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Isotope data of metarodingites and associated rocks from the Lanzo and the Bracco ophiolitic massifs: indications on the evolution of the Alpino-type ultramafic-mafic complexes

by *U. Rösli*[1], *S. Hoernes*[2] and *V. Köppel*[3]

Abstract

Rodingites are widespread in all the ultramafic rocks of Alpine and Apennine ophiolites. They normally consist of calcium silicate minerals such as garnet, vesuvianite, zoisite, prehnite and chlorite. In spite of their clear mineralogical composition, the origin of the rodingites, the time relations between the hydrothermal alterations and the evolution during Alpine orogeny is often hidden. To get new data on this problem $^{18}O/^{16}O$, $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, $^{208}Pb/^{204}Pb$ and $^{87}Sr/^{86}Sr$-ratios have been measured on whole rock and mineral samples from rodingites and from their associated rocks. The studied rocks have been sampled in the eclogitized Lanzo massif (Piedmont) and the very low-grade Bracco massif (Eastern Liguria), both in Northern Italy. Diffusional zoning at the border of the rodingites is a common feature.

For all samples lead isotope data still show typical MORB composition. The $^{87}Sr/^{86}Sr$-value of one sample from the Lanzo massif is strongly enriched in ^{87}Sr (0.70794) and is similar to the suggested values of Cretaceous sea water. The samples of the Bracco massif are only slightly enriched (0.70359–0.70416). Both regions seem to have equilibrated with almost pure sea water during a first stage of alteration. Whole rock $\delta^{18}O$-values of rodingites and serpentinites of the Lanzo massif (2.7–3.6‰) are considerably lower than original magmatic compositions (5.1–5.4‰), whereas the Bracco samples vary over a wide range (3.8–7.1‰). The temperatures and $\delta^{18}O^{f}_{H_2O}$ calculated give for the Lanzo massif 420 °C and 3.8–5.2‰, for the Bracco massif 250 °C and 5.9–8.6‰. At the latter locality the $\delta^{18}O^{f}_{H_2O}$ is highest in the serpentinites and lowest in the diffusion zones. These data indicate that both regions equilibrated in a second stage with water enriched in ^{18}O in comparison to oceanic water, probably at about 250 °C in a deeper level of the oceanic crust. Alpine overprinting was quite different for the two ophiolitic massifs. The Lanzo rocks first recrystallized at 450 ± 50 °C and at 12–18 kbar while, during a later phase, pumpellyite formed. In the Bracco massif pumpellyite, hydrothermal veins and the diffusion zones were formed, probably under the influence of sediment contaminated or meteoric water.

Keywords: Rodingite, isotope data, ophiolite, Alpine metamorphism, Lanzo massif, Bracco massif, Northern Italy.

1. Introduction

Rodingites were first mentioned by MARSHALL (1911) who described bright, dense masses of dikelike rocks in the serpentinites of Dun Mountain, New Zealand, near the river Roding. Today, the name "rodingite" is used for metasomatically altered rocks, rich in Ca and poor in alkalies, which are closely associated with serpentinized ultramafics (KEUSEN, 1972). The mineralogical and chemical compositions of rodingites have been described from ultramafic massifs worldwide (e.g. DAL PIAZ, 1967) and from the oceanic environment (HONNOREZ and KIRST, 1975). It has been postulated that rodingitization can take place in the oceanic environment as well as during later tectono-metamorphic processes (DAL PIAZ et al., 1980; COMPAGNONI and SANDRONE, 1978). However, up to now, there are only a few data concerning the exact environment of formation of these rocks, mainly due to the complicated chemical system without free quartz and to the thermodynamic stability of most rodingite minerals over a wide pressure and temperature range.

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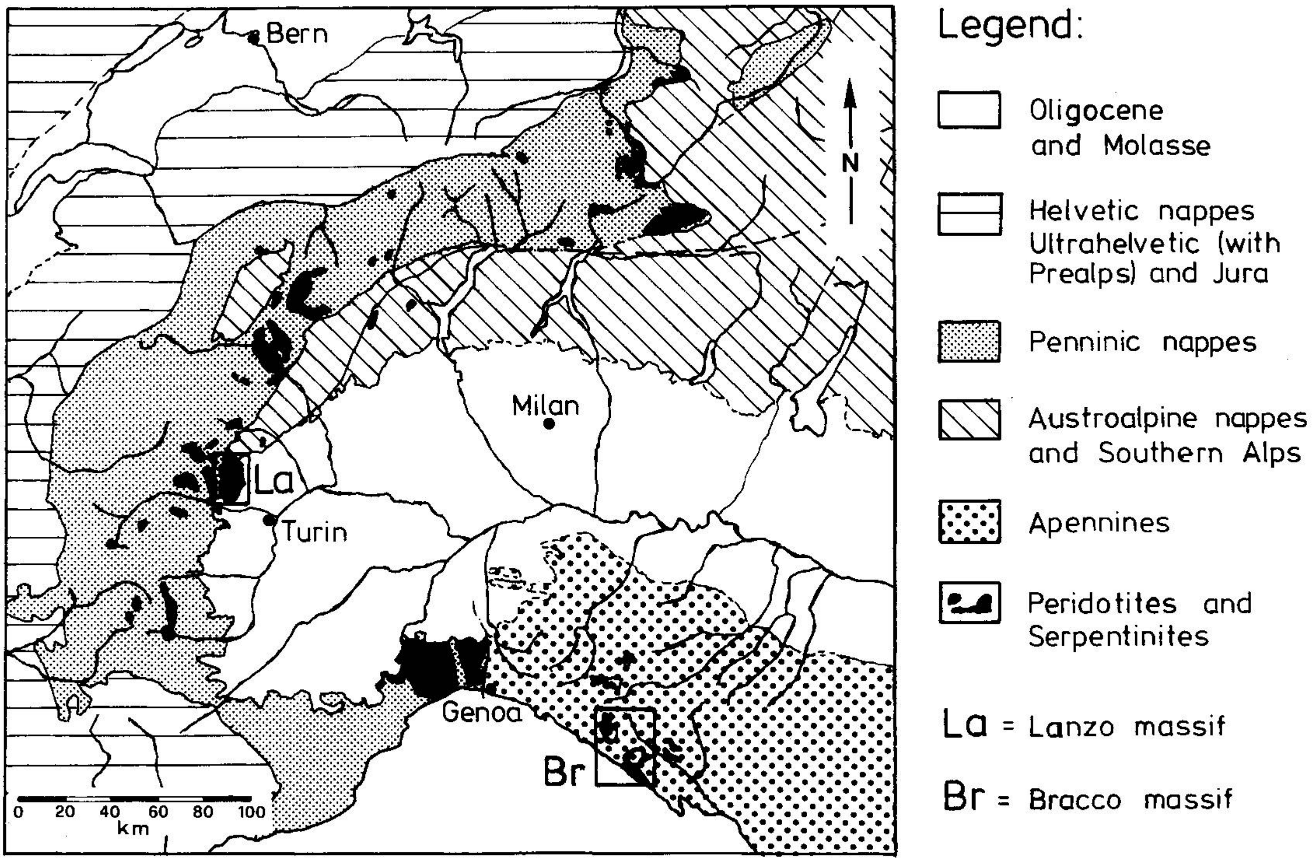

