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1 **Basement-cover decoupling during the inversion of a hyperextended basin: insights**
2 **from the Eastern Pyrenees**

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9 **Abstract**

10 Deformation processes related to early stages of collisional belts, especially the inversion of
11 rifted systems remain poorly constrained, partly because evidence of these processes is usually
12 obliterated during the subsequent collision. The Pyrenean belt resulting from the inversion of a
13 Cretaceous hyperextended rifted margin associated with a HT/LP metamorphism in the Internal
14 Metamorphic Zone (IMZ), is a good example for studying the early stage of orogenic
15 deformation. This study is focused on the Eastern Pyrenees where the relation between inverted
16 Mesozoic rifted basins and their basement are well-preserved. By using maximum temperatures
17 (T_{\max}) estimated by the Raman Spectroscopy of Carbonaceous Materials geothermometer and
18 structural data, we describe the spatial distribution of the various tectono-metamorphic units.
19 T_{\max} recorded in the sedimentary cover exposed to the north and to the south of a Paleozoic
20 basement block (Agly massif), exceed 550°C, while the Paleozoic metasediments and their
21 autochthonous Mesozoic cover show $T_{\max} < 350^\circ\text{C}$. The metamorphic sedimentary cover is
22 affected by ductile deformation, while the basement is only affected by brittle deformation.
23 Post-metamorphism breccias are observed between the basement and the metamorphic

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24 sedimentary cover, which are interpreted as a decollement level in the Upper Triassic
25 evaporites. Unlike previous models suggesting that the basement block separated two
26 metamorphic basins (Boucheville and Bas Agly) during rifting, we propose a large
27 displacement of a single metamorphic basin by a large thrust above the basement block. This
28 novel interpretation emphasizes the general allochthonous position of the former hyperextended
29 rift basin (IMZ) thrust along a decoupling layer.

30 **1. Introduction**

31 It is commonly accepted that the major part of collisional orogens often results from the
32 inversion and contraction of former continental rifted margins. However, the early stages of
33 orogenic processes, corresponding to the inversion of the rifted system, usually affected by
34 hyperextension, remains poorly constrained. The lack of knowledge on the early orogenic
35 processes stems from the obliteration of the hyperextended part of former rift systems, during
36 collision, by a strong metamorphic and tectonic overprint or because this domain is now buried
37 in the crustal root of belts. Therefore, the inherited rift-related geometries and thermal-
38 mechanical structure of the pre-orogenic crust and lithosphere play important roles in the
39 structural style of collisional orogens (e.g. [Lemoine et al., 1986](#); [Watts et al., 1995](#); [Mouthereau](#)
40 [et al., 2013](#); [Jammes & Huisman, 2012](#); [Manatschal et al., 2015](#)). Reconstructing the evolution
41 of collisional belts and especially quantifying the finite shortening requires a precise
42 understanding of their overall structure and kinematics, including the pre-orogenic position of
43 the thrust packages. Recent studies have shown that the occurrence of salt tectonics during the
44 pre-orogenic period can lead to significant overestimation of finite orogenic shortening (e.g.
45 [Jourdon et al., 2020](#); [Izquierdo-Llavall et al., 2020](#); [Labaume & Teixell, 2020](#)). The timing of
46 deposition of weak layers (salt-bearing rocks or shales), pre-, syn- or post-rift, is thus one of the
47 main factors that controls the overall architecture and the structural patterns of rifted basins
48 (e.g. [Jackson and Vendeville, 1994](#); [Rowan, 2014](#); [Duretz et al., 2019](#); [Jourdon et al., 2019](#),

49 2020). When the decoupling layer is pre-rift, it may act as a decollement layer leading to the
50 partitioning of the deformation between the basement and overburden units during rifting (e.g.
51 Jammes *et al.*, 2010; Rowan, 2014; Ducoux *et al.*, 2019; Coleman *et al.*, 2019; Izquierdo-Llavall
52 *et al.*, 2020; Labaume & Teixell, 2020). During contractional deformation, pre-rift weak layers
53 control the geometry of the thrust belt by favoring strain propagation toward the foreland thin-
54 skinned tectonics (e.g. Letouzey *et al.*, 1995; Jourdon *et al.*, 2020).

55 The Pyrenean orogen (Fig. 1) results from the inversion and integration of a Lower Cretaceous
56 hyperextended rift system with a pre-rift Triassic evaporitic layer (Lagabrielle and Bodinier,
57 2008; Jammes *et al.*, 2009; Lagabrielle *et al.*, 2010; Clerc *et al.*, 2012, 2013; Clerc and
58 Lagabrielle, 2014; Masini *et al.*, 2014; Tugend *et al.*, 2014; Lagabrielle *et al.*, 2016; Clerc *et*
59 *al.*, 2016, Duretz *et al.*, 2019; Jourdon *et al.*, 2019, 2020). This crustal thinning episode was
60 associated with a high-temperature and low-pressure (HT/LP) metamorphic event observed
61 within narrow stripes on the south edge of North Pyrenean Zone (NPZ), forming the so-called
62 Internal Metamorphic Zone (IMZ), where pre- to syn-rift sediments display a HT/LP
63 metamorphism up to 600°C and below 4kbar (Bernus-Maury, 1984; Golberg and Leyreloup,
64 1990; Vauchez *et al.*, 2013; Clerc *et al.*, 2015). The NPZ domain is limited by major crustal
65 faults, the North Pyrenean Frontal Thrust in the north and the North Pyrenean Fault along the
66 southern rim, considered as a major discontinuity between two lithospheric plates and assumed
67 to be active since the end of the Variscan orogenic cycle (Mattauer, 1968; Choukroune, 1976;
68 Choukroune and Mattauer, 1978; Choukroune and ECORS Team, 1989).

69 Several studies investigated the tectonic evolution of Pyrenean belt and proposed estimates of
70 the finite shortening based on N-S balanced geological sections (Choukroune and ECORS-
71 Pyrenees Team, 1989; Roure *et al.*, 1989; Muñoz, 1992; Teixell, 1998; Vergés *et al.*, 2002;
72 Martinez-Peña and Casas-Sainz, 2003; Beaumont *et al.*, 2000; Mouthereau *et al.*, 2014; Teixell
73 *et al.*, 2016; Ternois *et al.*, 2019) but the role of decoupling layer and the early orogenic

74 processes were not assessed in details in these reconstructions. Only a few previous works
75 described the early orogenic stages, suggesting the formation of an accretionary prism by the
76 inversion of the former distal part of the rift system (Mouthereau et al., 2014; Ford et al., 2016).
77 A recent study in the Western Pyrenees demonstrated that the early orogenic stage was
78 associated with the inversion of the hyperextended rift system, reactivating rift-related
79 extensional structures (Gomez-Romeu et al., 2019). In addition, the thick pre-rift Triassic
80 evaporites, which had induced salt-tectonics during the Cretaceous rifting and gravity-driven
81 post-rift deformation (e.g. Lopez-Mir et al., 2014; Saura et al., 2016; Teixell et al., 2016) acted
82 as a decollement during the subsequent Pyrenean shortening (Canérot, 1988; Canérot et
83 Lenoble, 1993; James & Canérot, 1999; Canérot, 2008; Ferrer et al., 2009 ; Jammes et al.,
84 2009 ; 2010; Ducoux et al., 2019; Labaume & Teixell, 2020; Izquierdo-Llavall et al., 2020).
85 The geometry of syn-rift structures is then difficult to assess because the former Cretaceous
86 hyperextended rift basins were strongly overprinted by collisional deformation, except in the
87 western and eastern Pyrenees. In particular, in the eastern Pyrenees, extension led to intense
88 crustal boudinage and decoupling of the sedimentary cover from the basement (Clerc and
89 Lagabrielle, 2014; Clerc et al., 2016). In this area, we have access to a complete section across
90 basins which belonged to the former distal portion of the rift system, making this region a good
91 candidate to study early orogenic processes corresponding to the inversion of the hyperextended
92 rift basins.

93 By using new maximum temperature (T_{\max}) reached by sedimentary rocks and structural data,
94 the aim of this study is to reconsider the bulk structure of the NE portion of the eastern Pyrenees
95 and investigate the role of a pre-rift weak evaporitic layer in the partitioning of the deformation
96 during contractional deformation. In the eastern Pyrenees, published rift restorations suggest
97 that the Agly massif was a high separating two basins (Vauchez et al., 2013; Clerc et al., 2015;
98 2016; Ternois et al., 2019; Odlum and Stockli, 2019); (1) the Boucheville basin sits on the rift

99 axis above exhumed lithospheric mantle, and (2) the Bas-Agly/St-Paul-de-Fenouillet basin sits
100 on hyperthinned to thinned continental crust. This architecture is questionable, because it does
101 not explain the HT/LP metamorphism observed in the present-day Bas-Agly syncline with
102 temperatures similar to the Boucheville Basin, nor the low-temperature recorded in the St-Paul-
103 de-Fenouillet syncline. We propose that the IMZ is allochthonous on top of the Agly North
104 Pyrenean massif and its sedimentary cover and that the Bas-Agly syncline is a northern klippe
105 of the IMZ. This interpretation suggests that the amount of Mesozoic sedimentary unit
106 displacement in the eastern Pyrenees may be larger than classically proposed. In this work, we
107 describe the succession of events that produce the present-day Pyrenean geometries, from the
108 Late Cretaceous rifting stage to the Cenozoic main horizontal shortening, showing how the syn-
109 rift geometry was reworked and how it controlled the finite structure of the belt.

110 2. Geological setting

111 2.1. The Pyrenean belt

112 The Pyrenean belt results from the collision between the Eurasian and Iberian plates from late
113 Cretaceous to Miocene time. The present structure of the Pyrenean belt shows an asymmetric
114 double-verging tectonic wedge above the northward underthrusting Iberian continental
115 lithosphere (Choukroune and ECORS Team, 1989; Roure *et al.*, 1989; Muñoz, 1992; Vergés *et*
116 *al.*, 1995; Teixell, 1998; Teixell *et al.*, 2016; Chevrot *et al.*, 2018). Across a maximum width of
117 150 km, the Pyrenees are traditionally divided into five distinct structural domains (Bertrand,
118 1940; Mattauer, 1968; Choukroune & Séguret, 1973; Castéras, 1974; Mattauer & Henry, 1974;
119 Mirouse, 1980; Boillot, 1984) (Figure 1). (1) The Ebro and (2) the Aquitaine basins respectively
120 correspond to the foreland and retro-foreland basins (e.g. Ford *et al.*, 2016; Angrand *et al.*,
121 2018). (3) The NPZ is a WNW-ESE elongated ribbon made of thrusts and folds deforming the
122 Mesozoic sedimentary succession of Cretaceous basins and locally associated with basement
123 blocks (North Pyrenean massifs). The NPZ is limited by the North Pyrenean Frontal Thrust

124 (NPFT) and the North Pyrenean Fault. (4) The Axial Zone, the topographic backbone of the
125 Pyrenees, is mainly composed of Paleozoic rocks deformed during the Variscan and Alpine
126 orogenic cycles (Mattauer, 1964; Mattauer and Seguret, 1966; Debat, 1969). Finally, (5) the
127 South Pyrenean Zone (SPZ) consists of Mesozoic to Eocene sedimentary rocks transported
128 southward over the Ebro basin above the South Pyrenean Frontal Thrust (SPFT) localized
129 within the Triassic evaporites (Séguret, 1972; Vergés and Muñoz, 1990; López-Mir *et al.*,
130 2014).

131 The architecture of the Pyrenees results from the inversion and integration in the orogen of early
132 Aptian to early Cenomanian rifted basins (Lagabrielle and Bodinier, 2008; Jammes *et al.*, 2009,
133 2010; Lagabrielle *et al.*, 2010; Clerc *et al.*, 2012, 2013; Masini *et al.*, 2014; Tugend *et al.*, 2014;
134 Lagabrielle *et al.*, 2016). Rifting has then evolved to hyperextension and mantle exhumation in
135 the NPZ and the Basque-Cantabrian Basin as suggested by several mantle exposures, then to
136 drifting further west in the Bay of Biscay (e.g. Fabries *et al.*, 1991, 1998; Lagabrielle and
137 Bodinier, 2008; Jammes *et al.*, 2009; Clerc *et al.*, 2012, 2013; Masini *et al.*, 2014; Tugend *et*
138 *al.*, 2014; Lagabrielle *et al.*, 2016). It is commonly accepted that hyperextension and mantle
139 exhumation are responsible for a high-temperature and low-pressure (HT/LP) metamorphism
140 (Ravier, 1959; Bernus-Maury, 1984; Azambre and Rossy, 1976; Golberg and Leyreloup, 1990;
141 Dauteuil and Ricou, 1989; Clerc and Lagabrielle, 2014; Clerc *et al.*, 2015; Ducoux, 2017;
142 Ducoux *et al.*, 2019). This HT/LP metamorphic event affecting the southern NPZ it is mainly
143 evidenced by marbles of the IMZ issued from pre-rift to syn-rift sediments, with peak
144 conditions up to 600°C and below 4kbar (Bernus-Maury, 1984; Golberg and Leyreloup, 1990;
145 Vauchez *et al.*, 2013; Clerc *et al.*, 2015). Available T_{\max} , obtained with the Raman Spectroscopy
146 of Carbonaceous Materials (RSCM) method along the IMZ, first suggested the existence of a
147 temperature gradient from west to east (Clerc *et al.* 2015), but more recent studies of the
148 westernmost orogen (i.e. Basque-Cantabrian Basin) showed otherwise with equally high

149 maximum temperatures in the east and west (Ducoux *et al.*, 2019). Published geochronological
150 data indicate ages ranging from Albian to Santonian (110–85 Ma) for this HT/LP metamorphic
151 event, demonstrating a short delay (10 My) between the end of rifting event (Cenomanian) and
152 the end of the HT/LP metamorphism (Santonian) (Albarède and Michard-Vitrac, 1978a;
153 Albarède and Michard-Vitrac, 1978b; Montigny *et al.*, 1986; Golberg and Maluski, 1988;
154 Golberg *et al.*, 1986; Bandet and Gourinard in Thiébaud *et al.*, 1988; Clerc *et al.*, 2015; Chelalou
155 *et al.*, 2016).

