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The rapid deglaciation of the Skagafjörður fjord, northern Iceland

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The Skagafjörður fjord in northern Iceland is located between the Tröllaskagi Peninsula in the east and the Skagi Peninsula in the west. The tributary valleys of the fjord originate in the highland area about 15 km north of the Hofsjökull icecap. The results of this work improve the knowledge of the deglaciation pattern in Skagafjörður and explore the adequacy of the ^{36}Cl cosmic ray exposure dating method in an Icelandic environment, where this method has rarely been applied to deglaciated surfaces. The ^{36}Cl dating method was applied to 13 rock samples taken on a transect from the coastal areas towards the highlands. All samples were obtained from rock outcrops with glacier-polished surfaces from the Last Glaciation and from one of the few well-preserved erratic boulders. The cosmogenic results, combined with previous radiocarbon results, indicate that the ice margin was situated in the outermost sector of Skagafjörður at approximately 17–15 ka BP. Subsequently, it retreated and occupied the central part of the fjord between 15 and 12 ka BP and then the innermost sector of the fjord about 11 ka BP. The samples collected between this position and the highlands show an average age of approximately 11 ka, indicating rapid deglaciation after the early Preboreal. These results agree with earlier studies of the deglaciation history of northern Iceland, reinforce previous deglaciation models in the area and enable a better understanding of glacial evolution in the North Atlantic from the Late Pleistocene to Holocene transition.

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The deglaciation of the Northern Hemisphere ice sheets is considered the primary cause of changes in sea surface temperature and salinity in the North Atlantic and of planetary climate change, due to its influence on the Atlantic meridional overturning circulation during the Late Pleistocene and Holocene (Clark *et al.* 2012; Renssen *et al.* 2012; Chen *et al.* 2015; Gil *et al.* 2015; Jaccard *et al.* 2016; Lippold *et al.* 2016; Xiao *et al.* 2017). The evolution of palaeo-ice sheets is increasingly attracting attention to interpret the changes observed in modern ice sheets in the Northern Hemisphere and their contribution to sea-level rise (Clark *et al.* 2009; Carlson & Clark 2012; Kleman & Applegate 2014; Stokes *et al.* 2015; Hughes *et al.* 2016; Liakka *et al.* 2016; Sinclair *et al.* 2016; Stroeven *et al.* 2016; Stokes 2017). In this context, knowledge of the temporal evolution of the Icelandic ice sheet (IIS) is of particular importance to understand the climatic interconnections between the Greenland and Eurasian ice sheets and the potential spatio-temporal variability across the northern North Atlantic. The position of Iceland is particularly important due to the meeting of the northward-flowing North Icelandic Irminger Current and the southward-flowing Arctic East Greenland Current (Malmberg 1985). These factors make the Icelandic glaciers and environment extremely vulnerable and sensitive to any change in ocean circulation patterns (Andrews & Giraudeau 2003; Xiao *et al.* 2017).

The IIS covered the whole island during the Last Glacial Maximum (LGM). The ice was approximately 2 km thick at its centre (Hubbard *et al.* 2006) and the grounded ice reached the shelf edge, about 150 km off the present coastline of northern Iceland, between 29.1 and 18.6 cal. ka BP according to radiocarbon ages of submarine glacigenic sediment cores (Andrews *et al.* 2000; Spagnolo & Clark 2009; see synthesis and discussion in Pétursson *et al.* 2015 and Patton *et al.* 2017). Approximately 18.6 cal. ka BP, the sea level started to rise around Iceland (Andrews *et al.* 2000; Ingólfsson & Norðdahl 2001). During the Bølling interstadial (14.7 to 14.1 ka BP), the IIS retreated rapidly, or rather collapsed, because of significant sea-level rise (Norðdahl & Ingólfsson 2015). During this time, the IIS covered the centre and southeast of the island (Norðdahl & Ingólfsson 2015; Pétursson *et al.* 2015). Changes in pollen, marine sediments and sea levels suggest a cold period and glacial advances at the end of the Bølling interstadial, around 14 cal. ka BP according to radiocarbon dates (Ingólfsson *et al.* 1997; Pétursson *et al.* 2015). Changes in pollen and radiocarbon dates collected in Lake Torfadalsvatn on the Skagi Peninsula demonstrated that shrub tundra vegetation was established in the Peninsula, with temperatures rising almost to present levels during the Allerød interstadial (13.9 to 12.9 ka BP; Rundgren 1995, 1999; Rundgren & Ingólfsson 1999).

The Younger Dryas stadial (12.9 to 11.7 ka BP) is reflected in Iceland by the clear IIS re-advance. In northern Iceland, the glaciers expanded and reached the present coastline and entered the fjords, but many interfluves remained ice-free during this period (see synthesis in

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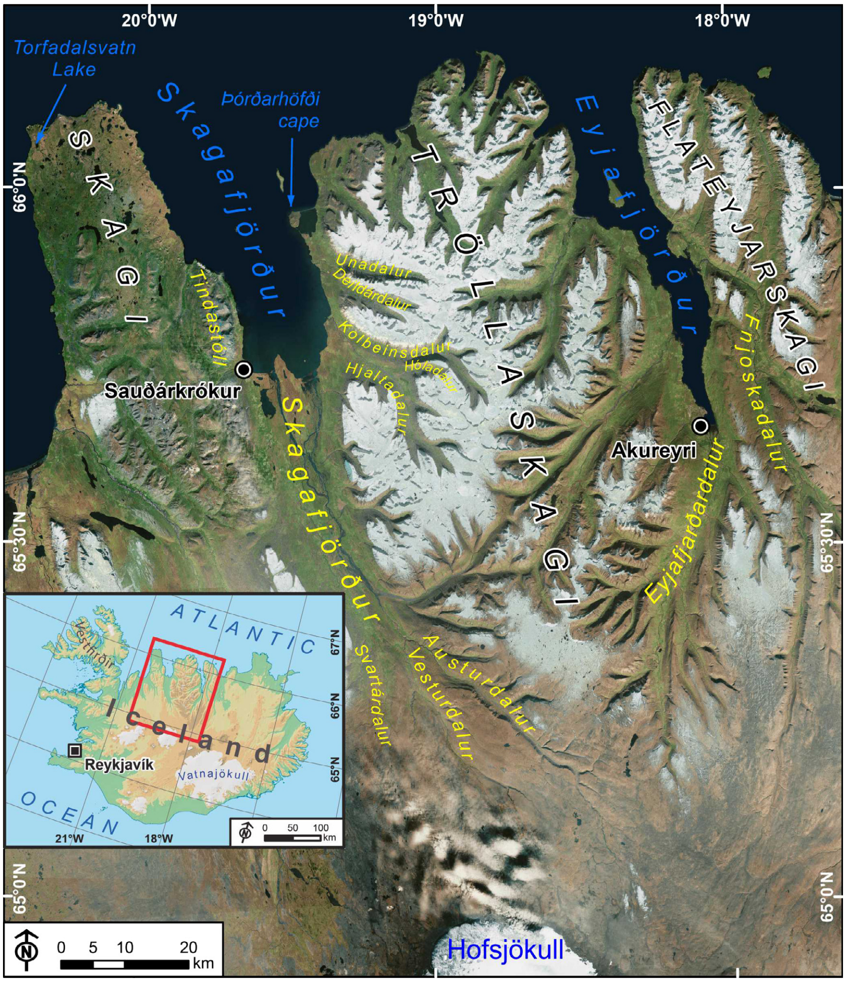

Fig. 1. Location of the study area, villages, main valleys, peninsulas and fjords cited in the text. See detailed map in Fig. 2.

Pétursson *et al.* 2015). The sudden cooling at the beginning of the Younger Dryas is demonstrated in the Skagi Peninsula in a change to grass tundra, which reverted to shrub tundra vegetation at the end of this stadial, according to pollen records (Rundgren 1995, 1999).

