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1 **Mantle heterogeneities produced by open-system melting and melt/rock reactions in**
2 **Patagonian extra-Andean backarc mantle (Paso de Indios, Argentina)**

3

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23

24 **ABSTRACT**

25 The Eocene basaltic extrusions in the Paso de Indios region (Chubut-Argentina) are one
26 manifestation of the extensional tectonism of the active margin of South America during the
27 Cenozoic. Ultramafic xenoliths embedded in these volcanics are mainly harzburgites and
28 lherzolites with subordinate pyroxenites, estimated equilibrium temperatures ranging from
29 853 ± 15 to $1057 \pm 32^\circ\text{C}$ and pressures in the spinel stability field.

30 Geochemical and modal evidences point to a multistage magmatic history with record of a
31 last reactional open-system episode associated to the influx of adakitic-like melts in a
32 orthopyroxene-rich, clinopyroxene-poor mantle column. The great variability of
33 clinopyroxene modal and geochemical composition in a ~ 20 km² area suggests extreme
34 variability of the physical parameters connected to melt infiltration and melt/rock reactions

35 processes at a very small scale superimposed on a mantle with an inherited meter scale
36 heterogeneity. Variations in the melt influx rate and residual porosity of the mantle column
37 produced different melt/rock reactions which could be summarized in two entangled main
38 reaction pathways: 1) $\text{opx}+\text{cpx}+\text{melt1} \rightarrow \text{ol}+\text{melt2}$ and 2) $\text{opx}+\text{melt1} \rightarrow \text{cpx}+\text{ol}+\text{melt2}$.
39 These reactions deeply modified the trace elements content of clinopyroxenes producing
40 variable enrichments in LREEs and LILEs related to both chromatographic and pure
41 incremental open system processes.
42 Petrological evidence suggests that the last reactional process occurred in the spinel stability
43 field overprinting a strongly depleted mantle that, in a previous stage, had experienced
44 extreme depletion in the garnet stability field, possibly under hydrous conditions.
45 The adakitic-like nature of the influxing melt associates this episode to the subduction
46 system along the western margin of South America, active at least since Late Triassic times.

47

48 **1. Introduction**

49 The South American sublithospheric mantle records a complex sequence of depletion and
50 refertilization episodes linked to the geodynamic evolution of supercontinents and large
51 igneous provinces (e.g. Mazzucchelli et al., 1995, 2016; Rivalenti et al., 1995, 2004a, 2007a,
52 2007b; Correia et al., 2012; Faccini et al., 2013; Giovanardi et al., 2018, 2019; Roverato et
53 al., 2019; Liu et al., 2019a; 2019b; Melchiorre et al., 2020). In this context, Patagonia
54 (southern Argentina, $\sim 38^{\circ}$ - 53° S) is a relatively large area, comprising part of the back-arc
55 region of the Andean arc, in which Pliocene-Quaternary alkaline basalt volcanism is
56 randomly distributed (Stern et al., 1990). Mantle xenoliths from more than twenty well
57 known of these basaltic outcrops, record geochemical evidence of depletion, metasomatism
58 and refertilization processes related to the influence of the subduction system active at least
59 since the Triassic along the western border of South America (Melchiorre et al., 2020 and
60 references therein).

61 The very first Patagonian mantle xenoliths ever studied were from the localities of Comallo
62 and Los Adobes in the North (Gelós and Hayase, 1979). Decades later, mantle xenoliths
63 collected and studied from the entire Patagonia, from the Payenia region in the North to the
64 Tierra del Fuego to the South, were found to be mainly equilibrated in the spinel stability
65 field (Bjerg et al., 1991, 2005; Barbieri et al., 1999; Bertotto, 2000; Acevedo and Quartino,
66 2004; Rivalenti et al., 2004a; Jalowitzki et al., 2010; Bertotto et al., 2013; Mazzucchelli et
67 al., 2016; Melchiorre et al., 2020). Garnet-bearing peridotites were identified only in
68 Prahuaníyeyu in northern Patagonia (Ntaflos et al., 2001; Ntaflos and Bjerg, 2006; Bjerg et

69 al., 2009) and in Pali Aike, in the **S**outh (Skewes and Stern, 1979; Douglas et al., 1987;
70 Kempton et al., 1999; Vannucci et al., 2002; Zaffarana et al., 2014).

71 In the area of Paso de Indios (PDI, northern Patagonia), several outcrops of Paleogene basalts
72 known as Cerro León, Cerro Matilde and Cerro Chenque (hereafter named León, Matilde
73 and Chenque) contain strongly heterogeneous mantle xenoliths in the spinel facies (Gelós
74 and Hayase, 1979; Alric et al., 1993; Labudía, 1994; Alric, 1996 Rivalenti et al. (2004a) with
75 HREE in clinopyroxene that suggest enhanced partial melting depletion (Rivalenti et al.,
76 2004a; this work). This is a unique case in the Patagonia region, where the mantle column
77 is generally completely overprinted by melt intrusions and refertilization events (Bjerg et al.,
78 1991, 2005; Barbieri et al., 1999; Bertotto, 2000; Acevedo and Quartino, 2004; Rivalenti et
79 al., 2004a; Jalowitzki et al., 2010; Bertotto et al., 2013; Mazzucchelli et al., 2016; Melchiorre
80 et al., 2020). A few granulitic xenoliths found in Chenque were interpreted as solid residues
81 after separation of granitic to granodioritic melts (Castro et al., 2011). A zircon U-Pb age
82 dates this event at 175.9 ± 4.9 Ma during a time of complex lithospheric evolution for
83 Patagonia in a geodynamic context preceding the rupture of Gondwana.

84 In spite of all these studies, the reported petrochemical information is insufficient to
85 characterize in detail the composition of the lithospheric mantle and the associated
86 petrological processes that affected the PDI mantle column. This is due to the low number
87 of samples and the fact that, among the several outcrops of the area, the studied xenoliths
88 are mainly from Chenque (Rivalenti et al., 2004; Bjerg et al., 2005). A more recent
89 petrological study, based on the major element composition of mineral phases of xenoliths
90 from all three basaltic outcrops in the PDI area has shown more variable geochemical and
91 petrological features than previously reported (Ponce et al., 2015). We have increased the
92 number of xenoliths studied in this area and analyzed the trace element composition of
93 clinopyroxenes from the xenoliths of Ponce et al (2015) and new xenoliths of this collection.
94 Combining new and old petrographic observations and major elements analyses with our
95 complete trace element dataset, we discuss evidence supporting open system partial melting
96 processes as responsible for the petrochemical characteristics of the mantle column below
97 the PDI area.

98

99 **2. Geological background**

100 The study area is located in the extra-Andean zone of north Patagonia near the town of Paso
101 de Indios (PDI, Chubut province, Argentina) (Fig. 1) and is part of the geological province
102 of the Patagónides (Ramos, 1999; Giacosa and Márquez, 1999). In this area, ultramafic

103 xenoliths have been reported in eight basaltic outcrops (Alric, 1996) and belong to the El
104 Buitre Formation (Ardolino and Franchi, 1993). We analyzed samples from three relicts of
105 Eocene lava flows and dykes with alkaline composition, namely Matilde, León and Chenque
106 (Fig. 1). These basalts cover and intrude Cretaceous sedimentary rocks and are partially
107 covered by Quaternary debris (Alric et al., 2002; Anselmi et al., 2004; Silva Nieto, 2005).
108 They are related to the Paleogene Rift of the Patagónides (Aragón et al., 2015), which formed
109 when extensional tectonics affected the active margin of southern South America from the
110 Paleocene to the Oligocene. This extensional event was triggered by the arrival of the
111 transform plate boundary at the subduction zone, with the tearing of the Farallón-Aluk plate
112 triggering the formation of a slab window (Aragón et al., 2011).

113 Chenque is a basaltic dyke, about 50 meters long, located at 68°56'37" W and 43°38'37" S
114 with an estimated age of 52 Ma based on the geochemical affinity with similar basaltic dykes
115 of the region (Alric et al., 2002). Matilde is a remnant of a lava flow at 68°55'27" W and
116 43°48'42" S with an Ar-Ar age of 49.35 ± 0.74 Ma (Alric, 1996). Lastly, León, partly covered
117 by basaltic debris and recent sediments, is 50.17 Ma old (± 0.41 ; Alric, 1996) and located at
118 69°0'14" W and 43°42'9" S (Fig. 1).

119

120 **3. Analytical methods**

121 In this work, we discuss new chemical data of eighteen samples (12 from Chenque, 2 from
122 Matilde and 4 from León) together with eleven samples (5 from Chenque, 5 from Matilde
123 and 1 from León) previously analyzed for major elements only (Ponce et al., 2015). Modes
124 are reported in Table 1.

125 As discussed below, the occurrence within the xenoliths of glass veins from the host basalt
126 has produced strong reaction at the contact with the mineral phases. However, given that our
127 aim is to reconstruct the composition and characteristics of the original mantle column
128 beneath the PDI area, all our data have been collected at the mineral core. Similar studies
129 have already demonstrated that the reaction of minerals with host basalt affected their
130 composition only at the rim thus preserving their original composition in the cores (e.g.
131 Laurora et al., 2001).

132 Whole rock major and trace element were analyzed by inductively coupled plasma atomic
133 emission spectrometry (ICP-AES) at the "Université de Bretagne Occidentale" (Brest,
134 France) following methods described in Cotten et al. (1995). To remove alteration effects,
135 the whole-rock powders were leached with 2.5N HCl for 10 min in an ultrasonic bath and
136 rinsed three times in ultrapure water. Relative standard deviations were <2% for major

137 elements, Rb and Sr and <5% for the other trace elements. Data are reported in
138 Supplementary Material 1, Table 1.

139 Major elements in mineral phases were determined at the Dipartimento di Scienze della
140 Terra of the Università di Milano (Italy) with a JEOL JXA-8200 electron microprobe using
141 a 15 kV accelerating voltage, 15 nA beam current, 1-3 μm beam diameter, 30 s counting
142 time on the peaks and 10 s on the background. Natural minerals (olivine for Mg; omphacite
143 for Na; ilmenite for Ti; rodonite for Mn; K-feldspar for K; anorthite for Al and Ca;
144 wollastonite for Si; fayalite for Fe and nicolite for Ni) and synthetic chromite were used as
145 standards. The results were corrected for matrix effects using the conventional ZAF method
146 provided by the JEOL suite of programs. Results are considered to be accurate within 2-6%.
147 Details on the methodology are reported in Giovanardi et al. (2020). Averages of each
148 sample are presented in Tables 2-5 and the entire dataset is reported in Supplementary
149 Material 2.

150 Trace-elements in clinopyroxene and orthopyroxene have been analyzed at Consiglio
151 Nazionale delle Ricerche - Istituto di Geoscienze e Georisorse (Section of Pavia, Italy) by
152 laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS) using a Perkin
153 Elmer ELAN DRC-e quadrupole mass spectrometer coupled with a Q-switched Nd:YAG
154 laser source Brilliant, whose fundamental emission in the infrared region (1064 nm) is
155 converted into 266 nm by two harmonic generators. The laser was operated at a repetition
156 rate of 10 Hz, with a pulse energy of ~ 35 mJ. Helium was used as carrier gas and was mixed
157 with Ar downstream of the ablation cell. Spot diameter was adjusted in the range of 40-80
158 μm . Data reduction was performed offline using the GLITTER software. For this study, the
159 NIST SRM 610 synthetic glass reference material was used as external standard and CaO
160 was used as internal standard for clinopyroxene. Precision and accuracy of the REE
161 concentration values were assessed through repeated analysis of the BCR2-g standard to be
162 better than $\pm 7\%$ and $\pm 10\%$, respectively at the ppm concentration level (further details are
163 reported in Zanetti et al., 2016). Average of each sample is reported in Table 6. The entire
164 dataset is reported in Supplementary Material 3.

