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1 **Eocene ultra high temperature (UHT) metamorphism in the Gruf complex (Central Alps):**
2 **constraints by LA-ICPMS zircon and monazite dating in petrographic context.**

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9 The Gruf complex in the Lepontine Alps is one of the rare occurrences of Phanerozoic UHT
10 metamorphism in the world but its age is still a matter of debate. Here we present LA-ICPMS
11 dating in petrographic context of zircon and monazite from an UHT restitic granulite. Zircons and
12 monazites are both included in large crystals and in retrograde symplectites. In such restitic rocks,
13 partial melting or fluid interactions are unlikely precluding resetting of the monazite chronometers.
14 Zircon cores yield Permian ages interpreted as age of charnockitisation. They are sometimes
15 surrounded by a narrow rim at 32 Ma. Monazites are strongly zoned, but all yield a 31.8 ± 0.3 Ma
16 age interpreted as the time of complete (re-)crystallisation during the UHT paragenesis. We propose
17 that the zircons dated a post Hercynian metamorphism which is responsible of the widespread
18 formation of granulites in the Southern Alps and the crust differentiation. This fluid-absent melting
19 event produced refractory lithologies such as restites in charnockites. We suggest that Gruf UHT
20 paragenesis is alpine in age and crystallised from these refractory lithologies. We conclude that the
21 lower restitic crust produced at the Permian time had the ability to achieve UHT conditions during
22 the fast exhumation and heating related to lithospheric thinning in Alpine time.

23 **Supplementary material:** Analytical procedures for monazite analysis and dating; table of major
24 elements of the minerals; table of the isotope data.

25 **Introduction**

26 The Gruf complex in the Lepontine Alps is one of the rare occurrences of Phanerozoic UHT
27 metamorphism in the world, discovered in the Val Codera by Cornelius (1916), Cornelius & Dittler
28 (1929), and described by Barker (1964), Wenk et al. (1974). Because occurrences of UHT
29 metamorphism are mainly of Precambrian age (eg Harley, 1998; Brown, 2007; Kelsey and Hand,
30 2015), this area is of major interest to understand the geodynamic signification of such extreme
31 metamorphic conditions. Indeed, the main difficulty in understanding the geodynamic significance
32 of UHT metamorphism is that Precambrian UHT granulites are often preserved in small-scale

33 lenses: they usually represent structural and metamorphic relics in polymetamorphic rocks
34 belonging to polycyclic terrains. The lack of large-scale tectonic structures associated to their
35 emplacement precludes a clear understanding of their geodynamic history. Thereby the few
36 Phanerozoic UHT occurrences for which the geological context is well constrained are precious
37 records that help to understand and interpret this type of metamorphism.

38 The age of the UHT metamorphism in the Gruf complex is currently a matter of debate. Based on
39 zircon U/Pb dating, Galli et al. (2012) have proposed a Permian age (282 – 260 Ma) for the
40 granulite facies fluid-absent biotite melting event. For these authors, the presence of orthopyroxene
41 inclusions in zircons confirms the Permian age of the charnockites and associated sapphirine-
42 bearing granulites. Zircon rims from the same samples yield 34-29 Ma ages interpreted as dating
43 the Alpine amphibolite facies migmatization. A different interpretation was suggested by Droop and
44 Bucher (1984) and proposed by Liati and Gebauer (2003). The latter authors considered that the
45 zircon Alpine rims (yielding a weighted mean age at 32.7 ± 0.5 Ma in their samples) grew during
46 the UHT metamorphic event, and that the sapphirine-bearing granulites were restites formed during
47 partial melting of the Permian granitoids. Moreover, Schmitz et al. (2009) measuring an age of 33.0
48 ± 4.4 by monazite chemical dating also agree with this interpretation.

49 These contrasting results imply two different models for the origin of the UHT metamorphism: 1) a
50 Permian UHT metamorphism can be linked to the post Hercynian high-thermal regime, associated
51 with lithospheric Permian-Triassic thinning and responsible for of the widespread formation of
52 granulites in the Austroalpine and South -Alpine continental crust (e.g.: Brodie et al., 1989;
53 Lardeaux and Spalla, 1991; Diella et al., 1992; Barboza and Bergantz, 2000; Muntener et al. 2000;
54 Schuster et al., 2001; Spalla and Marotta, 2007; Schuster and Stuewe 2008; Galli et al., 2013; Spalla
55 et al., 2014). This anomalously high thermal regime would be responsible for a pervasive melting
56 event, associated with melt loss leading to the genesis of a residual, refractory lower crust. Such
57 processes are interpreted to be responsible of the lower continental crust differentiation (e.g.
58 Vielzeuf and Holloway, 1988; Brown, 2008; Redler et al., 2013); 2) An Alpine UHT metamorphism
59 would have been driven by lithospheric thinning associated with slab breakoff and asthenospheric
60 upwelling ((e.g. Davies and von Blanckenburg, 1995; von Blanckenburg and Davies, 1995; Oalman
61 et al., 2016) which would provide the considerable amount of heat necessary to reach the UHT
62 conditions.

63 In this study, we combine for the first time zircon and monazite in situ dating in petrographic
64 context for the Gruf sapphirine granulite. The results provide a new opportunity to clarify the age of
65 the UHT event and its geodynamic context.

66

67 **Geological setting**

68 Penninic nappes in the Central Alps consist of variegated rocks of continental and oceanic origins.
69 This nappe stack is separated from the Southern Alps by the Periadriatic Lineament and by a thin
70 ribbon of Austroalpine crust verticalised along the “southern steep belt” (e.g. Schmid et al., 1996).
71 The axial portion of Central Alps has recorded a polycyclic metamorphic evolution: structural and
72 petrologic relics of Hercynian and Caledonian imprints have been described (Schaltegger, 1994;
73 Spalla et al., 2014 and refs therein). The Alpine overprint occurs with heterogeneous intensity, and
74 the separation of Alpine from pre-Alpine metamorphic imprints has remained for a long time
75 difficult due to the similar metamorphic conditions associated with these successive orogenic cycles
76 (e.g.: Niggli, 1974; Engi et al., 2004). Alpine metamorphism in the Central Alps is characterized by
77 a polyphasic metamorphic evolution characterised by an early high pressure- low- to intermediate-
78 temperature imprint, preserved as relic blueschist- and eclogite-facies assemblages, recorded during
79 the south-verging subduction. The second metamorphic imprint is characterised by assemblages
80 indicating a Barrow-type event interpreted as consequent to the continental collision. Isograds of the
81 Barrovian metamorphism define the “Lepontine metamorphic dome” (Trommsdorff, 1966; Todd
82 and Engi, 1997) and their concentric distribution indicates that metamorphic conditions increase
83 southwards from greenschists- to upper amphibolite-facies. In this southern part migmatitisation
84 conditions have been attained at about 700°C and 0.6-0.8 GPa between 32 and 22 Ma (Engi et al.,
85 1995; Burri et al, 2005; Berger et al, 2009; Rubatto et al., 2009). The Gruf complex (Fig. 1) is a
86 small tectonic unit of about 200 km², located in the southeastern part of the Lepontine dome, north
87 of the Insubric Line, and limited to the east by the calc-alkaline Tertiary intrusive stock of Bergell.
88 This intrusive massif is considered synchronous with crustal anatexis at 33 - 28 Ma (Berger et al.
89 1996). The Gruf complex is composed mainly of biotite-garnet-sillimanite-cordierite metapelitic
90 rocks and migmatitic orthogneisses and paragneisses, and has been recently considered as part of
91 the tectonic *mélange* of continental and oceanic units accreted together in the Alpine tectonic
92 accretion channel (Engi et al., 2001). During a remarkable field work in a very difficult terrain Galli
93 et al. (2012) described structural and petrologic characters of the complex in situ for the first time,
94 focusing especially the Mg-Al-rich sapphirine granulites. The latter form schlieren and residual
95 enclaves within sheet like bodies of charnockites and migmatitic orthogneisses (Galli et al., 2013).
96 The chemical composition of the Mg-Al-rich granulites is comparable to that of restitic
97 surmicaceous enclaves in granites (e.g. Montel et al., 1991) except for magnesium. Kelsey et al.
98 (2003) suggested that the production of Mg-Al-rich compositions through melt loss is improbable
99 because of the low Mg partition into melt. For these authors Mg-rich assemblages may source from
100 Mg-enriched protoliths. Consequently, the Gruf Mg-Al-rich sapphirine granulites may represent
101 resister lenses within charnockites/orthogneisses or restites from Mg-rich protolith included in the

102 charnockites.

103 Previous published geochronological studies are at the origin of the contrasted interpretations for
104 the geodynamic signification of the UHT conditions recorded in the Gruf complex. Liati and
105 Gebauer (2003) report SHRIMP weighted mean ages at 272.0 ± 4.1 Ma in zircon cores and at 32.7
106 ± 0.5 Ma in zircon rims from a sapphirine bearing granulite sample. The Permian ages are interpreted
107 as reflecting the age of the magmatic protolith whereas the Alpine ages are considered to represent
108 the age of the granulite facies metamorphism. For these authors, sapphirine-bearing granulites
109 represent restites formed during Alpine partial melting of Permian granitoids. Galli et al. (2011;
110 2012) proposed different interpretation based on similar geochronological data obtained with the
111 same method (Zircon SHRIMP analyses). For these authors 282-260 Ma ages obtained in
112 oscillatory zoned zircon cores represent melts generated through granulite facies fluid-absent biotite
113 melting at 920-940°C in metapelitic rocks, whereas 34-29 Ma ages in zircon rims date the Alpine
114 amphibolite facies migmatization. For these authors the charnockites associated with the sapphirine-
115 bearing granulites belong to the post-Hercynian European lower crust. Schmitz et al. (2009) applied
116 the method of 3D-Micro X-ray fluorescence analysis on monazite in thin section from a sapphirine-
117 bearing granulite. They obtained an age at 33.0 ± 4.4 Ma in monazites included in and intergrown
118 with HT minerals which they interpreted as the age of the high-temperature event. Finally, Oalman
119 et al. (2013, 2016) also suggest that UHT conditions were reached slightly before 32.5 Ma followed
120 by cooling from 30 to 19 Ma recorded by rutile $^{206}\text{Pb}/^{238}\text{U}$ ages.

121

122 **Petrography of the Mg-Al granulites**

123 Mg-Al-rich sapphirine granulites have been already carefully described by Barker (1964), Droop
124 and Bucher (1984), Galli et al., (2011) and Guevara and Caddick (2016). We will concentrate here
125 on the main petrological characters and mineral reactions, which appear to be of major interest to
126 link petrology and geochronology. The studied sample is a pebble which has been collected in upper
127 Val Codera (Fig. 1). Two domains have been recognized at the sample scale: one is a sapphirine-
128 bearing granulite (Fig. 2a) while the other has the typical mineralogy of a charnockite (Figs. 2d and
129 e). The boundary between these two domains is progressive. Following Galli et al (2011), we admit
130 that the granulitic domain could be a restite/schlieren of Mg-rich metapelite or perhaps a restite
131 within the charnockitic domain. The charnockitic paragenesis is composed of millimetric to pluri-
132 millimetric crystals of Opx - Bt - Kfs \pm Pl - Qz - Mnz - Zrn and Ap (Figs. 2d and e). Rare inclusions
133 of zircon, monazite, plagioclase and quartz are observed in the phenocrysts. In some places, clusters
134 of K-feldspar - quartz-apatite microcrysts are following the grain boundaries in the charnockitic
135 assemblage (Fig. 2e). They seem to represent incipient melting or residual melt after extraction. In

136 the charnockite domain, K-feldspar is orthose (XOr: 80%; XAb: 20%), plagioclase has an
137 intermediate composition with XAn: 40. The compositions of orthopyroxene and biotite are
138 substantially the same as in the granulitic domain of the rock.

