

# *New insights into late Devensian late glacial and early Holocene environmental change: two high-resolution case studies from SE England*

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

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10

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13 *isotopes;*

## 14 1. Introduction

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

29

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

41

42

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

45

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

57

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

60

### 61 **1.1. The sites: Langshot Bog and Elstead Bog B**

62 Langshot Bog (Figure 1) is located on the southern side of Chobham Common (51° 21'  
63 43.0"N 00° 35' 53.2"W; elevation: c. 30m a.s.l.) and is a 0.12km<sup>2</sup> area of superficial peat,  
64 situated on the Bracklesham Group. Chobham Common is a large expanse of heathland in  
65 Surrey and is the largest National Nature Reserve in SE England, managed by the Surrey  
66 Wildlife Trust. On Langshot Bog, species of *Sphagnum* occupy the bog surface (*Sphagnum*  
67 *compactum*, *S. papillosum*, *S. recurvum* and *S. palustre*) alongside other plants such as  
68 *Drosera rotundifolia*, *Narthecium ossifragum*, *Eriophorum angustifolium*, *Osmunda regalis*,  
69 and *Gentiana pneumonanthe*. Much of the bog surface is wooded, with *Betula pubescens*  
70 and *Alnus glutinosa* the dominant tree taxa. The modern-day vegetation across the Common

71 itself is dominated by *Calluna vulgaris*, *Erica cinerea*, and *Ulex minor* (plant nomenclature  
72 follows the flora of the British Isles Stace (2010)).

73  
74 *Figure 2. The location of Langshot Bog highlighting the coring transects and sampling location*  
75

76 Elstead Bog B (Figure 1) is a ~0.003km<sup>2</sup> peat basin located within pine woodland in Surrey  
77 (51° 10' 18.3"N 00° 43' 00.6"W; elevation: c. 52m a.s.l.), situated on gently downward  
78 sloping ground towards the Wey Valley to the north, with the ground rising steeply to the  
79 south. The site is nearby to the previously investigated Elstead Bog (Seagrief and Godwin,  
80 1960; Carpenter and Woodcock, 1981; Farr, 2008). The underlying geology is well sorted  
81 Folkestone Bed sands of the Lower Greensand (Carpenter and Woodcock, 1981). Soils are  
82 thin and podzolised, immediately resting on the Lower Greensand. Bog surface vegetation is  
83 dominated by *Molinia caerulea* and *Sphagnum* (*Sphagnum compactum*, *S. papillosum*, *S.*  
84 *recurvum*, and *S. palustre*), with *Betula pubescens* on stable areas. Around the bog margins,  
85 *Calluna vulgaris* and *Erica tetralix* prevail, with *Pinus sylvestris* in the wider woodland.

86

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

89

## 90 **2. Methods**

91 A stratigraphic coring survey was used to understand the size and shape of the two sites. At  
92 Langshot Bog, two transects, running east-west, and north-south, were sampled at 20m  
93 intervals, which resulted in 9 cores running west to east, and 13 cores running south to north  
94 (Figure 2). Cores were extracted until the corer hit sandy Bracklesham group sediments,  
95 identified within sediments in the base of each core, or observed whilst coring. At Elstead  
96 Bog B an east to west transect of 9 boreholes (5 metre spacing) and a north to south  
97 transect of 11 boreholes (10 metre spacing) cored down to the parent material, the Sandgate  
98 Formation (Figure 3). The deepest surveyed points were used as master coring locations. All

99 cores were collected using a Russian corer (or D-section corer), a hand coring device that  
100 cuts an undisturbed 500mm x 50mm section of core, a method widely used in  
101 palaeoenvironmental sampling because of the high quality of extracted sample (Moore et al.,  
102 1991) and the speed and ease of operation (Jowsey, 1966). At each site, two boreholes,  
103 situated no more than 30cm apart, allowed for each core section to have an overlap of  
104 10cm, and meant there was no disturbance of lower samples. Cores for further analysis  
105 were wrapped in rigid plastic downpipe and kept in cold storage (<4°C) at the University of  
106 Reading. Core stratigraphy was described by reference to the Troels-Smith classification  
107 scheme (Troels–Smith, 1955). Colour was noted from a Munsell Colour Chart and  
108 determination of organic content by loss-on-ignition (Bengtsson and Enell, 1986).

109

110 The alkali-soluble ('humic acid') and alkali- and acid-insoluble ('humin') fractions of bulk peat  
111 samples taken from above and below lithostratigraphic and biostratigraphic unit boundaries  
112 were dated by Accelerator Mass Spectrometry (AMS) at the Scottish Universities  
113 Environmental Research Centre (SUERC) and <sup>14</sup>CHRONO centre, Queen's University  
114 Belfast. The samples dated at Queen's University Belfast were pre-treated and measured  
115 following the methods described in Reimer et al. (2015). Despite six attempts trying both  
116 hydrogen reduction (four times), and zinc reduction (twice) the humin sample (UBA-26775)  
117 from Langshot Bog (137–138cm) failed to graphitise. Usually a sample fails to graphitise due  
118 to a higher than normal sulphur content and thus the laboratory increased the amount of  
119 silver present in the combustion tube but unfortunately this failed. Subsequent analysis of  
120 the sediment from this portion of the core identified very high concentrations of pyrite (FeS<sub>2</sub>).  
121 Pyrite releases sulphur oxides when combusted which can hamper the reduction of carbon  
122 dioxide to graphite (Proske et al., 2015).

