JOURNAL OF QUATERNARY SCIENCE (2018) 33(7) 763–776 ISSN 0267-8179. DOI: 10.1002/jqs.3058

JQS Journal of Quaternary Science QRA Quaternary Research Association

Deglaciation of coastal south-western Spitsbergen dated with *in situ* cosmogenic ^{10}Be and ^{14}C measurements

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Received 15 December 2017; Revised 23 May 2018; Accepted 11 June 2018

ABSTRACT: The Svalbard–Barents ice sheet was predominantly a marine-based ice sheet and reconstructing the timing and rate of its decay during the last deglaciation informs predictions of future decay of marine-based ice sheets (e.g. West Antarctica). Records of ice-sheet change are routinely built with cosmogenic surface exposure ages, but in some regions, this method is complicated by the presence of isotopic inheritance yielding artificially old and erroneous exposure ages for the most recent deglaciation. We present 46 ^{10}Be ages from south-western Spitsbergen that, when paired with *in situ* ^{14}C measurements ($n = 5$), constrain the timing of coastal deglaciation following the last glacial maximum. ^{10}Be and *in situ* ^{14}C measurements from bedrock along a ~400-m elevation transect reveal inheritance-skewed ^{10}Be ages, whereas *in situ* ^{14}C measurements constrain 400 m of ice-sheet thinning and coastal deglaciation at 17.4 ± 1.5 ka. Our *in situ* ^{14}C-dated transect, combined with three additional ^{10}Be-dated coastal sites, show that the south-western margin of the Svalbard–Barents ice sheet retreated out of the Norwegian Sea between ~18 and 16 ka. *In situ* ^{14}C measurements provide key chronological information on ice-sheet response to the last termination in cases where measurements of long-lived nuclides are compromised by isotopic inheritance. © 2018 John Wiley & Sons, Ltd.

KEYWORDS: cosmogenic nuclides; ice sheets; *in situ* ^{14}C; Quaternary; Svalbard.

Introduction

Geological records that constrain the timing and magnitude of ice-sheet demise during the last deglaciation provide important insights into the response of ice sheets to a warming climate. At its maximum extent during the last glacial cycle, the Svalbard–Barents ice sheet (SBIS) was part of the broader Eurasian ice sheet complex with a sea-level equivalent of ~24 m (Hughes *et al.*, 2016). Resting at the north-western limit of the SBIS, the Svalbard Archipelago is one of the few terrestrial locations within the primarily marine-based SBIS footprint. Accordingly, much of our current understanding of how the SBIS evolved through the last glacial cycle is based on archives of ice-sheet change present on Svalbard (Fig. 1; Landvik *et al.*, 1998, 2014; Hormes *et al.*, 2013; Ingólfsson and Landvik, 2013; Eccleshall *et al.*, 2016). Gauging how the SBIS decayed at the end of the last glaciation can help identify mechanisms of global climate change and inform ice sheet models used to explore the sensitivity of marine-based ice sheets to various climatic and glaciological parameters (Stokes *et al.*, 2015; Patton *et al.*, 2016).

Cosmogenic nuclide measurements are often used to develop detailed chronologies of ice sheet and glacier change (e.g. Balco, 2011; Granger *et al.*, 2013; Ivy-Ochs and Briner, 2014). On Svalbard, there have been several efforts to reconstruct SBIS behavior during the last glacial cycle using cosmogenic nuclides with mixed success. One limitation is that much of Svalbard does not host the quartz-bearing rocks required for ^{10}Be measurements, and thus the geographical scope of cosmogenic nuclide-based measurements is relatively restricted. Nonetheless, the first ^{10}Be ages from Svalbard were used to propose that ice-free regions in north-western Svalbard existed during the last glacial maximum (Landvik *et al.*, 2003). However, while ^{10}Be ages older than ~75 ka defined the maximum SBIS thickness in NW Svalbard, these old ^{10}Be ages do not preclude the presence of widespread and systematic isotopic inheritance on the landscape. Using the same approach, several ^{10}Be ages from western Svalbard and Nordaustlandet help constrain the dimensions of the SBIS during the last glacial cycle (Hormes *et al.*, 2011, 2013; Gjermundsen *et al.*, 2013; Landvik *et al.*, 2013; Fig 1). Most recently, ^{10}Be and ^{26}Al measurements from high-elevation bedrock suggest that Svalbard's alpine landscape has survived repeated glaciations through the Quaternary, suggestive of a minimally erosive ice sheet (Gjermundsen *et al.*, 2015). Although these studies place broad constraints on SBIS behavior through the last glacial cycle and longer, they also suffer from somewhat geographically and chronologically scattered ^{10}Be ages, making it difficult to develop detailed millennial-scale chronologies of ice-sheet change. Collectively, these studies highlight a landscape that is challenging for developing cosmogenic-nuclide based chronologies of ice-sheet change, probably

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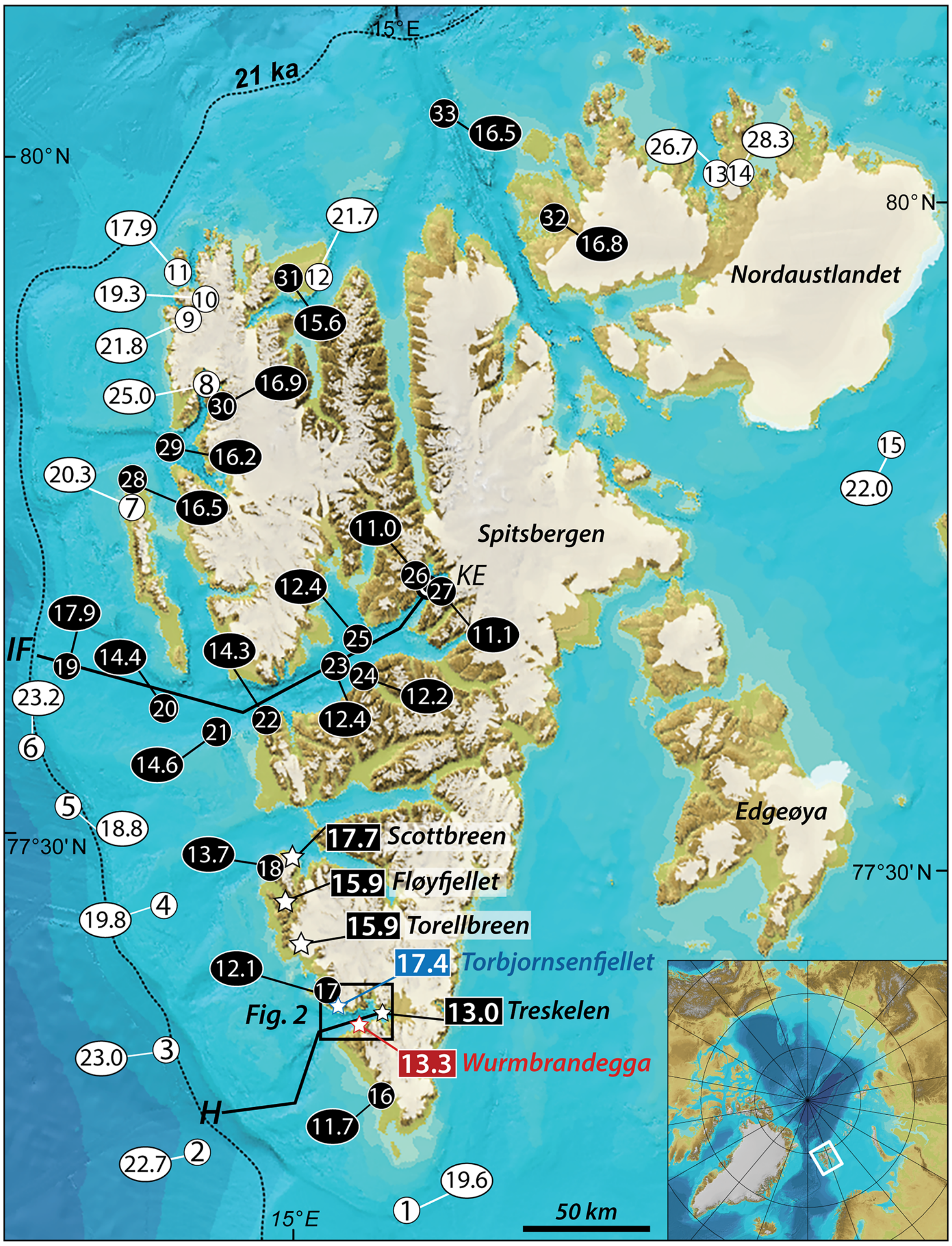

Figure 1. Spitsbergen with locations referred to in the text and the 21 ka ice limit from Hughes et al. (2016; dashed line). KE – Kapp Ekholm, H – Hornsund, and IF – Isfjorden. Previously published radiocarbon (cal ka BP) and ^{10}Be ages across Spitsbergen from previous studies constraining ice margin position are presented in two groups: (i) ages that mark the initial timing of retreat or thinning of the ice margin from the Last Glacial Maximum (LGM) extent (white ovals with black text), and (ii) oldest coastal ages that constrain deglaciation following the LGM maximum extent and deglaciation of the present coastline (black ovals/white text). Location numbers are linked to Table 4; see Table 4 for sample details and references. Locations and ^{10}Be ages marked with stars and boxes are from this study. Isfjorden and Hornsund transects from Fig. 4 are marked with black lines. Base and inset maps are from Jakobsson *et al.* (2012). [Colour figure can be viewed at wileyonlinelibrary.com]

due to the widespread presence of non-erosive cold-based ice and its variable imprint on the landscape (Landvik *et al.*, 2014).

Surface exposure dating in glacial landscapes relies on the assumption that cosmogenic nuclides that accumulated on the landscape before the most recent episode of exposure have been removed by ~2–3 m of subglacial erosion during the latest interval of ice cover. In settings dominated by warm-based and erosive ice, this assumption is typically valid, but at high-latitude locations, minimally erosive polythermal and cold-based ice can result in cosmogenic nuclide datasets that are influenced by isotopic inheritance (e.g. Håkansson *et al.*, 2008; Balco *et al.*, 2014; Young *et al.*, 2016). Isotopic inheritance occurs when ice is unable to erode through the ~2–3 m of rock required to reset the cosmogenic clock between periods of surface exposure and the resulting nuclide concentration is an aggregate of two or more distinct periods of exposure. Inheritance is also possible within landscapes where despite 2–3 m of erosion during the latest interval of glaciation, deep subsurface nuclide accumulation in periods of prolonged surface exposure between glaciations results in excess nuclide inventories that pre-date the most recent period of ice cover (Briner *et al.*, 2016).

