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**Dating deformation in the Gran Paradiso Massif (NW Italian Alps): implications for the exhumation of high-pressure rocks in a collisional belt**

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## Abstract

The Gran Paradiso massif, situated in the internal part of the Western Italian Alps, records a complex tectono-metamorphic history involving high-pressure metamorphism and subsequent exhumation during retrograde metamorphism. The exact timing of deformation and, consequently, the geodynamic evolution of this part of the Western Alps is still debated and is addressed here by the application of Rb/Sr geochronology, $^{40}Ar/^{39}Ar$ step heating and $^{40}Ar/^{39}Ar$ total fusion dating techniques. Geochronological results are presented from shear zone samples in the core of the Gran Paradiso massif (Piantonetto Valley), and in the area closer to the contact with the overlying Piedmont ophiolitic domain (south and southwest of Pont Valsavarenche). The shear zones operated during crustal thinning and exhumation of the Gran Paradiso massif. $^{40}Ar/^{39}Ar$ step heating results from shear zones in the Piantonetto Valley show acceptable plateau ages that are interpreted to represent two events of mica growth. Similar ages, and an additional younger age cluster, are recognised in the $^{40}Ar/^{39}Ar$ total fusion analyses, indicating that specific cleavage domains operated at 39.2±0.2, 36.5±0.6 and 33.3±0.4 Ma. P-T pseudosections show a progressive decrease in metamorphic conditions during deformation, suggesting that the age of incipient exhumation and the related deformation in the Piantonetto Valley is equal to or older than 39.2±0.2 Ma. In the Pont area, the last increments of deformation in a top-to-W shear zone postdate 36.6±0.6 Ma (Rb/Sr mineral data), whereas the present-day top-to-W contact of the Gran Paradiso massif with the overlying Piedmont domain is dated at 41.2±1.1 Ma (Rb/Sr multi-mineral isochron age). We propose a model that considers exhumation of the Gran Paradiso nappe at 41-34 Ma. During this period, the nappe was coupled with the Zermatt-Saas zone,

forming an extruding wedge. The kinematics associated with this wedge involved top-to-W shearing within the Gran Paradiso nappe (e.g. Pont area shear zones) and top-to-E shearing at the top of the extruding wedge (e.g. Orco shear zone). Subsequent deformation (after ~34 Ma) was characterised by coaxial strain involving orogenic-scale backfolding and backthrusting.

**Keywords:** Western Alps; Gran Paradiso Massif; Ar-Ar; Rb-Sr; dating deformation; exhumation.

## 1. Introduction

The exhumation history of continental-derived tectonic units that experienced subduction and collisional processes is a matter of ongoing research in structural geology and tectonics (e.g. Ring et al., 1999). Deeply subducted continental blocks are commonly metamorphosed at high pressure (HP) and ultra-high-pressure (UHP) conditions and are subsequently exhumed to shallower crustal levels. In some places, it appears that rapid exhumation has occurred shortly after peak (U)HP metamorphism (Gebauer et al., 1997; Rubatto and Hermann, 2001; Baldwin et al., 2004; Glodny et al., 2005), but tectonic mechanisms that control this early exhumation are still debated.

One place where these processes can be investigated is the Western Alps, where continental-derived units, commonly referred to as the Internal Crystalline Massifs (Monte Rosa, Gran Paradiso and Dora Maira, Figure 1), experienced (U)HP metamorphism during the Alpine orogeny (Chopin, 1984; Reinecke, 1991). The

(U)HP metamorphic evolution in these continental units culminated at roughly 35 Ma, as indicated by U-Pb data on (U)HP-related accessory phases (Gebauer et al., 1997; Rubatto and Gebauer, 1999; Gabudianu Radulescu et al., 2009). This final episode of (U)HP metamorphism was followed by exhumation, which in some cases (e.g. Dora Maira Massif) has been estimated to be relatively rapid, in the order of 1.6-3.4 cm/y (Gebauer et al., 1997; Rubatto and Hermann, 2001; Morag et al., 2008). The exhumation of the UHP units has been explained by a variety of mechanisms. For example, in a model proposed by Wheeler (1991), UHP rocks at the "pip" of the crustal wedge were brought to the surface by the simultaneous operation of normal and reverse shear zones above and below the pip. The role of buoyancy forces during exhumation in a subduction channel has been considered in many other models (e.g. Beaumont et al., 1996; Burov et al., 2001; Yamato et al., 2008). Somewhat more provocative models have suggested that episodic extensional faulting played a role during exhumation (Beltrando et al., 2010).

This paper aims to elucidate the exhumation-related deformation history of the Gran Paradiso Massif by providing geochronological constraints linked to structural observations from a network of mid-crustal shear zones that formed during progressive exhumation of this HP unit. We use a combination of Rb/Sr and $^{40}Ar/^{39}Ar$ dating techniques on micas taken from mylonitic shear zones within the Gran Paradiso metagranite and from the boundary between the Gran Paradiso nappe and the Piedmont domain. The shear zones formed during the main deformation phase of the Gran Paradiso nappe and are associated with the top-to-W shearing and a dominant flattening strain responsible for part of the exhumation of the nappe (Ring and Kassem, 2007).

## 2. Geological setting

The Gran Paradiso Massif, together with the Monte Rosa and Dora Maira Massifs, represents the most internal (lower-plate) continental element of the collisional belt in the Western Alps (Figure 1). These massifs are predominantly composed of pre-Alpine high-grade metamorphic rocks intruded by Permian granitoids (Bertrand et al., 2005; Ring et al., 2005), which have been largely converted during Alpine orogeny into augen-gneisses.

The Alpine tectonometamorphic history of the Gran Paradiso massif involved HP metamorphism and subsequent greenschist- to amphibolite facies re-equilibration during nappe stacking and foreland (Europe)-directed thrusting (Le Bayon and Ballèvre, 2006). The high-pressure mineral assemblages are preserved in eclogitic metabasites and in the kyanite-chloritoid-talc-bearing 'silvery micaschists' within the orthogneisses (Dal Piaz and Lombardo, 1986; Brouwer et al., 2002; Le Bayon and Ballèvre, 2006; Gabudianu Radulescu et al., 2011). Estimates of peak conditions of the eclogite-facies metamorphism cover a wide range of pressures (1.0 GPa $< P <$ 2.7 GPa) and temperatures (410° C $< T <$ 600° C) (Dal Piaz and Lombardo, 1986; Borghi et al., 1996; Brouwer et al., 2002; Wei and Powell, 2003; Meffan-Main et al., 2004; Wei and Powell, 2004; Le Bayon et al., 2006; Gabudianu Radulescu et al., 2009; Gasco et al., 2010). The timing of HP metamorphism is a matter of debate, with available geochronological data for the eclogite facies event varying from 43±0.5 Ma (Meffan-Main et al., 2004) to 33.7±1.6 Ma (Gabudianu Radulescu et al., 2009).