165

166 **4. Petrography**

167 The studied xenoliths are anhydrous spinel-facies peridotites and pyroxenites. Based on
168 modal analyses (by count pointing, at least 600 points for each sample) they can be classified
169 as lherzolite (13 samples), harzburgite (12 samples), websterite (3 samples) and olivine-
170 websterite (1 sample) (Fig. 2; Table 1). The predominant texture (after Harte, 1977) is

171 porphyroclastic, followed by the transitional coarse to porphyroclastic texture and, to a lesser
172 extent, the coarse and mosaic-porphyroclastic textures (Fig. 3A-D). Among the three
173 outcrops, Matilde xenoliths seem to be the most depleted, represented by harzburgites (five
174 samples) and lherzolites (two samples; Fig. 2). León xenoliths comprise lherzolites (two
175 samples), harzburgites (two samples) and one olivine-websterite (Fig. 2; Table 1). Xenoliths
176 from Chenque show the highest modal variability with lherzolites (nine samples),
177 harzburgites (five samples) and three websterites (Fig. 2; Table 1). Considering the modal
178 variability of the PDI xenoliths (Fig. 2B), peridotites with low olivine are orthopyroxene-
179 rich and clinopyroxene-poor compared to the modal composition of the primitive mantle
180 (PM, Warren, 2016;). We now briefly synthesize the petrographic characteristics of the
181 studied xenoliths, given that all details can be found in Ponce et al. (2015).

182 Olivine is present as large, mainly anhedral crystals up to 12 mm in length and small anhedral
183 to subhedral crystals up to 0.5-2 mm. The larger olivines are kink-banded and the smaller
184 are unstrained. Frequently, in samples of León and rarely in Matilde and Chenque, grains of
185 1-2 mm of unstrained olivine occupy embayments of large orthopyroxene suggesting
186 replacement (Fig. 3E).

187 Orthopyroxene is anhedral and up to 12 mm in diameter. Large grains commonly show a
188 thin regular pattern of clinopyroxene exsolution lamellae and rarely kink-bands less
189 developed than those of olivine. Vermicular orthopyroxene, sometimes associated with
190 clinopyroxene and rarely with small and unstrained olivine, frequently replaces large
191 strained olivine (Fig. 3F). Orthopyroxene shows partial to total replacement when in contact
192 with the host lava by clinopyroxene + olivine + glass ± orthopyroxene ± spinel group
193 minerals ± plagioclase (Fig. 3G). Within the xenolith, orthopyroxene can develop reaction
194 rims linked to veins of altered glass directly connected to the host basalt similar to what
195 reported above.

196 Clinopyroxene is anhedral, reaching up to 3 mm in diameter. Rarely, clinopyroxene has
197 cloudy appearance because of the occurrence of minute (<30 µm diameter) melt inclusions.
198 In olivine-rich harzburgites, clinopyroxene is interstitial, and sometimes is associated with
199 secondary olivine replacing primary orthopyroxene. Rarely, clinopyroxene shows spongy
200 rims with the development of a corona frequently linked to veins from the host basalt.

201 Spinel is mainly anhedral and reaches 2.7 mm in size. Holly-leaf shaped spinel is the most
202 common and frequently observed in the porphyroclastic samples. Less frequently subhedral
203 and euhedral spinels are included in olivine and are also present as rounded grains associated
204 with secondary olivine and clinopyroxene replacing larger, primary orthopyroxenes.

205 Spinel+ol±opx irregular symplectites could represent garnet breakdown (Supplementary
206 Material 4, Fig. 14) but according to Seyler et al. (2007) similar symplectites could also
207 originate from melt reaction percolation in the spinel facies.
208 As shown in Ponce et al. (2015), frequently the xenoliths bear veinlets of altered
209 (serpentinized) glass from the host basalt.

210

211 **5. Whole rock chemistry**

212 Five whole rock analyzes were obtained from the core portion (2-3 cm in diameter) of
213 xenolith samples larger than 10 cm. The abundance of MgO ranges between 42.6 and 44.0
214 wt.%. The compositional variation of MgO is within the range of those xenoliths previously
215 reported for Patagonia (Rivalenti et al., 2004a, 2004b; Bjerg et al., 2005) and MgO shows a
216 negative correlation with Al₂O₃, CaO, TiO₂ and Na₂O content (Supplementary Material 1,
217 Fig. 1). As a whole, PDI harzburgites and lherzolites are, on average, between the more
218 refractory spinel facies xenoliths from South America (Rivalenti et al., 2004a, 2004b; Bjerg
219 et al., 2005; Ntaflou et al., 2007). Further discussion about whole rock data can be misleading
220 because of the disturbing influence of intergranular glass films or veins (see Supplementary
221 Material 1 and 4, Table 1). The infiltrating melt shows strong reactivity with the xenoliths
222 minerals as suggested by the extreme variability in the chemical composition of the veins
223 (Supplementary Material 4 Table 1; SEM analysis) and consistent with reactive porous flow
224 of initially ne-normative infiltrating melt that evolved through silica saturation (Rivalenti et
225 al., 2004b). However, this interaction is limited to the rims of the minerals (Laurora et al.,
226 2001; Rivalenti et al., 2004b) and therefore the cores preserve the original chemical
227 composition.

228

229 **6. Mineral chemistry**

230 Previous work on the PDI xenoliths showed that primary and secondary phases are quite
231 homogenous with negligible variations in major element composition (Ponce et al., 2015;
232 Ponce, 2016). A similar behavior is found in our samples for both major and trace elements.
233 For further discussion see Supplementary Material 4. We thus report and discuss averages
234 only (olivine in Table 2; spinel in Table 3; orthopyroxene in Table 4 and clinopyroxene in
235 Table 5) while the entire dataset of single analyses is reported as Supplementary Material 2.

236

237 *6.1. Olivine*

238 The forsterite (Fo, calculated as $Mg^{2+}/(Mg^{2+}+Fe^{2+}_{tot})\cdot 100$ molar ratio) content varies
239 between 90.6-92.3%, within the previously reported variability range of the PDI xenoliths
240 (Rivalenti et al., 2004a). The content of NiO and CaO varies between 0.31-0.46 wt.% (61
241 analyses) with four analyses up to 0.86 wt.% and 0.03-0.22 wt.%, respectively (Table 2).
242 Harzburgitic olivines commonly have higher MgO and NiO (Fig. 4-A), but lower CaO than
243 lherzolites. Olivines from León are more restitic and have higher Fo than Matilde olivines.
244 Chenque olivines overlap the variability fields of León and Matilde.

245

246 6.2. Spinel

247 Spinel Cr# (calculated as $Cr^{3+}/(Cr^{3+}+Al^{3+}_{tot})\cdot 100$ molar ratio) varies between 16.6 and
248 77.5% and Mg# between 58.1 and 79.1% (Fig. 4-B). According to the classification of
249 Schulze (2001), minerals are mainly spinel, but in lherzolites, harzburgites and ol-
250 websterites from León and one harzburgite from Matilde they are magnesiochromite (Fig.
251 4-B). New samples from León show the highest Cr# values while the Chenque spinels have
252 the highest Mg# (Fig. 4-B).

253

254 6.3. Orthopyroxene

255 Orthopyroxene is enstatite and its composition is in the range En_{86-92} , Ws_{1-2} and Fs_{7-12} and
256 Mg# (calculated as $Mg^{2+}/(Mg^{2+}+Fe^{2+}_{tot})\cdot 100$ molar ratio) varies between 82.4 and 93.0%.
257 The content of MgO in the majority of the orthopyroxenes ranges from 28.88 to 35.03 wt.%
258 with Q150 websterite of Chenque with the lowest MgO content (28.88 wt.%) (Table 4).
259 Websterites and olivine-websterites have the lowest Mg# values, between 81.4 and 87.8%
260 with the exception of sample L90 (Fig. 4-C; Table 4). No significant distinction has been
261 observed between orthopyroxenes of the three outcrops. Overall, our orthopyroxenes fall
262 within the range defined by Melchiorre et al. (2020; and references therein) for North
263 Patagonia mantle peridotites (Fig. 4-C) with the exception of the following three samples
264 one lherzolite from León (sample L92) and one lherzolite and one harzburgite from Chenque
265 (samples Q155 and Q109, respectively).

266

267 6.4. Clinopyroxene

268 Clinopyroxene is diopside with En_{47-51} , Wo_{44-48} and Fs_{2-7} composition. It is characterized by
269 an increase of Na_2O and Al_2O_3 with a decrease of CaO. Na_2O varies in the range of 0.40-
270 1.55 wt.%, TiO_2 is between 0.01-0.38 wt.%, Cr_2O_3 between 0.22-1.27 wt.% and Mg#
271 between 87.2-95.6 (Table 5). As for the orthopyroxenes, our clinopyroxenes mainly fall in

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272 the compositional field defined for North Patagonia by Melchiorre et al. (2020; and
273 references therein) (Fig.4-D), with the exception of four clinopyroxenes from Chenque and
274 two from León, therefore expanding the clinopyroxene compositional field reported for
275 mantle xenoliths in this region.

276

277 **7. Clinopyroxene trace element chemistry**

278 Clinopyroxene trace element composition of each sample is very homogenous and contrasts
279 with a marked inter-sample heterogeneity. Here, we focus on the REE patterns that well
280 picture the extreme variability in element distribution. We recognized five significant groups
281 based on clinopyroxenes REE patterns (Fig. 5).

282 **Group I** clinopyroxenes are depleted in LREE and were recognized only in one lherzolite
283 (Q155) from Chenque. This sample pattern approaches the model curves for 11-14% melting
284 in the spinel field (Fig. 5-A). This sample is the only one approaching a near fractional
285 melting trend.

286 **Group II** clinopyroxenes are characterized by patterns enriched in LREE and relatively flat
287 to weakly concave upward. They comprise clinopyroxenes from one lherzolite (M53) and
288 one harzburgite (M55) from Matilde, one lherzolite (L91), one harzburgite (L69) and one
289 olivine-websterite (L90) from León, and one lherzolite (Q103) and one harzburgite (Q108)
290 from Chenque. The LREE enrichment is clearly due to melt/rock reaction processes.
291 However, a high degree of depletion can be inferred for a possible protolith if assuming
292 HREEs as representative, or close to, the pre-reactive composition. These degrees of
293 depletion require up to 20% non-modal partial melting of a primitive mantle (Fig. 5-B). Two
294 clinopyroxene patterns from PDI xenoliths published by Rivalenti et al (2004a) have M- to
295 H-REE that correspond to PM clinopyroxene composition (Fig. 5-B). The LREE of group II
296 clinopyroxene are strongly fractionated, with La/Yb_N between 1.3 and 68.0. Trace elements
297 in clinopyroxenes with fractionated HREE show negative anomalies in Ba and Ti
298 accompanied by depletion in Nb and Ta and Zr and Hf compared to “flat” samples (Fig. 5-
299 B). Conversely, they have higher concentration for Th and U with respect to Group V.
300 Matilde samples also show negative Pb negative anomaly.

301 **Group III** clinopyroxenes have V/U-shaped patterns and are found only in xenoliths from
302 Chenque (samples Q65, Q90 and Q101) and Matilde (samples M57, M67, M79 and M82).
303 Lherzolites show mainly V-shaped patterns with a common minimum in Nd and are the most
304 enriched in HREE (Fig. 5-C; Sm/Yb_N between 0.04 and 0.19). Harzburgites show U-shaped
305 patterns with more fractionated and less abundant HREE (Fig. 5C; Sm/Yb_N between 0.40

306 and 7.5). Matilde clinopyroxenes are enriched in LREE and depleted in HREE with respect
307 to those of Chenque. They also have a higher concentration of incompatible elements such
308 as Rb, Th, U and Ba and lower Ti compared to Chenque (Fig. 5-C). Similarly to Group II,
309 these clinopyroxenes are enriched in Th and U and have negative anomalies in Ba, Pb and
310 Ti.

311 **Group IV** clinopyroxenes have sinusoidal patterns and belong to three lherzolites (Q105,
312 Q107, Q154) and one harzburgite (Q109) from Chenque (Fig. 5-D). They exhibit HREE
313 patterns consistent with a primitive mantle residual clinopyroxene depleted by 8 to 20% of
314 melt extraction. Among the clinopyroxenes showing REE patterns different from the
315 primitive mantle (groups II-V), samples with sinusoidal REE patterns show the lowest
316 abundances with values equal or below the primitive mantle concentration (Fig. 5-D). They
317 show Ba negative anomaly and a variable positive to negative Pb anomaly. Ti, which in other
318 groups shows negative anomalies, is flat or slightly depleted.

