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## Volcanoes and climate: the triggering of preboreal Jökulhlaups in Iceland

Van Vliet-Lanoë Brigitte <sup>1,\*</sup>, Knudsen Oskar <sup>2</sup>, Guðmundsson Agust <sup>3</sup>, Guillou Hervé <sup>4</sup>, Chazot Gilles <sup>1</sup>, Langlade Jessica <sup>5</sup>, Liorzou Celine <sup>1</sup>, Nonnotte Philippe <sup>1</sup>

<sup>1</sup> Géosciences Ocean, UMR 6538 Brest University, CNRS, ue@b, IUEM, Pl.N.Copernic, 29280, Plouzané, France

<sup>2</sup> Klettur Consulting Engineers, Bildshofda 12, 102, Reykjavik, Iceland

<sup>3</sup> Jarðfræðistofan ehf, 200, Hafnarfjörður, Iceland

<sup>4</sup> CNRS-CEA, UMR 8212 LSCE, Domaines CNRS, Bât. 12, Av. de la terrasse, 91198, Gif/Yvette, France

<sup>5</sup> Microsonde de l'Ouest, Ifremer-Centre de Bretagne Technopole Brest Iroise, BP 70, 29280, Plouzané, France

\* Corresponding author : Brigitte Van Vliet-Lanoë, email address : [brigitte.vanvlietlanoë@univ-brest.fr](mailto:brigitte.vanvlietlanoë@univ-brest.fr)

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### Abstract :

The Early Holocene (12–8.2 cal ka) deglaciation and pulsed warming was associated in Iceland with two major generations of jökulhlaups around the Vatna ice-cap (Vatnajökull), at ca 11.4–11.2 cal ka and ca 10.4–9.9 cal ka, and major tephra emissions from the Grímsvötn and Bárðarbunga subglacial volcanoes. The earliest flood events were recorded inland during the Middle Younger Dryas and their deposits were overlain by the Early Preboreal Vedde Ash (11.8 cal ka). The first Holocene flood events (ca 11.4–11.2 cal ka) are issued from a glacial advance. The second, and major, set of floods was partly driven by the Erdalen cold events and advances (10.1–9.7 10Be ka) initially issued from the Bárðarbunga (10.4, 10.1–9.9 ka) and Grímsvötn volcanoes (Saksunarvatn tephra complex, ca. 10.2–9.9 cal ka). These floods were also fed by the residual glacio-isostatic depressions below the Vatnajökull that enabled the storage of meltwaters in large subglacial lakes or aquifers until ca. 9.3 cal ka. This storage was enhanced by ice-damming and permafrost, especially during the twinned Erdalen events. Due to the glacio-isostatic rebound, the general slope was nearly flat, and the valley was partly filled with sediments until ca 10.8 cal ka. Temporary lacustrine deposits in this valley resulted from the very broad splay of waters as for the ca 11.2 cal ka and ca 10.1–9.9 cal ka flood, due to regional permafrost. These floods had a potential duration of several months as they were mostly fed by climate-driven meltwater. The maximal volume evacuated by these events did not greatly exceed  $1 \times 10^6 \text{ m}^3 \text{ s}^{-1}$  from the flood-affected transverse profile of the valleys that remain partly filled with sediments.

**Keywords :** Holocene, Deglaciation, Iceland, Geomorphology, Glacial, Flood, Sedimentology, Tephra, Glacio-isostatic rebound, Permafrost, Saksunarvatn event, Askja S

## 47 1. Introduction

48 Large-scale outburst flows were a common phenomenon that accompanied the termination of the last  
49 glaciation in non-volcanic regions (e.g. Carling 2013). Meltwater storage in or at the surface of a  
50 glacier favours jökulhlaup occurrence (Rushmer 2006; Carrivick et al. 2009). Jökulhlaups may be also  
51 triggered by volcanism. As volcanism and melting are enhanced by deglaciation (Jull and McKenzie,  
52 1996; Slater et al. 1998; MacLennan et al. 2002; Sinton et al. 2005), these events are suspected  
53 having promoted both jökulhlaups and more explosive volcanism (Höskuldsson et al. 2006; Van Vliet-  
54 Lanoë et al. 2007; Carrivick et al. 2009).

55

56 The analysis of the Early Holocene (ICS 2018) deglaciation in Iceland is a favourable place to  
57 understand their interrelations due to the radial organization of the drainage around the Vatnajökull  
58 (Fig. 1), the largest Late Holocene ice-cap.

59 Major active volcanic systems, with an extensive morphological impact, exist beneath the Vatnajökull  
60 (Fig. 1B) in the form of the Grímsvötn and Bárðarbunga Volcanoes (Björnsson 2017). The Vatnajökull  
61 rests on porous volcanogenic sediments on the western side of the Grímsvötn caldera, and on the  
62 impermeable bedrock below the Brúarjökull glacier and to the east (Flowers et al. 2003). Subglacial  
63 eruptions, associated with tephra outfalls from the Grímsvötn calderas, were responsible for numerous  
64 hazardous jökulhlaups to the north and south (Thorarinsson 1974; Björnsson 1992; Thordarson and  
65 Larsen 2007). The relationship between jökulhlaups and the last deglaciation is not analysed in  
66 Iceland, even though most of the terminal moraines in the south have been carefully described by  
67 Kaldal and Víkingsson (1990) and a chronology of jökulhaup events has been extracted from the  
68 lacustrine record of the deglaciation by Geirsdóttir et al. (2000).

69

70 The interval analysed in this paper covers the Younger Dryas to the Holocene Thermal Optimum (12.8  
71 - 8 cal ka). The period from Termination Ib (11.8 cal ka, ICS 2018) to the Thermal Optimum (8–6 cal  
72 ka) was characterised in the south of Iceland (Fig. 1a) by a series of jökulhlaups (Geirsdóttir et al.  
73 2000). In the southernmost Iceland, the first major Early Holocene eruption after Termination Ib and  
74 the deposition of the Vedde Ash was the Saksunarvatn Tephra (Mangerud et al. 1986). This was  
75 issued from the Grímsvötn Volcano at ca. 10.28 cal ka BP (Tephabase), and was a potential trigger of  
76 major jökulhlaups. This volcano generated a large jökulhlaup via the Skeiðarárjökull Glacier (Fig.1B),

77 which delivered turbidites to the Mýrdalsjökull Submarine Canyon at ca 10.15 cal ka (Fig.1B, Lacasse  
78 et al. 1998). The source of the largest Holocene jökulhlaup events in the north is commonly attributed  
79 to the Grímsvötn and Kverkfjöll calderas (Sæmundsson 1973; Carrivick 2007; Carrivick et al. 2009)  
80 with another source in the Bárðarbunga subglacial caldera (Björnsson and Einarsson 1991; Waitt  
81 2002; Fig. 1A, 2, 3A). These volcanoes were also responsible for jökulhlaups during the Eemian and  
82 the Holocene in the south, e.g. along the Jökulsá á Fjöllum and Þjorsá–Ytrí Rangá Rivers (Van Vliet-  
83 Lanoë et al. 2018).

84  
85 In the North, most studies up to now have been focused on the main channel of the Jökulsá á Fjöllum  
86 (Fig.1a) and its morphologies (e.g. Middle Holocene floods: Kirkbride et al. 2006; synthesis in Baynes  
87 et al. 2015), but little attention has been paid to traces existing outside the canyon, especially on the  
88 nearby plateau, or upstream of adjacent dry valleys. This valley experienced multiple jökulhlaups of  
89 varying sizes following Termination Ib, as evidenced by a sequence of prehistoric flood deposits in the  
90 river canyon (e.g. Björnsson and Kristmannsdóttir, 1984) and on the upper terraces (this paper).

91  
92 In this paper, we attempt to synthesize the existing published data, complemented with new  
93 observations that constrain the timing of the Holocene deglaciation and the early jökulhlaup events. It  
94 will be related to the Preboreal glacial advances and retreats, and their respective dynamic conditions.  
95 We first searched for key sedimentary sections with available chronological markers constraining  
96 intimate as tephra the deglaciation history inland and on the continental shelf. The gathered data,  
97 including the permafrost information, was correlated with the climatic framework of the Early Holocene.  
98 Following this, we have tried to constrain the impact of the Early Holocene deglaciation on the  
99 Vatnajökull hydrological system and jökulhlaup recurrences. Control of glacio-isostasy and its impact  
100 on the volcanism was analysed, with a special focus on the complex Saksunarvatn tephra.

101

## 102 **2. Methodology**

103 Most of this work was performed on digital satellite images and aerial photos black and white and  
104 colour provided mostly by Landmælingar Íslands (National Land Survey of Iceland) and occasionally  
105 Google Earth (GE) for various dates, complemented by the photographic interpretations of Kaldal and  
106 Víkingsson (1990), Sæmundsson et al. (2012) and Sigurgeirsson et al. (2015). In terms of glaciation,

107 attention was primarily paid to fluted morphologies, “pitted” or hummocky moraines, surge moraines  
108 and normal terminal moraines (Fig. 1B), in an attempt to place them within a chronological framework.  
109 The development of specific periglacial morphologies (pattern grounds, rock glaciers) was used to  
110 delimit the glacier boundaries (Guðmundsson 2000). Erosional and sedimentary morphologies  
111 induced by flood activities (e.g. Maizel 2009; Baynes et al. 2015) were analysed using the same  
112 images as for the glacial morphology: channeled scablands with scourings, canyons and plunge  
113 basins, dry perched valleys, dismantled surfaces of lava flows, potholes, pillars and rough surfaces on  
114 consolidated sediments, streamlined residual landforms, fluvial dunes. These morphologies were  
115 used to map the maximal flood extent (Fig. 1A).

116  
117 The tephra record used herein (Table 1) refers to the published data from Icelandic lakes and soils,  
118 and the marine- and ice-core records (Greenland and Iceland shelf; Wohlfarth et al. 2006; Koren et al.  
119 2008; Lind and Wastegård 2011; Óladóttir et al. 2011; Larsen et al. 2012; Voelker and Hafliðason  
120 2015; Guðmundsdóttir et al. 2016). Tephra and sediments were analysed in bulk by ICP–MS AES to  
121 determine their sources, with single-grain microprobe and laser ICP–MS being used to determine the  
122 source of rhyolitic tephra. Tephra-derived ages are given in c. X cal ka (BP), to avoid repetition of error  
123 margins, most of them being established on age models.

124

### 125 **3. Paleoclimate and volcanic context**

#### 126 *3.1. Deglaciation context and climate*

127 The onset of the Younger Dryas (YD, 12.9 cal ka) was driven both by Oceanic circulation perturbation  
128 (Condron and Winsor 2012) and by a major solar minimum (Andresen et al. 2000; Goslar et al 2000).  
129 Since the YD, the Northern Hemisphere deglaciation has proceeded with brief, discrete cooling  
130 events, at c. 11.3–11.1 cal ka, 10.3, 9.3 and 8.2 cal ka (Fig.3), associated with the advance of glaciers  
131 (North Greenland Ice Core Project [NGRIP], 2004). Four episodes of glacier advance fit the cooling  
132 events that have been recognised in the ocean (Bond et al. 2001), in Scandinavia, the Færoe Islands and  
133 Greenland (Rasmussen et al. 2006; Matthews et al. 2008; Geirsdóttir et al. 2009; Kobashi et al. 2017):  
134 the Preboreal oscillation (11.3–11.1 cal ka; the 10.3–10.2 cal ka / 10.1–9.7 <sup>10</sup>Be ka cooling or Erdalen  
135 cooling events (Fig.3) allegedly driven by volcanic activity (Linde and Wastegård, 2011; Rasmussen  
136 et al. 2014) and a c. 9.3 cal ka cooling linked to a solar low that lasted for over 50 years (<sup>10</sup>Be; Björck

137 et al. 1997; Bos et al. 2007). The youngest, the classical “8.2 cal ka” cooling event was the longest  
138 and coolest lasting for the next two to four centuries (Matero et al. 2017). This cooling seems triggered  
139 by solar activity (Stuiver and Braziunas 1988; Bond et al. 2001; Vonmoos et al. 2006), by a major  
140 meltwater pulse, issued from the Laurentide Ice Sheet collapse (Matero et al. 2017) with some impact  
141 of volcanic activity (Kobashi et al. 2017).

142 The Early Holocene pulsed warming was associated with the restoration of the North Atlantic and  
143 Irminger Currents, which increased the temperature and precipitation resulting in an accelerated ice-  
144 sheet retreat in Iceland (Jennings et al. 2000). The main Icelandic ice-sheet retreated rapidly across  
145 the highlands at that time (Geirsdóttir et al. 2009). After the “8.2 cal ka” cooling, the glaciers almost  
146 disappeared in Iceland. The thermal optimum was apparently ca. 2°C – 3°C higher than today  
147 (Andresen et al. 2007; Geirsdóttir et al. 2009; Langdon et al. 2010).

148 The Younger Dryas cooling has a duration of 1 ka (Condrón and Winsor 2012), long enough for  
149 promoting glacier advances as demonstrated for the North by the <sup>36</sup>Cl dating of rockglaciers and  
150 coastal deposits (Andrès et al. 2016; Andrès et al. 2019), but not for restoring the ice-sheet, as has  
151 been commonly proposed (Norðdhal and Petursson, 2005; Patton et al. 2015) on the base of  
152 radiocarbon dated coastal deposits. This restricted extent is demonstrated in the west of the island by  
153 the limited glacioisostatic rebound of the YD (Brader et al. 2015). The YD seems to have been  
154 associated with the spreading of ice-lobes in the south (Geirsdóttir et al. 2009; Van Vliet-Lanoë et al.  
155 2018) although the tidal glacier or ice-streams, calving into the coastal bays, occupied the valleys in  
156 the west and the north (Fig. 1; Jennings et al. 2007; Geirsdóttir et al. 2009; Andrès et al. 2019).

157 The durations of the pulsed Holocene cooling events were relatively short, with ca 150 years for the  
158 “10.3 cal ka” twinned events, 50 to 200 years for the “9.3 cal ka” event and 200 to 400 years for the  
159 “8.2 cal ka” event (Rasmussen et al. 2006; Matero et al. 2017). They are considered to have been  
160 insufficient to have favoured ice-sheet development or major valley glacier advances during the  
161 Holocene. However, they were sufficient enough to have promoted glacier surges (Meier and Post,  
162 1969).

163

### 164 *3.2 Glaciers and permafrost*

165

### 166 *3.3 Subglacial volcanoes*

167 The Grímsvötn volcanic system is believed to have been the most active volcano during the Holocene  
168 (Björnsson 1974; Larsen et al. 1998; Óladóttir et al. 2011), with an eruption frequency quite constant  
169 through time, and higher activity cycles, 60 to 80 years long. This volcano appears to have two  
170 reservoirs – a deeper one at 15 km depth and a shallow one at 3 km depth (Reverso et al. 2014)  
171 probably very reactive to deglaciation events via unloading (Hoskuldsson et al. 2006). Its large  
172 calderas, 3.6 km<sup>3</sup> in volume, is a major zone of meltwater storage, and a source for jökulhlaups,  
173 especially along the southern coast (Björnsson 2002). The detailed geochemistry of the Grímsvötn  
174 eruptions (single-grain laser ICP–MS) is not suitable for dating, as very few variations were recorded  
175 during the Holocene (Óladóttir et al. 2014; Thordarson 2014).

176  
177 The Bárðarbunga Volcano, located at the NW edge of Vatnajökull, forms a wide and elevated caldera  
178 (1850 masl), completely glacier covered and is located along the extensive Veiðivötn fissure system,  
179 which parallels the Grímsvötn fissure system. The reservoir is located at a depth of 12 km  
180 (Guðmundsson et al. 2016). During the Holocene, the eruption frequency was five eruptions per  
181 century (Óladóttir et al. 2011). Its wide caldera, 14.4 km<sup>3</sup> in volume has been a major source for  
182 jökulhlaups, especially in the southern embayment (Björnsson 2002). The Bárðarbunga Volcano had  
183 probably already been emerging from the ice-sheet since Bölling times, based on the tephra record  
184 (Table 1), producing five significant eruptions, at c.11.35, 11 – 10.8, 10.4, 10.1 and 9.9 cal ka  
185 (Óladóttir et al. 2011).

186

### 187 *3.4. The Askja rhyolitic tephra*

188 Questions exist in the literature concerning the Askja 10/S Tephra, first described in Iceland and  
189 observed in Eyafjörður above the Vedde Ash (Sigurðsson and Sparks 1978; Askja 10) and below the  
190 Reitsvík 8 Tephra (c.10.2 cal ka; Larsen et al. 2002; Guðmundstóttir et al. 2016). The classic Askja S  
191 Tephra age is now 10.8 cal ka (Wolfarth et al. 2006; Bronk et al. 2015; Table 1). In the Early Holocene,  
192 the Askja Volcano emitted probably at least five rhyolitic tephra, at c. 9.4 cal ka, c. 10.4 cal ka (10.5 –  
193 10.35 cal ka; Lind and Wastegård 2011), 10.8 cal ka (Askja S), c. 11.5 cal ka (Ott et al. 2016), and c.  
194 12.5 and 12.9 cal ka (Jones et al. 2017). The Krapla 10 Tephra (Sæmundsson et al. 2012), very likely  
195 equates to the Reitsvík 8 Tephra (10.2 cal ka; Guðmundsdóttir et al. 2016), based on the available  
196 geochemical analysis (Krafla Phase 3; Jonásson 1994) but is hard to distinguish.

