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1 **THERMAL EVOLUTION OF PERMIAN POST-OROGENIC EXTENSION AND JURASSIC RIFTING**  
2 **RECORDED IN THE AUSTRALPINE BASEMENT (SE SWITZERLAND, N ITALY)**

3  
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20 **HIGHLIGHTS**

- 21 • New Ar-ages are reported from the Alpine Tethys necking zone.  
22 • A slow (5-10 °C/Myr) post-Variscan cooling is identified in the Grosina unit.  
23 • A rapid (25-100 °C/Myr) Jurassic cooling is identified in the necking zone.  
24 • A re-heating is suspected during the necking of the Alpine Tethys rifted margin.  
25 • The thermal evolution is constrained by Monte Carlo simulations based on DiffArgP.

26 **ABSTRACT**

27           The thermal evolution of the continental lithosphere is intimately linked to its tectonic  
28 history. In the present contribution, we explore the thermal evolution of the Austroalpine Campo  
29 and Grosina units that composed the former necking zone of the Adriatic rifted margin and that are  
30 now exposed in the Eastern Alps (SE Switzerland and N Italy). The Campo unit and the overlying  
31 Grosina unit are separated by a Jurassic horizontal extensional structure, the Eita shear zone. We  
32 report new muscovite and biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages for the Grosina unit (~261 Ma for muscovite and  
33 ~246-245 Ma for biotite) and for the Campo unit (~218-178 Ma for muscovite and ~214-171 Ma  
34 for biotite). No geologically meaningful amphibole ages were obtained because of contamination by  
35 extraneous argon in these minerals. The new results are combined with published ages obtained  
36 with different chronometers (U–Pb, Sm–Nd, Rb–Sr, K–Ar,  $^{40}\text{Ar}/^{39}\text{Ar}$ ). This dataset reveals that the  
37 two units record different thermal evolutions from the Permian post-orogenic extension to the  
38 Jurassic rifting. Ages of cooling initiation and associated cooling rates are estimated by Monte  
39 Carlo simulations based on DiffArgP. The Grosina unit records a Permo-Triassic slow cooling of 5-  
40 10 °C/Myr initiated during Early Permian in relation to the post-Variscan relaxation of the  
41 isotherms and cooling of the lithosphere that caused a large-scale thermal subsidence. By contrast,  
42 the Campo unit records a fast cooling of 25-100 °C/Myr associated to the tectonic exhumation of  
43 basement rocks during the Jurassic rifting and the formation of the Alpine Tethys rifted margin. This  
44 fast cooling appears delayed from major basement deformation phases, and likely follows a period  
45 of crustal re-heating.

46           **Keywords:** Post-Variscan; Alpine Tethys; Rifting; Necking zone; Austroalpine;  $^{40}\text{Ar}/^{39}\text{Ar}$   
47 ages.

## 48 1. INTRODUCTION

49 The interplay between tectonics, thermal evolution and topography is well exemplified by  
50 two events in an extensional cycle: the late orogenic phase (transition from an orogenic thick crust  
51 to an equilibrated ~35 km thick continental crust), and a rift event (thinning of a ~35 km thick crust  
52 to form a rifted margin and ultimately an oceanic domain).

53 The last stages of most orogens are characterized by the thinning of the continental crust.  
54 The reasons for the reduction in crustal thickness is still a matter of debate for some cases, with  
55 competing scenarios that include (see Rey et al., 2001 for a review): detachment or convective  
56 removal of the crustal and/or lithospheric root, slab break-off, change in far-field stress conditions  
57 or crustal strength. However, all scenarios are accompanied by a transient rise of the asthenosphere  
58 that can cause a rapid uplift together with magmatism and prograde metamorphism. As the  
59 lithosphere-asthenosphere boundary returns to thermal equilibrium, the continental crust gently  
60 subsides (e.g., Platt and England, 1994). This is the case for the Variscan orogeny in W Europe. It is  
61 characterized by the decrease in topography together with a thinning of the continental crust to  
62 reach a normal thickness of  $35 \pm 5$  km before 310-300 Ma (Burg et al., 1994) that is immediately  
63 followed (but with a cogenetic link that remains debated) by a period of intense magmatic activity  
64 together with regional high temperature prograde metamorphism (Fig. 1a; Schuster and Stüwe,  
65 2008). Extensional deformation is attested by both the presence of detachment faults at the surface  
66 (Pohl et al., 2018) and by several deformation structures in the basement, mostly associated to the  
67 emplacement of massive volumes of mantle-derived melts during the Early Permian (e.g., Petri et  
68 al., 2018). However, exhumation of deep crust should have been limited, as most of Permian lower  
69 crustal rocks recorded a post-HT isobaric cooling (e.g., Müntener et al., 2000; Petri et al., 2016).  
70 This is followed by a period of (relative) tectonic quiescence, characterized by a conductive cooling  
71 of the lithosphere and the associated thermal subsidence (Fig. 1ab; e.g., Prijac et al., 2000). The

72 detailed thermal evolution of this phase remains to be calibrated, notably the rate at which the  
73 continental crust is thermally re-equilibrated from the Permian to the Triassic.

74         During the formation of rifts and rifted margins, rocks are vertically displaced across the  
75 lithosphere by tectonics, but part of the topography (or the bathymetry) remains controlled by the  
76 transient thermal architecture of the crust and the lithosphere. This subsidence can be linked to the  
77 rifting history in the pure shear rift model of McKenzie (1978). However, the depth-uniform  
78 thinning implied by pure shear deformation results in lithospheric cooling only, and the more  
79 complex thermal and uplift/subsidence evolution of rifted margins can only be explained by the  
80 alternative models that integrate depth-dependant thinning. Several rift systems worldwide are  
81 indeed characterized by transient subaerial exposure and erosion (Fig. 1c; Esedo et al., 2012),  
82 followed by a rapid subsidence as soon as the thermal support lowers (Fig. 1c; e.g., Chenin et al.,  
83 2019). In remnants of the Alpine Tethys rifted margins, such an evolution is recorded in several  
84 settings. The Briançonnais domain delimited by the Valais and the Piemont-Liguria basins, is  
85 characterized by a phase of emersion attested by the presence of karst followed by a rapid  
86 subsidence during an early post-rift stage (Lemoine et al., 1986; and Mohn et al., 2011 for a  
87 review). A phase of uplift and potential basement exposure has been identified in zones of crustal  
88 necking located between (future) proximal and distal domains (Fig. 1c) by the occurrence of major  
89 unconformities and of siliciclastic sedimentary deposits sourced from the necking domain that are  
90 synchronous to major crustal thinning (i.e., crustal necking). So far 3 zones have been recognized in  
91 the Alps: the Mont Blanc massif (Ribes et al., 2020), the Gozzano-Invorio-Arona area in the  
92 Southern Alps (Decarlis et al., 2017), and suspected for the Campo and Grosina units in the Eastern  
93 Alps (Mohn et al., 2012). At depth, several lines of evidence seem to indicate that distal and  
94 necking domains are thermally supported by a short lived syn-rift phase of heating that stopped  
95 during late rifting (documented in the Southern Alps; Ewing et al., 2015; Smye and Stöckli, 2014),  
96 but details of this event (amplitude, duration) and notably its lateral extent remain unclear.

97           Constraining the thermal history of reactivated rifted margins remains, however, strongly  
98 hampered by the accessibility to deep crustal levels that escaped subsequent thermal overprint  
99 during the Alpine orogeny (Mohn et al., 2022). In the western end of the Eastern Alps, this can be  
100 achieved only by  $^{40}\text{Ar}/^{39}\text{Ar}$  thermochronometers having closure temperatures above 300 °C, higher  
101 than the maximum temperature reached during the Alpine orogeny. Their use in pre-Alpine  
102 basement units are however complexified by a polyphase history with multiple deformation events  
103 and heating/cooling cycles. Early works identified groups of basement cooling ages related to both  
104 the Variscan (~ 390 to 290 Ma) and the Alpine (~ 60 to 20 Ma) orogenic cycles (e.g., Hunziker et  
105 al., 1992), i.e., essentially to compressive tectonic deformation phases, and authors referred to ages  
106 lying between these two bonds as “intermediate” or “mixed” ages. In recent years, several groups of  
107 ages have been proven to record extensional tectonic events too: (1) the Permian post-orogenic  
108 evolution that is characterized by widespread magmatism and prograde metamorphism associated  
109 with crustal and lithospheric extension notably documented in the Alpine realm (Manzotti et al.,  
110 2018, 2012; Petri et al., 2017; Schuster and Stüwe, 2008; Wyatt et al., 2022), and (2) the Jurassic  
111 rifting and opening of the Alpine Tethys, which resulted in the exhumation of middle to lower  
112 crustal levels (Mohn et al., 2012) with a thermal evolution that remains to be constrained. For both  
113 contexts, several key questions remain open, that can be answered by analysing the thermal  
114 evolution of the involved basement units.

115           The aim of the present contribution is to characterize the nature and the temperature-time  
116 evolution of the basement of the Adriatic necking zone from the Paleozoic Variscan to the Permian  
117 post-Variscan extension to the Jurassic rifting. More precisely, the goal is twofold: (1) tracking a  
118 possible post-Variscan (Permian) exhumation and/or cooling (Fig. 1ab), and (2) characterizing the  
119 exhumation of the middle-crust during rifting and its related thermal evolution (Fig. 1c). To do so,  
120 we compile structural and petrological data for both the Grosina and Campo units (Eastern Alps, N  
121 Italy and SE Switzerland) that were part of the former Adriatic necking zone and constrain their

122 temporal and thermal evolution with  $^{40}\text{Ar}/^{39}\text{Ar}$  dating. These new ages are compared to published  
123 ages and the temperature-time evolution is constrained through Monte Carlo random simulations  
124 based on DiffArgP (Warren et al., 2012a; Wheeler, 1996).

## 125 **2. TECTONIC, GEOCHRONOLOGICAL AND GEOLOGICAL FRAMEWORK**

### 126 **2.1. The evolution of the Austroalpine units**

127 We focus on the Austroalpine nappes exposed along the western end of the Eastern Alps  
128 (Fig. 2a). Before being encompassed in the Alpine orogen, these nappes underwent a polyphase pre-  
129 Alpine history including the Variscan and post-Variscan evolution, and the rifting and opening of  
130 the Alpine Tethys.

#### 131 ***2.1.1. Variscan and post-Variscan evolution***

132 Evidence for Variscan high-pressure (HP) metamorphism and deformation are mostly  
133 reported in the western Austroalpine nappes, where crustal units reached peak metamorphic  
134 conditions during the Carboniferous. In the study area, pre-Alpine amphibolite-facies  
135 metamorphism is documented by several authors (e.g., Petri et al., 2016), and is presumed to be  
136 Variscan in age, although undated. However, evidence for Variscan-related magmatism is reported  
137 in the Bernina unit with Carboniferous plutonism of a calc-alkaline affinity (Spillmann and Büchi,  
138 1993; von Quadt et al., 1994) with associated volcanics.

139 During the Permian post-Variscan event, the continental crust was affected by a prominent  
140 magmatic and metamorphic episode that deeply modified its lithological composition and crustal  
141 architecture.

142 (1) In the Permian upper crust, alkaline felsic magmas were emplaced and extruded between  
143 295 and 288 Ma (Bernina unit; Spillmann and Büchi, 1993; von Quadt et al., 1994). The  
144 relation between volcanics and tectonic structures remains unclear.

145 (2) In the same age range, the Permian middle crust (Campo and Grosina units) was intruded by  
146 both granitoids and mafic bodies (Fig. 2b), notably the Sondalo gabbroic complex (Petri et

147 al., 2018, 2017; Tribuzio et al., 1999). Locally, mafic intrusions are surrounded by a  
148 migmatitic contact aureole reaching granulite-facies conditions in metasedimentary septa  
149 (Braga et al., 2001; Petri et al., 2016).

150 (3) The Permian lower crust was affected during the Permian by magmatic underplating through  
151 the emplacement of tholeiitic gabbros associated with felsic granulites (Malenco and Lower  
152 Margna units; Hermann et al., 2001; Müntener et al., 2000), indicating an emplacement and  
153 contact metamorphism at around 10 kbar at 281-278 Ma (U-Pb on zircon; Hansmann et al.,  
154 2001).

### 155 **2.1.2. Jurassic rifting history**

156 The post-orogenic and pre-rift Permian lithospheric architecture was disrupted during the  
157 Jurassic by a rifting event that led to the opening of the Alpine Tethys and the formation of the  
158 Adriatic rifted margin. Following Mohn et al. (2012), this process occurred in three stages:

159 (1) The stretching phase (220-185 Ma), which is linked to the formation of rift basins bounded  
160 by widely distributed high-angle normal faults (Eberli, 1988). The related normal faults are  
161 typically listric and rooted at mid-crustal levels (Bertotti, 1991). During this stage, extension  
162 in the brittle upper crust is decoupled from deformation in the mostly mafic and therefore  
163 stronger lower crust. The subcontinental mantle and the lower crust cooled at that time, as  
164 recorded by  $^{40}\text{Ar}/^{39}\text{Ar}$  ages on amphibole from the Malenco unit (Hermann et al., 1997; Villa  
165 et al., 2000).

166 (2) The necking phase (185-175 Ma; thinning phase of Mohn et al., 2012) corresponds to the  
167 localisation of extension within the future distal margin (Froitzheim and Eberli, 1990), and  
168 the continental crust is thinned to ~10 km by the complete obliteration of the ductile middle  
169 crust. The Briançonnais domain starts to be uplifted (e.g., Decarlis et al., 2013), and the  
170 initially ductile middle crust starts to be exhumed at the seafloor along “necking” faults  
171 (Chenin et al., 2019; Mohn et al., 2012).



172 (3) The exhumation phase (175-165 Ma) initiates when the crust in the distal domain has been  
173 thinned below 10 km and is entirely brittle. Detachment faults cut into the lower crust and  
174 the mantle enabling its exhumation to the seafloor (e.g., Mohn et al., 2012). The final  
175 exhumation faults are cross cutting a crustal section that has been reduced during the  
176 previous necking phase.

177 The final architecture of the distal margin is the result of the superposition of all  
178 extensional phases on the inherited pre-rift Permian crust. Rock units that were prior to rifting on  
179 top of each other were juxtaposed and occurred at the end of the rifting stage next to each other as a  
180 consequence of rift-related thinning and exhumation. The evolution of the thinning and exhumation  
181 is documented by the  $P$ - $T$ - $t$  evolution of these rocks as well as by the syn-tectonic sediments  
182 associated with these events (e.g., Petri et al., 2019). In this work, we are mainly focusing on the  
183 evolution of the middle crust from the pre-rift Permian stage to the Jurassic rifting stage. These  
184 rocks were exhumed in between the distal and the proximal domains, in the so-called “necking”  
185 zone, and are today exposed in the Grosina and Campo units.

### 186 **2.1.3. Alpine reactivation**

187 From Late Cretaceous onward, the Jurassic Adriatic margin was reactivated and disrupted  
188 by Alpine convergence. The stacking of the nappes occurred from Late Cretaceous to Tertiary, by  
189 multistage thrusting of proximal domains over more distal domains of the margin. The first major  
190 deformation phase, the Da1 phase (Trupchun phase of Froitzheim et al., 1994) consists of a W- to  
191 NW-directed thrusting occurring between 100 and 76 Ma (see Froitzheim et al., 1994; Handy et al.,  
192 1996), and related to the coeval closure of the Meliata-Vardar domain further East and the onset of  
193 the closure of the Alpine Tethys (Froitzheim et al., 1996; Mohn et al., 2011). This is followed by the  
194 Da2 phase (Duncan-Ela phase of Froitzheim et al., 1994), an extensional phase developing normal  
195 faults and associated recumbent folds (Froitzheim et al., 1994) during Late Cretaceous times (80-67  
196 Ma; Handy et al., 1996). The subsequent Da3 phase (Blaisun phase of Froitzheim et al., 1994) is

197 expressed by large-scale upright open-folds with sub-horizontal E-W trending axis in upper  
198 greenschist facies conditions (Hermann and Müntener, 1992). Da3 has been estimated to be Eocene  
199 in age and is associated with the continental collision between Adria and Europe, but the recently  
200 published ages of Picazo et al. (2019) in the Malenco-Margna nappes shift the prograde  
201 metamorphic event to ca. 63 Ma (U-Pb on rutile) followed by a retrogression at ca. 55 Ma (U-Pb on  
202 titanite). The complete nappe stack is finally deformed by the Engadine and the Insubric Lines, both  
203 related to the Periadriatic fault system.

## 204 **2.2. The Campo–Grosina units**

205 The study area encompasses the Campo and the Grosina units that are part of the middle  
206 Austroalpine units (following Mohn et al., 2011), exposed in SE Switzerland and N Italy. Both units  
207 are separated by the sub-horizontal Eita Shear Zone (ESZ) that can be mapped continuously over  
208 more than 20 km (Figs 2,3; Meier, 2003; Mohn et al., 2012; Petri et al., 2019).

