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1 **Fast exhumation rate during late orogenic extension: the new timing of the Pilat**
 2 **detachment fault (French Massif Central, Variscan belt)**

3

4 Abbreviated title: New timing of the Pilat detachment fault (FMC)

5

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19

20 **Abstract:** A leucogranite sill in the footwall of the extensional Pilat shear zone (PSZ) Eastern

21 French Massif Central was emplaced and sheared in a short time interval, between

22 301.8±3.1 and 303.0±1.6 Ma (zircon U–Pb and mica ⁴⁰Ar/³⁹Ar ages, respectively). Extension

23 ended at 298.5±1.5 Ma (⁴⁰Ar/³⁹Ar on mica) as shown by a non deformed dike cross cutting

24 the sheared zone. On the hanging wall of the fault, the opening of the St. Etienne Basin filled

25 by coarse clastic sedimentation confirms the existence of a relief south of the PSZ. U-Pb

26 dating of a lithic-rich ignimbrite at the bottom of the basin and of a volcanic layer at the top

27 delivered ages of 310.5±3.4 Ma and 299.3±1.3 Ma, respectively. Close U-Pb and ⁴⁰Ar/³⁹Ar

28 ages from syn- to post-tectonic granites indicate very fast cooling (~150 °C/Myr) and

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29 exhumation rate (6 mm/yr) corresponding to the emplacement of granites in the middle
30 crust immediately followed by their fast tectonic exhumation and cooling into the upper
31 crust. The development of the Pilat fault in the thermally softened crust is responsible for
32 the rapid exhumation of the Velay dome, a Metamorphic Core Complex localized in the
33 footwall of the sheared zone. Here we propose that its rapid exhumation was due to
34 localized sub crustal lithospheric mantle dripping. These new data show that there was only
35 one extensional phase in the PSZ coeval with the opening of the St. Etienne coal Basin.
36 Finally, the new chronology obtained in this study is challenging previous ages suggesting 20
37 Ma activity for the Pilat shear zone.

Keywords: Variscan belt, Pilat shear zone, Velay, lithospheric dripping

1. Introduction

43 Since 1980 many studies have focused on normal faults observed in continental
44 collision settings (Molnar and Tapponnier, 1977; Malavieille 1993). This has led to the
45 concept of syn- to post-orogenic collapse where normal faults reflect an orogenic scale
46 process that contributes to smooth out the relief: the overthickened, and hence hot and
47 with a low viscosity, crust spreads out, while in the upper crust, the extensional deformation
48 is absorbed on low angle detachment faults above metamorphic core complexes (Wernicke
49 et al., 1987; Malavieille, 1987; Echtler and Malavieille, 1990). To explain such structures,
50 several models invoke either delamination, slab break-off (Sacks and Secor, 1990; Beaumont
51 et al., 2001) or the flow of a partially molten lower crust (Kusznir and Matthews, 1988; Block
52 and Royden, 1990; McKenzie and Jackson, 2002). These tectonic processes accompanying an

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53 elevation of the relief interact with erosion and sedimentation shaping the surface
54 morphology and the relief. To well characterize the extensive context in an orogeny it is
55 necessary to obtain geochronological and structural data. These data would then allow (1) to
56 determine if several extensional events occurred, particularly if ductile deformation were
57 followed by brittle deformation, (2) to place the extension in the magmatic and regional
58 tectonic evolution.

59 In the French Massif Central it has been proposed that two major events of
60 extensional tectonics occurred successively (Fig. 1A) during the Early Carboniferous (325-315
61 Ma) and the Late Carboniferous (305-275 Ma) respectively called syn- and post-orogenic
62 extension (Burg et al., 1991; Faure et al., 2009). The Late Carboniferous stage is linked to the
63 formation of several intramontane basins containing abundant clastic deposits with coal
64 intercalations. In this context the St Etienne basin is interpreted as a supra detachment basin
65 above the Pilat low angle shear zone on top of the large Velay granitic dome (Malavieille et
66 al., 1990; Gardien et al., 1997). However, existing geochronological data suggest that the
67 Pilat shear zone is Early Carboniferous in age dated at 322 and 313 Ma (Caen-Vachette et al.,
68 1984; Vitel 1988; Costa, 1990) while structural studies agree with the Late Carboniferous
69 extensive period (Malavieille et al., 1990; Gardien, 1990).

70 To settle this controversy, we carried out a structural and geochronological study to
71 constrain the deformation in the Pilat shear zone and the timing of sedimentation in the St
72 Etienne basin. Our results allow us to specify if this evolution corresponds to a transient or
73 ending episode of the Variscan orogeny in the French Massif Central.

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75 **1.1 Thickening and extensional tectonic episodes in the French Massif Central**

76 The Variscan orogeny (420–300 Ma) results from the collision between Laurussia and
77 Gondwana after the closure by subduction of several oceanic domains (Matte, 1986, 2001;
78 Domeier and Torsvik, 2014; Franke et al., 2017). One of the most metamorphic and thickest
79 parts of the belt crops out in the French Massif Central and in the Armorican Massif.
80 Petrological data imply that the crust reached at least 45 km in thickness (Lardeaux et al.,
81 1989; Gardien et al., 1990) after the collision phase dated at c.a. 360-350 Ma (Gay et al.,
82 1981; Maluski and Monié, 1988, Costa et al., 1993; Gardien et al., 2011; Do Couto et al.
83 2016; Chelle-Michou et al., 2017) following a subduction phase (Gardien et al., 1988; Mercier
84 et al., 1990). Thereafter, two successive stages of crustal extension affected the French
85 Massif Central during the last stages of its evolution. The first one, during the Early
86 Carboniferous period (Namuro- Westphalian, 325-315 Ma-equivalent to the Serpukhovian of
87 the Internationale Stratigraphic Chart (Gradstein et al., 2012), has been defined based on
88 geochronological and structural data obtained on syntectonic granites and country
89 metamorphic rocks. This stage is characterized by a NW-SE stretching direction in almost the
90 entire French Massif Central (Fig. 1B) (Faure, 1995; Roig et al., 1998; Roig and Faure, 2000;
91 Talbot et al., 2005a, 2005b). At that time compression and transpression were still active in
92 the external parts of the belt (Montagne Noire, Maures-Tanneron, Sardinia) (Lagarde et al.,
93 1994; Feybesse et al., 1995; Carosi and Palmeri, 2002; Rolland et al., 2009; Roger et al., 2015,
94 Trap et al., 2017; Poujol et al., 2017). The second extensional stage occurred during the late
95 Carboniferous-Early Permian period (Stephanian-Autunian, 305-282 Ma equivalent to the
96 Upper Pennsylvanian and lower Cisuralian of the international stratigraphic chart) and is
97 defined as a post collision extension associated with the ascent of migmatitic domes and the
98 opening of Carboniferous intramontane coal basins. Throughout the French Massif Central,
99 this stage is characterized by a NNE-SSW stretching direction (Fig. 1B) (Malavieille et al.,

100 1990; Malavieille 1993; Burg et al., 1994). However, precise data on how (mode and rate)
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104 **1.2 Regional Geologic setting**

105 The St Etienne basin is a ~33 km long trending NE-SW basin filled by continental detrital
106 sediments of Carboniferous age (Fig. 1). The sedimentary succession starts with a coarse-
107 grained breccia and conglomerates locally channelized in a micaceous sandy matrix. These
108 are followed by a mixed well-rounded conglomeratic and sandy sequence with interlayered
109 silts and coal deposits (Becq-Giraudon et al., 1996; Brugier et al., 2003). The plant and spore
110 fossils contained in the coal series of the S^t Etienne basin were used to define the Stephanian
111 stratotype, the continental equivalent of the Moscovian, the Kasimovian, and the basal part
112 of the Gzhelian (Gradstein et al., 2012). To the north the sedimentary series rests
113 unconformably on the metamorphic series of the Mont du Lyonnais while to the south the
114 NE-SW trending St Chamond fault is the basin boundary that separates it from the Pilat Unit
115 (Fig. 1C). The lithology of the Pilat Unit consists of aluminous micaschists, leptynitic gneisses
116 and orthogneisses in which leucogranite pods are variably involved in the main tectonic
117 structures (Malavieille et al., 1990; Gardien 1990). Geochemical analysis indicate that the
118 granites were generated by the partial melting of pelitic sediments (Vitel 1988). The unit
119 underwent a high-temperature and low-pressure metamorphism superimposed on a former
120 medium pressure and temperature metamorphic stage related to a top to the south nappe
121 stacking predating the normal faulting (Gardien 1990; Gardien et al., 1997). The main
122 structure of the Pilat Unit is characterized by a foliation plane mainly parallel to the
123 lithological limits, dipping 5 to 30° to the north (Figs. 1C and 2) and supporting a stretching

124 lineation trending N160° to N30° (Fig. 1C) dipping of 0 to 30° towards the north (Fig. 2). The
125 foliation plane consists of quartzo-feldspathic layers interbedded with mica-sillimanite rich
126 layers wrapping garnet and staurolite relicts and bear a biotite lineation. In the southern part
127 of the Unit, numerous kinematics indicators: S-C plane relationships, shear bands, feldspar
128 asymmetric tails and pressure shadows show top-to-the-north sense of shear compatible
129 with a kilometer-scale normal shear zone: The Pilat Shear Zone (PSZ) (Malavieille et al.,
130 1990). To the south, the Velay granito-migmatitic dome is found at the base of the PSZ (Figs.
131 1C and 2). At the Gouffre d'Enfer locality a leucogranite sill intrusive in the micaschist is
132 interpreted as synkinematic with the top-to-the-north sense of shear (Malavieille et al.,
133 1990). Its whole rock Rb/Sr date of 322±9 Ma was interpreted as Bashkirian-Moscovian
134 (Namuro-Westaphalien: 325-315 Ma) age of the ductile shearing (Caen-Vachette et al., 1984;
135 Vitel, 1988). Such age was corroborated by a biotite ³⁹Ar/⁴⁰Ar date of 313±6 Ma (Costa,
136 1990) of a lenticular orthogneiss located in the southern part of the Pilat unit (sample PI11,
137 Figs 1 and 2). Biotite and muscovite from an aluminous micaschist (sample PI2) surrounding
138 the lenticular orthogneiss yield dates of 296.0±3.1 and 299.7±3.1 Ma respectively
139 interpreted as the minimum age for the extensional deformation (Costa, 1990). The St
140 Etienne basin, Gzhelian in age (Stephanian: 305-299 Ma) would have opened later,
141 synchronously with the motion of a second detachment that crosscuts the former shear
142 zone hanging wall and connects to the first one at depth (Malavieille et al., 1990). This stage
143 was dated at 290.9±3.1Ma in a granite (³⁹Ar/⁴⁰Ar on biotite Costa, 1990) crosscut by the St
144 Chamond fault delimiting the St Etienne basin to the south (sample PI4, Figs 1 and 2).
145 Altogether these geochronological data suggested that the motion of the Pilat fault lasted at
146 least 30 Ma from the ductile deep to superficial brittle stages. Alternatively, Mattauer and
147 Matte (1998) interpret the St Etienne coal basin as a pull apart basin along the right-lateral

148 St Chamond Fault crosscutting the PSZ. Indeed, based on the previous geochronological data
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3 149 and structural analysis, the model proposed by Mattauer and Matte (1998) proposed that
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5 150 the opening of the Carboniferous basin of St Etienne was dissociated from the activation of
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8 151 the Pilat normal fault. They put forward two stages for the opening of the basin i) a pull
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10 152 apart system which formed on a dextral detachment relay allowing a thick sedimentation in
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13 153 the basin and ii) followed by a compressional stage occurring in its southeastern part and
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15 154 responsible for the deformation within the basin.
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156 **2. Material and Method**

157 **2.1 Sampling description (Figure 3)**

158 To better constrain the age of the Pilat extensional “system” we have sampled granites and a
159 micaschist from the Pilat Unit (refer to figures 1 and 2 for sample location). We have
160 sampled the previously dated syntectonic leucogranite from the Gouffre d’Enfer (PL06, Figs
161 3a, b) and complemented with samples from two other granites from the Pilat Unit. These
162 new sample sites are moderately affected (pegmatites PL49 and PL51) and unaffected
163 (microgranite PL43) by extensional tectonics, respectively. Dating the three granites allowed
164 us not only to constrain the duration of the Pilat normal fault but also the magmatic activity
165 associated with the late evolution of the belt. We have selected a plane of a brittle normal
166 fault (PL48) crosscutting a microgranite PL 43 to date the extensional superficial brittle stage.
167 The fault plane contains a mineral stretching lineation down-dip on the fault plane
168 underlined by white mica that are amenable to dating.

