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Crustal Contributions to Late Hercynian Peraluminous Magmatism in the Southern Calabria–Peloritani Orogen, Southern Italy: Petrogenetic Inferences and the Gondwana Connection

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Sensitive high-resolution ion microprobe (SHRIMP) analyses of zircon from granites of the medium-high grade Aspromonte–Peloritani Unit, Calabria–Peloritani Orogen (CPO), southern Italy, show that one of the minor trondhjemites (313·7 ± 3·5 Ma) represents the earliest identified occurrence of Late Hercynian peraluminous igneous rocks in the CPO, predating the emplacement of the more common peraluminous leucogranodiorites by about 14 Myr. Some of the trondhjemite zircon grains contain small cores with ages of about 2·45 Ga, 625 Ma and 490 Ma, consistent with the presence of a sediment component in the magma. A newly dated leucogranodiorite (300·2 ± 3·8 Ma) is rich in inherited zircon. Cores with ages of about 2·36 Ga, 870 Ma, 630 Ma, 545 Ma and 460 Ma are overgrown by two generations of Hercynian igneous zircon, the first with moderate to high Th/U (up to 1·67), and the second with low Th/U (<0·1). The overgrowths probably crystallized from magmas of two compositions, the first metaluminous and the second peraluminous. This could indicate either magma mixing or, more probably, crystallization in a single, evolving magma. In either case, the leucogranodiorite magma is considered to have been the product of anatexis of a metasedimentary source. Differences in the inherited zircon age spectra, and the relatively small amount of inheritance in the trondhjemite, indicate that the trondhjemite and leucogranodiorite are unlikely to be genetically related. The ages of the inherited zircons are consistent with the sedimentary component in both magmas being derived from North Africa, with a possible contribution from Pan-African granitoids similar to those exposed in southern Calabria.

KEY WORDS: *Calabria–Peloritani Orogen; Hercynian peraluminous magmatism; inheritance; trondhjemite; SHRIMP zircon ages*

INTRODUCTION

The last stages of the Hercynian orogeny in southern Europe were characterized by the emplacement of large volumes of felsic to intermediate magmas, creating granite batholiths and isolated plutons that form the backbone of the main circum-Mediterranean segments of the Hercynian Belt (Bonin *et al.*, 1993; Bea *et al.*, 2003; Vilà *et al.*, 2005). In the Calabria–Peloritani Orogen (CPO) of southern Italy, part of the central Mediterranean segment of the Hercynian orogen, there are two main late to post-Hercynian granite associations; a predominant metaluminous to weakly peraluminous suite and a less abundant strongly peraluminous one (Rottura *et al.*, 1993, and references therein). The granites of the metaluminous suite are the main components of large composite batholiths in

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central–northern Calabria. In contrast, the smaller, strongly peraluminous, granodioritic to leucogranitic bodies of the minor suite crop out throughout the CPO and are the most widespread granite association in its southernmost part, within the Aspromonte–Peloritani Unit of southern Calabria and northeastern Sicily. The strongly peraluminous granites have been interpreted as either typical S-type granites (D'Amico *et al.*, 1982; Rottura *et al.*, 1990) or as of mixed mantle–crust origin (Rottura *et al.*, 1991, 1993). The isotopic ages for granites of both suites range from *c.* 303 to *c.* 290 Ma (mineral and whole-rock Rb–Sr, zircon and monazite U–Pb; Borsi & Dubois, 1968; Borsi *et al.*, 1976; Schenk, 1980; Del Moro *et al.*, 1982; Graessner *et al.*, 2000).

A third granite type that has been largely overlooked, despite being widely distributed within the Aspromonte–Peloritani Unit in NE Peloritani and southern Calabria, is trondhjemitic in composition. The trondhjemites have been linked to the late Hercynian magmatism on the basis of field, petrographic and geochemical evidence (Atzori *et al.*, 1984*a*; Fiannacca *et al.*, 2005), but their ages have not previously been measured. One proposal has been that the trondhjemites are in fact Hercynian granites that have been altered by alkali metasomatism (Fiannacca *et al.*, 2005).

To unravel the complex tectono-metamorphic evolution of the Calabria–Peloritani segment of the Hercynian chain (now incorporated into an Alpine–Apennine nappe system) and in particular the petrogenesis of the granites, it is necessary to determine the sequence of igneous events. The history and structure of the CPO is the result of pre-Hercynian to Alpine events that also affected many other European basement terranes, starting at least as early as the Early Palaeozoic (Stampfli & Borel, 2002; von Raumer *et al.*, 2002, 2003).

Here we report sensitive high-resolution ion microprobe (SHRIMP) measurements of zircon U–Th–Pb ages from a leucogranodiorite and a trondhjemite from the Aspromonte–Peloritani Unit of the Aspromonte Massif and the northeastern Peloritani Mountain Belt, respectively. Dating by ion microprobe avoided some of the problems commonly encountered in isotope dilution thermal ionization mass spectrometry (ID-TIMS) zircon analysis; for example, biasing of the ages by Pb loss or the presence of inheritance. Furthermore, imaging of the sectioned zircon grains by cathodoluminescence (CL) prior to analysis made it possible to target discrete zircon components (e.g. inherited and melt-precipitated zircon) and to avoid analysis of inclusions or altered domains. It was also possible to analyse zircon crystallized at different stages of the Hercynian igneous episode.

GEOLOGICAL SETTING

The Peloritani Mountains and the Aspromonte Massif are located in northeastern Sicily and southern Calabria, respectively (Fig. 1). They represent the southernmost part of the Calabria–Peloritani Orogen, an arcuate belt connecting the Southern Apennine and Maghrebid chains. The evolution and geodynamic significance of the Calabria–Peloritani Orogen remain the subject of numerous contrasting interpretations, principally because of the complexity produced by multiple dynamothermal events and the difficulty in correlating between the several segments of the orogen (Peloritani, Aspromonte, Serre, the Coastal Chain and Sila). Pre-Mesozoic crystalline nappes that crop out in the CPO have been variously interpreted as fragments of the neo-Tethyan continental margin of either Europe (Ogniben, 1973; Bouillin *et al.*, 1986; Knott, 1987) or Africa (Haccard *et al.*, 1972; Alvarez, 1976; Amodio-Morelli *et al.*, 1976; Grandjacquet & Mascle, 1978), as a part of a micro-continent formerly located between the two margins (Guerrera *et al.*, 1993; Bonardi *et al.*, 1996, 2001; Perrone, 1996; Critelli & Le Pera, 1998), or as a result of the accretion of three crustal micro-blocks (e.g. Vai, 1992).

The Peloritani Mountains and the Aspromonte Massif are composed of a pile of south-verging Alpine nappes, consisting mostly of Hercynian basement rocks with fragments of Meso-Cenozoic cover rocks (Lentini & Vezzani, 1975; Pezzino *et al.*, 1990; Ghisetti *et al.*, 1991). The Peloritani Mountains have been subdivided into two domains characterized by different tectono-metamorphic histories (Atzori *et al.*, 1994; Cirrincione & Pezzino, 1994). The Lower Domain is exposed in the southern part of the Peloritani Belt and consists of three tectonic units, each composed of very low to low-grade metamorphic sequences and a Meso-Cenozoic sedimentary cover. The overlying Upper Domain, in the northeastern part of the belt, consists of two tectonic units: (1) the phyllitic Mandanici Unit, which was affected by low- to medium-grade Hercynian metamorphism and an Alpine sub-greenschist- to greenschist-facies metamorphic overprint and (2) the uppermost Aspromonte–Peloritani Unit, composed of Hercynian amphibolite-facies rocks (dominated by biotite paragneisses and augen gneisses, with minor micaschists, amphibolites and marbles). The metamorphic rocks of the Aspromonte–Peloritani Unit have been intruded by numerous late Hercynian peraluminous granites, and both the metamorphic and igneous rocks have been locally affected by a medium-pressure, low-temperature Alpine overprint that produced pseudotachylites and belts of cataclastic to mylonitic rocks.

The Aspromonte Massif consists of three crystalline tectonic units (Crisci *et al.*, 1982; Bonardi *et al.*, 1984; Pezzino *et al.*, 1990). The lowermost unit (the Madonna di Polsi Unit; Pezzino *et al.*, 2008, and references therein) consists of low- to medium-grade rocks in a structural position equivalent to that of the Mandanici Unit in the Peloritani Mountains. The Madonna di Polsi Unit has recently been recognized as the Mesozoic sedimentary cover of the

