

Origin of basalts by hybridisation in andesite-dominated arcs

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25 **ABSTRACT**

26 Mafic magmas are common in subduction zone settings, yet their high density
27 restricts their ascent to the surface. Once stalled in the crust, these magmas may differentiate,
28 assimilate crust and other melts and mushes to produce hybridised intermediate magmas. The
29 Soufriere Hills Volcano on Montserrat is a ‘type locality’ for these hybridisation processes
30 and yet, just 3 km south of the crater, voluminous basalts have erupted from the South
31 Soufriere Hills volcano within the same time period as the Soufriere Hills Volcano was
32 erupting hybrid andesites (131 - 128 ka). Basaltic South Soufriere Hills magmas have 48 - 53
33 wt% SiO₂ and 4 - 6 wt% MgO. They were hot (970 - 1160 °C), volatile-rich (melt inclusions
34 contain up to 6.2 wt% H₂O) and were stored at 8 – 13 km prior to eruption (based on olivine
35 and pyroxene-hosted melt inclusion volatile geochemistry). Melt inclusions do not preserve
36 basaltic liquids: they are andesitic to rhyolitic in composition, related to one another by a line
37 of descent controlled by simple closed-system fractionation. Whole rock compositions,
38 however, are best described by a hybridisation model involving “back”-mixing of andesitic to
39 rhyolitic melts with mafic crystal phases such as magnetite, olivine, orthopyroxene and
40 clinopyroxene. Phenocryst zoning illustrates repeated mixing events between evolved melts
41 and mafic phenocrysts, which, when coupled with the heterogeneity of crystal compositions,
42 strongly suggests that although the bulk composition is basalt (containing Fo₈₀ olivine), they
43 were assembled from disparate ingredients, likely derived from mafic crystal mushes and
44 more evolved melt lenses of variable composition. The mixing events occur days to weeks
45 prior to eruption. We propose that the South Soufriere Hills basaltic magmas, with their
46 higher bulk density over andesites from neighbouring volcanoes, ultimately may have been
47 eruptible owing to both the transtensional tectonics imposed by offshore grabens (related to
48 the oblique subduction of the Lesser Antilles) and to surface unloading caused by large scale
49 edifice collapse. Our observations support the idea that compositional changes in arcs might

50 reflect not only changes in source compositions, but also effects caused by patterns in crustal
51 strain and tectonics.

52 **INTRODUCTION**

53 Intermediate magmas are generated by intensive crustal magmatic processing
54 involving crystallisation, assimilation and mixing ([Anderson, 1976](#); [Eichelberger, 1978](#);
55 [Rudnick, 1995](#); [Eichelberger et al. 2006](#); [Reubi and Blundy, 2009](#); [Kent et al., 2010](#);
56 [Melekhova et al., 2013](#)). Mafic magmas are implicated in these processes through recharging
57 of magma bodies by mingling at the interface and by large-scale overturn in magma
58 reservoirs ([Pallister et al., 1992](#); [Bateman, 1995](#)). These processes are well-illustrated by
59 volcanoes in the Lesser Antilles arc where andesitic lavas containing mafic enclaves are
60 commonly erupted. Andesites may erupt preferentially due to their relatively low density
61 compared to the denser mafic lavas that are “trapped” at depth by a density filter mechanism
62 ([Plank and Langmuir, 1988](#)). Rheological and lithological barriers may also inhibit the
63 propagation of a basaltic melt ([Eichelberger, 1978](#); [Dufek and Begantz, 2005](#); [Karlstrom et](#)
64 [al., 2009](#); [Kent et al., 2010](#)). Indeed, intermediate to rhyolitic magma reservoirs can obstruct
65 the passage of mafic magma, explaining why basaltic eruptions often only reach the surface
66 on the periphery of silicic volcanoes ([Hildreth, 1981](#)). An interesting variant on this process is
67 illustrated on Montserrat, where basalts were erupted from the South Soufrière Hills (SSH)
68 volcano over the same broad time interval as crystal-rich andesites (with rhyolitic melts) were
69 being erupted from the Soufrière Hills Volcano (SHV) located less than 3 km away. This
70 raises the question as to what mechanisms allow eruption of felsic and mafic volcanic rocks
71 in such close proximity.

72 More detailed study of the SSH is also of interest because, while there is strong
73 evidence that andesites are generated largely by mixing of repeated injections of mafic
74 magma into high level silicic magma chambers ([Anderson, 1976](#); [Eichelberger, 1978](#);

75 [Eichelberger et al. 2006](#); [Reubi and Blundy, 2009](#); [Kent et al., 2010](#)), the petrogenesis and
76 history of the mafic magmas is not well understood and may itself be complex. The density
77 filter trap ([Plank and Langmuir, 1988](#)) means that mafic enclaves from SHV are the only
78 evidence of deeper, mafic magmas that are available for petrologic analysis and in many
79 cases these mafic inclusions have experienced varying degrees of intrusion, quenching and
80 degassing that obscures their earlier characteristics. Thus, a study of closely spaced (in
81 distance and time) andesitic and basaltic volcanism at SHV and SSH has the potential to
82 reveal more detail regarding the nature of basaltic magmas resident in the mid- to upper-
83 crust, and can provide insights into the relative importance of magma mixing and
84 fractionation in controlling the composition of all arc volcanic rocks, and how this relates to
85 processes of magma storage, hybridisation, eruption triggering and growth of the arc crust.

86 In this paper we present new whole rock and melt inclusion analyses of basaltic to
87 andesitic lavas erupted from the SHV. We compare their geochemical characteristics to the
88 andesites erupted from SHV and examine the geochemistry of individual phenocrysts phases
89 to characterise compositional gradients related to normal crystal growth during cooling and
90 also due to mixing. We assess whether their compositions could have been generated by
91 simple processes of fractional crystallisation alone or whether mixing between disparate
92 liquid and mush components is necessary. Mineral melt thermometry has been used (from
93 two-pyroxenes and plagioclase-glass pairs) and barometry (using H₂O-CO₂ systematics of the
94 melt inclusions) to estimate pre-eruptive storage conditions. We use the relaxed
95 compositional steps across olivine crystals to infer pre-eruptive mixing timescales between
96 felsic liquids and mafic crystals. Using all of the available petrological and geochemical data
97 we develop a model for the generation of hybrid basalts on Montserrat and how they are
98 assembled and speculate as to the possible reasons for extraction and eruption of higher
99 density hybrid magmas relating to tectonics and unloading.

100

101 **Geological background**

102 The Lesser Antilles, like many arcs, comprises predominantly andesitic volcanic
103 islands with relatively few basaltic centres. For example, in the northern and central islands
104 (Saba to St. Lucia) <10% of the erupted volcanic rocks are basaltic. Where basaltic rocks are
105 present, they generally occur as small-volume centres adjacent to much larger andesitic
106 volcanoes ([Westercamp and Mervoyer, 1976](#); [Rea and Baker, 1980](#); [Macdonald et al. 2000](#)).
107 This is exemplified on Montserrat ([Fig. 1](#)), where andesite lavas are predominant ([Rea,](#)
108 [1974](#)), with a single isolated basaltic centre (SSH) in the southernmost part of the island.
109 Apart from the SSH, basalt occurrences are restricted to mafic inclusions within andesites.
110 There is abundant petrological evidence (particularly from the currently active SHV) to show
111 that the erupted andesites are hybrids formed over long timescales (10^3 to 10^4 years) by
112 multiple recharges of deeply-sourced mafic magmas into large reservoirs of crystal-rich
113 andesite magmas prior to ascent to the surface ([Murphy et al., 2000](#); [Humphreys et al., 2009](#);
114 [Plail et al., 2014](#)).

115 The SSH basalts are, however, sufficiently geochemically distinct from the SHV
116 basaltic enclaves suggesting that they reflect different magma sources and processes, such as
117 increased relative contribution from slab fluids over subducted sediments ([Zellmer et al.,](#)
118 [2003](#); [Cassidy et al., 2012](#); [2014](#)), and thus provide information regarding magmas forming
119 within the arc that are not generally observed, at least in an identifiable form, at the surface.
120 Indeed, the SSH volcanic rocks represent some of the most mafic lavas in the northern Lesser
121 Antilles arc (47 wt% SiO₂; 6 wt% MgO), with the exception of the high-Mg basalts in
122 Martinique ([Westercamp and Mervoyer, 1976](#)). These geochemical differences are not
123 simply related to temporal evolution of the volcanism on Montserrat, because Ar-Ar dating
124 and stratigraphic relationships clearly indicate that the SSH and the SHV were both active in

125 the interval 130 ± 5 ka (with SHV-type rocks forming the basal unit to the main SSH
126 lithologies), and the predominant andesitic volcanic rocks of the island were emplaced before
127 and after eruption of the SSH (Harford et al., 2002; Cassidy et al., 2012).

128 The island of Montserrat is located in the northern part of the Lesser Antilles; a 750
129 km long chain of volcanic islands formed as a result of the slow (2 cm yr^{-1}) subduction of the
130 North American plate beneath the Caribbean plate (Fig. 1) (Wadge, 1984; DeMets et al.,
131 2000). The oblique nature of this subduction means that the northern part of the arc is
132 influenced by transtensional forces that have led to intra-plate deformation (Feuillet, 2000;
133 Feuillet et al., 2010). Montserrat lies on crust ~ 30 km thick that sits on an asthenospheric
134 mantle wedge that extends to ~ 130 km in depth (Wadge and Shepherd, 1984). The island
135 comprises four volcanic centres: Silver Hills (2600-1200 ka), Centre Hills (950-550 ka), SHV
136 (282 ka to present) and SSH (131-128 ka) (Harford et al., 2002). All these volcanic centres
137 (except for the mafic-dominated SSH) are andesitic in composition, but their erupted
138 products all contain abundant inclusions of mafic magma (Rea, 1974; Murphy et al., 2000;
139 Zellmer et al., 2003; Barclay et al., 2010; Plail et al., 2014).

140 The SHV centre has been studied in most detail and is comprised of phenocrysts of
141 orthopyroxene, plagioclase and amphibole in a rhyolitic glass, with clear evidence for magma
142 mixing and mingling (Murphy et al., 2000; Humphreys et al., 2009; Humphreys et al., 2013).
143 Under-plating of the crystal-rich andesite by wet mafic magma causes instabilities to form at
144 the interface, forming enclaves (Plail et al., 2014; Edmonds et al., 2014), interspersed with
145 sporadic magma overturn events that thoroughly mix the magmas (Woods and Cowan, 2009),
146 distributing widely dispersed mafic components (Fe-rich plagioclase microlites, K-rich glass;
147 Humphreys et al., 2010) into the andesite body. The petrography and geochemistry of the
148 mafic enclaves of the SHV is also best explained by a mixing process between a mafic end
149 member (which varies in composition with time owing to lower crustal cryptic amphibole

150 fractionation) and variable amounts of rhyolitic melt hosting up to 20 vol% phenocrysts of
151 plagioclase, amphibole and magnetite, although not in bulk rock proportions.

152 While there have been a large number of studies on the andesites of Montserrat,
153 petrological work on the SSH basalts is more limited. [Murphy et al. \(2000\)](#) report that the
154 mineral assemblage consists of plagioclase, olivine, clinopyroxene, and titanomagnetite. The
155 SSH exposures comprise a range of rock suites from lava flows, to scoria, to reworked
156 volcanoclastic material ([Cassidy et al., 2014](#)), with some more mafic enclaves and some lava
157 flows containing cumulate xenoliths of orthopyroxene and plagioclase, similar to those
158 described by [Kiddle et al. \(2010\)](#). The SSH exposures can be divided into two units on the
159 basis of their distinct trace element and isotopic compositions: SSH Suite A has lower Sr/La
160 and Sm/Zr ratios, but higher Zr/Er ratios and more radiogenic Pb isotope compositions than
161 Suite B ([Cassidy et al., 2014](#))

162

163 **METHODS**

164 *Samples*

165 Samples of SSH rocks were collected along the south coast of Montserrat ([Fig. 1;](#)
166 [Table 1](#)). Splits were crushed using an agate Mortar and powdered for whole rock analysis
167 and thin sections were also cut for electron microprobe (EMPA) and scanning electron
168 microscope (SEM) analysis. Fractions of samples were crushed coarsely and crystals of
169 enstatite, augite and olivine were picked from the 125-250 μm grain size fraction. The
170 crystals were ground and polished to expose melt inclusions and mounted in indium for
171 secondary ion mass spectrometry (SIMS) analysis. All the inclusions analysed were natural
172 quenched, 40-200 μm in size and were not necked or breached by cracks.