The regional foliation is generally flat-lying and defines a broad dome-structure. A stretching lineation trending approximately E-W is commonly present on the main foliation and is associated with a top-to-W sense of shear (Brouwer et al., 2002; Kassem and Ring, 2004; Le Bayon and Ballèvre, 2006). The main foliation developed during the growth of albite (albite-epidote amphibolite facies, Le Bayon and Ballèvre, 2006; Gasco et al., 2010). It overprints an earlier HP structural fabric, which is locally preserved in mafic and metasedimentary lithologies as a folded schistosity containing relics of garnet, phengite, glaucophane and chloritoid (Le Bayon and Ballèvre, 2006; Gasco et al., 2010). Estimates for P-T conditions during the albite-epidote amphibolite facies overprint range from 0.4-0.6 GPa, 500-550°C (Borghi et al., 1996; Brouwer et al., 2002) to 0.95-1.15 GPa, 600-625°C (Gasco et al., 2010). These variations may reflect different structural levels within the nappe.

## 3. Structure and metamorphism of the studied shear zones

Shear zones were sampled in two different areas within the Gran Paradiso nappe: (1) the Piantonetto Valley in the structurally deep central part of the massif, and (2) the central-western area of the nappe, between Col del Nivolet and Pont Valsavarenche, hereafter referred to as the Pont area (Figure 1). The Pont area is situated in the structurally higher parts of the massif close to the nappe contacts with overlying units.

### *3.1. Piantonetto Valley*

The Piantonetto Valley is a km-scale low-strain domain of massive- to weakly-foliated metagranites in the core of the Gran Paradiso nappe (Callegari et al., 1969). In this area, Alpine deformation is localised and is expressed by discrete (few millimetres to a few decimetres) ductile shear zones that cut the largely undeformed metagranite (Menegon et al., 2006; Menegon and Pennacchioni, 2010) (Figure 2a).

The shear zones are characterised by a mylonitic foliation defined by a millimetric-scale banding. The cleavage domains consist of a mixture of fine-grained (50-200 μm) biotite and white mica. These mica-rich layers alternate with polymineralic feldspar + quartz layers and polycrystalline ribbons of dynamically recrystallised quartz (Figure 2c-d). The metamorphic assemblage along the mylonitic foliation consists of quartz, K-feldspar, plagioclase ($An_{0-5}$ and $An_{17-27}$), biotite, white mica, ± titanite ± epidote ± garnet ± ilmenite. White micas are characterised by a relatively large range of celadonitic component (Si content 3.1-3.4 a.p.f.u, Figure 3).

Three samples (GPP2, GPP3 and GPP5*)* from shear zones in the Piantonetto Valley have been analysed by isotopic dating techniques (Table 1). The shear zones have a wide range of orientations and stretching lineations. Typically, this lineation is shallowly plunging and trends E-W. However, in shear zones that are steeply dipping, the plunge of the stretching lineations is also relatively steep (pitch >60°). Most commonly, shear zones accommodate an overall E-W to NW-SE transtension with a sense of transport of individual shear zone ranging from dip-slip (normal) to strike-slip (Menegon and Pennacchioni, 2010).

Detailed structural analysis on the Piantonetto Valley shear zone has been presented by Menegon and Pennacchioni (2010). These authors have shown a number of observations, which seem to indicate a strong flattening component with a sub-vertical direction of maximum shortening in the km-scale low-strain domain of Piantonetto Valley. Strain analysis coupled with bulk-rock chemical analysis indicate (1) that subhorizontal shear zones formed by general shear with subvertical maximum shortening direction under constant-volume conditions, and (2) that the pods of weakly foliated metagranite between sub-parallel high-strain zones were subjected to strong flattening strain (Menegon et al., 2006; Menegon and Pennacchioni, 2010). In addition, moderately- to steeply dipping shear zones locally show crenulations with sub-horizontal axial planes, which are parallel to the weak foliation in the host metagranite. The formation of the discrete shear zones and the weak foliation in the metagranites is attributed to the top-to-W main deformation phase of the nappe (Menegon and Pennacchioni, 2010). The amount of vertical shortening in the Piantonetto domain has been estimated at about 60%, in accordance with measurements reported by Kassem and Ring (2004) from large portions of the Gran Paradiso nappe.

The metamorphic conditions during the development of the localised shear zones in the Piantonetto Valley have been attributed to lower amphibolite facies metamorphism (Menegon and Pennacchioni, 2010). We have estimated the P-T conditions during deformation from pseudosections (Figure 4), using the NCKFMASH phase topological relation. The calculations consider the bulk composition of two mylonitic samples (GPP2 and GPP5) and their associated syn-kinematic mineral assemblages (Table 2). Whole-rock chemical analyses of major and trace elements were performed by wavelength dispersive X-Ray fluorescence (WD-XRF) analysis with a Philips PW2400 equipped with rhodium tube at the Department of Geosciences, Padova University. Powder samples were mixed and diluted at 1:10 with $Li_2B_4O_7$ and $LiBO_2$ flux, and melted into glass beads. Loss on ignition (LOI) was determined from weight lost after ignition at 860°C for 20 minutes and at 980°C for two hours. FeO was determined with permanganometry using a rhodium tube. The phase diagram involves the following minerals and solution models: biotite (TCC), mica (CHA1), feldspar_B and melt (HP). All other phases are considered as pure phases with an activity of 1. The calculations were performed in Perplex_07

(Connolly, 2005), using solution_file.dat as the solution model definition file. The P-T conditions of interest are at the boundary between the fields Bt + Pl + Mica + Zo + Sa + Qtz + $H_2O$ and Bt + Pl + Mica + Sa + Qtz + $H_2O$, because epidote is progressively consumed in the shear zones due to the reaction epidote + albite à□ oligoclase (Menegon et al., 2006).

Compositional isopleths of plagioclase ($An_{\%}$) and white mica (Al mole % and X Mg) were plotted into the assemblages to determine accurately the P-T fields where mineral chemistry match the composition obtained from electron microprobe analysis. In the studied samples, the composition of recrystallised plagioclase is highly variable and ranges from $An_{17}$ to $An_{27}$. The compositional isopleths of mica representative of samples GPP2 and GPP5 are X Mg 37.5, Al mole% 32 and X Mg 39, Al mole% 41, respectively (Table 2). Only recrystallised mica along the mylonitic foliation has been considered in the modelling.

The P-T pseudosections (Figure 4) estimate conditions of 470-520° C and 0.5 GPa for sample GPP5, and 500-550° C and 0.8-1.0 GPa for sample GPP2. In the context of published P-T paths for the Gran Paradiso unit (Gasco et al., 2010, and references therein), P-T conditions recorded by the Piantonetto shear zones are consistent with the late stages of exhumation of the Gran Paradiso nappe.

