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Shortening of the European Dauphinois margin (Oisans Massif, Western Alps): new insights from RSCM maximum temperature estimates and $^{40}\text{Ar}/^{39}\text{Ar}$ *in situ* dating.

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Abstract

The Oisans Massif, located in the external zones of the western Alps, experienced significant shortening during the Alpine collision. While a series of major top-to-the west shear zones was recently described, the general low grade of the metamorphism has not attracted much petrological and geochronological studies. This paper provides combined temperature and age constraints on the evolution of the Oisans Massif. Temperature was estimated with the Raman Spectrometry of Carbonaceous Material (RSCM) method and chlorite geothermometry. Maximum temperature reached by the Mesozoic cover (T_{max}) from Grenoble to the Galibier pass (E-W) and from Saint Jean de Maurienne to Embrun (N-S) yielded almost constant ca. 330°C temperatures all through the Massif. Temperatures however strongly decrease either westward toward the top of the Vercors sedimentary sequence or eastward toward the Penninic Frontal Thrust. Age constraints were retrieved using $^{40}\text{Ar}/^{39}\text{Ar}$ *in situ* analyses performed on variously strained samples from Alpine shear zones. Over strain gradients, incipient Alpine recrystallizations progressively develop at the expense of former Variscan parageneses. A combined textural and EPMA approach permitted to identify newly formed chlorite and phengite that unequivocally grew in response to deformation or to low grade metamorphism. Chlorites recorded temperatures from ca. 350 to 150°C during the activity of shear zones. In parallel, $^{40}\text{Ar}/^{39}\text{Ar}$ *in situ*

experiments enabled dating deformation using both synkinematic phengites crystallized below the closure temperature of white-micas and former Variscan muscovite whose isotopic system have been, at least partially re-opened. Activity of top-to-the-west shear zones responsible for the shortening and thickening of the Oisans massif thus occurred between 34-33 and 25 Ma. Integrating these new age-constraints, Tmax estimates, and published geological data on the Oisans Massif and neighbouring areas allow proposing a new shortening scenario for the external zones. From ca. 34-33 Ma, the Oisans Massif was buried as a rigid block below more internal units and reached temperature of ca. 330°C. Shearing progressively localized along a series of top-to-the-W shear zones in the basement until 25 Ma. Deformation then localised along a major thrust at the base of the ECM massif and propagated along the basement-cover interface below the Vercors massif after 16Ma.

1 Introduction

The involvement of basement units in the external zones of collisional orogens is still poorly understood because the associated deformation is not intense and metamorphic conditions not severe enough to erase the effects of earlier deformation episodes, rendering the syn-collision deformation difficult to recognize. The deformation and kinematic history of these continental units thus often remains partly enigmatic and the rheology of the crust in the early stages of collision poorly known.

The Oisans Massif, which belongs to the External Crystalline Massifs (ECM) of the Western Alps is a key-place to understand the first stages of continental collision. Major basement brittle/ductile shear zones reworking the earlier Variscan fabric have been recognized which have accommodated most of the collisional shortening (Wibberley, 1999, 2005; Bellahsen et al., 2012, 2014; Bellanger et al., 2014; Boutoux et al., 2014a), but precise P, T, time data are still lacking to better constrain the sequence of shortening and thickening.

The Oisans Massif also comprises a Mesozoic sedimentary cover, mainly deposited during the Early Jurassic rifting episode, which has not seen the Variscan metamorphic conditions and can thus be used to assess the Alpine history. Bellahsen et al. (2012) and Bellanger et al. (2014) have shown that localised west-verging Alpine shear zones developed in the basement during collision while the cover was strongly folded above without a major décollement in between. The geometry and kinematics of these shear zones is thus now established and it remains to estimate the P-T conditions of their development and to date them.

In this contribution, we first provide a map of the maximum temperature ($T_{\max}$) reached by the cover between Grenoble and the Galibier pass (E-W) and between Saint Jean de Maurienne and Embrun (N-S) (Fig. 1) using the RSCM method (Beyssac et al., 2002). It constrains the thermal history as well as the burial dynamics around the Oisans Massif. Then, we compare these $T_{\max}$ with temperatures obtained in both basement and cover, using two chlorite geothermometers (Cathelineau and Nieva, 1985; Cathelineau, 1988; Vidal et al., 2001, 2005, 2006). In a second step, we provide EPMA analyses of white micas coupled to detailed microstructural and textural analysis of samples collected from various shear zones, in order to document the nature of these white micas, i.e. inherited vs. synkinematic. Finally, in a third step we performed *in situ* dating of white micas using the $^{40}\text{Ar}/^{39}\text{Ar}$ method. These results constrain the scenario of collisional shortening in the external zones of the Western Alps.

2 Geological setting

2.1 The Oisans Massif within the Western Alps

The Belledonne, Grandes Rousses, and Oisans massifs belong to the external zones and more particularly to the Dauphinois zone of the Western Alps (Fig. 1). External zones, which mainly recorded the Liassic rifting stage and the Oligo-Miocene collision event, were affected by a low-grade metamorphism after the end of Eocene. These massifs are parts of the External Crystalline Massifs (noted ECM, thereafter) that correspond to uplifted Variscan basement covered by Mesozoic to Cenozoic sedimentary series. The ECM were underthrust below an orogenic prism made of stacked continental (Briançonnais zone) and oceanic (Piemontais zone) nappes transported along the Penninic Frontal Thrust (PFT). The PFT, also called the Crustal Pennine Thrust or the Embrunais Basal Thrust in Dumont et al. (2011, 2012) was active with top-to-the-(north-)west motion during the Early Oligocene (Ceriani et al., 2001; Tricart et al., 2001, 2006). The internal zones, east of the PFT (Fig. 1), corresponds to the subducted continental margin of Europe and the Ligurian Ocean. They are pervasively affected by high-pressure low-temperature (HP/LT) metamorphism, which occurred between 70 and 35 Ma (Rosenbaum and Lister, 2005; Berger and Bousquet, 2008). During the Palaeozoic, the Oisans Massif experienced a complex tectonometamorphic history (Debon and Lemmet, 1999; Guillot and Menot, 1999; Von Raumer et al., 1999; Guillot et al., 2009). High-grade Carboniferous metamorphism has strongly overprinted previous tectonometamorphic records. The Oisans Massif mainly consists of late Variscan

granites and earlier Devonian-Dinantian plutonic and younger volcano-sedimentary sequences, including Viséan conglomerates, volcano-sedimentary and coal intercalations (Barfty and Pcher, 1984). The Belledonne and Rochail massifs were then structured by nappe stacking during the Visan and later affected by late Carboniferous to early Permian late-orogenic gravitational collapse (Guillot and Menot, 1999), contemporaneous with partial-melting and plutons emplacement between 330-290 Ma and the formation of extensional Stephano-Westphalian continental basins (i.e. in Grandes Rousses, Belledonne and La Mure massifs) (Barfty and Pcher, 1984; Barfty et al., 1984, 1988; Guillot et al., 2009). This zone was ultimately eroded during Permian to Triassic times as evidenced by the continental Permo-Triassic deposits (sandstones and conglomerates together with coal intercalations) and middle to late Triassic shallow marine sediments (dolomitic limestone with evaporitic layers). These formations remained attached to the basement during Alpine deformation, and the contact can thus be used as a reference surface to quantify both the extensional strain during the rifting episode and the later compressional deformation during collision (Dumont et al., 2008, 2011; Bellahsen et al., 2012, 2014; Bellanger et al., 2014; Boutoux et al., 2014a).

During early Liassic times, the post-Variscan basement was stretched during the rifting episode (Dumont, 1988; Dumont et al., 1998), as evidenced by basaltic flows ("spilites") (Laurent, 1992) and a series of large-scale fault-bounded tilted blocks, limited by major normal faults that still shape the present-day geometry of the external zones (Barfty and Gidon, 1983; Lemoine et al., 1986; Lemoine and De Graciansky, 1988). During the early Liassic, the Meije block (Fig. 3) was a horst, as evidenced by a condensed sedimentary cover and shallow paleobathymetric conditions (Roux et al., 1988; Bouillin et al., 1997; Corna et al., 1997).

After the rifting stage, the Liguro-Piemontais oceanic domain formed between the European and Apulian continental domains from the Bajocian onward, as evidenced by the age of the first sediments deposited above the ocean floor (De Wever and Caby, 1981; De Wever et al., 1987). Subduction of the Ligurian domain, which started during late Cretaceous times was followed by subduction of the leading edge of the European continental margin under the Apulian microplate as evidenced by the HP to UHP metamorphism in the internal zones (e.g., Chopin, 1984; Agard et al., 2001). The more proximal margin was subducted to shallow depths and shortened and the deformation propagated outward.

The "Pyreneo-Provenal phase", Late Cretaceous to Early Eocene in age, responsible for E-W folds and thrusts especially in the southern part of the external zone, is the first post-

147 Variscan shortening event (Lemoine, 1972; Caby, 1973). Ante-Senonian E-W fold axes
 148 described in the Dévoluy massif, were recently interpreted as due to submarine sliding
 149 (Michard et al., 2010). The Nummulitic series are deposited within the foreland flexural
 150 basin from Lutetian/Bartonian to Priabonian/Rupelian (Sinclair, 1997; Du Fornel et al., 2004
 151 Mulder et al., 2010). They unconformably overly reverse faults associated to N-S
 152 shortening, south of the Oisans Massif (Ford, 1996). The timing and the kinematics of other
 153 thrusts north of the Oisans Massif, such as the Meije or Combeynot thrusts, are debated,
 154 whether they are pre- (Barbier, 1963; Bravard and Gidon, 1979; Ford, 1996), or syn- to post-
 155 Nummulitic (Beach, 1981b; Dumont et al., 2011; Bellanger et al., 2014). The only evidence
 156 of N-S to NW-SE Oligocene shortening within the basement is restricted to a narrow area
 157 around the PFT (Bellanger et al., 2014).

158 The main Alpine collisional tectonometamorphic event in the external zones was thus due to
 159 its burial under the internal units immediately after the deposition of a Rupelian olistostrome
 160 topping the Nummulitic series, the so-called "schistes à blocs" (Kerckhove, 1969; Gidon and
 161 Pairis, 1980; Sinclair, 1997; Mulder et al., 2010). In the Oisans Massif, E-W shortening was
 162 responsible for folding in the cover with axial plane cleavage development in the west
 163 (Gratier and Vialon, 1980; Bellahsen et al., 2012; Bellanger et al., 2014). The metamorphic
 164 peak reached 300 to 350°C for 1-5 kbar (Waibel, 1990; Jullien and Goffé, 1993; Crouzet et
 165 al., 1999, 2001; Ceriani et al., 2003), which implies a tectonic overburden of around 10km
 166 when considering a normal geothermal gradient. The temperature in the Bourg d'Oisans area
 167 (Crouzet et al., 1999, 2001) reached 330°C after the end of deformation. The metamorphic
 168 peak was dated with large uncertainties around 26-24 Ma by K-Ar method in fine clay
 169 fraction from the cover (Nziengui, 1993), zircon fission-tracks ages ranging between 27 Ma
 170 (alt. 3065m) and 13 Ma (alt. 1310m) (van der Beek et al., 2010). A deformation stage was
 171 dated between 34 and 27 Ma by $^{40}\text{Ar}/^{39}\text{Ar}$ phengite ages from strike-slip shear zones related
 172 to E-W shortening in the eastern part of the Oisans Massif (Simon-Labric et al., 2009).
 173 Bellahsen et al. (2012) and Bellanger et al. (2014) have further proposed that the E-W
 174 shortening in the Oisans Massif was mainly accommodated by kilometer-scale top-to-the-
 175 west reverse shear zones, which are summarized in the next section.

176

177 ***2.2 Inherited or newly formed: the Oisans shear zones revisited***

178 Contrary to the internal zones where deformation is too intense to preserve former
 179 structures, the amount of shortening of the Oisans Massif was less than ca. 30-40% during
 180 the Alpine collision (Dumont et al., 2008; Bellahsen et al., 2012, 2014; Bellanger et al.,

2014). As a consequence, inherited structures are still preserved, which allows studying relationships between inherited and Alpine newly formed structures. The detailed structural evolution of the Oisans Massif, which is beyond the scope of the present paper, was recently investigated by Bellahsen et al. (2012) and Bellanger et al. (2014) in order to describe the distribution of the deformation in the basement.

The Oisans Massif presents unambiguous evidence of intense shortening either at small (outcrop) scale or at large (kilometric) scale. Maps of shortening axes of small-scale structures together with transport directions over the whole Oisans Massif reveal only one major E-W shortening stage (Fig. 2). Strain localization in the basement leads to the formation of a set of newly-formed, top-to-the-west, reverse shear zones that fold the basement-cover interface to the west and form large basement thrusts near the Penninic Frontal Thrust (Bellanger et al., 2014). Alpine shear zones can be separated from the Variscan deformation by the lower-temperature, more localised character of the deformation that cross-cuts a former high-temperature, pervasive Variscan fabrics. Moreover, an Alpine age is ensured by the deformation with consistent top-to-the-west kinematics within the Triassic-Liassic tegument (Bartoli et al., 1974; Barféty and Pêcher, 1984; Bellahsen et al., 2012; Bellanger et al., 2014).

