U–Pb zircon ages for post-Variscan volcanism in the Ligurian Alps (Northern Italy)

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Abstract: Ages of Permian volcanic rocks from the Ligurian Briançonnais domain (Western Italian Alps) have been determined by laser ablation inductively coupled plasma mass spectrometry U–Pb dating of zircon. Three major volcanic units yielded zircons that were dated: calc-alkaline rhyolites (285.6 ± 2.6 Ma), andesites (with inherited cores yielded ages c. 476 Ma and older) and voluminous rhyodacites–rhyolites (272.7 ± 2.2 Ma). Following an amagmatic, sediment-starved time gap of c. 14 Ma, alkaline volcanic activity is recorded, at the top of the sequence, by K-alkaline rhyolites dated at 258.5 ± 2.8 Ma. The Ligurian segment of the Southern Variscan belt records transtensional and then extensional tectonics associated with the volcanic activity. The switch from calc-alkaline to alkaline activity corresponds to the transition from a post-orogenic to an anorogenic setting in the Southern Variscides; it may represent progressive and increasing delamination of the continental lithosphere, accompanied by partial melting of the lithospheric mantle.

Supplementary material: Technical and data acquisition parameters, U–Th–Pb isotope analyses and calculated ages of zircons from samples and standard, and trace element compositions of selected zircons are available at: http://www.geolsoc.org.uk/SUP18322.

The Variscan belt of Western Europe was created by the polyphase collision of Laurussia, Gondwana and Gondwana-derived microplates (Hun Terranes), during the suturing of Early Palaeozoic oceanic basins (Stampfli et al. 2002; Von Raumer et al. 2003). The Palaeozoic record of the southernmost segment of the Hun Terranes, involved in the Variscan orogenesis from Late Devonian times, is now preserved in the Ligurian Alps, Provence, Sardinia–Corsica and the Southern Alps. The Variscan terranes consist of: (1) pre-Variscan basements characterized by medium- to high-grade metamorphism related to the collisional stage and synorogenic intrusions; (2) late Variscan volcanic-sedimentary sequences and intrusions.

The Late Palaeozoic volcano-sedimentary stratigraphic record of the Ligurian Alps (Cabella et al. 1988; Cortesogno et al. 1993, Figs 1 and 2), reveals that in this region volcanic rocks were predominant over intrusive and sedimentary products, in contrast to the intrusion-dominated Sardinia–Corsica and Helvetic domains (Schaltegger 1997; Cortesogno et al. 1998; Paquette et al. 2003; Cocherie et al. 2005), and also in contrast to the Southern Alps, where sedimentation in rapidly subsiding intramontane basins overwhelmed the volcanic products (Cassinis et al. 2003).

At a regional scale, bimodal (rhyolitic, andesitic and rhyodacitic–dacitic) calc-alkaline igneous activity ended in the Early Permian, with felsic and transitional basaltic activity extending into the Late Permian. The post-Variscan alkaline magmatism in Sardinia and Corsica dates to the Early Triassic (Deroin & Bonin 2003; Cocherie et al. 2005); the Ligurian sequence ended with a high-K rhyolitic event that is not yet dated.

The major features of the Variscan belt are relatively well known, but discontinuous. Here, we present new radiometric age determinations that significantly advance our understanding of the late- to post-orogenic volcano-sedimentary processes of the Ligurian Alps.

Until now, only unconfirmed palaeontological data have been used to assess the timing of post-Variscan magmatism in the Ligurian Alps. Precise geochronology is fundamental to constraining the evolution of this mostly volcanic suite. A new palaeogeographical reconstruction based on palaeomagnetic data (Muttoni et al. 2003) ascribes the Palaeozoic terranes of the Ligurian Alps during the Permian to the boundary of Gondwana-land and Laurussia, as previously envisaged by Stampfli (1996). This results in a new perspective on the key role played by the Ligurian Permian successions in our understanding of the post-collisional evolution leading to the break-up of Pangaea.

We determined spot U–Pb ages and trace-element compositions of zircon separates by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) to characterize the successive steps of post-orogenic Variscan history and to make correlations with the adjacent areas.

Geological background

In the Ligurian Alps (Fig. 1), Late Palaeozoic volcano-sedimentary successions are preserved within all Alpine tectonic units of the Briançonnais and Pre-Piedmont domains, which represent respectively the European lithosphere and its transition to the neo-Tethyan Ocean during Jurassic times (Vanossi et al. 1986). Whereas the Pre-Piedmont volcano-sedimentary sequences are generally detached (Dallagiovanna 1988), the Briançonnais domains rest unconformably on the pre-Variscan basement (Cortesogno et al. 1993) and are truncated by a Late Permian unconformity.

The Early Palaeozoic basement consists of two lithological...
complexes: (1) Cambro-Ordovician paragneisses, micaschists, metarhyolites, metagranitoids, migmatites, eclogites and amphibolites, characterized by polyphase metamorphic evolution and protolith ages for the felsic rocks of $507^{12/8}$ and $494^{12/5}$ Ma (sensitive high-resolution ion microprobe (SHRIMP) U–Pb data, Gaggero et al. 2004); (2) Mid-Ordovician metagranitoids (protolith ages $473^{11/1}$, $468^{11/1}$ and $459.4^{2/9}$ Ma, Gaggero et al. 2004), metagabbros (protolith age of $469^{6/6}$ Ma, Gaggero et al. 2004) and eclogites (protolith ages of $468^{4/1}$ and $446^{2/2}$ Ma, LA-ICP-MS U–Pb method, Giacomini et al. 2007) affected by a single phase of pre-Alpine schistosity.

