



HAL
open science

Paleotemperature investigation of the Variscan southern external domain: the case of the Montagne Noire (France)

Clément Montmartin, Michel Faure, Hugues Raimbourg

► To cite this version:

Clément Montmartin, Michel Faure, Hugues Raimbourg. Paleotemperature investigation of the Variscan southern external domain: the case of the Montagne Noire (France). Bulletin de la Société Géologique de France, 2020, 10.1051/bsgf/2020043 . insu-03086940v1

HAL Id: insu-03086940

<https://insu.hal.science/insu-03086940v1>

Submitted on 23 Dec 2020 (v1), last revised 31 Mar 2021 (v2)

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 **Paleotemperature investigation of the Variscan southern external**
2 **domain: the case of the Montagne Noire (France)**

3 **Étude de paléotempérature dans le domaine méridional externe varisque: le cas de la**
4 **Montagne Noire (France)**

5 Clément Montmartin*, Michel Faure, Hugues Raimbourg

6 *Institut des Sciences de la Terre d'Orléans (ISTO), UMR 7327, Université d'Orléans-CNRS-*
7 *BRGM, Campus Géosciences, 1A Rue de la Férollerie 45071 Orléans Cedex 2, France*

8 * Corresponding author, clement.montmartin@univ-orleans.fr, 02 38 49 25 73

9 **Keywords:**

10 Raman Spectroscopy of Carbonaceous Material (RSCM) – Spectrométrie Raman sur Matière
11 Carbonée (RSCM)

12 Granite-migmatite dome - Dôme granite-migmatite

13 Montagne Noire - Montagne noire

14 Variscan orogen - Orogenèse Varisque

15 French Massif Central - Massif Central Français

16 Recumbent folds – Plis couchés

17 **Résumé**

18 La Montagne Noire, située dans la partie sud du Massif Central français, représente la partie
19 nord de l'avant-pays varisque. La zone est divisée en trois parties. Le dôme granite-
20 migmatite de la zone axiale est entouré de séries sédimentaires paléozoïques pas ou
21 faiblement métamorphisées. Les flancs nord et sud du dôme de la Montagne Noire sont
22 déformés par des plis d'échelle kilométrique, déversés vers le sud sud-est. La méthode de
23 spectrométrie Raman sur matière carbonée (RSCM), réalisée dans les roches de bas grade
24 métamorphique du flanc sud de la Montagne Noire, a donné des températures comprises
25 entre 400°C près du dôme et 230°C dans le domaine sud. Trois géothermomètres Raman

26 ont été utilisés pour couvrir cette gamme de température. Ces températures RSCM sont
27 qualitativement conformes aux estimations précédentes basées sur la cristallinité de l'illite,
28 l'altération de la couleur des conodontes et les inclusions fluides effectuées dans la même
29 zone, qui démontrent une augmentation de la température vers le dôme.

30 Les isothermes traversent les différents contacts de nappe et sont orientées
31 parallèlement à la marge sud de la zone axiale. Cette distribution de température suggère
32 que la structure thermique a été acquise lors de la mise en place du dôme de la zone axiale.
33 La structure thermique acquise lors de la mise en place des plis couchés et de
34 l'enfouissement des séries sédimentaires est ainsi totalement effacé par le dôme.

35 De plus, dans un domaine relativement éloigné du dôme de la zone axiale, les
36 mesures RSCM ont donné des températures significativement plus élevées que la
37 cristallinité illite. Cette divergence indique une sensibilité plus élevée du RSCM face à la
38 cristallinité illite aux événements thermiques de courte durée, probablement en raison d'une
39 cinétique plus efficace de la réaction de carbonisation. D'autre part, de fortes températures
40 RSCM analysées loin de la Zone Axiale, entre 300°C et 360°C, pourraient être expliquées
41 par la présence de plutons granitiques sous le bassin d'avant-pays.

42 **Abstract**

43 The Montagne Noire located in the southern part of the French Massif Central represents the
44 northern part of the South-Variscan Foreland. It is subdivided into three parts. The granite-
45 migmatite Axial Zone dome is surrounded by non- or weakly metamorphosed Paleozoic
46 sedimentary series. Both northern and southern flanks of the Montagne Noire dome are
47 deformed by km-scale, south to southeast facing recumbent folds and thrusts sheets. The
48 Raman Spectroscopy of Carbonaceous Material (RSCM) method, carried out in the low-
49 grade metamorphic rocks of the southern flank of the Montagne Noire, yielded temperatures
50 comprised between 400°C near the dome, and 230°C in the southern domain. Three Raman
51 geothermometers were used to cover this temperature range. RSCM temperatures comply

52 qualitatively with previous estimates based on illite crystallinity, conodont colour alteration,
53 and fluid inclusions carried out in the same area, which document a metamorphic
54 temperature increase towards the dome.

55 The isotherms cut across the different nappe contacts and are oriented parallel to the
56 southern margin of the Axial Zone. This temperature distribution supports the idea that the
57 thermal structure was acquired during the Axial Zone dome emplacement. The thermal
58 structure acquired during the recumbent folds emplacement and burial of the sedimentary
59 series is totally overprinted by the doming event.

60 In addition, in a domain relatively remote from the Axial Zone dome, the RSCM
61 measurements yielded significantly higher temperatures than illite crystallinity. This
62 discrepancy points to a higher sensitivity of RSCM to short-lived thermal events than illite
63 crystallinity, possibly because of more efficient kinetics of the carbonization reaction. On the
64 other hand, high RSCM temperatures analysed far from the Axial Zone, between 300°C and
65 360°C could be explained by the presence of granitic plutons under the foreland basin.

66 **1. Introduction**

67 The knowledge of quantitative constraints, such as pressure, temperature, duration of
68 heating, strain, strain rate, exhumation rate, and uplift rate, is essential to understand the
69 formation and evolution of a mountain belt. In recent years, many studies (e.g. Beysac et al.,
70 2002; Gerya and Stöckhert, 2002) dealt with these issues thanks to significant advances in
71 petrology and geochronology. In the inner domain of an orogen, temperature and pressure
72 conditions experienced by metamorphic rocks can be approached by petrological
73 investigations based on mineral parageneses. On the contrary, in the outer domain, where
74 the deformation and associated metamorphism are usually less developed than in the inner
75 domain, the evaluation of the thermo-barometric conditions is difficult, since the size of syn-
76 tectonic metamorphic minerals is generally small and conditions of thermodynamic equilibria
77 are often not achieved (Frey, 1987; Merriman and Frey, 1999; Merriman and Peacor, 1999).

78 The measurement of low temperature or low pressure values is therefore delicate, and often
79 bears a large error. However, several methods have been developed to overcome this
80 problem and to reach a quantitative knowledge of the conditions of low-grade
81 metamorphism, including illite crystallinity, conodont alteration colour, calcite-dolomite
82 thermo-barometer (e.g. Kübler, 1968; Dunoyer de Segonzac et al., 1968; Epstein et al.,
83 1977; Frey, 1987). More recently, the Raman Spectroscopy on Carbonaceous Material
84 (RSCM) method based on carbonaceous material crystallinity has been developed for the
85 temperature range 330-600°C (Beysac et al., 2002). Its applicability has been further
86 extended towards low temperatures (Rahl et al., 2005; Lahfid et al., 2010; Kouketsu et al.,
87 2014), using different calibrations of the Raman spectra based on other geothermometers,
88 such as vitrinite reflectance, fluid inclusions microthermometry or illite crystallinity (Rahn et
89 al., 1995, Hara et al., 2013) or indirect data on temperature provided by fission-tracks and U-
90 Th/He method (Rahl et al., 2005).

91 The applicability of the RSCM method is a priori restricted to the conditions of its
92 calibration. In all the different studies that focused on RSCM at low temperature, the case
93 studies were chosen in subduction or collision settings such as Sanbagawa Belt in Japan
94 (Kouketsu et al., 2014) and the Glarus Alps in Switzerland (Lahfid et al., 2010), providing
95 examples of metamorphic gradients. Therefore, in these cases, the thermal imprint on the
96 rocks is related to burial then exhumation within a subduction zone or a collision belt, i.e. the
97 thermal event lasts for at least one or several million years. Heating by a magmatic intrusion
98 is potentially much shorter-lived and the inter-compatibility of the different geothermometers
99 in such a case is not guaranteed (Velde and Lanson, 1993).

100 When compared to other geothermometers, the biases linked to the kinetics of the
101 thermal recording by the carbonaceous material with respect to other mineral reactions
102 should be considered (Velde and Lanson, 1993; Mullis et al., 2017). It appears that duration
103 of heating heavily impacts maturity processes and is a key parameter that controls the
104 measured temperature (Wada et al., 1994; Mählmann, 2001; Le Bayon et al., 2011; Mori et
105 al., 2017). Moreover, the geothermometers based on carbonaceous material differ from fluid

106 inclusion or metamorphic mineral ones mostly because no water is needed for the maturation
107 of carbonaceous material (Velde and Lanson, 1993; Le Bayon et al., 2011). Thus, even if the
108 whole rock experienced the same thermal history, disparities might be observed between the
109 different methods.

110 Further complexity arises when dealing with multi-metamorphic events, as in the case
111 of post-orogenic, large-scale heating and pluton emplacement postdating collisional
112 metamorphism, for instance in the Jebilet Massif in Moroccan Variscan Belt (Delchini et al.,
113 2016). A contact metamorphism linked to a magmatic intrusion may overprint the crystallinity
114 of carbonaceous material inherited from earlier stages, as it has already been documented in
115 several studies. For instance, Mori et al. (2017), measured temperatures imposed by thin
116 (50m) magmatic bodies such as the Great Whin Sill in the UK, overprinting the regional low-
117 grade metamorphism (<300°C). On a larger scale, Aoya et al. (2010) focussed on contact
118 metamorphic aureoles around granitic plutons in Japan within two poorly metamorphosed
119 accretionary complexes (<400°C). Hilchie et al. (2014) have studied the South Mountain
120 Batholith intrusion thermicity of Nova Scotia emplaced under a regional greenschist facies
121 metamorphism. In each case, RSCM temperatures decrease with the distance to the
122 intrusive bodies. Finally, shear heating between structural units could also influence the
123 organic matter organization (e.g. Mori et al., 2015). The interpretation of RSCM data in
124 domains with a poly-stage metamorphic evolution is therefore often ambiguous, as the
125 RSCM measurements combine the contribution of regional syn-tectonic metamorphism
126 coeval with crustal thickening due to nappe stacking and the later thermal overprint related to
127 a magmatic event.

