Centennial climate variability in the British Isles during the mid-late Holocene

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Abstract

Multi-millennial climate changes were relatively minor over the mid-late Holocene in the British Isles, because orbitally forced insolation changes were smaller than those at higher latitudes. Centennial climate variability is thus likely to have exerted a greater influence on the environment and human society of the region. Proxy climate records from the British Isles covering the last 4500 years are assembled and re-evaluated with the aim of identifying centennial climate variability reflected by multi-proxy indicators. The proxies include bog oak populations, peatland surface wetness, flooding episodes from fluvial deposits, speleothem band width and oxygen isotopes, chironomids from lake sediments and sand and dune deposition. Most proxies reflect water balance rather than temperature alone, and records predominantly reflect warm season climate. A series of twelve key periods of enhanced precipitation-evaporation (P-E) are identified by their presence in two or more proxy records. Variability in P-E is much greater than that shown by temperature proxies and there is no necessary association between warm/cool and dry/wet periods. Although the data for temperature are less robust than those for P-E, a series of key temperature changes are proposed based on speleothem $\delta^{18}O$ and chironomid inferred July temperature records; relatively cool before c. 3100 yr BP, warmer (3100-2000 yr BP), cool (2000-1250 cal yr BP), warm (1250-650 cal yr BP), and cool (650 cal yr BP onwards). Some key increases in P-E (2750, 1650, 550 cal yr BP) show a strong correspondence with 'Bond cycles' in ocean proxy records for increased ice-rafted debris, decreased summer sea surface temperatures and sometimes decreased North Atlantic Deep Water circulation. Other higher frequency changes in P-E are also strongly related to SST variability. Whilst some of the main changes to cooler SSTs and increased P-E are approximately coincident with reduced solar output, most are not and thus must be the result of the internal dynamics of the ocean and atmosphere. Future work should concentrate on firmly establishing the pattern of temperature change, improving

chronological accuracy and precision in existing records and improving process-based understanding of proxies.

1. Introduction

The late Quaternary was a period of considerable climate variability. The major long term fluctuations between temperature minima during glacials and temperature maxima during interglacial periods were punctuated by a large number of much more abrupt swings in climate occurring over shorter timescales. The last glacial period holds abundant evidence of many such dramatic changes expressed as Dansgaard-Oeschger cycles and Heinrich events (Lowe and Walker 1997). Particular attention has focused on the last glacial-interglacial transition (LGIT) when the temporal precision, spatial coverage and multi-dimensional detail of climate reconstructions is high, and there is the greatest dynamism in the global climate system as a result of deglaciation. The North Atlantic region, including the British Isles, was close to the locus of both the mechanisms and evidence of these very large changes (Lowe et al. 2008; Walker et al. 1999; Walker et al. 2003; Walker 1995). In comparison to the large scale climate variability occurring during the LGIT, the last c.11,500 calendar years looks relatively benign and stable, perhaps best exemplified by the relatively flat oxygen isotope curve from the Greenland ice cores (Figure 1a). Why then is the Holocene of interest as a period of study? There are four principal reasons for studying Holocene climate variability.

First, not all records of Holocene climate change show the relative stability suggested by the ice cores (Mayewski *et al.* 2004). Tropical areas especially are known to have undergone large magnitude changes, some of which alter the regional environment dramatically over a relatively short period of time (Claussen 2008). Temperate and high latitude regions have also undergone significant changes, especially in terms of hydrological change (Verschuren and Charman 2008). A greater range of climate variability is becoming increasingly apparent as a growing number of climate variables are reconstructed across a larger number of locations, and the signal to noise ratio is increased as a result of improved methodologies. Second, the boundary conditions during the Holocene, and particularly the late Holocene, are much more similar to those of today and the near-future than those pertaining during the last glacial. Orbital configurations converge towards those of the present today throughout the Holocene and ice sheet dynamics play a less important role in the climate system. Third, the level of detail available for the Holocene is very much larger than for former times. The resolution, spatial distribution and range of ocean and terrestrial climate variables that can be reconstructed is much greater than for the late Pleistocene, and many records can also be integrated with and calibrated by instrumental climate data covering recent centuries. Finally, and perhaps most importantly, this is an era when societal development became dependent on climate to a far greater extent than had previously been the case, especially following the development of sedentary agriculture and complex civilisations with high population density. Climate changes had a dramatic effect on past societies and will continue to do so in the future (Bradley, 2008). The Holocene is therefore an interesting, data rich period of time where improving understanding of the nature and causes of climate variability is perhaps most relevant to present and future societal well-being. However, Holocene palaeoclimatology is not without its challenges, especially concerning understanding of centennial and decadal climate variability (see section 3 below).

The primary aim of this paper is to review key evidence for Holocene climate variability of the British Isles. The emphasis is on quantitative and semi-quantitative continuous records of change, where the rate and magnitude of change can be assessed over multi-centennial time periods. In doing so, the intent is to look back on progress made since the last major review of the Quaternary of this region (Shotton 1977), focus especially on centennial variability during the mid-late Holocene as the period that much of the richest new evidence covers, and to reflect on the context of the British Isles in relation to changes in the North Atlantic over the same period. A particularly thorny challenge in reviewing the evidence is reconciling the different proxy records now available within this relatively small region. Finally, there is an attempt to evaluate the possible role of different potential forcings, both external (orbital, solar and volcanic) and internal (ocean and atmospheric), in determining Holocene climate variability of the British Isles.

2. Millennial temperature change

The main outlines of Holocene climate change in the British Isles had already been suggested well before the late 1970s. Following the classic Scandinavian sub-division of the Holocene formalised by Mangerud et al. (1974), Godwin's then well-established pollen zones provided the broad framework for understanding of climate, vegetation and environmental change in the British Isles. Godwin (1975; 1977) still essentially focused on the concept of fixed pollen zones as the basis for changes through time, applicable over a wide area of the British Isles. However, at the same time, others were starting to develop an appreciation of the time-transgressive nature of vegetation change even in a small region such as mainland Britain (Birks 1977), which would ultimately lead to pollen mapping of larger spatial patterns on a continental scale, demonstrating the migrational response of trees to changing climate (Huntley and Birks 1983), and showing significant spatial variability and complexity even within the British Isles (Birks 1989). While Godwin and many others focused on biotic and landscape changes, Lamb (1965) was already developing much improved climate reconstructions, culminating in the publication of a comprehensive summary of climate change in the British Isles and beyond in 1977, coincidentally the same year as the Shotton volume (Lamb, 1977). Based on the evidence available at the time, the Holocene was divisible into three main periods (Figure 1d); initial postglacial warming followed by a period of greatest warmth between around 8500 to 5500 cal yr BP and then a decline in temperature to the present day. Two features stand out in this reconstruction. First, there was clearly an appreciation of greater centennial variability emerging for periods where there was a larger amount of data available, especially in the last 1000 years. Second, there was rather little difference between the summer and winter reconstructions in terms of the overall temporal patterns, although the magnitude of change was different. Apart from the last 1000 years where documentary data were used extensively by Lamb, inferences were mainly based on pollen and, to a lesser extent, plant macrofossils. Although inevitably limited in accuracy, there have been few attempts to improve upon this reconstruction using the more recent data from the full range of data sources.

Quantitative reconstructions have tended to be based on individual sites and proxies with no serious attempts at large scale synthesis. However, Davis et al. (2003) used pollen data to provide reconstructions for millennial scale change across Europe, with the British Isles falling within their northwest and central western sectors (Figure 1b and 1c). It is not entirely clear whether temperature variability for the British Isles as a whole is most similar to the northwestern or the central western pattern, but it seems reasonable to assume that it lies somewhere between these two reconstructions, as the division between these regions occurs at 55^{o}N (approximately the Solway Firth). The overall changes in summer temperatures are similar to Lamb's earlier reconstruction, with the same broad phases, but there are key differences in the seasonality and timing of change. Of particular significance is the difference between summer and winter changes. Both show increasing trends in the early and mid-Holocene to around 6000 cal yr BP, but summer temperatures decline thereafter while winter temperatures continue to increase or show little change. By 6000 cal yr BP, maximum summer warmth was perhaps up to 1.8^{o}C above present (the 1961-1990 mean) in the north but may well have been less than this in the south, and winter temperatures may still have been up to 1^{o}C cooler than today even during the so-called climatic optimum of the mid-Holocene. The seasonal contrasts are more pronounced in the north western sector (covering Scotland) of Davis et al.'s reconstruction than in the central west sector which includes England, Wales and Ireland. This reflects the strength of the influence of changing summer insolation, which was the main driver of the observed multi-millennial climate variability in the north. In terms of broad scale European climate, the notion of a Holocene climatic optimum thus seems to apply only to northern summer temperatures. Europe as a whole did not experience a mid-Holocene climatic optimum, as warmth in the north was balanced by cooler conditions in the south (Davis et al., 2003). For the British Isles, it seems likely that multi-millennial temperature change was rather moderate after about 5000 cal yr BP. While these data suggest some shorter term variability, the noise created by chronological imprecision and uncertainty of correlation, migrational lags and human impact precludes any assessment of the magnitude and timing of this. For sub-millennial change, data from other sources must be examined.

3. Centennial climate change in the British Isles

The detection and quantification of centennial climate variability is difficult in the British Isles for a number of reasons. It seems likely that temperature changes in the mid to late-Holocene may not have been dramatic over long timescales because insolation forced change was weaker than in the high latitudes. Solar and volcanic forcing, together with ocean and atmospheric processes and non-linear responses to external and internal forcing were the principal determinants of decadal to centennial climate change. If the magnitude of change was relatively small, then the sensitivity of proxies used to reconstruct change has to be much greater and the uncertainty in the reconstruction technique needs to be minimal. The key problem and challenge for Holcene palaeoclimate reconstruction in the British isles is thus one of signal:noise ratio, combined with the question of chronological precision and accuracy.