U–Pb dating (Giacomini et al. 2007) of the metamorphic recrystallization of zircons from the Mid-Ordovician complex yielded ages from $392^{13/7}$ to $374^{18/8}$ Ma, probably corresponding to eclogite-facies conditions during the Eo-Variscan collisional stage, and ages between $346$ and $320$ Ma, associated with the amphibolite-facies overprint during uplift. A zircon rim yielded a concordant age of $305^{6/6}$ Ma related to the final stages of exhumation (lower amphibolite to greenschist facies), consistent with Rb/Sr dating of white mica and biotite (ages between $327$ and $297$ Ma) by Del Moro et al. (1981). Also, $^{40}$Ar–$^{39}$Ar ages between $313$ and $302$ Ma were obtained for white micas separated from the paragneisses and orthogneisses of the basement (Barbieri et al. 2003).

The igneous activity

In spite of Alpine nappe-stacking (Seno et al. 2005, Fig. 1), the stratigraphic setting of the Late Palaeozoic succession records the development of volcanic and sedimentary products from the innermost to the outermost zones, but with highly variable relative thickness caused by irregular palaeomorphology, characterized by grabens and half-grabens. The geochemistry of the volcanic rocks was characterized by Cabella et al. (1988) and Cortesogno et al. (1992, 1998).

**The early episodes: the basal ignimbrite (Case Lisetto Metarhyolites) and the andesite (Eze Fm.)**

The volcano-sedimentary sequence begins with fine to coarse arkosic and quartzitic continental deposits, formed by reworking of the underlying orthogneissic basement, or with ignimbritic rhyolites with minor rhyodacites (Case Lisetto Metarhyolites, Fig. 2). The ignimbrite is characterized by coarse ($1–3$ cm) K-feldspar and quartz phenocrystals (porphyritic index (PI) $30–40$) with plagioclase ($0–10$ mm) and biotite ($0.5–10$ mm). There are rare xenoliths of quartzites, garnet micaschists and muscovite–biotite metapelites.

Above, conglomeratic–sandy–pelitic, fluvial–lacustrine deposits occur, hosting flora ascribed to the Late Westphalian–Stephanian interval (Bloch 1966). The early calc-alkaline rhyolite also includes pyroclastic agglomerates, and another ignimbrite (Osiglia Porphyries), dated at $278^{3.4}$ Ma (SHRIMP U–Pb data, Molina 2002), is intercalated with the continental sediments. In some places, fine-grained sediments contain intercalations of black phyllites and light-coloured metapelites.

Andesite lavas (up to $200$ m in thickness) and fine to very fine (rarely very coarse) tuffs with sub-alkalic to calc-alkaline
The dykes, flows and tuffs of the Eze Fm. are andesites to lati-andesites, with an aphanitic to cryptocrystalline matrix, sometimes a fluidal texture, and, locally, high vesicularity. The PI ranges from 20–30 to 60. Phenocrysts include zoned plagioclase, which is sometimes glomeroporphyritic, and, more rarely, hornblende and biotite, showing dehydration reactions. Millimetre-sized euhedral to skeletal ilmenite and acicular to prismatic apatite are abundant. Coarse, rounded zircon is scarce. Rounded xenoliths of granoblastic quartz occasionally occur. Magnetite and ilmenite are probable secondary phases. Carbonatation is diffuse and results in the development of calcite and ankerite.

Small intrusive bodies (probable sills, domes and dykes of the Borda Granodiorites) occur in the inner Briançonnais Alpine units. Their geochemistry suggests a similarity to the Case Lisetto Metarhyolites; the SHRIMP U–Pb dates indicate an intrusion age between 300 and 294 Ma (Gaggero et al. 2004).

**The main episode: the rhyolite ignimbrite (Lithozone C)**

This is by far the most voluminous unit, with an area of >1000 km² in the Briançonnais sectors (Melogno Porphyroids Fm.). This formation consists of calc–alkaline ignimbrites, divided into three lithozones. From bottom to top, Lithozone A is composed of thin and discontinuous layers of frequently silicified, fine-grained rhyolites. Lithozone B comprises up to 100 m of coarse- to very coarse-grained rhyolite–rhyodacite ignimbrites, and locally rhyodacite or dacite breccias and granophyres. Lithozone C is the most common facies and consists of up to several hundred metres of mainly rhyolitic, fine- to medium-grained ignimbrites. The various volcanic facies were distinguished as single episodes within the major acidic episode of Lithozone C. As a whole, the rhyolitic to rhyodacitic ignimbrites show fiamme structures and quartz- and mica-rich pockets, probably derived from reworking or intense hydrother-
mal alteration. Quartz, K-feldspar and plagioclase phenocrystals occur in variable quantities, ranging from fine to medium grain size (0.5–10 mm). Biotite shows dehydration textures as a result of its breakdown during ascent of the magma. Ilmenite, rutile and apatite are common accessory phases. The abundant glassy groundmass has been transformed to phengite and often preserves pumice and fiamme structures. Scant gneissic xenoliths are included.

