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► To cite this version:

Stéphanie Duchene, Jean-Luc Severac, Nadège Samalens, Youssef Driouch, Ahmed Ntarmouchant, et al.. Multi-stage metamorphism recorded in crustal xenoliths from Permian dykes of the region of Mrirt (Moroccan Central Massif). *Journal of African Earth Sciences*, 2022, 194, 10.1016/j.jafrearsci.2022.104636 . insu-03863527

HAL Id: insu-03863527

<https://insu.hal.science/insu-03863527v1>

Submitted on 22 Jul 2024

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1 **Multi-stage metamorphism recorded in crustal xenoliths from Permian dykes of the region**
2 **of Mrirt (Moroccan Central Massif)**

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27
28
29 **Abstract**
30

31 The Permian magmatic rocks from Morocco contain crustal xenoliths that sample the Variscan
32 crust in a context of widespread magmatism. A series of such xenoliths was collected in Permian
33 dykes of the Central Massif, in the region of Mrirt. The metapelitic xenoliths are silica poor to
34 intermediate ($44 < \text{SiO}_2 < 57$ wt.%) and alumina-rich ($17 < \text{Al}_2\text{O}_3 < 34$ wt.%) and are notably enriched
35 in some HFSE (Nb, Ta, Ti) and some transition elements (Cr, V, W, Ni). Their petrographic
36 evolution depicts a multi-stage evolution from an early, subsolidus, metamorphic history related
37 to regional metamorphism, of which biotite, garnet and sillimanite are the witnesses toward a late
38 thermal evolution coeval with the entrapment in the magma, marked by pervasive partial melting
39 and development of peritectic spinel and cordierite together with K-feldspar and ilmenite. The
40 overall presence of corundum, which relates to the high Al_2O_3 content, accounts for an initial
41 stage of partial melting and magma escape, prior to the entrapment in the magma. Textural and
42 chemical observation suggests further xenolith digestion and melt flux from the xenoliths toward
43 the magma during ascent. Thermodynamic modelling allowing the determination of the pressure-
44 temperature history of each xenolith shows that the initial pressure varies from *ca.* 1.5 to *ca.* 6
45 kbar, which, considering lithostatic pressure, corresponds to sampling depths of *ca.* 5 to 25 km.
46 The temperatures recorded by the parageneses coeval with partial melting in the presence of
47 spinel are in the range 800-900 °C whatever the pressure, which accounts for rapid heating due to
48 thermal equilibration with the magma. U-Th/Pb dating of monazite by EMPA gives a poorly
49 resolved Permian age of 293 ± 25 Ma for metamorphism in the xenoliths, in accordance with the
50 stratigraphic age of the host rock. The pressure conditions are similar to those recorded for
51 regional metamorphism in the other Variscan outcrops in Morocco, namely the Jebilet, the
52 Rehamna and the Aouli-Mibladen granitic complex, while the maximal temperatures are much
53 higher. The xenoliths thus appear as typical of the Late Variscan geological evolution of the deep

54 crust in the Moroccan Mesetas, where abundant magmatism was responsible for local crustal
55 heating of a crust previously affected by regional metamorphism.

56

57

58 **Highlights**

59

- 60 • Metapelitic xenoliths sample Moroccan Variscan crust from 5 to 25 km depth
- 61 • Xenoliths record a first stage of regional metamorphism
- 62 • High-temperature metamorphism at low-pressure is due to xenoliths entrapment in the
63 magma
- 64 • EMPA monazite dating confirms the Permian age of the microgranodioritic dykes

65

66

67

68 **Keywords:** crustal xenoliths, Moroccan Central Massif, metamorphism, thermal history,
69 thermodynamic modelling, magma-xenolith interactions

70

71

72 **1. Introduction**

73 The presence of granulite facies rocks and migmatites in metamorphic rocks formed at
74 middle to lower crustal level in mountain belts accounts for high geothermal gradients ($>30^{\circ}/\text{km}$).
75 Such gradients cannot be sustained by normal radioactivity in a crust of standard thickness. Their
76 occurrence claims either for an increase of crustal thickness (England and Thompson, 1984),
77 which induces an increase in radioactive heat production, or for additional heat sources
78 associated to orogenic processes. Specifically, the injection in the crust of magmas of mantle
79 origin may explain significant increase of the temperature (*e.g.* Juez-Larré and Ter Voorde, 2009)
80 whereas magma transfer ensures near-isothermal conditions in middle-low crustal level and high
81 gradients at upper crustal levels (Depine *et al.*, 2008). Magmatic and metamorphic rocks that
82 outcrop together in high temperature orogenic crust are invaluable witnesses of crustal heating
83 but prograde metamorphism is often obliterated by later retrogression associated to a long
84 exhumation history in a thermally unstable orogenic crust. By contrast, crustal xenoliths
85 transported to the surface in volcanic rocks have been shown to record the pressure-temperature
86 conditions prevailing in the crust in the very moment of sampling by the magmas, thanks to their
87 rapid ascent toward the surface. A remarkable example is that of the metamorphic xenoliths
88 trapped in Neogene lavas of SE Spain (*e.g.* Cesare, 2000; Acosta-Vigil *et al.*, 2010; Álvarez-
89 Valero and Kriegsman, 2007), that record, from west to east, both crustal thinning (decreasing
90 pressure) and crustal heating (increasing temperature) and image crustal extension and increasing
91 heat flux related to the opening of the Alborán Sea. Nevertheless, xenoliths also undergo strong
92 albeit short heating during their ascent (Álvarez-Valero *et al.*, 2015), possible partial melting and
93 digestion, that may hamper the record of the *PT* conditions prevailing during magma injection in
94 the crust (*e.g.* Díaz-Alvarado *et al.* 2011; Johnson *et al.*, 2011).

95

96 The Variscan crust underwent high-temperature, granulite facies metamorphism associated to
97 magmatism in the Ivrea zone in the Alps (Kunz *et al.*, 2018; Petri *et al.*, 2017), Calabria
98 (Graessner and Schenk, 2001; Fornelli *et al.*, 2011; Duchene *et al.*, 2013), Corsica (Corsini and
99 Rolland, 2009; Rossi *et al.*, 2015) and Pyrenees (Guitard *et al.*, 1995; Siron *et al.*, 2020). In
100 Morocco, there are only limited outcrops of the Variscan median crust, that mainly shows
101 *MP/MT* greenschist to amphibolite facies metamorphism. However, high temperature granulitic

102 conditions are retrieved from abundant crustal xenoliths sampled both by Permian and Neogene
103 magmatic rocks (Bouloton *et al.*, 1991; Bouloton and Gasquet, 1995; Moukadiri and Bouloton,
104 1998; Ntarmouchant, 2003; Dostal *et al.*, 2005; Elabouyi *et al.*, 2019) that appear as a window
105 into the underlying crust. Here we propose to study a set of crustal metapelitic xenoliths trapped
106 in Permian dykes from the Moroccan Central Massif, in the late stages of the Variscan orogeny,
107 in order to decipher the relative contributions to the metamorphic record of i) general
108 metamorphism and ii) heating during magma emplacement. More specifically, through
109 petrographic observation and thermodynamic modelling, we intend to retrieve the petrogenetic
110 evolution of xenoliths, their pressure – temperature evolution from regional metamorphism
111 toward entrapment in the magma during its ascent and explore the intensity of thermal and
112 chemical crustal – melt interaction.

113

114

115 **2. Geological context**

116 In Morocco, witnesses of the Variscan belt are exposed in the Mauritanides, in the Anti-
117 Atlas and in the Meseta domain (Fig. 1a) as summarized by Simancas *et al.* (2005, 2009),
118 Hoepffner *et al.* (2005) and Michard *et al.* (2008, 2010). It is believed that these domains
119 represent the former Gondwana passive margin and are the southern termination of the Variscan
120 belt (Simancas *et al.*, 2005, 2009; Michard *et al.*, 2010; Martínez Catalán *et al.*, 2021). This
121 passive margin was involved into the orogeny when closure of the Rheic ocean eventually led to
122 the collision between Armorica and Gondwana. No remnant of an oceanic suture exists in the
123 Moroccan Variscan belt, so that several models for the orogeny have been proposed, from that of
124 an intracontinental orogeny (Piqué and Michard, 1989) to a subduction –related orogeny due to
125 either south-eastward subduction of the Rheic ocean under the northern margin of Gondwana
126 (Michard *et al.*, 2010) or north-westward subduction of an eastern Paleotethyan oceanic domain
127 (Roddaz *et al.*, 2002; Stampfli and Borel, 2002) or a succession of both (Martinez Catalán *et al.*,
128 2021).

129

130 The Central Massif is one of the Paleozoic massifs that constitute the Western Meseta domain,
131 west of the Middle Atlas high. It is composed of a sequence from Cambrian to Permian
132 sediments. The study area is located close to the locality of Mrirt (Fig. 1b), close to the Tighza

133 Massif and is described in detail in Ntarmouchant (2003). The oldest Paleozoic, undifferentiated
134 Cambrian-Ordovician sediments are detrital sandstones and pelites. Silurian sediments are black
135 to purple shales which represent the decollement level for a series of Devonian-Carboniferous
136 duplex system (Ben Abbou *et al.*, 2001). The overlying Devonian sediments are intercalations of
137 carbonates and pelites. The detrital Visean-Namurian stratas, progressively turning into
138 continental deposits, are discordant on the Ordovician to Devonian sediments and have been
139 interpreted as syn-tectonic sediments (Ben Abbou *et al.*, 2001; Roddaz *et al.*, 2002; Accotto *et*
140 *al.*, 2020) deposited in a foreland basin. Finally, the Hercynian structures are sealed by Upper
141 Stephanian-Autunian continental redbeds that unconformably overlay the Visean-Namurian
142 sediments and are affected by late Hercynian faulting.

143
144 Numerous gabbroic and doleritic bodies intrude the Carboniferous sedimentary layers and
145 form vein swarms parallel to the Hercynian structures. Since they are folded together with the
146 Carboniferous sediments, they are considered as late-Visean to Serpukovian in age
147 (Ntarmouchant, 2003). Later magmatism takes the form of microgranitic dykes or granitoid
148 plutons that crosscut the Carboniferous stratas and structures, but are mainly oriented parallel to
149 the schistosity. A final pulse of magmatic activity is represented by microgranodioritic enclave-
150 rich dykes that crosscut the regional schistosity and all granitic bodies. As they do not cross the
151 mesozoic stratas, those dykes are considered as Permo-Triassic in age.

152
153 The following chronology of magmatic events can be depicted in the Western Meseta domain. 1)
154 The early magmatic stage, characterized by a bimodal association, is constrained at 358-330 Ma
155 by U-Pb zircon dating in the Jebilet Massif (358-336 Ma, Delchini *et al.* 2018; 331.18-
156 329.67 Ma, Essaifi *et al.*, 2003). 2) The emplacement of syn-tectonic granodioritic to granitic
157 bodies is dated by SHRIMP U-Pb dating on magmatic zircons from xenoliths by Dostal *et al.*
158 (2005) at 285-328 Ma and by $^{40}\text{Ar}/^{39}\text{Ar}$ ages on a syn-tectonic granite and associated contact
159 metamorphism between 280 and 295 Ma. These ages confirm earlier whole-rock Rb/Sr ages by
160 Mrini *et al.* (1992) in the Jebilet massif and in the Oulmès and Ment area in the Central Massif. In
161 the Tichka area, the granitic stocks have been dated by U-Pb on zircon at 295-280 Ma (Rossi *et*
162 *al.*, 2016). 3) Late to post-tectonic magmatism is generally considered as Permo-Triassic, based
163 on whole-rock Rb-Sr data at 268 ± 6 Ma obtained on the Rehamna by Mrini *et al.* (1992) and

164 white micas $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages (275 Ma) on a post-tectonic pluton from the Rehamna by
165 Chopin *et al.* (2014), whereas U-Pb lower intercept age on a zircon rim from granulitic xenolith
166 obtained by Dostal *et al.* (2005) suggests Jurassic activity at 235 ± 8 Ma. 4) In the Mrit area, a K-
167 Ar age of 286 Ma obtained on a granitic dyke crosscut by microgranodioritic dyke (Cheilletz,
168 1984) provides an upper age limit, whereas the age for the latest dykes is constrained by Th/Pb
169 on monazite in the associated Pb-Zn-Ag veins located 2 km south of the sampling area at
170 254 ± 16 Ma (Rossi *et al.*, 2016).

171
172 The Paleozoic series of the Moroccan Variscan belt have been metamorphosed under greenschist
173 to amphibolite facies conditions (Michard *et al.*, 2008; 2010). The highest metamorphic
174 conditions are observed along a NE-SW shear zone (the Western Meseta Shear Zone) that
175 crosscut the Central, the Rehamna and the Jebilet massifs (Michard *et al.*, 2008). The syn-
176 tectonic metamorphic *PT* conditions in the Rehamna domain, reaching locally 620 °C, 6 kbar
177 (Wernert *et al.*, 2016), correspond to a *MP/MT* gradient of *ca.* 35°C/km (Aghzer and Arenas,
178 1995; 1998). Similar temperatures (628 ± 50 °C) are recorded in the Jebilet area (Delchini *et al.* ,
179 2016). In the study area however, metamorphic conditions do not exceed greenschist facies
180 conditions (365 ± 7 °C, Lahfid *et al.*, 2019). The age for *MP-MT* metamorphism is constrained by
181 few direct radiometric in the Rehamna by U-Pb dating on monazite and ^{40}Ar - ^{39}Ar dating on
182 amphibole and micas (Chopin *et al.*, 2014; Wernert *et al.*, 2016). The microstructural analysis
183 associated to *in situ* dating depicts a multistage metamorphic history, beginning in the late
184 Carboniferous (minimum ^{40}Ar - ^{30}Ar age of *ca.* 310 Ma) with a prograde metamorphic phase
185 associated to thickening followed by a syn-convergence exhumation phase at 310-295 Ma and
186 ending in the Permian (292-275 Ma), while late tectonic granites were emplaced. Contact
187 metamorphism related to syn- to post-tectonic magmatism is described in the Rehamna (Michard
188 *et al.*, 2010) and in the Jebilet (Delchini *et al.*, 2016), where maximum temperature reaches
189 600 °C for maximal pressure of 3 kbar. In the Central Massif, metamorphic aureoles
190 characterized by the presence of andalusite – biotite – K-feldspar association have been described
191 around the syn-tectonic calc-alkaline plutons and the post-tectonic leucogranitic bodies in Zaër,
192 Oulmès and Ment (Boushaba and Cailleux, 1992; Dahmani, 1995), which leads to pressure
193 estimates of 1.3 to 2 kbar and maximal temperatures of 600 °C.

194

195

196 Moreover, metamorphic xenoliths of sedimentary origin and of high metamorphic grade
197 (amphibolite to granulite facies conditions) have been sampled by late- to post-hercynian
198 magmatic bodies. In the Jebilet massif, late-hercynian granodioritic dykes contain both partially
199 molten quartz, feldspar, biotite bearing gneisses and K-feldspar, plagioclase, cordierite, garnet,
200 spinel, aluminosilicate, corundum and ilmenite bearing peraluminous enclaves (Bouloton *et al.*,
201 1991; Bouloton and Gasquet, 1995). In the quaternary basalts of the Middle Atlas,
202 metasedimentary enclaves are characterized by a quartz-feldspar-sillimanite-garnet association,
203 the rarity of biotite and the development of hercynite and orthopyroxene as a product of
204 incongruent melting in the granulite facies (Moukadiri and Bouloton, 1998; van Hoeflaken,
205 2011). In the Zaër granitic pluton (a late-hercynian intrusion of the Central Massif), Haïmeur *et*
206 *al.* (2003) describe a diversity of metapelitic enclaves, generally characterized by the presence of
207 quartz, biotite, feldspar and cordierite. Several types of xenoliths are distinguished based on the
208 presence or absence of aluminosilicate (andalusite partly replaced by fibrous sillimanite) and of
209 corundum, pyroxene and spinel relics. Corresponding *PT* conditions vary from 600-650 °C, 2 to
210 2.5 kbar for andalusite bearing samples and 690-770 °C, 4 to 5 kbar in the spinel bearing
211 xenoliths from the Zaër pluton (Haïmeur *et al.*, 2003), 750-800 °C, 3.5 to 3.9 kbar in the Jebilet
212 (Bouloton *et al.*, 1991) and up to 900 °C, 9-11 kbar in the recent lavas from Middle Atlas
213 (Moukadiri and Bouloton, 1998). Note also that granulite facies conditions (830-870 °C; 6 kbar)
214 are reported for high-grade migmatites and enclaves associated to mafic to acidic magmatism
215 from the Eastern Meseta domain (Elabouyi *et al.*, 2019).

216

217 A set of crustal xenoliths has been collected in the locality of Kariane, 2 km east of Mrirt. Two
218 dykes of 1 to 2 meters in thickness intrude the Ordovician metasediments, with a broadly N40
219 orientation (Ntarmouchant, 2003) and can be followed over a distance of *ca.* 2 km (Fig. 2a). The
220 emplacement depth of these dykes is not constrained. However, dykes of similar petrographic
221 type and stratigraphic age have been shown to crosscut both Variscan plutons and the Paleozoic
222 metasediments in the eastern Moroccan Massif Central (Ntarmouchant, 2003), or the contact
223 metamorphic aureoles in the Tighza district (Rossi *et al.*, 2016) and Oulmès and Ment plutons,
224 for which pressures were estimated at 2 and 1.3 kbar respectively (Dahmani, 1995; Delchini *et*
225 *al.*, 2016), which converts to depths of 5 to 7.5 km (considering lithostatic pressure and a density

226 of 2700 kg/m³). Moreover, metamorphic temperatures of the Paleozoic metasediments and the
227 geothermal gradient prevailing during the Variscan collision have been estimated from RSCM in
228 the Zaian mountains, and indicate depth of burial of 5 to 8.5 km (Lahfid *et al.*, 2019). We
229 consider this as an upper limit for the emplacement depth of the dykes. One of the two dykes of
230 the Kariane locality is of granitic composition and comprises few metasedimentary xenoliths.
231 Only one of those xenoliths has been collected for this study (2Kar8). The second dyke is a
232 microgranodioritic dyke which crosscuts the former one and is considered as post-orogenic since
233 it also crosscuts the regional schistosity (Fig. 2b). It comprises numerous granitic and
234 metasedimentary enclaves (metagreywackes and metapelites) (Ntarmouchand, 2003). The
235 metasedimentary enclaves (Fig. 2c and d) are of small size (less than 10 cm), often elongated in
236 shape and are non-oriented with respect to the dyke orientation. They can present or not an
237 internal schistosity and also present a mineralogic variability (*e.g.* presence or not of andalusite).
238 Twenty-one samples of metapelitic xenoliths from the microgranodioritic dyke have been
239 collected at different places along the dyke and studied in thin sections. A representative set of
240 those samples will be described in the following.

241

242 **3. Methods**

243

244 Samples collected in the field are centimeter size xenoliths (4-10 cm in length) included in their
245 microgranodioritic or microgranitic host rock. Each sample has been prepared in order to separate
246 i) one slice for thin section across the xenolith ii) one fragment (10 – 50 g) of xenolith for whole-
247 rock analysis and iii) in eight samples, one fragment of the host rock (*ca.* 100 g) for whole-rock
248 analysis. Thin sections have been made in the Géosciences environnement Toulouse laboratory
249 (Toulouse, France). Samples selected for whole-rock analyses were crushed with a jaw crusher
250 and pulverized in an agate mill. Five samples have been selected for detailed petrological
251 investigation on the basis of petrographical observation by optical microscopy and SEM.