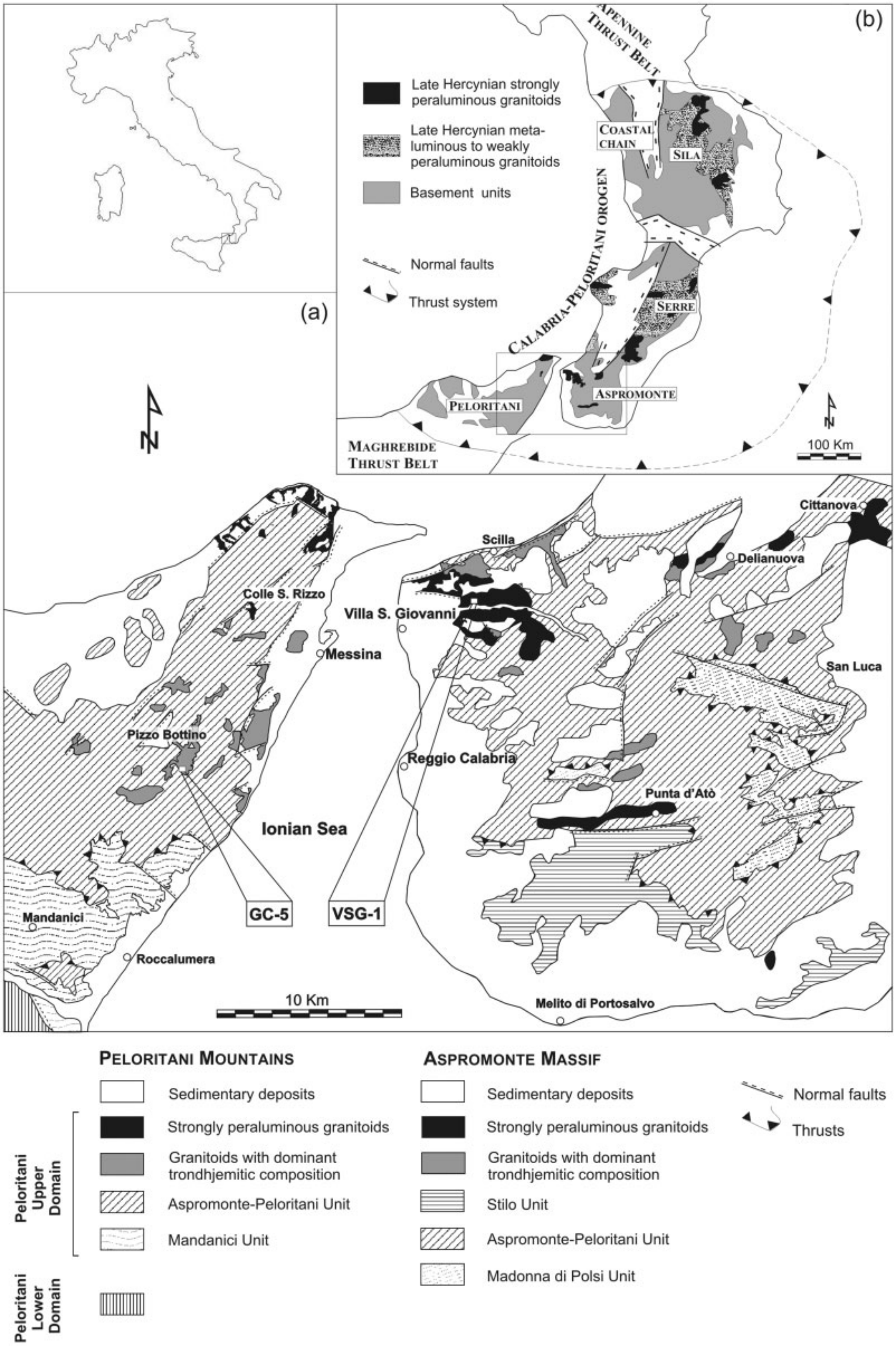

Fig. 1. (a) Geological sketch map of the western Aspromonte Massif and northeastern Peloritani Mountains showing the main tectonic units, the distribution of late Hercynian granitoids (after Atzori *et al.*, 1983; Lentini *et al.*, 2000; Ortolano *et al.*, 2005) and the locations of trondhjemite sample GC-5 and leucogranodiorite sample VSG-1. (b) Distribution of crystalline basements units and Late Hercynian granitoids in the Calabria–Peloritani Orogen and location of the study area.

Mandanici Unit [Pezzino *et al.* (2008) have shown that it experienced only Alpine metamorphism]. The middle unit is the same Aspromonte–Peloritani Unit that crops out in the Peloritani Mountains. The uppermost Stilo Unit, which is absent in the Peloritani Mountains, is composed of low greenschist- to low amphibolite-facies metapelites intruded by late Hercynian peraluminous granites.

Previous ID-TIMS monazite U–Pb ages (Graessner *et al.*, 2000) from upper crustal amphibolite-facies paragneisses in the Aspromonte–Peloritani Unit in southern Calabria record a metamorphic peak at *c.* 295 Ma (620°C at *c.* 250 MPa for the base of the upper crust), coeval with metamorphism in the lower crust (690–800°C at 550–750 MPa; Schenk, 1984; Graessner *et al.*, 2000) and nearly synchronous with the intrusion of the granites at 304–290 Ma (U–Pb and Rb–Sr ages; Borsi & Dubois, 1968; Borsi *et al.*, 1976; Schenk, 1980; Del Moro *et al.*, 1982; Graessner *et al.*, 2000). During the Hercynian metamorphism of the Aspromonte–Peloritani Unit in the Peloritani Mountains, conditions reached 500–680°C at 300–500 MPa (Ioppolo & Puglisi, 1989; Atzori *et al.*, 1984*b*; Messina *et al.*, 1996), similar to the peak P–T conditions of 650–675°C and 390–500 MPa estimated for the rocks of the Central Aspromonte Massif (Ortolano *et al.*, 2005).

Late Hercynian magmatism was widespread throughout the CPO, when several granite complexes were emplaced into the upper–middle crust. Those granites belong to two different groups, a main calc-alkaline suite of metaluminous to weakly peraluminous rocks forming large batholiths, and a less extensive strongly peraluminous suite. The former, representing *c.* 70% of the exposed granite, has a broad compositional range (48–70% SiO_2). Biotite tonalites and granodiorites are the dominant rock types. The strongly peraluminous granites have a more restricted range of compositions (67–76% SiO_2) and contain two micas, with or without Al-silicates. The granites are late to post-tectonic and were probably emplaced along extensional ductile shear zones (Rottura *et al.*, 1990). The younger, unfoliated to weakly foliated, strongly peraluminous and calc-alkaline granites intruded in a brittle domain, whereas the older, strongly foliated calc-alkaline granites were emplaced at a deeper structural level (Rottura *et al.*, 1990). The calc-alkaline suite has been interpreted as resulting from the interaction of mantle-derived magmas with crustal rocks (Rottura *et al.*, 1991). The granites of the strongly peraluminous suite have been interpreted as either typical S-type granites (D'Amico *et al.*, 1982; Rottura *et al.*, 1990) or as having a mixed mantle–crust origin (Rottura *et al.*, 1991, 1993). Only peraluminous granites, of granitic to granodioritic and trondhjemitic composition, crop out within the Aspromonte–Peloritani Unit. They occur as small plutons and stocks, and as discordant to sub-concordant dykes up to several metres wide. Preliminary studies of some of the trondhjemite bodies have concluded that they possibly originated in association with late Hercynian peraluminous magmatism (Atzori *et al.*, 1984*a*; Lo Giudice *et al.*, 1985; Fiannacca *et al.*, 2005).

PREVIOUS GEOCHRONOLOGY

Most of the geochronological results from metamorphic and igneous rocks from the crystalline basement of the southern sector of the CPO have indicated a temporal link to the Hercynian orogeny.

Schenk (1990) reported ID-TIMS zircon and monazite U–Pb analyses from felsic granulites of the Serre Massif, southern Calabria. Highly discordant zircon analyses defined a discordance line with a concordia intercept of 300 ± 10 Ma. Several monazite analyses were concordant between *c.* 296 and 289 Ma. These ages are consistent with zircon and monazite ages of *c.* 295 Ma previously obtained from the basement of the Serre Massif (Schenk, 1980) and possibly record the peak of static granulite-facies metamorphism. Bonardi *et al.* (1991) interpreted muscovite Rb–Sr ages of *c.* 314 Ma from rocks of the Aspromonte–Peloritani Unit in southern Calabria as recording the static growth of staurolite, cordierite and andalusite porphyroblasts. U–Pb monazite ages from upper and lower crustal paragneisses from the same unit probably record peak metamorphism at 295–293 ± 4 Ma (Graessner *et al.*, 2000). There is also geochronological evidence in the upper crustal gneisses for early Hercynian events; for example, a biotite Rb–Sr age of 330 Ma (Bonardi *et al.*, 1987) and a poorly defined zircon U–Pb lower concordia intercept age of 377 ± 55 Ma (Schenk, 1990). No evidence for these early events has yet been found in the deep crustal rocks.

The augen gneisses and associated biotite paragneisses in the Peloritanian sector of the Aspromonte–Peloritani Unit appear to have shared a common metamorphic history. Mica Rb–Sr ages of 280–292 Ma have been interpreted as recording cooling after the Hercynian metamorphism (Atzori *et al.*, 1990); ^{40}Ar–^{39}Ar and Rb–Sr dating of amphibole, biotite and muscovite from different outcrops of amphibolite and augen gneiss in northern Peloritani yielded minimum metamorphic ages of 340–300 Ma (De Gregorio *et al.*, 2003).

Hercynian magmatism in southern Calabria spanned the period 298 ± 5 to 270 ± 5 Ma (Rb–Sr whole-rock and mineral ages, zircon U–Pb ages; Borsi & Dubois, 1968; Borsi *et al.*, 1976; Schenk, 1980; Del Moro *et al.*, 1982). The Villa San Giovanni granitoids have given biotite and muscovite Rb–Sr cooling ages of *c.* 286–282 Ma (Del Moro *et al.*, 1982). A zircon U–Pb age of 298 ± 5 Ma measured on a metamorphosed mafic sill intruded into the lower crust possibly records magmatism late in the granulite-facies metamorphism (Schenk, 1980). A similar zircon age, 295 ± 2 Ma, has also been obtained from a large, high-level tonalitic body (Schenk, 1980). A blastomylonitic quartz dioritic gneiss situated between the lower crustal unit and the tonalite shows evidence for biotite recrystallization and Pb loss from zircon at 283 ± 3 Ma, consistent with the quartz diorite having occupied a shear zone during the initial stage of uplift. A poorly defined age of '<314 Ma' has been reported for the Serre granodiorite by Graessner *et al.* (2000, and references therein), whereas two peraluminous granites, from Serre and Aspromonte, have yielded ID-TIMS zircon ages of $303-302 \pm 0{\cdot}6$ Ma (Graessner *et al.*, 2000). Intrusion of large volumes of granitic magma at this time has been suggested as the possible source of heat leading to the peak static metamorphism of the Calabrian crust (Graessner *et al.*, 2000). The only Hercynian granitoids from the Peloritanian sector of the CPO to have been dated are those from Capo Rasocolmo, which have yielded a Rb–Sr whole-rock age of 293 ± 9 Ma and Rb–Sr mica cooling ages of 287–285 Ma (Del Moro *et al.*, 1982).