While similar Alpine shear zones of Argentera or Mont Blanc massifs were recently dated from the Oligo-Miocene (Rolland et al., 2008; Sanchez et al., 2011a; Cenki-Tok et al., 2013), the only recent constraints on the timing of the Alpine deformation were performed on strike-slip shear zones of the eastern Oisans Massif (Simon-Labrie et al., 2009). The most prominent structures accommodating most of the shortening, W-directed reverse shear zones remain therefore undated.

3 Aims and methods

To put age-constraints on reverse shear zones developed during collision in the Oisans Massif, deformed rocks of four main shear zones were sampled across the W- E profile. These are, from west to east: the Col de Cluy, Plan du Lac, Col du Lac and Combeynot shear zones (Fig. 2), all sharing a common and consistent E-W to SE-NW stretching lineation and an overall top-to-the-west sense of shear. *In situ* $^{40}\text{Ar}/^{39}\text{Ar}$ laser ablation method (Maluski and Monié, 1988) was chosen as it allows linking radiochronology to petrology and particularly to textures/structures (Agard et al., 2002; Müller, 2003; Augier et al., 2005; Gerber, 2008). This link is particularly important as

different parameters such as temperature, deformation or fluid circulations may drastically impact the behaviour of the K/Ar system. It is therefore crucial to decipher the respective contribution of these parameters to isolate the signal related to deformation.

To ensure a reliable T record, temperatures were constrained at the scale of the Oisans Massif and the neighbouring areas using the Raman Spectrometry of Carbonaceous Material method (RSCM; Beyssac et al., 2002; Lahfid et al., 2010). These data were supplemented by local temperature estimates using the chemical composition of chlorite in the chlorite-quartz-water equilibrium (Vidal et al., 2001, 2005, 2006) either to control RSCM results or to put direct constraints within the shear zones. In parallel, impact of the deformation was evaluated by a combined textural/petrological study of the dated samples and particularly the behaviour of white-micas. Besides, the potential impact of the fluid percolations was evaluated by a careful study of output isotopic ratios. The next sections are dedicated to a brief description of the methods used in this study.

3.1 Raman spectrometry of carbonaceous material (RSCM)

The Raman Spectrometry of Carbonaceous Material method (Beyssac et al., 2002) is based on the quantitative study of the degree of graphitization of carbonaceous material (CM), which is a reliable indicator of metamorphic T. Because of the irreversible character of graphitization, CM structure is not sensitive to the retrograde reactions during exhumation and therefore depends on maximum T (noted T_{max}, thereafter) reached during metamorphism (Beyssac et al., 2002). T_{max} can be determined in the range 200-650°C by using two empirical geothermometers which link the degree of graphitisation measured from Raman Spectra (RS) to the highest temperature ever experienced by the rock (Beyssac et al., 2002; Lahfid et al., 2010). Lahfid et al. (2010) geothermometer was used when the D4 defect band was present in the RS, for the low-grade rocks. Absolute precisions of ca. 50°C between 650 and 300°C and ca. 25°C between 330 and 200°C are related to uncertainties on petrological data used for the calibration. However, relative uncertainties on T_{max} appear however much smaller, probably around 10-15°C and variations of that order of magnitude may be detected (Beyssac et al., 2004, 2007; Gabalda et al., 2009; Vitale-Brovarone et al., 2013).

All Raman spectra were measured and analysed following procedures described in Beyssac et al. (2002, 2004 and 2007) and Lahfid et al. (2010). Raman spectra were obtained using a Renishaw InVIA Reflex microspectrometer (IMPMC, Paris, France) with a 514 nm argon laser. RSCM analyses were conducted on thin sections prepared on CM-rich metasediments

particularly abundant in the cover series in the studied area, cut in the structural X-Z plane (i.e. orthogonal to foliation and parallel to lineation). To avoid defects on the CM related to thin-section preparation, analyses were all performed below the surface of the section by focusing the laser beam beneath a transparent crystal (i.e. dominantly quartz, calcite, albite and occasionally phyllosilicates). Before each session, the spectrometer was calibrated with a silicon standard. 10-20 spectra were routinely recorded for each sample to smooth out the inner structural heterogeneity of CM within samples.

3.2 Microprobe and independent temperature estimates

Mineral analyses were performed with a Cameca SX50 electron microprobe (ISTO-BRGM, Orléans). Analytic conditions were 15 kV accelerating voltage and 10 nA beam current using Fe₂O₃ (Fe), MnTiO₃ (Mn, Ti), diopside (Mg, Si), CaF₂ (F), orthoclase (Al, K), anorthite (Ca), albite (Na) as standards. Elements, which are measured in %weight of oxide, are SiO₂, Al₂O₃, Fe₂O₃, MgO, MnO, CaO, Na₂O, K₂O, TiO₂, F and Cl. Chemical composition of chlorite and the white-micas were particularly studied. Composition of chlorite has been used to retrieve independent T constraints using a thermodynamic approach (Vidal et al., 2001, 2005, 2006). Besides, composition of white-micas was used to distinguish different mineral generations to characterise the dated minerals.

Chlorite, which is abundant in the studied low-grade metamorphic rocks, was used as a geothermometer. Temperatures were estimated by two ways: i) traditional empirical geothermometry based on temperature dependency of the Si, Al and vacancy content of the tetrahedral and octahedral sites within chlorite (Cathelineau and Nieva, 1985; Cathelineau, 1988), and ii) a thermodynamical approach (Vidal et al., 2001, 2005, 2006) that integrates multiple substitutions (FeMg₋₁, Al_{IV}Al_{VI}Si₋₁Mg₋₁, (Mg, Fe²⁺)₃□₋₁Al₂ where □ is the vacancy). In the last case, metamorphic temperatures that prevailed during the formation of chlorite were estimated upon 4 equilibria at fixed water activity and pressure (a_w=1 and P=4kbar (Simon-Labric et al., 2009)). Standard deviations are routinely close to ±10°C (from ±4°C to ±32°C).

3.3 In situ ⁴⁰Ar/³⁹Ar dating

The ⁴⁰Ar/³⁹Ar *in situ* laser ablation method was first described by Schaeffer et al. (1977) and successively upgraded by York et al. (1981) and Maluski and Monié (1988). Procedure was detailed in Agard et al. (2002), Augier et al. (2005) or Turrillot et al (2011).

282 We recall here the main stages for $^{40}\text{Ar}/^{39}\text{Ar}$ *in situ* sample preparation and analytical
283 procedure.

284 Measurements were all performed at Géosciences Montpellier (France) with an analytical
285 system that consists of: (a) an IR-CO₂ laser of 100 kHz used at 5-15% during 60-80 sec; (b)
286 a beam shutter for selection of exposure times; (c) a lenses system for beam focusing; (d) a
287 camera connected to the monitor for beam focusing and selection of ablated zones; (e) a
288 steel chamber with a copper plate and the four samples on. This chamber is depressurized to
289 10^{-8} to 10^{-9} bar; (f) an inlet line for purification of gases including two Zr-Al getters; (g) a
290 Argus VI (Thermo-Fisher) multi-collector mass spectrometer from; (h) a software which
291 controls the timing of extraction/purification and the data acquisition. Rock sections of 1
292 mm thick, which had been used to make the petrographic thin-sections, were double
293 polished to c. 1 μm . Whole section and detailed-area photographs of both the rock section
294 and corresponding thin section were taken for an accurate selection of suitable areas for
295 laser experiments. All samples were ultrasonically rinsed in ethanol and distilled water,
296 wrapped in pure aluminium foils and then irradiated in the Triga Mark II nuclear reactor
297 (Pavia, Italia) with several aliquots of the Fisch Canyon sanidine standard ($28.03 \pm 0.08\text{Ma}$,
298 (Jourdan and Renne, 2007) as flux monitors. After irradiation, both the monitors and the
299 sections were placed on a Cu-holder inside the sample chamber and heated for 48 h at 150–
300 200 °C.

301 For each experience, argon was released from a 100 x 300 μm mean surface corresponding
302 very often to aggregates of white micas. Incision of the sample did not exceed 10-20 μm
303 depending on the three-dimensional orientation of the white mica crystals. For each
304 experiment, ages have been obtained after correction with blanks, mass discrimination,
305 radioactive decay of ^{37}Ar and ^{36}Ar and irradiation-induced mass interferences. They are
306 reported with 1σ uncertainty and were evaluated assuming an atmospheric composition for
307 the initially trapped argon [i.e. $(^{40}\text{Ar}/^{36}\text{Ar})_i = 295.5$]. The proportion of ^{38}Ar and ^{37}Ar
308 isotopes that result from interference with chlorine and calcium during irradiation reveals
309 the contribution of Cl and Ca-rich minerals or inclusions to the overall argon release and so
310 to the calculated age. A blank analysis was done every two or three analyses to evaluate the
311 argon background within the whole analytical system. Isotopic data are computed with the
312 ArArCalc freeware (Koppers, 2002) based on the algorithms described in (York, 1968;
313 Faure, 1986; McDougall and Harrison, 1988; Renne et al., 1998; Min et al., 2000).

314 Along with a strong dependency with temperature and deformation, the behaviour of the
315 K/Ar system in white micas can be partially altered by fluid circulations. As fluid

316 circulations are hardly detected *a priori*, output isotopic ratios ($^{36}\text{Ar}/^{39}\text{Ar}$, $^{37}\text{Ar}/^{39}\text{Ar}$ and
317 $^{38}\text{Ar}/^{39}\text{Ar}$ ratios) were carefully checked.

318

319 **4 Temperature record within the sedimentary cover**

320

321 **4.1 A large-scale RSCM Tmax cover in the sedimentary series**

322

323 ***First order thermal features***

324 Our dataset consists of 70 samples collected across the external zones, from Vif to
325 the Penninic Frontal Thrust (E-W transect) and from Saint Jean de Maurienne to Chorges
326 (N-S transect) in various lithologies and all structural units of the studied area except
327 Variscan basement rocks (Table 1, Fig. 4). A more systematic sampling was performed
328 along the E-W transect where the thermal structure is so far poorly documented. Results are
329 all presented in maps of various scales and projected along structure-orthogonal cross-
330 sections. For the sake of comparison, results of previous studies including illite-crystallinity
331 maps (Aprahamian, 1974; Barlier et al., 1974) or RSCM data in bounding areas (e.g.
332 Gabalda et al., 2009) are also shown (Fig. 4).

333 At first glance, an unexpected major result of this study is the relative homogeneity of the
334 Tmax value at the scale of study area, with an average value of ca. 330°C (329±8 for 27
335 Tmax values along an E-W cross-section), from the Lower Jurassic series close to Vif, all
336 the way to the Galibier pass, at the eastern edge of the study area (Fig. 4, 5). Conversely,
337 lower Tmax ca. 200-220°C is recorded to the south either in the Mesozoic series or Eocene
338 flysch (north of Saint Michel de Chaillol; Fig. 4). Moreover, Tmax results recorded further
339 south along an E-W section between Chorges and l'Argentière-la-Bessée (Fig. 4) highlight a
340 drastic southward lateral change in the thermal structure. There, Tmax values present an
341 overall upward temperature increase from 253±10°C to more than 330±5°C in the vicinity of
342 the Penninic Frontal Thrust. Besides, additional important temperature variations were
343 recorded either to the west of the study area (Belledonne and La Mure massifs), to the east
344 in the vicinity of the Penninic Frontal Thrust (Fig. 4, 5). These variations are described here
345 in more details.

346

347 ***Thermal structure of the western bound of the Belledonne and La Mure massifs***

348 To the west of the Belledonne and La Mure massifs, Tmax values decrease upward
349 the stratigraphic succession from 334±6°C-329±8 in the Liassic, 325±6-286±5°C in

calcareous Dogger and 227 ± 23 – $<200^\circ\text{C}$ in upper Jurassic black-shales (Fig. 4, 5) highlighting a ca. 130°C temperature variation. In details, T_{max} values gradually decrease sequence along a ca. $30^\circ\text{C}/\text{km}$ gradient but also highlight a localised temperature decrease near the top of the calcareous Dogger (296 ± 5 – $286\pm 5^\circ\text{C}$) and the base of the upper Jurassic (227 ± 23 – $204\pm 14^\circ\text{C}$). This ca. 80°C temperature gap can be followed 40 km strike, which is consistent with a normal sense west-dipping tectonic contact located within the upper Jurassic black-shales (Oxfordian). However, due to poor local outcrop conditions, this contact has not been observed in the field.

In the same area, the base of thick Jurassic sequences of the Vercors yielded T_{max} value close to 335°C ($334\pm 6^\circ\text{C}$) while the thin Jurassic cover of the Belledonne and La Mure massifs yielded only ca. 280°C ($279\pm 9^\circ\text{C}$). In this case, T_{max} distribution evidenced an east-verging shear zone/thrust consistent with the description of a large-scale backthrust in the field by Butler (1989). The Mesozoic sequence of the Vercors therefore appears i) transported eastward above the La Mure thrust which roots in Triassic or Liassic units, and ii) is partly dismembered by an extensional structure (Fig. 4, 5).