Whereas long-lived or stable nuclides such as ^{10}Be ($t_{1/2} = 1.387$ Ma; Chmeleff *et al.*, 2010) must be removed from the landscape via sufficient subglacial erosion, *in situ* ^{14}C is unique because its relatively short half-life ($t_{1/2} = 5730$ years) allows for previously accumulated *in situ* ^{14}C to decay away to undetectable levels after ~30 ka of simple burial of a surface by ice without the aid of subglacial erosion. *In situ* ^{14}C measurements are perhaps most powerful when paired with ^{10}Be to resolve complex exposure–burial histories (e.g. Goehring *et al.*, 2011), but *in situ* ^{14}C measurements are also particularly attractive in environments characterized by minimally erosive ice that is not capable of resetting the cosmogenic clock between periods of exposure. Measuring several nuclides in conjunction yields a more complete quantitative understanding of ice-sheet fluctuations over multiple time-scales, but *in situ* ^{14}C measurements are also well-suited to constrain the timing of the last deglaciation in settings where long-lived nuclides such as ^{10}Be run a much higher risk of carrying inheritance from prior exposure (e.g.

Briner *et al.*, 2014; Johnson *et al.*, 2017). Despite its potential, *in situ* ^{14}C is rarely used because of the difficulty of extracting ^{14}C from the mineral quartz in geological samples (e.g. Lifton *et al.*, 2001; Balco *et al.*, 2016).

We present 46 ^{10}Be ages, five *in situ* ^{14}C ages and five ^{26}Al ages from four sites in south-western Spitsbergen to constrain the timing of coastal deglaciation following the last glacial maximum. A component of our ^{10}Be ages are influenced by isotopic inheritance, but our population of ^{10}Be ages is sufficiently large to constrain the timing of coastal deglaciation in south-western Spitsbergen. Combined *in situ* ^{10}Be, ^{14}C and ^{26}Al in five bedrock samples along a ~400-m elevation transect reveal that ^{10}Be and ^{26}Al concentrations yield an ambiguous timing of deglaciation and a complex long-term exposure history but corresponding *in situ* ^{14}C measurements robustly constrain millennial-scale ice-sheet thinning.

Setting and methods

Our study area is Hornsund (76.97°N, 15.70°E), located in south-western Spitsbergen (Figs 1 and 2). Several independent glaciers feed into the primary Hornsund channel and, at present, ~800 km^2 of Hornsund's ~1200-km^2 drainage basin is glaciated with glacier retreat during the observational record averaging ~70 m a^{-1} (Błaszczyk *et al.*, 2013). At the head of Hornsund, a tidewater glacier is currently located ~13 km east of its late Holocene maximum extent, which is marked by a prominent moraine that was emplaced at Treskelen just before 1.9 ± 0.3 ka based on recent ^{10}Be ages ($n = 4$; Philipps *et al.*, 2017). During the last glacial cycle, Hornsund hosted an SBIS outlet glacier that was part of the western margin of the SBIS. The western SBIS is thought to have advanced out to the continental shelf three times during the last glacial cycle during Marine Isotope Stage (MIS) 5d, MIS 5b, and MIS 2 with retreat from the outer shelf underway as early as ~23–20 ka (Mangerud *et al.*, 1998; Jessen *et al.*, 2010; Hormes *et al.*, 2013; Eccleshall *et al.*, 2016). A single minimum limiting radiocarbon age from near Hornsund indicates that by ~12.1 cal ka BP ice was less extensive than it is today (Birkenmajer and Olsson, 1998; Hormes *et al.*, 2013). The long-term pattern of SBIS advance and retreat in western Svalbard is largely based on the Kapp Ekholm sediment section located at the inner reaches of Isfjorden,

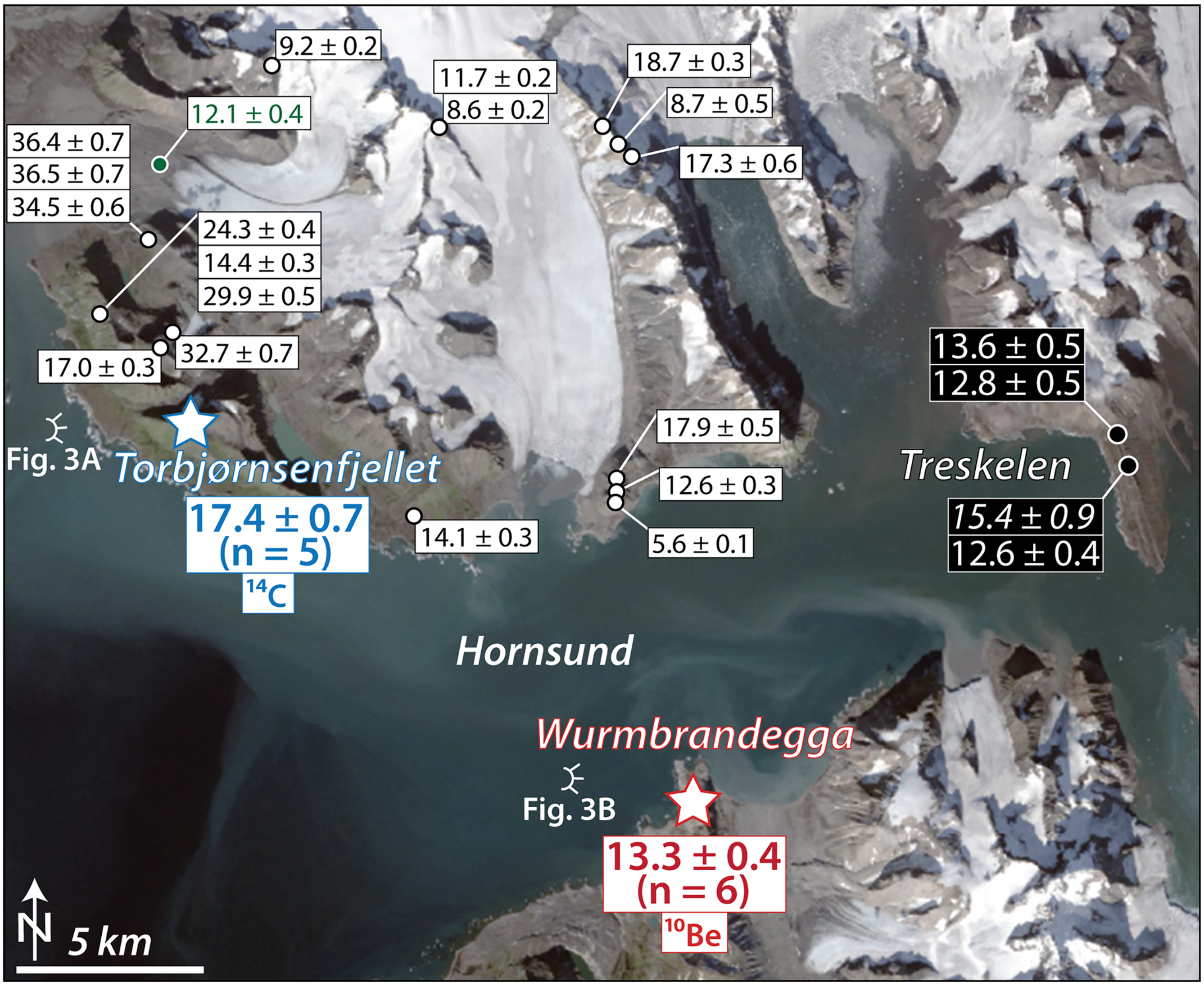

Figure 2. Ice margin constraints at Hornsund (Norwegian Polar Institute; toposvalbard.npolar.no/; collected in 2011). Individual ^{10}Be ages from bedrock are shown in black text with white boxes; up-fjord erratics on the Treskelen Peninsula are displayed in white text with black boxes. Only the mean *in situ* ^{14}C and ^{10}Be age from the Torbjørnsenfjellet ($n = 5$) and Wurmbrandegga ($n = 6$; one outlier removed) elevation transects are shown (Fig. 3). The minimum-limiting radiocarbon age just north of Hornsund is in green (U-2972; 12 100 ± 320 cal ka BP; Birkenmajer and Olsson, 1998; Table 4). [Colour figure can be viewed at wileyonlinelibrary.com]

which displays alternating units of glacial till and marine sediments (Fig. 1). Because Kapp Ekholm is situated only ~14 km from modern ice, the marine sediment units mark intervals when Svalbard glaciers were probably not much larger than today (Mangerud *et al.*, 1998; Eccleshall *et al.*, 2016).

We collected 46 samples for ^{10}Be dating, five samples for ^{26}Al measurement and five samples for *in situ* ^{14}C measurements along the south-western coast of Spitsbergen. Thirty-four ^{10}Be samples are from the Hornsund region and are divided into four distinct groups (Figs 1 and 2): (i) a series of nunatak and bedrock ridges ($n=18$), (ii) an elevation transect at Torbjørnsenfjellet on the north side of Hornsund (bedrock; $n=5$; Fig. 3A), (iii) an elevation transect at Wurmbrandegga on the south side of Hornsund (bedrock; $n=7$; Fig. 3B), and (iv) boulders perched on a bedrock ridge adjacent to, but beyond, the late Holocene ice extent at Treskelen ($n=4$). We measured *in situ* ^{14}C and ^{26}Al in each bedrock sample from the Torbjørnsenfjellet elevation transect. We also collected four bedrock samples from surfaces immediately outboard of the late Holocene terminal moraine (<150 m) at Scottbreen (77.54°N, 14.36°E) located ~70 km north-west of Hornsund (Fig. 1). Lastly, we collected four ridgeline bedrock samples from Fløyfjellet (77.41°N, 14.09°E) located ~60 km north-west of Hornsund and four bedrock samples from summits near Torellbreen (77.31°N, 14.09°E) located between Fløyfjellet and Hornsund (Fig. 1).

Samples were collected in 2013 and 2014 with a hammer and chisel, and a Trimble GeoXT and Tempest antenna GPS receiver with a vertical uncertainty of ± 0.5 m was used to record sample location and elevation. A handheld clinometer was used to measure topographic shielding by the surrounding topography. ^{10}Be samples were processed at the Buffalo Cosmogenic Laboratory ($n=29$), Lamont-Doherty Earth Observatory (LDEO) Cosmogenic Nuclide Laboratory ($n=13$; $n=5$ ^{26}Al samples), and the Scottish Universities Environmental Research Centre (SUERC; $n=4$) following standard extraction methods for ^{10}Be and ^{26}Al (Schaefer *et al.*, 2009). *In situ* ^{14}C samples were processed at LDEO and measured ^{14}C concentrations are blank-corrected using a long-term laboratory blank (Goehring *et al.*, 2014; see Tables 2 and 3). Accelerator mass spectrometric measurements for LDEO and Buffalo samples were made at Lawrence Livermore National Laboratory – Center for Accelerator Mass Spectrometry, and the remaining samples were measured at SUERC (Table 1). Surface exposure ages were calculated using the Arctic (^{10}Be and ^{26}Al) and western Greenland (^{14}C) production rate calibration datasets (Young *et al.*, 2013, 2014) and 'Lm' scaling (Lal, 1991; Stone, 2000). Ages are calculated using version 3 of the exposure age calculator found at https://hess.ess.washington.edu/, that implements an updated treatment of muon-based nuclide production (Balco *et al.*, 2008; Balco, 2017). We do not correct

