

Zeitschrift: Schweizerische mineralogische und petrographische Mitteilungen =
Bulletin suisse de minéralogie et pétrographie

Band: 77 (1997)

Heft: 3

Artikel: Pb isotope and trace element signatures of polymetamorphic rocks from
the Silvretta nappe, a comparison

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DOI: <https://doi.org/10.5169/seals-58488>

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Pb isotope and trace element signatures of polymetamorphic rocks from the Silvretta nappe, a comparison

by Viktor Köppel¹, Werner Hansmann¹ and Marino Maggetti²

Abstract

The polymetamorphic crystalline basement of the Silvretta nappe is composed of paragneisses, amphibolites, eclogites and igneous rocks which are broadly subdivided into an older, Late Precambrian to Cambrian group comprising gabbros, diorites, tonalites and granites and a younger, Ordovician group of mainly granites. Lead isotope signatures of major mineral phases from the magmatic rocks complement the whole rock trace element data and support in most cases the petrogenetic conclusions obtained from trace element signatures. Similar proportions of mantle and crustal lead were observed in meta-volcanic and basic to intermediate meta-igneous rocks with geochemical signatures indicative of a MORB, IAB and I-type VAG origin of their protoliths. Higher proportions of crustal lead were observed in an IAB-type gabbro. A MORB type eclogite and an S-type augengneiss contained crustal lead. The lead of a younger S-type COLG granitic gneiss is dominated by crustal lead, and metasediments were found to contain crustal lead with the highest μ value observed in the region.

The data pattern suggests that the magmas which originated from a depleted mantle source were contaminated by a crustal Pb-component from the metasediments. The lead isotope ratios of meta-igneous rocks indicate an increasing contribution of crustal lead as cratonisation progressed and thereby reflect the evolution of the crystalline core of the Silvretta nappe from a primitive island arc to a mature crust.

Keywords: Pb isotopes, trace element data, depleted mantle, crustal lead, cratonisation, Austroalpine.

Introduction

While geochemical characteristics of magmatic rocks commonly serve to distinguish different tectonic settings, lead isotope signatures reflect the principal reservoirs of the lithosphere in which their magmas were generated. These major reservoirs are the upper crust where high μ - ($= {}^{238}\text{U}/{}^{204}\text{Pb}$) and high W-values ($= {}^{232}\text{Th}/{}^{204}\text{Pb}$) prevail, the lower crust with low μ - and high W-values, and the normal, i.e. more or less depleted mantle with low μ - and W-values. In contrast to most trace elements, lead concentrations of mantle derived rocks are low and their Pb-isotope ratios are therefore sensitive to contamination and therefore offer the opportunity to detect minor contaminations.

The formation of new continental crust commonly begins with the formation of primitive island arcs along subduction zones. At this stage

mantle material and sediments will be involved, possibly also fragments of old continental crust. Active island arcs will progressively cratonise and accrete to continental margins. The lead isotope pattern of the accompanying magmatism will reflect this trend. In the early formed rocks the mantle signature of lead will remain discernible while at later stages, when sediments and possibly older crustal material are increasingly involved in orogenic processes, the rocks will obtain a transient lead isotope signal of the average continental crust. The lead isotopes will then evolve depending on the μ - and W-values of the newly formed rocks (e.g. DOE and ZARTMAN, 1979).

The composition of the lead isotope signal at the time of crystallization or metamorphic recrystallization is best represented in minerals with very low μ - and W-values such as feldspars from igneous rocks or galenas from associated sulfide deposits. Such lead-isotopic ratios reflect the time

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integrated μ - and W-values of a relatively large volume of the crust and thus they are reasonably representative for the continental crust. The evolution of average continental crustal lead has been modelled, for example, by STACEY and KRAMERS (1975).

This study aimed at testing whether or not lead isotopes of polymetamorphic igneous rocks yield information that is compatible with their geochemical signatures. Previous lead isotope studies of metamorphic rocks demonstrated that the mobility of lead across lithological boundaries is generally restricted to a few tens of meters or even less (CUMMING et al., 1987; PERETTI and KÖPPEL, 1986). CURTI (1987) demonstrated that the lead in mica-poor metapsammities from the Monte Rosa area did not homogenise under Alpidic lower am-

phibolite facies conditions. Mobility is, however, greatly enhanced by fluid flow in shear zones and in rocks with abundant water bearing phases such as micas.

Ten samples representative of the major lithologic units of the Silvretta basement rocks were selected for lead isotope analyses (Tab. 3). They comprise a MORB-type eclogite and an amphibolite, two gabbros with an island-arc signature, a metatonalite and metadiorite with a volcanic arc signature, an older, S-type augengneiss with a volcanic arc and a younger S-type augengneiss with a collisional granite signature. Two paragneiss samples were included. Their major and trace element compositions are listed in table 1 and 2. The characterization of the older and younger orthogneisses is based on the SiO_2 -Zr/TiO₂ (FLOYD and WINCHES-