The IIS retreated at the end of the Younger Dryas and re-advanced again in the early Preboreal, around 11.2 ka BP (see synthesis in Pétursson *et al.* 2015). During this period, the IIS advanced into the fjords in northern Iceland (Ingólfsson *et al.* 1997; Norðdahl & Einarsson 2001; Norðdahl & Pétursson 2005). After 11.2 ka BP, the glaciers retreated rapidly (Kaldal & Víkingsson 1990; Andrews *et al.* 2000; Norðdahl & Einarsson 2001; Geirsdóttir *et al.* 2002, 2009). By 10.3 ka BP, when the Saksunarvatn tephra was deposited, much of the centre and north of Iceland was already deglaciated (Stötter *et al.* 1999; Caseldine *et al.* 2003; Larsen *et al.* 2012; Striberger *et al.* 2012). Research in the Vestfirðir Peninsula, in the northwest of Iceland where, Drangajökull ice cup still remain, reports a local icecap during the LGM, which remained during the whole deglaciation of this Peninsula, with evidence derived from cosmogenic dating (Principato *et al.* 2006). About when the Saksunarvatn tephra was deposited, the dimensions of this local icecap were suggested to have been similar to its present dimensions, according to radiocarbon dating of pollen and macrofossil plants covered by the Saksunarvatn tephra in some Vestfirðir Peninsula lakes (Eddudóttir *et al.* 2015, 2016; Harning *et al.* 2016). However, different behaviour of Drangajökull and later deglaciation at southeast of Drangajökull have been suggested, supported by ^{14}C and tephrochronology ages in lake sediment cores (Schomacker *et al.* 2016). Recent work, supported by ^{14}C dates from entombed dead vegetation and lake sediment records, disputes this partial contradiction and supports deglaciation of the southeast part of Drangajökull during the Lateglacial or Early Holocene (Harning *et al.* 2018). Furthermore, cosmogenic exposure dating of glacial landforms and boulders in coastal areas around Drangajökull indicates that extensive outlet glaciers to the north and northeast of the present Drangajökull icecap persisted at least until *c.* 9 ka BP (Brynjólfsson *et al.* 2015). Despite our improved understanding of the deglaciation in Iceland, several issues remain uncertain, for example the deglaciation pattern in the northern part of the island, especially around the Tröllaskagi Peninsula.

The cosmic ray exposure (CRE) dating method has been applied successfully to study the geochronological deglaciation pattern of European ice sheets and mountains (see synthesis in Ivy-Ochs 2015; Vasskog *et al.* 2015; Chenet *et al.* 2016; Hughes *et al.* 2016; Stroeven *et al.* 2016; Wirsig *et al.* 2016; Dede *et al.* 2017; Palacios *et al.* 2017b; Stokes 2017); however, it is still rarely applied to improve deglaciated surface studies in Iceland (Principato *et al.* 2006; Brynjólfsson *et al.* 2015). This may be because quartz is absent in basalts, the most prevalent rocks in Iceland, and the Be^{10} dating method cannot be applied; also, the intensive erosion and weathering hinders finding original deglaciated surfaces and these surfaces may occasionally have been covered by ash layers shielding them from cosmic radiation. However, the ^{36}Cl dating method has been applied successfully in basalts (Swanson & Caffee 2001; Phillips 2003; Principato *et al.* 2006; Licciardi *et al.* 2008; Schimmelpfennig 2009; Schimmelpfennig *et al.* 2009, 2011; Brynjólfsson *et al.* 2015); the CRE results can be validated by other methods previously applied, mainly radiocarbon dating. Moreover, some well-preserved protruding glacial landforms can be observed in Skagafjörður where ash was almost immediately blown away after recent eruptions, which enables them to be targeted for CRE dating.

This study aimed to present a new data set of ^{36}Cl cosmogenic exposure dates to improve our understanding of the deglaciation pattern in the Skagafjörður from during the Lateglacial until the Holocene. Although ^{36}Cl production systematics are complex, the ^{36}Cl production rate parameters have been improved recently (Schimmelpfennig *et al.* 2009; Marrero *et al.* 2016a, b). For these reasons, this work also aimed to explore the possibility of obtaining an adequate geochronology of deglaciation in Iceland by applying the new ^{36}Cl age models.

Study area

Location

The Skagafjörður is located in central northern Iceland, between the Tröllaskagi Peninsula to the east and the Skagi Peninsula to the west (Fig. 1). This U-shaped fjord is ~25 km wide at the mouth and 12 km wide at the current coastline. The broad, flat valley bottom gently narrows inland. About 30 km south of the present coastline, the valley continues up to the highland plateau in three tributary valleys: Svartárdalur to the west, Vesturárdalur in the centre and Austurdalur to the east (Fig. 1). These valleys cut into the highland plateau at 700–800 m a.s.l. north of the Hofsjökull icecap. To the west of Skagafjörður, a 700–1000 m high mountain chain continues from the highland plateau in the south towards the Skagi Peninsula. To the east, a higher and broader mountain chain, the Tröllaskagi Peninsula, separates the Skagafjörður and Eyjafjörður fjords. This mountain massif is up to 1500 m high and 40–50 km wide with numerous tributary valleys. The Peninsula has been sculpted by glacial erosion with glacially carved fjords, valleys and cirques, with over 150 alpine glaciers at present (Sigurðsson & Williams 2008). The bedrock in the area dates to 16–3.3 Ma (Moorbath *et al.* 1968; Watkins & Walker 1977; McDougall *et al.* 1984) and is mostly composed of jointed basaltic lava flows 2–30 m thick, often separated by lithified sedimentary horizons 30–50 m thick (Sæmundsson 1979; Sæmundsson *et al.* 1980; Hjartarson & Sæmundsson 2014). The upper parts of the valley slopes, at 600–900 m a.s.l., are often very steep, while the lower parts are gentler and generally covered by glacial deposits or debris talus. The slopes are also affected by numerous rock slope failures and deep-seated gravitational slope deformation, where the red interbed layers act as decollement levels (Jónsson 1976; Cossart *et al.* 2014; Feuillet *et al.* 2014). These macro-mass movements have been described as the paraglacial response of the deglaciation during the Early Holocene (Mercier *et al.* 2013, 2017; Cossart *et al.* 2014; Coquin *et al.* 2015; Decaulne *et al.* 2016), although in fact, some are still active (Wangensteen *et al.* 2006).

Previous knowledge of the Skagafjörður deglaciation patterns

During the Last Maximum Glaciation extent in Iceland, the north part of Skagafjörður was covered by ice extending far beyond the present coastline, confirmed by radiocarbon dating of glacigenic marine sediments (see synthesis in Pétursson *et al.* 2015 and Patton *et al.* 2017). The striation pattern in the Skagafjörður area indicates an ice-flow direction parallel to the main fjord in a northward direction and a predominantly westward direction from the main tributary valleys on the eastern side of the fjord (Víkingsson 1978; Bourgeois *et al.* 2000). Pollen, diatom and organic carbon analyses in the sedimentary series of the Torfadalsvatn and other nearby lakes on the outermost part of the Skagi Peninsula (Rundgren 1995; Rundgren & Ingólfsson 1999) have revealed that deglaciation was already underway in the Allerød interstadial. Severe cooling occurred during the Younger Dryas, with a significant impact on vegetation, but the Skagi Peninsula remained ice-free (Rundgren 1995; Rundgren & Ingólfsson 1999). The Younger Dryas ice extent in Skagafjörður is not well marked, except for radiocarbon-dated shells around 12 cal. ka BP and for Vedde Ash distribution (12.1 cal. ka BP). From this information, some authors have deduced the glacier margin to have been on the lowlands close to the present coastline (Norðdahl & Pétursson

