

Durham Research Online

Deposited in DRO:

07 February 2017

Version of attached file:

Accepted Version

Peer-review status of attached file:

Peer-reviewed

Citation for published item:

Scarlato, P. and Mollo, S. and Del Bello, E. and von Quadt, A. and Brown, R.J. and Gutierrez, E. and Martinez-Hackert, B. and Papale, P. (2017) 'The 2013 eruption of Chaparrastique volcano (El Salvador) : effects of magma storage, mixing, and decompression.', *Chemical geology.*, 448 . pp. 110-122.

Further information on publisher's website:

<https://doi.org/10.1016/j.chemgeo.2016.11.015>

Publisher's copyright statement:

© 2016 This manuscript version is made available under the CC-BY-NC-ND 4.0 license
<http://creativecommons.org/licenses/by-nc-nd/4.0/>

Additional information:

Use policy

The full-text may be used and/or reproduced, and given to third parties in any format or medium, without prior permission or charge, for personal research or study, educational, or not-for-profit purposes provided that:

- a full bibliographic reference is made to the original source
- a [link](#) is made to the metadata record in DRO
- the full-text is not changed in any way

The full-text must not be sold in any format or medium without the formal permission of the copyright holders.

Please consult the [full DRO policy](#) for further details.

Manuscript Number: CHEMGE9903R1

Title: The 2013 eruption of Chaparrastique volcano (El Salvador): Effects of magma storage, mixing, and decompression

Article Type: Research paper

Keywords: Chaparrastique volcano; Central American volcanic arc; magma mixing; magma decompression

Corresponding Author: Professor. Silvio Mollo,

Corresponding Author's Institution: Sapienza - Università di Roma

First Author: Piergiorgio Scarlato

Order of Authors: Piergiorgio Scarlato; Silvio Mollo; Elisabetta Del Bello; Albrecht von Quadt; Richard J. Brown; Eduardo Gutierrez; Bettina Martinez-Hackert; Paolo Papale

Abstract: On December 29, 2013, an isolated vulcanian-type eruption occurred at Chaparrastique volcano (El Salvador) after 12 years of inactivity. The eruption was classified as VEI 2 and produced an ash plume with a maximum height of ~9 km. Textural and compositional data from phenocrysts from the erupted products have been integrated with geochemical and isotopic information from bulk rocks to elucidate the magmatic processes responsible for the reawakening of volcanic activity. Phenocrysts consist of Fo-rich poikilitic olivines hosting high-Mg titanomagnetites, and Fo-poor olivines coexisting with low-Mg titanomagnetites. Mineral-melt equilibria suggest an origin for the distinct phenocryst populations by mixing between a high-T (~1,130-1,150 °C), basaltic magma with fO₂ (NNO buffer) typical of the lower crust in arc systems and a low-T (~1,060-1,080 °C), basaltic andesitic magma with fO₂ (NNO+1 buffer) commonly encountered in shallower, more oxidized crustal reservoirs. Thermobarometry based on Fe-Mg exchange between orthopyroxene and clinopyroxene constrains the crystallization before eruption at relative low-P (~150-250 MPa) and low-T (~1,000-1,050 °C). Mixing between two chemically distinct magmas is also evidenced by the occurrence of reverse zoned plagioclase phenocrysts with resorbed sodic cores and re-growth of sieve-textured calcic mantles. Conversely, plagioclase rims exhibit disequilibrium compositions addressed to decompression kinetics (~10⁻³ MPa/s) driven by rapid magma ascent to the surface (~0.03 m/s).

Major and trace element modelling excludes fractional crystallization as the primary mechanism controlling the bulk rock variability, whereas geochemical data align along a mixing trend between two end-members representative of the primitive basalt and the differentiated basaltic andesite. Trace element and isotope data indicate that the primary source of magmatism is an enriched MORB-like mantle with the contribution of fluxes of metasomatic fluids and/or melts produced by the subducted slab. The role played by slab-fluid inputs of carbonate origin and slab-melts from the hemipelagic sediments seems to be minimal.

Assimilation/contamination processes of magmas by crustal rocks are also

negligible. In contrast, the geochemical signature of magmas is greatly influenced by slab-derived aqueous fluids produced prevalently by progressive dehydration of marine sediments and altered basaltic crust.

1 **The 2013 eruption of Chaparrastique volcano (El Salvador): Effects of magma**
2 **storage, mixing, and decompression**

3

4 Piergiorgio Scarlato¹, Silvio Mollo^{1,2}, Elisabetta Del Bello¹, Albrecht von Quadt³, Richard J.
5 Brown⁴, Eduardo Gutierrez⁵, Bettina Martinez-Hackert⁶, Paolo Papale¹

6

7 ¹Istituto Nazionale di Geofisica e Vulcanologia, Via di Vigna Murata 605, 00143 Rome, Italy

8 ²Dipartimento di Scienze della Terra, Sapienza-Università di Roma, P.le Aldo Moro 5, 00185 Roma,
9 Italy

10 ³Department of Earth Sciences, Institute of Geochemistry and Petrology, ETH, 8092 Zurich,
11 Switzerland

12 ⁴Department of Earth Sciences, Durham University, Science Site, DH1 3LE, UK

13 ⁵Observatorio ambiental, Ministerio de Medio Ambiente y Recursos Naturales, El Salvador

14 ⁶SUNY Buffalo State, Department of Earth Sciences and Science Education, Buffalo, USA,

15

16

17 Corresponding author:

18 Silvio Mollo

19 Sapienza-Università di Roma

20 Dipartimento di Scienze della Terra

21 P.le Aldo Moro 5

22 00185 Roma, Italy

23 e-mail silvio.mollo@uniroma1.it

24

25

26

27 **Abstract**

28 On December 29, 2013, an isolated vulcanian-type eruption occurred at Chaparrastique
29 volcano (El Salvador) after 12 years of inactivity. The eruption was classified as VEI 2 and
30 produced an ash plume with a maximum height of ~9 km. Textural and compositional data from
31 phenocrysts from the erupted products have been integrated with geochemical and isotopic
32 information from bulk rocks to elucidate the magmatic processes responsible for the reawakening of
33 volcanic activity.

34 Phenocrysts consist of Fo-rich poikilitic olivines hosting high-Mg titanomagnetites, and Fo-
35 poor olivines coexisting with low-Mg titanomagnetites. Mineral-melt equilibria suggest an origin
36 for the distinct phenocryst populations by mixing between a high- T (~1,130-1,150 °C), basaltic
37 magma with fO_2 (NNO buffer) typical of the lower crust in arc systems and a low- T (~1,060-
38 1,080 °C), basaltic andesitic magma with fO_2 (NNO+1 buffer) commonly encountered in shallower,
39 more oxidized crustal reservoirs. Thermobarometry based on Fe-Mg exchange between
40 orthopyroxene and clinopyroxene constrains the crystallization before eruption at relative low- P
41 (~150-250 MPa) and low- T (~1,000-1,050 °C). Mixing between two chemically distinct magmas is
42 also evidenced by the occurrence of reverse zoned plagioclase phenocrysts with resorbed sodic
43 cores and re-growth of sieve-textured calcic mantles. Conversely, plagioclase rims exhibit
44 disequilibrium compositions addressed to decompression kinetics (~ 10^{-3} MPa/s) driven by rapid
45 magma ascent to the surface (~0.03 m/s).

46 Major and trace element modelling excludes fractional crystallization as the primary
47 mechanism controlling the bulk rock variability, whereas geochemical data align along a mixing
48 trend between two end-members representative of the primitive basalt and the differentiated basaltic
49 andesite. Trace element and isotope data indicate that the primary source of magmatism is an
50 enriched MORB-like mantle with the contribution of fluxes of metasomatic fluids and/or melts
51 produced by the subducted slab. The role played by slab-fluid inputs of carbonate origin and slab-
52 melts from the hemipelagic sediments seems to be minimal. Assimilation/contamination processes

53 of magmas by crustal rocks are also negligible. In contrast, the geochemical signature of magmas is
54 greatly influenced by slab-derived aqueous fluids produced prevalently by progressive dehydration
55 of marine sediments and altered basaltic crust.

56

57 Keywords: Chaparrastique volcano; Central American volcanic arc; magma mixing; magma
58 decompression;

59

60

61

62

63

64

65

66

67

68

69

70

71

72

73

74

75

76

77

78

79 **1. Introduction**

80 El Salvador is one of the most seismically-active regions on Earth (Fig. 1a), being located
81 along the Central American volcanic front that is defined by ~50 major volcanic centers, including
82 composite or clustered volcanoes, distributed nearly continuously along 1,200 km from western
83 Guatemala to central Panama (e.g., Carr et al., 1979; Carr, 1984; MacMillan et al., 2004). The
84 Volcanic Cordillera in El Salvador is a line of volcanoes parallel to the coast and the offshore
85 subduction zone. El Salvador has 23 volcanoes and 5 volcanic fields. At present, only a few
86 volcanoes are active (outgassing and rarely ash emissions) such as the Santa Ana, San Salvador,
87 San Miguel and Izalco volcanoes (e.g., Agostini et al., 2006 and references therein).

88 Chaparrastique volcano, also known as San Miguel, is a composite volcano in central-
89 eastern El Salvador, ~15 km southwest of the city of San Miguel. A rural population of ~70,000
90 people live within 10 km of the summit crater of Chaparrastique, and San Miguel is the second
91 largest city in El Salvador. On December 29, 2013, after 12 years of inactivity, Chaparrastique
92 volcano erupted at 10:30 local time (16.30 GMT), prompting the evacuation of more than ~5,000
93 people living in ~3 km radius around the volcano (Martinez-Hackert et al., 2015). The eruption was
94 a vulcanian-type explosion that lasted 2.5 hours (Fig. 1b). The explosion was classified as VEI 2
95 and produced an ash plume with a maximum height of ~9 km (Martinez-Hackert et al., 2015). The
96 eruption column generated heavy ash fall in nearby areas downwind, especially in the towns of
97 Chinameca and San Jorge. A short lived pyroclastic density current generated from an early
98 Vulcanian explosion travelled ~500 m down the flanks of the volcano and entered coffee
99 plantations.

100 On January 2014, in response to a request of support by the government of El Salvador, the
101 INGV (Istituto Nazionale di Geofisica e Vulcanologia, Italy) organized a task force in close
102 collaboration with volcanologists from MARN (Ministerio de Medio Ambiente y Recursos
103 Naturales, El Salvador). A campaign survey was conducted over a period of ten days, with the aim
104 to install a monitoring network on the flanks of the volcano (Bonforte et al., 2015; Granieri et al.,

105 2015; Scarlato et al., 2014). At the same time, a suite of twelve eruptive products (i.e., fallout
106 tephra) was collected in correspondence of the seismic monitoring stations of VSMG, RANC, and
107 LACA located, respectively, at 1 km north, 2 km east, and 3 km south-east to the crater (Fig. 1c).

108 Here we present textural and mineral chemistry data from phenocrysts integrated with bulk
109 rock geochemical and isotopic data, with the aim to elucidate magmatic processes that triggered the
110 2013 eruption. Mineral-melt equilibria, and major and trace element modelling define the pre-
111 eruptive crystallization conditions of the system and indicate that magma mixing and magma
112 decompression were the most important mechanisms for triggering the eruption. The magma source
113 has been also constrained by trace element and isotope systematics. Magmas were generated by
114 partial melting of an enriched MORB-like mantle wedge with a dominant contribution of aqueous
115 fluids derived from the subducted slab.

116

117 **2. Geological setting**

118 The Pacific coastline of much of Central America is marked by a line of active and quiescent
119 volcanoes known as the Central American Volcanic Arc (CAVA) (e.g. Aubouin et al. 1982; Carr et
120 al. 1990, 2004; Leeman et al. 1994; Protti et al. 1995). To the south of Central America the Cocos
121 Plate, which underlies an area of the east Pacific, is being subducted beneath the Caribbean Plate
122 along the Middle American trench. The volcanoes result from the upward movement of magma
123 generated along the subduction zone between the Cocos and Caribbean tectonic plates. El Salvador,
124 and neighbouring Central American states, lies on the Caribbean Plate. Frequent earthquakes also
125 occur along the plate boundary. From Guatemala to northern Costa Rica, ~25 Ma old crust (formed
126 at the East Pacific Rise) is subducted at an angle varying from 55° in Guatemala and El Salvador to
127 65° in Nicaragua and northern Costa Rica (Protti et al. 1995; Syracuse and Abers 2006). The crustal
128 thickness, angle of subduction, and convergence rate contribute to the thermal structure of the
129 subduction zone, which is a major control on the generation of magmas in Central America (Davies
130 and Stevenson, 1992). The sediment input consists of a ~200-m-thick layer of hemipelagic clay

131 overlying a ~250-m-thick layer of carbonate ooze (Aubouin et al. 1982; Plank and Langmuir 1998).
132 The calc-alkaline compositions of volcanic rocks from CAVA record systematic along-strike
133 variations in many geochemical tracers of the slab, including Ba/La (Carr et al., 1990), U/Th
134 (Patino et al., 2000), B/Be (Leeman et al., 1994), and $\delta^{18}\text{O}$ (Eiler et al., 2005). For most of these
135 tracers, the peak in inferred slab flux occurs in the central portion of the CAVA, in Nicaragua, and
136 falls to nearly the global minimum to the southeast, at the terminal sector of volcanoes of Costa
137 Rica. Where the crust is thinner, high-Mg basalts are more abundant (e.g., Nicaragua). Conversely,
138 in correspondence of a thick and old continental crust, low-Mg basaltic lavas occur and are less
139 abundant (e.g., central and western Guatemala). Moreover, the continental crust of southern Central
140 America is relatively young and thin, compared to the crust of other continental margins, making it
141 less radiogenic (Carr 1984; Donnelly et al. 1990).

142 Chaparrastique volcano in eastern El Salvador is a 2,130 m-high symmetrical stratovolcano
143 with a broad, deep crater complex, which has frequently changed morphology during historic
144 eruptions (Escobar, 2003). Chaparrastique is considered one of the most active volcanoes in El
145 Salvador, with 26 eruptions in the past 500 years. Seismic activity is high, while surface activity
146 consists of pulsating gas emissions that ascend to maximum heights of ~200 m. On December 29,
147 2013, the volcano erupted ~80 vol.% of crystal-free juvenile material (shards of fresh glass and
148 crystal fragments) associated with ~20 vol.% of lapilli and scoria clasts. The eruption formed a new
149 spatter cone in the central crater. In the days that followed, the SO_2 emissions were estimated to
150 reach 2,800 tons per day. The increasing tremor and gas emissions also suggested migration of hot
151 fluids (gas, water, and magma) into the shallow plumbing system of the volcano. Before such event,
152 the eruptive history of Chaparrastique was characterized by the occurrence of modest VEI 1–2
153 eruptions (search for Chaparrastique at the MARN's webpage <http://www.marn.gob.sv>), mainly
154 attributed to spattering and scoria cones at the flanks, and minor Strombolian and ash eruptions in
155 the summit crater (Chesner et al., 2004). Since the 29 December eruption, Chaparrastique has
156 experienced periods (from May to July 2014) of increased seismic and degassing activity, landslides

157 inside the crater, and lahars. Occasionally, small steam/phreatomagmatic explosions and minor ash
158 falls have occurred, but no substantial explosive events have been reported. However, the level of
159 alert remains high and explosions of small to moderate size similar to the 29 December event
160 remain a possibility.

