



Deglacial to postglacial palaeoenvironments of the Celtic Sea: lacustrine conditions versus a continuous marine sequence

Journal:	<i>Boreas</i>
Manuscript ID:	BOR-011-2013
Manuscript Type:	Original Article
Date Submitted by the Author:	21-Jan-2013
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Keywords:	Celtic Deep Basin, NW European continental shelf, sea-level, molluscs, microfossils, glacialmarine, glacialacustrine

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1 Deglacial to postglacial palaeoenvironments of the Celtic Sea: lacustrine conditions versus a
2 continuous marine sequence

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2
3 14 **Abstract**
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5 15 Recent work on the last glaciation of the British Isles has led to an improved understanding of the
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7 16 nature and timing of the retreat of the British-Irish Ice Sheet (BIIS) from its southern maximum (Isles
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9 17 of Scilly), northwards into the Celtic and Irish seas. However, the nature of the deglacial
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11 18 environments across the Celtic Sea shelf, the extent of subaerial exposure, and the existence (or
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13 19 otherwise) of a contiguous terrestrial linkage between Britain and Ireland following ice retreat
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15 20 remains ambiguous. Multiproxy research, based on the analysis of BGS vibrocores from the Celtic
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17 21 Deep Basin (CDB), seeks to address these issues. CDB cores exhibit a shell-rich upward fining
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19 22 sequence of Holocene marine sand above an erosional contact cut in laminated muds with
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21 23 infrequent limestones. Molluscs, *in situ* foraminifera, and marine diatoms are absent from the basal
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23 24 muds, but rare damaged freshwater diatoms and foraminiferal linings occur. Dinoflagellate cysts and
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25 25 non-pollen palynomorphs evidence diverse, environmentally incompatible floras with temperate,
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27 26 boreal, and Arctic glacimarine taxa co-occurring. Such multiproxy records can be interpreted as
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29 27 representing a retreating ice margin, with reworking of marine sediments into a lacustrine basin.
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31 28 Equally, the same record may be interpreted as recording similar conditions within a semi-enclosed
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33 29 marine embayment dominated by meltwater export and deposition of reworked microfossils. Since
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35 30 assemblages from these cores contrast markedly with proven glacimarine sequences from outside
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37 31 the CDB, a glacialacustrine interpretation is favoured for the laminated sequence, truncated by a Late
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39 32 Weichselian transgressive sequence fining upwards into fully marine conditions. Reworked rare
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41 33 intertidal molluscs from immediately above the regional unconformity provide a minimum date
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43 34 ~13.9 cal ka BP for commencement of widespread marine erosion. Though suggestive of
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45 35 glacialacustrine conditions, the exact nature and timing of laminated sediment deposition within the
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47 36 CDB, and the implications this has on (pen)insularity of Ireland following deglaciation, remain
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49 37 elusive.
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39 **Keywords:** Celtic Deep Basin, Northwest European continental shelf, sea-level, molluscs,
40 microfossils, glacimarine, glacialustrine

For Review Only

1. Introduction

Elucidating the nature and timing of deglaciation in the Celtic and Irish seas at the end of the Weichselian Cold Stage and their associated sea-level histories has significant implications for understanding the development of insularity between Britain and Ireland along with the concomitant effects that this had on the evolving biogeographies of these regions (Stuart 1977, 1995; Devoy 1986, 1995; Coxon & Waldren 1995). Much work has been conducted in the northern part of this area (Irish and Malin seas; e.g. Clark *et al.* 2004; Thomas *et al.* 2004; McCabe *et al.* 2005; Roberts *et al.* 2007; van Landeghem *et al.* 2009; Roberts *et al.* 2011) and increasingly along the coasts of the Celtic Sea (e.g. Ó Cofaigh & Evans 2001a; Evans & Ó Cofaigh 2003; Hiemstra *et al.* 2006; McCarroll *et al.* 2010). However, important evidence for ice occupation and subsequent deglaciation at the southern limit of the British-Irish Ice Sheet (BIIS) and the implications this has for ice sheet dynamics and stability, and sediment delivery to the North Atlantic, lies offshore on the floor of the Celtic Sea (Pantin & Evans 1984; Evans 1990; Scourse *et al.* 1990, 1991, 2009; Tappin *et al.* 1994). Furthermore, establishing the deglacial to postglacial marine evolution of shelf seas is important given the role such systems play in global ocean productivity (Wollast 1991; Austin & Scourse 1997; Marret & Scourse 2002; Scourse *et al.* 2002; Scourse in press), as well as CO₂ ocean-atmosphere exchange during climate transitions (Rippeth *et al.* 2008).

Despite a number of observational and glacial isostatic adjustment (GIA) modelling simulations, the extent to which the shallow continental shelf of the Celtic Sea was exposed subaerially following deglaciation and during the Holocene marine transgression remains unresolved (Devoy 1983, 1985, 1995; Lambeck 1995, 1996; Wingfield 1995; Lambeck & Purcell 2001; Peltier *et al.* 2002). Fundamentally, did deglaciation occur in a tidewater setting followed by uninterrupted postglacial marine conditions, or was there a contiguous land-linkage existed between Britain and Ireland, either during, or subsequent to, deglaciation as suggested by GIA simulations (e.g. Lambeck 1995)?

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3 66 To address this question, and to elucidate the deglacial environments of the Celtic Sea,
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5 67 twelve previously-collected British Geological Survey (BGS) vibrocores (Fig. 1; Table 1) were analysed
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7 68 using a multiproxy litho- and biostratigraphic approach (organic-walled, siliceous, and calcareous
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9 69 microfossils; calcareous macrofossils). A chronological framework is provided by 42 accelerator mass
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11 70 spectrometry (AMS) radiocarbon dates (Table 2). Cores were selected to sample a SW-NE transect
12
13 71 across the Celtic Deep Basin (CDB; a bathymetric low on the Celtic Sea shelf; Fig. 1), ideally
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15 72 positioned to test the two competing hypotheses.
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21 74 **2. Regional Setting**

22 75 Lying to the south of St George's Channel, between southern Ireland, southwest Britain, and
23
24 76 northwest France, and extending out to the shelf break (Fig. 1), the Celtic Sea occupies some
25
26 77 186,000 km² of the Northwest European Continental Shelf. Water depths range from 60 to 200 m,
27
28 78 the central and northern shelf being generally flat and featureless, the result of bevelling of the inner
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30 79 shelf due to eustatic transgressions and regressions under cool temperate and low arctic Pleistocene
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32 80 conditions (Tappin *et al.* 1994). In the northern part of the study area, the SW-trending Celtic Deep
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34 81 Basin (CDB) attains water depths of up to 130 m, while the Haig Fras granite batholith shoals to
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36 82 within 40 m of modern sea-level in the central Celtic Sea. The CDB follows the Caledonian structural
37
38 83 trend present in the northern Celtic Sea, modified by glacial erosion since the middle-Pleistocene
39
40 84 (Tappin *et al.* 1994). The northern and central areas also feature linear enclosed deeps <5 km wide
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42 85 and 30 km long, considered to represent jökulhlaup erosion (Wingfield 1989, 1990) or subglacial
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44 86 tunnel-valley formation (Jeffery 1990). In the southern Celtic Sea, the smooth profile of the shelf is
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46 87 interrupted by numerous, generally NW-SE orientated, parallel linear sand ridges (LSR) attaining
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48 88 heights of up to 60 m, lengths of >200 km, and wavelengths of 10-15 km. The precise origin of these
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50 89 features is disputed (e.g. Praeg *et al.* 2011) though formation under conditions of lowered eustatic
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52 90 sea-level and high tidal bed shear stress during the Pleistocene appears to be the developing
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3 91 consensus (Bouysse *et al.* 1976; Pantin & Evans 1984; Reynaud *et al.* 1985, 1999, 2003; Belderson *et*
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5 92 *al.* 1986; Marsset *et al.* 1999; Uehara *et al.* 2006; Scourse *et al.* 2009).

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8 93 The northern and central parts of the study area were occupied by grounded glacial ice of
9
10 94 the Irish Sea Ice Stream (ISIS) during the Last Glacial Maximum (LGM), marking the southernmost
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12 95 extent of the British – Irish Ice Sheet (BIIS) during this interval (Scourse & Furze 2001; Sejrup *et al.*
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14 96 2005; Ó Cofaigh & Evans 2007; Chiverrell & Thomas 2010; Fig. 1). Thermoluminescence, radiocarbon,
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16 97 and cosmogenic radionuclide exposure dating of glacially-influenced sediments and landforms in the
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18 98 Isles of Scilly suggests ice retreat from the northern margins of the archipelago ~21 ka BP with rapid
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20 99 retreat northwards into the central Irish Sea (Anglesey) under mostly subaqueous conditions in
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22 100 <3000 years (Scourse 1991a, b; Scourse & Furze 2001; Hiemstra *et al.* 2006; van Landeghem *et al.*
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24 101 2009; McCarroll *et al.* 2010). The maximal position of the ice margin west of Scilly remains poorly
25
26 102 constrained, but the occurrence of over-consolidated diamicton beneath laminated waterlain
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28 103 glacial sediments in BGS core 49/-09/044 was used by Scourse *et al.* (1990, 1991; core CS-3 this
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30 104 study) to infer that the southern limit of the grounded ISIS ice lay near to, but south of, this location
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32 105 (Fig. 1) terminating in a tidewater margin around 49°30'N at approximately the modern 135 m
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34 106 isobath. Scourse *et al.* (1990, 1991; and Scourse & Furze 2001) suggest the rapid surging advance of
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36 107 a thin (~100 m) ice lobe across the northern and central Celtic Sea from the BIIS's previously
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38 108 accepted southern margin just south the CDB (Fig. 1). Such thin surging behaviour is supported by
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40 109 evidence from the SE coast of Ireland (Ó Cofaigh & Evans 2001a, b), ice-streaming in the Irish Sea
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42 110 (Roberts *et al.* 2007), and limited glacial erosion present in the northern Isles of Scilly (Hiemstra *et al.*
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44 111 2006; McCarroll *et al.* 2010). The ISIS tidewater terminus in the central Celtic Sea is considered a
45
46 112 significant sediment source for the formation of LSRs, though the exact relationship between LSR
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48 113 development and ice sheet behaviour remains equivocal (Reynaud *et al.* 1985, 1999, 2003; Marsset
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50 114 *et al.* 1999; Scourse *et al.* 2009). Nonetheless, palaeotidal modelling indicates megatidal amplitudes
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52 115 in the Celtic Sea over the deglacial interval with high bed stress leading to significant sediment
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54 116 transport (Scourse *et al.* 2009).

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3 117 Late Pleistocene sediments of the Celtic and Irish seas include extensive subglacial and
4
5 118 waterlain facies associated with Weichselian ISIS advance and retreat. Multiple tabular Quaternary
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7 119 glacial units underlie much of the northern study area with the Upper Till Member of the
8
9 120 Cardigan Bay Formation – a stiff clayey diamicton with infrequent shell debris – being interpreted as
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11 121 a late Weichselian lodgement till (Tappin *et al.* 1994). This is considered to be correlative with the
12
13 122 late Devensian Scilly Till (Scourse 1991a). The late Weichselian to early Holocene Western Irish Sea
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15 123 Formation overlies the Cardigan Bay Formation, composed primarily of two facies within the
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17 124 northern Celtic Sea: the Chaotic and Mud facies. The Chaotic Facies is a sandy gravel deposit
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19 125 containing abraded shell debris and is interpreted as a glacial marine or glacial lacustrine ice-proximal
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21 126 deglacial unit, whereas the Mud Facies is represented by tabular stratified units of black to grey-
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23 127 green shelly silt with sparse small dropstones and microbiota indicative of cold distal glacial marine
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25 128 conditions (Tappin *et al.* 1994). Holocene seabed sediments in the northern region are represented
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27 129 by the Surface Sands Formation (Pantin 1977, 1978) divided into an upper member - SL1 (modern
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29 130 marine), conformably overlying the lower SL2 member. SL2 is considered to be a transgressive facies
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31 131 comprising a basal gravel lag and a rich shallow-water temperate fauna. It rests unconformably on,
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33 132 and truncates, underlying Quaternary and pre-Quaternary sediments.
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38 133 Sequences from the central and southern region are more complex. The Melville Formation
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40 134 (Evans 1990) comprises late Weichselian to earliest Holocene shallow marine sands with minor shell
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42 135 and gravel beds and occasional muddy horizons. It is in this formation that the LSR are developed,
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44 136 but it also includes sporadic and discontinuous massive and laminated glacial deposits. Core VE
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46 137 49/-09/044 (CS-3 this study), recovered from the flank of one such ridge (Scourse *et al.* 1990, 1991;
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48 138 this study) shows a laminated waterlain mud conformably overlying a stiff diamicton and was used
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50 139 to help constrain the SW extent of the ISIS. The exact stratigraphic relationship between these
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52 140 sediments and those shelly sands and gravels comprising the LSR from which it was taken remains
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54 141 unclear (Reynaud *et al.* 2003; Marsset *et al.* 1999; Scourse *et al.* 2009; Praeg *et al.* 2011).
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56 142 Nonetheless, Scourse *et al.* (1990, 1991) defined two glacial components of the Melville
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3 143 Formation: Facies A, the Melville Till, an overconsolidated mud with abundant fine gravel and
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5 144 pebbles and an absent, or infrequent, reworked temperate and arctic macrobiota; and Facies B, the
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7 145 Melville Laminated Clay, conformably overlying Facies A, being a plastic laminated mud with a rich *in*
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9 146 *situ* ostracod and exclusively arctic foraminifera. Whereas Facies A, the Melville Till is interpreted as
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11 147 a lodgement till (= Cardigan Bay Formation Upper Till Member in the northern region), Facies B, the
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13 148 Melville Laminated Clay is considered a distal glacial marine facies deposited under shallow quiescent
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15 149 conditions (Scourse *et al.* 1990, 1991). Both have been interpreted to correlate with late Devensian
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17 150 glacial deposits on the Isles of Scilly. However, whilst these interpretations may hold true for the
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19 151 southern Celtic Sea, it should be noted that no paleontological investigation has been undertaken on
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21 152 the critical core VE 49/-09/044 (CS-3) where both facies occur together. Surface sediments in the
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23 153 southern region strongly resemble those in the northern sector, Layers A and B of Evans (1990)
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25 154 being equivalent to SL1 and SL2 (respectively) of the Surface Sands Formation of Pantin (1977,
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27 155 1978).

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32 156 It is these units of the Melville, Cardigan Bay, Western Irish Sea, and Surface Sands
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34 157 formations that form the basis of this study; the glacial marine or glacial lacustrine nature of the Melville
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36 158 Laminated Clay/Cardigan Bay Mud Facies being critical in testing the hypothesis of a subaerially-
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38 159 exposed landbridge between Britain and Ireland, south of the CDB, during and subsequent to
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40 160 deglaciation.

