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► To cite this version:

Arnaud Agranier, René C. Maury, Laurent Geoffroy, François Chauvet, Bernard Le Gall, et al.. Volcanic record of continental thinning in Baffin Bay margins: Insights from Svartenhuk Halvø Peninsula basalts, West Greenland. *Lithos*, 2019, 334, pp.117 - 140. 10.1016/j.lithos.2019.03.017 . hal-03485028

HAL Id: hal-03485028

<https://hal.science/hal-03485028>

Submitted on 20 Dec 2021

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1 **Volcanic record of continental thinning in Baffin Bay**
2 **margins: insights from Svartenhuk Halvø Peninsula basalts,**
3 **West Greenland**

4

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17

18 **Keywords**

19 Plume-related basalts

20 Enriched mantle sources

21 Partial melting

22 West Greenland Volcanic Province

23 North Atlantic Igneous Province

24 Volcanic passive margins

25

26 8088 words, 52773 characters

27

28 **Abstract**

29

30 We report major, trace elements and Pb, Hf, Nd and Sr isotopes in 61-54 Ma old basalts
31 from Svartenhuk Halvø Peninsula (Greenland). This area corresponds to the
32 northernmost exposure of the West Greenland volcanic province, of which the
33 emplacement marks the continental thinning and breakup process leading to the opening

34 of Baffin Bay during the Eocene. Its wedge structure displays typical characteristics of
35 inner seaward-dipping-reflectors (SDR) with an exposed volcanic sequence thicker than
36 7 km. Our results cover the entire volcanic sequence starting with an earliest V1 unit
37 reflecting rather low degree partial melts (transitional basalts), which are geochemically
38 imprinted by continental crust contamination. This unit is followed by much thicker and
39 mainly tholeiitic V2 and V3 basaltic lava piles, and ends with a rather thin V4 unit,
40 consisting of alkali basalts and associated trachytes. Pressures and temperatures of melt
41 extraction were estimated based on major element concentrations and rare earth
42 elements patterns. Our results suggest that melts were extracted from the garnet-spinel
43 transition zone at greater depths for the initial V1 and the final V4 lavas (alkali to
44 transitional basaltic compositions) than for the main V2 and V3 mostly tholeiitic lava
45 piles (at 2 ± 0.5 GPa and $1350 \pm 100^\circ\text{C}$). Isotope signatures suggest that mantle sources
46 of the melts were controlled by the mixing of ambient upper mantle and Icelandic
47 plume-type materials. The proportion of ambient upper mantle involved in the lava
48 source appears to increase together with melting rates and the upward propagation of
49 melting zone during V3 emplacement, suggesting that melting in this area progressed as
50 follows: 1. Initiation of partial melting (low degrees) of deep-seated Icelandic plume-
51 type material (V1 and V2); 2. Upward propagation of the melting area within the
52 shallower mixed upper mantle (V3); 3. Progressive decrease of “ambient upper mantle”
53 involvement, deepening of magma extraction and decrease of partial melting associated
54 with the emplacement of V4 lavas. We show that these changes are consistent with the
55 tectonic stages of continental thinning that preceded the Eocene breakup in Baffin Bay.

56

57 **1. Introduction**

58 Volcanic passive margins (VPM) characterize continental breakup associated with
59 the emplacement of large igneous provinces (White and McKenzie, 1989; Coffin and
60 Eldholm, 1994). In detail, initial mantle melting is thought to start prior to significant
61 lithosphere extension (Courtilot et al., 1999), leading to the eruption of voluminous
62 flood basalts. Then, focusing of mantle melting combined with lithosphere extension
63 and consecutive syn-magmatic break-up leads to the formation of volcanic passive
64 margins (Geoffroy et al., 2015). The corresponding volcanic sequences, although
65 emplaced subaerially, are usually found as immersed seaward-dipping reflectors (SDR)
66 sequences beneath post-rift sediments, a common occurrence throughout the North
67 Atlantic Igneous Province (NAIP; Saunders et al., 1997). Such sequences, linked to the

68 opening of the Baffin Bay oceanic basin, are exceptionally well-exposed along the West
69 Greenland coast, where they are of Paleocene-Eocene age and up to 5-7 km thick
70 (Larsen and Pulvertaft, 2000; Larsen and Pedersen, 2009; Larsen et al., 2015; Chauvet
71 et al., accepted). Petrologic and geochemical studies of these volcanic successions as
72 well as those of similar sequences exposed in the East Greenland margin raise several
73 important (and still debated) petrogenetic and tectono-magmatic questions, e.g. (1) do
74 the considerable volumes of erupted picrites indicate abnormally high temperatures of
75 partial melting of plume-related materials (Francis, 1985; Larsen and Pedersen, 2000,
76 2009; Hole and Millett, 2016); (2) does the heterogeneity of the mantle sources of
77 Paleocene-Eocene mafic magmas foreshadow that of the present-day Icelandic hotspot
78 (Holm et al., 1993; Peate and Stecher, 2003; Barker et al., 2006); and finally (3) does
79 the temporal evolution of melting rates and depths record the successive stages of
80 continental extension and rupture in the East (Larsen and Saunders, 1998; Abdelmalak
81 et al., 2012, 2018) and West Greenland (Tegner et al., 1998; Barker et al., 2006)
82 margins as well as in other parts of the NAIP (Kerr, 1994; Geoffroy, 2005; Meyer et al.,
83 2007; Saunders et al., 2007)?

84 Svartenhuk Halvø Peninsula represents the northernmost exposure of the West
85 Greenland Volcanic Province. The Svartenhuk wedge typically displays the structural
86 and stratigraphic characteristics of inner-SDR wedges observed on deep seismic
87 reflection profiles from VPM worldwide (e.g. Planke et al. 2000; Stica et al., 2014;
88 Geoffroy et al., 2015). The thickness of the exposed volcanic sequence may reach up to
89 7 km and at least one third of it is made of picrites (Larsen and Pulvertaft, 2000). The
90 aims of the present study are (1) to analyze magmatic processes such as mantle source
91 heterogeneity, depths and rates of partial melting, and crustal contamination that
92 occurred during the c. 10 m.y. of its emplacement (i.e. from 62.5-61 Ma to 54-53 Ma;
93 Larsen et al., 2015; Chauvet et al., accepted), and (2) to discuss their temporal evolution
94 during the successive stages of the continental stretching and thinning margin system
95 related to the opening of the Baffin Bay oceanic basin.

96
97

98 **2. Geological setting**

99

100 *2.1. The Labrador-Baffin oceanic domain and West Greenland Volcanic Province*

101 The Labrador-Baffin axis is part of a wider Phanerozoic divergent system
102 including the NE-Atlantic area between Eurasia and Greenland (Fig. 1). A continental
103 and diachronous rift system developed between Greenland and Canada during the
104 Cretaceous. This a-magmatic extension led to continental breakup with mantle
105 exhumation in the southern Labrador Sea. A now extinct oceanic spreading axis in the
106 Labrador Sea led to the drifting apart of Greenland from North America since the Late
107 Cretaceous (Roest and Srivastava, 1989) or Paleocene (Chalmers and Laursen, 1995)
108 and ceased to operate during the Oligocene (~ magnetochron C13, Roest and Srivastava,
109 1989). The SW-NE-trending left-lateral Ungava transform-fault system (UFS)
110 connected the Labrador Sea to Baffin Bay farther North (Fig. 1).

111 In the northern Labrador Sea, the oldest undisputed oceanic magnetic anomaly has
112 a mid-Paleocene age (Chalmers and Laursen, 1995), i.e. C27n (62.2 – 62.5 Ma,
113 according to Ogg, 2012). During the Paleocene, the spreading directions were oriented
114 ~N060E, i.e. sub-orthogonal to earlier rifted structures on the conjugate margins. Both
115 the oceanic fracture zones and magnetic anomalies in the Labrador Sea indicate that a
116 major shift in kinematics occurred during the Early Eocene, between C25 and C24,
117 when spreading directions rotated from N060E to N010E (Roest and Srivastava, 1989;
118 Geoffroy et al., 2001; Fig. 1).

119 In Baffin Bay, the age of the earliest accretion is still debated since no obvious
120 linear magnetic anomalies can be identified on its floor, although there is a consensus
121 for the cessation of spreading during the Oligocene (Roest and Srivastava, 1989; Oakey
122 and Chalmers, 2012). Two major N-S-trending fracture zones in the central part of the
123 Baffin Bay (Fig. 1) suggest that oceanic spreading may have been coeval with the ~N-S
124 reorientation of the kinematic trend in the Labrador Sea during Eocene (Chalmers and
125 Pulvertaft, 2001; Geoffroy et al., 2001; Suckro et al., 2012; Funck et al., 2012). Other
126 authors (Oakey and Chalmers, 2012; Suckro et al., 2012; Hosseinpour et al., 2013)
127 consider that earlier oceanic-type crust of mid-Paleocene age could exist within Baffin
128 Bay. However, none of these studies took into account the age of continental stretching
129 and thinning recorded along the Baffin Bay continental passive margins.

130 The conjugate passive margins in Baffin Bay (Fig. 1) are clearly of volcanic type
131 in the South of the basin (Geoffroy et al., 2001; Skaarup et al., 2006; Block et al., 2012).
132 The inner part of the SE-Baffin volcanic passive margin (VPM) is largely exposed from
133 latitudes 69°N to 72°N in the Disko-Svartenhuk area (Fig. 1). An inner-SDR (Planke et
134 al., 2000) developed orthogonally to the regional stretching direction during the

135 Paleogene (Geoffroy et al., 2001; Abdelmalak et al., 2012) and is fully exposed at
136 Svartenhuk where its total thickness reaches up to 7 km. Recently published Ar-Ar ages
137 (Larsen et al., 2015; Chauvet et al., accepted) that range from 62.5-61 Ma to 54-53 Ma
138 for Svartenhuk basalts provide new constraints for the tectono-magmatic evolution of
139 the Baffin Bay margin system.

140

141 *2.2. The Svartenhuk Halvø Peninsula*

142 The tectonics of the Svartenhuk SDR has already been investigated (Larsen and
143 Pulvertaft, 2000; Geoffroy et al., 2001; Abdelmalak et al., 2012). Until recently, the age
144 of the regional strain remained a matter of debate due to the lack of age determinations
145 on lavas in this area (see discussion in Larsen and Pulvertaft, 2000). However, new sets
146 of $^{40}\text{Ar}/^{39}\text{Ar}$ ages recently published by Larsen et al. (2015) and Chauvet et al.
147 (accepted) provide new constraints on the timing of melting processes at the scale of the
148 NAIP. In Svartenhuk Peninsula, the thickness of the exposed volcanic sequence may
149 reach up to 7 km (Larsen and Pulvertaft, 2000). Its dense fault network results from a
150 seaward-migrating strain (Geoffroy et al., 2001). The Arfertuarssuk Fault represents the
151 exposed master fault bounding and controlling the development of the Svartenhuk
152 wedge (Fig. 2).

153 In the present study and in the companion paper by Chauvet et al. (accepted)
154 dealing with the same sample set, the volcanic formations making up the SDR wedge in
155 Svartenhuk Halvø have been subdivided into four successive sequences (labeled V1 to
156 V4 upwards; Fig. 2) (Fig. 2). The boundaries between these sequences are consistent
157 with the geological and petrographic distinctions of Larsen (1983) and Larsen and
158 Grocott (1992).

159 Coeval volcanics crop out on Disko Island, Nuussuaq Peninsula and Ubekendt
160 Ejland Island, where they form a plurikilometric pile of ultramafic to mafic Paleogene
161 lavas crosscut by sheeted intrusions (e.g. Pedersen et al., 2002). The stratigraphy and
162 age of the W-Greenland basalts were defined in these areas as follows (Storey et al.,
163 1998 and references therein). The Vaigat Fm. consists of a lower sequence dominated
164 by hyaloclastites and pillow lavas overlain by picrites and Mg-rich subaerial basalts.
165 The earliest subaqueous volcanics unconformably overlie the remnants of an Early
166 Cretaceous to Paleogene paleobasin (Fig. 2). The overlying Maligât Fm. is dominated
167 by feldspar-phyric and aphyric basalts associated with a sharp change in lava chemistry
168 (Larsen and Pedersen, 2009). Larsen and Pulvertaft (2000) suggested that the lava

169 stratigraphy established in the Disko-Nuussuaq area could be applied to the Svartenhuk
170 area. As such, the V1 and V2 formations correlate with the southern Vaigat Formation,
171 while the “Svartenhuk Fm.” (V3 in Fig. 2) would be equivalent to the Maligât Fm. (Fig.
172 3). In Svartenhuk Halvø, a 1-2 km thick succession of transitional to alkaline basalts
173 (V4 unit) lies above the V3 unit. A stratiform sill-like trachytic body marks the
174 boundary between the V3 and V4 units. Larsen and Grocott (1992) considered these
175 basalts and trachyte as the uppermost part of the Svartenhuk Fm. According to Larsen et
176 al. (2015), this succession belongs to a newly introduced Naqerloq Fm. (Fig. 2),
177 dominated by transitional light rare Earth element (LREE) enriched basalts.

178 From observations on the Disko and Nuussuaq areas, two main volcanic pulses
179 have been identified and further correlated by radiometric dating and paleomagnetic
180 data (Storey et al., 1998; Riisager and Abrahamsen, 1999; Riisager et al., 2003). The
181 earliest pulse was mid-Paleocene in age and spans chrons 27n and 26r (62.5-59.2 Ma
182 after Ogg, 2012). The second pulse occurred during the early Eocene at C24r (54-
183 57 Ma), i.e., after a hiatus of 2 m.y. or more. An Early Eocene age was obtained from
184 the basaltic succession of the Kanísut Mb. (Storey et al., 1998), that crops out
185 exclusively West of the Itilli fault, and which is seaward tilted at about 15° (Fig. 2,
186 Riisager et al., 2003). These basalts lie unconformably above the Vaigat-Maligat
187 basalts, which are in turn deeply down-faulted westward (Larsen et al., 2015). It is
188 noteworthy that early Eocene ages were also obtained for dykes crosscutting the
189 Paleocene-related series in Disko and Nuussuaq (Storey et al., 1998) and also in the
190 Svartenhuk Peninsula (Geoffroy et al., 2001). Larsen et al. (2015) proposed that the
191 Paleocene magmatic pulse in West Greenland spans not only C27n-C26r but also up to
192 chron C25r (59-57.7 Ma), based on four $^{40}\text{Ar}/^{39}\text{Ar}$ ages ranging from 60.3 ± 1.4 Ma to
193 58.0 ± 0.6 Ma on Svartenhuk Fm. basalts from the northern part of the Svartenhuk
194 peninsula. This implies that the Svartenhuk Fm. was mostly contemporaneous with the
195 Maligât Fm. but that its upper part was younger. These authors also outlined that the
196 transitional to alkaline basaltic successions exposed along the Arfertuarssuk Fault in
197 Svartenhuk Halvø (V4 unit) and on the western parts of the Ubekendt Ejland, Nuussuaq
198 peninsula and Hareøen are all early Eocene in age (Naqerloq and Erqa Fms). Chauvet et
199 al. (accepted) point out that both in W-Nuussuaq and Svartenhuk the Eocene lavas are
200 part of the syn-extension inner SDR, i.e. that continental crust thinning and stretching
201 still occurred during the Eocene (fig. 2).