209 The hanging wall of the ESZ, the Grosina unit, is essentially made of Bt-bearing  
210 orthogneiss (also known as “Cristallino di Grosina”, Figs 3a-d,4ab) associated with garnet-  
211 staurolite-bearing metapelite with rare occurrence of andalusite, and crosscut by rare Grt-bearing  
212 granodiorite dykes. Orthogneiss is derived from an Ordovician protolith (see Boriani et al., 2005)  
213 that is preserved from subsequent deformation phases in a few locations only (e.g., Fig. 4a), and  
214 comparable to other pre-Variscan basement units in the Alps (e.g., Strona-Ceneri Zone, Southern  
215 Alps; Zurbriggen et al., 1997). The structural evolution of the Grosina unit, although not well  
216 constrained yet, involves three main fabrics. (1) A sub-vertical foliation (Sg1) is only locally  
217 preserved due to the pervasive overprint during the development of (2) a sub-horizontal planar  
218 fabric (Sg2) formed at amphibolite-facies grade associated with a well-defined lineation with a NW-  
219 SE trend (Figs 3b,4b), with metric sheath folds in the metasediments. Orthogneiss, metasediment  
220 and granodioritic dykes are deformed by local greenschist facies shear zones (Figs 3cd,4c; Petri et  
221 al., 2019), notably approaching the ESZ in which chlorite-rich orthogneiss presents (3) a sub-

222 horizontal Sg3 fabric (Fig. 3d). This new fabric is parallel to the Sg2 fabric and shows a similar  
223 NW-SE trending lineation. The complete section is crosscut by undated and undeformed dykes  
224 (andesite?), potentially coeval to the ones described in the underlying Campo basement (Bianchi  
225 Potenza et al., 1985) and related to the Cenozoic magmatism along the Periadriatic faults system.  
226 Mohn et al. (2012) presented the only ages available for the Grosina basement with  $^{40}\text{Ar}/^{39}\text{Ar}$  dating  
227 on muscovite, ranging between 273 and 260 Ma (Fig. 10a, Supplementary Material S5 and S7).

228         The Campo unit, located in the footwall of the ESZ, contains amphibolite-facies  
229 metasediments (Fig. 4d) derived from an Ordovician protolith (Bergomi and Boriani, 2012) with  
230 paragneiss, micaschist, and minor intercalations of marble and amphibolite. The unit is affected by a  
231 metamorphism reaching up to 6 kbar and 650 °C, synchronous with the development of a N-W  
232 steeply dipping foliation overprinted by a N-E trending sub-vertical Sc2 fabric (Fig. 3aef; for  
233 complete structural description of the Campo unit, see Petri et al., 2018, 2016). Both likely formed  
234 during the Variscan orogeny. It was subsequently intruded by several plutons during the Permian  
235 (300-270 Ma, U-Pb on zircon and Sm-Ng ages; e.g., Petri et al., 2017; Tribuzio et al., 1999),  
236 notably the Sondalo gabbroic complex (Figs 2a,4f) with associated contact metamorphism reaching  
237 upper amphibolite to granulite-facies (Fig. 4e; Braga et al., 2001; Petri et al., 2016), and the  
238 development of a high-temperature Sc3 fabric at the contact of the complex with the host rocks  
239 (Petri et al., 2018). Locally, undeformed Permian pegmatitic dykes crosscut the Sc2 foliation (Fig.  
240 3ef) and were dated at 255-250 Ma (Sm-Nd ages; Sölva et al., 2003). Rb–Sr ages on muscovite  
241 from pegmatites and granitoids range between 277 and 251 Ma and between 263 and 97 Ma on  
242 biotite (see supplementary Table S5 for references; Fig. 10b).  $^{40}\text{Ar}/^{39}\text{Ar}$  and K–Ar ages from other  
243 lithologies are spanning between 217 and 187 Ma for muscovite and from 189 to 122 Ma for biotite  
244 (Fig. 10b, Supplementary Material S5 and S7). Hundred meters below the ESZ, the Sc2 planar  
245 fabric of the Campo unit is rotated by Fc4 recumbent folding with E-W trending axes and shallow

246 W-dipping axial planes, affecting also the Permian pegmatites (Fig. 3e). The rest of the Campo unit  
247 is little affected by the ESZ-related deformation.

### 248 **2.3. The Eita shear zone**

249 Both Fc4 recumbent folding in the Campo unit and Sg3 fabric in the Grosina unit intensify  
250 approaching the ESZ, and overprint Permian fabrics and pegmatites: the activity of the ESZ is  
251 therefore post-Permian (Mohn et al., 2012; Petri et al., 2019, 2018, 2016).

252 The Campo–Grosina units are thrust during late-Cretaceous (Da1; Froitzheim et al., 1994)  
253 by the overlying Filladi di Bormio unit (Fig. 2). This unit is made of prograde greenschist facies  
254 metapelites of unknown age (Boriani et al., 2011). At the contact with the Filladi di Bormio,  
255 micaschists from the Campo unit are retrogressed into greenschist facies rocks.

256 The ESZ was previously interpreted as a late-Cretaceous Da1 Alpine thrust (Meier, 2003).  
257 However, the ESZ being cross-cut by Da1 thrusts in several locations (Fig. 2) and with new  
258 geochronological arguments, Mohn et al. (2012) interpreted this structure as a Jurassic extensional  
259 shear zone active during rifting (200–180 Ma) with possible, but only minor Alpine reactivation.  
260 The superposition of both units through the activity of the ESZ has therefore been established  
261 before the first Alpine deformation occurred (Da1), hence during rifting.

### 262 **3. ANALYTICAL METHODS**

263 Former geochronological investigations were focused along the contact between the  
264 Campo and the Grosina units (Meier, 2003; Mohn et al., 2012) in order to constrain the age of  
265 activity of the ESZ. By contrast, the precise characterization of exhumation and cooling history of  
266 these units with modern dating techniques are lacking. To this end, we stepped away from the ESZ  
267 and selected amphibole, muscovite and biotite grains from various lithologies in both units to carry  
268 out petrological characterization and  $^{40}\text{Ar}/^{39}\text{Ar}$  step-heating dating (sample location indicated in Fig.  
269 2, sampling coordinates in Table 1, and Supplementary Tables S2-S4). Mineral abbreviations follow  
270 IUGS recommendations after Siivola and Schmid (2007).

### 271 3.1. Mineral composition

272 Major element mineral compositions were determined for dated samples at the Institute of  
273 Earth Sciences of the University of Lausanne using a JEOL JXA-8530F HyperProbe EPMA  
274 equipped with five WDS detectors, and operating at 15 kV acceleration voltage and 15 nA.  
275 Particular attention was paid to the celadonite (Si content) and paragonite ( $Pg = Na / (Na + K)$ )  
276 substitutions in muscovite and its Ti content that are pressure and temperature sensitive. Biotite was  
277 monitored through  $X_{Mg} = Mg / (Fe_t + Mg)$  and its Ti content to estimate temperature; and amphibole  
278 was monitored through its Al, Na and Ca/K content. Thermobarometric estimates were derived  
279 when applicable using the Wu and Chen (2015) Ti-in-muscovite and Henry (2005) Ti-in-biotite  
280 thermometers. The Ernst and Liu (1998) thermobarometer based on the Al and Ti content of  
281 amphibole was applied to MORB-type lithologies, the Schmidt (1992) Al-in-hornblende barometer  
282 was applied to tonalitic lithologies. Measured mineral compositions and calculated  
283 thermobarometric estimates are reported in Fig. 5, Table S1, and compiled in Table 1.

### 284 3.2. $^{40}Ar/^{39}Ar$ geochronology

285 Multigrain  $^{40}Ar/^{39}Ar$  isotopic analyses were performed on muscovite, biotite, and  
286 amphibole by laser step-heating at the Faculty of Earth and Life Sciences, VU Amsterdam. Minerals  
287 were separated at the mineral separation laboratory of the Faculty of Earth and Life Sciences, VU  
288 Amsterdam after crushing, sieving between 250 and 500  $\mu m$  and cleaning ultrasonically with  
289 demineralized water. Desired minerals were concentrated using a vibrating table, heavy liquids, and  
290 a magnetic separator to be finally selected by hand picking under the binocular. Mineral packages  
291 wrapped in Al-foil were co-irradiated with DRA sanidine standard packages wrapped in Cu-foil as  
292 fluence monitor at the NRG-Petten HFS facility in the Netherlands for 18 hours.

293 Measurements were performed after ca. 6 months to allow  $^{37}Ar$  decay in Ca-rich  
294 amphiboles. Samples were pre-heated to  $\sim 200$  °C to remove undesirable atmospheric argon. Total  
295 fusion and step-heating, for standards and samples, respectively, were carried out by a Synrad CO<sub>2</sub>

296 laser combined to a Raylease scan-head system to scan the totality of the mounted grains. Isotopic  
297 ratios were evaluated using a MAP 215-50 mass spectrometer with both secondary electron  
298 multiplier and Faraday collector. Blanks were measured every two (muscovite and biotite) to four  
299 (amphibole) samples measurements and before and after each experiment. Mass discrimination was  
300 made through  $^{40}\text{Ar}/^{36}\text{Ar}$  air pipette aliquots measurement.

301         Ages were calculated using ArArCalc (Koppers, 2002) using the decay constants of Steiger  
302 and Jäger (1977), and the atmospheric  $^{40}\text{Ar}/^{39}\text{Ar}$  ratio of  $298.56 \pm 0.1\%$  (Lee et al., 2006). For each  
303 sample, the J irradiation parameter was determined through linear interpolation between each  
304 standard. Standard calibration of 24.99 Ma from Wijbrans et al. (1995) was used for VU89  
305 experiments, whereas calibration of 25.42 Ma from Kuiper et al. (2008) was used for experiments  
306 VU97D. Raw dating analytical results are provided in Supplementary Table S2 for muscovite,  
307 Supplementary Table S3 for biotite and Supplementary Table S4 for hornblende. Results are  
308 represented as step-release spectra and inverse isochrons in Figs 6-9, in which errors are reported at  
309  $2\sigma$  (95% level of confidence), together with the mean standard weighted deviation (MSWD).

310         Ages are subsequently discussed using Monte Carlo simulations to generate random  
311 temperature-time paths used to calculate synthetic argon ages in micas with DiffArgP (Warren et al.,  
312 2012a), a version of DiffArg (Wheeler, 1996) that integrates the P-dependency of muscovite  
313 diffusion parameters (Harrison et al., 2009), and the diffusion parameters of Harrison et al. (1985)  
314 for biotite. A total of 10,000 paths per scenario were generated, defined by 4 nodes for the Grosina  
315 unit (G1 to G4) and up to 5 nodes for the Campo unit (C1 to C5). Nodes are located in T-t-P boxes  
316 constrained from geological arguments (Fig. 11b) and that may overlap. Each node is randomly  
317 picked within a box, with the only rule that age should always decrease (e.g., from node G1 to node  
318 G2). For each path, calculated muscovite and biotite bulk ages (cylindrical grains with 0.25 mm  
319 radius) were compared to the measured ages through a merit function compiled using the  
320 probability density function based on the population of measured ages and associated uncertainties.

321 Paths reaching high probabilities ( $p$ ) translate a high fit between the modelled age and the set of  
322 measured ages (Fig. 12); paths with  $p < 0.05$  were discarded. The model space is also explored by  
323 age of key nodes for the temperature-time evolution and key cooling rates (Fig. 12). More details  
324 are reported in Supplementary Text S8.

#### 325 4. SAMPLE PETROLOGY

326 We selected two samples from the Grosina unit and eight samples from the Campo unit  
327 including granodiorite, diorite, amphibolites and metapelites (microphotographs of samples are  
328 provided in Fig. 4 and Supplementary Material S6). All samples were selected for having well  
329 preserved minerals that crystallized before or during the emplacement of the Sondalo gabbroic  
330 complex, and the weakest possible post-Permian deformation microstructures.

##### 331 4.1. Grosina unit

332 Two samples were collected from the Grosina unit: *Sample GM119* is a Bt-granodiorite  
333 that underwent only minor Dg2 deformation. Quartz presents lobate grain boundaries, but large  
334 euhedral and inclusion-free biotite and feldspar preserved their magmatic structure assemblage (Fig.  
335 4a). Biotite grains have a low  $X_{Mg}$  (0.37-0.43), with core richer in Ti (0.15-0.21 a.p.f.u.) than rim  
336 (0.08-0.18 a.p.f.u.), reaching temperature estimates of  $681 \pm 16$  °C and  $617 \pm 50$  °C, respectively.

337 Additionally, *sample GM264* is a Bt–Ms paragneiss with minor garnet content with Bt–Ms  
338 crystals aligned in the Sg2 foliation alternating with porphyroblastic feldspar. Quartz presents an  
339 equant microstructure with numerous pinning structure on mica layers indicative of pervasive static  
340 recrystallization. A few muscovite grains present high Si-content ( $> 3.1$  a.p.f.u.) with grain core  
341 having a slightly higher Si content than the rim, *Pg* is rather variable (0.07-0.15 a.p.f.u.). The Ti-  
342 content of muscovite is low ( $< 0.03$  a.p.f.u.), resulting in similar temperature estimates between  
343 cores ( $578 \pm 20$  °C) and rims ( $561 \pm 33$  °C). Biotite is characterized by low Ti-content ( $< 0.10$   
344 a.p.f.u.) and a moderate  $X_{Mg}$  ( $\sim 0.5$ ), with no compositional difference between core and rim;  
345 calculated temperature estimate reaches  $570 \pm 8$  °C.

## 346 4.2. Campo unit

347 Three samples of metasediments have been selected from the Campo unit. *Sample CP003b*  
348 is a restitic migmatite sampled in the contact aureole of the Sondalo gabbroic complex. The sample  
349 presents the Sc3 foliation synchronous to the emplacement of the Sondalo gabbro; muscovite is  
350 randomly oriented in quartz-rich levels and appears post-kinematic and post peak metamorphic  
351 conditions. Quartz forms polygonal aggregates and presents weak undulous extinction. Muscovite  
352 presents a few sillimanite inclusions but no compositional zoning, a high  $Pg$  ( $\sim 0.13$  a.p.f.u.) and Ti-  
353 content (0.04-0.07 a.p.f.u.), and a low Si-content ( $\sim 3.02$  a.p.f.u.), constraining a temperature of  $691$   
354  $\pm 33$  °C.

355 *Sample CP028c* is a micaschist collected close to the Sondalo gabbro. Quartz shows very  
356 weak development of sub-grain rotation microstructure, whereas plagioclase is not deformed.  
357 Porphyroblastic muscovite and biotite are aligned in the Sc2 fabric. They are kinked in places and  
358 recrystallized in micrometric grains. Muscovite porphyroblasts have very low  $Pg$  ( $\sim 0.07$ ) and Si-  
359 content ( $\sim 3.00$  a.p.f.u.), and rather high concentration in Ti (0.05 a.p.f.u.), without any difference  
360 between core and rim. Temperature estimates reach  $695 \pm 8$  °C. Biotite porphyroblasts are  
361 homogeneously rich in Ti ( $\sim 0.20$  a.p.f.u.) and have a moderate  $X_{Mg}$  ( $\sim 0.46$ ), constraining a  
362 temperature of  $698 \pm 16$  °C.

363 *Sample CM260* is a Grt-Ms-Bt micaschist with Ms-Bt aggregates aligned in the main pre-  
364 Permian Sc2 fabric of the Campo basement. Micas are inclusion-free, and a few large biotite  
365 crystals are included in garnet. Biotite is locally slightly chloritized, but quartz presents equant  
366 textures and is pinned to micas grains. Muscovite grains have moderately low Si (3.04 a.p.f.u.) and  
367 Ti (0.03 a.p.f.u.), and high  $Pg$  (0.14-0.16) with rims being slightly richer in  $Pg$  than cores.  
368 Temperature estimates are  $572 \pm 32$  °C for cores and  $562 \pm 44$  °C for rims. Biotite has low  $X_{Mg}$  ( $\sim$   
369 0.40) and Ti (0.12 a.p.f.u.), without presenting compositional difference between core and rim;  
370 temperature estimates reach  $611 \pm 14$  °C.



371 Two amphibolites were selected. *Sample CP010g* presents hornblende aligned in the pre-  
372 Permian Sc2 fabric. Quartz and plagioclase are present in polygonal aggregates, whereas  
373 isogranular amphibole composition points to a magnesio-hornblende, with low Al-, Na- and Ti-  
374 content ( $Al^{tot} = 1.99$  a.p.f.u.,  $Na^{M4} = 0.14$  a.p.f.u. and  $Ti = 0.07$  a.p.f.u., respectively), high Ca/K  
375 ratio ( $\sim 31$ ) and no difference between core and rim. In the lack of proper mineral assemblage, no  
376 thermobarometric estimates were performed.

377 *Sample CM234* is an amphibolite with hornblende aligned in the Sc2 fabric too, but in the  
378 vicinity of the ESZ. Amphibole is euhedral and forms continuous beds separated by plagioclase  
379 aggregates, both showing no evidence for dynamic recrystallization. Amphibole composition is a  
380 tschermakitic to magnesian hornblende, with moderate Al- content ( $Al^{tot} = 2.30$  a.p.f.u.) and low Na-  
381 and Ti-content ( $Na^{M4} = 0.10$  a.p.f.u.,  $Ti = 0.08$  a.p.f.u.), with no difference between amphibole core  
382 and rim. The rock assemblage does not allow for proper thermobarometric estimates.

