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Geochronological constraints for the time of metamorphism in the Gruf Complex (Central Alps) and implications for the Adula-Cima Lunga nappe system

Anthi Liati[1] and *Dieter Gebauer*[1]

Abstract

The Gruf Complex, Central Alps, consists of high-grade, upper amphibolite- to granulite-facies metamorphic rocks. Previous petrological studies of sapphirine-bearing granulites indicate maximum P–T conditions of ca. 10 kbar, 830 °C for the granulite-facies metamorphism, applicable to the whole Gruf Complex. We dated by SHRIMP zircon domains from a sapphirine-bearing granulite in Val Codera. The rock consists of a major dark-colored part (restite) and a minor light colored part (leucosome). Zircon occurs in both parts as relatively large, euhedral and prismatic crystals. Cathodoluminescence (CL) imaging reveals a quite uniform zircon population, identical in both the restitic part and in the leucosomes, with large oscillatory (magmatic) domains, often surrounded by metamorphic rims with bright, homogeneous CL-characteristics. No inherited cores were found inside the oscillatory domains. Five spots on the oscillatory domains of zircons from both the restite and the leucosomes yielded a weighted mean age of 272.0 ± 4.1 Ma (95% c.l.), interpreted to reflect the crystallization time of the magmatic protolith. The consistent Permian ages obtained from different zircon crystals, together with the large crystal size, the uniformity in zircon morphology and CL-characteristics, as well as the lack of inherited cores point to a magmatic protolith. This is at variance with the generally accepted view that these rocks are of metapelitic origin. Six spots from the metamorphic rims yielded a weighted mean age at 32.7 ± 0.5 Ma. This age is identical to granulite-facies ages of crustal and mantle rocks in different parts of the Adula-Cima Lunga nappe system (e.g., Alpe Arami, Cima di Gagnone). The new SHRIMP data strongly favour the view that the sapphirine-bearing granulites are restites formed during Alpine partial melting of Permian granitoids. Moreover, they are in agreement with the previous view that the Gruf Complex belongs to the Adula-Cima Lunga nappe system, i.e. that it is part of the European margin. Thus, this area of the Central Alps compares well with the geotectonic position of the Dora Maira and the Monte Rosa nappes in the Western Alps, which have recently been interpreted to belong also to the European margin, tectonically emplaced from below and to the south of the Briançonnais nappes.

Keywords: Gruf Complex, Adula nappe, Central Alps, SHRIMP-dating, U–Pb geochronology, zircon, granulites.

1. Introduction

The metamorphic rocks of the Central Alps occur in a series of nappes originating from the Adriatic plate in the south, the European plate in the north and intervening ocean basin(s) and micro-continent(s). The rocks formed during subduction/collision and exhumation processes of Hercynian basement and its largely Mesozoic sedimentary cover rocks during the Alpine orogeny, i.e., since the Late Cretaceous (e.g., Gebauer, 1999). The geodynamic reconstruction of the different nappes depends largely on reliable geochronological data of the metamorphic rocks. An effective geochronological technique used to unravel the complicated geological history of these rocks (including successive magmatic, sedimentary and multiple metamorphic episodes) is the cathodoluminescence (CL)-controlled ion microprobe dating (SHRIMP) on distinct zircon domains. The effectiveness of this technique relies mainly on the fact that zircon commonly forms multiple domains which correspond to different stages of its growth/recrystallization history, as Pb diffusion in crystalline zircon seems to be negligible, at least for most crustal temperatures (e.g., Lee et al., 1997; Cherniak and Watson, 2000). The different domains of the zircon crystals can be identified by imaging its internal structure in CL. These domains can be individually dated by SHRIMP even if they are as narrow as 10 µm. Thus, in rocks metamorphosed under upper amphibolite-facies conditions or higher (see section 6) one expects to obtain at least two ages by SHRIMP dating of zircon: the age of the protolith, preserved in the core domain of the zircon

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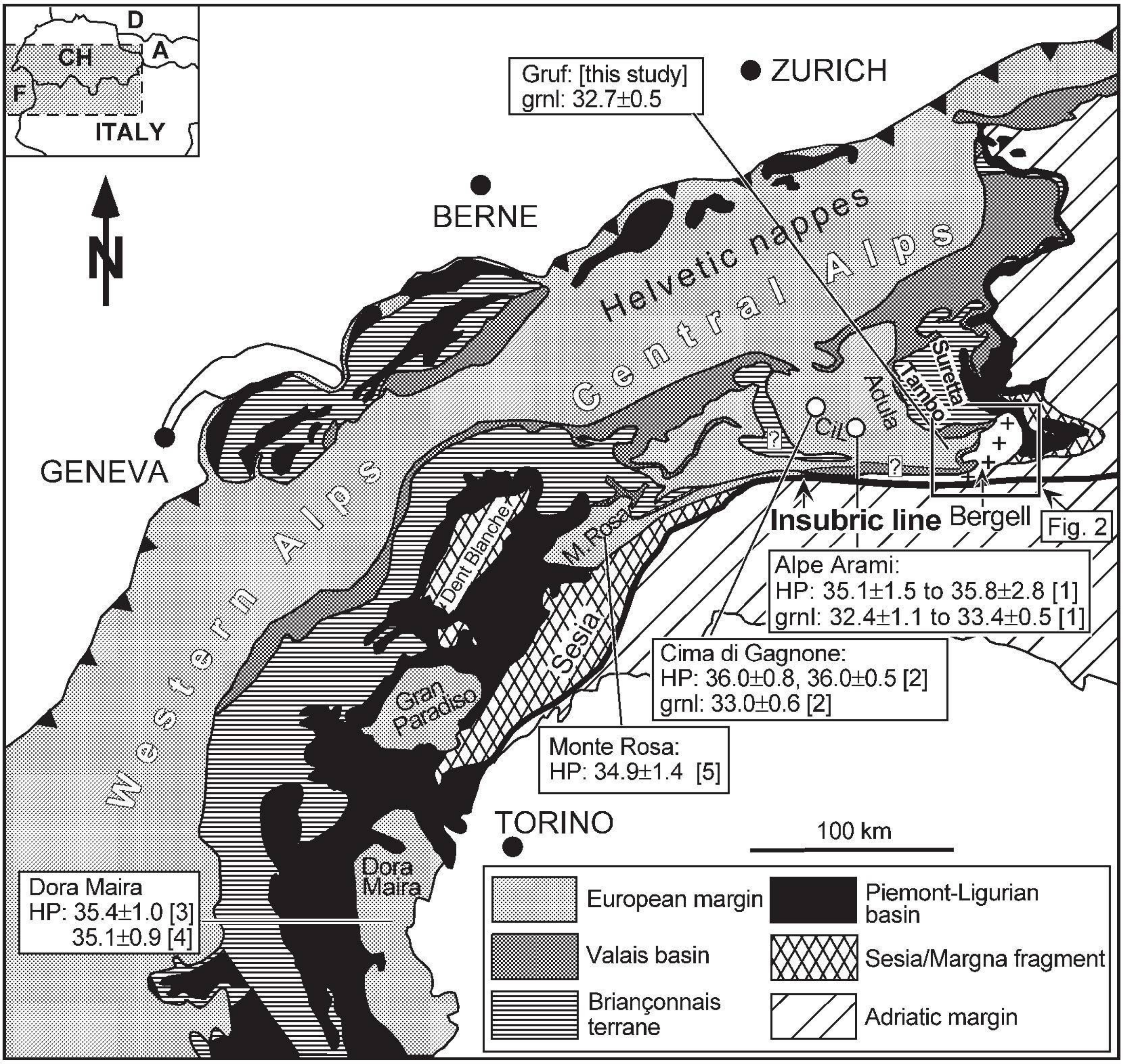