128 The present study aims to decipher the thermal evolution recorded in a cold foreland
129 domain of a collisional orogen overprinted by a hot metamorphic dome (e. g. Gèze, 1949;
130 Schuiling, 1960; Demange, 1982; Echtler and Malavieille, 1990; Soula et al., 2001;
131 Alabouvette et al., 2003; Franke et al., 2011). The study area is located in the southern outer
132 zone of the Variscan belt, in the Montagne Noire of the French Massif Central (Fig.1). The
133 bulk paleotemperature field is derived from 73 RSCM measurements of graphite crystallinity

134 on weakly deformed silts and pelites from Paleozoic formations that constitute the southern
135 Variscan fold-and-thrust belt of the Massif Central. These new results are discussed in the
136 tectonic framework of this segment of the Variscan orogen.

137 **2. Geological Background**

138 The Variscan belt develops from South Spain to PolandThe French Massif Central is a stack
139 of nappes formed through a long and complex polyorogenic evolution (for details see Faure
140 et al., 2005, 2009; 2017). The SE part of the Massif exposes a succession of thrusts and
141 nappes formed from Visean to Bashkirian during the late collisional intracontinental evolution
142 (the D3 event of Faure et al., 2009). In the Montagne Noire area, the southernmost part of
143 Massif Central, weakly metamorphosed sedimentary series develop in a southeast-verging
144 fold-and-thrust belt (Fig. 1). There, kilometre-scale recumbent folds affect the entire
145 sedimentary series, including the Visean-Serpukhovian flysch. Over the whole belt, the
146 crustal thickening resulting from the successive emplacement from North to South of large-
147 scale nappes (UGU, LGU, PAU), was followed by pervasive melting affecting the tectonic
148 pile. Middle to Late Carboniferous migmatites and granites are widespread in the Massif
149 Central (Fig. 1).

150 **3. The outer Variscan domain in the Montagne Noire**

151 **3.1. General structure**

152 The Montagne Noire is subdivided into southern and northern flanks separated by the Axial
153 Zone (e.g. Gèze, 1949; Arthaud, 1970; Fig. 2). The former two zones consist of Cambrian,
154 Ordovician, Devonian and Carboniferous sedimentary rocks folded and thrust to the south
155 from the Visean to the Serpukhovian. The Axial Zone is a granite-migmatite dome emplaced
156 within the folded series. Paleozoic formations are well exposed in the Montagne Noire
157 southern flank (Fig. 3). Detailed lithological descriptions are available in Gèze (1949); Feist
158 and Galtier (1985); Álvaro et al. (1998); Vizcaïno and Álvaro (2001) and Alabouvette et al.
159 (2003). The Visean-Serpukhovian series, well exposed on the southern flank, consists of

160 syn-tectonic gravity flow deposits such as greywacke turbidites and olistostromes, which
161 recycle pre-Viséan sedimentary rocks of the advancing nappe pile (Engel et al., 1978, 1980-
162 1981; Feist and Galtier, 1985; Poty et al., 2002; Vachard and Aretz, 2004; Cózar et al., 2017;
163 Vachard et al., 2017). During the Carboniferous, the progressive deepening of the
164 sedimentary basin argues for a foreland trough, filled by syntectonic deposits, and coeval
165 with the formation of recumbent folds. The entire Paleozoic series has been deformed into
166 several km-scale, southeast-verging, thrust sheets and recumbent folds (Arthaud 1970;
167 Echtler and Malavieille, 1990).

168 3.2. Tectonic subdivisions

169 3.2.1. *The Southern Flank*

170 The upper recumbent fold, or Pardailhan unit, is the highest tectonic unit of the southern
171 flank of the Montagne Noire. It is composed of Cambrian to Devonian sedimentary rocks with
172 an inverted stratigraphic order, i.e. the geometrically upper part consists of Early Cambrian
173 sandstone, and the lower part is formed by Devonian limestone boudins that delimits the
174 basal shear zone separating the upper recumbent fold from the lower one. From north to
175 south, perpendicularly to the fold axes, the inverted limb of the Pardailhan recumbent fold
176 develops along 12 km. This inverted sedimentary series is subdivided into three km-scale
177 recumbent anticlines overturned to the south (Gèze, 1949; Arthaud, 1970; Alabouvette et al.,
178 1982; 2003).

179 Structurally below the upper recumbent fold, the lower one, also named Mt-Peyroux
180 unit, can be observed both in the eastern and western sides of the upper recumbent fold
181 (Fig. 2). The western part of the lower recumbent fold is called the Minervois unit. The
182 highest part of this recumbent fold is occupied by Ordovician flysch while Viséan turbidites
183 crop out in the lower part, indicating a stratigraphic inversion. NE-SW upright antiforms and
184 synforms deform the inverted series of the entire nappe stack. The poorly exposed contact
185 between the lower recumbent fold and the foreland basin led to controversial interpretations.
186 According to Arthaud (1970), and Alabouvette et al. (1982, 2003) the inverted limb is in

187 contact with Viséan-Serpukhovian turbidites. However, according to (Engel et al., 1978,
188 1980-1981), the flysch sequence is part of the lower recumbent (Mt. Peyroux) fold. The
189 normal and inverted limbs of that fold nappe are connected by a D1 fold hinge. Whatever the
190 structural interpretation, this does not change the RSCM results presented below, and their
191 interpretations. The presence of Paleozoic olistoliths in the Viséan-Serpukhovian turbidites
192 basin in which the recumbent folds are emplaced documents the syn-sedimentary character
193 of this event. It provides an important time constraint for the tectonic evolution of the
194 southern flank: the nappe thrusting is contemporary or older than 318Ma (Engel et al., 1978,
195 1980-1981; Alabouvette et al., 1982, 2003; Vachard et al., 2017).

196 Below the eastern part of the lower (Mt Peyroux) recumbent fold, characterized by
197 overturned Ordovician to Early Carboniferous series, the Mts de Faugères unit consists of
198 Devonian and Early Carboniferous sedimentary rocks deformed by southeast-verging
199 recumbent folds. The Para-autochthon unit is composed of Devonian to Middle
200 Carboniferous sedimentary rocks exposed in the normal stratigraphic order. This domain is
201 observed between the lower recumbent fold and the Axial Zone metamorphic rocks. Close to
202 the Axial Zone, for instance near Saint-Pons (Fig. 2), the Para-autochthonous unit consists of
203 Devonian marbles and micaschists (Engel et al., 1980-1981; Feist and Galtier, 1985; Poty et
204 al., 2002; Vachard and Aretz, 2004; Cózar et al., 2017; Vachard et al., 2017).

205 The autochthonous turbiditic basin represents the foreland basin into which the Mt-
206 Peyroux and Mts-de-Faugères recumbent folds were emplaced. The basin substratum is
207 unknown but might probably be similar to that observed in the northern para-autochthonous
208 unit. To the west, the basin underlies the stack of recumbent folds, and to the south, it is
209 hidden below the Cenozoic formations.

210 *3.2.2. The Axial Zone*

211 The Axial Zone is composed of orthogneiss, paragneiss, amphibolite, micaschist, and rare
212 marble that experienced a partial melting giving rise to migmatites and granites. The Axial
213 Zone presents a dome architecture of about 90 km long, and 20 km wide, with a N70E long

214 axis (Gèze, 1949; Faure and Cottureau, 1988; Echtler and Malavieille, 1990; Matte et al.,
215 1998; Van den Driessche and Brun, 1992; Demange, 1993; Alabouvette et al., 2003). The
216 Eocene Pyrenean Mazamet fault divides the western part of the dome in two parts, the Agout
217 and Nore massifs in the North and South, respectively. The orthogneiss yields zircon U-Pb
218 Ordovician to Devonian ages at 472 ± 2.8 Ma, 456 ± 3 Ma, 450 ± 6 Ma, 455 ± 2 Ma, and 416
219 ± 5 Ma interpreted as those of the granite protolith (Roger et al., 2004; Cocherie et al., 2005;
220 Franke et al., 2011; Pitra et al., 2012; Trap et al., 2017). The country rocks protoliths of these
221 orthogneiss are interpreted as Neoproterozoic to Cambrian or Ordovician (Alabouvette et al.,
222 2003). Several geochronological ages on the migmatite and anatectic granites range
223 between 333 Ma and 294 Ma (Hamet and Allègre, 1976; Faure et al., 2010; Franke et al.,
224 2011; Poilvet et al., 2011; Roger et al., 2015; Trap et al., 2017) with pegmatite until 282 Ma
225 (Doublie et al., 2015). The abundance of granitoids and migmatite argue for widespread
226 thermal event. The Axial Zone experienced two metamorphic and tectonic events. The early
227 one is recorded only in rare mafic eclogitic restites enclosed in the migmatite. Though zircon
228 in eclogite yields a ca. 315 Ma age (Faure et al., 2014; Whitney et al., 2015), this can hardly
229 be interpreted as the age of the high pressure event since the enclosing migmatite yields
230 ages older than the eclogite blocks. A hydrothermal event might be responsible for zircon
231 recrystallization. The second, and main one is coeval with the dome formation (Schuiling,
232 1960; Bard and Rabeloson, 1973; Thompson and Bard, 1982; Soula et al., 2001; Faure et
233 al., 2014; Fréville et al., 2016; Trap et al., 2017).

234 The metasedimentary rocks that form the outer envelope of the granite-migmatite
235 dome underwent a pervasive high temperature/medium pressure (HT/MP) metamorphism
236 coeval with the dome formation (Soula et al., 2001; Faure et al., 2014; Fréville et al., 2016).
237 From South to North, chlorite, biotite, garnet, andalusite, staurolite, and sillimanite isograds
238 appear successively. As already pointed out, chlorite, biotite, and muscovite belonging to the
239 low temperature part of the HT/MP metamorphism develops in the various lithotectonic units
240 close to the Axial Zone, such as the para-autochthonous, Mts-de-Faugères, Mt-Peyroux and

241 Pardailhan units (Arthaud, 1970; Franke et al., 2011; Fréville et al., 2016). The Axial Zone
242 HT/MP metamorphism started after the emplacement of the thrusts and recumbent folds but
243 continued still afterwards into the Permian (Alabouvette et al., 2003; Franke et al., 2011; Pitra
244 et al., 2012).

245 3.2.3. *The Northern Flank*

246 The Montagne Noire northern flank is composed of Late Neoproterozoic (Ediacaran) to
247 Silurian sedimentary formations subdivided into several south-directed folds and thrusts
248 considered as equivalent to those of the southern flank. The deformation age is estimated of
249 Visean age (342-333 Ma; Gèze, 1949; Alabouvette et al., 2003; Doublier et al., 2006). The
250 southern part of the northern flank, east and SW of the Lacaune fault (Fig. 2), exposes
251 biotite, garnet, andalusite micaschists comparable to the metapelites of the southern
252 envelope of the Axial Zone (Demange, 1982). This metamorphism that superimposes to the
253 fold-and thrust structure was coeval with the HT/MP event related to the Axial Zone doming
254 (Doublier et al., 2006). The contact between the northern flank and Axial Zone is the Mts-de-
255 Lacaune NE-dipping ductile normal fault that accommodated the dome emplacement (Van
256 den Driessche and Brun, 1989). Furthermore, a brittle normal fault controlled the opening of
257 the Late Carboniferous (Gzhelian) Graissessac coal basin. In the following, the northern flank
258 will not be considered.