197

198 *3.5 The Saksunarvatn basaltic tephra*

199 The most common and thick tephra during the deglaciation is the Saksunarvatn Tephra (ca. 10.3 cal  
200 ka; Mangerud et al. 1986; Fig.3). Dates for this very large deposit ( $> 15 \text{ km}^3$ ) cover several eruptive  
201 events. In the Icelandic lake cores, there are up to six Early Holocene tephra layers with a Grímsvötn  
202 chemical composition (Jóhannsdóttir, 2007; Fig.2A). The Saksunarvatn Tephra has not been recorded  
203 in the Jökulsá á Fjöllum, although it is present in the Eyafjörður Fjord (Fig.1A; Larsen and Eiríksson,  
204 2007) and on the northern shelf (Eiríksson et al. 2000). In Iceland and Norway (Guðmundsdóttir et al.  
205 2016; Andresen et al. 2007; Birks et al. 1996; Grönvold et al. 1995; Fig 1a), Saksunarvatn/Grímsvötn-  
206 type ash sedimentation began at c. 10.24 cal ka and continued up to 9.9 cal ka. In marine cores, four  
207 aerial Grímsvötn eruptions were recorded from 10.30 to 9.85 cal ka (marine age-model; Lacasse et al.  
208 1998; Guðmundsdóttir et al. 2012; Voelker and Hafliðason 2015). In Greenland ice cores, the ranges  
209 are from 10.24 - 10.12 ka ice-core ka (Rasmussen et al. 2006). It thus seems that the major Grímsvötn  
210 eruption took place at ca 10.24 cal ka, followed by serial eruptions until 10.12 cal ka. The Mýrdal  
211 Canyon turbidite, identified in the south (10.15 cal ka) possibly corresponded to a large subglacial  
212 flood.

213

214 *3.6 Holocene and historical jökulhlaups*

215 The Jökulsá á Fjöllum River is incised by many dry channels issued from jökulhlaups, mapped by  
216 Sæmundsson et al. 2012). Most of the studies were conducted along the canyon. From 1600 until  
217 1934 AD, one large jökulhlaup per decade was issued from the Grímsvötn volcanic system into the  
218 Jökulsá á Fjöllum River (Björnsson and Kristmannsdóttir 1984), fitting more or less with the basic  
219 decennial frequency of eruptions (Óladóttir et al. 2011). Some Holocene events have been detected in  
220 the same river between 7.10 and 2.0 cal ka (Sæmundsson 1973; Tómasson 1973; Elíasson 1977;  
221 Waitt 2002; Kirkbride et al. 2006; Baynes et al. 2015), and specifically at c.. 4.6 cal ka, 3 cal ka and 2  
222 cal ka.

223 In the SW, the main jökulhlaup activities have been assigned to 12.0 – 11.2 cal ka and c. 10.3 – 9.9  
224 cal ka, essentially from Hestvatn Lake stratigraphy (Geirsdóttir et al. 2000). In the east of this sector  
225 (Þjórsá–Tungnaá river system), we have previously shown that, from 155 to 8 ka, most of the  
226 jökulhlaups were issued from both the volcanoes (Van Vliet-Lanoë et al. 2018).



227 Historical jökulhlaups have also been analysed near the southern edge of the Vatnajökull, particularly  
228 in relation to the Skeiðarárjökull Glacier (southern Vatnajökull; Maizels, 1991, 1997; Snorrasson et al.  
229 1997; Russell et al. 2001).

230

## 231 **4. Results**

### 232 **4.1. Tephrostratigraphy**

233 Tephrostratigraphy and geochemistry were used to clarify regionally the age attributions of different  
234 events observed on aerial photography and controlled in the field. All the geochemical data are shown  
235 in Table 2.

236

237 Askja S is the most frequently recorded tephra of the observed series (see also Supplement, Figs. 3,  
238 4). The observed Askja S Tephra seems to be a mixture of rhyolitic and dacitic lavas, whilst the Askja E  
239 rhyolite is mixed with basaltic components that were issued from the Askja/Bárðarbunga and Torfa  
240 Volcanoes (Supplement Table 2).

241 We proposed the following nomenclature to simplify the reading in time of the observed tephra: the  
242 Askja E tephra, at ca 10.4 cal ka and the Askja S, corresponding to the classical ca 10.8 cal ka tephra,  
243 both predate the Erdalen Event; the Vedde Ash (11,8 cal ka) and the Askja PB tephra at ca 11.5 cal  
244 ka predate most of the Preboreal events; the Askja YD tephra at ca 12.5 and 12.9 cal ka related to the  
245 YD events were not observed in sections in the 4 investigated Vatnajökull outlets systems.

246

### 247 **4.2 The Jökulsá á Fjöllum system (Fig.2B)**

248

249 The Röndin coastal section records the whole Late Glacial, followed by the transgression of the  
250 Termination Ib (from radiocarbon analysis; Norðdhal and Pétursson, 2005). These deposits lap onto  
251 local Eemian deposits (Van Vliet-Lanoë et al. 2005). The canyon of the Jökulsá á Fjöllum River  
252 (Supplement Fig. 2) seems to be ancient (Sæmundsson, 1973; Wait, 2002) and was filled with ca. 100  
253 m of lithified glaciofluvial and lacustrine sediments, overlaid by Saalian and Eemian deposits, forming  
254 a progradation in a pre-existing incision. These Eemian deposits crop out at several places in the  
255 valley Van Vliet-Lanoë et al. 2001, 2005; Fig. 2B- yellow stars; Supplement Fig. 2).

256

257 4.2.1 Left bank – Asbyrgi, Vestudalur and Hnausar

258 This sector has been classically analysed along the canyon from a morphological point of view (see §  
259 3.6). Near the Asbyrgi dead-end, commonly attributed to prehistorical events, on the left bank, large  
260 flood evidence exists outside the canyon, in the form of vegetated fluvial megaripple fields (west of  
261 the Klappir terrace; Fig. 4), at 170 and 202 masl, with a train of smaller megaripples (east of  
262 Lambafell; Fig. 4).

263 In continuation of Asbyrgi / Klappir main terrace, the Vesturdalur one and the canyon are occupied by  
264 the Hljóðaklettur sublacustrine rootless basalt injections (Vesturdalur site, left bank; Hjartardóttir and  
265 Einarsson, 2017) estimated to be c. 10.8 ka in age, based on the Askja S Tephra analyses  
266 (Sigurgeirsson, 2016). They form a lava lake on the rocky valley bottom, close to the present river  
267 level, at an altitude of 145 – 150 masl (Fig. 2B; Supplement Fig. 3). The terrace on the left bank is  
268 topped at a 191 masl altitude by ca. 5 m of black tephra deposits and rootless cones with an  
269 Askja/Bárðarbunga geochemical signature (Table 2, sample Vest D2, see Supplement for pictures).  
270 This tephra being older than the overlying Askja E Tephra (10.4 cal ka; Table 2: Vest D3; Supplement  
271 Fig.3), seems to be local (lapilli from the rootless cones) but also associated with the Askja S Tephra  
272 (10.8 cal ka). On the other side of the valley, the fissural eruption feeding the Hljóðaklettur system  
273 consists of aerial cinder cones on the highest terrace, at 325 –350 masl, partly eroded by jökulhlaups.  
274 It issued from the Askja Volcanic system, and is subsynchronous with the Askja S Tephra (10.8 cal ka;  
275 Sigurgeirsson, 2016).

276  
277 At the Hnausar site (Figs 2b-c, 4), 5 km south of Vestudalur site, new observations in quarries and  
278 roadworks have revealed jökulhlaup deposits perched high, at 392 masl, in a juxtaglacial position  
279 (kame terrace at 396 masl. These deposits are covered by three tephra in loesses: a Bárðarbunga  
280 basaltic tephra, a rhyolitic Vedde Ash (c. 11.8 cal ka) (Table 2), and a thin Askja tephra, probably PB  
281 (c. 11.5 ka; Table 2; Fig. 5B; Supplement Fig. 3E), These tephra are locally reworked by slack water  
282 deposits, and finally buried by grey and orange loesses, latterly reworked by a thin loessic stacked  
283 moraine (Fig. 5B). Down to Hólmatungur, these deposits are remolded by a large series of  
284 palaeolacustrine beaches attesting to an ephemeral palaeolake (Van Vliet-Lanoë et al. 2005) between  
285 380 and 200 masl down to the Vestudalur terrace (Figs. 4, 5A). Two kilometers upstream of Hnausar,  
286 a complex morainic arc, at 390 masl, has been pierced by a flood along the Sauðadalur (Fig. 4 -5D).

287 Further upstream, an esker that issued from Lake Eilífisvötn (Fig. 4) joins the valley at Norðurfjöll, 10  
 288 km further south, at ca. 400 masl, and has been partly eroded by flood waters below 380 masl.

289

290 On the right-bank, in front of Vesturdalur, the Hólssandur site (Kvensöðull) forms a wide terrace above  
 291 the jökulsá canyon (Fig. 4). New observations have revealed highly visible asymmetrical-triangle-  
 292 shaped remains of a morainic arc (Fig. 6C) that have been strongly eroded by floods. These relicts  
 293 correspond to the Hnausar terminal moraine (Figs. 1B, 4). A fresh and broad splay of fine basaltic  
 294 sand (Fig. 6A, B) was deposited at 2 km to the north, as a field of large fluvial dunes, up to 380 masl  
 295 and about 220 m above the present-day river (at 160 masl). These dunes lap onto intact orange  
 296 loesses that contain a rhyolitic tephra, the Askja E (10.4 cal ka) and are surrounded by a dismantled  
 297 lava flow from the c.10.8 cal ka Askja dyke, the cones of which have been abraded by jökulhlaups to c.  
 298 280 masl. The dune splay was formed on two successive phases, one subhorizontally laminated, the  
 299 second incising and reworking the first deposit with megaripples. The end of alluvial sedimentation is  
 300 recorded in impounded levels at the southern ends of the deposits. These deposits argue for a very  
 301 limited slope at the time of the floods, and a flood trajectory located to the east of the canyon.

302

#### 303 *4.2.2 Hrossaborg tuff ring* (close to the road N1)

304 The Hrossaborg tuff ring (Fig. 2B) is located in a graben in the middle of a valley. It consists of a  
 305 phreatomagmatic cone, truncated and abraded by jökulhlaups (Alho, 2003), resting on a terrace, at  
 306 380 masl. One kilometer upstream, and to the west of the Hrossaborg crater, we observed the well  
 307 visible asymmetrical-triangle-shaped 10 m high remains (Fig. 7A) of a morainic arc that has been  
 308 strongly eroded by jökulhlaups. It is probably one of the traces within the valley axis of a younger  
 309 advance compared to Hnausar, between Dettifoss and the Heirðubreid (Fig.2B).

310

#### 311 *4.2.3 Uppermost valley of the Jökulsá á Fjöllum* (10 km north of the Vatnajökull)

312 In the uppermost valley of the Jökulsá á Fjöllum, the Vaðalda Volcano is sparsely splayed on its west  
 313 flank by fresh basaltic tephra, to a limited altitude, 200 m above the Dyngjuvatn Lake surface. Field  
 314 and image observations have revealed that fresh tephra fills the lake that occupies the crater  
 315 depression (Fig. 7D). An associated train of large, flat slackwater dunes, subhorizontally laminated  
 316 and now consolidated (25 m wide, 50 cm high; Fig. 7D), covers the Dyngjuvatn palaeolake surface up

317 to 688 masl. This deposit seems related to the rapid drainage of a larger lake, 'dammed' between the  
318 Askja caldera and Vaðalda Volcano. The lake was issued from the Dyngjujökull Glacier (Fig. 1). The  
319 dark, fine basaltic sand splay extends to the north, through Vikursandur, to the foot of the  
320 Heiðubreiðartögl tuya, ending 1 km north of the confluence of the Jökulsá and Kreppa Rivers. It  
321 reappears downstream, where it continues forming discrete terraces. These are higher in the  
322 topography than the jökulhlaup features formerly described which splayed closer to the present course  
323 of the Jökulsá River (Alho 2003, Alho et al. 2005; Carrivick et al.2013),.

324 The Kreppa River runs from the Brúarjökull outlet glacier (Fig. 2B), via the Kverká River, to the  
325 extension of a major and incised subglacial plain (Björnsson, 2009). Downstream, in front of the  
326 Heirðubreið Tuya, a dead valley diverges from the Kreppa River to the East, piercing a subglacial  
327 volcanic ridge, and producing a giant, fresh basaltic crevasse splay with megaripples (20 m wide). This  
328 abandoned channel is now drained by the Arnardalsá River, at 3 km to the North of a relatively fresh,  
329 thin terminal moraine (Fig. 7B), which was issued erratically, straight from the Brúarjökull Glacier and  
330 draining it. This moraine is only present on the eastern side of the Arnardalsalda hill, where it covers  
331 the trace of an ancient braided channel. It was pierced by a jökulhlaup and further incised by  
332 retrogressive scablands (Supplement Fig.1). .

333

#### 334 *4.2.4 Synthesis for the Jökulsá á Fjöllum valley*

##### 335 *Advances and deglaciation*

336 The Bölling deglaciation occurred quite far inland, south of the Þeistareykir Volcano (Figs 1, 3), prior to  
337 10.8 cal ka (Askja S Tephra; Sæmundsson, 1992, 2002; Sigurgeirsson, 2016). This means that the YD  
338 glaciers occupied the valleys, as shown by the presence of the Vedde Ash in a juxtaglacial position at  
339 Hnausar (Fig. 5B), but the plateaus were only locally occupied by isolated ice-domes. The  
340 tephrostatigraphy justifies an early deglaciation for the surge moraine, morainic arc and formation of  
341 the Hnausar palaeolake, followed by the injection of the Hljóðaklettur rootless eruption (Figs 5, 6)  
342 estimated at c. 11 – 10.8 cal ka (Sæmundsson et al. 2012; Sigurgeirsson, 2016) issued from the Askja  
343 fissure system. The position of the main Preboreal moraine (ca 11.2 cal ka) was thus immediately  
344 south of Hólssandur and at Hnausar.

345 The morainic arc preserved upstream close to the Hrossaborg is necessarily younger than Hnausar,  
346 and it possibly corresponds to a glacial advance at ca 10.3 cal ka. This is henceforth called the

347 Hrossaborg Advance. The upstream of Morðrudalur, the single morainic arc issued straight from the  
348 Brúarjökull outlet glacier, despite the hilly morphology, remains untouched by large jökulhlaups  
349 (Fig.7B), excepted to the West. This advance is older than the Little Ice Age and potentially  
350 corresponds to the "9.3 cal ka" advance. The traces of this advance – probably a sudden surge – in  
351 the upper valley, are hereafter called the Kreppa Advance.

352

### 353 *Jökulhlaups*

354 Some early jökulhlaups occurred before the deposition of the Vedde Ash; some of them had already  
355 eroded the surface prior to the occurrence of the 11-10.8 cal ka cinder cones, but after the Preboreal  
356 Advance. The rupture of the Hnausar morainic arc developed more or less synchronously with the  
357 palaeolake formed during the 11.2 cal ka deglaciation, after the deposition of the Askja PB tephra  
358 (11.5 cal ka). The recent incision of the canyon was reaching the valley bottom probably prior to the  
359 infilling of the Younger Dryas glacial advance, but the re-incision by the Holocene floods had not yet  
360 been achieved at 11 – 10.8 cal ka, during the setting of the Askja S lava lake (Sigurgeirsson, 2016).

361 The Hólssandur fluvial dunes are younger than the setting of the Askja S fissure eruption. The  
362 geochemistry of the dunes (samples Hols 1, 2, Table 2) yielded a Bárðarbunga source. It is very  
363 similar to that of the upstream fluvial dunes at the foot of the Vaðalda Volcano (Dingjuvatn Dunes β)  
364 from which the northern termination resembles the Hólssandur fluvial ones. Both dunes partly  
365 rework, and lap onto the orange loesses at Hólssandur that included the Askja E Tephra (c.10.4 cal ka;  
366 Table 2, Hols 3). The only Bárðarbunga tephra that fit with this stratigraphy are those at 10.1 and 9.9  
367 cal ka from the database (Table 1), the 9.9 cal ka event being the largest volcanic event  
368 (Guðmundsdóttir et al. 2016). The Hrossaborg terminal moraine remained strongly eroded by  
369 jökulhlaups that should fit the same event. The Kreppa terminal moraine (ca 9.3 cal ka) although  
370 eroded to the West was just pierced by a younger flood issued from the Kverka River related to a  
371 major volcanic eruption close to 9.2 or 9.1 cal ka from the Grimsvötn volcano. It probably corresponds  
372 to the event formerly described (Alho et al. 2005; Carrivick et al. 2013).

