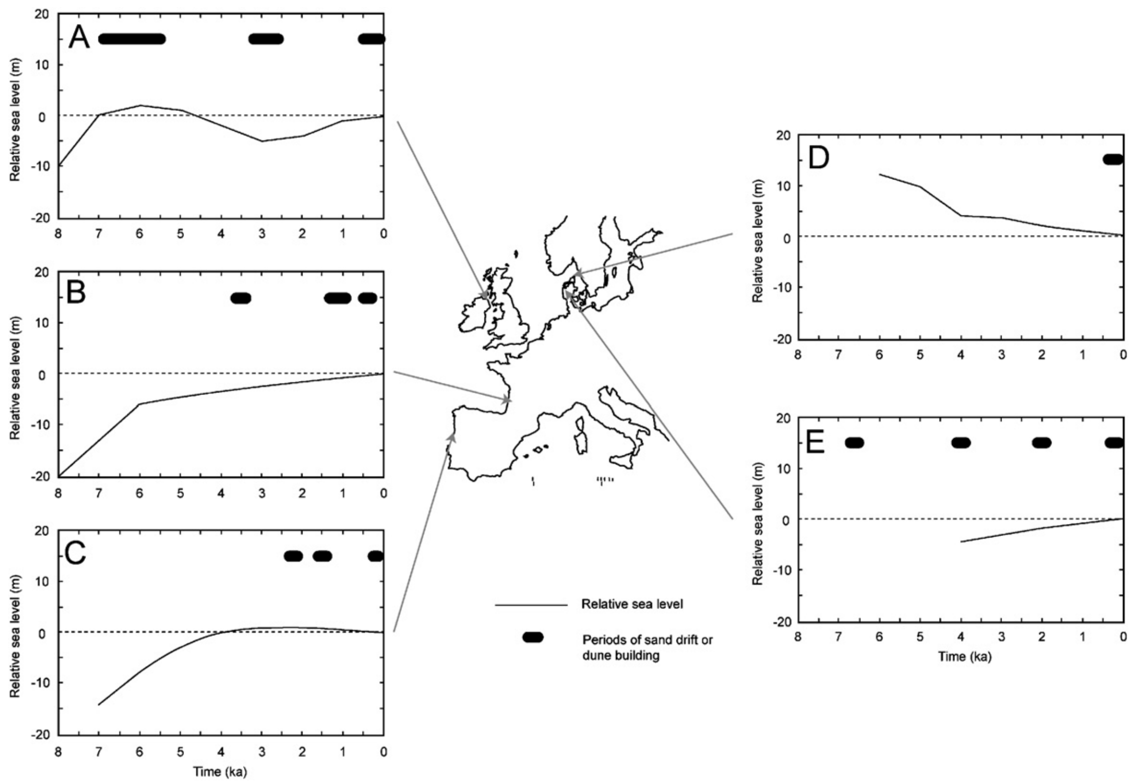

Fig. 2. Relative sea level (RSL) curves over the last 8 ka, with timings of sand drift marked with heavy line, for (A) Northern Ireland (RSL data from Carter and Wilson, 1993; dune data from Wilson et al., 2004); (B) Aquitaine southwest France (RSL curve from Lambeck, 1997; dune data from Clarke et al., 1999, 2002; (C) Portugal (RSL date from Granja and De Groot, 1996; dune data from Clarke and Rendell, 2006); (D) Skagen, Denmark (RSL data from Clemmensen et al., 2001c; dune data from Clemmensen and Murray, 2005); (E) Jutland, Denmark (RSL data from Gehrels et al., 2006; dune data from Clemmensen et al., 2001a, b).

period, and dated by archaeological artefacts 0.7–0.5 ka (AD 1300–1500) (Borja et al., 1999; Zazo et al., 2005).

Additional evidence of aeolian sand mobilisation is provided by the sand grain content of ombrotrophic peat bogs near Halland in western Sweden (Björck and Clemmensen, 2004; de Jong et al., 2006; de Jong, 2007). Variations in aeolian sand influx into peat bogs indicated mobilisation during the period 0.4–0.1 ka (AD 1600–1900) and the record of sand movement is strongly correlated with that for the Vejers dunefield (Clemmensen et al., 1996) in western Jutland, Denmark (de Jong et al., 2006), suggesting a regional response to storm impacts.

Thus, across western Europe, during the last 1000 years, independent dating evidence points to spatially variable, episodic sand mobilisation. The most notable aeolian sand drift activity was concentrated in the historic period 0.5–0.1 ka (AD 1500–1900) which spans the LIA.

