WAS THE VALAISAN BASIN FLOORED BY OCEANIC CRUST? EVIDENCE OF PERMIAN MAGMATISM IN THE VERSOYEN UNIT (VALAISAN DOMAIN, NW ALPS)

Marco Beltrando*,✉, **Daniela Rubatto***, **Roberto Compagnoni**** and **Gordon Lister***

* *Research School of Earth Sciences, The Australian National University, 0200 Canberra ACT, Australia.*
** *Dipartimento di Scienze Mineralogiche e Petrologiche, Via Valperga Caluso 35, 10125 Torino, Italy.*
✉ *Corresponding author, e-mail: marco.beltrando@anu.edu.au*

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ABSTRACT

The Versoyen Unit (Western Alps) and its mafic rocks have been long considered the remnants of the oceanic crust that supposedly floored the Valaisan basin during the Cretaceous. Here we present U-Pb dating of zircons from a metaleucogabbro and a metagranite from the Versoyen Unit challenging this view. Magmatic zircon cores yield Permian ages of 267±1 and 272±2 Ma, respectively, which are interpreted as dating the crystallization of the magmas. Older inherited crystals and rare Cretaceous zircon rims (~110-100 Ma) are also present. The young rims are characterized by very high U and REE contents. We speculate that the Cretaceous ages are related to a thermal/fluid event possibly induced by the opening of the Valaisan basin. The proposed Permian age for the Versoyen magmatism, together with the lack of geochronological evidence for a Cretaceous oceanic crust in the Valaisan domain *sensu stricto*, may force to reconsider the oceanic nature of the Valaisan Basin. We propose a model in which the Versoyen Unit is unrelated to and pre-dates the extensional tectonics that led to the formation of the Valaisan Basin and the Cretaceous deposition of sediments on this Permian basement. The Permian ages for the Versoyen intrusives correlate with extensive Permian intra-plate magmatism related to lithospheric stretching prior to the break-up of Pangea. The Versoyen Unit becomes the most external Alpine terrane that displays traces of this Permian basic magmatism. Traces of Cretaceous magmatism are preserved in the more internal Chiavenna and Balma units, located in the Central and Western Alps, respectively. However, several lines of evidence suggest that such units may have been unrelated to the Valaisan Basin. Therefore, we propose a new palaeogeographic scenario for the western Tethys, where two independent basins, the Valaisan Basin and the Chiavenna/Balma Ocean, were located between the Briançonnais micro-Plate and the European Plate *sensu stricto*.

INTRODUCTION

The tectonic evolution of mountain belts is affected by the geometry of the convergent plate boundaries and by the relative distribution of oceanic and continental plates and micro-plates (e.g. Royden, 1993; Lister et al., 2001; Wallace et al., 2005). This justifies the importance attributed to palaeogeographic reconstructions that aim at establishing the plate pattern along convergent margins.

Palaeogeographic reconstructions of the Mediterranean area prior to the collision between the European and African Plates are commonly characterized by alternating continental ribbons and narrow oceanic domains (Trümpy, 1980; Platt, 1986; Gealey, 1988; Stampfli, 1993; Froitzheim, 2001; Stampfli et al., 2002; Rosenbaum and Lister, 2005). Such a configuration resulted from the dismembering of Pangea in the Mesozoic, induced by the opening of the Alpine Tethys, with its various branches isolating intervening continental ribbons (Fig. 1).

Several reconstructions suggest that Lower Mesozoic rifting culminated in the opening of the Piemonte Ocean at ca. 160 Ma (Fig. 1b) (Lemoine et al., 1987; Stampfli et al., 1998; Stampfli et al., 2002) followed, at ca. 120 Ma, by the opening of the Valaisan Basin in a more north-westerly position (Fig. 1c) (Trümpy, 1980; Gealey, 1988; Stampfli et al., 1998; Stampfli et al., 2002). Whereas the formation of the Tethys oceanic crust during the Jurassic is well documented by radiometric data (e.g. Ohnenstetter et al., 1981; Borsi et al., 1996; Bill et al., 1997; Rubatto et al., 1998; Costa and Caby, 2001; Rubatto and Hermann, 2003; Stucki et al., 2003; Liati et al., 2005), controversies exist as to whether crustal stretching in the Valaisan Domain resulted in the formation of a narrow strip of oceanic crust.

Various hints of the presence of oceanic crust in the Valaisan Domain are found in the Western and Central Alps. The Misox zone in the Central Alps (Steinmann and Stille, 1999) and the Versoyen Unit in the Western Alps (Antoine, 1971; Cannic et al., 1995) (Fig. 2), which can be unambiguously attributed to the Valaisan Domain on the basis of their current position in the Alpine belt, contain traces of mafic magmatism. In the Tasna Nappe, in the Central Alps (Fig. 2), serpentinites are directly overlain by Cretaceous sediments (Florineth and Froitzheim, 1994). Furthermore, mafic magmatism contemporaneous to the opening of the Valaisan Basin has been reported from the Chiavenna Unit, in the Central Alps, and the Balma Unit, in the Western Alps (Fig. 2, Liati et al., 2002 and 2003).

However, only the Versoyen Unit displays a lithological association characteristic of oceanic crust, with mafic intrusive and extrusive rocks, minor serpentinites and marine sediments. The location of the Versoyen Unit, in contact with the Tertiary Valaisan sediments, also allows to avoid ambiguities related to the uncertain structural position of the Chiavenna Unit and the Balma Unit (see below). Therefore, in order to test the possibility of the existence of Cretaceous oceanic crust in the Valaisan Basin, U-Pb dating of the basic magmatism found in the Versoyen Unit has been undertaken.

GEOLOGICAL SETTING

The Valaisan Domain extends from the Tauern Window, in the Eastern Alps, to the Western Alps (Fig. 2). It is tec-

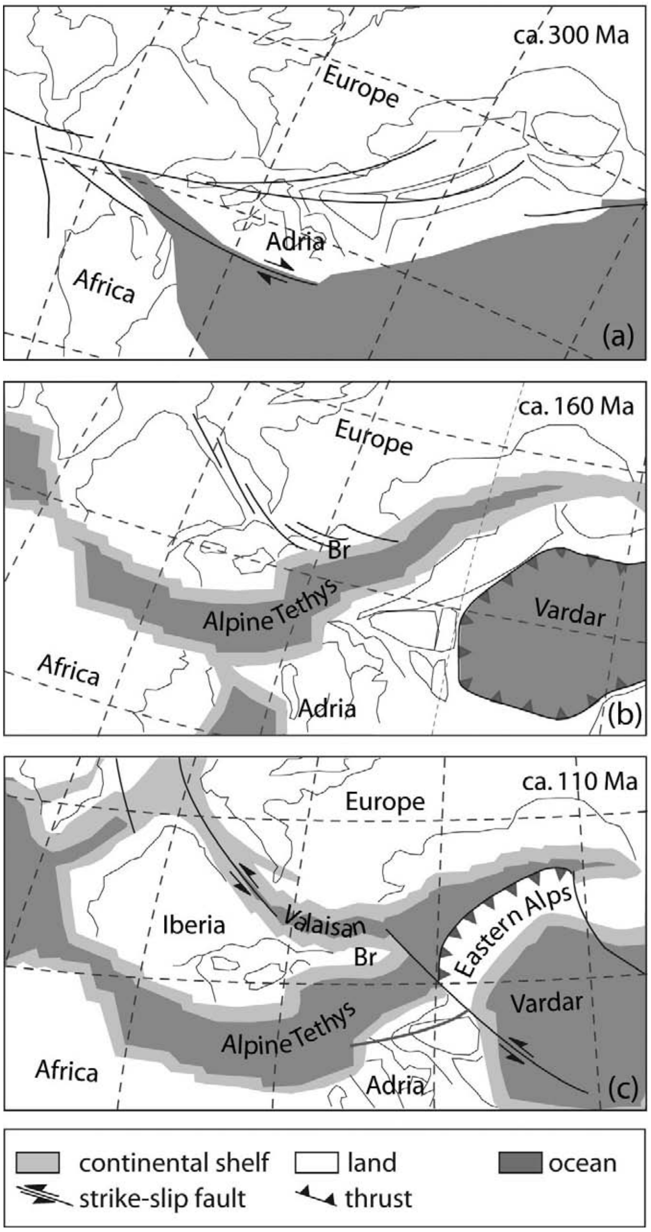

Fig. 1 - Palaeogeographic reconstruction of the Mediterranean area (a) at the Carboniferous-Permian boundary, (b) in the Late Jurassic and (c) in the Middle Cretaceous (modified from Stampfli et al., 2002). Br- Briançonnais.

tonically intercalated between the more external and underlying European Plate and the more internal Briançonnais Domain, both composed of polymetamorphic basement rocks intruded by Palaeozoic plutons and overlain by a Permian to Tertiary sedimentary cover. As a consequence of their current position in the Alpine edifice the rocks of the Valaisan Domain are interpreted as deriving from a basin that was located between the European Plate and the Briançonnais Domain (e.g. Trümpy, 1980). This basin might have extended from the Piemontese-Ligurian Ocean to the Bay of Biscay (Fig. 1, Stampfli, 1993; Stampfli et al., 2002; Schmid et al., 2004), or tapered out to the west of the area that later became part of the Alps (Trümpy, 1980; Gealey, 1988; Jeanbourquin, 1995).

The study area is located in the Western Alps, in proximity of the Petit St. Bernard Pass, at the French-Italian border. This domain, which was initially labelled as the "Nappe des Brèches de Tarentaise" (Barbier, 1948), has been subsequently divided in "Internal" and "External"" Valaisan (Fig. 3; Fügenschuh et al., 1999). This subdivision is based on stratigraphic, metamorphic and structural differences.

The stratigraphic argument hinges on the assumption that the mafic magmatism of the Versoyen Unit is of Cretaceous age and that the entirety of the metasediments of the Valaisan Domain are Mesozoic in age. Given the results of our study (see below), this assumption will have to be reconsidered. According to this view, the External Valaisan represents the thinned European margin where Mesozoic-Tertiary sediments were deposited on continental crust, while in the Internal Valaisan sedimentation took place over newly formed oceanic crust.

In the External Valaisan continental basement rocks are overlain by Permo-Triassic continental deposits and a Triassic-Liassic carbonate platform (pre-rift sequence of Antoine, 1971; Fig. 4). Occasionally, a poorly dated syn-rift sedimentary sequence is found above the pre-rift sediments (Antoine, 1971). These sediments are unconformably overlain by a post-rift sedimentary sequence dated at ca. 112-40 Ma (Antoine et al., 1993).

