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Retreat dynamics of the eastern sector of the British–Irish Ice Sheet during the last glaciation

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ABSTRACT: The findings of BRITICE-CHRONO Transect 2 through the North Sea Basin and eastern England are reported. We define ice-sheet marginal oscillation between ~31 and 16 ka, with seven distinctive former ice-sheet limits (L1–7) constrained by Bayesian statistical analysis. The southernmost limit of the North Sea Lobe is recorded by the Bolders Bank Formation (L1; 25.8–24.6 ka). L2 represents ice-sheet oscillation and early retreat to the northern edge of the Dogger Bank (23.5–22.2 ka), with the Garret Hill Moraine in north Norfolk recording a significant regional readvance to L3 at 21.5–20.8 ka. Ice-marginal oscillations at ~26–21 ka resulted in L1, L2 and L3 being partially to totally overprinted. Ice-dammed lakes related to L1–3, including Lake Humber, are dated at 24.1–22.3 ka. Ice-sheet oscillation and retreat from L4 to L5 occurred between 19.7 and 17.3 ka, with grounding zone wedges marking an important transition from terrestrial to marine tidewater conditions, triggered by the opening of the Dogger Lake spillway between 19.9 and 17.5 ka. L6 relates to ice retreat under glacimarine conditions and final ice retreat into the Firth of Forth by 15.8 ka. L7 (~15 ka) represents an ice retreat from Bosies Bank into the Moray Firth.

KEYWORDS: British–Irish Ice Sheet; ice-sheet flow dynamics; MIS2; North Sea Lobe; palaeoglaciology

Introduction

This paper reviews the results of the BRITICE-CHRONO Transect 2 through the North Sea Basin and onto the onshore areas of eastern England in the context of pre-existing knowledge on the behaviour of the British–Irish Ice Sheet (BIIS) in the region. It is a critical region for palaeoglaciological reconstruction because it was characterised by confluence between the BIIS and the Fennoscandinavian Ice Sheet (FIS), where ice flow was highly dynamic (and possibly surge-influenced) and ice streams appear to have switched on and off through the last glacial cycle (cf. Balson & Jeffrey 1991; Eyles *et al.* 1994; Carr *et al.* 2006; Hubbard *et al.* 2009; Sejrup *et al.* 2009, 2016; Boston *et al.* 2010; Graham *et al.* 2011; Clark *et al.* 2012). In the north and central North Sea, subglacial landforms and deposits are overlain by glacimarine sediments that record marine inundation following deglaciation. In contrast, in the south, the shallow seas of the Dogger Bank and offshore areas of Yorkshire, Lincolnshire and Norfolk contain a record of deglaciation in a terrestrial setting, with subglacial landscapes overprinted by glacilacustrine, glacifluvial outwash and Holocene marine sediments (e.g. Dove *et al.* 2017; Roberts *et al.* 2018).

Recent hypotheses suggest that ice-sheet coalescence between the BIIS and FIS over the central North Sea occurred throughout the Last Glacial Maximum (LGM; 30–23 ka), but was followed by regional decoupling and rapid flow instability of the BIIS (Sejrup *et al.* 2000; Graham *et al.* 2007, 2010; Bradwell *et al.* 2008, 2019). This was triggered first by the retreat of the Norwegian Channel Ice Stream (NCIS) and second by the catastrophic northerly drainage of the Dogger Lake complex between 19 and 17.5 ka in the southern North Sea (Sejrup *et al.*, 2016). However, despite many years of work, several key questions remain unanswered. Working in the North Sea region, Transect 2 of the BRITICE-CHRONO project specifically aimed to explore several key questions: 1) how far did the BIIS extend at the LGM?; 2) how, when and where did the BIIS and FIS uncouple?; 3) did marine inundation of the northern North Sea destabilise the BIIS and thereby trigger rapid east-to-west retreat?; and 4) what were the

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major BIIS controls on North Sea ice flow dynamics and deglaciation patterns?

The findings of Transect 2 are summarised here based upon new data from a number of localised case studies that were designed to further interrogate the existing glacial geomorphology, Late Quaternary stratigraphy and geochronology in key locations around the British sector of the North Sea Basin (Evans *et al.* 2017, 2018, 2019; Bateman *et al.* 2018; Roberts *et al.* 2018, 2019), as well as concurrent reports on glacial reconstructions prepared outwith, but at the same time complementary to, the BRITICE-CHRONO project (e.g. Murton *et al.* 2009; Bateman *et al.* 2008, 2015; Fairburn & Bateman 2016; Dove *et al.* 2017; Fig. 1). The findings reported in these case studies are now critically reviewed in the context of the regional reconstructions of BIIS palaeoglaciology as it pertains to the eastern margins of the ice sheet. The North Sea is subdivided into two sectors: the northern sector, extending from the Firth of Forth to the Moray Firth and east to Fladen Ground and Viking Bank/Plateau; and the southern sector, extending from the Firth of Forth to Great Fisher Bank and southwards into the southern North Sea Basin towards Dogger Bank and the Yorkshire, Lincolnshire and Norfolk coasts.

Glacial land systems of the North Sea region

The North Sea Basin during the Late Quaternary cold stages was characterised by subaerial conditions as a result of the low stands of glacioeustatic sea levels (e.g. Waelbroeck *et al.* 2002). Chronostratigraphic control relating to this period of time is provided by the critical exposure at Sewerby, near Bridlington on the East Yorkshire coast. The stratigraphy here records beach sedimentation during Marine Isotope (MIS) Sub-Stage 5e, followed by MIS 4 periglacial colluvial and aeolian sedimentation on an exposed marine cliff and platform, and capped by the MIS 2 Skipsea Till and overlying outwash (Lamplugh 1888; Boylan 1967; Bateman & Catt 1996; Bateman *et al.* 2015). During the last cold stage, culminating in the LGM of MIS 2, it has been proposed that the BIIS and FIS coalesced, although this reconstruction has not been without controversy. Recent research has addressed the pattern and chronology of ice-sheet occupancy of the North Sea region during MIS 2, identifying complex and dynamic interaction between the FIS and the BIIS (cf. Nygård *et al.*, 2005; Carr *et al.*, 2006; Bradwell *et al.*, 2008, 2019; Sejrup *et al.*, 2009, 2015; Hjelstuen *et al.* 2018). Multiple advances and recessions, as well as ice-flow directional switches driven by shifting ice dispersal centres, have been recorded from complex glacial landform and stratigraphic assemblages related to the central BIIS (e.g. Livingstone *et al.*, 2008, 2010a, b, 2015; Davies *et al.* 2009, 2011, 2012; Evans *et al.*, 2009; Roberts *et al.* 2013) and consequently are now firmly established elements of the BIIS palaeoglaciology (cf. Clark *et al.* 2012; Livingstone *et al.*, 2012; Chiverrell *et al.* 2013; Hughes *et al.* 2014; Sejrup *et al.*, 2015).

The interaction of lobate ice streams in the central and eastern parts of the BIIS during MIS 2 is recorded by substantial arcuate moraine sequences and intervening glacial lake plains (Clark *et al.* 2004, 2012, 2018; Evans *et al.* 2005; Fig. 1). The most significant of these ice bodies is the North Sea Lobe (NSL), which flowed southwards for some 400 km from southern Scotland down the eastern margin of the North Sea Basin (Catt & Penny 1966; Boulton *et al.* 1985) to terminate in north Norfolk (Pawley *et al.* 2006; Moorlock *et al.* 2008; Roberts *et al.* 2018; Evans *et al.* 2019; Fig. 1). The NSL also coalesced with the Tweed, Tyne Gap and Stainmore ice streams, forming ice-dammed lakes such as Lake Tyne, Lake Wear and Lake Tees during phases of ice-marginal recession (Agar 1954; Smith 1981; Plater *et al.* 2000; Davies *et al.* 2009, 2019; Livingstone *et al.* 2012, 2015; Yorke *et al.* 2012). Around the North Yorkshire Moors, ice-dammed lakes were created in Kildale and the Vale of Pickering by onshore flow of the NSL at its maximum extent during MIS 2 (Kendall 1902; Evans *et al.* 2017; Lincoln *et al.* 2017; Fig. 2). In the Vale of Pickering, the NSL plugged the valley mouth with substantial morainic topography (Filey and Flamborough moraines; Kendall 1902; Farrington & Mitchell 1951), thereby permanently reversing the flow of the River Derwent (King 1965; Evans *et al.* 2017). The most substantial lake created between the Pennine-sourced ice and the NSL existed in the Vale of York/Humber lowland region, where the 125 km long Vale of York ice lobe flowed from the Stainmore Gap and Tees lowlands to terminate within 10 km of the onshore-flowing NSL, plugging the Humber Estuary to construct the Ellingham/Ferriby/Horkstow moraine complex and thereby creating Glacial Lake Humber (Gaunt 1974, 1981; Straw 1979; Bateman *et al.* 2000, 2008, 2015, 2018; Fig. 2). Although Lake Humber highstands of up to ~30 m OD (Ordnance Datum) have been proposed (e.g. Gaunt *et al.* 1992; Fairburn & Bateman 2016), there is unequivocal evidence of glacilacustrine deposits (Hemingbrough Formation) only up to 8 m OD in the Humber lowlands (Thomas 1999; Ford *et al.* 2008). Consequently, a potentially deeper lake (cf. Straw 1979) is regarded as an ephemeral feature during MIS 2 by Evans *et al.* (2018), likely because it was difficult to maintain by the NSL plugging the Humber Gap, and hence highstand Lake Humber was repeatedly terminated by catastrophic drainage in response to ice margin floatation.

Along the east coast of England, the advance of the NSL against the higher coastal topography and its competition with easterly flowing inland ice is recorded by multiple tills. Stratified intra- and inter-till beds such as those at Sandy Bay (Eyles *et al.* 1982), Whitburn (Davies *et al.* 2009) and Holderness (Evans *et al.* 1995; Evans & Thomson 2010) record both switches in ice flow provenance as well as phases of sedimentation in subglacial canal fills that in places fed into ice-marginal lakes. At some locations, multiple tills are separated by stratified sediments of subaerial origin and hence were suitable for optically stimulated luminescence (OSL) dating, best exemplified at sites like Sandsend, north of Whitby (Roberts *et al.* 2013) and Seaham, on the Durham coast (Roberts *et al.* 2019). At Sandsend the stratigraphy displays evidence of an early NSL advance in the form of a bedrock glacitectonite and till which is separated from an upper till, documenting a later readvance, by upward-coarsening stratified sediments of an ice-marginal lake and sandur. Similarly, on the Holderness lowlands, onshore oscillations of the NSL have been proposed by Evans and Thomson (2010) and Boston *et al.* (2010) based upon the regional stratigraphic architecture and through framing genetic interpretations within the principles of advance and retreat tills inherent within the ice-sheet sub-marginal sedimentation model of Boulton (1996a, b). These oscillations have often been associated with surging behaviour (e.g. Lamplugh 1911; Eyles *et al.* 1994; Boston *et al.* 2010) but the lack of a diagnostic surging land system signature (Evans & Rea 1999, 2003) has cautioned against an unequivocal classification of the NSL as a predominantly surging lobe (Evans & Thomson 2010). In addition to the two advances recorded by the Skipsea Till margin (Catt & Penny 1966; Catt 2007) and the Holderness hummocky drift (Eyles *et al.* 1994; Evans & Thomson 2010), a late stage readvance by the NSL is recorded by the Withernsea Till at Dimlington (Catt & Penny 1966). Throughout the oscillations and final recession of the NSL from the coast of eastern England, sedimentation has been dominated by

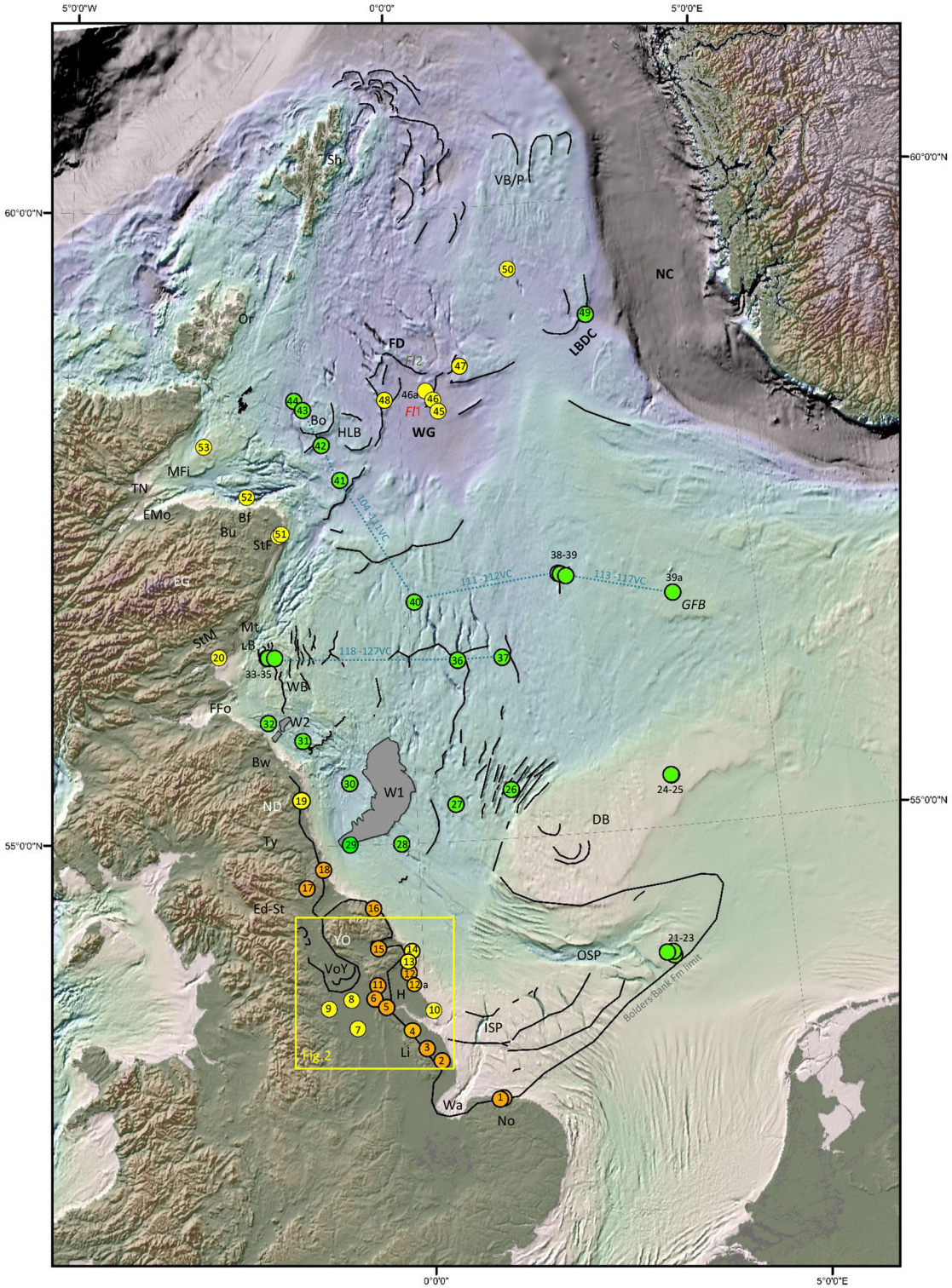

Figure 1. Bathymetry (EMODnet) and topography of the North Sea region, including key site locations and ice margin positions, as demarcated by moraines or the edges of till sheets (black lines) and grounding zone wedges (grey outlined areas labelled W1 and W2). Data derived from multiple published sources and BRITICE version 2 database (cf. Clark *et al.* 2018). BRITICE-CHRONO offshore and onshore sites are denoted by green and orange symbols, respectively. Legacy sites are denoted by yellow symbols. (Bf – Banff; Bw – Berwick; Bo – Bosies Bank; Bu – Buchan; DB – Dogger Bank; EG – East Grampians; EMo – East Morayshire; Ed-ST – Eden/Stainmore; FFo – Firth of Forth; FD – Fladen deeps; Fl1 and Fl2 – Fladen Moraines; GFB – Great Fisher Bank; HLB – Halibut Bank; H – Holderness; Hu – Humber; ISP – Inner Silver Pit; Li – Lincolnshire; LBDC – Ling Bank Drainage Channel; MFi – Moray Firth; No – Norfolk; ND – Northumberland; NC – Norwegian Channel; OSP – Outer Silver Pit; StF – St Fergus; StM – Strathmore; Wa – The Wash; TN – Tarbert Ness; TY – Tyne Gap; VoY – Vale of York; VB/P – Viking Bank/Plateau; WB – Wee Bankie; WG – Witch Ground; W1 and W2 – Grounding zone wedges; YO – Yorkshire). Onshore site locations are designated by numbers used in combination with place names in Fig. 4a and offshore site locations are designated by numbers used in combination with core names in Fig. 4b. These site numbers are used to locate dates in Table 1. Blue dashed lines represent locations of transects in Fig. 5.

the progradation and aggradation of coarse-grained subaqueous glacilacustrine deposits (e.g. Holderness, Fig. 1; Evans & Thomson 2010) and/or ice-contact glacifluvial deposits associated with the development of glacier karst (e.g. Lincolnshire coastal lowlands; Evans *et al.* 2019). This has provided numerous suitable targets for OSL dating (see Table 1 and references therein), which are critically reviewed below as part of a regional palaeoglaciological reconstruction.

