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New insights into Late Devensian Lateglacial and early Holocene environmental change: two high-resolution case studies from SE England

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1. Introduction

Despite a number of palaeoenvironmental records across south-east (SE) England, they are unevenly distributed and concentrated in coastal regions, especially the Kent and East Sussex coast (Kerney et al., 1980; Thorley, 1981; Waller, 1993; Waller and Marlow, 1994; Preece and Bridgland, 1998; Waller et al., 1999; Waller and Hamilton, 2000; Waller and Long, 2010), and the Solent (Scaife, 1980, 1987) (Figure 1). There is also a considerable bias towards records from floodplain locations, particularly along the lower reaches of the River Thames (Gibbard and Hall, 1982; Branch and Lowe, 1994; Wilkinson et al., 2000; Corcoran et al., 2011; Batchelor et al., 2012; Branch et al., 2012; Batchelor et al., 2014; Batchelor et al., 2015; Batchelor et al., 2019) and the Upper River Thames and River Kennett (Healy et al., 1992; Lewis et al., 1992; Parker and Robinson, 2003; Chisham, 2004). There are only a small number of sites from inland, lowland wetland locations that have provided well-dated palaeoenvironmental records, including Spartum Fen (Parker, 2000), Bagshot (Groves, 2008), Conford (Groves et al., 2012) and Elstead Bog (Farr, 2008), although they do not all provide a continuous Lateglacial and early Holocene sequence.

Current understanding from the limited available data suggests that four main biostratigraphic zones exist (Branch and Green, 2004; Groves, 2008; Groves et al., 2012) during the Lateglacial and early Holocene (Table 1). Pre 11,500 cal. BP, there was open scrubby tundra style vegetation comprising *Juniperus* type, *Betula* and *Salix* woodland that often continued into the early Holocene. From 11,500–11,200/10,500cal. BP *Pinus* became dominant in dryland areas alongside *Betula*. On the wetland edge *Salix* continued to be the dominant taxon, alongside Cyperaceae. Between 10,200–8900 cal. BP mixed deciduous woodland of *Quercus*, *Betula*, *Corylus* and *Ulmus* developed in previously *Pinus* dominated areas; *Salix* dominated the wetland edge. From 8900–7000 cal. BP mixed deciduous woodland dominated drier areas, with the emergence of *Tilia*, the loss of *Betula,* with *Alnus*-carr woodland formation in the wetter areas.

Table 1. Palaeoenvironmental sites in SE England with evidence for one or more of the biostratigraphic zones

This observed framework requires testing and refinement however due to the uneven distribution and paucity of sites. Here we present a comparison and discussion of the pollen-stratigraphical and sedimentary results from two new unique Lateglacial and early Holocene sequences in SE England: Langshot Bog and Elstead Bog B, which are integrated with data from Elstead Bog (Carpenter and Woodcock, 1981; Farr, 2008), other sites in SE England (Figure 1) and more widely in Britain and NW Europe. These two sites develop our knowledge on the nature of vegetation succession, and temporal variability in vegetation dynamics, during the Lateglacial and early Holocene, and provide an opportunity to contribute to current debates about vegetation change as a response to known periods of abrupt climate change and human activity (Barton and Roberts, 2004; Barton, 2009; Rasmussen et al., 2014).

Figure 1. Location map of the two study sites and other key palaeoenvironmental sites in SE England.

1.1. The sites: Langshot Bog and Elstead Bog B

Langshot Bog (Figure 1) is located on the southern side of Chobham Common (51° 21' 43.0"N 00° 35' 53.2"W; elevation: *c.* 30m a.s.l.) and is a 0.12km^2 area of superficial peat, situated on the Bracklesham Group. Chobham Common is a large expanse of heathland in Surrey and is the largest National Nature Reserve in SE England, managed by the Surrey Wildlife Trust. On Langshot Bog, species of *Sphagnum* occupy the bog surface (*Sphagnum compactum, S. papillosum, S. recurvum* and *S. palustre*) alongside other plants such as *Drosera rotundifolia, Narthecium ossifragum, Eriophorum angustifolium*, *Osmunda regalis,* and *Gentiana pneumonanthe*. Much of the bog surface is wooded, with *Betula pubescens* and *Alnus glutinosa* the dominant tree taxa. The modern-day vegetation across the Common itself is dominated by *Calluna vulgaris, Erica cinerea,* and *Ulex minor* (plant nomenclature follows the flora of the British Isles Stace (2010)).

Figure 2. The location of Langshot Bog highlighting the coring transects and sampling location

Elstead Bog B (Figure 1) is a ~0.003km^2 peat basin located within pine woodland in Surrey (51° 10' 18.3"N 00° 43' 00.6"W; elevation: *c.* 52m a.s.l.), situated on gently downward sloping ground towards the Wey Valley to the north, with the ground rising steeply to the south. The site is nearby to the previously investigated Elstead Bog (Seagrief and Godwin, 1960; Carpenter and Woodcock, 1981; Farr, 2008). The underlying geology is well sorted Folkestone Bed sands of the Lower Greensand (Carpenter and Woodcock, 1981). Soils are thin and podzolised, immediately resting on the Lower Greensand. Bog surface vegetation is dominated by *Molinia caerulea* and *Sphagnum (Sphagnum compactum, S. papillosum, S. recurvum,* and *S. palustre)*, with *Betula pubescens* on stable areas. Around the bog margins, *Calluna vulgaris* and *Erica tetralix* prevail, with *Pinus sylvestris* in the wider woodland.

Figure 3. The location of Elstead Bog B showing the coring transects and the site of Elstead Bog

2. Methods

A stratigraphic coring survey was used to understand the size and shape of the two sites. At Langshot Bog, two transects, running east-west, and north-south, were sampled at 20m intervals, which resulted in 9 cores running west to east, and 13 cores running south to north (Figure 2). Cores were extracted until the corer hit sandy Bracklesham group sediments, identified within sediments in the base of each core, or observed whilst coring. At Elstead Bog B an east to west transect of 9 boreholes (5 metre spacing) and a north to south transect of 11 boreholes (10 metre spacing) cored down to the parent material, the Sandgate Formation (Figure 3). The deepest surveyed points were used as master coring locations. All cores were collected using a Russian corer (or D-section corer), a hand coring device that cuts an undisturbed 500mm x 50mm section of core, a method widely used in palaeoenvironmental sampling because of the high quality of extracted sample (Moore et al., 1991) and the speed and ease of operation (Jowsey, 1966). At each site, two boreholes, situated no more than 30cm apart, allowed for each core section to have an overlap of 10cm, and meant there was no disturbance of lower samples. Cores for further analysis were wrapped in rigid plastic downpipe and kept in cold storage (<4˚C) at the University of Reading. Core stratigraphy was described by reference to the Troels-Smith classification scheme (Troels–Smith, 1955). Colour was noted from a Munsell Colour Chart and determination of organic content by loss-on-ignition (Bengtsson and Enell, 1986).

