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1 **Eocene continental breakup in Baffin Bay**

2

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22

23 **Keywords**

24 - Baffin Bay opening

25 - West Greenland Volcanic Province

- 26 - Seaward Dipping Reflectors
- 27 - Volcanic passive margin
- 28 -  $^{40}\text{Ar}/^{39}\text{Ar}$  and K-Ar dating
- 29 - Paleomagnetism

## 30 **Abstract**

31

32 We question the timing of continental breakup and early oceanization in Baffin Bay, North-  
33 East Atlantic. North of the Ungava fault zone, the breakup was syn-magmatic and led to the  
34 development of conjugate volcanic passive margins (VPMs). We investigated the innermost part of  
35 the W-Greenland VPM where a remarkable inner-SDR is fully exposed in the Svartenhuk area. Our  
36 new radiometric ages and paleomagnetic data from syn-tectonic basaltic lavas indicate that  
37 continental stretching and thinning spanned the C26r to, at least, the C24r period, giving an Eocene  
38 lower boundary age for continental breakup in Baffin Bay. These results contradict the proposed  
39 flooring of Baffin Bay by a Paleocene oceanic crust older than C24n and also question the accretion of  
40 oceanic crust before C22. We confront our results to the dynamics of the northward oceanic-rift  
41 propagation across the Ungava transform fault system, and we suggest that plate breakup in Baffin  
42 Bay occurred ~8 m.y. later than in N-Labrador Sea as a result of the thermal and mechanical barrier  
43 effect induced by the Ungava transform zone.

44

## 45 **1. Introduction**

46

47 The extinct Labrador-Baffin oceanic axis (Fig. 1) is part of a wide Phanerozoic divergent  
48 system including the NE-Atlantic Ocean, resulting from the Tertiary continental breakup of Eurasia,  
49 Greenland and North-America (Fig. 1a). On both sides of Greenland, the Paleogene plate breakup was  
50 preceded and accompanied by abnormal mantle melting (Saunders et al., 1997). The Labrador-Baffin  
51 axis is a typical example of oceanic rift propagation across an inherited transform fault (the Ungava  
52 fault zone) in connection with abnormal mantle melting. A recent analysis by Koopmann et al. (2014)  
53 suggested that early transform faults may act as rift propagation barriers and could enhance the  
54 amount of syn-breakup magmatism. We hereafter summarize the tectonic evolution of both the  
55 Labrador Sea and the Baffin Bay based on published data, as well as the related age uncertainties  
56 investigated in this study.

57

58           Following minor Late-Triassic to Late Jurassic magmatic dilatation (Watt, 1969; Larsen et al.,  
59 2009), the earliest synrift successions in the Labrador Sea are Early Cretaceous in age (Chalmers and  
60 Pulvertaft, 2001; Dickie et al., 2011; Jauer et al., 2015). They are represented by the volcanism of the  
61 Alexis formation and by wedge-shaped deposition of the Upper Bjarni and Appat sequence within the  
62 half grabens formed along the Labrador shelf (Fig. 2; see Chalmers, 2012 for a review). In the  
63 northern part of the Labrador Sea, a second episode of tectonic stretching spanned the Late  
64 Cretaceous until, possibly, the earliest Tertiary (Chalmers and Pulvertaft, 2001). Here, the oldest  
65 undisputed oceanic magnetic anomaly is C27n (Chalmers and Laursen, 1995), i.e. 62.2 – 62.5 Ma  
66 according to Ogg (2012). The “transitional” crust (Fig. 1c) bearing magnetic lineaments older than  
67 C27n has been modeled either as magma injected stretched continental crust (Chalmers, 1991,  
68 Chalmers and Laursen 1995), as slivers of stretched continental crust lying on exhumed serpentinized  
69 mantle (Chian and Loudon, 1994) or as formed by slow seafloor spreading between C33 and C27 (Fig.  
70 2; Strivastava and Roest, 1999).

71

72           The initial spreading directions were oriented ~N060E, i.e. sub-normal to the earlier  
73 continental rifted structures (Roest and Strivastava, 1989). Between C25 and C24, the oceanic  
74 fracture zones and magnetic anomalies indicate a major shift in kinematics in the Labrador Sea (Fig.  
75 1). Spreading directions rotated from N060E to N010E (Roest and Strivastava, 1989; Fig. 1b). Oceanic  
76 spreading ceased in the Labrador Sea during the Oligocene (C13, Roest and Srivastava, 1989; Fig. 1a).

77

78           The SW-NE-trending left-lateral Ungava transform-fault system (UFS) connects the Labrador  
79 Sea to Baffin Bay (Fig. 1c). This area acted as a transtensive fault zone up to C25, becoming  
80 transpressive after C25 (Chalmers and Pulvertaft, 2001; Geoffroy et al., 2001; Suckro et al., 2013).  
81 Related folding structures are seismically well imaged on both sides of the Davis Strait High (Fig. 1c,  
82 e.g. Chalmers, 2012). The compressional deformation ceased in mid-Eocene times (Dalhoff et al.,  
83 2003).

84           Continental extension within Baffin Bay was also polyphase, defining an early continental  
85 sedimentary rift between Greenland and Baffin Island (Altenberndt et al., 2014; Gregersen et al.,  
86 2016; Fig. 1c). Extension would have also started during the Aptian-Albian (Kome Fm. in Fig. 2) and  
87 renewed in latest Campanian (see Dam et al., 2009 and Chalmers, 2012 for a review).

88



89 Linear gravity lows located in the central part of the Baffin Bay are interpreted as extinct  
90 spreading centers (Whittaker et al., 1997; Chalmers and Pulvertaft, 2001; Fig. 3a), but the offshore  
91 Baffin Bay depths are floored with non-interpretable magnetic anomalies (Fig. 3b). Crustal  
92 thicknesses and velocities from seismic refraction data are nevertheless consistent with the presence  
93 of oceanic crust in the deepest part of Baffin Bay (Fig. 1c; Keen et al., 1974; Suckro et al., 2012; Funck  
94 et al., 2012). Two major N-S-trending fracture zones in the basin (Fig. 3a) suggest also that oceanic  
95 spreading, if any, may have been coeval with the ~N-S reorientation of the kinematic trend in the  
96 Labrador Sea during the Eocene (Fig. 2; Geoffroy et al., 2001).

97  
98 No absolute dating of any acknowledged oceanic crust exists in Baffin Bay, where there is no  
99 widely recognized pattern of linear magnetic anomalies (Fig. 3b). However, recent contributions  
100 (Oakey and Chalmers, 2012; Suckro et al., 2012; Hosseinpour et al., 2013) propose that oceanic-type  
101 crust of mid-Paleocene age (C27) would exist within the Baffin Bay. Oakey and Chalmers (2012) show  
102 that the pattern of transform faults in Baffin Bay is consistent with that of the Labrador Sea according  
103 to a common Euler rotation pole during the C24-C13 period. These authors state that older basement  
104 surrounding their postulated Eocene-Oligocene oceanic domain in the Baffin Bay is Paleocene. They  
105 further reinterpret shipborne magnetic profiles to reveal a Paleocene spreading center and  
106 associated magnetic anomalies, and thus draw limits for both Eocene and Paleocene oceanic crust  
107 domains in the Baffin Bay (Fig. 1c). However, none of the related studies takes into account the age of  
108 continental stretching recorded within the Baffin Bay continental passive margins.

109  
110 A precise location of the continent-ocean boundary and a refined estimate of the age of the  
111 onset of continental extension and breakup in Baffin Bay are critical constraints to understand the  
112 dynamics of the propagation of continental breakup between Greenland and NAM plates from apart  
113 the Ungava fault zone (Fig. 1). In this contribution, we investigate from a detailed onshore study the  
114 precise timing of syn-magmatic continental extension along the SE-Baffin Bay passive margin in order  
115 to generate a time constraint on continental breakup and earliest oceanic accretion in Baffin Bay.

116

## 117 **2. Geological setting**

118

### 119 **2.1. The volcanic passive margins**

120

121 Volcanic passive margins (VPMs) are characterized by significant mantle melting and magma  
122 emplacement coeval with continental extension and final breakup (e.g. White et al., 1987). The  
123 upper-crustal section of conjugate VPMs is associated with two kinds of SDRs (Seaward-dipping  
124 seismic reflectors; Planke et al., 2000), both of them developing seaward with time: the inner SDR  
125 wedges and the outer SDR series (e.g. Planke et al., 2000; Franke et al., 2010). Inner-SDRs (Planke et  
126 al., 2000) represent wedges of syn-tectonic lavas developed over continentward-dipping detachment  
127 faults (e.g. Geoffroy, 2005; Quirk et al., 2014; Geoffroy et al., 2015). They represent the uppermost  
128 crustal section at VPMs and are clearly associated with continental stretching and thinning. They  
129 develop in the necking zones of VPMs (e.g. Geoffroy, 2005). The detachment faults bounding those  
130 inner-SDRs root at the top of a thick high-velocity lower crust (HVLC in Funck et al., 2016) of  
131 continental origin. This magma-injected lower crust shows a ductile behavior during crustal extension  
132 (e.g. Geoffroy et al., 2015; Clerc et al., 2015).

133

134 The interpretation of outer-SDRs is far more conjectural. No observable continentward-  
135 dipping faults bound outer-SDRs (e.g. McDermott et al., 2018). They form regular convex upward  
136 reflector sequences that deepen downward and merge along a planar low-dipping and seismically  
137 opaque area separating the extrusive sequences from the underlying mafic lower crust (Quirk et al,  
138 2014; Type IIa in McDermott et al., 2018). The nature of the lower crust is unclear as there is  
139 insufficient data for an unequivocal characterization. Linear magnetic anomalies are recorded over  
140 domains covered with SDRs (e.g. Stica et al., 2014; Collier et al., 2017). Therefore, defining the  
141 continent-ocean boundary at the VPMs is a complex issue, as it could be located far away from the  
142 seaward limit of the necked part of the continental crust, i.e. within the so-called “oceanic crust  
143 domain” (Geoffroy et al., 2015).

144

## 145 **2.2. The Baffin Bay VPMs**

146

147 South of the Upernavik Escarpment (UE in Fig. 1c), the passive margins of Baffin Bay are  
148 volcanic passive margins (Geoffroy et al., 2001; Gregersen et al., 2013). Inner SDRs have been  
149 identified off SE Baffin Island and Central West Greenland (Fig. 1c; Skaarup et al., 2006; Block et al.,  
150 2012; Abdelmalak et al., 2018). Their age and along-strike extent are presently unknown. However,  
151 the inner part of the SE-Baffin Bay VPM is well-exposed in the Disko-Svartenhuk region, which is part  
152 of the so-called Western Greenland Volcanic Province (WGVP, Fig. 4). In this area, continental  
153 extension, coeval with regional magmatism, has never been dated with precision. Consequently, the  
154 timing of oceanic breakup in Baffin Bay is not constrained.

155

156 The WGVP corresponds to a variably thick pile of ultramafic to mafic lavas (from few  
157 hundred metres up to 7 km) crosscut by sheeted intrusions (e.g. Pedersen et al., 2002). The northern  
158 part of the WGVP represents the landwardmost exposure of the SE-Baffin Bay passive margin. In the  
159 inner part of the margin, the lava pile forms a flat-lying basaltic plateau up to 2 km thick. A crustal  
160 flexure has developed to the West, associated with a 4 to 7 km-thick lava wedge (Larsen and  
161 Pulvertaft, 2000). This wedge represents a syn-magmatic roll-over anticline developed over a major  
162 continentward-dipping normal fault, i.e. a fully-exposed inner-SDR (Geoffroy et al., 1998; 2001,  
163 Abdelmalak et al., 2012). The Arfertuarssuk Fault represents the master fault bounding the  
164 Svartenhuk SDR (Figs. 4 and 5).

165

### 166 **2.3. Stratigraphy and previous dating**

167

168 In agreement with former studies (e.g. Larsen and Grocott, 1992), we divided the Svartenhuk  
169 volcanic formations into four successive sequences (V1 to V4, from bottom to top) (Fig. 5). Our V1  
170 and V2 formations correlate with the lower (subaqueous) and upper (subaerial) Vaigat picritic  
171 sequences, respectively. The V3 and V4 units correspond to the Svartenhuk and Naqerloq formations  
172 (Larsen et al., 2015), respectively (Fig. 4).

173

174 Recent  $^{40}\text{Ar}/^{39}\text{Ar}$  datings are available from the Paleogene lava flows of W-Greenland (Larsen  
175 et al., 2015; Fig. 4a). The corresponding sampling was aimed at dating the Tertiary volcanic activity  
176 and was not focused on the timing relationships between magmatism and extension. Nevertheless,  
177 these data significantly improved the regional stratigraphic chart, based on former absolute datings  
178 and paleomagnetism (Figs. 4a and 4b; Storey et al., 1998; Riisager and Abrahamsen, 1999; Riisager et  
179 al., 2003).