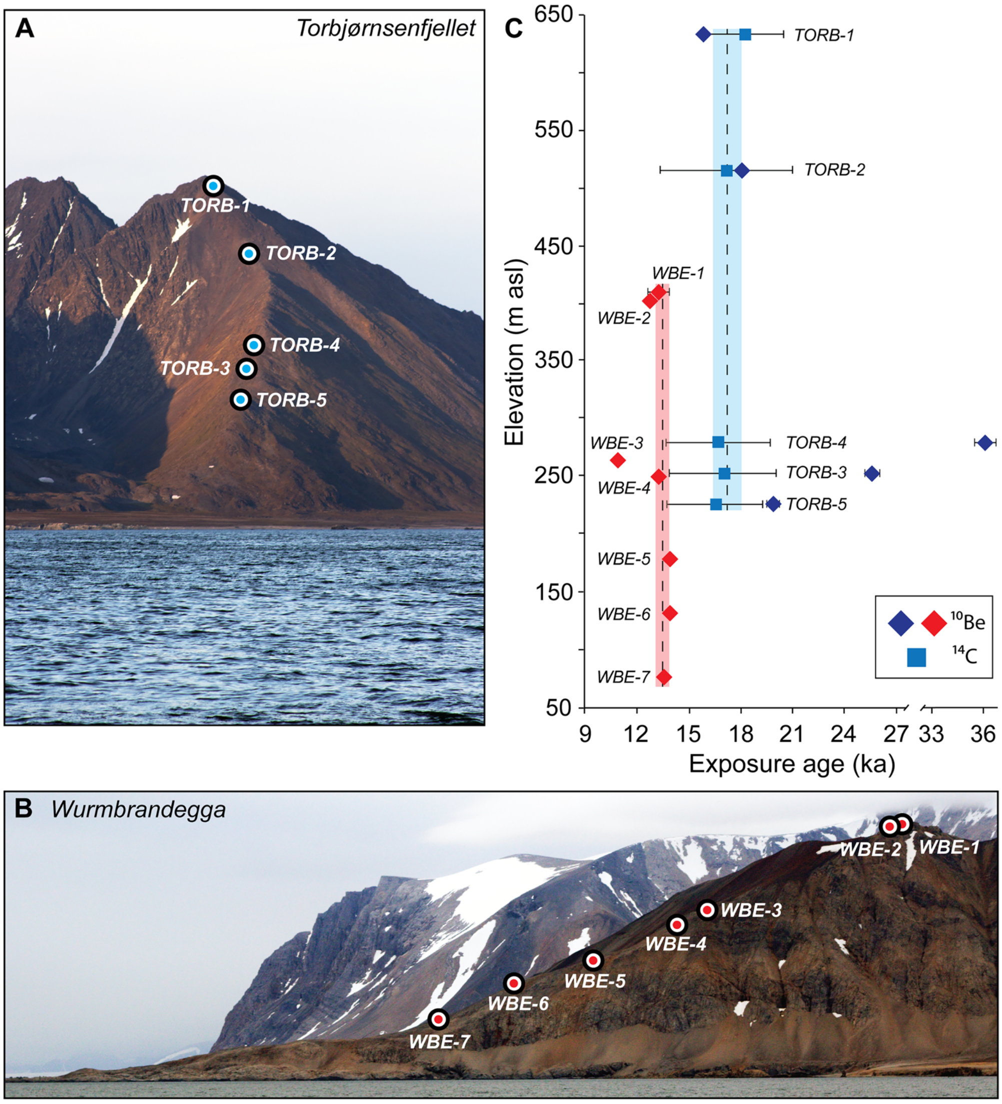

Figure 3. Elevation transects in the Hornsund region. (A) Field photograph of Torbjørnsenfjellet with sample locations. (B) Field photograph of Wurmbrandegga with sample locations. (C) Comparison of paired ^{10}Be and *in situ* ^{14}C ages at Torbjørnsenfjellet, and the ^{10}Be-dated Wurmbrandegga elevation transect located ~12 km up-fjord from Torbjørnsenfjellet. [Colour figure can be viewed at wileyonlinelibrary.com]

Table 1. Sample information and ^{10}Be and ^{26}Al data.

Sample	Sample type	Latitude (DD)	Longitude (DD)	Elevation (m asl)	Thickness (cm)	Shielding correction	Quartz (g)	^{9}Be carrier added (g)	Carrier conc. (p.p.m.)	^{10}Be/^{9}Be ratio ± 1σ (10^{-13})*	^{26}Al/^{27}Al ratio ± 1σ (10^{-13})	^{10}Be conc. (atoms g^{-1})†	^{10}Be uncertainty (atoms g^{-1})	^{10}Be age – Lm (ka)	^{26}Al conc. (atoms g^{-1})	^{26}Al uncertainty (atoms g^{-1})	^{26}Al age – Lm (ka)	^{26}Al /^{10}Be	Lab.
Scottbreen																			
SCO-14-12	Bedrock	77.5538	14.4423	142	1.0	0.9955	39.9946	0.6092	372.5	2.4648 ± 0.0638		92 616	2421	18.8 ± 0.5					Buffalo
SCO-14-13	Bedrock	77.5557	14.4447	119	2.0	0.9973	30.4387	0.6082	372.5	1.7251 ± 0.0325		85 543	1629	17.9 ± 0.3					Buffalo
SCO-14-14	Bedrock	77.5563	14.4438	104	1.0	0.9982	35.0566	0.6072	372.5	1.8143 ± 0.0343		77 270	1488	16.3 ± 0.3					Buffalo
SCO-14-15	Bedrock	77.5563	14.4436	115	2.0	0.9982	20.1923	0.6106	372.5	1.1563 ± 0.0228		84 618	1738	17.8 ± 0.4					Buffalo
Fløyfjellet																			
FLO-14-01	Bedrock	77.4124	14.0924	265	2.0	0.9968	15.0139	0.1818	1037	1.0766 ± 0.0174		90 104	1962	16.2 ± 0.4					LDEO
FLO-14-02	Bedrock	77.4123	14.0880	252	2.0	0.9996	12.0044	0.1816	1037	0.8357 ± 0.0157		87 311	2063	15.9 ± 0.4					LDEO
FLO-14-03	Bedrock	77.4133	14.0883	234	2.0	0.9988	14.9806	0.1815	1037	0.9870 ± 0.0227		82 634	2392	15.3 ± 0.4					LDEO
FLO-14-04	Bedrock	77.4141	14.0867	229	2.0	0.9985	15.2578	0.1818	1037	1.0567 ± 0.0198		87 026	2053	16.2 ± 0.4					LDEO
Torellbreen																			
55-PLO-1	Bedrock	77.2604	15.1582	812	5.0	0.9469	15.1596	0.6073	372.5	1.5003 ± 0.0420		149 906	4185	17.3 ± 0.5					Buffalo
70-ORV-1	Bedrock	77.3097	14.6887	637	2.0	0.9774	13.4797	0.6071	372.5	1.6248 ± 0.0322		179 656	3614	*23.0 ± 0.5*					Buffalo
70-ORV-2	Bedrock	77.3097	14.6887	637	2.0	0.9998	20.0332	0.6064	372.5	1.7188 ± 0.0320		127 826	2408	16.0 ± 0.3					Buffalo
70-ORV-3	Bedrock	77.3099	14.6775	587	2.0	0.9992	20.0732	0.6054	372.5	1.4715 ± 0.0273		108 796	2049	14.3 ± 0.3					Buffalo
Hornsund nunataks/bedrock																			
70-NORD-1	Bedrock	77.0939	15.6894	751	2.0	0.9999	20.2885	0.6065	372.5	2.2582 ± 0.0367		166 380	2734	18.7 ± 0.3					Buffalo
55-VAR-1	Bedrock	77.0890	15.7062	698	5.0	0.9999	5.1862	0.6073	372.5	0.2472 ± 0.0144		72 048	4209	8.7 ± 0.5					Buffalo
55-VAR-2	Bedrock	77.0885	15.7082	680	4.5	0.9967	13.4856	0.6076	372.5	1.2506 ± 0.0449		140 255	5036	17.3 ± 0.6					Buffalo
66-SLY-1	Bedrock	77.0939	15.5151	656	2.32	1.0000	26.0395	0.6122	372.5	1.6440 ± 0.0290		95 058	1855	11.7 ± 0.2					Buffalo
65-SLY-2	Bedrock	77.0939	15.5152	657	1.47	1.0000	19.6358	0.6106	372.5	0.9339 ± 0.0177		70 586	1395	8.6 ± 0.2					Buffalo
70-BARA-1	Bedrock	77.1086	15.3442	868	2.0	1.0000	27.8512	0.6112	372.5	1.7000 ± 0.0301		90 701	1689	9.2 ± 0.2					Buffalo
68-BRATT-1	Bedrock	77.0674	15.2118	110	2.43	1.0000	35.7727	0.6066	372.5	4.0805 ± 0.0793		170 651	3361	36.4 ± 0.7					Buffalo
66-BRATT-2	Bedrock	77.0674	15.2114	110	0.94	1.0000	40.0399	0.6127	372.5	4.5705 ± 0.0881		173 339	3347	36.5 ± 0.7					Buffalo
66-BRATT-3	Bedrock	77.0674	15.2115	109	1.26	1.0000	29.0404	0.6116	372.5	3.1365 ± 0.0559		163 393	3007	34.5 ± 0.6					Buffalo
66-GUL-1	Bedrock	77.0522	15.1845	548	1.27	1.0000	27.0770	0.6123	372.5	3.2105 ± 0.0546		179 608	3156	24.3 ± 0.4					Buffalo
66-GUL-2	Bedrock	77.0535	15.1838	580	1.78	1.0000	40.1016	0.6104	372.5	2.9056 ± 0.0522		109 342	2034	14.4 ± 0.3					Buffalo
67-GUL-3	Bedrock	77.0515	15.1852	526	1.47	1.0000	23.4779	0.6092	372.5	3.3705 ± 0.0541		215 620	3495	29.9 ± 0.5					Buffalo
67-Jahn-1	Bedrock	77.0440	15.2406	600	2.0	1.0000	30.1245	0.6104	372.5	2.6393 ± 0.0449		131 504	2262	17.0 ± 0.3					Buffalo
68-JAHN-2	Bedrock	77.0461	15.2449	510	2.0	1.0000	12.4195	0.6068	372.5	1.9395 ± 0.0378		231 315	4600	32.7 ± 0.7					Buffalo
FAN-14-01	Bedrock	77.0103	15.7024	388	3.0	1.0000	10.8458	0.1804	1037	0.6877 ± 0.0131		78 935	1865	12.6 ± 0.3					LDEO
FAN-14-02	Bedrock	77.0145	15.7017	397	3.0	0.9970	15.1240	0.1820	1037	1.3551 ± 0.0272		112 784	2976	17.9 ± 0.5					LDEO
FAN-14-03	Bedrock	77.0097	15.7015	380	3.0	0.9990	14.6232	0.1814	1037	0.4088 ± 0.0095		34 889	898	5.6 ± 0.1					LDEO
ARD-01	Bedrock	77.0068	15.4910	61	2.89	0.9965	20.3730	0.1829	1037	1.0092 ± 0.0193		62 610	1479	14.1 ± 0.3					LDEO
Torbjørsenfjellet																			
14TORB-1	Bedrock	77.0265	15.2679	633	1.87	0.9773	17.5809	0.1824	1024	1.7664 ± 0.0285	4.3881 ± 0.3316	124 966	2031	16.0 ± 0.3	921 682	79 016	17.6 ± 1.5	7.38 ± 0.64	LDEO
14TORB-2	Bedrock	77.0263	15.2601	515	3.04	0.9781	16.7093	0.1819	1024	1.7027 ± 0.0276	3.8344 ± 0.1548	126 381	2068	18.3 ± 0.3	910 442	46 629	19.6 ± 1.0	7.20 ± 0.39	LDEO
14TORB-3	Bedrock	77.0271	15.2312	252	2.45	0.9951	31.2907	0.1828	1024	3.5131 ± 0.0538	7.9293 ± 0.2429	140 213	2153	25.8 ± 0.4	866 472	31 816	23.7 ± 0.9	6.18 ± 0.25	LDEO
14TORB-4	Bedrock	77.0273	15.2343	279	1.53	0.9910	14.8254	0.1819	1024	2.4246 ± 0.0394	4.1403 ± 0.1666	203 065	3313	36.3 ± 0.6	1 339 280	65 019	35.8 ± 1.8	6.60 ± 0.34	LDEO
14TORB-5	Bedrock	77.0276	15.2281	225	2.09	0.9936	36.0443	0.1821	1024	3.0806 ± 0.0516	6.3974 ± 0.3634	106 472	1789	20.1 ± 0.3	614 729	39 491	17.2 ± 1.1	5.77 ± 0.38	LDEO
Wurmbrandegga																			
WBE-14-01	Bedrock	76.9349	15.7808	412	2.0	1.0000	25.7482	0.6087	372.5	0.8823 ± 0.0360		85 508	3646	13.2 ± 0.6					Buffalo
WBE-14-02	Bedrock	76.9350	15.7805	406	2.0	1.0000	25.1815	0.6075	372.5	0.8342 ± 0.0160		81 229	1732	12.7 ± 0.3					Buffalo
WBE-14-03	Bedrock	76.9377	15.7748	260	2.0	0.9880	23.7882	0.6080	372.5	1.0545 ± 0.0230		59 715	1418	10.9 ± 0.3					Buffalo
WBE-14-04	Bedrock	76.9380	15.7745	248	2.0	0.9727	30.3079	0.6063	372.5	1.2143 ± 0.0212		70 532	1348	13.2 ± 0.3					Buffalo
WBE-14-05	Bedrock	76.9392	15.7718	181	2.0	0.9883	30.6524	0.6105	372.5	1.1370 ± 0.0194		69 812	1316	13.8 ± 0.3					Buffalo
WBE-14-06	Bedrock	76.9404	15.7702	133	2.0	0.9885	15.0869	0.6043	372.5	1.3564 ± 0.0276		65 557	1419	13.7 ± 0.3					Buffalo
WBE-14-07	Bedrock	76.9414	15.7680	78	2.0	0.9884	15.0632	0.6071	372.5	1.2637 ± 0.0238		60 686	1234	13.4 ± 0.3					Buffalo