The isotopic record of the pre-Hercynian evolution of the southern CPO is dominated by evidence for a late Pan-African (600–500 Ma) crust-forming event. Zircon U–Pb ages measured on rocks from many different levels within the southern Calabrian crust by Schenk & Todt (1989) and Schenk (1990) include:

(1) a 553 ± 27 Ma intrusion age (based on discordant zircon) for a granulite-facies calc-alkaline metabasite;
(2) a poorly defined 622 ± 120 Ma intrusion age for I-type granitic gneisses;
(3) a 516 ± 25 Ma lower intercept age (interpreted as the intrusion age) for S-type granitic gneisses; the upper intercept age, *c.* 2·3 Ga, was interpreted as the age of inheritance;
(4) detrital zircon from an unmetamorphosed (probably Devonian) siltstone defining a discordance line with intercepts of 550 ± 50 Ma and *c.* 2·5 Ga; dominance of the Pan-African component indicates that the orthogneisses described above might be the source of some of the detritus.

Neoproterozoic $^{40}Ar-^{39}Ar$ hornblende ages and latest Palaeoproterozoic U–Pb titanite ages have been reported from the Peloritani Mountains (De Gregorio *et al.*, 2003). ID-TIMS zircon dating of felsic porphyries from the Peloritani Lower Domain has identified a mid-Ordovician (*c.* 455 Ma) crust-forming event, and inherited zircon with ages of *c.* 2·02 and 1·15 Ga has been found in andesite from the same area (Trombetta *et al.*, 2004). Recent secondary ionization mass spectrometry (SIMS) zircon dating by Micheletti *et al.* (2007) has identified Late Neoproterozoic to Early Cambrian ages (562 ± 15, 547 ± 7, 540 ± 4, 539 ± 16 and 526 ± 10 Ma) for the igneous protoliths of five Calabrian augen gneisses (two from the Aspromonte–Peloritani Unit), as well as Archaean, Palaeoproterozoic and Neoproterozoic inheritance.

SAMPLES

One sample of trondhjemite and one of leucogranodiorite from the Aspromonte–Peloritani Unit were selected for SHRIMP dating to compare the features of their zircon populations (e.g. morphology, zoning, compositional range), and the relative ages of both intrusion and inheritance. These are the first *in situ* zircon ages obtained from Hercynian granitic rocks of the CPO and the first direct age measurements on the Calabro-Peloritanian trondhjemites. They provide new insights into previously unrecognized late Hercynian magmatism in this sector of the Hercynian Belt.

Pizzo Bottino trondhjemites

Petrographic and geochemical characteristics of the Peloritani trondhjemites have been reported in detail by Atzori *et al.* (1984*a*), Lo Giudice *et al.* (1985) and Fiannacca *et al.* (2005).

The Pizzo Bottino trondhjemites (PB*t*) are mostly coarse- to very coarse-grained rocks. Recrystallization of varied intensity has overprinted the original igneous features, some of which are preserved in places as structural relics. Rock-forming minerals are dominantly oligoclase plagioclase and quartz (up to 90% by volume), with small amounts of biotite, muscovite and microcline. Accessory phases include apatite, zircon, sillimanite, Fe–Ti oxides and rare monazite and garnet. Tiny metasedimentary enclaves composed of muscovite + sillimanite, and biotite + muscovite ± sillimanite ± quartz ± plagioclase ± apatite, are common in some places. No mafic microgranular enclaves have been observed. Plagioclase mostly occurs as anhedral to subhedral megacrysts up to 5–6 cm long. In some samples there are millimetre-sized plagioclase crystals with igneous features (e.g. euhedral elongated habit, idiomorphic oscillatory zoning and simple twinning). Quartz mainly occurs as discrete medium to large anhedral grains or as polycrystalline aggregates; anhedral or rounded quartz also occurs within plagioclase megacrysts. Microcline occurs as rare interstitial patches or, more commonly, as homoaxial scattered inclusions in large plagioclase grains. Quartz and microcline inclusions in plagioclase megacrysts are sometimes so abundant as to resemble a chessboard texture. This texture, as well as the plagioclase and/or myrmekite growth at the expense of both older microcline and plagioclase observed in some trondhjemite samples, has been interpreted as a replacement texture related to alkali metasomatism (Fiannacca *et al.*, 2005, and references therein). Biotite, other than in the polymineralic aggregates, sometimes occurs as essentially monomineralic clots. It (and muscovite) also occurs as discrete euhedral or subhedral plates of varied size, enclosed in plagioclase. Accessory sillimanite and garnet are interpreted as being restitic or xenolithic in origin.

The PB*t* have 71–76% SiO_2, high Al_2O_3 and Sr, low Ba, Nb, Y, Ni and Cr, and very low K_2O/Na_2O and Rb/Sr ratios. They are mildly peraluminous (A/CNK mostly 1·0–1·1, rarely >1·2). The Zr contents are in the range

25–167 ppm and the rare earth element (REE) contents are varied; light REE (LREE) 10–100 times chondrites; La_N/Yb_N 14–20; Eu anomalies negative to highly positive ($Eu/Eu^* = 0{\cdot}6–13{\cdot}9$). $(^{87}Sr/^{86}Sr)_i$ and $(\varepsilon_{Nd})_i$, calculated at 290 Ma, are in the range 0·7073–0·7076 and −6·7 to −6·9, respectively (Fiannacca *et al.*, 2005). Trondhjemites from the Peloritani Mountains have been variously interpreted as the products of the isochemical metamorphism of arkoses (Atzori *et al.*, 1974), the partial melting of biotite paragneisses (Atzori *et al.*, 1984*a*) and the fluid-assisted metamorphic differentiation or metasomatic alteration of metasediments (D'Amico *et al.*, 1972; Lo Giudice *et al.*, 1985). Fiannacca *et al.* (2005) suggested that the trondhjemites from the Pizzo Bottino area originated by alkali metasomatism of original late Hercynian peraluminous granitoids.

Sample GC5 (38°06′02″N, 15°26′16″E; road cut, mountain road, 17 km SW of Messina, 300 m south of Puntale Tammurinaru; Fig. 1) is of trondhjemite collected from the Pizzo Bottino body. The sample is coarse- to very coarse-grained, with centimetre-scale plagioclase megacrysts. Approximate modal abundances are: plagioclase (50 vol. %), quartz (38 vol. %), K-feldspar (4 vol. %), biotite (4 vol. %), muscovite (3 vol. %), sillimanite (1 vol. %). Accessory phases include apatite, zircon, monazite, Fe–Ti oxides and epidotes. The chemical composition of the sample is listed in Table 1.

Villa S. Giovanni leucogranodiorites

Petrographic and geochemical features of the Villa S. Giovanni granitoids (VSG*g*) have been thoroughly documented by Messina *et al.* (1974), D'Amico *et al.* (1982), Del Moro *et al.* (1982) and Rottura *et al.* (1990, 1993). The textures are hypidiomorphic and inequigranular as a result of the occurrence of variously sized plagioclase, quartz and K-feldspar. Zoned plagioclase (cores $An_{35–53}$, rims $An_{16–35}$, with a core–rim difference commonly >10% An), quartz and biotite occur within a 'matrix' composed of oligoclase, quartz, microcline and muscovite ± fibrolite. Metamorphic aggregates of probable xenolithic origin composed of muscovite + fibrolitic sillimanite ± cordierite ± quartz ± plagioclase ± apatite ± opaque minerals or clusters of muscovite + biotite with relics of fibrolite are common within the leucogranodiorites. No mafic enclaves have been observed.

The VSG*g* are all strongly peraluminous (A/CNK ≥1·1) and characterized by moderate to high SiO_2 (67–76%), high Al_2O_3, Ba and Sr; low Rb and mafic components ($FeO_t + MgO + TiO_2 = 1{\cdot}5–3{\cdot}9\%$), and varied CaO, total alkalis and K_2O/Na_2O. The REE patterns are highly fractionated (average $La_N/Yb_N = 34$; $Yb_N = 3{\cdot}7$) with moderate negative or no Eu anomalies ($Eu/Eu^* = 0{\cdot}6–1{\cdot}1$), consistent with equilibration with garnet-bearing residua, possibly analogous to the garnet–sillimanite-rich metapelitic rocks of the Calabrian lower crust, which some

Table 1: Major, minor and trace element composition of leucogranodiorite VSG-1 and trondhjemite GC-5

Sample:	VSG-1	GC-5
wt %		
SiO_2	73·97	74·93
TiO_2	0·10	0·10
Al_2O_3	14·45	15·06
Fe_2O_{3TOT}	1·35	0·70
MnO	0·03	0·02
MgO	0·28	0·31
CaO	1·09	2·31
Na_2O	3·70	4·96
K_2O	3·87	1·16
P_2O_5	0·20	0·12
LOI	0·81	0·76
Total	99·85	100·43
A/CNK	1·18	1·11
ppm		
Ba	598	609
Rb	134	35
Sr	180	510
Y	10·0	9
Zr	57	142
Nb	10	2
Cr	−20	−20
Ni	−20	−15
Th	6·1	8·0
U	1·4	1·7
La	20·0	26·0
Ce	38·0	53·9
Pr	4·22	6·12
Nd	15·6	23·6
Sm	3·2	4·1
Eu	0·72	0·85
Gd	2·5	3·3
Tb	0·4	0·4
Dy	1·7	1·8
Ho	0·3	0·3
Er	0·7	0·8
Tm	0·09	0·11
Yb	0·5	0·8
Lu	0·07	0·10

Analyses performed by ICP and ICP-MS (following the procedure 4-Litoresearch), at Actlabs, Ontario, Canada. Reported relative errors are 5% or less for major elements and about 5–15% for most minor and trace elements. Numbers preceded by (–) indicate contents lower than detection limit, expressed by the same numbers. LOI, loss on ignition.

researchers have interpreted as restitic (Schenk, 1990; Caggianelli *et al.*, 1991; Fornelli *et al.*, 2002). The VSG granitoids have a range of initial (at 290 Ma) $^{87}Sr/^{86}Sr$ ratios (0·7098–0·7115) and ε_{Nd} (–7·0 to –8·4; Rottura *et al.*, 1990), consistent with derivation from a mature crustal source.