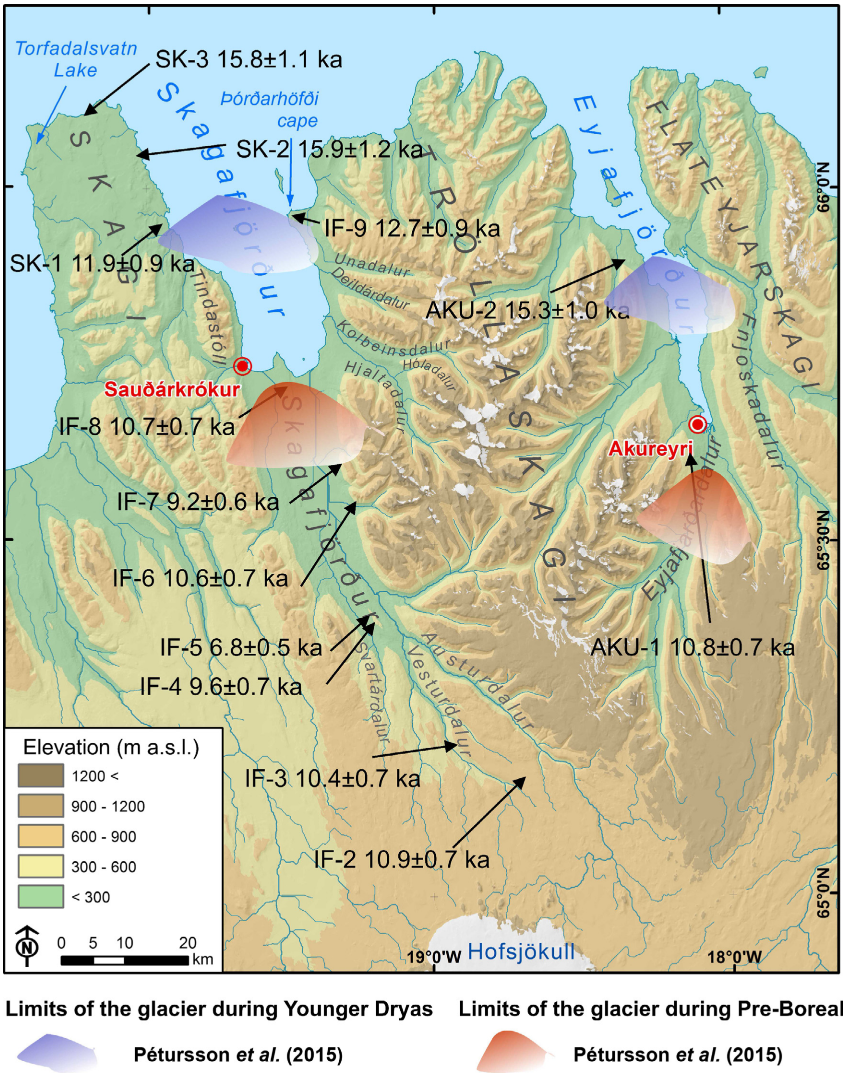

Fig. 2. Location of sampling sites and ice margins at different ages, according to previous deglaciation models (Pétursson *et al.* 2015).

2005; Norðdahl *et al.* 2008), but others consider that the glacier front extended a few km beyond the current coastline (Pétursson *et al.* 2015). It is not known either where the Skagafjörður glacier was located during the Preboreal; however, the glacier probably terminated on land, occupying the main valleys of Skagafjörður by 11.2 cal. ka BP, according to fossil marine molluscs, with a relative sea level at around 40–50 m above the present level, when the Skagi Peninsula was free of ice as observed from lake sediments (Ingólfsson *et al.* 1997; Norðdahl & Pétursson 2005). After the Preboreal, the temperature rose during the first half of the Holocene (Caseldine *et al.* 2006; Geirsdóttir *et al.* 2013). In fact, the expansion of birch in Tröllaskagi had begun by 10 ka BP, reaching its maximum between 8 and 5 ka BP, that is during the Holocene Thermal Maximum (Wastl *et al.* 2001; Caseldine *et al.* 2006).

Material and methods

Research into the deglaciation pattern in the Skagafjörður was approached through a number of fieldwork campaigns during the summers of 2010, 2012, 2014 and 2015, using fieldwork and photo-interpretation to identify the most suitable landforms for sampling purposes, including glacially polished bedrock outcrops, moraines and erratic boulders. The main tributary valleys on the eastern side of the Skagafjörður, namely Hjaltadalur, Hóladalur, Kolbeinsdalur, Deildardalur and Unadalur (Fig. 1) were surveyed during four summer periods, but no appropriate glacial landforms to apply CRE dating methods were observed. No bedrock outcrops or outstanding blocks are evident in the relief of these tributary valleys, with the exception of those generated by postglacial rock avalanches. Sampling in Skagafjörður was carried out from

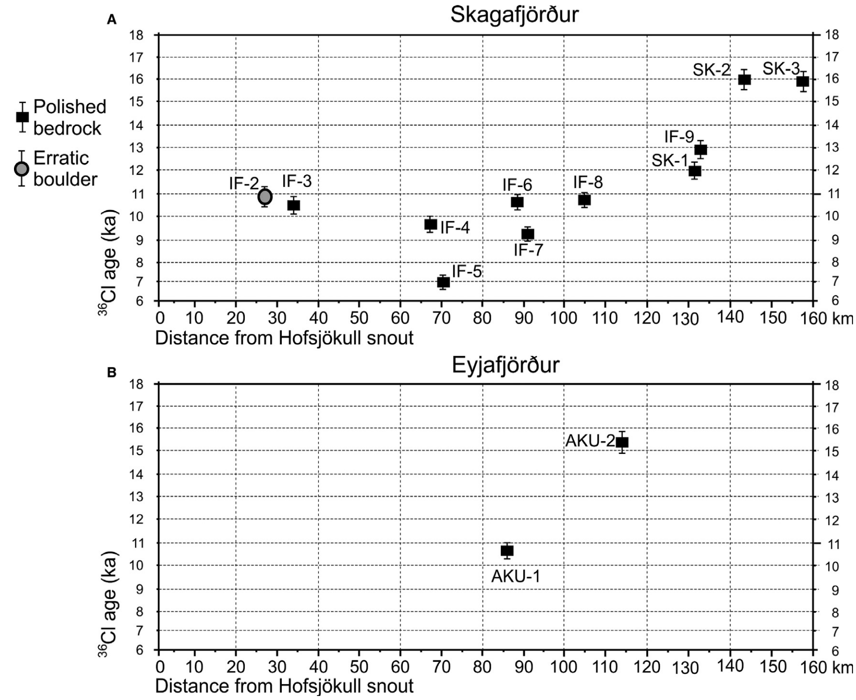

Fig. 3. Cosmogenic exposure dates and their distribution in relation to distance from the northern margin of the Hofsjökull icecap in Skagafjörður (A) and in Eyjafjörður (B).

the highlands to the north of the Hofsjökull icecap to the coastal areas and that also extended along the fjord shoreline onto the Skagi Peninsula, a total distance of 170 km (Figs 2, 3; Table 1). There are only 10 glacially polished bedrock outcrops suitable for sampling along this transect (Fig. 2). A sample was collected from each of these outcrops and another from an erratic boulder. This boulder was practically the only one found that showed signs of not having been altered by postglacial processes. Additional samples were collected from the Eyjafjörður to compare the deglaciation pattern of the two valleys: one located in the innermost part of the fjord and the other one approximately 30 km to the north from this position (Fig. 2). Samples were collected using a hammer and chisel from above 90 m a.s.l., as no higher shorelines above that altitude have been observed after deglaciation in the area (Pétursson *et al.* 2015), which could have disturbed the reception of cosmic radiation by the sampled surfaces.