169 To provide the first radio-isotopic age constraints on the volcano-sedimentary series of the
170 St. Etienne Carboniferous Basin we have selected two volcanic samples belonging to the
171 base (PL92) and to the top (PL90) of the sedimentary series (Fig. 1C). One sample (PL90)

172 corresponds to a cinerite interbedded with coal deposits outcropping at the Ricamary site
173 south of St. Etienne providing a minimum age for the coal basin. The second sample (PL92) is
174 a previously described rhyolitic dyke on the 1/50 000 geological map of St. Etienne
175 (Chenenvoy et al., 1970) outcropping at the base of the basin east of St. Chamond. We have
176 used the U/Pb by LA-ICP-MS method on separated zircons from the granites and the volcanic
177 rocks. We have also selected a micaschist sample south of the St. Chamond fault (PL94) to
178 verify the existence of a second superficial detachment that has been previously proposed as
179 responsible for the opening of the St. Etienne Coal Basin (Fig. 5 in Malavieille et al., 1990).
180 We have used the laser probe $^{40}\text{Ar}/^{39}\text{Ar}$ method on separated micas from the granites and
181 the micaschist.

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183 **2.1.1 The Studied samples**

184 The Gouffre d'Enfer leucogranite (sample PL06) forms a NE-SW oriented sill intruding the
185 southwestern and lower part of the micaschists of the PSZ (Figs. 1C and 2). The leucogranite
186 displays an equant texture free of deformation which passes locally to an orthogneissic or
187 even to a highly deformed rock with a mylonitic texture. The mylonitized zones show an N-S
188 trending lineation (Figs. 3a and d) defined by elongated biotite \pm sillimanite crystals hosted
189 by an E-W to N30 oriented foliation plans (Fig. 3e). Shear criteria as feldspar asymmetric tails
190 and S-C plane relationships (Fig. 3b) indicate top to the north sense of shear (Pitiot, 1984;
191 Vitel, 1988; Malavieille et al., 1990). The leucogranite and the micachists display concordant
192 boundaries and tectonic structures as foliation and shear planes (Fig. 3c). Previous structural
193 studies conclude that the leucogranite emplacement is contemporaneous with the regional
194 deformation (Malavieille et al., 1990). The ductile structures are cut by brittle normal faults
195 and sinistral strike-slip faults oriented N45 to N90 (Figs. 2 and 3e). The PL06 sample is

196 composed of quartz forming ribbons, feldspar, biotite, and muscovite oriented in both S and
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3 197 C planes. The basal and deepest part of the intrusion is marked by the occurrence of
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5 198 sillimanite elongated in the foliation plans and shear bands.

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7 199 The most southerly collected granite (sample PL43, Fig. 1C and 2) corresponds to a vertical
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10 200 two-mica microgranite dyke cross cutting a migmatitic orthogneiss (Fig. 3f). In the migmatitic
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13 201 orthogneiss (sample PL44) the relationships between the foliation plans and the shear bands
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15 202 indicate a normal motion toward the North (Fig. 3h). The microgranite, made of millimeter
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18 203 size muscovite and biotite associated with quartz, plagioclase, and K-feldspar (Fig. 3g) shows
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21 204 no evidence of ductile deformation. However, the microgranite and the orthogneiss are
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23 205 crosscut by a series of sub-vertical fractures and slip planes with white micas and quartz
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26 206 striations (Fig. 3i). Sample PL48 is one of these fault planes oriented N80, dipping 50° toward
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28 207 the south and bearing N-S elongated white mica dipping toward the south.

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31 208 We have collected two samples of a pegmatitic leucogranite sill (PL 49 and PL 51) which
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33 209 crops out in the middle part of the Pilat micaschist Unit (Fig. 3j) and whose orientation and
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36 210 dip are subparallel (angular shift from 5° to 10°) to the foliation of the host micaschists (Fig.
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39 211 3l). The weak deformation affecting the pegmatite is shown by the development of a weak
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41 212 foliation plane emphasized by recrystallized quartz ribbons as well as slight undulose
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44 213 extinctions in muscovite and feldspar porphyroclasts. Late high angle brittle normal faults
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46 214 crosscut the pegmatite (Fig. 3k and m) and the fault planes show striations outlined by
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49 215 quartz and white micas. The orientation and dip of the fault planes as well as the striations
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52 216 indicate a brittle N-S direction of extension (Fig. 3n and o).

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54 217 Sample PL94 is a micaschist taken from the northern part of the Pilat Unit above the PSZ,
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57 218 near the St. Chamond fault that bounds the St. Etienne Carboniferous basin to the south.

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59 219 The main paragenesis in this sample consists of an association of muscovite + quartz +
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220 feldspars elongated in the foliation plane and surrounding more or less retrogressed garnet
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3 221 and staurolite relicts (Gardien, 1990). Here the foliation is either vertical or dipping 50° to
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5 222 the west, and striking N-S to N20°. The foliation is crossed by different generations of quartz-
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7 223 chlorite veins (Fig. 4a).

9
10 224 Near the city of Grand Croix (Fig. 1), a lithology has been previously described as a rhyolite
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12 225 dyke (1/50000 geological map of St. Etienne, Chenevoy, 1970). We re-sample this unit (PL
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14 226 92) in the same outcrop and in our view, it does not show a clear dyke crosscutting
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16 227 relationship with the metasediments, but rather consists of blocks of green rocks within a
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18 228 conglomeratic series. Microscopic observations show that the rock is composed of an
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20 229 assemblage of very small quartz grains, pyrophyllite and chlorite replacing a glassy matrix.
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22 230 The latter contains phenocrysts of quartz, plagioclase, and highly altered micas as well as
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24 231 many small fragments (diameter < 0.5 mm) of metamorphic rocks including quartzite and
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26 232 micaschist (Figs. 4b, c and d). Thus, in this work we consider sample PL92 as a lithic-rich
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28 233 ignimbrite rather than a rhyolitic dyke.

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31 234 A sample of sediments of the St Etienne basin was picked near St. Etienne city. This site
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33 235 exposes stratified sedimentary rocks showing alternations of coal and ash layers. Sample
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35 236 PL90 was taken in an ash layer containing non-altered micas, iron oxides and zircon (Fig. 4e).

36 237

37 238 **2.2. Analytical methods**

38 239 **2.2.1. U/Pb (LA-ICP-MS) on separated zircon**

39 240 Rock samples were crushed and < 400 µm sieved fractions were processed using a Wilfley
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41 241 shaking table, a Frantz magnetic separator and heavy liquids (methylene iodide) to extract
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43 242 zircons. Zircons grains were handpicked under a binocular microscope, mounted in epoxy
44
45 243 and polished to expose their interior. Zircons grains were imaged with cathodoluminescence

244 (CL) on a JEOL CamScan scanning electron microscope (using a 10 kV accelerating voltage) at
245 the University of Geneva, Switzerland for sample PL06 Appendix 1 (Supplementary data) and
246 at the University of Clermont-Ferrand, France for samples PL92 and PL90.

247 U-Th-Pb isotope data were measured by laser ablation inductively coupled mass
248 spectrometry (LA-ICP-MS) at the Laboratoire Magmas & Volcans (Clermont-Ferrand, France)
249 using a Resonetics M-50 193 nm Excimer laser system coupled to a Thermo Element XR
250 sector field ICP-MS. Helium carrier gas was supplemented with N₂ for sensitivity
251 enhancement [Hurai et al., 2010; Paquette et al., 2014]. The laser operated with a spot size
252 of 27 μm, a repetition rate of 3 Hz, and a fluence of 2.5 J/cm². Reduction of raw data was
253 carried out using the GLITTER[®] software package (van Achterbergh et al., 2001). Isotope
254 ratios were corrected for laser-induced and instrumental mass fractionation via sample-
255 standard bracketing using the GJ-1 zircon reference material (Jackson et al., 2004). Data
256 were not corrected for common Pb. ²⁰⁷Pb/²⁰⁶Pb vs. ²³⁸U/²⁰⁶Pb diagrams were generated
257 using the Isoplot/Ex v. 2.49 software of Ludwig (2001). 2σ level error ellipses incorporate
258 both internal and external uncertainties. The 91500 (Wiedenbeck et al., 1995) zircon
259 reference material (132 analyses: 1063.9 ± 2.4 Ma) was analyzed along with the samples to
260 independently monitor the external precision and accuracy of the measurements. Detailed
261 analytical data for samples PL06, PL90 and PL92 are given in the Appendix 2 (Supplementary
262 data).

263

264 **2.2.2. ⁴⁰Ar/³⁹Ar dating on separated micas**

265 Rock samples were broadly crushed, washed, dried, and finally sieved to select the 200-400
266 μm fraction. Unaltered grains of mica were then carefully sorted under the binocular lens,
267 then cleaned with alcohol and distilled water. The picked micas were irradiated in the TRIGA

268 reactor of Pavia (Italia) together with several monitors of Taylor Creek sanidine at $28.34 \pm$
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2 269 0.28 Ma (Renne et al., 1998). After irradiation, the samples were loaded in a CO₂ laser
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5 270 extraction-purification line connected to a multi-collector mass spectrometer for argon
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8 271 analyses (Argus VI by ThermoFisher at the Montpellier University, France) equipped with
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10 272 four Faraday cups and one CDD multiplier for ³⁶Ar. Argon isotope intensities were acquired
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13 273 simultaneously on the five detectors in ten runs. Air aliquots from an automated pipette
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15 274 system were analyzed each sample to monitor mass discrimination and detector bias. Step-
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18 275 heating degassing of the micas was done by regularly incrementing the laser power to
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21 276 recover the distribution of argon isotopes in the mineral and to reconstruct its age spectrum.
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23 277 For each spectrum, two to five mica grains were simultaneously degassed depending on
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26 278 their size. Ages reported on the spectra have been obtained after correction with blanks,
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28 279 mass discrimination, radioactive decay of ³⁷Ar and ³⁹Ar, irradiation-induced mass
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31 280 interferences and atmospheric contamination. They are reported with two sigma
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33 281 uncertainty and were evaluated using a ⁴⁰Ar/³⁶Ar value of atmospheric argon of 298.6 (Lee et
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35
36 282 al., 2006) and the decay constants of Steiger and Jäger (1977). Different decay constants
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39 283 have been more recently proposed (Min et al., 2000; Renne et al., 2011) but until now there
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41 284 is no full consensus for their use in ⁴⁰Ar/³⁹Ar dating (Fleck et al., 2019; Schaen et al., 2020).
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43
44 285 The raw data obtained during the acquisition are processed under ArArCalc (Koppers, 2002)
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46 286 to calculate the ages of the age spectra and isochron plots. A summary of argon results is
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49 287 given in table 1. Detailed analytical data are given in the Appendix 3 (Supplementary data).
50
51 288 The micas collected for dating the rocks were analyzed with a SEM coupled to an EDX at the
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53
54 289 ENS-Lyon. The chemical composition in mass percentages of biotite and muscovite are
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56
57 290 comparable to those found in granites. The results of the analyses see Appendix 4
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59 291 (Supplementary data) reveal a homogeneous chemical composition from the core to the

292 edge. The Al_2O_3 contents vary from 35 to 38%, the K_2O varies from 10 to 11% and their Si
293 contents are constant (~6.15). The analyzed biotite grains (PL06 and PL43) have an Al_2O_3
294 contents which vary from 20 to 23%. Biotite from sample PL43 contain 28% of FeO while
295 biotite from PL06 is poorer in FeO (19%) and richer in MgO (~8%), alkalis (K_2O) vary from 8.5
296 to 11%. Their Si / Al ratios are homogeneous in the range of 1.6 to 1.8. The maps of element
297 distributions presented in Appendix 4 (Supplementary data) confirmed that the dated biotite
298 and muscovite are not zoned and do not present traces of alteration.