156 The end of rifting was rapidly followed by the onset of convergence across the Pyrenean belt
157 during the Late Santonian, as suggested by tectonic and sedimentological data (Garrido-Megias
158 & Rios 1972; Muñoz, 1992; Vergés *et al.*, 1995; Teixell, 1998; Vergés & García-Senz, 2001;
159 García-Senz 2002; McClay *et al.*, 2004; Biteau *et al.*, 2006; Mouthereau *et al.*, 2014). After a
160 quiet tectonic period during the Paleocene (Danian to early Selandian, Ford *et al.*, 2016), the
161 main collisional phase occurred in Eocene-Oligocene times (Muñoz 1992, 2002; Vergès *et al.*
162 2002; Mouthereau *et al.*, 2014), ultimately leading to the present-day structure. The collisional
163 phase ended during the Chattian in the NPZ (Ortiz *et al.*, 2020) but remained active in the
164 southern Pyrenees until the early Miocene (Muñoz *et al.*, 1992; Hogan and Burbank, 1996;
165 Teixell, 1996; Millán Garrido *et al.*, 2000; Millán Garrido, 2006; Jolivet *et al.*, 2007; Oliva-
166 Urcia *et al.*, 2015; Labaume *et al.*, 2016). Finally, after the main collisional event, since the
167 middle Oligocene, the Eastern Pyrenees were affected by extensional deformation associated
168 with the opening of the Valencia Trough and Gulf of Lion (e.g. Gorini *et al.*, 1993; Gorini *et*
169 *al.*, 1994; Mauffret *et al.*, 1995; 2001; Roca *et al.*, 1999; Vergés and Garcia-Senz, 2001; Roca,
170 2001; Wehr *et al.*, 2018; Etheve *et al.*, 2018; Jolivet *et al.*, 2020).

171

172 2.2. The eastern North Pyrenean Zone

173 The architecture of the eastern NPZ consists of the Agly Massif, the easternmost North
174 Pyrenean massif (Fig. 1), bounded by the Boucheville Basin to the south and the St-Paul-de-
175 Fenouillet and Bas-Agly synclines to the north (Fig. 2). The Agly massif, which represents a
176 condensed crustal section, is composed of plutonic and metamorphic rocks affected by a
177 Variscan HT/LP metamorphism (granulite to greenschist facies) (Barnolas *et al.*, 1996; Guille
178 *et al.*, 2018; Olivier *et al.*, 2008; Wickham & Oxburgh, 1986). This metamorphic event is
179 characterized by a strong geothermal gradient, locally higher than 100°C/km (Fonteilles 1976;
180 Vielzeuf 1984; Delay 1990, Berger *et al.*, 1993). The upper parts of the massif are composed
181 by thick Ediacarian to Devonian metasediments and by granitoids (Fonteilles, 1970; Berger *et*
182 *al.*, 1993; Casas and Palacios, 2012). The deepest parts contain thin sheets of paragneisses and
183 orthogneisses derived from granitic sills emplaced at ca. 540 Ma (Guille *et al.*, 2019).
184 Deformation and late Variscan HT/LP metamorphism that affect these para and ortho-derived
185 series have been variably interpreted as distributed mylonitic shear zones during the Variscan
186 orogeny between 305 and 296 Ma (e.g Olivier *et al.*, 2004, 2008; Wickham & Oxburgh, 1986;
187 Vanardois *et al.*, 2020) or during a late Variscan extensional event related to a major detachment
188 fault in the middle crust (Bouhalier *et al.*, 1991). Part of this high-strain mylonitic deformation
189 has alternatively been attributed to the Cretaceous rifting event (Passchier, 1984; Costa and
190 Maluski, 1988; St Blanquat *et al.*, 1990; Delay, 1990; Delay & Paquet, 1989; Paquet and
191 Mansy, 1991; Vauchez *et al.*, 2013; Clerc *et al.*, 2016; Odlum and Stockli, 2019). Low-grade
192 Ordovician to Devonian metasediments (mainly metapelites, shales and carbonates) compose
193 the northeastern part of the Agly massif. They were weakly metamorphosed in the chlorite zone
194 of the regional HT/LP metamorphism (Berger *et al.*, 1993), while high-grade gneissic and
195 granulitic rocks are observed in the southwestern domain (Fig. 2). The Agly basement was also
196 affected by hydrothermal alteration during the rifting episode, as revealed by 98±2 Ma
197 albitization within the St. Arnac pluton (Poujol *et al.*, 2010). This massif is finally crosscut by

198 numerous N110°E-striking faults displaying down-dip slickenlines attributed to Pyrenean
199 collision (Olivier *et al.*, 2004).

200 The Mesozoic sedimentary sequence of the area is exposed in the St-Paul-de-Fenouillet and
201 Bas-Agly synclines, and in the Boucheville basin. It starts with unconformable continental
202 Permian to Middle Triassic deposits, only observed in the southern margin of the Mouthoumet
203 massif. These sediments are missing over the Agly massif and the northern margin of the Axial
204 Zone (Fig. 2). Shallow marine Upper Triassic sediments are carbonates, clays and evaporites
205 (Keuper facies) passing upward to Liassic to late Aptian shallow marine limestones. The Albian
206 period is then marked by the opening of the large and interconnected basins Albian to
207 Cenomanian “Flysch Noir” basin (Debroas & Souquet 1976; Debroas, 1990; Berger *et al.*,
208 1997), later dismembered during convergence and now preserved in the St-Paul-de-Fenouillet,
209 Boucheville and Bas-Agly synclines (Fig. 2). Post-rift late Cenomanian-Turonian flysch
210 deposits are transgressive on underlying older syn-rift sequences (Debroas 1990; Debroas in
211 Ternet *et al.*, 1997), and were continuously deposited until the early-Coniacian, onlapping the
212 northern part of Axial Zone and the North Pyrenean massifs (Debroas, 1987; 1990). The
213 Pyrenean convergence was responsible for the positive inversion of the rifted basins from the
214 Santonian to the Eocene (Bessière *et al.*, 1989). From Miocene to Present, the eastern part of
215 the Pyrenean range was characterized by the development of extensional structures due to the
216 opening of the Gulf of Lion (Berger *et al.*, 1993).

217 The Boucheville unit is a large synclinorium (Choukroune, 1970, Chelalou *et al.*, 2016)
218 showing evidence of early extensional ductile deformation coeval with the rifting event
219 (Chelalou *et al.*, 2016). The Boucheville syncline and the eastern part of the Agly massif are
220 separated from the Axial Zone by the sub-vertical North Pyrenean Fault (Fig. 2). To the
221 northeast of the Agly massif, the Bas-Agly syncline overthrusts the eastern part of the St-Paul-
222 de-Fenouillet syncline (Fig. 2) that is limited, to the north, by the south-dipping NPFT (Fig. 2).

223 The northern border of the Agly massif, corresponding to the southern Bas-Agly syncline,
224 shows the remains of a décollement of the pre-rift cover (Durand-Delga, 1964; Légiér *et al.*,
225 1987; Clerc and Lagabrielle, 2014; Clerc *et al.*, 2016) with tectonic lineations and drag folds
226 indicating a top to the NNE displacement (Légiér *et al.*, 1987; Vauchez *et al.*, 2013). The Bas-
227 Agly/St-Paul-de-Fenouillet and the Boucheville synclines are interpreted in most restorations,
228 as two separate rifted basins on either side of the Agly massif, the latter corresponding to a
229 tilted basement block (Vauchez *et al.*, 2013; Clerc *et al.*, 2015; 2016; Ternois *et al.*, 2019;
230 Odum and Stockli, 2019). In these rift restorations, the Boucheville syncline sits on the rift axis
231 above exhumed lithospheric mantle, while the Bas-Agly and St-Paul-de-Fenouillet synclines
232 sit on hyperthinned to thinned continental crust.

233 3. Materials and Methods

234 To determine the distribution of the HT/LP metamorphism and then compare with the
235 deformation pattern, we sampled 9 Mesozoic rocks of the Bas-Agly and the St-Paul-de-
236 Fenouillet synclines for determining the maximum recorded temperatures with the RSCM
237 method (Beysac *et al.* 2002a; Lahfid *et al.* 2010). This data set completes the previous works
238 focused on the HT/LP metamorphism (Clerc *et al.*, 2015; Chelalou, 2015). This analytical
239 method allows characterizing the structural evolution of carbonaceous material (CM), reflecting
240 a transformation from disordered to well-ordered CM during metamorphism (Wopenka and
241 Pasteris, 1993). The relation of this increasing graphitization with temperature was quantified,
242 leading to a tool to determine peak temperatures attained by metamorphic rocks (Beysac *et al.*,
243 2002a). Since graphitization is an irreversible process, the RSCM method gives the temperature
244 peak (Pasteris and Wopenka, 1991; Beysac *et al.*, 2002a). This is the basis of the RSCM
245 geothermometer, which was calibrated in the range between 330 and 650°C by Beysac *et al.*
246 (2002a) and extended to the range between 200 and 320°C by Lahfid *et al.* (2010). In this study,
247 we applied these two calibrations to estimate paleotemperatures in carbonates and pelitic

248 metasedimentary rocks from the Paleozoic and Upper Cretaceous series of the study area.
249 Raman analysis protocol is described in supplementary material. The entire results are
250 presented in [Table SM1](#) in supplementary materials and [Figures 2 and 3](#) (T_{\max} values of this
251 study are display in red) with the details of the data acquisition protocol.

252 **4. Thermal pattern of the eastern NPZ**

253 4.1. Distribution of the HT/LP metamorphism

254 The eastern part of NPZ has been extensively studied and numerous studies propose
255 temperature estimates for the HT/LP Pyrenean metamorphic event with different methods
256 including the RSCM approach (e.g [Golberg, 1987](#); [Golberg and Leyreloup, 1990](#); [Clerc *et al.*,
257 2015](#); [Chelalou, 2015](#); [Chelalou *et al.*, 2016](#)). The Boucheville syncline is largely affected by
258 this HT/LP metamorphism and has recorded rather homogeneous peak temperatures comprised
259 between 530 and 580°C ([Golberg et Leyreloup, 1990](#); [Chelalou *et al.*, 2016](#)) ([Figs. 2 and 3](#)).
260 Locally, near the southern edge of the basin, within pre-rift Jurassic sediments, T_{\max} are below
261 500°C ([Chelalou *et al.*, 2016](#)). According to [Clerc *et al.* \(2015\)](#), the eastern part of the Agly
262 massif, made of Ordovician to Devonian metasediments, experienced T_{\max} 351±3°C measured
263 in Silurian metasediments and close to the contact with the south margin of the Bas-Agly
264 syncline ([Fig. 2](#)). Our data confirm this trend, with a T_{\max} of 351±12°C measured a few
265 kilometers westward. This range of T_{\max} shows that the thermal event has not affected or has
266 not exceeded 350°C in the upper part of the Agly massif.

267 Remnants of the metamorphic Mesozoic cover lying on the central-western part of Agly
268 basement ([Berger *et al.*, 1993](#); [Fonteilles *et al.*, 1993](#)), more precisely in the Serres de Verges
269 area, correspond to Jurassic and Lower Cretaceous brecciated metasediments. We measured in
270 this unit a T_{\max} of 424± 32°C obtained in foliated clasts ([Fig. 2](#)). This value of T_{\max} shows high-
271 grade metamorphic conditions of the Mesozoic sedimentary cover, located above high-grade

272 Variscan metamorphic rocks of the Agly massif. Thus, there is no observed metamorphic gap
273 between the breccia and the basement. Measured T_{\max} in the Mesozoic metasediments of the
274 eastern Agly massif are instead higher than in Paleozoic rocks, the latter being only affected by
275 low-grade metamorphism under low greenschist-facies conditions.

276 We measured T_{\max} higher than 550°C in the Cretaceous core of the Bas-Agly syncline. The
277 highest T_{\max} estimated in this syncline correspond to respectively $570 \pm 27^\circ\text{C}$ and $566 \pm 15^\circ\text{C}$
278 in Valanginian and Aptian metacarbonates. A T_{\max} of $534 \pm 19^\circ\text{C}$ is recorded in black Albian
279 marbles in the central part of the syncline (Fig. 2). In a previous study (Chelalou, 2015), a
280 similar range of T_{\max} comprise between 530 to 559°C was characterized. In the southern Bas-
281 Agly syncline, close to the contact with the Agly massif and documented by previous studies,
282 T_{\max} decreases around 350°C with a local value of $494^\circ\text{C} \pm 7^\circ\text{C}$ in anhydrite-rich Upper Triassic
283 sediments (Clerc et al., 2015). Intermediate temperatures comprised between 450 and 400°C
284 (Golberg and Leyreloup, 1990) indicate a progressive southward decrease of metamorphic
285 grade (Fig. 2). In the Bas-Agly, the highest temperatures are thus recorded in the central part of
286 the syncline.

287 In the Albian marls of the St-Paul-de-Fenouillet syncline, we measured a local T_{\max} value of
288 246°C and two T_{\max} lower than 200°C. These temperatures corresponding to low-grade
289 metamorphism, are in agreement with previous data obtained in this syncline, which are
290 comprised between 200 and 290°C (Chelalou et al., 2016). In the eastern part of this basin,
291 close to the contact with the Bas-Agly syncline, we also obtained T_{\max} lower than 200°C (Fig.
292 2) confirming that the whole basin has not been heated, which marks a major difference with
293 both the Bas-Agly and Boucheville synclines.