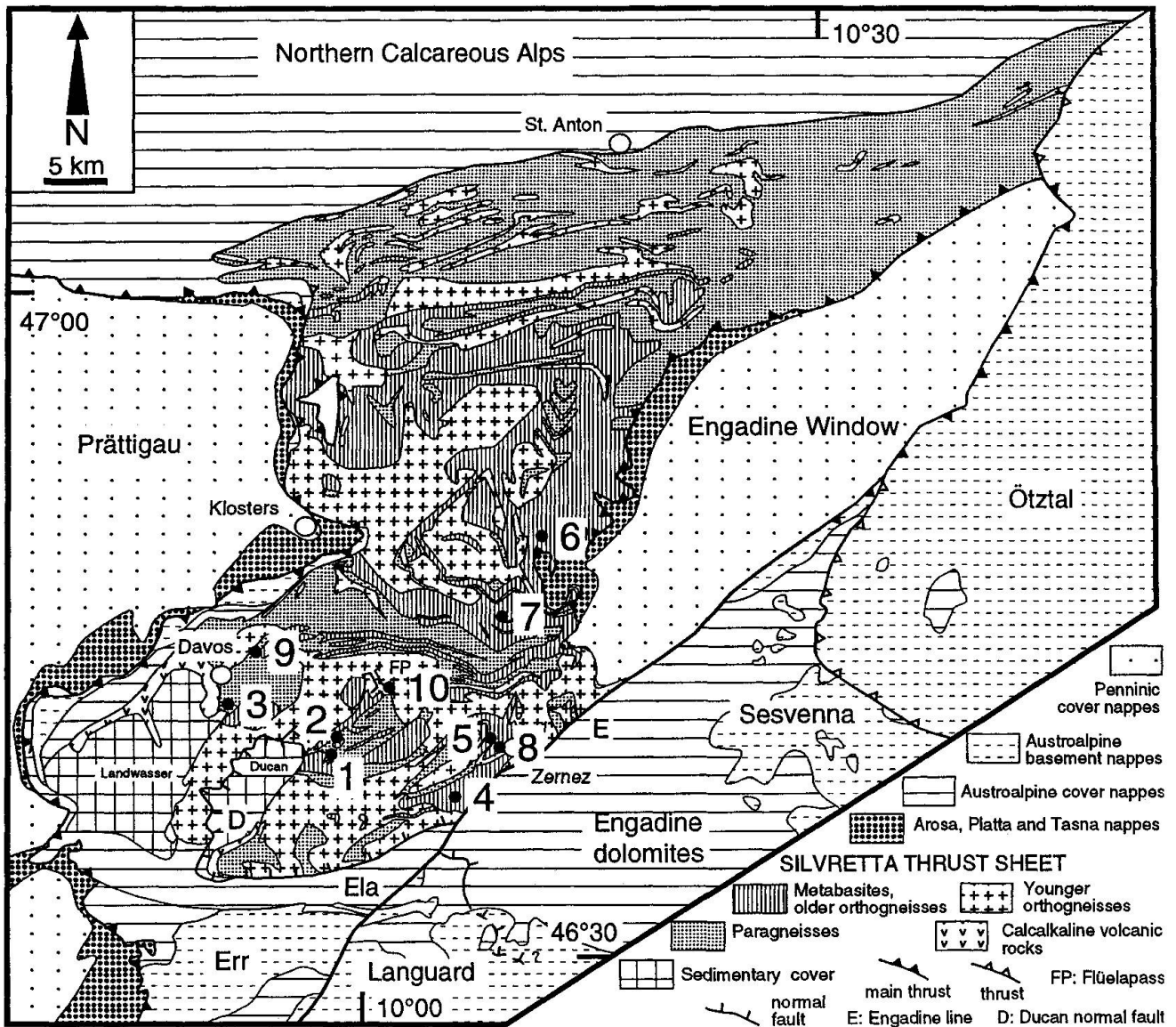


Fig. 1 Generalised geologic map of the Silvretta thrust sheet (after MAGGETTI and FLISCH, 1993).

Sample numbers: 1 = RB 64, 2 = RB 92, 3 = Ma 640, 4 = Ma 124a, 5 = ST 221, 6 = Ma 1061, 7 = Ma 252, 8 Ma 193, 9 = Ma 943, 10 = Si 586.

Tab. 1 XRF Analyses.

Type Sample weight%	A Ma 640	E Ma 124a	O St 221	O Ma 1061	O Ma 252	O Ma 193	O Ma 943	Y Si 586
SiO ₂	49.19	48.50	49.83	54.72	57.84	67.73	72.99	75.86
TiO ₂	2.26	1.68	0.42	0.60	0.53	0.29	0.39	0.11
Al ₂ O ₃	14.82	16.10	12.12	19.02	16.33	16.10	13.49	12.67
Fe ₂ O ₃	6.29	3.46	1.52	2.57	2.27	1.21	0.44	0.22
FeO	7.25	9.01	8.91	4.42	5.71	2.26	2.07	0.96
MnO	0.34	0.21	0.21	0.14	0.14	0.09	0.04	0.03
MgO	5.94	7.21	16.21	4.93	4.10	1.59	1.04	0.30
CaO	6.98	11.11	5.55	8.64	8.08	4.21	1.23	0.42
Na ₂ O	4.68	3.72	1.87	3.27	2.71	4.18	2.64	3.03
K ₂ O	0.58	0.14	0.42	0.22	0.80	1.41	3.68	0.96
P ₂ O ₅	0.31	0.14	0.02	0.16	0.11	0.12	0.10	0.13
L.O.I.	1.08	0.04	2.75	1.76	0.63	1.06	1.75	1.08
Total	100.08	100.36	99.83	100.45	99.35	100.25	99.86	99.77
Fe _{tot} (Fe ₂ O ₃)	14.35	13.65	11.42	7.48	8.71	3.63	2.74	1.28
ppm								
Nb	19	6	1	2	—	—	7	6
Zr	109	98	18	54	55	105	142	67
Y	41	66	2	12	13	9	21	17
Sr	147	47	249	618	238	334	148	41
Rb	2	3	4	2	30	38	107	203
Th	—	1	1	—	—	7	14	14
Pb	—	10	1	3	—	—	25	17
Ga	21	16	9	15	17	18	15	13
Zn	113	101	80	73	112	65	44	13
Cu	46	6	1	1	22	18	19	14
Ni	36	63	61	45	20	15	22	2
V	342	321	169	213	238	62	39	8
Cr	77	199	747	115	52	25	26	9
Ba	127	31	83	159	318	726	769	100