139 The primary crystals of the residual/resister granulite are millimetric to plurimillimetric. The peak
140 UHT paragenesis was: Al rich-Opx - Sil - Spr - Bt - Grt - Crd - Rt - Ap - Zrn - Mnz (Fig. 2a). Here
141 again some tiny inclusions are rarely present. Detailed observations of the mineral associations
142 reveal complex relationships with (at least) 2 generations of Al-rich orthopyroxene, sillimanite,
143 cordierite and sapphirine ± spinel ± biotite, garnet and rare inclusions of quartz, rutile and apatite.
144 Secondary minerals are abundant. Except for orthopyroxene, the chemical composition of the
145 minerals does not change much and the crystals are weakly zoned (Table A supplementary
146 material). Garnet is almost pure almandine - pyrope solid solution with a slight decrease of pyrope
147 relative to almandine at the rim of the crystals (from the core to the rim: XPy: 48-46 %; XAlm: 42-
148 45 %). Grossular and spessartine contents are low: XGrs: 2.2-3.1% and XSp: ≤1%. Opx is Al₂O₃-
149 rich and zoned (core: 7.5-8.6% and rim: 5.5 to 7 % Fig. 3). Biotite is Mg- (XMg: 75-80%) and Ti-
150 rich (TiO₂: 2.6-3.6%). Cordierite composition is homogeneous both in primary minerals and in
151 symplectites. Galli et al. (2011), Oalman et al. (2013), Guevara and Caddick (2016) estimated the
152 conditions of the primary paragenesis in these granulites at T = 920-940 °C and P= 0.85-0.95 GPa.

153 The secondary symplectites in the granulites are varied and complex, with Spr, Sil, Crd, Opx ± Sp.
154 Spinel is hercynite - spinel solution, homogeneous in composition. Two main reactions dominate
155 (Fig. 2b-c): Sil 1 phenocrysts are surrounded by Spr 2 + Crd 2 in contact with Opx ± Bt; the garnet
156 is destabilized into symplectites of Opx 2 + Crd 2. These two reactions indicate a pressure decrease.
157 Al₂O₃ content in Opx 2 is between 7.4 to 5.9 % similar to the rims of the primary phenocrysts (Fig.
158 3) whereas the XMg is slightly lower than in the primary crystals. Primary Opx probably continues
159 to grow at the beginning of the retrograde evolution during decompression, while garnet is
160 destabilized. This suggests that primary paragenesis and retrograde symplectites are the product of a
161 single metamorphic event. These retrograde textures demonstrate a re-equilibration of the UHT
162 peak assemblages at lower metamorphic conditions: 720-740 °C at 6.5-7.5 kbar which starts with a
163 decompression and continue with a temperature decrease (Galli et al. 2011; Oalman et al. 2013;
164 Guevara and Caddick, 2016). These conditions are similar to those inferred for the migmatization
165 occurring during for the Tertiary Barrovian regional metamorphism (Burri et al, 2005; Engi et al.,
166 1995).

167 Both zircons and monazites are included in the large crystals from the UHT assemblage as well as
168 in the late symplectites.

169

170 **Zircon and monazites textures**

171 Zircons are subhedral elongated and/or resorbed rounded crystals 30-100 μm long. They contain
172 rare and very tiny inclusions among which biotite, white mica and apatite were only unequivocally
173 identified by Raman spectrometry. Sillimanite has not been observed. Zircons included in the
174 primary phenocrysts are resorbed grains and seem to be relictual (Fig. 4a). Cathodoluminescence
175 images (CLI) show that most of the grains display a large inner domain with usually oscillatory or
176 rarely complex zoning (Fig. 4b - c). In zircon included in the cordierite or biotite, this inner domain
177 is sometimes surrounded by a thin rim ($<5 \mu\text{m}$ to $15 \mu\text{m}$) with euhedral faces and bright in CDI
178 (Fig. 4c). U and Th contents are variable ($150 < \text{U} < \sim 4860$ ppm and $4 < \text{Th} < 1180$ ppm; Table B
179 supplementary material) with no correlation with the internal structure nor ages (see the next
180 section). As a result, Th/U ratios are also variable ($0 < \text{Th}/\text{U} < 0.50$) and not related to the different
181 zircon domains (i.e. oscillatory zoned inner domain or external rim).

182 Monazites are 50 to 150 μm in size and are present in the core of large Spr, Opx, Crd crystals or
183 form clusters of small grains (10 to 40 μm) in the late symplectites. The shape of the grains varies
184 from large rounded grains to small elongated ones with rarely indented grain boundaries (Fig 5b
185 and 6a), without any correlation with the textural position. Rare inclusions are needles of sillimanite
186 (Fig. 6a). All the grains are strongly zoned (Table 1). Chemical variations of the LREE-, Th- and
187 Ca-contents broadly follow the brabantite substitution ($2\text{REE}^{3+} = \text{Th}^{4+} + \text{Ca}^{2+}$). ThO₂ content varies
188 from 1.3 to 10.8 wt%. Y₂O₃ and UO₂ contents are highly variable (from 0.1 to 5.5 wt% and from
189 0.2 to 3.5 wt%, respectively) and inversely correlated. These variations are clearly related to the
190 zonation of the sample (Fig. 7a): monazites in the charnockite domain are Y-rich and U-poor
191 whereas the opposite is true in the granulite domain. Variations in LREE, MREE and Y₂O₃ contents
192 are also correlated to the sample zonation: monazites from the charnockite domain are enriched in
193 Gd₂O₃ and Y₂O₃ whereas monazites from the granulite domain are enriched in La₂O₃ and poor in
194 Gd₂O₃ and Y₂O₃ (Fig. 7 b - c). Chemical variations are observed from grain to another but also
195 within single crystal (Fig. 6 a - b). Some of the grains display very tiny Y-rich overgrowths (Fig.
196 6a).

197 **U-Th-Pb geochronology**

198 Monazite and zircon have been analysed in thin section by LA-ICPMS in order to control the
199 position of the grain (and thus the measured ages) relative to the textures observed in the sample.
200 Zircons and monazites were ablated using a Resonetics Resolution M-50 equipped with a 193 nm
201 Excimer laser system coupled to an Agilent 7500 cs ICP-MS with a 1 Hz repetition rate, a fluence
202 of 7 J/cm² and a spot size of 7 μm for monazite dating and 3 Hz repetition rate, a fluence of 2.6
203 J/cm² and a spot size of 20 μm for zircon dating. Helium carrier gas was supplemented with N₂

204 prior to mixing with Ar for sensitivity enhancement (Paquette et al., 2014). Detailed analytical
205 procedures are reported in Paquette and Tiepolo (2007) and Didier et al. (2015) and in
206 Supplementary – Analytical procedures. Monazite can be dated using both U-Pb and Th-Pb decay
207 schemes. Here we will only consider $^{208}\text{Pb}/^{232}\text{Th}$ ages for the following reasons. Firstly, ^{232}Th is
208 largely predominant in monazite, allowing small spots (7 μm) to be performed during laser ablation.
209 Secondly, U decay series could be in disequilibrium in young monazites (Schärer, 1984), resulting
210 in overestimated $^{206}\text{Pb}/^{238}\text{U}$ ages. Because secular equilibrium among the intermediate daughters of
211 ^{232}Th occurs after about 30 years, it seems reasonable to assume that initial ^{208}Pb is absent. Thirdly,
212 ^{232}Th is so abundant that ^{208}Pb originating from common Pb is usually negligible compared to
213 radiogenic ^{208}Pb . To reinforce this hypothesis, we only consider here the concordant $^{208}\text{Pb}/^{232}\text{Th}$ –
214 $^{206}\text{Pb}/^{238}\text{U}$ ages (Fig. 8a). Whatever the textural position of the monazite grains or the chemical
215 composition of the monazite domain, the $^{208}\text{Pb}/^{232}\text{Th}$ ages are between 29.6 ± 1.0 and 34.1 ± 1.1
216 Ma, allowing to calculate a weighted average age of 31.8 ± 0.3 Ma (MSWD = 3,2, N = 51; Fig. 8b;
217 Supplementary material: Table B).

218 $^{206}\text{Pb}/^{238}\text{U}$ ages measured in zircon from our sample are scattered from 30 ± 1 to 304 ± 12 Ma (only
219 U/Pb concordant ages; Fig. 8c). The youngest ages have been obtained in the bright rims. Plotted in
220 a Tera Wasserburg diagram, we see that these ages are in total agreement with the ages measured in
221 the monazite grains (Fig. 8c). The oldest ages are upper Carboniferous-lower Permian and have
222 been obtained in the inner oscillatory zoned domains. Two reasons can be provided to explain
223 intermediate ages between the oldest and the youngest ones: i) they reflect mixing between the
224 Carboniferous core and the narrow Cenozoic rim (<5 μm to 15 μm) due to analysis of distinct
225 adjacent domains; ii) they result from Pb loss. Indeed, in some of the grains the U content is high
226 (U > 1000 ppm). Such a high U content could be responsible for radiation damage of the crystal
227 lattice especially in Paleozoic zircons and induce Pb loss. This results in younger core compared to
228 the rim as observed in Figure 4b. With the aim of limiting these two effects and to better restrict the
229 range of the oldest ages, we decide to only consider the ages which were measured i) in the zircon
230 crystals which do not present the luminescent rim on the CLI, ii) measured in domains with U
231 content ≤ 1000 ppm. By doing this the scattering of the $^{206}\text{Pb}/^{238}\text{U}$ ages remains between 247 ± 6 and
232 304 ± 12 Ma (Fig. 8d).

233

234 Discussion

235 ■ Ages recorded in zircon

236 Two groups of ages have been recorded in the zircons: between 247 ± 6 and 304 ± 12 Ma in the
237 core and ca 30 Ma in the rim. These ages are in total agreement with the results obtained by Liati

238 and Gebauer (2003) and Galli et al. (2012). Following these last authors, who found the
239 charnockitic paragenesis included within the zircon cores, the ages measured in the oscillatory
240 zoned inner domain could be interpreted as the age of the charnockitisation. As no older inherited
241 domains have been observed in our zircons, we favour this interpretation. This charnockitisation
242 can be developed under the Permian-Triassic high-thermal regime (Galli et al., 2013) responsible of
243 the widespread formation of granulites in the Southern Alps and Austroalpine (Bertotti et al., 1993;
244 Spalla and Marotta, 2007; Spalla et al., 2014 and references herein). The scattering of the ages
245 between 247 to 304 Ma in a single sample is rather surprising and, as previously suggested, can not
246 be only attributed to mixing ages. Such a large spread of zircon ages is known in other Austroalpine
247 and Southalpine units (Marotta et al., 2016 and refs therein) representing lower continental crust:
248 Ivrea zone (Vavra and Schaltegger, 1999; Peressini et al., 2007), Malenco (Müntener et al., 2000),
249 Sondalo (Tribuzio et al., 1999), Valtournanche (Manzotti et al, 2012). Vavra and Schaltegger (1999)
250 suggest that the U-Pb system in zircons from the Ivrea Zone could have been disturbed by different
251 types of alteration and Pb-loss processes during the Permian, Triassic and possibly early Jurassic
252 times. These perturbations were differently interpreted as consequent to the early stages of
253 Mesozoic rifting related to thermal and/or decompression pulses during extensional unroofing in the
254 Permian (Marotta et al., 2009), as for example in Valtournanche, where the pre-Alpine evolution is
255 associated with LP-HT metamorphism related to Permian-Triassic lithospheric thinning (Manzotti
256 et al, 2012; Manzotti & Zucali, 2013). Our results, although slightly older than the previously
257 published ages (but within 2σ error bar), are in agreement with these interpretations.