319 **Group V** clinopyroxenes are characterized by “flat” patterns and are represented by a
320 number of clinopyroxenes from Chenque (samples Q58, Q77, Q80, Q150, Q152, Q153 and
321 QEug), Matilde (sample MH) and León (samples L82 and L92) (L_{AN}/Yb_N from 1.01 to 1.88)
322 (Fig. 5-E). The HREE fit with a fractional melting depletion between 10 and 17%, whereas
323 the MREE and LREE deviate from the model (Fig. 5-E). All three websterites from Chenque
324 belong to this group. All samples show negative anomalies in Ba and slightly negative
325 anomalies for Ti. Thorium, U, Nb and Ta vary from depleted to enriched with respect to the
326 primitive mantle, with the lower values in websterite and the higher values in peridotites.

327

328 **8. Geothermobarometry**

329 We carried out geothermometrical calculations based on two pyroxene equilibria (Brey and
330 Köhler, 1990; Taylor, 1998) and Ca-in orthopyroxene (Brey and Köhler, 1990).
331 Temperatures were calculated for each couple with core analyses of crystals in contact with
332 each other. Averages for each sample are reported in Table 7. An extensive discussion about
333 the precision and accuracy of the geothermometers used is reported in Nimis and Grütter
334 (2010). Hereafter we discuss only the estimated temperatures of Ca-in orthopyroxene
335 geothermometer to compare our data with those of Patagonian spinel peridotites reported in
336 the literature (e.g. Rivalenti et al., 2004a; Bjerg et al., 2005; Bertotto et al., 2013).

337 At Chenque, temperatures vary between $853 \pm 15^\circ\text{C}$ (websterite Q150) and $975 \pm 5^\circ\text{C}$
338 (lherzolite Q155) with an average of $905 \pm 34^\circ\text{C}$. Chenques' lherzolites have higher
339 estimated T compared to the harzburgites (between $879\text{-}975^\circ\text{C}$ and $868\text{-}930^\circ\text{C}$, respectively;

340 Table 7). In Matilde, temperatures range from 859°C (harzburgite M55) to 1057 ± 32°C
341 (harzburgite M57) with an average of 903 ± 76°C. The average estimated T is higher in
342 lherzolite (908 ± 66°C) than in harzburgite (902 ± 87°C) but they are comparable within
343 errors. In León, temperatures range between 860 ± 44°C (lherzolite L91) and 981°C
344 (lherzolite L92) with an average T of 914 ± 45°C. Similarly to Chenque and Matilde,
345 lherzolite in León shows higher average T with respect to harzburgite (920 ± 86°C and 899
346 ± 10°C, respectively).

347 The equilibrium pressures were estimated based on the Mercier (1980) clinopyroxene
348 geobarometer and ~~vary~~ varied between 1.4 ± 0.3 and 2.9 ± 0.1 GPa in Chenque, between 1.4
349 and 3.7 ± 1.3 GPa in Matilde and between 1.6 and 2.7 ± 0.1 GPa in León (Table 7 and Fig.
350 6). We note here that in the literature (see discussion in Yang et al., 1998), it is well known
351 that pressure estimations in the spinel field could be affected by large errors (e.g. 3.7 ± 1.3
352 GPa) and that the best estimates are given by Mercier (1980) geobarometer.

353 The comparison of the entire PDI data set (Ponce et al., 2015; and this work) with estimated
354 temperatures for Patagonian spinel peridotites reported in literature shows that the lowest
355 temperatures are indeed those of the PDI samples (Fig. 6). Bjerg et al. (2005) previously
356 reported a T of 1030°C for a PDI harzburgite (CH16-A), and Rivalenti et al. (2004a)
357 estimated a T interval between 839 and 1197°C for eight mantle xenoliths in this area. The
358 overall range of temperatures of the studied xenoliths varies between 853 ± 15 and 1057 ±
359 32°C, with an average of 906 ± 47°C.

360

361 **9. Discussion**

362 *9.1. Mantle source and melting processes*

363 As shown by the mineral modal composition, the studied peridotites are markedly depleted.
364 The abundance of harzburgites and lherzolites with less than 6% clinopyroxene suggests that
365 the mantle column below the PDI area is significantly more depleted than the primitive
366 mantle reference (Warren, 2016; Fig. 2). This is true for most of the Patagonian mantle
367 xenoliths described in literature (Laurora et al., 2001; Rivalenti et al., 2004a, 2007a, 2007b;
368 Bjerg et al., 2005) suggesting that the regional mantle is strongly depleted. However, the age
369 of this depletion process is still controversial. According to Schilling et al. (2008), xenoliths
370 from the North Patagonian Massif (comprising those of the PDI area) have a range in
371 ¹⁸⁷Os/¹⁸⁸Os similar to modern oceanic peridotites with T_{RD} and T_{MA} ages varying from 0.03
372 to 0.98 Ga, indicative of a relatively recent lithospheric mantle formation process from a
373 convecting mantle. Granulite xenoliths from the PDI area have been dated by U-Pb SHRIMP

374 zircon ages to 175.9 ± 4.9 Ma (Castro et al., 2011), which is interpreted as the formation
375 time of this residual lithosphere.

376 The refractory character of the mineral associations is consistent with the bulk composition,
377 with MgO abundances similar to the more depleted spinel facies xenoliths of Patagonia
378 (Supplementary Material 1 Table 1). Accordingly, the major elements composition of
379 mineral phases is mainly Al-poor and Mg and Cr-rich (Tables 2 to 5).

380 The chemical composition of mineral phases and the variation in modal composition of the
381 PDI xenoliths do not follow a melting trend. In comparison with the modal and mineral
382 composition of global abyssal residual mantle (Warren et al., 2016), a reference for near
383 fractional (extraction-dominated) melting processes in the mantle, and in particular with
384 peridotites from the Vema Lithospheric Section, representative of the integrated melting
385 events over a single mantle parcel (Brunelli et al., 2006; Cipriani et al., 2009), the PDI mantle
386 rocks show an excess of clinopyroxene, Yb and higher Cr# for both clinopyroxene and spinel
387 (Fig. 7). PDI pyroxenes appear to be more residual than abyssal peridotites, having higher
388 Mg# and lower Al_2O_3 (Fig. 7), but unexpectedly, they have lower Cr_2O_3 than abyssal rocks
389 (Fig. 7). Cr behaving as compatible during melting is expected to be higher for these Al and
390 Mg values. At the same time, clinopyroxene Na_2O is higher than in the abyssal rocks.

391 These relationships are consistent with extensive melt-rock reactions resulting in a
392 decoupling between modal and chemical compositions of the xenoliths (Fig. 7). High Na
393 contents and positive Na-Cr correlations in clinopyroxene are indicative of reactions with
394 Na-rich/Cr-poor melts under near-batch conditions as shown by Seyler and Brunelli (2018).

395 The lack of correlation between modal composition and mineral chemistry is shown by both
396 major and trace elements (see Table 1; Fig. 7). For instance, no correlation between mode
397 and chemical indicators of the degree of depletion is observed when grouping samples by
398 clinopyroxene abundance (Fig 7A, B). The observed modal-chemical distribution falls
399 outside of the residual mantle array used to calibrate melting proxies such as Cr# or Yb_N
400 (e.g. Warren, 2016), therefore a relative degree of melting cannot be inferred for our sample
401 suite. Overall the PDI xenoliths indicate a remarkable depletion of the mantle column prior
402 to the younger metasomatic/refertilization processes generating the observed REE patterns
403 and the modal distribution.

404 Moreover, the variation of modal composition of the PDI xenoliths cannot be reproduced by
405 the simple melting of a PM or DM mantle (Fig. 2B). With the exclusion of the websteritic
406 samples, the two pyroxenes are related to olivine by complex trends, with a general

407 decreasing of orthopyroxene by increasing olivine and the persistence of clinopyroxene even
408 at very high olivine contents (Fig. 2B).

409 At low olivine contents, orthopyroxene exceeds the expected composition for the modal
410 variation of a residual mantle after melt extraction in the spinel facies. This suggests the
411 presence of inherited local orthopyroxene-rich and clinopyroxene-poor regions in the PDI
412 mantle before the last refertilization process. High modal orthopyroxene can result from
413 fluid-assisted partial melting by injection of Si-rich melts as those originated in a subducted
414 slab (e.g. Mazzucchelli et al., 2009; Zanetti et al., 2016; Jing et al., 2019; Giovanardi et al.,
415 2020).

416 The general modal decrease in orthopyroxene and clinopyroxene at increasing olivine
417 follows the “normal” depletion trend in the spinel field: $opx+cpx+melt1 \rightarrow ol+melt2$.

418 However, steep trends of decreasing orthopyroxene and increasing clinopyroxene suggest
419 that the last melting event was marked by melt/rock interaction in which clinopyroxene
420 crystallized at the expense of orthopyroxene. The instability of orthopyroxenes is confirmed
421 by the petrographic observations of secondary olivine replacing large orthopyroxene in
422 olivine rich peridotites (Fig. 3E, H). [This](#) evidence suggest that the main melt/rock reaction
423 during the melting process follows the reaction: $opx+melt1 \rightarrow ol+melt2\pm cpx$.

424 Experimental studies have shown that under hydrous conditions the orthopyroxene stability
425 field is reduced by the expansion of olivine and clinopyroxene fields in the spinel-facies (e.g.
426 see Fig. 3 in Asimow and Longhi, 2004). We stress therefore that the decreasing opx/cpx
427 modal trends have been possibly generated under hydrous conditions.

428 The low clinopyroxene Cr content can be explained by the formation of pyroxenes by
429 reaction with Cr-poor melts (as all mantle-derived melts are) and thus sustaining further the
430 process of clinopyroxene crystallization during melt/rock reaction. This reasoning is
431 however complicated by the fact that also orthopyroxene is depleted in Cr and Al (Table 4).

432 If only clinopyroxene is generated during melting, assuming a short equilibration time, we
433 should expect the orthopyroxene to record a more “residual” fingerprint (i.e. higher Cr for a
434 given Al content). However, the two pyroxenes are clearly equilibrated to low Cr, thus
435 suggesting an efficient re-equilibration during melt/rock reaction hiding the different genesis
436 of the two minerals. The low Cr content can also be a clue to the origin of the orthopyroxene
437 enrichment. It would point to a reactive origin in [the](#) presence of melt, such as an open-
438 system melting in a hydrous environment deep in the garnet stability field. This suggests that
439 melting before the last event largely occurred in the garnet-facies followed by
440 decompression in the spinel stability field where the orthopyroxene consuming reaction

441 occurred. This scenario is compatible with a mantle parcel rising within a suprasubduction
442 setting being percolated by hydrous melts derived from deep slab dehydration. A first
443 extreme depletion episode with large orthopyroxene crystallization in the garnet facies is
444 followed by a shallower (spinel-facies) event where orthopyroxene is frequently consumed
445 in favor of clinopyroxene. The low Al, Fe and high Mg contents described are therefore
446 related to depletion processes associated with long integrated melt extraction. The exception
447 shown by Cr might be a proof of the progressive clinopyroxene crystallization and at the
448 same time a consequence of the orthopyroxene enrichment in the source before the last
449 melt/rock reaction event.

450 Overall the melting process hints at the last melt/rock reaction and melt extraction events
451 occurring in open-system conditions. This is also suggested by the clinopyroxene trace
452 element composition. Mildly incompatible trace elements are overall depleted; the patterns
453 variability and the partial enrichment of the most incompatible elements may mask the
454 overall paucity of the incompatible content (Fig. 5). During melt/rock reaction the HREE are
455 barely modified and usually hard to be enriched under different P-T conditions and chemical
456 composition of the infiltrating melt (Ozawa, 2001; Brunelli et al., 2014). The overall
457 depletion of the PDI pyroxenes rules out the possibility of pure melt trapping as the leading
458 process. A large pattern and enrichment variability can instead be created during conditions
459 of open-system melting with melt infiltration (see model patterns in Brunelli et al., 2014). In
460 an open-system scenario the HREE is generally depleted by progressive melt extraction with
461 negligible effect on their abundances by the infiltrating melt.