L.O.I. = loss on ignition, P = paragneiss, E = eclogite, A = amphibolite, O = older orthogneiss, Y = younger orthogneiss

TER, 1978) and the Zr/TiO₂-Nb/Y (WINCHESTER and FLOYD, 1977) discrimination diagrams.

Geologic setting

The Silvretta thrust sheet (Fig. 1) is part of the pre-Permian basement of the Alps. MAGGETTI et al. (1990) distinguish four major groups of crystalline rocks: (1) paragneisses, (2) eclogites and amphibolites, (3) older orthogneisses and (4) younger orthogneisses (= "Flüelagranitic association"). The complex geologic evolution of this Gondwana polymetamorphic basement was summarised by MAGGETTI and FLISCH (1993). Detailed geochemical studies were undertaken during the last years by the Silvretta working group of the University of Fribourg. In the following, we present a short synthesis of the results.

The paragneisses comprise dominantly biotite-plagioclase gneisses and micaschists. BOLLIN et al. (1993 a, b) interpreted these rocks as cyclic

greywacky and pelitic turbidites which were deposited onto a shallow continental shelf. BIINO et al. (1994) reported Sm-Nd model ages of approx. 1.7 Ga for biotite-plagioclase gneisses and micaschists which represents an "average" source age of the detritus. Detrital zircons yielded upper intercept ages of 1.5 to 1.9 Ga, which also reflect an "average" Proterozoic source age of the sediments (GRAUER and ARNOLD, 1968).

The metasedimentary rocks are closely associated with eclogites and amphibolites that were derived from N- and T-type tholeiitic protoliths (MAGGETTI et al., 1987; MAGGETTI and GALETTI, 1988). According to MAGGETTI et al. (1987) the intercalations of these metabasic rocks with terrigenous paragneisses represent an original composite layering of igneous basic rocks and sediments. The geotectonic setting could correspond to a divergent oceanic plate boundary that was located close to a continent which provided detritus or, less likely, to a rifting stage of a continental plat-

Tab. 2 REE analyses, ppm, Ma 640 and Ma 252 by INA (MAGGETTI et al., 1987), all others by ICP-MS (SARM CRPG, Vandœuvre, France).

	Ma 640	Ma 124a	ST 221	Ma 1061	Ma 252	Ma 193	Ma 943	Si 586
La	18.4	5.73	1.742	8.12	7.4	12.03	30.41	6.17
Ce	41.0	15.3	3.929	19.8	16.7	22.34	60.74	15.08
Pr	—	1.75	0.544	2.81	—	2.38	6.99	1.72
Nd	24.8	7.93	2.26	11.99	7.8	8.24	27.18	6.42
Sm	6.95	2.34	0.681	2.82	1.90	1.66	5.2	1.84
Eu	2.27	0.72	0.517	0.96	0.71	0.62	1.36	0.15
Gd	7.5	3.39	0.615	2.41	—	1.427	4.66	1.80
Tb	1.29	0.8	0.121	0.259	0.305	0.193	0.66	0.41
Dy	—	5.89	0.757	2.112	—	1.19	3.57	2.79
Ho	—	1.46	0.186	0.434	—	0.252	0.79	0.62
Er	—	3.75	0.499	1.125	—	0.72	2.05	1.77
Tm	—	0.57	0.094	0.156	—	0.107	0.29	0.33
Yb	4.1	3.85	0.617	1.039	1.29	0.789	2.07	2.23
Lu	0.65	0.62	0.096	0.151	0.221	0.139	0.34	0.34

form with formation of new oceanic crust. Similar gneiss-amphibolite associations throughout the Alpine domain were interpreted as indicative of either incipient oceanic rifting or back arc environments (e.g. BOUCHARDON et al., 1989; PAQUETTE et al., 1989; MESSIGA et al., 1992; MILLER and THOENI, 1995). Another possibility is that the protoliths of the amphibolites and the paragneisses were originally generated in different geotectonic environments. The metabasic rocks may have accreted together with continental-derived sediments (= paragneisses) in a wedge along an active plate margin, and both were subsequently subducted. However, this model can hardly explain the intimate interlayering of both amphibolites and paragneisses.

Major and trace element chemistry of the amphibolite (Ma 640) and the eclogite (Ma 124a) point to MORB compositions (Fig. 2, Tab. 1, 2). The precursor rocks correspond to basic liquids (PEARCE, 1983). Both are, however, enriched in LIL-Elements (Fig. 3). MAGGETTI and GALETTI

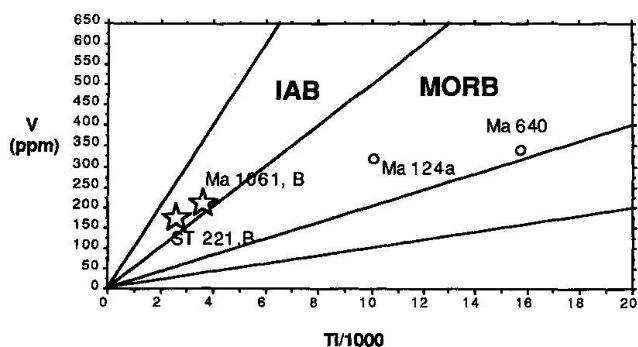


Fig. 2 V-Ti/1000 plot (SHERVAIS, 1982; HODDER, 1985). MORB = mid ocean ridge basalt, IAB = island arc basalt.