161

162 **3. Methods**

163 Textural and microchemical analyses were carried out at the HP-HT Laboratory of
164 Experimental Volcanology and Geophysics of the Istituto Nazionale di Geofisica e Vulcanologia in
165 Rome (Italy). Images were collected using the backscattered electron (BSE) mode of a field
166 emission gun-scanning electron microscopy (FE-SEM) Jeol 6500F equipped with an energy-
167 dispersive spectrometer (EDS) detector. Microchemical analyses were performed using an electron
168 probe micro-analyzer (EPMA) Jeol-JXA8200 combined EDS-WDS (five spectrometers with twelve
169 crystals) with 15 kV accelerating voltage and 10 nA electric current. For glasses, a slightly
170 defocused electron beam with a size of 3 μm was used, with a counting time of 5 s on background
171 and 15 s on peak (Table 1S). For crystals, the beam size was 1 μm with a counting time of 20 and
172 10 s on peaks and background, respectively (Table 1S). The following standards have been adopted
173 for the various chemical elements: jadeite (Si and Na), corundum (Al), forsterite (Mg), andradite
174 (Fe), rutile (Ti), orthoclase (K), barite (Ba), apatite (P), spessartine (Mn) and chromite (Cr). Sodium
175 and potassium were analyzed first to prevent alkali migration effects. The accuracy of the
176 microprobe was measured through the analysis of well-characterized synthetic oxides and mineral
177 standards. Data quality was ensured by analyzing these test materials as unknowns. Based on
178 counting statistics, analytical uncertainties relative to their reported concentrations indicate that
179 precision was better than 5% for all cations.

180 Bulk rock analyses for major and trace elements, as well as oxygen isotope analyses were
181 conducted at Actlabs (Activation Laboratories Ltd.). These analyses are reported in the
182 supplementary material (Table 2S), together with measured and certified analyses of international

183 reference materials. Major elements were analyzed by lithium metaborate/tetraborate fusion – ICP-
184 OES (inductively coupled plasma optical emission spectrometry). Sample washing was performed
185 to remove any organic, loosely adhered, and cementing material. Washing included soaking in a hot
186 mixture of HCl and H₂O₂ as well as cleaning in acetone using ultrasound. The analyses were
187 performed in a batch system. Each batch contained a method reagent blank, certified reference
188 material and 17% replicates. Samples were mixed with a flux of lithium metaborate and lithium
189 tetraborate and fused in an induction furnace. The molten material was immediately poured into a
190 solution of 5% nitric acid containing an internal standard, and mixed continuously until completely
191 dissolved (about 30 minutes). The samples were run on a Thermo Jarrell-Ash ENVIRO II ICP.
192 Calibration was performed using 7 prepared USGS and CANMET certified reference materials.
193 FeO was determined through titration, using a cold acid digestion of ammonium metavanadate, and
194 hydrofluoric acid in an open system. Ferrous ammonium sulphate was added after digestion and
195 potassium dichromate was the titrating agent. This cold digestion dissolved silicates and some
196 sulphides. Water content was determined by gravimetric method. About 0.3 g sample was thermally
197 decomposed in a resistance furnace in a pure nitrogen environment at 110 °C (moisture, H₂O⁻)
198 followed by decomposition at 1,000 °C (interstitial water, H₂O⁺), using an ELTRA CW-800,
199 directly releasing H₂O. Trace elements were measured by lithium metaborate/tetraborate fusion –
200 ICP-MS (inductively coupled plasma mass spectrometry). Samples fused as reported above were
201 diluted and analyzed by Perkin Elmer Sciex ELAN 6000, 6100 or 9000 ICP/MS. Three blanks and
202 five controls (three before sample group and two after) were analyzed per group of samples.
203 Oxygen isotopes are measured by reaction with BrF₅ at about 650 °C in nickel bombs. The
204 fluorination reaction converts O in the minerals to O₂ gas, which is subsequently converted to CO₂
205 gas using a hot C rod. All reaction steps are quantitative. Bulk rock isotopic analyses are performed
206 on a Finnigan MAT Delta, dual inlet, isotope ratio mass spectrometer. The data are reported in the
207 standard delta notation as per mil deviations from V-SMOW (hereafter reported to as SMOW).
208 External reproducibility is ±0.19‰ (1 s).

209 The isotopic compositions of Sr and Nd (Table 2S) were measured at the ETH Zürich using
210 a Triton Plus thermal ionization mass spectrometer (TIMS). Details of analytical techniques,
211 accuracy and precision can be found in von Quadt (1997). The isotope analysis set for Sr and Nd
212 was performed on bulk rock powder (50–100 mg). Values for $^{87}\text{Sr}/^{86}\text{Sr}$ were internally corrected for
213 fractionation using a $^{88}\text{Sr}/^{86}\text{Sr}$ value of 8.37521. Repeated measurements of the NBS 987 standard
214 for $^{88}\text{Sr}/^{86}\text{Sr}$ gave a mean value of 0.710252 ± 0.000002 (2σ). $^{143}\text{Nd}/^{144}\text{Nd}$ values were internally
215 corrected for mass fractionation using a $^{146}\text{Nd}/^{144}\text{Nd}$ value of 0.7219. No correction on Nd was
216 performed due to good segregation of Nd and Sm during column chemistry. Repeated
217 measurements on the Nd Merck standard for $^{146}\text{Nd}/^{144}\text{Nd}$ gave a mean value of $0.511730 \pm$
218 0.000001 (2σ). The analytical uncertainties associated to the Sr and Nd isotopic compositions of our
219 samples are in the range of 0.000006-0.000019 and 0.000002-0.000012, respectively.

220

221 **4. Mineralogy, petrology and geochemistry**

222 Mineral chemistry, petrology and geochemistry was performed in twelve samples of juvenile
223 material and scoria clasts ejected during the 2013 eruption. The juvenile material consists of white-
224 grey coarse ashes and dark-grey accretionary lapilli (i.e., CH1, CH2, CH4, CH5, CH8, CH10, CH11,
225 and CH12 samples), and are associated to centimetre-sized scoria clasts containing phenocrysts and
226 microlites surrounded by matrix glass (i.e., CH3, CH6, CH7, and CH9 samples). Mineral and
227 matrix glass analyses of scoria clasts are reported in Table 1S submitted as supplementary material.
228 Bulk rock analyses of the juvenile material and scoria clasts are reported in Table 2S submitted as
229 supplementary material. Major and trace elements were measured for all the bulk rock samples,
230 whereas radiogenic and stable isotopes were analyzed only for CH1, CH2, CH4, CH9, and CH12
231 samples that represent the entire bulk rock compositional variation.

232

233 4.1 Textures of scoria clasts

234 The paragenesis of scoria clasts is dominated by phenocrysts of olivine and plagioclase (~23 vol.%),
235 and minor pyroxene and spinel (~7 vol.%). The groundmass consists of plagioclase, pyroxene, and
236 spinel microlites (~12 vol.%), surrounded by matrix glass (~58 vol.%). The relatively low microlite
237 content suggests that, at the time of eruption, late crystallization was hampered by rapid quenching
238 of magma.

239 The habit of olivine is generally euhedral with well-formed planar edges. Larger olivine
240 phenocrysts are poikilitic and enclose sub-rounded spinels (Fig. 2a), whereas smaller olivines are
241 accompanied by spinel phenocrysts with irregular boundaries and glass embayments resulting from
242 the agglomeration of bleb-like grains (Fig. 2b).

243 Pyroxene crystallized as both sub-euhedral orthopyroxene and clinopyroxene.
244 Orthopyroxenes are less frequently observed but sometimes occur as large phenocrysts partly
245 enclosing small clinopyroxenes (Fig. 2c).

246 Plagioclase is characterized by complex zoning in which the cores, mantles and rims are
247 separated by sharp interfaces. BSE images reveal (1) dark-grey, partly corroded cores, (2) light-grey,
248 coarsely sieve-textured mantles, and (3) grey, euhedral to sub-euhedral rims (Fig. 2d-f). Notably,
249 detailed BSE images of the plagioclase rims show that their final euhedral habit results from the
250 attachment of a number of groundmass microlites growing as tabular crystals (Fig. 2g-i). These
251 small, elongated plagioclases attach preferentially parallel preferentially parallel to the longest
252 crystal dimension that normally corresponds to the [100] crystallographic direction contributing to
253 the textural maturation of the rim. In contrast, plagioclase mantles are characterized by the
254 occurrence of micrometre-sized glass channels and hollows, sometimes, in full communication with
255 the matrix glass when the growth of the crystal rim results incomplete.

256

257 4.2 Mineral and glass chemistry of scoria clasts

258 Olivine is normally zoned with forsterite (Fo) contents decreasing from core-to-rim. Larger
259 poikilitic olivines enclosing spinels show Fo₇₇₋₇₈ cores and Fo₇₄₋₇₆ rims (Fig. 2a). The olivine core is

260 also more calcic (0.51-0.63 wt.% CaO) than the rim (0.32-0.56 wt.% CaO). Conversely, smaller
261 olivines mutually touching with spinels in the matrix glass are weakly zoned with Fo₇₄₋₇₅ cores and
262 Fo₇₂₋₇₃ rims (Fig. 2b).

263 Spinel is titanomagnetite with variable ulvospinel content (Usp). Smaller, sub-rounded
264 titanomagnetites enclosed in olivine are Usp₂₄₋₂₇ (Fig. 2a), whereas single, isolated titanomagnetites
265 are Usp₂₁₋₁₉ (Fig. 2b). Titanomagnetites in poikilitic olivines are also more primitive (~6-7 wt.%
266 MgO) than those found in the matrix glass (~4-5 wt.% MgO).

267 Pyroxene is both orthopyroxene and clinopyroxene. Orthopyroxene exhibits weak chemical
268 variation with a narrow enstatite (En) content of En₆₀₋₆₃ (Fig. 2c). Clinopyroxene is augite (cf.
269 Morimoto et al., 1988) with relative low-CaO concentration (<17 wt.%) and variable diopside (Di)
270 content ranging from Di₃₇ to Di₄₆ (Fig. 2c).

271 Among all the minerals, plagioclase phenocrysts are characterized by important
272 compositional changes reflecting the complex textural characteristics. Plagioclases are strongly
273 reversely zoned with resorbed sodic cores surrounded by dusty sieve-textured calcic mantles. This
274 suggests an origin by resorption and rapid re-growth due to mixing between two compositionally
275 distinct magmas. The anorthite (An) content sharply increases from core (An₅₃₋₆₅) to mantle (An<sub>80-
276 84</sub>). Conversely, at the final stage of the crystal growth, a thin-to-thick rim occurs with chemistry
277 (An₇₇₋₈₀) intermediates between core and mantle compositions (Fig. 2d, e, f). Importantly, the rim
278 chemistry is almost identical to that of groundmass plagioclases, suggesting coeval crystallization
279 (Fig. 2g, h, i).

280 The matrix glass is 52.7-54.3 wt.% SiO₂ and 3.5-4.6 wt.% Na₂O+K₂O, corresponding to the
281 composition of a basaltic andesite (i.e., Le Bas et al., 1986). However, Mg# (100 × molecular
282 MgO/FeO_{tot} + MgO) values are highly variable ranging from 25 to 43. Indeed, the glass analyses
283 are not homogeneous showing concentration gradients especially next to the well-developed rims of
284 large plagioclase phenocrysts (see discussion below). Within such chemical variability, the

285 composition of the matrix glass does not differ substantially from that of micrometre-sized glass
286 channels and glass inclusions within plagioclase phenocrysts.

287

288 4.3 Bulk rock geochemistry

289 On anhydrous basis, the products are calc-alkaline basalts (CH1-CH2) and moderately
290 (CH3-CH6) to more differentiated (CH7-CH12) basaltic andesites (i.e., Le Bas et al., 1986; Fig. 3a)
291 with medium-K affinity (i.e., LeMaitre, 2002; Fig. 3b). Bulk rocks align along one single trend in
292 which both silica and alkali explore limited compositional variations (Fig. 3a-b). From CH1 to
293 CH12, MgO and CaO linearly decrease from ~7 to ~3 wt.% and from ~11 to ~8 wt.%, respectively,
294 reflecting the more evolved character of the samples.

295 The chondrite-normalized pattern (Sun and McDonough, 1989) of REE (rare earth elements;
296 Fig. 3c) exhibits sub-parallel trends shifting towards progressive REE enrichments from CH1 to
297 CH12. Slight enrichments in LREE (light rare earth elements) relative to HREE (heavy rare earth
298 elements) are also observed. Bulk rocks are characterized by weak negative Eu anomaly ($\text{Eu}/\text{Eu}^* =$
299 $0.86\text{-}0.92$) and relatively low values of normalized La/Yb ratios (2.8-3.1). The primordial mantle-
300 normalized pattern (Sun and McDonough, 1989) of trace elements (Fig. 3d) shows typical features
301 of arc magmas, with LILE (large ion lithophile elements) to HFSE (high field strength elements)
302 ratios higher than the primitive mantle values. This is confirmed by low Nb and Ta concentrations
303 and correspondingly high Ba/Ta (2,900-4,200) and La/Ta (62-80) ratios. The incompatible elements
304 generally increase with increasing SiO_2 . Moreover, trace element patterns show positive spikes of
305 Ba, U, K and Sr, and negative Nb, Ta, Ti, Hf and Zr anomalies. The compatible TE (transition
306 elements) such as Ni (11-20 ppm) and Cr (20-58) are low and their concentration decreases from
307 CH1 to CH12, in accordance with decreasing MgO.

308 Radiogenic isotope compositions of bulk rocks are constrained in narrow intervals of
309 $0.70377\text{-}0.70387$ and $0.51291\text{-}0.51299$ for $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, respectively. According
310 to isotope data from other volcanoes in El Salvador, the lack of sensible isotopic variations with the

311 degree of magma evolution indicates that crustal contamination processes are likely to be minimal
312 (Agostini et al., 2006).

313 Oxygen-isotope ratios range between 5.4 and 5.6 ‰, showing values comparable to most
314 mantle peridotites ($\sim 5.5 \pm 0.2\text{‰}$; e.g., Matthey et al., 1994). However, it is important to note that the
315 oceanic lithosphere is highly heterogeneous in $\delta^{18}\text{O}$ and may contribute with slab-derived
316 components with oxygen-isotope ratios (0-6‰) equal to or lower than typical upper mantle
317 peridotites, i.e., hydrothermally altered lower oceanic crust and ultramafic rocks such as
318 serpentinites (Cocker et al., 1982; Staudigel et al., 1995).

319

320 **5. Discussion**

321 5.1 Magma crystallization conditions

322 In order to decipher the crystallization conditions of the erupted products, the olivine-based
323 thermometer of Putirka (2005a) with a relatively low uncertainty of ± 27 °C has been used. Olivine-
324 melt equilibria are particularly useful for temperature estimations because the Fe-Mg exchange
325 reaction is nearly constant over a wide range of temperatures, bulk compositions and $f\text{O}_2$ conditions
326 (i.e., $^{ol-melt}K_{\text{Fe-Mg}} = 0.30 \pm 0.03$), and because the olivine Fo content is highly sensitive to the
327 thermal path of magma (e.g., Roeder and Emslie, 1970). Assuming that olivine is the liquidus phase,
328 its early-formed core is expected to be in equilibrium with the original magma represented by the
329 bulk rock analysis (cf. Putirka 2008). Through this approach, it is found that Fo₇₈₋₇₇ olivines are in
330 equilibrium with the primitive basalts (MgO > 6 wt.%), yielding onset temperatures of 1,135-
331 1,146 °C (Fig. 4a). Conversely, Fo₇₄₋₇₅ olivines are in equilibrium with the more evolved basaltic
332 andesites (MgO < 4 wt.%), recording onset temperatures of 1,067-1,084 °C (Fig. 4a). The
333 equilibrium condition is attained when the total Fe of the melt is partitioned between Fe²⁺ and Fe³⁺
334 through the formulation of Kress and Carmichael (1991), over buffering conditions relative to NNO
335 (nickel-nickel oxide) and NNO+1, respectively for Fo₇₈₋₇₇ and Fo₇₄₋₇₅ olivines. These different redox
336 states have been predicted by the spinel-melt oxygen barometer of Ariskin and Nikolaev (1996),

337 using high-Mg titanomagnetites hosted in Fo₇₈₋₇₇ poikilitic olivines (Fig. 2a) and low-Mg
338 titanomagnetites coexisting with Fo₇₄₋₇₅ olivines (Fig. 2b). The distinct textural and chemical
339 characteristics for olivine and titanomagnetite suggests mixing between a high-*T*, mafic melt with
340 *f*O₂ typical of the lower crust of volcanic arcs (Brounce et al., 2014) and a low-*T*, differentiated melt
341 with *f*O₂ commonly encountered in shallower, slightly more oxidized crustal reservoirs (see below).