365

366 ***Thermal structure in the vicinity of the Penninic Frontal Thrust***

To the east of the study area, close to the Lautaret and Galibier passes, T_{max} values decrease gradually from ca. 330°C in the easternmost Mesozoic cover of the Oisans Massif to ca. 260°C in the Helmintoids flysch (Tête Noire area) despite the presence of numerous tectonic contacts including the PFT. T_{max} results therefore align well along an apparent 20 – 30°C.km^{-1} gradient suggesting either that the T_{max} imprint was post-tectonic or that extensional movements are not restricted to the PFT and rather distributed over numerous small-scale, normal-sense structures. In any case, in their current position, Briançonnais units did not experience T_{max} higher than ca. 270°C which fall well below the ca. 330°C temperatures characterising the internal Dauphinois. Conversely, a ca. 40°C localized temperature drop is recorded further south, across the PFT (around L'Argentière-la-Bessée, Fig. 4) where the Nummulitic series yielded ca. $330\pm 5^\circ\text{C}$ while Briançonnais units only experienced $291\pm 6^\circ\text{C}$. There, this result is consistent with both field and low-temperature thermochronology evidences of the reactivation of the PFT as an extensional structure (Tricart et al., 2001, 2004, 2006, 2007; Ceriani and Schmid, 2004).

381

382 ***4.2 Independent temperature record using chlorite geothermometry***

Four samples were selected from the sample collection of the sedimentary cover for the RSCM approach (see location on Table 2 and on Fig. 2). The main aim of this approach is to estimate the temperature recorded by the cover series using independent geothermometers based on the chemical composition of chlorites (Cathelineau and Nieva, 1985; Cathelineau, 1988; Vidal et al., 2001, 2005, 2006). Results are all presented on figure 6.

Chlorites present different habits and occur in various textures. They are either dispersed in the matrix (samples A-2-2, A-4-232, see location Fig. 2), crystallized within veins (samples A-1-345(2)) or along the foliation and top-to-the-W shear bands (sample A-1-321). Irrespectively to their structural position and to the potential impact of the protolith bulk chemical composition, the two geothermometers yielded quite consistent results for all samples within both the analytical and calibration errors and the within-sample internal dispersion. However, the thermodynamical approach generally yield more dispersed temperature results between ca. 210 and 350°C than empirical geothermometry that show an unimodal distribution centred on the narrow ca. 270-330°C range (Fig. 6). Temperature ranging from 150 to 350°C was retrieved from sample A.2.2, 150-340°C from sample A-4-232, 230-380°C for sample A-1-321 and 240-390°C for sample A-1-345 (Fig. 6). Interestingly, RSCM T_{max} fall systematically within the high temperature value indicating i) the cross-validation of the methods and that ii) chlorites, at variance with the RSCM, do record other discrete metamorphic increments. However, textural observations did not allow distinguishing the relative timing of successive chlorite growth events during the metamorphic evolution (prograde or retrograde (re-)equilibrations).

5 Phyllosilicate habits, chemical composition and chlorite temperature (T_{Chl}) record for dated samples

The crystallisation-deformation relationships of chlorite and white-micas that were first recognised in the field were then strengthened by detailed textural observations in thin-section. Mineral chemical compositions were used to constrain i) the chemistry of dated minerals and ii) the physical conditions by chlorite geothermometry. Together with textural observations, these results are of prime importance to gather accurate *a priori* constraints and discussion elements for *in situ* laser-probe investigations.

5.1 First order common features: guidelines for the $^{40}\text{Ar}/^{39}\text{Ar}$ in situ approach

At the scale of the outcrop, rocks present routinely large white-mica crystals. Conversely, aggregates of very fine-grained white-micas occur only in deformed zones. At the scale of thin-sections, deformed samples present two contrasted habits of white-micas. Large, up to 2mm variously oriented white-mica flakes appear closely associated with a former high-temperature foliation and high-temperature parageneses (Fig. 7). These crystals coexist with aggregates of fine-grained white-micas growing either at the expense of former large crystals or in cracks of various scales within feldspars. Besides, this population presents a close association with deformation structures and particularly with top-to-the-west shear bands within proto-mylonites, mylonites or phyllonites. This strong correlation has been observed at the scale of strain gradients along which large, well-developed white-mica clasts tend to disappear whereas fine-grained blasts form the frame of the rock (Fig. 7). Interestingly, EPMA analyses reveal the existence of two texturally controlled groups of white-mica of contrasted compositions reminding the results of Lanari et al. (2012) in the Briançonnais domain, further east. The first group, which plots close to the muscovite end-member composition, is mainly represented by the large white-micas clasts population carried by an earlier foliation (Fig. 7). Besides, the Na-content of this population is systematically higher than 0.05 a.p.f.u. (atoms per formula unit) while the Si content is lower than 3.1 a.p.f.u. (Fig. 7). The second group, mostly corresponding to fine-grained aggregates of white-micas, has a phengite chemistry, i.e. between muscovite and celadonite end-member due to Tschermak substitutions. This group presents variable silica-content scattered between 3.2 and 3.6 a.p.f.u. and Na-content systematically lower than 0.05 a.p.f.u. (Fig. 7, 8). However, the small size of crystals and large internal variations of compositions suggest a partial mixing between two end-members.

EPMA analysis of the chemical compositions confirm the presence of two unrelated generations of white-micas recognised on the basis of textural observations and characterized by contrasted chemical compositions. Being conspicuously involved with the greenschist facies partial reequilibration of former high-temperature parageneses and closely associated with a top-to-the-W asymmetry, phengites therefore appear as a synkinematic mineral particularly suitable to put age-constraints on the deformation. Conversely, the habits of pre-kinematic clasts and the muscovite composition of most of the large white-mica population are likely inherited crystals pertaining to Variscan parageneses.

Besides, temperature, which is the most important factor controlling Argon diffusion, must be estimated to make a difference between cooling and deformation (crystallisation) ages. Similarly to temperature estimates performed in the cover series, temperature recorded by basement rocks was estimated using two different chlorite geothermometers. Results are all presented in figure 8.

5.2 Textural and chemical characteristics of dated samples

Following this structural and chemical guideline, white-micas present in the dated samples were carefully studied. Besides, chemical composition of chlorite was used to constrain the temperature experienced by the rocks. Results are all presented from west to east on figure 8.

Col de Cluy shear zone

Samples from Col de Cluy are Variscan micaschists and gneisses with greenschist to amphibolite facies associations with quartz, white-micas, feldspar, biotite and/or chlorite. Rocks show generally only an incipient recrystallization. Former biotite and K-feldspar are still preserved and locally abundant. White-micas, which are present in all samples, mostly occur as ca. 1-2mm weakly deformed flakes and locally as clasts involved in shear bands. This population coexists with fine-grained white-micas that occur as rare isolated aggregates. Grain-size is generally less than 10 x 50 μm and white-mica crystals are often closely associated with chlorite. EPMA analyses of the white-micas mostly plot within a rather narrow composition range close to the muscovite end-member. Analyses of fine-grained white-micas plot along continuous pattern from muscovite (Si(3.05); Na(0.09;0.12)) to low-silica content phengite (Si(3.15;3.25); Na(0;0.02)) (Fig. 8). Chlorite occurs as numerous small-scale aggregates. Chlorite shows variable composition with 70 to 85% of clinocllore and XMg values close to 0.45 for variable silica content. Geothermometric results yield quite contrasting results. While the thermodynamic approach yielded dispersed temperature between 150 to 370°C, empirical geothermometry showed quite consistent ca. 230 to 320°C temperature results (Fig. 8). Being devoid of conspicuous Alpine re-equilibrations, these samples appear, at a first glance poorly suitable to allow dating the deformation and were therefore not used for the $^{40}\text{Ar}/^{39}\text{Ar}$ *in situ* approach.

483 *Plan du Lac shear zone*

484

485 Thanks to good outcrop conditions allowing the recognition of the strain distribution,
 486 the Plan du Lac shear zone was particularly studied. Four samples were selected over the
 487 strain gradient. Sample A-4-74b is a proto-mylonite sample showing a rather weak Alpine
 488 deformation by comparison to samples A-4-68 and 68b that are phyllonites. Sample A-4-66b
 489 was selected from an intermediate position within the strain gradient. All samples show a
 490 clear top-to-the-west asymmetry except sample A-4-66b picked along a localized top-to-the-
 491 E shear zone.

492 All samples derive from a quartz, feldspar, biotite and white mica Variscan gneiss.
 493 Chlorite pervasively replaces biotite either as static pseudomorphs in low strain domains or
 494 as synkinematic crystallisations in mylonites. In low strain domain, white mica occur as
 495 large, up to ca. 1-2mm clasts with clear disequilibrium textures while fine-grained white-
 496 mica are rare and clearly secondary. Conversely, mylonite and phyllonite are almost devoid
 497 of large clasts and exclusively made of fine-grained white-micas. EPMA composition of
 498 white-micas reveals two distinct poles with very few intermediate values (Fig. 8). The first
 499 pole that corresponds to muscovite (Si(3.15); Na(0.06;0.08)) and the second one to a silica-
 500 rich phengite (Si(3.2;3.4); Na(0;0.02)) (Fig. 8). Interestingly, mineral composition shows a
 501 conspicuous dependency to deformation. Mylonite and phyllonite samples yield mostly
 502 phengitic white-micas. Conversely, less-deformed proto-mylonite sample yield texturally-
 503 controlled chemical compositions with muscovite compositions for the large clasts and
 504 phengite composition for the fine-grained blasts.

505 Chlorite is locally abundant. It occurs in the matrix, growing at the expense of
 506 biotite, crystallized in syn-deformation veins or associated with fine-grained white-micas
 507 along shear planes. Most analyses yield consistent results with 70 to 90% of clinocllore
 508 content, an XMg close to 0.45 for a variable silica content (Fig. 8). Regardless of strain
 509 intensity, structural position and habits, most results cluster between 260 to 290°C (Fig. 8).

510

511 *Col du Lac shear zone*

512

513 Sample A-1-355 was collected from the core of a top-to-the-W shear band
 514 characterized by a thin layer of phengite-quartz phyllonites. This sample derives from a
 515 Variscan high-grade gneiss still preserved a few decimetres apart on the same outcrop. The
 516 rock is made of quartz, feldspar and biotite partly turned into chlorite. Interestingly, white

mica is absent from the mineralogical composition of the protolith. Small-scale but abundant white-mica aggregates from the phyllonites are therefore newly-crystallized as the result of strain and/or fluid-induced destabilization of the K-feldspar (e.g. Gueydan et al., 2003). Accordingly, EPMA analyses of the white-micas all yielded phengitic compositions with a very low dispersion (Si(3.2;3.4); Na(0;0.02)).

Chlorite occurs as small-scale aggregates and rare dispersed larger crystals that may derive from former biotites. Chlorite shows almost constant composition close to the clinocllore end-member with XMg values around 0.6. Most geothermometric results cluster between 250 and 290°C for both geothermometers (Fig. 8).

Combeynot shear zone

Sample A-1-345 was selected in the most deformed outcropping ductily-deformed basement rocks of the Combeynot shear zone. Mylonites are scarce and proto-mylonites, derived from orthogneiss, which belongs to a volcano-sedimentary complex (Barbier et al., 1973), are often directly overthrust onto the sedimentary cover. White micas occur generally as large clasts surrounded by very fine-grained blasts. Aggregates of fine-grained white-micas also occur in tension gashes in K-feldspar. Composition of white-micas shows a rather continuous pattern from muscovite (Si(3.05;3.15); Na(0.06;0.08)) to phengite (Si(3.25;3.5); Na(0;0.02)) (Fig. 8). Despite a clear partial re-equilibration within Alpine physico-chemical conditions, Alpine white-micas occur as very fine grained aggregates. Chlorite is rare, small and often closely associated with white-micas. Thus, most chlorites show mixed composition with white-micas (i.e. K-rich chlorite) and were thus discarded. Data are thus only presented in table 2.

Synkinematic fine-grained aggregates of phengite appear particularly adapted to put age-constraints on the activity of the shear zones. Accordingly, samples from Col de Cluy shear zone which may record only a weak Alpine re-equilibration have not been selected from the *in situ* $^{39}\text{Ar}/^{40}\text{Ar}$ approach. For all other samples, success of the study relies on our ability to effectively analyze newly formed crystals whose typical size is often small (150µm max.).

Temperatures recorded in the shear zones yielded very consistent results close to 250-300°C which falls below RSCM T_{max} recorded in the cover series. Chlorite therefore probably recorded retrograde metamorphic increments during exhumation and cooling of the Oisans

Massif. It is noteworthy that temperatures recorded at peak-T conditions also fall below the lower bound of the closure temperature of white-micas (e.g. Dodson, 1973; Cliff et al., 1998; Villa, 1998).

