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Monazite U–Th–Pb EPMA and zircon U–Pb SIMS chronological constraints on the tectonic, metamorphic, and thermal events in the inner part of the Variscan orogen, example from the Sioule series, French Massif Central

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Damien Do Couto, Michel Faure, Romain Augier, Alain Cocherie, Philippe Rossi, et al.. Monazite U–Th–Pb EPMA and zircon U–Pb SIMS chronological constraints on the tectonic, metamorphic, and thermal events in the inner part of the Variscan orogen, example from the Sioule series, French Massif Central. *International Journal of Earth Sciences*, 2016, 105, pp.557-579. 10.1007/s00531-015-1184-0 . insu-01167544

HAL Id: insu-01167544

<https://insu.hal.science/insu-01167544v1>

Submitted on 25 Jun 2015

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1 Monazite **U-Th-Pb** EPMA and zircon U-Pb SIMS chronological constraints on
2 the tectonic, metamorphic and thermal events in the inner part of the Variscan
3 orogen, example from the Sioule series, French Massif Central

4

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15 **Abstract**

16 In the northern Variscan French Massif Central, the Sioule metamorphic series exposes from
17 top to bottom the tectonic superposition of the Upper Gneiss Unit (UGU), Lower Gneiss Unit
18 (LGU), and Para-autochthonous Unit (PAU). The nappe stacking developed throughout two
19 **prograde** syn-metamorphic events: D₁ is a top-to-the SW shearing coeval with a **probable**
20 Devonian migmatization, and D₂ is a top-to-the NW shearing **event**. Both events were
21 completed before the unconformable deposition of the undeformed and unmetamorphosed
22 “Tufs anthracifères” formation, dated at ca 330 Ma (Late Viséan). Furthermore, the UGU
23 experienced a high pressure metamorphism ascribed to a D₀ event during which eclogite or
24 granulite crystallized in several parts of the UGU. Monazite U-Th-Pb, and zircon U-Pb SIMS

25 datings were carried out in order to constrain the ages of these D₀, D₁, and D₂ tectonic-
26 metamorphic events. These new geochronological results are placed in a P-T-t diagram
27 constructed for the UGU, LGU, and PAU. Monazite sampled in UGU, LGU, PAU rocks
28 yields similar 365-350 Ma ages consistent with the D₂ event dated in other places of the
29 French Massif Central. A zoned monazite grain from a granulitic paragneiss yields 416±15
30 Ma and 362±14 Ma ages interpreted as those of the D₀ and D₂ events, respectively. Zircon
31 from the same granulitic paragneiss yields SIMS ages at 343±2 Ma, and 328±2 Ma that are
32 interpreted as recrystallization processes associated with post-thickening thermal events,
33 possibly recording the onset of orogenic collapse of the Northern Massif Central. It is worth
34 to note that neither monazite nor zircon recorded the D₁ event.

35

36 **Key words** Variscan orogeny, monazite chemistry, monazite U-Th-Pb chemical dating,
37 zircon SIMS geochronology, granulite facies, monazite chemistry, French Massif Central

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41 **Introduction**

42 Dating **the syn-**metamorphic rocks and their related ductile fabrics remains a challenge.

43 **Indeed, many** of the common metamorphic phases such as biotite, muscovite, feldspar, and

44 amphibole can be dated by the $^{40}\text{Ar}/^{39}\text{Ar}$ method, however it is well acknowledged that this

45 chronometer is highly sensitive **to any late thermal perturbation**, thus the radiometric ages

46 obtained by this method **may represent cooling ages or "re-opening" ages rather than genuine**

47 crystallization age of the dated minerals (**see Jourdan et al 2014 for a recent review**). Zircon is

48 also a common **accessory** phase used for U-Pb geochronology. Due to analytical

49 improvements, **punctual** measurements using SIMS, SHRIMP or LA-ICP-MS technique,

50 zircon yields precise age, however as for the $^{40}\text{Ar}/^{39}\text{Ar}$ method, the significance of zircon U-

51 Pb geochronology must be carefully considered for dating metamorphic rocks since zircon

52 may crystallize or recrystallize at various stages of the rock history, including initial

53 magmatism formation stage for igneous protoliths. Monazite that may crystallize either from

54 magma or during metamorphic reactions appears as one of the **most useful** phase to assess the

55 timing of **tectono-**metamorphic events. **Although separated monazite grains can be dated by**

56 **SIMS or LA-ICP-MS**, the *in-situ* approach that preserves the textural relationships **between**

57 **minerals** allows the deciphering of superimposed magmatic and/or metamorphic high-

58 temperature stages. **Furthermore**, the size of the analytical spot (*ca* 20 μm) for LA-ICP MS

59 measurements generally prevents this detailed investigation **as metamorphic monazite is often**

60 **small-sized, chemically zoned or partly dissolved**. Nevertheless, though involving a non-

61 isotopic technique, **with a *ca* 5 to 15 Ma uncertainty, larger than the isotopic methods**, the

62 monazite U-Th-Pb EPMA method can provide one of the best estimate of the metamorphic

63 mineral crystallization ages in polymetamorphic terranes.

64 **Due** to the abundance of metapelitic and felsic magmatic protoliths, the Variscan belt

65 appears as a good example to determine the ages of successive tectono-metamorphic events,

66 in polymetamorphosed and polydeformed domains, particularly when thermo-barometric
67 conditions preclude the use of temperature-sensitive geochronological methods such as
68 $^{40}\text{Ar}/^{39}\text{Ar}$. This paper aims to apply the monazite U-Th-Pb EPMA chemical, and zircon U-Pb
69 SIMS dating methods to the Sioule metamorphic series, in the Northern part of the Variscan
70 French Massif Central since **in this area**, the tectonic-metamorphic succession is **already** well
71 established, but radiometric constraints still rare, **as presented in the following sections (e.g.**
72 **Faure et al. 1993, 2002; Schulz et al. 2001; Schulz 2009).**

73

74

75 **Geological outline of the Variscan French Massif Central**

76 The stack of nappes

77 The French Massif Central (FMC thereafter) is an important segment of the Variscan Belt that
78 results of a complex interplay between two main continents, namely Gondwana, and
79 Laurussia, and several intervening microcontinents such as Avalonia, Mid-German
80 Crystalline Rise, and Armorica (e.g. Matte 2001; Cocks 2000; Faure et al. 2005; Murphy et al.
81 2011). It is now well acknowledged that the bulk architecture of the French Massif Central,
82 alike the southern part of the Massif Armoricain or the Vosges-Black Forest massif, is a stack
83 of metamorphic nappes developed diachronously from Late Silurian to Early Carboniferous at
84 the expense of the north Gondwana margin (Matte 1986; Ledru et al. 1989; Faure et al. 2005,
85 2009; Lardeaux et al. 2014; Fig . 1).

86 From bottom to top, and South to North, the following units are recognized (Figs. 1, 2). **1)**

87 *The foreland basin* develops in the southernmost part of the Massif Central. It is an **Early**
88 Carboniferous (Visean-Serpukhovian) turbiditic basin that extends southward below the
89 coastal plain of the Mediterranean Sea. **2) The Paleozoic Fold-and-Thrust Belt** consists of
90 Early Cambrian to Visean unmetamorphosed sedimentary rocks displaced to the South as

91 thrust sheets or kilometre-scale recumbent folds well. There, the Early Devonian (*i. e.*
92 **Lochkovian, 416-411 Ma**) unconformity argues for an Early Paleozoic tectonic event (at ca
93 420-400 Ma), which is also recorded to the North of the Massif Central in the Lower and
94 Upper Gneiss metamorphic units (cf. below). **3) The Para-autochthonous Unit that**
95 overthrusts the Paleozoic Fold-and-Thrust Belt is formed by greenschist to epidote-
96 amphibolite facies metapelites, metagreywackes, quartzites, and a small amount of marbles
97 and metabasites. In the northern Massif Central, the Para-autochthonous Unit crops out in
98 tectonic windows surrounded by the Lower Gneiss Unit. In the Cévennes area, some granitic
99 plutons, now transformed into orthogneiss, intruding the series, yield Cambrian to Early
100 Ordovician zircons (Bé Mézème et al. 2006). **4) The Lower Gneiss Unit (LGU)** is composed
101 of metagreywackes, metapelites, and metarhyolites intruded by Cambrian to Early Ordovician
102 porphyritic alkaline granitoids transformed into augen orthogneiss during the tectonic-
103 metamorphic events. The LGU underwent a Middle Devonian metamorphism leading to
104 crustal melting coeval with a ductile shearing dated in the Limousin at 375-370 Ma (Faure et
105 al. 2008) and attributed to the D₁ event (cf below). Moreover, the LGU also experienced a
106 medium pressure/medium temperature metamorphism characterized by a biotite-garnet-
107 staurolite assemblage, dated at ca. 360-350 Ma (Costa 1989; Melleton et al. 2009), and
108 attributed to the D₂ event. **5) The Upper Gneiss Unit (UGU)** forms the overlying nappe whose
109 the **main protoliths are metapelites and metagrauwackes**. This unit contains high-pressure
110 rocks, such as eclogites or **high pressure (HP)** granulitic orthogneiss that may locally reach the
111 coesite-eclogite facies (Santallier 1981; Pin and Vielzeuf 1988; Lardeaux et al. 2001; **Berger**
112 **et al., 2010**). The upper part of the UGU consists of Devonian migmatites, yielding zircon U-
113 Pb ages **at ca** 385-380 Ma, attributed to the D₁ event, and formed by the partial melting of
114 pelitic and quartzo-feldspathic rocks (Duthou et al. 1994; Faure et al. 2008). These rocks
115 contain amphibolite blocks derived from retrogressed eclogites that did not experience partial

116 melting. The HP metamorphism is dated at ca 420-400 Ma by U-Pb and Sm-Nd methods in
117 several locations (Pin and Peucat 1986; Paquette et al. 1995; Faure et al. 2005).

118 The five units described above represent the classical nappe pile of the FMC in which the
119 metamorphism increases from bottom to top. Furthermore, in the northern part of the French
120 Massif Central, the undeformed and unmetamorphosed “Tuffs anthracifères” series, which
121 consists of terrigenous rocks (sandstone, shales, conglomerates), coal measures, and felsic
122 magmatic rocks (dacites and rhyolites), is dated as Late Viséan (*ca* 330 Ma, Bruguier et al.
123 1998). These rocks unconformably cover Late Devonian-Early Carboniferous plutonic or
124 metamorphic rocks of the UGU and LGU. The “Tuffs anthracifères” series provides a major
125 time constraint for the evolution of the French Massif Central. As the “Tuffs anthracifères”
126 series is coeval with the development of the Fold-and-Thrust Belt, it supports the north-south
127 diachronism in the deformation at the scale of the entire massif.

128

129 The polyphase tectono-metamorphic evolution

130 The Variscan geodynamic evolution of the French Massif Central remains disputed. Some
131 authors (e.g. Matte 1986; Ledru et al. 1989) consider that the formation of the FMC results of
132 a continuous north-directed **oceanic then continental** subduction from early Silurian to early
133 Carboniferous. **In contrast**, other authors (e.g. Pin 1990; Faure et al. 1997, 2008, 2009) argue
134 for a polycyclic evolution. The Early Silurian to Early Devonian events correspond to the Eo-
135 Variscan cycle due the collision of the Armorica microcontinent with the north Gondwana
136 margin. From Middle Devonian to Early Carboniferous, the Variscan belt develops due to the
137 southward subduction of the Rheic Ocean below Gondwana followed by the collision of
138 Gondwana with Laurussia. In spite of differences between the two geodynamic scenarii, most
139 of authors agree on the main tectonic, metamorphic, and magmatic events. Six tectono-
140 metamorphic events are recognized (Faure et al. 2009).

141 The earliest event, D₀, corresponds to the high to ultra high-pressure metamorphism
142 recorded in the eclogite facies rocks of the UGU. Most of the HP rocks derive from mafic
143 protoliths, but also HP relics are also found in orthogneiss are locally found. In the Lyonnais
144 area, the UGU eclogites record pressure and temperature conditions of 1.8-2 GPa and 650-
145 750°C, respectively (Lardeaux et al. 2001). On the basis of the HP metamorphism, the D₀
146 event took place around 415 Ma (Pin and Peucat 1986).

147 The Early to Middle Devonian, tectonic-metamorphic event D₁, is coeval with the
148 crustal melting of the metapelites and orthogneiss of the UGU and LGU, whereas mafic rocks
149 are preserved as amphibolite restitic blocks in migmatites. The migmatisation is dated as
150 Middle Devonian (385-375 Ma) in the Limousin or Lyonnais areas (Duthou et al. 1994, Faure
151 et al. 2008 and enclosed references). P-T estimates from garnet-plagioclase and garnet-biotite
152 pairs in migmatites, indicate metamorphic conditions of 0.7±0.05 GPa and 700±50°C
153 respectively (Mercier et al. 1991; Roig and Faure 2000). The flat-lying foliation exhibits a
154 NE-SW striking (N30 to N60E) stretching lineation marked by fibrolitic sillimanite or biotite
155 in migmatites, and biotite or amphibole in amphibolites. Shear criteria indicate a top-to-the-
156 SW displacement (Ledru et al. 1989; Faure et al. 1997; Roig and Faure 2000). Amphibolite in
157 the northern Massif Central yields biotite and hornblende ⁴⁰Ar/³⁹Ar cooling ages ranging from
158 390 Ma to 381Ma (Costa and Maluski 1988; Boutin and Montigny 1993). In northeastern
159 Massif Central, or Morvan area, Middle Devonian unmetamorphosed and undeformed
160 sedimentary rocks unconformably overlying migmatites with eclogite relics suggest that, at
161 least in the northern part of the Massif Central, the HP rocks were already exhumed before the
162 Middle Devonian (Faure et al. 1997).

163 In the LGU, UGU, the dominant fabric is a NW-SE trending stretching lineation
164 developed either on flat-lying or steeply dipping foliations, which is attributed to the D₂ event
165 (Faure et al. 2005, 2008). Thermo-barometric constraints are well established in the Limousin

166 and Rouergue areas (Duguet et al. 2006; Bellot and Roig 2007). In the UGU, the syn-D₂
167 minerals, crystallized along the NW-SE lineation, indicate 0.7-1 GPa and 600-700°C
168 metamorphic conditions. Paragneiss belonging to the LGU yield nearly similar conditions of
169 0.8-1 GPa and 550-600°C. In the LGU, amphibolites that never experienced the high-pressure
170 event indicate 0.8-1 GPa and 700-800°C metamorphic conditions (Santallier 1981). The D₂,
171 event, characterised by a top-to-the-NW ductile shearing, is widely documented from the
172 southeast Massif Central up to the Southern part of the Massif Armoricain. This deformation
173 is stratigraphically dated as pre-Early Viséan (i.e. older than 345 Ma) by the unconformity
174 recognized in the Brévenne area (e.g. Leloix et al., 1999 and enclosed references, Fig. 1). The
175 age of the syn-D₂ MP/MT metamorphism is constrained by ⁴⁰Ar/³⁹Ar method on biotite,
176 muscovite and amphibole, and by EPMA on monazite between 360 and 350 Ma (Costa 1989;
177 Leloix et al. 1999; Melleton et al. 2009).