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### *3.2. Pont area*

The three samples from the Pont area (Table 1) are from high strain shear zones within the upper parts of the Gran Paradiso Massif south of Pont Valsavarenche (samples GP03-161, GP03-163) and from a quartzitic mylonite east of Col del Nivolet

(ALP04-01). South of Pont, Alpine deformation is distributed heterogeneously (Kassem and Ring, 2004). The most penetrative structure is the pervasive, sub-horizontal foliation on which a WNW-trending stretching lineation occurs. The mylonitic foliation is defined by metamorphic differentiation associated with mica-rich layers alternating with banded quartz and fine-grained recrystallised feldspar layers (Figure 5a). The metamorphic assemblage consists of quartz, K-feldspar, albite, white mica, occasional biotite and opaques. Both samples GP03-161 and GP03-163 have similar mineralogy and microstructures. Shear sense indicators are generally top-to-W, but conjugate top-to-E kinematic indicators also occur. Cross-cutting relationships are not clear, suggesting that both sets of shear bands developed simultaneously, possibly as a result of top-to-W shearing and a superimposed vertical flattening component. This interpretation is consistent with the pronounced flattening strain in the area (Kassem and Ring, 2004). The average amount of vertical flattening is 66% (Ring and Kassem, 2007). There is also evidence for a crenulation cleavage that deforms the pervasive foliation. The new cleavage is commonly at a small angle to the earlier pervasive cleavage but dipping more steeply than the earlier one. Folding associated with the crenulation cleavage tends to have E-dipping axial planes and top-to-W vergence.

Sample ALP04-01 is an ultramylonite from the contact between the Gran Paradiso massif and the Piedmont domain. The mylonitic foliation is shallowly dipping to the NW and a stretching lineation marked by elongation of quartz aggregates is oriented E-W. Top-to-W sense of shear, indicated by white mica fish, is recognised in thin sections (Figure 5b). The sample shows a deformation texture indicative of complete recrystallisation of quartz and white mica phases during deformation. Its composition is dominated by quartz (~80 vol-%), which suggests that the rock is a metamorphosed quartz-dominated psammite. The metamorphic assemblage comprises, beside quartz, abundant phengite and paragonite, and minor quantities of an epidote-group phase, primary chlorite, garnet, apatite and rutile. Phengite has elevated Si contents (between 3.25 and 3.30 per formula unit, Figure 3), and the paragonite component in phengite (Na/[Na+K]) is low, with only 0.01 to 0.08. In average metapelitic compositions, the occurrence of epidote-group phases is restricted to temperatures below ~500°C, and in the field of coexistence of paragonite and phengite, the paragonite component in phengite decreases with increasing pressure (Massonne and Schreyer, 1987). At temperatures near 500°C, the above Na-in-phengite composition is indicative of pressures in the range of 10 kbar or higher (cf. Keller et al., 2005). We note that this is a first-order approximation, because sample ALP04-01 is a quartzite and not a metapelite. However, given the fact that this rock also includes white mica, chlorite and garnet (suggesting a sedimentary protolith with a pelitic component), we do not expect large deviations in the pressure estimates from an average metapelitic composition. The high Si contents, and possibly also the presence of rutile, give additional hints at equilibration of the assemblage at considerable pressures in the deep crust.

## 4. Rb/Sr geochronology

Rb/Sr data were generated at GFZ Potsdam. _ For Rb/Sr dating of ductile deformation we used both the internal mineral isochron approach on bulk mineral separates (e.g. Glodny et al., 2002) and Rb/Sr microsampling (Müller et al., 2000) (Figure 6). Sample selection was based on thin-section observations, combined with sample-specific structural evidence from the field. For bulk mineral separation work, we selected samples of small size (approximately 20-100 g), texturally recording high-strain ductile deformation. Mineral isochron work was focused on samples containing white mica as a high Rb/Sr phase. The Rb/Sr isotope system of white mica is thermally stable to temperatures higher than 500°C-550°C (Freeman et al., 1997; Villa, 1998) but may be fully reset by dynamic recrystallisation at temperatures as low as 350°C (Müller et al., 1999). Careful study of the correlation between microtextures and isotopic signatures has shown that complete synkinematic recrystallisation in mylonites is usually accompanied by isotopic re-equilibration (Müller et al., 1999; Müller et al., 2000; Cliff and Meffan-Main, 2003; Glodny et al., 2008). Therefore, white mica-based Rb/Sr mineral isochron data from penetratively deformed rocks can be used to date the waning stages of mylonitic deformation, given that deformation occurred at or below the temperature range for diffusional Sr-isotopic resetting of white mica, i.e., 500 to 550°C, which was generally the case in the studied shear zones. Wherever possible, white mica was analysed in several, physically different fractions (in terms of magnetic properties and/or grain size). This was done in order to detect possible Sr isotope inhomogeneities resulting from (1) isotopic inheritance; (2) long-term or incomplete dynamic recrystallisation; (3) diffusional Sr redistribution; and (4) alteration processes. This approach ensures control on the possible presence of unequilibrated, pre-deformational white mica relics (Müller et al., 1999).

In sample GPP2, abundant eye-shaped feldspar textural relics and microlithons are interspersed in a very fine-grained, biotite-rich mylonitic matrix. For this sample, we therefore applied in-situ Rb/Sr microsampling, using a microdrill on a polished thick (50 μm) section (Figure 6). The microdrill was mounted to a microscope, allowing the extraction of structurally-controlled microsamples from the mylonites.

Weights of individual cores were between 0.19 and 0.39 mg (supplementary material Table S1).

Mineral concentrates of white mica, quartz-feldspar, apatite, epidote and biotite were extracted from the mylonites. Mica concentrates were ground in ethanol in an agate mortar, and then sieved in ethanol to obtain pure, inclusion-free separates. A detailed description of the Rb/Sr analytical procedure is outlined in Glodny et al. (2008). Isochron parameters were calculated using the Isoplot/Ex program of Ludwig (1999). Decay constants are those recommended by Steiger and Jäger (1977). Standard errors of ± 0.005% for $^{87}Sr/^{86}Sr$ ratios and of ±1.5% for Rb/Sr ratios, as derived from replicate analyses of spiked white mica samples, were applied in isochron age calculations (cf. Kullerud, 1991). For microdrill cores, standard errors of ± 0.01% for $^{87}Sr/^{86}Sr$ ratios and of ± 2.5% for Rb/Sr ratios were used, to allow for possible enhanced sample-to-blank ratios. Individual analytical errors were generally smaller than these values. Rb/Sr analytical data are given in the supplementary material Table S1 and isochron results are shown in Figure 6b and Figure 7. All analytical uncertainties are reported at the 2σ confidence level.

## 5. $^{40}Ar/^{39}Ar$ geochronology

$^{40}Ar/^{39}Ar$ analysis was done using both step-heating and in situ total fusion techniques. For step heating analyses, we used aggregates of muscovite and biotite extracted from sheared lamellae. The samples were washed in distilled water in an ultrasonic bath for more than 30 minutes, and then washed in absolute ethanol. The selected samples, picked under a binocular microscope, were then loaded into a 21-pit aluminium disk along with the neutron fluence monitor Fish Canyon Sanidine (age

28.201 ± 0.046 Ma, Kuiper et al., 2008), following the geometry illustrated in Vasconcelos et al. (2002). The irradiation disks were closed with aluminium covers, wrapped in aluminium foil and vacuum heat sealed into quartz vials. They were irradiated for 14 hours in the Cadmium-lined B-1 CLICIT facility, a TRIGA-type reactor, Oregon State University, USA. Irradiation correction factors are: $(^{36}Ar/^{37}Ar)Ca$ = (2.64 ± 0.02) x 10-4, $(^{39}Ar/^{37}Ar)Ca$ = (7.04 ±0.06) x 10-4, and $(^{40}Ar/^{39}Ar)K$ = (8 ± 3) x 10-4. All the reported ages (supplementary material Table S2) are calculated using the decay constants of Steiger and Jäger (1977).