The alkali-soluble (‘humic acid’) and alkali- and acid-insoluble (‘humin’) fractions of bulk peat samples taken from above and below lithostratigraphic and biostratigraphic unit boundaries were dated by Accelerator Mass Spectrometry (AMS) at the Scottish Universities Environmental Research Centre (SUERC) and ^{14}CHRONO centre, Queen’s University Belfast. The samples dated at Queen’s University Belfast were pre-treated and measured following the methods described in Reimer et al. (2015). Despite six attempts trying both hydrogen reduction (four times), and zinc reduction (twice) the humin sample (UBA-26775) from Langshot Bog (137–138cm) failed to graphitise. Usually a sample fails to graphitise due to a higher than normal sulphur content and thus the laboratory increased the amount of silver present in the combustion tube but unfortunately this failed. Subsequent analysis of the sediment from this portion of the core identified very high concentrations of pyrite (FeS_2). Pyrite releases sulphur oxides when combusted which can hamper the reduction of carbon dioxide to graphite (Proske et al., 2015).

The samples dated at SUERC were pre-treated using methods outlined in Stenhouse and Baxter (1981), combusted following Vandeputte et al. (1996), graphitised as described by Slota et al. (1987), and measured by AMS (Freeman et al., 2010). Internal quality assurance procedures and international inter-comparisons (Scott, 2003; Scott et al., 2010) indicate no laboratory offsets and validate the measurement precision quoted. Details of the dated samples, radiocarbon ages, and associated stable isotopic measurements are provided in Tables 2-3. The radiocarbon results are conventional radiocarbon ages (Stuiver and Polach, 1977). A weighted mean (Ward and Wilson, 1978) has been taken as providing the best estimate for the age of formation of the dated horizons (Tables 2–3) before inclusion in the age-depth model. The age-depth model (Figure 5) has been constructed using the program OxCal v4.2 (Bronk Ramsey, 2009; Bronk Ramsey and Lee, 2013) and the atmospheric calibration curve for the northern hemisphere published by Reimer et al. (2013). The P_Sequence Poisson process model (Bronk Ramsey, 2008) employs a variable k parameter (Bronk Ramsey and Lee, 2013) with the overall age-depth model defined as P_Sequence (1,1,U(-2,2)), with k_0 (the base k parameter) =1cm^{-1}, the interpolation rate = 1cm^{-1} (output from the model given every 1cm), and variability in k allowed between a factor of 10^{-2} and 10^{2}.

Sub-samples for pollen analysis were derived from a 1cm^{3} volumetric sampler (Bennett and Willis, 2002). At Langshot Bog and Elstead Bog B, sampling at an interval of 2cm provided a very high-resolution vegetation record in respect to core length. At Langshot Bog this equated to an average time-depth resolution of 50 years per sample (25 years cm^{-1}). Samples were prepared for pollen analysis with the addition of the exotic pollen marker *Lycopodium* (Stockmarr, 1971) and then underwent heavy liquid flotation (Branch et al., 2005). Acetolysis removed extraneous organic matter (Moore et al., 1991). A Leica DME microscope at ×400 and oil immersion ×1000 magnification was used for all identifications. Identification involved the use of the University of Reading reference collection and pollen keys (Moore et al., 1991; Reille, 1995), and pollen nomenclature follows Moore (et al. 1991). Pollen percentage values were calculated in Tilia version 1.7.16 (Grimm, 2011) based upon a pollen sum of ~300 total land pollen (TLP). This count excluded aquatic species and spores; subsequently calculated as a percentage of their totals + TLP. Microscopic charred particles were recorded quantitatively using a modified method from Robinson (1984) where MCP were counted relative to the trees, shrub and herb pollen count at the same time as pollen grains (Simmons and Innes, 1996; Innes et al., 2004; Innes et al., 2010). Different size classes for the charcoal were not implemented, due to the potential for fragmentation during the pollen preparation process, potentially artificially increasing charcoal counts in small size classes (Innes et al., 2004; Innes et al., 2010). Tilia.graph (Grimm, 2011) was used to draw pollen diagrams, and constrained incremental sum of squares clustering (CONISS) was used to help divide the percentage pollen diagram into a number of local pollen assemblage zones (Grimm, 1987). Pollen influx values were calculated for the Langshot Bog sequence, based upon the pollen concentration (grains cm^{-3}) and deposition time (years cm^{-1}). Values are expressed as a pollen deposition rate (grains cm^{-2}year^{-1}) and represent the number of grains incorporated into the unit area of sediment each year.

Samples for stable isotope analysis (δ^{18}O and δ^{2}H) were extracted at the same locations as the pollen samples from the Langshot Bog sequence. Individual culm and leaf macrofossils of *Carex* were extracted from the core and checked under a Leica S6D zoom-stereo microscope at x10 magnification to ensure homogeneity throughout the sequence. These leaf macrofossil samples were dried and weighed into silver capsules and analysed for δ^{2}H and δ^{18}O on a Thermo-Fisher Delta V Advantage IRMS fitted with a TC Flash EA at the University of Reading in the Chemical Analytical Facility. The IRMS data was corrected for both drift and stretch corrections using international standards following blank corrections (Eley et al., 2014). Precision on independent standards run alongside the unknown samples was ±0.15‰ for δ^{18}O and 1.58‰ for δ^{2}H.

3. Results

3.1. Langshot Bog: Lithostratigraphy, chronology and stable isotopes

The transect survey at Langshot Bog indicates that the unconsolidated sediments running east–west have a depth of 150–210cm. The north–south sequence has an undulating profile of shallow and deeper cores up to 242cm in depth. The deepest location on this north–south transect (51° 21' 43.0"N 00° 35' 53.2"W) was chosen as the master core sequence for Langshot Bog (Figure 4).

Figure 4. Stratigraphic results from the two Langshot Bog transects.

The sediments at the base of the sequence (unit 1) are sand-rich, representing the local Bracklesham geology at the site, with low organic matter values (~40%). Organic sediment accumulation (unit 2) started prior to *12,570–11,910 cal. BP* (*95% probability*[1], *LB1/2*, Figure 5) and lake sediment has been observed at the base of other sequences from the site, immediately overlying the Bracklesham Beds. Palaeo-hydrological studies often interpret shifts in sedimentary organic matter δ^2H and δ^{18}O values in terms of palaeo-precipitation (Pagani et al., 2006; Schefuß et al., 2011; Magill et al., 2013). This relies on the assumption that any temporal change in the isotopic composition of plant δ^2H and δ^{18}O values of the sedimentary archives reflects shifts in the isotopic composition of source water through time (Leider et al., 2013). The trends observed within the Langshot Bog isotopic data appear to show that the δ^{18}O and δ^2H records are not synchronous and are decoupled, matching trends observed previously in other stable-isotope studies (Eley et al., 2014) where the δ^2H data reflects the plant community, whilst a climatic signal is provided by the δ^{18}O data. During this period of peat formation, it is likely that small pools were present across the site occupying natural hollows in an undulating sandy land surface. A short period of decreasing evaporation (as indicated by depleted δ^{18}O values) may indicate slightly drier conditions to

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[1] [1] Highest Posterior Density intervals which describe the posterior distributions derived from age-depth modelling are given in italics (and quoted at 95% probability) to distinguish them from simple calibrated dates that are given in plain text (and are quoted at 2σ).

allow peat formation to begin, although the low humification of the peat in this unit suggests that the bog surface was still relatively wet.