The problem of low signal:noise ratio can be tackled in several ways; looking for a greater range of climate variables, considering seasonality and focusing on proxies that are less likely to be subject to influences of migrational lag and human impact than pollen. Despite the large number of British and Irish Quaternary scientists with an interest in Holocene climate reconstruction, relatively few of them choose to work on records from the British Isles. This is explicable by a) the range of deposits undisturbed by human activity is limited, and b) a number of techniques that are well-established for other areas of the world are difficult or impossible to apply to the British Isles. For example, tree ring records that have been so successful in reconstructing summer temperatures at high latitudes (Grudd *et al.* 2002) are less sensitive to temperature in temperate regions, and show a more complex relationship with a broader range of climate variables (Kelly *et al.* 2002).

Sources of data currently available are principally based on peat deposits, bog oaks, speleothems, flood deposits, sand and dune deposition, and more recently chironomids (Table 1, Figure 2). These records are selected for this review on the basis of a number of different criteria; the range of climate variables represented, continuity and coverage of the mid-late Holocene, and chronological information. For the purposes of this review, I focus on centennial changes and make the assumption that temporal variability at this scale is coherent at the scale of the British Isles. Although some studies have suggested regional variability of proxy records at smaller spatial scales is detectable (Macklin et al., 2005; Langdon and Barber, 2004; Charman et al., 2006), there are insufficient records to attempt analysis of all proxies for smaller regions. Furthermore, it could be argued that many of the records are not adequately replicated within regions to allow secure conclusions on regional differences in temporal trends. A second assumption is that the chronologies in the published studies are adequate for correlation between records. This is more difficult to establish, but if the focus is on centennial variations, this is probably not unreasonable, although individual records must be critically assessed for hypothesised correlations. Finally, I have assumed that there is a reasonable conceptual basis for the interpretation of the proxy. This is true in an *a priori* sense, but it will become clear that the detail of interpretation requires revision when different proxies are considered. Indeed, this is one of the benefits of assembling multi-proxy data sets.

4. Proxies, climate and processes

A key issue in interpretation of proxy records is an understanding of their response to climate changes. At the coarsest scale, the proxies are responsive to either temperature or precipitation, but most of the available proxy records for the British Isles are primarily a response to some combination of the two, often via changes in precipitation-evapotranspiration (P-E) balance. The seasonality of the proxy-climate relationship may also be important. However, although there is often a reasonable basis for suggesting that records represent changes between wet/cool and warm/dry conditions, the precise relationship of the relationship is poorly known for almost all proxies, most often because the influence of climate is mediated by ecological and physical systems which themselves are imperfectly understood. Comparing and attempting to reconcile two or more proxy records can help constrain the reconstruction of the overall climate system, but it can also force a re-evaluation of the records and yield an improved understanding of the proxy response to climate.

One example of records requiring better process-based understanding is provided by the palaeoclimate record from Irish bog oaks. Conventionally, the record of bog oak numbers has been interpreted as reflecting hydrological changes on the bog surface, associated with changing precipitation-evaporation balance. High numbers of bog oaks are interpreted as reflecting dry growing conditions and low numbers suggest greater waterlogging of soils and hence a reduction in the numbers of trees present (Figure 3a, Turney et al., 2005; Turney et al., 2006). This interpretation is illustrated in Figure 3a and a limited test of the hypothesised link with bog surface wetness is provided by Turney et al. (2005). However, there are now much more extensive records of bog surface wetness reconstructions with which to assess this interpretation of the bog oak data, some of which are shown here (Figure 3c-e). These records represent a total of 17 separate reconstructions of bog surface wetness from northern England, Scotland, Northern Ireland and Ireland and show a coherent pattern of multi-centennial P-E variability across the British isles for the mid to late-Holocene (see section 5 below). However, the periods of increasing P-E do not coincide with the periods of low bog oak numbers, originally interpreted as being the periods of high surface wetness (Figure 3a).

This could be explained by chronological uncertainty in the peat records, as suggested by Turney et al (2005), in comparing a single bog surface wetness record with the bog oak data. However, this can now be discounted because of the much greater number of records shown in Figure 3 and also because the dating precision of some of these records is of the order of ±50 years at 2 sigma (Swindles et al. 2007). A further alternative explanation is that the bog oak record is primarily a function of ecological dynamics, which either mediated or even predominated in the relationship between climate and oak populations. Certainly the ecology of these non-analogue populations is complex and perhaps difficult to explain simplistically (Barratt, 2007). Finally, and perhaps most plausibly, the interpretation of the bog oak record needs to be adjusted. The bog oak curve represents the numbers of oaks growing in any particular cross-dated year in the chronology. When conditions are dry, one might expect to see an expansion in the oak population. Conversely, in a period of increasing wetness, populations would decline and there would be more limited recruitment of new trees. Thus, it is not the peaks or troughs in the bog oak numbers that are important, it is whether the population is increasing (rising curve) or declining (falling curve). Declining populations are associated with increasing mean age as recruitment is slowed or stopped altogether. Reinterpreting the bog oak data in this way (Figure 3a and 3b), there is much greater correspondence between the two proxies. Clearly there are still differences between the records and this may be explicable by local scale variability or uneven sampling effects in the bog oak record, or by chronological imprecision and noise in the bog surface wetness record. However, this provides a good demonstration of the need to carefully evaluate process-based understanding of proxy records and the complementarity of records.

5. Precipitation and P-E related records

The records in Table 1 are divisible into two main groups; those reflecting precipitation and P-E, and those reflecting temperature as the dominant signal. There are surprisingly few records that can convincingly be shown to be primarily driven by temperature, given that it is often the climate variable that is usually referred to in relation to recent, present and future climate change (IPCC 2007). To those proxies primarily reflecting P-E, data on dune building and sand deposition are added as an indicator of storminess (Table 1).

Bog surface wetness records are a long-established source of late Holocene data for the British Isles. Since the work of Barber (1981), there have been numerous records developed from a range of locations, with a particular concentration in northern England (see Barber and Charman, 2003, for a review). Peat is particularly well-suited to centennial scale climate reconstruction as it accumulates at a reasonable rate (typically 5-20 yrs cm^{-1}) and contains a range of potential proxy evidence including plant macrofossils, testate amoebae, peat humification, biomarkers and stable isotopes. The first three proxies are now fairly well established but the last two are still being developed so there are no published reconstructions of long-term Holocene change based on them. All three of these proxies reflect changes in the hydrological status of the peatland surface, often termed ‘bog surface wetness’. Proxy BSW reflects changes in P-E balance over the extended summer or growing season period, because during the rest of the year the bog is saturated, and biological activity and decay is at a minimum. There are now dozens of published reconstructions of bog surface wetness for the British Isles, so there is potentially a very large amount of data available (e.g. see records referred to in Chambers and Blackford, 2001; Barber and Charman, 2003; Charman et al., 2006). To date however, there have been few attempts to provide composite records primarily because of the difficulty of dealing with imprecise chronologies. Typically, data on the main wet shifts are tabulated to assess clustering of dates of main biostratigraphic changes (e.g. Hughes et al., 2000; Barber and Charman, 2003). For the purposes of this paper, the compilation of data from northern Britain (Charman et al., 2006) is used as a representative curve for the British Isles, as it appears to represent changes across the British Isles reasonably well (e.g. Figure 3c to 3e) and periods of major changes are also shown in tabulated data (Hughes, 2000; Barber and Charman, 2003). Three key changes were identified by Charman et al. (2006); commencing at c.3600, 2760 and 1600 cal BP, but all continuing for periods of c. 100-200 years and often followed by wet phases before a return to drier conditions. These key periods are also reflected in other independent records from northern England (Barber et al., 1994; Hughes et al., 2000) and southern Scotland (Chambers *et al.* 1997). Records spanning only the later periods also often identify the middle part of the first millennium AD as being a period of change to wetter conditions from areas including Wales and Ireland (Blackford and Chambers, 1991; Barber et al., 2000; Blundell et al. 2007). In addition to the higher magnitude, most widespread changes to wetter conditions, there are a number of other smaller magnitude changes of particular note in the bog surface wetness records at c. 3060, 2050, 1260, 860, 550 and 270 cal yr. The change at c. 550 cal yr BP is particularly well replicated in many records in northern England and southern Scotland.

As discussed above, the numbers of Irish bog oaks are also thought to represent warm season P-E are (Turney et al., 2005). The declines in bog oak numbers occur over a range of timescales and here only periods of consistent change over >50 years are identified. Periods of decline cover 50-400 years (Figure 4) and occur at (rounded to nearest decade) 4060-3900, 3400-3310, 3120-2870, 2590-2430, 2240-2120, 1670-1270, 780-730, 540-390 cal yr BP. A number of these periods coincide with rising water tables shown by the bog surface wetness record, although there are fewer declines in bog oak numbers. None of the declines in bog oak numbers occur during phases dominated by dry bog surface conditions, supporting the contention that the main driver of bog oak populations is water balance. Particularly strong associations are shown for the periods 3120-2870, 2590-2430, 2240-2120, 1670-1270 and 540-390 cal yr BP. The major long term decline at 1670 cal BP is associated with rising water table and continues through the ensuing wet phase shown by the bog record. Several other periods of declining bog oak numbers are contained within longer periods of water table rise (3400-3310, 780-730 cal yr BP). Furthermore, several periods of rising water table occur during times when bog oak populations were already very low (1250-1100, c. 200 cal yr BP) and others are associated with more minor population reductions (c. 3800 and c.1800 cal yr BP) or slow downs in the rate of bog oak expansion (c. 2760, c. 2000 cal yr BP). The muted change in the bog oak record around 2700-2800 cal yr BP is notable, as this is well-established as one of the key changes in bog surface wetness in Ireland and elsewhere (van Geel et al., 1996; Swindles et al., 2007). It seems likely that there were other factors involved in bog oak population change, such that a significant climate shift was only sufficient to slow population expansion. Alternatively these may be periods when the assumption of even sampling effort over time does not hold true. Although bog oak numbers are increasing during this period, the mean age is also increasing (Figure 3b), suggesting an ageing population with no or slow recruitment. The high numbers of bog oaks retrieved for this period may even be a consequence of improved preservation conditions brought about by more waterlogged conditions.