Mass spectrometric analyses were performed at the University of Queensland Argon Geochronology (UQ-AGES) laboratory. _ After a post-irradiation decay period, the step heating samples were analysed by laser $^{40}Ar/^{39}Ar$ heating, following procedures detailed in Vasconcelos et al. (2002). Before analysis, the rock grains and fluence monitors were baked-out under vacuum at ~200 °C for ca. 12 hours. Each sample was heated incrementally with a continuous-wave Ar-ion laser with a 2 mm wide defocused beam. The fraction of gas released was cleaned through a cryocooled cold-trap (T = -125 °C) and two C-50 SAES Zr-V-Fe getters, and analysed for Ar isotopes in a MAP215-50 mass spectrometer equipped with a third C-50 SAES Zr-V-Fe getter. Full system blanks and air pipettes were determined before and after each sample. Automation and analytical procedures followed are described in Deino and Potts (1990) and Vasconcelos et al. (2002). The data were corrected for mass discrimination, nucleogenic interferences, and atmospheric contamination following the procedures in Vasconcelos et al. (2002), using the software "MassSpec Version 7.527" developed by Alan Deino of the Berkeley Geochronology Centre. An $^{40}Ar/^{36}Ar$ value of 298.56 ± 0.31 for atmospheric argon was used for the calculation of the mass spectrometer discrimination (Renne et al., 2009). J-factors for each Al-disk were determined by the laser total fusion analyses of 15 individual aliquots of neutron fluence monitor, each aliquot consisting of one to three crystals of Fish Canyon sanidine. The mass spectrometer sensitivity was calculated based on the analysis of an air pipette (1.634 x $10^{-13}$ moles $^{40}Ar$) on the Faraday detector (4.257 mV) equipped with a 1 x $10^{11}$ Ohms resistor, yielding a Faraday sensitivity of 3.84 x $10^{-9}$ moles/nA. The current multiplier sensitivity measured on a Balzers 217 Electron Multiplier, operated with a gain of ~145,000 is ~4.5 x $10^{-14}$ moles/nA.

For the total fusion analysis, a 5 mm mini diamond core bit was used to cut two 5 mm diameter by 1 mm thick sample disks from each of the larger hand specimen samples GPP2, GPP3 and GPP5 (Figure 9). The sample disks were rigorously cleaned in an ultrasonic bath using acetone, distilled water and ethanol in succession and dried. The two 5 mm disks from each sample were placed into aluminium irradiation disks along with Fish Canyon sanidine standards and irradiated for 14 hours in the same reactor as the step heating samples. After a decay period, selected sites in each sample were analysed by $^{40}Ar/^{39}Ar$ total fusion laser heating following procedures detailed in Vasconcelos et al. (2002). Laser heating/ablation was done using an excimer laser with a minimum focal diameter of about 40 microns. The excimer laser system used has a wavelength of 248 nM, a user adjustable pulse rate of zero to 50 Hz, a pulse duration of 0.5 to 10 nS, a maximum pulse energy of 30 mJ at 50 Hz and a maximum average power of approximately 2 watts at a pulse frequency of 50 Hz. Sites for analysis in each sample were selected using incident light microscopy. A binocular microscope image was captured for each 5 mm diameter disk for registration and location of individual sample areas in each disk, and images captured of potential areas to be analysed by laser ablation. Sites on the disks selected for analysis were located using a co-focal video camera on the excimer laser sampling system. Single or multiple laser targets were programmed depending upon size and geometry of the sample site selected. The gas released during the laser ablation was cleaned and analysed using exactly the same methods and procedures as used for the gas fractions released by the step heating experiments.

## 6. Interpretation of the geochronological results

$^{40}Ar/^{39}Ar$ analytical data for all the step heating and total fusion analyses are given in the supplementary material Table S2. The results for the step heating analyses are presented in Figure 8. The total fusion results with the sample locations are plotted on the images of the 5 mm sample disks (Figure 9), and a statistical analysis combining the step heating and total fusion results is shown in Figure 10.

### *6.1. Piantonetto Valley samples*

$^{40}Ar/^{39}Ar$ step heating analysis of mica rich domains in sample GPP2 show acceptable plateau ages of 39.3±0.5 Ma and 36.4±0.6 Ma (Figure 8a,b). Results of total fusion analyses show a range of ages at 33-55 Ma, with possible peaks at ~39 and 36 Ma (Figure 9a,b). The micro-texture of this sample shows indications for possibly pre-deformational textural relics or at least for inhomogeneous late stage deformation. However, in both Rb/Sr (see below) and $^{40}Ar/^{39}Ar$ geochronology, no clear isotopic pre-Alpine relics were detected for this sample, and textures thus record an entirely Alpine deformation history.

Rb/Sr micro-sampling in sample GPP2 did not result in a good-fit isochron (Figure 6b). However, the clear correlation between Rb/Sr ratios and Sr isotopic compositions for individual cores can be used to calculate a regression line corresponding to an age of 30.1±5.4 Ma. Exclusion of cores 11 and 12 (Figure 6a) which originate from microlithons and therefore possibly are affected by isotopic inheritance does only slightly change the apparent age (29.9±4.9 Ma, MSWD = 8.2, Figure 6b). This indicates that the observed Sr-isotopic disequilibria are not primarily related to isotopic inheritance. A possible alternative reason for Sr isotopic disequilibria may arise from the fact that the Rb/Sr micro-sampling was performed in a biotite-rich domain of the sample. It cannot be ruled out that biotite, as the main carrier of radiogenic Sr in this domain, was affected by post-deformational alteration or incipient weathering. Biotite is known to be easily oxidised or chloritised during such processes, which tends to disturb and usually rejuvenate apparent Rb/Sr (and K-Ar) ages (Clauer et al., 1982; Jeong et al., 2006; Chen et al., 2007). If this assertion is correct, the apparent age of 30.1±5.4 Ma can be interpreted as a minimum age for the end of ductile deformation in this sample or, alternatively, as a minimum age of pre-alteration closure of the Rb/Sr system in biotite of this sample. These two alternatives are equally compatible with the $^{40}Ar/^{39}Ar$ results. We therefore conclude that the age for the end of deformation is bracketed by the younger of the two $^{40}Ar/^{39}Ar$ step heating plateau ages (36.4±0.6 Ma) and the Rb/Sr micro-sampling-based age (30.1±5.4 Ma).