In the Internal Valaisan, the same post-rift sedimentary sequence is interpreted to have originally overlain Cretaceous oceanic crust, supposedly preserved in the Versoyen Unit (Fig. 4). The Versoyen Unit is composed of pillow lavas and mafic tuffs and a succession of mafic sills and laccoliths, interbedded with metasediments, mostly graphitic calcschists (Fig. 4; Antoine, 1971). Intrusion breccias are sometimes found at the contact between the mafic sills/cumulates and the metasediments (Fig. 5a) and contact metamorphism affects the metasediments in the immediate proximity of the intrusive bodies (Loubat, 1968). Flat Rare Earth Element patterns from the mafic cumulates fall within the range of T-MORB (Cannic et al., 1995). Rare serpentinites, preserving relicts of ortho- and clinopyroxenes, are also found (Fig. 4). This association of mafic cumulates, pillow lavas, serpentinites and black shales has been previously interpreted as recording the formation of oceanic crust on the floor of the Valaisan Basin during the Cretaceous (e.g. Fügenschuh et al., 1999).

Two metagranitoids, the Hautecour and the Punta Rossa (Figs. 3 and 4), are also found in the Versoyen Unit. They have been interpreted as parts of the thinned European margin that was subsequently stretched and intruded by mafic magma during the opening of the Valaisan Basin in the Cretaceous, although the possibility of them being allochthonous terranes has also been proposed (Fügenschuh et al., 1999). A discontinuous layer of polygenic conglomerate containing pieces of metagranitoid as well as of mafic rocks and graphitic schists of the Versoyen Unit has been observed along the southern margin of the Punta Rossa body (Fig. 4).

The Petit St. Bernard Unit, which is located at the interface between the Versoyen Unit and the Briançonnais (Fig. 3), is also attributed to the Internal Valaisan (Bousquet et al., 2002), although its contact with the Versoyen Unit is probably of tectonic origin (Fügenschuh et al., 1999).

As anticipated above, the distinction between the Internal and External Valaisan is also drawn on structural and metamorphic grounds. The contact between the Internal and External Valaisan Units is marked by a thrust that formed during an early phase of deformation related to the Alpine orogeny (D_1 of Fügenschuh et al., 1999) and was subsequently folded during D_2 and D_3.

Detailed fieldwork has revealed that the Versoyen Unit

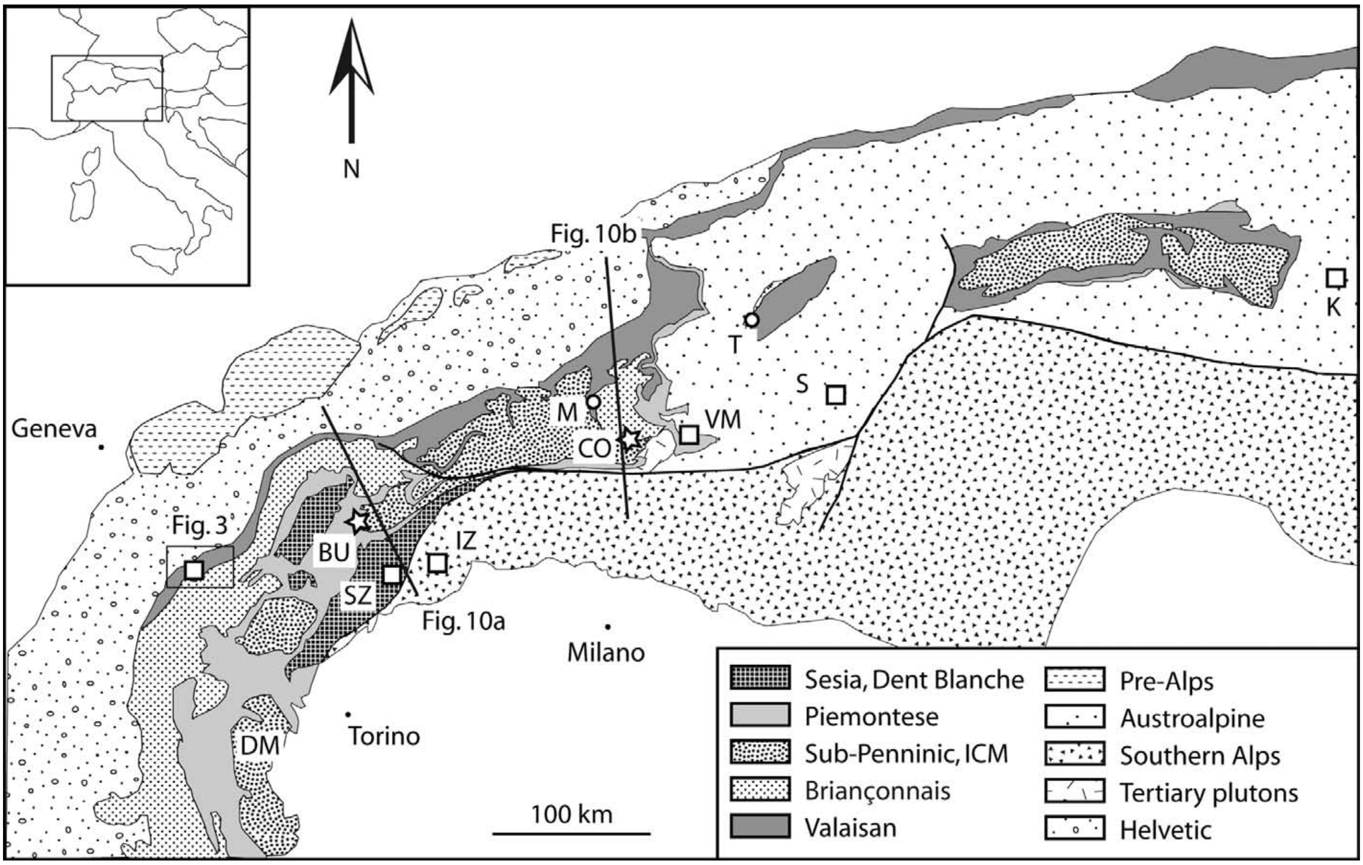

Fig. 2 - Simplified tectonic map of the Alps. Stars indicate the localities were Middle Cretaceous basic magmatism has been found. Dots indicate the position of the Misox Zone and Tasna nappe. Squares indicate localities where Permian basic magmatism has been found (see text for explanation). ICM- Internal Crystalline Massifs; BU- Balma Unit; CO- Chiavenna ophiolites; M- Misox Zone; T- Tasna nappe; IZ- Ivrea Zone; VM- Val Malenco; S- Sondalo gabbroic complex; SZ- Sesia Zone; K- Koralpe; DM- Dora Maira. The positions of the cross sections (Fig. 9) are also shown.

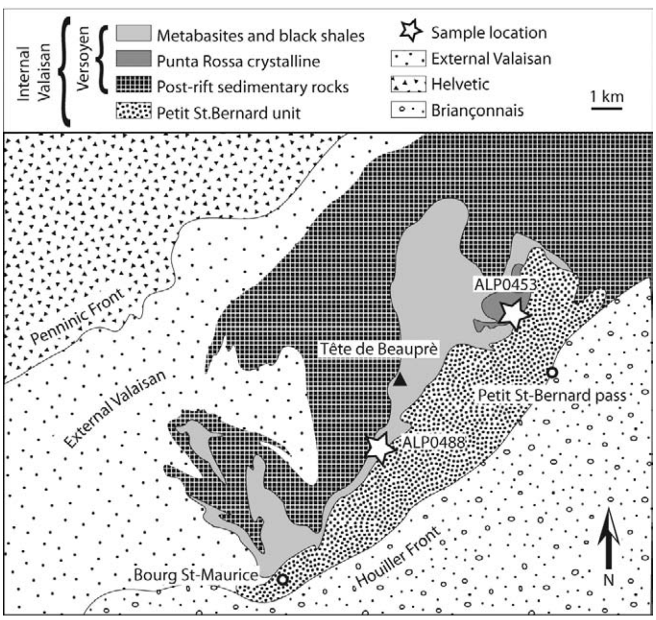

Fig. 3 - Geologic map of the Versoyen and External Valaisan Units in the Petit St. Bernard area (simplified from Fügenschuh et al., 1999). Also shown are the sample locations.

displays a high degree of internal deformation related to Alpine tectonics. The mafic sills and laccoliths are normally found at the core of tight folds, which correspond to the F_1 of Fügenschuh et al. (1999) (Fig. 4). F_1 folds are observed to deform carpholite-quartz veins, indicating that the studied units attained peak pressure conditions prior to F_1 (as also noted by Fügenschuh et al., 1999). This inference is confirmed by the presence of mineral assemblages indicative of greenschist facies conditions in the axial planar foliation to F_1 folds, which is particularly well developed in the graphitic schists. F_1 folds are also observed to deform the contacts between the Punta Rossa metagranitoid and the surrounding graphitic schists (Fig. 4). Subsequent deformation caused the Punta Rossa and the other Versoyen lithologies to be folded together in a km-scale recumbent synform (Fig. 4) related to the F_2 of Fügenschuh et al. (1999). In the Punta Rossa area, later F_3 folds resulted only in minor modifications of the acquired geometry.

The polygenic conglomerate, the Punta Rossa metagranitoid and the rest of the Versoyen Unit underwent the same deformation history during Alpine times. Therefore, it seems plausible to propose that Punta Rossa granitoid was tectonically juxtaposed to the rest of the Versoyen Unit in pre-Alpine times. This view would allow to interpret the polygenic conglomerate, which is now highly deformed, as a tectonic breccia that formed in response to fault movement.

The extent of Alpine metamorphism represents a further element of differentiation between the Internal and External Valaisan. Alpine tectonism resulted in HP metamorphism of the Petit St. Bernard and Versoyen Units at 15-17 kbar and ~350ºC (Bousquet et al., 2002). The Alpine overprint reached only greenschist facies conditions (8 kbar) in the External Valaisan Unit (Goffé and Bousquet, 1997).

