Contribution of magmatism, partial melting buffering and localized crustal thinning on the late-variscan thermal structure of the Agly massif (French Pyrenees)

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Abstract

Low pressure-high temperature (LP-HT) metamorphism, with geothermal gradients in the order of 50 to 100°C/km, is a common feature of the late evolution of collisional orogens. These abnormal thermal conditions may be the results of complex interactions between magmatism, metamorphism and deformation. The Agly massif, in the French Pyrenees, preserves the metamorphic footprints of the late Variscan thermal structure of an almost continuous section from the upper and middle continental crust. The upper crust is characterized by a very high geothermal gradient of ~55 °C/km, evolving from greenschist to amphibolite facies, while the middle crust, exposed in a gneissic core, exhibits granulite facies conditions with a near isothermal geothermal gradient (<8°C/km) between 740 and 790°C.

The abnormal and discontinuous crustal geothermal gradient, dated at *c.* 305 Ma on syn-granulitic monazite by LA-ICPMS, is interpreted to be the result of magmatic intrusions at different structural levels in the crust: the Ansignan charnockite (*c.* 305 Ma) in the deepest part of the gneissic core, the Tournefort granodiorite (*c.* 308 Ma) at the interface between the gneissic core and the upper crust and the Saint-Arnac granite (*c.* 304 Ma) in the upper section of the massif. The heat input from these magmas combined with the thermal buffering effect of the biotite dehydration-melting reaction resulted in the near-isothermal geothermal gradient in the gneissic core (melt-enhanced geotherm). The higher geothermal gradient (> 50°C/km) in the upper crust is only due to conduction between the hot middle crust and the Earth's surface.

The estimated maximum finite pressure range suggests that about 10 to 12 km of crust are exposed in the Agly massif while the present day thickness does not exceed 5 to 6 km. This pressure / depth gap is consistent with the presence of several normal mylonitic shear zones that could have contributed to the subtraction of about 5 km of the rock pile. Monazite U-Th-Pb ages carried out on monazite overgrowths from a highly mylonitized sample suggest that this vertical thinning of the massif occurred at *c.* 296-300 Ma.

This later Variscan extension might have slightly perturbed the 305 Ma geothermal gradient, resulting in an apparent higher conductive geothermal gradient in the upper crust. Although the Agly massif has been affected by Cretaceous extension and Eocene Alpine compression, we suggest that most of the present-day thickness of the column rock was acquired by the end of the Paleozoic.

1 INTRODUCTION

Low pressure-high temperature metamorphism (LP-HT) in the continental crust is defined by high geothermal gradient (60-150°C/km) and temperature reaching about 600-800 °C, at pressure below 0.4 GPa. These conditions are usually reached in contact metamorphism settings, resulting in hornfels and possible partial melting of pelitic rocks close to the contact with the intrusive rocks. Nevertheless, numerous examples for LP-HT metamorphic terranes have been ascribed to regional metamorphic environments (De Yoreo, Lux, Guidotti, Decker, & Osberg, 1989; Kim & Cho, 2003 Gibson, 1991; Okay et al., 2013; Yardley, Barber, & Grey, 1987). Such high geothermal gradients, much higher than steady state gradients for thickened crust, require a considerable heat flux to reach such thermal state in the continental crust and are more common in Archean cratons (Brown, 2007; 2009). They have been commonly ascribed to crustal underplating of mantle-derived basic magma (Thompson & England, 1984), intrusion of large felsic magma in the middle and upper crust (De Yoreo et al., 1989; Lux, De Yoreo, Guidotti, & Decker, 1986) or Metamorphic Core Complexes (Coney, 1980).

These regional LP-HT metamorphic gradients are commonly reported in gneissic domes and often associated with magmatism and partial melting (De Yoreo et al., 1989; Kim & Cho, 2003). Over the past decades, numerous studies have highlighted the important role of partial melting, thermal buffering and advective heat transfer due to melt migration towards the structurally upper levels to produce this particular thermal structure of the crust (Depine, Andronicos, & Phipps-Morgan, 2008; Riel, Mercier, & Weinberg, 2016). One of the key feature of these studies is that the combined effect of radioactive heat production, high mantle heat source, heat diffusion, latent heat of melting and heat advection by melt transfer result in discontinuous thermal structure of the continental crust. The upper crust is characterized by high to very high LP-HT geothermal gradient. In contrast the middle to lower crust is partially molten and characterized by a low to very low geothermal gradient. This thermal structure has been modeled and defined by Depine, Andronicos, & Phipps-Morgan (2008) as a melt-enhanced geotherm. This discontinuous crustal thermal structure has been described in various crustal settings including collisional setting, in the French Massif Central (Trap, Roger, Cenki-Tok, & Paquette, 2016) and Himalaya (Groppo, Rolfo, & Indares, 2012), and continental arc settings such as the Andean active margin in Ecuador (Riel et al., 2016).

Deciphering the thermal structure of the crust at a specific geological time (i.e. the geothermal gradient) relies on the estimation of pressure and temperature conditions as well as dating peak metamorphism for different rocks along a metamorphic field gradient-related cross-section. Unfortunately, this task is not straightforward due to potential kinetics effect for amphibolite facies rocks that makes part of the upper crust. Indeed, estimated peak temperature from equilibrium thermodynamic modeling of phase relation may not reflect the “real” temperature because of kinetic factors that delay the onset of nucleation and growth of metamorphic phases (Carlson, Pattison, & Caddick, 2015; Pattison & Tinkham, 2009; Waters & Lovegrove, 2002). Additionally, field metamorphic gradient may not be only the result of heat transport but could also be modified by deformation processes. Heterogeneous deformation leads to low strain domains preserving relict and metastable assemblages partially re-equilibrated while high strain domains are fully re-equilibrated. Localized crustal shear zones can also modify metamorphic gradient by juxtaposing terrains of different grades, disrupting the observed metamorphic gradient and leading to transient geotherms.

LP-HT metamorphism is a common feature in the Variscan belt, especially in Pyrenean massifs: St Barthélémy (Saint Blanquat, Lardeaux, & Brunel, 1990; Saint Blanquat, 1993), Canigou (Gibson, 1991), Aston-Hospitalet (Barnolas & Chiron, 1996; Denèle, Olivier, Gleizes, & Barbey, 2007). This LP-HT metamorphism is usually associated with the development of gneissic domes and a widespread late Variscan magmatism between 300-320 Ma (Cochelin et al. 2017, ref therein). Different geodynamic models have been proposed that might be responsible for the LP-HT gradients: (1) early extension with thinning during continental rifting associated with mantle-derived magmas (Wickham & Oxburgh, 1985; Wickham, Oxburgh, Reading, & Vissers, 1987), (2) late extension after crustal thickening (Van den Eeckhout & Zwart, 1988) or (3) compressional setting with thickening associated with numerous intrusions in the crust and partial melting (De Yoreo et al., 1989; Lux et al., 1986).

In the easternmost part of the French Pyrenees, the Agly massif is one of these gneiss domes that preserves a complex HT-LP late Variscan tectono-metamorphic evolution (figure 1a). However, its present-day geometry is not only due to Variscan tectonics but also results from the superposition of late Cretaceous extension and Eocene Alpine compression (Clerc & Lagabrielle, 2014; Vauchez et al., 2013). The late Variscan metamorphism event is characterized by a high geothermal gradient in the upper crust of about 80 to 100°C/km (Fonteilles, 1976; Guitard, Vielzeuf, & Martinez, 1996; Vielzeuf 1984) and up to 150°C/km is some localities (Delay, 1989), partial melting in the middle crust and different generations of magmatic intrusions (Delay, 1990; Fonteilles, 1976; Guitard et al., 1996). A detailed review of the bibliography reveals that quantitative thermobarometry has never been published on the metamorphic gradient of the whole Neoproterozoic-Paleozoic sequence. Furthermore, the available P-T estimations are restricted to the gneissic core, and mostly to the deepest part near the Ansignan charnockite.

This paper is aimed at defining quantitatively the P-T conditions along the metamorphic gradient, the age and thermal structure of the Agly massif during the Late-Variscan orogeny. A suite of samples located at different structural levels were collected along an E-W cross section to constrain peak P-T conditions via phase relation forward modeling. Two distinct lithologies have been used: Ordovician pelites from the upper crustal level and Proterozoic pelites (kinzigites) from the genissic core, in order to minimize uncertainties on thermobarometric estimates and highlight possible metamorphic discontinuities along the metamorphic gradient. In-situ LA-ICPMS U-Th-Pb monazite dating has been used to constrain the timing of peak metamorphism and compared to previously documented magmatic ages. Finally, we will discuss and propose a conceptual model taking into account the role of partial melting, heat advection during magmatism and localized deformation responsible for the finite shape of the HT-LP discontinuous Variscan geotherm.

2 GEOLOGICAL SETTING

2.1 The Agly massif in the Pyrenees

The Pyrenees mountain belt was formed during the Tertiary collision of the Iberian microplate and the Eurasian plate (e.g. Choukroune, 1992). The Axial and Northern Pyrenean zones are characterized by a suite of exhumed Variscan massifs (e.g. Aston, Hospitalet, Canigou, Trois-Seigneurs, Agly), that consist of metasediments, para- and orthogneisses, migmatites and plutonic rocks, thrusted and juxtaposed to Mesozoic and Cenozoic covers.