462 The compositional variability observed also suggests a variation of the apparent degree of
463 melting that cannot be justified in terms of lateral variations of temperature due to the small
464 distance between the eruptive centers. Moreover, the modal and chemical variabilities are
465 extremely high within the same locality (Figs 2 and 5). Lack of any correlation between
466 modal and chemical composition may suggest that we are observing the results of a process
467 acting at a small scale, for both geometry (size of reacting cells) and composition (variability
468 of the percolating melts with respect to the composition of the percolated cell).

469 It means that percolating processes act over a short-scale (meter), thus making impossible to
470 disentangle single events. Partial melting may be locally enhanced by progressive melt
471 channeling which includes dissolution of pyroxenes, compositionally mimicking the melt-
472 depletion tendency (Kelemen, 1990). In this context, the occurrence of dunites in the PDI
473 area (Rivalenti et al., 2004a; Bjerg et al., 2005) can be interpreted as representing extreme

474 terms of pyroxene dissolution due to the channeling of silica sub-saturated melts to relatively
475 low pressure (Mazzucchelli et al., 2009).

476

477 9.2. Melt/rock reactions

478 As previously discussed, evidence shows that the xenolith endured at least two reactive
479 events during the last melting process along with the infiltration of the host-basalt at low
480 pressure (Fig. 3; Supplementary Material 4). The infiltration of the host basalt occurred
481 through decompressional fractures originated during the dismembering of the mantle and the
482 extraction of the xenoliths during mechanical relaxation. In such conditions, the carrying
483 melt modified only the chemical composition of the xenoliths' minerals at the rim, in contact
484 with the host basalt or with intruding veins (Laurora et al., 2001; Rivalenti et al., 2004b; Fig.
485 3G; Supplementary Material 4).

486 The last main melting process within the PDI mantle column produced instead an
487 homogenization of the mineral composition as observed by the similar composition of
488 primary and secondary minerals within the same sample (Supplementary Material 4). The
489 fluids/melts modified the trace element imprint as observed in the variable REE and
490 extended patterns of clinopyroxene, mostly incongruent with the fractional melting model
491 (Fig. 5). As previously discussed, it is difficult to disentangle the melt/rock reactions and the
492 melting process.

493 Group I sample seems the least affected by melt/rock reactions as seen in the REE pattern
494 and modal composition. A similar pattern was recognized in clinopyroxenes from group 1
495 of Agua Poca (Bertotto et al., 2013), which also show comparable Th and U concentrations.
496 The clinopyroxene trace elements concentration, with almost flat HREE and MREE, is
497 consistent with non-modal fractional melting in spinel-facies conditions (Fig. 5). Modeling
498 of the melt in equilibrium with the clinopyroxene of Group I (Fig. 8-A; clinopyroxene/melt
499 partition coefficient are from Ionov et al., 2002) is consistent with a depleted melt.

500 Group V 'flat' patterns are characterized by uniform values of L- and MREE and are enriched
501 compared to the non-modal fractional melting modeled clinopyroxene composition. This
502 behavior suggests a more prolonged melt/rock interaction episode compared to Group II, III
503 and IV xenoliths, with the residual mantle of this region attaining relative equilibrium with
504 the interacting melt. Similar parallel, almost horizontal patterns have been observed in
505 clinopyroxenes from Cerro de los Chenques (160 km SW of PDI; Rivalenti et al., 2007a)
506 and from Cerro Chenque (north of the Somuncurá plateau, Mundl et al., 2015). Potential
507 melts in equilibrium with the clinopyroxene of Group V are the best candidates for the

508 original composition of this migrating melt because of the longer melt/rock interaction with
509 respect to the other clinopyroxenes. The melt in equilibrium with Group V is calculated
510 using the clinopyroxene/melt partition coefficients of Green et al. (2000) and Hauri et al.
511 (1994). The resulting composition is similar to Patagonian arc basalts or adakitic-like melt
512 (Fig. 8-B).

513 We modeled melt-rock reactions in the PDI mantle column using the approach proposed by
514 Brunelli et al. (2014) based on the equation system of Ozawa and Shimizu (1995) and Ozawa
515 (2001). In particular we will address the general variations in modal composition and REE
516 patterns of Group II, III and IV. Used parameters are F = degree of melting, β = influx melt
517 rate, ϕ_p = critical mass porosity and ϕ = actual porosity. ϕ/F is the porosity/melting ratio and
518 is used as proxy for the type of melting process: a pure batch melting has $\phi/F = 1$ (where all
519 the produced melt is retained in the system) while pure fractional melting has $\phi/F = 0$ (where
520 the system porosity does not permit the retention of the melt). Mineral melting coefficients
521 (X) determine their dissolution (positive) or crystallization (negative). Details and equations
522 are in Brunelli et al. (2014).

523 The mantle original modal composition was estimated as: olivine 58.3%, orthopyroxene
524 35.0%, clinopyroxene 4.7% and spinel 2.0%. Modes for dry melting are from Brunelli et al.
525 (2006), for hydrous melting are from Grove and Till (2019) (garnet-facies) and from Bizimis
526 et al. (2000) (spinel-facies). The processes were modeled using melt compositions of
527 Patagonian adakites (Stern and Kilian, 1996) similar to melts in equilibrium with Group V
528 clinopyroxene. Modelling the process previously described requires the creation of a large
529 clinopyroxene excess by reaction and contextual melt extraction. Melt extraction is required
530 by the progressive depletion of HREE and Al and the increase of Mg. Orthopyroxene
531 consumption appears to have a random correlation with clinopyroxene modal content. This
532 suggests a significant variation of the reaction coefficients of the process at different
533 moments.

534 The stochastic distribution of the different pattern shapes with respect to the modal
535 distribution suggests that these rocks experienced a short scale melt/rock reaction,
536 overprinting an extremely variable protolith and with extremely variable melt composition.
537 Therefore, a general model cannot be extrapolated. Only part of the process can be modeled
538 congruently. Here, we focus on some significant examples showing how different patterns
539 can possibly be generated setting some constraints on the composition of the melt and the
540 intensive process parameters (ϕ - F - β -reaction mode). A general character that can be
541 assumed is the depleted nature of the protolith in term of incompatible elements. We hence

542 assume a depleted starting clinopyroxene in the calculations (similar to Group I) and low
543 modal clinopyroxene in the starting assemblage, assuming that the majority of this mineral
544 is created by reaction as already discussed. For this reason, melting coefficients are always
545 high for orthopyroxene and low or negative for clinopyroxene. Chromatographic effects
546 must also be considered as shown by V-shaped REE patterns in Group III (Fig. 5). We thus
547 create models in spinel-facies conditions considering the chromatographic effects and pure
548 incremental open system (Fig. 9). The latter represents the end member case in which the
549 inflowing melt always has the same composition, thus a case representative of a process in
550 which melt is supplied by an external reservoir (as for instance a melt channel) that adds
551 fresh melt of constant composition at each step. In this configuration chromatographic
552 effects are therefore not visible.

553 The main melting trend, i.e. depletion in pyroxenes and crystallization of olivine, requires
554 positive coefficients for orthopyroxene and clinopyroxene. Positive coefficients mean
555 consumption of both phases during the reaction. This model follows the upper limit of the
556 maximum pyroxenes trend with respect to olivine thus considering high orthopyroxene
557 content in the source (Fig. 9A). The most favorable case is given by $X_{Opx} = 0.7$ and $X_{Cpx} =$
558 0.1 . For this case, ϕ/F and β can only have intermediate values, both set at 0.40 and using
559 the adakite as influxing melt. Modal variations are consistent with the main pyroxenes
560 depletion trends (Fig. 9A). Resulting REE patterns for chromatographic effects show
561 extreme enrichments in LREE (Fig. 9A). The chromatographic model matches the REE
562 patterns of the most enriched samples of Group II and III (Fig. 9A).

563 To simulate the modal trend enriching the system in clinopyroxene at the expense of
564 orthopyroxene, which crosscut the main trend at high orthopyroxene content, we need to set
565 the X_{Cpx} as negative. Patterns become significant at $\beta > 0.2$ and a low $\phi/F = 0.3$ (thus
566 increasing the fractional melting): the resulting pattern considering the chromatographic
567 effect shows a sinusoidal shape (Fig. 9B). Under this configuration, increasing the ϕ/F value
568 (thus increasing the batch process) results in relatively more enriched patterns with a shape
569 changing progressively to concave while the modal variation remains constant (Fig. 9C).
570 These patterns could match the Group IV for high fractional melting ($\phi/F = 0.2$; Fig. 9B) and
571 Group III for a higher batch process ($\phi/F = 0.6$; Fig. 9C). Similar sinusoidal patterns as Group
572 IV have been recognized in two xenoliths from Prahuaniyeu: Pra1 of Mundl et al. (2015)
573 and Pra68 from Group 3 of Bjerg et al. (2009). Concave LREE-enriched patterns were
574 instead recognized in clinopyroxenes from Prahuaniyeu, Pra304 from Group 1 of Bjerg et
575 al. (2009), and Pr301 and 303 from north of the Somuncurá plateau (Mundl et al., 2015).

576 If we consider the pure incremental open system with the same conditions and high batch
577 process (i.e. $\phi/F=0.6$), the resulting pattern shows almost flat HREE (Fig. 9C). This pattern
578 is similar to most of Group II samples. Similar patterns in Patagonia were recorded, among
579 others, in clinopyroxenes from Puesto Díaz (north of the Somuncurá plateau; Mundl et al.,
580 2015) and in clinopyroxenes from Group 2 of Bjerg et al. (2009) from Prahuaniyeu
581 (Somuncurá plateau). Rivalenti et al. (2007a) proposed a numerical simulation of a
582 chromatographic enrichment for similar patterns in Cerro de los Chenques xenoliths (160
583 km SW of PDI).

584 The pattern relationships observed before remain valid at low orthopyroxene content:
585 changes in the melting coefficients produce the same changes in the pattern morphology
586 (Fig. 9D-G). This is true for the vertical pyroxene trend (Fig. 9D), the pyroxene dispersion
587 toward (extremely) low orthopyroxene contents with some residual clinopyroxene (Fig. 9E)
588 or the extreme case that considers depletion of both pyroxenes (Fig. 9F-G).

589

590 *9.3 Regional events*

591 A regional model for the metasomatic/refertilization events of northern Patagonia reported
592 the occurrence of two distinct melt percolations in the mantle column (Melchiorre et al.
593 2020). According this model, the first metasomatic event occurred after partial melting of
594 the mantle column by percolation of a tholeiitic depleted basalt, which depleted the
595 clinopyroxene in LREE. The second event was induced by percolation of alkaline to
596 transitional SiO₂-undersaturated melts, which produced limited refertilization with LREE
597 enrichment in clinopyroxene. These two events have been related to the Patagonian plateau
598 magmatism where large volumes of tholeiitic lavas are followed by limited pulses of alkaline
599 melts (Gorring et al., 2003; D’Orazio et al., 2004; Kay et al., 2007, 2013; Melchiorre et al.,
600 2020). These events, however, are not identical in all the northern Patagonian regions but
601 change locally (Melchiorre et al., 2020; and references therein). This is the case of the PDI
602 area.

603 As previously discussed, the last melting event in the studied area and their associated
604 melt/rock reactions overprinted the previous mantle composition. In particular, several
605 different REE patterns and modal compositions could be obtained from the same melting
606 event when melt influx rate and porosity change. This is taken to the extreme if the mantle
607 column was, as certainly was, already heterogeneous by previous processes. Modal and
608 chemical variations of mantle xenoliths of PDI could be obtained from an open-system
609 melting process due to the influx of an adakitic melt. In addition, the LREE-depleted samples

610 (Group I), previously interpreted as related to a tholeiitic metasomatism (Melchiorre et al.,
611 2020), could be explained by fractional melting in spinel conditions. These evidences
612 indicate that the PDI mantle column has been completely overprinted by the alkaline/adakitic
613 melting event (Fig. 10). The tholeiitic event, if occurred, is now completely overprinted.
614 Similar geochemical features from several other Patagonian xenoliths (e.g. Puesto Díaz,
615 Prahuaniyeu, Cerro de los Chenques, Cerro Chenque; Rivalenti et al., 2007a; Bjerg et al.,
616 2009; Mundl et al., 2015) suggest that open-system melting could be responsible for mantle
617 heterogeneity in the Patagonian region.

