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Pre-Variscan deformation, metamorphism and magmatism in the Strona-Ceneri Zone (southern Alps of northern Italy and southern Switzerland)

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Abstract

The Strona-Ceneri Zone comprises the deformed and metamorphosed remains of a Late Proterozoic or Paleozoic subduction-accretion complex that was intruded by Ordovician granitoids. Xenoliths within these granitoids include relics of MOR-basalt that experienced eclogite facies metamorphism (690–750 °C, > 15–16 kbar) and paragneiss that underwent D1 deformation at amphibolite facies conditions. Pre-intrusive exhumation of the eclogites involved nearly isothermal decompression to 9–11 kbar sometime before their equilibration under prograde D2 amphibolite facies conditions (570–630 °C, 7–9 kbar). D2 mylonitization at these conditions coincided with and outlasted the intrusion of large volumes of Ordovician granitoids. Most of these granitoids stem from mixing of mantle-derived melts with aluminous metasediments in the Strona-Ceneri Zone. Pure S-type granitoids, the so-called Ceneri granitoids, derived from partial, biotite-dehydration melting of such metasediments under fluid-depleted, granulite facies conditions. These granitoid magmas then intruded and were emplaced syntectonically at the current erosional surface in the Strona-Ceneri Zone.

The metasediments of the Strona-Ceneri Zone were probably deposited during Cadomian orogenesis and/or Cambro-Ordovician rifting. Eclogitization of oceanic crust and subsequent uplift is either Cadomian or Ordovician age. Ordovician magmatism and related D2 deformation and metamorphism are ostensibly related to subduction and/or collision along the northern margin of Gondwanaland.

Keywords: Strona-Ceneri Zone, southern Alps, Paleozoic tectonics, Ordovician granitoids, anatexis, subduction, exhumation.

1. Introduction

The Strona-Ceneri Zone[1] is a SW–NE striking, pre-Alpine basement unit in the western part of the southern Alps (Fig. 1) that comprises mainly fine to medium grained gneisses and schists of sedimentary origin (metagreywacke and metapelite), banded amphibolites and metagranitoids (descriptions in BÄCHLIN, 1937; SPICHER, 1940; GRAETER, 1951 and BORIANI et al., 1977). The position of the Strona-Ceneri Zone between pre-Alpine, amphibolite to granulite facies rocks of the Ivrea Zone to the NW and unmetamorphosed, Permo-Mesozoic sediments to the S and SE has led some workers to interpret it as the originally intermediate part of a tilted crustal section in which progressively deeper crustal levels are exposed from SE to NW (e.g., FOUNTAIN, 1976; FOUNTAIN and SALISBURY, 1981). However, other workers, most notably BORIANI et al. (1990), have long maintained that the Strona-Ceneri Zone was never tilted and that the subvertical dip of its compositional banding and schistosity is a Variscan feature.

Despite an abundance of geochronological

[1] *The name "Strona-Ceneri Zone" was first coined by SCHMID (1968) and is synonymous with the term "Serie dei Laghi" of BORIANI et al. (1977). The reader is referred to table 1 of ZINGG (1983) for a listing of different names for this unit in the local literature.*

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and geochemical studies in the Strona-Ceneri Zone, little is known about its pre-Variscan history, primarily due to a dearth of combined structural and petrological studies. The metagranitoids yield Ordovician intrusive ages (U–Pb zircon, PIDGEON et al., 1970; KÖPPEL and GRÜNENFELDER, 1971; RAGETTLI, 1993; and this work; Rb–Sr whole rock, BORIANI et al., 1982/83), but the age of their deformation is poorly constrained. Either Ordovician or Early Carboniferous deformational ages are possible given the observation that the penetrative schistosity in the Strona-Ceneri Zone overprints the Ordovician metagranitoids but is itself folded by Middle Carboniferous, km-scale folds, so-called "Schlingen" in the local literature (Fig. 1; ZURBRIGGEN et al., 1998). Moreover, the metagranitoids contain inclusions of deformed schist and gneiss that betray a complex, pre-intrusive history (BORGHI, 1989; BORIANI et al., 1990). The age and tectonic setting of these structures remains enigmatic. Some local authors favour a convergent margin setting for the Strona-Ceneri Zone in Cambro-Ordovician time (SCHMID, 1993 and references therein), but this interpretation contrasts with large scale tectonic reconstructions calling for rifting (ZIEGLER, 1990) or transcurrent tectonics (VAI et al., 1984) in southern and eastern Alpine units during this time.

The genesis of Ordovician magmatism in the Strona-Ceneri Zone lies at the heart of the early Paleozoic enigma outlined above. The Ordovician metagranitoids make up more than 20% of the Strona-Ceneri Zone (Fig. 1) and comprise both mantle- and crustal-derived intrusive types (BO-

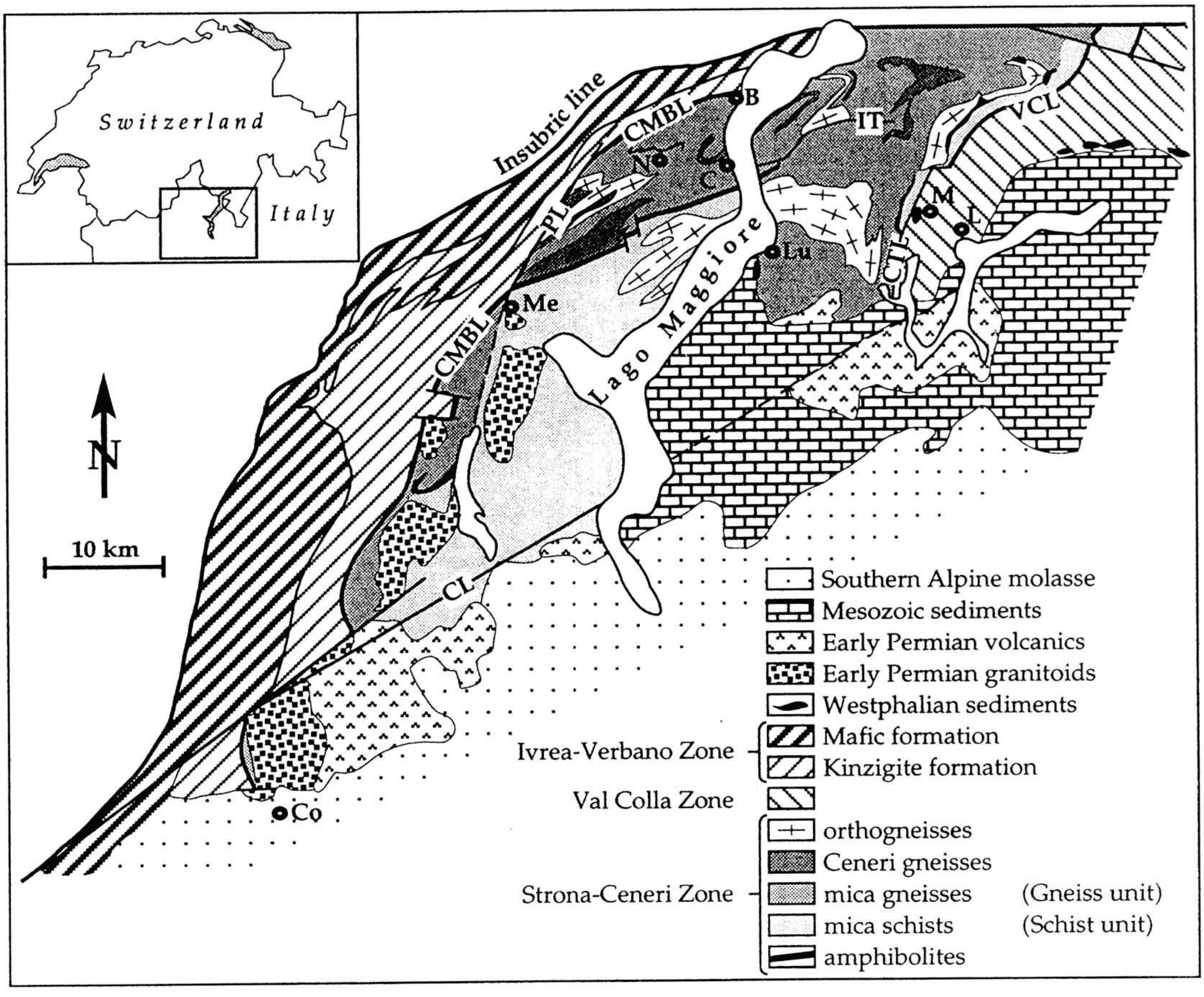

Fig. 1 Geological map of the western part of the southern Alps (modified from BORIANI et al., 1977; ZINGG et al., 1990). Tectonic lineaments: CMBL: Cossato-Mergozzo-Brissago Line, PL: Pogallo Line, VCL: Val Colla Line, CTL: Caslano-Taverne Line, CL: Cremosina Line, IT: Indemini thrust. Towns or villages mentioned in text: Me: Mergozzo, Ni: Nivetta, Ca: Cannobio, Co: Cossato, Br: Brissago, Li: Luino, Ma: Manno, Lu: Lugano, MC: Monte Ceneri. Note locality of figure 6.

RIANI et al., 1982/83). The origin of one of these types, the Ceneri granitoids (known as "Ceneri gneiss" in the local literature; BÄCHLIN, 1937), remains controversial. This is partly because the Ceneri granitoids have both sedimentary and magmatic features, leading people to regard it variously as a metasediment (REINHARD, 1953), a metasomatically altered metasediment (BÄCHLIN, 1937; GRAETER, 1951; BORIANI et al., 1982/83; GIOBBI ORIGONI et al., 1982, BORIANI et al., 1990) or an intrusive rock derived from the melting of sediments (PREISWERK and REINHARD, 1934; BORIANI, 1968; GIOBBI MANCINI and POTENZA BIANCHI, 1972; HANDY, 1986; ZURBRIGGEN, 1996). Clearly, characterizing this magmatism and associated metamorphism is a key to reconstructing the early Paleozoic tectonic setting of the Strona-Ceneri Zone.

This paper presents evidence that the Strona-Ceneri Zone contains relics of Proterozoic and/or