383 Finally, three magmatic rocks from the Sondalo gabbroic complex were selected. *Sample*  
384 *CP031a* is a garnet-bearing granodiorite collected at the margin of the Sondalo gabbroic complex.  
385 The rock preserves its magmatic structure and mineralogy, with subhedral plagioclase and biotite  
386 and xenomorphic amphibole. Thin polygonal quartz aggregates lie along plagioclase grain  
387 boundaries with no evidence for sub-solidus deformation. Amphibole contains in places biotite  
388 inclusions and has a composition of a ferro-tschermakitic hornblende with Al-poor and Ti-rich cores  
389 ( $2.29$  a.p.f.u. and  $0.09$  a.p.f.u., respectively) compared to rims ( $2.67$  a.p.f.u. and  $0.05$  a.p.f.u.,  
390 respectively), with a homogeneous Ca/K ratio ( $\sim 21$ ). Cores are slightly richer in Na ( $Na^{M4} = 0.19$ )  
391 than rims ( $Na^{M4} = 0.16$ ). The application of the Schmidt (1992) barometer constrains a P estimate of  
392 9 kbar (Table 1). Biotite contains a few inclusions of opaque minerals, is not zoned and has a low  
393 Ti-content ( $0.07$ - $0.14$  a.p.f.u.) and moderate  $X_{Mg}$  ( $0.44$ ), pointing to a temperature of  $581 \pm 40$  °C.  
394 This temperature estimate is slightly off the range of temperature of the calibration of Schmidt  
395 (1992), and the resulting pressure is higher than what was estimated in Petri et al. (2016).

396 *Sample CP032c* is an amphibole-bearing gabbro from the centre of the Sondalo gabbro,  
397 preserving a poikilitic magmatic structure with interstitial amphibole rimming pyroxene cores with  
398 magnetite inclusions. Amphibole is an iron-rich pargasite, with a high Ti-content (Ti = 0.34 a.p.f.u.),  
399 and a moderate Na- and Al-content ( $\text{Na}^{\text{M4}} = 0.25$  a.p.f.u.,  $\text{Al}^{\text{tot}} = 2.29$  a.p.f.u.). Conditions of  
400 equilibration of the amphibole are constrained by the Ernst and Liu (1998) thermobarometer to 950  
401 °C for 6 kbar, in agreement with Petri et al. (2016).

402 *Sample CP034c* is a granodiorite from the centre of the Sondalo gabbro sampled close to a  
403 metapelitic xenolith. It contains large euhedral garnet and subhedral biotite and plagioclase showing  
404 a preferred orientation: the preservation of magmatic minerals and the absence of evidence for sub-  
405 solidus recrystallization indicates that the fabric formed at supra-solidus conditions. Biotite grains  
406 show no compositional difference between core and rim, with low Ti (0.07-0.08 a.p.f.u.) and high  
407  $X_{\text{Mg}}$  (0.60-0.62), constraining a temperature of  $563 \pm 10$  °C.

## 408 5. ARGON DATING

409 In most cases, the analytical data were too disturbed to define “plateau” ages that normally  
410 require at least three consecutive steps within their  $2\sigma$  errors encompassing more than 70% of the  
411 total released  $^{39}\text{Ar}$ : sub-plateau were defined by correlating the age spectrum and the estimated  
412 K/Ca ratio of each step. In case of suspected excess argon shown e.g., by older steps in the  
413 beginning of the spectrum, the inverse isochron method was used, allowing the identification of the  
414 extraneous source of argon. We analysed muscovite (Fig. 7) and biotite (Figs 8,9) from both the  
415 Campo and the Grosina units. In the absence of amphibole-bearing lithologies in the Grosina unit,  
416 amphiboles (Fig. 6) were only dated in the Campo unit.

## 417 **5.1. Grosina unit**

### 418 **5.1.1. Muscovite ages**

419 *Muscovite GM264*. The highly disturbed step-release spectra give a total age of  $262 \pm 3$  Ma  
420 with a very high Mean Standard Weighted Deviation (MSWD). The inverse isochron age reaches  
421  $261 \pm 2$  Ma (Fig. 7a), and is preferred over the step-release spectra.

### 422 **5.1.2. Biotite ages**

423 *Biotite GM119*. Eight out of 14 measurements define a “sub-plateau” at  $280 \pm 3$  Ma (Fig.  
424 8a). Measurement points are reasonably well grouped close to the  $^{36}\text{Ar}/^{40}\text{Ar} = 0$  line in the inverse  
425 isochron age but present a trend pointing to excess argon. Moreover, ages are decreasing with the  
426 increasing laser power usually indicating biotite–chlorite intermixing resulting in  $^{39}\text{Ar}$  recoil from  
427 biotite to chlorite during irradiation (e.g., Lo and Onstott, 1989). However, the chlorite content here  
428 is likely too low to be identified with high  $^{37}\text{Ar}/^{39}\text{Ar}$  ratios. It is therefore recommended to use the  
429 inverse isochron age of  $246 \pm 13$  Ma.

430 *Biotite GM264*. Fifteen of the 21 steps define a sub-plateau at  $246 \pm 1$  Ma, coeval with the  
431 total fusion age (Fig. 8b). As the spectrum is quite disturbed, the isochron age of  $245 \pm 2$  Ma should  
432 be used rather than the “plateau” or total fusion age.

## 433 **5.2. Campo unit**

### 434 **5.2.1. Amphibole ages**

435 *Hornblende CP010g*. Isotopic measurements show strong excess  $^{40}\text{Ar}$  in the mineral (Fig.  
436 6a) with hardly realistic ages. Both total fusion age of  $510 \pm 4$  Ma and inverse isochron age of  $283$   
437  $\pm 90$  Ma are geologically meaningless due to the high excess argon content.

438 *Hornblende CP031a*. The spectrum is highly disturbed, with old ages in the beginning of  
439 the experiment, decreasing down to 183.50 Ma for the last step. More than 40 % of the released  $^{39}\text{Ar}$   
440 constitute steps older than the age of emplacement of the Sondalo gabbro determined by U-Pb on  
441 zircon (Fig. 10; Petri et al., 2017), confirming a strong component of excess argon (Fig. 6b). On the

442 inverse isochron plot, data are moderately aligned between air and a single radiogenic reservoir: the  
443 mean inverse isochron constrains an age of  $205 \pm 9$  Ma, and data can be bracketed between an old  
444 and a young envelope with inverse isochron ages of  $210 \pm 4$  Ma and  $177 \pm 13$  Ma, respectively.

445 *Hornblende CP032c*. Isotopic measurements show strong excess  $^{40}\text{Ar}$  (Fig. 6c); the total  
446 fusion age of  $493 \pm 4$  Ma, the inverse isochron age of  $344 \pm 8$  Ma and every single measurement  
447 step are older than the emplacement age of the Sondalo gabbroic complex determined by Sm–Nd or  
448 U–Pb (Fig. 10), pointing to unrealistic ages.

449 *Hornblende CM234*. Isotopic measurements show strong excess  $^{40}\text{Ar}$  in the mineral (Fig.  
450 6d) with continuously decreasing step ages. Both total fusion age of  $493 \pm 4$  Ma and the inverse  
451 isochron age of  $135 \pm 20$  Ma constrained by very scattered data points are geologically  
452 meaningless.

### 453 **5.2.2. Muscovite ages**

454 *Muscovite CP003b*. Five consecutive steps out of 13 steps constrain an age of  $217 \pm 2$  Ma  
455 close to the total fusion age of  $216 \pm 2$  Ma (Fig. 7b). Data are very scattered on the inverse isochron  
456 plot suggesting a strong mixing between air and radiogenic argon, indicating an age of  $218 \pm 2$  Ma.  
457 This may suggest weathered muscovite (e.g., illitization; de Jong et al., 2001) but with only air  
458 mixed with one radiogenic  $^{40}\text{Ar}$  source; the inverse isochron age should hence be preferred.

459 *Muscovite CM260*. Eighteen consecutive steps loosely define an age of  $182 \pm 2$  Ma, coeval  
460 with the total fusion age of  $182 \pm 1$  Ma (Fig. 7c). Inverse isochron confirms that the sample contains  
461 only one argon reservoir with a minor air component and gives an age of  $182 \pm 2$  Ma that is  
462 preferred over the step-release spectrum.

463 *Muscovite CP028c*. The first 4 consecutive steps loosely define an age of  $176 \pm 4$  Ma, with  
464 a total fusion age of  $178 \pm 2$  Ma (Fig. 7d). On the inverse isochron, data are well grouped and  
465 define an age of  $178 \pm 5$  Ma. In the view of the disturbed step-released spectrum, the inverse  
466 isochron age is preferred.

467 **5.2.3. Biotite ages**

468 *Biotite CP034c.* Ages become progressively older during the experiment, pointing to a  
469 partial argon loss. Eight out of 14 steps define an age of  $216 \pm 2$  Ma, with a total fusion age of  $212$   
470  $\pm 2$  Ma (Fig. 8c). The inverse isochron plot shows scattered data points, with an inverse isochron at  
471  $214 \pm 3$  Ma. The inverse isochron age is preferred to the sub-plateau age, because of the high  
472 MSWD of the latter.

473 *Biotite CP031a.* Seven consecutive steps selected from 14 steps constrain a reasonable  
474 plateau of  $173 \pm 2$  Ma with a MSWD = 0.7, totalling more than 70 % of the released  $^{39}\text{Ar}$  (Fig. 8d).  
475 Despite the spread of data in the inverse isochron diagram, the plateau age is reasonable. The total  
476 fusion age is  $170 \pm 2$  Ma and the inverse isochron age is  $171 \pm 2$  Ma. The plateau age is therefore  
477 preferred.

478 *Biotite CM260.* Six out of 20 steps defines a “sub-plateau” age of  $171 \pm 1$  Ma close to the  
479 total fusion age of  $172 \pm 1$  Ma and preferred over the inverse isochron age of  $171 \pm 2$  Ma (Fig. 9a).  
480 Both the inverse isochron plot pointing to a quite strong contribution of air in the biotite and the  
481 decreasing age indicate that biotite is slightly chloritized (Lo and Onstott, 1989), in agreement with  
482 petrographic observations. This is however not identified by the  $^{37}\text{Ar}/^{39}\text{Ar}$  ratio of these steps due to  
483  $^{37}\text{Ar}$  concentrations below detection limits.

484 *Biotite CP028c.* Ten consecutive steps out of 14 steps define a sub-plateau age of  $174 \pm 2$   
485 Ma (MSWD = 8) and a total fusion age of  $173 \pm 3$  Ma (Fig. 9b). The inverse isochron presents  
486 scattered data but with a consistent age of  $174 \pm 2$  Ma. The plateau age is here preferred.

487 **6. DISCUSSION**

488 **6.1. Significance of measured ages**

489 **6.1.1. Muscovite ages**

490 Dated muscovite from the Campo unit have a large *Pg* and Ti compositional range,  
491 pointing to various temperatures of equilibrium, as confirmed by the thermometry calculations

492 (Table 1). Despite variable composition, ages of muscovite CM260 and muscovite CP028c are  
493 similar within error. This is however not the case for muscovite CP003b that is 30 Myr older, with  
494 an identical Si and Ti content, and a *Pg* between the two other samples. All muscovite grains are,  
495 however, not occurring in the same structural position: the older muscovite CP003b is growing  
496 post-deformation and post-peak metamorphic conditions in a quartz-rich layer. Conversely, the two  
497 other samples present grains that crystallized before or during Permian times based on the following  
498 elements: (1) both CM260 and CP028c samples have preserved muscovite grains aligned in the Sc2  
499 fabric that predates the intrusion of the Sondalo gabbroic complex, (2) muscovites show no  
500 microstructural evidence for recrystallization in CM260, and only large porphyroclasts were  
501 selected in CP028c, and (3) muscovite still preserves its peak-metamorphic compositions as attested  
502 by applied thermometry (695 °C and 567 °C, respectively). This is in line with estimates obtained  
503 by biotite thermometry (see below) as well as for the regional and contact metamorphism around  
504 the Early Permian Sondalo gabbro (Petri et al., 2017, 2016). These temperature estimates are  
505 notably above the expected closure temperature of the argon system in muscovite. Therefore, the  
506 overlapping 182 and 178 Ma ages of muscovite CM260 and muscovite CP028c can be interpreted  
507 as cooling ages. This is not straightforward for the older muscovite CP003b that can also be  
508 interpreted as a crystallization age given its structural position. Interestingly, this 217 Ma age is  
509 coeval to the older ages of amphibole in the lower crustal Malenco unit (Villa et al., 2000), pointing  
510 to a potential role of deformation-induced fluid mobilization during early rifting stages. An  
511 alternative scenario would be that the older muscovite ages are caused by a strong component of  
512 extraneous argon, related to the Permian emplacement of the Sondalo gabbro.

513           The age of muscovite GM264 from the Grosina unit is clearly different than the age of  
514 muscovite CM260 and muscovite CP028c from the Campo unit, highlighting an age contrast  
515 between the two units. In both units, mineral compositions are variable and not correlated with ages.  
516 Step-release spectra are rather disturbed for all dated muscovite grains, notably from the Grosina

517 unit. Yet, the inverse isochron plot presents rather clustered measurements, pointing to systematic  
518 single radiogenic argon reservoirs, with low air contamination thus validating the various ages.

### 519 **6.1.2. Biotite ages**

520 Three of the four biotite ages from the Campo unit lie in the 174-171 Ma range; biotite  
521 CP034c appears as an outlier age, evaluated at 216 Ma. All samples present either the pre-Permian  
522 Sc<sub>2</sub> fabric (biotite from micaschists CM260 and CP028c), or preserve Permian magmatic textures,  
523 microstructures (biotite from igneous CP034c and CP031a). The equilibrium temperature estimates  
524 based on biotite composition are overlapping with muscovite thermometry for the two  
525 metasediments (611 °C and 698 °C for CM260 and CP028c, respectively; Table 1), and are  
526 consistent with estimates from Petri et al. (2016). The biotite temperature estimates for the two  
527 granodiorites (581 °C and 563 °C) are lower than expected for the crystallization temperature of  
528 biotite for such melt composition. This is very likely due to the fact that Henry (2005) Ti-in-biotite  
529 thermometer was calibrated with metapelites only. For all samples, biotite thermometry reaches  
530 temperatures largely above the expected closure temperature of the argon system in biotite. The  
531 three youngest samples encompass the full range of biotite composition, showing that such mineral  
532 compositions had no effect on apparent argon retention. This is however not the case for the older  
533 biotite CP034c that was separated from a Permian granodiorite, and not from a metasediment; it has  
534 the composition of a phlogopite with much higher  $X_{Mg}$ , likely caused by a higher bulk rock Mg-  
535 number. This strong compositional difference may increase argon retention, increasing in turn the  
536 closure temperature for the biotite grains of this sample (potentially a few tens of °C; Harrison et  
537 al., 1985). This effect of biotite composition on Ar retention is, however, not identified for annite-  
538 rich biotite ( $X_{Mg}$  below 0.5), in agreement with Grove and Harrison (1996). Based on the structural  
539 control, the preserved high-temperature mineral textures and compositions, we can state that all  
540 biotite ages from the Campo unit can be interpreted as cooling ages, but with a potentially higher

541 closure temperature for sample CP034c. Alternatively, this older age for CP034c can be due to  
542 incorporation of extraneous Ar, as discussed hereafter.

543         Regarding the Grosina unit, both biotite ages are coeval at 246-245 Ma. Related grains are  
544 occurring in two different and likely not coeval structural contexts, minerals being either aligned in  
545 Sg2 (micaschist GM264) or still preserving magmatic biotite from the granitic protolith  
546 (orthogneiss GM119). Temperatures estimated from both biotite and muscovite of sample GM264  
547 are notably identical (569 °C and 570 °C, respectively). Derived ages are therefore interpreted as  
548 cooling ages. Measured biotite have variable Ti-content but rather low  $X_{Mg}$ , similar to biotite from  
549 the Campo unit (except biotite CP034c). Therefore, a simple difference in mineral composition does  
550 not explain the observed difference in age between the Campo and the Grosina units.

### 551         **6.1.3. Amphibole ages**

552         Whereas hornblende CP032c reaches pressure estimates consistent with previous works,  
553 hornblende CP031a leads to slightly higher P that may be caused by a high-Al bulk rock  
554 composition compared to the tonalitic calibration of Schmidt (1992), which is also attested by the  
555 presence of Grt in this sample. The paragenesis of the two other amphibole-bearing samples,  
556 however, excluded the use of Al-in-hornblende barometry. Despite structural and metamorphic  
557 control, most amphibole ages led to meaningless ages, being strongly affected by extraneous argon  
558 contamination. Among the four, only hornblende CP031a was prone to derive a geologically  
559 relevant age using the inverse isochron, bracketed between  $210 \pm 4$  and  $177 \pm 13$  Ma.

### 560         **6.1.4. Possible reasons for excess $^{40}\text{Ar}$**

561         The assumption that the argon system was initially open and above the “closure  
562 temperature” is expected to be valid for long term residence in the middle and lower continental  
563 crust for standard temperature and pressure condition (e.g., Warren et al., 2012b). We however  
564 noticed that several of our samples present excess-Ar based on several points: (1) amphibole  
565 sampled in but also around the Permian Sondalo gabbro have older  $^{40}\text{Ar}/^{39}\text{Ar}$  ages than the age of



566 intrusion of the pluton determined by U–Pb on zircon and Sm–Nd ages (Fig. 10b; Petri et al., 2017;  
567 Tribuzio et al., 1999), (2) amphibole systematically presents dates decreasing during the  
568 experiments with increasing %<sup>39</sup>Ar released, and (3) amphibole have moderately to highly scattered  
569 measurement points in inverse isochron plots. The two older mica ages recovered from the Campo  
570 unit (muscovite CP003B and biotite CP034c) may also be older by excess argon, although  
571 alternative scenarios are also plausible (see discussion above).