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44 162 **3. Materials and Methods**

45 163 **3.1 Core materials and lithostratigraphy**

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49 164 Twelve vibrocores collected by the BGS were selected for the present study, constituting a SW-NE
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51 165 transect across the CDB (cores CD-1 to -9) and adjacent shelf (cores CS-1 to -3; Fig.1; Table 1), to
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53 166 provide a representative regional modern-to-deglacial stratigraphy (Fig. 2). Cores prove the Western
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55 167 Irish Sea Formation Mud Facies (Tappin *et al.* 1994) in the northern and central Celtic Sea and the till
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57 168 and laminated clay members of the Melville Formation (Pantin & Evans 1984; Scourse *et al.* 1990,
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3 169 1991) in the central/southern region. Cores were described lithostratigraphically and subsampled for
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5 170 grain size and macro- and microfossils. Grain size analysis comprised oven-drying at low temperature
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7 171 (45°C), weighing, wet-sieving at 63 µm, oven-drying (45°C), and subsequent dry-sieving at 63, 125,
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9 172 250, 500, and 1000 µm, and where necessary at 2, 4, and 8 mm to calculate % grain size frequency.
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11 173 Due to the variable degree of desiccation since collection, initial wet weights, water content, and
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13 174 shear strength could not be assessed. Descriptions of two cores (VE 51/-07/199 = CD-7 and VE 49/-
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15 175 09/044 = CS-3) examined for this study have been previously published, including lithostratigraphy
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17 176 and (predominantly) Holocene micropalaeontology for CD-7 (Scourse & Austin 1994; Austin &
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19 177 Scourse 1997; Scourse *et al.* 2002; Marret *et al.* 2004) and lithostratigraphy for CS-3 (Scourse *et al.*
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21 178 1990, 1991). Stratigraphic units are numbered from the sea-bed surface down, descriptions being
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23 179 given with the deepest first.
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29 181 **3.2 Calcareous macro- and microfossils**

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31 182 Samples investigated for calcareous macrofossils (e.g. molluscs) and microfossils (calcareous
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33 183 foraminifera, ostracods) were processed as for grain size analysis (see §3.1). Sample fractions >500
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35 184 µm for macrofossil investigation were examined visually and under low-power microscopy. In the
36
37 185 majority of samples, all identifiable calcareous macrofossils were picked out. In particularly rich
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39 186 samples, materials were split by micro-splitter. All macrofossil abundances are given as individuals
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41 187 per 100 g dry sediment (ind./100 g). Macrofossils were identified using Tesch (1947), Tebble (1976),
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43 188 Thompson & Brown (1976), Graham (1988), and Hayward & Ryland (1998). Molluscan nomenclature
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45 189 follows Appletans *et al.* (2011). Calcareous microfossils were noted from subsamples from
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47 190 lithostratigraphic units correlated with the laminated clay and till members of the Melville Formation
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49 191 in an attempt to determine the lacustrine versus marine nature of this unit. Calcareous microfossils
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51 192 were extremely rare (1-2 individuals per dry g of sediment = ind./g), highly abraded, damaged, and
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53 193 generally not identifiable to species level.
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3 195 **3.3 Organic-walled and siliceous microfossils**
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5 196 Organic-walled microfossils (dinoflagellate cysts = dinocysts, other non-pollen palynomorphs = NPPs,
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7 197 pollen) were examined from laminated clay and till members of the Melville Formation (and
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9 198 Western Irish Sea Formation mud facies). Samples were weighed, oven-dried at low temperature
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11 199 (45°C), weighed, and wet-sieved at 10 µm following the addition of *Lycopodium clavatum* tablets for
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13 200 calculation of concentrations (ind./g). The >10 µm fraction was processed for organic-walled
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15 201 microfossils (Marret & Zonneveld 2003), encompassing repeated treatments with 10% cold HCl and
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17 202 cold 38% HF acids. Residues were mounted in safranin-stained glycerine jelly and examined
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19 203 systematically under high-power microscopy (x 400). A minimum of 100 dinocysts was counted from
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21 204 each sample wherever possible and any co-occurring NPPs and pollen were noted. Species
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23 205 identifications primarily follow Rochon *et al.* (1999) and Head *et al.* (2005). Most dinocysts were
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25 206 identified to species level, with the exception of *Brigantedinium* spp. (grouped *Brigantedinium*
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27 207 *cariacoense*, *Brigantedinium simplex*, *Brigantedinium* sp. - cysts without visible/present
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29 208 archaeopyle), *Protopteridinium* spp. (folded/torn round brown cysts), and *Spiniferites* spp. Within
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31 209 *Spiniferites*, an unknown form, *Spiniferites* sp. 1 was distinguished, whereas spiny brown cysts were
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33 210 tentatively identified as *Islandinium minutum*?.
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37 211 Several samples from cores CS-3 and CD-1,-2, and-3 were analyzed for diatoms. These were
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39 212 prepared in accordance with standard protocols (Battarbee *et al.* 2001). Sediments were processed
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41 213 by adding 30% H₂O₂ and heating gently to oxidize organics. Sample dilutions were mounted on
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43 214 microscope slides using Naphrax™. Identification and enumeration of diatoms was carried out using
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45 215 high-power microscopy (1000x). Taxonomic identification follows Krammer & Lange-Bertalot (1991,
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47 216 1997). Due to the estimated low amounts of carbonate in these sediments, HCl was not used in
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49 217 sample preparation. Consequently, infrequent calcareous nanofossils were apparent in examined
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51 218 samples. Furthermore, processing did not remove all organic matter, as evidenced by foraminiferal
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53 219 linings and freshwater algae present in some samples.
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221 **3.4 Radiocarbon assay**

222 Forty-two marine molluscan samples were submitted for radiocarbon dating in three discrete
223 batches (Table 2); two batches to the UK Natural Environmental Research Council Radiocarbon
224 Facility – University of Arizona AMS Laboratory (lab code AA), and one batch to the National Ocean
225 Sciences AMS Facility at Woods Hole Oceanographic Institution (lab code OS). Dates were calibrated
226 using CALIB 6.0 (Stuiver *et al.* 2010) based on the marine calibration dataset Marine09 (Reimer *et al.*
227 2009). A ΔR value of -33 ± 93 years (Reimer *et al.* 2002) for the Late Holocene Celtic Sea and adjacent
228 Irish coast was applied to all dates. Dates are reported in the text as calibrated median probability
229 ages before present (AD 1950).

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231 **4. Results**

232 Three regionally-extensive lithostratigraphic units were identified from the examined cores, being
233 encountered throughout the study area. This typical tripartite stratigraphy (Fig. 2) is characterised by
234 a basal sequence of massive to laminated muds (Unit III) marked by small infrequent lonestones and
235 an apparent absence of molluscs. Truncated by a regionally extensive unconformity, this is overlain
236 by bio- and litho-clastic gravels (Unit II) beneath silty sands (Unit I), rich in whole and comminuted
237 marine macrofossils, and comprising the modern seafloor sediments. Variations on this simple
238 tripartite sequence were encountered in several cores (Fig. 2): CD-2 showed an additional unit of
239 gravelly sand (Unit IIIa) beneath Unit III laminated muds; CD-3 exhibited thin interbeds of organic-
240 rich laminated silt (Unit IIa) within a particularly extensive sandy gravel sequence considered to be
241 Unit II; CD-8 proved the upper two units (units I and II) underlain by massive shelly and gravelly
242 sands (Unit IIb); CS-3 displayed a massive well-consolidated stony diamicton (Unit IIIb) beneath Unit
243 III.

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245 **4.1 Unit III (laminated-massive mud)**

246 **4.1.1 Lithostratigraphy**

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3 247 Remarkably uniform across the study area, Unit III is a greyish brown, faintly to well laminated
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5 248 (infrequently massive), sandy to silty clay (Fig. 2). Where well developed, laminae form couplets of
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7 249 silt (to fine sand) and clay approximately 1-2 mm in thickness. Small (typically <30 mm) subangular
8
9 250 to rounded lithic clasts – lonestones – occur in variable concentrations in all cores (CD-8 and -9 do
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11 251 not recover this unit). Unit III varies in thickness from 94 to 475 cm from the upper erosional contact
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13 252 with Unit II to the core base. In only two instances (CD-2, CS-3) are sediments beneath typical Unit III
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15 253 muds penetrated (Fig. 2). In CD-2, a poorly sorted gravelly sand containing clay and lithic clasts and
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17 254 highly abraded bioclastic debris (Unit IIIa) occurs below the laminated mud. In CS-3, a 15 cm thick
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19 255 subunit (Unit IIIb) of overconsolidated clast-rich diamicton, previously described by Scourse *et al.*
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21 256 (1990) as a lodgement till, occurs at the core base beneath Unit III.
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26 257 27 258 4.1.2 Biostratigraphy

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29 259 In stark contrast to units I and II, Unit III is marked by the near-absence of identifiable macrofossils
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31 260 (Fig. 3). Notably, macrofossil fauna identified from the very top of Unit III (CD-5, 070-075 cm)
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33 261 represent burrow infill and interstitial breccia matrix derived from the overlying Unit II sandy gravels
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35 262 and is thus not considered *in situ*. CD-2 237-242 cm contained exceptionally rare identifiable
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37 263 macrofossil materials (one heavily abraded bryzoan fragment, two balanoid plates) considered not
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39 264 to be contamination from overlying units. In CD-2, a poorly sorted gravelly sand (Unit IIIa), occurring
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41 265 beneath Unit III, contained infrequent extremely abraded bioclastic fragments (0.44 ind./100 g;
42
43 266 including barnacle plates, bryzoan fragments, extremely rare echinoid spines, and a single abraded
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45 267 valve of *Nucula* sp.; no microfossil assay conducted). No macrofossils were evident in the basal
46
47 268 diamicton (Unit IIIb) from CS-3, however Scourse *et al.* (1990) assign this subunit to Facies A of the
48
49 269 Melville Till from which rare abraded, broken temperate and arctic macrofossils have been recorded
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51 270 (*Hiatella* sp. and balanoid plates, considered reworked; Scourse *et al.* 1990).
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55 271 Four Unit III samples from CDB cores CD-1, -2, and -3 and the open shelf core CS-3 were
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57 272 investigated for siliceous microfossils. These showed exceptionally rare freshwater diatoms, though
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3 273 observed numbers (1-12 valves) were insufficient to warrant meaningful statistical counts. One
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5 274 sample (CD-3) proved barren of diatoms. The common planktonic diatom *Aulacoseira distans*, and
6
7 275 the widely distributed and exclusively freshwater epipellic diatom *Neidium ampliatum* was observed
8
9 276 with good preservation, whereas a large fragment of the epipellic freshwater diatom *Pinnularia*
10
11 277 *microstauron* occurred in one sample (CD-1). No marine diatoms, in whole or fragments, were
12
13 278 observed in any of the investigated samples. Other microfossils present in diatom slides include
14
15 279 siliceous stomatocyst of the freshwater algae Chrysophyceae, and organic microfossils such as
16
17 280 *Botryococcus*, *Pediastrum*, and foraminiferal linings. External calcified plates (coccoliths) from
18
19 281 calcareous marine nanoplankton (Prymnesiophyta) were observed in all samples consistent with
20
21 282 their reworking from Cretaceous chalk outcrops on the floor of the Celtic and Irish seas (Evans 1990;
22
23 283 Scourse *et al.* 1990; Tappin *et al.* 1994). No meaningful differences in the diatom assemblages were
24
25 284 evidenced between the CDB and open shelf Unit III samples.
26
27

28
29 285 Dinocyst abundances are low in Unit III, with most samples exhibiting concentrations <650
30
31 286 cysts/g and diversities of <10 species (max. 18 species; Fig. 4A). Only one sample (CD-7) shows
32
33 287 conspicuously high dinocyst abundances, reaching ~2780 cysts/g. Dinocyst assemblages are
34
35 288 overwhelmingly dominated by phototrophic taxa, in particular *Bitectatodinium tepikiense* whose
36
37 289 relative abundances typically reach >40%. This species is accompanied by lesser proportions of
38
39 290 *Lingulodinium macherophorum* and *Operculodinium centrocarpum*, as well as *Spiniferites* spp.
40
41 291 (*Spiniferites lazus*, *Spiniferites* cf. *membranaceus*) and *Islandinium minutum*?. *Brigantedinium* spp.
42
43 292 and other protoperidinoid dinocysts are much reduced and only present in noticeable amounts in
44
45 293 samples from CS-3. Absolute abundances of acritarchs and zoomorphs are much reduced in Unit III
46
47 294 (Fig. 4B). Foraminiferal lining concentrations are particularly low (<35 ind./g), apart from one,
48
49 295 potentially contaminated sample at the Unit II/Unit III transition (CD-2, 190 cm). Invertebrate eggs
50
51 296 and, in particular, invertebrate mouthparts are also much reduced or absent in Unit III. In contrast,
52
53 297 freshwater chlorophycean algae (*Botryococcus*, *Pediastrum*) are present in most Unit III samples in
54
55 298 variable amounts. *Corylus* pollen concentrations are low whereas *Pinus* pollen and *Pteridium* spores
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57
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1
2
3 299 show variable absolute abundances, sometimes outnumbering Unit II values (Fig. 4B).
4
5 300 Concentrations of pre-Quaternary materials such as dinocysts (e.g. *Achomosphaera andalousiensis*)
6
7 301 and spores are high. Unit III samples from the open shelf (CS-3) show similar dinocyst abundances
8
9 302 and assemblage structures to those from within the CDB, although *Brigantedinium* spp. are elevated
10
11 303 and *Spiniferites* spp. appear reduced (Fig. 4A). Three samples from the Unit IIIb basal diamicton in
12
13 304 CS-3 (Fig. 4A; Facies A - Melville Till of the Melville Formation *sensu* Scourse *et al.* 1990), below the
14
15 305 typical Unit III laminated muds, show extremely rare dinocysts or are barren. Where present,
16
17 306 dinocysts exhibit exceptionally low concentrations and diversities. Nonetheless, similar to typical
18
19 307 Unit III deposits, a mixture of cool-temperate and cosmopolitan taxa occurs. Though markedly
20
21 308 impoverished, these Unit IIIb diamicton populations nevertheless closely resemble some dinocyst
22
23 309 assemblages from typical Unit III laminated mud samples. Other organic-walled microfossils are
24
25 310 generally absent in the diamicton, apart from sporadic occurrences at very low concentrations
26
27 311 (foraminiferal linings, invertebrate eggs, *Botryococcus*, pollen; Fig. 4B). It should be noted that
28
29 312 Scourse *et al.* (1990) consider this diamicton to be a subglacial deposit, thus inferring that all
30
31 313 included microfossils should be reworked.
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315 4.1.3 Chronology

316 No dates are available from Unit III due to the absence of dateable *in situ* organic materials. The
317 oldest date from the overlying Unit II of 13.9 cal ka BP (CD-8, *S. elliptica*, AA-36251; Table 2) provides
318 a minimum age on the deposition of Unit III. If the Unit IIIb basal diamicton present below Unit III in
319 CS-3 is considered a lodgement till correlative and coeval with the Scilly Till (*sensu* Scourse *et al.*
320 1990), an approximate maximum age of 21-22 cal ka BP (McCarroll *et al.* 2010) for Unit III deposition
321 can be inferred.
322

322

323 4.2 Unit II (bioclastic gravel)

324 4.2.1 Lithostratigraphy

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2
3 325 Unit II, occurring in all cores, is typically a massive, silty sandy gravel, frequently upward fining (Fig.
4
5 326 2). Whole shells and fragments are common throughout, increasing upwards. The lithic component
6
7 327 includes rounded to subrounded gravel-sized clasts (~5-20 mm in size). In all cases, Unit II rests
8
9 328 unconformably on underlying materials, with an erosional surface marked by brecciation and
10
11 329 bioturbation. Clay clasts derived from erosion of underlying Unit III are frequent towards the base of
12
13 330 Unit II. Iron and manganese staining around lithic and biogenic clasts is also common towards the
14
15 331 unit base. Unit thicknesses are variable (8-264 cm), as are colours, though greyish brown to olive
16
17 332 yellow is typical.

18
19
20 333 In two cores (CD-1 and -3) Unit II is interrupted by thin (<20 cm) sandy silt interbeds (Unit IIa;
21
22 334 Fig. 2). In CD-1 this single interbed is massive and shell-rich whilst the three interbeds in Unit II CD-3
23
24 335 are laminated, apparently organic-rich, and dark in colour. An additional variation (Unit IIb) is noted
25
26 336 from CD-8 (Fig. 2), where massive sand (>85 cm thick) unconformably underlies the typical Unit II
27
28 337 sequence. These sands are gravelly towards the base, containing blackened lithic and shell clasts,
29
30 338 and clay balls, but fine upward into a clean and well sorted sand with little to no silt.

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33 339

34 340 4.2.2 Biostratigraphy

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36
37 341 Macrofossil assemblages from Unit II are broadly similar to Unit I (Fig. 3). Species diversities are high
38
39 342 (18-47; typically >30), populations dominated by subtidal boreal-temperate taxa typical of sandy
40
41 343 substrates. As in Unit I, bivalves dominate (max. 483 ind./100 g, CS-1) with prosobranch gastropods
42
43 344 (max. 90 ind./100 g, CD-1) and echinoderms (max. 182 ind./100 g, CD-1). Dominant species include
44
45 345 bivalves *A. ephippium*, *P. ovale*, and *T. ovata*, and the echinoderm *E. pusillus*. Though present in Unit
46
47 346 I, high numbers of the bivalve *S. elliptica* and the prosobranch gastropod *Gibbula tumida*
48
49 347 characterise Unit II, along with high concentrations of serpulid polychaete material (common species
50
51 348 including *Pomatoceros triqueter* and *Ditrupa arietina*) and calcareous bryzoan debris. Abundant
52
53 349 balanoid plates (max. 372 per 100 g, CD-9) frequently outnumber echinoderm spicules (max. 163 per
54
55 350 100 g, CD-2). In contrast to Unit I, infrequent small (1-5 mm) prosobranch gastropods with intertidal,
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1
2
3 351 or restricted low water spring tide to shallow subtidal affinities (Graham 1988; Peacock 1993) occur
4
5 352 in the majority of Unit II samples (max. 19 ind./100 g, CD-9). These include *Rissoa parva*, *Rissoella*
6
7 353 *opalina*, *Pusillina sarsi*, *Tornus subcarinatus*, and *Cingula cingillus*. Silty interbeds (Unit II a; CD-1 and
8
9 354 -3) and underlying massive sands (Unit IIb; CD-8) possess macrofossil assemblages similar to those in
10
11 355 more typical Unit II deposits, but display much lower species diversities (<10 taxa) and
12
13 356 concentrations (max. 39 ind./100 g, CD-3; Fig. 3).