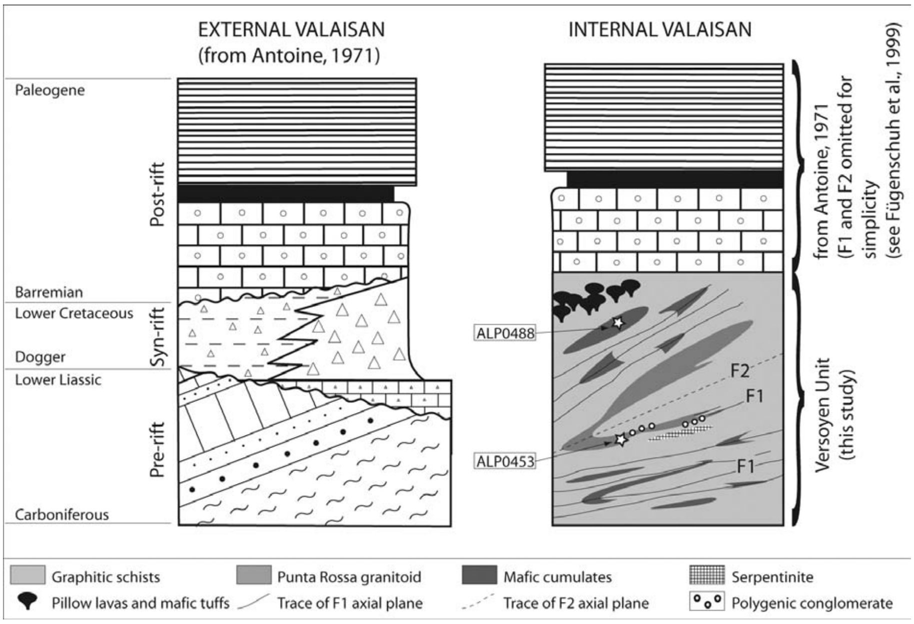

Fig. 4 - Lithostratigraphy of the External and Internal Valaisan. The upper part of the Internal Valaisan column and the External Valaisan column are from Antoine (1971). Deformation in the Mesozoic-Tertiary sediments of the Internal and External Valaisan is not shown and can be found in Fügenschuh et al. (1999). Alpine folding in the Versoyen Unit, which has been the subject of this study, is shown together with the location of the dated samples. Dots in the lower limb of the F2 fold indicate the structural position of the polygenic breccia found at the contact between the Punta Rossa metagranitoid and the other Versoyen lithologies.

EXISTING AGE CONSTRAINTS

Two absolute age constraints has been provided for the Versoyen Unit. Isotope dilution zircon U-Pb analysis has been performed by Schärer et al. (2000) on a leucogabbro, resulting in a lower intercept of the concordia at 309±6 Ma. However, this bulk method did not allow observation of the internal structure of the zircons and distinction of any possible different domains within the same grain that are evident in cathodoluminescence images (see below). Therefore, the published ages are likely to be mixed ages. More recently, Bussy et al. (2005) provided SHRIMP concordia ages of 337.0±4.1 Ma for zircons separated from a gabbroic body of the Vesoyen Unit. Such Carboniferous ages have been interpreted to indicate the time of crystallization of the gabbro (Bussy et al., 2005).

Jurassic ages of the Petit St. Bernard marbles have commonly been extrapolated to supposedly similar lithologies found in the Versoyen Unit. However, as noted by Schärer et al. (2000), several lines of evidence suggest a secondary origin for the marbles of the Versoyen Unit through solution and re-precipitation in a metamorphic environment.

SAMPLE DESCRIPTION

A number of rock types from the Versoyen Unit have been sampled for zircon separation. Samples were collected from the mafic sills, which constitute the majority of the mafic rocks of the area, the graphitic schists at the immediate contact with the sills, a layered gabbro complex and the Punta Rossa metagranitoid. Only the latter two lithologies yielded zircon crystals.

Sample ALP0488 is a metaleucogabbro found in a layered magmatic complex (Fig. 5b and c), which provides the best example of possible Valaisan oceanic crust. This 400 m-thick complex is characterized by mafic and ultramafic cumulates at the base progressively evolving towards the upper part into more doleritic layers (Schärer et al., 2000). Ophitic and cumulate textures can be observed in the largest and less deformed parts of the laccolith. The sampled layer is ca. 40-50 cm thick and is under- and overlain by more mafic layers (Fig. 5c). Alpine metamorphism resulted in the obliteration of the original mineralogy, although the magmatic structure was retained (Fig. 5e). Aggregates of chlorite+calcite+ankerite and minor albite are replacing the original pyroxenes. The matrix is composed of an aggregate of fine grained albite+white mica+chlorite+opaques (Fig. 5e). Tourmaline is a common accessory mineral and is found scattered throughout the rock. Although the Versoyen rocks are reported to have reached eclogite facies conditions (Bousquet et al., 2002) no traces of this early high pressure event are preserved in the studied sample, whose mineral assemblage is indicative of pervasive re-equilibration under greenschist facies conditions. However, fresh carpholite in segregation quartz veins and white mica+chlorite pseudomorphs after carpholite are commonly found in the graphitic schists belonging to the Versoyen Unit.

Sample ALP0453 is a leucocratic metagranitoid from the Punta Rossa body (Figs. 3, 4 and 5c). The rock is largely composed of quartz+albite+white mica (Fig. 5f). Magmatic relicts of quartz and plagioclase are often observed to recrystallize along their margins through subgrain rotation recrystallization in response to Alpine deformation (Fig. 5f). Metamorphic white micas define two foliations. Chlorite is observed as a late accessory phase. The lack of epidote or other Ca-bearing phases indicates that the original plagioclase was albitic. This observation, together with the occurrence of white mica inclusions in magmatic plagioclase and quartz, suggests that the Punta Rossa metagranitoid originated from a Si-rich, most likely peraluminous, magma. The occurrence of jadeite in samples from the Punta Rossa body has been taken as an indication of high pressure metamorphism (Bousquet et al., 2002), but no traces of high pressure minerals were found in ALP0453.

ANALYTICAL TECHNIQUES

Zircons for U-Th-Pb analysis were prepared as mineral separates, mounted in epoxy and polished down to expose the grain centres. Cathodoluminescence (CL) investigation of zircon was carried out at the Electron Microscope Unit,

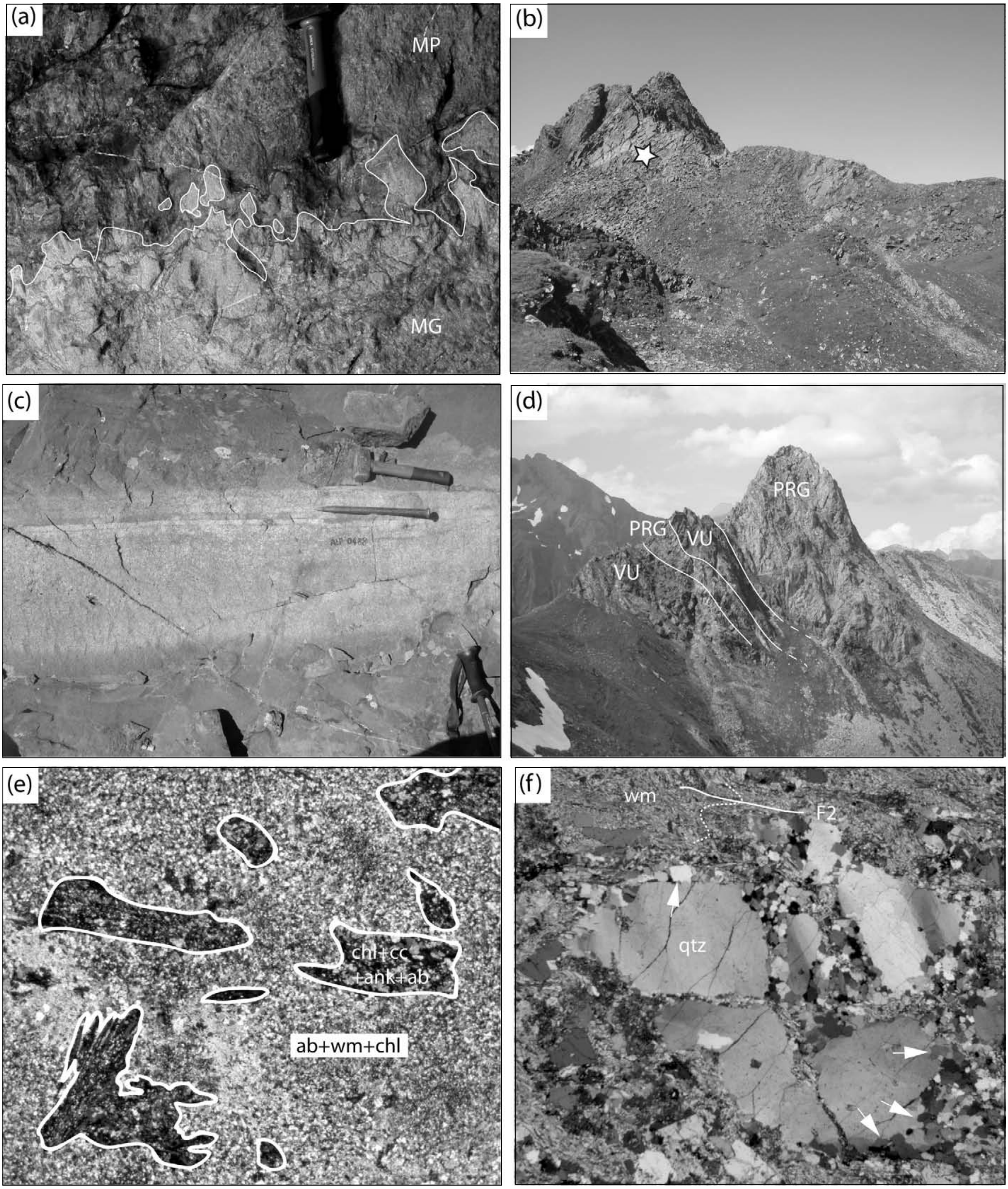

Fig. 5 - (a) Intrusion breccia at the contact between a mafic layers and the metapelites of the Versoyen Complex, suggesting a primary nature for their contact. Field occurrence (b, c, d) and optical microscope images (e, f) of the dated rock samples. (b) Layered gabbro complex with the metaleucogabbro layer (c) from which sample ALP0488 was collected. The original magmatic structure of the metaleucogabbro (ALP0488) (e) is still recognizable. The original pyroxenes (dark sites) are now replaced by a chlorite+calcite+ankerite+(minor) albite assemblage, whilst the original plagioclase is now replaced by fine grained aggregates of albite+white mica+chlorite+opaques. The metamorphic assemblage indicates low-T greenschist facies conditions. (d) The Punta Rossa metagranitoid (PRG) from which ALP0453 was collected. Notice the tight F2 folding with the Versoyen graphitic schists (VU). (f) The original magmatic quartz crystals display wavy extinction and subgrain rotation recrystallization along their margins. Two foliations are visible in the picture, both defined by white micas. Field of view in (e) and (f) is 1.5 cm.