Del Moro *et al.* (1982) interpreted the Rb–Sr isotopic compositions of the VSG leucogranodiorites as indicating their derivation from a heterogeneous metasedimentary source. They inferred that two crustal components, characterized by different Sr concentrations and isotopic compositions, were involved in their genesis. According to Rottura *et al.* (1990), the VSG granitoids originated from a LREE-enriched, garnet-bearing, crustal source that was dominantly quartzo-feldspathic, rather than pelitic. They argued that the strongly peraluminous granitoids of the CPO have an S-type signature in terms of their mineralogical composition, enclave population and zircon typology, and that their geochemical features are analogous to those of late to post-collisional granites. Contrary to previous petrogenetic interpretations, Rottura *et al.* (1993) asserted that the granitoids of Villa S. Giovanni (and Capo Rasocolmo) could not be considered S-type granites in the sense of White *et al.* (1986) as they had a mixed origin involving components derived from both the mantle and crust. They proposed that the late Hercynian peraluminous plutons originated following magmatic underplating of the continental crust, parental calc-alkaline magmas having been strongly modified by crustal assimilation and mixing with lower crustal melts.

The sample of leucogranodiorite selected for zircon analysis (VSG-1; 38°12′39″N, 15°41′45″E; road cut, Campo–Fiumara provincial road, 10 km NE of Reggio Calabria, large curve near to the eastern entrance of S. Nicola; Fig. 1) is medium-grained, with the approximate mode: plagioclase (35 vol. %), quartz (35 vol. %), K-feldspar (15 vol. %), biotite (8 vol. %), muscovite (6 vol. %), sillimanite (1 vol. %). Accessory phases include apatite, zircon, monazite, Fe–Ti oxides and epidotes. The chemical composition of the sample is listed in Table 1.

SHRIMP ZIRCON U–TH–PB ANALYSES

Sample preparation and analytical procedures

Approximately 1 kg of each rock was crushed to chips in a jaw crusher, screened to >5 mm, washed in water in an ultrasonic bath, and dried. The chips were crushed in a tungsten carbide swing mill to <250 μm. The powder was deslimed and dried, then the heavy minerals were separated using tetrabromoethane and methylene iodide. Zircon was concentrated using an isodynamic magnetic separator.

Zircon yields from both samples were very small for granitic rocks; only a few hundred grains. About 80 grains from each sample were chosen at random and separated by hand for mounting in epoxy resin with zircon standards TEMORA II ($^{206}Pb^*/^{238}U = 0{\cdot}06683$) and SL13 (U = 238 ppm). After sectioning and polishing, the grains were photographed in transmitted and reflected light, then imaged by CL using a Hitachi S-2250N scanning electron microscope at the Australian National University (Fig. 2). The mount was coated with gold in preparation for SHRIMP analysis.

The zircons were analysed during a single analytical session on the ANU SHRIMP II ion microprobe using procedures based on those described by Williams & Claesson (1987). A 2·5 nA, 10 kV primary beam of O_2^- ions was focused to a probe of *c.* 25 μm diameter. Positive secondary ions were extracted from the sample at 10 kV, and the atomic and molecular species of interest analysed at *c.* 5000 mass resolution using a single ETP electron multiplier and peak switching. The Pb isotopic composition was measured directly, without correction for the small mass dependent mass-fractionation (*c.* 0·25% per a.m.u.). Interelement fractionation was corrected using the TEMORA II reference zircon, using a Pb/U–UO/U power law calibration equation (Claoué-Long *et al.*, 1995). The uncertainty in the Pb/U calibration was 0·46%. Pb, U and Th concentrations were measured relative to SL13. Common Pb corrections were very small (most <0·3 ppm total Pb), so all were made assuming that the common Pb was all laboratory contamination of Broken Hill galena Pb composition ($^{204}Pb/^{206}Pb = 0{\cdot}0625$, $^{207}Pb/^{206}Pb = 0{\cdot}962$, $^{208}Pb/^{206}Pb = 2{\cdot}23$; Cumming & Richards, 1975). Corrections for the plots and isotopic data table were made using ^{204}Pb. Corrections for the calculation of mean $^{206}Pb/^{238}U$ ages used ^{207}Pb, assuming the analyses to be concordant. Uncertainties in the plots and data table are 1σ. Uncertainties in the calculated mean $^{206}Pb/^{238}U$ ages are 95% confidence limits (namely $t\sigma$, where t is 'Student's t') and include the 0·46% uncertainty in the Pb/U calibration. Ages were calculated using the constants recommended by the IUGS Subcommission on Geochronology (Steiger & Jäger, 1977). The U-Th-Pb analyses are listed in Table 2 and plotted on concordia diagrams in Figs 3–5.

Trondhjemite GC-5

Zircon morphology and zoning

Sample GC-5 contains medium-sized (mostly 50–100 μm diameter), pink to pale purple, transparent, prismatic zircon grains (mostly grain fragments), commonly with well-preserved crystal faces and simple pyramidal terminations. Aspect ratios are 1–5. Inclusions are relatively rare. Many grains have numerous fractures, possibly accounting for the rarity of intact crystals. A few grains contain a core visible under an optical microscope. CL imaging (Fig. 2) revealed relatively simple growth

Table 2: Ion microprobe U–Th–Pb isotopic data for zircon from leucogranodiorite VSG-1 and trondhjemite GC-5

Grain. spot	Pb* (ppm)	U (ppm)	Th (ppm)	Th/U	^{204}Pb/^{206}Pb	±	f206%†	^{208}Pb*/^{206}Pb	±	^{208}Pb*/^{232}Th	±	^{206}Pb*/^{238}U	±	^{207}Pb*/^{206}Pb	±	Apparent ages (Ma) 208/232	±	206/238	±	207/206	±	Preferred age‡	±
Leucogranodiorite VSG-1																							
Cores																							
5·2	102	206	211	1·02	1·87E–05	2·82E–05	0·03	0·2779	0·0047	0·1089	0·0026	0·4010	0·0060	0·1511	0·0014	2090	48	2174	27	2358	16	2358	16
7.2	18	121	33	0·27	6·54E–05	1·21E–04	0·11	0·0838	0·0053	0·0448	0·0032	0·1458	0·0045	0·0723	0·0024	886	61	877	25	994	68	873	25
2.2	9	85	42	0·49	3·27E–04	2·82E–04	0·52	0·1390	0·0136	0·0302	0·0031	0·1069	0·0032	0·0626	0·0049	600	61	654	19	696	174	654	18
20.1	63	563	440	0·78	7·36E–06	7·92E–06	0·01	0·2419	0·0049	0·0310	0·0008	0·0999	0·0011	0·0601	0·0009	616	15	613·7	6·6	606	34	613·8	6·6
6.2	20	220	94	0·43	1·66E–04	1·03E–04	0·27	0·1313	0·0069	0·0273	0·0015	0·0889	0·0013	0·0562	0·0024	544	30	549·0	7·9	460	98	550·4	8·0
15.1	34	362	198	0·55	2·00E–05	2·00E–05	0·03	0·1725	0·0034	0·0281	0·0006	0·0891	0·0009	0·0596	0·0019	561	13	550·3	5·2	591	69	549·6	5·3
17.1	32	386	48	0·12	4·86E–05	3·18E–05	0·08	0·0343	0·0018	0·0248	0·0014	0·0890	0·0010	0·0602	0·0016	496	27	549·5	5·8	610	60	548·6	5·8
14.1	22	234	139	0·59	1·50E–04	1·09E–04	0·24	0·1821	0·0059	0·0272	0·0010	0·0884	0·0012	0·0562	0·0021	542	19	546·3	7·1	462	85	547·6	7·1
13.1	27	300	137	0·46	2·17E–05	2·00E–05	0·04	0·1474	0·0056	0·0284	0·0012	0·0879	0·0011	0·0586	0·0012	565	23	543·3	6·4	554	45	543·2	6·4
11.1	22	262	67	0·25	1·98E–04	8·66E–05	0·32	0·0828	0·0041	0·0283	0·0015	0·0871	0·0014	0·0547	0·0020	564	30	538·5	8·5	401	83	540·6	8·5
19.1	19	201	125	0·62	2·00E–05	2·00E–05	0·03	0·1983	0·0051	0·0277	0·0009	0·0869	0·0013	0·0597	0·0027	552	17	536·9	7·9	592	102	536·0	8·0
4.1	10	139	70	0·50	4·25E–04	2·06E–04	0·68	0·1502	0·0102	0·0218	0·0019	0·0729	0·0035	0·0520	0·0038	437	38	454	21	286	176	456	21
Igneous																							
12.1	28	629	37	0·06	1·53E–05	2·02E–05	0·02	0·0205	0·0015	0·0168	0·0012	0·0484	0·0006	0·0525	0·0011	337	25	304·7	3·9	307	49	304·6	3·9
10.1	23	349	590	1·69	2·00E–05	2·00E–05	0·03	0·5247	0·0158	0·0149	0·0005	0·0481	0·0007	0·0501	0·0013	299	10	302·7	4·4	201	60	303·4	4·4
5.1	45	1012	30	0·03	1·34E–05	1·70E–05	0·02	0·0098	0·0009	0·0160	0·0014	0·0483	0·0004	0·0535	0·0012	321	28	303·8	2·6	351	51	303·4	2·6
7.1	5	108	59	0·54	6·32E–04	3·78E–04	1·01	0·1583	0·0159	0·0138	0·0014	0·0475	0·0012	0·0438	0·0069	277	29	299·0	7·4	—	—	301·8	7·4
9.1	34	638	513	0·80	6·33E–05	4·16E–05	0·10	0·2542	0·0071	0·0151	0·0005	0·0478	0·0006	0·0513	0·0014	303·3	9·2	301·1	3·4	255	65	301·4	3·5
3.1	27	456	547	1·20	1·21E–04	8·97E–05	0·19	0·3777	0·0082	0·0150	0·0004	0·0477	0·0007	0·0506	0·0027	300·7	8·4	300·1	4·6	225	129	300·7	4·6
6.1	20	449	71	0·16	2·41E–04	2·13E–04	0·39	0·0473	0·0079	0·0141	0·0024	0·0474	0·0006	0·0510	0·0036	284	47	298·2	3·7	239	170	298·7	3·6
16.1	24	531	72	0·14	7·02E–05	3·99E–05	0·11	0·0433	0·0028	0·0151	0·0010	0·0474	0·0006	0·0523	0·0012	302	20	298·4	3·6	299	53	298·4	3·6
18.1	17	347	216	0·62	3·01E–04	3·44E–04	0·48	0·1817	0·0136	0·0138	0·0011	0·0472	0·0007	0·0492	0·0056	277	21	297·3	4·2	156	247	298·3	3·9
1.1	42	727	845	1·16	2·63E–05	3·00E–05	0·04	0·3610	0·0061	0·0147	0·0003	0·0473	0·0005	0·0524	0·0011	295·1	5·9	298·1	2·8	304	48	298·1	2·8
8.1	27	422	705	1·67	1·62E–05	2·42E–05	0·03	0·5123	0·0077	0·0146	0·0003	0·0474	0·0006	0·0546	0·0012	291·9	5·8	298·8	3·4	395	52	298·1	3·4
2.1	28	498	499	1·00	5·35E–05	7·38E–05	0·09	0·3243	0·0060	0·0153	0·0004	0·0472	0·0006	0·0533	0·0018	306·5	7·2	297·6	3·7	339	80	297·3	3·8