342 With respect to plagioclase crystallization, the equilibrium model of Namur et al. (2012)
343 indicates that the more sodic (An₅₃₋₆₅) core and the more calcic (An₈₀₋₈₄) mantle approach to
344 equilibrium with more evolved (CaO ≤ 8 wt.%) and primitive (CaO ≥ 9 wt.%) bulk rocks,
345 respectively (Fig. 4b). Conversely, the equilibrium condition is never found between the
346 intermediate (An₇₇₋₈₀) plagioclase rim and the overall bulk rock compositions. The plagioclase rim
347 is also in disequilibrium with the surrounding matrix glass (Fig. 4b) and data deviate from the one-
348 to-one straight line, responding to moderate An enrichments in the plagioclase rim (Δ An = 16-27%).
349 The thermometer of Putirka (2005b) with uncertainties of ±36 °C, yields crystallization
350 temperatures of 1,051-1,063 °C, 1,119-1,129 °C, and 1,098-1,117 °C for plagioclase core, mantle
351 and rim, respectively. These predictions account for minimum H₂O-melt concentration of 2 wt.%
352 derived by the difference to 100% of the total of electron microprobe analyses of matrix glasses (cf.
353 Devine et al., 1995). It has been also tested that H₂O contents up to 4 wt.% changes the temperature
354 estimates by ±32 °C that, however, remains within the relatively high uncertainty of the
355 thermometer. The occurrence of low-*T*, partly resorbed cores surrounded by high-*T*, sieve-textured
356 mantles (Δ *T* ~70 °C) confirms mixing between two compositionally distinct magmas. It is therefore
357 inferred that a colder, basaltic andesitic magma reservoir was intruded by a hotter, basaltic magma
358 rising from depth. The increase in An content for the plagioclase mantle supports the hypothesis
359 that the new magma was compositionally more primitive and possibly hotter and/or wetter. The
360 textural evidence of dissolution and patchy regions at the inner part of the plagioclase corroborates
361 the mixing mechanism (e.g., Tsuchiyama and Takahashi, 1983; Tsuchiyama, 1985; Tepley et al.,
362 2000). Therefore, more evolved Fo₇₄₋₇₅ olivines and An₅₃₋₆₅ plagioclases crystallized from a colder

363 basaltic andesitic magma that, subsequently, mixed with the hotter basaltic magma rising from
364 depth and carrying more primitive Fo₇₈₋₇₇ olivines. Pre-existing less calcic An₅₃₋₆₅ plagioclases in
365 contact with the recharging basaltic magma experienced an increase in temperature that resulted in
366 resorption textures (Fig. 3d-f). While the recharging and host magmas attempted to hybridize, high-
367 temperature gradients lead to relatively rapid crystallization of sieve-textured, more calcic An₈₀₋₈₄
368 mantles with the contribution of the more primitive component of the mixing system (e.g., Hibbard,
369 1981; Shcherbakov et al., 2010). Coherently with this scenario, the weak negative europium
370 anomaly of the bulk rocks excludes substantial plagioclase fractionation at depth (Fig. 3c).

371 By means of the two-pyroxene model of Putirka (2008), it is found that orthopyroxene and
372 augite exhibit Fe-Mg exchange values (i.e., $^{cpx-opx}Kd_{Fe-Mg} = 1.09 \pm 0.14$; Fig. 4c, d) indicative of
373 equilibrium crystallization from the hybrid magma from which pyroxenes formed immediately
374 before eruption. Temperature and pressure estimates based only on mineral chemistry (i.e.,
375 independently on melt composition) provide values of 1,012-1,053 °C (uncertainty of ± 38 °C) and
376 158-253 MPa (uncertainty of ± 200 MPa), respectively (Fig. 4c-d). The temperature estimates are
377 lower than those obtained for the more evolved Fo₇₄₋₇₅ olivine and An₅₃₋₆₅ plagioclase, evidencing a
378 late stage of pyroxene crystallization from the more oxidized hybrid magma. According to Grove
379 and Juster (1989), if fO_2 of calc-alkaline magmas increases to NNO+1, the crystallization of
380 orthopyroxene + augite occurs at the expense of pigeonite (never observed in samples from this
381 study) along a single reaction boundary in which olivine + liquid = orthopyroxene + augite +
382 plagioclase. In addition, the formation of a clinopyroxene with relative high-Ca contents is also
383 limited by previous incorporation of Ca in plagioclase.

384 Considering the wide spectrum of equilibrium experiments from the literature, the calc-
385 alkaline melts studied by Foden and Green (1992) show compositions comparable to those from
386 this study. At $P \geq 500$ MPa and H₂O < 2 wt.%, plagioclase is the liquidus phase followed by olivine
387 or, alternatively, pyroxene if $P \geq 700$ MPa. However, at $P \geq 500$ MPa and H₂O ≥ 2 wt.%, the
388 stability field of liquidus olivine expands significantly, so that plagioclase does not crystallize from

389 the melt. On the other hand, at $P = 200$ MPa and $H_2O = 2-4$ wt.%, the crystallization sequence
390 consists of olivine, olivine + plagioclase, and olivine + plagioclase + pyroxene (Fig. 5). This latter
391 phase assemblage closely agrees with the textural relationships and saturation temperatures of
392 natural minerals from this study, evidencing primary nucleation and growth of olivine + plagioclase,
393 and subordinated formation of pyroxene. The lack of amphibole in the eruptive products is also in
394 accordance with a low- P crystallization path in which the amphibole saturation surface is
395 encountered only at $T < 950$ °C (Fig. 5).

396 Generally, El Salvador arc lavas crystallize at $P \leq 500$ MPa, NNO-NNO+1 buffer, and 3–4
397 wt.% H_2O (Agostini et al., 2006). The shallow residing conditions of magmas are corroborated by
398 the low La/Yb ratios of bulk rocks, excluding development of large magma chambers in a thick
399 crust with HREE-bearing minerals. Moreover, melt- H_2O contents of 2-4 wt.% are frequently
400 documented for arc magmas in Central America (e.g., Sisson and Layne, 1993; Roggensack et al.,
401 1997; Roggensack, 2001), accounting for the remarkable fluid contribution to the mantle wedge
402 from the subducting slab (e.g., Patino et al., 2000). Wade et al. (2006) observed that the correlation
403 between melt- H_2O contents and Ba/La ratios may be significant. On this basis, Sadofsky et al.
404 (2007) proposed a regression fit in which the ratio of Ba/La can be used as a proxy for H_2O in
405 magma. Through this equation it is determined that Ba/La ratios of bulk rocks from this study
406 correspond to original melt- H_2O contents of 3.3-3.7 wt.%, coherently with previous estimates for El
407 Salvador arc lavas by Agostini et al., (2006), and the low- P crystallization sequence experimentally-
408 derived by Foden and Green (1992). Notably, LOI (Loss-On-Ignition) from bulk rock analyses
409 (Tables 2S) ranges from 1.1-2.1 wt.% (considering total iron as FeO) suggesting overall outgassing
410 of magma during eruption to the surface and pressure drops down to atmospheric condition (Aloisi
411 et al., 2006; Toramaru et al., 2008; Rutherford, 2008; Gonnermann and Manga, 2013).

412

413 5.2 Magma mixing vs. magma fractional crystallization

414 Magma mixing is recognized as a widespread process in calc-alkaline, arc-related magmas that
415 contributes to magma chamber evolution and is a likely trigger for eruptions (Sparks et al. 1977;
416 Eichelberger et al. 2000). Previous petrological studies on volcanoes along the CAVA have
417 documented the role of magma mixing as the cause of substantial changes in major and trace
418 element compositions (e.g., Streck et al., 2002; Richer et al., 2004). In particular, magma mixing
419 has been observed for a number of volcanoes in El Salvador, generally between basaltic andesitic
420 and andesitic magmas (e.g., San Vicente volcano), and between basaltic andesitic and rhyodacitic
421 magmas (e.g., Ilopango volcano), leading to a variety of complex rock textures characterized by
422 corroded/resorbed plagioclases and bimodal compositional distributions of phenocrysts (e.g., Carr
423 et al., 2007 and references therein).

424 If magma mixing occurred at the time of Chaparrastique eruption, bulk rock compositions
425 are expected to plot on tie-lines between CH1 and CH12 end-members, resulting to be rather
426 different from the curved trajectories caused by fractional crystallization. To test this hypothesis,
427 MELTS (Ghiorso and Sack, 1995) simulations have been conducted using the most primitive CH1
428 basalt as a starting liquid (Table 3s submitted as supplementary material). Thermodynamic runs
429 started from the superliquidus temperature of 1,170 °C and continued along a fractional
430 crystallization path that terminated at 1,020 °C. The pressure, melt-H₂O content and oxygen
431 fugacity were set at 200 MPa, 3 wt.%, and NNO+1, respectively, as the best input parameters
432 reproducing the natural phase assemblage. Results show that the curved MELTS trajectories do not
433 match in full with the natural Chaparrastique compositions that, in turn, align along tie-lines
434 between CH1 and CH12 end-members (Fig. 6a). This is also true for the behaviour of trace
435 elements modelled by the Excel spreadsheet of Ersoy and Helvacı (2010). The original partition
436 coefficients set in the spreadsheet were used to derive the fractional crystallization vectors for REE
437 (Fig. 6b), HFSE (Fig. 6c), and TE (Fig. 6d). While these vectors describe an implausible magmatic
438 evolution for the Chaparrastique products, the bulk rock variability is successfully modelled by
439 mixing (straight) lines joining CH1 and CH12 end-members (Fig. 6b-d).

440

441 5.3 Magma decompression and ascent to the surface

442 Plagioclase phenocrysts from Chaparrastique eruption exhibit partly resorbed An₅₃₋₆₅ cores that are
443 the herald of early equilibrium in a basaltic andesitic magma chamber, and sieve-textured An₈₀₋₈₄
444 mantles interpreted as markers of further basaltic recharge (Fig. 2d, e, f and Fig. 4b). If the mixing
445 efficiency is low, the influx of mafic magma, possibly transferring heat and H₂O, causes inverse
446 zoning of plagioclase due to dissolution and overgrowth of highly calcic crystals (e.g., Tsuchiyama
447 1985; Davidson and Tepley 1997). Conversely, if the system rapidly re-equilibrates to form a
448 homogeneous hybrid melt, an overall normally zoned trend is resumed with newly-formed calcic
449 cores grading into sodic rims (e.g., Izbekov et al. 2004; Neill et al., 2015). Apparently, this latter
450 mechanism seems to explain the final growth of An₇₇₋₈₀ plagioclase rims and the crystallization of
451 groundmass microlites (Fig. 3d-i). However, these late-stage plagioclases are not in equilibrium
452 with the hybrid melt if its composition is assumed to correspond with the final matrix glass (Fig.
453 4b). Moreover, chemical profiles carried out far enough away from the plagioclase surface reveal
454 that the matrix glass is not homogeneous (Table 4s), due to the development of chemical gradients
455 at the crystal-melt interface (Fig. 7a).

456 Numerous cooling and decompression studies explored the effect of magma mixing and
457 undercooling on the crystallization kinetics of plagioclase (e.g., Kirkpatrick et al., 1976; Smith &
458 Lofgren, 1983; Muncill & Lasaga, 1987; Sato, 1995; Hammer & Rutherford, 2002; Couch, 2003;
459 Pupier et al., 2008; Fiege et al., 2015). According to experimentally-derived data (Mollo et al., 2011,
460 2012a), if crystal growth rate is large compared with the rate of component diffusion in the melt, a
461 diffusive boundary layer develops causing enrichments in elements incompatible with the
462 plagioclase crystal lattice (Fig. 7a). Concentration-dependent partitioning produces plagioclases
463 that respond to chemical gradients in the melt, so that crystals incorporate high Ti, Mg, and Fe
464 proportions (Fig. 7a) and become preferentially enriched in An with respect to equilibrium
465 crystallization (Fig. 4b). The effect of undercooling due to temperature decrease (cf. Mollo et al.,

466 2011) and/or H₂O exsolution (cf. Brugger & Hammer, 2010) leads to the formation of
467 disequilibrium An-rich plagioclases in response to the drastic change of the topology of the Ab-An
468 plagioclase binary loop in which the temperature difference between liquidus and solidus is
469 minimized (cf. Iezzi et al., 2014). It is interesting to note that disequilibrium plagioclases
470 crystallizing from rapidly decompressed calc-alkaline melts in laboratory show An enrichments
471 ($\Delta\text{An} = 12\text{-}30\%$; data from Brugger & Hammer, 2010) very close to those ($\Delta\text{An} = 16\text{-}27\%$)
472 measured in this study. The control of undercooling is also evidenced by the attachment of
473 groundmass microlites to form the plagioclase rim (Fig. 3g-i). This crystal growth mechanism has
474 been documented under kinetically-controlled conditions by *in-situ* experimental observations
475 (Schiavi et al., 2009), crystal size distribution analysis (Iezzi et al., 2011; Lanzafame et al., 2013),
476 and electron-microprobe mapping (Iezzi et al., 2014). The attachment mechanism is possibly
477 explained by heterogeneous nucleation events on the pre-existing crystal surface (Mollo et al.,
478 2012b; Vetere et al., 2015) and/or by the emerging theory of aggregation by self-orientation of sub-
479 micrometric crystals (Teng et al., 2013 and references therein).