Assuming the changes registered by bog surface wetness and bog oak populations represent significant variability in water balance, large changes in runoff must also have occurred throughout the late Holocene. Records of fluvial change are more difficult to assemble and interpret than those based on peat and bog oaks, as fluvial activity is represented by a series of events in deposits rather than as a continuous sequence of change. There are further complications with interpretation of fluvial sediments and whether they represent flooding or internal dynamic changes associated with non-climatic factors, as well as the problem of erosion and selective preservation of the record. However, through careful screening and compilation of radiocarbon dates from fluvial deposits, a record of flood frequency has been developed (Macklin and Lewin, 2003; Macklin et al., 2005; Johnstone et al., 2006). The combined probability distribution of all the radiocarbon ages is expressed as a probability difference curve where peaks represent periods with increased flood frequency (Figure 4c). The record extends over the entire Holocene but here we focus here on the mid-late Holocene for comparisons with other records. The data have been divided into upland and lowland catchments (Macklin et al., 2005; Johnstone et al., 2006) but the flood peaks presented here are those from the 'all British rivers' data set of Johnstone et al (2006, their Table 2).

A clear pattern emerges from this comparison. All the peaks in flood frequency occur during periods of rising peatland water tables (Figure 4c). The only exception is the flood peak at 660 cal yr BP, which is just before the start of the rise in water tables. Accepting that the reconstructed peat surface wetness is a function of extended summer water balance (Charman 2007), the correspondence with flood peaks implies that centennial changes in flood frequency as recorded in fluvial sediments may also be dominated by spring-autumn flooding. Moreover, there is a tendency for the flood peaks to occur at the start of the rises in water table rather than during the main or end of the wet phases. This suggests that the fluvial system has a very rapid response time and increased flooding is a near-instantaneous response to changes in prevailing longer-term conditions. However, higher flood frequency does not always persist through phases of high peatland water tables. This may be explained in terms of a change in equilibrium of the fluvial system. An initial rapid catchment response to long term variability is followed by a period of stabilisation and adjustment to a new equilibrium.

One further source of P-E related data from the British Isles is the speleothem record from northern Scotland, which was the first to demonstrate the potential of annually resolved records from this archive (Baker et al., 1993). Several speleothems now form a 3,600 year series based on annual band width variations (Proctor et al., 2000; 2002). The annual band width shows strong correlations with instrumental climate data over the last 150 years, positively correlated with annual temperature and negatively correlated with annual precipitation. Because variation in temperature over the calibration period is small, precipitation is the main control on band width (Proctor et al. 2000). The underlying process thought to drive speleothem growth rate in this region is related to CO_2 production in the soil above the cave. Extended and enhanced dry, warm periods produce more soil CO_2 leading to increased speleothem growth rates. During wet periods, speleothem growth is reduced and may reach a minimum level during sustained wet phases such that the linear relationship between precipitation and band width breaks down (Proctor et al. 2002).

Following this basic reasoning one might expect to see correspondence between the other P-E records (Figs a to c) and the speleothem band width record (Fig 4d). While this is the case in a number of instances, there is no simple correlation between the common periods of increasing wetness in oak, bog and fluvial records and speleothem band width. Because of the potential insensitivity to very wet periods, most emphasis should be given to the periods of high or increasing speleothem band width. Centennial averages of speleothem band width are rising and greatest around 3600-3500, 3250-3050, 2550-2000, 1370-1220, and after 400 cal yr BP. The phases 2550-2000 and after 400 cal yr BP are characterised by a series of strong peaks but also contain some shorter periods of lower values.

A number of the main periods of increasing dryness shown by the speleothems are in antiphase with the wetness phases from the other proxies. The period between 2300 and 600 cal yr BP shows a series of dry (speleothem) and wet (other proxies) phases in succession. In particular, the lowest speleothem growth rates occur during the marked wet phase 1600-1400 cal yr BP. Some of the earlier wet phases also occur during periods of low or declining speleothem growth – after 3500 cal yr BP and following 3100 cal yr BP. Notable periods where mismatches occur are at 2800 cal yr BP, where speleothem band thickness shows little change or a slightly rising trend, 2550 cal yr BP where speleothem band width is low to start with but then shows a strong change to drier conditions during the wet phase inferred from the other proxies. There are also differences in the detail of changes after 600 cal yr BP; the switch to wetter conditions at c.550 cal yr BP shown by other proxies is not evident in the speleothem, and there is a long term increase in band width (dry conditions) during the broadly wet period shown by the peatland record. The imperfect correspondence between the speleothem and other proxies may have a number of explanations. First, the speleothem record is from a single location, perhaps not representative of trends across the British Isles more generally. However, the broader scale correlations between the speleothem band width and the winter NAO and North Atlantic SSTs suggest that the record does reflect wider atmospheric circulation patterns that also influence most of western Europe (Proctor et al., 2002). Second, the other P-E indicators may incorporate chronological errors sufficient to make correlations uncertain for centennial timescales. This also seems unlikely given the agreement between the other proxies and the fact that a number of the suggested anti-phase correlations are longer term trends (e.g. the last 600 years). Instead, it seems more likely that the speleothem signal is dominated by slightly different seasonality than the other records, with probably a greater influence of autumn and winter precipitation rather than summer P-E balance alone. Two factors support this hypothesis. First, the wettest part of the year is the autumn and early winter in northwest Scotland, with the largest rainfall amounts arising from strong westerly and southwesterly airflow (Roy 1997), making it more likely that this period of the year dominates the annual precipitation signal. Second, strong correlations with the winter NAO suggest a stronger link with winter precipitation rather than summer conditions (Trouet et al., 2008).

One further line of evidence that may also be linked to the P-E records is that of dune formation and sand deposition (Fig 4d). High windspeeds across the whole of the British Isles are most often linked to the strength of westerly and southwesterly airflows, even in eastern regions (Mayes and Sutton 1997). Thus, it might be expected that increased sand deposition and dune formation would occur during phases of enhanced westerly airflow, also the most likely cause of increased precipitation. Dated records of dune formation are widely distributed but most age estimates (radiocarbon and luminescence) are from the Outer Hebrides (Figure 2). The two phases shown on Figure 4d are an approximate average of those suggested by the authors of original papers as being the dominant phases of sand deposition and dune formation (Table 2) and can only be regarded as indicative. Despite this the agreement on the two phases of enhanced activity between c. 1600 and 1100, and between c. 500 and 200 cal yr BP is strong, suggesting that these are reasonably robust estimates of periods of increased wind speed. Dune formation in earlier periods is less well constrained and conflicting between locations. The earlier periods of dune formation (2800 and 4000 cal yr BP) on the east coast of England are tentatively identified by Orford et al. (2000) on the basis of a small number of age estimates. The enhanced activity during the periods 3800-3300 cal yr BP in the Outer Hebrides (Gilbertson et al., 1999) and 3400-2400 cal yr BP in Northern Ireland (Wilson et al., 2004) are more secure, but suggest patterns of dune formation may not have been synchronous during earlier times. The multi-centennial nature of the phases of dune formation means that all of them tend to overlap both wet and dry periods in the more highly resolved P-E records. However, it is notable that the start of each of the two latter phases coincides with the start of the most prominent shifts to wetter conditions over the last two millennia.

Despite the chronological uncertainties in some individual records and radiocarbon dates, it is possible to propose a series of increasingly wet phases for the British Isles (Figure 4), based on periods of when two or more proxies indicate increased P-E. Changes to wetter conditions occur throughout the last 4500 years at the times indicated in Table 3. These do not appear to be regularly spaced throughout the last 4500 years and are almost certainly of variable magnitude, although no attempt is made here to quantify overall changes. The earlier part of the record is least robust, with fewer proxies and less agreement between proxies until after 3000 cal yr BP. Although coverage is poorer prior to 3000 cal yr BP, there is some evidence for rather drier climates during this period, with a significant increase in the number of wet events between 3000 and 2000 cal yr BP. The largest and most consistent change is in the period 1670-1390 cal yr BP when all P-E proxies show a significant increase. It is also one of only two periods of enhanced dune deposition.

6. Temperature records

Reconstructions of temperature variability, where precipitation has no or only a minor influence, are difficult to achieve in areas such as the British Isles. Proxies that are temperature sensitive in other parts of the world are not applicable to temperate regions. For example, tree rings have been widely applied to high latitude and high altitude locations where growth rate is dominated by the thermal regime, but in temperate regions such as the British Isles, trees have a more complex relationship with water balance and temperature (Briffa and Atkinson, 1997). Millennial temperature variability has been reconstructed using pollen (see above, Davis et al., 2003) but higher frequency changes are rather poorly known. Many of the ideas on late Holocene temperature variability are based on inferences from other parts of Europe and inappropriate proxies such as glacier advance and retreat (which also contain a precipitation signal), and sometimes tenuous documentary data from Roman and other early sources (Lamb 1995). Documentary data from more recent sources is more reliable so that a good deal is known about temperature variability for the last 500 years (Luterbacher *et al.* 2004), but direct evidence for centennial variability over the late Holocene is still subject to considerable uncertainties.