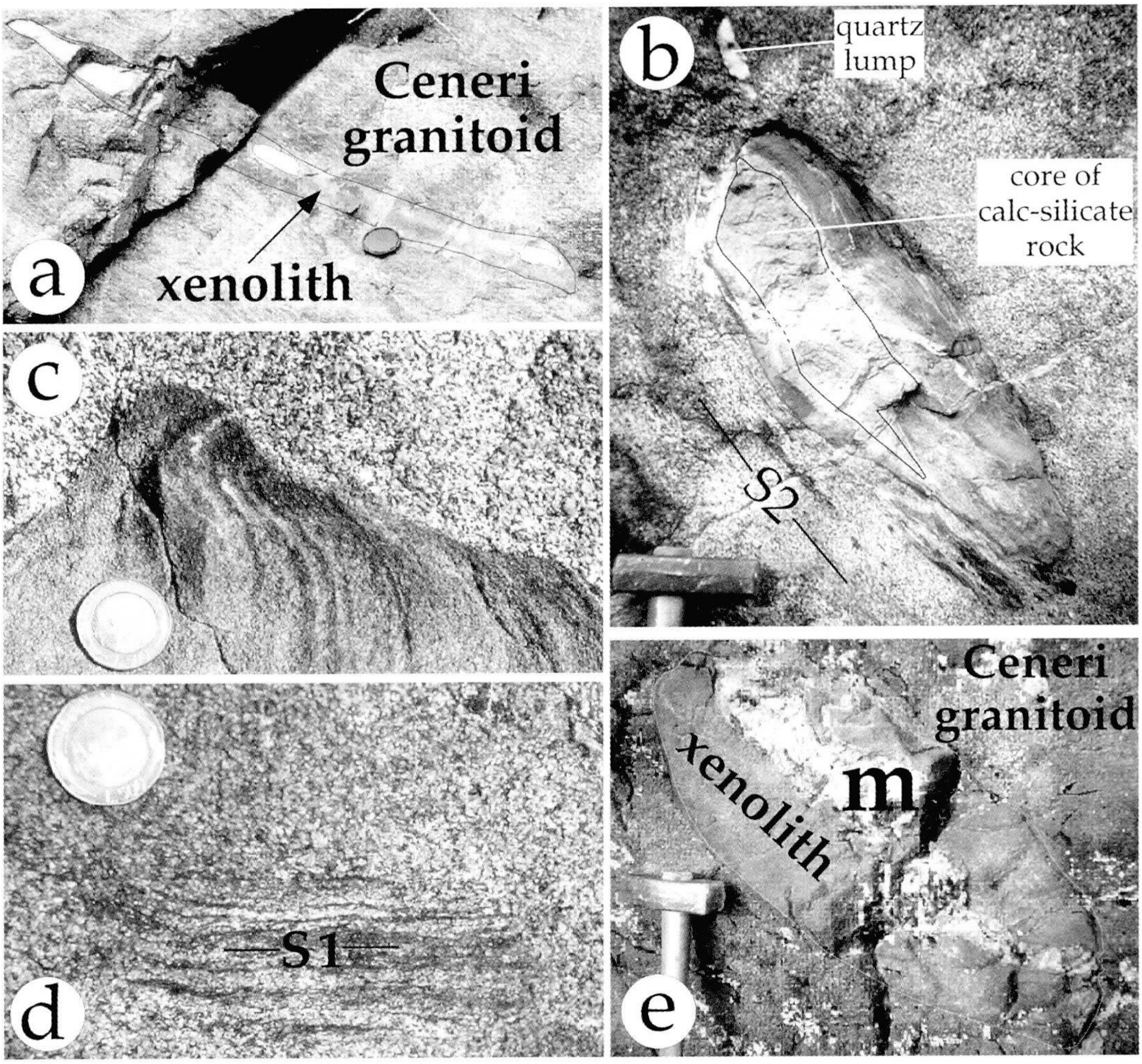

Fig. 2 Xenoliths in the Ceneri granitoid: (a) D1 boudin preserved within xenolith of fine grained biotite gneiss; xenolith is oriented parallel to S2 schistosity of Ceneri granitoid (Swiss coordinates: 697 180 / 103 430); (b) calc-silicate nodule as a xenolith within weakly foliated (S2) Ceneri granitoid; note that the S2 schistosity of the Ceneri granitoid truncates the mineral zonation in the nodule (Swiss coordinates: 696 970/103 235); (c) massive, magmatic fabric of Ceneri granitoid cuts the S1 schistosity in xenolith of fine grained Bt-gneiss (Swiss coordinates: 696 880 / 103 120); (d) partially melted xenolith within a massive, magmatic fabric of undeformed Ceneri gneiss; restitic biotite-rich bands and schlieren are the only remnants of the former biotite gneiss (Swiss coordinates: 696 880/103 120), (e) xenolith of fine grained biotite gneiss containing a pegmatitic dike (marked with "m") that is truncated by Ceneri granitoid (Swiss coordinates: 696 910/103 100).

early Paleozoic continental margins. In the first section, we outline structural arguments for the relative age of tectonometamorphic events in the Strona-Ceneri Zone. This forms the basis for the interpretation of thermo-barometric estimates in terms of pressure-temperature-time paths in part two. Geochemical and geochronological data advanced in the third part of the paper yield insight into the origin and emplacement mechanisms of the Ordovician granitoids. Considered together, this information allows us to shed new light on the pre-Variscan evolution of the Strona-Ceneri Zone.

2. Structural relationships

The Ordovician granitoid gneisses in the Strona-Ceneri Zone are excellent structural markers because they allow one to classify structures as pre-, syn-, or post-magmatic.

2.1. PRE-MAGMATIC STRUCTURES

Pre-intrusive D1 structures are unequivocally identifiable in fine grained xenoliths within the granitoids (Fig. 2). These xenoliths have an S1 schistosity that forms the axial plane of F1 folds and locally envelops boudinaged layers (Fig. 2a). Such structures are best observed in the central parts of Ceneri granitoid bodies where the S2 schistosity (described below) is absent or only weakly developed and massive magmatic fabrics are preserved (Figs 2 c and d). The schistosity in metapelitic xenoliths comprises Bt-Ms ± Grt ± Ky ± St (ZURBRIGGEN, 1996, p. 81), a mineral assemblage that is diagnostic of pressure-dominated amphibolite facies conditions. Sometimes S1 can also be discerned in the hinge zones of tight, F2 folds within fine to medium grained, biotite-rich paragneiss. This early, pre-granitoid foliation has been termed S_{1a} by HANDY (1986) and S_x by BORIANI et al. (1990).

An early phase of high pressure deformation and metamorphism is recorded by foliated eclogitic pods within amphibolite lenses. These lenses are intercalated with micaceous schists and gneisses in the Monte Gambarogno area (located north of the Tertiary Indemini thrust, see Fig. 1; BULETTI, 1983; BORGHI, 1988). The fact that the relictic eclogites are boudinaged within the S2 schistosity and show a retrograde amphibolite facies overprint toward their contacts with the surrounding micaceous gneisses indicates that high pressure metamorphism pre-dated regional D2 deformation. This inference is confirmed by the occurrence of xenoliths of eclogitic garnet-amphibolite within Ordovician granitoids (BORGHI, 1989). Unfortunately, the relative age of this eclogite facies deformation and amphibolite facies D1 deformation cannot be established due to the lack of cross-cutting relationships.

2.2. SYN- TO LATE-MAGMATIC STRUCTURES

In granitoid augen gneisses, euhedral feldspar phenocrysts display a strong shape preferred orientation that defines a magmatic to submagmatic foliation. In most places, however, this foliation is obliterated by the S2 schistosity.

The S2 schistosity forms the axial plane of tight to isoclinal F2 folds and is the predominant foliation of the Ordovician granitoids, and indeed of the entire Strona-Ceneri Zone. This schistosity is generally best developed at and within the rims of the granitoid bodies and within the adjacent country rock. It is associated with dynamically recrystallized quartz and feldspar aggregates and therefore formed during high-temperature, solid-state flow. These dynamic, solid-state microstructures are partly annealed (Fig. 3b; Plates 4.10 a–c in HANDY, 1986; Fig. 5c in HANDY and ZINGG, 1991; Fig. 16.3 in SCHMID and HANDY, 1990) and the S2 schistosity contains stable, amphibolite facies mineral parageneses, indicating that amphibolite facies conditions outlasted mylonitic flow.

Several lines of evidence suggest that the emplacement of the Ordovician granitoids was coeval with D2 deformation in the entire Strona-Ceneri Zone: (1) Some granitoids locally preserve a magmatic foliation defined by the shape preferred orientation of feldspar phenocrysts. This relictic foliation is concordant with the S2 schistosity in the adjacent country rocks as well as with the contact between granitoids and country rocks. Such mutual concordance is generally considered to be a reliable geometric criterion for syn-tectonic intrusion (e.g., PATERSON et al., 1989). (2) The high aspect ratio of the Ordovician granitoids and their orientation parallel to the strike of the S2 foliation are characteristic features of plutons whose emplacement was tectonically and structurally controlled (HOLLISTER and CRAWFORD, 1986; PATERSON et al., 1989; HUTTON and EBBERT, 1995). (3) Metasedimentary country rocks of the granitoids yield lower intercept U–Pb detrital zircon ages (510 ± 50 Ma, PIDGEON et al., 1970; 430–560 Ma, KÖPPEL and GRÜNENFELDER, 1971) that fall within the range of U–Pb zircon and monazite ages (426–515 Ma, RAGETTLI, 1993) and Rb–Sr whole rock ages (466 ± 5 Ma, BORIANI et al., 1982/83) for the granitoid gneisses (see section

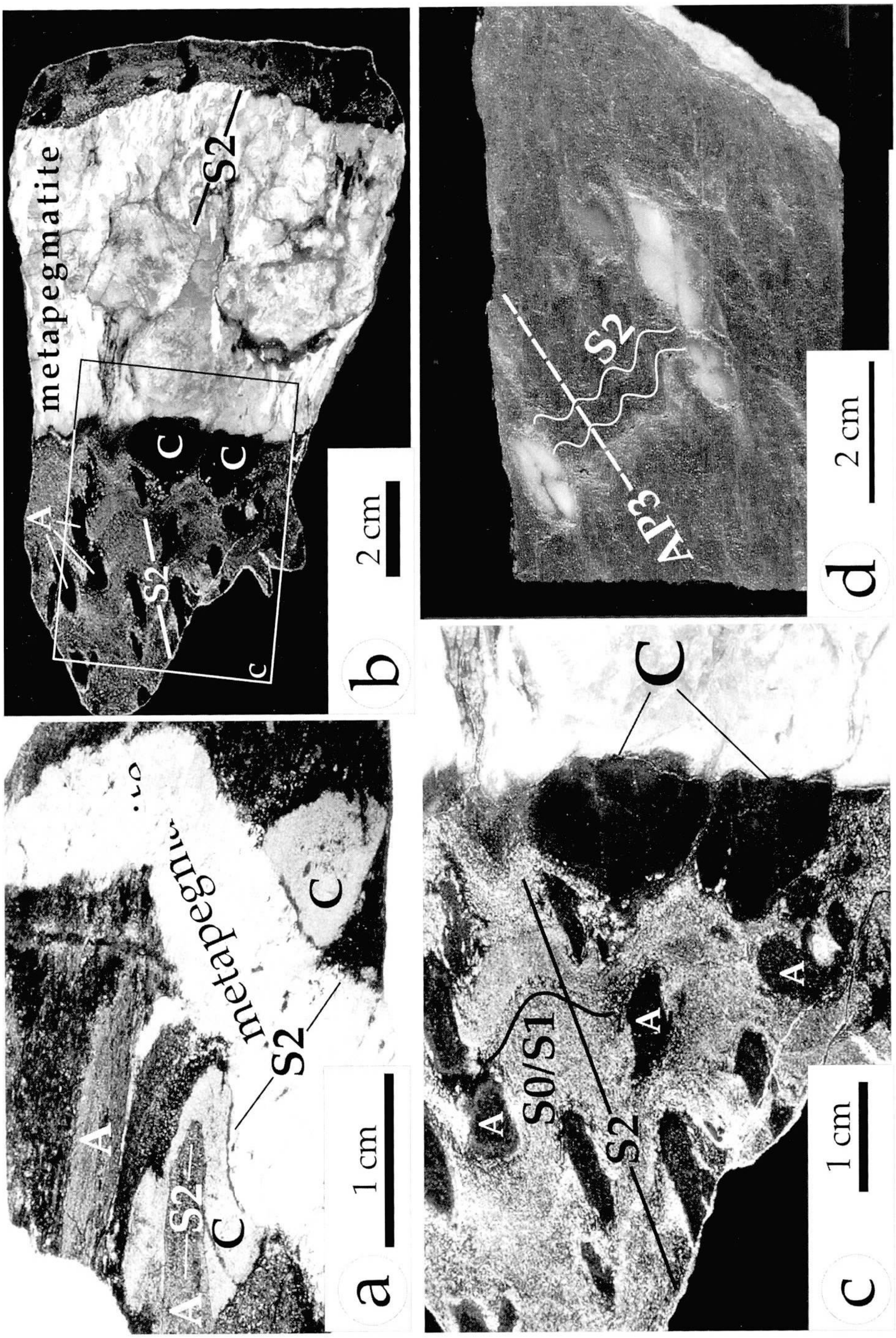

Fig. 3 Relationship between deformation and metamorphic minerals. (a) Al-silicate nodules (A) oriented parallel to S2; S2 deforms both country rock and pegmatite dike; chiastolite (C) overgrows Al-silicate nodules; (b) deformed pegmatite dike showing flaser microstructure associated with D2 deformation; (c) detail of (b) showing deformed S0 and/or S1 at high angles to S2. Figures 3 a–c originate from the same hand-specimen, Swiss coordinates: 697 000/103 935). (d) Fine grained mica schist containing chiastolite pseudomorphs. Note how the chiastolite pseudomorphs discordantly overgrew the older S2 schistosity and therefore lack any penetrating D2 deformation. The chiastolite pseudomorphs are elongate parallel to the axial planes of F3 Schlingen folds (Swiss coordinates: 705 090/108 430).

on Ordovician magmatism below). The concordant granitoid ages are interpreted as intrusive ages, whereas the ages from the metasediments date a severe isotopic disturbance that is reasonably inferred to be the age of the syn- to post-D2, amphibolite facies metamorphism in the country rock (KÖPPEL, 1974; HUNZIKER and ZINGG, 1980). Therefore, emplacement of the Ordovician granitoids was syn-tectonic with respect to the amphibolite facies, S2 schistosity. We interpret other mineral isotopic systems that yield Carboniferous ages (K–Ar and Ar–Ar hornblende, Rb–Sr white mica, BORIANI et al., 1982/83; BORIANI and VILLA, 1996; K–Ar biotite, MCDOWELL, 1970) to record amphibolite facies metamorphism associated with Variscan D3 deformation. This deformation occurred at similar temperatures, but under less hydrous conditions and lower pressures than during D2 (see discussion below).