Fig. 1 Map of the Western and Central Alps showing the different paleogeographic units (modified from Froitzheim et al., 1996. Units of disputed origin in the Central Alps are denoted with a question mark). Upper Eocene–Lower Oligocene HP- and granulite-facies metamorphic ages are summarized in this map (see text). Based on a series of radiometric data (see summary by Gebauer, 1999) and on a recently suggested tectonic model by Froitzheim (2001), the internal massifs, Dora Maira, Gran Paradiso, and Monte Rosa, are given here as having a European provenance. Abbreviations: grnl—granulite; CiL—Cima Lunga. (1) Gebauer (1996); (2) Gebauer (1994); (3) Gebauer et al., (1997); (4) Rubatto and Hermann (2001); (5) Rubatto and Gebauer (1999).

crystal, and the age of metamorphism, obtained by dating metamorphic domains usually developed at the outermost part of the crystal.

An important factor that may lead to disturbance of the isotopic systems applied in several mineral chronometers to infer metamorphic ages is the presence of post-metamorphic fluids. This seems to be the case, at least for part of the metamorphic rocks in the broad area of study (Adula-Cima Lunga nappe and Gruf Complex), which is situated in the so-called "Lepontine area" of the Central Alps. The term "Lepontine" has been introduced to describe the high-grade Tertiary metamorphism of rocks in the area between the Gotthard massif in the north and the Insubric line in the south, and between the Simplon area in the west and the Bergell area in the east (e.g., Wenk, 1956; Trommsdorff, 1966; Frey and Ferreiro Mählmann, 1999 and references therein). The Lepontine metamorphism was believed to have taken place at ca. 38 Ma (Jäger, 1973), referring to both the southern European margin and the Briançonnais terrane. The latter represents the northeastern, peninsula-like part of the Iberian microconti-

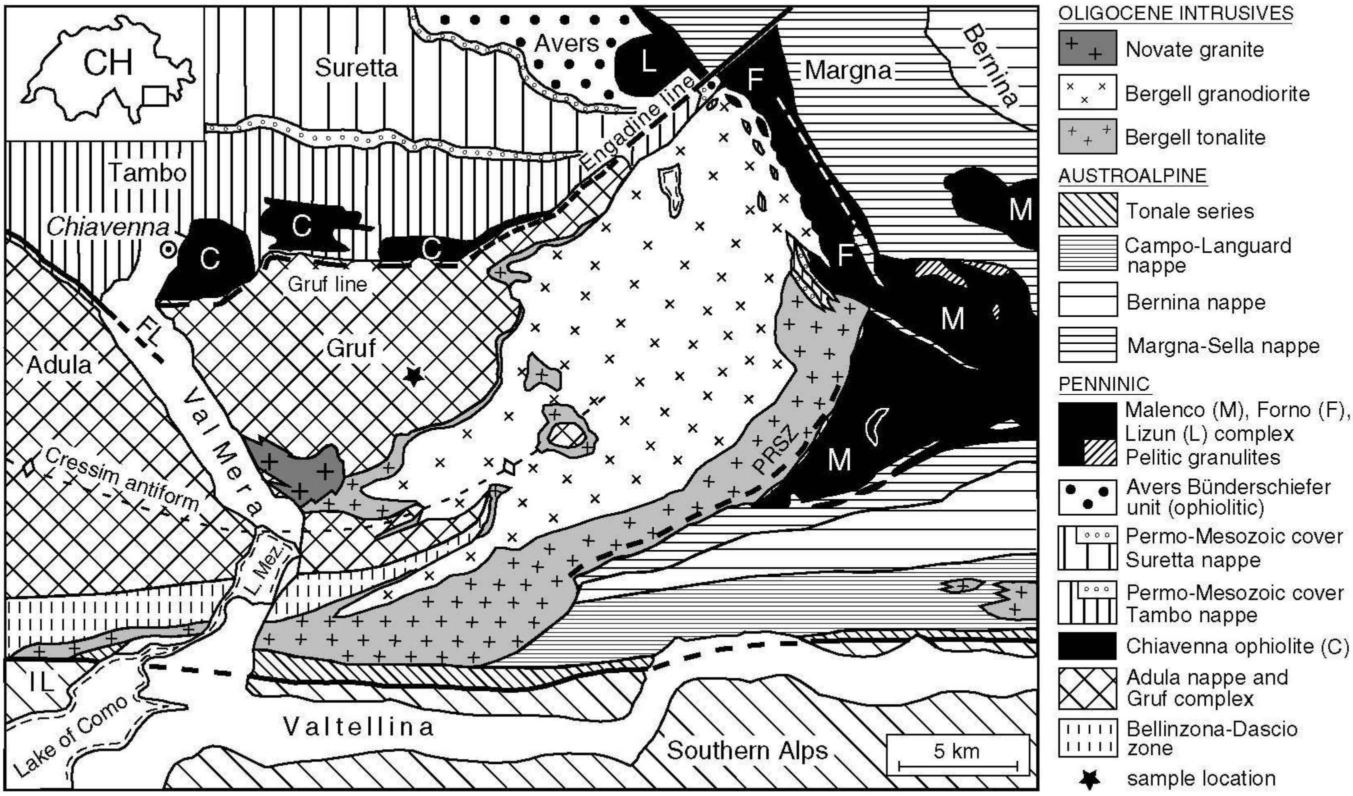

Fig. 2 Tectonic map of part of the Central Alps, with the Gruf Complex (modified from Schmid et al., 1996a). FL—Forcola Line; IL—Insubric Line; PRSZ—Preda Rossa Shear Zone; C—Chiavenna; F—Forno; M—Malenco; L—Lizun.

nent, which also includes Sardinia and Corsica (e.g., Stampfli, 1993). Recent geochronological work revealed a temperature peak at ca. 33 Ma for units belonging to the southern European margin (see summary by Gebauer, 1999). The presence of high amounts of fluids, connected to the intrusions of the Bergell and Novate plutons, contributed substantially to fluid induced recrystallization in the regional metamorphic rocks (e.g., Villa, 1998; Liati et al., 2000), which led to partial or complete resetting of ages. As a consequence, many of the existing geochronological data and interpretations have no clear geological significance and are therefore inconclusive.

The area of study, the Gruf Complex (Figs. 1 and 2), is characterized by high-temperature (HT) regional metamorphism and intense migmatization. A well-established age for the HT metamorphism of this unit is still missing. Within the framework of the present paper, we dated by SHRIMP zircon domains from a sapphirine-bearing granulite of the Gruf Complex. Our primary aim was to determine the age of metamorphism and examine how it correlates with that of metamorphic rocks from the Adula nappe and other nappes of the Alps. An additional goal of our work was to clarify (by means of zircon size, morphology, CL-characteristics and age uniformity) whether the protolith of the sapphirine-granulites, generally characterized as "metapelites", was indeed a sedimentary rock or if it was of igneous origin. In the case of a sedimentary protolith, we would determine the age of the youngest detrital zircon and obtain the minimum age for the sedimentation, and if the protolith were an igneous rock, we would determine the age of crystallization.