123

124 The samples dated at SUERC were pre-treated using methods outlined in Stenhouse and  
125 Baxter (1981), combusted following Vandeputte et al. (1996), graphitised as described by  
126 Slota et al. (1987), and measured by AMS (Freeman et al., 2010). Internal quality assurance

127 procedures and international inter-comparisons (Scott, 2003; Scott et al., 2010) indicate no  
128 laboratory offsets and validate the measurement precision quoted. Details of the dated  
129 samples, radiocarbon ages, and associated stable isotopic measurements are provided in  
130 Tables 2-3. The radiocarbon results are conventional radiocarbon ages (Stuiver and Polach,  
131 1977). A weighted mean (Ward and Wilson, 1978) has been taken as providing the best  
132 estimate for the age of formation of the dated horizons (Tables 2–3) before inclusion in the  
133 age-depth model. The age-depth model (Figure 5) has been constructed using the program  
134 OxCal v4.2 (Bronk Ramsey, 2009; Bronk Ramsey and Lee, 2013) and the atmospheric  
135 calibration curve for the northern hemisphere published by Reimer et al. (2013). The  
136 P\_Sequence Poisson process model (Bronk Ramsey, 2008) employs a variable  $k$  parameter  
137 (Bronk Ramsey and Lee, 2013) with the overall age-depth model defined as P\_Sequence  
138 (1,1,U(-2,2)), with  $k_0$  (the base  $k$  parameter) =  $1\text{cm}^{-1}$ , the interpolation rate =  $1\text{cm}^{-1}$  (output  
139 from the model given every 1cm), and variability in  $k$  allowed between a factor of  $10^{-2}$  and  
140  $10^2$ .

141

142 Sub-samples for pollen analysis were derived from a  $1\text{cm}^3$  volumetric sampler (Bennett and  
143 Willis, 2002). At Langshot Bog and Elstead Bog B, sampling at an interval of 2cm provided a  
144 very high-resolution vegetation record in respect to core length. At Langshot Bog this  
145 equated to an average time-depth resolution of 50 years per sample ( $25\text{ years cm}^{-1}$ ).

146 Samples were prepared for pollen analysis with the addition of the exotic pollen marker  
147 *Lycopodium* (Stockmarr, 1971) and then underwent heavy liquid flotation (Branch et al.,  
148 2005). Acetolysis removed extraneous organic matter (Moore et al., 1991). A Leica DME  
149 microscope at  $\times 400$  and oil immersion  $\times 1000$  magnification was used for all identifications.  
150 Identification involved the use of the University of Reading reference collection and pollen  
151 keys (Moore et al., 1991; Reille, 1995), and pollen nomenclature follows Moore (et al. 1991).  
152 Pollen percentage values were calculated in Tilia version 1.7.16 (Grimm, 2011) based upon  
153 a pollen sum of  $\sim 300$  total land pollen (TLP). This count excluded aquatic species and  
154 spores; subsequently calculated as a percentage of their totals + TLP. Microscopic charred



155 particles were recorded quantitatively using a modified method from Robinson (1984) where  
156 MCP were counted relative to the trees, shrub and herb pollen count at the same time as  
157 pollen grains (Simmons and Innes, 1996; Innes et al., 2004; Innes et al., 2010). Different  
158 size classes for the charcoal were not implemented, due to the potential for fragmentation  
159 during the pollen preparation process, potentially artificially increasing charcoal counts in  
160 small size classes (Innes et al., 2004; Innes et al., 2010). Tilia.graph (Grimm, 2011) was  
161 used to draw pollen diagrams, and constrained incremental sum of squares clustering  
162 (CONISS) was used to help divide the percentage pollen diagram into a number of local  
163 pollen assemblage zones (Grimm, 1987). Pollen influx values were calculated for the  
164 Langshot Bog sequence, based upon the pollen concentration (grains cm<sup>-3</sup>) and deposition  
165 time (years cm<sup>-1</sup>). Values are expressed as a pollen deposition rate (grains cm<sup>-2</sup>year<sup>-1</sup>) and  
166 represent the number of grains incorporated into the unit area of sediment each year.

167

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

178

### 179 **3. Results**

#### 180 **3.1. Langshot Bog: Lithostratigraphy, chronology and stable isotopes**

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

186

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

188

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

---

<sup>1</sup> 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  $2\sigma$ ).

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

207

208

209 Peat formation continued in unit 3 until *12,010–11,640 cal. BP (95% probability, LB2/3,*  
210 *Figure 5)*. Shallow bog pools and wet peaty conditions were present at this point. The wet  
211 bog surface conditions indicated by the vegetation community correlate with the isotope  
212 records through this period (Figure 6), which indicate a marked shift to wetter conditions  
213 (depleted  $\delta^2\text{H}$ ) with increased evaporation (enhanced  $\delta^{18}\text{O}$ ). A small influx of sands and silts  
214 in units 2 and 3 are thought to represent in-washing from the higher elevated slopes around  
215 the edge of the basin, potentially caused by strong storm events, corresponding with wetter  
216 conditions or a loss of surface vegetation. An increase of trees on the surface of the bog  
217 (unit 4) from *12,010–11,640 cal. BP (95% probability, LB2/3, Figure 5)*, suggested from  
218 wood peat (unidentified) found within this unit, correlates with a distinctive shift in the  $\delta^{18}\text{O}$   
219 values indicating lower levels of evaporation. This may be a response to an early Holocene  
220 drying out of the surface of the bog. A period of relatively stable peat accumulation then  
221 continues from the onset of the Holocene until *9790–9470 cal. BP (95% probability, 96cm,*  
222 *Figure 5)*. During this time, localized changes in wetness are highlighted within the isotope  
223 data, which shows a period of instability, with overall trends indicating increased evaporation  
224 and slightly wetter conditions as this unit progresses. Two events within this period of  
225 instability stand out, with decreased evaporation at *11,350–10,715 cal. BP (95% probability,*  
226 *162cm, Figure 5)* and at *10,850–10,520 cal. BP (95% probability, 140cm, Figure 5)*, although  
227 any response from the vegetation ( $\delta^2\text{H}$  values) appears muted. An increase in humification  
228 in unit 5, aligned with an increase in *Substantia humosa*, indicates a further drying out of the  
229 bog surface from *9790–9470 cal. BP (95% probability, 96cm, Figure 5)* to *8430–8350 cal.*  
230 *BP (95% probability, wm65, Figure 5)*. The  $\delta^2\text{H}$  record identifies a broad drying trend,  
231 alongside a reduced amount of evaporation suggested by the  $\delta^{18}\text{O}$  profile. An increase in  
232 mineral influx and decline in organic matter is attributed to an increase in burning events