Peat formation continued in unit 3 until *12,010–11,640 cal. BP* (*95% probability, LB2/3,* Figure 5). Shallow bog pools and wet peaty conditions were present at this point. The wet bog surface conditions indicated by the vegetation community correlate with the isotope records through this period (Figure 6), which indicate a marked shift to wetter conditions (depleted δ^2H) with increased evaporation (enhanced δ^{18}O). A small influx of sands and silts in units 2 and 3 are thought to represent in-washing from the higher elevated slopes around the edge of the basin, potentially caused by strong storm events, corresponding with wetter conditions or a loss of surface vegetation. An increase of trees on the surface of the bog (unit 4) from *12,010–11,640 cal. BP* (*95% probability, LB2/3,* Figure 5), suggested from wood peat (unidentified) found within this unit, correlates with a distinctive shift in the δ^{18}O values indicating lower levels of evaporation. This may be a response to an early Holocene drying out of the surface of the bog. A period of relatively stable peat accumulation then continues from the onset of the Holocene until *9790–9470 cal. BP (95% probability, 96cm,* Figure 5*).* During this time, localized changes in wetness are highlighted within the isotope data, which shows a period of instability, with overall trends indicating increased evaporation and slightly wetter conditions as this unit progresses. Two events within this period of instability stand out, with decreased evaporation at *11,350-10,715 cal. BP* (*95% probability, 162cm,* Figure 5) and at *10,850-10,520 cal. BP* (*95% probability, 140cm,* Figure 5), although any response from the vegetation (δ^2H values) appears muted. An increase in humification in unit 5, aligned with an increase in *Substantia humosa,* indicates a further drying out of the bog surface from *9790–9470 cal. BP* (*95% probability, 96cm,* Figure 5) to *8430–8350 cal. BP* (*95% probability, wm65,* Figure 5). The δ^2H record identifies a broad drying trend, alongside a reduced amount of evaporation suggested by the δ^{18}O profile. An increase in mineral influx and decline in organic matter is attributed to an increase in burning events

(Figure 6), causing increased runoff and in-washing of mineral matter into the basin. A hiatus in the sedimentary record is identified between units 5 and 7, thought to have occurred from *8430–8350 cal. BP* (*95% probability, wm65,* Figure 5) (unit 6), where there is a large influx of mineral material.

Figure 5. Bayesian age-depth model of the chronology of the sediment sequence at Langshot Bog (P_Sequence model (k=0.01–100); Bronk Ramsey 2008). For each of the radiocarbon dates two distributions have been plotted: one in outline, which is the result of simple calibration, and a solid one, which is based on the chronological model used. Figures in brackets after the sample numbers are the individual indices of agreement which provide an indication of the consistency of the radiocarbon dates with the prior information included in the model (Bronk Ramsey, 1995).

Table 2. Radiocarbon dates from Langshot Bog. Replicate measurements have been tested for statistical consistency and combined by taking a weighted mean before calibration as described by Ward & Wilson (1978; T'(5%)=3.8, v=1).

Figure 6. Composite stratigraphy, isotopic results, and age-depth model for Langshot Bog.

3.2. Langshot Bog: Pollen Stratigraphy

The Lateglacial at Langshot Bog (Figure 7: Local Pollen Zone (LB-1)) shows vegetation indicative of a cold, tundra-style landscape of shrubland and short turf grassland with Ericaceae and Cyperaceae. It was species-poor with natural burning events and characterised by a very sparse scrubby scatter of *Betula* and *Juniperus* type. These species are cold tolerant, which is important as this zone can be chronologically correlated with the start of the Loch Lomond Stadial (Walker et al., 1994), dating from *13,210–12,430 cal. BP* (*95% probability, 241cm,* Figure 8) to *12,570–11,910 cal. BP* (*95% probability, LB1/2,* Figure 8). Ericaceae is present due to its ability to regenerate after fires (Averis, 2013), which based on the charcoal record would have been regular occurrences throughout this period. Regular fires have been observed at other sites in the UK and NW Europe during the last Glacial period (Kolstrup, 1992; Edwards et al., 2000) although understanding their onset has been difficult, a point explored further in the discussion. Ericaceae, and *Betula*, are resistant to wind exposure and are likely to indicate the presence of small heathland patches as observed elsewhere in the UK (Bennett, 1983; Walker et al., 2003). *Juniperus* type colonises unstable ground surfaces and can survive in poor soils with low nutrient content (Averis, 2013). The low percentages (predominantly <5%) suggest that dwarf *Juniperus* type was present, and alongside *Artemisia* indicate year-round temperatures would have been relatively cold (Kolstrup, 1980). *Pinus* might have been present on both drier and wetter soils but the low influx values (Figure 9) may also indicate long distance pollen transport.

Figure 7. Langshot Bog, Chobham Common: pollen percentage diagram.

LB-2 is a similar tundra-style vegetation mosaic, and the dating has enabled correlation with the mid–late Loch Lomond Stadial: *12,570–11,910 cal. BP* (*95% probability, LB1/2,* Figure 8) to *12,010–11,640 cal. BP* (*95% probability, LB2/3,* Figure 8). A consistent influx of *Salix* suggests it is underrepresented in the pollen percentage data and that it would have formed a component of the local woodland cover, around the wetland margins. The relatively open landscape present at this time suggests that *Salix* and *Betula* are likely to have been dwarf species, with both *Betula nana* and *Salix herbacea* identified in British Lateglacial contexts (Blackburn, 1952; Birks, 1965; Beerling, 1998). There was no attempt to do size frequency measurements on the *Betula* (Birks, 1968), as no distinct difference was observed and therefore, we cannot be definitive about the *Betula* species present during this time. *Juniperus* type increases, potentially due to a reduction in burning, and there is a small expansion of various herbaceous taxa, including *Filipendula* and Poaceae, which could quickly colonise bare ground on the wetland margin along with *Artemisia*. An increase in the pollen influx of *Equisetum, Sphagnum* and *Typha latifolia* in this zone represents the growth of these species in small pools or damp areas of this undulating landscape.