178 The syn-metamorphic D₃ deformation is recognized only in the Para-autochthonous
179 Unit of the southern part of the Massif Central that experienced a first deformation at *ca* 340-
180 335 Ma (Caron 1994; Faure et al. 2005). Further south, south-verging recumbent folds and
181 thrusts of the Paleozoic series dated of Late Viséan-Early Namurian (*ca* 325 Ma) by the syn-
182 orogenic sediments of the foreland basin, are also related to the D₃ event. This progressive
183 southward younging of the nappe tectonics is coeval with the development of a flat-lying
184 foliation and a N-S to NE-SW striking stretching lineation. In contrast to the southern part of
185 the Massif Central, in its northern part, the Late Viséan deformation develops under brittle
186 conditions. It is represented by a NW-SE stretching, related to the emplacement of NE-SW
187 striking dykes of basalt, dacite, and microgranite in the Tuff Anthracifères series. This
188 deformation is interpreted as the onset of the syn-orogenic extension (Faure 1995).

189 Ductile normal faults, such as Argentat or Nantiat faults, coeval with Serpukhovian-
190 Bashkirian (or Namurian-Westphalian in W. European stratigraphy, *i.e.* 325-315 Ma)

191 synkinematic granite emplacement, argue for an extensional tectonic phase named the D₄
192 event (Faure 1995). As shown by petrofabric and AMS studies, most of the Late
193 Carboniferous leucogranite and porphyritic monzogranite plutons and their metamorphic
194 aureoles exhibit a well-defined NW-SE trending mineral or stretching lineation (Talbot et al.
195 2005; Joly et al. 2007). The tectonic-plutonic D₄ event, widespread in the entire Massif
196 Central, is interpreted as a syn-orogenic extensional process since at this time, compression
197 was still active in the northern and southern outer zones of the Variscan belt, such as Ardenne
198 and Montagne Noire-Pyrénées, respectively.

199 The last phase in the structural evolution of the French Massif Central is a Late
200 Carboniferous (Kasimovian-Ghzelian, or Stephanian) D₅ late-orogenic extensional event that
201 is distinguished from the previous one by the NNE-SSW strike of the maximum stretching
202 direction. Intramontane coal basins develop either as half-grabens or left-lateral pull-aparts
203 depending on the strike and kinematics of the bounding fault that controls the basin opening
204 and sedimentary infill.

205

206 **The Sioule metamorphic series**

207 The bulk architecture

208 Located in the northern part of the French Massif Central, the Sioule metamorphic series is
209 bound to the west by the Late Carboniferous Sillon Houiller sinistral wrench fault, and to the
210 east by the Oligocene Ebreuil and Limagne grabens (Figs. 1, 2). The Tréban Massif and the
211 S^{te}-Christine wrench fault are the northern and southern limits of the Sioule metamorphic
212 series, respectively (Fig. 2). The metamorphic series is intruded by two granitic plutons. The
213 Pouzol-Servant porphyritic microgranite is a biotite granite dated at *ca* 330 Ma (Pin, 1991).
214 The Echassières leucogranite is a small, differentiated, peraluminous pluton dated *at ca* 310
215 Ma (Duthou and Pin 1987; Cheillets et al. 1992). The Sioule series is a classic example of an
216 inverted metamorphic sequence in the Massif Central (Grolier 1971). Three lithological and

217 structural units have been described (Richard 1938; Grolier 1971; Faure et al. 1993). From top
218 to bottom, they are: a) an upper unit composed of biotite-sillimanite gneiss, and biotite-
219 sillimanite± cordierite migmatite developed at the expense of the underlying gneiss that
220 enclose a few meter-sized boudins of kyanite-garnet-plagioclase high-pressure granulitic
221 paragneiss partly retrogressed under amphibolite facies conditions (Ravier and Chenevoy
222 1979). This unit is correlated with the Upper Gneiss Unit (UGU) of the French Massif Central
223 (cf section 2); b) an intermediate unit formed by biotite-muscovite-sillimanite-K feldspar-
224 garnet paragneiss corresponding to the Lower Gneiss Unit (LGU), and c) a biotite-staurolite-
225 garnet micaschist that crops out around the Echassières pluton. This unit, which is structurally
226 below the two micas paragneiss of the LGU, **lacks of orthogneiss and anatectic melts, it is**
227 **therefore** correlated with the Para-autochthonous Unit (PAU).

228

229 Structural outline of the Sioule metamorphic series

230 In the study area, the observed structures, and the related kinematics account well for a
231 multistage tectonic evolution. In agreement with previous works (Grolier 1971; Feybesse and
232 Teygey 1987; Faure et al. 1993; Schulz et al. 2001; Schulz 2009), the foliation is folded by
233 NNW-SSE trending km-scale upright synform and antiform (Fig.2). A WNW-ESE striking
234 stretching and mineral lineation is well developed in the foliation plane of the PAU, LGU, and
235 UGU. Microscope-scale shear criteria, such as sigmoidal micas, asymmetric pressure shadows
236 around feldspar or garnet, and S-shaped inclusion trails in garnet consistently indicate a top-
237 to-the WNW shearing (Faure et al. 1993; Schulz 2009). These microstructural features are
238 relevant to the D₂ event of the Massif Central.

239 In the UGU, in addition to the main NW-SE D₂ deformation, a NNE-SSW striking
240 mineral and stretching lineation is also recognized (Grolier 1971; Feybesse and Teygey 1987;
241 Faure et al. 1993). This deformation, characterized by elongated biotite-sillimanite

242 aggregates, stretched K-feldspar, and axes of intrafolial folds parallel to the stretching
243 lineation, is associated with top-to-the SSW kinematic indicators. A Barrovian-type
244 metamorphism with biotite-sillimanite-K-feldspar assemblage is coeval with this deformation
245 event that appears less developed, and older, than the NW-SE D₂ one. In the upper part of the
246 UGU, the biotite-sillimanite±cordierite migmatite exhibits also the NNE-SSW mineral and
247 stretching lineation, with similar top-to-the-SW kinematics as those observed in the
248 underlying biotite-sillimanite gneiss. By correlation with the general tectonic-metamorphic
249 pattern of the FMC, this early phase is correlated with the D₁ event, thus a Middle Devonian
250 age, dated from 385 to 375 Ma is suggested (Faure et al. 2009) but not documented yet.

251 Furthermore, top-to-the-WNW and top-to-the-ESE shearing, characterized by shear
252 bands cutting through the regional foliation, sericite pressure shadows upon cataclazed
253 staurolite or garnet porphyroclasts, develop around the Echassières pluton. These kinematic
254 indicators, coeval with greenschist facies metamorphic conditions, are related to the
255 emplacement of the Echassière leucogranite pluton in an extensional setting. Thus this
256 deformation phase corresponds to the D₄ event (Faure et al. 2009).

257 In contrast with other areas of the French Massif Central, mafic eclogites formed
258 during the earliest D₀ event are not recognized in the Sioule area. Nevertheless, kyanite-
259 garnet-plagioclase-rutile high-pressure granulitic gneiss, garnet-pyroxenite, and rare
260 serpentinite boudins are mapped in the UGU (Grolier 1971; Ravier and Chenevoy 1979).
261 These rocks argue for an early HP event probably equivalent to the D₀ one. Moreover, in
262 paragneiss, as shown by the symplectitic texture, these HP garnet and kyanite grains are
263 surrounded by a biotite-sillimanite or muscovite assemblage representative of an amphibolite
264 facies retrogression developed during the D₁ or D₂ events (Fig. 3). As in most areas of the
265 Massif Central, the planar and linear fabrics related to the D₀ event have been erased by
266 subsequent metamorphic recrystallization or deformation events. But since syn-D₀ minerals,

267 such as kyanite, core of garnet, and possibly monazite are preserved, the rocks may still be
268 suitable to yield the age of the HP event.

269 Available time constraints in the Sioule metamorphic series are currently limited
270 (Faure et al., 2002; Schulz, 2009). The Late Visean Pouzol-Servant microgranite and the
271 Bashkirian-Moscovian Echassières leucogranite intrude the already metamorphosed and
272 ductilely deformed series, and thus provide the upper time limit for the D₀, D₁, and D₂
273 tectono-metamorphic events. An ⁴⁰Ar/³⁹Ar survey has been carried out in the metamorphic
274 units of the Sioule series (Faure et al. 2002). All the minerals, whatever the rock type or the
275 tectonic unit, yielded similar ca 335-330 Ma ages that were interpreted as the Late Visean
276 thermal overprint since these ages are similar, within the errors, to the age of the “Tuffs
277 Anthracifères” event. Thus, the ages of the different pre-Late Visean tectono-metamorphic
278 events in the Sioule series remain presently unknown.

279

280 **Sampling and petrological description**

281 Twenty-four samples representative of the different metamorphic assemblages cropping out in
282 the Sioule series have been petrologically studied in order to bring the time constraints on the
283 metamorphic evolution of this series. After detailed microscopic observations, five samples
284 were selected due to their suitable parageneses and textures to ensure the presence of
285 accessory minerals such as monazite and zircon developed during the D₁ and D₂ tectono-
286 metamorphic events. The sample location in the Sioule nappe pile, and the GPS position, are
287 provided in Figure 2 and Table 1, respectively. The PAU is represented by a staurolite-garnet
288 micaschist (SO3) located to the East of the Echassières pluton. The LGU is represented by a
289 two-mica paragneiss (SO17) to the SE of the same pluton. The UGU is represented by a
290 biotite-sillimanite gneiss (SO7), a migmatitic gneiss (SO11), and a biotite-kyanite-garnet
291 paragneiss (SO15), located on both limbs of the antiform. The microstructural descriptions

292 herein refer to thin sections cut parallel to the stretching lineation (X) and perpendicular to the
293 foliation plane (Z), i.e. in the XZ plane of the finite strain ellipsoid.

294 *Sample SO3*, belonging to the Para-autochthonous unit, consists of quartz, feldspar,
295 biotite, muscovite, garnet, staurolite, sericite, and chlorite. In the field, structural observations
296 show a low-angle ESE plunging foliation, and a N105E to N115E striking mineral lineation
297 (Fig. 3a2). The foliation cut by shear bands, and sigmoidal quartz veins indicate a top-to-the-
298 SE sense of shear (Fig. 3a1). Under the microscope, cataclazed garnet and staurolite
299 surrounded by an asymmetric mantle of sericite developed during retrogression of the main
300 assemblage indicate the same sense of shear (Fig 4a). Furthermore, garnet exhibits sigmoidal
301 inclusions of quartz and opaque minerals indicating an earlier foliation. The asymmetry of
302 these inclusions indicates a synkinematic crystallisation of these porphyroblasts during a
303 noncoaxial deformation that is not in agreement with the top-to-the-SE sense of shear, but
304 instead argues for a top-to-WNW sense of shear developed during the D₂ event. **According to**
305 **Schulz et al (2001)**, the thermobarometry from the PAU micaschist indicates a prograde P-T
306 evolution with increasing temperatures from 450°C to 600°C under slightly increasing
307 pressure from 0.7 to 0.8 GPa, followed by a P and T decrease indicated by the formation of
308 retrogressive sericite during the decompression. Scan Electron Microscope (SEM)
309 backscattered electron (**BSE**) observation of sample SO3 reveals the presence of 50 to 100 µm
310 sized, homogeneous, monazite grains included into biotite or crystallized along grain
311 boundaries.

312 *Sample SO17* is a two-mica paragneiss presenting traces of incipient migmatization
313 belonging to the LGU with a segregation of thick quartz-feldspar and thin biotite and
314 sillimanite layers. The flat lying foliation contains a NW-SE striking mineral and stretching
315 lineation related to the D₂ event. The mineral assemblage consists of quartz, feldspar, biotite,
316 muscovite, chlorite, and accessory minerals including sillimanite, Fe-Ti-oxides, and monazite.

317 Muscovite often presents a clear secondary character, growing without preferred orientation
318 while biotite is partially altered into chlorite in line with a late thermal overprint (Faure et al.
319 2002). Biotite contains large subeuhedral monazite grains of *ca* 80-120 μm size. Monazite
320 also occurs as *ca* 80-100 μm euhedral inclusions in biotite or quartz. Despite the existence of
321 these two occurrences, in SEM BSE imaging, the monazite grains exhibit homogeneous
322 internal textures without zoning.

323 *Sample SO7* is a biotite-sillimanite migmatitic paragneiss belonging to the UGU
324 collected near Moulin du Bouchet (Fig. 2). This rock is composed of quartz, feldspar, garnet,
325 biotite, muscovite, rare staurolite and fibrolitic sillimanite. Field observations of coarse-
326 grained quartz-feldspar leucosome parallel to the foliation indicate that this rock underwent a
327 limited partial melting. The biotite-sillimanite assemblage defines a pervasive foliation and a
328 N50E striking mineral and stretching lineation that can be assigned to the regional D₁
329 deformation. However, as already pointed out by Schulz (2009), the subeuhedral shape of the
330 garnet argues for a static growth after the development of the D₁ structures. According to
331 Schulz et al. (2001), the biotite-sillimanite gneiss of Moulin de Bouchet, which is located
332 close to our sample SO7, underwent a prograde metamorphic evolution at low pressure (0.4-
333 0.6 GPa) and high temperature (650°C-700°C). Thus, this biotite-sillimanite paragneiss might
334 have also experienced the D₂ event, although typical D₂ NW-SE stretching lineation is not
335 observed. In this rock, all the analyzed monazite grains, observed in SEM BSE mode, are
336 homogeneous inclusions in biotite (Fig. 4b).

337 *Sample SO11*, collected at Pont de Menat (Fig. 2) is a migmatitic paragneiss belonging
338 to the upper part of the UGU. This rock is composed of alternations of coarse-grained quartz-
339 feldspar leucosome and fine-grained biotite-fibrolitic sillimanite-garnet melanosome. Field
340 observation shows the occurrence of a pervasive deformation characterized by a flat-lying
341 foliation and a N120E trending mineral and stretching lineation assigned to the D₂ event (Fig.

342 3 c). All the analyzed monazite grains observed in SEM BSE mode are homogeneous
343 inclusions in biotite (Fig. 4c).