180

181 According to these results, the ages of volcanic activity in W-Greenland span from C27n  
182 (late Paleocene) to C24r (Early Eocene). The earliest pulse is mid-Paleocene in age, spanning chrons  
183 27n to 25r (Riisager and Abrahamsen, 1999; Riisager et al., 2003; Larsen et al., 2015), i.e. 62.5 to 57.7  
184 Ma (after Ogg, 2012). Both Storey et al. (1998) and Larsen et al. (2015) suggest a hiatus of 1-2 m.y. or  
185 more at the end of this period (Fig. 4b). The second magmatic pulse, associated with transitional to  
186 alkaline basaltic successions, is dated back to 57-54 Ma, i.e. Early Eocene, during C24r (Storey et al.,  
187 1998; Riisager et al., 2003; Larsen et al., 2015). Early Eocene ages were also obtained for dykes  
188 crosscutting the Paleocene-related series in Disko and Nuussuaq (Storey et al., 1998) and in the  
189 Svartenhuk Peninsula (Geoffroy et al., 2001; Fig. 4).

190

### 191 **2.4. Svartenhuk structure**

192

193 The tectonics of the Svartenhuk inner-SDR has already been investigated (Geoffroy et al.,  
194 2001; Larsen and Pulvertaft, 2000). We present an original cross-section across the southern part of  
195 the Peninsula (Fig. 5). The Svartenhuk flexure (Fig. 6a) developed in the hanging wall of the  
196 Arfertuarssuk fault (Fig. 5). At a smaller scale, the subaerial V2-V3 successions are repeatedly  
197 downfaulted by normal faults, all of them facing the continent. Extensional fans are observed at all  
198 scales with lavas exhibiting an increase in dip with depth (Geoffroy et al., 2001). Internal angular  
199 unconformities are very common within the lava wedge. Southwestward, i.e. toward the ocean, the  
200 average dip of lavas tends to increase at sea level together with a progressive passive rotation to low  
201 dips of early syn-magmatic normal faults (Fig. 5; Fig. 6b). The eroded V4 unit is located at the top of  
202 the Svartenhuk wedge in apparent conformity with the V3 unit (Fig. 5, Fig. 6c-d). The V4 basalts are  
203 tilted seaward and downthrown (together with underlying V3 to V1 lava piles) against the  
204 Arfertuarssuk Fault (Fig. 5 and Fig. 6d).

205

206 The structural and stratigraphic architecture of the entire volcanic succession shows a  
207 seaward-migration of syn-magmatic extension (Geoffroy et al., 2001; Fig. 5). Another SDR appears to  
208 have developed westward of the Arfertuarssuk Fault (Fig. 5). Although available offshore seismic  
209 reflection data from this area are of poor penetration, we are confident that series of SDRs continue  
210 far offshore, like across the E-Greenland VPM (e.g. Dahl-Jensen et al., 1997; Brooks, 2011) and,  
211 elsewhere, across VPMs worldwide (e.g. Planke et al., 2000; Franke et al., 2010; Stica et al., 2014).  
212 SDRs are not only described far away from Svartenhuk in Baffin Bay (Suckro et al., 2012; Abdelmalak  
213 et al., 2018) but also along the conjugate Baffin Island VPM (Skaarup et al., 2006; Abdelmalak et al.,  
214 2018). As it represents the innermost part of the “syn-rift” formations of the SE-Baffin Bay passive  
215 margin, the Svartenhuk SDR offers a unique opportunity to assign a precise age to the SE-Baffin Bay  
216 margin continental extension.

217

### 218 **3. Radiometric data**

219

#### 220 **3.1. Sampling strategy**

221

222 Selected rocks for radiometric dating were collected from 12 basaltic flows (Fig. 7a), one  
223 trachytic intrusion (S29-E2), 4 sills and 7 dykes (Fig. 7b). From the 12 flows (Fig. 7a), one sample came  
224 from V1 (S40-E1), 5 from V2, 2 from the lower V3 and 4 from the topmost V4 sequence. Reliable  
225 dating was precluded by the massive hydrothermal alteration of samples from the upper half of V3  
226 above the Arfertuarssuk Fault. Among the 5 selected sills, one sill crosscuts the V2 unit (S22-H2), two  
227 crosscut the base of V3 (S36-E5 and S8-E13), another crosscuts the upper part of the V3 unit (S26-H1)  
228 and the trachytic intrusion (S29-E2) was emplaced between the V3 and V4 units. The sampled dykes  
229 show different trends and dips, some being parallel to the SDR flexure axis, while others are not  
230 (Geoffroy et al., 2001). Three dykes strike parallel to the NW-SE trending flexure and extensional  
231 faults (S40-E6, S14-H2, 23H4), and are thus considered to have been emplaced under the same stress  
232 field. The other dykes strike obliquely (S22H4, S27-E4) or are sub-normal (S20-E18, S2-E8) to the  
233 flexure axis (see the different directions of dykes in Fig. 7b).

234

### 235 **3.2. Dating methods**

236

237 With the exception of porphyritic trachyte S29-E2, dated on handpicked alkali feldspar  
238 phenocrysts, all datings were performed on the separated groundmass of the lavas. The techniques  
239 of groundmass separation and preparation are described in Guillou et al. (1998). All procedures  
240 concerning the sample preparation, as well as the unspiked K-Ar and  $^{40}\text{Ar}$ - $^{39}\text{Ar}$  methods, are detailed  
241 in the supplemental online material (Section methods).

242

243 New unspiked K-Ar ages are reported in Table 1a. Along with the radiometric dating, major  
244 element analyses were carried out to estimate the degree of alteration of selected samples and their  
245 groundmass. The loss-on-ignition values (L.O.I.) usually range from 0 to 2 wt.%. However, despite a  
246 thorough removal of secondary phases, the groundmass of the V2 unit picritic lavas yielded L.O.I.  
247 values spanning 2.5 to 5.1 wt.%. These values indicate that the corresponding lavas have undergone a  
248 pervasive alteration. Owing to the very low K contents and high modal abundance of olivine  
249 phenocrysts in these basalts, the ages obtained from the V2 samples were considered unreliable and  
250 excluded from our set of dating results (Fig. 7a).

251

252 Full analytical results of trachyte S29-E2 dating by the  $^{40}\text{Ar}/^{39}\text{Ar}$  method are given in Suppl.  
253 Table 1. Step-heating  $^{40}\text{Ar}/^{39}\text{Ar}$  plateau age data for groundmass samples are reported in Table 1b  
254 (with full analytical results in Suppl. Table 2). We were unable to obtain an age for sample S27-E12  
255 because of alteration, and ages could not be determined for samples S33-E9 and S40-E1 because of  
256 recoil effects during irradiation. Only the one step-heating experiment (S35-E5) yielded concordant  
257 spectra with 100% of the gas defining age plateaus (Suppl. Fig. 1). The six other plateau ages were  
258 defined for gas release rates from 60 % to 87 % (Table 1b). The  $^{40}\text{Ar}^*$  contents in samples S30-E1a,  
259 S30-E1b, S22-H4 and S21-H1 range from 68 % to 99 %, with typical values from 90 % to 99 % for the  
260 plateau steps. For samples S23-H4, S35-E5 and S42-E5, which are less potassium-rich than the others,  
261  $^{40}\text{Ar}^*$  ranges from 22 % to 90 %, with typical values of 60 % to 90 % for the plateau steps.

262

263 Plateau ages, isochron regressions and probability of fit estimates were calculated using  
264 ArArCalc (Koppers, 2002). The uncertainties on individual plateau ages as well as the inverse  
265 isochrons are given at  $2\sigma$  (Suppl. Fig. 1, Table 1b). The  $^{40}\text{Ar}/^{36}\text{Ar}$  intercept values defined for the  
266 associated isochrons are atmospheric, within the error bars. The total fusion ages are similar to the  
267 plateau or isochron ages when the margin of error is taken into account. This feature indicates that  
268 the effect of argon loss or excess argon is almost negligible for most samples. With the isochron  
269 approach, we make no assumption regarding the trapped component, and combine estimates of  
270 analytical precision and internal disturbance of the sample (scatter around the isochron). Therefore,  
271 the isochron ages (Table 1b) are preferred over the weighted mean plateau ages.

272

### 273 **3.3. Analysis of results**

274

275 The measured K-Ar ages range from  $60.38 \pm 1.34$  to  $53.85 \pm 2.84$  Ma, discarding the age of  
276 63 Ma obtained on the highly olivine-phyric sill S36-E5 which crosscuts the V3 unit (Fig. 7b). The  
277  $^{40}\text{Ar}/^{39}\text{Ar}$  isochron ages and the unspiked K-Ar ages of samples S21-H1 and S35-E5 are equivalent at  
278 the 95 % confidence level (Fig. 7c). This comparison is useful to assess the reliability of the ages  
279 obtained via the unspiked K-Ar method. However, a comparison between Table 1a and 1b reveals  
280 some inconsistencies in the K-Ar data with respect to stratigraphic constraints as well as differences  
281 between K-Ar and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages (Fig. 7c and samples with asterisk \* Table 1a).

282

283 For instance, the  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of  $61.08 \pm 0.56$  Ma and  $59.78 \pm 0.41$  obtained for the base of  
284 V2 (S42-E5) and for the base of V3 (S35-E5), respectively (Table 1), indicate that the youngest K-Ar  
285 ages of the stratigraphically lowermost samples S40-E1 (V1), as well as the V3-related sample S33-E9,  
286 are too young (Fig. 7c). This underestimation is most likely due to argon loss through glass hydration  
287 and alteration. Sample S36-E5 comes from a 15-m-thick highly porphyritic sill. Incomplete degassing  
288 of the magma prior to crystallisation might be a source of excess  $^{40}\text{Ar}^*$ , thus explaining the apparent  
289 old K-Ar age of  $\sim 63$  Ma. For the topmost V4 succession, the  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of S30-E1 ( $54.39 \pm 0.6$  Ma)  
290 and S21-H1 ( $55.20 \pm 0.79$  Ma) are preferred to the older unspiked K-Ar ages of samples S30-E1, S30-  
291 E3 and S30-E4 (Figs. 7a-c).

292

293 The overestimation of the K-Ar ages is probably related to excess  $^{40}\text{Ar}^*$ , as shown from the  
294 step-heating experiments of S30-E1. We note that the 4 and 3 first increments from experiment A  
295 and B, respectively (Suppl. Fig. 1), when excluded from the plateau calculations, yield apparent ages  
296 significantly older than the plateau ages. This feature may be due to a component of extraneous  
297 argon in some aliquots.

298

299 The unspiked K-Ar and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages obtained for the Svartenhuk intrusive system are  
300 clustered into two main groups, Paleocene and Eocene. The Eocene K-Ar ages yielded by dykes and  
301 sills are consistent, at  $2\sigma$ , with the  $^{40}\text{Ar}/^{39}\text{Ar}$  isochron age of dyke S22-H4 ( $\sim 54.4 \pm 1$  Ma, Table 1). The  
302 Paleocene K-Ar age yielded by dyke S40-E6 is consistent with the  $^{40}\text{Ar}/^{39}\text{Ar}$  isochron ages of dyke 23-  
303 H4 ( $\sim 59 \pm 1$  Ma). Most of the intrusions are Early Eocene, with ages spanning a wider range than for  
304 unit V4, from 57 to 53 Ma (Table 1). The dykes are either parallel (e.g. S14-H2) or oblique (e.g., S20-  
305 E18) to the flexure and clearly fed the eroded top of the SDR. Two dykes only (S40-E6 and S23-H4),  
306 striking parallel to the regional flexure, were emplaced synchronously with the Paleocene volcanic  
307 series.

308

## 309 **4. Paleomagnetic data**

310

### 311 **4.1. Sampling**

312

313 Samples for paleomagnetic studies were drilled in the V2 and V3 units, alongshore the  
314 southern coast of Svartenhuk. The V1 sequence was previously sampled to the East of Svartenhuk by  
315 Riisager et al. (2003) along a 200 m high continuous vertical section (profiles 3 & 4 on Fig. 4a). A total  
316 of 244 cores were drilled from 30 lava flows and 11 intrusions (Figs. 7a-b). Among the sampled flows,  
317 19 were sampled within the V2 unit and the remaining 11 belong to the lower and upper parts of the  
318 V3 unit. All cores were drilled in the central part of the lava flows and dykes to minimize weathering  
319 effects and to avoid the thermal influence of overlying flows or nearby intrusions. The cloudy and  
320 foggy weather precluded the systematic use of a sun compass for orientation of the drilling cores. All  
321 measurements were corrected using the 2012 IGRF-based regional declination of  $-37^\circ$ .

322

#### 323 **4.2. Methods**

324

325 A total of 232 samples were stepwise thermally demagnetized. A “pilot demagnetization”  
326 was performed for one or two samples per flow/dyke with 15 successive steps to determine the  
327 thermal spectrum of each flow. These results show that the magnetic mineralogy is very stable over a  
328 large range of temperatures, while the Curie temperature is close to  $580^\circ\text{C}$  indicating that the main  
329 magnetic carrier phase is low-Ti magnetite (Fig. 8a-b).

330

331 Demagnetization was performed on 5 to 10 samples per flow/dyke (6 samples on average)  
332 using 12 to 14 steps, from room temperature up to a maximum of  $600^\circ\text{C}$  (Suppl. Tables 3-4). Thermal  
333 demagnetization was conducted in air using a PYROX furnace with the heating space controlled by  
334 three separate heating coils with independently controlled thermocouples maintaining the thermal  
335 gradient under  $2^\circ\text{C}$  at  $500^\circ\text{C}$  over the entire heating zone. Cooling was obtained with air circulation  
336 over the samples in a magnetic field of less than 1 nT.