259 3.3. The tectonic outline of the southern flank

260 The structural analysis of the Montagne Noire southern flank revealed two main tectonic
261 events (Arthaud, 1970). The first deformation stage corresponds to the emplacement of the
262 south verging recumbent folds (F1) and thrusts. The bedding and cleavage relationships
263 demonstrate the south to southeastward vergence of the F1 folds (Arthaud, 1970; Fig. 4a). A
264 recent study proposes a nappe thrusting towards the northeast (Chardon et al., 2020).
265 Whatever the tectonic interpretation, in spite of 15 km of subhorizontal displacement, and the
266 ca 1.5 km overload due to the stratigraphic inversion, the Paleozoic series poorly exhibit a

267 ductile syn-metamorphic deformation. A slaty cleavage is only visible in the hinges of F1
268 folds (Arthaud, 1970). The metamorphic grade reached by the upside down Paleozoic series
269 in the Upper (Pardailhan) and Lower (Mt-Peyroux) recumbent folds is below the biotite grade,
270 which firstly appear near the dome, in the para-autochthonous unit.

271 The second deformation event is an upright folding phase (F2). Field observations
272 document the refolding of the F1 recumbent folds by the F2 upright ones with NE-SW striking
273 axes (Arthaud, 1970; Alabouvette et al., 2003). Close to the Axial Zone, a subvertical axial
274 planar cleavage develops in the F2 folds, whereas away from the Axial Zone dome, the F2
275 folds are rather open and devoid of cleavage (Fig. 4b). The F2 upright folding corresponds to
276 a NW-SE shortening that formed large- scale antiforms and synforms including the Axial
277 Zone dome. A N70 mineral and crenulation lineation appears in the Carboniferous formation
278 at about 5 km south of the dome and northward (Harris et al., 1983). Sericite is also
279 recognized north of this point (close to sample H11 - Fig. 2). However, Franke et al. (2011)
280 interpreted this lineation as an extension direction linked to the pull apart emplacement of the
281 Axial Zone. Anyhow, these metamorphic and structural features argue for a syntectonic
282 metamorphism linked to the doming (Soula et al., 2001; Franke et al., 2011).

283 **4. The RSCM method**

284 The Raman spectroscopy on carbonaceous material (RSCM) method was carried out to
285 estimate the paleotemperature field of the low grade, weakly deformed, sedimentary rocks of
286 the southern flank of the Montagne Noire. The method is based on the irreversible
287 transformation of carbonaceous matter towards graphite structure, using empirical
288 correlations to correlate Raman spectra of carbonaceous matter into maximum temperature
289 of heating, in the range 150-650°C (Beysac et al., 2002, Lahfid et al., 2010).

290 **4.1. Analytical settings**

291 The Raman measurements were carried out on a Renishaw inVia reflex system belonging to
292 the ISTO-BRGM analytic platform. The Wire 3.4 software was used for the data acquisition.

293 The argon-ion laser source excitation of 514.5nm was set at a power of about 5% of its
294 capacity (2.5mW). The monochromatic ray was coupled to a reflection microscope with an
295 x100 objective. Before each series of measurement, the spectrometer was calibrated using
296 an internal silica standard for the wavenumber ($520,4 \text{ cm}^{-1}$) and the signal intensity (at least
297 30 000 counts per second). Fifteen carbonaceous matter spectra were systematically
298 acquired for each sample within the thin section in order to obtain a representative and
299 reliable temperature. Only organic particles located below transparent quartz, a few μm
300 under the thin section surface, were analysed, to avoid any mechanical damaging of their
301 crystalline structure due to thin section preparation and polishing. The acquisition duration
302 was set to at least 60s and adapted depending on the quality of the spectrum.

303 4.2. RSCM data processing

304 Several empirical calibrations were proposed to correlate the carbonaceous matter Raman
305 spectra to temperature (Beysac et al., 2002; Rahl et al., 2005; Aoya et al., 2010; Lahfid et
306 al., 2010; Kouketsu et al., 2014). Carbonaceous matter presents characteristic Raman bands
307 in the wavenumber range between $1100\text{-}1800 \text{ cm}^{-1}$ and $2500\text{-}3100 \text{ cm}^{-1}$. The first order
308 region of the carbonaceous matter spectrum for $1100\text{-}1800 \text{ cm}^{-1}$ provided the data of this
309 study.

310 Depending on its crystallinity (hence on the temperature experienced), carbonaceous
311 matter presents a variable number of Raman peaks, which can be used to discriminate
312 between the different calibration methods. Since the sedimentary rocks of the Montagne
313 Noire southern flank are weakly to un-metamorphosed, most of the Raman spectra were
314 processed with the Lahfid et al. (2010) method. Accordingly, each spectrum was
315 decomposed into five bands (Lorentzian functions), with a well-constrained position: G (1590
316 cm^{-1}) D1 (1350 cm^{-1}), D2 (1620 cm^{-1}), D3 (1515 cm^{-1}) and D4 (1250 cm^{-1}) (Lahfid et al., 2010;
317 Sadezky et al., 2005). From this decomposition, the RSCM temperature was computed using
318 the formula proposed by Lahfid et al. (2010) which is valid between 200°C and 320°C , with
319 an accuracy of $\pm 30^\circ\text{C}$ (Fig. 5a):

$$320 \quad T^{\circ}C = \frac{RA1 - 0.3758}{0.0008} \quad \text{with} \quad RA1 = \frac{D1 + D4}{D1 + D2 + D3 + D4 + G}$$

321 Samples with higher crystallinity have fewer Raman peaks and were processed
 322 accordingly using the method proposed by Beyssac et al. (2002) that allows an estimation of
 323 temperature in the range between 330 up to 650°C, with a $\pm 50^{\circ}C$ uncertainties (Fig. 5b)

$$324 \quad T^{\circ}C = (-445 * R2 + 641) \quad \text{with} \quad R2 = \frac{D1}{D1 + D2 + G}$$

325 Considering the low metamorphic grade of the analysed samples, only the low range
 326 of this method was used ($<400^{\circ}C$). Since the Beyssac et al. (2002) and Lahfid et al. (2010)
 327 methods do not exactly overlap with each another; we also used a third thermometer
 328 (Kouketsu et al., 2014). This method does not correspond to peaks area ratios but focuses
 329 on the peak full width at half maximal (FWHM) of the D1 and D2 band. Because of the
 330 difficulty to distinguish the D2 band from the G band, only the formula about the former band
 331 was retained. According to this method, the temperature is derived from Raman peaks as:

$$332 \quad T^{\circ}C = -2.15 * (FWHM - D1) + 478$$

333 The Kouketsu et al. (2014) method leads to temperatures comprised between 150 and
 334 $400^{\circ}C$ with a $\pm 30^{\circ}C$ uncertainty.

335 4.3. Uncertainty in RSCM temperatures

336 The different RSCM methods are calibrated against several geothermometers. The higher
 337 range temperatures method proposed by Beyssac et al. (2002) is mostly based on mineral
 338 assemblages. However, vitrinite reflectance, illite crystallinity, chlorite geothermometers were
 339 compared with RSCM spectra both in Kouketsu et al. (2014) and Lahfid et al. (2010) studies.
 340 Furthermore, thermal modelling, garnet-chlorite geothermometers were also used for
 341 calibration in Kouketsu et al. (2014) while Lahfid et al. (2010) used calcite-dolomite
 342 thermometry, quartz-chlorite isotopic thermometry and fluid inclusion. These calibrations
 343 were carried out on domains that were affected by a single metamorphic event.

344 The maximum error (1σ) on R2 ratio obtained by Beyssac et al. (2002) is ± 0.08
 345 corresponding to a $\pm 36^{\circ}C$ intra-sample variability. Adding the uncertainty linked to the

346 calibration of RSCM with reference temperatures, the authors considered that the uncertainty
347 on the estimation of temperature was about $\pm 50^{\circ}\text{C}$.

348 In the low temperature range, Lahfid et al. (2010) method uses the ratio of Raman
349 peak spectra area, while Kouketsu et al. (2014) method uses the width of the peaks, to
350 derive the paleotemperature. Kouketsu et al. (2014) estimated that the difference in
351 estimated temperature due to the use of the two different methods is of the order of 50°C .

352 4.4. RSCM acquisition

353 The temperatures derived from the RSCM method were obtained with a minimum of 10 to 15
354 spectra for each sample in order to ensure their validity. Some of the grains of carbonaceous
355 matter present within a sediment might originate from the erosion of a sediment that has
356 already been affected by a prior metamorphic event. Since the organic matter only records
357 the highest temperature, these inherited grains might yield higher conditions of temperature
358 than their host sediments, if they come from higher-grade material. This case is often
359 suspected when a few CM grains give much higher temperatures than most grains in a
360 sample. In such a case, we disregarded the few anomalous grains and estimated the
361 temperature from the statistical distribution concentrating most of the grains (Beysac et al.,
362 2002).

363 The 122 samples analysed cover the major part of the Montagne Noire southern flank
364 in order to get a general picture of the paleotemperature field reached by the Paleozoic
365 series. Almost all lithologies were tested for the RSCM analysis. Several samples were
366 polished at $35\mu\text{m}$ for optical microscope observation (Fig. 6) but most of the measurements
367 were carried out on $200\text{-}300\mu\text{m}$ -thick thin sections. The thin sections were prepared
368 regardless of the mineral preferred orientation of the carbonaceous matter because of the
369 poorly developed carbonization in these low metamorphic grade series (Aoya et al., 2010).

370 The RSCM method reliability depends on the richness of the carbonaceous matter
371 enclosed in the sample. In the field, the dark lithologies, i.e. the Early Ordovician and Viséan-
372 Serpukhovian turbidites, turned out to be the most appropriate lithologies (Fig. 6a, b, d, f). On

373 the contrary, the Early Cambrian Marcory green sandstone, and the Early Cambrian and
374 Devonian carbonates did not yield sufficient organic matter (Fig. 6c, e), and thus, were
375 avoided for sampling.

376 4.5. Sampling Strategy

377 The main goal of the present study was to establish the paleotemperature field associated
378 with the regional tectonic-metamorphic events. Several hypotheses may be formulated.,

379 The temperature development was coeval with the recumbent folding and thrusting
380 event for this purpose, samples were collected in each structural unit except for the South
381 Minervois and Mts-de-Faugères units. Higher temperatures can be expected at deeper
382 tectonic levels. Furthermore, areas distant from the Axial Zone dome where the metamorphic
383 and structural imprint of the Axial Zone dome upon the recumbent folds seems absent
384 (Doublier et al., 2006) represent attractive sites. For instance, in the Ordovician sandstone-
385 siltstone formation of the Upper (Pardailhan) recumbent fold, only the S1 slaty cleavage is
386 observed (Fig. 4a).