252

253 Petrological studies by SEM and EPMA were carried out at the Géosciences environnement
254 Toulouse laboratory (GET, Université de Toulouse, France). Backscattered electron images as
255 well as EDS maps were obtained using the JEOL JSM 6360 LV Scanning Electron Microscope
256 (SEM) coupled to an energy dispersive spectrometry (EDS) analysis system Bruker Quantax 800

257 at the GET laboratory. The analytical conditions used were 20 kV for the acceleration voltage,
258 while a tungsten filament was used as the electron source. Semi-quantitative analytical
259 composition were extracted from selected micro-domains in the chemical maps using the Bruker
260 Hypermap software. Electron Probe Micro Analyses (EPMA) were obtained using the
261 CAMECA SX 50 on polished thin sections using a beam current of 20 nA and an acceleration
262 voltage of 15 keV for major elements. Monazites were analysed for U-Th-Pb dating with an
263 accelerating voltage of 15 kV and 300 nA probe current. The age and associated confidence
264 interval were calculated with the Rpackage NileDam (Montel *et al.* 1996; Seydoux-Guillaume *et*
265 *al.*, 2012; Villa-Vialaneix *et al.*, 2013). Analysis of the reference crystal MOACYR gave ages of
266 487 ± 25 Ma, in accordance with the recommended value of 474 ± 1 Ma (Seydoux-Guillaume *et*
267 *al.*, 2002), later revised to 504.3 ± 0.2 Ma (Gasquet *et al.* 2010) and 507.7 ± 1.3 Ma (Gonçalves
268 *et al.*, 2016).

269
270 Major elements and trace elements bulk-rock analyses were obtained by ICP-OES and ICP-MS
271 respectively in the Service d'Analyse des Roches et des Minéraux (SARM), Nancy (France),
272 following the analytical procedures described by Carignan *et al.* (2001).

273
274 Isochemical phase diagrams (pseudosections) were calculated in order to constrain metamorphic
275 conditions, *i.e.* pressure and temperature range, of the aluminous xenoliths. Mineral equilibrium
276 calculations were made in the NCKFMASHMT (Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–
277 H₂O–MnO–TiO₂) system using Perple_X 07 software (Connolly, 2005; Connolly, 2009) with the
278 Holland and Powell (1998) database updated in 2002. Except for sample *Ma10*, the composition
279 of the system was fixed at a local scale from SEM EDS maps. One pseudosection at the whole-
280 rock scale has been calculated for sample *Ma10*. We have estimated the water content from *T*-
281 *X*_{H₂O} diagrams by choosing *X*_{H₂O} leading to the best agreement with observed assemblages as
282 described in the petrography section, in particular the low volume of hydrous phases (namely
283 biotite and cordierite). The solid solutions models are: biotite (Tajčmanová *et al.*, 2009); chlorite
284 (Holland and Powell, 1998); garnet (Holland and Powell, 1998); melt (Holland and Powell, 2001;
285 White *et al.*, 2001); cordierite (Berman and Aranovich, 1996); plagioclase (Newton *et al.*, 1980);
286 K-feldspar (Waldbaum and Thompson, 1968); ilmenite (White *et al.*, 2000); spinel (Holland and

287 Powell, 1998); staurolite (Holland and Powell, 1998) and white mica (Coggon and Holland,
288 2002). Phase abbreviations are from Whitney and Evans (2010).

289

290 **4. Petrography**

291 *4.1. Xenolith Ma3*

292 Sample *Ma3* (Table 1) is characterized by a foliation underlined by biotite, sillimanite and
293 elongated garnet (Fig. 3a and b, Fig. 4a). Crystals of biotite are sometimes skeletal (Fig. 3a and
294 4a). Garnet forms elongated corroded crystals included in alkali-feldspar and contains inclusions
295 of ilmenite, biotite and numerous sillimanite needles oriented parallel to the main foliation (Fig.
296 3a and 4a). Spinel is rimmed by alkali-feldspar and often contains an inner foliation parallel to
297 the main foliation (Fig. 3b). Cordierite appears as large clear invasive crystals in textural
298 equilibrium with feldspar, locally enclosing internal schistosity marked by sillimanite fibers and
299 containing ilmenite grains (Fig. 3a).

300

301 Sample *Ma3* thus records a multi-stage metamorphic history. Biotite, sillimanite and garnet are
302 syn-kinematic minerals and constitute an early mineral paragenesis. Garnet and spinel are both
303 contemporaneous to late compared to the foliation, which they both enclose and crosscut in
304 several places. Feldspar and cordierite are post-kinematic minerals. The overgrowth of K-
305 feldspar at alkali feldspar rims (Fig. 4a and g) as well as the presence of interstitial alkali feldspar
306 enclosing euhedral porphyroblasts of spinel (Fig. 3b) suggest they grew from a melt, in
307 equilibrium with cordierite.

308

309 *4.2. Xenolith Ma10*

310 Euhedral porphyroblasts of andalusite and garnet are dispersed in a matrix composed of
311 feldspar, biotite, spinel and ilmenite (Fig. 3c and d) (Table 1). Andalusite shows a chiastolitic
312 form and appear surrounded by spinel and plagioclase coronas (Fig. 3c and d). In places,
313 andalusite crystals are fractured and filled with the same mineral association as in the matrix (Fig.
314 3d; Fig. 4c and i). Rare corundum spreads in the matrix or appears enclosed in garnet or
315 andalusite (Fig. 4c and i). Feldspars are either alkali feldspar or plagioclase. Plagioclase is
316 observed at the rim or in the cracks of andalusite crystals (Fig. 4b, c, h and i). Plagioclase forms a
317 continuous strip in contact with broken prisms of andalusite whereas K-rich feldspar is scattered

318 in the middle of the fracture (Fig. 4c and i). Tiny crystals of orthoclase are included within the
319 corona of cordierite around garnet. Sillimanite crosscuts and partly replaces andalusite (Fig. 3d),
320 locally also clearly crosscutting the feldspar bearing microcracks (not shown). Garnet crystals are
321 rimmed by cordierite and alkali feldspar (Fig. 3d and 4b). Spinel occurs around and within
322 andalusite crystals and as inclusion in feldspar (Fig. 3d).

323
324 To conclude, andalusite, garnet, corundum, alkali feldspar, biotite and ilmenite appear as an early
325 mineral association, whereas plagioclase, spinel, sillimanite and cordierite belong to later
326 parageneses. Several lines of evidence account for the presence of a melt. These include i) the
327 presence of feldspar-rich veinlets (Fig. 4c and i), ii) the clear spatial association of feldspar
328 surrounding euhedral porphyroblasts of spinel in the vicinity of aluminosilicate, which is
329 interpreted to record a melting reaction with spinel as peritectic product (*e.g.* Powell & Downes,
330 1990; Johnson *et al.*, 2004; Álvarez-Valero *et al.*, 2007) (Fig. 3d), iii) the zoning pattern in
331 plagioclase, characterized by a outwards decrease of the anorthitic content in the coronas
332 surrounding andalusite crystals and a decrease of the anorthitic content away from the edges in
333 the microfractures, which witnesses fractional growth from a liquid (Fig. 4i). iv) The presence of
334 ternary feldspars (*e.g.* Or₂₀, An₁₅, Ab₆₅ (Fig. 5a) which suggest that they formed under partial
335 melting conditions at *ca.* 900°C (Elkins and Grove, 1990).

336 337 4.3. Xenoliths Ma6 and Kar2b

338 Large euhedral fragmented crystals of andalusite are scattered in a fine-grained matrix of
339 plagioclase (Fig. 3e) (Table 1). Andalusite is rimmed by a spinel-plagioclase association (Fig. 3e
340 and f, Fig. 4d). Rare biotite (not shown) is scattered in the matrix. Acicular corundum appears as
341 inclusions within andalusite (Fig. 3f and h; Fig. 4d). Locally, fibers of sillimanite develop upon
342 andalusite (Fig. 3e and h). Large grains of cordierite are observed in the matrix, always separated
343 from andalusite by a spinel-plagioclase association (Fig. 3e).

344
345 Thus, corundum and andalusite are considered as early mineral phases, whereas sillimanite,
346 cordierite, plagioclase and spinel belong to later parageneses. The presence of plagioclase-spinel
347 association at andalusite rim and the presence of interstitial anhedral plagioclase enclosing
348 euhedral porphyroblasts of spinel (Fig. 3g) are interpreted as the witnesses of former partial

349 melting reactions including spinel as a peritectic phase (e.g. Powell & Downes, 1990; Johnson *et*
350 *al.*, 2004; Álvarez-Valero *et al.*, 2007).

351

352 4.4. Xenolith *Ma5*

353 Grey abundant spinel appears within large prisms of euhedral plagioclase (Fig. 3i and j; Table 1)
354 and scattered in a matrix composed of plagioclase and rare and biotite (Fig. 3l) (Table 1).

355 Corundum forms generally rounded crystals (Fig. 3k and 4f; Fig. 4l). Corroded garnet crystals
356 (Fig. 3l) form isolated relics rimmed by cordierite and enclosing ilmenite. The rounded shape of
357 garnet and corundum and the plagioclase matrix forming in places large euhedral crystals suggest
358 the former presence of melt. Anhedral interstitial cordierite (Fig. 4f) forms around feldspar and
359 some corundum crystals are rimmed by spinel and/or cordierite (Fig. 3k). This is broadly similar
360 to sample *Ma10*, except for the absence of andalusite in sample *Ma5*. However, the former
361 presence of andalusite is suggested by the squared shape of plagioclase-rich domains (Fig. 3i)
362 that can be interpreted as pseudomorphs of andalusite crystals.

363

364 The early mineral paragenesis therefore comprises garnet, corundum, biotite and ilmenite,
365 whereas spinel, cordierite and plagioclase belong to later parageneses. The presence of a former
366 partial melt is indicated by i) the euhedral twinned plagioclase crystals (Fig. 3j), ii) the presence
367 of interstitial anhedral plagioclase and/or alkali feldspar enclosing euhedral porphyroblasts of
368 corundum (Fig. 3j and k), iii) rational crystal faces of feldspar crystals and iv) formation of
369 overgrowths of spinel and cordierite on grains adjacent to feldspar (Fig. 3k) (Johnson *et al.*,
370 2010).

371

372 5. Mineral chemistry

373 *Feldspars*

374 In xenolith *Ma3*, only alkali-feldspar is present and crystals present chemical zoning (Or
375 62 to 90 mol.%, Table 2, Fig. 5a). Alkali-feldspar domains are formed by Na-rich crystals with
376 interstitial K-rich domains (Fig. 4g). In xenolith *Ma10*, feldspars composition is highly variable
377 from plagioclase (An 35 to 45 mol.%) to alkali feldspar (Or 47 to 92 mol.%) with several
378 intermediate ternary feldspar compositions (Table 2, Fig. 4h and 5a). Feldspar in xenoliths *Ma6*
379 and *Kar2b* is mainly plagioclase (An 26 to 74 mol.% and 15 to 45 mol.% respectively, Table 2,

380 Fig. 5a) associated with corundum and spinel in replacement of andalusite (Fig. 4j and k). Rare
381 alkali feldspar (Or 57 to 88 mol.%) is also present in sample *Ma6*. In xenolith *Ma5*, feldspar is a
382 homogeneous (Table 2, Fig. 4l and 5a), with An content comprised between 45 and 53 mol.%.

383

384 *Spinel*

385 Most of crystals of spinel are grey and belong to hercynitic spinel (Table 3). X_{Fe} varies
386 from sample to sample but appears mostly homogeneous in each xenolith. The lowest X_{Fe} (0.57 to
387 0.64) is observed in xenolith *Ma5*, the highest in xenolith *Ma6* (0.70 to 0.74), whereas it is
388 intermediate in samples *Ma3* (0.67-0.72), *Ma10* (0.63-0.79) and *Kar2b* (0.64-0.69). The ZnO
389 content reaches 0.5 wt.%.

390

391 *Garnet*

392 Garnet is present in xenoliths *Ma3*, *Ma10* and *Ma5*. They are almandine-pyrope garnets,
393 with composition ranging from $X_{Alm} = 64$ to 73 % in *Ma3* and from $X_{Alm} = 59$ to 67 % in *Ma10*
394 (Table 4). Garnet from *Ma10* is generally spessartine - richer than garnet from *Ma3* (X_{Sps} up to
395 6 %), with spessartine content presenting typical bell-shaped profile (not shown).

396

397 *Biotite*

398 Biotite is generally Ti-rich (TiO_2 3 to 7 wt.% in *Ma10* and up to 8 wt.% in *Ma3*) (Table 4, Fig.
399 5b). The lowest TiO_2 is observed in sample *Ma5* ($TiO_2 < 5$ wt.%). X_{Mg} varies from 0.45 in *Ma3*,
400 0.4 to 0.56 in *Ma10* and 0.49 to 0.56 in *Ma5*.

401

402 **6. Petrological and geochemical relationships between host rocks and xenoliths**

403

404 *6.1. Host rocks and xenolith contacts*

405 The microgranodiorite host rock has a porphyritic texture with phenocrysts of plagioclase,
406 amphibole, biotite and quartz in a fine-grained matrix composed of quartz, plagioclase, K-
407 feldspar and rare biotite (Fig. 6a). In sample *Ma3* (Fig. 6b), the contact is sharp. A few xenocrysts
408 of biotite from the xenolith are nevertheless present in the host rock near the contact, accounting
409 for mechanical exchange between the host magma and the xenolith. In sample *Ma10* (Fig. 6c),
410 the contact is characterized by a two mm large rim along the xenolith limit where abundant

411 pinnite replaces cordierite. In sample *Ma6*, the contact is underlined by the presence of large
412 andalusite crystals broadly similar to those of the core of the xenoliths and preserving spinel
413 relics (Fig. 6d and e). Such andalusite crystals are sometimes fully enclosed in the host rock,
414 featuring mechanical exchange between the host magma and the xenolith. In sample *Ma5* (Fig.
415 6f), the contact is sharp, with a two mm large rim along the xenolith limit. In the rim, the host
416 rock is characterized by reduced crystal size, the lack of phenocrystals, and the absence of biotite.
417

418 6.2. Geochemistry of host rocks and xenoliths

419 Major and trace elements have been analyzed in the xenoliths and their host rocks (Table
420 5). Xenoliths have a low to intermediate SiO₂ content of 44 to 57 wt.% (Table 5, Fig. 7a). They
421 are characterized by a high although variable Al₂O₃ content (18-34 wt.%) that reaches very high
422 values (*ca.* > 30 %) in the silica poorer samples. Alkali content is high (5 < Na₂O+K₂O < 9 wt.%).
423 Host rocks have a SiO₂ content of 62 to 68 wt.% (Table 5). In the SiO₂-Na₂O+K₂O diagram (Fig.
424 7b, Cox *et al.*, 1979), 6 samples fall in the field of quartz diorite. One sample, *2Kar 8*, coming
425 from a different dyke, has the composition of a granite. Compared to the host rock, the xenoliths
426 are enriched in a number of trace elements (Fig. 8a and 9), notably HFSE (Nb, Ti) and some of
427 the transition elements (Cr, V, W, Ni). By contrast, there is no element for which there is a
428 systematic enrichment of the host rock compared to the xenolith.

429
430 Host rocks present very similar decreasing REE spectra, with (La/Yb)_N in the range 13-16 (Fig.
431 8b and 9a). The microgranitic sample *2Kar 8* presents a slight Eu anomaly. Xenolith samples also
432 display decreasing REE spectra, broadly parallel, either above or below the host-rock spectra,
433 with 5 samples presenting (La/Yb)_N values similar to that of the host rock and 4 samples
434 presenting (La/Yb)_N scattered between 5 and 30 (Fig. 8b). Note that the (La/Yb)_N ratio is not
435 correlated to the SiO₂ content of the xenolith. The overall REE content varies by an order of
436 magnitude in the xenoliths, the more enriched samples (*Ma3, Ma10, Ma5, 2Kar5, Kar2b*) having
437 the lowest SiO₂ content (44-51 wt.%). The extended spectra highlight negative anomalies for both
438 the host rocks and the xenoliths in Nb, Ta, Ti and P, and a positive anomaly in Pb. No correlation
439 was found between the chemical composition of the host rock and corresponding xenolith.

440

441 7. Mineral equilibria modeling

442

443 7.1. *Ma3 xenolith*

444 A pseudosection has been drawn for the local composition of a microdomain in sample
445 *Ma3* (EDS map Fig. 6a and Fig.10). The microdomain is fully representative of the mineral
446 assemblages at the thin section scale. The mineral assemblage (biotite + melt + plagioclase + K-
447 feldspar + spinel + garnet + sillimanite + ilmenite) is stable in the pseudosection for a
448 temperature range of 850-900°C and for a pressure higher than 5 kbar. Isopleths curves for garnet
449 and biotite are reported in the pseudosection and are in agreement with such *PT* range. They
450 define a narrow range between 850 and 915 °C. Modes at this temperature correspond to K-
451 feldspar 50 wt.%, biotite 10 wt.%, garnet 15wt.%, ilmenite 5 wt.%, melt 10 wt.%, plagioclase
452 <5 wt.%, sillimanite <5 wt.% and spinel 5 wt.%, in agreement with petrographic observations.
453 The fraction of melt calculated at 850-900 °C is estimated as 15-25%. Partial melting may have
454 started at a temperature of *ca.* 725 °C in the stability field of garnet and sillimanite for a pressure
455 above 5 kbar. The presence of cordierite accounts for a decompression below 5 kbar. These
456 characteristics are accounted for by a prograde path at a pressure above 5 kbar from 725 to 850-
457 900 °C, followed by a decompression path toward lower pressure and temperature conditions, as
458 shown in Fig. 10.

459

460 7.2. *Ma10 xenolith*

461 The pseudosection (Fig. 11) for sample *Ma10* has been calculated for whole rock
462 composition (recalculated from Table 5 for a water content of 0.54 wt.% determined from *T-X_{H2O}*
463 section) since none of the microdomain was representative of the sample. In the pseudosection
464 the mineral assemblage (biotite + cordierite + plagioclase + K-feldspar + garnet + spinel +
465 sillimanite + ilmenite+melt) is reached for temperatures between ~800 to 850 °C and pressures
466 between 4 to 6 kbar. The upper and lower temperatures are constrained by the presence of biotite
467 and spinel in the mineral assemblage respectively, whereas pressure is constrained by the garnet-
468 cordierite stability field. The isopleths for plagioclase, biotite and garnet compositions better
469 constrain the temperature to ~850 °C (Fig. 11). Modes at 850 °C correspond to biotite 15 wt.%,
470 plagioclase 10 wt.%, K-feldspar 20 wt.%, sillimanite 30 wt.%, garnet 10 wt %, cordierite
471 <5 wt.%, spinel <5 wt.%, melt <5 wt% and ilmenite <5 wt.%, and are in agreement with
472 petrographic observations. In addition, the observations of i) corroded garnet grains and

473 cordierite growth in the presence of biotite and spinel and ii) spinel at andalusite/sillimanite rims
474 (Fig. 3c) indicate a temperature increase above 4 kbar to reach the peak *PT* conditions.
475 Andalusite chiastolite attests for an early metamorphism (*PT* conditions outside of the range of
476 the pseudosection). Sillimanite may have grown at the expense of andalusite during heating.
477 Sillimanite fibers crosscutting the feldspar-rich veins and the modes calculated from the model
478 (Supplemental material 2) suggest as an alternative that cooling may then have produced both
479 crystallization of feldspar from the melt and growth of sillimanite fibers fringing andalusite
480 crystals.