202

203 2.3. *Svartenhuk volcanic facies*

204 We investigated and sampled 45 sites mostly located along the coastal cross-
205 section of the Svartenhuk volcanic margin (Fig. 1). Hyaloclastic sub-aquatic breccias
206 and associated pillowed lavas formed in a lacustrine environment (Escher and
207 Pulvertaft, 1995) are characteristic of the lower member of the Vaigat Fm (V1 unit,
208 sampling sites 40 and 38). They are up to 200 m thick and dip 20-30° westwards. These
209 dips are mostly of tectonic origin but may also be due to dipping prograding foresets
210 (Jerram, 2002), which have been identified for instance in the Naajaat syn-volcanic
211 lake, Nuussuaq Basin (Pedersen et al., 1998). Svartenhuk hyaloclastic breccias consist
212 of basaltic glassy clasts and pillow fragments, usually of centimeter- to decimeter-size,
213 set in a matrix of yellowish glass particles altered to clays and palagonite. Locally,
214 preserved pillow piles form cliffs up to 10-15 m high which delineate small high points
215 on the shoreline. Individual basaltic pillows, 10 cm to 1 m large, display typical glassy
216 margins and radial fractures together with flow features consistent with their
217 emplacement over moderate slopes. Attempts to recover geochemically fresh glassy
218 basaltic rims were unsuccessful, but the cores of large pillows proved to be rather well
219 preserved from post-magmatic alteration.

220 Lava flows from V2, V3 and V4 units were exclusively emplaced in subaerial
221 environments as suggested by lack of pillow lavas and hyaloclastites and common
222 occurrence of pahoehoe surface features. Their most common facies, especially in V3
223 and V4 units, is that of individual tabular flows, 1 to 10 m thick. They generally show a
224 basal scoriaceous autoclastic breccia, which overlies a reddish thermal metamorphic
225 sole. These breccias are, in turn, overlain by a massive basaltic part often displaying
226 rough columnar-jointing patterns, and less frequently sheet flow patterns. The rather
227 common occurrence, in these massive levels, of subvertical vesicle pipes, sometimes
228 feeding horizontal vesicle sheets (Caroff et al., 2000), suggests that the Svartenhuk
229 Peninsula basaltic magmas were water-rich at the time of their emplacement. Tuff
230 horizons usually less than 50 cm thick are occasionally interbedded between these
231 tabular flows, and probably correspond to subaerial ash and/or lapilli fall deposits.
232 Intercalated reworked (volcanic-sedimentary) deposits are much less common than in
233 Disko and Nuussuaq (Larsen and Pedersen, 2000). They correspond to basaltic pebble
234 conglomerates of probable laharic origin, filling local topographic channels. Their
235 maximal thickness is less than 10 m (sites 44 and 45), and they are sometimes overlain
236 by layered clay horizons less than 1 m thick.

237 Vaigat Fm picrites (V2 unit) and less commonly Vaigat and Svartenhuk Fms
238 tholeiitic basalts (V2 and V3) display a more specific emplacement facies, i.e. a
239 compound-braided flow facies (Jerram, 2002). The corresponding lava piles, up to 30 m
240 thick (site 39), have a typical grayish color also characteristic of Disko and Nuussuaq
241 picrites, (Larsen and Pedersen (2009), and are usually deeply altered and converted into
242 assemblages of post-magmatic minerals (mostly zeolites and carbonates). They consist
243 of thin (usually less than 50 cm thick) anastomosed individual layers emplaced as sheets
244 and lobes of pahoehoe lava breakouts. These individual lobes are elongated (up to
245 several meters) and usually hollow; all the transitions are observed between meter-size
246 lobes and typical lava tubes. Well preserved small pahoehoe ropy lava “tongues” often
247 occur within the central cavity of these lobes or tubes, and provide good markers of
248 flow emplacement patterns. In V2 and V3 units they point usually upwards due to the
249 tectonic tilting of these sequences. Massive (and geochemically better preserved)
250 porphyritic layers 50 cm to 2-3 m thick are common within the lobe/tube piles. They
251 display intrusive relationships with respect to them (locally chilled margins) and are
252 thought to result from inflation processes within the pahoehoe compound flow
253 sequences (Jerram, 2002; Self et al., 1997). In the studied picritic lava sequences, both
254 pahoehoe lobes/tubes and intrusive (inflation-related) sheets commonly contain large
255 vesicles and vesicle pipes typical of water-rich magma emplacement features (Caroff et
256 al., 2000). These facies are often ascribed to relatively low volume, low viscosity
257 eruptions occurring during early stages of emplacement of continental flood basalts
258 (Jerram, 2002).

259 Most dykes range in width from 10-20 cm to 2-3 m, and display characteristic
260 jointing patterns perpendicular to their chilled margins (which are glassy to microlitic)
261 and have coarser-grained, often porphyritic, vesicle-rich central parts. These dykes tend
262 to form swarms associated with fracture zones, and composite (multiple) dykes are
263 common. Sills are usually easily distinguished from lava flows by their larger and more
264 regular columnar jointing patterns. The thickest ones (> 10 m) display evidence for
265 gravitational settling of olivine phenocrysts, which accumulate above their lower chilled
266 margin.

267

268 **3. Sampling and analytical methods**

269

270 *3.1. Sampling strategy*

271 Sampling was conducted in 45 sites mostly located along the coastal cross-section
272 of the volcanic margin (Table 1 and Fig. 2), from the basal lacustrine hyaloclastites and
273 pillow basalts of the Vaigat Fm (labelled V1 in the geochemical discussion below; sites
274 40 and 38) towards successively: the subaerial picritic and basaltic flows of the main
275 Vaigat Fm (labelled V2; sites 37, 39, 41, 42, 20, 43, 45, 44, 5, 3, 6, 7, 1, 27, 25, 28);
276 then those of the overlying Svartenhuk Fm (labelled V3; sites 4, 2, 36, 35, 8, 31, 32, 11,
277 10, 34, 12, 9, 33); and finally the trachytic intrusion (site 29) and the subaerial alkali
278 basaltic lava flows from the top of the Svartenhuk Fm, recently attributed to the
279 Naqerloq Fm by Larsen et al. (2015) (site 30, labelled V4 below). Additional samplings
280 were done for V2 (sites 19, 22, 23, 16, 14, 15, 13), V3 (site 26) and V4 (site 21). Major,
281 trace element and isotopic measurements were only carried out on samples from lava
282 flows, dykes and sills which were relatively fresh based on their macroscopic and
283 microscopic features, and from which zeolite- and carbonate-bearing vesicles were
284 removed by handpicking. The corresponding analytical database includes 5 samples
285 from V1, 53 from V2, 28 from V3 and 5 from V4 lava flows, together with 47 intrusive
286 (sills, dikes) samples (Tables 1, 2, 3 and 4). Most of the pillow fragments collected from
287 V1 are altered (Loss on Ignition higher than 1.9 wt%), and a number of compound-
288 braided basaltic flows (pahoehoe lobes and lava tube piles, see below), especially from
289 V2, are too rich in secondary minerals (zeolites and carbonates) to have retained their
290 primary magmatic features. Such samples were primarily collected to investigate these
291 post-magmatic mineral associations found as vesicle fillings and vein/fracture deposits
292 (57 samples from V2 and V3).

293

294 *3.2. Analytical methods*

295 All samples were subsequently crushed in stainless steel and agate mortars. Major
296 element concentrations were obtained using an ICP-AES Jobin Yvon Ultima 2 at the
297 University of Brest, after a HF-HNO₃ digestion, boric acid neutralization and dilution in
298 nitric acid, as described in Cotten et al. (1995). Trace element concentrations were
299 measured with a Thermo® Element2 HR-ICP-MS in Brest. 15 mg of samples were
300 repeatedly (twice) dissolved at 120°C in 2 mL Teflon Savilex® beakers using in 1mL
301 HF + 0.5 mL HClO₄. After total evaporation, samples were diluted 10000 times using
302 0.5N HNO₃ and analyzed using both medium and low resolution on the Thermo®
303 Element2 as discribed by Mougél et al. (2014). All samples were externally calibrated
304 using international standards BCR2, BHVO2 and BIR-1 (Li and Lee, 2006). These

305 standards demonstrated external reproducibilities better than 5–10% depending on the
306 element and concentration as documented in Supplementary Table 1. Pb, Hf, Sr and Nd
307 isotope compositions were obtained on single fractions (0.5-1 g of samples) leached for
308 one hour at 100°C in 6mL of 6N HCl, and subsequently dissolved in 3mL HF + 1mL
309 HNO₃ following the protocol described by Mougél et al. (2014). Hf, Nd and Pb were
310 purified following the protocol of Blichert-Toft et al. (2005) (fluoride centrifugation,
311 and column chromatography). Sr was separated from the Nd fraction using Eichrom Sr-
312 specific resin following the protocol by Moynier et al. (2010), adapted from Fietzke and
313 Eisenhauer (2006). Pb, Hf and Nd isotopes were analysed using a Thermo® Neptune
314 MC-ICP-MS at Ifremer-Brest. Sr isotopes were obtained using a Thermo Triton TIMS
315 at the University of Brest. 2σ internal errors for Hf and Nd are reported in Table 4 and
316 never exceed the last significant digit (1E-5) in each analysis. External reproducibility
317 of Pb isotopes is estimated based on the repeated measurements since 2008 of a mixed
318 standard solution and never exceeded 60 ppm for ²⁰⁶Pb/²⁰⁴Pb, 150 ppm for ²⁰⁷Pb/²⁰⁴Pb
319 and 300 ppm for ²⁰⁸Pb/²⁰⁴Pb. Measured values were compared with repeatedly analyzed
320 international standards NBS981 (Todt et al., 1996), JMC475 (Blichert-Toft et al., 1997)
321 and La Jolla48 (Tanaka et al., 2000) shown in Supplementary Table 2. Duplicated Sr,
322 Nd, Hf, Pb isotope analyses of one of our samples (S39E1) are shown in Supplementary
323 Table 3.

324

325 **4. Petrographic and geochemical results**

326

327 *4.1. Petrographic features and lava classification*

328 Lava samples were classified according to their position in the Total Alkali-Silica
329 plot (TAS; Le Bas et al., 1986; Fig 3a). With the exception of a trachytic intrusion from
330 V4 (site 29), all of them plot within or near the fields of microbasalts and basalts (SiO₂=
331 43-53 wt%, K₂O+Na₂O=0.1-5 wt%). A large majority of them are silica-saturated or -
332 oversaturated, and plot below the alkaline-subalkaline boundary of Macdonald and
333 Katsura (1964). Therefore, these samples, which include all the massive or pillowed
334 flows of units V1, V2 and V3, the sills and about half the population of dykes can be
335 termed as tholeiitic or transitional picrites and basalts. Alkali basalts plotting above this
336 boundary include the V4 basaltic flows (site 30) and the other half of the analyzed
337 dykes. Some V3 basalts plot near or slightly above Macdonald and Katsura's (1964)

338 boundary (Fig. 3a) and can be considered as transitional, as well as V1 basalts which
339 plot definitively below this limit but display strong enrichments in highly incompatible
340 elements (see below).

341 Pillowed tholeiitic basalts from V1 unit display groundmass textures ranging from
342 glassy (devitrified) rims to microlitic cores. They are subaphyric, with 4-8 modal%
343 phenocrysts and microphenocrysts of olivine (mostly altered to smectite and chlorite)
344 and fresh plagioclase ranging in size from 0.1 to 0.5 mm. Their groundmass is variably
345 altered into clays and palagonite-type mineral aggregates. Zeolites and carbonates are
346 found as vesicle- and vein-filling materials.

347 Picrites and picritic basalts (microbasalts in Fig. 3a) are much more common in
348 subaerial Vaigat Fm flows (V2) than in the overlying main unit of Svartenhuk Fm (V3).
349 They are highly porphyritic, with up to 60 modal% centimeter- to millimeter-size
350 olivine phenocrysts, containing occasional magnesiochromite inclusions, and
351 subordinate (and smaller) plagioclase phenocrysts and microphenocrysts. The
352 percentage of olivine phenocrysts and their size tend to decrease together with
353 decreasing MgO contents of bulk rocks, while the relative abundance of plagioclase
354 increases. Clinopyroxene microphenocrysts are uncommon in picrites, and this mineral
355 follows plagioclase in the crystallization sequence of V2 and V3 lavas. The picritic
356 basalts from these two units are rather similar in thin section. Their groundmass is
357 usually intergranular or microlitic. Neither native iron, nor graphite has been identified
358 in the studied picrites. This suggests a weaker contamination in these samples
359 (originally selected for their primitive features) than that described for some more acidic
360 samples from Disko and Nuussuaq (Larsen and Pedersen, 2009).

361 Tholeiitic basalts from V2 and V3 are the most common petrographic type
362 encountered in the studied sequences. Their phenocryst contents are highly variable and
363 decrease together with MgO contents of bulk rocks, from 25-20 modal% (porphyritic
364 basalts) to less than 5 modal% (subaphyric basalts). Olivine is followed by calcic
365 plagioclase, and then sometimes by minor amounts of augitic clinopyroxene in their
366 crystallization sequence. Plagioclase phenocrysts dominate over olivine in the most
367 evolved (silica-rich) samples, which are especially common in the Svartenhuk Fm main
368 unit (V3). Clinopyroxene phenocrysts mostly occur in this V3 unit, and are uncommon
369 in V2 basalts. The groundmass of tholeiitic basalts usually display microlitic to
370 intergranular textures. Subdoleritic textures occur in sills and in the cores of thick flows
371 and dykes.

372 Post-magmatic mineral associations are commonly found in V2 and V3 tholeiitic
373 picrites and basalts. They occur as vesicle- and vein-fillings, and include a variety of
374 zeolites, carbonates (calcite, Mg-calcite, aragonite, dolomite), and clays (smectites,
375 chlorites).

376 Alkali basalts from Naqerloq Fm. (V4) and from a number of dykes display
377 microlitic to intergranular groundmass textures. They contain about 10 to 15 modal%
378 phenocrysts, which are, by order of decreasing abundance, calcic plagioclase (up to 0.5
379 cm large), diopsidic clinopyroxene and magnesian olivine. They contain significantly
380 smaller amounts of post-magmatic minerals than V1, V2 and V3 lavas.

381 The fresh alkali trachyte collected at site 29 (V4 unit) is highly porphyritic, with
382 ~40 modal% centimeter-size sanidine phenocrysts, and lesser amounts of smaller
383 titanomagnetite, olivine and green clinopyroxene phenocrysts. Its groundmass, almost
384 exclusively made of alkali feldspar laths, displays a fluidal texture.

385

386 *4.2. Major element trends*

387 The Loss on Ignition (LOI) values (Table 2) range from 0.58 wt% to 10 wt%,
388 with an average of 2.8 wt%. These rather high values are accounted for by the common
389 occurrence of post-magmatic water-rich minerals (zeolites and clays) and of carbonates
390 in the studied lavas. The variations of SiO₂ and MgO in the studied basaltic flows (V1
391 to V4) are plotted against their Ni, CaO and alkali element contents. The higher MgO
392 contents of V2 lavas (Table 2) evidence the dominance of picrites (MgO > 12 wt%) and
393 Mg-rich basalts in the Vaigat subaerial flows. An upward increase of TiO₂ is also
394 observed in the stratigraphic pile: contents higher than 2 wt%, usually considered
395 typical of High-Ti basalts from Large Igneous Provinces (Pik et al., 1998; Shellnutt and
396 Jahn, 2011), are only observed in some V3 evolved tholeiitic and transitional basalts
397 and in V4 alkali basalts (Table 2). Conversely, all V1 and V2 basalts and a majority of
398 V3 tholeiites are typical Low-Ti basalts (TiO₂ < 2 wt%, Table 2). Similar features have
399 been reported by (Larsen and Pedersen, 2009) for the Disko and Nuussuaq areas, where
400 the lower Vaigat Fm is exclusively made of Low-Ti basalts while High-Ti lavas occur
401 within the overlying Maligât Fm (lateral equivalent of Svartenhuk Fm according to
402 Escher and Pulvertaft (1995).