For sample GPP3, $^{40}Ar/^{39}Ar$ step heating data yield plateau ages of 36.2±1.2 and 37.0±1.0 Ma, respectively (Figure 8c,d), constraining a well-defined, deformation-induced reset of the K-Ar isotopic system of the mica at that time. The total fusion results show ages in the range of 36-47 Ma, excluding one older age of ~100 Ma (Figure 9c,d). The $^{40}Ar/^{39}Ar$ data indicate degassing steps and degassing spectra, with apparent ages up to ~80 Ma (Figure 8c), pointing to the presence of pre-deformational, unequilibrated mica isotopic relics in the sample. This is evident also in the Rb/Sr results for white mica and feldspar for that sample, from which an apparent Rb/Sr mineral age of ~214 Ma can be calculated (supplementary material Table S1). Without additional insights into white mica grain-size *vs.* age correlations, this age value only reveals presence of pre-Alpine white mica relics, the Sr (and Ar) of which was not equilibrated with neighbouring phases during the Alpine overprint.

For sample GPP5, step heating $^{40}Ar/^{39}Ar$ results yield an age of 38.7±0.7 Ma with evidence for earlier history and recrystallisation (Figure 8e,f). The total fusion analyses show broad distribution of ages from 60 to 30 Ma (Figure 9e,f), but there is a peak at ~33.3±0.4 Ma that might correspond to the age of the last partial recrystallisation (see Figure 10m). This sample was not analysed for Rb/Sr isotope systematics.

A statistical analysis combining all the $^{40}Ar/^{39}Ar$ results (both step heating plateau ages and total fusion ages) in the three samples from the Piantonetto Valley (GPP2, GPP3 and GPP5) is presented in Figure 10. Two distinct ages of 39.1±0.7 and 36.8±0.7 are recognised in the analysis of step-heating results (Figure 10a,g). These ages, which are calculated from probability plots, are within the error of the calculated isochron ages (Figure 10b,h), and are therefore regarded as robust constraints for mica growth.

The interpretation of the total fusion results is more problematic. Our results show a large number of older ages (labelled in white in Figure 9), and these were most likely affected by inherited Ar. Therefore, the only way of evaluating the reliability of the total fusion ages is by comparing these results to the plateau ages, and attributing an age to each of the events identified (colour code in Figure 9). Once this is done, we can see a relatively good consistency between the step heating results and the total fusion ages (Figure 10c-f and i-l), thus further supporting the robustness of the two events. In addition, a third event of mica growth is interpreted from the probability plot of total fusion ages from sample GPP5, showing a younger age cluster at 33.3±0.4 Ma (Figure 10m), which is within error of the isochron age (Figure 10n).

We conclude that the mylonites in the Piantonetto Valley were affected by three distinct events of mica growth. The timing of these events, calculated from the probability plots of both step heating plateau ages and total fusion ages, is 39.2±0.2 Ma, 36.5±0.6 Ma and 33.3±0.4 Ma (Figure 10o).

### *6.2. Pont area samples*

Sample GP03-163 did not yield a valid Rb/Sr mineral isochron. This is because of the contrasts between the Sr-isotopic signatures of different white mica grain size fractions, with a higher apparent age for the larger mica crystals. This can be interpreted as evidence for isotopic inheritance and incomplete recrystallisation of the white mica during the last ductile deformation episode (Freeman et al., 1998; Glodny et al., 1998; Müller et al., 1999; Glodny et al., 2008). With presence of augen-like feldspar, incomplete deformation-induced recrystallisation in this sample is also evident from the microtexture. The apparent Rb/Sr age for the mineral pair quartz-feldspar and the smaller (160-125 μm) white mica grains of 36.6±0.6 Ma (Figure 7a), therefore, is interpreted as a maximum age for the end of ductile deformation in this sample. Isotopic inheritance and incomplete dynamic recrystallisation of the white mica population is equally revealed by the $^{40}Ar/^{39}Ar$ data for that sample. Here, no plateau age was obtained, and some degassing steps reveal ages as young as ~45 Ma

(Figure 8g). Combined Rb/Sr and $^{40}Ar/^{39}Ar$ evidence, therefore, indicates that the youngest increments of ductile deformation in this sample postdate 36.6±0.6 Ma.

Sample GP03-161 shows Rb/Sr and $^{40}Ar/^{39}Ar$ results comparable to those of sample GP03-163. For GP03-161, there is again a systematic covariation between white mica grain sizes and apparent ages in the Rb/Sr system. Small crystals (200-125 µm) yield a lower apparent age than larger crystals (Figure 7b). The apparent age of ~80 Ma is thus interpreted as a maximum age for late increments of deformation. $^{40}Ar/^{39}Ar$ step heating spectra for the same sample equally show isotopic disequilibria within the white mica population, and no plateau age is obtained. The nominally youngest degassing steps point to ages between 30 and 35 Ma (Figure 8h,i), which indicate that the latest increments of deformation-induced recrystallisation fell in this age interval or occurred even later.

$^{40}Ar/^{39}Ar$ results for ALP04-01 do not show well-defined plateaus (Figure 8j,k). We interpret the nominally 'youngest' steps of both runs (at around 40 to 45 Ma) as a maximum age for the latest increments of ductile shear. In contrast, Rb/Sr results for this sample yield a good-fit multi-mineral isochron corresponding to an age of 41.2±1.1 Ma (MSWD = 2.2, Figure 7c). There is no systematic correlation between white mica grain size and apparent ages for three different white mica fractions, indicating that complete inter-mineral isotopic re-equilibration between all investigated phases (apatite, epidote, and different white mica fractions) occurred at 41.2±1.1 Ma. Therefore, this age is interpreted as accurately dating the end of ductile deformation in this sample.

### 6.3. Significance of the new $^{40}Ar/^{39}Ar$ data

Careful examination of the complete $^{40}Ar/^{39}Ar$ dataset (Figures 8-10 and Table S2) reveals that there is a broad range of apparent ages, spanning an age interval from the Oligocene (~30 Ma) to the Cretaceous (~100 Ma). Furthermore, three different clusters of ages can be recognised in the combined plateau and total fusion ages (Fig. 10o). Since the typical mica grain sizes analysed in our samples were rather uniform, between 50 and 200 µm, the age differences are not attributed to grain size differences. It therefore appears that the spread of apparent ages cannot be explained as a simple consequence of cooling.

Warren et al. (2012a) have recently published nomograms obtained by numerical modelling of Ar diffusion in white mica. These nomograms describe the extent of Ar loss from a white mica grain as a function of time, temperature, pressure and grain size. This modelling was done for an ideal, open system matrix, allowing for instantaneous removal of any Ar atom as soon as it reaches the grain boundary. When applying this set of nomograms to our $^{40}Ar/^{39}Ar$ data and associated geological context of the deformation zones (mica grain size 50-200 µm, 470-550°C, 0.5-1 GPa), we find that the new $^{40}Ar/^{39}Ar$ data are not compatible with the modelling assumption of open system diffusion. For example, in an open system one would expect that a white mica grain 100 µm in size would, at conditions of 500°C and 1 GPa, loose almost all its Ar in less that 2 Ma. This modelling result is in stark contrast to our observation of apparent ages for ~100 µm mica grains spanning an age range much wider than 2 Ma. Our P-T estimates for the amphibolite- to greenschist-facies shear zones of the Gran Paradiso are consistent with previously published results (e.g. Borghi et al., 1996; Gasco et al., 2010) and are considered reliable. Therefore, the only explanation for the wide range of apparent $^{40}Ar/^{39}Ar$ ages is that the modelling assumption of open system behaviour of the rock matrix of our mica was not universally met during the Alpine history of the Gran Paradiso. Instead, either closed system behaviour must have prevailed most of the time during the Alpine history, or our set of apparent age data must be affected by variable abundance of excess Ar. In any case, the removal of grain boundary Ar, which is usually mediated by fluids, must have been inefficient through most of the Alpine history, probably indicating that either fluid systems acted only on small scales (cf. Warren et al., 2012b) or that fluids were present only episodically during the Alpine deformation history.