continued

Table 1. (*Continued*)

Sample	Sample type	Latitude (DD)	Longitude (DD)	Elevation (m asl)	Thickness (cm)	Shielding correction	Quartz (g)	^{9}Be carrier added (g)	Carrier conc. (p.p.m.)	^{10}Be/^{9}Be ratio ± 1σ (10^{-13})*	^{26}Al/^{27}Al ratio ± 1σ (10^{-13})	^{10}Be conc. (atoms g^{-1})†	^{10}Be uncertainty (atoms g^{-1})	^{10}Be age – Lm (ka)	^{26}Al conc. (atoms g^{-1})	^{26}Al uncertainty (atoms g^{-1})	^{26}Al age – Lm (ka)	^{26}Al/^{10}Be	Lab.
Treskelen																			
TR-01	Erratic	77.0227	16.2057	156	2.6	0.9619	20.2170	0.12670	1664	0.957 ± 0.0326		64 813	2462	13.6 ± 0.5					SUERC
TR-02	Erratic	77.0227	16.2055	156	3.4	0.9922	28.8390	0.12660	1664	1.3057 ± 0.0511		62 423	2650	12.8 ± 0.5					SUERC
TR-04	Erratic	77.0160	16.2162	111	1.9	0.9981	16.5260	0.12810	1664	0.8705 ± 0.0477		72 734	4247	15.4 ± 0.9					SUERC
TR-05	Erratic	77.0161	16.2160	110	1.4	0.9979	25.5210	0.12710	1664	1.1076 ± 0.0349		59 853	2117	12.6 ± 0.4					SUERC

*Samples processed at Buffalo and LDEO were measured at Lawrence Livermore National Laboratory – Center for Accelerator Mass Spectrometry (CAMS). Samples processed at SUERC were measured at SUERC AMS Laboratory. All ratios are reported relative to 07KNSTD with a reported ratio of 2.85×10^{-12} using a ^{10}Be half-life of 1.36×10^6 years (Nishiizumi *et al*., 2007).

†^{10}Be concentrations are blank corrected using batch-specific process blanks. LDEO samples are blank corrected using values of 8259 ± 3148 atoms (TORB samples) and 3842 ± 539 atoms ($n = 2$; FLO, FAN, ARD samples); ^{10}Be concentrations for all SUERC samples are blank corrected using a value of 37 999 ± 537 atoms; Buffalo samples are blank corrected suing values of 48599 ± 5588 atoms (SCO-13, SCO-15, GUL-3, JAHN-1 samples), 33523 ± 9022 atoms (ORV-1, ORV-2, ORV-3, NORD-1 samples), 56575 ± 24289 atoms (BRATT-1, JAHN-2 samples), 29873 ± 1944 atoms (BRATT-2, BRATT-3, GUL-1, GUL-2, SLY-1 samples), 33 430 ± 5180 atoms (SCO-12, SCO-14, SLY-2 samples), 60 094 ± 10 954 atoms (BARA-1, WBE-03, WBE-04, WBE-05, WBE-06, WBE-07 samples), and 37 093 ± 9868 atoms (WBE-01 and WBE-02 samples).

All ages are calculated using version 3 of the calculator code found at https://hess.ess.washington.edu/ (Balco *et al*., 2008; Balco, 2017), the 'Arctic' ^{10}Be production rate (Young *et al*., 2013), standard atmosphere and pressure 'std', a rock density of 2.65 g cm^{-3}, and assumes no erosion.

nuclide concentrations for snow-cover or erosion; samples are primarily from windswept locations and many sampled surfaces displayed primary glacial features. In addition, we make no correction for isostatic rebound because the effects of uplift on nuclide production are probably offset by atmospheric compression, although these effects are difficult to quantify (Staiger *et al*., 2007). Individual exposure ages are presented and discussed with 1-sigma analytical uncertainties, and when comparing our results to independently dated records of ice sheet or environmental change, the production rate uncertainty is propagated through in quadrature.

Results

Four individual ^{10}Be ages at Scottbreen are 18.8 ± 0.5, 17.9 ± 0.3, 16.3 ± 0.3 and 17.8 ± 0.4 ka (all bedrock). At Fløyfjellet, four individual ^{10}Be ages are 16.2 ± 0.4, 16.2 ± 0.4, 15.9 ± 0.4 and 15.3 ± 0.4 ka (all bedrock), and at Torellbreen, four additional ^{10}Be ages are 23.0 ± 0.5, 17.3 ± 0.5, 16.0 ± 0.3 and 14.3 ± 0.3 ka (all bedrock). At Hornsund, the coastal nunatak and ridgeline ^{10}Be ages span 5.6 ± 0.1 to 36.5 ± 0.7 ka ($n = 18$; bedrock). The Wurmbrandegga elevation transect has ^{10}Be ages ranging between 10.9 ± 0.3 and 13.8 ± 0.3 ka ($n = 7$; bedrock), and the Torbjørnsenfjellet elevation transect has ^{10}Be ages ranging from 16.0 ± 0.3 to 36.3 ± 0.6 ka ($n = 5$; bedrock). The up-fjord ^{10}Be ages from erratic boulders perched on bedrock range from 12.6 ± 0.4 to 15.4 ± 0.9 ka ($n = 4$). Lastly, ^{26}Al and *in situ* ^{14}C ages at Torbjørnsenfjellet range from 17.2 ± 1.1 to 35.8 ± 1.8 ka and 16.7 ± 2.9 to 18.5 ± 2.7 ka, respectively (Tables 1–3). Measured ^{26}Al/^{10}Be ratios range from 7.38 ± 0.64 to 5.77 ± 0.38.

Multiple nuclides constrain the timing of deglaciation and erosion regimes

Scottbreen, Fløyfjellet and Torellbreen

The mean of four ^{10}Be ages at Scottbreen is 17.7 ± 1.2 ka (production rate uncertainty included) and it is tempting to use this age constraint as the timing of local coastal deglaciation. However, these four samples are all from bedrock locations that are in close proximity to one another (~300 m distance and 38 m elevation) and are located immediately outside Scottbreen's historical maximum extent. Landscapes positioned near glacial maxima that spend a large proportion of a glacial cycle ice free may be affected by small, uniform amounts of isotopic inheritance that yield consistent, but slightly too old ^{10}Be ages (Briner *et al*., 2016). Because the Scottbreen samples are from such close proximity to each other, it is possible these ^{10}Be bedrock ages simply constrain a local ^{10}Be inventory that contains a small, and uniform, amount of isotopic inheritance. In addition, the age 17.7 ± 1.2 ka would indicate that Scottbreen retreated within its late Holocene maximum extent rather early following the last glacial maximum. Nonetheless, while small amounts of inheritance may be influencing our Scottbreen ^{10}Be ages, we tentatively use 17.7 ± 1.2 ka as the timing of local deglaciation. At Fløyfjellet, all ^{10}Be ages overlap at 1-sigma, indicating that deglaciation occurred at 15.9 ± 0.7 ka. The four ^{10}Be ages at Torellbreen are more scattered (Table 1), but the individual ^{10}Be ages of 14.3 ± 0.3, 16.0 ± 0.3 and 17.3 ± 0.5 ka (mean = 15.9 ± 1.5 ka) are consistent with the timing of deglaciation at Fløyfjellet and Scottbreen. Combined, ^{10}Be ages indicate that deglaciation at Scottbreen, Fløyfjellet and Torellbreen occurred at 17.7 ± 1.2, 15.9 ± 0.7 and 15.9 ± 1.5 ka, respectively (Fig. 1).

Table 2. *In situ* ^{14}C extraction details.