403 Major element variations in the studied sequence are dominated (and largely
404 overprinted) by processes linked to accumulation or removal/fractionation of olivine
405 phenocrysts. Indeed, as shown in Figure 3B, addition or removal of magnesian olivine

406 are consistent with the covariation of CaO and MgO: olivine-rich picritic basalts are
407 CaO-poor while evolved basalts are CaO-rich. These variations are partly accounted for
408 by local gravitational settling effects involving olivine phenocrysts in thick lava flows
409 or sills. However, such shallow processes during and after emplacement are unlikely to
410 occur in thin pahoehoe units and meter-thick tabular flows which make up most of the
411 Vaigat and Svartenhuk Fms sequences. Therefore, fractional crystallization involving
412 olivine separation/accumulation must have occurred at depth, possibly within thick sills
413 or genuine underlying magma chambers. In addition, the least Mg-rich basalts, mostly
414 sampled from dykes, display a co-linear decrease of CaO and MgO (Fig. 3B), that is
415 consistent with concomitant removal/ fractionation of olivine plus augite and/or calcic
416 plagioclase. Such variations might result from differential motions of crystal-rich and
417 crystal-poor liquids during dyke emplacement (“Bagnold effect”, e.g. Komar, 1972),
418 and thus cannot be taken as evidence for deep-seated fractional crystallization involving
419 olivine, augite and plagioclase.

420

421 *4.3. Trace element variations*

422 Compatible trace elements Ni, Co and Cr, which are hosted by olivine and
423 magnesiochromite, display positive correlations with MgO contents (Fig. 3B). Their
424 contents decrease from picrites to evolved basalts, and their behaviors are mainly
425 controlled by olivine removal/accumulation processes.

426 Incompatible element concentrations vary in a more complex fashion, as
427 illustrated by Rare Earth Element (REE) plots normalized to CI chondrite (Fig. 4). V1
428 transitional basalts display the most fractionated REE patterns and the highest contents
429 in Light REE (LREE; e.g. La, Ce), up to 80 times the primitive mantle (PM). V4 alkali
430 basalts also display fractionated REE patterns, although their LREE enrichment remains
431 lower than that of V1 lavas. V2 and V3 REE patterns are fan-shaped, and display
432 variable enrichments in LREE, those of picritic samples being the lowest. The REE
433 patterns of dykes are also fan-shaped, and they encompass the variability found within
434 lava flows (Fig. 4). Alkali basaltic dykes show fractionated patterns and LREE
435 enrichments similar to those observed for V4 alkali basaltic flows. A second family of
436 dykes (and the analyzed sills) has rather flat patterns which resemble those of picrites;
437 and a third dyke family shows moderately LREE enriched patterns similar to those of
438 V2 and V3 evolved tholeiitic basalts. Large Ion Lithophile Elements (LILE: Cs, Rb, Ba,
439 Sr) display somewhat erratic behaviors, with highly variable enrichments or depletions

440 generating positive or negative spikes in incompatible multi-element patterns
441 normalized to the primitive mantle (Figure 4B). Given the high solubility of these
442 elements into water, such variations can be considered as post-magmatic and attributed
443 to alteration by surface waters and/or percolation of hydrothermal fluids through the
444 studied lava piles. The elements less affected by post-magmatic processes (ie rare earth
445 elements, high field strength elements...) define patterns similar to that of typical OIB
446 (Fig. 5).

447 Highly incompatible elements (e.g. Th and La) versus MgO (not shown) in V2
448 and V3 lavas define a rough negative correlation (not shown) consistent with the
449 “dilution” of incompatible elements in bulk rocks as a result of accumulation of
450 incompatible element-poor olivine phenocrysts in picritic and Mg-rich lavas. Compared
451 to V2 and V3, V4 alkali basalts and alkaline dykes display higher Th and La contents
452 (Fig. 4) consistent with their more fractionated REE and multi-element patterns, as well
453 as higher Nb/La ratios (Fig. 5). These features might indicate either lower degrees of
454 partial melting, or a more incompatible element enriched source (with respect to V2 and
455 V3 ones), or a combination of both. Finally, despite their transitional character, some
456 V1 pillowed lavas (together with a few dykes) show the highest La, Th and Pb contents
457 found in the studied lava sequence (Figs. 4 and 5), and consistent with their highly
458 fractionated REE and multi-element patterns. Once again, these features could be
459 attributed to a more enriched source, lower degrees of melting, or both.

460

461 *4.4. Sr, Nd, Pb and Hf isotopes*

462 Lead isotopes define two distinct trends (Figs. 6 and 7). $^{208}\text{Pb}/^{204}\text{Pb}$ range from
463 36.7 to 39, $^{207}\text{Pb}/^{204}\text{Pb}$ from 15 to 15.5 and $^{206}\text{Pb}/^{204}\text{Pb}$ from 16.13 to 19.26. A principal
464 component analysis calculated on these Pb isotopes reveals the existence of only two
465 significant principal components (the first one accounting for 94.2% of the total
466 variability, principal component 2 for about 5.3%, and principal component 3 is
467 insignificant). This result indicates that a mixture of three end-members is sufficient to
468 explain the isotope variability in this area (Blichert-Toft et al., 2005). Hafnium isotopes
469 are scattered as well, and feature both radiogenic and unradiogenic signatures
470 ($^{176}\text{Hf}/^{177}\text{Hf}$ from 0.28247 to 0.2832; $\epsilon_{\text{Hf}} = -10.81$ to $+15.95$). They also define a three
471 end-members-pattern corresponding to two distinct mixing trends (Fig. 6). Nd isotopes
472 correlate with Hf data ranging from $^{143}\text{Nd}/^{144}\text{Nd} = 0.51222$ ($\epsilon_{\text{Nd}} = -8.15$) to

473 $^{143}\text{Nd}/^{144}\text{Nd}=0.51309$ ($\epsilon_{\text{Nd}}=8.9$). Even if most of the samples have “mantle-like” Sr
474 isotopic compositions ($^{87}\text{Sr}/^{86}\text{Sr} < 0.704$), 16 of them display significantly more
475 radiogenic Sr signatures (up to $^{87}\text{Sr}/^{86}\text{Sr} = 0.708$). Ten out of those 16 samples also have
476 unradiogenic Pb ($^{206}\text{Pb}/^{204}\text{Pb} < 17$), and trace element patterns (low Nb, Ta and high Pb
477 contents, Fig. 5) typical of continental crust contamination (Rudnick and Fountain,
478 1995; Rudnick and Gao, 2003). These samples include all of those collected from the
479 basal unit V1.

480 The two mixing trends identified in Sr, Pb and Hf isotope spaces, are also visible
481 on diagrams using isotopic and REE ratios (e.g., $^{206}\text{Pb}/^{204}\text{Pb}$, La/Sm, Sm/Yb and La/Yb,
482 Fig. 8). They are also related to the stratigraphic position of the samples, i.e. their height
483 within the lava pile. Indeed, all the samples from basal units V1 and V2 lie within the
484 same trend (Figs. 6 and 7) connecting a component “Sk1” (with low $^{207}\text{Pb}/^{204}\text{Pb}$ and
485 “intermediate” $^{208}\text{Pb}/^{204}\text{Pb}$) to a moderately high $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ component
486 “Sk2”. Complementarily, all samples from units V3 and V4, together with those from
487 sills and dykes, plot along a trend connecting “Sk2” to a component “Sk3” (with high
488 $^{208}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$).

489

490 **5. Discussion**

491

492 *5.1. Trace element and isotopic evidence for a minor role of crustal contamination*

493 The highly incompatible elements Nb and Ta display positive or negative peaks in
494 multi-element patterns (Fig. 4B) which are unlikely of post-magmatic origin
495 considering the insoluble character of these High Field Strength Elements (HFSE).
496 Nb/La ratios show a tentative negative correlation with SiO_2 , whereas Pb contents are
497 positively correlated with silica for V1 and V2 lavas (Fig. 5). These features likely
498 reflect crustal contamination effects in these two lowermost units, since silica-rich
499 continental crust is selectively depleted in Nb and Ta and enriched in Pb with respect to
500 elements of similar incompatibility, e.g. La (Tang et al, 2019 and references therein)..

501 Component Sk1 is clearly distinct from either “typical” mantle-reservoirs (Hart
502 et al., 1992), or any North Atlantic volcanic rocks. Its radiogenic Sr, unradiogenic Hf
503 and Nd, as well as the low Nb, Ta, and high Pb contents (Fig. 5) of lavas plotting close
504 to it (Figs. 6 and 7) suggest that the origin of this component is related to continental
505 crust contamination of mantle-derived melts by materials similar in composition to

506 Greenland basement late Archean orthogneisses and granodiorites analyzed by Larsen
507 and Pedersen (2009). The fact that these contamination effects are limited to the oldest
508 units V1 and V2 is consistent with the hypothesis of an uprise of the younger melts
509 (constituting units V3 and V4) through a plumbing system coated with earlier basaltic
510 lavas (e.g. Arndt and Christensen, 1992) or exhausted of fertile material by previous
511 melting events (e.g. Meade et al, 2014).

512

513 *5.2. Partial melting conditions*

514 Si and Mg contents can be used as proxies in order to estimate pressure and
515 temperature of melt extraction in the mantle (Lee et al., 2009). Only a limited number of
516 samples are “primitive enough” (i.e. experienced only olivine fractionation), to
517 adequately estimate their pristine major element concentrations by simple olivine
518 addition (Albarède, 1992; Lee et al., 2009). Moreover, some of them, with MgO
519 contents >15 wt%, need to be afterwards corrected for olivine accumulation (back to
520 equilibrium with their olivine phenocrysts Mg#~0.83, Supplementary Table 4) and
521 olivine fractionation (back to equilibrium with olivine of mantle-like composition
522 Mg#~0.89). Thermobarometric estimates thus obtained on ten samples from units V2
523 and V3 suggest temperatures and pressures of $\sim 1350 \pm 100$ °C and 2.5 ± 0.5 GPa,
524 respectively. These results, although consistent with those recently reported by Hole and
525 Natland (2019), need to be considered with caution, given the number of
526 approximations in the thermobarometric approach, as well as uncertainties concerning
527 the precise geological setting at that time (crustal thickness). The estimated temperature
528 range suggests that the mantle beneath Svartenhuk was not very hot compared with
529 other published evaluations for the NAIP (e.g. 1480-1550 °C, Hole and Millett, 2016;
530 see also Herzberg et al. 2007 and references therein). Our estimated P-T conditions
531 overlap the spinel/garnet transition zone in the mantle (2.2 GPa at 1300 °C, O’Neill,
532 1981; Robinson and Wood, 1998), and it is therefore probable that Svartenhuk Fm.
533 lavas were extracted from melting columns involving both garnet- and spinel-bearing
534 peridotites, as often proposed for NAIP magmas (Kerr, 1994; Kerr et al, 1999; Saunders
535 et al., 2007; Hole and Millett, 2016).

536 Even if the “fan-shaped” rare earth elements (REE) patterns observed on figure 4
537 are partly the result of different rates of partial melting, LREE/HREE fractionation can
538 also be accounted for by the presence of residual garnet in the source (e.g. Frey et al.
539 1978, Thirlwall et al., 1994). In figure 8 plots, Pb isotopes, La/Sm and Sm/Yb ratios are

540 compared with the average slope of REE spectra quantified by La/Yb ratios. The upper
541 diagram ($^{206}\text{Pb}/^{204}\text{Pb}$ vs La/Yb) reflects the effects of crustal contamination (units V1
542 and V2) on geochemical compositions (see section 5.1). The middle diagram documents
543 the concomitant effects of crustal contamination and degrees of partial melting on
544 LREE/intermediate REE relative abundances. The lower plot provides insights
545 concerning the amount of garnet-bearing peridotites involved in the melting column.
546 Altogether, these three diagrams suggest that:

547 i/ Lavas derived from rather high degrees of partial melting all belong to units V2
548 and V3. Basalts from units V1 and V4 were generated by smaller degrees of
549 mantle melting (as suggested by their alkali to transitional signatures in Figs. 4
550 and 5). Precise amounts of melting are difficult to assert considering the
551 uncertainties concerning the initial composition(s) of mantle source(s).
552 Nevertheless, models of garnet- (shown in red) and spinel-bearing (shown in
553 green) batch lherzolite melting (see Gurenko and Chaussidon, 1995 for details on
554 initial compositions, McKenzy and O'nion, 1991 for partition coefficients; and
555 Thirlwall et al.1993 for mineral proportions in the melting) show that V1 and V4
556 lavas would be derived from melting degrees lower than 5%, and most V2 and
557 V3 lavas from clearly higher ones (Fig. 8).

558 ii/ On average, basalts from units V2 and V3 feature Sm/Yb ratios 50% lower
559 than those of samples from units V1 and V4. This stronger fractionation between
560 intermediate and heavy REE in V1 and V4 suggests a larger involvement of
561 garnet peridotites (and therefore greater melting depths) at the beginning and the
562 end of the magmatic history than during the emplacement of the main V2-V3
563 volcanic sequences. These results are in agreement with previous petrogenetic
564 studies of West Greenland picritic and Mg-rich basalts demonstrating that their
565 main sequences derived predominantly from spinel-bearing peridotites (Gill et al.,
566 1991; Ryabchikov et al., 1998; Larsen and Pedersen, 2000, 2009).

567 Therefore, we propose that small degrees of melting leading to V1 lavas were
568 initiated at depths greater than 75 km below a relatively thick lithosphere. Subsequently,
569 these melts propagated upwards into the spinel peridotite zone in connection with
570 lithospheric thinning. The melting rates increased consistently during this process,
571 forming V2 and V3 lavas. Finally, the melting zone deepened again, and melting rates
572 decreased, thereby generating V4 alkali basalts.

573

574 *5.3. Mantle source heterogeneity and its temporal evolution*

575 The isotope diversity reported in this study covers a range similar to that
576 previously reported for the neighboring Disko Island (Larsen and Pedersen, 2009) and
577 the only differences concern the extent of crustal contamination (Fig. 5). The least
578 contaminated samples (identified on Figs. 5 and 8) also largely overlap with samples
579 from Iceland and the adjacent ridges, as well as with those from East Greenland (figs. 6
580 and 7). The isotopic diversity within the Iceland plume has been widely discussed
581 (Chauvel and Hémond, 2000; Thirlwall et al., 2004; Kokfelt et al., 2006; Maclennan,
582 2008; Kitawaga et al., 2008) and is denoted in figure 7 by both depleted (D, Kitawaga et
583 al., 2008; ID-1 and ID-2, Thirlwall et al., 2004), and enriched (E-1 and E-2, Kitawaga et
584 al., 2008; IE-1 and IE-2, Thirlwall et al., 2004) Icelandic internal components (Fig. 7).
585 In figures 6 and 7, Hf, Sr and Pb compositions of component “Sk2” are very similar to
586 those of OIB and E-MORB related to the Iceland hotspot, and more especially to a
587 mixture of the Icelandic inner components “D” (for depleted) and “E-1” (for Enriched
588 1) of Kitawaga et al. (2008) or “ID1” – “ID2” (for Iceland Depleted 1 and 2) and “IE1”
589 (for Iceland enriched 1) of Thirlwall et al. (2004).