(continued)

Table 2: Continued

Grain. spot	Pb* (ppm)	U (ppm)	Th (ppm)	Th/U	$^{204}Pb/^{206}Pb$	±	f206%†	$^{208}Pb^*/^{206}Pb$	±	$^{208}Pb^*/^{232}Th$	±	$^{206}Pb^*/^{238}U$	±	$^{207}Pb^*/^{206}Pb$	±	Apparent ages (Ma)						Preferred age‡	±
																208/232	±	206/238	±	207/206	±		
Trondhjemite GC-5																							
Cores																							
11.2	234	591	164	0·28	3·51E–06	2·94E–06	0·01	0·0894	0·0017	0·1185	0·0027	0·3689	0·0038	0·1591	0·0014	2264	49	2024	18	2446	14	2446	14
6.2	17	160	76	0·48	1·16E–04	1·09E–04	0·19	0·1343	0·0139	0·0289	0·0031	0·1024	0·0031	0·0648	0·0033	575	62	629	18	769	112	626	18
5.2	9	109	29	0·27	2·00E–05	2·00E–05	0·03	0·0993	0·0064	0·0296	0·0021	0·0794	0·0020	0·0587	0·0020	589	41	492	12	557	76	491	12
Igneous																							
7.1	95	1942	326	0·17	4·90E–05	2·05E–05	0·08	0·0528	0·0020	0·0162	0·0006	0·0516	0·0004	0·0535	0·0007	325	13	324·1	2·4	348	28	323·9	2·4
8.1	106	2253	248	0·11	2·00E–05	2·00E–05	0·03	0·0324	0·0010	0·0148	0·0005	0·0504	0·0003	0·0517	0·0007	298	10	316·9	2·0	271	32	317·3	2·0
2.1	73	1555	191	0·12	4·03E–05	3·05E–05	0·07	0·0378	0·0015	0·0155	0·0006	0·0503	0·0004	0·0529	0·0008	310	12	316·2	2·5	326	34	316·1	2·5
10.1	99	2059	371	0·18	2·50E–05	1·85E–05	0·04	0·0561	0·0012	0·0157	0·0004	0·0503	0·0003	0·0535	0·0006	313·9	7·2	316·3	1·7	348	25	316·0	1·7
6.1	158	3466	103	0·03	2·01E–06	2·67E–06	0·01	0·0092	0·0005	0·0155	0·0009	0·0500	0·0003	0·0526	0·0007	311	17	314·8	1·6	312	30	314·8	1·6
11.1	81	1734	243	0·14	1·96E–05	2·12E–05	0·03	0·0414	0·0032	0·0147	0·0012	0·0498	0·0003	0·0533	0·0007	296	23	313·1	1·6	340	28	312·9	1·6
4.1	169	3492	948	0·27	5·74E–05	2·78E–05	0·09	0·0805	0·0016	0·0147	0·0003	0·0497	0·0002	0·0523	0·0007	295·7	5·9	312·7	1·3	296	31	312·9	1·3
5.1	141	2930	723	0·25	2·34E–05	1·13E–05	0·04	0·0767	0·0012	0·0155	0·0003	0·0497	0·0002	0·0526	0·0007	310·0	5·2	312·6	1·5	310	29	312·6	1·5
9.1	62	1341	165	0·12	5·97E–05	4·58E–05	0·10	0·0364	0·0020	0·0146	0·0008	0·0495	0·0004	0·0526	0·0011	293	16	311·4	2·3	311	49	311·5	2·3
3.1	73	1540	282	0·18	5·14E–05	3·27E–05	0·08	0·0558	0·0017	0·0151	0·0005	0·0495	0·0003	0·0525	0·0008	303·2	9·1	311·4	1·6	305	36	311·4	1·6
1.1	71	1538	193	0·13	3·02E–05	1·79E–05	0·05	0·0390	0·0014	0·0153	0·0006	0·0492	0·0003	0·0529	0·0007	308	11	309·4	1·7	323	28	309·4	1·7

1σ errors.
*Corrected for common Pb of Broken Hill galena composition using ^{204}Pb.
†Percentage of total ^{206}Pb that is common ^{206}Pb.
‡Preferred age estimate, based on $^{206}Pb/^{238}U$ (corrected for common Pb assuming concordance) for apparent ages <1·5 Ga and $^{207}Pb/^{206}Pb$ (corrected for common Pb using ^{204}Pb) for ages >1·5 Ga.

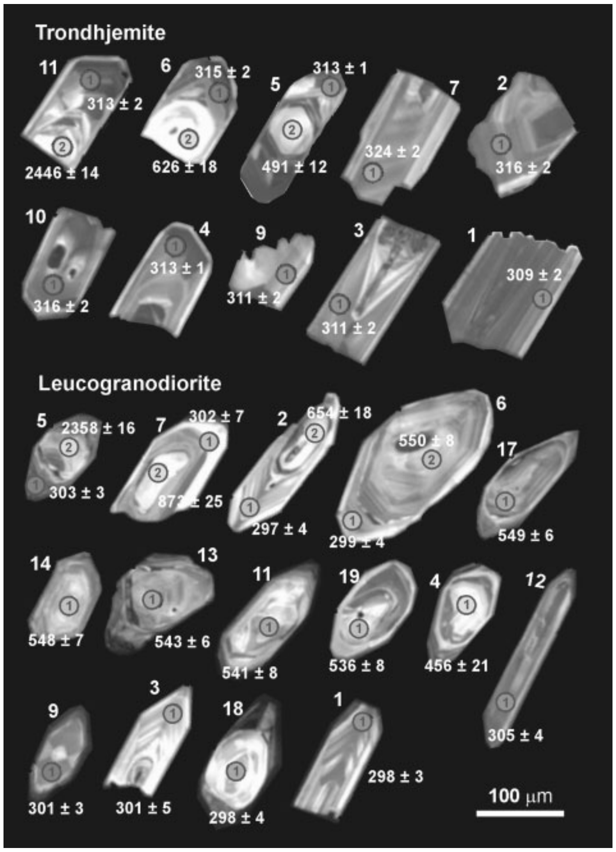

Fig. 2. Cathodoluminescence (CL) images of selected zircon grains from trondhjemite GC-5 and leucogranodiorite VSG-1 with analytical sites and measured ages (see Table 1). The trondhjemite zircon grains are predominantly crystallized from the melt phase of the magma, with small or no inherited cores. The leucogranodiorite zircon grains consist mostly of a large inherited core of Neoproterozoic age surrounded by a thin igneous overgrowth.

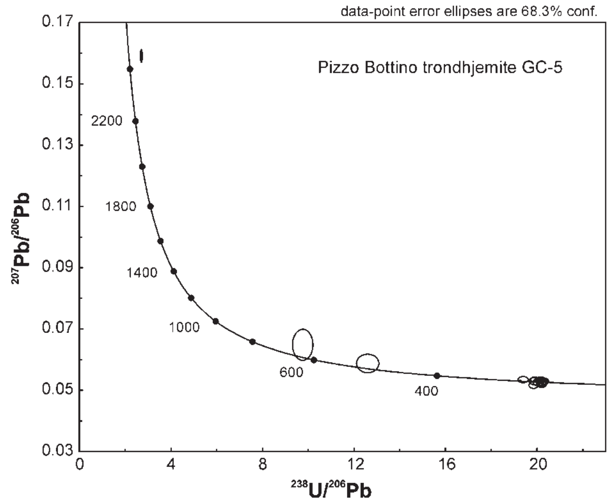

Fig. 3. Concordia diagram showing the whole range of SHRIMP U–Pb analyses of inherited and igneous zircon from Pizzo Bottino trondhjemite sample GC-5. Figure 5 shows the igneous zircon analyses in detail.

structures dominated by prism-parallel oscillatory growth zoning. Cores, commonly small, unzoned, rounded or angular, and more strongly luminescent than the igneous zircon, are present in about 10% of the grains. Many grains have a thin outermost zone or patches or tips that are strongly luminescent. The boundary between the luminescent material and the rest of the grain in some cases crosscuts the zoning, consistent with the luminescent zircon being the product of late local recrystallization.