Sample	Date extracted	Quartz (g)	V_{CO_2} (cc STP)	V_{dilute} (cc STP)	CAMS no.	F_m measured	^{14}C blank-corrected (atoms g^{-1})	^{14}C age – Lm (ka)	^{14}C age – LSD (ka)
14TORB-1	3/4/15	5.1234	0.2576 ± 0.0030	1.5901 ± 0.0184	170000	0.0259 ± 0.0002	208 248 ± 8111	18.5 ± 2.7	19.1 ± 2.9
14TORB-2	26/2/16	3.4877	0.4839 ± 0.0056	1.8295 ± 0.0211	173883	0.0153 ± 0.0002	183 882 ± 12 146	17.6 ± 4.0	18.0 ± 4.3
14TORB-3	22/4/15	5.0427	0.7704 ± 0.0089	1.4295 ± 0.0165	170152	0.0214 ± 0.0002	146 890 ± 7942	17.3 ± 3.2	17.3 ± 3.2
14TORB-4	22/2/16	5.0307	0.3336 ± 0.0038	1.6875 ± 0.0194	173884	0.0186 ± 0.0001	149 829 ± 8358	16.9 ± 3.1	16.9 ± 3.1
14TORB-5	5/5/15	5.1089	0.5360 ± 0.0062	0.9921 ± 0.0115	173356	0.0298 ± 0.0003	141 823 ± 7643	16.7 ± 2.9	16.6 ± 2.9

All samples are blank corrected using an LDEO long-term value of $112.55 \pm 36.83 \times 10^3$ ^{14}C atoms ($n = 23$).
Sample weight, gas volume after carbon extraction from quartz (V_{CO_2}) and after addition of ^{14}C-free dilution gas (V_{dilute}), cc STP cubic centimeters at standard temperature and pressure, measured fraction modern (F_m the $^{14}C/^{13}C$ ratio of the sample vs. that of a standard, both corrected to d^{13}C 25‰ VPDB and to 1950 CE), blank-corrected ^{14}C concentrations with analytical uncertainties. We report ^{14}C ages using 'Lm' (Lal, 1991; Stone, 2000) and 'LSD' (Lifton et al., 2014) scaling.

The consistency between the timing of deglaciation at Scottbreen (17.7 ± 1.2 ka), Fløyfjellet (15.9 ± 0.7 ka) and Torellbreen (15.9 ± 1.5 ka), which all post-date the last glacial maximum, is suggestive of ^{10}Be ages that are accurately recording the timing of deglaciation and are not influenced by inheritance. Moreover, ^{10}Be ages at all three locations are solely from bedrock, suggesting that a warm-based SBIS was able to erode through the ~2 m of rock required to reset the cosmogenic clock along Spitsbergen's south-western coast. Ultimately, we cannot rule out that our ^{10}Be measurements from Scottbreen, Fløyfjellet and Torellbreen are systematically influenced by deep subsurface nuclide accumulation during periods of prolonged surface exposure and contain a small amount of isotopic inheritance (Briner *et al.*, 2016). However, similar ^{10}Be ages that are influenced by systematic deep subsurface nuclide accumulation would probably require near-identical exposure histories and total erosion depths during periods of ice cover across all three sites. We prefer the more likely scenario where the consistency in ^{10}Be ages at Scottbreen, Fløyfjellet and Torellbreen simply reflects the similar timing of deglaciation across these sites and the erosional efficiency of a warm-based SBIS (Fig. 1).

Hornsund

Constraining the timing of initial deglaciation at Hornsund is more challenging. Here, we consider ^{10}Be ages within the context of their morphostratigraphic position. A series of ages from coastal nunataks and ridges range from 5.6 ± 0.1 to 36.5 ± 0.7 ka, show no clear trend with elevation as would be expected with glacier thinning, and adjacent samples from similar elevations often have drastically different ages (Fig. 2; black text/white boxes; Table 1). A number of coastal ages are older than 20 ka and there appears to be a mode of ^{10}Be ages centered at ~30–35 ka (Fig. 2). These ages could constrain an initial pulse of ice-sheet thinning or an episode of MIS 3 deglaciation followed by non-erosive MIS 2 burial. However, three ^{10}Be ages of 34.5 ± 0.6, 36.4 ± 0.7 and 36.5 ± 0.7 ka are from ~110 m asl, whereas slightly inland there are ^{10}Be ages of 18.7 ± 0.3and 17.3 ± 0.6 ka from ~750 and 680 m asl, respectively. It is possible that these younger ages inland were affected by post-deglaciation mass wasting events, but our ages of >30 ka that pre-date the last deglaciation rest in a region where subglacial erosion during periods of ice cover was probably not as intense as in the primary Hornsund channel (Fig. 2). The large spread in ^{10}Be ages and a ^{10}Be age–elevation relationship that violates simple morphostratigraphy is probably due to a combination of isotopic inheritance in low-erosion zones, combined with sampling bedrock surfaces where the original post-deglaciation surface has not been preserved. Radiocarbon ages from marine sediments in a variety of settings indicate that the ice sheet extended ~70 km west of Hornsund to the continental shelf edge until ~23 ka (Fig. 1). Moreover, a recent synthesis of Eurasian ice sheet extent suggests that ice in the Barents sector was at its maximum between ~23 and 20 ka, making it unlikely that the Hornsund mouth deglaciated at or before ~23–21 ka as suggested by several our ^{10}Be ages older than 23 ka (i.e. 35 ka mode in ^{10}Be ages; Figs 1 and 2; Hughes *et al.*, 2016).

Torbjørnsenfjellet

^{10}Be and *in situ* ^{14}C measurements from the Torbjørnsenfjellet elevation transect constrain the timing of initial deglaciation of the Hornsund fjord mouth. In descending elevational order: the highest elevation sample (TORB-1; 633 m asl) has a ^{10}Be age of 16.0 ± 0.3 ka followed by ^{10}Be ages of 18.3 ± 0.3

Table 3a. *In situ* ^{14}C blank data.

Sample	V_{CO_2} (cc STP)	V_{dilute} (cc STP)	CAMS no.	F_m measured	^{14}C (10^3 atoms)
Blank 11-20-14	0.01643 ± 0.00019	1.446 ± 0.017	168812	0.0048 ± 0.0001	107.03 ± 13.60
Blank 1-15-15	0.01295 ± 0.00015	1.391 ± 0.016	168813	0.0048 ± 0.0001	100.87 ± 13.18
Blank 3-10-15	0.01202 ± 0.00014	1.351 ± 0.015	169702	0.0061 ± 0.0001	153.24 ± 12.92
Blank 4-17-15	0.01346 ± 0.00016	1.424 ± 0.016	170151	0.0045 ± 0.0001	89.66 ± 13.48
Blank 9-16-15	0.01318 ± 0.00015	1.373 ± 0.016	172629	0.0063 ± 0.0001	164.18 ± 13.17
Blank 3-2-16	0.01401 ± 0.00016	1.403 ± 0.016	173886	0.0040 ± 0.0001	67.01 ± 13.22

We report all blank measurements completed since November 2014. The previous LDEO long-term blank was $118.09 \pm 39.28 \times 10^3$ ^{14}C atoms and included blanks up to Septmeber 2013 (Young *et al.*, 2014). The updated LDEO long-term blank value that includes the above measurements is $112.55 \pm 36.83 \times 10^3$ ^{14}C atoms ($n = 23$).

Table 3b. LDEO CRONUS-A *in situ* ^{14}C data.

Sample	Quartz (g)	V_{CO_2} (cc STP)	V_{dilute} (cc STP)	CAMS no.	F_m measured	^{14}C (10^3 atoms g^{-1})
CRONUS-A-3-24-15	4.8795	0.1122 ± 0.0013	1.470 ± 0.017	169935	0.0826 ± 0.0003	735.89 ± 12.27
CRONUS-A-4-16-16	3.6393	0.0553 ± 0.0006	1.439 ± 0.017	173885	0.0618 ± 0.0004	706.23 ± 14.53
CRONUS-A-5-19-16	3.6502	0.0512 ± 0.0006	1.423 ± 0.016	174602	0.0624 ± 0.0002	696.69 ± 13.78

CRONUS-A measurements since October 2013. Values are consistent with a previosuly reported long-term CRONUS-A value of $655.17 \pm 30.87 \times 10^3$ ^{14}C atoms g^{-1} (Young *et al.*, 2014). We note, however, that beginning with the CRONUS-A-3-24-15 extraction, we started working with a new aliquot of the CRONUS-A quartz standard.

ka (TORB-2; 515 m), 36.3 ± 0.6 ka (TORB-4; 279 m), 25.8 ± 0.4 ka (TORB-3; 252 m) and 20.1 ± 0.3 ka (TORB-5; 225 m). The oldest ages rest in the middle of the elevation transect and the youngest age is also the highest elevation sample (Fig. 2). This ^{10}Be age–elevation distribution reveals that ^{10}Be ages are not accurately recording the timing of glacier thinning and deglaciation because ^{10}Be ages do not get younger with decreasing elevation nor are they statistically indistinguishable, the latter of which would suggest rapid deglaciation of all sample sites (i.e. within dating resolution). Corresponding *in situ* ^{14}C ages, however, display a much different age–elevation relationship. Paired ^{10}Be and *in situ* ^{14}C ages for TORB-1 and TORB-2, our highest elevation samples, statistically overlap at 1-sigma, indicating these samples do not contain inherited ^{10}Be (Fig. 3; Table 1). The mid-transect samples with ^{10}Be ages of 36.3 ± 0.6 ka (TORB-4) and 25.8 ± 0.4 ka (TORB-3) have significantly younger ^{14}C ages of 16.9 ± 3.1 and 17.3 ± 3.2 ka, indicating that they contain a ^{10}Be inventory equating to 19.4 ± 3.2 and 8.5 ± 3.2 ka of excess ^{10}Be that accumulated during a previous period(s) of surface exposure. The lowest elevation sample (TORB-5) has a ^{10}Be age of 20.1 ± 0.3 ka and a ^{14}C age 16.7 ± 2.9 ka. We note that the 1-sigma analytical uncertainties of our *in situ* ^{14}C measurements range from 4 to 7%, but that these measurements equate to exposure ages with internal uncertainties that range from ~15 to 23%. Uncertainties range from ~15 to 23% because our measured concentrations intersect the ^{14}C production–time curve where small changes in ^{14}C concentration equate to large changes in exposure age as one approaches surface saturation (nuclide production = decay; Table 2). Regardless, our *in situ* ^{14}C measurements are able to quantify inherited ^{10}Be in three of our five transect samples and, moreover, *in situ* ^{14}C ages are statistically indistinguishable and average 17.4 ± 0.7 ka ($n = 5$; Fig. 2; Fig. 3; Table 2). When accounting for the uncertainty in the production-rate calibration dataset, our ^{14}C measurements reveal glacier thinning across our sample sites at the Hornsund mouth at 17.4 ± 1.5 ka (7.5%; Young *et al*, 2014).