590 Alternatively, Sk3 seems to correspond to the Iceland E-2 enriched component of
591 Kitawaga et al. (2008) (or “IE-1” of Thirlwall et al., 2004) that point toward the
592 common component C (Hanan and Graham, 1996) in all isotope spaces. This enriched
593 component has been proposed to be intrinsic to the Icelandic hotspot (Hanan and
594 Schilling, 1997; Kitawaga et al., 2008; Thirlwall et al., 2004; Peate et al., 2010). The
595 presence of distinct Icelandic isotopic components has also been described in Eastern
596 Greenland (Barker et al., 2006). Nevertheless, in that region, the different signatures
597 appear in separate areas (Milne, Rømer and Skræterne), located hundreds of kilometers
598 away from each other. In the present study, we report distinct signatures on samples all
599 coming from a restricted segment (<40 km along the coast) and covering a time span
600 shorter than 8 m.y. Such diversity can therefore hardly be imagined as the result of the
601 tapping of distinct parts of an Iceland type mantle plume. Instead, we suggest that the
602 Sk3 signature in our dataset corresponds to a trend toward the C component of Hanan
603 and Graham (1996). This component has been recognized as being widely disseminated
604 within the upper mantle feeding the Arctic segments of the mid-Atlantic ridges and is
605 therefore believed to correspond to an ubiquitous ingredient of the North Atlantic upper
606 mantle (Blichert-Toft et al., 2005).

607 Nb, Zr and Y concentrations also suggest consistency between uncontaminated
608 Svartenhuk samples and Iceland plume geochemical signatures. Indeed, Fitton et al.
609 (1997) used the comparison of Nb/Y and Zr/Y in order to decipher whether the
610 magmatic sources of basalts are more likely related to the Iceland plume or to the
611 neighboring MORB mantle. They proposed a ΔNb index $= 1.72 + \log(\text{Nb}/\text{Y}) -$
612 $1.92\log(\text{Zr}/\text{Y})$; where ΔNb is negative in MORB samples, but positive in Iceland
613 plume-derived material. Even if this approach has been criticized as possibly biased by
614 differences in pressures of melting (Hanan et al., 2000; Stracke et al., 2003), it is worth
615 noting that most samples from this study feature positive ΔNb values. Indeed, in Figure
616 9, only some samples from units V2 and V3 display negative ΔNb (Fig. 9), suggesting
617 that the contribution of Icelandic material in lava sources was lower during these
618 intermediate stages than during the emplacement of the earliest and latest V1 and V4
619 units. Moreover, a plume imprint on geochemical signatures of Svartenhuk lavas is also
620 consistent with the extremely primitive helium isotopic compositions recently published
621 by Péron et al. (2018) for olivines from our V3 samples S33E7 and S35E7 ($R/R_a=31.0$
622 and 34.9, respectively), similar to those previously reported for samples from Disko
623 Island and Baffin Island (Graham et al., 1998; Jackson et al., 2010).

624 Altogether, these features suggest that the Sk2-Sk3 trend (common to both V3 and
625 V4 units) could be accounted for by a mixture of Icelandic-type material (Sk2) and a
626 mixed ambient upper mantle reservoir (Sk3). The respective proportions of these two
627 components appear to evolve through time, since only samples coming from unit V3
628 carry isotope signatures suggesting the involvement of significant amounts of C-like
629 material, whereas the uppermost V4 lavas seem to tap preferentially Iceland hotspot
630 material. Furthermore, even if geochemical signatures of V1 and V2 are partly
631 controlled by crustal contamination, they also point solely toward the Iceland-like
632 signature. This sequence of magmatic events is summarized in figures 10 (petrogenetic
633 constraints) and 11 (speculative tectonic interpretation). During the emplacements of
634 basal units V1 and V2 (61 to 59.5 Ma old, Chauvet et al., accepted), initially low, but
635 temporally increasing degrees of melting of enriched Icelandic plume material
636 (displaying Sk2 isotopic signatures) generated basaltic melts at depths starting beneath
637 75 km, and these magmas were contaminated during their ascent through the continental
638 crust. During V3 emplacement (59.5-57.7 Ma), the melting process affected a wider
639 range of lithologies in the spinel stability field and propagated to neighboring

640 (shallower) materials from the upper mantle (involving material displaying component
641 Sk3 isotope signatures), with little or undetectable crustal contamination. Finally, during
642 V4 emplacement, i.e. between 57.7 and 53.3 Ma (Chauvet et al., accepted), the melting
643 zone deepened again and was once more mostly restricted to the enriched mantle
644 material (Figs. 10 and 11).

645

646 *5.4. Baffin Bay breakup and tectonic controls of mantle melting*

647 This sequence of events as reported by combining geochemical and tectonic data can
648 be interpreted as the progressive initiation of decompression melting of an upwelling
649 mantle below a progressively thinning lithosphere in a continental extension setting
650 (Fig. 11). The thinning observed between the emplacement of V1 and V2-V3 lavas
651 likely results from a combination of the tectonic stretching of the upper part of the
652 lithosphere and of the thermal erosion of its base by convecting hot mantle (Fig. 11).
653 Such a combination has been proposed as an explanation of the magmatic history of
654 Mull and Skye islands (Kerr, 1994; Kerr et al., 1999; Saunders et al., 1997). In these
655 islands, as well as in Giant's Causeway basalts (Barrat and Nesbitt, 1996), temporally
656 increasing melting rates and decreasing melting pressures similar to those described
657 in Svartenhuk are observed, despite rather different tectonic contexts characterized by
658 lower extension rates (Meyer et al., 2007). A temporal decrease of lithospheric
659 thickness seems the most likely factor controlling such changes in NAIP basaltic
660 magma compositions (Ellam, 1992; Kerr, 1994).

661 The decreasing involvement of the ambient mantle (component Sk3) contribution from
662 V3 to V4 (Fig. 10), and the concomitant deepening and decrease of partial melting rates
663 (back to alkali basaltic compositions) in this most proximal part of the volcanic margin,
664 may be (at least partly) attributed to the gradual westward shift of extension with time
665 (Geoffroy et al., 2015; Chauvet et al., accepted). Indeed, the innermost Svartenhuk
666 margin progressively became passive although extension still continued (as attested by
667 syn-tectonic control of V4 emplacement), albeit at lower rates (Chauvet et al, accepted).
668 It is also possible that the accumulation beneath the lithosphere of refractory mantle
669 residues left after melt extraction blocked the ascent of hot and fertile mantle sources
670 and resulted in deeper, lower-degree melting as previously suggested for East Greenland
671 flood basalts (Fram and Leshner, 1993, 1997).

672

673 **6. Conclusions**

674 Geochemical data obtained on 93 samples of basaltic to trachytic rocks from the
675 Svartehuk peninsula allow us to document the c. 7 m.y. long volcanic activity
676 accompanying the continental thinning process that preceded the Eocene opening of
677 Baffin Bay, the chronological and tectonic aspects of which are discussed in a
678 companion paper (Chauvet et al., accepted). The sequence of magmatic events started
679 with (i) the deep (>75 km) low degree partial melting of an Iceland plume-like source,
680 followed by (ii) shallower and higher degrees of melting of a mixture of Iceland hotspot
681 and “ambient upper mantle” materials, and finally ended with (iii) lower degrees of
682 melting of a deeper source dominated again by Icelandic-type signatures. The first
683 melting event could have been triggered by the adiabatic ascent of the Iceland plume
684 below a ~75-km-thick lithosphere, inherited from the previous Mesozoic extension. The
685 OIB-type signature of V1 lavas is clear despite the crustal contamination processes that
686 they experienced during their ascent. The following melting events (V2 to V3 lavas)
687 occurred at decreasing pressures and were coeval with syn-magmatic extension in the
688 upper crust (Chauvet et al., accepted) and therefore a subsequent reduction in
689 lithosphere thickness (Figs. 10 and 11). Thus, V2-V3 melts were a direct consequence
690 of the convective upwelling of Icelandic OIB-type mantle in the necked sub-lithospheric
691 area (Boutiller and Keen, 1999), associated with lithospheric thinning resulting from a
692 combination of tectonic stretching and thermal erosion of the Svartehuk lithosphere.
693 The later decrease in magmatic activity (emplacement of V4 lavas) was concomitant
694 with a progressive focusing of syn-magmatic strain towards the central axis of the
695 proto-Baffin Bay during the Eocene. This shift in extension was likely accompanied by
696 a progressive cooling (and thus, thickening) of the proximal W-Greenland lithosphere
697 over a still convecting asthenosphere. The combined decrease in melting rates and the
698 return to enriched source materials suggest a further deepening of the melting zone,
699 possibly linked to the accumulation beneath the lithosphere of refractory mantle
700 residues.

701

702 **Acknowledgments**

703 This study is part of the Petrobras-UBO Volcabasin program. The field work was
704 carried out during a five-week expedition in the summer of 2012 and benefited from the
705 logistic support of the research vessel Porsild from the Arctic Station in Godhavn
706 (Greenland) and the French polar yacht Vagabond. Philippe Nonnotte’s help was
707 especially appreciated before and during the fieldwork in Greenland. We also thank the

708 Vagabond family, who provided pleasant and efficient assistance, both at sea and in the
709 field. AA and RCM thank Claire Bassoullet, Philippe Nonnotte, Céline Liorzou and
710 Bleuenn Gueguen for their help during the analytical work. AA thanks David Graham
711 and Barry Hanan for insightful discussions. We are grateful to Elishevah Van Kooten
712 for her careful reading. The manuscript was considerably improved following the
713 suggestions of Editor A.C. Kerr and the thorough reviews of Drs. R. Gooday and N.
714 Arndt.

715

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1060
1061

1062 **Figure captions**

1063 Figure 1: Simplified structural map of Greenland and Northern Atlantic Igneous
1064 Province, modified after Chauvet et al (accepted). The studied area (red square) is
1065 located on the western volcanic margin of Greenland.

1066

1067 Figure 2: The 45 sampling sites (numbers) located on the simplified geological map
1068 of Svartenhuk Peninsula (West Greenland), modified after Chauvet et al. (accepted).
1069 Locations of cross sections AA', BB' and CC' are shown on the map.

1070

1071 Figure 3: Major element compositions of samples from units V1 (purple triangles),
1072 V2 (beige circles), V3 (blue hexagons) and V4 (yellow triangles). A: Total alkali –
1073 Silica plot (Le Bas et al., 1986). The thick black line (Macdonald and Katsura, 1964)
1074 separates the fields of subalkaline basaltic series (below) from alkali basaltic series
1075 (above).

1076 B: CaO vs SiO₂ and Ni vs SiO₂ plots. Compositions range from basaltic to trachytic
1077 (the latter in V4 unit only) following olivine and plagioclase + pyroxene fractionation
1078 trends. Dashed line identifies primitive Ni concentrations (higher than 400 ppm) and
1079 dotted line denotes samples with MgO=12 wt%, consistent with olivine phenocryst
1080 compositions (Mg#=0.83, supplementary table 1).

1081

1082 Figure 4: Multielement patterns of Svartenhuk samples. A: Rare earth element (REE)
1083 patterns normalized to CI Chondrite (McDonough and Sun, 1995) for samples from
1084 units V1 (pink), V2 (beige), V3 (blue) and V4 (yellow). Samples from V1 and V4
1085 have comparable REE patterns with significant light REE enrichments compared with

1086 heavy REE, typical of alkali and transitional basaltic lavas. Samples from units V2
1087 and V3 have more diverse signatures ranging from alkali to tholeiitic (LREE
1088 depleted) patterns. B: Multielement plot normalized to Primitive mantle (McDonough
1089 and Sun, 1995) compiling all samples (V1=purple lines, V2= beige lines, V3= blue
1090 lines, V4= yellow lines) and the median compositions of each group (V1= purple
1091 triangles, V2= beige circles, V3= blue hexagons, V4= yellow triangles).

1092

1093 Figure 5: Plot of SiO₂ vs Pb concentrations and Nb/La ratios. Continental crust
1094 contamination typically generates SiO₂ and Pb enrichments (Rudnick and Fountain,
1095 1995) associated with Nb depletions (Tang et al, 2019 and references therein)
1096 illustrated here by low Nb/La ratios. Crustal contamination trends (black arrows) can
1097 be identified within samples from units V1 and V2 only.

1098

1099 Figure 6: Pb, Hf, Sr and Nd isotopic data. Samples from units V1 to V4 (see figure 3
1100 for color codes) are plotted together with basalts from the North Atlantic ridge : blue
1101 crosses (Agranier et al., 2005; Blichert-Toft et al., 2005), Iceland : gray squares:
1102 (Breddam, 2002; Chauvel and Hémond, 2000; Hanan and Schilling, 1997; Hanan et
1103 al., 2000, Kitawaga et al., 2008; Kokfelt et al., 2009; Kuritani et al., 2011; Peate et al.,
1104 2010, 2009; Sims et al., 2013; Stracke et al., 2003; Thirlwall et al., 2004; Willbold et
1105 al., 2009) and East Greenland: orange crosses (Barker et al., 2006). Samples from this
1106 study plot between three end-members materialized by letters Sk1, Sk2 and Sk3. The
1107 Sk1-Sk2 trend can be accounted for by crustal contamination (Sk1 features radiogenic
1108 Sr, and unradiogenic Pb, Hf and Nd); while Sk2-Sk3 largely overlaps with mantle like
1109 signatures (Kitawaga et al. 2008) including North Mid-Atlantic Ridge and Iceland
1110 basalts. Its trend in all diagrams appears to point toward the common component C of
1111 Hanan and Graham (1996).

1112

1113 Figure 7: Pb isotopic data. Samples from this study are represented together with basalts
1114 from Iceland (gray squares, see figure 6 for references) and East Greenland (Barker et
1115 al., 2006) (orange crosses). The Iceland plume inner components: D, E-1 and E-2
1116 (Kitawaga et al., 2008; Thirlwall et al., 2004) are also located in this diagram. Apart
1117 from continental crust trend (segment Sk1-Sk2), all samples plot within in a triangle
1118 delineated by D, E-1 and E-2.

1119

1120 Figure 8: Plots of $^{206}\text{Pb}/^{204}\text{Pb}$, La/Sm, Sm/Yb against La/Yb. Pb isotopes (upper
1121 diagram) evidence crustal contamination for some V1 and V2 lavas and mantle-like
1122 compositions for most of the sample set. La/Sm as well as Sm/Yb provide insights
1123 about melting rates. On the Sm/Yb plot are reported for comparison the results of
1124 simple batch melting models considering garnet bearing (red line) and spinel bearing
1125 (green line) depleted lherzolite as proposed for Iceland by Gurenko and Chaussidon
1126 (1995), partition coefficients by McKenzie and O’Nions (1991) and mineral
1127 proportions by Thirlwall et al. (1993). Numbers from 0.02 to 0.1 denote F values. The
1128 higher Sm/Yb values observed for most of V1 and V4 lavas suggest a stronger
1129 involvement of garnet in their sources.

1130

1131 Figure 9: Plots of $^{206}\text{Pb}/^{204}\text{Pb}$, La/Sm and Sm/Yb against $\Delta\text{Nb} = 1.74 + \log(\text{Nb}/\text{Y}) -$
1132 $1.92 \log(\text{Zr}/\text{Y})$ (Fitton et al., 1997). ΔNb is considered as a bench marker of Iceland
1133 plume affinity. Only samples from units V2 and V3 display negative ΔNb , indicative
1134 of the contribution of “normal” upper mantle devoid of plume related enriched
1135 component. The thick black arrows illustrate the temporal evolution of Icelandic
1136 material contribution from V1 to V4.