There are only two available records of continuous change that cover the late Holocene, based on speleothem oxygen isotope variations (McDermott et al., 2001) and chironomids from a lake sediment record (Langdon et al., 2004). Only the latter record is likely to be principally driven by temperature, and even here other factors such as land use change may have influenced the more recent part of the record. Chironomids have been used extensively as indicators of summer temperature changes for Lateglacial and early Holocene climate reconstructions including the British Isles (Marshall et al., 2007; Brooks and Birks, 2001), but mid and late-Holocene reconstructions are fewer (Velle et al., 2005). The principal difficulties are in the signal:noise ratio, because temperature changes are similar in magnitude to the reconstruction errors, and also in eliminating confounding influences such as land use and water quality change (Brooks and Birks, 2001). The record from Talkin Tarn, Cumbria (Langdon et al., 2004), is the only record published for the British Isles (Figure 5a), although several other records are likely to be forthcoming in future (Brown, 2006). It is relatively low temporal resolution but valuable because it is the only published quantitative temperature reconstruction for this period of time. Given the resolution and some problems with dating this sequence, inferences should be regarded with some caution, but the pattern of change is broadly of a decline from temperatures similar to present from just before c. 5000 cal yr BP, subsequently rising from c. 3300 cal yr BP. A sustained period of higher temperatures c. 3000-2000 cal yr BP is interrupted by several abrupt drops in temperature but as these are only individual data points there is considerable uncertainty in the timing and existence of these changes. A cooler period between 2000 and 1000 cal yr BP is followed by rising temperatures to the present day, with some decline or stabilisation after c. 700-800 cal yr BP.

There are no other records that can be directly compared to that from Talkin Tarn; the $\delta^{18}O$ speleothem from Crag Cave, southwest Ireland is perhaps the only other unequivocally temperature-sensitive record from the British Isles (McDermott et al., 2001). Oxygen isotope records in speleothems have been interpreted in a variety of ways (Fairchild et al., 2006) but the Crag cave record is seen as changes in air temperature combined with shifts in the $\delta^{18}O$ of the water source, because the changes in $\delta^{18}O$ are too large to be ascribed to temperature alone (McDermott et al., 2001). The record is well dated by uranium series age estimates and the record is very high resolution, although here the centennial variability is emphasised (Figure 5b). Low/high $\delta^{18}O$ reflects lower/higher air temperatures. It is not clear whether there is a seasonal influence on the record but there is most likely to be some effect from rainfall throughout the year. The pattern over the last 6000 years, is for cooler conditions 5900-5000 yrs BP, a brief warm phase 5000-4600 yrs BP, followed by sustained cool conditions until c. 3100 when a rise in temperature led to sustained high temperatures until c. 2100 yrs BP. The cooler phase that followed lasted until c. 1250 yr BP. Warmer conditions are inferred until c. 650 yrs BP. McDermott et al (2001) suggest that over the last 2000 years, these phases are similar to the periods known as the Roman Warm Period, Dark Ages Cold Period, the Medieval Warm Period and the Little Ice Age.

Comparing the Talkin Tarn and Crag Cave records (Figure 5), there are clear similarities in multi-centennial trends. Given the differences in the nature of the proxies, seasonality, dating methods and other possible influences on the records, the coincidence in terms of these trends is notable. After 5000 cal yr BP, the broad scale trends are for relatively cool (before c. 3100 yr BP), warm (3100-2000 yr BP), cool (2000-1250 cal yr BP), warm (1250-650 cal yr BP), cool (650 cal yr BP onwards). Mismatch on the timing of the initial decline is likely due to chronological errors in the Talkin Tarn record (extrapolated from upper sediments), and the lack of large fluctuations during 3-2.5 ka BP in the high resolution Crag Cave record suggests that the higher frequency changes based on single samples in the Talkin Tarn record may not reflect broader scale temperature variability.

Given the lack of available records against which to cross check the trends suggested here, the temperature phases defined above should be viewed as working hypotheses rather than firm conclusions. There is a clear need to develop further similar records from the British Isles or perhaps other temperature records based on different proxies.

7. Temperature and precipitation (-evapotranspiration) variability

Comparison between P-E and temperature trends can shed light on the nature of climate change and help identify changes in atmospheric circulation that must have been the proximal cause of observed variability. At millennial timescales, it is perhaps surprising that the strength of the long term mid-late Holocene temperature decline is not as marked as traditionally thought (Lamb, 1977). This may partly reflect the fact that the records shown here only cover the last 6000 (Figure 5) or 4500 years (Figure 4), but is also likely a result of smaller precessional change at mid-latitudes (Davis et al., 2003, Wanner et al., in press). Thus, centennial variability is the most important of the late-Holocene climate changes that can be reliably resolved with existing records for the British Isles. In comparing the P-E and temperature records, a key question is whether P(-E) and T change at the same time, or whether some or all changes in water balance are independent of temperature and thus presumably primarily a result of precipitation change alone rather than being a function of changing evapotranspiration.

From 4500 to c. 3100, temperature was relatively low and reasonably stable, although there may have been a temporary downturn around 4000 (Fig 5a) or 4300 (Fig 5b) cal yr BP. P-E records also reflect a generally drier climate; bog oak numbers are generally high, there are few flood peaks and peatland water tables are perhaps lower than average. The temporary downturn in temperature at or just before 4000 cal yr BP is present in bog oak and peatland records. However, a strong wet phase shown by peatland, fluvial and speleothem records at or just after 3500 cal yr BP is not present in the temperature records. There is an upward temperature trend between c.3100-2900 cal yr BP, followed by sustained higher temperatures until c. 2000 cal yr BP. Initially this is accompanied by a change to drier conditions (c. 3200-3000 cal yr BP in speleothem, peat and bog oaks), but subsequently, there is a large increase in wet phases in the P-E records, with rather wet conditions throughout much of the warm period 3100-2000 cal yr BP. It is possible there were temperature declines during this period as shown by the chironomid record, but these are equivocal, being based on only two samples and not shown by the speleothem $\delta^{18}O$ record.

The decline in temperatures from c. 2000 cal yr BP and sustained until c. 1250 cal yr BP, is also a period of large moisture variations. Initially, colder temperatures are accompanied by increased wetness in all proxies, but the main increase in P-E comes later at c.1600 cal yr BP, following a brief drier period. The warmer phase 1250-c.6-700 cal yr BP is also initially drier but again includes two short wetter phases around 1250-1150 and 850-700 cal yr BP. Finally the period after c.600 cal yr BP with cooler, but fluctuating temperatures, is also shown by the wetness proxies, although as mentioned above, the general shift to wider speleothem band thickness runs counter to this overall trend.

Overall, it seems that the initial shifts in temperature trends are also often shown by P-E proxies. In particular, the main changes to warmer (3100, 1250 cal yr BP) and cooler (4300-4000, 2000, 600 cal yr BP) conditions are also initially reflected in some or all of the P-E records. However, a number of the largest shifts in P-E are not accompanied by temperature shifts and the P-E record shows a much greater variability than temperature. Notable changes to higher P-E not reflected in temperature records occur from around c. 3550, 2750, 2550, 2250, 1650, and 850 cal yr BP (nearest 50 years from Table 3) . This may suggest that the forcing for these events is different to that during periods when P and T appear to be changing synchronously. The lack of correspondence between P-E and T during the majority of the late Holocene implies that the P-E records are more a function of P variability than changes in E, an argument supported by simulation of changes in the warm season P-E deficit resulting from T and P variability (Charman, 2007).

8. Climate forcing and dynamics

The discussion so far has focused on analysis of the evidence for late Holocene climate variability shown by terrestrial records from the British Isles. A reasonably coherent picture has emerged from comparison of multi-proxy records reflecting precipitation, temperature and, to a lesser extent storminess, despite the difficulties of chronologies, complex and poorly known climate-proxy process links and differences in seasonality. Ultimately, however, it is important to understand not just what past climate changes took place but how these recorded climate changes relate to regional atmospheric circulation and potential forcing. At multi-millennial scales, orbital forcing is the primary influence on temperature trends such as those shown in Figure 1 (b and c). Over the mid-late Holocene, there has been a decline in summer insolation and seasonality (June-December difference in insolation) at mid-high northern latitudes resulting in the slow decline in summer temperatures (Wanner et al., in press).

However, there is clearly much greater shorter term variability shown by many records and these changes cannot be explained by orbital cycles. Potential external forcings are solar variability, volcanism, greenhouse gases and land cover changes. Although all of these may be important in parts of the global climate system (Wanner et al., in press), it seems likely that it is solar variability and perhaps volcanic forcing that may have had the greatest influence on the climates of the British Isles. However, internal dynamics of the atmosphere/ocean/land system can also give rise to systematic climate variability, at least as large in magnitude as observed changes over the past 500 years (Bengtsson et al., 2006). The location of the British Isles on the western edge of the Atlantic Ocean may make the regional climate particularly susceptible to changes in ocean circulation and sea surface temperature.

Figure 6 shows some key ocean records and solar variability as indicated by atmospheric ^{14}C, compared to the summarised precipitation (-evapotranspiration) and temperature records for the British Isles. Two problems arise in attempting to make these comparisons. First, until recently the resolution of many ocean records was limited by low sediment accumulation rates. However, there are an increasing number of higher resolution records from regions with rapidly accumulating sediments which go some way to resolving this problem. Second, the chronologies for ocean sediments are often less accurate and precise than those of terrestrial proxies, partly because of lower numbers of age estimates in some records (but see Berner et al., 2008 and Figure 6), but also because of the added uncertainty of the marine reservoir effect. Both these problems make robust correlations with terrestrial records more difficult.