337

338 After each step, the samples were measured using a 755R-2G cryogenic magnetometer  
339 equipped with high homogeneity pick-up coils. Both the furnace and the magnetometer were housed  
340 within the mu-metal shielded room of LSCE Gif-sur-Yvette. Possible mineralogical changes during  
341 heating were monitored by measuring the low field susceptibility value after each thermal  
342 demagnetization step. The thermal demagnetization was stopped when less than 10 % of the initial  
343 Natural Remanent Magnetization (NRM) was left or when magnetic mineral changes occurred due to  
344 destabilization of the magnetization.

345



346 The final direction of the Characteristic Remanent Magnetization (ChRM) was determined  
347 using Principal Component Analysis (PCA) (Kirschvink, 1980), that allowed us to define the mean  
348 angular deviation (MAD, Suppl. Tables 3-4). In the great majority of cases, apart from a viscous  
349 component removed after the very first demagnetization step (100-140 °C), demagnetization  
350 diagrams indicated univectorial behaviors (Figs. 8c-8e). However, five samples from dyke S6A, six  
351 samples from dyke 27B and one sample from flow S6C displayed more complex demagnetization  
352 patterns and could not be interpreted (Fig. 8f). Overall, reliable results were obtained from 232  
353 samples out of 244.

354  
355 MAD values were lower than 5° for 226 samples, and ranged between 5 and 10° for the 6  
356 remaining samples (Suppl. Tables 3-4). Mean-site directions calculated using Fisher's statistics were  
357 based on 3 to 7 independent ChRM directions (Suppl. Tables 5-6). Some of the ChRM directions could  
358 be excluded from the mean direction of individual flows as these data points were clearly outliers  
359 (Suppl. Tables 3). The mean-flow directions were defined with a precision angle  $\alpha_{95}$  ranging from 2° to  
360 10.7° with a mean value of 5° (Suppl. Tables 5; Suppl. Fig. 2). The results obtained from the dykes  
361 were more complicated. Two of the dykes (S27B and S6A) yielded a poorly defined mean direction  
362 with a precision angle  $\alpha_{95}$  of 18° and 40°, respectively (Suppl. Table 4, Suppl. Fig. 3). The directions  
363 obtained from the other dykes were characterized by  $\alpha_{95}$  values varying between 3.4° and 11° (Suppl.  
364 Table 4).

365

### 366 **4.3. Data analysis**

367

368 The new paleomagnetic dataset indicates that all the lava flows and intrusions from the  
369 Svartenhuk volcanic wedge were emplaced during a single or several distinct periods of reversed  
370 polarity of the magnetic field (Suppl. Table 5). Solely dyke S7A yielded a normal polarity ChRM (Suppl.  
371 Fig. 3; Suppl. Table 6). Our results are consistent with previous aeromagnetic results in this area  
372 (Rasmussen, 2002). After tilt correction, the resulting mean paleomagnetic direction of lava flows is  
373 Dec = 151.4°, Inc = -64.9° (n/N=30/30,  $\alpha_{95}$  = 6.2°; Fig. 9a). The ChRM directions of the dykes are more  
374 scattered than those obtained from the flows (Fig. 9b). Restoring dykes to the vertical gives ChRM  
375 directions similar to those from the basaltic flows (Fig. 9b).

376

377 Our paleomagnetic mean directions for the untilted V2 and V3 lavas are very close to those  
378 obtained by Riisager et al. (2003) from the sub-horizontal V1 unit (Fig. 9). Incidentally, this indicates  
379 that the dip of lava flows within the SDR wedge is of tectonic origin and not due to paleo-slopes.

380

381 Plotting declinations and inclinations from the bottom to the top of the SDR reveals  
382 homogeneity (Fig. 9c). The ChRM directions show deviations less than 32° regarding the mean vector.  
383 The deviations in our data set could reflect short-time excursions of the geomagnetic field as  
384 otherwise suggested by Riisager et al. (2003).

385

## 386 5. Discussion

387

388 Our set of K-Ar and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages indicates that the Svartenhuk SDR lavas were emplaced  
389 between 61.5 and 54 Ma (Fig. 7c). According to Ogg's (2012) time scale, these ages suggest that the  
390 reversed magnetization inferred from our paleomagnetic data correspond to magnetochrons C26r  
391 and C24r (Fig. 10). These results are consistent with former data from the WGVP. The normal polarity  
392 interval recognized by Riisager and Abrahamsen (1999) in the basal Vaigat Formation at Vaigat Fjord  
393 (Fig. 4) would indicate that the Paleogene volcanic activity in the exposed WGVP spanned C27n to  
394 C24r (Storey et al., 1998; Riisager and Abrahamsen, 1999; Riisager et al., 2003; Skaarup and  
395 Pulvertaft, 2007; Larsen et al., 2015; Fig. 10). In Svartenhuk, the youngest (Eocene) ages are related to  
396 the V4 unit. Due to pervasive hydrothermal alteration we did not obtain reliable ages from the top of  
397 the Late Paleocene V3 unit. However, our datings bracket the eruption of the Vaigat picrites V2  
398 between 61.5 and 59.5 Ma, i.e. during the Middle Paleocene (Selandian).

399

400 With the exception of early dykes S40E6 and S23H4, which were injected during the  
401 Paleocene, most of the sampled intrusions in Svartenhuk yielded early Eocene ages consistent with  
402 former observations in this area (Geoffroy et al., 2001) and farther south in Nuussuaq and Disko  
403 (Storey et al., 1998; Larsen et al., 2015, Fig. 10). Paleomagnetic results allow us to refine the age of  
404 those intrusions as being restricted to C24r (57-54 Ma; Ogg, 2012; Fig. 10).

405

406 The magmatic development of the inner SDR until the Eocene is of particular importance.  
407 Although a 1-2 m.y. pause in magmatic activity may have occurred during the development of the  
408 SDR wedge, the Svartenhuk SDR wedge developed throughout the Paleocene, until the Early Eocene.  
409 The V4 alkaline basalts appear to be coeval with the thick lava piles of enriched basalts recognized in  
410 western Nuussuaq and Hareøen Island (Kanísut Member) and in western Ubekendt Ejland (Nûk  
411 takisôq Member and Erqua Fm., Fig. 10). We investigated the structure of the oblique margin  
412 segment of Nuussuaq during several field trips (Geoffroy et al., 2001; Abdelmalak et al., 2012). The  
413 original sections from Fig. 11 clearly outline that the Eocene Kanisut Member (Larsen et al., 2015) is  
414 also part of the syn-tectonic magmatic wedge in Nuussuaq (Guan, 2018).

415

416 Normal-polarity basalts of probable C24n ages do exist offshore to the West of Nuussuaq  
417 and Ubekendt Ejland (Rasmussen, 2002; Fig. 3b). In addition, Eocene sub-aerial basalts, spanning  
418 C24r to C22r (~56.5 to ~50 Ma), have been drilled off W-Greenland (Delta-1 well, >110 km WSW of  
419 Svartenhuk Halvø, Fig. 1c; Nelson et al., 2015). These basalts present petrologic and geochemical  
420 affinities with the Eocene Kanisut and Svartenhuk V4 formations and they overlie the margin crustal  
421 necking zone (Hosseinpour et al., 2013; Suckro et al., 2012). In addition, SDRs have recently been  
422 imaged in the Central-East Baffin Bay 100 km seaward and 200 km to the NW of the Svartenhuk SDR  
423 (Abdelmalak et al., 2018; Fig. 1c). These SDRs are located close to the Delta 1 Eocene basalts (Fig. 1c),  
424 within the oceanic crust domain inferred to be Paleocene by Oakey and Chalmers (2012). The  
425 occurrence of SDRs of probable Eocene age in the central part of the Baffin Bay might lead to envision  
426 a larger volume and possibly an extended duration for the syn-rift magmatism.

427  
428 As indicated by dyke orientations (e.g. Fig. 7b) and by stress-inversion of a large amount of  
429 reliable fault-slip data, the C24r period was associated with instability in the stress field at the scale of  
430 the WGVP, with a shift in the minimum stress  $\sigma_3$  from N060E to N010E (Geoffroy et al., 2001;  
431 Abdelmalak et al., 2012). Our additional datings of dykes globally support this view, with the ~EW  
432 dykes being younger (Fig. 12). This stress field was coeval and consistent with the C24 kinematic  
433 change in the Labrador-Baffin system (Geoffroy et al., 2001). However, the NS crustal extension had a  
434 minor influence on the ongoing tectonic development of the SDR in the NE-SW trend. The related  
435 deformation is better expressed along NW-SE trending fjords, suggesting a reactivation of older  
436 structures of probable Cretaceous age. During our field and airborne surveys, we also observed some  
437 ~NW-SE dykes (associated with the N060E extensional regime) cross-cutting ~E-W dykes (associated  
438 with the N010E extensional regime). The syn-magmatic extensional stress regime with  $\sigma_3$  trending  
439 N010E, which appeared later, was probably a transient phenomenon.

440  
441 Since it is widely admitted that significant continental extension ceases at the onset of  
442 seafloor spreading, our results strongly question the previously proposed Paleocene age for the  
443 earliest oceanic accretion in the southern Baffin Bay (e.g. Oakey and Chalmers, 2012; Hosseinpour et  
444 al., 2013). The development of an inner SDR along the innermost part of the West Greenland volcanic  
445 margin during C24r is inconsistent with the occurrence of continental breakup in Baffin Bay before  
446 C24n. More recent SDRs of possible C22 age (see above) are observed offshore in the Baffin Bay  
447 suggesting that continental breakup could be much younger than C24n. By comparing the SDR  
448 observations of Block et al. (2012), Suckro et al. (2012) and Abdelmalak et al. (2018), located ~180  
449 km off Svartenhuk, and nearby Delta-1 observations of subaerial basalts (Fig. 1c; Nelson et al., 2015),  
450 we infer that the crust in this area was buoyant and thicker than any normal oceanic crust, and  
451 therefore of possible continental affinity.

452

453 The present-day knowledge of the crustal architecture on the southern Baffin Bay passive  
454 margins is exclusively based on two refraction profiles striking at a high angle to the crustal thinning  
455 gradient (profiles 4 and 5 in Fig. 1c). Unfortunately, no seismic refraction or wide-angle seismic  
456 reflection lines were shot along the dip of the exposed W-Greenland VPM (Geoffroy et al., 2001). In  
457 Baffin Bay, the potential oceanic crust domain, i.e. the area without recognized SDRs (inner or outer)  
458 and with a constant oceanic-type crustal thickness in the range of  $7.1 \pm 0.8$  km (White et al, 1992), is  
459 very limited. Figure 13 displays the results of two distinct and recent inversions for crustal thickness in  
460 Baffin Bay (Hosseinpour et al., 2013 and Welford et al., 2018). The oceanic-type crust (thickness lower  
461 than 8 km) is restricted to two small domains lying at the NW and SE tips of the main Baffin Bay  
462 transform, respectively. These domains are considerably more restricted in size than those formerly  
463 suggested (Oakey and Chalmers, 2012) and reported in Fig. 13a. We infer that the age of the oceanic  
464 crust in Baffin Bay, where it exists, ranges from C24n to C13 or, much more probably, from C22 to C13.  
465

466 We consequently suggest that the oceanic-type floor in Baffin Bay is considerably narrower  
467 than previously proposed, with an accretion period limited to a max.  $\sim 20$  m.y. time span. According to  
468 this interpretation, the southern Baffin Bay opened ca. 8 Ma after the proposed earliest oceanic  
469 accretion in the northern Labrador Sea and also after the N060E to NS kinematic reorganisation  
470 between North America and Greenland.

471  
472 Such a diachronism in oceanic rift propagation from the N-Labrador Sea to Baffin Bay could  
473 be the result of the thermal and mechanical barrier effect caused by the Ungava transform zone (Fig.  
474 1). Koopman et al. (2014) developed a numerical model suggesting that rift-parallel mantle flow is  
475 delayed at transform faults in segmented breakup systems. Transforms would act as rift-propagation  
476 barriers and would enhance considerably the amount of syn-breakup magmatism at the tip of the  
477 propagating rift. Funk et al. (2007) demonstrated the occurrence of high velocity igneous lower crust  
478 4 to 8 km-thick (7.4 km/s) beneath the Ungava fault zone itself. These authors considered that the  
479 Ungava leaky transform zone would have acted as a barrier guiding the migration of hot material  
480 towards both the adjacent northern Labrador Sea and southern Baffin Bay VPMs.  
481

482 Our data confirm that the Ungava fault zone acted as a mechanical barrier to the Labrador  
483 axis propagation for at least 8 m.y. (see above). However, the passive margins south of the Ungava  
484 fault zone appear to be less magma-rich (e.g. Peace et al., 2016) than those located to the north.  
485 These observations contradict part of the outcomes of Koopmann et al. (2014). Instead, our results  
486 suggest that mantle melting occurred on both sides of the Ungava fault zone and was possibly coeval  
487 with oceanic accretion to the south (N-Labrador Sea) and continental extension to the north (Baffin  
488 Bay) of this fault zone. The origin of this magmatism is beyond the scope of this paper. However,  
489 higher mantle temperature to the north of the Ungava fault zone (and possibly beneath it) and/or  
490 higher mantle fertility in the area may have enhanced mantle melting. In addition, we note that there  
491 is a very sharp transition from volcanic to non-volcanic passive margins to the north of the  
492 conspicuous Upernavik escarpment (UE in Fig. 1c; Whittaker et al. 1997). This escarpment bounds  
493 southwards the preserved Rae Archean lithosphere from a basement reworked during the  
494 Paleoproterozoic (Henriksen et al., 2009). This major NE-SW structure within the Baffin Bay could  
495 have played a significant role in the localization of asthenospheric melting at depth.