1137

1138 Figure 10: Sketch summarizing the four main stages of Svartenhuk volcanism deduced
1139 from geochemical and petrologic constraints. Sk1, Sk2 and Sk3 represent the three
1140 geochemical reservoirs defined in figures 6 and 7. The tectonic evolution of the
1141 lithosphere (crustal thinning) is more precisely depicted in figure 11. During V1 stage,
1142 decompression melting affects Iceland plume-like material ascending through the
1143 asthenosphere below an already thinned lithosphere. During V2, the ongoing crustal
1144 thinning leads to an uprise of the isotherms and therefore to the upward propagation of
1145 partial melting and to its increase. During the emplacement of these two first units,
1146 geochemical signatures (benchmarked by Pb isotopes on the bottom diagram) can be
1147 accounted for by the simple binary mixture of components Sk1 and Sk2. During V3, the
1148 melting has propagated within Sk3 reservoir. Finally, during V4, the melting deepens
1149 again and is once more mostly restricted to reservoir Sk2. See text for details.

1150

1151 Figure 11: Tentative tectonic sketches depicting Svartenhuk magmatic evolution in
1152 relation with the evolution of South Baffin Bay during the Paleogene, at the times of
1153 V1, V2-V3 and V4 emplacements. Sk1, Sk2 and Sk3 refer to crustal and mantle

1154 reservoirs discussed in the text and shown in figure 10. Convection geometry is purely
1155 conceptual, and based on small scale convection models (Petersen et al., 2018; Simon et
1156 al., 2009). In the V1 sketch (top left), the occurrence of Cretaceous sediments, shown in
1157 light green suggests that crustal thinning started during the Cretaceous. The melting
1158 zones are indicated with coloured dots. Red boxes show the location of our field and
1159 petrologic investigations. The shaded areas on the left of some diagrams underline the
1160 speculative character of these interpretations when extrapolated to areas located away
1161 from our study area.

1162

1163 Table 1: Sample locations. K/Ar and Ar/Ar ages are from Chauvet et al (accepted)

1164 Table 2: Major element data (oxide wt%, total iron as Fe₂O₃, LOI: Loss On Ignition).

1165 Analytical methods are presented in the text.

1166 Table 3: Trace element data (ppm). Analytical methods are presented in the text.

1167 Table 4: Pb, Hf, Sr and Nd isotopic data. Analytical methods are presented in the text.

1168

1169 Supplementary Table 1: ICP-MS repeated trace element analyses for standards BHVO2,
1170 BCR-2 and BIR1.

1171 Supplementary Table 2: MC-ICP-MS repeated Nd, Hf and Pb isotope analyses of
1172 standards JNDi-1, JMC475 and NIST981, respectively.

1173 Supplementary Table 3: Replicated analyses of Pb, Hf, Nd and Sr isotopic compositions
1174 of sample S39E1.

1175 Supplementary Table 4: Olivine compositions obtained on Cameca SX100 electron
1176 microprobe at Service Commun “Microsonde Ouest” (Plouzané), at 15 kV
1177 accelerating voltage with a probe current of 12 nA.