344 *Sample SO15* is a kyanite-garnet \pm rutile paragneiss coming from Saugères (Fig. 2).
345 This rock is composed of biotite, feldspar, quartz, garnet, and kyanite. Garnet and kyanite are
346 1-5 mm sized clasts included in a well-foliated granoblastic matrix composed of quartz,
347 feldspar, and small biotite flakes (Fig. 3d). The kyanite-garnet assemblage that formed in
348 high-pressure granulite facies underwent an amphibolite facies retrogression as shown by
349 biotite overgrowths around garnet porphyroclasts (Ravier and Chenevoy 1979; Fig. 4d).
350 Monazite inclusions in garnet and in biotite exhibit a core-rim chemical zoning.

351

352 **Monazite EPMA chemical dating**

353 Analytical method and strategy

354 Due to its high U and Th contents, and negligible initial Pb content (Parrish 1990), it is
355 now well acknowledged that a crystallization (or recrystallization) age can be determined
356 from monazite on the basis of its chemical composition in U, Th, and Pb determined by
357 electron probe microanalysis (EPMA). Such a simple and fast technique has been successfully
358 applied to several plutonic or metamorphic rocks (e.g. Suzuki and Adachi, 1991; Montel et
359 al., 1996; Cocherie et al., 1998, 2005; Cocherie and Albarède, 2001; Schulz et al., 2009; Faure
360 et al., 2010, and enclosed references). Monazite is a common accessory mineral in
361 metamorphic rocks, especially in metapelites. It crystallizes in a wide range of P-T conditions,
362 from greenschist facies (Cabella et al. 2001; Rasmussen et al. 2001) to amphibolite (Smith
363 and Barreiro 1990; Kohn and Malloy 2004) and granulite facies (Bé Mézème et al. 2006; Mc
364 Farlane and Frost 2009) in rocks metamorphosed under Barrovian P-T gradient during
365 prograde or retrograde paths. Previous studies have shown also that monazite can crystallize
366 in blueschist (Janots et al. 2006) or even eclogite facies (Nagy et al. 2002). Consequently,

367 analyzed monazite grains must be linked with petrological studies to delimit their P-T
368 crystallization domain. The chemical composition of monazite is an important parameter in
369 order to discuss the radiometric ages.

370 *In situ* chemical U-Th-Pb dating of monazite has been developed over the past fifteen
371 years allowed by technological progress performed on electron microprobe analyses (EPMA;
372 Suzuki and Adachi 1991; Montel et al. 1996; Cocherie et al. 1998, 2005). The method uses
373 the radiochronometer between U and Th parent and Pb radiogenic elements, considering the
374 restricted lead diffusion in monazite lattice below 900°C (Cocherie et al. 1998). The
375 determination of U, Th and Pb contents can be performed directly by EPMA *in situ* analyses
376 on polished thin section after textural optical microscope observation of crystallisation-
377 deformation relationships between the constitutive minerals, and monazite internal structure
378 determination using a SEM in BSE mode. This approach provides petrological constraints on
379 the analyzed grains and thus allows us to discuss the significance of the calculated ages in the
380 crystallisation-deformation history of the rock.

381 SEM BSE was used to define heterogeneous chemical compositional domains of
382 monazite. In metamorphic rocks, contrasted chemical domains can be explained by the
383 superimposition of metamorphic events that record a polyphase evolution during orogenic
384 processes. Theoretical procedure to reduce the data and to calculate the average age from
385 individual spot analyses is given by Cocherie and Albarède (2001), Cocherie et al. (2005), and
386 Cocherie and Legendre (2007). The analysis of monazite grains directly from polished thin
387 sections was performed by a Cameca SX 50 EPMA co-operated by BRGM-CNRS-Orléans
388 University. An accelerating voltage of 20 kV, and a beam current of 100 nA were chosen as
389 operating conditions. Synthetic phosphates, glasses and natural minerals were used as
390 standards. The counting times (2/3 on peak and 1/3 on background) varies from 20s for Si, P,
391 Ca, Nd, Gd, Y, Th, Ce and La, to 100s for U and 120s for Pb. Considering such parameters,

392 the detection limit for Th, U and Pb is 150 ppm (Cocherie and Albarede, 2001). X-ray
393 element maps were also recorded with an accelerating voltage of 20 kV and beam current of
394 100 nA, using the wavelength-dispersive spectrometers. Details on the analytical procedure
395 are the same as those used in Cocherie et al. (2005) and Faure et al. (2008, 2010). Age
396 calculations were done using the Isoplot/Ex (2.49) program of Ludwig (2001) and a Microsoft
397 Excel add-in program for determining U-Th-Pb ages from EPMA measurements (Pommier et
398 al. 2002). The 2σ errors given on individual ages depend on U, Th, and Pb contents, and are
399 calculated by propagating the uncertainties of these elements (at 95% confidence level).
400 Because the mean age of an homogeneous age population is directly derived from the isochron
401 plot Th/Pb vs U/Pb, only the uncertainty on these two parameters (Th/Pb and U/Pb) are
402 required to calculate the uncertainty on the mean age.

403 The strategy followed for this study was to derive time constraints for the monazite
404 included into the foliation forming biotite or in contact with it, in order to settle the
405 crystallisation time of the main foliation. Consequently, we concentrated on subeuhedral to
406 euhedral monazite grains. We focused on the dating of well-defined chemical domains so as
407 not to shuffle different heterogeneous domains that would yield different crystallization
408 periods.

409

410 Results

411 *Chemical composition of monazite grains.* SEM BSE images reveal chemically
412 homogeneous monazite grains for the four samples SO3, SO7, SO11, and SO17. Conversely,
413 the SO15 kyanite-garnet granulitic paragneiss sample presents zoned monazite grains. This
414 zonation corresponds to a core/rim zonation as described by Bé Mézème et al. (2006), mostly
415 due to the spatial distribution of U, Th and Pb. The EPMA analysis allows us to quantitatively
416 characterize the chemical composition of the dated grains. The average U/Pb and Th/Pb ratio

417 values, details concerning the number of spot analyses, and analyzed grains are provided in
418 Table 1. **Representative monazite chemical compositions with uncertainties for samples SO3,**
419 **SO7, SO11, SO15, and SO17 are given in Table 2.** For the granulitic gneiss (sample SO15),
420 core and rim chemical data have been separated. Monazite chemical data for each samples are
421 plotted in the Monazite (2REEPO_4) - Cheralite ($\text{CaTh}(\text{PO}_4)_2$) - Huttonite (2ThSiO_4) diagram
422 (Linthout 2007; Fig. 5a), in order to show the relative substitution between **end-member**
423 monazite phase and the cheralite or huttonite components (Fig. 5a). The cheralite substitution
424 is always low, as already observed for various metamorphic environments (e.g. Franz et al.
425 1996; Cabella et al. 2001), and huttonite substitution is absent or negligible. In samples SO3,
426 SO7, SO11 and SO17, monazite grains appear as homogeneous domains in BSE images, and
427 the analytical data scatter along the cheralite-monazite axis showing clearly a common
428 substitution. In sample SO15, the analyzed monazite grain is composed of two domains in the
429 BSE images, with a light (i.e. dense) core and a dark (i.e. light) rim. Data plotted in the
430 Monazite (2REEPO_4) - Cheralite ($\text{CaTh}(\text{PO}_4)_2$) - Huttonite (2ThSiO_4) diagram (Fig. 5a) show
431 a relatively high Ce-monazite component into the core of the grain (**white** diamond) and a
432 cheralite enriched rim (**black** diamond), corresponding to two different chemical domains.
433 The $(\text{Th}+\text{U}+\text{Si}+\text{Pb}) = f(\text{REE}+\text{P})$ plot, modified from Franz et al. (1996), confirms the spread
434 of the analytical data for samples SO3, SO7, and SO11 along the CaThREE_2 axis
435 corresponding to the cheralite substitution (Fig. 5b). Monazite chemical data from the
436 staurolite micaschist SO3 (Fig. 5b) indicate a limited huttonite substitution pointed out by
437 small vertical variation in $\text{Th}+\text{U}+\text{Si}+\text{Pb}$. Chemical data of monazite from biotite-sillimanite
438 gneiss SO7 and SO11 show a more significant huttonite substitution than for SO3. The core
439 and the rim of dated monazite grain from sample SO15 are characterized by two distinct sets
440 of chemical compositions. The rim is richer in cheralite component than the core, even if
441 huttonite substitution in each domain seems to be similar (Fig. 5b). X-ray maps depicting the

442 semiquantitative distribution of Th, Ca, and U confirm the chemical differences between the
443 core and the rim of SO15 grain (Fig. 5d). These maps show that the monazite rim of monazite
444 is enriched in Th and Ca compared to the core. The U content is relatively equivalent in the
445 core and rim, except for an enriched fringe in the rim. This domain was not analyzed because
446 of holes on the polished surface. Despite the presence of several small-sized domains
447 indicated by the Ca and Th contents, the entire core is homogeneous in Th and U
448 composition.

449
450 *U-Th-Pb chemical dating.* In this section, the ages of the analyzed monazite grains
451 from samples SO3, SO 17, SO7, SO11, SO15 are presented following the tectonic order from
452 bottom to top (Fig. 2).

453 Staurolite micaschist SO3. Data points are highly scattered on the Th/Pb vs. U/Pb
454 diagram, leading to a small error envelope. The intercept ages are similar within errors, since
455 the U-Pb age (intercept with U/Pb axis), and the Th-Pb age (intercept with Th/Pb axis) are at
456 $379 +54/-62$ Ma, and $357+26/-23$ Ma, respectively. The calculated MSWD of 1.3 indicates
457 that the population of 73 analyses is statistically homogeneous within the analytical errors
458 (Wendt and Carl 1991). As a consequence, it is possible to calculate the mean age at the
459 centroid where the precision is the best within the error envelope. Thus, the monazite grains
460 yield a mean age of 363 ± 8 Ma (at 2σ confidence level; Fig. 6a).

461 Two-mica gneiss (SO17). Data points are weakly scattered on the Th/Pb vs. U/Pb
462 diagram, leading to a large error envelope. The intercept ages are similar within errors, since
463 the **intercept** U-Pb age and the **intercept** Th-Pb age are at $399 +44/-53$ Ma, and $328+31/-26$
464 Ma, respectively. The calculated MSWD of 0.63 indicates that the population of 38 analyses
465 is statistically homogeneous within the analytical errors (Wendt and Carl 1991). As a
466 consequence, it is possible to calculate the mean age at the centroid where the precision is the

467 best within the error envelope. Thus, the monazite grains yield a mean age of 351 ± 5 Ma (at
468 2σ confidence level; Fig. 6b).

469 Biotite-sillimanite gneiss (SO7). Data points are **relatively** scattered on the Th/Pb vs.
470 U/Pb diagram, leading to a **small** error envelope. The intercept ages are similar within errors,
471 since the **intercept** U-Pb age and the **intercept** Th-Pb age are at $331+39/-46$ Ma, and $371+33/-$
472 28 Ma, respectively. The calculated MSWD of 1.19 indicates that the population of 73
473 analyses is statistically homogeneous within the analytical errors (Wendt and Carl 1991) **and**
474 **as** a consequence, it is possible to calculate the mean age at the centroid. Thus, the monazite
475 grains yield a mean age of 355 ± 7 Ma (at 2σ confidence level; Fig. 6c).

476 Migmatitic gneiss (SO11) of Pont de Menat. Data points on the Th/Pb vs. U/Pb
477 diagram are less scattered than for previous sample, **which may** lead to a larger error envelop.
478 **However**, the large number of analyses ($n= 174$) decreases the envelop size. The intercept
479 ages are similar within errors, since the **intercept** U-Pb age and the **intercept** Th-Pb age are at
480 $342+45/-51$ Ma, and $373+24/-21$ Ma, respectively. The calculated MSWD of 0.98 indicates
481 that the population of 174 analyses is statistically homogeneous within the analytical errors
482 (Wendt and Carl 1991). As a consequence, it is possible to calculate the mean age at the
483 centroid. Thus, the monazite grains yield a mean age of 363 ± 4 Ma (at 2σ confidence level;
484 Fig. 6d).

485 Kyanite-garnet granulitic paragneiss (SO15) of Saugères. **Monazite** in sample SO15
486 exhibits a core-rim zonation of the grain, **and in consequence**, EPMA data were processed
487 separately for the different domains in order to calculate two isochron ages. 47 and 21
488 analyses were performed in the core and rim of a single grain, respectively. The 47 analyses
489 from the core on the Th/Pb vs. U/Pb diagram are **clustered**, leading to a large error envelope.
490 The intercept ages are similar within errors, since the **intercept** U-Pb age and the **intercept** Th-
491 Pb age are at $517+281/-322$ Ma, and $390+81/-57$ Ma, respectively. The calculated MSWD of

492 1.15 indicates that the population of the 47 analyses is statistically homogeneous within the
493 analytical errors (Wendt and Carl 1991). As a consequence, it is possible to calculate a mean
494 age at the centroid. Thus, the core of this monazite grain yields a mean age of 416 ± 15 Ma (at
495 2σ confidence level; Fig. 6e).

496 Moreover, the 21 analyses performed from the monazite rim are well-scattered,
497 leading to a small error envelope. The intercept ages are similar within errors, since the
498 intercept U-Pb age and the intercept Th-Pb age are at $423+74/-88$ Ma, and $338+36/-30$ Ma,
499 respectively. The calculated MSWD of 0.93 indicates that the population of 21 analyses is
500 statistically homogeneous within the analytical errors (Wendt and Carl 1991). As a
501 consequence, it is possible to calculate the mean age at the centroid where the precision is the
502 best within the error envelope. Thus, the monazite grains yield a mean age of 362 ± 14 Ma (at
503 2σ confidence level; Fig. 6f).