173

174 *Whole rock analysis*

175 Major elements were analysed by X-ray Fluorescence (XRF) analysis of glass beads
176 prepared by fusion of a mixture of 0.5 g subsamples and lithium tetraborate in a ratio of 1:10.
177 Analyses were undertaken using a Philips Magix Pro WD-XRF at the National Oceanography
178 Centre (NOC), Southampton, UK. Error and external accuracy was generally <2%.

179

180 *Microanalysis (EMPA, SEM and SIMS)*

181 Concentrations of H₂O and CO₂ in glass were obtained by SIMS on a Cameca IMF 4f
182 ion microprobe at the NERC microanalytical facility at the University of Edinburgh, using a
183 15kV primary beam of O⁻ ions (Hauri et al., 2002; Blundy and Cashman, 2008). Positive
184 secondary ions were accelerated to 4500 eV, with an offset of -75eV (for ¹H and trace
185 elements) and -50eV (for ¹²C) (± 20eV) to reduce transfer of molecular ions. A 50 µm raster
186 was performed for three minutes prior to the start of each analysis, and a primary beam
187 current of 5-6 nA used with a non-rastered, oval-shaped beam covering a 15-20 µm area on
188 single spots within the boundaries of the melt inclusions. Peak positions were verified before
189 each analysis. The following elements were analysed by counting for 3 s in each of a 10
190 cycle run: ¹H, ²⁵Mg, ³⁰Si. These counts were then normalised to ³⁰Si and converted to
191 concentrations using a calibration curve populated by glass standards. The relative ion yield
192 for H correlates with SiO₂ content, such that plotting ¹H/³⁰Si versus H₂O yields a single
193 working curve for glasses of variable SiO₂ content. CO₂ concentrations, however, require a
194 correction for SiO₂ content.

195 Carbon was measured independently of ¹H, using the same beam conditions, but with
196 a 50 µm image field to improve transmission at moderate mass resolution, which was
197 sufficient to resolve ²⁴Mg²⁺ at the ¹²C peak position for background olivine measurements
198 and inclusion analyses. ¹²C was analysed for 3 s in each of 20 cycle runs in which ²⁴Mg²⁺,
199 ²⁸Si²⁺ and ³⁰Si were also measured. During data processing, the first 5 cycles of the ¹H

200 analyses and the first 10 cycles of the ^{12}C data were discarded to avoid the effects of surface
201 contamination on the samples which may have survived the cleaning process. Instrumental
202 backgrounds were minimized by allowing samples held in epoxy to outgas in a separate
203 vacuum for at least ten hours prior to use in the SIMS instrument. The full list of glass
204 standards used is shown in [suppl. Table 1](#). The accuracy and precision were monitored
205 throughout the sessions by repeat analysis of the standards as unknowns: for H_2O analyses
206 these were <9% and <6% respectively; and for CO_2 <11% and <8% respectively. The
207 average CO_2 and H_2O backgrounds over seven sessions were 56 ppm and 0.03 wt%
208 respectively. There is lack of variation between Al_2O_3 and MgO in melt inclusions
209 compositions, suggests that they do not follow the vectors anticipated for post-entrapment
210 crystallisation of the host mineral.

211 The major element and volatile (S, Cl and F) compositions of the glasses, inclusions
212 and phenocrysts were determined using the Cameca SX100 electron microprobe at the
213 University of Cambridge. Quantitative determinations of elements were made using the
214 wavelength dispersive system with TAP, PET and LIF crystals. A range of metal, oxide and
215 silicate (e.g. jadeite, wollastonite) standards was used for calibration of the spectrometers.
216 All analyses used an accelerating voltage of 15kV. For olivine, pyroxene and plagioclase a
217 spot size of 4 μm and a 100 nA beam current was used. For glasses, a 10 μm spot was used
218 with a beam current of 60 nA for Cl, F, S, P, Cr and Ni, and 4 nA for all other elements, with
219 counting times of 50-200 s per analysis. During glass measurements, Na peaks were counted
220 first to avoid significant migration during the run. In addition to calibration of each X-ray
221 line, a series of secondary reference standards (olivines, pyroxenes, feldspars and glasses)
222 were measured daily to check accuracy, precision and totals. Standards used were periclase
223 for Mg, jadeite for Na, fused Si for Si, rutile for Ti, fayalite for Fe, K-feldspar for K
224 corundum for Al, apatite for P, and pure metals for Cr and Mn. Repeat analyses of standards

225 were used to estimate the precision of An, Mg# and Fo measurements. Forsterite content of
226 the St. John's Island Olivine standard was determined with a precision of $2\sigma=0.46$ mol %
227 (n=33). Precision of Mg# of clinopyroxene was similar to the precision of forsterite content
228 in olivine. Anorthite content in the Anorthite55 standard was determined with a precision of
229 $2\sigma=1.01$ mol % (n=46).

230 Accuracy was generally better than 5% for most elements, based on repeat analyses of
231 EMPA secondary standard 2390-5 and by comparison with reference concentrations for the
232 standard, with the exception of TiO₂, K₂O, P₂O₅ and Cl, which were better than 20-35 %.
233 Detection limits for S, Cl and F were 40, 38 and 170 ppm, respectively, and precision was
234 typically < 5% for all oxides, with the exception of MnO, P₂O₅ and F, which was better than
235 20%.

236 Backscattered SEM images were taken at the NOC, using a LEO 1450VP (variable
237 pressure) SEM. Carbon-coated samples were imaged at 15 kV, a working distance of 10 mm
238 and a nominal probe current of 50–500 pA, using both secondary electron (SE) and
239 backscattered electron (BSE) detectors.

240

241 **RESULTS**

242 The whole rock samples are black to grey in colour, and poorly to moderately
243 vesicular (6-38%, average 20%). The SSH samples have bulk rock compositions ranging
244 from basalt to andesite (47-58% SiO₂) (Table 1; Fig. 2). Also shown in Figure 2 are the
245 compositions of andesites and mafic enclaves erupted from the SHV during 1995-2010,
246 together with previously published data from SSH (Murphy et al., 1998, 2000; Horwell et al.,
247 2001; Zellmer et al., 2003; Humphreys et al., 2009, 2010; Cassidy et al., 2012). Relative to
248 the SSH, the SHV volcanic rocks are more silicic, ranging from basaltic andesite to dacite

249 (53-68% SiO₂), but the SHV contain mafic enclaves that range from basaltic to basaltic
250 andesite (49-55% SiO₂).

251 The SSH lavas are highly crystalline, with 31-53 vol.% phenocrysts and
252 microphenocrysts (>100 μ m) and 47-69% microlites. Plagioclase is the most abundant
253 crystal phase (up to 61 vol.% of the crystal assemblage), followed by orthopyroxene (15
254 vol.%), olivine (11 vol.%) and clinopyroxene (10 vol.%), with titanomagnetite and rare
255 amphibole in the basaltic andesite samples (SSH5B) comprising the remaining 3 vol.% (Fig.
256 3). The microlite crystal size fraction comprises a similar assemblage, however with less
257 olivine present.

258 *Olivine petrography*

259 On average, olivines form the largest crystals (mean size 390 μ m; range \pm 100 μ m)
260 and are often euhedral to subhedral. They are commonly fractured and slightly altered
261 (slightly reddened along cracks, visible in plane polarised light). The forsterite contents
262 (molar Fo% = Mg/(Mg+Fe) x 100) range from 56 to 80 mol. % (Figs. 4 and 5), with two
263 main peaks in olivine core compositions (Fo₇₂₋₈₀ in Group 1; Fo₅₆₋₆₈ in Group 2) and two
264 peaks in olivine rim compositions that are slightly less forsteritic than the cores. Most of the
265 olivines are normally zoned or unzoned, but some exhibit reverse zoning (Fig. 6; Fig. 7),
266 suggesting multiple magma bodies which have experienced mixing. The reverse-zoned
267 olivines have core compositions of Fo₇₁₋₈₀, compared to Fo₅₆₋₈₀ in the normally-zoned
268 olivines (Figs. 5 and 6c). There is a negative correlation between olivine forsterite contents
269 and CaO and MnO concentrations, with generally higher Fo% and lower Ca and Mn contents
270 in the cores (Fig. 5). These correlations are significant at >95% confidence, with P-values
271 <0.05. This correlation is especially strong between MnO and Fo% with a R² value of 0.9,
272 but this correlation is less apparent with CaO and Fo% (R² of 0.36). The normally-zoned
273 crystals show a trend of increasing Fo% from rim to core, mirrored by decreasing CaO and

274 MnO profiles (Figs. 6a and 6b). Figure 6d illustrates an olivine with reverse zoning towards
275 the outer edge of the crystal, with a thin (<20 μm) band of normal zoning at the rim and no
276 visible overgrowth. The core of this crystal has a constant forsterite composition of Fo₇₂,
277 except for the outer 50 μm . The increase in forsterite content in the reverse zone is
278 positively correlated with CaO, but negatively correlated with MnO content.

279 *Plagioclase petrography*

280 Plagioclase crystals range in size from microlites (<15 μm) to phenocrysts (>500
281 μm), with the latter commonly showing both normal and oscillatory zoning, as well as sieve
282 textures (Fig. 3). Anorthite contents (An mol.% = Ca/(Ca+Na) x 100) range from 49-97%
283 (Fig. 4). The feldspars are commonly normally-zoned, but with rare reverse-zoned
284 phenocrysts also present (Fig. 7), suggesting a complex set of magmatic processes have
285 occurred. The plagioclase crystals can be separated into two main groups based on their
286 anorthite compositions. The cores and reverse-zoned rims are anorthitic (An₇₉₋₉₇), while the
287 rims of the normally-zoned plagioclase are more albitic (An₅₂₋₇₀) and are generally richer in
288 MgO, FeO and TiO₂ than the more anorthitic cores and rims (Fig.8). Complex dissolution
289 and resorption is also seen in some crystals (Fig. 8b).

290 *Pyroxene petrography*

291 The average size of the orthopyroxene crystals is 142 $\mu\text{m} \pm 100 \mu\text{m}$. They are
292 commonly zoned and often occur as overgrowths on olivine (e.g. Fig. 6c). Magnesium
293 number (Mg# = Mg/(Mg+Fe) x 100) ranges from 60-74 (Figs. 4 and 9), and all are enstatite
294 in composition. Enstatite TiO₂ and Al₂O₃ contents generally decrease with decreasing Mg#
295 (Fig. 9), but do not correlate significantly with Al/Ti ratios. The enstatite shows common
296 reverse zoning and some normal zoning, but rare unzoned crystals are also present (Fig. 7).

297 Clinopyroxenes have an average crystal size of 176 $\mu\text{m} \pm 100 \mu\text{m}$, with Mg#
298 ranging from 58-80. The majority of the clinopyroxenes are augite, but some cores are

309 diopside . The augites are commonly zoned, but rare unzoned crystals also exist. Some of
300 the clinopyroxene occurs as pigeonite overgrowths on the olivines (Figs. 6 and 10). Plots of
301 Mg# versus minor elements (Fig. 9) show that the clinopyroxenes contain higher
302 concentrations of TiO₂, Al₂O₃ and Al/Ti ratios than the enstatites. A traverse of a normally-
303 zoned crystal shows complex saw tooth zoning (Fig. 10b) that is particularly oscillatory in the
304 last 70 μm toward the rim, which occurs along with a sharp increase in Al₂O₃ and TiO₂ and a
305 decrease in both Mg# and Al/Ti ratios.

306 *Melt inclusion geochemistry*

307 The melt inclusions are pristine, up to 90 μm in diameter, with no vapour bubbles and
308 no daughter crystal phases. They span a range in compositions from andesitic to rhyolitic,
309 with 58.2-72.6 wt.% SiO₂, 0.45-2.6 wt.% K₂O and 0.01-2.8 wt.% MgO (Fig. 2; Table 2).
310 Their H₂O contents range from 1.50-6.19 wt.%, with CO₂ contents of 20-313 ppm (Fig. 11).
311 CO₂ and S concentrations decrease with increasing melt SiO₂ contents, ranging from 395
312 ppm S and 313 ppm CO₂ at 58.2 wt.% SiO₂, to 18 ppm S and 20 ppm CO₂ at 72.6 wt.% SiO₂.
313 Cl shows a positive relationship with SiO₂, ranging from 2500 ppm at 58.2 wt.% SiO₂ to
314 3610 ppm at 72.6 wt.% SiO₂.

315

316 **DISCUSSION**

317 The range of compositions and textures in mineral, whole-rock and melt inclusion
318 chemistry suggests that the SSH mafic magma petrogenesis was just as complex as that
319 observed for the SHV andesitic volcanic system on Montserrat and involved the assembly of
320 multiple components. Here we discuss the origin of these components by considering the
321 pressure-temperature conditions of magma storage, fractional crystallization and magma
322 mixing that are reflected in the crystal and melt phases in the SSH erupted products, as well
323 as the conditions required for the eruption of these products at the surface.