554

555 **6 $^{40}\text{Ar}/^{39}\text{Ar}$ age constraints**

556 Our dataset consists of 199 *in situ* $^{40}\text{Ar}/^{39}\text{Ar}$ ages performed on 7 samples collected
557 over three main Alpine shear zones, namely (from west to east), the Plan du Lac, Col du
558 Lac, and the Combeynot shear zones (Table 3, see location on Fig. 2). To ensure the
559 coverage of the different strain facies recognized in the field, a more systematic sampling
560 was performed over the Plan du Lac shear zone. Results of the *in situ* $^{40}\text{Ar}/^{39}\text{Ar}$ approach are
561 all shown and discussed in Fig. 9 as apparent ages for which an initial $^{40}\text{Ar}/^{39}\text{Ar}$ ratio of
562 295.5 is assumed for the non-radiogenic argon component. These age results are described
563 in the light of the textural/structural observations, chemical composition results and the
564 output isotopic ratios ($^{36}\text{Ar}/^{39}\text{Ar}$, $^{37}\text{Ar}/^{39}\text{Ar}$ and $^{38}\text{Ar}/^{39}\text{Ar}$) informative of potential
565 disturbances of the K/Ar system. For small-scale phengite aggregates located in the vicinity
566 of chlorite, calcium or fluid inclusions-bearing minerals, the release of the ^{38}Ar and ^{37}Ar
567 isotopes interfering with chlorine and calcium revealed that argon was extracted from both
568 phengites and K-poor minerals (Brereton, 1970; McDougall & Harrison, 1988). The quoted
569 errors represent one sigma deviation and do not include uncertainty on the monitor. Errors
570 are routinely inferior to ca. 1 Ma while larger error (ca. 2 to 15 Ma) are generally due to both
571 a weak argon signal and a high atmospheric contribution (Table 3).

572 At a first glance, an important conclusion is that the successive phengite generations
573 recognized on the basis of textures (e.g. Fig. 7), compositions (Figs 8) effectively yield
574 successive ages. Fine-grained aggregates of synkinematic phengites yield Alpine ages *sensu*
575 *lato*. Conversely, white-micas characterized by muscovite chemical compositions and/or
576 presenting clear pre-kinematic (i.e. inherited) habits yielded pre-alpine ages or even
577 sometimes Variscan ages. Interestingly, a clear dependency to the intensity of the
578 deformation may even be detected either at the scale of the sample or at the scale of the
579 strain gradient within a shear zone (e.g. Fig. 10). Therefore, most ages are tentatively
580 interpreted either as, at least partially, re-opened Variscan isotopic systems for “old” ages or
581 as genuine crystallization ages for the “young” ca. 35-25 Ma ages. More locally, results
582 however show important age variations that may also record fluid percolations.

583

584 **6.1 Plan du Lac shear zone**

585 In order to study the potential impact of the finite Alpine deformation, four samples
 586 were selected over a single strain gradient toward the core of the Plan du Lac shear zone.
 587 The relative structural position of samples is given on figure 10.

588

589 *A-4-74b (Proto-mylonite)*

590 In this sample, discrete ca. 1mm-thick shear bands partially overprint a high-
 591 temperature inherited texture composed by large, brittlely fractured feldspar clasts and more
 592 ductilely stretched quartz and muscovite crystals (Fig. 10). Conversely, this sample contains
 593 Alpine phyllosilicates including phengite and chlorite either carried along a weak vertical
 594 foliation, or crystallized along top-to-the-W shear bands. A total of 40 apparent *in situ*
 595 analyses were performed to cover all situations. Apparent *in situ* ages are scattered between
 596 207 ± 5.2 Ma and 24.3 ± 1.8 Ma (Fig. 10). 22 analyses corresponding to the oldest ages are
 597 correlated with high $^{38}\text{Ar}/^{39}\text{Ar}$ ratio and atmospheric content (Table 3) but ages fall within
 598 the age range of uncontaminated phengites. Interestingly, the core of the shear bands
 599 provides ca. 25 Ma consistent ages (24.3 ± 1.8 Ma and 26.7 ± 0.6 Ma).

600

601 *A-4-66b (Top-to-the-E Mylonite)*

602 This sample is characterized by a schistosed high-grade gneiss traversed by a 3 mm-
 603 thick top-to-the-E shear band (Fig. 10). While Variscan fabrics are still well developed in
 604 most of the rock volume, Alpine synkinematic recrystallizations are almost complete within
 605 shear bands. There, former muscovites and feldspars are absent and a ca. 2mm thick
 606 phengitized fringe is developed (Fig. 10). 34 *in situ* analyses were performed. No
 607 relationship between apparent ages and the $^{37}\text{Ar}/^{39}\text{Ar}$ and $^{38}\text{Ar}/^{39}\text{Ar}$ variations were detected,
 608 except for the 3 three apparent ages older than ca. 100Ma, which have high $^{38}\text{Ar}/^{39}\text{Ar}$ values
 609 (Table 3). Analyses show a first order textural control of the age. Large muscovites, at least
 610 partially preserved within the matrix yielded 301.4 ± 6.6 and 62.3 ± 1.4 Ma. Conversely,
 611 small-scale phengites within the shear band yielded 30 alpine ages ranging between $40.5 \pm$
 612 1.1 Ma and 29.5 ± 0.6 Ma plus an older age of 47.6 ± 1.2 Ma (Fig. 10). Interestingly, while
 613 ages depict a clear first order younging toward and within the shear band, older ages occur
 614 within the core of the shear band over a ca. 1-2mm-thick fringe (Fig. 10). In this case, a late
 615 gradient of ^{40}Ar excess seems to overprint an initial age gradient related to deformation.

616

617 *A-4-68 (Phyllonite)*

Sample A-4-68 is a phyllonite sample. 15 *in situ* analyses were performed and yielded ages ranging between 93.5 ± 2.0 and 22.4 ± 0.9 Ma (Table 3). Most of apparent ages cluster between 32 and 22 Ma (Fig. 10). There is no relationship of apparent ages with $^{37}\text{Ar}/^{39}\text{Ar}$ unlike with $^{38}\text{Ar}/^{39}\text{Ar}$, suggesting some interaction with chlorine-rich fluids that may occur as trapped inclusions (Fig. 10).

A-4-68b (Phyllonite)

Sample A-4-68b was cut in two distinct rock slabs (section 8 and 5) selected only a few centimeters distant; section 5 being a little less deformed than section 8. Analyses performed on section 5 yielded 21 apparent ages, ranging between 114.2 ± 4.3 and 25.9 ± 0.5 Ma (Fig. 10). Analyses present a strong relationship between apparent ages and $^{37}\text{Ar}/^{39}\text{Ar}$, $^{38}\text{Ar}/^{39}\text{Ar}$ and atmospheric content. In the absence of Ca-rich and Cl-rich phases in the chip, contamination may be rather related with fluid inclusions trapped in the analysed phengites (Table 3). Despite contamination by Ca-rich and Cl-rich fluids, analyses yielded ages of about 25-27 Ma. Interestingly, analyses performed on section 8 yielded 44 apparent ages, which do not show any evidence of a direct relationship with the $^{37}\text{Ar}/^{39}\text{Ar}$, $^{38}\text{Ar}/^{39}\text{Ar}$ and atmospheric content. These ages are scattered between 37.2 ± 1.0 Ma and 23.2 ± 0.6 Ma (with two old ages of 46.4 and 47.2 Ma) and define mainly two age groups at 31-33 Ma (7 ages) and 25-27 Ma (18 ages) (Fig. 10).

6.2 Col du Lac shear zone (Sample A-1-355, phyllonites)

Sample A-1-355 was collected from the core of a top-to-the-W shear band characterized by a ca. 1 meter-thick layer of phyllonites. This sample derives from a Variscan high-grade gneiss still preserved a few decimetres apart on the same outcrop and turned into a phengite-quartz rock in the most deformed zone. It contains few chlorites and no remnant of inherited muscovites present in the undeformed protolith. The 30 apparent ages are scattered between 110 ± 7 Ma and 15 ± 10 Ma and 25 of them range between 42.68 ± 2.22 and 23.88 ± 1.6 Ma (Table 3). No correlation between the apparent age and $^{38}\text{Ar}/^{39}\text{Ar}$ and $^{37}\text{Ar}/^{39}\text{Ar}$ ratios was observed. However, some analyses, particularly the “old”, up to 50 Ma ages present high $^{38}\text{Ar}/^{39}\text{Ar}$ values and/or an important atmospheric contribution (Table 3). Interestingly, these old ages were found in micas from C'-type shear bands and therefore argue for a ^{40}Ar excess. These results are quite similar to those obtained in the most deformed parts of sample A-4-66b. Despite this potential contamination, analyses with low

$^{38}\text{Ar}/^{39}\text{Ar}$ ratio and low atmospheric content yielded 15 internally consistent apparent ages scattered between 32.9 ± 0.9 and 24.8 ± 0.6 Ma (Fig. 10).

6.3 Combeynot shear zone (*Sample A-1-345, proto-mylonite*)

Sample A-1-345 was selected in the most deformed outcropping ductilely-deformed basement rocks of the Combeynot shear zone. Mylonites are scarce and protomylonites, derived at the expense of orthogneiss (former leucogranite) are often directly overthrust onto the sedimentary cover. It is the less deformed sample of the data set. None of the 15 *in situ* analyses provide Alpine age (Table 3, Fig. 10). Instead, apparent ages are distributed between 389 ± 27 Ma and 156 ± 3 Ma with no clear textural control. Part of the apparent ages is correlated with the $^{38}\text{Ar}/^{39}\text{Ar}$ ratio, which suggests a possible contribution of Cl-rich phases or fluid inclusions. High $^{37}\text{Ar}/^{39}\text{Ar}$ ratios (Table 3) are probably due to the degassing of Variscan epidote or titanite, but again no systematic correlation between the variation of these ratios and the age dispersion was observed.

7. Discussion

7.1 Reliability, significance of the RSCM *T*_{max} mapping

RSCM *T*_{max} estimates has proven an efficient mean to put extensive temperature constraints on the thermal structure of coherent units or on the importance of tectonic contacts (Beyssac et al., 2002; Augier et al., 2005; Negro et al., 2006; Gabalda et al., 2009; Angiboust et al., 2011). In the Oisans Massif, RSCM *T*_{max} fall in the same temperature range as more punctual results issued from other methods available in literature. The most important result is the uniformity of *T*_{max} values at the scale of study area, with an average value of ca. 330°C with a relative error that may not exceed 10-15°C (Beyssac et al., 2004). Thermopaleomagnetic studies indicate that cooling started from 335°C in the north of the Emparis formations and in the Bourg d'Oisans basin, after the main tectonic phase at ca. 25 Ma (Ménard and Rochette, 1992; Crouzet et al., 1999, 2001) (Fig. 11). The stability field of paragenesis observed in the Jurassic cover (quartz-cookeite-margarite-pyrophyllite) that most likely recorded peak-metamorphism conditions in the Emparis basin is restricted from 270 to 340°C and from 1 to 5kbar (Jullien and Goffé, 1993) (Fig. 11). Besides, the overall

684 thermal structure of the Oisans Massif as recorded by the distribution of RSCM $T_{\max}$ is
 685 consistent with illite-crystallinity maps (Aprohmanian, 1974; 1988; Barlier et al., 1974) (Fig.
 686 4).

687 Related to the irreversible character of graphitization, and the insensitivity of the CM
 688 structure to retrograde reactions (e.g. Beyssac et al., 2002; Lahfid et al., 2010; Aoya et al.,
 689 2010), RSCM record the highest temperature ever undergone by the rock sample. In
 690 principle, RSCM $T_{\max}$ results issued from a continuous section may therefore be
 691 diachronous and represent the superimposition of distinct thermal events.

692 In the Bourg d'Oisans and Emparis basins, metamorphic peak-T conditions of ca. 335°C
 693 where reached at ca. 25 Ma (i.e. Crouzet et al., 1999, 2001). As the restored sedimentary
 694 cover of this area is generally thinner than ca. 2 km (Barféty et al., 1972; Barbier et al.,
 695 1973; Crouzet et al., 2001; Bellahsen et al., 2012), these temperatures are clearly ascribed to
 696 an Alpine tectonometamorphic event and particularly to the tectonic burial of the
 697 Dauphinois domain under internal units (Crouzet et al. 2001; Dumont et al., 2008). Besides,
 698 the progressive temperature decrease toward the top of the overthrusting units close to the
 699 Galibier pass is consistent with a late to post-tectonic temperature imprint, as suggested by
 700 thermopaleomagnetic results (Crouzet et al., 1999, 2000). Interestingly, the upward RSCM
 701 $T_{\max}$ evolution indicates a ca. 20-30°C.km⁻¹ gradient (Fig. 4C), which is consistent with the
 702 gradient around 20 to 25°C.km⁻¹ proposed by Crouzet et al. (1999) but lower than the ca. 40-
 703 50°C.km⁻¹ gradient deduced from fluid inclusion data (Gratier et al., 1973; Nziengui, 1993)
 704 (Fig. 12). However, high geothermal gradients deduced from fluid inclusions may result
 705 from an underestimated pressure due to reequilibration of fluid inclusion densities during an
 706 isothermal decompression as proposed by Boullier (1999) or Crouzet et al. (2001).

707 Conversely, in the sedimentary cover west of the Belledonne Massif, $T_{\max}$ values show
 708 large temperature variations from 335°C to less than 200°C (Fig. 4, 5). Tilted with the
 709 bedding, this ca. 130°C $T_{\max}$ upward sequence decrease may witness a post-metamorphic
 710 peak westward tilting of the sedimentary succession (Fig. 4, 5). Moreover, $T_{\max}$ recorded at
 711 the base of the succession is consistent with the ca. 8 km thickness of the Mesozoic series
 712 assuming a ca. 40°C.km⁻¹ geothermal gradient (Philippe et al., 1998). Besides, this result is
 713 consistent with the 60 to 80 mW.m⁻² basal heat flow estimated in this area (Deville and
 714 Sassi, 2006) and a bulk thermal conductivity close to 1.5 W.m⁻¹.K⁻¹ (Beardsmore and Cull,
 715 2001).