2. Geological setting

The Gruf Complex (GC), Central Alps, occurs east of Val Mera and is structurally situated below the Bergell pluton (Fig. 2). It consists of high-grade, upper amphibolite- to granulite-facies metamorphic rocks, mainly migmatitic quartz-feldspar gneisses. Pelitic schists and gneisses, as well as amphibolites and calc-silicate marbles are subordinate. The GC has been correlated with the zone of Bellinzona (Wenk, 1973). However, in most of the recent interpretations (e.g., Schmid et al., 1996a; Frey and Ferreiro Mählmann, 1999) the GC is considered as SE continuation of the Adula nappe, which is exposed west of Val Mera (Fig. 2). The Adula-Cima Lunga nappe, the largest basement nappe in the eastern part of the Central Alps, consists of pre-Mesozoic orthogneisses, pelitic schists and metabasic/ultrabasic rocks, as well as metamorphosed Mesozoic marls, limestones, dolomites and sandstones (see e.g., Frey and Ferreiro Mählmann, 1999 for a summary and references). Especially in the upper parts of the Adula nappe and the corresponding parts of the thinner Cima Lunga nappe, strong imbrication of different tectonic slices is observed and interpreted as "lithospheric mélange" (Trommsdorff, 1990). A different early P–T evolution has been invoked for different tectonic slices in this part of the Southern Steep Belt (Engi et al., 2001). According to Schmid et al. (1996a, b) the original paleogeographic position of the Adula-Cima Lunga nappe is at the distal European margin, north of the Valais ocean. This view is supported also by geochronological arguments (e.g., Gebauer, 1999), as well as by recent geodynamic reconstructions (Froitzheim, 2001). Metamorphism in the Cima Lunga nappe (Alpe Arami and Cima di Gagnone) has been dated at the uppermost Eocene (ca. 35 Ma: high-pressure) to lowermost Oligocene (ca. 33 Ma: granulite-facies; Gebauer, 1994, 1996).

Regarding the time of metamorphism of the GC, different opinions exist, based mainly on radiometric and structural/geological data from the Gruf Complex itself and from adjacent metamorphic rocks, as well as from the neighboring plutonic rocks of Bergell and Novate. The few radiometric data do not lead to a clear and straightforward conclusion for the time of metamorphism. This is partly true due to poor, insufficient and scattered Rb–Sr data, lack of correction for disequilibrium and, as a result of this, lack of demonstration of concordancy in the case of the early monazite data, as well as problems with the interpretation of the "ages". Gulson (1973), based on whole-rock Rb–Sr data, interpreted the migmatites as reflecting Hercynian migmatization. Similarly, Huber (1999) and Huber et al. (1999) suggest pre-Alpine (late Permian to Early Mesozoic) ages for the granulite-facies metamorphism in the GC, based on structural/deformational arguments. Hänny et al. (1975) report Rb–Sr whole-rock and U–Pb monazite data for migmatites from Val Bodengo, on the western side of Val Mera (southern Adula nappe), generally assumed to have the same origin and metamorphic history as those of the GC. These authors claim that the migmatites are of Paleozoic age and have experienced high-grade metamorphic overprint during the Tertiary (thermal maximum at about 20–25 Ma). An Alpine age of ca. 38 Ma has been suggested for the peak, granulite-facies metamorphism (Droop and Bucher-Nurminen, 1984), probably because at that time, the Lepontine event was believed to have taken place at ca. 38 Ma (Jäger, 1973). The same authors report an age of ca. 30 Ma for the time when the metamorphic rocks were at depths corresponding to ca. 4–5 kbar. Migmatization has been considered as contemporaneous or slightly post-dating the final emplacement of the Bergell pluton (Davidson et al., 1996; Berger et al., 1996). It is noted that ages of 31.88 ± 0.09 Ma and 31.5 ± 0.35 Ma are reported by von Blackenburg (1992) for the older parts of the Bergell pluton (tonalite) and 30.03 ± 0.17 Ma for the younger ones (granodiorite). An Alpine age of the migmatites has been inferred also on the basis of structural arguments (Hafner, 1993).

3. Sample description and petrological data

3.1. Sample description

The sapphirine-bearing granulites that we dated here by SHRIMP were discovered by Cornelius (1916) in form of blocks, east of Bresciadega, in Val Codera (Fig. 2). Wenk et al. (1974) report another locality of these rocks (again in form of blocks) in Val Codera (southern end of Val Conco). Our sample derives from a block in the river gravel (locality with Swiss coordinates 761'40/125'40). The "in situ" outcrop must be in the inaccessible rock cliffs at the catchments area of this local creek. This rock-type received special attention and has been petrologically studied by Ackermand and Seifert (1969), and in great detail by Droop and Bucher-Nurminen (1984).

The rock sample studied here is medium- to coarse-grained, predominantly dark-colored with a minor light-colored part forming a fine, irregular network with local development of small "pockets", up to 2–3 mm across. It does not have the typical appearance of a banded migmatite, in that no leucosome-melanosome-paleosome relationships are observed. The present rock probably represents the refractory product of advanced melting (restite) of a rock, from which the major part of the extracted melt was transported away. Only a small part of the melt crystallized "in situ" giving rise to the few and small leucosomes. This idea is in agreement with the Mg–Al rich, refractory nature of these granulites (see also Droop and Bucher-Nurminen, 1984; Clemens, 1990). The dark-colored, main part of the rock consists predominantly of garnet (commonly up to 1 cm large), orthopyroxene, biotite, sapphirine, spinel and many very fine-grained symplectites in form of coronae, mainly around garnet (see Droop and Bucher-Nurminen, 1984). The nature of the minerals taking part in the symplectites could not be defined here microscopically, due to their very fine grain size. The light-colored minor part consists mainly of K-feldspar, plagioclase, quartz, disseminated biotite, and minor cordierite.

Zircon is a relatively abundant accessory mineral and has been observed as inclusion in sapphirine, orthopyroxene, garnet, biotite, spinel and the symplectites of the dark-colored part, as well as inside plagioclase and K-feldspar of the light-colored part of the rock. As discussed later (section 7), zircon seems to have been already present in the precursor of the granulites and was partially recrystallized during high temperature metamorphism, thereby developing metamorphic rims, which we dated here by SHRIMP.

It is worth mentioning that the term "pelite", which implies a sedimentary protolith is widely used to define the granulites studied (e.g., Droop and Bucher-Nurminen, 1984; Huber, 1999). However, as discussed in the present paper, these rocks very likely originated from an igneous precursor (see sections 4 and 7).

3.2. Petrological data

The petrology of the granulites dated here has been studied by Droop and Bucher-Nurminen (1984). This detailed petrological work provides a good basis for the present study. Therefore, our own work in these terms is restricted to the petrographic description of the dated sample, as given above.