294 A map of Cretaceous HT/LP metamorphism distribution has been interpolated from the
295 available T_{\max} and field geological data in the eastern part of the ZNP (Fig. 3). It shows a clear

296 thermal contrast between the Boucheville and the Bas-Agly synclines on the one hand, and the
297 St-Paul-de-Fenouillet syncline on the other hand. Both the Bas-Agly and Boucheville synclines
298 show lateral temperature gradients within the sediments that cannot be attributed to differential
299 burial. Actually, the highest temperatures are observed in the syn-rift sediments, while pre-rift
300 sediments experienced only moderate temperatures (~400°C for the Boucheville syncline and
301 between 350 and 400°C for Bas-Agly syncline) (Fig. 3). Consequently, HT/LP metamorphic
302 isograds are oblique on the bedding in those basins and on different structures observed in this
303 area. A thermal gap (>250°C) is observed across the Tautavel thrust fault bounding the Bas-
304 Agly and St-Paul-de-Fenouillet synclines, connecting northward with the NPFT (Figs. 2 and
305 3).

306 To summarize the distribution of the thermal imprint of the Cretaceous HT/LP metamorphism,
307 the Boucheville and Bas-Agly synclines show the highest grade of metamorphism with T_{\max}
308 exceeding 500°C. The underlying eastern Agly massif experienced T_{\max} never exceeding
309 350°C, a peak-temperature colder than these two metamorphic Mesozoic synclines.
310 Furthermore, this contrast of T_{\max} between basement and metamorphic Mesozoic sedimentary
311 cover is attested by remnants of Mesozoic sediments located above the Agly massif, where T_{\max}
312 exceeds 400°C. Thermal contrast is most significant between the Boucheville/Bas-Agly-
313 synclines and the St-Paul-de-Fenouillet syncline where T_{\max} does not exceed 250°C.

314 4.2. Deformation of basement versus Mesozoic sedimentary cover

315 Below the basal contact of the Bas-Agly syncline, the north-eastern margin of the Agly massif
316 is made of low metamorphic grade sediments (Fig. 4). Silurian sandstones and feldspathic
317 shales that composed the Paleozoic basement of the northeastern Agly massif still show
318 graptolites and preserved turbiditic sedimentary figures. This upper part of the basement is only
319 affected by a series of N110°E localized metric-scale shear bands (Fig. 5), some of which
320 corresponding to N-verging normal faults (Fig. 5a). We also observe reverse faults with a top-

321 to-the S sense of shear, suggested by folding and reverse offset of sandstone layers (Fig. 5b).
322 The northeast Agly massif is thus affected by various brittle and brittle-ductile shears, but no
323 pervasive ductile deformation is observed (Fig. 5), in agreement with the measured low T_{\max} .
324 Our observations are in agreement with previous studies (Légier *et al.*, 1987; Vauchez *et al.*,
325 2013). Only the contact between low-grade metasediments and high-grade metamorphic rocks
326 underneath are associated with ductile deformation previously attributed to the Variscan event
327 (Berger *et al.*, 1933 and references therein).

328 In contrast with the Paleozoic low-grade metasediments of the Agly Massif affected by brittle
329 deformation, the Rhaetian and Liassic sediments of the Bas-Agly syncline show an intense
330 ductile deformation with stretching lineation and plurimetric-scale boudinage (Figs. 4 and 6),
331 with a NNE-ward sense of shear (Vauchez *et al.*, 2013). The base of this sedimentary cover is
332 associated to the development of north-verging recumbent metric folds as well (Fig. 6a and b).
333 It is worth noting the presence of Upper Triassic evaporites at the base of the sedimentary pile.
334 The contrasted deformation observed in the different units confirmed the distribution of T_{\max}
335 with higher temperature in the cover than in the basement.

336 As for the Bas-Agly, this ductile deformation is also observed in the Boucheville syncline
337 equally affected by the HT/LP metamorphism. This syncline is strongly overprinted by
338 Pyrenean-related compressional deformation responsible for the development of N110°-
339 striking cleavage (Fig. 4). This cleavage is coeval with the development of the main faults (as
340 the North Pyrenean Fault) which bound the Boucheville syncline, attributed to the late
341 convergence (Berger *et al.*, 1993). Concerning the Bas-Agly syncline, only its southern margin
342 is folded and associated to the late subvertical cleavage. This deformation zone is located along
343 the extension of the Latour-de-France Fault, responsible of the late thrusting of the Agly massif
344 over the Bas-Agly. This cleavage is not clearly expressed in the rest of the Bas-Agly syncline.
345 However, this subvertical cleavage extends into the St-Paul-de-Fenouillet syncline until the

346 NPFT (Fig. 4). No cleavage is observed to the north of the North Pyrenean Frontal Thrust. Only
347 salt tectonics-related structure is preserved (Fig. 4).

348 The contact between the Mesozoic sediments and the crystalline/metamorphic basement
349 (orange faults in Fig. 4) is mainly observed in the eastern Boucheville syncline and along the
350 southern margin of the Bas-Agly syncline. This flat-lying contact is characterized by the
351 presence of breccia recognized on the top of the Agly Massif and at the base of Jurassic series
352 in the Bas-Agly. Furthermore, the surface of this contact is subsequently deformed and
353 truncated by a fault (Durand-Delga, 1964). The contact between the Bas-Agly and the St-Paul-
354 de-Fenouillet corresponds to the Tautavel fault which consist to a low angle thrust connected
355 to the NPFT (Fig. 4).

356 The Boucheville and Bas-Agly syncline are always limited by tectonic contacts with the
357 basement. On the opposite, the St-Paul-de-Fenouillet syncline is lying over the northern margin
358 of the Agly massif without any significant deformation (Fig. 4), suggesting that it could be the
359 autochthonous Mesozoic sedimentary cover of the Agly massif.

360 **5. Discussion and interpretation: tectonic consequences of the IMZ thermal** 361 **structure**

362 5.1. Architecture of the eastern North Pyrenean Zone

363 The geological cross-sections of figure 7 show the structure of the eastern part of the IMZ and
364 its relationships with NPZ and the Agly massif (Fig. 7). The Boucheville Basin corresponds to
365 an asymmetric synclorium (e.g. Choukroune, 1976; Chelalou *et al.*, 2016) bounded by two
366 steeply-dipping major faults, including the North Pyrenean Fault to the south. These faults
367 affect both the basement and the IMZ (Figs. 4 and 7a) and can be extended eastward within the
368 Agly massif where they offset Variscan metamorphic isograds (Fig. 7a). These steeply-dipping
369 faults are intersected by late normal faults linked to the formation of grabens filled by Miocene-

370 Quaternary sediments (Figs. 4 and 7b). Variscan structures within the Agly massif have been
371 deformed during the Pyrenean compression, and are affected by WNW-ESE striking reverse
372 faults (Olivier *et al.*, 2004).

373 West of Latour-de-France, the contact between the Bas-Agly syncline and the Agly massif is
374 overturned and faulted (Latour-de-France Fault) while, further east, the Bas-Agly cover is flat-
375 lying on the top of Paleozoic series with a tectonic contact (Figs. 2 and 4). The northern limb
376 of the Bas-Agly syncline overthrusts the St-Paul-de-Fenouillet syncline through the Tautavel
377 Fault that extends and connects with the NPFT further north (Fig. 4). West of Maury, the low-
378 temperature Mesozoic sequence of the St-Paul-de-Fenouillet syncline, which is the cover of the
379 Paleozoic basement of the Agly massif, is still present, only partly detached thanks to the
380 Triassic evaporites.

381 Isograds of metamorphism are represented on the geological cross sections and show the
382 distribution of HT/LP Cretaceous metamorphism in the eastern NPZ (Fig. 7). In the Boucheville
383 Basin, temperature isograds are clearly intersected by steeply-dipping faults that bound the
384 basin (Fig. 7a). Moreover, a large gap of temperature across the Tautavel Fault is evidenced on
385 both cross sections (Fig. 7). Folding observed in the Mesozoic sedimentary cover results partly
386 from the late compressional event, but some folds may result from the halokinesis active during
387 the rifting event. Temperature isograds are overturned, intersected and shifted by late
388 deformational structures and faults. However, the distribution of metamorphism in these
389 synclines is rather heterogenous and shows lateral temperature gradients within the basin.
390 Isograds in the Bas-Agly syncline clearly intersect bedding. This observation was already made
391 in the Western Pyrenees, especially in the Basque-Cantabrian Basin and in the Chaînons
392 Béarnais area, where the obliquity of isotherms on the structure is due to a pre-metamorphic
393 folding related to salt tectonics (Ducoux *et al.* 2019; Izquierdo-Llavall *et al.*, 2020).

394 5.2. Significance of temperature and deformation contrasts across tectonic contacts.

395 The central parts of the Bas-Agly and Boucheville synclines show significantly higher
396 temperature $>450^{\circ}\text{C}$, while the St-Paul-de-Fenouillet syncline shows temperatures $<250^{\circ}\text{C}$.

397 The temperature gap between the Bas-Agly and St-Paul-de-Fenouillet synclines is abrupt and
398 corresponds to the Tautavel thrust fault. On the other hand, the metamorphic contrast is
399 significant between Paleozoic shales with low-grade greenschist facies conditions (*HT/LP*
400 Variscan metamorphism) of the northern Agly massif and Mesozoic sediments with higher
401 T_{max} . Consequently, the Bas-Agly and Boucheville synclines were affected by Cretaceous *HT*-
402 metamorphism, coeval with mantle exhumation, but not the overlying northern Agly massif
403 which never experienced temperature exceeding 350°C . T_{max} measured in the Paleozoic
404 metasediments of the northern Agly massif can either date back to the Paleozoic or to the
405 Cretaceous rifting event, but the measured temperature shows that the basement has seen lower
406 temperatures than the overlying cover. Note that we do not discuss here the southern part of the
407 Agly massif where the *HT/LP* metamorphism has been attributed to the Variscan event
408 (Vielzeuf and Kornprobst, 1984) or to the Cretaceous *HT/LP* episode (Odlum and Stockli,
409 2019). It is commonly admitted that the regional *HT/LP* that affects the NPZ is coeval with the
410 Lower Cretaceous crustal thinning event (Ravier, 1959; Bernus-Maury, 1984; Azambre and
411 Rossy, 1976; Golberg and Leyreloup, 1990; Dauteuil and Ricou, 1989; Clerc and Lagabrielle,
412 2014; Clerc *et al.*, 2015). In the Boucheville synclinorium, the paleogeothermal gradient during
413 this *HT/LP* metamorphic event has been estimated between 70 and $80^{\circ}\text{C}/\text{km}$ (Chelalou *et al.*,
414 2016). Recent low-temperature thermochronological study indicates that the Agly block cooled
415 below 250°C during the syn- to post rifting (117-90 Ma) (Odlum & Stockli, 2019; Ternois *et*
416 *al.*, 2019), indicating significant lower temperatures than bounding metamorphic sedimentary
417 basins that recorded temperatures $>500^{\circ}\text{C}$ during the post-rift period (95-90 Ma) (Clerc *et al.*,

418 2015) . Thus, the HT/LP metamorphism during the Cretaceous would have been confined to
419 the southern Agly massif, like further west in the Aulus Basin for instance.

420 With reference to the distribution of HT-metamorphism, the deformation in the different units
421 is also contrasted. Intense ductile deformation is observed in the Bas-Agly and Boucheville
422 synclines that recorded the highest temperatures, while the underlying Silurian strata
423 corresponding to Paleozoic cover of the Agly Massif, where temperature are lower, are much
424 less deformed and certainly not mylonitized. Remnants of a Mesozoic cover are locally
425 preserved on top of the Agly Massif, such as cataclastic breccias containing clasts of HT
426 ductilely deformed marbles sealed by unmetamorphic cement. These observations and the
427 temperature higher than 400°C in marble clasts in the Serre de Vergés area confirm that the
428 breccia is a post-metamorphic event.

429 In previous studies, the contact between the Agly massif and the Bas-Agly syncline has been
430 interpreted as the result of basal truncation of a detached Mesozoic cover (Durand-Delga, 1964;
431 Légier *et al.*, 1987) and recently reinterpreted as a north-dipping extensional detachment, the
432 Agly massif being in the position of a metamorphic core complex (Vauchez *et al.*, 2013; Clerc
433 and Lagabrielle, 2014) where extensional deformation near the contact in the Mesozoic cover
434 would be expressed by the boudinage and drag folds (Clerc *et al.*, 2016). In this interpretation,
435 the Bas-Agly syncline corresponds to the detached sedimentary cover of the Agly massif,
436 decoupled via a décollement layer in Upper Triassic sediments (Vauchez *et al.*, 2013; Clerc *et*
437 *al.*, 2016). EBSD and field study in carbonate layers of the base of the Bas-Agly syncline
438 suggest that this N- to NE trending ductile shearing was developed at a temperature around
439 400°C (Vauchez *et al.*, 2013). In the opposite way, a recent study suggests that the Lower
440 Cretaceous exhumation of the southern part of the Agly massif by extension was accommodated
441 by a top-to-the south detachment (Odlum & Stockli, 2019).

442 The extensional model and the pre-orogenic restorations of [Vauchez et al \(2013\)](#), [Clerc et al.](#)
443 [\(2016\)](#) and more recently the crustal-scale model of [Ternois et al. \(2019\)](#) and [Odlum & Stockli](#)
444 [\(2019\)](#) do not easily explain the thermal imprint measured in the central part of the Bas-Agly
445 syncline. In particular they do not explain the higher T_{\max} measured in the Bas-Agly syncline.
446 [Figure 8](#) shows two typical situations of a metamorphic core complex and a thrust in terms of
447 deformation and thermal structure. First of all the contrast in T_{\max} is the opposite in the two
448 situations. A detachment with a significant displacement will inevitably lead to cold
449 (superficial) over hot (deep) units, while a thrust will show the opposite arrangement ([Fig. 8a](#)).
450 Then, all detachments show a continuum of ductile-to-brittle deformation within a distinct shear
451 zone that is mostly observed in the lower plate, below the detachment ([Crittenden et al., 1980](#);
452 [Davis and Lister, 1988](#)) and the lower plate display more ductile deformation while the upper
453 plate shows mostly brittle deformation, which is the exact opposite to what is observed in the
454 study area. The deformation observed in the Paleozoic sediments of the top of the Agly massif
455 is weak and certainly does not correspond to a major shear zone.