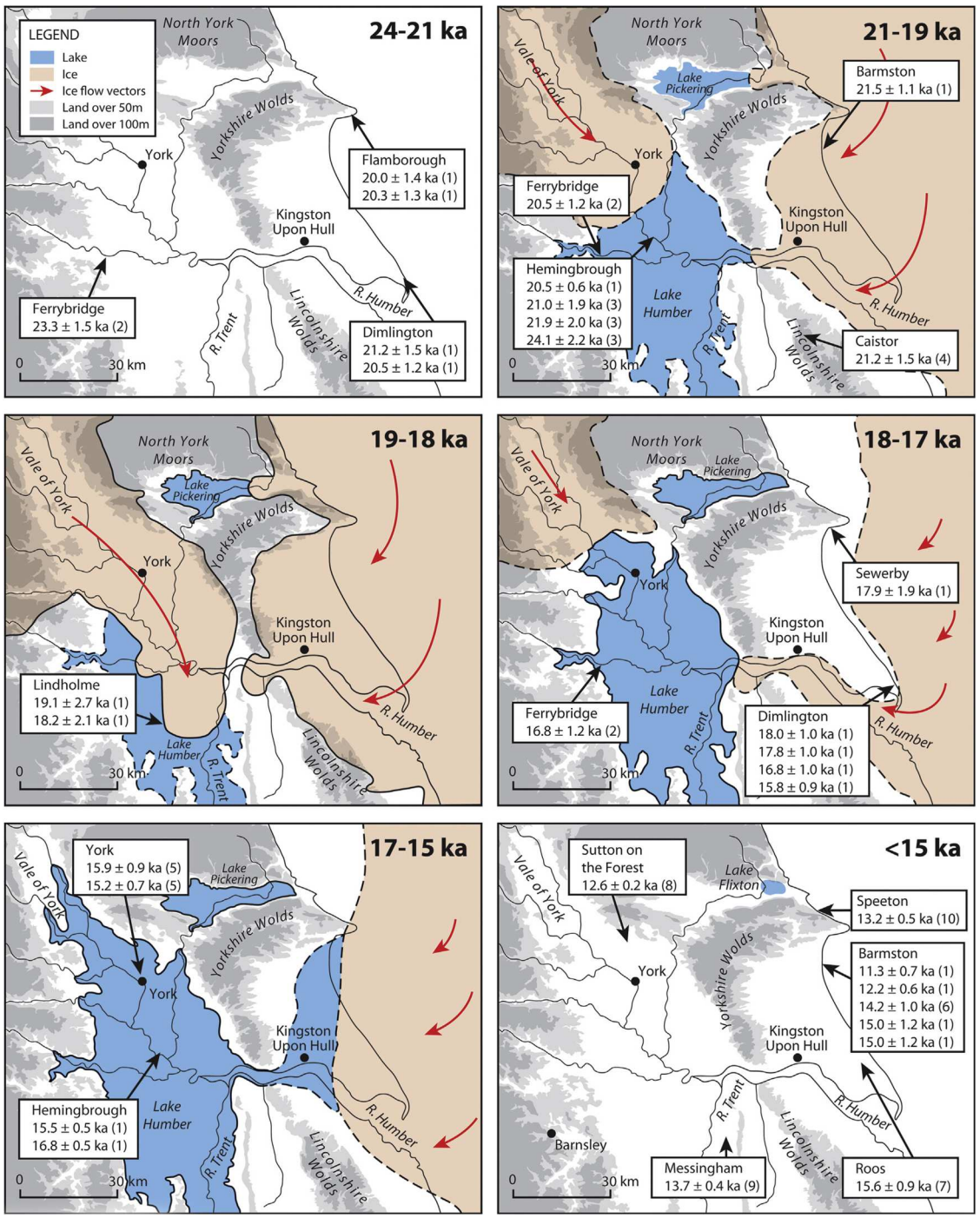

Figure 2. Palaeoglaciology of East Yorkshire and Vale of York/Humber Lowlands (from Bateman *et al.* 2015), showing the relative ice movements of the Vale of York and North Sea lobes of the BIIS. Ice margins are illustrative only and are based, where available, on dated stratigraphic and geomorphic evidence indicating either ice-free or ice-inundated conditions. Within time slices, ice margins may have moved significantly from that shown. Lake Humber margins are based on those of Fairburn and Bateman (2016) and for simplicity do not show internal emergent high ground. Lake Pickering extent is based on Clark *et al.* (2004) with the vestige (Lake Flixton) based on Palmer *et al.* (2015). Ages are based on OSL and radiocarbon dates and are from the following sources (bracketed numbers following ages in location boxes): 1. Bateman *et al.* (2015); 2. Bateman *et al.* (2008); 3. Murton *et al.* (2009); 4. Bateman *et al.* (2000); 5. Fairburn and Bateman (2016); 6. Allen (1980); 7. Beckett (1981); 8. Matthews (1970); 9. Bateman *et al.* (2001); 10. Straw (1979).

Recent investigations on the floor of the North Sea have identified significant landform-sediment assemblages that are critical to the reconstruction of the former dynamics and oscillations of the NSL during MIS 2 in offshore locations (Fig. 1; Dove *et al.* 2017; Roberts *et al.* 2018, 2019; Emery *et al.* 2019). In the central North Sea, the prominent feature of the Dogger Bank has been heavily glacitectonised, which Phillips *et al.* (2017) relate to complex thrust moraine construction by a dynamic NSL margin during MIS 2 ice-sheet recession. The shallow seafloor to the west and southwest of the Dogger Bank also contains evidence of several later phases of ice advance and retreat related to the flow of the NSL along the depression located between the Dogger Bank and the Yorkshire/Lincolnshire coasts down to north Norfolk. Roberts *et al.* (2018) propose that the early phases of the LGM were characterised by the interaction between the southern margin of the BIIS and Dogger Lake, located in the eastern sector of the central North Sea. This water body initiated flow instability in the NSL, which was consequently subject to rapid ice advance and retreat. In contrast, during the later stages of the LGM the flow of the NSL was restricted to the western side of the Dogger Bank

Table 1. Dates used in compiling the chronology of ice recession in this paper, organised from south to north. Highlights are blue for OSL, yellow for TL and green for ^{14}C. Where more than one OSL measurement approach has been taken for the same sample, SA denotes single aliquot measurements and SG denotes single grain measurements.

Site N° (see Fig. 1)	Site name/ core name	OSL/^{14}C Sample code	Lat/ Long	Source	Depth (m)	Previously reported OSL (ka) or ^{14}C (cal ka BP)	Revised or new age (cal ka BP)	Bayesian median modelled age (ka)
1	Garret Hill	Shfd15033	52.954	Roberts *et al.* 2018	1.1	21.5±1.3		21.5
1	Garret Hill	Shfd15034	–0.974	Roberts *et al.* 2018	1.5	22.8±1.8		22.8
2	Thoresthorpe	Shfd15154	53.273	Evans *et al.* 2019	2.2	17.4±1.1		17.4
2	Thoresthorpe	Shfd15155 (SG)	–0.189	Evans *et al.* 2019	2.4	19.3±1.4		19.6
		Shfd15155 (SA)				19.3±2.2		19.6
3	Welton-le-Wold	Shfd13074	53.374	Evans *et al.* 2019	3.5	19.5±2.6		19.5
			–0.068					
4	North Kelsey Road, Caistor	Back 1	53.494	Bateman *et al.* 2000	2.9	22.7±1.4	24.1±1.5	24.5
4	North Kelsey Road, Caistor	Back 2	–0.353	Bateman *et al.* 2000	1.9	17.7±1.2	19.1±1.1	19.6
5	North Ferriby	Shfd13072	53.712	Bateman *et al.* 2018	3.4	21.7±2.0	27.0±2.2	27.6
5	North Ferriby	Shfd13071	–0.516	Bateman *et al.* 2018	2.4	22.5±1.6	29.5±1.7	29.5
5	North Ferriby	Shfd13073		Bateman *et al.* 2018	1.9	19.4±1.8	21.7±2.0	21.7
6	North Cave	Shfd13060	53.781	Evans *et al.* 2018	0.8	11.1±0.8		
6	North Cave	Shfd13058	–0.668	Evans *et al.* 2018	8.0	37.6±3.3	38.6±3.3	37.1
6	North Cave	Shfd13059		Evans *et al.* 2018	6.4	34.7±3.4	41.8±3.7	37.9
6	North Cave	SUERC-77600	58.781	Evans *et al.* 2018		49.0±3.6	C14 yrs	
6	North Cave	SUERC-79023	–0.672	Evans *et al.* 2018		41.6±1.4	C14 yrs	
7	Lindholme	Shfd10072	53.545	Bateman *et al.* 2015	1.2	19.1±2.7	19.9±3.4	21.9
7	Lindholme	Shfd10071	–0.930	Bateman *et al.* 2015	0.95	18.2±2.1	18.9±3.0	18.9
8	Hemingbrough	Shfd11093	53.776	Bateman *et al.* 2015	14.1	20.5±0.6	22.3±0.7	22.0
8	Hemingbrough	Shfd11087	–0.977	Bateman *et al.* 2015	1.9	16.8±0.5	19.2±0.5	19.6
8	Hemingbrough	Shfd11088		Bateman *et al.* 2015	2.6	15.5±0.5	18.0±0.6	18.0
8	Hemingbrough	HEMA01	53.775	Murton *et al.* 2009	1.5	21.9±2.0		21.9
8	Hemingbrough	HEMA02	–0.976	Murton *et al.* 2009	40	21.0±1.9		21.9
8	Hemingbrough	HEMA03		Murton *et al.* 2009	4.0	24.1±2.2		24.1
9	Ferrybridge	Shfd02012	53.713	Bateman *et al.* 2008	1.1	16.6±1.2	19.0±1.4	
9	Ferrybridge	Shfd02014	–1.286	Bateman *et al.* 2008	1.2	20.5±1.2	20.1±1.2	21.2
9	Ferrybridge	Shfd02013		Bateman *et al.* 2008	2.6	23.3±1.5	24.4±1.6	24.5
10	Dimlington	Shfd07113 (SA)	53.673	Bateman *et al.* 2015	21.6	17.8±1.0	21.2±1.2	21.1
		Shfd07113 (SG)	–0.103			16.9±0.9	20.2±1.2	21.2
10	Dimlington	Shfd07114 (SA)		Bateman *et al.* 2015	20.8	17.6±0.9	21.0±1.1	21.2
		Shfd07114 (SG)				18.0±0.9	21.5±1.1	21.2
10	Dimlington	Shfd07115 (SG)		Bateman *et al.* 2015	20.0	15.7±0.8	18.8±1.0	18.8
10	Dimlington	Shfd07116 (SA)		Bateman *et al.* 2015	18.7	17.0±0.9	20.3±1.1	21.1
		Shfd07116 (SG)				17.7±1.0	21.2±1.1	21.2
10	Dimlington	Shfd09062 (SG)		Bateman *et al.* 2015	26.0	21.1±1.4	23.501.5	21.2
10	Dimlington	Shfd09063 (SG)		Bateman *et al.* 2015	26.0	20.5±1.2	22.8±1.4	21.2
10	Dimlington	Birm-108	53.673	Penny *et al.* 1969	-	21.8±0.3		21.9
10	Dimlington	I-3372	–0.109	Penny *et al.* 1969	-	21.9±0.4		22.0
11	Catwick/Gembling	Shfd13064	53.886	Bateman *et al.* 2018	6.3	22.8±1.8	23.5±1.8	23.1
11	Catwick/Gembling	Shfd13065	–0.623	Bateman *et al.* 2018	2.0	23.4±1.7	23.6±1.7	23.1
12	Barmston	Shfd10213	54.016	Bateman *et al.* 2015	4.7	11.3±0.7	12.0±0.7	-
12	Barmston	Shfd10214 (SA)	–0.215	Bateman *et al.* 2015	2.1	15.0±1.0	18.0±1.1	19.6
		Shfd10214 (SG)				15.0±1.2	17.6±1.4	-
12	Barmston	Shfd10215 (SA)		Bateman *et al.* 2015	7.5	18.6±1.4	20.4±1.6	20.2
		Shfd10215 (SG)				21.5±1.6	23.6±1.9	23.6
12	Barmston	Shfd10216		Bateman *et al.* 2015	4.7	12.2±0.6	12.9±0.7	-
12	Barmston	Shfd11007 (SA)		Bateman *et al.* 2015	1.7	16.4±1.0	20.4±1.6	20.4
		Shfd11007 (SG)				15.0±1.2	17.5±1.4	-
12a	Skipsea	Shfd13069	53.979	Bateman *et al.* 2018	7.0	16.5±1.0	18.3±1.2	18.3
12a	Skipsea	Shfd13070	–0.199	Bateman *et al.* 2018	6.0	15.9±1.3	17.6±1.4	17.6
13	Sewerby	Shfd05250	54.115	Bateman *et al.* 2015	3.75	17.9±1.9	18.6±1.9	19.6
			–0.168					

(*Continued*)

Table 1. (*Continued*)

14	Flamborough	Shfd10084	53.104	Bateman *et al.* 2015	18.4	20.0±1.4	20.9±1.4	21.9
14	Flamborough	Shfd10085	–0.119	Bateman *et al.* 2015	6.7	20.3±1.3	21.1±1.3	21.9
15	East Heslerton	Shfd13054	54.178	Evans *et al.* 2017	8.6	17.6±1.0	18.4±1.0	19.6
15	East Heslerton	Shfd13055	–0.597	Evans *et al.* 2017	5.5	17.3±1.0	18.4±1.0	19.6
15	East Heslerton	Shfd13056		Evans *et al.* 2017	2.0	15.8±0.90	16.9±1.0	-
15	East Heslerton	Shfd13057		Evans *et al.* 2017	1.0	10.1±0.7	11.0±0.8	-
16	Sandsend	Shfd13068	54.495	This paper	12.0		23.4±1.7	23.1
16	Sandsend	Shfd13067	–0.647	This paper	16.8		26.1±2.1	26.1
16	Sandsend	Shfd13066		This paper	21.8		23.8±1.8	23.1
18	Seaham	Shfd14064	54.811	Roberts *et al.* 2019	5.4	19.8±1.8		19.0
18	Seaham	Shfd14065	–1.313	Roberts *et al.* 2019	7.4	19.1±1.9		19.0
18	Seaham	Shfd14066		Roberts *et al.* 2019	8.0	19.9±2.3		19.0
17	Maiden Castle, Durham City	Shfd14067	54.668 –1.544	This paper	2.7		16.2±0.8	16.2
17	Maiden Castle, Durham City	Shfd14068		This paper	1.0		10.1±1.2	-
20	Tay/Forth Estuary (Shiells)	SRR-391	56.279 –3.164	Harkness & Wilson 1979	-	16.4±0.2		16.0
20	Tay/Forth Estuary (Gallowflat)	AA-37787	56.374 –3.276	Peacock 2002	Loc 3, Div B	16.7±0.1		16.7
20	Tay/Forth Estuary (Inchcoonans)	Beta-111508	56.397 –3.231	Peacock 2002	No.3, Div A	16.0±0.1		16.0
20	Tay/Forth Estuary (Gallowflat)	CAMS-77912	56.374 –3.276	Peacock 2002	Loc 3, Div B	16.8±0.1		16.9
20	Tay/Forth Estuary (Barry Clay Pit)	OxA-1704	–2.736 –2.736	Hedges *et al.* 1989	-	16.9±0.2		16.8
22	North Sea 178VC	Shfd15179	54.013 3.351	Roberts *et al.* 2018	0.15	25.8±2.4		27.2
23	North Sea 179VC	Shfd15180	54.012 3.271	Roberts *et al.* 2018	0.25	31.7±2.1		31.7
24	North Sea150VC	Shfd15175	55.405	Roberts *et al.* 2018	0.16	29.5±1.9		29.5
25	North Sea 151VC	Shfd15176	3.544	Roberts *et al.* 2018	0.13	26.2±2.1		27.3
26	North Sea 155VC	Shfd15178	55.386 1.319	Roberts *et al.* 2018	0.35	23.1±2.3		23.1
27	North Sea 137VC	UCIAMS-176372	55.293 0.542	Roberts *et al.* 2019	5.52	19.8±0.2		19.9
30	North Sea 132VC	SUERC-68010	55.491 –0.919	Roberts *et al.* 2019	1.44	19.6±0.2		19.4
32	North Sea 128VC	SUERC-68009	55.979 –2.040	Roberts *et al.* 2019	2.80	17.0±0.2		16.9
33	North Sea 118VC	SUERC-68001	56.495	Roberts *et al.* 2019	2.40	17.9±0.2		17.9
33	North Sea 118VC	SUERC-68007	–2.046	Roberts *et al.* 2019	2.40	16.6±0.2		16.6
34	North Sea 119VC	SUERC-68883	56.494 –2.029	This paper	2.00		18.0±0.2	19.7
35	North Sea 120VC	SUERC-68008	56.492 –1.938	This paper	0.43		17.6±0.2	18.0
39a	North Sea 117VC	SUERC-72877	56.835 3.814	This paper	0.20		13.4±0.1	Not in Bayesian
40	North Sea 111VC	SUERC-68000	56.904 0.088	This paper	3.44		17.6±0.2	18.8
42	North Sea 106VC	SUERC-67993	58.167	This paper	4.66		14.5±0.3	14.5
42	North Sea 106VC	SUERC-67997	–1.192	This paper	5.36		15.1±0.2	15.4
43	North Sea 105VC	SUERC-67992	58.448 –1.460	This paper	1.55		15.1±0.2	15.1
44	North Sea 104VC	SUERC-67990	58.522	This paper	2.55		15.1±0.2	15.1
44	North Sea 104VC	SUERC-67991	–1.590	This paper	3.23		15.2±0.1	15.2
45	Fladen/Witch Ground -BH 04/01	Ua- 32760	58.403 0.577	Graham *et al.* (2010)	17.70	19.6±0.4		18.3
45	Fladen/Witch Ground -BH 04/01	Ua-32761		Graham *et al.* (2010)	17.70	16.8±0.5		16.8
46	Fladen/Witch Ground -BGS 77/2	Ua-831	58.492	Sejrup *et al.* (1994)	8.98	17.5±0.3		17.7
46	Fladen/Witch Ground -BGS 77/2	TUa-149	0.505	Sejrup *et al.* (1994)	9.00	23.1±0.3		23.1

(*Continued*)

Table 1. (*Continued*)