Droop and Bucher-Nurminen (1984) suggest that partial melting processes took place during metamorphism of the presumed metapelitic precursors of the granulites and that the actual granulites represent restites after extraction and migration of high amounts of melt from the original rock. This view is in accordance with recent studies on anatectic migmatites (e.g., Watt et al., 1996; Kriegsman, 2001; White and Powell, 2002), which suggest absence of retrogression in the granulites due to melt extraction and migration. In the studied rocks from the GC, features of retrogression are not observed. Maximum metamorphic conditions of ca. 10 kbar, 830 °C are given by Droop and Bucher-Nurminen, who suggest that these P–T conditions apply for the whole GC, since there is good evidence that partial melting occurred "in situ" (e.g., absence of late stage tectonic emplacement of the sapphirine-bearing granulites, absence of shearing along restitic granulite and leucosome). This suggestion is in line with the metamorphic history of metapelitic gneisses and ultramafic rocks of the GC (Bucher-Nurminen and Droop, 1983). Formation of the granulites is attributed to a single metamorphic cycle (Droop and Bucher-Nurminen, 1984) and is related to the "Lepontine" event, which at that time was believed to be at ca. 38 Ma (Jäger, 1973). Re-evaluation of the metamorphic conditions for the granulites, based on more recent mineralogical and petrological data of the participating mineral phases (e.g., Harley, 1998 and references therein) may result in some deviation from above P–T conditions. However, this possible discrepancy would have no influence on the interpretation of the age data of the rock dated in this study.

3.3. Criteria for the selection of the sapphirine-granulite for dating the granulite-facies metamorphism in the GC

In view of the different rock types that occur in the GC and their metamorphic history, we would like to emphasize here the reasons why we chose the sapphirine-granulite for dating: (1) Leucosomes within migmatized ortho- or paragneisses do not always contain syn-genetic zircon crystals (or crystal domains), i.e. crystallized newly within the partial melt. However, metamorphic zircon domains are ubiquitous in granulites and thus can be dated easily by SHRIMP (e.g., Vavra et al., 1999; Schaltegger et al., 1999). (2) The restitic nature of the sapphirine-bearing granulites and the fact that they remained "dry" during decompression (absence of retrogression; Droop and Bucher-Nurminen, 1984) are very good conditions for the preservation of the first, high-grade age signature in the metamorphic zircon domains. Thus, in these rocks we would not expect age resetting (e.g., by Pb-loss), due to post-metamorphic fluid circulation and/or reheating related to later plutonic activity (Bergell and Novate plutons, see also below). This is different for rocks containing hydrous minerals: partial Pb-loss and age resetting to various degrees due to the presence of high amounts of fluids have been observed in a metabasic rock (metagabbro) of the neighboring Chiavenna unit, at the immediate northern contact to the GC (Prata area; Liati et al., 2004). Thus, with the choice of a dry granulite sample, the possibility of multiple younging of ages at 32–30 Ma (age of the Bergell tonalite and granodiorite) and/or at 24–25 Ma (age of the Novate granite), and/or after the intrusion of the Novate granite (Liati et al., 2000) can be minimized. (3) Previous SHRIMP work on zircons from a leucosome (both newly crystallized oscillatory zoned, as well as granulite-type zircons or zircon domains) at Alpe Arami and from a paragneiss close to peridotites at Cima di Gagnone (both Cima Lunga nappe) were interpreted to reflect granulite-facies conditions at 32.4 ± 1.1 Ma and 33.0 ± 0.6 Ma, respectively (Gebauer, 1994, 1996). At Alpe Arami, part of the dated zircons (oscillatory zoned) crystallized newly in the leucosome, while at Cima di Gagnone the morphology and CL-characteristics of the metamorphic zircon domains are of granulite-type (e.g., Vavra et al., 1999). Apart from the CL-evidence that the metamorphic zircon domains are of granulite-type, characteristic minerals indicating a granulite-facies stage are not reported for these high-grade quartzofeldspathic rocks. It was therefore of special interest for us to compare the existing SHRIMP data from the Cima Lunga nappe with the age of granulite-type zircon domains from the sapphirine-bearing Gruf rocks, where granulite-facies metamorphism is petrologically well documented.

4. Description of the zircons, cathodoluminescence characteristics and interpretation

We separated zircons from both the dark- (restite) and light-colored (leucosome) fractions of the mineral concentrates of these granulites, in order to examine whether newly formed zircon crystals are present in the leucosomes and if so, whether these zircons would show any difference in metamorphic age to the ones from the restite. The morphological, as well as the CL characteristics of the zircon crystals, however, proved to be identical for both the restitic part and the leucosomes. Zircons are usually large, ranging between ca. 400–170 µm in length and ca. 160–40 µm in width. They are commonly elongate, prismatic, euhedral to subhedral with rounded outlines, a common feature in granulite-facies rocks (e.g., Vavra et al., 1999). The large size and abundance of zircon crystals already preclude a pelitic origin for these granulites, since in pelites zircons, if present, are commonly very small (<30 µm). CL-imaging revealed a very uniform zircon population. All crystals consist of a large oscillatory zoned inner domain with closely spaced lamellae (Fig. 3), typical for zircons precipitating in a melt. No inherited cores (either detrital or formed during previous melting events) were found within these oscillatory zoned domains. The presence of abundant zircons with uniform magmatic-type CL-pattern and morphology also suggests that the protolith of this rock was igneous and not sedimentary, as previously assumed. This view is further strengthened by the consistency of the protolith ages obtained from the oscillatory zoned domains (see below).

The large oscillatory zoned domains are, in most cases, surrounded by rims that range in thickness from several µm to ca. 50 µm and show bright, homogeneous CL intensitiy (Fig. 3). The rims probably derived from recrystallization processes during metamorphism. This interpreta-