480 An important consequence of the diffusion-controlled growth is that the partition coefficient
481 measured between the advancing plagioclase surface and the diffusive boundary layer changes upon
482 the effect of chemical gradients in the melt. The partition coefficient is expressed as:

483

$$484 \quad K_i = C_i^{xls} / C_i^{melt} \quad (1)$$

485

486 where C_i^{xls} is the concentration of a chemical element i in the crystal and C_i^{melt} is the concentration
487 of the same element in the melt. If the compositions of both crystal and melt result from interface-
488 controlled growth and are supposed to be in equilibrium with respect to element exchange at the
489 crystal-melt interface, Eqn. (1) provides the ‘true’ partition coefficient K_i^{true} . Conversely, the effect
490 of diffusion-controlled growth is to change K_i^{true} in an ‘apparent’ partition coefficient K_i^{app} when the

491 chemistry of the advancing crystal surface reflects the chemical concentrations in the diffusive
 492 boundary layer. During plagioclase disequilibrium growth driven by undercooling, it has been
 493 experimentally-derived that $K_{Fe}^{app} > K_{Fe}^{true}$ and $K_{Mg}^{app} > K_{Mg}^{true}$ (Mollo et al., 2011). For the case of
 494 natural plagioclase phenocrysts, the lattice strain equation of Dohmen & Blundy (2014) represents a
 495 powerful tool to predict whether or not the partitioning of Mg is controlled by equilibrium or
 496 disequilibrium growth conditions. The model was derived by thermodynamic principles and allows
 497 to predict K_{Fe}^{true} for a wide range of equilibrium plagioclase and melt compositions. The partitioning
 498 of Mg measured between sodic core (or calcic mantle) of plagioclase and CaO-poor (or CaO-rich)
 499 bulk rocks yields values close to those predicted by the lattice strain equation of Dohmen & Blundy
 500 (2014). It is worth noting that $\ln K_{Mg}$ increases linearly with $1/T$, according to thermodynamic
 501 principles of cation partitioning (Fig. 7b). In contrast, partition coefficients measured as the ratio
 502 between plagioclase rim and matrix glass show values much higher than those expected for
 503 equilibrium crystallization (Fig. 7b), as for the case of $K_{Mg}^{app} > K_{Mg}^{true}$ due to the effect of
 504 undercooling (Mollo et al., 2011). If the advancing plagioclase rim does not violate the condition of
 505 crystallization into an infinite melt reservoir, the value of K_{Mg}^{app} can be modelled as a function of
 506 melt and crystal parameters (Smith et al., 1955):

507

$$\begin{aligned}
 508 \quad K_{Mg}^{app} = & 1/2 \left\{ 1 + \operatorname{erf} \left[\sqrt{G^{pl} x / D_{Mg}^{melt}} / 2 \right] \right\} \\
 & + (2K_{Mg}^{true} - 1) \exp \left\{ \left[-K_{Mg}^{true} (1 - K_{Mg}^{true}) (G^{pl} x / D_{Mg}^{melt}) \right] \operatorname{erfc} \left[(2K_{Mg}^{true} - 1) / 2 \sqrt{G^{pl} x / D_{Mg}^{melt}} \right] \right\}
 \end{aligned} \tag{2}$$

509

510 where K_{Mg}^{true} is the equilibrium partition coefficient provided by the formulation of Dohmen &
 511 Blundy (2014), G^{pl} is the plagioclase growth velocity that is unknown, D_{Mg}^{melt} is the diffusivity of
 512 Mg in the melt (i.e., 9×10^{-11} cm²/s derived by the model of Zhang et al., 2010), and x is the

513 distance the plagioclase interface moves into the melt (i.e., the linear dimension of the plagioclase
514 rim). Results based on the algebraic rearrangement of Eqn. (2) show that K_{Mg}^{app} measured in this
515 study can be ascribed to rapid plagioclase growth velocities of $\sim 2\text{--}7 \times 10^{-8}$ cm/s (Fig. 7b; see also
516 Table 5s for the entire numerical dataset). Decompression-induced crystallization experiments of
517 Brugger & Hammer (2010) have demonstrated that, in calc-alkaline melts, the growth velocity of
518 plagioclase scales with the decompression path of magma (ΔP). It is found that $\sim 10^{-3}$ MPa/s ΔP
519 corresponds to $\sim 10^{-8}$ cm/s G^{pl} . If magma decompression started at $P = 200$ MPa (Fig. 5) over a
520 timescale of $\sim 10^{-3}$ MPa/s ΔP , the ascent velocity of magma feeding eruption was ~ 0.03 m/s (i.e.,
521 100 MPa ≈ 3 km), in agreement with values estimated for a number of volcanic eruptions
522 worldwide (Aloisi et al., 2006; Toramaru et al., 2008; Rutherford, 2008; Gonnermann and Manga,
523 2013; Mollo et al., 2015).

524

525 5.4 Magma genesis and slab inputs

526 While partial melting of the mantle wedge is the predominant source of magmatism in Central
527 America, the geochemical signature of arc magmas is highly influenced by the nature of the
528 subducted slab components (e.g., Carr et al., 2004). An enriched MORB-like mantle has been
529 identified as the primary source for CAVA magmas with the contribution of fluxes of metasomatic
530 fluids and/or melts produced by the crustal units of the subducting Cocos Plate, i.e. the basal altered
531 MORB unit, the middle carbonate unit, and the upper hemipelagic unit (Patino et al., 2000). A
532 generalized approach for identifying the transfer agents between the subducted slab and the mantle
533 wedge is based on the knowledge that ratios of some specific incompatible elements weakly change
534 over the range of mantle melting and/or during assimilation \pm fractional crystallization processes. In
535 contrast, the same incompatible element ratios are greatly affected by the contribution of the
536 subducting slab, being distinctive indicators of sediment input into magma genesis (e.g., Cameron
537 et al., 2003).

538 Considering the different geochemical signature of the hemipelagic and carbonate sediments
539 subducted beneath Central America, the use of U/La and Ba/Th ratios has the advantage to clearly
540 separate the two slab components (Patino et al., 2000). These ratios correspond to soluble/insoluble
541 element pairs with the important feature that U/La is higher in the hemipelagic sediments and Ba/Th
542 is higher in the carbonate sediments (Fig. 8a). Using partition coefficients from literature and the
543 modal batch melting equation, Cameron et al. (2013) modelled the role played by melts and fluids
544 derived from the subducted slab for the partial melting of the enriched MORB source (Fig. 8a).
545 Modelling results indicate that the majority of primitive magmas at the CAVA have relatively low
546 U/La and Ba/Th ratios and cluster near the enriched MORB source. Ba-rich fluids from the
547 carbonate sediments produce magmas with clear slab-fluid signatures ($>0.1\%$) corresponding to
548 high Ba/Th ratios and low U/La ratios observed in Nicaragua and addressed to an increased dip of
549 the subducting Cocos Plate (Patino et al., 2000). In contrast, as the degree of differentiation
550 proceeds, volcanoes in Costa Rica and El Salvador are weakly influenced by slab-fluid inputs of
551 carbonate origin (Fig. 8a). In El Salvador, Ba/Th and U/La ratios of magmas show a low
552 contribution of melts ($<0.04\%$) from the subducted hemipelagic sediments and a relative high
553 participation of fluids ($<2\%$) from the mature oceanic crust (Fig. 8a).

554 Additional important information concerning the genesis of Central American magmatism
555 can be retrieved by using ratios of elements that are not highly fluid mobile in arc systems, such as
556 REE and HFSE, in order to better evaluate the role of slab components with melt-like properties
557 (Fig. 8b). Volcanic products in Nicaragua form a nearly vertical trend at low La/Sm extending
558 towards the highest Ba/La ratios with respect to other CAVA magmas (Fig. 8b). Taken together
559 with the discussion of Ba/Th and U/La ratios, this finding corroborates the addition to the mantle
560 wedge of variable amounts of Ba-rich fluids derived from subducted carbonate sediments.
561 Conversely, volcanic products in Costa Rica are characterized by the highest La/Sm and La/Yb
562 ratios due to remarkable LREE enrichments (Fig. 8b) but among the lowest Ba/La and Ba/Th ratios,
563 suggesting either involvement of an enriched mantle wedge or a melt-like component derived from

564 the subducted slab (Sadofsky et al., 2007). Volcanic products in El Salvador are intermediate
565 between the geochemical signature of magmas in Nicaragua and Costa Rica, being characterized by
566 relative low La/Sm and moderate Ba/La ratios due to pronounced slab-fluid signals from marine
567 sediments (Fig. 8b).

568 For discriminating between crustal and source contamination of CAVA magmas, Cameron
569 and Walker (2006) modelled the oxygen and strontium isotopes in order to define mixing
570 trajectories between the sediment-modified mantle and crustal end-members (Fig. 8c). For most
571 rocks in Guatemala produced by back-arc volcanism behind the volcanic front, there is clear
572 evidence of crustal contamination (>25%). This accounts for the increased continental crust
573 thickness and the highly radiogenic nature of the old crust (cf. Carr et al., 1990; Walker et al., 1995).
574 In contrast, in Central America, crustal assimilation causes weak changes in the geochemical
575 characteristics of magmas (Fig. 8c). This is particularly true for volcanoes in El Salvador,
576 Nicaragua, and Costa Rica, emplaced in a thin and young continental crust (cf. Eiler et al., 2005)
577 and characterized by magmatism with source contamination signature (<5%).

578 Sr and Nd systematics presented by Cameron et al. (2002) confirm that magmas in
579 Guatemala extend in the direction of high degrees of crust contamination (Fig. 8d). On the other
580 hand, volcanic products in El Salvador, Nicaragua, and Costa Rica define an array with positive
581 slope (Fig. 8d) as the result of a mix between an enriched MORB source and the subducted slab
582 (Feigenson and Carr, 1986; Carr et al., 1990). The slab source itself represents a mix between a
583 depleted MORB source and sediments plus altered basaltic crust (Fig. 8d). Boron isotope data by
584 Tonarini et al. (2007) for El Salvador products acknowledge the importance of slab-derived aqueous
585 fluids as metasomatizing agents of the mantle wedge. These aqueous fluids are mostly produced by
586 dehydration of marine sediments and altered basaltic crust of the subducted slab, but fluxes of
587 additional fluids result also from progressive dehydration of the serpentized mantle wedge dragged
588 beneath the volcanic arc (Tonarini et al., 2007).

589

590 **6. Conclusions**

591 This study documents for the first time the mineralogical, petrological and geochemical
592 characteristics of products erupted at during the 2013 eruption of Chaparrastique volcano. Mineral-
593 melt equilibria and bulk rock modelling indicate that the eruption was likely triggered by the input
594 of hotter, basaltic magma from depth into a colder, basaltic andesitic reservoir residing at shallow
595 crustal levels. Magma mixing is supported by the textural and compositional characteristics of
596 olivine and titanomagnetite reflecting different crystallization histories, as well as by the occurrence
597 of inversely zoned plagioclases showing partly resorbed cores surrounded by sieve-textured mantles.
598 Furthermore, the final growth of plagioclase was controlled by the effect of undercooling due to
599 rapid decompression during magma ascent to the surface. Trace element and isotope data tend to
600 exclude crustal contamination, suggesting that magmas originated by partial melting of an enriched
601 MORB-like mantle wedge metasomatized prevalently by slab-derived aqueous fluids.

602

603 **Acknowledgments**

604 The authors are grateful to B. S. Ellis, G. Iezzi, and D. B. Dingwell for their useful and constructive
605 suggestions. Francesca Forni and Andrea Cavallo are acknowledge for assistance during sample
606 selection and microprobe analysis, respectively. Part of this work was supported by MIUR project -
607 Premiale NORTH (New hORizons of the Technology applied to experimental researches and
608 geophysical and volcanological monitoring). The research activities of the HP-HT laboratory of the
609 INGV were supported by the European Observing System Infrastructure project (EPOS). RB
610 acknowledges a NERC Urgency Grant (NE/M005003/1).

611

612 **References**

613 Agostini, S., Corti, G., Doglioni, C., Carminati, E., Innocenti, F., Tonarini, S., Manetti, P., Di
614 Vincenzo, G., Montanari, D., 2006. Tectonic and magmatic evolution of the active volcanic

615 front in El Salvador: insight into the Berlín and Ahuachapán geothermal areas. *Geothermics* 35,
616 368-408. doi:10.1016/j.geothermics.2006.05.003

617 Aloisi, M., Bonaccorso, A., Gambino, S., 2006. Imaging composite dike propagation (Etna, 2002
618 case). *Journal of Geophysical Research* 111, B06404. doi:10.1029/2005jb003908

619 Ariskin, A., Nikolaev, G., 1996. An empirical model for the calculation of spinel-melt equilibria in
620 mafic igneous systems at atmospheric pressure: 1. Chromian spinels. *Contributions to
621 Mineralogy and Petrology* 123, 282-292. doi:10.1007/s004100050156

622 Aubouin, J., Stephan, J., Roump, J., Renard, V., 1982. The middle american trench as an example
623 of a subduction zone. *Tectonophysics* 86, 113-132. doi:10.1016/0040-1951(82)90063-4

624 Bonforte, A., Hernandez, D., Gutiérrez, E., Handal, L., Polío, C., Rapisarda, S., and Scarlato, P.,
625 2015. The unrest of S. Miguel volcano (El Salvador, CA): installation of the monitoring
626 network and observed volcano-tectonic ground deformation. *Natural Hazards and Earth
627 System Sciences Discussions*, 3, 6117-6148. doi:10.5194/nhessd-3-6117-2015

628 Brounce, M., Kelley, K., Cottrell, E., 2014. Variations in $Fe^{3+}/\Sigma Fe$ of Mariana Arc Basalts and
629 Mantle Wedge fO_2 . *Journal of Petrology* 55, 2513-2536. doi:10.1093/petrology/egu065

630 Brugger, C., Hammer, J., 2010. Crystallization Kinetics in Continuous Decompression
631 Experiments: Implications for Interpreting Natural Magma Ascent Processes. *Journal of
632 Petrology* 51, 1941-1965. doi:10.1093/petrology/egq044

633 Cameron, B., Walker, J., Carr, M., Patino, L., Matias, O., Feigenson, M., 2003. Flux versus
634 decompression melting at stratovolcanoes in southeastern Guatemala. *Journal of Volcanology
635 and Geothermal Research* 119, 21-50. doi:10.1016/s0377-0273(02)00304-9

636 Cameron, B.I., Walker, J.A., 2006, Diverse volcanism in southeastern Guatemala: The role of
637 crustal contamination, in: eds. Rose et al., *Geological Society of America Special Paper* 412,
638 *Volcanic Hazards in Central America*, 121-139. doi:10.1130/2006.2412(07)

639 Carr, M., 1984. Symmetrical and segmented variation of physical and geochemical characteristics
640 of the central american volcanic front. *Journal of Volcanology and Geothermal Research* 20,

641 231-252. doi:10.1016/0377-0273(84)90041-6

642 Carr, M., Feigenson, M., Bennett, E., 1990. Incompatible element and isotopic evidence for tectonic
643 control of source mixing and melt extraction along the Central American arc. *Contributions to*
644 *Mineralogy and Petrology* 105, 369-380. doi:10.1007/bf00286825

645 Carr, M., Rose, W., Mayfield, D., 1979. Potassium content of lavas and depth to the seismic zone in
646 central america. *Journal of Volcanology and Geothermal Research* 5, 387-401.
647 doi:10.1016/0377-0273(79)90025-8

648 Carr, M. J., Feigenson, M. D., Patino, L. C. and Walker, J. A., 2004. *Volcanism and Geochemistry*
649 *in Central America: Progress and Problems*, in Eiler, J. (Ed.), *Inside the Subduction Factory*,
650 *American Geophysical Union, Washington, D. C.* doi: 10.1029/138GM09

651 Carr, M., Patino, L., Feigenson, M., 2007. Petrology and geochemistry of lavas, in Bundschuh, J.,
652 Alvarado, G. (Eds.), *Central America: Geology, Resources, and Hazards* Taylor & Francis,
653 London, pp 565-590

654 Chesner, C. A., Pullinger C., and Escobar C. D., 2004. Physical and chemical evolution of San
655 Miguel volcano, El Salvador. *Geological Society of America Special Papers*, 375, 213-
656 226. doi: 10.1130/0-8137-2375-2.213

657 Cocker, J., Griffin, B., Muehlenbachs, K., 1982. Oxygen and carbon isotope evidence for seawater-
658 hydrothermal alteration of the Macquarie Island ophiolite. *Earth and Planetary Science Letters*
659 61, 112-122. doi:10.1016/0012-821x(82)90043-7

660 Couch, S., 2003. Experimental investigation of crystallization kinetics in a haplogranite system.
661 *American Mineralogist* 88, 1471-1485. doi:10.2138/am-2003-1011