258 The second group of ages between 30 ± 1 Ma and 34 ± 2 Ma has been recorded in euhedral rims
259 from some of the matrix zircons. Most of the zircons included in the primary phenocrysts do not
260 present such a younger rim (Fig 4 b). Furthermore, these results are similar to the $^{208}\text{Pb}/^{232}\text{Th}$ ages
261 measured in the adjacent monazites, whatever the textural position of the latter. Thus, these results
262 argue for a zircon recrystallisation event at ca. 32 Ma simultaneous with the crystallisation (or re-
263 crystallisation, see the discussion below) of the monazite grains.

264

265 ▪ What do we learn from the monazite ages ?

266 Contrary to the zircon results, monazites from the studied sample provide narrow age estimation at
267 31.8 ± 0.3 Ma (between 29.6 ± 1.0 and 34.1 ± 1.1 Ma). The ages are the same in all the different
268 textural positions of the monazite (included in large Spr, Opx or Crd crystals or in the Crd-Bt matrix
269 or as cluster of small grains within secondary symplectites), or the chemical zoning or the granulitic
270 (Fig. 6a) or charnockitic (Fig. 6b) paragenesis. The Y-rich overgrowth observed in some monazite
271 grains which could be related to the garnet destabilisation do not yield younger ages (Fig. 6a).

272 Similarly, no older ages (i.e. Paleozoic) have been recorded by the monazites. Usually the shape of
273 the monazite grains suggests equilibrium with, on one hand, the large primary crystals recording the
274 peak of the UHT conditions and on the other hand, with the secondary symplectites during the
275 beginning of the retrograde evolution. This demonstrates that the Spr-Opx-Sil UHT paragenesis as
276 well as the retrograde symplectites in the restitic granulites equilibrated at ca. 32 Ma, in agreement
277 with earlier works (Liatì and Gebauer, 2003, Schmitz et al., 2009, Droop and Bucher, 1984). The
278 Alpine UHT event is recorded by the monazite and by the zircon rims.

279 Galli et al. (2011, 2012, 2013) disagree with this interpretation. In their model, the Alpine age
280 recorded in the zircon rims represent the age of the migmatization of the Gruf lithologies in the
281 upper amphibolite facies conditions. Moreover, they interpreted that the ages recorded at 33.0 ± 4.4
282 Ma by monazite (Schmitz et al., 2009) are supposedly the result of the monazite resetting during
283 fluid-assisted migmatization (Galli et al. 2012). We cannot support this interpretation for two
284 reasons. First, many examples exist in the literature that show that monazite is a robust
285 geochronometer able to retain age records during anatexis processes (Paquette et al. 2004; Rubatto
286 et al. 2006; Stepanov et al. 2012; Rubatto et al. 2013; Didier et al. 2014; Didier et al. 2015) or even
287 UHT conditions (Korhonen et al. 2013; Rocha et al. 2017). The main causes of possible disturbance
288 of the Th-U-Pb systems in monazite are dissolution/recrystallization processes during melt or
289 hydrous fluid interaction (eg: Cherniak et al., 2004; Kelsey et al. 2012). Fluid interaction in the
290 present context is difficult to argue and in any case supposes an external input (see the next point
291 below). Dissolution/recrystallization processes during melt interaction strongly depend upon the
292 melt composition and its H₂O content (Montel, 1986; Rapp and Watson 1986; Spear and Pyle
293 2002). Monazite solubility is low in meta- or peraluminous melts and thus monazite is able to resist
294 to crustal anatexis. To conclude, if fluid-assisted migmatization in the amphibolite conditions
295 occurred during the Alpine times following a Permian UHT event, we would expect to measure
296 Permian ages in part of the monazite domains or grains, which is not the case. “Resetting” cannot
297 be an explanation for the absence of Permian age records in monazite in the present conditions.

298 The second point questions the migmatization processes in such refractory lithologies. As reported
299 by Galli et al. (2011), Mg-Al granulites occur as restites or schlieren in the charnockites and thus
300 represent highly refractory rocks formed during partial melting processes in the Permian times.
301 Partial melting of such refractory lithologies is improbable as also suggested by McDade and
302 Harley (2001) unless a high amount of fluid is added. It is very unlikely that this fluid-assisted
303 migmatization event in the amphibolite facies conditions could be responsible for all the textures
304 observed in the UHT granulites. There is no evidence in the retrograde reactions of significative
305 fluid input as argued by Galli et al. (2011). In contrast we propose that all the metamorphic textures

306 in the Mg-Al granulites correspond to the UHT conditions during the Alpine time, as demonstrated
307 by the age recorded in the monazites equilibrated with the UHT mineral assemblages.

308 ▪ A model for the evolution of the Gruf granulites and charnockites

309 Based on the petrological observations and geochronological results obtained in our sample we
310 propose the following model for the time evolution of the Gruf granulites and charnockites. The
311 oldest ages we obtained in the zircons from our UHT granulite sample are between 304 ± 12 Ma
312 and 247 ± 6 Ma. These ages, despite slightly older, are similar to those measured by Galli et al.
313 (2012) and Liati and Gebauer (2003) in the Gruf UHT granulites. Older ages at 513 ± 8 Ma and 453
314 ± 8 Ma (weighted average ages in magmatic zircon cores) have only been reported in an enclave-
315 rich biotite orthogneiss in the Gruf complex (Galli et al. 2012). Our results suggest that oscillatory
316 zoned zircons from the UHT granulites crystallized in the Permian-Triassic times during a partial
317 melting event. This Paleozoic anatectic event occurred under the granulite facies conditions, as
318 shown by the presence of the charnockitic paragenesis included in the zircons cores (Galli et
319 al.2013), inducing the formation of both charnockitic melt and restitic/resister rocks, the precursor
320 of the sapphirine-bearing granulites. Unfortunately, the few tiny inclusions observed in the zircon
321 grains do not allow to precise what was the mineral assemblage in this precursor during the
322 Paleozoic event. We suggest that the precursor of the sapphirine-bearing granulites was a biotite –
323 cordierite rich selvage probably containing also orthopyroxene and garnet could have been added.
324 Considering the abundance of the monazites in our sample, it is likely that monazites were present
325 during the Permian times. During the Alpine times, UHT metamorphism occurred during a short-
326 lived event (less than 5 Ma) between 30 to 34 Ma recorded by both zircon rims and monazites.
327 UHT metamorphic conditions at 920-940 °C/ 0.85-0.95 GPa (Galli et al., 2011) are responsible for
328 the complete re-crystallization of both the charnockites and the Bt-Cd-Grt(?) -Opx(?) -bearing rocks
329 and enables the development of UHT parageneses appearance: Al rich-Opx - Sil - Spr - Bt - Grt –
330 Crd – Rt – Ap- Zrn - Mnz in the Mg-Al granulite and Opx, Bt, Kfs, \pm Pl, Qz, Zrn, Mnz and Ap in
331 the charnockite. In both the charnockites and the Mg-Al granulites, this UHT event is responsible
332 for the partial dissolution of the Paleozoic zircons and the growth of the Alpine rims. Because no
333 Paleozoic ages were found in monazites from our samples (as well as in the previous monazite
334 study, Schmitz et al. 2009), crystallisation processes for the Alpine monazites are more difficult to
335 assess. Two processes can be proposed: crystallisation from a mineral precursor such as allanite or
336 apatite, or complete recrystallisation of an earlier monazite generation, possibly late Carboniferous
337 – Permian in age. We have no textural evidence of allanite nor apatite-monazite relation in our
338 sample suggesting that these minerals participate to the reaction inducing monazite crystallisation.
339 On the other hand we did not observed nor suspected the presence of older domains in the

340 monazites that we analysed, suggesting the partial recrystallisation from previous monazite.
341 However, the differences in chemical composition, but identical ages, between the monazites of
342 charnockite and granulite domains (Fig. 7) suggest a local equilibrium between monazite and the
343 charnockite versus granulite UHT mineral assemblages during the Alpine times. We thus propose
344 that Alpine monazites result from the complete recrystallisation of an earlier monazite generation,
345 probably of Paleozoic age, during the crystallisation of UHT parageneses.

346

347 ▪ Geodynamic implications and the origin of the UHT metamorphism in the
348 central Alps

349 The oldest ages we obtained in the zircon cores between 304 ± 12 Ma and 247 ± 6 Ma interpreted as
350 the age of the charnockitisation, unknown elsewhere in the Penninic domain of the Central Alps
351 (Galli et al, 2011),), developed during Permian-Triassic high thermal regime responsible for HT-LP
352 metamorphism widespread in the Austroalpine and Southalpine domains (Galli et al., 2013, Barboza
353 and Bergantz, 2000; Lardeaux and Spalla, 1991; Diella et al., 1992; Bertotti et al., 1993; Muntener
354 et al. 2000; Marotta et al., 2016). These granulites are interpreted as residual and refractory rocks
355 following partial melting events and extensive melt loss. This process is responsible for the
356 differentiation of the continental crust via fluid-absent melting reactions involving muscovite and
357 biotite (and amphibole in metabasic rocks) destabilisation. As a consequence, the precursor of the
358 Al-Mg granulites of the Gruf complex acquired its restitic character during this post Hercynian
359 event.

360 UHT metamorphism occurs at temperatures above the fluid-absent melting in most crustal rocks
361 which is an endothermic process that consumes heat and buffers the temperature (Vielzeuf and
362 Holloway, 1988; Stüwe 1995). Thus partial melting limits heat production avoiding a fertile crust to
363 attain UHT. Nevertheless, UHT conditions are more easily reached in refractory/restitic rocks and
364 develop preferentially in terrains that previously underwent metamorphism and melt loss (Clark et
365 al., 2011; Kelsey and Hand, 2015). The Gruf complex is the sole unit in Penninic Domain of Central
366 Alps recording the Permian-Triassic granulitic metamorphism, which permitted its rocks to reach
367 and record UHT conditions during the Alpine cycle.