387 The thermal effect was related to the late-stage emplacement of the migmatites and
388 plutons in the Axial Zone dome. The strategy to assess the dome effect, as suggested by
389 Wiederer et al. (2002) and Franke et al. (2011) studies, was totally different. In the southern
390 flank, upright folds coeval with a steep S2 cleavage (Fig. 4b) structurally represents this
391 event. The occurrence of a N70E striking mineral lineation marked by white mica (sericite)
392 becomes more and more difficult to recognize south of the dome. Consequently, high
393 temperatures should be recorded near the dome while they must decrease away from it,
394 southwards. Nord-south sections were sampled in order to test this hypothesis.

395 Finally, a last possibility can be considered. The axial dome exhumation itself results
396 of a regional horizontal flow that affected the entire crust. Thus, a hot crust could be located
397 below the nappes piles. Beside the dome thermal impact, a high temperature event
398 overprinting the nappe stacking can be expected far from the Axial Zone.

399

400

401 **5. Results**

402 Despite the low-grade metamorphism of the area, both high and low temperature methods
403 were used during this study. The results are well distributed in the 230 to 400°C range
404 covered by both methods (Fig. 5 and 7). Seventy-three temperature values were calculated
405 in the Montagne Noire southern flank. Forty-eight RSCM spectra were processed using
406 Lahfid et al. (2010) method, while 25 spectra were computed through the Beyssac et al.
407 (2002) method (Table. 1). Thirty-six temperatures values were also reassessed by the
408 FWHD-D1-linked thermometer (Kouketsu et al., 2014; Table. 2). The results, plotted in Fig.8,
409 reveal a decreasing of maximum temperatures towards the south. It appears that the
410 metamorphic zonation cuts across the boundaries between the tectonic units and the limbs
411 of recumbent folds close the dome and a more homogenous in further distance. In contrast
412 to one of our working hypotheses, the RSCM temperatures do not indicate any thermal gap
413 on both sides of the tectonic contacts.

414 **6. Discussion**

415 6.1. Temperature field acquired by RSCM method

416 In this study, the entire Montagne Noire southern flank has been investigated by the RSCM
417 method. Most of the data set obtained from the sedimentary rocks records a gradient
418 consistent over the whole investigated area, visible in both normal and inverted sedimentary
419 series (Fig. 8, 9 cross sections 1, 2, 3, 4, 5). Globally, the temperature decreases from the
420 southern edge of the Axial Zone dome towards the southeast in the Paleozoic sedimentary
421 rocks (Fig. 10). The maximum temperature is located in the micaschists series that form the
422 dome envelope with a value close to 400°C. The lowest temperature, about 220°C, was
423 obtained in the Viséan turbidites. In the western part of the southern flank, the gradient is

424 less apparent that in the eastern part. It decreases from 400°C to 315°C towards the
425 Cenozoic sedimentary rocks.

426 The RSCM method was also used within the dome micaschist envelope (Fréville et
427 al., 2016). The measured temperatures, ranging between 450°C and 600°C are higher than
428 those obtained in the sedimentary rocks of the southern flank (Fig. 11). The temperature
429 gradient is steeper in the dome northern edge than in the micaschist series of the eastern
430 area. Indeed, the deformation is more intense in the dome northern edge, as also shown by
431 the tight arrangement of the metamorphic isograds (Alabouvette et al., 1993; Fréville et al.,
432 2016). This is probably a consequence of a brittle shearing that poste-dates the dome
433 exhumation (Thompson and Bard, 1982). The data set obtained via the RSCM method
434 documents a continuous evolution from the dome core southward with a temperature
435 decrease from 580°C to less than 300°C (Fig. 11). Since the dome foliation is plunging under
436 the sedimentary series, samples horizontally away from the dome are also vertically distant
437 from the dome.

438 In case of a thermal field predating the emplacement of the recumbent folds, a
439 temperature discontinuity should coincide with the contacts between different units. In case
440 of a thermal event coeval with nappe thrusting and tectonic thickening, the gradient should
441 be consistent with the structural position in the tectonic pile. Indeed, it has been shown that
442 temperatures recorded in terrigenous sedimentary series deposited in foreland basins
443 involved in fold-and-thrust belts of orogens are related to tectonic thickening. For instance, in
444 the northern Variscan foreland basin of the Ardennes massif, the temperature of 200 to
445 300°C, measured by illite crystallinity, vitrinite reflectance, conodont alteration index, and
446 fluid inclusion methods, has been linked to the burial of sedimentary series below the
447 Variscan thrust front (Fielitz and Mansy, 1999).

448 The RSCM temperature map of the Montagne Noire southern flank (Fig. 10) is clearly
449 incompatible with these assumptions. Since the isotherms cut across the different fold units
450 irrespective of their contacts, the bulk thermal structure of the Montagne Noire southern flank
451 cannot be linked to the pre-doming thickening stage. Moreover, the RSCM temperature is not

452 correlated with the structural position of the analyzed samples since the highest nappes of
453 the tectonic pile record-temperatures similar to those measured in the para-autochthonous
454 units (Fig. 10). Thus, the link of the temperature field with nappe emplacement, and tectonic
455 burial is unlikely. The measured temperature is also not in agreement with shear heating
456 along major contacts, as that observed in Japan (Mori et al., 2015).

457 In contrast, there is a clear temperature gradient decreasing with the distance to the
458 dome. Whatever the thrusting direction of the nappe, southeastward (Arthaud, 1970) or
459 northeastward (Chardon et al., 2020), the interpretation of a thermal effect related to the
460 Axial Zone dome is therefore preferred to explain the temperature field as suggested by
461 previous authors (Wiederer et al., 2002; Doublier et al., 2005; Franke et al., 2011). At first
462 order, the RSCM method records the HT metamorphic event already recognized close to the
463 Axial Zone over a very broad area including its entire southern flank. It has been shown that
464 RSCM investigations around an intrusion provide high temperatures away from the
465 metamorphic aureole (e. g. Hilchie et al., 2014; Beyssac et al., 2019).

466 However, it is quite unlikely that the heat propagation away from the dome may
467 develop temperatures above 300°C as those observed south of St-Pons at ca 10km in the
468 southernmost part of the upper recumbent fold. We suggest that the temperature field is
469 linked with an hypothetical hot crust below the recumbent folds.

470 6.2. Anomalies in the temperature field

471 Two temperature anomalies are visible within the N-S temperature gradient (Fig. 8). To the
472 east of the study area, within the Visean-Serpukhovian turbiditic basin, several samples
473 present relatively high temperatures around 360°C, similar to those found near the dome
474 micaschists (Figs. 6, 9 cross section 5). However, in this area, the Axial Zone dome is not
475 exposed. The other temperature anomaly is located in the southern part of the Pardailhan
476 upper recumbent fold (Fig. 9 cross section 3). The gradient is decreasing southward from ca
477 380°C to 270°C, but at the very end of the exposed Paleozoic series, the temperature

478 increases to reach ca 320°C. Since this area is located about 15 km away from the Axial
479 Zone dome south margin, it is unlikely that these high temperatures result from the direct
480 dome thermal influence.

481 Three hypotheses might explain these singularities.

482 First, the effect of a late fault might be taken in account as it could reorganize the
483 distribution of paleotemperatures. It is especially noticeable for the anomaly within the upper
484 recumbent fold (Pardailhan unit). The N-S striking sinistral fault at the east of the points H25
485 an H26 Fig. 2) separates high temperature in the west >315°C (H27-H28-H29) from lower
486 temperatures in the east <290°C (H36-H39-H40; Fig. 8; Fig. 9). This effect is also shown by
487 the bending of the 300°C isotherm (Fig. 10).

488 Second, the circulation of fluids, related to the nappes thickening or the doming may
489 be considered. Several quartz veins, developed during these circulations, appear within the
490 basin and close to the tectonic contacts (Guiraud et al., 1981) or are linked with the
491 micaschist /sedimentary rocks contact as observed in Salsigne (Lescuyer et al., 1993).

492 Finally, at the regional scale, the district is riddled by Late Carboniferous granitic
493 intrusions, such as the Sidobre or Folat plutons (Fig. 2), thus the presence of a hidden pluton
494 below the sedimentary rocks cannot be discarded. Granitic plutons emplaced in an orogen
495 external zone and in gneissic domes have been already documented in the Nappe and
496 External zones of Sardinia (Carmignani et al., 1994; Carosi et al., 1998) and more recently
497 by RSCM investigation in Morocco (Delchini et al., 2016). Thus, the existence of several
498 hidden gneiss-granite-migmatite massifs underlying the southern flank stack of recumbent
499 folds and the turbiditic basin must be considered. Moreover, a pegmatite dyke, 2 km north of
500 the anomaly in the basin, dated at 282 Ma (Doublie et al., 2015), could be the witness of one
501 of these intrusions. This hypothesis is closely related with the presence of a hot crust which
502 can explain the isotherm pattern in the easternmost part of the southern flank, more than
503 15km away from the dome (Fig. 10).

504 6.3. Comparison of the RSCM results with other thermometers

505 *6.3.1. General pattern of the temperature field*

506 The illite crystallinity has been widely studied in the eastern part of the Montagne Noire
507 southern flank to unravel low-grade metamorphic gradients. In our study area, two previous
508 works (Engel et al., 1980-1981; Doublier et al., 2015) documented a north to south
509 decreasing thermal gradient (Fig. 12). The latter study used the Árkai index (Árkai, 1991;
510 Guggenheim et al., 2002) and the Kübler index (Kübler, 1964) to estimate a relative indicator
511 of temperature. Furthermore, Wiederer et al. (2002) compared the results obtained via the
512 Weber index (Weber 1972) in Engel et al. (1980-1981) to the diagenetic domains defined in
513 Frey and Robinson, (1999). In agreement with our own results, these works show the same
514 N-S variations as those documented by the RSCM method.

515 In the same area, the conodont color alteration has been used to estimate the
516 paleotemperature repartition (Wiederer et al., 2002). The method is based on the color of
517 apatite crystals that form the conodonts (Epstein et al., 1977). During a metamorphic event,
518 the crystallinity of organic matter trapped within the apatite grains increases, allowing a
519 possible correlation with the metamorphic grade experienced by the fossil. As shown in Fig.
520 13, the dome thermal effect is also visible by this method. The intensity of the metamorphism
521 is globally higher to the northwest and decreases away to the southeast (Fig. 13).