Fig. 1 The ophiolite occurrences of the Alps and the Northern Apennines with the two investigated areas in Northern Italy: La = Lanzo massif (Piedmont); Br = Bracco massif (Liguria).

There are a few stable isotope studies on rodingites from ophiolite complexes (Magaritz and Taylor, 1976; Wenner, 1979). These authors postulated multiple phases of fluid-rock interaction where antigorite and some rodingites were formed during a first phase. In a second phase the ultramafic rocks were transformed to chrysotile-lizardite serpentinites whereas the rodingites remained unchanged. However, as most of these samples represent tectonic inclusions in ultramafic rocks, the exact environment of formation and rodingitization cannot be determined. Rodingites from the Semail Ophiolite, Oman, are enriched in ^{18}O and seem to have reacted with water which had exchanged with mafic and ultramafic rocks prior to rodingitization (Gregory and Taylor, 1980).

The aim of this study was to obtain new data on the rodingitization process, to distinguish between oceanic and Alpine metasomatic processes and to understand the development of rodingites during the Alpine orogeny. For this purpose rodingites with a well-defined origin, derived from magnesio-gabbro dikes, their precursors and their associated rocks were sampled. In order to estimate the influence of the Alpine overprinting two ophiolitic massifs with a different Alpine metamorphic history were studied: the Lanzo massif (Piedmont) which sustained an advanced to complete high-pressure/low-temperature recrystallization, and the Bracco massif (Liguria) which shows only an incipient very low-grade recrystallization, both in Northern Italy (Fig. 1). Detailed mineralogical, chemical and oxygen isotope analyses were performed and Pb- and Sr-isotopes were analyzed in order to determine the influence of oceanic water.

2. The Lanzo massif

2.1. GEOLOGY

The Lanzo massif is situated about 25 km west of Turin in Northern Italy (Fig. 1). It consists mainly of spinel-plagioclase lherzolites with a small amount of "in situ"-gabbros and has been investigated in detail by Nicolas (1966) and Boudier (1978). The lherzolites seem to originate from a sub-continental lithosphere, upwelled during crustal thinning and incorporated into the early oceanic crust after the continental break up. The subsequent evolution is typically oceanic, with the

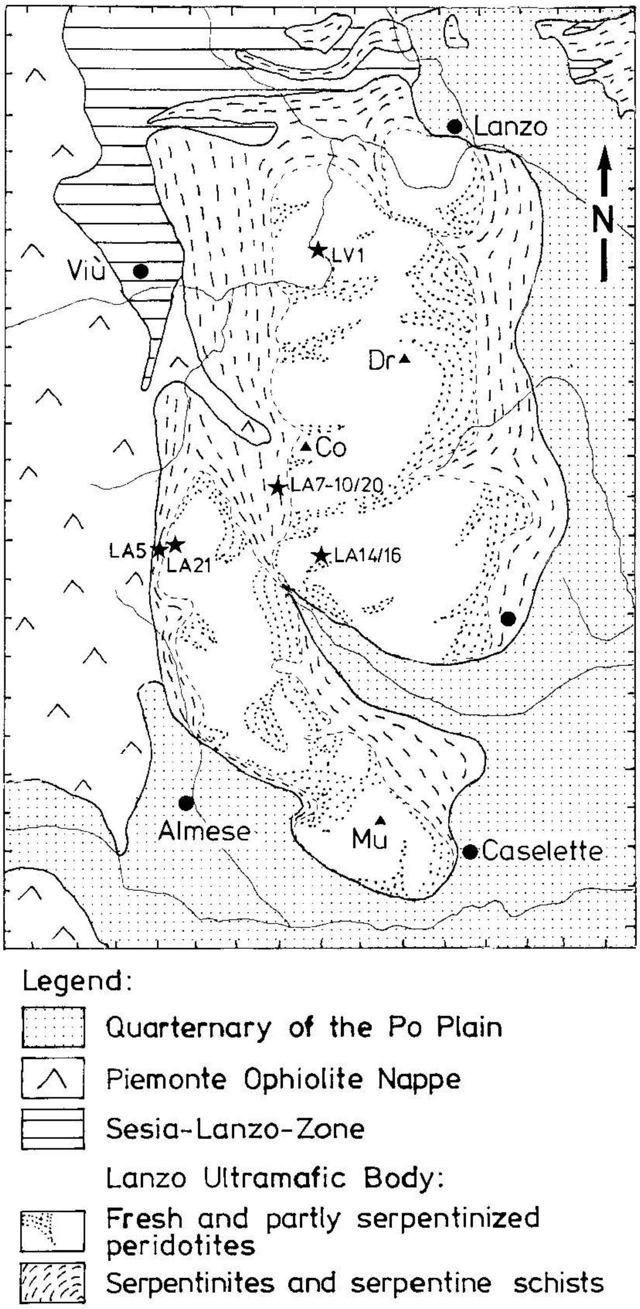

Fig. 2 Tectonic map of the Lanzo massif in Piedmont with the sample locations (compiled after Boudier, 1976; Pognante et al., 1985; Carta Geologia d'Italia, 1:100 000, F.55, Susa and F.56, Torino). Border dashes represent a distance of 1 km. Dr = M. Druina, Co = M. Colombano, Mu = M. Musiné.

intrusion of troctolitic to plagiogranitic dikes, with an early oceanic high-temperature/low-pressure metamorphism and a late injection of basaltic dikes similar to N-MORB (Pognante et al., 1985; Bodinier et al., 1986). During the closure of the Piedmont ocean, the rocks of the Lanzo massif were subducted and recrystallized under eclogite-facies conditions at P = 12–18 kbar and T = 450 ± 50 °C (Pognante and Kienast, 1987; Kienast and Pognante, 1988). This event was finally followed by a greenschist-facies metamorphism (Nicolas, 1969; Compagnoni and Sandrone, 1979). Fission track ages on zircons, which relate to a closure temperature of about 250 °C, yield ages of 90–95 Ma (Upper Cretaceous) for plagiogranites (Carpena et al., 1986).