496

## 497 **Conclusions**

498

499 Volcanic activity in Svartenhuk was coeval with the tectonic development of the innermost  
500 SDR of the SE-Baffin Bay VPM. Our paleomagnetic and unspiked K-Ar and Ar-Ar dating results show  
501 that post-Mesozoic continental extension spanned the Paleocene until the Eocene (C24r). The inner  
502 position of the Svartenhuk SDR suggests that it recorded the earliest extension stage with regards to  
503 the seaward development of the VPM through time. Therefore, given the additional data from the  
504 Delta-1 well and available seismic data (Abdelmalak et al., 2018), we infer that oceanic accretion in  
505 Baffin Bay started certainly after C24r and probably as late as C22. All oceanic spreading operated  
506 with a ~NS trend. Admitting that seafloor spreading stopped at C13 (Roest and Strivastava, 1989;  
507 Oakey and Chalmers, 2012), the Baffin Bay oceanic stage would have lasted 20 m.y., from 50 Ma to 30  
508 Ma. The Ungava fault zone acted as a barrier for oceanic rift propagation but not as a boundary for  
509 mantle melting. We suggest that the Upernavik discontinuity (and its prolongation to the East of  
510 Baffin Island) could have influenced the dynamics of mantle melting and magma output in the crust.

511 Targeted marine surveys on both sides of Greenland are required to better define the  
512 continent-ocean transitions in the NE-Atlantic and Baffin Bay basins. Such constraints are essential for  
513 a better understanding of continent-ocean boundaries at magma-rich passive margins and, at a wider  
514 scale, to improve our knowledge of the geodynamics of continental breakup between the NAM and  
515 Eurasia plates.

516

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518

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533

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718 **Figure captions**

719

720 **Figure 1.** Regional bathymetric, geological and tectonic framework. (a) The North Atlantic Igneous  
721 Province (modified from Larsen and Saunders, 1998). (b) Bathymetry of the Labrador Sea, Davis Strait  
722 and Baffin Bay (GEBCO\_08 Grid, <http://www.gebco.net>). The Greenland Plate motion path relative to  
723 the North American Plate is shown as black lines and squares for 3 locations of the Greenland coast  
724 at magnetochrons C27 to C13 (from Oakey and Chalmers, 2012). The northward reorganization  
725 started at C25 (late Paleocene). CD: Cape Dyer, D: Disko, N: Nuussuaq, S: Svartenhuk. (c) Geological  
726 map of the Labrador-Baffin margins and oceanic domains modified after Chalmers and Oakey (2007)  
727 and Oakey and Chalmers (2012). SDR mapping from Keen et al. (2012), Skaarup et al. (2006), Funck et  
728 al. (2012), Suckro et al. (2012) and Abdelmalak et al. (2018). UFZ and HFZ: Ungava and Hudson fault  
729 zones respectively; DSH: Davis Straight High. Seismic refraction lines from: (1) Funck et al. (2007); (2)  
730 Gerlings et al. (2009); (3) Suckro et al. (2013); (4) Funck et al. (2012); (5) Suckro et al. (2012). Orange  
731 circles: well locations where Paleogene volcanics were drilled: (a) Delta-1, (b) Hellefisk-1, (c) Nukik-2,  
732 (d) Gjoa G-37, (e) Raleigh N-18, (f) Hekja O-71. UE : Upernavik Escarpement.

733

734 **Figure 2.** Stratigraphic columns and major tectonic events from the Labrador (Hopedale and Saglek)  
735 basins, southern West Greenland offshore basins and the Nuussuaq Basin onshore central West  
736 Greenland. Stratigraphic columns are modified from Chalmers (2012). “*DSH transpression*” applies to  
737 the folding phase recorded along the Davis Strait High (Fig. 1c) after the N-S kinematic reorganization  
738 of the Labrador seafloor spreading. In Nuussuaq basin, volcanic formations are V: Vaigat, M: Maligat,  
739 S: Svartenhuk, N: Naqerloq, E: Erqa.

740

741 **Figure 3.** Free air gravity (a) and magnetic (b) anomalies over Baffin Bay and Davis Strait. Free air  
742 anomaly is from Sandwell et al. (2014) and the magnetic anomaly at sea-level grid is from Meyer et  
743 al. (2017) EMAG2-V3. The linear gravity lows are interpreted as extinct spreading centers in the Baffin  
744 Bay (Whittaker et al., 1997). Their location and the extent of the Eocene and Paleocene oceanic crust  
745 domains according to Oakey and Chalmers (2012) are also reported on the magnetic grid.

746

747 **Figure 4.** Simplified geological map (a) of West Greenland (adapted from Abdelmalak et al., 2012)  
748 showing the locations of available palaeomagnetic data (white circles) and dated rock samples  
749 (colored dots). Riisager et al. (2003) palaeomagnetic profiles 1-4 are from the Vaigat Fm., profile 5  
750 from the Maligat Fm., profile 6 from the Kanisut Mb. (b) Stratigraphy and ages of the Tertiary volcanic  
751 succession according to Larsen et al. (2015).

752

753 **Figure 5.** Geological map and structure of Svartenhuk Peninsula, modified from Larsen (1983) and  
754 Larsen and Grocott (1992). The cross-sections illustrate the structure of the Tertiary volcanic  
755 successions along profiles AA', BB' and CC'. CBF: Cretaceous basin Boundary Fault.

756

757 **Figure 6.** Selected views of the Svartenhuk inner-SDR structure. The inset map (same legend as for  
758 Fig. 3) shows the location and orientation of each panorama. A.F.: Arfertuarssuk Fault. (a) Helicopter  
759 view of the crustal flexure which developed in the hanging wall of the A.F. (b) Example of a tilted  
760 continentward-dipping normal fault and associated splay faults within the V2 unit (southern coast of  
761 Svartenhuk Halvø). (c) Eroded V4 lavas dipping towards the Arfertuarssuk Fault. In the background,  
762 V3 unit lava outcrop along the fault footwall. (d) 3D view (Spot 6 images draped on the Aster DEM) of  
763 the A.F. hanging wall with V3 and V4 units tilted SW-ward (i.e. oceanward).

764

765 **Figure 7.** Location maps of samples used for radiometric dating and palaeomagnetic studies (a.  
766 basaltic flows; b. planar intrusions). Ages reported in whites boxes are  $^{40}\text{Ar}/^{39}\text{Ar}$  ages ( $\pm 2\sigma$ ). Zones I to  
767 VI refer to sampling areas. The dykes are mapped from Spot 6 satellite images. (c) Stratigraphy of the  
768 Svartenhuk lavas based on  $^{40}\text{Ar}/^{39}\text{Ar}$  and K-Ar ages.

769

770 **Figure 8.** Paleomagnetic results. (a) and (b) Pilot demagnetization (15 steps) of samples from flows  
771 and dykes, respectively. (c) to (f) Representative demagnetization diagrams obtained using stepwise  
772 thermal demagnetization. Black (white) circles represent projections onto the horizontal (vertical)  
773 plane. Dashed lines indicate the best fit line obtained using principal component analysis. The  
774 declination, inclination, and maximum angular deviation (MAD) values are given for each of these fits.  
775 Gray symbols on the diagram reported in f illustrate the viscous component removed after the very  
776 first demagnetization steps.

777

778 **Figure 9.** Stereoplots of the characteristic remanent magnetization (ChRM) mean directions obtained  
779 from stepwise demagnetizations of (a) flows and (b) dykes. The mean directions calculated using the  
780 Fisher statistics are reported as blue crosses and the red circles show the ellipse of confidence ( $\alpha_{95}$ ).  
781 The average directions obtained by Riisager et al. (2003) in the Svartenhuk Halvo V1 unit and in the  
782 Kanisut Mb. of Nuussuaq are plotted for comparison. (c) Variations in the flow-mean paleomagnetic  
783 directions, bottom to top, in Svartenhuk lavas. The average declinations and inclinations calculated  
784 from all data correspond to the gray shaded lines. The y axis refers to the relative position of samples  
785 along the southern coast of Svartenhuk (see Fig. 5a for location).

786

787 **Figure 10.** Synthesis of ages obtained from flows and intrusions of the Western Greenland Volcanic  
788 Province, from the southern Davis Strait to central West Greenland. Time scale is from Ogg (2012).  
789 Ages are plotted with confidence interval of  $\pm 1\sigma$ . In each area, data are displayed from the left to  
790 right as a function of stratigraphic order. VF - Vaigat Formation, MF - Maligât Formation. SF –  
791 Svartenhuk Formation, Na - Naujánguit Mb., Or - Ordlingassoq Mb., RD - Rinks Dal Mb., No -  
792 Nordfjord Mb. & Niaqussat Mb. (Nunavik unit), Ka - Kanísut Mb., SF LM and SF UM – Lower and  
793 Upper Mb. of the Svartenhuk Formation; (1) Storey et al. (1998) ages modified by Skaarup and  
794 Pulvertaft (2007); (2) Larsen et al. (2015); (3) Geoffroy et al. (2001) mean age calculated from 5  
795 different dykes; (4) Williamson et al. (2001); (5) Nelson et al. (2015).

796

797 **Figure 11.** Structure of the Tertiary volcanics to the west of the Itilli fault in the Nuussuaq peninsula.  
798 Insert: location of the profiles. See also Fig. 4.

799

800 **Figure 12.** Age versus structural attitude of dykes sampled in the Svartenhuk Halvø. Uncertainties for  
801 K-Ar and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages are given at the  $2\sigma$  level. Strikes and dips of dykes are shown between  
802 brackets.

803

804 **Figure 13.** Baffin Bay crustal models. (a) Oakey and Chalmers (2012)'s crustal model of Fig. 1c  
805 compared with crustal thickness within Baffin Bay derived from gravity inversion models; (b)  
806 Hosseinpour et al. (2013); (c) Welford et al. (2018).