(1988) explained this by the mobile behaviour of the alkalis during post-magmatic alterations. Lithophile elements can also be enriched in magmas generated from a metasomatised subcontinental lithosphere, or by intracrustal contamination during ascent of basic magmas derived from a "normal" source. PAQUETTE et al. (1989) explained similar LILE enrichments of eclogites from the Alpine External Massifs by 4–5% crustal contamination of tholeiitic magmas which were emplaced into thinned continental crust during the initial stages of oceanic rifting.

Already in pre-Ordovician time the metasediments and basic rocks were multiply metamorphosed, first under amphibolite and eclogite and then again under amphibolite facies conditions (MAGGETTI and GALETTI, 1988). After decompression, they were intruded by the protoliths of the older orthogneisses that include ultramafic, gabbroic, dioritic to tonalitic and granitic rocks.

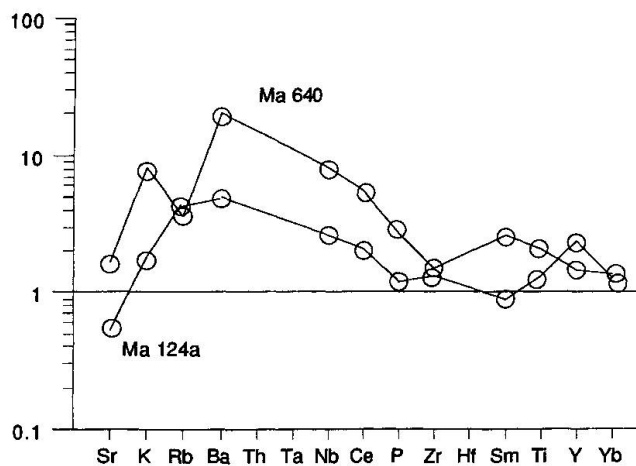


Fig. 3 N(normal)-MORB normalised plot of metabasites.

Zircon U–Pb ages of these rocks indicate crystallization ages of 610 Ma for a diorite and 530–520 Ma for gabbros, tonalites (SCHALTEGGER et al., this volume) and granitic gneisses (MÜLLER et al., 1995, 1996). The tholeiitic basic and the intermediate members of this suite exhibit I-type characteristics (MAGGETTI et al., 1990) whereas the granitic members are of S-type origin (POLLER, 1994 a, b) (Fig. 4). This magmatism has an overall island (volcanic) arc affinity (Fig. 5) (MAGGETTI et

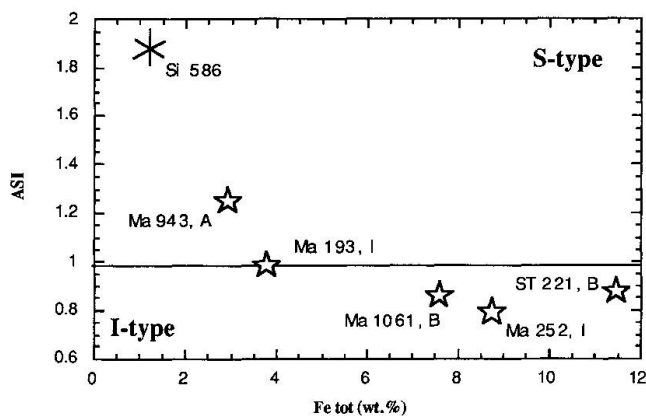


Fig. 4 ASI- Fe_{tot} plot to distinguish I- and S-type granitoids after NORMAN et al. (1992). ASI = aluminum saturation index.

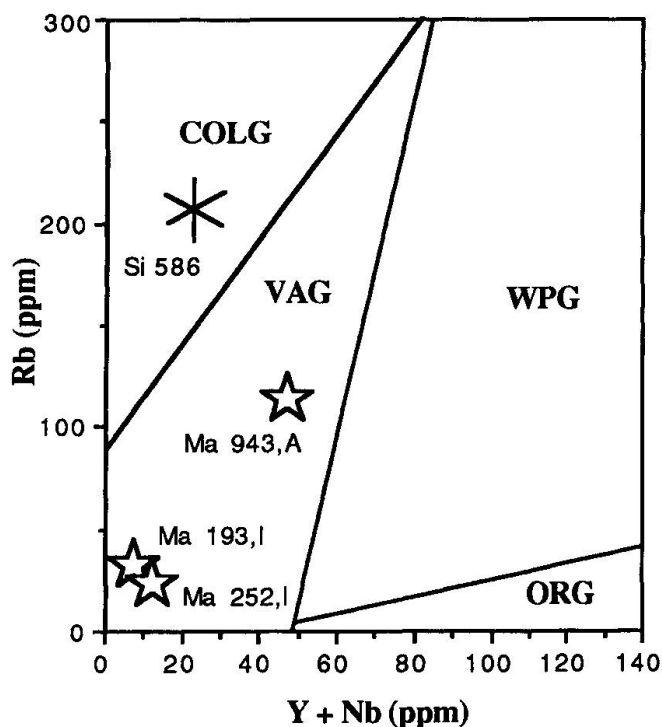


Fig. 5 Rb-(Y+Nb) discrimination plot for different tectonic regimes (PEARCE et al., 1984). COLG = syn-collisional granites, ORG = ocean ridge granites, VAG = volcanic arc granites, WPG = within plate granites.

al., 1990; POLLER, 1994 a, b). A detailed discussion of the trace element geochemistry is given by SCHALTEGGER et al. (this volume).