The Agly massif is the easternmost Variscan massif of the Northern Pyrenean Zone in France (Figure 1a). The southern boundary of the massif is an alpine NW-SE-trending and vertical fault

(Bélesta-Trilla fault) that brings into contact Variscan magmatic and metamorphic rocks with the Boucheville syncline, consisting of Mesozoic sedimentary rocks (up to Albian) metamorphosed at high temperature during Cretaceous time (~95 Ma) (Delay, 1990 and references therein). On its northwestern edge, the Variscan Agly massif is locally thrusted northward over the Mesozoic cover (Triassic formation) of the Saint-Paul de Fenouillet syncline (Figure 1). In contrast, the same contact towards the northeast has been recently interpreted as a detachment fault that juxtaposed the Agly massif with the Mesozoic Agly syncline that has been metamorphosed during Cretaceous time (Vauchez et al., 2013). Although it is well acknowledged that both northern contacts have been tectonically reactivated during the Tertiary orogeny, there is a controversy about the age of this extensional tectonic history that could be either Paleozoic or Cretaceous (Vauchez et al., 2013; Olivier, 2013). This controversy has been rejuvenated by two recent studies using (U-Th)/He (Ternois et al., 2019) and U-Pb (Odlum & Stockli, 2019) in both zircon and apatite. The first study concluded that the Agly massif acquired most of its present-day geometry during Paleozoic times while the second argued that most of the thinning occurred during pre-Alpine hyper-extension.

Distinct geodynamic models have been proposed to account for these structural, metamorphic and magmatic features. Bouhallier, Choukroune, and Ballèvre (1991) proposed a late Variscan extensional setting after crustal thickening, while Paquet and Mansy (1991) rather suggested that crustal thinning may have occurred during Cretaceous time. On the other hand, Delay (1990) argued for a late Variscan compressional setting with heat provided by a charnockitic intrusion in the lower part of the massif. More recently, a bulk compression tectonics associated with a transpressive setting has been proposed to explain the structural evolution of the Agly massif (Olivier, Gleizes, & Paquette, 2004; Olivier, Gleizes, Paquette, & Muñoz Saez, 2008). It appears from this review that there is no consensus on the geodynamic model, and this situation may have arisen partly because the geothermal gradient (thermal structure) and the age of metamorphic and deformation events have not been well constrained.

2.2 Lithological units

Since the pioneer work of Fonteilles and Guitard (1964) and Fonteilles (1970; 1976), the Variscan Agly massif has been subdivided into two main units: (1) A metamorphosed Paleozoic sequence, mostly exposed in the northeastern part of the massif with a thickness of ~2.5 to 3 km, that structurally overlies (2) a high grade gneissic core of presumed late Proterozoic age in the southwestern part of this area. Both units are intruded by Variscan to late Variscan plutonic bodies (Ansignan charnockite, Tournefort diorite and the Saint-Arnac granite). The two main units were classically interpreted as a sedimentary cover on top of an older basement, respectively (Fonteilles & Guitard, 1964), but this interpretation is revised in this contribution.

The upper part of the Paleozoic sequence consists of Devonian limestones with minor and discontinuous Silurian black phyllites and locally Late Ordovician (Caradoc) volcano-clastic rocks. The lower part of the sequence is a monotonous and thick succession composed of an alternation of siltstone and sandstone layers (Jujols group of Laumonier (1998)). Recent biostratigraphic age data obtained on comparable rocks, on the southern slope of the Canigou massif (in the Pyrenean Axial Zone, south of the Agly massif), suggest that the Jujols group may be of Late Cambrian - Early Ordovician age and that most of the Early to Middle Ordovician sequence was eroded prior to the deposition of the Late Ordovician volcano-clastic rocks during the Sarde tectonic event (Casas & Palacios, 2012). The entire Paleozoic sequence has been affected by a low pressure-high temperature (LP-HT) metamorphism that increases downward from lower greenschist facies conditions at the top, leading to the transformation of Devonian and Silurian rocks into marbles and calc-schists, respectively, to upper amphibolite facies conditions with partial melting involved, at the bottom of the Jujols group (Fonteilles, 1970).

The "high grade gneissic core" is heterogeneous and consists of partially-melted metapelites (locally garnet-cordierite-sillimanite-bearing gneiss) and metagreywackes interbedded with discontinuous and highly deformed layers of calc-silicate gneisses and marbles. These meta-sedimentary are inter-layered with orthogneisses of variable thickness (on the order of meters) which were emplaced as granitic sills during Early Cambrian at *c.* 530-540 Ma (Tournaire Guille, Olivier, Paquette, Bosse, & Guillaume, 2019). In the western part of the massif, the uppermost part of the gneissic core is intruded by the kilometer thick Riverolles orthogneiss. The "gneissic core" is separated from the Paleozoic Jujols group by a discontinuous intercalation of white metacarbonate (the so-called "base marble"), which in some cases is present few meters inside the gneisses. All these observations suggest that the metasediments of the "gneissic core" of the Agly massif are a highly metamorphosed equivalent to the Late Neoproterozoic sediments of the Canaveilles group, observed for instance south of the Agly massif, in the Canigou dome (e.g. Cocherie et al., 2005; Laumonier, 1998). The Late Neoproterozoic age of the sediments was first suggested by Autran, Fonteilles, & Guitard (1966) and Vitrac-Michard and Allegre (1975), but they argued for a cover and pre-Variscan basement relationships (see review in Padel, 2016).

The Paleozoic sequence (Jujols group) and the gneissic core (Canaveilles group) were intruded by late Variscan plutons. In the southwestern part of the massif, the deepest part of the gneissic core was intruded by the Ansignan charnockite. This intrusion is a 500 to 800 m thick laccolith parallel to slightly discordant to the main foliation and consists of an orthopyroxene-bearing granodiorite with megacrysts of K-feldspar defining an internal magmatic foliation. The age of magmatic emplacement of this intrusion was initially constrained at 314 ± 6 Ma by Respaut and Lancelot (1983) on monazite grains, and more recently re-evaluated at 307 ± 3 Ma (Tournaire Guille et al., 2019) on zircons. The upper part of the gneissic core was intruded by the meter-thick Cassagnes sills at the same time (308 ± 3 Ma, Tournaire Guille et al., 2019), and may be equivalent to the Ansignan intrusion except that it is orthopyroxene-free and was emplaced in an upper structural level. In the northwestern part of the Agly massif, the Paleozoic sequence was intruded by the late Variscan Tournefort Diorite and the Saint-Arnac granite at 308 ± 1 Ma and 304 ± 5 Ma, respectively (Olivier et al., 2008). While the Tournefort diorite was emplaced at the base of the Jujols Group sequence, the Saint-Arnac granite intruded the entire Jujols group. Emplacement of the late Variscan Tournefort diorite and Saint-Arnac granite induced a local contact metamorphism with the development of partial melting around the diorite and late muscovite crystallization around the granite contact and in the metasediment septas present in the granite.

2.3 Tectono-metamorphic evolution

The Paleozoic sequence shows a progressive downward increase of metamorphic conditions, with the following succession of isograds: chlorite out, biotite in, cordierite in, andalousite in, sillimanite in, muscovite out and finally water saturated partial melting. Based on semi-quantitative phase diagram projections and the estimation of the thickness of the Paleozoic sequence, the late-Variscan geothermal gradient has been estimated between 130 and 150 $°C \cdot km^{-1}$ (Fonteilles, 1970) or at 80 $°C \cdot km^{-1}$ (Delay, 1990).

The gneissic core recorded a more homogenous upper amphibolite to granulite facies conditions of metamorphism. Nevertheless, the gneissic core is classically divided into two structural levels based on their metamorphic assemblages (Fonteilles, 1970): the Belesta gneisses forming the upper part with K-feldspar-sillimanite-bearing assemblages, and the Caramany gneisses in the deepest part with orthopyroxene-bearing assemblages. Metamorphic conditions estimated on the Ansignan charnockite or nearby lithologies have been estimated at 0.6 ± 0.1 GPa and 750 ± 50 °C (Andrieux, 1982), 0.5 ± 0.05 GPa and 800 ± 100 °C (Vielzeuf, 1984) or 0.4 ± 0.05 GPa and 600-650 °C (Delay, 1990).

2.4 Deformation

The metamorphic rocks of the upper Paleozoic sequence and the lower gneissic core share a predominant strain fabric observed in the entire Agly massif that corresponds to a moderately to shallowly dipping HT/LP foliation, holding a conspicuous N20 trending mineral and stretching lineation (e.g. Bouhallier et al., 1991; Olivier et al., 2004). Similarly, the Ansignan charnockite and the Cassagnes intrusives show a well-developed magmatic foliation, with a minor component of solid-state deformation. This magmatic foliation is parallel to the main LP/HT foliation of the surrounding gneisses (Delay, 1989; Olivier et al., 2004).

In addition to the main foliation, the gneissic core is characterized by the development of low angle extensional mylonitic shear zones, mainly visible in the eastern part of the massif, dipping either toward the north or toward the south and associated with a strong stretching lineation that trends N20 (Delay, 1989). The main foliation and the mylonitic shear zones show a similar normal and top-to-the north kinematics attributed to a same tectono-metamorphic event responsible for the thinning of the Variscan crust (Vanardois et al. in press). Olivier et al. (2004; 2008) suggest a local thinning during a late Variscan transpressive regime with the mylonitic normal shear zones that developed in response to late doming. In contrast, Bouhallier et al. (1991) documented a major detachment zone located between the Paleozoic sequence and the gneissic core (the Caladroy-Latour shear zone) that would be responsible for the subtraction of about 10 km of crustal material and consequently the juxtaposition of the low pressure-medium temperature Paleozoic sequence with the moderate pressure-high temperature gneissic core. The age of mylonitic shearing (either in a context of crustal thinning or transpression) is still a matter of debate. For Delay (1990) and Bouhallier et al. (1991), the extension is contemporaneous with the HT-LP metamorphism and therefore would be late Variscan (*c.* 315 Ma). On the contrary, Olivier et al. (2008) argued that the root of the ~304 Ma Saint-Arnac and Tournefort plutonic body underwent northward normal shearing suggesting that the development of mylonitic shear zones took place just after 304 Ma. In contrast, Paquet & Mansy (1991) suggested that the thinning responsible for the loss of ~ 10 km of crustal material occurred during Mid-Cretaceous time (110-100 Ma).