46	Fladen/Witch Ground -BGS 77/2	TUa-276		Sejrup *et al.* (1994)	12.70	18.0±0.2		17.9
46	Fladen/Witch Ground -BGS 77/2	TUa-458		Sejrup *et al.* (1994)	16.40	21.3±0.2		21.3
46	Fladen/Witch Ground -BGS 77/2	TUa-459		Sejrup *et al.* (1994)	17.55	17.7±0.2		17.7
46	Fladen/Witch Ground -BGS 77/2	TUa-141		Sejrup *et al.* (1994)	18.00	19.5±0.3		19.5
46a	Fladen/Witch Ground -BGS 58+00/ 111	OxA-1317	58.567 0.402	Long *et al.* 1986 Hedges *et al.* 1988	2.4	15.9±0.3		15.9
46a	Fladen/Witch Ground -BGS 58+00/ 111	OxA-1318		Long *et al.* 1986 Hedges *et al.* 1988	2.8	25.2±0.4		25.2
46a	Fladen/Witch Ground -BGS 58+00/ 111	OxA-1319		Long *et al.* 1986 Hedges *et al.* 1988	4.0	16.7±0.3		16.7
46a	Fladen/Witch Ground -BGS 58+00/ 111	OxA-1320		Long *et al.* 1986 Hedges *et al.* 1988	5.2	15.6±0.2		15.7
46a	Fladen/Witch Ground -BGS 58+00/ 111	OxA-1321		Long *et al.* 1986 Hedges *et al.* 1988	5.2	15.9±0.2		15.9
47	Fladen GS140-05-15GC	Beta-226955	58.926 0.803	Sejrup *et al.* (2015)	1.10	14.2±0.1		14.2
47	Fladen GS140-05-15GC	Beta-226956		Sejrup *et al.* (2015)	1.36	14.6±0.2		14.6
48	Fladen GS140-05-14GC	Beta-225422	58.578 0.726	Sejrup *et al.* (2015)	1.68	16.8± 0.1		16.8
48	Fladen GS140-05-14GC	Beta-225420		Sejrup *et al.* (2015)	1.80	15.6±0.1		15.6
48	Fladen GS140-05-14GC	Beta-225418		Sejrup *et al.* (2015)	2.01	16.0±0.1		16.0
48	Fladen GS140-05-14GC	Beta-216741		Sejrup *et al.* (2015)	2.46	16.2±0.1		16.1
49	North Sea 097VC	SUERC-67989	59.083 2.913	This paper	4.07		14.5±0.3	Not in Bayesian
50	North Sea B2001	TUa-274	58.390 1.728	Sejrup *et al.* (1994)		37.0±0.7		Not in Bayesian
51	St Fergus, Aberdeen coast	Beta-101953	57.471 –1.816	Peacock 1999		17.1±0.1		Not in Bayesian
51	St Fergus, Aberdeen coast	Lu-3028	57.456 –1.852	Hall and Jarvis 1989		17.6±0.3		Not in Bayesian
52	Moray Firth – core 73/35	SRR-322	57.767 –2.317	Harkness & Wilson 1979	-	19.3±0.9		18.8
52	Moray Firth – core 73/35	SRR-323		Harkness & Wilson 1979	-	19.7±0.7		19.3
52	Moray Firth – core 73/35	SRR-324		Harkness & Wilson 1979	-	17.0±0.7		16.5
52	Moray Firth – core 73/35	SRR-325		Harkness & Wilson 1979	-	20.2±0.8		19.7
53	Moray Firth – core 74/18	SRR - 627	58.167 –2.950	Harkness & Wilson 1979	-	13.9±0.1		14.0

because it constituted a substantial topographic feature. This topographic confinement, in combination with the decoupling of the BIIS and FIS further north, resulted in the NSL advancing down the western margin only of the North Sea Basin to the north Norfolk coast where it emplaced the late stage tills. In this area of the southern North Sea, Dove *et al.* (2017) record a series of broad sedimentary wedges and associated moraines (sub-marginal till wedges) that document inset marginal locations of the NSL, classified as stillstand margins 1–4, and formerly regarded as the Bolders Bank Formation (Jansen *et al.* 1979; Balson & Jeffrey 1991; Cameron *et al.* 1992). This represents a record of northerly retreat punctuated by readvances, indicative of an oscillating NSL margin. Margin 1 aligns with the onshore limit of the NSL, demarcated by the Garret Hill Moraine in north Norfolk (Pawley *et al.* 2006; Moorlock *et al.* 2008; Roberts *et al.* 2018; Evans *et al.* 2019) and Margin 2 appears to continue on land as the Hogsthorpe-Killingholme Moraine of the Lincolnshire Marsh (Straw 1979; Evans *et al.* 2019). Margins 3 and 4 are inset within Margin 2

and appear to arc northwards to align with the moraines on Holderness.

The impact of the NSL along the Durham and Northumberland coasts and its interaction with inland ice sources is evidenced by various features. Onshore multiple till sequences separated by glacifluvial sediments record ice moving eastward into the North Sea from inland sources (e.g. the Tyne Ice stream) before being displaced by the southward-flowing NSL (Eyles *et al.*, 1982; Davies *et al.*, 2009; Livingstone *et al.*, 2015; Fig. 1). Just offshore, partially excavated, streamlined, mixed-bed terrain shows the NSL dominating the latter stages of the last glacial cycle, with ice streams such as the Tyne and Tweed either in recession or being deflected south (Everest *et al.* 2005; Livingstone *et al.*, 2010a; Roberts *et al.*, 2019). There is also a clear set of ice-contact, marginal landforms that document the initial decoupling of the NSL and inland ice (Staines 2009; Teasdale, 2013). Livingstone *et al.* (2015; cf. Davies *et al.* 2019) date this uncoupling at 18.5–18.3 ka based on surface exposure ages in the western Tyne Gap. The uncoupling also led to the development of several ice-dammed lakes along the Durham coast (e.g. glacial lakes Tees, Edder Acres and Wear).

Further north, the onshore regions adjacent to the Firth of Forth, Firth of Tay, Strathmore and the Moray Firth were dominated glaciologically by large ice streams during the last glacial cycle (Jamieson 1906; Linton 1959, 1962; Sissons 1963; Golledge and Stoker 2006; Hughes *et al.* 2010, 2014; Fig. 3). Many early workers identified an initial east/southeast ice flow across the Strathmore region sourced from the Grampian mountains, with a later switch in ice flow direction to the northeast along the Strathmore valley suggesting a more southerly source of ice or perhaps topographic influence on flow direction as the ice thinned (Bremner 1943; Charlesworth 1956; Synge 1956; Armstrong *et al.* 1985; Merritt *et al.* 2003). Golledge and Stoker (2006) assessed the subglacial bedform signature in the Strathmore area and concluded that the landforms represented a former ice stream (Strathmore Ice Stream), which was probably coalescent with a large ice stream in the Firth of Forth (Firth of Forth Ice Stream). Indeed, the geomorphology of the onshore corridors adjacent to the Firth of Forth (i.e. the Edinburgh and Kircaldy coasts) also exhibit streamlined bedforms in the form of streamlined bedrock bumps, crag and tails and drumlins, and these demonstrate ice broadly flowing east and bending southeast towards the Northumberland coast (Sissons 1971; Evans and Hansom 1996; Fig. 3). Additional evidence for ice streaming offshore was linked to deeply incised channels/tunnel valleys and seafloor glacial lineations running southwest–northeast from Montrose. Golledge and Stoker (2006) also suggested that the Wee Bankie Moraine complex may represent a shear margin moraine, although recent work suggests it is more likely to represent the onshore retreat of Firth of Forth ice during deglaciation (Roberts *et al.* 2019).

Following the work of Golledge and Stoker (2006), Hughes *et al.* (2010, 2014) constructed a comprehensive map of bedform types (e.g. drumlins, mega-scale glacial lineations, crag and tails) in this region and reconstructed complex changes in ice dynamics based on flowset mapping. This showed both the Strathmore Ice Stream and the Firth of Forth Ice Stream responding to changes in ice thickness, ice divide migration and topographic influences during the last glacial cycle, but clearly demonstrated that the Strathmore Ice Stream and Firth of Forth Ice Stream were key sources of ice heading northeast, east and southeast into the central North Sea. The behaviour of the Firth of Forth Ice Stream offshore is poorly understood during the build-up to the LGM, but advances have been made in understanding the role of Firth of Forth ice in controlling the NSL; indeed it is highly likely that the NSL was an offshore extension of the Firth of Forth Ice Stream. This is supported by the imprint of the Bolders Bank Formation, subglacial bedforms, tunnel valleys and moraines on the seafloor as well as onshore stacked tills and ice margin geomorphology down the east coast of Britain and in the vicinity of Dogger Bank (Dove *et al.* 2017; Roberts *et al.* 2018; Emery *et al.* 2019), all of which supports previous work and demonstrates that Scottish ice was flowing from the Firth of Forth into the southern North Sea (Boulton *et al.* 1985; Davies *et al.* 2009; Evans and Thompson 2010; Busfield *et al.* 2015; Dove *et al.* 2017).

Methods

Sampling and stratigraphy

The sediments and stratigraphies of all locations (Figs 4–6) were analysed either from natural coastal cliff and quarried exposures in onshore settings or from seismic and bathymetric data and sediment cores in offshore settings. Sedimentological analysis followed a lithofacies approach with sediments classified on the basis of colour, particle size, clast lithologies and sedimentary structures (cf. Evans and Benn 2004), and observations and data were compiled on vertical profile logs and in some cases scaled photo mosaics. Detailed information included primary sedimentary structures, bed contacts, sediment body geometry, sorting and texture and secondary structures, as well as data on clast form and lithology and palaeocurrents where appropriate. Further details in terms of site-specific analyses are available in Evans *et al.* (2017, 2018, 2019), Bateman *et al.* (2018) and Roberts *et al.* (2018, 2019). The sediment–landform associations identified both onshore and offshore were placed in a glacial land-systems context established for the region by earlier research (Eyles *et al.* 1982; Evans *et al.* 1995; Bradwell *et al.* 2008; Davies *et al.* 2009; Boston *et al.* 2010; Evans & Thomson 2010; Roberts *et al.* 2013; Sejrup *et al.* 2015; Dove *et al.* 2017) as well as a regional lithostratigraphic context (cf. Straw 1979; Cameron *et al.* 1987, 1992; Balson & Jeffrey 1991; Eyles *et al.* 1994; Carr *et al.* 2006; Pawley *et al.* 2006; Catt 2007; Bateman *et al.* 2008; Sejrup *et al.* 2009; Davies *et al.* 2011).

Geochronology

Geochronology was secured using a combination of OSL and radiocarbon dating (Table 1). All BRITICE-CHRONO OSL ages were based on extracted and cleaned quartz grains of 180–250 μm diameter, which were prepared under controlled light conditions as per Bateman and Catt (1996) and measured using an automated Risø luminescence reader. Equivalent dose (D_e) estimates were measured unless stated otherwise using the SAR protocol of Murray and Wintle (2000, 2003) on small multigrain aliquots containing ~20 quartz grains each. Up to 80 small multigrain aliquots were measured for each sample in order to have a representative D_e dose distribution.

For the new Maiden Castle and Sandsend OSL ages presented as part of this paper (Table 2), measurements with the SAR protocol included an additional second recycling point, using IR stimulation prior to OSL measurement, in order to detect and exclude aliquots with any feldspar contamination. Within the SAR protocol, a preheat of 220 °C and 180 °C for 10 s was used to measure the samples from Maiden Castle and Sandsend, respectively. These temperatures were selected as the most appropriate, based on a dose recovery preheat temperature test performed on samples Shfd14068 (Maiden Castle) and Shfd13067 (Sandsend), which recovered to a given ratio consistent with unity. D_e values were accepted only if the relative uncertainty on the natural test–dose response was less than 20%, the recycling and the IR depletion ratio, including uncertainties, were within 20% of unity, and the recuperation

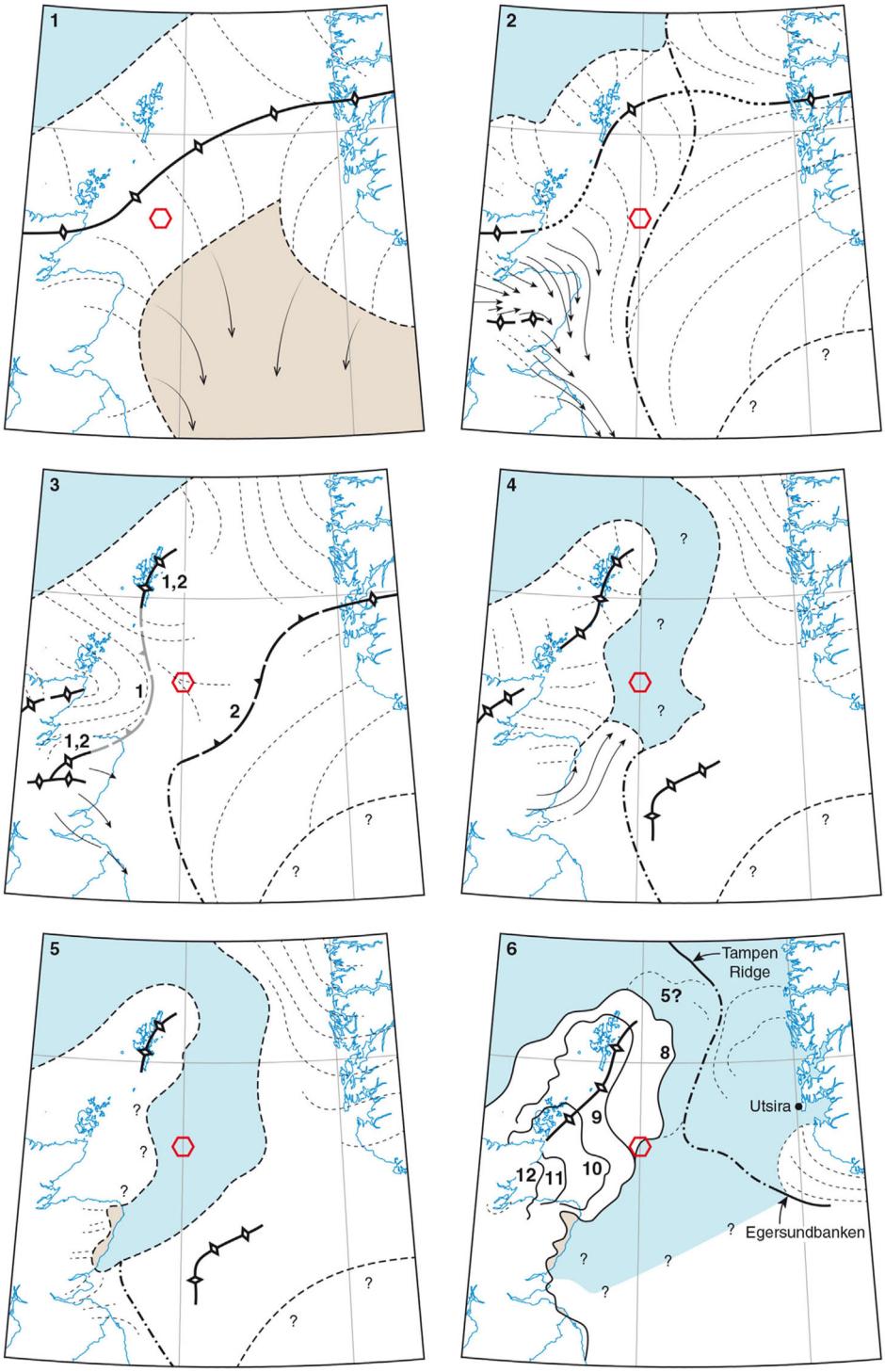

Figure 3. Palaeoglaciology of eastern Scottish coast and adjacent offshore area (from Merritt *et al.* 2017): 1) reconstruction of Boulton *et al.* (1985); 2) local LGM (stage 2); 3) considerable thinning and glacial reorganisation during stages 3 and 4; 4) re-expansion following initial marine ingression into Witch Ground Basin during stage 5, followed by probable coalescence of BIIS and the Norwegian Channel Ice Stream at the Tampen Ridge (see panel 6 for limits); 5) marine ingression into Lunan Bay during stage 7; 6) stages 8–12 and other limits (after Clark *et al.* 2012). Red hexagons show the centre of the Witch Ground Basin. Ice divides shown by thick black lines with open diamonds; single filled ticks denote ice divide receding by ice stream headward scavenging.

was less than 5%. These criteria lead to D_e distributions containing ~60 independent values per sample. These distributions are highly scattered, with a number of the independent values in the high dose region, indicating that these samples were affected by incomplete bleaching (i.e. insufficient exposure to daylight during transport to reset the luminescence signal). This is the case for all but sample Shfd14068, which is characterised by a normal distribution. In this latter case, the final D_e was estimated through the Central Age Model (CAM, Galbraith *et al.* 1999). For the samples affected by incomplete bleaching, the internal–external uncertainty model (IEU, Thomsen *et al.*, 2007) has been applied in order to base the final D_e estimate on the dose population most likely to belong to well-bleached grains. To apply the IEU approach, an a-value of 0.2 and b-value of 1.5, as determined from a dose recovery test, have been used.

Total dose rates were calculated based on gamma contributions, measured using an EG&G MicroNomad gamma field

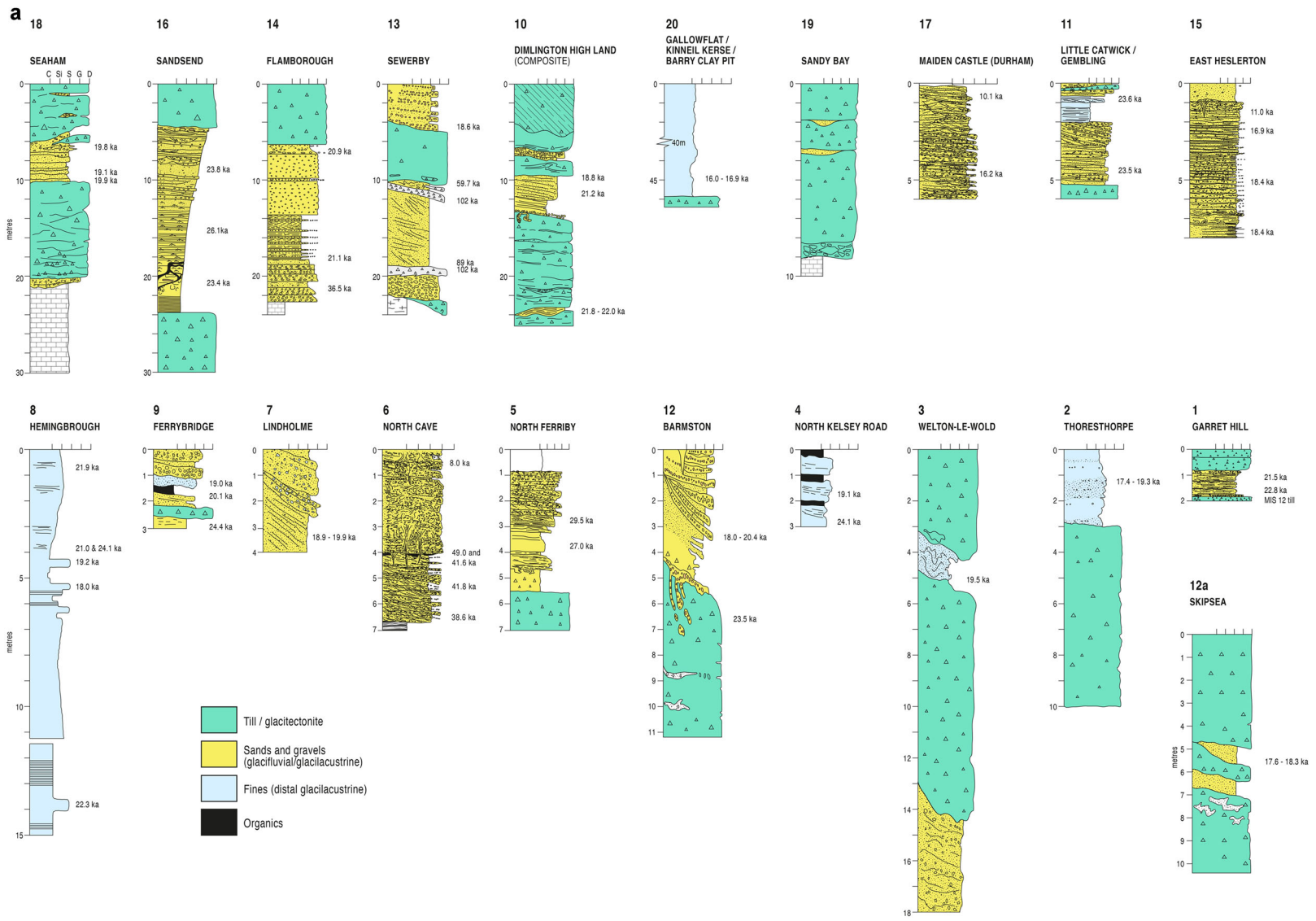

Figure 4. Vertical profile logs (from various sources) compiled at the locations critical to the palaeoglaciological reconstructions in this paper: a) onshore sites; b) and c) offshore core sites (green units are till/glacitectonite). For site locations see Fig. 1. Dates and date ranges are compiled from Table 1, with additional dates included for the onshore sites of Flamborough and Sewerby for broader context, as reported by Bateman *et al.* (2015).