618

619 **10. Conclusions**

620 The abundance of harzburgites and lherzolites with less than 6% clinopyroxene by volume
621 (~70% of all samples) confirms that the mantle column below the Paso de Indios area is
622 significantly depleted with respect to the modal composition of the primitive mantle.
623 Petrological evidence of garnet breakup suggests that melting started in garnet-facies and
624 continued in spinel facies driven by the influx of an adakitic melt. This melt is related to the
625 subduction process active along the western margin of South America since Late Triassic
626 times.

627 The melting process was open-system, inducing melt/rock reactions that produced modal
628 and compositional heterogeneities on a small scale. These processes completely overprinted
629 the original mantle composition and previous processes. Heterogeneities were produced by
630 different degrees of melt influx and rock porosity, which are the main factors controlling the
631 nature of the melting process from complete fractional to mainly batch (up to 60%).

632 Among the several Patagonian localities with outcropping subcontinental mantle xenoliths,
633 the PDI xenoliths are unique. Here, the mantle can be studied in different stages, from the
634 initial pre-subduction depletion to the late superimposed melt/rock reaction events during
635 open-system melting.

636

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644

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958 **Figure Captions**

959
 960 Fig. 1. Geologic map of the study area and location of Chenque, León and Matilde hills,
 961 modified after Ponce et al. (2015). The three localities are framed in red.
 962

963 Fig. 2. A: Modal composition of the studied PDI xenoliths. Star (PM) primitive mantle after
 964 Warren (2016). The grey field represents the variation of anhydrous Patagonian xenoliths
 965 from Rivalenti et al. (2004a) and Bjerg et al. (2005). B: pyroxenes vs olivine contents of PDI
 966 lherzolite xenoliths. B: pyroxenes vs olivine contents of PDI harzburgite xenoliths. Filled
 967 symbols are new samples from this work, white-filled symbols are samples from Ponce et
 968 al. (2015). PM values are from Warren (2016).
 969

970 Fig. 3. Photomicrographs of PDI mantle xenoliths under cross-polarized light. A)
 971 Porphyroclastic texture of sample Q77. B) Transitional coarse to porphyroclastic texture of
 972 harzburgite MH. C) Transitional coarse to porphyroclastic texture of harzburgite L82a and
 973 D) mosaic-porphyroclastic texture of lherzolite Q65. E) Olivine relict within orthopyroxene
 974 in harzburgite M55. F) Interstitial secondary orthopyroxene growing at olivine borders in
 975 harzburgite Q101. G) Contact of the xenolith with the host basalt and development of a
 976 reaction corona (harzburgite sample M67). H) secondary orthopyroxene engulfing a primary
 977 olivine (lherzolite M53). I) SEM image of basalt-xenolith contact and reaction zone in
 978 lherzolite L91. L) spongy clinopyroxene and associated glass veins in harzburgite L82a. Ol:
 979 olivine; opx: orthopyroxene; cpx: clinopyroxene; spl: spinel.
 980

981 Fig. 4. (A) NiO vs Mg# in olivine from the PDI area; (B) Cr# vs Mg# of spinels and fields
 982 of the spinel classification of Schilze (2001); Al₂O₃ vs. Mg# for orthopyroxenes (C) and
 983 clinopyroxenes (D). Grey fields are the compositional variations determined by Melchiorre
 984 et al. (2020) and references therein for peridotitic and pyroxenitic xenoliths of North
 985 Patagonia.
 986

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987 Fig. 5. C1 chondrite (Anders and Grevesse, 1988) normalized REE elements patterns and
988 primitive mantle (PM, Sun and McDonough 1989) normalized spiderdiagrams of
989 clinopyroxenes. A) group I: LREE depleted patterns; B) group II: LREE enriched patterns;
990 C) group III: “U” and “V”-shaped patterns; D) group IV: sinusoidal patterns and E) group
991 V: flat patterns to weakly LREE enriched;. The pattern of a primitive mantle-cpx (solid black
992 line) of Sun and McDonough (1989) and patterns (gray lines) after 1, 3, 5, 10, 15 and 20%
993 of depletion by partial non-modal melting, are also shown. For comparison, data from PDI
994 xenoliths from literature (Rivalenti et al., 2004a) are reported as dashed lines. Trace elements
995 extended diagrams are reported with values normalized to the Primitive Mantle composition
996 recalculated from Pyrolite (PM; McDonough and Sun, 1995).

997
998 Fig. 6. Pressure (GPa) vs. Temperature (°C). Temperatures were calculated with the Ca-
999 inOpx geothermometer of Brey and Köhler (1990). L: lherzolite; H: harzburgite; W:
1000 websterite and olivine-websterite. Data from other Patagonian localities are reported for
1001 comparison: Prahuaniyeu is from Bjerg et al. (2009); Tres Lagos is from Mazzucchelli et al.
1002 (2016); Agua Poca is from Bertotto et al. (2013); Río Negro, Gobernador Gregores and Pali
1003 Aike are from Bjerg et al. (2005).

1004
1005 Fig. 7. A: modal clinopyroxene content vs spinel Cr# diagram. B: modal clinopyroxene
1006 content vs Yb in clinopyroxene normalized to the C1 chondrite (Anders and Grevesse, 1988).
1007 C: Mg# vs Na₂O (wt.%) in clinopyroxene. D: Al₂O₃ vs Cr₂O₃ (wt.%) in clinopyroxene. E:
1008 Cr₂O₃ vs Na₂O (wt.%) in clinopyroxene. F: Forsterite (Fo) vs NiO (wt%) in olivine. Data
1009 from abyssal peridotite (Warren et al., 2016) and VEMA mantle rocks (Brunelli et al., 2006;
1010 Cipriani et al., 2009) are reported for comparison. Red arrows identify the fractional melting
1011 trend.

1012
1013 Fig. 8. Potential melts in equilibrium with PDI clinopyroxenes from A) Group I and B)
1014 Group V, normalized to the Primitive Mantle composition recalculated from Pyrolite (PM;
1015 McDonough and Sun, 1995). Melts are calculated using the clinopyroxene/melt partition
1016 coefficient of Ionov et al. (2002) for Group I and of Green et al. (2000) and Hauri et al.
1017 (1994) for Group V. Host basalts (Chenque, Matilde and León) from Alric (1996) are
1018 reported for comparison. The grey fields are: dark grey - arc basalts from the Patagonian
1019 region from Hickey et al. (1986) and D’Orazio et al. (2003); light grey - adakites from the
1020 Austral Volcanic Zone from Stern and Kilian (1996).

1021
1022 Fig. 9. Open-system model for the PDI xenoliths.

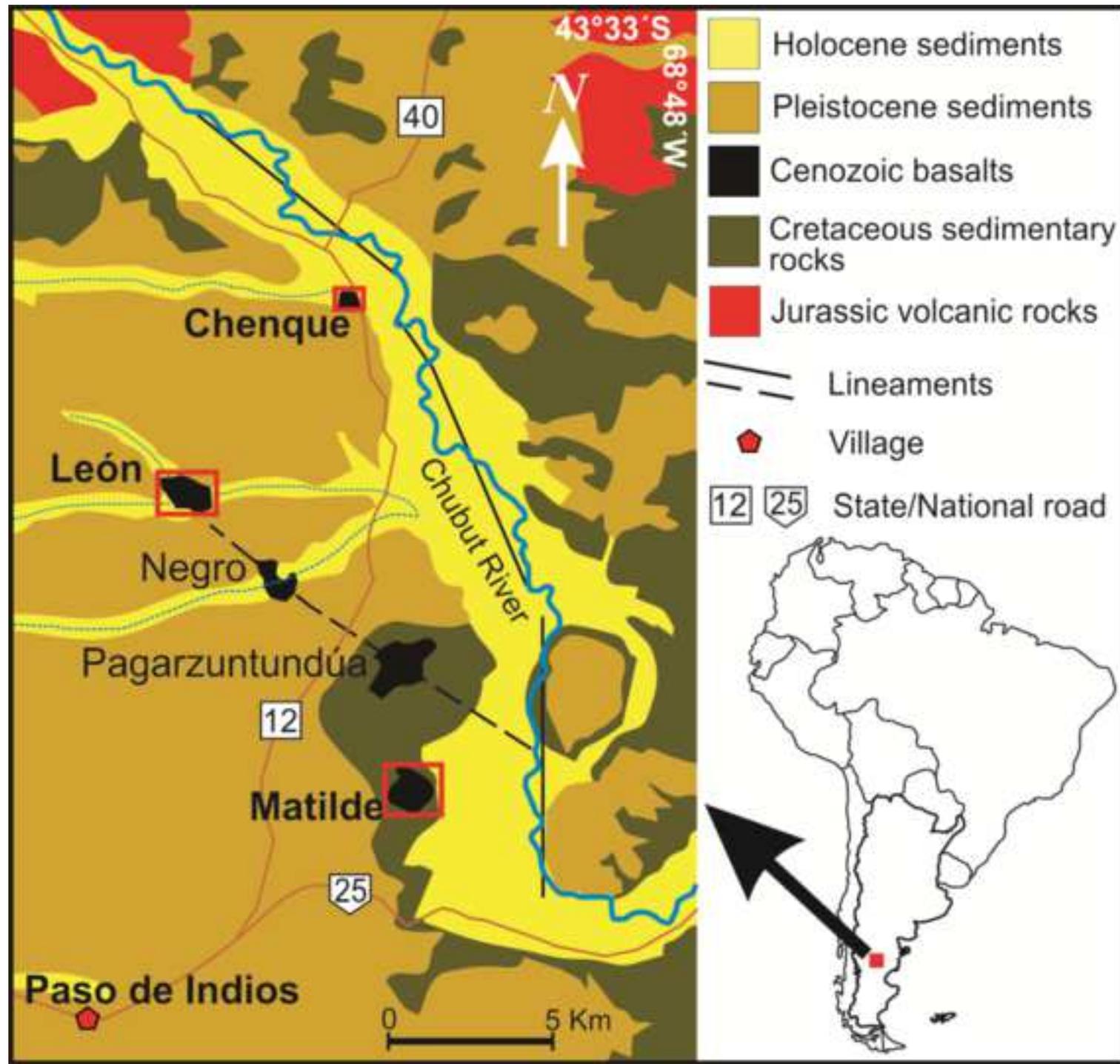
1023
1024 Fig. 10: Tectonic Model for the PDI mantle evolution during adakitic-like melt infiltration
1025 in a orthopyroxene-rich, clinopyroxene-poor mantle column. After the Cretaceous flat slab
1026 event different tectonic configurations were proposed to explain the alkaline features of the
1027 Paleocene-Eocene magmas, that are either the opening of a slab window (Iannelli et al.
1028 2017; Fernandez Paz et al.. 2018), or a pure a slab rollback of the Farallon plate (Muñoz et
1029 al.. 2000; Echaurren et al.. 2016, 2019), or a complete detachment of the Aluk plate and the
1030 formation of a transform margin (Aragón et al.. 2011; 2015). In the figure, for simplicity, we
1031 used for the asthenospheric mantle ascent two possibilities: with subduction and opening of
1032 a slab window (Fernández Paz et al.. 2018) or without subduction (Aragón et al. 2011, 2015).

