Lateglacial and early Holocene climates of the Atlantic margins of Europe: stable isotope, mollusc and pollen records from Orkney, Scotland

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ABSTRACT

The margins of mainland Europe, and especially those areas coming under the influence of North Atlantic weather systems, are ideally placed to record changing palaeoclimates. Cores from an infilled lake basin at Crudale Meadow in Mainland, Orkney, revealed basal deposits of calcareous mud ('marl') beneath sedge peat. Stable isotope, palynological and molluscan analyses allowed the establishment of palaeoenvironmental changes through the Devensian Lateglacial and the early Holocene. The $\delta^{18}O_{marl}$ record exhibited the existence of possibly four climatic oscillations in the Lateglacial (one of which, within event cf. GI-1c, is not often commented upon), as well as the Preboreal Oscillation and other Holocene perturbations. The cold episodes succeeding the Preboreal Oscillation were demarcated conservatively and one of these (event C5, ~11.0 ka) may have previously been unremarked, while the putative 9.3 and 8.2 ka events seem not to produce corresponding palynologically visible floristic changes. The events at Crudale Meadow are consistent with those recorded at other sites from Britain, Ireland and elsewhere, and can be correlated with isotopic changes shown by the Greenland ice cores. The multi-proxy approach enriches the environmental reconstructions from the site, although the synchronicity of the response of the various proxies is sometimes equivocal, depending upon the time period concerned, taphonomy, and the nature of the deposits. The site may contain the most northerly Lateglacial isotope record from northwest Europe, and it has yielded one of the best archives for the demonstration of abrupt early Holocene events within Britain.

Keywords:

Palaeoclimates, isotopes, palynology, molluscs, Orkney, Britain, Ireland, Europe, North Atlantic, Greenland

1. Introduction

Stable isotopes from lake deposits have long been used for climate reconstructions (e.g. Leng and Marshall, 2004) and their use alongside complementary proxies such as pollen, chironomids and Mollusca have strengthened insights into the processes and patterns of global palaeoclimates (e.g. Eicher and Siegenthaler, 1976; Ammann et al., 1983; Böttger et al., 1998; van Asch et al., 2012). Lateglacial successions have been a particular focus because of their sedimentological suitability and the marked oscillatory nature of the climate records they contain (O'Connell et al., 1999; von Grafenstein et al., 2000; Jones et al., 2002; Lang et al., 2010b; van Raden et al., 2013). Holocene deposits have perhaps received less attention. This is partly a function of the availability of suitable material for geochemical analysis, and partly of contemporary research foci. Instances of Holocene or combined Lateglacial and Holocene investigation are to be found, however, and given current concerns which emphasise a need to comprehend climate change since glacial times, they are arguably increasingly valuable (cf. Whittington et al., 1996; Ahlberg et al., 2001; Garnett et al., 2004a; Diefendorf et al., 2006; Eastwood et al., 2007; Marshall et al., 2007; Daley et al., 2011).

The margins of Continental Europe, and especially those areas coming under the influence of North Atlantic weather systems, so important in driving climate variability, are ideally placed to reflect changing palaeoclimates. In this respect, Ireland and Britain are particularly well placed to produce multi-proxy records of environmental change, as attested by the suite of sites around Lough Gur (Ahlberg et al., 1996; O'Connell et al., 1999; Diefendorf et al., 2006; van Asch et al., 2012), Gransmoor (Walker et al., 1993), Clettnadal (Whittington et al., 2003), Wester Cartmore (Edwards and Whittington, 2010), and more widely across Europe (cf. Birks et al., 2000, 2012; Brooks and Langdon, 2014). This applies equally, of course, to single proxy studies (e.g. chironomids – Lang et al., 2010a; Brooks et al., 2012; Brooks and Langdon, 2014). In spite of an abundance of publications globally, relatively few palaeo-isotope studies have been carried out in Britain (e.g. Turney et al., 1997, 1998; Walker et al., 2003; Garnett et al., 2004a; Marshall et al., 2007; Daley et al., 2007; Candy et al., 2015), and Scotland, a key northerly location, has seen a single comprehensive isotope investigation (for Lundin Tower – Whittington et al., 1996) and another which produced outline details of $\delta^{13}C$ at three sites (Borrobol, Tynaspirit West, Whitrig Bog – Turney et al., 1997; Turney, 1999) (Fig. 1).

What has been lacking is a near-coastal research site in an oceanic context. Such a site might be anticipated to provide a sensitive record of environmental change, although competing site attributes and external climatic factors could always mute responses in various proxies at different times (cf. Whittington et al., 2003). A site in Orkney provides an opportunity to pursue the aims of multi-proxy climate and wider environmental enquiry for deposits of both Lateglacial and early Holocene age. This paper adds not just another comprehensive multi-proxy data set, including stable isotopes, to the few available from Britain and the rest of the Continental Atlantic margins of Europe, but it also presents evidence which reinforces environmental correlates with the Greenland ice core records and raises the issue of cold oscillations in both Lateglacial and early Holocene times which are little commented upon. Critically, as far as we are aware, the site provides the most northerly Lateglacial isotope record from northwest Europe, and one of the best instances for the demonstration of abrupt early Holocene events within Britain.

2. **The site**

The Orkney Islands archipelago (58°43-59°23’ N; 2°22-3°04’W) is found 16 km north of Caithness on the Scottish mainland and 78 km southwest of the Shetland Islands. The area is subject to the ameliorating effects of the North Atlantic Drift, but wind speeds and exposure are high – in many respects, its present-day weather and climatic characteristics are intermediate between those of mainland Scotland and Shetland (Berry, 2000). The site of Crudale Meadow is in the west of the island of Mainland, the largest island of the Orkney Islands (Fig. 1). It lies 1.7 km from the west coast and 6.1 km NNE of the town of Stromness in an area floored by sandstones and siltstones of the Yesnaby Sandstone Group of the Lower Old Red Sandstone (Mykura, 1976). Thin tills on the surrounding slopes derive partly from the carbonate-rich dolomitic siltstones, shales and sandstones of the Middle Old Red Sandstone Lower Stromness Flags. The glacial history of Orkney is not particularly well understood. Striae from the Late Devensian icesheet are reported in the hills to the west of Crudale Meadow (Wilson et al., 1935), and the whole of Orkney was apparently overridden by ice in the late Devensian. This ice sheet flowed across the islands from the North Sea in a southeast to northwest direction towards the Atlantic (Hall, 1996), and maximum ice extent seems to have occurred at around 18 ka BP. The origins of this ice are unclear – the islands may have been over-ridden by the westward advance of the Scandinavian ice sheet, or the presence of a smaller Scandinavian ice sheet in the North Sea might have deflected outflows from the Scottish ice sheet, causing them to cross Orkney in an east-west direction (Hall, 1996; Carr et al., 2006). The first high ground seems to have emerged from the ice around 15 ka BP, but there is no evidence of flow reversal (west to east) and extensive areas of hummocky moraine suggest active east-west flow until termination. Local glaciers subsequently formed during the Younger Dryas (Loch Lomond Stade) in the hills of Hoy (based on ^{10}Be exposure ages), with an equilibrium line altitude as low as 91 m (area-weighted mean of 141 m for two corries), although the rest of the islands remained ice free (Ballantyne et al., 2007).

The poorly drained, infilled lake basin at Crudale Meadow (~12 ha in extent and 9 m a.s.l) contains a valley mire with some open water areas, and drains eastwards via a minor stream into the Loch of Stenness. The mire was investigated by Moar (1969) who named it Yesnaby, and subsequently by Bunting (1994) who termed it Crudale Meadow. Based on the location reported by Moar, Yesnaby (National Grid Reference: HY 237152; 59°01’00.08” N, 3°19’48.35”W) lay some 0.1 km WNW from Crudale Meadow (HY 238151; 59°00’58.03” N, 3°19’43.20”W).

The surface vegetation of the mire (plant nomenclature follows Stace [2010]) is dominated by *Phragmites australis* and Cyperaceae spp., along with other fen taxa including *Menyanthes trifoliata, Ranunculus* spp., *Narthecium ossifragum, Hydrocotyle vulgaris, Potentilla palustris, Caltha palustris, Filipendula ulmaria* and *Pedicularis palustris*. Wet areas contain *Sphagnum* spp.

The adjacent slopes and dryland areas are in use for rough grazing and feature a heathland mosaic with *Calluna vulgaris, Empetrum nigrum, Erica cinerea*, Poaceae spp., *Angelica* sp., *Potentilla erecta*, *Plantago lanceolata*, *Succisa pratensis, Hypericum* sp., *Rumex* sp., *Cirsium sp.* and *Senecio* sp.

3. Methods

3.1. Fieldwork and core storage

Cores were obtained from the centre of the Crudale Meadow basin with a modified 5 cm diameter Livingstone piston corer (Wright, 1967) and led to the recovery of 5.76 m of deposits. The core reported here is a parallel one to that reported in Bunting (1994). Core sections were stored, wrapped in plastic sheeting over aluminium foil, at 4º C until sub-sampled for pollen, stable isotopes, loss-on-ignition (LOI) and mollusc analyses.

3.2. Lithostratigraphy

Generalised lithostratigraphy was assessed using the Troel-Smith (1955) scheme. The deposits were assessed for organic content by LOI (4h at 550 ºC) on samples taken at 2 cm intervals. Continuous profiles of uncalibrated volume magnetic susceptibility measurements were obtained with a Bartington Instruments magnetic susceptibility meter and core scanning loop sensor (Thompson and Oldfield, 1986). Although susceptibility can be related to allochthonous inputs of minerogenic material, it was used here to potentially assist in the location of tephra-rich strata.

3.3. Palynology

The core was sampled from 576-198 cm for pollen analysis at a maximum interval of 4 cm. The pollen sum obtained at each level to a depth of 460 cm was at least 500 identified land pollen (TLP) grains, but below that depth it sometimes only proved feasible to reach a total of 300 grains. Each grain was assessed as to its preservation status on a hierarchical index of perfect, crumpled/folded, broken, pitted/thinned. Pollen concentrations were made possible by the use of *Lycopodium* tablets added during the pollen preparation which followed the method of Faegri and Iversen (1989). Pollen type nomenclature followed Stace (2010), amended after Bennett et al. (1994) and Bennett (2015). The computer program TILIA v.1.7.16 (Grimm, 1991-2011) was used for the production of pollen diagrams. Pollen diagram zonation was aided by CONISS (Grimm, 1987).

3.4. Mollusc analysis

The core was cut into 5 cm sections which were soaked overnight and washed through a 250 μm sieve (this rather than the more usual 500 μm sieve was used in order to collect the most juvenile instars of ostracods; although ostracods were found in most samples, it did not prove possible to have them analysed). The residues were oven dried at 40 ºC and sorted under a x10-16 binocular microscope. Shells were counted using the convention of Sparks (1961) where every gastropod apex is recorded as an individual and bivalve umbo totals were halved to give a minimum number of individuals present. Taxonomic nomenclature followed Kerney (1999).

3.5. Isotopic analysis

Individual samples (1 cm intervals) of bulk carbonate were gently dry sieved at 100 μm to separate shells and ostracods from authigenic calcite (e.g. Leng et al., 2010) and then dried, powdered and treated for 4 hours in low temperature oxygen plasma to remove organic contaminants.

Carbonate isotopic compositions were determined on CO_2 released by overnight reaction with 100% H_3PO_4 at 25 °C using a VG SIRA 10 mass spectrometer calibrated via NBS 19 standard. XRD analyses were also performed on seven samples.

Mollusc shells, belonging to *Lymnaea peregra* (one of the two commonest taxa present), where available, were carefully cleaned in an ultrasonic bath and then dried and powdered. They were analysed by means of an automatic carbonate micro-treatment device (90 °C 500 s reaction time in H_3PO_4) attached to a VG PRISM 2 mass spectrometer. Usually one to three shells provided the sample.

Samples for the measurement of organic matter $\delta^{13}C$ (10 cm intervals) were treated with dilute HCl to remove carbonate and washed in distilled water to reach neutral pH and dried. The CO_2 from the organic matter was obtained by combusting the samples in an evacuated quartz tube at 850 °C with an excess of cupric oxide. NBS 22 gives $\delta^{13}C$ of -29.7‰ by this method.

3.6. Core chronology

^{14}C AMS assay was carried out on six samples, with pre-treatment (2M HCl, 80°C, 8 hours followed by quartz tube combustion of samples to CO_2) at the NERC Radiocarbon Facility in East Kilbride and the analysis of graphite targets at the University of Arizona NSF-AMS Laboratory.

The susceptibility profile assisted in the location of a cryptotephra layer centred at 384 cm depth (no additional tephras were sought; the site is under consideration for a forthcoming tephra-based project at Royal Holloway, University of London [Rhys Timms pers. comm]). The sedimentary matrix was subjected to H_2O_2 digestion and the resultant dry residues were mounted in a conductive phenolic resin ('bakelite') using a hot press. Stubs were polished and carbon-coated. Electron microprobe analysis was carried out in the Department of Earth Sciences at Cambridge

University using a CAMECA SX50 electron microscope fitted with three wave-dispersive spectrometers and a Link ANIOOOO energy-dispersive spectrometer with PAP matrix correction software. Analysis was carried out using an accelerating voltage of 20 kV, a beam strength of 10 nA and a beam diameter of 10μm with the spot slightly defocused. 20 shards were analysed and a mixture of minerals, natural oxides and pure metals were used as standards.

Where ice core data from NGRIP and other sites are used, the Greenland Ice Core Chronology 2005 (Rasmussen et al., 2006; Lowe et al., 2008; Walker et al., 2009; Blockley et al., 2012) is employed, with age estimates expressed as GICC05 age b2k, where b2k is years before the AD 2000 datum (Rasmussen et al., 2006).

4. Presentation of results

4.1. Lithostratigraphy

Details of the lithostratigraphy are presented in Table 1 and shown in Figure 2 (et seq.). The sedimentary succession falls into four distinct categories. From the base up to 520 cm there are bands of shelly calcareous mud (hereafter termed ‘marl’) with an interleaving of silts. From 520-500 cm a stratum of minerogenic material was found, with the lowest LOI values for the entire depositional sequence, above which, up to 299 cm, there is a further almost continuous accumulation of shelly marl. The sequence is completed up to the ground surface by sedge peat.

4.2. Palynology

Pollen and spore data are presented as percentages of TLP and as concentrations (Figs. 3 and 4; Supplementary Information, Figs. 1-3). The pollen diagram has been divided into nine local pollen assemblage zones (LPAZs; summarized in Table 2) identified as CRU- followed by a number, of which two are further subdivided into subzones designated by lower case letters. The pollen profile closely parallels those from Yesnaby (Moar, 1969) and, unsurprisingly, the parallel core from Crudale Meadow (Bunting, 1994), but is of higher resolution than either, which led to more taxon identifications (56, 68 and 92 respectively for approximately similar palynomorph counts) despite the shorter time period under investigation, and includes some preservation data. Given these earlier investigations, the interpretation here is shortened accordingly and considers the core up to 194 cm, close to the marl/peat boundary.