The E-W trending fault that bounds the southern limb of the Mesozoic Agly syncline (north-eastern part of the Agly massif) has recently been interpreted as an ancient Mid-Cretaceous detachment fault (~100 Ma), with a top-to-the-north shearing sense (Vauchez et al., 2013). This contact may have contributed to the exhumation of the Agly massif during the Iberian microplate rotation relative to Eurasia and opening of the Bay of Biscay during or after Albian (Vauchez et al., 2013). Vauchez et al. (2013) also suggested that some of the extensional mylonitic shear zones observed in the Variscan basement, and described above, may also be related to this Mid-Cretaceous extensional event. Furthermore, this Mid-Cretaceous deformation is coeval with an intense episode of fluid circulation and fluid-rock interactions that are responsible for the local albitization of the Saint-Arnac granite, the paragneisses of the gneissic core and, locally, the Paleozoic sequence (Poujol, Boulvais, & Košler, 2010).

Most of the tectonic contacts bounding the Agly massif to the north and the south have been reworked during the Tertiary orogeny corresponding to the N-S convergence between Iberia and Eurasia during Eocene time. For instance, the reactivation of the western part of the upper Cretaceous detachment, described above, corresponds to the northward thrusting of the Agly massif over the Mesozoic cover of the Saint-Paul de Fenouillet syncline (Figure 1). Alpine deformation is also visible within the Agly massif where Mesozoic carbonate cover and Early Eocene Breccias are pinched into the high-grade rocks between steeply dipping N110 trending Alpine faults (Figure 1b).

3 METAMORPHIC GRADIENT ESTIMATION

As introduced in the previous section, the N20 pervasive foliation of the main metamorphic event (LP/HT) defines a single tectono-metamorphic event throughout the entire massif that can be linked to the Charnockite emplacement at 305 Ma. This lineation is similar to the stretching lineation of all mylonitic shear zones. Thus, sampling along the metamorphic gradient should be in a N-S direction in order to highlight the possible role of mylonitic shear zones on the finite metamorphic gradient shape. However, the occurrence of major East-West trending post-Mesozoic steep faults and Pyrenean thrusts that may have juxtaposed blocks of different structural level precludes a simple north-south sampling. Furthermore, in the central part of the Agly massif, where the deepest part located around the Ansignan charnockite is exposed, the metamorphic gradient in the Paleozoic sequence is obliterated by the St Arnac granitic intrusion. For these reasons, we have chosen to sample the different structural levels along an E-W cross section (Figure 2), from Força Real on the Eastern side of the massif, towards the lowest structural levels near the Ansignan charnockite on the western side. The location of this cross section is represented on the map (Figure 1b). The GPS coordinates for each sample can be found in Table 1.

3.1 Petrology

3.1.1 Metamorphic evolution of the Late Cambrian - Early Ordovician sequence (Jujol group)

The Agly massif preserves a typical LP-HT metamorphic gradient in the cover as suggested in previous studies (Fonteilles, 1970; Guitard et al., 1996; Vielzeuf, 1984) and referred to here as the upper Paleozoic sequence (Jujol group). The sequence of mineral isograds is biotite-in, closely followed by chlorite-out, cordierite-in, andalusite-in, sillimanite-in, muscovite-out/K-feldspar-in and onset of partial melting. We describe here the different reactions that could lead to the appearance and disappearance of these index minerals based on the petrography and available petrogenetic grids (Pattison & Tracy, 1991; Spear & Cheney, 1989). For clarity, all reactions are represented in the simplified KFMASH system. Consequently, the role of plagioclase and other Ca-bearing phases is not taken into account to describe the main phase transitions.

The first observable isograd of the LP-HT metamorphic gradient is the biotite-in. This reaction results in the appearance of small biotite flakes in the fine-grained chlorite-muscovite bearing schists. Biotite content continuously increases with increasing metamorphic grade. The low-grade chlorite-muscovite schists preserve evidences of a bedding that corresponds to an alternation of millimeter to centimeter thick phyllosilicates-rich and quartz-rich layers (Figure 3a).

The very fine-grained nature of the chlorite schists makes difficult the petrographic observation, and although we have not observed K-feldspar, we suggest that the first appearance of biotite is the result of the simplified reaction:

$$\mathrm{Chl + Kfs = Ms + Bt + Qtz + H_2O} \qquad (1)$$

This reaction is consistent with the observation of micro-lithons bearing chlorite and surrounding preferred orientation of muscovite and biotite parallel to the HT/LP foliation (caption Figure 3a, see S cleavage).

In the field, the continuous increase of biotite content results in a striking change in color and grain size from a greenish and very fine-grained chlorite-muscovite schist to a darker and coarser biotite-muscovite schist (Figure 3b). The darker color is mostly due to the presence of graphite derived from organic matter, present until partial melting.

The next isograd corresponds to the appearance of cordierite porphyroblasts together with the complete disappearance of primary chlorite (Figure 3c). This isograd is immediately followed by the appearance of andalusite in the stability field of cordierite. Preferred orientation of cordierite and andalusite porphyroblasts underline the orientation of the main N20 stretching lineation. In high-aluminum pelites, several chlorite dehydration reactions can be responsible for the crystallization of cordierite at low pressure. All of them involve muscovite, chlorite and quartz as reactant phases and biotite and H_2O as products:

$$\mathrm{Chl + Ms + Qtz = Bt + Crd + H_2O} \qquad (2)$$

$$\mathrm{Chl + Ms + Qtz = Bt + Crd + And + H_2O} \qquad (3)$$

$$\mathrm{Chl + Ms + Qtz = Bt + And + H_2O} \qquad (4)$$

Based on field observations, cordierite appears before andalusite. This together with the pattern of inclusions inside andalusite visible in figure 3d (i.e. only biotite and quartz without core to rim differences) suggest that cordierite grew outside the stability field of andalusite due to reaction (2). Nevertheless, this interpretation would require a more Mg-rich bulk composition. Several factors can explain this compositional discrepancy. Measured bulk X_{Mg} composition varies from ~ 0.4 to 0.5 (see AFM diagram figure 4), which is very close to both chlorite and biotite that are close to 0.5 in X_{Mg}. Thus, small retrograde changes in X_{Mg} of these phases can potentially override the prograde compositions and make interpretations based on AFM more difficult. Additionally, iron-sulfide is usually present in such rocks, but bulk rock composition was not projected since no major element of oxides were constrained. This would change the X_{Mg} of the equilibrium volume considered, and most likely would be a function of temperature since iron sulfides usually vary along the pyrite-pyrrhotite join.

The pattern of inclusions inside andalusite together with the progressive disappearance of cordierite in the stability field of andalusite is interpreted as the consequence of the reaction:

$$\text{Ms} + \text{Crd} = \text{Bt} + \text{And} + \text{Qtz} + \text{H}_2\text{O} \qquad (5)$$

The next encountered isograd is the appearance of fibrolitic sillimanite (Figure 3e), that forms millimeter-size nodules elongated within the main stretching direction. In one sample, we found evidence of polymorphic breakdown of andalusite into sillimanite (reaction 6), having the two minerals coexisting but never in direct contact as observed by many authors (e.g. Carmichael, 1969). However, the sillimanite isograd mapped in the field rather corresponds to the muscovite breakdown reaction (reaction 7) that produced large cluster of fibrolite, which are easier to observe in the field.

$$\text{And} = \text{Sil} \qquad (6)$$

$$\text{Ms} + \text{Qtz} = \text{Kfs} + \text{Sil} + \text{H}_2\text{O} \qquad (7)$$

The last observed isograd in the upper Paleozoic sequence corresponds to the onset of partial melting that is characterized by the formation of quartzo-feldspatic leucosomes parallel to the HT/LP main foliation (Figure 3f). In the field, the melt-in isograd is always located a few tens of meter away from the muscovite-out isograd that corresponds to reaction (7). Onset of partial melting is assumed to be water saturated or near water saturation. Partially molten micashists host tourmaline-bearing pegmatitic bodies that range in thickness from a few centimeters up to a several decameters. In the vicinity of these pegmatites, the micaschists are characterized by the crystallization of late euhedral muscovite that cross-cut the main foliation (Ms2 in figure 3e).

In the field, the thickness of the partially molten micaschist zone represents only a few decameters to a hundred meters. Then the molten micaschists evolve into a biotite-rich quartzo-feldpathic gneiss characterized by minute porphyroblast of K-feldspar. This lithology has been referred by Delay (1990) as the “pearled gneiss”. Partial melting evidences in this lithology are rare and when present consist of discrete dismembered leucosomes. We suggest that the onset of partial melting resulted in a strong fractionation of water into the melt leading to local fluid-undersaturation and therefore subsolidus conditions. Furthermore, in most locations, including Caladroy, the “pearled gneiss” is strongly deformed and consists in mylonitized partial melted micaschists. Bouhallier et. al (1991) described a high strain zone in Caladroy having an abrupt transition between the micaschists and the gneisses. The transition between micaschists and pearled gneisses is also characterized by a discontinuous intercalation of a marble layer, the so-called “base marble”, that can be located either at the interface between the partially molten micaschists and the gneiss or within the gneiss itself. To conclude, we propose that the “pearled gneiss” is derived from the same Early Ordovician protolith as the overlying micaschist and schist, and is produced by concomitant water-undersaturated metamorphism and strain localization.