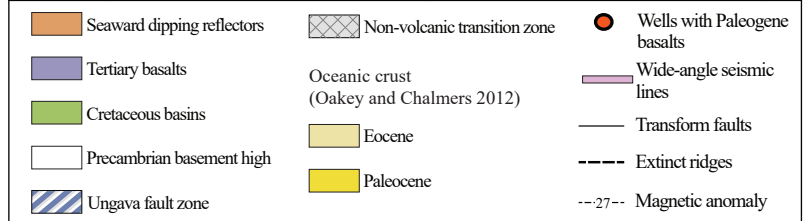
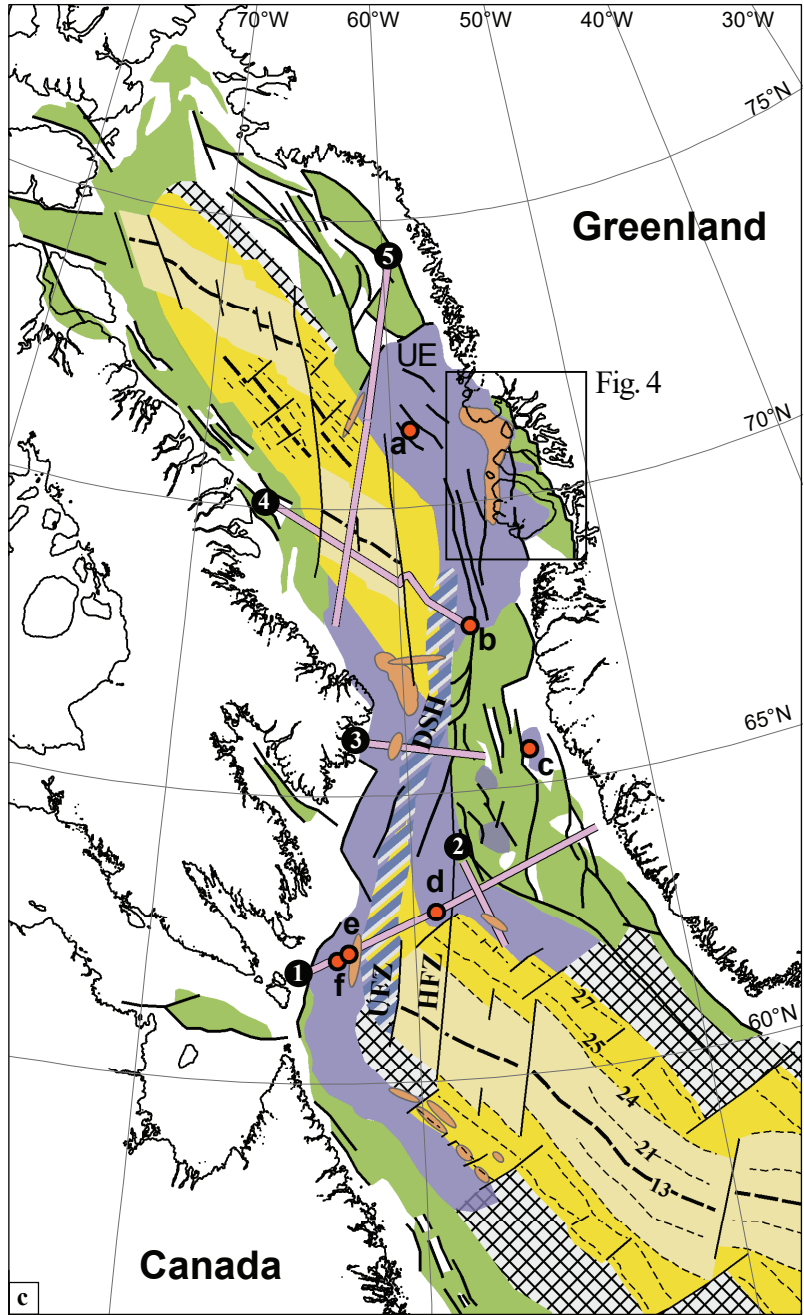
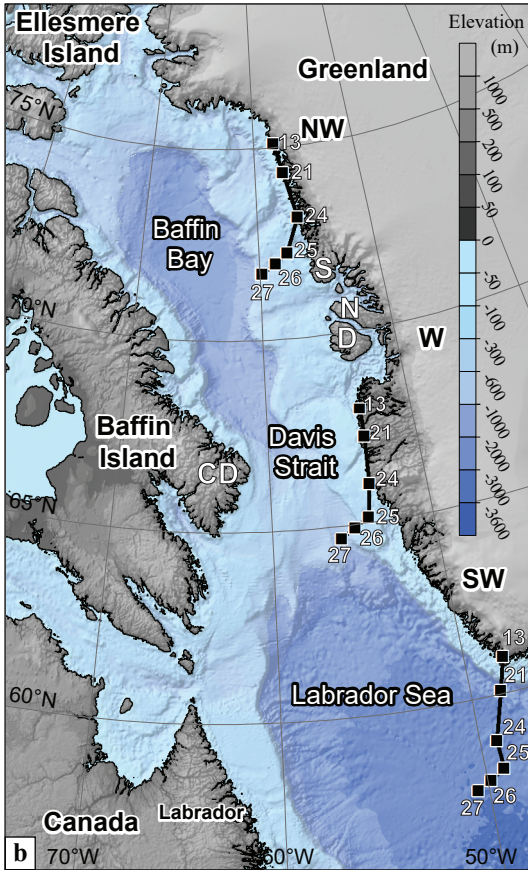
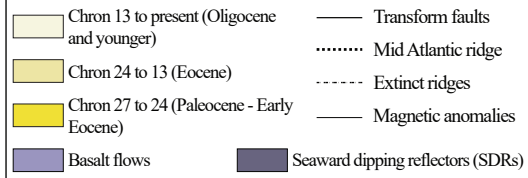
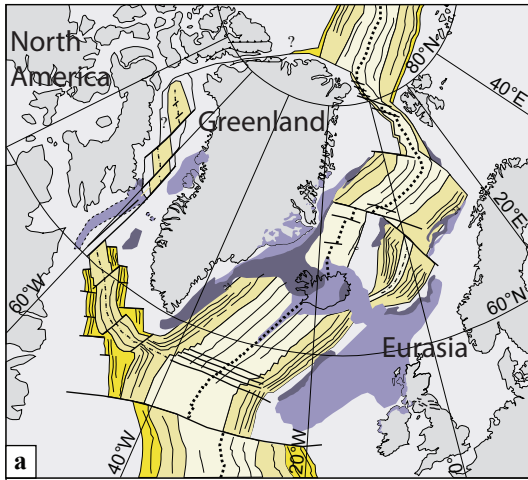
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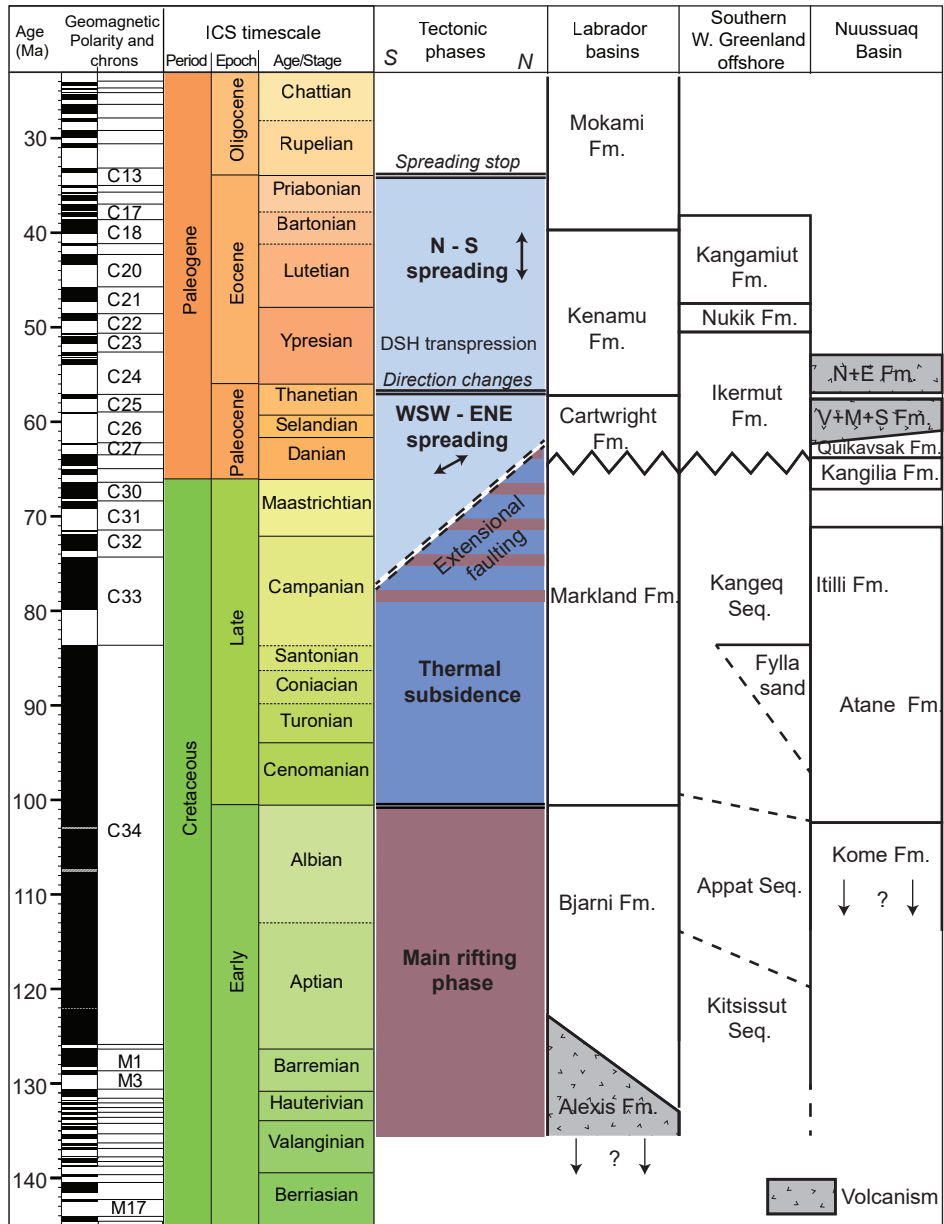
808 **Table caption**

809

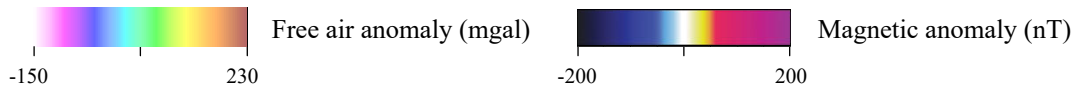
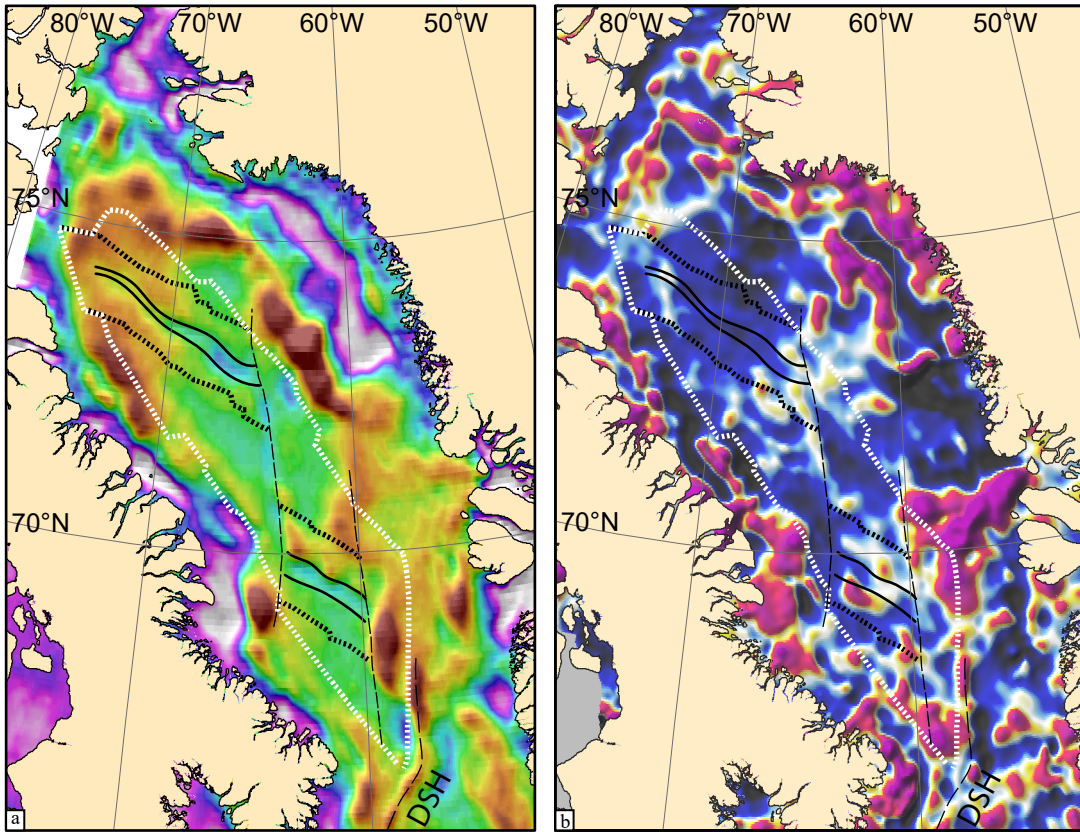
810 **Table 1.** (a) Details of replicate analyses and K/Ar age determinations (shown in stratigraphic order).  
811 Values reported with an asterisk have not been included in the data analysis. Strike/dip of dykes are  
812 indicated in brackets. (b) Summary of  $^{40}\text{Ar}/^{39}\text{Ar}$  data from incremental heating experiments. Ages are  
813 calculated relative to the 28.201 Ma Fish Canyon sanidine standard.