504

505 **Zircon U-Pb SIMS dating**

506 Analytical procedure

507 Zircon concentrates were separated from a *ca* 5 kg of the granulitic paragneiss sample SO15
508 using standard density and magnetic separation techniques. Zircon grains, together with
509 standard zircons TEMORA 2, 91500, and an in-house standard zircon Qinghu, were mounted
510 in an epoxy mount that was then ground and polished to section the crystals in half for
511 analysis. Zircons were imaged with transmitted and reflected light photomicrographs as well
512 as cathodoluminescence (CL) to reveal their internal structures, and the mount was vacuum-
513 coated with high-purity gold prior to SIMS analyses. Measurements of U, Th and Pb isotopes
514 were conducted using a Cameca IMS 1280 SIMS at the Institute of Geology and Geophysics,
515 Chinese Academy of Sciences in Beijing. Analytical procedures are similar to those reported
516 by Li et al. (2009). The primary O_2^- ion beam spot is about 20×30 μm in size. Positive

517 secondary ions were extracted with a 10 kV potential. In the secondary ion beam optics, a 60
518 eV energy window was used, together with a mass resolution of *ca* 5400 (at 10% peak
519 height), to separate Pb⁺ peaks from isobaric interferences. A single electron multiplier was
520 used in ion-counting mode to measure secondary ion beam intensities by peak jumping mode.
521 Analyses of the standard zircon TEMORA 2 were interspersed with unknown grains. Each
522 measurement consists of 7 cycles. Pb/U calibration was performed relative to zircon standard
523 TEMORA 2 (²⁰⁶Pb/²³⁸U age = 417 Ma, Black et al. 2004); U and Th concentrations were
524 calibrated against zircon standard 91500 (Th = 29 ppm, and U = 81 ppm, Wiedenbeck and
525 Allen 1995). Measured compositions were corrected for common Pb using non-radiogenic
526 ²⁰⁴Pb, and an average of crustal composition (Stacey and Kramers 1975) was used for the
527 common Pb. Data reduction was carried out using the Isoplot/Ex v. 2.49 program (Ludwig
528 2001). In order to monitor the external uncertainties of SIMS U-Pb zircon measurements, an
529 in-house standard zircon Qinghu was alternately analyzed as an unknown together with other
530 unknowns. Twenty measurements on Qinghu zircon (Table 3) yield a Concordia age of 160 ±
531 1 Ma (2SE, MSWD of concordance = 0.92), which is in good agreement within errors with
532 the recommended U-Pb age of 159.5 ± 0.2 Ma (Li et al. 2009). Zircon U-Pb age data are
533 presented in Table 3.

534

535 Results

536 **Fifty-five** zircon grains from the SO 15 granulitic paragneiss have been analyzed (Table 3).
537 **All but four zircons are concordant within analytical uncertainties (discordance degree <4%).**
538 **The dated zircons can be divided into two groups in terms of their morphological feature, ages**
539 **and Th/U ratio. In group 1, zircons are mostly sub-rounded to multifaceted grains and yield**
540 **ages between 324 and 349 Ma. They are characterized by low Th/U ratio (<0.1) and faint CL**
541 **luminescence (Table 3, Fig. 7), similar to metamorphic zircons (Corfu et al. 2003). Contrarily,**

542 Group 2 zircons are dated between 368 Ma and 1.99 Ga (Table 3). They are mostly oval
543 shaped grains showing concentric zoning under CL (Fig. 7), and relatively high Th/U ratio
544 (mostly >0.1), similar to the magmatic zircons (Corfu et al. 2003). Among them, four zircons
545 (spots @06, 26, 27 and 37) are highly discordant, with $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 1629 and
546 1994 Ma. On the probabilistic histogram of U-Pb ages for 51 concordant analyses (Fig 8),
547 Group 1 zircons form one major peak at 342 Ma ($n = 24$) and a subordinate peak at 327 Ma (n
548 = 6), corresponding to Concordia age of 343 ± 2 Ma and 328 ± 4 Ma, respectively, on a U-Pb
549 Concordia plot (Fig. 9). Thus, Group 1 zircons are most likely metamorphic in origin. The
550 remaining 21 Group 2 zircons display a scattered age pattern ranging from 370 to 800 Ma,
551 which might be detritus zircons of magmatic origin with or without later metamorphic and
552 hydrothermal overprinting.

553

554 **Discussion**

555 Many studies have pointed out the relationships between the age and the chemical
556 composition of monazite grains in polymetamorphic terranes (e. g. Martins et al., 2009; Kelly
557 et al., 2012). In samples SO3, SO7, SO11 and SO17, SEM BSE images of monazite grains
558 revealed homogeneous grains. The chemical compositions of the analysed monazite grains
559 cluster near the Monazite-(Ce) pole and scatter along the cheralite axis (Fig. 5a and b). The
560 weak data dispersal along the cheralite substitution line indicates that the analysed grains
561 represent a rather homogeneous population. Monazite grains from sample SO15 that exhibit a
562 light (i.e. dense) core and a dark (i.e. light) rim in BSE images, are scattered between the
563 cheralite ($\text{CaTh}(\text{PO}_4)_2$) and huttonite (2ThSiO_4) substitutions (Fig. 5b). Compared with the
564 other analysed samples, such a difference is explained by a high REE (La, Ce, Pr, Nd) content
565 and absence of Y in the core of monazite grains from sample SO15 (Table 2). The analysed
566 spots in both the core and rim parts delimit two clusters with different chemical composition.

567 It is noteworthy that i) the chemistry of the SO15 monazite rim is close to the chemistry of the
568 other analysed monazite grains, and ii) the reaction leading to the rim crystallisation occurred
569 contemporaneously to the crystallisation of the other monazite grains in samples SO3, SO7,
570 SO11 and SO17. Datings carried out in this study have been made on grain populations with
571 homogeneous chemical composition that make us confident about the age interpretation.

572 The kyanite-garnet granulitic gneiss SO15 preserves a core domain within monazite
573 yielding a *ca* 415 Ma age. This Early Devonian age can be assigned to the D₀ event
574 recognized in the UGU in several places in the Massif Central (Pin and Peucat 1986; Pin and
575 Vielzeuf 1988; Paquette et al. 1995). All the dated samples, whatever the tectonic unit to
576 which they belong, yielded similar within the error, monazite U-Th-Pb ages at *ca* 365-350 Ma
577 (Fig. 10). Since they share common structural features, including a similar NW-SE striking,
578 D₂, stretching lineation, which is well developed in samples SO3, SO11, SO17, and SO15, the
579 363-351 Ma ages yielded by the monazite in these rocks can be attributed to the D₂ tectono-
580 metamorphic event. Similar ages were already reported for the D₂ tectono-metamorphic event
581 in other areas of the French Massif Central, such as the Lot series (Costa 1989) or Limousin
582 (Melleton et al. 2009). Moreover, these ages are consistent, **within the errors**, with the
583 emplacement age of the Guéret and S^t-Gervais d'Auvergne plutons (Figs 1, 2) that **exhibit a**
584 **NW-SE striking magmatic lineation** (Faure et al. 2005; Cartannaz et al. 2007). These plutons,
585 dated at 360-350 Ma, **emplaced** at the end of the D₂ event (Berthier et al. 1979; Cartannaz et
586 al. 2007; Faure et al., 2009). The SO11 D₁ migmatite sample, deformed during the D₂ event,
587 yields also a monazite U-Th-Pb age at *ca* 363 Ma, which is consistent with the Late
588 Devonian-Early Carboniferous age of the D₂ event.

589 It is worth to note that although the D₁ migmatization is preserved in samples SO11,
590 and SO7, **the P-T conditions of 0.4-0.6 GPa and 550-600°C, calculated in the UGU from a**
591 **sample close to our sample SO7 (Schulz et al. 2001, Schulz 2009) are not in agreement with**

592 the D₁ event, as defined in other areas of the French Massif Central (e.g. Santallier 1981;
593 Mercier et al. 1991; Lardeaux et al. 2001; Bellot and Roig 2007), but instead represent a low
594 pressure-medium temperature event similar to the D₂ one as observed in the Limousin, Lot or
595 Rouergue areas (Duguet et al. 2006; Bellot and Roig 2007; Melleton et al 2009). Thus this
596 biotite-sillimanite paragneiss might have also experienced the D₂ event, even if the typical
597 NW-SE D₂ stretching lineation is not observed in this rock. Furthermore, the NE-SW striking
598 D₁ mineral lineation is clearly observed in sample SO17. In these three rocks, the analyzed
599 monazites do not record the Devonian migmatization dated between 385 Ma and 375 Ma in
600 Lyonnais (Duthou et al. 1994), Limousin (Faure et al. 2008) and S. Brittany (Cocherie et al.
601 2005). This result can be variously interpreted. Firstly, monazite was unable to crystallize
602 under the D₁ metamorphic conditions; **secondly, monazite effectively crystallized, but it may**
603 **have been dissolved during the HT event, and then neocrystallized in biotite during the**
604 **prograde D₂ event;** thirdly, due to the limited number of analyzed grains, our sampling missed
605 any monazite grain formed during the D₁ event. Though the first hypothesis appears unlikely,
606 since aluminous composition protoliths are suitable for monazite crystallization, the other two
607 explanations cannot be discarded. Thus, the 365-350 Ma monazite ages obtained from
608 samples SO3, SO17, SO7, SO11, and the monazite rim from sample SO15 are attributed to
609 the neocrystallization or the complete recrystallization of monazite during the D₂ event,
610 leading to the total loss of any early monazite. **In order to get a better understanding of the**
611 **thermo-barometric and chemical conditions that allow monazite crystallization or dissolution**
612 **with respect to other phases such as garnet and biotite, a detailed analysis of the chemical**
613 **elements exchanges during the succession of the crystallization-deformation events would be**
614 **necessary. On the contrary to Limousin or Maures areas (e. g. Duguet et al. 2006; Bellot and**
615 **Roig 2007; Oliot et al. 2015), such a study is not available yet for the Sioule metamorphic**
616 **series.**

617 Besides, zircon from the granulitic paragneiss yields U-Pb SIMS ages younger than
618 the monazite ones at 343 ± 2 Ma, and 328 ± 2 Ma. The latest is close to the whole-rock Rb/Sr
619 age of the Pouzol-Servant microgranite at 330 Ma (Duthou and Pin 1987; Pin 1991), and also
620 to the cooling ages yielded by the $^{40}\text{Ar}/^{39}\text{Ar}$ analyses on biotite, muscovite and amphibole for
621 the entire nappe-stack of the Sioule metamorphic series, either in the Upper Gneiss, Lower
622 Gneiss, and Para-autochthonous Units (Faure et al. 2002; Fig. 11). As pointed out in the
623 “Geological Setting” section, the *ca* 330 Ma age is that of the “Tufs anthracifères”
624 magmatism widespread in the northern part of the French Massif Central (Fig. 11). The
625 $^{40}\text{Ar}/^{39}\text{Ar}$ ages of the Sioule metamorphic rocks, younger than the Late Visean unconformity,
626 have been interpreted as a thermal reset provided by the high heat flux during the onset on the
627 extensional tectonics in the northern part of the French Massif Central. Thus we suggest here
628 that the zircon age at 328 ± 4 Ma might be due to recrystallizations processes, possibly
629 assisted by hydrothermal fluids, also as a consequence of the Late Visean thermal reset.

630 **Indeed, the LGU and PAU rocks exhibit several evidence of retrogression such as chlorite**
631 **developed at the expense of biotite, sericite surrounding staurolite, (Fig 4a) and albite**
632 **overprinting the foliation. These minerals argue for a late metamorphic stage associated with**
633 **fluid circulation. In the present state of knowledge, detailed petrological investigations to**
634 **characterize this hydrothermal event are not available.** The 343 ± 2 Ma age yielded by 24
635 zircon grains is more difficult to interpret in the geological framework. This age
636 corresponding to the majority of analyzed concordant grains is different, **within the error,**
637 from the 328 ± 4 Ma age. Furthermore, such a Middle Visean age is **rarely** reported in the
638 northern part of the Massif Central. **However, to the East of the study area, in the Lyonnais,**
639 **syn-kinematic granites emplaced along a NE-SW striking dextral strike-slip fault yield biotite**
640 **and muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ ages at *ca* 346 ± 3 Ma and 349 ± 3 Ma, respectively (Costa et al. 1993).**
641 **Such a NE-SW strike-slip shearing is not identified in the Sioule area, but the NW-SE striking**

642 **S^{te}-Christine dextral strike slip fault develops in the southern boundary of the Sioule series**
643 **(Fig. 2). One possible interpretation would be to relate this *ca* 343 Ma (i. e. Middle Viséan)**
644 **age to fluid circulations coeval with a transcurrent shearing event too. Importantly, the zircon**
645 **SIMS measurement did not recorded the 365-350 Ma D₂ ages.**

646 Previous monazite U-Th/Pb ages (Schulz 2009) range between 343 Ma and 327 Ma,
647 i.e. much younger than those reported in this study. These dates were attributed to two distinct
648 events, in agreement with the ⁴⁰Ar/³⁹Ar ages (Faure et al. 2002), the monazite chemical ages
649 younger than 340 Ma were considered as due to the Late Viséan **hydrothermal** event related to
650 the “Tufs anthracifères” magmatism. Conversely, ages older than 340 Ma were considered as
651 minimum ones associated to a top-to-the-SE tectonic event. However, structural
652 investigations demonstrated that the top-to-the-SE shearing was coeval with the emplacement
653 of the Echassières pluton at *ca* 310 Ma (Feybesse and Teygey 1987; Faure et al. 1993). The
654 discrepancy between our results and the previous ages by Schulz (2009) can also be explained
655 considering the data reduction method. The previous study uses frequency histograms to
656 process the analytical data associated with very large, routinely *ca* 15 Ma, errors. Moreover,
657 the frequency diagram method provides a mean age systematically younger than the one given
658 by the Th/Pb= f(U/Pb) isochron method (Cocherie and Albarède 2001). For example, the
659 **weighted average age** computed here for the monazite core and rim of sample SO15 (Fig. 5c)
660 also yields ages at 392±16 Ma, and 359±18 Ma younger than 416±8 Ma, and 362±14 Ma
661 yielded by the isochron method. In addition, the results provided by the isochron method are
662 generally closer to the ICP-MS isotopic measurements than those provided by the histogram
663 method (Cocherie et al. 1998; Cocherie and Albarède 2001), and thus are preferred here. As
664 mentionned in the “Sioule metamorphic series” section, in the study area, two pre-Late Viséan
665 events, namely D₁ and D₂, are recognized. However, the age of the D₁ and D₂ range between

666 385-375 Ma and 360-350 Ma, respectively. Therefore, the *ca* 340 Ma monazite ages cannot
667 be ascribed to a single tectono-metamorphic event.