456 Then the question arises about the significance of the observed ductile deformation at the base
457 of the Bas-Agly syncline. The N-NE trend of stretching lineations and calcite fabrics near the
458 base of the Bas-Agly syncline described by [Vauchez et al. \(2013\)](#) attests for a northward
459 displacement of the metamorphic unit corresponding to the current Bas-Agly syncline, but it
460 does not prove that it corresponds to an extensional deformation. Here, the large thermal
461 contrast between the hot metamorphic Mesozoic cover and the colder Paleozoic basement
462 underneath is opposite and thus more compatible with a thrust than a detachment ([Fig. 8b](#)).

463 Finally, models proposed by [Vauchez et al \(2013\)](#), [Clerc et al. \(2016\)](#), [Ternois et al. \(2019\)](#) and
464 [Odlum & Stockli \(2019\)](#) describe the Boucheville and Bas-Agly synclines as two narrow
465 disconnected basins on either side of the Agly basement massif, but this assumption conflicts
466 with sedimentological observations indicating that Aptian, Albian and Lower Cretaceous

467 pelagic sediments (corresponding to “Flysch noirs”) were likely deposited in a unique large
468 basin and not in two narrow separated basins (Olivier, 2013).

469 The main question then concerns the link between the Mesozoic sedimentary cover and
470 Paleozoic rocks of the Agly massif. A section along the northern margin of the Agly massif
471 further west, near Maury, shows that the St-Paul-de-Fenouillet syncline is the poorly
472 metamorphosed Mesozoic cover of the Agly Paleozoic basement, in agreement with the low
473 temperatures observed in the St-Paul-de-Fenouillet Basin. When moving east, this sedimentary
474 cover is substituted by the Bas-Agly syncline that is more intensely metamorphosed and the
475 contact between the St-Paul-de-Fenouillet syncline and the Bas-Agly corresponds to a well-
476 characterized thrust that connects to the north with the North Pyrenean Frontal thrust. Then,
477 two hypotheses can be proposed: (1) the Bas-Agly syncline may be the eastern equivalent of
478 the St-Paul-de-Fenouillet Basin with a higher metamorphic grade and the two basins were
479 carried along the basal decollement in Upper Triassic evaporites until they got into close
480 contact; (2) alternatively, the Bas-Agly syncline may be the western termination of a rift
481 segment developing eastward, if we consider the eastern termination of Pyrenees as a relay or
482 accommodation zone where the rift axes are shifted. The Boucheville and Bas-Agly synclines
483 may corresponded to V-shaped terminations of two rifted basins, as the western Pyrenees for
484 instance (e.g. Lescoutre & Manatschal, 2020); (3) the Bas-Agly syncline is not the cover of the
485 Agly massif and corresponds instead to a displaced outlier of a more internal and hotter unit
486 where high temperature is due to mantle exhumation as elsewhere in the NPZ.

487 By combining structural observations and new T_{\max} dataset we then propose an alternative
488 interpretation whereby the Bas-Agly syncline corresponds to a northern klippe of the
489 Boucheville Basin thrust over the Agly massif and a part of St-Paul-de-Fenouillet during the
490 Pyrenean shortening. In this interpretation, the Bas-Agly would connect above the Agly massif
491 with the more internal Boucheville syncline, where high temperatures are recorded.

492 The top-north detachment described by [Vauchez et al. \(2013\)](#) in the northeastern part of the
493 Agly Massif is then reinterpreted as a top-to-the north thrust instead, assisted by regional-scale
494 well known Upper Triassic evaporites layer. This interpretation explains the distribution of
495 metamorphism, the observed temperature contrasts recorded within the different synclines and
496 the distribution of deformation.

497 5.3 A 2D crustal-scale geodynamic model

498 [Figure 9](#) presents a basin-scale restoration model showing a possible evolution of the eastern
499 end of the NPZ from the rifting stage to the end of the Pyrenean orogeny. Based on the model
500 of [Ternois et al. \(2019\)](#) for the crustal architecture, we propose a different interpretation for the
501 sedimentary cover, in which the Bas-Agly syncline represents an allochthonous unit initially
502 connected with the Boucheville syncline above a body of exhumed mantle during the
503 Cretaceous rifting, where both formed a single basin and then transported northward by a thrust.
504 The first step represents the rift template before the onset of the convergence during the
505 Santonian with the inversion of the hyperextended rifted margins formed a span of 30 Myrs
506 between early Aptian to early Cenomanian stages ([Fig. 9a](#)). Rifting was associated to
507 hyperthinning of the lithosphere accommodated by the development of a large-scale top-to-the-
508 south detachment fault located at the base of the basin which induced mantle exhumation and
509 HT/LP metamorphism in the pre- and syn-rift sediments. Such model of crustal boudinage was
510 already proposed by [Clerc et al. \(2016\)](#), but we show here that the northern side of the Agly
511 massif was not affected by ductile deformation during rifting. On the opposite, the Cretaceous
512 basin was affected by intense ductile deformation during this episode, especially in the northern
513 part of the Boucheville Basin ([Chelalou et al., 2016](#)). The syn-rift geometry supposes that the
514 Boucheville, Bas-Agly and St-Paul-de-Fenouillet basins are connected and are filled mainly by
515 Albian-Cenomanian deposits. The Agly massif represents a paleo-topographic high, but it was
516 never significantly eroded before the Late Cretaceous, because there is no evidence of Paleozoic

517 clasts in Aptian to Cenomanian deposits (Olivier *et al.*, 2013). It is only covered by a thin
518 succession of syn-rift sediment (Fig. 9a). We suggest that the Agly massif separates two basins
519 probably interconnected laterally farther west and probably farther east. There is no argument
520 to bury the Agly paleo-high beneath a thick pile of syn-rift sediments. On the contrary, there is
521 a possibility that this paleo-high was only buried beneath post-rift successions. This assumption
522 is supported by recent thermochronological data (Odlum & Stockli, 2019; Ternois *et al.*, 2019).
523 The eastern NPZ consist in rift and salt architecture with salt pillows, diapirs and raft of the
524 syn-rift sediment further north, driven by gravity gliding of the Mesozoic cover. In this model,
525 syn-rift sediments were deposited over the pre-rift sediments that experienced gravity gliding
526 and were juxtaposed above exhumed mantle and a thin continental crust corresponding to the
527 future Boucheville and Bas-Agly basins where HT/LP metamorphism developed (Fig. 9a). We
528 assume a juxtaposition of pre- and syn-rift sediments and exhumed mantle; but we cannot
529 exclude the possibility that the Mesozoic basin was sitting on a very thin continental crust and
530 not directly on the exhumed mantle, because there is no evidence of mantle clast and/or
531 volcanism in this area. During the onset of convergence, the hyperextended domain is firstly
532 reactivated and inverted up to collision of both margins Gómez-Romeu *et al.* (2019).

533 The second step corresponds to the Late Cretaceous convergence with the inversion of the two
534 hyper-extended margins after underthrusting of the exhumed mantle and reactivation of the
535 necking domain associated to the development of the south-verging thrusts located in the Axial
536 Zone and the NPFT (Fig. 9b). A major thin-skinned flat-lying thrust rooted in the Upper Triassic
537 decollement level allowed the northward displacement of the metamorphic part of the basin on
538 top of the St-Paul-de-Fenouillet Basin that recorded only low-grade metamorphism. We
539 hypothesize that the inherited detachment at the base of the basin associated to the decollement
540 layer has been reactivated as a thrust. Early salt pillows and diapirs could be responsible for the
541 localization of the major thin-skinned thrust, corresponding to the current Tautavel fault. In

542 addition, this thrust is probably responsible for the development of breccias at the base of the
543 pre-rift cover. Indeed, these breccias may be a witness of the decollement level localized
544 between the basement and the Mesozoic sedimentary cover (Clerc *et al.*, 2016). It is more
545 difficult to estimate the amount and the direction of this general displacement. The displacement
546 and the thermal contrast seem to decrease toward the east, and Bas-Agly syncline seems to be
547 interconnected with the St-Paul-de-Fenouillet syncline. It may be possible that the deformation
548 softened eastward, especially as this area corresponding to a well-known transfer zone
549 corresponding to the Corbières virgation (Tugend *et al.*, 2014). In the rest of the Pyrenees, the
550 IMZ where higher T_{\max} are recorded, is always interconnected with non-metamorphic basins to
551 the north where North Pyrenean massifs are missing or not outcropping (Clerc *et al.*, 2015).
552 Recent study indicates that the thermal imprint of the metamorphism gently decreases when
553 moving away from the former distal part of the rift system (Ducoux *et al.*, 2019).

554 The Paleocene corresponds to a period of tectonic quiescence, the convergence rate strongly
555 decreasing (Desegaulx and Brunet, 1990; Ford *et al.*, 2016; Rougier *et al.*, 2016; Machiavelli
556 *et al.*, 2017; Grool *et al.*, 2018; Dielforder *et al.*, 2019; Ternois *et al.*, 2019), as emphasized by
557 the absence of cooling at that age in the Agly blocks (Ternois *et al.*, 2019) and by a minimal
558 deposition of clastic sediments in the Aquitaine retro-foreland basin (Desegaulx and Brunet,
559 1990, Ford *et al.*, 2016). This tectonic quiescence could be attributed to the transition period
560 between the closure and inversion of both hyperextended domain and the beginning of crustal
561 thickening related to the collision strictly speaking.

562 The third step corresponds to the onset of the main collision phase and the beginning of crustal
563 thickening involving thick-skinned tectonic style and the development of crustal-scale thrust
564 faults in the pro-wedge domain (Ternois *et al.*, 2019) (Fig. 9c). A major part of the deformation
565 is then accommodated within the basement by the reactivation of former normal faults in the
566 retro-wedge and by the neoformation of thrusts in the pro-wedge. The progressive crustal

567 thickening is responsible for basement exhumation, especially the Agly massif and sedimentary
568 cover as well (Fig. 9c). Recent thermochronological data indeed indicate a rapid exhumation of
569 the Agly massif near the surface during the Eocene, while the NPZ continues to shorten and
570 deformation migrates northward (Ternois *et al.*, 2019). These recent data are in agreement with
571 other published thermochronological data which consistently suggest such a timing: 49 ± 4 to
572 35 ± 3 Ma in Mauléon Basin (Western Pyrenees) (Vacherat *et al.*, 2014); 50 to 35 Ma for North
573 Pyrenean massifs in the central Pyrenees (Vacherat *et al.*, 2016); 44 to 35 Ma in the Axial Zone
574 (Fitzgerald *et al.*, 1999). In our model, the exhumation of the Agly block along the main thrust
575 caused a shift of the primary thrust (i.e. Tautavel fault, Figs. 2, 4 and 6), and a folding of the
576 Mesozoic cover that generated the current shape of the Bas-Agly syncline.

577 The last step corresponds to the end of collision with nappe stacking of the Axial Zone (Fig.
578 9d). The shortening is mainly localized on the future SPFT at depth and on the NPFT. From the
579 middle Eocene to the early Oligocene, the Axial Zone records higher exhumation rates
580 (Whitchurch *et al.*, 2011; Rushlow *et al.*, 2013; Mouthereau *et al.*, 2014; Bosch *et al.*, 2016;
581 Labaume *et al.*, 2016), while the exhumation of the Agly massif ceases (Gunnell *et al.*, 2009;
582 Ternois *et al.*, 2019). The NPZ is finally tilted and uplifted by the exhumation of the Axial Zone
583 related to crustal stacking. Consequently, the Boucheville and Bas-Agly synclines are now
584 separated by the Agly Massif (Fig. 9d). Only small remnants of sedimentary cover are observed
585 on the top of Agly massif. In this configuration, the North Pyrenean Fault corresponds to an
586 inherited structure related to the rifting event and was only reactivated as a reverse fault during
587 the end of collision. The main part of the deformation was accommodated by thick-skinned
588 deformation expressed by the stacking of basement nappes, corresponding to the present-day
589 Axial Zone. Thin-skinned deformation of the sedimentary cover was accommodated by the
590 Upper Triassic decoupling layer. Consequently, the distribution of the deformation is drastically
591 different between the basement and the sedimentary cover when an efficient decoupling layer

592 is present (e.g. [Jammes et al., 2010](#); [Jourdon et al., 2020](#)). The presence of the Triassic evaporite
593 layer thus has considerably impacted the tectonic history of the eastern Pyrenees. First during
594 the rifting stage and then during the Pyrenean shortening.

595 After the main collision, the eastern Pyrenees was affected by normal faults associated with the
596 opening of Valencia Trough since the middle Oligocene (e.g. [Roca et al., 1999](#); [Roca, 2001](#);
597 [Etheve et al., 2018](#); [Jolivet et al., 2020](#)).