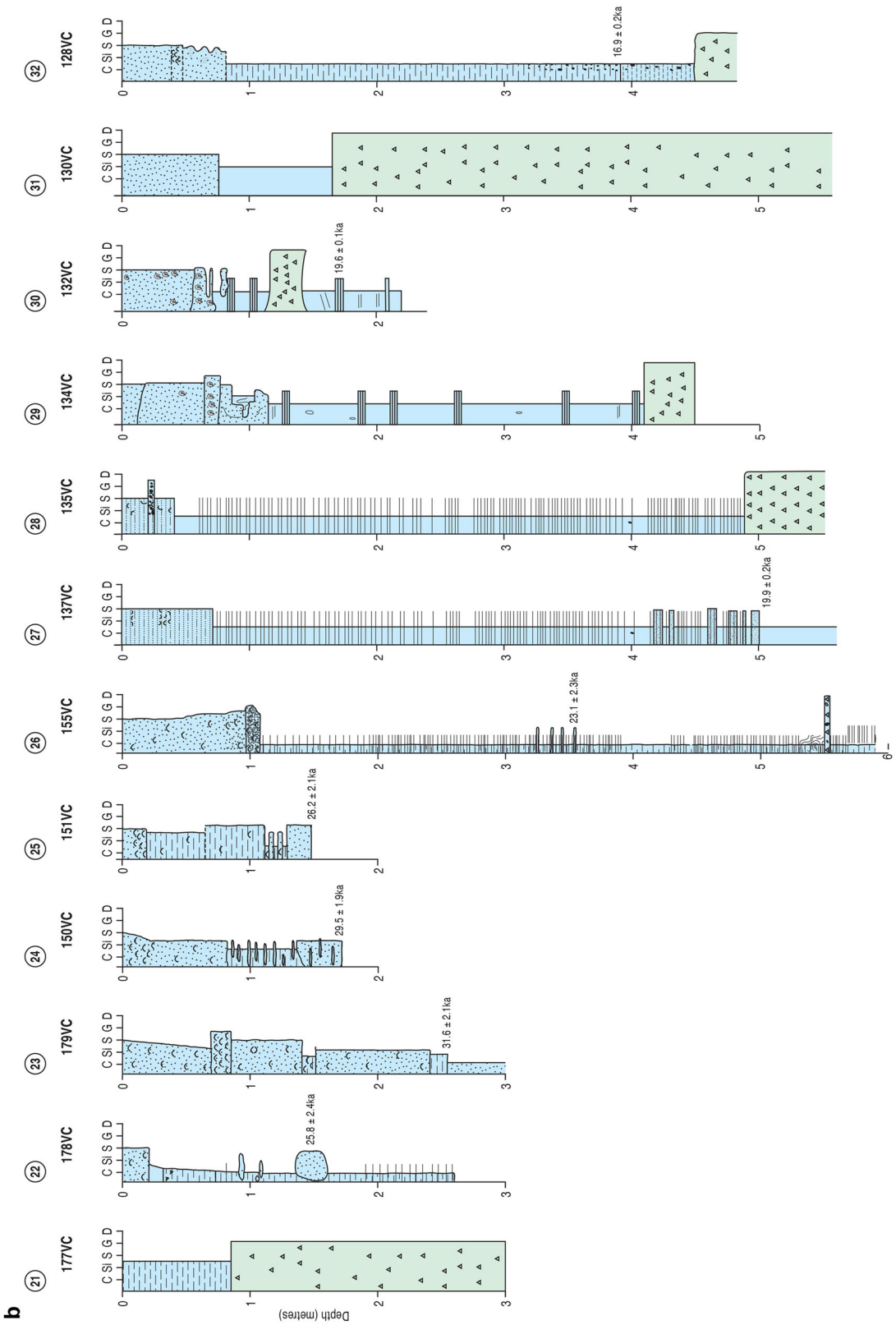

Figure 4. Continued

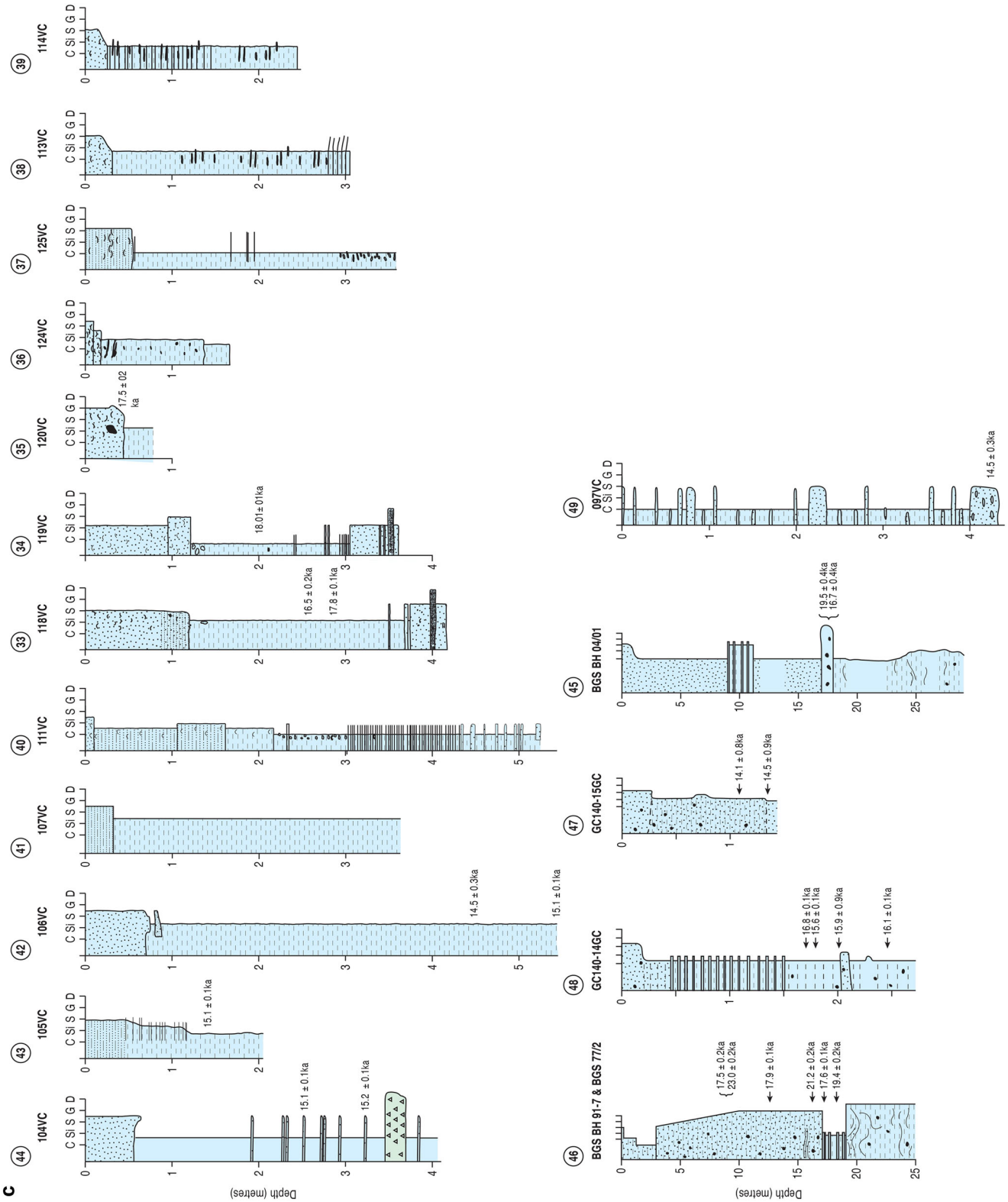

Figure 4. Continued

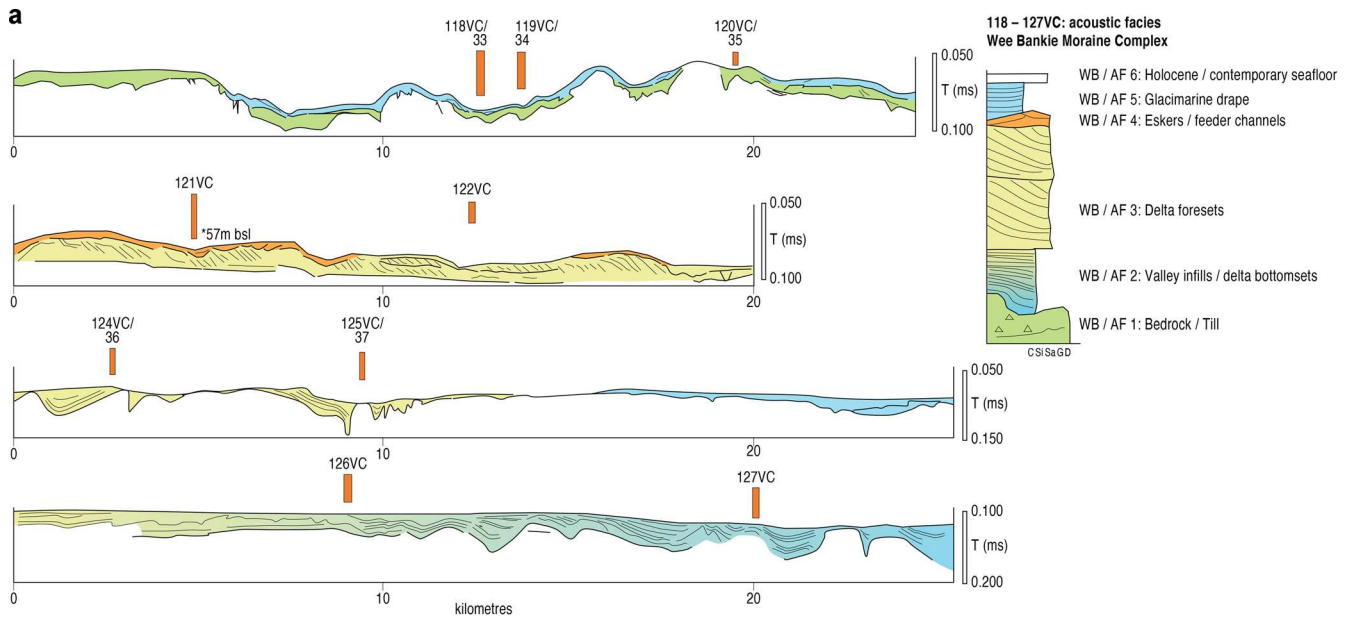

Figure 5. Offshore stratigraphic architecture and selected vertical profile logs based on acoustic and core data. Transect lines are located by blue dashed lines in Fig. 1: a) Acoustic stratigraphy to the east of Great Fisher Bank associated with core locations 112VC–114VC where the seafloor is marked by a distinctive scarp/step formed by glacitectonised lake sediments and a subaqueous ice-contact fan; hence, this may mark a former ice margin; b) Between core locations 111VC and 104VC the seafloor displays a series of subglacial tills draped by glacimarine sediments with inset moraines marking grounding line retreat towards Bosies Bank between 17.5 and 15.0 cal. ka BP; c) Between core locations 127VC and 118VC an east-to-west acoustic profile shows deglacial glacimarine sediments transitioning to shallower ice-marginal delta complexes as the ice sheet retreated west into the Firth of Forth. The Wee Bankie Moraine complex is thus composed of both drowned deltas to the east as well as push moraines to the west. Cores 118VC–120VC provide deglacial ages for the Wee Bankie Moraine complex of 18.0 to 16.5 cal. ka BP.

spectrometer and beta contributions based on elemental analysis using inductively coupled plasma mass spectroscopy (ICP-MS) at SGS laboratories Ontario, Canada. The dose rates from radionuclide concentrations were derived from the conversion factors of Guérin *et al.* (2011). A linear accumulation of deposits has been assumed in order to calculate the contribution of cosmic radiation according to a varying burial depth (based on Prescott and Hutton, 1994). Finally, the total dose rates as shown in Table 2 were calculated according to attenuation caused by moisture and grain size. D_e estimates and further details associated with the OSL ages from Maiden Castle and Sandsend are presented in the Supplementary Information.

Accurately determining the average palaeomoisture since burial has proved challenging, as groundwater levels through time have changed markedly for most samples due to ice and permafrost degradation, sea-level changes and formation and emptying of proglacial lakes. Hence, present-day moisture levels are thought unlikely to reflect average palaeomoisture levels. Consequently, sites and samples were evaluated based on whether evidence suggested they had been saturated for most or a substantial part of their burial history in which case palaeomoisture values of 27% and 23% were applied, respectively. If samples were only partially saturated or minimally saturated during burial, values of 15% or 10% were applied, respectively. To cover fluctuations, wide uncertainties of ± 5% on all water content were adopted to include variations related to porosity and compaction. For consistency, these four possible palaeomoisture values were applied to all BRITICE-CHRONO OSL samples. For example, samples at Garret Hill, Norfolk (Shfd15033 and Shfd15034), at 9 m OD and in glacifluvial sands underlain by free-draining chalk, were assigned palaeomoisture values of 10%. In contrast, samples from Thoresthorpe, Lincolnshire (Shfd15154 and Shfd15155), at 12 m OD and in ice-marginal lacustrine sediments, which even today have a perched near-surface groundwater table, were assigned saturated moisture values of 27%. Of particular note were samples collected from eroded shore-line cliff sections in sand units between diamictons. Initially many of these were assessed as minimally saturated (10%) or partially saturated (20%). However, re-evaluation has established that prior to cliff erosion such intercalated sands within diamictons would have acted either as aquifers or have been below the groundwater table. High regional water tables are evidenced by the ongoing presence of Hornsea Mere, found in a topographic low on the diamicton. As a result, palaeomoisture values have been adjusted and in many cases increased. Moreover, previously published ages have been recalculated to better reflect what is believed to be true palaeomoisture levels (see Table S1) and to be consistent with the regional OSL age dataset (Table 1). For full details of OSL measurement protocols, approaches to D_e data analysis, age estimation and associated details, readers should consult the source publication where they were originally published as shown in Table 1.

As a result of the above revisions, the model proposed by Bateman *et al.* (2018) of initial NSL incursion into the Humber Gap at 21.6 ka (Stage 1) is revised to ~29 ka and its retreat offshore from the present terrestrial record, originally dated at ~18 ka (Stage 2), is revised to ~20 ka. Impoundment of Lake Humber occurred at least by 24 ka until 18 ka with Lake Pickering being at least partially coeval with this. Full withdrawal of the NSL from the Yorkshire region occurred prior to ~19 ka (previously ~15 ka; Stage 4 of Bateman *et al.* 2018).

Radiocarbon dates were obtained from a mixture of paired bivalves, mixed benthic foraminifera samples and shell fragments. The shell fragments and bivalves were cleaned with deionised water and dried at 40 °C in an oven.

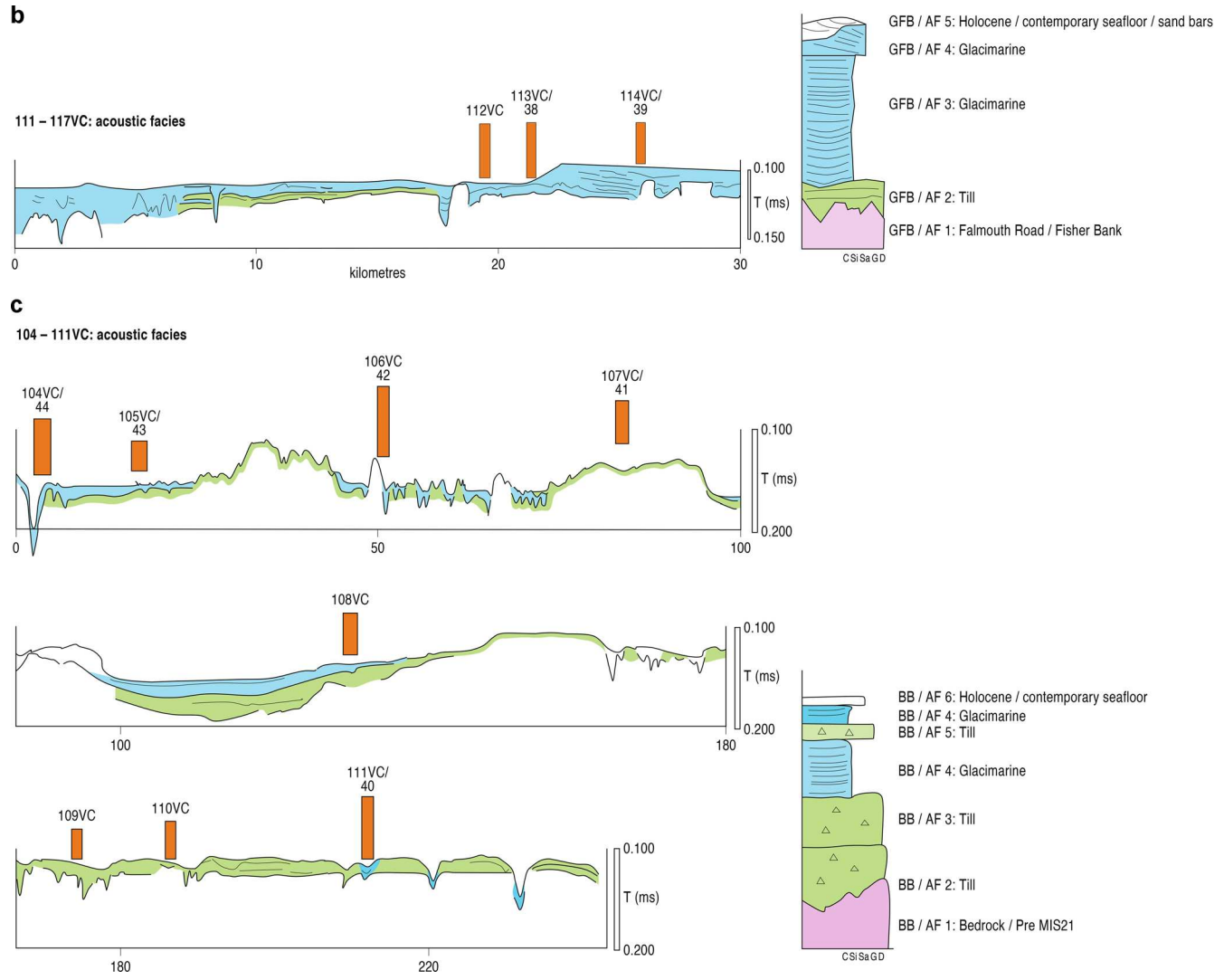

Figure 5. Continued

Foraminifera were sieved through 500 µm, 180 µm and 63 µm sieves and dried at 40 °C and then whole, unabraded specimens were picked dry from the 500–180 µm fraction. Conventional ages were calibrated using the Marine13 curve with an inbuilt marine reservoir correction of 400 years and a ΔR of 0 years (Calib v7.0 software; Reimer *et al.*, 2013) and reported here as the calibrated 1σ median result. Importantly, it is likely that samples would be subject to large and variable local ΔR during the LGM and the Lateglacial period.