373 .

### 374 **4.3 Skjálfandi–Fnjóskadalur**

375 The valley of the Skjálfandaflljót is a little-studied area (Figs. 2 and 9), but it is the regular path of  
376 jökulhlaups (Björnsson, 2017). The upper valley is partly masked by the Middle Holocene lavas from  
377 the Trölladyngja Volcano (< 4.2 cal ka; Sigurgeirsson et al. 2015; Fig. 2A).

378

#### 379 *4.3.1 Lower Valleys of Skjálfandi and Fnjóskadalur*

380 This part is the most studied (Fig.8) on very few sections. We have thus described new ones and  
381 analysed the tephra. The estuary has been ice-free since the Preboreal (11.2 cal ka; Norðdhal and  
382 Pétursson 2005). The Bárðardalur Valley (Fig. 2A) was mostly incised in the bedrock, as close to  
383 Fosshóll scabland, with lateral moraines bordering the lower valleys. Eemian, hyaloclasite-derived,  
384 bioturbated marine deposits crop upstream out as pillars, along the valley bottom to 240 masl, from  
385 the entrance of the lateral Fnjóskadalur.

386 At the eastern entrance to Fnjóskadalur, thick, massive deposits of fresh basaltic gravels are buried  
387 under the recent loess along with the Askja S Tephra (11–10.8 cal ka; Fig. 10E). Inside the  
388 Fnjóskadalur, the 'pitted' moraine at the western end of the lake (at Stórutjarnaskóli) is considered to  
389 be from the Preboreal advance (Figs. 8, 9A; Ingólfosson et al. 1997). It is truncated by a terrace at 120  
390 masl from the Fnjóskadalur palaeolake, but to the north, there is evidence of it being in a juxtaglacial  
391 position, with iceberg thermokarst filled with fresh basaltic sands (Bárðarbunga, 11.3 cal ka, Table 1;  
392 Table 2: Ljos2). These deposits also reworked an Askja tephra much older than the Askja S (Askja  
393 PB, c. 11.5 cal ka), deforming very coarse, rounded gravel. It is usually covered by loesses, including  
394 the Askja S tephra. The moraine morphology resembles the collapse features described by Fard  
395 (2002) on jökulhlaup eskers, and by Olszewski and Weckwerth (1999) on sandurs. The lower alluvial  
396 unit of the kame along the pitted moraine contains both reworked Vedde and Skógar Tephra (Table 2:  
397 Ljos 1). Downstream of Fnjóskadalur, below Hólls, the side of the valley is pasted by lacustrine and  
398 alluvial deposits resting on coarse till, including several layers of fresh basaltic sands reworking the  
399 Vedde ash (Fig. 9C). These deposits are covered conformably by deformed loesses that incorporated  
400 the in situ but deformed Askja PB Tephra at its base (c.11.5 cal ka; Fig.9E).

401 Two main morainic arcs are preserved in the parallel Aðaldalur Valley. The Fragarness arc is a  
402 disrupted terminal moraine (Fig. 8). The related Palmholt Quarries (Fig. 8; Supplement Fig 4), host  
403 broad, flat-bottomed kettle holes, which are invaded by deformed flood gravel layers and reworked  
404 basaltic tephra. The morainic material is sandwiched between two generations of jökulhlaups, the

405 lower one reworking the Vedde tephra. The more recent generation accommodates decametric kettle  
406 holes. The late loessic sedimentation is similar to these of Fnjóskadalur including Askja Tephra. A later  
407 morainic arc, partly eroded, exists upstream, at the Laugar village.

408

#### 409 *4.3.2 Upper Valley, the Bárðardalur and Vosnaskarð depression*

410 The morphology in the Bárðardalur Valley is only erosional and partly masked by Middle Holocene  
411 lavas from the Trölladyngja Volcano (< 4.2 cal ka; Sigurgeirsson et al. 2015; Fig.2B). Hníflar  
412 palagonite pinnacles are flood-sculpted in the central part of the Vosnaskarð depression (Fig.2B). A  
413 palaeolake occupied this depression, at a higher altitude (1100 masl) than the Vaðalda Volcano and  
414 palaeolake Hágongulón (see § 4.5).

415 Fresh glacial striae are oriented towards the Skjálfandaflljót from the upper Sprengisandur (Fig. 2A).  
416 One terminal moraine is observed NW of Svartárvatn Lake, restricted to the right side of the canyon of  
417 the Bárðardalur Valley, a (Fig. 2A; Supplement Figs. 5A - 5B). The canyon was probably infilled by  
418 rather old tills and lacustrine deposits, as preserved on the left bank of the canyon at  
419 Halldórsstaðaskógur (Fig. 2A, Supplement Fig. 5C). Sandy basaltic deposits are splayed on the  
420 plateau, upstream and around the Svartárvatn Lake. The stratigraphy at Svartárvot clearly indicates a  
421 Bárðarbunga Tephra (probably 10.4 cal ka; Table 2), followed by the Reistvik 8 Tephra (10.2 cal ka;  
422 Guðmundsdóttir et al. 2016). SE of the Svartárvatn Lake, the Reitsvik 8 Tephra (Table 2, Svartár) is  
423 preserved in the upper part of the consolidated orange loess, but has been laterally eroded in a NE  
424 direction by jökulhlaups.

425

426 Another structure is visible to the NE of Svartárvot, on the plateau, near the Sellandafjall Tuya. These  
427 structure shaped the surface of the Kárkárbotnar Formation which is consolidated and faulted  
428 palaeofan considered as Late Glacial and initially scoured after the YD Advance (Sigurgeisson et al.  
429 2015). An alluvial splay of granules and sands shaped the surface of Kárkárbotnar Formation into  
430 fluvial megadunes (two generations) and residual hillocks (Figs. 2A and Supplement 12) based on  
431 aerial views. This surficial morphology is associated with an esker (Fig Supplement.12 E) that  
432 propagated subglacially to Myvatn Lake from the north of Krákárbotnar, parallel to the present Kráka  
433 River. The final evidence of flooding is here the splayed sands stopping at the western foot of the  
434 Sellandafjell and associated with the sculpting of elongated islands in the loess cover

435 (Supplement.12). From these data, and the location of the terminal moraine in the Bárðardalur Valley,  
436 the Kárkábotnar initial superficial scouring could be generated by the same events as these recorded  
437 at the Svartárvatn Lake, a part of the flood(s) deriving towards the Myvatn Lake to the east (Fig. 2A).

438

#### 439 4.3.3 Synthesis for Skjálfandi–Fnjóskadalur- Myvatn-Vosnarskard system

440

##### 441 *Glacial advances and retreats*

442 In the Bárðardalur and Adaðalur Valleys, (Fig. 8), the basal fluvial sediments have reworked the  
443 Vedde Ash as also observed at Laufas and Kaupangur in Eyjafjörður (Sigurðsson and Sparks 1978;  
444 Van Vliet-Lanoë et al. 2005). Ljosavatn and Palmholt quarries reveal dismantled glacier ice tongues  
445 that represent a rapid advance or a surge. The Skjálfandi terminal moraine was located in the vicinity  
446 of Skollahnjúkur as suggested by a perched pitted morphology at 200 masl (Fig.8). This surge can be  
447 attributed to the "11.2 – 11.4 cal ka" Preboreal event (tephrostratigraphy; Ingólfsson et al. 1997),  
448 formed before the deposition of the Askja PB and S Tephra. By deduction, the "10.3 cal ka" terminal  
449 moraine could be restricted to these at the Laugar village. Upstream, the lateral moraine immediately  
450 NW of Svartárvatn Lake most probably yields an age of ca 9.3 cal ka, based on tephra (younger than  
451 the Reistvik 8 Tephra at 10.2 cal ka). This is the proposed Svartárvatn Advance.

452

##### 453 *Jökulhlaups*

454 The oldest basal generation of flood immediately postdates the Vedde Ash and represents an early  
455 stage in the Preboreal deglaciation. The main (second) generation of floods predates the Askja S  
456 Tephra (10.8 cal ka), thus yielding 11.2 - 11.1 cal ka in age. The third generation took place before the  
457 first Bárðarbunga eruption (10.4 cal ka; Guðmundsdóttir et al. 2016) and the overlying Askja E Tephra  
458 (10.4 cal ka). The latter generation (last terrace) is logically related to eruptions of the Bárðarbunga at  
459 c. 10.1 or 9.9 cal ka (Table 1; Guðmundsdóttir et al. 2016). Flood deposits from the Skalfandafliót  
460 River were also washed into Fnjóskadalur and Eyjafjörður via the Dalsmynni Canyon (Hólls and  
461 Finnastadir Farm; Figs. 4, 8D). This path for floods explains the water-lain facies of the "10.3 cal ka  
462 Saksunarvatn Tephra" observed at Reitsvik (Guðmundsdóttir et al. 2016).

463 The surface granular components of the Krákárbotnar palaeofan (Sigurgeirsson et al. 2015) had  
464 probably the same source and age as the deposits at Svartárvatn, based on the orientation to the NE of



465 the megadunes. All these floods seem to issue from the Dingjujökull outlet glacier (Table 1, Fig.2B)  
466 and to correspond to the Bárðarbunga 10.4, 10.1 and mostly 9.9 cal ka tephra. These floods probably  
467 had a long duration (weeks or more), as they were mainly fed by deglaciation, justifying their lateral  
468 displacement to the east by the Coriolis force feeding the Myvatn basin. Re-incision in Preboreal  
469 glacial sediments of the Bárðardalur valley after the Preboreal glacial advances occurred much  
470 later, prior to 7 cal ka, but probably from ca 9.3 cal ka.

471

#### 472 **4.4 Jökulsá á Brú and the East**

##### 473 4.4.1: Halslón palaeolake

474 To the NE of the Vatnajökull, sparse evidence of jökulhlaups has been recorded in the Halslón Dam  
475 sedimentary fill that was issued from the western Brúarjökull outlet glacier, via Kringislá (Knudsen and  
476 Marren, 2002). The stratigraphy of the Holocene events has evolved from the ones previously  
477 published (Knudsen and Marren, 2002; Van Vliet-Lanoë et al. 2010). The LIA moraine is located north  
478 of the Kringslaranni, on both sides of the dam lake, followed by successive termini. A terminal moraine  
479 older than the LIA is visible North of Hals, on the right bank of the lake (Fig.1b), fitting the 9.3 cal ka  
480 advance.

481 Jökulhlaups reworked first the Younger Dryas Grímsvötn Tephra, filling troughs with massive deposits  
482 (Van Vliet-Lanoë et al. 2010), that were further deformed by a first glacial advance. An Askja E Tephra  
483 (10, 2 cal ka comparable to Vestur R3 in Table 2; Fig. 7C) has been deformed by a second glacial  
484 advance. The Saksunarvatn Tephra was water-lain in the highest terrace and deformed as a  
485 thermokarst depression in a juxtaglacial position (Fig. 7D). Laterally, it is undeformed, and post-dated  
486 by a faulted geyserite at  $10.09 \pm 0.28$ ka BP (U–Th uncorrected dating; Van Vliet-Lanoë et al. 2007;  
487 Fig. 7E).

488 The Jökulsá á Brú River is deeply incised down from the Brú village, similar to the canyon of the  
489 Jökulsá á Fjöllum. Two remains of terminal moraines were preserved. Downstream, a ‘pitted’ moraine  
490 is visible on the left bank, at 1.5 km south of Selland, below a large mass flow on the right bank. This  
491 is probably an end moraine from the Botarheiði that evolved into a rock glacier. Further down, another  
492 1.5 km long “pitted” end moraine crops out near Krókavatn (Fig. 1b).

493

##### 494 4.4.2 The Eyabakkajökull

495 This glacial outlet shows traces of stacked, permafrosted ice-thrust ridges, comparable to several  
496 Svalbard glaciers (Boulton et al. 1999; Sund et al. 2009; Waller et al. 2012). They formed in relation  
497 with the 1890 surge and commonly attributed to hydraulic surging (Sharp, 1985; Kaldal and Vikingsson,  
498 2000; Björnsson et al. 2003; Schomacker et al. 2014, Fig. 11). The southern eastern part is a rather  
499 thin diamicton and associated with a negative inprint of ice crevices and a narrow sharp plastic edge  
500 and stacked folds of a surge fitting the 1890 one (Schomacker et al, 2014). The northernmost / outer  
501 part of this moraine seems to be partly older, seeing the thermal cracking and vegetation cover which  
502 are similar to the situation on the recent volcanic ridge at Veiðivötn (1477 AD eruption). For the nearby  
503 Brúarjökull Glacier, the presence of a glacier-deformed Örfæfajökull Volcano pumice (1362 AD  
504 eruption; Benediktsson, 2012) suggests a setting during the coldest part of the Little Ice Age, from  
505 1600 AD to 1800 AD (Guðmundsson et al. 1997). The Múlajökull Glacier had its maximal advance  
506 between AD 1717 and 1760 (Benediktsson et al. 2015). The persistence of relict morainic arcs within a  
507 palsa bog immediately downstream of the Eyabakkajökull terminal should indicate a long, unglaciated  
508 interval, at least from the 9.3 cal ka advance. Furthermore, a field of open-system hydrolaccolite  
509 occupies the western side of the valley, attesting of the limited glacial extent to this side. Several  
510 jokulhaups have pierced the arc, mostly to the West, in association with an esker, but it does not seem  
511 possible to build a chronology.

512 Observations were performed downstream at the Lögurinn Lake (Fig. 1): the deglacial sedimentation  
513 covered an interval from 10.2 cal ka (Strinberger et al. 2012; Guðmundsdóttir et al. 2016; Norðdahl et  
514 al. 2019), fitting an age of 11.2 cal ka for the terminal moraine north of Eglisstaðir (Fig. 1).

515

#### 516 4.4.3 Synthesis.

##### 517 *Glaciers*

518 At the Halslón Lake, the older Grímsvötn Tephra has been reworked in flood troughs in association  
519 with the Bölling deglaciation (Fig. 10). The Boreal advance is slightly younger than 10.4 cal ka (Askja  
520 E Tephra) at the foot of the Kárahnjúkar Tuya (Fig. 7C). This corresponds to a "10.3 cal ka" advance,  
521 synchronous with the Hrossaborg and Laugar events. Laterally, the "9.3 cal ka" terminal moraine from  
522 the Brúarjökull Glacier reached the deformed Saksunarvatn Tephra in the middle of the length of the  
523 Halslón Lake, equating with the Kreppa–Svartarvatn Advance to the west.

524 To the East, the Eggilstadir moraine yields c. 11.2 cal ka. The moraines relics in the palsa bog thus  
525 yields c. 9.3 cal ka; the outer Eyabakkajökull stacked vegetated moraine with ice wedging and  
526 thermokarst (mostly to the East) yields a LIA age and the Eyabakkajökull thin recent stacked moraine  
527 yields a 1890 AD age.

528

#### 529 *Jökulhlaups*

530 Megafloods first seem to occur prior to the YD. The Saksunarvatn Tephra splayed in a flood, being  
531 preserved intact in juxtaglacial position from a pre-existing surge lateral moraine, yielding probably c.  
532 10.3 cal ka.

533

#### 534 *4.5 The South-West: The Þjorsá–Kadavísl–Tungnaá outlet*

535 The South-western main outlet from the Vatnajökull is drained by at least 3 valleys (Fig.1a). The  
536 Þjorsá–Kadavísl–Tungnaá Valleys are clearly shaped by jökulhlaups over the last two interglacials  
537 (Van Vliet-Lanoë et al. 2018). This is proven by the occurrence of floods erosional forms (Supplement  
538 Fig.6). The Late Glacial sedimentary prism is rather thin, mostly located in the jökulhlaup incisions,  
539 sometimes infilled with sediments associated with the tidal deposits of the Preboreal marine  
540 transgression. Regional field descriptions are rare, and most of the data are being extracted from lake  
541 cores (Geirsdóttir et al. 2009). To the north of this system, data are non-existent or unpublished.