233 (Figure 6), causing increased runoff and in-washing of mineral matter into the basin. A hiatus  
234 in the sedimentary record is identified between units 5 and 7, thought to have occurred from  
235 8430–8350 cal. BP (95% probability, *wm65*, Figure 5) (unit 6), where there is a large influx of  
236 mineral material.

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

244

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

248

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

250

### 251 **3.2. Langshot Bog: Pollen Stratigraphy**

252 The Lateglacial at Langshot Bog (Figure 7: Local Pollen Zone (LB-1)) shows vegetation  
253 indicative of a cold, tundra-style landscape of shrubland and short turf grassland with  
254 Ericaceae and Cyperaceae. It was species-poor with natural burning events and  
255 characterised by a very sparse scrubby scatter of *Betula* and *Juniperus* type. These species  
256 are cold tolerant, which is important as this zone can be chronologically correlated with the  
257 start of the Loch Lomond Stadial (Walker et al., 1994), dating from 13,210–12,430 cal. BP  
258 (95% probability, 241cm, Figure 8) to 12,570–11,910 cal. BP (95% probability, LB1/2, Figure  
259 8). Ericaceae is present due to its ability to regenerate after fires (Averis, 2013), which based  
260 on the charcoal record would have been regular occurrences throughout this period. Regular  
261 fires have been observed at other sites in the UK and NW Europe during the last Glacial  
262 period (Kolstrup, 1992; Edwards et al., 2000) although understanding their onset has been  
263 difficult, a point explored further in the discussion. Ericaceae, and *Betula*, are resistant to  
264 wind exposure and are likely to indicate the presence of small heathland patches as

265 observed elsewhere in the UK (Bennett, 1983; Walker et al., 2003). *Juniperus* type colonises  
266 unstable ground surfaces and can survive in poor soils with low nutrient content (Averis,  
267 2013). The low percentages (predominantly <5%) suggest that dwarf *Juniperus* type was  
268 present, and alongside *Artemisia* indicate year-round temperatures would have been  
269 relatively cold (Kolstrup, 1980). *Pinus* might have been present on both drier and wetter soils  
270 but the low influx values (Figure 9) may also indicate long distance pollen transport.

271

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

273

274 LB-2 is a similar tundra-style vegetation mosaic, and the dating has enabled correlation with  
275 the mid-late Loch Lomond Stadial: 12,570–11,910 cal. BP (95% probability, LB1/2, Figure 8)  
276 to 12,010–11,640 cal. BP (95% probability, LB2/3, Figure 8). A consistent influx of *Salix*  
277 suggests it is underrepresented in the pollen percentage data and that it would have formed  
278 a component of the local woodland cover, around the wetland margins. The relatively open  
279 landscape present at this time suggests that *Salix* and *Betula* are likely to have been dwarf  
280 species, with both *Betula nana* and *Salix herbacea* identified in British Lateglacial contexts  
281 (Blackburn, 1952; Birks, 1965; Beerling, 1998). There was no attempt to do size frequency  
282 measurements on the *Betula* (Birks, 1968), as no distinct difference was observed and  
283 therefore, we cannot be definitive about the *Betula* species present during this time.

284 *Juniperus* type increases, potentially due to a reduction in burning, and there is a small  
285 expansion of various herbaceous taxa, including *Filipendula* and Poaceae, which could  
286 quickly colonise bare ground on the wetland margin along with *Artemisia*. An increase in the  
287 pollen influx of *Equisetum*, *Sphagnum* and *Typha latifolia* in this zone represents the growth  
288 of these species in small pools or damp areas of this undulating landscape.

289

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

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

305

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

308

309 As temperatures warm during the early Holocene (Davis et al., 2003), thermophilous species  
310 are observed in abundance in LB-5 from 10,440–10,230 cal. BP (95% probability, LB4/5,  
311 Figure 8). *Corylus* rapidly expands at this point, along with the arrival of *Ulmus*, potentially  
312 indicating a reduction in precipitation (Hall, 1978). A rapid decline in the influx of *Pinus* pollen  
313 indicates an inability to compete with newly arriving species (Birks, 1989). An increase in  
314 microscopic charcoal could be evidence for human induced woodland clearance, potentially  
315 for attracting animals to the water source, although drier conditions would also have led to  
316 an increased chance of natural burning events. A sustained increase in *Calluna vulgaris*  
317 suggests small heathland areas may have been present in the landscape. Although present  
318 in small amounts in LB-5, *Quercus* expands slightly later than the other thermophilous taxa,  
319 at the start of LB-6, from 9540–9480 cal. BP (95% probability, wm95; Figure 8). Additionally  
320 during LB-6, *Corylus* started to form the understorey vegetation (Godwin, 1975) as the

321 canopy of the deciduous woodland grew denser. An increase in humification, aligned with an  
322 increase in *Substantia humosa*, indicates a drying of the bog surface, supported by  
323 enhanced  $\delta^{18}\text{O}$  values representing increased evaporation. An increase in Poaceae as well  
324 as the presence of other herbaceous (*Plantago media/major*) and shrub (Ericaceae, *Calluna*  
325 *vulgaris*) species suggests the continued development of clearings within the woodland,  
326 although it is not known whether these clearings were permanent or intermittent in nature.  
327 Burning events increased and could be due to either anthropogenic or natural causes,  
328 although the increase in Poaceae and relative stability of many arboreal species may  
329 indicate selective burning of woodland species to create cleared areas, likely helped by, or  
330 focused around the fall of senescent trees. The mixed open deciduous woodland remains  
331 established until 8430–8350 cal. BP (95% probability, wm65, Figure 8). The subsequent  
332 hiatus means timings past this point are uncertain and zones LB-7, 8, and 9 are not  
333 discussed further.