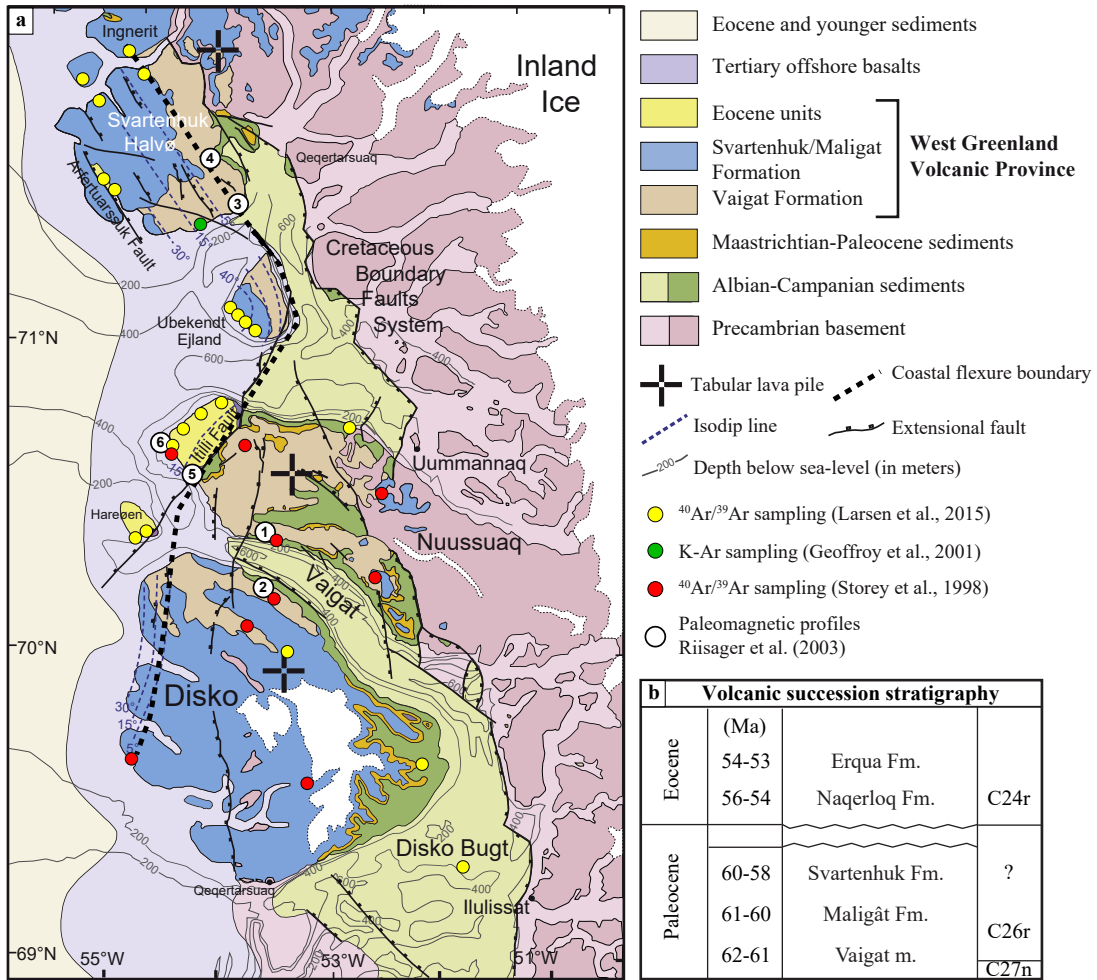




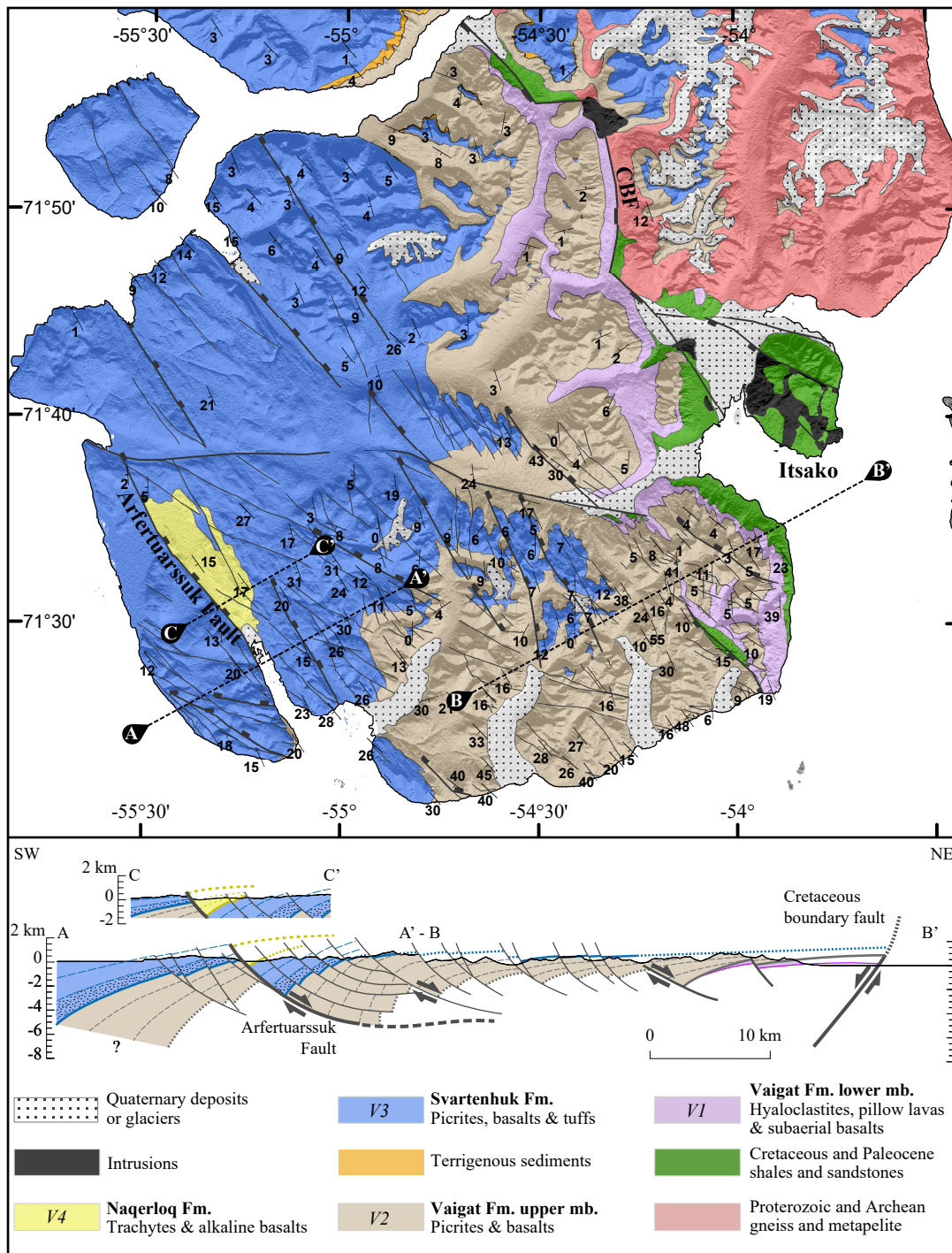




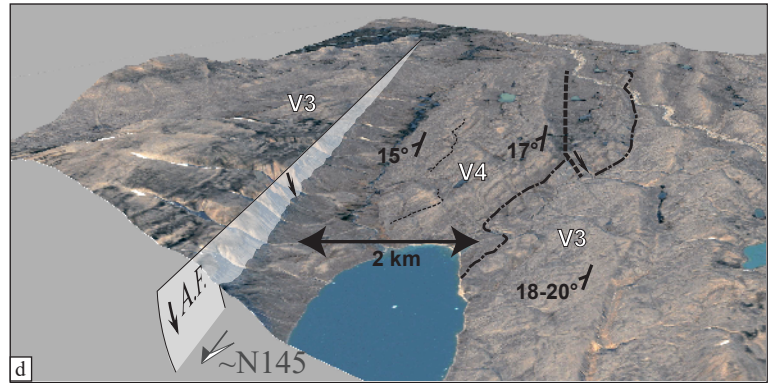
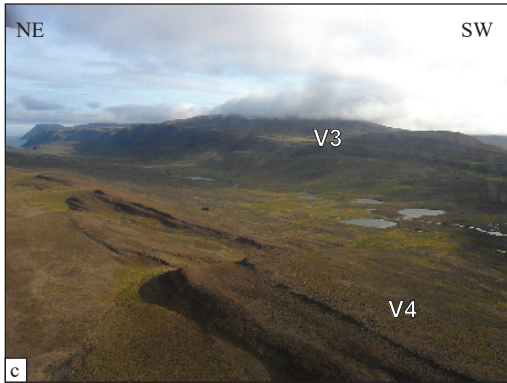
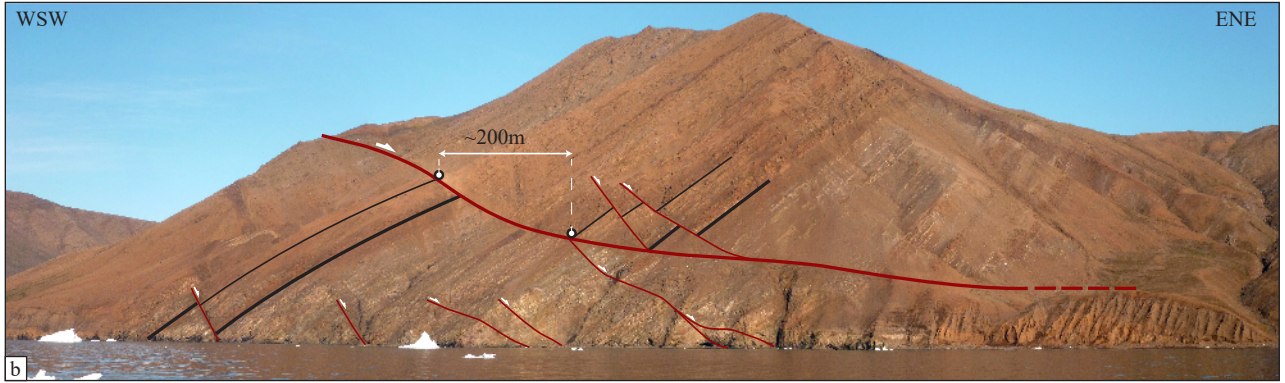
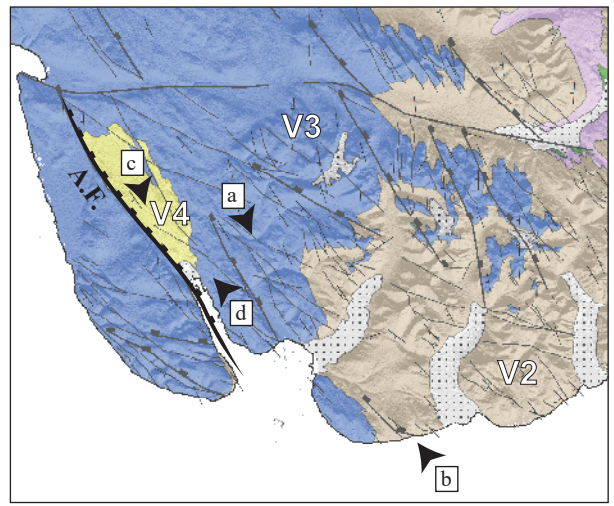
Oakey and Chalmers (2012)	Chalmers (2012)
..... Paleocene oceanic crust (chrons 27n to 24n)	==== linear gravity low
..... Eocene oceanic crust (chrons 24n to 13n)	----- Transform fault

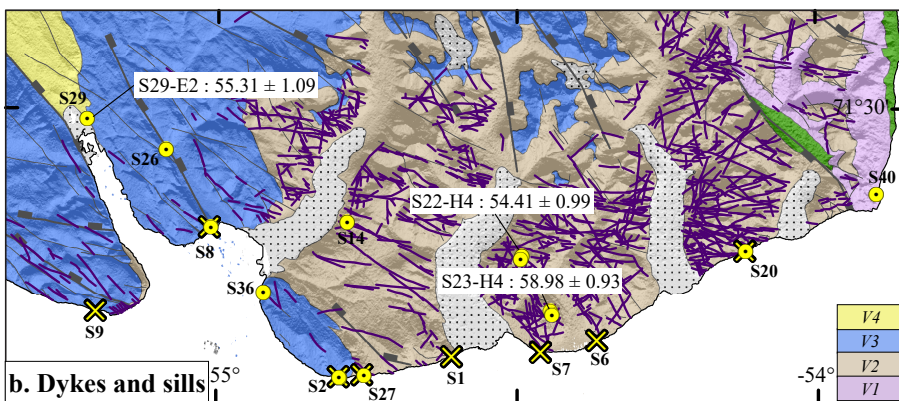
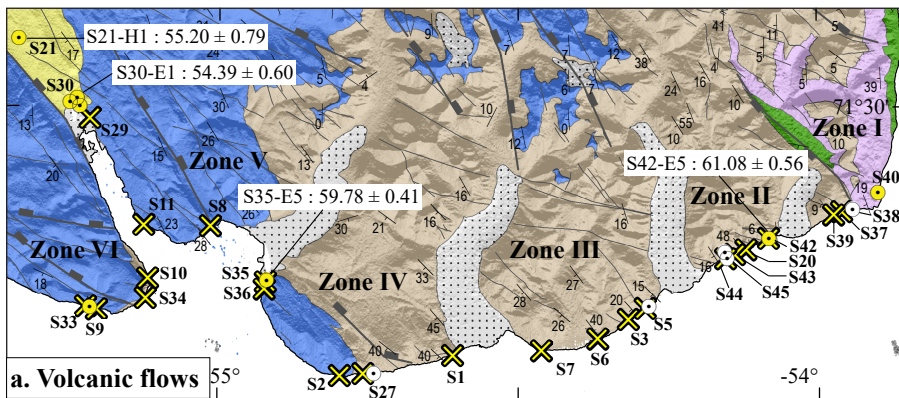


- Eocene and younger sediments
  - Tertiary offshore basalts
  - Eocene units
  - Svartenhuk/Maligat Formation
  - Vaigat Formation
  - Maastrichtian-Paleocene sediments
  - Albian-Campanian sediments
  - Precambrian basement
- Tabular lava pile      Coastal flexure boundary  
 Isodip line      Extensional fault  
 Depth below sea-level (in meters)
- $^{40}\text{Ar}/^{39}\text{Ar}$  sampling (Larsen et al., 2015)
  - K-Ar sampling (Geoffroy et al., 2001)
  - $^{40}\text{Ar}/^{39}\text{Ar}$  sampling (Storey et al., 1998)
  - Paleomagnetic profiles Riisager et al. (2003)









**c. Svartenhuk volcano-stratigraphy**

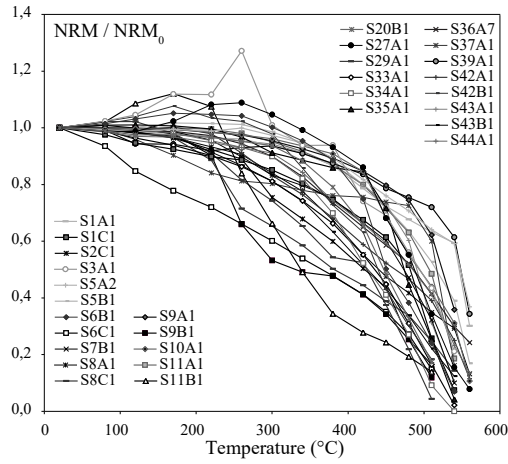
	Samples	<sup>40</sup> Ar/ <sup>39</sup> Ar ages	K-Ar ages
Naqerolq Fm.	S30-E4 & S30-E3		58.73 ± 1.53 57.96 ± 1.20
	S30-E1	54.39 ± 0.60	59.07 ± 1.16
	S21-H1	55.20 ± 0.79	55.86 ± 1.25
?			
Svartenhuk Fm.	S33-E9 & S35-E5	59.78 ± 0.41	57.78 ± 1.88 59.15 ± 0.85
Vaigat Formation	S42-E5	61.08 ± 0.56	
	S40-E1		59.11 ± 0.83