598 5.4 Rift inheritances control the finite orogenic architecture

599 Classical models describing the geodynamic evolution of the Pyrenean range do not address the
600 early convergence and the onset of the collision of the two hyper-extended margins. Before the
601 two last decades, the Pyrenean rift was described as a narrow rift and hyperextension and mantle
602 exhumation were thus not considered in geodynamic models (e.g. [Choukroune et al., 1990](#);
603 [Munoz, 1992](#); [Beaumont et al., 2000](#); [Canérot, 2016](#)). Some recent models suggested that the
604 distal part of the rift system, corresponding to deep basins sitting on the exhumed mantle,
605 formed an accretionary prism during the subsequent early inversion (e.g. [Mouthereau et al.,](#)
606 [2014](#); [Ford et al., 2016](#)). The early convergence setting is characterized by thin-skinned
607 tectonics in the sedimentary pile which corresponded to the former hyperextended domain
608 (mantle exhumation) ([Fig. 9](#)). A decollement propagates at the interface between Mesozoic
609 sediments and the basement (Paleozoic rocks and subcontinental mantle) within the Upper
610 Triassic evaporites. We further suggest that most of the distal rifted basin is transported over
611 the former necking domain along the former Cretaceous south-dipping detachment ([Fig. 9](#)).
612 This tectonic process is in agreement with recent studies promoting large displacements of
613 rifted basins along inherited decoupling layer ([Gomez-Romeu et al., 2019](#); [Labaume & Teixell,](#)
614 [2020](#)). Rift-related inheritances (crustal structures and rheology) may fundamentally control the
615 development and the finite structure of orogens by controlling the depth of the decoupling layer
616 and its propagation (e.g., [Tugend et al., 2014](#); [Lacombe & Bellahsen, 2016](#); [Gomez-Romeu et](#)

617 al., 2019; Tavani et al., 2020; Lescoutre & Manatschal, 2020). Inversion of the hyperextended
618 domains depends of the depth of the efficient decoupling layer. The positioning of the
619 decoupling layer is variable and can be located at the interface between the basement and the
620 sedimentary units (e.g. Bellahsen et al., 2012; Muñoz et al., 2013; this study) or in the
621 serpentinized exhumed mantle which corresponds to the weakest part of rifted margins (Pérez-
622 Gussinyé et al., 2001; Péron-Pinvidic et al., 2008), where deformation may preferentially
623 initiate during the contractional deformation (Péron-Pinvidic et al., 2008; Lundin and Doré,
624 2011; Tugend et al., 2014 and 2015a; Gomez-Romeu et al., 2019). In the necking and proximal
625 domains, the continental crust is thicker, and the decoupling layer consequently develops in the
626 middle-lower crust (e.g. Pfiffner, 2017; Jourdon et al., 2019; Tavani et al., 2020). Therefore,
627 during contractional deformation thick-skinned tectonics is prevailing, and inherited rift-related
628 crustal structures may be reactivated (e.g. Froitzheim et al., 1988; Letouzey, 1990; Mitra et
629 Mount, 1998; Brown et al., 1999; Domènech et al., 2016) or intersected by syn-tectonic
630 structures (e.g. Bellahsen et al., 2012 ; Bellanger et al., 2014; Branellec et al., 2016). The
631 tectonic history of the eastern Pyrenees as discussed in this paper thus shows the succession of
632 two periods of the contractional deformation that led to the full inversion of the rifted margins.
633 Other sections, further west, except for the Basque-Cantabrian basin with the Nappe des
634 Marbres (Ducoux et al. 2019; Lescoutre and Manatschal, 2020), show a narrower IMZ where
635 most units are pinched and vertical, thus obliterating the details of the shortening history. The
636 Agly section and the Nappe des Marbres sections are thus key to unravel the early stages of the
637 Pyrenean shortening with two stages, a first thin-skinned episode with decollement of the basin
638 above the Triassic evaporites and a second thick-skinned episode leading to the present-day
639 structure.

640 **Conclusions**

641 The eastern North Pyrenean Zone is thus an ideal locality to study the tectonic inversion and
642 integration of hyperextended rifted system (Internal Metamorphic Zone), associated with a pre-
643 rift decoupling layer, in an orogenic belt. By applying structural data and measured maximum
644 temperatures (T_{\max}) estimated with the Raman Spectroscopy of Carbonaceous Materials
645 (RSCM) method, we show that a pre-kinematic decoupling layer has strongly impacted the
646 architecture of the Pyrenean belt, as well as the disposition of its sedimentary basins.

647 Concerning the thermal pattern, the Agly massif corresponded to a basement block separating
648 two metamorphic areas, the Boucheville syncline to the south and the Bas-Agly syncline to the
649 north. The Bas-Agly and Boucheville synclines are both affected by the Cretaceous *HT/LP*
650 metamorphism with T_{\max} exceeding 550°C. Remnants of Mesozoic sediments observed on top
651 of the Agly massif is also affected by the *HT/LP* Pyrenean metamorphism with temperature
652 higher than 400°C but lower than in the Bas Agly syncline. A strong thermal contrast exists
653 between the Bas-Agly/Boucheville synclines and the St-Paul-de-Fenouillet syncline located to
654 the north. A thermal gap is also observed between Bas-Agly/Boucheville synclines and the
655 northeastern part of the Agly massif.

656 Structural observations confirm the T_{\max} distribution measured in the Agly massif and Bas-Agly
657 syncline. The northeast part of the Agly massif composed by Silurian to Devonian shales
658 exposed various brittle and brittle-ductile shears without no pervasive ductile deformation,
659 while the Mesozoic sediments of the Bas-Agly syncline show an intense ductile deformation
660 with metric-scale boudinage and recumbent metric folds.

661 To explain thermal and structural gaps and the similarities between the Boucheville and Bas-
662 Agly synclines, we propose that the Bas-Agly syncline may correspond to an allochthonous
663 unit thrust northward on top of the Agly massif and the St-Paul-de-Fenouillet syncline during
664 early stages of horizontal shortening (thin-skinned tectonics). The pre-orogenic Upper Triassic

665 evaporites decoupling layer and the low-angle normal faults inherited from the early Cretaceous
666 rifting allowed the transport of the Mesozoic cover above a north-verging thrust and the
667 substitution of the normal cover of the Agly massif by the metamorphic Bas-Agly syncline.
668 Consequently, the Boucheville and Bas-Agly basins were initially parts of a single basin during
669 the hyperextension rifting phase.

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1152 **Figures captions**

1153 **Figure 1:** Simplified structural map of the Pyrenean belt. Only main “alpine” thrusts and faults
1154 are represented: North-Pyrenean Frontal Thrust (NPFT), South-Pyrenean Frontal Thrust
1155 (SPFT), and North-Pyrenean Fault. The Internal Metamorphic Zone (IMZ, orange in the figure)
1156 is located in the North Pyrenean Zone (NPZ) and crops out along the strike of the belt. The
1157 study area is located in the eastern part of this belt.

1158 **Figure 2:** Geological map of the eastern part of the North Pyrenean Zone (localization on Fig.
1159 1) (after [Crochet *et al.*, 1989](#); [Berger *et al.*, 1997](#); [Bles and Berger, 1982](#); [Fonteilles *et al.*, 1993](#))
1160 with T_{\max} measured with Raman spectrometry ([Chelalou, 2015 \(in blue\)](#); [Clerc *et al.*, 2015 \(in](#)
1161 [black\)](#) and this study (in red)) and thermobarometry estimates ([Golberg and Leyreloup, 1990](#)).
1162 Localisation of Mesozoic sequences remnants on the Agly massif are pointed out by numbers:

1163 (1) Serre de Cors; (2) Serre de Verges; (3) Roc de Lansac; (4) Lake Caramany and (5) Agly
1164 dam.

1165 **Figure 3:** Isometamorphic map of the HT/LP metamorphism distribution in the eastern part of
1166 the North Pyrenean Zone.

1167 **Figure 4:** Structural map of the eastern part of the North Pyrenean Zone (localization on Fig.
1168 1) (after [Crochet *et al.*, 1989](#); [Berger *et al.*, 1997](#); [Bles and Berger, 1982](#); [Fonteilles *et al.*, 1993](#)).
1169 This map displays the different tectono-stratigraphic units and the detail of structural framework
1170 (trajectories of foliation and cleavage; lineations and folds). Early convergence-related fault are
1171 drawn in orange and late convergence-related fault in red.

1172 **Figure 5:** field photographs and interpretations of the observed deformation in the Paleozoic
1173 sediments of the Agly massif. a) top-to-the N normal shears. b) top-to-the S reverse shears.

1174 **Figure 6:** field photographs and interpretations of the observed deformation in Mesozoic Bas-
1175 Agly sedimentary cover. a) big picture on Rhaetian and Liassic sediments of the southern Bas-
1176 Agly, modified after [Clerc *et al.* \(2016\)](#). b) zoom on recumbent fold with top-to-the north
1177 verging. c) pure shear deformation and hectometer-scale boudinaged of Liassic sediments.

1178 **Figure 7:** Geological cross sections through the eastern part of the IMZ illustrating the
1179 relationships between the Bas-Agly syncline and surroundings units. Legend is similar to Figure
1180 2.

1181 **Figure 8:** Comparison of thermal structure for two endmembers: (a) a detachment system
1182 forming a MCC and (b) a thrust system. Structures are sole in the middle-lower continental
1183 crust and isotherms represent static maximum temperatures reached by rocks.

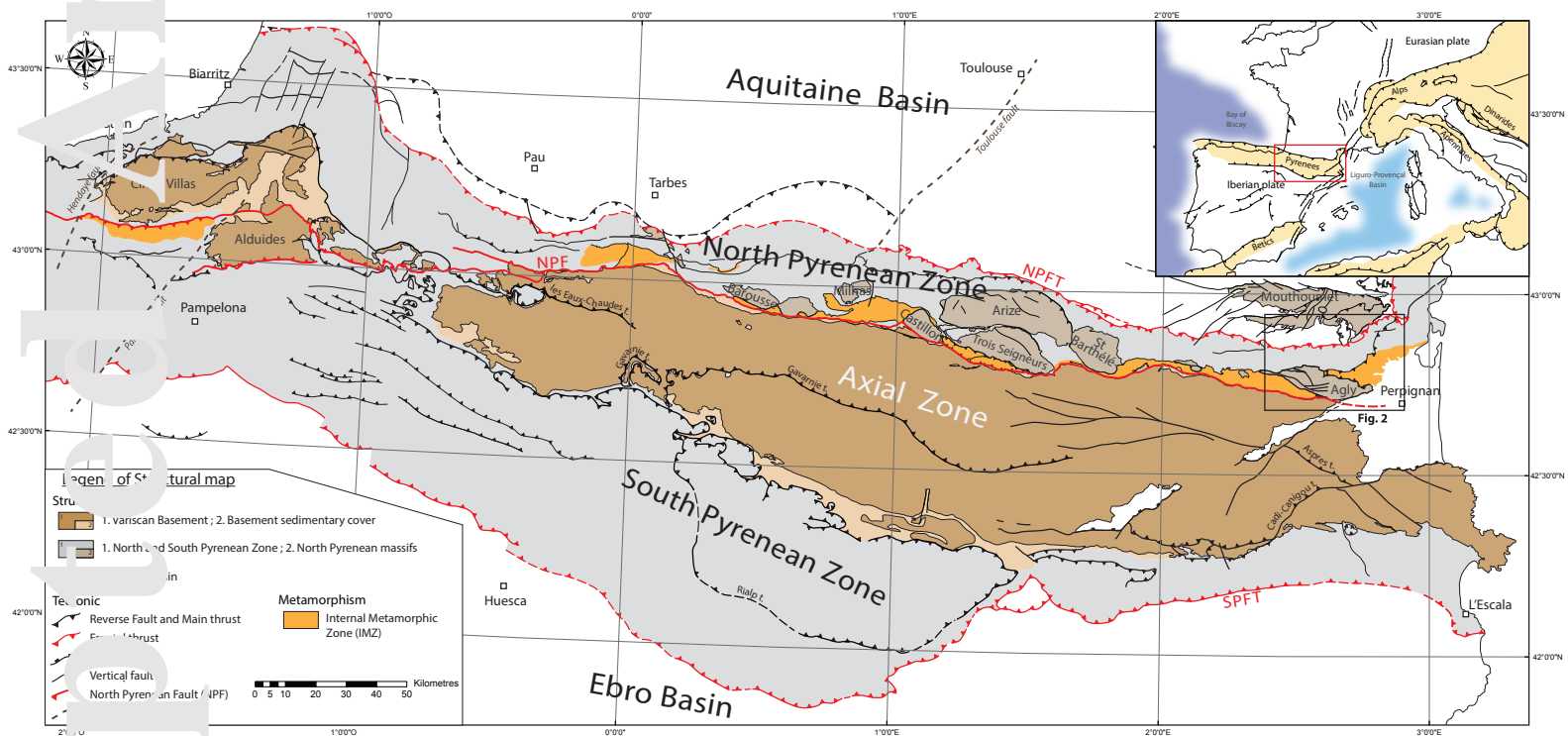
1184 **Figure 9:** Geodynamical model of the eastern part of the NPZ illustrating a structural evolution
1185 from the rifting event to the present, modified after [Ternois *et al.* \(2019\)](#). Four steps of

1186 restoration are depicted. SFPT: South Pyrenean Frontal Thrust; NPF: North Pyrenean Fault;