Bond et al. (2001) suggested that climate variability over multi-centennial timescales is primarily driven by solar variations with subsequent effects on ocean circulation via North Atlantic Deep Water (NADW) formation, and pointed to synchronous temperature changes in ocean and terrestrial proxies over the last 3 cycles (last 3000 years). However, the terrestrial evidence assessed was extremely limited in nature and coverage. The ice rafted debris record (IRD) for the last 4500 years covers the last 4 Bond cycles (Bond et al., 2001). Comparison between the IRD record and the records of NADW from two ocean cores (Figure 6, Bianchi and McCave, 1999; Hall et al., 2004) show some support for the hypothesised relationship between iceberg discharge and reduced NADW circulation, but the picture is not a clear one. During cycles 0 (c. 600 cal yr BP) and 2 (c. 2700 cal yr BP), the records are in agreement but during other cycles, records show confused or opposing trends. The two records of NADW show opposite trends in cycles 3 (c.4200 cal yr BP) and 1 (c.1600 cal yr BP), and Hall et al. (2004) remarked upon the apparent *anti*-correlation between NADW proxies and the IRD record during some phases of the Holocene (Hall et al., 2004, page 1534). The (non)coherence of these records is difficult to establish with current resolution and chronological accuracy, but a range of higher resolution, often well-dated records of ocean water temperature have become available from different locations in the North Atlantic. These allow links between ocean temperature and terrestrial climate records to be assessed. Careful consideration has to be given to both seasonality of the proxy and the ocean depth which they represent (Jansen et al., 2008). Recent diatom records provide some of the most highly resolved records of North Atlantic summer SSTs (Berner et al., 2008; Ran et al., 2008). The record of Berner et al. (2008) from south of Iceland (Site LO-09-14) is shown in Figure 6. This is also broadly consistent with the Mg/Ca temperature reconstruction of Came et al. (2007) from site ODP 984 in the same region, but the Berner et al. record is higher resolution. The last four Bond IRD cycles appear to be present in this record with pronounced drops in SST of 1-1.5^{o}C at c. 4500-4200 (their 'Holocene climate event' HCE11), 2800-2600 (HCE14), 1700-1600 (HCE17) and 900-800 cal yr BP (HCE 18), although a later (unmarked by Berner et al.) event at c.650 cal yr BP is closer in time to Bond cycle 0. A recent record from the north of Iceland also shows increased sea ice extent (Ran et al., 2008) and increased abundance of Arctic diatom taxa close to these times, supporting the idea that there were periods when SSTs declined over a large area of the North Atlantic.

Although there is clearly doubt over the correlations with NADW proxies, it is certainly plausible that there have been periods when large scale changes in ocean circulation and SSTs occurred, at approximately the time periods suggested by Bond et al. (2001). The period around 2800-2700 cal yr BP is notable by the magnitude and coherence of the signal in all proxies. Similarly, the changes at c. 600 cal yr BP and c. 1500 cal yr BP are reasonably coherent, although there is some mismatch in the exact timing and magnitude of change. The shift at around 4200 cal yr BP is less well marked and signals are equivocal in the NADW proxies. Such large scale changes would be expected to have an impact on the climate of the British Isles. An obvious question is whether these periods are the same as those identified in the terrestrial proxies from the British Isles?

The evidence for a major change at 4200-4000 cal yr (Cycle 3) is a little muted, partly because several of the records do not cover or provide only poor coverage of this period. However, a minor shift is shown by both temperature and P-E proxies (except flooding) close to this time. Evidence form other peatland records (Barber and Charman 2003) and changes in pine forest distribution (Gear and Huntley, 1991) also support the idea that there was a change to cooler and wetter conditions at this time. At 2800-2600 cal yr BP (Cycle 2), where the ocean proxies suggest the largest and most coherent changes in the North Atlantic, there is a strong signal in P-E records but a more equivocal response in temperature. In fact this event occurs during periods of generally higher inferred temperatures and the increase in P-E may not be linked to a temperature decrease. There is also a strong coherence between P-E proxies and ocean changes registered at 1600-1500 cal yr BP (Cycle 1), again in the absence of significant temperature change in terrestrial proxies, as temperatures were already at a low before this time. Finally, the start of the LIA period (Cycle 0) is shown by all proxies with a shift to cooler and wetter conditions in the British Isles at the same time as reduced SSTs, NADW circulation and increased IRD. Overall, there is good evidence that P-E in the British Isles increased during the main periods of increased IRD, decreased SSTs, sometimes accompanied by reduced NADW flow. This was accompanied by reduced temperature during the LIA and perhaps during the 4200 cal BP event.

However, it is notable that not all the significant changes in the climate of the British Isles coincide with the timing of Bond cycles. They do, however, often coincide with changes in SSTs shown by Berner et al. (2008) for site LO-09-14. The double drop in SSTs at 3600-3500 and 3400-3300 cal yr BP (HCEs 12 and 13) and similar declines at 2500-2300 (HCE 15) and 2100-1900 (HCE 16) cal yr BP are broadly in phase with periods of enhanced P-E in the British Isles, and the latter change is also coincident with the downturn in temperature (Figure 6). Furthermore, other changes in SST not noted by Berner et al. (2008) may be synchronous with P-E changes, such as the drops in SSTs at c.1250 and c.700 cal yr BP.

Bond cycles and other climate changes have often been associated with solar variability as primary drivers (Bond et al., 2001), and the broad-scale changes in SSTs tend to support this hypothesis (Ran et al., 2008). Furthermore, evidence has accumulated from terrestrial climate archives in Europe that solar variability has been associated with some of the main climate changes over much of the Holocene (Magny, 1993; Blackford and Chambers, 1995; van Geel et al., 1996; Chambers and Blackford, 2001; Mauquoy et al., 2002; Blaauw et al., 2004; Magny, 2004; Holzhauser et al., 2005). It thus seems likely that solar variability has been an important influence on natural climate variability in the British Isles, western Europe and the North Atlantic region. Several key climate transitions in the British Isles are certainly coincident with decreased solar output as reflected in atmospheric ^{14}C (Figure 6). However, it is notable that of the four Bond cycles in the last 4500 years, only the change at 2800-2600 cal yr BP and the transition to the LIA are associated with major declines in solar output. There is no exceptional excursion in ^{14}C at 1600-1500 or 4200-4000 cal yr BP. Furthermore, the higher frequency variability reflected both in the P-E records and the high resolution SST records suggest there are other events of equal magnitude that are unconnected with solar variations. Thus, although there is now strong evidence that solar variability is one of the ultimate forcings driving some of the climate changes manifest in these records, it cannot explain all of the observed changes, and claimed direct linkages between solar variability and terrestrial climate records may be rather weak (Wanner et al., in press).

Ocean processes are clearly important during some periods of time (Figure 6); the transitions at 4200-4000 and 1600-1500 cal yr BP are examples of periods when IRD increases, SSTs decrease, NADW may have decreased and P-E rises sharply in the British Isles, all without a solar trigger. Until there are higher resolution records of NADW and IRD available, it is not possible to assess whether the other periods of decreasing SSTs/increasing P-E are also associated with similar circulation changes. There are also other periods where solar anomalies are strong and there is an associated terrestrial and ocean surface perturbation, with no evidence of ocean circulation change (e.g. around 3500 cal yr BP when SSTs decline and P-E rises). It is conceivable that during these phases, solar forcing is not transmitted through North Atlantic circulation but via atmospheric processes alone.

A more complex relationship between solar forcing and internal climate system dynamics is supported by changing spectral peaks in wavelet analysis of solar and ocean proxies (Debret *et al.* 2007). IRD and NADW proxies show a near-1500 year periodicity during the late Holocene but solar variability (^{14}C) shows only a much longer periodicity in the second half of the Holocene. A 1600 year periodicity shown by the Na+ record of the GISP2 ice core is thought to represent the actual periodicity of late Holocene North Atlantic climate more accurately, with the ultimate forcing identified as ocean circulation, as suggested by Broecker et al. (1999).

It is now apparent that a simplistic view of late Holocene climate change and solar forcing is not supported by the available data, despite the different ways in which many records can be interpreted given our current level of understanding of proxies, temporal precision of records, and the precision and accuracy of chronologies. A simple linear link between solar variability and terrestrial climate change cannot be demonstrated, and it seems more likely that a combination of solar activity (and perhaps large volcanic eruptions), together with internal variability of the thermohaline circulation, and atmospheric phenomena such as ENSO and the NAO, and feedback processes within the earth system need to be invoked to explain the observed changes in the British isles and elsewhere (Wanner et al., in press).

9. Conclusions

The main purpose of this paper is to summarise late Holocene climate change for the British Isles using multi-proxy data of sufficient temporal resolution to identify centennial variability. It is clear from a range of proxies that there have been a number of century-scale climate changes over the last 4500 years and that these do not always fit with conventional views of climate periods, which have often focused on temperature, assumed a necessary correspondence between temperature and precipitation, and neglected the importance of seasonality in proxy records. Most available records for the region are proxies for water balance rather than temperature *per se*, and therefore reflect precipitation-evapotranspiration, probably with a greater influence of precipitation over evapotranspiration. The British Isles have experienced much more important changes in precipitation (-evaporanspiration) than they have in temperature. P-E records show significant high frequency changes over the last

4500 years, whereas temperature probably changed over longer periods, and changes in P-E and temperature did not always occur simultaneously. Changes in storminess also occurred during at least two key phases of change shown by dates of dune building and sand movement. There is good evidence that terrestrial climate variability in the British Isles is linked to changes in the North Atlantic, including Bond cycles when IRD flux was increased, SSTs were lower and NADW circulation was probably reduced. A strong relationship between North Atlantic SSTs and P-E in the British Isles throughout the late Holocene suggests that ocean variability was more important for precipitation than for temperature changes in the British Isles.