662 Davidson, J., Tepley, F., 1997. Recharge in Volcanic Systems: Evidence from Isotope Profiles of
663 Phenocrysts. *Science* 275, 826-829. doi:10.1126/science.275.5301.826

664 Davies, J., Stevenson, D., 1992. Physical model of source region of subduction zone volcanics.
665 *Journal of Geophysical Research* 97, 2037-2070. doi:10.1029/91jb02571

666 Devine, J., Gardner, J., Brack, H., Layne, G., Rutherford, M., 1995. Comparison of microanalytical

667 methods for estimating H₂O contents of silicic volcanic glasses. *American Mineralogist* 80,
668 319-328. doi:10.2138/am-1995-3-413

669 Dohmen, R., Blundy, J., 2014. A predictive thermodynamic model for element partitioning between
670 plagioclase and melt as a function of pressure, temperature and composition. *American Journal*
671 *of Science* 314, 1319-1372. doi:10.2475/09.2014.04

672 Donnelly, T. W., D. Beets, M. J. Carr, T. Jackson, G. Klaver, J. Lewis, Maury, R., H. Schellekens,
673 A. Smith, G. Wadge & D. Westerkamp, 1990, History and tectonic setting of Caribbean
674 magmatism: In: Dengo, G. (Ed.), *The Geology of North America*, v. H., The Caribbean
675 Region, chapter 13, p. 339 – 374

676 Eichelberger, J., Chertkoff, D., Dreher, S., Nye, C., 2000. Magmas in collision: Rethinking
677 chemical zonation in silicic magmas. *Geology* 28, 603. doi:10.1130/0091-
678 7613(2000)28<603:micrcz>2.0.co;2

679 Eiler, J., Carr, M., Reagan, M., Stolper, E., 2005. Oxygen isotope constraints on the sources of
680 Central American arc lavas. *Geochemistry, Geophysics, Geosystems* 6.
681 doi:10.1029/2004gc000804

682 Ersoy, Y., Helvacı, C., 2010. FC–AFC–FCA and mixing modeler: A Microsoft® Excel® spreadsheet
683 program for modeling geochemical differentiation of magma by crystal fractionation, crustal
684 assimilation and mixing. *Computers & Geosciences* 36, 383-390.
685 doi:10.1016/j.cageo.2009.06.007

686 Escobar, C. D., 2003. San Miguel Volcano and its Volcanic Hazards, El Salvador. Unpublished
687 thesis, Michigan Technological University (USA).

688 Feigenson, M., Carr, M., 1986. Positively correlated Nd and Sr isotope ratios of lavas from the
689 Central American volcanic front. *Geol* 14, 79. doi:10.1130/0091-
690 7613(1986)14<79:pcnasi>2.0.co;2

691 Fiege, A., Vetere, F., Iezzi, G., Simon, A., Holtz, F., 2015. The roles of decompression rate and
692 volatiles (H₂O + Cl ± CO₂±S) on crystallization in (trachy-) basaltic magma. *Chem. Geol.*, 411,

693 310-322.

694 Foden, J., Green, D., 1992. Possible role of amphibole in the origin of andesite: some experimental
695 and natural evidence. *Contr. Mineral. and Petrol.* 109, 479-493. doi:10.1007/bf00306551

696 Ghiorso, M., Sack, R., 1995. Chemical mass transfer in magmatic processes IV. A revised and
697 internally consistent thermodynamic model for the interpolation and extrapolation of liquid-
698 solid equilibria in magmatic systems at elevated temperatures and pressures. *Contribution to
699 Mineralogy and Petrology* 119, 197-212. doi:10.1007/bf00307281

700 Gonnermann H. M., Manga M., 2013. Dynamics of magma ascent in the volcanic conduit, in:
701 Fagents, S., Gregg, T., Lopes, R. (Eds.), *Modeling volcanic processes*. Cambridge University
702 Press, Cambridge.

703 Granieri, D., Salerno, G., Liuzzo, M., La Spina, A., Giuffrida, G., Caltabiano, T., Giudice, G.,
704 Gutierrez, E., Montalvo, F., Burton, M. R., Papale, P., 2015. Emission of gas and atmospheric
705 dispersion of SO₂ during the December 2013 eruption at San Miguel volcano (El Salvador,
706 Central America). *Geophysical Research Letters*, 42(14), 5847-5854.
707 doi:10.1002/2015GL064660.

708 Grove, T., Juster, T., 1989. Experimental investigations of low-Ca pyroxene stability and olivine-
709 pyroxene-liquid equilibria at 1-atm in natural basaltic and andesitic liquids. *Contributions to
710 Mineralogy and Petrology*, 103, 287-305. doi:10.1007/bf00402916

711 Hammer, J., Rutherford, M., 2002. An experimental study of the kinetics of decompression-induced
712 crystallization in silicic melt. *Journal of Geophysical Research* 107.
713 doi:10.1029/2001jb000281

714 Hibbard, M., 1981. The magma mixing origin of mantled feldspars. *Contributions to Mineralogy
715 and Petrology*, 76, 158-170. doi:10.1007/bf00371956

716 Iezzi, G., Mollo, S., Shahini, E., Cavallo, A., Scarlato, P., 2014. The cooling kinetics of plagioclase
717 feldspar as revealed by electron-microprobe mapping. *American Mineralogist* 99, 898-907.
718 doi:10.2138/am.2014.4626

719 Iezzi, G., Mollo, S., Torresi, G., Ventura, G., Cavallo, A., Scarlato, P., 2011. Experimental
720 solidification of an andesitic melt by cooling. *Chemical Geology* 283, 261-273.
721 doi:10.1016/j.chemgeo.2011.01.024

722 Izbekov, P., Eichelberger, J., Ivanov, B., 2004. The 1996 Eruption of Karymsky Volcano,
723 Kamchatka: Historical Record of Basaltic Replenishment of an Andesite Reservoir. *Journal of*
724 *Petrology* 45, 2325-2345. doi:10.1093/petrology/egh059

725 Kirkpatrick, R., Robinson, G., Hays, J., 1976. Kinetics of crystal growth from silicate melts:
726 Anorthite and diopside. *J. Geophys. Res.*, 81, 5715-5720. doi:10.1029/jb081i032p05715

727 Kress, V., Carmichael, I., 1991. The compressibility of silicate liquids containing Fe₂O₃ and the
728 effect of composition, temperature, oxygen fugacity and pressure on their redox states.
729 *Contributions to Mineralogy and Petrology* ,108, 82-92. doi:10.1007/bf00307328

730 Lanzafame, G., Mollo, S., Iezzi, G., Ferlito, C., Ventura, G., 2013. Unraveling the solidification
731 path of a pahoehoe “cicirara” lava from Mount Etna volcano. *Bull. Volcanol.* 75, 703.
732 doi:http://dx.doi.org/10.1007/s00445-013-0703-8.

733 Le Bas, M., Maitre, R., Streckeisen, A., Zanettin, B., 1986. A Chemical Classification of Volcanic
734 Rocks Based on the Total Alkali-Silica Diagram. *Journal of Petrology* 27, 745-750.
735 doi:10.1093/petrology/27.3.745

736 Le Maitre, R., 2002. *Igneous rocks*. Cambridge University Press, Cambridge, U.K.

737 Leeman, W., Carr, M., Morris, J., 1994. Boron geochemistry of the Central American Volcanic Arc:
738 Constraints on the genesis of subduction-related magmas. *Geochimica et Cosmochimica Acta*
739 58, 149-168. doi:10.1016/0016-7037(94)90453-7

740 MacMillan, I., Gans, P., Alvarado, G., 2004. Middle Miocene to present plate tectonic history of the
741 southern Central American Volcanic Arc. *Tectonophysics* 392, 325-348.
742 doi:10.1016/j.tecto.2004.04.014

743 Martinez-Hackert, B., Bajo, J., Escobar, D., Gutierrez, E., Mixco, L., Hernandez, W., 2015,
744 Chaparrastique (San Miguel) Volcano Eruptions since Dec. 29th, 2013, El Salvador, Abstract

745 V23A-3087, presented at 2015 Fall Meeting, AGU, San Francisco, Calif., 14-18 Dec.

746 Matthey, D., Lowry, D., Macpherson, C., 1994. Oxygen isotope composition of mantle peridotite.
747 Earth and Planetary Science Letters 128, 231-241. doi:10.1016/0012-821x(94)90147-3

748 Mollo, S., Giacomoni, P., Andronico, D., Scarlato, P., 2015. Clinopyroxene and titanomagnetite
749 cation redistributions at Mt. Etna volcano (Sicily, Italy): Footprints of the final solidification
750 history of lava fountains and lava flows. Chemical Geology, 406, 45-54.
751 doi:10.1016/j.chemgeo.2015.04.017

752 Mollo, S., Misiti, V., Scarlato, P., Soligo, M., 2012a. The role of cooling rate in the origin of high
753 temperature phases at the chilled margin of magmatic intrusions. Chemical Geology, 322-323,
754 28-46. doi:10.1016/j.chemgeo.2012.05.029

755 Mollo, S., Iezzi, G., Ventura, G., Cavallo, A., Scarlato, P., 2012b. Heterogeneous nucleation
756 mechanisms and formation of metastable phase assemblages induced by different crystalline
757 seeds in a rapidly cooled andesitic melt. Journal of Non-Crystalline Solids 358, 1624-1628.
758 doi:10.1016/j.jnoncrysol.2012.04.010

759 Mollo, S., Putirka, K., Iezzi, G., Del Gaudio, P., Scarlato, P., 2011. Plagioclase-melt
760 (dis)equilibrium due to cooling dynamics: Implications for thermometry, barometry and
761 hygrometry. Lithos 125, 221-235. doi:10.1016/j.lithos.2011.02.008

762 Morimoto, N., 1988. Nomenclature of Pyroxenes. Mineralogy and Petrology 39, 55-76.
763 doi:10.1007/bf01226262

764 Muncill, G.E., Lasaga, A.C., 1987. Crystal-growth kinetics of plagioclase in igneous systems: One-
765 atmosphere experiments and application of a simplified growth model. American Mineralogist
766 72, 299-311

767 Namur, O., Charlier, B., Toplis, M., Vander Auwera, J., 2011. Prediction of plagioclase-melt
768 equilibria in anhydrous silicate melts at 1-atm. Contributions to Mineralogy and Petrology 163,
769 133-150. doi:10.1007/s00410-011-0662-z

770 Neill, O., Larsen, J., Izbekov, P., Nye, C., 2015. Pre-eruptive magma mixing and crystal transfer

771 revealed by phenocryst and microlite compositions in basaltic andesite from the 2008 eruption
772 of Kasatochi Island volcano. *American Mineralogist* 100, 722-737. doi:10.2138/am-2015-4967

773 Patino, L., Carr, M., Feigenson, M., 2000. Local and regional variations in Central American arc
774 lavas controlled by variations in subducted sediment input. *Contributions to Mineralogy and*
775 *Petrology* 138, 265-283. doi:10.1007/s004100050562

776 Plank, T., Langmuir, C., 1998. The chemical composition of subducting sediment and its
777 consequences for the crust and mantle. *Chemical Geology* 145, 325-394. doi:10.1016/s0009-
778 2541(97)00150-2

779 Protti, M., McNally, K., Pacheco, J., González, V., Montero, C., Segura, J., Brenes, J., Barboza, V.,
780 Malavassi, E., Güendel, F., Simila, G., Rojas, D., Velasco, A., Mata, A., Schillinger, W., 1995.
781 The March 25, 1990 ($M_w = 7.0$, $M_L = 6.8$), earthquake at the entrance of the Nicoya Gulf,
782 Costa Rica: Its prior activity, foreshocks, aftershocks, and triggered seismicity. *J. Geophys.*
783 *Res.* 100, 20345. doi:10.1029/94jb03099

784 Pupier, E., Duchene, S., Toplis, M., 2007. Experimental quantification of plagioclase crystal size
785 distribution during cooling of a basaltic liquid. *Contributions to Mineralogy and Petrology* 155,
786 555-570. doi:10.1007/s00410-007-0258-9

787 Putirka, K., 2005a. Mantle potential temperatures at Hawaii, Iceland, and the mid-ocean ridge
788 system, as inferred from olivine phenocrysts: Evidence for thermally driven mantle plumes.
789 *Geochemistry, Geophysics, Geosystems* 6, n/a-n/a. doi:10.1029/2005gc000915

790 Putirka, K., 2005b. Igneous thermometers and barometers based on plagioclase + liquid equilibria:
791 Tests of some existing models and new calibrations. *American Mineralogist* 90, 336-346.
792 doi:10.2138/am.2005.1449

793 Putirka, K., 2008. Thermometers and barometers for volcanic systems. In: Putirka, K.D., Tepley, F.
794 (Eds.), *Minerals, Inclusions and Volcanic Processes. Reviews in mineralogy and geochemistry*
795 69, pp. 61–120.

796 Richer, M., Mann, C., Stix, J., 2004. Mafic magma injection triggers eruption at Ilopango Caldera,

797 El Salvador, Central America. In: W.J. Rose, J.J. Bommer, D.L. López, M.J. Carr and
798 J.J. Major (eds.): Natural hazards in El Salvador. Geological Society of America, Special Paper
799 375, Boulder CO, 2004, pp.175–189. doi:10.1130/0-8137-2375-2.175

800 Roeder, P., Emslie, R., 1970. Olivine-liquid equilibrium. Contributions to Mineralogy and
801 Petrology 29, 275-289. doi:10.1007/bf00371276

802 Roggensack, K., 2001. Unraveling the 1974 eruption of Fuego volcano (Guatemala) with small
803 crystals and their young melt inclusions. Geology 29, 911. doi:10.1130/0091-
804 7613(2001)029<0911:uteofv>2.0.co;2

805 Roggensack, K., Hervig, R., McKnight, S., Williams, S., 1997. Explosive Basaltic Volcanism from
806 Cerro Negro Volcano: Influence of Volatiles on Eruptive Style. Science 277, 1639-1642.
807 doi:10.1126/science.277.5332.1639

808 Rutherford, M., 2008. Magma Ascent Rates. Reviews in Mineralogy and Geochemistry 69, 241-271.
809 doi:10.2138/rmg.2008.69.7

810 Sadofsky, S., Portnyagin, M., Hoernle, K., van den Bogaard, P., 2007. Subduction cycling of
811 volatiles and trace elements through the Central American volcanic arc: evidence from melt
812 inclusions. Contributions to Mineralogy and Petrology 155, 433-456. doi:10.1007/s00410-007-
813 0251-3

814 Sato, H., 1995. Textural difference between pahoehoe and aa lavas of Izu-Oshima volcano, Japan -
815 an experimental study on population density of plagioclase. Journal of Volcanology and
816 Geothermal Research 66, 101-113. doi:10.1016/0377-0273(94)00055-1

817 Scarlato, P., Kattan, C., Papale, P., Gresta, S., 2014, International cooperation during volcanic
818 crisis: an example from the Italy-El Salvador monitoring system installed at Chaparrastique
819 volcano, El Salvador. Cities on Volcanoes 8.