3.2. POST-MAGMATIC STRUCTURES

The Ordovician granitoids are folded by, and therefore clearly pre-date, Mid-Carboniferous F3 folds in the Strona-Ceneri Zone (Fig. 1). The relationship between late- and post-magmatic deformation and metamorphism is particularly well exposed near Cannobio on the Lago Maggiore, at the contact between the Ordovician Ceneri granitoid and its micaceous country rocks (location in Fig. 1). There, a pegmatite dike cuts and therefore post-dates an earlier foliation (So and/or S1) and schistosity in the fine grained gneisses making up the country rock (Figs 3 a–c). However, pegmatitic dikes are truncated by and therefore predate Ordovician Ceneri granitoids (Fig. 2e). We infer that the schistosity in both the dike and the adjacent biotite gneiss is the regionally developed, S2 schistosity because this schistosity also affects the adjacent Ceneri granitoid and must therefore be late- to post-magmatic in age. Large (1–2 cm long) Al-silicate nodules comprising very fine grained (0.05–0.15 mm) aggregates of Qtz-Plg-Bt-Ms-Grt-Sil-Ky lie within the S2 schistosity (Fig. 3) and are therefore interpreted to have grown prior to or during D2 deformation. Finger-sized pseudomorphs of kyanite after chiastolite (marked C in Fig. 3) overgrow both the Al-silicate nodules and the S2 schistosity. The chiastolites are clearly deformed by F3 folds (Fig. 3d), indicating that andalusite grew sometime between D2 shearing and F3 folding. The chiastolites terminate at the dike margins (Figs 3 a–c), presumably because the dike was not sufficiently aluminous for their growth. The partial replacement of the chiastolite, first by kyanite and then by white mica (Fig. 4a), is interpreted to reflect renewed burial during D3 followed by syn- to post-D3 exhumation under hydrous, greenschist facies conditions.

As an aside, we note that ZURBRIGGEN (1996) interpreted the discordant dike in figure 3 to postdate the growth of chiastolite. However, this interpretation is not consistent with the cross-cutting relationships shown in figure 3d.

BIGIOGGERO and BORIANI (1975) and BORIANI et al. (1982/83) interpreted the Al-silicate nodules from this same outcrop as evidence for contact metamorphism adjacent to the pegmatite dikes in relatively shallow crustal levels during Ordovician time. We are inclined to doubt this interpretation, because the Al-silicate nodules are not restricted to contact aureoles of D2-deformed pegmatitic dikes (also, see discussion in ZINGG, 1983, p. 381).

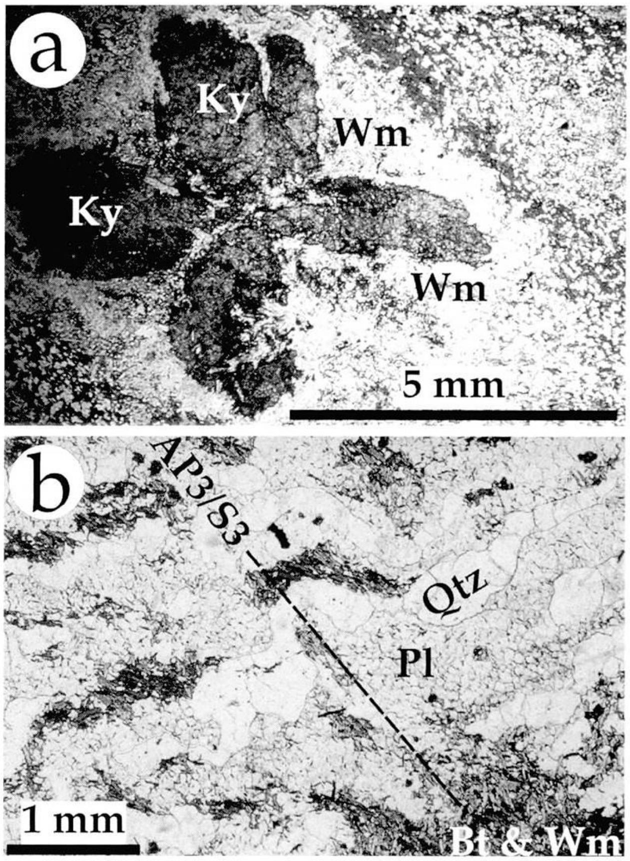

Fig. 4 Microstructures: (a) Chiastolite (viewed perpendicular to its long axis) replaced by kyanite (Ky) that is rimmed by white mica (Wm); sample found as boulder on old forest path near Scaiano (Swiss coordinates: 702 470/106 400); (b) Ceneri granitoid from hinge zone of the Monte Giove D3 fold (Swiss coordinates: 694 725/102 450); quartz ribbons (Qtz) and aggregates of polygonized plagioclase (Pl) define S2, which is deformed by open F3 folds; small grains of biotite (Bt) and white mica (Wm) oriented perpendicular to S2 define the S3 schistosity parallel to the axial planes of F3 folds (Swiss coordinates: 694 725/102 450).

Figure 4b shows the typical microstructure of a gneissic Ceneri granitoid (BORIANI, 1970). Clusters of mica, elongate aggregates of tiny, polygonal plagioclase grains, and quartz ribbons define the S2 schistosity. Note, however, that the mica grains define a new, S3 schistosity oblique to S2, and parallel to the axial plane of an F3 fold. In the less competent mica schists making up the country rock, S3 completely obliterates S2, particularly in the axial planes of F3 folds (ZURBRIGGEN, 1996).

3. Pressure-temperature conditions and evolution

3.1. D1 METAMORPHISM

The P-T conditions of amphibolite facies D1 deformation in the fine to medium grained inclusions within the Ceneri gneiss are unconstrained due to the lack of appropriate mineral assemblages for geothermometry or -barometry. However, the eclogitic garnet-amphibolites in the Gamborogno area reveal different stages of pre-D2 metamorphism.

An initial high-pressure event is preserved in the central parts of the garnet-amphibolite lenses, where garnets contain numerous inclusions of omphacitic pyroxene, isofacial hornblende, and rutile. Microprobe profiles through garnet cores show flat zonation patterns with $Alm_{52-53}Prp_{17}$ $Grs_{27-29}Sps_{1-2}$, whereas the garnet rims often show a distinct increase in X_{Mg} (see Tab. 1). Grt-Cpx thermometry (ELLIS and GREEN, 1979) and Grt-Hbl thermometry (GRAHAM and POWELL, 1986) on garnet inclusions yield temperatures of 690–750 °C. Pressure estimates vary from a minimum of 15–16 kbar, deduced from the reaction Ab <==> Qtz + Cpx (Jd_{40-43}; cp. HOLLAND, 1980 and Tab. 1), and a maximum of 22–24 kbar constrained by the amphibole-out isograd of CARSWELL (1990).

Subsequent decompression and uplift of the eclogites is documented by symplectitic intergrowths of pyroxene and plagioclase around garnet via the reaction Grt + Omph <==> Cpx (Jd_{8-13}) + Plg(An_{5-15}) + Qtz. The jadeite content of the pyroxenes indicates pressures of 9–11 kbar for the formation of these symplectites, and the width of the pyroxene and plagioclase lamellae (3–5 mm) indicates temperatures of 690–750 °C if the thermometric method of JOANNY et al. (1991) is used. At such high temperatures, we would expect an Fe–Mg exchange equilibrium between the newly formed pyroxenes and the garnet rims, as well as the net transfer reaction: Grt + Qtz <==> Cpx + Plg. This is only locally observed, however, and many of the symplectites appear to be in disequilibrium with the garnet. Only locally do some rim sections of the garnets appear to have re-equilibrated with the adjacent clinopyroxene-plagioclase symplectites (CPS in Tab. 1). For these assemblages, pressures of 11–13 kbar at 700–760 °C are estimated using the Grt-Cpx-Plg-Qtz geobarometer of NEWTON and PERKINS (1986) and the Grt-Cpx geothermometer of ELLIS and GREEN (1979). Similar P-T conditions are obtained with the Grt-Hbl geothermometer of GRAHAM and POWELL (1986) and the Grt-Hbl-Plg-Qtz geobarometer of KOHN and SPEAR (1990) applied to newly grown green-brown hornblende and plagioclase at the garnet rims (see HP in Tab. 1). These results indicate that the decompression of the eclogites was nearly isothermal.

The margins of the garnet-amphibolite lenses do not record the high pressure event or even the subsequent isothermal decompression. Garnets from these amphibolites are compositionally zoned, with bell-shaped Mn and Ca profiles indicating a complete mineral chemical remobilization of the eclogitic protolith. This suggests that the margins of the boudinaged eclogitic layers underwent total re-equilibration during the prograde part of subsequent D2, amphibolite facies metamorphism.

3.2. D2 METAMORPHISM

In order to obtain reliable estimates of D2 metamorphic conditions, we sampled parts of the Strona-Ceneri Zone between the Lago Maggiore and Lago d'Orta (Fig. 1) where the amphibolite to greenschist facies, D3 structural and metamorphic overprint is minor or nonexistent. There, the alignment of the critical mineral assemblages Bt-Ms-Chl-Grt-Plg(An_{20}) ± Rt ± Ilm and Bt-Ms-Grt-St-Plg(An_{16-20}) ± Rt ± Ilm within the S2 schistosity, as well as the occurrence of kyanite in quartzites are diagnostic of D2 amphibolite facies conditions at elevated pressures. In Grt-St mica-schist, large garnets (> 1 cm) have bell-shaped Mn- and Ca-zonation patterns, and are progressively Fe- and Mg-richer from their cores to their rims. Where the highest X_{Mg}-values (0.85–0.86) are recorded at the rims, the garnets have a composition of $Alm_{76}Prp_{13}Grs_8Sps_3$. At the outermost parts of some rims, a decrease of the pyrope content at the expense of almandine and spessartine components points to retrograde cooling. Small garnets (0.1–0.2 mm) are found as inclusions in large plagioclase grains. These small garnets yield similar X_{Mg}-values and compositions as the afore-

Tab. 1 Representative microprobe analyses of mineral assemblages related to the eclogite facies event and the subsequent uplift of a garnet-amphibolite from the Mt. Gambarogno area. Abbreviations: CPS (clinopyoxene-plagioclase symplectite), HP (hornblende-plagioclase intergrowth), Uvar (uvarovite), Ce (celsiane), Cel (celadonite), other abbreviations are from KRETZ (1983). Mineral analyses were performed with a CAMECA SX50 electron beam microprobe with four spectrometers at the GFZ-Potsdam. Major and minor elements were determined at 15 kV acceleration voltage with a beam current of 20 nA and counting times of 20–30 seconds, depending on the element. The CAMECA standard set was used as a reference and the PAP program for matrix correction.