481

482 7.3. *Ma6 xenolith*

483 The composition of a microdomain (Fig. 4d and j), silica poor and alumina rich, has been
484 used to compute a pseudosection (Fig. 12). It is representative in terms of petrographic
485 characteristics of the sample at the thin section scale. The H₂O content was fixed to be just low
486 enough to account for the absence of biotite at the solidus, in order to match the observations, *i.e.*
487 no biotite on the microdomain and rare biotite at the xenolith scale. In the pseudosection, partial
488 melting takes place above 800 °C thanks to the consumption of K-feldspar and, at low pressure
489 (<3 kbar) produces small amounts of spinel (*e.g.* Fig. 3g) while consuming cordierite
490 (Supplemental material 3). The observed assemblage corresponds to temperatures higher than
491 900 °C, as constrained by partial melting and the absence of K-feldspar. Modes at 900 °C
492 correspond to plagioclase 30 wt.%, corundum 15 wt.%, cordierite 20 wt.%, melt 10 wt.%, spinel
493 10 wt.%, K-feldspar 5 wt.%; sillimanite 10 wt.% and ilmenite <0.4 wt.%, in agreement with
494 petrographic observations. Moreover, the presence of andalusite resorbed and rimmed by a spinel
495 – plagioclase association accounts for an initial metamorphic stage at a temperature lower than
496 700 °C and a pressure lower than 2 kbar. Andalusite then remains in a metastable state, without
497 being replaced by sillimanite in this microdomain. To summarize, sample *Ma6* thus record partial
498 melting at low pressure (< 2 kbar) with a temperature increase from less than 700 °C to more than
499 900 °C.

500

501 7.4. *Kar2b xenolith*

502 The composition corresponding to the microdomain of Fig. 4e and k, silica poor and alumina
503 rich, has been used to compute the pseudosection of Fig. 13. The petrographical features are

504 representative of the sample at the sample scale. The H₂O content was fixed to be just low
505 enough to account for the absence of biotite at the solidus, in order to match the observations, *i.e.*
506 no biotite on the microdomain and rare biotite at the xenolith scale.

507 The absence of K-feldspar suggests its almost total consumption by partial melting reaction at *ca.*
508 900 °C, whereas the absence of garnet constrains pressure to below 4 kbar (Fig. 13). At such low
509 pressure, partial melting would produce spinel (Fig. 3h) at the expense of cordierite
510 (Supplemental material 4). Modes at this *PT* range correspond to plagioclase 45 wt.%, cordierite
511 10 wt.%, sillimanite 10 wt.%, corundum 10 wt.%, spinel 10 wt.%, melt 15 wt.%, ilmenite
512 <1 wt.%. Moreover, metastable andalusite grains account for an initial temperature below
513 650 °C, at a pressure that is further constrained by the absence of garnet to below 2 kbar. As a
514 whole, the observations suggest an increase of temperature from *ca.* 650 °C to more than 900 °C,
515 at a pressure lower than 2 kbar. This evolution is fully consistent with the evolution estimated for
516 sample *Ma6*. Cooling may then have produced both crystallization of feldspar from the melt and
517 possibly sillimanite fibers growing on andalusite crystals (Supplemental material 4).

518

519 7.5. *Ma5 xenolith*

520 A pseudosection (Fig. 14) has been drawn for the local composition of a microdomain in
521 sample *Ma5* (Fig. 4f and l). It differs from the sample at the sample scale in that there is no garnet
522 relict. In the pseudosection, the mineral assemblage (cordierite + plagioclase + spinel +
523 corundum + ilmenite ± biotite) is stable for temperature of higher than 800°C and for pressure
524 lower than 5 kbar. Mineral modes at 850 °C correspond to plagioclase 55 wt.%, cordierite
525 9 wt.%, spinel 10 wt.%, ilmenite 2 wt.%, K-feldspar 7 wt.%, corundum 1 wt.% and melt
526 16 wt.%. These proportions are in close agreement with petrographic observations. The
527 petrographic observation of spinel corona surrounding corroded corundum (Fig. 3k) argues for an
528 increase of spinel proportions along with decrease of corundum above a temperature of 800 °C
529 (Supplemental material 5). Although not visible in the microdomain, the presence of garnet
530 relics elsewhere in the xenolith records previous mineral equilibrium at lower temperature and/or
531 higher pressure (650 to 800 °C for pressures in the range 2.5 – 5 kbar).

532

533 8. Ages constraints

534 A few small (<10 µm) monazite crystals have been found in xenoliths *Ma3* and *Ma6*. These were
535 too small and rare for LA-ICPMS analyses but U-Pb dating was attempted by the U-Th-Pb
536 chemical dating method by EPMA. The diameter of the interaction volume below the electron
537 beam (4 µm) allowed only 7 analysis to be done in sample *Ma6* and 21 in sample *Ma3*. In sample
538 *Ma6*, individual data give ages lower than 230 Ma that do not define a single population. By
539 contrast, in sample *Ma3*, the analyses, although slightly dispersed, provide an age of
540 293 ± 25 Ma, or, if restricted to a more coherent population of 11 analyses, 304 ± 3 Ma (Fig. 15).

541

542 **9. A petrogenetic model for the crustal xenoliths from the Moroccan Central Massif**

543

544 *9.1. Evidence for prograde multi-stage evolution and partial melting in the xenoliths*

545 Regional metamorphism is best represented in sample *Ma3* where biotite, sillimanite and garnet
546 underline the foliation and indicate pressure above *ca.* 5 kbar. In low pressure xenoliths (*Ma6* and
547 *Kar2b*), euhedral crystals of andalusite, randomly scattered in the matrix, belong to an early
548 paragenesis and may have formed during contact metamorphism prior to the entrapment or at the
549 onset of heating in the dyke.

550 The peak temperature parageneses formed during ongoing partial melting, as shown by i) textural
551 observations, *i.e.* the presence of feldspar-rich veinlets (*e.g.* Fig. 4c) and magmatic textures of
552 feldspar associations (Fig. 3b, g and j), ii) chemical zoning of feldspar which witnesses fractional
553 growth from a liquid (Fig. 4g, k), iii) the presence of ternary feldspar (Fig. 5a) that suggest that
554 they formed under partial melting conditions at *ca.* 900 °C (Elkins and Grove, 1990).

555 In high pressure samples (*Ma3*, *Ma10*, *Ma5*), post-kinematic mineral associations formed during
556 partial melting overprint foliated domains. These include coronas of cordierite + ilmenite + alkali
557 feldspar around garnet (Fig. 3e, Fig. 4b and h) and spinel crystals crosscutting initial foliation
558 (Fig. 3b) or as coronas around andalusite pseudomorphs (Fig. 3i). In low-pressure samples,
559 secondary minerals associated to partial melting include coronas of spinel and plagioclase around
560 and inside andalusite (Fig. 3f, Fig. 4e and k), and secondary sillimanite developed upon
561 porphyroblasts of andalusite (Fig. 3e). The new products (including melt, neoblasts of spinel,

562 cordierite, K-feldspar, ilmenite) formed from the (garnet) - biotite – andalusite (sillimanite)
563 assemblages. (Cesare 2000; Álvarez-Valero *et al.*, 2007; Álvarez-Valero and Waters, 2010).

564 Partial melting reactions relevant to the conditions observed in the xenoliths of the Moroccan
565 Central Massif have been reproduced in experiments designed to reproduce assimilation of
566 metasedimentary xenoliths in a granodioritic magma at lower crustal level (4 kbar) under water
567 undersaturated conditions by Díaz-Alvarado *et al.* (2011). During heating and partial melting,
568 they have shown that the initial parageneses of metapelitic xenoliths have evolved to form
569 cordierite, spinel and K-feldspar. These minerals may be formed according to the well-studied
570 reactions (*e.g.* Bouloton *et al.*, 1991; Vielzeuf and Holloway, 1988; Patiño Douce and Johnson,
571 1991; Johnson *et al.*, 2004; Álvarez-Valero *et al.* 2007) :



574 or, in garnet-free samples (Buick *et al.*, 2004) :



576 Experiments undertaken at 3 kbar from 700 ° to 1000 °C on water-rich or water-poor metapelites
577 in order to explain contact metamorphism assemblages (Buick *et al.*, 2004) also display mineral
578 assemblages and melting reactions relevant to the present study. In water poor samples, partial
579 melting begins slightly below 750 °C while muscovite, quartz, andalusite, biotite and plagioclase
580 remain from the protolith and sillimanite appears as well as new crystals of cordierite, biotite and
581 plagioclase. Spinel appears in the experiments at *ca.* 800 °C and orthopyroxene at 1000 °C. Melt
582 production is progressive with a main production between 800 and 950 °C when melt fraction
583 reaches *ca.* 50 %. An association of spinel + secondary plagioclase surrounding relicts of
584 andalusite is noteworthy above 800 °C. Moreover, this secondary plagioclase appears richer in
585 An, and melt is peraluminous granitic in composition with normative corundum. This evolution
586 can be compared to that of the more silica-rich sample *Ma3*, the others having different, silica-
587 depleted compositions. In sample *Ma3*, the paragenesis (Bt, Grt, Crd, Pl, Kfs, Spl, Ilm and melt)
588 is comparable to those observed in the experiments between 800 and 900 °C, whereas the

589 comparison with the thermodynamic model (Fig. 10) allows to reduce this interval to 850-900 °C
590 for a pressure of *ca.* 5 kbar.

591 The melting reactions can be deciphered by a comparison of textural observation and
592 thermodynamic modeling. For the composition of the microdomain in *Ma3* (Fig. 10), at high
593 pressure (6 kbar), melting begins at *ca.* 750 °C (Fig. 16a) and involves consumption of biotite,
594 plagioclase, quartz and produces low amount of garnet (reaction 1). The appearance of spinel is
595 linked to a second major melt-producing reaction at 900°C involving the consumption of
596 sillimanite and biotite and the production of ilmenite together with spinel. This leads to the
597 observed mineral assemblage in sample *Ma3* if we except late cordierite that appears at the
598 expanse of garnet during decompression and cooling. At lower pressure (2.5 kbar, Fig. 16b),
599 thermodynamic modeling shows that melting reaction begins at lower temperature (*ca.* 680 °C)
600 and produces cordierite instead of garnet. The appearance of spinel as a peritectic phase takes
601 place at *ca.* 850 °C and melt is produced at the expense of biotite, plagioclase, K-feldspar and
602 cordierite. Note also that partial melting leads to the disappearance of quartz at high temperature
603 at 775 to 825 °C depending at low and high pressure respectively. However, for the composition
604 of the microdomain in sample *Ma5* (Fig. 14), melting is slightly different (Fig. 16c). It is shifted
605 to higher temperature (800 °C) and involves a single process of consumption of biotite,
606 plagioclase and cordierite but also corundum and production of spinel, K-feldspar and ilmenite.
607 According to the simulation, the growth of secondary cordierite in low temperature sample *Ma5*
608 and *Kar2b* would not be directly related to partial melting but rather to an earlier solid-state
609 reaction consuming biotite, sillimanite and small amounts of spinel to form cordierite, K-feldspar
610 and corundum.

611

612 9.2. The significance of corundum and the depleted protolith of the xenoliths

613 Corundum has been observed in all samples except *Ma3*. These samples are characterized by
614 their poor to intermediate SiO₂ and high Al₂O₃ content (44 to 57 wt.% SiO₂, 17.5 wt.% to
615 33.8 wt.% Al₂O₃). Corundum belongs to the early mineral parageneses since it appears as
616 inclusion in andalusite and garnets. However, it remains euhedral and stable in the late, high
617 temperature parageneses together with feldspar and spinel. Thermodynamic calculations show

618 that the presence of corundum is controlled by the chemical composition rather than by the
619 pressure-temperature conditions. Thus, in the simulations, corundum is absent when SiO₂ is
620 higher than 50 wt.% and Al₂O₃ lower than 30 wt.% (Fig. 10), present over the whole *PT* range
621 when SiO₂ is lower than 40 wt.% and Al₂O₃ higher than 45 wt.% (Fig. 12 and 13) and restricted
622 to limited low-pressure domains (<5 kbar) at high temperature (>600 °C) for intermediate SiO₂
623 and Al₂O₃ contents (Fig. 11 and 14). Reaction textures preserved in spinel–corundum–feldspar-
624 rich rocks show corundum partially replaced by spinel, which is consistent with numerical
625 modelling that shows corundum consumption and spinel production above 800°C (Fig. 16c) and
626 has been already described by Johnson *et al.* (2010). It has been suggested that the presence of
627 corundum is favoured by high H₂O content (Johnson *et al.*, 2010). However, according to
628 thermodynamic modelling corresponding to the composition of sample *Ma5* (Fig. 16d), an
629 increase in H₂O content (from 0 to the observed value of 0.56 wt.%) would rather correspond to
630 the disappearance of corundum. At the composition corresponding to the studied microdomain,
631 corundum is present at a pressure of 3kb from 550 °C to 980 °C with a maximum modal
632 abundance of 6% at *ca.* 850 °C.

633

634 *9.3. Magma – xenoliths interactions and the origin of the refractory nature of the xenoliths*

635 There are two lines of evidence for chemical – mechanical magma-xenolith interactions in the
636 samples from the Moroccan Central Massif. The first one is the petrographic observation of some
637 digestion of the xenolith by the enclosing magma. This is marked by the presence of biotite and
638 andalusite xenocrysts in the host rock at the contact with xenoliths in a number of samples (Fig.
639 3b, d and e). The second one is the observation of a linear relationship between the transition
640 element and SiO₂ content in the host rock sampled in the vicinity of the xenoliths, which we
641 tentatively interpret as a witness of chemical mixing between transition element enriched
642 xenoliths and the enclosing magma (Fig. 8a). Such digestion of the xenoliths by the enclosing
643 magma, and partial melting and melt loss due to heating at their contact, has been long observed
644 and taken as a marker of crustal contamination during magma ascent. In particular, this process
645 has been recently addressed on Díaz-Alvarado *et al.* (2011), which explained the shift from calc-
646 alkaline magmas toward peraluminous granite composition by the assimilation of crustal xenolith

647 that underwent partial melting and retained a solid fraction containing in particular the peritectic
648 phases formed by partial melting reactions.

649
650 Such a process requires substantial amount of melt loss, which is difficult to evidence from the
651 chemistry and the mineralogy of the residue. Indeed, as recalled by Johnson *et al.* (2010), melt
652 loss at a given temperature would not change significantly mineral proportion and chemical
653 composition of the residue. Hence, the refractory nature of the xenolith, marked by the abundance
654 of aluminosilicate phases, spinel and corundum, the low SiO₂ and large Al₂O₃ content, must be
655 largely inherited from the protolith. Considering the partition coefficient from Bea *et al.*, (1994)
656 relevant for partial melting of crustal material, we have calculated the evolution of the trace
657 element content of the residues for samples *Ma3* (model 1) and *Ma5* (model 2) in the case of non
658 modal batch melting, *i.e.*

$$659 \quad \frac{C_l}{C_0} = \frac{1}{D + F(1 - P)}$$

660 where C_l is the concentration of the element of interest in the partial melt, C_0 is the concentration
661 in the initial solid, D is the bulk partition coefficient of the solid and P is the partition coefficient
662 weighted by the proportion that each mineral contributes to the melt. Mineral proportion at the
663 onset of cooling and the proportion of the mineral that contributes to the melt at a final
664 temperature of 900°C are taken from the thermodynamic calculation (Table 6). The two samples
665 represent contrasted situation where, in the case of *Ma3*, melting is dominated by the
666 consumption of biotite and quartz and garnet growth in the solid as a peritectic phase, while in
667 the case of *Ma5*, cordierite and Kfs contribute significantly to melting in addition to biotite, in the
668 absence of garnet. In the first case, the bulk partition coefficients for REE vary from 0.30 (Sm) to
669 12.17 (Lu) while in the second case, they vary from 0.89 (Sm) to 2.78 (La) (Table 6), *i.e.* in the
670 first case, the LREE and MREE will behave as slightly incompatible elements whereas the HREE
671 will behave as compatible elements, whereas in the second case, all REE but Sm and Yb will
672 behave as compatible elements during partial melting. In both cases, the bulk partition
673 coefficients for transition elements are much higher than 1 and are generally higher in the case of
674 *Ma3*. As a result, there will be very little variation in the trace element content of the residual
675 solid C_s compared to the initial solid C_0 during partial melting at the considered temperature,
676 where the melt fraction does not exceed 20%. The C_s/C_0 is in the range 0.7 to 1.23 and 0.96 to

677 1.1 respectively for REE for model 1 and model 2, and is in the range 1.14 to 1.23 and 1.05 to
678 1.15 for transition elements (Table 6). This is much lower than the observed variations among
679 xenoliths (Fig. 9, Table 5), where the REE content varies by a factor of 4.5 to 7.5 depending on
680 the REE and the transition element content by a factor of 2.1 to 2.9. Thus, the chemical diversity
681 among xenolith cannot be attributed to different degrees of partial melting and melt extraction
682 due to their incorporation in the magma. The diversity in major and trace element composition
683 must be attributed to the diversity of their metapelitic protoliths. Thus, the protolith from which
684 xenoliths have been derived must have undergone first partial melt extraction to explain their
685 variable depletion in SiO₂ and enrichment in Al₂O₃ compared to standard metapelites. This must
686 have taken place before entrapment in the magmas, *i.e.* during either earlier regional
687 metamorphism or contact metamorphism related magma emplacement, as will be discussed
688 below. This conclusion is in line with those proposed by other authors (*e.g.* Aguilar *et al.*, 2016)
689 that multiple melt extraction is necessary to shift the composition of the solid residue toward
690 marked refractory composition.

691

692

693 **10. Xenoliths as witnesses of metamorphic conditions of late Hercynian metamorphism in** 694 **Moroccan Meseta**

695

696 Three types of xenoliths can be distinguished based on their pressure-temperature record (Fig.
697 17). The first type, represented by samples *Ma3* and *Ma10*, record pressures higher than 5 kbar
698 (as demonstrated by the presence of garnet) for temperatures exceeding 800°C (as demonstrated
699 by the presence of spinel associated to partial melting). A decompression path at high
700 temperature is recorded in both samples by the transformation of garnet into cordierite. The
701 second type, represented by sample *Ma5*, records intermediate pressure of ca 3 kbar for a
702 temperature of 750 to 800 °C. Few garnet relics surrounded by a corona of cordierite testify for
703 limited heating / decompression path. At last, samples *Ma6* and *Kar2b* exemplify a third type of
704 xenoliths, that record low pressure (<2.5 kbar), as demonstrated by the abundance of andalusite
705 relics and the absence of garnet. The transformation of an initial andalusite bearing assemblage
706 into a sillimanite-corundum-spinel-cordierite assemblage associated to partial melting records a
707 major heating from below 700°C to *ca.* 900°C at constant pressure.