1187 LTFF: La-Tour-de-France Fault; TF: Tautavel Fault; NPFT: North Pyrenean Frontal Thrust.

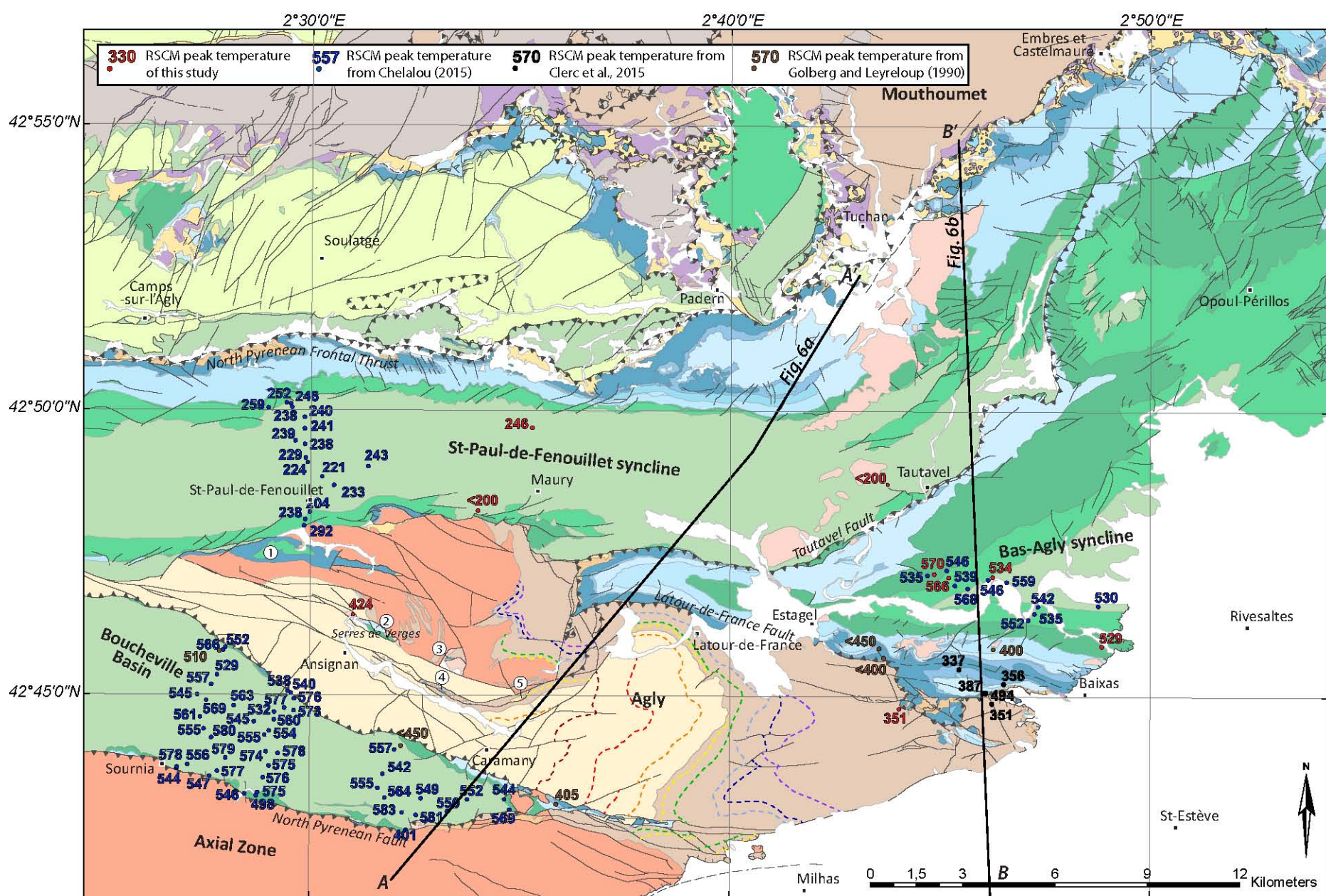
1188 See text for explanations.

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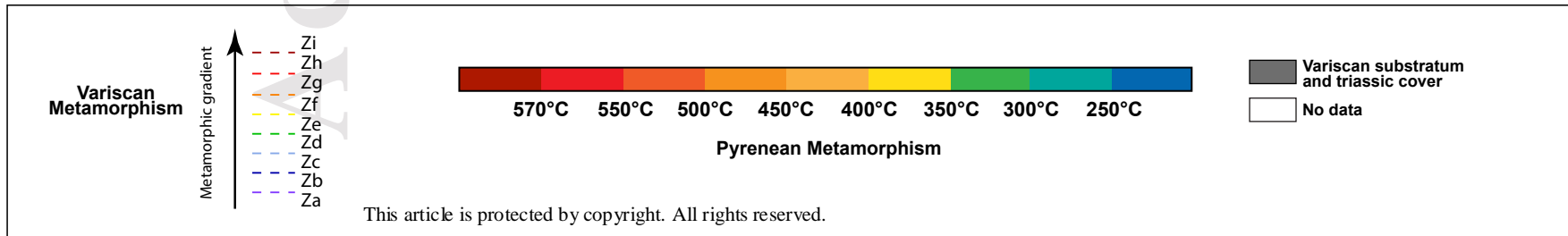
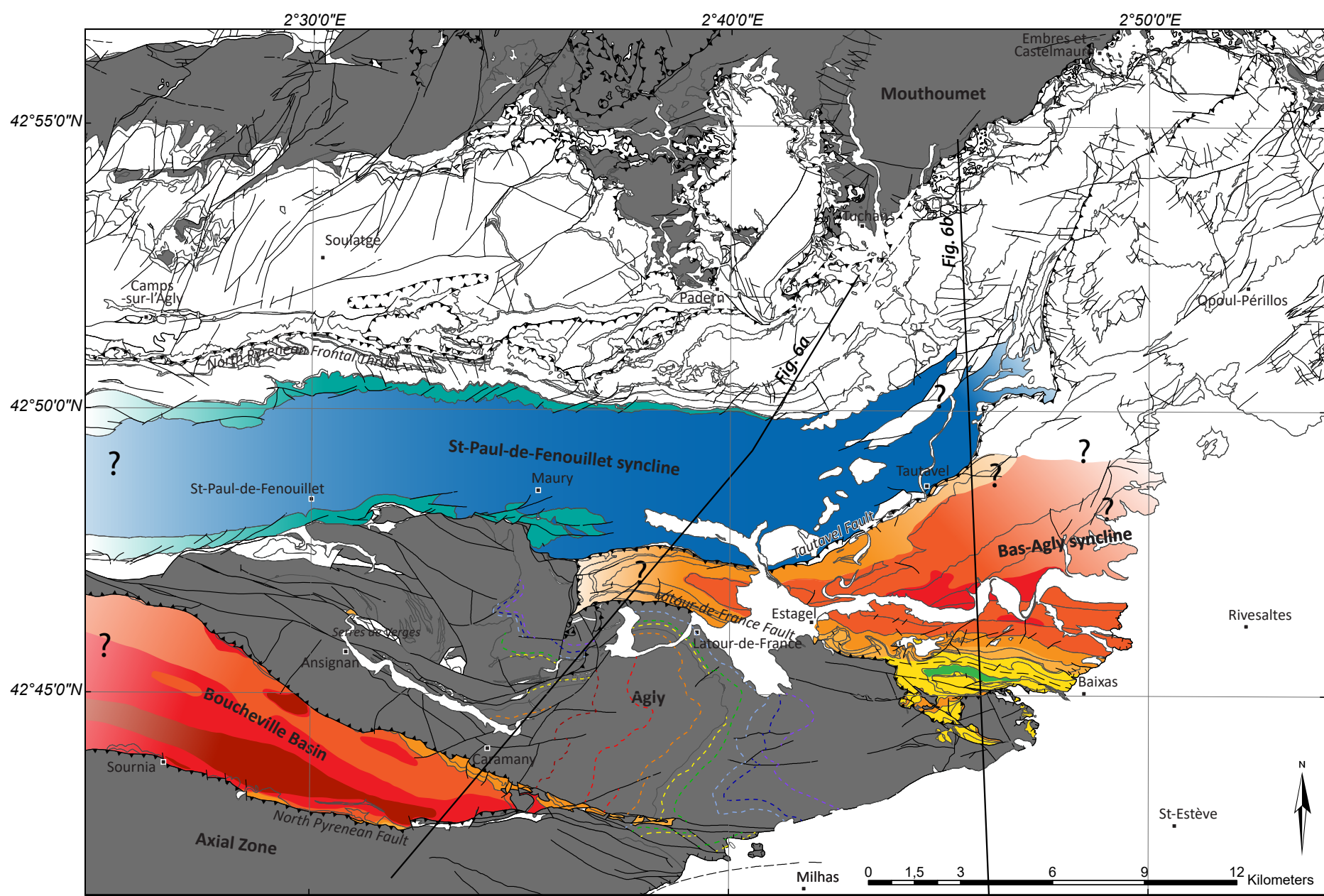
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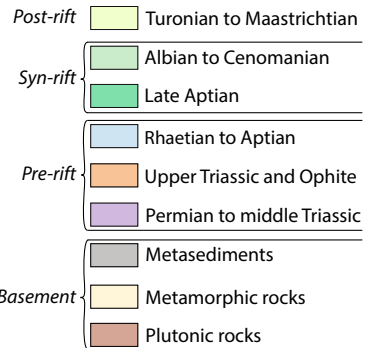
● 330 RSCM peak temperature of this study
 ● 557 RSCM peak temperature from Chelalou (2015)
 ● 570 RSCM peak temperature from Clerc et al., 2015
 ● 570 RSCM peak temperature from Golberg and Leyreloup (1990)

<ul style="list-style-type: none"> Quaternary Paleogen (day) Turonian to Maastrichtian (Sandstone, marls and limestone) Albian to lower Cenomanian (Marls) Aptian (Marls and limestones) 	<ul style="list-style-type: none"> Berriasian to Baremian (Limestones) Oxfordian to Tithonian (Limestones) Bajocian to Callovian (Limestones and dolomites) Bajocian to Tithonian (Limestones and dolomites) Pliensbachian to Aalenian (marls and limestones) 	<ul style="list-style-type: none"> Hettangian to Sinemurian (Limestones and dolomites) Rhaetian (Clays, limestones, sandstones, dolomites and marls) Upper Triassic (Marls, gypsum, sandstones and dolomites) Middle to Lower Triassic (Conglomerates, sandstones, limestones, marls and dolomites) ③ Remnants of Mesozoic sequences 	<ul style="list-style-type: none"> Carboniferous (Lydian stone, limestones, conglomerates) Ordovician to Devonian (Micaschists, shales, sandstone and limestone) Plutonic rocks (Granits) Agly paragneisses and orthogneisses 	<p>Variscan Metamorphism</p> <p>Metamorphic gradient ↑</p> <ul style="list-style-type: none"> Zi Zh Zg Zf Ze Zd Zc Zb Za
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Legend of structural map

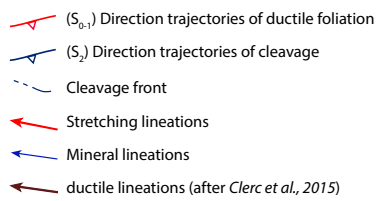
Tectono-stratigraphic Units



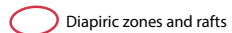
Details of structures

Pyrenean Structures

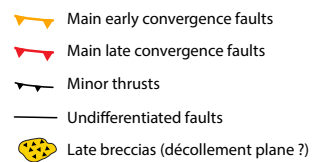
Structural data



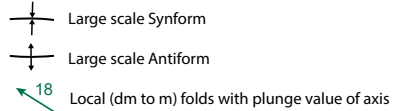
Salt tectonic-related structures



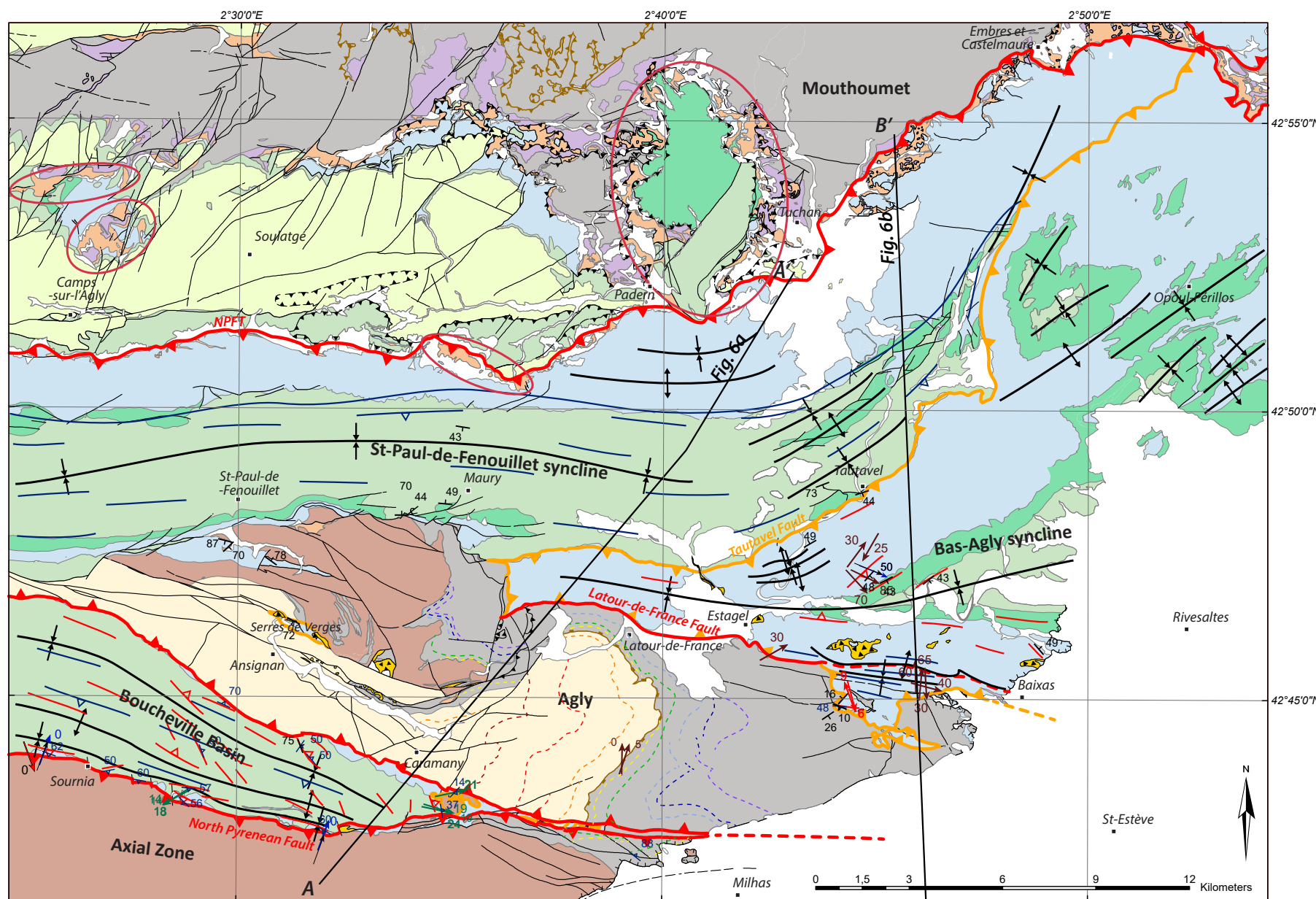
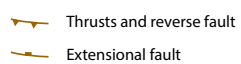
Faults