Garnet wt%	IZ93-72 core	IZ93-72 rim near CPS	IZ93-72 rim near HP
SiO_2	38.62	38.12	38.07
TiO_2	0.07	0.43	0.01
Al_2O_3	21.78	21.39	21.50
Cr_2O_3	0.02	0.02	0.01
Fe_2O_3	0.01	0.00	0.21
MgO	4.40	4.38	4.91
CaO	10.26	10.15	9.59
MnO	0.70	0.75	0.51
FeO	24.47	23.64	24.08
Na_2O	0.04	0.07	0.01
Total:	100.38	98.93	98.89
Cations (O = 12)			
Si	3.000	2.999	2.995
Ti	0.004	0.025	0.000
Al	1.994	1.983	1.993
Cr	0.001	0.001	0.001
Fe^{3+}	0.001	0.000	0.013
Mg	0.510	0.514	0.576
Ca	0.854	0.856	0 808
Mn	0.046	0.050	0.034
Fe^{2+}	1.589	1.555	1.584
Na	0.006	0.010	0.001
Total:	8.005	7.994	8.003
Endmembers:			
Uvar	0.1	0.1	0.0
Adr:	0.2	1.3	0.4
Grs:	28.2	27.4	26.5
Alm:	53.0	52.3	52.8
Sps:	1.5	1.7	1.1
Prp:	17.0	17.3	19.2

Plagioclase wt%	IZ93-72 rim CPS	IZ93-72 rim HP
SiO_2	64.50	60.41
Al_2O_3	21.67	24.39
MgO	0.16	0.02
CaO	2.86	5.98
MnO	0.01	0.06
FeO	0.10	0.10
BaO	0.00	0.00
Na_2O	9.59	8.21
K_2O	0.05	0.05
Total:	98.95	99.22
Cations (O = 8)		
Si	2.866	2.707
Al	1.135	1.288
Mg	0.011	0.001
Ca	0.136	0.287
Mn	0.000	0.002
Fe	0.004	0.004
Ba	0.000	0.000
Na	0 826	0.713
K	0.003	0.003
Total:	4.981	5.006
Endmembers:		
An:	14.1	28.6
Ab:	85.6	71.1
Or:	0.3	0.3
Ce:	0.0	0.0

Clino-pyroxene wt%	IZ93-72 core incl. in Grt	IZ93-72 rim CPS
SiO_2	55.12	53.33
TiO_2	0.10	0.09
Al_2O_3	8.67	2.29
Cr_2O_3	0.04	0.01
Fe_2O_3	0.21	0.75
MgO	9.66	12.62
CaO	13.80	20.87
MnO	0.06	0.04
FeO	5.75	7.35
Na_2O	5.46	1.56
K_2O	0 00	0.01
Total:	98.87	98.91
Cations (O = 6)		
Si	2.000	1.993
Ti	0.003	0.003
Al	0.371	0.101
Fe^{3+}	0.006	0.021
Cr	0.001	0.000
Mg	0.523	0.703
Ca	0.537	0.836
Mn	0.002	0.001
Fe^{2+}	0.174	0.230
Na	0.384	0.113
K	0.000	0.000
Total:	4.000	4.000
Endmembers		
Jd:	41.1	9.7
Acm:	0.6	2.2
Aug:	58.3	88.1

Hornblende wt%	IZ93-72 core incl. in Grt	IZ93-72 rim HP
SiO_2	40.03	40.26
TiO_2	2.31	0.80
Al_2O_3	16.33	16.42
Fe_2O_3	3.17	2.66
Cr_2O_3	0.06	0.06
MgO	9.57	9.90
CaO	9.87	10.87
MnO	0.01	0.15
FeO	12.15	12.56
Na_2O	3.25	3.07
K_2O	0.12	0.35
Total:	96.87	97.10
Cations (O = 23)		
Si	5.967	6.010
Ti	0.259	0. 090
Al	2.869	2.889
Fe^{3+}	0.355	0.299
Cr	0.007	0.007
Mg	2.127	2.203
Ca	1.576	1.739
Mn	0.002	0.019
Fe^{2+}	1.515	1.568
Na	0.939	0.890
K	0.023	0.066
Total:	15.639	15.780

mentioned rims of the large garnet grains and are believed to have grown syntectonically at or near the peak of D2 metamorphism. Peak metamorphic temperatures for the D2 event estimated with the Grt-Bt-geothermometer of WILLIAMS and GRAMBLING (1990), the Grt-St-geothermometer of PERCHUK (1969) and the Grt-Chl-geothermometer of GHENT et al. (1987) were in the range of 570–610 °C. Metamorphic pressures were at 7.7–8.6 kbar according to the Grt-Ms-Bt-Plg-geobarometer of POWELL and HOLLAND (1988) or the GRIPS geobarometer of BOHLEN and LIOTTA (1986). A prograde, clockwise P-T path during D2 can be inferred from inclusions of plagioclase (An_{20-22}) in the grossular-rich cores of the big garnets (cp. Tab. 2). Pressures of about 8 kbar at temperatures of 550–600 °C are estimated from feldspar thermometry and the phengite geobarometer of MASSONNE (1991) on core sections of phengites aligned within the S2 schistosity of gneissic Ceneri granitoids. These phengites have an Si-content of about 3.3 in their cores and of 3.05–3.1 at their rims, values that are consistent with those which BORIANI et al. (1990) obtained from phengites in metasediments from the Valle Cannobina. Peak metamorphic P-T conditions of 7–8 kbar at 590–630 °C were obtained with the Grt-Hbl geothermometer of GRAHAM and POWELL (1986) and the Grt-Hbl-Plg-Qtz geobarometer of KOHN and SPEAR (1990) applied to D2-assemblages at the margins of the eclogitic garnet-amphibolite lenses of the Monte Gam-

Tab. 2 Representative microprobe analyses of D2-mineral assemblages from a garnet-staurolite micaschist (Grt-Ms), a garnet amphibolite (Grt-Am), and a Ceneri gneiss (CG); see table 1 for mineral abbreviations and analytical details.

Garnet wt%	IZ93-50 Grt-Ms core	IZ93-50 Grt-Ms near rim	IZ93-50 Grt-Ms rim	IZ94-72 Grt-Am rim
SiO_2	36.84	37.34	37.22	37.75
TiO_2	0.10	0.00	0.01	0.11
Al_2O_3	21.43	21.50	21.19	21.39
Cr_2O_3	0.03	0.03	0.02	0.03
Fe_2O_3	0.42	0.17	0.00	0.34
MgO	2.26	3.33	3.09	3.83
CaO	5.12	2.82	0.92	8.02
MnO	2.71	1.11	1.67	0.50
FeO	31.61	34.46	36.11	27.85
Na_2O	0.01	0.00	0.02	0.02
Total:	100.52	100.77	100.23	99.83
Cations (O = 12)				
Si	2.951	2.974	2.995	2.983
Ti	0.006	0.000	0.001	0.006
Al	2.023	2.018	2.009	1.991
Cr	0.002	0.002	0.001	0.002
Fe^{3+}	0.025	0.010	0.000	0.020
Mg	0.270	0.395	0.370	0.451
Ca	0.439	0.241	0.079	0.679
Mn	0.184	0.075	0.114	0.034
Fe^{2+}	2.117	2.295	2.430	1.840
Na	0.002	0.000	0.004	0.004
Total:	8.019	8.011	8.003	8.008
Endmembers:				
Uvar:	0.09	0.10	0.05	0.08
Adr:	0.29	0.11	0.04	0.46
Grs:	14.22	7.80	2.55	22.07
Alm:	70.33	76.34	81.19	61.27
Sps:	6.11	2.50	3.80	1.12
Prp:	8.98	13.15	12.37	15.01

Plagio-clase wt%	IZ93-50 Grt-Ms incl. in Grt	IZ93-50 Grt-Ms rim	IZ94-72 Grt-Am rim
SiO_2	63.57	64.99	58.72
Al_2O_3	22.41	22.20	25.07
MgO	0.00	0.00	0.00
CaO	4.39	4.13	7.30
MnO	0.03	0.03	0.01
FeO	0.52	0.06	0.14
BaO	0.00	0.05	0.05
Na_2O	9.09	9.57	7.17
K_2O	0.38	0.10	0.07
Total:	100.37	101.17	98.50
Cations (O = 8)			
Si	2.810	2.840	2.658
Al	1.168	1.143	1.338
Mg	0.000	0.000	0.000
Ca	0.208	0.193	0.354
Mn	0.001	0.001	0.000
Fe	0.019	0.002	0.005
Ba	0.000	0.001	0.001
Na	0.779	0.811	0.629
K	0.021	0.006	0.004
Total:	5.006	4.997	4.989
Endmembers:			
An:	20.61	19.11	35.85
Ab:	77.28	80.19	63.69
Or	2.10	0.56	0.38
Ce:	0.00	0.14	0.08

Biotite wt%	IZ93-50 Grt-Ms core
SiO_2	36.36
TiO_2	1.45
Al_2O_3	18.10
Cr_2O_3	0.00
MgO	11.43
CaO	0.00
MnO	0.15
FeO	18.45
BaO	0.05
Na_2O	0.13
K_2O	9.22
H_2O	3.72
F	0.48
Cl	0.06
Total:	99.60
Total cor:	99.38
Cations (O = 12)	
Si	2.749
Ti	0.082
Al	1.613
Cr	0.000
Mg	1.288
Ca	0.000
Mn	0.010
Fe	1.167
Ba	0.002
Na	0.019
K	0.889
Total:	7.817
OH	1.877
F	0.115
Cl	0.008

Musco-vite wt%	IZ93-50 Grt-Ms core	IZ93-162 CG core	IZ93-162 CG rim
SiO_2	47 17	50.36	46.95
TiO_2	0.51	0.59	0.84
Al_2O_3	36.36	29.97	34.74
Cr_2O_3	0.01	0.00	0.01
MgO	0.60	2.45	1.10
CaO	0.01	0.00	0.01
MnO	0.01	0.03	0.04
FeO	0.87	1.76	1.47
BaO	0.00	0.16	0.25
Na_2O	2.78	0.23	0.31
K_2O	7.42	10.36	9.58
Total:	95.74	95.90	95.31
Cations (O = 11)			
Si	3.080	3.324	3.111
Ti	0.025	0.029	0.042
Al	2.798	2.332	2.713
Cr	0.000	0.000	0.001
Mg	0.059	0.241	0.108
Ca	0.001	0.000	0.001
Mn	0.001	0.002	0.002
Fe	0.048	0.097	0.082
Ba	0.000	0.004	0.007
Na	0.352	0.029	0.040
K	0.618	0.872	0.810
Total:	6.980	6.931	6.915
Endmembers:			
Mrg:	0.1	0.0	0.1
Pg:	36.2	3.2	4.7
Cel:	8.0	32.4	11.11
Ms:	55.7	64.3	84.2

Stauro-lite wt%	IZ93-50 Grt-Ms rim
SiO_2	28.67
TiO_2	0.75
Al_2O_3	53.67
Cr_2O_3	0.03
MgO	1.57
CaO	0.00
MnO	0.43
FeO	12.79
CuO	0.00
ZnO	0.50
BaO	0.00
Na_2O	0.00
K_2O	0.01
Total:	98.41
Cations (O = 23)	
Si	3.945
Ti	0.077
Al	8.703
Cr	0.003
Mg	0.321
Ca	0.000
Mn	0.050
Fe	1.472
Cu	0.000
Zn	0.051
Ba	0.000
Na	0.000
K	0.002
Total:	14.625

Horn-blende wt%	IZ94-72 Grt-Am rim
SiO_2	42.86
TiO_2	0.73
Al_2O_3	53.34
Fe_2O_3	2.78
Cr_2O_3	0.01
MgO	8.93
CaO	10.83
MnO	0.09
FeO	13.14
Na_2O	1.33
K_2O	0.37
Total:	96.41
Cations (O = 23)	
Si	6.376
Ti	0.081
Al	2.690
Fe^{3+}	0.311
Cr	0.002
Mg	1.981
Ca	1.726
Mn	0.012
Fe^{2+}	1.635
Na	0.338
K	0.070
Total:	15.267

barogno area. These P-T conditions were also determined for the metapelitic host rocks that surround the eclogitic garnet-amphibolite lenses. Representative mineral analyses for the D2 event are presented in table 2.