368 Zircon and monazite ages, together with their microstructural relationships with granulite facies
369 minerals, show that the typical UHT paragenesis crystallised at Alpine time during a short-lived
370 event (less than 5 Ma) between 34 to 30 Ma. These results are in agreement with the garnet
371 diffusion modelling proposed by Galli et al. (2011) also suggesting that UHT metamorphism was
372 brief. This UHT metamorphism is contemporaneous with the Bergell tonalite-granodiorite intrusion

373 dated between 33 and 28 Ma (Berger et al. 1996) and the beginning of the anatectic event in the
374 Southern part of the Central Alps at 32 Ma (Köppel et al, 1981; Berger et al, 2009; Rubatto et al.,
375 2009). Secondary symplectites in the Mg-Al granulites just following the UHT peak conditions are
376 equilibrated in the same conditions as the Lepontine migmatization (e.g. Engi et al. 1995; Burri et
377 al. 2005). However, the 38-34 Ma UHP metamorphism in the Lepontine domain, is quickly
378 followed by a fast exhumation, then the UHT event and the final amphibolite facies migmatization
379 (Brouwer et al. 2004). This fast exhumation is related to a short-lived (less than 4 Ma) episode of
380 lithospheric thinning associated with a rise of hot asthenospheric material (Brouwer et al., 2004)
381 from 34 Ma to 32–30 Ma (Beltrando et al., 2010). Oalman et al. (2016) proposed that similar
382 lithospheric thinning associated with slab breakoff or roll back and asthenospheric upwelling could
383 also be responsible for the short-lived Alpine UHT metamorphism in the Gruf complex. A similar
384 geodynamic context is proposed for another example of Phanerozoic UHT metamorphism (16Ma)
385 of the island of Seram (Indonesia) where the UHT conditions were produced by slab rollback–
386 driven lithospheric extension and the exhumation of hot subcontinental lithospheric mantle
387 (Pownall et al, 2014; Pownall, 2015). Harley (2016) proposed that shorter duration UHT granulite
388 event can be formed as a consequence of severe lithospheric thinning and crustal extension
389 accompanied by voluminous magmatism in arc settings affected by subduction roll-back. Brouwer
390 et al. (2004) notice that heating by slab detachment is fast and transient (more than 100°C in up to
391 10 million years,), whereas radiogenic heating requires time spans of the order of tens of millions of
392 years and cessation of the subduction process. Finally, Perchuk et al. (2017) used numerical models
393 to explain ultra-hot orogeny and showed that UHT conditions at the bottom of the crust might be
394 produced by lateral propagation of the hot asthenospheric front during plate convergence associated
395 with lithospheric delamination. In this model, a close relationship of HT-UHT metamorphism with
396 tonalitic magmatism is proposed which could explain the relation between the Gruf UHT
397 metamorphism and the Bergell intrusion, which is one of the main Periadriatic igneous bodies
398 emplaced during Late Alpine times, characterised by an enriched mantle source (von Blanckenburg
399 et al., 1998), and interpreted as triggered by post-collisional slab break off.

400

401 **Conclusions**

402 In situ and in context zircon dating on zircon from a restitic granulite within charnockites give both
403 Permian-Triassic (304- 247 Ma) and Alpine ages (ca 34-30 Ma). $^{208}\text{Pb}/^{232}\text{Th}$ ages measured in
404 monazite from the same sample yield a weighted average age of 31.8 ± 0.3 Ma (MSWD = 3.2; N =
405 51) interpreted as the time of complete (re-)crystallisation of the monazite in equilibrium with the
406 UHT paragenesis. The Alpine UHT event was thus recorded by zircon rims and monazites. The

407 oldest ages we obtained in the zircons between 304 ± 12 Ma and 247 ± 6 Ma are interpreted as the
408 age of the charnockitisation during which the precursor of the Al-Mg granulites of the Gruf
409 complex acquired its restitic character. Our results show that the Gruf complex is the sole Penninic
410 unit in the Central Alps recording the post Hercynian granulitic metamorphism which permits to
411 reach and record UHT conditions during the Alpine cycle. The typical UHT paragenesis crystallised
412 at Alpine time during a short-lived event (less than 5 Ma) related to lithospheric thinning associated
413 with a rise of hot asthenospheric material.

414 **Acknowledgments**

415

416 **References**

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643

644 **Figures captions**

645 **Fig.1:** Sketch map of the Gruf complex in the eastern Central Alps after Galli et al., (2011);
646 Asterisk locates upper Val Codera. The sketch is located in the Alpine chain (inset).

647 **Fig. 2:** Microphotographs of the parageneses and textures of the granulitic rock (Plane-
648 (PPL) and cross (XPL)-polarized light; (a): The Spr bearing granulitic paragenesis (PPL); (b):
649 Complex retrograde symplectites in the Spr bearing portion of the rock granulite with spr₂, sp, cord,
650 sil₂. Opx + Crd symplectites replace Grt (PPL); (c): Spr₂+crd resulting of the retrograde reaction
651 Opx + Sil = Spr + Crd (PPL); (d): the charnockitic paragenesis (XPL); (e): textures of incipient
652 melting or residual melt after extraction in the charnockite portion of the rock (XPL). Abbreviations
653 in text and figures are from Whitney and Evans (2010).

654 **Fig. 3:** Al₂O₃ profiles in the primary Opx (C: core and R: rim) and Al₂O₃ concentrations
655 (contents values) in the Opx from the Opx + Crd symplectite (sym) around garnet.

656 **Fig. 4 (a):** Resorbed crystals of zircons included in primary phenocrysts, Spr and Sil (PPL);
657 (b): Cathodoluminescence images of zircons in the Crd – Bt matrix with oscillatory zoning; (c):
658 zircon grains surrounded by highly luminescent rim (<5 μm to 15 μm) with euhedral faces with
659 Alpine or intermediate ages (²⁰⁶Pb/²³⁸U ages; 2σ)

660 **Fig. 5 (a):** Monazite inclusions in the phenocrysts (PPL) of Sapphirine and Al-rich
661 orthopyroxene of the Spr bearing domain of the rock; (b): Monazite inclusions in an Opx of the
662 charnockite portion of the rock; note the tiny overgrowths (rich in Y) in the BSE image (crystal
663 T14M1). Circles: ²⁰⁸Pb/²³²Th ages in Ma (with ±1.1 Ma for all data).

664 **Fig. 6 (a):** BSE images of figure 2b and of the monazites in Spr, Crd and in clusters in late
665 symplectites; cathodoluminescence of a zircon in Crd. ThO₂ and Y₂O₃ X-ray maps (spots; in wt%)
666 in the crystal T50M1 and ²⁰⁸Pb/²³²Th ages (circles) ages in Ma (with ±1.1 Ma for the monazites).
667 Black inclusions are needles of sillimanite; arrows: diamond-shaped basal sections.

668 **Fig. 6 (b):** ThO₂ and Y₂O₃ X-ray maps (spots; in wt%) and BSE image and ²⁰⁸Pb/²³²Th ages
669 (circles) ages in Ma with ±1.1Ma for a monazite in the charnockite portion of the rock (crystal
670 T12M1).

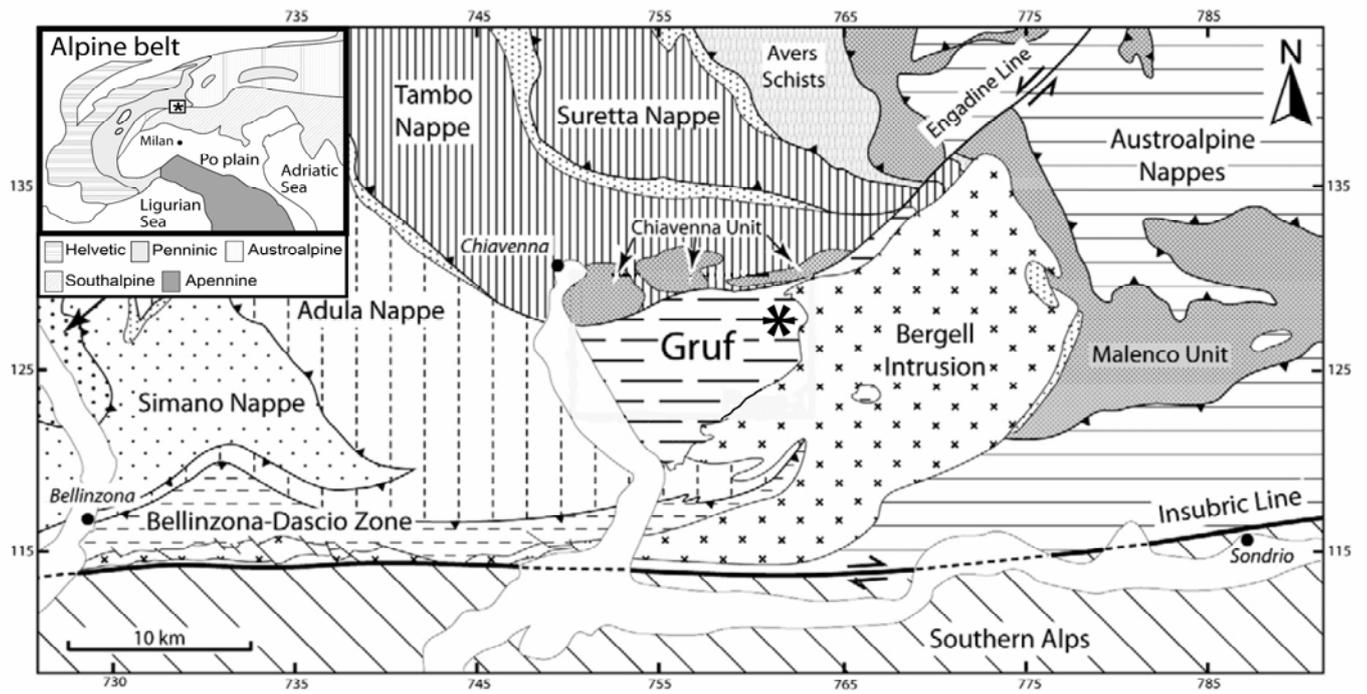
671 **Fig. 7:** Chemical composition of the monazites; **(a):** Ternary diagram showing the relative
672 Y₂O₃, ThO₂ and UO₂ (wt%) contents; **(b):** Y₂O₃ versus Gd₂O₃; **(c):** La₂O₃ versus Gd₂O₃.

673 **Fig. 8:** Geochronological results; **(a):** ²⁰⁸Pb/²³²Th versus ²⁰⁶Pb/²³⁸U diagram for the
674 monazite; **(b):** Weighted average ²⁰⁸Pb/²³²Th ages in monazite; **(c):** Tera Wasserburg diagram
675 showing the zircon and monazite data (for more clarity, only concordant U/Pb data are shown); **(d):**
676 Tera Wasserburg diagram showing U/Pb results for zircon cores with U content ≤ 1000 ppm and
677 zircon grain without apparent Alpine rim in CLI.

678 **Table 1:** Electron microprobe analyses of the monazites. G = granulite domain; Ch =
679 charnockite domain.

Table 1 : analyses of monazites.