334

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

336

### 337 **3.3. Elstead Bog B: Lithostratigraphy and chronology**

338 Two transects helped understand the morphological characteristics of the basin at Elstead  
339 Bog B and a 221cm core was extracted at their crossover point (51° 10' 18.3"N 00° 43'  
340 00.6"W); which is the deepest part of the basin (Figure 10). The core consists of 32 separate  
341 lithological units (Figure 11). Three radiocarbon determinations were obtained from the  
342 Elstead Bog B core (Figure 11) with the samples taken from locations in the core where LOI  
343 indicated high organic matter content (>10%), however, no age-depth model could be  
344 created for the sequence (Simmonds, 2017, 261–262). An 8410–8915 year interval (95%  
345 probability) between the dates at 168–169cm and 138–139cm suggests the existence of a  
346 hiatus in the sequence, tentatively placed at 158cm due to a large reduction in organic  
347 matter and corresponding lithological shift at this depth. The age of the sample at 168–  
348 169cm, 11,820–11,400 cal. BP (WM-168, 2 $\sigma$ ), indicates that there are early Holocene, and

349 potentially Lateglacial, deposits surviving. The presence of the hiatus at 158cm means  
350 discussion will be focused on material below this level. Due to the low organic matter content  
351 of the material at the base of the sequence no further samples were submitted for  
352 radiocarbon dating.

353

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

355

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

360

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

375

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

377



### 378 3.4. Elstead Bog B: Pollen Stratigraphy

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

400

401 A rapid increase in *Betula* is observed at the transition from the Lateglacial to the Holocene  
402 at 11,820–11,400 cal. BP (WM-168; 2 $\sigma$ ) at Elstead Bog B (Figure 12). As the climate began  
403 to ameliorate pioneering taxa such as *Betula* and *Pinus* would have formed woodland  
404 (Bennett, 1983; Birks, 1989). *Salix* would have been present in wetter areas (Averis, 2013),  
405 and *Equisetum* and *Sparganium erectum* would also have grown in damp shallow water

406 margins of the lake (Cook, 1961; Clapham et al., 1987). Their presence indicates a reduction  
407 in the size of the water body and the gradual encroachment of peat. The decline in Poaceae,  
408 Cyperaceae and other herbaceous taxa suggests that the woodland was closed with little or  
409 no understorey (Godwin, 1975).

410

411 *Figure 12. Elstead Bog B: Pollen percentage diagram.*

412

#### 413 **4. Discussion**

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

418

419 Both Langshot Bog and Elstead Bog B formed as small pools within hollows in the naturally  
420 undulating sandy Bracklesham Beds and Lower Greensand, respectively. A lack of ramparts  
421 suggest these sites were not formed from the collapse of cryogenic mounds (pingos), as  
422 thought to be the case at Elstead Bog (Carpenter and Woodcock, 1981). Unusually, the  
423 radiocarbon dated sedimentary records indicate that the onset of organic sedimentation at  
424 both sites occurred during the Late Devensian Lateglacial and can be chronologically  
425 correlated with the Loch Lomond Stadial. This indicates that there was sufficient vegetation  
426 and low levels of landscape erosion permitting organic sedimentation. Importantly, this is in  
427 contrast to records from the north and west of Britain where sediments are often highly  
428 mineral rich, with no organic component, such as at Mill House 1, The Flasks 69 (Innes et  
429 al., 2009), Whitrig Bog (Mayle et al., 1997), Llyn Gwernan (Lowe and Lowe, 1989), Llanilid  
430 (Walker et al., 2003) and Sluggan Bog (Lowe et al., 2004). At other sites in SE England,  
431 organic sedimentation started prior to the Loch Lomond Stadial, during the Windemere  
432 Interstadial, including those at Holywell Coombe (Preece and Bridgland, 1999). It is unclear  
433 why earlier sediments are not present at Langshot Bog and Elstead Bog B, but it may be due

434 to the survey and sampling strategy, drier climatic conditions preventing sediment  
435 accumulation, the absence of impeded drainage (forming a 'closed' basin) until later in the  
436 Lateglacial or sediment erosion by fluvial or aeolian processes. These processes could have  
437 been exacerbated by climatic fluctuations in the Lateglacial Interstadial, as observed at  
438 Holywell Coombe (Preece and Bridgland, 1999). Organic sediment deposition during the  
439 Stadial at Langshot Bog and Elstead Bog B probably occurred due to a combination of  
440 factors, reflecting both regional climatic and geological influences. A warmer climate during  
441 the Loch Lomond Stadial in SE England is indicated by molluscan records from Holywell  
442 Coombe (Kent) with mean summer temperatures of ~16°C (Rousseau et al., 1998)  
443 compared to mean summer temperatures of ~10-11°C at Llanilid (Wales) (Walker et al.,  
444 2003), 6–7.5°C at Hawes Water (northern England) (Marshall et al., 2002) and ~7.5-9°C at  
445 Whitrig Bog (Scotland) (Brooks and Birks, 2000), based upon coleopteran and chironomid  
446 data. Warmer southern temperatures may have meant a muted response to the reversal to  
447 colder climatic conditions, which would mean that erosive processes would not take hold as  
448 quickly or as significantly as it did in other areas, where species-poor slopes brought  
449 unweathered minerogenic material into the basins (Walker et al., 2003).