3.3. P-T PATHS FOR THE STRONA-CENERI ZONE

The P-T paths in figure 5 were constructed from the available information above on the relative age and physical conditions of the D1 and D2 events in the Strona-Ceneri Zone. The pre-D2, high pressure event documented by mineral assemblages in the eclogites (Fig. 5a) probably marks late Proterozoic and/or early Paleozoic subduction of oceanic crust, as the garnet-amphibolites containing the retrograded eclogites have

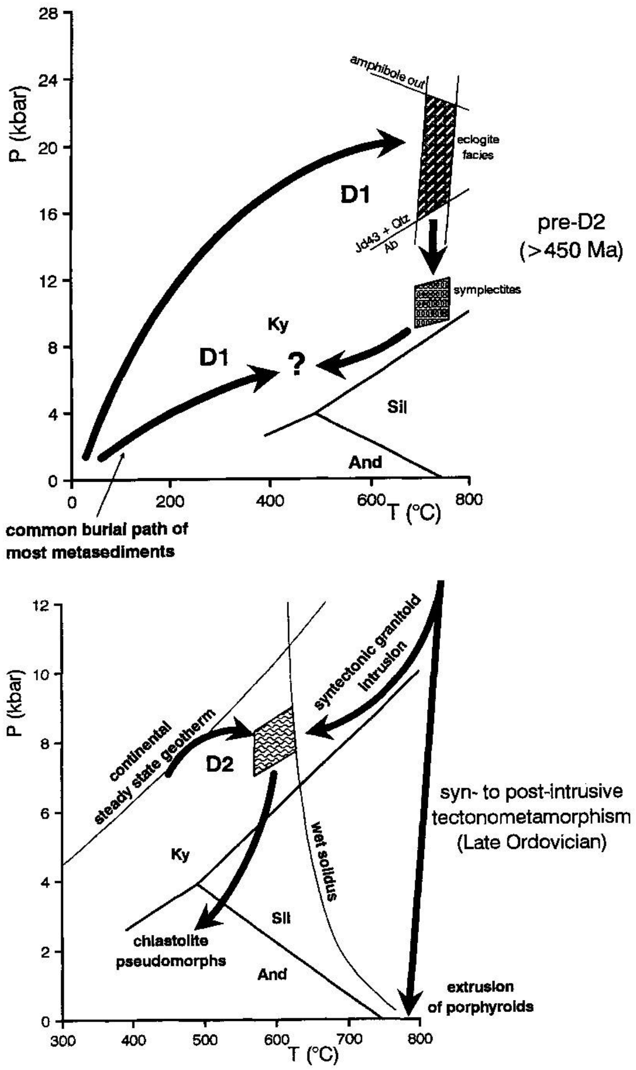

Fig. 5 P-T paths for the Strona-Ceneri Zone: (a) pre-D2 evolution from eclogitic garnet amphibolites of the Gambarogno area; (b) D2 evolution.

a distinct MORB composition (Buletti, 1983; Giobbi Origoni et al., 1997). Subsequent exhumation of this subducted oceanic crust was probably rapid given the nearly isothermal decompression following the baric peak of metamorphism. The D1 amphibolite facies deformation preserved in the fine grained gneiss inclusions may be unrelated to the eclogite facies event and it is likely that the Strona-Ceneri Zone experienced a much more complicated pre-D2 evolution than indicated by the P-T path for the eclogites in figure 5a. Indeed, a single P-T path for D1 is unreasonable given the possibility discussed in the final section that the metasedimentary and amphibolitic rocks of the Strona-Ceneri Zone were imbricated within an early Paleozoic subduction-accretion complex.

The solid-state D2 deformation in the Strona-Ceneri Zone syn-dated prograde, amphibolite facies metamorphism, but pre-dated decompression and uplift, all in Ordovician time (Fig. 5b). Unfortunately, the kinematics of D2 deformation are obscured by the strong D3 overprint (see Franz et al., 1996), so the mechanisms of D2 uplift remain speculative. The occurrence of post-D2 chiastolites, however, points to a decompression after the peak of D2 metamorphism (Fig. 5b).

Obviously, poor constraints on the timing of pre-D2, eclogite and amphibolite facies deformation(s) and of the post-D2 growth of andalusite remain the main obstacle to reconstructing the pre-Variscan evolution of the Strona-Ceneri Zone. Also, P-T paths alone are not diagnostic of the tectonic setting of a tectonometamorphic event. We therefore turn to the Ordovician granitoids for some geochemical and petrological clues regarding the conditions and tectonic setting of events up to and including D2.

4. Ordovician magmatism

We classified the Ordovician granitoids of the Strona-Ceneri Zone into three groups that differ in abundance, mineralogy and bulk composition: (1) Hornblende-bearing tonalites make up 15% of all granitoids in the Strona-Ceneri Zone and have Sr-initial ratios lower than 0.706 (Boriani et al., 1995), reflecting their mantle heritage; (2) hornblende-free granodiorites and biotitic augen gneisses make up, respectively, 45% and 10% of all granitoids and have Sr initial values greater than 0.708 (Boriani et al., 1995), betraying a clear crustal signature; (3) the third type making up the remaining 30% of granitoids comprises the so-called Ceneri gneisses (based on their type-locality at the Monte Ceneri pass, location in Fig. 1), and has peculiar structural, mineralogical and geochemical characteristics. As pointed out below, some parts of these granitoid bodies are unfoliated, so we will refer to them more generally as Ceneri granitoids instead of gneisses.

The Ceneri granitoid is a Ms-Bt-Qtz-Plg rock that sometimes bears garnet, sillimanite, kyanite and K-feldspar. It forms elongate bodies (Fig. 1) and is usually gneissic due to the D2 and D3 overprints (Fig. 2b). This structural overprint is strongest towards the margins of the bodies, where the contacts are concordant with the S2 schistosity in the gneissic country rock, but truncate discordant pegmatite dikes like that shown above in figure 3. The centers of some Ceneri granitoid bodies preserve a massive, magmatic fabric (Figs 2 c, d). The S2 schistosity forms foliation triple points at the ends of some bodies (Fig. 6), indicating that the relictic magmatic cores of these bodies were more competent than the surrounding gneisses during the latter stages of D2 deformation. Both gneissic and massive varieties of the Ceneri granitoid contain quartz lumps (Fig. 2b), xenoliths of fine grained biotite gneiss ("gneiss minuti" in the local literature), zoned

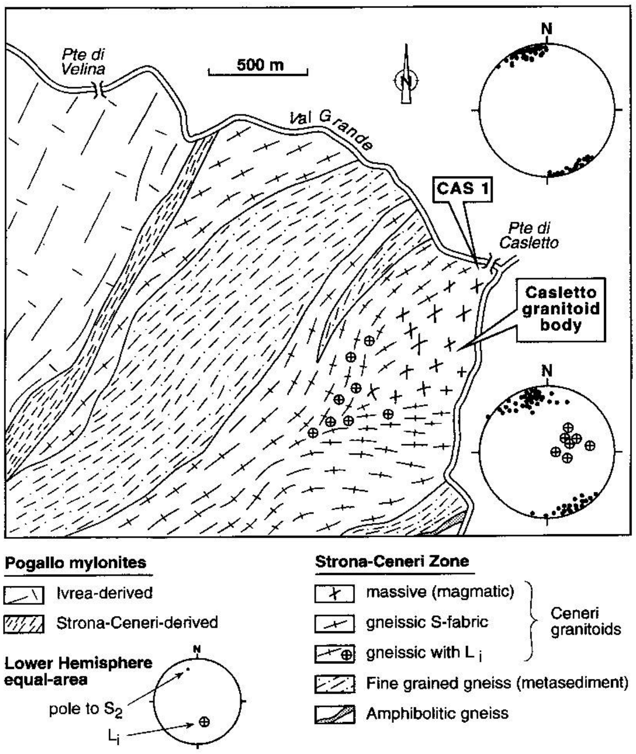

Fig. 6 Map of part of the Ceneri granitoid body at Ponte di Casletto (see Fig. 1 for location). Note trend of S2 schistosity in and around the granitoid body (see text for explanation). Intersecting S2 surfaces yield an intersection lineation, Li. CAS1 is sample of gneissic Ceneri granitoid dated by KÖPPEL and GRÜNENFELDER (1971). Map from figure 2.02 in HANDY (1986).

calc-silicate nodules (Fig. 2b), and biotitic enclaves and schlieren (see also BORIANI and CLERICI RISARI, 1970). All of these features suggest a magmatic origin for the Ceneri granitoid. Yet, the bulk composition of the Ceneri granitoid is that of a pelitic greywacke and is identical to that of the fine grained biotite gneiss in the Strona-Ceneri Zone (ZURBRIGGEN, 1996, see also Fig. 8). Thus, the Ceneri granitoid has the chemical and mineralogical characteristics of a metasediment, but the structural features of a deformed intrusive rock.

We interpret the Ceneri granitoid to stem from partial melting of a strongly peraluminous source rock under water-undersaturated, granulite facies conditions. The source rock is inferred to be the fine grained biotitic gneiss. Further to the evidence outlined above for a magmatic origin, several lines of circumstantial evidence support this interpretation: (1) The Ceneri granitoid and its xenoliths of fine grained biotitic gneiss have the same composition and both contain zoned calc-silicate nodules. (2) The discordant pegmatites found in the vicinity of the Ceneri granitoid bodies pre-date, rather than post-date the Ceneri granitoid bodies. Although the absolute age of these pegmatites is unknown, their proximity to the Ceneri granitoid bodies suggests that they are related to the latter in time and space. The early age of the pegmatites with respect to the granitoid bodies is consistent with an origin as an eutectic melt associated with incipient partial melting, rather than as a late stage product of fractional crystallization. The pegmatite dikes probably originated by partial melting in the roof of the Ceneri granitoid bodies and intruded the country rocks before themselves being cut by the rising bodies in the final stages of emplacement. (3) The Ceneri granitoids contain lumps of quartz (Fig. 2b), a typical feature in anatectically derived granitoids. CHAPPELL et al. (1987) explain such lumps as restitic quartz that originated as vein quartz in metasedimentary protoliths. (4) The P-T estimates for the D2 event and lack of evidence for extensive partial melting of the gneissic country rock during D2 indicate that the Ceneri granitoid bodies did not form by in-situ anatexis at the D2 paleodepth (26–33 km; Fig. 5b) presently exposed to erosion in the Strona-Ceneri Zone. They must therefore have originated at greater depths and correspondingly higher temperatures and pressures.

Because the biotitic gneiss protolith of the Ceneri granitoid clearly underwent amphibolite facies metamorphism during D1, we can assume that the porosity of this protolith was already low at the time of D2 anatexis and therefore that the protolith contained an insufficient amount of interstitial water to sustain partial melting. We therefore propose that anatexis of the Ceneri protolith involved dehydration melting of micas in the fine grained biotitic gneisses at water-undersaturated conditions and at yet higher temperatures and pressures (CLEMENS and VIELZEUF, 1987). According to VIELZEUF and HOLLOWAY (1988), fluid-absent, biotite dehydration melting of a pelitic greywacke rock within a narrow temperature range of 850–875 °C can yield up to 40 vol.% of melt in the form of S-type granitoids. This is more than enough melt to form interconnected weak layers in the anatectic rock (e.g. ARZI, 1978), enabling the anatectic masses to rise and intrude levels corresponding with the currently exposed surface of the Strona-Ceneri Zone. Assuming a geothermal gradient of 25 °C/km for the peak of the D2-event, one can estimate that anatexis at around 860 °C occurred at pressures of about 11 kbar, corresponding to a granitoid source depth of approximately 35 km. Granulite facies melting temperatures in the granitoid source area are sufficiently high to account for the granodioritic to tonalitic composition of most Ceneri granitoids, a range of compositions which is far from that of a minimum melt in the pelitic greywacke system (Fig. 8c). These temperature and pressure conditions are also in accord with the occurrence of kyanite and garnet in the neosomes of partially melted xenoliths within Ceneri granitoids (ZURBRIGGEN, 1996).