522 A fluid inclusion study of quartz veins (see location on Fig. 11) has been carried out
523 (Guiraud et al., 1981). The analyzed rocks lie along the basal tectonic contact of the
524 Pardailhan upper recumbent fold, called “queue de cochon” (pig tail) contact (Gèze, 1949)
525 composed of Devonian limestone boudins surrounded by Ordovician sandstone and pelite.
526 These 1- to 10-cm sized quartz lenses were inferred to be related to fluid circulation coeval
527 with the emplacement of the upper recumbent fold (Guiraud et al., 1981). Analyses show a
528 temperature around $275\pm 25^{\circ}\text{C}$, which is in agreement with our results at ca 300°C , but
529 higher than those provided by illite crystallinity at 200°C (Doublier et al., 2015). This 300°C
530 temperature has been related to the thickening event (Guiraud et al., 1981), however, fluid
531 circulation might have occurred also during the doming. The tectonic contact between the

532 upper and lower recumbent folds has been reworked during doming, as indicated by E-W
533 striking slickenlines on flat lying surfaces, and N70E striking fibers infilling tension gashes
534 (Arthaud, 1970; Sauniac, 1980; Harris et al., 1983). Therefore, the deformation observed
535 along the contact suggests that the ca 300°C temperature might be related to the Axial Zone
536 dome thermal effect or even to a late event as documented by Aerden (1998) and Franke et
537 al. (2011).

538 *6.3.2. Discrepancy in the low T range between RSCM temperature and other thermometers*

539 For the sake of comparison between the different geothermometers, both conodont
540 color alteration index and illite crystallinity data were converted in temperatures using to the
541 tables provided by Merriman and Frey (1999) and Wiederer et al. (2002) for low grade
542 metamorphism and the equation (1) from Zhu et al. (2016) work. Several studies have shown
543 a good correlation between illite crystallinity (IC, estimated from Kübler Index, KI) and
544 organic geothermometers such as vitrinite reflectance (VR) (Underwood et al., 1993;
545 Mukoyoshi et al., 2009; Fukuchi et al., 2014). This is true in basinal settings (Baludikay et al.,
546 2018) but also in collision/subduction settings (Rahn et al., 1995). In the Montagne Noire,
547 and despite qualitatively convergent trends indicating a north-south gradient from the Axial
548 Zone dome to the sedimentary southern flank, the measured temperatures vary significantly,
549 depending on the method considered.

550 As example of these differences, the temperature range estimated using the
551 conodont approach is the widest, with temperatures comprised between 75°C and 475°C
552 against 85°C to 300°C for IC and 230°C to 400°C for RSCM. Furthermore, for a given
553 location, the results are sometimes different between the two illite crystallinity studies by
554 Engel et al. (1980-1981) and Doublier et al. (2015). This is remarkable at the contact
555 between the lower recumbent fold and the Visean basin where the data from Engel et al.
556 (1980-1981) are 50°C higher than those from Doublier et al. (2015). It can be explained by
557 the use of a standardized and more precise procedure in the study of Doublier et al. (2015)
558 while Engel et al. (1980-1981) employed an early version of the method known to have

559 interlaboratory standardization issues (Merriman and Peacor, 1999). Comparing with RSCM
560 result, other geothermometers provide temperature lower by at least 70°C. Baludikay et al.
561 (2018) reported a general overestimation of RSCM temperatures (using Kouketsu et al.,
562 2014 calibration) with respect to other geothermometer in low-grade sediments ($T < 200^\circ\text{C}$)
563 from intra-cratonic basins. The discrepancy is here much higher, as illite crystallinity data
564 from Doublier et al. (2015) and RSCM data from this study show a difference up to 175°C
565 (Fig. 14). A first possible source of these differences lies in the large uncertainty in the
566 conversion of any of the metamorphic/organic indicators into temperatures. A second source
567 of error is related to the complex geological history of the domain treated here, involving
568 several tectonic and heating stages, which might have affected to a variable extent the
569 organic and mineral signals (García-López et al., 2001).

570 Nonetheless, irrespective of the comparison between temperatures themselves, there
571 are also large differences in the temperature gradients derived from the different
572 geothermometers. Between 1 to 6 km away from the dome, the illite crystallinity seems to
573 reach a plateau around 275°C (dashed line in Fig. 14). This plateau could correspond to the
574 temperature above which small muscovite or sericite have completely replaced the original
575 clays and would be the upper limit of the method, while carbonaceous matter crystallinity
576 (hence its Raman spectra) continue its evolution at higher temperature. Hence, for
577 comparing the RSCM temperature with illite crystallinity temperature, we shall exclude the
578 highest temperature domain in the vicinity of the dome and consider the section at a distance
579 from 5 km to 15 km from the dome. In this distance range, the gap between RSCM
580 temperatures and illite crystallinity temperatures increases with the distance to the dome
581 (Fig. 14). This divergence is apparently irrespective of the RSCM calibration used (Fig. 15a,
582 b) or the calibration to convert KI in temperature (T):

583 -(i) With Merriman and Frey (1999) table: the difference between KI-derived and
584 RSCM-derived temperatures varies from ~75°C to ~150°C (Lahfid et al., 2010) and from
585 ~80°C to ~120°C (Kouketsu et al., 2014) at 5 km and 14 km from the dome, respectively.

586 -(ii) With Zhu et al. (2016) equation: the T difference varies from ~75°C to ~175°C
587 (Lahfid et al., 2010) and from ~80°C to ~150°C (Kouketsu et al., 2014) in the same section.
588 Therefore, these disparities suggest that organic matter record the dome exhumation thermal
589 impact on a much broader area than illite crystallinity.

590 To interpret these discrepancies, one has to consider the nature of the physical and
591 chemical processes involved in carbonaceous matter and illite maturation. Velde and Lanson
592 (1993), Belmar et al. (2002) and Mählmann et al. (2012) suggested that the carbonaceous
593 matter records short thermal episodes, such as magmatic intrusions, more easily than clays.
594 This contrasted record could be linked to the fact that organic matter maturation beyond its
595 low-T cracking is principally isochemical, while illite and other clays evolution requires
596 chemical reactions involving elemental exchanges (Velde and Lanson, 1993). The samples
597 used to calibrate RSCM with illite crystallinity (Kouketsu et al., 2014) were collected in an
598 accretionary complex in SW Japan (Hara et al., 2013). In this regional metamorphism
599 context, both illite and carbonaceous material had sufficient time to mature and tend
600 towards some “equilibrium” state. With a duration longer than 10Ma, several
601 geothermometers based on various mineralogical assemblages yield the same temperature
602 (Le Bayon et al., 2011; Mullis et al., 2017).

603 Hence, the disparities between illite crystallinity and RSCM in Montagne Noire
604 suggest that the dome emplacement, which controls IC and RSCM evolution, was a relatively
605 short event in terms of heat source. Accordingly, it left a more significant imprint on
606 carbonaceous matter than on illite.

607 One can note that the much smaller-scale T anomaly in the basin, recorded by RSCM but
608 not by illite crystallinity, can be similarly explained by hidden magmatic bodies that provided
609 a local heat source for a short period of time. We therefore propose that a regional high heat-
610 flow linked to a hot crust is present over a long period 330-295Ma, as recorded by Illite and
611 CAI. However, local events resulting from this regional heat such as plutonic intrusion and on
612 a larger scale the migmatization mostly affected the carbonaceous material.

613 Moreover, the high RSCM, illite and conodont temperatures observed in the basin far from the
614 eastern terminaison of the dome suggest that a sinistral movement occurred during the dome
615 emplacement. It is then possible that the tectonic origin of the dome emplacement took place
616 during a sinistral transpression as suggested by Chardon et al. (2020). This view is also in
617 agreement with a N-S shortening responsible for the double anticline present within the Axial
618 Zone dome (Matte et al., 1988; Malavieille, 2010).

619 **7. Conclusion**

620 The 75 RSCM measurements acquired in this study combined with the results of several
621 previous studies, derived from different methods, allow us to reconstruct the thermal history
622 of the Montagne Noire southern flank. The RSCM measurements, processed either by Lahfid
623 et al. (2010) or Beyssac et al. (2002) methods, cover the whole metamorphic range present
624 in the sedimentary rocks. In agreement with the RSCM results, the conodont color alteration,
625 and the illite crystallinity geothermometers revealed a temperature gradient within the
626 Paleozoic formations decreasing from the Axial Zone dome toward the SE. This RSCM
627 temperature gradient represents the effect of the Axial Zone dome overprint on the
628 sedimentary rocks as suggested by previous studies on a smaller area (Engel et al., 1980-
629 1981; Wiederer et al., 2002; Doublier et al., 2015) and also observed in the Montagne Noire
630 northern side (Doublier et al., 2006).

631 Moreover, an underneath heating by the hot crust might also explain the relatively
632 high temperatures, $>300^{\circ}\text{C}$, recorded by RSCM, IC and CAI in the very southern part of the
633 upper recumbent fold and in the east of the southern flank.

634 In spite of a common North to South decreasing temperature gradient, the different
635 thermometers present quantitative discrepancies in the estimated temperature. IC
636 temperatures decrease away from the dome to temperatures below 150°C , while RSCM
637 temperatures remain above 250°C . In this case, the fast kinetics of carbonaceous material
638 crystalline evolution, with respect to illite evolution, could then account for the record of the
639 heating event over a larger area by RSCM than by IC.

640 The link between the bulk temperatures and the recumbent folding is not supported
641 by this study. Furthermore, anomalies are observed within the thermal gradient. Since these
642 anomalies are not documented by the IC method, the most likely explanation would be that
643 hidden bodies such as granitic plutons, would have emplaced beneath the Paleozoic
644 sediments after the dome exhumation. In addition, the presence of a hot crust underlying the
645 Paleozoic series cannot be discarded.

646 Finally, the temperatures obtained in the southern part of the Upper recumbent fold,
647 alike those measured by fluid inclusion study (Guiraud et al., 1981), are compatible with the
648 Axial Zone doming. The RSCM approach suggests that the thermal effects related to the
649 thickening event has been totally overprinted by the dome thermal effect.

650 As already been investigated in the Rheno-Hercynian zone in the northern Variscan
651 branch (e.g. Fielitz and Mansy, 1999; Doublier et al., 2012), the study of thermicity in
652 foreland basins can be extended to other areas within the Variscan belt. For instance,
653 Mouthoumet massif (Kretschmer et al., 2015), Balearic or southern Sardinia-areas are
654 suitable places to get a general picture of the Variscan thermicity.

655 **Acknowledgments**

656 This study has been funded by the Labex VOLTAIRE, and the Observatoire des Sciences de
657 l'Univers en Région Centre (OSUC). Jean-Gabriel Badin, Sylvain Janiec, and Ida di Carlo are
658 thanked for their assistance during the thin section preparation and Raman spectroscopy
659 measurements. We also thank Stanislas Sizaret for his help during sampling session.

660 Constructive comments by J. Malavieille, R. Carosi, and an anonymous reviewer are deeply
661 acknowledged.

662

663 **References**

664 Aerden DG. 1998. Tectonic evolution of the Montagne Noire and a possible orogenic model
665 for syncollisional exhumation of deep rocks, Variscan belt, France. *Tectonics*, 17(1),
666 62-79.