A number of previously reported legacy dates were available to augment the chronology established by BRITICE-CHRONO Transect 2 (Table 1). These are integrated into the various sections below where pertinent and used alongside the BRITICE-CHRONO dates in order to compile a regional reconstruction of ice-sheet dynamics in the North Sea region.

The glacial geomorphology and stratigraphy of the North Sea and eastern England have been used to define seven distinctive former ice-sheet limits (Fig. 6). This deglaciation sequence provides a prior model (i.e. the hypothetical 'relative order' of events) for Bayesian statistical analysis of the independent dating of these sediments (e.g. Chiverrell *et al.* 2013). The complexity of the region necessitates dividing the glacial retreat sequence and geochronology into two models describing first a southern sequence, from Norfolk to the Firth of Forth, and second a northern sequence, from the Fladen Moraines to the Moray Firth. Both Bayesian prior models were developed independently and included all geochronology (Bronk Ramsey 2008, 2009a, b; Bronk Ramsey and Lee 2013). These two prior models were determined independently of the geochronological data and here constitute the net ice-marginal pattern of retreat integrating small-scale readvances and stillstands. The Bayesian modelling used uniform phase sequence models that were punctuated by boundaries and were coded using OxCal 4.3 (Bronk Ramsey and Lee 2013). The statistical analyses use Markov Chain Monte Carlo sampling to build up a distribution of possible solutions, generating probabilities called posterior density estimates, and these are the product of both the prior or relative-order model and the likelihood (measured age) probabilities for each sample. Each retreat zone was coded in the models as Phases, containing grouped dating information for sites that shared a common relationship with the preceding and subsequent zones. Phases were separated by the Boundary command and generated a modelled age probability distribution output for the major ice limits.

Both sequence models were run in outlier mode in order to assess outliers in time using a Student's *t*-distribution ($P < 0.05$) to describe the outlier distribution and a scaling of 10^0–10^4 years (Bronk Ramsey, 2009b). The models were run initially to

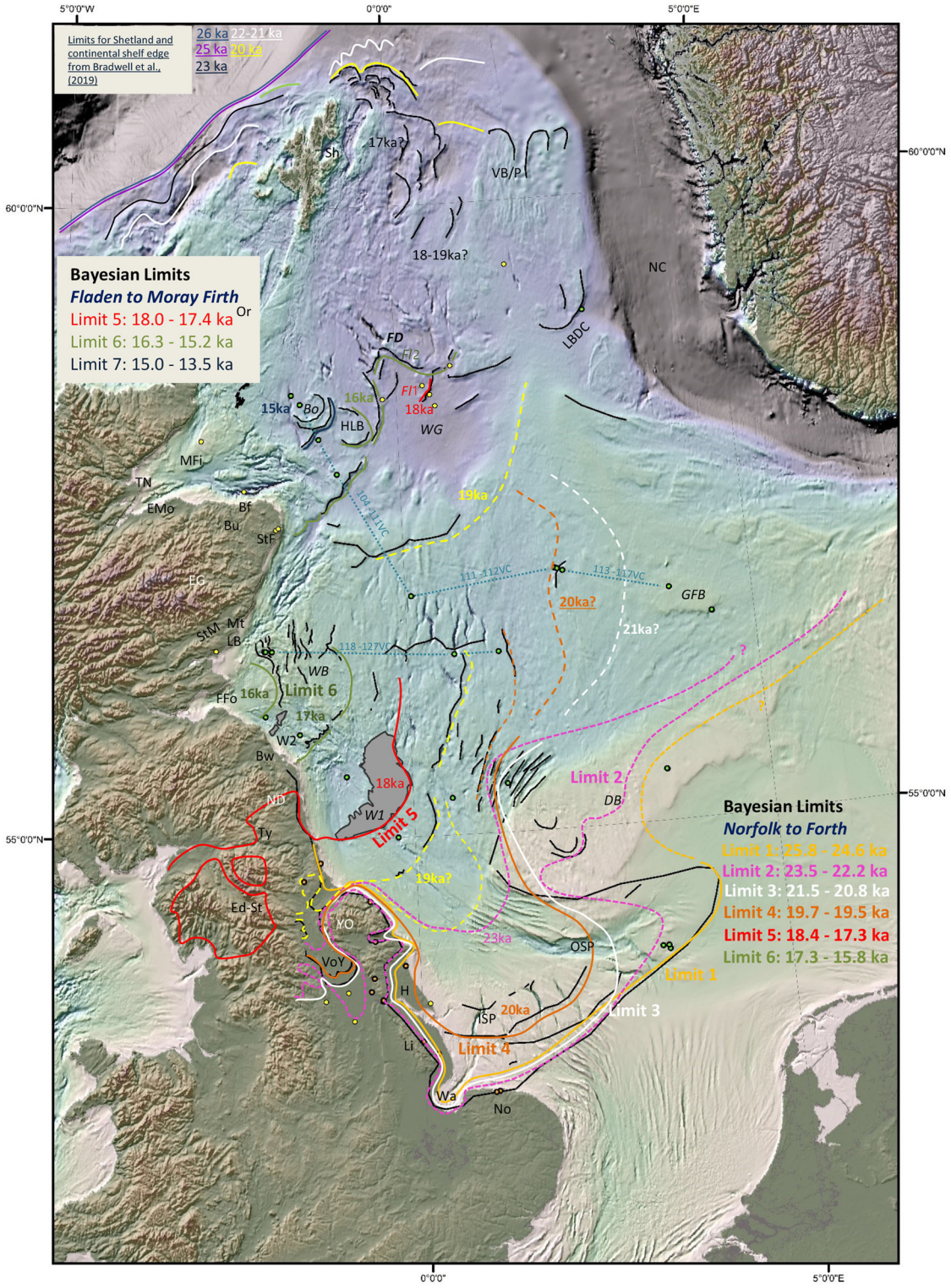

Figure 6. Bathymetry and topography of the North Sea region, showing ice-sheet advance and retreat patterns for the period 25.8–13.5 ka with ice limits determined by Bayesian analysis. Black lines demarcate major moraines and/or till limits. The main Bayesian Limits 1–7 are anchored by major moraine or grounding zone wedge complexes (grey coloured areas W1 and W2 are grounding zone wedges). Morphostratigraphic controls clearly show that Limits 2, 3, 4 and 5 are associated with regional readvances between ~24 and 18 ka. The Limit 4 to Limit 5 marks a transition between glaciterrestrial and glacimarine conditions in the southern and central North Sea. In the northern North Sea deglacial isochrones for Shetland and the continental shelf edge are sourced from Bradwell *et al.* (2019), also derived using Bayesian modelling.

give obvious outliers a probability scaling of $P = 1$ and using a high generic probability of all other ages being outliers ($P < 0.2$). This process identified a series of statistical outlier ages. These outliers were not disregarded automatically, but explanations were sought in terms of stratigraphical context or geochronological measurement (e.g. poor resetting of the OSL signal). Subsequent iterative runs of the Bayesian modelling gradually reduced the individual outlier probabilities, with $P < 0.1$, $P > 0.2$, $P > 0.5$ and $P > 0.95$ reflecting an increasing order of outlier severity. Ultimately, the Bayesian analyses produced conformable age models, with overall model agreement indices of 179% (Norfolk to Firth of Forth) and 132% (Fladen to Moray), both exceeding the >60% threshold advocated by Bronk Ramsey (2009a). Agreement indices were

Table 2. Summary of previously unreported OSL estimated ages and associated information for three samples from Sandsend (North Yorkshire) and two samples from Maiden Castle (County Durham). Depth, water content and the calculated contribution to the total dose rates are summarised. The table also includes information on the total number of aliquots measured with OSL which passed the acceptance criteria and the overdispersion (OD) of the resulting dose distribution. Reported equivalent doses (D_e) for all samples except Shfd14068 were derived using the IEU approach. The exception (Shfd14068) involved the application of the CAM instead of IEU, as the D_e population was normally distributed and therefore not affected by incomplete bleaching.

	Lab code	Depth (m)	w (%)	β dose rate (Gy/ka)	γ dose rate (Gy/ka)	Cosmic dose rate (Gy/ka)	Total dose rate (Gy/ka)	Aliquots accepted (measured)	OD (%)	Equivalent dose (Gy)	Age (ka)
Sandsend	Shfd13066	21.8	23	0.66 ± 0.06	0.57 ± 0.04	0.03 ± 0.001	1.27 ± 0.07	29 (29)	55	30.1 ± 1.7	23.8 ± 1.8
	Shfd13067	16.8	23	0.67 ± 0.05	0.56 ± 0.04	0.04 ± 0.002	1.28 ± 0.007	31 (94)	54	33.5 ± 2.1	26.1 ± 2.1
	Shfd13068	12.0	23	0.47 ± 0.04	0.56 ± 0.04	0.05 ± 0.003	1.10 ± 0.05	28 (73)	63	25.9 ± 1.4	23.4 ± 1.7
Maiden Castle	Shfd14067	2.7	10	0.52 ± 0.04	0.46 ± 0.03	0.13 ± 0.006	1.16 ± 0.05	39 (98)	91	18.7 ± 2.7	16.2 ± 0.8
	Shfd14068	1.0	10	0.57 ± 0.04	0.60 ± 0.03	0.18 ± 0.009	1.32 ± 0.05	34 (98)	46	13.3 ± 1.2	10.1 ± 1.0

calculated for individual age determinations, identifying samples that did not agree with the model; individual agreement indices should usually be over 60%. The Bayesian modelling has generated modelled ages for boundaries between seven ice limits (L1–7) related to ice advance and retreat from Norfolk into the Firth of Forth and also further north into the Moray Firth (Figs 6 and 7).

Glacial geomorphology and chronological controls

Southern sector

Defining and dating pre-LGM ice extent

Evidence of terrestrial environmental conditions immediately prior to MIS 2 is contained within fluvioperiglacial fan deposits on the east side of the Glacial Lake Humber basin (North Cave, Fig. 4a; Evans *et al.* 2018). The earliest of these deposits date to MIS 3, based on OSL ages of 38.6–41.8 ka (Shfd 13058 and Shfd 13059) and radiocarbon ages of 41.6–49.0 ka (SUERC-77600 and SUERC79023; Table 1). Importantly, the fan deposits were prograded to near or below 0 m OD and hence indicate that Glacial Lake Humber did not occupy the area until later, during MIS 2, at which time the lake waters only reached 8 m OD. The basal pair of ages from North Cave constrains ice-free conditions in the region prior to ice advance and predates the MIS 2 ice advance by an unknown period. The coding in the Bayesian analysis uses an "After" constraint for these ages, which focuses the modelling towards the subsequent chronology that directly dates proglacial settings associated with the Limit 1 NSL advance.

The southernmost extent of MIS 2 glaciation in eastern England has traditionally been associated with the outer edge of the Skipsea Till in Holderness (Catt & Penny 1966; Catt 2007), the Upper Marsh Till in Lincolnshire (Straw 1957, 1958, 1961, 1979) and the Hunstanton Till (Holkham Member, Lewis 1999) in north Norfolk (Suggate & West 1959; Straw 1960, 1979), all of which are lithostratigraphically equivalent (Boston *et al.* 2010). Moraines associated with this limit are not particularly widespread but are developed instead immediately inboard of the till limits; for example, the Holderness hummocky drift, extending from Gembling in the north to Catwick and then to Mill Hill in the south, (Eyles *et al.* 1994; Evans & Thomson 2010) and the Hogsthorpe–Killingholme Moraine of the Lincolnshire Marsh (Straw 1979), both of which have been explained as readvance positions. Additionally, the Horkstow Moraine in the Humber Estuary and the Stickney Moraine and Hunstanton Till (Holkham Member) in The Wash have been equated with an early Devensian (MIS 3/4) glaciation by Straw (1979) and related to his Lower Marsh Till. However, Evans *et al.* (2019) have pointed out that the Lower and Upper Marsh Tills have never been observed together in a single exposure and early depictions of the last glaciation limit are drawn at the more extensive, Lower Marsh Till position (e.g. Clayton 1957). Therefore, Evans *et al.* (2019) propose that the Stickney and Horkstow moraines represent the MIS 2 limit, in agreement with the age determination of Bateman *et al.* (2018). On the north Norfolk coast, the Garret Hill Moraine (Pawley *et al.* 2006) is associated with the Hunstanton Till limit and hence regarded as MIS 2 in age (Evans *et al.* 2019; Roberts *et al.*, 2019).

Limit 1 – NSL advance (pre-31–25 ka)

The nature of the glacial landforms of the southern North Sea have recently been reported by Dove *et al.* (2017) and dated by Roberts *et al.* (2018). The most extensive MIS 2 limit (Limit

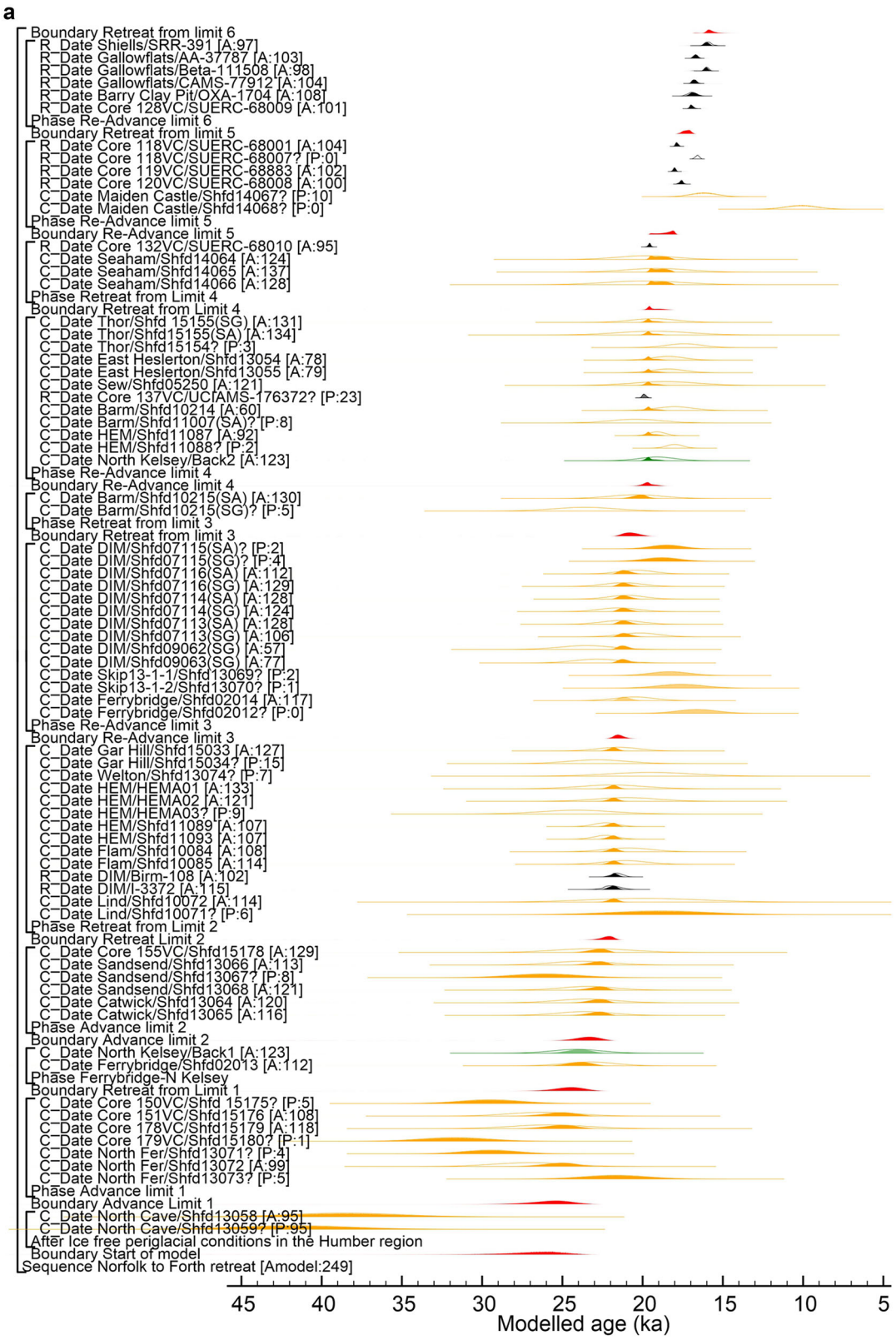

Figure 7. Bayesian models for the dating of ice retreat from (a) Norfolk to the Firth of Forth and (b) the Fladen 1 Moraine to the Moray Firth. The model structure shown uses OxCal brackets (left) and keywords that define the relative order of events (Bronk Ramsey, 2009a, b). Each original distribution (open symbol) represents the relative probability of each age estimate with posterior density estimate (solid symbol) generated by the modelling. Shown are ^{14}C ages (black/grey), luminescence ages (orange) and boundary ages produced in the modelling (Red). Outliers are denoted by their probability (e.g. [P:5]) of being an outlier indicated by low values <5 (95% confidence). Overall model agreement indices (e.g. [Amodel:100]) are shown alongside the agreement indices for individual ages (e.g. [A:66]), identifying their fit to the model (>60% is the widely used agreement threshold for 'good' fit (Bronk Ramsey, 2009a, b)).