As serpentinization is closely related to rodingitization, it was possible to sample magnesiogabbro dikes and the lherzolitic country rocks, as well as their rodingitized equivalents and the surrounding serpentinites (Fig. 2). They all originate from dikes with clear intrusive characteristics with sharp, straight contacts to the country rocks, cutting the peridotitic layering.

2.2. MINERALOGY

Magmatic clinopyroxene in the Lanzo rodingites is rarely altered, whereas plagioclase has mostly been replaced by a complex mixture of garnet, vesuvianite, diopside, zoisite and chlorite. Olivine domains consist mainly of chlorite. In two samples only garnet and chlorite are found. All minerals are well crystallized, sometimes with mosaic texture. They consist mainly of their Mg-CaAl-endmember with a small amount of bi- and trivalent iron (Rösli, 1988). Garnet is water-free, indicating a high temperature of formation (Yoder, 1950), probably a recrystallization during the eclogitic overprinting. Pumpellyite occurs in distinct zones of pronounced shearing and in veins. In one sample of an unrodingitized gabbro dike (LA10), these late pumpellyite veins crosscut the eclogitic mineral assemblage of jadeite and omphacite which indicates a second phase of interaction with Ca-rich fluids during a late Alpine stage. At the contact between the eclogitized dike and the country rock several distinct zones are developed. A zone of sheared chlorite and some diopside occurs at the contact and is followed by a zone of chlorite with idiomorphic diopside and a zone of antigorite with idiomorphic diopside which grades into the normal antigoritic country rock. The serpentinites contain several generations of antigorite together with magnetite and show a massive or, locally, a foliated texture (Fig. 2).

3. The Bracco massif

3.1. GEOLOGY

The ophiolite bodies of the Bracco region are situated about 50–60 km east of Genova (Fig. 1). The rodingites have been sampled in two dif-

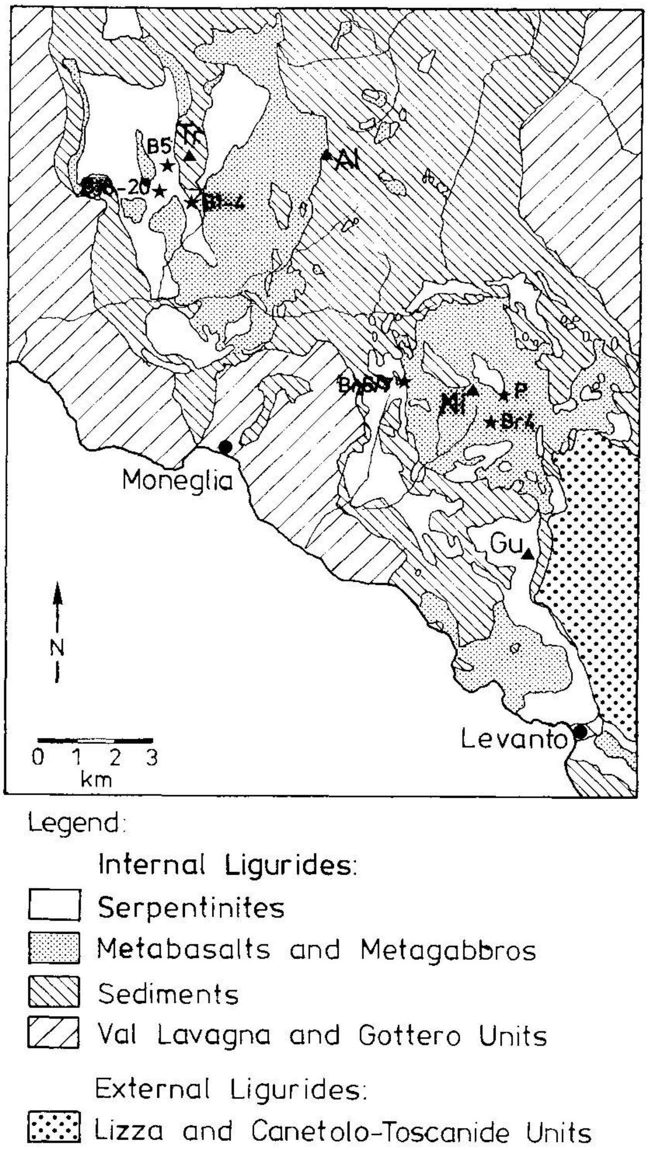

Fig. 3 Tectonic map of the Bracco massif in Eastern Liguria with sample locations (compiled after DECANDIA and ELTER, 1972; PAGANINI et al., 1972; CORTESOGNO et al., 1987). Sample numbers without the prefix Li; P = all samples LiP from M. Pascile. Tr = M. Tregin; Al = M. Alpe; Gu = M. Guaitarola.

ferent serpentinite occurrences: in the Bargone body, west of Monte Tregin; and in a small lense at Monte Pascile, east of Monte San Nicolao (Fig. 3). The spinel-plagioclase lherzolites and the serpentinites of Bargone belong tectonically to the Internal Ligurides and show, in contrast to the Lanzo massif, typical suboceanic composition, evolved in a mature oceanic ridge system (BECCALUVA et al., 1984). However, the clinopyroxenes of the ultramafics from Monte Pascile (RÖSLI, 1988) show the same characteristics as those from the Lanzo massif (POGNANTE et al., 1985) and from the External Ligurides (BECCALUVA et al., 1984), indicating a similar origin for this small ultramafic lens and therefore probably a tectonic emplacement within the Internal Ligurides.

The oceanic evolution for both serpentinite bodies is similar to that of the Lanzo massif (CORTESOGNO et al., 1987), but in addition there are at least two oceanic low-temperature metamorphic events preserved (CORTESOGNO and OLIVERI, 1974; CORTESOGNO et al., 1975; CORTESOGNO, 1980; CORTESOGNO and LUCCHETTI, 1982). The Alpine metamorphic overprint is limited to the prehnite-pumpellyite facies.

In both Bracco serpentinite occurrences rodingites derived from magnesiogabbro dikes are found. However, since the lherzolites with the exception of some small outcrops (BECCALUVA et al., 1984) are completely serpentinized, no outcrop of the unaltered magnesiogabbroic precursor was found. Therefore, three samples of unrodingitized magnesiogabbros outside of the serpentinites were also investigated.