The end of the stadial is indicated by a rise in *Betula* from *12,010–11,640 cal. BP* (*95% probability, LB2/3,* Figure 8), and a decline in *Juniperus* type and Cyperaceae (LB-3). A transition from *Betula*-dominated to *Pinus/Betula* woodland occurs during the very early

Holocene at *11,810–11,150 cal. BP* (*95% probability, LB3/4,* Figure 8)*,* and represented a period of drier bog surface conditions as *Sphagnum* declines (LB-4). At this time, woodland began to expand in the local area, indicated by an increase in the total tree pollen sum and tree pollen influx data (Figure 13). Although *Pinus* appears dominant, the influx data (Figure 9) indicates *Betula* did not decline as rapidly as the percentage data implies. This mixed *Pinus/Betula* woodland lasts until *10,440–10,230 cal.* BP (*95% probability, LB4/5,* Figure 8) and was dense, due to a lack of an herbaceous understory, with woodland also present on drier parts of the bog itself. A small increase in *Corylus* across LB-4 indicates continued warming of the climate in the early Holocene (Birks, 1989) as *Corylus* prefers warmer conditions, extended growth seasons and a less variable climate (Tallantire, 2002). However, the very low pollen influx values indicate these are either scattered trees rather than *Corylus* woodland or indicate long-distance transport of *Corylus* pollen.

Figure 8. Probability distributions of key events from Langshot Bog (derived from the age-depth model described in Figure 5) and their relationship to key climatic events.

As temperatures warm during the early Holocene (Davis et al., 2003), thermophilous species are observed in abundance in LB-5 from *10,440–10,230 cal. BP* (*95% probability, LB4/5,* Figure 8)*. Corylus* rapidly expands at this point, along with the arrival of *Ulmus*, potentially indicating a reduction in precipitation (Hall, 1978). A rapid decline in the influx of *Pinus* pollen indicates an inability to compete with newly arriving species (Birks, 1989). An increase in microscopic charcoal could be evidence for human induced woodland clearance, potentially for attracting animals to the water source, although drier conditions would also have led to an increased chance of natural burning events. A sustained increase in *Calluna vulgaris* suggests small heathland areas may have been present in the landscape. Although present in small amounts in LB-5, *Quercus* expands slightly later than the other thermophilous taxa, at the start of LB-6, from *9540–9480 cal. BP* (95% probability, wm95; Figure 8). Additionally during LB-6, *Corylus* started to form the understorey vegetation (Godwin, 1975) as the canopy of the deciduous woodland grew denser. An increase in humification, aligned with an increase in *Substantia humosa*, indicates a drying of the bog surface, supported by enhanced $\delta^{18}O$ values representing increased evaporation. An increase in Poaceae as well as the presence of other herbaceous (*Plantago media/major*) and shrub (Ericaceae, *Calluna vulgaris*) species suggests the continued development of clearings within the woodland, although it is not known whether these clearings were permanent or intermittent in nature. Burning events increased and could be due to either anthropogenic or natural causes, although the increase in Poaceae and relative stability of many arboreal species may indicate selective burning of woodland species to create cleared areas, likely helped by, or focused around the fall of senescent trees. The mixed open deciduous woodland remains established until *8430–8350 cal. BP* (*95% probability, wm65,* Figure 8). The subsequent hiatus means timings past this point are uncertain and zones LB-7, 8, and 9 are not discussed further.

Figure 9. Langshot Bog, Chobham Common: selected pollen taxa influx.

3.3. Elstead Bog B: Lithostratigraphy and chronology

Two transects helped understand the morphological characteristics of the basin at Elstead Bog B and a 221cm core was extracted at their crossover point (51° 10' 18.3"N 00° 43' 00.6"W); which is the deepest part of the basin (Figure 10). The core consists of 32 separate lithological units (Figure 11). Three radiocarbon determinations were obtained from the Elstead Bog B core (Figure 11) with the samples taken from locations in the core where LOI indicated high organic matter content (>10%), however, no age-depth model could be created for the sequence (Simmonds, 2017, 261–262). An *8410–8915 year* interval (*95% probability*) between the dates at 168–169cm and 138–139cm suggests the existence of a hiatus in the sequence, tentatively placed at 158cm due to a large reduction in organic matter and corresponding lithological shift at this depth. The age of the sample at 168–169cm, 11,820–11,400 cal. BP (WM-168, 2σ), indicates that there are early Holocene, and potentially Lateglacial, deposits surviving. The presence of the hiatus at 158cm means discussion will be focused on material below this level. Due to the low organic matter content of the material at the base of the sequence no further samples were submitted for radiocarbon dating.

Figure 10. Stratigraphic results from the two transects at Elstead Bog B.

Table 3. Radiocarbon dates from Elstead Bog B. Replicate measurements have been tested for statistical consistency and combined by taking a weighted mean before calibration as described by Ward & Wilson (1978; T'(5%)=3.8, v=1), except for those in bold that are not consistent at the 5% level.

The basal lithostratigraphic units (1–4) of fine and coarse sands indicate the bedrock geology of the Lower Greensand (Gallois, 1965). Organic matter and organic lake sediment (units 5 and 7) are interspersed with sand (units 6, 8, and 9), potentially at a time when there was a slight destabilisation of the slopes on the edges of the bog, either through increased storminess or changes in vegetation cover (Orme et al., 2016). A period of lake sedimentation is observed (units 10–13) and continued until the transition from the Lateglacial to the early Holocene. High values of *Potamogeton* subgenus *Potamogeton* type at this time suggest a small pool of still or slow flowing water (Fitter, 1987). A gradual shift from *Potamogeton* subgenus *Potamogeton* type to *Myriophyllum alterniflorium* suggests a slight shallowing and infilling of this pool towards the onset of the Holocene (Fitter, 1987). The beginning of the Holocene (units 14–18) is characterised by an increase in herbaceous peat growth. *Sparganium* would have grown in the shallow water margins of the water body (Cook, 1961), and its presence indicates a reduction in the size of the pool and the gradual development of peat encroaching from the margins.

Figure 11. Elstead Bog B: stratigraphy, organic matter, and radiocarbon dates.

3.4. Elstead Bog B: Pollen Stratigraphy

The Lateglacial vegetation record from Elstead Bog B (EBB-1) indicates a cold, open landscape (Figure 12). *Salix* and *Alnus* would have been present on moist to wet ground with *Pinus* and *Betula* present across a range of soil types. As suggested at Langshot Bog, *Salix* and *Betula* may have been dwarf species. Ericaceae would have been present on moist slopes, present here due to its ability to regenerate after fire events. Herbaceous taxa *Ranunculus* type, Poaceae and *Artemisia* would have been scattered across disturbed parts of the landscape (Clapham et al., 1987; Averis, 2013). During EBB-2, high values of *Potamogeton* subgenus *Potamogeton* type suggest a body of still or slow flowing water (Fitter, 1987) with Cyperaceae on the edges of this water body (Chapin and Chapin, 1980). The dry-ground flora remains relatively consistent with the most significant development being the arrival of *Juniperus* type, suggesting lower year-round temperatures. The radiocarbon chronology suggests that EBB-1 and EBB-2 may be correlated to the onset of the Loch Lomond Stadial, with both a lowering of temperatures and increased precipitation (Lowe and Walker, 2015). In EBB-3, the relatively high *Juniperus* type values suggest that the climate was still cold (Rawat and Everson, 2012), and the landscape continued to be relatively species poor. *Calluna vulgaris* represents a scrubby heathland element and has previously been identified in Late Devensian Lateglacial records (Godwin, 1975). High levels of Poaceae, Cyperaceae and *Rumex* undiff. suggest that disturbed open ground is present (Pennington et al., 1977), with a shift from *Potamogeton* subgenus *Potamogeton* type to *Myriophyllum alterniflorium* indicative of shallowing of the water body because of warming at the end of the Lateglacial.