3.1.2 Kinzigites of the gneissic core

The gneissic core, located structurally below the metamorphosed upper Paleozoïc sequence (Jujol group), is mainly composed of paragneisses, orthogneisses and metacarbonates, the latter being more abundantly present in the deeper part of the massif (Figure 2). In this study, we have chosen to focus our petrological analysis on kinzigites (bt-sil-crd-grt bearing gneisses) that occur as discontinuous meter-scale layers at different structural levels in the gneissic core (Figure 2). These rocks are good candidates for calibrating P-T peak conditions across the entire gneissic core using P-T pseudosections. The studied samples show the same peak metamorphic mineralogical assemblage that consists of cordierite + garnet + sillimanite + biotite + alkali feldspar + quartz ± plagioclase (see samples 10Ag04, 10Ag11 and 11Ag03, Figure 4). Sillimanite occurs either as elongated prismatic grains, which define, with biotite, the main foliation wrapping around garnet and cordierite porphyroblasts (Figure 4a), or small needle inclusions in garnet, together with quartz and biotite; sillimanite commonly exhibits both textural positions in the same sample. Although evidences of partial melting are unclear at the thin-section scale due to post-peak deformation, field observations clearly suggest that these rocks were partially molten. All these minerals are stable at peak metamorphic conditions, and are consistent with biotite dehydration melting reaction:

$$\mathrm{Bt + Sil + Qtz = Grt + Crd + Kfs + H_2O\ or\ L} \quad (8)$$

Some studied kinzigites also show late shear zones where a high-temperature retrogression assemblage of Bi + Sill is observed. Samples 10Ag04, located near the Trilla bridge, shows that the peak metamorphic assemblage is partially overprinted by brittle deformation that results in the cataclasis of garnet porphyroblasts (Figure 4b). Similarly, sample 11Ag03, near the Cuxous castle, is affected by numerous extensive ductile to brittle-ductile shear zones without low-temperature retrogression, except for a strong retrogression of cordierite and K-feldspar (Figure 4c).

4 THERMOBAROMETRY

To constrain the P-T conditions along the metamorphic gradient, pseudosections were computed using two thermodynamic modelling programs, Perple_X (Connolly & Petrini, 2002, Connolly, 2005) and Domino/Theriak (De Capitani & Petrakakis, 2010). Whole rock analyses are from XRF analyses and are presented in table 1. Since no mineral zoning and very limited retrogression was observed, the bulk rock XRF analyses represent a good estimate of the equilibrium volume for both the HT-LP rocks and the rocks from the gneissic core.

4.1.1 Late Cambrian – Early Ordovician LP-HT gradient

In the case of the Late Cambrian - Early Ordovician sequence, the bulk chemistry is rather homogeneous (Figure 5). The effect of bulk composition variability on phase relations and the location of major phase boundaries are thus very limited (Figure 5b). Therefore, phase relations were computed with the bulk composition of sample 10Ag16, in the middle of the sample composition cluster (white dot, Figure 4). The P-T phase diagram section produced in NCKFMASHT (Figure 6a) shares similarities with the P-T projection (petrogenetic grid) of Pattison & Tracy (1991) that was used by Guitard et al. (1996) to constrain semi-quantitatively the LP-HT metamorphic gradient in the Agly massif for the first time. In this contribution we rejuvenate this approach quantitatively with more petrological constraints and using recent thermodynamic databases.

Figures 6b, c and 7 show that the metamorphic P-T conditions are bracketed in a temperature range from 500-520°C (chlorite breakdown, reaction (2) outside the stability field of andalusite) to 660-

680°C (muscovite breakdown, reaction (7) and the onset of partial melting at ~ 0.35 GPa.

Defining the exact temperature (and pressure) of intermediate samples characterized by the appearance of andalusite and cordierite is more challenging because of the controversy regarding the exact P-T slope and location of reaction (5) that involves both andalusite and cordierite. In the computed NCKFMASHT phase diagram section (Figure 6a), the positive slope of the reaction (5), here represented by the field Ms + Crd + Bt + And + Qtz + Pl + Ilm (labelled "Ms Crd Bt And" in figure 6a), is in contradiction with field observations made in many low pressure contact aureoles as well as the Agly massif, where cordierite appears just before andalusite (Pattison, Spear, Debuhr, Cheney, & Guidotti, 2002; Pattison & Vogl, 2005; Pattison & Debuhr, 2015). As can be seen in figure 6b, this inconsistency is not related to bulk compositional variations, although the position of this equilibrium varies with X_{Mg} bulk rock composition.

The reaction 5 has low entropy change, thus it is very sensitive to the quality of the thermodynamic data. To overcome this known problem in low pressure metapelites, Pattison & Debuhr (2015) used a modified Berman (1988) thermodynamic database (Pattison et al. 2002; Pattison & Debuhr, 2015; Spear & Cheney, 1989; Spear & Pyle, 2010) instead of the Holland and Powell (1998) or (2011) version; this database is referred to here as SPaC14 as in Pattison & Debuhr (2015). Their approach was (1) to simplify the chemical system (excluding non-essential components like Na_2O, CaO and TiO_2) to minimize uncertainties related to poorly constrained thermodynamic constant of pure end-member, and (2) to use simplified ideal site mixing solid-solutions. Following this approach, we have computed the phase diagram section in a simplified KFMASH system shown in figures 6c and 6d using the 1998 version of the Holland and Powell thermodynamic database, updated in 2002, and the SPaC14 database, respectively. The bulk composition used is the same as in figure 6a, except that Na is projected through albite, Ca through anorthite and Ti through ilmenite, resulting in a small variation in the molar amount of Al_2O_3, SiO_2 and FeO contents of the equilibrium volume considered in the KFMASH system. As shown by Pattison & Debuhr (2015), the use of the SPaC14 database results in a phase diagram with a negative slope for the Crd + And + Bt + Ms equilibria (5), that is more consistent with the appearance of cordierite and then andalusite along a prograde path at ~ 0.3 GPa. In figures 6c and 6d, it is interesting to note that andalusite is predicted to be stable at low temperature in equilibrium with chlorite and chloritoid (i.e. before the appearance of cordierite). This feature is commonly described in contact aureoles, for example in the Bugaboo aureole (Pattison & Debuhr, 2015) or the Little Cottonwood (Wohlers & Baumgartner, 2012), but is in contradiction with field observations in the Agly massif. In the case of the Agly massif, we suggest that the stabilization of andalusite and chloritoid is the result of excess aluminum in the calculated bulk composition after albite and/or anorthite projection. At these P-T conditions, paragonite is very likely to be stable but it is not taking into account in this modeling, resulting in a higher Al_2O_3 content of the effective bulk composition.

The main petrological constrains for the metamorphic gradient of the Paleozoic sequence are: 1) chlorite breakdown and crystallization of biotite and cordierite, 2) crystallization of andalusite with or without cordierite present, 3) andalusite/sillimanite polymorphic transformation, 4) muscovite breakdown without cordierite stable, and 5) melting outside the stability field of muscovite. Minimum pressure in the low-temperature side of the gradient (~550 °C) cannot be constrain using solely the phase diagram. However, the thickness of rocks above the Paleozoic schists is inferred to be approximatively 10 km (Olivier et al. 2004). Assuming a density of the crust of 2.7 g/cm^3 would give a pressure around 0.27 GPa. The transition from andalusite to sillimanite, breakdown of muscovite and partial melting occur on a very short distance in the field and a very limited temperature range (620 to 670 °C) suggest that the maximum pressure is limited by the location of the invariant point (i) at 0.35 GPa. Assuming a linear gradient, these two values (0.27 GPa at 550 °C and 0.35 GPa at 670 °C, allow us estimating the gradient at 55°C/km (2076 °C/GPa with 2.7 g/cm^3). Such value is in very good agreement with a conductive gradient between partially molten schists at 670 °C and 0.35 GPa (i.e. a 13 km depth), and the surface since the resulting gradient would be ~52 °C/km. This high value is however lower than the previous qualitative estimates of 80 to 100 °C/km (Fonteilles, 1976; Delay, 1989; Guitard et al., 1996; Vielzeuf 1984).

4.1.2 Kinzigites from the gneissic core

To constrain the P-T conditions in the gneissic core, we have selected 10 kinzigite layers from various structural levels. The pristine kinzigite samples preserve the same peak mineral assemblage produced by biotite dehydration-melting reaction, which has a steep slope in P-T space (Figure 8a). In addition, divalent cations composition of garnet was used to better constrain P-T conditions and highlight continuous or discontinuous changes in pressure and temperature along the cross-section. A phase diagram section was computed for each sample with Perple_X using whole rock XRF composition. Garnet compositions were determined using electron microprobe with wavelength dispersive spectrometry (WDS) during 3 sessions. The first two sessions were conducted with a JEOL Superprobe 8200 at the Institute of Earth Sciences at the University of Lausanne, the third session was on a Cameca SX100 at the University of Lille 1. Synthetic and natural minerals were used for standardization. Average compositions and variations for the flat core of garnets for each sample can be found in table 2, full analyses can be download with the supplementary materials.

There is no discernable zonation at the level of precision of EPMA analyses (Figure 7), except for small increase in Fe/(Fe + Mg) at the rim, due to retrograde interdiffusion with biotite. Spessartine content can locally increase at the rim when garnet is retrogressed by biotite due to a very high preference for incorporation of Mn in garnet compare to biotite (Kohn & Spear 2000). The assumed equilibrium composition used to constrain P-T conditions were always coming from garnet cores. Figure 8a is an example of pseudosection obtained for the sample 11Ag03. Almandine and pyrope content in garnet are suitable to constrain P-T conditions because the slope of their isopleths is either temperature or pressure dependent in the quadrivariant garnet – melt – bearing stability fields located on both sides of the main equilibrium (8). There is generally a very good agreement between almandine and pyrope compositions for all samples, with isopleths usually located close from each other (Figure 8a). Spessartine and grossular content are very low, i.e. below 5 % of the X site of garnet, except for 4 samples (Table 2), thus they do not represent a good constraint for P-T condition due to a limited accuracy of the thermodynamic database for these low contents.