Analyses	G			G			G			Ch			G		Ch		G		G	G	G	G
	2	7	14	17	18	23	25	32	33	34	42	45	49	53	56	57	58	59	61	64		
	T50 - M2	T50-M1		i saphirine	T50 - cluster			T14 M1			T13 M1		T12-M1		T10bis-M2		T10 bis	T10bisM1	T18-i OPX	T33-i OPX		
SiO2	0.17	0.27	0.12	0.19	0.21	0.10	0.16	0.25	0.31	0.47	0.22	0.38	0.39	0.59	0.41	0.32	0.19	1.34	0.19	0.32		
P2O5	30.15	29.51	29.77	29.54	29.93	29.99	29.67	30.20	29.96	29.65	30.29	29.57	30.02	29.09	29.60	29.57	29.80	27.70	29.65	30.12		
Ce2O3	29.92	28.92	30.59	29.74	29.23	30.28	29.79	25.18	25.44	25.61	28.36	25.20	24.73	25.71	29.57	29.48	29.22	30.89	29.94	27.14		
La2O3	13.58	13.48	14.31	13.26	13.14	13.30	13.59	10.70	11.09	11.10	12.51	11.39	10.70	11.22	13.47	13.39	13.20	14.78	13.59	12.29		
Pr2O3	3.30	3.18	3.10	3.39	3.19	3.45	3.21	2.78	2.98	2.87	2.93	3.01	2.77	2.89	3.08	3.09	3.23	3.65	3.05	3.02		
Nd2O3	11.49	10.97	11.95	11.58	11.36	11.32	11.06	9.88	10.44	10.49	10.75	8.71	10.32	11.27	11.65	11.66	10.92	11.99	10.52	11.02		
Sm2O3	1.95	1.46	1.54	1.66	1.74	2.17	1.66	2.14	1.90	2.03	1.92	1.64	1.85	1.82	1.73	1.52	1.75	1.19	1.77	1.74		
Gd2O3	1.33	0.44	0.69	1.09	1.54	1.58	1.06	2.44	1.91	2.05	1.39	1.01	2.04	1.60	1.07	1.11	1.01	0.64	0.91	1.82		
Y2O3	0.39	0.13	0.35	0.31	0.50	0.69	0.32	4.78	3.78	3.24	0.98	0.47	3.89	2.63	0.27	0.58	0.91	0.14	0.65	2.14		
CaO	1.34	1.75	1.16	1.39	1.45	1.15	1.46	1.61	1.76	1.77	1.83	2.90	1.78	1.67	1.31	1.45	1.53	0.44	1.53	1.49		
ThO2	5.79	8.43	5.15	6.42	6.07	3.92	6.73	7.84	8.75	9.32	6.78	10.82	9.59	10.07	7.17	5.74	5.19	6.96	5.48	8.00		
UO2	0.98	0.33	0.50	0.78	0.83	1.35	0.75	0.74	0.41	0.32	2.05	3.50	0.40	0.26	0.45	1.29	2.01	0.20	1.78	0.37		
PbO	0.03	0.00	0.00	0.04	0.01	0.00	0.00	0.00	0.03	0.00	0.01	0.02	0.00	0.01	0.00	0.00	0.01	0.00	0.03	0.01		
Total	100.41	98.87	99.23	99.38	99.20	99.30	99.48	98.54	98.76	98.93	100.00	98.63	98.49	98.83	99.76	99.21	98.96	99.90	99.08	99.47		
Si	0.007	0.011	0.005	0.007	0.008	0.004	0.007	0.010	0.012	0.018	0.009	0.015	0.015	0.023	0.016	0.013	0.007	0.054	0.008	0.012		
P	0.996	0.991	0.995	0.990	0.997	0.999	0.992	0.995	0.991	0.984	0.998	0.990	0.993	0.975	0.987	0.988	0.995	0.941	0.992	0.995		
Ce	0.427	0.420	0.442	0.431	0.421	0.436	0.431	0.359	0.364	0.368	0.404	0.365	0.354	0.373	0.426	0.426	0.422	0.454	0.433	0.388		
La	0.195	0.197	0.208	0.194	0.191	0.193	0.198	0.154	0.160	0.161	0.179	0.166	0.154	0.164	0.196	0.195	0.192	0.219	0.198	0.177		
Pr	0.047	0.046	0.045	0.049	0.046	0.049	0.046	0.039	0.042	0.041	0.042	0.043	0.039	0.042	0.044	0.044	0.046	0.053	0.044	0.043		
Nd	0.160	0.155	0.169	0.164	0.160	0.159	0.156	0.137	0.146	0.147	0.149	0.123	0.144	0.159	0.164	0.164	0.154	0.172	0.149	0.153		
Sm	0.026	0.020	0.021	0.023	0.024	0.029	0.023	0.029	0.026	0.027	0.026	0.022	0.025	0.025	0.023	0.021	0.024	0.016	0.024	0.023		
Eu	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000		
Gd	0.017	0.006	0.009	0.014	0.020	0.021	0.014	0.031	0.025	0.027	0.018	0.013	0.026	0.021	0.014	0.015	0.013	0.009	0.012	0.023		
Y	0.008	0.003	0.007	0.006	0.010	0.015	0.007	0.099	0.079	0.068	0.020	0.010	0.081	0.055	0.006	0.012	0.019	0.003	0.014	0.044		
Ca	0.056	0.074	0.049	0.059	0.061	0.048	0.062	0.067	0.074	0.075	0.076	0.123	0.074	0.071	0.055	0.062	0.064	0.019	0.065	0.062		
Th	0.051	0.076	0.046	0.058	0.054	0.035	0.060	0.069	0.078	0.083	0.060	0.097	0.085	0.091	0.064	0.052	0.047	0.064	0.049	0.071		
U	0.008	0.003	0.004	0.007	0.007	0.012	0.007	0.006	0.004	0.003	0.018	0.031	0.004	0.002	0.004	0.011	0.018	0.002	0.016	0.003		
Pb	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000		
Summe	2.000	2.001	2.001	2.002	1.999	2.000	2.001	1.997	1.999	2.001	1.998	2.000	1.995	2.001	1.999	2.003	2.001	2.006	2.003	1.995		

**Figure 1**

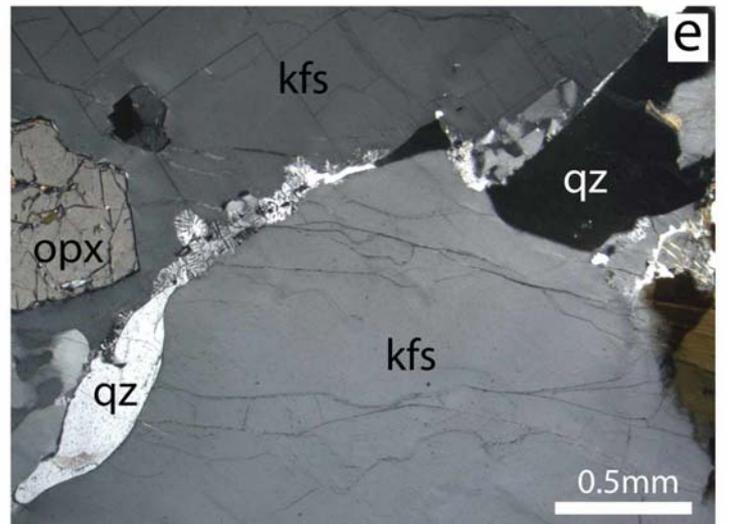
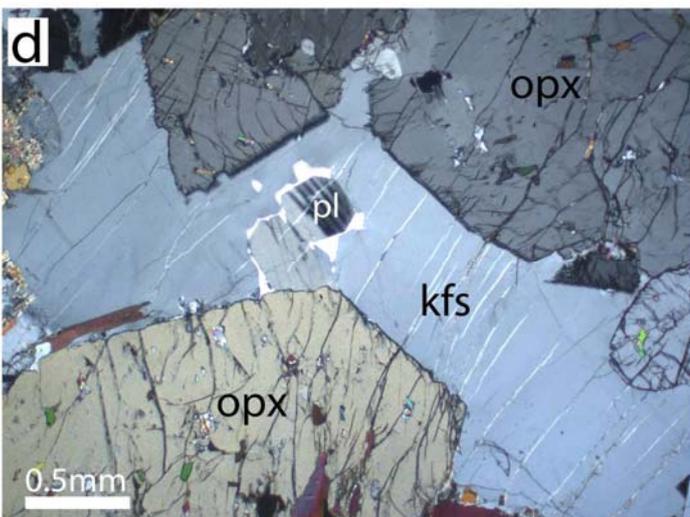
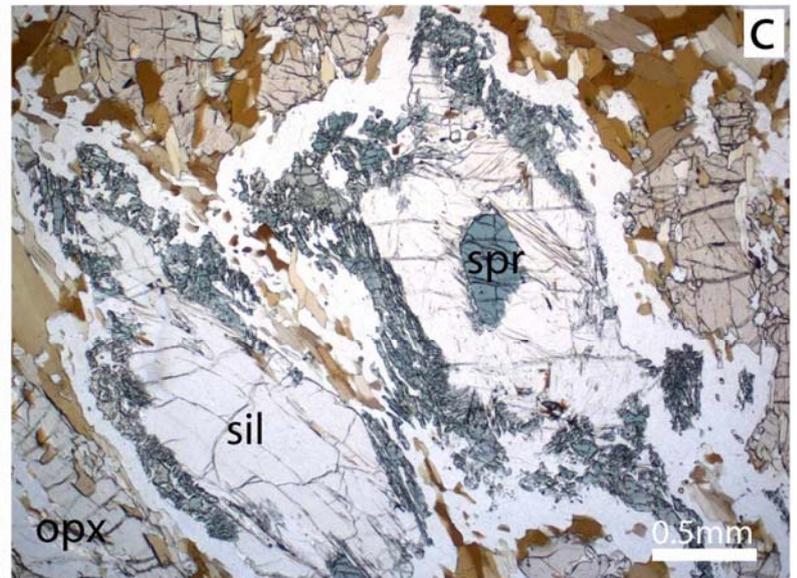
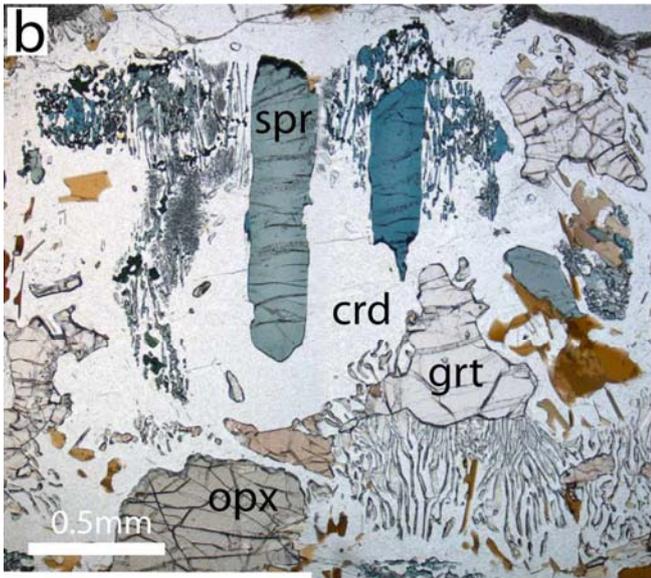
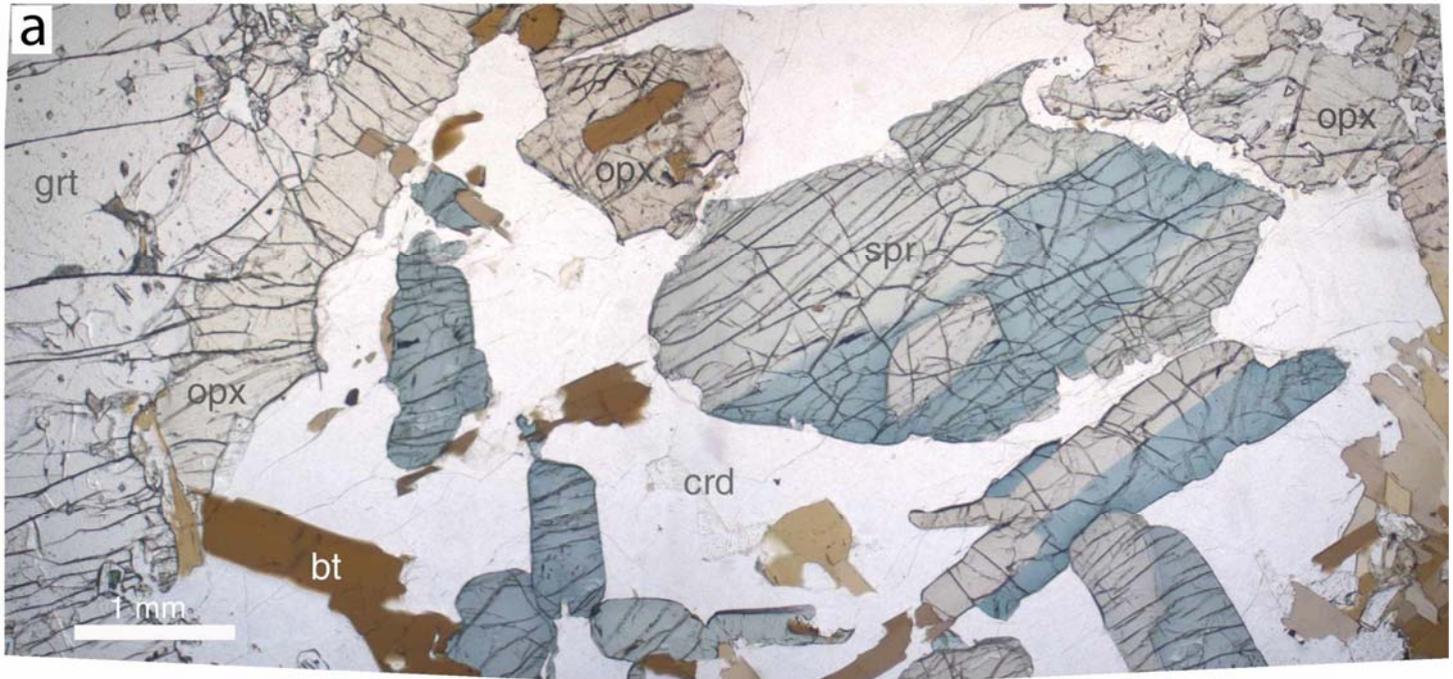