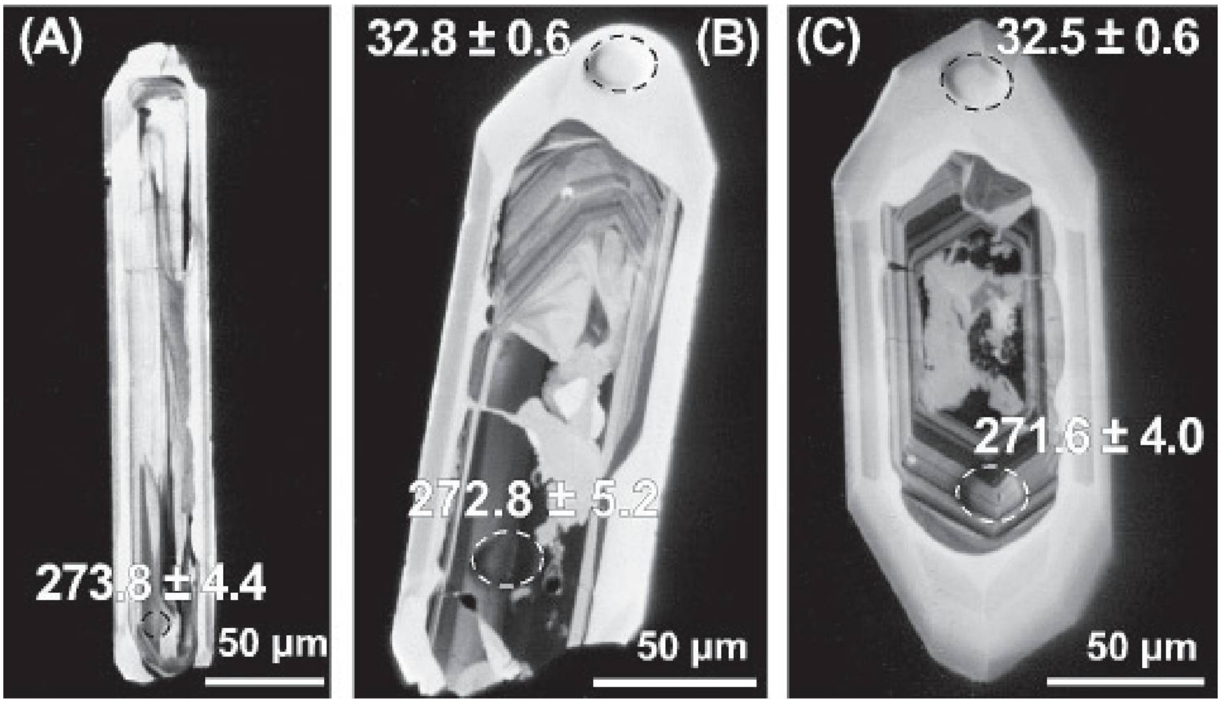

Fig. 3 Cathodoluminescence (CL) pictures of representative zircon crystals from the sapphirine-bearing granulite (A, B from the restitic parts; C from the leucosome) of the Gruf Complex. (A) zircon consists entirely of a magmatic domain of Permian age with no metamorphic rim. In (B) and (C), a relatively broad metamorphic rim (bright in CL) is observed around the co-magmatic, oscillatory zoned domain. Note the bright CL domain in the middle/lower part of crystal (B) possibly formed by recrystallization due to fluid circulation along fractures through the crystal. Note also the ghost oscillatory zoning in the rim domain of crystal (C), indicating that the original size of the magmatic crystal extended to the actual size of the crystal and that the metamorphic rim formed by recrystallization, and not by new zircon growth.

tion is based mainly on the preservation of the oscillatory zoning pattern in most rim domains, which is "inherited" from the original igneous crystal. This pattern is visible in some parts of the metamorphic rims as "ghost oscillatory zoning", with alternating dark and light bands, parallel to those preserved in the igneous core (e.g., Fig. 3C: left and right parts of the rim). Mineral inclusions were not observed in the zircon rims, which is in line with the interpretation that the metamorphic domains probably formed by recrystallization of preexisting zircon crystals. Newly grown zircon crystals could not be observed in the leucosomes of the dated granulite, in agreement with previous observations reported for other leucosomes in different units of the European Hercynides (Gebauer et al., 1989) or for a migmatite suite of the Kirtomy nappe, Scotland (Watt et al., 1996). In other cases, however, co-genetic entire zircon crystals or overgrowths within the leucosomes of migmatized leucocratic gneisses are described (e.g., Gebauer, 1996; Oliver et al., 1999; Liati and Gebauer, 1999; Ordoñez Casado et al., 2001).

The fact that zircons from both the restitic part and the leucosomes of the sapphirine-granulites are identical in terms of morphology, internal structure as imaged by CL and, as shown later, also in regard to protolith and metamorphic age, leads us to the following interpretation: zircon crystals were widespread in the igneous precursor of the granulite and were subsequently recrystallized, as soon as the rock reached partial melting and granulite-facies conditions (see also section 6). Some of the zircons were left behind in the restitic fraction of the rock, while others were extracted from the original rock together with the melt.

5. Analytical techniques and data evaluation

5.1. Analytical techniques

The SHRIMP-data presented in this paper were obtained on SHRIMP II at the Geological Survey of Canada in Ottawa, following the standard operating techniques. A spot size of ca. 20 µm was used for the measurements. For data collection, 7 scans through the critical mass range were made. For further details on the SHRIMP technique see Stern (1997).

All CL pictures were photographed from a split screen on a CamScan CS 4 scanning electron microscope (SEM) at ETH, Zurich, operating at 13 kV and equipped with an ellipsoidal mirror located close to the sample within the vacuum chamber, in order to increase the CL-signal. In general, weak CL-emission (dark-colours in the picture) reflects high amounts of minor and trace elements, strong CL-emission (light colours in the picture) reflects low amounts of minor and trace elements, including Uranium. Thus, the U contents can be roughly predicted via CL.

CL images of the zircon crystals were obtained before SHRIMP-dating, in order to distinguish between different domains and avoid mixing of ages, as well as after the SHRIMP-analyses, to check the exact location of the SHRIMP-spot.

5.2. Data evaluation

For the calculation of the Pb–U ratios, the data were corrected for common Pb using the ^{207}Pb correction method. The data are graphically presented on Tera-Wasserburg (TW) diagrams (Tera and Wasserburg, 1972), where total ^{207}Pb/^{206}Pb vs. the calibrated total ^{238}U/^{206}Pb is plotted. For the data presentation, we used the program "Isoplot/Ex" by Ludwig (2000). In the diagrams, the error axes on the ellipses are at the 2σ level.

The amount of common Pb was calculated using the isotope composition of common Pb obtained from the model of Cumming and Richards (1975). The mean ages are given as weighted mean; the error of the weighted mean is at the 95% confidence level. For single analyses (Table 1), the 1σ error is given.

6. Results of SHRIMP dating and interpretation

Nine data points were obtained from the oscillatory zoned core domains of eight different zircon crystals (Table 1). On a TW diagram, five analyses plot on a mixing line with common Pb and calibrated total ^{238}U/^{206}Pb as end members intersecting the concordia at a weighted mean age of 272.0 ± 4.1 Ma (Fig. 4A). This age is interpreted to reflect the crystallization time of the magmatic protolith.

One analysis (No. 7 in Table 1) was not taken into consideration for the weighted mean calculation because of its very high U content (4053 ppm), although it overlaps within analytical uncertainty (2σ) with analyses 1–5. SHRIMP analyses with U contents higher than ca. 2000 ppm may yield erroneously high "ages" (see e.g., Mc Laren et al., 1994 or Williams et al., 1996). Two further analyses (Nos. 8 and 9) plot to the right of the mixing line. They yield younger – but also Permo-Triassic – "ages", probably due to partial Pb-loss attributed to post-crystallization events. Such events could be either younger Permo-Triassic thermal pulses, a feature identified commonly in the poly-episodic rifting of the Permo-Triassic period in the area of the Alps (see summary by

Table 1 U, Th, Pb SHRIMP data for magmatic and metamorphic zircon domains from the sapphirine-granulite of the Gruf Complex.