14
15
16 357 Dinocyst data are only available from Unit IIa (Fig. 4A) interbeds present in CD-3, not from
17
18 358 the more typical silty sands and gravels of Unit II widespread through the CDB (see Marret *et al.*
19
20 359 2004 for typical Unit II dinocyst records; CD-7). These specific samples were taken to elucidate the
21
22 360 palaeoenvironmental origin and the intertidal/ estuarine vs. deeper marine character of these
23
24 361 interbeds. Dinocyst concentrations in these samples are generally low (~150-560 ind./g), diversities
25
26 362 ranging from 12 to 23 species. Assemblages are predominantly composed of protoperidinoid taxa,
27
28 363 including *Protoperidinium* spp., *Quinquecuspsis concreta*, and *Votadinium calvum*. These
29
30 364 heterotrophs are accompanied by phototrophic *Spiniferites* spp. (including *Spiniferites* sp. 1 and
31
32 365 *Spiniferites* cf. *membranaceus*). Other NPPs (Fig. 4B), including invertebrate remains (eggs,
33
34 366 mouthparts), foraminiferal linings, and freshwater algae are present in low to moderate amounts.
35
36 367 *Corylus* and *Pinus* pollen are abundant, whereas pre-Quaternary dinocysts and spores are rare (both
37
38 368 <20 ind./g; Fig. 4B).

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42 370 4.2.3 Chronology

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45 371 Excluding deposit-feeding bivalves (e.g. *Nucula* spp.), which may exhibit exaggerated ages due to
46
47 372 uptake of "old" carbonate depleted in ¹⁴C ("Portlandia Effect" *sensu* England *et al.* 2012), dates from
48
49 373 Unit II span a wide range of ages, from 4.0 to 13.9 cal ka BP. We exclude a date of 0.4 cal ka BP (CS-1,
50
51 374 *Corbula gibba*, AA-36188; Table 2) considered anomalously young relative to other dates from this
52
53 375 horizon. Age inversions and wide age spreads within the same horizon are frequent. For example,
54
55 376 CD-4 where dates of 10.3 cal ka BP (OS-79220) and 6.7 cal ka BP (AA-32278) occur at the same
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3 377 sample depth, and CD-3 with a date of 10.9 cal ka BP (AA-36242) above dates of 9.3 to 10.0 cal ka BP
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5 378 (AA-36243; AA-36244; OS-79103; Table 2). Nonetheless, the majority of accepted dates (20) are >9
6
7 379 cal ka BP, with seven dates being ≥ 12.7 cal ka BP. Of these seven older dates, five occur within the
8
9 380 massive shelly sands (Unit IIb) underlying Unit II in CD-8, including the oldest dates (13.9 cal ka BP;
10
11 381 OS-79120 and AA-36251) encountered in this study. Previously published dates from CD-7 (Scourse
12
13 382 & Austin 1994; 9.4 and 13.1 cal ka BP) agree with this general chronology.

14
15 383 Out of the total accepted 31 dates from Unit II (including IIa and IIb), seven were derived
16
17 384 from obligate intertidal to shallow water taxa (Table 2). These seven dates span the entire age range
18
19 385 encountered for Unit II though over half are ≥ 12.7 cal ka BP. Such intertidal taxa dates should be
20
21 386 viewed with caution, as all from within the CDB must be considered redeposited given their
22
23 387 occurrence in a bathymetric basin that could not have been subjected to intertidal conditions during
24
25 388 late Pleistocene - early Holocene lower sea-levels. As a basin, the CDB would have either formed a
26
27 389 subtidal (glaci)marine embayment or an isolated lacustrine system under such conditions (see §5.2).

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32 33 391 **4.3 Unit I (upper silty sand)**

34 35 392 *4.3.1 Lithostratigraphy*

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37 393 Unit I is a predominantly an upward-fining massive silty sand with infrequent finer silty interbeds in
38
39 394 some cores (CD-2 and -9). Thickness varies from 10 to 195 cm, whilst colour ranges from olive grey
40
41 395 to brownish grey. Contact with underlying Unit II is typically gradational although some cores (CD-1,
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43 396 -4, -8) show an abrupt transition. Throughout the area, Unit I is exceptionally rich in bioclastic
44
45 397 material, constituting the modern seabed sediment.

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48 398

49 50 399 *4.3.2 Biostratigraphy*

51
52 400 Calcareous marine macrofossils are universally abundant (Fig. 3), with a maximum of 39 species per
53
54 401 individual sample (CD-4, 50-55 cm). Although variations are apparent, assemblages are typically
55
56 402 dominated by the bivalves *Anomia ehippium*, *Parvicardium ovale*, and *Timoclea ovata*,

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2
3 403 accompanied by the prosobranch gastropod *Turritella communis* and the clypeasteroid echinoderm
4
5 404 *Echinocyamus pusillus*. Bivalves such as *Spisula elliptica*, *Hiatella arctica* and *Nucula sulcata* also
6
7 405 occur frequently. Bivalve concentrations reach a maximum of 2314 ind./100 g (CD-4) whereas
8
9 406 prosobranch gastropods are less abundant (max. 133 ind./100 g, CD-2). Large quantities of
10
11 407 unidentified echinoderm spicules (max. 1160 ind./100 g) and barnacle (balanoid) plates (max. 515
12
13 408 /100 g) typify most samples, as do infrequent fish otoliths (likely from *Trisopterus minutus*). In some
14
15 409 samples (CD-5) shell fragments and byssus threads of the fan mussel *Atrina fragilis* occur in notable
16
17 410 quantities. The thecosomat pteropod *Limacina retroversa* is also locally abundant (CD-4 and -2; 312
18
19 411 and 227 ind./100 g, respectively).
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21
22 412

23 413 4.3.3 Chronology

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25 414 The few direct dates available from Unit I (Table 2) indicate an early to mid Holocene
26
27 415 commencement of deposition. Chronology is further constrained by numerous dates from
28
29 416 underlying Unit II (see §4.2.3), suggesting asynchronous onset of Unit I deposition throughout the
30
31 417 region. Previously published molluscan and bulk benthic foraminiferal dates from Unit I in CD-7
32
33 418 (Scourse & Austin 1994; Scourse *et al.* 2002) range from ~8 to 3 cal ka BP, in agreement with
34
35 419 chronologies described in the present study.
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39 421 5. Interpretations and Discussion

40 422 5.1 Interpretation of units

41
42 423 The shallow stratigraphy of the CDB and adjacent shelf documents the terminal Pleistocene to
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44 424 Holocene palaeoenvironmental evolution of the Celtic Sea. Typically, short (<6 m) vibrocores prove
45
46 425 three lithostratigraphic units (Fig. 2): (i) an upper shelly silty sand deposit (Unit I) constituting
47
48 426 modern seabed sediments and containing a rich cool-temperate marine macrofauna, resting
49
50 427 conformably on (ii) a regionally extensive gravel lag (Unit II), itself containing abundant boreal-
51
52 428 temperate marine macro- and microfossils. The majority of cores also prove (iii) a waterlain massive
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3 429 to laminated glacially-influenced mud unit (Unit III) marked by infrequent small limestones,
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5 430 interpreted as dropstones; an absence of macrofossils, calcareous microfossils, and marine diatoms;
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7 431 and low-abundance and disparate palynomorph assemblages. Unit III is truncated by an extensive
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9 432 erosional surface over which Unit II occurs. Rarely, cores prove deposits beneath Unit III, including
10
11 433 glacial diamicton (Unit IIIb; CS-3) and a poorly sorted gravelly sand (Unit IIIa; CD-2).

12
13
14 434 Unit I is considered equivalent to SL1 (in the north) and Layer A (central and south), whilst
15
16 435 Unit II equates to SL2 (Tappin *et al.* 1994) and Layer B (Evans 1990). The glacial waterlain muds of
17
18 436 Unit III are correlative with the Western Irish Sea Formation Mud Facies (Tappin *et al.* 1994) and the
19
20 437 Laminated Clay of the Melville Formation (Pantin & Evans 1984; Evans 1990; Scourse *et al.* 1990).
21
22 438 The Unit IIIb diamicton at the base of CS-3 has previously been assigned to the southern region
23
24 439 Melville Till of the Melville Formation (Scourse *et al.* 1990, 1991), equivalent to the Upper Till
25
26 440 Member of the Cardigan Bay Formation (Scourse 1991a; Tappin *et al.* 1994). Although limited core
27
28 441 penetration restricts the interpretation of the Unit IIIa gravelly sand at the base of CD-2, its
29
30 442 relationship to overlying units suggests a correlation with the Chaotic Facies of the Western Irish Sea
31
32 443 Formation. (Tappin *et al.* 1994).

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36 37 445 5.1.1. Unit III

38
39 446 Lithostratigraphically, the laminated muds of Unit III represent glacially-influenced waterlain
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41 447 sediments deposited predominantly from suspension, with a minor component of small ice-rafted
42
43 448 dropstones. Similar deposits have been recorded from deglacial glacialmarine and glaciallacustrine
44
45 449 settings boarding the Celtic and Irish seas (Fletcher & Siddle 1998; Hambrey *et al.* 2001; Thomas *et*
46
47 450 *al.* 2004; Etienne *et al.* 2006) and from numerous other locations including the deglacial margins of
48
49 451 the Laurentide Ice Sheet (e.g. Freeman-Lynde *et al.* 1980; Kerr 1987; MacLean *et al.* 1989; Andrews
50
51 452 *et al.* 1991; Pieńkowski *et al.* 2012). Deposition under highly turbid, quiescent, low energy conditions
52
53 453 and an absence of bioturbation is evidenced by the frequently laminated nature of this unit (Syvitski
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55 454 1991; Ó Cofaigh & Dowdeswell 2001). Though clearly “glaciaqueous”, the lithostratigraphy alone
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3 455 (this study) is insufficient to conclusively distinguish between marine or lacustrine conditions.
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5 456 Nonetheless, given the regional uniformity of this unit and its occurrence in the previously described
6
7 457 CS-3, it is considered synonymous with Facies B and the Melville Laminated Clay of the Melville
8
9 458 Formation of Scourse *et al.* (1990). Notably, the Melville Laminated Clay is considered glacialmarine,
10
11 459 defined from southern Celtic Sea BGS cores, with rich, *in situ* arctic ostracods, foraminifera, and
12
13 460 molluscs (Scourse *et al.* 1990).

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15
16 461 In marked contrast to previously described Melville Laminated Clay samples (Scourse *et al.*
17
18 462 1990), Unit III samples examined in this study are characterised by an absence of identifiable
19
20 463 microfossils (Fig. 3), apart from one sample (CD-2) containing extremely sparse and highly abraded
21
22 464 biogenic fragments possibly glacially reworked from earlier marine sediments. Similarly, microfossils
23
24 465 are generally sparse. Only extremely rare freshwater diatoms are recorded, whereas marine diatoms
25
26 466 are absent. Organic-walled microfossils (Fig. 4) typically exhibit low dinocyst concentrations and a
27
28 467 near-absence of other NPPs, especially foraminiferal linings. Dinocyst assemblage structures also
29
30 468 differ compared to studied Unit II interbeds, phototrophs dominating over heterotrophs. The
31
32 469 principal Unit III dinocyst *B. tepikiense* is indicative of fully marine, temperate-polar conditions. It
33
34 470 predominantly dwells in regions with seasonal sea-ice cover of <4 months/year and summer sea-
35
36 471 surface temperatures (SST) of 10-20°C (Marret & Zonneveld 2003), including waters off Iceland, the
37
38 472 Faroe Islands, and the east coast of Canada (Harland 1983; Mudie 1992; Matthiessen *et al.* 2005). In
39
40 473 the modern Celtic Sea, *B. tepikiense* is characteristic of cool, stratified regions (Marret & Scourse
41
42 474 2002; *cf.* Marret *et al.* 2004), but never reaches relative abundances of >5%, in marked contrast to
43
44 475 its dominance in Unit III. *B. tepikiense* is furthermore prominent in Late Glacial Interstadial
45
46 476 sediments of the NW European continental margin (Harland 1994). Of the two other prominent
47
48 477 dinocyst taxa in Unit III, *O. centrocarpum* is cosmopolitan, tolerant of large fluctuations in physical
49
50 478 parameters (including sea-ice), and abundant in the cold to temperate North Atlantic, whereas *L.*
51
52 479 *macherophorum* prefers temperate to tropical regions with SST of >10°C. Its highest abundances are
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54 480 reported from coastal waters off NW Africa and the Iberian Peninsula (Marret & Zonneveld 2003). In
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2
3 481 general, the dinocyst assemblages in Unit III indicate cool to temperate marine conditions, but show
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5 482 a mixture of both fully coastal/marine (*B. tepikiense*, *S. membranaceus*, *S. lazus*) and brackish water-
6
7 483 tolerant (*S. ramosus*), sea-ice tolerant (*B. tepikiense*, *O. centrocarpum*, *Selenopemphix quanta*,
8
9 484 *Islandinium minutum*) and intolerant (*L. macherophorum*, *S. membranaceus*), and temperate-polar
10
11 485 (*B. tepikiense*) as well as thermophilic (*S. mirabilis*) taxa (Marret & Zonneveld 2003). The overall cool
12
13 486 to temperate character of this assemblage is inconsistent with the demonstrably glacially-influenced
14
15 487 sediments. The occurrence of sea-ice-intolerant and thermophilic dinocysts and the relatively high
16
17 488 species diversity is particularly problematic, and thus points to a notable reworked component.
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19 489 Furthermore, the paucity of other NPPs, especially zoomorphs, is striking. The presence of such
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21 490 palynomorphs should be expected in glacial marine sediments, even at times of adverse, ice-proximal
22
23 491 conditions (e.g. Mudie *et al.* 2006; Pieńkowski *et al.* 2012).

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25
26 492 It has been suggested that such mixed dinocyst assemblages resemble those from modern
27
28 493 temperate saltmarsh and estuarine environments, such as the Humber Estuary (Marret pers. obs.)
29
30 494 and the Massachusetts coast (Pospelova *et al.* 2004, 2005). However, despite the occurrence of
31
32 495 common taxa such as *O. centrocarpum*, *Spiniferites* spp., *L. macherophorum* and *I. minutum* both in
33
34 496 Unit III and reported estuarine samples (Pospelova *et al.* 2004, 2005), the foremost taxon present in
35
36 497 Unit III, *B. tepikiense*, is absent. Critically, the laminated and dropstone-rich glacially-influenced
37
38 498 sediments and the near-complete absence of any marine/estuarine macrofossils, foraminifera,
39
40 499 ostracods, and diatoms strongly argue against an estuarine origin for Unit III deposits.