324 *Pre-eruptive temperature, pressure and volatile content*

325 Temperature estimates of the magma reservoir conditions are derived from the two-
326 pyroxene thermometry and plagioclase-whole rock equilibria after applying the equilibrium
327 test (where $K_D = 1.09 \pm 0.14$ for pyroxene and 0.1 ± 0.11 for plagioclase) (Table 3; Putirka,
328 2008). The calculated temperature range of 970-1170 °C is hotter than the estimates of the
329 temperature for the neighbouring SHV magma reservoir, which is thought to reside at $840 \pm$
330 40 °C based on experimental studies and pyroxene thermometry, heated by mafic magmas
331 with temperatures of 900 ± 100 °C (Devine et al., 1998; Barclay et al., 1998; Murphy et al.
332 2000; Devine et al. 2003; Humphreys et al., 2009) (Table 3). The SSH temperatures reported
333 here were calculated on different samples and give a wide temperature range, which supports
334 our argument that the erupted magma comprises components assembled from multiple
335 magma bodies with differing storage conditions.

336 The melt inclusion data were used to estimate equilibration pressures using
337 Volatilecalc (Newman and Lowenstern, 2002; Table 4). Most of the calculated pressures
338 (using a temperature of 1000 °C) range from 194-267 MPa, which equates to depths of 8.4-
339 11.6 km (using an upper crustal density of 2300 kg/m³; Hautmann et al., 2013), with one
340 sample yielding a pressure of 25 MPa and a depth of 1.2 km. By comparison the magma
341 stored beneath SHV is thought to reside in a dual reservoir system, one at 5-6 km depth, and
342 the other at 10-12 km depth (Devine et al., 1998; Murphy et al., 1998; Barclay et al., 1998;
343 Elsworth et al., 2008; Paulatto et al., 2010). With the exception of the low H₂O measurement
344 (1.5 wt.%), which likely represents a melt inclusion that either equilibrated at shallow depth
345 (1.2 km) or has lost H⁺ by diffusive equilibration (Gaetani et al., 2012), the H₂O contents in
346 the SSH melt inclusions lie at the upper range of H₂O contents (1.0-6.3 wt.%) measured in
347 SHV melt inclusions (Humphreys et al., 2009; Mann et al., 2013; Edmonds et al., 2014).
348 Thus, the high anorthite contents in the cores of the SSH plagioclase crystals (up to An₉₇) are

349 most likely due to the high dissolved H₂O contents (water contents exert a first order control
350 on anorthite content and can elevate the anorthite contents to >An₉₀; Figure 4 in [Lange et al.,](#)
351 [2009](#)).

352 *Melt inclusion chemistry*

353 With one exception, H₂O contents are approximately constant over the entire range of
354 K₂O, SiO₂ and MgO concentrations ([Fig. 11](#)). At depths of 8-12 km, the exsolved vapour is
355 likely to be CO₂-rich ([Blundy et al., 2010](#)), and the invariant water contents may thus reflect
356 that the source of magmas hosting the phenocrysts erupted at SSH had similar primary H₂O
357 contents ([Tables 2 and 4](#)). Cl concentrations are positively correlated with those of SiO₂,
358 consistent with Cl behaving incompatibly with little or no degassing. Both CO₂ and S
359 contents decrease with increasing SiO₂, indicating that these volatiles were progressively
360 partitioned into a vapour phase as melts evolved. This is consistent with experimental data
361 that suggests that oxidised arc rhyolites are associated with high vapour-melt partition
362 coefficients for sulphur ([Clemente et al., 2004](#); [Zajacz et al., 2012](#)). Similar melt inclusion
363 trends have been observed in melt inclusion suites from Grenada which range from basalt to
364 rhyolite and thought to be related by fractional crystallisation ([Devine, 1995](#)), as well as
365 other examples from Kermadec arc ([Haase et al., 2006; 2011; Barker et al., 2013](#)), South
366 Sandwich islands ([Pearce et al., 1995](#)), Mt Shasta ([Grove et al., 2003](#)) and from experimental
367 studies ([Sisson et al., 2005](#)).

368 [Figure 2](#) illustrates a comparison of melt inclusion and whole rock data from SSH and
369 SHV with models of fractional crystallisation at pressures of 100-200 MPa under moderately
370 oxidizing conditions using the AlphaMelts/RhyoliteMELTS model ([Ghiorso and Sack 1995](#);
371 [Gualda et al., 2012](#)). Two different scenarios are considered, the first models fractional
372 crystallisation from a mafic bulk rock starting composition, and the second starts the model

373 from the most mafic melt inclusion composition. In the first, the starting composition is
374 defined by the most mafic of the SSH whole rocks (~47% SiO₂). The input parameters
375 include a fixed pressure (100 or 200 MPa), a starting temperature of 1200°C (as defined by
376 the two pyroxene thermometer above, and close to the calculated liquidus temperature from
377 RhyoliteMELTS) and an oxygen fugacity, f_{O_2} , buffered at QFM+2 or NNO (Devine et al.,
378 1998; [Murphy et al., 2000](#)). The melt was then cooled at 50 °C intervals to simulate isobaric
379 fractional crystallization involving olivine, plagioclase, magnetite, augite, enstatite and
380 amphibole ([Table 2, Fig. 2](#)).

381 Regardless of the pressure or f_{O_2} , simple isobaric fractional crystallization predicts
382 non-linear liquid lines of descent that fail to reproduce the simple linear trends defined by the
383 majority of the whole rock data. Hence, the range in whole rock data from both SSH and
384 SHV are best described by a hybridization model in which the rocks are mixtures between
385 andesitic to rhyolitic melts and mafic crystal phases, as observed in many other arc volcanic
386 settings ([Davidson et al., 2005](#); [Reubi and Blundy, 2009](#); [Kent et al, 2010](#); [Cashman and](#)
387 [Blundy, 2013](#); [Humphreys et al., 2013](#); [Cooper and Kent, 2014](#)).

388 In contrast, a fractional crystallisation history can explain most of the melt inclusions
389 from SSH, and a significant proportion of those from SHV. These melt inclusions do not lie
390 on the linear trend defined by the whole rock data. For the melt inclusions, the best fit to the
391 AlphaMelts/RhyoliteMELTS model ([Ghiorso and Sack 1995](#); [Gualda et al., 2012](#)) is provided
392 by a scenario in which the starting composition is defined by the most mafic of the SSH melt
393 inclusions (58.7% SiO₂). The input and cooling parameters are the same as for the first
394 modelling scenario above and, again, the effects of pressure and f_{O_2} do not yield major
395 variation in the liquid line of descent ([Fig. 2](#)).

396 To summarise, the melts are related to one another by fractionation crystallisation and
397 likely evolve in closed systems in storage lenses in the crust. The bulk basaltic lavas are

398 “assembled” by mixing liquids along this line of descent with mafic crystal mushes
399 containing mixtures of plagioclase, olivine and clinopyroxene. The whole rocks therefore
400 represent hybrids or mixtures between melts and mush components. In detail, it can be
401 observed that most of the melt inclusion liquids are in equilibrium with their host crystals
402 (Table 2), which means that at the time of melt entrapment, the crystal and its carrier liquid
403 were in equilibrium. The crystals are strongly zoned however, and the melts are therefore not
404 necessarily in equilibrium with other parts of the crystal, or with other crystals in the magma.

405 The melt inclusions were trapped over a pressure range corresponding to depths of
406 between 8 and 12 km (Table 4). We speculate that the more mafic liquids are sourced from
407 the deeper parts of the magma reservoir system. In contrast to SHV, the crystal assemblage at
408 SSH is markedly more mafic, likely derived from deeper in the crust. For the basalts of the
409 SSH, the depths recorded from volatile solubilities in melt inclusions suggest that melt
410 entrapment occurs at the deeper end of the range estimated for the SHV system Edmonds et
411 al., (2014), thus preserving a greater range of melt inclusion compositions (from andesite to
412 rhyolite), further suggesting that in general melts become more evolved upward through the
413 crust. This is supported by a broad negative correlation in the melt inclusion data, between
414 SiO₂ and equilibration pressure ($R^2= 0.45$), indicating that the least evolved compositions
415 were generally formed at deeper depths.

416 It is important to note that the record of pressures recorded by the melt inclusions is
417 itself subject to bias. The depths of melt entrapment are probably governed not only by the
418 physical dimensions of the reservoir but also and perhaps more importantly by the conditions
419 under which melt inclusions form, which requires both high degrees of undercooling and a
420 period of isothermal crystal growth (Kohut and Nielsen, 2004; Kent et al., 2008). Mafic
421 phenocrysts may have not experienced sufficient undercooling, until mixing, by which time

422 the compositions had been modified by time isothermal crystallisation occurs and melt
423 inclusions become trapped (Koleszar et al., 2012).

424 Mixing is well documented in other arc systems. A notable example of the mixing
425 process described above is associated with the Mount St Helens dacite, where temperature
426 fluctuations of 20-40 °C were a consequence of incremental, or pulsed assembly of crustal
427 magma bodies wherein each pulse interacts with ancestral, stored magmas, accounting for
428 much of the plagioclase zoning and textural complexity seen in the erupted magmas
429 (Cashman and Blundy, 2013). These authors suggest that magma storage systems under most
430 arc volcanoes are dominated by similar processes, where crystal mushes are fed by hotter,
431 slightly more mafic magma, coupled with episodes of magma ascent from one storage region
432 to another. The presence of common enclaves of cumulate material, such as gabbro and
433 pyroxenite, in the SSH lavas (Cassidy et al., 2014) is also consistent with the remobilisation
434 of plutonic material. The way in which the model we propose differs from this fundamental
435 mixing scenario is that we propose “back-mixing” to generate mafic bulk compositions by
436 mixing more evolved melts with mafic mushes, illustrating the importance of not only
437 mushes, but also regions of andesitic to rhyolitic liquids in magma reservoirs for generating
438 bulk compositions.

439 *Textural evidence for mixing*

440 The olivine, plagioclase and pyroxene phenocryst compositional profiles all record
441 normal and reverse zoning, suggesting a combination of growth zoning and magma mixing
442 (Figs. 6, 7, 8 and 11). Major element mineral chemistry is modified during growth in
443 response to cooling, melt compositional changes and magma reservoir conditions; including
444 pressure, temperature, volatile content and f_{O_2} (Housh & Luhr 1991; Nelson & Montana,
445 1992; Sisson and Grove, 1993; Couch et al., 2003a, 2003b; Streck, 2008; Cashman and
446 Blundy 2013). Minor element concentrations are particularly useful for discriminating

447 between magma mixing and growth zoning, as they are almost entirely a function of melt
448 composition and are largely unaffected by changes in magma storage conditions (Ruprecht
449 and Worner, 2007; Aigner-Torres et al., 2007).

450 Zoning profiles in plagioclase crystals shows that anorthite contents are negatively
451 correlated with Fe, Mg and Ti (Fig. 8), with magma crystallisation and differentiation
452 yielding less An-rich compositions, and increases in magma temperature or water content
453 raising An contents. Although Fe partitioning in plagioclase strongly depends on crystal
454 composition, and melt temperature and f_{O_2} (Longhi et al., 1976; Sugawara, 2001; Aigner-
455 Torres et al., 2007), melt composition has the greatest effect on Fe plagioclase content
456 (Ginibre et al., 2002). By comparison, experimental data show a clear negative correlation
457 between Ti and An% that is largely independent of temperature, and Mg partitioning depends
458 weakly on An content (Bindeman et al., 1998) and temperature (Longhi et al., 1976; Aigner-
459 Torres et al 2007). Therefore, changes in An content, temperature, f_{O_2} alone cannot fully
460 replicate the observed increases in Fe, Mg and Ti observed at the rim of the crystals (Fig. 8).
461 Rather, these observations suggest that the increases in these elements must be due, at least in
462 part, to disequilibrium crystallisation prior to eruption as a result of mixing with melts
463 enriched in Fe, Mg and Ti. This interpretation is supported by the kernel density plots of
464 anorthite content (Fig. 4), where two populations of cores are evident, as well as a large range
465 of anorthite values at the rims. The population of cores with An₇₆₋₉₅ likely represents deeper,
466 more stable plagioclase crystallisation, but the cores with lower anorthite contents (An₅₀₋₆₅)
467 may represent plagioclase crystals that evolved in a shallower (lower PH₂O), more evolved,
468 magma body. Zoning profiles (Figure 8a) show cores with high anorthite contents (An₈₆) and
469 increasingly albitic rims (down to An₅₆) with a corresponding increase in Fe, Mg and Ti
470 contents.. This zoning profile is consistent with a plagioclase from a wet mafic mush being
471 mixed into a more evolved melt at lower pressures.