4.3. Mollusc analysis

Many small bivalves (<250 μm) were sorted from the samples. These were mostly identifiable only to genus and provide the bulk of the high totals for *Pisidium* spp. The low numbers of species at most levels in the sequence meant that the usual diagram of molluscan percentages by depth has not been drawn. Instead, total molluscan numbers for each level are shown in Figures 2 and 5.

4.4. Isotope analysis

The isotopic results (Figs 2, 6 and 7) are presented using the conventional δ‰ notation, with reference to the V-PDB standard. The isotopic differences between untreated and treated samples of bulk carbonate (Supplementary Information, Table 1) were sometimes greater than analytical reproducibility, thus justifying the removal of organic contaminants. Mean analytical reproducibility of duplicate analyses (n=39) on treated samples yielded ±0.08‰ and ±0.14‰ for carbon and oxygen isotope ratios respectively. The XRD analyses showed that calcite was the only carbonate mineral.

4.5. Chronology

Radiocarbon (^{14}C) data from Crudale Meadow are presented in Table 3 and tephra microprobe data appear as Supplementary Information, Fig. 4.

5. Discussion of the Crudale Meadow results

A robust absolute chronology is absent for Crudale Meadow, and this is discussed below (sections 5.5 and 6). This is not an unusual occurrence when dealing with calcareous deposits where various stratagems have been employed to overcome local inadequacies in dating (e.g. Ahlberg et al., 1996; Garnett et al., 2004b; van Asch et al. 2012; van Raden et al., 2013). It is very clear on litho-, bio-stratigraphic, and partially isotopic grounds, and in comparison with an extensive corpus of research at local through to sub-continental scales, that the core from Crudale Meadow encompass both Lateglacial and Holocene age deposits. These are most obviously demonstrated here via the palynological evidence (section 5.2), but in the other proxies also, even if to a lesser extent. In order to facilitate discussion, ‘classical’ nomenclature (e.g. Allerød, Younger Dryas) is employed in the first instance.

5.1. Lithostratigraphy

The sharp falls in LOI (from ~568 and 524 cm) correspond to inputs of minerogenic, and especially silty, deposits. This also accords with the magnetic susceptibility curve where high values are recorded in the basal silts and marls (maximum of 33.55 units at 518 cm). The susceptibility measurements within the overlying marl are much lower (typically around 0.1 units), other than at the level where a tephra peak was suspected (0.75 units at 384 cm). The inputs of minerogenic material are typical of erosional episodes and can be shown to correspond to pollen- and mollusc-inferred cold phases (see below).

5.2. Palynology

The basal zone (CRU-1) is dominated by Poaceae (up to 50% TLP) accompanied by *Betula* (24%), some of which is likely to have been *B. nana* (Whittington et al., 2003) and *Salix* (10%). The *Pinus sylvestris* pollen (up to 8%) is assumed to represent wind-transported grains from Scandinavia or southern Britain rather than redeposited elements or contamination (cf. Donner, 1957; Cundill and Whittington, 1983; Tzedakis et al. 2013). The relatively open nature of the landscape is emphasized by the abundance of pollen from such herb taxa as *Artemisia,* Asteroideae/Cardueae undiff., Lactuceae and *Rumex acetosa* and the expanding curve for *Empetrum nigrum*. The number of unidentifiable/unidentified pollen and spores is high, the numbers of pitted/thinned *Betula* and Poaceae grains increase through the zone, and the total fossil pollen concentration values are at their lowest for the whole profile. This zone may equate to at least part of a temperate climatic event (cf. the Bølling), moving into a cooling one (cf. the Older Dryas) towards the end of this Lateglacial zone.

During zone CRU-2, a major change occurs due to the rises in *Empetrum* (up to 58% TLP) and Cyperaceae (17%) and the accompanying decline in Poaceae. Among the minor taxa there is some decline in *Salix* and *Rumex*. In the middle of the zone, various taxa show falls in percentages (e.g. *Betula*, *Empetrum nigrum*) or increases (e.g. Cyperaceae, *Artemisia*) and there are complementary rises in pitted/thinned grains of *Betula* and Poaceae. Pollen concentrations rise slightly through the zone. CRU-2 would seem to denote warming, but with an intriguing cooling episode within it (see section 6). The LPAZ could represent the Allerød event.

Zone CRU-3 has initial expansions in a range of taxa such as *Betula*, Poaceae and Cyperaceae, followed by a reduction in all of them along with a marked fall in *Empetrum nigrum*.

The end of the zone sees a resurgence in *Betula* and *E. nigrum*, rising Lactuceae and *Salix* curves and a fall in Cyperaceae. There seem to be fluctuations in vegetation which are not well resolved (cf. Hoek, 2001), but which could include the presence of the Intra-Allerød Cold Period (Gerzenzee oscillation; Andresen et al., 2000, Yu and Eicher, 2001) and a warm amelioration within CRU-3.

Poaceae retains its numerically dominant position in zone CRU-4, but it declines throughout as does *Empetrum nigrum* from an initial peak, and expansions occur in *Salix* (*S. herbacea* leaf fragments are present), *Artemisia*, Asteroideae/Cardueae undiff., Lactuceae, Caryophyllaceae, *Huperzia selago* and *Selaginella selaginoides*. This pollen assemblage is indicative of cold open landscapes in which heliophilous herbs thrived and total pollen concentrations fall along with marked increases in the proportions of pitted/thinned *Betula* and Poaceae pollen grains and unidentifiable palynomorphs. The severity of the environmental attributes of this LPAZ is characteristic of the Younger Dryas interval.

Zone CRU-5 marks the beginning of major floristic change at the site. Thermophilous woodland elements *Betula* and *Corylus avellana*-type expand, Poaceae continues to be well represented overall, and there are collapses in typically open land taxa which characterise cold environments (e.g. *Artemisia*, Asteroideae/Cardueae undiff., Lactuceae, *Huperzia selago* and *Selaginella selaginoides*). *Quercus* and *Ulmus* are present in trace amounts, but were probably growing further south on the British mainland. The marl deposits are clearly reflecting the early Holocene environment around Crudale Meadow.

Subzone CRU-5b is a notable oscillation in which *Empetrum nigrum* expands to 45% TLP along with a rise in *Myriophyllum alterniflorum* to its profile maximum (10%) and there are relative and concentration rate falls in *Corylus avellana*-type, *Salix* and Poaceae. This seems to denote a prolonged cold phase with more catchment-scale erosion (LOI declines) – the rise in *M. alterniflorum* at this point probably reflects the ingress of base-rich minerogenic material to the then lake. The fall in presumed anemophilous *Pinus sylvestris* pollen may show that the cold episode was of wide geographical scope and the most likely candidate is the Preboreal Oscillation (PBO) of ca. 11400 cal. BP (Björck et al., 1997; Bos et al., 2007).

The start of LPAZ CRU-6 sees a marked decline in *Empetrum nigrum,* and increases for *Pinus sylvestris, Filipendula* and Pteropsida (monolete) indet. Poaceae remains at high levels throughout. The zone is subdivided at the point where expansions occur in *Ulmus* and *Equisetum* (Supplementary Information, Fig. 1), with falls in *Betula* and *Rumex*, followed by a consistent decline in *Empetrum nigrum.* The zone seems to be indicative of birch-hazel scrub and scattered pine with, variously, a tallherb dryland and/or mire flora on adjacent slopes rich in grasses, meadowsweet, horsetails and ferns.

Zone CRU-7 is especially characterised by declines in *Betula* and Poaceae and increases for *Corylus avellana*-type, *Quercus*, *Ulmus*, *Alnus glutinosa* and *Salix.* These changes are clearly paralleled in the concentration diagram (Supplementary Information, Fig. 2). The assemblage is probably reflecting an increased presence of deciduous woodland within the pollen catchment area, while the expansion of *C. avellana*-type (which reaches 75% TLP) could also denote the spread of *Myrica gale* (cf. Edwards, 1981) in nearby damp areas in which *Salix* and *Filipendula* were major components. Given the magnitude of the fall in Poaceae (from 19% down to ~3%), a decline of grasses on the mire itself is probably indicated.

The main features of zone CRU-8 are a continuous decline in *Corylus* values, a major sustained expansion in Pteropsida (monolete) indet. spores (to 121% TLP) along with those of *Drypoteris filix-mas*-type, and an increase in the pollen of *Pinus sylvestris* (reaching 25%). The spectra are probably reflecting the vegetation on adjacent slopes including an increasing fern element.

Zone CRU-9 is represented by several spectra within the sedge peat which succeeds the deposition of marl. A hiatus may exist at the CRU-8/9 boundary or the palynomorph catchment areas may have changed dramatically from one dominated by adjacent dryland taxa, including microfossil components from incoming streams and slopewash, to one in which mire taxa – in this case Poaceae and Cyperaceae – are over-represented.

5.3. Mollusc analysis

At most levels the shells were well preserved, although below 516 cm, within the Lateglacial deposits, some corrosion had occurred. Where shells were damaged in this way identification was not easy; otherwise, except for the abundant juveniles of *Pisidium* spp. noted earlier, the molluscs presented few problems of identification to species level. The total fauna consists of eight determinable species of which only *Lymnaea peregra* and *P. nitidum* are present at most levels. The other taxa are restricted to levels below 400 cm. Between 501 and 516 cm (cf. the Younger Dryas) no Mollusca were recorded.

The presence of two rare species of *Pisidium* requires comment. *P. obtusale lapponicum* is a boreal and arctic sub-species of *P. obtusale* (Ellis, 1978; Kuiper et al., 1989). Six shells from levels between 466 and 491 cm are characteristic of *P. obtusale lapponicum* while a further twenty shells have the less globular form of *P. obtusale*. Also occurring are six valves identified as the arctic-alpine species *P. vincentianum* (Ellis, 1978). Except for one valve from 451-456 cm, all these specimens are from levels below 526 cm where shell preservation is poor.

The species in the fauna are mostly pioneer forms with highly developed dispersal capabilities (Kerney, 1999). They are all pond species and there is no evidence of any water movement from either springs or streams entering the waterbody. As pioneers, the Mollusca at Crudale Meadow are tolerant of a variety of water conditions, both of temperature and substrate, and the present-day distribution of all species ranges to high latitudes in Europe (Kuiper et al., 1989; Ökland, 1990; Kerney, 1999). All taxa are of permanent water rather than ephemeral pools liable to drying; this is in contrast to Quoyloo Meadow (O’Connor and Bunting, 2009) where the Lateglacial section of the profile contained few molluscs and the Holocene section had frequent individuals associated with shallow water areas and probable desiccation. Water depths at Crudale Meadow were no more than one metre. The occurrence of large numbers of specimens of *Gyraulus crista* at levels between 431 and 495 cm indicates macrophytic vegetation growth in the pool as the molluscan species are typical of well-vegetated water (Ökland, 1990).

The change in the environment through time is indicated by the variation of molluscan totals (Fig. 5) rather than by a sequential development of the fauna as the new species colonize the site. The first Mollusca appear at 578 cm with *Lymnaea peregra* and *Pisidium nitidum* occurring in most levels up to 521 cm. As these two species are tolerant of a wide variety of water temperatures and conditions, little can be deduced from their presence alone. The scattered occurrence of valves of *P. vincentianum* points to cold water conditions. The fluctuation of absolute numbers of shells between 576-531 cm, from 2 to 45, may suggest environmental change, but in the absence of a greater diversity of species it is difficult to suggest a reason for this.

Between 521 and 495 cm, shells are absent or nearly so. The presence of low molluscan totals after the steady presence of shells below 521 cm indicates that between 521 and 496 cm a clear environmental change took place. Two possibilities may explain this change: an episode of drying, making conditions unsuitable for aquatic Mollusca, or an episode in which winter ice persisted into the summer and prevented the oxygenation of the water from the atmosphere, thus making molluscan life impossible. This latter possibility is a major control over the distribution of Mollusca in high latitudes at present (Ökland, 1990). Of the two possible causes, a complete desiccation of the waterbody seems unlikely. There is no evidence in the lithostratigraphy for a hiatus in deposition or a weathering horizon that might have developed, if the pond had dried up. Similarly, there is no trace of colonization of the site by terrestrial Mollusca as might occur if the waterbody disappeared (cf. Horne, 2000). Therefore, it seems probable that an extreme cold event, which prolonged ice cover in the summer, was responsible for the extinction of the fauna in these levels. This would be consistent with similar situations reported from northern England (Keen et al., 1984, 1988; Jones et al., 2000) and Sweden (Hammarlund and Keen, 1994) and in each case attributed to the Younger Dryas ice advance.

Above 491 cm and up to 436 cm, the fauna is at its most diverse and exhibits the highest numbers of individuals in the whole core. This phase probably indicates an improvement in conditions allowing the re-immigration of the Mollusca into the site at the beginning of the Holocene.

From 436-322 cm, molluscan numbers and diversity both decline. For much of the span only three species occur and numbers are below 30 individuals. This change to less diverse fauna marks a further environmental change, but the exact nature of this is difficult to determine. Climatic fluctuations during the first two millennia of the Holocene are now well documented and these would affect the temperature, water levels and vegetation cover (Yu and Harrison, 1995; Hughes et al. 2000; Jones et al., 2000) of the waterbody and thus have repercussions on molluscan growth.

Above 322 cm, numbers of individuals are again high, suggesting that conditions for molluscan existence were good. It is, however, difficult to account for the small number of species above 322 cm if conditions at Crudale had become favourable. In contrast, mid-Holocene faunas from Orkney (de la Vega-Leinart et al., 2000) do show diversity in the species assemblage, suggesting that the events at Crudale Meadow which caused the impoverishment of the fauna were perhaps local in their effect (a similar pattern of relative species poverty is apparent at Quoyloo Meadow; O'Connor and Bunting 2009). De la Vega-Leinart's site at Bay of Skaill was a much larger basin in a more lowland open setting, whereas Crudale and Quoyloo Meadows are both small basins with small catchments and probably had only limited open water surrounded by wider belts of marsh, fen or mire, which may have inhibited Mollusca richness.

5.4. Isotope analysis

Different sources of evidence indicate that the marl samples are free from clastic carbonate. The siliclastic-rich intervals contain no carbonate (as confirmed by XRD), suggesting that the water-body's surroundings do not supply clastic carbonate. Further, this is strongly supported by the absence in the XRD analyses of dolomite which should be present if some carbonate were derived from tills which contain partly dolomitic siltstones of the Middle Old Red Sandstone. $\delta^{13}C_{org}$, $\delta^{13}C_{marl}$ and $\delta^{13}C_{shell}$ display similar trends (Fig. 6a), indicating that their ultimate source of carbon was identical, i.e. the same dissolved inorganic carbon (DIC) of the waterbody, and both $\delta^{18}O_{marl}$ and $\delta^{18}O_{shell}$ also have the same general trend (Figs. 6b and 7). The falsely old radiocarbon analyses (section 5.5) are consistent with this – a carbon source containing dissolved old carbonate would be expected to yield significantly lower $^{14}C/^{12}C$ than would be found in contemporary, atmospherically-derived sources.