708

709 As a whole, the xenoliths have thus sampled a crustal segment from *ca.* 5 to 25 km (Fig. 17),
710 considering a lithostatic pressure and a density of 2700 kg.m⁻³. The overall pressure range
711 recorded by the xenoliths matches the pressure range recorded by metamorphic rocks
712 outcropping in the Rehamna (6 to 9 kbar; Aghzer et Arenas, 1998; Wernert *et al.*, 2016), in the
713 Jebilet (1.5 to 5 kbar; Delchini *et al.*, 2016) and in the Eastern Meseta (6 kbar, Elabouyi *et al.*,
714 2019) and by the crustal xenoliths in late Hercynian magmas from the Jebilet (3 kbar, Bouloton *et*
715 *al.*, 1991), in the Middle Atlas (9-11 kbar; Moukadiri and Bouloton, 1998), and in the Zaër pluton
716 (2-5 kbar, Haïmeur *et al.*, 2003). The absence of xenoliths recording a pressure higher than
717 11 kbar suggests either that the lower crustal levels have not been sampled by the magmas
718 coming from the underlying mantle or that the Hercynian orogenic crust was only moderately
719 thickened (less than 40 km) in the area.

720 Based on geochemical and isotopic data, it has been argued that the analogue Permian
721 granodioritic magmas have a mixed mantle-crustal origin with limited crustal component
722 (⁸⁷Sr/⁸⁶Sr up to 0.705; Mrini *et al.*, 1992, for the Central Massif and the Rehamna; Gasquet *et al.*,
723 1992 for the Tichka Massif; Youbi *et al.*, 2001 and Dostal *et al.*, 2005 for the dykes of the
724 Jebilet). This is confirmed by the trace element data on the microgranodiorite and microgranite
725 dykes from Kariane that show a marked negative Nb–Ta anomaly that can best be explained by
726 crustal contamination in a late orogenic context.

727
728 The *PT* conditions recorded by the xenoliths corresponding to the maximum temperature are
729 consistently in the range 800 to 900 °C, regardless of the pressure. Moreover, in xenoliths *Ma3*
730 and *Ma10*, which record the highest pressure (*ca.* 6 kbar), the temperature associated to the
731 earliest metamorphic stage (>850 °C) is higher than the metamorphic temperature at the same
732 pressure reported from regional metamorphism associated to the orogenic stages in the D1- D2
733 deformation events in the Rehamna (Aghzer et Arenas, 1998; Wernert *et al.*, 2016), in the Jebilet
734 (Delchini *et al.*, 2016) and in the Eastern Moulouya (Elabouyi *et al.* 2019) (Fig. 17). This
735 suggests that the maximal temperature records thermal metamorphism associated to the
736 incorporation into the magma, as already suggested by Bouloton *et al.* (1991). The peak-
737 temperature range (800-900 °C) recorded by the xenoliths lies in between the solidus (700-
738 800 °C) and liquidus (900-1000 °C) temperatures calculated for the host magmas (Table 5) over

739 the pressure range of 1-6 kbar with MELTS-Excel sheet (Gualda and Ghiorso, 2015). Given the
740 small size of the xenoliths (<10 cm), thermal equilibrium between the host magma and the
741 xenolith should be achieved in less than one day. On the other hand, the persistence of andalusite
742 in a metastable form suggests a short metamorphic history. This is consistent with the
743 characteristic cooling time t of *ca.* 10 days that can be estimated for a dyke of half-thickness d of
744 1 m considering a thermal diffusivity K of *ca.* $1 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ and a conductive model where
745 $t = d^2/(4K)$ (Turcotte and Schubert, 2003). The pressures and temperatures recorded by the
746 xenoliths in Kariane are in good agreement with those reported in xenoliths in other places in
747 Morocco (Fig. 17). The xenoliths in Oulad Ouaslam (Jebilet) described by Bouloton *et al.* (1991)
748 record PT conditions of *ca.* 750 °C, 3 kbar, similar to xenolith *Ma5*. The xenoliths from the Zaër
749 pluton (Central Massif) studied by Haïmeur *et al.* (2003) record PT conditions of 600-650 °C, 2
750 to 2.5 kbar for andalusite-bearing xenoliths, similar to xenoliths *Ma6* and *Kar2b*, while
751 corundum-spinel-bearing xenoliths record PT conditions of 690-770°C, 4 to 5 kbar, similar to
752 samples *Ma3* and *Ma10*. The PT conditions retrieved from migmatitic terrains associated to both
753 M-I type and S-type magmatism in the so-called Colline à Essaim d'Enclaves of the Eastern
754 Meseta by Elabouyi *et al.* (2019) record an isobaric heating history with maximal PT condition in
755 full agreement with those recorded by samples *Ma3* and *Ma10*. At last, granulitic xenoliths from
756 Plio-Quaternary basalts from the Middle Atlas record the highest pressures (9-11 kbar) for
757 temperatures of *ca.* 900 °C similar to those of samples *Kar2b* and *Ma6*. In the case of the
758 xenoliths from the Kariane locality, the age of the magmatic event responsible for the entrapment
759 of the xenoliths in the dykes and the associated metamorphism is not well constrained. It can be
760 attributed to the Permian both on stratigraphic grounds and on previous radiometric dating
761 (Cheilletz, 1984; Rossi *et al.*, 2016). The ages obtained in the present study on monazites from
762 the xenolith (EPMA age of $293 \pm 25 \text{ Ma}$) are consistent with those earlier age estimates.

763
764 The retrieval of the PT conditions that prevailed before the incorporation of the xenolith into the
765 magma is difficult to assess. Regional metamorphism or syn-tectonic contact metamorphism is
766 attested in the xenoliths from the Kariane locality by the oriented sillimanite-biotite-garnet
767 association in sample *Ma3*, but the pressure-temperature record is obliterated by the temperature
768 increase due to the entrapment in the magma. The early andalusite-bearing paragenesis in
769 samples *Ma6* and *Kar2b* records pressures and temperature conditions of *ca.* 600 °C for a

770 pressure lower than 2.5 kbar, which is consistent both with the *PT* estimates of some of the
771 xenoliths from the Zaër Pluton (Haïmeur *et al.*, 2003) and with the temperature and pressure
772 associated to contact metamorphism in amphibolite facies in metapelites from the Jebilet Massif
773 by Delchini *et al.* (2016). However, in sample *Ma10*, the presence of andalusite chiastolite
774 attesting for an early contact metamorphism implies *a priori* a pressure lower than 4 kbar, which
775 is not consistent with the peak *PT* conditions (>4.5 kbar). This discrepancy may be explained
776 either by a burial of the sample after contact metamorphism and before entrapment in the magma
777 (orange line in Fig. 17), or if we admit a high-pressure high-temperature position for the silicate
778 triple point (4.5 ± 0.5 kbar, 550 ± 35 °C) (Pattison, 1992; Cesare *et al.*, 2003) rather than the
779 position of the triple point from Holland and Powell (1998) data base used in the thermodynamic
780 models (Fig.17). In the latter case, slightly higher *P* and *T* conditions for the early andalusite-
781 bearing stages are retrieved both for samples *Ma6* and *Kar2b* (*ca.* 700 °C for a pressure lower
782 than 2.5 kbar) and the prograde evolution of sample *Ma10* can be explained by isobaric heating
783 from 550°C toward 800 °C at *ca.* 4.5 kbar (dashed orange line in Fig. 17). There is no available
784 dating of those initial stages of metamorphism, which can represent either the early stage of
785 heating and incorporation in the Permian dykes, or a former stage of contact metamorphism of
786 the metasedimentary protoliths due to the intrusion of Carboniferous to Permian granitoids, such
787 as described in the Jebilet, in the Rehamna, or in the neighbouring granitoid of Ment (*e.g.*
788 Dahmani, 1995; Chopin *et al.*, 2014; Delphini *et al.*, 2016). Both regional metamorphism of the
789 protolith and a pre-entrapment contact metamorphism and could have promoted early partial
790 melting and melt extraction from the metasedimentary protolith and explain their refractory
791 nature.

792

793 **11. Conclusion**

794

795 A suite of metapelitic xenoliths enclosed in a micro-quartz diorite dyke from the Moroccan
796 Central Massif records granulite facies metamorphism, with maximal temperatures of 800 °C to
797 900 °C. Peak pressure conditions varying from *ca.* 1.5 to 6 kbar depending on the xenolith.
798 Stratigraphic constrains assign a Permian age for the micro-quartz diorite dyke, whereas U-Th/Pb
799 EPMA dating on monazite in the xenoliths gives a date of 293 ± 25 Ma which, although poorly
800 defined, confirms a late Variscan age for this metamorphism. The xenoliths therefore sample the

801 first 25 km of the Variscan crust. However, the recorded maximal temperatures are systematically
802 higher than the temperatures of the regional metamorphism associated to the Variscan tectono-
803 metamorphic history. It is therefore concluded that the high temperature recorded in the xenoliths
804 results from heating by the hot granodioritic host magmas.

805
806 The chemistry of the metapelitic xenoliths is generally silica-poor, aluminum-rich, which results
807 in peculiar metamorphic quartz-absent assemblages comprising aluminosilicate (andalusite \pm
808 sillimanite), corundum, spinel, cordierite and sometimes garnet. Although partial melt may have
809 escaped the xenolith during entrapment in the magma, it is shown that melt loss cannot explain
810 the refractory conditions observed in the xenoliths. The protoliths must therefore have been
811 preconditioned by regional metamorphism or previous contact metamorphism at high
812 temperature condition by a first depletion due to partial melting and melt loss. The remnants of
813 this initial metamorphic stage can be found in the presence of andalusite, biotite and garnet
814 crystals in some xenoliths.

815

816

817 **Acknowledgements**

818 This project has been founded by CNRS –INSU - SYSTER program, by the University of
819 Toulouse through an international partnership with University Sidi Mohamed Ben Abdellah in
820 Fez, Morocco. We are grateful to the head and members of the departement of geology of the
821 University Sidi Mohamedb Ben Abdellah of Fès for their support and welcome during meetings
822 and field work. Many thanks to Antonio Álvarez-Valero, Ingrid van Namen and Joost van
823 Hoeflaken who participated to this early field work and sampling in Morocco. This publication
824 has been improved by the careful and constructive reviews of Bernardo Cesare and an
825 anonymous reviewer.

826

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1100
1101

1102 **Figure Captions**

1103

1104 **Fig. 1 Geological maps** (a) Tectonic map of Morocco and location of the study area (b)
1105 Geological map of the study area (d'après Ntarmouchant, 2003)

1106 **Fig. 2 Field photographs of the sampling area (Kariane locality)** (a) Field view of the
1107 microgranodioritic dyke (b) Contact between the dyke and the Ordovician metasediments
1108 (c, d) Two examples of crustal xenoliths in the microgranodioritic host rock

1109 **Fig. 3 Photomicrographs showing the assemblages and microstructures within the**
1110 **xenoliths.** (a) *Ma3* Foliation underlined by sillimanite, skeletal biotite and elongated corroded
1111 garnet grains with sillimanite needles underlining and internal schistosity. Note the large
1112 cordierite flake enfolding all other mineral phases. (b) *Ma3* Note the presence of spinel
1113 overprinting the foliation. (c) *Ma10* Large chiascolitic andalusite prisms surrounded by a spinel +
1114 plagioclase corona. (d) *Ma10* Broken enbay crystal partly transformed into sillimanite and
1115 surrounded by a spinel-rich corona; cordierite corona surrounding garnet; plagioclase corona
1116 surrounding spinel. (e) *Ma6* Crystal of andalusite partially replaced by sillimanite and
1117 plagioclase; plagioclase contains spinel and separates andalusite from cordierite (f) *Ma6*
1118 Andalusite pseudomorph including corundum and transformed into plagioclase. (g) *Ma6*
1119 Plagioclase-spinel matrix displaying magmatic texture. (h) *Kar2b* Corundum and spinel within
1120 plagioclase replacing andalusite. (i) *Ma5* Plagioclase pseudomorphose of a crystal of andalusite
1121 (square section is recognizable). (j) *Ma5* Plagioclase euhedral twinned crystals with Spl
1122 inclusions. (k) *Ma5* Corundum rimmed by cordierite and spinel. (l) *Ma5* Corroded garnet grain
1123 rimmed by brown cordierite flakes. Mineral abbreviations from Whitney and Evans (2010).

1124 **Fig. 4 BSE images and EDS maps of the xenoliths.** (a,g) *Ma3*. Note oriented inclusions of
1125 sillimanite and ilmenite in garnet and the K enrichment at alkali feldspar boundaries (b) *Ma10*
1126 Feldspar – cordierite rims at garnet crystal. Note the complex alkali-feldspar / plagioclase
1127 concentric pattern (c) *Ma10* Plagioclase – alkali feldspar association filling microcracks in
1128 andalusite crystal (d) *Ma6* Andalusite crystals resorbed into a plagioclase – spinel association (d)
1129 *Kar2b* Andalusite crystals resorbed into a plagioclase – spinel association (f) *Ma5* Note resorbed
1130 corundum in a plagioclase matrix and the presence of cordierite. Quantitative EDS maps (g, j, k,
1131 l) have been used to quantify local composition for thermodynamical modelling . Mineral
1132 abbreviations from Whitney and Evans (2010).

1133 **Fig. 5 Mineral chemistry** a) Ternary diagram for feldspar with solvus isotherms at 5 kbar from
1134 Elkins and Grove (1990) (b) $Ti+Al^{VI} = f(Fe+Mg)$ diagram for biotite.

1135
1136 **Fig. 6 Photomicrographs of the host rock and contacts with the xenoliths** (a)
1137 *Microgranodioritic host rock*. Note the porphyritic texture with phenocrysts of plagioclase,
1138 amphibole, biotite and quartz in a fine-grained matrix composed of quartz, plagioclase, K-
1139 feldspar and rare biotite. (b) *Ma3* Sharp contact, with a few xenocrysts of biotite from the
1140 xenolith in the host rock. (c) *Ma10* Sharp contact between xenolith and host rock underlined by
1141 a two mm large zone where abundant pinitite replaces cordierite. (d, e) *Ma6* Host rock / xenolith
1142 contact underlined by the presence of large andalusite crystals sometimes fully enclosed in the
1143 host rock. (f) *Ma5* Host rock / xenolith contact marked a localized (<1 mm large) chloritized
1144 band. Mineral abbreviations from Whitney and Evans (2010).

1145 **Fig. 7 Major element characteristics of xenoliths and host rocks** (a) Harker diagrams showing
1146 Al_2O_3 , Na_2O+K_2O and TiO_2 variation as a function of SiO_2 (b) Cox classification diagram for
1147 plutonic rocks (Cox *et al.* 1979), plotted with GCDKit software (Janoušek *et al.*, 2006). Note that
1148 sample Kar2b corresponds to the host rock of the corresponding studied xenolith.

1149 **Fig. 8 Trace element characteristics of xenoliths and host rocks** (a) Harker diagrams showing
1150 selected transition element variation as a function of SiO_2 . (b) La/Yb as a function of SiO_2 .

1151 **Fig. 9 Spidergrams for xenoliths and host rocks** (a) REE elements normalized to chondrite
1152 (Nakamura *et al.*, 1974) (b) Trace element spectra normalized to primitive mantle (McDonough
1153 and Sun, 1995). Diagrams plotted with GCDKit software (Janoušek *et al.*, 2006)

1154 **Fig. 10 Pressure-temperature pseudosection for sample *Ma3***. Composition in weight percent
1155 from EDS mapping of the microdomain shown in Fig. 4g. Also shown are the relevant isopleths
1156 for plagioclase (X_{Ab}), garnet (X_{Alm}) and biotite (X_{Mg} , Ti-Bio pfu) corresponding to the observed
1157 compositions (Tables 2 and 4), and the upper stability of cordierite. Pressure-temperature field
1158 corresponding to the paragenesis observed in the microdomain is underlined in yellow. The
1159 suggested prograde and retrograde paths are indicated by white dashed arrows. Mineral
1160 abbreviations from Whitney and Evans (2010).

1161 **Fig. 11 Pressure-temperature pseudosection for sample *Ma10***. Composition in weight percent
1162 from whole-rock composition (Table 5). Also shown are the relevant isopleths for plagioclase

1163 (X_{Ab}), garnet (X_{Alm} , X_{Grs} , X_{Prp}) and biotite (X_{Mg}^{Bt}) corresponding to the observed compositions
1164 (Tables 2 and 4), and the upper stability of cordierite. Pressure-temperature field corresponding to
1165 the paragenesis observed in the microdomain is underlined in yellow. The suggested prograde
1166 path is indicated by a white dashed arrow. Mineral abbreviations from Whitney and Evans
1167 (2010).

1168 **Fig. 12 Pressure-temperature pseudosection for sample *Ma6*.** Composition in weight percent
1169 from EDS mapping of the microdomain shown in Fig. 4j. Also shown are the lower stability of
1170 garnet, the upper stability of K-feldspar and the solidus. Pressure-temperature field corresponding
1171 to the initial and final parageneses observed in the microdomain is underlined in yellow. Mineral
1172 abbreviations from Whitney and Evans (2010).

1173 **Fig. 13 Pressure-temperature pseudosection for sample *Kar2b*.** Composition in weight percent
1174 from EDS mapping of the microdomain shown in Fig. 4k. Also shown are the lower stability of
1175 garnet, the upper stability of K-feldspar and the solidus. Pressure-temperature field corresponding
1176 to the initial and final parageneses observed in the microdomain is underlined in yellow. Mineral
1177 abbreviations from Whitney and Evans (2010).

1178 **Fig. 14 Pressure-temperature pseudosection for sample *Ma5*.** Composition in weight percent
1179 from EDS mapping of the microdomain shown in Fig. 4l. Also shown is the lower stability of
1180 garnet. Pressure-temperature field corresponding to the assemblage observed in the microdomain
1181 is underlined in yellow. Mineral abbreviations from Whitney and Evans (2010).

1182 **Fig. 15. U-Th-Pb chemical ages of monazite for sample *Ma3*.**

1183 **Fig. 16 Evolution of modal proportions of minerals from thermodynamic modeling.**
1184 Cumulated modal proportion as a function of temperature for sample *Ma3* at 6 kbar (a) and 2.5
1185 kbar (b) and for sample *Ma5* at 3 kbar (c). (d) Modal proportion of corundum as a fonction of T
1186 and $X(H_2O)$ in sample *Ma5*, with $X=1$ corresponding to 0.56 wt.% H_2O .

1187 **Fig. 17 Summary of pressure and temperature estimates for xenoliths of the Kariane**
1188 **locality.** Pressure and temperature conditions for amphibolite to granulite facies Variscan
1189 metamorphism in Morocco estimated from previous work in regional metamorphism, contact
1190 metamorphism and in xenoliths from other localities are also shown. Position of the triple point
1191 for Al_2SiO_5 according to (HP) Holland and Powell (1998) or (P) Pattison (1992).