Whereas paired ^{10}Be-^{14}C measurements at Torbjørnsenfjellet constrain the amount of isotopic inheritance and timing of deglaciation to 17.4 ± 1.5 ka, paired ^{26}Al–^{10}Be measurements offer a long-term perspective of surface exposure and burial of the Torbjørnsenfjellet ridgeline. TORB-1 and TORB-2, which have statistically identical ^{10}Be and *in situ* ^{14}C ages, have $^{26}Al/^{10}Be$ ratios consistent with constant exposure (Fig. 4; Table 1). Our mid-transect samples (TORB-4 and TORB-3) contain inherited ^{10}Be and measured $^{26}Al/^{10}Be$ ratios suggest some degree of prolonged burial (Fig. 4). Although the TORB-5 ^{10}Be and *in situ* ^{14}C ages overlap at 2-sigma, suggestive of continuous exposure, the corresponding $^{26}Al/^{10}Be$ ratio of 5.77 ± 0.38 is inconsistent with constant exposure. Because all the transect *in situ* ^{14}C ages are statistically identical and constrain the most recent period of exposure, we use the average *in situ* ^{14}C age to quantify the exposure–burial history of the Torbjørnsenfjellet samples sites before the last ~17.4 ka. Specifically, we subtract 17.4 ka worth of exposure from each paired ^{26}Al–^{10}Be measurement suggestive of non-continuous exposure (i.e. Al/Be <6.75; TORB-3, -4, -5) to quantify pre-17.4 ka exposure and burial at each sample location (Fig. 4). At TORB-5, our approach results in subtracting more ^{26}Al than what was measured, but points to the presence of a small amount of excess ^{10}Be, equating to 3.4 ± 2.9 kyr of exposure. The corrected ratios for our mid-transect samples reveal a significant amount of pre-17.4 ka surface burial equating to ~54–570 kyr, albeit with large uncertainties (Fig. 4).

Wurmbrandegga and Treskelen

At the Wurmbrandegga elevation transect, which is located ~12 km up-fjord from Torbjørnsenfjellet and therefore must have deglaciated at or after 17.4 ± 1.5 ka, six of seven ^{10}Be ages are indistinguishable and indicate that this portion of Hornsund deglaciated at 13.3 ± 0.6 ka (Figs 2 and 3). Unlike the Torbjørnsenfjellet transect, we find no evidence that the samples at Wurmbrandegga contain inherited ^{10}Be, which is suggestive of a greater total erosion depth than the Torbjørnsenfjellet site during periods of ice cover. Finally, our easternmost ^{10}Be ages from perched boulders positioned ~13 km up-fjord from Wurmbrandegga at Treskelen indicate that final deglaciation of the fjord occurred by 13.0 ± 0.7 ka (n = 3) after removal of an older outlier (15.4 ± 0.9; Fig. 2). Combined, our ^{10}Be and *in situ* ^{14}C ages suggest that the Hornsund mouth deglaciated at 17.4 ± 1.5 ka, with ice remaining near the fjord mouth for several thousand years before complete deglaciation between 13.3 ± 0.6 and 13.0 ± 0.7 ka (Figs 2 and 5). Alternatively, our ice-margin chronology allows for initial deglaciation of the Hornsund mouth at 17.4 ± 1.5 ka followed by continued fjord deglaciation, an ice-margin re-advance beyond the Wurmbrandegga transect, followed by inner fjord deglaciation between 13.3 ± 0.6 and 13.0 ± 0.7 ka. However, we are unaware of any sediment packages within the fjord that are suggestive of a significant re-advance of the Hornsund glacier.

The Hornsund outlet glacier during the last glacial cycle

^{10}Be and in situ ^{14}C ages versus traditional radiocarbon constraints

In situ ^{14}C ages indicate that the Hornsund outlet glacier thinned ~400 m at 17.4 ± 1.5 ka, consistent with the timing of coastal deglaciation at Scottbreen (17.7 ± 1.2 ka), Fløyfjellet (15.7 ± 0.7 ka) and Torellbreen (15.9 ± 1.6 ka; Fig. 1). However, previously published records suggest that deglaciation of the western Svalbard coast occurred much later. A series of radiocarbon ages from marine sediments place the SBIS margin out on the shelf edge between ~23 and 20 ka with deglaciation of the SW Spitsbergen coast constrained to ~13.7–11.7 cal ka BP based on minimum-limiting radiocarbon ages from raised marine sediments (Figs 1 and 2; Landvik *et al.*, 1998; Hormes *et al.*, 2013). It is possible that all of our deglaciation ages constrain initial ice-sheet thinning and unroofing of our sampling sites before deglaciation of the coastal lowlands. However, with the exception of the Torellbreen site located north of Hornsund, all of our deglaciation ages span relatively low elevations that should capture the timing of coastal deglaciation (Table 1). Alternatively, there is a possible ~5 kyr offset in the timing of Hornsund deglaciation as defined by our *in situ* ^{14}C ages versus the minimum-limiting radiocarbon of 12.1 cal ka BP. And, the deglaciation age provided by the Wurmbrandegga elevation transect located only ~12 km up-fjord from the outer coast is 13.3 ± 0.6 ka; ~4 kyr younger than the Hornsund mouth *in situ* ^{14}C age, but still older than the minimum-limiting 12.1 cal ka BP deglaciation age. One explanation is that the minimum-limiting 12.1 cal ka BP age does not closely constrain the timing of deglaciation and that deglaciation occurred at either 17.4 ± 1.5 ka or just before 13.3 ± 0.6 ka. Or, the Hornsund outlet glacier thinned to at least 225 m asl at 17.4 ± 1.5 ka (lowest sample in elevation transect) but occupied the fjord mouth for another several thousand years before final deglaciation. A final possibility is that our *in situ* ^{14}C measurements contain inherited *in situ* ^{14}C from a previous period of exposure that occurred before 13.3 ± 0.6 ka or 12.1 cal ka BP.

Modeling ^{10}Be and in situ ^{14}C inventories

To assess our measured ^{10}Be and *in situ* ^{14}C inventories at Torbjørnsenfjellet, we use the Svalbard glaciation curve over the last glacial cycle to model the possible ^{10}Be and *in situ* ^{14}C accumulation history at our sample sites. The Svalbard glaciation curve (Fig. 5) suggests a dynamic SBIS advanced onto the continental shelf several times over the last glacial cycle; between these glacial maxima, the SBIS retreated back to an ice configuration similar to today (Mangerud *et al.*, 1998; Eccleshall *et al.*, 2016; Figs 1 and 5). Notably, the SBIS occupied the shelf or shelf edge during MIS 6, MIS 5d, MIS 5b and MIS 2, with an additional, although probably not as extensive, advance during MIS 4. Although this Svalbard glaciation history is largely based on the Kapp Ekholm section at the head of Isfjorden (Fig. 1), a non-finite radiocarbon age of >40 ka from shell fragments suggests that Hornsund may have deglaciated at least once before the most recent episode of deglaciation (Landvik *et al.*, 1998).

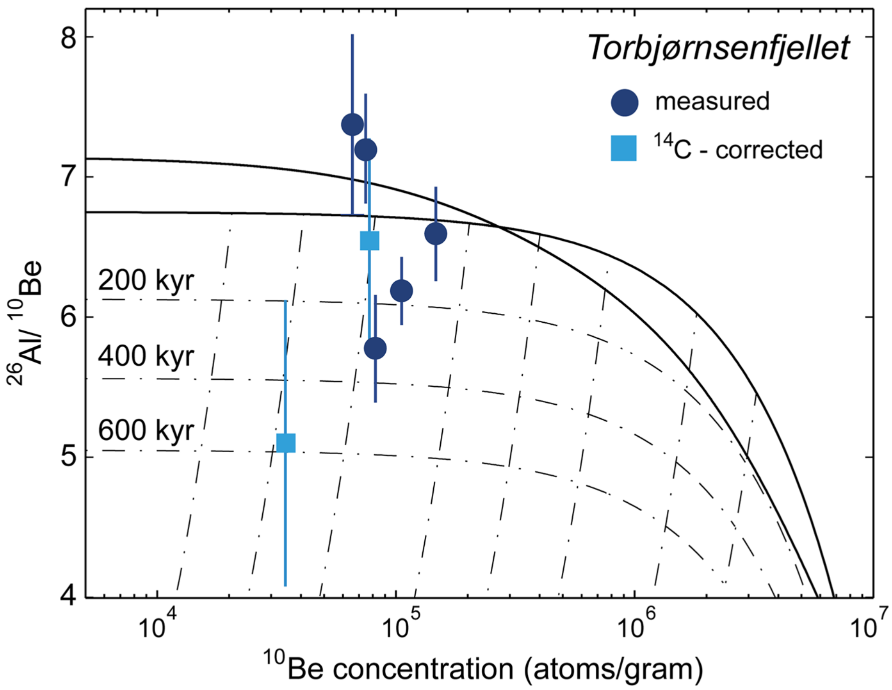

Figure 4. Measured ^{26}Al/^{10}Be ratios against ^{10}Be concentration at Torbjørnsenfjellet. Also plotted is the ^{14}C-corrected ^{26}Al/^{10}Be ratio for samples TORB-3 and TORB-4 using the average ^{14}C age of 17.4 ± 0.7 ka from the Torbjørnsenfjellet transect (i.e. the ratio at 17.4 ka). The present-day ratios for TORB-1 and TORB-2 (>= 6.75) suggest no previous sample burial. The corrected ratio for TORB-5 overlaps zero, which is due to subtracting more ^{26}Al than was measured. However, the corrected ^{10}Be concentration in TORB-5 suggests a small amount of previous exposure. [Colour figure can be viewed at wileyonlinelibrary.com]

This template of SBIS advance and retreat provides a unique opportunity to assess the likelihood of our coastal Torbjørnsenfjellet bedrock sites yielding significant inherited ^{10}Be coupled with *in situ* ^{14}C inventories that have minimal or no inheritance. We assume no prior nuclide inventory at our bedrock sites at the termination of MIS 6 (Fig. 5). Next, we allow surface exposure and nuclide accumulation during MIS 5e, 5c, 5a and 3, burial and nuclide decay during MIS 5d, 5b, 4 and 2, and assume a 'true' deglaciation of ~13.3 ka as defined by our nearby Wurmbrandegga ^{10}Be ages. We want to assess the maximum amount of potential inherited nuclides at Torbjørnsenfjellet so we assume no bedrock erosion during periods of ice cover; only nuclide decay via burial decreases the nuclide inventory. When modeling ^{10}Be and *in situ* ^{14}C concentrations using these assumptions, our Torbjørnsenfjellet bedrock sites would have ^{10}Be and *in situ* ^{14}C concentrations equating to exposure ages of ~81 and 14.6 ka, respectively (Fig. 5).