Figure 2

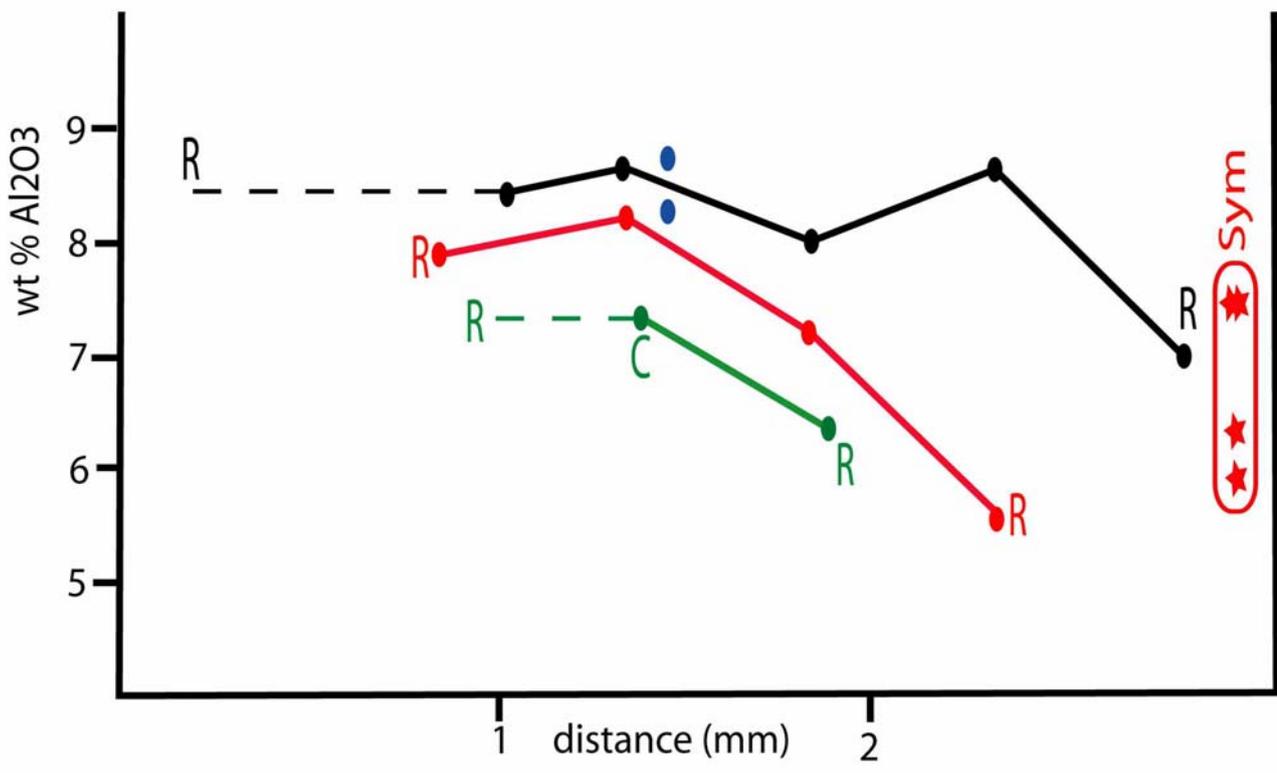


Figure 3

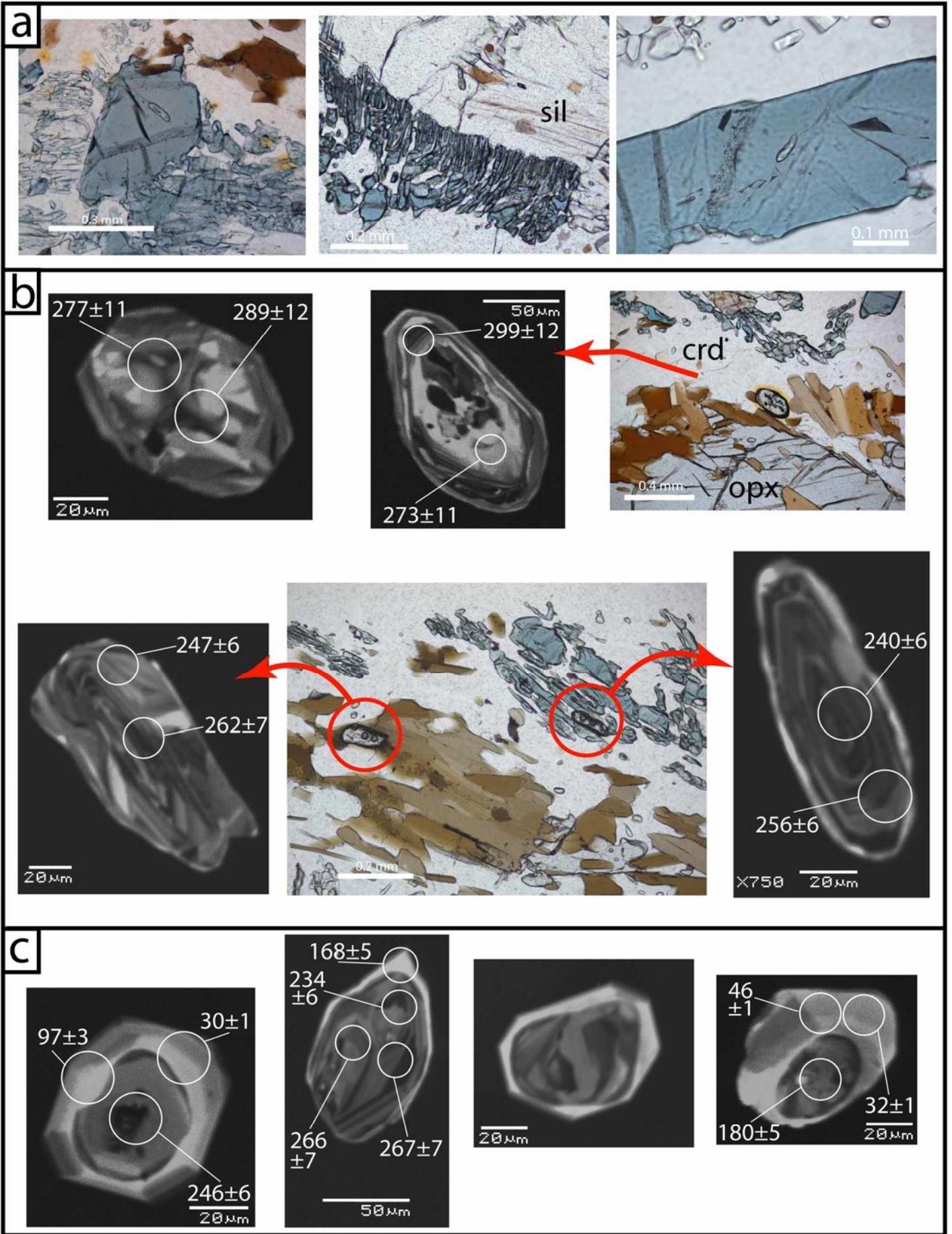


Figure 4

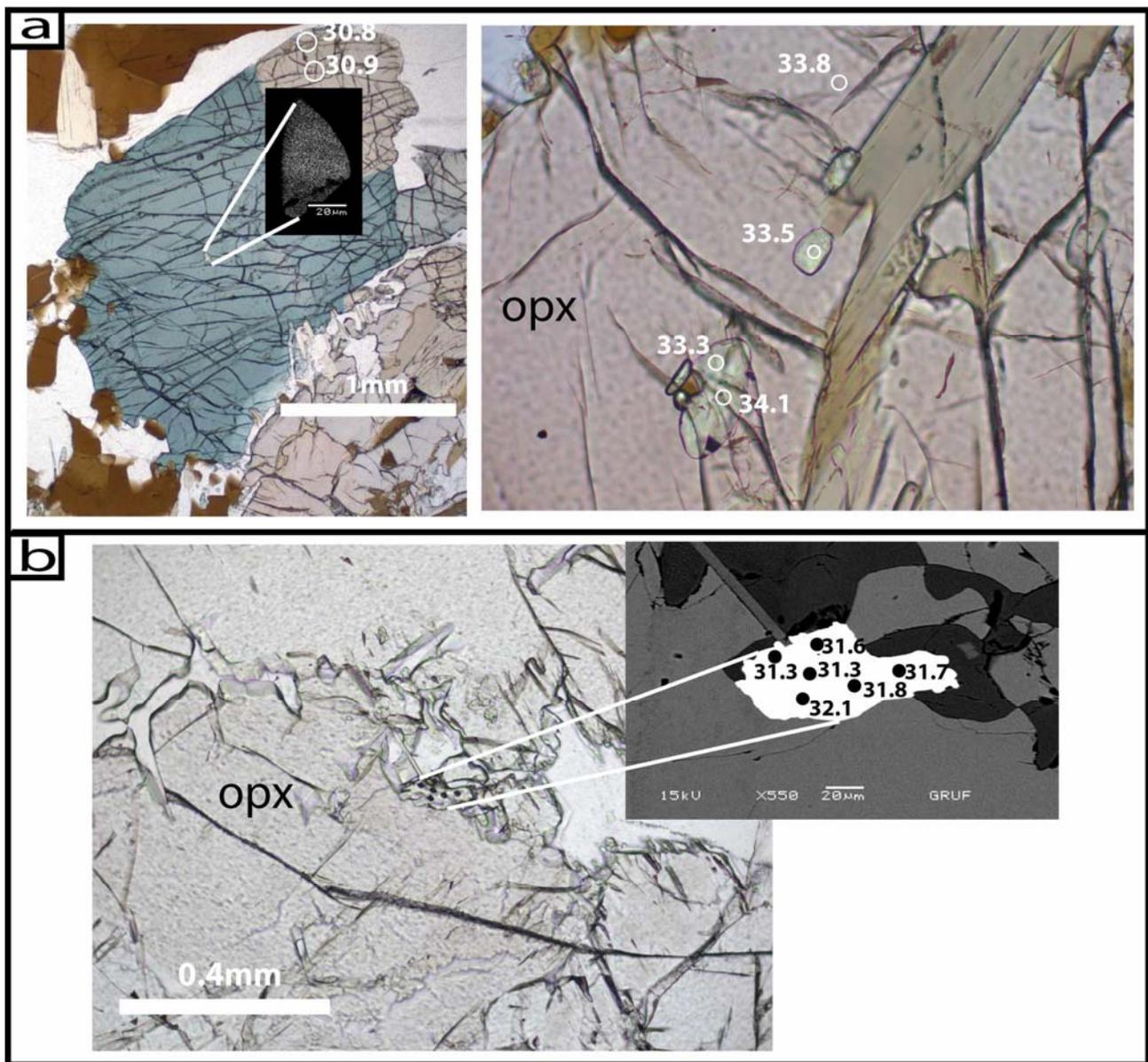


Figure 5