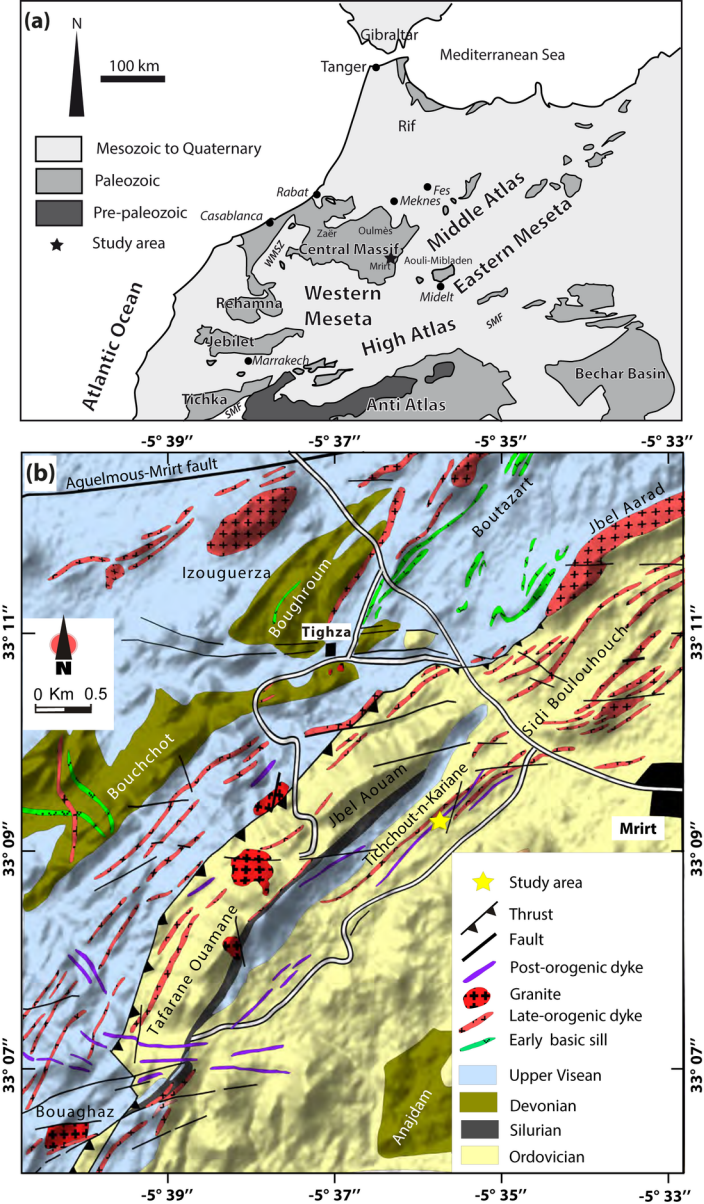


Figure 1

2-column fitting image

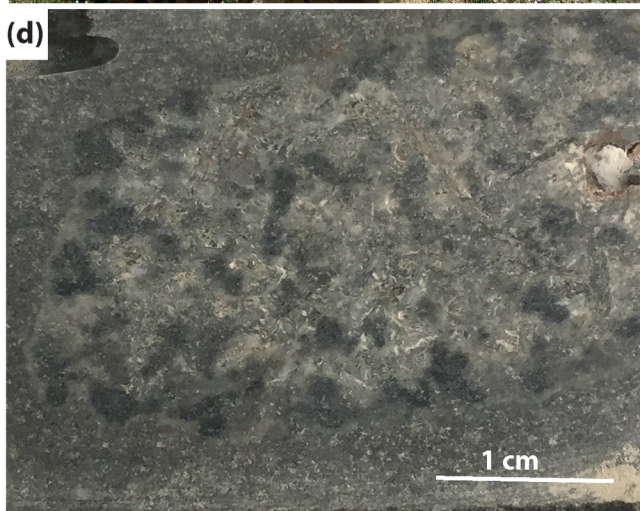
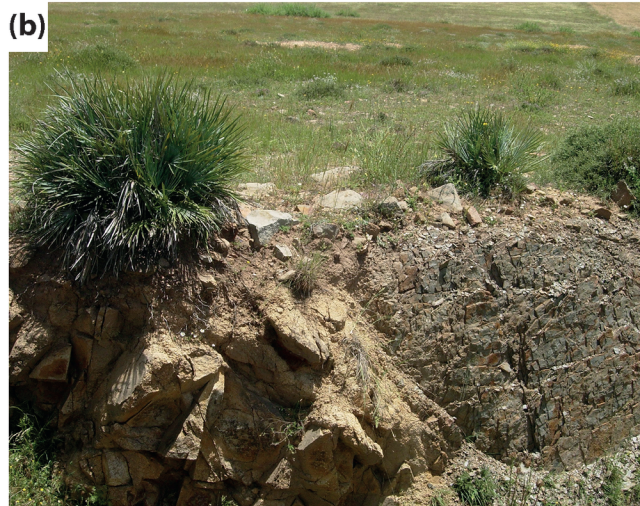


Figure 2

2-column fitting image

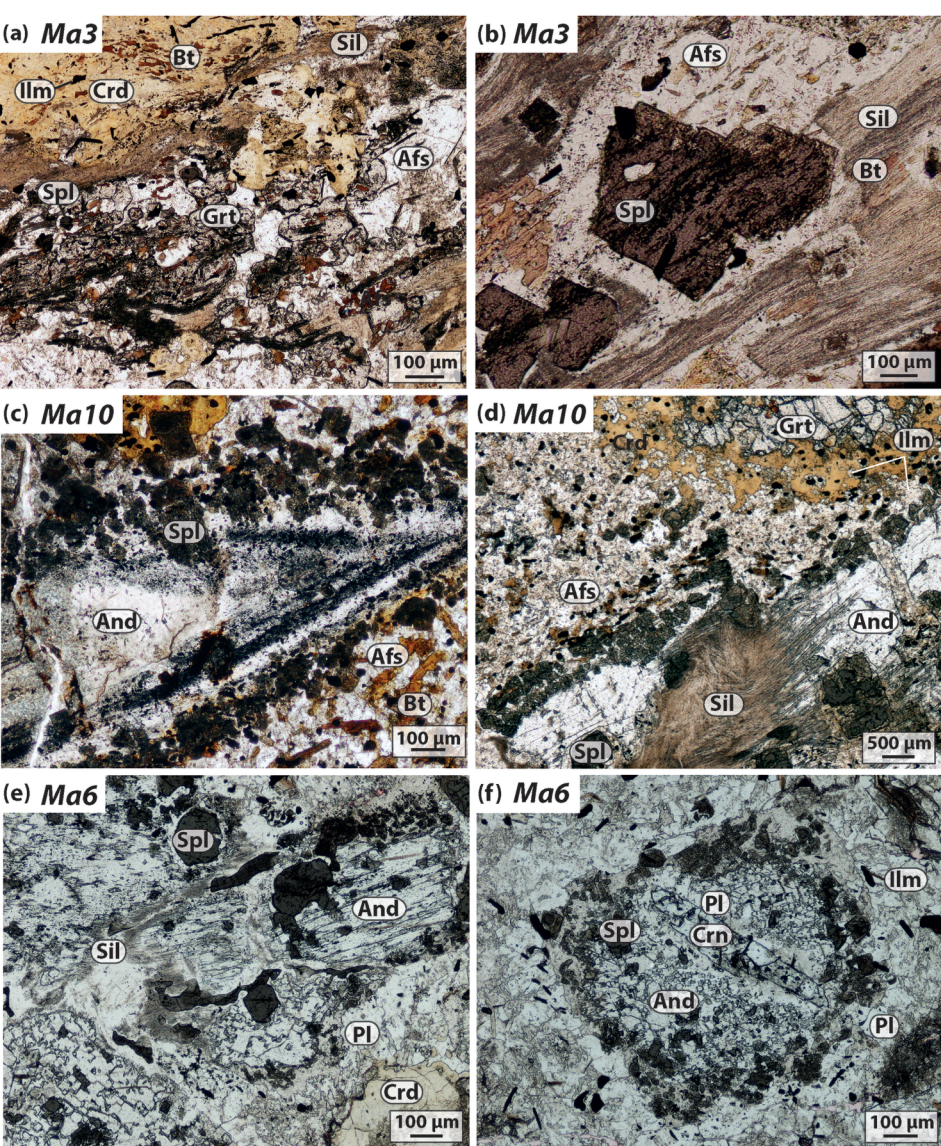
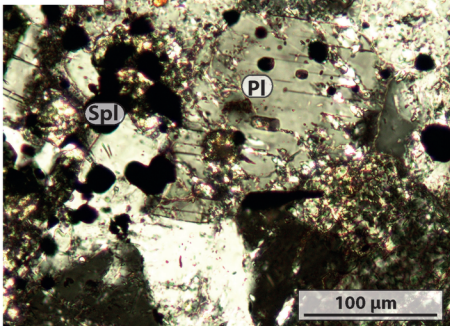
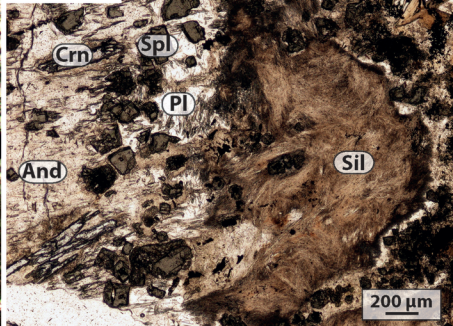


Figure 3

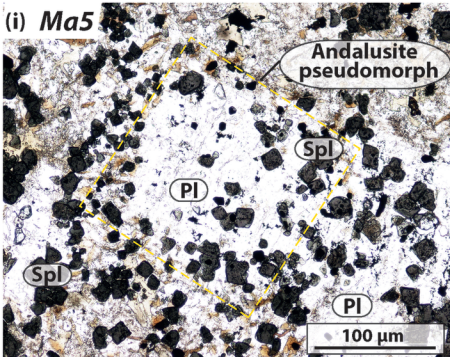
(g) *Ma6*



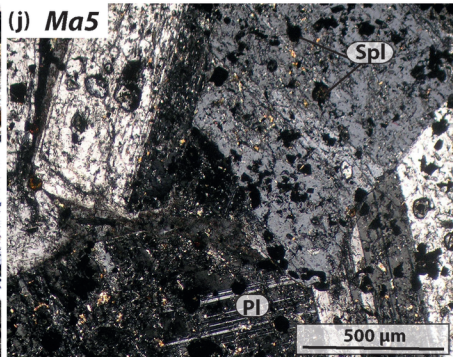
(h) *Kar2b*



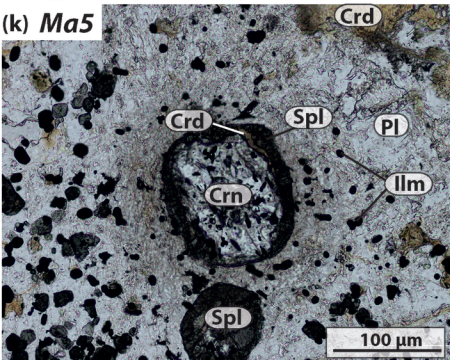
(i) *Ma5*



(j) *Ma5*



(k) *Ma5*



(l) *Ma5*

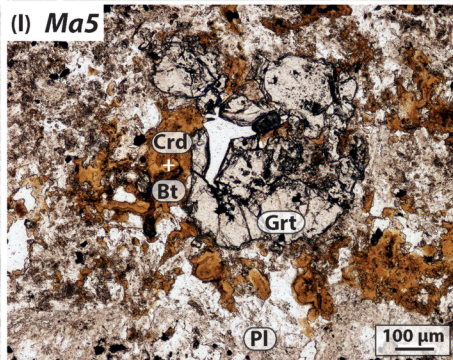


Figure 3 (continue)

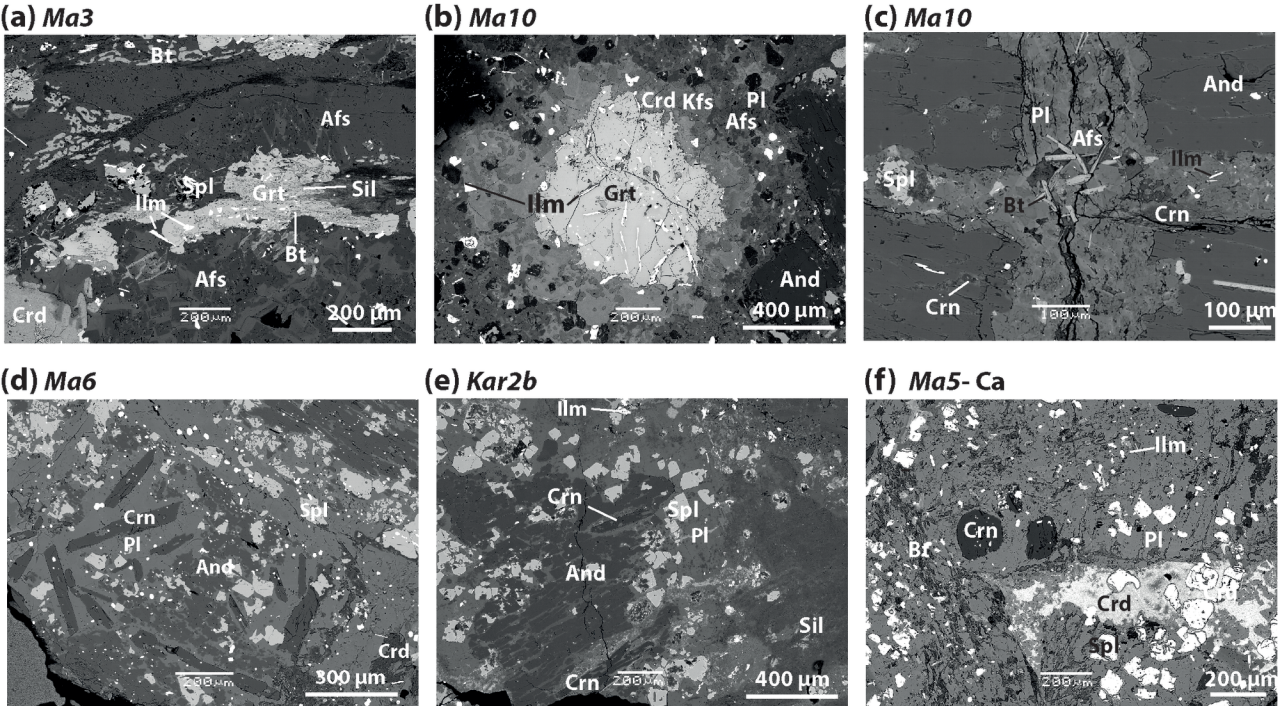


Figure 4

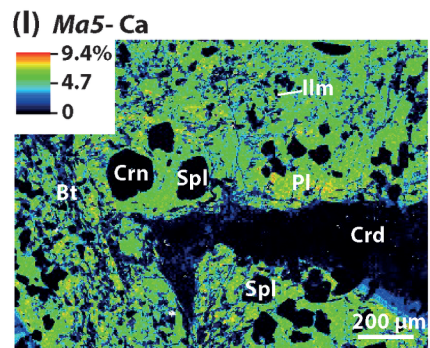
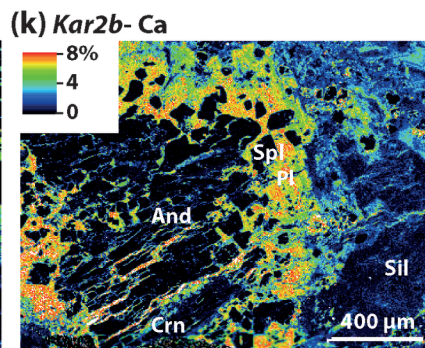
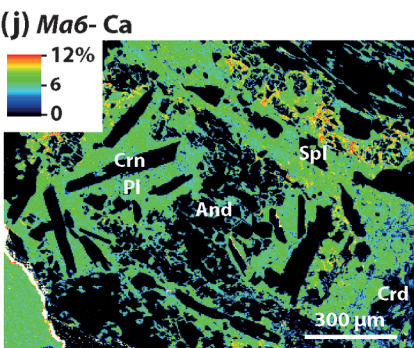
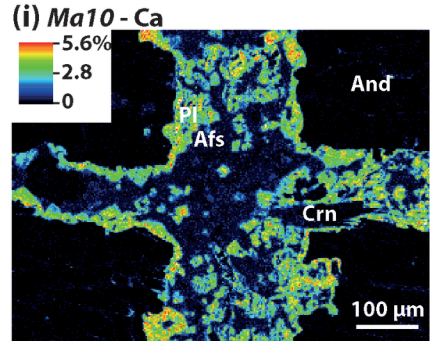
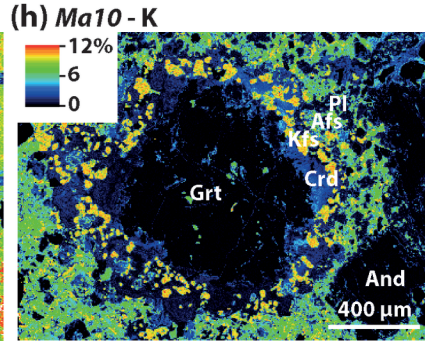
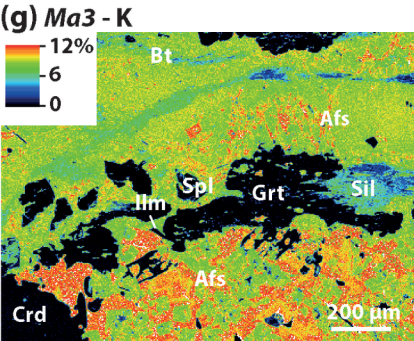


Figure 4 (continue)

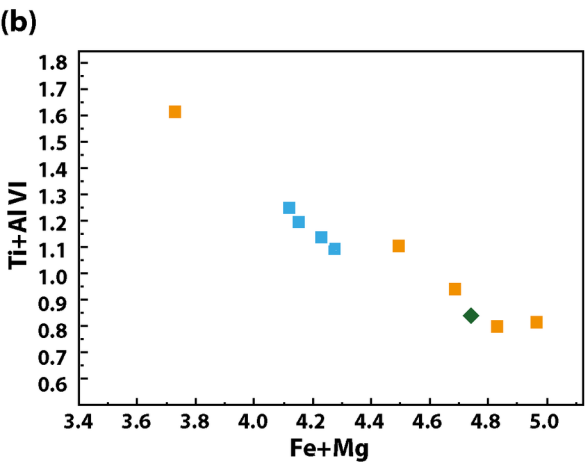
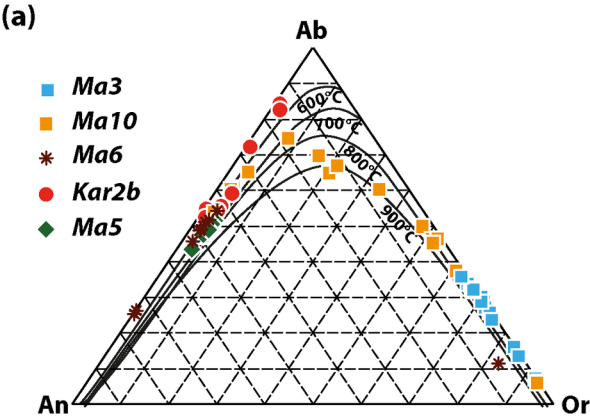