The similarity in the isotopic trends of shells and marl is clearly indicative that the record is robust and uncontaminated by clastic carbonates. However, the difference between isotopic composition of marls and shells (~1.5-2‰ and ~4‰ for oxygen and carbon respectively) needs some further explanations. Freshwater shells are usually aragonitic (Zanchetta et al., 1999; Leng and Marshall, 2004), and aragonite is ~0.6‰ ^{18}O-enriched compared to calcite (e.g. Tarutani et al., 1969). The rest of the difference can be explained by shell vital offset compared to isotopic equilibrium conditions, and to difference in timing of deposition which influences temperature of precipitation. Under favourable conditions *Lymnaea peregra* seems to grow continuously throughout the year, giving an environmental signal that mixes temperature variation with changes in the isotopic composition of water (White et al., 1999), and so $\delta^{18}O_{shell}$ also records the lower temperature during winter. The differences observed here are in agreement with other cases where bulk carbonate and shell oxygen isotopic composition are considered (e.g. Filippi et al. 1997; Böttger et al. 1998; Zanchetta et al., 2007). Similarly the differences found between the $\delta^{13}C_{shell}$ and $\delta^{13}C_{marl}$ are in agreement with those observed in other studies. Indeed usually the difference between the $\delta^{13}C_{shell}$ and $\delta^{13}C_{marl}$ is <0, and variable, depending on the freshwater species considered (e.g. Shanahan et al. 2005, Zanchetta et al., 2007) but values of -4 to -6 ‰ are common (Zanchetta et al., 2007). This difference depends on the relative importance of respired versus environmental CO_2 (i.e. water DIC isotopic composition) in molluscs (McConnaughey and Gillikin, 2008), amplified by the local effect during bio-induced precipitation of endogenic calcite from a DIC locally ^{12}C-depleted by photosynthetic activity (Leng and Marhall, 2004). In addition, the molluscs may record a larger variation in the lake DIC throughout the year (as noted above for oxygen isotopic composition) than marl, which usually precipitates during the warmer part of the year.

The $\delta^{18}O_{marl}$ record shows a long-term trend (presented in Fig. 6c by the 6th order regression) punctuated by several oscillations of varying amplitude and duration. To give importance to coherent and persistent oscillations, the curve in Figure 6c is averaged every 2 cm so as to smooth the possible effects of bioturbation. It is important to note that although the smoothing process depicts the $\delta^{18}O_{marl}$ curve as continuous other than for the major break at C3 (cf. Younger Dryas), there are also minor breaks centred elsewhere which usually coincide with the presence of silty layers (hence no measurements on marl were possible). Like C3, the deposits inferred to contain records of cold episodes (C1, C2, and C5 [C=cold]) contain levels with neither marl nor shells, and this may reflect temperature extremes.

We assume that the oxygen isotope composition of marl mainly reflects the dominant effect of the change in isotopic composition of rainfall recharging the lake, rather than changes in the oxygen isotope fractionation factor between calcite and water with temperature (e.g. Leng and Marshall, 2004). Indeed, at middle latitudes there is a significant relationship between the isotopic composition of rainfall and air temperature (Rozanski et al., 1993), with the ratio of $\Delta\delta^{18}O/T$ in precipitation much higher than the change in the oxygen isotope fractionation factor between calcite and water. This would indicate that a significant part of the variability in isotopic composition of lake calcite is related to changes in air temperature and associated effects on the isotopic composition of lake water (Leng and Marshall, 2004). This seems the most appropriate assumption for the location of this site, and follows a similar approach used for interpreting records in Ireland (Ahlberg et al., 1996; Diefendorf et al., 2006) and England (Marshall et al., 2002, 2007; Daley et al., 2011). We consider that this first order interpretation is also justified because many European carbonate $\delta^{18}O$ records located at middle latitudes substantially parallel the Greenland ice isotopic record (e.g. Marshall et al., 2002, 2007) (section 6), which is accepted as mostly reflecting temperature changes. However, other additional factors would affect the final isotopic composition of lacustrine carbonates such as changes in the hydrological budget of the lake, and/or changes in the pattern of precipitation and rainout linked to changes in atmospheric circulation (Marshall et al., 2007). In particular, as the waterbody (at least today) has no clear inflow channel, though was presumably partly groundwater-fed through carbonate-rich tills, and its outflow stream was a minor, channelized drainage feature, it is likely that evaporation may have played some important role during particular phases, enhancing or dampening the effect of changes in isotopic composition of rainfall and temperature. For instance, two egregious high and apparently aberrant values of $\delta^{18}O_{marl}$ at 461 and 252 cm may derive from episodes of enhanced evaporation of the lake water, rather than representing temperature extremes, indicating drier rather than warmer conditions (– other proxies do not contradict this). Interestingly, these two extreme values (confirmed by replicate measurements) are associated with an interval of general increase of $\delta^{18}O_{marl}$ values, suggesting increasing evaporation associated with warmer conditions. It is also important to note that the isotopic record of marl is mostly indicative of spring/summer conditions when algal bloom is expected (Leng and Marshall, 2004). Being aware that the interpretation of $\delta^{18}O_{marl}$ in terms of simple changes in isotopic composition of precipitation driven by temperature would be a simplification, in general we assume that increase in $\delta^{18}O_{marl}$ corresponds to climatic improvement and a decrease in $\delta^{18}O_{marl}$ to climatic deterioration.

The carbonate-^{18}O depletion peaks C1 and C2 below 520 cm, inferred to represent lowered temperature during the early stages of sediment accumulation in the basin (Figs 2 and 6c), are flanked by three ^{18}O-enrichment peaks (maxima at 562, 550 and 543 cm). These are assumed to represent rises in temperature, so that the basal record at Crudale Meadow indicates considerable variations in temperature. As the sediments between 520 and 500 cm (C3) are devoid of carbonate, indicating a substantial episode of low temperatures, the basal oscillations would seem to be related to the widely recognised Lateglacial multi-phase Bølling-Older Dryas-Allerød-Younger Dryas sequence.

The strong carbonate-^{18}O enrichment of marl after C3 would indicate the marked rise in Holocene temperatures which took place following the cold stade, while the subsequent oscillation (C4) is suggestive of Preboreal cooling (Björck et al. 1997; Bos et al., 2007). The recovery from that episode was again interrupted (event C5) – bearing in mind that the level of $\delta^{18}O_{marl}$ values at 461 cm is regarded as an isotopic rather than a climatic excursion. There followed an overall inferred rise in temperature which was later to undergo a decline (after about 325 cm), with a period (C6) where unstable conditions with cooler events marks the end of this period. After a new climatic recovery, colder conditions may then be present at the end of the record (C7). It is possible that cold periods C6 and C7 represent the 9.3 and 8.2 ka events respectively (Rasmussen et al., 2007; Yu et al., 2010; sections 5.6 and 6). Both events have been previously inferred from British deposits (e.g. Edwards et al., 2007; Marshall et al., 2007; Lang et al., 2010a; Daley et al., 2011), though relatively infrequently for the 9.3 ka event. We feel it important to stress the uncertainty associated with the isotopic perturbations in the Holocene section of the profile. The 2 cm averaged $\delta^{18}O_{marl}$ isotope curve contains many short fluctuations, but the demarcation of more than the discussed oscillations in the upper section of the core would be somewhat arbitrary. Had smoothing not been applied to the data, it would have been possible, for instance, to include further cold oscillations.

The $\delta^{13}C_{marl}$ record is similar to, although with higher resolution than, the $\delta^{13}C_{org}$ and $\delta^{13}C_{shell}$ records (Fig. 6a). From 576 cm to 522 cm, the $\delta^{13}C_{marl}$ shows a progressive decrease. A marked $\delta^{13}C_{marl}$ rise occurs above 500 cm, which is rapidly followed by a long-term trend of progressive ^{13}C depletion. The $\delta^{13}C_{marl}$ decrease is not continuous, but is marked by sections with almost constant values separated by phases of sharp decline. The general trend of $\delta^{13}C_{marl}$, as well as that of $\delta^{13}C_{org}$, merits comment. The broad covariance of these records is clear evidence that the same carbon pool was utilised by both organic matter and inorganic carbonate, *viz*. the dissolved inorganic carbon of the lake. The ~25‰ offset between the two records is characteristic of the C3 mode of photosynthesis (cf. Fig.6 of Whittington et al., 1996). Most of the important changes in the $\delta^{13}C$ values, especially during the early Holocene, are, to some extent, correlated with the local pollen zone boundaries (Figs 3-4). This suggests that a link exists between changes in the terrestrial environment and carbon isotope composition of the lake DIC. One plausible mechanism to explain this link is the change in the amount of transfer of soil CO_2 into the lake. Soil CO_2 usually has low $\delta^{13}C$ values, which originate from the oxidation of organic matter and root and bacterial respiration (Deines, 1980; Cerling, 1984). A tight linkage exists between plant productivity, specific vegetation type, climate condition (i.e. temperature and amount of rainfall) and soil CO_2 production (i.e. soil respiration) (Brook et al., 1983; Raich and Schlesinger, 1992). Changes in type and extent of vegetation can therefore modulate the soil CO_2 production and the availability of the CO_2 leached and delivered to the lake systems (e.g. Aravena et al., 1992; Benson et al., 1996; Lezine et al., 2010) as well as soil recovery and development after the Lateglacial cold stages. The lowest rates of soil respiration occur in the coldest and driest biomes. Under these conditions the amount of CO_2 delivered to a lake is strongly reduced and equilibration of lake water with atmospheric CO_2 may occur along with consumption of CO_2 by biological activity, both of which produce high $\delta^{13}C_{marl}$ (e.g. Leng and Marshall, 2004). The progressive decrease of $\delta^{13}C_{marl}$ during the Holocene and the Lateglacial interstadial may be due to ongoing increases in inputs of ^{13}C-depleted soil CO_2, controlled by vegetation abundance and climate condition. This can also apply at a fine scale when Lateglacial oscillations are considered. If this hypothesis has some basis there should be analogues in other lake records, and studies at other sites in Orkney's West Mainland (e.g. Peat Moss No. 27 ['Lime gyttja'], Cairston [Erdtman, 1924], The Loons [Moar, 1969], Loch of Skaill/Pow [Keatinge and Dickson, 1979], Quoyloo Meadow [Bunting, 1994, O'Connor and Bunting, 2009]) could be advantageous. This would be difficult to carry out on much of the Scottish mainland owing to the removal of marl for use as a soil dressing during the period of the Agricultural Improvement Movement which gathered momentum from the middle of the eighteenth century AD (Whittington, 1975). The relationship between $\delta^{13}C_{marl}$ and $\delta^{13}C_{shell}$ (Figs 6a and 7) has a scatter of values consistent with a component of low ^{13}C carbon contributing to metabolic precipitation of shell carbonate (cf. Parkinson et al., 2005).

5.5. Chronology

It was appreciated that the calcareous nature of the deposits at Crudale Meadow for the period under consideration, and the general lack of terrestrial plant macrofossils, were likely to be problematic for a radiocarbon-based chronology (cf. Karrow et al., 1984; McDonald et al., 1991; Garnett et al., 2004b) due to a contribution of carbon from calcareous sediment resulting in older radiocarbon ages. The peat immediately overlying the marl and the near-adjacent marl beneath (200.25-202.75cm) were radiocarbon dated in the hope that this would allow reasonable age estimates for the marl, along with an additional bulk sediment (428cm) date on marl further down the profile. Three samples of possible aquatic plant remains from 523-526cm, close to the start of inferred Lateglacial (cf. Younger Dryas) deposits were also dated.

The ^{14}C dates at Crudale Meadow are clearly too old and have been affected by the hardwater error arising from the calcareous sediments in the catchment area (cf. Harkness, 1979). A linear regression (R^2=0.998 but biased by closely-spaced multiple data points at either end of the depth range with only one marl date in between) produces an intercept at 0 cm depth of 4777 ^{14}C yr BP. The uppermost date of 9630-9300 cal BP (95.4% probability range; Table 3) is palynologically unlikely as it relates to a period when woodland had been much reduced, which in Orkney generally dates to after the start of Neolithic activity (i.e. post-5800 cal BP; Keatinge and Dickson, 1979;

Bunting, 1994, 1996; de la Vega-Leinart et al., 2000; Farrell et al., 2014). It seems probable that there is carbonaceous contamination of the peat above the marl or a hiatus at the stratigraphic boundary. The underlying marl date (10250-9890 cal BP) might otherwise have been acceptable as displaying a half millennium age offset, but not when considered with the Lateglacial age of 16230-15630 cal BP for the marl sample centred on 428 cm, which is clearly associated with Holocene deposits. The Lateglacial fares no better in that the series of three dates from possible aquatic plant remains (a supposition supported by the very negative $\delta^{13}C$ values) resulted in age estimates of >18000 cal BP rather than the ca. 13000 cal BP of the Younger Dryas/Allerød boundary (Table 4). The data are not suitable for deposition modelling, producing poor agreement indices with the Bayesian statistical approach in OxCal (Bronk Ramsey, 2015). Although the contribution of old carbon from calcareous sediments cannot be assumed to have remained constant over the deposition period or across the different sample types, applying a correction of 4777 ^{14}C years to all sample data results in a calibrated age range of 12790-13460 cal BP for depths 523-526 cm, not inconsistent with ca. 13000 cal BP of the Younger Dryas/Allerød boundary boundary. By interpolation, the Saskunavatn Ash at 384 cm (see below) would have a radiocarbon-based age estimate of c.9000-10500 cal BP compared with the eruption date of ~10297 cal BP. However, these age estimates are highly speculative, and the unquantifiable influence at this site of the old calcareous carbon on the radiocarbon data discouraged further radiometric dating.

Microprobe analysis revealed the cryptotephra at 384 cm to be a good fit with the Saksunarvatn Ash layer (Mangerud et al., 1986; Bunting, 1994). The tholeiitic basaltic composition of the tephra is considered to denote an origin in the Icelandic Grimsvötn or Kverfjöll complex and it is widely distributed (Davies et al., 2002). The Crudale data points (Supplementary Information, Fig. 4) are quite scattered (cf. Bramham-Law et al., 2013), but they do fall over the means of tephras found at Saksunarvatn (Faroes) and, in terms of proximity, this ash layer is recorded from other northern Scottish sites such as Dallican Water, Shetland (Bennett et al., 1992) and Loch Ashik, Isle of Skye (Pyne-O'Donnell, 2007).

5.6 Synthesis

Between that basal portion of the profile inferred to represent the Bølling (CRU-1) and the Younger Dryas (CRU-4), there are two marked cold oscillations reflected in lowered $\delta^{18}O$ values of the phases C1 and C2 (cf. Older Dryas and Intra-Allerød Cold Period) with intervening warm events. The LOI and, to a lesser extent the magnetic susceptibility records, reflect high minerogenic inputs, typical of less stable, solifluction processes over a prolonged period embracing the cold phases. The vegetation was open throughout and dominated by herbaceous and low shrub taxa characteristic of unstable substrates. Molluscs, including pioneer and arctic species, were present in low numbers and were often corroded. There was a possible cold oscillation centred upon 548 cm (between C1 and C2) with fluctuating but lowered $\delta^{18}O$ values corresponding with a reduction in *Empetrum nigrum* and perhaps *Betula*, along with expansions in *Artemisia*-type and Asteroideae/Cardueae undiff.