450

451 Organic sedimentation during the Loch Lomond Stadial permits a reconstruction of the  
452 vegetation history during this period, which is unusual in SE England. Significantly, similar  
453 successional trends between Elstead Bog B and Langshot Bog, sites over 20km apart,  
454 suggest these basins provide a coherent regional signal of vegetation history. The two  
455 records indicate that herbaceous taxa (Poaceae, *Artemisia*, Caryophyllaceae) would have  
456 been present across much of the open ground with Cyperaceae, *Sphagnum*, *Equisetum* and  
457 *Typha latifolia* indicating reed and sedge swamp and small pools at Langshot Bog, whilst  
458 Elstead Bog B would have been a small open water body during the Stadial. This  
459 herbaceous dominated landscape is consistent with that observed from river valley and  
460 coastal wetlands in SE England (Figure 13). There is also evidence for the development of  
461 patchy heathland, with Ericaceae, *Calluna vulgaris* and *Empetrum nigrum* ssp. *nigrum* type

462 present. Dwarf-shrub heath is found in both low and high arctic vegetation communities  
463 (Razzhavin, 2012) and may offer a modern analogue for this landscape. Such a dwarf-shrub  
464 heathland community is also present locally at Bagshot (Groves, 2008) and Gatcombe Withy  
465 Bed (Scaife, 1987), wider afield in Britain at Llanilid (Walker et al., 2003), The Flasks 69  
466 (Innes et al., 2009) and is recorded in other north-western Europe pollen diagrams  
467 (Verbruggen, 1979; Huntley and Birks, 1983; Bohncke et al., 1988; Walker et al., 1994). The  
468 presence of *Empetrum nigrum* ssp. *nigrum* type indicates a cool and cold climate as it (in  
469 addition to *Calluna vulgaris* and many Ericaceae species) tolerates windy exposed habitats  
470 and cold climatic conditions (Averis, 2013). Arboreal taxa primarily consisted of *Betula*  
471 (potentially the dwarf species *Betula nana*) with smaller amounts of *Pinus*, alongside *Alnus*  
472 and *Corylus*. The presence of *Pinus*, *Alnus* and *Corylus* warrant further examination, as their  
473 presence during this period is not universally observed at sites in NW Europe. These  
474 species may be present in the landscape, or may appear in the record due to contamination,  
475 long-distance transport, or reworking of older material (Cundill and Whittington, 1983;  
476 Edwards and Whittington, 2010). In this study, contamination is unlikely due to the use of a  
477 Russian corer, with material for analysis obtained from the centre of the core. Analysis of a  
478 pollen trap in the laboratory also indicated that there was no contamination by atmospheric  
479 pollen. Reworking of pollen also seems unlikely, as a limited range of unusual taxa are  
480 present, with reworking likely to lead to a broader range of anomalous pollen. There were  
481 also no obviously deteriorated pollen grains of specific taxa from either site, and these  
482 species have been identified at other sites in a regional setting. Studies of *Pinus* in  
483 Lateglacial contexts in Britain and Scandinavia have highlighted its susceptibility to long-  
484 distance pollen transportation (Day, 1996; Barnekow, 1999; Kullman, 2002; Birks et al.,  
485 2005) and without the presence of *Pinus* macrofossils, the exact relationship between long-  
486 distance transportation and local growth cannot be defined (Birks and Birks, 2000).  
487 However, studies have shown that percentage pollen counts of 5% (Bennett, 1995;  
488 McGeever and Mitchell, 2016) can indicate local growth, and in some cases this can be as  
489 low as 0.4% (Froyd, 2005). With percentage values around 5% during this period at both

490 Elstead Bog B and Langshot Bog, and its identification in Lateglacial deposits at a range of  
491 other SE England sites (Figures 13–14) it is likely that there would have been small levels of  
492 local growth in sheltered areas, alongside some level of long-distance transportation. The  
493 distinct size differentiation in the *Pinus* grains observed at Langshot Bog is thought to relate  
494 to natural variations within the pollen morphology. The species is likely to be *Pinus*  
495 *sylvestris*, as it is the only species of *Pinus* identified in Pleistocene/Holocene Britain  
496 (Bennett, 1984), and has a wide variation in total grain size (Desprat et al., 2015). The  
497 presence of *Alnus* at Elstead Bog B highlights its importance in this vegetation community  
498 and may have been present at the wetland/dryland interface (The Museum of London, 2000;  
499 Figures 13–14). *Alnus* is now becoming a more established component of the Lateglacial  
500 vegetation cover at sites across SE England dating to both the Windermere Interstadial and  
501 the Loch Lomond Stadial (Figure 14), including at Minchery Farm and Spartum Fen (Parker  
502 and Preston, 2014), West Silvertown (Wilkinson et al., 2000), Bramcote Green (Thomas et  
503 al., 1996) and Ockley Bog (Simmonds, 2017). Macrofossil records of *Alnus* have also been  
504 identified across Britain from as early as 14,700-14,000 cal. BP at Turker Beck, north-east  
505 England (Young et al., In Press), indicating that the presence of *Alnus* at these sites is not  
506 solely explained by long-distance transport of *Alnus* pollen. The presence of *Alnus* in SE  
507 England suggests that microclimates would have been hospitable enough to allow for its  
508 survival. This is also indicated by *Corylus*, present during the Lateglacial at Langshot Bog  
509 (>30% probability at c. 12,800 cal. BP: Figure 14) and at the very end of the Loch Lomond  
510 Stadial at Elstead Bog B (>90% from c. 12,000 cal. BP). *Corylus* has also been identified  
511 during the Loch Lomond Stadial at Minchery Farm and Spartum Fen (Parker and Preston,  
512 2014) and Thursley Bog (Simmonds, 2017). The presence of *Corylus* is significant because  
513 although pollen values are relatively low (Figure 14), it may indicate that small stands were  
514 able to survive the climatic conditions of the stadial in SE England, and could represent a  
515 refugia location for the species in NW Europe. However, its presence during this period has  
516 also been attributed to long-distance transport (Birks and Mathewes, 1978), and without the  
517 identification of *Corylus* macrofossils in SE England, this could also be applicable here.