	Sample	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$f^{206}Pb$ (%)	$^{207}Pb/^{206}Pb$ (uncorrected)	± error (1σ)	$^{238}U/^{206}Pb$ (uncorrected)	± error (1σ)	$^{206}Pb/^{238}U$ (radiogenic)	± error (1σ)	AGE (Ma) $^{206}Pb/^{238}U$	± error (1σ)
	magmatic (protolith) zircon domains													
1.	COD1a-17.2	1409	706	0.517	64	0.09	0.0526	0.0009	23.11	0.45	0.0432	0.0008	272.8	5.2
2.	COD1a-10.1	406	164	0.418	18	0.02	0.0523	0.0004	23.03	0.38	0.0434	0.0007	273.8	4.4
3.	COD1a-4.1	132	55	0.428	6	0.24	0.0521	0.0012	23.19	0.87	0.0430	0.0016	272.0	10.0
4.	COD1b-14.1	463	69	0.154	19	0.04	0.0520	0.0005	23.32	0.33	0.0429	0.0006	270.6	3.8
5.	COD1b-4.2	1379	299	0.224	57	0.04	0.0513	0.0004	23.25	0.35	0.0430	0.0006	271.6	4.0
6.	*COD1a-4.2*	*577*	*87*	*0.156*	*25*	*0.14*	*0.0527*	0.0005	*21.45*	*0.37*	0.0466	0.0008	*293.6*	*4.9*
7.	*COD1b-7.2*	*4053*	*785*	*0.200*	*176*	*0.05*	*0.0524*	0.0001	*22.08*	*0.30*	0.0453	0.0006	*285.4*	*3.8*
8.	*COD1a-6.1*	*517*	*129*	*0.258*	*21*	*0.02*	*0.0521*	0.0006	*24.47*	*0.37*	0.0409	0.0006	*258.0*	*3.8*
9.	*COD1a-12.2*	*606*	*120*	*0.205*	*22*	*0.12*	*0.0582*	0.0025	*26.98*	*1.22*	0.0370	0.0017	*232.5*	*10.4*
													WM: 272.0± 4.1	
	metamorphic rims													
10.	COD1a-19.1	631	127	0.209	3	0.07	0.0583	0.0015	197.5	3.5	0.0051	0.0001	32.1	0.6
11.	COD1a-6.2	1448	12	0.009	7	0.43	0.0509	0.0006	194.9	2.9	0.0051	0.0001	32.8	0.5
12.	COD1a-17.1	318	161	0.523	2	1.29	0.0519	0.0020	195.0	3.8	0.0051	0.0001	32.8	0.6
13.	COD1a-12.1	730	94	0.133	3	2.62	0.0612	0.0009	186.9	3.3	0.0052	0.0001	33.8	0.6
14.	COD1b-7.1	909	142	0.161	4	0.72	0.0502	0.0014	197.1	3.2	0.0050	0.0001	32.5	0.5
15.	COD1b-4.1	583	119	0.210	3	1.11	0.0530	0.0018	196.1	5.3	0.0050	0.0001	32.5	0.9
													WM: 32.7 ± 0.5	

Notes:
1. Pb* is radiogenic Pb; **2.** $f^{206}Pb$ denotes the percentage of ^{206}Pb that is common Pb; **3.** WM: weighted mean (error is at the 95% c.l.). **4.** Analyses in *italics* were not considered for the calculation of the weighted mean (see text). **5.** zircons with the signature COD1a are from the restite, those with the signature COD1b from the leucosome.

Schaltegger and Gebauer, 1999) or metamorphic overprint in the Tertiary (see below). Finally, one analysis (No. 6) plots to the left of the mixing line of Fig. 4A and yields a higher "age" at 293.6 ± 4.9 Ma, slightly deviating, within analytical uncertainty (2σ), from the weighted mean age. Post-SHRIMP CL studies reveal that this spot lies entirely on a crack. Thus, some accumulation of radiogenic Pb in the crack may have been the cause for this slightly higher age.

Six spots were analyzed from the metamorphic rims of six different zircon grains from both the leucosomes (analyses named COD1b in Table 1), as well as the restitic part (analyses named COD1a in Table 1). On a TW diagram, they all plot on a mixing line with common Pb and calibrated total $^{238}U/^{206}Pb$ as end members (Fig. 4B). The weighted mean age of these analyses is 32.7 ± 0.5 Ma, which is interpreted as the age of the granulite-facies metamorphism. Late action of fluids and resetting to younger ages due to Pb-loss is very unlikely for these "dry" sapphirine granulites (see section 3).

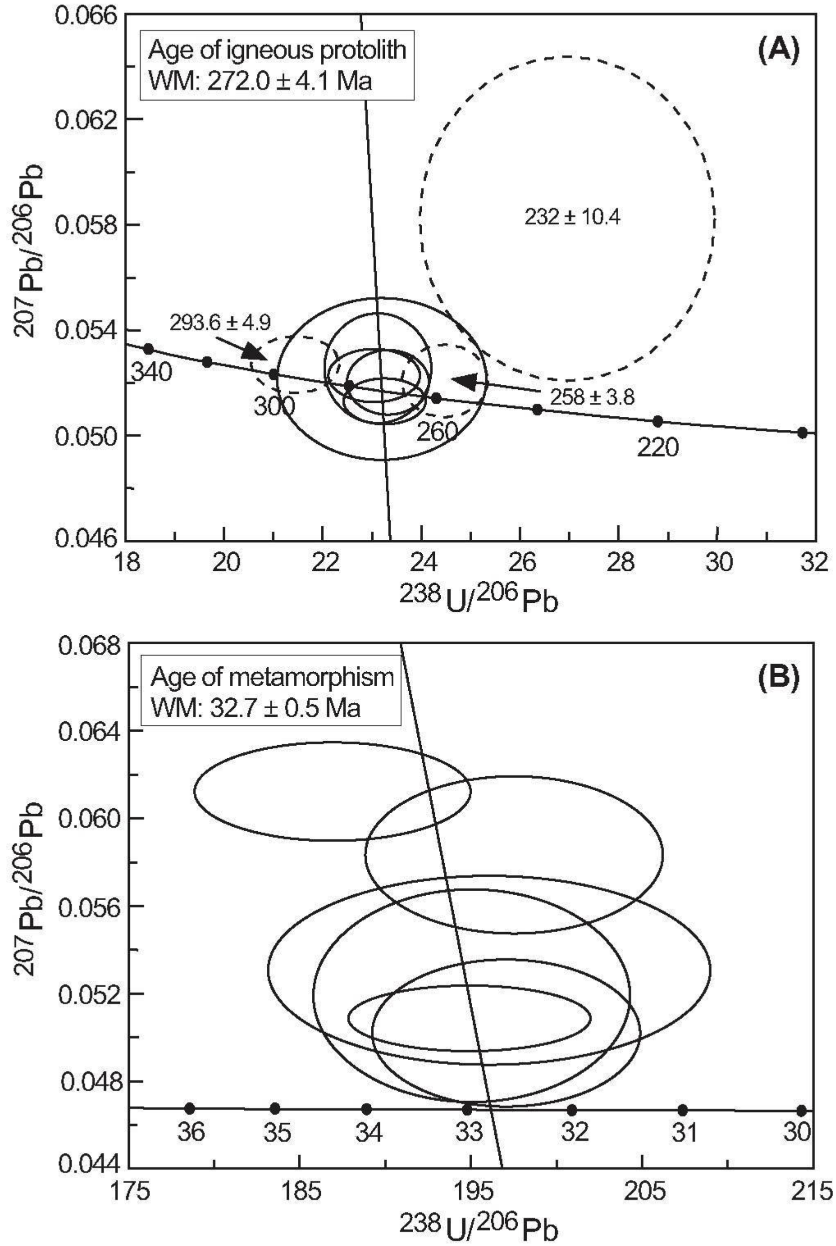

Fig. 4 Tera-Wasserburg (TW) diagrams with data of zircons from: (A) the co-magmatic, oscillatory zoned zircon domains. Five analyses plot along a mixing line with common Pb (composition according to the model of Cumming and Richards, 1975) and radiogenic $^{238}U/^{206}Pb$ as end members intersecting the concordia at 272.0 ± 4.1 Ma, interpreted as the age of crystallization of the magmatic protolith. Two outliers (dashed ellipses) on the right side of the mixing line are due to post-crystallization Pb-loss. One outlier on the left side of the mixing line is probably due to excess radiogenic Pb in a crack inside the analyzed spot (see text). (B) the metamorphic rims. All analyzed points plot along a mixing line intersecting the concordia at 32.7 ± 0.5 Ma, interpreted as the age of the granulite-facies metamorphism (see text). WM: weighted mean; error at the 95% confidence level; error axes of the ellipses are 2σ errors.