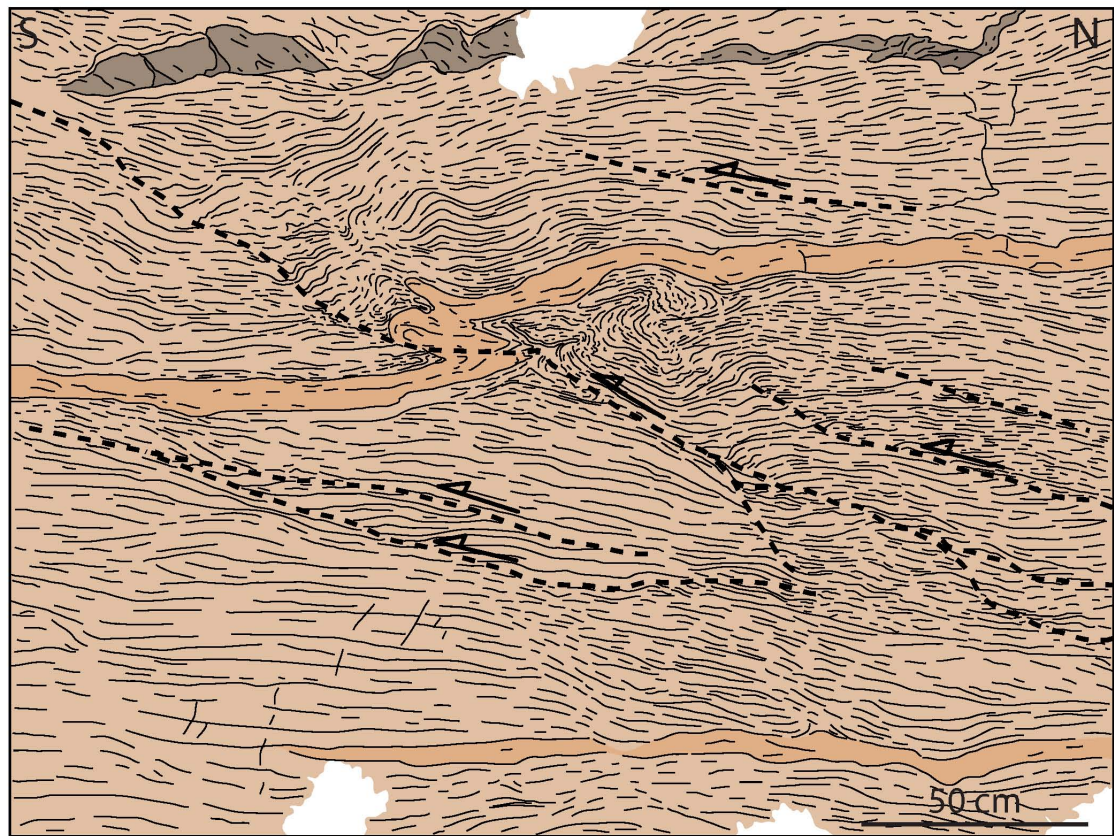
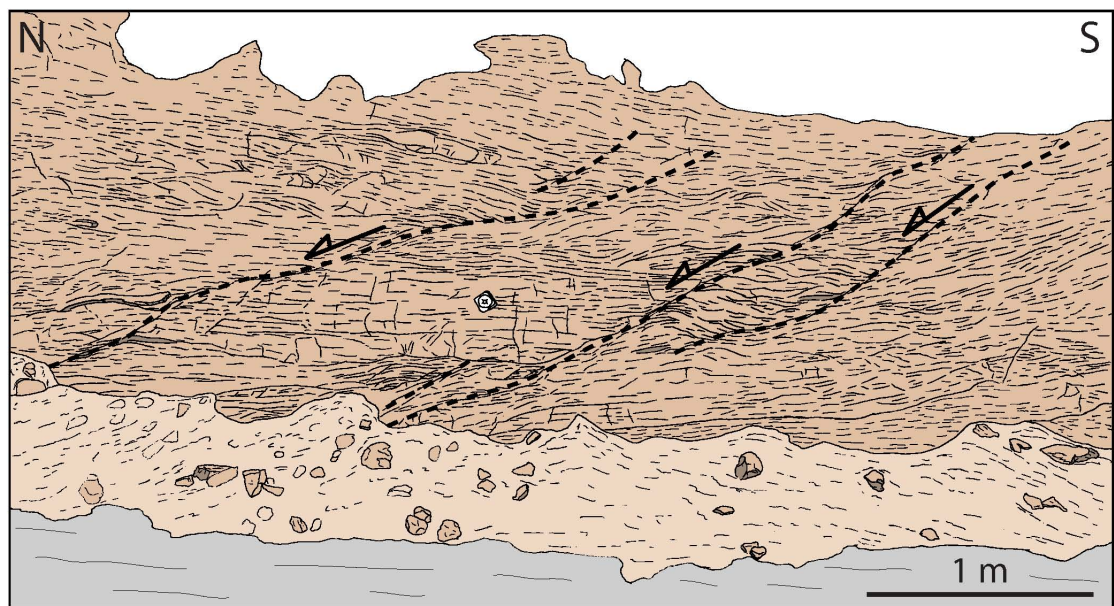


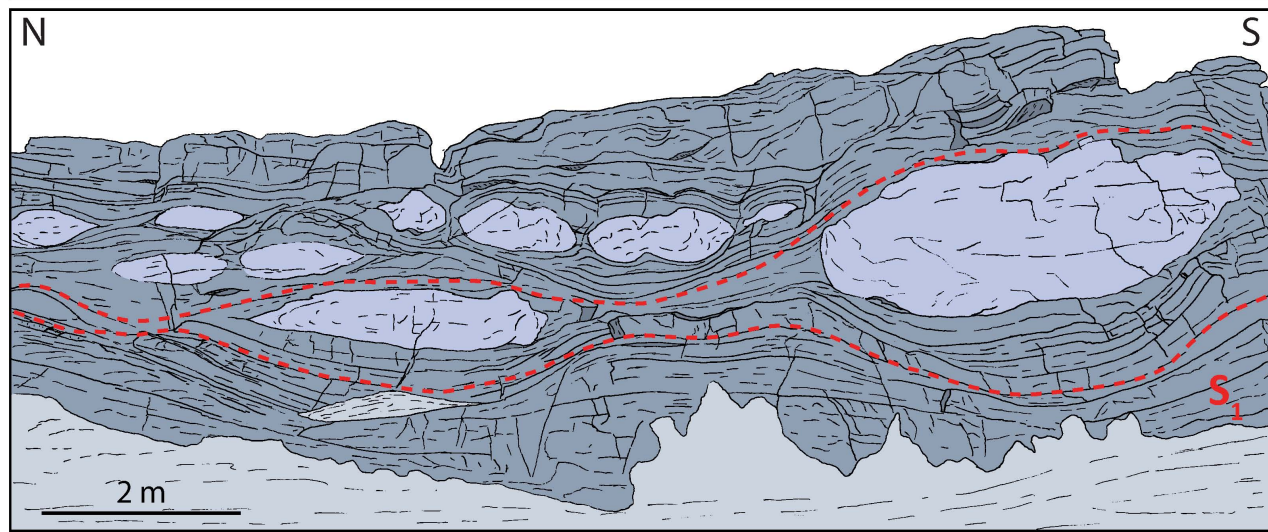
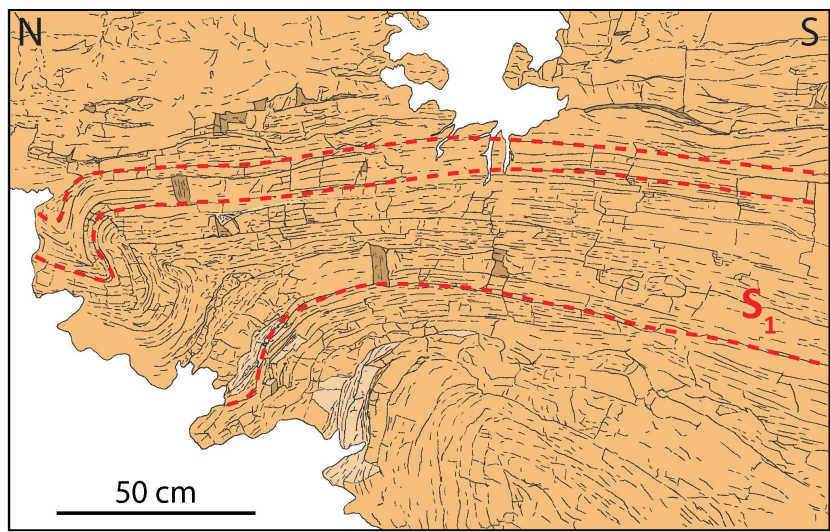
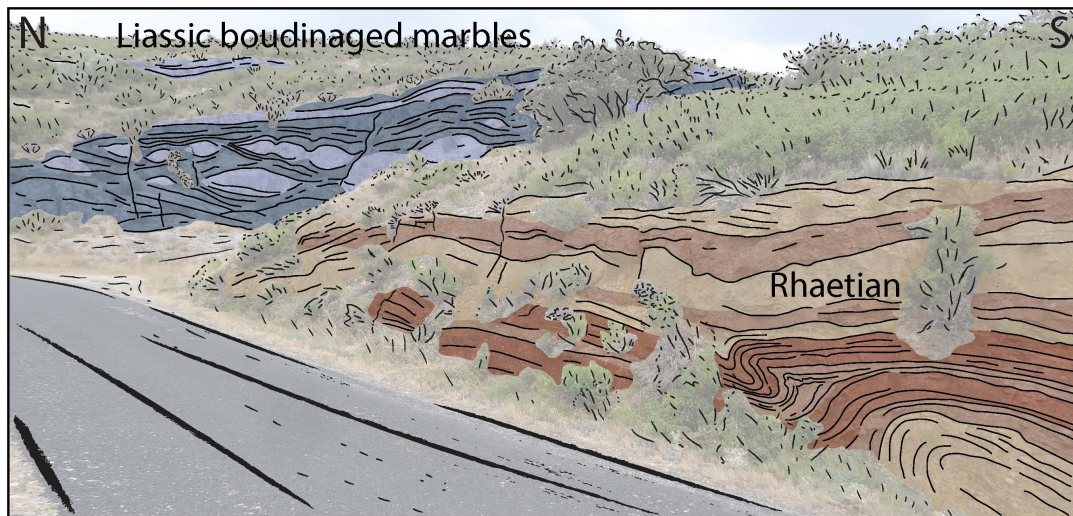
Folds



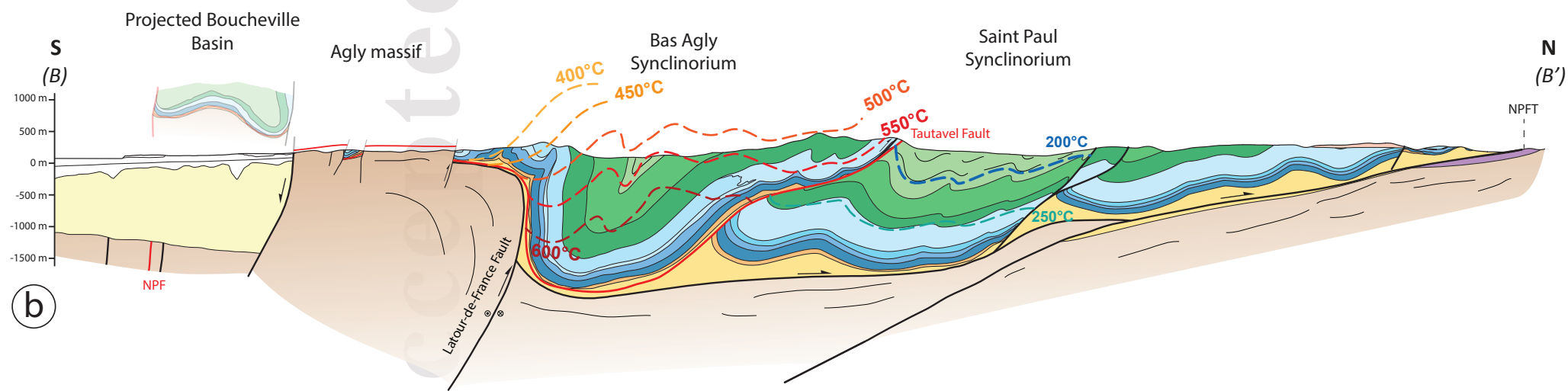
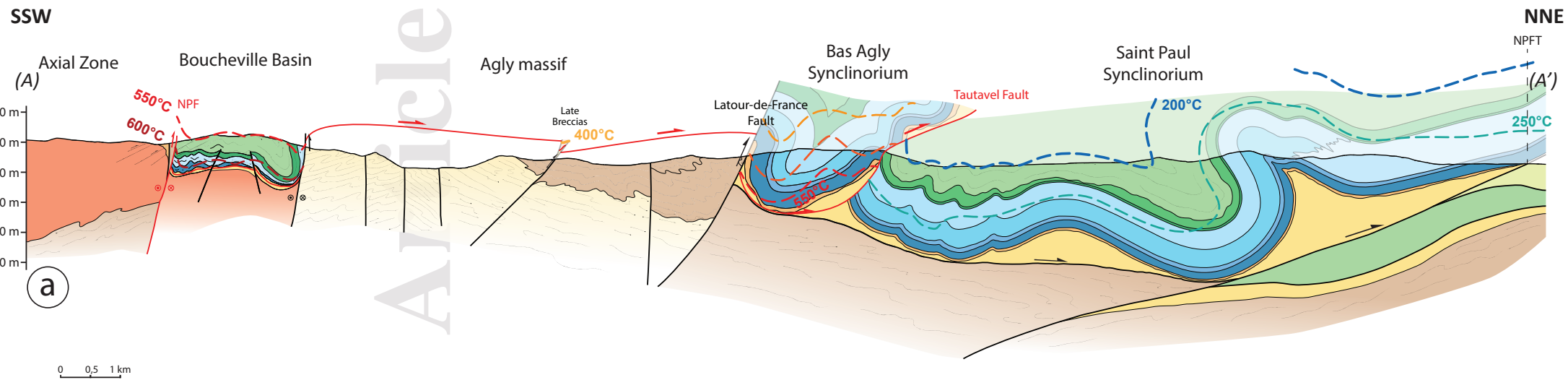
Variscan Structures