The ^{36}Cl cosmogenic nuclide dating method was selected as the bedrock consists predominantly of basalt, and phenocrysts could not be identified in the samples. Laboratory procedures for ^{36}Cl analysis of whole rock samples (Zreda *et al.* 1999; Phillips 2003) were followed. Whole rock samples were crushed and pulverized using a roller grinder. The samples were then sieved to separate the 150–850 μm fraction and leached in deionized water and nitric acid (HNO_3) to remove atmospheric Cl. Next, the samples were dissolved in a hot mixture of hydrofluoric (HF) and nitric acids, and silver nitrate ($AgNO_3$) was added to the solution to precipitate silver chloride (AgCl). Sulphur was separated by adding barium nitrate ($Ba(NO_3)_2$) and consequently precipitating barium sulphate ($BaSO_4$). A spike of isotopically enriched ^{35}Cl was added during the dissolution process. This isotope dilution mass spectrometry enabled the natural ^{35}Cl/^{37}Cl ratio that is fixed at 3.1 to be changed and thus enabled the determination of the Cl concentration in the sample (Ivy-Ochs *et al.* 2004; Desilets *et al.* 2006) and of the sample Cl content from isotope dilution accelerator mass spectrometry (AMS) measurements. The AMS analysis of the ^{36}Cl/^{35}Cl and ^{37}Cl/^{35}Cl ratios in the AgCl targets was carried out at the PRIME Laboratory (Purdue University, USA). Aliquots of rock were analysed for major elements and trace elements at the Activation Laboratories (Ancaster, Canada). These analyses were necessary to calculate the relative contributions from the different composition-dependent ^{36}Cl production pathways (Schimmelpfennig *et al.* 2009).

The exposure ages were calculated using the spreadsheet for *in situ* ^{36}Cl exposure age calculations of Schimmelpfennig (2009) and Schimmelpfennig *et al.* (2009). We used the cosmogenic ^{36}Cl production rates for Ca spallation by Stone *et al.* (1996) (48.8±3.4 atoms ^{36}Cl (g Ca)$^{-1}$ a^{-1}), but we also used a production rate of 57.3±5.2 atoms ^{36}Cl (g Ca)$^{-1}$ a^{-1} (Licciardi *et al.* 2008); for K spallation by Schimmelpfennig *et al.* (2014) (148.1±7.8 atoms ^{36}Cl (g K)$^{-1}$ a^{-1}); for Ti spallation by Fink *et al.* (2000) (13±3 atoms ^{36}Cl (g Ti)$^{-1}$ a^{-1}); and for Fe spallation by Stone *et al.* (2005) (1.9 atoms ^{36}Cl (g Fe)$^{-1}$ a^{-1}). Finally, we also applied the epithermal

Table 1. Geographical locations of samples, topographical shielding factor and sample thickness of ^{36}Cl samples from Skagafjörður (northern Iceland).

Location	Sample ID	Sample type	Distance from the current Hofsjökull margin (km)	Latitude °N (DD)	Longitude °W (DD)	Elevation (m a.s.l.)	Topographical shielding factor	Thickness (cm)
Skagafjörður fjord and highlands	IF-2	Erratic boulder	26	65.1568	18.6599	704	0.91	2.0
	IF-3	Polished bedrock	36	65.1984	18.8703	645	1.00	4.0
	IF-4	Polished bedrock	69	65.4413	19.3074	148	0.98	4.0
	IF-5	Polished bedrock	70	65.4418	19.3075	152	0.97	4.0
	IF-6	Polished bedrock	89	65.5979	19.3332	122	0.97	2.5
	IF-7	Polished bedrock	92	65.6209	19.3479	139	1.00	3.0
	IF-8	Polished bedrock	104	65.7164	19.4914	129	1.00	2.5
	IF-9	Polished bedrock	135	65.9603	19.4862	138	0.98	2.5
Skagi Peninsula	SK-1	Polished bedrock	132	65.9047	19.8990	162	0.98	2.0
	SK-2	Polished bedrock	146	65.9971	19.9819	137	0.98	2.0
	SK-3	Polished bedrock	161	66.0969	20.2680	95	0.95	4.0
Eyjafjörður fjord	AKU-1	Polished bedrock	96	65.6448	18.1101	202	0.93	2.0
	AKU-2	Polished bedrock	124	65.8988	18.3137	210	0.87	5.0

neutron production rate from fast neutrons in the atmosphere at the land/atmosphere interface by Marrero *et al.* (2016b) (695±185 neutrons (g air)$^{-1}$ a^{-1}).

The *in situ* accumulation of ^{36}Cl depends also on various factors including latitude, elevation, sample thickness, surrounding topography and snow cover. The elevation-latitude scaling factors for nucleonic and muonic production were evaluated using the time-invariant 'St' model of Stone (2000). The topographical shielding factor was calculated using the Topographic Shielding Calculator v1.0 of CRONUS-Earth Project (2015).

The results obtained from the Schimmelpfennig *et al.* (2009) spreadsheet allowed us to change the Ca spallation production rate according to either that of Stone *et al.* (1996), derived from feldspar phenocrysts isolated from basaltic calibration samples, or Licciardi *et al.* (2008), derived from basaltic whole rock calibration samples. The comparison of the results from the two production rates are compared in Table 2. In both production rates, the composition of the calibration samples was dominated by Ca (Stone *et al.* 1996; Licciardi *et al.* 2008). The calibration sites of the Licciardi *et al.* (2008) production rate are located in southwestern Iceland (see also Licciardi *et al.* 2006), permanently affected by a low pressure cell (the 'Icelandic Low'). For this reason, Licciardi *et al.*'s production rate already accounts for this anomaly. Stone *et al.*'s production rate was calibrated in Tabernacle Hill (Utah, central-western United States) and hence needs to be corrected to be applied in Iceland.

Therefore, we recalculated the nucleonic and muonic production scaling factors applied to the Stone *et al.* (1996) production rate, considering the Stone (2000) scaling scheme and the local atmospheric pressure instead of the standard atmosphere (1013.25 hPa at sea level). In fact, Dunai (2010) advises including any long-term atmospheric pressure anomaly at least for Holocene exposures. Thus, we applied the local atmospheric pressure at sample locations derived from the 1006.9 hPa sea-level value (Akureyri meteorological station; Icelandic Meteorological Office 2018) assuming a linear variation of temperature with altitude. We discuss these ages throughout the text and the figures, with the analytical uncertainties, which are almost identical to those from Licciardi *et al.* (2008). In fact, the ages calculated with the spallation production rate for Ca = 48.8±3.4 atoms ^{36}Cl (g Ca)$^{-1}$ a^{-1} (Stone *et al.* 1996) and the scaling factor calculated with the local pressure for Iceland turn out to be 2.7–6.8% higher than those calculated with the spallation production rate for Ca = 57.3±5.2 atoms ^{36}Cl (g Ca)$^{-1}$ a^{-1} (Licciardi *et al.* 2008). The total uncertainty and internal uncertainty are included in Table 2. We contrast our ages with others mainly from lake and littoral sediments and tephras, dated by radiocarbon. The radiocarbon ages cited in the text were recalibrated with the latest calibration curve at the time of the cited publication.

Results

In total, 13 samples were collected on a north–south transect from the Hofsjökull icecap north to the Skagi Peninsula. The ages obtained from these 13 samples, using the proposed, method range from 15.9±1.2 to 6.8±0.5 ka as shown in Table 2 and Figs 2, 3.

Two samples, SK-2 located 146 km (137 m a.s.l.) and SK-3 located 161 km (95 m a.s.l.) north of the Hofsjökull icecap, were collected from polished ridges on the Skagi Peninsula and yielded ages of 15.9±1.2 and 15.8±1.1 ka, respectively (Figs 2, 3). Sample IF-9 was collected from a polished surface on the summit of the Þórðarhöfði cape, located on the eastern side of the Skagafjörður, approximately 104 km north of the Hofsjökull icecap, and yielded an age of 12.7±0.9 ka (Figs 2, 3, 4). On the opposite side of the fjord, sample SK-1 was collected from a polished ridge,

Table 2. ^{36}Cl exposure ages from Skagafjörður (northern Iceland). Spike enriched in ^{35}Cl (99.66%). Dilution spike $^{35}Cl/^{37}Cl$ ratio is 293.8±3.1. Ages are reported for 0 mm/ka erosion rate. Two age uncertainties are reported: the first figure shows analytical uncertainties only and the second (in parentheses) includes analytical uncertainties and production rate errors (one standard deviation). (*) is scaled total sample specific ^{36}Cl production rate without radiogenic (including ^{36}Cl production rate by spallation of Ca, K, Fe and Ti; by capture of epithermal neutrons; by capture of thermal neutrons; and by capture of slow negative muons). (1) = Data calculated from ^{36}Cl production rate from Ca spallation = 48.8±3.4 atoms $^{36}Cl\ (g\ Ca)^{-1}\ a^{-1}$ from Stone *et al.* (1996) and scaling factors calculated with local atmospheric pressure; (2) = Data calculated from ^{36}Cl production rate from Ca spallation s = 57.3±5.2 atoms $^{36}Cl\ (g\ Ca)^{-1}\ a^{-1}$ from Licciardi *et al.* (2008) and scaling factors calculated with no atmospheric pressure correction (Stone 2000). Analytical data of ^{36}Cl samples from Skagafjörður are provided in Table S1.