Assembling multi-proxy data provides a means of re-evaluating process understanding of individual proxies and demonstrates a need for more careful consideration of seasonality of in the proxy-climate signal. The compilation of multi-proxy records from within regions such as the British Isles is not without its problems. Chronological precision and accuracy of non-annually resolved records remains a key difficulty, although the use of large data sets or replicate records can overcome some of these problems and allow the identification of century-scale patterns with greater confidence. However, there are still considerable uncertainties in correlation between records, especially during the more high frequency changes in the P-E records. Replication of individual proxy records is also still important to gain more complete and reliable estimates of past changes. Within the British Isles, temperature reconstructions that are unaffected by precipitation variability are especially in need of further work to test the patterns of change suggested here.

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Table 1: Key proxy records of Holocene climate change for the British Isles, with references selected for Figures 3-5 listed. See Figure 2 for locations. of references included in Figures 3-5 and Table 2.

Archive & location	**Proxy**	**Climate parameter(s)**	**Time period**	**Sources referred to here**
Speleothem NW Scotland	Annual band width	Annual T and P (mostly P)	3500-present	Proctor et al 2002
Speleothem S Ireland	$\delta^{18}O$	Annual T (and P source)	Holocene	McDermott et al. 2001
Peat N England & Scotland	Surface wetness (macrofossils, testate amoebae, peat humification)	Warm season P-E	Holocene, most mid-late	Charman et al. (2006) Stacked record of 12 cores
Bog oaks N Ireland	No. dated trees, Mean age	Warm season P-E (via bog surface wetness)	7500-present	Turney et al. 2005
Fluvial Britain	Flood deposit dates	P-E	Holocene	Johnstone et al. 2006
Dune and sand deposits British Isles	OSL, ^{14}C ages of sand layers	Storminess	Holocene (esp late)	See Table 2
Lake NW England	Chironomids	Mean July T	6000-present	Langdon et al. 2004

Table 2: Periods of dune formation and sand deposition based on radiocarbon and OSL ages in published sources covering the last 4500 years. Average ages are only estimated for periods where there is consistency between locations. Average ages are approximate start and end dates of phases of enhanced dune deposition, as identified by authors in discussion sections and summary diagrams. Phases based on multiple age estimates. All radiocarbon ages are calibrated to cal yr BP. *Lagoon sediments (Carrownisky) only.

Source and location	**Cal yr BP**			
	Later	LIA	Dark ages	Earlier
Bateman and Godby, 2004 East Anglia	30-200	335-400, 500	1100-1600	
Gilbertson et al., 1999 Outer Hebrides		200-600	1300-1700	3800-3300
Dawson et al., 2004 (main phase) Outer Hebrides		250-550		
*Delaney and Devoy, 1995 Western Ireland		100-500		
Orford et al., 2000 Norfolk and Northumberland		200-500	1000-1500	2800, 4000
Wilson et al., 2004 N. Ireland		650-50		3400-2400
Approx. 'average' periods		**220-510**	**1130-1600**	

Figure 1: Records of LGIT and Holocene climate change. a) The oxygen isotope curve from Greenland ice cores GISP2, GRIP and NGRIP (Greenland Summit Ice Cores CD-ROM. 1997, Vinther et al., 2006, Rasmussen et al., 2006, NGRIP dating group, 2006). b) and c) Seasonal temperature changes reconstructed from pollen data for the north west and central west European sectors (Davis *et al.* 2003). MTCO – Mean temperature of the coldest month, MTWA – Mean temperature of the warmest month. d) Temperature curves estimated by Lamb (1977) from palaeobotanical and documentary evidence, ovals represent estimates of uncertainty based on both palaeobotanical indicators and chronologies. In a), and d) timescales adjusted to cal yr BP from original published data (estimated for d).

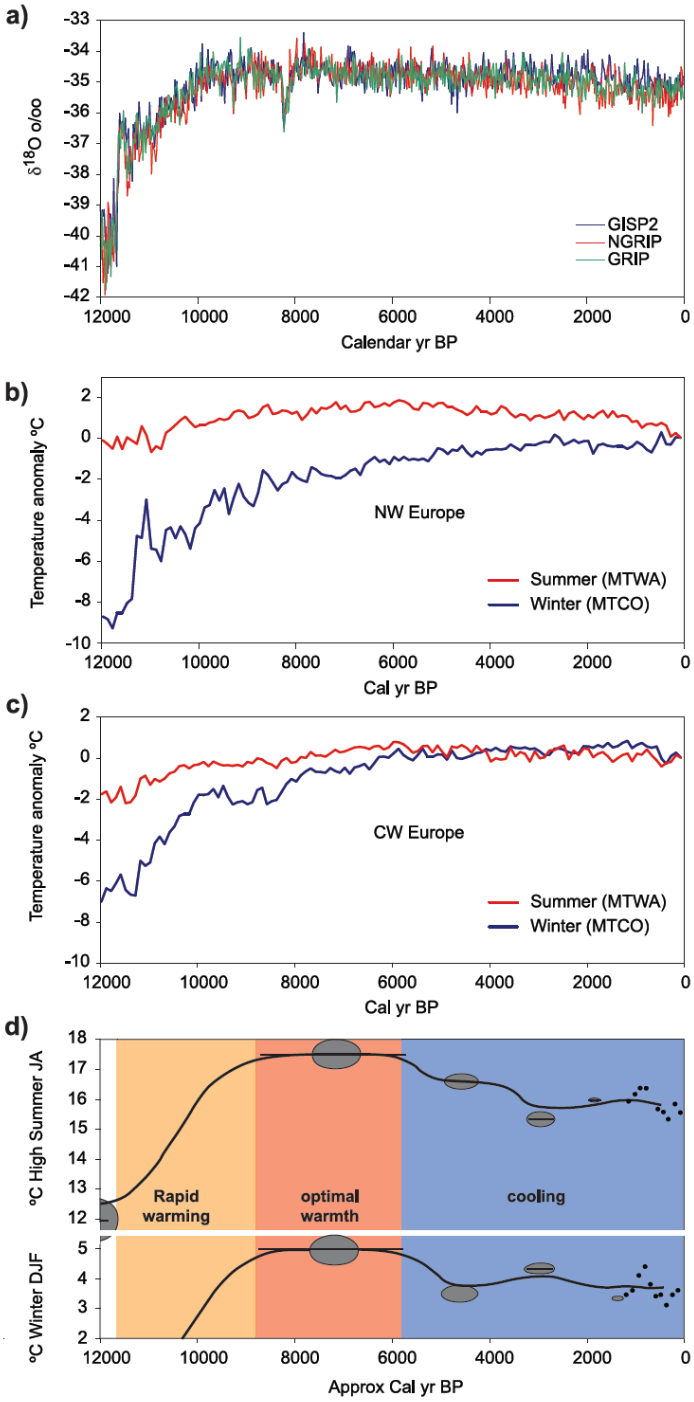

Figure 2: Location of main records referred to in Table 1 and shown in Figures 3-6, according to proxy.