667

668 Alabouvette B, Arthaud F, Bambier A, Freytet P, Paloc H. 1982. Notice explicative de la carte
669 géologique au 1/ 50 000 de Saint-Chinian, n°1014. Bureau de Recherches
670 Géologiques et Minières, Orléans, France. 1-44.

671

672 Alabouvette B, Demange M, Sauvel C, Vautrelle C. 1993. Notice explicative de la feuille Saint-
673 Pons à 1/50 000. Édition du Bureau de Recherches Géologiques et Minières, Orléans.

674

675 Alabouvette B, Demange M, Guérangé-Lozes J, Ambert P. 2003. Notice explicative de la carte
676 géologique au 1:250000 de Montpellier. Bureau de Recherches Géologiques et
677 Minières, Orléans, France.

678

679 Álvaro JJ, Courjault-Radé P, Chauvel JJ, Dabard M P, Debrenne F, Feist R, Pillola GL, Vennin
680 E, Vizcaïno D. 1998. Nouveau découpage stratigraphique des séries cambriennes des
681 nappes de Pardailhan et du Minervois (versant sud de la Montagne Noire, France).
682 *Géologie de la France*, 2, 3-12.

683

684 Aoya M, Kouketsu Y, Endo S, Shimizu H, Mizukami T, Nakamura D, Wallis S. 2010. Extending
685 the applicability of the Raman carbonaceous - material geothermometer using data
686 from contact metamorphic rocks. *Journal of Metamorphic Geology*, 28, 895-914

687

- 688 Árkai P. 1991. Chlorite crystallinity: an empirical approach and correlation with illite
689 crystallinity, coal rank and mineral facies as exemplified by Palaeozoic and Mesozoic
690 rocks of northeast Hungary. *Journal of Metamorphic Geology*, 9, 723–734.
691
- 692 Arthaud F. 1970. Etude tectonique et microtectonique comparée de deux domaines
693 hercyniens : les nappes de la Montagne Noire (France) et l'anticlinorium de
694 l'Iglesiente (Sardaigne). *Université des Sciences et Techniques du Languedoc*, 175
695 pp.
696
- 697 Baludikay BK, François C, Sforza MC, Beghin J, Cornet Y, Storme JY, Fagel N, Fontaine F,
698 Littke R, Baudet D, Delvaux D, Javaux EJ. 2018. Raman microspectroscopy, bitumen
699 reflectance and illite crystallinity scale: comparison of different geothermometry
700 methods on fossiliferous Proterozoic sedimentary basins (DR Congo, Mauritania and
701 Australia). *International Journal of Coal Geology*, 191, 80-94.
702
- 703 Bard JP, Rambeloson R. (1973). Métamorphisme plurifacial et sens de variation du degré
704 géothermique durant la tectogenèse polyphasée hercynienne dans la partie orientale
705 de la zone axiale de la Montagne Noire (Massif du Caroux, Sud du Massif Central
706 français). *Bulletin de la Société géologique de France*, 7(5-6), 579-586.
707
- 708 Berger G, Alabouvette B, Guérangé-Lozes J, Demange M, Ambert P. 2001. Carte géol. France
709 (1/250 000), feuille Montpellier (38). Orléans : BRGM. Notice explicative par B.
710 Alabouvette M, Démange J, Guérangé-Lozes P. Ambert (2003), 164 pp.
711
- 712 Beyssac O, Goffé B, Chopin C, Rouzaud JN. 2002. Raman spectra of carbonaceous material
713 in metasediments: a new geothermometer. *Journal of metamorphic Geology*, 20, 859-
714 871.
715

- 716 Beyssac O, Pattison DRM, Bourdelle F. 2019. Contrasting degrees of recrystallization of
717 carbonaceous material in the Nelson aureole, British Columbia and Ballachulish
718 aureole, Scotland, with implications for thermometry based on Raman spectroscopy of
719 carbonaceous material, *J. Metamorphic Geol.*, 37, 71-95.
- 720
- 721 Belmar M, Schmidt ST, Frey M, Ferreiro-Mählmann R, Mullis J, Stern W. 2002. Diagenesis,
722 low-grade and contact metamorphism in the Triassic–Jurassic of the Vichuquén–
723 Tilicura and Hualané–Gualleco Basins, Coastal Range of Chile. *Schweizerische*
724 *Mineralogische und Petrographische Mitteilungen*, 82, 375-392.
- 725
- 726 Carmignani L, Carosi R, Di Pisa A, Gattiglio M, Musumeci G, Oggiano G, Carlo Pertusati P.
727 1994. The hercynian chain in Sardinia (Italy). *Geodinamica Acta*, 7(1), 31-47.
- 728
- 729 Carosi R, Perillo M, Pertusati PC. 1998. Structural evolution of the Southern Sulcis
730 Metamorphic Complex (SW Sardinia, Italy). *C. R. Acad. Sci. Paris, Sciences de la terre*
731 *et des planètes/Earth & Planetary Sciences*, 326, 505-512, Paris, Elsevier Gauthier-
732 Villars, 1998.
- 733
- 734 Chardon D, Aretz M, Roques D. 2020. Reappraisal of Variscan tectonics in the southern
735 French Massif Central. *Tectonophysics*, 787.
- 736
- 737 Cocherie A, Baudin T, Autran A, Guerrot C, Fanning CM, Laumonier B. 2005. U-Pb zircon (ID-
738 TIMS and SHRIMP) evidence for the early Ordovician intrusion of metagranites in the
739 late Proterozoic Canaveilles Group of the Pyrenees, and the Montagne Noire (France).
740 *Bull. Société géologique de France*, 176, 269-282.
- 741

- 742 Cózar P, Izart A, Vachard D, Coronado I. 2017 A mid-Tournaisian-late Viséan carbonate ramp
743 reconstructed from nappes and olistolites in the southern Montagne Noire (France).
744 *Sedimentary Geology*, 358, 148-175.
- 745
- 746 Delchini S, Lahfid A, Plunder A, Michard A. 2016. Applicability of the RSCM geothermometry
747 approach in a complex tectono-metamorphic context: The Jebilet massif case study
748 (Variscan Belt, Morocco). *Lithos*, 256, 1-12.
- 749
- 750 Demange M. 1982. Etude géologique du massif de l'Agoût, Montagne Noire, France. Thèse
751 d'Etat, Paris 6, 1050p.
- 752 Demange M. 1993. Que signifie la faille des Monts de Lacaune (Montagne Noire, France) ?
753 Implications quant au problème de la patrie des nappes. *Comptes rendus de*
754 *l'Académie des sciences. Série 2*, 317, 411-418.
- 755
- 756 Doublier MP, Potel S, Wemmer K. 2006. Age and grade of metamorphism in the eastern Monts
757 de Lacaune—implications for the collisional accretion in Variscan externalides (French
758 Massif Central). *Geodinamica Acta*, 19, 391-407.
- 759
- 760 Doublier MP, Potel S, Franke W, Roache T. 2012. Very low-grade metamorphism of
761 Rhenohercynian allochthons (Variscides, Germany): facts and tectonic
762 consequences. *International J. Earth Sciences*, 101, 1229–1252.
- 763
- 764 Doublier MP, Potel S, Wemmer K. 2015. The tectono-metamorphic evolution of the very low-
765 grade hangingwall constrains two-stage gneiss dome formation in the Montagne Noire
766 (Southern France). *Journal of Metamorphic Geology*, 33, 71-89.
- 767
- 768 Dunoyer de Segonzac G, Ferrero J, Kübler B. 1968. Sur la cristallinité de l'illite dans la
769 diagenèse et l'anchimétamorphisme. *Sedimentology*, 10, 137-143

- 770
- 771 Echtler H, Malavieille J. 1990. Extensional tectonics, basement uplift and Stephano-Permian
772 collapse basin in a late Variscan metamorphic core complex (Montagne Noire,
773 Southern Massif Central). *Tectonophysics*, 177, 125-138.
- 774
- 775 Engel W, Feist R, Franke W. 1978. Synorogenic gravitational transport in the Carboniferous of
776 the Montagne Noire (S-France). *Zeitschrift der deutschen geologischen Gesellschaft*,
777 461-472.
- 778 Engel W, Feist R, Franke W. 1980-81. Le Carbonifère anté-stéphanien de la Montagne Noire:
779 rapports entre mise en place des nappes et sédimentation. *Bull. B.R.G.M.* 2, 341-389.
- 780
- 781 Epstein AG, Epstein JB, Harris LD. 1977. Conodont color alteration: an index to organic
782 metamorphism. *Geological Survey of America, Professional paper* 995, 1-27.
- 783
- 784 Faure M, Cottureau N. 1988. Données cinématiques sur la mise en place du dôme migmatique
785 carbonifère moyen de la zone axiale, de la Montagne Noire (Massif Central, France).
786 *Comptes rendus Acad. Sci. Série 2*, 307, 1787-1794.
- 787
- 788 Faure M. 1995. Late orogenic carboniferous extensions in the Variscan French Massif Central.
789 *Tectonics*, 14, 132-153.
- 790
- 791 Faure M, Bé Mézème E, Duguet M, Cartier C, Talbot JY. 2005. Paleozoic tectonic evolution of
792 medio-Europa from the example of the French Massif Central and Massif Armoricaïn.
793 *Journal of the virtual Explorer*, 19, 1-25.
- 794
- 795 Faure M, Lardeaux JM, Ledru P. 2009. A review of the pre-Permian geology of the Variscan
796 French Massif Central. *Comptes Rendus Geoscience*, 341, 202-213.
- 797

- 798 Faure M, Cocherie A, Bé Mézème E, Charles N, Rossi P. 2010. Middle Carboniferous crustal
799 melting in the Variscan belt: New insights from U–Th–Pb tot monazite and U–PB zircon
800 ages of the Montagne Noire Axial Zone (southern French Massif Central). *Gondwana*
801 *Res.* 18, 653–673.
802
- 803 Faure M, Cocherie A, Gaché J, Esnault C, Guerrot C, Rossi P, Lin W, Qiuli L. 2014. Middle
804 Carboniferous intracontinental subduction in the Outer Zone of the Variscan belt
805 (Montagne Noire Axial Zone, French Massif Central): multimethod geochronological
806 approach of polyphase metamorphism. *Multimethod geochronological approach of*
807 *polyphase metamorphism. Geological Society, London, Special Publications*, 405(1),
808 289-311.
809
- 810 Faure M, Li XH, Lin W. 2017. The northwest-directed "Bretonian phase" in the French Variscan
811 Belt (Massif Central and Massif Armoricaïn): a consequence of the Early Carboniferous
812 Gondwana-Laurussia collision. *C. R Géoscience*, 349, 126-136.
813
- 814 Feist R, Galtier J. 1985. Découverte de flores d'âge namurien probable dans le flysch à
815 olistolites de Cabrières (Hérault). Implication sur la durée de la sédimentation
816 synorogénique dans la Montagne Noire (France méridionale). *Comptes Rendus Acad.*
817 *Sciences. Série 2*, 300, 207-212.
818
- 819 Fielitz W, Mansy JL. 1999. Pre-and synorogenic burial metamorphism in the Ardenne and
820 neighbouring areas (Renohercynian zone, central European Variscides).
821 *Tectonophysics*, 309, 227-256.
822
- 823 Franke W, Doublier MP, Klama K, Potel S, Wemmer K. 2011. Hot metamorphic core complex
824 in a cold foreland. *International Journal of Earth Sciences*, 100, 753-785.
825