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

521 **LB–Langshot Bog; EBB–Elstead Bog B; 1–Ockley Bog** (Simmonds, 2017); **2–Thursley**  
 522 **Bog** (Simmonds, 2017); **3–Bagshot** (Groves, 2008); **4–Elstead Bog A** (Farr, 2008); **5–**  
 523 **Bramcote Green** (Thomas et al., 1996; Branch and Green, 2004); **6–Nutfield Marsh** (Farr,  
 524 2008); **7–Runnymede Bridge** (Scaife, 2000a); **8–Farm Bog** (Jennings and Smythe, 2000);  
 525 **9–Holywell Coombe** (Preece and Bridgland, 1999); **10–West Silvertown** (Branch and  
 526 Lowe, 1994; Wilkinson et al., 2000); **11–Colnbrook** (Gibbard and Hall, 1982); **12–New Ford**  
 527 **Road** (Corcoran et al., 2011); **13–Meridian Point** (Corcoran et al., 2011); **14–Spartum Fen**  
 528 (Parker and Preston, 2014); **15–Northmoor** (Shotton et al., 1970); **16–Standlake** (Sandford,  
 529 1965); **17–Queenborough** (Pratt et al., 2003); **18–Munsley Peat bed** (Scaife, 1980, 1982);  
 530 **19–Intertidal peats in Hampshire** (Long et al., 2000); **20–The Isle of Wight** (Scaife, 1987);  
 531 **21–Bouldnor Cliff** (Momber et al., 2011); **22–Borthwood Farm** (Scaife, 1987); **23–**  
 532 **Yarmouth Marsh** (Scaife, Unpublished-b); **24–Panel Bridge** (Waller, 1993); **25–Enfield**  
 533 **Lock** (Chambers et al., 1996); **26–Cothill** (Day, 1991); **27–Farlington Marsh** (Scaife,  
 534 2000b); **28–The Lower Brede and Tillingham** (Waller and Long, 2010); **29–West Quay**  
 535 **Road** (Godwin, 1940; Nicholls and Scaife, 2008); **30–George V Docks** (Godwin, 1940); **31–**  
 536 **Portsmouth Harbour** (Godwin, 1945; Nicholls and Scaife, 2008); **32–London Cable Car**  
 537 (Batchelor et al., 2015); **33–Fleet and Rownhams** (Scaife, Unpublished-a); **34–New Pond**  
 538 (Scaife, 2001); **35–Newbury Sewage Works** (Carter, 2001; Chisham, 2004); **36–**  
 539 **Woolhampton** (Carter, 2001; Chisham, 2004); **37–Thatcham Reedbeds** (Carter, 2001;  
 540 Chisham, 2004); **38–Minchery Farm** (Parker, 2000); **39–Cranes Moor** (Barber and Clark,  
 541 1987); **40–Church Moor** (Barber and Clark, 1987); **41–Warwick Slade Bog** (Barber and  
 542 Clark, 1987); **42–Wateringbury** (Kerney et al., 1980; Garnett et al., 2004); **43–Stonehenge**  
 543 (Scaife, 1995); **44–Durrington** (French et al., 2012); **45–Chatham Dockyard and the**  
 544 **Medway Tunnel** (Scaife, Unpublished-c); **46–Eton Rowing Lake** (Parker and Robinson,  
 545 2003); **47–Uxbridge** (Wessex Archaeology, 2006); **48–Riverside Way** (Wessex  
 546 Archaeology, 2006); **49–Ferry Lane** (Scaife, 2004); **50–Three Ways Wharf** (Lewis, 1991;  
 547 Lewis et al., 1992; Lewis and Rackham, 2011); **51–Lewes I and II** (Waller and Hamilton,  
 548 2000; Waller and Long, 2010); **52–Ufton Green** (Chisham, 2004).

549  
 550

551 During the Loch Lomond Stadial, the presence of micro-charcoal has provided a record of  
 552 regional burning at both Elstead Bog B pre 11,820–11,400 cal. BP (WM-168, 2 $\sigma$ ) and  
 553 Langshot Bog from 13,210–12,430 cal. BP (95% probability, Start of Sequence, Figure 8),  
 554 although a large decline is recorded at the start of the Loch Lomond Stadial from Langshot  
 555 Bog at 12,940–12,420 cal. BP (95% probability, 236cm; Figure 5). Burning is observed  
 556 across NW Europe, in Scotland (Edwards et al., 2000) and Denmark (Kolstrup, 1992) where  
 557 multi-causal models have been suggested for these fires, including aridity, anthropogenic

558 activity and taphonomic processes due to a lack of data highlighting a single cause. Across  
559 SE England during the Lateglacial; at Minchery Farm (Parker and Preston, 2014) burning  
560 was attributed to lightning strikes, and at Holywell Coombe it was similarly identified as the  
561 product of natural, local fires (Preece and Bridgland, 1999). In both cases the fires were  
562 attributed to natural causes due to a paucity of regional archaeological evidence, as is also  
563 thought to be the case at Elstead Bog B and Langshot Bog. Evidence for the natural  
564 occurrence of wildfires in tundra-style ecosystems has been observed from modern studies  
565 (Gowlett, 2016), and the decreasing frequency of fires at Langshot Bog during the Loch  
566 Lomond Stadial is attributed to declining summer temperatures (Wein, 1976; Racine et al.,  
567 1985), and may also reflect changes in fuel availability as woodland declined. These burning  
568 events are significant, because burning in modern tundra ecosystems has shown that post-  
569 fire soils can be 1–4°C warmer, with a deeper active layer than unburned soil (Rocha and  
570 Shaver, 2011), possibly creating improved growing conditions for warmth loving anomalous  
571 taxa, such as *Corylus* and *Alnus*.