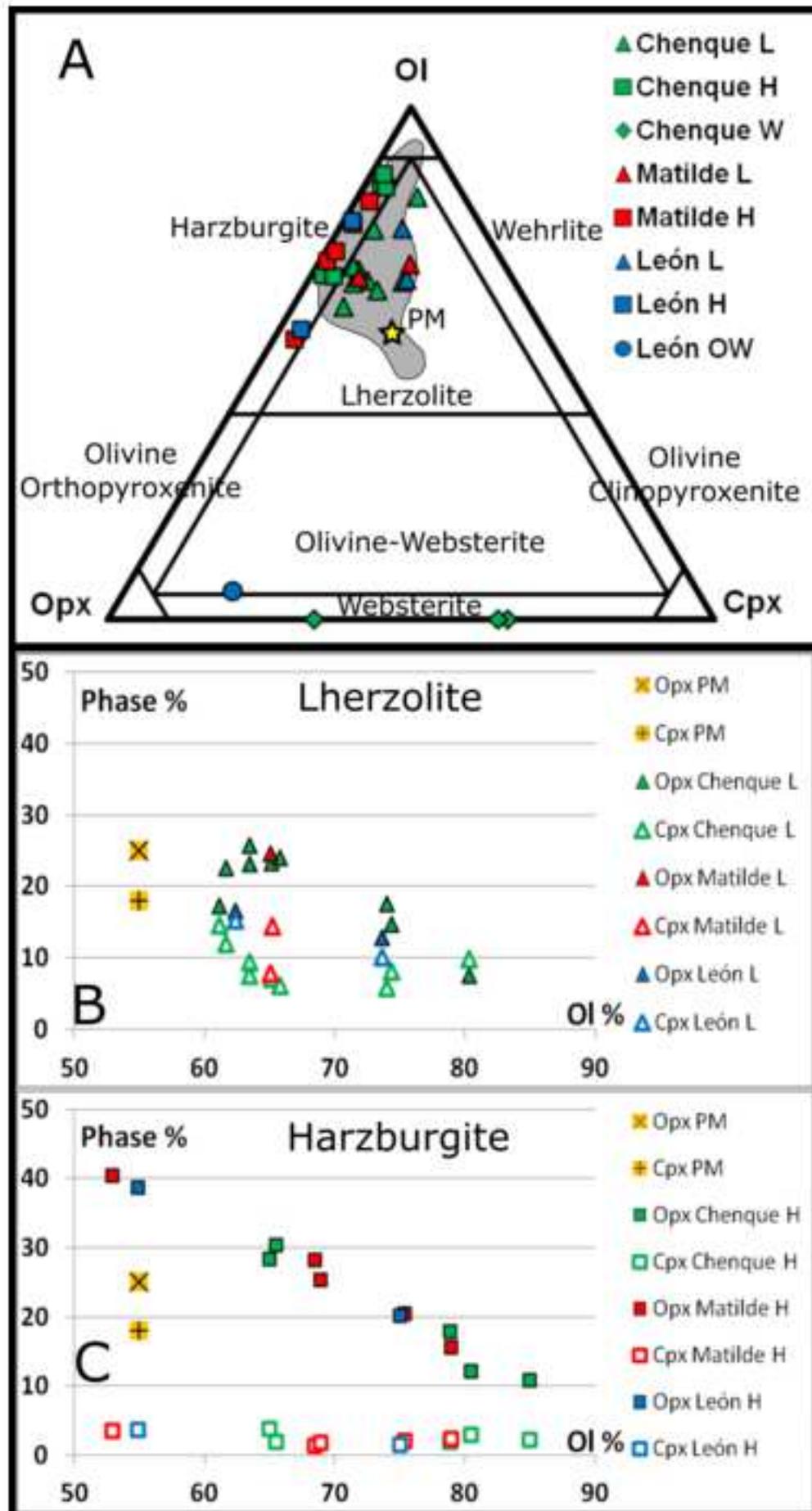
1033 **Table**

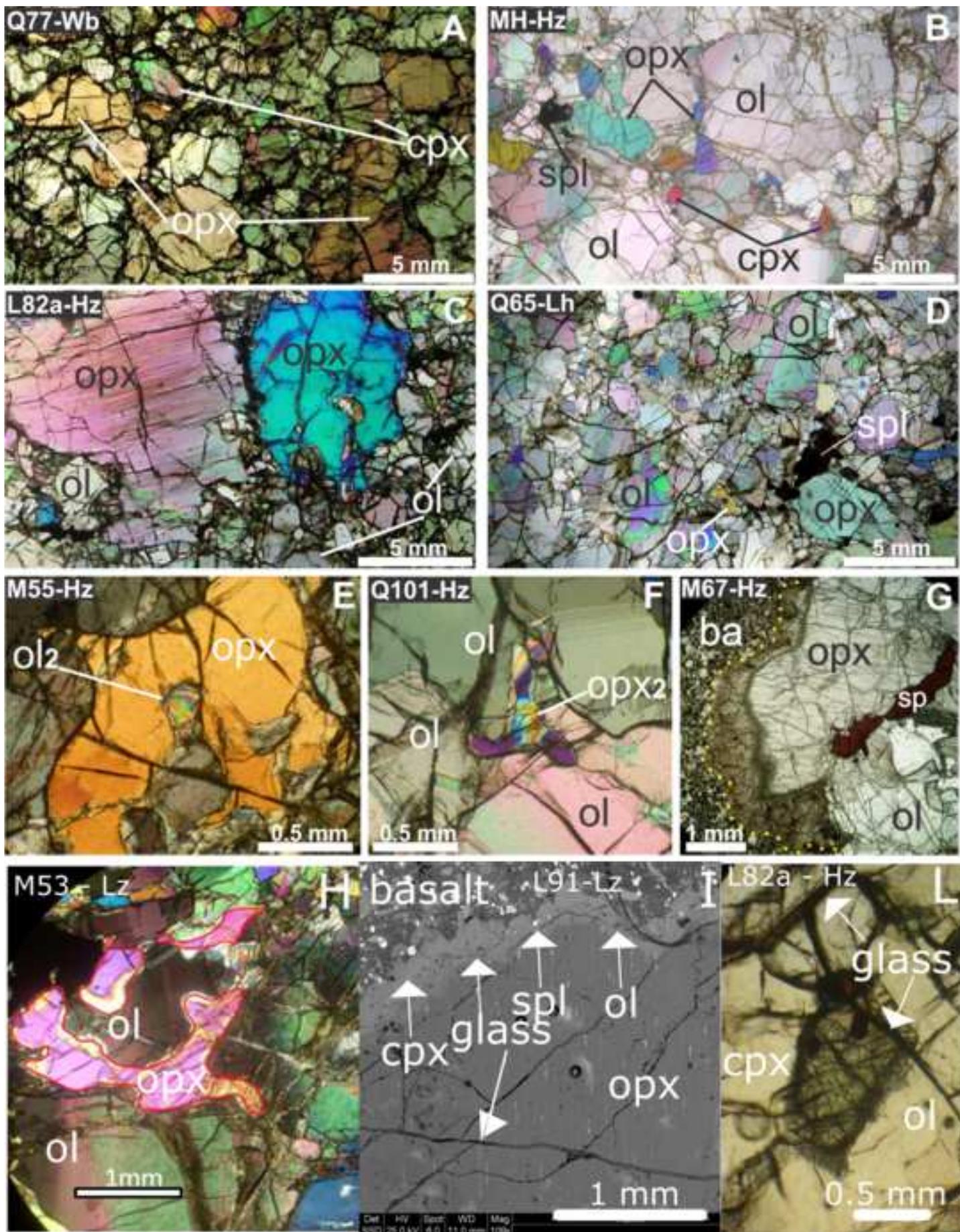
1034 Table 1. Modal composition of the PDI xenoliths. Q are from Chenque, M are from
1035 Matilde and L are from León. H: harzburgite; L: lherzolite; W: websterite and OW:
1036 olivine-websterite.