3.2. MINERALOGY

The magmatic clinopyroxenes in the Bracco rodingites are preserved. Magmatic plagioclase reflects a lower metamorphic overprinting and is altered to garnet with a high substitution of hydroxyl in the lattice, vesuvianite and in a few samples prehnite. The olivine domains consist mainly of chlorite. All these newly formed minerals are very fine grained and often inhomogeneous in composition. In one sample (LiP2) pumpellyite

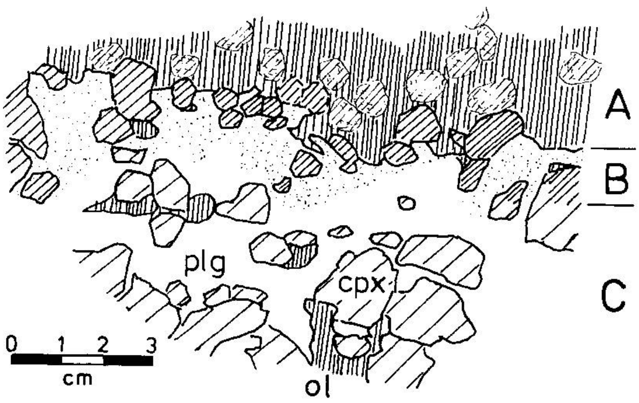

Fig. 4 Scetch of sample LiB4, a rodingitized magnesiogabbro dike from the Bracco massif. A: Blackwall consisting of chrysotile/lizardite and garnet, grown over the original serpentinite; B: Rodingite border zone, rodingitic mineral assemblage overgrown by chlorite; C: Central rodingite with magmatic mineral domaines, plagioclase (plg) is overgrown by garnet, vesuvianite and diopside, magmatic clinopyroxene (cpx) is preserved and olivine (ol) is overgrown by chlorite.

Tab. 1 Lead isotope compositions and the amount of total lead of plagioclase (plg) and clinopyroxene (cpx) domains from magnesiogabbros and rodingites. Sample description see tables 3 and 4.

Probe	Domain	C$^{(1)}$	^{206}Pb/^{204}Pb $^{(2)}$	^{207}Pb/^{204}Pb $^{(2)}$	^{208}Pb/^{204}Pb $^{(2)}$	Pb [ppm]
Lanzo massif (3):						
LA16	plg	+	17.968 ±0.005	15.482 ±0.005	37.666 ±0.021	3.10
LA14	plg	+	17.889 ±0.004	15.456 ±0.007	37.504 ±0.019	1.70
LV1	plg	+	17.862 ±0.009	15.490 ±0.003	37.471 ±0.016	0.34
LA5a	cpx	+	17.948 ±0.020	15.462 ±0.020	37.557 ±0.040	0.48
	plg	–	18.032 ±0.009	15.464 ±0.023	37.586 ±0.017	2.80
Bracco massif:						
LiBr5	cpx	–	18.175 ±0.050	15.525 ±0.050	37.740 ±0.100	0.27
	cpx $^{(4)}$	–	18.327 ±0.014	15.548 ±0.009	37.896 ±0.023	0.11
	plg	+	18.142 ±0.013	15.520 ±0.110	37.758 ±0.022	0.20
LiB18	cpx	+	18.113 ±0.030	15.495 ±0.030	37.701 ±0.090	0.04
	plg	–	18.061 ±0.040	15.503 ±0.040	37.530 ±0.100	0.05
LiB19	cpx	–	18.062 ±0.020	15.493 ±0.020	37.582 ±0.030	0.05
	cpx $^{(4)}$	+	18.236 ±0.030	15.496 ±0.028	37.870 ±0.060	0.06
	plg	+	19.055 ±0.033	15.660 ±0.024	38.624 ±0.060	0.30
LiP1	cpx	–	18.066 ±0.017	15.474 ±0.017	37.590 ±0.041	0.05
	plg	+	18.047 ±0.027	15.492 ±0.023	37.611 ±0.051	0.09
LiP4e	cpx	–	18.036 ±0.013	15.501 ±0.008	37.694 ±0.031	0.07
	cpx $^{(4)}$	–	18.249 ±0.040	15.496 ±0.030	37.751 ±0.080	0.05
	plg	–	17.920 ±0.010	15.470 ±0.010	37.507 ±0.020	0.17

reproducibility of the standards: ±0.05% for ^{206}Pb/^{204}Pb and ^{207}Pb/^{204}Pb, ±0.1% for ^{208}Pb/^{204}Pb.

(1) With (+) or without (–) fractionation correction of 1.0 ‰ per mass unit. The samples could partly only be measured at temperatures above 1500° C, and the correction was used in dependence of this behaviour.

(2) The error is 1σ of the average.

(3) The following, in Tab. 3 not described sample was measured: LA16: plagiogranite, genetically connected with the magnesiogabbros.

(4) The second value is a repetition of the measurement, on a newly prepared sample.

and prehnite occur in plagioclase domains which have preserved twinning structures and may represent late stage formation (RÖSLI, 1988). Pumpellyite and prehnite are also found in late veins.

The country rock of these rodingites is a chrysotile-lizardite serpentinite with magnetite and relic clinopyroxene. At the contact between serpentinite and rodingites mineralogically different zones occur locally. The outermost rim of the rodingite is enriched in chlorite, followed by various zones with garnet/chlorite and garnet/chrysotile-lizardite where garnet replaces magnetite and clinopyroxene (Fig. 4).

In the unrodingitized magnesiogabbros which are not in contact with serpentinite, the magmatic clinopyroxenes are transformed to tremolite to a small degree. Magmatic olivine is wholly altered to a corona consisting of chlorite or chrysotile/lizardite in the center which is surrounded by a zone of tremolite with an external zone of chlorite at the contact to plagioclase domains. In sample LiBr4, magmatic plagioclase is partially preserved with incipient alteration to sericite and albite. In the highly altered sample, LiBr7, plagioclase is entirely replaced by sericite, chlorite, albite and prehnite, so that the whole rock potassium content reaches 2.31 wt% and sodium is depleted to 0.12 wt% (RÖSLI, 1988).