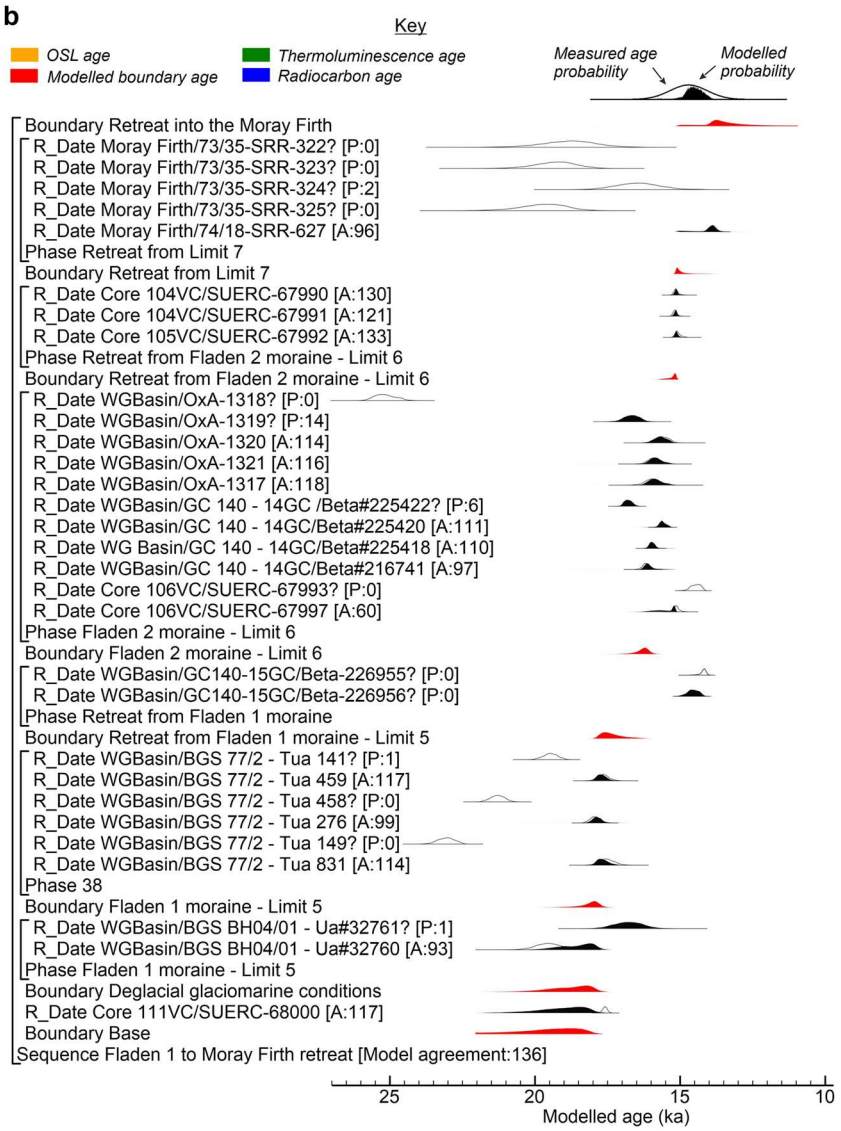

Figure 7. Continued

1) is recorded by till at the base of cores 138–141VC and 177VC and dated to 29–30 ka in cores 178VC and 179VC, all located south of Dogger Bank (Roberts *et al.*, 2018; Table 1; Figs 1 and 4b). The glacilacustrine sediments of the Dogger Bank date the Dogger Lake to 31.7–25.8 ka (Shfd15175, Shfd15176, Shfd15179, Shfd15180, Table 1). The western end of the lake was glacially overridden by ice that extended to ~54°N, co-incident with the southern limit of the Bolders Bank Formation (subglacial till). The overriding of the glacilacustrine deposits resulted in compressive glacitectonism to form the Dogger Bank. This phase of overriding and subsequent northwards recession appears to have been rapid, being complete by 23.1 ± 2.3 ka (Shfd15178, Table 1), and resulted in the construction of moraine complexes and ice-marginal deposits on both the southern and northern edges of the newly formed Dogger Bank. Retreat from Limit 1 is recorded by outwash over till in cores 178VC and 179VC, lake sediments in cores 150VC and 151VC (Roberts *et al.* 2018), and glacilacustrine sediments and outwash on the proximal side of the North Ferriby moraine (Fig. 4a; Bateman *et al.* 2018). The Bayesian analysis identifies OSL age Shfd13073 as too young relative to subsequent locations in the model, and the remaining six ages divide into two clusters: an older group (Shfd13071, Shfd15180 and Shfd15175) is centred on 29–30 ka and a younger group (Shfd13072, Shfd15176 and Shfd15179) centred on 26–27 ka. The contexts for the two groups intermix within sequences and, given the potential for poor resetting of the OSL signal in proglacial outwash, the younger group is the more likely to be valid. This produced good agreement (>60%) in the model with the other four ages treated as complete outliers ($P = 1$) (Shfd13071/N Ferriby, Shfd15175 and Shfd15180/Dogger Bank), and constrains the advance to Limit 1 as post-25.8 ± 1.4 ka and pre-24.6 ± 1.0 ka (Fig. 7a). Assessing the influence of imposing the "After" constraint on the North Cave ages (Fig. 7a) was explored by coding those ages instead as a Phase, which also produced a conformable model (not shown) and constrains the Limit 1 advance to post-28.1 ± 2.0 ka and pre-25.0 ± 1.1 ka. The parsimonious interpretation is that ice advance into the region occurred after 28.1 ± 2.0 ka, reaching the maximum regional extent at 25.8 ± 1.4 ka to 24.6 ± 1.0 ka.

Limit 2 – NSL oscillation (~26–24 ka) and retreat (~22 ka)

A subsequent advance to Limit 2 at 24–26 ka is evidenced by a buried moraine beneath the seafloor at core site 142VC and till patches over Dogger Bank (LF DB4b of Roberts *et al.* 2018). Onshore, the equivalent evidence for this advance is the lower

tills at Sandsend (Roberts *et al.* 2013) and Little Catwick/Gembling (Bateman *et al.* 2018; Fig. 4a). At Sandsend, OSL ages of 23.8 ± 1.8 ka and 23.4 ± 1.7 ka (Shfd13066 and Shfd13068; Table 1) date the uppermost sand and gravel outwash in a deglacial coarsening-upward lacustrine-to-fluvial sequence and thereby provide a minimum date on the underlying till. The Little Catwick/Gembling stratigraphy is typical of the advance and retreat/readvance Skipsea Till sequence identified by Evans and Thomson (2010) using regional stratigraphic reconstructions, with the lower till at the site appearing at a more advanced marginal position of the NSL than that preserved in the sediments at Flamborough, Barmston and Dimlington. At Little Catwick/Gembling and other sites further north, subaqueous fans and lake rhythmites deposited into a northern arm of Lake Humber were deposited over Skipsea Till at 23.5 ± 1.8 ka and 23.6 ± 1.7 ka (Shfd13064 and Shfd13065; revised from 22.8–23.4 ka in Bateman *et al.* 2018; Table 1). This advance also appears to equate with the earliest impoundment of Glacial Lake Humber, dated at 24.1 ± 1.5 ka (Bateman *et al.* 2000; revised from 22.7 ± 1.4 ka) using thermoluminescence (TL) at North Kelsey Road, Caistor (Fig. 4a; Back 1; Table 1) and 24.4 ± 1.6 ka (Bateman *et al.* 2008; revised from 23.3 ± 1.5 ka) using OSL of sub-till silts at Ferrybridge (Fig. 4a; Shfd02013; Table 1).

Retreat from Limit 2 appears to have taken place at around 23 ka and is constrained offshore by glacilacustrine sediments that on-lap recessional moraines along the northern edge of Dogger Bank. These sediments show no evidence of subsequent glacial overriding and are OSL-dated in core 155VC to 23.1 ± 2.3 ka (Shfd15178; Table 1). Onshore, this ice recession is constrained chronologically by the deglacial outwash between tills at Sandsend dated to 23.4 ± 1.7 ka and 23.8 ± 1.8 ka and at Catwick/Gembling dated to ~23.5 ka (see above; Fig. 4a). The advance and retreat phase boundaries for Limit 2 from the Bayesian model are dated 23.5 and 22.2 ka, respectively (Fig. 7a).

Limit 3 – NSL advance (~21.5 ka) and subsequent retreat

A significant readvance to Limit 3 and subsequent retreat are recorded to the west and southwest of Dogger Bank and relate to the NSL flowing between Dogger Bank and the Yorkshire/Lincolnshire coasts down to north Norfolk, where the Garret Hill Moraine was constructed onshore. Here a maximum age on the emplacement of the Hunstanton Till (Holkham Member) and the associated Garret Hill Moraine (Pawley *et al.* 2006) is reported by Roberts *et al.* (2018) and Evans *et al.* (2019) at 21.5 ± 1.3 ka (Shfd15033; Table 1) from glacitectonised glacifluvial outwash overlying MIS 12 till (Fig. 4a). The outwash dates to 22.8 ± 1.8 ka (Shfd15034; Table 1) and dates the attainment of Limit 3 by the NSL in this area. Retreat from Limit 3 is marked offshore of north Norfolk and Lincolnshire, by a series of superimposed, lobate-shaped till wedges that post-date the initial phases of the emplacement of the Bolders Bank Formation (Dove *et al.* 2017). These demonstrate that the NSL underwent a series of quasi-stable oscillations during recession.

Further north, in Lincolnshire, an OSL age of 19.5 ± 2.6 ka (Shfd13074; Table 1) on glacitectonised ice-dammed lake sediments from beneath the Marsh Till at Welton-le-Wold (Fig. 4a) provides a maximum age on the advance to the Stickney and Horkstow moraines. In Yorkshire, this event is recorded by the emplacement of the Skipsea Till over the organics at Dimlington (Fig. 4a), radiocarbon-dated by Penny *et al.* (1969) at 22 cal ka BP (Birm-108 and I-3372; Table 1) and OSL-dated to 23.5 ± 1.6 ka and 22.8 ± 1.4 ka (Shfd09062 and Shfd09063; revised from 21.2 and 20.5 ka in Bateman *et al.* 2015; Table 1), and at Flamborough Head (Fig. 4a) by sub-till sands that were OSL-dated by Bateman *et al.* (2015) and now have revised ages of 20.9 ± 1.4 ka and 21.1 ± 1.3 ka (Shfd10084, Shfd10085; Table 1). Dates recording the continued existence of Glacial Lake Humber at this time (Table 1) have been reported from Hemingbrough (Fig. 4a) and range from 24.1 ± 2.2 ka to 21.0 ± 1.9 ka (HEMA01-HEMA03; Table 1), although HEMA03 at 24.1 ka shows as a clear outlier (Fig. 7a).

At Lindholme (Fig. 2), Bateman *et al.* (2015) report ages of 18.2 ka and 19.1 ka (Shfd10071 and Shfd10072; revised here to 18.9 ± 3.0 ka and 19.9 ± 3.4 ka; Table 1) on glacial outwash sands in the vicinity of the proposed ice limit at Wroot (Fig. 4a). This appears to be the short-lived maximal limit of an ice lobe that extended down the Vale of York (Friend *et al.* 2016; Fig. 12 of Bateman *et al.* 2018). More prolonged stillstands are recognised by the extensive moraines at Escrick and York.

The age of retreat from Limit 3 is further documented at Dimlington on Holderness, in relation to inter-till stratified deposits, and at Hemingbrough in relation to Glacial Lake Humber deposits. A range of ages from 18.5 to 22.8 ka (Shfd07113-Shfd07115 and Shfd09062-Shfd09063; Table 1) on sands between the traditionally classified Skipsea and Withernsea tills at Dimlington (Catt & Penny 1966; Rose 1985; Boston *et al.* 2010; Evans & Thomson 2010; Fig. 4a) provide a chronological control on NSL recession from the East Yorkshire coast and the sedimentation of ice-contact to proglacial subaqueous fans and proximal bottomsets (LFA 2 of Evans & Thomson 2010). At Hemingbrough (Fig. 4a), OSL ages of 19.2 ± 0.5 and 18.0 ± 0.6 (Shfd11087 and Shfd11088; revised from the 16.8 and 15.5 of Bateman *et al.* 2015; Table 1) indicate that Glacial Lake Humber still existed at this time (NB. Shfd11088 is an outlier in the Bayesian analysis; Fig. 7a) and hence the Humber Estuary remained blocked even though the Holderness ice margin had receded enough to allow the development of an extensive lake to the north; this is therefore most likely a northern arm of Lake Humber (Fig. 12 of Bateman *et al.* 2018). The advance and retreat phase boundaries for Limit 3 from the Bayesian model are dated 21.5 and 20.8 ka, respectively (Fig. 7a).

Limits 4 and 5 – NSL oscillations and retreat (~20–17.2 ka)

A long-recognised minor readvance on the East Yorkshire coast is defined here as Limit 4 and is demarcated by the emplacement of the Withernsea Till in the area around Dimlington (Catt & Penny 1966; Rose 1985; Evans & Thomson 2010). Although this readvance has traditionally been emphasised because it is stratigraphically documented by the Withernsea Till, it is only one of a number of potential onshore readvances (oscillations) by the margin of the NSL recorded in interbedded sequences of tills and ice-contact lake deposits on Holderness (Boston *et al.* 2010; Evans & Thomson 2010). Moreover, Boston *et al.* (2010) indicate that the Withernsea Till is lithologically internally complex and cannot be unequivocally differentiated geochemically from the equally complex Skipsea Till. Hence the same readvance could be recorded north of Dimlington but in till that has traditionally been equated with an earlier "Skipsea Till advance". This is verified by OSL dates of 23.6 ± 1.9 ka (single grain) and 20.4 ± 1.6 ka (single aliquot) from sands and gravels that have been incorporated into the base of the Skipsea Till at Barmston (Fig. 4a; Shfd10215, Table 1; revised from 18.6 ka and 21.5 ka of Bateman *et al.* 2015). These sediments provide a maximum age on the emplacement of the upper till along the Holderness coast and hence are derived from the chronos-

tratigraphically equivalent inter-till sand dates from Dimlington, which clearly predate the Withernsea Till at its type site. Offshore, the thick accretionary till wedges identified by Dove *et al.* (2017) record the dynamic, quasi-stable oscillation of the NSL during the onset of this period (i.e. Limit 4).

Retreat from Limit 4 was marked by the withdrawal of the NSL from the eastern England coast into the central North Sea. At Barmston this is dated at 20.4–18.0 ka based on single aliquot OSL ages from sands and gravels overlying the Skipsea Till (Shfd10214 annd Shfd11007, Table 1; revised from 15.0 ka and 16.3 ka of Bateman *et al.* 2015). The deglaciation of the Lincolnshire Marsh is clearly recorded geomorphologically by the superimposition of glacier karst features (eskers and ice-walled lake plains) on overridden moraine ridges, collectively known as the Hogsthorpe–Killingholme Moraine (Evans *et al.* 2019). This phase of downwasting is dated by OSL ages of 19.3 ± 1.4–17.4 ± 1.1 ka (Shfd15154 and Shfd15155; Table 1) from sands in one of the many ice-walled lake plains at Thoresthorpe (Fig. 4a). At Sewerby, north of Holderness (Fig. 4a), glacifluvial outwash gravels and sands overlying the Skipsea Till are OSL-dated to 18.6 ± 1.9 ka (Shfd05250, Table 1; revised from the 17.9 ka of Bateman *et al.* 2015) and record the final phase of terrestrial deglacial sedimentation in the area. Further north, the later stages of Glacial Lake Pickering sedimentation at East Heslerton (Fig. 4a) are dated to 18.4 ka (Shfd13054 and Shfd13055; revised from the 17.6 ka of Evans *et al.* 2017; Table 1). For both lakes Humber and Pickering, the later stages of sedimentation were likely conditioned as much by moraine-damming as by ice-damming. Lake Pickering was impounded by the Filey and Flamborough Head moraine complexes and Lake Humber by the Elloughton/Ferriby/Horkstow moraine complex until its breaching and incision (cf. de Boer *et al.* 1957; Gaunt *et al.* 1971; Gaunt 1974; Bateman *et al.* 2018). Retreat offshore is documented by the sediments of core 137VC (Fig. 4b), within which a radiocarbon age of 19.8 ± 0.2 cal yrs BP (UCIAMS-176372; Roberts *et al.* 2019; Table 1) from deglacial glacimarine sediments represents a minimum age for deglaciation. The advance and retreat phase boundaries for Limit 4 from the Bayesian model are dated 19.7 and 19.5 ka, respectively (Fig. 7a).

Further north, at Seaham on the Durham coast (Fig. 4a), OSL ages of 19.9 ± 2.3–19.1 ± 1.9 ka (Shfd14064-Shfd14066; Roberts *et al.* 2019; Table 1) can be used to chronologically constrain both the retreat from Limit 4 and the subsequent advance to Limit 5, because they are derived from glacifluvial outwash sands and gravels between two tills (Fig. 4a). In addition to the upper till at Seaham, the sediments of offshore core 132VC also record the advance to Limit 5 and include glacimarine deposits dated to younger than 19.6 ± 0.2 cal yrs BP (SUERC-68010; Roberts *et al.* 2019; Table 1) overlain by till. This ice-sheet margin is demarcated by a grounding zone wedge (W1 of Roberts *et al.* 2019) offshore of the Durham coast (Fig. 1). The advance and retreat phase boundaries for Limit 5 from the Bayesian model are dated 18.4 and 17.3 ka, respectively (Fig. 7a).