the Australian National University, with an HITACHI S2250-N scanning electron microscope working at 15 kV, ~60 μA and ~20 mm working distance. Additionally, analysis of mineral inclusions in zircon was made with a JEOL 6400 SEM (Electron Microscopy Unit, ANU) with an electron dispersive spectrometer using an acceleration voltage of 15 kV and a beam current of 1 nA. U-Th-Pb analyses were performed using a sensitive, high-resolution ion microprobe (SHRIMP II and SHRIMP RG) at the Research School of Earth Sciences. Instrumental conditions and data acquisition for zircon analysis were generally as described by Compston et al. (1984). The data were collected in sets of seven or six scans throughout the masses. The measured $^{206}Pb/^{238}U$ ratio was corrected using reference zircon from Temora, Australia (TEM, 417 Ma, Black et al., 2003). A zircon of known composition (SL 13) has been used to determine the U content of the target. The data were corrected for common Pb on the basis of the measured $^{207}Pb/^{206}Pb$ ratios as described in Compston et al. (1992). The common Pb composition was assumed to be that of Stacey and Kramers (1975) for the appropriate ages. Age calculations were done using the software Isoplot/Ex (Ludwig, 2000). Isotopic ratios and single ages are reported with 1 sigma errors, whereas mean ages are concordia ages at 95% confidence level.

The trace element content of zircon has been measured by LA-ICP-MS at the Research School of Earth Sciences, Australian National University, using a pulsed 193nm ArF Excimer laser with 100 mJ energy at a repetition rate of 5 Hz (Eggins et al., 1998) coupled to an Agilent 7500 quadrupole ICP-MS. Spot size varied between 19 and 24 μm. The NIST 612 glass was used for external calibration, whereas stoichiometric Si was used as internal standard. Most analyses were located on top of the SHRIMP pits. Contamination from inclusions or zones of different composition was detected by monitoring several elements and integrating only the relevant part of the signal.

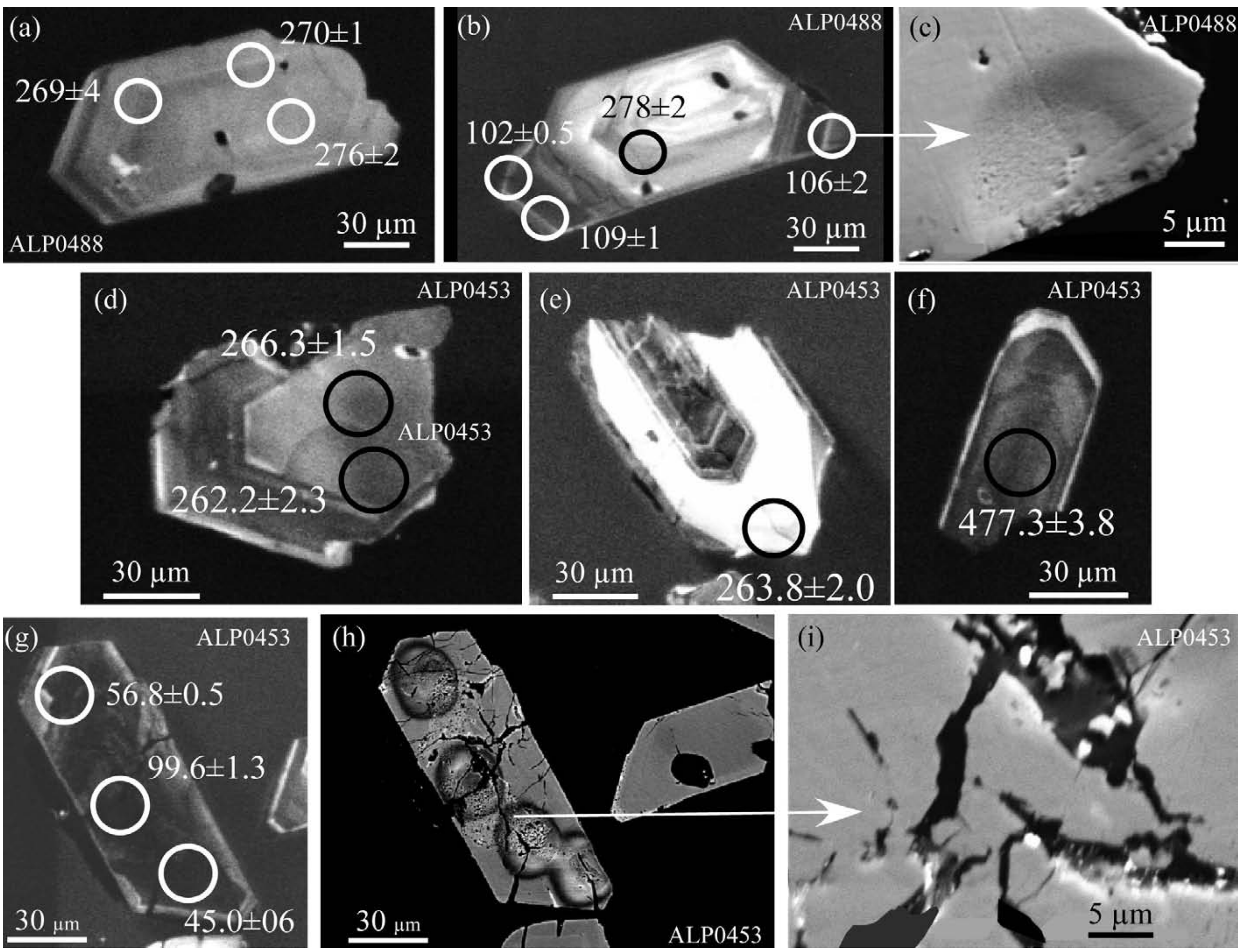

Fig. 6 - Cathodoluminescence and secondary electron images of the analysed zircon crystals. (a) and (b) are CL images of zircon crystals from the metaleucogabbro (ALP0488). The crystal shown in (a) and the core of the crystal shown in (b) are interpreted to be of magmatic origin. Notice the dark U-rich rims in (b). SEM images of the rims (c) have been obtained to check for the presence of cracks or U-oxides (cf. h and i). No inclusions are present and the porosity is interpreted as resulting from the impact of the primary beam on the sample, since undisturbed areas are not porous. (d), (e) and (f) are CL images of zircon crystals from the Punta Rossa metagranite (ALP0453). Crystals (d) and (e) are interpreted to be of magmatic origin. Crystal (f) represents a rare example of an old inherited core. (g) is a CL image and (h) and (i) are SE images of a zircon crystal from ALP0453, which fortuitously samples exclusively the dark, U-rich zircon generation. Measured ages (g) display a wide scatter. SEM images (h, i) reveal the presence of a dense network of fractures. The bright spots in the cracks (i) are U-oxides. Therefore, the scatter of ages is interpreted as resulting from the heterogeneous distribution of U-oxides, inclusion trails and cracks. The same features are observed in the rims of ALP0488.

RESULTS

Metaleucogabbro

The zircons separated from sample ALP0488 (Fig. 6a, b, and c) are euhedral and mostly elongated (aspect ratio 3:1). Their size ranges between 100 and 200 μm. CL investigation revealed the presence of cores and rims in most crystals (Fig. 6b). The cores are characterized by euhedral shape and oscillatory zoning, features typical of magmatic zircons. The 10-30 μm thick rims are euhedral, CL-dark and display some broad band zoning. Inclusions of biotite and ilmenite are often found in the zircon cores. The U content of cores normally ranges between 900 and 2300 ppm (Table 1). The rims are characterized by very high U content, ranging between 4500 and 8600 ppm (Table 1). Th/U ratios are mainly restricted to the 0.13-0.18 range.

Cores yield $^{206}Pb/^{238}U$ ages between 259 and 278 Ma. Nine analyses out of twelve define a major cluster with a concordia age of 272±2 Ma (95% confidence level; Fig. 7a, Table 1). The three remaining analyses are slightly younger (Table 1) and are excluded for suspected Pb loss. Variably older core ages of 477±4, 452±4, 336±3 Ma have also been found.

Only a few rims were sufficiently large to allow the SHRIMP analyses and they yielded apparent ages scattering between ca. 109 and 44 Ma (Table 1). Backscattered images reveal the presence of a dense network of fractures containing quartz and variable quantities of Zr and U oxides (see Fig. 6h and i, taken on zircons from ALP0453. Only one sample is shown here for brevity). This could explain the wide scatter of apparent ages obtained within individual rims that look homogeneous in CL images. However, three analysis performed on crack and inclusion free rims from a grain yielded consistent ages of 106±2, 109±1 and 102±1 Ma (errors are 1 sigma; Fig. 6b and c; Table 1).

Chondrite-normalized REE analyses reveal that the rims are characterized by greater Y and REE contents than the cores, particularly for LREE (Fig. 7a and Table 2). The cores display larger Ce and Eu anomalies. The Ti content of the cores is in the 2-4 ppm range, whilst the rims contain ca. 9 ppm of Ti. However, in these samples the application of the Ti-in-zircon thermometer (Watson et al., 2006) is considered unreliable, since the Si-buffer for the crystallization of the cores and the Si- and Ti-buffer for the formation of the rims are unknown. The relatively low Ti content in the zircon cores could also be related to the fact that zircon is a late crystallizing phase in low-temperature residual liquids of mafic composition.

Table 1 - U, Th, Pb SHRIMP data for zircons from a metaleucogabbro and the Punta Rossa metagranitoid of the Versoyen Unit.