3.3. Sand movement prior to 1000 years ago

Independent dating evidence for earlier periods of sand movement is patchier, both spatially and temporally. In Denmark luminescence dates of 2.3, 1.5 and 6.6 ka, and 4.0 and 2.0 ka were obtained for the older units in the Vejers and Hvidbjerg dunefields, respectively (Clemmensen et al., 2001a), while the oldest episode of dune accretion occurred at 8.3 ka (Dalsgaard and Odgaard, 2001). These dune-building episodes have taken place against a backdrop of both rising and falling relative sea level (RSL) during the Holocene (Fig. 2). In the most northerly part of Jutland, at Skagen, sea level has fallen in excess of 11 m since the start of the Holocene (Clemmensen et al., 2001c) whereas at Skallingen on the west coast of central Jutland, sea level has risen 4 m in 4000 years in response to glacio-isostatic subsidence (Gehrels et al., 2006). In addition to the dune record in Denmark, the ombrotrophic peat bog records from western Sweden indicate that periods of sand influx occurred 4.8, 4.2, 2.2–2.8, 1.5 and 1.1 ka (de Jong et al., 2006).

In England, dunes formed in the mid Holocene around 5.8–5.7 ka on an emergent sand bank at Sefton, NW England (Pye and Neal, 1993) and around 5.0 ka in an embayment at Gwithian in Cornwall, SW England (Lewis, 1992). Later, between 3.3 and 2.3 ka, low dunes and sand sheets developed at Sefton (Pye and Neal, 1993). In the Outer Hebrides of Scotland periods of sand movement have been dated at 3.7–3.3 ka and 1.7–1.3 ka (Gilbertson et al., 1999), while in Orkney and Shetland sand movement occurred 1.2–1.1 ka (Sommerville et al., 2003). Multiple periods of Holocene dune emplacement been dated in Northern Ireland at 5.5–6.9, 4.5, 3.9, 3.4–2.4 and 1.1 ka (Wilson et al., 2004) during both regressive and transgressive RSL regimes (Carter and Wilson, 1993).

The start of active development of the Brittany dunefields in northern France has been dated to the Iron Age (ca. 2.4 ka) (Guilcher et al., 1992), although the isolated cliff top dune fields of western Brittany began developing 4.4–4.2 cal kyr BP, on the basis of ^{14}C dating of the underlying palaeosol (Haslett et al., 2000). These perched dunes were isolated from their local beach sediment supply, by erosion of the underlying Pleistocene deposits to form a low cliff line, as sea levels have continued to rise (Lambeck, 1997). In southwest France, a rising sea level trend impacting upon a sandy barrier coastline has resulted in rollover dune formation, as the Dune de la Grave has evolved in the Dune de Pyla (or Pilat), the largest dune in western Europe (Bressolier et al., 1990). Sea level data show the rate of sea level rise slowing after 6.0 ka, when sea level was at −5.0 to −4.0 m (Lambeck, 1997), and was probably within 1 m of present levels by 2.2 ka (Mellalieu et al., 2000). Coastal dunes line the 210 km stretch of the Aquitaine coast to the south of the Gironde estuary, reaching up to 10 km inland (Tastet and Pontee, 1998). Dune building occurred 1.3–0.9 ka and (Clarke et al., 2002) with intervening stabilisation and peat formation and age of 3.6–3.4 ka have been obtained for sands intercalated between coastal peats. In Portugal, several periods of dune-building activity have been identified along the Atlantic coast at 9.7, 8.2, 2.2 and 1.5 ka (Clarke and Rendell, 2006). This region shows sea level trends at or close to present levels over the last 5 ka (Granja and De Groot, 1996).

4. Discussion

Historical records and independent dating results both confirm that widespread activity of coastal dunefields across western Europe took place during the LIA (ca. AD 1570–1900) (Matthews and Briffa, 2005). Within this period, low solar activity, during the Maunder (AD1645–1715) and Dalton (AD 1790–1830) Minima, has been related to changes in Atlantic storm tracks (van der Schrier and Barkmeijer, 2005), anomalously cold winter and summer temperatures in Scandinavia (Bjerknes, 1965), and the repositioning of the polar front and changing sea ice cover (Ogilvie and Jónsson, 2001). The southward extent of sea ice (Vinje, 2001), particularly in the Northern Hemisphere winter, would have resulted in an increased thermal gradient across the North Atlantic and both a southward displacement, and an increase in the frequency and intensity, of storm tracks (Raible et al., 2007). Although few of the instrumental records of storminess extend back far into the LIA (see Table 1), independent corroboration of increased storminess during the LIA is provided by the record of Na^+ ion concentration in the Greenland GISP2 ice core (Meeker and Mayewski, 2002; Dawson et al., 2003).