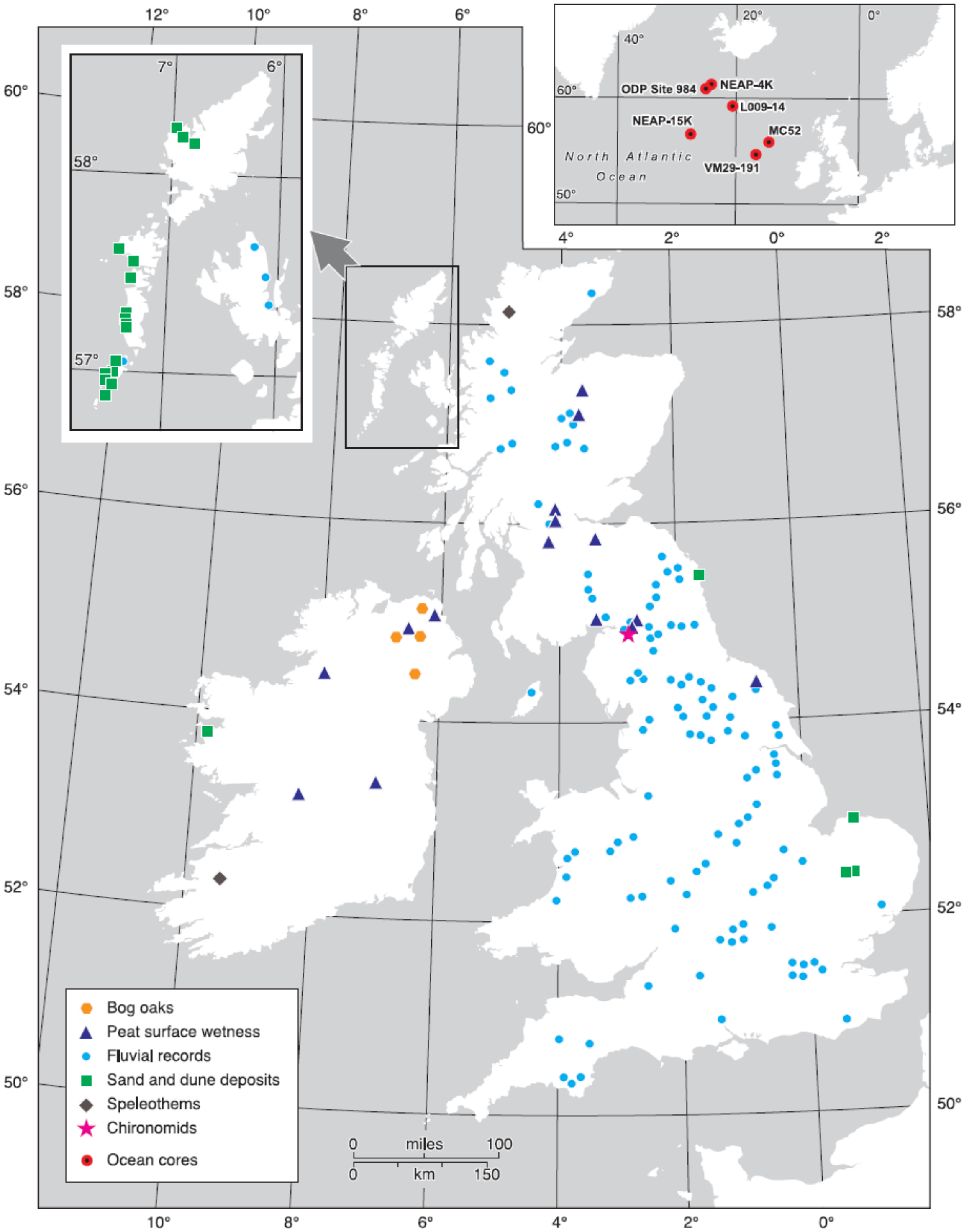

Figure 3: The Irish bog-oak record and records of bog surface wetness from Ireland and northern Britain. a) Bog oak numbers and b) Mean bog oak age. Lighter shading (blue) shows wet phases as originally inferred by Turney et al (2005) and heavy (green) shading shows those inferred here defined by periods of declining bog oak numbers. c) to e) Bog surface wetness records; c) Northern Britain composite record of 12 water table reconstructions from testate amoebae analysis (Charman et al. 2006), d) Start and duration of north of Ireland bog wet phase based on three precisely dated multiproxy records (Swindles et al. 2007), e) Southern Ireland multi-proxy surface wetness record based on two sites (Blundell et al. 2008). In c) to e) periods of increasing wetness (rising water tables) are highlighted by the shading.

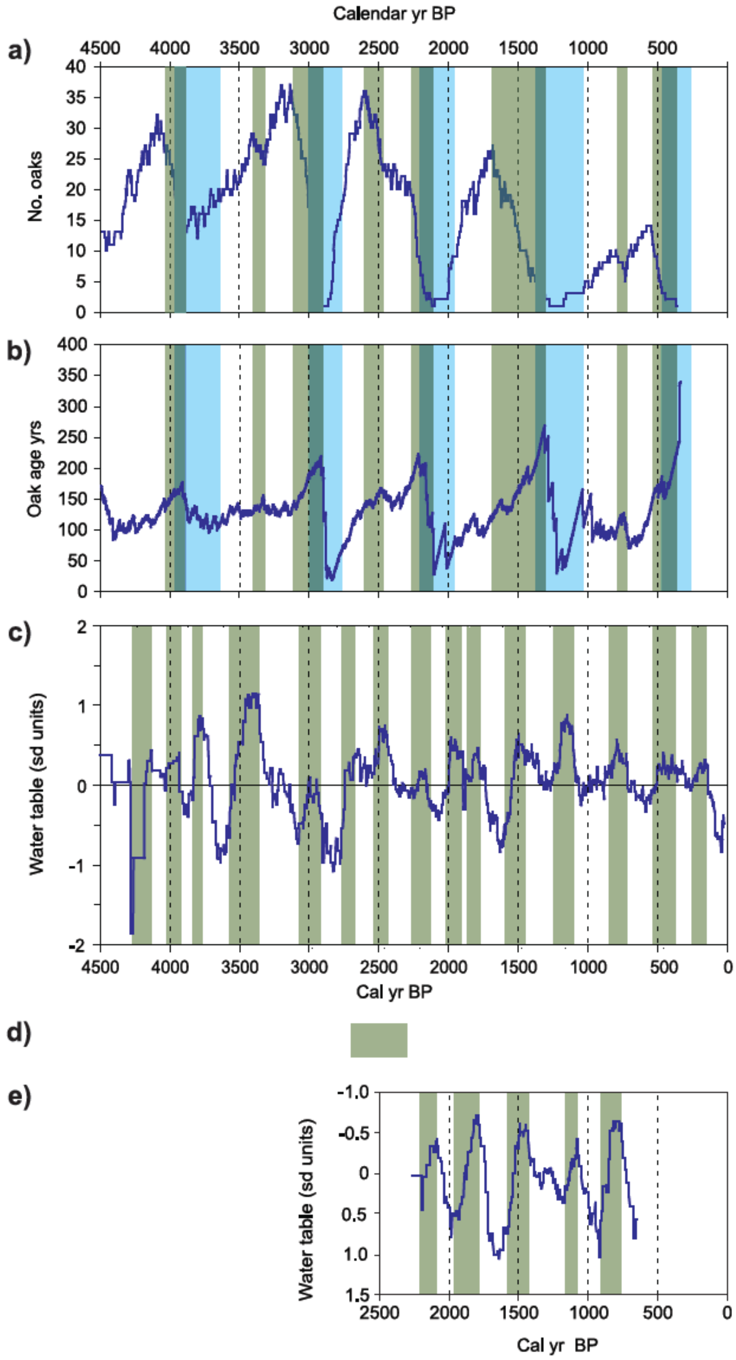

Figure 4: Precipitation and precipitation-evapotranspiration records. a) Bog oak numbers, b) peatland surface wetness. a) and b) are as Figure 3. c) Fluvial records of flood deposits expressed as a probability difference function with peaks in flooding identified by arrows (Johnstone et al. 2006). d) Speleothem band width record of Proctor et al (2002), with 100 year mean (red curve). Horizontal bars show key periods of dune formation (see Table 2). Vertical green bands identify common periods of increasingly wet conditions shown by two or more proxies (See Table 3). Arrows indicate dry (pale, yellow) and wet (darker, green) directions of axes.

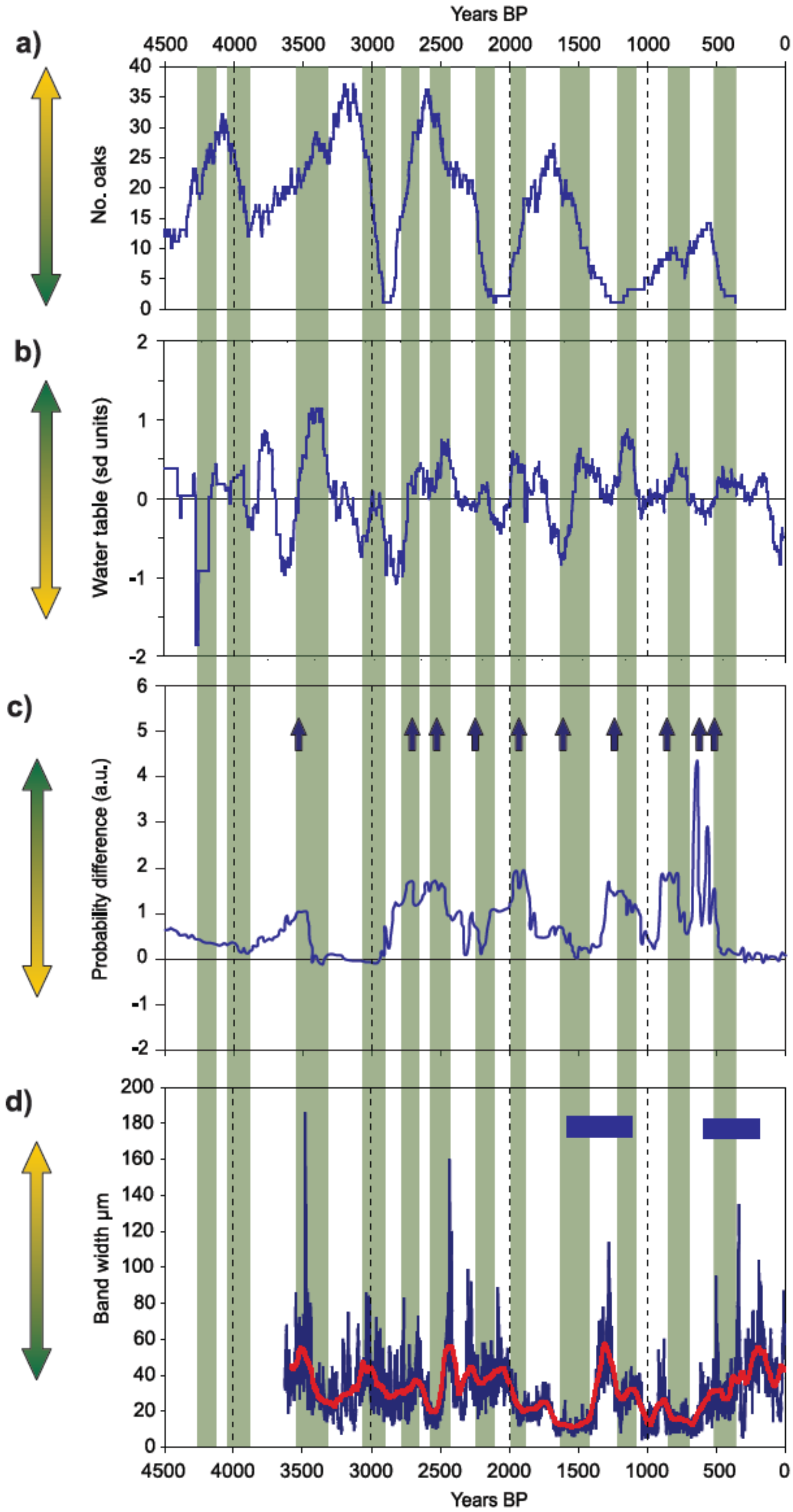

Figure 5: Temperature related proxy records. a) lake sediment reconstruction based chironomid head capsules (Langdon et al., 2004). Thin (red) line shows individual data points and thicker (black) line shows three-point smoothed record. b) Oxygen isotopes from Crag Cave speleothem (McDermott et al. 2001), showing only the last 6000 years of the record. 10 point smoothed record shown to highlight approximately centennial trends (data points are every 10-18 years). Age estimates used to derive age model shown on both records (radiocarbon for a, and U series for b). Red/blue arrow indicates direction of warm/cold on axis.