299 **3. Geochronological Results**

300 **3.1. U-Pb dates**

301 The Gouffre d'Enfer syntectonic granite (Sample PL06) contains poorly translucent zircons
302 with varied shapes (elongated or rounded). They are composed of inherited relict cores
303 surrounded by oscillatory zoned rims of magmatic origin (Fig. 5; Photo Zr PL06). There is no
304 textural evidence for metamorphic zircon overgrowth. Dating the zircon rims provides two
305 main populations of concordant U-Pb dates at ca. 528 Ma and 301.8 ± 3.1 Ma (lower
306 intercept, $n = 14$, Fig. 5) respectively. An intermediate group of concordant dates at ca. 470
307 Ma has also been measured and most likely represents an Ordovician inherited component
308 in the source of the granite.

309 The volcanic layer (sample PL 90) provided translucent homogeneous needle shape
310 zircons grains with high length to width ratios suggesting rapid growth (Fig. 5). Euhedral
311 grains show oscillatory euhedral zoning consistent with a magmatic origin (Fig. 5). Some
312 crystals display rounded gap growth corresponding to former gas bubble footprints (Zr40,
313 Zr48, and Fig. 5). No inherited relict cores have been identified. Such morphology is
314 consistent with an igneous origin of the zircons. Most points are concordant and yield a
315 lower intercept date of 299.3 ± 1.3 Ma.

316 The zircon population from the lithic-rich ignimbrite (PL92) is homogeneous,
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3 317 characterized by translucent crystals with potentially corroded and inherited cores (Zr14, Fig.
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5 318 5) surrounded by thick euhedral oscillatory zoning rims. A wide range of concordant dates
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7 319 are obtained from 850 to 310 Ma as expected for a volcano clastic rock. For such rock type,
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10 320 the emplacement age is constrained by the youngest zircon population defining a lower
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13 321 intercept date of 310.5 ± 3.4 Ma ($n = 7$). All the youngest dates were obtained on euhedral
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15 322 grains showing oscillatory without any inherited relict cores (grains Zr31, Zr45 and Zr48, Fig.
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18 323 5).

20 324 **3.2. $^{40}\text{Ar}/^{39}\text{Ar}$ ages**

23 325 It should be noted that all the studied micas provide well defined $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages
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25 326 (Table 1 and Fig. 6) for 95 to 100% of the ^{39}Ar released, except for biotite PL43 (Fig. 6 B and C)
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28 327 for which the plateau age was calculated over 60% of the argon release. The intercept ages
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30
31 328 in the $(^{36}\text{Ar}/^{40}\text{Ar})$ versus $(^{40}\text{Ar}/^{39}\text{Ar})$ correlation plot is similar to the plateau ages with
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33 329 atmospheric $^{40}\text{Ar}/^{36}\text{Ar}$ initial ratios. All experiments show Ca-derived ^{37}Ar close to the
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36 330 detection limit and very low $^{38}\text{Ar}_{\text{Cl}}/^{39}\text{Ar}_{\text{K}}$ values. The biotite grains selected from the syn-
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38
39 331 kinematic Gouffre d'Enfer granite (PL06) are elongated grains in both foliation planes and
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41 332 shear bands. They yield a plateau age of 303.0 ± 1.6 Ma (Fig. 6A). The white micas selected
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43
44 333 from the pegmatites (PL 49 and PL51) moderately affected by the shear deformation
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46 334 provided plateau ages of 301.5 ± 1.6 Ma and 302.9 ± 1.6 Ma, respectively (Fig. 6E and F).
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48
49 335 Muscovite and biotite grains from the post kinematic leucogranite dyke (PL 43) display
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51 336 $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 298.5 ± 1.5 Ma and 302.9 ± 1.5 Ma respectively (Fig. 6B and C). The
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53
54 337 elongated white micas on the brittle normal fault plane (PL48) give an age of 303.9 ± 1.6 Ma
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57 338 (Fig. 6D). Finally, the muscovite grains from the micaschist from the northern part of the
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59 339 Pilat series (PL94) yielded a much older age of 332.8 ± 1.7 Ma (Fig. 6G).

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341 **4. Discussion**

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342 **4.1. Age of the Pilat Shear Zone motion**

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343 U-Pb ages on Gouffre d'Enfer Granite (sample PL06) yield two distinct populations at

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344 530-460 Ma and at 302 ± 3 Ma (Table 2). The two age populations were obtained on distinct

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345 crystals; thus, two interpretations are possible: i) It may be considered that the oldest ages

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346 may correspond to inherited zircons from surrounding rocks or ii) to the granite source rock,

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347 this age being close to the Cambrian age of 545 ± 14 Ma (Rb-Sr method) obtained by Caen-

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348 Vachette [1979] on a leptynite from the Pilat Unit. The youngest population may be

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349 interpreted as the emplacement age of the granite. However, assuming that the source of

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350 Gouffre d'Enfer granite contains zircons of 530-460 Ma, the partial melting of this rock

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351 would have probably dissolved some zircons, while other grains included in minerals that are

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352 not involved in the fusion reaction would remain intact. The CL images (Fig. 5) show a very

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353 dark border around the old zircons, this border too small to be measured can however be

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354 synchronous with the youngest zircon population thus we consider that the oldest zircons

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355 are inherited from the source of the granite. The $^{40}\text{Ar}/^{39}\text{Ar}$ dating of synkinematic biotite

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356 for the same rock yields an age of 303.0 ± 1.6 Ma (Table 2) that can receive two different

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357 interpretations. It could correspond to the retention of radiogenic argon at temperature of

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358 320-350 °C during the cooling of the granite at relatively fast rates (Harrison et al., 1985).

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359 Alternatively, this age could be considered as a crystallization age in agreement with many

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360 studies showing that mica ages recorded by deformed rocks are mostly under the control of

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361 recrystallization and dissolution/precipitation effects, with a minor influence of volume

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362 diffusion processes (Chopin and Maluski, 1982; Di Vincenzo et al., 2004; Augier et al., 2005;

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363 Villa, 2014; Bosse and Villa, 2019). However, because the syn-kinematic granite has

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emplaced in hot surrounding rocks at temperatures above 700°C, it is likely that the meaning of the biotite age at 303 Ma lies between the two interpretations. The closure temperature in the zircon is considered as ~ 900°C (Cherniak and Watson, 2001), much higher than those of sialic magmas crystallization. We will thus adopt the peak metamorphic temperature at 750-700°C which was recorded by the surrounding metamorphic rocks as the zircon growth temperature. Considering the uncertainties on the ages obtained from the two dating systems, the radiometric data of both chronometers are overlapping. Thus, the cooling between 700°C and 355°C could have occurred either in 2 My or instantaneously. Both hypotheses however indicate that the Gouffre d'Enfer leucogranite experienced a quasi-instantaneous cooling at an average rate of 150 °C/My or more.

The micaschists of the Pilat Unit underwent a low pressure-high temperature (5-6 kbar, 700 °C) metamorphic stage coeval with the development of the PSZ and the emplacement of the syn-kinematic Gouffre d'Enfer granite (stage 2 in the P-T evolution of the Fig. 5 in Gardien, 1990). The granite and the surrounding micaschists were exhumed to the surface before or slightly before the opening of the St. Etienne basin as they are found as blocks in the basal conglomerates of the St Etienne sedimentary basin. Assuming a depth of 18-15 km for the granitic magma emplacement and its final crystallization at 5-7 km in the upper crust as indicated by the presence of retrograde andalusite and pyrophyllite in the micaschists, (stage 3 in the P-T evolution of the Fig 5 in Gardien, 1990) this very rapid cooling is equal to an exhumation rate of ~6 mm/yr. Such result is not incongruous insofar rapid exhumation of deep crustal levels in mountain systems is generally related to extension after thermal softening of a thickened continental lithosphere (Johnston et al., 2000).

The $^{40}\text{Ar}/^{39}\text{Ar}$ biotite age of the Gouffre d'Enfer granite is like the biotite age of 302.9±1.5 obtained on the post kinematic leucogranite dyke (PL 43) coming from the

388 footwall of the fault. Muscovite $40\text{Ar}/39\text{Ar}$ ages of 301.5 ± 1.6 Ma and 302.9 ± 1.6 Ma obtained
389 on pegmatitic synkinematic granites (samples PL 49 and PL51 from the PSZ) and of
390 303.9 ± 1.6 Ma of the lineation of a brittle fault plane (PL48) are statistically similar to the age
391 of the biotite. They are considered to narrowly date their crystallization according to the
392 references cited above showing that white mica ages in crustal scale shear zones are better
393 interpretable to record their (re) crystallization at temperatures up to 550°C . Therefore,
394 these $40\text{Ar}/39\text{Ar}$ ages on syn- and post-kinematic biotite and muscovite that crystallized at
395 different crustal levels suggest that the rocks of the Pilat fault footwall were exhumed to the
396 upper crust during a relatively short-lived tectonic event. Muscovite from the post kinematic
397 leucogranite dyke PL 43 gives an age at 298.5 ± 1.5 Ma younger than the biotite age from the
398 same body. This age inversion with respect to the generally assumed closure temperatures
399 of both micas suggests that muscovite could record a crystallization age related to the late
400 infiltration of fluids in the leucogranite along fractures.

401 The PSZ bordering the northern part to the Velay dome and the granite sills emanating from
402 it (as for example the Gouffre d'Enfer granite) corresponds to the footwall of an extensional
403 fault system allowing the exhumation of the Velay dome. The exhumation of the dome
404 created a relief that was eroded, resulting in coarse-grained clastic sedimentation preserved
405 as boulders in the conglomeratic and brecciated formations at the base of the sedimentary
406 sequence filling the Late Carboniferous basin.

4.2. Relationship between the PSZ and the St Etienne Basin

408 The basin corresponding to the hangingwall of the fault was filled by sediments
409 comprising very coarse conglomerates and sandstones, including migmatite and granite
410 from the Velay dome and micaschists from the Pilat Unit. The nature of the clasts and their
411 imbrication (Fig. 7) indicate a southern origin confirming that the main relief bordering the

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412 Stephanian coal basin was the Pilat-Velay area. The U-Pb age of 310.5 ± 3.4 Ma obtained on
413 the volcano clastic block (PL92) is compatible with the age Westphalian - Middle
414 Pennsylvanian of the Grand Croix conglomeratic series in which the sample PL92 was taken.
415 This new dating highlights that magmatism in the lower crust responsible for the formation
416 of the granite-migmatite Velay dome and the associated surface volcanism are synchronous
417 with the basin formation whose initiation occurred as early as 310 Ma. The new age confirms
418 that some basins across the French Massif Central could have started to open as early as the
419 Middle Pennsylvanian as for example the Decazeville and the Bosmoreau basins [Bruguier et
420 al., 2003] located along the Sillon Houiller and the Argentat fault respectively (Fig. 1B).

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421 The sedimentary series that have filled the St Etienne basin ended with a fine-grained
422 volcano-detrital sequence alternating with coal layers. The ash sample (PL90) taken at the
423 summit of the series gives a U-Pb age of 299.3 ± 1.3 Ma consistent with the Autunian—Middle
424 Cisuralian age of the surrounding sandstone and shale sediments of the Ricamary site. This
425 age agrees with 297 Ma K–Ar illite age (Bruguier et al., 2003) obtained for the sedimentation
426 in the Carboniferous Bosmoreau basin along the Argentat Fault (Fig. 1B). The St. Etienne coal
427 Basin correspond to an intracontinental rift that opened during Carboniferous time.
428 Whatever the global tectonic context, intracontinental rifting is often synonymous with
429 volcanic activity characterized by the formation of volcanoes in the grabben of the rift, very
430 often related to the astheospheric rise at the base of the thinned crust. Our
431 geochronological data show that the St. Etienne basin begins to open with such magmatic
432 activity as evidenced by the presence of the ignimbrite (PL92) dated at 310 Ma. Between 305
433 Ma and 300 Ma, extension and magmatic activity were active as indicated by both the
434 emplacement of the syntectonic granite (PL06) from the PSZ and the deposit of the cinerite
435 (PL90) sampled at the top of the sedimentary series of the St Etienne basin. Finally, the

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3 437 tectonic granite (Gay et al.), 313 Ma for an orthogneiss sill in the PSZ and 290 Ma for an
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5 438 undeformed granite (PI4, Fig. 1) cut off by the St Chamond fault (Costa 1990) suggest a long-
6
7 439 lasting activity of the PSZ. This evolution over 20 Ma would therefore comprise two stages, i)
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9
10 440 an early ductile and deep stage related to the motion of a low angle normal fault (the PSZ)
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13 441 and the exhumation of the Velay granite, ii) a second more surficial stage corresponding to
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15 442 the opening of the Carboniferous basin associated with the development of a second-high
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18 443 angle normal fault connecting in depth to the former ductile fault (Fig. 5d Malavieille et al.,
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21 444 1990). This second fault correspond to the St. Chamont fault which intersects the USG
22
23 445 gneiss to the east and surrounded by the Carboniferous basin and the Pilat series to the
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25
26 446 West (Fig. 1). Our geochronological data are in conflict with such a scheme. Indeed, we have
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28 447 demonstrated the synchronism between the ductile and fragile deformation and thus the
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31 448 opening of the Carboniferous Basin. The St. Chamond fault which limits the basin to the
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33 449 southwest and intersecting the USG gneisses to the east is therefore contemporary with the
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36 450 PSZ further suggesting a single deformation step and potentially a single extensive structure.
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39 451 However, in the absence of a detailed structural study on the fault of St. Chamond, we
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41 452 cannot exclude that the latter is a second fragile and more recent fault superimposed on the
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44 453 PSZ. This could explain the presence of a sheet of coal caught between micachists and
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46 454 controlled by vertical faults bearing horizontal striations showing that these are strike-slip
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49 455 planes (Mattauer and Matte, 1998). Finally, the ages obtained from the PSZ and those of the
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52 456 Carboniferous St Etienne basin suggest that the opening of the basin is coeval with the
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54 457 development of the PSZ contrary to what Mattauer and Matte (1998) proposed.