We argue, for the following reasons, that in most of our samples, the wide range of $^{40}Ar/^{39}Ar$ ages is due to isotopic inheritance and incomplete deformation-driven isotopic resetting, assisted by prevalence of closed system behaviour. First, wall rocks of the shear zones contain pre-deformational white mica, and there is textural evidence that some of this mica forms unequilibrated relics in the deformation zones. Textural inheritance is usually associated with at least partial isotopic inheritance (Villa, 2010). Second, while excess Ar is ubiquitous in almost all high-pressure (eclogite facies) terranes of the Alps, there is no reason for an *a-priori* suspicion of presence of excess Ar in the amphibolite- to greenschist facies shear zones studied here. Third, we argue that the clear distinction between the three multiple-data-based clusters of ages in shear zones (near 39 Ma, 36 Ma and 33 Ma, Fig. 10o) is better explained by three distinct geological events than by presence of excess Ar, which would be expected to result in a more random distribution of apparent ages. Fourth, the very large variability of the celadonitic component of white mica from the shear zones (Si = 3.1-3.4 a.p.f.u., Figure 3) suggests that the white mica population is composed of crystals or crystal domains that formed at varying P-T conditions during the exhumation history of the shear zones. Fifth, isotopic inheritance is equally seen in some of the Rb-Sr datasets (e.g., sample GP03-161, Figure 7b, and discussion above).

Based on the collective sets of arguments, we conclude that our new $^{40}Ar/^{39}Ar$ data from shear zones mainly record mica growth stages during Alpine deformation or partial to complete isotopic inheritance, with isotopic signatures locally preserved through the Alpine temperature and deformation history. This interpretation is consistent with findings in other high-pressure terranes of white mica Ar dates recording different burial and exhumation stages (Maurel et al., 2003; Gray and Foster, 2004; Putlitz et al., 2005; Di Vincenzo et al., 2006). The only sample that clearly exhibits excess Ar is sample ALP04-01, which demonstrates a textural equilibration and a perfectly equilibrated Rb-Sr systematics (Figure 7c), but has nevertheless yielded erratic $^{40}Ar/^{39}Ar$ step heating results (Figure 8j, k).

Independent age constraints can be used to distinguish between inherited or partially inherited pre-deformational mica ages, and mica ages reflecting deformation episodes in the exhumation related shear zones. We argue that the Rb-Sr isochron age for sample ALP04-01 of 41.2±1.1 Ma (Fig. 7c) can be regarded as a near approximation of the age of HP metamorphism in the Gran Paradiso (see section 7.1 below). Exhumation-related deformation therefore must be younger than that. Indeed, a number of Rb-Sr mineral data on post-HP, retrogressive, and deformationally reworked samples from the Gran Paradiso has been published (Inger and Ramsbotham, 1997, and our own data; Meffan-Main et al., 2004), which consistently yield ages between 39 and 30 Ma. We therefore assume, based on the available Rb-Sr data, that deformation related to Alpine exhumation started at near 39 Ma, an assumption which is consistent with our 'oldest' cluster of $^{40}Ar/^{39}Ar$ apparent ages (Figure 10o) and also with previous $^{40}Ar/^{39}Ar$ data on similar samples (Chopin and Maluski, 1980). We therefore argue that the three clusters of $^{40}Ar/^{39}Ar$ ages equal to or younger than ~39 Ma represent real geologic ages rather than artefacts of undetected excess Ar. We interpret the three clusters as reflecting three major episodes of deformation in our sample set, with associated local, short term fluid ingress facilitating Ar transport (cf. Kelley, 2002; Villa, 2010), and consequently, short term open system behaviour of mica. The individual age peaks of the three age clusters (Figure 10o) may be slightly biased by occasionally incomplete resetting of either pre-deformational mica or of mica formed in an earlier deformation episode and being incompletely reset during a later deformation episode. Nevertheless, we regard these clusters as evidence for exhumation-related deformation starting at ~39 Ma reactivated until ~33 Ma, when the rocks reached lower greenschist facies conditions.

## 7. Discussion

### *7.1 The age of HP metamorphism in the Gran Paradiso massif*

Our new geochronological data indicate that activity along the Piantonetto and Pont shear zones occurred in the time span of 40-30 Ma, with incremental stages recorded at ~39, ~36 and ~33 Ma. These ages are within the range of the recently reported U-Pb SHRIMP ages of 37.4 Ma (monazite) and 33.7 Ma (allanite), which were interpreted as prograde to peak eclogite-facies metamorphism (Gabudianu Radulescu et al., 2009). However, based on our P-T estimates (Figure 4), it seems that activity along the Piantonetto shear zones occurred during decreasing pressure conditions, suggesting that at least the peak pressures of metamorphism occurred near or prior to ~39 Ma. Our pseudosection analysis provides further support for the decreasing pressure conditions during deformation. We recognise that sample GPP2, which deformed under relatively high P conditions (0.8-1.0 GPa), predominantly records the early event at 39.2±0.2 Ma, whereas sample GPP5 that deformed at lower P conditions (0.5 GPa), records all the three events at 39.2±0.2, 36.5±0.6 and 33.3±0.4 Ma.

We have re-evaluated the data presented by Gabudianu Radulescu et al. (2009) and recognised that the initial, inherited Pb in the different allanites analysed by the above authors, has a complex isotopic composition. It is likely that the inherited Pb has originated from at least two sources, as evident from the absence of a clear linear trend in a $^{208}Pb/^{232}Th$ *vs.* $^{238}U/^{206}Pb$ correlation diagram. Therefore, we think that the correction procedure for initial Pb in the allanites was oversimplified, and that, consequently, the uncertainty of the age value was underestimated. In addition, inclusion of the apparent outliers of the presented dataset in a U-Pb age calculation shifts the apparent age to near 35 Ma, which suggests that the absolute age of that allanite was underestimated. Zircon fission track ages in the range of 30-33 Ma

(Hurford et al., 1989; Malusà et al., 2005) further suggest that the ~33.7 Ma U-Pb allanite age of Gabudianu Radulescu et al. (2009) was biased and probably underestimated the true age of HP metamorphism. The most reliable age value for an increment of HP metamorphism so far is a monazite-based U-Pb SHRIMP age from the silvery micaschists of 37.4 ± 0.9 Ma, which was interpreted by Gabudianu Radulescu et al. (2009) as a prograde high-pressure metamorphic stage.