Up to some hundreds of metres of fine to very coarse pyroclastic fall deposits, with minor dacite ignimbrites, cover Lithozone C in the outer Briançonnais Alpine units.

The late episode: the K-rich rhyolite ignimbrite (Lithozone D)

The Latest Palaeozoic igneous activity is represented by an ignimbrite event (Lithozone D) of K-alkaline affinity. It consists of about 200 m of massive, welded, varicoloured rhyolites with high modal K-feldspar within a glassy groundmass. Several ignimbrite eruptions formed the latest K-rich episode. Within these rocks, differences in thickness, modal phenocrysts and groundmass indicate different volcanic facies. From the bottom up, these are as follows.

The blackish lower Dn facies is subordinate in volume, discontinuous and does not exceed 20 m thickness. It consists of rhyolites with centimetre-sized phenocrysts of K-feldspar, greatly exceeding the modal quartz (PI 30–40). Biotite is scarce, whereas ilmenite, apatite and zircon are common. The groundmass consists of vitrophylic lava fragments, granophyric ortholiths and pumices. Secondary mineralization (carbonates, silex, pyrite, tourmaline, hematite) is developed.

The Dr reddish facies corresponds to welded rhyolites with eutaxitic textures, and is up to 100 m in thickness. The PI is ≃30; the phenocrysts range in size from 1 to 20 mm; K-feldspar exceeds quartz, and biotite is scarce. Minor zircons occur as an accessory phase. Vitrophyric to microcrystalline lava fragments occur in the groundmass, which also contains hematite and local pyrite concentrations.

The Dv greenish facies is pervasively schistose and the ignimbritic textures are hardly recognizable; it is up to 60 m thick, and in some places is eroded by the overlying Verrucano conglomerates. Fine- to medium-grained (0.5–5 mm) K-feldspar phenocrysts predominate over quartz (PI 40–80). Biotite is rare, but there is abundant ilmenite, zircon and apatite. The groundmass is green owing to the presence of secondary phengite. Ankerite nodules, probably corresponding to present-day caliches, probably mark pristine exposed surfaces within the ignimbrite apron.

The continental volcano-sedimentary sequence is truncated by an erosional surface at the base of the Uppermost Permian Verrucano conglomerates grading to Triassic quartzites.

Geochronology

Methods

Zircons for U–Pb geochronology were first separated using standard magnetic techniques and heavy liquids. Zircon grains without fractures and as free from inclusions as possible were selected by hand-picking, embedded in epoxy resin and polished to 0.25 μm using diamond paste. Prior to age determination, the internal structure of the zircons was investigated with back-scattered electron (BSE) microscopy and cathodoluminescence (CL) with a Philips XL30 electron microscope, at the Earth Science Department, Siena University, Italy. The in situ U–Pb geochronology and trace element abundances were determined with excimer laser ablation (ELA)-ICP-MS at CNR–Istituto di Geoscienze e Georisorse (IGG), Unità di Pavia. The laser ablation instrument consists of an ArF excimer laser microprobe at 193 nm (Geolas2000-Microlas) and a high-resolution sector-field ICP-MS system (Element I from ThermoFinnigan). The analytical method followed that described by Tiepolo (2003). The instrumental and laser-induced U–Pb fractionations were corrected adopting the 1065 Ma 91500 zircon (Widenbeck et al. 1995) as an external standard. The spot size was set to 20 or 10 μm and laser fluence to 12 J cm⁻². Data reduction was performed using the Glitter software package (van Achterbergh et al. 2001). Time-resolved isotope ratios were carefully inspected to detect perturbations related to inclusions, fractures or mixing of different age domains. In all these cases, ablation intervals were set to represent only the initial unaltered portion of the signal. The reproducibility of the standards was propagated to all determinations according to the equation of Horstwood et al. (2003); after this procedure, analyses are considered accurate within quoted errors. The 295 Ma zircon 02123 (Ketchum et al. 1995) was, however, analysed in every analytical run together with unknowns to monitor the instrumental accuracy. The mean accuracy can be evaluated at about 1.7%. Isoplot/EX 3.0 software (Ludwig 2000) was used to calculate the concordia ages and draw concordia plots. When data points refer to the measurement of the same population, the weighted-mean concordia age was considered and the goodness of fit is reported as mean square weighted deviation (MSWD). The main parameter that distinguished the populations was the CL property of the zircon domains. All errors are given at the 2σ level.

Zircon features

The U–Pb analysis was carried out on zircons separated from two samples of Case Lisetto Metarhyolites (samples LIS1, LIS2), one from Eze Fm. (EZV), one from Lithozone C (LIC), and three from Lithozone D (LIDn, LIDv, LIDr) (Figs 1 and 2).