1178

Figure 1

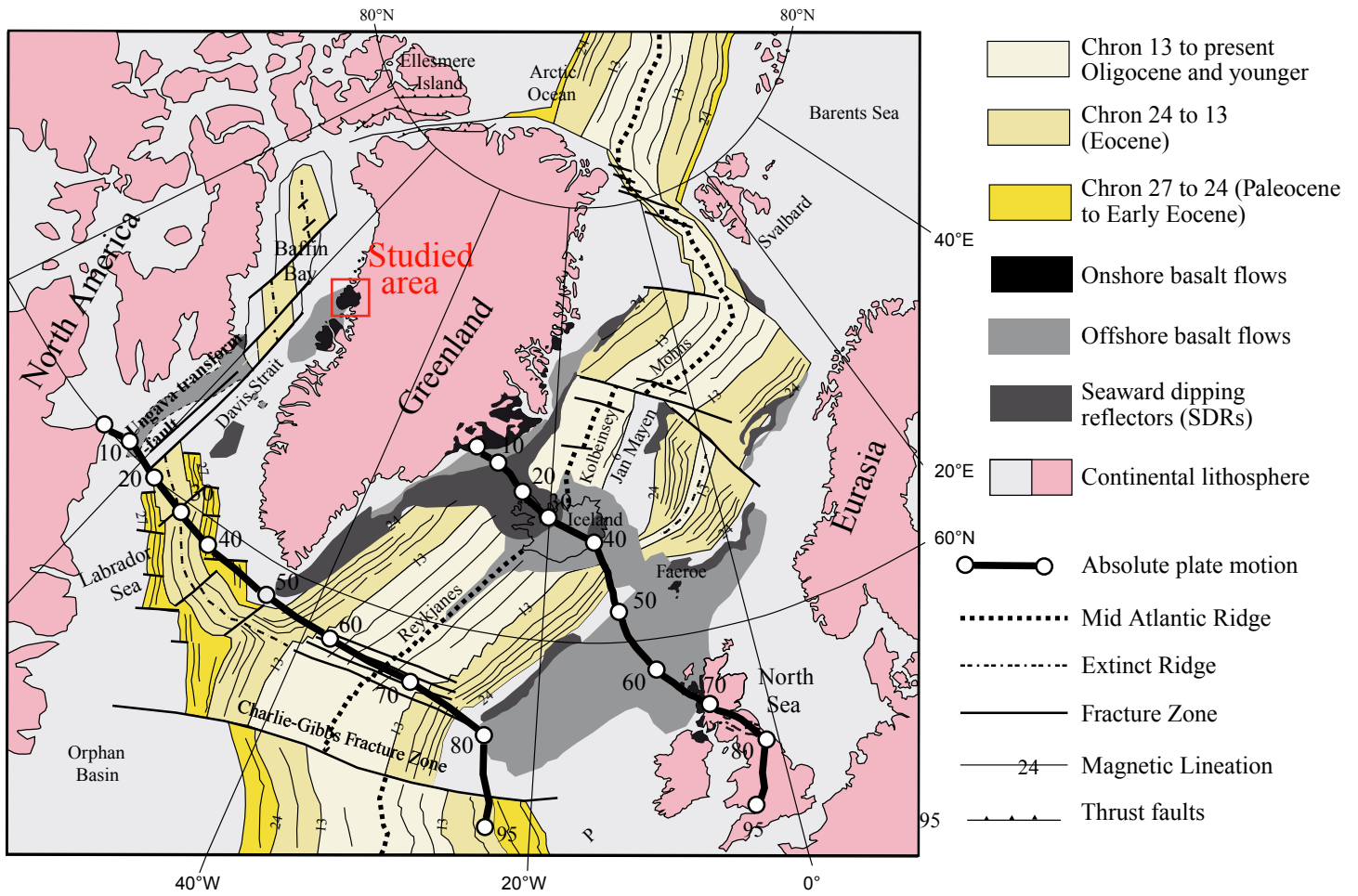
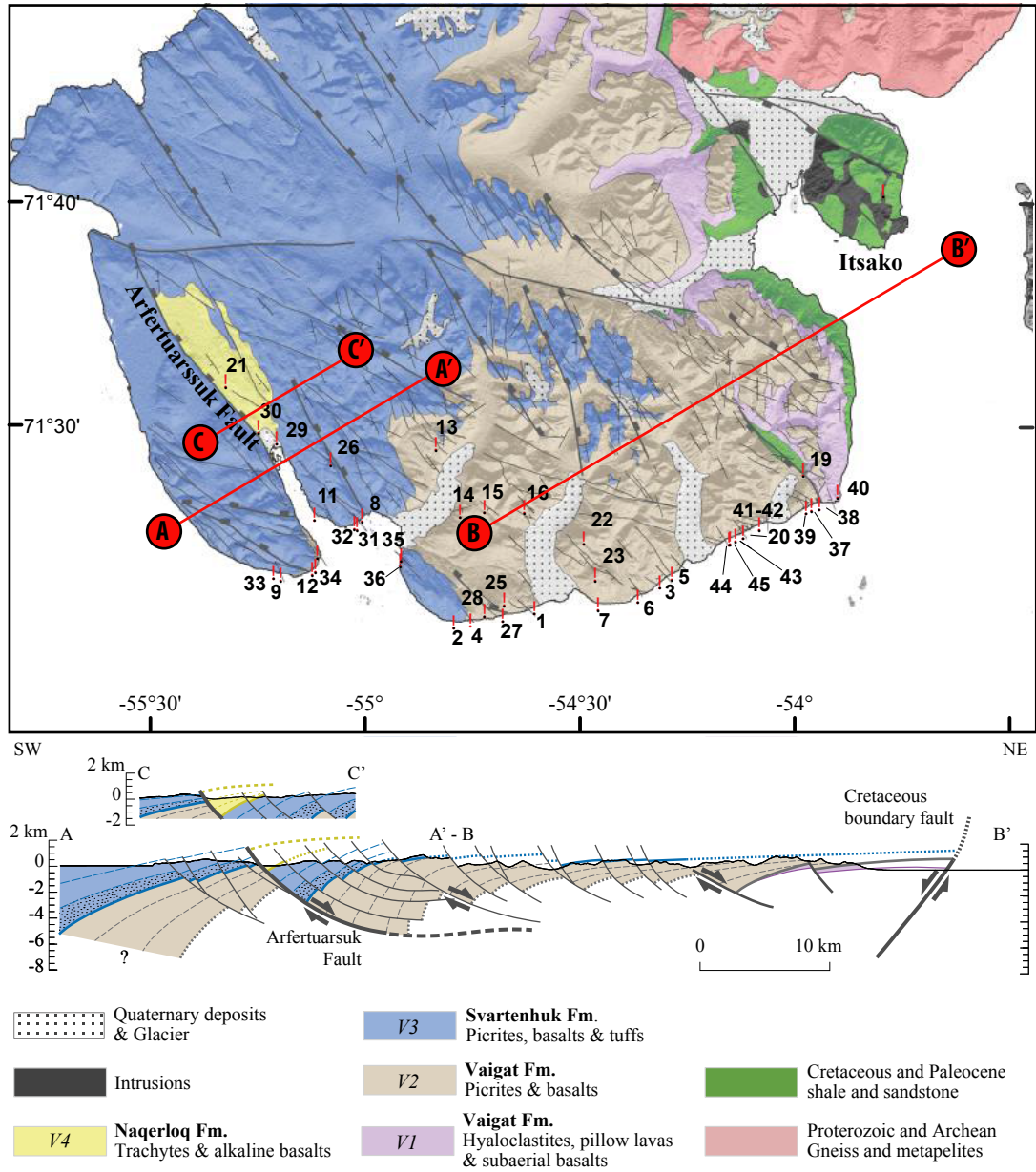
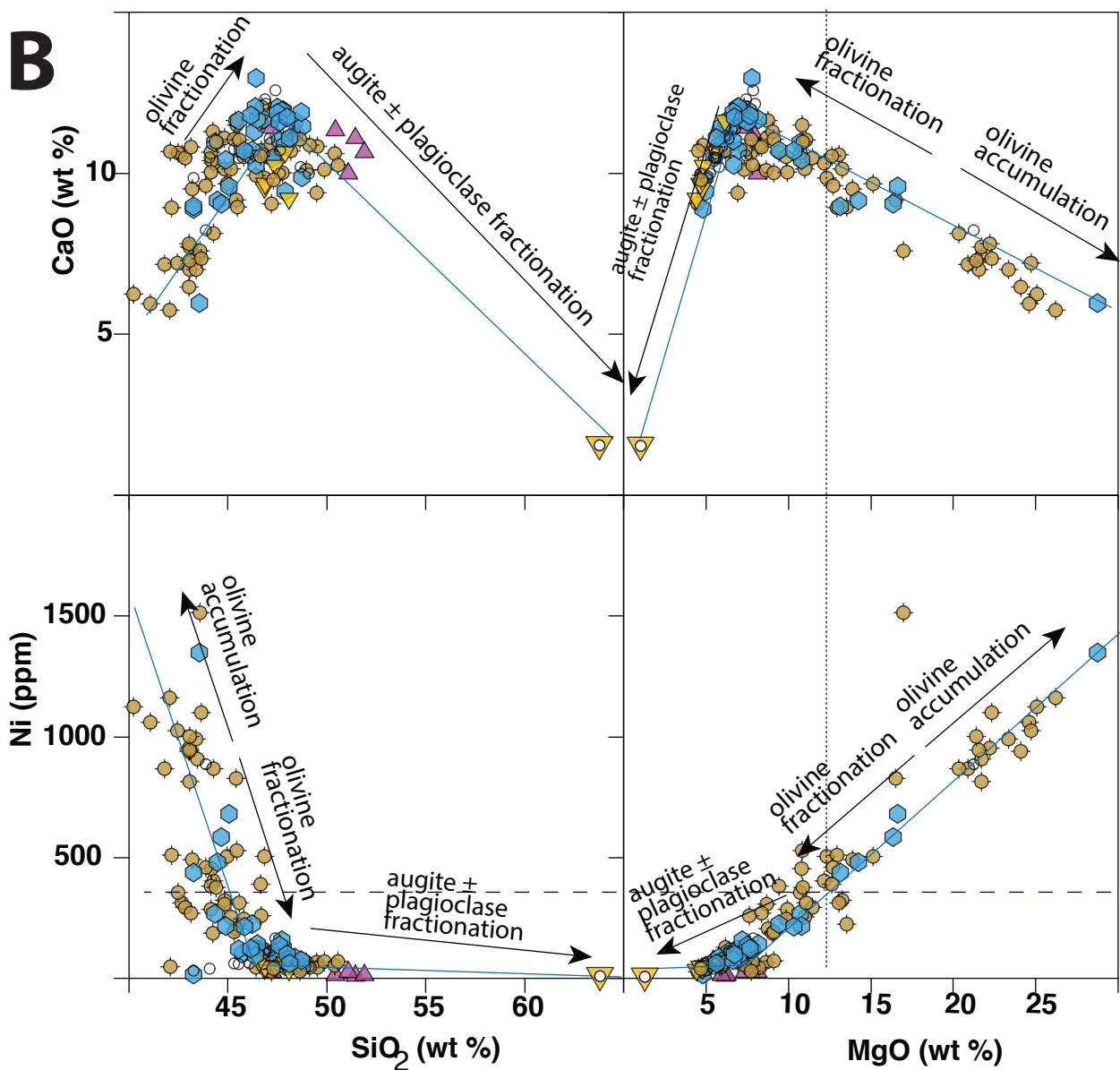
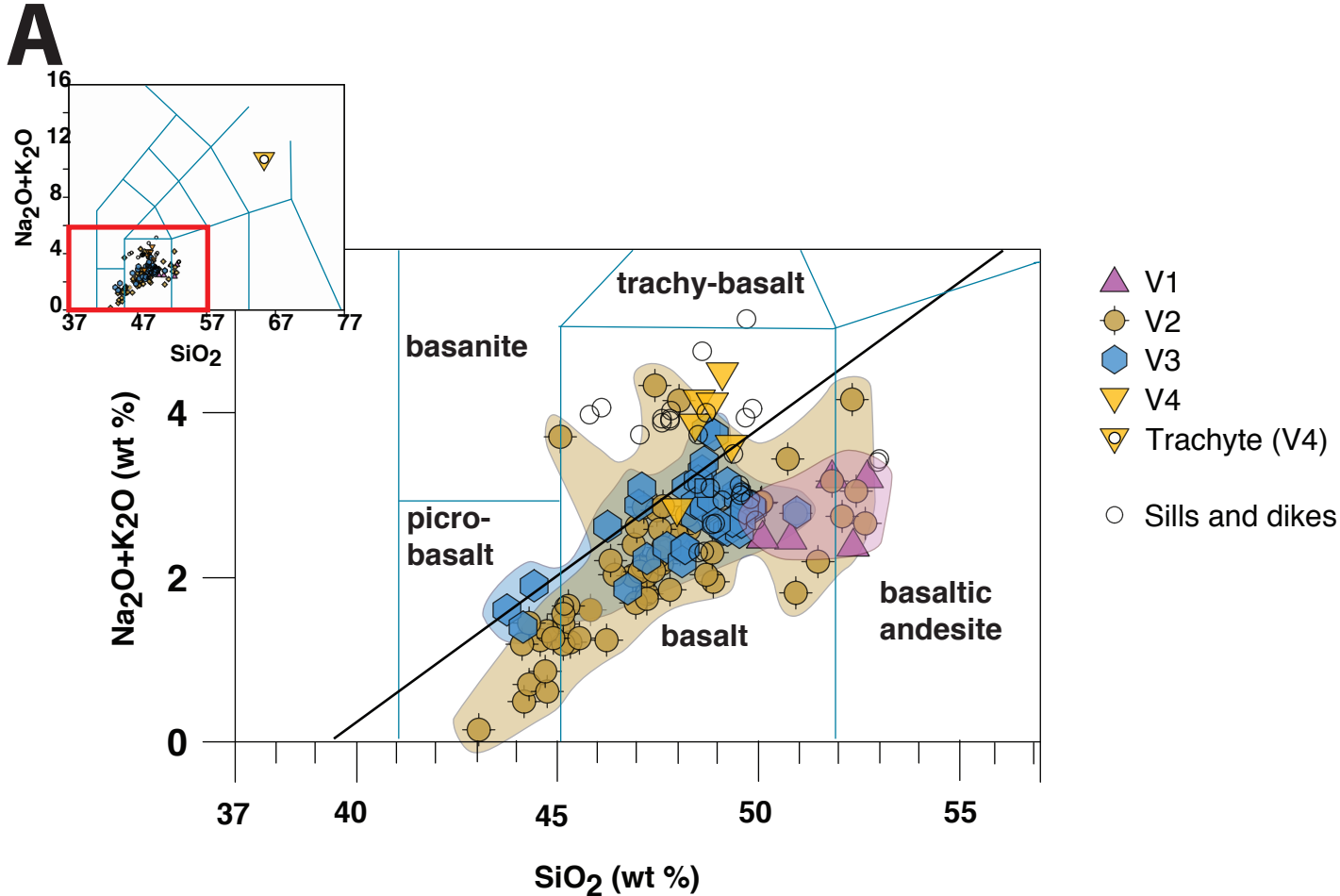
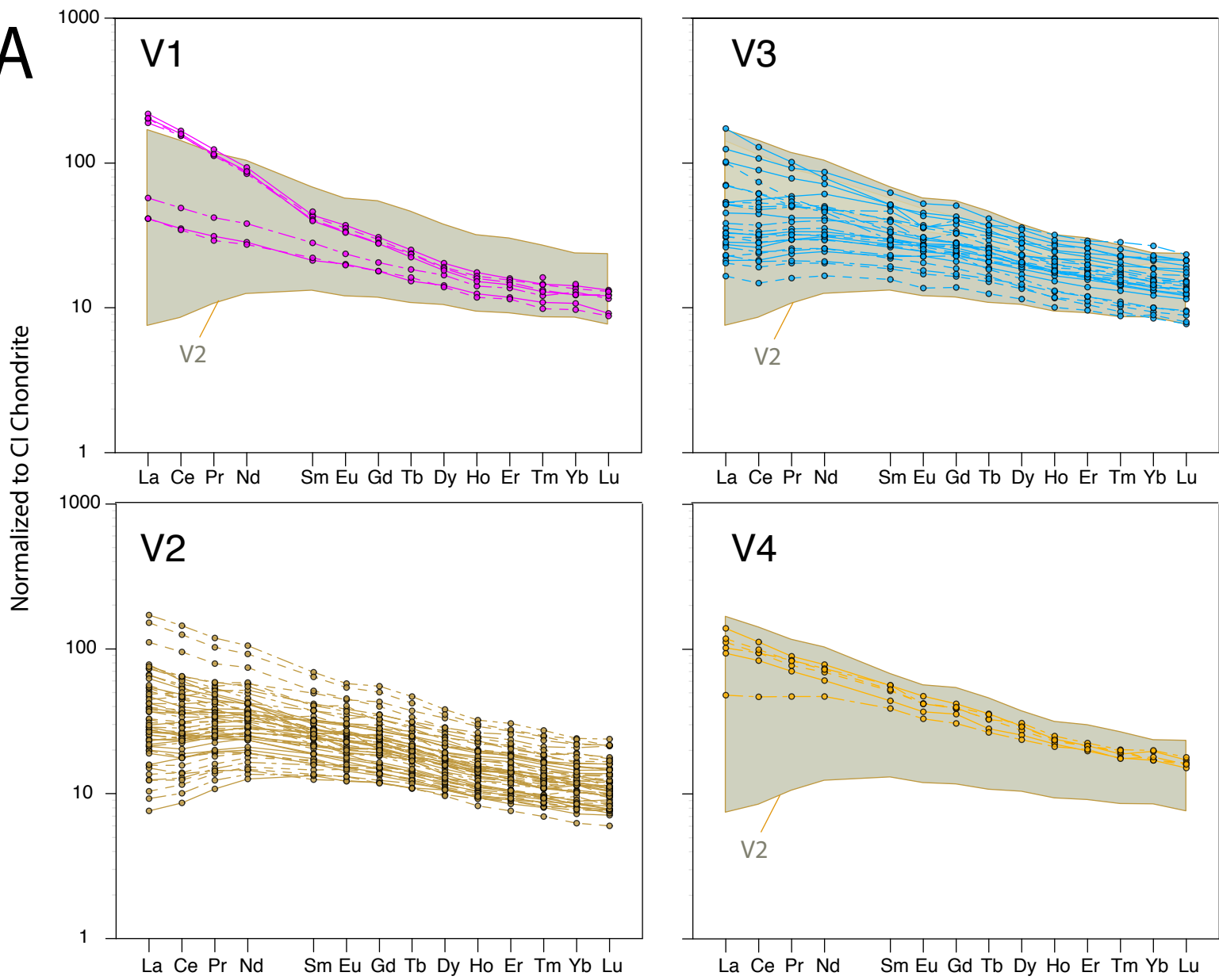
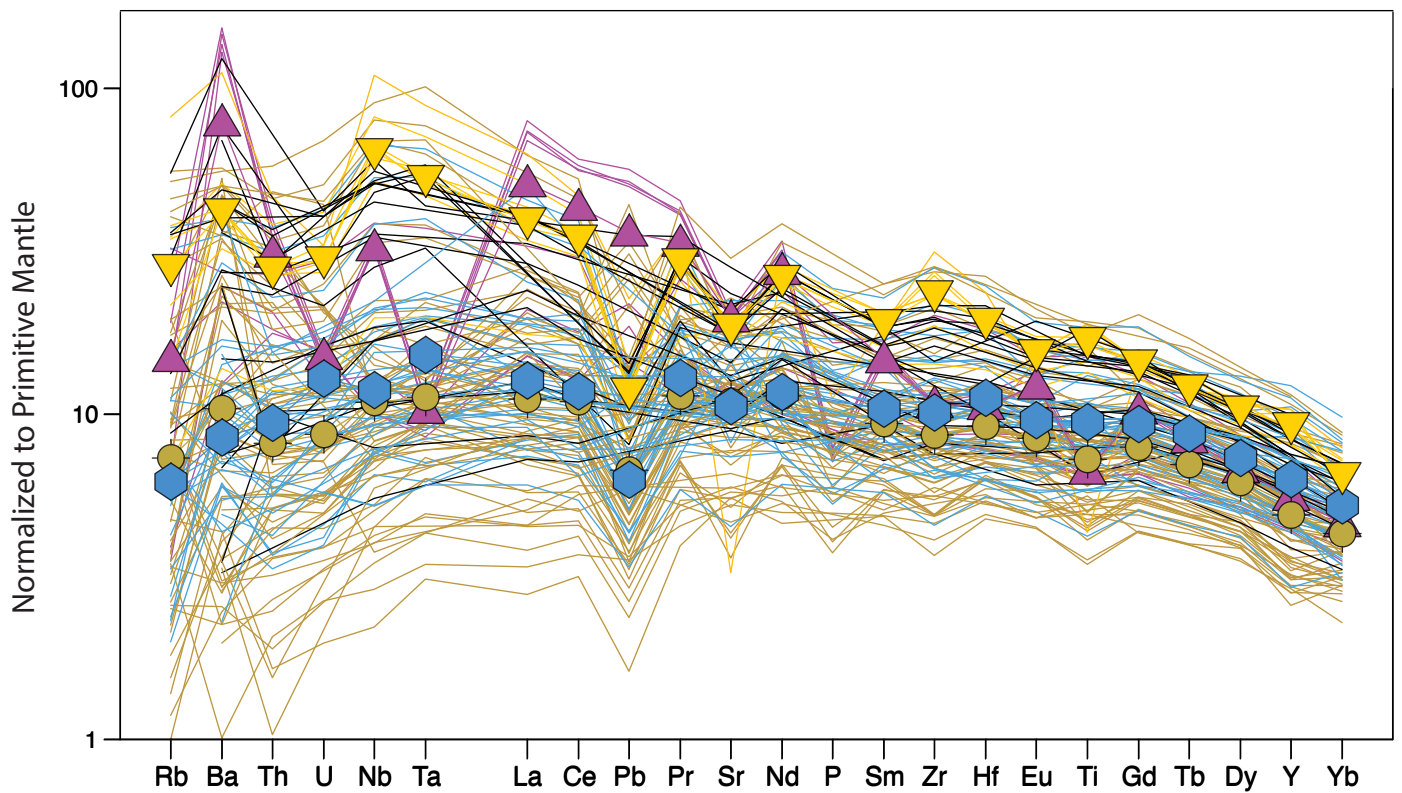
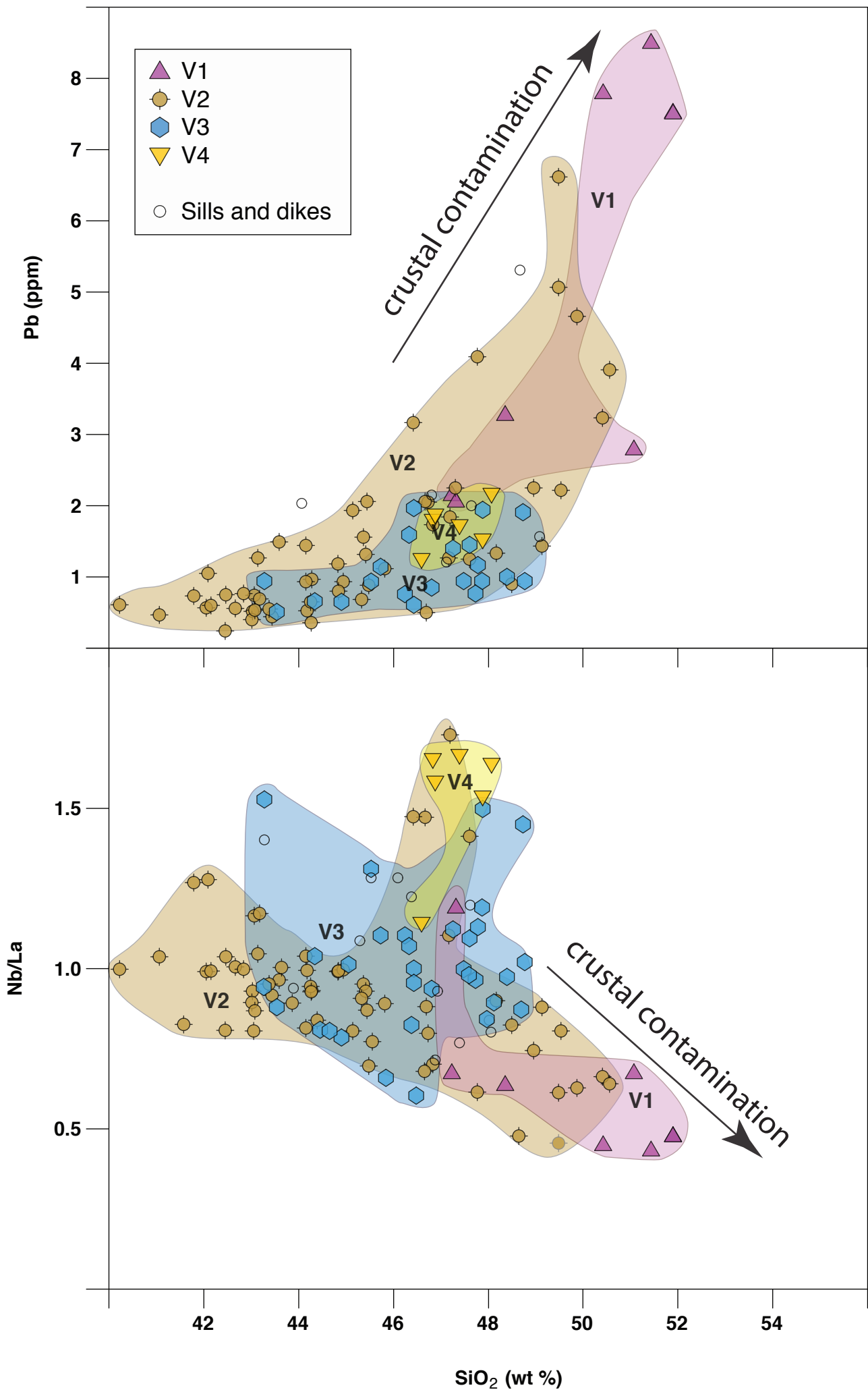