The Crudale Meadow site is no longer a lake, and the inability to assess the limnological characteristics of the site discourage us from attempting quantitative reconstructions of past temperature and precipitation based upon the isotope data (cf. Eicher and Siegenthaler, 1976; Ahlberg et al., 2001; Marshall et al., 2002; Leng and Marshall, 2004). Nevertheless, the collective environmental analyses point to a period of extreme cold between 520 and 500 cm. Marl was no longer deposited in the basin, being replaced by siliciclastic sediments with a very low LOI and a continuing strongly positive magnetic susceptibility response. Molluscs were extinguished not because the waterbody had dried up, but probably due to the likely existence of an ice cover which extended well into the summer months. From the pollen analyses, the distinct nature of LPAZ CRU-4, with the lowest pollen concentration values for the whole profile and its strong representation of *Salix, Artemisia,* Asteroideae/Cardueae undiff. and Lactuceae undiff., confirms the severe nature of the climate and the existence of a tundra. The pollen preservation analyses

(Supplementary Information, Fig. 3) suggest that this climatic severity was greater than any of the cold episodes during the preceding interstadial period. All of these episodes show increased pollen degradation, but during the Younger Dryas all of the pollen of the main taxa are at some stage suffering from either pitting or thinning. This suggests that the soils around the basin were being severely eroded, allowing the redeposition of soil pollen.

What appears to be a secure defining of the Younger Dryas stratigraphic event strongly supports the argument that the marl deposits above 500 cm belong to the Holocene. The end of the Loch Lomond Stade is known to have been marked by a rapid rise in temperature and the $\delta^{18}O$ record reveals this with a sharp carbonate-^{18}O enrichment peak between C3 and C4. LOI values increase sharply as organic soil development proceeds apace and the markedly lowered susceptibility curve denotes a commensurate reduction in magnetite-enriched eroded soils reaching the basin. LPAZ CRU-5a reveals increases in Poaceae and *Betula*, but perhaps more significantly the establishment of the forerunner of a continuous *Corylus* curve as hazel, a pioneer thermophilous shrub, migrates into the pollen catchment area. During this LPAZ the molluscan representation is numerically at its highest for the whole of its record (Figs 2 and 5).

During the earliest Holocene, the strong climatic oscillation of the inferred PBO (C4) affected the land on both sides of the North Atlantic, although its age is difficult to determine due to the presence of two radiocarbon plateaux (c. 11300–11150 cal yr BP). In an investigation of this event, Björck et al. (1997) found that evidence for it varied widely between sites. There seems little doubt that the lacustrine sediments at Crudale Meadow record this event. Figure 6c shows that at 491 cm, following upon the sharp recovery of temperature at the start of the Holocene, there is a rapid decline, followed by a further rise that culminates in the $\delta^{18}O$ value at 463 cm. During the inferred cold phase (C4), there is a significant response in both the molluscan and the palynological records. The number of shells declines from its peak of 1069 at the height of the immediate post-Younger Dryas temperature rise to only 74. The CRU-5a/5b subzones boundary shows that the continuous expansion of *Corylus*, which might be expected after the taxon had become established, is considerably delayed. There is also a marked expansion in values for *Empetrum* which is sustained until the cold was ameliorated later in CRU-5b. The pollen preservation status is again severely affected with all of the main taxa showing high percentages of pitted and thinned exines.

The recovery in the $\delta^{18}O$ curve in the latter part of subzone CRU-5b sees the start of increased values for the thermophilous trees *Betula* and *Corylus avellana*-type, a rise in *Myriophyllum alterniflorum* which may owe its resurgence to warming rather than inputs of base-rich minerogenic material to the then lake as surmised for the preceding peak in the taxon, and the start of an increase in total palynomorph concentrations. The $\delta^{18}O$ peak at 461 cm is followed by a sharp fall which, at its minimum, is matched by declines in warm pollen taxa (*Betula, C. avellana*-type, *M. alterniflorum*). Accepting the tephra peak at 384 cm as denoting Saksunarvatn Ash deposition, then the preceding isotopically determined oscillation C5 (perhaps ca. 11000 cal yr BP or shortly thereafter) can be assigned to a cold phase in the ice-core and some regional records (discussed in section 6), though this does not seem to have been generally recognised.

Following this, the environmental records from Crudale Meadow cover the rest of the marl deposits of the Holocene. The $\delta^{18}O$ record appears to suggest that temperatures ameliorated and maintained an equilibrium up to the depth of about 330 cm. From that time a general cooling developed, with periods of $\delta^{18}O$ minima (C6-C7), with some recovery in between. The general directional increase in the oxygen isotopic values is matched by that of the molluscan record up to 320 cm. As noted previously, this is a time of low numbers of shells accompanied by low diversity of species. Explanation for this phenomenon is still no further advanced. On the contrary, the pollen record does show a response to what appears to be a period of greater warmth. The delay in the expansion of *Corylus* as a result of the Preboreal cooling is overcome and the taxon’s pollen values increase steadily throughout LPAZ CRU-6, while those for *Empetrum* go into continuous decline.

The possible interruption caused by cold oscillation C6 (which may represent the 9.3 ka event) has no demonstrable impact upon vegetational succession. It occurs as part of the decline in temperature levels noted earlier and beginning around 325 cm. At the latter depth, there is a sudden fall of over 20% in the LOI value which may be related to the fact that Mollusca begin to increase considerably in abundance – although there are no obvious indications of a shallowing of the water body which might have been thought responsible for this. The palynological record also reveals changes. *Betula* increases and from the start of zone CRU-7, *Salix, Corylus avellana*-type (cf. *Myrica* gale in this part of the profile) and *Filipendula* expand and Poaceae declines abruptly, suggesting an extension of mire at the expense of grass-dominated habitats. The great expansion in the representation of Pteropsida (monolete) indet. could indicate shady areas within birch stands, if not within mire communities; it seems unlikely to indicate palynomorph redeposition because of the continuing deposition of the marl.

The effect of an overall if unremarkable decline in temperature over the final part of the marl deposits appears to have little effect on either the status of the Mollusca or the pollen record. Thus during the period of the emplacement of these final marl deposits, it would appear that any temperature change (either cooling or warming) had little effect on the vegetation cover at Crudale Meadow (cf. C7, the possible 8.2 ka event; section 5.7). It was not until the marl deposits were replaced by sedge peat that the vegetational landscape changed. In terms of both percentages and concentrations, *Corylus avellana*-type pollen values became negligible while both Cyperaceae and Poaceae increased (Bunting, 1994). Given the stratigraphic and palynological changes, the zone CRU-8/9 boundary may signify a hiatus in sediment accumulation – a similar pattern in lithobiostratigraphy is evident at nearby Quoyloo Meadow (O'Connor and Bunting, 2009).

5.7 Lead-lag relationships

The isotope and palynological data at Crudale Meadow are sufficiently detailed to assess the phenomenon of lead-lag relationships (e.g. Coope et al., 1998; Ammann et al., 2000; Hoek, 2001; Colombaroli et al., 2007; Edwards and Whittington, 2010). Figures 3 and 4 reveal the correspondence between the $\delta^{18}O_{marl}$ record and selected pollen taxa. The latter can be considered as individual taxa, or collectively when grouped within LPAZs and subzones. Attempts to define the onset and termination of climatic events can be 'an ambiguous task' (Lowe et al. 2008, note to their Table 1), especially in the absence of deuterium excess data which can provide the clearest indications of climate change (cf. Rasmussen et al., 2006), but are not readily available from the archives described for Crudale. The decision is taken here to demarcate the start of cold oscillations as the mid-point between a preceding $\delta^{18}O_{marl}$ enrichment peak and the immediately following point on the slope towards a minimum in the $\delta^{18}O_{marl}$ curve. Conversely, the start of the warm oscillation is taken to be the mid-point between the minimum in the $\delta^{18}O_{marl}$ curve and the immediately succeeding point on the slope towards a maximum in the $\delta^{18}O_{marl}$ curve.

It might be expected that vegetation would respond more rapidly to marked down-turns in temperature than to warming trends owing to the slower migration rates and inertia of many individual plant taxa and communities (Wick, 2000; Von Holle et al., 2003). Figure 4 shows that the start of the cold oscillations C1, C2 and C4 precedes vegetational changes as denoted by zone and subzone boundaries, while C3 and C5 are associated with inferred vegetational changes which are mostly synchronous within the resolution constraints of the pollen and isotope data (see also Supplementary Information Figs 1-3).

The level of synchronicity of response evident in the cases of the C4 and C5 oscillations is of considerable interest as it relates to apparent vegetational changes in response to two inferred cold episodes, only one of which – the Preboreal Oscillation, taken here to be reflected in phase C4 – is recognised. Of course, the use of zone boundaries in this way 'averages' changes in that some taxa may already be falling or increasing across the designated isotope boundaries and there is always the issue of stratigraphic integrity. If individual taxa are examined, then three lagging zones still have pollen types which respond penesynchronously with the start of reduction in the $\delta^{18}O_{marl}$ curve – *Empetrum nigrum* and Asteroideae/Cardueae undiff. at both C1 and C2, and *Corylus avellana*-type and Cyperaceae at C4. With regard to warming episodes, then there is little consistency of response in taxa – a feature also seen from a collation of Lateglacial sites in eastern Scotland (cf. Table 6 in Edwards and Whittington, 2010).

The Holocene isotope-palynological events later in the profile are less distinct and probably more short-lived. As indicated earlier, the palaeoflora does not seem to display clear responses to inferred temperature oscillations (C6, C7) which are only tentatively referable to the 9.3 and 8.2 ka events respectively.

6. Broader comparisons

A considerable body of isotope data is becoming available for Britain and Ireland as well as complementary climate data in the form of chironomid evidence (cf. comparative profiles in Daley et al., 2011; Brooks et al., 2012; van Asch et al., 2012). In addition, most reports also make comparisons with data from the Greenland ice cores and, indeed, adopt 'event' terminology (Lowe et al., 2008) to some extent (e.g. Whittington et al., 1996; Brooks and Birks, 2000; Garnett et al., 2004a; Diedendorf et al., 2006; Marshall et al., 2007). The dangers of homotaxial error and nomenclatural confusion (cf. De Klerk, 2004; Railsback et al., 2015) are inevitably present when inferring environmental history which makes use of both 'event' and traditional climatostratigraphic nomenclature (cf. Björck et al., 1998; Walker et al., 1999; Edwards et al., 2000; Ilyashuk et al., 2009; Edwards and Whittington, 2010), although as advised in Lowe et al. (2008, p. 7), this seems the most straightforward way to proceed. Here we make brief comparisons with such evidence while noting that commentary could, of course, be extended to records from further afield (e.g. Hammarlund and Keen, 1994; Drummond et al., 1995; Haflidason et al., 1995; Gulliksen at al., 1998; Mayer and Schwark, 1999; Hammarlund et al., 2002; Andrews et al., 2006; Bohncke and Hoek, 2007; Magny et al., 2007; Fletcher et al., 2010; van Raden et al., 2013).

An indicative set of Lateglacial $\delta^{18}O$ records from the British Isles (Fig. 8), together with researcher-demarcated events, reveals changes referrable to cold oscillations at Crudale Meadow which are similar to those found, for instance, at Hawes Water and Loch Inchiquin, and more clearly distinguishable than those found in more complacent records (as at Lundin Tower, Lough Gur and Fiddaun). The Orkney data would seem to represnt the most northerly Lateglacial isotopic archive yet obtained from northwest Europe.

This is not the place to attempt a discussion of quantitative and geographical differences (cf. Daley et al., 2011 which considered patterns from within a far more restricted timeframe than here), but there seem to be a sufficient number of 'tie-points' which are reinforced when fuller litho- and biostratigraphic environmental datasets are considered for the various sites (and cf. Section 5.6 here). The chironomid-inferred temperature reconstructions from the north Scottish sites of Loch Ashik and Abernethy Forest (Fig. 9) strengthen this further with pronounced mean July temperature reductions assigned to several climatic events within the Lateglacial. Although Brooks et al. (2012) acknowledge the poor absolute dating controls outwith the tephra layers for portions of the profiles, they still make tentative correlations with the Greenland NGRIP ice core records, as do most authors. In keeping with this, while accepting the potential shortcomings, we have done the same for Crudale Meadow (Figs 8 and 9; the NGRIP comparisons for the site are listed in Table 4 and shown on Figs 2-4).

The putative cold oscillation centred upon 548 cm (sections 5.2, 5.6, and termed 'C?' on Fig. 9) may be the same episode within GI-1c that may be discerned in the isotope records for Fiddaun, Lough Inchiquin, and Hawes Water and has been noted for Switzerland by Lotter et al. (2012). At Abernethy Forest, the chironomid curve has an estimated cold oscillation of about 1.9 °C dated to 13680±190 cal BP (626 cm, Fig. 9) which is tentatively equated with a cold excursion of intermediate amplitude at 13640±160 GICC05 yr BP in event GI-1c within the NGRIP record.

Isotopic records from Holocene profiles are less frequent and Figure 10 presents three profiles together with Crudale Meadow. Along with Hawes Water, the data from the Orkney site would seem to be unusual in providing good isotopic evidence for abrupt events within the early Holocene. A common feature of the curves is their high frequency fluctuating nature. In spite of this, the researchers have demarcated events reflecting major isotope excursions which are partly based on supplementary information (e.g. ^{14}C dates at Knockadoon South and Wateringbury; U-series dates at Hawes Water HWLC1). The patterns at Knockadoon South (a littoral core within Lough Gur) are complicated by the possible existence of a mid-Holocene hiatus at about 5.0 m (pre-5135 BP) caused by lowered lake levels (Ahlberg et al., 2001).

As was indicated earlier (sections 5.4, 5.6, 5.7), oscillation C5, possibly dating to around 11000 cal BP, seems to be unremarked, although similar isotope excursions may be evident at Knockadoon South, Wateringbury and Hawes Water (Fig. 10) and in NGRIP, GRIP and DYE-3 Holocene $\delta^{18}O$ records (Rasmussen et al., 2006; Vinther et al., 2006;Walker et al., 2012). In contrast, post-PBO isotopically-determined warming is inferred at Loch Inchiquin as from ~10800 cal yr BP (Diefendorf et al., 2006), and this is related to a supposed decrease in pack ice following changes in the position of the Polar Front.

Given the caveats concerning the correlation of stratigraphic events, and accepting an approximate chronology involving cross-correlation with other time-based proxies (cf. Jones et al., 2002; Rasmussen and Anderson, 2005; Edwards and Whittington, 2010), it is possible to gauge the utility of dating inferences and estimates via correlation with data from the Greenland ice cores. The collation of data from the NGRIP ice-drilling programme (Johnsen et al., 2001; North Greenland Ice Core Project members, 2004), combined with existing GRIP and DYE-3 ice core records, new isotopic data and Bayesian re-modelling (Lowe et al., 2008; Walker et al., 2009, 2012), permit a refined time-stratigraphic correlation of palaeoenvironmental events during the Lateglacial in the North Atlantic region (Table 4). In Figure 11, the GICC05 chronology is used to assign age estimates to the climate events inferred from the Crudale Meadow palaeoenvironmental data, giving primacy to the $\delta^{18}O_{marl}$ curve and assuming that the tephra peak at 384 cm represents Saksunarvatn Ash with a GICC05 age of 10347 b2k. The resulting curve is presented for indicative purposes only and is considered to be a reasonable fit to the data for the Lateglacial and earliest Holocene section of the core, with marl formation ceasing shortly before 8000 b2k, which also reinforces the suggestion of ^{14}C contamination or more likely a hiatus (section 5.5) immediately above this.