Our model suggests that using the Svalbard glaciation curve as a template for surface exposure and burial results in ^{10}Be and *in situ* ^{14}C concentrations similar to what we measured – old ^{10}Be ages coupled with younger *in situ* ^{14}C ages. In addition, our approach results in a small amount of inherited *in situ* ^{14}C at our Torbjørnsenfjellet sample sites. Rather, the modeled duration of MIS 2 burial is not long enough to completely remove the accumulated MIS 5e to MIS 3 inventory of *in situ* ^{14}C (Fig. 5). Because the duration of MIS 2 burial results in a small amount of inherited *in situ* ^{14}C in our model experiment, we next assume that our measured *in situ* ^{14}C concentrations are influenced by inheritance and then calculate the timing of deglaciation needed to result in our measured *in situ* ^{14}C concentrations. With this approach, our measured *in situ* ^{14}C concentrations are achieved if deglaciation occurs at ~15.6 ka, which is suggestive of a 'true' deglaciation age that is ~2 ka younger than our measured *in situ* ^{14}C ages. However, our model set up is tuned to allow for the maximum amount of inherited *in situ* ^{14}C because we assume no glacial erosion during periods of ice cover, which is not supported by our measured ^{10}Be ages, which vary across the transect and imply varying degrees of glacial erosion. The modeled exposure–burial histories result in ^{10}Be concentrations that equate to ~81–85 ka, whereas our measured ^{10}Be ages from the Torbjørnsenfjellet transect range from ~16 to ~36 ka (Fig. 3C). This disparity between the modeled and measured ^{10}Be ages suggests that the Torbjørnsenfjellet sample sites either experienced some degree of glacial erosion that stripped away ^{10}Be (and *in situ* ^{14}C) from the bedrock sites and/or our sample sites experienced less surface exposure during the last glacial cycle than what we modeled using the Svalbard glaciation curve.

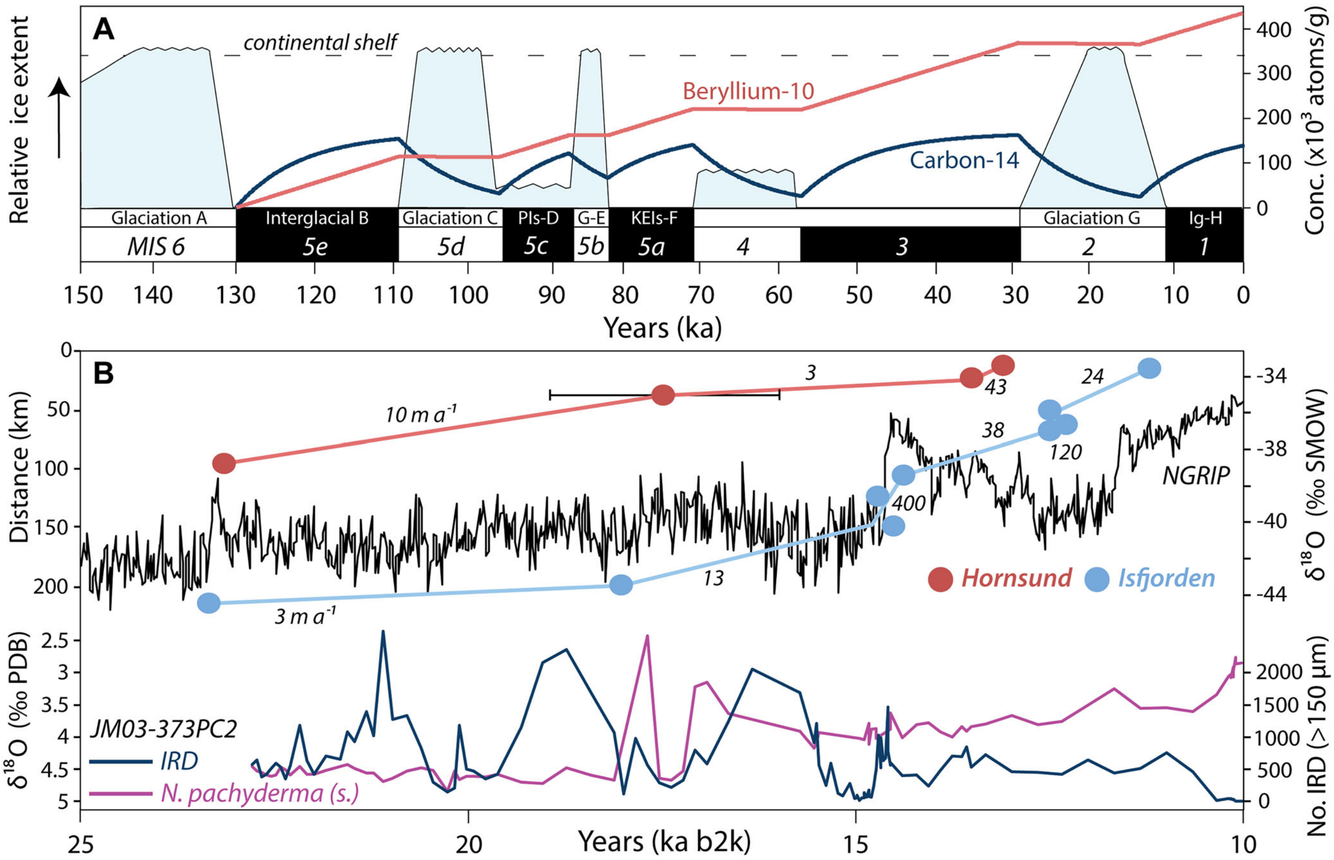

Figure 5. (A) Svalbard glaciation curve (Mangerud *et al.*, 1998; Eccleshall *et al.*, 2016). The timing of glacier advance has been tuned to coincide with the Marine Isotope Stage (MIS) that each advance is thought to correlate to as discussed in Eccleshall *et al.* (2016). MIS definitions are from Lisiecki and Raymo (2005). PIs-D: Phantomodden insterstadial D; G-E: Glaciation E; KEIs-F: Kapp Ekholm interstadial F; Ig-H: interglaciation H. Shown are the modeled proof-of-concept ^{10}Be and *in situ* ^{14}C concentrations at 252 m asl (TORB-3) assuming that our Torbjørnsenfjellet sample sites become exposed and buried following the Svalbard glaciation curve and assuming a 'true' deglaciation age of 13.3 ka. The resulting *in situ* ^{14}C concentration equates to an exposure age of ~14.6 ka. (B) Retreat chronologies for the Hornsund and Isfjorden sectors of the SBIS compared to the NGRIP δ^{18}O record (North Greenland Ice Core Project members, 2004) and a record of ice-rafted detritus and δ^{18}O (*N. pachyderma s.*) from core JM03-373PC2 located south-west of Hornsund (Rasmussen *et al.*, 2007; Jessen *et al.*, 2010; see location No. 2 on Fig. 1). Numbers between data points are the calculated net (minimum) ice-margin retreat rates using the age constraint midpoint. Symbols are larger than the uncertainties with the exception of the *in situ* ^{14}C-based Torbjørnsenfjellet data point at 17.4 ± 1.5 ka. The JM03-373PC2 age model has been re-calibrated using Calib 7.1 and a reservoir correction of 440 years (Stuiver *et al.*, 2018; see Rasmussen *et al.*, 2007 and Jessen *et al.*, 2010 for details). [Colour figure can be viewed at wileyonlinelibrary.com]

Table 4. Summary of deglacial radiocarbon ages.

Location (Fig. 1)	Latitude	Longitude	^{14}C years	1-sigma uncertainty	cal a BP	Setting (Site; Lab ID; material; description)	Reference
1	76	16	16 750	110	19 610 ± 170	JM02-460PC; AAR-8764; *N. pachyderma*; hemipelagic deposits above till	Rasmussen *et al.* (2007)
2	76.4	13.1	19 310	140	22 660 ± 170	JM03-373PC2; AAR-8773; *N. pachyderma*; hemipelagic deposits above debris flow	Rasmussen *et al.* (2007)
3	76.333	12.600	19 630	150	23 040 ± 230	JM03-374PC; AAR-8766; *N. pachyderma*; hemipelagic deposits above debris flow	Jessen *et al.* (2010)
4	77.220	12.625	16 880	80	19 750 ± 170	NP90-46; Beta-71988; *E. excavatum*; marine sediment above diamicton	Cadman (1996)
5	77.617	9.936	16 035	130	18 780 ± 135	NP90-36; Tua-845; *N. pachyderma*; above till, glaciomarine with IRD of Svalbard origin	Elverhøi *et al.* (1995)
6	77.817	9.093	19 815	120	23 240 ± 190	NP90-39; Tua-557; *N. pachyderma*; above till, glaciomarine with IRD of Svalbard origin	Elverhøi *et al.* (1995)
7*	78.7588	10.7463	NA	NA	20 290 ± 2120	JL00-31 Leefjellet; ^{10}Be exposure age	Landvik *et al.* (2013)
8*	79.2388	11.8139	NA	NA	25 010 ± 1010	Langskipet; ^{10}Be exposure age; 611 m a.s.l.	Gjermundsen *et al.* (2013)
9*	79.4640	11.3937	NA	NA	21 840 ± 950	Kaf-1; ^{10}Be exposure age; 836 m a.s.l.	Gjermundsen *et al.* (2013)
10*	79.6013	11.7866	NA	NA	19 340 ± 1260	Average of Ovo-3 & Ovo-4; ^{10}Be exposure ages; 687 m and 730 m a.s.l.	Gjermundsen *et al.* (2013)
11*	79.7216	10.9483	NA	NA	17 860 ± 2040	Average of 99-01 & 99-05 Danskøya; ^{10}Be exposure ages; 77 m and 74 m a.s.l.	Landvik *et al.* (2003); Gjermundsen *et al.* (2013)
12*	79.7375	13.6142	NA	NA	21 670 ± 1310	R4; ^{10}Be exposure age; 85 m a.s.l.	Gjermundsen *et al.* (2013)
13*	80.2088	22.4817	NA	NA	26 730 ± 3910	Bluffen-2; ^{10}Be exposure age; 165 m a.s.l.	Hormes *et al.* (2011)
14*	80.2073	22.5102	NA	NA	28 270 ± 2140	Bluffen-3; ^{10}Be exposure age; 123 m a.s.l.	Hormes *et al.* (2011)
15	79.095	25.095	18 640	100	21 990 ± 170	GC06; NA; bulk sediments; mud above diamict	Hogan *et al.* (2010)
16	76.7	16.4	10 660	220	11 730 ± 400	Werenskioldbreen; U-2831; *Mya truncata* (reworked); glacier margin near medial moraine	Birkenmajer and Olsson (1998)
17	77.08	15.18	10 790	160	12 100 ± 320	Werenskioldbreen; U-2972; *Mya truncata* (reworked); esker in glacier forefield	Birkenmajer and Olsson (1998)
18	77.55	14.03	12 350	145	13 690 ± 170	Wedel Jarlsberg/Dyrstadalen; Ua-1081; shell fragment; beach gravels at 50.5 m a.s.l.	Salvigsen and Elgersma (1991)
19	78.188	9.943	15 255	180	17 930 ± 230	NP90-21; Tua-359; *N. pachyderma*; marine sediment above diamicton	Elverhøi *et al.* (1995)
20	78.022	11.857	12 835	100	14 360 ± 250	NP-90-25; Tua-553; unidentified mollusc; glaciomarine mud over massive till	Svendsen *et al.* (1996)
21	78.047	12.988	12 985	145	14 630 ± 330	88-02; Tua-42; *Nucula tenuis*; mud with dropstones on firm diamicton	Svendsen *et al.* (1992)
22	78.071	13.759	12 740	190	14 260 ± 360	Linnevatnet; Ua-732, shells; marine sediment above diamicton	Mangerud and Svendsen (1990)
23	78.293	14.803	10 975	60	12 390 ± 150	JM98-818-PC; Tua-5191; Foraminifera; glacimarine sediments on glaciomarine diamicton	Forwick and Vorren (2007)
24	78.277	15.260	10 835	140	12 180 ± 180	NP87-144; Ua-757; mollusc unidentified	Elverhøi *et al.* (1995)
25	78.380	15.479	11 025	90	12 430 ± 150	90-03 PC; Tua-442; Foraminifera; laminated glacimarine mud on top of till	Elverhøi *et al.* (1995)
26	78.565	16.428	10 085	115	10 980 ± 160	NP90-01-PC; Tua-186; Bryzoyoa; marine mud on top of laminated deglaciated mud	Svendsen *et al.* (1996)
27	78.552	16.540	10 240	60	11 130 ± 80	Kapp Ekholm V-H8; Ua-35635; *Hiatella arctica*; marine sands on LGM melt-out/ subglacial till	Hormes *et al.* (2013)
28*	78.8400	10.6023	NA	NA	16 470 ± 1900	JL-00-10; ^{10}Be exposure age; 325 m a.s.l.	Landvik *et al.* (2003)
29	79.022	11.104	13 960	120	16 180 ± 190	NP90-9-PC3; WHG-941; mixed benthic forams; laminated marine mud over glacial till	Landvik *et al.* (2013)
30*	79.1914	12.0009	NA	NA	16 930 ± 1120	Flakstor; ^{10}Be exposure age; 217 m a.s.l.	Gjermundsen *et al.* (2013)
31*	79.7447	13.0695	NA	NA	15 570 ± 790	R10; ^{10}Be exposure age; 84 m a.s.l.	Gjermundsen *et al.* (2013)
32*	80.0367	18.7056	NA	NA	16 810 ± 1220	Flora-2; ^{10}Be exposure age; 220 m a.s.l.	Hormes *et al.* (2011)
33	80.355	16.299	14 165	135	16 470 ± 230	NP94-51SC2; mixed benthic foraminifera; outer shelf	Koç *et al.* (2002)