Figure 5

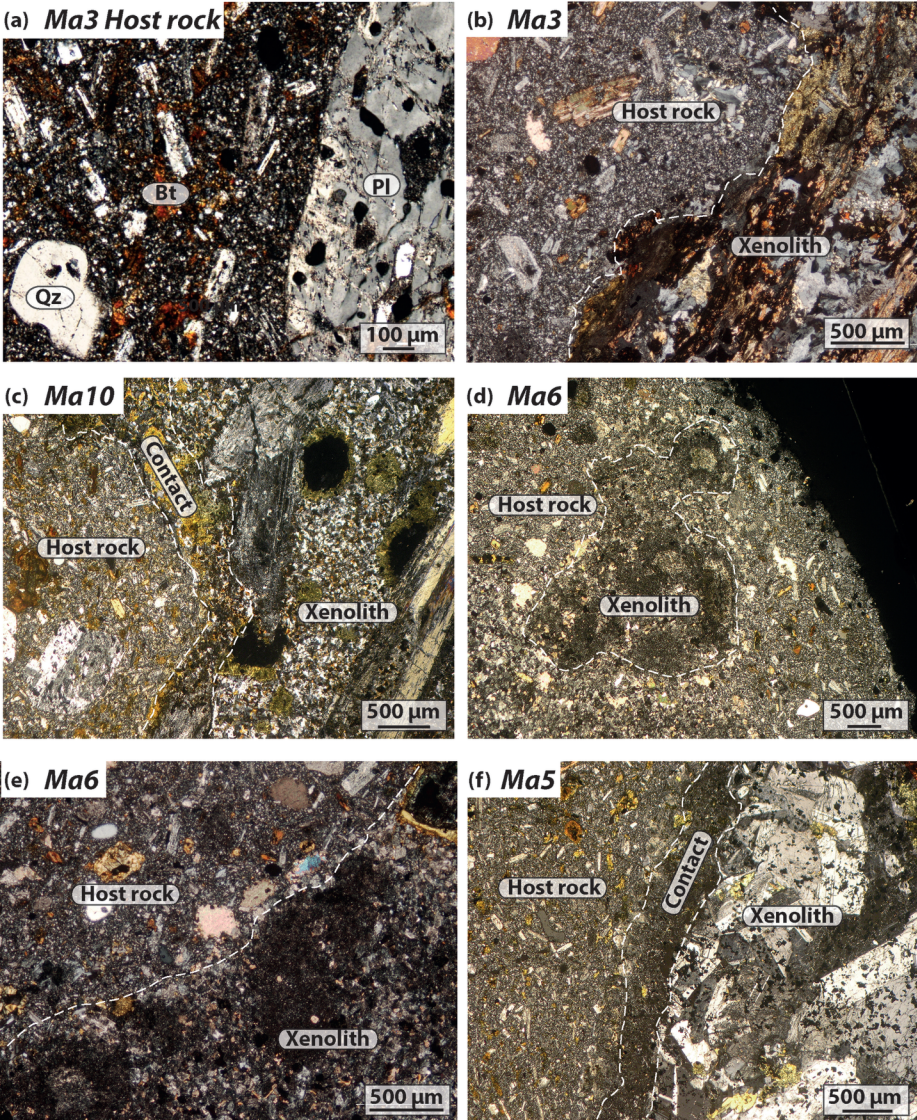


Figure 6

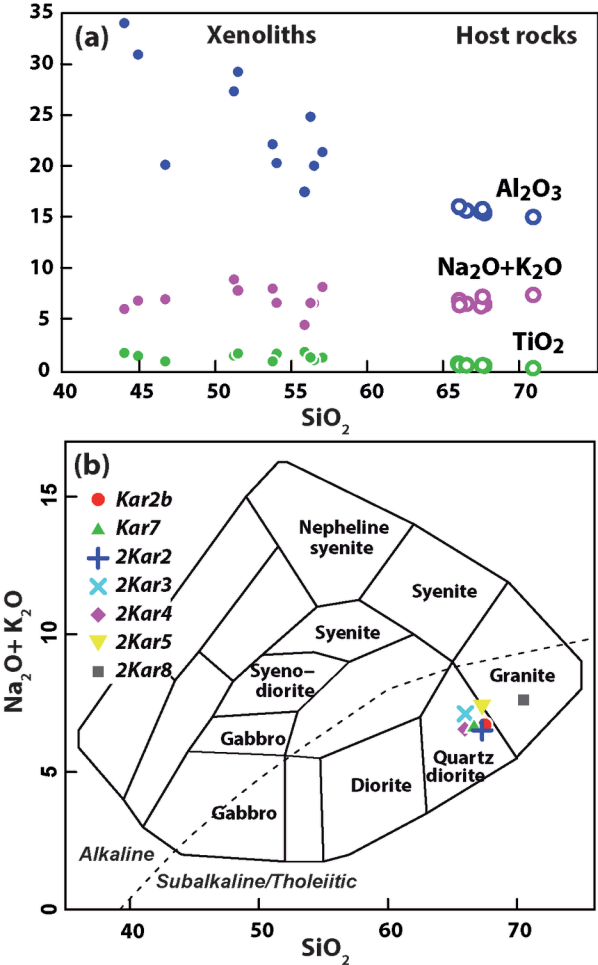


Figure 7

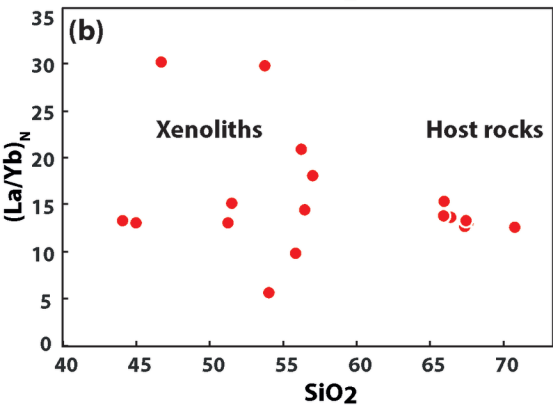
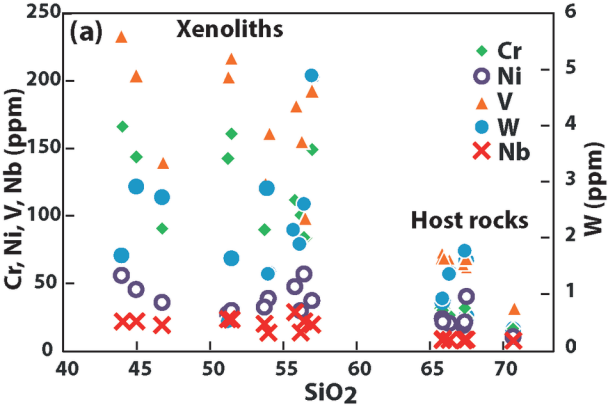


Figure 8

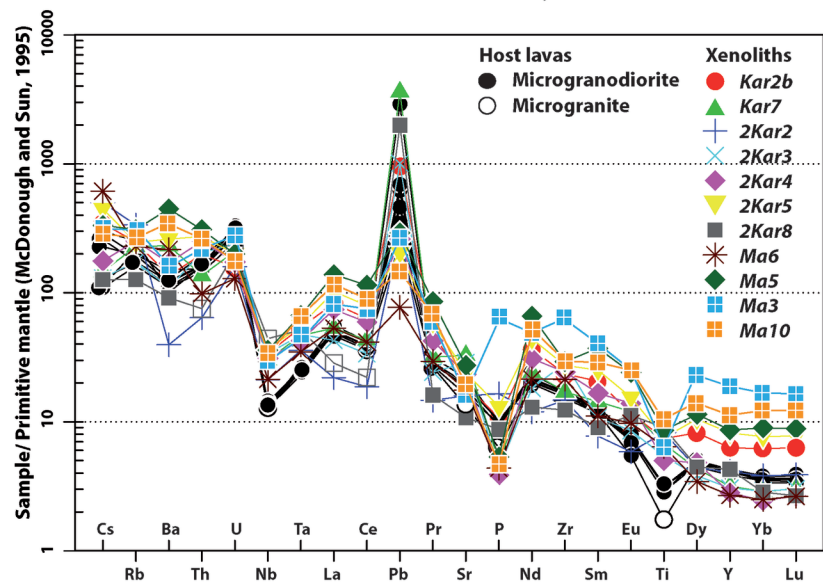
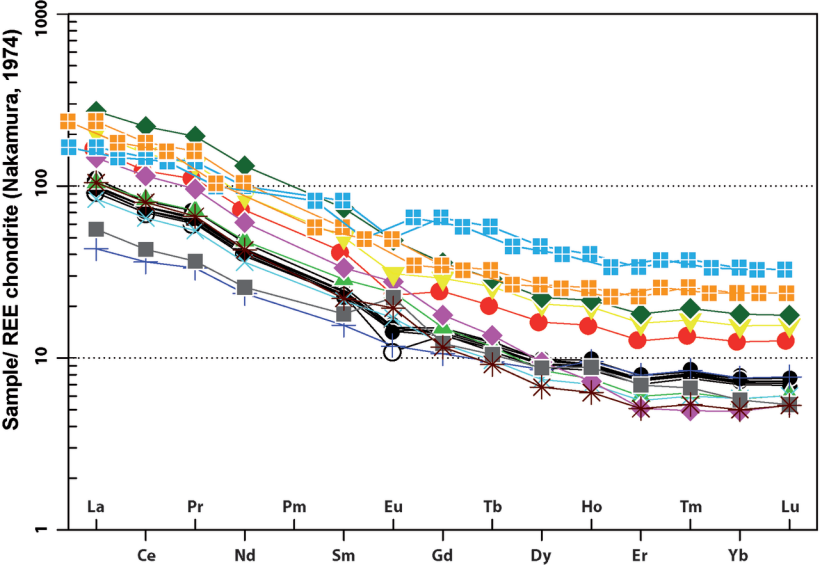
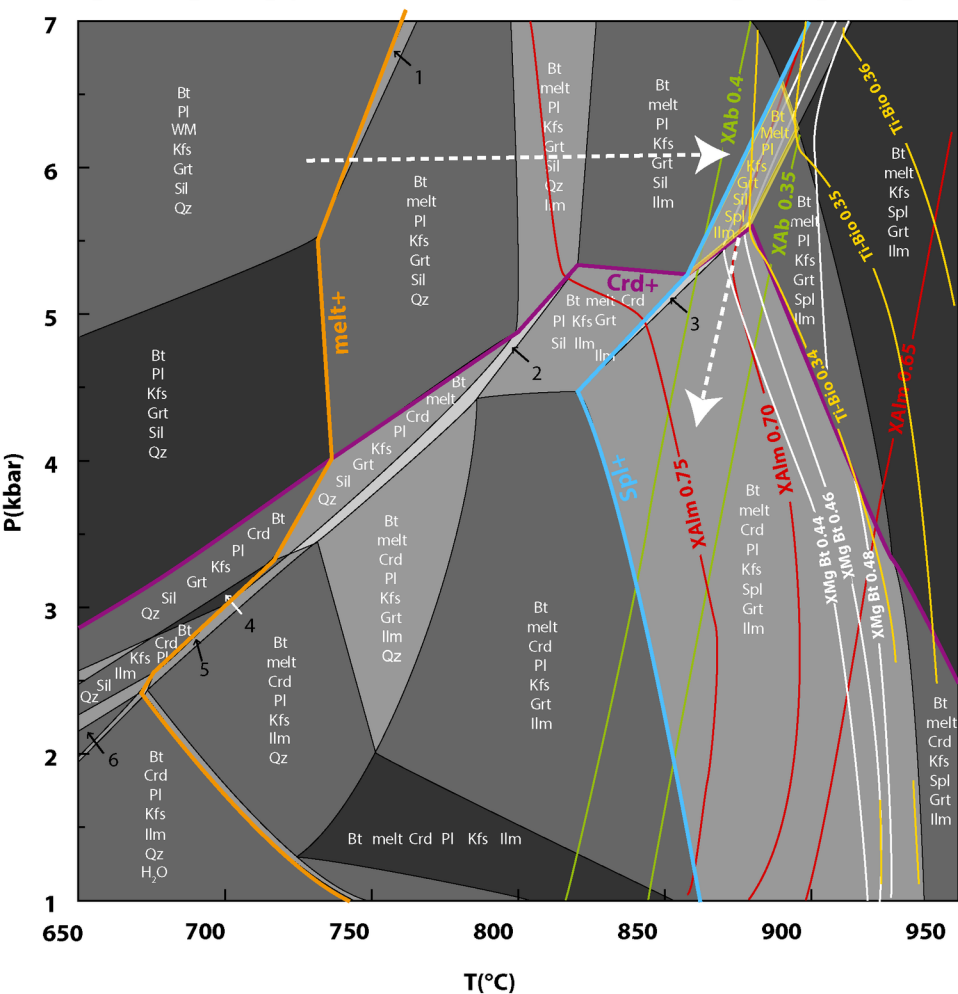


Figure 9

Sample Ma3 (Microdomain)

SiO₂ 55.40, TiO₂ 1.22, Al₂O₃ 21.07, FeO 7.85, MnO 0.14, MgO 1.85, CaO 0.41, Na₂O 1.64, K₂O 9.46, H₂O 0.95



Additional fields

1. Bt melt WM Pl Kfs Grt Sil Qz
 4. Bt Crd Pl Kfs Ilm Sil Qz

2. Bt melt Crd Pl Kfs Grt Sil Ilm

3. Bt melt Crd Pl Kfs Spl Grt Sil Ilm
 5. Bt melt Crd Pl Kfs Ilm Sil Qz
 6. Bt WM Crd Pl Kfs Ilm Qz

Figure 10

Sample Ma19 (Whole-rock)

SiO₂ 46.20; TiO₂ 2.08; Al₂O₃ 35.47; FeO 6.50; MnO 0.13; MgO 1.81; CaO 0.70; Na₂O 2.03; K₂O 4.52; H₂O 0.54

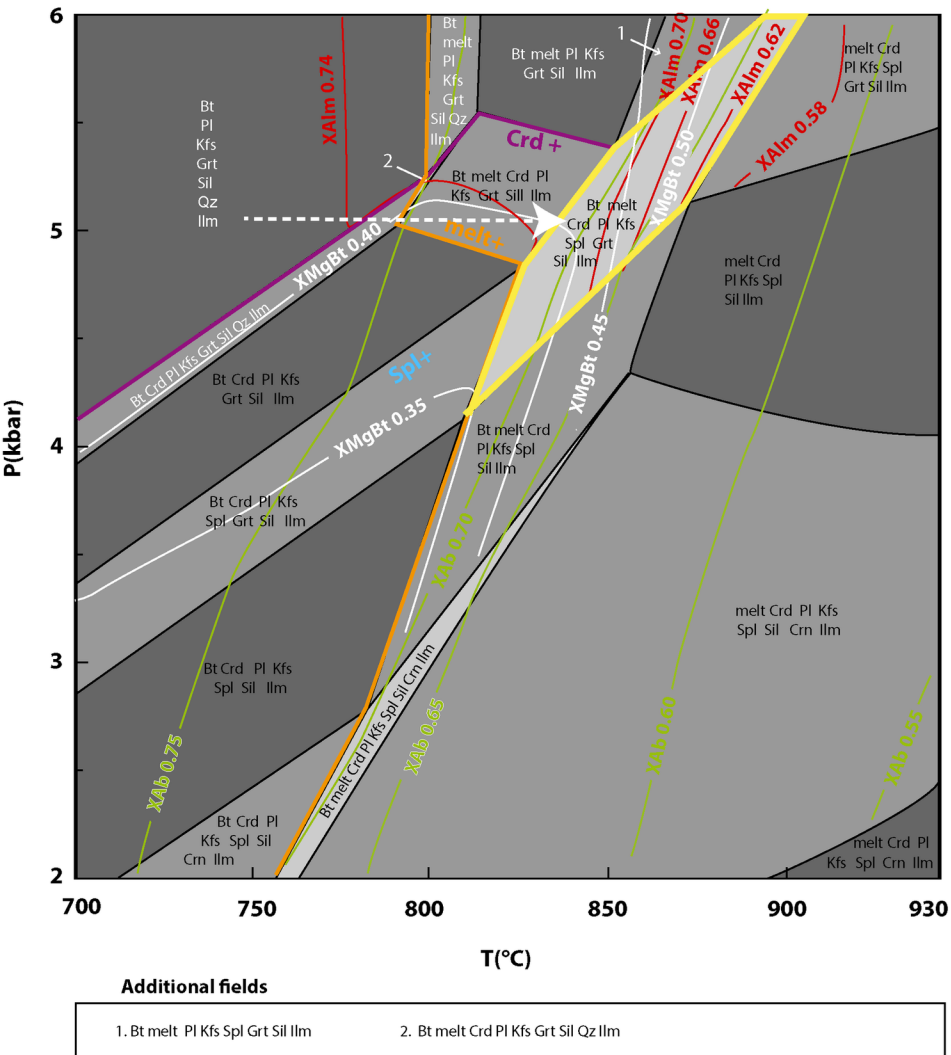


Figure 11

Sample Ma6 (Microdomain)

SiO₂ 38.72; TiO₂ 1.06; Al₂O₃ 45.69; FeO 5.25; MnO 0.06; MgO 1.06; CaO 5.14; Na₂O 2.37; K₂O 0.65; H₂O 0.1

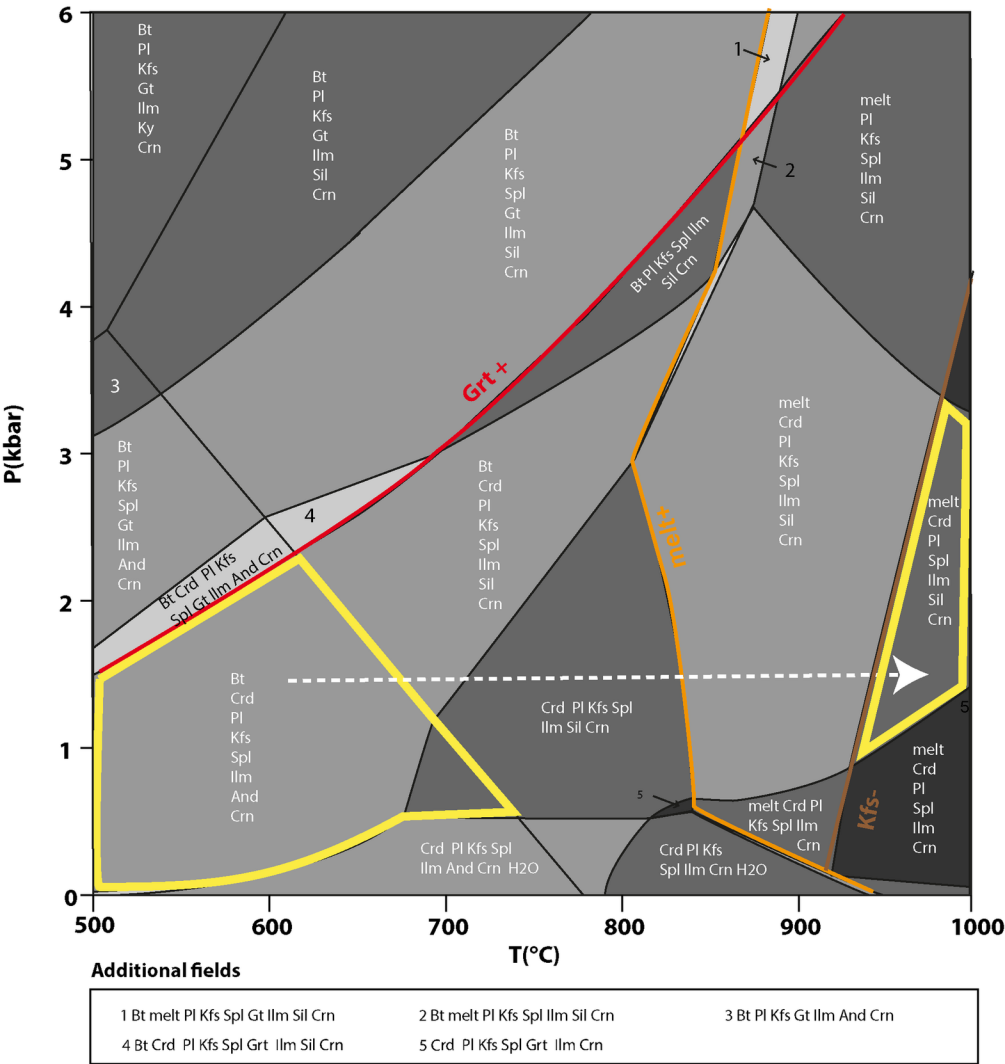


Figure 12

Sample Kar2b (Microcrone)