55 56 57 458 **4.3. Thermal evolution of the crust**

58 59 459 **4.3.1. Petrological constraints**

460 A common technique to estimate cooling and exhumation rates uses cooling ages
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3 461 from isotopic systems with different closure temperatures. This technique requires the
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5 462 knowledge of the thermal gradient associated with the studied context and be ensured that
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7 463 it remains stable during the exhumation of the rocks. During the continental collision, the
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9
10 464 Pilat series were buried at a depth of 22-25 km and underwent a prograde metamorphism of
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12
13 465 medium P and T conditions (8-9 kbar, 650°C; Gardien, 1990). This event occurred before 335
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15 466 Ma in the Pilat Unit (this study) and between 340 Ma and 360 Ma in the Lyonnais series
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18 467 (Costa et al., 1993) and corresponds to the main nappe stacking event in the eastern part of
19
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21 468 the French Massif Central. Subsequent evolution under low pressure and high temperature
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23 469 conditions (5-6 kbar, 700-750°C) implies a high thermal gradient of 50°C/km (Gardien et al.,
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26 470 1990; Gardien, 1990) which caused the partial melting of the lower part of the series and the
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28 471 formation of migmatites mainly located to the south and below the Pilat series (Fig. 1 and 2).
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31 472 This partially melted zone gradually becomes toward the south the Velay Dome which
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33 473 consists of migmatites and granites themselves formed at the expense of metapelites and
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36 474 orthogneiss (Downes et al., 1997; Ledru et al. 2001; Barbey et al., 2015) during a high
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39 475 temperature ~ 850 °C and low pressure 4-5 kbar metamorphic stage (Montel et al., 1992;
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41 476 Barbey et al., 1999, 2015). The crystallization of these magmas is dated between 310 and
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43
44 477 300 Ma (Barbey et al., 2015; Chelle-Michou et al., 2017).

46 478 The peraluminous Gouffre d'Enfer granite (sample PL06) was produced by the
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49 479 melting of pelitic protoliths (Vitel, 1988) and is thought to represent an extension of the
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52 480 Velay dome (Gay et al., 1981; Vitel, 1988). The 302 Ma age of this syn tectonic granite (U-Pb
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54 481 on zircon, this study) clearly shows that the high temperature stage metamorphism was
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57 482 coeval with the northward extensional tectonics of the Pilat Unit. The new zircon U-Pb ages
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59 483 close to the $^{40}\text{Ar}/^{39}\text{Ar}$ mica ages (303-301 Ma) of the syn-to post-tectonic granites (Samples

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3 484 PL 49, 51) from the same series characterizes a very rapid cooling of the magmatic rocks
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5 485 during their exhumation.

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7 486 To test cooling rate of this specific place we performed a 1D thermal modelling of
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9 487 both the footwall (the PSZ) and the hanging wall (The Carboniferous Basin of St Etienne) of
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11 488 the extension area. To do so we will use the geochronological, thermobarometrical and
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13 489 geometrical data of the Gouffre d'Enfer granite emplaced in the footwall of the PSZ.

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16 17 18 491 **4.3.2. 1D Thermal modelling (Figure 8)**

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20 492 The thermal evolution of the crust affected by both extension and granite sill
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22 493 intrusion has been estimated from a 1D finite difference model which considers at the
23
24 494 surface the possible sedimentation and at depth the latent heat of crystallization of the
25
26 495 magma. The heat equation with production and advection terms has been solved classically
27
28 496 by a Crank Nicholson scheme (Wilmott et al., 1995) on a grid with 100 m nodes space. We
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30 497 fixed a constant temperature of 0 °C at the top of the crust and the lower boundary of the
31
32 498 model is also kept at a constant temperature of 750 °C. Pure shear thinning of the crust is
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34 499 modelled by a linear increase of the upward velocities. The point at the surface has a null
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36 500 velocity. Geochronological and thermobarometrical data (see geochronological results) show
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38 501 that thinning was fast. In less than one My, the granite emplaced between 18 to 15km depth
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40 502 and was exhumed at a depth of 5 km with a velocity around 6mm/ year. To model such
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42 503 average exhumation velocity, an upward velocity of 0.015m/y was set at 18 km depth
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44 504 decreasing linearly to 0 at the surface. The initial geotherm is computed using an internal
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46 505 heat radioactive heat production of $1e-6$ W/m³ and a surface heat flow of 90 mW/m²
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48 506 consistent for a thickened crust (Davies & Davies, 2010). The parameters that have been
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50 507 tested are the thickness of the intrusion, the strain rate, and the sedimentation rate.

508 First, we consider a magma layer of 2 km wide corresponding to the dimensions of the
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 3 509 Gouffre d'Enfer granite that intruded instantaneously in the footwall of the fault at a depth
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 5 510 of 18 km depth. The intrusion is assumed to be contemporaneous to the crustal thinning and
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 7 511 is thus strained during its emplacement. The crystallization of the magma releases heat in
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 10 512 proportion to the cooling following equation of Bussod and Williams (1991):

$$(1) \frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} - v \frac{\partial T}{\partial z} + A - H$$

16 514 With, $T(z)$ is the temperature at depth z , t is the time, κ is the thermal diffusivity, v is the
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 19 515 advective velocity, A is the heat production by radioactivity and H is heat produced by
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 21 516 magma crystallization.

$$(2) H = \frac{L}{C(Tl - Ts)} \frac{\partial T}{\partial t}$$

27 518 With H = energy of crystallization, L = latent heat of fusion, C = heat capacity, Tl = liquidus
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 30 519 temperature, Ts = solidus temperature. Equation (1) is solved by a forward finite difference
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 32 520 scheme. The latent heat of crystallization, L , is computed at each time step. The latent heat
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 35 521 of crystallization (L) is $28 \cdot 10^4$ J/kg, the solidus temperature (Ts) is 660 °C the liquidus
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 37
 38 522 temperature (Tl) is 900 °C and the heat capacity (C) is 1000 J/kg.

523 4.3.2.1 Results

43 524 The results show that when a 2 km thick or wide layer of magma is intruded at 18 km depth
 44
 45 525 at a temperature of 750°C, the magma layer (footwall) crystallizes and cools in less than 500
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 48 526 ky when the crust is not thinned (Fig. 8A). A second case has been tested in which the crust
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 51 527 is thinned by pure shear. In this case too, the thermal anomaly produced by the magma is
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 53 528 relaxed in less than 500 ky (Fig. 8A). The modelling shows that in both cases (with or without
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 56 529 shearing and with or without exhumation) the extremely rapid cooling of a granitic sill is
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 58 530 achievable. However, the key point we can learn from the results is that because of the

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531 tightness of the isotherms the thermal field is dominated by the strain effect rather than by
532 the intrusion effect. Thus, the strain effect maintains the crust in hot thermal regime
533 allowing it to flow easily. These results allow us addressing the question of whether the Pilat
534 Shear Zone is a response to the formation of the Velay dome or is consequence. In other
535 words, either the partial melting of the continental crust responsible for the Velay formation
536 has favoured the extensive regime or this extension has induced partial melting of the
537 continental crust by adiabatic decompression. Our numerical modelling is more in
538 agreement with the second hypothesis corresponding to a dominant passive rifting model.
539 To supports this model, we also can invoke the under plating of mantle derived magmas,
540 the so called vaugnerites (Laurent et al., 2017) whose intrusion contemporaneous with the
541 Velay dome formation would have increased the gravitational forces associated with the
542 delamination of the lithosphere under the Velay dome.

543 **4.3.4 High thermal regime, the possible heating sources**

544 The high thermal regime associated with the extensional tectonic in the Pilat Unit
545 showing a widespread melting of the rocks and the formation of anatectic granites has been
546 well tested (Montel et al., 1992; Gardien et al., 1997; Ledru et al., 2001; Laurent et al., 2017).
547 The mechanism classically invoked for such context is the removal of the orogenic root by
548 delamination or rollback (Laurent et al., 2017), inducing a rapid rise of the asthenosphere
549 and a significant heating of the orogenic crust. Consequently, the magmatism is produced by
550 decompression-induced melting of the previously thickened crust and the asthenospheric
551 mantle. This lasting high heat flux after delamination and a very low-density lower crust and
552 mantle maintain the internal part of the Variscan Belt at relatively elevated altitudes with
553 respect to the thickness of the lithosphere. Hence, some authors (Becq-Giraudon et al.,
554 1996) relying on sedimentary structures and paleontological data suggested that at the

555 period of the extensional tectonics in the Pilat Unit and the opening of the St Etienne Coal
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2 556 Basin the elevation of this part of the belt could have been ~ 4000 m.
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5 557 **4.3.5. Exhumation mechanism**

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7 558 The fast exhumation rates of the Pilat metamorphic Unit and the Velay dome require
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9 geological factors including not only a heat source to provide the partial melting of the lower
10 559 to middle orogenic crust but also driving forces allowing a rock uplift. In that case, the
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12 560 presence of melt in the crust would induce the localisation of strain within shear zones, the
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14 561 latter being a suitable structural geometry able to exhume the deep melted crust. These
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16 562 conditions are present in a scenario involving the removal of part of the lower lithosphere
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18 563 beneath the Velay dome.
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25 565 Paragneiss and orthogneiss are the dominant constituents of the nappes pile formed
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27 during the continental collision at ~345 Ma. They are known to be partially molten during
28 566 the Carboniferous (345-315 Ma) as evidenced by the formation of migmatites at medium
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30 567 pressure-high temperature conditions (Montel et al., 1992; Ledru et al., 1994; Bé Mézème et
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32 568 al., 2005; Faure et al., 2009; Barbey et al., 2015), to the point of forming plutons like the
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34 569 Tournon granite, the Margueride granite, the Civrieux granite, all dated between 335 and
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36 570 320 Ma (Laurent et al., 2017 and ref therein). Hence, anatectic domes such as the Velay
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38 571 dome could have developed earlier in the Variscan belt history, especially because extension
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40 572 was already active since the Moscovian. Thus, as such dome appears during the late
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42 573 Carboniferous (Gzhelian), removal of the lithospheric mantle remains, therefore, the most
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44 574 plausible cause of the sudden Pilat/Velay uplift. The Velay a poly-diapir dome corresponds to
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46 575 a multistage emplacement of granites superimposed on extensional structures (Laurent et
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48 576 al., 2017]. During this process, the dome was shifted horizontally southward (Lagarde et al.
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50 577 1994; Ledru et al., 2001) carrying during its migration rafts of metamorphic units mainly
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1 affected by northward normal fault (Gardien and Lardeaux, 1991; Gardien, 1993; Chelle-
2 Michou et al., 2017). Thus, granites forming the dome intruded the main detachment fault
3 whose ~35 km amplitude (Fig 1A) (Malavieille et al., 1990) and 10 to 30° angle towards the
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5 581 North can accommodate the rise of the dome from 20 km depth.
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10 583 **4.3.6. Thermo-mechanical evolution**

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13 584 Granites and migmatites resulting from partial melting of the continental crust and
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15 585 the lithospheric mantle are abundant in the Variscan chain testifying of the high thermal
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18 586 regime reached during its evolution (Zwart, 1967; Letterrier, 1978; Cuney et al., 1990;
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21 587 Bussien et al., 2008; Von Raumer et al., 2013, Laurent et al., 2017). However, among these
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23 588 plutons very few have the characteristics of domes (i.e., concentric shape of foliations,
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26 589 horizontal structures at the roof progressively tilted to the vertical at its margin, detachment
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28 590 faults bordering the pluton and responsible for its exhumation). The Velay massif has a dome
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31 591 feature (Burg and Vanderhaeghe, 1993; Lagarde et al., 1994; Ledru et al., 2001) reflecting
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33 592 gravitational instabilities that develop at the base of the orogenic crust coeval to the late
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36 593 extensional evolution of the belt (Van den Driesche and Brun, 1992; Gardien et al., 1997;
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39 594 Ledru et al., 2001). The mechanisms invoked for the development of these domes consistently
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41 595 imply lithospheric tear (Ledru et al., 2001; Laurent et al., 2017). Whether it is slab break-off
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43
44 596 or slab retreat remains an open question. Both mechanisms require that the lithosphere
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47 597 shears horizontally or vertically in its thickness over a large distance. Moreover, the strong
48
49 598 coupling between the crust and the lithospheric mantle causing the tearing of the slab
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52 599 affects a wide area in the overriding domain. For example, the tearing of the Indian
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54 600 lithosphere is referred to for the formation of MCC North of the Shyok Suture Zone (SSZ) over
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57 601 more than 150 km in the South Karakorum and south Tibet (Rolland et al., 2001). In the
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5 602 French Massif Central, the very restricted size of the Velay dome (~7000 km²) suggests a
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8 603 punctual removal of the lithospheric mantle such as dripping.