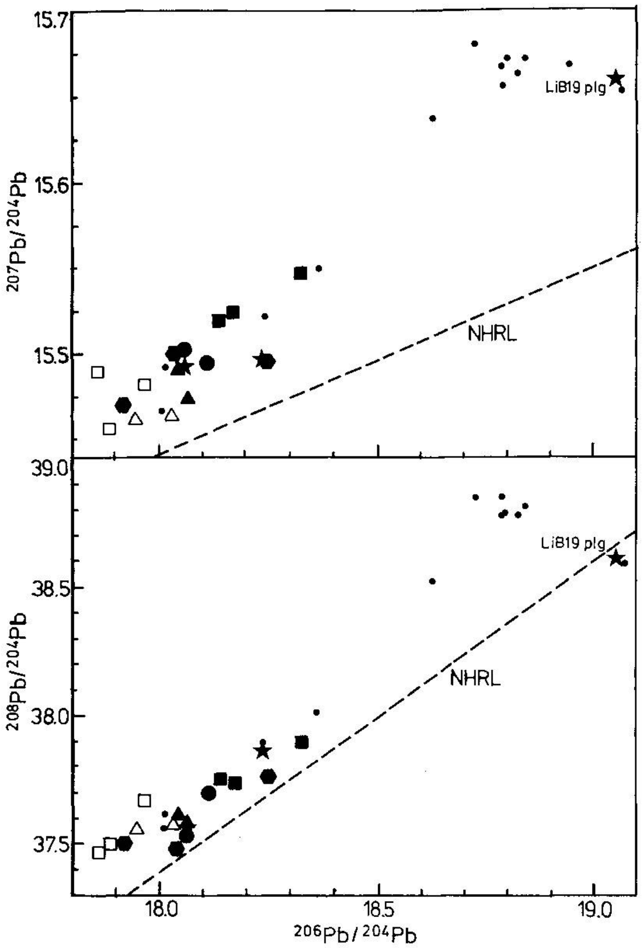

Fig. 5 ^{206}Pb/^{204}Pb – ^{207}Pb/^{204}Pb – ^{208}Pb/^{204}Pb diagram (see Tab. 1). Open symbols denote the samples from the Lanzo massif, close symbols the samples from the Bracco massif. Small dots are samples of amphibolites, sediments and PbS-ores from the Monte del Forno Unit (PERETTI and KÖPPEL, 1986). NHRL = Northern Hemisphere reference lines (HART, 1984).

4. Pb- and Sr-isotopes

Analytical procedure. The magmatic plagioclase and clinopyroxene domains were separated macroscopically, cleaned ultrasonically, crushed and handpicked. The chemical and mass-spectrometric procedures are described by CUMMING et al. (1987). About 0.5 g of sample material was used. So the blank contribution of 500 to 800 picograms of Pb was still negligible and no blank corrections were applied to the data. The isotopic ratios were measured on a Finnigan-MAT 261 mass-spectrometer.

4.1. Pb-ISOTOPES

The analytical data are listed in Tab. 1 and presented in Fig. 5. In the samples from the Lanzo massif the total lead content increase from the troctolitic gabbro LV1 (0.34 ppm) to the plagiogranite LA16 (3.1 ppm) and is related to the degree of differentiation. In contrast, the samples from the Bracco massif show very low Pb-contents (0.04–0.30 ppm) which seem to be metasomatically depleted. The high error of the analyses (Tab. 1) is due to this low lead content, whereas differences observed in repeats (LiBr5, LiB19, LiP4e) indicate sample inhomogeneities.

These data (Fig. 5) plot mostly near the Northern Hemisphere reference lines (HART, 1984) and thus are comparable with those of MOR-basalts (TATSUMOTO, 1978; HAMELIN et al., 1984a; DUPRÉ and ALLÈGRE, 1980) and also with data on rocks from other Alpine ophiolitic complexes such as gabbros from the In Zecca ophiolite in Corsica (HAMELIN et al., 1984b) and amphibolites from the Monte del Forno Unit, Switzerland/Italy (PERETTI and KÖPPEL, 1986). In view of the scatter of the data from MOR-basalts which define the Northern Hemisphere reference line and which encompasses the major part of our data field it seems reasonable to assume that the lead of the magnesiogabbro dikes from the Lanzo and from the Bracco massif represents uncontaminated MORB lead.

A trend towards an enrichment in radiogenic lead is recognizable in plagioclase of sample LiB19 and in clinopyroxene and plagioclase of LiBr5. A comparable trend was found in amphibolites, metasediments, sulphide veins and Mn-mineralizations associated with metacherts from the Monte del Forno Unit (PERETTI and KÖPPEL, 1986). This evolution was interpreted as a contamination with metamorphic fluids during the regional greenschist-facies metamorphism. In the Bracco massif the restricted occurrence of radiogenic lead indicates a local alteration phenomena, which is probably related to the Alpine prehnite-pumpellyite facies overprinting.

4.2. Sr-ISOTOPES

Four ^{87}Sr/^{86}Sr- and one ^{87}Rb/^{86}Sr-ratios (LiP1) have been determined (Tab. 2). The ^{87}Rb/^{86}Sr-composition as well as the Rb- and the Sr-content are very low in comparison to oceanic tholeiite basalts (TATSUMOTO et al., 1965). The difference between the measured and the initial ^{87}Sr/^{86}Sr-ratio calculated for a probable age of 170 m.y. (BIGAZZI et al., 1972) is quite small, and therefore the Sr-isotopes can be compared with other data.

Typical ^{87}Sr/^{86}Sr-values of basalts and gabbros from the Mid-Atlantic Ridge lie between 0.70233 and 0.70359 (TATSUMOTO et al., 1985; BONATTI et al., 1970), and the average composition of all

Tab. 2 Strontium isotope composition of the plagioclase domains of four sample from the Lanzo and the Bracco massif.

Probe Rock	LA5a rodingite	LiBr5 gabbro	LiP1 rodingite	LiP4e rodingite
minerals	zoisite	albite sericite	prehnite garnet	prehnite
$^{87}Sr/^{86}Sr$ (1)	0.70794 ±12	0.70416 ±18	0.704046 ±7	0.70359 ±16
$^{87}Rb/^{86}Sr$			0.002809	
Sr [ppm]			30.7	
Rb [ppm]			$30.6 \cdot 10^{-3}$	
$\delta^{18}O$ [‰]	zoi: 3.4	plg: 12.5 (2)	pre: 6.2	pre: 6.8
$\delta^{18}O^{f}_{H_2O}$ [‰](3)	5.3	6.7	6.2	6.8

(1) $^{87}Sr/^{86}Sr$ as average of minimal 18 blocks. The error is $\pm 2\sigma \cdot 10^{-5}$.