Location	Sample ID	Sample mass (g)	Mass of ^{35}Cl spike solution (mg)	Analytical stable isotope ratio ($^{35}Cl\ (^{35}Cl+^{37}Cl)^{-1}$)	Analytical $^{36}Cl/^{35}Cl$ ratio (10^{-15})	Measured ^{36}Cl concentration (10^4 atoms $^{36}Cl\ g^{-1}$)	^{36}Cl production rate (*) (atoms $^{36}Cl\ g^{-1}\ a^{-1}$)		Age calculated with spallation production rate for Ca = 48.8±3.4 atoms $^{36}Cl\ (g\ Ca)^{-1}\ a^{-1}$ (Stone *et al.* 1996) (ka) (1)	Age calculated with spallation production rate for Ca = 57.3±5.2 atoms $^{36}Cl\ (g\ Ca)^{-1}\ a^{-1}$ (Licciardi *et al.* 2008) (ka) (2)
							(1)	(2)		
Skagafjörður and highlands	IF-2	30.54	1.0084	6.67±0.0350	101.29±3.49	12.3±0.4	11.3	11.7	10.9±0.7 (1.0)	10.6±0.7 (1.0)
	IF-3	30.69	1.0102	7.57±0.0490	107.88±3.72	11.6±0.4	11.3	11.7	10.4±0.7 (0.9)	10.0±0.7 (1.0)
	IF-4	30.10	0.9960	13.20±0.0910	85.81±3.35	6.8±0.3	7.1	7.5	9.6±0.7 (0.9)	9.1±0.6 (0.9)
	IF-5	30.45	0.9906	14.23±0.2180	63.46±2.22	4.8±0.2	7.1	7.5	6.8±0.5 (0.6)	6.5±0.4 (0.6)
	IF-6	31.30	1.0031	27.78±0.2180	108.54±4.28	6.9±0.3	6.6	7.1	10.6±0.7 (1.0)	9.9±0.7 (1.0)
	IF-7	30.19	0.9807	19.55±0.0770	95.73±3.31	6.6±0.2	7.3	7.8	9.2±0.6 (0.8)	8.6±0.6 (0.9)
	IF-8	30.28	1.0096	10.31±0.0890	74.94±2.60	6.7±0.2	6.3	6.6	10.7±0.7 (0.9)	10.2±0.7 (1.0)
	IF-9	31.11	0.9953	5.19±0.0200	63.30±2.64	10.3±0.4	8.0	8.3	12.7±0.9 (1.3)	12.4±0.9 (1.3)
Skagi Peninsula	SK-1	30.14	0.9974	26.61±0.1740	113.21±5.86	7.5±0.4	6.3	6.8	11.9±0.9 (1.2)	11.2±0.9 (1.2)
	SK-2	30.01	0.9635	6.57±0.0630	92.52±4.60	11.1±0.6	7.0	7.3	15.9±1.2 (1.5)	15.2±1.2 (1.6)
	SK-3	30.09	1.0350	9.87±0.0870	98.62±3.75	9.3±0.4	5.9	6.2	15.8±1.1 (1.4)	15.0±1.0 (1.5)
Eyjafjörður	AKU-1	30.40	0.9296	8.22±0.0320	83.17±2.88	7.8±0.3	7.3	7.7	10.8±0.7 (1.0)	10.3±0.7 (1.0)
	AKU-2	30.50	0.9548	17.15±0.1250	133.46±4.36	9.2±0.3	6.1	6.5	15.3±1.0 (1.4)	14.4±1.0 (1.4)

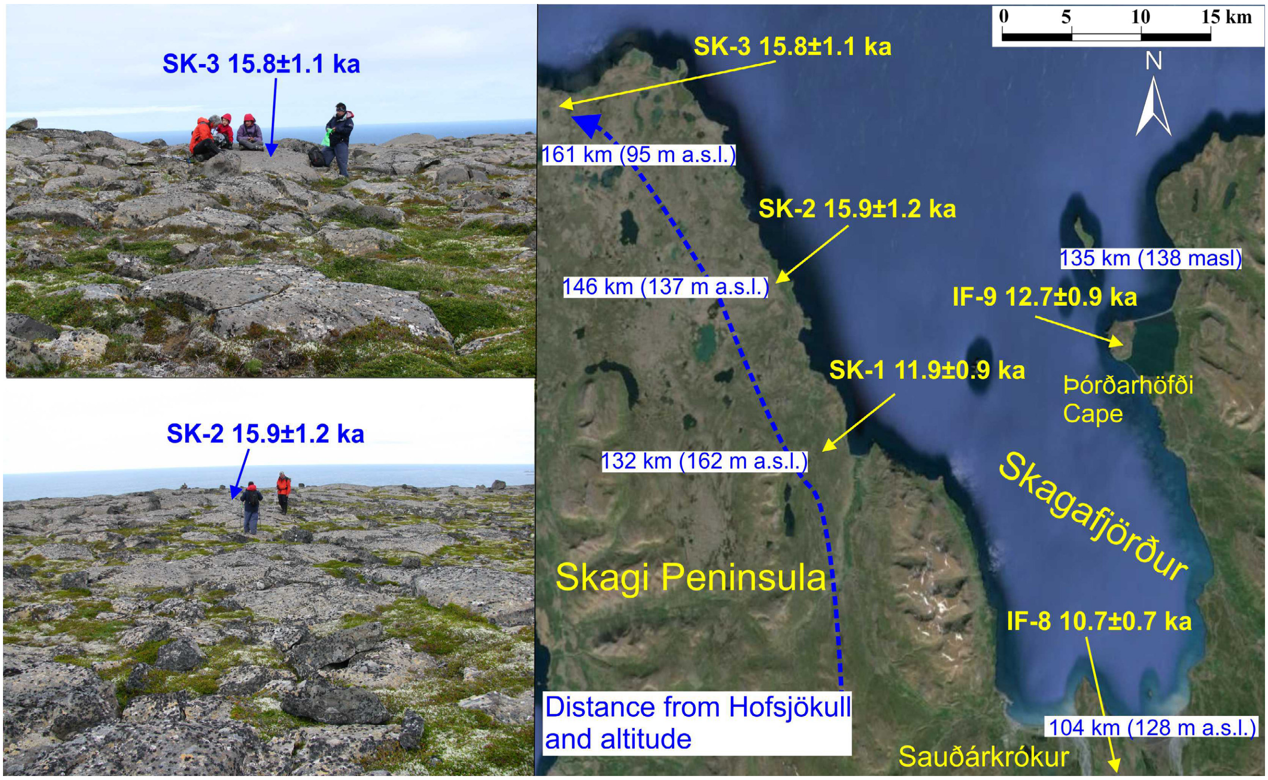

Fig. 4. Polished surfaces on the Skagi Peninsula (for location, see Fig. 2).

approximately 132 km and 162 m a.s.l., north of the Hofsjökull icecap, yielding an age of 11.9±0.9 ka.

Sample IF-8 was collected from an ice-polished ridge at 129 m a.s.l., by the present shoreline of the Skagafjörður fjord, approximately 92 km north of the Hofsjökull icecap, and yielded an age of 10.7±0.7 ka (Figs 2, 3).

Two other samples (IF-6 and IF-7) were collected from a polished lowland ridge at the bottom of the valley, approximately 90 km north of the Hofsjökull icecap at 139 and 122 m a.s.l., respectively. Sample IF-6 was collected on the top of the ridge and yielded an age (10.6±0.7 ka) similar to IF-2 and IF-3. Sample IF-7, collected approximately 20 m below the IF-6 site, yielded a younger age (9.2±0.6 ka) (Figs 2, 3, 5).