A synthesis of estimated P-T conditions is shown in figure 8c and shows a range in P-T conditions from 0.49 GPa, 730°C to 0.65 GPa, 790°C. However, there is good agreement between the pressure estimates from thermodynamic modelling and the relative structural position (i.e. depth) of samples in the gneissic core (Figure 8c). Samples located in the upper part of the gneissic core above the Ansignan charnockite intrusion have pressures ranging between 0.49 and 0.55 GPa (Figure 8c). The minimum pressures of 0.50 to 0.52 GPa were obtained on samples 10Ag03, 10Ag09, 11Ag03 and 11Ag04 which are located a few tens to hundreds of meters below the sheared contact with the Paleozoic sequence. The difference in pressure with the deepest sample form the upper Paleozoic sequence (partially melted micaschist: 0.35 GPa at 670°C) is about 0.15 GPa.

Sample 10Ag11 located in the western central part of the massif and structurally in the middle of the gneissic core gave P-T conditions of 0.57 GPa and 750°C. In contrast, samples 10Ag01 and 10Ag02 that are the deepest samples, below the Ansignan charnockite, are characterized by pressures estimates of 0.66 and 0.62 GPa for temperatures of 770-800°C, respectively. However, sample 10Ag04, that is also located below the charnockite and may belong to the same kinzigite layer as 10Ag01 and 10Ag02, gave inconsistent pressure estimate of 0.52 GPa. If we disregard sample 10Ag04, then the difference in pressure between samples located in the deepest level and 10Ag11 is around 0.1 GPa (~2.7 km), which contrasts significantly with the observed thickness on cross-section that does not exceed 200 m.

5 U-Th-Pb MONAZITE GEOCHRONOLOGY

5.1 Studied samples

Three kinzigites (10Ag02, 10Ag04 and 11Ag03) have been collected in order to constrain the timing of metamorphism. These samples are located at different structural levels and show various degree of deformation and retrogression. The deepest samples, 10Ag02 (Agly lake) and 10Ag04 (near the Trilla bridge), come from the same structural level (Figure 1 and 2). They are pristine kinzigites that consist of garnet, cordierite, sillimanite, quartz and feldspar with minor biotite. Sample 10Ag04 shows a more penetrative foliation with localized thin shear zones where garnet is fractured and partially dismembered (Figure 9b). Sample 11Ag03 comes from the uppermost level of the gneissic core, near the castle of Cuxous. It is characterized by the development of a mylonitic foliation, normal shear sense ductile shear zones and brittle fractures that affect sillimanite and K-feldspar porphyroclasts (Figure 9c). Peak metamorphic phases like K-feldspar and cordierite are almost completely pinnitised in this last sample.

5.2 Analytical methods

In situ U-Th-Pb isotopic data were obtained by laser ablation inductively coupled plasma spectrometry (LA-ICP-MS) at Laboratoire Magmas & Volcans (Clermont-Ferrand, France). The analyses involved the ablation of minerals with a Resonetics Resolution M-50 powered by an ATL Atlex Excimer laser system operating at a wavelength of 193 nm. For monazite, spot diameters of 9 µm were used, associated with repetition rate of 1 Hz and laser fluence of 9 /cm^2. The ablated material was carried by helium and then mixed with nitrogen and argon before injection into the plasma source of an Agilent 7500 cs ICP-MS equipped with a dual pumping system to enhance sensitivity (Paquette et al., 2014). The alignment of the instrument and mass calibration were performed before every analytical session using the NIST SRM 612 reference glass, by inspecting the signal of ^{238}U and by minimising the ThO+/Th+ ratio (< 1%). The analytical method for isotope dating is similar to that developed and reported in Paquette and Tiepolo (2007) and detailed in Hurai et al. (2010) and Gasquet (2010). The signals of ^{204}(Pb+Hg), ^{206}Pb, ^{207}Pb, ^{208}Pb, ^{232}Th and ^{238}U masses were acquired. The occurrence of common Pb in the sample was monitored by the evolution of the ^{204}(Pb+Hg) signal intensity, but no common Pb correction was applied owing to the large isobaric interference from Hg. The ^{235}U signal was calculated from ^{238}U on the basis of the ratio ^{238}U/^{235}U= 137.88. Single analyses consisted of 30 seconds of background integration with the laser off, followed by 60 seconds integration with the laser firing and a 20 second delay to wash out the previous sample and prepare for the next analysis.

Data were corrected for U-Pb fractionation occurring during laser sampling and for instrumental mass discrimination (mass bias) by standard bracketing with repeated measurements of the Trebilcock monazite (Tomascak et al., 1996) reference material. Repeated analyses of the monazites Itambé and Bananeira (Gonçalves et al., 2016) reference materials, treated as an unknown independently control the reproducibility and accuracy of the corrections. Data reduction was carried out with the software package GLITTER® from Macquarie Research Ltd (Jackson et al., 2004; van Achterbergh et al., 2001). For each analysis, the time resolved signals of single isotopes and isotope ratios were monitored and carefully inspected to verify the presence of perturbations related to inclusions, fractures, mixing of different age domains or common Pb. Calculated ratios were exported and reported in $^{208}Pb/^{232}Th$ vs. $^{206}Pb/^{238}U$ "Concordia" diagrams generated using the Isoplot/Ex v. 2.49 software package of Ludwig (2001). Owing to the lack of correlation between Th/Pb and U/Pb uncertainties, a low correlation coefficient value of 0.1 was considered in the calculations. The concentrations of U-Th-Pb were calibrated relative to the values of the GJ-1 zircon (Jackson et al. 2004) and Trebilcock (Tomascak et al., 1996) reference materials.

5.3 Results

5.3.1 Pristine kinzigite from the gneissic core (10Ag02 and 10Ag 04)

Monazites in sample 10Ag02 (Agly lake) are located in the matrix and range in size from 20 to 100 µm. Two distinct compositional domains are defined based on Y distribution: a Y rich core and Y poor rim (Figure 10a). Th and U content is constant between both domains. Six monazite grains were analyzed and most of the analyses are concordant in a $^{232}Th/^{208}Pb$ vs $^{238}U/^{206}Pb$ concordia diagram (table 3, Figures 10a and 11a). Y-rich cores give the oldest ages, with $^{232}Th/^{208}Pb$ ages ranging from 240 ± 7 Ma to 311 ± 9 Ma. The four oldest analyses were selected to calculate a concordant age at 305 ± 4 Ma ($MSWD_{(C+E)} = 1.5$). Y-poor rims show a spread of concordant ages that range for *c.* 307 to 232 Ma, which precludes the calculation of concordant age (Figure 11a, red circles).

Monazites in sample 10Ag04 are either present as inclusions in garnet or grains in the matrix. They range in size from 50 to 150 µm (Figure 10b). Two compositional domains are defined based on Y and Th/U compositions. Monazite included in garnet (m2 and m5) and the core of monazite in the matrix (m3, m6, m8, m9) are characterized by high and variable Y content with a very constant Th/U of 6.5 ± 1 (Table 3, Figure 10b). Monazite in the matrix are all characterized by small overgrowths (10 to 20 µm) with very low Y content (m1, m3, m4, m6, m8, m9) and a very high and variable Th/U, ranging from 10 to 34 (Table 3, Figure 10b). Grain m4 is homogeneous in composition, with low Y content and high Th/U ratio, it is interpreted to belong to the same generation as the overgrowths based on its composition. Monazite included in garnet do not show low Y and high Th/U overgrowth suggesting that the formation of the overgrowths is post-garnet crystallization and monazite entrapment. Eight monazite grains were dated. Seventeen isotopic analyses belonging to the five monazites with high Y content and low Th/U ratio were performed. On a $^{232}Th/^{208}Pb$ vs $^{238}U/^{206}Pb$ concordia diagram, all sixteen analytical points are concordant and allow to calculate a concordia age of 306 ± 2 Ma ($MSWD_{(C+E)}$ = 1.05), which is consistent with the concordia age obtained on 10Ag02 (Figure 11a and b). The constant Th/U ratio and the occurrence of monazite included in garnet suggest that this age correspond to the age of the peak metamorphism. The second compositional domain corresponding to the low Y and high Th/U overgrowth, yield a second set of twelve concordant dates with a concordia age of 300 ± 2 Ma ($MSWD_{(C+E)}$ = 0.71) (Figure 11d).

5.3.2 Deformed kinzigite of the upper gneissic core (11Ag03)

Five monazites located in the matrix were mapped. They range in size from 50 to 120 µm. Most of the grains show homogeneous, large, Y-depleted core surrounded by an Y rich overgrowth that do not exceed 10 µm in size (Figure 10c). Thorium content and distribution are homogeneous at the grain scale. Monazite 1 differs from the others because it preserves an irregular very low Th core (table 3 and Figure 10c). Three of the monazite grains analyzed (m1, m2 and m3, Figure 9c) show a cluster of nine concordant dates that allow calculating a concordia age of 296 ± 2 Ma ($MSWD_{(C+E)}$ = 1.3) (Figure 11c). A single analysis has been obtained on the Th poor core of monazite 1 (#16d: $^{232}Th/^{208}Pb$ age of 319 ± 10 Ma, table 3) and was not considered for the calculation of the concordia age. The 296 Ma age is consistent within errors with the age obtained on the high Th/U monazite of sample 10Ag04 (300 ± 2 Ma) and can either be interpreted as the age of fluid percolation during the mylonitic deformation or fluid exsolution from crystalizing melt of migmatites. Although monazite m3 has the same compositional features as the other grains, it shows a spread of concordant dates in the $^{232}Th/^{208}Pb$ vs $^{238}U/^{206}Pb$ concordia diagram. This spread in age along the concordia has been observed in all analyzed samples, especially 10Ag02 and 11Ag03 (Figure 11a and c), it could reflect a younger partial lead loss. We speculate that this isotopic disturbance could be the result of localized fluid percolation during a Cretaceous event (ca. 100Ma) that affects the Pyrenees range (Vitrac-Michard & Allègre, 1975, Boulvais et al., 2007, Poujol et al., 2010).