Spot name	U (ppm)	Th (ppm)	Th/U	Common Pb (%)	Uncorrected $^{207}Pb/^{206}Pb$ (% err)	Uncorrected $^{238}U/^{206}Pb$ (% err)	Age $^{206}Pb/^{238}U$ (Ma) ±1σ	CL domain
Metaleucogabbro								
1,3	907	168	0,18	0.0	0.0514 (1.1)	22.68 (0.7)	278±2	*core*
7,1	1277	302	0,24	0.0	0.0518 (0.8)	23.77 (0.7)	266±2	*core*
7,2	1260	282	0,22	0,08	0.0510 (1.1)	23.21 (0.4)	272±1	*core*
8,1	1019	201	0.20	0.0	0.0511 (0.9)	22.90 (0.9)	276±2	*core*
8,2	1051	204	0,19	0.0	0.0517 (1.2)	23.47 (1.5)	269±4	*core*
8,3	1200	335	0,28	0,2	0.0530 (1.3)	23.34 (0.4)	270±1	*core*
9,1	2328	530	0,23	0.0	0.0516 (0.6)	23.42 (0.8)	270±2	*core*
11,1	1042	243	0,23	0,12	0.0514 (0.9)	23.19 (0.7)	272±2	*core*
11,2	1278	315	0,25	0,54	0.0565 (1.1)	23.01 (0.7)	273±2	*core*
2,2	298	27	0,09	0.0	0.0562 (1.1)	12.98 (0.8)	477±4	*core*
4,2	165	63	0,38	0,69	0.0590 (1.5)	13.69 (0.9)	452±4	*core*
6,1	2180	4413	2,02	0.0	0.0533 (0.5)	18.95 (0.7)	336±3	*core*
7,3	842	157	0,19	0,1	0.0529 (1.5)	24.39 (0.5)	259±1	*core (Pb loss)*
10,1	2664	573	0,21	0,08	0.0519 (0.6)	24.15 (1.0)	261±3	*core (Pb loss)*
1,7	765	176	0,23	0,08	0.0510 (1.7)	24.02 (0.6)	263±1	*core (Pb loss)*
1,1	4654	990	0,21	6.9	0.1122 (12)	56.08 (0.7)	106±2	*rim*
1,2	4436	717	0,16	0.11	0.0524 (0.8)	58.33 (0.7)	109±1	*rim*
1,5	5529	889	0,16	1.2	0.0606 (0.9)	62.03 (0.3)	102±1	*rim*
5,1	8671	1530	0,18	3,4	0.0738 (3.5)	138.58 (0.8)	44.7±0.4	*rim*
5,2	6544	1532	0,23	5.0	0.0869 (1.7)	92.70 (0.8)	66.6±0.7	*rim*
5,4	6073	885	0,14	7,9	0.0568 (0.9)	67.15 (0.9)	94.8±0.8	*rim*
Metagranitoid								
3,2	3200	982	0,31	0.0	0.0519 (0.5)	23.61 (0.7)	268±2	*core*
8,2	1029	199	0,19	0,1	0.0517 (1.4)	23.40 (0.5)	269±1	*core*
10,2	2372	370	0,16	0.0	0.0516 (0.8)	24.11 (0.9)	262±2	*core*
10,3	2606	421	0,16	0.0	0.0514 (0.8)	23.71 (0.6)	266±1	*ccre*
13,1	2471	557	0,22	0.0	0.0511 (0.6)	23.60 (0.8)	270±2	*core*
16,1	992	250	0,25	0,08	0.0525 (1.0)	23.93 (0.7)	264±2	*core*
22,1	1896	586	0,31	0,16	0.0516 (1.0)	23.80 (0.4)	264±1	*core*
23,1	2479	564	0,23	0,04	0.0514 (0.9)	23.41 (0.4)	270±1	*core*
25,1	3294	612	0,19	0,03	0.0520 (0.8)	23.54 (0.7)	268±2	*core*
26,1	1718	412	0,24	0.0	0.0509 (1.1)	23.45 (0.6)	269±2	*core*
14,1	976	362	0,37	0,12	0.0576 (1.0)	13.03 (0.8)	477±4	*core*
15,1	2278	459	0.20	0,05	0.0523 (0.6)	24.72 (0.9)	255±2	*core (Pb loss)*
13,2	3410	724	0,21	0,08	0.0516 (0.8)	24.67 (0.3)	256±2	*core (Pb loss)*
1,3	1865	272	0,15	0.0	0.0518 (1.0)	24.19 (0.6)	261±2	*core (Pb loss)*
1,2	2775	375	0,13	0.0	0.0507 (0.7)	24.09 (0.7)	262±2	*core (Pb loss)*
3,1	6879	829	0,12	0,9	0.0546 (0.8)	90.34 (0.8)	70.3±0.6	*rim*
7,1	7689	1107	0,14	1,6	0.0598 (0.7)	139.88 (1.2)	45.0±0.6	*rim*
7,2	7649	1149	0,15	2,2	0.0643 (0.6)	110.35 (0.8)	56.8±0.5	*rim*
7,3	6824	2887	0,42	2,6	0.0687 (3.3)	64.49 (1.1)	99.6±1.3	*rim*

Table 2 - Trace element abundance (in ppm) in the zircons of the metaleucogabbro and metagranite from the Versoyen Unit as determined by LA-ICP-MS analysis.

Sample	*ALP0488*	*ALP0488*	*ALP0488*	*ALP0488*	*ALP0488*	*ALP0488*	*ALP0488*	*ALP0453*	*ALP0453*	*ALP0453*	*ALP0453*	*ALP0453*
position	*rim*	*rim*	*core*	*core*	*core*	*core*	*core*	*rim*	*rim*	*rim*	*rim*	*core*
P	3042	3004	894	1610	1709	2057	1738	4020	4609	4798	3034	1541
Ti	8,83	9,59	3,20	2,44	bdl	2,56	3,88	44,61	36,11	9,11	8,13	10,05
Sr	51,40	41,82	0,95	0,73	0,79	1,03	0,95	64,34	71,85	95,74	69,50	1,00
Y	5461	6390	2689	1957	2511	2639	2793	8629	9263	8842	6320	1955
Nb	27,26	30,82	2,65	2,60	2,95	3,24	5,12	123,14	90,76	47,19	31,45	7,51
La	1,10	1,53	bdl	0,02	bdl	0,28	bdl	8,68	9,56	3,24	0,47	0,11
Ce	13,35	15,58	5,15	2,83	3,94	4,18	5,83	83,48	79,87	20,42	12,23	3,11
Pr	0,716	0,861	0,102	0,032	0,082	0,197	0,058	7,866	7,507	1,316	0,187	0,078
Nd	4,04	4,93	3,29	0,80	1,12	1,81	1,62	39,25	39,42	7,23	2,31	1,26
Sm	9,06	10,74	8,25	2,72	3,77	4,40	5,31	35,71	41,73	13,58	9,10	3,23
Eu	0,205	0,455	0,280	0,060	0,089	0,102	0,148	5,682	6,256	0,581	0,077	0,132
Gd	72,0	87,2	51,3	24,4	33,0	34,3	42,2	159,9	193,2	116,8	82,6	26,0
Tb	32,9	39,4	19,1	10,8	14,2	15,0	17,3	62,3	72,6	55,3	39,2	11,2
Dy	479	565	251	160	197	217	241	798	905	798	562	165
Ho	188,0	221,0	93,7	65,0	78,8	86,5	94,2	291,2	320,7	302,7	216,5	66,1
Er	861	1023	419	332	397	429	452	1329	1449	1380	994,0	337
Tm	195,4	230,1	92,0	75,3	83,6	99,4	102,4	304,0	324,7	306,8	221,5	76,7
Yb	1807	2182	852	743	777	974	989	2901	3075	2858	2080	775
Lu	284	345	140	125	144	160	156	454	474	442	324	128
Hf	11028	11679	8096	8707	7065	8999	8672	10611	11580	11146	11132	9139
Ta	13,4	14,2	1,3	1,5	1,4	1,7	2,5	39,4	38,4	21,7	15,3	16,9
Th	1078	819	254	103	203	165	268	1137	1290	1598	1407	116
U	5187	4796	851	672	799	939	1135	6905	7927	8111	6324	793

Metagranite

Zircons separated from sample ALP0453 are characterized by broad-band oscillatory zoning (Fig. 6d and e) of likely magmatic origin. Euhedral, thin (≤ 30 μm) U-rich rims are rarely found. As in the previous sample, the U content of the rims is extremely high (4500-9000 ppm, Table 1). Cores yield ages between 256±2 and 270±2 Ma (1 sigma) The main cluster of analyses defines a concordia age of 267±1 Ma (95% confidence level; Fig. 8b), whereas four younger analyses are excluded for suspected Pb loss. Only one older core has been found, which is euhedral and yielded an age of 477±4 Ma (Fig. 6f). The apparent ages of four rims range between 100 and 45 Ma. However, the analyses were obtained from heavily fractured domains (Fig. 6g, h and i).

As for ALP0488, the U-rich rims are characterized by higher Y and REE contents than the cores (Table 2). The positive Ce anomaly and the negative Eu anomalies are more significant in the cores than in the rims (Fig. 7b). Highly variable Ti contents ranging from 8 to 45 ppm have been measured in the rims, whilst the cores are characterized by 10 ppm of Ti. Such contents result in estimated temperatures of 720-900°C for the rims and 740°C for the cores (Watson et al., 2006). These results for the core should be considered a minimum estimate given that the Ti-buffer is unknown. The high variability of Ti content in the rims points towards the presence of a possible contaminant phase, which has not been detected with optical, CL and SE imaging or anomaly in the elements analysed with the LA-ICP-MS.

DISCUSSION

Data Interpretation

The large majority of analysed zircons yielded Permian ages, resulting in concordia ages of 267±1 Ma for the metagranite and 272±2 Ma for the metaleucogabbro. The euhedral shape, oscillatory zoning, high Th/U ratios, as well the REE pattern with the enrichment in HREE and the negative Eu anomaly, point towards a magmatic origin for the Permian cores. For both rocks the Permian age is therefore interpreted as dating the crystallization of the magmatic protolith. Since the studied metaleucogabbro (Fig. 5b and c) is part of one of the several mafic layered complexes that constitute the Versoyen Unit, its age is considered representative of the age of formation of the laccoliths. Its mineralogy and relationship with the more mafic layers indicate that the dated sample is one of the most evolved products of the magmatic episode that resulted in the observed intrusions. It could be argued that the more mafic layers and the more acidic ones were formed during distinct magmatic episodes of possibly very different ages. However, a continuous evolution from more mafic olivine-bearing layers to more acidic ones is observed. This feature is consistent with fractional crystallization of mafic magmas during a single magmatic episode. Interaction with the surrounding pelitic rocks may account for the presence of rare inherited cores, with ages of 336±3, 452±4 and 477±4 Ma.