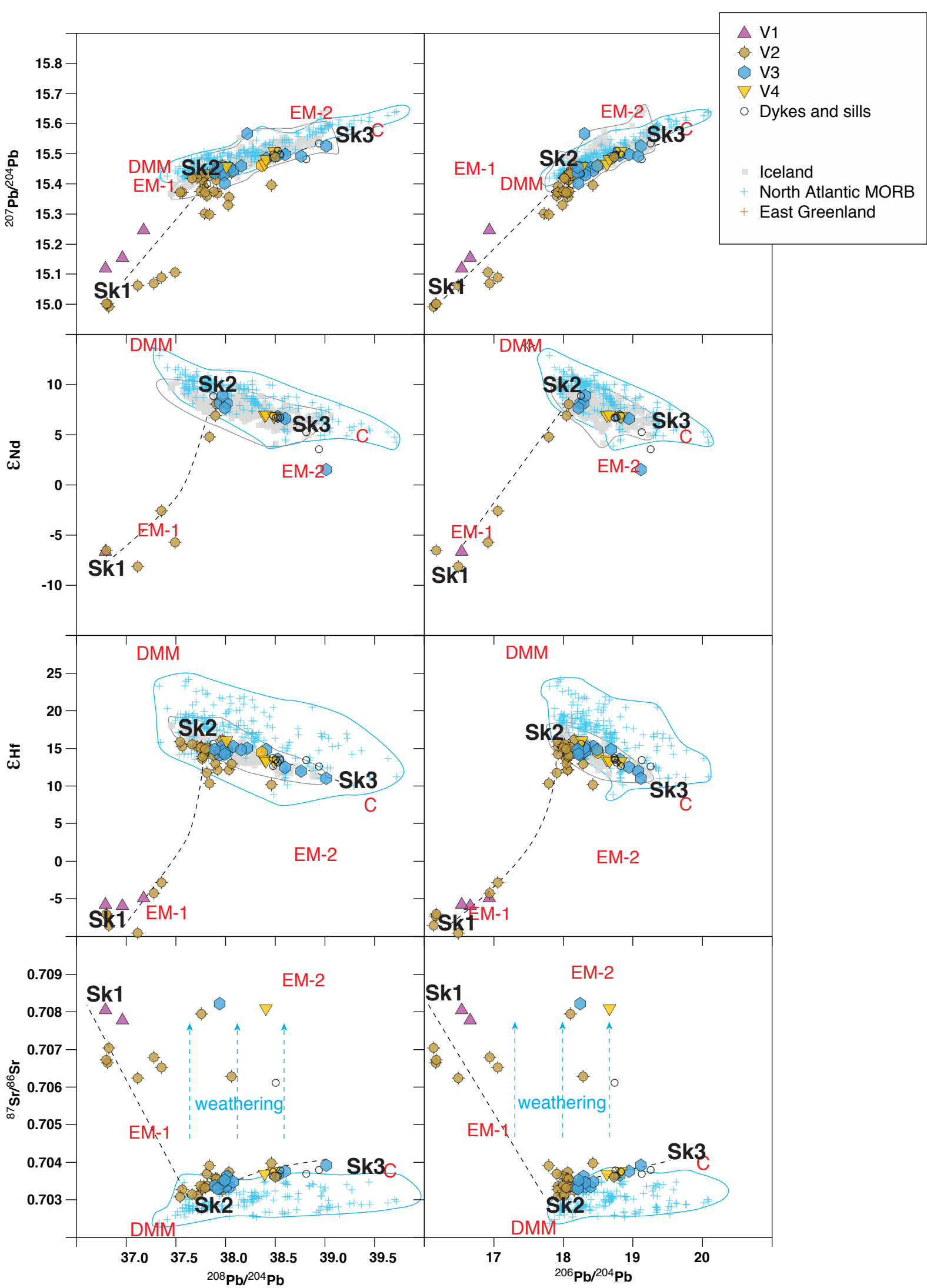
Figure 2

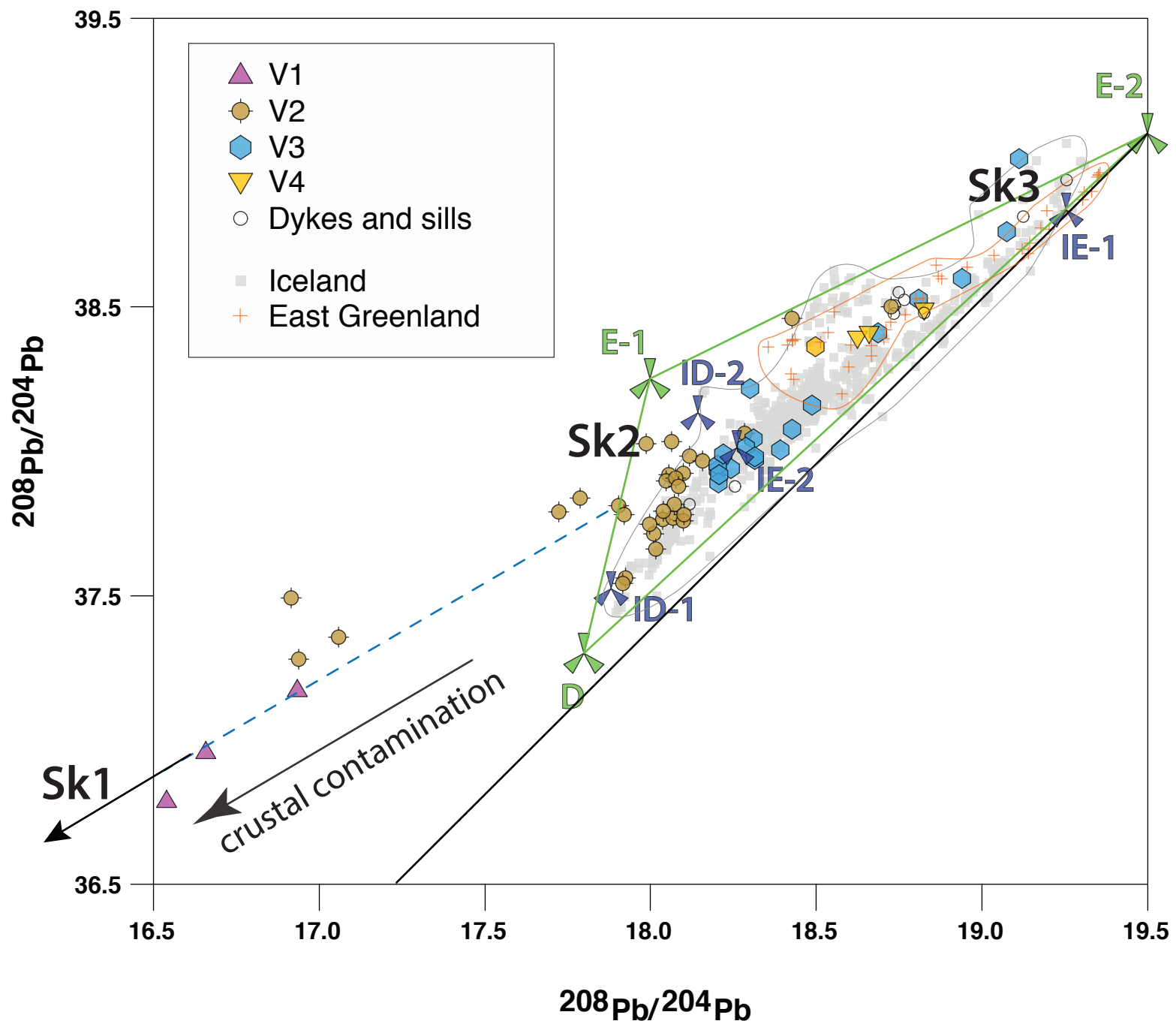


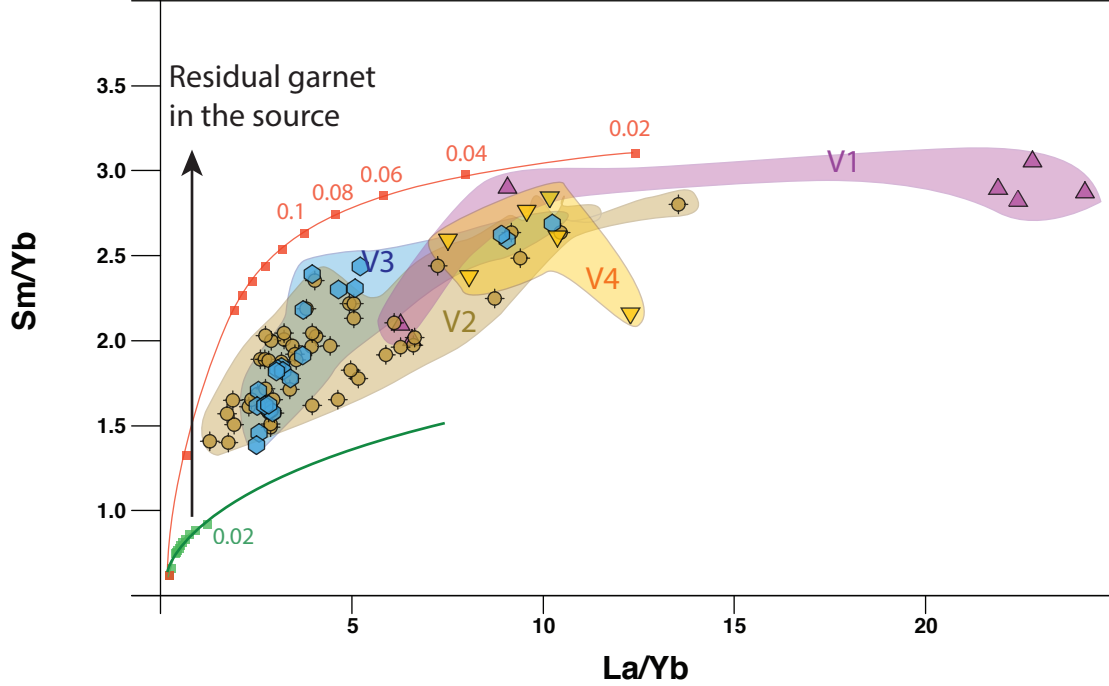
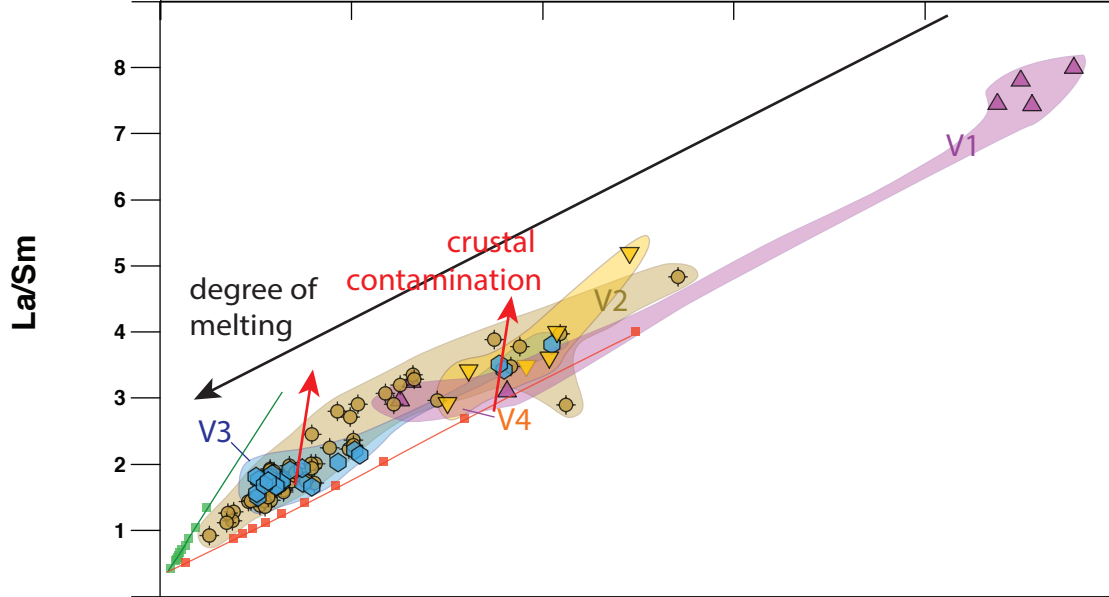
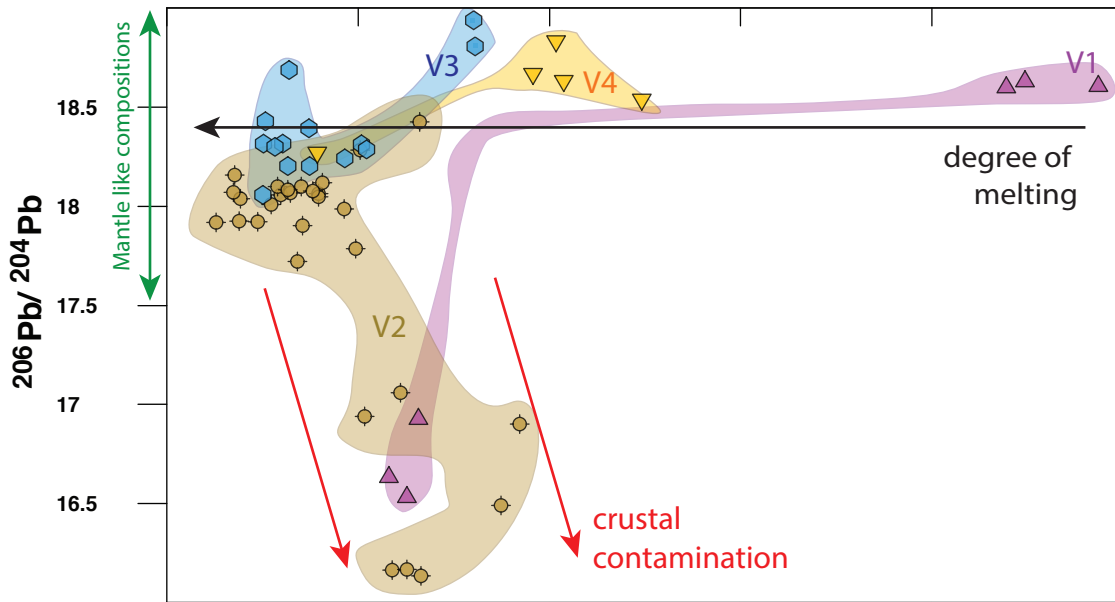