572         We suggest that the high <sup>40</sup>Ar/<sup>36</sup>Ar ratio of the Sondalo gabbro original melt is the first  
573 possible source of excess-Ar in the pluton and its country rocks: excess-Ar was mobilized due to  
574 heat introduced into the unit by the mantle-derived mafic intrusion that disturbed <sup>40</sup>Ar/<sup>39</sup>Ar ratios.  
575 As the melt sources in mantle partial melting, it has a low K-content and associated low radiogenic  
576 Ar production. Its <sup>40</sup>Ar/<sup>36</sup>Ar ratio is consequently very close to the high ratio of the mantle (e.g.,  
577 Staudacher et al., 1989). The second possible source may be related to the intrusion of such a melt  
578 in a K-rich environment. Although Ar is expected to leave and diffuse out of the mineral, the higher  
579 environmental Ar inhibited this process and potentially triggered back diffusion from the external  
580 reservoir into the mineral. This process was documented in country rocks few meters around  
581 intrusions (Hyodo and York, 1993). However, in the Campo unit, disturbed analyses are located up  
582 to 5 km away from the intrusion, questioning the extent of the gabbro below the nowadays exposed  
583 surface of the Campo unit, or the ability of a postulated excess <sup>40</sup>Ar “wave” to affect host rocks.

584         Micas are expected to be more affected by extraneous argon incorporation due to the  
585 higher partition coefficient between extraneous fluids and micas (Kelley, 2002) compared to  
586 amphibole. The amphiboles of the Campo unit were, however, significantly more affected by  
587 excess-Ar than the micas. This is likely due to the more sensitive character of amphibole caused by  
588 a low <sup>40</sup>K content and therefore a lower internal radiogenic <sup>40</sup>Ar concentration (Kelley, 2002). In  
589 addition, this extraneous <sup>40</sup>Ar contamination occurred at a time micas were still residing above their  
590 closure conditions. Only muscovite CP003b and biotite CP034c, being ca. 30 Myr and ca. 40 Myr

591 older than other dated muscovites and biotites from the Campo unit are potentially retracing this  
592 high, inherited  $^{40}\text{Ar}/^{36}\text{Ar}$  ratio. In this scenario, the older “sub-plateau” age presented in Figures 7b  
593 and 8c may be due to extraneous  $^{40}\text{Ar}$  incorporation homogeneously distributed in the mineral  
594 lattice (Foland, 1983) potentially brought by fluids (Kelley, 2002).

## 595 **6.2. Age difference between the Campo and the Grosina units**

596 The Grosina and the Campo units apparently underwent contrasting evolutions (Fig. 10).  
597 The Grosina unit shows consistent muscovite ages between 273 and 261 Ma older than the biotite  
598 ages at 248–246 Ma. By contrast, the Campo unit has a poorly constrained amphibole age (between  
599  $210 \pm 4$  and  $177 \pm 13$  Ma) partly overlapping muscovite ages spanning between 182 and 176 Ma.  
600 Similar muscovite ages of the Campo unit further south were dated by Meier (2003) at  $188 \pm 4$  Ma.  
601 Dated biotite separates from the Campo unit are slightly younger, ranging between 174 and 171 Ma  
602 ( $\pm 1$ – $2$  Ma). Figure 10ab clearly illustrates that the Campo unit has ages significantly younger than  
603 the Grosina unit. The transition between the two basement units, marked in the field by the ESZ, is  
604 characterized by a sharp change in  $^{40}\text{Ar}/^{39}\text{Ar}$  ages (Fig. 10c). In both units, dated minerals were  
605 separated from high-temperature assemblages inferred to be at least Carboniferous or Permian in  
606 age. In agreement with petrological and structural considerations, we consider these ages record the  
607 cooling rather than the crystallization of minerals.

## 608 **6.3. Cooling and exhumation history of the Campo and Grosina basements**

609 These new ages, together with a compilation of published U–Pb, Sm–Nd, Rb–Sr, K–Ar  
610 and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages performed on both the Grosina and Campo basements (see Fig. 10,  
611 Supplementary Material S5 and S7), constitute a consistent dataset to estimate the respective  
612 cooling history of the two units. Deciphering the thermal history and cooling rates is hampered by  
613 the fact that closure temperatures are sensitive to the cooling rate itself, mineral composition, and  
614 grain size. To a first approximation, geochronological data can be linked to the closure temperature  
615 of the probed chronometer/mineral system (details presented in Supplementary Figure S7). Two

616 distinct events of cooling can be identified (Fig. 10 and Supplementary Material S7): a Permo-  
617 Triassic cooling, and a subsequent Jurassic cooling. We characterized these two cooling events  
618 using a Monte Carlo -based approach (Figs 11,12).

### 619 **6.3.1. Thermal evolution during Permian post-orogenic extension**

620 From our dataset, the Permian cooling phase is best recorded in the Grosina unit with the  
621  $261 \pm 2$  Ma muscovite age, and the  $246 \pm 13$  and  $245 \pm 2$  Ma biotite ages, indicating a cooling  
622 during Permo-Triassic times. At a regional scale, this period corresponds to the end of the intense  
623 magmatic and metamorphic event immediately following the Variscan orogeny and its collapse  
624 (Fig. 13ab; Petri et al., 2017). This period is associated with the emplacement of both mafic and  
625 felsic intrusives from the lower continental crust to the upper continental crust (Hermann and  
626 Rubatto, 2003; von Quadt et al., 1994), and regional high-temperature metamorphism (see  
627 compilation in Fig. 11; Hermann and Rubatto, 2003; Petri et al., 2016). The cessation of the HT  
628 regional conditions implies a progressive thermal relaxation and deepening of the isotherms that  
629 caused a thermal subsidence permitting the deposition of Triassic shallow-marine sediments (Fig.  
630 13c; Eberli, 1988).

631 These constrains were used to design the temperature-time paths of the Grosina unit used  
632 as entry values for Monte Carlo DiffArgP simulations (Fig. 11). An Early Permian node G1 is  
633 located at  $550 \pm 150$  °C and 3 kbar that is expected to be the thermal maximum of the Grosina unit  
634 based on the mineralogical assemblage and the presence of andalusite in the metapelites that are  
635 interpreted to be Permian in nearby Austroalpine basement units (Petri et al., 2016; Schuster and  
636 Stüwe, 2008). Node G1 is followed by node G2, marking the end of the Permian cooling. The age  
637 and temperature of this stage is largely unknown and we opted to evaluate several plausible paths  
638 within a broad thermal window ( $230 \pm 30$  Ma,  $250 \pm 150$  °C). The period from the middle Permian  
639 to the end of the Triassic is characterized by a slow and progressive drowning of the continental  
640 surface, indicating that no major tectonic deformation occurred (no major crustal thinning). This

641 implies that continental basement units were likely not exhumed. As a consequence, we maintain P  
642 constant at 3 kbar. Node G3 simulates final rift-related exhumation close to the surface ( $170 \pm 10$   
643 Ma,  $100 \text{ }^\circ\text{C}$ , 0.5 kbar). The selected timing is based on both stratigraphic and thermochronological  
644 data documenting the evidence for final rifting stage and basement exhumation (Mohn et al., 2010;  
645 Müntener and Hermann, 2001; Petri et al., 2019). Node G4 at 0 Ma has similar P-T conditions than  
646 node G3.

647         The 10,000 paths that were generated (Fig. 12a) converge around 260 Ma and around 240  
648 Ma, but are not converging for node G3 because of its large difference in age with our dataset  
649 combined to the low T-t gradient of the path. The initial temperature conditions are consistent with  
650 qualitative P-T estimates of 550-600  $^\circ\text{C}$  that can be deduced from the regional presence of  
651 andalusite of probable Early Permian age in the Grosina unit. Although the highest best fit is found  
652 for a variable age of node G2 (245-215 Ma), the calculated cooling rate is very constant at 5-10  
653  $^\circ\text{C}/\text{Myr}$  (Fig. 12a); the highly variable age of node G2 translates a low difference in cooling rates  
654 towards the Jurassic node G3.

655         The Permo-Triassic cooling event can also be reported in the Campo unit using  
656 chronometers with higher closure temperatures (Supplementary Figure S7). The difference between  
657 Sm–Nd and Rb–Sr ages (11 and 34 Myr of difference for samples SO5/11 and SO5/1, respectively)  
658 from the Campo unit led Tribuzio et al. (1999) and Braga et al. (2001) to conclude that rocks from  
659 the Sondalo gabbro cooled slowly at a rate of  $10 \text{ }^\circ\text{C}/\text{Myr}$  after its emplacement during the Permian.  
660 This is confirmed by the Rb–Sr ages on muscovite being 14 Myr older than Rb–Sr ages on biotite  
661 (sample PB2 of Del Moro and Notarpietro, 1987) and 35 Myr older than K–Ar ages on muscovite  
662 (sample 10-M of Hanson et al., 1966). The spread of Rb–Sr ages may also support a slow cooling  
663 (see Fig. 10b), but mixed ages caused by subsequent partial reopening cannot be excluded. The  
664 cooling history of the Grosina basement is only documented by  $^{40}\text{Ar}/^{39}\text{Ar}$  dating with a gap of 16  
665 Myr between muscovite and biotite from the sample GM264 (Figs 7a,8b). Although old age pairs

666 are pointing to fast cooling immediately after the emplacement of the Sondalo gabbroic intrusion,  
667 post-intrusion cooling rates for samples collected in the Grosina and the Campo units are below 10  
668 °C/Myr (sample 10-M of Hanson et al., 1966 and GM264). We can thus conclude that the Campo–  
669 Grosina units record the same cooling “event” in Permian to Triassic times, which has some  
670 implications for the subsequent thermal history (see discussion below).

671           The low cooling rates here reported (5-10 °C/Myr) are in the same order of cooling rates  
672 that are expected during the Permian event (~ 5 °C/Myr; Schuster and Stüwe, 2008). They are also  
673 similar to those reported from middle to lower crustal units from the Southern Alps (~ 1-5 °C/Myr  
674 in the Ivrea-Verbano Zone; Hurford, 1986; Siegesmund et al., 2008; Wolff et al., 2012), although  
675 there, the cooling may have been slowed down by Triassic magmatism as well as by tectonically-  
676 and magmatically activated re-heating during the Jurassic (see discussion below).

677           Cooling can be the result of several tectonic events. The first hypothesis is related to a  
678 crustal thinning and exhumation associated to the opening of the Meliata oceanic domain that  
679 occurred at around 240 Ma (Stampfli and Kozur, 2006). This is yet unlikely, as the initiation of our  
680 cooling (node G1) clearly pre-dates Meliata opening, and Triassic extensional deformation remains  
681 very limited in our study area. The second hypothesis is related to a post-orogenic thermal  
682 equilibration. The same type of rather low cooling rates has been retrieved by modelling the  
683 Mesozoic thermal subsidence of the Paris basin, that lies on the European plate outside the Alpine  
684 domain and that did not undergo Jurassic rifting (Prijs et al., 2000): at around 2-4 °C/Myr in the  
685 lower continental crust, and 1-3 °C/Myr in the middle crust (considering the first 60 Ma after  
686 thermal maximum only). We therefore suggest that our Ar-Ar cooling ages from the Grosina unit  
687 records this slow thermal equilibration of the lithosphere after the Variscan orogeny. This cooling  
688 implies a thickening of the lithosphere that will progressively subside without faulting at the surface  
689 (McKenzie, 1978). This subsidence permitted the deposition of thick continental Permo-Triassic  
690 sediments progressively superseded by marine sediments. Cooling rates below 10 °C/Myr seem

691 characteristic to post-orogenic deep crust recording conductive (isobaric) lithospheric cooling, as  
692 reported for Proterozoic orogens ( $< 10$  °C/Myr; Blackburn et al., 2011; Busch et al., 1996; Schmitz  
693 and Bowring, 2003; Scibiorski et al., 2016), and rarely documented but similar for Phanerozoic  
694 orogenic deep crust ( $\sim 1$ -20 °C/Myr; Bohlen and Mezger, 1989; Grove and Harrison, 1999; Hurford,  
695 1986; Siegesmund et al., 2008; Wolff et al., 2012). In addition, these values are significantly low  
696 compared to those of deep crust undergoing syn-orogenic exhumation (isothermally decompressed  
697 lower crust showing values in the range of 25-100 °C/Myr; e.g., Štípská et al., 2004). The rate of the  
698 conductive cooling of the lower crust appears noteworthy one order of magnitude smaller than what  
699 can be calculated for a cooling oceanic lithosphere using results of Stein and Stein (1992; averaging  
700 the first 60 Myr of cooling), even if tuning the crustal thermal conductivity to continental values  
701 ( $2.5 \text{ W.m}^{-1}.\text{K}^{-1}$ ).

702         The observed cooling rates suggest that isobaric/conductive cooling is more important than  
703 faster cooling associated to tectonic exhumation. This statement may however not be valid for  
704 shallower crustal levels located in the footwall of Permian extensional detachment faults (Pohl et  
705 al., 2018).

### 706         **6.3.2. Thermal evolution during Jurassic rifting**

707         From our  $^{40}\text{Ar}/^{39}\text{Ar}$  dataset, the Jurassic, rift-related cooling seems to be only recorded in  
708 the Campo unit. The new hornblende, muscovite and biotite ages presented in this study span a  
709 narrow temporal range within ca. 4 Myr between hornblende and biotite (sample CP031a) and  
710 between muscovite and biotite (sample CP028c). It points to a cooling of the Campo basement  
711 during the Jurassic that appears to be rapid, with estimates based on theoretical closure temperatures  
712 reaching up to 30-45 °C/Myr (Supplementary Figure S7). Such rapid cooling is also indicated by  
713 the near-constant age of consecutive steps for biotites, translated by lower MSWD values of the  
714 sub-plateau ages for the Jurassic cooling (Campo unit) compared to the Permian cooling (Grosina  
715 unit; Figs 7,9). We better constrained the temperature-time evolution of the Campo unit by testing

716 two scenarios built from regional constrains (Fig. 11). Rifting related to the Alpine Tethys opening  
717 initiates during Norian to Late Sinemurian (stretching phase of Mohn et al., 2010) with the  
718 formation of high-angle normal faults in the upper crust (Ortler unit; Eberli, 1988), pre-dated by the  
719 first Ar-Ar ages in the underlying lower crust and mantle, that is interpreted as the initiation of their  
720 exhumation and related deformation (Villa et al., 2000). This leads us to define a model initiation at  
721 250 Ma (node C1), before Jurassic rifting initiated. At that time, the temperature is inferred to be  
722 above muscovite expected closure temperature ( $500 \pm 100$  °C) at a depth constrained by pre-rift  
723 metamorphic conditions in the Campo unit (Petri et al., 2016). Deformation starts being localized in  
724 the future necking zone and distal domain at 190-175 Ma (Mohn et al., 2012; Ribes et al., 2020), the  
725 youngest pre-tectonic sedimentary deposit in the distal domain (before rift localization) being the  
726 Agnelli Formation in the Err-Bernina units (185 Ma). This constrains the initiation of the final  
727 exhumation and related cooling at node C3 ( $185 \pm 15$  Ma;  $450 \pm 150$  °C; 3 kbar). Deformation in  
728 the necking zone is expected to stop at ca. 170 Ma, and hyperextension takes over as the mantle is  
729 exhumed in the distal domain at ca. 165 Ma (Müntener and Hermann, 2001; Villa et al., 2000). We  
730 therefore propose that at 160 Ma, the deformation and cooling history of the necking zone is ended,  
731 constraining a node C4 at  $170 \pm 10$  Ma, 100 °C and 0.5 kbar, and a node C5 at 0 Ma and similar  
732 conditions than node C4. The evolution between node C1 and node C3 is however rather  
733 unconstrained, as the thermal history of the necking zone involves cooling potentially interrupted by  
734 heating phase(s). This is suspected for instance in the Ivrea-Verbano Zone (Southern Alps; Ewing et  
735 al., 2015; Smye and Stöckli, 2014), potentially helped by synchronous magmatism (e.g., Galli et al.,  
736 2019). The age and temperature increase of such pulse(s) are, however, unknown. A regional  
737 thermal peak should be reached before 175 Ma, which corresponds to the period during which the  
738 Briançonnais domain and the necking domains are uplifted and emerged (Decarlis et al., 2017;  
739 Mohn et al., 2012; Ribes et al., 2020). The oldest bound of this thermal pulse may correspond to the  
740 time when the lower crust and the mantle started to rise between 220-190 Ma (Fig. 11). We

741 therefore designed a node C2 at  $205 \pm 15$  Ma,  $300 \pm 100$  °C, and 3 kbar. To test if a re-heating is  
742 required, this node C2 is set inactive in a first scenario (Fig. 11 and Fig. 12b; simple cooling from  
743 node C1 to node C3, followed by exhumation-related cooling from node C3 to node C4), and is  
744 activated in a second scenario to force re-heating from node C2 to node C3, before final exhumation  
745 and related cooling from node C3 to node C4 (Fig. 11 and 12c). Each scenario was explored by  
746 10,000 individual paths (Fig. 12). The only interpretable amphibole age (hornblende CP031a)  
747 carries a large uncertainty, and has therefore been excluded from the rest of the discussion.