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10 604 The geochronological data and more precisely the argon ages provide arguments in
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12
13 605 favour of this localized delamination of the mantle under the Velay. Indeed, the similarity of
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15
16 606 the micas ages of samples coming from the footwall of the PSZ and from the PSZ itself can be
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18
19 607 due to a reset of Ar ages even away from the deformation zone due to fluid ingress (Villa,
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21
22 608 2014) and metamorphic recrystallization. However, the absence of zonation and
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25 609 retromorphism on the dated micas grains (see appendix 2) does not argue in favour of
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28 610 resetting the Ar-Ar system in connection with fluid circulation. An explanation for the
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31 611 velocity and synchronism of exhumation on either side of the PSZ is that the lithospheric
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34 612 rebound linked to the mantle dripping provoke the sudden upward of the whole system,
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37 613 including the footwall and the hangingwall of the PSZ. This rebound has the effect of
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40 614 canceling the diachronism of the exhumation between the foot wall and the hanging wall
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43 615 because it is much faster than the exhumation velocity of the Velay along the Pilat
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46 616 detachment fault.

4.3.6.1 Related Evolution of the coal Basin

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49 617 The effect of sedimentation on the hanging wall of the fault has also been tested. As
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51
52 618 it concerns the upper crust, only the first five kilometers of the crust have been modelled (fig
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55 619 8B). At the beginning of the experiment, however, the model was limited to the uppermost
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58 620 2.5 km of the crust. The velocity field is kept constant and equal to the sedimentation rate
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61 621 except for the first point at the surface for which velocity is nil. The thermal parameters used
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64 622 in the model are given in Table 3.

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67 623 The effect of sedimentation rate on the thermal state of the upper crust has been tested.
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70 624 Two very small sedimentation rates of 0.005 mm/yr and 0.002 mm/yr, respectively, have

626 been applied (Fig.8B). The results show that sedimentation has a clear cooling effect on the
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2 627 base of the basin and therefore on the roof of the footwall. Indeed, the base of the basin is
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5 628 at a temperature of 60 °C after sedimentation of 5000 m at a velocity of $5 \cdot 10^{-3}$ m/yr during 1
6
7 629 My while it is only at 40 °C after sedimentation of 2000 m at a velocity of $2 \cdot 10^{-3}$ m/yr during
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10 630 1 My. The results show that sedimentation has a clear cooling effect at the base of the basin
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12 631 and therefore at the roof of the footwall. Indeed, the base of the basin is at a temperature of
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15 632 60 °C after sedimentation of 5000 m at a velocity of $5 \cdot 10^{-3}$ m/yr during 1 My while it is only
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18 633 at 40 °C after sedimentation of 2000 m at a velocity of $2 \cdot 10^{-3}$ m/yr during 1 My. The
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20 634 modelling shows that the supply of cold material (sediments) on top of the crust prevents
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23 635 the heat diffusion throughout the basin. Thus, a cold thermal gradient prevail in the upper
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26 636 crust (hanging wall) during extension eventually preventing the resetting of the geo
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28 637 chronometers during extensional. The northern part of the Pilat metamorphic series located
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31 638 just below the St. Chamond fault (fig.1) where the basin opened and the sediments were
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33 639 deposited, have preserved the $^{40}\text{Ar}/^{39}\text{Ar}$ age of ~335 Ma that could be assigned to the
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36 640 stacking nappes stacking event during the continental collision
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39 641 In addition, these results show that independently of the rate of sedimentation, the
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41 642 temperature at the base of the basin remains low precluding thermal degradation of the
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44 643 organic matter (120-200°C). Thus, according to the model, the temperatures obtained during
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46 644 the sedimentation would prevent the formation of coal. It suggests that the formation of the
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48
49 645 coal layers is not in relation with the formation of the PSZ and the exhumation of Velay. The
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51 646 history of the maturation of the organic matter should therefore be the subject of a specific
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54 647 and detailed study in relation to the history of sedimentation in the basin.
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57 648 **5. Conclusions**

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649 The extensional ductile shear zone of the Pilat has been dated through the Gouffre d'Enfer
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2 650 syn-kinematic leucogranite at 301.8 ± 3.1 Ma by U-Pb on zircon and at 303.0 ± 1.6 Ma by the
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5 651 $^{40}\text{Ar}/^{39}\text{Ar}$ method on biotite.

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7 652 Argon ages on biotite and muscovite from synkinematic to postkinematic granites
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10 653 from the Pilat shear zone allow the following extensional fault system to be dated from
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12 654 ductile (syn kinematic pegmatites PL 49 and PL51) to brittle (post kinematic leucogranite
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15 655 dyke PL 43 and fault plane PL48) stages between 301.5 ± 1.6 Ma and 303.9 ± 1.6 Ma.

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18 656 The ages obtained in this study suggest a "rejuvenation" of 20 Ma of the PSZ and thus
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20
21 657 of the extensional tectonic on the northern border of the Velay Dome connecting it to the
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23 658 generalized Northeast-Southwest Late Carboniferous to Permian extension in the Variscan
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26 659 Belt. By rejuvenating the PSZ we demonstrate that it not a transient extensive episode but
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28 660 the ending extensional phase affecting the Variscan belt. These new geochronological data
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31 661 indicate a very fast cooling and exhumation rates (150 °C/Myr and ~ 6 mm/yr respectively) of
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33 662 the Gouffre d'enfer granite whose emplacement in the upper crust is due to the motion of
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36 663 the extensional PSZ also responsible for the exhumation of the Velay Dome. We interpret
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39 664 such fast exhumation rate as the result of punctual lithospheric dripping process beneath
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41 665 the Velay dome.

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44 666 Finally, for the first time we provide absolute ages (U-Pb on zircon) for the opening
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46 667 and filling of the St. Etienne carboniferous basin ranging between 310.5 ± 3.4 Ma
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49 668 (conglomeratic base to the sequence) and 299.3 ± 1.3 Ma (ash layer to the top of the
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51 669 sequence), respectively.

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1927 **Figure captions**

1928 Figure 1 **A**: The French Massif central in France. The black arrows represent the Namuro-

1929 Westphalian extension oriented NW-SE, the red arrows represent the Carboniferous-Early Permian

1930 extension oriented NNE-SSW **B**: Location of the main Carboniferous and Permian basins in the French

1931 Massif Central. 1 = St Etienne basin, 2= Decazeville Basin, 3 = Bosmoreau Basin. AF = Argentat Fault,

1932 SHF = Sillon Houiller Fault. Kinematic of the extensional Namurian Westphalian (black arrow) and

1933 Stephanian Early Permian (red arrow) stages, C: Geological and structural map of the Pilat series and

1934 of the St Etienne coal basin. A-B and C-D are the cross sections in figure 2. Location of the studied

1935 samples together with the radio-chronological available ages in black (this study) and in red (Costa,

1936 1990).

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1938 Figure 2: Northwest-Southeast oriented sections (located in Fig 1C) across the St Etienne coal basin,

1939 the Pilat metamorphic series and the Velay migmatitic dome, PSZ: Pilat shear Zone. Location of the

1940 studied samples. Note the red arrows for PL90, PL92 and PL4. Indeed, those samples are not localized

1941 on the trajectory of the cross-sections, they were reported by projection on the cross-section AB for

1942 PL90 and PL4 and, on the cross-section BC for PL92, to report their position with respect to the PSZ

1943 and the other dated samples.

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945 Figure 3: Outcrop view of the syntectonic Gouffre d'Enfer granite (PL06) **a**: northward mineral
946 lineation (biotite/sillimanite) on the foliation plane, **b**: Plane-polarized optical image of the granite
947 showing the foliation plane / shear band relationship, **c**: Gouffre d'Enfer leucogranite intrusive in the
948 micaschists, **d** and **e**: Schmidt stereonet, lower hemisphere, of foliation/lineation and fault/striation
949 data in the Gouffre d'Enfer leucogranite **f**: Outcrop view of the microgranite dyke (PL43) crosscutting
950 migmatitic orthogneiss and **g**: cross-polarized optical image of the microgranite, **h**: Plane-polarized
951 optical image of the migmatitic orthogneiss showing the foliation plane / shear band relationship, **i**:
952 Outcrop view of the migmatitic orthogneiss **J,k, l, m**: Outcrop view of the pegmatitic granitic sill
953 (PL49) cross cut by brittle normal faults; **n** and **o**: Schmidt stereonet lower hemisphere of the
954 schistosity (S) and fault data in the pegmatitic sill PL49.

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956 Figure 4: **a**) Outcrop view of the northern micaschists series PL94 crosscut by Quartz-chlorite vein; **b**,
957 **c** and **d**: Crossed polarized optical image and details of the rhyolite sample (PL 92); **e**: Crossed
958 polarized optical image of the cinerite (PL 90). Qtz = quartz, Mus = muscovite, Bio = biotite, Plg =
959 plagioclase, ves = vesicle.

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961 Figure 5: LA-ICPMS zircon U-Pb dating results from the Pilat Unit samples. **a**) Tera-Wasserburg
962 diagrams of the dated zircons from the Gouffre d'Enfer granite (sample PL06), **b**) representative
963 zircon CL-images for each dated sample and the position of the SHRIMP spot is highlighted. **c**) Tera-
964 Wasserburg diagrams of the dated zircons from the cinerite (sample PL90), **d**) representative zircon
965 CL-images for each dated sample and the position of the SHRIMP spot is highlighted. **e**) Tera-
966 Wasserburg diagrams of the dated zircons from the ignimbrite (sample PL92), **f**) representative zircon
967 CL-images for each dated sample and the position of the SHRIMP spot is highlighted.

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969 Figure 6: Argon ages for granites and micaschist samples from the Pilat Unit. **A**: biotite age spectrum

970 of the granite PL06; B: biotite age spectrum of the microgranite dyke PL43, C: Muscovite age
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2 971 spectrum of the microgranite dyke PL43; D: Muscovite age spectrum of sample PL48; E: Muscovite
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4 972 age spectrum of the pegmatitic granite sill PL49; F: Muscovite age spectrum of the pegmatitic granite
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7 973 sill PL51; G: Muscovite age spectrum of the micaschist PL94.
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9 974
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11 975 Figure 7: Rose diagrams of the river current directions from pebble imbrications of the conglomerate
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14 976 in the basal sediments, from cross sections 1 (A) and 2 (B) localized in figure 1.
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18 978 Figure 8(A): Cooling of a magmatic intrusion at 18km depth with an initial temperature of 750 °C, one
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21 979 geotherm each 50ky; left) without thinning of the crust. The thermal anomaly is less than 0.5 Myr;
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23 980 right) with a pure shear thinning of the crust (strain rate of $4 \cdot 10^{-12} \text{ s}^{-1}$), the thermal anomaly is also
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25 981 less than 0.5 Myr. What is dominant on the thermal field is the strain effect more than the intrusion.