6 DISCUSSION

Our new geochronological data combined with phase relation modeling suggest that peak metamorphism in the gneissic core was achieved at *c.* 305 Ma and that the massif was affected by deformation between 300 Ma and 296 Ma. The age of metamorphism in the upper Paleozoic sequence was not directly and quantitatively constrained by geochronology but we assume it is related to the same late Variscan event. In the following, we will see that the finite metamorphic field gradient observed in the Agly massif results from the superposition of a thermal (magmatic and metamorphic) event (305 Ma) and a later tectonic (296-300 Ma) event.

6.1 Geothermal gradient during the main metamorphic event (*c.* 305 Ma)

Figure 12 shows the reconstructed metamorphic gradient for the entire rock column of the Agly massif. It can be clearly seen that the geothermal gradient of the Agly massif is not linear and does not result from a purely conductive heat transfer. The middle crust preserved in the gneissic core of the Agly massif shows a very low to near isothermal gradient (~ < 8°C/km), while the low pressure assemblages in the upper crust (i.e. andalusite-sillimanite transition, cordierite stable) defines a significantly higher geothermal gradient (> 50°C/km) in the Late Cambrian – Early Ordovician sequence. This thermal structure of the continental crust has been previously described in high-temperature and mid-crustal granulite terranes (Hollister, 1982; Patiño-Douce et al. 1990; De Yoreo et al. 1991; Thompson & England, 1984; Thompson, 1999). Two different tectonic settings have been called upon to explain such abnormal geothermal gradients. First, rifting resulting in an extremely thin continental crust, with a possible mantle derived basaltic sill intrusion in the lower crust providing sufficient heat to reach such high temperatures. This has been proposed in the Pyrenean belt by Wickham and Oxburgh (1985; 1987), partly based on stable isotopes signature in the uppermost crust (equivalent of the Late Cambrian - Early Ordovician sequence in the Agly massif). The second alternative model is referred as the “melt-enhanced geotherm” (Depine et al. 2008) for which the combined effect of (1) upward heat advection associated with melt transfer and (2) thermal buffering in the middle crust due to dehydration-melting reactions (either biotite or hornblende breakdown reaction). In the case of the Agly massif several arguments are in favor of a melt-enhanced thermal structure at *c.* 305 Ma:

1) Contemporaneous magmatism and metamorphism at ca. 305 Ma.

The main metamorphic event responsible for the abnormal geothermal gradient at 305 Ma is coeval with a widespread magmatic event between 308 and 304 Ma (Figures 13a and b). In the deepest part of the Agly crustal section, the Ansignan charnockite and the Cassagnes sills were emplaced at 307 ± 3 Ma and 308 ± 3 Ma, as a 500 to 800 m laccolith and decameter sills, respectively, parallel with the main gently dipping foliation developed at peak metamorphic conditions. The upper part of the gneissic core and the micaschist belonging to the Paleozoic sequence are intruded by the kilometer thick Tournefort diorite at 308 ± 1 Ma and St Arnac granite at 304 ± 5 Ma that is exposed on a 25 km^2 surface (Olivier et al., 2008).

2) Melt source and upward advection of heat by melt

The Ansignan charnockite is a composite laccolith intrusive that consists of a massive ~ 500 to 800 m sill of orthopyroxene ± garnet and K-feldspar megacrysts granite (i.e. the charnockite) associated with norite bodies up to 300 m in its core and garnet-bearing leucocratic granites in its upper margins. The source of the charnockite and associated magmatic rocks has been a matter of debate in the 1980's (Fonteilles, 1981; Fourcade, 1981; Pin, 1989). All these studies agreed that the charnockite was related to the partial melting of continental crust and more particularly metasediments. Pin (1989) who conducted the most extensive isotopic analysis on the different lithologies, suggested that the charnockitic magma without garnet (the more widespread facies of the charnockitic complex), characterized by εNd_{300Ma}, ranging from -6.4 and -4.4 and $^{87}Sr/^{86}Sr_i$ = 0.7079 was produced by partial melting of metasediments due to the intrusion of contaminated mafic magmas, corresponding to the norite bodies (εNd_{300Ma} = -2.5 to -3.4 and $^{87}Sr/^{86}Sr_i$ = 0.7062-0.7067). These charnockitic magmas are characterized by a strong HREE fractionation (Fourcade, 1981; Tournaire-Guille et al., 2019, Figure 11b) suggesting that melts, produced by a high degree of partial melting, were in equilibrium with a garnet-rich restite at a pressure greater than 6 kbar (see below and Figure 13a). Chemical exchange between the mantle-derived mafic magmas with lower crust-derived magmas resulted in a moderate Nd ans Sr isotopic hybridization. Pin (1989) invocated a second stage corresponding to the ascent and emplacement of this hybridized magma (norite and charnockite) into upper crustal levels, corresponding to the present day gneissic core of the Agly massif, at conditions estimated at 0.5 ± 0.05 GPa (~18 to 22 km) at 750 ± 100°C (Vielzeuf, 1984; 1986 and Figures 13a and b). The latent heat of crystallizing magmas induced partial melting of the surrounding metasedimentary host-rock and the formation of the garnet-bearing charnockite and leucogranites commonly observed at the margins of the intrusions. These garnet-bearing magmas are characterized by a stronger crustal signature with the lowest εNd_{300Ma} = -6.4 and -7.3 and the highest-$^{87}Sr/^{86}Sr_i$ = 0.7109 – 0.7127 recorded in the Ansignan intrusive complex.

The Cassagnes granodiorite, that consists of K-feldspar-bearing megacrystic sills of about 10 meters thick, has a close age at 308 ± 3 Ma and very similar major / trace element composition to the

Ansignan charnockite (see Figures 10 and 11 in Tournaire-Guille et al., 2019). Therefore, we suggest that both magmas have the same origin and source depth. The main difference between them is that Cassagnes sills do not contain orthopyroxene nor garnet probably because they were emplaced in a much shallower crustal level than the Ansignan granodiorite. Based on P-T constrains on kinzigite located near the Cassagnes sills (10Ag06 and 11Ag03) on the eastern part of the gneissic core, we suggest that they were emplaced at 0.5 GPa at 740°C corresponding to a depth of ~ 18 km (Figures 13a and b).

The Tournefort diorite and Saint-Arnac granite are calc-alkaline magmas, similar to the other major late-Variscan intrusions in the Pyrenees like the granite of Millas, Quérigut, Saint Laurent, Maladetta or Mont-Louis. They are interpreted as the result of the emplacement of magmas in the upper crust issued from partial melting of a deep old magmatic continental crust or hybridization of deep seated crustal magmas with mantle-derived basalt intrusive and related to continental strike slip and exhumation tectonics during the late Variscan collision (Lagarde et al., 1992).

It is clear from field observations that the Saint-Arnac granite and Tournefort diorite were not produced in-situ but rather transferred from an unknown depth in the deep crust to the upper Paleozoic sequence. Because the Tournefort diorite intruded andalusite-bearing micaschist, we estimate the maximum depth of emplacement at about 11 km based on the 0.3 GPa pressure estimated on the host rock (Figures 12 and13a). Field evidence, like surrounding rocks xenoliths, magma mingling and diorite enclaves in the granite, suggests that the Saint Arnac pluton is located at the same depth in the micaschist. The granite intruded all the Ordovician micaschist and schist series up to the Devonian marbles found as decameter xenolith in the wall of the granite. We therefore suggest that the Saint Arnac granite has a minimum thickness of ~2500 m, corresponding to the estimated thickness of the Ordovician series. The Saint-Arnac granite would have been emplaced in the depth bracket of 9.5 and 12 km (Figure 13a). This estimation differs significantly with Olivier et al. (2008), who suggested that the granite is 5 km thick with a roof located at a shallow depth of 2.7 to 4 km (0.1 to 0.15 GPa). The emplacement of both intrusions is associated with a local contact metamorphism that is characterized by partial melting around the diorite and crystallization of large muscovite flakes in the surrounding micaschists around the granite and micaschist xenoliths.