820 Schiavi, F., Walte, N., Keppler, H., 2009. First in situ observation of crystallization processes in a
821 basaltic-andesitic melt with the moissanite cell. Geology 37, 963-966. doi:10.1130/g30087a.1

822 Shcherbakov, V., Plechov, P., Izbekov, P., Shipman, J., 2010. Plagioclase zoning as an indicator of

823 magma processes at Bezymianny Volcano, Kamchatka. *Contributions to Mineralogy and*
824 *Petrology* 162, 83-99. doi:10.1007/s00410-010-0584-1

825 Sisson, T., Layne, G., 1993. H₂O in basalt and basaltic andesite glass inclusions from four
826 subduction-related volcanoes. *Earth and Planetary Science Letters* 117, 619-635.
827 doi:10.1016/0012-821x(93)90107-k

828 Smith, R., Lofgren, G., 1983. An analytical and experimental study of zoning in plagioclase. *Lithos*
829 16, 153-168. doi:10.1016/0024-4937(83)90012-9

830 Smith, V., Tiller, W., Rutter, J., 1955. A mathematical analysis of solute redistribution during
831 solidification. *Can. J. Phys.* 33, 723-745. doi:10.1139/p55-089

832 Sparks, S., Sigurdsson, H., Wilson, L., 1977. Magma mixing: a mechanism for triggering acid
833 explosive eruptions. *Nature* 267, 315-318. doi:10.1038/267315a0

834 Staudigel, H., Davies, G., Hart, S., Marchant, K., Smith, B., 1995. Large scale isotopic Sr, Nd and
835 O isotopic anatomy of altered oceanic crust: DSDP/ODP sites 417/418. *Earth and Planetary*
836 *Science Letters* 130, 169-185. doi:10.1016/0012-821x(94)00263-x

837 Streck, M., Dungan, M., Malavassi, E., Reagan, M., Bussy, F., 2002. The role of basalt
838 replenishment in the generation of basaltic andesites of the ongoing activity at Arenal volcano,
839 Costa Rica: evidence from clinopyroxene and spinel. *Bulletin of Volcanology* 64, 316-327.
840 doi:10.1007/s00445-002-0209-2

841 Sun, S., McDonough, W., 1989. Chemical and isotopic systematics of oceanic basalts: implications
842 for mantle composition and processes. *Geological Society, London, Special Publications* 42,
843 313-345. doi:10.1144/gsl.sp.1989.042.01.19

844 Syracuse, E., Abers, G., 2006. Global compilation of variations in slab depth beneath arc volcanoes
845 and implications. *Geochemistry, Geophysics, Geosystems* 7, n/a-n/a.
846 doi:10.1029/2005gc001045

847 Teng, H.H., 2013. How ions and molecules organize to form crystals. *Elements*, 9, 189–194

848 Tepley, F., Davidson, J., Tilling, R., ARTH, J., 2000. Magma Mixing, Recharge and Eruption

849 Histories Recorded in Plagioclase Phenocrysts from El Chichon Volcano, Mexico. *Journal of*
850 *Petrology* 41, 1397-1411. doi:10.1093/petrology/41.9.1397

851 Tonarini, S., Agostini, S., Doglioni, C., Innocenti, F., Manetti, P., 2007. Evidence for serpentinite
852 fluid in convergent margin systems: The example of El Salvador (Central America) arc lavas.
853 *Geochemistry, Geophysics, Geosystems* 8, 9. doi:10.1029/2006gc001508

854 Toramaru, A., Noguchi, S., Oyoshihara, S., Tsune, A., 2008. MND(microlite number density) water
855 exsolution rate meter. *Journal of Volcanology and Geothermal Research* 175, 156-167.
856 doi:10.1016/j.jvolgeores.2008.03.035

857 Tsuchiyama, A., 1985. Dissolution kinetics of plagioclase in the melt of the system diopside-albite-
858 anorthite, and origin of dusty plagioclase in andesites. *Contributions to Mineralogy and*
859 *Petrology* 89, 1-16. doi:10.1007/bf01177585

860 Tsuchiyama, A., Takahashi, E., 1983. Melting kinetics of a plagioclase feldspar. *Contr. Mineral.*
861 *and Petrol.* 84, 345-354. doi:10.1007/bf01160286

862 Vetere, F., Iezzi, G., Behrens, H., Holtz, F., Ventura, G., Misiti, V., Cavallo, A., Mollo, S., Dietrich,
863 M., 2015. Glass forming ability and crystallization behaviour of sub-alkaline silicate melts.
864 *Earth-Science Reviews* 150, 25-44. doi:10.1016/j.earscirev.2015.07.001

865 vonQuadt, A., 1997. U–Pb zircon and Sr–Nd–Pb whole-rock investigations from the continental
866 deep drilling (KTB). *Geologische Rundschau* 86, S258-S271. doi:10.1007/pl00014659

867 Wade, J., Plank, T., Melson, W., Soto, G., Hauri, E., 2006. The volatile content of magmas from
868 Arenal volcano, Costa Rica. *Journal of Volcanology and Geothermal Research* 157, 94-120.
869 doi:10.1016/j.jvolgeores.2006.03.045

870 Walker, R. J., Morgan, J. W., & Horan, M. F., 1995. Osmium-187 enrichment in some plumes:
871 evidence for core-mantle interaction?. *Science*, 269(5225), 819-822. doi:
872 10.1126/science.269.5225.819

873 Zhang, Y., Ni, H., Chen, Y., 2010. Diffusion Data in Silicate Melts. *Reviews in Mineralogy and*
874 *Geochemistry* 72, 311-408. doi:10.2138/rmg.2010.72.8

875

876 **Figure captions**

877 Fig.1. Schematic map showing the Central American volcanic arc (CAVA), where volcanism
878 occurs above the plate boundary between the subducting Cocos plate and the overriding Caribbean
879 plate (a). The ash plume of the Chaparrastique vulcanian-type eruption captured by the Suomi NPP
880 satellite (b). Seismic monitoring stations of VSMG, RANC, and LACA located, respectively, at 1
881 km north, 2 km east, and 3 km south-east to the volcanic crater where twelve rock samples were
882 collected (c).

883

884 Fig.2. Textural and compositional characteristics of the scoria clasts. Larger Fo-rich poikilitic
885 olivines include sub-rounded Mg-rich, Usp-rich titanomagnetites (a). Smaller Fo-poor olivines are
886 accompanied by the Mg-poor, Usp-poor titanomagnetites with incomplete textures and glass
887 embayments resulting from the agglomeration of bleb-like grains (b). Orthopyroxenes are less
888 frequently observed and sometimes occur as larger crystals partly enclosing smaller clinopyroxenes
889 (c). Plagioclases are characterized by dark-grey, partly corroded An₅₃₋₆₅ cores, light-grey, coarsely
890 sieve-textured An₈₀₋₈₄ mantles, and grey, euhedral to sub-euhedral An₇₇₋₈₀ rims (d, e, f). Plagioclase
891 rims results from the attachment of a number of groundmass microlites growing as tabular crystals
892 (g, h, i). Ol, olivine. Timt, titanomagnetite. Opx, orthopyroxene. Cpx, clinopyroxene. Pl,
893 plagioclase. Fo, forsterite. Usp, ulvospinel. En, enstatite. Di, diopside. An, anorthite.

894

895 Fig.3. TAS (total alkali vs. silica; Le Bas et al., 1986) diagram showing calc-alkaline basalts (CH1-
896 CH2) and moderately (CH3-CH6) to more differentiated (CH7-CH12) basaltic andesites (a).
897 According to the SiO₂ vs. K₂O diagram (LeMaitre, 2002), the Chaparrastique volcanic products
898 have medium-K affinity (b). The chondrite-normalized pattern (Sun and McDonough, 1989) of
899 REE (rare earth elements) exhibits sub-parallel trends shifting towards progressive REE
900 enrichments from CH1 to CH12 (c). The primordial mantle-normalized pattern (Sun and

901 McDonough, 1989) of trace elements shows typical features of arc magmas, with LILE (large ion
902 lithophile elements) to HFSE (high field strength elements) ratios higher than the primitive mantle
903 values (d).

904

905 Fig.4. Predictions for the crystallization temperatures of olivine phenocrysts using the thermometer
906 of Putirka (2005a) and assuming equilibrium with the bulk rock compositions through the Fe-Mg
907 exchange of Roeder and Emslie (1970) (a). Test for equilibrium using the plagioclase-based model
908 of Namur et al. (2012). Temperature estimates for plagioclase are obtained by the thermometer of
909 Putirka (2005b) (b). According to Putirka (2008), the Fe-Mg exchange between clinopyroxene and
910 orthopyroxene suggests equilibrium crystallization conditions. The two-pyroxenes equilibrated at
911 relative low temperatures (c) and low pressures (d).

912

913 Fig.5. Phase equilibrium experiments of Foden and Green (1992) conducted at $P = 200$ MPa and
914 reproducing the crystallization sequence of olivine, olivine + plagioclase, and olivine + plagioclase
915 + pyroxene that is observed in the natural products. The experimental temperatures of Foden and
916 Green (1992) closely agrees with the saturation temperatures predicted by olivine-, plagioclase-,
917 and pyroxene-based thermometers, evidencing primary nucleation and growth of olivine and
918 plagioclase, and subordinated formation of pyroxene. The lack of amphibole in the eruptive
919 products is in accordance with low- P crystallization conditions in which the amphibole saturation
920 surface is encountered only at $T < 950$ °C.

921

922 Fig.6. Comparison between major oxides of bulk rocks from this study and MELTS (Ghiorso and
923 Sack, 1995) simulations conducted using as starting liquid the most primitive CH1 basalt, and
924 initiated from the superliquidus temperature of 1170 °C and continued along a fractional
925 crystallization path (Table 3s submitted as supplementary material). Dotted lines indicate the
926 appearance of the mineral phase as suggested by MELTS thermodynamics. Ol, olivine. Pl,

927 plagioclase. Cpx, clinopyroxene. Opx, orthopyroxene. Sp, spinel (a). Mixing and fractional
928 crystallization processes modelled by the Excel spreadsheet of Ersoy and Helvacı (2010). The
929 original partition coefficients set in the spreadsheet were used to derive the fractional crystallization
930 vectors of REE (c), HFSE (b), and TE (d). While these vectors describe an implausible magmatic
931 evolution for the Chaparrastique products, mixing (straight) lines joining CH1 and CH12 end-
932 members successfully model the bulk rock variability.

933

934 Fig.7. Chemical profiles carried out far enough away from the plagioclase surface (see also Table
935 4s) reveal showing chemical gradients at the crystal-melt interface causing enrichments of cations
936 incompatible with plagioclase crystal lattice (a). The lattice strain equation of Dohmen & Blundy
937 (2014) predicts that, upon the effect of equilibrium crystallization, $\ln K_{Mg}$ increases linearly with
938 $1/T$. The partitioning of Mg measured between plagioclase sodic core (or calcic mantle) and CaO-
939 poor (or CaO-rich) bulk rocks yields values that closely resemble those predicted by the
940 thermodynamic equation. In contrast, partition coefficients measured as the ratio between
941 plagioclase rim and matrix glass show values much higher than those expected for equilibrium
942 crystallization, as for the case K_{Mg}^{app} (i.e., disequilibrium growth) $>$ K_{Mg}^{true} (i.e., equilibrium growth)
943 (b). Assuming that the advancing plagioclase rim does not violate the condition of crystallization
944 into an infinite melt reservoir, the value of K_{Mg}^{app} can be modelled as a function of melt and crystal
945 parameters and can be ascribed to rapid plagioclase growth velocities of $\sim 2-7 \times 10^{-8}$ cm/s (c).

946

947 Fig.8. U/La vs. Ba/Th diagram based on soluble/insoluble element pairs. Using these element ratios,
948 Cameron et al. (2002) modelled the role played by melts and fluids (grey trajectories) derived from
949 the subducted slab for the partial melting of the mantle wedge (a). La/Sm vs. Ba/La diagram based
950 on ratios of elements that are not highly fluid mobile in arc systems, such as the REE and HFSE.
951 This diagram allows to evidence the contribution of to the mantle wedge of melt-like components

952 derived from the subducted slab (b). $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $\delta^{18}\text{O}$ diagram that serve to discriminate crustal and
953 source contamination. Cameron and Walker (2006) modelled the mixing trajectories between the
954 sediment-modified mantle (blue trajectory) and crustal contaminants (green trajectory) (c). $^{87}\text{Sr}/^{86}\text{Sr}$
955 vs. $^{143}\text{Nd}/^{144}\text{Nd}$ diagram shows evidence of crustal contamination for rocks in Guatemala that are
956 behind the volcanic front. However, most of the volcanic products in Central America define an
957 array with positive slope that represents a mix between the enriched MORB source and the
958 subducted slab (d). EM, enriched MORB source. OC, oceanic crust. CS, carbonate sediments. HS,
959 hemipelagic sediments. DM, depleted MORB source.

Figure1
[Click here to download high resolution image](#)

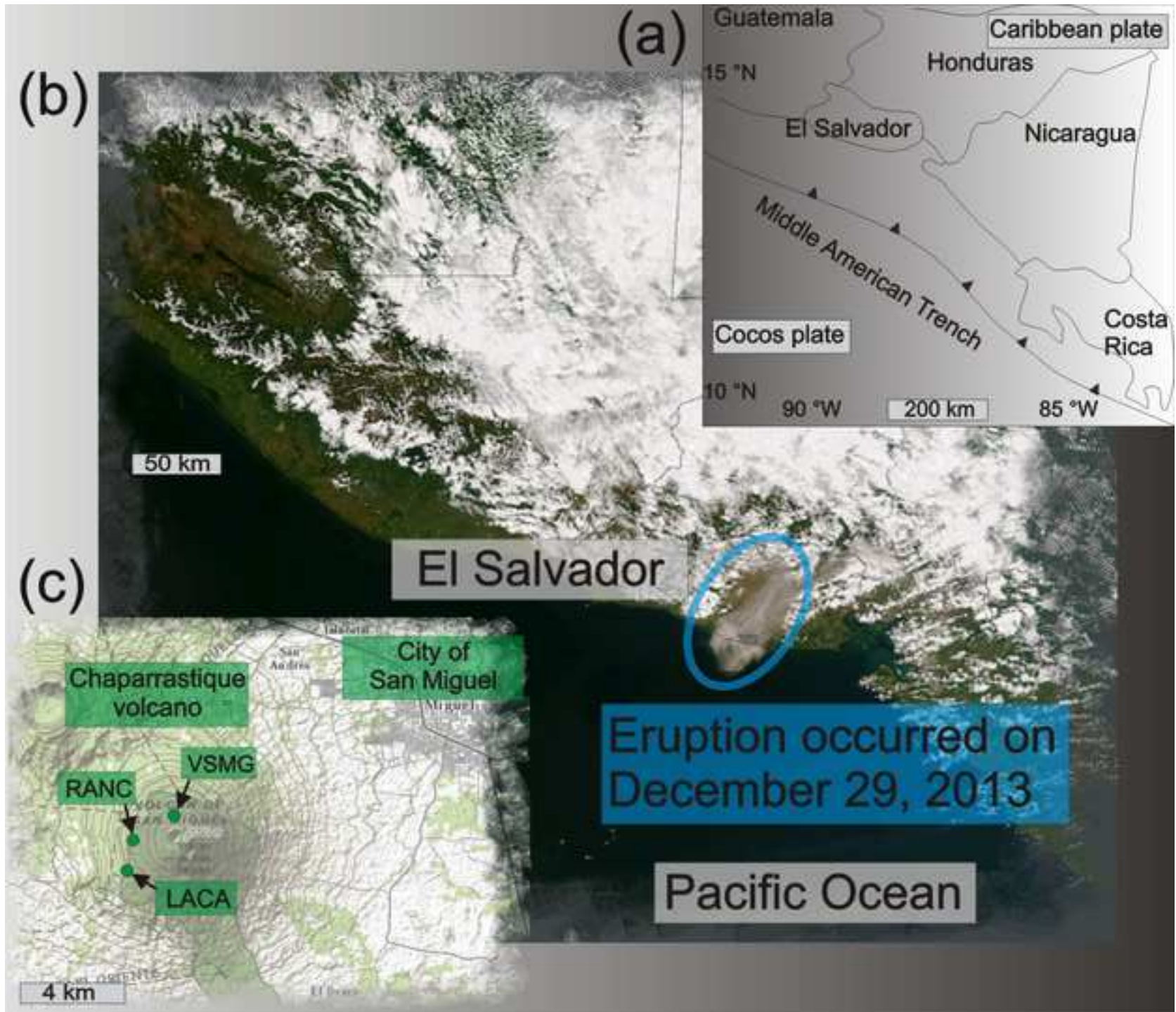


Figure 2

[Click here to download high resolution image](#)