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28 982 (B): Thermal evolution of the upper crust with sedimentation effects. The sedimentation has a clear
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30 983 cooling effect by bringing cold sediment at the top of the basin. Left) The base of the basin is at a
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32 984 temperature of 60°C after a sedimentation of 5000m at a velocity of $5 \cdot 10^{-3} \text{ m/yr}$ during 1 Myr and
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34 985 right) at a temperature of 40°C after a sedimentation of 2000m at a velocity of $1 \cdot 10^{-3} \text{ m/yr}$ during 1
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36 986 Myr. These temperatures are incompatible with coal formation comprises between 150°C and 200°C.
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40 987 **Tables captions**

41
42 988 Table 1: Synthesis of the Ar-Ar ages obtained on the Pilat samples.

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44 989 Table 2: Synthesis of the ages obtained on the Pilat samples.

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46 990 Table 3: Thermal data used in the numerical simulation.
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50 51 992 **Supplementary Data**

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54 993 Appendix 1 : cathodo-luminescence pictures of the zircons from Gouffre d'Enfer syntectonic granite
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56 994 (sample PL06)
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995 Appendix 2a : U-Pb zircon data set of PL06 Gouffre d'Enfer syntectonic granite ; 2b : U-Pb zircon data
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2 996 set of samples PL90 and 92 respectively a cinerite and an ignimbrite from the carboniferous basin of
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4 997 St. Etienne.
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7 998 Appendix 3 : Detailed argon data of the biotite from samples PL06 and PL43, of the muscovite from
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9 999 samples PL43, PL48, PL49, 51 and, PL94.
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11 1000 Appendix 4 : Composition map and punctual analyzes of the micas dated in this study.
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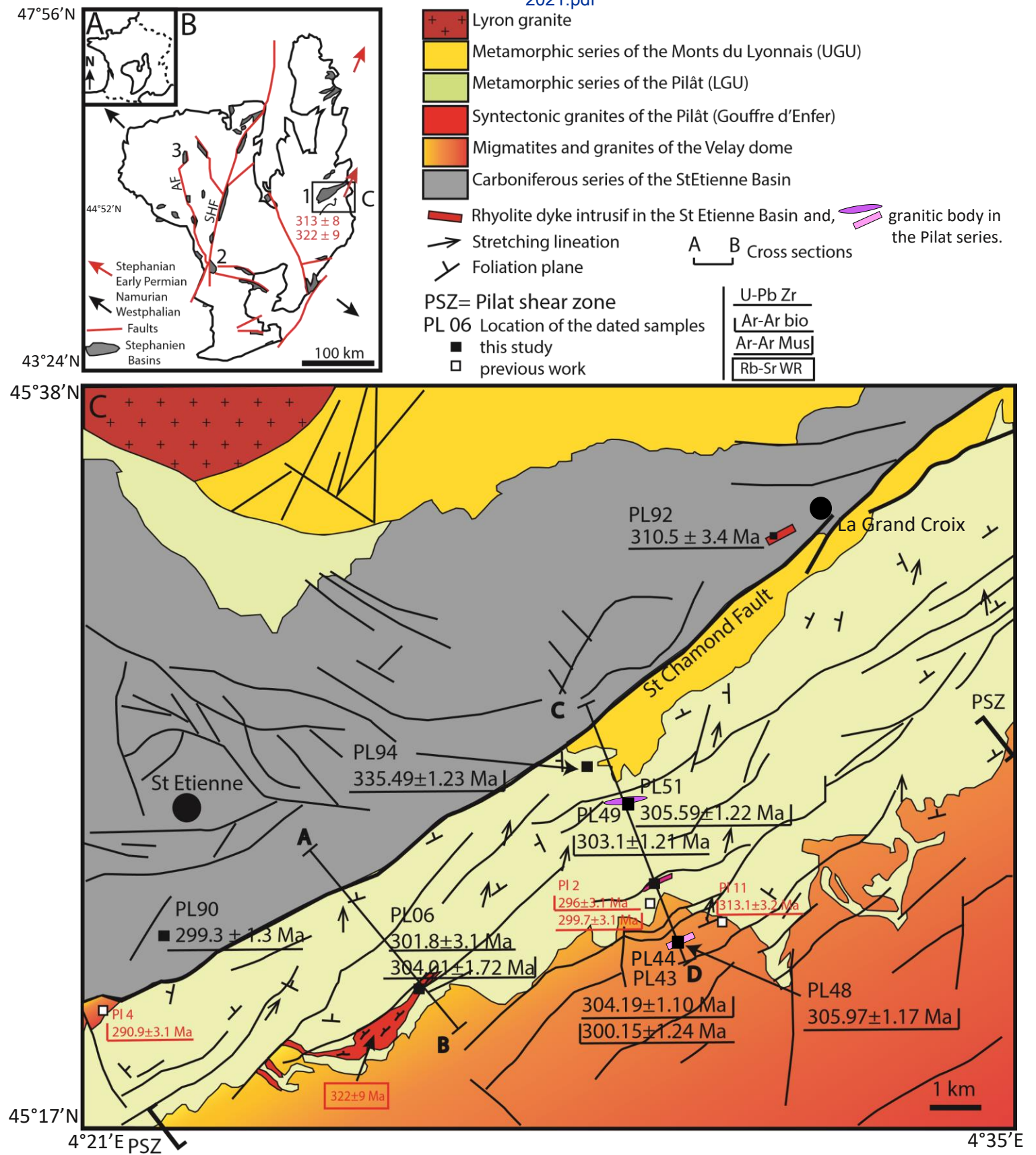