The zircons from the rhyolite–rhyodacite ignimbrites range in length from 100 to 280 μm and are subhedral to euhedral. Most of the zircons are short and prismatic to acicular; width-to-length ratios are about 1.2–1.5, but sub-rounded grains also occasionally occur. Prismatic to acicular morphology is commonly related to rapid crystallization, typical of volcanic rocks or high-level intrusions (Hoskin & Schaltegger 2003). Apatite and fluid inclusions are relatively common. Zircons in andesites are rare and relatively small, never exceeding 80 μm in length.

The BSE and CL images (Fig. 3) reveal a complex internal structure. Most of the crystals are composite, with low-luminescence xenocrystic cores surrounded by overgrowths. The cores may be zoned (oscillatory or patchy, especially in Lithozone D) or, more frequently, structureless. The overgrowths are variable in thickness and show well-developed oscillatory zoning (or rarely sector zoning). These features are typical of zircon growth under magmatic conditions (Corfu et al. 2003). Irregular surfaces, truncating the internal zoning, commonly separate the xenocrystic cores from the overgrowths. Occasionally, homogeneous high-luminescence patches, obscuring the primary texture, extend from the rim inwards towards the core of the zircon. Rarely, faint oscillatory zoning can be distinguished, suggesting that the zircon underwent low-temperature processes (Corfu et al. 2003). Analyses focused on the magmatic textures, because this study was concerned with the magmatic evolution.
**U–Pb results**

The early episodes: the basal ignimbrite (Case Lisetto Metarhyolites). Laser ablation was carried out at 47 spots on 42 zircon grains from samples LIS1 and LIS2, and 24 U–Pb concordant ages were obtained. The population primarily selected for the analyses was the magmatic overgrowths with oscillatory zoning. Laser ablation was also carried out at a few spots on the xenocrystic cores and in the high-luminescence patches. Three analyses of the inherited cores yielded ages of 289 ± 12, 296 ± 11 and 305 ± 11 Ma. Fifteen ages from the overgrowths with oscillatory zoning range from 273 ± 11 to 293 ± 11 Ma, with a weighted-mean concordia age of 285.6 ± 2.6 Ma (MSWD = 1.05, P = 0.31; Fig. 5a), and a single analysis yielded a younger age of 263 ± 11 Ma. Five spots on the high-luminescence patches gave scattered younger ages at 207 ± 8, 242 ± 8, 251 ± 11, 256 ± 11 and 261 ± 11 Ma.

Six trace-element analyses were also performed on the zircons (Fig. 4). The chondrite-normalized REE pattern was characterized by strong fractionation between heavy REE (HREE) and light REE (LREE), with a positive Ce anomaly and an occasional negative Eu anomaly. This trace element signature is typical of zircon growth under igneous conditions (e.g. Rubatto & Gebauer 2000; Rubatto 2002; Hoskin & Schaltegger 2003). We did not find significant compositional differences between the oscillatory-zoned overgrowths and the high-luminescence patches, nor unequivocal trends in the U and Th contents or the Th/U ratio. We interpreted the 285.6 ± 2.6 Ma concordant age for the population with oscillatory-zoned overgrowths as the timing of zircon crystallization in the magma. The isolated ages from xenocrystic cores of 289 ± 12, 296 ± 11 and 305 ± 11 Ma suggest the recycling of slightly older rocks (e.g. Borda granodiorites, c. 294–300 Ma, Gaggero et al. 2004). The dispersion of
data in the high-luminescence patches towards younger ages suggests multiple resetting of the U–Pb system, but the texture yields little information on the magmatic evolution.

The early episodes: the andesite (Eze Fm.). The rarity and small size of zircons from sample EZE prevented us from obtaining a good geochronological analysis of this formation. Twelve LA spots on seven zircon grains yielded seven concordant ages between 476 ± 15 and 769 ± 21 Ma, strongly suggesting the presence of xenocrystic zircon fragments.

The main episode: the upper rhyolitic ignimbrite (Lithozone C). Twenty-five LA spots on 22 zircon grains from sample LIC gave 20 U–Pb concordant ages. Two low-luminescence xenocrystic cores yielded ages of 809 ± 21 and 655 ± 17 Ma. Fourteen analyses performed on the magmatic overgrowths with oscillatory zoning gave ages ranging from 263 ± 8 to 285 ± 8 Ma. This zircon population yield a weighted-mean concordia age of 272.7 ± 2.2 Ma (MSWD = 0.98, P = 0.32; Fig. 5b). One fractured zircon, which gave a significantly younger age at 250 ± 7 Ma, was not considered in the discussion. Three LA spots on the high-luminescence patches yielded ages at 247 ± 7, 247 ± 9 and 252 ± 7 Ma.

Three trace-element analyses were performed on the overgrowths with oscillatory zoning and on the high-luminescence patches (Fig. 4). Trace-element patterns were very similar and are comparable with those of zircons in the Case Lisetto Metarhyolites. This suggests zircon growth under igneous conditions. The mean concordia age at 272.7 ± 2.2 Ma is thus the best estimate of the timing of zircon crystallization in the magma. The older ages of the cores suggest the recycling of zircon from older rocks. It should be noted that the magma rose through Ordovician basement rocks (Gaggero et al. 2004). Likewise, in the Case Lisetto Metarhyolites, the younger ages obtained from the high-luminescence patches were interpreted as related to U–Pb system resetting.