Analytical results

Zircon with simple igneous zoning, interpreted as having crystallized from the melt fraction of the magma, was analysed from 11 grains. Cores in three of those grains were also analysed. The results are listed in Table 2 and plotted on concordia diagrams in Figs 3 and 5.

The igneous zircon has consistently high U contents (1340–3490 ppm) and low to very low Th/U (0·27–0·03). Radiation damage from U decay is very probably the reason for the pronounced coloration of the grains. In contrast, the U contents of the cores are moderate to low (590–110 ppm) and Th/U slightly higher (0·27–0·48). All analyses of the igneous zircon are concordant within analytical uncertainty, but radiogenic $^{206}Pb/^{238}U$ is more dispersed than expected for analyses of zircon of a single age (MSWD = 7·4). Most of the scatter is due to the high $^{206}Pb/^{238}U$ measured on the three areas with highest U contents (>2930 ppm). This is due to a matrix effect previously documented in SHRIMP analyses of other high-U zircon (Williams & Hergt, 2000). Correcting for the effect (2% enhancement of measured $^{206}Pb/^{238}U$ per 1000 ppm U over 2300 ppm) substantially reduced the scatter, but did not eliminate it (MSWD = 3·6). Most of the remaining scatter was due to one analysis being much higher than the rest (7·1). There was no obvious textural or analytical reason for the high value, but omitting it reduced the scatter almost to insignificance (MSWD = 1·9), although one analysis (1.1) remained slightly but significantly lower than the rest. Omitting that analysis also, the remaining nine analyses had equal radiogenic $^{206}Pb/^{238}U$ within error (MSWD = 1·3), giving a weighted mean age of 313·7 ± 3·5 Ma ($t\sigma$).

The cores yielded a range of apparent ages. Two core analyses are concordant within error at 491 ± 12 (σ) and 626 ± 18 Ma (σ), respectively. The third is *c.* 17% discordant at an inferred age of 2·45 ± 0·01 Ga (σ).

Leucogranodiorite VSG-1

Zircon morphology and zoning

Sample VSG-1 contains fine (25–100 μm diameter), subhedral to euhedral, transparent, colourless, prismatic zircon grains with well-preserved crystal faces. Aspect ratios are 2–7. Many of the grains have

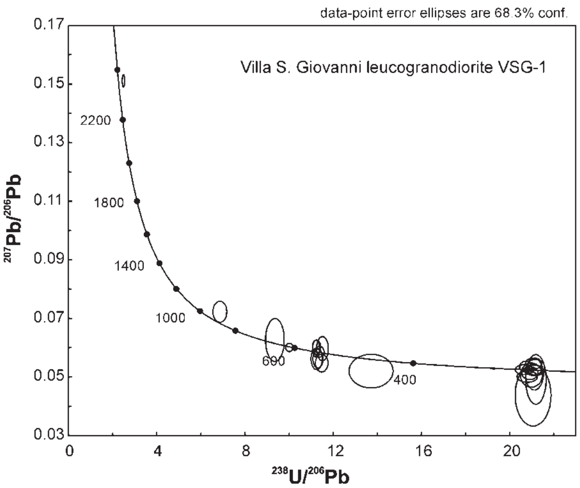

Fig. 4. Concordia diagram showing the whole range of SHRIMP U–Pb analyses of inherited and igneous zircon from Villa S. Giovanni leucogranodiorite VSG-1. Figure 5 shows the igneous zircon analyses in detail.

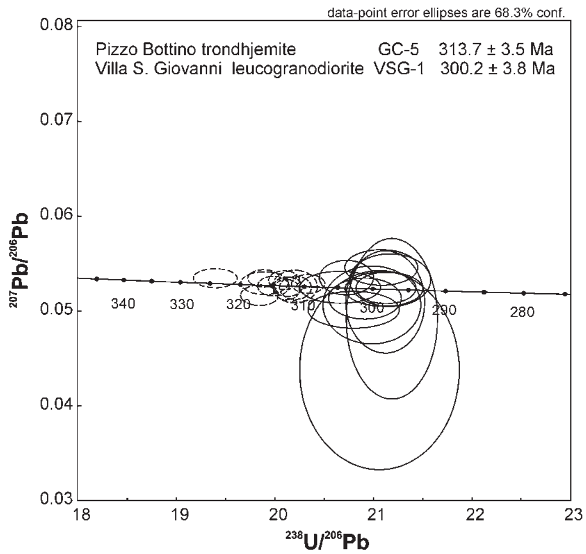

Fig. 5. Concordia diagram showing SHRIMP U–Pb analyses of igneous zircon from the Calabria–Peloritani trondhjemite (dashed lines) and leucogranodiorite (continuous lines) samples in detail. The uncertainties for the reported ages are 2σ.

well-developed {211} crystal faces as commonly seen on zircon grains containing an inherited core (Williams, 2001). Few cores are visible under an optical microscope; however, CL imaging (Fig. 2) revealed a variety of zoning textures, some rather complex. Nearly all grains contain an obvious core. Some grains are composed mainly of zircon with simple oscillatory zoning and contain only a very small core. A very few are mostly sector zoned. Most zircon grains consist of a large core, accounting for more than 80% of the grain diameter, surrounded by a thin overgrowth with weak oscillatory zoning. In many such cases the overgrowth formed only the pyramids at the tips of the grain. Such overgrowths commonly have much weaker luminescence than the cores. Zoning in the cores ranges from indistinct to convolute to oscillatory, with oscillatory zoning predominating. Many cores were rounded or angular fragments of previously larger grains.

Analytical results

Igneous zircon with simple oscillatory or sector zoning, either as core-free grains or overgrowths, was analysed from 12 grains. Cores from four of these grains were also analysed, plus cores from eight other grains. The analytical results are listed in Table 2 and plotted on concordia diagrams in Figs 4 and 5.

The igneous zircon has a wide range of U contents (110–1010 ppm) and Th/U (0·03–1·69), the differences in composition in part correlating with differences in zoning pattern and luminescence (high trace element contents commonly suppress zircon CL). For example, the five analyses with high Th/U ($\geq$1) came from grains with simple zoning and small or no cores. The analysis with highest U (5·1) came from an overgrowth with very weak luminescence, and that with lowest U (7·1) from an overgrowth with very strong luminescence. As a consequence of the generally low U contents (and hence low radiogenic Pb contents), the uncertainties in the analyses of igneous zircon from the leucogranodiorite were larger than those for equivalent zircon from the trondhjemite. The analyses are tightly clustered, however, regardless of the range of zoning patterns. All 12 measurements of radiogenic $^{206}Pb/^{238}U$ are equal within error (MSWD = 0·5), giving a weighted mean age of 300·2 ± 3·8 Ma ($t\sigma$).

The cores yielded a wide range of apparent ages (2360–456 Ma) similar to that of the cores from the trondhjemite. In contrast to the latter, however, most of the leucogranodiorite cores (seven of 12) yielded a very narrow range of concordant apparent ages, all very close to the inferred age of the Precambrian–Cambrian boundary (Gradstein *et al.*, 2004). In fact, the $^{206}Pb/^{238}U$ of those seven cores was the same within analytical uncertainty (MSWD = 0·5), giving a weighted mean age of 546·0 ± 8·6 Ma ($t\sigma$). These cores record a single episode of zircon growth in the region from which their host sediment was ultimately derived. It is just possible that the two cores from the leucogranodiorite and one from the trondhjemite at *c.* 630 Ma together reflect another, slightly earlier, period of zircon growth in that region.

INTERPRETATION

Late Hercynian magmatic zircon

The trondhjemite

The age obtained from the igneous zircon from the Pizzo Bottino trondhjemite, 313·7 ± 3·5 Ma, identifies that unit as the oldest dated Hercynian igneous rock in the Calabria–Peloritani Orogen. It suggests that there was a significant, previously unrecognized, late Hercynian igneous event in the southern CPO. Until now, the peraluminous granitic magmatism in the region has been considered, based on the available age measurements, to be younger than 304 Ma (zircon and monazite ID-TIMS U–Pb ages, Rb–Sr whole-rock and mica ages, $^{40}Ar/^{39}Ar$ muscovite and hornblende ages; Schenk, 1980, 1990; Del Moro *et al.*, 1982; Graessner *et al.*, 2000).

In addition, the new age reported here makes it necessary to review the proposal that the trondhjemites in the Peloritani Mountains are the products of alkali metasomatism of late Hercynian granites (Fiannacca *et al.*, 2005). Notwithstanding the clear metasomatic features displayed by the trondhjemitic rocks, it now appears that they were emplaced at least 10 Myr before the strongly peraluminous granitoids of the southern CPO, their presumed protoliths. There are two possibilities as a consequence. First, the trondhjemites and strongly peraluminous granites are not genetically related. This would be consistent with previous petrographic and geochemical arguments that the rocks could not be related by any known magmatic process (Fiannacca *et al.*, 2005). Second, the trondhjemites represent the first appearance of peraluminous granites, which were subsequently metasomatized, either before or during the major production of peraluminous magmas at *c.* 300 Ma. The inheritance patterns of the zircons from both rock types provide useful information in this regard.