7. Conclusions

To the best of our knowledge, the oxygen isotope evidence from Crudale Meadow is the most northerly record from the Lateglacial in northwest Europe and, arguably, only the second study from Britain which provides convincing isotopic evidence for abrupt events within the early Holocene. The data demonstrate that the north Atlantic fringe of Britain shares its Lateglacial and early Holocene environmental history with sites from Britain, Ireland and further afield, as well as permitting correlations with the Greenland ice cores.

The investigation of the oxygen isotope content of marl along with molluscan and higher resolution palynological records, has produced similarities with inferred environmental and climate events from northern Scotland. This may be seen from studies, variously, of pollen, isotopes and lithology at Grunna Water (Edwards et al., 2000) in Shetland, the Fife sites of Lundin Tower, West Lomond and Wester Cartmore (Whittington et al., 1996; Edwards and Whittington, 1997, 2010), as well as the chironomid archives from Loch Ashik and Abernethy Forest (Brooks et al., 2012). Comparisons are now shown to be more secure with the addition of the multi-proxy evidence from a location in Orkney. Confidence in the general features of isotope proxy changes is reinforced by having two $\delta^{18}O$ records (marl and shell), and three $\delta^{13}C$ records (marl, shell and organic matter). The plausible age-depth curve provides some chronological reassurance concerning site history, even if it is not possible to be too categorical about this.

The possible cold oscillation within LPAZ CRU-2 (cf. GI-1c), supported by isotope data, is hinted at from sites elsewhere and is a phenomenon that would repay further investigation. Crudale Meadow would seem to possess a convincing Preboreal Oscillation, in contrast to its absence at Clettnadal in Shetland (where insect and diatom data were also available) – perhaps for taphonomic reasons or conceivably due to greater oceanicity of the more northerly archipelago (Whittington et al., 2003). Several post-PBO cold episodes were demarcated tentatively at Crudale Meadow, and while uncertainty surrounds the designation of putative 9.3 and 8.2 ka events, another (C5), tentatively assigned to ca. 11 ka, has not to the best of our knowledge been recognised elsewhere, although similar patterns are evident in various isotope records from Britain, Ireland and Greenland. If meaningful, the failure to detect or note excursions may be down to such factors as sample resolution, site taphonomy or investigator expectation.

Acknowledgements

Thanks are due to Colin Cameron for the pollen preparations and the loss on ignition tests; NERC for the provision of radiocarbon dates (NRCF010001); Alison Sandison for artwork; and W. Cutness for permission to work at Crudale Meadow. GZ thanks the Italian National Council for Research for grants to support two 6-month visits to SUERC, which is supported by NERC and a group of Scottish Universities. Aspects of the research reported herein benefitted from the Scottish Alliance for Geoscience, the Environment and Society (SAGES). The authors greatly appreciate the comprehensive comments of two anonymous referees and the constructive advice of the editor.

This paper is dedicated to the memory of David Keen.

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Figure captions

Fig. 1. A. Sites from Britain and Ireland mentioned in the text; B. The location of Orkney and Shetland in the North Atlantic Ocean and sites mentioned in the text; C. The position of Crudale Meadow on the island of Mainland.

Fig. 2. Summary diagram of lithology, loss-on-ignition, volume magnetic susceptibility, stable isotopes ($\delta^{18}O_{marl}$, $\delta^{13}C_{marl}$), mollusc numbers and pollen zones for Crudale Meadow. Inferred 'cold' (C) events are shaded and labelled (see text and caption to Fig. 6c for further details).

Fig. 3. Selected taxa percentage (TLP) pollen and spore and isotope ($\delta^{18}O_{marl}$) data for Crudale Meadow. Inferred 'cold' (C) events are shaded and labelled.

Fig. 4. Selected taxa percentage (TLP) pollen and spore and isotope ($\delta^{18}O_{marl}$) data for the Lateglacial and early Holocene from Crudale Meadow. Inferred 'cold' (C) events are shaded and labelled.

Fig. 5. Mollusc data for Crudale Meadow. For maximum clarity, see the on-line colour version of the diagram.

Fig. 6. Stable isotope data for Crudale Meadow: (a) $\delta^{13}C_{marl}$, $\delta^{13}C_{shell}$ and $\delta^{13}C_{org}$ data; (b) $\delta^{18}O_{shell}$ and $\delta^{18}O_{marl}$ data with 6th order polynomials; (c) $\delta^{18}O_{marl}$ record with data presented as 2-point running means (corresponding mostly to 2 cm) with a 6th order polynomial and inferred 'cold' (C) events shaded and labelled. With the exception of C3 (for which no carbonate is available), all other

shaded areas are sections of the curve containing $\delta^{18}O_{marl}$ data outside -2σ of the mean; no data outside -2σ occur in unshaded regions.

Fig. 7. $\delta^{13}C_{marl}$ versus $\delta^{13}C_{shell}$ data with linear regression line for Crudale Meadow.

Fig. 8. Comparison of $\delta^{18}O$ records from selected Irish (Fiddaun [van Asch et al., 2012]; Lough Inchiquin [Diefendorf et al., 2006]; Lough Gur [Ahlberg et al., 1996]) and British (Hawes Water [Lang et al., 2010b]; Lundin Tower [Whittington et al., 1996]; Crudale Meadow [this paper]) sites with the Greenland ice core NGRIP record [Rasmussen et al., 2006]. Event abbreviations based on those suggested within the original publications (or their synonyms): All, Allerød; Bø, Bølling; GI, Greenland Interstadial; IACP, Intra Allerød Cold Period; OD, Older Dryas; YD, Younger Dryas.

Fig. 9. Comparison of NGRIP $\delta^{18}O$ isotope record (Rasmussen et al., 2006; Lowe et al., 2008; Walker et al., 2012) with those from Crudale Meadow (this paper) and Lundin Tower (Whittington et al., 1996) and chironomids-inferred temperature reconstructions from Loch Ashik and Abernethy Forest (Brooks et al., 2012). Event abbreviations based on those suggested within the original publications (or their synonyms): All, Allerød; Bø, Bølling; GI, Greenland Interstadial; IACP, Intra Allerød Cold Period; OD, Older Dryas; PBO, Preboreal Oscillation; YD, Younger Dryas. The cold oscillations (C1-5, C?) from Crudale Meadow are also indicated tentatively on the NGRIP profile.

Fig. 10. Comparison of selected $\delta^{18}O$ isotope records from Ireland (Knockadoon South [Ahlberg et al., 2001]) and Britain (Wateringbury [Garnett et al., 2004a]; Hawes Water HWLC1 [Marshall et al., 2007]); Crudale Meadow [this paper]).). Event abbreviations based on those suggested within the original publications (or their synonyms): IACP, Intra Allerød Cold Period; PBO, Preboreal Oscillation; YD, Younger Dryas. C1-7 indicate cold oscillations in the Crudale Meadow record.

Fig. 11. Tentative age-depth curve (2nd order polynomial) for Crudale Meadow constructed by assigning GICC05 chronology age estimates (Lowe et al., 2008) to inferred climate events based upon lithostratigraphic, isotopic and palynological data from the site. All terms are used within the text; / signifies boundaries between events.

Figure captions for Supplementary Information Figures

Supplementary Information, Fig. 1. Pollen and spore percentage (TLP sum) diagram for Crudale Meadow. + signifies <2.0%.

Supplementary Information, Fig. 2. Selected taxa pollen and spore concentration (palynomorphs per cm^{-3} wet of sediment) diagram for Crudale Meadow.

Supplementary Information, Fig. 3. Selected taxa pollen and spore preservation diagram for Crudale Meadow. Each preservation category sums to 100% for each taxon.

Supplementary Information, Fig. 4. Tephra microprobe data from Crudale Meadow compared to mean values for Saksunarvatn Ash (Mangerud et al., 1986; Bramham-Law et al., 2013), Faroe Islands and tephras inferred to be of the same eruption from Torfadalsvatn (Tv4), Iceland (Björck et al., 1992) and Dallican Water, Shetland (Bennett et al., 1992).

Table 1

The depositional sequence at Crudale Meadow

Depth from surface (cm)	Depositional type
201-194	Gyttja
202-201	Gyttja/marl transition
499-202	Marl with shells
500-499	Gyttja
508-500	Grey clayey silt
514-508	Grey clayey silt with organic inclusions
523-514	Grey clayey silt
537-523	Marl
559-537	Marl with shells
561.5-559	Grey clayey silt
576-561.5	Marl with shells

Table 2

Local pollen assemblage zones for the Crudale Meadow profile

LPAZ	Major taxa	Depth (cm) below surface
CRU-9	Poaceae	200-194
CRU-8	*Corylus*-*Pinus*-Pteropsida	236-200
CRU-7	*Corylus*-*Betula*-Pteropsida	308-236
CRU-6b	*Corylus*-*Pinus*- Poaceae	376-308
CRU-6a	*Corylus*-*Pinus*- Poaceae-*Filipendula*	428-376
CRU-5c	*Betula*-*Corylus*-*Empetrum*-Poaceae	460-428
CRU-5b	*Betula*-*Empetrum*-Poaceae	488-460
CRU-5a	*Betula*-Poaceae	504-488
CRU-4	*Salix*-*Empetrum*-*Artemisia*-Asteroideae-Lactuceae	524-504
CRU-3	Poaceae-Cyperaceae	540-524
CRU-2	*Empetrum*-Cyperaceae	556-540
CRU-1	*Betula*-*Salix*-Poaceae	576-556

Table 3

Radiocarbon dates[a] from Crudale Meadow

Lab code	Depths (cm)	Material	^{14}C BP (1 σ)	$\delta^{13}C$	Cal. BP [a]
AA-36189	201.25-200.25	Gyttja, close to marl boundary	8495±80	-29.3	9630-9300
AA-36190	202.75-201.75	Marl	8960±65	-27.9	10250-9890
AA-36191	428.50-427.50	Marl	13245±95	-24.0	16230-15630
AA-54791	523	Aquatic? plant remains	14950±130	-34.3	18510-17870
AA-54792	525	Aquatic? plant remains	15170±120	-34.4	18710-18110
AA-54793	526	Aquatic? plant remains	15089±81	-34.4	18570-18080

[a] ^{14}C ages calibrated using Oxcal v4.2.4 Bronk Ramsey (2015) and IntCal13 atmospheric curve (Reimer et al., 2013). 95.4% probability ranges shown, rounded to nearest 10 years. BP refers to AD 1950.

Table 4

Devensian Lateglacial ice core record nomenclature for the period covered by the Crudale Meadow profile and the duration of the oscillations (after Lowe et al., 2008 and Walker et al., 2009)

Events	Start date (GICC05 age b2k)	Duration of events (years)	$\delta^{18}O_{marl}$ curve depths (cm)	cf. Classical nomenclature
Holocene	11700		504.0[a]	
GS-1	12896	1193	523.5	Younger Dryas
GI-1a	13099	203	534.5	Allerød
GI-1b	13311	212	542.5	(Intra- Allerød Cold Period)
GI-1c	13954	643	559.5	Allerød
GI-1d	14075	121	561.0	Older Dryas
GI-1e	14692	617		Bølling

[a] Pollen-derived

Figure 1

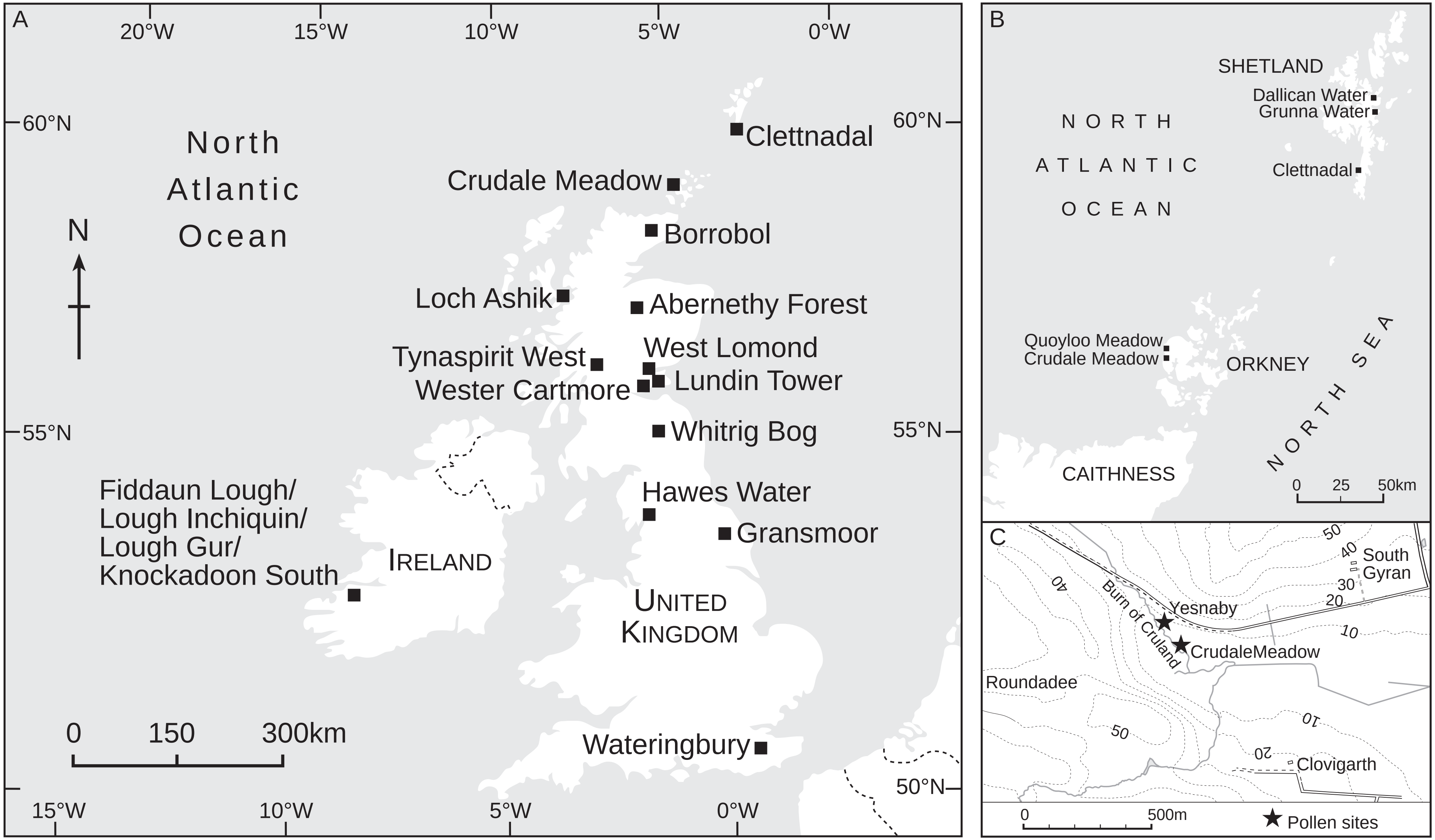

A
20°W
15°W
10°W
5°W
0°W
60°N
55°N
50°N
15°W
10°W
5°W
0°W
North Atlantic Ocean
N
Clettnadal
Crudale Meadow
Borrobol
Loch Ashik
Abernethy Forest
Tynaspirit West
West Lomond
Lundin Tower
Wester Cartmore
Whitrig Bog
Hawes Water
Gransmoor
Fiddaun Lough/
Lough Inchiquin/
Lough Gur/
Knockadoon South
IRELAND
UNITED KINGDOM
Waterinbury
0 150 300km
B
SHETLAND
Dallican Water
Grunna Water
Clettnadal
NORTH ATLANTIC OCEAN
Quoyloo Meadow
Crudale Meadow
ORKNEY
NORTH SEA
CAITHNESS
0 25 50km
C
South Gyran
Yesnaby
Burn of Cruland
CrudaleMeadow
Roundadee
Clovigarth
50
40
30
20
10
0 500m
Pollen sites