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42
43 500 In summary, Unit III is a waterlain distal glacial deposit containing an exceptionally
44
45 501 sparse reworked macrofossil assemblage and lacking key *in situ* marine indicators such as
46
47 502 foraminifera, ostracods, and diatoms. If organic-walled microfossils are assumed to be entirely (or to
48
49 503 a large extent) *in situ*, the cool-temperate palaeoenvironmental conditions they indicate appear at
50
51 504 odds with the remainder of the evidence. These inconsistencies preclude a simple
52
53 505 palaeoenvironmental interpretation of this unit. Rather, two hypotheses are proposed: firstly, that
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55 506 Unit III represents a shallow ice-distal glacial marine facies with a general absence of *in situ* flora and
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1
2
3 507 fauna, aside from the sea-ice and brackish-water tolerant dinocyst component (which itself may also
4
5 508 be reworked). Secondly, Unit III represents a glacialacustrine facies, wherein all marine macro- and
6
7 509 microfossils, including the dinocyst component are glacially (and/or glacialfluviially) reworked from
8
9 510 pre-existing temperate, boreal, and arctic marine deposits during deglaciation.
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11

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13 512 5.1.2 Unit II

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16 513 Unit II of the central and northern Celtic Sea is interpreted as a regionally extensive lithic and
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18 514 bioclastic lag associated with an unconformity truncating underlying basal laminated muds of Unit III
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20 515 (or other pre-Holocene sediments), above which it occurs. Oldest dates from Unit II provide a
21
22 516 minimum age for the onset of lag generation of 13.9 cal ka BP. This indicates commencement of
23
24 517 widespread erosion across the continental shelf associated with rising late Weichselian to early
25
26 518 Holocene sea-levels. Sedimentology and macrofossil biostratigraphy suggest a highly productive
27
28 519 shallow subtidal marine environment marked by storm-wave and current re-suspension (Peacock
29
30 520 1993; Wingfield 1996; Kidwell 1998, 2002), its typical upward-fining nature and gradation into the
31
32 521 overlying silty sands of Unit I consistent with increasing Holocene water depths. Radiocarbon dates
33
34 522 from Unit II span much of the early to mid Holocene (Table 2), with apparent age inversions and
35
36 523 wide ranges of dates from the same depth intervals suggesting considerable time-averaging,
37
38 524 condensation, and sediment starvation – conditions expected in a shallow high-energy environment
39
40 525 (Fürsich & Aberhan 1990; Kowalewski *et al.* 1998; Kidwell 1998, 2002).

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43
44 526 Macrofaunas are typically subtidal boreal to temperate, taxa currently encountered on the
45
46 527 NW European continental shelf (Tebble 1976; Graham 1988; Hayward & Ryland 1998). Organic-
47
48 528 walled microfossils from interbed samples (CD-3) show dinocysts typical of the NW European
49
50 529 continental shelf dominated by temperate taxa (Fig. 4A; Dodge & Harland 1991; Marret & Scourse
51
52 530 2002), including protoperidinoids such as *Q. concreta* and *V. calvum*, also present in the modern-
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54 531 day, seasonally stratified sector of the Celtic Sea (Marret & Scourse 2002). Round brown cysts
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56 532 (*Protoperidinium* spp., *Brigantedinium* spp.) produced by several motile dinoflagellate species, are
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3 533 also prominent in this unit. These have been shown to occur in high abundances in regions of
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5 534 enhanced productivity (Marret & Zonneveld 2003). Species of the genus *Spiniferites*, also abundant
6
7 535 in Unit II, have been found in high numbers along the modern British coastline (Reid, 1975). Unlike
8
9 536 modern populations in the Celtic Sea (Marret & Scourse 2002), dinocysts concentrations from Unit II
10
11 537 are low, however this may be a function of the coarse nature of the sediment and shallow turbulent
12
13 538 conditions preventing cyst deposition as opposed to reflecting *in situ* productivity. It should be noted
14
15 539 that these microfossil samples are from interbeds only, and thus may not be representative of
16
17 540 “typical” Unit II sandy gravels. However, macrofossil concentrations from Unit IIa interbeds are also
18
19 541 markedly low relative to typical Unit II deposits. Nonetheless, both dinocyst assemblages and
20
21 542 concentrations in Unit IIa closely resemble those previously described from Unit II (CD-7, Marret *et*
22
23 543 *al.* 2004). Although no comparable NPP data are available from the study area, the presence of
24
25 544 foraminiferal linings, invertebrate remains, and other palynomorphs in Unit II (Fig. 4B) is comparable
26
27 545 to other marine settings (e.g. de Vernal *et al.* 1992; McCarthy *et al.* 2003), supporting dinocyst-based
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29 546 interpretations.
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32
33 547 The presence of infrequent small obligate intertidal and shallow water prosobranch
34
35 548 gastropods is inconsistent with CDB physiography, where either subtidal marine or lacustrine
36
37 549 conditions should have persisted following deglaciation (Eyles & McCabe 1989, 1991; Lambeck 1991,
38
39 550 1993, 1995, 1996; Lambeck & Purcell 2001; Scourse & Furze 2001; Uehara *et al.* 2006). This material
40
41 551 is considered allochthonous, advected into the CDB from nearby intertidal areas by stormwave and
42
43 552 tidal currents. High tidally-driven bottom current velocities resulting in post mortem transport and
44
45 553 selective destruction is also indicated by benthic foraminifera previously described from this unit
46
47 554 (CD-7, foram zone F1b, Scourse *et al.* 2002) and supported by palaeotidal simulations of bed stress
48
49 555 evolution (Scourse *et al.* 2009). Despite the possibility of intertidal conditions on the higher
50
51 556 continental shelf following deglaciation (Lambeck 1991, 1993, 1995, 1996; Lambeck & Purcell 2001),
52
53 557 macrofaunas from Unit II do not conclusively indicate such an environment. Indeed, fewer obligate
54
55 558 intertidal individuals are found in open shelf samples from Unit II than within the CDB, macrofaunas
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2
3 559 being almost universally composed of subtidal taxa (Fig. 3). If the littoral zone did progress across
4
5 560 the central Celtic Sea shelf during rising late glacial and Holocene sea-levels, no definitive *in situ*
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7 561 deposits have been identified in this study.
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9 562

10 563 *5.1.3 Unit I*

11 564 Unit I represents a boreal-temperate subtidal shelf sea deposit recording increasing water depths
12
13 565 throughout the Holocene. Some cores (CD-1 and -3) show possible truncation at the top of Unit I, a
14
15 566 phenomenon also recorded from CD-7 (Austin & Scourse 1997; Scourse *et al.* 2002). Furthermore, it
16
17 567 is from this unit CD-7 (Unit I this study; "Lithozone 3" *sensu* Scourse *et al.* 2002; "Lithozone 1" *sensu*
18
19 568 Marret *et al.* 2004) that the onset of Celtic Sea seasonal stratification as a function of increasing
20
21 569 water depth has been described on the basis of foraminifera and dinocysts (Scourse & Austin 1994;
22
23 570 Austin & Scourse 1997; Marret *et al.* 2004), a threshold change from mixed to stratified watermasses
24
25 571 that occurred between 8990 and 8440 cal yr cal BP (Scourse *et al.* 2002). This environmental shift in
26
27 572 Unit I is confirmed by the occurrence of the pteropod *L. retroversa* – a planktonic species associated
28
29 573 with cool boreal, stratified waters (Bathman *et al.* 1991; Gallagher *et al.* 1996; Gowen *et al.* 1998).
30
31 574 Where present (CS-3, CD-2, CD-4), high *L. retroversa* numbers are achieved after the establishment
32
33 575 of seasonal stratification. Its decline in the late Holocene is likely a function of increased water
34
35 576 temperatures rather than a breakdown in seasonal stratification. Molluscan (and other) macrofossils
36
37 577 are all subtidal and of boreal to temperate affinities, found on the modern NW European continental
38
39 578 shelf (Tebble 1976; Thompson & Brown 1976; Graham 1988; Hayward & Ryland 1998).
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47 580 *5.2 Competing hypotheses*

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49 581 The clay-rich and frequently laminated sediments of Unit III, marked by sporadic small dropstones,
50
51 582 suggest distal, glacially-influenced subaqueous suspension-driven deposition with a minor seasonal
52
53 583 ice or iceberg-driven rafted component. However, the marine or lacustrine nature of such an
54
55 584 environment is difficult to ascertain especially considering the contradictory nature of the contained
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2
3 585 microfossil assemblages and absence of macrofossils. It is on this unit, however, that a glacimarine
4
5 586 or a glacialacustrine interpretation hinges.
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7 587 A temperate estuarine and saltmarsh origin for the Celtic Sea laminated clay sequences (Unit
8
9 588 III) can be discounted on litho- and bio-stratigraphic grounds (see §5.1.1 above). Furthermore, a fully
10
11 589 glacimarine origin (deep water, full salinities) is unlikely given the absence of critical *in situ* macro-
12
13 590 and micro-fossils. Barren glacimarine sediments have been recorded from several modern tidewater
14
15 591 glacier and fjord delta settings in Alaska and Greenland (e.g. Cowan & Powell 1990; Gilbert *et al.*
16
17 592 1998, 2002; Desloges *et al.* 2002) where high sedimentation rates prevent the establishment of
18
19 593 benthic communities (Korsun & Hald 1998; Jaeger & Nittrouer 1999). In all such cases, however, a
20
21 594 steep environmental gradient exists from immediately proximal to more distal settings; bioturbated
22
23 595 sediments being encountered within only four kilometres of the ice front or fjord head. Nor should
24
25 596 planktonic material be excluded from deposition in such settings, especially where tidal pumping is
26
27 597 significant (Cowan *et al.* 1998, 1999; Gilbert *et al.* 2002). Indeed, ice-proximal sediments including
28
29 598 water-lain diamictos and laminated sequences from the deglacial Barents Sea (Murdmaa *et al.*
30
31 599 2006) and the Canadian High Arctic (Pieńkowski *et al.* 2012), where rapid deposition from sediment
32
33 600 plumes can be inferred, are nonetheless marked by a sparse but recognizable *in situ* glacimarine
34
35 601 microbiota. In the Celtic Sea, *in situ* arctic and glacimarine species have been noted from the south
36
37 602 where they occur in lithostratigraphically similar units (Melville Laminated Clay; Scourse *et al.* 1990)
38
39 603 to Unit III. Fauna reported by Scourse *et al.* (1990) include ostracods (*Rabulimys mirabilis*, *Krithe*
40
41 604 *glacialis*, *Cytheropteron montrosiense*), abundant foraminifera (*Islandiella helenae*, *Islandiella*
42
43 605 *islandica*) and the boreo-arctic mollusc *Yoldiella fraterna*. These definitive glacimarine indicators all
44
45 606 occur south of the postulated LGM southern maximum for the BIIS on the Celtic Sea shelf, laminated
46
47 607 and massive clays north of the ice limit (Unit III this study) being effectively barren of macrofossils,
48
49 608 foraminifera, and marine diatoms, and containing ecologically incompatible palynomorphs. Whilst
50
51 609 the transition from fossil-rich sediments in the south to near-barren in the central and northern
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53 610 Celtic Sea may indeed reflect a steep glacimarine environmental gradient, the spatial consistency of
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3 611 Unit III laminated muds both within and between cores may challenge this interpretation.
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5 612 Furthermore, the absence of any successional changes in Unit III micro- or macrobiota that could be
6
7 613 expected during progressive northward ice retreat from the Celtic and Irish seas is also problematic.
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9
10 614 Two competing hypotheses thus remain that can be invoked to explain the formation of the
11
12 615 Unit III laminated clays present in the CDB and across the broader northern and central shelf of the
13
14 616 Celtic Sea: a low sea-level model with deposition of such laminated sequences in a glaciallacustrine
15
16 617 system impounded within the CDB and in other smaller satellite water bodies on a subaerially
17
18 618 exposed shelf; or a higher sea-level model with deposition of fine grained sediments in a broad
19
20 619 shallow glacialmarine embayment, where very high sedimentation rates effectively prevent primary
21
22 620 production.
23

24 621

25 622 5.2.1 The glaciallacustrine hypothesis

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27
28 623 In this scenario, low relative sea-level in the central Celtic Sea (Lambeck 1991, 1993, 1995, 1996;
29
30 624 Lambeck & Purcell 2001; Uehara *et al.* 2006) results in sufficient continental shelf exposure during
31
32 625 and subsequent to deglaciation to isolate the CDB and other bathymetric lows from the North
33
34 626 Atlantic. Rapid sediment plume deposition and iceberg rafting associated with the northward retreat
35
36 627 of the ice margin results in emplacement of massive and laminated mud sequences with a minor
37
38 628 dropstone component (Unit III). *In situ* productivity within the glaciallacustrine system is inhibited due
39
40 629 to high turbidity, rapid sediment accumulation rates, and low temperatures (Korsun & Hald 1998;
41
42 630 Jaeger & Nittrouer 1999). Thus, aside from an extremely sparse freshwater diatom component, all
43
44 631 biota noted from Unit III are glacially reworked from older temperate to glacialmarine sediments
45
46 632 overridden by the ISIS. Water depths within such a postulated CDB lake would have been shallow,
47
48 633 not exceeding 30 m (based on modern bathymetry) whilst the rhythmicity seen in some of the Unit
49
50 634 III muds would be a function of variable sediment and meltwater input events (Cowan *et al.* 1999; Ó
51
52 635 Cofaigh & Dowdeswell 2001). The extent of such a glaciallacustrine system is hard to determine, but
53
54 636 would have occupied the CDB and could have conceivably extended through St George's Channel in
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3 637 to the Irish Sea in contact with a rapidly retreating ice margin. Separate or connected lesser lake
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5 638 basins would also have occurred on the exposed shelf within the ISIS ice limit. Assuming lake
6
7 639 isolation and shelf exposure during ice retreat, lacustrine conditions would have persisted from ~21-
8
9 640 22 cal ka BP (assuming Celtic Sea and Isles of Scilly ice maxima are coeval; McCarroll *et al.* 2010) until
10
11 641 the reconnection with the North Atlantic due to the eustatically-driven transgression sometime prior
12
13 642 to 13.9 cal ka BP. With the establishment of marine conditions, tidal current and stormwave erosion
14
15 643 and winnowing of Unit III glacialacustrine sediments would have been significant (Uehara *et al.* 2006;
16
17 644 Neill *et al.* 2009; Scourse *et al.* 2009). Wingfield (1996) estimates that up to 15 m of stony
18
19 645 glacialmarine/glacialacustrine muddy sediment may be lost before a sufficient armouring lag of
20
21 646 winnowed dropstones and accumulated molluscan fragments (Unit II) develops that inhibits further
22
23 647 erosion.
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29 649 5.2.2 The glacialmarine hypothesis

30
31 650 Assuming the retreat of the ISIS from the Celtic Sea shelf in a tidewater setting, the CDB and
32
33 651 adjacent shelf would have formed a shallow (maximum 30-40 m given modern bathymetry and
34
35 652 elevation of southern sill; Peltier *et al.* 2002), broad glacialmarine embayment fed by meltwater from
36
37 653 the retreating ice margin characterized by direct iceberg calving. As in the glacialacustrine hypothesis,
38
39 654 exceptionally high sedimentation rates can be expected, with fine-grained deposition from sediment
40
41 655 plumes and minor dropstone rafting. Rates of 5-10 mm per annum have been reported for modern
42
43 656 fjords (Cowan and Powell 1991; Gilbert *et al.* 1998; Desloges *et al.* 2002) with extreme rates of >4
44
45 657 mm per day recorded in systems influenced by surging glaciers (Gilbert *et al.* 2002). As in the
46
47 658 glacialacustrine hypothesis, such high sedimentation rates and turbidity would have significantly
48
49 659 curtailed any *in situ* primary productivity preventing the establishment of macrofaunal benthic
50
51 660 communities (Korsun & Hald 1998; Gordillo & Aitken 2001). The majority of microfossils would thus
52
53 661 be glacially/glaciofluvially reworked from earlier Irish and Celtic sea temperate and glacialmarine
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55 662 sequences. Under glacialmarine conditions, some microfossils could be expected to have been
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3 663 advected from more distal areas on the open shelf (*cf.* Murdmaa *et al.* 2006), however, given the
4
5 664 durability and resistance to damage of organic-walled microfossils (in particular dinocysts)
6
7 665 distinguishing between *wholly* reworked mixed temperate and glacimarine, and *dominantly*
8
9 666 reworked mixed temperate and glacimarine (but with a minor *in situ* glacimarine component)
10
11 667 assemblages is problematic. Salinities may also have been low due to significant meltwater flux into
12
13 668 the basin. Nonetheless, with a connection to the North Atlantic, a tidal influence within the
14
15 669 embayment should be expected, though high bed shear stresses at this time would have most likely
16
17 670 prevented the deposition of glacimarine fines (Uehara *et al.* 2006; Scourse *et al.* 2009). Subsequent
18
19 671 sea-level increase associated with enhanced stormwave and tidal current erosion would have
20
21 672 resulted in erosion and winnowing of Unit III and generation of the Unit II lag deposit in a similar
22
23 673 fashion to the glacialacustrine hypothesis.
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26
27 674 To accommodate the apparent uniformity seen in the Unit III litho- and bio-stratigraphy,
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29 675 such a glacimarine model must be characterised by the rapid break-up and retreat of the ISIS
30
31 676 northwards from the basin rather than a gradual and sequential ice retreat. Successional changes
32
33 677 with increasing ice-distal conditions (Thomsen & Vorren 1986; Korsun & Hald 1998; Jaeger &
34
35 678 Nittrouer 1999; Gilbert *et al.* 2002) are not evidenced in the Unit III waterlain glacialic sediments.
36
37 679 Indeed, the relative rapid retreat of the ice margin from the Isles of Scilly to Anglesey (<3000 years;
38
39 680 Scourse *et al.* 2009b; McCarroll *et al.* 2010) is also consistent with this hypothesis. A previously
40
41 681 proposed (Scourse *et al.* 1990) deglacial iceshelf during Unit III deposition is hard to justify on
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43 682 glaciological and palaeotidal grounds. Early deglacial water depths would have been too shallow and
44
45 683 with too greater tidal amplitude (Uehara *et al.* 2006; Scourse *et al.* 2009) to support a floating ice
46
47 684 margin (Benn & Evans 2010) until the mid to late Holocene (e.g. Heyworth & Kidson 1982; Shennan
48
49 685 *et al.* 2006; Brooks *et al.* 2008; Massey *et al.* 2008). Furthermore, Unit III sediments are not
50
51 686 consistent with extensive undermelt and rapid coarse clastic deposition from a debris-rich mid-
52
53 687 latitude iceshelf (Powell 1984; Syvitski 1991, Ó Cofaigh & Dowdeswell 2001; Post *et al.* 2007).
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689 **5.3 Critique of Glacimarine and Glacilacustrine Hypotheses**