The leucogranodiorite

Dating of the igneous zircon from the Villa S. Giovanni leucogranodiorite indicates that it was emplaced at $300{\cdot}2 \pm 3{\cdot}8$ Ma. This age is consistent with recent ID-TIMS U–Pb ages of $303–302 \pm 0{\cdot}6$ Ma measured on igneous monazite from the peraluminous granites of Punta d'Atò and Cittanova, which also crop out in the Aspromonte Massif (Graessner *et al.*, 2000). The geochronological information now available indicates that the late Hercynian, strongly peraluminous magmatic activity in southern Calabria resulted in the virtually simultaneous intrusion, at *c.* 300 Ma, of several relatively small, discrete plutons. The age of $300{\cdot}2 \pm 3{\cdot}8$ Ma is about 12–14 Myr older than the Rb–Sr cooling ages of *c.* 286–282 Ma obtained by Del Moro *et al.* (1982) for the Villa S. Giovanni granitoids. A similar age difference was reported for the Cittanova granites by Graessner *et al.* (2000), where the U–Pb emplacement ages are 8–15 Myr older than the Rb–Sr mica ages obtained by Del Moro *et al.* (1982).

The use of SIMS has made it possible to obtain evidence for the crystallization of two different generations of igneous zircon in the same rock. The variety of textures and compositions seen in the zircon overgrowths from the leucogranodiorite is unusual, and suggests that the overgrowths crystallized from melts of different compositions, either in different magmas or a single, rapidly evolving magma chamber. The overgrowths that are sector zoned or have high Th/U (≥ 1) probably crystallized in a magma (?metaluminous) with an intermediate SiO_2 content (see Hoskin, 2000). The weakly luminescent overgrowths with higher U and very low Th/U ($<0{\cdot}1$) probably crystallized from a magma with relatively high SiO_2 (?peraluminous), and possibly in the presence of a Th-rich phase such as monazite. The change in igneous zircon composition might mark the point in the evolution of the magma at which monazite became a stable mineral phase.

There is no measurable difference in age between the two types of overgrowths, but textural relationships suggest that the low-Th/U zircon is the younger. The high-Th/U zircon commonly forms simply zoned grains that, in some cases, are overgrown by a thin layer of weakly luminescent (high-U) zircon. The reverse is never the case. If the leucogranodiorite formed from a mixed magma, then the peraluminous magma was the minor component. If the shift in zircon composition was due to an evolution in magma composition, then most of the zircon crystallization occurred before the magma became peraluminous and/or monazite began to crystallize.

The initial undersaturation of monazite may be explained by the interplay between Ca and Al upon crystallization (e.g. Dini *et al.*, 2004). An excess of Ca over Al stabilizes minerals such as apatite, and for this reason crystallization of apatite tends to occur early in metaluminous melts. Crystallization of monazite occurs early in originally peraluminous Ca-poor melts, but it may also occur at a late stage in melts that initially had high CaO/Al_2O_3, but became peraluminous after crystallization of Ca-rich phases such as An-rich plagioclase. This interpretation is consistent with the occurrence of $An_{35–53}$ plagioclase cores in the leucogranodiorite that have been interpreted as high-temperature early magmatic phases, in accordance with crystallization experiments and phase relation models for felsic magmas (Rottura *et al.*, 1993, and references therein). It might be argued that the grains with low-Th/U overgrowths are xenocrysts. If so, they were incorporated into the magma after most of the igneous zircon had crystallized, as no low-Th/U zircon was identified as cores. This contrasts with observations in some other cases where zircon grown during the metamorphism predating magmatism is measurably older than zircon crystallized from the magma (e.g. Zeck & Williams, 2002).

Inherited zircon

The trondhjemite

The low abundance of zircon in the trondhjemite is consistent with the low Zr content of the rock (142 ppm). The presence of inheritance, however, shows that either the magma was nevertheless zircon saturated or was too cool to dissolve pre-existing or assimilated zircon in the time between magma genesis and crystallization after emplacement. The zircon grains show no textural evidence that might suggest zircon loss by partial dissolution during metasomatism.

The large difference in the ages of the three dated cores and their range of shapes suggest that they are detrital zircon grains incorporated into the magma either at source or during emplacement. The presence of thick igneous overgrowths on the cores argues strongly that they were derived from the magma source. The scarcity of inheritance indicates that the sedimentary component in the magma was very small, or contained little detrital zircon, or that the initial magma was not saturated in zircon and much of the older zircon originally incorporated has been dissolved. The last option is unlikely, given the relatively small amount of zircon that eventually crystallized when the magma cooled and that, assuming complete melting, the trondhjemite magma would have been zircon saturated below *c.* 785°C (Watson & Harrison, 1983). A fourth possible explanation might be that the magma segregated early in the metamorphic history, as proposed by Zeh *et al.* (2003) for an inheritance-poor granite from Central Germany. Watt *et al.* (1996) and Rubatto *et al.* (2001) also have argued that melt extracted rapidly from the source area might be in disequilibrium with the restite and therefore low in Zr as a result of incomplete dissolution of older accessory phases. This scenario would fit with the relative ages of the trondhjemite and leucogranodiorite. It is not consistent, however, with the common observation that zircon precipitated from the partial melts of metasediments has low Th/U (e.g. Williams, 2001; Rubatto *et al.*, 2001), which is not the case for the igneous zircon in the Pizzo Bottino trondhjemite.

A major outcome of this work is the demonstration that inherited zircon is much less abundant in the trondhjemite (<10 vol. % of the total zircon) than in the leucogranodiorite (>50 vol. % of the total zircon). The difference between the amount and the age distributions of the inherited zircon in the 'older' trondhjemite and the 'younger' leucogranodiorite argues against the two granitoids being genetically related. On the contrary, it appears very likely that the two were derived from different source regions and/or under different melting conditions.

Three dated inherited zircon grains are not sufficient for any meaningful speculation on the origin of their source sediment. However, it is self evident that the source region for that sediment must have contained Palaeoproterozoic (*c.* 2·45 Ga), Neoproterozoic (*c.* 625 Ma) and Early Palaeozoic (*c.* 490 Ma) components.

The leucogranodiorite

Pre-magmatic zircon could be derived from the interaction between the granitic magma and its wall-rock during ascent or through late-stage contamination at the emplacement level, or it might be inherited from the magma source. Large zircon cores are so common in the Villa S. Giovanni leucogranodiorite and so thickly overgrown by igneous zircon that it is unlikely that they represent wall-rock contamination. Also, felsic to intermediate magmas do not have sufficient heat capacity to cause substantial wall-rock melting if intruded into cold crust. The very large amount of inherited zircon suggests that the magma was low temperature (the zircon saturation temperature of the magma would have been <715°C; Watson & Harrison, 1983), although Watson (1996, and references therein) has pointed out that the amount of zircon dissolved during partial melting, and the subsequent Zr content of the melt, is controlled by several other factors, such as the zircon content of the source, the extent to which zircon is armoured in stable restitic phases, the melting and extraction rate, and the melt composition. The high relative abundance of inherited zircon in the VSG leucogranodiorite shows that it was a 'cold' granite, both in the sense of Chappell *et al.* (1998) and of Miller *et al.* (2003). The magma temperature was probably less than 800°C. Miller *et al.* (2003) argued that 'cold' granites, as they defined them, are probably generated at temperatures too low for dehydration melting involving biotite or hornblende, and require fluid influx or abundant muscovite dehydration melting. In contrast, Chappell (2004) has emphasized the role played by source composition, and the dependence of magma temperature on the availability of the principal components of minimum melt, namely quartz, albite, K-feldspar and water. Low-temperature granites form only when the source contains sufficient of these components to produce a critical melt fraction (*c.* 35% partial melt) below *c.* 800°C. If it does not, then the temperature must rise further before the magma can mobilize.

The inherited zircon found in the VSG leucogranodiorite can be subdivided into five age groups: *c.* 460 Ma, 546·0 ± 8·6 Ma, *c.* 630 Ma, *c.* 870 Ma and *c.* 2·36 Ga. The estimate of 630 Ma is based on two analyses and the estimates of *c.* 460 Ma, 870 Ma and 2·36 Ga on single analyses. There are insufficient core analyses to allow specific identification of the source of the sediment, but there are enough to characterize the source in general terms and for a consideration of the petrogenetic and geological implications.

The broad range of ages found in the zircon cores may be explained by the melting of a source containing several zircon populations, and this in turn is consistent with a source containing metasediment. The source was not necessarily metasediment alone. The augen gneisses from southern Calabria, for example, contain zircon (inherited, igneous and recrystallized) with a comparable range of ages (Micheletti *et al.*, 2007). Even though felsic augen gneisses represent fertile source rocks for leucogranitic magmas (e.g Ollo de Sapo gneisses; Castro *et al.*, 1999, 2000), melt fractions approaching 5 vol. % can be produced by dehydration melting at 600 MPa and 800°C following total consumption of a large amount of muscovite. To obtain a melt fraction of *c.* 8 vol. %, a temperature of at least 850°C, leading to biotite melting, is required. The Calabrian augen gneisses contain muscovite only locally (Micheletti *et al.*, 2007), so are an unlikely source for a low-temperature granite magma. In addition, the REE pattern of the VSG leucogranodiorite is highly fractionated, suggesting derivation from a source containing garnet. Like muscovite, garnet occurs only locally in the augen gneisses. Finally, the ε_{Nd} values of the leucogranodiorite are strongly negative (−7·0 to −8·4), similar to the values for the metasedimentary rocks of southern CPO ($\varepsilon_{Nd} = -5{\cdot}4$ to −11·4; Rottura *et al.*, 1990, 1993), but lower than the values of −3·2 to −5·4 obtained for the Calabrian augen gneisses (Micheletti *et al.*, 2007).