A rapid increase in *Betula* is observed at the transition from the Lateglacial to the Holocene at 11,820–11,400 cal. BP (WM-168; 2σ) at Elstead Bog B (Figure 12). As the climate began to ameliorate pioneering taxa such as *Betula* and *Pinus* would have formed woodland (Bennett, 1983; Birks, 1989). *Salix* would have been present in wetter areas (Averis, 2013), and *Equisetum* and *Sparganium erectum* would also have grown in damp shallow water margins of the lake (Cook, 1961; Clapham et al., 1987). Their presence indicates a reduction in the size of the water body and the gradual encroachment of peat. The decline in Poaceae, Cyperaceae and other herbaceous taxa suggests that the woodland was closed with little or no understorey (Godwin, 1975).

Figure 12. Elstead Bog B: Pollen percentage diagram.

4. Discussion

Contextualising these two new palaeoenvironmental records in their contemporary timescape (Figure 13) helps develop our understanding of sedimentary history, vegetation succession, climatic change and human activity during the Lateglacial and early Holocene in both SE England and more widely across NW Europe.

Both Langshot Bog and Elstead Bog B formed as small pools within hollows in the naturally undulating sandy Bracklesham Beds and Lower Greensand, respectively. A lack of ramparts suggest these sites were not formed from the collapse of cryogenic mounds (pingos), as thought to be the case at Elstead Bog (Carpenter and Woodcock, 1981). Unusually, the radiocarbon dated sedimentary records indicate that the onset of organic sedimentation at both sites occurred during the Late Devensian Lateglacial and can be chronologically correlated with the Loch Lomond Stadial. This indicates that there was sufficient vegetation and low levels of landscape erosion permitting organic sedimentation. Importantly, this is in contrast to records from the north and west of Britain where sediments are often highly mineral rich, with no organic component, such as at Mill House 1, The Flasks 69 (Innes et al., 2009), Whitrig Bog (Mayle et al., 1997), Llyn Gwernan (Lowe and Lowe, 1989), Llanilid (Walker et al., 2003) and Sluggan Bog (Lowe et al., 2004). At other sites in SE England, organic sedimentation started prior to the Loch Lomond Stadial, during the Windemere Interstadial, including those at Holywell Coombe (Preece and Bridgland, 1999). It is unclear why earlier sediments are not present at Langshot Bog and Elstead Bog B, but it may be due to the survey and sampling strategy, drier climatic conditions preventing sediment accumulation, the absence of impeded drainage (forming a ‘closed’ basin) until later in the Lateglacial or sediment erosion by fluvial or aeolian processes. These processes could have been exacerbated by climatic fluctuations in the Lateglacial Interstadial, as observed at Holywell Coombe (Preece and Bridgland, 1999). Organic sediment deposition during the Stadial at Langshot Bog and Elstead Bog B probably occurred due to a combination of factors, reflecting both regional climatic and geological influences. A warmer climate during the Loch Lomond Stadial in SE England is indicated by molluscan records from Holywell Coombe (Kent) with mean summer temperatures of ~16°C (Rousseau et al., 1998) compared to mean summer temperatures of ~10-11°C at Llanilid (Wales) (Walker et al., 2003), 6–7.5°C at Hawes Water (northern England) (Marshall et al., 2002) and ~7.5-9°C at Whitrig Bog (Scotland) (Brooks and Birks, 2000), based upon coleopteran and chironomid data. Warmer southern temperatures may have meant a muted response to the reversal to colder climatic conditions, which would mean that erosive processes would not take hold as quickly or as significantly as it did in other areas, where species-poor slopes brought unweathered minerogenic material into the basins (Walker et al., 2003).

Organic sedimentation during the Loch Lomond Stadial permits a reconstruction of the vegetation history during this period, which is unusual in SE England. Significantly, similar successional trends between Elstead Bog B and Langshot Bog, sites over 20km apart, suggest these basins provide a coherent regional signal of vegetation history. The two records indicate that herbaceous taxa (Poaceae, *Artemisia,* Caryophyllaceae) would have been present across much of the open ground with Cyperaceae, *Sphagnum*, *Equisetum* and *Typha latifolia* indicating reed and sedge swamp and small pools at Langshot Bog, whilst Elstead Bog B would have been a small open water body during the Stadial. This herbaceous dominated landscape is consistent with that observed from river valley and coastal wetlands in SE England (Figure 13). There is also evidence for the development of patchy heathland, with Ericaceae, *Calluna vulgaris* and *Empetrum nigrum* ssp*. nigrum* type present. Dwarf-shrub heath is found in both low and high arctic vegetation communities (Razzhavin, 2012) and may offer a modern analogue for this landscape. Such a dwarf-shrub heathland community is also present locally at Bagshot (Groves, 2008) and Gatcombe Withy Bed (Scaife, 1987), wider afield in Britain at Llanilid (Walker et al., 2003), The Flasks 69 (Innes et al., 2009) and is recorded in other north-western Europe pollen diagrams (Verbruggen, 1979; Huntley and Birks, 1983; Bohncke et al., 1988; Walker et al., 1994). The presence of *Empetrum nigrum* ssp*. nigrum* type indicates a cool and cold climate as it (in addition to *Calluna vulgaris* and many Ericaceae species) tolerates windy exposed habitats and cold climatic conditions (Averis, 2013). Arboreal taxa primarily consisted of *Betula* (potentially the dwarf species *Betula nana*) with smaller amounts of *Pinus,* alongside *Alnus* and *Corylus*. The presence of *Pinus*, *Alnus* and *Corylus* warrant further examination, as their presence during this period is not universally observed at sites in NW Europe. These species may be present in the landscape, or may appear in the record due to contamination, long-distance transport, or reworking of older material (Cundill and Whittington, 1983; Edwards and Whittington, 2010). In this study, contamination is unlikely due to the use of a Russian corer, with material for analysis obtained from the centre of the core. Analysis of a pollen trap in the laboratory also indicated that there was no contamination by atmospheric pollen. Reworking of pollen also seems unlikely, as a limited range of unusual taxa are present, with reworking likely to lead to a broader range of anomalous pollen. There were also no obviously deteriorated pollen grains of specific taxa from either site, and these species have been identified at other sites in a regional setting. Studies of *Pinus* in Lateglacial contexts in Britain and Scandinavia have highlighted its susceptibility to long-distance pollen transportation (Day, 1996; Barnekow, 1999; Kullman, 2002; Birks et al., 2005) and without the presence of *Pinus* macrofossils, the exact relationship between long-distance transportation and local growth cannot be defined (Birks and Birks, 2000). However, studies have shown that percentage pollen counts of 5% (Bennett, 1995; McGeever and Mitchell, 2016) can to indicate local growth, and in some cases this can be as low as 0.4% (Froyd, 2005). With percentage values around 5% during this period at both