748 Both scenarios for the Campo unit clearly converge for node C3 at around 180-175 Ma for  
749 400-450°C when the final cooling is initiated and the distribution of path's probability points to an  
750 achievement of cooling from 175 to 160 Ma. This indicates that our dataset is clearly supporting a  
751 final cooling initiated between 180-175 Ma, with rather fast cooling rates of 25-100 °C/Myr (Fig.  
752 12bc). This age slightly precedes the cessation of activity of necking detachment faults and a  
753 drowning of the Tethyan necking zones reported from several studies (Southern Alps: Decarlis et  
754 al., 2017; Eastern Alps: Mohn et al., 2012; External Crystalline Massifs: Ribes et al., 2020) that  
755 occurred in Toarcian-Aalenian times.

756 The thermal evolution of the necking zone before node C3 remains more ambiguous. Our  
757 Monte Carlo simulations do not converge for both nodes C1 and C2 (when active, Fig. 12c). We  
758 know that the maximal temperature of node C3 should be likely limited, as petrological  
759 observations from the Campo unit do not report any evidence from syn-rift high temperature  
760 metamorphism, but the initial temperature (position of node C2) and the amount of heating cannot  
761 be directly constrained. However, the Campo unit also underwent the Permo-Triassic cooling of the  
762 lithosphere (nodes C1 to C2 in scenario 2, Fig. 12c), which is supported by other geochronometers  
763 than  $^{40}\text{Ar}/^{39}\text{Ar}$  (see discussion above) and that, considering the pre-rift depth of the unit, should lead  
764 to pre-rift temperature slightly above the temperatures of the Grosina unit (~200-300 °C). Such  
765 range of temperatures is not permitted by the scenario 1 as pre-rift temperatures are kept high



766 during the Permo-Triassic (nodes C1 to C3 in Fig. 12b). Hence, the scenario 2 with the presence of  
767 a node C2 is favoured. This suggests that a syn-rift heating of basement rocks in the necking zone  
768 has occurred, but the maximum temperature (node C3) remains limited to 400-450 °C and the  
769 amount of re-heating (nodes C2 to C3) to maximum 200 °C. The re-heating should be achieved by  
770 180 Ma, which is the maximal best-fit age of node C3: this 180 Ma limit is also reported from the  
771 Southern Alps necking zone by Beltrando et al. (2015). The potential effect of thermal blanketing  
772 expected by thick sedimentary deposits (e.g., Mauléon basin, Pyrenees; Lescoutre et al., 2019;  
773 Saspiturry et al., 2020) can here be excluded to cause this heating because of a strong sediment  
774 starvation. This heating may instead concomitantly result from the rise of the asthenosphere due to  
775 rifting, from the ascent of the lower crust in the adjacent future distal domain (Fig. 13e), and from  
776 the activity of deeper extraction faults (see Petri et al., 2019 for details). It may also result in a  
777 thermal support allowing to transiently exhume above sea-level pre-rift sediments as reported from  
778 the Southern Alps by Decarlis et al. (2017) and potentially basement rocks as suspected in the Mont  
779 Blanc massif by Ribes et al. (2020) during the Sinemurian-Pliensbachian (~190-185 Ma). This  
780 indicates that the necking zone uplift or even emersion is caused by a flexural response of the  
781 lithosphere to rifting (Chenin et al., 2020), but may also be thermally supported.

782         Noteworthy, the two older ages in the Campo unit ( $218 \pm 2$  Ma muscovite age and  $214 \pm 3$   
783 Ma biotite age) can be explained in this scenario either (1) by the preservation of cooling ages from  
784 the Permo-Triassic cooling phase (path from nodes C1 to C3), (2) by an increased argon retentivity  
785 in biotite due to its high Mg-content but no difference in muscovite composition has been detected  
786 for the sample with older muscovite age, (3) by a crystallization that may have occurred close to the  
787 argon closure temperature, or (4) by cryptic and homogeneous extraneous argon incorporation.

#### 788 **6.4. Age of the Eita Shear Zone (ESZ)**

789         In contrast to the Permo-Triassic cooling history of the Grosina unit, the final cooling of  
790 the Campo unit was achieved during Jurassic necking. The associated fast cooling rates (25-100

791 °C/Myr) are characteristic of mid-crustal units affected by rift-related exhumation of rocks to  
792 shallow and cooler crustal levels (e.g., Beltrando et al., 2013; Bertotti et al., 1999; Decarlis et al.,  
793 2017). The presence of brittle deformation structures along the Grosina detachment indicates that  
794 the Grosina unit was in the top few kilometres of the crust at the time necking occurred. The pre-rift  
795 position and syn-rift evolution of the Campo and the Grosina units can also be discussed following  
796 their thermal evolution. The Grosina unit was very certainly below  $\sim 330$  °C before rifting started  
797 (below the  $^{40}\text{Ar}/^{39}\text{Ar}$  closure temperature of biotite). The contrasted thermal evolution of the Grosina  
798 unit and the Campo unit during the Jurassic indicates that, in a pre-rift section, both units were not  
799 as close as they are nowadays. The superposition likely occurred through the activity of the Jurassic  
800 ESZ that was interpreted as an extraction fault by Petri et al. (2019). In addition, the ESZ was active  
801 in greenschist facies conditions while being sub-horizontal and therefore it cannot correspond to a  
802 structure that exhumed efficiently rocks. This suggests that the exhumation of the Campo and  
803 Grosina units may have not occurred along the ESZ only, but was accommodated by other structures  
804 such as the Grosina detachment (Fig. 13ef), interpreted by Mohn et al. (2012) as a necking  
805 detachment. In this scenario, the ESZ is thus not directly responsible for the cooling of the Campo  
806 unit, leaving the effective age of the shear zone rather unconstrained. We can however postulate that  
807 it was active during the stretching and early necking phase (Fig. 13d) from late Triassic to early  
808 Jurassic, synchronous to the initiation of the rise of mantle and lower crust documented in the  
809 Lower Platta and Malenco units (Villa et al., 2000).

810           Observations in the distal domain (Err and Bernina units) indicate that deformation related  
811 to crustal necking was active between 190-175 Ma (Ribes et al., 2020), before being superseded by  
812 hyperextension in the distal domain from 175 Ma to 165 Ma. The cooling of the Campo unit is  
813 likely initiated around 180-175 Ma (position of node C3), which lies on the upper age limit of  
814 necking zones activity, and was achieved between 175-165 Ma. This end of cooling is coeval to the  
815 subsidence and the drowning of the necking zones documented in the post-Toarcian sediments

816 (Ribes et al., 2020). This implies that a complete heating-cooling cycle is recorded in the necking  
817 zone during its formation and that cooling of the necking zone may substantially postdate the main  
818 phase of necking but predates mantle exhumation that occurred at 165 Ma.

## 819 7. CONCLUSION

820 This contribution discussed the structural and thermal evolution of two basement units (i.e.  
821 Campo-Grosina units) in the Austroalpine nappes preserving the relics of the former necking zone  
822 of the Adriatic rifted margin. We report new  $^{40}\text{Ar}/^{39}\text{Ar}$  ages from these two units that record a  
823 contrasted cooling history:

824 -  $^{40}\text{Ar}/^{39}\text{Ar}$  ages from the Grosina unit lie at  $261 \pm 2$  Ma for muscovite and at  $246 \pm 13$  Ma  
825 and  $245 \pm 2$  Ma for biotite.

826 -  $^{40}\text{Ar}/^{39}\text{Ar}$  ages from the Campo unit are clearly younger, between 182 to 178 Ma for  
827 muscovite and between 174 and 171 Ma for biotite, with occasional older ages at  $218 \pm 2$  Ma for  
828 muscovite and at  $214 \pm 3$  Ma for biotite.

829 - In the Campo unit, 3 out of 4 dated amphiboles led to geologically meaningless dates, and  
830 the only interpretable  $^{40}\text{Ar}/^{39}\text{Ar}$  interpretable age carries a large uncertainty with an inverse isochron  
831 age between  $210 \pm 4$  Ma and  $177 \pm 13$  Ma. These amphiboles suffered from both a low potassium  
832 content, and strong incorporation of extraneous argon caused by the intrusion of the Permian  
833 Sondalo gabbroic complex in the Campo unit.

834 The cooling history of the two units was modelled using a Monte Carlo approach based on  
835 DiffArgP, which allowed to constrain:

836 - a post-Variscan Permo-Triassic slow cooling history of 5-10 °C/Myr in the Grosina unit,  
837 that is characteristic of the conductive cooling of the continental lithosphere after an orogenic  
838 phase.

839 - a Jurassic rapid cooling of 25-100 °C/Myr in the Campo unit that is characteristic for  
840 tectonically-exhumed basement rocks. This exhumation was likely initiated at around 185 Ma, the

841 associated cooling being delayed by a few millions of years. A re-heating of the necking zone is  
842 suspected, but its characteristics remain to be better constrained.

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#### 850 **SUPPLEMENTARY MATERIAL**

851 **Supplementary Table S1:** Representative mineral composition of dated samples

852 **Supplementary Table S2:**  $^{40}\text{Ar}/^{39}\text{Ar}$  on muscovite geochronology data

853 **Supplementary Table S3:**  $^{40}\text{Ar}/^{39}\text{Ar}$  on biotite geochronology data

854 **Supplementary Table S4:**  $^{40}\text{Ar}/^{39}\text{Ar}$  on hornblende geochronology data

855 **Supplementary Table S5:** Compilation of pre-Alpine ages from the Campo and Grosina  
856 units

857 **Supplementary Figure S6:** Microphotographs of samples used for geochronology and  
858 mineral composition analyses

859 **Supplementary Figure S7:** Compilation of published and new ages for the Campo and  
860 Grosina units

861 **Supplementary Text S8:** DiffArgP modelling strategy

862 **FIGURES**

Sample name	Coordinates	Rock type	Assemblage	Mineral	Mineral chemistry			Core			Rim			Total			40Ar/39Ar dates	
					Amp Altot Bt Xmg Ms Si	Amp NaM4 Ms Pg	Amp Ti Bt Ti Ms Ti	T (°C)	P (kbar)	n	T (°C)	P (kbar)	n	T (°C)	P (kbar)	n	Ma	± 2σ
<b>Grosina unit</b>																		
GM119	46.39640°N 10.23004°E	Granodiorite	Bt-Pl-Kfs-Qtz-Ilm	Biotite	0.37-0.43	0.08-0.21		681 ± 16	/	10	617 ± 50	/	10	649 ± 49	/	20	246 ± 13 Ma	
GM264	46.34994°N 10.17277°E	Grt-paragneiss	Bt-Ms-Grt-Pl-Kfs-Qtz-Ilm	Biotite Muscovite	0.5	< 0.10		569 ± 6	/	10	572 ± 9	/	10	570 ± 8	/	20	245 ± 2 Ma	
					> 3.1	0.09-0.15	< 0.03	578 ± 20	/	10	561 ± 33	/	10	569 ± 28	/	20	261 ± 2 Ma	
<b>Campo unit</b>																		
CP003b	46.35977°N 10.33070°E	Migmatitic micashist	Ms-Bt-Grt-Sil-Pl-Kfs-Qtz-Ilm	Muscovite	3.02	0.13	0.04-0.07	693 ± 37	/	10	689 ± 30	/	10	691 ± 33	/	20	218 ± 2 Ma	
CP010g	46.36312°N 10.24878°E	Amphibolite	Hbl-Pl-Ilm	Amphibole	1.99	0.14	0.07	/	/	/	/	/	/	/	/	/	Ex <sup>40</sup> Ar	
CP028c	46.33248°N 10.31080°E	Bt-Ms micashist	Bt-Ms-Pl-Qtz-Ilm	Biotite Muscovite	0.46	0.2		696 ± 16	/	10	699 ± 17	/	10	698 ± 16	/	20	174 ± 2 Ma	
					3.00	0.7	0.05	696 ± 5	/	10	694 ± 11	/	10	695 ± 8	/	20	178 ± 5 Ma	
CP031a	46.33479°N 10.31815°E	Grt-granodiorite	Bt-Hbl-Grt-Pl-Qtz-Ilm	Amphibole Biotite	2.29-2.67	0.16-0.19	0.05-0.09	/	8 ± 1	10	/	10 ± 1	10	/	9 ± 1	20	210-177 ± 4-13 Ma	
					0.44	0.07-0.14		585 ± 36	/	10	576 ± 45	/	10	581 ± 40	/	20	173 ± 2 Ma	
CP032c	46.35123°N 10.35480°E	Amp-gabbro	Hbl-Cpx-Opx-Pl-Mag	Amphibole	2.29	0.25	0.34	950 ± 20	6 ± 2	10	940 ± 30	6 ± 3	9	940 ± 30	6 ± 2	19	Ex <sup>40</sup> Ar	
CP034c	46.34941°N 10.36257°E	Grt-granodiorite	Bt-Grt-Pl-Ilm	Biotite	0.60-0.62	0.07-0.08		563 ± 10	/	10	563 ± 9	/	10	563 ± 10	/	20	214 ± 3 Ma	
CM234	46.33479°N 10.31815°E	Amphibolite	Amp-Pl-Ilm	Amphibole	2.3	0.1	0.08	/	/	/	/	/	/	/	/	/	Ex <sup>40</sup> Ar	
CM260	46.35939°N 10.25048°E	Grt-micashist	Bt-Ms-Grt-Pl-Kfs-Qtz-Ilm	Biotite Muscovite	0.4	0.12		611 ± 17	/	10	611 ± 11	/	10	611 ± 14	/	20	171 ± 1 Ma	
					3.04	0.14-0.16	0.03	572 ± 32	/	10	562 ± 44	/	10	567 ± 38	/	20	182 ± 2 Ma	

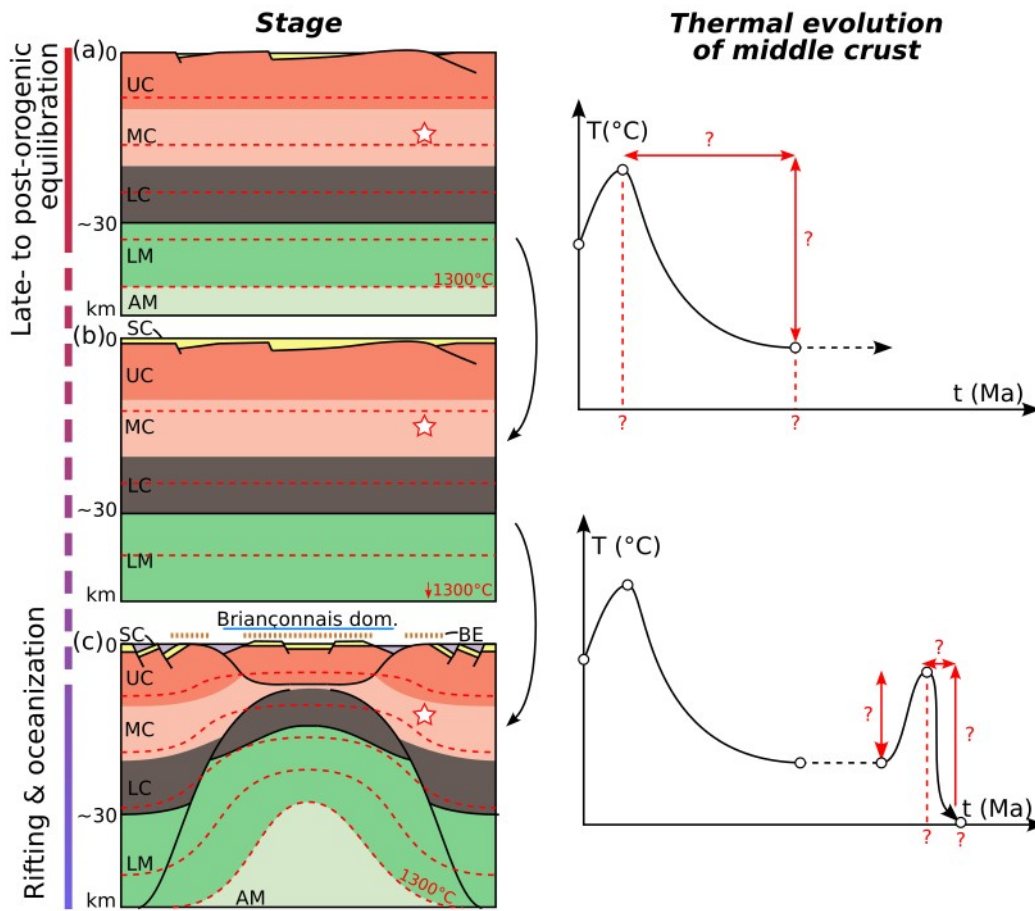
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**Table 1:** Sample list, coordinates, rock type, mineral assemblage and composition with average values of thermobarometric estimates, and

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<sup>40</sup>Ar/<sup>39</sup>Ar dates. See text for details and Supplementary Material S1 for detailed mineral composition and thermobarometric estimates.