668 Our new monazite chemical ages can be used to place time constraints in the pressure
669 (P)-temperature (T) plot in order to settle the P-T-t paths experienced by each unit. P-T
670 estimates from the Sioule area (Schulz et al. 2001) are consistent with P-T conditions from
671 other areas of the Massif Central (e.g. Santallier 1981; Mercier et al. 1991; Lardeaux et al.
672 2001; Bellot and Roig 2007; Faure et al. 2008). The UGU, LGU, and PAU underwent
673 different P-T paths depending on their location in the French Massif Central. In the Lyonnais
674 area, the coesite bearing eclogite argue for P conditions about 2.7-2.8 GPa (Lardeaux et al.
675 2001) whereas such ultra-high pressure conditions are not recorded (or not preserved) in the
676 Sioule area. On the basis of petrological studies in the Limousin and Artense areas (Santallier
677 1981, Mercier et al. 1991), peak-conditions of *ca* 2 GPa for *ca* 800°C are inferred for the
678 Sioule granulite (Fig. 12). Sample SO15 places a time constraint at *ca* 415 Ma on this HP
679 event. On the contrary, in the Limousin area, the LGU reached a maximum P value of 1.1-1.2
680 GPa for 500 °C (Bellot and Roig 2007; Faure et al. 2008). During their exhumation, the UGU
681 and LGU followed nearly isothermal decompression paths coeval with migmatization, and
682 subsequent cooling during the D₁ event at 385-375 Ma. The minimum P-T conditions
683 experienced by the UGU and LGU during this eo-Variscan cycle are not precisely settled.
684 Since andalusite replacing sillimanite or kyanite is never observed in the UGU or LGU, we
685 assume that the two units were not exhumed close to the surface but remained buried in the
686 upper crust, **even if aluminosilicates might have remained stable during exhumation.**
687 Conversely, the PAU never experienced the D₁ metamorphism nor migmatization. This unit
688 records maximum P and T values of 0.7-0.8 GPa, and 500-600°C, respectively. Our monazite
689 age of sample SO3 at 363±8 Ma constrains the age of the MP/MT metamorphism coeval with
690 the top-to-the-NW D₂ event. During the same late Devonian-Early Carboniferous period, the

691 UGU and LGU experienced a second metamorphic cycle leading to maximum P-T conditions
692 of 0.7-0.8 GPa, and 500-600°C similar to those of the PAU. Our monazite U-Th-Pb ages at
693 362Ma, 363 Ma, 355 Ma, and 351 Ma, for samples SO10, SO11, SO7, and SO17,
694 respectively represent the age of the D₂ event recorded in the UGU and LGU (Fig. 12).

695 After the completion of the two Variscan metamorphic cycles, the Sioule metamorphic
696 series, like most of the areas in the northern part of the FMC, experienced a huge late
697 **orogenic** magmatic episode, **at ca** 335-330 Ma, represented both by deep seated plutonism,
698 dyke emplacement, and aerial volcanism. As already proposed to account for the ⁴⁰Ar/³⁹Ar
699 ages, the hydrothermalism was responsible for the complete resetting of the zircon U-Pb
700 isotopic systems during the Tufts Anthracifères event **but the U-Th-Pb chronometer was not**
701 **affected by this late thermal event.**

702

703 **Conclusion**

704 This geochronological study in a key area of the Variscan French Massif Central, casts light
705 on the importance of structural and petrological constraints when dealing with age
706 determination of thermal, tectonic and metamorphic events in orogens. **These new radiometric**
707 **ages** yielded by monazite and zircon included in the macroscopic foliation **allow us to place**
708 **time constraints on the succession of superimposed deformation and related metamorphic**
709 **events in the Sioule area. The early Variscan HP/HT D₀ event occurred at 416±15 Ma and the**
710 **MP/MT D₂ event, coeval with the top-to-the NW shearing took place between 365 Ma and**
711 **350 Ma. On the contrary, the ca 380 Ma D₁ event has not been recorded by monazite. Our**
712 **study emphasizes that** radiometric ages must be carefully discussed in the light of the regional
713 geological setting. Particularly, the knowledge of the thermo-barometric and hydrothermal
714 processes must be a pre-requisite for the understanding of the geochronological data.

715

716 **Acknowledgements**

717 Field, and analytical expenses for this study have been supported by the French national
718 program “Carte géologique de la France au 1/50 000” led in BRGM, and by a National
719 Natural Science Foundation of China (NNSFC) grant 41273070. Ida di Carlo and Olivier
720 Rouer are acknowledged for their help for the acquisition of SEM and EPMA analyses. Jean-
721 Marc Lardeaux and Emilien Oliot are thanked for their constructive reviews.

722

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1023

1024 Figure Captions

1025 Fig. 1. A: Location of the French Massif Central in the European Variscan framework. B:

1026 Structural map of the Massif Central (modified from Ledru et al., 1989, and Faure et al.,

1027 2009). D₁, D₂, D₃ refer to the stretching lineation and kinematics referred to the Early

1028 Devonian, Early Carboniferous, and Middle Carboniferous tectono-metamorphic events,

1029 respectively. SGA: S^t-Gervais d’Auvergne pluton

1030

1031 Fig. 2. Structural map, and NW-SE cross-section of the Sioule series with sampling sites and

1032 microstructural data, after Grolier (1971), Feybesse and Teygey (1987), and Faure et al.

1033 (1993).

1034

1035 Fig. 3. Examples of representative macroscopic field structures of the Para-autochthonous (a),

1036 Lower Gneiss (b) and Upper Gneiss (c, d) Units. a1: outcrop of biotite-staurolite micaschist

1037 of sample SO3, with top-to-the-SE sigmoidal quart veins, a2: close-up of the mineral lineation

1038 marked by elongated biotite and staurolite porphyroblasts. This NW-SE lineation results of

1039 the superposition of D₂ and D₄ events. b) D₁ NE-SW striking isoclinal folds at Chouvigny. c)

1040 Biotite-sillimanite paragneiss with S1 flat-lying foliation and L1 mineral lineation, close to
1041 sample SO7. d) hand sample of garnet-kyanite granulitic paragneiss at Saugères (sample SO
1042 15). Note the biotite rim around garnet developed during the retrogression of the granulite
1043 facies metamorphic assemblage into amphibolite facies ones.

1044

1045 Fig. 4. Polarized light micrographs of analyzed samples showing the textural relationships of
1046 the dated monazite grains with other minerals (see location in Fig. 2). a: PAU micaschist
1047 sample SO3 from the Para-autochthonous Unit showing D₂ staurolite mantled by a sericite
1048 rim and deformed during the D₄ event. b: biotite-garnet-sillimanite gneiss (sample SO7) from
1049 the Upper Gneiss Unit with monazite included in biotite. c: biotite-sillimanite migmatite
1050 (sample SO11) foliated and lineated during the D₂ event, the dated monazite is included in
1051 biotite. d: kyanite-garnet granulitic paragneiss (sample SO15) with garnet surrounded by a
1052 biotite rim developed during the D₂ event.

1053

1054 Fig. 5. a: plot of the analyzed monazites in the $2\text{REEPO}_4\text{-CaTh(PO}_4)_2\text{-2ThSiO}_4$ system
1055 (Linthout, 2007) showing that the grains are close to the Ce-monazite pole. b: plot of the
1056 analyzed monazite in the Th+U+Si+Pb vs REE+P diagram (Franz et al., 1996). c: **weighted**
1057 **average age** of the core and rim of monazite contained in kyanite-garnet paragneiss of
1058 Saugères (sample SO15). d: X-ray compositional map for Th, Ca, U of the analyzed monazite
1059 grain in the granulitic paragneiss sample SO15, showing the chemical zoning with a Th, Ca,
1060 and U enriched rim. White and black squares represent EPMA analyse points for core and
1061 rim, respectively.

1062

1063 Fig. 6. Th/Pb vs U/Pb plots with SEM BSE pictures of dated monazites, sample location
1064 shown on Fig. 2, and GPS in Table 1. a, b, c, b: Upper Gneiss Unit. a) core of monazite grain

1065 from SO 15 granulite (Saugères); b) rim of monazite grain from SO15 granulite (Saugères); c)
1066 biotite-sillimanite migmatitic gneiss SO11 from Pont de Menat; d) biotite-sillimanite
1067 migmatitic paragneiss SO7 from Moulin de Bouchet; e) biotite-staurolite micaschist SO3
1068 from the Para-autochthonous Unit; f) two-mica paragneiss SO17 from the Lower Gneiss Unit.

1069

1070 Fig. 7. Cathodoluminescence (CL) images for representative zircons sample SO 15. The white
1071 ellipses in the CL images represent the spots of SIMS U-Pb analysis. SIMS spots are 30 μm in
1072 length for scale.

1073

1074 Fig. 8. Probability histogram of U-Pb age for 51 concordant analyses showing the two groups
1075 of zircon ages. Group 1 zircons have a main peak at 342 Ma and a secondary one at 327 Ma.

1076

1077 Fig. 9. Zircon U-Pb Concordia age of granulitic paragneiss sample S015.

1078

1079 Fig. 10. Synthetic map summarizing the available radiometric ages obtained through different
1080 methods, whole rock Rb/Sr, biotite and muscovite $^{40}\text{Ar}/^{39}\text{Ar}$, monazite U-Th-Pb, and zircon
1081 U-Pb SIMS for the Sioule series.

1082

1083 Fig. 11. Synoptic chart of the radiometric ages from the Sioule series presented in this paper,
1084 compared with the ages of the D₁, D₂, D₃ tectono-thermal events for the French Massif
1085 central, and the Guéret and S^t-Gervais biotite granites, and Pouzol-Servant microgranite.

1086 (1): Average biotite, muscovite, and hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ ages from Faure et al. (2002); (2):
1087 zircon U-Pb age of the Pouzol-Servant microgranite (Pin 1991); (3): zircon U-Pb SHRIMP and
1088 monazite U-Th-Pb chemical ages of the Guéret-St-Gervais plutons (Cartannaz et al. 2007).

1089

1090 Fig. 12. P-T-t paths for the Upper Gneiss Unit (UGU), Lower Gneiss Unit (LGU), and Para-
1091 Autochthonous Unit (PAU) modified from Mercier et al. 1991, Lardeaux et al., 2001; Schulz
1092 et al., 2001; Bellot and Roig 2007; Faure et al. 2007; Schulz 2009) emphasizing the
1093 polycyclic evolution. During the eo-Variscan cycle, the UGU experienced the D₀ high to
1094 ultra-high pressure metamorphism, and Devonian migmatization during its D₁ exhumation.
1095 The LGU underwent burial under lower pressure conditions than the UGU, then Devonian
1096 exhumation and migmatization during D₁. During the Late Devonian-Early Carboniferous D₂
1097 event, the PAU followed its first burial and exhumation cycle, whereas the LGU and UGU
1098 experienced a second metamorphic cycle before its Late Visean exhumation.

1099

1100 Table 1. Dated sample locations and analytical data

1101

1102 Table 2. Electron microprobe analyses of dated monazites

1103

1104 Table 3. Zircon U-Pb zircon analyses of granulitic paragneiss sample SO15

Figure 1
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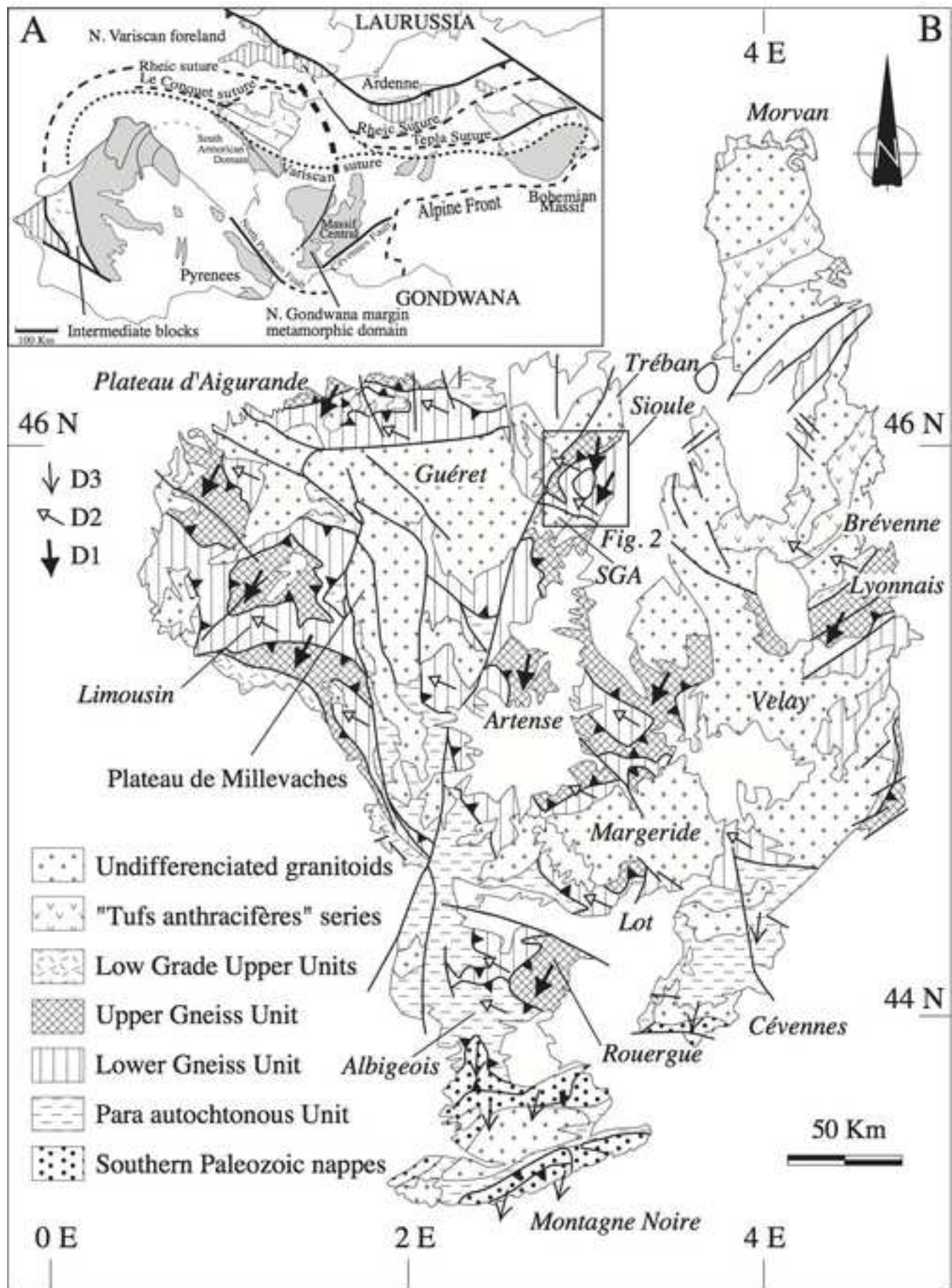