Elstead Bog B and Langshot Bog, and its identification in Lateglacial deposits at a range of other SE England sites (Figures 13–14) it is likely that there would have been small levels of local growth in sheltered areas, alongside some level of long-distance transportation. The distinct size differentiation in the *Pinus* grains observed at Langshot Bog is thought to relate to natural variations within the pollen morphology. The species is likely to be *Pinus sylvestris,* as it is the only species of *Pinus* identified in Pleistocene/Holocene Britain (Bennett, 1984), and has a wide variation in total grain size (Desprat et al., 2015). The presence of *Alnus* at Elstead Bog B highlights its importance in this vegetation community and may have been present at the wetland/dryland interface (The Museum of London, 2000; Figures 13–14). *Alnus* is now becoming a more established component of the Lateglacial vegetation cover at sites across SE England dating to both the Windermere Interstadial and the Loch Lomond Stadial (Figure 14), including at Minchery Farm and Spartum Fen (Parker and Preston, 2014), West Silvertown (Wilkinson et al., 2000), Bramcote Green (Thomas et al., 1996) and Ockley Bog (Simmonds, 2017). Macrofossil records of *Alnus* have also been identified across Britain from as early as 14,700-14,000 cal. BP at Turker Beck, north-east England (Young et al., In Press), indicating that the presence of *Alnus* at these sites is not solely explained by long-distance transport of *Alnus* pollen. The presence of *Alnus* in SE England suggests that microclimates would have been hospitable enough to allow for its survival. This is also indicated by *Corylus,* present during the Lateglacial at Langshot Bog (>30% probability at c. *12,800 cal. BP*: Figure 14) and at the very end of the Loch Lomond Stadial at Elstead Bog B (>90% from c. *12,000 cal. BP*). *Corylus* has also been identified during the Loch Lomond Stadial at Minchery Farm and Spartum Fen (Parker and Preston, 2014) and Thursley Bog (Simmonds, 2017). The presence of *Corylus* is significant because although pollen values are relatively low (Figure 14), it may indicate that small stands were able to survive the climatic conditions of the stadial in SE England, and could represent a refugia location for the species in NW Europe. However, it presence during this period has also been attributed to long-distance transport (Birks and Mathewes, 1978), and without the identification of *Corylus* macrofossils in SE England, this could also be applicable here.

Figure 13. Vegetation synthesis for the Lateglacial and early Holocene period (13,000-8000 cal. BP) in SE England. Numbers refer to sites as follows:

LB–Langshot Bog; EBB–Elstead Bog B*;* ***1–Ockley Bog*** *(Simmonds, 2017);* ***2–Thursley Bog*** *(Simmonds, 2017);* ***3–Bagshot*** *(Groves, 2008);* ***4–Elstead Bog A*** *(Farr, 2008);* ***5–Bramcote Green*** *(Thomas et al., 1996; Branch and Green, 2004);* ***6–Nutfield Marsh*** *(Farr, 2008);* ***7–Runnymede Bridge*** *(Scaife, 2000a);* ***8–Farm Bog*** *(Jennings and Smythe, 2000);* ***9****–**Holywell Coombe*** *(Preece and Bridgland, 1999);* ***10–West Silvertown*** *(Branch and Lowe, 1994; Wilkinson et al., 2000);* ***11–Colnbrook*** *(Gibbard and Hall, 1982);* ***12–New Ford Road*** *(Corcoran et al., 2011);* ***13–Meridian Point*** *(Corcoran et al., 2011);* ***14–Spartum Fen*** *(Parker and Preston, 2014);* ***15–Northmoor*** *(Shotton et al., 1970);* ***16–Standlake*** *(Sandford, 1965);* ***17–Queenborough*** *(Pratt et al., 2003);* ***18–Munsley Peat bed*** *(Scaife, 1980, 1982);* ***19–Intertidal peats in Hampshire*** *(Long et al., 2000);* ***20–The Isle of Wight*** *(Scaife, 1987);* ***21–Bouldnor Cliff*** *(Momber et al., 2011);* ***22–Borthwood Farm*** *(Scaife, 1987);* ***23–Yarmouth Marsh*** *(Scaife, Unpublished-b);* ***24–Panel Bridge*** *(Waller, 1993);* ***25–Enfield Lock*** *(Chambers et al., 1996);* ***26–Cothill*** *(Day, 1991); 27**–Farlington Marsh*** *(Scaife, 2000b);* ***28–The Lower Brede and Tillingham*** *(Waller and Long, 2010);* ***29–West Quay Road*** *(Godwin, 1940; Nicholls and Scaife, 2008);* ***30–George V Docks*** *(Godwin, 1940);* ***31–Portsmouth Harbour*** *(Godwin, 1945; Nicholls and Scaife, 2008);* ***32–London Cable Car*** *(Batchelor et al., 2015);* ***33–Fleet and Rownhams*** *(Scaife, Unpublished-a);* ***34–New Pond*** *(Scaife, 2001);* ***35–Newbury Sewage Works*** *(Carter, 2001; Chisham, 2004);* ***36–Woolhampton*** *(Carter, 2001; Chisham, 2004);* ***37–Thatcham Reedbeds*** *(Carter, 2001; Chisham, 2004);* ***38–Minchery Farm*** *(Parker, 2000);* ***39–Cranes Moor*** *(Barber and Clark, 1987);* ***40–Church Moor*** *(Barber and Clark, 1987);* ***41–Warwick Slade Bog*** *(Barber and Clark, 1987);* ***42–Wateringbury*** *(Kerney et al., 1980; Garnett et al., 2004);* ***43–Stonehenge*** *(Scaife, 1995);* ***44–Durrington*** *(French et al., 2012);* ***45–Chatham Dockyard and the Medway Tunnel*** *(Scaife, Unpublished-c);* ***46–Eton Rowing Lake*** *(Parker and Robinson, 2003);* ***47–Uxbridge*** *(Wessex Archaeology, 2006);* ***48–Riverside Way*** *(Wessex Archaeology, 2006);* ***49–Ferry Lane*** *(Scaife, 2004);* ***50–Three Ways Wharf*** *(Lewis, 1991; Lewis et al., 1992; Lewis and Rackham, 2011);* ***51–Lewes I and II*** *(Waller and Hamilton, 2000; Waller and Long, 2010);* ***52–Ufton Green*** *(Chisham, 2004).*