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**Fig. 1:** Schematic thermal evolution of the continental lithosphere in the middle crust of a

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future necking zone from (a) a post-orogenic stage characterized by active extension (high angle

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normal faults and detachment faults), (b) a conductive lithospheric cooling, and (c) subsequent

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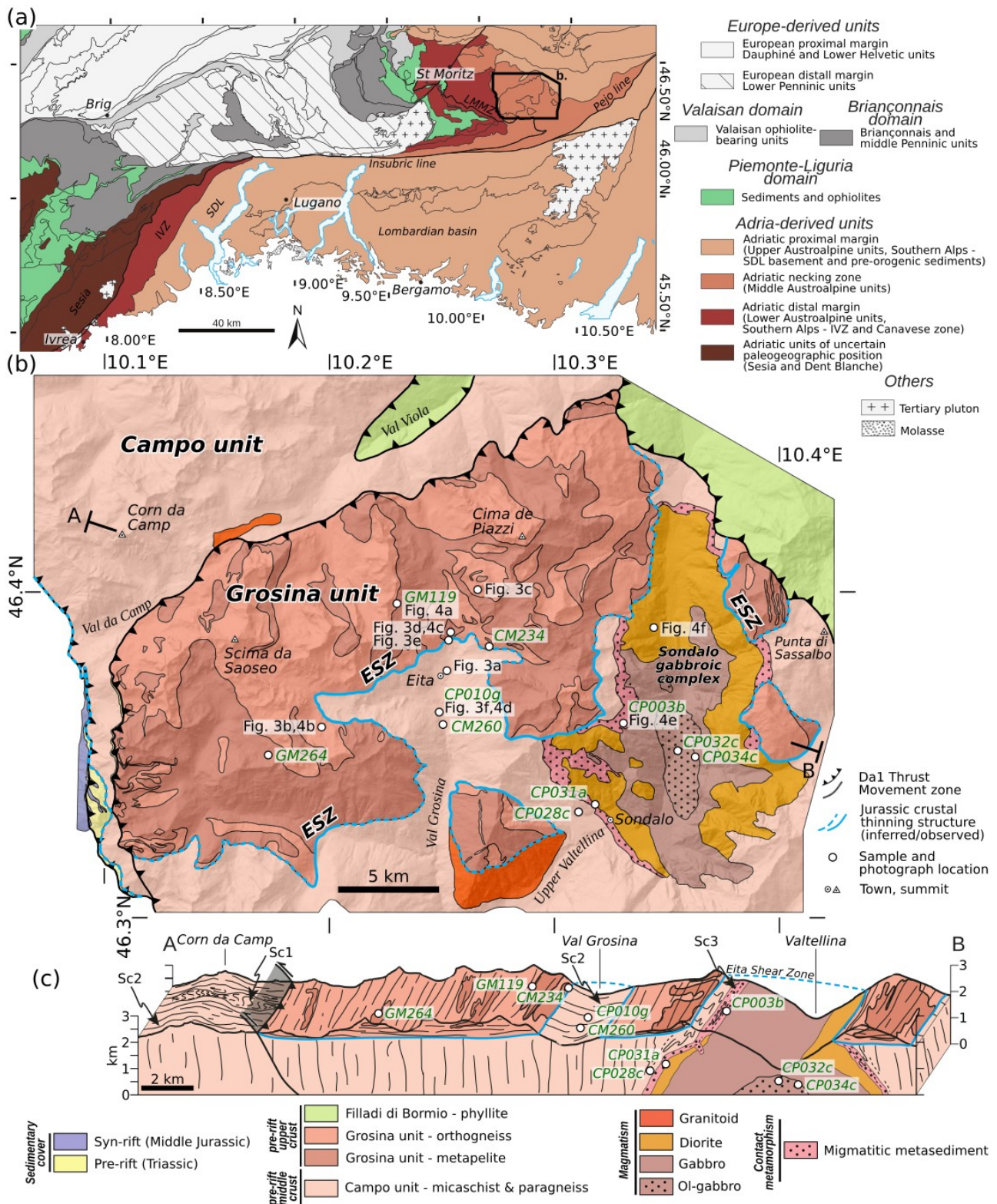
rifting and crustal necking. Sedimentary basins and thermal evolution not to scale. BE: basement

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exposure/erosion; SC: sedimentary cover; UC: upper crust; MC: middle crust; LC: lower crust; LM:

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lithospheric mantle; AM: asthenospheric mantle.



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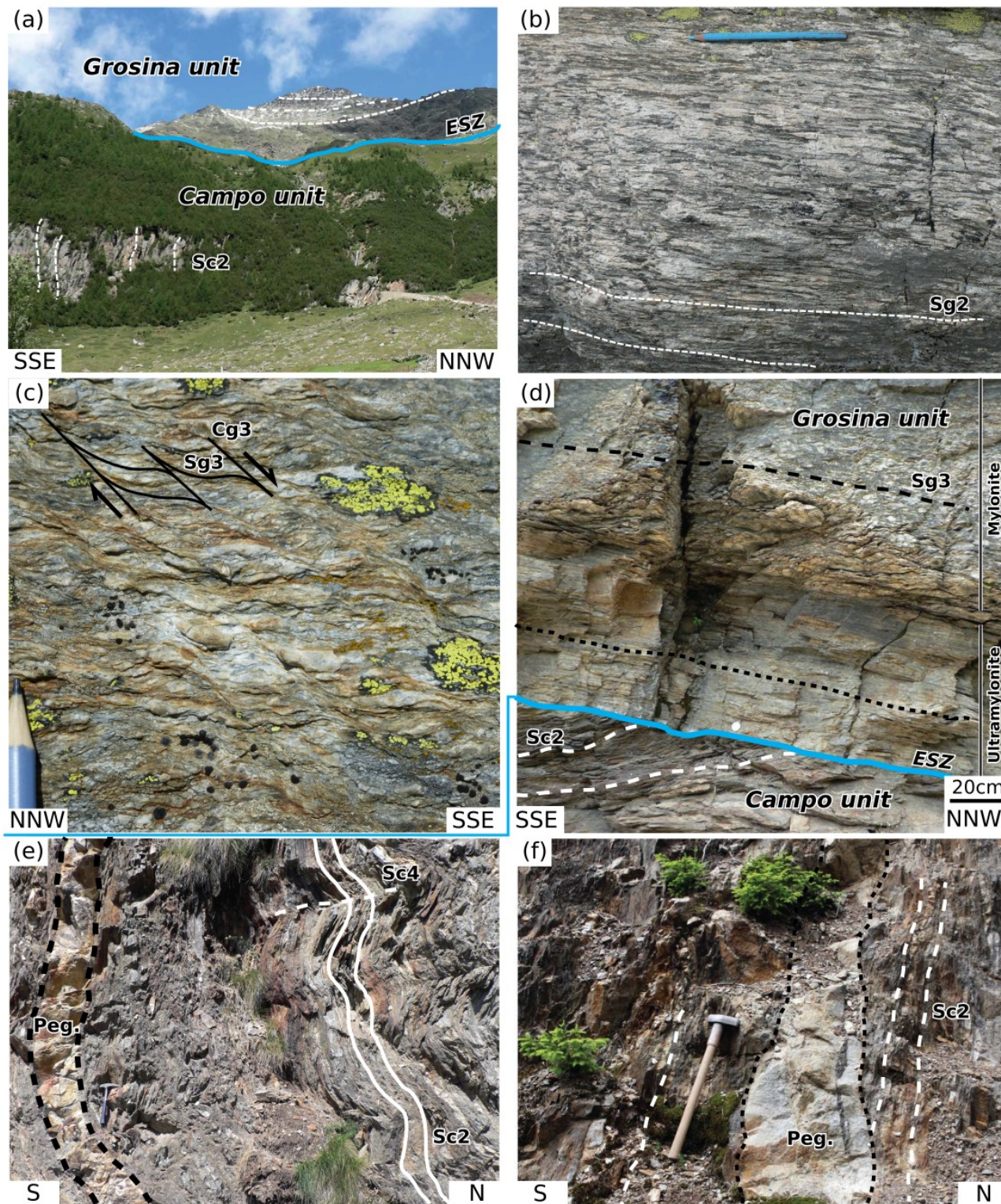
**Fig. 2:** (a) Tectonic map of the central part of the Alps reporting the paleogeographic origin

875 of the different units, modified from Petri et al. (2019). IVZ: Ivrea-Verbano Zone; SDL: Serie dei

876 Laghi; LMMZ: Lunghin-Mortirolo Movement Zone. (b) Geological map of the study area modified

877 from Mohn et al. (2011), the 1:10,000 and 1:25,000 geological maps of Italy and personal

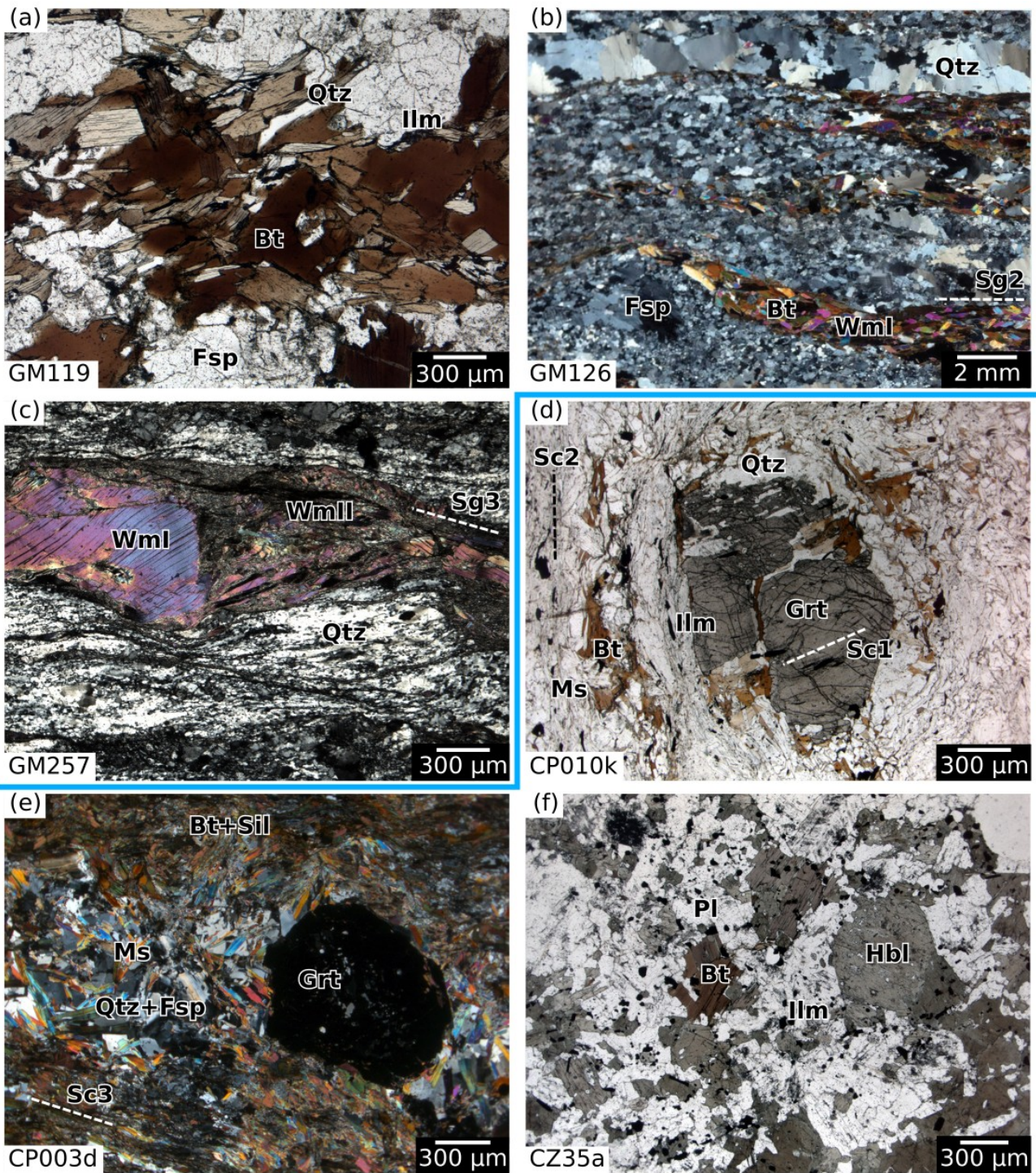
878 observations, and (c) schematic cross-section. Location of samples used for geochronology is  
 879 reported in green on the map and located or projected on the cross-section; location of photographs  
 880 is reported in black.



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882           **Fig. 3:** Field photographs. (a) Panoramic view of the Campo/Grosina units close to the Eita  
883 village. Note the vertical Sc2 foliation of the Campo unit and the horizontal foliation of the Grosina  
884 unit. Grosina unit: (b) Mylonitic orthogneiss showing the sub-horizontal amphibolite-facies Sg2  
885 foliation; (c) greenschist facies S-C structures associated to the Sg3 foliation. (d) Contact between  
886 the Campo metasediments with tilted Sc2 foliation and the Grosina orthogneiss with Sg3  
887 greenschist facies fabric along the ESZ. Campo unit: (e) Micaschist and pegmatitic dykes  
888 presenting Fc4 folds; (f) pegmatitic dyke cross-cutting the Sc2 planar fabric of metasediments.



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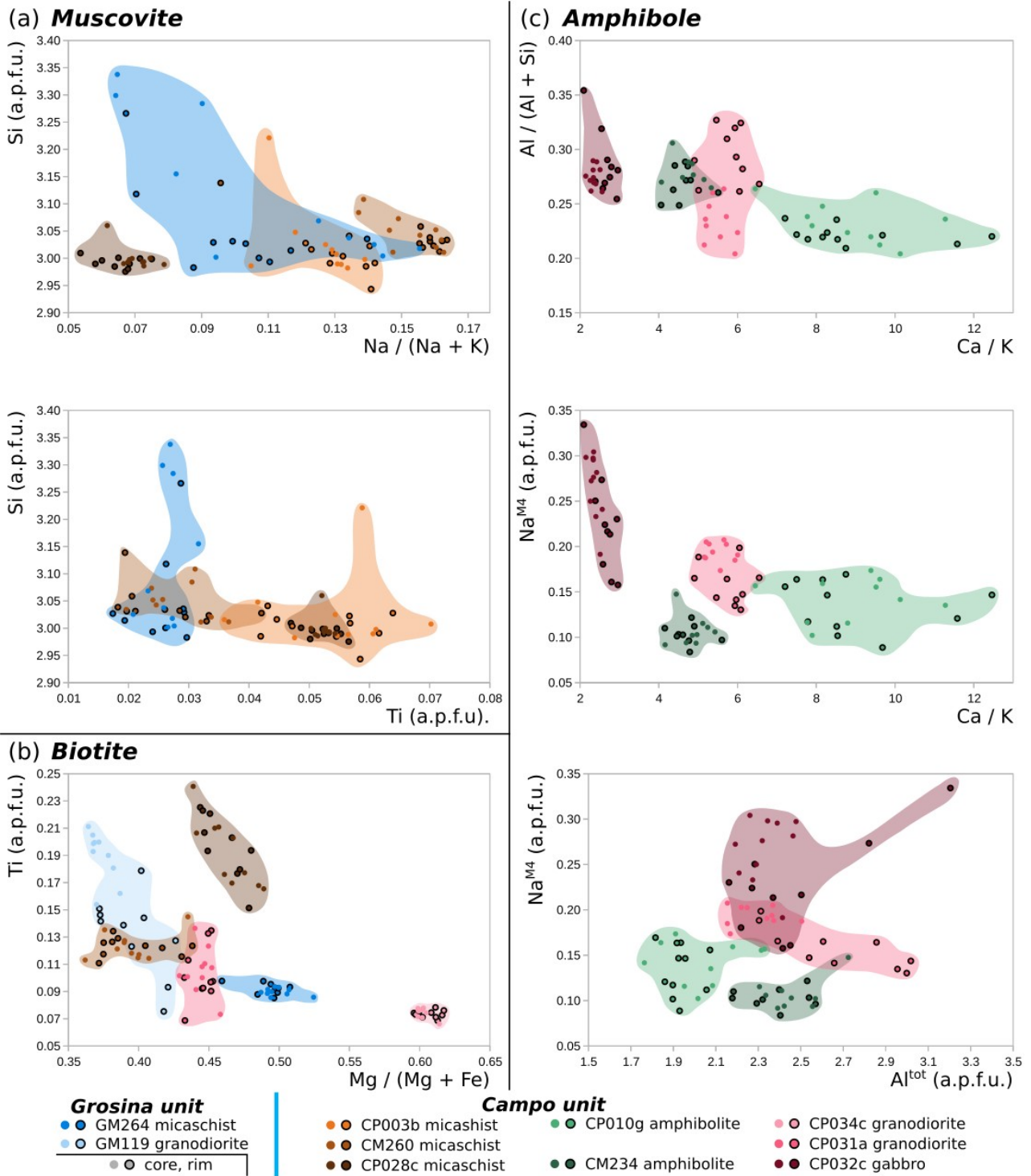
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**Fig. 4:** Photomicrographs illustrating various lithologies for (a-c) the Grosina unit and (d-f) the Campo unit. (a) Undeformed granodiorite with preserved magmatic biotites. (b) Amphibolite-facies mylonitic orthogneiss with metamorphic biotite and white mica I. Note the quartz recrystallization through grain-boundary migration. (c) Greenschist facies recrystallization of porphyroclastic white mica I into white mica II. (d) Garnet-bearing micaschist with metamorphic