Fig. 1

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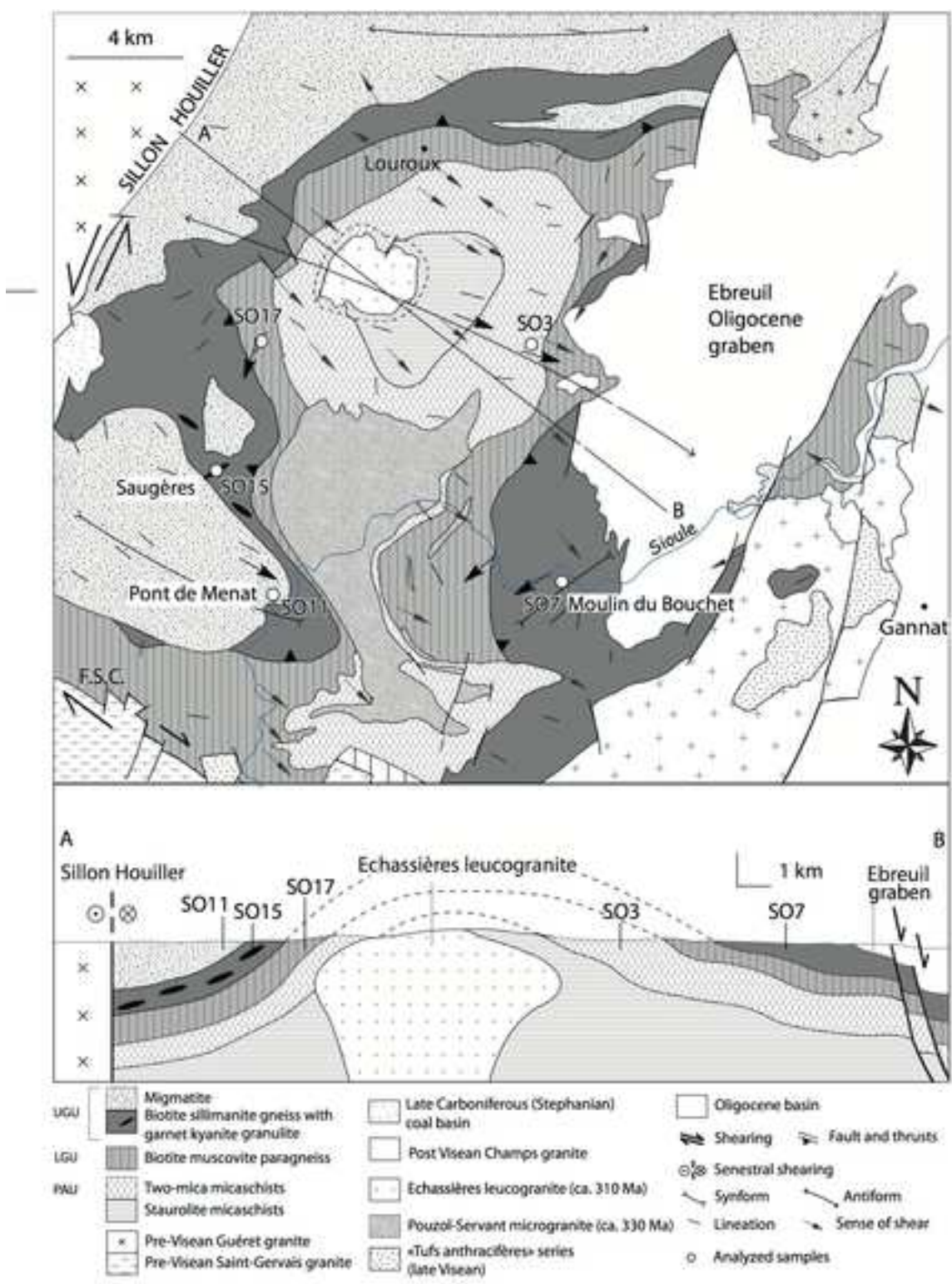


Fig. 2

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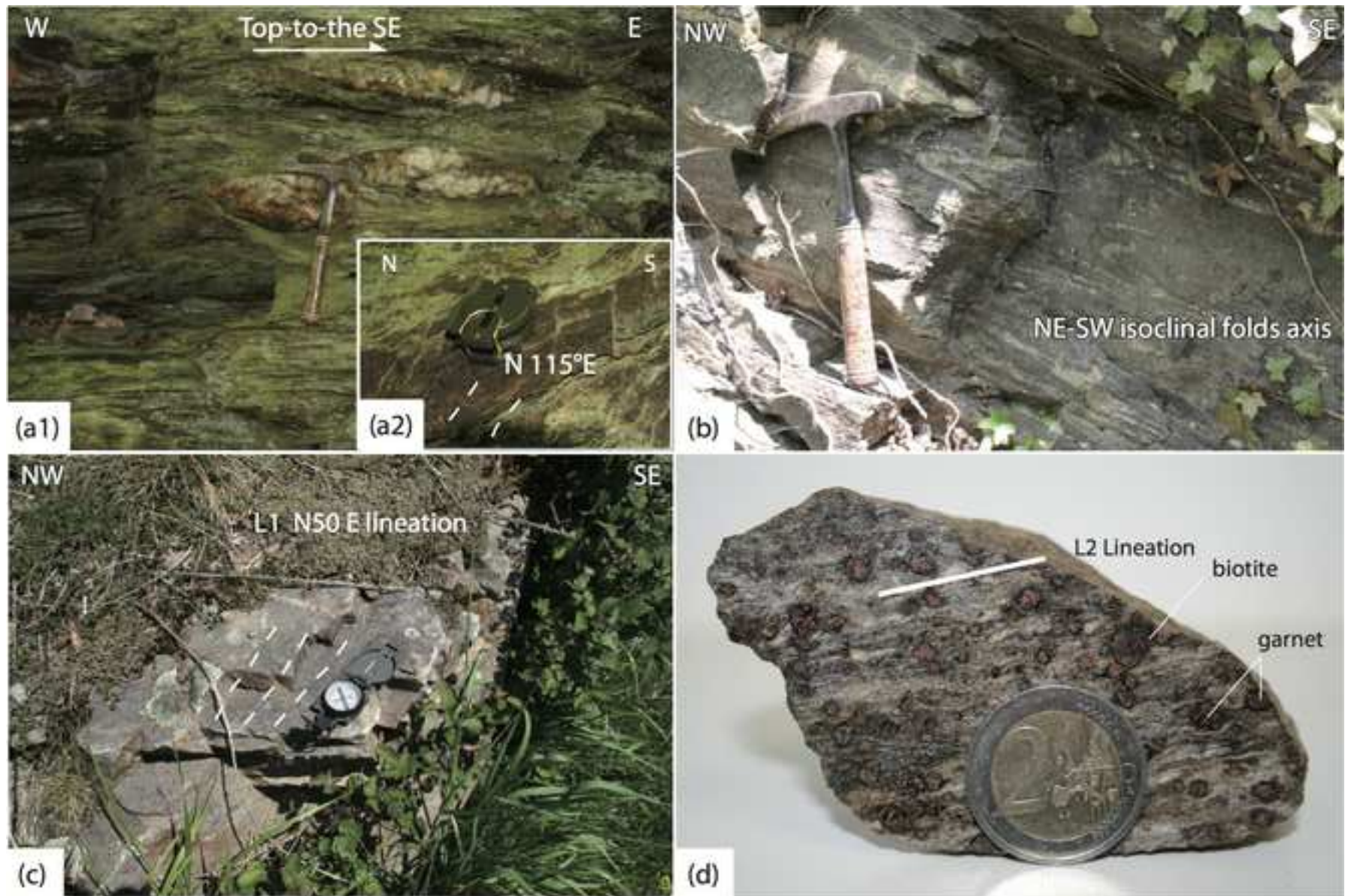


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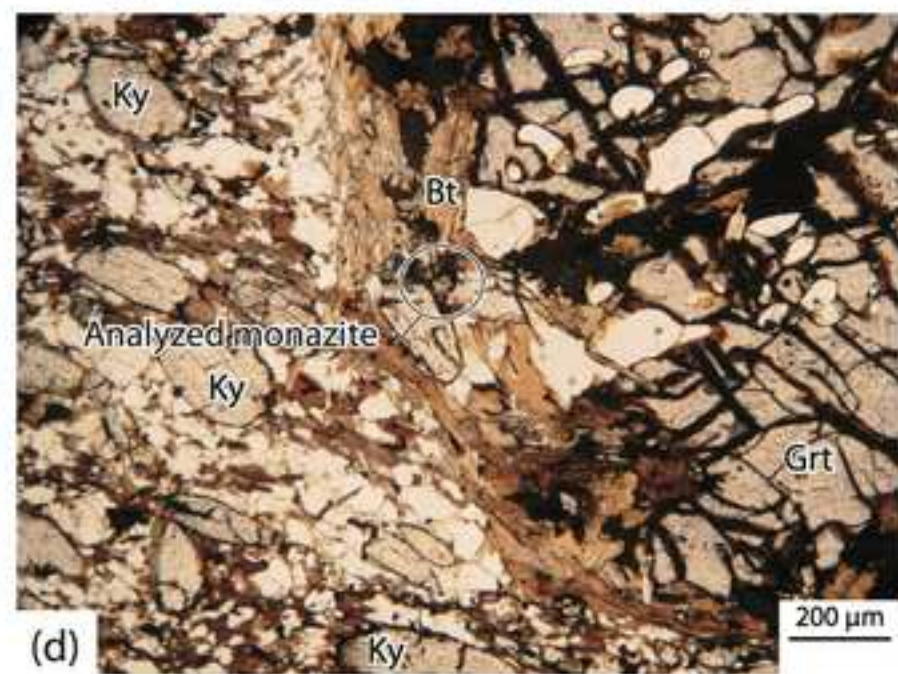
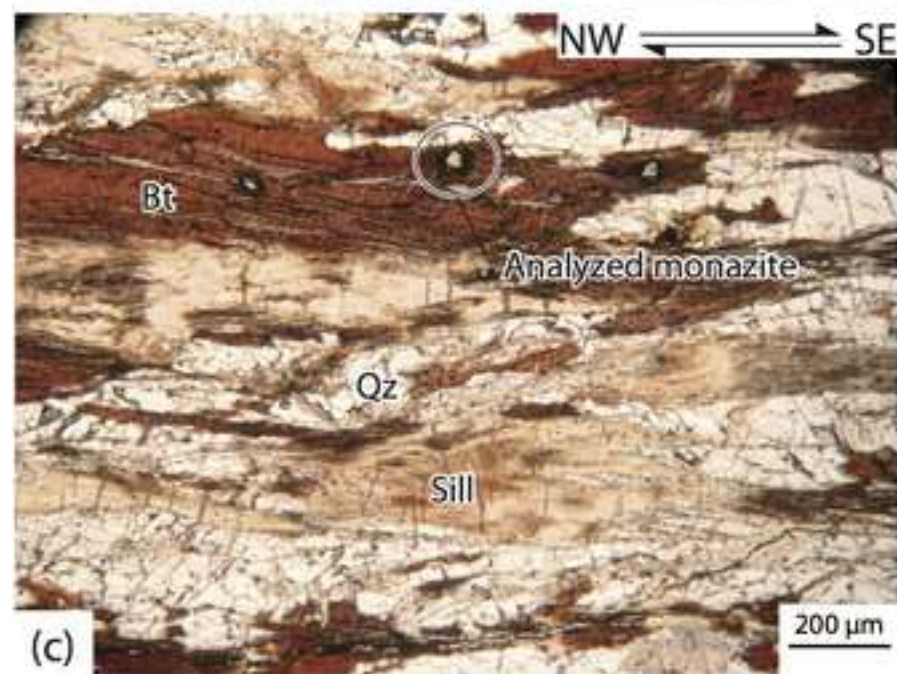
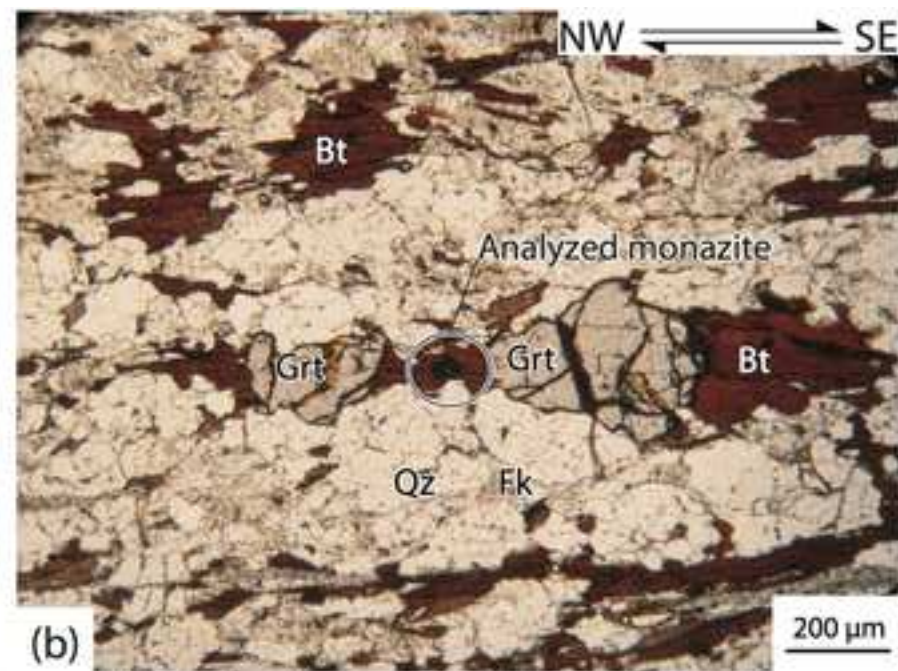
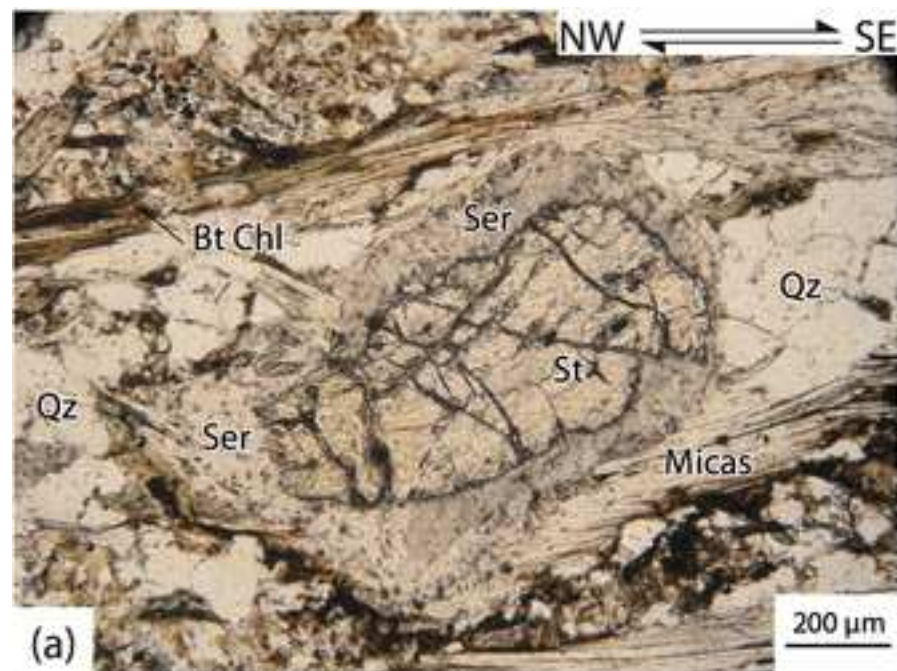