542

#### 543 *4.5.1 Akbraut and Varghóll Quarries*

544 The section at Akbraut (Figs 1, 10) is located on a terrace perched at 90 masl on the east bank of the  
545 Þjorsá River. It is situated south of the Búði moraine complex, dated at ca. 11.2 cal ka (Geirsdóttir et  
546 al. 2009), and north of the Pula–Mykjunes terminal moraines (YDII, in situ Vedde Ash; Van Vliet-Lanoë  
547 et al. 2018). It has an infill of coarsely-stratified, unsorted and unconsolidated cobble material, set by  
548 jökulhlaups, which are incised into the Eemian and Weichselian deposits. These gravels are covered  
549 by stratified lacustrine silts, deformed by moderate seismic activity, as is also the case in the section at  
550 Hrepprhólar (Fig. 1), where they are related to Termination Ia. A second unsorted diamicton was  
551 deposited, which includes, at its base, fresh, dark basaltic sands that are coarsely stratified, and  
552 sheared by glaciotectionism (Vedde β, 11.8 cal ka, Van Vliet-Lanoë et al. 2019). This is covered by 1

553 m of stratified marine silts (from Termination 1b), including Vedde pumice-rich beds, with small load  
554 casts. This marine unit is further truncated by more recent jökulhlaup gravely deposits.

555

556 The Varghóll–Akbraut moraines correspond, in aerial view, to thin terminal moraines that are later than  
557 the YD, but older than the major Búðí terminal moraine. These moraines present surge-type  
558 characteristics, with a stacked folded pattern. The suggested age, from the presence at the base of  
559 reworked Vedde Ash, is ca. 11.4 – 11.3 cal ka, followed by the 11.2 cal ka Búðí Advance.

560 To the West, “pitted” jökulhlaup abraded rocks (Supplement Fig. 7) crop at 230 masl on the Skarðsfjall  
561 (Fig.1A), disconnected from the natural drainage system. It suggests a discharge of water emerging  
562 from the surface of a thin decaying glacier lobe during the late YD or the Early Preboreal, probably  
563 from a crevice net.

564

#### 565 4.5.2 Upstream watershed of Þjórsá River

566 Another terminal moraine was mapped by Kaladal and Vikingsson (1990) to the north, east of the  
567 Sultartangalón Lake. It is associated with an eroded esker deposit (Fig.12A). Herein, this terminus will  
568 be referenced as the Búðarhals Advance. At the hydroelectric station of the Þjórsá Lake, the base of  
569 the Late Glacial to the Holocene canyon infill is extremely rich in basaltic tephra, resting on an Eemian  
570 fluvial formation, as in the Upper Ytrí–Rangá Valley (Van Vliet-Lanoë et al. 2018).

571 Upstream, the Þjórsá River is connected to the artificial Hágögulón Lake (Fig. 1A). This lake’ site was  
572 occupied several times by palaeolakes. At the eastern foot of the Nyðri Háganga Tuya, ca. 100 m  
573 above the original valley floor, the upper raised beach (at 887 masl) is covered by scattered fresh  
574 basalt fragments that were issued from the Bárðarbunga Volcano. This upper raised beach is at the  
575 same altitude as the tephra filled Vaðalda crater, east of the Askja Volcano (Figs 1, 2B). Another  
576 palaeolake outlet is perched at 860 masl, and is connected to a lake level located at 80 m above the  
577 valley floor, associated with a retrogressive dry valley, incised in an Eemian glaciofluvial deposit  
578 (Supplement Fig. 8E). Many jökulhlaup-polished blocks (Supplement Fig. 8A-D) are visible at 835  
579 masl, 65m above the valley floor, at a palaeo-outlet of the palaeolake. It seems that this palaeolake  
580 infilled several times, with up to 100 m of water for the oldest highstand, with evidence of flood  
581 bursting (incision). Damming was generated by westward ice surging from the Tungnaárjökull and  
582 Sylgjujökull outlet glaciers (Fig. 1, extent in yellow). This is evidenced by the morainic system that

583 ends close to the Kisa River to the west (Fig. 7C; Kaldal and Vikingsson 1990), issuing from the north  
 584 of the Sauðafell hill (North of Veidivötn). This is associated with evidence of lakes with terraces, and a  
 585 field of open system pingos in silty sands, close to 850 masl, at SW of the lake. It is overlapped finally  
 586 to the south by the Þjorsá Lava (8.6 cal ka). This terminus will be herein referenced as the Kisa  
 587 Advance, surging from the Tungnaárjökull and Syglujökull outlet glaciers.

588

#### 589 4.5.3 Synthesis for the Þjorsá–Kadavísl–Tungnaá outlet

590 Following the Glacial Termination 1b, several glacial advances occurred in the south. The twinned  
 591 advances of the YD I – Pula moraine (c. 12.8 ka cal) – and YD II– Mykjunes moraine (c. 11.7 – 11.5  
 592 cal ka) – formed a major glacial terminus (Van Vliet-Lanoë et al. 2018).

593 Observations at Varghóll suggest a retreat from Mykjunes to the Búði arc, with pulsed surges  
 594 associated with jökulhlaups and iceberg discharges (Figs 10, 11) that correlate with a c. 11.4 – 11.3  
 595 cal ka surge, the Varghóll – Akbraut Advance just after the YD deglaciation. The Búði Terminus fits the  
 596 “11.2 cal ka” Preboreal advance (e.g. Geirsdóttir et al. 2009). The next glacial advance should  
 597 correspond, in the south, to the Búðarháls Advance on the eastern side of the Kaldavísl River. A later  
 598 trace of glacial surge emerges from the East, between the Sandfell in the south and the Hágöngulón  
 599 lake sector in the north, to the Kisa River, but this seems to have been untouched by jökulhlaups. Its  
 600 source was hindered by the Þjorsá Lava, at 8.7 cal ka (Halldórsson et al. 2008), and seems to have  
 601 issued from the Syglujökull and Tungnarjökull outlet glaciers, which normally surge westward  
 602 (Björnsson et al. 2003). This places at ca. 9.3 cal ka the Kisa Advance which is an erratic advance.  
 603 This indicates an age of ca 10.3 cal ka for the Búðarháls Advance.

604

## 605 5. Discussion

606 To understand the various potential controls on jökulhlaup genesis, we first discuss the climate  
 607 evolution and deglaciation history for each outlet of the Vatnajökull. Second, we analyse the  
 608 connection between glacio-isostasy and the potential storage of aquifers, or of subglacial lakes, at the  
 609 level of the Vatnajökull. On this base, it will be possible to discuss the dynamics that drove the Early  
 610 Holocene jökulhlaups.

611

### 612 5.1 Timing of the Glacial Advances (Fig. 1 b)

613 After the deglaciation of the YD and the deposition of the Vedde ash (ca 11.8 cal ka), the Early  
 614 Holocene glacial advances of the Vatnajökull spreaded at ca 11.5-11.4, 11.3-11.2, 10.3 and 9.3 cal ka,  
 615 in concert with climate evolution in the other regions surrounding the North Atlantic (see § 3.1.).

616 The Varghóll – Akbraut in the South, Hnaussar, Ljósavatn and the Fragarness - Palmholt terminal  
 617 moraines in the North (ca 11.4 - 11.3 cal ka) correspond to the first Early Preboreal advance. The Búði  
 618 (c. 11,2 cal ka) is a second Early Preboreal Advance. The later Búðarhals Advance in the South is  
 619 estimated to be synchronic with the Erdalen Events and the Saksunarvatn complex tephra, at c.10.3-  
 620 10.2 cal ka as well as the Hrossaborg, Laugar and Eglišstaðir termini in the North. For the independent  
 621 Langjökull on the West Volcanic Zone, the Saksunarvatn Tephra has been found below a lava flow  
 622 (Jóhannsdóttir 2007; Eason et al. 2015), also suggesting an early deglaciation, and isolation of the  
 623 southern and eastern margins of the glacier.

624 The Kreppa, Kisa, Svartárvatn and outer Eyabakkajökull Advances seem to be equivalent to the “9.3  
 625 cal ka” Cooling Event. They are characterized as erratic in their flow direction, compared to the Late  
 626 Glacial and Preboreal Advances. This late advance is also recognised around the Drangajökull Glacier  
 627 (NW peninsula; Harning et al. 2016).

628 The “8.2 cal ka” event has not been officially recorded in lacustrine records in the NE of the island  
 629 (Striberger et al. 2012); however, it is clearly recorded in cores at Hvitárvatn (Langjökull; Larsen et al.  
 630 2012; Fig. 1) and is also extractable from the <sup>3</sup>He measurements of volcano summit lava’s tracing the  
 631 local deglaciation (Liccardi et al. 2007). This means that the glacial limits for this pulse were mostly  
 632 inside the present extent of the glacier.

633

### 634 **5.2 Timing of the Jökulhlaups (Fig.2 B)**

635 In the Jökulsá á Fjöllum, major jökulhlaups thus occurred from our data in relation to the Younger  
 636 Dryas deglaciation (Hnaussar kame terrace), probably between 11.8 and 11.4 cal ka, centred on 10.1 –  
 637 9.9 cal ka (Hölssandur) and ca 9.1 cal ka south of the Hrossaborg. The major Icelandic jökulhlaups in  
 638 the literature relate to the onset of canyon clearance of the sedimentary infilling. After 9.3 cal ka, some  
 639 events were still large, but were better channeled.

640 It thus seems that most of the early jökulhlaup events in the Þjórsá, Skalfandi, Jökulsá á Fjöllum and  
 641 Jökulsá a Brú valleys occurred from the Late Younger Dryas deglaciation, some having bursted just  
 642 prior to the Early Preboreal Advance (at 11.2 cal ka). The jökulhlaup mentioned at Kjölur (Kajafell

643 Volcano, Langjökull) by Tómasson (1973) also seems to correspond to a first-generation surge (at  
644 12.0 to 11.2 cal ka), as it is associated with iceberg kettle holes and two eskers, all overflowed by the  
645 Preboreal lava (Eason et al. 2015).

646

647 In the South, after the Preboreal, the majority of flood events occurred after the Búðarháls Advance  
648 ("10.3 cal ka"), but prior to the Kisa Advance (" 9.3 cal ka"). These field data fit the results from the  
649 Hestvatn Lake record (Fig.1; Geirsdóttir et al. 2000) that assigned two major periods to the flood  
650 activity around 12.0 - 11.2 cal ka, and again around 10 – 9.9 cal ka as it is also the case for the  
651 Jökulsá á Fjollum watershed.

652

### 653 **5.3 Permafrost and surging**

654 Outlet glaciers of ice caps that periodically surge after long quiescent phases, undergo large and  
655 sudden pulses accompanied by terminus advance (Harrisson et al. 2015; Benn et al. 2019).  
656 Polythermal glaciers, often associated with permafrost are prone to slow surging (Benn and Evans,  
657 2011). The temperate glaciers in Iceland exhibit surges with a sudden onset, extremely high (tens of  
658 meters/day) maximum flow rate and an abrupt termination, associated with a discharge of the intra-  
659 glacier stored water (Björnsson et al. 2003). It can be triggered by an enhanced climate-driven melting  
660 (Stibberger et al. 2011) or a volcanic meltwater supply. Ice breakage at the glacier surface or perched  
661 outlets, with the local surface dismantling and iceberg splay, can be induced by constrained  
662 overpressure (Roberts et al. 2000). The downward locking of hydraulic pressure can result from the  
663 impeded drainage by permafrost development that seals the snout and margins of the glacier, frozen  
664 to the bed during cooling events, as in the polythermal glaciers in Svalbard (Fig. 12D; Lonne et al.  
665 2016).

666 The impact of the permafrost seems evident for the Eyarbakkajökull outlet and the Kreppa, Kverká and  
667 Jökulvísl River outlets of the Brúarjökull Glacier (evidenced by the recently pitted terraces). This  
668 system with surface dismantling is also valid for the outlets of the Breiðamerkurjökull River, in  
669 association with concertina eskers (Knudsen, 1995), for the Gígjulvísl and Skeiðará Rivers  
670 (Skeiðarárjökull Glacier) and the uppermost Tugnaá River.

671 As the Preboreal climate was rather cold especially during the two first Bond events, particularly in the  
672 North-East (about - 10°C lower than today; Rasmussen et al. 2011), we might expect the impact of

673 permafrost damming to be one of the main triggers of jökulhlaups. The preserved pre-jökulhlaup  
674 morphology suggests such permafrost prior to the principal Early Holocene flood, also deformed by  
675 tectonic faulting to the east (Fig. 14C). The degradation of ice bodies in the terminal moraine during a  
676 warming event could have reduced the internal stability of the dam and, therefore, easing flood  
677 emergence (GAPHAZ 2017). Sealing of the glacier tongue by permafrost could have favoured the  
678 retention of meltwater in the subglacial lakes or intraglacial aquifers, especially if the surface and  
679 bottom slopes were very low (see § 5.2), increasing the probability for glacial surging for both  
680 polythermal and temperate glaciers (Benn et al, 2019). Proglacial icing accumulations from eskers  
681 could also have induced damming. This was likely the case for the glacial tongues reaching the lower  
682 Skálfandi River (at Ljósavatn and Palmholt) and the lower Þjórsá River (at Varghóll), around 11.2 cal  
683 ka. After 10.4 cal ka, jökulhlaups were largely splayed on the deglaciated plateaus. The occurrence of  
684 permafrost in the watershed probably limited the vertical incision of any flood, promoting a lateral  
685 extension of such floods in the valleys, as it took place with the first mega-jökulhlaup, responsible for  
686 the Hólssandur hydraulic dunes (Jökulsá á Fjöllum), or the Svartárvatn–Krákárbotnar jökulhlaup (at  
687 Skjálfandaflljot).

688

#### 689 ***5.4 Impact of the glacio-isostatic rebound***

690 As the c. 10.8 cal ka Askja S Tephra and the c. 10.3 cal ka Saksunar events happened relatively soon  
691 after Termination Ib, an early glacial rebound from the coast to the inland, should be expected  
692 attenuating the slope of the lower to middle Jökulsá Valley. The distance from Röndin to the  
693 Hrossaborg moraine is about 50 km, and is being uplifted at rates similar to those in the south of the  
694 isle, c. 40 to 60 mm yr<sup>-1</sup> (Le Breton et al. 2010). Present-day isostatic uplift around the Vatnajökull is 9  
695 to 25 mm yr<sup>-1</sup> (Pagli et al. 2007). The persistence upstream of an extended Vatnajökull suggests that  
696 this sector is still downwarped in relation to the remaining ice thickness, and the presence of the  
697 hotspot (due to lower viscosity). An initial subsidence of 500 m is expected at the LGM for a 1500 m  
698 thick ice sheet. Supposing that half of the ice-sheet thickness (ca. 750 m) has already melted at 10.3  
699 cal ka, we could theoretically expect a residual subsidence of 250 m in the central part of the system,  
700 and half of that (125 m) in the outlet zone of the Hrossaborg and Buðardalur surging tongues. This  
701 simple but certainly overestimated approach allows imaging of the potential extent of subglacial lakes  
702 or aquifers (Fig. 13) which are retained below the residual thinned flat ice-sheet (slope ca 4%,



703 subglacial volcano excepted), and adapted from the subglacial topography described by Björnsson  
704 (2017). Due to the speed of the rebound (ca 25 - 40 mm yr<sup>-1</sup> from the present-day values) imposed by  
705 the fractured substratum (Höskuldsson et al. 2006; Le Breton et al. 2010), this isostatic rebound had  
706 very little chance of strongly modifying the geometry of the meltwater catchment areas for the brief  
707 interval (10<sup>2</sup> yrs) considered, but the isostatic subsidence was certainly resorbed after 1 ka, as in the  
708 “9.3 cal ka” deglaciation.

709

### 710 **5.5 Volcanic activity, jökulhlaups and unloading**

711 Glacial unloading could directly influence the activity of volcanoes by adiabatic crustal melting  
712 (Eksinhol et al. 2019). The Grímsvötn and Bárðarbunga Volcanoes were partly merging from the ice  
713 sheet since 18 cal ka and mostly from 11.8 cal ka (Van Vliet-Lanoë et al, 2019). With the ongoing  
714 rapid warming from 11.8 cal ka (Fig.3), the glacial unloading of the western Vatnajökull allowed the  
715 supply in magma in the deep subcrustal reservoirs of the NVZ and EVZ (Hartley and Thordarson,  
716 2013). This supply fed first the Grímsvötn and Bárðarbunga reservoirs (see tephra record, Table 1).  
717 The Askja Volcano is presumed deglaciated around 10.3 cal ka only (Hjartarson, 2003; Hartley and  
718 Thordarson, 2013), but it was fully emerged from 11.4 cal ka. The PB Tephra was followed several  
719 smaller eruptions (Fig.10; See Supplement Fig.4), before the major Askja S. The Saksunarvatn  
720 Tephra (10.24 – 10.12 cal ka) has a more complex story.

721

722 The potential triggering effect on eruptions of small unloading events, such as water discharge from a  
723 lake and ice thickness variations, has been demonstrated when the underlying magma chamber is  
724 close to failure conditions comparable to the static stress change induced by earthquakes (1-10 kPa;  
725 Albino et al. 2010). Today, ice loss reaches 6 m yr<sup>-1</sup> for the Vatnajökull (<http://Vedur.is>). The climatic  
726 unloading reaches 60 kPa after 100 years of warming, with a potential effect on the upper reservoir of  
727 the Grímsvötn Volcano.