A considerably older HP age of 43±0.5 Ma was reported by Meffan-Main et al. (2004) from micaschists in the northern Gran Paradiso massif. However, this age was determined based on a single Rb/Sr two-point isochron in a disequilibrium assemblage. A similar age has not been reproduced, and therefore this age and its interpretation should be regarded with caution.

Our geochronological results do not provide direct constraints on the timing of HP metamorphism, but suggest that eclogite-facies metamorphism probably took place prior or at ~41 Ma, which is the age of deformation in Col del Nivolet (ALP04-01). The mylonitic foliation in this locality is associated with deformation at the contact between the Gran Paradiso nappe and the Zermatt-Saas zone and most likely represents the juxtaposition of the two units. Based on field evidence (Gasco et al., 2009; Gasco and Gattiglio, 2010), coupling of the two units occurred after the eclogite facies stage, suggesting that HP metamorphism took place at or prior to ~41 Ma. However, based on our data, the possibility for younger activity along this broad shear zone cannot be ruled out.

Deformation in the Piantonetto Valley shear zones was active at ~39 Ma and was associated with nearly isothermal decompression from 39-33 Ma. We interpret this deformation as evidence for thinning during exhumation, meaning that the age of

HP metamorphism cannot be younger than ~39 Ma. This interpretation is consistent with age constraints on the timing of the epidote-amphibolite facies overprint estimated at 39-34 Ma from previous Rb/Sr geochronology (Inger and Ramsbotham, 1997; Meffan-Main et al., 2004). The slight discrepancy between our interpretation for the timing of HP metamorphism (equal to or >39.2±0.2 Ma) and the U-Pb-monazite-based HP age (37.4±0.9 Ma) of Gabudianu Radulescu et al. (2009) remains an open question. Our preferred interpretation is that our $^{40}Ar/^{39}Ar$ mica fabric age of 39.2±0.2 Ma dates an early stage of decompression and exhumation related deformation, meaning that it postdates peak pressure. We also argue that the monazite and allanite-forming reaction sequence in the samples of Gabudianu Radulescu et al. (2009) were possibly driven by the last increments of temperature increase in the samples. Assuming a clockwise P-T path for the Gran Paradiso rocks (Le Bayon et al., 2006; Gabudianu Radulescu et al., 2011), one would presume that peak pressure may predate peak temperature, consistently with the 39.2±0.2 Ma decompression related deformation age (this study) and the slightly younger 37.4±0.9 Ma U-Pb monazite ages (Gabudianu Radulescu et al., 2009).

### *7.2. Exhumation of the Gran Paradiso Massif*

Our geochronological data likely constrain the timing of flattening during exhumation of the Gran Paradiso Massif. As pointed out by Ring and Kassem (2007), the exhumation of the Gran Paradiso Massif was assisted by pronounced pure shear flattening, as indicated by the penetrative flat-lying foliation. This vertical shortening was locally accommodated by shear zones, such as the conjugate sets of localised shear zones in the Piantonetto Valley, which are associated with vertical shortening of

~60% (Menegon and Pennacchioni, 2010). The shear zones are characterised by a sub-horizontal foliation and an E-W trending flat-lying stretching lineation. The dominant sense of shear is top-to-W, but orthorhombic fabrics are common. These observations are in agreement with results by Kassem and Ring (2004), whereby the sub-horizontal orientation of the regional foliation in the Gran Paradiso Massif was interpreted to indicate heterogeneous flattening strain during exhumation superimposed on the already formed nappe stack. The finite-strain study of orthogneisses in the central western and northern sectors of the massif (Kassem and Ring, 2004) indicates strong vertical shortening of 36-88% at the time of the main foliation development.

The actual exhumation mechanism may have involved an early stage of deformation in a ductile extrusion wedge (Wheeler, 1991), associated with a general shear flow and asymmetric (top-to-W) structures (Figure 11a). This was followed by later backfolding and backthrusting at the exhuming nappe front (e.g. Butler and Freeman, 1996) and vertical ductile thinning (Ring and Kassem, 2007) (Figure 11b). The extrusion wedge model has been shown as a viable exhumation mechanism for high-pressure rocks in the Aegean Sea region and the Tauern Window of the Eastern Alps (Ring et al., 2007a; Ring et al., 2007b; Ring and Glodny, 2010). It demands alternating shear sense across the extruding wedge, i.e. top-to-W at the bottom of the wedge and top-to-E at the top (Figure 11a). The top-to-E movement at the top would be due to normal faulting accommodating lateral upward extrusion of the wedge below. However, as shown by Ring and Glodny (2010), the normal fault is basically a geometric effect of extrusion and does not necessarily record lithospheric extension.

In this context, we propose that the exhumation of the Gran Paradiso nappe was facilitated by an extruding wedge (see also Ganne et al., 2006). Within the wedge, the Gran Paradiso nappe was coupled together with the Zermatt-Saas zone, and the dominant sense of shear was top-to-W (Figure 11a). The upper boundary of the extruding wedge was the contact between the Zermatt-Saas zone and Combin unit, which is represented, for example, by the Orco shear zone (southeast of the Gran Paradiso Massif), and is characterised by top-to-E kinematics (Gasco et al., 2009) (Figure 11a). According to Gasco et al. (2009), the Orco shear zone is possibly the continuation of the Combin Fault (Ballèvre and Merle, 1993) and the Gressoney shear zone, which accommodated normal-sense top-to-E tectonic transport at 45-36 Ma (Reddy et al., 1999). It is therefore suggested that top-to-W shearing in the central and western part of the Gran Paradiso Massif (this study) was contemporaneous with top-to-E shearing in the Combin Fault (Reddy et al., 1999).

On the basis of these observations, we propose that the Combin Fault initially facilitated the upward extrusion of the Zermatt-Saas zone. While the latter was exhumed, it was thrust on top of the Gran Paradiso Massif at ~41 Ma (Rb/Sr mineral age of sample ALP04-01), causing HP metamorphism in the Gran Paradiso Massif. At about the same time (i.e. near ~41 Ma and prior to ~39 Ma), the Gran Paradiso Massif was decoupled from the downgoing plate and became part of the extruding wedge (together with the overlying Zermatt-Sass zone). Exhumation and deformation of the Gran Paradiso Massif resulted in the amphibolite-facies overprint and pronounced flattening strain in the middle crust. The strong flattening strain observed in the Gran Paradiso Massif indicates that the extruding wedge was ductile, and accommodated thinning during exhumation. Our results show that ductile thinning, represented by the Piantonetto Valley shear zones, was active at 39-33 Ma.

We envisage that the Gran Paradiso Massif was part of the extruding wedge until about 34 Ma, when this wedge had reached middle crustal depths and backthrusting/backfolding commenced at its front (Figure 11b). The younger ages presented in this study may represent this deformation, which was simultaneous with deformation along the Entrelor shear zone (Freeman et al., 1997) and the Mischabel back-fold (Barnicoat et al., 1995). These structures have been interpreted as a SE-verging backfold and a backthrust, respectively (Barnicoat et al., 1995; Freeman et al., 1997), although contrasting structural observations from the Entrelor shear zone have also supported top-to-NW sense of shearing (Caby, 1996; Bucher et al., 2003). Assuming that the timing of deformation at ~35 Ma was indeed associated with backfolding and backthrusting, we think that the initiation of this deformation represents a stage in the orogenic development when topography became pronounced, and the Alps underwent a more coaxial deformation at the orogen-wide scale.