690 Both proposed hypotheses broadly accommodate the available data, making a determination
691 between either model problematic. Both predict rapid ice-distal laminated to massive fine grained
692 subaqueous sedimentation with a minor ice rafted coarse component, deposited under low energy
693 conditions. Both also require rapid deglaciation of the Celtic Sea to explain the lack of lateral or
694 vertical facies variation evident in the glacial waterlain sediments. Broadly analogous conditions
695 (though under greater water depths) persisted in the deglacial Barents Sea (Murdmaa *et al.* 2006),
696 and such a rapid deglaciation in a marine or lacustrine setting has been previously inferred for the
697 Celtic and southern Irish Sea (Scourse *et al.* 2009; McCarroll *et al.* 2010). This apparent rapid ice
698 retreat may indeed favour the glacimarine hypothesis whereby rising glacio-eustatic sea-levels
699 induce the destabilization and break-up of the ISIS during deglaciation, a forcing mechanism not
700 accommodated by a glacilacustrine model. Nonetheless, the presupposition of a deglacial tidal
701 connection to the North Atlantic across the Celtic Sea shelf is hard to reconcile with the fine-grained
702 nature of the Unit III sediments during a period when enhanced tidal bed shear vectors are
703 anticipated (Uehara *et al.* 2006; Neill *et al.* 2009; Scourse *et al.* 2009), whilst an iceshelf can be ruled
704 out on glaciological and palaeotidal grounds.

705 Biostratigraphically, both models predict the deposition of significant quantities of
706 palynomorphs, reworked from pre-existing marine and glacimarine sediments overridden by the
707 advancing ISIS, with minimal *in situ* production under high turbidity and low salinity. In both
708 scenarios, larger, calcareous, or more delicate materials are considered less likely to be preserved.
709 Glacigenic sediments, including glacilacustrine sequences, containing reworked marine micro- and
710 macro-fossils are common along the Celtic and Irish sea margins. Tills on the Llŷn Peninsula contain
711 reworked foraminifera derived Pleistocene temperate and glacimarine sequences overridden by the
712 ISIS (Austin & McCarroll 1992) whilst similar foram-containing deposits along the eastern Irish coast
713 are also now considered subglacial till (McCarroll 2001) rather than glacimarine (Haynes *et al.* 1995).
714 At Killiney, Ireland, marine mollusc in presumed ice-proximal glacimarine diamictons (*sensu* Eyles &

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2
3 715 McCabe 1989), are now considered reworked within terrestrial tills (Rijsdijk *et al.* 2010).
4
5 716 Glacilacustrine sequences on the Isle of Man (Thomas *et al.* 2004) show either barren or disparate
6
7 717 reworked foraminiferal assemblages, as do ice-marginal lacustrine sequences from Glacial Llyn Teifi
8
9 718 in SW Wales (Riding 1997; Fletcher & Siddle 1998; Hambrey *et al.* 2001; Etienne *et al.* 2006). There is
10
11 719 thus clear evidence for glacial transport and lacustrine redeposition of reworked marine biota
12
13 720 associated with the glaciation of the Irish and Celtic seas. In the present study, this is supported by
14
15 721 the occurrence of a sparse reworked dinocyst and NPP assemblage within the basal Melville Till
16
17 722 diamicton (Unit IIIb) in CS-3 which strongly resemble those from Unit III (Melville Laminated Clay)
18
19 723 waterlain laminated muds. An ice-proximal glaciaqueous, as opposed to subglacial lodgement, origin
20
21 724 of the diamicton cannot be dismissed given the liquidity index of the samples (Lambert & Khwaja
22
23 725 1978) at the base of core CS-3 where overconsolidation during core penetration is likely. The
24
25 726 apparent absence of *any* macro or micro marine indicators, aside from highly preservable
26
27 727 palynomorphs with a strong potential for reworking, and the presence of extremely rare freshwater
28
29 728 diatoms further argues in favour of a lacustrine system. Additionally, even under conditions of very
30
31 729 rapid marine deglaciation, lateral environmental gradients influencing benthic and planktonic
32
33 730 community structure, and thus preserved macro- and microfossil assemblages, should be expected
34
35 731 (Korsun & Hald 1998; Jaeger & Nittrouer 1999). This is especially true of large areas such as the
36
37 732 central and northern Celtic Sea where, under rapid ice retreat, ice distal locations may have been up
38
39 733 to 200 km from the calving margin. Furthermore, with increasingly distal glacialmarine conditions a
40
41 734 succession from early colonizers to an established boreo-arctic flora and fauna should be observed
42
43 735 vertically through Unit III samples (Thomsen & Vorren 1986). Neither of these patterns is evident.
44
45 736 Nonetheless, given the potential for very low salinities, high sedimentation rates, and high
46
47 737 meltwater efflux, and thus minimal *in situ* production, the glacialmarine hypothesis cannot be wholly
48
49 738 ruled out on biostratigraphic grounds.
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54 739 Timing of deglaciation in the Irish and Celtic seas, and the sea-levels in which it occurred,
55
56 740 remain debatable. A deglacial marine limit of 30 m asl is described by Clark *et al.* (2004) and McCabe
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2
3 741 *et al.* (2005) from Kilkeel on the northeast Irish coast with deglacial ^{14}C dates between 21 and 19 ka
4
5 742 BP. This, however, is exceptionally early, and hard to reconcile with other data suggesting much later
6
7 743 deglaciation this far north (McCarroll *et al.* 2010). Furthermore, the Kilkeel chronology is based on
8
9 744 bulk foraminiferal AMS ^{14}C dates in an environment where the admixture of resuspended reworked
10
11 745 pre-LGM material is highly likely (*cf.* Austin & McCarroll 1992; Thomas *et al.* 2004; Pieńkowski *et al.*
12
13 746 submitted), as indicated by re-deposited molluscs in the same unit (Rijsdijk *et al.* 2010). McCabe *et*
14
15 747 *al.* (2005) suggest a later lowstand between 18 and 16.7 ka BP (but 13.4 cal ka BP and 30 m below
16
17 748 modern sea-level in Belfast Lough according to Kelley *et al.* 2006) attributable to marked isostatic
18
19 749 rebound, though again, the chronology must be regarded with caution. From North Wales into the
20
21 750 Celtic Sea, deglaciation takes place with sea-levels lower than present on both Irish and British
22
23 751 coastlines (e.g. Rijsdijk *et al.* 2010; Roberts *et al.* 2011). This is counter to earlier work suggesting
24
25 752 glacial and deglacial sea-levels higher than present along the Irish coast of the southern Irish Sea in
26
27 753 particular (e.g. Eyles & McCabe 1989; McCabe & Ó Cofaigh 1995; McCabe *et al.* 1998; McCabe 2008).
28
29 754 Well constrained dates from the central Irish Sea place deglaciation around 19 ka BP with dates of
30
31 755 18.1-19.2 cal ka BP for Holyhead Mountain (Anglesey) and 19 ^{14}C ka BP for Dundalk Bay (McCarroll *et*
32
33 756 *al.* 2010).

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37 757 GIA models (Lambeck 1991, 1993, 1995, 1996; Lambeck & Purcell 2001) suggest the
38
39 758 significant exposure of continental shelf in the Celtic Sea following deglaciation and the persistence
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41 759 of an isolated CDB lacustrine system. Deglaciation northwards through the Irish Sea in contact with a
42
43 760 large standing water body is evidenced by swath bathymetry from north of Anglesey (van
44
45 761 Landeghem *et al.* 2009). Given the modern bathymetry, even accounting for isostatic rebound and
46
47 762 Holocene erosion and deposition, such conditions might have extended southwards through St
48
49 763 Georges Channel into the CDB. Furthermore, deglacial dates of 20.9-22.1 cal ka BP (Scourse & Furze
50
51 764 2001; McCarroll *et al.* 2010) from the northern margin of the Isles of Scilly (southern ISIS limit)
52
53 765 provide an approximate age on Celtic Sea deglaciation. Both hypotheses can be accommodated by
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55 766 the available deglacial and sea-level data from the Irish and Celtic Sea basins. The age of 13.9 cal ka
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3 767 BP from CD-8 for the onset of marked marine erosion and bioclastic lag generation is a minimum
4
5 768 value given the potential for time averaging and condensation in such lag sediments. Remnant (and
6
7 769 confluent) British and Irish ice remained over the North Channel between NE Ireland and SW
8
9 770 Scotland until after 16 cal ka BP (e.g. Shennan *et al.* 2006).

11 771 Geomorphically, the glacialacustrine hypothesis is harder to accommodate than the
12
13 772 glacialmarine model. While the glacio-isostatic rebound models of Lambeck (1995, 1996, in Uehara *et*
14
15 773 *al.* 2006), Lambeck & Purcell (2001), and Peltier *et al.* (2002) predict an isolation basin in the CDB
16
17 774 during low postglacial sea-levels, consistent with a glacialacustrine interpretation of Unit III in this
18
19 775 area (cores CD-1 to -9), similar Unit III deposits on the open shelf south of the CDB sill are harder to
20
21 776 explain (cores CS-1 to -3). If glacialacustrine, an extensive area of subaerially exposed shelf
22
23 777 characterised by moraine-dammed and kettle lake basins peripheral to the main CDB system must
24
25 778 be invoked, extending southwards to at least the location of CS-3 (VE 49/-09/044) and the modern
26
27 779 120m isobath. Despite the fact that there exists little stratigraphic or seismic evidence from the
28
29 780 Celtic Sea to support this, the latest Lambeck GIA model (in Uehara *et al.* 2006; Scourse *et al.* 2009)
30
31 781 does depict sufficiently extensive subaerial shelf exposure from 21 ka BP until after 15 ka BP lending
32
33 782 some credence to this hypothesis. Furthermore, the model predicts marine ingression and final
34
35 783 breaching of the CDB's southern sill shortly before 14 ka BP, consistent with the oldest marine dates
36
37 784 of 13.9 cal ka BP from the erosional lag of Unit II. This, too, is reconcilable with the deglacial records
38
39 785 farther north in the Irish Sea if problematic deglacial foraminiferal dates are discounted.
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46 787 **5.3 Implications**

47 788 **5.3.1 Irish Biogeography**

48 789 A lacustrine system (and thus a Celtic Sea landbridge) before 13.9-14 cal ka BP has direct
49
50 790 implications for Ireland's postglacial biogeography. Irish faunas from the late glacial and early
51
52 791 Holocene are impoverished compared to those from the same time period in Britain (Stuart 1977,
53
54 792 1995; Yalden 1982; Stuart & van Wijngaarden-Bakker 1985; Woodman *et al.* 1997; Coard &
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3 793 Chamberlin 1999). The impoverishment and arrival times of different species into Ireland has been
4
5 794 used to argue both for and against the existence of landbridges. Some authors (Stuart & van
6
7 795 Wijngaarden-Bakker 1985; Dobson 1994; Stuart 1995; Lynch 1996; Stewart & Lister 2001;
8
9 796 Mascheretti *et al.* 2003; Teacher *et al.* 2009) conclude that no terrestrial link existed since before the
10
11 797 last cold stage. A reduced pre-Midlandian/Devensian fauna either persisted in cryptic ice-free refugia
12
13 798 (unsupported by glaciological evidence), or was derived from terrestrial taxa crossing marine
14
15 799 channels between Britain and Ireland.

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17
18 800 The potentially glacialacustrine sequences of the Celtic Sea, in combination with latest GIA
19
20 801 model of Lambeck (in Uehara *et al.* 2006) counter arguments of Irish insularity, suggesting an
21
22 802 ephemeral isthmus prior to the Younger Dryas (~13 000 BP). The distribution of taxa in southern
23
24 803 Ireland and mainland Europe favours an early southern Celtic Sea immigration corridor (Preece *et al.*
25
26 804 1986; Seyd 1992; Gleed-Owen 1997; Woodman *et al.* 1997) lending credence to the glacialacustrine
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28 805 hypothesis.

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31 806

32 33 807 5.3.2 Celtic Sea Linear Sand Ridges (LSRs)

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35 808 If the glacialacustrine model proposed here is accepted, it implies the existence of not only a CDB lake
36
37 809 system, but numerous shallow peripheral kettle and moraine-dammed lakes across the recently
38
39 810 deglaciated subaerial shelf. The occurrence of glaciaqueous units interpretable as glacialacustrine in
40
41 811 BGS vibrocores VE 49/-09/044 (Scourse *et al.* 1990, 1991; core CS-3 this study) and VE 50/-07/141
42
43 812 (core CS-1 this study) south of the CDB on the open shelf is consistent with this model. Furthermore,
44
45 813 a glacialacustrine interpretation of Unit III muds in core CS-3 (VE 49/-09/044) has implications for
46
47 814 understanding the development of the Celtic Sea LSR system. This vibrocore was recovered from the
48
49 815 flanks of one of the most northerly LSRs and its interpretation as a lodgement till and glacialmarine
50
51 816 sequence draping an earlier LSR deposit is critical for the hypothesis that LSR development predated
52
53 817 the MIS2 advance of the ISIS (Bouysse *et al.* 1976; Pantin & Evans 1984; Belderson *et al.* 1986;
54
55 818 Scourse *et al.* 1990, 1991; Reynaud *et al.* 1995, 1999). However, recent modelling and seismic data
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3 819 suggest that though some LSRs may indeed be cored by older deltaic and estuarine sediments, the
4
5 820 main period of LSR development was between 20 and 12 ka BP under rising eustatic sea-levels and
6
7 821 sediment cannibalization from the extensive midshelf depocentres of the Fleuve Manche estuary
8
9 822 and the ISIS fan (Lericolais *et al.* 2003; Reynaud *et al.* 2003; Scourse *et al.* 2009).

10
11 823 The laminated sediments of Unit III in core CS-3 (VE 49/-09/044; Melville Laminated Clay of
12
13 824 Scourse *et al.* 1990, 1991) are clearly different to similar deposits to the south, with an absence of
14
15 825 marine macrofossils, and microfossil assemblages suggestive of glacialacustrine redeposition. If this
16
17 826 core thus records glacialacustrine rather than glacialmarine conditions, and if those glacialgenic
18
19 827 sediments (Units III and IIIb; Melville Laminated Clay and Melville Till) outcrop from the *side* of the
20
21 828 LSR rather than drape it, then a subaerial interval on the central Celtic Sea shelf followed by
22
23 829 inundation and LSR development as outlined by Scourse *et al.* (2009) is plausible and consistent with
24
25 830 the glacialacustrine model. However, until more detailed high resolution seismic data and long core
26
27 831 records are available from the area this aspect of the hypothesis will remain untested.