During a second period of magmatism (475–440 Ma) the protoliths of the so-called younger orthogneisses comprising gabbros and granitoids intruded (FLISCH, 1987; LIEBETRAU, 1995). The latter show S-type signatures (Fig. 4) (LIEBETRAU and NÄGLER, 1994; LIEBETRAU, 1995) and are related to collisional regimes (Fig. 5), documenting in the view of LIEBETRAU (1995) differing metasedimentary sources rather than a transition from an oceanic-continent to continent-continent collisional regime.

Nd model ages of 900 to 1000 Ma were reported of basic to intermediate igneous rocks of the older suite (SCHALTEGGER et al., this volume) and of ca. 1700 Ma for older as well a younger granitic gneisses (POLLER, 1994b; POLLER et al., 1994; LIEBETRAU, 1995). These results probably reflect variable mixtures of 500–600 Ma depleted mantle material with crustal components with a mean age of > 1700 Ma. During late Carboniferous the whole Silvretta basement underwent amphibolite facies metamorphism and pervasive deformation (FREI et al., 1995; PROSPERT and BIINO, 1996).

In summary, the metasediments of the Silvretta nappe contain mid-Proterozoic and older components which were deposited in a late Proterozoic oceanic trough. Subduction of this oceanic crust gave rise to two major magmatic cycles in Cambrian and Ordovician times. Cratonization ended with an intensive Variscan amphibolite facies metamorphism. Trace element and REE compositions show the influence of both mantle and crust.

Analytical procedure

The hand-picked mineral separates (plag, bi, musc, kf, hbl) were treated with hot (120 °C) 2N HCl for 60–80 minutes. After removal of the acid, they were rinsed 2× with distilled water and dried. By this treatment only biotite was markedly corroded. Samples weighing between 25–120 mg were dissolved in 4 ml of a mixture of 6N HCl and 2.3N HF. After complete dissolution the clear solutions were evaporated to dryness and redissolved in 2 ml of 6N HCl. A ^{205}Pb tracer was added to the solution which then was again evaporated to dryness. The evaporate was dissolved in a 12:1 mixture of 1.5N HBr and 2N of HCl.

Pb was separated using ion-exchange columns filled with 0.5 ml of DOWEX AG1x8 resin (100–200 mesh) and applying an HBr procedure. The purified resin was preconditioned with 2 resin

Tab. 3 Fractionation corrected lead-isotope ratios and lead concentrations of mineral separates.

Sample	$^{206}\text{Pb}/^{204}\text{Pb} \pm$	$^{207}\text{Pb}/^{204}\text{Pb} \pm$	$^{208}\text{Pb}/^{204}\text{Pb} \pm$	ppm Pb
RB 64 paragneiss (Knotengneiss), N Scalettapass 789 690/174 510				
plag (V)	18.297 .011	15.661 .012	38.580 .050	8.19
bi (V)	19.032 .013	15.695 .015	39.657 .048	5.20
RB 92 paragneiss (Knotengneiss), Dischmatal, Scalettariser 790 000/175 400				
bi (V)	18.639 .012	15.673 .015	38.991 .047	4.60
plag (V)	18.266 .009	15.654 .007	38.398 .046	20.80
Ma 640 amphibolite (MORB), Sertig, Rhynerhorn 782 250/177 900				
hbl (V)	20.474 .025	15.706 .029	40.278 .070	0.39
repeat new sample	20.486 .012	15.712 .014	40.291 .048	0.42
plag (V)	18.937 .013	15.597 .014	38.874 .054	1.59
Ma 124a eclogite (MORB), Inntal, Punt Nova 800 900/172 700				
fsp (pV)	18.178 .013	15.610 .016	38.338 .049	4.04
px (pV)	18.298 .013	15.600 .016	38.412 .050	2.95
ST 221 older orthogneiss, gabbro, Val Sarsura 798 250/176 050				
plag (pV ?)	18.157 .013	15.585 .015	37.804 .045	1.98
Ma 1061 older orthogneiss, gabbro, Val Tuoi 805 480/188 720				
hbl (V)	18.602 .011	15.576 .014	38.127 .055	0.37
repeat, new sample	18.584 .014	15.567 .016	38.101 .048	0.42
plag(V)	18.005 .011	15.539 .013	37.668 .045	6.75
Ma 252 older orthogneiss, metadiorite, Val Sagliains 799 400/186 000				
plag (V)	19.805 .011	15.664 .013	38.007 .045	5.21
hbl (V)	21.385 .015	15.751 .015	38.795 .049	12.03
Ma 193 older orthogneiss, bi-bearing metatonalite, Inntal, God Praditsch 802 350/176 500				
bi (V)	19.321 .018	15.612 .015	37.835 .037	0.53
repeat	19.314 .034	15.606 .027	37.822 .070	0.70
plag (V)	19.215 .013	15.601 .015	37.742 .048	5.34
Ma 943 older orthogneiss, augengneiss type Mönchalp, E Tällikopf, Mönchalp 789 800/189 500				
bi (V)	137.663 .039	21.939 .006	42.638 .050	7.50
plag (V)	18.881 .026	15.646 .022	42.208 .058	0.34
musc (V)	18.135 .011	15.598 .012	38.642 .045	3.85
Si 586 younger orthogneiss, augengneiss type Flüela, Radönt 793 220/178 760				
musc (V)	19.636 .013	15.700 .016	38.650 .047	2.23
kf (V)	18.356 .011	15.638 .013	38.099 .045	36.89