SiO₂ 35.76; TiO₂ 0.32; Al₂O₃ 49.11; FeO 6.46; MnO 0.07; MgO 1.75; CaO 3.17; Na₂O 1.71; K₂O 0.19; H₂O 0.14

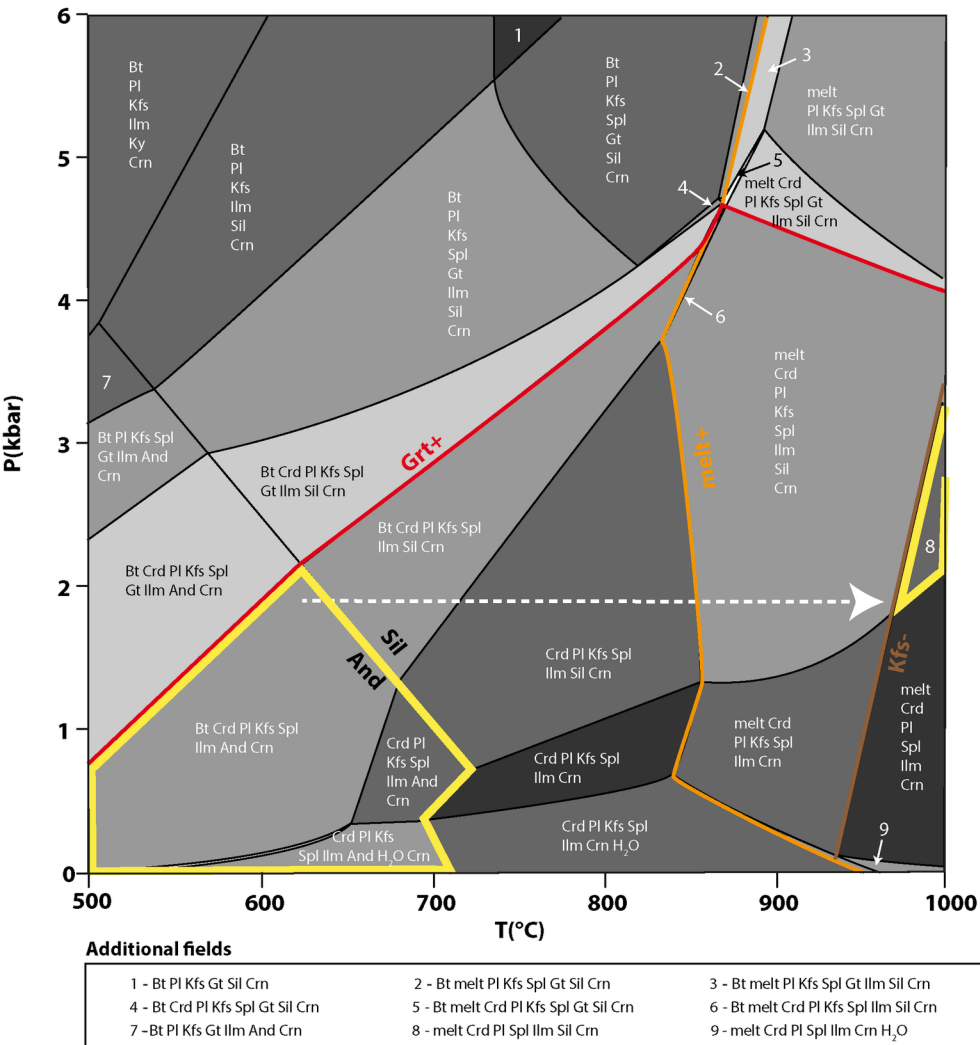


Figure 13

Sample Ma5 (Microdomain)

SiO₂ 46.84; TiO₂ 1.44; Al₂O₃ 32.11; FeO 5.73; MnO 0.10; MgO 2.18; CaO 5.34; Na₂O 3.95; K₂O 1.77; H₂O 0.53

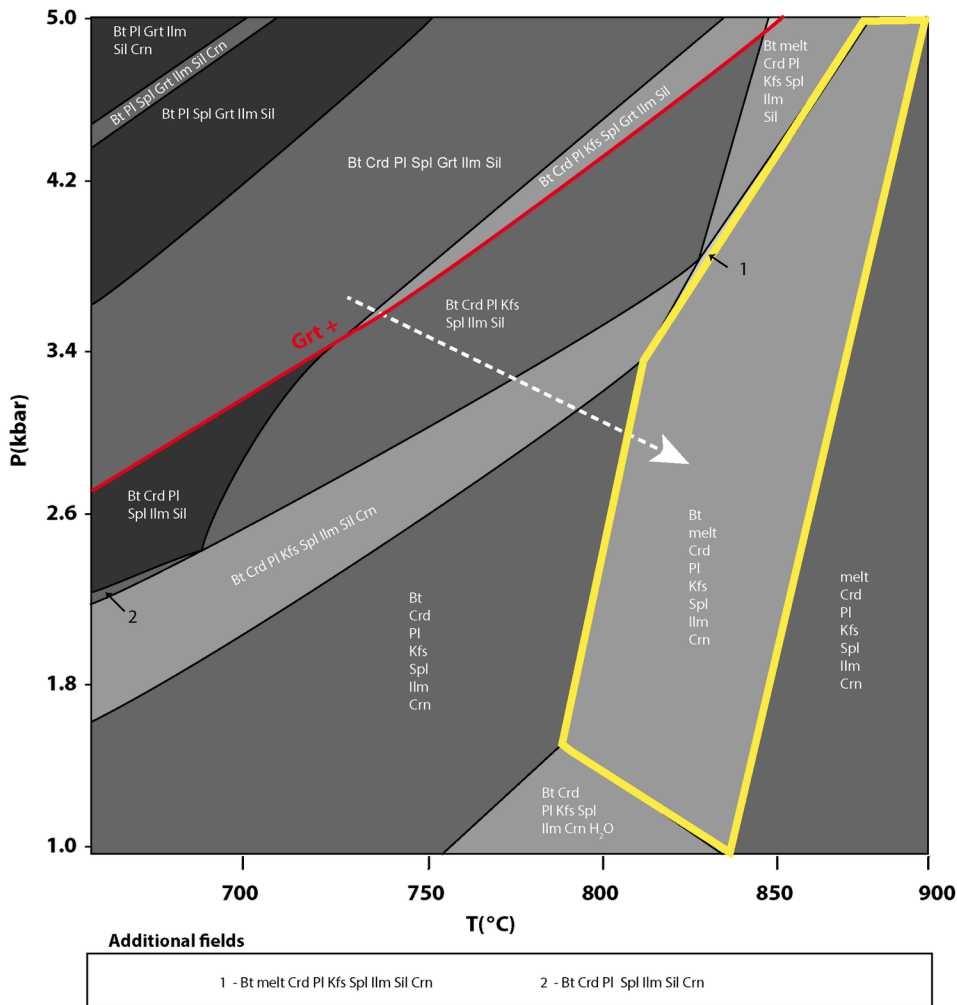


Figure 14

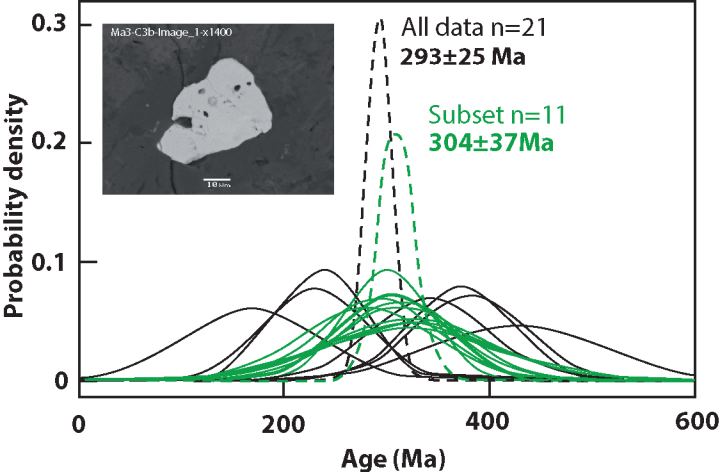


Figure 15

Single column fitting image

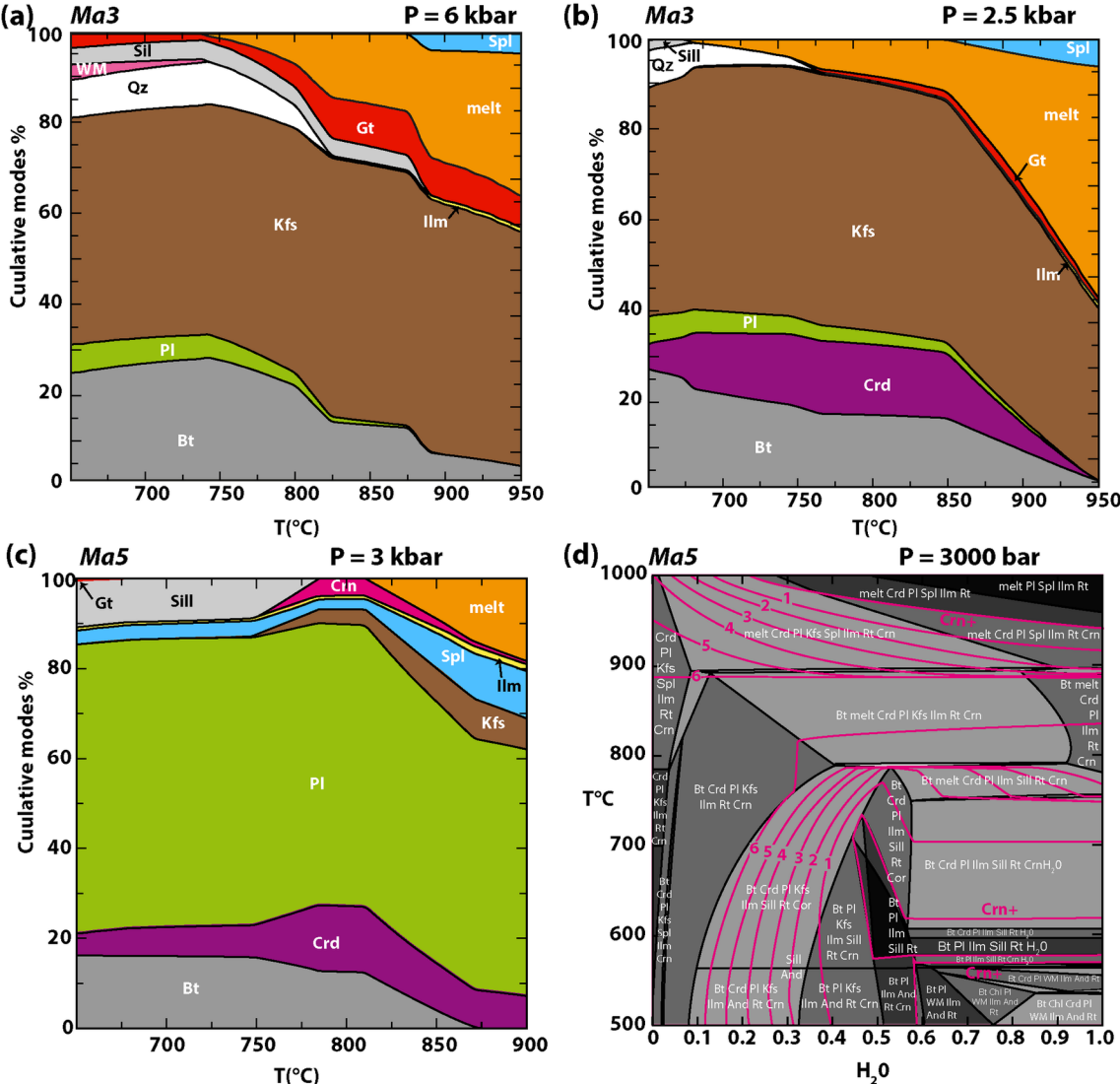
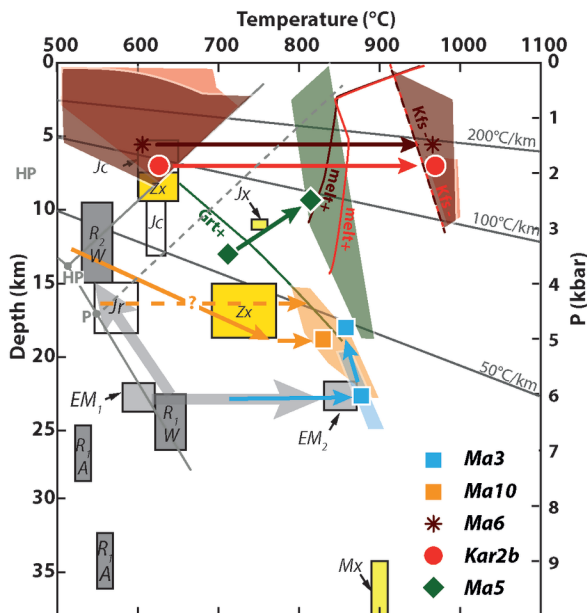


Figure 16

2-column fitting image



- Rehamna.*
Metamorphic condition : peak pressure (1) and exhumation (2) from Aghzer et Arenas (1998) (A) and Wernert et al. (2016) (W).
- Jebilet.*
Regional (r) or contact (c) metamorphic conditions from Delchini et al. (2016)
- Eastern Moulouya.*
Regional metamorphic conditions (1) and peak temperature due to magma injection (2) from Elabouyi et al. (2019)
- Zaër pluton.*
Metamorphic conditions in xenoliths, from Haimeur et al. (2003)
- Oulad Ouaslam granite (Jebilet).*
Metamorphic conditions in xenoliths, from Bouloton et al. (1991)
- Quaternary basalts (Middle Atlas).*
Granulitic xenoliths, from Moukadiri and Bouloton (1998)

Figure 17

Table 1

Summary of petrological description of xenoliths

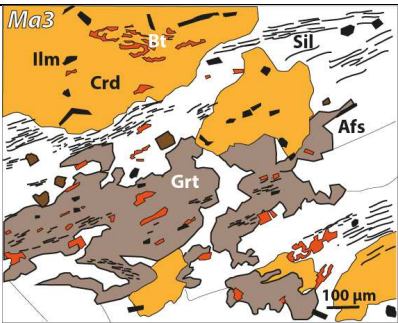
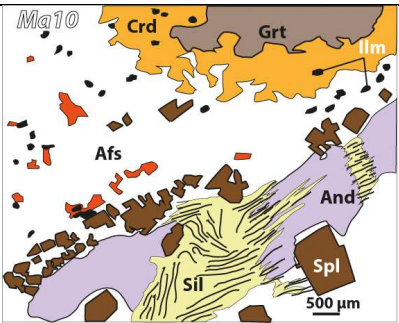
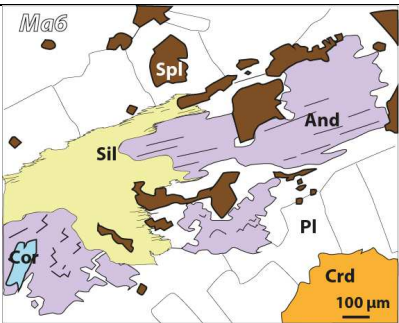
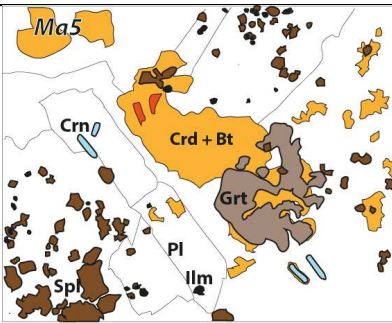
	<i>Ma3</i>	<i>Ma10</i>	<i>Ma6 and Kar2b</i>	<i>Ma 5</i>
				
Xenolith chemistry	SiO ₂ 46.76 % Al ₂ O ₃ 20.26 % ΣREE 1317 ppm	SiO ₂ 44.07 % Al ₂ O ₃ 33.84 % ΣREE 1297 ppm	SiO ₂ 56.24 % (<i>Ma6</i>) 44.99 % (<i>Kar2b</i>) Al ₂ O ₃ 24.91 % (<i>Ma6</i>) 30.99 % (<i>Kar2b</i>) ΣREE 522 ppm (<i>Ma6</i>) 873 ppm (<i>Kar2b</i>)	SiO ₂ 44.99 % Al ₂ O ₃ 30.99 % ΣREE 1495 ppm
Pre-entrapment metamorphism remnants	Biotite, sillimanite, garnet	Andalusite, garnet	Andalusite, corundum	Garnet, corundum, biotite, ilmenite
Contact metamorphism	Garnet, spinel, alkali feldspar, biotite, ilmenite, melt Late cordierite, K-feldspar	Plagioclase, alkali feldspar, biotite, spinel, ilmenite, sillimanite, cordierite, melt	Spinel, plagioclase, biotite, sillimanite, cordierite, melt	Spinel, cordierite, corundum, ilmenite, plagioclase, Afs, melt
Mineral chemistry	Afs Or 62-90 Spl X _{Fe} 0.67-0.72 Grt X _{Alm} 0.64-0.73 Bt X _{Mg} 0.45 TiO ₂ 6-8%	Pl An 35-45 Afs Or 47-92 Spl X _{Fe} 0.63-0.79 Grt X _{Alm} 0.59-0.67 Bt X _{Mg} 0.4-0.56 TiO ₂ 3-7%	Pl An26-74 (<i>Ma6</i>) An15-45 (<i>Kar2b</i>) Afs Or 57-88 Spl X _{Fe} 0.64-0.69 (<i>Ma6</i>) X _{Fe} 0.70-0.74 (<i>Kar2b</i>)	Pl An 45-53 Spl X _{Fe} 0.57-0.64 Bt X _{Mg} 0.49-0.56 TiO ₂ < 5%
Contact <i>P-T</i> conditions	<i>P</i> > 5 kbar <i>T</i> 850-900°C	<i>P</i> 4-6 kbar <i>T</i> 850 °C	<i>P</i> < 2 kbar <i>T</i> > 900 °C	<i>P</i> < 5 kbar <i>T</i> 800-850 °C