472 A history of mixing is supported by the presence of two distinct groups in olivine core
473 compositions (Figs. 4 and 5): Group 1, Fo₇₂₋₈₀ and Group 2, Fo₅₆₋₆₈. These groups suggest
474 mixing between two distinct magma batches, or with the entrainment of more forsteritic
475 olivines from a crystal mush into a more evolved crystal-rich magma. The olivine crystals
476 (both Group 1 and Group 2) exhibit both normal (most common) and reverse zoning at the
477 rim of the crystal (Figs. 5c, 6, 7 and 12a). Many of the Group 1 olivines exhibit normal
478 zoning at the rims, consistent with magma from a primitive mush entrained into a more
479 evolved storage system. This hypothesis is illustrated by the zoning profile in Figure 6b,
480 which contains shows a Group 1 olivine with a lower forsterite, but higher Ca and Mn rim.
481 Simple fractional crystallisation would reduce the CaO content along with Fo content, but
482 while Ca and Mn partitioning are not directly affected by melt f_{O_2} and temperature (Dunn,
483 1987; Libourel, 1999), Ca concentration of olivines is strongly dependent on the alkali
484 composition of the melt (Jurewicz and Watson, 1988; Libourel, 1999). Mixing of the Group
485 1 olivines into an evolved melt with a higher alkali content may therefore explain the
486 observed increased Ca content with decreasing Fo. The reverse zoning observed in some of
487 the Group 2 olivines is consistent with olivine from the partially crystalline andesite being
488 exposed to more mafic compositions and hotter temperatures of the intruding magma.

489 Pyroxene Mg# can change in response to changes in melt composition or f_{O_2} (Streck
490 et al., 2002). Thus, the saw-tooth major element zoning in Figure 10 is likely related to a
491 combination of open system fractionation and recharge (Ginibre et al., 2002; Ruprecht and
492 Worner, 2007), while the relatively large increases in Mg# approaching the rims (the outer 40
493 μm) of a fraction (~5%) of the pyroxenes are consistent with a change in the composition
494 and/or temperature of the intruding mafic magma (Fig. 10). Indeed, similar orthopyroxenes
495 have been erupted at SHV eruption since May 1996, with well-developed reverse zoned rims
496 (10–25 μm) (Murphy et al., 2000).

498 The mixing of a phenocryst into a melt of a different composition would lead to a
499 sharp step in the mineral composition crystallising at the rim, assuming that conditions for
500 crystal growth are maintained and that the mixing event results in an instantaneous, rather
501 than gradual, change in the composition of the host melt. This sharp step then relaxes over
502 time, via diffusion, as the interior of the crystal begins to equilibrate with its new host melt
503 composition. The resulting diffusion profiles may be used to estimate the timescales between
504 magma mixing and eruption, by assuming a particular temperature (Costa and Chakraborty,
505 2004; Morgan et al., 2004; Costa and Dugan, 2005; Costa et al., 2008). This diffusion
506 chronometric approach has been applied to reverse zoning profiles in our SSH samples (we
507 cannot apply it to normal zoning profiles, because it is difficult to distinguish mixing-driven
508 disequilibrium from fractionation-dependent growth zoning in this case). We use the DIPRA
509 model (Girona and Costa, 2013) for both forsterite and Mn zoning, at 1000°C. The shapes of
510 the compositional profiles in the reverse zones at the rims of two olivines (Fig. 5b) are
511 consistent with relaxation of an initial compositional step over 10 to 60 days (Supplementary
512 figures 1 and 2). This timescale is similar to that estimated from compositional profiles in
513 Fe-Ti oxides induced by heating in SHV lavas, where andesite remobilisation by mafic
514 intrusions occurred days to weeks prior to eruptions (Devine et al., 2003). A timescale of
515 days to weeks between mixing and eruption is comparable to the short pre-eruptive mixing
516 timescales calculated at Ceboruco, Quizapu, Nea Kamini and Mount Unzen volcanoes (days
517 to months; Nakamura, 1995; Chertkoff and Gardner, 2004; Ruprecht and Cooper, 2012;
518 Martin et al., 2008). Other mixed systems at Trident, Taupo and Volcan San Pedro, give
519 longer timescales (months to decades; Coombs et al. 2000; Costa and Chakraborty, 2004;
520 Millet et al., 2014). Our results imply a relatively short period between the assembly of the
521 SSH magmas and their ascent and eruption at the surface.

522 *Formation of basaltic magmas at the SSH*

523 Basalts are often thought to represent relatively unmodified primary melts from the
524 mantle. However observations in this study from whole rock trends, melt inclusions,
525 fractional crystallisation modelling and phenocryst zoning attest to a hybridisation model
526 similar to that previously inferred for the formation of andesites at many intermediate
527 systems. Magma mixing commonly occurs between mafic and felsic melts to form andesitic
528 compositions, following the recharge filtering model of Kent et al. (2010). However, the
529 basaltic whole rock compositions of SSH are generated mixing components from multiple
530 magma bodies, comprising andesitic to rhyolitic melt compositions and mafic mineral phases.
531 The SSH preserves a wide range of melt inclusion compositions unlike the SHV which
532 comprises only limited range of evolved rhyolitic melt inclusions. This is likely a
533 consequence of the deeper mixing of multiple different magma bodies and the lack of a
534 further shallow crystallisation stage, which would otherwise increase the likelihood of
535 preserving silicic melt inclusions through the incorporation of crystals derived from a shallow
536 crystal mush.

537 *Tectonic control for the eruption of basalts*

538 While many of the observations relating to magma mixing as a control over whole
539 rock and melt inclusion compositions have been well-documented in arc volcanic rocks
540 ([Reubi and Blundy, 2009](#)), they do not explain the closely-spaced and near coeval eruption of
541 basaltic and andesite lavas at SSH and SHV ~130 ka. In this context, it is noteworthy that the
542 volatile contents of melt inclusions and geophysical investigations of SHV support the
543 existence of two upper crustal magma chambers; one at 10-12 km that feed into a shallower
544 chamber at 5-6 km depth that serves as the source of the erupted material ([Devine et al.,
545 1998; Barclay et al., 1998; Elsworth et al., 2008; Humphreys et al., 2009; Paulatto et al.,
546 2010; Mann et al., 2013; Edmonds et al., 2014](#)). We hypothesise that the eruption of more

547 mafic rocks at SSH was because these lavas were assembled directly from a magma chamber
548 of similar depth (8-12 km) to the deeper of the two chambers below SHV, but without
549 passing through the shallower chamber. But what allows the SSH basalts to bypass this
550 shallow density filter?

551 In general, eruption of basaltic compositions in dominantly andesitic settings requires
552 a favourable stress field (Hildreth, 1981). Indeed, density is not the only factor which limits
553 the ascent of mafic magmas; structural controls imposed by lithology and rheological
554 boundaries within the crust can also act to slow and sometimes stall magma ascent
555 (Eichelberger, 1978; Dufek and Bergantz, 2005; Karlstrom et al 2009; Kent et al., 2010).
556 Faulting systems may promote the ascent of denser magmas, particularly within an
557 extensional and therefore decompressional regime. Volcanoes are also commonly found
558 along major strike-slip faults, such as the great Sumatran fault zone, the Sulawesi fault and
559 the Liquiñe–Ofqui fault zone (LOFZ) in Chile (Bellier and Sébrier, 1994; Lécuyer et al.,
560 1997; Cembrano and Lara, 2009). In these areas, local extensional features are associated
561 with individual volcanoes, and it is suggested that a causal relationship exists between
562 extension and volcanism or intrusion (Moore, 1979; Aydin and Nur, 1982; Hutton and
563 Reavey, 1992; Tibaldi, 1992; Milia and Torrente, 2003; Spinks et al. 2005; Brogi et al. 2010;
564 Davis et al. 2010). In addition, there is evidence that tectonics can strongly control the
565 composition of magmas. For instance, at the Taupo volcanic zone basaltic volcanism occurs
566 at the intersection between major faults and caldera boundaries (Cole et al., 1990; Millet et al.
567 2014) whereas, more intermediate magmatism occurs in areas which have experienced less
568 crustal extension (Allan et al. 2013 ; Millet et al. 2014), following the recharge filtering
569 process. Transtensional faults in the neighbouring island to Montserrat, Guadeloupe, which
570 lies along the same en echelon fault system, are thought to control the location of volcanism
571 and may be the cause for the frequent sector collapses on the island (Mathieu et al., 2011).

572 Transtensional tectonics in this region may not only control the source of these magmas
573 (Cassidy et al., 2012), but may also lead to localised faulting that thus provides a pathway for
574 these higher density mafic magmas, that would otherwise be trapped within the crust (Fig.
575 12). Over time, however, the crust in these areas may impose lithostatic control as the
576 eruption of the basalts thickens the crust. As a result, later magmas would be required to
577 undergo differentiation by crystal segregation to become buoyant enough to erupt at the
578 surface (Plank and Langmuir, 1988; Devine, 1995), thus increasing the likelihood of
579 generating more evolved andesites. This is supported by numerical modelling from Pinel and
580 Jaupart (2000), which predicts that as the edifice grows the ascension of lower density
581 magma is favoured, thus promoting stalling in the crust and magma differentiation. Hence,
582 the eruption of basaltic lavas may be characteristic of the early products of new eruption
583 centres where extensional tectonics are operative in arc settings. This may be the case for
584 many volcanic regions which comprise early phases of basaltic activity before evolving into
585 mature andesitic systems, including northern Japan (Katsui et al., 1978; 1979); central south
586 Chile (Lopez-Escobar et al., 1977), New Zealand (Price et al., 2005), the Aleutians and
587 Alaska (Marsh, 1980; Myers and Marsh, 1981). The role of transtensional tectonics is
588 strengthened by the observation that both Redonda and Kahouanne, two adjacent islands to
589 Montserrat which lie on the same transtensional fault systems (Fig. 1), also produce mafic
590 volcanism. These seamounts represent the emergence of new volcanism in the Lesser
591 Antilles, and again suggest that early arc volcanism in this region may be controlled by
592 tectonics, until further growth of the edifice inhibits the ascent of high density mafic magmas,
593 producing the commonly observed andesitic volcanoes. Although fault structures thus
594 provide a possible mechanism for promoting the ascent of the SSH magmas, this alone does
595 not explain the timing of SSH basaltic magmatism. Basaltic eruptions have not been
596 identified at other periods in Montserrat's history. The conditions favourable to basaltic

597 eruptions at SSH thus appear to have been transient, and are unique in the currently identified
598 history of Montserrat. The SSH doesn't clearly correspond to an initial phase of volcanism, in
599 the sense of the birth of a new volcanic centre, since the event is bracketed by andesite
600 eruptions at the adjacent SHV, and there have been no subsequent eruptions (since 130 ka) at
601 SSH. We know of no reasons why fault activity at the time of SSH volcanism would have
602 been enhanced relative to other periods in Montserrat's history. Thus, although fault
603 structures may have promoted ascent of dense mafic magmas at this location, this alone does
604 not provide a satisfactory explanation for the timing of the SSH episode of basaltic
605 volcanism. Other processes affecting crustal stress conditions, such as collapse of the
606 volcanic edifice, may help explain the precise timing of SSH volcanism.

607

608 **CONCLUSIONS**

609

610 There is now abundant evidence that arc andesites are generated by hybridisation
611 processes, involving the mixing of felsic melts and abundant crystal phases, for instance at
612 the SHV on Montserrat, Mt St Helens and at Mount Hood (USA). Arc basalts, on the other
613 hand, are commonly attributed to simple closed-system fractionation. Our study of the SSH,
614 shows that olivine-bearing basalt petrogenesis can be just as complex as the generation of
615 andesites at the SHV, implying that basalts in arcs may have a less simple history than is
616 commonly assumed on account of the hybridisation processes explored in this study.

617 This study also shows how two volcanoes active at similar times and located very
618 close to each other can erupt different bulk compositions. Basalts erupted from the SSH in
619 Montserrat were stored under different magmatic conditions to the andesites of the SHV, yet
620 underwent similar magmatic processes of mixing, recharge and cumulate entrainment prior to
621 eruption. The range of magmatic temperature estimates (970 - 1160°C), reservoir depth

622 estimates (8-12 km), coupled with crystal and whole rock compositions, strongly indicates
623 the presence of multiple magma bodies, which interact and feed basaltic eruptions. Melt
624 inclusion data, phenocryst chemistry and fractional crystallisation modelling suggests that
625 mixing and crystal entrainment were involved in the petrogenesis of the SSH mafic magmas.
626 The SSH magmatic system seems to match the deeper mafic-proposed SHV magma
627 reservoir, but geophysical and petrological studies suggest that this deeper SHV system is
628 much larger in volume than the shallow SHV reservoir. This is in contrast with the SSH, the
629 results here show evidence for small, discrete pockets of crystal mushes with melt batches,
630 which might appear in geophysical surveys as one large reservoir. We suggest that ascent of
631 mafic magmas can be promoted by tectonics, which may ascend along faults or under
632 specific stress conditions (i.e. post collapse).