V = Variscan, pV = pre Variscan

The quoted errors are 2σ standard deviations and include the error introduced by the fractionation correction.

volumes of 1N HBr. After the sample was loaded onto the column, the resin was flushed with 3 volumes of 1N HBr and 1 volume of 2N HCl. Pb was then extracted by adding 3 volumes of 7.5N HCl. Pb was purified by repeating the same procedure using 0.25 ml of resin volume and accordingly reduced acid volumes. The analytical Pb blank contribution of 0.2 ng was in all cases < 0.5% of the total Pb and thus was negligible.

Pb was loaded on Re filaments and analyzed with a Finnigan MAT 261 thermal ion mass spec-

trometer. Fractionation effects on the isotope ratios were corrected by a factor

$$c = 1 + fxDamu$$

($f = 1.3 \pm 03$ permil [2σ]; Damu = difference in atomic mass units between numerator and denominator of the isotope ratio), determined by replicate analysis of the NBS standard reference material 981.

Experience has shown that occasionally lead isotopes from samples may fractionate during mass spectrometric analysis far in excess of the

fractionation observed when analysing the NBS standard lead (KNILL, 1996). This excessive fractionation is not recognisable by a trend of increasing isotopic ratios during analyses. We therefore analysed two minerals of most of the rock samples. Extremely flat or steep slopes in the $^{207}\text{Pb}/^{204}\text{Pb}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams would then either indicate that one of the samples fractionated excessively or else that the mineral pair was not in isotopic equilibrium.

Results and discussion

Lead isotope analyses are listed in table 3 and the data points are plotted in figure 6. Because of their polymetamorphic nature we did not attempt to correct the isotopic ratios for in situ decayed U and Th and therefore did not determine their U and Th concentrations. The discussion of the present-day lead isotope ratios is therefore focused on the coupled uraniumogenic $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios. In most cases the feldspar lead is comparatively unradiogenic which agrees with the common observation that the U/Pb ratio in feldspars is lower than in other silicates and thus approximates the initial lead composition at the time of feldspar (re-)crystallization. An exception is the feldspar-lead of sample Ma 943 which is more radiogenic than that of the muscovite. According to microscopic evidence most of the minerals (re-)crystallized during the Variscan metamorphism, exceptions are the eclogite (Ma 124a) and the plagioclase of the gabbro (ST 221). The

plagioclase of the eclogite forms a symplectitic intergrowth with hornblende which is interpreted as a retrograde reaction product during uplift.

The tie lines (Fig. 6) between coexisting mineral pairs represent isochrons if the minerals contained the same initial common lead at the time of crystallization and if their U-Pb system remained closed in the past. Because of the relatively small spread of the data the error of the slopes of the tie-lines is high and thus prevents the calculation of meaningful ages. For example, the slope of sample Si 586 ($s = 0.0456 \pm .016$) would still be compatible with a Variscan isochron age of 330 Ma ($s = 0.0530$). The scatter of the slopes in figure 6 can be entirely attributed to the analytical uncertainties stated in table 3. Only the biotite of sample Ma 943 yielded a sufficiently high $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of 137.7 to allow in combination with the coexisting plagioclase and muscovite the calculation of a $^{207}\text{Pb}/^{204}\text{Pb}$ age of 330 ± 5 Ma. The age reflects the Variscan metamorphism in the Silvretta nappe (FREI et al., 1995). The highly radiogenic lead is probably caused by the presence of a uranium rich phase in the analyzed sample.

The isotope signatures of the data points cover a spectrum ranging from low $^{207}\text{Pb}/^{204}\text{Pb}$ (low μ) ratios relative to $^{206}\text{Pb}/^{204}\text{Pb}$ to relatively high $^{207}\text{Pb}/^{204}\text{Pb}$ ratios (high μ) (Fig. 6). The lead of the paragneisses yielded the highest μ values (9.9 to 10) which indicates a long residence time of the lead in a relatively uranium rich environment. Such a signature is typical for paragneisses of the East Alpine nappes and of the Southern Alps (KÖPPEL et al., 1993).

Most of the basic to intermediate members of the older orthogneiss suite contain a mantle lead component, i.e. the metagabbro (Ma 1061), the metatonalite (Ma 193) and within the analytical uncertainty also the metadiorite (Ma 252). The volcanic-arc gabbro (ST 221) and the S-type augengneiss (Ma 943) contain lead that is isotopically similar to average crustal lead. Of the two samples from the amphibolite-eclogite group, the lead of the amphibolite (Ma 640) is only little contaminated by crustal lead. The Pb-concentrations of its the major constituents (0.4 to 1.6 ppm, Tab. 3) suggest a somewhat higher whole rock lead concentration relative to N-type MORB (< 1 ppm). The amphibolite is significantly enriched in LILElements (Fig. 3). The eclogite (Ma 124a) consisted prior to retrograde reequilibration of ca. 35% garnet and 55% omphacite plus minor amounts of zoisite and rutile. In contrast to the amphibolite, the eclogite displays an average crustal lead signature. The lead concentration of the pyroxene and retrograde plagioclase (3 to 4 ppm, Tab. 3) also point to higher whole rock lead concentra-