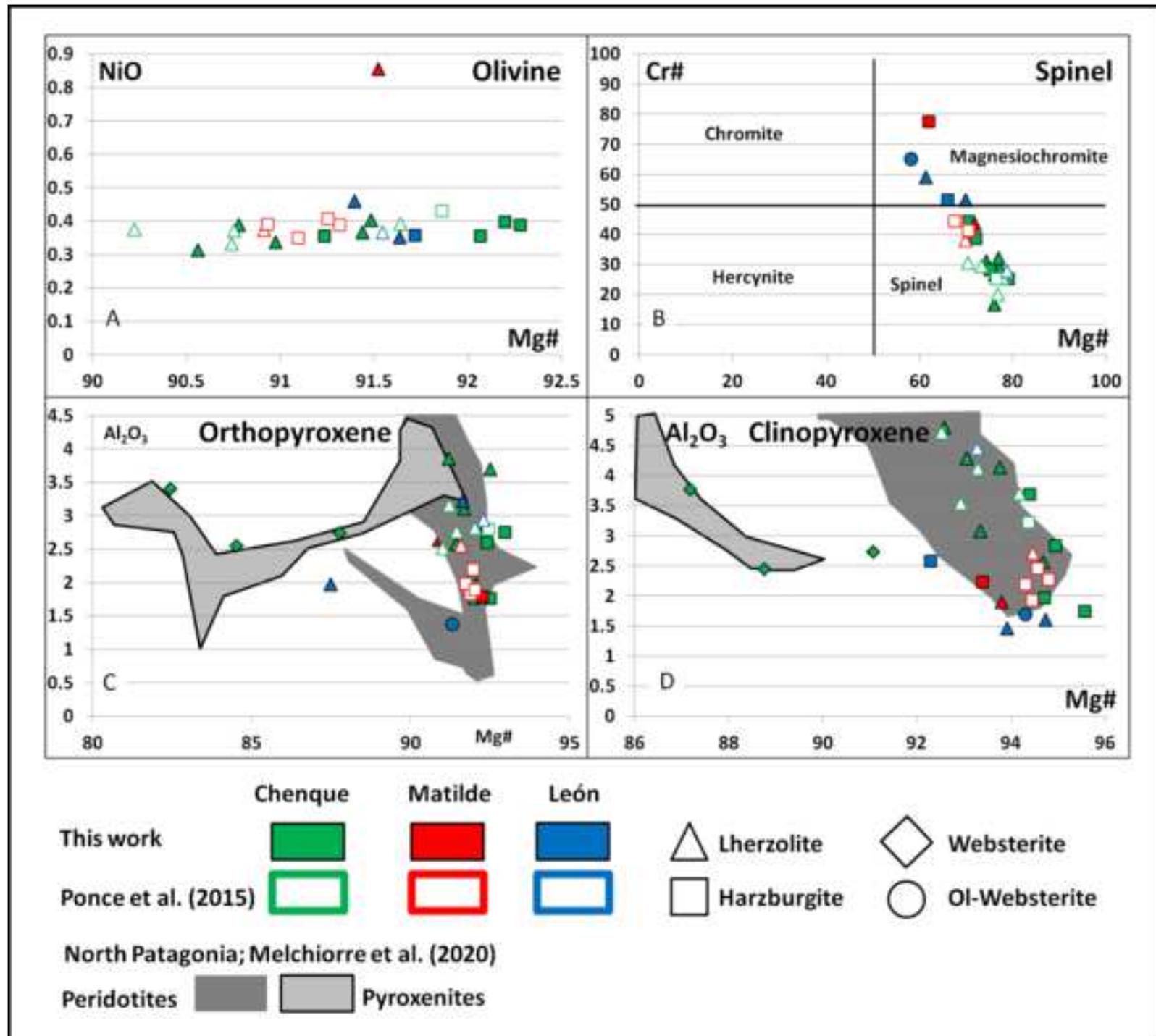
1037
1038 Table 2. Average olivine major element compositions. L: lherzolite; H: harzburgite; SD is
1039 standard deviation. b.d.l.: below detection limit. Values are wt.%.
1040
1041 Table 3. Average spinel major element compositions. L: lherzolite; H: harzburgite; OW:
1042 olivine websterite; SD is standard deviation. b.d.l.: below detection limit. Values are wt.%.
1043
1044 Table 4. Average orthopyroxene major element compositions. L: lherzolite; H: harzburgite;
1045 OW: olivine websterite; W: websterite; SD is standard deviation. b.d.l.: below detection
1046 limit. Values are wt.%.
1047
1048 Table 5. Average clinopyroxene major element compositions. L: lherzolite; H: harzburgite;
1049 OW: olivine websterite; W: websterite; SD is standard deviation. b.d.l.: below detection
1050 limit. Values are wt.%.
1051
1052
1053 Table 6. Average clinopyroxene trace element composition. L: lherzolite; H: harzburgite;
1054 OW: olivine websterite; W: websterite; SD is standard deviation. b.d.l.: below detection
1055 limit. Values are in ppm.
1056
1057 Table 7. Temperature (T-°C) and pressure (P-GPa) estimations from PDI xenoliths after data
1058 collected by Ponce et al. (2015) and this work. Type of lithologies is as follow: L: lherzolite;
1059 H: harzburgite; OW: olivine websterite; W: websterite. Geothermometer and geobarometer
1060 references: BKN = two-pyroxene thermometer (Brey and Köhler, 1990); Ca-Opx = Ca in
1061 orthopyroxene thermometer (Brey and Köhler, 1990); TA98 = two-pyroxene thermometer
1062 (Taylor, 1998); PMc = clinopyroxene barometer (Mercier, 1980). SD is standard deviation.
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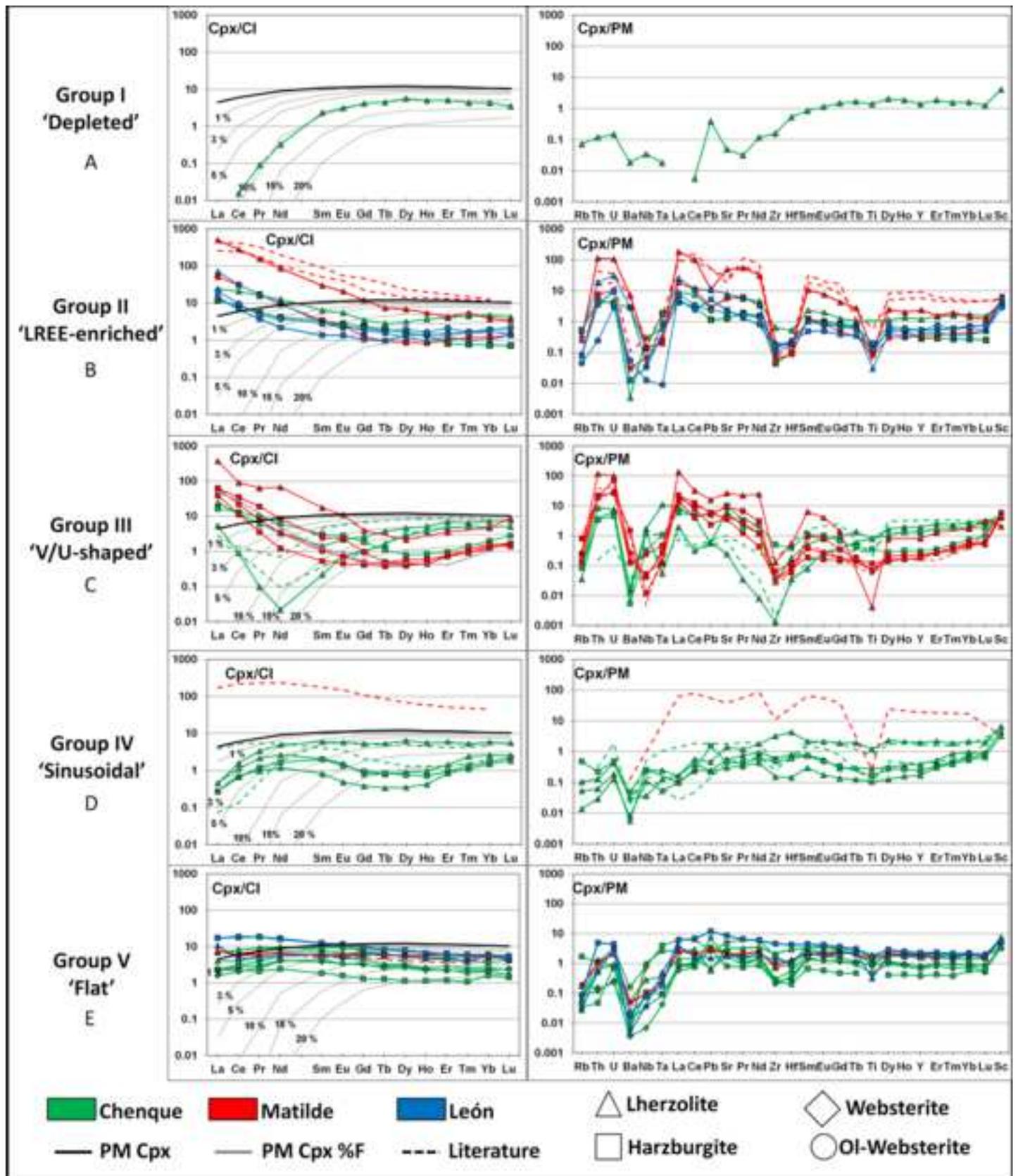
Figure 1