The late episode: the K-rich rhyolitic ignimbrite (Lithozone D). In samples LIDn and LIDv from high-K rhyolites, 23 U–Pb concordant ages were obtained from 49 LA spots on 22 zircon grains. One xenocrystic core gave an age of 396 ± 12 Ma. The other analysed cores (11 analyses) range from 267 ± 9 to 282 ± 8 Ma. This zircon population yielded a weighted-mean concordia age of 274.1 ± 2.6 Ma (MSWD = 1.4, P = 0.25; Fig. 5c). One zircon fragment gave a significantly younger age at 234 ± 8 Ma and was not considered in the discussion. Ten analyses were performed on the magmatic overgrowths with oscillatory zoning. Age results on this population range from 262 ± 8 to 254 ± 7 Ma with a weighted-mean concordia age of 258.5 ± 2.8 Ma (MSWD = 1.6, P = 0.21; Fig. 5d). This age probably dates the zircon crystallization during the igneous event of Lithozone D, as confirmed by trace-element analyses on selected zircons (Fig. 4). The older ages of the cores suggest recycling of zircon from the basement (e.g. 396 ± 12 Ma) and older volcanic rocks (e.g. Lithozone D). The ignimbritic character of the products, use of the same faults during emplacement and storage in adjacent magmatic reservoirs probably accounts for the recycling of older igneous material. In this sample, the late-stage patches were not analysed.

Tectonic inferences from igneous events: towards a geodynamic model

The geochronological data, from both this work and the literature, suggest an Early Permian age for the calc-alkaline magmatism (Borda Granodiorites c. 300–294 Ma; Case Lisetto Metarhyolites 285.6 ± 2.6 Ma; Osiglia Porphyries 278 ± 3.4 Ma; Lithozone C 272.7 ± 2.2 Ma) and a Late Permian age for the alkaline Lithozone D (258.5 ± 2.8 Ma). There is a temporal gap of c. 14 Ma between the emplacement of the lithozones C and D that lacks a magmatic or sedimentary record.

On a global scale, a comparison between the emplacement ages of the volcano-sedimentary succession of the Ligurian Alps and the post-Variscan igneous activity in the adjacent domains (Table 1) shows the evolution of the late- to post-collisional stages of the Southern Variscan belt. On the whole, the evolution was not continuous through time, but was stepwise.

Early Carboniferous collision

The Variscan phase (Fig. 6) corresponds to the period of metamorphism from amphibolite to greenschist facies, dated, in
| Region | Early Carboniferous | Late Carboniferous | Early Permian | Mid- to Late Permian |
|--------|--------------------|-------------------|--------------|---------------------|
| Sardinia (a), Corsica (b), Provence (c), Calabria (d), Kabylies (e) | | | | |
| | c. 350–330: granulite facies (a) (1 Am) | c. 324–317, 308–300: lower amphibolite facies, W Maures, E Maures (c) (9 Am–b) | 298 ± 4: SOS Canales pluton (a) (1 Rw) | 259 ± 3: CA dykes (a) (26 Rb) |
| | c. 346: Al-rich granitoids (b) (2 Z3) | c. 320–315: amphibolite facies (a) (1 Am) | 297 ± 5: Camarat pluton (c) (9 Rw) | 254 ± 5, 248 ± 9: basaltic dykes (a) (4 Ab) |
| | | c. 345 ± 3: Middle Bornes Unit (c) (28 Z3) | c. 313–304: Tanneron massif orthogneiss (c) (24 M) | |
| | | c. 338 ± 8: Hermitan granites (c) (28 Z3) | c. 338: Al-rich granitoids (b) (2 Z3) | |
| | | | 334 ± 3, 334 ± 10 to 324 ± 5: Reverdit tonalite, Plan de la Tour granites (c) (28 Z3, 9–28 Rw–Z3) | |
| | | | c. 331–325, 320–306: upper amphibolite facies, W Maures, E Maures (c) (9 Aa) | |
| | | | | |
| External Massifs: Aar–Gotthard (a), Aiguilles, Rouges–Mont Blanc (b), Pelvoux–Belledonne (c), Argentera (d) | | | | |
| | 343 ± 11: Rochail pluton (c) (15 Z3) | c. 312–308: Schollenen–Dussel–Fruttstock–Voralp diorites + granites (a) (30 ZTA3) | 298 ± 2 to 294 ± 1: granites + granodiorites (a) (30 ZTA3) | 283 ± 279: A-type granites + mafic intrusion (b) (7 Z1) |
| | 334 ± 2.5: Giuv syenite–Punteglias diorites + granites (a) (30 ZTA3) | 307 ± 2: Vallorcine gabbros + granites + rhyolites (b) (16 Z3M) | 293–285: central granite (d) (14 Rw) | |
| | 304 ± 3: Mont Blanc granites (b) (16 Z3M) | 303 ± 4 to 299 ± 2: rhyolitic tuffs (a) (31 Z3) | | |
| | 332 ± 2: Gessois della Barr monzonites (d) (17 Z2-3) | 302 ± 2: Turbat granites (c) (15 Z3) | | |
| | 333 ± 2: Pormenaz monzonite (b) (19 Z3M) | | | |
| | 324 ± 12: amphibolite facies (b, c, d) (29 KAA, 17 Z2–3) | | | |
| Penninic zone | 310 ± 2: Mt Mort paragneiss (18 M) | 304 ± 3: Cavour leucograniotes (37 Z3) | 290 ± 2: Malanaggio diorite (37 Z3) | |
| | 323 ± 8: Costa Citrin granites (12 Z3) | | c. 290–270: Sanfront granites (37 Z3) | |
| | | | c. 283–268: Freidour granite (37 Z3) | |
| | | | c. 282–267: Goli d’Aget rhyolites (18, 38 Z3) | |
| | | | c. 279–269 Sangone granite (37 Z3) | |
| | | | c. 272 ± 4: Mattmark gneiss (23 Z2) | |
| Ligurian Briançonnais | 333 ± 7: amphibolite facies (10 Z1) | c. 310–300: greenschist facies (10 Z1) | 269 ± 2: Randa orthogneiss (18 Z3) | |
| | | | 268 ± 1: Truzzo granites, Roffna rhyolites (33 Z3) | |
| | | c. 300–294: Borda granodiorites (11 Z2) | | |
| | | c. 285.6 ± 1.3: Case Lisetto Metarhyolites | | |
| | | 278 ± 3.4: Osiglia Porphyrites (22 Z2) | | |
| | | 272.7 ± 1.1: Lithozone C | | |
| | | | | |