All radiocarbon ages were calibrated with Calib 7.1 and the Marine13 database (Stuiver *et al.*, 2018), and use a ΔR value of 107 ± 52 years (Mangerud *et al.*, 2006; Hormes *et al.*, 2013). Calibrated ages and their uncertainties have been rounded to the nearest decade. Note that locations marked with an asterisk report ^{10}Be ages and are reported as years of exposure, not years BP, and have been re-calcualted using v3 of the exposure age calculator found at https://hess.ess.washington.edu/ (see main text).

The minor differences in our measured versus modeled ^{10}Be and *in situ* ^{14}C inventories is probably due to variable glacial erosion during periods of ice cover and/or uncertainty in the exact duration that our bedrock sites experienced ice cover and nuclide decay versus the exact duration of ice-free conditions and nuclide accumulation. Despite these differences, our model captures the overall pattern of older ^{10}Be ages influenced by isotopic inheritance coupled with *in situ* ^{14}C ages that are much younger. The maximum inheritance scenario that does not account for glacial erosion results in *in situ* ^{14}C concentrations with only a small degree of inheritance (~2 ka). Glacial erosion during periods of ice cover, as suggested by the spread in ^{10}Be ages, would not only remove a portion of previously accumulated ^{10}Be, but also remove *in situ* ^{14}C resulting in bedrock sample sites that do not contain previously accumulated ^{14}C before the most recent period of exposure. Thus, it is unlikely that our *in situ* ^{14}C ages are influenced by inheritance and that the 'true' age of deglaciation at the fjord mouth is 17.4 ± 1.5 ka (Fig. 2).

Coastal deglaciation of western Spitsbergen

Our ^{10}Be and *in situ* ^{14}C ages indicate that the region of coastal south-western Spitsbergen south of Isfjorden deglaciated between ~18 and 16 ka (Fig. 1). This timing of coastal deglaciation in south-western Spitsbergen is broadly consistent with the timing of coastal deglaciation in north-western Spitsbergen, which is constrained to ~17.9–15.6 cal ka BP based on the oldest published coastal radiocarbon ages, and low-elevation ^{10}Be ages from the region (Fig. 1; Table 4). Although several older ^{10}Be ages from north-western Spitsbergen range from ~25.0 to 19.3 ka (Fig. 1), these ages are mainly from high-elevation nunataks and probably record initial ice sheet thinning rather than coastal deglaciation. Two older ^{10}Be ages of 26.7 ± 3.9 and 28.3 ± 2.1 ka from erratics in Nordaustlandet may record the timing of early coastal deglaciation considering their low elevations (165 and 123 m asl), but their anomalously old ages could simply represent the presence of cold-based ice and ^{10}Be inheritance as suggested by the original authors (Hormes *et al.*, 2011; Fig. 1; Table 1). It appears that the timing of coastal deglaciation in north-western and south-western Spitsbergen (between Hornsund and Scottbreen) is perhaps similar, but deglaciation through the Isfjorden trough occurred much later (Figs 1 and 5). A relatively dense transect of radiocarbon ages extending from the outer shelf to near the modern ice margin in Isfjorden indicates that although the timing of initial retreat from the shelf edge is similar to the timing of initial ice-margin retreat west of Hornsund (~23 ka), ice at Isfjorden remained near the shelf edge until as late as ~17.9 cal ka BP (Fig. 1). In addition, the Isfjorden mouth did not deglaciate until ~14.3 cal ka BP, 1.5–3.5 kyr later than the timing of coastal deglaciation in south-western Spitsbergen (Fig. 1). A record of ice-rafted detritus (IRD) located immediately south-west of Hornsund reveals peaks in IRD at ~18.7 and ~16.3 ka, suggestive of increased calving and ice-margin retreat at these times (Fig. 5; Rasmussen *et al.*, 2007; Jessen *et al.*, 2010). This increase in IRD deposition is broadly correlative with the timing of coastal deglaciation as constrained by our ^{10}Be and *in situ* ^{14}C measurements (~18–16 ka); however, the resolution of our record, and in particular *in situ* ^{14}C-based age of deglaciation (17.4 ± 1.5 ka), prevents us from linking the timing of deglaciation to any one IRD peak (Fig. 5).

The timing of coastal deglaciation appears to have differed between Isfjorden and south-western Spitsbergen, but the relatively sparse number of radiocarbon constraints between the outer shelf and coast at Hornsund prevents us from determining if ice in this sector stayed near the shelf edge for several thousand years or gradually retreated between ~23 and 17.4 ± 1.5 ka (Figs 1 and 5). Existing age constraints from Isfjorden and Hornsund allow us to place millennial- to centennial-scale retreat of the ice margin into a long-term context (Fig. 5). At Hornsund, the SBIS retreated at a millennially averaged rate of ~10 m a^{-1} between ~23 and 17.4 ka, and ~3 m a^{-1} between 17.4 and 13.3 ka. However, the 17.4 ± 1.5 ka constraint along the Hornsund transect at Torbjørnsenfjellet requires ~400 m of ice-sheet thinning in addition to constraining the lateral retreat of the ice margin. Following deglaciation at Wurmbrandegga, the ice margin retreated between ~13.3 and 13.0 ka at a rate of ~43 m a^{-1} (Fig. 5). At Isfjorden, initial ice-margin retreat occurred at ~3 m a^{-1} between 23.2 cal ka BP and 17.9 cal ka BP followed by a slightly faster rate of retreat of ~13 m a^{-1} between ~17.9 and 14.5 cal ka BP. Afterwards, retreat rates increased significantly between ~14.5 and 14.3 cal ka BP (~400 m a^{-1}), with another pulse of rapid deglaciation centered on ~12.3 cal ka BP (~120 m a^{-1}; Fig. 5). We note that these retreat rates should be considered minimum or net retreat rates because our methods are limited in their ability to identify pulses of fast ice retreat and almost certainly smooth over episodes of faster ice retreat.

It appears that initial retreat of the western margin of the SBIS occurred as early as ~23 ka, synchronous with the initial rise in boreal summer insolation at ~24–23 ka and consistent with the onset of initial retreat of the southern margin of the Laurentide Ice Sheet (e.g. Ullman *et al.*, 2015), but pre-dating any significant rise in eustatic sea level. However, SBIS retreat rates remained relatively slow until at least 17.9 ka and perhaps as late as ~14.5 ka as suggested by the Isfjorden recession chronology. Indeed, whereas initial retreat was contemporaneous with rising summer insolation, elevated retreat rates were not achieved until several millennia later, contemporaneous with rising temperatures as recorded in Greenland ice cores at the onset of the Bølling–Allerød (Fig. 5).

Conclusions

Deglaciation of the south-western Spitsbergen coast probably occurred between ~18 and 16 ka based on new ^{10}Be and *in situ* ^{14}C measurements from four separate sites along ~60 km of south-western Spitsbergen. ^{10}Be measurements in bedrock along a ~400-m elevation transect display varying amounts of isotopic inheritance and are unable to constrain the timing of deglaciation or ice-sheet thinning, but complimentary *in situ* ^{14}C measurements are statistically identical and mark an episode of ice-sheet thinning at 17.4 ± 1.5 ka. Following coastal deglaciation, the middle of Hornsund deglaciated by 13.3 ± 0.6 ka with complete fjord deglaciation by 13.0 ± 0.7 ka. Our dataset indicates that the timing of coastal deglaciation in south-western Spitsbergen was significantly earlier than previous estimates based on a limited number of minimum-constraining radiocarbon ages. Previously published age constraints, coupled with our new ^{10}Be and *in situ* ^{14}C ages, suggest that the western coast of Spitsbergen between Hornsund and Scottbreen deglaciated between ~18 and 16 ka, and that deglaciation of Isfjorden occurred much later with the fjord mouth deglaciating at ~14.3 ka. Initial retreat of the western SBIS margin appears to have occurred at ~23 ka followed by relatively variable and asynchronous retreat between the Isfjorden and Hornsund sectors of the SBIS, thus highlighting the dynamic nature in which ice sheets recede. Lastly, *in situ* ^{14}C measurements offer the ability to rectify ambiguous ^{10}Be-based datasets influenced by isotopic inheritance in order to extract key chronological information.

Acknowledgements. We are grateful for logistical support from the Polish Polar Station at Hornsund and two Arctic field grants to Endre Gjermundsen from the Svalbard Science Forum by the Norwegian Research Council. A portion of this work was performed under the auspices of the U.S. Department of Energy by the Lawrence Livermore National Laboratory under contracts DE-AC52-07NA27344. Fieldwork would not have been possible without the help of Ellinor Falkgjerdet, Lina Gislefoss, Carl Petter Nielsen and Bjørn Andrè Skjæret. We thank one anonymous reviewer for several suggestions that helped improve the manuscript. This is LDEO contribution #8225 and LLNL-JRNL-743296.

Abbreviations. IRD, ice-rafted detritus; LDEO, Lamont-Doherty Earth Observatory; MIS, Marine Isotope Stage; SBIS, Svalbard–Barents Ice Sheet; SUERC, Scottish Universities Environmental Research Centre.

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