- 826 Fréville K, Cenki-Tok B, Trap P, Rabin M, Leyreloup A, Régnier JL, Whitney DL. 2016. Thermal
827 interaction of middle and upper crust during gneiss dome formation: Example from the
828 Montagne Noire (French Massif Central). *Journal of Metamorphic Geology*, 34, 447-
829 462.
- 830
- 831 Frey M. Very low grade metamorphism of clastic sedimentary rock. In : Frey M, Ed. *Low*
832 *temperature metamorphism*. Chapman and Hall, New York, 1987, pp. 9-58.
- 833
- 834 Frey, M., Robinson, D., (eds) 1999, *Low-grade metamorphism*. Blackwell, 313 pp.
- 835
- 836 Fukuchi R, Fujimoto K, Kameda J, Hamahashi M, Yamaguchi A, Kimura G, Hamada Y,
837 Hashimoto Y, Kitamura Y, Saito S. 2014. Changes in illite crystallinity within an ancient
838 tectonic boundary thrust caused by thermal, mechanical, and hydrothermal effects: an
839 example from the Nobeoka Thrust, southwest Japan, *Earth, Planets and Space*,
840 66:116, 1-12.
- 841
- 842 García-López S, Bastida F, Aller J, Sanz-López J. 2001. Geothermal palaeogradients and
843 metamorphic zonation from the conodont colour alteration index (CAI). *Terra Nova*,
844 13(2), 79-83.
- 845
- 846 Gerya T, Stöckhert B. 2002. Exhumation rates of high-pressure metamorphic rocks in
847 subduction channels: The effect of Rheology. *Geophysical Research Letters*, 29, 1261-
848 1264.
- 849
- 850 Gèze B. 1949. *Etude géologique de la Montagne Noire et des Cévennes méridionales*.
851 *Soc.Géol. Fr. Mém.* 62, 1–125.
- 852

- 853 Guggenheim S, Bain DC, Bergaya F, Brigatti MF, Drits VA, Eberl DD, Formoso MLL, Galán
854 E, Merriman RJ, Peacor DR, Stanjek H, Watanabe T. 2002. Report of the Association
855 Internationale pour l'Etude des Argiles (AIPEA) Nomenclature Committee for 2001:
856 Order, disorder and crystallinity in phyllosilicates and the use of the 'Crystallinity
857 Index'. *Clay Minerals*, 37, 389-393.
- 858
- 859 Guiraud M, Sauniac S, Burg JP. 1981. Précision sur les conditions Pression-Température lors
860 de la mise en place de la nappe de Pardailhan (Montagne Noire), par la détermination
861 des inclusions fluides. *Comptes-rendus de l'Académie des Sciences, Série II*, 292, 229-
862 232.
- 863
- 864 Hamet J, Allègre CJ. 1976. Hercynian orogeny in the Montagne Noire (France): Application
865 of Rb87-Sr87 systematics. *Geological Society of America Bulletin*, 87, 1429-1442.
- 866
- 867 Hara H, Kurihara T, Mori H. 2013. Tectono-stratigraphy and low-grade metamorphism of
868 Late Permian and Early Jurassic accretionary complexes within the Kurosegawa belt,
869 Southwest Japan: Implications for mechanisms of crustal displacement within active
870 continental margin. *Tectonophysics*, 592, 80-93.
- 871
- 872 Harris LB, Burg JP, Sauniac S. 1983. Strain distribution within the Pardailhan nappe
873 (Montagne Noire, France), and structure of its basal thrust zone: implications for
874 events associated with nappe emplacement. *J. Struct. Geol.* 5, 431-440.
- 875
- 876 Hilchie LJ, Jamieson RA. 2014. Graphite thermometry in a low-pressure contact aureole,
877 Halifax, Nova Scotia. *Lithos*, 208, 21-33.
- 878

- 879 Kouketsu Y, Mizukami T, Mori H, Endo S, Aoya M, Hara H, Nakamura D, Wallis S. 2014. A
880 new approach to develop the Raman carbonaceous material geothermometer for low-
881 grade metamorphism using peak width. *Island Arc*, 23, 33–50.
- 882
- 883 Kretschmer S, Franke W, Wemmer K, Köngigshof P, Gerdes A. 2015. Tectono-metamorphic
884 evolution of the Mouthoumet Massif (Variscides, S-France): Interference of orogenic
885 accretion and crustal extension: *Géologie de la France*, 1, 78.
- 886
- 887 Kübler B. 1964. Les argiles, indicateurs de métamorphisme. *Revue Institut de la Française*
888 *de Pétrole*, 19, 1093–1112.
- 889
- 890 Kübler B. 1968. Evaluation quantitative du metamorphism par la cristallinité de l'illite. *Bull.*
891 *Centre Rech. Pau-SNPA*, 2, 385-397.
- 892
- 893 Lahfid A, Beyssac O, Deville E, Negro F, Chopin C, Goffé B. 2010. Evolution of the Raman
894 spectrum of carbonaceous material in low-grade metasediments of the Glarus Alps
895 (Switzerland). *Terra Nova*, 22, 354–360.
- 896
- 897 Le Bayon R, Brey GP, Ernst WG, Mählmann RF. 2011. Experimental kinetic study of organic
898 matter maturation: time and pressure effects on vitrinite reflectance at 400 C. *Organic*
899 *Geochemistry*, 42, 340-355.
- 900
- 901 Lescuyer JL, Bouchot V, Cassard D, Feybesse JL, Marcoux E, Moine B, Piantone P, Teygey
902 M, Tollon F. Le gisement aurifère de Salsigne (Aude, France) : Une concentration
903 syntectonique tardivarisque dans les sédiments détritiques et carbonatés de la
904 Montagne Noire: *Chronique de la Recherche Minière*, n° 512, (1993) pp 3–73.
- 905

- 906 Mählmann RF. 2001. Correlation of very low-grade data to calibrate a thermal maturity model
907 in a nappe tectonic setting, a case study from the Alps. *Tectonophysics*, 334, 1-33.
908
- 909 Mählmann RF, Bozkaya Ö, Potel S, Le Bayon R, Šegvić B, Nieto F. 2012. The pioneer work
910 of Bernard Kübler and Martin Frey in very low-grade metamorphic terranes: paleo-
911 geothermal potential of variation in Kübler-Index/organic matter reflectance
912 correlations. A review. *Swiss Journal of Geosciences*, 105(2), 121-152.
913
- 914 Malavieille J. 2010. Impact of erosion, sedimentation and structural heritage on the structure
915 and kinematics of orogenic wedges: analog models and case studies. *GSA Today*, 20,
916 4–10.
917
- 918 Mattauer M, Laurent P, Matte P. 1996. Plissement hercynien synschisteux post-nappe et
919 étirement subhorizontal dans le versant sud de la Montagne noire (Sud du Massif
920 central, France). *Comptes rendus de l'Acad. des Sciences. Série 2*, 322, 309-315.
921
- 922 Matte P, Lancelot J, Mattauer M. 1998. La zone axiale hercynienne de la Montagne Noire n'est
923 pas un "metamorphic core complex" extensif mais un anticlinal post-nappe à cœur
924 anatectique. *Geodinamica Acta*, 11, 13-22.
925
- 926 Merriman R, Frey M. Patterns of very low-grade metamorphism in metapelitic rocks. In :
927 Robinson D, Frey M, Eds. *Low-grade metamorphism*, 1999, pp. 61-108.
928
- 929 Merriman R, Peacor DR. Very low-grade metapelites: mineralogy, microfabrics and measuring
930 reaction progress. In : Robinson D, Frey, M Eds. *Low-grade metamorphism*, 1999, pp.
931 10-60.
932

- 933 Mori H, Wallis S, Fujimoto K, Shigematsu N. 2015. Recognition of shear heating on a long-
934 lived major fault using Raman carbonaceous material thermometry: implications for
935 strength and displacement history of the MTL, SW Japan. *Island Arc*, 24, 425-446.
936
- 937 Mori H, Mori N, Wallis S, Westaway R, Annen C. 2017. The importance of heating duration
938 for Raman CM thermometry: evidence from contact metamorphism around the Great
939 Whin Sill intrusion, UK. *Journal of Metamorphic Geology*, 35, 165-180.
940
- 941 Mukoyoshi H, Hirono T, Hara H, Sekine K, Tsuchiya N, Sakaguchi A, Soh W. 2009. Style of
942 fluid flow and deformation in and around an ancient out-of-sequence thrust: An example
943 from the Nobeoka Tectonic Line in the Shimanto accretionary complex, southwest
944 Japan, *Island Arc*, 18, 333-351.
945
- 946 Mullis J, Mählmann RF, Wolf M. 2017. Fluid inclusion microthermometry to calibrate vitrinite
947 reflectance (between 50 and 270° C), illite Kübler-Index data and the
948 diagenesis/anchizone boundary in the external part of the Central Alps. *Applied Clay
949 Science*, 143, 307-319.
950
- 951 Pitra P, Poujol M, Van Den Driessche J, Poilvet JC, Paquette JL. (2012). Early Permian
952 extensional shearing of an ordovician granite: The saint-eutrope “c/s-like” orthogneiss
953 (montagne noire, French massif central). *Comptes Rendus Géoscience*, 344(8), 377-
954 384.
955
- 956 Poilvet JC, Poujol M, Pitra P, Van Den Driessche J, Paquette JL. (2011). The Montalet granite,
957 Montagne Noire, France: An Early Permian syn-extensional pluton as evidenced by
958 new U-Th-Pb data on zircon and monazite. *Comptes Rendus Géoscience*, 343(7), 454-
959 461.