716 Minor and local fluid circulations along the basement-cover interface cannot be ruled out
 717 but an overall thermal imprint related to large-scale fluid circulations is unlikely as available

studies show that the fluid flow operated only at small scale in the studied area (Henry et al., 1996; Kirschner et al., 1999; Boutoux et al., 2014b).

7.2 Significance of the $^{40}\text{Ar}/^{39}\text{Ar}$ in situ ages

Geochronological data and particularly $^{40}\text{Ar}/^{39}\text{Ar}$ data are traditionally interpreted as recording closure of the isotopic system at a specific temperature depending on the isotopic diffusion properties of the mineral (e.g. Dodson, 1973; Cliff et al., 1998; Villa, 1998; Harrison et al., 2009). Accordingly, many studies therefore assume that the closure temperature of the K/Ar system for muscovite argon retention is around 350-450°C depending on cooling-rates, grain-size and other controlling factors such as the geometry of the diffusion domain (e.g. Wijbrans & McDougall, 1986; Hames & Bowring, 1994; Reddy et al., 1996; Villa, 1998). Using recent results of modeling, the closure temperature for muscovite would fall in the range of 400-420°C for a 10-30°C.Ma⁻¹ cooling-rate and a diffusion radius of 100µm (Harrison et al., 2009). However, closure temperature for phengite is still discussed as it may largely vary as a function of both intrinsic (e.g. grain-size) and extrinsic (cooling-rates, temperature, pressure and deformation or fluid circulations) parameters. Recent studies suggest that phengite is more retentive for argon than muscovite and that phengite ages are interpretable as crystallisation ages to temperature as high as 500-550°C (e.g. Dunlap, 1997; Agard et al., 2002; Müller, 2003; Mulch and Cosca, 2004; Augier et al., 2005; Mulch et al., 2005; Gerber, 2008; Rolland et al., 2008, 2009; Sanchez et al., 2011a). Temperature experienced during the Alpine orogeny by dated samples did not exceeded 330°C ± 30°C since they initially cooled at the end of the Variscan Orogeny. Ages of the fine-grained synkinematic phengites therefore likely represent crystallization ages. Conversely, large, inherited clasts of muscovite may reflect Variscan cooling ages in the absence of additional disturbing factors. Stratigraphic constraints indicate that the flexural basin of the external zones was diachronously filled up from Lutetian to Bartonian (45 to 33Ma) (Du Fornel et al., 2004; Mulder et al., 2010; Salles et al., 2011). Burial of this basin below internal units occurred during Rupelian (34 to 28 Ma; Fry, 1989; Sinclair, 1997; Ford and Lickorish, 2004; Mulder et al., 2010). In Champsaur and Saint-Antonin basins, amphiboles reworked from Oligocene volcanism coeval with deposition yielded ca. 34-31 Ma Priabonian ages by $^{40}\text{Ar}/^{39}\text{Ar}$ methods (Féraud et al., 1995; Ivaldi et al. 2003), giving a maximum age for the burial of the basins under the

internal nappes. In addition, part of the Oisans Massif formed a 500 m high relief in the Oligocene (Gupta and Allen, 2000) and dated samples, restored in their inferred initial position, were probably less than 2km deep, between ca. 40 and 34 Ma. It therefore appears unlikely that phengites were formed under these P-T conditions and the ca. 34-33 Ma age clearly fixes the upper bound of the possible ages for the deformation of the Oisans Massif. 97 *in situ* experiments yielded ages younger than 34 Ma.

Interestingly, young ages are clearly correlated with the intensity of fine-grained white-micas blastesis particularly important in high-strain zones. It is therefore proposed that ages younger than 34 Ma reflect synkinematic crystallisation ages and thus deformation ages, in a broad sense. In addition, apparent ages show a bimodal distribution with a first age group settled at ca. 33-30 Ma and a second group at ca. 28-25 Ma ages (Fig. 11). These results are consistent with results of a conventional $^{40}\text{Ar}/^{39}\text{Ar}$ approach performed on wrench shear zones of the same area (Simon-Labric et al., 2009). However, our interpretation differs. Based on age spectra scattered between 32 and 24 Ma, Simon-Labric et al. (2009) proposed the superimposition of two distinctive shearing events (one at ca. 32 and one ca. 24 Ma). In this study, the use of an *in situ* mean over strain gradients allow linking deformation ages and the amount of finite strain experienced by the rocks. Accordingly, proto-mylonite and mylonite samples yielded older, ca. 33-30Ma ages while highly strained samples (phyllonites) yielded younger ca. 28-25 Ma ages, suggesting that the deformation progressively localized along the phyllonite bands thus leaving large rock volumes devoid of further recrystallizations. In any case, intense E-W shortening started with or just after the burial of external zones under the internal nappes around ca. 34-33 Ma and was still active at ca. 25 Ma. Synchronous deformation ages of either top-to-the-W reverse shear zones or wrench shear zones thus confirm the small-scale mutual cross-cutting relationships observed in the field (Bellanger et al., 2014).

102 *in situ* experiments yielded ages mostly clustering between 100 and 40 Ma with ages as old as 380 Ma. Older than ca. 34 Ma and presenting sometimes pre-Alpine or even genuine Variscan ages, these apparent ages clearly do not record alpine collisional deformation. Relying on temperature estimates throughout the studied area, the temperature concept predicts that all white-micas sharing common pre-Alpine textures and inherited muscovite chemical composition should yield the same age corresponding to the time elapsed since initial cooling at the end of the Variscan Orogeny. Yet this is not the case for most analyses. On the contrary, the correlation of *in situ* $^{40}\text{Ar}/^{39}\text{Ar}$ ages and finite strain intensity at any scale reveals that ages are texturally-controlled and therefore independent of closure

785 temperature. This deviation to the closure temperature concept could be simply explained in
 786 terms of presence of ^{40}Ar excess either due to (1) partial deformation-induced resetting of
 787 Variscan muscovites (inherited argon), or (2) genuine incorporation of extraneous argon
 788 related to fluid circulations. Besides, a straightforward yet possible source of geologically
 789 meaningless ages could be simply explained by mixtures between inherited muscovite and
 790 synkinematic phengite argon reservoirs during lasering.

791 At first glance, as stated above, an overall correlation with the intensity of the finite strain
 792 can be drawn between samples. Weakly strained samples (from the Combeynot or the Plan
 793 du Lac shear zones) yielded rather old, sometime only pre-Alpine ages while phyllonites
 794 yielded much younger ages that tend to crystallization ages of newly formed crystals. Such
 795 control by the deformation is also perceptible at the scale of some dated rock samples with
 796 an overall age decrease toward the core of a shear band from ca. 300 Ma to less than 30 Ma
 797 in less than 1 centimetre (Fig. 11). It appears therefore tempting to ascribe to deformation
 798 the ability to facilitate argon diffusion and thus to partially reset (i.e. re-open) inherited
 799 muscovite isotopic system. Intensity of deformation, whether it occurs by grain-size
 800 reduction during mylonitization or by presence and distribution of high-diffusivity pathways
 801 lattice-modifying microstructures would therefore be the main controlling factor explaining
 802 the age scatter between an inherited cooling ages and the age of a tectonometamorphic event
 803 (i.e. Scaillet et al., 1990; West and Lux, 1993; Hames & Cheney 1997; Kramar et al. 2001,
 804 2003; Mulch et al. 2002). As suggested by the linear relationships observed between some
 805 apparent ages and corresponding $^{38}\text{Ar}/^{39}\text{Ar}$ ratios arguing for interactions with Cl-rich fluids,
 806 minor excess argon cannot be simply ruled out. Besides, a limited yet conspicuous input of
 807 extraneous argon within and in the vicinity of shear bands is clearly detectable for some
 808 samples (Fig. 10). In this case, results of absolute chronology are markedly inconsistent with
 809 the relative chronology established independently on the basis of microtextures. For
 810 example, mylonite sample A-4-66b presents a series of ca. 50-40 Ma old ages occurring
 811 within the core of the shear band while other analyses show a clear age decrease toward the
 812 shear band. Excess argon was moreover reported within fluid inclusions from Oisans
 813 basement quartz veins (Nziengui, 1993). In this study seven K/Ar experiments on white-
 814 micas from the sedimentary cover yielding initially 48 to 30 Ma ages were corrected to 27-
 815 23 Ma by removing excess ^{40}Ar content from fluid inclusions. If excess argon exists, it
 816 should have a homogeneous composition at the scale of the whole study area, which seems
 817 particularly unlikely. Excess argon is therefore present but does not explain the first order
 818 features of the results.

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822 **7.3 A new tectonic scenario**

823

824 This study brings primarily T-t-d (Temperature-time-deformation) results that further
 825 constrain the tectonometamorphic evolution of the external zones of the Western Alps. In
 826 this section, these results are all integrated into a compilation of published P-T-t-d results
 827 (Fig. 11 and 12) in order to discuss a new tectonic scenario (see also in Bellahsen et al.,
 828 2014) (Fig. 13).

829 Among the new results, this study constrains the thermal structure of a large volume of the
 830 external basement. Results, despite local variations, show a constant T_{max} around 330°C.
 831 This constant temperature along several tens of kilometres suggests that the basement and
 832 the cover preserved in half-grabens have been first quickly underthrust with rather limited
 833 internal shortening. Deformation ages from this study and from Simon-Labric et al. (2009)
 834 are consistent with a burial of the external zones around between 34 and 25 Ma (Fig. 12,
 835 13). The main shortening seems to have occurred at or around the metamorphic peak (Fig.
 836 12) as suggested in Boutoux et al. (2014b). The distribution of isotherms obtained with
 837 thermopaleomagnetic method, which cross-cut the folds, indicates that the main shortening
 838 had ended before cooling started in the Bourg d'Oisans basin (Crouzet et al., 1999). The
 839 sampled profile may therefore correspond to an east-dipping isotherm (around 330°C)
 840 during underthrusting of the Oisans Massif, which would then suggest that the whole area
 841 behaved as a more or less rigid block below the internal units during part of its burial
 842 history, before it was ultimately shortened. Similarly, the geothermal gradient observed
 843 close to the Lautaret and Galibier passes (around 20-30°C.km⁻¹) seems to seal the PFT
 844 activity dated between 32 and 27 Ma (Freeman, 1998). Thus, deformation mainly occurred
 845 just prior or during metamorphic-peak conditions and seems, in any case, to have ceased
 846 before the onset of the main cooling event. During this time period, and around the
 847 metamorphic peak conditions, the external zone crust underwent homogeneous shortening
 848 until approximately 25 Ma. Shortening was progressively localized into phyllonite bands. At
 849 25 Ma, the crust was progressively decoupled from the subducting lithosphere and started to
 850 exhumate. The PFT was then inactivated. These results argue against a progressive westward
 851 thrust slice development that lead to the exhumation of units after that they have reached

852 diachronously temperatures close to 330°C. However, this would imply that the basement
 853 top has been heated up uniformly from 50-100°C to 300-350°C uniformly over ca. 70km,
 854 which in turn suggests a rather slow burial.

855 At the scale of individual shear zones, high-strain samples (phyllonite) are younger
 856 (28-25Ma) than proto-mylonite and mylonite (33-30Ma). This suggests that deformation
 857 progressively localized along phyllonite bands. Moreover, ages of deformation (33-25 Ma)
 858 reveal that the deformation is contemporaneous all along the profile from the PFT to the
 859 Plan du Lac shear zone, i.e. over more than ca. 30km. This synchronicity of deformation
 860 shows that shortening in the ECM was distributed in space and time and confirms that the
 861 crust was relatively weak during shortening as proposed in Bellahsen et al. (2012). This is in
 862 agreement with the formation of newly formed shear zones rather than normal faults
 863 reactivation.

864 Later, shortening reached the Vercors fold and thrust belt during Burdigalian to
 865 Langhian times (Fig. 13). This shortening appears linked to crustal ramps below the
 866 Belledonne Massif, which suggests that the deformation localized in frontal parts after the
 867 end of the main shortening phase of the ECM at around 20 Ma (Ménard, 1979; Butler, 1989;
 868 Philippe et al., 1998). Between 25 and 20 Ma, the Oisans Massif has undergone fast cooling
 869 at a rate of 33°C.Ma⁻¹ (Crouzet et al., 2001) and was exhumed at a rate of 0.4 km.Ma⁻¹ (van
 870 der Beek et al., 2010). Final cooling, whose age is estimated from fission tracks ages
 871 (Vernon et al., 2008), occurred after 15 Ma, which corresponds to the age of thrusting within
 872 the Vercors Massif. During this more external fold and thrust belt shortening, only slight
 873 deformation occurred in the Oisans Massif (Dumont et al., 2008). The new argon data
 874 reported here thus complete the scenario of burial and first shortening at ca. 33 Ma,
 875 progressive localisation of shear zones until 25 Ma, and final exhumation above a crustal
 876 ramp until after 15-20 Ma.