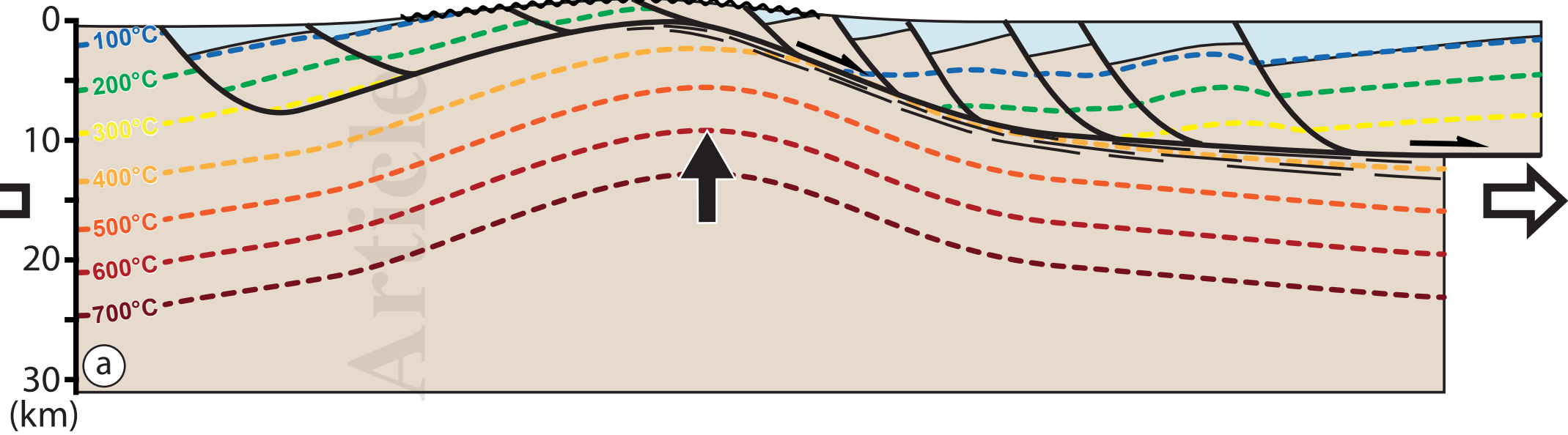




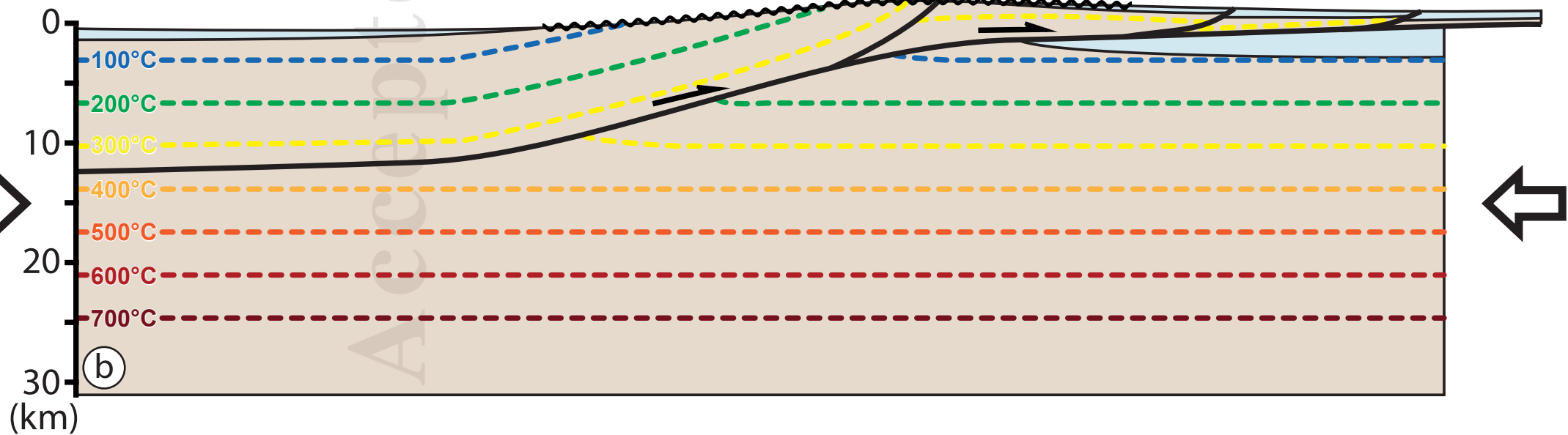
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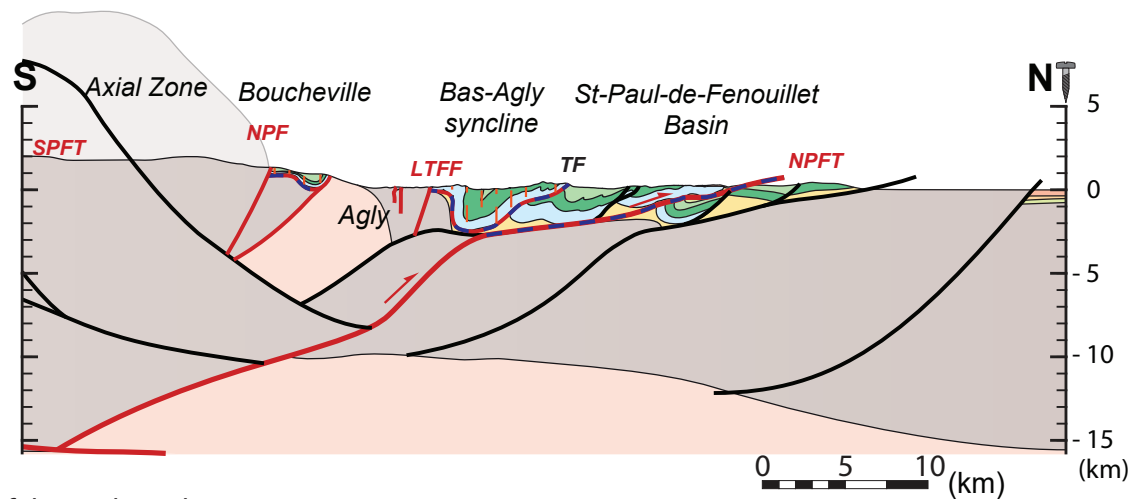
Typical thermal structure related to a Detachment system



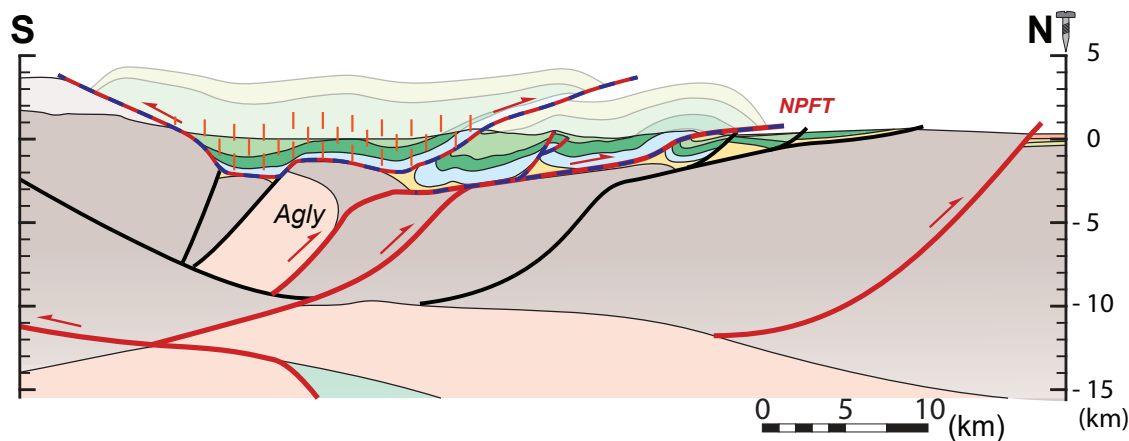
Typical thermal structure related to a Thrust system



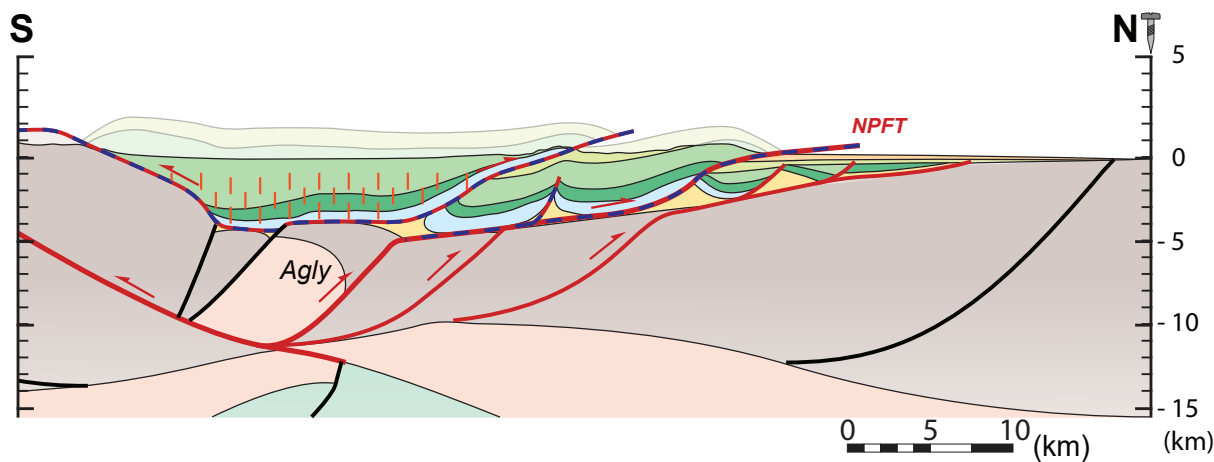
d) Late Eocene (35Ma) to Present-day - Orogenic crustal thickening



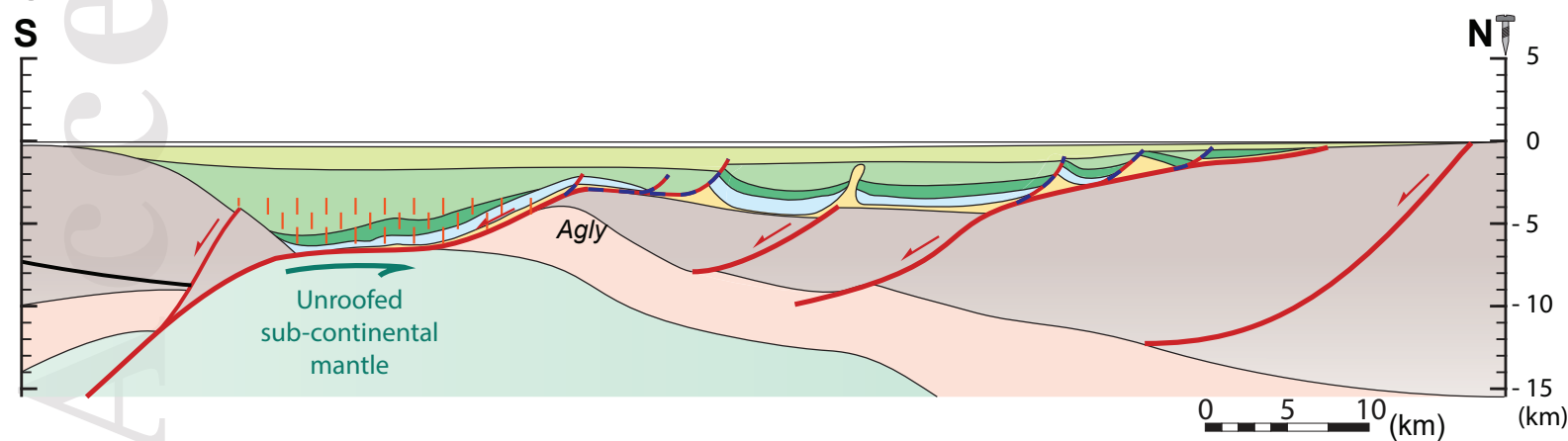
c) Bartonian (40Ma) - Reactivation of the necking domain



b) Late Maastrichtian (66 Ma) - Early convergence



a) Early Santonian (84 Ma) - Rifting templates and onset of the convergence



Lithostratigraphic units

Basement		Pre-rift	Post-rift
Upper crust	Jurassic	Late Cenomanian to late Coniacian	
Lower crust	Upper Triassic		
Lithospheric mantle	Late Albian to mid-Cenomanian	Syn-orogenic	
Astenospheric mantle	Early Aptian to early Albian	Maastrichtian	Eocene

Metamorphism

HT metamorphic rocks

Tectonic

- Abandoned faults
- Active faults
- Decollement and salt-assisted faults