Paleomagnetic samples  
 Dated samples

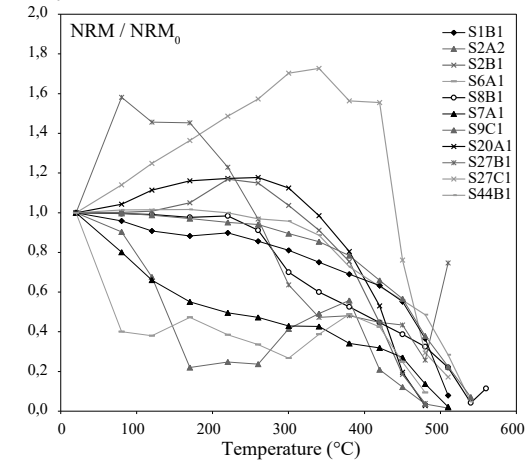
Dyke  
 Unreliable dating

Lavas  
 Sediments  
 Lavas and hyaloclastites

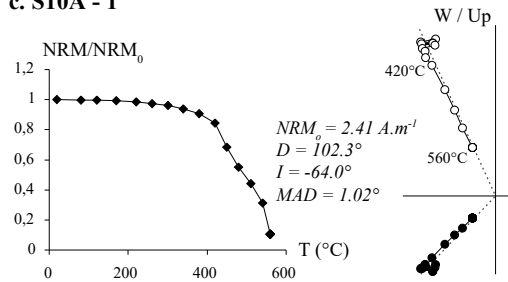
**a. Flows**



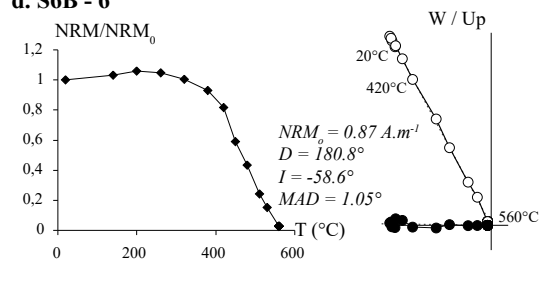
**b. Dykes**



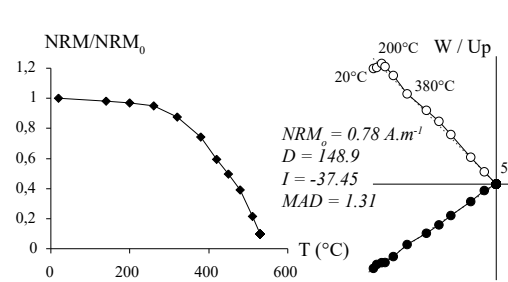
**c. S10A - 1**



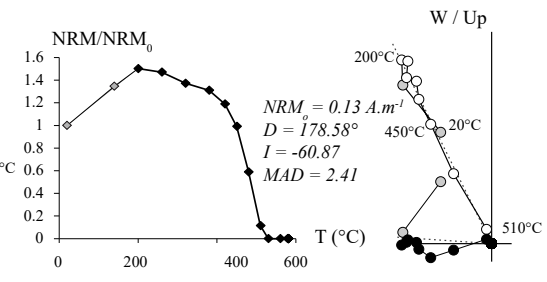
**d. S6B - 6**



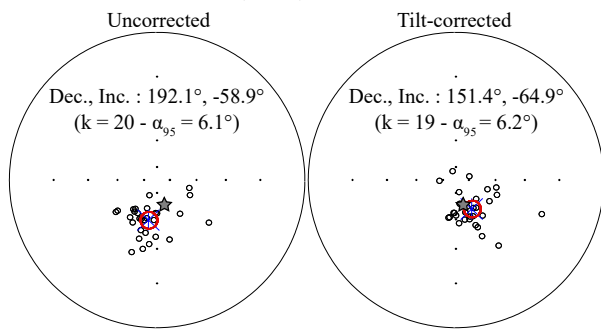
**e. S6C - 9**



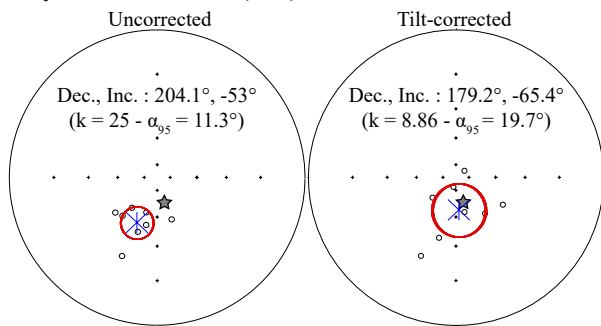
**f. S1B - 3**



**a. Flow-mean CHRMs (30/30)**

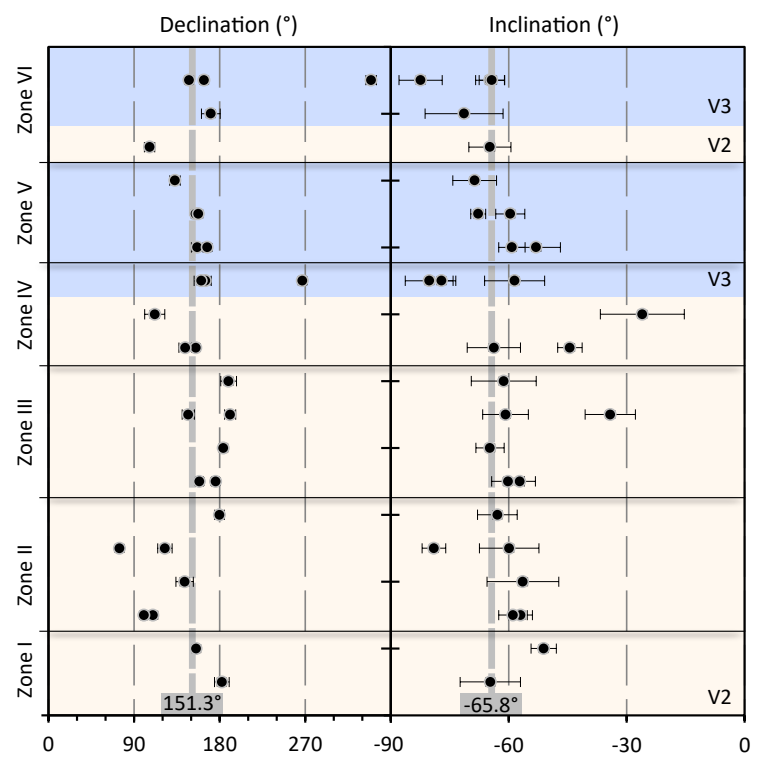


**b. Dyke-mean CHRMs (8/11)**

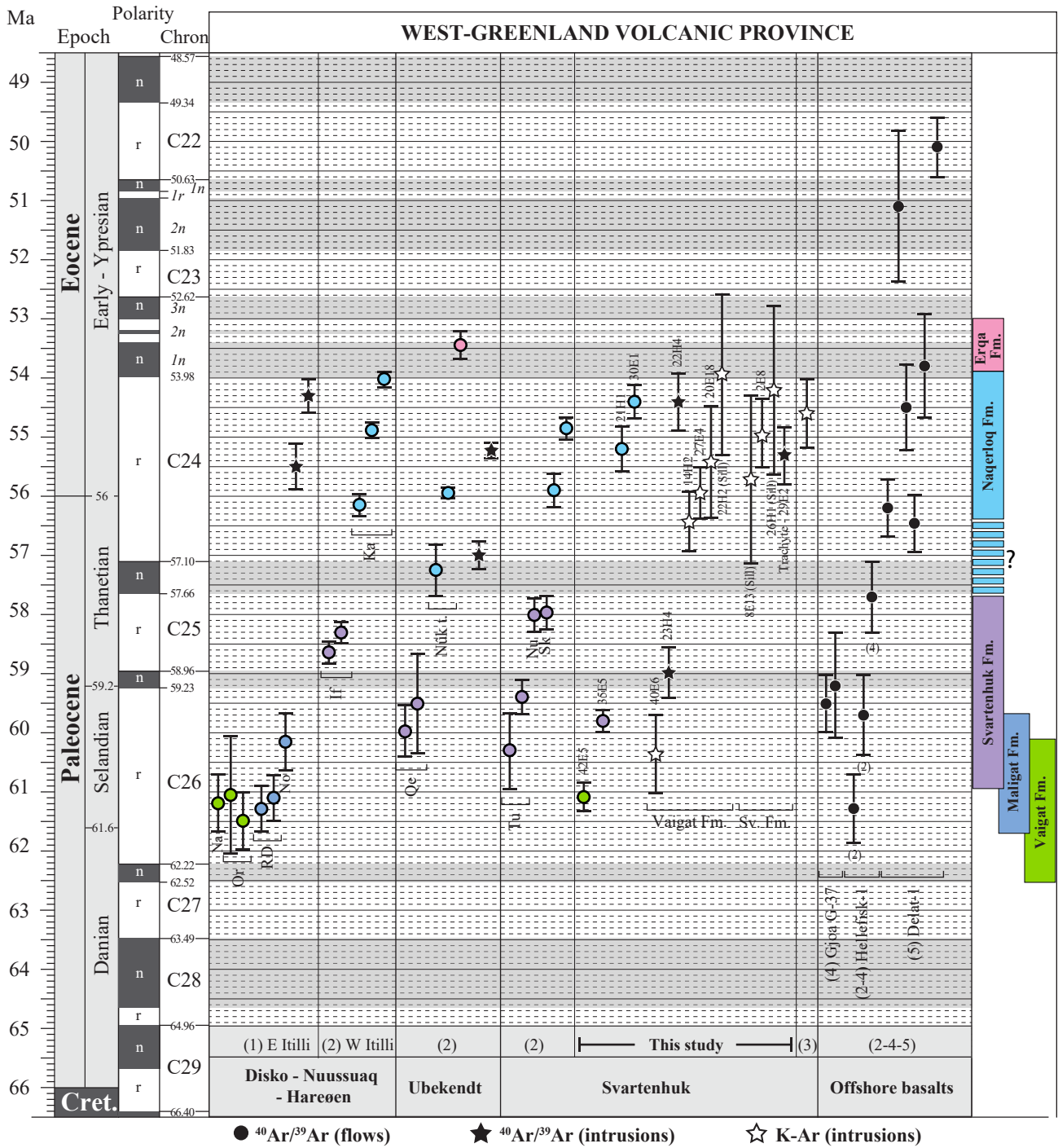


★ ChRM directions from Riisager et al. (2003) profiles in Svartenhuk (Vaigat Fm.) and Nuussuaq (Kanisut Mb.): N = 44; Dec., Inc. : 163.1°, -70.0° (k = 29.1 -  $\alpha_{95}$  = 4.1°)

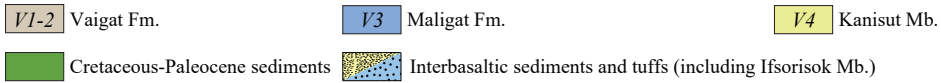
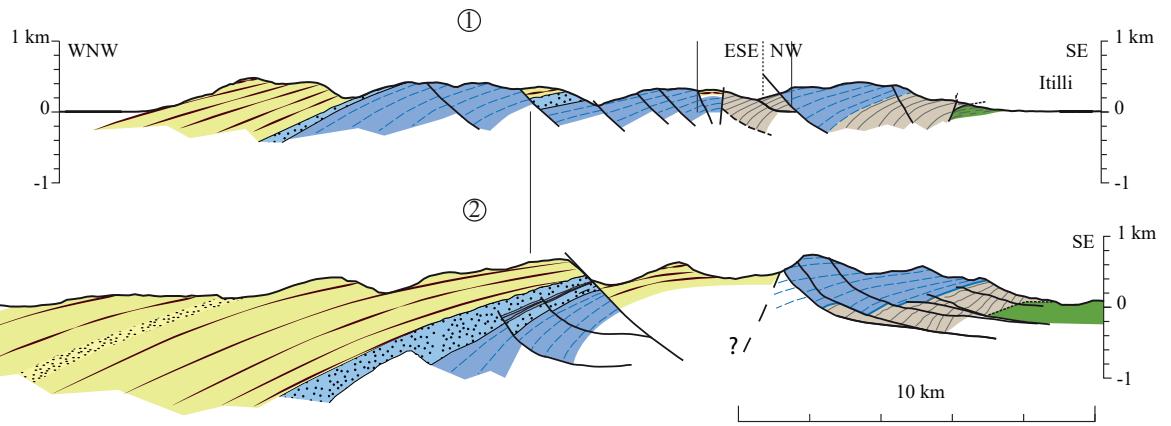
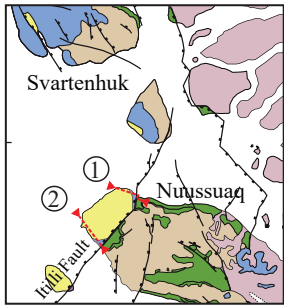
**c. Paleomagnetic flow-mean directions (tilt-corrected)**