Fig 4

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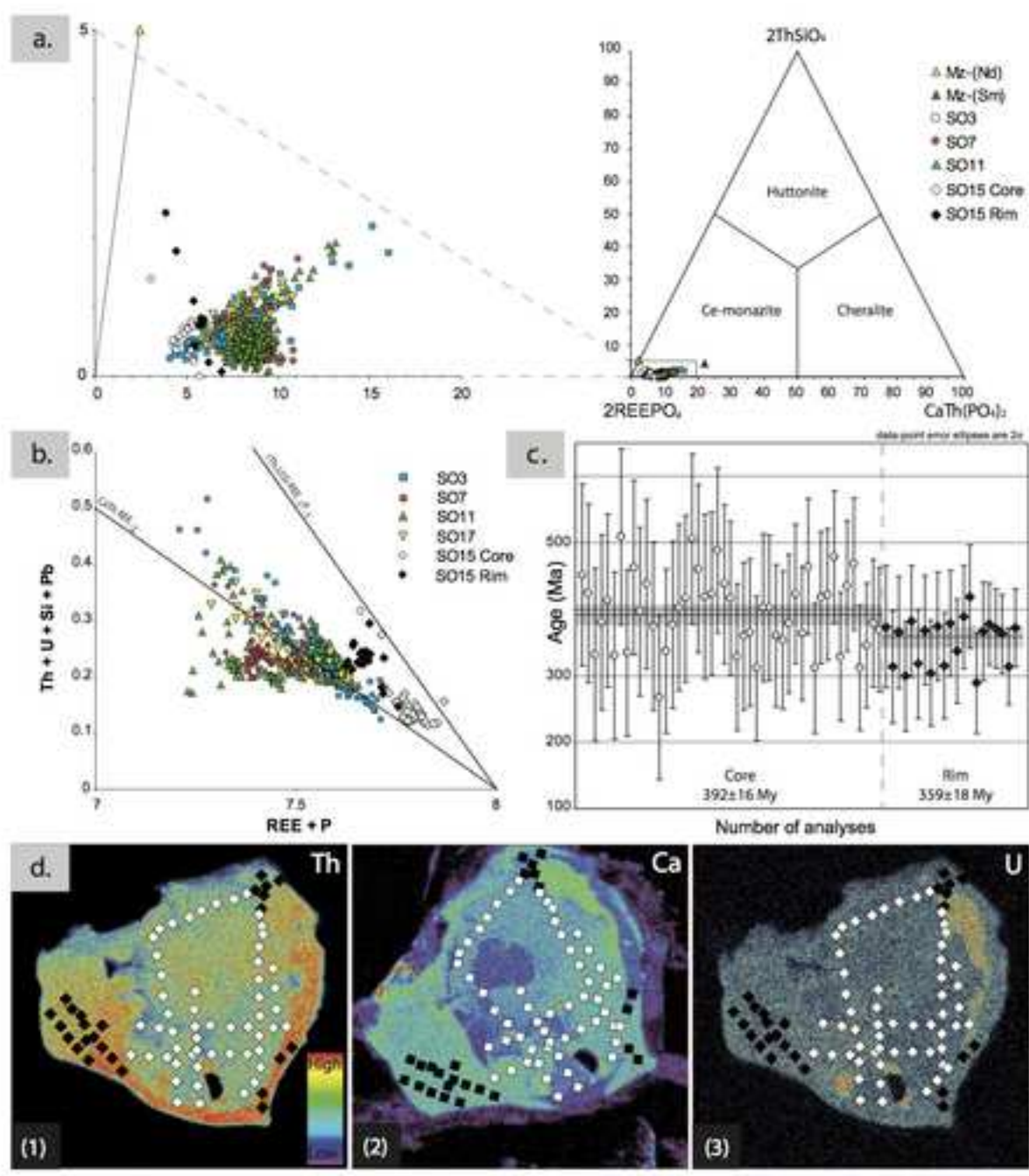


Fig. 5

Figure 6

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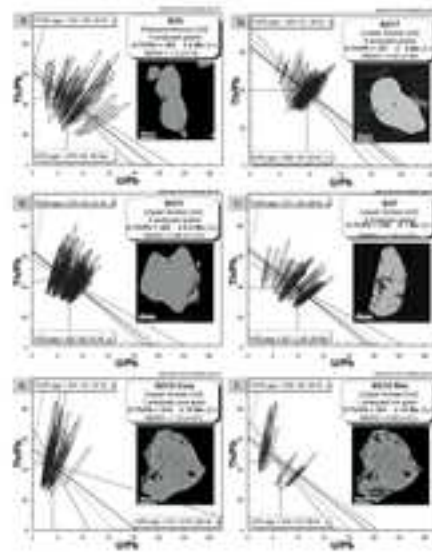


Fig 6

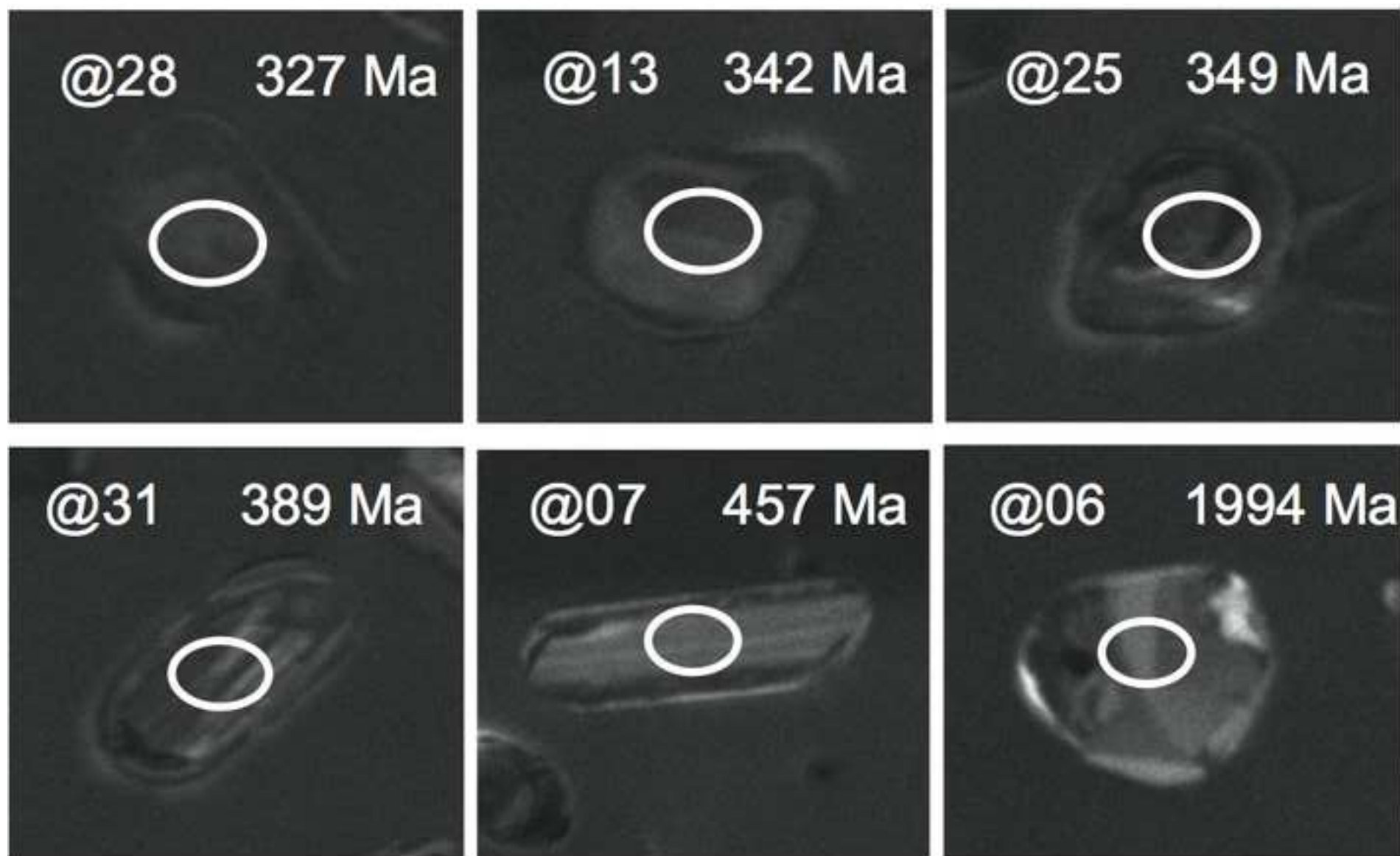


Fig. 2

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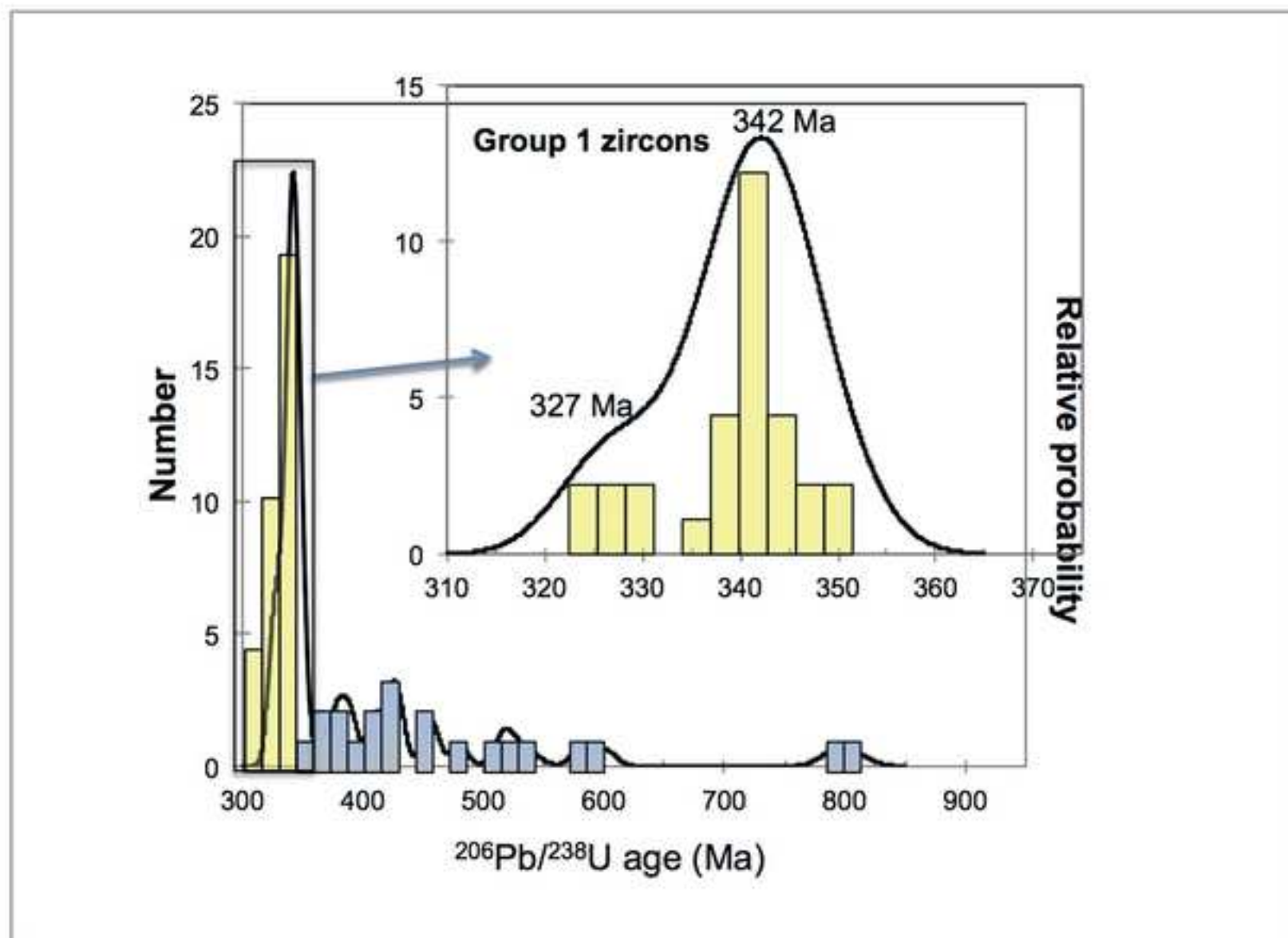


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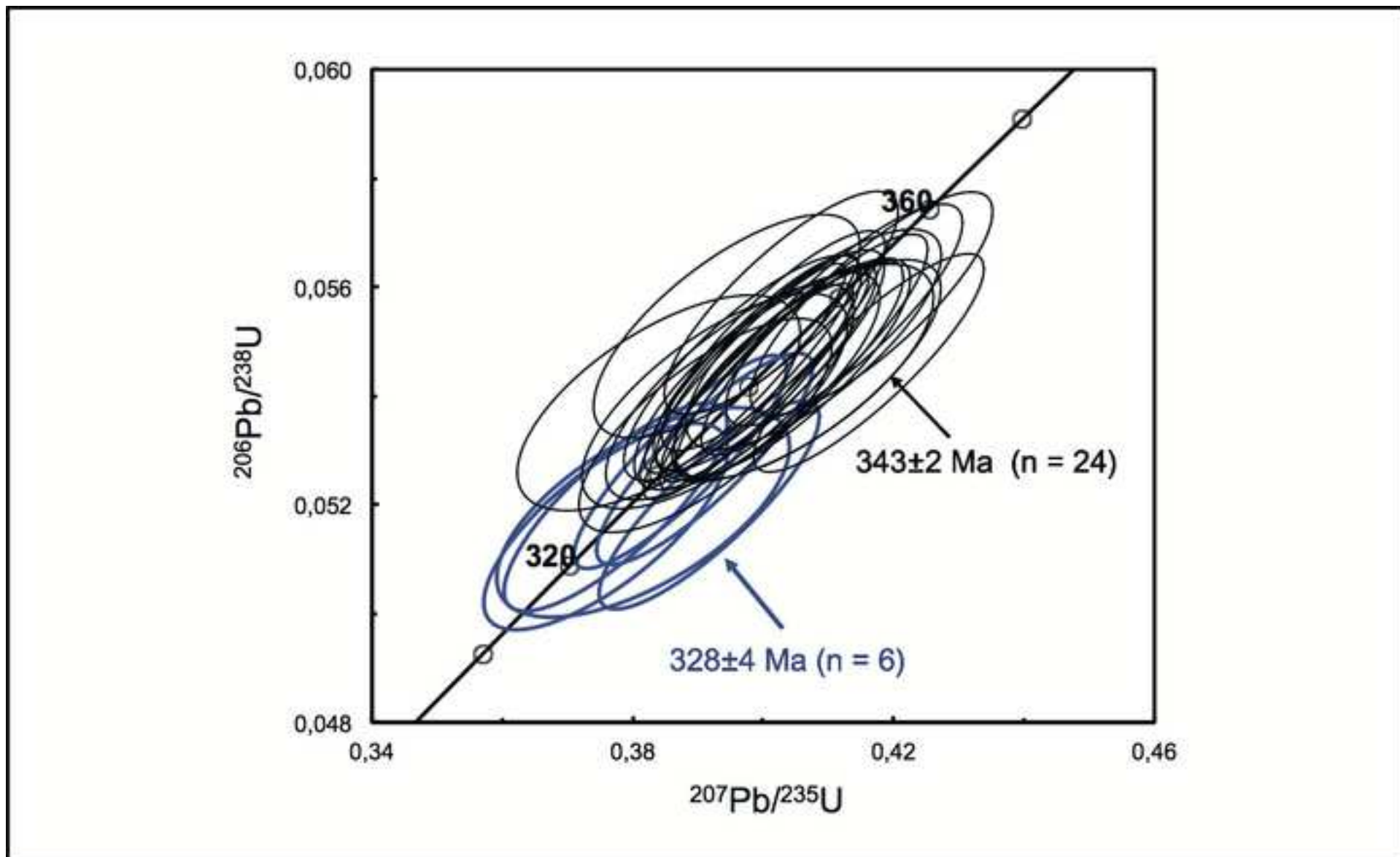