A**B**

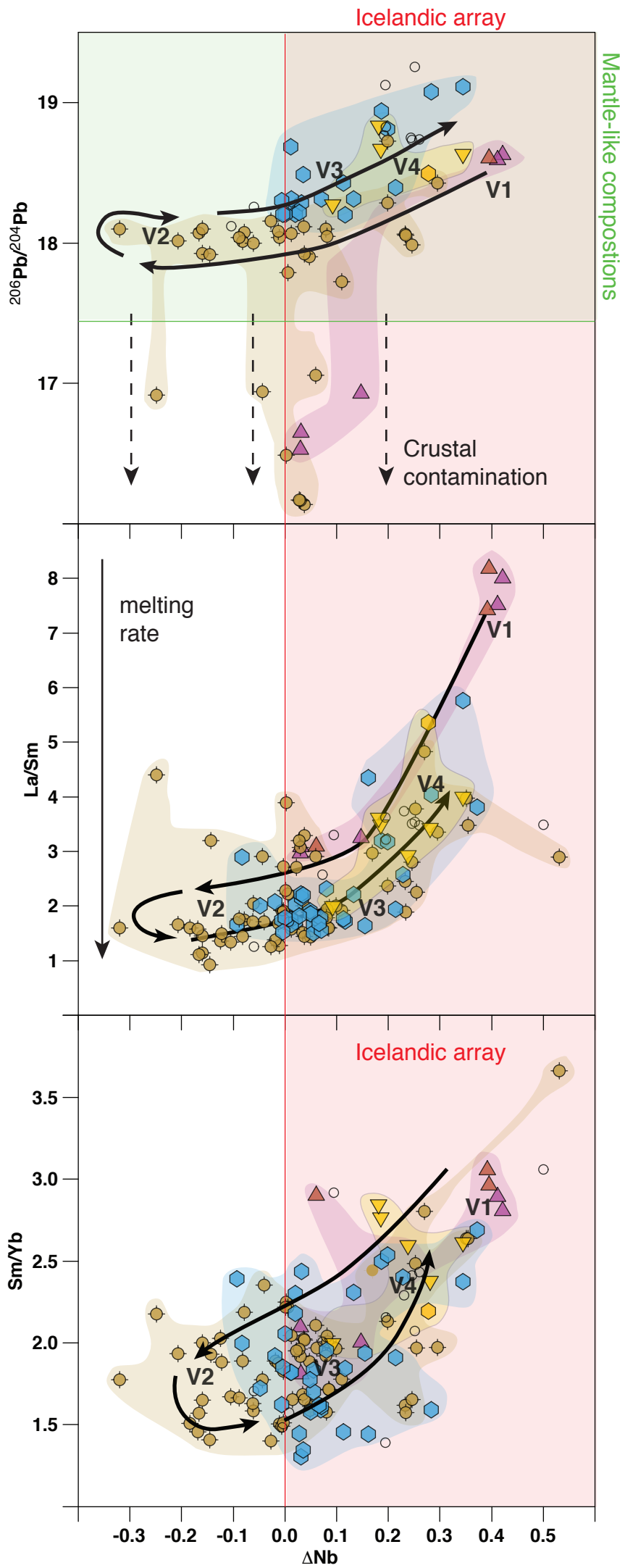


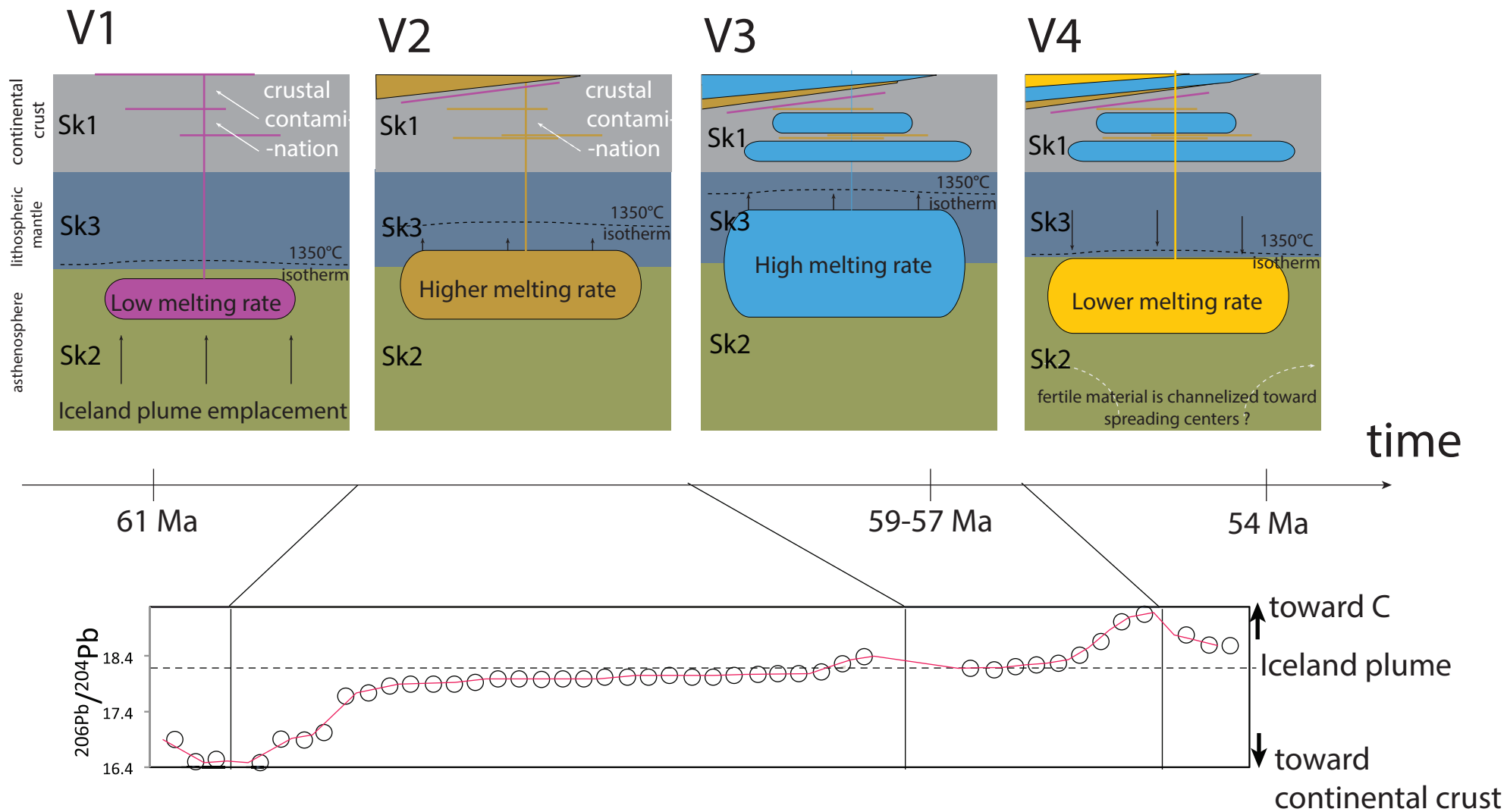


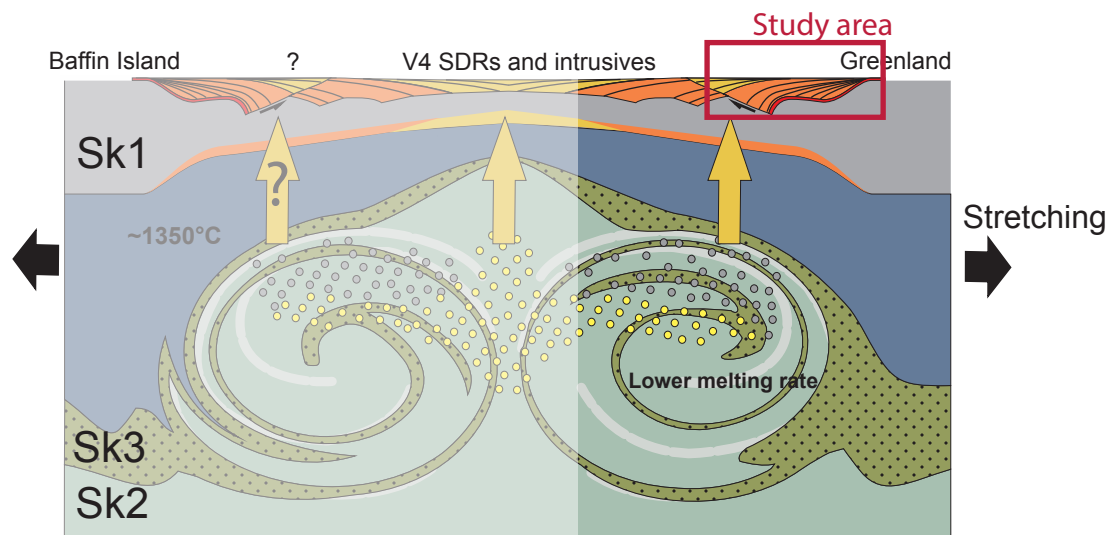
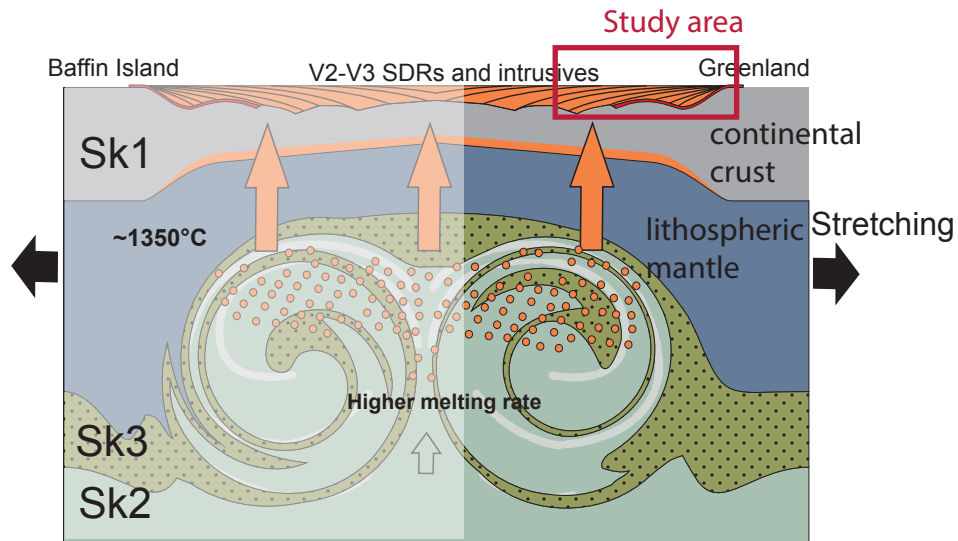
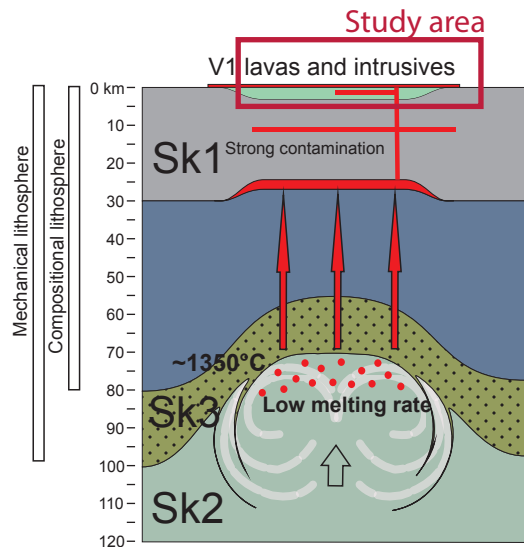




La/Yb







Sample names	longitude	latitude	Fm	³⁹ Ar- ⁴⁰ Ar age K-Ar age (My)
	Lon	lat	Fm	
S19E1	-53.974	71.470	V1	
S19E2	-53.974	71.470	V1	
S38E2	-53.936	71.446	V1	
S38E8	-53.936	71.446	V1	
S40E1	-53.899	71.455	V1	59.11 ± 0.83
S40E2	-53.899	71.455	V1	
S40E4	-53.893	71.453	V1	
S40E5	-53.893	71.453	V1	
S40E6	-54.067	71.426	V1	60.38 ± 1.34
S1E1	-54.605	71.370	V2	
S1E2	-54.605	71.370	V2	
S1E3	-54.605	71.370	V2	
S1E4	-54.606	71.370	V2	
S1E5	-54.604	71.368	V2	
S1E6	-54.607	71.369	V2	
S3E1	-54.300	71.386	V2	
S4E1	-54.679	71.363	V2	
S5E11	-54.282	71.395	V2	
S5E5	-54.288	71.394	V2	
S5E6	-54.287	71.394	V2	
S6E10	-54.359	71.374	V2	
S6E11	-54.367	71.378	V2	
S6E6	-54.364	71.378	V2	
S6E9	-54.366	71.378	V2	
S7E10	-54.458	71.371	V2	
S7E11	-54.457	71.371	V2	
S7E4	-54.461	71.372	V2	
S7E5	-54.461	71.372	V2	
S7E6	-54.456	71.370	V2	
S7E7	-54.459	71.371	V2	
S7E8	-54.455	71.370	V2	
S7E9	-54.459	71.371	V2	
S15H4	-54.719	71.438	V2	
S17H1	-54.218	71.835	V2	
S17H2	-54.224	71.834	V2	
S20I4	-54.117	71.425	V2	
S20E10	-54.109	71.422	V2	
S20E15	-54.118	71.425	V2	
S20E17	-54.119	71.425	V2	
S20E7	-54.112	71.425	V2	
S22H3	-54.486	71.417	V2	
S27E10	-54.738	71.356	V2	
S27E12	-54.740	71.359	V2	
S27E13	-54.741	71.359	V2	
S27E15	-54.751	71.356	V2	
S27E17	-54.751	71.356	V2	
S27E18	-54.752	71.359	V2	
S27E20	-54.752	71.356	V2	
S37E1	-55.204	71.389	V2	
S38E10	-53.942	71.446	V2	
S38E6?	-53.936	71.446	V2	
S39E1	-53.974	71.449	V2	
S41E1	-54.066	71.432	V2	
S41E3	-54.066	71.432	V2	
S42E1	-54.074	71.432	V2	
S42E2	-54.074	71.432	V2	
S42E4	-54.082	71.431	V2	
S42E5	-54.082	71.431	V2	61.08 ± 0.56
S43E1	-54.133	71.423	V2	
S43E2	-54.134	71.423	V2	
S43E4	-54.136	71.423	V2	
S43E5	-54.136	71.423	V2	
S43E6	-54.139	71.422	V2	
S44E10	-54.163	71.419	V2	
S44E2	-54.154	71.420	V2	
S44E4	-54.156	71.420	V2	
S44E7	-54.157	71.420	V2	
S44E8	-54.159	71.420	V2	
S44E9	-54.160	71.419	V2	
S45E10	-54.151	71.420	V2	
S45E4	-54.150	71.421	V2	
S45E9	-54.151	71.420	V2	
S2E2	-54.791	71.358	V3	
S2E3	-54.792	71.358	V3	
S2E5	-54.793	71.358	V3	
S2E7	-54.791	71.355	V3	
S2E8	-54.791	71.355	V3	54.98 ± 1.16
S2E9	-54.796	71.358	V3	
S8E10	-55.006	71.437	V3	
S8E13	-55.013	71.437	V3	55.73 ± 2.49
S8E6	-55.004	71.438	V3	
S8E9	-55.003	71.436	V3	
S9E1	-55.201	71.399	V3	
S11E10	-55.102	71.436	V3	
S11E12	-55.104	71.437	V3	
S11E7	-55.110	71.435	V3	
S26H1	-55.086	71.479	V3	54.20 ± 2.88
S26H3	-55.088	71.479	V3	
S33E1	-55.206	71.393	V3	
S33E12	-55.219	71.393	V3	
S33E4	-55.206	71.393	V3	
S33E7	-55.208	71.393	V3	
S33E8	-55.210	71.393	V3	57.78 ± 1.88
S34E2	-55.118	71.401	V3	
S34E5	-55.117	71.403	V3	
S34E6	-55.120	71.399	V3	
S34E7	-55.120	71.399	V3	
S35E1	-54.920	71.408	V3	
S35E11	-54.916	71.410	V3	
S35E12	-54.916	71.410	V3	
S35E3	-54.919	71.408	V3	
S35E5	-54.918	71.409	V3	59.78 ± 0.41 59.15 ± 0.85
S35E7	-54.917	71.409	V3	
S35E8	-54.917	71.410	V3	
S36E1	-54.922	71.404	V3	
S36E4	-54.922	71.403	V3	
S36E5	-54.922	71.403	V3	
S21H1	-55.338	71.541	V4	55.20 ± 0.79 55.86 ± 1.25
S21H2	-55.340	71.541	V4	
S29E2	-55.218	71.495	V4	
S30E1	-55.249	71.502	V4	54.39 ± 0.60 59.07 ± 1.16
S30E2	-55.236	71.503	V4	
S30E3	-55.244	71.503	V4	57.96 ± 1.20
S30E4	-55.241	71.502	V4	58.73 ± 1.53
S11E9	-55.118	71.435	dike	
S13H1	-54.823	71.489	dike	
S13H5	-54.823	71.489	dike	
S14H1a	-54.773	71.438	dike	
S14H2	-54.773	71.438	dike	56.43 ± 1.06
S15H3	-54.719	71.438	dike	
S16H1	-54.635	71.441	dike	
S16H2	-54.635	71.441	dike	
S16H4	-54.635	71.441	dike	
S20E18	-54.108	71.422	dike	55.42 ± 1.91
S22H2	-54.491	71.423	sill	53.85 ± 2.84
S22H4	-54.486	71.417	dike	54.41 ± 0.99
S23H1	-54.456	71.390	dike	
S23H4	-54.456	71.390	dike V3	58.98 ± 0.93
S23H5	-54.456	71.390	dike	
S24H1	-53.783	71.679	Sill Itsako	55.31 ± 1.09
S27E4	-54.752	71.356	dike	55.95 ± 0.88
S27E8	-54.738	71.356	dike	
S41E2	-54.060	71.426	dike	