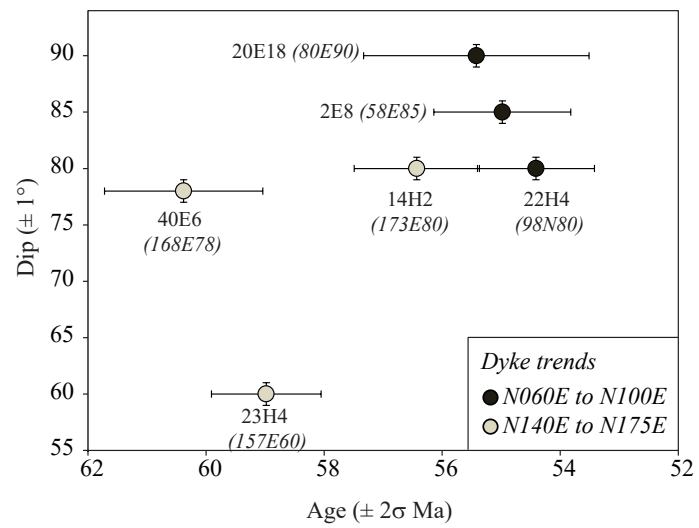


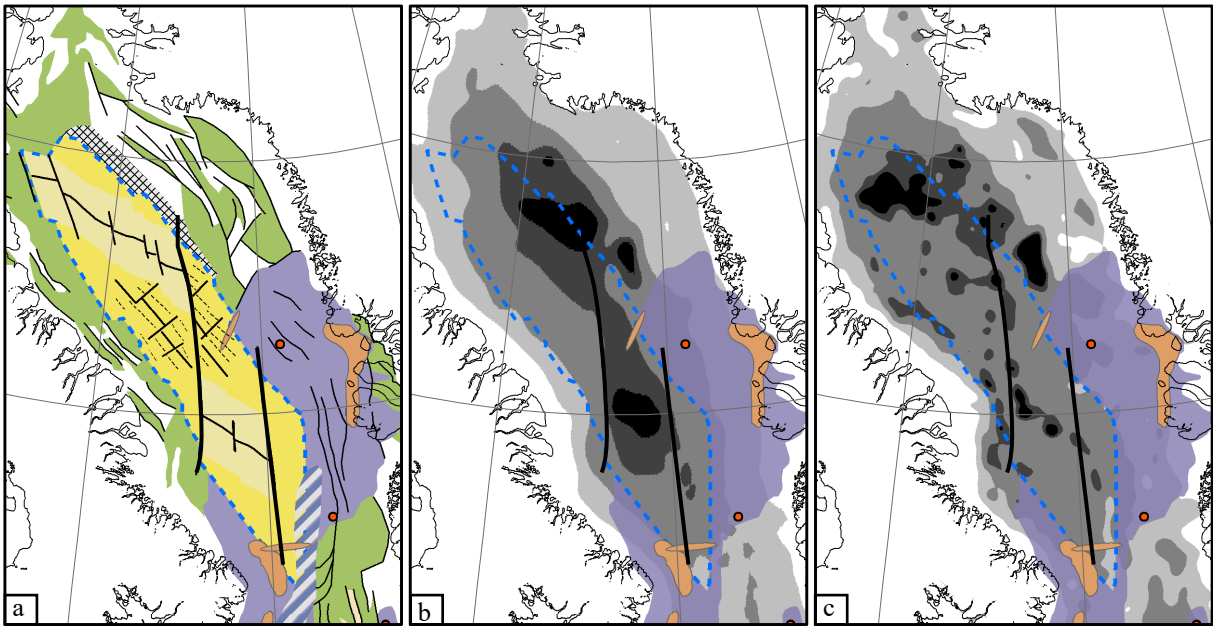




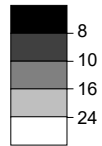








Crustal thickness (km)



SDRs

Tertiary basalts

Main transform fault

Wells with Paleogene basalts

Continent-ocean boundary from Oakey and Chalmers (2012)

**Table 1.** (a) Details of replicate analyses and K/Ar age determinations (classified in a stratigraphic order). Values reported with an asterisk have not been taken into account. Strike/dip of dykes is indicated in brackets. (b) Summary of  $^{40}\text{Ar}/^{39}\text{Ar}$  data from incremental heating experiments. Ages are calculated relative to the 28.201 Ma Fish Canyon sanidine standard.

a. K/Ar									
Sample	Lat. ( $^{\circ}\text{N}$ )	Long. ( $^{\circ}\text{W}$ )	Experiment no.	K (wt%) $\pm 1 \sigma$	Mass molten (g)	$^{40}\text{Ar}^*$ (%)	$^{40}\text{Ar}^*$ ( $10^{-11}$ mol./g) $\pm 1 \sigma$	$^{40}\text{Ar}^*$ weighted mean ( $10^{-11}$ mol./g) $\pm 1 \sigma$	Age (Ma) $\pm 2 \sigma$
<b>Flow</b>									
S40-E1	71.45477	53.89866	8631	0.470 $\pm$ 0.003	0.32556	41.975	4.863 $\pm$ 0.024		
			8647	« ..... »	0.31922	34.937	4.933 $\pm$ 0.025	4.898 $\pm$ 0.018	<b>59.11 <math>\pm</math> 0.83*</b>
S35-E5	71.40867	54.91839	8629	0.428 $\pm$ 0.004	0.30871	63.746	4.461 $\pm$ 0.023		
			8677	« ..... »	0.31594	26.988	4.524 $\pm$ 0.023		
			8716	« ..... »	0.31737	51.665	4.407 $\pm$ 0.22	4.463 $\pm$ 0.013	<b>59.15 <math>\pm</math> 0.85</b>
S33-E9	71.39358	55.21253	8634	0.205 $\pm$ 0.004	0.34632	61.668	2.083 $\pm$ 0.011		
			8650	« ..... »	0.50091	50.705	2.091 $\pm$ 0.011	2.087 $\pm$ 0.008	<b>57.78 <math>\pm</math> 1.88*</b>
S30-E1	71.50236	55.24889	8630	0.821 $\pm$ 0.013	0.32107	87.934	8.755 $\pm$ 0.043		
			8646	« ..... »	0.37134	80.374	8.542 $\pm$ 0.042		
			8717	« ..... »	0.41709	84.681	8.532 $\pm$ 0.043	8.550 $\pm$ 0.025	<b>59.07 <math>\pm</math> 1.16*</b>
S30-E3	71.50233	55.24106	8777	0.604 $\pm$ 0.008	0.30554	88.082	6.181 $\pm$ 0.031		
			8793	« ..... »	0.30494	74.402	6.158 $\pm$ 0.031	6.169 $\pm$ 0.022	<b>57.96 <math>\pm</math> 1.20*</b>
S30-E4	71.50233	55.24106	8654	0.834 $\pm$ 0.014	0.29949	80.107	8.621 $\pm$ 0.043		
			8670	« ..... »	0.2519	83.044	8.647 $\pm$ 0.043	8.634 $\pm$ 0.030	<b>58.73 <math>\pm</math> 1.53*</b>
S21-H1	71.53605	55.33622	8792	0.947 $\pm$ 0.013	0.29972	81.802	9.280 $\pm$ 0.046		
			8813	« ..... »	0.38571	94.611	9.355 $\pm$ 0.047	9.317 $\pm$ 0.035	<b>55.86 <math>\pm</math> 1.25</b>
<b>Sill</b>									
S22-H2	71.42281	54.49116	8800	0.136 $\pm$ 0.005	0.51055	22.029	1.310 $\pm$ 0.007		
			8806	« ..... »	0.43979	17.128	1.268 $\pm$ 0.007	1.289 $\pm$ 0.005	<b>53.85 <math>\pm</math> 2.84</b>
S36-E5	71.40308	54.92181	8635	0.222 $\pm$ 0.003	0.33573	39.503	2.468 $\pm$ 0.013		
			8651	« ..... »	0.50249	40.015	2.480 $\pm$ 0.013		
			8714	« ..... »	0.51316	35.472	2.448 $\pm$ 0.013	2.465 $\pm$ 0.007	<b>62.92 <math>\pm</math> 0.92*</b>
S26-H1	71.47894	55.08561	8628	0.228 $\pm$ 0.008	0.30007	30.645	2.185 $\pm$ 0.011		
			8644	« ..... »	0.50213	40.288	2.167 $\pm$ 0.011	2.176 $\pm$ 0.008	<b>54.20 <math>\pm</math> 2.88</b>
S8-E13	71.43780	55.01043	8805	0.190 $\pm$ 0.006	0.30856	22.17	1.856 $\pm$ 0.010		
			8821	« ..... »	0.30692	65.418	1.873 $\pm$ 0.010	1.865 $\pm$ 0.007	<b>55.73 <math>\pm</math> 2.49</b>
<b>Dyke</b>									
S40-E6	71.45472	53.89972	8773	0.556 $\pm$ 0.008	0.30171	22.037	5.959 $\pm$ 0.030		
			8789	« ..... »	0.30652	23	5.881 $\pm$ 0.031	5.921 $\pm$ 0.021	<b>60.38 <math>\pm</math> 1.34</b>
S14-H2	71.44078	54.78275	8782	0.764 $\pm$ 0.009	0.2251	76.085	7.475 $\pm$ 0.037		
			8798	« ..... »	0.3026	59.493	7.721 $\pm$ 0.039	7.594 $\pm$ 0.027	<b>56.43 <math>\pm</math> 1.06</b>
S27-E4	71.35933	54.75433	8823	0.519 $\pm$ 0.004	0.37472	76.323	5.092 $\pm$ 0.026		
			8839	« ..... »	0.38755	71.082	5.137 $\pm$ 0.026	5.114 $\pm$ 0.018	<b>55.95 <math>\pm</math> 0.88</b>
S20-E18	71.42483	54.11903	8657	0.654 $\pm$ 0.015	0.32671	37.396	6.431 $\pm$ 0.033		
			8673	« ..... »	0.29308	58.455	6.338 $\pm$ 0.032	6.383 $\pm$ 0.023	<b>55.42 <math>\pm</math> 1.91</b>
S2-E8	71.35819	54.79525	8781	0.621 $\pm$ 0.008	0.35445	78.369	5.892 $\pm$ 0.030		
			8797	« ..... »	0.3219	57.36	6.145 $\pm$ 0.031	6.0122 $\pm$ 0.021	<b>54.98 <math>\pm</math> 1.16</b>

b. $^{40}\text{Ar}/^{39}\text{Ar}$										
Sample	Experiment no.	wt. (mg)	K/Ca (total)	Total Fusion Age (Ma)	Increments used ( $^{\circ}\text{C}$ )	$^{39}\text{Ar}$ (%)	Isochron Analysis N	MSWD	$^{40}\text{Ar}/^{36}\text{Ar} \pm 2 \sigma$ intercept	Age $\pm 2 \sigma$ (Ma)
<b>Flow</b>										
S42-E5, groundmass (71.43083 $^{\circ}\text{N}$ - 54.08153 $^{\circ}\text{W}$ )	FG-1121 to FG-1129	101	0.06	60.00 $\pm$ 0.34	652-1016	80.9	7 of 9	1.08	293.4 $\pm$ 4.9	<b>61.08 <math>\pm</math> 0.56</b>
	FG-999 to FG-1009	138	0.12	59.60 $\pm$ 0.37	646-1098	100.0	11 of 11	1.59	293.2 $\pm$ 3.5	<b>59.78 <math>\pm</math> 0.41</b>
S21-H1, groundmass (71.53605 $^{\circ}\text{N}$ - 55.33622 $^{\circ}\text{W}$ )	FG-1151 to FG-1159	104	0.28	55.80 $\pm$ 0.76	839-1034	60.7	5 of 9	1.67	328.8 $\pm$ 114.0	<b>55.20 <math>\pm</math> 0.79</b>
	FG-1111 to FG-11120	106	0.34	55.46 $\pm$ 0.88	867-1100	70.3	5 of 9	1.89	307.9 $\pm$ 86.6	54.38 $\pm$ 0.90
S30-E1, groundmass (71.50236 $^{\circ}\text{N}$ - 55.24889 $^{\circ}\text{W}$ )	FG-1140 to FG-1148	100	0.30	55.59 $\pm$ 0.79	898-1092	60.2	5 of 9	1.21	380.0 $\pm$ 105.0	54.40 $\pm$ 0.81
	weighted mean plateau and isochron ages from two experiments: simple mean plateau and isochron ages from two experiments: <b>54.39 <math>\pm</math> 0.60</b>									
<b>Dyke</b>										
S23-H4, groundmass (71.39363 $^{\circ}\text{N}$ - 54.44445 $^{\circ}\text{W}$ )	FG-1160 to FG-1171	101	0.10	58.79 $\pm$ 0.80	839-1085	66.7	6 of 11	2.10	295.3 $\pm$ 10.3	<b>58.98 <math>\pm</math> 0.93</b>
	FG-1185 to FG-1195	106	0.24	54.74 $\pm$ 0.92	839-1225	86.5	8 of 11	1.60	300.5 $\pm$ 17.0	<b>54.41 <math>\pm</math> 0.99</b>
<b>Trachyte</b>										
S29-E2, feldspars (71.49475 $^{\circ}\text{N}$ - 55.21839 $^{\circ}\text{W}$ )	N1268-01 to N1268-08									
	7 of 8 2.00 319.5 $\pm$ 99.0 <b>55.3 <math>\pm</math> 1.09</b>									