572

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

576

577 The onset of the Holocene at 11,700 cal. BP (Walker et al., 2012) is marked by a rapid  
578 increase in *Betula* woodland, and low levels of other arboreal taxa including *Juniperus* type,  
579 *Salix*, *Pinus*, *Corylus* and *Carpinus* type at Langshot Bog and Elstead Bog B. A decline in  
580 herbaceous and heathland taxa are thought to be related to the reduced amount of light  
581 penetrating a denser *Betula* dominated canopy. Climatic warming at the start of the  
582 Holocene was rapid, with a rise in mean annual temperatures of 1.7–2.8°C per 100 years  
583 (Lowe and Walker, 2015) and in SE England, mean summer temperatures could have  
584 increased by up to 7.5°C in comparison to the Loch Lomond Stadial (Coope et al., 1998).  
585 The isotopic data from Langshot Bog indicates that after the initial early Holocene shift to

586 warmer conditions, with lower levels of evaporation, there was a prolonged period of  
587 climatic instability lasting until 9790–9470 cal. BP (95% probability, 96cm, Figure 5) as  
588 woodland expansion continued, and herbaceous taxa declined. Due to the truncated  
589 sequence at Elstead Bog B, similar trends to that observed at Langshot Bog, such as this  
590 woodland expansion, are only observed in the record from Elstead Bog (Farr, 2008; Figure  
591 15). Comparable vegetation trends, with a decline in herbaceous taxa alongside the  
592 dominance of *Betula* and *Pinus* woodland are also observed at Panel Bridge (Waller, 1993),  
593 Holywell Coombe (Kerney et al., 1980; Preece and Bridgland, 1998), and Wateringbury  
594 (Kerney et al., 1980), as well as many of the sites on the Hampshire Coast (Figure 13).

595

596 Two potential climatic events were identified in the early Holocene from the  $\delta^{18}\text{O}$  isotopic  
597 data at Langshot Bog. The 11,350–10,715 cal. BP (95% probability, 162cm, Figure 5) period  
598 of decreased evaporation may relate to the Pre Boreal Oscillation (PBO) at 11,300–11,150  
599 cal. BP (Björck et al., 1997) or ice rafted debris (IRD) event 8 (11,100 years BP) (Bond et al.,  
600 1997). Large-scale syntheses highlight a variable vegetation response to the PBO event  
601 around the Nordic Seas (Björck et al., 1997) and evidence for early Holocene IRD Events  
602 (Bond et al., 1997) are also variable both within SE England and around the Atlantic fringe of  
603 Europe (Wilkinson et al., 2000; Whittington et al., 2003; Garnett et al., 2004; Farr, 2008;  
604 Groves, 2008; Ghilardi and O’Connell, 2013), related to a paucity of (continuous) records, a  
605 lack of multi-proxy studies and poor sediment deposition rates. However, some evidence is  
606 observed in north-west Europe for either the PBO or the 11,100 years BP event from oxygen  
607 isotope values of Lough Inchiquin, Ireland (Diefendorf et al., 2006) and Crudale Meadow  
608 (Whittington et al., 2015), tufa at Wateringbury (Garnett et al., 2004) and from Hawes Water  
609 at both 11,400 cal. BP and 11,200 cal. BP (Lang et al., 2010). At Kingbeekdal in The  
610 Netherlands, cooling is identified between 11,400 and 11,300 cal. BP in both oxygen  
611 isotopes and the pollen record, with a decrease in *Betula* and subsequent opening of the  
612 woodland (Bohncke and Hoek, 2007). The second period of decreased evaporation, dated



613 to 10,850-10,520 cal. BP (95% probability, 140cm, Figure 5) does not appear to tie in with  
614 IRD Event 7 (10,300 years BP) (Bond et al. 1997) but lower temperatures (based on  
615 chironomid data) are observed from 10,400 cal. BP at Hawes Water (Lang et al., 2010).  
616 Cooler, moister and less seasonal conditions observed from lake records in the Faroe  
617 Islands at 10,600, 10,450 and 10,300 cal. BP (Hannon et al., 2003; Andresen et al., 2007)  
618 and an advance in glacier limits was also observed in Western Norway, dated to 10,235 cal.  
619 BP (Nesje et al., 2004). These studies all serve to highlight widespread climatic instability  
620 across a broad period. Identifying these events at Langshot Bog through the isotopic data is  
621 significant, as neither period of climatic instability appears clearly in either in the pollen  
622 record or  $\delta^2\text{H}$  values, although general climatic instability could have assisted the  
623 widespread *Pinus* woodland development during this period. These results suggest that  
624 there may have been a climatic downturn at Langshot Bog around both the PBO and  
625 10,850-10,520 cal. BP (95% probability, 140cm, Figure 5), but the magnitude or duration of  
626 either event was not enough to trigger widespread vegetation change.