728

729 Thermal analysis at the Kverkfjöll Volcano has indicated that a jökulhlaup must have taken place a few  
730 days after the initial subglacial lava emplacement (Höskuldsson et al. 2006). Maximal ice melting rates  
731 of the order of 10<sup>3</sup> m s<sup>-1</sup> have been indicated. The magma must have reached water content close to  
732 saturation at emission allowing the onset of a phreatomagmatic eruption after lake drainage

733 (Höskuldsson et al. 2006). It also lowered the melting temperature of magma and the spreading of  
734 extensive lava flows (Wylie et al. 2000), as observed immediately after deglaciation. This interpretation  
735 suggests that jökulhlaups reactivate eruptions by lowering the pressure on the magma chamber  
736 (Höskuldsson et al. 2006; Albino et al. 2010).

737

738 The expected succession for deglacial events are thus: 1) subglacial eruption; 2) lake formations; 3)  
739 jökulhlaup and iceberg discharge, or possible glacial surge; and 4) phreatomagmatic ash emission.  
740 This succession could occur rather rapidly, within a few weeks, or be recurrent as a part of the 10-year  
741 cycle for Grímsvötn volcanic activity and is most likely for summer events, especially if the perturbation  
742 is within a range of 7 kms from the centre of the reservoir (Albino et al. 2010). Climatic glacial  
743 unloading, at present values, seems to be enough to induce a first eruption of the upper reservoir.

744

#### 745 *5.6 Climatic melt control and jökulhlaups*

746 Jökulhlaups have been recorded from the Alleröd cooling (ACE Fig.3), even during brief events in the  
747 Younger Dryas, especially in the NE of the isle, commonly starved in precipitation. The onset of  
748 warming would have raised the water level in subglacial lakes, favouring surges (e.g. Russell et al.  
749 2001). From 10.25 cal ka BP, warming was significant, producing temperatures similar to those of  
750 today (see § 3.1). Restoration of the Irminger Current favoured a rise in precipitation. Rapid ice melting  
751 could have been driven by intense late-summer rainfall, causing the thinning of large ice-sheet at low  
752 altitudes (Doyle et al. 2015). Surface melt-lakes represent today ca. 1 km<sup>3</sup> of the decaying part of the  
753 Greenland ice-sheet (Fitzpatrick et al. 2014). The depression below the Brúarjökull outlet glacier  
754 (Brúarjökull Lake; Fig. 16) is the largest, but with an "impermeable" bedrock. This glacier surged first.  
755 Traces indicate that surging certainly issued from the west of the Brúarjökull in both of the latter  
756 events, the "10.3" and "9.3 cal ka" ones. The surge responsible for the Hrossaborg Advance ("10.3 cal  
757 ka") clearly dammed from the north the depression south of the Vaðalda and Askja Volcanoes. This  
758 explains the Dingjuvatn Dunes and the splay of dark tephra on the western flank and in the crater of  
759 Vaðalda, to 200 m above the valley floor, emitted by ca 10.1 and/or 9.9 cal ka eruption of the  
760 Bárðarbunga. The Kreppa Advance ("9.3 cal ka") mostly dammed the Jökulsá á Fjöllum, and was  
761 probably responsible for the high flood described by Alho et al. (2007). Both flood series (c. 10.1–9.9  
762 cal ka and < 9.3 cal ka) were initially linked to ice-damming. Today, most outlet glaciers in Iceland are

763 thin, plastic and temperate-based surging glaciers (Bjornsson, 2017). These could be more  
 764 susceptible to recurrent overpressure compared to the colder surging glaciers (Ben and Evans, 2010)  
 765 that existed in Northern Iceland during the Preboreal and that probably needed higher overpressures  
 766 to breach a permafrosted dam.

767

768 As the '10.3 cal ka' events seem to have been rather synchronous with the Askja, Grímsvötn and  
 769 Bárðarbunga eruptions, a supplementary melt is expected from the splay of ash on the surface of the  
 770 glaciers. A lowering of surface albedo could have further increased the melting efficiency by 60%  
 771 (Vogfjörd et al. 2005; Möller et al. 2013). The low albedo measured in 2005 is related to the 2004  
 772 Grímsvötn eruption (Möller et al. 2013) that was immediately followed by a jökulhlaup (Vogfjörd et al.  
 773 2005). This is also valid for the Gjálp eruption in 1996 (Guðmundsson et al. 1997; Björnsson, 1998;  
 774 Russell and Knudsen, 1999). Insolation has been rising since 10.29 cal ka, to a high between 10.20  
 775 cal ka and 10.05 cal ka, with a maximum value of ‰  $\delta^{14}\text{C}$  at 10.13 cal ka (Stuiver et al. 1998; Fig.2),  
 776 enhancing ice sheet collapse. Intervals of positive mass balance for the glaciers have conversely  
 777 lowered the volume and frequency of recent jökulhlaups (Guðmundsson et al. 1995). It thus seems  
 778 that the first large eruption took place subsynchronously with the breakage of the ice-dams north of the  
 779 Vaðalda, at ca. 10.2 cal ka, in agreement with ice-core dating. The Bárðarbunga Tephra, observed at  
 780 the base of the Svartárkot Tephra sequence, probably fit the ca 10.45 cal ka event (Guðmundóttir et  
 781 al. 2016), or possibly another event closer to 10.25 cal ka, masked by the volume of Saksunarvatn  
 782 Ash. In the following years, insolation rose to a maximum, and melting was enhanced by the first large  
 783 ash splay. Local eruptions blasted from Grímsvötn with the discharge of the caldera lake, with a  
 784 frequency close to the usual 60-year cycle during the century to 10.12 cal ka, thus filling the subglacial  
 785 lakes.

786

## 787 **5.6 The complex history of the Saksunarvatn event and associated jökulhlaups**

788 *Phase 1 – Glacial rupture and lake drainage ca. 10.3 cal ka:* the aerial Vaðalda Lake (Fig.10) was ice-  
 789 dammed by the surge of the Bruarjökull (Kreppa advance, to 200 m water depth) and fed by climatic  
 790 melt. But it is too far from both the Grímsvötn and Bárðarbunga Craters to have influenced the  
 791 reservoirs mechanically. Subglacial lakes as the Brúarjökull Lake or aquifers, also climate-fed, have  
 792 expanded probably to the edges of the Grímsvötn caldera. An initial breaching or lifting of the

793 permafrosted glacier margin down from the Vaðalda Lake, followed by the drainage of subglacial lakes  
794 / aquifers may have been associated with unloading and the first volcanic event, close to the cold  
795 Erdalen Events – “10.3 cal ka” – in relation to the shallowest magmatic chamber. The other possibility  
796 also exists: with this subglacial eruption, the input of warm water would have increased suddenly as  
797 today (Snorrason et al. 1997; Björnsson 1998; Russel et al. 2002), the subglacial water pressure,  
798 allowing hydraulic fracturing, buoyancy and subsequent ice-dam rupture.

799

800 *Phase 2 – Main volcanic activity ca. 10.25 cal ka:* Based on the sizes of the Brúarjökull and Vaðalda  
801 Lakes, it is plausible that the deepest reservoir at Grímsvötn could have been triggered by this forced  
802 drainage, promoting the main phreatomagmatic ash eruption, the main Saksunarvatn event and the  
803 Skeidarar jökulhlaup (Lacasse et al. 1998). Massive basaltic ash deposition occurred during the  
804 diminution of the flood, and on all the slackwater deposits along the upper Jökulsá á Fjöllum Valley.  
805 The drainage of the caldera to the south (via the Tungnaá River) probably occurred with a second  
806 eruption, as similar to the occurrence of the delayed triggering of the Bárðarbunga eruption (at 10.15  
807 cal ka). Several high floods thus emerged during the ca 10.24 – 10.15 cal ka period in the whole  
808 Jökulsá Rivers of the Vatnajökull. Furthermore, these events seem to have had a limited erosional  
809 capability, probably due to low-angled slopes and the persistence of permafrost, particularly during the  
810 Preboreal.

811

812 *Phase 3 – The Bárðarbunga and Grímsvötn late response, 10.15–9.9 cal ka:* The 10.15 cal ka  
813 eruption of the Bárðarbunga Volcano occurred in a similar way to that at Grímsvötn (10.25 cal ka) – by  
814 the disturbance of deep magmatic reservoirs. This interpretation probably also fits the ongoing  
815 deglaciation between the volcano and the western Tungafells outlet glacier, allowing massive storage  
816 in the natural-aerial Hágöngulón and Vonarskarð Lakes (Fig.13), resulting in flood escapes through  
817 the Vonarskarð Lake to Skjálfandi. Icebergs were apparently included in the jökulhlaup deposits, along  
818 with ash-rich deposits from the slack waters. The major jökulhlaup at 9.9 cal ka had already occurred  
819 by the onset of the next cooling, due to insolation, and is recorded in the Þjórsá Valley on the Akbraut  
820 90 m terrace (last superficial event; Figs 1,10), at Hólssandur (Kvensodull dunes) and at Dingjuvatn  
821 (close to the Askja), as also in surface the Krákárbotnar palaeofan (Bárðarbunga source).

822

## 823 **5.6 Hypothetical mega-jökulhlaups**

824 The two potential mega-jökulhlaups were those (1) responsible for the limited Varghóll surge (South,  
825 ca. 11.5 cal ka) and the Hnausar surge and bursting (Jökulsá á Fjöllum; ca. 11.5 cal ka) and (2) the  
826 "Hólssandur hydraulic dunes" event (Jökulsá á Fjöllum, Barðardalur, Þjorsá Rivers; ca 10.1–9.9 cal  
827 ka). These were of the rupture type, with a permafrost infiltrated terminal moraine.

828  
829 Volcanic melt-induced floodwaters were likely rapidly transmitted down via fractures nets in the ice and  
830 the bedrock, with peak flooding lasting only for a few days (Björnsson, 1998). In the case of the  
831 climatic-melt-and-ice-rupture type, the lake discharge increased slowly to reach peak flow. Such floods  
832 are cold, dense and progressive (Snorrason et al. 1997; Flowers et al. 2003) in connection with  
833 climate warming. They probably took a much longer time to diminish, due to the more diffused  
834 geometry of the path of subglacial water migration. If a diffuse aquifer was formed, constrained by  
835 permafrost, the duration of the drainage might have been several weeks, limiting the size of the  
836 maximal outburst.

837  
838 Evidence for large prehistoric peak discharges of  $0.2 - 1.0 \cdot 10^6 \text{ m}^3 \text{ s}^{-1}$  in the Jökulsá á Fjöllum Valley  
839 has been presented by Sæmundsson (1973), Tómasson (1973, 2002) and Waitt (2002), among  
840 others. For the upper valley, Alho et al. (2005) and Carrivick et al. (2013) estimated a maximal  
841 discharge of up to  $0.9 \cdot 10^6 \text{ m}^3 \text{ s}^{-1}$  along the entire Jökulsá á Fjöllum for the "9.3 cal ka" events. At the  
842 level of Hrossaborg, we obtained a peak discharge of  $0.71 \cdot 10^6 \text{ m}^3 \text{ s}^{-1}$  to erode its top (50 m water  
843 depth) by using the flood-affected transverse profile of the valleys and the same flow speed (Alho et al.  
844 2005:  $2.0 \text{ m s}^{-1}$ ). Modeling by Gylfadóttir et al. (2017) estimated a flood discharge of  $0.1 \cdot 10^6 \text{ m}^3 \text{ s}^{-1}$ , for  
845 a flood level at the footh of the Hrossaborg. At the level of Hólssandur (Kvensodull, 10.1 – 9.9 cal ka  
846 events), we obtained a peak discharge of  $1.65 \cdot 10^6 \text{ m}^3 \text{ s}^{-1}$  with the present-day morphology, and  $1.08$   
847  $\cdot 10^6 \text{ m}^3 \text{ s}^{-1}$ , assuming an infilled canyon to 200 masl (Hólmatungur–Vestudalur terrace). At Ásbyrgi,  
848 Waitt (2002) estimated it to be  $0.7 \cdot 10^6 \text{ m}^3 \text{ s}^{-1}$ . Our observations give a similar value for an infilled  
849 canyon.

850  
851 The Preboreal jökulhlaups in Iceland were not much bigger than the others, or for the Jökulsá Rivers,  
852 not largely exceeding a maximal volume of  $1 \cdot 10^6 \text{ m}^3 \text{ s}^{-1}$ . This implies a limited significance for climate-

853 driven meltwater storage in subglacial aquifers and lakes, and their specific drainage contributions, as  
854 also stressed for the recent warming (Flowers et al. 2003). Since the volume of the Bárðarbunga  
855 caldera is  $3.6 \cdot 10^6 \text{ m}^3$ , the emptying of this basin is not sufficient to provide such a discharge for hours,  
856 but it obviously contributed to flood peaking. The volume of climate-driven meltwater has in turn  
857 impacted the loading / unloading of magmatic chambers to trigger eruptions, controlling the maximal  
858 discharge of floods obviously much less than do volcanogenic events. All volcanic / climatic  
859 combinations are possible, but during deglacial times, the duration of flood events was probably very  
860 long, of the order of a few months following the season. For these reasons, they seem less efficient as  
861 a bedrock erosion agent than commonly published. This is not true for soft sedimentary infillings.

862  
863 The canyons incisions for the different outlets of the Vatnajökull result apparently from recurrent  
864 periods of activity throughout the Quaternary. This is shown by the preservation of the very old and/or  
865 last interglacial sediments close to the valleys bottoms. The excavation of the successive  
866 unconsolidated glacial infills, perhaps reached several times the basement during the early Holocene.  
867 The Bárðardalur sedimentary infilling was fully re-excavated before the Svartarvatn advance (ca 9.3  
868 cal ka). The Jökulsá á Fjöllum valley already was deeply excavated prior to the re-infilling by the  
869 Preboreal advance (ca 11.3 cal ka); the last re-excavation started with the ca 9.9 cal ka event, as  
870 probably also in the Bárðardalur. It corresponds to the largest splayed jökulhlaups in connection to the  
871 ca 9.9 cal ka major eruption of the Bárðarbunga Volcano (Guðmundsdóttir et al. 2016) and perhaps  
872 the emptying of the whole caldera: a profound channel scours the northern side of the caldera  
873 (Björnsson, 2017). The  $^3\text{He}$  dating provided by Baynes et al. (2015) of the surface exposure of the  
874 terraces from the “9.9 cal ka” event represents the clearance by steps of the Jökulsá á Fjöllum valley.  
875 This erosion within the canyon is more efficient to incise the basement due to constricted floods.

## 876 877 5.7 Generalisation

878 The expected succession for deglacial events is not the systematic rule. A subglacial major eruption is  
879 rarely immediately followed by the formation of a large lake, a surging in form of jökulhlaup or glacial  
880 surge and a final emission of phreatomagmatic ash. Some steps are often missing or delayed in time.  
881 The Preboreal megafloods revealed in Iceland that climate-driven flood events from ice-cap internal  
882 storage are generally long, with relatively limited discharges, but potentially large volumes. This can be

883 compared to dam breakage as for the Missoula Lake (Wait, 1985) or for the Proglacial Lake in Altai  
884 (Rudoy et al, 1993). The piking of the flood discharge occurs as well with dam breakage as with  
885 subglacial eruption. Surging of glaciers is frequent today as in Preboreal times. These floods are  
886 generally associated with major deglaciation events, excepted when the subglacial volcanic activity is  
887 raised by a major glacio-isostatic unloading. Wide water lateral-splay are most probably linked to  
888 permafrost persistence in the valley and do not necessarily imply larger flood volumes.

889 The canyons incision are for a major part an inheritance of quaternary glaciations. This morphology is  
890 commonly recycled with recurrent glaciations, following the same flow lines. It has been also  
891 demonstrated for the ice-dammed Missoula Lake (Waitt, 1985; Clague et al. 2003). The erosion  
892 capability of megafloods have been exaggerated, even more efficient in constricted conditions. The  
893 youngest incision in glacial environment proceeds by clearance of the successive glacial pulse  
894 accompanying the deglaciation under control of the glacio-isostatic rebound. Only the full interglacial  
895 flows will incise efficiently the bedrock.