Two other samples (IF-4 and IF-5) were collected on a polished ridge at 150 m a.s.l., which divides the Vesturárdalur valley longitudinally, approximately 70 km north of the Hofsjökull icecap. These samples yielded ages of 9.6±0.7 and 6.8±0.5 ka, respectively (Figs 2, 3). The southernmost sample in the series (IF-2) was collected from an erratic boulder located 26 km from the Hofsjökull icecap at 704 m a.s.l., which yielded an age of 10.9±0.7 ka (Figs 2, 3, 6). Approximately 10 km to the north, at 645 m a.s.l., a sample (IF-3) was collected from a glacially polished bedrock outcrop, yielding a similar age (10.4±0.7 ka) to IF-2 (Figs 2, 3).

Two more samples were collected in the Eyjafjarðardalur valley. Sample AKU-1, collected from a polished ridge at the mouth of this valley (~96 km north of the Hofsjökull icecap and 202 m a.s.l.), yielded an age of 10.8±0.7 ka (Figs 2, 3B). Sample AKU-2, also collected from a polished ridge 30 km further north in the Eyjafjörður fjord and 210 m a.s.l. (Fig. 2), yielded an age of 15.3±1.0 ka (Fig. 7).

Discussion

Various glacigenic landforms, especially moraines, have been mapped and related to the deglaciation history of Iceland. However, only a few of these landforms have been dated, mainly due to the lack of organic material for radiocarbon dating, especially in northern Iceland (Kaldal & Víkingsson 1990; Norðdahl & Pétursson 2005; Norðdahl *et al.* 2008). The lake records are the most reliable, as they capture continuous sedimentation during deglaciation, and the boundary between glacial and non-glacial sediment can be dated by both radiocarbon and tephrochronology (Rundgren 1995; Rundgren & Ingólfsson 1999; Larsen *et al.* 2012; Striberger *et al.* 2012; Harning *et al.* 2016, 2018; Schomacker *et al.* 2016). ^{36}Cl CRE dating methodology has rarely been applied in Iceland to deglaciated surfaces (Principato *et al.* 2006; Brynjólfsson *et al.* 2015). Therefore, the possible application of the ^{36}Cl CRE dating method to reconstruct the deglaciation history of the Skagafjörður fjord and contrast with lake records from surrounding areas opens up new perspectives in our understanding of this deglaciation.

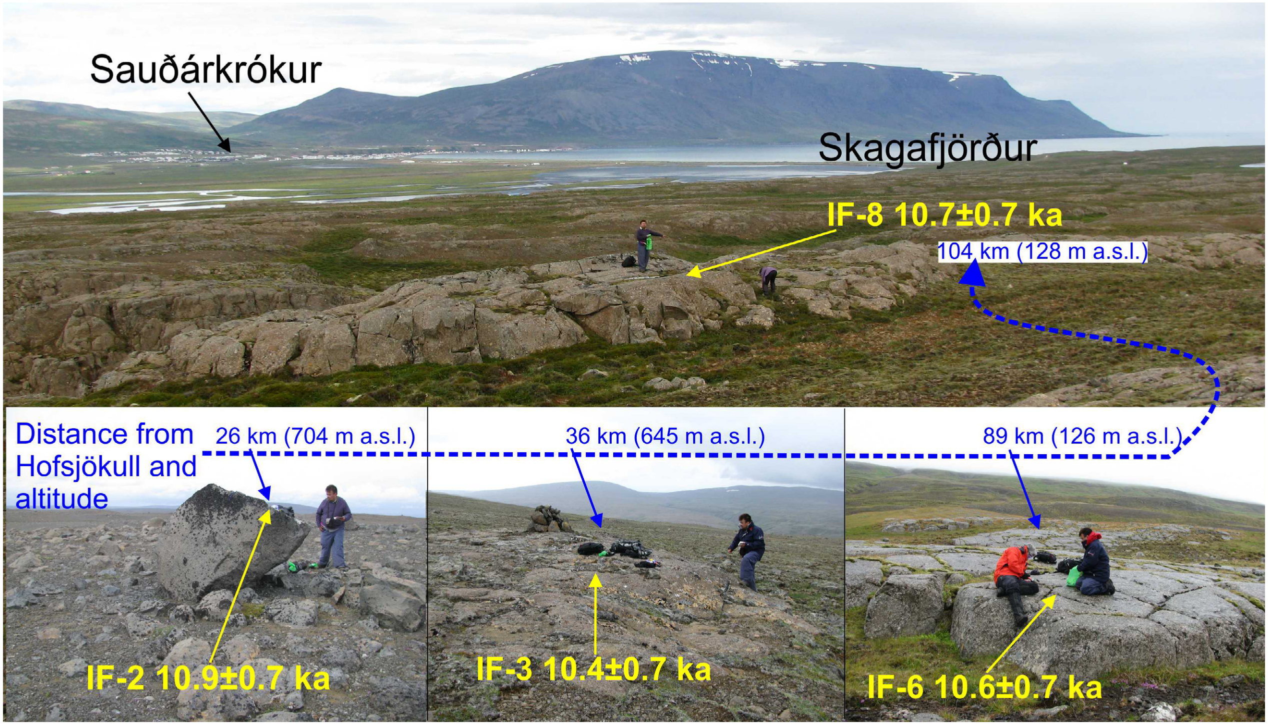

Fig. 5. Polished surfaces and an erratic boulder at various sites sampled within the Skagafjörður (for location, see Fig. 2).

During the Skagafjörður sampling, it was difficult to find landforms meeting the appropriate sampling criteria as many of the slopes and the bottom of the fjord are covered by soils, tephra and other sediments, which shield the glacial surfaces from cosmic radiation and thus disturb the ^{36}Cl surface production. In addition, much of the area has been affected by landslides, solifluction or floods, which has led to the destruction of glacial surfaces. No clear moraine formations have been observed, and practically all the protruding outcrops that conserve glacier polishing were sampled. Taking these circumstances into account, the only way to analyse the reliability of the results is to observe their geomorphological location within the sampled transect, from the highlands to the outermost part of the fjord.

The results obtained in most of the samples are properly sorted according to the geomorphological logic, with the younger ones being closer to the current glacier front. Our results obtained on the outermost part of the Skagi Peninsula (SK-2 and SK-3) suggest that it was deglaciated after 17 ka. Previous ages obtained by radiocarbon dating in marine sediments confirm that the glacier had retreated to the mid-shelf before 18.5 cal. ka BP (Andrews *et al.* 2000). Radiocarbon dates in a nearby lake (Rundgren 1995; Rundgren & Ingólfsson 1999) also demonstrated that the mouth of the fjord was deglaciated by 15 ka. The radiocarbon and cosmogenic ages suggest that the IIS limit was most likely situated at the mouth of the Skagafjörður, before 17–15 ka BP, following the rapid rise in sea level from *c.* 18.6 ka BP (Andrews *et al.* 2000). Previous data from northern Iceland, based on criteria of radiocarbon-dated lacustrine and littoral sediments and tephrochronology dating, also support this ice-margin location (Kaldal & Víkingsson 1990; Eiríksson *et al.* 2000; Norðdahl & Einarsson 2001; Norðdahl *et al.* 2008). Nevertheless, the limited number of samples and the wide range of uncertainty obtained in this work do not allow the exact location of the ice margin to be defined. The 17–15 ka BP period is coincident with the Oldest Dryas (stadial GS-2a, between 17.5 and 15 ka BP). There are no existing references that can clarify the potential impact of cooling on the Icelandic glaciers in this period, but an overall trend of continental deglaciation following the end of the LGM was abruptly interrupted in many parts of Europe during the Oldest Dryas (Eiríksson *et al.* 2000; Ivy-Ochs 2015; Palacios *et al.* 2017a). During this period, there was a significant reduction in the meridional overturning circulation (McManus *et al.* 2004). In addition to these climatic proxies, other factors must be taken into account in stabilizing the ice sheet between 17 and 15 ka BP, such as topography and the role that the coastline at that time may have played in its stabilization, as happened with other ice sheets (Stokes *et al.* 2014). In fact, several studies have suggested an IIS stepwise retreat (see review in Geirsdóttir *et al.* 2009), and recent work in the Vestfirðir Peninsula in northwest Iceland also demonstrates that the