877 Exhumation of the ECM was also accompanied by extensional reactivation of the PFT since
 878 the late Oligocene (Tricart et al., 2007) (Fig. 13). Thus, during the late Oligocene and early
 879 Miocene, the ECM was a crustal unit bordered by a normal fault to the east and a reverse
 880 fault to the west, transported with low internal deformation, similarly to the models of syn-
 881 collisional rock exhumation described by Chemenda et al. (1995) (Fig. 14). Moreover,
 882 numerous veins witnessing for fluids circulation have been documented between 17 to 5Ma
 883 in the Aar Massif (Mullis et al., 1994; Mullis, 1996; Challandes et al., 2008), in the Mont
 884 Blanc Massif (Leutwein et al., 1970; Marshall et al., 1998) or in the Belledonne Massif

(Gasquet et al., 2010), indicating that this event was coeval with a fluid-rich process which affected all the orogen.

8. Conclusion

Following the description of Alpine shear zones responsible for shortening of the Oisans External Massif of the Western Alps by Bellahsen et al. (2012) and Bellanger et al. (2014), the alpine thermal structure of the massif is established here thanks to the RSCM method and the shear zones were dated. The main outcome is that the interface between the cover and the basement has recorded a maximum temperature of $\sim 330^{\circ}\text{C}$, constant across the whole massif and that the shortening was accommodated by top-to-the west shear zones that progressively localized from 34 to 25 Ma, before the ECM were finally exhumed above a crustal ramp after 15-20 Ma, which in turn controlled shortening in the deformed cover further west (Vercors Massif).

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1510

1511 **Figure captions**

1512

1513 Figure 1: (a) Large-scale tectonic map of the Western Alps. Indicated is the location of the
1514 figure 2 that embraces the whole studied area. (b) Geological cross-section across the
1515 studied area close to the 45th parallel modified from Bellanger et al. (2014).

1516

1517 Figure 2: Simplified structural map of the studied area modified from Bellanger et al.
1518 (2014). Indicated is the location of the studied samples either from the Mesozoic cover for
1519 RSCM and chlorite geothermometry or from the basement shear zones for chlorite
1520 geothermometry and for the $^{40}\text{Ar}/^{39}\text{Ar}$ *in situ* approach. See tables for precise locations of
1521 samples.

1522

1523 Figure 3: Deformation features of the Oisans Massif at decreasing scales. (a) kilometer-scale
1524 view as illustrated by a geological cross-section from the Penninic Frontal Thrust (PFT) to
1525 La Mure Massif. Note the distribution of the Alpine shear zones and the overall eastward
1526 increasing of the finite strain intensity; (b) landscape-scale view as illustrated by a picture of
1527 a part of the Plan du Lac shear zone in the Vénéon Valley. Note the anastomosed
1528 relationships of the shear planes, the folded shape of the basement-cover interface and the
1529 pinched part of Mesozoic cover. Field-sketch diagrams represent macroscopic features of

the deformation from mylonite to phyllonite at outcrop-scale. (c) Thin-section pictures showing small-scale deformation features of from proto-mylonite to phyllonite strain-stage. Note the clear increase of the phengite fraction with strain increasing.

Figure 4: First-order results from the Raman Spectrometry of Carbonaceous Material approach. (a) RSCM Map showing mapping of the $T_{\max}$ results and inferred isotherms. The isotherm shape was drawn from interpolated map and from the distribution of the Illite crystallinity; (b) Illite crystallinity map modified from Aprahamian (1974) and Barlier et al. (1974). (c) Projection of the RSCM $T_{\max}$ over a classical E-W cross-section from the west of the Belledonne Massif to the Penninic Frontal Thrust. (d and e) RSCM $T_{\max}$ results vs. longitude and elevation along the cross-section (c). Data represented with smaller symbols to the east of the Penninic Frontal Thrust come from Beyssac et al. (2002); Gabalda et al. (2009); Lanari et al. (2012) and Plunder et al. (2012). Detailed results are given in table 1.

Figure 5: Local-scale features of results from the Raman Spectrometry of Carbonaceous Material approach. (a) Distribution of the $T_{\max}$ to the west of the Belledonne Massif from RSCM (circle) and vitrinite reflectance results (star; data from Deville and Sassi (2006) transformed into temperatures using the equation proposed in Allen and Allen (2005). The distribution of the $T_{\max}$ in the sedimentary pile is aligned along a ca. $30\text{-}50^{\circ}\text{C.km}^{-1}$ gradient. Besides, temperature gaps across narrow zones highlight localized tectonic contacts. (b) Distribution of the RSCM $T_{\max}$ results close to the Lautaret and Galibier passes. The structural data come from Depardon (1979), Beach (1981a, 1982), Ceriani et al. (2001), and Bellanger (2013). Note that the distribution of the $T_{\max}$ within the tectonic pile is consistent with a ca. $20\text{-}30^{\circ}\text{C.km}^{-1}$ post-tectonic field gradient. Detailed results are given in table 1.

Figure 6: Chlorite composition and results from chlorite geothermometry the Mesozoic cover. (a) XMg vs. Silica content diagram. (b) Clinocllore/Amesite/Sudoite ternary plot. Note that chlorite composition is first controlled by the bulk composition of the protolith. (c) Internal consistency of the geothermometric results. Chlorite geothermometry results using

both thermodynamic and empirical approaches (Cathelineau and Nieva, 1985, Vidal et al., 2005, 2006) are compared to RSCM data. Chlorite probably records a part of the cooling (retrograde) evolution. Detailed results are given in table 2.

Figure 7: Example of microstructural and composition of sample A.4.64 used as a guideline for dated samples. (a) Contrasted habits of white-micas as seen on several close-up views from the same thin-section. (b) White-mica chemical composition in ternary plots and XMg vs. Silica content diagram. Fine-grained, often syn-kinematic white-micas present in C-type and S-type foliation have phengite compositions. Well-developed white-mica clasts have a muscovite composition.

Figure 8: White-micas composition and chlorite geothermometry from dated samples (basement shear zones). For each of the dated shear zone, presented is the composition of white-micas as a function to their textural position in Si vs. Na content diagram and Muscovite/Celadonite/Pyrophyllite ternary plot together with results of chlorite geothermometry using both thermodynamic and empirical approaches (Cathelineau and Nieva, 1985, Vidal et al., 2001, 2005, 2006). Detailed results are given in table 2. (a) corresponds to results for the Col de Cluy shear zone while (b), (c) and (d) correspond to Plan du Lac, Col du Lac and Combeynot shear zones, respectively. Due to a very poor dataset, results for the Combeynot shear zone are not shown.

Figure 9: Time-chart showing all results of the *in situ* $^{40}\text{Ar}/^{39}\text{Ar}$ approach. Detailed results are given in table 3. Stratigraphic age for the Nummulitic flysch deposit is from Mulder et al. (2010); zircon and apatite fission-tracks results from Van der Beek et al. (2010).

Figure 10: Pictures showing the distribution of *in situ* laser ablation of the $^{40}\text{Ar}/^{39}\text{Ar}$ approach. Note the overall dependency of the K/Ar system to the finite strain intensity. Detailed results are given in table 3.

Figure 11: Pressure-Temperature-deformation-time evolution of the Oisans Massif. (a) Pressure-Temperature (P-T) constraints for the Oisans Massif. The P-T data come from (Gratier et al., 1973; Bernard, 1978) and references therein (Thermobarometry from fluids inclusions and paragenesis from cover and basement veins), (Jullien and Goffé, 1993) (conventional metamorphic petrology from the Emparis basin Mesozoic cover, (Crouzet et al., 1999) (thermopaleomagnetism from the Bourg d'Oisans basin), (Ceriani et al., 2003) (Zircon fission track and Kübler index and b-cell dimension of white mica from the Nummulitic cover of the Cheval noir unit), this study (RSCM from Mesozoic and Nummulitic cover). Temperature-deformation-time (T-d-t) constraints for the Oisans Massif. The T-d-t data come from (1) (Nziengui, 1993); (2) (Jullien and Goffé, 1993); (3) this study; (4) (Crouzet et al., 1999); (5) (Ménard and Rochette, 1992); (6) (Crouzet et al., 2001); (7) (van der Beek et al., 2010); (8) (Sabil, 1995; van der Beek et al., 2010; Beucher et al., 2012).

Figure 12: Large-scale compilation of existing time-constraints for the External zone and for the Western Alps close to the 45th parallel. (1) (Riche and Trémolières, 1987) (2) (Clauzon, 1990) (3) (Philippe et al., 1998) (4) (Du Fornel et al., 2004) (5) (Mulder et al., 2010) (6) (Féraud et al., 1995) (7) (Marignac et al., 1997) (8) (Gasquet et al., 2010) (9) (Simon-Labric et al., 2009) (a: Combeynot sinistral strike-slip shear zone; b: Pelvoux dextral strike-slip shear zone; c: Ailefroide dextral strike-slip shear zone); (10) this study; (11) (Nziengui, 1993) (12) (Crouzet et al., 1999, 2001) (13) (van der Beek et al., 2010) (14) (Sabil, 1995; van der Beek et al., 2010; Beucher et al., 2012) (15-16) (Fügenschuh and Schmid, 2003) (17) (Ceriani and Schmid, 2004) (18) (Ceriani et al., 2003; Fügenschuh and Schmid, 2003) (19) (Fügenschuh and Schmid, 2003) (20) (Féraud et al., 1995) (21) (Corsini et al., 2004) (22) (Sanchez et al., 2011a) (23) (Bigot-Cormier et al., 2000) (24) (Bigot-Cormier et al., 2000; Bogdanoff et al., 2000; Sanchez et al., 2011b) (25) (Pfiffner, 1992) (26) (Ruffini et al., 1997) (27-29) (Leutwein et al., 1970) (30) (Marshall et al., 1998) (31) (Crespo-Blanc et al., 1995) (32) (Kirschner et al., 1996) (33-34) (Rolland et al., 2008) (35) (Krummenacher and Evernden, 1960) (36) (Leutwein et al., 1970) (37-40) (Leloup et al., 2005) (41) (Baggio et al., 1967) (42-43) (Seward and Mancktelow, 1994; Leloup et al., 2005) (44-45) (Fügenschuh and Schmid, 2003) (46) (Freeman et al., 1998) (47) (Beucher, 2009) (48-49) (Fügenschuh and Schmid, 2003; Tricart et al., 2007) (50) (Ellenberger, 1958) (51) (Ganne, 2003) (52) (Gerber, 2008) (53) (Malusà et al., 2005) (54-56) (Ganne, 2003) (57-61) (Gerber, 2008) (62)

(Strzeczynski et al., 2012) (63) (Freeman et al., 1997) (64) (Agard et al., 2002) (65) (Duchêne et al., 1997) (66) (Rubatto and Hermann, 2003) (67) (Monié and Philippot, 1989) (68-69) (Cliff et al., 1998) (70-71) (Schwartz et al., 2007; Tricart et al., 2007; Beucher, 2009) (72) (Agard et al., 2002) (73) (Duchêne et al., 1997) (74) (Gebauer et al., 1997) (75) (Rubatto and Hermann, 2001) (76 et 78) (Tilton et al., 1991) (77) (Scaillet et al., 1992) (79) (Gebauer et al., 1997) (80) (Beucher et al., 2012) (81 et 83) (Chopin and Maluski, 1980) (82) (Ganne, 2003) (84) (Meffan-Main et al., 2004) (85-86) (Hurford and Hunziker, 1989; Hurford et al., 1991).

Figure 13: Lithospheric reconstructions of the Alps along a cross-section close to the 45th parallel. See text for explanations.

Table captions

Table 1: Detailed results from the RSCM approach. Given are the GPS position (latitude (Lat.) and longitude (Long.) in decimal degrees (WGS84) and the elevation in meter), the tectonic unit, the stratigraphic age, the lithology, the number of Raman spectra (Sp.), the R2 or the RA1 ratio, the RSCM temperature (mean and standard deviation). R2 ratio, in regular font (Beyssac et al., 2002) or the RA1 ratio, in italic font (Lahfid et al. 2010) was preferred depending on the shape of the spectra and the presence of the D4 peak (Lahfid et al., 2010).

Table 2: Detailed results from the chlorite geothermometry approach. Given are the GPS position (latitude (Lat.) and longitude (Long.) in decimal degrees (WGS84) and results of thermodynamic and empirical approaches (Cathelineau and Nieva, 1985, Vidal et al., 2005, 2006). Results are presented separately for the Mesozoic cover and the basement (Alpine shear zones).

Table 3: Detailed results from the $^{40}\text{Ar}/^{39}\text{Ar}$ approach. Given are the GPS position (latitude (Lat.) and longitude (Long.) in decimal degrees (WGS84) in decimal degrees (WGS84) and different output isotopic ratio.