896

## 897 **6. Conclusions**

898 Deglaciation events are almost synchronic in Iceland with the surrounding north Atlantic regions, as far  
899 the accuracy of the dating may allow correlations; they are under control of the Irminger Sea Current.  
900 The Preboreal jökulhlaups in Iceland fully correlate with both the deglaciation events and the  
901 subglacial volcanic activity. This succession, initially triggered by climate change, and responsible for  
902 superficial melting and volcanic activity, led to cascading retroactive events. The succession occurred  
903 at least twice close to 11.5 – 11.3 cal ka and around 10.3 to 10.1 – 9.9 cal ka. Minor events existed  
904 during the Alleröd / Younger Dryas, and also probably occurred in association with the “9.3 cal ka”  
905 deglaciation, with an already restricted ice-cap mass. These Preboreal jökulhlaups were not much  
906 larger than the others, and never largely exceeded a flood of  $1 \cdot 10^6 \text{ m}^3 \text{ s}^{-1}$ . They splayed on partially  
907 frozen ground with a limited incision capability, often driven to the East by the Coriolis forces. Ongoing  
908 glacial rebound to 10.3 cal ka temporarily lowered the slopes of the valleys, limiting the clearance and  
909 incision of the canyons. The Saksunarvatn Tephra mostly marked the end of the main phase of  
910 deglaciation, at 10.3 – 9.9 cal ka, in relation to the temperature rise to the thermal optimum. It also  
911 signaled the onset of interglacial activity for the Barðarbungá and Grímsvötn Volcanoes. These close  
912 interrelations between climate and volcanic activity for generating jökulhlaups of long duration during

913 major deglaciation events can be easily applied to other volcanic deglaciated regions such as  
914 Western Antarctica, Alaska or Oregon.

915

916

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925

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1312

1313 **FIGURES CAPTIONS**

1314

1315 Figure 1: A) Early Holocene jökulhlaup trajectories mapped with field observations and photographic  
1316 interpretation. B) Cartographic illustration of the deglaciation history of Iceland, based on  
1317 Kaldal and Víkingsson (1990) and recent geological maps (Sæmundsson et al. 2012;  
1318 Sigurgeirsson et al. 2015), complemented by new photographic and/or field observations,  
1319 tephrostratigraphy and correlations explained in the text.

1320

1321 Figure 2 Map of the maximal extent (Early Holocene) of floods in A) the Skjálfandafjót and B) the  
1322 Jökulsá á Fjöllum Rivers. Light blue corresponds to the mapped largest flood extent, probably

1323 9.9 cal ka. Dark blue corresponds to the Late Holocene jökulhlaup pathways. Present-day  
 1324 glacial limits. Explanations and dating in § 4 Data.

1325

1326 Figure 3: Chronology of the recorded tephra and potential floods in relation to climate, 8–15 ka BP:  
 1327 potential impact of deglaciation induced by climate warming. Arrows – potential jökulhlaup  
 1328 events; SKA – Skjálfandafhlöt River; JF – Jökulsá á Fjöllum River; SDJ – Skeiðararjökull  
 1329 Glacier, BRU: Jökulsá á Brú

1330 A) Warming potentially linked to solar activity, derived from  $\delta^{14}\text{C}$  (Stuiver and Braziunas,  
 1331 1988). Entire Saksunarvatn Tephra record from the literature shown in grey.

1332 B) Regional mean temperatures from the NGRIP  $\delta^{18}\text{O}$  isotope curve (NGRIP, 2004). For  
 1333 tephra sources, see Table 1; SAKS – range of the Saksunarvatn event.

1334 Tephra from Grímsvötn Volcano in black; Bárðarbunga Volcano in red; tephra from Askja  
 1335 Volcano in white, other rhyolitic tephra in yellow. The position of glacial advances are shown in  
 1336 blue. PBE–Preboreal cooling event; AL– Alleröd; ACE– Alleröd cooling event; BCE– Bölling  
 1337 cooling event.

1338

1339 Figure 4: map of the largest floods and morainic arc (partly eroded) preserved in the lower Jökulsá á  
 1340 Fjöllum.

1341

1342 Figure 5. Hnausar.(images from Google Earth [GE]). A) Deglaciation paleolake. B) Glacial surge (11.2  
 1343 cal ka). C) Piercing by jökulhlaup. D) Tephrostratigraphic record.

1344

1345 Figure 6: Evidence of jökulhlaups in the Hólssandur Plateau at 380 m (images GE). A) Large, flat  
 1346 lozenge-shaped hydraulic dune at Kvensöðull. B) Details of the dunes, showing evidence of  
 1347 surface megaripples, and terrace levels from the end of the flood. C) Evidence south of A of a  
 1348 residual terminal moraine and cinder cones, to the west, both eroded by floods, Note s, eroded  
 1349 by floods.

1350

1351 Figure 7: Evidence of terminal moraines on the Jökulsá á Fjöllum and Þjórsá–Tungnaá Rivers. North  
 1352 is upwards. A) Relict of a terminal moraine, SW of the Hrossaborg, eroded by floods – the

1353 Hrossaborg Advance, c. 10.3 cal ka (image Landmælingar Island [LmIs]). B) Terminal moraine  
 1354 that was issued from the Bruarjokul outlet glacier, NE of the Vaðalda – the Kreppa Advance, c.  
 1355 9.3 cal ka pierced by jökulhlaup (image from LmIs). C) Kisa Advance, c. 9.3 cal ka, overlapped  
 1356 in the east by the Þjórsá Lava (8.7 cal ka; image LmIs ). D) Megaripples in basaltic sands (25  
 1357 m wide) at the SW foot of the Vaðalda Volcano (Sample Dingjuvatn D3; image LmIs). The  
 1358 North is upward.

1359  
 1360 Figure 8: Paths of the early Holocene jökulhlaup in the Skalfandaflljót, Fnjoskadalur and Eyafjörður.  
 1361 Notice the location of disrupted glacial tongue that hypothetically (?) also reached the lower  
 1362 Skalfandaflljót, but have been now eroded.

1363  
 1364 Figure 9: Ljósavatn and Háslón records. A-B) Ljósavatn pitted moraine. Evidence of jökulhlaup  
 1365 deposits, with iceberg kettle hole in juxtaglacial position The final infilling reworks an Askja  
 1366 tephra much older than the Askja S (Askja PB), Stórutjarnaskóli Quarry. B) Loess section  
 1367 above the Preboreal tillite; the Askja PB is involved with the loadcatsing in the grey loess. C-D-  
 1368 E: Upper Háslón Lake record. C) Deglacial braided sandur, reworking the Askja E Tephra; D)  
 1369 Twinned Saksunarvatn Tephra deposits in juxtaglacial position, deformed by late, but limited,  
 1370 glaciotectionics. E) Undeformed Saksunarvatn Tephra, buried by jökulhlaup deposits and loess  
 1371 in a lateral valley (Van Vliet-Lanoë et al. 2010).

1372  
 1373 Figure 10: Composite record of main sections, with reference to the main sources of tephra. JL –  
 1374 jökulhlaup deposits, GI – tillite. Iceberg scours at Varghóll and iceberg thermokarst at  
 1375 Ljósavatn. Classic thermokast (kettle hole) at Palmholt. Sections are located on figures 2-4-9-  
 1376 10.

1377  
 1378 Figure 11: A, B) Detail of the N moraine of the Eyjabakkajökull Glacier (images LmIs), the thick white  
 1379 arrow indicate 1890 stacked “push” moraines following Schomaker et al. 2014). C) Potential  
 1380 evidence of stacked, permafrosted moraines (Hólssandur). D) Stacked, permafrosted moraine,  
 1381 Usherbreen Glacier, Svalbard (image courtesy of J.O. Hagen).

1382

1383 Figure 12: The South-West A) Esker at Búðarháls, eroded by a jökulhlaup. B–D) Varghóll section, with  
1384 terminal ridges (thick arrows on C; images Lmls), iceberg scours (B) incising a basaltic tephra  
1385 deposit containing Vedde Ash pellets (stars), and a tillite stacking pattern (D).

1386

1387 Figure 13: Routes (black) of the early jökulhlaups in relation to the Holocene glacial advances (10.3 ka  
1388 – white, 9.3 ka – yellow). Potential subglacial lake / aquifers extents are shown in blue, related  
1389 to a 125 m and 250 m residual glacio-isostatic subsidence. Black stippled line – present-day  
1390 glaciers, white stippled line – 10.3 cal ka glaciers

1391

1392 Table 1: Chronology of the significant eruptions of Grímsvötn, Bárðarbunga, Askja, Katla and Krafla  
1393 volcanoes from Late Glacial to 8.2 cal ka.

1394 Table 2: Geochemical analysis of the tephra (ICPMS AES and \* microprobe average; Microprobe on  
1395 Supplement table 1)

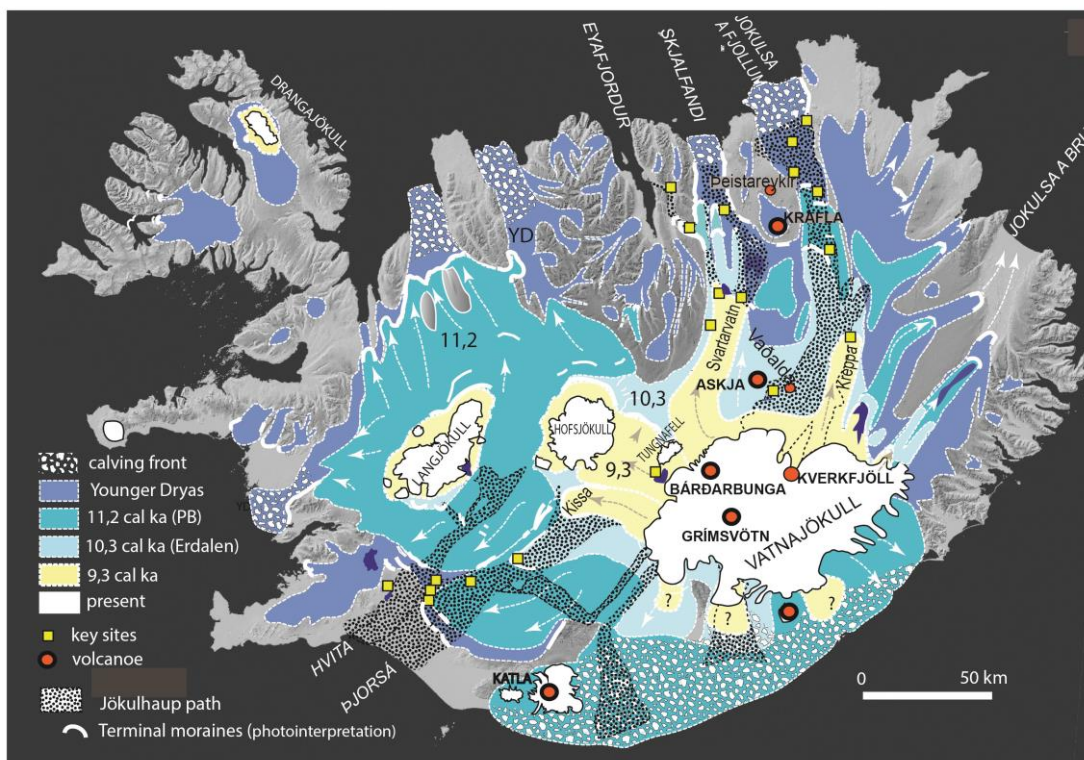
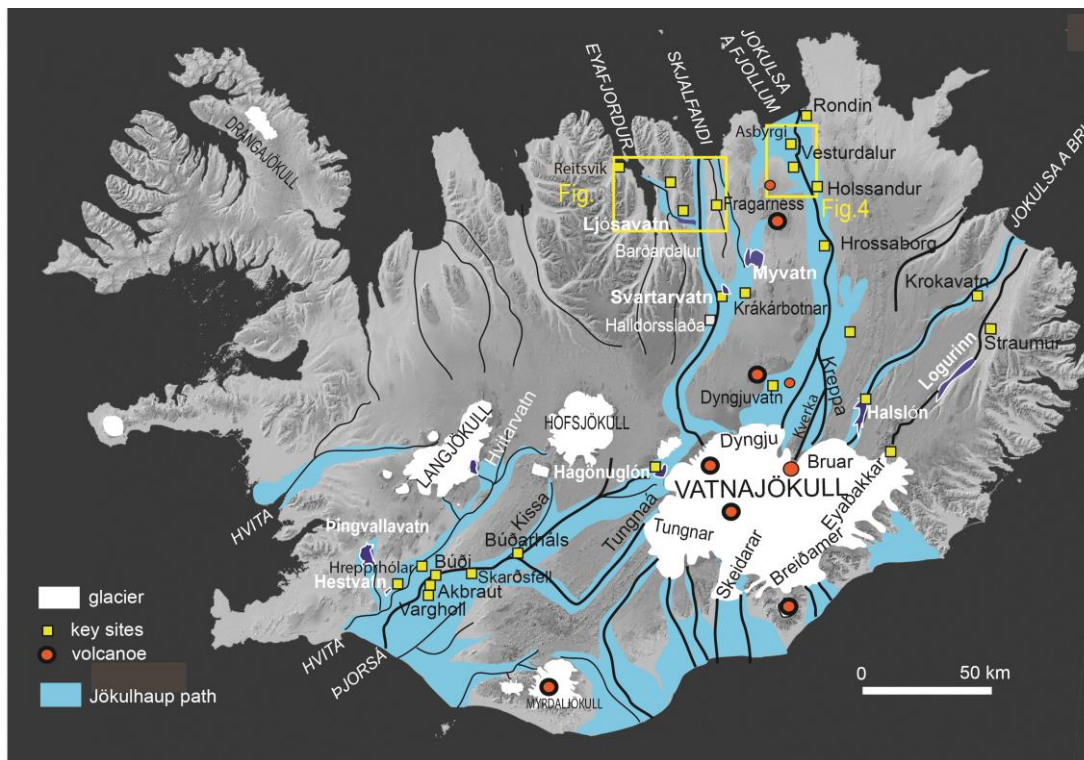
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1397 **Supplement available on**

1398 [https://static-content.springer.com/esm/art%3A10.1007%2Fs00531-020-01833-](https://static-content.springer.com/esm/art%3A10.1007%2Fs00531-020-01833-9/MediaObjects/531_2020_1833_MOESM1_ESM.pdf)

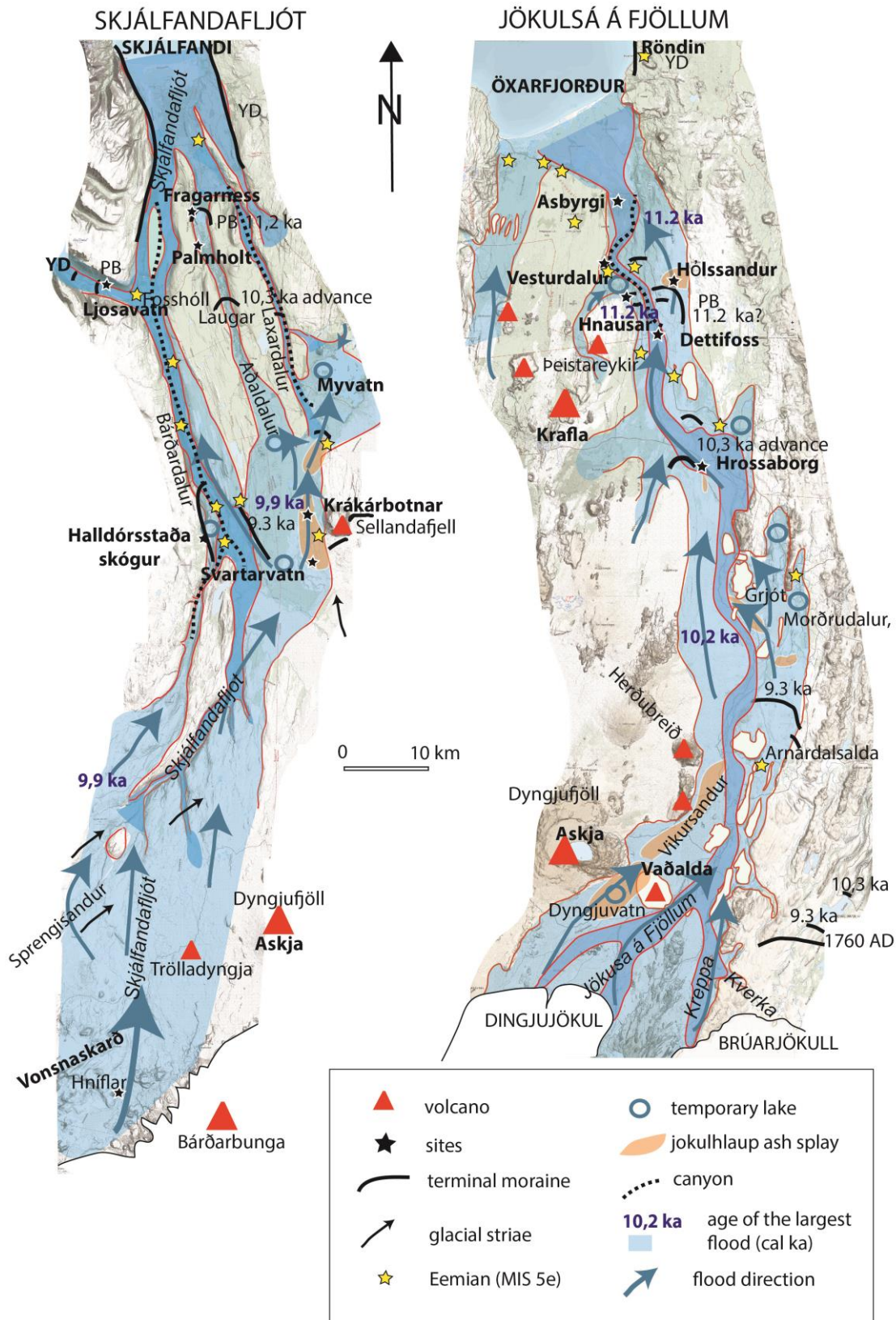
1399 [9/MediaObjects/531\\_2020\\_1833\\_MOESM1\\_ESM.pdf](https://static-content.springer.com/esm/art%3A10.1007%2Fs00531-020-01833-9/MediaObjects/531_2020_1833_MOESM1_ESM.pdf)