Table 2

Representative electron microprobe analyses of feldspars in xenoliths

	<i>Ma3</i>	<i>Ma3</i>	<i>Ma3</i>	<i>Ma10</i>	<i>Ma10</i>	<i>Ma10</i>	<i>Ma6</i>	<i>Ma6</i>	<i>Ma6</i>	<i>Kar2b</i>	<i>Kar2b</i>	<i>Kar2b</i>	<i>Ma5</i>	<i>Ma5</i>	<i>Ma5</i>
SiO ₂	66.03	64.78	64.52	55.72	63.49	65.32	54.02	65.8	59.23	65.9	57.77	62.1	57.12	55.09	55.97
TiO ₂	0.00	0.28	0.06	0.03	0.11	0.00	0.00	0.07	0.00	0.09	0.18	0.14	0.03	0.06	0.02
Al ₂ O ₃	18.8	18.9	18.87	28.71	19.21	22.10	29.25	18.37	25.49	22.13	26.6	24.39	27.00	28.62	27.89
Fe ₂ O ₃	0.1	0.11	0.02	0.25	1.56	0.00	0.25	0.06	0.22	0.38	0.50	0.13	0.12	0.18	0.10
CaO	0.31	0.00	0.16	9.24	0.07	3.03	11.44	0.2	5.39	3.28	8.93	5.83	9.23	11.05	10.13
Na ₂ O	4.2	0.74	3.24	6.31	0.61	8.98	4.70	2.89	7.71	10.24	6.26	8.53	5.93	5.02	5.52
K ₂ O	10.93	15.32	11.73	0.46	14.35	1.21	0.34	12.55	1.02	0.20	0.57	0.19	0.65	0.49	0.59
Total	100.36	102.32	98.62	100.73	100.31	100.64	100.01	99.94	99.19	102.23	100.9	101.4	100.08	100.50	100.23
Si	2.99	2.99	3.03	2.58	2.96	3.01	2.51	3.05	2.79	3.01	2.67	2.86	2.65	2.55	2.59
Al	1.00	1.03	1.05	1.57	1.06	1.20	1.60	1.00	1.41	1.19	1.45	1.32	1.48	1.56	1.52
Ti	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00
Fe ²⁺	0.00	0.00	0.00	0.01	0.05	0.00	0.01	0.00	0.01	0.01	0.02	0.00	0.00	0.01	0.00
Ca	0.02	0.00	0.01	0.46	0.00	0.15	0.57	0.01	0.27	0.16	0.44	0.29	0.46	0.55	0.50
Na	0.37	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K	0.63	0.90	0.70	0.03	0.85	0.07	0.02	0.74	0.06	0.01	0.03	0.01	0.04	0.03	0.03
Ab	36.32	6.84	29.33	53.84	6.05	78.42	41.79	25.67	67.87	84.04	54.11	71.82	51.75	43.85	47.97
An	1.48	0.00	0.80	43.57	0.38	14.62	56.22	0.98	26.22	14.88	42.65	27.13	44.52	53.34	48.65
Or	62.20	93.16	69.87	2.58	93.57	6.95	1.99	73.35	5.91	1.08	3.24	1.05	3.72	2.82	3.37

Chemical formulae were calculated on the basis of 8 oxygens

Table 3

Representative electron microprobe analyses of spinel in xenoliths

	<i>Ma3</i>	<i>Ma3</i>	<i>Ma10</i>	<i>Ma10</i>	<i>Ma6</i>	<i>Ma6</i>	<i>Kar2b</i>	<i>Kar2b</i>	<i>Ma5</i>	<i>Ma5</i>
SiO ₂	0.13	0.05	0.08	0.21	0.22	0.15	0.19	0.03	0.04	0.19
TiO ₂	0.21	0.26	0.34	0.36	0.14	0.19	0.32	0.37	0.21	0.42
Al ₂ O ₃	60.01	59.23	61.14	62.47	59.38	60.64	60.43	60.57	60.51	60.84
Cr ₂ O ₃	0.04	0.13	0.07	0.17	0.04	0.06	0.03	0.04	0.17	0.09
Fe ₂ O ₃	1.00	2.51	1.05	0.00	0.92	0.00	0.93	0.96	1.48	0.21
FeO	29.31	29.04	27.84	29.59	31.52	29.59	30.08	27.87	24.66	27.98
MnO	0.00	0.04	0.25	0.24	0.27	0.34	0.53	0.40	0.20	0.23
MgO	7.72	7.88	9.05	8.47	6.32	7.22	7.43	8.76	10.56	8.80
ZnO	0.00	0.00	0.11	0.15	0.00	0.53	0.22	0.00	0.07	0.33
Total	98.42	99.14	99.93	101.66	98.81	98.72	100.15	99.00	97.90	99.09
Si	0.00	0.00	0.00	0.01	0.01	0.00	0.01	0.00	0.01	0.00
Ti	0.01	0.01	0.01	0.01	0.00	0.00	0.01	0.01	0.01	0.01
Al VI	1.97	1.93	1.96	1.97	1.96	1.97	1.96	1.96	1.96	1.96
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ³⁺	0.02	0.05	0.02	0.00	0.02	0.01	0.02	0.02	0.02	0.02
Fe ²⁺	0.68	0.67	0.63	0.66	0.74	0.69	0.69	0.64	0.69	0.64
Mn ²⁺	0.00	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Mg	0.32	0.33	0.37	0.34	0.26	0.30	0.30	0.36	0.30	0.36
Zn	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00
XFe ²⁺	0.68	0.67	0.63	0.66	0.74	0.70	0.69	0.64	0.69	0.64

Chemical formulae were calculated on the basis of 4 oxygens

Table 4

Representative electron microprobe analyses of garnet and biotite in xenoliths

	Garnet				Biotite							
	<i>Ma3</i>	<i>Ma3</i>	<i>Ma10</i>	<i>Ma10</i>	<i>Ma3</i>	<i>Ma3</i>	<i>Ma3</i>	<i>Ma10</i>	<i>Ma5</i>	<i>Ma5</i>	<i>Ma5</i>	
SiO ₂	39.00	37.45	37.94	37.23	SiO ₂	34.62	36.23	34.67	36.05	35.4	35.17	36.17
TiO ₂	0.21	0.18	0.14	0.39	TiO ₂	7.98	6.21	8.14	4.29	5.15	3.77	4.91
Al ₂ O ₃	21.98	20.88	22.4	21.04	Al ₂ O ₃	17.33	16.18	16.71	16.81	15.56	17.15	15.71
Cr ₂ O ₃	0.09	0.00	0.00	0.00	Cr ₂ O ₃	0.01	0.18	0.3	0.06	0	0.02	0.08
Fe ₂ O ₃ (c)	0.19	0.91	1.49	1.06	FeO	16.87	18.45	17.68	17.23	18.81	17.35	18.13
FeO	30.37	32.61	26.85	31.05	MnO	0.07	0.09	0	0.21	0.02	0	0.1
MnO	0.57	0.35	1.60	2.69	MgO	8.8	8.6	8.89	11.12	10.19	12.28	11.34
MgO	8.62	6.14	8.57	5.86	CaO	0	0.03	0.06	0.02	0	0.1	0.15
CaO	0.4	0.81	1.36	1.32	Na ₂ O	0.37	0.45	0.44	0.47	0.56	0.5	0.43
Total	101.42	99.33	100.35	99.66	K ₂ O	8.78	8.98	8.71	8.28	8.99	8.27	8.41
					BaO	0.11	0.18	0.23	0.47	0	0.6	0.35
Si	2.99	2.98	2.93	2.96	NiO	0.04	0	0.06	0	0.04	0.08	0
Ti	0.01	0.01	0.01	0.02	F	1.24	1.28	1.76	1.22	2.64	1.93	2.02
Al IV	0.01	0.02	0.07	0.04	Cl	0.15	0.25	0.16	0.28	0.78	0.21	0.65
Al VI	1.97	1.94	1.97	1.93	Total	96.37	97.11	97.81	96.51	98.14	97.43	98.45
Cr	0.01	0.00	0.00	0.00								
Fe ³⁺	0.01	0.05	0.09	0.06	Si	2.585	2.703	2.566	2.686	2.629	2.604	2.660
Fe ²⁺	1.95	2.17	1.73	2.00	Ti	0.448	0.349	0.453	0.240	0.288	0.210	0.272
Mn ²⁺	0.04	0.02	0.11	0.18	Al IV	1.415	1.297	1.434	1.314	1.362	1.396	1.340
Mg	0.98	0.73	0.99	0.69	Al VI	0.110	0.125	0.024	0.163	0.000	0.101	0.022
Ca	0.03	0.07	0.11	0.11	Fe ²⁺	1.054	1.151	1.094	1.074	1.168	1.074	1.115
					Mn ²⁺	0.004	0.006	0.000	0.013	0.001	0.000	0.006
Pyrope	32.80	24.00	33.59	23.00	Mg	0.980	0.956	0.981	1.235	1.128	1.355	1.243
Almandin	64.87	73.00	59.02	67.00	Ca	0.000	0.002	0.005	0.002	0.000	0.008	0.012
Spessartine	1.24	1.00	3.56	6.00	Na	0.054	0.065	0.063	0.068	0.081	0.072	0.061
Andradite	0.54	0.00	3.83	0.00	K	0.836	0.855	0.822	0.787	0.852	0.781	0.789
Uvarovite	0.27	0.00	0.00	0.00	Ba	0.003	0.005	0.007	0.014	0.000	0.017	0.010
Grossulaire	0.29	2.00	0.00	4.00	Ni	0.002	0.000	0.004	0.000	0.002	0.005	0.000
XMg	0.34	0.26	0.36	0.24	F	0.293	0.302	0.412	0.287	0.620	0.452	0.470
					Cl	0.019	0.032	0.020	0.035	0.098	0.026	0.081
					XMg	0.482	0.454	0.473	0.559	0.491	0.558	0.527

Chemical formulae were calculated on the basis of 12 oxygens for garnet and 11 oxygens for biotite

Table 5

Whole rock analyses of selected host rocks and xenoliths (Major elements in wt.%, trace elements in ppm)

	Host rocks		Xenoliths										
	Kar2b	2Kar8	Kar2b	Kar 7	2Kar 2	2Kar 3	2Kar 4	2Kar 5	2Kar 8	Ma6	Ma5	Ma3	Ma10
SiO ₂	67.58	70.76	44.99	57.01	54.03	56.47	53.76	51.25	55.84	56.24	51.51	46.76	44.07
Al ₂ O ₃	15.49	15.11	30.99	21.46	20.39	20.12	22.22	27.39	17.58	24.91	29.29	20.26	33.84
Fe ₂ O ₃	5.15	2.74	9.41	6.22	8.75	8.98	8.08	5.92	12.66	5.31	6.01	9.55	6.20
MnO	0.22	0.04	0.25	0.24	0.16	0.18	0.10	0.15	0.12	0.11	0.22	0.15	0.13
MgO	1.82	1.00	2.67	2.37	3.17	4.33	2.55	2.26	4.15	1.78	1.57	3.31	1.74
CaO	2.37	2.34	2.94	2.88	4.62	1.97	4.04	2.13	2.91	3.40	1.53	3.58	0.67
Na ₂ O	3.18	3.57	2.98	2.68	3.49	4.01	4.45	3.86	2.82	3.49	2.60	2.74	1.94
K ₂ O	3.39	3.94	3.98	5.62	3.25	2.69	3.69	5.17	1.78	3.25	5.36	4.38	4.29
TiO ₂	0.60	0.36	1.56	1.41	1.78	1.15	1.03	1.60	1.95	1.42	1.81	1.22	2.00
P ₂ O ₅	0.20	0.14	0.23	0.10	0.37	0.09	0.08	0.28	0.19	0.10	0.12	1.36	0.09
L.O.I	2.62	3.38	3.22	3.25	6.45	4.35	2.16	3.86	5.81	7.93	3.62		
Ba	812	842	895	1531	260	725	1126	1724	603	1426	2954	1108	2390
Be	2.41	3.05	5.60	4.00	2.62	2.70	2.79	4.33	3.47	10.87	7.92	4.77	10.03
Cd	16.82	0.48	1.56	1.23	0.58	2.03	0.31	0.34	1.50	0.25	0.42	0.52	0.16
Ce	65.03	60.44	105.80	72.16	31.31	56.36	99.05	133.10	37.03	69.62	191.70	124.10	151.70
Co	21.44	4.16	19.09	11.74	29.04	24.20	16.83	14.15	16.92	13.27	19.60	25.04	18.59
Cr	25.10	16.86	143.90	149.50	57.32	84.55	90.22	142.90	112.0	100.90	160.90	91.68	167.20
Cs	2.71	2.37	7.23	3.07	10.52	2.72	3.70	9.43	2.66	12.86	6.98	6.82	6.36
Cu	17.77	9.41	6.00	8.78	6.27	18.66	18.26	0.00	40.17	14.38	7.84	7.01	5.82
Dy	3.37	3.15	5.54	2.90	2.94	2.57	3.25	7.07	3.00	2.32	7.70	15.10	9.02
Er	1.77	1.66	2.83	1.35	1.79	1.27	1.15	3.59	1.56	1.14	4.05	7.43	5.01
Eu	1.11	0.84	1.79	1.85	0.90	1.31	2.14	2.38	1.73	1.50	3.72	3.75	3.73
Ga	20.52	20.20	37.71	28.55	28.21	28.76	32.22	34.25	34.53	29.58	33.02	26.96	40.14
Gd	4.01	3.73	6.73	4.15	2.92	3.33	4.91	8.02	3.35	3.19	9.87	17.63	9.40
Ge	1.45	1.34	1.90	1.58	1.55	1.65	1.88	2.19	1.74	2.04	3.01	2.00	4.50
Hf	4.56	4.59	6.92	5.12	3.72	8.27	7.24	7.69	3.42	5.94	8.51	20.46	8.48
Ho	0.68	0.62	1.09	0.53	0.67	0.49	0.51	1.38	0.62	0.44	1.52	2.76	1.76
La	32.77	30.12	53.55	34.47	14.25	27.73	48.09	66.34	18.48	34.49	89.82	54.26	76.65
Lu	0.26	0.25	0.43	0.21	0.26	0.21	0.18	0.53	0.18	0.18	0.60	1.09	0.80
Mo	3.70	0.00	1.22	0.66	1.78	1.22	1.03	5.15	0.00	0.80	0.89	0.50	0.52
Nb	8.89	8.20	22.32	19.87	13.94	22.19	20.45	23.92	29.03	14.03	23.70	19.07	21.88
Nd	26.86	24.64	45.64	30.20	14.93	22.70	38.62	55.05	16.25	26.83	82.41	61.38	64.28
Ni	40.64	10.95	45.82	37.63	39.26	57.29	32.79	27.70	47.96	30.10	30.30	35.97	55.67
Pb	431.85	51.84	142.78	542.61	43.15	151.1	39.06	30.79	299.0	11.57	43.81	41.64	22.47
Pr	7.33	6.74	12.28	8.05	3.73	6.16	10.76	14.91	4.09	7.46	21.78	15.25	17.46
Rb	103.60	128.1	148.60	136.90	201.3	104.8	151.20	147.30	76.03	137.70	185.50	189.90	169.00
Sm	4.91	4.71	8.39	5.82	3.15	4.37	6.77	10.23	3.66	4.51	15.09	16.35	11.43
Sr	360.1	263.1	387.8	652.9	316.4	573.6	464.3	477.7	214.5	350.8	545.0	312.7	377.0
Ta	0.99	1.03	2.00	1.73	1.33	1.65	1.58	2.06	1.91	1.29	2.47	1.73	2.40
Tb	0.58	0.53	0.94	0.55	0.45	0.46	0.64	1.22	0.49	0.43	1.36	2.68	1.50
Th	13.07	15.44	15.88	10.89	5.13	16.30	20.07	21.83	6.03	7.89	24.65	18.77	22.16
U	6.51	5.88	3.16	4.18	3.19	4.79	3.46	4.03	4.10	2.58	4.00	5.81	3.62
V	62.05	32.99	204.1	192.9	162.2	99.54	124.8	205.2	182.6	156.3	216.8	140.5	234.0
W	1.59	0.41	2.94	4.91	2.94	2.65	1.40	22.49	2.17	1.94	1.62	2.74	1.66
Y	18.41	18.34	27.05	13.63	17.70	13.18	12.13	35.19	18.43	11.57	37.15	79.46	45.80
Yb	1.68	1.59	2.73	1.27	1.68	1.28	1.08	3.39	1.25	1.10	3.96	7.22	5.16
Zn	1636.0	73.21	776.1	348.7	154.4	260.1	257.6	339.2	440.3	123.3	120.6	200.4	244.0
Zr	174.20	160.10	246.60	179.00	154.80	299.50	255.40	287.10	129.60	224.20	307.10	685.50	301.50
ΣREE	529	487	873	570	305	451	735	1083	352	522	1495	1317	1297

Table 6

Non modal partial melting models

	Model 1 <i>Ma3</i>			Model 2 <i>Ma5</i>		
		6		3		
Pressure <i>P</i> (kbar)		738		810		
Solidus temperature T_0 (°C)		900		900		
Peak temperature <i>T</i> (°C)		900		900		
Phase proportions (wt %)	at T_0	in the melting assemblage		at T_0	in the melting assemblage	
Bt	0.31	0.55		0.14	0.43	
Crd	0.00			0.14	0.21	
Pl	0.05	0.11		0.59	0.23	
Kfs	0.48			0.03		
Qz	0.10	0.23		0		
Sil	0.05	0.11		0		
Gt	0.02			0		
Sp	0.00			0.04		
Ilm	0.00			0.01		
Crn	0.00			0.06	0.13	
melt	0					
Final melt fraction (wt%)		21		15		
	D_{bulk}	P	Cs/C_0	D_{bulk}	P	Cs/C_0
La	0.72	0.54	0.94	2.78	1.08	1.11
Ce	0.61	0.46	0.90	2.34	0.91	1.10
Pr	0.64	0.51	0.91	2.55	1.01	1.11
Nd	0.40	0.33	0.78	1.55	0.63	1.07
Sm	0.30	0.19	0.71	0.89	0.38	1.00
Eu	1.27	0.36	1.07	1.85	0.70	1.08
Gd	0.52	0.28	0.88	1.29	0.57	1.05
Tb	1.26	0.42	1.07	1.92	0.94	1.08
Dy	2.11	0.31	1.14	1.33	0.72	1.05
Ho	3.69	0.29	1.18	1.34	0.82	1.05
Er	8.39	0.34	1.22	1.63	1.17	1.07
Tm	9.91	0.30	1.23	1.43	1.07	1.05
Yb	7.69	0.16	1.22	0.76	0.61	0.96
Lu	12.17	0.26	1.23	1.44	1.32	1.05
V	24.64	44.08	1.23	11.05	34.25	1.15
Cr	13.74	23.59	1.22	6.17	18.45	1.13
Ni	3.23	5.59	1.14	1.98	5.02	1.05
Nb	7.82	13.91	1.20	3.41	10.55	1.09