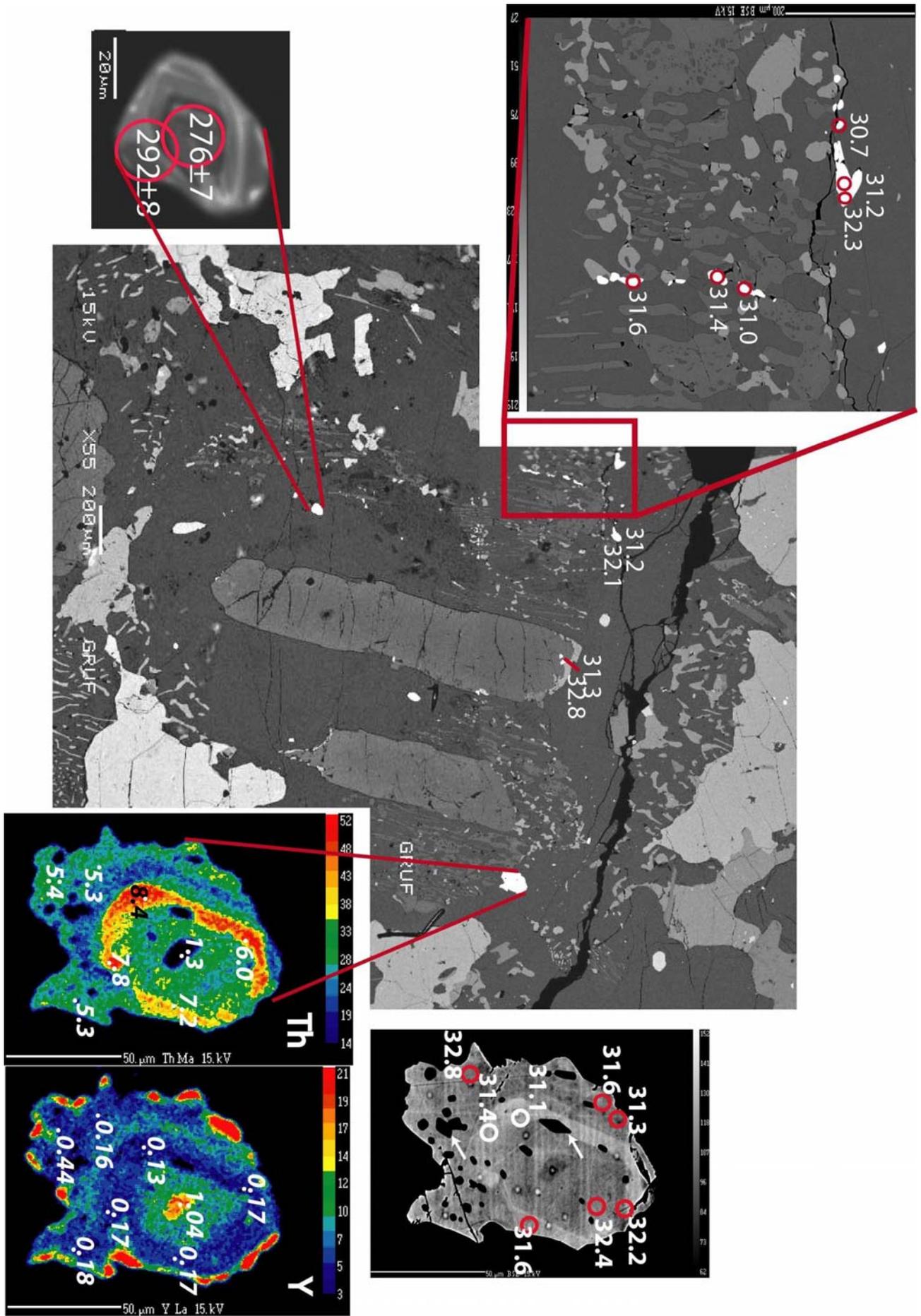


Figure 6a

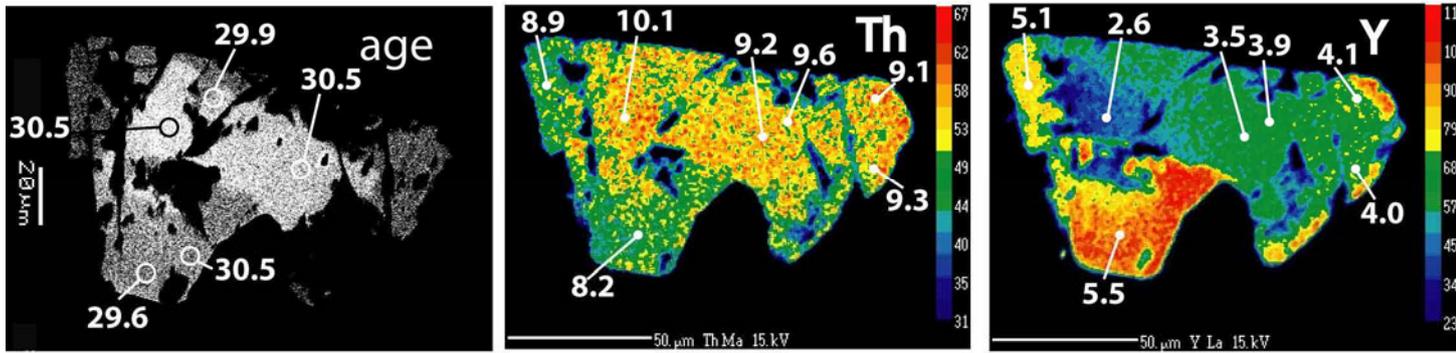


Figure 6b

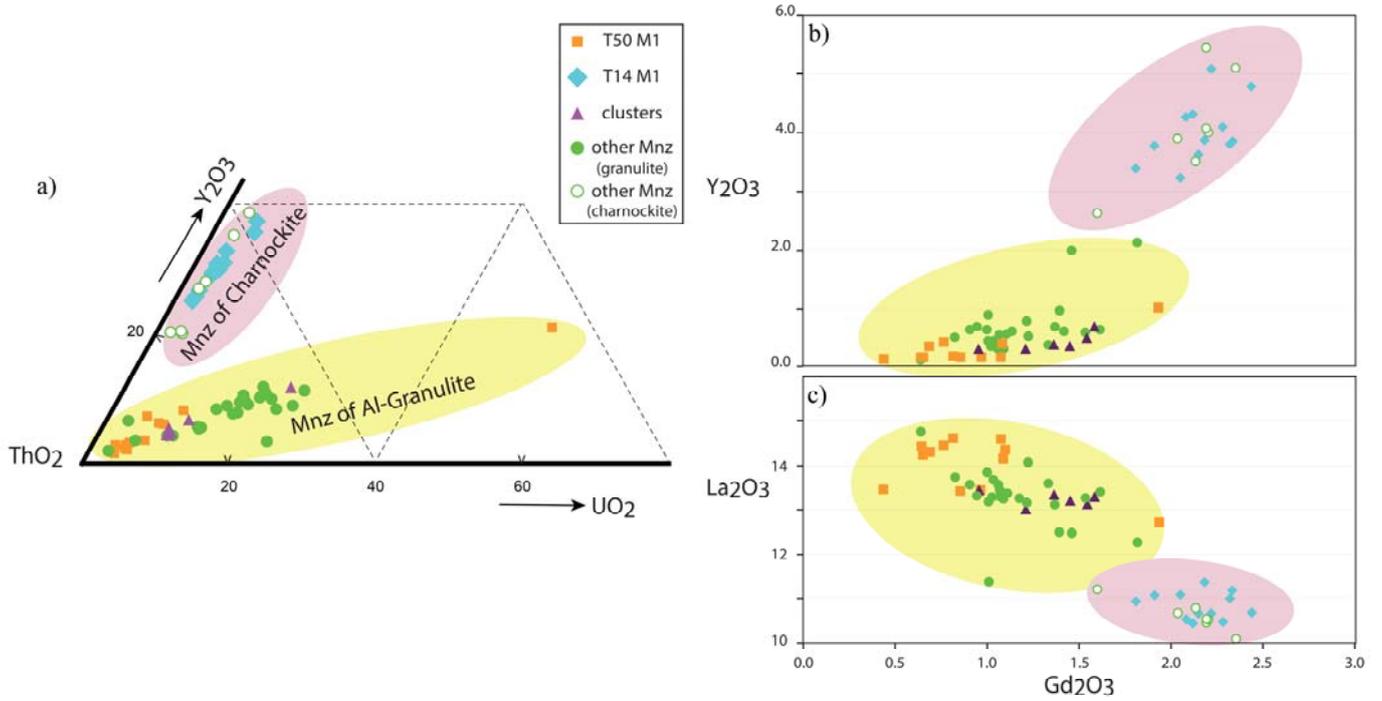


Figure 7

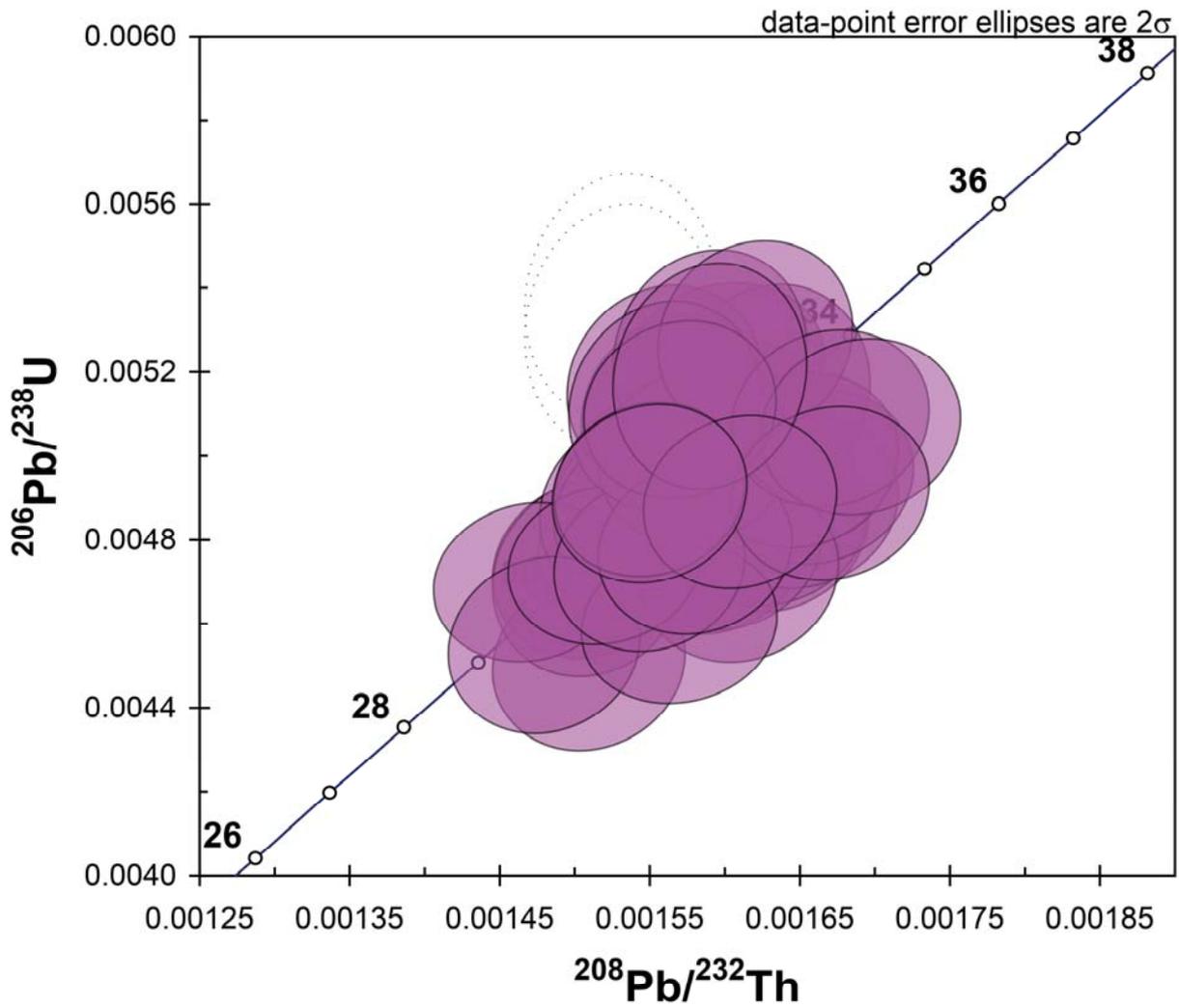