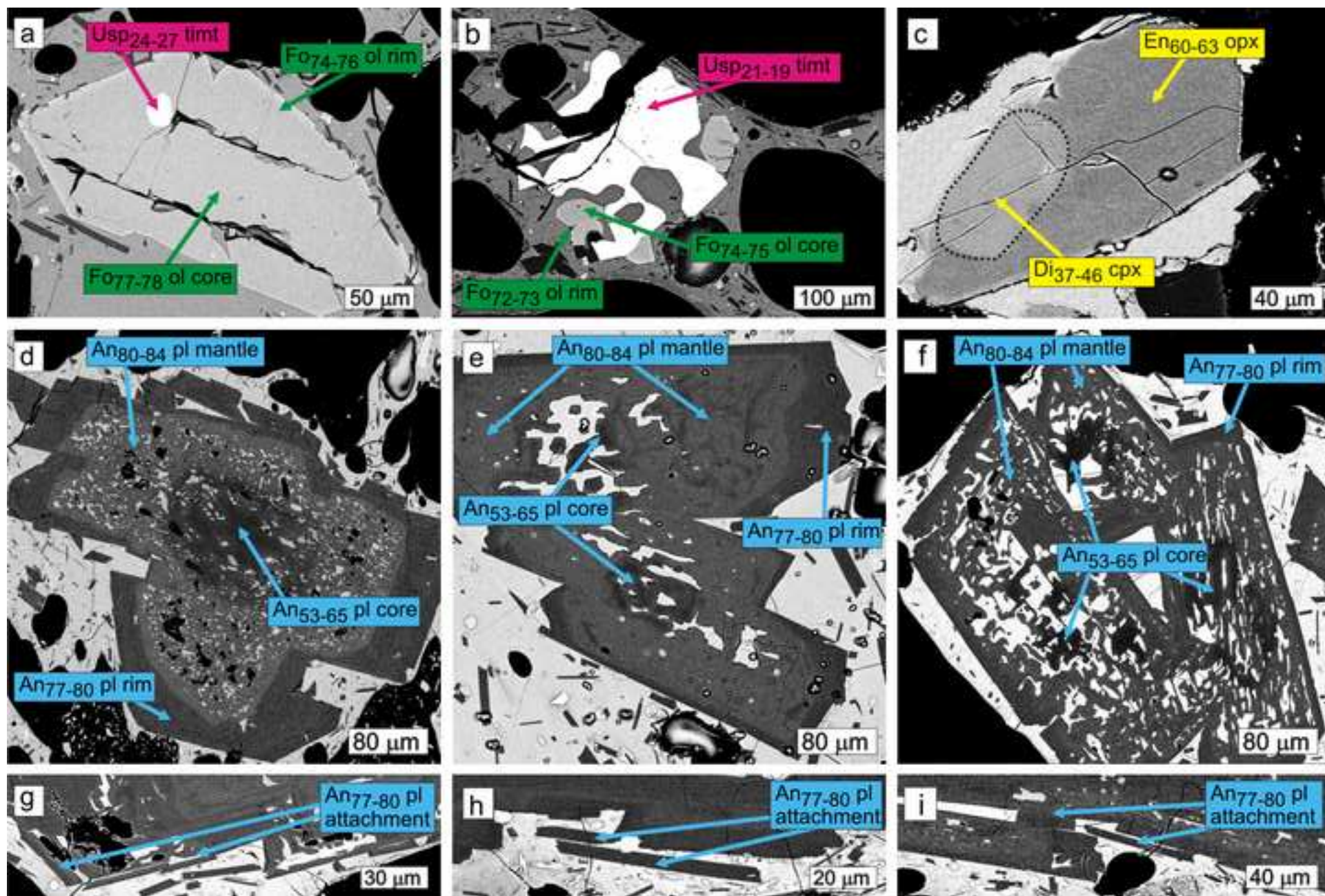


Figure3

[Click here to download high resolution image](#)

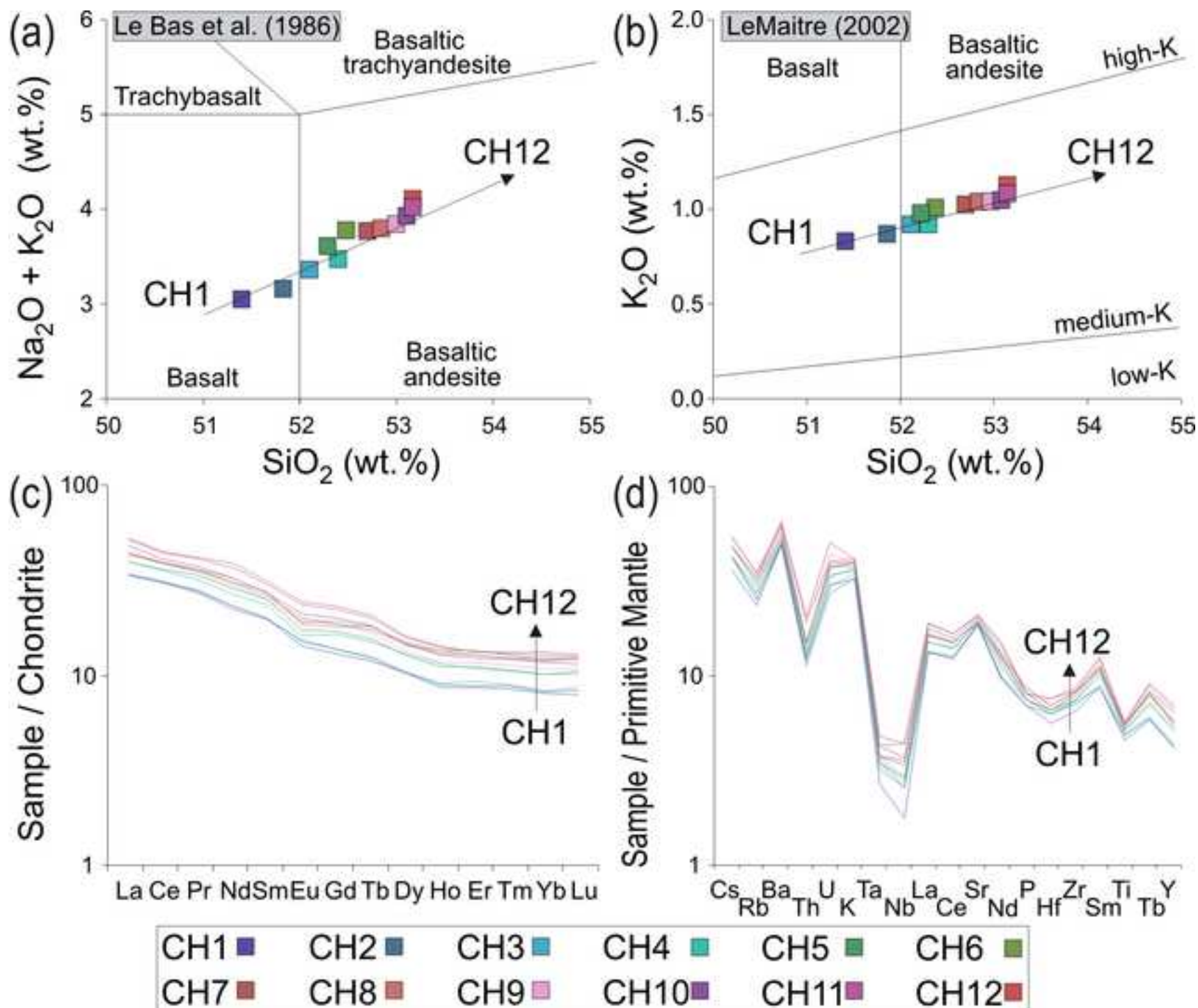


Figure 4

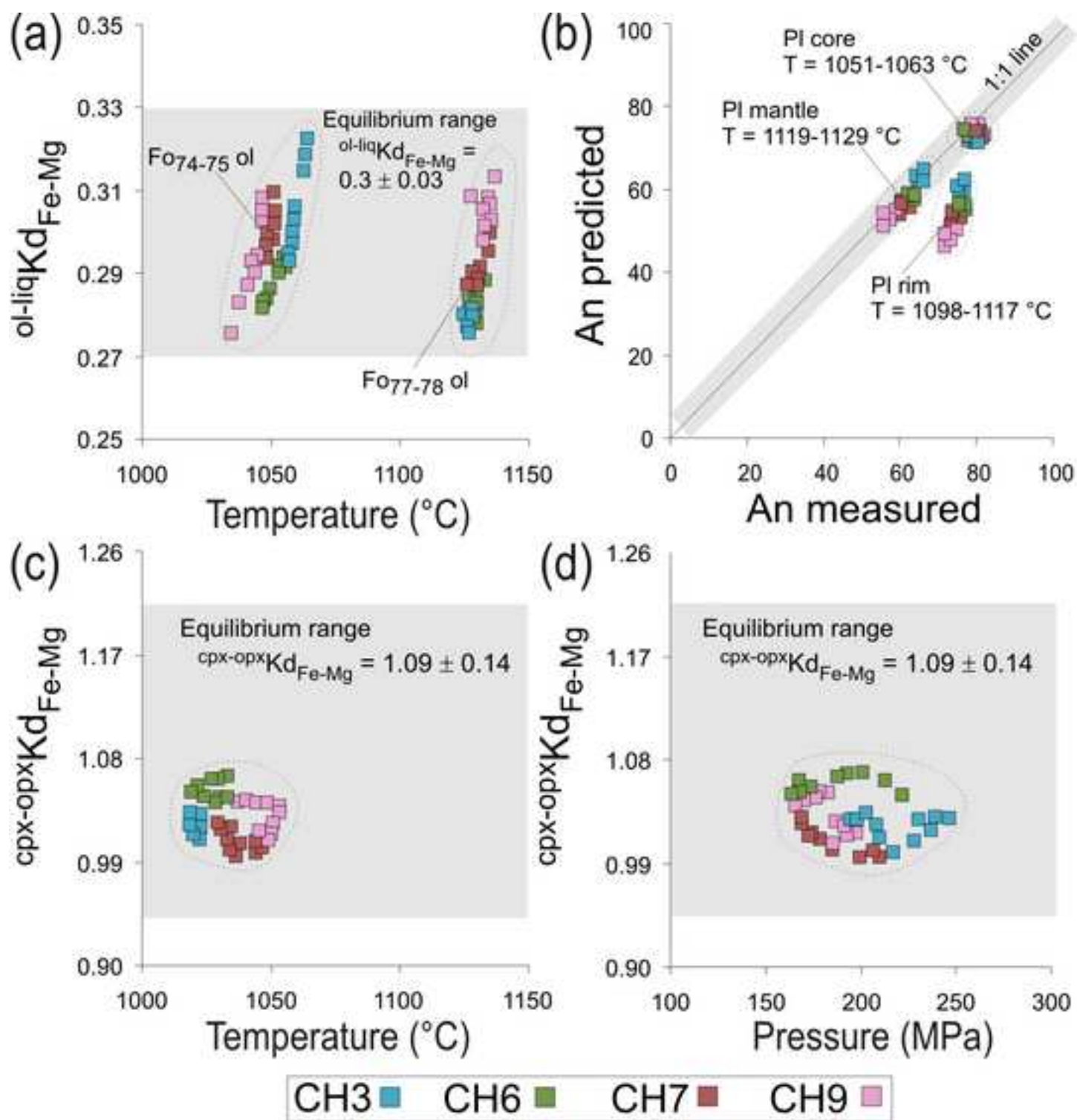
[Click here to download high resolution image](#)

Figure5
[Click here to download high resolution image](#)

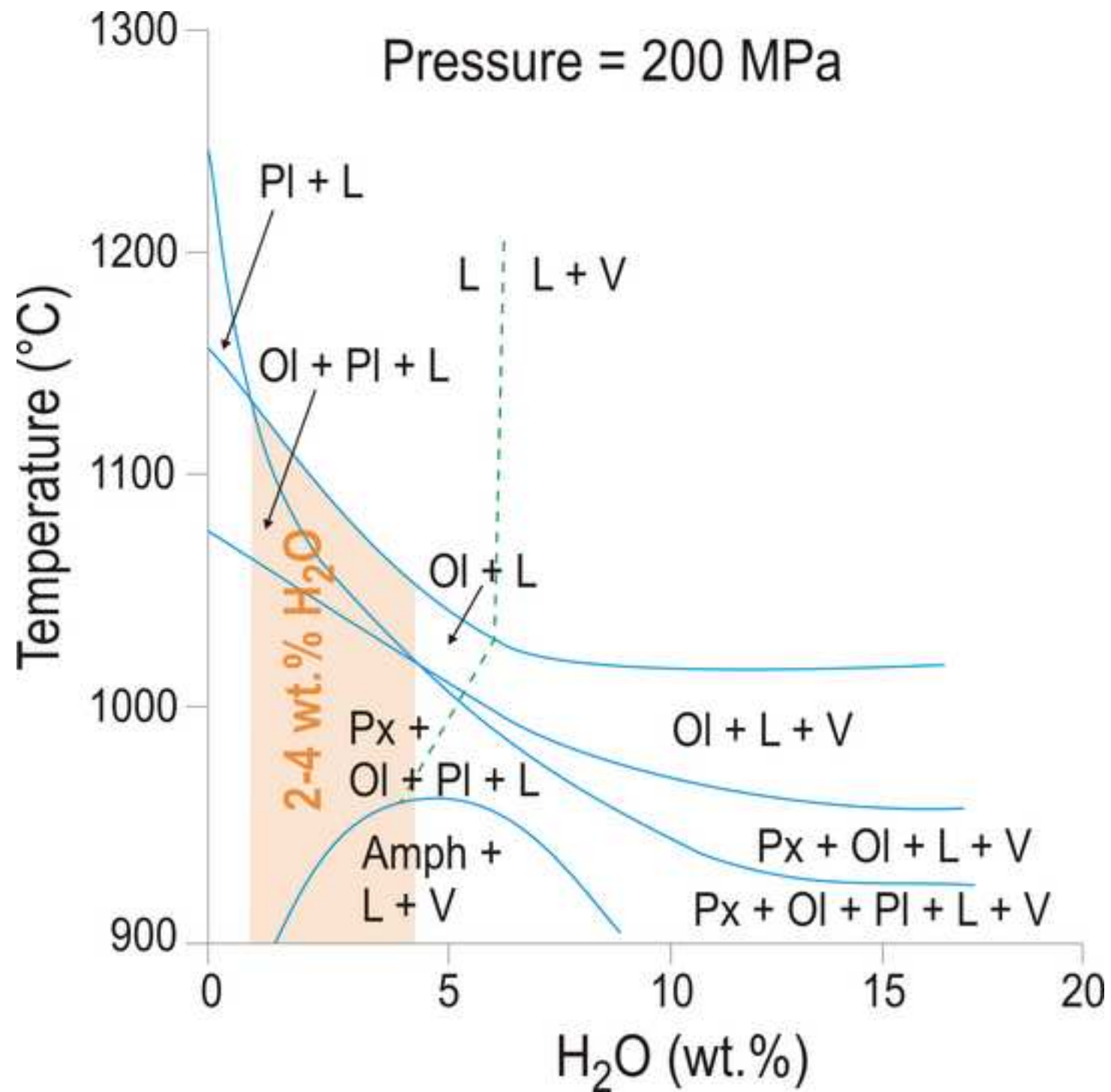


Figure6
[Click here to download high resolution image](#)