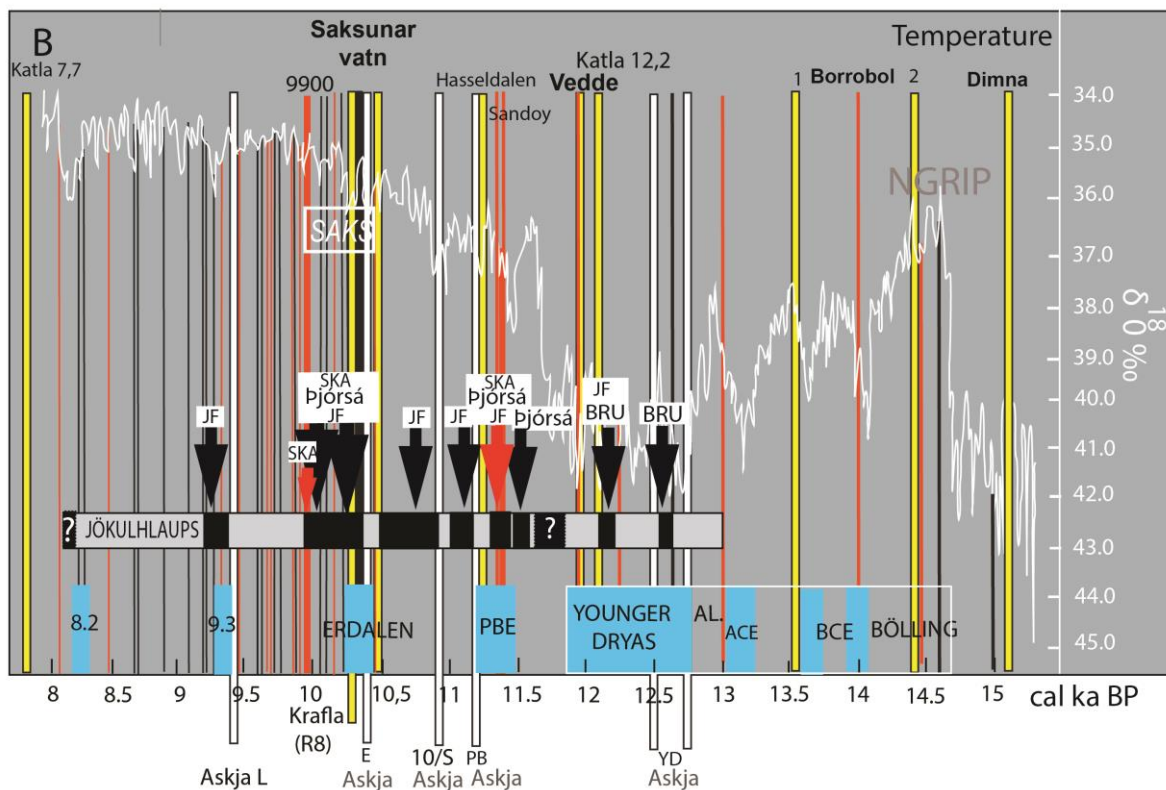
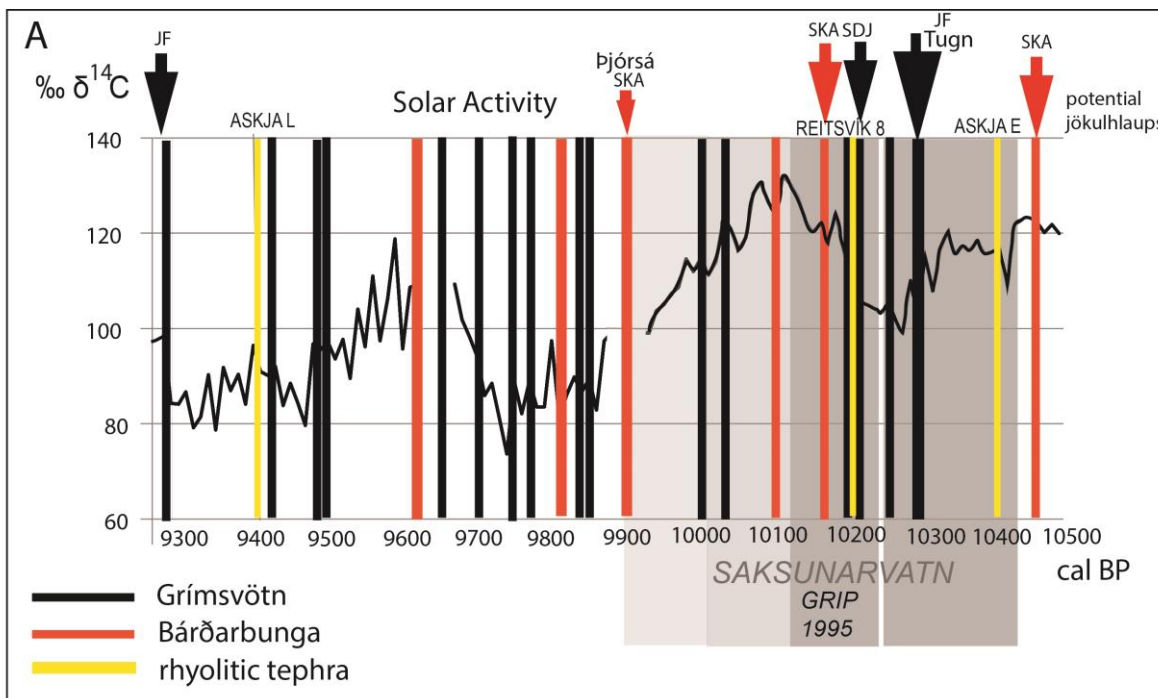
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1401 Fig.1



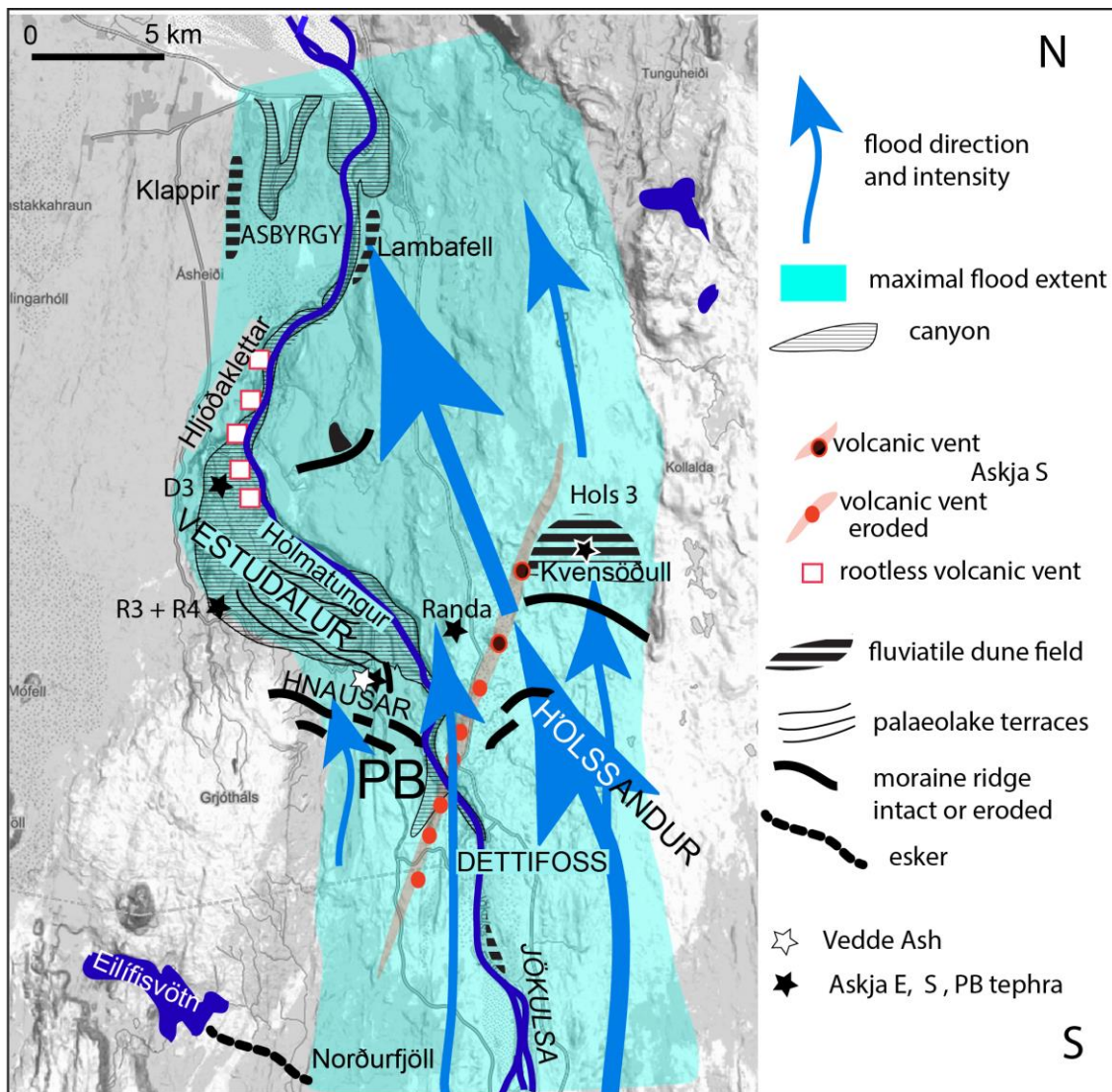
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1403 Fig.2



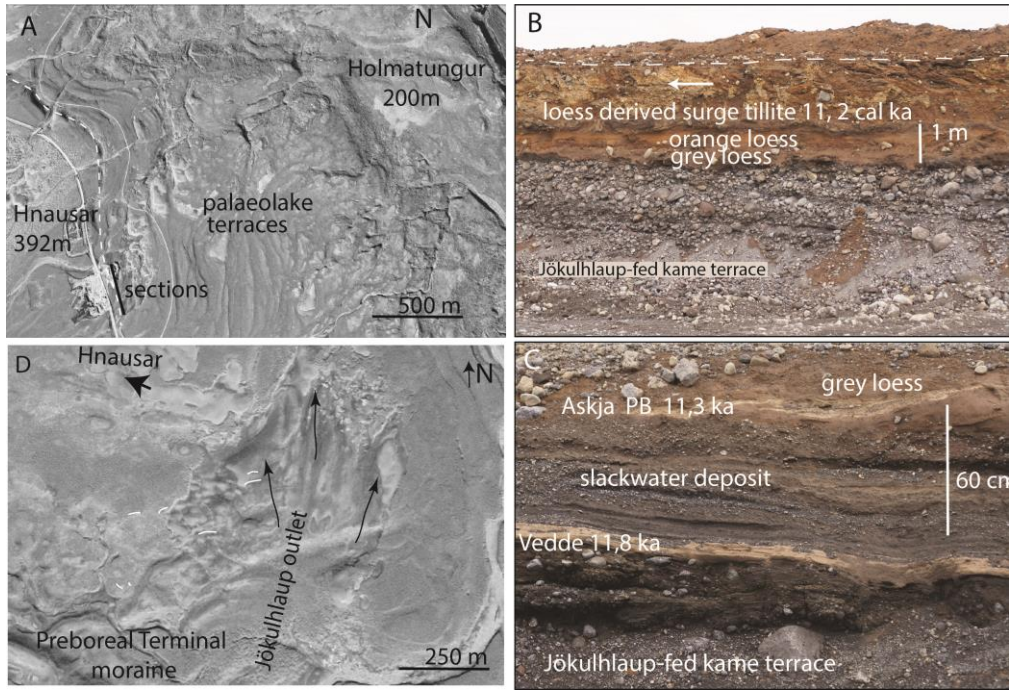
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1405 Fig.3



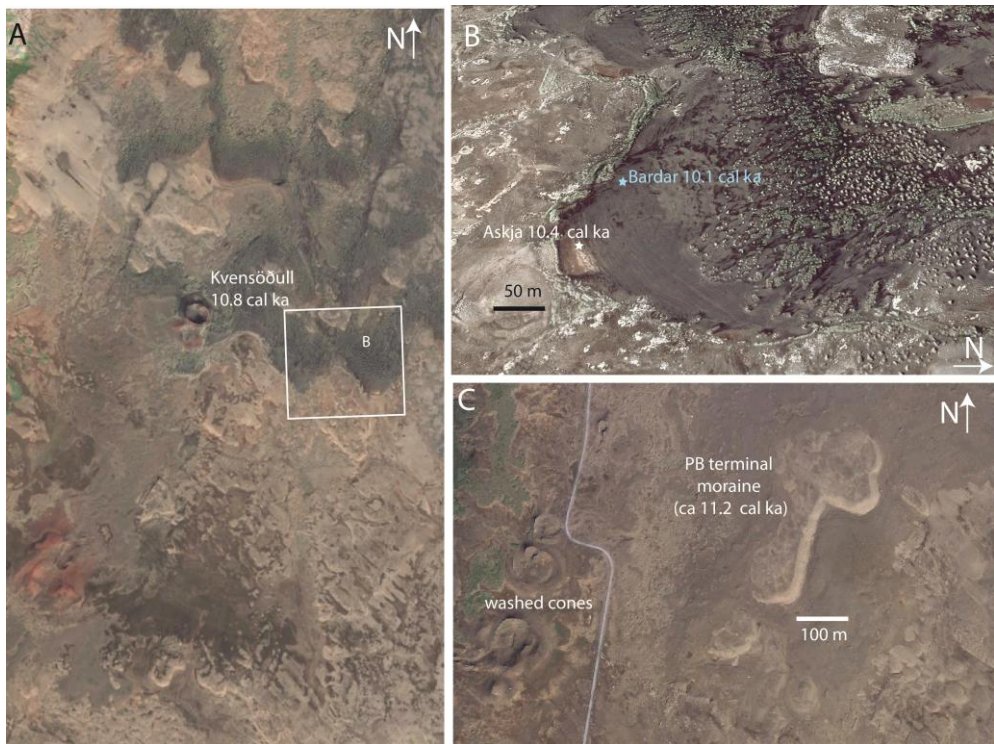
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1407 Fig.4