(2) $\delta^{18}O$ for plagioclase calculated from the whole rock $\delta^{18}O$ (9.6 ‰, Tab. 4) and the volumetric proportion of plagioclase of 60 vol%.

(3) $\delta^{18}O^{f}_{H_2O}$ calculated for LA5a at T = 450° C, for the samples from the Bracco massif at T = 250° C.

MOR-basalts is about 0.70280 (FAURE and POWELL, 1972). The $^{87}Sr/^{86}Sr$-data of the Bracco massif rodingites are slightly enriched in ^{87}Sr in comparison to MOR-data, whereas the rodingite from Lanzo (0.70794) is strongly enriched in ^{87}Sr. Its composition is quite similar to Cretaceous sea water with an average $^{87}Sr/^{86}Sr$ of 0.7076 (MCCULLOCH et al., 1981).

4.3. DISCUSSION

Similar high $^{87}Sr/^{86}Sr$-values together with typical MORB-compositions for lead isotope values were also found in oceanic basalts where sea water was responsible for the more or less intensive alteration (BARRETT, 1983). Sea water shows a Sr-content of about 8 ppm and a low lead content of 0.002–0.03 ppb (TATSUMOTO, 1978). Therefore the Pb-isotope composition of gabbroic rocks will not change much by interaction with sea water whereas the $^{87}Sr/^{86}Sr$-ratios will increase markedly. Water in equilibrium with crustal rocks such as granites ($^{87}Sr/^{86}Sr$ up to 0.74) or with sedimentary rocks, such as limestones, schists, pelites and sandstones, could also cause an enrichment in ^{87}Sr. However, the interaction with such waters should also have an effect on lead isotope compositions.

For the investigated samples the very low lead content, the magmatic values in the Pb-isotopes and the enrichment in ^{87}Sr can probably be attributed to a metasomatic exchange with ocean water.

5. Oxygen isotopes

Analytical procedure: Oxygen isotope analyses were performed on whole rock samples and on hand-picked mineral separates. Mineral separation was performed using standard techniques.

The chemical composition of the analysed minerals was determined with an ARL-SEMQ Microprobe. As magnetite was closely intergrown with fine chrysotile/lizardite it was impossible to purify it completely. For this reason, the Si-content of the magnetite separates was determined by AAS. From these data the content of serpentine minerals was calculated and the measured $\delta^{18}O$-values were corrected accordingly.

A further complication was the determination of the OH-content in garnet. It was estimated indirectly using two different methods. On the basis of microprobe-analyses the missing cations in tetraedral position were calculated. To substantiate these values the garnets were separated and the refraction indexes as well as the lattice constants (by a Jagodzinski-camera) were determined on the purified samples (RÖSLI, 1988). The OH-content, respectively Si/(Si + H)/4, was estimated graphically according to experimental data (HUCKENHOLZ and FEHR, 1982). Unfortunately, these methods gave different OH-contents, especially in the case of the andradite rich garnets. As the real OH-contents could not be evaluated, both values were used in further calculations and specified as low OH-content (from microprobe-analysis) or high OH-content (from n and a_0 determinations) (see Tab 5).

Tab. 3 Oxygen isotope analyses on whole rock samples and minerals from the Lanzo massif. Sample localities see Fig. 2

Probe Number	Description (1)	$\delta^{18}O$ Whole Rock	$\delta^{18}O$ Minerals (2)			
LA14	mg-gabbro	5.4 ±0.1				
LV1	mg-gabbro	5.4 ±0.0				
LA21	peridotite	5.3 ±0.1				
LV1d	serp. peridotite	5.0 ±0.1				
LA20	massive serp, 70 m	5.4 ±0.0				
LA10	mg-gabbro (3)	4.1 ±0.1				
LA10c	contact serp		chl	2.9 ±0.0		
LA10c1	contact serp, 1 cm	2.7 ±0.2	dio	2.9 ±0.5	chl	2.6 ±0.0
LA10c2	contact serp, 3 cm	3.5 ±0.1	dio	2.9 ±0.2	ant	3.2 ±0.1
LA10d	serp, 1 m	3.5 ±0.3				
LA9	rodingite	2.9 ±0.2	ves	1.6 ±0.2	chl	3.1 ±0.2
LA9a	contact serp	3.1 ±0.1				
LA7	rodingite	3.2 ±0.2	gar	2.3 ±0.2	chl	2.5 ±0.1
LA8	rodingite	3.6 ±0.5	gar	1.8 ±0.2	chl	2.9 ±0.1
LA5	rodingite	3.4 ±0.1	zoi	3.4 ±0.0	pum	2.5 ±0.1
LA5e	sheared serp	3.0 ±0.1				

$\delta^{18}O$-values as average ± standard deviation from minimal two analyses. The big standard deviation for some minerals is due to difficulties with the liberation of oxygen.

(1) serp=serpentinite; with distance of serpentinite samples from the next rodingite dike.

(2) mineral abreviations: gar=garnet, ves=vesuvianite, dio=diopside, zoi=zoisite, pum= pumpellyite, chl=chlorite, ant=antigorite.

(3) with cross-cutting pumpellyite veins.

For the liberation of oxygen, a modified TAYLOR and EPSTEIN-type fluorine line was used (TAYLOR and EPSTEIN, 1962). To overcome the difficulties with blanks, we purified the fluorine as described by ASPREY (1976). Mass spectrometric analyses were carried out using the SIRA-9, a triple collector 90 degree, 9-cm-radius instrument by VG Isogas.

5.1. RESULTS

The Lanzo massif: The analytical data are listed in Tab. 3. The $\delta^{18}O$ whole rock values (Fig. 6) of unaltered magnesiogabbros and peridotites from the Lanzo massif (between 5.3 and 5.4‰) are only slightly depleted in ^{18}O compared to an average mantle value of 5.7 ± 0.2‰ (MUEHLENBACHS, 1986), which could indicate some exchange with sea water at very high temperatures. Only one sample of serpentinite (LA20) shows the same composition as the unaltered peridotites, whereas the rodingites as well as the other serpentinites are depleted in ^{18}O (2.7–3.6‰) (Fig. 6). The unrodingitized magnesiogabbro dike with cross-cutting pumpellyite veins (LA10) shows an intermediate $\delta^{18}O$ value of 4.1‰. A profile from the dike through the serpentinite contact zones to the adjoining massive serpentinite (LA20) is shown in Fig. 7. In comparison with the peridotite and the