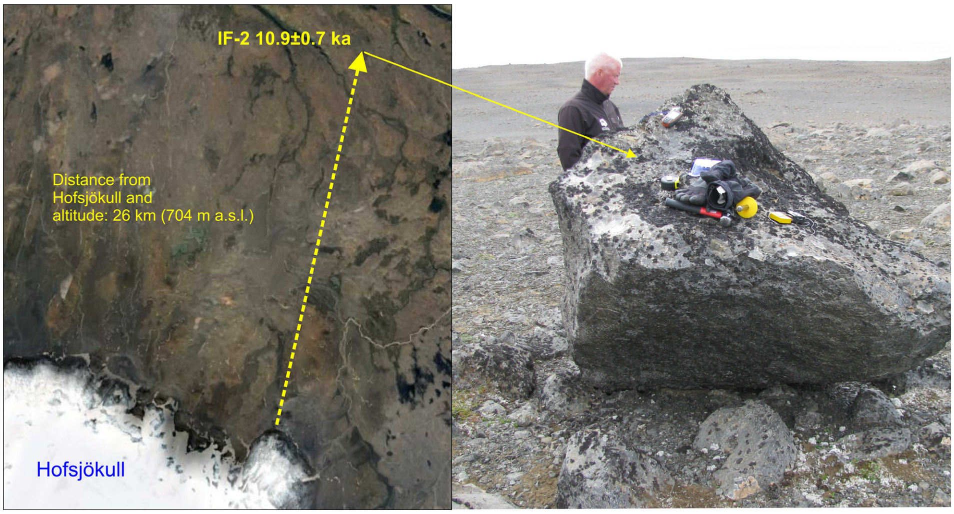

Fig. 6. Erratic boulder located 26 km north of the northern margin of the Hofsjökull icecap at 704 m a.s.l. (for location, see Fig. 2).

IIS response to climatic forcing is not linear (Brynjólfsson *et al.* 2015). Patton *et al.* (2017) relate the IIS retreat patterns in the north not only to the climate evolution but also to basal topography and sea-level changes.

The destabilization of the IIS itself, due to a rapid sea-level rise, may have caused its collapse with a rapid retreat just around 15 ka (see synthesis in Norðdahl & Ingólfsson 2015; Pétursson *et al.* 2015; Patton *et al.* 2017). Ingólfsson & Norðdahl (2001) first suggested that this global sea-level rise was caused by the intensive melting of the northern ice sheets and provoked massive calving in the IIS and its collapse. Our two samples located further to the south on each side of the fjord about 130 km from the Hofsjökull icecap (IF-9 and SK-1) give 12.7±0.9 and 11.9±0.9 ka, respectively. These ages would place the retreat of the ice margin from the central part of the fjord around 15–11.5 ka BP and support a rapid and intensive retreat after 15 ka BP (around 25 km), but once again, the small number of samples and the range of uncertainty do not give us the exact location of the ice margin. These ages are also coincident with the deglaciation of the Torfadalsvatn dated by radiocarbon and the expansion of shrub tundra vegetation in the Skagi Peninsula according to pollen collected in the lake sediments between 13.9 and 12.9 ka BP, with bioclimatic conditions similar to the present (Rundgren 1995, 1999; Rundgren & Ingólfsson 1999). Extensive subaerial lava flows to the east of Tröllaskagi confirm the large deglaciation areas in the north of Iceland during this period (Norðdahl & Pétursson 2005). Norðdahl *et al.* (2008) and Pétursson *et al.* (2015) modelled the glacier limit during the Allerød interstadial and during the Younger Dryas maximum in Skagafjörður, following the radiocarbon ages, mollusc fauna and the shoreline distribution criteria. According to these models, the SK-1 and IF-9 samples are within the Bølling deglaciation areas but just outside of the Younger Dryas advance. A similar situation for our samples is obtained from the more complex numerical model proposed by Patton *et al.* (2017) (Fig. 2).

The results of our samples do not allow us to define the position of the ice margin in Skagafjörður during the Younger Dryas period that led to major biosphere degradation in the north of Iceland (Rundgren 1995, 1999) and sea-ice conditions (Xiao *et al.* 2017). Previous publications have demonstrated that the glaciers reoccupied valleys and reinvaded the fjords to their middle sector during this period in northern Iceland, leaving terminal moraines on the bottoms, according to shoreline distribution radiocarbon dating (Pétursson *et al.* 2015). The very few radiocarbon and tephrochronology dates that exist at present along the shores of northern Iceland fjords show that they were already ice-free at the end of the Younger Dryas period (Norðdahl & Haflidason 1992; Ingólfsson *et al.* 1997; Eiríksson *et al.* 2000; Pétursson *et al.* 2015).

The ages of samples IF-2 (10.9±0.7 ka), IF-3 (10.4±0.7 ka), IF-4 (9.6±0.7ka), IF-6 (10.6± 0.7ka), IF-7 (9.2±0.6 ka) and IF-8 (10.7±0.7 ka) are all statistically the same considering their range of uncertainty (Figs 2, 3, 5, 8). Anomalously, the age of IF-5 (6.8±0.5 ka) is younger than the samples collected closer to the present glacier. We consider the sample IF-5 an outlier, which could be explained by previous burial. The other six samples located between the present shoreline and the Hofsjökull icecap (IF-2, IF-3, IF-4, IF-6, IF-7 and IF-8) all give similar ages and show an arithmetic mean of 10.2 ka. The absence of the Skogar-Vedde Tephra, dated to 12.1 cal. ka BP, and the presence of the Saksunarvatn tephra, dated to 10.3 cal. ka BP (Stötter *et al.* 1999; Caseldine *et al.* 2006; Rasmussen *et al.* 2006), suggest that this area was most probably

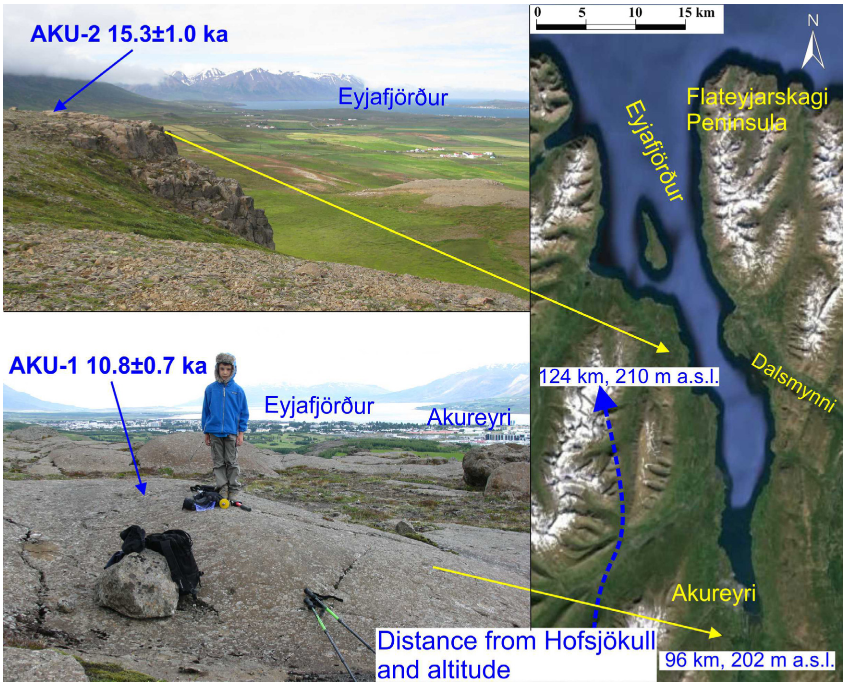

Fig. 7. Polished surfaces at the Eyjafjörður, and CRE samples taken from the head and middle of the fjord.