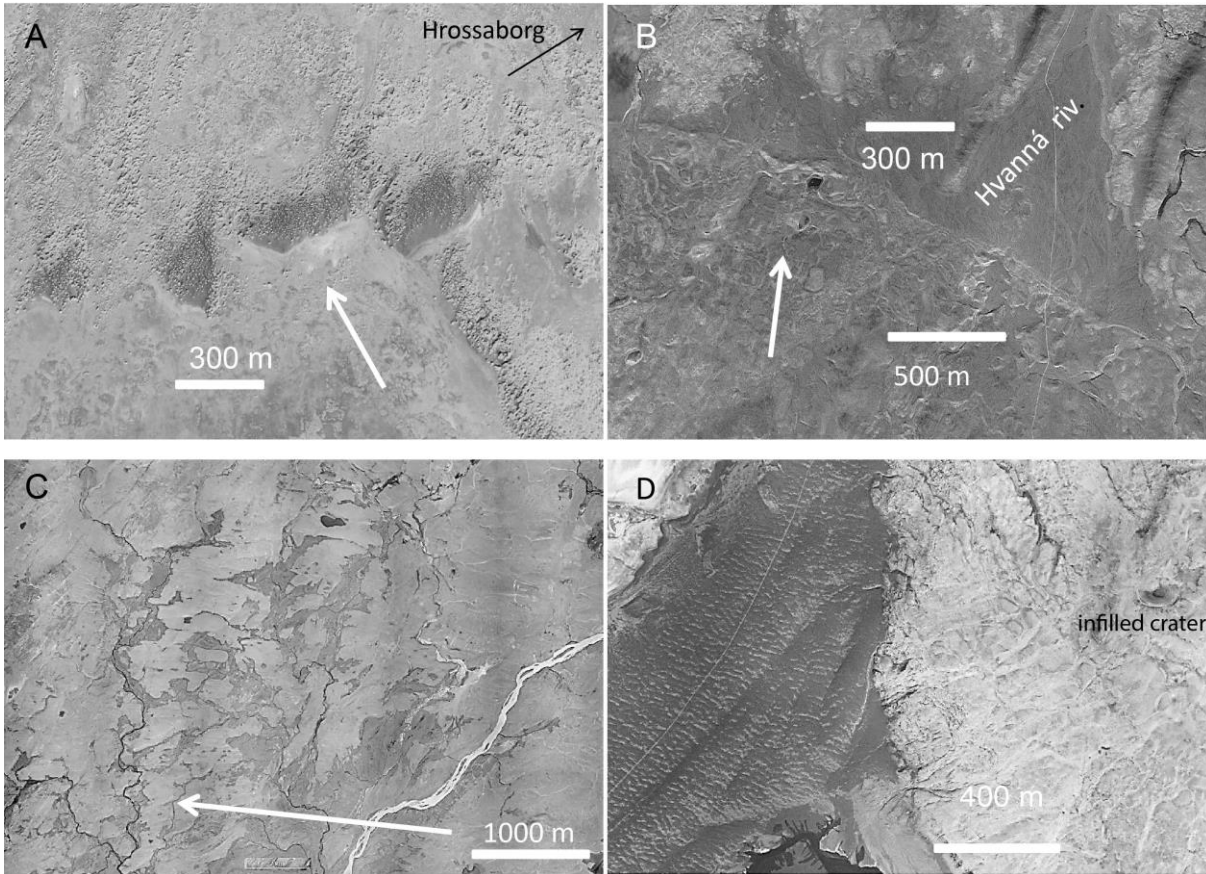
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1409 Fig.5



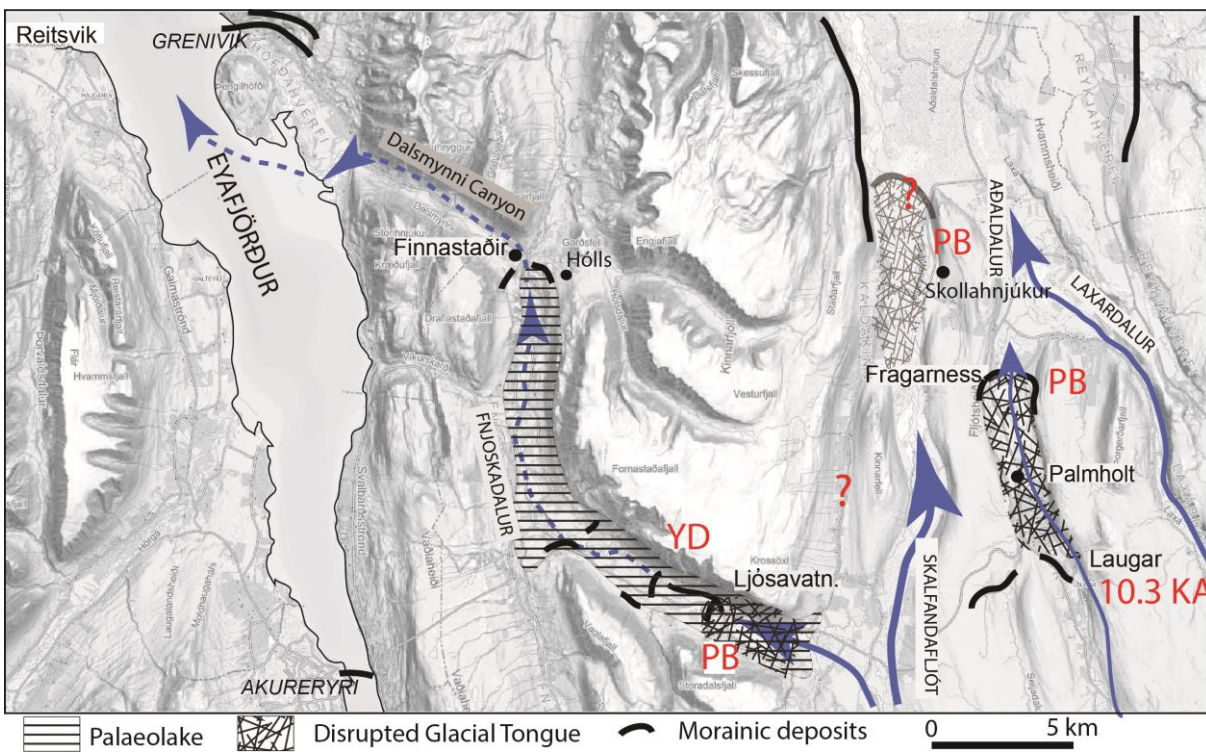
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1411 Fig.6



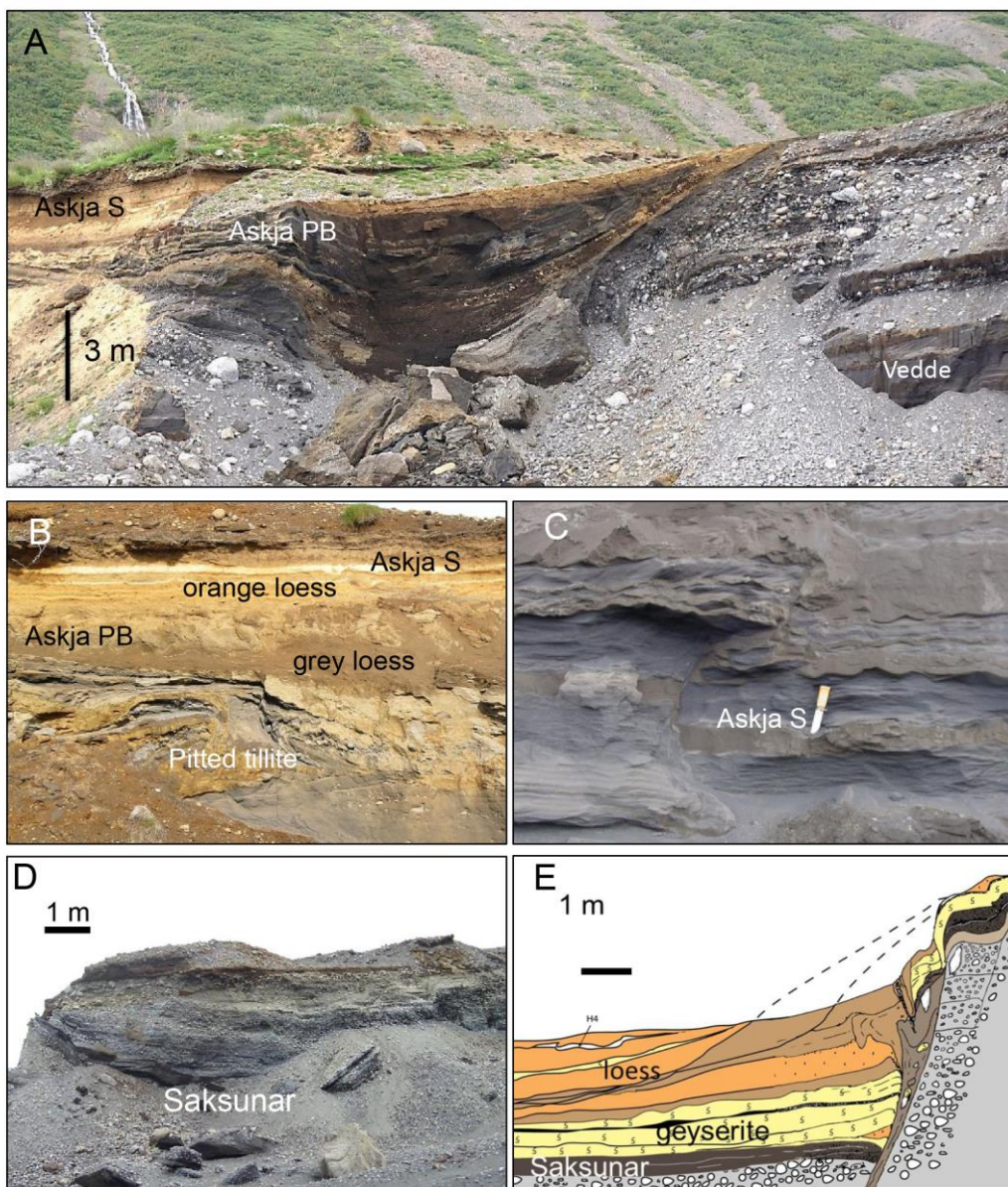
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1413 Fig.7



1414

1415 Fig.8



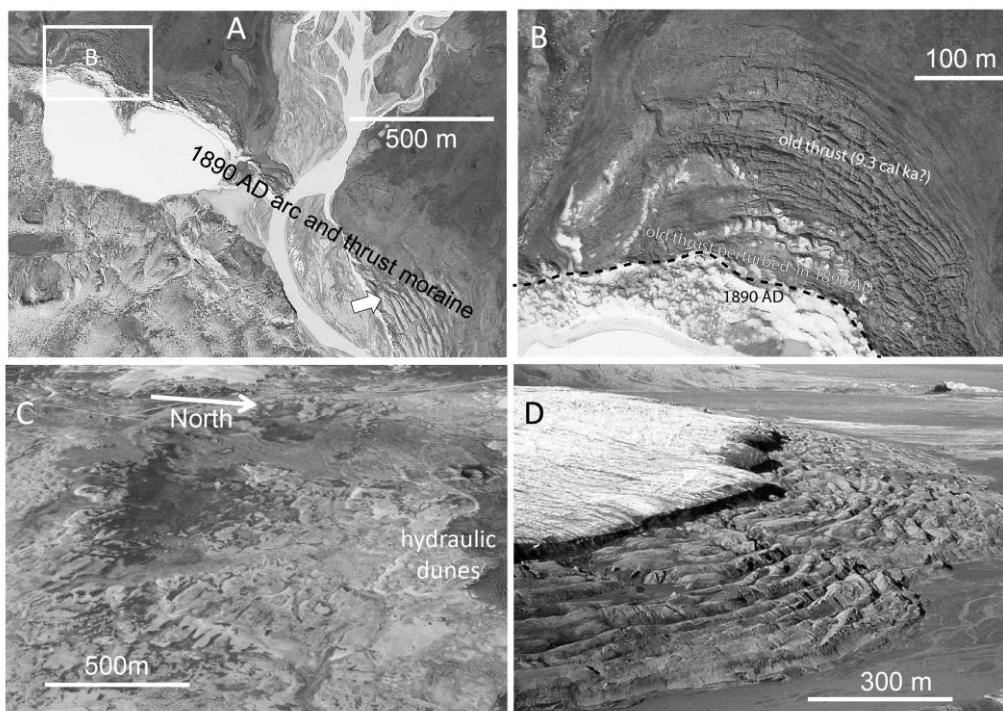
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 1417

Fig.9

1418

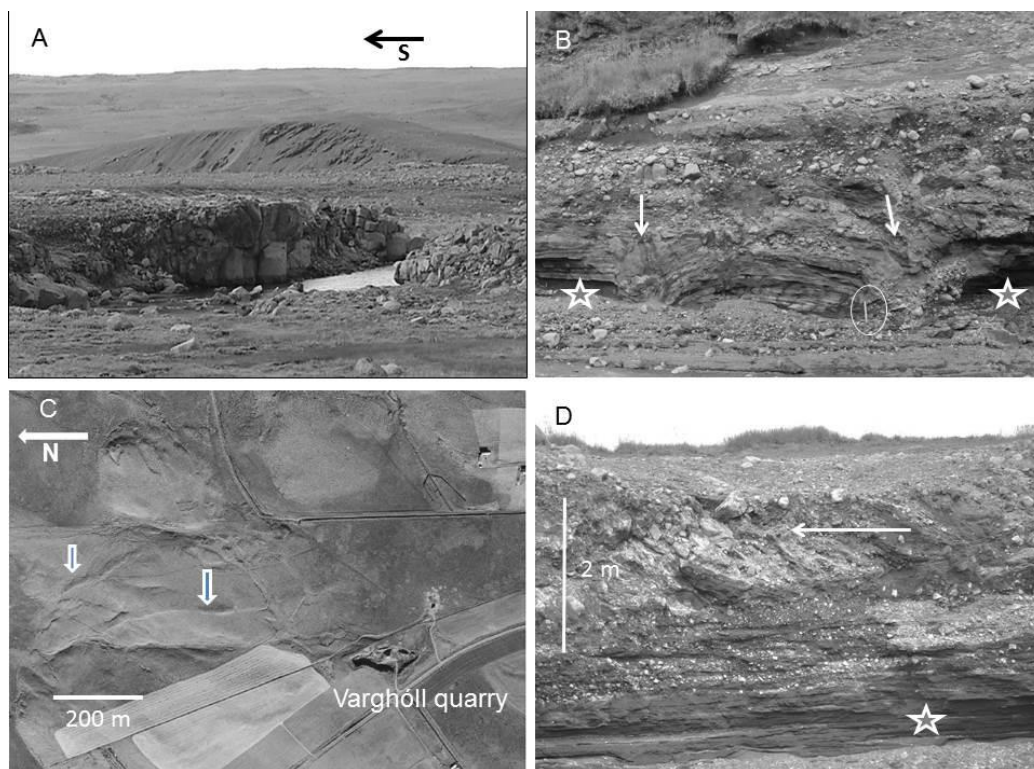






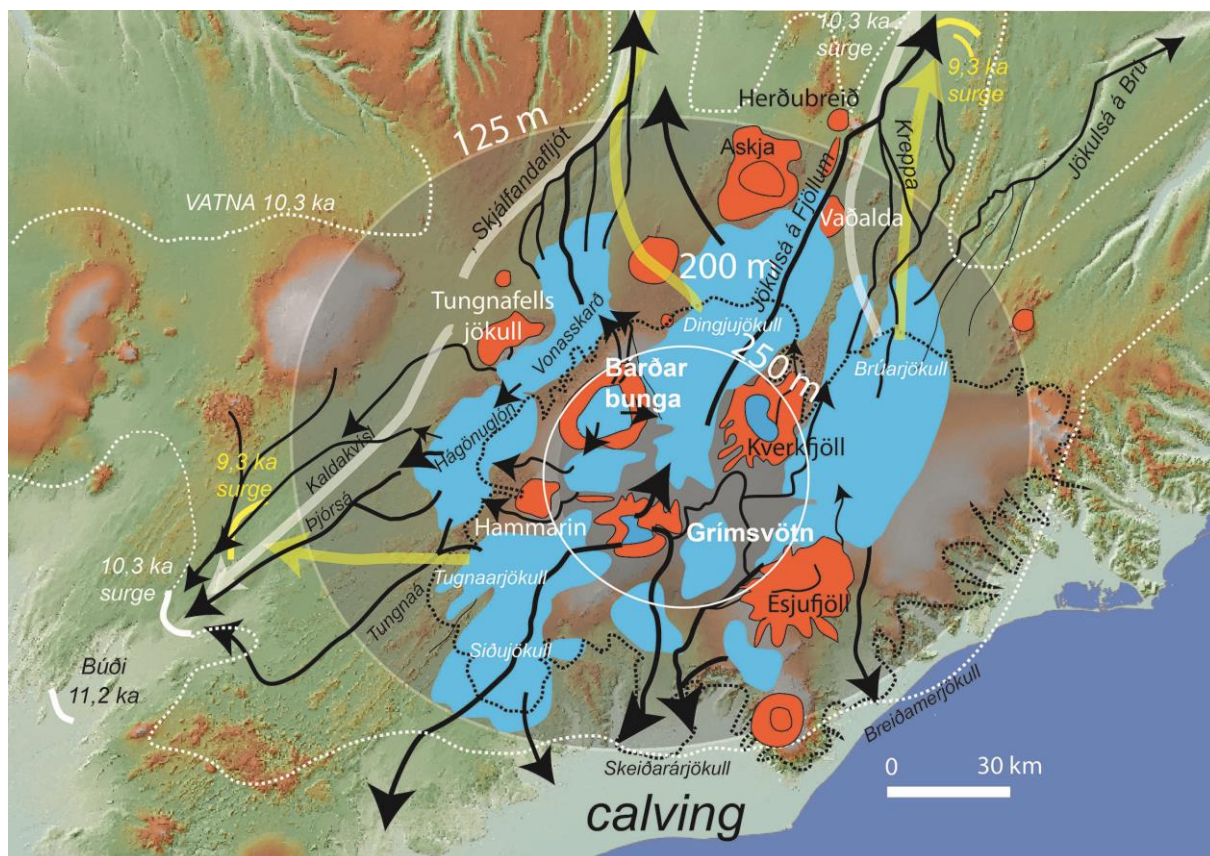
1421

1422 Fig.11



1423

1424 Fig.12



1425

1426 Fig.13