The oldest zircon core found in the Punta Rossa metagranite also yielded an age of 477±4 Ma. Pre-Permian cores are characterized by little or no zoning under CL and are interpreted as xenocrysts or inherited from the magma source.

The ages measured on zircon rims present a wide scatter between 109±1 and 45.0±0.6 Ma (Table 1). SEM images revealed widespread occurrence of fractures containing Zr and U oxides (Fig. 6). We suggest that the high U content of the rims resulted in partial metamictization of the crystal lattice and solution and reprecipitation of Zr and U as oxides (Fig. 6i). The uneven distribution of the oxides, which is controlled by the presence of cracks, results in marked local heterogeneities in the U-Pb ratio and therefore in the calculated dates. It is noteworthy that the three spots that were placed in fracture and inclusion-free areas yielded ages of 102±1, 106±2 and 109±1 Ma (Fig. 6b and c).

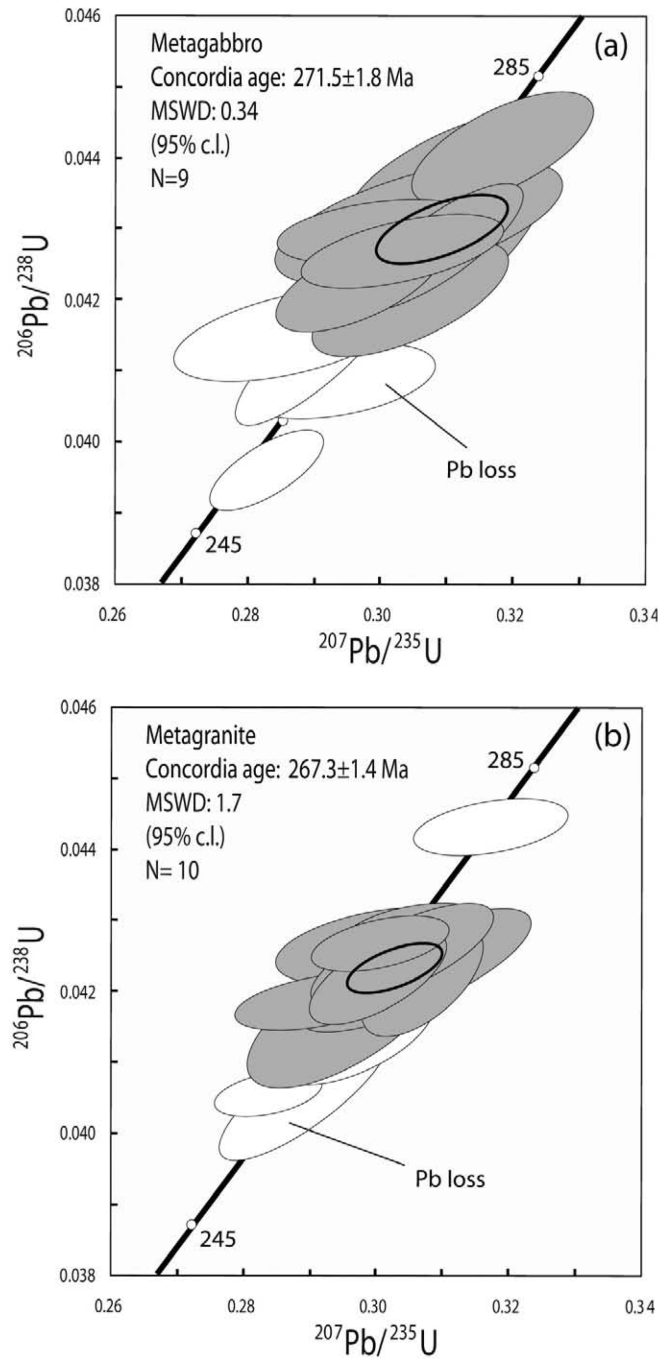

Fig. 7 - Concordia diagrams for SHRIMP analyses of zircons from (a) the metaleucogabbro layer and (b) the Punta Rossa metagranite from the Versoyen Unit. White ellipses indicate data points that were not used to calculate the concordia ages. Ellipses with thicker outline represent the concordia ages. The ellipses of the individual data points represent 2σ errors.

The rims appear to be euhedral overgrowths of the magmatic cores. Because the Cretaceous rims grew significantly after the Permian cores, we suggest that they are the product of a metamorphic process. The higher REE content of the rims with respect to the cores suggests that the rims did not form through low-T recrystallization of the older cores, a process that generally results in loss of REE (e.g. Hoskin and Black, 2000; Rubatto, 2002). The REE enrichment rather suggests that the rims crystallized from a fluid/melt phase. Indeed, high REE zircons were found in fluid-related veins in the Sesia zone (Rubatto, 2002). We tentatively suggest that the interaction between the Versoyen rocks and fluids of crustal origin in the Cretaceous may account for the zircon rim growth and the enrichment in incompatible elements, such as U and, possibly, B: the zircon rims are rich in U and tourmaline, a boron-bearing mineral, is observed in the metaleucogabbro. It is difficult to assess at what stage of the evolution of the rock the tourmaline crystallization took place, but we speculate it may have formed as a consequence of interaction with external fluids, since tourmaline is never found as a primary phase in gabbros.

An alternative interpretation of our data would be that the 110-100 Ma rims found in the metaleucogabbro formed through mixing of an intruding mafic magma and the crustal host rocks in the Cretaceous. In this case the zircon cores would be inherited from the rocks that interacted with the mafic magma during its ascent and emplacement and only the thin rims would have recorded the emplacement age. However, we regard this possibility unlikely, since the high Zr solubility in mafic magmas (Watson and Harrison, 1983) would not allow this degree of inheritance to be preserved.

We also consider unlikely the possibility that the U-rich rims record the Alpine metamorphism of the Versoyen Unit. Cretaceous HP metamorphism is well known in the Eastern Alps (Thöni and Miller, 1996), but in the Western Alps this age corresponds to extensional tectonics, as indicated by the sedimentological record (e.g. Antoine, 1971; Trümpy, 1980).

Therefore, with the available data we propose that the leucogabbro and the Punta Rossa granite intruded in the Permian, at 272±2 and 267±1 Ma, respectively. Subsequent-

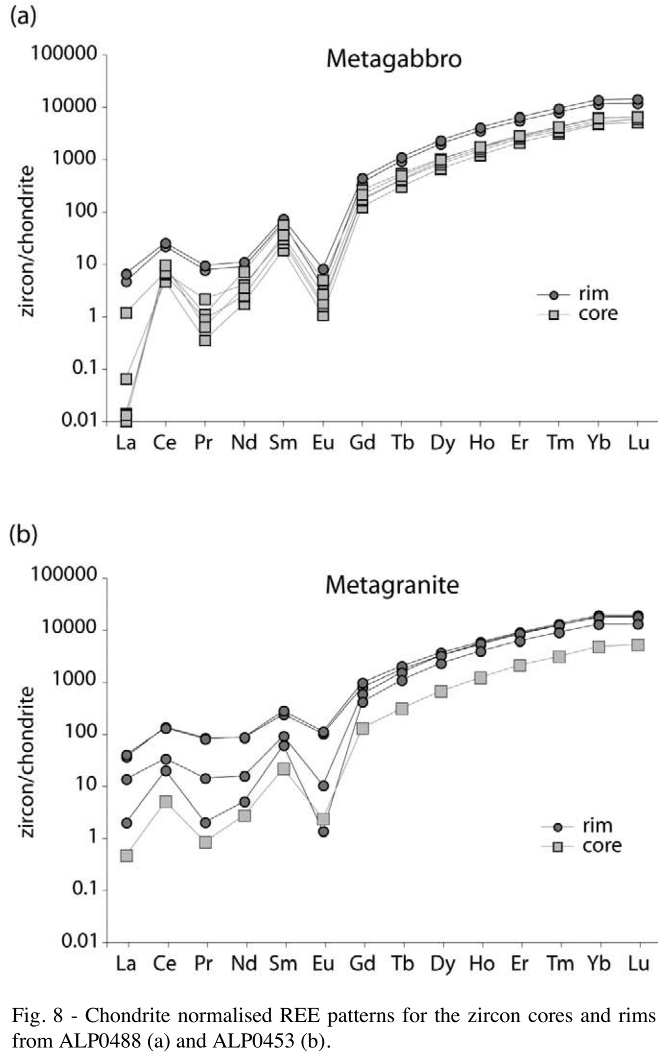

Fig. 8 - Chondrite normalised REE patterns for the zircon cores and rims from ALP0488 (a) and ALP0453 (b).

ly, in the Cretaceous, at ca. 110-100 Ma, a thermal/fluid event (see below) resulted in the formation of zircon rims in both rock types.

Regional significance

Existing studies tend to attribute the magmatism of the Versoyen Unit to the formation of oceanic crust at the base of the Valaisan Basin in the Cretaceous. According to this view, lithospheric thinning resulted in upwelling of asthenosphere and its subsequent partial melting in a rift or mid-ocean ridge environment (e.g. Antoine, 1971; Fügenschuh et al., 1999). This model is based on the assumption that the observed lithological association of layered mafic complexes, pillow lavas, serpentinites and mafic tuffs, which is observed in ophiolite sequences worldwide, formed in the Cretaceous. Sedimentological data from the Valaisan Domain provide indirect support to this hypothesis, showing that crustal extension resulted in the formation of the Valaisan sedimentary basin at that time (e.g. Florineth and Froitzhem, 1994). However, our geochronological investigation suggests that the intrusive rocks of the Versoyen Unit crystallized in the Permian. In the absence of any other age data to support Cretaceous magmatism in the Versoyen Unit, we need to propose and alternative scenario: the Versoyen Unit represents the basement on which the Valaisan sediments were deposited, but the magmatism within the Versoyen Unit is unrelated to and pre-dates the extensional tectonic that led to the formation of the basin itself.

It may be argued that the Versoyen sequence represents a section of oceanic crust that formed in the Permian. However, the rare serpentinites preserve relicts of ortho-and clinopyroxene, suggesting an origin through hydration of sub-continental lherzolitic mantle. Such fertile mantle could have not been the source of the basic magmatism observed in the Versoyen Unit. Therefore, the serpentinites were probably part of the basement in which the mafic magmas intruded. At this stage it is difficult to establish the original relationship between the mafic intrusive and extrusive rocks, since no radiometric age constraints are available for the pillow lavas and mafic tuffs. However, the high degree of deformation of the Internal Valaisan calls for great caution when attempting to extract a pre-Alpine stratigraphic section of the Versoyen Unit from the presently observable lithological sequence. The presence of tectonic contacts within the Versoyen Unit itself, as shown in this study and in Fügenschuh et al. (1999), re-enforces the need for this cautious approach.