633

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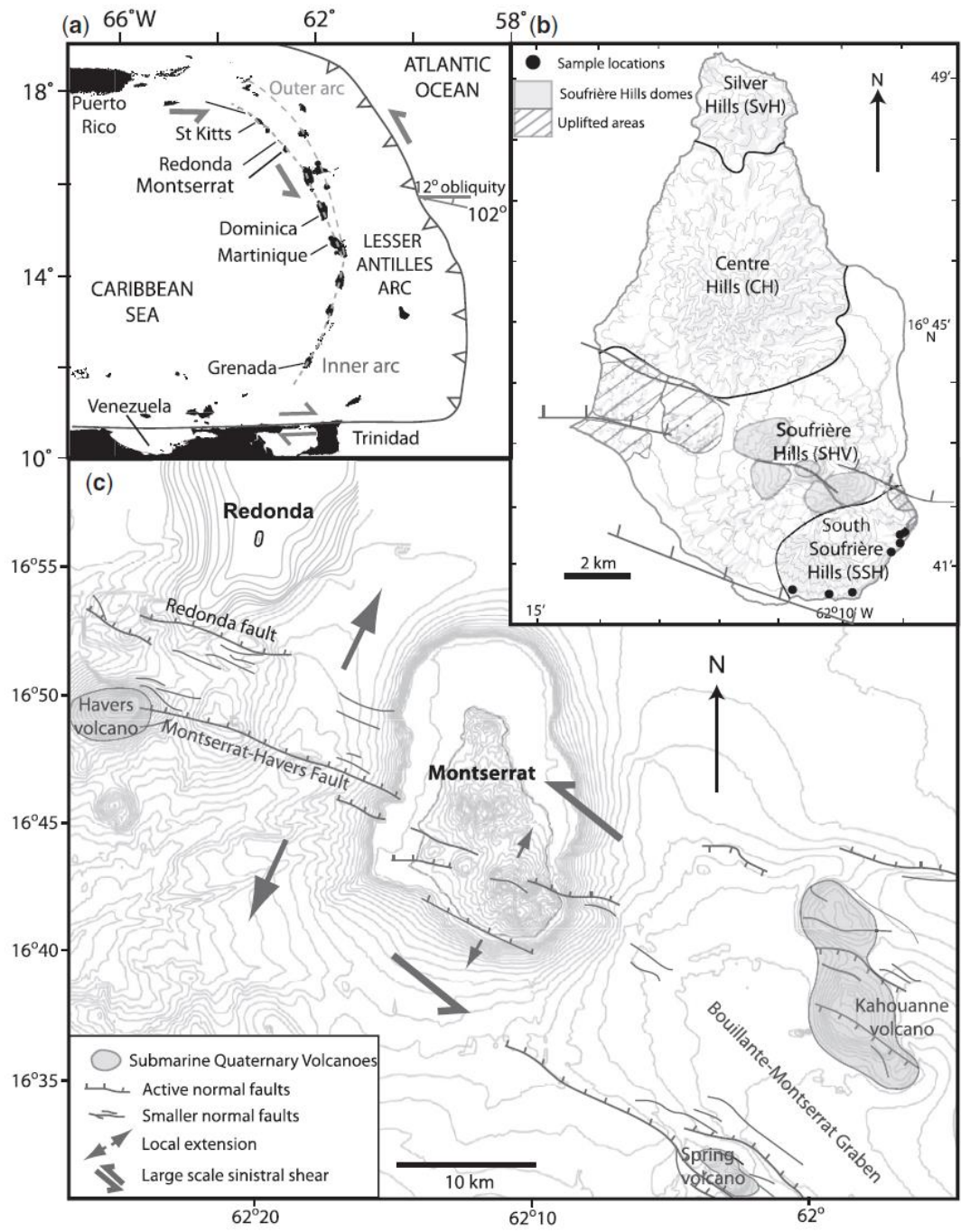


Fig. 1. (a) Regional map of the Lesser Antilles, showing oblique subduction and stresses. (b) The four volcanic centres of Montserrat, with locations of the sampled rocks from the SSH for this study indicated by the filled circles. (c) Submarine and sub-aerial faults and transensional stresses in the region.

Table 1: XRF bulk-rock data along with the standard Japanese Andesite 2 (JA-2), for localities and stratigraphic units sampled at the SSH

Latitude (N):	16-67672	16-67622	16-69131	16-69226	16-69005	16-69005	16-6769	JA-2	2 RSD (%)
Longitude (W):	62-1693	62-1683	62-1487	62-1479	62-1488	62-1488	62-179		
Stratigraphic unit:*	SSH B	SSH A	SSH A	SSH B	SSH A	SSH A	SSH A		
Sample name:	SSH3	SSH4	SSH5B	SSH10	SSH7G	SSH7B	SSH1F		
<i>Major elements (wt %)</i>									
SiO ₂	49.22	50.87	58.77	48.06	47.84	50.19	52.69	56.35	0.2
TiO ₂	1.00	0.87	0.59	0.97	0.90	0.77	0.87	0.65	1.7
Al ₂ O ₃	18.67	19.20	17.95	18.89	19.30	19.67	17.63	15.64	2.1
Fe ₂ O ₃	10.90	9.58	7.58	10.65	10.31	9.02	8.76	6.12	0.4
MnO	0.18	0.18	0.19	0.19	0.18	0.18	0.18	0.11	1.2
MgO	5.93	4.83	3.19	5.33	5.33	4.58	4.17	7.51	1.6
CaO	10.93	10.65	7.27	10.96	11.52	10.39	9.04	6.36	1.3
Na ₂ O	2.45	2.73	3.52	2.53	2.41	2.69	3.21	3.18	3.0
K ₂ O	0.63	0.69	0.60	0.59	0.53	0.34	0.77	1.85	3.3
P ₂ O ₅	0.12	0.13	0.15	0.10	0.09	0.11	0.21	0.15	3.7
Total	100.0	99.7	99.8	98.3	98.4	97.9	97.5	99.2	

*According to Cassidy *et al.* (2012, 2014).

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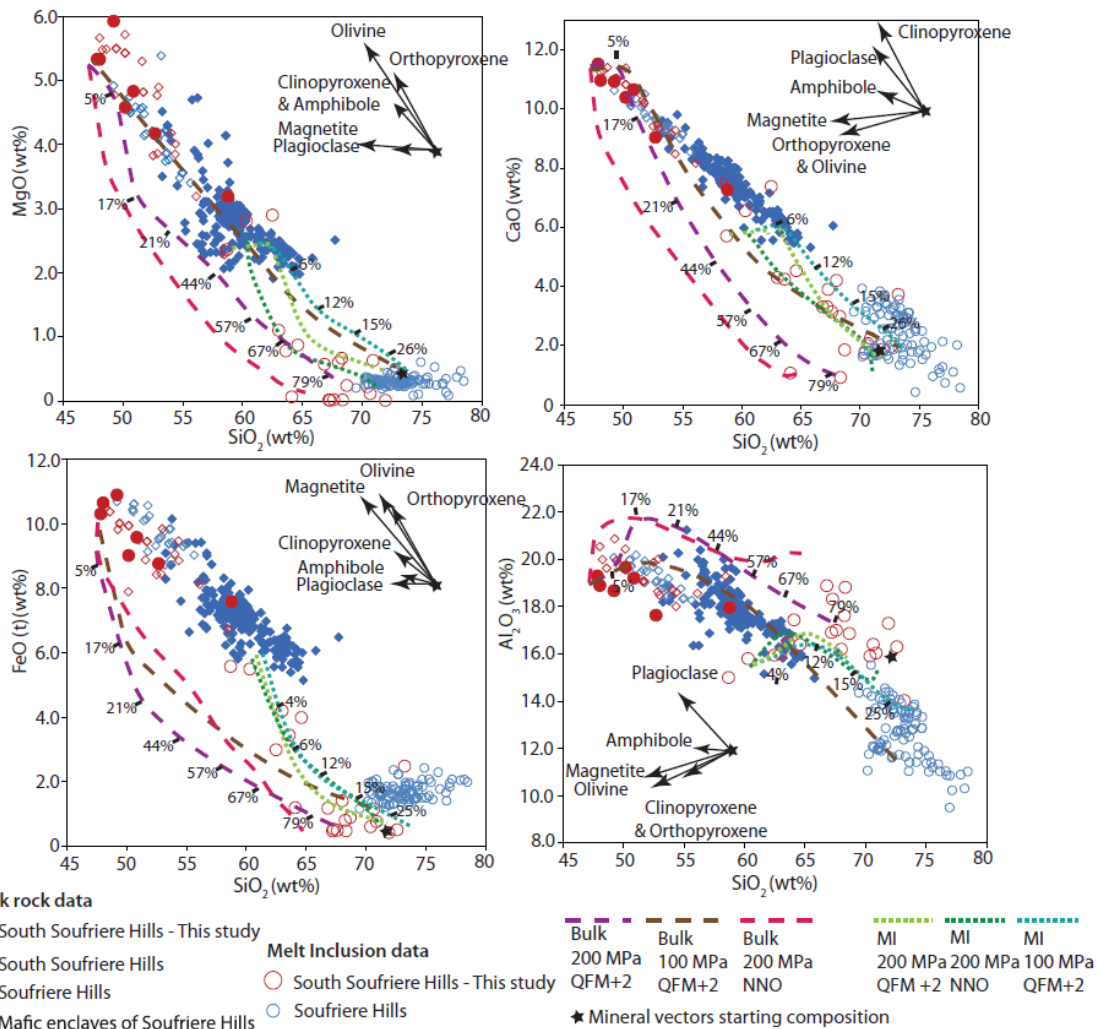


Fig. 2. Whole-rock variation diagrams from the Soufrière Hills, the mafic enclaves within the Soufrière Hills (SHV), the South Soufrière Hills (SSH), and the SSH samples used in this study. Data sources include Murphy *et al.* (1998, 2000), Horwell *et al.* (2001), Zellmer *et al.* (2003) and Cassidy *et al.* (2012). Dashed lines indicate fractional crystallization modelling under variable pressure and f_{O_2} conditions.

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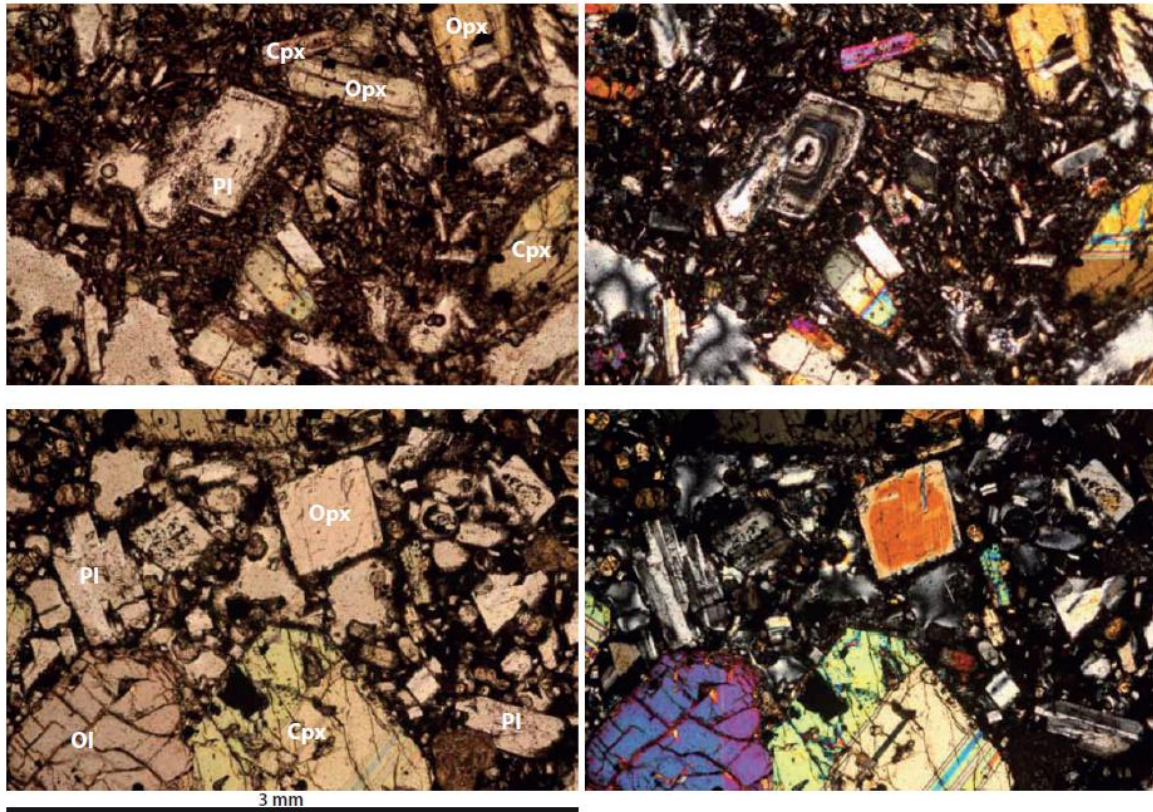


Fig. 3. Photomicrographs of two representative basalts, in plane-polarized light (left) and cross-polars (right). Each slide is 3 mm across. Some mineral phases are labelled: Cpx, clinopyroxene; Opx, orthopyroxene; Ol, olivine; Pl, plagioclase. Samples shown are 6_SSH1F (top) and 9_SSH4. The high crystallinity, large phenocrysts and features such as oscillatory zoning (e.g. plagioclase in top right photograph), normal zoning, twinning and sieve textures should be noted. Olivine is commonly large and fractured in appearance. Some pleochroism is present in clinopyroxenes.