At the scale of the Agly massif, all these magmatic intrusions represent a significant volume that is difficult to establish with confidence but may be greater than 30% of the basement (Figure 1). Their migration from a deep source, currently not accessible in the exposed Agly massif, towards upper crustal levels constitutes a very efficient and dynamic heat advection process. The crystallization of late muscovite in micaschists around the Saint-Arnac granite and partial melting surrounding the

Tournefort diorite are consistent with magmatic heat advection in the previously LT Paleozoic sequence. In the gneissic core, the upward heat advection during melt transfer and heat released during the crystallization of magmas would increase temperature. This is in good agreement with the observation of high temperatures ranging from 730 °C to 790 °C for pressure from 0.49 to 0.65 GPa (Figure 13b) for the entire gneissic core. At these depths, with a steady state-geotherm, the temperatures would only be about 400-600 °C (figure 13a). The juxtaposition of high-grade rocks from the lower crust as the explanation of the metamorphic gradient observed in the Agly is thus a very unlikely explanation. These high temperatures cannot be achieved without the additional heat source from magmas, even with a high mantle heat flux (Figure 13a). We suggest that the low gradient at about 750 °C in the gneissic core is the combination of melt migration, plutons formation and thermal buffering by the biotite-dehydration melting reaction that produced garnet +/- cordierite. In contrast, the geothermal gradient in the upper crust, above the shallower intrusion to the surface, is significantly higher (~ 55 °C/km) and correspond to a diffusive geothermal gradient (compare Figures 13a and b). The orthogneisses of the gneissic core might not have been as affected as the metapelites by the thermal buffering since they are less fertile in terms of melt production (Schorn, Diener, Powell, & Stüwe, 2018). Nevertheless, they represent only a marginal portion of the gneissic core (Figure 1). Thus, the temperatures computed from metapelites, buffered by the biotite dehydration-melting reaction, can be interpreted as representative of the thermal state of the gneissic core.

6.2 Thinning of the massif

Evidences for a major thinning of the Agly massif have been well accepted for decades now, especially based on the difference between depth inferred from previous geothermobarometric studies and thickness of the rock column of the massif in its present geometry. This thickness has been inferred with paleo-pressure calibrations to be of about 11 km between the top of the Paleozoic Jujol group and the bottom of the Charnockite and represents now a lithologic pile of only 5-6 km of exposed rocks. Nevertheless, the timing and importance of this thinning is still a matter of debate, as recently outlined by the succession of publications (Olivier at al. 2004; 2008; Olivier 2013; Vauchez et al. 2013a; 2013b). Even more recently, two studies have rejuvenated this debate with new data, using (U-Th)/He (Ternois et al., 2019) and U-Pb (Odlum & Stockli, 2019) in both zircons and apatite. The first study founded that the Agly massif never experienced temperature higher than 250 °C during Cretaceous rifting while the second study argued that the gneissic core was still at temperatures higher than 450 °C before rifting. In terms of the thinning itself and concerning the late Variscan tectonics only, two different deformation settings have been proposed.

First, the presence of one major subtracting zone responsible for the main thinning of the massif was proposed by Bouhaillier et al. (1991) and Delay (1990). This detachment zone would act as a major normal fault located at the boundary between the Late Cambrian – Early Ordovician sequence (Jujol group) and the gneissic core. It has been referred to the Latour-Caladroy mylonitic zone (Figure 1 and labelled “Caladroy SZ” on the cross-section, figure 2) since many mylonites with normal shear behavior have been observed between the city of Latour-de-France and the castle of Caladroy. In contrast, Olivier et al. (2008) argued that the thinning of the massif was accommodated by numerous normal shear zones over the entire massif.

The reconstruction of the geothermal gradient in the Jujol group and in the gneissic core from the present study clearly shows a pressure gap of 0.15 GPa between the bottom of the Late Cambrian – Early Ordovician sequence and the uppermost level of kinzigite (Figure 12). A simple pressure conversion into depth (using an average density of 2.7), indicates that about 5.5 km of rocks may have been substracted by the Latour-Caladroy mylonites (Figure 12c). The fact that most of the deformation and subtraction was localized at the boundary between the partially molten micaschists and the gneissic core is not surprising since partial melting induces a major strength drop when melt proportion reaches 7 vol% (Rosenberg & Handy, 2005). The onset of partial melting triggered by muscovite dehydration reaction is known to produce under 10% of melt (White, Powell, & Holland, 2001), such value would then be in good agreement with deformation experiments.

We also suggest that thinning and associated rock subtraction is not restricted to the Latour-Caladroy mylonite but also occur within the gneissic core. Pressures estimated for the deepest kinzigite level close to the Charnockite are 0.62-0.66 GPa (10Ag01 and 10Ag02, Figure 12). This would give a pressure difference between the shallowest and the deepest kinzigite levels for the gneissic core of about 0.10 to 0.15 GPa, or a thickness of about 3.7 to 5.5 km. Since the present-day gneissic core thickness do not exceed 2.5-3 km, we suggest that the numerous mylonites observed throughout the entire gneissic core might be responsible for the remaining thinning of the massif (as quoted Ansignan and gneissic core mylonites on Figure 12c).

Monazite overgrowths with highly variable Th/U ratio from mylonitized kinzigites yield ages of about 296-300 Ma (samples 11Ag03 and 11Ag04; Figures 11c and 11d). These overgrowths may have been produced by fluid-assisted dissolution/ precipitation during localized fluid percolation in shear zones pathways. Furthermore, the highly variable Th/U ratio is also a good evidence that monazite rims were not in equilibrium rock-wide but rather locally equilibrated with a fluid. The origin of the fluid responsible for the dissolution/precipitation of the overgrowth could be the result of crystalizing melt from migmatites. This would require to be near the solidus at significantly lower temperature and pressure than 0.6 GPa and 750-790 °C determined at 305 Ma. Such drop in pressure and temperature would likely be the result of tectonic assisted exhumation, such that the age at 296-300 Ma would be the age of the thinning of the massif. To conclude, we suggest that the thinning of the Agly massif occurred between 300 and 296 Ma after peak metamorphism and the main magmatic event at *c.* 305 Ma. The Cretaceous thinning during rifting prior to the Pyrenean orogen would not have significantly disturbed the geometry acquired at the end of the Variscan orogeny, in agreement with the (U-Pb)/He on zircons and apatite data from Ternois et al. (2019). Nevertheless, blocs might have slightly moved along the E-W Pyrenean fault that crosscut the massif, especially the southernmost bloc where the Ansignan charnockite is located, that most probably moved to slightly upper levels along the E-W fault located in the Caramany lake.

7 CONCLUSION

The detailed petrological and geothermobarometric data from the present study shows that the Agly massif has experienced an abnormal thermal state during late-Variscan orogeny. This has resulted in a very high geothermal gradient of 55 °C/km in the upper crust (above 13 km depth) and a near isothermal gradient in the middle crust. New monazite U-Th-Pb geochronology show that this metamorphic event took place at *c.* 305 Ma. This abnormal geothermal gradient is the consequence of the coeval emplacement of several intrusions at different structural level during a limited time period *c.* 304-308 Ma. The heat provided by these intrusions allowed the middle crust to reach partial melting temperatures at depth below 13 km, and very high temperatures, i.e. ~750°C, between 18 and 25 km depth and be responsible of a “regional contact metamorphism”. The biotite dehydration-melting reaction, strongly endothermic, buffered temperature in the middle crust, resulting in the observed near isothermal geothermal gradient. The high geothermal gradient in the upper crust is mostly due to heat conduction between the anomalously hot middle crust and the cold surface.

This gradient was latter disturbed by normal mylonitic shear zones, resulting in the thinning of the massif from a thickness of 15 km to a final 5-6 km thickness. Most of this thinning was concentrated within a major detachment zone located at the onset of partial melting, between the Late Cambrian – Early Ordovician sequence and the gneissic core. This detachment zone was responsible for the subtraction of 5-6 km of rocks thickness. The remaining thinning most likely took place at several levels inside the gneissic core, resulting in the abundance of small mylonitic zones. The tectonic juxtaposition of a very hot middle crust (i.e. ~750 °C) and a colder upper crust probably increased the temperature of the later. Nevertheless, due to the already existing thermal structure resulting of the metamorphic event at *c.* 305 Ma, temperatures in the upper crust were already higher than normal, close to 660 °C at the contact with the gneissic core. The temperature increase due to thinning was only marginal and well below the precision of any present geothermobarometric method. This thinning event could be linked to the second age observed in monazite overgrowths and determined to be at *c*. 296-300 Ma. Nevertheless, this should be the focus of a detailed geochronologic study, aiming specific deformed monazites inside strongly deformed samples to directly date the thinning event.

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APPENDIX

Chemical system, software and activity-composition relationship model used for the thermodynamic modelling.

Figure	Chemical system	Software	Thermodynamic database	Solid Solution	a-X models abbreviation	Reference
Fig. 6a and 6b	NCKFMASHT	Perple_X	Holland and Powell (2002)	K-feldspar Plagioclase Melt White Mica Garnet Chlorite Cordierite Biotite Staurotide	San Pl(h) melt(HP) Pheng(HP) Gt(WPPH) Chl(HP) hCrd Bio(HP) St(HP)	Waldbaum and Thompson (1968) Newton et al. (1980) White et al. (2001) Holland & Powell (1998) White et al. (2005) Holland & Powell (1998)
Fig. 6c	KFMASH	Perple_X	Holland and Powell (2002)	K-feldspar Plagioclase Melt White Mica Garnet Chlorite Cordierite Biotite Orthopyroxne Staurotide	San Pl(h) melt(HP) Mica(CHA) Gt(WPPH) Chl(HP) hCrd Bio(HP) Opx(HP) St(HP)	Waldbaum and Thompson (1968) Newton et al. (1980)White et al. (2001) Coggon & Holland (2002) White al. (2005) Holland et al. (1998) ideal mixing Holland et al. (1998) Powell and Holland (1999)

						ideal mixing
Fig. 6d	KFMASH	Domino/ Theriak	SPac14	same as Fig 6c	Ideal site mixing	
Fig. 8a and 8b	NCKFMASHT	Perple_X	Holland and Powell (2002)	K-feldspar Plagioclase Melt White Mica Garnet Chlorite Cordierite Biotite Staurotide	San Pl(h) melt(HP) Mica(CHA Gt(WPPH) Chl(HP) hCrd Bio(HP) St(HP)	Waldbaum and Thompson (1968) Newton et al. (1980) White et al. (2001) Coggon & Holland (2002) White al. (2005) Holland et al. (1998) ideal mixing Holland et al. (1998) Ieal mixing

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Figure captions:

Figure 1: Geographic and geologic setting of the Agly massif; a) situation of the Agly massif in the Pyrenees in respect to all Variscan massifs, modified after Olivier et al. (2004); b) geological map of the massif produced during this study; the E-W cross section is discontinuous to take into account the displacement along the Planèzes fault, which was most likely active during very late Variscan stage.