- 960 Poty É, Aretz M, Barchy L. 2002, Stratigraphie et sédimentologie des «calcaires à Productus»
961 du Carbonifère inférieur de la Montagne noire (Massif central, France). Comptes
962 Rendus Geoscience, 334, 843-848.
- 963
- 964 Rahl JM, Anderson KM, Brandon MT, Fassoulas C. 2005. Raman spectroscopic carbonaceous
965 material thermometry of low-grade metamorphic rocks: calibration and application to
966 tectonic exhumation in Crete, Greece. Earth and Planetary Science Letters, 240, 339-
967 354.
- 968
- 969 Rahn M, Mullis J, Erdelbrock K, Frey M. 1995. Alpine metamorphism in the north Helvetic
970 flysch of the Glarus-Alps, Switzerland. Eclogae Geologicae Helvetiae, 88, 157-178.
- 971
- 972 Roger F, Respaut JP, Brunel M, Matte P, Paquette JL. 2004. Première datation U-Pb des
973 orthogneiss œillés de la zone axiale de la Montagne noire (Sud du Massif central):
974 nouveaux témoins du magmatisme ordovicien dans la chaîne Varisque. Comptes
975 Rendus Geoscience, 336, 19-28.
- 976
- 977 Roger F, Teyssier C, Respaut JP, Rey PF, Jolivet M, Whitney DL, Brunel M. 2015. Timing of
978 formation and exhumation of the Montagne Noire double dome, French Massif Central.
979 Tectonophysics, 640, 53-69.
- 980
- 981 Sadezky A, Muckenhuber H, Grothe H, Niessner R, Pöschl U. 2005. Raman
982 microspectroscopy of soot and related carbonaceous materials: spectral analysis and
983 structural information. Carbon, 43, 1731-1742.
- 984
- 985 Sauniac S. 1980. Utilisation des exsudats de quartz comme critères de reconnaissance d'un
986 régime cisailant: exemple de la base de la nappe de Pardailhan (versant sud de la
987 Montagne Noire). Rev. Géol. Dyn. Geog. Phys, 22, 177-186.

- 988
- 989 Schuiling R. 1960. Le dôme gneissique de l'Agoût (Tarn et Hérault). Mémoire de la
990 Société Géologique de France, 91.
- 991
- 992 Soula JC, Debat P, Brusset S, Bessière G, Christophoul F, Déramond J. 2001. Thrust-
993 related, diapiric, and extensional doming in a frontal orogenic wedge: example of the
994 Montagne Noire, Southern French Hercynian Belt. *Journal of Structural Geology*, 23,
995 1677-1699.
- 996
- 997 Thompson PH, Bard JP. 1982. Isograds and mineral assemblages in the eastern Axial Zone,
998 Montagne Noire (France): implications for temperature gradients and P–T history.
999 *Canadian Journal of Earth Sciences*, 19, 129-143.
- 1000
- 1001 Trap P, Roger F, Cenki-Tok B, Paquette JL. 2017. Timing and duration of partial melting and
1002 magmatism in the Variscan Montagne Noire gneiss dome (French Massif Central).
1003 *International Journal of Earth Sciences*, 106, 453-476.
- 1004
- 1005 Underwood M, Byrne T, Hibbard JP, DiTullio L. A comparison among organic and inorganic
1006 indicators of diagenesis and low-temperature metamorphism, Tertiary Shimanto Belt,
1007 Shikoku, Japan. In : Underwood M, Ed. *Thermal evolution of the Tertiary Shimanto*
1008 *Belt, southwest Japan: an example of ridge-trench interaction*, 1993, pp. 45-61.
- 1009
- 1010 Vachard D, Aretz M. 2004. Biostratigraphical precisions on the Early Serpukhovian (Late
1011 Mississippian), by means of a carbonate algal microflora (cyanobacteria, algae and
1012 pseudo-algae) from La Serre (Montagne Noire, France). *Geobios*, 37, 643-666.
- 1013

- 1014 Vachard D, Izart A, Cózar P. 2017. Mississippian (middle Tournaisian-late Serpukhovian)
1015 lithostratigraphic and tectonosedimentary units of the southeastern Montagne Noire
1016 (Hérault, France). *Géologie de la France*, 1, 47-88
1017
- 1018 Van Den Driessche J, Brun JP. (1989). Un modèle cinématique de l'extension paléozoïque
1019 supérieur dans le Sud du Massif Central. *Comptes rendus de l'Académie des*
1020 *sciences. Série 2, Mécanique, Physique, Chimie, Sciences de l'univers, Sciences de*
1021 *la Terre*, 309(16), 1607-1613.
1022
- 1023 Van Den Driessche J, Brun JP. 1992. Tectonic evolution of the Montagne Noire (French Massif
1024 Central): a model of extensional gneiss dome. *Geodinamica Acta*, 5, 85-97.
1025
- 1026 Velde B, Lanson B. 1993. Comparison of I/S transformation and maturity of organic matter at
1027 elevated temperatures. *Clays and Clay Minerals*, 41, 178-178.
1028
- 1029 Vizcaïno D, Álvaro JJ. 2001. The Cambrian and lower Ordovician of the southern Montagne
1030 Noire: a synthesis for the beginning of the new century. *Ann. Soc. géol.Nord (2° sér.)*
1031 8, 185–242.
1032
- 1033 Wada H, Tomita T, Matsuura K, Tuchi K, Ito M, Morikiyo T. 1994. Graphitization of
1034 carbonaceous matter during metamorphism with references to carbonate and pelitic
1035 rocks of contact and regional metamorphisms, Japan. *Contributions to Mineralogy*
1036 *and Petrology*, 118, 217-228.
1037
- 1038 Weber K. 1972. Notes on the determination of illite crystallinity. *Neues Jahrbuch für*
1039 *Mineralogie, Monatshefte*, 6, 267-276.
1040

- 1041 Whitney DL, Roger F, Teyssier C, Rey PF, Respaut JP. 2015. Syn-collapse eclogite
1042 metamorphism and exhumation of deep crust in a migmatite dome: the P–T–t record
1043 of the youngest Variscan eclogite (Montagne Noire, French Massif Central). *Earth
1044 and Planetary Science Letters*, 430, 224-234.
- 1045
- 1046 Wiederer U, Koenigshof P, Feist R, Franke W, Doublier MP. (2002). Low-grade metamorphism
1047 in the Montagne Noire (S-France): Conodont Alteration Index (CAI) in Palaeozoic
1048 carbonates and implications for the exhumation of a hot metamorphic core complex.
1049 *Schweizerische Mineralogische und Petrographische Mitteilungen*, 82, 393–407.
- 1050
- 1051 Zhu C, Rao S, Hu S. 2016. Application of illite crystallinity for paleo - temperature
1052 reconstruction: A case study in the western Sichuan basin, SW China. *Carpathian
1053 Journal of Earth and Environmental Sciences*, 11(2), 599-608.
- 1054
- 1055

1056 **Figure captions**

1057 **Fig. 1.** Variscan massifs map with location of the Viséan-Serpukhovian migmatites. The red
1058 rectangle represents the study area (modified from Faure et al., 2010).

1059 **Fig. 2.** Simplified structural map of the Montagne Noire showing the granite-migmatite dome
1060 of the Axial Zone, and the sedimentary northern and southern flanks (modified from Gèze,
1061 1949; Arthaud, 1970; Faure et al., 2010). The recumbent folding in the south area is coeval
1062 with the sedimentation in the Viséan-Serpukhovian basin. The stratigraphy presents a
1063 reverse order in most part of these series. The figure was made from the vectorised map
1064 250 000 of Montpellier by the BRGM (Berger et al., 2001).

1065 **Fig. 3.** Paleozoic stratigraphic log, with the abundance in organic matter shown on the left
1066 column (modified from Arthaud, 1970; Engel et al. 1980-1981; Álvaro and Vizcaïno, 1998;
1067 and Vizcaïno and Álvaro, 2001).

1068 **Fig. 4.** Pictures showing the two main deformation phases in the Montagne Noire southern
1069 flank. (a). Steep inverted limb of a S-verging fold coeval with the formation of the km-scale
1070 recumbent folds (F1), near Ferrals-les-Montagnes. Sample H52 was picked up for RSCM
1071 study. (b). Upright fold with axial planar cleavage (S2) coeval with the Axial Zone doming (F2)
1072 and location of the sample H41, close to S^t-Pons. The folded layers are S0-S1 surface where
1073 (S1) is coeval with recumbent folding (F1).

1074 **Fig. 5.** Peakfitting of samples H52 (a), and H75 (b) by low temperature (Lahfid et al., 2010)
1075 Sand high temperature method (Beyssac et al., 2002), respectively.

1076 **Fig. 6.** Thin sections observed with optical microscope in polarized non-analysed light showing
1077 the lithology diversity, (a). Mudstone in the Viséan Turbidite (sample G8), (b). Siltstone in the
1078 Early Ordovician rocks (sample G1), (c). Early Cambrian green sandstone (Marcory formation,
1079 sample H32), (d). Mudstone in the Early Ordovician rocks (sample G2), (e). Middle Cambrian
1080 fine-grained sandstone (sample H22), (f). Early Ordovician mudstone (sample G3).

1081 **Fig. 7.** Graph showing the repartition of our data with respect to the two methods used. The
1082 right column shows the evolution of the spectra aspect and the deconvolution chosen with
1083 D1 band (blue), D2 band (red), D3 band (yellow), D4 band (orange) and G band (green).

1084 **Fig. 8.** Geological map of the Montagne Noire with the Tmax measured in the southern flank.
1085 Totally, 74 samples from the recumbent folds and turbiditic basin have been analysed by the
1086 RSCM method in the present study.

1087 **Fig. 9.** Projections of the analyzed samples on five cross sections (located in figure 8) of the
1088 southern flank of the Montagne Noire. Temperatures are globally increasing toward the Axial
1089 Zone dome.

1090 **Fig. 10.** Temperature map determined by the RSCM Tmax results of this study. The isotherms
1091 crosscut the tectonic units of the southern flank. Roughly, the temperature decreases from the
1092 NW to the SE away from the Axial Zone dome. The isotherms cut the micaschists between
1093 Salsigne and St-Pons because of a lack of data.

1094 **Fig. 11.** Comparison of temperatures in °C acquired via RSCM method in this study and
1095 Fréville et al. (2015) and Guiraud et al. (1981) temperature from fluid inclusion.

1096 **Fig. 12.** Illite crystallinity data from previous studies (Engel et al., 1981; Doublier et al., 2015)
1097 based on the crystalline organization of illite analysed via X-ray diffraction converted from the
1098 Kübler index into °C (Merriman and Frey, 1999; Zhu et al., 2016). The Fig. 14 is based on the
1099 data from the red rectangle.

1100 **Fig. 13.** Conodont index alteration method from previous study (Wiederer et al., 2002). The
1101 colour of apatite crystals indicates its metamorphic degree. Wiederer et al. (2002) have
1102 studied more than three hundred conodonts.

1103 **Fig. 14.** Comparison of Lahfid et al. (2010) and Kouketsu et al. (2014) thermometers, arrows
1104 represent the major linear trends.

1105 **Fig. 15.** (a). Plot of the variation of measured RSCM temperatures using different methods
1106 with respect to the distance to the dome. (b). Histograms of frequencies of the full width at
1107 half maximum of the band D1 (FWHM) acquired using the Kouketsu et al. (2014) method.

1108 **Table captions**

1109 **Table. 1.** Tmax results obtained by RSCM according to Beyssac et al. (2002) and Lahfid et al.
1110 (2010) methods

1111 **Table. 2.** Tmax results obtained by RSCM according to Kouketsu et al. (2014) method