Assignment of the 32.7 ± 0.5 Ma age of the metamorphic zircon rims to the granulite-facies metamorphism is based on a series of arguments and observations acquired over the past decade of CL-controlled ion microprobe dating of complex (poly-) metamorphic zircons from orogens worldwide. It has been empirically ascertained that zircon starts responding to recrystallization only at or above T conditions corresponding to the upper amphibolite/granulite or high-temperature eclogite-facies (e.g., Vavra et al., 1999 and references therein). Thus, the age obtained from the metamorphic rims of the zircon probably reflects the time when the rocks were at or close to the thermal peak. This either coincides with or post-dates the pressure peak. Since the rates of exhumation are usually fast (>1–2 cm/y), as identified in several (high-pressure [HP])-HT rocks from orogens worldwide (e.g., Gebauer et al., 1997; Amato et al., 1999; Liati and Gebauer, 1999; Cartwright and Barnicoat, 2002 and references therein), the time difference between temperature and pressure peak is relatively short. Therefore, the age data obtained from metamorphic zircon crystals (or crystal domains) of (HP)-HT rocks, is likely to correspond to the peak of metamorphism, with respect to both P and T.

Above arguments are supported by the results of SHRIMP-dating of zircons drilled out directly from thin sections, therefore offering a good petrological control, in e.g., HP granulites of South Bohemia (Kröner et al., 2000). The zircons dated “in situ” on the thin section were associated with mineral assemblages characteristic of different temperature stages along the P–T path. They all showed identical metamorphic ages within analytical uncertainties. These ages were also identical to the ones obtained from the zircon separates of the whole rock. The above authors explain these results on the basis of fast decompression processes that usually characterize such terranes and imply that formation of metamorphic zircon domains during HP and decompressional stages along the P–T path occurred within the time scale of the SHRIMP-age errors. Another plausible interpretation may be that the metamorphic domains of the zircons from these Bohemian granulites formed originally at (or close to) the thermal peak and remained since then unaffected during later stages of metamorphism, probably due to the absence or very limited amounts of fluids.

7. Discussion

The SHRIMP data obtained within the framework of the present study, in combination with the zircon morphology and CL-patterns, as well as with SHRIMP data from other metamorphic rocks of the Central and Western Alps, led us to the following implications.

7.1. Protolith of the sapphirine-granulites

The sapphirine-bearing granulites from the GC are very likely not of pelitic origin, as presumed so far, but rather Hercynian magmatic rocks. This conclusion is based on: (a) the observation that zircons in pelites are either absent or extremely small (<30 μm), in contrast to the much larger (400–170 μm long) and abundant zircons that occur in the studied granulites; (b) the uniform sub- to euhedral morphology and oscillatory zoned (magmatic) CL-patterns of the relatively large domains of the numerous zircon crystals recovered from this rock. Such uniformity is expected in zircons that crystallize in a magmatic rock. If the parent rock is of sedimentary origin, one typically finds detrital cores of round shape and various CL-patterns; (c) absence of inherited cores; and (d) the consistent Permian magmatic ages. In a rock of sedimentary origin, one expects to find detrital zircon cores with a variety of ages. Magmatic rocks of Permian age are common in the Alps and are related to the large-scale rifting event that took place during the Permo-Triassic in the Alps and the European Variscides, usually in multiple phases (e.g., summary by Schaltegger and Gebauer, 1999).

7.2. Time of metamorphism in the GC and correlation with age data in the Adula nappe

The 32.7 ± 0.5 Ma age of the zircon rims, interpreted here as the time of granulite-facies metamorphism, strongly supports the view that these rocks formed during Alpine partial melting (compare Droop and Bucher-Nurminen, 1984). The identical morphology, CL-characteristics and age of the metamorphic zircon rim domains from both the restitic parts and the leucosomes of the granulites suggest that high-grade metamorphism at 32.7 ± 0.5 Ma caused partial recrystallization of the originally magmatic zircon crystals. Recrystallization at the zircon rims and along cracks led to formation of metamorphic domains around the magmatic cores and on either side of the cracks (Fig. 3B, C). New growth of zircon did not take place in the restites nor in the leucosomes probably because recrystallization of zircon took place before melt extraction. In that case, one part of the zircon crystals remained inside the restitic fraction and one part was mechanically transported by the extracted melt. This interpretation is in

Table 2 Upper Eocene-Lower Oligocene SHRIMP ages in the Adula-Cima Lunga-Gruf unit (Central Alps) attributed to HP and granulite-facies metamorphism and comparison with (ultra) HP ages of the Dora Maira and Monte Rosa units (Western Alps).

Area	Rock-type	HP age (Ma)	granulite age (Ma)	Source
Alpe Arami	garnet-peridotite	35.4±0.7	33.4±0.5	(1)
	pyroxenite	35.1±1.5	32.5±0.8	(1)
	eclogite	35.4±0.8		(1)
	orthogneiss leucosome	35.8±2.8	32.4±1.1	(1)
Cima di Gagnone	peridotite	36.0±0.8		(2)
	gneiss	36.0±0.5	33.0±0.6	(2)
Dora Maira	gneiss	35.4±1.0		(3)
	gneiss	35.1±0.9		(4)
Monte Rosa	gneiss	34.9±1.4		(5)
Gruf Complex	sapphirine-granulite		**32.7±0.5**	**this study**

(1) Gebauer (1996); (2) Gebauer (1994); (3) Gebauer et al. (1997); (4) Rubatto and Hermann (2001); (5) Rubatto and Gebauer (1999).

line with the idea that melt extraction and migration are significantly faster when compared to the rates of zircon dissolution in granitic melts (e.g., Watt and Harley, 1993), which prevents new growth of zircon.

Resetting of ages, subsequent to the granulite-facies metamorphism at 32.7 ± 0.5 Ma, is not observed in the metamorphic zircon domains, probably because of limited fluid circulation in these rocks after the T peak. This observation is compatible with the absence of retrograde features in the granulites.

The 32.7 ± 0.5 Ma metamorphic age attributed to the granulites of the GC is identical to previously reported granulite-facies zircon ages of crustal and mantle rocks in various parts of the Adula-Cima Lunga nappe system (e.g., Alpe Arami or Cima di Gagnone; Gebauer, 1994, 1996). Table 2 and Fig. 5 summarize SHRIMP data of the granulite-facies overprint and the HP metamorphism at Alpe Arami, Cima di Gagnone, and the GC. As shown in Fig. 5, the SHRIMP age obtained here for the Gruf granulites matches perfectly with the ages reported for granulite-facies metamorphism in peridotites, pyroxenites and gneisses from Alpe Arami and Cima di Gagnone. This overlap of ages supports previous structural/geological arguments that the GC is part of the Adula-Cima Lunga nappe (e.g., Schmid et al., 1996a).