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32 33 833 *5.3.3 Deglacial Styles*

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35 834 The maximum extent of the LGM BIIS in the Celtic Sea is yet to be established (Sejrup *et al.* 2005)
36
37 835 though it is thought to have terminated in a tidewater margin south of 49°30'N (Scourse *et al.* 1990,
38
39 836 1991; Scourse & Furze 2001). While surging behaviour across a deformable bed has been invoked for
40
41 837 the development of the ISIS lobe in the Celtic Sea (Scourse & Furze 2001; Ó Cofaigh & Evans 2001a,
42
43 838 b, 2007; Roberts *et al.* 2007; McCarroll *et al.* 2010), a glacialacustrine hypothesis would imply the
44
45 839 retreat of a terrestrial margin northwards across the Celtic Sea shelf before the establishment of a
46
47 840 lacustrine calving margin in the CDB and a connected southern Irish Sea lake system. However, once
48
49 841 established in the topographic depression of the CDB and St George's Channel, calving of the ISIS
50
51 842 may have been particularly rapid as deeper water conditions (a function of bathymetry and
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53 843 increased isostatic depression) would be encountered northward. Rapid deglaciation is indicated by
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55 844 a mere 3000 year difference between Isles of Scilly (20.9-22.1 cal ka BP) and Holyhead Mountain,
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3 845 Anglesey (18.1-19.2 cal ka BP; McCarroll *et al.* 2010). Van Landeghem *et al.* (2009) presents detailed
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5 846 swath bathymetry evidence of subaqueous deglacial bedforms from west of Anglesey consistent
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7 847 with ice retreat in both lacustrine and marine settings.
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10 11 849 5.3.4 Unresolved Issues

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13 850 Both the proposed glacialmarine and glaciallacustrine hypotheses for the deglacial and post glacial Celtic
14
15 851 Sea remain speculative. In particular, the high preservation potential of organic-walled microfossils
16
17 852 (primarily dinoflagellate cysts and foram linings) makes the determination of low productivity
18
19 853 glacialmarine and glaciallacustrine systems subject to the resuspension and redeposition of older glacial
20
21 854 reworked marine sediments problematic. We are unaware of any systematic assemblage structure
22
23 855 or taphonomic studies of organic-walled microfossils from marine-derived tills and glaciallacustrine
24
25 856 sediments around the margins of the Celtic and Irish seas or analogous environments. The degree to
26
27 857 which such microfossils can be reworked with little or no morphological modification thus remains
28
29 858 undetermined.
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33 859 If the glacialmarine as opposed to the glaciallacustrine model is accepted, the absence of *in situ*
34
35 860 micro- and macro-palaeontological materials from the Unit III mud deposits within the BISS limits as
36
37 861 opposed to the rich *in situ* glacialmarine faunas in similar deposits beyond needs to be explained,
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39 862 especially given the inferred time-transgressive nature of such deposits during ice retreat.
40
41 863 Furthermore, the degree to which Unit III has been eroded and winnowed, potentially resulting in
42
43 864 the loss of indicative (and conceivably fossiliferous) ice-distal glacialaqueous horizons remains
44
45 865 unquantified. There also remains a mismatch between palaeotidal models (Uehara *et al.* 2006)
46
47 866 suggesting elevated peak bed shear stresses during the immediate deglacial period and the fine-
48
49 867 grained nature of those deglacial sediments (Unit III this study, Melville Laminated Clays of Scourse
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51 868 *et al.* 1990, 1991).
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55 869 Clearly, there is a pressing need to further refine and test the competing hypotheses of
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57 870 Celtic Sea shelf deglacial and postglacial evolution. In particular, if the Unit II muds are indeed
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3 871 lacustrine within the central and northern Celtic Sea regions, how extensive was the implied
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5 872 subaerially exposed shelf, and when did the final eustatically-driven separation between Britain and
6
7 873 Ireland occur? Building on this study, and the earlier work of Scourse *et al.* (1990, 1991), a detailed
8
9 874 multiproxy micro- and macro-palaeontological analysis of all units of the Melville Formation
10
11 875 (including the laminated clays and till) from the “proven” glacimarine units on the shelf edge
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13 876 through to the Unit III sequences in the northern CDB is required to examine if lateral biofacies
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15 877 variations are or are not apparent. Additionally, detailed microstratigraphic analysis of the laminated
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17 878 sequences across the Celtic Sea shelf, as well as geochemical or biogeochemical analysis of those
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19 879 glaciaqueous deposits may further advance the exploration of the lacustrine and marine deglacial
20
21 880 hypotheses.

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24 881 Until then, this study represents the first systematic, critical analysis of Celtic Sea deglacial to
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26 882 postglacial environments highlighting the complexities inherent where a major icestream overriding
27
28 883 previous marine sediments terminates in, and retreats from, a mid-shelf setting. Given the far-
29
30 884 reaching implications for the Quaternary of the British Isles associated with either hypothesis, this
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32 885 work raises crucial questions regarding the style and timing of deglaciation, and sea-levels in which
33
34 886 this occurred.

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38 39 888 **6. Conclusions**

40
41 889 The character of late Pleistocene deglaciation in the central and northern Celtic Sea (including the
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43 890 Celtic Deep Basin) remains unresolved. Based on the absence of (glaci)marine macrofossils, diatoms,
44
45 891 and foraminifera, and the presence of redeposited and environmentally incompatible dinocysts in
46
47 892 regionally extensive deglacial waterlain deposits, two competing hypotheses are proposed: 1) a
48
49 893 glacimarine model with the rapid retreat of an Irish Sea Ice Stream calving margin in contact with the
50
51 894 sea northwards across the shelf into St George’s Channel; 2) a glacialacustrine model with initially
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53 895 subaerial ice retreat exposing a system of ephemeral lacustrine basins across the shelf and the
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55 896 northward calving of the ice margin in an extensive Celtic Deep lake basin, potentially extending

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3 897 northwards into the southern and central Irish Sea. In the latter model, all encountered
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5 898 palynomorphs are considered redeposited from glacial (and or glacialfluviially) reworked older marine
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7 899 sequences in the Celtic and Irish seas.
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10 900 Based on the absence of any dated (molluscan) macrofossils older than 13.5-14.2 cal ka BP
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12 901 from extensive overlying erosional lag deposits and the apparent agreement with the GIA
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14 902 simulations of Lambeck (in Uehara *et al.* 2006 and Scourse *et al.* 2009), the glacialacustrine model is
15
16 903 favoured. However, until sufficient data is available to test either hypothesis further, such an
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18 904 interpretation must remain speculative. Given the important implications that this has for
19
20 905 understanding the late Quaternary environmental evolution of the NW European shelf, there
21
22 906 remains a need for further research to establish the nature and timing of Celtic Sea deglaciation.
23
24 907

26 908 **Acknowledgments**

28
29 909 MFAF gratefully acknowledges the receipt of a NERC PhD research studentship grant
30
31 910 (GT04/97/289/ES) and two NSERC-funded radiocarbon allocations (746/0898; 814/0999) in support
32
33 911 of this research. Also gratefully acknowledged is the award of an RSACAF grant and a Professional
34
35 912 Development Conference Attendance grant from Grant MacEwan University to MFAF. Staff at the
36
37 913 NERC Radiocarbon Facility in East Kilbride, Scotland, in particular Charlotte Bryant, Brian Miller, and
38
39 914 Margaret Currie must be thanked for their valued discussions regarding radiocarbon dating issues.
40
41 915 Thanks must also be extended to the British Geological Survey for making core materials central to
42
43 916 this study available, in particular Helen Glaves, Graham Tulloch, and the late Robin Wingfield.
44
45 917 Douglas Peacock, Vera Pospelova, Kenneth Mertens, John England, and David Evans are also highly
46
47 918 deserving of thanks for fruitful discussions essential to the development of this paper. AJP wishes to
48
49 919 thank Charlie Schweger and Harvey Friebe at the University of Alberta for the generous use of their
50
51 920 Environmental Archaeology Laboratory in the preparation and processing of dinoflagellate cysts and
52
53 921 other non-pollen playnomorphs. This is a contribution to the Climate Change Consortium of Wales
54
55 922 (C3W).
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For Review Only

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3 1310 **List of Figures and Tables**

4
5 1311 **Fig. 1** Map of study area and inset of northern and central Celtic Sea continental shelf showing
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7 1312 bathymetry and core locations.

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11 1314 **Fig. 2** Stratigraphies of British Geological Survey vibrocores from the northern and central Celtic Sea
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13 1315 used in this study. For core locations and details, refer to Fig. 1 and Table 1. Lithostratigraphic codes
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15 1316 are modified from Eyles *et al.* (1983). Dm = matrix-supported diamicton; F = fines; Gm = matrix-
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17 1317 supported gravel; S = sand; h = horizontal laminations; l = fine laminations; (l) = very faint fine
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19 1318 laminations; m = massive; -s = shelly.

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23 1320 **Fig. 3** Absolute abundances of macrofossil groups in analyzed British Geological Survey vibrocores
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25 1321 (Fig. 1) from the Celtic Sea. Core samples are grouped by regionally-extensive stratigraphic units I, II,
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27 1322 and III present in the study area, including subunits.

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31 1324 **Fig. 4** Results of organic-walled microfossil analyses on the Celtic Sea samples. (A) Dinoflagellate
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33 1325 cysts. (B) Other non-pollen palynomorphs and pollen and spores. Core samples are grouped by
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35 1326 regionally-extensive stratigraphic units I, II, and III present in the study area, including subunits.

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39 1328 **Table 1** Details of British Geological Survey vibrocore examined in the present study, including core
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41 1329 number, location, and core code used in the text.

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45 1331 **Table 2** Details of radiocarbon dates used in this study.

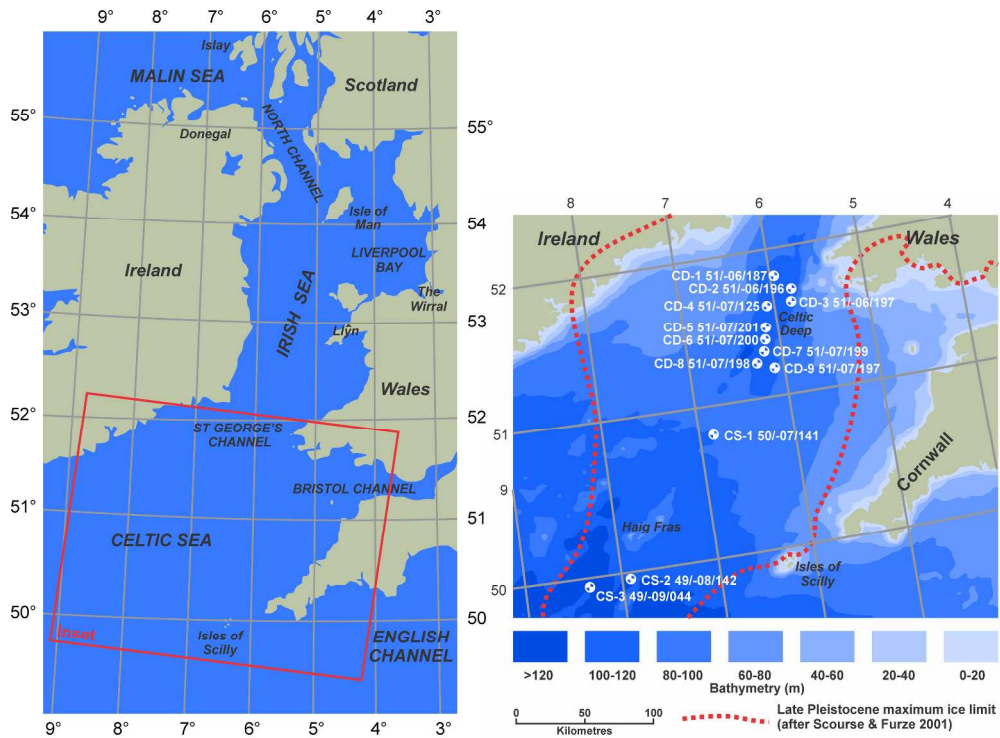


Fig. 1 Map of study area and inset of northern and central Celtic Sea continental shelf showing bathymetry and core locations.
124x90mm (600 x 600 DPI)

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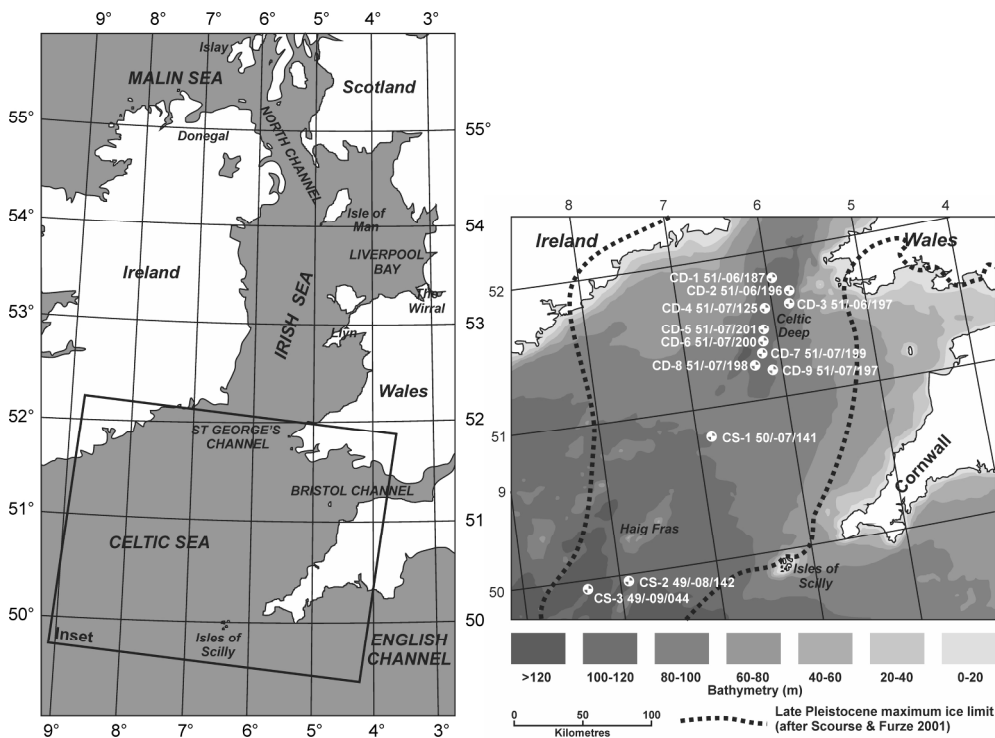


Fig. 1 Map of study area and inset of northern and central Celtic Sea continental shelf showing bathymetry and core locations.
124x90mm (600 x 600 DPI)