There appears to be broad agreement between historic records of sand movement and luminescence ages of sand deposition in Denmark, Ireland, Britain, France and Portugal, and in the radiocarbon-dated blown sand record in the coastal ombrotrophic peat bogs of western Sweden. Mobilisation of sand appears to be associated with periods of persistent storminess or, in some cases, individual Atlantic storm events (Lamb and Frydendahl, 1991). However, while storms are transient features, lasting hours or days, identifiable within the instrumental records, the statistical uncertainties associated with both luminescence and ^{14}C dates only enable sand movement to be related to particular decades or centuries (Sommerville et al., 2003). The errors on luminescence dates, typically of the order of 8–10%, mean that identification with individual storms or series of storms is a matter of probability rather than certainty (e.g. Wintle et al., 1998). If persistent sand movement reflects sequential storm impacts then, given the size of large low pressure systems relative to the size of western Europe, we would not expect a synchronous European-wide response to spatially discrete storm tracks (Lozano et al., 2004). There is some evidence that persistence in the spatial location of North Atlantic storm tracks, as controlled by the NAO (Lozano et al., 2004; Meyers and Pagani, 2006), may result in particular periods of sand mobilisation at a regional level (Clarke and Rendell, 2006).

Prior to the LIA, and therefore beyond the range of instrumental and most documentary records, knowledge of the timing of sand movement relies almost exclusively on the results of luminescence dating. Pre-LIA Holocene cooling events (9.5, 8.2, 5.9, 4.3, 2.8 and 1.4 ka) hypothesised by Bond et al. (1997) are associated with southward extension of sea ice in the North Atlantic (Bond et al., 2001). The 8.2 ka event (Alley et al., 1997) which represents a significant period of abrupt cooling within the early Holocene (Rohling and Pälike, 2005) is associated with sand accretion in Denmark (Dalsgaard and Odgaard, 2001), Scotland (Wilson, 2002) and Portugal (Clarke and Rendell, 2006), thus providing evidence linking storminess to cooling events in the North Atlantic. As yet there is no strong evidence that significant dune building occurred at 5.9, 4.3 and 2.8 ka across the coasts of western Europe, although further research may clarify this.

Although strong winds are a necessary condition for sand mobilisation, an abundant supply of sand is also required for dune building (Carter, 1991). While changing sea levels have an influence on sand supply (Carter, 1991), single events such as storm surges (Regnauld et al., 2004; Aagaard et al., 2007) and tsunamis (e.g. Hindson and Andrade, 1999) can provide sediment through material driven onshore by wave action. By contrast, anthropogenically driven changes in river catchments, particularly dam construction, have the effect of reducing sediment supply via estuaries (Carter, 1991). Beyond 5 ka, dune development at coastal sites in western Europe is likely to have been strongly influenced by regional patterns of RSL change which in turn control sediment supply to beaches for onward aeolian transport (Pye, 1984; Short, 1988; Carter, 1990; Tooley, 1990). While RSL curves show considerable spatial variation across western

Europe, dune-building episodes, particularly during the LIA, appear to have occurred irrespective of the speed and direction of RSL change (Fig. 2). While RSL may be changing on longer timescales of centuries or millennia, at shorter timescales sand movement will simply be a function of the interaction of sand supply, wind velocities above the threshold for sand movement and vegetation cover, all of which are affected by storminess.

5. Conclusions

The instrumental record of Atlantic storminess is limited to the last 200 years and demonstrates considerable temporal and spatial variability. The consensus that emerges is one of comparatively unchanged storminess with, for example, the final two decades of the 20th century having storminess comparable to that in the 1900s. The analysis of documentary records, discontinuous instrumental data and proxy records indicate that the period of the LIA (AD 1570–1900) included periods of enhanced storminess relative to present. This increased storminess coincided with numerous episodes of sand drift and dune building along the western European coast, as demonstrated by both documentary records and independent dating of sand movement. Sand drift and dune building reflect the persistence of strong winds associated with storms and such episodes can be used as proxies for storminess at a decadal to centennial resolution. The Holocene record of sand drift in western Europe includes episodes of movement corresponding to periods of Northern Hemisphere cooling (Bond et al., 1997), particularly 8.2 ka, and provides the additional evidence that these periods, like the LIA, were also stormy. Additional dating is needed to establish a more complete record of Holocene aeolian activity and to resolve further this promising proxy of past storminess. Given that Atlantic storminess has remained more or less unchanged over that last 150 years, modern dune management strategies which consider dune deforestation, driven by an increasing focus on 'naturalness', may give rise to a recurrence of sand drift problems for coastal settlements, agriculture and infrastructure. Predictions of increased storm frequency and sea level rise by the end of the current century (Lozano et al., 2004), related to anthropogenically driven climate change (Carnell et al., 1996), mean that in locations where coastal sand supply is abundant, dune management may instead need to focus on further stabilisation measures.

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