**Table 1.** Geochronological data for the Late Palaeozoic igneous rocks and metamorphism of Southern Variscan belt (all ages in Ma)
Table 1. (continued)

|                | Early Carboniferous | Late Carboniferous | Early Permian | Mid- to Late Permian |
|----------------|---------------------|--------------------|--------------|---------------------|
| Austroalpine (a), Southern Alps (b) | 300 ± 12: Sondalo troctolites (b) (34 S) | 289 ± 2, 288 ± 4: Arolla gneiss, Anzasca gabbro (a) (35 Z3) | 287 ± 2, 285 ± 4: Ivrea Mafic Cp (b) (21 Z3) | 260 ± 1: Mt Collon lamprophyric dykes (a) (21 Aa) |

Data from present work are shown in italics. Numbers in parentheses indicate literature data: (1) Di Vincenzo et al. (2004) and references within; (2) Paquette et al. (2003); (3) Cocherie et al. (2005); (4) Gaggero et al. (2007); (5) Del Moro et al. (1993); (6) Del Moro et al. (1996); (7) Renna et al. (2007); (8) Peucat et al. (1996); (9) Mordon et al. (2000); (10) Giacomini et al. (2007) and references within; (11) Gaggero et al. (2004); (12) Bertrand et al. (1998); (13) Ring et al. (2005); (14) Ferrara & Malaroda (1969); (15) Guerrot & Debon (2000); (16) Bussy et al. (2000); (17) Rubatto et al. (2001); (18) Bussy et al. (1996); (19) Bussy et al. (2001); (20) Schaltegger & Brack (2007); (21) Monjoie et al. (2007) and references within; (22) Molina (2002); (23) Liani et al. (2001); (24) Demoux et al. (2008); (25) Pittau et al. (2002); (26) Traversa et al. (2003); (27) Zheng et al. (1992); (28) Moussavou (1998); (29) Menot et al. (1987); (30) Schaltegger & Corfu (1992); (31) Schaltegger & Corfu (1995); (32) Barth et al. (1994); (33) Marques et al. (1998); (34) Tribuzio et al. (1999); (35) Bussy et al. (1998); (36) Hansmann et al. (2001); (37) Bussy & Cadoppi (1996); (38) Derron et al. (2006). Dating methods: A, 40Ar–39Ar (a, amphibole; b, biotite; f, feldspar; m, muscovite; p, plagioclase); AL, U–Th–Pb on allanite; KA, K/Ar (a, amphibole); M, U–(Th)–Pb on monazite; R, Rb–Sr (b, biotite; w, whole-rock); S, Sm/Nd; Z, U–Pb on zircon (Z1, LA-ICP-MS; Z2, SHRIMP; Z3, ID-TIMS); ZTA3, U–Pb on zircon–titanite–allanite. CA, Calc-alkaline; C0, Cordilleran; C1, Cordilleran, Late Devonian; C2, Cordilleran, Early Carboniferous; C3, Cordilleran, Late Carboniferous; C4, Cordilleran, Early Permian; C5, Cordilleran, Mid- to Late Permian.

Liguria, to c. 346–320 Ma (Giacomini et al. 2007). In Corsica–Sardinia, Provenç and the External Massifs, large volumes of igneous material, contemporaneous with Early Carboniferous tectonic shortening and thrusting (Carmignani et al. 1994; Carosi & Oggiano 2002; Fernandez et al. 2002; Di Vincenzo et al. 2004), were emplaced in several episodes between c. 346 and 324 ± 5 Ma. In the Massif Central, synorogenic igneous activity (c. 360–320 Ma), controlled by a local extensional regime, migrated from north (Late Devonian) to south (Early Carboniferous) (Echtler & Malavieille 1990; Faure 1995; Bellot & Roig 1994). The earliest volcanic episode overlying the basement can be dated to 293–308 Ma on metamorphic micas (Macera et al. 1998). In the External Massifs, between 280 and 284 Ma, large intrusive bodies were emplaced within strike-slip continental volcano-sedimentary basins. Also, in Corsica–Sardinia and in Provenç, in the Late Carboniferous–Early Permian.