Fig. 1

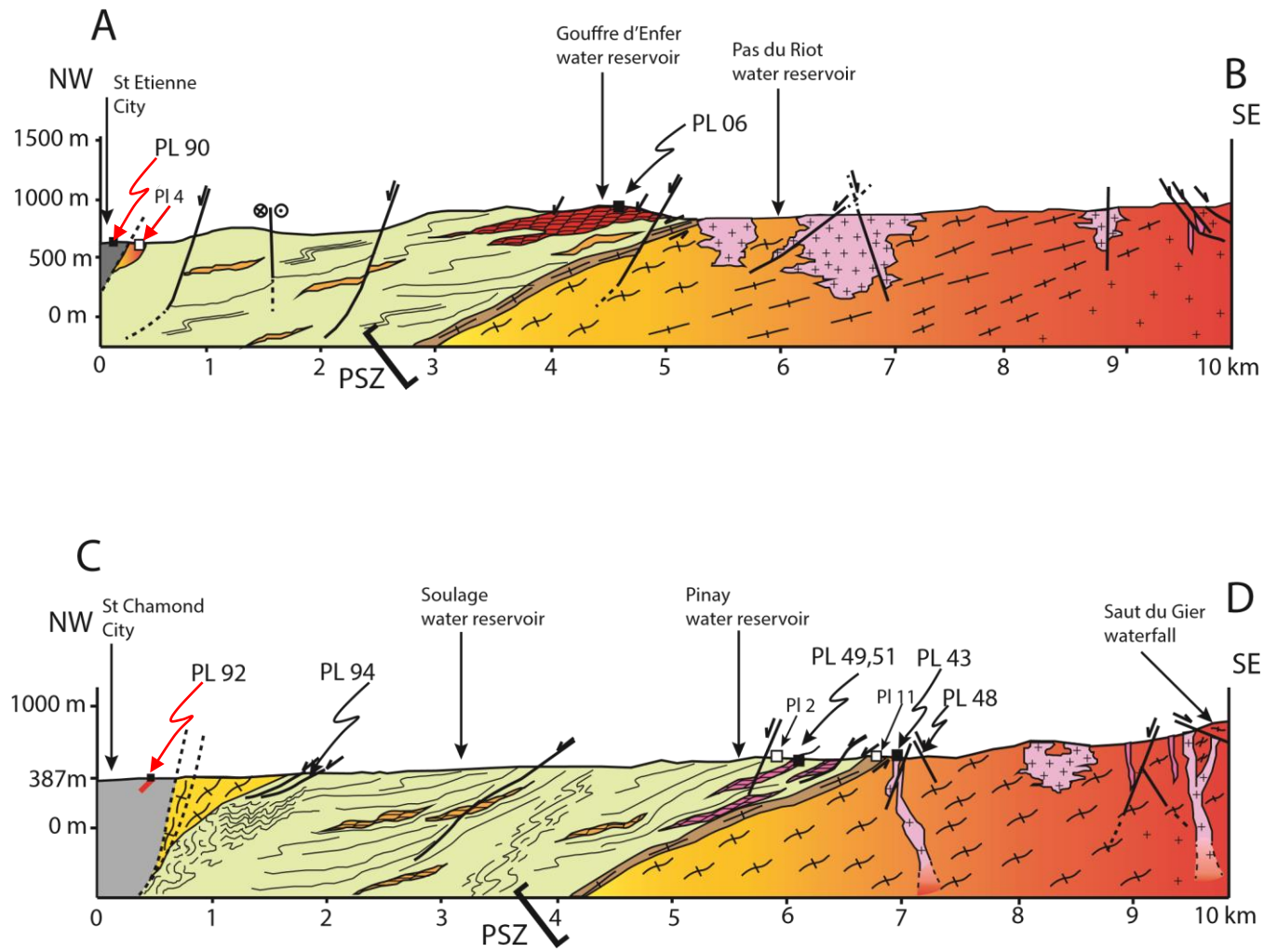


Fig. 2

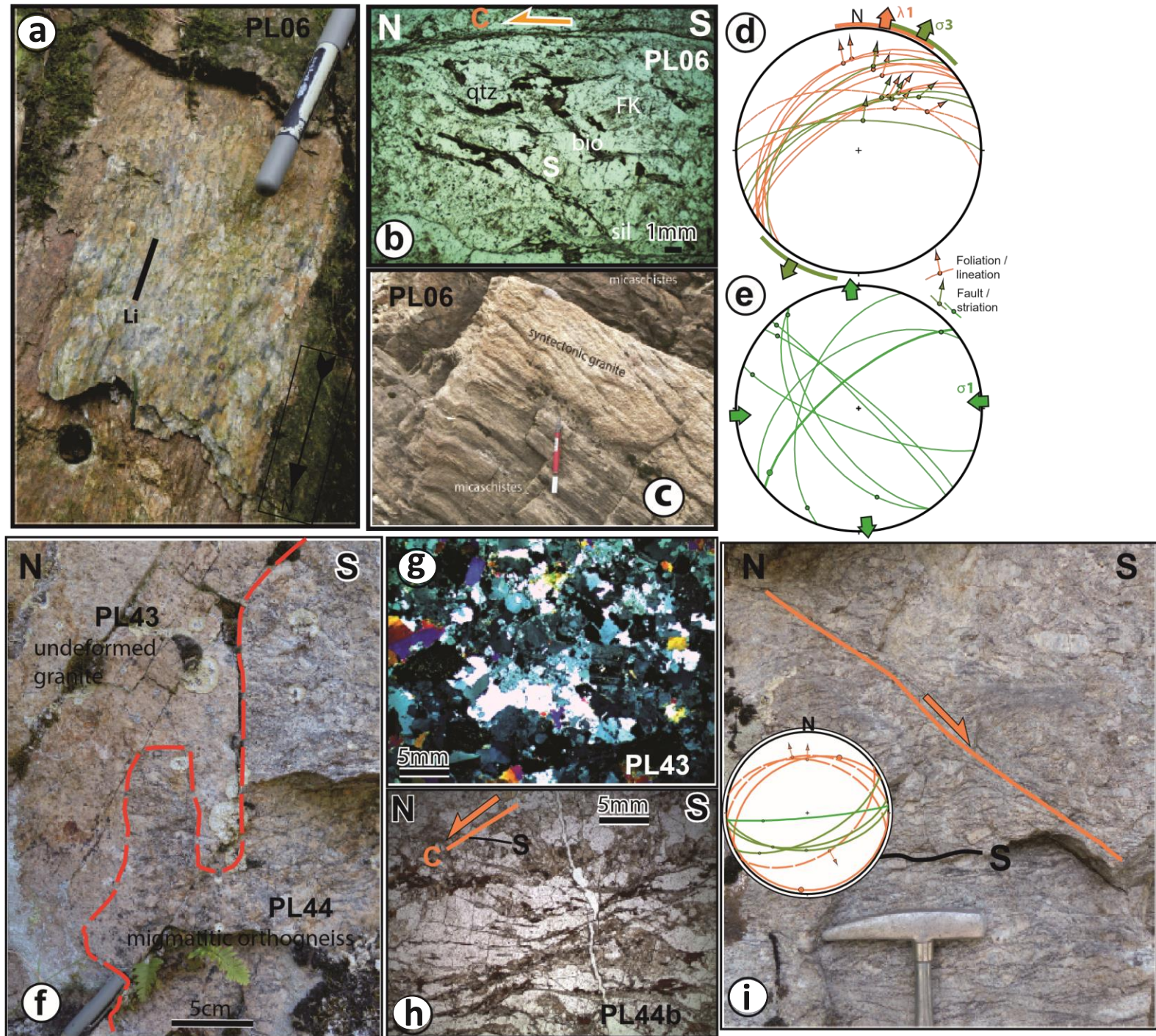


Figure 3

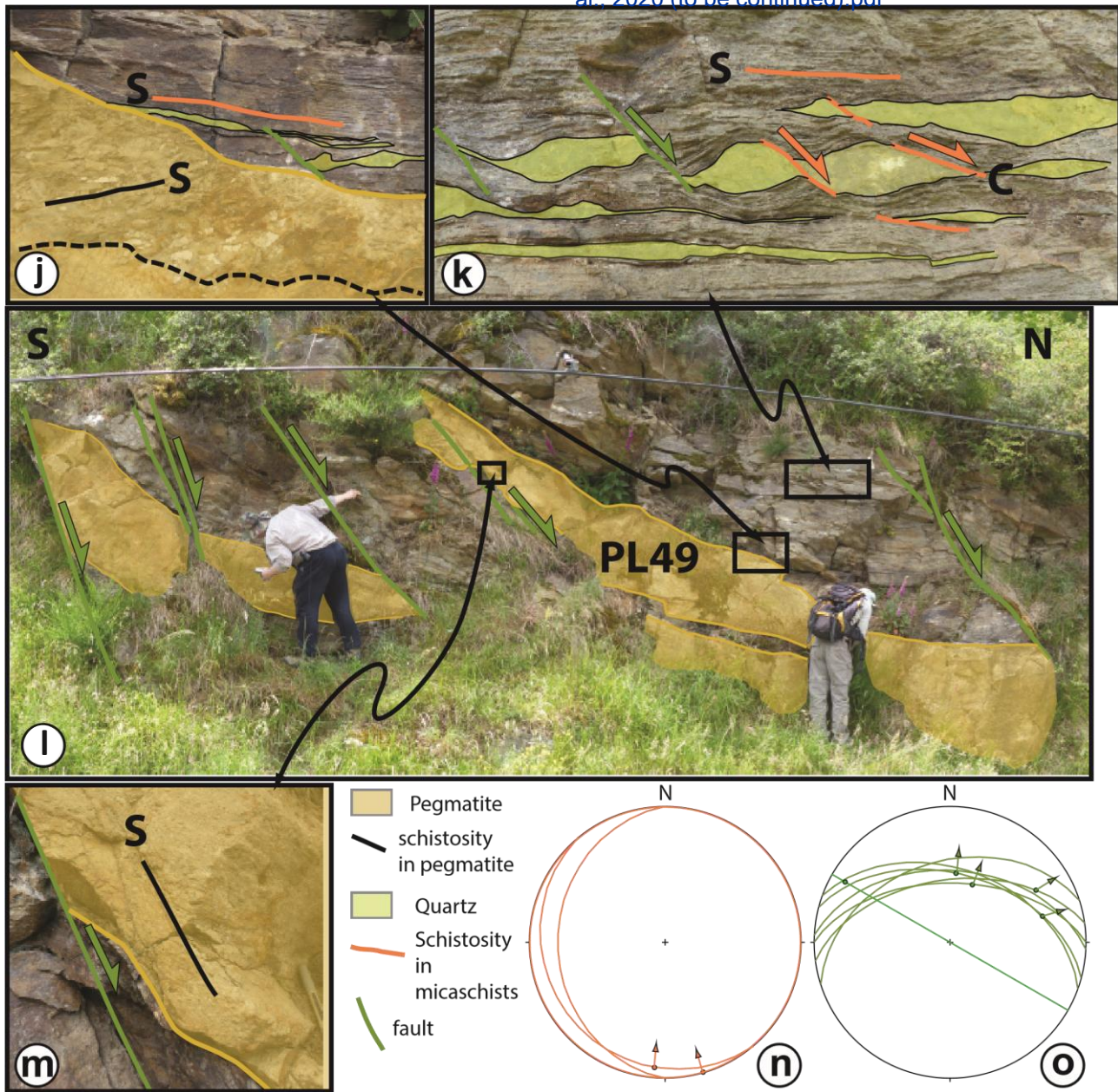


Figure 3 (to be continued)