Limit 6 – NSL retreat (~19–16.5 ka)

To the south of core 111VC, there is a spectacular series of north–south-aligned and infilled tunnel valleys (Figs 1 and 6). These transition westward into a series of seafloor ridges that have for many years been described as the Wee Bankie Moraine complex (Thomson & Eden 1977; Stoker *et al.* 1985; Gatliff *et al.* 1994; Merritt *et al.* 2003). Six distinctive acoustic facies (Wee Bankie (WB)/AF1-5) can be identified (Fig. 5a). There is a clear lower erosion surface across the base of the moraine complex. Below this (WB)/AF1 is generally acoustically transparent and massive, but in several places sub-horizontally orientated sediment packages occur below this undulatory erosion surface. These may be pre-MIS 2 sediments (Yarmouth Road Formation, Fisher Formation) but could also be MIS 2 tills. None were cored as part of the BRITICE-CHRONO project, hence further investigation is required. The tunnel valley infill (WB/AF2) contains several cut and fill sequences composed of grey, laminated clays and silts (see core 125VC; Fig. 5a).

Immediately to the west of core location 125VC, the architecture of a valley infill shows shallow delta foresets in a valley-marginal location, prograding into delta bottomsets (Fig. 5a, WB/AF2; 127VC–118VC). The upper surface of this delta (if marine in origin) provides a possible control on early sea level (91 m below OD) in the central basin. From this point westwards, the seafloor morphology changes, with a series of drowned ice-contact deltas (WB/AF3; Fig. 5a) marking ice recession into the Firth of Forth. Between cores 121VC and 124VC, the delta foresets are particularly prevalent in the acoustic stratigraphy and, again, if marine in origin their upper surfaces could provide a control on deglacial sea level (e.g. core site 121VC is at 57 m below OD) (see later discussion). These deltas form the core of the outer Wee Bankie Moraine complex and in places the acoustic architecture also shows feeder tunnels/channels and eskers connecting to the deltas (WB/AF4; Fig. 5c). However, west of core site 120VC the architecture of the seafloor changes, with large, bedrock and till-cored ridges (WB/AF1) forming prominent moraines. These are draped by laminated clays and silts containing marine fauna (WB/AF5) and in cores 118VC–120VC, radiocarbon-dating of mixed foram samples and a mollusc sample of *Nuculana pernula* provided minimum deglacial ages for the inner Wee Bankie Moraine complex of 17.9 ± 0.2, 16.6 ± 0.2, 18.0 ± 0.2 and 17.6 ± 0.2 cal ka BP, respectively (SUERC-68001 and 68007; Roberts *et al.* 2019 and SUERC-68008 & 68883; Table 1). These ages are supported by onshore dates from the Tay and Forth estuaries, where glacimarine sediments associated with the Errol Beds Formation show that ice had moved west of the present coastline prior to 16.9 ± 0.2–16.0 ± 0.1 ka (SRR-391, AA-37787, Beta-111508, CAMS-77912 aand OxA-1704, Table 1; see Harkness & Wilson 1979; Hedges *et al.* 1989; Peacock & Browne 1998; Peacock 2002; Hughes *et al.*, 2011, for overview).

In summary, retreat from Limit 5 is recorded by dates from glacimarine sediments in offshore cores 118VC–120VC. Radiocarbon ages from these deglacial deposits range from 18.0 to 16.6 cal yrs BP (Table 1). They lie beyond the W2 grounding zone wedge situated offshore of Berwick and were used by Roberts *et al.* (2019) to demarcate Limit 6. A minimum OSL date of 16.2 ± 0.8 ka from glacifluvial outwash at Maiden Castle in Durham City (Fig. 4a; Tables 1, S1; Shfd14067) records the final stages of sedimentation in the Wear valley. Retreat from Limit 6 is also recorded by ice distal glacimarine sediments from core 128VC, collected from inside the W2 grounding zone wedge offshore of Berwick. A radiocarbon date of 17.0 ± 0.2 cal yrs BP from these sediments (SUERC-68009; Roberts *et al.* 2019; Table 1) is compatible with those from the early deglacial raised marine Errol Beds Formation of the Tay and Forth estuaries, which indicate ice withdrawal from the Scottish east coast in this area prior to 17.0–16.0 cal yrs BP (Gallowflat/Kinneil Kerse/Barry Clay Pit, Fig. 4a; Harkness & Wilson 1979; Hedges *et al.* 1989; Peacock 2002; Table 1). The retreat phase boundary for Limit 6 from the Bayesian model is dated to 15.8 ka (Fig. 7a).

Northern sector

Defining and dating pre-LGM ice extent

The glacial history of eastern Scotland is complex and much debated. For many years it was argued that the eastern areas of Buchan remained ice free during MIS 2 (Synge 1956, 1963; Sutherland 1984), but more recently it has been demonstrated that the regions of Buchan and Moray were fully glaciated by ice sourced from the northwest and west and later bounded offshore by large ice streams (Strathmore and Moray Firth ice streams) that flowed into the northern and central North Sea during the later phases of the last glacial cycle. Along the southern edge of the Moray Firth, a number of sites with sandy tills indicate a northwest-to-southeast ice advance across the Moray Firth between ~45.4 and 37.8 ka based on OSL dating of gravel fans (Duller *et al.* 1995). Shelly tills across the lowlands of Buchan and Moray post-date the early sandy tills in this region (Peacock and Merritt, 2000a), and together with erratic trains and striae support southeasterly flow with thick ice over the eastern Grampians and Strathmore. Merritt *et al.* (2017) assign this to 32–24 ka and propose full shelf-edge glaciation of the northern North Sea by this time, with ice over Caithness, Orkney and Shetland coalescent with the FIS. Between 24 and 23 ka, Merritt *et al.* (2017) suggest ice thinning and flow reorganisation over eastern and northeast Scotland. This evidence is linked to flow-set mapping and till superimposition with ice flowing east-northeast/northeast (Merritt & Auton 1993; Fletcher *et al.* 1996; Hall *et al.* 2016a, b). This possibly marks the early onset of the Moray Firth Ice Stream but geochronological control is weak.

The offshore region east of the Firth of Forth is covered in a diverse mosaic of Quaternary sediments (Gatliff *et al.* 1994; Davies *et al.* 2011). The older (>MIS 6) Quaternary sediments infill a complex series of anastomosing tunnel valleys that run north to south (Huuse & Kristensen 2016). Those older sediment sequences include the Yarmouth Road Formation, Fisher Formation and Coal Pit Formation (Davies *et al.* 2011). With respect to constraining ages on LGM onset, there is a single radiocarbon age of 36.8 ± 0.7 ka (SRR-637) from a piece of lignitised wood in cold shallow marine sediments associated with the Aberdeen Ground Beds, 300 km east of the Firth of Forth (Holmes 1977).

In the northern sector of the North Sea, the onset of MIS 2 glaciation is constrained by several radiocarbon ages in the Fladen and Witch Ground areas, which limit ice build-up to post-33 ka. In particular, Graham *et al.* (2010) report reworked forams and shell fragments in overrun glacimarine sediments in core BGS BH 04/01, which range from 35.2 to 32.4 ka. Associated with this, Graham *et al.* (2010) report the production of mega-scale glacial lineations (MSGL) as ice streamed northwest along the suture zone between the FIS and BIIS, though the exact position and longevity of coalescence between the BIIS and FIS remain contentious. Sejrup *et al.* (1994) also report reworked shells in till to the east of the Witch Ground to 37.0 ± 0.7 (Table 1; TUa-274). Hughes *et al.* (2016) reconstruct coalescence between ~27 and 19 ka based on the available chronology. In contrast, Patton *et al.* (2016, 2017) limit coalescence to a brief window ~23–22 ka based on a 3D thermomechanical model validated against geomorphological, geochronological and sea-level datasets. Sejrup *et al.* (2016) suggest a longer coalescence period between 23 ka and 19 ka.

In the central sector of the North Sea (east of the Firth of Forth), glacigenic sediment sequences and landforms relating to MIS 2 glaciation are very poorly constrained. The Wee Bankie Moraine complex (and associated sediments) is situated ~50 km east of St Andrews, and marks ice retreat from the central North Sea, but beyond this the ice extent in the central North Sea during MIS 2 remains largely unexplored. Stoker *et al.* (1985) and Cameron *et al.* (1987) report the widespread occurrence of the Forth Formation in this region, which is a deglacial phase glacimarine drape that covers much of the seafloor, but there are few reports of subglacial or ice-marginal sediments that fix BIIS ice extent or provide evidence for coalescence with the FIS.

Between ca. 22 and 19 ka, Merritt *et al.* (2017) suggest Moray Firth ice thickened along the Moray and Buchan coasts, as the Strathmore Ice Stream began to feed ice northwards from the inner Firth of Forth. This is evidenced by the red tills, striae and subglacial bedforms along the eastern margin of the ice stream. It is highly likely that the BIIS and FIS remained coalescent across the central and northern North Sea at this time. Radiocarbon dates from the Witch Ground Basin cores (Sejrup *et al.* 2015) do not necessarily support early marine inundation between 25 and 19 ka and the Wee Bankie Moraine complex relates to a later period of ice-marginal stabilisation (Roberts *et al.* 2019; see *Discussion* below). The latter stages of recession in the region are also marked by ice thinning along the southern Moray and Buchan coasts and the development of ice-marginal lakes (glacial lakes Ugie and Ythan; Merritt *et al.* 2017). The east Grampian sector of the BIIS receded under cold-based conditions at this time, as evidenced by abundant ice-marginal meltwater channels.

Deglaciation of the central North Sea (18–15.2 ka; retreat through Limits 5 and 6)

The timing and nature of deglaciation in the northern North Sea is complex. Along the western margin of the NCIS between Ling Bank and the Viking Bank/Plateau, grounding zone wedges suggest ice initially deglaciated southward as the NCIS waned (Figs 1 and 5). However, the Tampen Ridge (Fig. 3) indicates re-expansion of the NCIS early during deglaciation. It has recently been hypothesised that the Ling Bank may have acted as a spillway for the decanting of the Dogger Lake (the Dogger Lake Uncoupling Event or DLUE) northwards as the FIS and BIIS uncoupled (Sejrup *et al.* 2016), and a new chronology based on an oxygen isotope anomaly related to the meltwater plume has fixed this event at 18.7 ka (Hjelstuen *et al.* 2018). This would mark the initial uncoupling of the BIIS and FIS at ~19 ka and suggests that the BIIS and FIS uncoupled along a south-to-north trajectory. Cores collected during the BRITICE-CHRONO cruise JC123 in the North Sea confirm that subglacial tills to the north of Ling Bank are draped by glacimarine sediments, but the oldest constraining date on deglaciation was 14.5 ± 0.3 ka (SUERC-67989) from core 097VC from the northern edge of the Ling Bank (Table 1). However, the early onset and timing of marine inundation can be established using several lines of other evidence. Drowned cuspate foreland shorelines at 106 m below OD south of the Ling Bank further allude to very early marine inundation of this sector of the North Sea. This is consistent with BRITICE-CHRONO sea-level reconstructions generated using the output of a regional ice-sheet model incorporated into a glacial-isostatic adjustment model (Ely *et al.* 2019), the model being validated using ice margins identified herein for the BIIS and the *DATED* ice margins for the FIS (Hughes *et al.* 2016) (Fig. 8).

The drowned deltas reported from further west on the seafloor between cores 121VC and 125VC (Figs 5c and 8) also help to constrain the nature and timing of marine inundation during initial deglaciation. As with the cuspate foreland, deltas 1 and 2 must represent drowned landscapes as they do not intersect relative sea level (Fig. 8). In order for it to be preserved, the cuspate foreland must have been instantaneously drowned

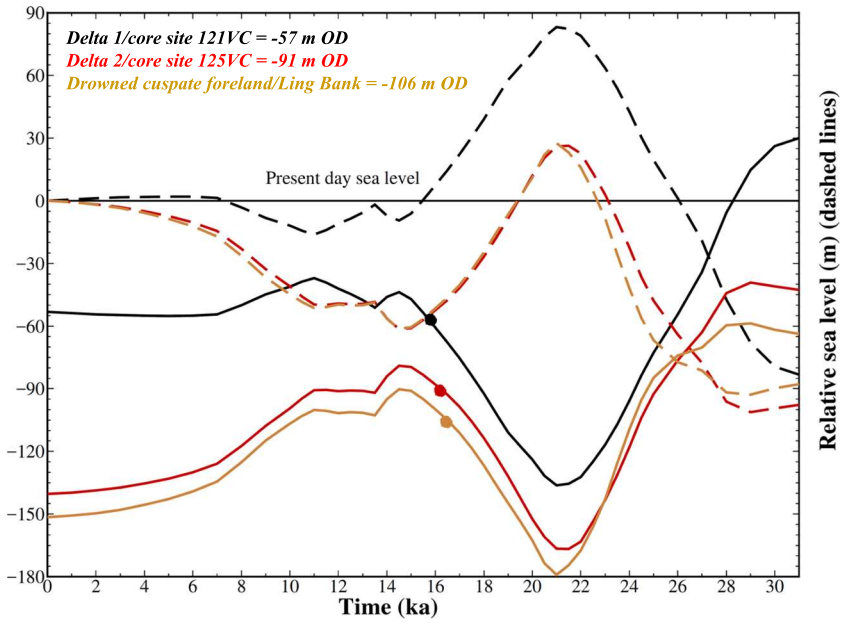

Figure 8. Predicted relative sea level (dashed) and water depth (solid) at three sites: core 121VC (black line; 56.42°N,1.64°W), core 125VC (red line; 56.68°N,1.3°E) and Ling Bank (brown line; 58.75°N,3.5°E). Model output is generated from a glacial-isostatic adjustment model updated to include the latest regional ice-sheet reconstruction (BIIS and FIS) from BRITICE-CHRONO. The water depth is calculated by subtracting the predicted relative sea level from present-day bathymetry.

following BIIS/FIS decoupling; in contrast, the deltas are dominated by foreset-type acoustic stratigraphy, inferring ice-contact sedimentation into deep water as the BIIS grounding line retreated to the west. Taking into account our new regional framework for regional deglaciation, the cuspate foreland and deltas can be assigned an age of ~19–17ka (Figs 5c and 8).

The initial retreat of the BIIS westward through the Fladen and Witch Ground regions is constrained by moraines and glacigenic debris flows that mark ice margin oscillation. They are dated using two cores (BGS 77/2 and BGS BH 04/01; Graham *et al.* 2010; Sejrup *et al.* 1994, 2015; Fig. 4b) but age reversals hinder a simple interpretation. Throughout both cores both foram and shell ages reverse consistently. This relates either to the reworking of older material, contamination by old carbon or bioturbation. In core BGS BH 04/01, a grounding line efflux unit has two dates 19.6 ± 0.4 and 16.8 ± 0.1 cal ka BP (Ua-32760 and Beta-225422; Table 1). These are very similar to two dates from the lower glacimarine unit in BGS 77/2, which date to 19.5 ± 0.3 and 17.7 ± 0.2 cal ka BP (TUa-141 and TUa-459). Taken together they bracket the Fladen 1 Moraine (Sejrup *et al.* 2015; Figs 1, 6) and would fit with a DLUE. To the northeast and northwest of the Fladen 1 Moraine, a more prominent moraine (Fladen 2; Fl2 in Figs 1 and 6) is bracketed by three cores in glacimarine sediment (BGS core 58 + 00/111; Witch Ground Basin 58°34.5′N, 00°24.1′E; Long *et al.*, 1986); GS140-05-14GC and GS140-05-15GC; Sejrup *et al.*, 2015), all of which provide minimum constraints on deglaciation (15.9 ± 0.2, 16.2 ± 0.1 and 14.6 ± 0.2 cal ka BP, respectively) (Table 1). It should be noted that GS140-05-14GC is characterised by age reversals. To the south of GS 140-05-15GC, a glacigenic debris flow, potentially sourced from the east, has been linked to a "Tampen Readvance" margin by Sejrup *et al.* (2015) but geomorphologically this is difficult to substantiate. To summarise, the key offshore moraine margins that constrain BIIS/FIS uncoupling and grounding line recession under glacimarine conditions are Fladen 1 and 2 which lie in a time window from 19.5 cal ka BP to 16.2 cal ka BP (Sejrup *et al.* 1994, 2015; Graham *et al.* 2010). Here we equate moraine margins Fladen 1 and 2 (Sejrup *et al.* 2015) with Limits 5 and 6 based on their chronological overlap. The advance and retreat phase boundaries for Limit 5 (Fladen 1) are dated to 18.0–17.4 ka using the Bayesian model, while Limit 6 (Fladen 2) is dated to 16.3–15.2 ka (Fig. 7b).

South of the Witch Ground Basin region, in the central North Sea, a series of cores (JC123: 111VC–117VC) and geophysical data from BRITICE-CHRONO cruise JC123 (Fig. 5b) shed further light on possible ice-marginal positions as the BIIS decoupled from the FIS in the vicinity of the Great Fisher Bank (Fig. 6). No subglacial tills were recovered in these cores but glacimarine and possible glacilacustrine sediments drape the seafloor. Six acoustic facies (Great Fisher Bank (GFB)/AF 1–6) were mapped in relation to these cores. GFB/AF1 is a clear lower reflector, possibly of Yarmouth Roads Formation or Fisher Formation origin. Overlying this are a series of on-lapping semi-transparent units of variable thickness (GFB/AF2; Figs 5b and 6). They are often incised by infilled channels/valleys and in places form distinctive 'bumps' on the seafloor, with inclined internal reflectors indicative of thrusting and stacking; hence they are interpreted as moraines within which GFB/AF2 is composed of subglacial till (Fig. 5b, 111VC–117VC). Draping these sediments and infilling numerous valleys is GFB/AF3, a stratified lithofacies composed of laminated fines. In cores 112VC–114VC (Figs 5b and 6), toward the eastern end of the area, an olive grey clay with thin sand lenses and occasional soft sediment deformation structures occurs and is devoid of fauna. At the location of cores 112VC–114VC the seafloor is marked by a distinctive scarp/step with a steeply dipping western face and gently dipping eastward slope (Figs 4b and 5b). The geophysical data from this location show this scarp feature to be composed of folded or dissected GFB/AF3, overlain by gently dipping GFB/AF4 and topped by GFB/AF5 (a silty sand). It has the internal architecture and surface profile of a subaqueous ice-contact fan (e.g. Cheel & Rust 1982; Powell 1990; Lønne 1995). The lack of fauna in cores 112VC–114 VC suggests GFB/AF3 may be of glacilacustrine origin. The Great Fisher Bank, slightly further east (Fig. 1), on the other hand, is not an ice-contact feature but rather a large sand bar, composed exclusively of Holocene sand; many areas of the seabed are truncated by migrating sand bars. A peat sample from a short core (117VC) just below the seafloor returned a radiocarbon date of 13.4 ± 0.1 cal ka BP (SUERC-72877, Table 1),

constraining the timing of marine inundation in this shallow eastern sector of the North Sea Basin.