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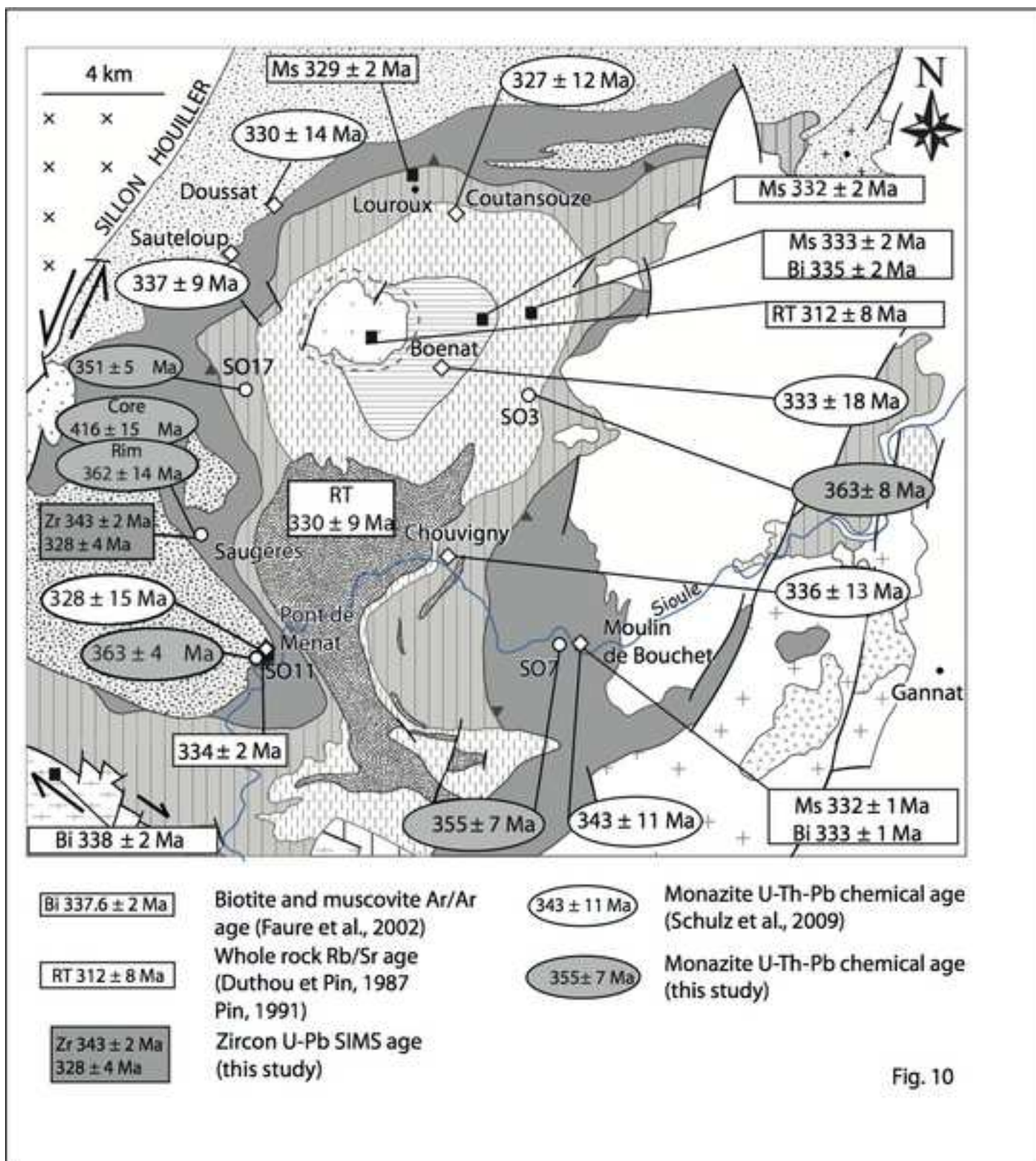


Fig. 10

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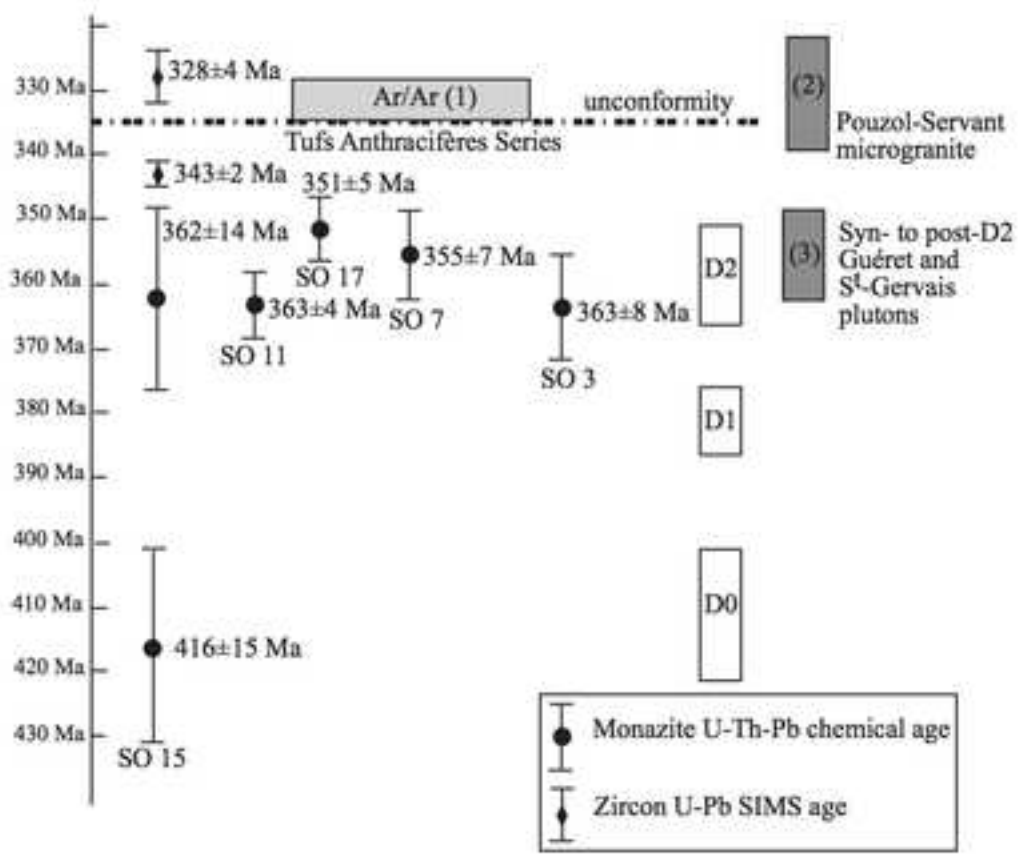


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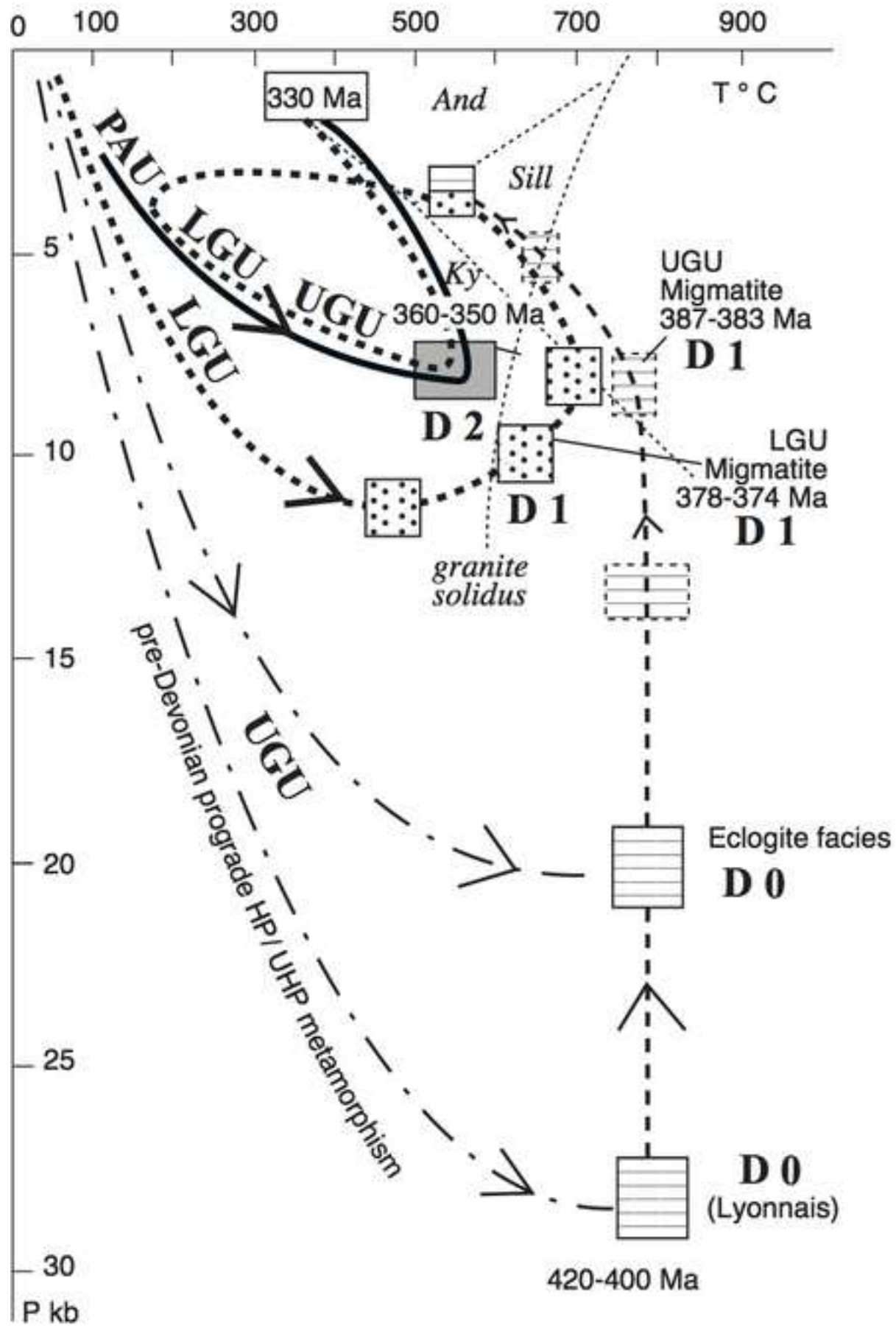


Fig. 12

Table 2. Electron microprobe analyses of dated monazites

Sample	UO ₂	PbO	Nd ₂ O ₃	CaO	P ₂ O ₅	ThO ₂	Gd ₂ O ₃	Ce ₂ O ₃	SiO ₂	Sm ₂ O ₃	La ₂ O ₃	Y ₂ O ₃	Pr ₂ O ₃
SO3	0.55	0.10	12.27	0.88	29.27	4.25	1.34	29.27	0.67	2.21	14.08	0.99	3.51
	±	±	±	±	±	±	±	±	±	±	±	±	±
SO7	0.18	0.03	0.54	0.27	0.5	1.63	0.28	0.94	0.99	0.16	0.78	0.26	0.15
	±	±	±	±	±	±	±	±	±	±	±	±	±
SO11	0.91	0.13	11.16	1.04	28.82	4.45	1.57	28.33	0.28	1.98	13.9	2.11	3.29
	±	±	±	±	±	±	±	±	±	±	±	±	±
SO15	0.29	0.02	0.45	0.11	0.58	0.59	0.15	0.51	0.13	0.16	0.31	0.64	0.1
	±	±	±	±	±	±	±	±	±	±	±	±	±
SO15 Rim	0.61	0.1	11.9	0.99	29.39	4.48	1.32	28.85	0.25	2.03	14.92	0.74	3.44
	±	±	±	±	±	±	±	±	±	±	±	±	±
SO15 Core	0.15	0.01	0.32	0.11	0.92	0.68	0.29	0.94	0.11	0.14	0.6	0.86	0.11
	±	±	±	±	±	±	±	±	±	±	±	±	±
SO17	0.22	0.06	12.34	0.65	29.33	3.05	0.51	32.19	0.15	1.62	16.55		3.65
	±	±	±	±	±	±	±	±	±	±	±	0	±
SO17	0.07	0.01	0.34	0.49	0.49	0.39	0.08	0.37	0.17	0.12	0.52		0.09
	±	±	±	±	±	±	±	±	±	±	±		±
SO17	0.44	0.09	12.44	0.81	29.1	4.39	0.76	30.78	0.32	1.93	15.86		3.57
	±	±	±	±	±	±	±	±	±	±	±	0	±
SO17	0.3	0.02	0.23	0.14	0.48	0.44	0.14	0.59	0.15	0.15	0.32		0.08
	±	±	±	±	±	±	±	±	±	±	±		±
SO17	0.95	0.15	10.98	1.35	28.04	5.67	1.55	27.26	1.11	1.88	13.36	2.56	3.23
	±	±	±	±	±	±	±	±	±	±	±	±	±
SO17	0.13	0.03	0.55	0.24	1.03	1.72	0.13	0.74	2.79	0.17	0.39	0.28	0.13
	±	±	±	±	±	±	±	±	±	±	±	±	±

Table 2