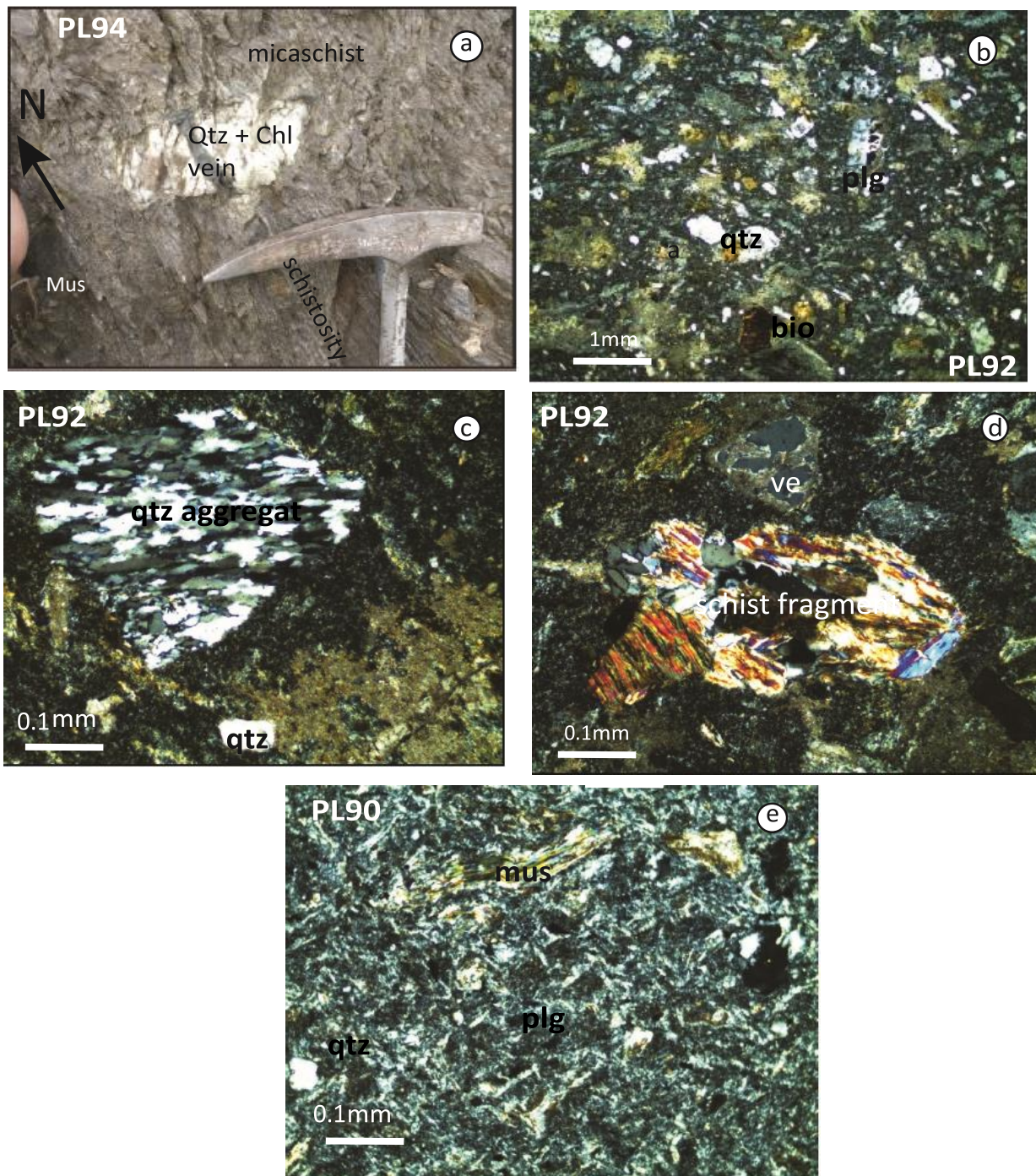


figure 4

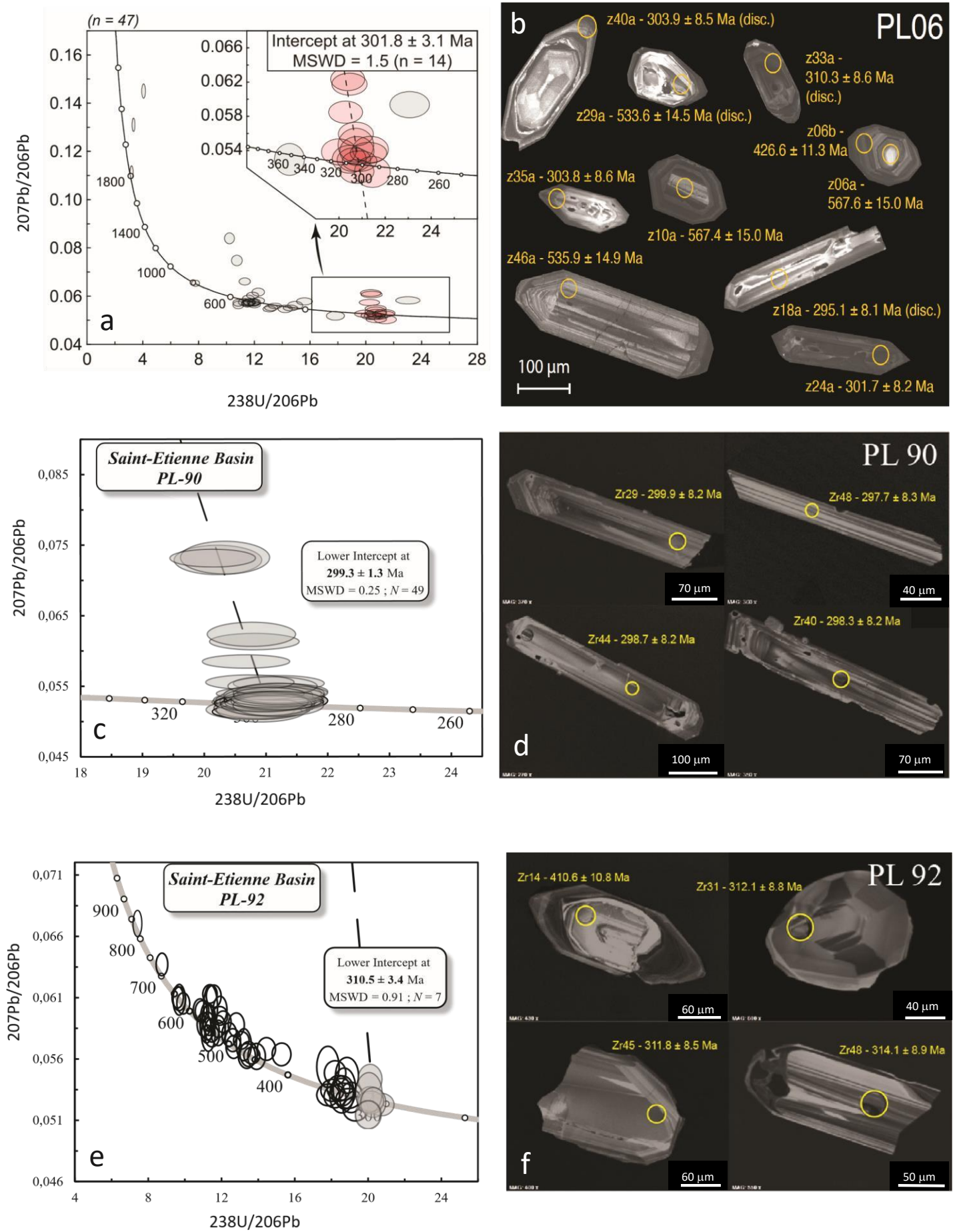
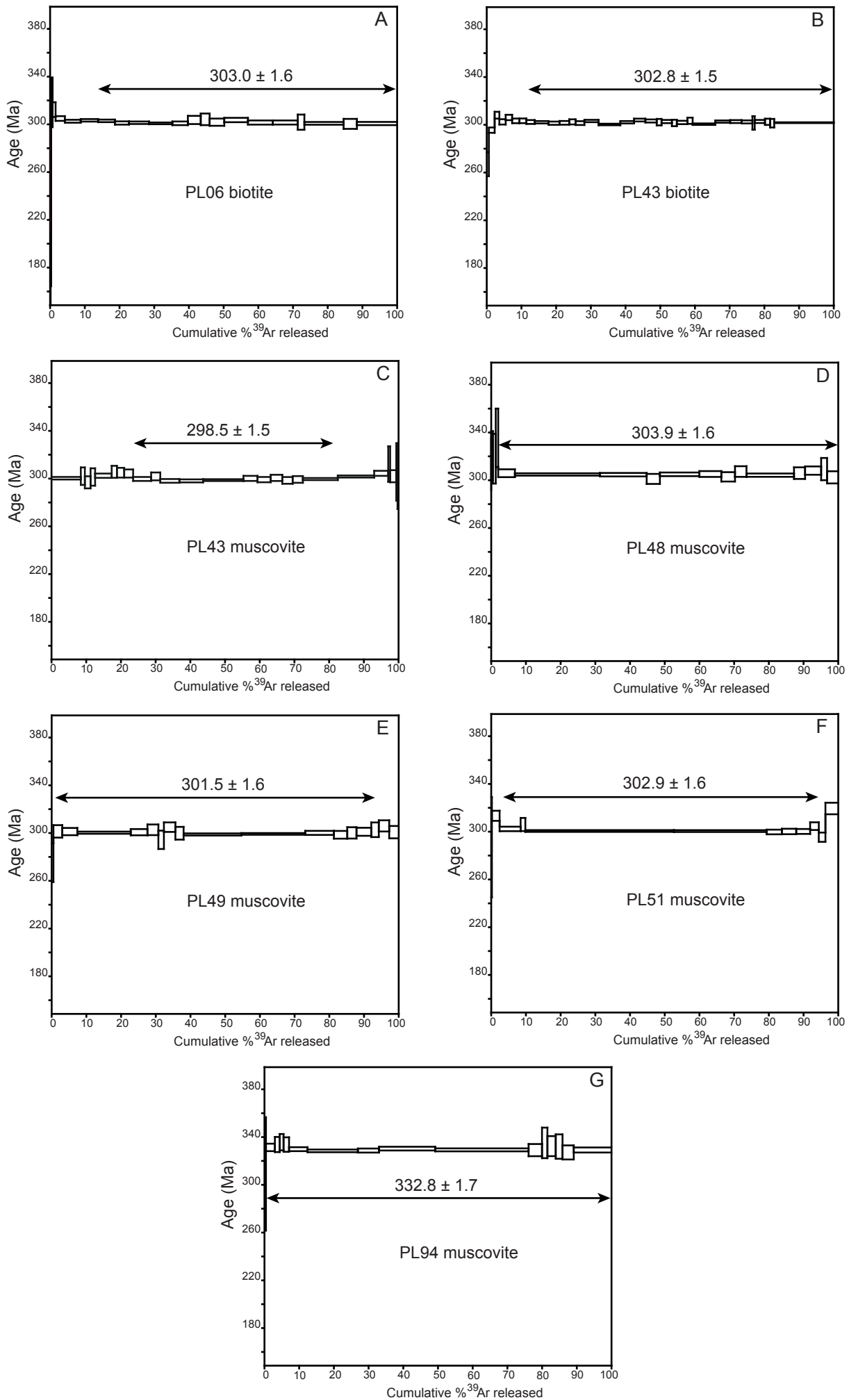


Figure 5



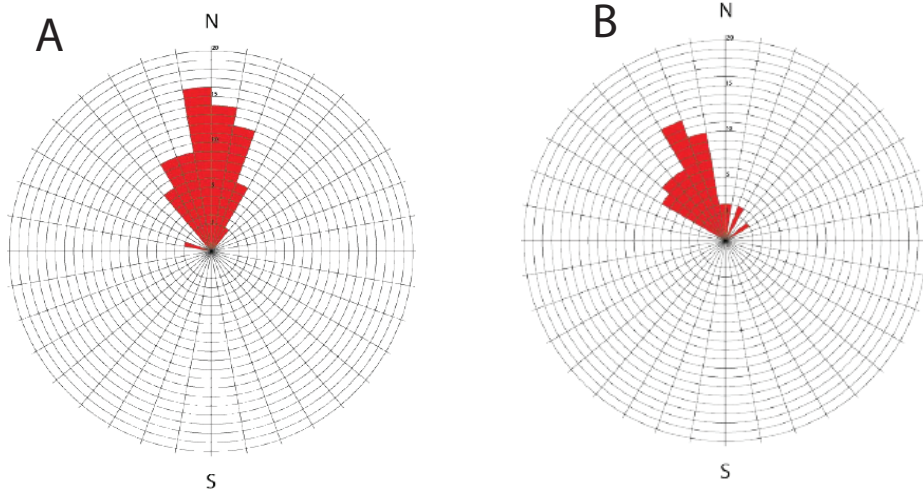


Figure 7

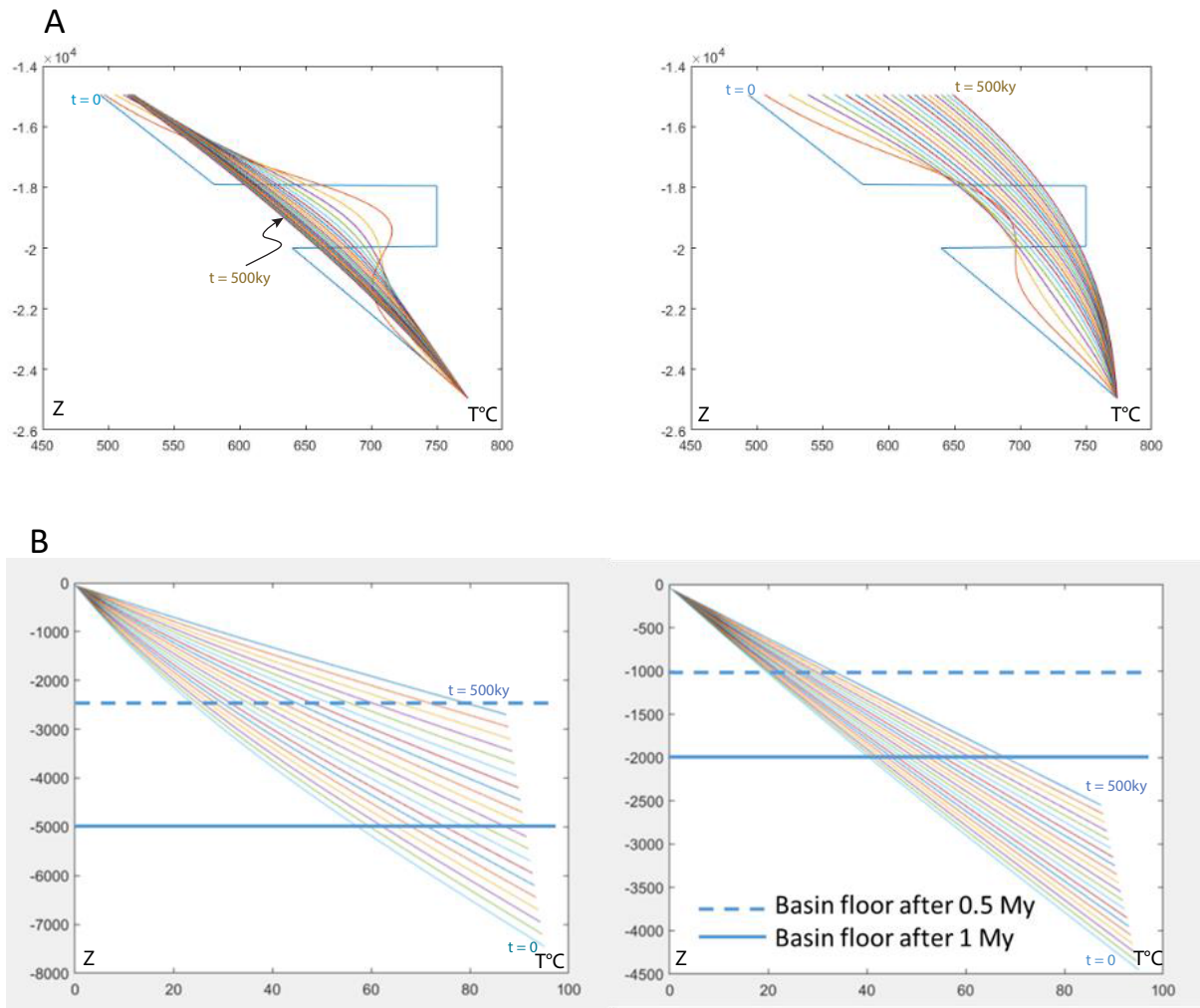


Figure 8

Sample	Total gas age	Plateau age	%³⁹ Ar (steps)	MSWD	Intercept age	⁴⁰Ar/³⁶Ar	MSWD
PL06 bio	303.1 ± 1.6	303.0 ± 1.6	86.2 (8-22)	1,6	302.9 ± 1.2	300 ± 24	1,74
PL43 bio	302.7 ± 1.5	302.9 ± 1.5 *	88.6 (8-32)	3,25	302.3 ± 1.3	350 ± 70	3,17
PL43 mus	299.8 ± 1.5	298.5 ± 1.5	58.7 (9-19)	1,24	297.5 ± 2.3	373 ± 142	1,25
PL48 mus	303.9 ± 1.6	303.9 ± 1.6	97.5 (5-17)	1,02	303.1 ± 1.3	398 ± 183	1,10
PL49 mus	301.9 ± 1.6	301.5 ± 1.6	93.8 (2-16)	1,58	301.4 ± 1.2	302 ± 5	1,52
PL51 mus	303.8 ± 1.6	302.9 ± 1.6	92.0 (3-10)	2,05	302.6 ± 1.1	304 ± 6	0,99
PL94 mus	333.1 ± 1.7	332.8 ± 1.7	100 (1-16)	1,14	332.6 ± 1.3	306 ± 16	1,28

Table 1

Sample	Locality	Latitude	Longitude	Facies	Mineral	Method	Age (Ma)	±
PL06	Gouffre d'Enfer	45.395008°	4.447512°	Deformed leucogranite	Zircon	U/Pb	301.8	3.1
					Biotite	Ar/Ar	304.01	1.72
PL43	Saut du Gier	45.399960°	4.548759°	Undeformed granite	Biotite	Ar/Ar	304.19	1.1
					Muscovite	Ar/Ar	300.15	1.24
PL48	Scie du Bost	45.399960°	4.548759°	Fault plane	Muscovite	Ar/Ar	305.97	1.17
PL49	Moulin de Cezignieux	45.419400°	4.529815°	Weakly deformed pegmatite	Muscovite	Ar/Ar	303.01	1.21
PL51	Moulin de Cezignieux	45.419400°	4.529815°	Weakly deformed pegmatite	Muscovite	Ar/Ar	305.59	1.22
PL90	La Ricamarie	45.415652°	4.371722°	Cinerites	Zircon	U/Pb	299.3	1.3
PL92	La Grand-Croix	45.506422°	4.565688°	lithic-rich ignimbrite	Zircon	U/Pb	310.5	3.4
PL94	Marie de Châteauneuf	45.322053°	4.385608°	Micaschists	Muscovite	Ar/Ar	335.49	1.23

Table 2

k	thermal conductivity, 2.5 W/m/°
ρ	rock density, 2800kg/m ³
c	heat capacity of rocks, 1400 J/kg
χ	thermal diffusivity of rocks, 1. 10 ⁻⁶ m/s ²
A	internal heat production by radioactivity, 1.10 ⁻⁶ W/m ³
Q	surface heat flow, W/m ²
C	latent heat of crystallization, 280. 10 ³ J/kg
Tl	liquidus temperature, 900°C
Ts	solidus temperature, 700°C

Table 3