An anatectic origin of the Ceneri granitoids at granulite facies conditions is consistent with the partial and total resetting of U–Pb isotopic systems, respectively, in zircons and monazites from all three types of Ordovician granitoids. Four zircon fractions from a small Ceneri granitoid body in the Valle Cannobina (Swiss coordinates 694 410/102 200) yield a U–Pb discordia, with an apparent lower intercept age of 479 ± 24 Ma and an upper intercept age at 2029 ± 287 Ma (see Fig. 7 and Tab. 3). We interpret the lower intercept age as the age of crystallization just following anatexis, whereas the upper intercept age is an inherited zircon age. The data above confirm the U–Pb ages of KÖPPEL and GRÜNENFELDER (1971, 1978) and RAGETTLI (1993) from other Ceneri granitoid bodies (e.g., Ponte di Casletto, Fig. 6): U–Pb zircon ages are either discordant, with lower and upper intercept ages, respectively, at 450–515 Ma and 1.6–2.8 Ga, or concordant at 456 ± 4 and 459 ± 2 Ma. Monazites yield concordant U–Pb ages ranging from 426 ± 5 to 450 ± 10 Ma. The Ordovician anatectic age of the Ceneri granitoids falls well within the 430–500 Ma age range of the other granitoids (types 1 and 2 above) in the Strona-Ceneri Zone (PIDGEON et al., 1970; KÖPPEL and GRÜNENFELDER, 1971; BORIANI et al., 1982/83; and RAGETTLI, 1993). There is broad consensus that these ages are magmatic crystallization ages (KÖPPEL, 1974; HUNZIKER and ZINGG, 1980; BORIANI et al., 1982/83). The isotopic evidence re-

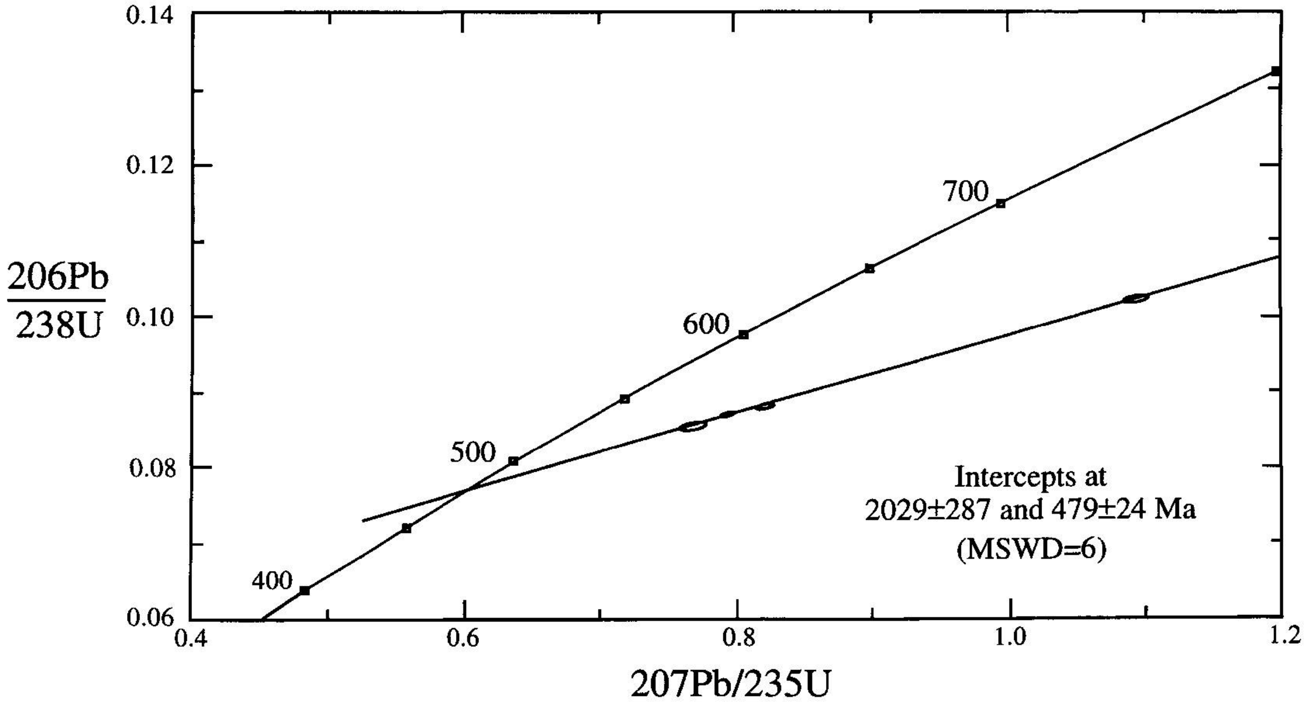

Fig. 7 U–Pb zircon age data of a Ceneri granitoid (Swiss coordinates: 694 410/102 200). See table 3 for listing of data.

Tab. 3 U–Pb analytical results for zircons of a Ceneri granitoid from Valle Cannobina (Swiss coordinates 694 410/102 200).
a) Mineral concentrates were obtained using Wilfley table, heavy liquids, and Frantz isodynamic magnetic separator. Concentrates were purified by hand under the binocular. The individual analyses were performed on euhedral, unbroken crystals. Zircon was digested with 52% HF in Parr bombs at 240 °C for 4 days. Lead and uranium were separated using the procedure of KROGH (1973) and MANHÉS et al. (1978), respectively. U and Pb were loaded with H_3PO_4 and silica gel on separate Re single-filaments and measured on a VG54 sector multi collector mass spectrometer using Faraday collectors and Daly probe. Pb^+ was analysed at 1260 °C and UO_2^+ at 1400 °C. Mass-discrimination was estimated from multiple measurements of lead standard NBS 981 and natural uranium. Mass fractionations were corrected with 0.1%/AMU for lead and uranium. Analytical precisions on the 2s level, including uncertainties for the mass fractionation, are generally better than 0.1%, except for ratios involving ^{204}Pb.
b) Ratio corrected for mass-discrimination and isotopic tracer. Not corrected for ^{206}Pb excess (SCHÄRER, 1984).
c) Corrected for mass-discrimination, isotopic tracer contribution, 30 pg of Pb blank, 1 pg of U blank and initial common Pb. Ages using the constants recommended by IUGS (STEIGER and JÄGER, 1977).

Sample (a)	Weight (mg)	Concentrations (ppm) U	Concentrations (ppm) Pb_{total}	$^{206}Pb/^{204}Pb$ measured (b)	Radiogenic Pb (at%) (c) ^{206}Pb	^{207}Pb	^{208}Pb	Atomic ratios (c) $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	Apparent ages (Ma) (c) $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
IZ 93-165													
Zircon													
oval, 60–80 μm	1	810.99	66.95	6 909.22	88.898	5.804	5.298	0.085333	0.768206	0.065292	528	579	783
short prismatic, > 63 μm	1.841	741.52	62.67	6 444.3	88.548	5.844	5.608	0.087024	0.791969	0.066003	538	592	806
prismatic, > 80 μm	2.482	705.17	60.59	15 878.1	88.138	5.983	5.878	0.08806	0.824226	0.067884	544	610	865
prismatic, rounded cores, < 160 μm	0.803	695.7	70.41	4 143.51	86.749	6.739	6.511	0.102085	1.093459	0.077685	627	750	1138

Tab. 4 Geochemical data of various rock samples from the Strona-Ceneri Zone. Z190R (restitic biotite-rich enclave in Ceneri granitoid, marked "M" in Fig. 8c), Z37 (mica schist), Z6 and Z36 (fine grained micaceous gneisses) Z48 and Z82 (micaceous gneiss), Z194H (fine grained micaceous gneiss with Al-silicate nodules). Data from ZURBRIGGEN (1996).

	Ceneri granitoids													
sample	M35	Z1	Z7	Z8	Z112	Z190C	Z192C	Z97	Z100	Z101	Z103	Z198	Z201	Z213
SiO2	68.17	67.40	66.89	68.01	68.06	67.65	68.03	66.49	65.41	65.69	65.60	74.35	65.47	65.42
TiO2	0.46	0.61	0.75	0.68	0.60	0.63	0.67	0.68	0.64	0.70	0.68	0.29	0.74	0.71
Al2O3	15.48	15.79	15.41	15.34	16.15	15.52	15.36	15.62	17.38	16.46	17.13	13.07	15.84	16.86
Fe2O3 total	4.27	4.78	5.42	5.02	4.56	5.03	4.63	5.40	5.51	5.71	5.09	1.90	5.23	5.20
MnO	0.07	0.07	0.08	0.07	0.08	0.07	0.07	0.05	0.08	0.06	0.08	0.02	0.07	0.07
MgO	1.97	2.02	2.16	1.96	1.40	2.11	2.04	2.08	2.07	1.76	2.03	0.77	2.11	2.20
CaO	2.00	1.71	1.27	1.05	1.03	1.38	1.29	1.32	0.72	1.09	0.99	1.44	2.33	1.20
Na2O	3.41	2.98	3.03	2.68	2.64	2.88	3.01	2.42	2.46	2.25	2.44	2.67	4.26	3.09
K2O	2.65	2.57	3.41	3.74	3.81	3.50	3.41	3.24	4.01	3.52	4.26	5.00	2.67	3.63
P2O5	0.25	0.27	0.18	0.16	0.09	0.07	0.15	0.26	0.12	0.35	0.13	0.37	0.38	0.12
l. o. i.	1.23	2.17	1.17	1.22	1.69	0.97	1.23	2.60	1.97	2.41	1.78	0.58	0.87	1.46
total (wt.%)	99.94	100.38	99.77	99.91	100.09	99.81	99.89	100.16	100.36	100.00	100.21	100.46	99.96	99.95
Ba	485	630	590	746	672	911	643	746	907	867	998	1245	189	647
Cr	49	61	70	63	55	65	64	77	73	66	64	22	60	65
Cu	19	18	13	23	17	14	17	16	29	21	27	14	23	19
Ga	16	18	18	17	18	16	17	17	20	18	17	10	18	21
Nb	9	13	14	12	15	11	11	14	12	15	7	6	17	14
Ni	25	27	30	35	21	22	18	35	19	34	35	19	22	24
Pb	11	17	15	20	17	9	15	7	17	12	13	33	15	24
Rb	96	85	158	143	193	143	137	93	161	172	127	112	140	150
Sr	217	193	165	170	158	167	187	160	211	197	185	236	213	206
Th	2	11	12	15	20	8	11	14	14	8	9	8	9	10
V	84	101	99	93	83	101	128	118	111	102	109	32	104	129
Y	19	27	20	30	31	16	28	23	28	30	23	25	50	25
Zn	64	68	83	67	55	55	60	56	76	102	73	33	65	65
Zr	72	145	185	193	170	171	166	189	138	146	126	133	159	154