A pelitic or semipelitic source is more consistent with the occurrence of exclusively metapelitic enclaves, and with the petrographic, major and trace element, and isotopic features of the leucogranodiorite. It also is more conducive to low-temperature melting. Further, muscovite dehydration melting of a metasedimentary source can be envisaged as the main process responsible for the production of a felsic monazite-bearing magma. This reaction has been shown to involve dissolution of apatite and, through raising the P_2O_5 and LREE contents of the melt, crystallization of monazite (Zeng *et al.*, 2005). The above petrogenetic considerations are at variance with the mixed mantle–crust origin for the VSG granitoids proposed by Rottura *et al.* (1993). On the other hand, a purely crustal origin for the leucogranodiorite agrees with the model, based on geochemical, isotopic and mass balance data, recently proposed by Caggianelli *et al.* (2003) for the coeval two-mica leucogranites from the Sila Massif (northern Calabria), and with the earlier models of Del Moro *et al.* (1982) and Rottura *et al.* (1990).

The distinct 546·0 ± 8·6 Ma inherited component corresponds in age to widespread granitic magmatism and metamorphism in many terranes of southern Europe that was related to the late collisional stages of the Pan-African orogeny at the northern Gondwanan margin, or to the transition to a passive continental margin (Villeneuve & Cornée, 1994; Zulauf *et al.*, 1999; Murphy *et al.*, 2001; Linnemann *et al.*, 2004; Zeck *et al.*, 2004; Gasquet *et al.*, 2005; Micheletti *et al.*, 2007).

The age of *c.* 630 Ma is also common among the Pan-African terranes. It is considered to mark a major period of igneous activity related to subduction and arc construction (e.g. Linnemann & Romer, 2002), or to ocean closure and arc–continent collision (Gasquet *et al.*, 2005) or to late collisional stages (Villeneuve & Cornée, 1994).

These Neoproterozoic inheritance ages suggest a sedimentary source produced by erosion of a Pan-African orogenic belt situated at the northern Gondwana margin in the late Neoproterozoic or early Palaeozoic. Zircon with ages of 650–550 Ma dominates the early Palaeozoic sediments of Gondwana, particularly those originating from North Africa (e.g. Avigad *et al.*, 2007). This fits very well with the U–Pb ages measured by Schenk & Todt (1989) and Schenk (1990) on detrital zircon from an unmetamorphosed (probably Devonian) siltstone at Stilo (Stilo Unit, southern Calabria). They found essentially a two-component mixture, mainly of Pan-African age (550 ± 50 Ma) with some Archaean ages (*c.* 2·5 Ga). The new results reported here, together with Neoproterozoic to Early Cambrian zircon U–Pb ages obtained for the igneous protoliths of the augen gneisses from Calabria (Micheletti *et al.*, 2007), suggest that the Pan-African belts in North Africa, possibly including the *c.* 550 Ma granites of southern Calabria, were the principal source area for the sediments.

The single Palaeoproterozoic detrital zircon age (*c.* 2·36 Ga) is consistent with previous SIMS and TIMS zircon U–Pb data (Schenk, 1990; De Gregorio *et al.*, 2003; Trombetta *et al.*, 2004; Micheletti *et al.*, 2007) that indicate the presence of Palaeoproterozoic components (probably detrital) in the basement of the south Italian sector of the Hercynian belt.

The age of *c.* 870 Ma could provide information on the possible Amazonian or West African provenance of the sediments, as the presence or absence of 1·1–0·9 Ga (Grenvillian) zircon has been proposed as one of the most important criteria for the recognition of a West African or Amazonian provenance of peri-Gondwanan terranes (Friedl *et al.*, 2000; Linnemann *et al.*, 2004). Sediments with a West African provenance are characterized by a predominance of 3·4–2·8 and 2·2–1·8 Ga zircon from the older basement, and the absence of Grenvillian (*c.* 1·2–1·0 Ga) zircon (e.g. Nance & Murphy 1994). In particular, Linnemann *et al.* (2004) pointed out that the presence of a *c.* 900 Myr gap (*c.* 1·7–0·8 Ga) in the igneous activity in the West African section of the northern Gondwana margin is considered by many workers to be the best fingerprint for the West African provenance of the Precambrian basement of peri-Gondwanan terranes detected in the circum-Atlantic Palaeozoic orogens.

On the other hand, *c.* 1·0 Ga zircon is commonly considered to indicate provenance from Grenvillian or Sunsas–Rondonian orogens of the Amazon cratons. Zeck *et al.* (2004) assumed a possible provenance from the Amazon craton for zircon from the Piedrahita orthogneiss (western–central Iberia), based on the dating of two zircon grains at *c.* 980 and *c.* 830 Ma. It is necessary to remember, however, that Grenvillian components have also been recognized in Palaeozoic sediments of the Arabian Platform, suggesting that the occurrence of *c.* 1·0 Ga zircon is not necessarily related to Amazonian provenance (Linnemann *et al.*, 2004, and references therein). It may also indicate a potential Arabian or East African source area for the crustal sequences of some peri-Gondwanan terranes. The *c.* 870 Ma inherited zircon in the VSG leucogranodiorite therefore might represent a late Grenvillian component of the CPO basement, as already indicated by the 1154 ± 10 Ma inherited zircon found in meta-andesites from the Lower Domain of the Peloritani Mountains (Trombetta *et al.*, 2004). Micheletti *et al.* (2007), however, still favoured a West African provenance for the source material of the Pan-African granitoids of Calabria. The most likely source of sediment that contains abundant Pan-African but very little Grenvillian zircon is the Early Palaeozoic sediments of North Africa, which possibly have an ultimate origin in East Africa (Williams *et al.*, 2002). This issue remains to be resolved.

The single *c.* 460 Ma zircon age, if there was no radiogenic Pb loss, shows that at least some of the source sediment for the peraluminous granites was Mid-Ordovician or younger. Crust-forming events of Caradoc age are documented in the southernmost sector of the CPO (south Peloritan Mountains), where they are represented by 456–452 Ma orogenic andesites to rhyolites (Trombetta *et al.*, 2004). Ordovician zircon cores have been reported from the Calabrian augen gneisses (Micheletti *et al.*, 2007) but, as the overgrowths in the same zircon suites are of Late Neoproterozoic age, these Ordovician dates must be underestimates as a result of radiogenic Pb loss. The presence of the Mid-Ordovician component in the source of the VSG leucogranodiorite might reflect derivation of some of the source sediment from the Ordovician of the south Peloritan Mountains or, more simply, from the same Pan-African Calabrian granitoids already indicated as main components of the source area.

CONCLUSIONS

SHRIMP U–Pb dating of zircon from the Pizzo Bottino trondhjemite shows that trondhjemites might represent the initial stage of Hercynian peraluminous plutonic magma production in the Calabria–Peloritani segment of the Hercynian Belt. The emplacement age of 313·7 ± 3·5 Ma obtained for the trondhjemite predates the bulk of the late Hercynian plutonism in both the southern and northern CPO by about 14 Myr. Moreover, the contrasting zircon inheritance patterns in the studied trondhjemite and leucogranodiorite suggest an independent genesis. The small amount of inherited zircon in the trondhjemite is in marked contrast to the ubiquitous occurrence and large size of inherited cores in zircon from the leucogranodiorite. The two rock types probably originated from different source regions and/or under different melting conditions. The presence of a sedimentary component in the source of the trondhjemite magma is suggested by the dispersed ages of the zircon cores, their irregular shapes and the thick overgrowths of igneous zircon. The paucity of inherited zircon, however, most probably indicates that the sediment component in the magma was either very small or zircon poor. The SIMS zircon dating of a Peloritanian trondhjemite has provided important new information, but the petrogenesis of the trondhjemites remains problematic, mainly because of their intense post-emplacement alteration (Fiannacca *et al.*, 2005).

Emplacement of the Villa S. Giovanni leucogranodiorite has been dated at 300·2 ± 3·8 Ma. This first high-precision SIMS age for Hercynian igneous rocks of the CPO agrees well with the recent ID-TIMS monazite and xenotime intrusion ages of strongly peraluminous granites from Calabria (Graessner *et al.*, 2000). The measured age confirms that the late Hercynian strongly peraluminous magmatism in southern Calabria resulted in the nearly simultaneous intrusion, at *c.* 300 Ma, of several relatively small discrete plutons. In addition, the presence in the leucogranodiorite of two igneous zircon generations differing in texture and composition has been interpreted as the result of crystallization from melts of different compositions, one monazite undersaturated (?metaluminous) and the other monazite saturated (?peraluminous). These were derived either from different magma batches or, more probably, through the fractional crystallization of a single magma. This question might be solved in future by Hf and/or O isotopic analysis of the high- and low-Th/U zircon growth phases.

The VSG leucogranodiorite contains abundant inherited zircon of Palaeoproterozoic and Neoproterozoic age, consistent with its derivation from a metasedimentary source. The inherited zircon can be divided into five age groups: *c.* 2·36 Ga, *c.* 870 Ma, *c.* 630 Ma, 546·0 ± 8·6 Ma and *c.* 460 Ma. Except for the age of *c.* 870 Ma, which might be exotic to the North African craton, the ages measured suggest a North African provenance for the sedimentary source of the leucogranodiorite, particularly considering that the most common ages of *c.* 630 Ma and 546·0 ± 8·6 Ma are the dominant detrital zircon ages in the early Palaeozoic sediments of North Africa.

The main inherited component, at 546·0 ± 8·6 Ma, is the same age as the granitic protoliths of augen gneisses from southern Calabria (Micheletti *et al.*, 2007), which also share with the Villa S. Giovanni leucogranodiorite a comparable set of inherited and recrystallized zircon components. As direct derivation of the leucogranodiorite through partial melting of the augen gneisses appears (on petrographic, geochemical and Nd isotopic evidence) to be unlikely, much of the detritus in the metasedimentary source of the leucogranodiorite was possibly derived from the erosion of Pan-African granitoids similar to those in the southern Calabria–Peloritani Orogen.

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