During the Loch Lomond Stadial, the presence of micro-charcoal has provided a record of regional burning at both Elstead Bog B pre 11,820–11,400 cal. BP (WM-168, 2σ) and Langshot Bog from *13,210–12,430 cal. BP* (*95% probability, Start of Sequence,* Figure 8), although a large decline is recorded at the start of the Loch Lomond Stadial from Langshot Bog at *12,940–12,420 cal. BP* (*95% probability, 236cm*; Figure 5). Burning is observed across NW Europe, in Scotland (Edwards et al., 2000) and Denmark (Kolstrup, 1992) where multi-causal models have been suggested for these fires, Including aridity, anthropogenic activity and taphonomic processes due to a lack of data highlighting a single cause. Across SE England during the Lateglacial; at Minchery Farm (Parker and Preston, 2014) burning was attributed to lightning strikes, and at Holywell Coombe it was similarly identified as the product of natural, local fires (Preece and Bridgland, 1999). In both cases the fires were attributed to natural causes due to a paucity of regional archaeological evidence, as is also thought to be the case at Elstead Bog B and Langshot Bog. Evidence for the natural occurrence of wildfires in tundra-style ecosystems has been observed from modern studies (Gowlett, 2016), and the decreasing frequency of fires at Langshot Bog during the Loch Lomond Stadial is attributed to declining summer temperatures (Wein, 1976; Racine et al., 1985), and may also reflect changes in fuel availability as woodland declined. These burning events are significant, because burning in modern tundra ecosystems has shown that post-fire soils can be 1–4°C warmer, with a deeper active layer than unburned soil (Rocha and Shaver, 2011), possibly creating improved growing conditions for warmth loving anomalous taxa, such as *Corylus* and *Alnus.*

Figure 14. Schematic diagram showing the presence of selected taxa (Pinus, Alnus and Corylus) and the timing of their first sustained presence in inland SE England. Pinus values of less than 5% are not shown due to the long-distance transport of Pinus pollen.

The onset of the Holocene at 11,700 cal. BP (Walker et al., 2012) is marked by a rapid increase in *Betula* woodland, and low levels of other arboreal taxa including *Juniperus* type, *Salix*, *Pinus, Corylus* and *Carpinus* type at Langshot Bog and Elstead Bog B. A decline in herbaceous and heathland taxa are thought to be related to the reduced amount of light penetrating a denser *Betula* dominated canopy. Climatic warming at the start of the Holocene was rapid, with a rise in mean annual temperatures of 1.7–2.8°C per 100 years (Lowe and Walker, 2015) and in SE England, mean summer temperatures could have increased by up to 7.5˚C in comparison to the Loch Lomond Stadial (Coope et al., 1998). The isotopic data from Langshot Bog indicates that after the initial early Holocene shift to warmer conditions, with lower levels of evaporation, there was a prolonged period of climatic instability lasting until *9790–9470 cal. BP* (*95% probability, 96cm,* Figure 5) as woodland expansion continued, and herbaceous taxa declined. Due to the truncated sequence at Elstead Bog B, similar trends to that observed at Langshot Bog, such as this woodland expansion, are only observed in the record from Elstead Bog (Farr, 2008; Figure 15). Comparable vegetation trends, with a decline in herbaceous taxa alongside the dominance of *Betula* and *Pinus* woodland are also observed at Panel Bridge (Waller, 1993), Holywell Coombe (Kerney et al., 1980; Preece and Bridgland, 1998), and Wateringbury (Kerney et al., 1980), as well as many of the sites on the Hampshire Coast (Figure 13).

Two potential climatic events were identified in the early Holocene from the $\delta^{18}O$ isotopic data at Langshot Bog. The *11,350-10,715 cal. BP* (*95% probability, 162cm,* Figure 5) period of decreased evaporation may relate to the Pre Boreal Oscillation (PBO) at 11,300–11,150 cal. BP (Björck et al., 1997) or ice rafted debris (IRD) event 8 (11,100 years BP) (Bond et al., 1997). Large-scale syntheses highlight a variable vegetation response to the PBO event around the Nordic Seas (Björck et al., 1997) and evidence for early Holocene IRD Events (Bond et al., 1997) are also variable both within SE England and around the Atlantic fringe of Europe (Wilkinson et al., 2000; Whittington et al., 2003; Garnett et al., 2004; Farr, 2008; Groves, 2008; Ghilardi and O'Connell, 2013), related to a paucity of (continuous) records, a lack of multi-proxy studies and poor sediment deposition rates. However, some evidence is observed in north-west Europe for either the PBO or the 11,100 years BP event from oxygen isotope values of Lough Inchiquin, Ireland (Diefendorf et al., 2006) and Crudale Meadow (Whittington et al., 2015), tufa at Wateringbury (Garnett et al., 2004) and from Hawes Water at both 11,400 cal. BP and 11,200 cal. BP (Lang et al., 2010). At Kingbeekdal in The Netherlands, cooling is identified between 11,400 and 11,300 cal. BP in both oxygen isotopes and the pollen record, with a decrease in *Betula* and subsequent opening of the woodland (Bohncke and Hoek, 2007). The second period of decreased evaporation, dated to *10,850-10,520 cal. BP* (*95% probability, 140cm,* Figure 5) does not appear to tie in with IRD Event 7 (10,300 years BP) (Bond et al. 1997) but lower temperatures (based on chironomid data) are observed from 10,400 cal. BP at Hawes Water (Lang et al., 2010). Cooler, moister and less seasonal conditions observed from lake records in the Faroe Islands at 10,600, 10,450 and 10,300 cal. BP (Hannon et al., 2003; Andresen et al., 2007) and an advance in glacier limits was also observed in Western Norway, dated to 10,235 cal. BP (Nesje et al., 2004). These studies all serve to highlight widespread climatic instability across a broad period. Identifying these events at Langshot Bog through the isotopic data is significant, as neither period of climatic instability appears clearly in either in the pollen record or δ^2H values, although general climatic instability could have assisted the widespread *Pinus* woodland development during this period. These results suggest that there may have been a climatic downturn at Langshot Bog around both the PBO and *10,850-10,520 cal. BP* (*95% probability, 140cm,* Figure 5), but the magnitude or duration of either event was not enough to trigger widespread vegetation change.

Figure 15. Elstead Bog: selected taxa percentage pollen diagram, after Farr (2008).