The pertinence of the Punta Rossa metagranitoid to the Versoyen Unit is somewhat unclear. The conglomerate found at the contact between the granitoid body and the rest of the Versoyen rocks may be interpreted as a tectonic breccia, indicating the presence of a pre-Alpine fault between the Punta Rossa granitoid and the schists and mafic rocks of the Versoyen Unit. However, the proposed similar crystallization ages for the metaleucogabbro and the metagranite indicate that they were part of the same magmatic cycle. Therefore, we suggest that they both intruded into the Valaisan basement during an event of bimodal magmatism in the Permian. Permian bimodal magmatism has been found in other terranes that later became part of the Alps (Fig. 2), namely the Sesia zone (Rubatto et al., 1999), the Ivrea zone (e.g. Handy et al., 1999) and the Val Malenco area (Hermann et al., 1997). Mafic magmatism of similar age has been reported from the Tasna Nappe (Froitzheim and Rubatto, 1998), in the Central Alps, the Sondalo gabbroic complex (Tribuzio et al., 1999), Koralpe in the Eastern Alps (Fig. 2; Thöni and Jagoutz, 1992) and the External Liguride Units, in the Northern Apennines (Meli et al., 1996). Acidic magmatism with similar age is also found in a number of western Alpine units, among which the Mont Blanc Massif (Bussy et al., 1996), the Gran Paradiso Massif (Bertrand et al., 2000), the Monte Rosa Massif (Lange et al., 2000) and the Dora Maira Massif (Gebauer et al., 1997). This extensive Permian magmatism has been interpreted in the context of extensional tectonism coupled with magmatic underplating, activity of extensional shear zones and incipient exhumation of the lower crust at ca. 270-290 Ma (Fig. 9a; e.g. Rutter et al., 1993; Mayer et al., 2000).

Notably, the Versoyen Unit represents the most external Alpine terrane that has recorded this widespread episode of mafic magmatism (Fig. 2). This finding calls for a re-assessment of the scale of this episode of lithospheric stretching. Prior to this study, the only traces of Permian mafic magmatism (Thöni and Jagoutz, 1992; Meli et al., 1996; Hermann et al., 1997; Handy et al., 1999; Rubatto et al., 1999; Tribuzio et al., 1999) were limited to rock units that later became part of the Adria micro-continent, a promontory of Africa, as a result of the disruption of Pangea (Fig. 9b). The Versoyen Unit, instead, became part of the European Plate, therefore indicating that it was originally located in a more north-westerly position. Hence, the asthenosphere upwelling that fed the basic magmatism probably occurred at a broader scale (Fig. 9a) than hitherto proposed.

The interpretation of the regional significance of the 110-100 Ma old thin rims is more difficult. Stampfli et al. (1998) proposed that the opening of the Valaisan "Ocean" occurred at ca. 130-90 Ma based on the sedimentological record of adjacent areas. Mafic magmatism has been reported from the Chiavenna ophiolites and the Balma Unit at ca. 93 Ma (Liati et al., 2002 and 2003). The existence of some MORB-type magmatic activity in the Mediterranean area in the Late Cretaceous (Talerico, 2001), together with the indication that crustal stretching was occurring in the Valaisan area (Florineth and Froitzheim, 1994; Froitzheim et al., 1996), suggests that the Versoyen Unit may have been affected by a thermal pulse at that time. Such thermal pulse may have triggered partial melting of crustal rocks and/or metamorphic mineral reactions with subsequent fluid mobilization, as possibly recorded by the zircon rims. This event, which occurred at ca. 110-100 Ma, affected an older crustal section that had formed by the Late Permian. The Versoyen Unit was probably part of the basement rocks on which the Valaisan sediments were deposited. It cannot be excluded that lithospheric thinning culminated in the formation of oceanic crust in the Valaisan Basin, but geochronological evidence has not been found in the Valaisan Domain to this date.

Towards a new Cretaceous palaeogeography?

In the light of the new Permian ages for the magmatism of the Versoyen Unit, the possibility of the Valaisan Basin being floored by oceanic crust comes into question. No relicts of true oceanic crust have yet been found in the Valaisan Domain *sensu stricto* (see below) in the Western and Central Alps. In the Tasna Nappe, which is part of the Valaisan Domain in the Engadine Window, crustal stretching resulted in exposure of serpentinites at the basin floor (Florineth and Froitzheim, 1994). However, the sub-continental nature of the serpentinized mantle rocks (Florineth and Froitzheim, 1994) suggests that they did not feed any basaltic magmatism. Furthermore, recent investigation

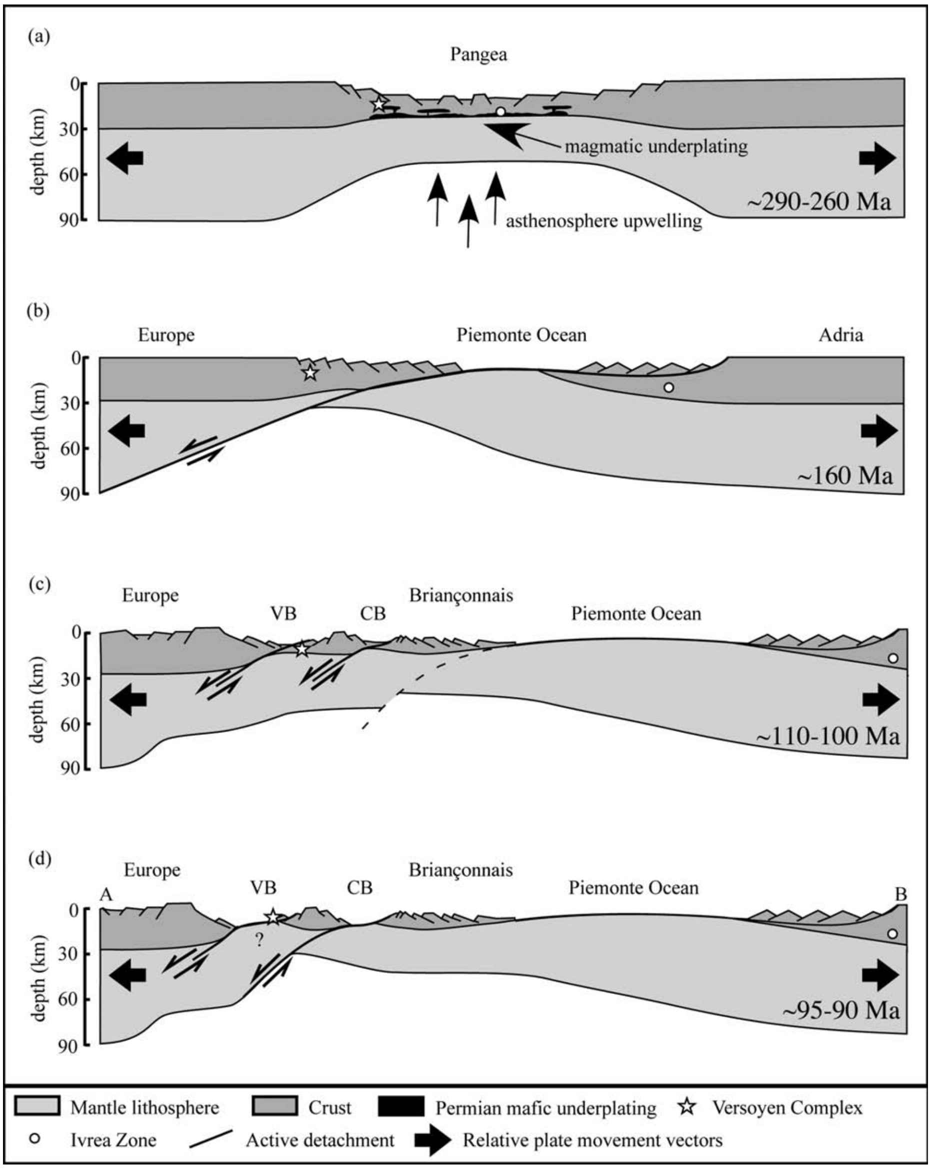

Fig. 9 - Schematic profiles from the European to the Adria Plate (NW to SE) depicting the evolution of the Central Mediterranean area from the Permian to the Cretaceous. (a) The magmatic rocks of the Versoyen Unit (star) intruded the basement in the Permian (ca. 290-260 Ma), at the same time as the Layered Gabbro Complex of the Ivrea Zone (circle, here used for reference only) probably in an intra-plate extensional setting resulting from the extensional collapse of the Hercynian orogen. In the Late Jurassic (b) the Versoyen Unit became part of the European Plate as a consequence of the opening of the Piemonte Ocean (geometry of the extensional system from Lemoine et al., 1987). During the Albian/Aptian (c) the Valaisan basin (VB) and the Chiavenna/Balma basin (CB) formed in an intra-continental transtensional tectonic regime. By that time opening of the Piemonte Ocean had ceased and thermal relaxation probably resulted in partial re-thickening of the lithosphere. (d) In the Middle Cretaceous (ca. 95-90 Ma), continuing extension may have resulted in the rupture of the continental crust and in the formation of oceanic crust in the CB. As a result, a ribbon of continental lithosphere, probably hosting the Dora Maira massif and the Adula Unit became isolated both from the European Plate *sensu stricto* and from the Briançonnais terrain. Notice that the dip of the detachments in (c) and (d) is purely speculative. Question mark in the Valaisan basin refers to the formation of oceanic crust.

(Liati et al., 2005) has shown that the intrusion of the mafic sills in the sediments of the Misox zone (Fig. 2) took place in the Jurassic and was related to the opening of the Piemonte Ocean.

Therefore, the Chiavenna Unit and the Balma Unit (Fig. 2) are the only two Alpine terranes that recorded basic magmatism and possible oceanic crust formation in the Cretaceous (Liati et al., 2003). Liati et al. (2003) used those results to suggest that extensional tectonics led to lithospheric thinning and ocean spreading in the Valaisan Basin. However, the attribution of the Chiavenna ophiolite and the Balma Unit to the Valaisan Basin may be challenged on several grounds. Indeed no lithological equivalent of the Chiavenna Unit and the Balma Unit is found in the Valaisan Domain *sensu stricto*. The Chiavenna Unit is composed of serpentinites, carbonates and metabasalts (Schmutz, 1976; Huber and Marquer, 1998; Talerico, 2001) and the Balma Unit of serpentinites, eclogites and amphibolites (Pleuger et al., 2005).