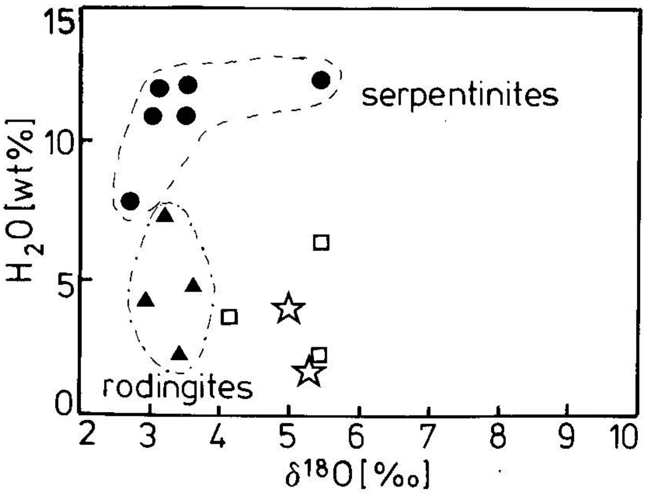

Fig. 6 $\delta^{18}O$ vs. H_2O (wt%) for the samples from the Lanzo massif. Squares = gabbros, triangles = rodingites, stars = peridotites, dots = serpentinites.

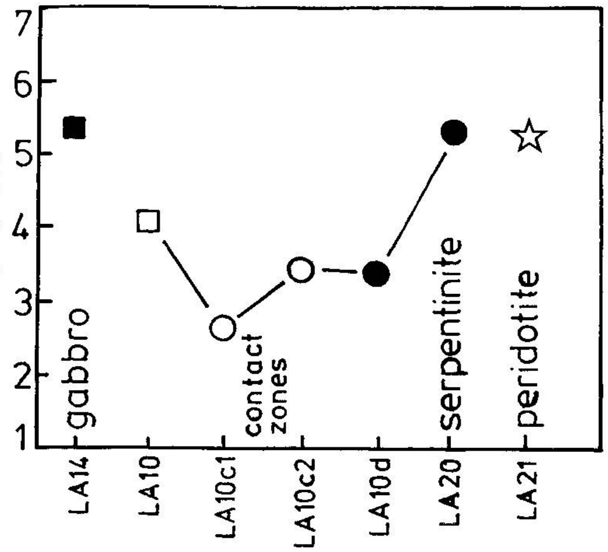

Fig. 7 $\delta^{18}O$ whole rock profile through unrodingitized gabbro dike LA10 with cross-cutting rodingitic veins, the border zones LA10c, c1 and c2 to the adjoining serpentinites LA10d and LA20. For distances from the serpentinite samples to the dike see Tab. 3. As references are given the unaltered gabbro sample LA14 and the peridotite LA21.

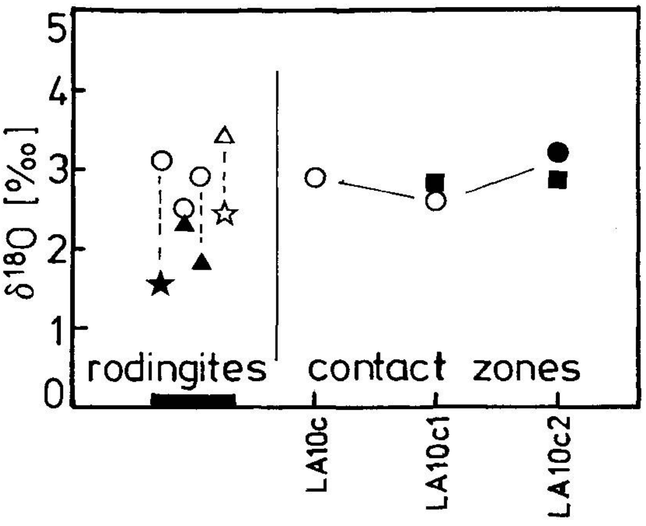

Fig. 8 The $\delta^{18}O$ mineral data in dependence of their position in the profiles from the Lanzo massif. Symbols are: filled circle = serpentine mineral, open circle = chlorite, filled square = diopside, filled triangle = garnet, filled star = vesuvianite, open triangle = zoisite, open star = pumpellyite.

massive serpentinite, the contact zones are clearly depleted in ^{18}O. The analysed minerals in the rodingites and in the contact zones vary little in oxygen isotope composition between 1.6 and 3.4‰ (Fig. 8).

The Bracco massif: The whole rock $^{18}O/^{16}O$ compositions of all samples vary between 3.8 and 9.9‰ (Tab. 4) and form several, clearly distinguishable groups (Fig. 9). The unrodingitized gabbros have high $\delta^{18}O$-values (8.1–9.9‰), which are related to the degree of alteration, as indicated by the high water content. The oxygen isotope compositions of the rodingites from Monte Pascile are somewhat lighter (6.3 to 8.6‰). These samples consist of prehnite, garnet and, in the case with the highest $\delta^{18}O$ (LiP2), of pumpellyite still showing the plagioclase twin lamellae. The rodingites from the Bargone body, with garnet and vesuvianite as principal constituents, are even more depleted in ^{18}O (3.8–6.4‰). The H_2O-content in all rodingite and gabbro samples is quite similar (2.29–6.18 wt%). The serpentinites, distinctly enriched in H_2O (8.22–13.12 wt%), show $\delta^{18}O$ compositions comparable with the rodingites from Bargone (4.2–7.1‰).

Large variations in oxygen isotope compositions between different outcrops are characteristic of the Bracco massif. Oxygen isotope profiles across the rodingite dikes and into the neighbouring serpentinites are shown in Fig. 10. The serpentinite with the highest $\delta^{18}O$ value of 7.1‰ (LiB5) is massive and can be taken as a reference sample for the primary serpentinization, scarcely affected by later shearing and alteration. The sheared serpentinites (LiB31, LiP12) and especially the contact zones are depleted in ^{18}O. Rather large differences exist between individual dikes. The entire profile LiB24 shows comparatively high $\delta^{18}O$ values (5.7–6.4‰), whereas those of LiB4 are lower (4.2–5.0‰). At the contact in profile LiB20 there is a chlorite-garnet zone enriched in ^{18}O (6.5‰), the other zones are comparable to LiB4. These features point to a local control of the fluid phases involved in the formation of the diffusion zones.