The sporadic occurrence of *Corylus,* during the Late Glacial and Early Holocene, alongside both coniferous (*Pinus*) and deciduous (*Betula*) taxa, may indicate the presence of brown earth soils, able to support this mixed woodland. This woodland continued at Langshot Bog until *10,440–10,230 cal. BP* (*95% probability, LB4/5,* Figure 8) and represents a pattern observed across SE England (Chambers et al., 1996; Preece and Bridgland, 1999; Scaife, 2001; Corcoran et al., 2011). From this time onwards*,* the records from Langshot Bog and Elstead Bog indicate woodland diversification: the colonisation of thermophilous species, specifically the rapid expansion of *Corylus*, followed by *Ulmus* and *Quercus* and a reduction in *Pinus* due to increased competition from deciduous trees. This pattern is recorded at inland and coastal wetlands in SE England including Wateringbury (Kerney et al., 1980), Holywell Coombe (Kerney et al., 1980; Preece and Bridgland, 1998), Lewes I and II

(Thorley, 1981), Panel Bridge (Waller, 1993), Brede Bridge (Waller and Marlow, 1994), Horsemarsh Sewer (Waller et al., 1999), Mount Caburn (Waller and Hamilton, 2000; Waller and Long, 2010), Queenborough (Pratt et al., 2003) and Tilling Green, Rye (Waller and Kirby, 2002). An increase in Poaceae, *Calluna vulgaris* and other herbaceous taxa implies clearings or openings in the woodland canopy in contrast to the dense *Pinus/Betula* woodland of the earliest Holocene.

The decline of *Pinus* and dominance of *Corylus* at Langshot Bog and across SE England warrants further discussion, because a number of studies have highlighted the importance of hazelnuts during the Mesolithic period both in Britain (Radley et al., 1974; Godwin, 1975) and in Europe (Holst, 2010). It has been suggested that Mesolithic hunter-gatherers may have contributed to this spread, by removing shading species to promote *Corylus* flowering and expansion (Mithen et al., 2001) or undertaking selective pruning to produce greater nut yield (Groß et al., 2018). A significant Mesolithic presence around the Elstead Bog sites is suggested from the archaeological record and therefore these human groups may have facilitated the expansion of *Corylus* around these sites. However, a lack of evidence for Mesolithic sites around the Langshot Bog site, and across the region more generally (Simmonds et al., 2019), together with a lack of burning and other anthropogenic indicators at the time of *Corylus* expansion at both Langshot Bog and Elstead Bog, suggests that the expansion is related to climatic amelioration and not human manipulation. This would include warmer conditions, extended growth seasons and a more stable climate (Tallantire, 2002), and therefore changes in vegetation composition and the formation of woodland clearings are unlikely to have been anthropogenic in origin. At Elstead Bog, there is no immediate decline in *Pinus* with the onset of deciduous woodland expansion, a pattern observed both at Conford (Groves et al., 2012) and in the Glynde Valley (Waller and Hamilton, 2000; Waller and Long, 2010) and attributed to the freely draining sandy geology. These substrates can often lead to low nutrient level soil, whereby invading species would have had less of an advantage over *Pinus*, allowing it to survive longer within the local woodland. Prolonged regular burning would have also influenced the *Pinus* record, as fires consume the humus soil layer, allowing *Pinus* to thrive in the bare mineral soil whilst other species struggle to survive (Richardson, 2000).

At Elstead Bog, a decline in several taxa is observed from 9290–9000 cal. BP (Wk-1422, 2σ) including *Betula*, *Corylus* and *Ulmus*, whilst *Pinus*, Poaceae and ferns increase. This indicates wetter conditions, possibly connected to either the 9400-year BP IRD Event (Event 6 - Bond et al., 1997), or the onset of the 9000-8000 cal. BP rapid climate change period (Mayewski et al., 2004). At this time, the isotopic record from Langshot Bog initially indicates a trend toward decreasing evaporation from *9540–9480 cal. BP* (95% probability, wm95) followed by a trend toward increased evaporation and wetter conditions from *9010-8520 cal. BP* (*95% probability, 76cm*, Figure 5) until at least *8430-8350 cal. BP* (95*% probability, wm65,* Figure 8). Again, palaeoenvironmental evidence for either climatic event from sites in SE England is sparse. However, across Britain, lower temperatures are observed at 9300 cal. BP from chironomid data at Hawes Water (Lang et al., 2010) and wetter conditions identified from Sphagnum records at 9200 cal. BP from Bolton Fell Moss (Barber et al., 2003). A climatic event following the 9400-year BP IRD event is also observed from a speleothem record in Ireland, where variability continues until 8800 cal. BP (McDermott et al., 2001). As with IRD Events 8 and 7, IRD Event 6 did not appear to result in identifiable alterations to the vegetation composition at Langshot Bog, and although some woodland decline was identified at Elstead Bog, it suggests these events were not definitively of sufficient duration or magnitude to result in expansive modification of the vegetation cover.

5. Conclusions

Palaeoenvironmental records from lowland, inland wetlands in SE England are rare, with most sites situated in riverine or coastal settings. There is also a paucity of sites containing organic deposits formed during the Loch Lomond Stadial. Langshot Bog and Elstead Bog B are therefore highly significant for furthering the understanding of vegetation dynamics during this period of climatic instability in Britain and NW Europe. The use of a multi-proxy approach, focusing on the Lateglacial and early Holocene, has led to the enhancement of previously identified biostratigraphic zones (Branch and Green, 2004; Groves, 2008; Groves et al., 2012) while providing greater detail and more robust timings for climatic events, vegetation change, species abundance and composition during this period. Significantly, there is evidence for *Pinus*, *Alnus* and *Corylus* as components of the Late Glacial vegetation around these sites. Although the presence of these species could be due to long-distance transport of pollen, it is not thought to fully account for the observed distributions, which are considered likely to have grown in sheltered parts of the landscape, perhaps offering insights into Glacial refugia locations. This is an area where the identification of plant macrofossils would assist greatly, as observed with *Alnus* at Turker Beck. Woodland development during the Holocene is characterised by the expansion of a dense *Betula* and *Pinus* forest. Mixed thermophilous woodland expansion is then observed, with a reduction in *Pinus*, apart from at sites with freely draining geologies or frequent burning events, where *Pinus* forms a component of the mixed woodland. The use of stable isotope records alongside pollen and sedimentological analysis has highlighted possible short-term climatic events, the first of which may chronologically correlate with the Pre-Boreal Oscillation, or the 11,100-year BP IRD event. A second, similar magnitude event seems too early for the 10,300-year BP IRD event, and may represent part of a period of climatic uncertainty, as represented by various studies in NW Europe which highlight climatic change across 10,800-10,200 cal. BP. A longer trend towards decreasing evaporation from *9540–9480 cal. BP* (95% probability, wm95) may indicate the 9400 year BP IRD event or early onset of the 9000-8000 year BP period of Rapid Climate Change, and although not represented clearly in the Langshot Bog pollen record, evidence is observed in the vegetation record at Elstead Bog for a decline in thermophilous taxa, potentially indicating a more severe or prolonged event than earlier periods of short-term climate change. These climatic shifts do not appear to be consistently observed in the vegetation record, and therefore the use of stable-isotope records for identifying these short-term shifts in climate is significant. Further studies should attempt to identify these short-term climatic shifts, which would allow for a greater understanding of the timing and impact of these early-Holocene events.

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