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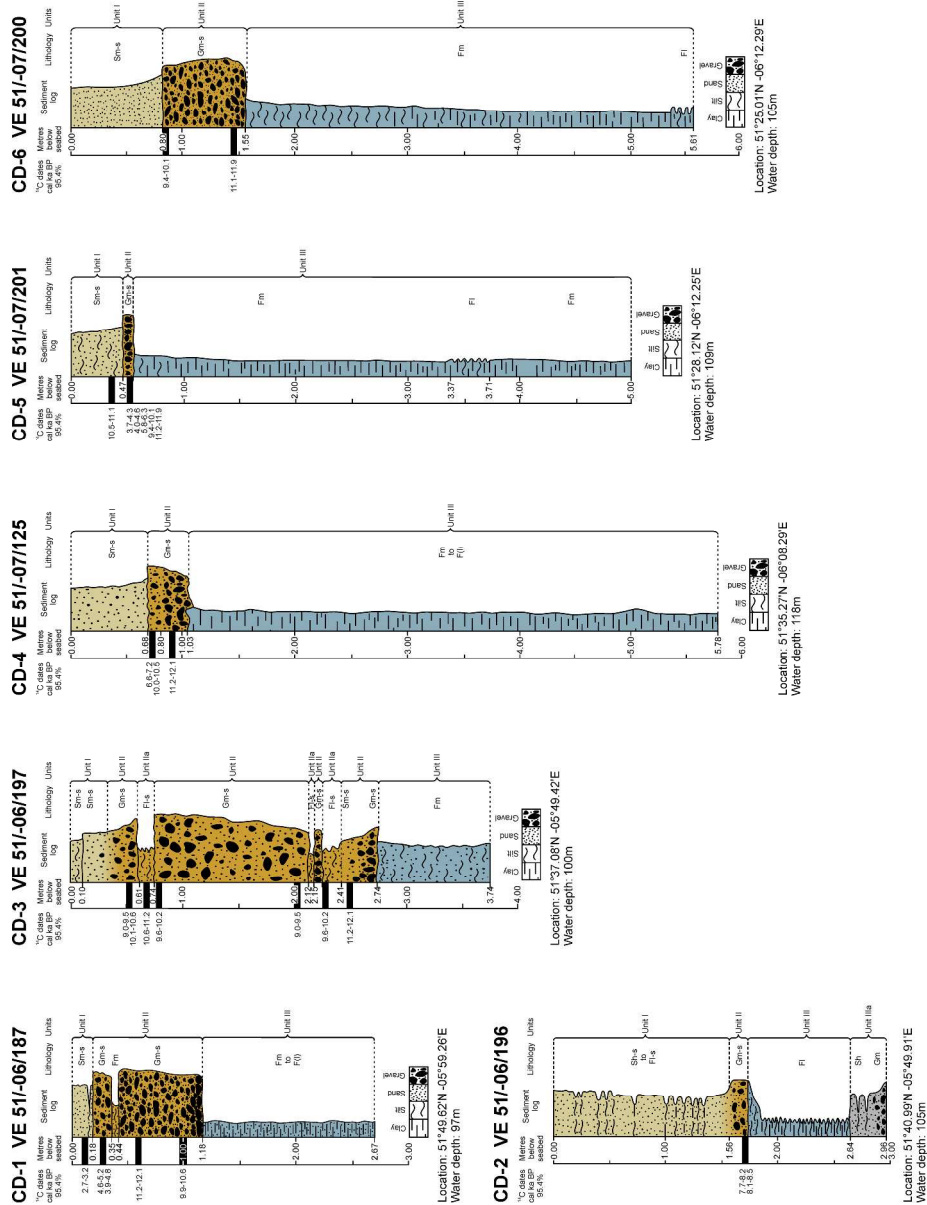


Fig. 2 Stratigraphies of British Geological Survey vibrocores from the northern and central Celtic Sea used in this study. For core locations and details, refer to Fig. 1 and Table 1. Lithostratigraphic codes are modified from Eyles et al. (1983). Dm = matrix-supported diamicton; F = fines; Gm = matrix-supported gravel; S = sand; h = horizontal laminations; l = fine laminations; (l) = very faint fine laminations; m = massive; -s = shelly.

222x292mm (600 x 600 DPI)

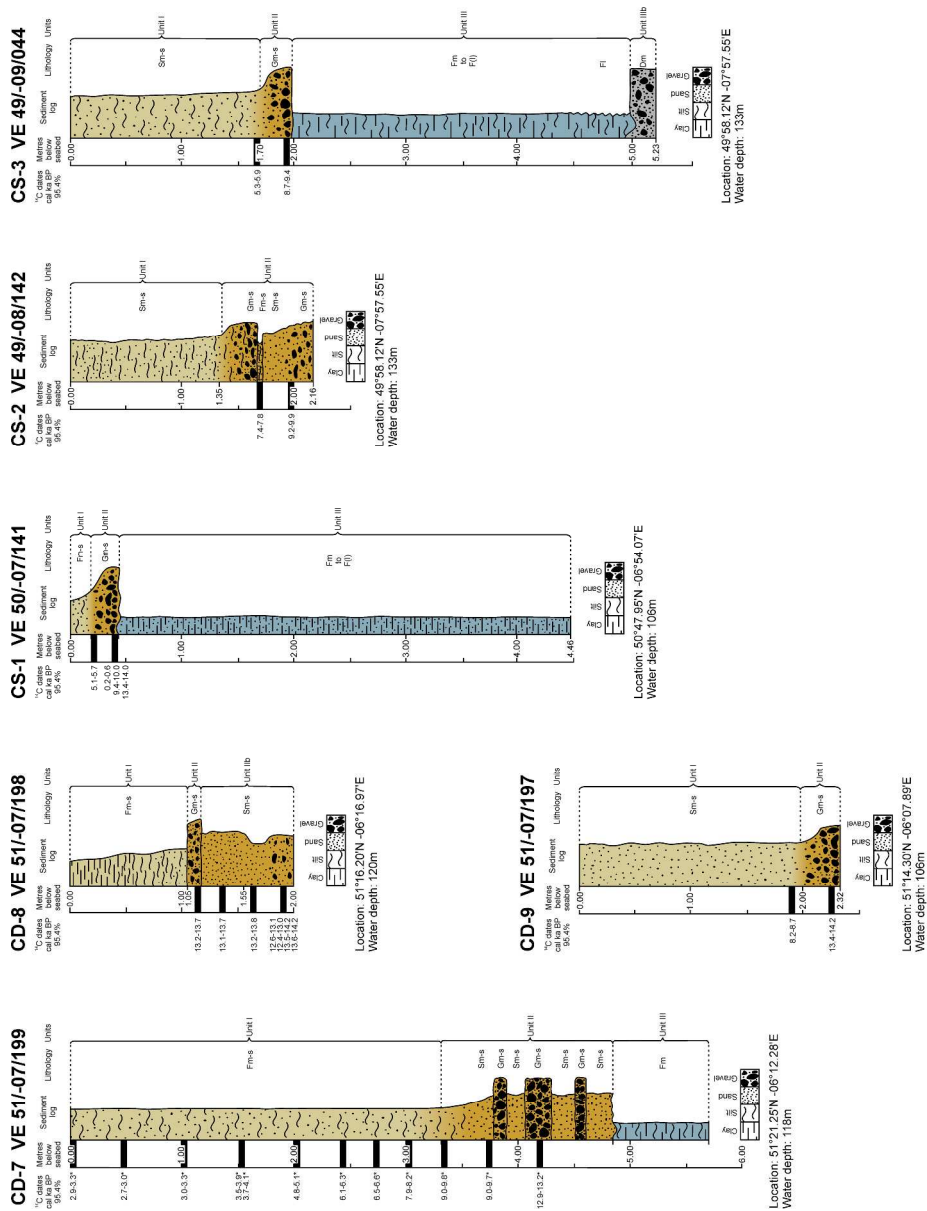


Fig. 2 Stratigraphies of British Geological Survey vibrocores from the northern and central Celtic Sea used in this study. For core locations and details, refer to Fig. 1 and Table 1. Lithostratigraphic codes are modified from Eyles et al. (1983). Dm = matrix-supported diamicton; F = fines; Gm = matrix-supported gravel; S = sand; h = horizontal laminations; l = fine laminations; (l) = very faint fine laminations; m = massive; -s = shelly.

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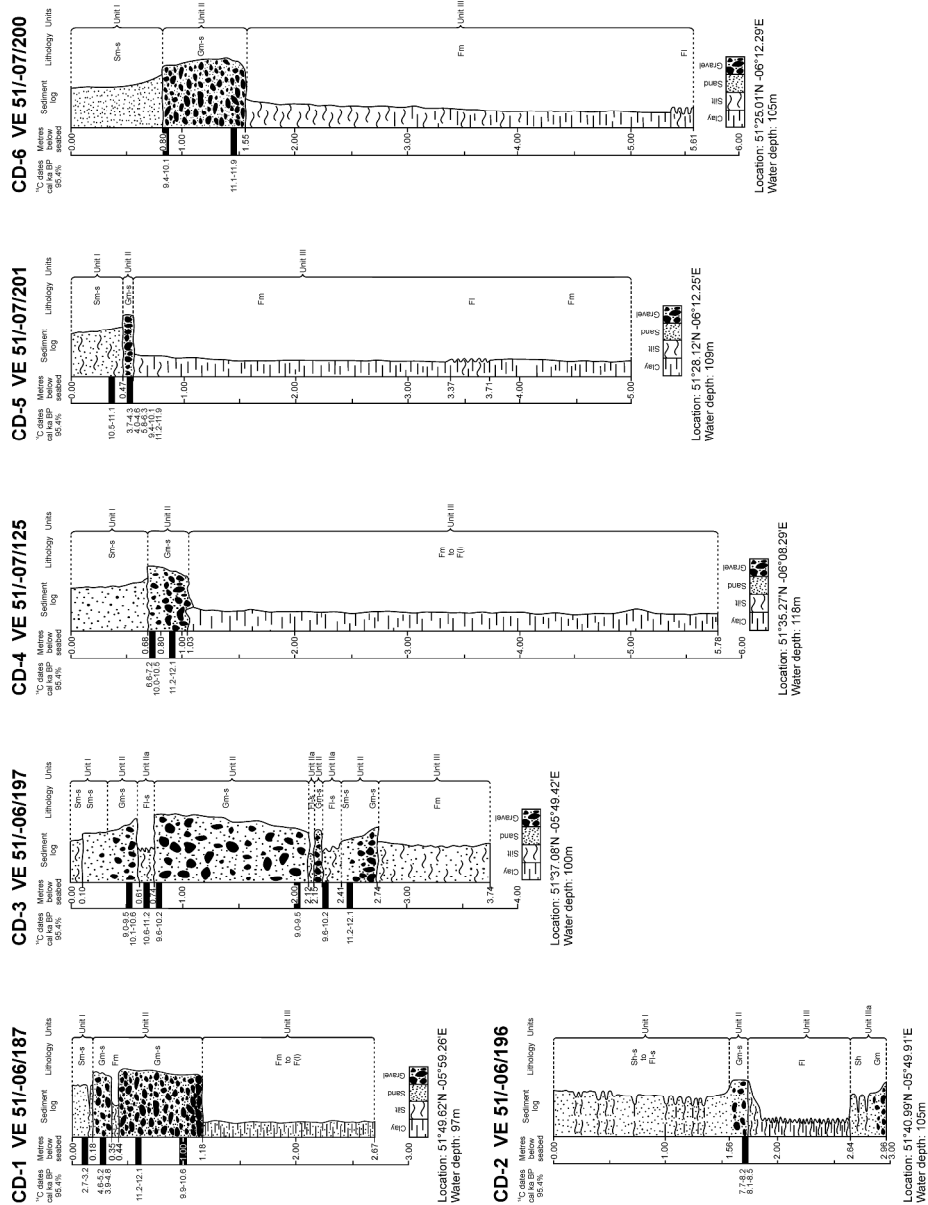


Fig. 2 Stratigraphies of British Geological Survey vibrocores from the northern and central Celtic Sea used in this study. For core locations and details, refer to Fig. 1 and Table 1. Lithostratigraphic codes are modified from Eyles et al. (1983). Dm = matrix-supported diamicton; F = fines; Gm = matrix-supported gravel; S = sand; h = horizontal laminations; l = fine laminations; (l) = very faint fine laminations; m = massive; -s = shelly.

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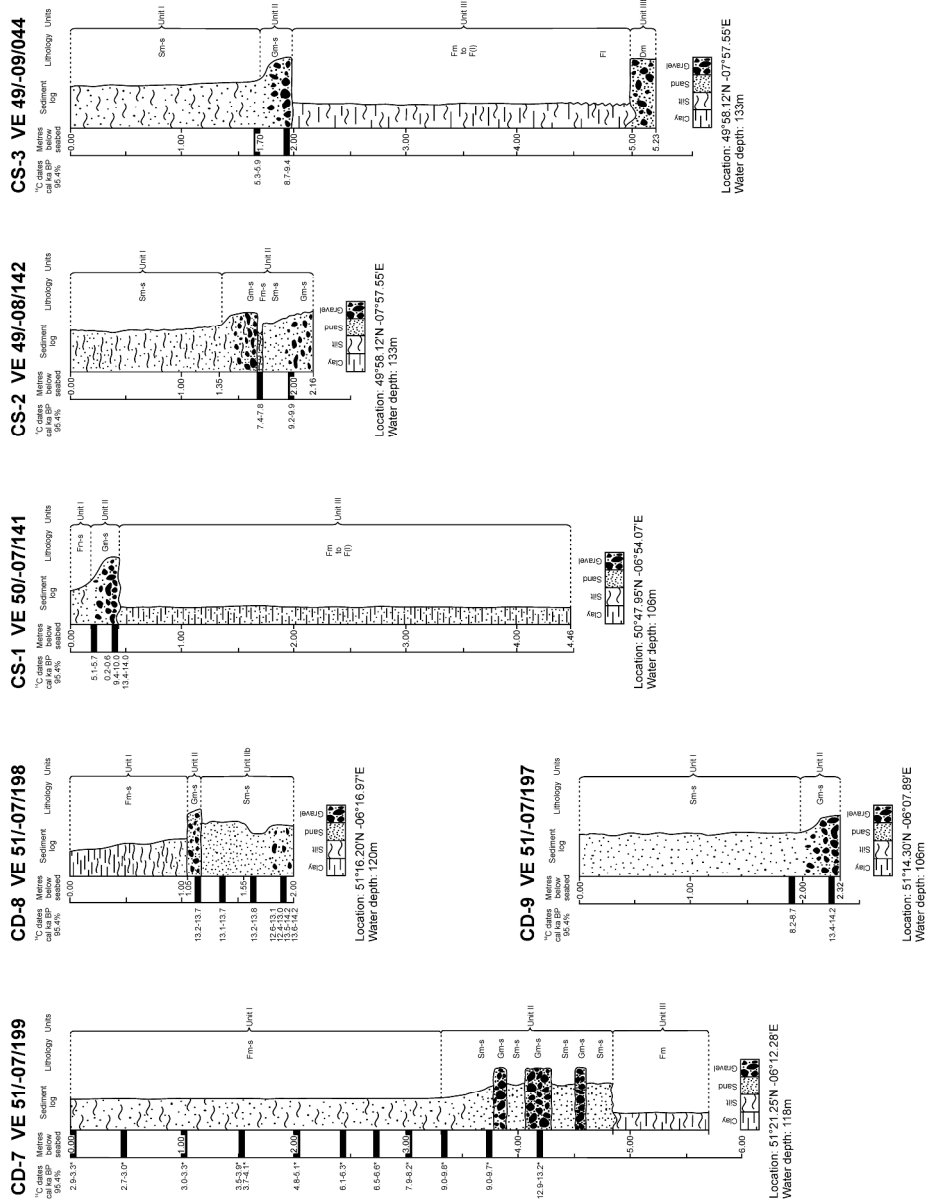


Fig. 2 Stratigraphies of British Geological Survey vibrocores from the northern and central Celtic Sea used in this study. For core locations and details, refer to Fig. 1 and Table 1. Lithostratigraphic codes are modified from Eyles et al. (1983). Dm = matrix-supported diamicton; F = fines; Gm = matrix-supported gravel; S = sand; h = horizontal laminations; l = fine laminations; (l) = very faint fine laminations; m = massive; -s = shelly.