	317	159.84	± 9.51	34.09	1.06	42.598	100.620	81.736	41	
	318	83.59	± 2.05	17.45	0.22	19.188	33.359	10.534	14	
	319	116.28	± 2.50	24.50	0.27	19.849	18.212	8.634	9	
	320	89.33	± 2.20	18.68	0.24	19.482	126.785	6.793	9	40
A-4-66b	19	37.78	± 3.17	7.79	0.33	21.743	41.626	26.734	47	
Plan du Lac	20	37.63	± 2.03	7.76	0.21	19.494	7.915	16.086	35	
Gneiss	26	38.95	± 1.54	8.03	0.16	15.253	0.760	11.468	27	
Lat.: 44.982410°	27	36.87	± 2.03	7.60	0.21	17.892	2.116	16.419	36	
Lon.: 6.147040°	29	301.42	± 6.56	66.91	0.79	35.325	0.000	34.137	13	
	30	62.29	± 1.41	12.93	0.15	16.109	16.173	6.088	11	
	33	129.05	± 3.03	27.29	0.33	27.668	8.263	15.628	14	
	34	36.00	± 0.98	7.42	0.10	12.959	0.000	5.512	16	
	37	35.72	± 1.89	7.36	0.20	22.241	35.724	14.517	34	
	39	32.67	± 1.63	6.73	0.17	17.296	12.892	12.831	33	
	40	32.08	± 1.10	6.60	0.11	14.308	7.733	7.737	23	
	42	30.72	± 1.09	6.32	0.11	14.967	0.000	7.800	24	
	43	32.87	± 1.14	6.77	0.12	13.521	3.686	8.014	23	
	45	33.49	± 0.77	6.90	0.08	13.770	2.480	3.254	11	
	46	33.55	± 0.82	6.91	0.09	13.321	3.816	4.022	13	
	48	29.17	± 0.68	6.00	0.07	14.767	13.296	2.745	10	
	49	29.58	± 0.67	6.08	0.07	15.319	13.327	2.398	9	
	51	31.49	± 0.65	6.48	0.07	13.435	4.182	1.473	5	
	52	31.98	± 0.70	6.58	0.07	14.403	2.899	2.587	9	
	54	33.12	± 0.79	6.82	0.08	14.390	5.062	3.737	12	
	55	39.81	± 0.89	8.21	0.09	14.321	2.465	3.589	10	
	57	37.33	± 2.48	7.70	0.26	18.621	8.016	20.685	41	
	58	31.65	± 2.39	6.51	0.25	19.490	4.940	20.049	44	
	60	35.03	± 1.33	7.22	0.14	18.049	2.470	9.941	26	
	61	34.70	± 1.35	7.15	0.14	18.266	3.038	10.078	27	
	63	37.84	± 0.96	7.80	0.10	15.423	5.192	5.144	15	
	64	40.50	± 1.06	8.35	0.11	18.151	4.933	6.022	16	
	66	36.59	± 1.26	7.54	0.13	18.187	8.598	8.838	23	
	67	38.66	± 1.45	7.97	0.15	18.954	7.837	10.731	26	
	69	47.58	± 1.20	9.83	0.13	16.492	6.142	6.583	15	
	70	125.56	± 3.47	26.52	0.38	36.088	10.440	22.749	20	
	72	31.89	± 0.88	6.56	0.09	16.576	0.000	4.987	16	
	73	30.51	± 0.74	6.28	0.08	14.076	2.988	3.610	13	
	74	30.90	± 0.72	6.36	0.07	14.087	5.008	3.139	11	34
A-4-68	245	61.32	± 2.22	12.72	0.23	19.501	25.355	16.475	26	
Plan du Lac	246	40.61	± 2.03	8.38	0.21	21.339	32.150	16.278	34	
Gneiss	247	31.80	± 0.80	6.55	0.08	15.058	19.614	4.127	14	
Lat.: 44.984950°	249	28.77	± 1.17	5.92	0.12	16.462	21.462	8.823	27	
Lon.: 6.143560°	250	93.51	± 2.10	19.58	0.23	24.615	5.839	9.355	12	
	251	74.73	± 1.81	15.56	0.19	24.938	42.041	9.522	14	
	253	24.28	± 1.26	4.99	0.13	15.279	87.896	9.612	32	
	254	25.98	± 1.67	5.34	0.17	16.846	9.266	13.651	39	
	255	49.36	± 2.18	10.21	0.23	26.893	34.100	17.096	31	
	257	32.08	± 1.17	6.60	0.12	21.853	29.097	8.412	25	
	258	24.47	± 0.77	5.03	0.08	16.339	36.424	4.936	19	
	259	32.06	± 1.34	6.60	0.14	24.525	32.552	10.214	28	
	264	25.79	± 0.79	5.30	0.08	16.313	21.900	5.101	19	
	265	22.35	± 0.90	4.59	0.09	15.449	29.685	6.636	26	
	266	36.85	± 0.87	7.60	0.09	16.856	28.362	4.120	12	15
A-4-68b (Ech8)	144	24.88	± 0.54	5.14	0.06	12.752	3.706	0.478	2	
Plan du Lac	145	27.16	± 1.41	5.62	0.15	17.908	0.000	0.000	0	
Gneiss	146	26.65	± 0.55	5.51	0.06	13.920	0.000	0.507	2	
Lat.: 44.984950°	148	25.74	± 0.56	5.32	0.06	14.891	0.000	0.727	3	
Lon.: 6.143560°	149	25.16	± 0.55	5.20	0.06	14.784	0.000	0.583	3	
	150	25.95	± 0.55	5.37	0.06	14.536	12.917	0.570	3	
	152	26.58	± 0.69	5.50	0.07	12.124	5.811	0.701	3	
	153	25.66	± 2.89	5.30	0.30	12.211	0.000	0.066	0	
	154	25.75	± 3.15	5.32	0.33	12.126	271.884	0.184	0	
	156	32.57	± 3.85	6.75	0.40	10.353	108.067	1.046	4	
	157	47.23	± 1.41	9.82	0.15	11.567	168.834	3.893	9	
	158	46.40	± 1.90	9.65	0.20	17.515	18.150	14.508	29	
	160	201.82	± 12.76	43.83	1.46	46.660	2794.213	115.611	42	
	161	25.34	± 0.63	5.24	0.07	14.586	8.934	0.558	3	
	162	28.20	± 0.64	5.83	0.07	15.229	181.960	2.092	8	
	164	25.22	± 0.59	5.21	0.06	14.405	2.375	2.040	9	
	165	28.43	± 1.02	5.88	0.11	14.576	75.735	7.377	24	
	166	27.73	± 1.10	5.74	0.12	14.654	0.432	8.144	26	
	168	25.92	± 0.58	5.36	0.06	10.834	14.811	0.693	3	
	169	25.30	± 0.53	5.23	0.05	12.533	13.464	0.821	4	
	170	32.25	± 0.74	6.68	0.08	14.533	15.877	2.877	10	
	172	31.89	± 0.86	6.61	0.09	14.772	25.511	4.845	16	
	173	26.67	± 0.61	5.52	0.06	13.389	0.000	2.188	9	
	174	32.11	± 0.84	6.65	0.09	14.983	227.616	4.170	13	
	176	26.71	± 0.62	5.52	0.07	10.443	40.681	2.076	9	
	177	28.76	± 0.68	5.95	0.07	14.683	11.784	2.458	9	
	178	30.74	± 0.66	6.36	0.07	12.792	56.401	1.576	6	
	180	31.38	± 2.22	6.50	0.23	15.375	6.264	18.517	42	
	181	26.39	± 0.57	5.46	0.06	13.365	44.315	1.762	7	
	182	26.68	± 0.86	5.52	0.09	16.617	31.735	5.459	20	

Sample	An.	Age (+2s)	40%/39(k)	Is	38Ar/39Ar (*1000)	37Ar/39Ar (*1000)	36Ar/39Ar (*1000)	% Atm	n
<i>A-1-345</i>	110	389.51 ± 27.17	89.23	3.46	109.716	5.672	280.279	48	
Combeynot	111	315.64 ± 11.36	70.80	1.39	62.828	8.248	100.136	29	
Porphyroid	113	199.95 ± 5.88	43.40	0.67	34.056	22.486	42.774	22	
Lat.: 45.026510°	114	330.96 ± 7.53	74.56	0.93	42.462	3.763	44.245	15	
Lon.: 6.404380°	116	294.73 ± 6.91	65.72	0.83	40.937	5.032	43.083	16	
	117	290.84 ± 6.72	64.78	0.81	36.516	0.000	40.250	15	
	119	231.14 ± 6.59	50.62	0.77	38.998	4.310	48.412	22	
	120	276.63 ± 9.00	61.37	1.08	50.417	8.596	74.129	26	
	122	231.91 ± 5.81	50.80	0.68	38.043	4.734	37.637	18	
	123	221.72 ± 6.55	48.43	0.76	40.747	2.831	49.034	23	
	125	169.08 ± 3.29	36.39	0.37	13.650	11.799	3.969	3	
	126	272.85 ± 5.81	60.46	0.69	29.630	24.865	17.269	8	
	128	252.54 ± 4.98	55.64	0.59	16.655	37.405	7.473	4	
	129	156.32 ± 3.22	33.52	0.36	21.734	4.032	7.403	6	
	130	206.61 ± 4.42	44.93	0.51	33.808	22.324	11.769	7	15
<i>A-1-355</i>	205	30.82 ± 1.00	6.38	0.10	18.330	67.910	5.718	18	
Col du Lac	206	38.08 ± 3.64	7.90	0.38	58.001	138.816	11.462	27	
Gneiss	207	32.87 ± 0.88	6.81	0.09	15.744	85.272	2.104	7	
Lat.: 45.017860°	209	27.30 ± 0.80	5.65	0.08	16.684	40.285	4.607	17	
Lon.: 6.273540°	210	27.06 ± 0.99	5.60	0.10	16.173	83.580	5.847	20	
	211	24.80 ± 0.76	5.13	0.08	15.227	32.113	3.441	14	
	213	24.80 ± 0.57	5.13	0.06	13.758	26.275	2.476	11	
	214	26.58 ± 1.41	5.50	0.15	30.685	83.420	4.539	17	
	215	26.02 ± 0.65	5.38	0.07	16.349	154.835	2.906	11	
	217	32.34 ± 1.05	6.70	0.11	16.960	124.361	7.162	21	
	218	27.13 ± 0.91	5.61	0.09	17.432	244.419	5.225	18	
	219	27.77 ± 1.05	5.74	0.11	19.465	251.880	6.967	23	
	221	26.22 ± 0.66	5.42	0.07	15.212	51.173	2.493	10	
	222	88.86 ± 8.58	18.70	0.93	22.432	121.326	71.317	52	
	223	58.72 ± 2.12	12.25	0.22	22.361	24.720	14.784	25	
	225	27.22 ± 0.88	5.63	0.09	15.473	93.917	5.594	20	
	226	31.72 ± 1.11	6.57	0.12	19.244	109.244	6.726	20	
	227	26.44 ± 2.01	5.47	0.21	18.452	89.430	16.213	42	
	229	41.30 ± 1.63	8.58	0.17	15.750	79.334	11.810	27	
	230	30.61 ± 1.13	6.34	0.12	19.183	111.046	5.922	19	
	231	23.88 ± 1.64	4.93	0.17	16.507	58.995	13.359	40	
	233	27.55 ± 1.44	5.70	0.15	10.904	120.782	9.622	29	
	234	26.44 ± 4.85	5.47	0.50	16.482	272.598	41.140	64	
	235	21.09 ± 4.19	4.36	0.44	23.129	37.242	36.304	67	
	237	14.54 ± 9.58	3.00	0.99	31.718	0.581	83.039	86	
	238	33.41 ± 2.36	6.92	0.25	22.418	113.398	18.933	41	
	239	42.68 ± 2.22	8.87	0.23	15.109	82.208	16.743	33	
	241	36.11 ± 2.27	7.49	0.24	19.655	164.103	18.251	38	
	242	109.75 ± 7.11	23.23	0.78	27.052	135.090	49.364	38	
	243	37.65 ± 2.33	7.81	0.24	24.920	87.031	18.521	38	30
<i>A-4-74b</i>	268	169.77 ± 3.55	36.31	0.40	27.306	43.850	13.453	10	
Plan du Lac	269	52.47 ± 1.07	10.86	0.11	16.745	55.616	2.240	5	
Gneiss	270	74.94 ± 2.66	15.61	0.28	19.667	1412.051	14.647	19	
Lat.: 44.984930°	272	26.69 ± 0.60	5.48	0.06	12.978	52.107	1.561	7	
Lon.: 6.125820°	273	24.76 ± 0.68	5.09	0.07	13.439	3.485	3.354	14	
	274	24.29 ± 1.75	4.99	0.18	16.268	13.714	14.242	41	
	277	105.95 ± 2.91	22.26	0.32	31.313	74.487	7.328	9	
	278	186.61 ± 4.40	40.10	0.50	26.968	0.000	22.227	14	
	279	158.63 ± 4.59	33.82	0.51	28.448	79.390	30.843	21	
	281	207.10 ± 5.24	44.76	0.60	34.788	41.001	32.855	18	
	282	132.95 ± 4.52	28.14	0.50	31.099	26.803	32.926	25	
	283	76.73 ± 3.65	15.99	0.39	12.256	154.469	25.925	31	
	285	126.70 ± 3.73	26.77	0.41	22.875	33.883	24.493	21	
	286	43.97 ± 5.66	9.08	0.59	22.167	48.539	48.845	59	
	287	57.14 ± 1.27	11.84	0.13	16.889	58.258	4.372	9	
	289	105.53 ± 2.37	22.17	0.26	21.393	37.534	6.227	7	
	290	131.78 ± 3.12	27.89	0.34	26.144	18.864	14.691	13	
	291	111.61 ± 6.66	23.49	0.72	37.316	48.608	54.146	40	
	293	40.96 ± 0.98	8.45	0.10	18.950	68.986	1.636	5	
	294	36.84 ± 0.82	7.59	0.09	16.349	5.038	2.719	9	
	295	54.01 ± 1.35	11.18	0.14	20.151	2.290	6.796	14	
	297	32.65 ± 0.93	6.72	0.10	16.406	321.626	4.222	13	
	298	36.43 ± 0.87	7.51	0.09	14.979	25.149	1.999	6	
	299	32.17 ± 0.80	6.62	0.08	15.178	20.601	3.209	11	
	301	40.05 ± 0.91	8.26	0.09	13.792	14.734	2.443	7	
	302	32.71 ± 0.81	6.73	0.08	13.812	102.166	1.986	7	
	303	38.28 ± 1.00	7.89	0.10	15.011	2.251	5.515	16	
	305	46.66 ± 1.09	9.64	0.11	17.635	42.403	4.734	12	
	306	47.53 ± 1.15	9.83	0.12	17.989	25.982	5.711	13	
	307	59.65 ± 1.67	12.37	0.18	19.478	125.491	9.956	18	
	309	25.22 ± 1.08	5.18	0.11	15.696	103.169	8.232	28	
	310	43.87 ± 1.14	9.06	0.12	16.736	4.395	6.180	15	
	311	86.77 ± 3.01	18.13	0.32	26.519	11.467	22.305	26	
	313	45.26 ± 1.45	9.35	0.15	20.073	12.845	9.898	22	
	314	34.80 ± 1.04	7.17	0.11	16.355	471.527	6.146	17	
	315	73.72 ± 3.24	15.35	0.34	28.247	13.271	25.751	32	