	M	metapegmatites							some representative metasediments					
sample	Z190R	Z10	Z64	Z85	Z98	Z102	Z189	Z194M	Z37	Z6	Z36	Z48	Z82	Z194H
SiO2	41.25	71.99	76.07	79.33	77.28	76.96	76.05	75.28	64.87	64.46	69.62	65.88	66.16	51.51
TiO2	2.38	0.08	0.03	0.05	0.03	0.10	0.04	0.04	0.81	0.70	0.90	0.72	0.67	0.92
Al2O3	21.27	16.67	14.12	12.59	13.98	14.05	14.70	15.74	15.62	16.16	13.65	16.77	15.73	23.06
Fe2O3 total	16.34	0.52	0.28	0.31	0.82	0.68	0.49	0.68	6.78	5.80	5.14	5.79	5.51	8.28
MnO	0.19	0.01	0.01	0.01	0.01	0.03	0.01	0.01	0.11	0.08	0.07	0.08	0.07	0.15
MgO	6.14	0.20	0.19	0.21	0.09	0.21	0.30	0.27	2.70	2.48	2.18	2.49	2.35	3.76
CaO	1.04	2.31	0.47	0.68	0.94	0.63	0.72	0.83	1.55	1.56	1.32	1.51	2.34	1.46
Na2O	1.96	6.83	1.53	4.25	4.26	1.87	3.20	1.78	2.59	2.92	3.07	2.29	4.00	2.79
K2O	7.00	0.68	5.75	1.84	1.60	4.03	2.98	3.62	2.80	3.36	2.35	2.62	2.50	4.84
P2O5	0.04	0.16	0.16	0.13	0.02	0.09	0.21	0.27	0.18	0.21	0.23	0.15	0.15	0.20
l. o. i.	2.48	0.42	1.20	0.82	0.96	1.35	1.33	1.78	1.89	2.01	1.43	1.58	0.99	3.02
total (wt.%)	100.07	99.86	99.82	100.22	99.99	99.98	100.03	100.30	99.90	99.75	99.95	99.87	100.48	99.98
Ba	688	96	1079	225	283	1435	515	427	911	620	517	699	454	794
Cr	289	1	7	4	10	10	9	14	78	67	75	75	51	108
Cu	59	18	14	17	23	16	15	20	14	34	15	22	15	14
Ga	54	13	6	9	13	10	11	13	21	18	14	19	16	31
Nb	32	4	1	1	3	3	2	2	12	13	16	13	11	13
Ni	45	14	12	13	14	12	13	11	43	39	25	18	30	51
Pb	1	30	6	14	35	21	14	9	6	9	12	10	13	9
Rb	551	17	160	43	34	75	78	59	115	139	75	106	105	168
Sr	90	243	116	127	128	154	124	85	181	214	209	232	281	179
Th	4	6	5	5	22	13	8	4	10	9	16	8	11	10
V	433	5	5	7	5	7	8	6	133	119	116	125	124	180
Y	7	9	1	4	75	14	1	4	32	29	34	27	27	32
Zn	192	13	4	4	6	6	6	6	93	87	60	103	76	115
Zr	310	33	4	14	64	99	6	14	199	164	303	152	161	159

viewed above for coeval magmatism and anatexis of metasediments corroborates field evidence of magma-mingling and -mixing between Ceneri granitoids and type 1 and 2 Ordovician granitoids in the Strona-Ceneri Zone, as observed in the form of hybridization zones tens to hundreds of meters wide in the summit region of M. Zottone in the Malcantone (WENGER, 1983).

Our interpretation of the Ceneri granitoid as a product of partial dehydration melting of a metasedimentary gneiss differs significantly from that of BORIANI et al. (1990, p. 108) who claim that the protolith of the Ceneri granitoid was a sandstone with conglomeratic layers containing interstitial water and that this porous sediment was granitized by the "magmatic residue" of the Ordovician granitoids. Their hypothesis is based mainly on a simple tectonic model of BÄCHLIN (1937) in which the sedimentary protolith of the Ceneri granitoid supposedly formed part of a stratigraphic sequence that was folded into a large syncline in the northern part of the Strona-Ceneri Zone. As evidence for a strictly sedimentary origin of the Ceneri granitoids, BORIANI et al. (1990) cite the sedimentary composition of the Ceneri granitoid, the occurrence therein of quartz- and

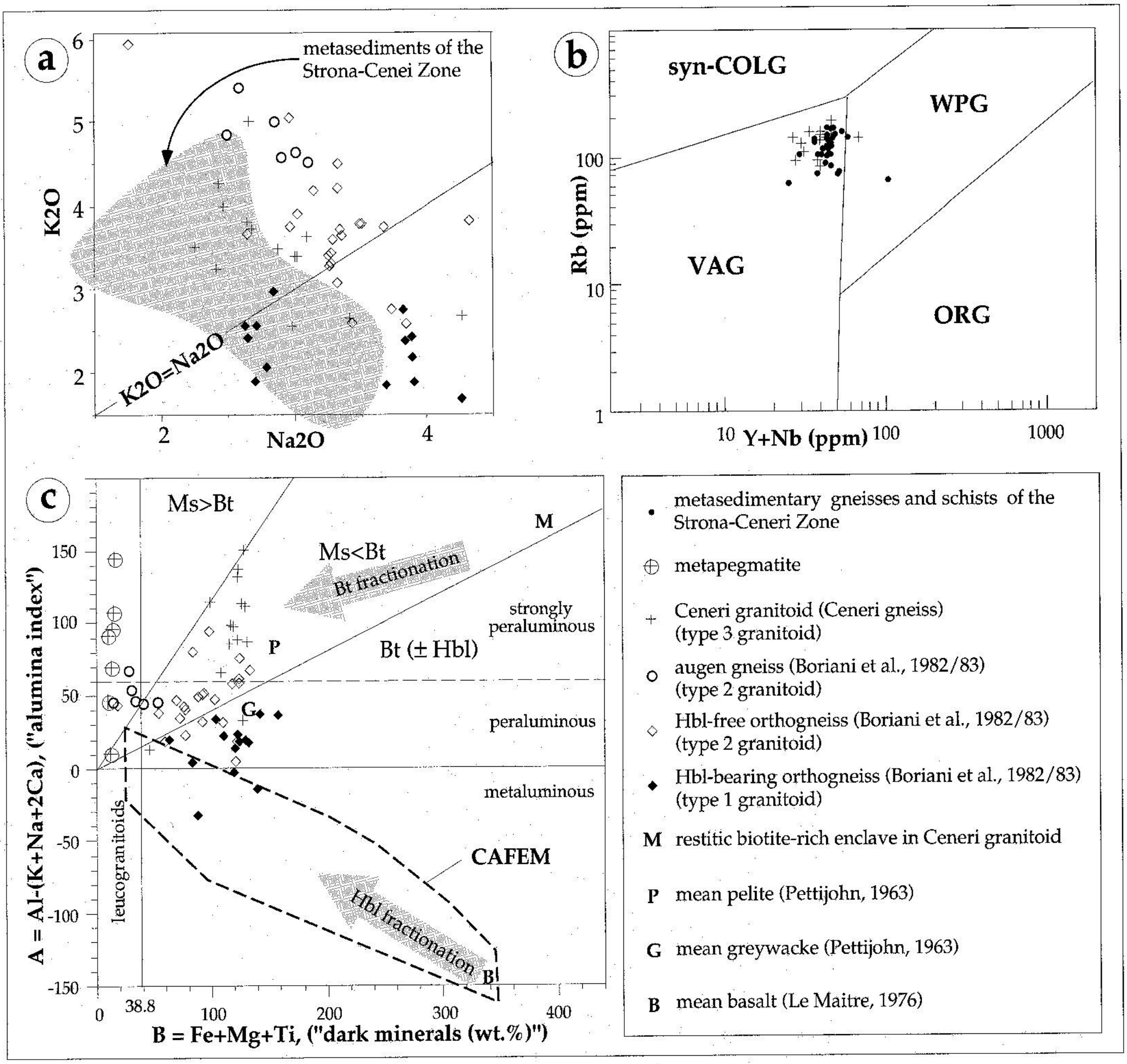

Fig. 8 Geochemistry of the Ordovician granitoids of the Strona-Ceneri Zone: (a) K_2O vs Na_2O diagram; (b) Rb vs Y+Nb discrimination diagram of PEARCE et al. (1984). Metasediments and Ceneri granitoids of the Strona-Ceneri Zone plot in the same part of the diagram. VAG: volcanic arc granites, syn-COLG: syn-collision granites, WPG: within plate granites, ORG: ocean ridge granites; (c) A-B diagram of DEBON and LE FORT (1988) for Ordovician granitoids and metasedimentary gneisses and schists of the Strona-Ceneri Zone. See text for explanation.

biotite-rich enclaves (which they interpret as sedimentary clasts), local gradations in grain size (interpreted as graded bedding), mineralogically zoned calc-silicate nodules (zonation interpreted to be of diagenetic origin). Although the authors' hypothesis cannot be dismissed out of hand, much of the evidence listed above is equivocal and can also be invoked to support an anatectic origin of the Ceneri granitoid. Besides the fact that BÄCHLIN'S (1937) model is incompatible with modern structural data, we consider it unlikely that a stratigraphic sequence can be recognized in the Strona-Ceneri Zone after three phases of deformation (D1–D3), each involving mylonitic shearing, tight to isoclinal folding and amphibolite facies regional metamorphism.

Figure 8c yields some critical information for the Ordovician magmatism in the Strona-Ceneri Zone. As a whole this magmatism is clearly peraluminous. Ceneri granitoids (type 3) with their metapegmatites (representing a first melt-composition) and biotite-rich restites (M in Fig. 8c) define a biotite fractionation trend which is a further indication for partial melting of a pelitic greywacke source rock. In comparison, some of the hornblende-bearing tonalites (type 1) reach into the uppermost CAFEM field (DEBON and LE FORT, 1988) typical for mantle-derived granitoids which evolved by fractionation of mafic minerals from a basaltic source rock. Note that peraluminous granitoids (type 2 and 3) and metaluminous CAFEM-suites roughly correspond in a descriptive sense to CHAPPELL and WHITE'S (1974) S- and I-type granitoids, respectively.

The origin of the Ceneri granitoid as an anatectically derived, S-type intrusive has consequences, first for the nature of Ordovician magmatism and second for the use of geochemistry to discern the tectonic setting of early Paleozoic events in the Strona-Ceneri Zone. Ordovician granitoids with a strong crustal signature (types 2 and 3 above) make up about 85% of all granitoids in the Strona-Ceneri Zone and are dominantly potassic, as shown in figure 8a. This figure also shows that the composition of the Ceneri granitoids is similar to that of metasediments in the Strona-Ceneri Zone. Only the mantle-derived, hornblende-bearing (type 1) granitoids are significantly more sodic. Mafic end-members such as diorites and gabbros are lacking (see also Fig. 8c). Thus, the main element chemistry of the Ceneri granitoids is largely inherited from their anatectic source rocks. The same appears to be true of the trace-element geochemistry. When plotted in the Rb vs Y+Nb discrimination diagram of PEARCE et al. (1984) in figure 8b, both the Ceneri granitoids and their metasedimentary source rocks fall in the field of volcanic-arc granitoids, close to the fields for within-plate granitoids and syn-collisional granitoids. Following BARKER et al. (1992) we note that PEARCE et al. (1984) include forearc intrusives in their volcanic arc granitoid category, thus precluding a use of this diagram to distinguish forearc from volcanic arc settings.

The main point here is that the inherited metasedimentary geochemical signature of the Ordovician granitoids reflects the convergent margin setting of the hinterland from which the metasediments were derived prior to both D1 and D2. Considered alone, therefore, geochemistry does not provide unequivocal information about the pre-Variscan tectonic setting of the Strona-Ceneri Zone. However, integrating the geochemical data with the radiometric, petrological and structural information allows to test tectonic models for early Paleozoic events in the Strona-Ceneri Zone.

5. Constraints on the Age and Nature of Pre-Variscan Tectonics in the Strona-Ceneri Zone

A sketch of the possible pre-Variscan evolution of the Strona-Ceneri Zone in figure 9 serves as a basis for discussing the existing constraints on the age and setting of sedimentation, metamorphism and magmatism. We note at the outset that the lack of radiometric ages for metamorphism prior to the D2 event renders any reconstruction of the early tectonic history of the Strona-Ceneri Zone highly conjectural.

The metasediments in the Strona-Ceneri Zone derive from a very heterogeneous Proterozoic crust, as documented by detrital zircon ages ranging from 2.4 Ga to 600 Ma (GEBAUER, 1993 and references therein). Erosion and sedimentation therefore occurred sometime between Late Proterozoic (600 Ma) and Ordovician D2 metamorphism. There is presently no unequivocal information concerning the depositional setting of the metasediments. The intercalation of psammitic gneiss with pelitic schist (turbiditic sequences?), very fine grained quartz-biotite gneiss (former radiolarites?) and local calc-silicate layers (former deep sea carbonates?) in the Strona-Ceneri Zone indicates that they were probably deposited in a submarine environment along a plate margin. A Cadomian convergent margin (Fig. 9a), a post-Cadomian/pre-Ordovician rifted margin (Fig. 9b; GEBAUER, 1993), or an Ordovician trench/arc complex (Fig. 9c; ZURBRIGGEN, 1996) along the northern periphery of Gondwanaland are all possible tectonic settings for the accretion of metasediments derived from 600 Ma old and older continental crust.