Furthermore, as shown in Fig. 10, the Valaisan Domain *sensu stricto* occupies a different structural position in the Alpine orogen with respect to the Balma Unit and the Chiavenna Unit. This difference is especially evident in the Western Alps (Fig. 10a). There, the Valaisan Domain *sensu stricto* crops out between the Briançonnais Domain and Helvetic Domain, which is part of the European Plate. The Balma Unit, instead, is overlain by the ophiolites of the Piemont Ocean and by a discontinuous layer of gneissic rocks, which have been attributed to the Briançonnais Domain (Pleuger et al., 2005), and underlain by the Monte Rosa Massif. The palaeogeographic attribution of the latter is uncertain (see Froitzheim, 2001, for a discussion), as it has been considered either as a part of the Briançonnais Domain (Escher et al., 1997), or as an independent terrane (Platt, 1986) or as a promontory of the European (Froitzheim, 2001) or African (Stampfli et al., 1998) Plates. No physical continuity exists between the Chiavenna ophiolites and the Valaisan Domain *sensu stricto* in the Central Alps, either (Fig. 10b). There, the Valaisan Domain crops out again at the interface between the Briançonnais Domain and the Helvetic Domain. The Chiavenna ophiolite, instead, is now sandwiched between the Briançonnais Domain and the migmatitic Gruf Unit (Huber and Marquer, 1998; Talerico, 2001). The latter is normally regarded as the south-eastern continuation of the Penninic Adula nappe (e.g. Frey and Ferreiro-Mählmann, 1999).

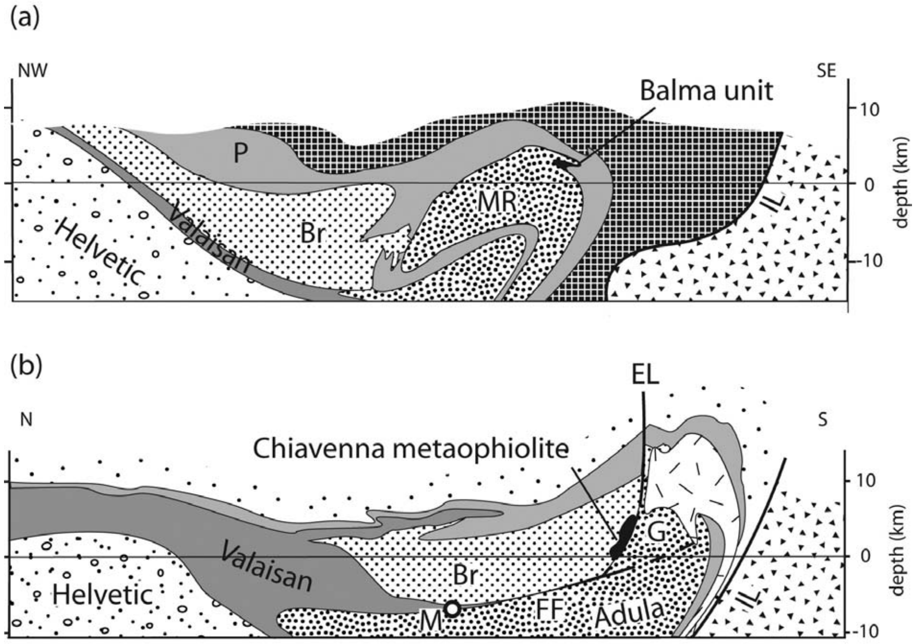

Fig. 10 - Schematic cross sections through (a) the Western Alps (modified after Schmid et al., 2004 and Pleuger et al., 2005) and (b) the Central Alps (modified after Huber and Marquer, 1998 and Schmid et al., 2004). Traces of cross sections are indicated in Fig. 2 (see Fig. 2 for legend). Br- Briançonnais domain, P- Piemonte ophiolites; G- Gruf Complex; MR- Monte Rosa; IL- Insubric Line, EL- Engadine Line; FF- Forcola Fault; M- Misox zone.

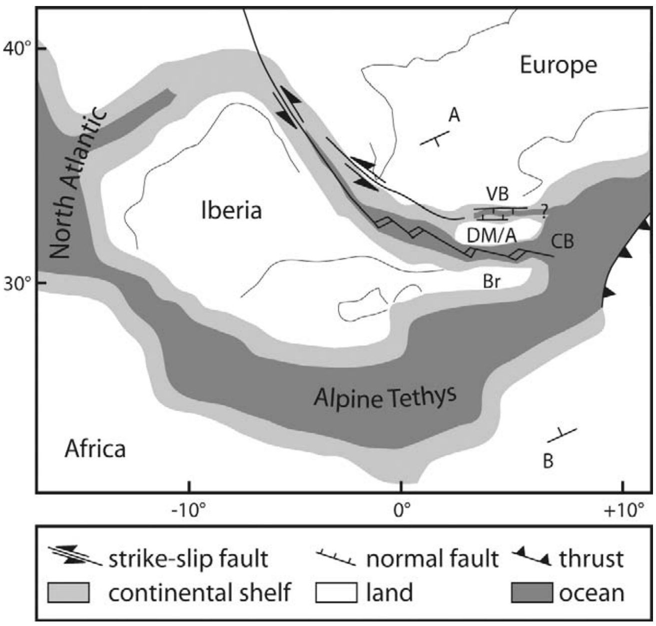

Fig. 11 - Simplified tectonic map of the north-western Tethys area in the Cretaceous. The Valaisan Basin (VB) and the Chiavenna/Balma Basin formed as a result of transtensional tectonics induced by the eastward drift of Iberia at ca. 110-90 Ma. Continued extension may have resulted in formation of oceanic crust in the Chiavenna/Balma Domain (CB) at ca. 95-90 Ma. The two basins isolated a continental ribbon containing the rock units that constitute the Dora Maira Massif and the Adula Unit (DM/A). The position of the cross section depicted in Fig. 9d is also indicated.

If the current position of rock units in the Alpine orogen is taken as an indication of the stacking order, which in turn reflects the palaeogeographic distribution of the different terranes, then the Chiavenna and Balma Units were more likely located in a much more internal position than the Valaisan Basin. This suggestion is supported by the recent estimates of the age of high-pressure metamorphism in the Monte Rosa Massif, where eclogite formation has been dated at 43-44 Ma (Lapen et al., 2007), at a time when sedimentation was still occurring in the Valaisan Basin *sensu stricto* (Stampfli et al., 1998). Therefore, the continental rock units now interposed between the Valaisan Domain *sensu stricto* and the Chiavenna/Balma Unit were probably part of a continental ribbon located in the Western Tethys between the two neighbouring basins (Figs. 9d and 11). This observation introduces the possibility that the Lower Cretaceous basic magmatism recorded in the Balma Unit and Chiavenna Unit took place in a different basin from the Valaisan Domain (Figs. 9 c, d and 11).

From the Late Jurassic to the Cretaceous, the tectonic evolution of the Alpine Tethys and the area located to the north of it was characterized by the opening of the northern part of the Central Atlantic (Fig. 1c; Stampfli, 1993). This resulted in the eastward drift of the Iberian Plate and its eastern continuation, the Briançonnais Domain. Stampfli (1993) proposed that the protracted relative movement between Iberia and the European Plates *sensu stricto* resulted in lithospheric stretching and, at least locally, in sea-floor spreading (Fig. 1c). Kinematic models (Srivastava et al., 1990; Malod, 1990; Sibuet and Colette, 1991; Srivastava and Verhoef, 1992) constrain ocean spreading to the 116-91 Ma time span. These results are in accordance with the estimated ages for the thermal event linked to the rising asthenosphere in the Pyrenean region (Peybernes and Souquet, 1984; Montigny et al., 1986), and also with the ages of sedimentation in the Valaisan Basin and basic magmatism in the Chiavenna/Balma basin/ocean (Liati et al., 2002 and 2003). Therefore, extensional tectonics were presumably active at the same time in both the Valaisan and Chiavenna/Balma areas as a consequence of the eastward drift of the Iberia Plate.

We suggest that protracted movement led to complete laceration of continental lithosphere and formation of oceanic crust in the Chiavenna/Balma basin/ocean in the Cretaceous. The Valaisan and Chiavenna/Balma Basins would have been separated by a continental ribbon with European affinity comprising rock units that during the Alpine orogenesis made up the Dora Maira Massif and the Penninic Adula Nappe, in the Central Alps. Figs. 9d and 11 contain the proposed palaeogeographic scenario for the Western and Central Alps in the Late Cretaceous. The Briançonnais terrain was located in a more internal position than the Chiavenna/Balma basin/ocean. Further to the northwest the Valaisan Basin separated the Dora Maira/Adula ribbon from the European Plate *sensu stricto*.

The Valais and Chiavenna/Balma extensional basins possibly merged together to the west into a single basin (the Pyrenean Ocean of Stampfli, 1993) that extended all the way to the Bay of Biscay (Fig. 11). To the east the two basins were probably branching off the Piemonte Ocean (Fig. 11), in a similar way to that envisaged by Trümpy (1980). Crustal extension in the Valaisan Basin may have been greater in the Central Alps, where it led to the exposure of serpentinites at the surface (Florineth and Froitzheim, 1994), than in the Western Alps.

CONCLUSION

The Valaisan Domain of the Western and Central Alps is commonly believed to represent the remnant of an Upper Jurassic-Cretaceous oceanic basin. However, the Punta Rossa metagranitoid and a metagabbro from a layered gabbro complex of the Versoyen Unit, which is part of the Valaisan Domain, contain magmatic zircon dated at 267±1 and 272±2, respectively. We interpret this age as crystallization ages and propose that magmatism in the Versoyen Unit was unrelated to the Cretaceous formation of the Valaisan Basin. During the Cretaceous, the Versoyen rocks likely underwent some degree of fluid metasomatism leading to rare zircon rims. These results, together with considerations based on existing literature, point towards a new scenario for the plate configuration along the southern margin of the European Plate. We propose that the Valaisan Basin and a more internal basin (the Chiavenna/Balma basin/ocean) formed independently as a result of regional transtension induced by the eastward movement of Iberia with respect to the European Plate. The geometry of extension probably involved anastomosing regional faults. Their activity resulted in the isolation of continental domains between intervening Mesozoic basins.

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