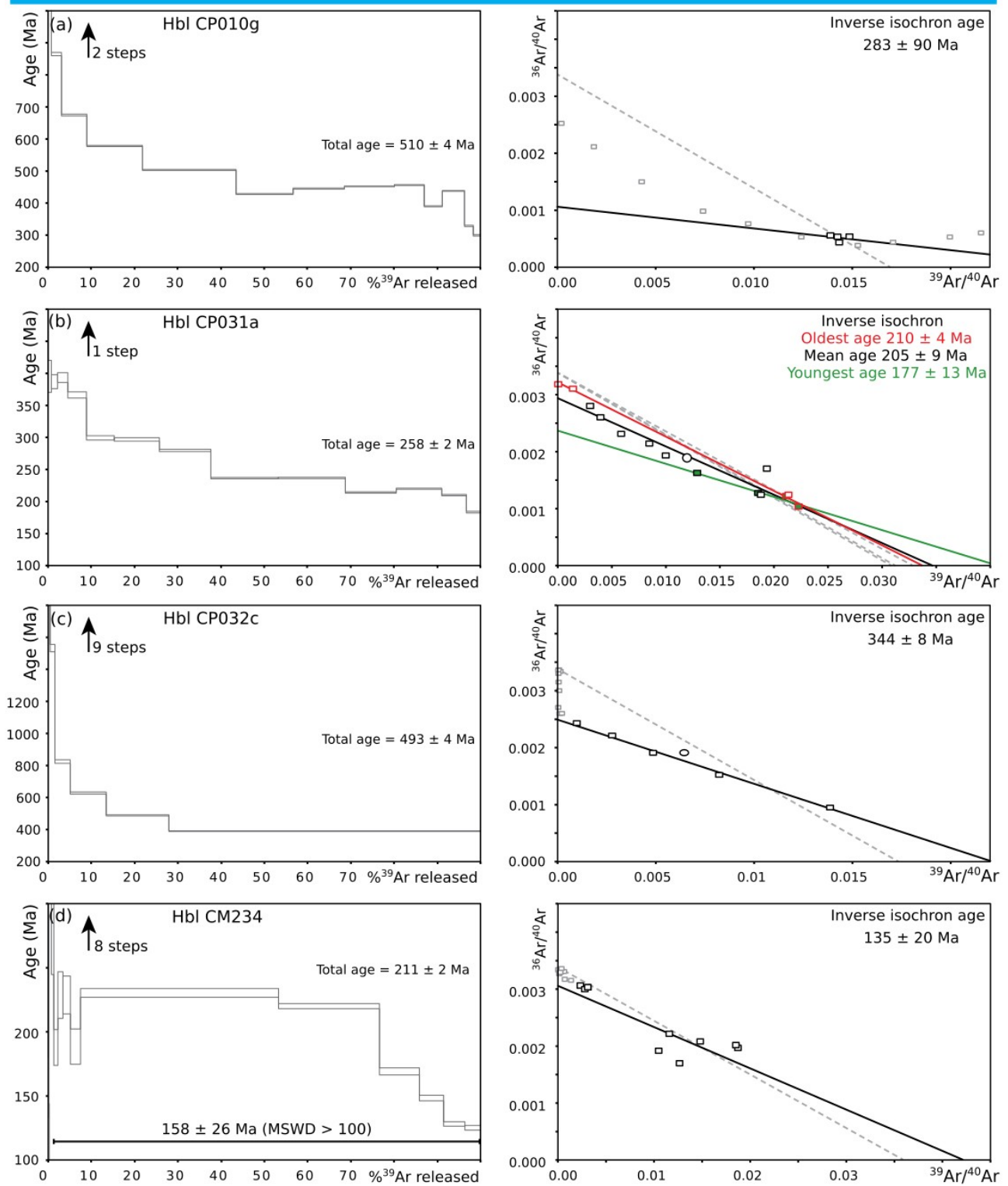
895 biotite and muscovite aligned along the Sc2 foliation. (e) Garnet–sillimanite–biotite migmatite in  
 896 the contact aureole of the Sondalo gabbro, with post-peak crystallization of muscovite in quartz–  
 897 feldspar–garnet leucocratic layer. (f) Magmatic hornblende and biotite from a diorite from the  
 898 Sondalo pluton.



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**Fig. 5:** Mineral composition plots for (a) muscovite, (b) biotite and (c) amphibole.



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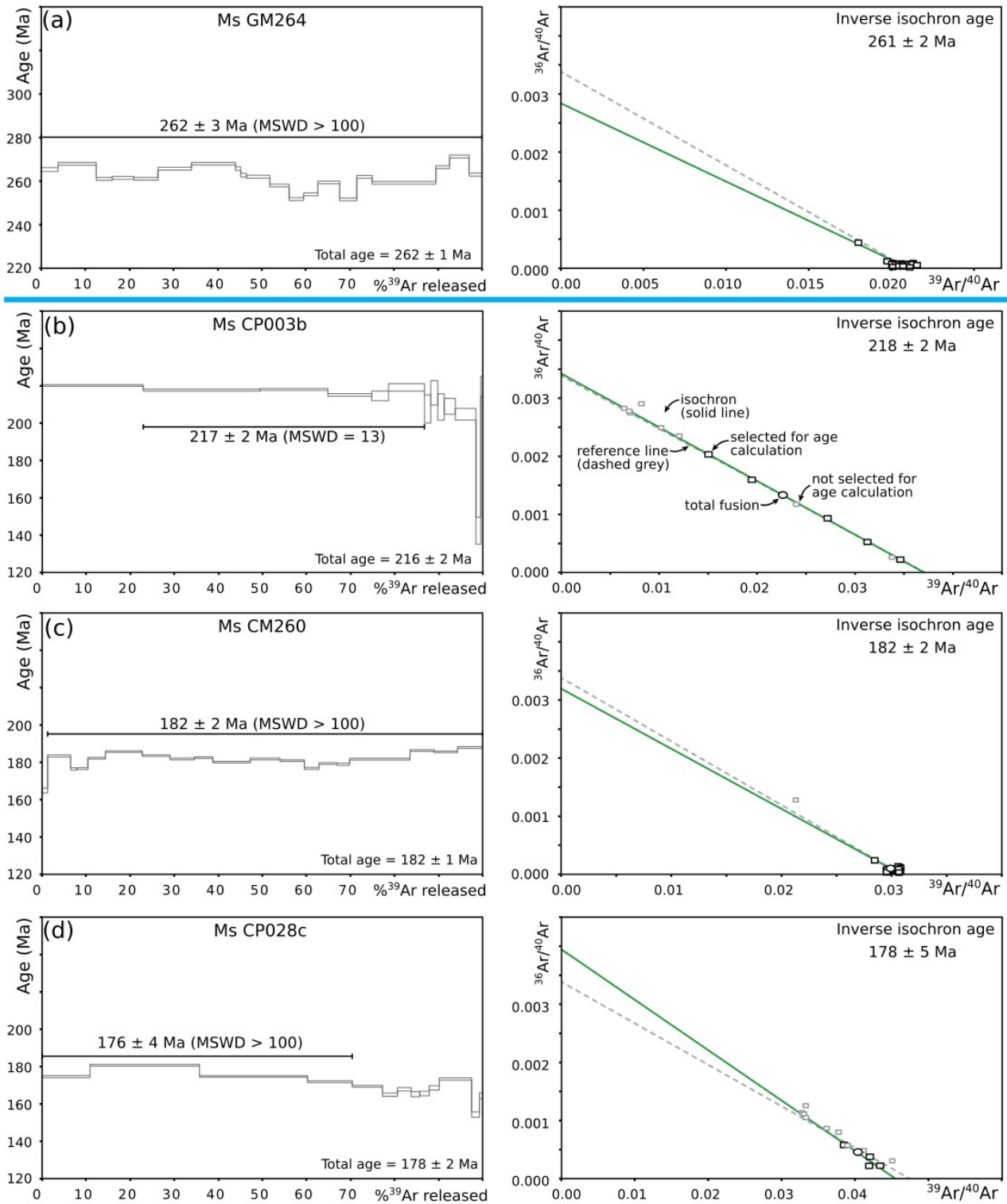
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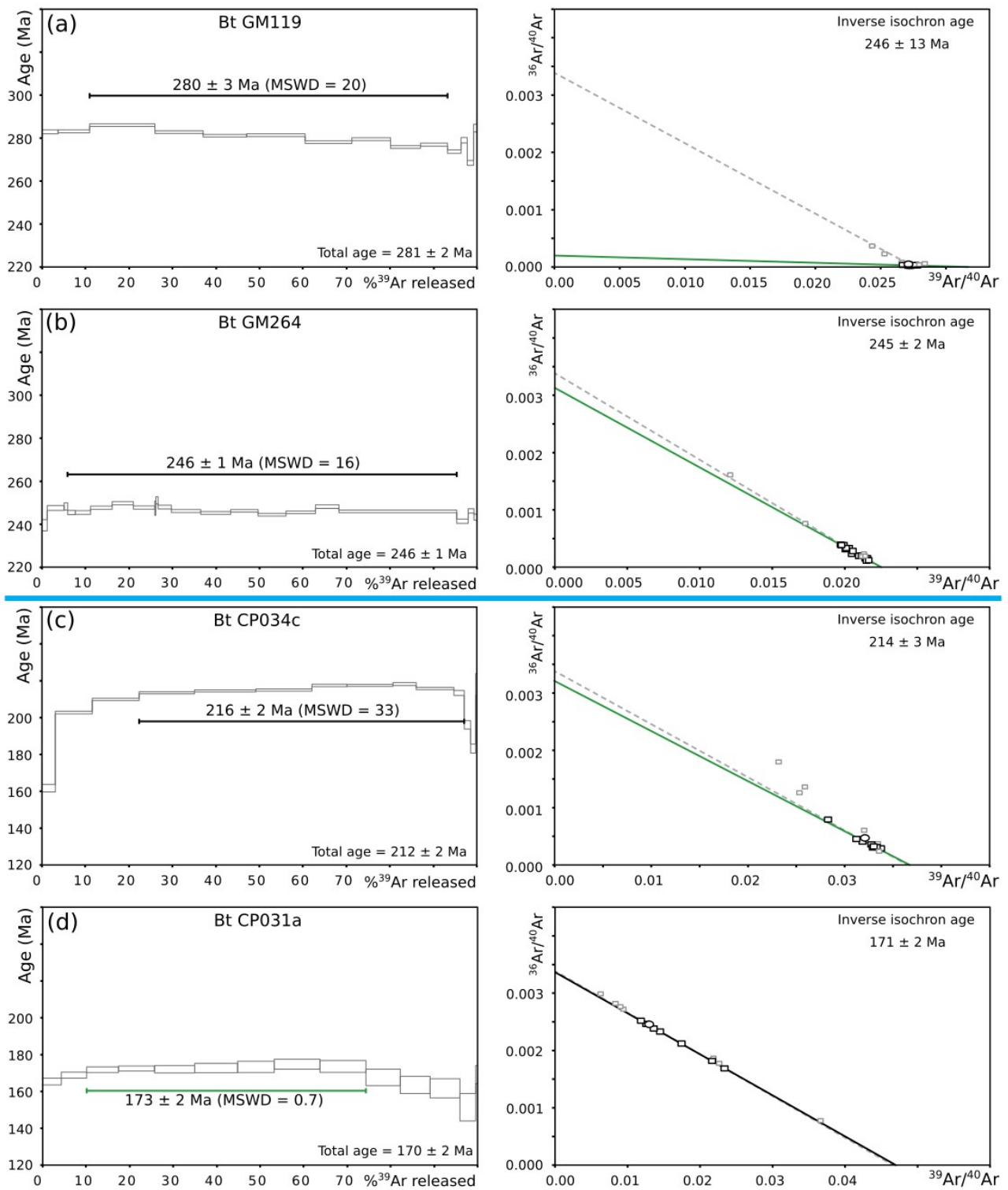
**Fig. 6:**  $^{40}\text{Ar}$ - $^{39}\text{Ar}$  step-release spectra and inverse isochron plots for hornblende from the Campo unit. Steps used to determine the “plateau” ages (if possible) are indicated by the horizontal black lines. Ages in bold and blue indicate the most reliable age between “plateau” age, total fusion age and inverse isochron age. Sample locations are shown in Fig. 2. Analysed mineral, sample and

906 experiment (in brackets) name are indicated (top side) along with the mean square weighted  
 907 deviation (MSWD) of the “plateau” age and the total fusion age (bottom right-hand side).



908  
 909 **Fig. 7:**  $^{40}\text{Ar}$ – $^{39}\text{Ar}$  step-release spectra and inverse isochron plots for muscovite from (a) the  
 910 Grosina unit and (b-d) the Campo unit. Steps used to determine the “plateau” ages (if possible) are

911 indicated by the horizontal black lines. Ages in bold and blue indicate the most reliable age between  
912 “plateau” age, total fusion age and inverse isochron age. Sample locations are shown in Fig. 2.  
913 Analysed mineral, sample and experiment (in brackets) name are indicated (top side) along with the  
914 mean square weighted deviation (MSWD) of the “plateau” age and the total fusion age (bottom  
915 right-hand side).



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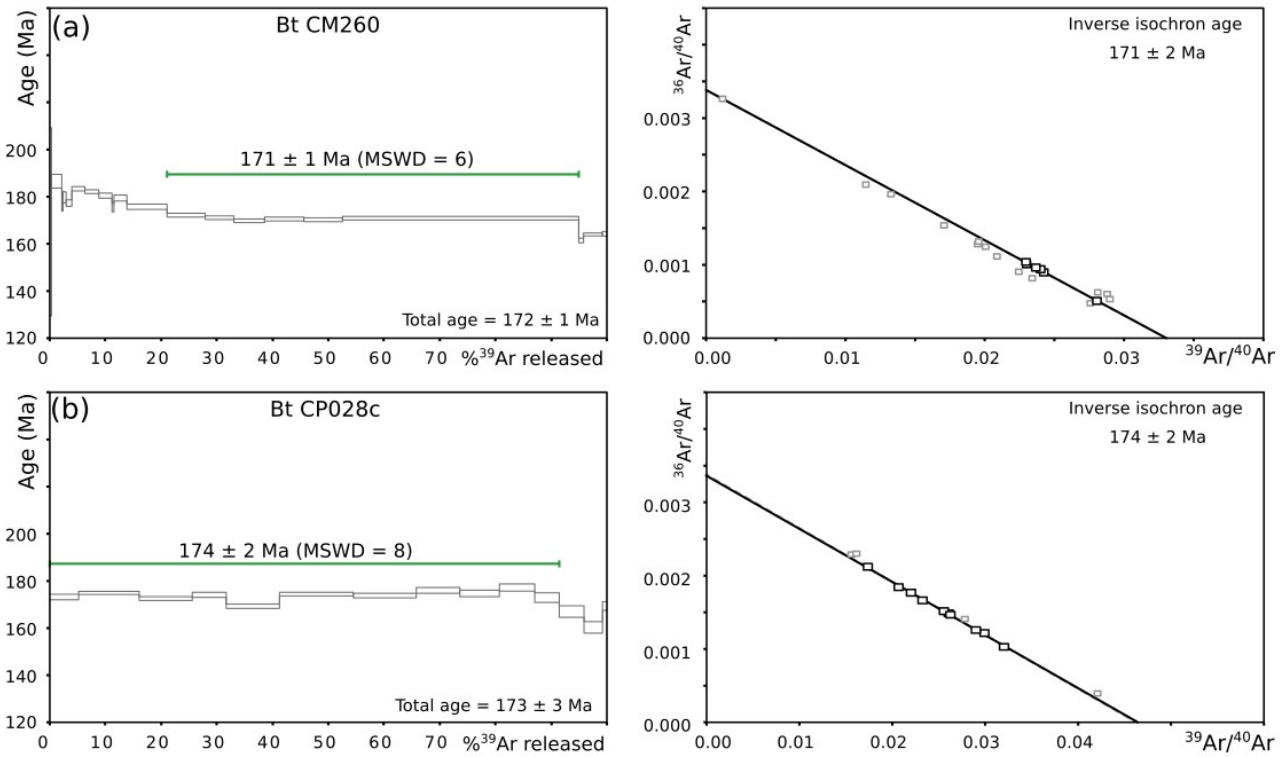
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**Fig. 8:**  $^{40}\text{Ar}$ - $^{39}\text{Ar}$  step-release spectra and inverse isochron plots for biotite from (a,b) the Grosina unit and (c,d) the Campo unit. Steps used to determine the “plateau” ages (if possible) are indicated by the horizontal black lines. Ages in bold and blue indicate the most reliable age between “plateau” age, total fusion age and inverse isochron age. Sample locations are shown in Fig. 2.