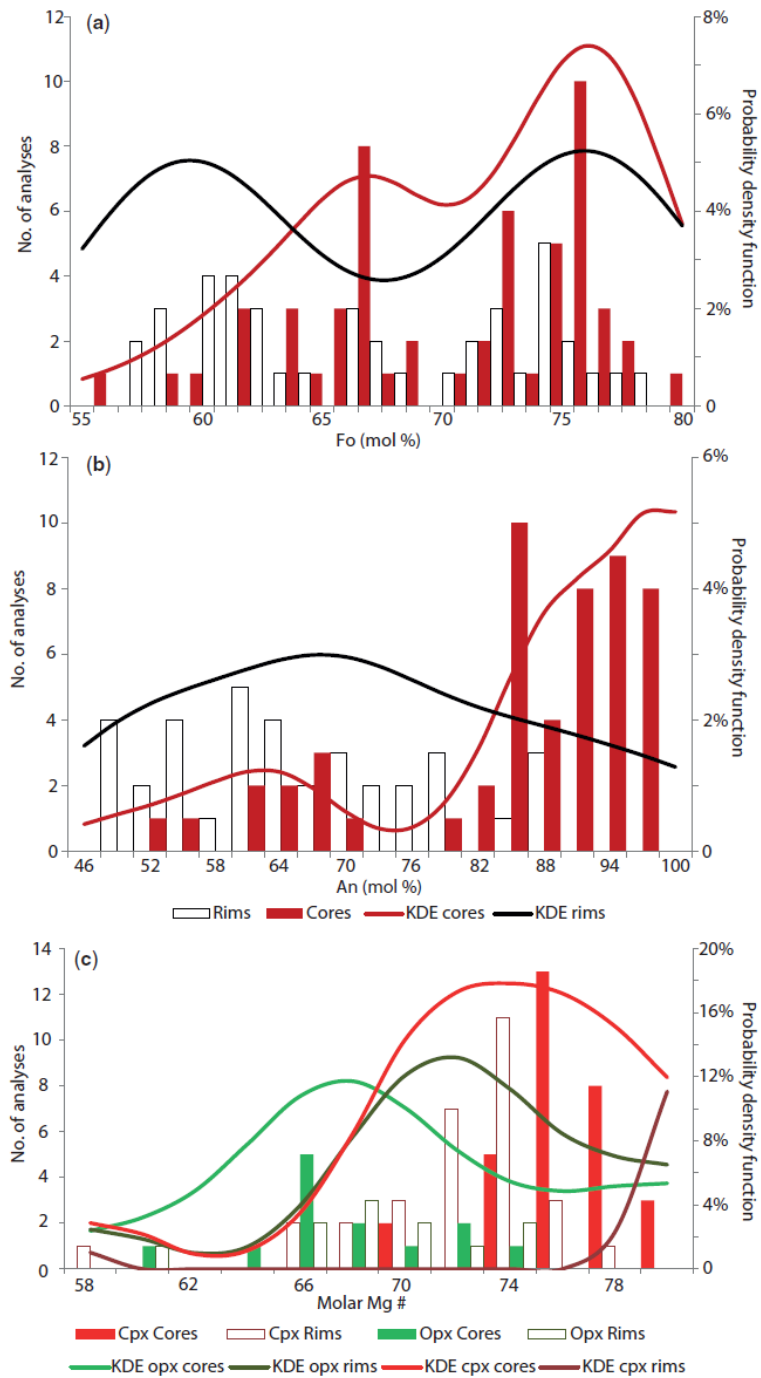


Fig. 4. Histograms to show the main crystal chemical ranges, including Kernel Density Estimation (KDE) curves, which correspond to the probability density function axis. (a) Distribution of forsterite content in olivines. (b) Anorthite distribution of cores and rims in plagioclase crystals. (c) Mg-number distribution of cores and rims in ortho- and clinopyroxene. The key for (a) and (b) is provided at the bottom of (b).

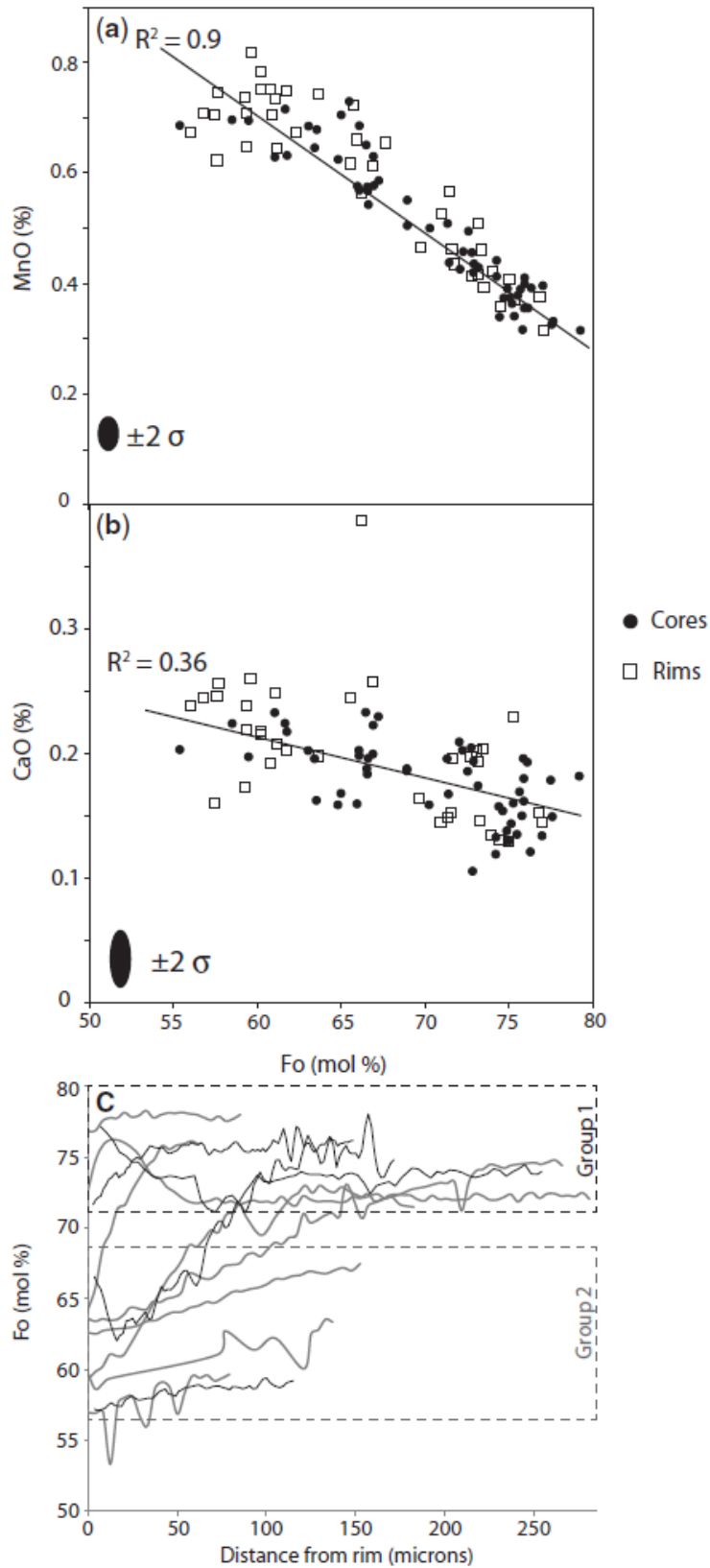


Fig. 5. (a, b) Point data plots of the chemical ranges in forsterite content against minor elements for olivines. (c) Zoning profiles for olivines. Black lines are traverses calculated by calibrated backscattered SEM images and grey lines are traverses measured directly with the electron microprobe.

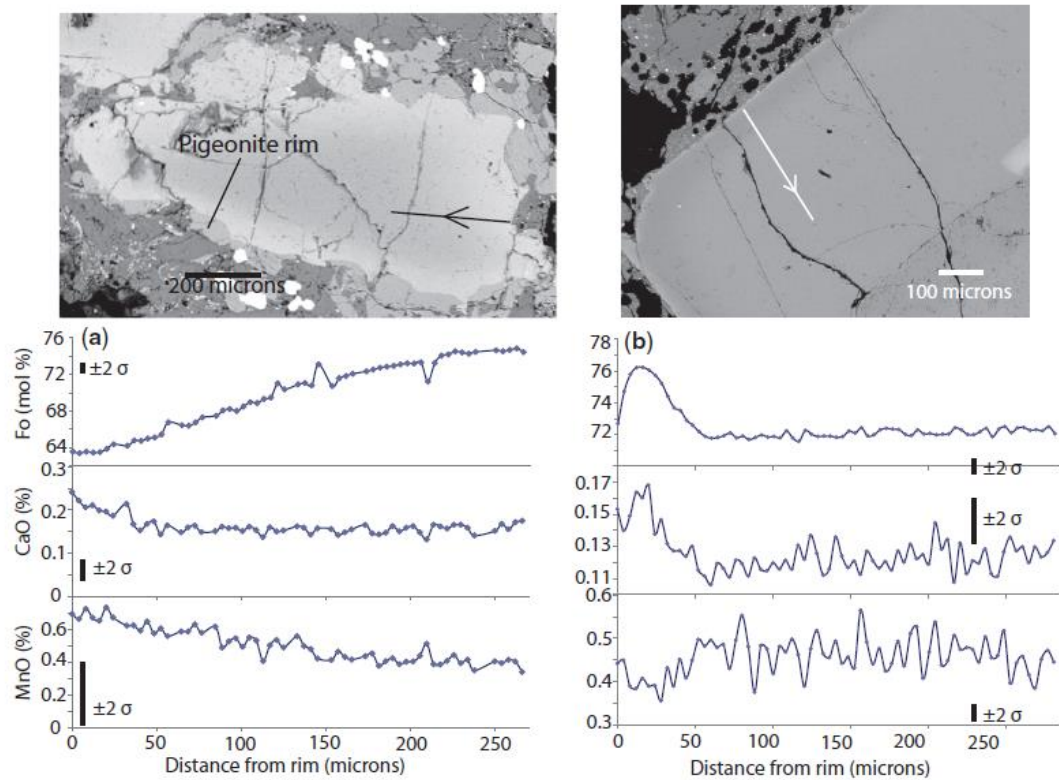


Fig. 6. (a) Normally zoned olivine profile (9_SSH4_ol01); (b) complex zoned rim of an olivine (15_SSH7B_OI02).