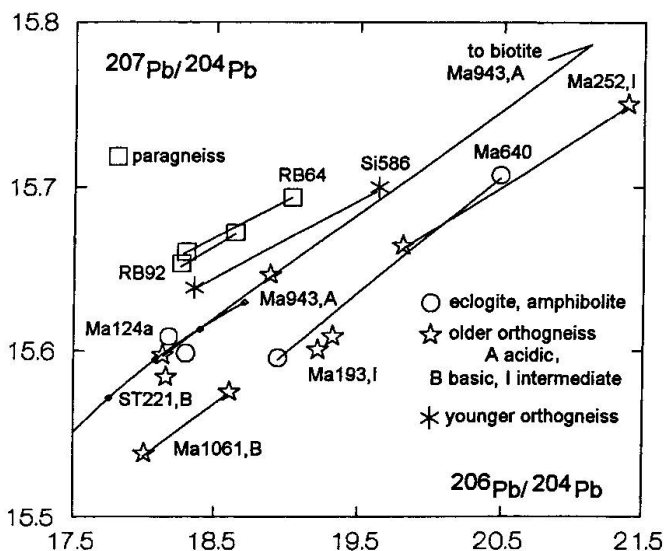


Fig. 6 (A) $^{206}\text{Pb}/^{204}\text{Pb}$ - $^{207}\text{Pb}/^{204}\text{Pb}$ -diagram with the growth curve for average Pb of the continental crust according to STACY and KRAMERS (1975) with time-marks at intervals of 100 Ma.

tions than one would expect in N-type MORB. The eclogite is also enriched in LIL-elements (Fig. 3), but less so than the amphibolite.

The data points do not reflect an origin of the lead solely from a depleted mantle. We have outlined in figure 7 the data field of the present-day MORB and the growth-curve for mantle lead according to ZARTMAN and HAINES (1988). The growth curve with $\mu = 40$ illustrates the lead-evolution of a mantle derived rock with an initial $^{206}\text{Pb}/^{204}\text{Pb} = 17.43$ and $^{207}\text{Pb}/^{204}\text{Pb} = 15.37$ and an age of 600 Ma which corresponds to the oldest reported age of igneous rocks of the Silvretta nappe (SCHALTEGGER et al., this volume). Evidently none of the samples contains a lead that was solely derived from a depleted mantle source. This would also hold if we had chosen a more radiogenic initial lead-composition, comparable to the relatively radiogenic lead of some of the present-day MORB. Note that in spite of the high μ value the growth curve plots below the mantle curve of ZARTMAN and HAINES (1988). This is a result of their model which accounts for recycling of crustal material with high $^{207}\text{Pb}/^{204}\text{Pb}$ into the mantle. To produce the $^{207}\text{Pb}/^{204}\text{Pb}$ ratios observed in

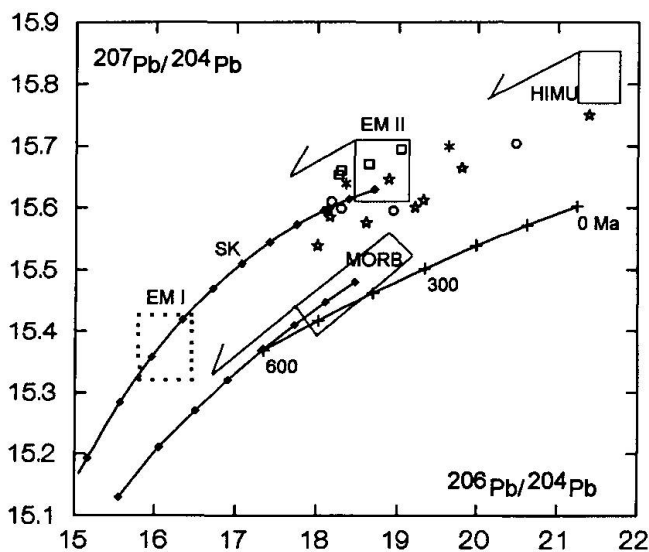


Fig. 7 $^{206}\text{Pb}/^{204}\text{Pb}$ – $^{207}\text{Pb}/^{204}\text{Pb}$ -diagram with the growth curve (SK) for average Pb of the continental crust according to STACY and KRAMERS (1975) and time-marks at 200 Ma intervals. MORB designates the field the present-day Pb-isotopic composition of mid-ocean ridge basalts. The mantle growth curve by ZARTMAN and HAINES (1988) indicates the direction along which the MORB field has moved in the past. A growth curve with time-marks at 100 Ma intervals and with $\mu = 40$ is shown which originates at the mantle growth curve 600 Ma ago. EM I, EM II and HIMU denote components of isotopically anomalous mantle lead. The arrows indicate the position of these fields ca. 600 Ma ago. Symbols as in figure 6.

the samples would require that mantle derived magmas with μ values between 20 and 25 were generated 2000 Ma ago. There is no evidence, however, of such high protolith ages of the I-type meta-granitoids or of the amphibolites. Moreover, the $^{206}\text{Pb}/^{204}\text{Pb}$ ratios would in this case vary between about 21.5 and 23.3.