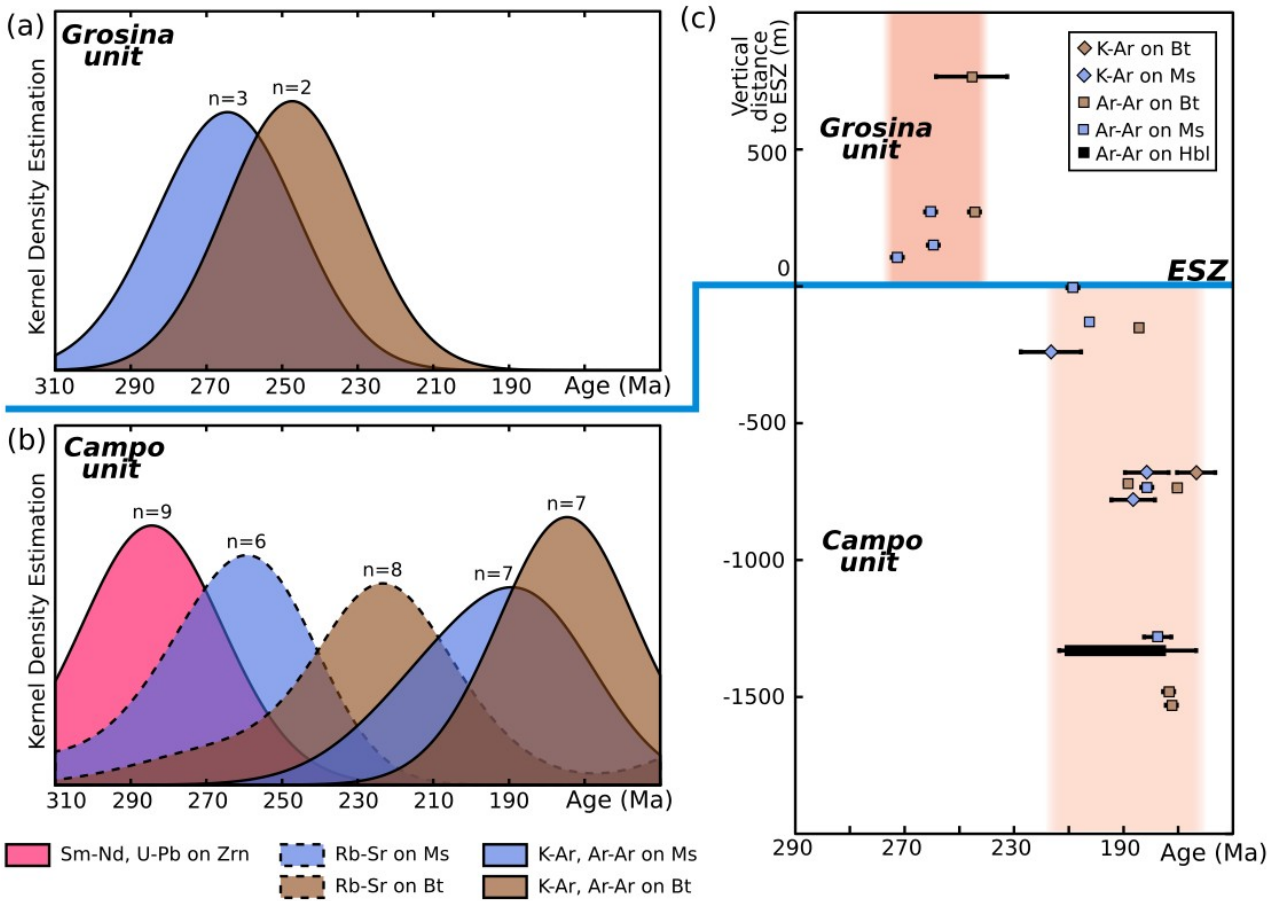
921 Analysed mineral, sample and experiment (in brackets) name are indicated (top side) along with the  
 922 mean square weighted deviation (MSWD) of the “plateau” age and the total fusion age (bottom  
 923 right-hand side).



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925 **Fig. 9:** <sup>40</sup>Ar–<sup>39</sup>Ar step-release spectra and inverse isochron plots for biotite from the Campo  
 926 unit. Steps used to determine the “plateau” ages (if possible) are indicated by the horizontal black  
 927 lines. Ages in bold and blue indicate the most reliable age between “plateau” age, total fusion age  
 928 and inverse isochron age. Sample locations are shown in Fig. 2. Analysed mineral, sample and  
 929 experiment (in brackets) name are indicated (top side) along with the mean square weighted  
 930 deviation (MSWD) of the “plateau” age and the total fusion age (bottom right-hand side).





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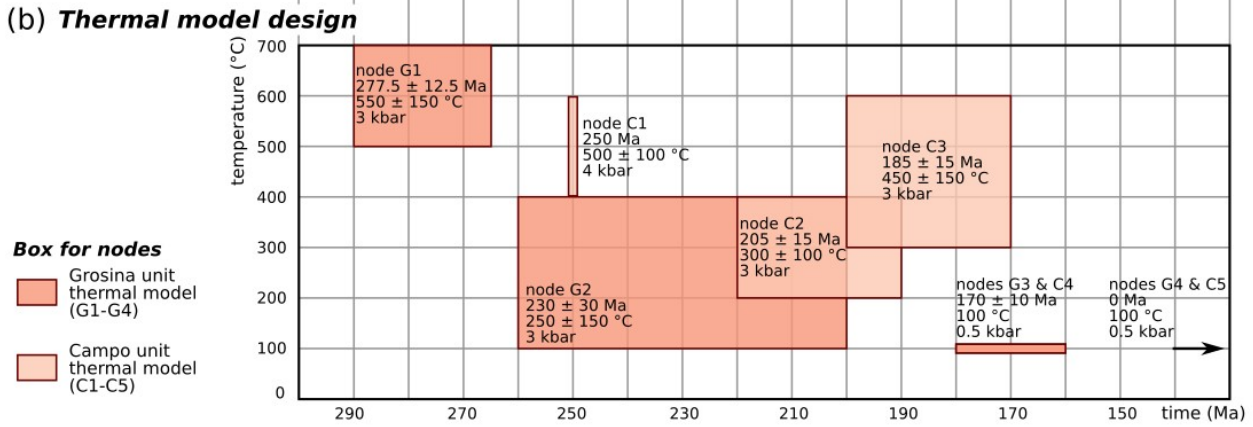
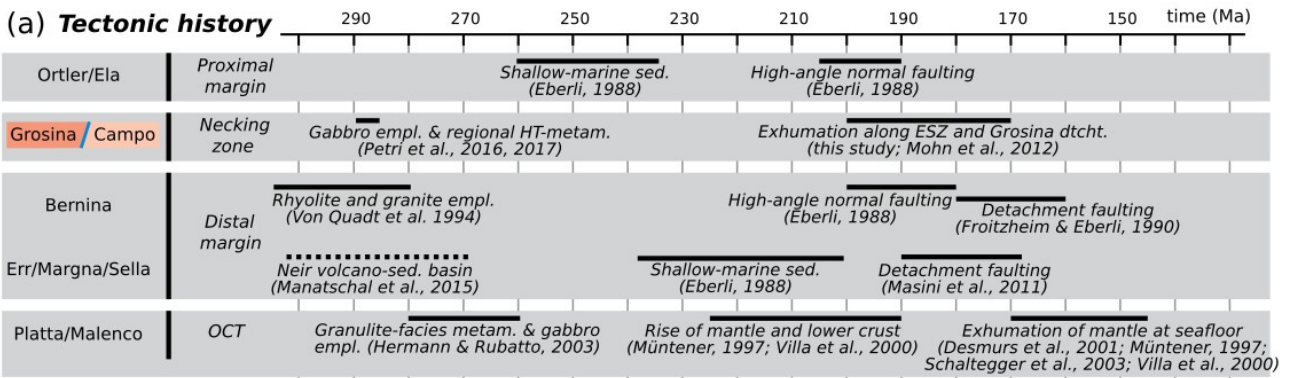
**Fig. 10:** Kernel Density Estimates (KDE) histogram of (a) the Grosina unit, (b) the Campo

unit and (c) age vs. vertical distance to the ESZ (one data point per sample). Plots were made using

data from this study and data from the literature. See text for details; complete dataset provided as

Supplementary Material S5. Errors are represented at a  $2\sigma$  level of confidence; KDE plots were

produced using IsoplotR (Vermeesch, 2018).



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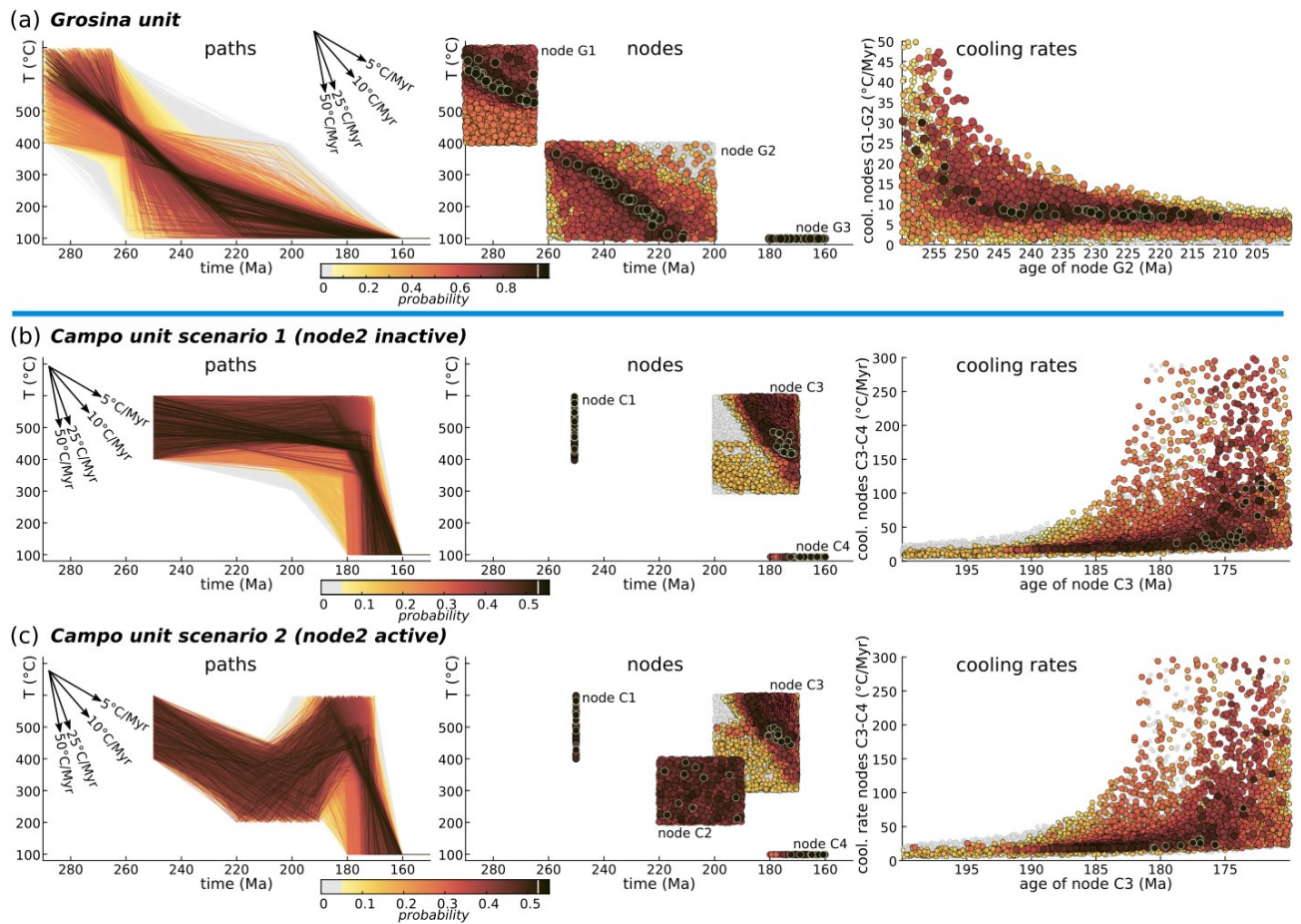
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**Fig. 11:** (a) Compilation of tectonic, magmatic and metamorphic phases in the Austroalpine nappes (Eastern Alps) used to construct (b) the thermal models that served as input for DiffArgP-based Monte Carlo simulations. Thermal models for the Grosina unit are defined by nodes G1 to G4; thermal models for the Campo unit are defined by nodes C1 to C5.



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**Fig. 12:** Paths, nodes and cooling rates of key nodes of DiffArgP-based Monte Carlo

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simulations for the thermal scenario of (a) the Grosina unit, (b) the Campo unit with node C2

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inactive involving a 2-phase cooling (slow pre-rift and early-rift, fast syn-rift) or (c) the Campo unit

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with node C2 active, forcing the pre-rift cooling to be interrupted by an Early Jurassic re-heating

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(between nodes C2 and C3). Each scenario was explored by 10,000 random paths which nodes

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characteristics (position and associated degree of freedom) are specified in Fig. 11. Paths with  $p <$

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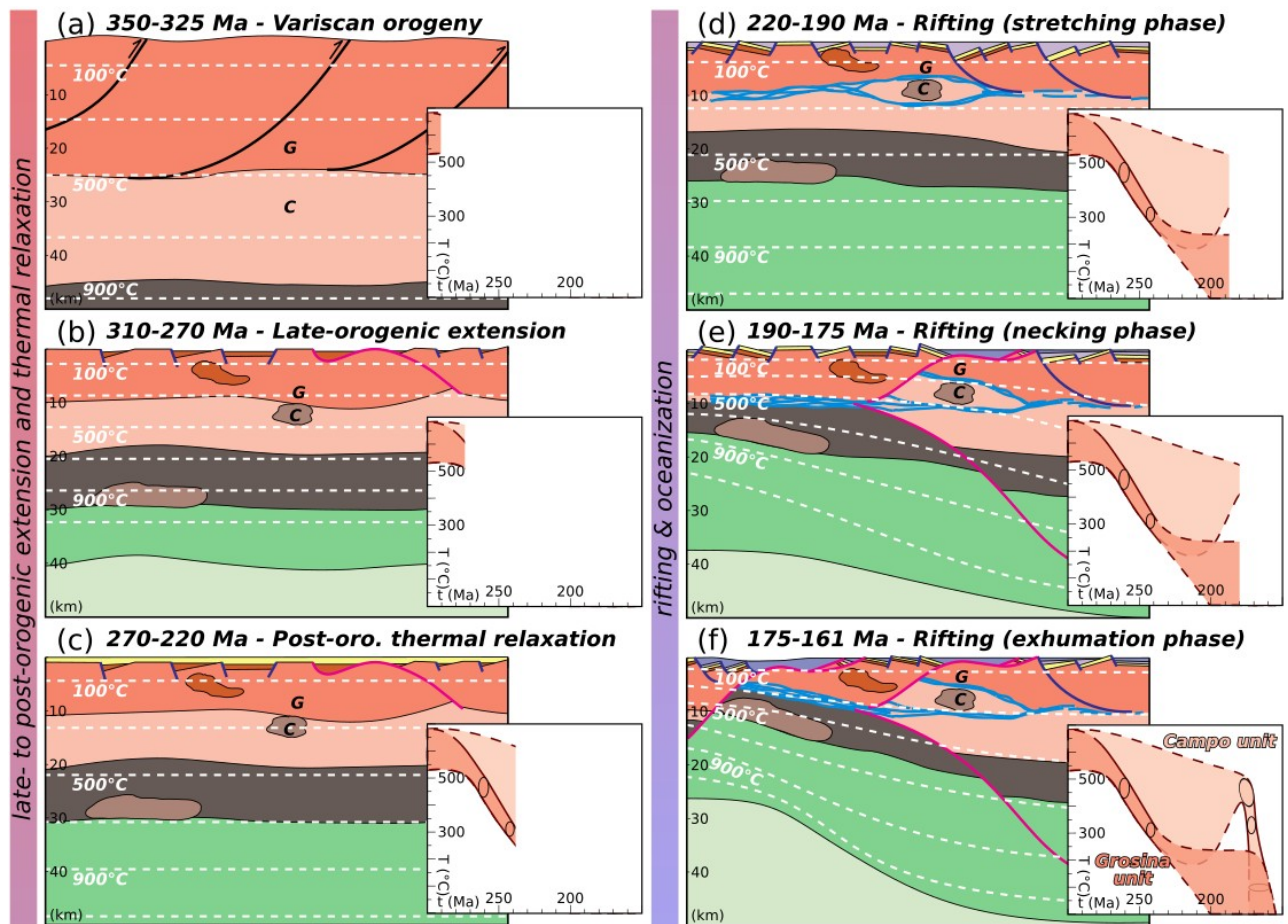
0.05 are discarded and grey-coloured, nodes from paths within the 5 highest probability % are

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circled in white. Path colour as well as nodes colour and size depend on the goodness of the

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temperature-time path to fit measured Ar-Ar ages from our study. See text for details.



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**Fig. 13:** Synthetic tectonic and thermal evolution of the Austroalpine nappes and notably

954 the Campo (light pink) and Grosina (darker pink) units, encompassing (a) the Variscan orogeny, (b)

955 late-orogenic collapse and extension during which the continental crust returns to a standard

956 equilibrated thickness (30-40 km), followed by (c) a period of thermal relaxation from Middle

957 Permian to late Triassic, until a new tectonic cycle develops with the opening of the Alpine Tethys

958 separated in three phases: (d) the stretching of the continental crust by high-angle normal faults and

959 extraction faults, (e) the necking of the continental crust with the concomitant activity of extraction

960 faults (see Petri et al., 2019 for details) and necking detachment faults and (f) exhumation of deep

961 continental crust and lithospheric mantle through multiple detachment faults. Modified from Mohn

962 et al. (2012) and Petri et al. (2019, 2017, 2016). Thermal evolution of the Campo (C) and Grosina

963 (G) units are constrained using the location of nodes and paths with the highest probabilities

964 reported in Fig. 12.

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