Late Carboniferous exhumation

In the Ligurian Briançonnais basement, the age of the low-grade Variscan amphibolite- to greenschist-facies metamorphism, associated with the final stages of exhumation (Fig. 7), is set at 310–300 Ma (Giacomini et al. 2007, and references therein). As the first volcanic episode overlying the basement can be dated to 285.6 ± 1.3 Ma (Case Lisetto Metarhyolites), the average exhumation rate is c. 1 mm a⁻¹. The exhumation mechanism is debatable, as these velocities could be accounted for by either purely isostatic or tectonically assisted kinematics. On the other hand, basement relics are not affected, as in Sardinia, by an HT–LP silimanite–cordierite–K-feldspar-andalusite overprint, coeval with the extensional structure dated at 293–308 Ma on metamorphic micas (Macera et al. 1989). Between 310–300 and 286 Ma the magmatic activity was represented only by small intrusions cutting through the basement (Borda Granodiorites, c. 300–294 Ma). The early magmatic episode should be considered to have originated in the decompressional melting of lithospheric mantle contaminated by the lower crust (Cortesogno et al. 1998).

Elsewhere in the Penninic domain, Late Carboniferous magmatism is poorly represented, whereas the continental deposits, associated with strike-slip tectonics, are well developed, such as within the coeval sediments of the ‘Zone Houiller Briançonnaise’ (Fabre 1961; Brousniche-Delcambre et al. 1995; Bertrand et al. 1998). In the External Massifs, between c. 312 and 294 ± 1 Ma (Table 1), large intrusive bodies were emplaced within strike-slip continental volcano-sedimentary basins. Also, in Corsica–Sardini and in Provenç, in the Late Carboniferous–Early Permian.
period (c. 305–295 Ma, Table 1) large igneous and volcanic bodies were emplaced within narrow basins (Cassinis et al. 2003, and reference therein). In the Massif Central the last granitic bodies to ascend along strike-slip faults are dated to c. 315 Ma in the north and to c. 305 Ma in the south (Echtler & Malavieille 1990; Bellot 2007; Brichau et al. 2007; Gebelin et al. 2007).

On the whole, migration of the extensional collapse and magmatism towards the southern part of the Variscan chain can be envisaged as being due to a generalized strike-slip regime within the whole belt (Arthaud & Matte 1977; Schaltegger 1999; Bellot 2007; Brichau et al. 2007; Gebelin et al. 2007).

Early Permian wrench tectonics and volcanism

The age of the Case Lisetto Metarhyolites (285.6 ± 2.6 Ma) marks the beginning of the volcanic-sedimentary cycle of the Ligurian Alps. This age differs from the previous Late Carboniferous stratigraphic age for the overlying sediments (Bloch 1966); on the other hand, it agrees with the Permian age of the volcanic rocks above (Lithozone C), and with the inferred exhumation rate of the basement and with the age of the greenschist-facies overprint of the basement at c. 310–300 Ma. However, the Carboniferous flora has proved not reliable for dating purposes (e.g. Massif Central; Bruguier et al. 2003, and references within).

On the whole, the remnants of the basement and volcanosedimentary sequence that escaped the Alpine disruption have allowed the reconstruction of an east–west polarity that indicates an asymmetrical, tectonically controlled basement sink, as demonstrated by the thick volcanic-dominated successions to the (present-day) west and the scarce volcanic record to the (present-day) east (Cortesogno et al. 1992), accompanied by a westwards

Fig. 7. Possible Late Carboniferous configuration (a) (modified after Stampfl 1996; Muttoni et al. 2003; von Raumer et al. 2003) and geodynamic setting (b) of the Southern Variscan belt. During the post-collisional stage, thermal instability of the continental lithospheric mantle and the weak link with the crust caused delamination of the orogenized lithosphere. Upwelling of hot asthenosphere induced magmatic underplating and partial melting of lower and middle crust (Black & Légeois 1993). Variscan terranes: AA, Austroalpine; Ca, Calabria; Co, Corsica; HE, Helvetic; Kb, Kabylies; Ma, Maures; MC, Massif Central; PE, Penninic; Py, Pyrenees; SA, Southern Alps; Sd, Sardinia. LB indicates the position of the Ligurian Briançonnais.
K$_2$O increase in the andesites. In Liguria, as well as in other collapsing segments of the belt (Sardinia, Southern Alps), the development of intramontane lacustrine basins occurred in a strike-slip regime (Figs 8 and 9). The volcanism between 285.6 ± 2.6 and 272.7 ± 2.2 Ma accompanied the collapse and subsidence of the lithospheric blocks. Furthermore, extensional collapse and related magmatism moved from the internal to the external zones of the Variscan front, as a result of the migration of the compression.