	Sample	Lat. (°)	Lon. (°)	Shear zone (or area)	Lithology	n	T σ		T σ		XFe ₃ σ		RSCM σ	
							Cathelineau and Nieva [1985]		Vidal et al. [2005, 2006]		Beyssac et al. [2002]			
Cover	A-2-2	45.16481	5.851862	Uriage	Limestone	25	294	17	252	61	6	5	306	13
	A-1-321	45.223812	6.138562	Bourg d'Oisans	Shale	11	296	9	313	55	6	6	345	9
	A-1-345(2)	45.00973	6.26403	Col du Lac	Limestone	24	297	12	294	45	7	5	335	5
	A-4-232	44.55958	6.47881	Embrun	Black shale	13	288	13	269	51	5	5	308	5
Basement	A-4-114	45.08019	6.11031	Col de Cluy	Chloritoschist	4	259	15	187	61	34	9		
						7	275	17	255	43	30	10		
	A-4-64	44.97745	6.15251	Plan du Lac	Gneiss	8	267	8	255	26	29	4		
						3	256	16	267	11	34	3		
						4	261	7	250	32	31	7		
						5	262	16	250	8	32	4		
	A-4-28a	45.0173	6.270507	Col du Lac	Gneiss	7	268	7	258	15	32	4		
						10	266	6	262	11	30	4		
	A-4-222c	45.01764	6.28244		Gneiss	2	265	4	271	10	42	3		
						2	262	2	263	9	46	0		
						4	269	1	283	4	38	1		
						1	266		268	-	40	-		
	A-1-355	45.01786	6.27354		Gneiss	17	255	8	267	13	56	5		
						5	239	18	263	9	62	3		
	A-1-345	45.02651	6.40438	Combeynot	Orthogneiss	2	263	1	244	1	31	0		

Table 2

Sample	Lat.	Lon.	Elev.	Tectonic unit	Stratigraphic age	Lithology	n	R2/RA/ ratio	Tmax (°C)	sd (°C)
	(°)	(°)	(m)							
A-1-001	45.004578	5.70136	560	Dauphinois	Upper Lias	Slate	14	0.70	329	8
A-1-004	44.87225	5.86169	948	Dauphinois	Upper Lias	Slate	25	0.58	258	7
A-1-007	45.069906	6.000484	806	Dauphinois	Lower Lias	Limestone	12	0.72	321	6
A-1-009	45.08103	6.04491	1513	Dauphinois	Lower Lias	Limestone	13	0.71	327	3
A-1-016	45.03812	6.131952	1257	Dauphinois	Lower Dogger	Slate	12	0.63	321	12
A-1-025	45.14811	6.31462	1516	Dauphinois	Upper Lias	Slate	12	0.69	335	6
A-1-029	45.032105	6.381495	1952	Dauphinois	Middle Dogger	Limestone	12	0.69	335	2
A-1-030	45.038448	6.398698	2128	Dauphinois	Lower Lias	Limestone	11	0.69	333	4
A-1-32	45.042092	6.388473	2215	Dauphinois	Priabonian	Flysch	15	0.61	298	4
A-1-37	45.291423	6.324367	1485	Dauphinois	Upper Lias	Slate	12	0.69	333	2
A-1-171	45.226435	6.190374	2119	Dauphinois	Lower Lias	Limestone	12	0.71	327	5
A-1-208	45.062326	6.126121	1761	Dauphinois	Lower Lias	Limestone	11	0.63	319	8
A-1-224	45.049632	6.102744	1166	Dauphinois	Lower Lias	Limestone	11	0.70	330	5
A-1-225	45.037	5.97841	1120	Dauphinois	Lower Lias	Limestone	10	0.70	331	3
A-1-318	44.984199	6.123207	982	Dauphinois	Upper Lias	Slate	14	0.69	333	9
A-1-320	45.127996	6.045468	801	Dauphinois	Upper Lias	Slate	12	0.69	336	9
A-1-321	45.223812	6.138562	1770	Dauphinois	Upper Lias	Slate	11	0.67	345	9
A-1-322b	45.21556	6.213118	1794	Dauphinois	Upper Lias	Slate	12	0.70	329	3
A-1-323	45.215115	6.252217	1461	Dauphinois	Upper Lias	Slate	12	0.69	336	8
A-1-324	45.192802	6.290928	1271	Dauphinois	Lower Lias	Limestone	12	0.69	332	3
A-1-326	45.200339	6.336937	1502	Dauphinois	Lower Dogger	Slate	11	0.63	322	10
A-1-327	45.218918	6.305905	1290	Dauphinois	Upper Lias	Slate	11	0.66	347	4
A-1-328	45.227265	6.303567	1240	Dauphinois	Lower Lias	Limestone	11	0.69	336	5
A-1-329	45.249661	6.28425	1286	Dauphinois	Lower Dogger	Slate	8	0.69	334	5
A-1-330	45.2608	6.367982	776	Dauphinois	Lower Dogger	Slate	12	0.66	347	6
A-1-331	45.291702	6.23121	1107	Dauphinois	Upper Lias	Slate	12	0.69	332	12
A-1-332	45.045302	6.313245	1464	Dauphinois	Upper Lias	Slate	12	0.69	332	8
A-1-333	45.040348	6.30609	1592	Dauphinois	Lower Dogger	Slate	12	0.71	326	4
A-1-334	45.039006	6.304639	1587	Dauphinois	Lower Dogger	Slate	11	0.70	328	4
A-1-337	45.052165	6.288855	1806	Dauphinois	Lower Dogger	Slate	13	0.70	330	4
A-1-342	45.076795	6.338908	2119	Dauphinois	Middle Dogger	Limestone	11	0.64	326	4
A-1-345(2)	45.00973	6.26403	3193	Dauphinois	Upper Lias	Slate	12	0.69	335	5
A-2-002	45.16481	5.851862	813	Dauphinois	Lower Dogger	Slate	12	0.62	306	13
A-2-003	45.162054	5.81894	393	Dauphinois	Middle Dogger	Limestone	10	0.62	306	6
A-3-019	44.526656	6.726225	1690	Embrunais-Ubaye	Cretaceous	Helm. Flysch	12	0.59	262	6
A-4-005	45.052408	6.390277	2355	Dauphinois	Priabonian	Flysch	12	0.61	296	9
A-4-006	45.05279	6.40051	2379	Dauphinois	Priabonian	Flysch	12	0.61	288	6
A-4-010	45.06035	6.4223	2758	Tête Noire	Cretaceous	Helm. Flysch	12	0.59	265	8
A-4-011	45.06062	6.42282	2772	Tête Noire	Cretaceous	Helm. Flysch	11	0.58	261	9
A-4-014	45.063536	6.412644	2670	Subbriançonnais	Eocene	Flysch	12	0.59	267	6
A-4-039c1	45.04773	6.13796	1072	Grandes Rousses	Carboniferous	Sandstone	15	0.63	319	7
A-4-042	45.044332	6.15657	1064	Dauphinois	Lower Dogger	Slate	15	0.69	335	14
A-4-044	45.042854	6.168657	1462	Dauphinois	Lower Dogger	Slate	12	0.70	329	5
A-4-055	45.0421	6.35506	1770	Dauphinois	Lower Dogger	Slate	12	0.69	335	3
A-4-057	45.016852	6.039292	1509	Dauphinois	Lower Lias	Limestone	12	0.69	334	2
A-4-121	44.95072	5.96606	978	Dauphinois	Lower Lias	Limestone	10	0.64	326	7
A-4-122	44.89179	5.8778	695	Dauphinois	Lower Lias	Limestone	12	0.61	298	10
A-4-123	44.87871	5.7535	793	Dauphinois	Lower Dogger	Slate	12	0.60	279	9
A-4-124	44.87459	5.71011	667	Dauphinois	Lower Lias	Limestone	14	0.64	334	6
A-4-126	44.96182	5.81198	1181	Dauphinois	Upper Lias	Slate	11	0.63	314	8
A-4-140	45.33314	6.33442	1137	Dauphinois	Upper Lias	Slate	11	0.70	331	3
A-4-196	44.789297	6.45698	1671	Dauphinois	Priabonian	Flysch	11	0.64	330	5
A-4-197	45.03796	5.7054	403	Dauphinois	Upper Lias	Slate	12	0.69	334	4
A-4-198	45.04444	5.69423	309	Dauphinois	Middle Dogger	Limestone	11	0.64	325	6
A-4-199	45.04834	5.68057	340	Dauphinois	Middle Dogger	Limestone	12	0.61	296	5
A-4-200	45.01018	5.63507	499	Subalpine chain	Lower Malm	Marls	12	0.54	208	9
A-4-201	45.00197	5.62223	741	Subalpine chain	Lower Malm	Marls	14		< 200	
A-4-223	45.013605	6.283496	2509	Dauphinois	Lower Lias	Limestone	11	0.68	338	4
A-4-231	44.63593	6.56532	912	Subbriançonnais	Eocene	Flysch	16	0.62	302	10
A-4-232	44.55958	6.47881	883	Vocontian domain	Dogger/Malm	Black Shale	13	0.62	308	5
A-4-233	44.58166	6.39937	1211	Vocontian domain	Dogger/Malm	Black Shale	11	0.59	271	11
A-4-235	44.5549	6.34229	1246	Vocontian domain	Dogger/Malm	Black Shale	11	0.58	253	10
A-4-236	44.714333	6.156877	1667	Dauphinois	Upper Lias	Slate	11	0.56	228	8
A-4-238	44.713257	6.167974	1960	Dauphinois	Upper Lias	Slate	12	0.54	202	4
A-4-239b	44.714389	6.168411	1982	Dauphinois	Priabonian	Flysch	13		< 200	
A-4-242	45.027604	6.312056	2228	Dauphinois	Lower Lias	Limestone	10	0.64	333	8
A-4-244	44.83378	5.93966	1002	Dauphinois	Upper Lias	Slate	11	0.57	237	7
A-4-247	44.844446	6.127314	2055	Dauphinois	Lower Lias	Limestone	12	0.69	334	2
A-5-1	44.867686	5.61239	847	Subalpine chain	Lower Malm	Marls	13	0.54	204	14
A-5-2	44.92326	5.63483	829	Dauphinois	Middle Dogger	Limestone	11	0.61	288	7
A-5-3	44.95121	5.60154	788	Subalpine chain	Upper Dogger	Marls	11		< 200	
A-5-4	44.9613	5.58629	701	Subalpine chain	Lower Malm	Marls	12		< 200	
A-5-5	44.92913	5.63472	819	Dauphinois	Middle Dogger	Limestone	11	0.60	286	5
A-5-6	44.9058	5.6258	902	Subalpine chain	Upper Dogger	Marls	15	0.56	227	23
A-4-154	45.25002	6.78357	1438	Subbriançonnais	Lias	Limestone	12	0.51	412	5
A-4-228	44.814212	6.567413	1147	Dent Parraché	Malm	Limestone	12	0.61	291	6

Table 1

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12

3: Crustal ramp and exhumation of the ECM

2: Internal shortening and progressive strain localisation

1: Burial with almost no shortening

European plate :

- Tertiary cover
- Mesozoic cover
- Upper crust

- Lower crust
- Mantle
- Penninic units

Apulian plate :

- Crust
- Mantle

Figure 13