Figure 8 a

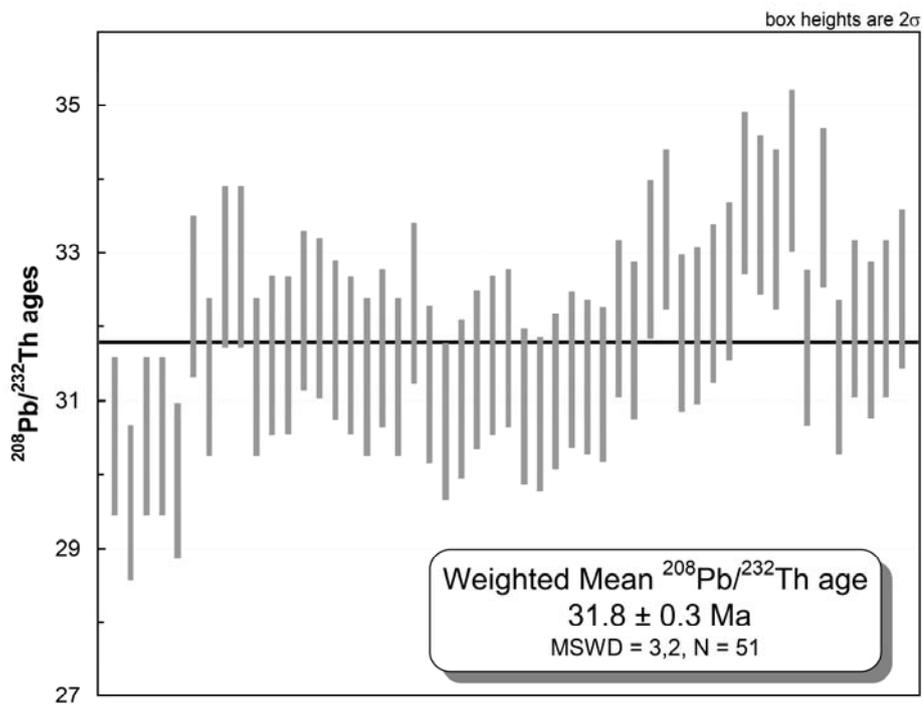


Figure 8 b

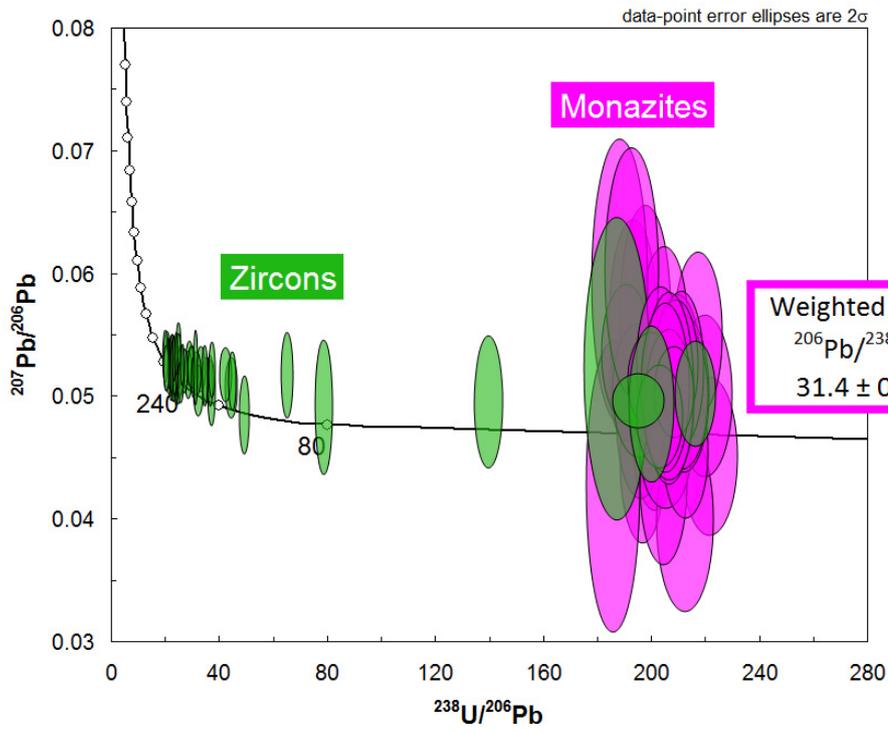


Figure 8c

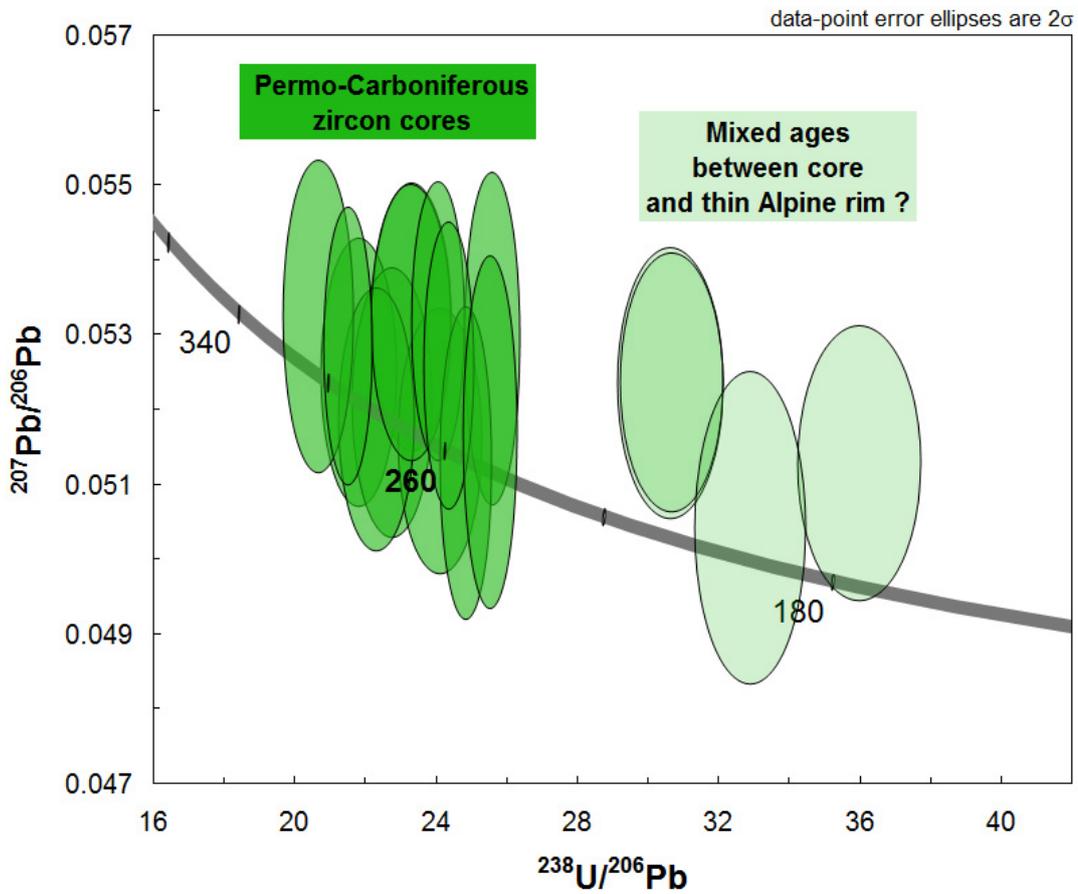


Figure 8 d