In the southernmost Variscides (Provenç, Corsica, Sardinia, the Penninic and Austroalpine zones, and the Southern Alps), the Early Permian magmatism (Table 1) can be referred to a generalized lithospheric thinning associated with a large-scale mechanism of crustal wrenching (Fig. 9). In this regard, recent palaeomagnetic data (Muttoni et al. 2003) have made it possible to infer a displacement of c. 2000 km between the northern and southern lithospheric blocks, in a dextral megashear zone active mainly during the Middle Permian between Laurussia and Gondwanaland. Therefore, Early Permian wrench tectonics at the surface in the southernmost Variscan sector could correspond to the shear at depth. The Ligurian Briançonnais, together with Maures, Sardinia, Corsica, the Penninic and Austroalpine zones and the Southern Alps, were located along a major WSW–ENE lineament at the end of the Variscan collision (Fig. 8). In the Ligurian Alps, the parental melts of the dominantly calc-alkaline igneous products have been ascribed to multiple sources (Cortesogno et al. 1998). In particular, the basal rhyolites have been envisaged as an early anatectic product; conversely, the andesite event probably originated in partial melting at the mantle–crust boundary, as indicated by the occurrence of garnet xenocrysts and HREE fractionation (Buzzi & Gaggero 2007). The interaction between the lower crust and crustal magmas probably accounted for the later conspicuous rhyolitic, rhyodacitic and dacitic activity.

**Late Permian K-rich volcanism and regional extension**

Following a c. 14 Ma Mid-Permian interval, alkaline volcanic activity began at 258.5 ± 2.8 Ma (Lithozone D). It developed within horst and graben structures related to a purely extensional regime (Cabella et al. 1988; Cortesogno et al. 1998). The Late Permian age of the uppermost rhyolites and the c. 14 Ma gap allow us to associate this event with the anorogenic volcanic phase, characterized (e.g. in Corsica) by well-defined geochemical features, interpreted as originating from partial melting of the upper mantle in ultradepth magma chambers, evolving into felsic residual liquids (Bonin 2007).

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**Fig. 8.** Possible Early Permian configuration (a) (modified after Stampfli 1996; Muttoni et al. 2003; Von Raumer et al. 2003) and geodynamic setting (b) of the Southern Variscan belt. Lithosphere delamination and related magmatism occurred along the main transcurrent margin. (Abbreviations as in Fig. 7.)
The switch from calc-alkaline to alkaline activity is marked by a Mid-Permian amagmatic gap, unlike in Sardinia, Corsica and Provence, where a chronological overlap is observed. The sudden end of calc-alkaline magmatism and the consequent gap in the sedimentary record may correspond to the main phase of the general strike-slip tectonic event, which is well developed in the Ligurian Briançonnais and, as in other sectors of the southern Variscides, was not accompanied by sedimentation (Cassinis et al. 2003; Virgili et al. 2006; Schaltegger & Brack 2007). In contrast, the age of 258.5 ± 2.8 Ma for the alkaline volcanic rocks of the Ligurian Alps matches the end of the plate reorganization and the change of geodynamic context from a post-orogenic to an anorogenic setting, and indicates the beginning of the Alpine cycle (Fig. 10). The literature suggests that lithospheric delamination could occur either by the rise of hot materials at mantle plumes, or, more suitable for our case, by strike-slip movements along shear zones (e.g. Liégeois et al. 2005). At the surface, the Lower Permian continental volcano-sedimentary sequences are generally separated from the uppermost Permian red beds by an erosional surface (Cassinis et al. 2003).

Finally, the dispersion of data in the high-luminescence patches of zircons towards younger ages (scattered between 207 ± 8 and 261 ± 11 Ma) suggests multiple resetting of the U–Pb system, probably as a result of the Late Permian and Triassic thermal anomalies.

Conclusions

U–Pb dating of the zircons from the volcanic succession constrains the duration of post-Variscan magmatic activity in the
Ligurian Briançonnais. In particular, the new data indicate the following.

(1) The calc-alkaline volcanism occurred over a time interval between c. 286 (Case Lisetto Metarhyolites) and 273 Ma (Lithozone C). As a consequence, the host sediments must be Early Permian rather than Late Carboniferous as inferred from pristine stratigraphy. Therefore the orogen collapse is on the whole a younger event, constrained also by the inferred exhumation rate.

(2) Within the continental volcanic series, the U–Pb dating indicates a gap of c. 14 Ma between calc-alkaline bimodal and K-alkaline volcanic activity dated at 258.5 ± 2.8 Ma (Lithozone D). The gap probably represents the transition from a thickened lithospheric setting during the post-Variscan collapse to the delaminated post- to anorogenic setting.

(3) It is likely that during the Early Permian the lithosphere rheology was modified, and plasticity enhanced, by extensive partial melting in the lower crust, which contributed to the production of the ryolites of Lithozone D. We suggest that the 14 Ma gap corresponds to the major displacement along the megashare zone between Laurussia and Gondwanaland; this is also supported by palaeomagnetic constraints (Muttoni et al. 2003).

(4) Finally, in contrast to the Early Permian calc-alkaline series, emplaced within transtensional pull-apart basins, the last K-alkaline products were emplaced in an extensional regime (horst and graben structures).

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