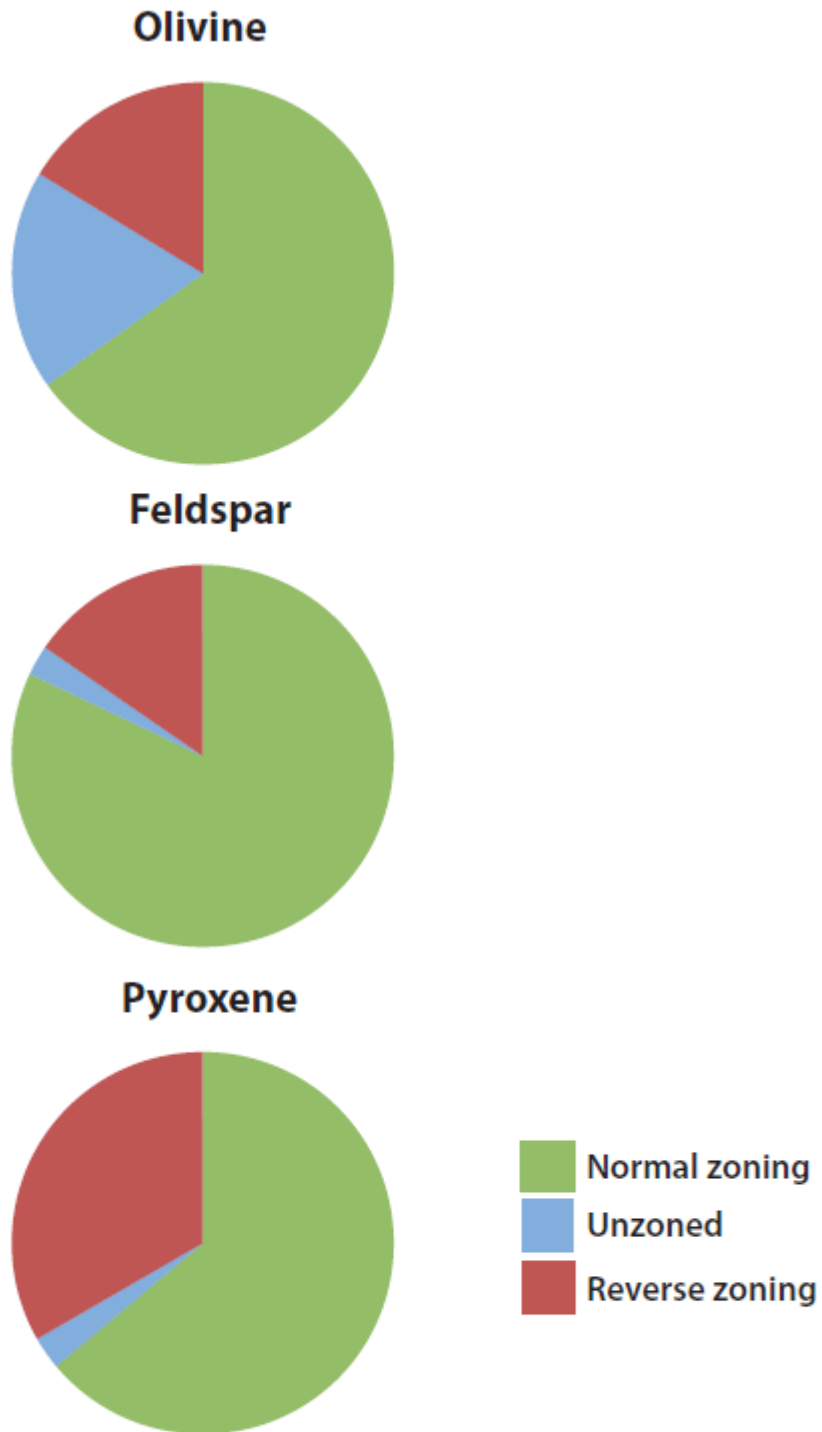


Fig. 7. Pie charts showing the proportion of crystals exhibiting normal, reverse or no zoning for olivine, plagioclase feldspars, and clino- and orthopyroxenes.

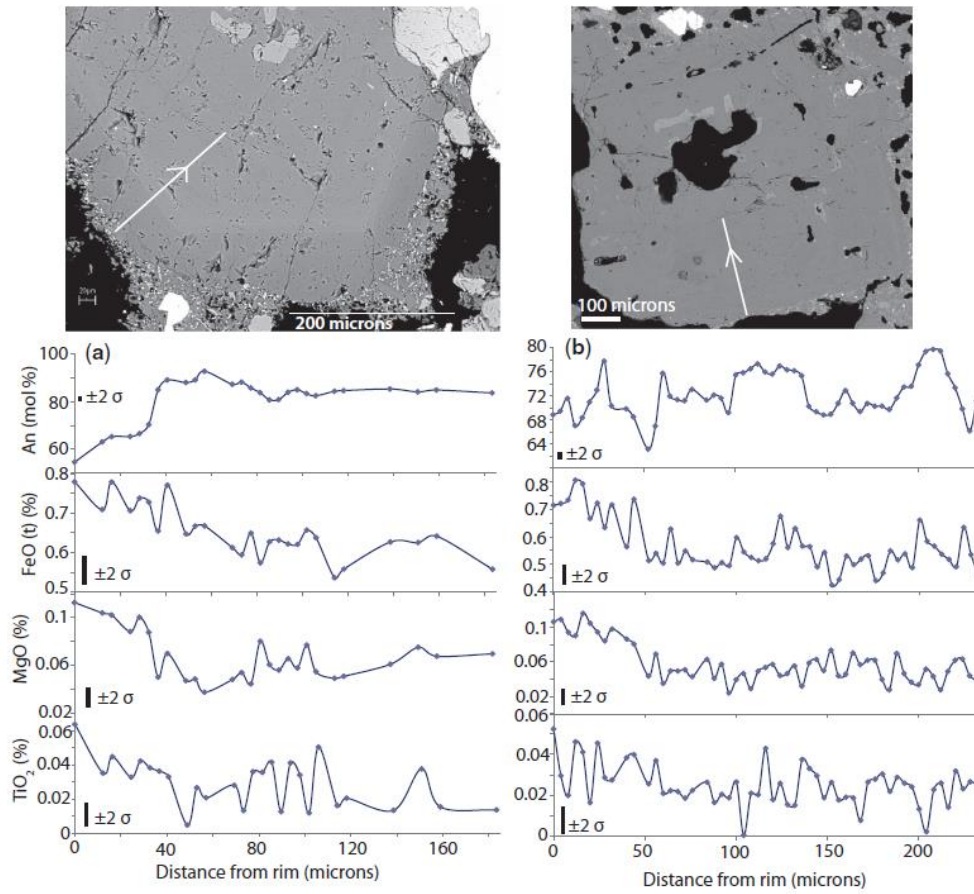


Fig. 8. Profiles of multi-element traverses showing plagioclase zoning: (a) s normal zoning of 9_SSH4Plag02; (b) demonstrates complex oscillatory zoning (15_SSH7BPlag02).

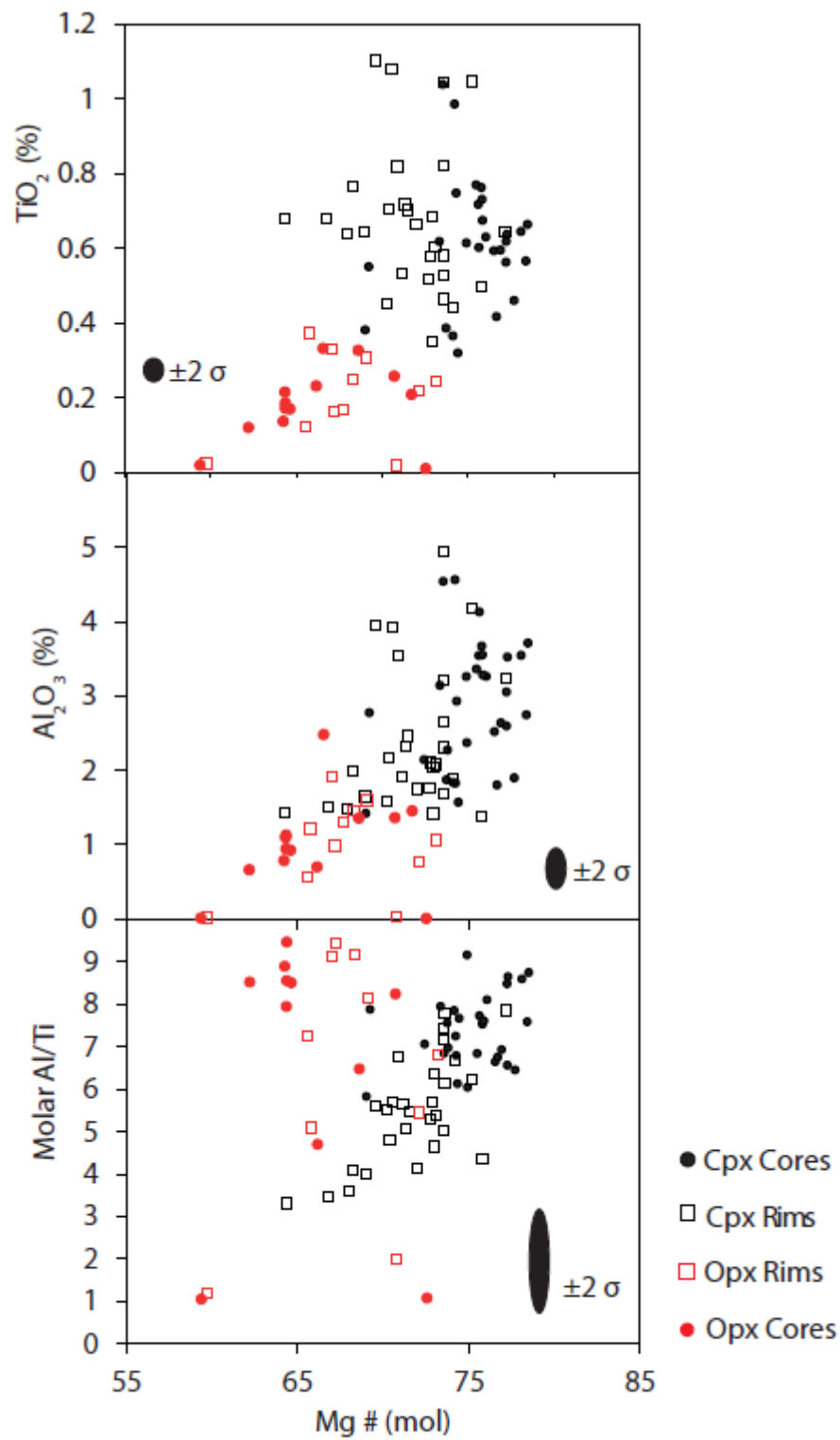


Fig. 9. Point data plots of the chemical ranges in Mg-number against minor elements for clino- and orthopyroxene.