Figure 2

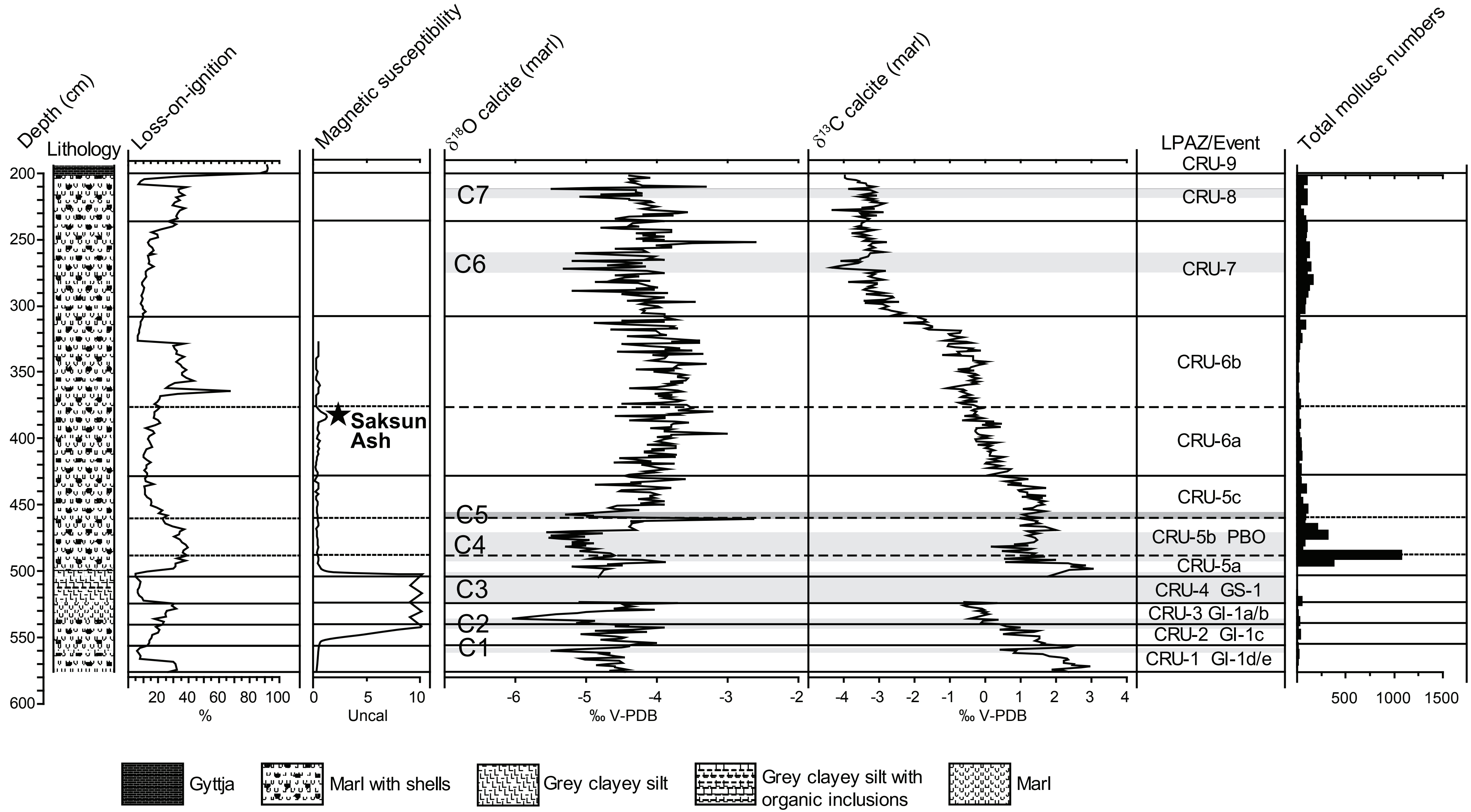
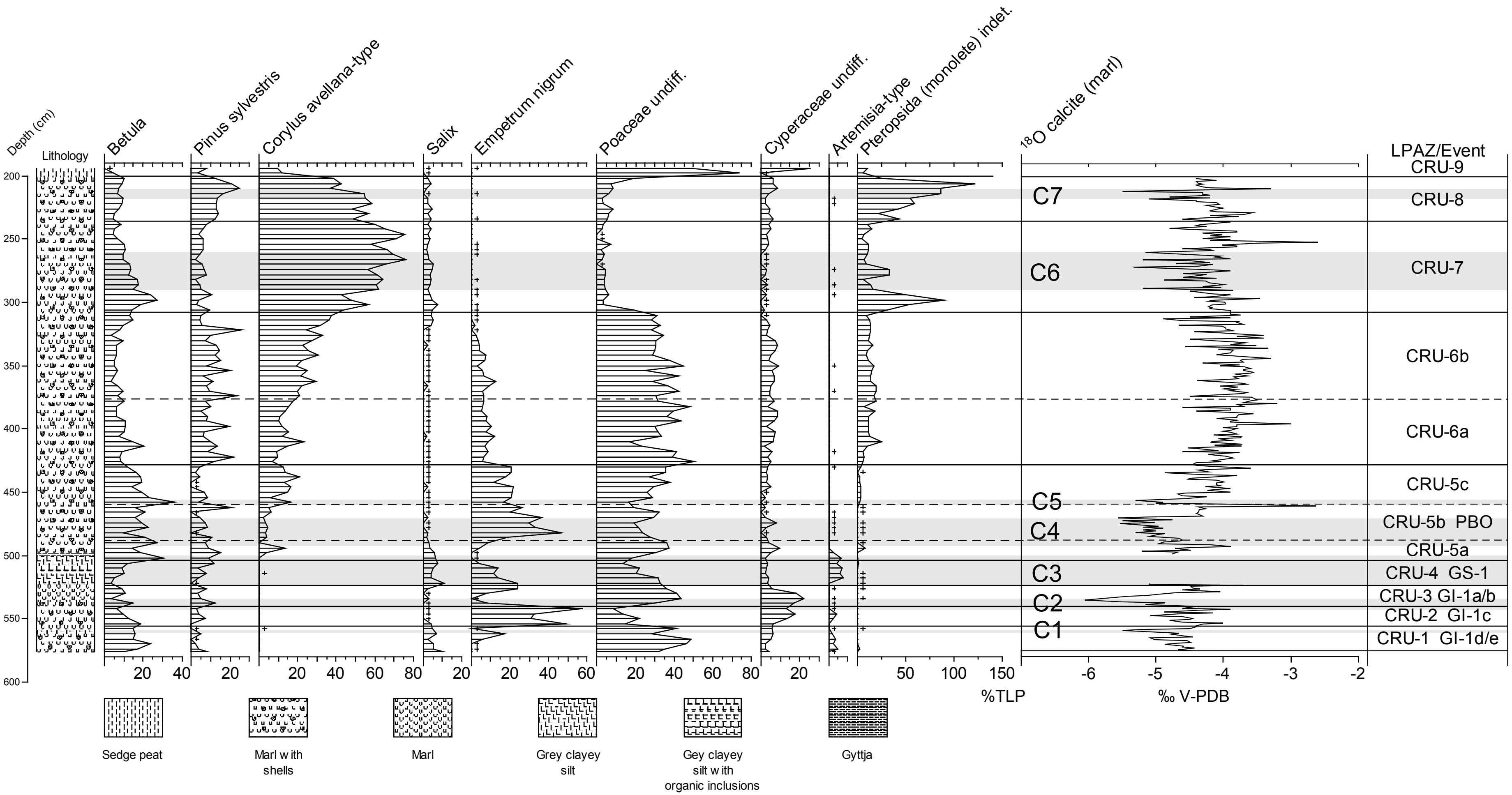
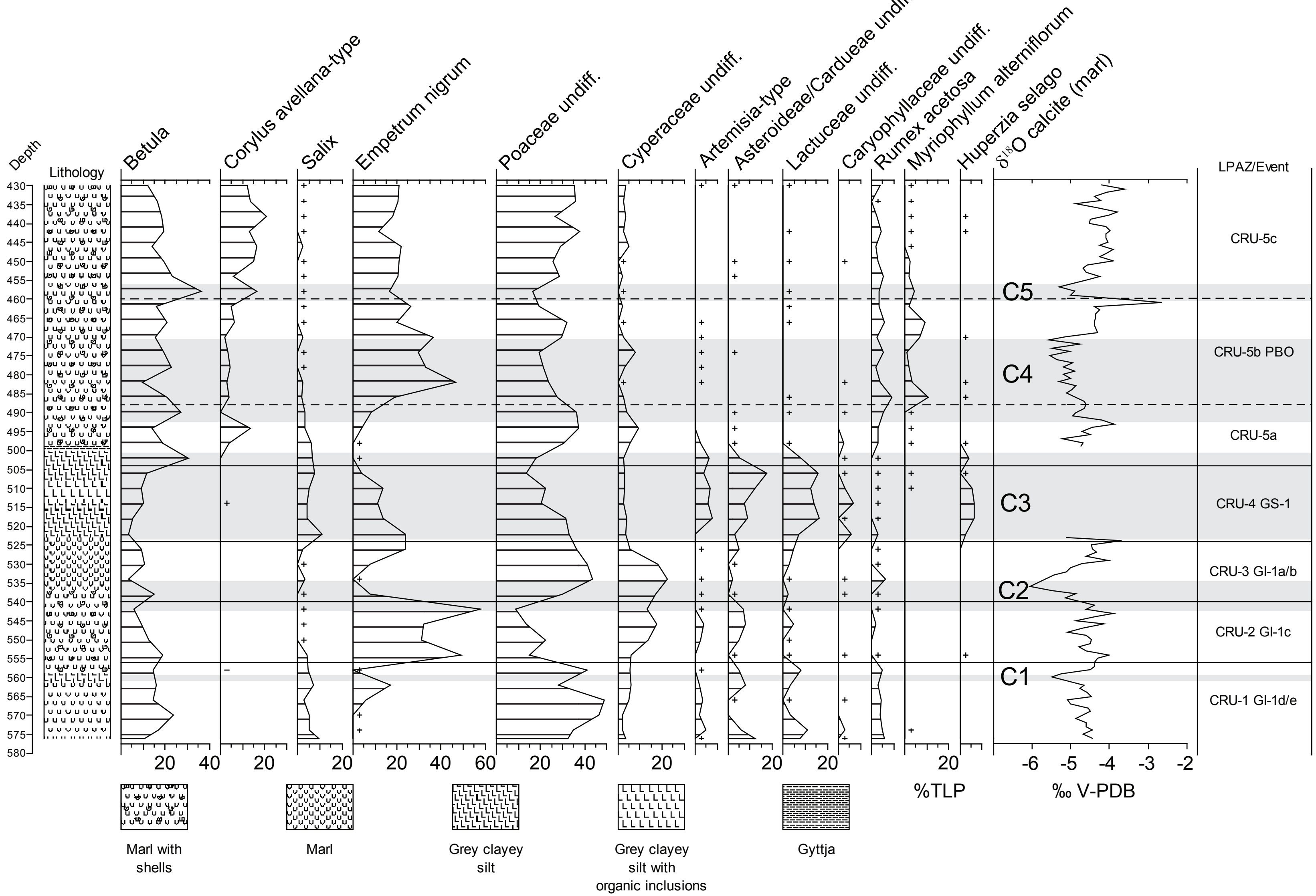

Depth (cm)
Lithology
Loss-on-ignition
Magnetic susceptibility
δ18O calcite (marl)
δ13C calcite (marl)
LPAZ/Event
Total mollusc numbers
C7
C6
C5
C4
C3
C2
C1
Saksun Ash
CRU-9
CRU-8
CRU-7
CRU-6b
CRU-6a
CRU-5c
CRU-5b PBO
CRU-5a
CRU-4 GS-1
CRU-3 GI-1a/b
CRU-2 GI-1c
CRU-1 GI-1d/e
%
Uncal
‰ V-PDB
‰ V-PDB
Gyttja
Marl with shells
Grey clayey silt
Grey clayey silt with organic inclusions
Marl

Figure 3



Depth (cm)
Lithology
Betula
Pinus sylvestris
Corylus avellana-type
Salix
Empetrum nigrum
Poaceae undiff.
Cyperaceae undiff.
Artemisia-type
Pteropsida (monolete) indet.
18O calcite (marl)
LPAZ/Event
CRU-9
CRU-8
CRU-7
CRU-6b
CRU-6a
CRU-5c
CRU-5b PBO
CRU-5a
CRU-4 GS-1
CRU-3 GI-1a/b
CRU-2 GI-1c
CRU-1 GI-1d/e
C7
C6
C5
C4
C3
C2
C1
%TLP
‰ V-PDB
Sedge peat
Marl with shells
Marl
Grey clayey silt
Gey clayey silt with organic inclusions
Gyttja

Figure 4



Depth
Lithology
Betula
Corylus avellana-type
Salix
Empetrum nigrum
Poaceae undiff.
Cyperaceae undiff.
Artemisia-type
Asteroideae/Cardueae undiff.
Lactuceae undiff.
Caryophyllaceae undiff.
Rumex acetosa
Myriophyllum alterniflorum
Huperzia selago
δ18O calcite (marl)
LPAZ/Event
CRU-5c
C5
CRU-5b PBO
C4
CRU-5a
C3
CRU-4 GS-1
CRU-3 GI-1a/b
C2
CRU-2 GI-1c
C1
CRU-1 GI-1d/e
%TLP
‰ V-PDB
Marl with shells
Marl
Grey clayey silt
Grey clayey silt with organic inclusions
Gyttja