Further to the west, core 111VC (Figs 4b and 5c) records a strong basal reflector overlain by several acoustic facies. This site is 150 km west of core site 112VC, hence cross correlation of acoustic facies is not feasible. The cores from 111VC heading north to 104VC are given the prefix Bosies Bank (BB)/AF1–6. Above the basal reflector several units are transparent to semi-transparent, massive and structureless (BB/AF2). They are also associated with possible moraines and interpreted as subglacial tills. At core site 111VC, a small valley incision was cored and this recovered 5 m of sediment. At the base of the core was a laminated clay/silt overlain by more massive grey/brown clays with sandy inclusions. This unit is assigned to BB/AF3 but contains marine fauna, indicating that it is glacimarine rather than glacilacustrine. A shell fragment provided a minimum deglacial age of 17.6 ± 0.2 cal ka BP (core 111VC; SUERC-68000; Table 1), consistent with our earlier interpretation that ice was retreating towards Limit 5 (Fladen 1).

Limit 7 (~15 ka)

Additional control on the retreat of ice through the Bosies Bank and into the Moray Firth areas comes from cores 104VC–111VC which span a 180 km transect that displays a series of inset moraines marking grounding line retreat towards Bosies Bank (Figs 1 and 5c). The seafloor in this region is composed of six acoustic facies. BB/AF1 is bedrock/pre-MIS 2. It is overlain by two acoustically transparent units BB/AF2 and BB/AF3 that have the geophysical hallmarks of subglacial till (Fig. 5b). In places these two facies form the cores of moraine complexes. Above this, stratified fines (BB/AF4) drape the underlying topography and on-lap the seafloor moraines. These are glacimarine sediments and contain *in situ* glacimarine fauna. Occasionally they are overlain by an upper diamicton (BB/AF5) that is interpreted as subglacial till relating to ice margin oscillation (see core 104VC). Above BB/AF4 and BB/AF5 the sequence is capped by a thin silty sand with ubiquitous shell fragments, which is a product of Holocene-to-recent seafloor processes (BB/AF6). Three cores provide minimum ages on ice retreat towards the Bosies Bank moraine complex. Cores 104VC–106VC contain four radiocarbon ages on *in situ* forams from glacimarine sediments overlying tills, which cluster at ~15.0 cal ka BP (SUERC-67990–67992 and 67997, Table 1). These tie in with ages from the Witch Ground Basin slightly further east where retreat began prior to 16.2 cal ka BP.

Graham *et al.* (2009) suggest ice was flowing east/south-eastwards towards the Witch Ground Basin prior to retreat to the Halibut and Bosies Bank moraine complexes (Fig. 1), although Finlayson *et al.* (2014) also show convincing evidence for MSGL orientated more southwest to northeast. Bosies Bank and Halibut Bank are related to a period of ice margin stability or readvance, or possibly a surge, with Graham *et al.* (2009) reporting evidence for ice-marginal glacitectonics as well as stagnation under rising sea-level conditions. West of Bosies Bank the seafloor is crosscut by a spectacular set of rock cut subglacial channels trending east to west. Further west, the Elgin Moraine (Peacock, 1968) and De Geer moraines along the coast at Tarbert Ness (Fig. 1) suggest glacimarine conditions along the Moray Firth Ice Stream margin as it retreated landward in a sea level that was at ~55 m OD (Finlayson *et al.* 2014). Radiocarbon dates on marine bivalves in glacimarine deposits at St Fergus suggest ice margin oscillation after 17.5 cal ka BP (Beta-101953, Lu-3028; Table 1). Four other radiocarbon ages from a core (Moray Firth/73/75) just north of the Banff coast limit ice retreat through this area to 20.2–17.0 cal ka BP (SRR-322–SRR-325; Table 1). However, the Bayesian analysis suggests that the majority of these are outliers and hence anomalously old (Fig. 7b). Offshore, core 74/18 in the centre of the Moray Firth also contains glacimarine sediments dated to 13.9 ± 0.1 cal ka BP (Harkness and Wilson, 1979), providing the only viable limiting deglacial age for this area according to the Bayesian model (Figs 6 and 7). The Bayesian model constrains Limit 7 to 15.0 ka with deglaciation of the inner Moray Firth by 13.5 ka.

Deglaciation of the Buchan coast under glacimarine conditions is inferred from glacimarine deposits and nearshore De Geer moraines lying just offshore between St Fergus and Lunan Bay (Bradwell *et al.* 2008), which suggest that the Strathmore Ice Stream began to retreat under marine conditions. Its northward trajectory at this late stage was likely a result of thinning and topographic confinement, or alternatively, it formed the northern sector of a much larger Firth of Forth piedmont lobe (Roberts *et al.* 2019). Previous reconstructions highlight glacimarine clays at Lunan Bay dated to 20.1–20.9 cal ka BP (McCabe *et al.* 2007) as indicative of ice-free conditions on the central Buchan coast, but this would signify a very early collapse of the Strathmore Ice Stream and early marine inundation of the northern and central North Sea Basin (Sejrup *et al.* 2015). This is incompatible with the regional offshore deglacial history presented here and with sea-level reconstructions by Shennan *et al.* (2006, 2018), which indicate that the area only became ice-free after ~16–17 ka.

Discussion: regional palaeoglaciological reconstruction

Based upon the details outlined above and their reconciliation with wider palaeoglaciological reconstructions pertaining to northern England and eastern Scotland we can now address the key research questions posed in the introduction to this study.

How far did the BIIS extend?

The southernmost limit of the NSL is recorded offshore by the Bolders Bank Formation south of the Dogger Bank (Limit 1) and onshore by the Garret Hill Moraine in north Norfolk (Limit 3). Limit 1 is dated to 31.4–25.3 ka by OSL dates overlying Bolders Bank Formation till in several cores (138VC–141VC, 178VC and 179VC; Roberts *et al.* 2018). Ice-marginal oscillations between 28 and 22 ka in the Dogger Bank region then produced several additional Bolders Bank Formation tills and it is these that overprint the seafloor down to Limit 3, forming the Hunstanton Till limit and Garret Hill Moraine on the Norfolk coast at 21.5–20.7 ka. Hence Limits 1 and 3 are partially to totally overprinted at the southern extent of the BIIS, as recently demonstrated by Dove *et al.* (2017). This overprinting is likely manifest also in the complex till stratigraphies and inset moraines along the Lincolnshire and East Yorkshire coast (i.e. Skipsea Till, Marsh Tills, Hunstanton Till, Stickney Moraine and Horkstow Moraine) and in the seafloor sedimentary architecture as the ice withdrew northwards towards Limit 4 (Dove *et al.* 2017). The existence of ice-dammed lakes along the east coast while the NSL occupied Limits 1–3 is reinforced by several OSL dates (e.g. 24.1–22.3 ka) limiting the development of Glacial Lake Humber (Bateman *et al.* 2000, 2015; Murton *et al.* 2009). In our model the Vale of York lobe occupied the York/Escrick moraines at ca. 21.5–20.7 ka (Limit 3) but potentially extended further south for a short period prior to this.

In the central North Sea, the extent of the BIIS remains largely unconstrained but has been regarded as coalescent with the FIS over the area of the Great Fisher Bank and along the western edge of the NCIS. Reported ages on this have varied, but include ice-sheet build-up from 35.2 ka to 32.4 ka (Graham *et al.* 2010) and coalescence in the time windows of

30–23 ka (Sejrup *et al.* 2016), ~27–19 ka (Hughes *et al.* 2016), and ~23–22 ka (Patton *et al.* 2016, 2017). Coalescence is supported by the production of MSGL recording the development of a suture zone between the FIS and BIIS with ice flow towards the northwest.

How, when and where did the BIIS and FIS uncouple?

Sejrup *et al.* (2016) proposed regional decoupling of the BIIS and FIS together with rapid flow instability between 19 and 17.5 ka, which they associated with the retreat of the NCIS and the catastrophic northerly drainage of the Dogger Lake complex. This was further constrained by Hjelstuen *et al.* (2018), who identified a distinct freshwater meltwater spike dated to 18.7 ka. However, Roberts *et al.* (2019) have questioned the veracity of these age estimates, as there are several deglacial radiocarbon ages derived from glacimarine sediments from the central North Sea that suggest marine inundation of this region by 19.9–19.5 ka (Table 1). The first evidence of the development of a BIIS ice-marginal feature in relation to decoupling is found at core sites 112VC–114VC, where a drowned ice-contact fan occurs on the seafloor and is associated with a glacial lake lying to the east of the Great Fisher Bank. This feature is not presently dated but is constrained by a minimum peat radiocarbon date of 13.4 ± 0.1 cal ka BP (SUERC-72877; Table 1).

Did marine inundation of the northern North Sea destabilise the BIIS and thereby trigger rapid east-to-west retreat?

The retreating BIIS margin is demonstrably marine tidewater in nature at and after the attainment of Limit 5 (18.4–17.3 ka; Bayesian model). Therefore, this question can be addressed only in relation to the later stages of retreat in both the central and northern North Sea. It should be noted that prior to this, the BIIS margin in the North Sea was exclusively terrestrially based and/or in contact with glacial lakes. The trigger for inundation may have been the combined effects of the NCIS retreating eastwards and the opening of the Dogger Lake spillway in the time window of 19.9–17.5 (see above; cf. Sejrup *et al.* 2016; Roberts *et al.* 2019).

In the southern sector of the North Sea, as the ice retreated north of the Yorkshire coast and Dogger Bank, there is the first clear evidence for glacimarine sediment draping tills as well as morainal banks and grounding zone wedges. These all clearly relate to the transition from terrestrial glaciation to fully marine ice margins at which the construction of grounding zone wedges in particular indicates re-equilibration in deepening marine water. The earliest dates on these landform-sediment assemblages come from cores 132VC and 137VC (19.8–19.6 ka). Further dating constraint is provided by the OSL ages of 19.9–19.1 ka from glacifluvial sediments between tills at Seaham. Recession of this ice margin continued into the Firth of Forth from Limits 5 to 6, also characterised by the continued deposition of glacimarine sediments over a mixed-bed subglacial imprint (Roberts *et al.* 2019). More importantly, only one further grounding zone wedge is reported off the coast of Berwick and with a deglacial date of 17.0 ± 0.2 cal ka BP (core 128VC), inferring rapid retreat in contact with marine water. Final deglaciation of the inner Firth of Forth under glacimarine conditions is recorded by the Errol Beds of the Tay and Forth estuaries, which indicate ice withdrawal in this area by 16.9–16.0 ka.

Further evidence of rapid recession in contact with marine conditions occurs in the Fladen and Witch Ground basins. For example, in cores BGS BH 04/01 and BGS 77/2, proximal glacimarine sediments are dated to 19.5–16.8 ka. These dates bracket the Fladen 1 Moraine, demonstrating uncoupling from the FIS, clear ice margin recession of the BIIS, and marine inundation. These ages overlap with the age of the DLUE at 18.7 ka. Further ice recession between Limits 5 and 6 is marked by the Fladen 2 Moraine, which is dated at approximately 16.2 ka. Ice retreat under marine conditions is further reinforced by evidence of ice-marginal recession through the Wee Bankie Moraine complex and into the Firth of Forth, with the ice-contact deltas on the seafloor marking ice retreat through deep water between 19 and 17 ka (Fig. 8). Continued recession from Fladen 2 to the Bosies Bank/Halibut Bank moraines is constrained by a cluster of dates around ~15 ka. These dates suggest that the Moray Firth Ice Stream may have occupied the inner continental shelf for slightly longer than ice in the Firth of Forth. Further west, as the Moray ice moved onshore, glacimarine sediments and De Geer moraines indicate the continued instability of the ice margin under tidewater conditions, although the timing of this is not well constrained (15–13 ka; see details above).

What are the major BIIS controls on North Sea ice flow dynamics and deglaciation patterns?

Firth of Forth ice was the dominant source for the NSL even during deglaciation. Smaller ice streams such as the Stainmore, Tyne Gap and Tweed have no significant imprint offshore in the North Sea and had uncoupled from the NSL by ~18 ka (Livingstone *et al.* 2015; Davies *et al.* 2019). Ice-marginal landforms running sub-parallel and inboard of the Northumberland coast demonstrate uncoupling of the western flank/suture zone of the NSL from inland ice streams (Teasdale, 2013), while the NSL retreated northward offshore and back into the Firth of Forth (Roberts *et al.* 2019). The deglaciation of the NSL was heavily influenced by the transition from terrestrial to marine conditions with a dynamic, quasi-unstable terrestrial margin, initially producing overprinted till sheets (Dove *et al.* 2017), before grounding zone wedges developed in response to further instability driven by marine inundation (Roberts *et al.* 2019). The possible 'surge-like' imprint of the NSL has been noted by many authors and this may be related to marginal instability. It is also possible that changes in the topographic configuration of the southern North Sea Basin in relation to the glacitectonism of the Dogger Bank, accentuated the dynamic behaviour of the NSL after ~23 ka (Roberts *et al.* 2018).

In the northern North Sea Basin, the Moray Firth drainage basin was extremely influential in determining ice retreat from the Fladen/Witch Ground areas westwards through the Bosies and Halibut Bank regions from 18 to 15 ka. However, the influence of ice sourced from the Orkneys and Shetland during early deglaciation of the northern North Sea requires further investigation. Bradwell *et al.* (2019) have dated initial retreat north of the Fladen/Witch Ground areas to ~20–21 ka. This is problematic if the DLUE occured at ~19 ka and is earlier than the deglacial dates discussed here for the Fladen 1 and 2 moraines (Limits 5 and 6). However, mapped ice margin configurations clearly show the influence of independent ice caps over Orkney and Shetland during the latter stages of the downwasting of the BIIS. Hence, the ice-sheet history of the central and southern North Sea was strongly dominated by the Firth of the Forth and evolution of the NSL from 31 to 21 ka. Between 21 and 15 ka the southern and central North Sea deglaciated, first under terrestrial conditions but the basin transitioned to marine conditions at ~19 ka. This triggered late stage instability and dynamic flow reorganisation, with Firth of

Forth ice dominating the central North Sea while Moray ice evacuated the Fladen/Witch Ground areas and retreated west; Shetland/Orkney ice deglaciated northward back to local ice centres fringing the northern North Sea.

Conclusions

The outcomes of the BRITICE-CHRONO Transect 2 through the North Sea Basin and onto the onshore areas of eastern England have facilitated a significantly refined palaeoglaciological reconstruction of the eastern sector of the BIIS and its relationship with the FIS during MIS 2, in a region where ice dynamics were strongly influenced by the transition from terrestrial to marine conditions (Fig. 6). The glacial landsystem imprints of this region comprise substantial arcuate moraine sequences and associated glacial lake plains representative of the interaction and onshore flow of lobate ice streams, the most significant being the NSL, which flowed some 400 km from southern Scotland down the eastern margin of the North Sea Basin to north Norfolk. The interdigitation of glacigenic (till and morainic deposits) and stratified sediments, in both onshore stratigraphic outcrops and offshore cores, provided numerous suitable targets for OSL and radiocarbon dating from which a chronology of ice-marginal recession and oscillation has been constructed. This glacial land-systems record and its geochronological control have constrained ice-sheet fluctuation in the time window ~31–16 ka and identified seven distinctive former ice-sheet limits (L1–L7), the latter defined by Bayesian statistical analysis which verified the use or exclusion of some dates where they represented obvious outliers (Fig. 7).

Limit 1 is the most extensive MIS 2 margin and is demarcated offshore by the Bolders Bank Formation south of the Dogger Bank. It relates to the advance of the NSL and dates to some time prior to 31–25 ka. Retreat from Limit 1 is dated at 26–29 ka. The advance and retreat phase boundaries for Limit 1 from the Bayesian model are dated to 25.8 ka and 24.6 ka, respectively. Limit 2 represents NSL oscillation at ~26–24 ka and retreat at ~22 ka. The advance and retreat phase boundaries for Limit 2 from the Bayesian model are dated 23.5 ka and 22.2 ka, respectively. Limit 3 records NSL advance dated to ~21.5 ka, including the construction of the Garett Hill Moraine in north Norfolk, and initial retreat, with the advance and retreat phase boundaries from the Bayesian model being calculated at 21.5 ka and 20.8 ka, respectively. Limits 4 to 5 relate to NSL oscillation and retreat, incorporating the long-established Withernsea Till emplacement event, and date to ~20–17.2 ka. The advance and retreat phase boundaries for Limit 4 from the Bayesian model are dated to 19.7 ka and 19.5 ka, respectively, and for Limit 5 are dated to 18.4 ka and 17.3 ka, respectively. Limit 6 relates to NSL retreat dating to ~19–16.5 ka and includes the construction of the W2 grounding zone wedge offshore of Berwick. The retreat phase boundary for Limit 6 from the Bayesian model is dated to 15.8 ka. Also related to Limits 5 and 6 is the deglaciation of the central North Sea which dates to 18–15.2 ka. This includes the uncoupling of the BIIS and FIS and grounding line recession under glacimarine conditions, as represented by moraines Fladen 1 and 2 (Limits 5 and 6, respectively). The advance and retreat phase boundaries for Limit 5 (Fladen 1) are dated to 18.0–17.4 ka, using the Bayesian model, while Limit 6 (Fladen 2) is dated to 16.3–15.2 ka. Finally, Limit 7 (~15.0 ka) represents ice retreat from Bosies Bank into the Moray Firth from the central North Sea. Radiocarbon ages from deglacial glacimarine deposits range from 19.6 to 16.4 ka but the majority of these are outliers. Final ice retreat into the Moray Firth is constrained by the Bayesian model to 15.0–13.5 ka.

The deglaciation of the North Sea region was heavily influenced by the transition from terrestrial to marine conditions with a dynamic, quasi-unstable terrestrial margin, initially producing overprinted till sheets, before grounding zone wedges developed in response to further instability driven by the later marine inundation. Previous proposals that the NSL was characterised by repeat surging behaviour is compatible with the marginal instability highlighted by the palaeoglaciological reconstructions presented in this study, but changes in the topographic configuration of the southern North Sea Basin in relation to the glacitectonism of the Dogger Bank may also have accentuated the dynamic behaviour of the NSL after ~23 ka.

Supporting information

Additional supporting information may be found in the online version of this article at the publisher's web-site.

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Data availability statement

All relevant data will be made available in the forthcoming BRITICE-CHRONO online data repository, or upon reasonable request from the lead author.

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