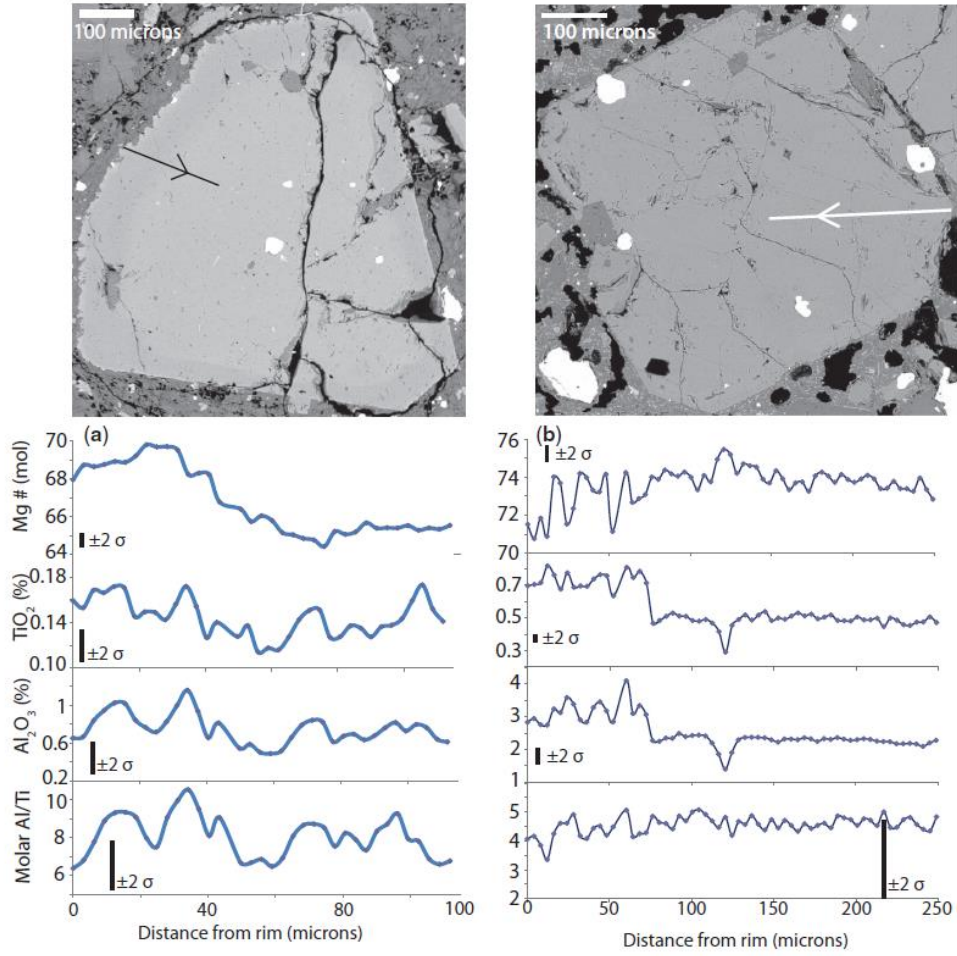
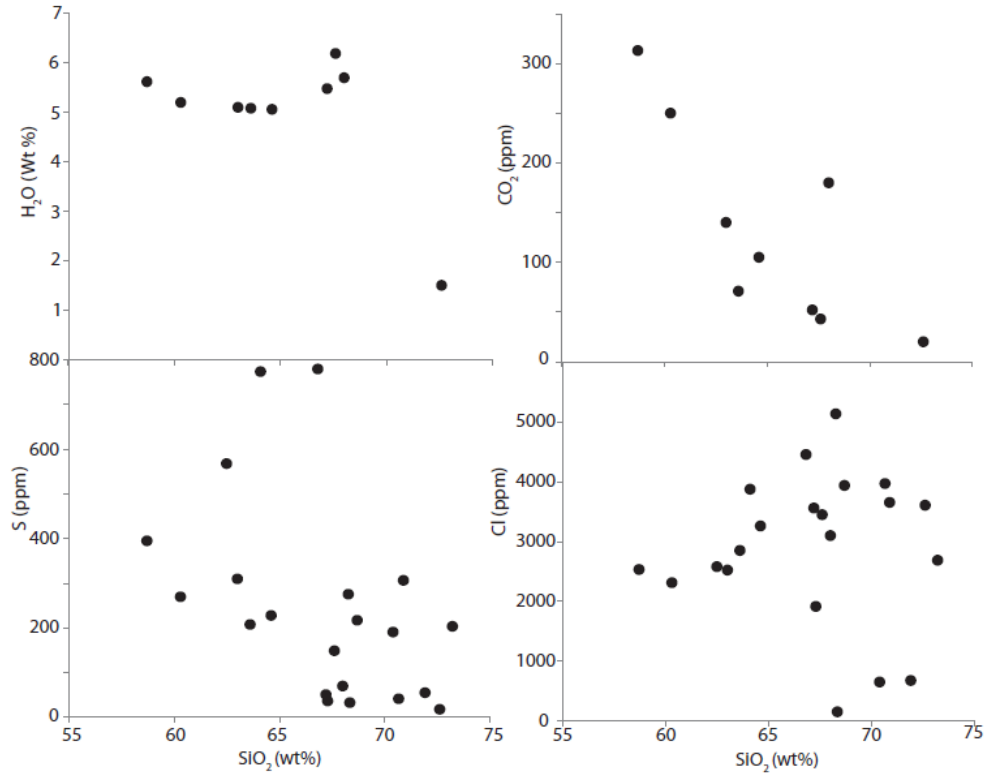


Fig. 10. Profiles of multi-element traverses showing pyroxene zoning: (a) reverse zoning of an orthopyroxene (12_SSH5Bopx09); (b) normal zoning of clinopyroxene (15_SSH7Bpyx02).



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Fig. 11. Melt inclusion plots, SiO₂ versus volatile contents.

Table 2: Major and volatile element compositions for melt inclusions analysed by electron microprobe and secondary ion mass spectrometry (SIMS) analysis

Sample	Phase	SiO ₂	MgO	Al ₂ O ₃	Na ₂ O	K ₂ O	CaO	TiO ₂	Cr ₂ O ₃	FeO _{tot}	Total	H ₂ O	CO ₂	S	Cl	Mg# MI	Mg# host	K _D
SSH3	Pyx	72.60	0.44	16.30	4.91	2.61	1.97	0.22	b.d.	0.50	100.20	1.50	20	18	3610	67	79	0.54
SSH4	Pyx	64.60	0.87	16.40	4.08	1.05	4.53	0.86	0.009	3.98	97.00	5.06	105	228	3260	33	62	0.31
SSH4	Pyx	63.60	0.78	16.50	4.31	0.97	4.24	0.59	0.008	3.44	95.00	5.08	71	208	2850	34	63	0.31
SSH10	OI	58.70	2.36	15.00	2.13	2.22	5.70	0.68	0.003	5.57	92.90	5.62	313	395	2530	49	77	0.29
SSH5B	Pyx	67.60	0.02	17.00	4.51	1.58	3.14	0.48	b.d.	0.48	95.40	6.19	43	149	3450	8.7	66	0.05
SSH5B	Pyx	67.20	0.01	16.90	4.49	1.60	3.31	0.70	b.d.	0.46	95.10	5.48	52	51	3560	4.8	65	0.03
SSH10	OI	60.30	2.80	15.80	1.94	0.32	6.56	0.81	b.d.	5.49	94.00	5.20	250	270	2310	54	80	0.29
SSH5B	Pyx	68.00	0.60	16.20	4.20	1.60	4.20	0.41	0.006	1.40	96.60	5.70	180	70	3100	50	76	0.31
SSH4	Pyx	63.00	1.10	16.00	3.30	0.90	4.30	0.62	0.008	4.20	93.40	5.10	140	310	2520	38	65	0.32
SSH10	Pyx	68.33	0.01	18.82	6.12	7.55	0.92	0.04	b.d.	0.47	102.28			33	141	4.8	72	0.02
SSH4	Pyx	71.91	0.01	17.29	8.11	0.67	1.87	0.06	b.d.	0.41	100.42			55	667	3.6	73	0.01
SSH4	Pyx	68.68	0.24	16.86	3.92	4.78	1.85	0.57	0.004	0.89	98.47			217	3940	32	74	0.17
SSH4	Pyx	70.39	0.22	15.91	6.04	2.03	1.75	0.17	0.002	0.60	97.30			191	642	40	73	0.24
SSH4	Pyx	68.26	0.66	17.62	5.23	2.53	2.99	0.36	0.013	0.79	99.21			276	5142	60	73	0.56
SSH3	Pyx	64.10	0.06	17.43	3.70	7.92	1.07	0.59	b.d.	1.18	96.74			774	3877	8.5	71	0.04
SSH3	Pyx	70.88	0.63	16.04	5.31	2.15	2.24	0.38	0.006	0.76	98.99			307	3657	60	77	0.45
SSH3	Pyx	67.29	0.01	18.31	5.70	2.48	3.90	0.03	0.014	0.52	98.46			37	1910	3.0	75	0.01
SSH10	Pyx	70.65	0.11	16.41	4.35	3.09	2.06	0.88	0.002	1.15	102.93			42	3973	15	70	0.08
SSH3	Pyx	66.81	0.57	18.88	5.83	2.05	3.31	0.57	b.d.	1.17	100.02			780	4460	46	74	0.31
SSH3	Pyx	62.49	2.90	15.95	3.48	3.17	7.38	0.70	b.d.	2.98	99.55			568	2577	63	73	0.65
SSH5B	Pyx	73.21	0.28	14.03	3.65	2.26	3.73	0.57	b.d.	2.48	101.03			204	2685	17	67	0.10

All major element oxide concentrations are given in units of wt %; CO₂, S and Cl are given in ppm. FeO_{tot} denotes the FeO oxide concentration assuming that all Fe in the sample exists as Fe²⁺. Mg# MI is the magnesium-number of the melt inclusion [molar Mg/(Mg + Fe)]; Mg# host is the magnesium-number of the crystal host (olivine, pyroxene—augite or enstatite); K_D is given by (X_{Fe}/X_{Mg})_{host}/(X_{Fe}/X_{Mg})_{melt} (in moles), which is equal to 0.3 ± 0.04. Values outside this range indicate disequilibrium between melt inclusion and host (Roeder & Emslie, 1970). b.d., below detection limit.

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Table 3: Temperatures and pressures estimated using results of [Putirka \(2008\)](#)

Sample	T (°C)	An% or Mg# value	Method
11_SSH10_cpx04	1043	70.7	Two pyx
11_SSH10_opx06	1024	70.4	Two pyx
6_SSH1F_cpx04	970	69.0	Two pyx
14_SSH7G_plag01	1166	89.7	Plag–melt
14_SSH7G_plag06	1166	94.0	Plag–melt

Table 4: Equilibration pressures for melt inclusions trapped in clinopyroxene and olivine phenocrysts using the saturation models of [Dixon et al. \(1997\)](#) in VOLATILECALC ([Newman & Lowenstern, 2004](#))

H ₂ O (wt %)	CO ₂ (ppm)	T (°C)	H ₂ O _v (mol %)	CO ₂ (mol %)	P (MPa)	Depth (km)
1.50	20	1000	86.8	13.2	26.8	1.2
5.06	105	1000	90.7	9.3	199.3	8.7
5.08	71	1000	93.5	6.5	193.9	8.4
5.62	313	1000	79.4	20.6	266.6	11.6
6.19	43	1000	97.0	3.0	247.1	10.7
5.48	52	1000	95.7	4.3	211.4	9.2
5.20	250	1000	81.1	18.9	233.7	10.2
5.70	180	1000	87.2	12.8	246.6	10.7
5.10	140	1000	88.1	11.9	208.0	9.0

Depths are estimated using a lithostatic pressure gradient of 23 MPa km⁻¹ ([Hautmann et al., 2013](#)).

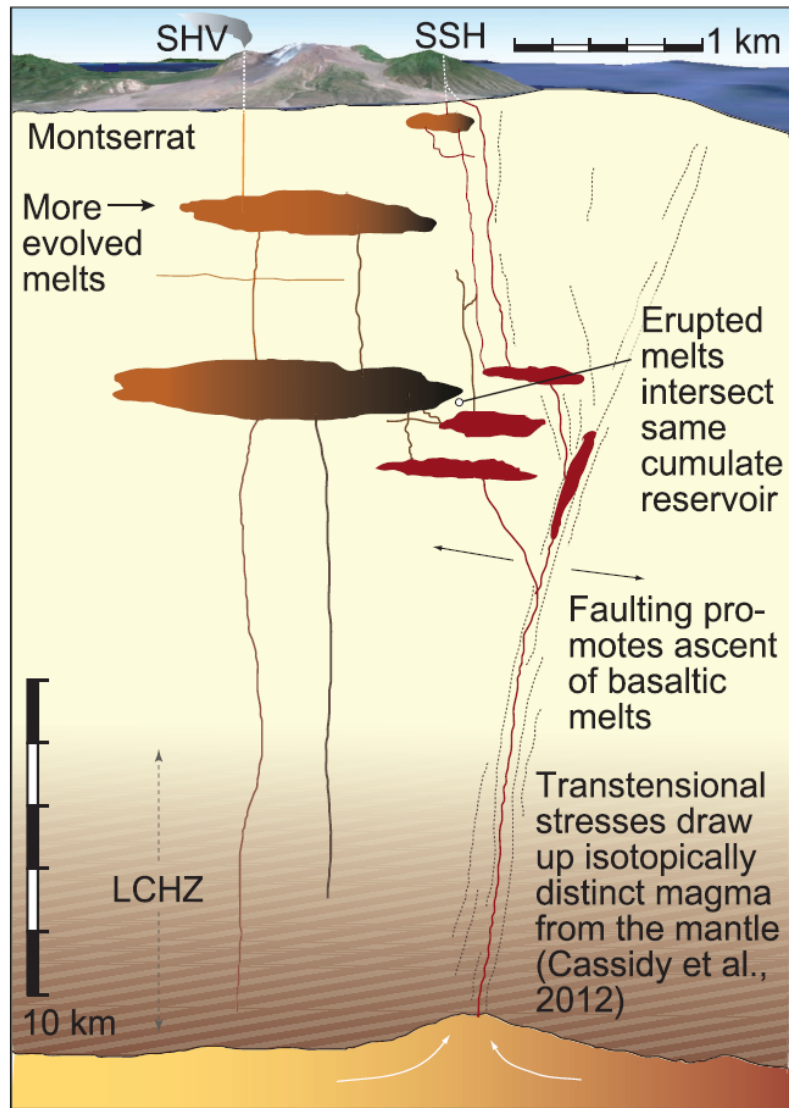


Fig. 12. Schematic figure showing how transtensional faulting can lead to the ascent of basaltic SSH magmas. The depths and processes involved in the generation of the SSH and SHV volcanism are shown. LCHZ, Lower Crustal Hot Zone of Annen *et al.* (2006).