The heat input responsible for the inferred ca. 33 Ma old granulite-facies metamorphism in parts of the Adula-Cima Lunga nappe (Alpe Arami, Cima di Gagnone) was interpreted to be mainly associated with subduction processes (Gebauer 1994, 1996). A connection between granulite-facies metamorphism in the GC and the heat related to the magmatic activity of the Bergell pluton cannot be excluded a priori, because of their immediate vicinity. Thus, Berger et al. (1996), for instance, suggest that emplacement of the western margin of the Bergell pluton occurred at temperatures leading to partial melting in the Adula-Gruf unit. However, the Bergell tonalite age (31.88 ± 0.09 Ma) is statistically clearly younger than the 32.7 ± 0.5 Ma age attributed to the granulite-facies metamorphism in the GC, even if one considers the unlikely case of extreme analytical uncertainty for the two ages. Intrusion of the syntectonic Bergell tonalite under amphibolite-facies conditions may have caused some local migmatization at the immediate contact with the previously migmatized lower crustal Gruf orthogneisses but this is of no regional significance. The heat source for the granulite-facies metamorphism in the GC, therefore, does not seem to be directly connected to the Bergell pluton, but is rather associated with regional metamorphic processes. Such is the case for Alpe Arami and Cima di Gagnone, ca. 10 km and 20 km, respectively, away from the contact to the Bergell tonalite, where the age of the high-temperature metamorphism (32.4 ± 1.1 and 33.0 ± 0.6 Ma, respectively), inferred to be of granulite-facies conditions (Gebauer, 1994, 1996), is identical to the granulite-facies age in the GC (32.7 ± 0.5 Ma, this work).

Taking into consideration that the Bergell tonalite was emplaced at 31.88 ± 0.09 Ma at 6 kbar (ca. 18 km; Davidson et al., 1996) and intruded the metamorphic rocks of the GC, and assuming that the granulites reached their maximum T and P (830 °C, 10 kbar, equivalent to ca. 30 km) at 32.7 ± 0.5 Ma, we can calculate an average exhumation rate of 12 km/0.92 Ma or 1.3 cm/y for the GC. This exhumation rate is in agreement with the exhumation rates of ca. 2cm/y (at lower crustal levels) suggested for Alpe Arami (Gebauer, 1996) or Cima di Gagnone (Gebauer, 1994). It is also in agreement with the >1–2 cm/y exhumation rates identified for metamorphic rocks in many orogens worldwide (see section 6 for references).

Exhumation rates immediately following the emplacement of the tonalites are difficult or impossible to assess due to the very restricted data set for this time period (see e.g. summary of Hansmann, 1996). The only data that may be used for the time interval 30–28 Ma are apatite fission track and K/Ar amphibole data from a tonalite boulder in the Gonfolite Lombarda (Giger and Hurford, 1989). The calculated average exhumation rate is 0.5–0.6 cm/y and applies for a depth range of 15–2 km. This exhumation rate applies for shallow levels and is, as expected, lower than the one inferred for the GC (1.3 cm/y), which refers to greater depths at earlier time (3–4 Ma). Other exhumation rates reported for the Bergell area were calculated for shallower crustal levels and for time intervals at younger ages, and therefore give lower values (see summary of Hansmann, 1996).

7.3. Geodynamic implications

The Gruf Complex probably belongs to the Adula-Cima Lunga nappe and is therefore part of the European margin. The paleogeographic and geotectonic position of this part of the Central Alps compares well with the Dora Maira Massif and the Monte Rosa nappe in the Western Alps. This correlation is in line with the view that the Dora Maira Massif and the Monte Rosa nappe belong to the European margin, tectonically emplaced from below and to the south of the Briançonnais nappes (Gebauer et al., 1997; Rubatto and Gebauer, 1999; Froitzheim, 2001). This view is at variance with the interpretation of the Dora Maira and Monte Rosa units as being of Briançonnais origin (e.g., Keller and Schmid 2002; see also review by Froitzheim, 2001 and references). The correlation between Adula-Cima Lunga-Gruf unit (Central Alps) and Dora Maira and Monte Rosa nappes (Western Alps) applies also for the metamorphic ages: Dora Maira and Monte Rosa record metamorphic ages of 35.4 ± 1.0 Ma and 34.9 ± 1.4 Ma, respectively, for HP metamorphism (Gebauer et al., 1997; Rubatto and Gebauer, 1999), identical to the HP metamorphic age at Alpe Arami and Cima di Gagnone (Table 2, Fig. 5). In the Dora Maira Massif, post-HP decompression ages at 32.9 ± 0.9 Ma (U–Th–Pb SHRIMP on titanite) are identified (Rubatto and Hermann, 2001). These ages are identical to those reported for the HT stage in the Adula-Cima Lunga nappe and determined here for the GC. Based on these data, and considering previous structural/geological arguments (e. g. Schmid et al., 1996a), we conclude that the GC together with Alpe Arami and Cima di Gagnone represent part of an Upper Eocene–Lower Oligocene Alpine metamorphic belt extending to the WSW to the Dora Maira

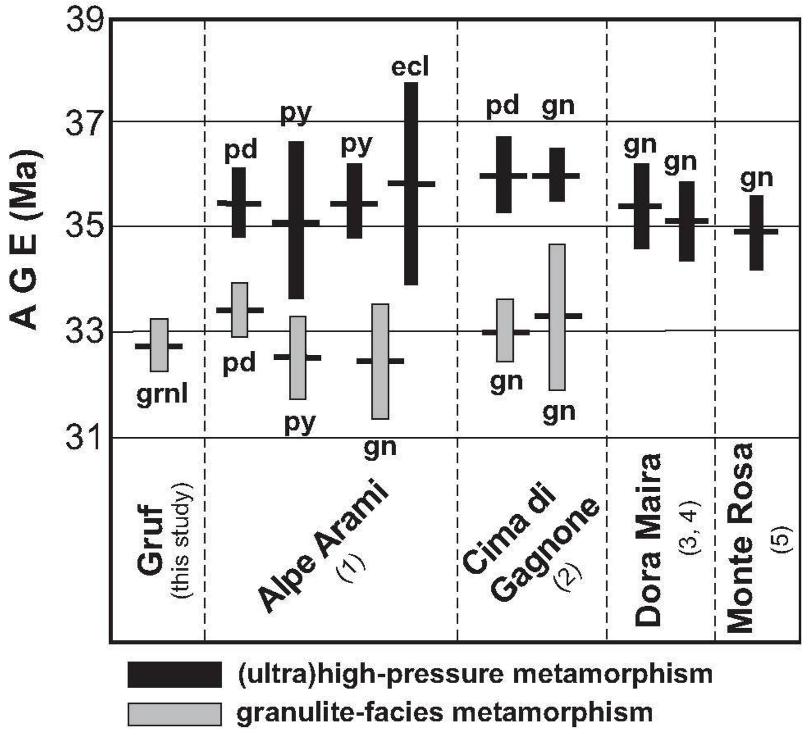

Fig. 5 Distribution of SHRIMP- (ultra) high-pressure and granulite-facies metamorphic ages in the Adula-Cima Lunga-Gruf unit of the Central Alps, and comparison with (ultra) high-pressure metamorphic ages in Dora Maira and Monte Rosa units (Western Alps). grnl—granulites; pd—peridotites; py—pyroxenites; ecl—eclogites; gn—gneisses. Age data are as follows: (1) Gebauer (1996); (2) Gebauer (1994); (3) Gebauer et al., (1997); (4) Rubatto and Hermann (2001); (5) Rubatto and Gebauer (1999).

Massif. The generalized map of the Western and Central Alps shown in Fig. 1 summarises the age distribution of HP and granulite-facies (overprinting) metamorphism of the units, assigned paleogeographically to the European margin. These units record the youngest regional metamorphic event in the Alps, which reached its peak between the uppermost Eocene (for the (U)HP metamorphism) and lowermost Oligocene (for the granulite-facies metamorphism).

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