Excluding a crustal lead contribution during magma genesis, ascent and cooling, the lead isotope pattern of the basic to intermediate members of the older orthogneisses thus suggests an origin from an anomalous mantle. An anomalously enriched mantle may result from metasomatism by dewatering reactions of subducted sediments (EM II, ZINDLER and HART, 1986) such as the paragneisses (Fig. 7). A further possibility would be a derivation from an EM I type mantle which was contaminated by metasomatism originating, for example, from subducted old lower crust which normally has low U/Pb and high Th/Pb ratios. Although such an origin cannot be ruled out, it appears unlikely because the uraniumogenic lead is not obviously retarded and the thorogenic ratios are not abnormally high. A contribution from a HIMU mantle source is unlikely in view of the relatively low $^{206}\text{Pb}/^{204}\text{Pb}$ but high $^{207}\text{Pb}/^{204}\text{Pb}$ ratios of most of the data (Fig. 7).

Alternatively, if the source region of the magma had approximately the isotopic characteristics seen in present-day MORB, a contribution of crustal lead from metasediments during magma genesis and emplacement would also explain the data. Assuming that the lead contaminant derived from sediments (RB 64 and RB 92), then their Pb-contribution to samples Ma 640, Ma 1061 and Ma 252 may be estimated from figure 6 to amount to 25% to 50% depending on the Pb-isotopic composition of the mantle lead. Assuming the same mixing components, the meta-tonalite (Ma 193) would then be somewhat less, and the meta-gabbro (ST 221) and the eclogite (Ma 124a) more contaminated (50–80%). Due to the contrasting Pb concentration of mantle derived magmas (approx. 1 ppm) and rocks of the continental crust (> 10 ppm) contamination by assimilation of sediments would have amounted to approximately 2% to 5% which is in reasonable agreement with the estimate of a maximum contribution of 10% on the basis of Sm/Nd isotopic systematics (SCHALTEGGER et al., this volume). A further possibility is contamination by submarine hydrothermal alteration in the presence of sediments. It certainly could have affected the basalts which were the protoliths of the amphibolites. Also the crustal lead isotope signature of the eclogite (Ma 124a) may be the result of a hydrothermal sea-floor alteration. The eclogites occur as alternating, 5 to

Tab. 4 Summary of geochemical characteristics.

Sample	V-Ti/1000	Zr/Y-Zr	Ti-Zr	Cr-Y	Rb-(Y+Nb)	Fe(tot)-ASI	Pb-Isotope signature
Eclogite Ma 124a	MORB	MORB(?)	MORB	MORB	-	-	average crust
Amphibolite Ma 640	MORB	MORB	MORB(?)	MORB(?)	-	-	mantle-crust
Gabbro Ma 1061	IAB	-?	IAB	IAB	-	I-type	mantle-crust
Gabbro ST 221	IAB	-?	IAB	-?	-	I-type	crust-mantle
Biotite bearing metatonalite Ma 193	-	-	-	-	VAG	I-type	mantle-crust
Metadiorite Ma 252*	-	-	-	-	VAG	I-Type	mantle-crust
Augengneiss (type Mönchalp) Ma 943	-	-	-	-	VAG	S-type	average crust
Augengneiss (type Flüela) Si 586	-	-	-	-	COLG	S-Type	crust

* The discrimination diagrams are not strictly applicable to this rock as its low quartz content originates probably from metamorphic reactions.

100 cm thick layers of differing eclogitic modal composition within a 500 m thick complex of metabasites which is surrounded by metasediments. Thus, the protoliths of the eclogites were most likely oceanic basalts.

Both augengneisses, the older (Ma 943) and the younger orthogneiss (Si 586) are, based on the SiO₂-ASI discrimination diagram, S-type granites. Accordingly, the uraniumogenic lead of the older orthogneiss rocks is of average crustal composition and is thus typical of lead of mature arcs, and the lead of the younger orthogneiss exhibits a slightly more pronounced crustal signature.

Conclusions

The lead isotope data support in most cases the interpretation of the geochemical signatures (Tab. 4). The isotopic signature of the amphibolite (Ma 640) indicates the presence of a mantle component in agreement with its geochemical characteristics that suggest a MORB origin. The geochemical signatures of the four I-type intrusives indicate an island arc basalt and volcanic arc setting which agrees with the presence of a mantle derived lead component in these samples. Based on the aluminum saturation index the VAG-type metagranite (Ma 943) and a syncollisional metagranite (Si 586) are S-type granites. Their isotope signature indicates a crustal origin of the lead and thus supports an S-type origin.

In spite of the polymetamorphic nature of the meta-igneous rocks, their geochemical and lead isotope signatures are in the majority of cases compatible. The only exception to this is the eclogite (Ma 124a) which contains an average crustal lead but the geochemical signatures are indicative

of a MORB origin. However, from its intimate association with metasediments one may plausibly conclude that hydrothermal sea floor alteration was responsible for the crustal signature of the lead.

The lead isotope data support the idea that new crust was generated first by mantle differentiation in an island arc situation and subsequently with increasing participation of crustally derived metasediments which contained early Proterozoic detritus. The lead isotope data do not indicate the presence of remnants of an old, for example early Proterozoic or even Archean crust.

Acknowledgements

We thank I. Ivanov-Bucher and W. Wittwer for preparing the mineral separates and R. Frei for reviewing the manuscript; his critical comments were of great help in preparing the final version of this paper. We are grateful to C. Prospero who contributed figure 1.

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Manuscript received February 5, 1997; revised manuscript accepted August 25, 1997.