## 8. Conclusions

New $^{40}Ar/^{39}Ar$ and Rb/Sr data from the Gran Paradiso Massif constrain the timing of deformation during exhumation. The top-to-W shear zone that separates the Gran Paradiso Massif from the Piedmont Domain (in Col del Nivolet) was active at 41.2±1.1 Ma. In more internal shear zones within the Gran Paradiso Massif (near Pont Valsavarenche and in the Piantonetto Valley), we dated conjugate shear zones that accommodated general shear with a combined component of flattening strain and top-to-W asymmetry. These shear zones were active at 39.2±0.2 Ma, 36.5±0.6 Ma and 33.3±0.4 Ma.

Based on the combined P-T and geochronological data from the Piantonetto shear zones, we conclude that about 15 km of exhumation were accommodated in the central Gran Paradiso Massif by vertical ductile thinning. We propose that the early stage of exhumation, at 41-34 Ma, was facilitated by a W-vergent extrusion wedge that consisted of both the Gran Paradiso nappe and the Zermatt-Saas zone. This deformation was followed after ~34 Ma by pure shear horizontal shortening, giving rise to backfolding, backthrusing and doming.

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**Figure captions**

Figure 1. (a) Geological map of the Western Alps showing the location of the Gran Paradiso Massif (modified after Rosenbaum and Lister, 2005). DM, Dora Maira; GP, Gran Paradiso; MR, Monte Rosa. (b) Geological map of the Gran Paradiso Massif and sampling locations (modified after Brouwer et al., 2002). VP, Piantonetto Valley.

Figure 2. Shear zones in the Piantonetto Valley. (a) Outcrop appearance of a narrow shear zone in granite showing pronounced strain localisation and associated mylonitic foliation. (b) Photomicrograph (crossed polarised light) of sample GPP2 showing a mylonitic foliation defined by compositional layering of biotite-rich layers and quartz-feldspar layers. White mica is also present. (c) Photomicrograph (crossed polarised light) of sample GPP3 showing anastomosing cleavage domains defined by white mica and biotite, wrapping quartz and feldspar layers. (d) Photomicrograph (parallel polarised light) of sample GPP5 showing the mylonitic SC foliation defined by layers of white mica and biotite. Mineral abbreviations: Qtz = quartz, Kfs = K-feldspar, WM = white mica, Bt = biotite.

Figure 3. Si *vs.* Al diagram of white micas in mylonitic foliation from the Piantonetto Valley (samples GPP2, GPP3 and GPP5) and from Col del Nivolet (sample ALP-04-01).

Figure 4. NCKFMASH P-T pseudosections (Perple_X, Connolly, 2005) of selected samples. Mineral abbreviations are after Kretz (1983) except mica. (a-b) P-T

pseudosections of GPP5 and GPP2, respectively. The following phases and solution models are considered: Bt (Bio(TCC)), Pl (feldspar_b), Mica (Mica(CHA1)), melt (melt(HP)). Zo, San and Qtz are considered as pure phases. (c-d). Chemical variation of mica. Isopleths of MgO/(MgO+FeO) X 100 and (Al2O3/SiO2) X 100 are used to constrain the equilibrium P-T condition (shown in black box). (e-f). Chemical variation of mica and plagioclase. Isopleths of (Al2O3/SiO2) X 100 and CaO/(CaO+Na2O) X 100 are used to constrain the equilibrium P-T condition (shown in black box). All calculations are in mole. The following chemical compositions were used: GPP5 (75.13% $SiO_2$, 9.17% $Al_2O_3$, 3.00% FeO, 1.19% MgO, 2.03% CaO, 3.53% $Na_2O$, 2.79% $K_2O$, 3.17% $H_2O$), GPP2 (75.47% $SiO_2$, 0.00% $TiO_2$, 9.55% $Al_2O_3$, 2.30% FeO, 1.07% MgO, 0.00% MnO, 1.52% CaO, 3.83% $Na_2O$, 3.08% $K_2O$, 3.19% $H_2O$).

Figure 5. Photomicrographs of samples from the western part of the study area. Images are taken in crossed polarised light. (a) Granitic mylonite GP03-161; (b) Quartzitic ultramylonite (ALP04-01) showing top-to-W (right hand) sense of shearing.

Figure 6. Micro-sample locations and Rb/Sr results of sample GPP2. (a) micro-drill locations; (b) Rb/Sr isochron.

Figure 7. Rb/Sr isochrons for samples (a) GP03-161, (b) GP03-163, and (c) ALP04-01.

Figure 8. $^{40}Ar/^{39}Ar$ step heating results for samples (a-b) GPP2, (c-d) GPP3, (f) GPP5, (g) GP03-163, (h-i) GP03-161, and (j-k) ALP-04-01.

Figure 9. Results of $^{40}Ar/^{39}Ar$ total fusion analyses for samples GPP2, GPP3 and GPP5. Analyses were focussed at the dark layers, which consist of a mixture of fine-grained (between 50 and 200 µm) biotite and white mica. The white layers consist of quartz and K-feldspar (see Figure 2). Circles indicate sampling localities and colours indicate distinct deformational event (green = Event 1, yellow = Event 2, Blue = Event 3). Analyses with excess argon are indicated by white circles.

Figure 10. Statistical analysis combing $^{40}Ar/^{39}Ar$ plateau ages and total fusion ages for the three samples from the Piantonetto Valley (GPP2, GPP3 and GPP5). The results are grouped into three distinct stages of mica growth (Event 1, Event 2 and Event 3). (a) Probability plot of plateau ages (Event 1); (b) isochron for plateau ages (Event 1); (c) probability plot of total fusion ages (Event 1); (d) isochron for total fusion ages (Event 1); (e) probability plot of combined plateau and total fusion ages (Event 1); (f) isochron for combined plateau and total fusion ages (Event 1); (g) probability plot of plateau ages (Event 2); (h) isochron for plateau ages (Event 2); (i) probability plot of total fusion ages (Event 2); (j) isochron for total fusion ages (Event 2); (k) probability plot of combined plateau and total fusion ages (Event 2); (l) isochron for combined plateau and total fusion ages (Event 2); (m) probability plot of total fusion ages (Event 3); (n) isochron for total fusion ages (Event 3); (o) probability plots of combined plateau and total fusion ages (Events 1,2 and 3).

Figure 11. Schematic cross sections illustrating the possible exhumation of the Gran Paradiso Massif. (a) An early stage of general shear deformation in an extrusion wedge, involving top-to-W shearing in the Gran Paradiso Massif and top-to-E shearing in the Combin Fault (b) Pure shear deformation

involving horizontal shortening, doming of the Gran Paradiso Massif and vertical thinning. CdV, Col del Nivolet; E, Entrelor shear zone; GP, Gran Paradiso Massif, Orco Shear Zone; PV, Piantonetto Valley.

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