Figure 2: Schematic E-W cross-section located on figure 1. The different shear zones (SZ) have been highlighted in dark black.

Figure 3: Petrography of the Late Cambrian – Early Ordovican samples, with increasing metamorphic grade from left to right and top to bottom. a) Polarized photomicrograph of a greenschist facies sample with sedimentary bedding and main schistosity; S_{n-1} microlithon shows the transition between chlorite grade and biotite grade. b) Polarized photomicrograph for a typical biotite grade sample, with biotite (Bt) and muscovite (Ms) in the main schistosity. c) Polarized photomicrograph of a cordierite grade sample, with a slightly rotated large poikilitic cordierite (Crd) porphyroblast with biotite, quartz and oxides inclusions. Biotite is also present in the matrix, in the main schistosity, as well as muscovite. d) Polarized photomicrograph of a cordierite-andalusite grade sample, with large poikilitic andalusite (And) porphyroblasts with biotite, quartz and oxides in inclusions. Biotite and muscovite are mainly present in the matrix aligned in the main schistosity,

interbedded with quartz (Qtz) and plagioclase (Pl). e) Polarized photomicrograph of a sillimanite grade sample, with typical fibrolite (Sil) patches, aligned in the main foliation. Muscovite (Ms2) is only present as secondary mineral, discordant to the main foliation. f) Polarized photomicrograph of a partially melted sample, with small aggregates of quartz (Qtz), alkali feldspar (Kfs) and plagioclase (Pl), representing the remaining melt pockets. Biotite (Bt) and sillimanite (Sil) are aligned in the main foliation.

Figure 4: Petrography for three kinzigite samples from different structural levels of the gneissic core, 10Ag11 located slightly above the Ansignan charnockite; 10Ag04 represents the deepest level, below the charnockite; 10Ag03 is the shallowest level, near the Cuxous castle at about 500m from the contact with the Late Cambrian – Early Ordovicain sequence. a) polarized photomicrograph for 10Ag11, with large garnet porphyroblasts and prismatic sillimanite (Sil) and biotite (Bt) aligned in the main foliation. Note the fresh cordierite grain on the lower right corner of the microphotograph. b) Polarized photomicrograph for sample 10Ag04, showing fractured garnet, fresh cordierite and monazite (Mnz) grains used for the geochronolical study. c) Polarized photomicrograph of the sample 11Ag03, showing highly deformed K-feldspars (Kfs), grain size reduction for prismatic sillimanite (Sil) and retrogressed cordierite (“Crd”). Note the grain size reduction in figure b) and c) in the shear bands.

Figure 5: Thompson-like AFM diagram projected from muscovite end-member. Black dots represent the XRF-analyses for the different samples of the Late Cambrian – Early Ordovician sequence. White dot corresponds to sample 10Ag16 considered as a representative composition (see text for explanations).

Figure 6: Thermodynamic modelling for the Late Cambrian – Early Ordovician sequence. Sample 10Ag16 was select since it is representative of the composition for the different samples of this sequence (white dot on figure 4). a) Pseudosection computed with Holland and Powell thermodynamic database (version 1998, updated in 2002) for an average composition of the schists in the NCKFMASHT system. Outlined on this pseudosection are: stability field of staurolite (light red area), upper limit of chlorite stability (green line), andalusite (And) to sillimanite (Sil) transition and muscovite breakdown (black lines), and hydrated granite solidus (red line) b) Compilation of

constrains for all samples in the system NCKFMASHT, the green lines represent the breakdown of chlorite, the yellow lines are the bounds for the stability field of staurolite, the black lines are the upper limit for the stability of muscovite, the blue lines represent the upper limit (in terms of pressure) of stability for cordierite, and red lines are the wet solidus curves; c) Pseudosection computed with Holland and Powell thermodynamic database (version 1998, updated in 2002) in the KFMASH system, with TiO2 projected through ilmenite, CaO projected through anorthite and NaO projected through albite. d) Pseudosection computed with Domino/Theriak and the SPaC145 thermodynamic database in the KFMASH system (using the same composition than for 6c). See text for more explanations.

Figure 7: a) BSE image of a garnet from sample 10Ag03 analyzed with EPMA for major elements determination. The bold black line represents the position of the transect analyzed. b) Garnet end-members composition along the transect shown on figure 8a. The garnet analyses highlight the absence of zonation for major elements composition, even near the rim of the grain. Note the discontinuous scale due to very high almandine content, above 80 % here and commonly above 65%for all garnet compositions.

Figure 8: thermodynamic modelling for the gneissic core, using Holland and Powell thermodynamic database (version 1998, updated in 2002) and NCKFMASHT system of components. a) Pseudosection for sample 11AgG03, located near the Castle of Cuxous, close to the Late Cambrian – Early Ordovician sequence. The shaded area in blue represent the small divariant field for the dehydration-melting reaction of biotite. Red and green lines are isopleths of garnet composition, almandine and pyrope contents, respectively. The bold lines for both contents represent the garnet composition for this sample (see table 2). The brown area between the two bold lines for isopleths represent the P-T conditions determined for this sample based on the assemblage and the garnet composition. b) Summary of all kinzigite samples P-T conditions determinations. As for figure 9a, red lines represent almandine contents of garnets for each sample, and green lines are pyrope contents, black lines are the upper and lower limits for the biotite dehydration-melting reaction. c) Cross section with the location and P-T conditions for all kinzigites samples. Those who are not strictly on the cross section (10Ag03 and 10Ag05) are projected based on their distance to the boundary between the gneissic core and the late Cambrian – early Ordovician sequence.

Figure 9: Large polarized photomicrographs for the 3 samples used for the geochronological study: a) 10Ag02; b) 10Ag04; c) 11Ag03. The position of the monazite grains analyzed for U-Th-Pb by LA ICP-MS are highlighted with yellow circles, with their associated names next to it. The limits of the field of view (dashed rectangles) used for the figure 6 for the petrological study for samples 10Ag04 (figure 5b) and 11Ag03 (figure 5c) are shown with dashed lines. Normal shear zones are highlighted for the highly deformed sample 11Ag03 using asymmetric arrows.

Figure 10: Yttrium and Thorium X-ray maps of the different monazite grains analyzed for the U-Th-Pb geochronology for samples 10Ag02 (A), 10Ag04 (B) and 11Ag03 (C). The white dots on Y maps represent the location of LA ICP-MS analyses, with the label number. For sample 10Ag04, monazite grains in inclusion in garnet are labelled with “grt” on top.

Figure 11: Concordia diagrams for all U-Th-Pb analyses of monazite grains. A) Analyses for sample 10Ag02, close to the Ansignan charnockite, all Y-rich cores define a statistical meaningful population with an age of 305 ± 4 Ma (2σ). B) Analyses of monazite for the sample 10Ag04, also close to the charnockite, taking only the spot representing Th/U ratio of 6.5 ± 1, they define a single population of ages with a mean of 306 ± 2 Ma (2σ), one very low concordant age has been excluded. C) Analyses of monazite for sample 11Ag03, the most deformed and retrogressed kinzigite sample, located in the upper part of the gneissic core. Nine of the analyses successfully represent a single population of ages, with a mean at 296 ± 2 Ma (2σ). The analyses of m3 have been excluded as well as the low Th core in m1 (figure 11c). Monazite grain 3 seems to be more affected by the Cretaceous deformations, with tailing of the ages toward younger values. D) Analyses of high Th/U zones of monazite grains from sample 10Ag04. 12 of these analyses clearly define a single population of ages with a mean value of 300 ± 2 Ma (2σ).

Figure 12: Reconstruction of the metamorphic gradient after the main Variscan metamorphic event with both the gneissic core samples and the late Cambrian – early Ordovician sequence.

Figure 13: reconstruction of the evolution of the geothermal gradient of the Agly massif during the Late-Variscan HT-LP metamorphic event. a) Intrusion of the magmatic bodies (grey boxes) at different structural levels of the Agly massif between 308 and 304 Ma, with a lower crustal source

for these magmas, see text. b) At the time of the main metamorphic event (*c*. 305 Ma), geothermal gradient is directly inferred from the P-T conditions determined from thermodynamic modelling for the gneissic core (red symbols) c) Reconstruction of the geothermal gradient after the main Late-Variscan thinning event, at *c*. 296-300 Ma. The late Cambrian – early Ordovician sequence is slightly heated due to the juxtaposition of the hot gneissic core. Blue areas represent the two main subtractive zones responsible for the thinning of the massif.

Mineral abbreviations around the alumino-silicate triple point are kyanite (ky), sillimanite (sil), andalusite (and). The two main dehydration reactions described in the text are highlighted, muscovite dehydration (ms -) and biotite dehydration melting (bt sil q = grt crd L).

Table 1: XRF analyses of all samples used in this study. GPS location are in the French system Lambert 2 étendu.

Table 2: Average garnet core analyses from EPMA for kinzigites samples of the gneissic core with their respective variations, computed as 1 standard deviation (1s). Structural formulae are computed based on 12 oxygens, the relative proportion of almandine (Alm), pyrope (Py), grossular (Gr) and spessartine (Sps).

Table 3: Summary of all U-Th-Pb data for monazite used in this study with their respective textural positions.

Table 4: Summary of P-T conditions for all kinzigite samples, determined from pseudosections (see figure 8b).