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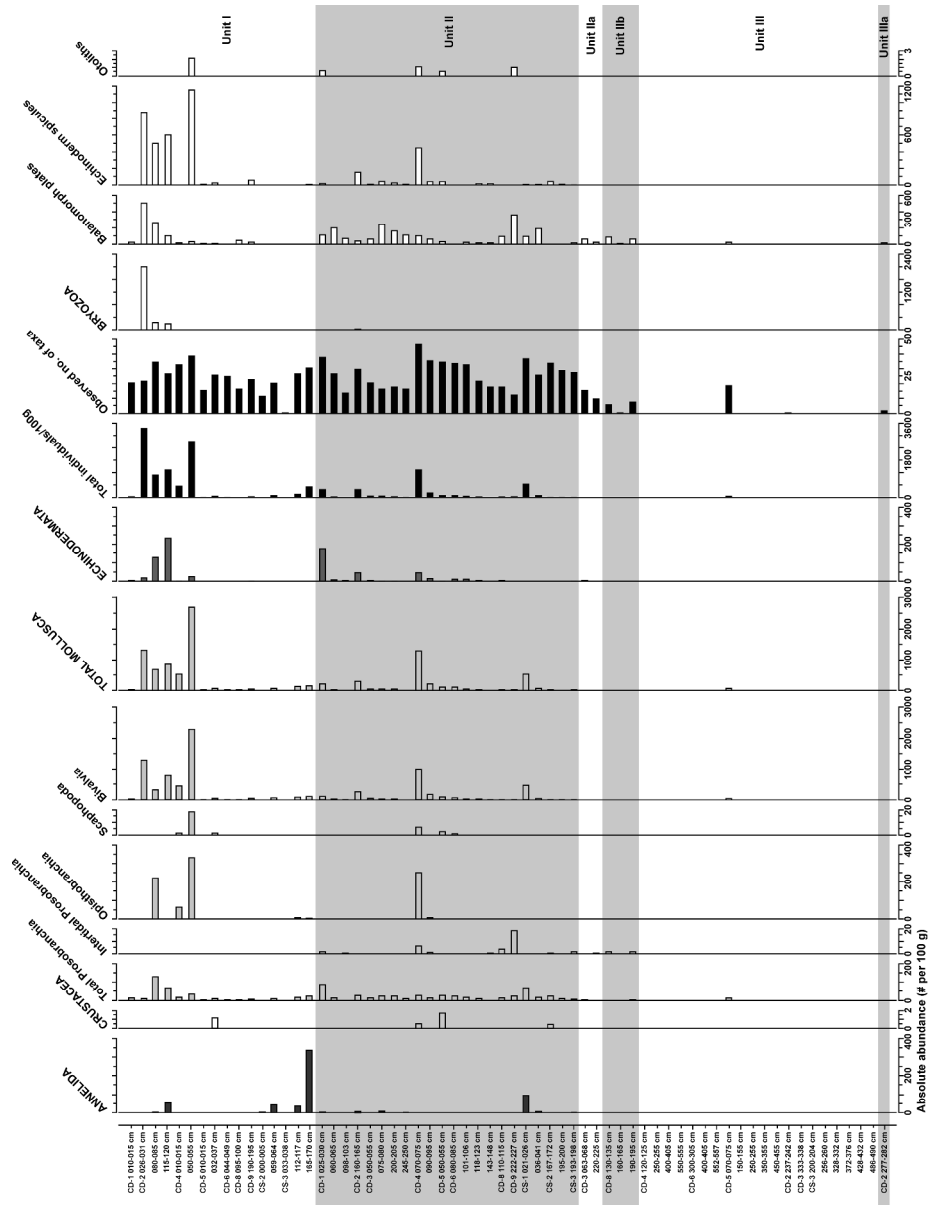


Fig. 3 Absolute abundances of macrofossil groups in analyzed British Geological Survey vibrocores (Fig. 1) from the Celtic Sea. Core samples are grouped by regionally-extensive stratigraphic units I, II, and III present in the study area, including subunits.
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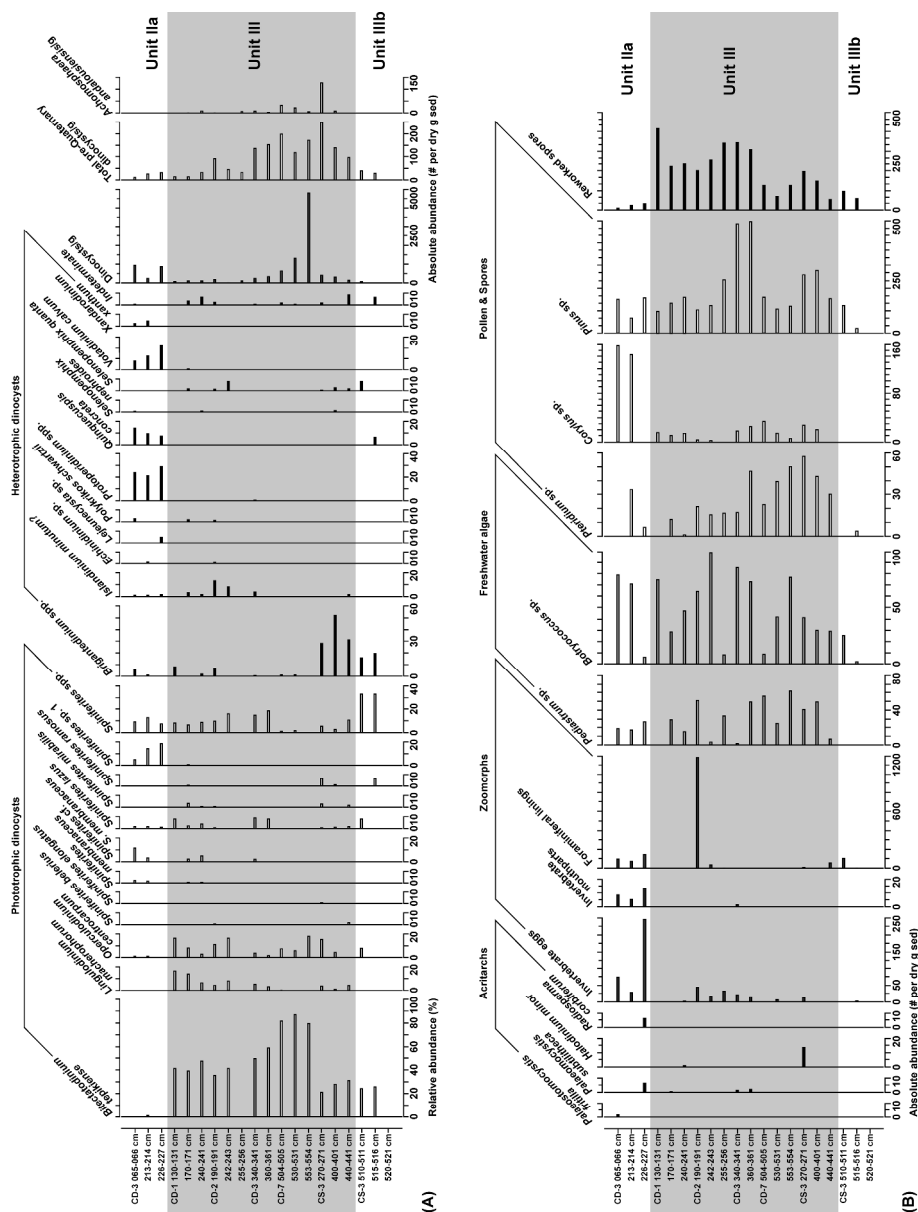


Fig. 4 Results of organic-walled microfossil analyses on the Celtic Sea samples. (A) Dinoflagellate cysts. (B) Other non-pollen palynomorphs and pollen and spores. Core samples are grouped by regionally-extensive stratigraphic units I, II, and III present in the study area, including subunits. 222x295mm (600 x 600 DPI)

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<i>Core Code</i>	<i>BGS Core Number</i>	<i>Lat. N</i>	<i>Long. W</i>	<i>Water Depth (m)</i>	<i>Core Length (cm)</i>	<i>Research Vessel</i>	<i>Recovery Date</i>
CD-1	VE 51/-06/187	51° 49.62'	05° 59.26'	97	267	MV Steelfish	8/12/1973
CD-2	VE 51/-06/196	51° 40.99'	05° 49.91'	105	296	MV Steelfish	8/13/1973
CD-3	VE 51/-06/197	51° 37.08'	05° 49.42'	100	374	MV Steelfish	8/13/1973
CD-4	VE 51/-07/125	51° 35.27'	06° 08.29'	118	578	MV Whitehorn	9/4/1982
CD-5	VE 51/-07/201	51° 28.12'	06° 12.25'	109	500	MV Whitehorn	9/8/1982
CD-6	VE 51/-07/200	51° 25.01'	06° 12.29'	105	561	MV Whitehorn	9/8/1982
CD-7	VE 51/-07/199	51° 21.25'	06° 12.23'	118	568	MV Whitehorn	9/8/1982
CD-8	VE 51/-07/198	51° 16.20'	06° 16.97'	120	200	MV Whitehorn	9/8/1982
CD-9	VE 51/-07/197	51° 14.30'	06° 07.89'	106	232	MV Whitehorn	9/8/1982
CS-1	VE 50/-07/141	50° 47.95'	06° 54.07'	106	446	MV Briarthorn	6/3/1974
CS-2	VE 49/-08/142	49° 58.12'	07° 57.55'	133	216	MV Emerald	6/18/1977
CS-3	VE 49/-09/044	49° 57.80'	08° 20.24'	127	523	MV Cape Shore	8/7/1978

Review Only

Core	BGS	Sample	Litho.	Laboratory			Median Prob. Age	Calibrated Ages Ranges (cal yrs BP)					
Code	Core Number	Depth (cm)	Unit	Material	Description	Number	$\delta^{13}C$	^{14}C Age	\pm	(cal yrs BP)	68.2% interval	95.4% interval	
4	CD-1	VE 51/-06/187	010-015	I	Bivalve	<i>Mimachlamys varia</i>	AA-36237	1.60	3045	50	2870	2980 - 2740	3180 - 2660
5			025-030	II	Gastropod	<i>Hydrobia ulvae</i> *	OS-79779	2.16	4630	25	4910	5040 - 4780	5220 - 4640
6			025-030	II	Bivalve	<i>Mimachlamys varia</i>	AA-36238	2.70	4100	50	4200	4360 - 4060	4480 - 3900
7			060-065	II	Bivalve	<i>Spisula elliptica</i>	AA-36239	1.60	10 445	75	11 620	11 860 - 11 360	12 050 - 11 230
8			098-103	II	Bivalve	<i>Spisula elliptica</i>	AA-36240	1.60	9390	95	10 280	10 450 - 10 150	10 550 - 9910
9	CD-2	VE 51/-06/196	160-165	II	Bivalve	<i>Timoclea ovata</i>	AA-36185	0.60	7450	55	7950	8060 - 7820	8170 - 7710
10			160-165	II	Bivalve	<i>Nucula sulcata</i> †	AA-36186	1.50	7810	60	8300	8400 - 8180	8530 - 8060
12	CD-3	VE 51/-06/197	050-055	II	Gastropod	<i>Tornus subcarinatus</i> *	OS-79219	1.39	8630	60	9320	9460 - 9200	9520 - 9030
13			050-055	II	Bivalve	<i>Venus casina</i>	AA-36241	0.90	9430	70	10 320	10 450 - 10 200	10 570 - 10 050
14			063-068	II a	Bivalve	<i>Ostrea edulis</i>	AA-36242	1.90	9925	90	10 900	11 100 - 10 740	11 170 - 10 580
15			075-080	II	Bivalve	<i>Aequipecten opercularis</i>	AA-36243	1.00	9140	70	9950	10 140 - 9800	10 220 - 9590
16			200-205	II	Bivalve	<i>Spisula elliptica</i>	AA-36244	1.40	8605	65	9290	9430 - 9160	9500 - 9010
17			220-225	II a	Bivalve	<i>Abra</i> cf. <i>alba</i> , left valve	OS-79103	-0.23	9110	40	9910	10 100 - 9770	10 180 - 9600
18			245-250	II a	Bivalve	<i>Spisula elliptica</i>	AA-36245	1.40	10 445	85	11 630	11 860 - 11 360	12 070 - 11 220
19	CD-4	VE 51/-07/125	070-075	II	Gastropod	<i>Crisilla semistriata</i> *	OS-79220	2.15	9390	70	10 280	10420 - 10 160	10 540 - 9960
20			070-075	II	Gastropod	<i>Turritella communis</i>	AA-32278	3.20	6345	60	6860	6990 - 6710	7150 - 6610
21			090-095	II	Bivalve	<i>Nucula turgida</i> †	AA-32279	1.60	10 460	80	11650	11 880 - 11 390	12 080 - 11 230
22	CD-5	VE 51/-07/201	032-037	I	Bivalve	<i>Nucula sulcata</i> †	AA-32282	1.10	9805	85	10 770	10 930 - 10 570	11 120 - 10 500
23			050-055	II	Gastropod	<i>Turritella communis</i>	AA-32283	2.90	4190	65	4310	4470 - 4150	4630 - 3980
24			050-055	II	Gastropod	<i>Turritella communis</i>	AA-32284	3.10	3960	65	4010	4150 - 3840	4330 - 3690
25			050-055	II	Gastropod	<i>Turritella communis</i>	AA-32285	3.20	5620	65	6060	6180 - 5930	6290 - 5790
26			050-055	II	Bivalve	<i>Spisula elliptica</i>	AA-32286	1.50	8955	90	9710	9870 - 9520	10 110 - 9440
27			050-055	II	Bivalve	<i>Nucula sulcata</i> †	AA-32287	1.30	10 330	90	11 480	11 650 - 11 220	11 920 - 11 150
28	CD-6	VE 51/-07/200	080-085	II	Bivalve	<i>Spisula elliptica</i>	AA-32280	1.60	8940	65	9680	9820 - 9510	10 070 - 9440
29			143-148	II	Bivalve	<i>Nucula sulcata</i> †	AA-32281	1.40	10 305	75	11 430	11 630 - 11 200	11 870 - 11 140
31	CD-8	VE 51/-07/198	110-115	II	Bivalve	<i>Nucula sulcata</i> †	AA-36248	1.60	11 935	80	13 420	13 570 - 13 280	13 720 - 13 190
32			130-135	II b	Bivalve	<i>Nucula sulcata</i> †	AA-36249	1.70	11 895	80	13 380	13 500 - 13 240	13 670 - 13 140
33			160-165	II b	Bivalve	<i>Spisula elliptica</i>	AA-36250	2.10	11 970	100	13 460	13 600 - 13 310	13 760 - 13 210
34			190-195	II b	Gastropod	<i>H. ulvae</i> x1, <i>R. parva</i> x1 *	OS-79120	0.69	12 400	50	13 890	14 000 - 13 770	14 150 - 13 630
35			190-195	II b	Gastropod	cf. <i>Rissoa parva</i> *	OS-79121	1.70	11 250	45	12 770	12 870 - 12 640	13 070 - 12 580
36			190-195	II b	Gastropod	<i>Hydrobia</i> cf. <i>ulvae</i> *	OS-79145	0.89	11 150	60	12 680	12 780 - 12 570	12 950 - 12 400
37			190-195	II b	Bivalve	<i>Spisula elliptica</i>	AA-36251	2.00	12 410	85	13 900	14 030 - 13 760	14 210 - 13 490
38	CD-9	VE 51/-07/197	190-195	I	Bivalve	<i>S. elliptica</i> x1, <i>T. ovata</i> x1	AA-36246	0.60	7945	65	8450	8570 - 8330	8730 - 8180
39			222-227	II	Bivalve	<i>Modiolus modiolus</i>	AA-36247	-0.20	12 350	120	13 840	14 020 - 13 660	14 190 - 13 430
41	CS-1	VE 50/-07/141	021-026	II	Bivalve	<i>Dosinia lupinus</i>	AA-36187	1.20	5050	50	5440	5550 - 5320	5680 - 5140
42			036-041	II	Gastropod	<i>Tornus subcarinatus</i> x2 *	OS-79139	1.48	8890	60	9620	9740 - 9470	9990 - 9370
43			036-041	II	Gastropod	<i>Rissoella diaphana</i> *	OS-79147	0.84	12 250	80	13 720	13 870 - 13 520	13 980 - 13 430
44			036-041	II	Bivalve	<i>Corbula gibba</i>	AA-36188	1.10	765	55	420	500 - 320	630 - 240

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CS-2	VE 49/-08/142	167-172	II	Bivalve	<i>Gari fervensis</i>	AA-32274	2.90	7085	60	7600	7690 - 7490	7820 - 7410
		195-200	II	Bivalve	<i>Spisula elliptica</i>	AA-32275	1.90	8785	80	9490	9640 - 9340	9850 - 9190
CS-3	VE 49/-09/044	165-170	I	Polychaete	<i>Ditrupa arietina</i>	AA-32276	2.00	5220	65	5620	5730 - 5480	5880 - 5340
		193-198	II	Bivalve	<i>Spisula elliptica</i>	AA-32277	1.50	8420	70	9070	9260 - 8940	9390 - 8720

Dates calibrated using CALIB 6.0 (Stuiver *et al.* 2010) and MARINE09 calibration curve (Reimer *et al.* 2009). $\Delta R = -33 \pm 93$ (Reimer *et al.* 2002)

AA dates = NERC - University of Arizona dates, year of assay 1998/99 OS dates = NOSAMS dates, year of assay 2010

Species marked by * are grazers (gastropods), species marked by † are deposit feeders (bivalves) and likely subject to enhanced radiocarbon age effects due to the ingestion of "old" carbonate (England *et al.* 2012), all other molluscs and polychaete dates are on filter-feeding taxa.

Species names according to World Register of Marine Species (Appeltans *et al.* 2011)