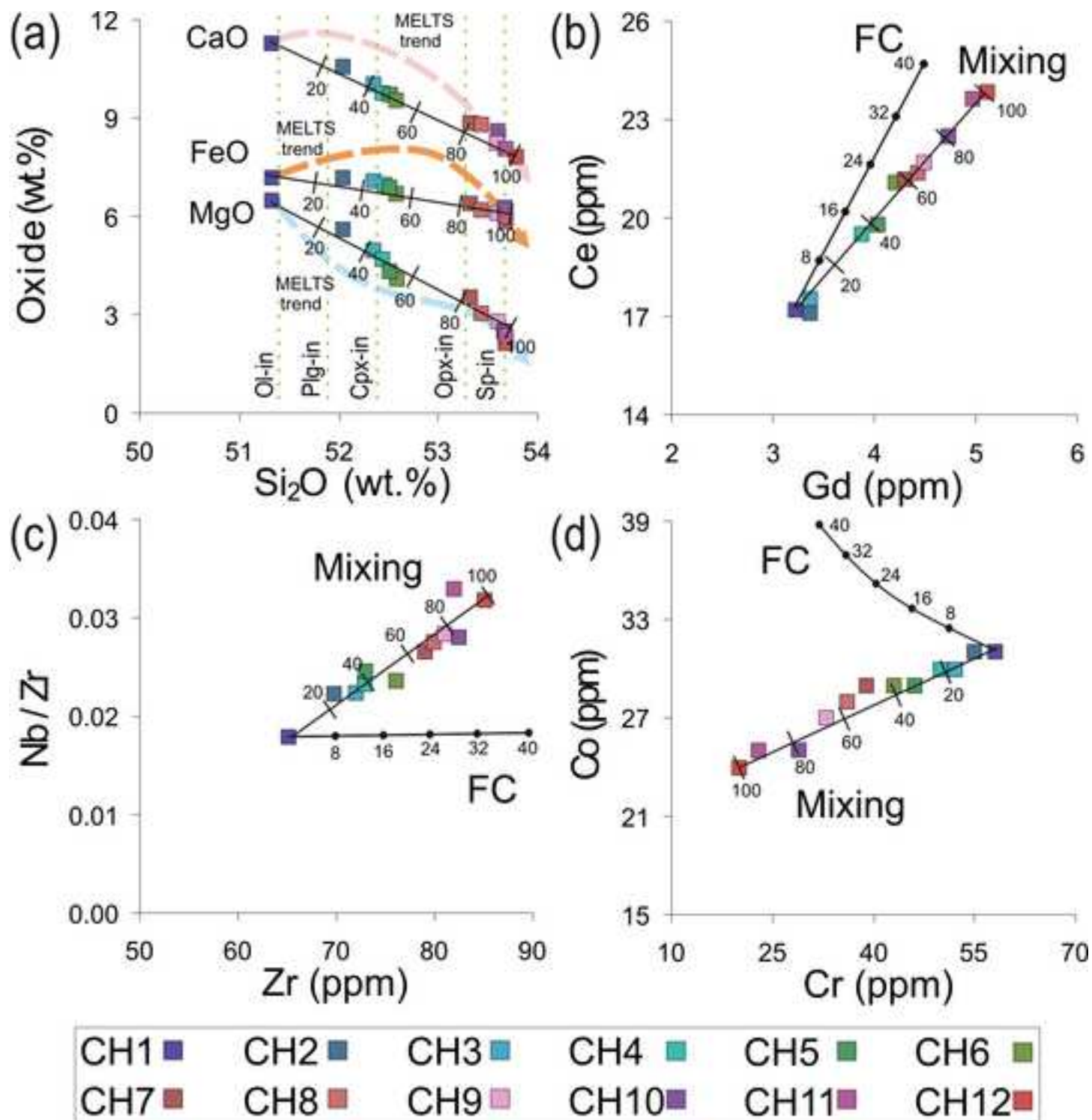


Figure 7
[Click here to download high resolution image](#)

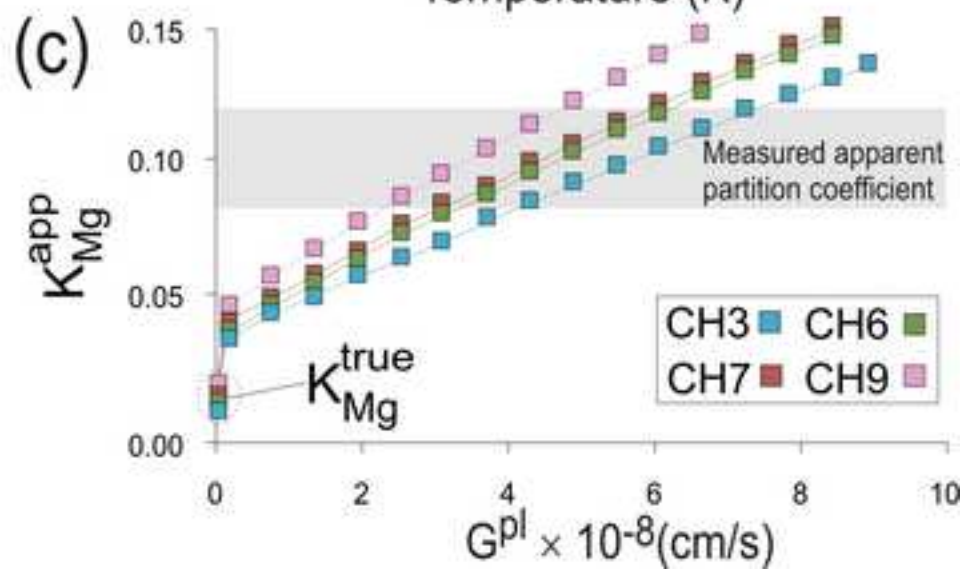
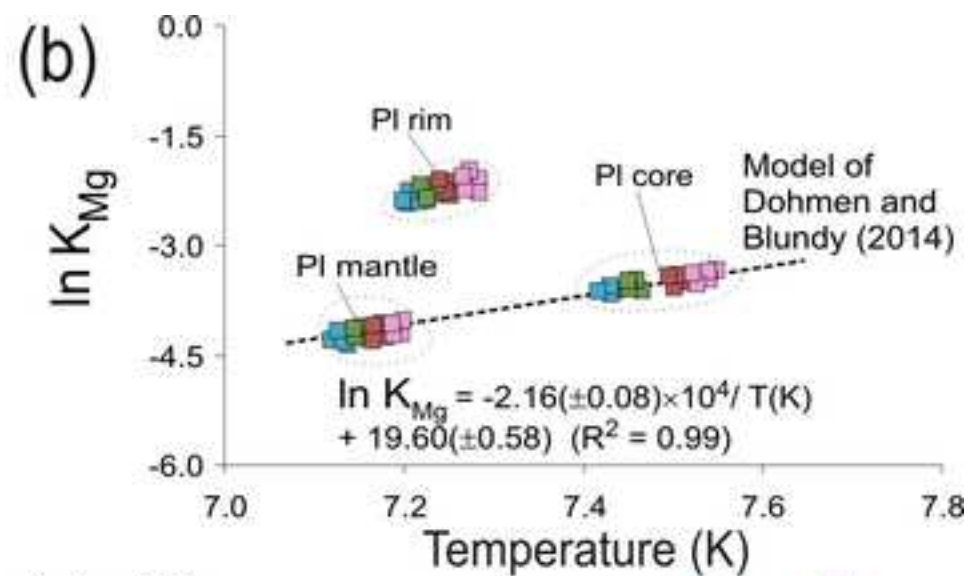
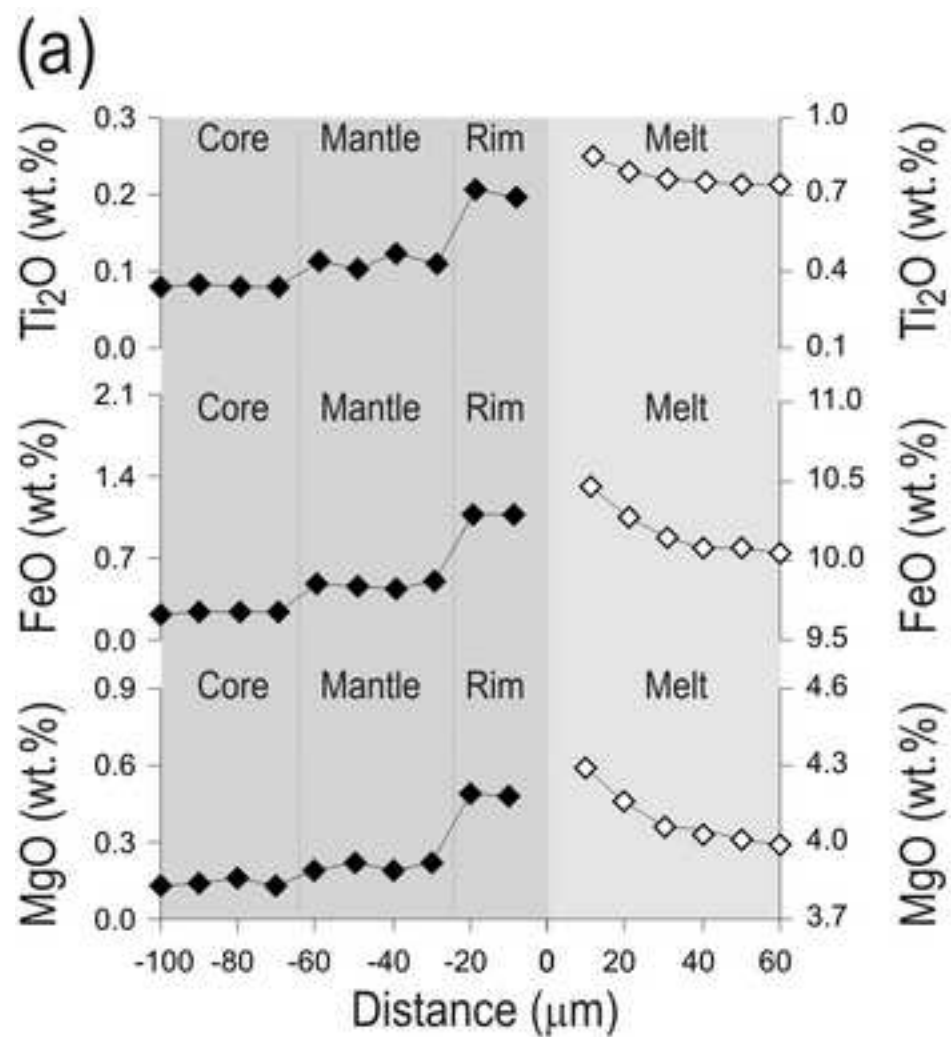


Figure 8

[Click here to download high resolution image](#)

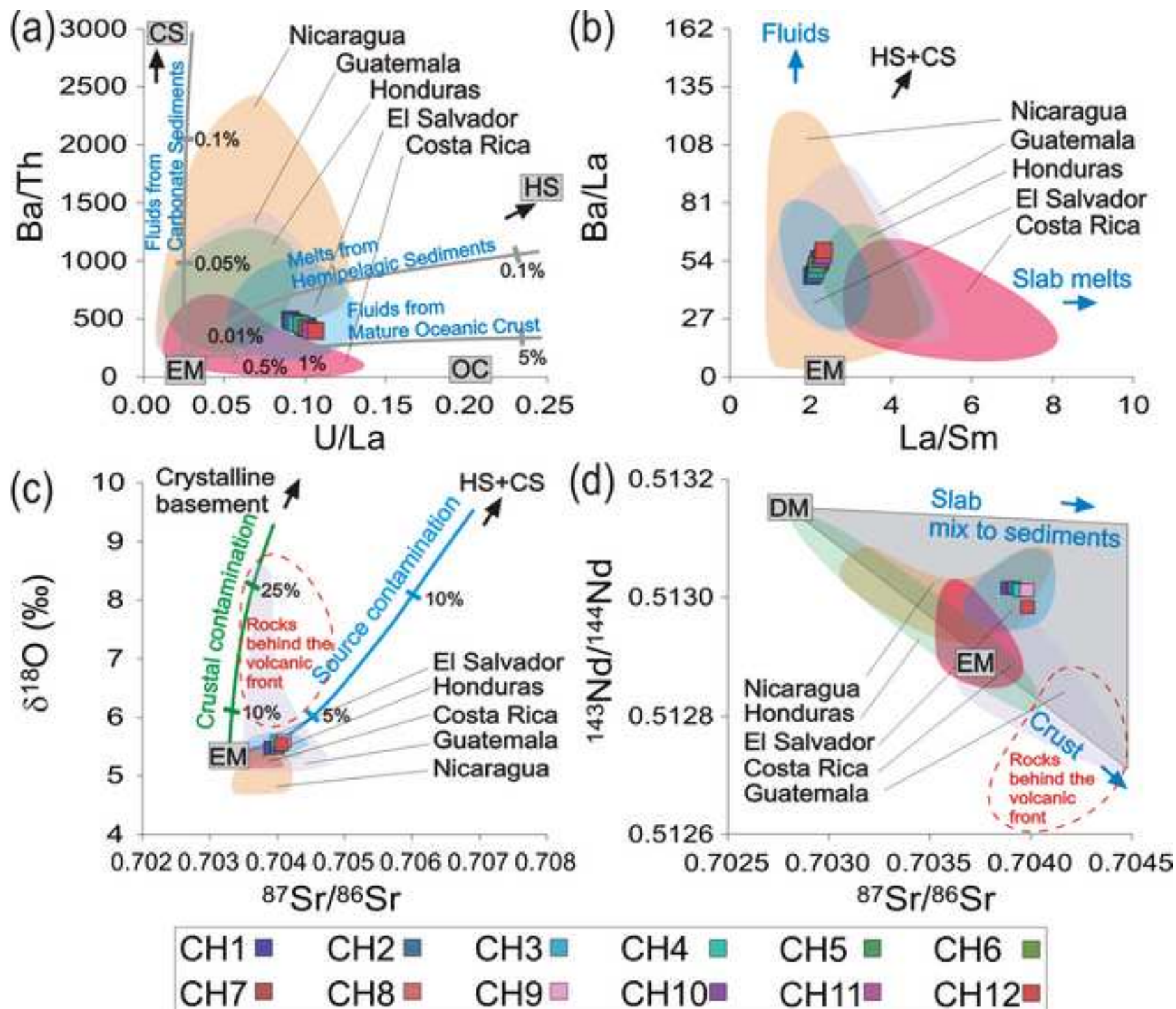


Table1S

[Click here to download Background dataset for online publication only: Table1S.xls](#)

Table2S

[Click here to download Background dataset for online publication only: Table2S.xls](#)

Table3S

[Click here to download Background dataset for online publication only: Table3S.xls](#)

Table4S

[Click here to download Background dataset for online publication only: Table4S.xls](#)

Table5S

[Click here to download Background dataset for online publication only: Table5S.xls](#)