The spatial and temporal relationship between D1 and D2 events in the Strona-Ceneri Zone is unconstrained. D1 deformation and amphibolite facies metamorphism must have occurred sometime between sedimentation and Ordovician D2 metamorphism/magmatism. Similarly, relic eclogites (eclogitic garnet-amphibolites) in the Strona-Ceneri Zone must post-date the (pre-Cadomian or Cambro-Ordovician?) age of the oceanic amphibolites (BULETTI, 1983; GIOBBI ORIGONI et al., 1997) from which they derived, but pre-date Ordovician D2 metamorphism under amphibolite facies conditions. Given these broad age constraints, burial and subsequent near-isothermal exhumation of these eclogites (Fig. 5a) is expected to have occurred during either Cadomian subduction (e.g., SCHMID, 1993), or Ordovician subduction (ZURBRIGGEN, 1996; see below). A Cadomian age for the eclogites would be consistent with ubiquitous Late Proterozoic to Early Paleozoic high-pressure metamorphism related to the Cadomian orogeny in pre-Mesozoic basement units across central and western Europe. On the other hand, the occurrence of 460–470 Ma eclogites in the Gotthard Massif (OBERLI et al., 1994), a Penninic basement unit north of the Strona-Ceneri Zone, and in the Eastern Alpine Ötztal-Stubai complex (HOINKES and THÖNI, 1993), suggests that pre-Variscan subduction, metamorphism and magmatism in the Alpine domain occurred later, within a short interval in Ordovician time. It is therefore quite conceivable that eclogite facies metamorphism, as well as both D1 and D2 tectonometamorphic phases in the Strona-Ceneri Zone all represent part of a single (Ordovician) orogenic event (ZURBRIGGEN, 1996).

Ordovician magmatism and amphibolite facies D2 deformation is depicted in figure 9c to have been related to subduction of oceanic crust and/or to collision with a previously rifted Gondwanan fragment. The exact nature of Ordovician tectonics is unclear, partly due to aforementioned

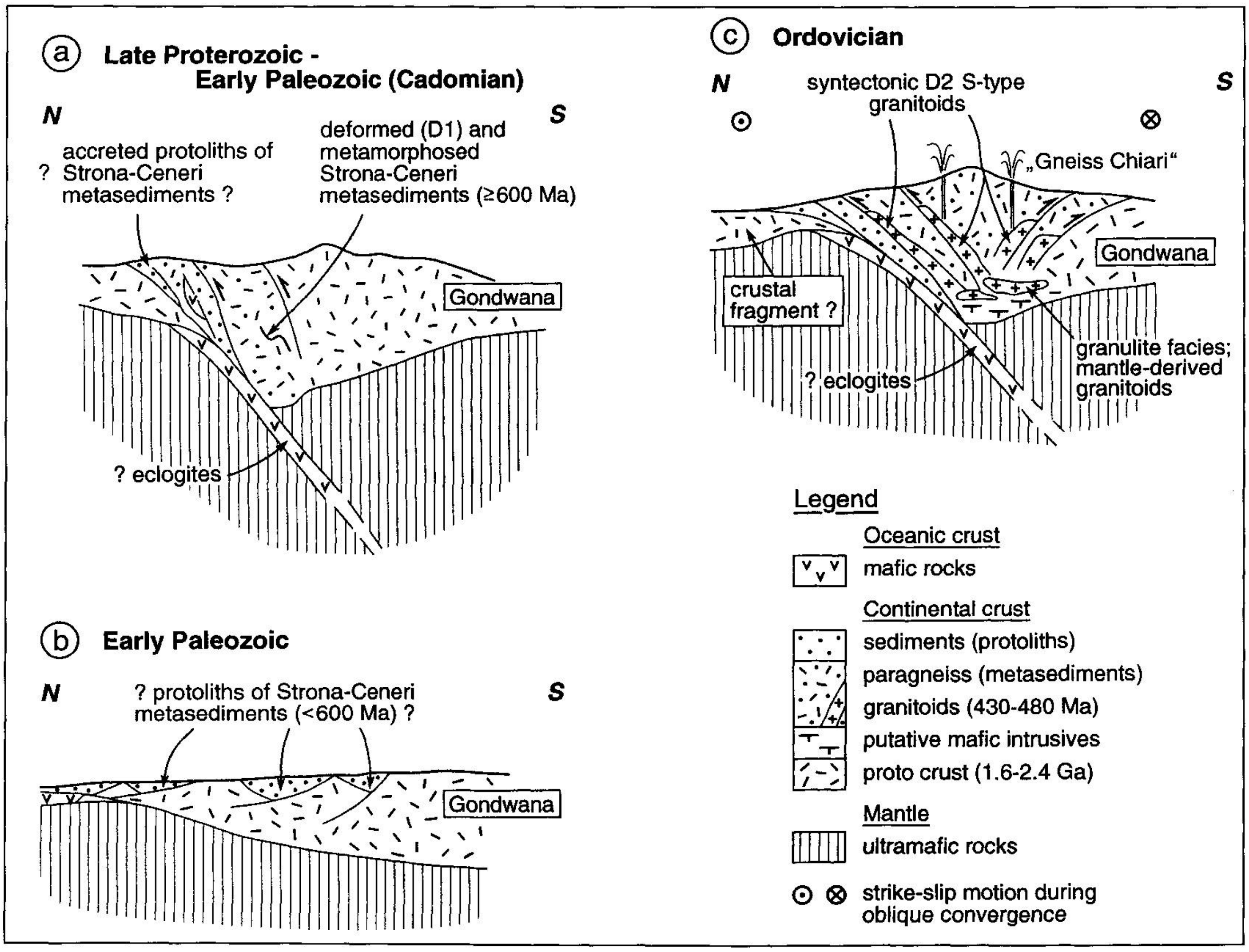

Fig. 9 Pre-Variscan tectonometamorphic evolution of the Strona-Ceneri Zone: (a) Cadomian orogenesis; (b) Early Paleozoic rifting; (c) Ordovician convergence and magmatism. See text for discussion.

lack of radiometric age determinations on the relic eclogites, but also because the Strona-Ceneri Zone represents only an intermediate segment of the Ordovician crust; the upper and lower parts of this crust are unexposed and may have been lost to erosion and subduction during Carboniferous to Early Permian time (HANDY et al., submitted). Oblique convergence during D2 is consistent with local structural arguments that the S2 schistosity in the Strona-Ceneri Zone was moderately to steeply dipping at its inception in Ordovician time (see ZURBRIGGEN, 1996). The setting depicted in figure 9c for the Strona-Ceneri Zone may have been analogous to that of the North American Cordilleran magmatic arcs, where oblique convergence and short episodes of rapid uplift or "surge" were accommodated by localized, melt-enhanced shearing (HOLLISTER and CRAWFORD, 1986).

Whatever its nature (subduction and/or collision), Ordovician tectonics in the Strona-Ceneri Zone occurred sometime from 430–480 Ma, the time span during which most pre-Variscan granitoids were emplaced. Given that the Ordovician crust in the southern Alps comprised mostly Cadomian and/or post-Cadomian peraluminous metasediments, both subduction and collision are viable settings for the generation of potassium rich, S-type Ceneri granitoids. Ordovician magmatism probably also affected upper levels of the southern Alpine crust; BORIANI and COLOMBO (1979) concluded that the "Gneiss Chiari" from the Val Colla Zone and the Orobic basement (east of the Strona-Ceneri Zone in Fig. 1) is a pre-Variscan alkali rhyolite, making it a possible volcanic end-member of the Ordovician magmatic suite in the southern Alps.

ZURBRIGGEN (1996) proposed an alternative convergent margin scenario for the Strona-Ceneri Zone in which accretion, eclogitization of MOR-basalt, polyphase (D1 and D2) deformation, anatexis and magmatism are all considered to have occurred in Ordovician time (see Fig. 11–5 in ZURBRIGGEN, 1996, p. 181). In his model, the detrital protoliths of the Strona-Ceneri metasediments were deposited in a forearc setting. Subduction of oceanic crust was concomitant with the accretion and prograde D1 deformation and metamorphism of these sediments. During subduction, mantle-derived magmas are believed to have intruded the base of the accretionary complex. These putative magmas initiated deep-crustal, syn-D2 anatexis and lead to the production of peraluminous S-type granitoids of the forearc type (BARKER et al., 1992). This scenario is similar to that proposed by CROOK (1980) for the magmatic and tectonometamorphic reconstitution of the Paleozoic accretionary complexes comprising the Lachlan fold belt of SE Australia.

We emphasize that the available evidence does not allow us to favour or reject outright any of the models described above. All of these scenarios are possible for the Strona-Ceneri Zone pending the acquisition of sorely needed geochronological data, particularly on the age of eclogitization and D1 metamorphism.

6. Conclusions

The evidence presented in this paper indicates that the main schistosity and associated metamorphism in the Strona-Ceneri Zone are Ordovician age, much older than the Carboniferous age previously proposed in some of the local literature (e.g., BORIANI et al., 1990). Whereas hornblende and mica ages in the Strona-Ceneri Zone yield Carboniferous metamorphic ages, the U–Pb isotopic systems in zircons and monazites as well as the Rb–Sr whole rock systems retain a selective memory of Ordovician magmatism and associated regional metamorphism (HUNZIKER et al., 1992 and references therein). This polyphase evolution resulting in superposed Carboniferous (Variscan) and Ordovician thermal regimes appears to be characteristic of most pre-Mesozoic basement units in the Alps.

Other pre-Mesozoic basement units in the Alps show similar pre-Variscan structures and lithologies to the Strona-Ceneri Zone. For example, the Paradis gneiss of the Gotthard Massif (ARNOLD, 1970), the Mönchalp granite of the Austroalpine Silvretta nappe (STRECKEISEN, 1928; POLLER, 1997) and the Winnebach granite of the Austroalpine Öztal-Stubai complex (HAMMER, 1925; KLÖTZLI-CHOWANETZ et al., 1997) are very similar in composition, structural characteristics and age to the Ceneri granitoid of the Strona-Ceneri Zone. OBERLI et al. (1994) present isotopic data from the Gotthard Massif pointing to a relatively short time interval of 470–440 Ma for subduction, eclogite facies metamorphism, granulite facies anatexis and magmatism. MERCOLLI et al. (1994) suggest a tectonic scenario for the Gotthard Massif in which Ordovician subduction of a Late Proterozoic accretionary prism is followed closely by eclogite, granulite and amphibolite facies metamorphisms and Late Ordovician magmatism. MAGGETTI and FLISCH (1993) propose a similar sequence of events for the Austroalpine Silvretta Nappe. In addition to Ordovician intrusives, the Silvretta Nappe comprises a group of "Older Orthogneisses" with ages around 520 up to about 610 Ma (SCHALTEGGER et al., 1997). The similar age range of the Mönchalp granite (lower intercept, 460 and 510 Ma U–Pb zircon ages; LIEBETRAU et al., 1994), one such Older Gneiss body in the Silvretta nappe, and the Ceneri granitoid in the Strona-Ceneri Zone indicate a similar Ordovician evolution for these units. FLISCH (1987) determined that the Silvretta nappe and Ötztal-Stubai complex have comparable rock types and an analogous polymetamorphic history. Therefore, all of these pre-Mesozoic basement units have undergone a similar, if not identical pre-Variscan polyphase evolution.

The ongoing discussion on the Paleozoic history of the Strona-Ceneri Zone manifests the fundamental difficulty of working in an old, polymetamorphic terrain for which only meager radiometric information on pre-Ordovician structures and mineral assemblages is available. Only further structural mapping combined with petrological and geochronological studies will resolve some of the long-standing issues addressed in this paper.

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