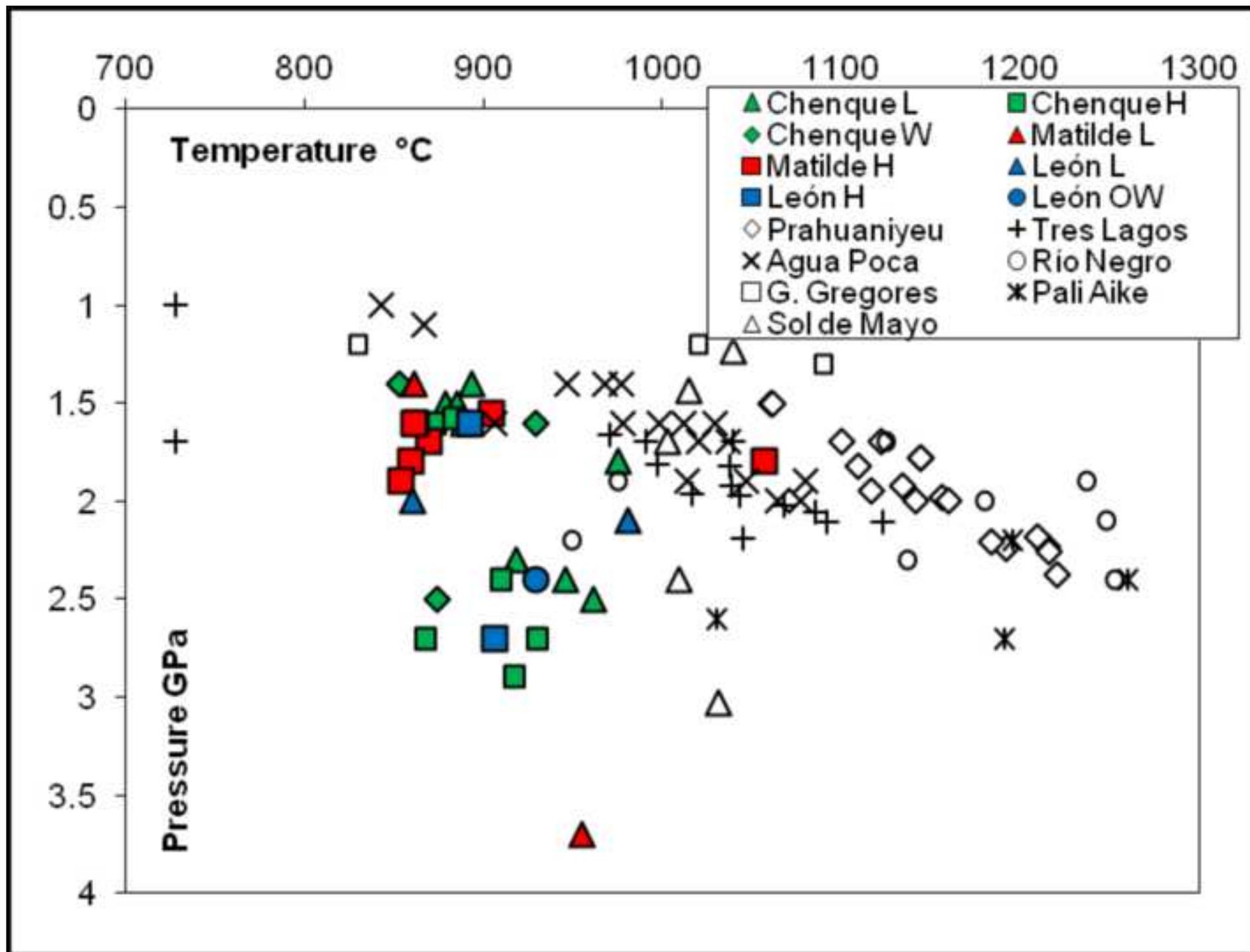


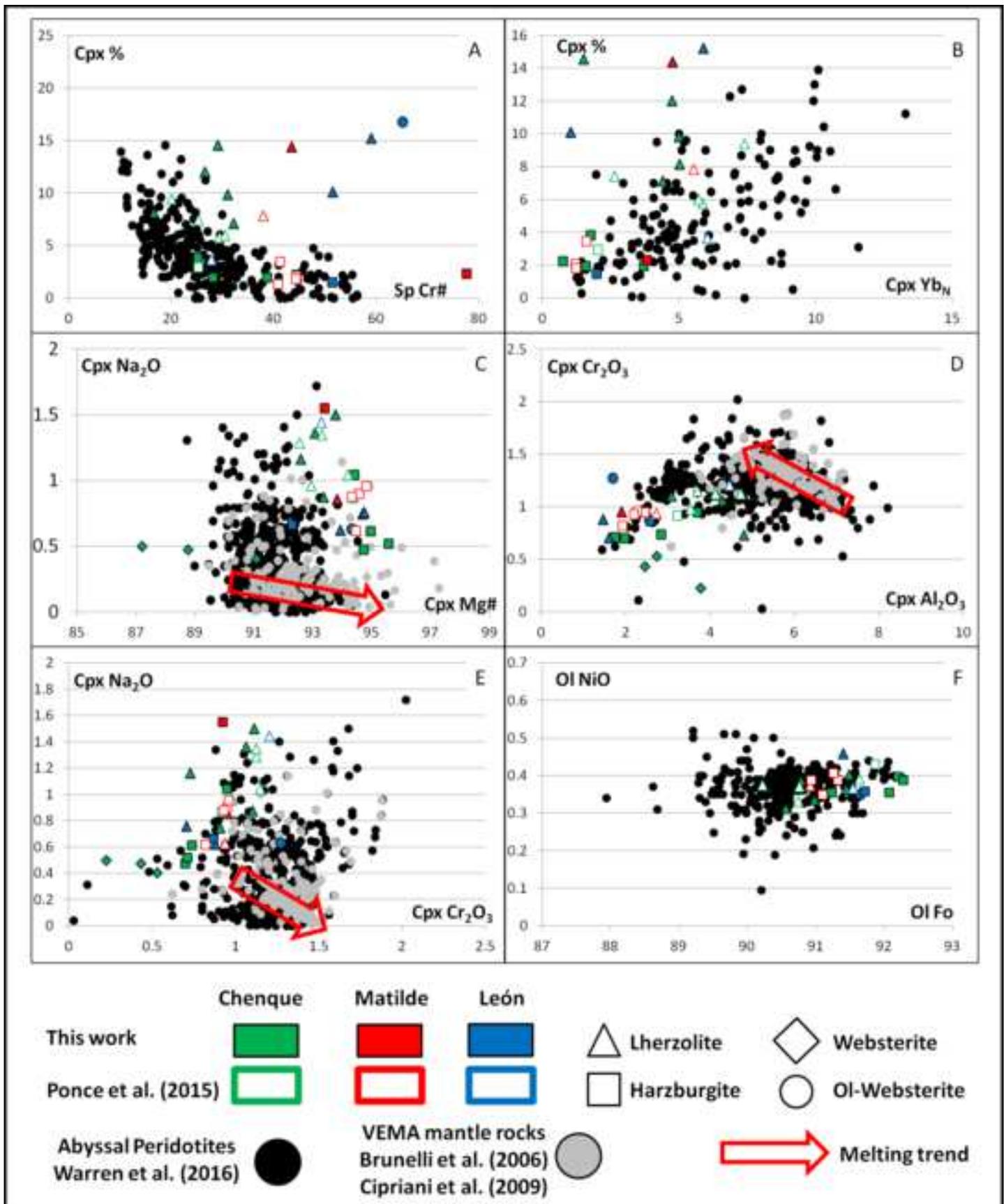


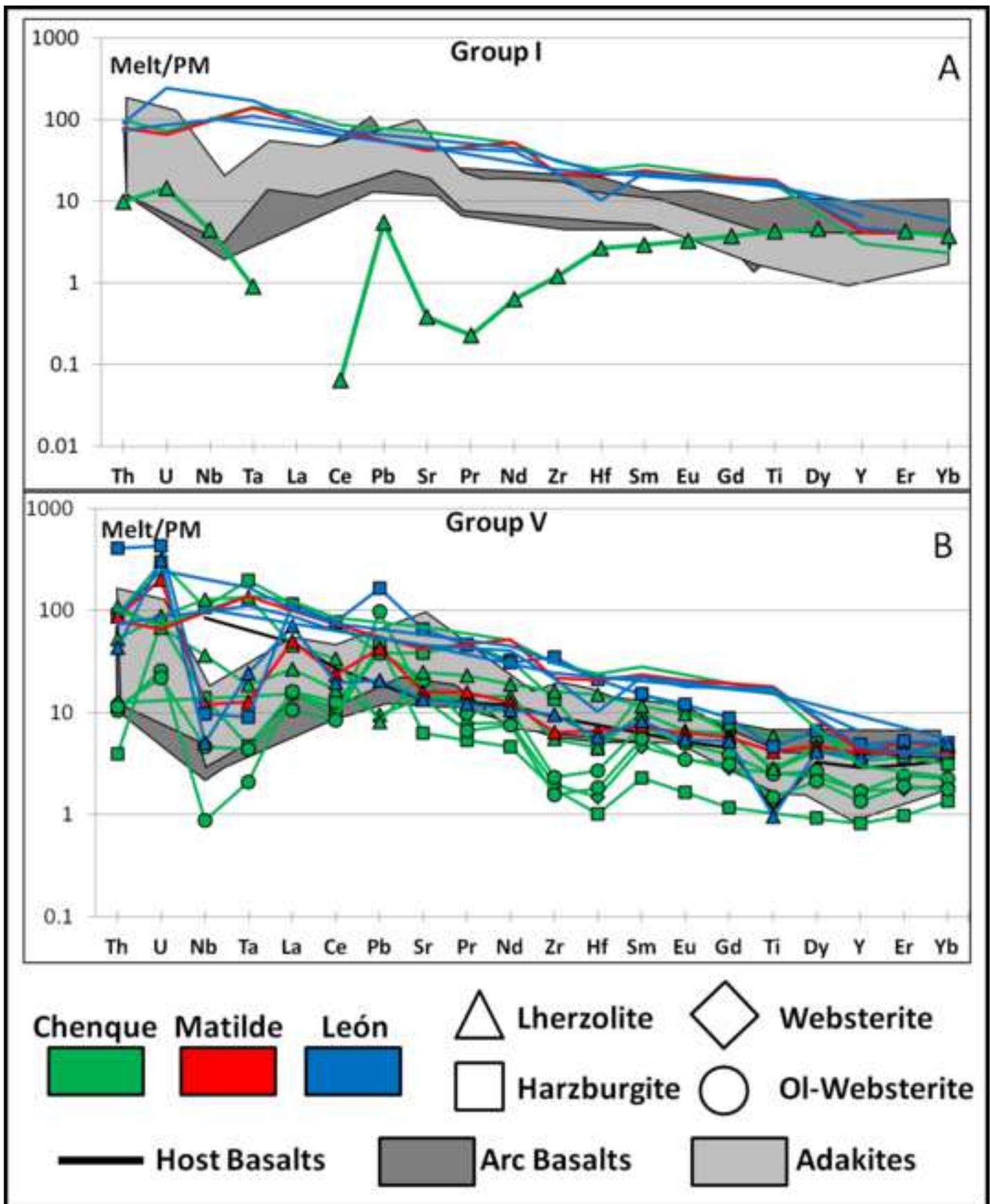


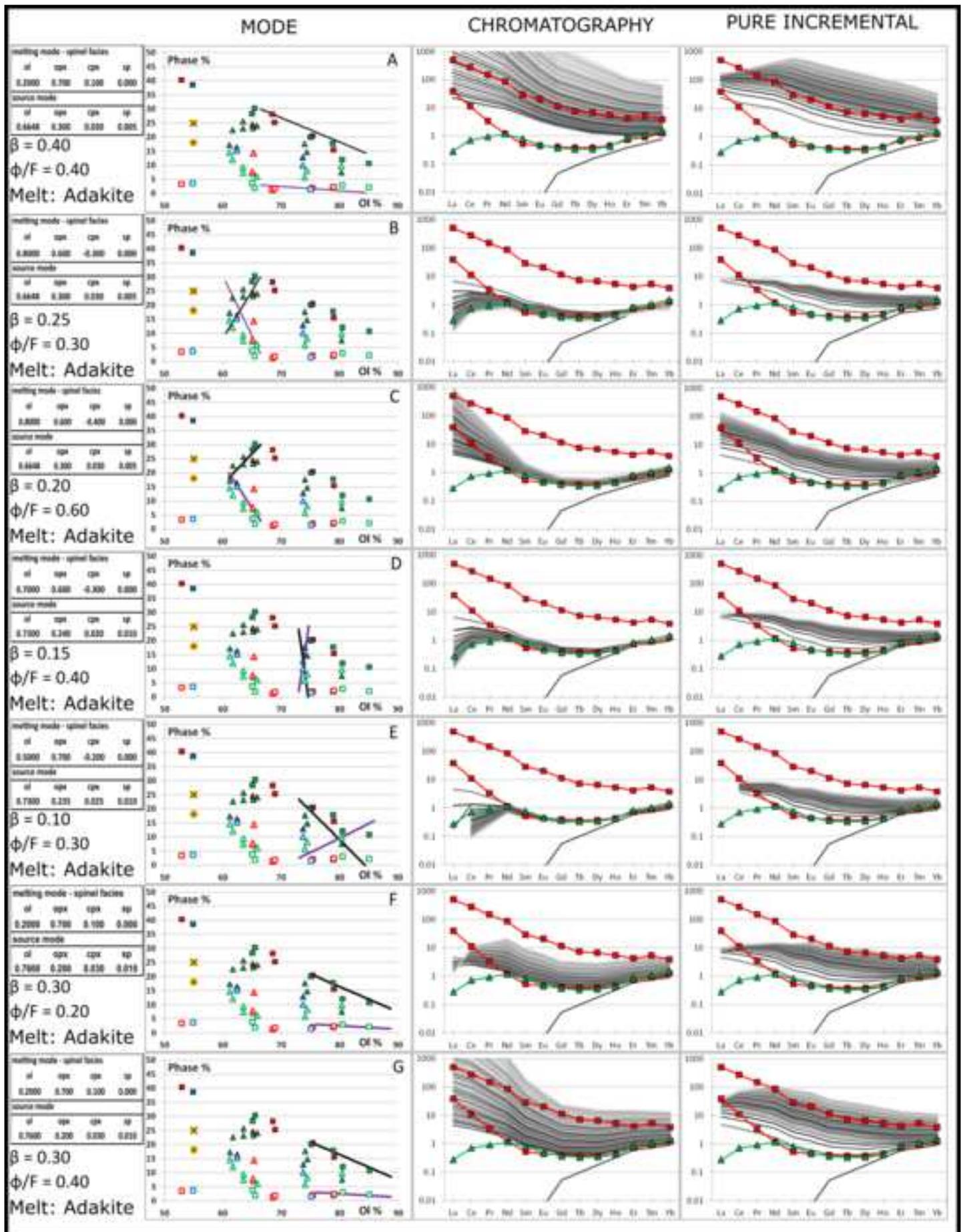


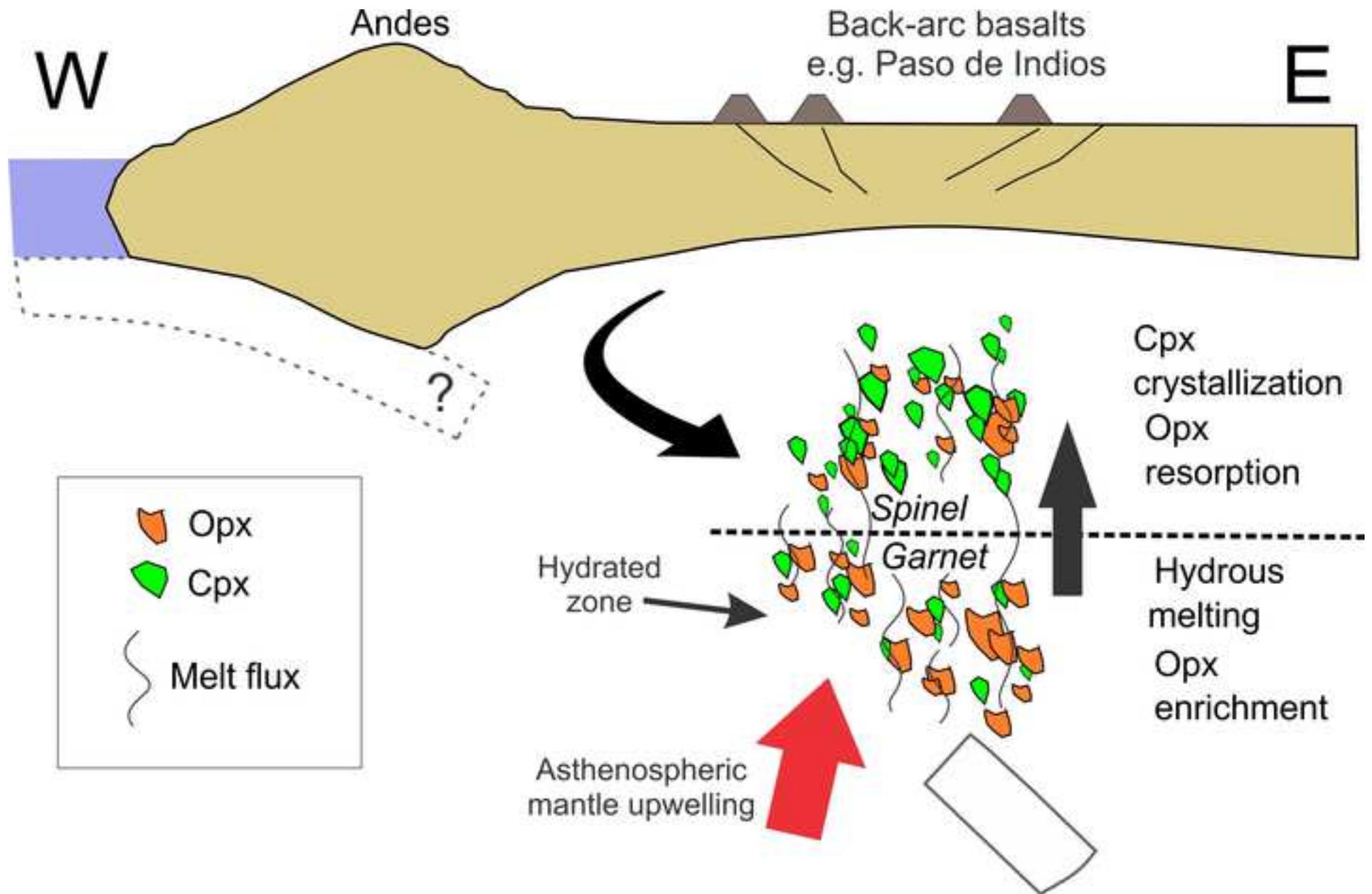














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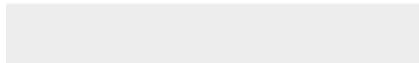
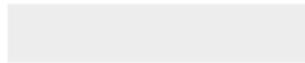




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