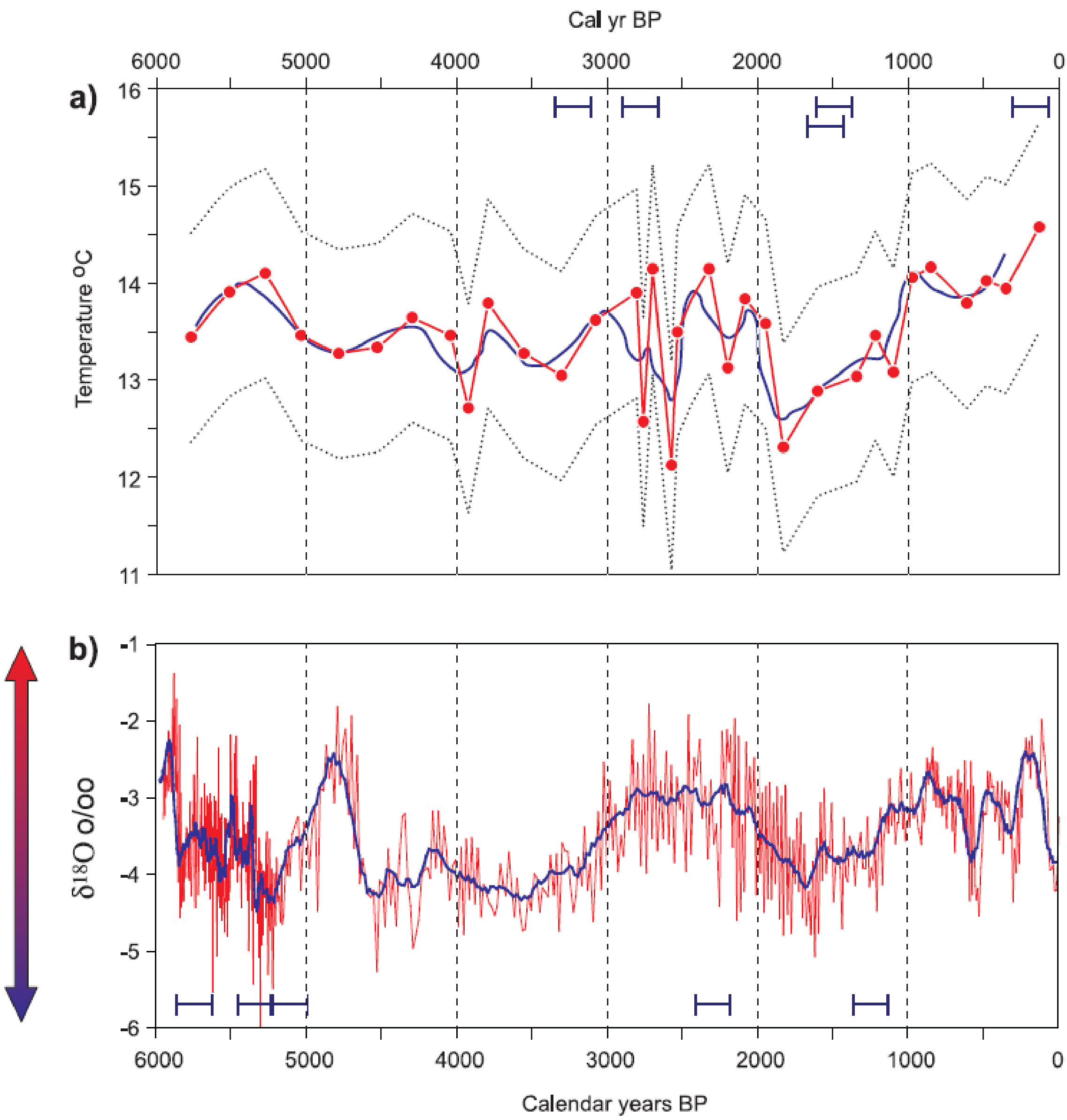

Figure 6: Summary of key ocean records (see Figure 2 for locations), together with representative records of terrestrial climate in the British Isles and solar variability. a) Ice rafted debris from MC52 and VM29-191 (Bond et al., 2001), with cycle numbers. b) Sortable silt mean size for (upper) NEAP4K (Hall et al., 2004) and (lower) NEAP15K (Bianchi and McCave, 1999). c) Diatom-based SST reconstruction for LO09-14 (Berner et al., 2008). Note inverted scale. Grey shaded bars are the Holocene Climate Events (HCEs) identified by Berner et al. d) Peatland surface wetness record (Charman et al., 2006) with key wet phases superimposed from Figure 4, arrows from fluvial record (Figure 4 and Johnstone et al., 2006). e) $\delta^{18}O$ record from Crag cave speleothem (McDermott et al., 2001) with main warm/cold phases superimposed from Figure 5. f) Residual ^{14}C as an indicator of solar variability from INTCAL04 (Reimer *et al.* 2004). Radiocarbon age estimates shown as bars for individual ocean records, except d), which has 20 age estimates included in the age-depth model from 4500 cal yr BP to present. Arrows indicate warm/cold as red/blue (pale/dark), and wet/dry as green/yellow (dark/pale) directions of axes.

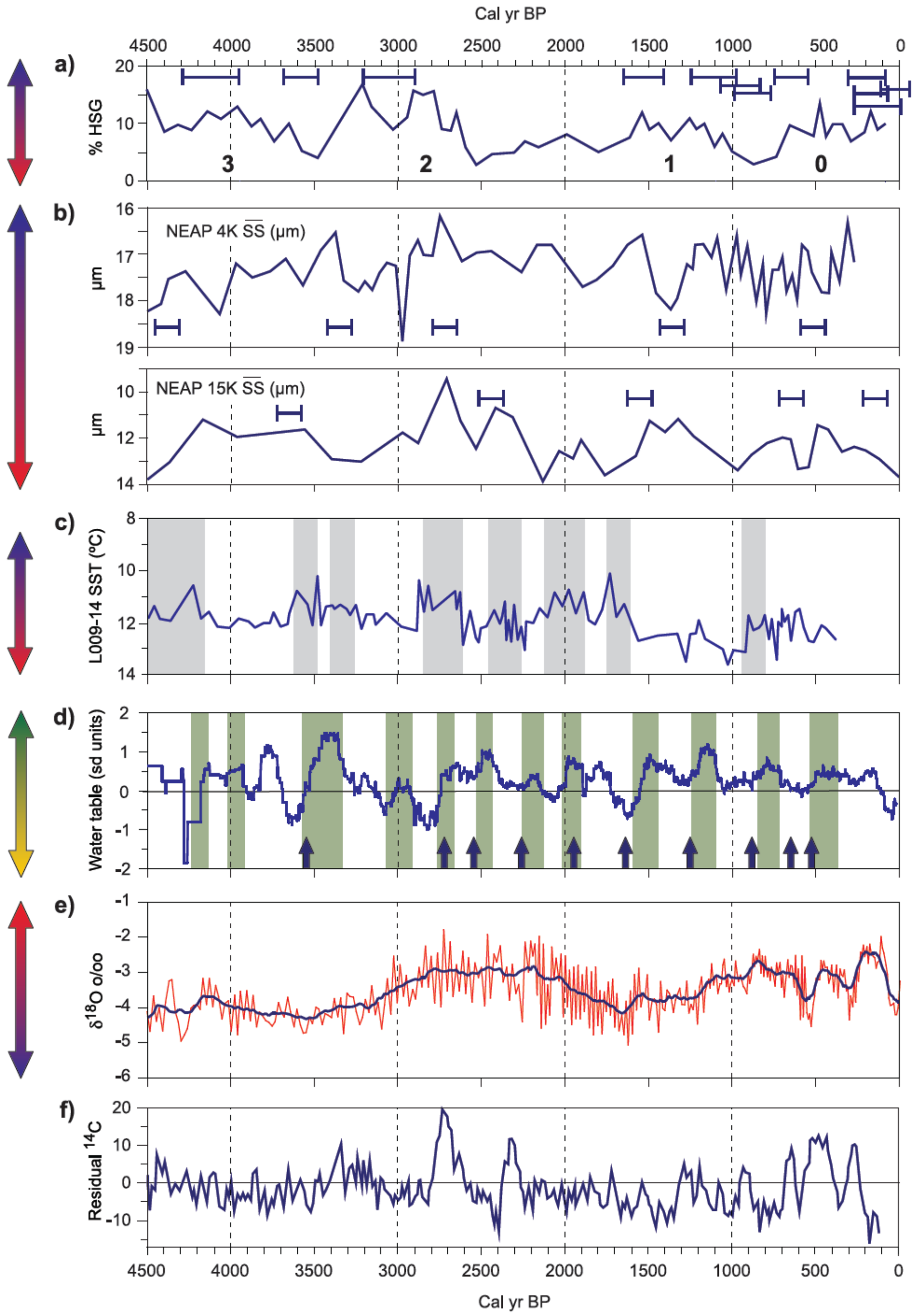

Table 3: Summary of periods of high P-E in the British Isles identified by centennial changes in two or more proxy records (Figure 4). Shaded cells identify proxies in which there is positive evidence for wetter conditions in each phase. Unshaded boxes show periods in which there is no clear evidence for change. X indicates proxy data suggesting drier conditions. See text for detailed discussion of each proxy. Age estimates for the start and finish of the phases are those shown on Figure 4, are approximate, and derived from dates in the different proxy records. *The end of this phase is not adequately defined by >1 proxy.

Wet phase	Bog oak	Peatland	Fluvial	Speleothem	Sand/Dune
*570-				X	
860-720					
1290-1140					
1670-1370					
2000-1870					
2260-2140					
2570-2460				?	
2770-2720					
3060-2870					
3550-3300					
4030-3870				N/A	
4260-4120				N/A	