627

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

629

630 The sporadic occurrence of *Corylus*, during the Late Glacial and Early Holocene, alongside  
631 both coniferous (*Pinus*) and deciduous (*Betula*) taxa, may indicate the presence of brown  
632 earth soils, able to support this mixed woodland. This woodland continued at Langshot Bog  
633 until 10,440–10,230 cal. BP (95% probability, LB4/5, Figure 8) and represents a pattern  
634 observed across SE England (Chambers et al., 1996; Preece and Bridgland, 1999; Scaife,  
635 2001; Corcoran et al., 2011). From this time onwards, the records from Langshot Bog and  
636 Elstead Bog indicate woodland diversification: the colonisation of thermophilous species,  
637 specifically the rapid expansion of *Corylus*, followed by *Ulmus* and *Quercus* and a reduction  
638 in *Pinus* due to increased competition from deciduous trees. This pattern is recorded at  
639 inland and coastal wetlands in SE England including Wateringbury (Kerney et al., 1980),  
640 Holywell Coombe (Kerney et al., 1980; Preece and Bridgland, 1998), Lewes I and II

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

647

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

669 regular burning would have also influenced the *Pinus* record, as fires consume the humus  
670 soil layer, allowing *Pinus* to thrive in the bare mineral soil whilst other species struggle to  
671 survive (Richardson, 2000).

672

673 At Elstead Bog, a decline in several taxa is observed from 9290–9000 cal. BP (Wk-1422,  $2\sigma$ )  
674 including *Betula*, *Corylus* and *Ulmus*, whilst *Pinus*, Poaceae and ferns increase. This  
675 indicates wetter conditions, possibly connected to either the 9400-year BP IRD Event (Event  
676 6 - Bond et al., 1997), or the onset of the 9000-8000 cal. BP rapid climate change period  
677 (Mayewski et al., 2004). At this time, the isotopic record from Langshot Bog initially indicates  
678 a trend toward decreasing evaporation from 9540–9480 cal. BP (95% probability,  $wm95$ )  
679 followed by a trend toward increased evaporation and wetter conditions from 9010-8520 cal.  
680 BP (95% probability, 76cm, Figure 5) until at least 8430-8350 cal. BP (95% probability,  
681  $wm65$ , Figure 8). Again, palaeoenvironmental evidence for either climatic event from sites in  
682 SE England is sparse. However, across Britain, lower temperatures are observed at 9300  
683 cal. BP from chironomid data at Hawes Water (Lang et al., 2010) and wetter conditions  
684 identified from Sphagnum records at 9200 cal. BP from Bolton Fell Moss (Barber et al.,  
685 2003). A climatic event following the 9400-year BP IRD event is also observed from a  
686 speleothem record in Ireland, where variability continues until 8800 cal. BP (McDermott et  
687 al., 2001). As with IRD Events 8 and 7, IRD Event 6 did not appear to result in identifiable  
688 alterations to the vegetation composition at Langshot Bog, and although some woodland  
689 decline was identified at Elstead Bog, it suggests these events were not definitively of  
690 sufficient duration or magnitude to result in expansive modification of the vegetation cover.

691

## 692 **5. Conclusions**

693 Palaeoenvironmental records from lowland, inland wetlands in SE England are rare, with  
694 most sites situated in riverine or coastal settings. There is also a paucity of sites containing  
695 organic deposits formed during the Loch Lomond Stadial. Langshot Bog and Elstead Bog B  
696 are therefore highly significant for furthering the understanding of vegetation dynamics

697 during this period of climatic instability in Britain and NW Europe. The use of a multi-proxy  
698 approach, focusing on the Lateglacial and early Holocene, has led to the enhancement of  
699 previously identified biostratigraphic zones (Branch and Green, 2004; Groves, 2008; Groves  
700 et al., 2012) while providing greater detail and more robust timings for climatic events,  
701 vegetation change, species abundance and composition during this period. Significantly,  
702 there is evidence for *Pinus*, *Alnus* and *Corylus* as components of the Late Glacial vegetation  
703 around these sites. Although the presence of these species could be due to long-distance  
704 transport of pollen, it is not thought to fully account for the observed distributions, which are  
705 considered likely to have grown in sheltered parts of the landscape, perhaps offering insights  
706 into Glacial refugia locations. This is an area where the identification of plant macrofossils  
707 would assist greatly, as observed with *Alnus* at Turker Beck. Woodland development during  
708 the Holocene is characterised by the expansion of a dense *Betula* and *Pinus* forest. Mixed  
709 thermophilous woodland expansion is then observed, with a reduction in *Pinus*, apart from at  
710 sites with freely draining geologies or frequent burning events, where *Pinus* forms a  
711 component of the mixed woodland. The use of stable isotope records alongside pollen and  
712 sedimentological analysis has highlighted possible short-term climatic events, the first of  
713 which may chronologically correlate with the Pre-Boreal Oscillation, or the 11,100-year BP  
714 IRD event. A second, similar magnitude event seems too early for the 10,300-year BP IRD  
715 event, and may represent part of a period of climatic uncertainty, as represented by various  
716 studies in NW Europe which highlight climatic change across 10,800-10,200 cal. BP. A  
717 longer trend towards decreasing evaporation from 9540–9480 cal. BP (95% probability,  
718 wm95) may indicate the 9400 year BP IRD event or early onset of the 9000-8000 year BP  
719 period of Rapid Climate Change, and although not represented clearly in the Langshot Bog  
720 pollen record, evidence is observed in the vegetation record at Elstead Bog for a decline in  
721 thermophilous taxa, potentially indicating a more severe or prolonged event than earlier  
722 periods of short-term climate change. These climatic shifts do not appear to be consistently  
723 observed in the vegetation record, and therefore the use of stable-isotope records for  
724 identifying these short-term shifts in climate is significant. Further studies should attempt to

725 identify these short-term climatic shifts, which would allow for a greater understanding of the  
726 timing and impact of these early-Holocene events.

727

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738

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