Figure 5 colour

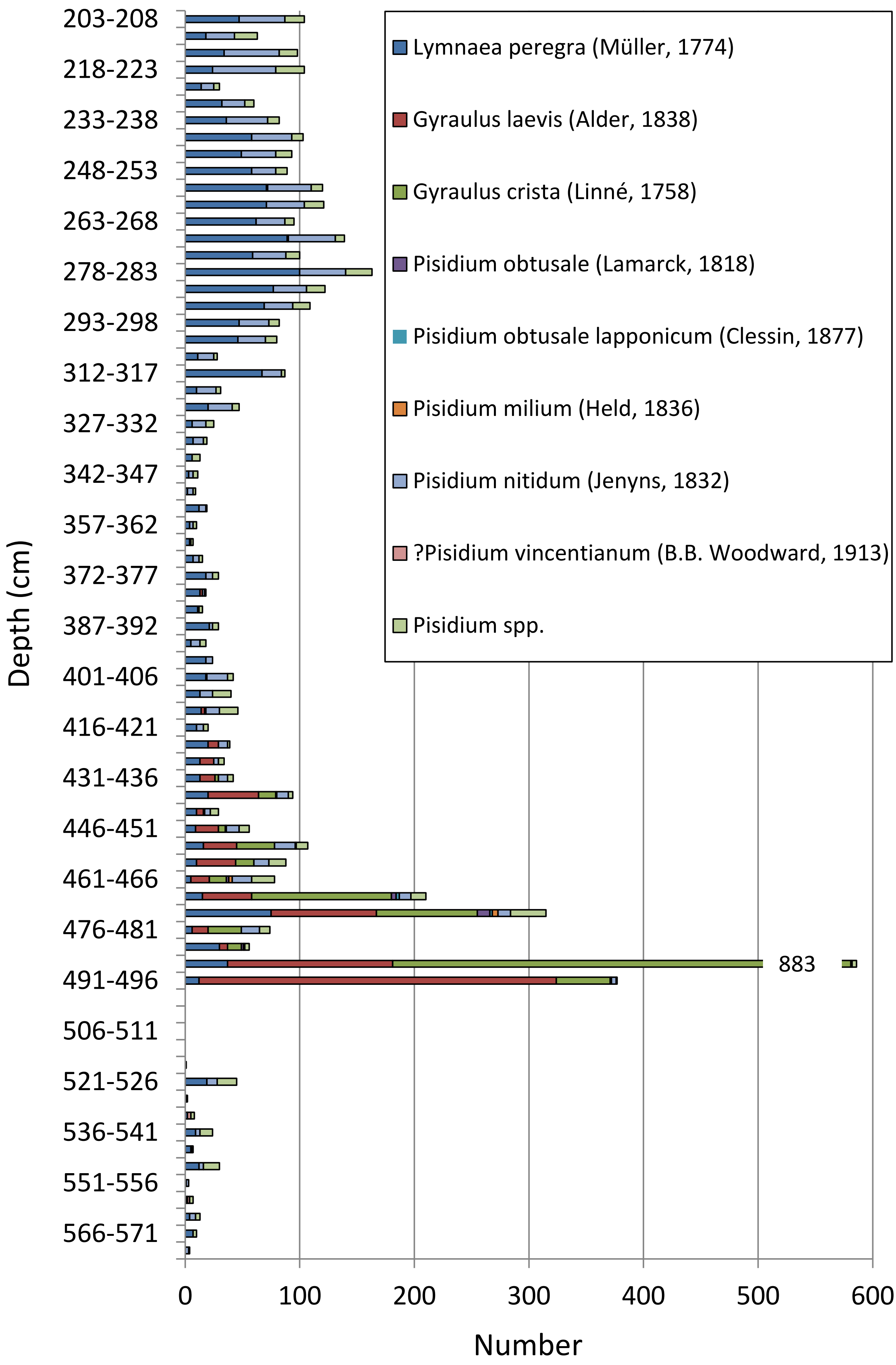


Figure 5 greyscale

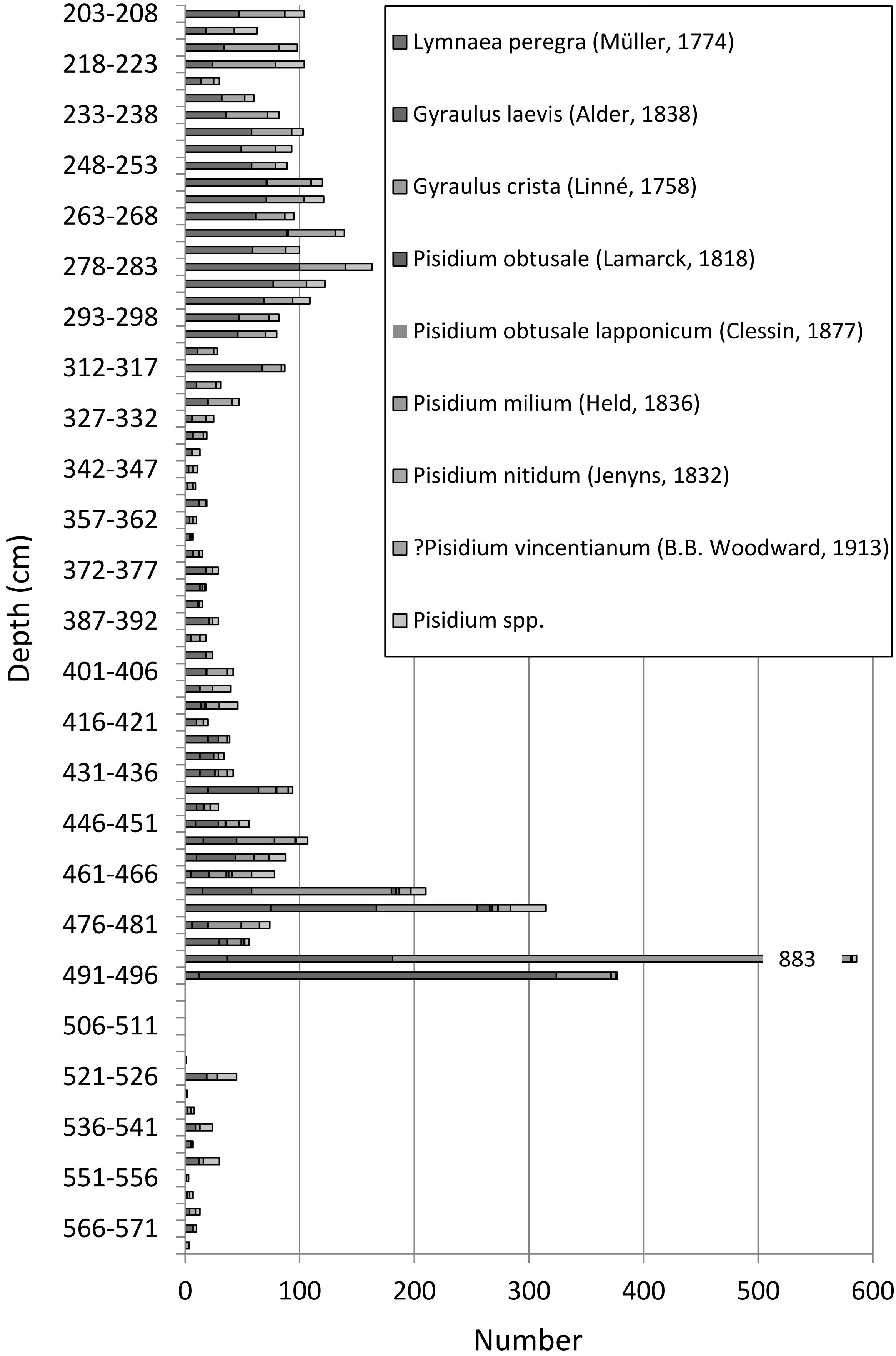
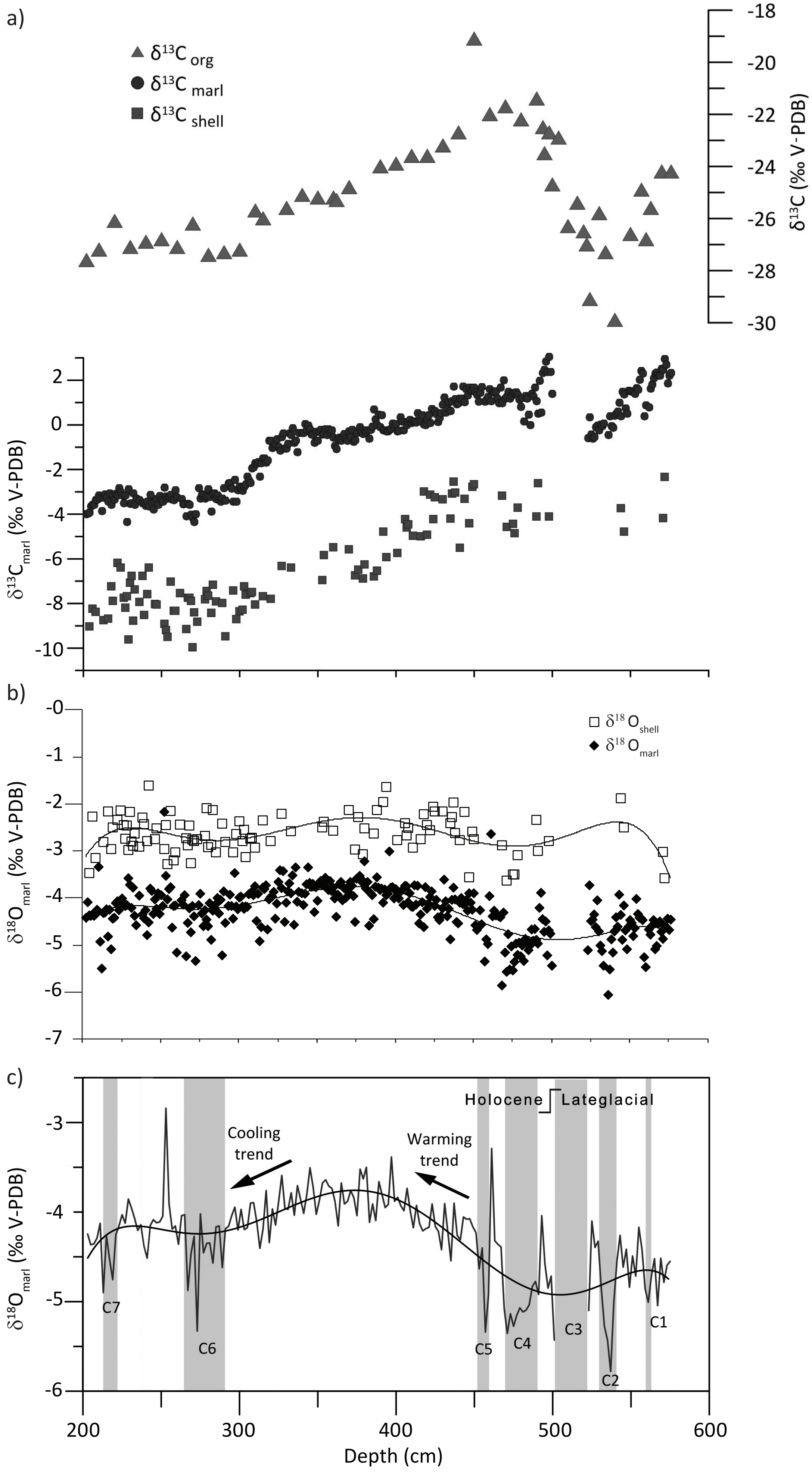
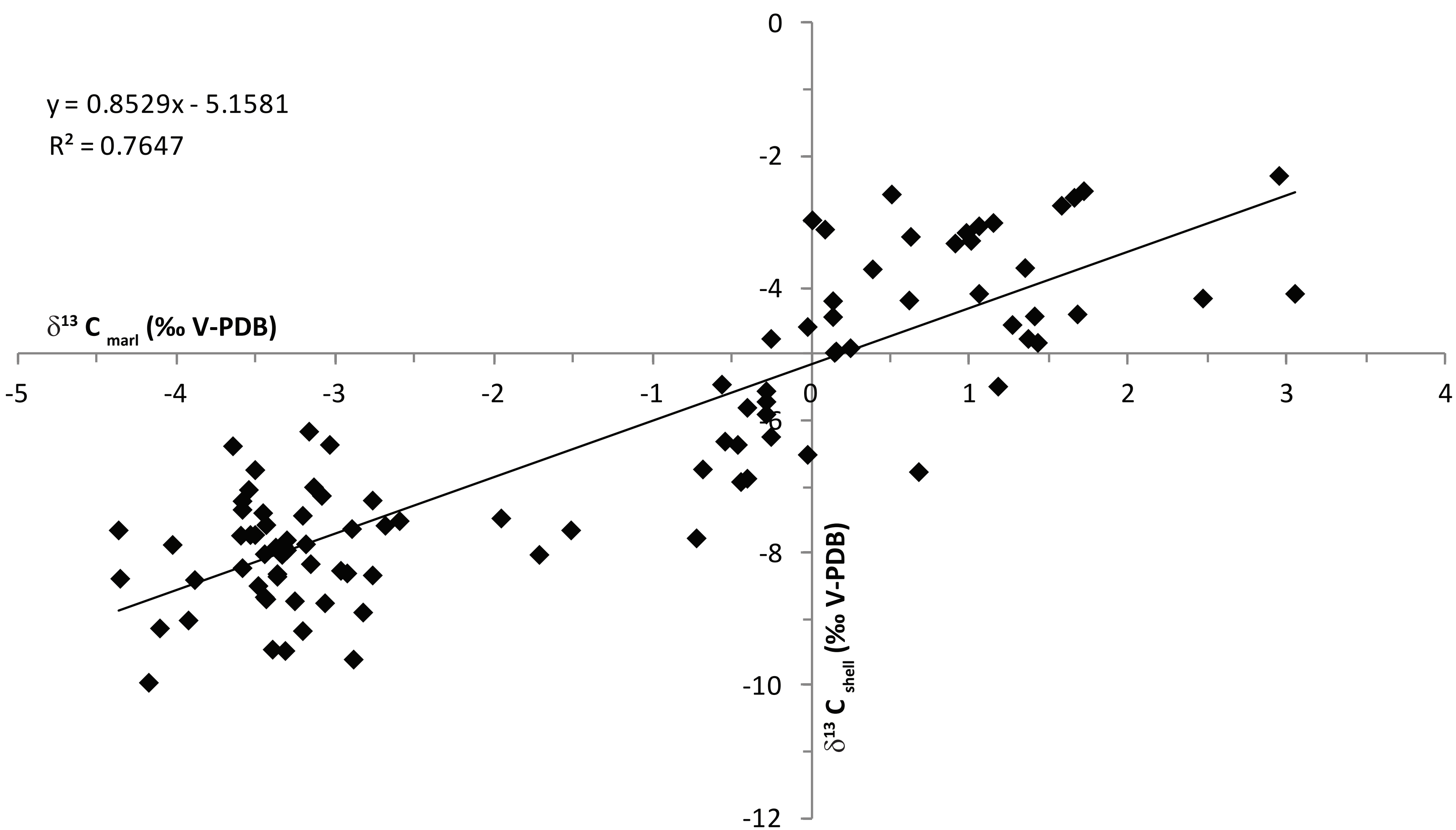