deglaciated after ~12.1 ka and before ~10.3 ka. Further evidence based on truncated raised shorelines and radiocarbon dating suggests that during the early Preboreal, a new re-advance occurred in northern Iceland during which the glacier margins advanced into the fjords (Ingólfsson *et al.* 1997; Norðdahl & Einarsson

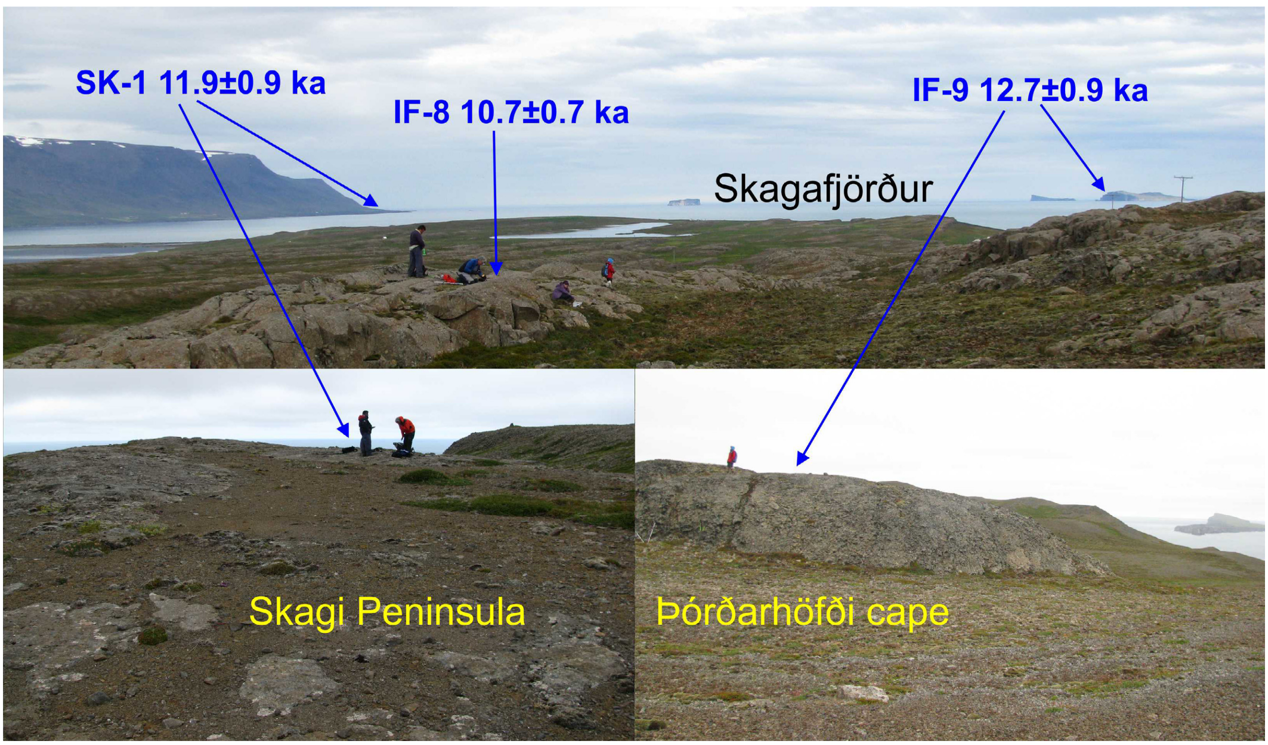

Fig. 8. Polished surfaces at the present coastline in Skagafjörður, close to Sauðárkrókur village and from the Skagi Peninsula to the west and the Þórðarhöfði semi-island to the east (for location, see Fig. 2).

2001; Norðdahl & Pétursson 2005; Pétursson *et al.* 2015). The results obtained for the age distribution of our samples between the ages of two well-known tephra deposits inland from the present coastline agree with early Preboreal truncated raised shorelines and radiocarbon dating (Ingólfsson *et al.* 1997). According to this information, the ice margin in Skagafjörður was located not far inland from the present shoreline during the early Preboreal period, when the Skagi Peninsula was mostly ice-free (Figs 1, 2).

The similar ages of the six samples, with an average of *c.* 10.2 ka, and the previous criteria, suggest abrupt deglaciation at the end of the early Preboreal period, of more than 90 km of glacier retreat in 1000 years, which is supported by previous deglaciation research in Iceland (Kaldal & Víkingsson 1990; Andrews *et al.* 2000; Norðdahl & Einarsson 2001; Geirsdóttir *et al.* 2002, 2009; Larsen *et al.* 2012; Pétursson *et al.* 2015). This rapid deglaciation process after the Preboreal agrees with the latest models developed of deglaciation in northern Iceland (Pétursson *et al.* 2015) (Fig. 2). Recently, similar results have been obtained from the study of the deglaciation of Drangajökull, in northwest Iceland, where radiocarbon dating of lake sediments demonstrates a rapid deglaciation of the north side of this icecap in the Early Holocene, reaching its present extent at 10.3 cal. ka BP (Harning *et al.* 2016).

It is possible that, in a complex glacier fluctuation scenario, there might be inherited ^{36}Cl on glacially polished bedrock outcrops, where a glacial advance may not eliminate the cosmogenic heritage accumulated previously. However, the chronological order of the samples most distant from those closest to the current glacier front is respected. In addition, the clustering of the ages of samples within the logical period limited by tephrochronology and by the radiocarbon dating of the lakes and littoral sediments appears to defend the idea of a lack of or limited cosmogenic inheritance in our samples.

The preliminary results from the only two samples in the Eyjafjörður agree with previous research on the deglaciation of this fjord. The obstruction of the Fnjóskadalur valley, tributary of the Eyjafjörður, started at 17 ka BP and ended by 12 ka BP, and it is estimated that the glacier margin was located at the head of the Eyjafjörður at 11 ka BP (Einarsson 1967, 1973; Norðdahl 1991). The samples AKU-2 (15.3±1.0) and AKU-1 (10.8±0.7 ka) are located in Eyjafjörður, on the opposite side of the mouth of the Fnjóskadalur and at the head of the fjord, respectively (Fig. 2). The results of these exploratory samples support the application of ^{36}Cl CRE methods in future research in Eyjafjörður.

Conclusions

The results of this work improve the knowledge of the glacial evolution of the Skagafjörður in northern Iceland by applying cosmogenic dating methods. The results agree with previous research based on other dating methods and with indicators of other palaeoclimatic proxies. The ages of samples SK-2 and SK-3, complemented with radiocarbon ages in lakes and marine sediments, suggest that the ice margin of the IIS was located at the outermost parts of the Skagafjörður at approximately 17 ka BP. The samples IF-8, IF-7, radiocarbon ages in lakes and littoral sediments and shoreline distribution confirm the effects of glacial collapse around 15 ka BP in Skagafjörður. Similar ages from six samples along a 70-km transect, together with tephrochronology and radiocarbon dating of lake sediments and truncated shoreline distribution, suggest that the ice margin was situated at the present coastline approximately at 11 ka BP and then retreated rapidly after the Preboreal period. Between approximately 11 and 10 ka BP, the glacier retreated proximal to the current margin of the Hofsjökull icecap in the central highland. These results agree with previous studies of the deglaciation history of northern Iceland and demonstrate the potential of the CRE dating method using ^{36}Cl to improve and define the glacial history of Iceland. The lack of contradictions between the methods applied in previous studies and the present study reinforces the models proposed previously. The inherent limitation in the lack of well-preserved glacial landforms, a characteristic feature of the Tröllaskagi Peninsula, should not discourage the application of CRE dating; on the contrary, an adequate sampling strategy based on geomorphological criteria and exhaustive fieldwork should offset this limitation.

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Supporting Information

Additional Supporting Information may be found in the online version of this article at http://www.boreas.dk.

Table S1. Analytical data of ^{36}Cl samples from Skagafjörður (northern Iceland).