Figure 6



a)
δ13C org
δ13C marl
δ13C shell
δ13C (‰ V-PDB)
δ13Cmarl (‰ V-PDB)
b)
δ18 Oshell
δ18 Omarl
δ18Omarl (‰ V-PDB)
c)
Holocene
Lateglacial
Cooling trend
Warming trend
C7
C6
C5
C4
C3
C2
C1
δ18Omarl (‰ V-PDB)
Depth (cm)

Figure 7



y = 0.8529x - 5.1581
R² = 0.7647
δ13 C marl (‰ V-PDB)
δ13 C shell (‰ V-PDB)
0
-2
-4
-6
-8
-10
-12
-5
-4
-3
-2
-1
0
1
2
3
4

Figure 8

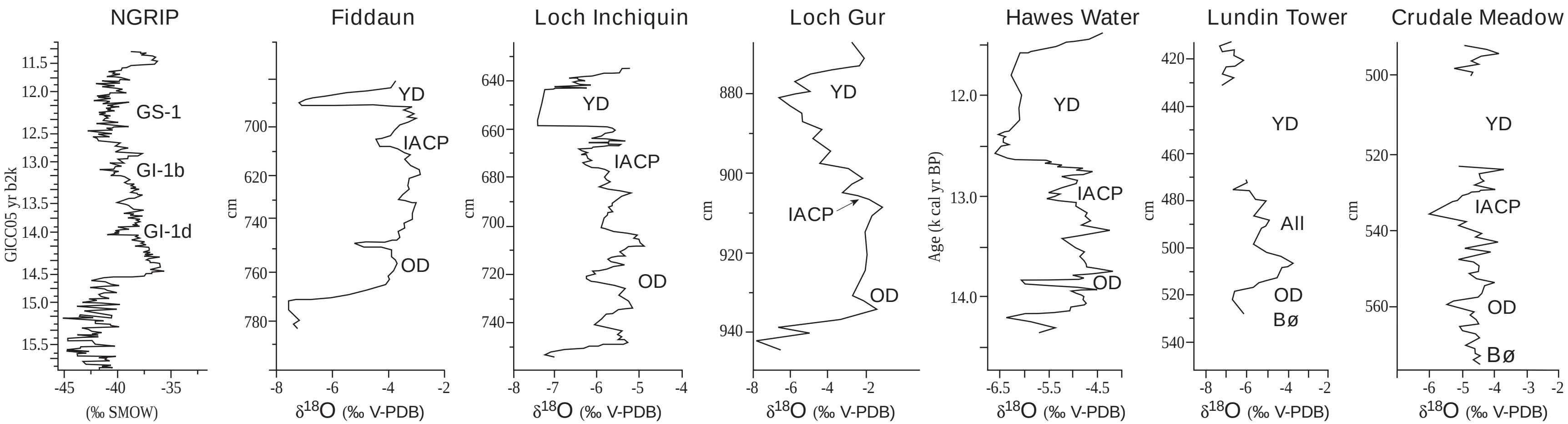
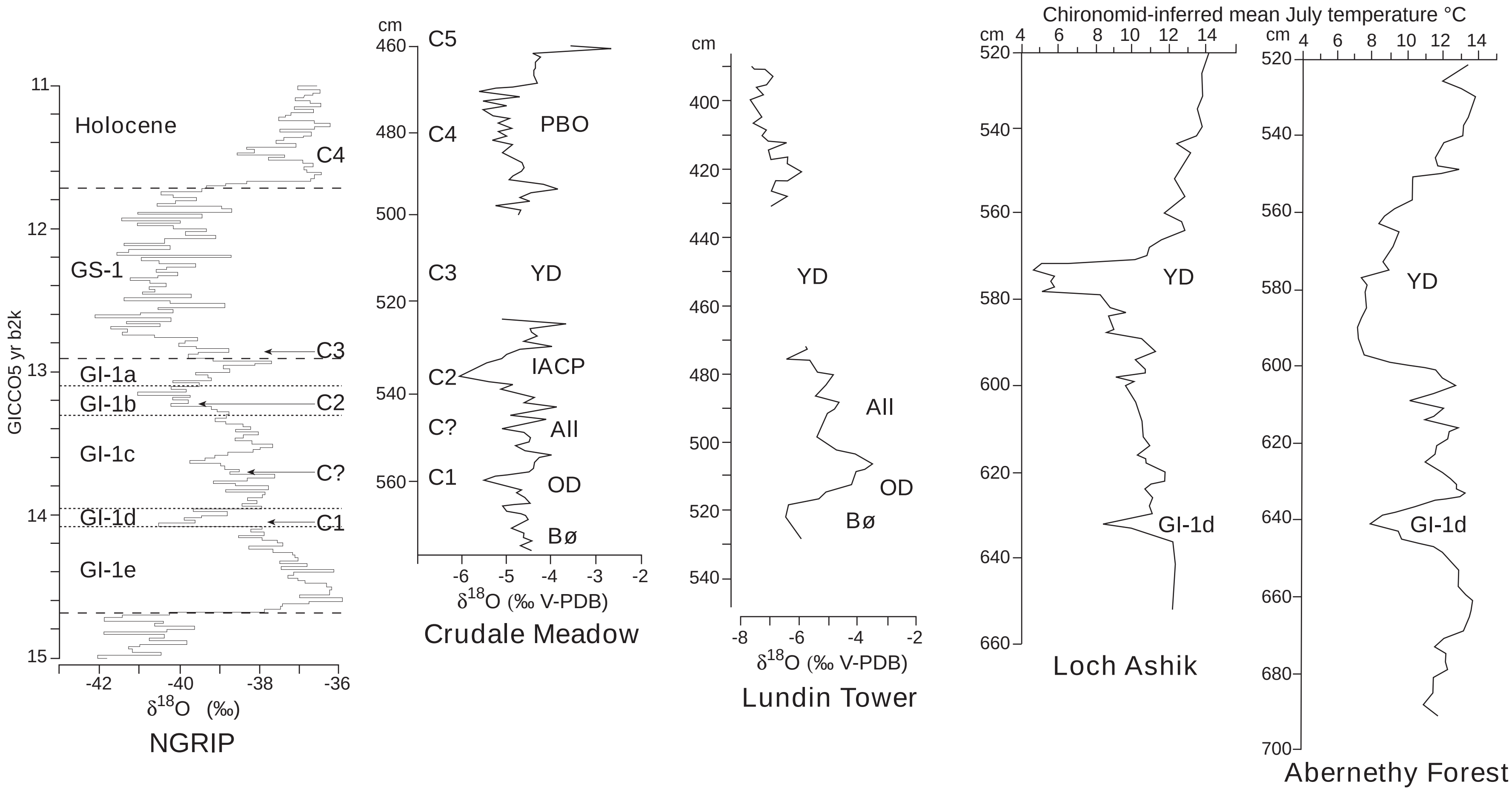
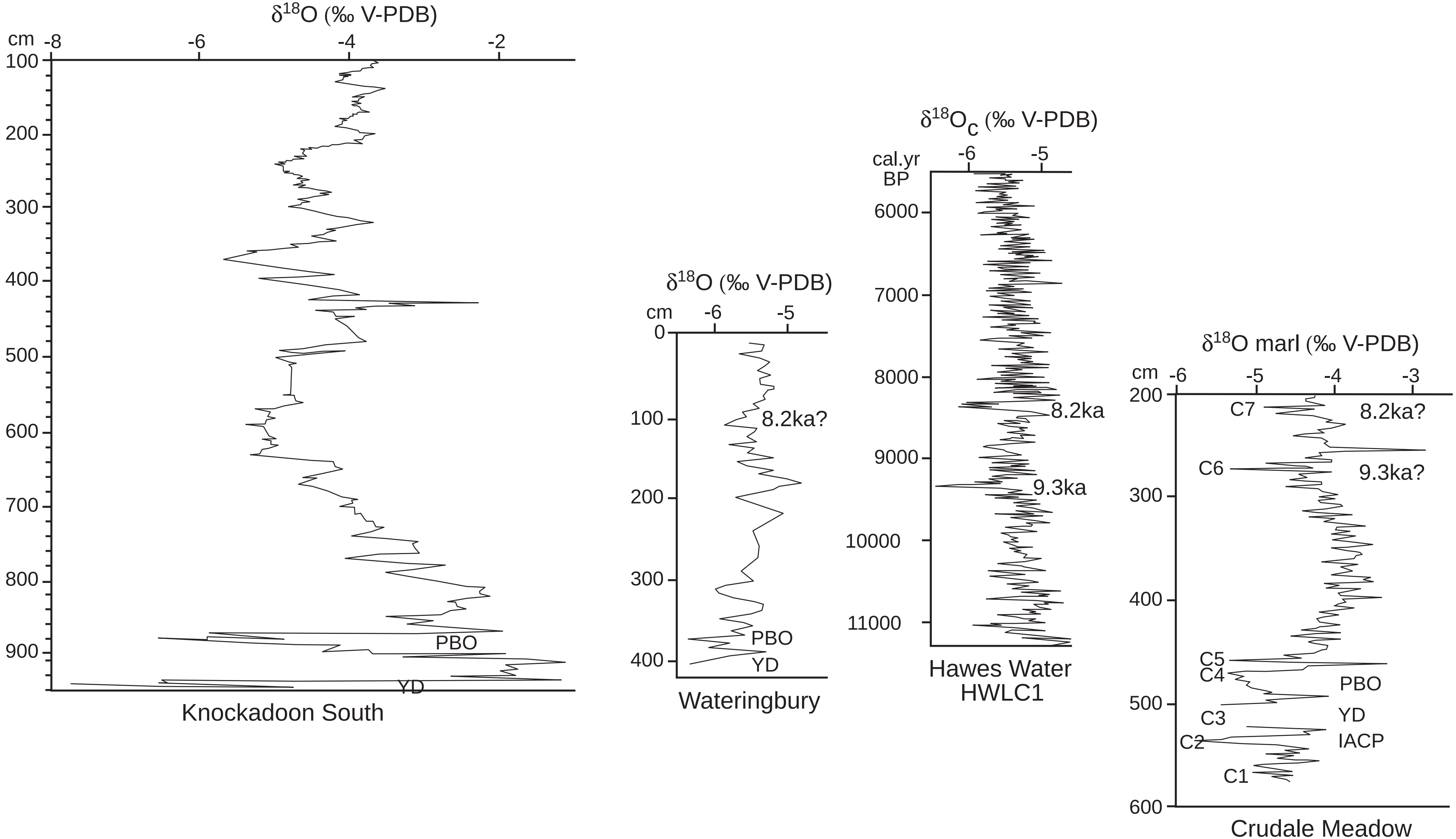

NGRIP
Fiddaun
Loch Inchiquin
Loch Gur
Hawes Water
Lundin Tower
Crudale Meadow
GICC05 yr b2k
(‰ SMOW)
GS-1
GI-1b
GI-1d
cm
Age (k cal yr BP)
δ18O (‰ V-PDB)
YD
IACP
OD
AII
Bø

Figure 9

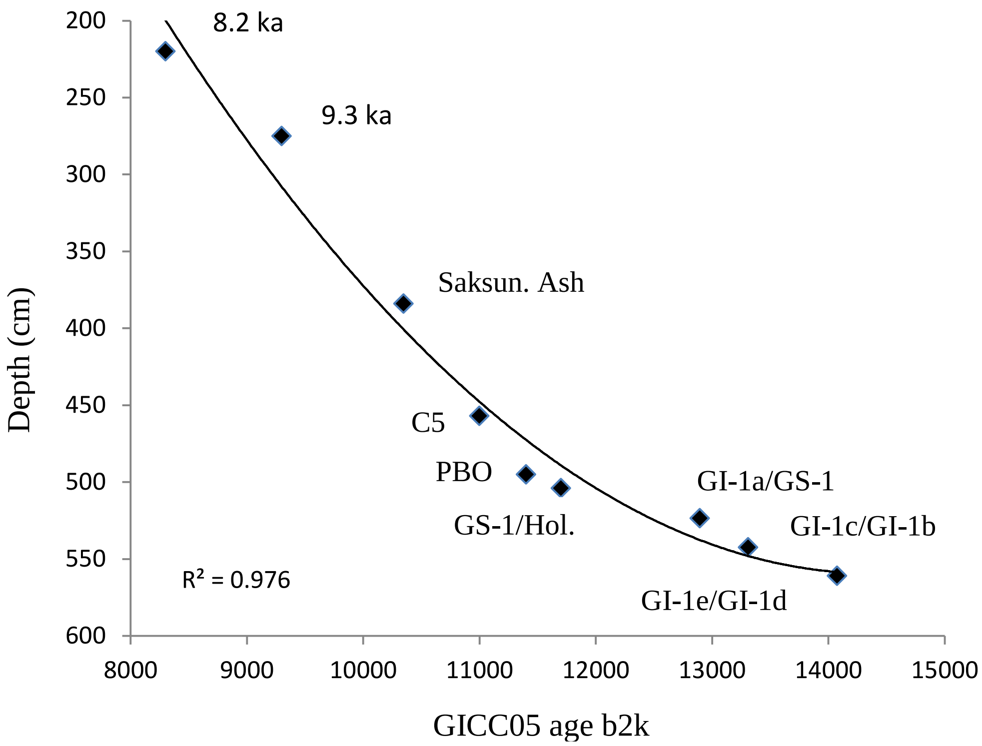


Holocene
GS-1
GI-1a
GI-1b
GI-1c
GI-1d
GI-1e
C4
C3
C2
C?
C1
GICCO5 yr b2k
δ18O (‰)
NGRIP
cm
C5
C4
C3
C2
C?
C1
PBO
YD
IACP
AII
OD
Bø
δ18O (‰ V-PDB)
Crudale Meadow
cm
YD
AII
OD
Bø
δ18O (‰ V-PDB)
Lundin Tower
Chironomid-inferred mean July temperature °C
cm
YD
GI-1d
Loch Ashik
cm
YD
GI-1d
Abernethy Forest



δ18O (‰ V-PDB)
cm
-8
-6
-4
-2
100
200
300
400
500
600
700
800
900
PBO
YD
Knockadoon South
δ18O (‰ V-PDB)
cm
-6
-5
0
100
200
300
400
8.2ka?
PBO
YD
Wateringbury
δ18Oc (‰ V-PDB)
cal.yr BP
-6
-5
6000
7000
8000
9000
10000
11000
8.2ka
9.3ka
Hawes Water HWLC1
δ18O marl (‰ V-PDB)
cm
-6
-5
-4
-3
200
300
400
500
600
C7
8.2ka?
C6
9.3ka?
C5
C4
PBO
C3
YD
C2
IACP
C1
Crudale Meadow

Figure 11



8.2 ka
9.3 ka
Saksun. Ash
C5
PBO
GS-1/Hol.
GI-1a/GS-1
GI-1c/GI-1b
GI-1e/GI-1d
R² = 0.976
Depth (cm)
GICC05 age b2k
200
250
300
350
400
450
500
550
600
8000
9000
10000
11000
12000
13000
14000
15000

Supplementary Data
Click here to download Supplementary Data: Supp. Inf. Figure 2.pdf

Supplementary Data

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