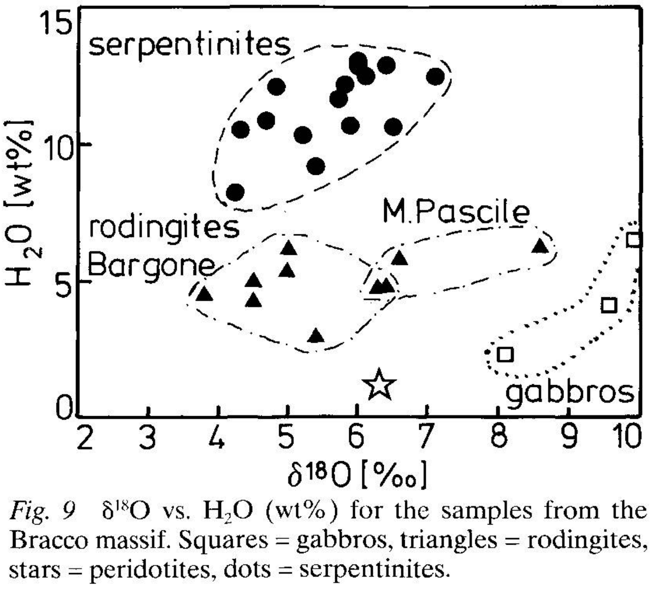

Fig. 9 $\delta^{18}O$ vs. H_2O (wt%) for the samples from the Bracco massif. Squares = gabbros, triangles = rodingites, stars = peridotites, dots = serpentinites.

Tab. 4 Oxygen isotope analyses on whole rock and minerals from the Bracco massif. Sample localities see Fig. 3.

Probe Number	Description (1)	δ^{18}O Whole Rock	δ^{18}O Minerals (2)			
LiBr4	mg-gabbro	8.1 ±0.1				
LiBr5	mg-gabbro	9.6 ±0.1				
LiBr7	mg-gabbro	9.9 ±0.3				
Li852	peridotite	6.3 ±0.0				
LiB5	massive serp	7.1 ±0.1	chr	7.6 ±0.0	mag	-1.1 ±0.2
LiB31	sheared serp	5.8 ±0.1				
LiB1	rodingite		gar$_{\text{Pl}}$	4.3 ±0.0	gar$_{\text{Ol}}$	3.0 ±0.1
LiB2	rodingite		gar	3.9 ±0.1		
LiB43	rodingite	4.5 ±0.2				
LiB42	border rodingite	5.0 ±0.1				
LiB41	blackwall	4.2 ±0.3	gar	1.7 ±0.1	chr	4.6 ±0.1
LiB4b	serp, 10 cm	4.3 ±0.2	gar	2.0 ±0.0	chr	5.0 ±0.2
LiB203	rodingite	4.5 ±0.1	ves	4.3 ±0.06		
LiB202	border rodingite	5.0 ±0.2				
LiB201	blackwall	6.5 ±0.1				
LiB20e	serp, 5 cm	4.7 ±0.1				
LiB20f	serp, 1 m	6.1 ±0.1				
LiB242	rodingite	6.4 ±0.2	ves	5.2 ±0.4	chl	6.4 ±0.3
LiB241	blackwall	5.9 ±0.3				
LiB24e	serp, 5 cm	5.7 ±0.4				
LiB24f	serp, 1.5 m	6.0 ±0.1				
LiB18	rodingite	5.4 ±0.1	gar	4.7 ±0.1		
LiB19	rodingite	3.8 ±0.1	ves	3.9 ±0.3		
LiP1	rodingite	6.3 ±0.2	pre	6.2 ±0.3	gar	5.1 ±0.3
LiP2	rodingite	8.6 ±0.2				
LiP4e	rodingite	6.6 ±0.1	pre	6.8 ±0.3		
LiP4c	sheared contact serp	4.8 ±0.1				
LiP7	contact serp	5.4 ±0.1				
LiP3	serp, 20 cm	5.2 ±0.3	gar	4.7 ±0.2	chr	6.5 ±0.1
					mag	-2.4 ±0.1
LiP9	serp, 3 m	6.4 ±0.1				
LiP12	serp, 30–40 m	6.0 ±0.6				

δ^{18}O-values as average ± standard deviation from minimal two analyses. The big standard deviation for some minerals is due to difficulties with the liberation of oxygen.

(1) serp: serpentinite, with distance of serpentinite samples from the next rodingite dike.

(2) mineral abreviations: gar=garnet, ves=vesuvianite, dio=diopside, pre=prehnite, chr =chrysotile, mag=magnetite. The subscripts pl (plagioclase) and ol (olivine) stand for the magmatic mineral domain.

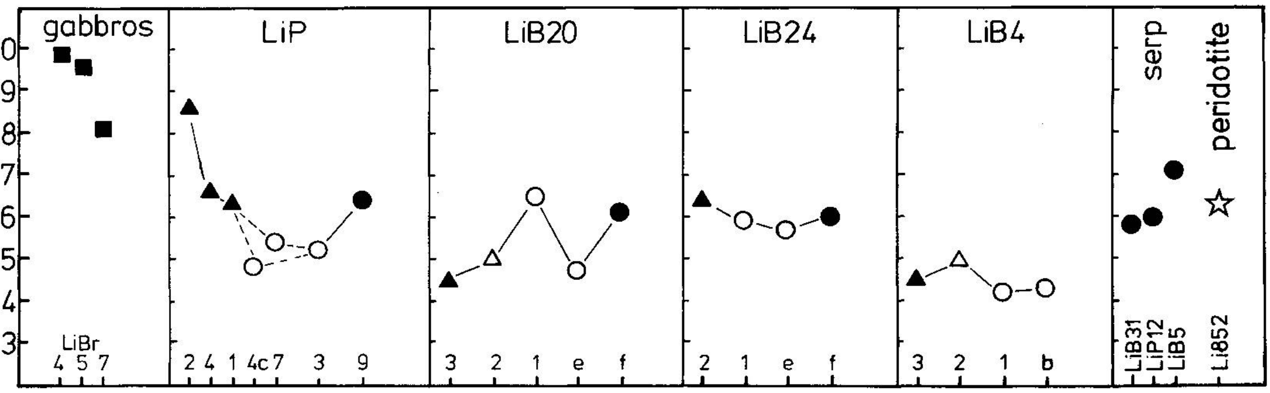

Fig. 10 $\delta^{18}O$ whole rock profile through the rodingite dikes from the Bracco massif. For distances from the serpentinite samples to the rodingite dikes see Tab. 4. As references are given the unrodingitized magnesiogabbros LiBr4, 5 and 7, the serpentinites LiB31, LiP12 and LiB5 and the peridotite Li852.

The oxygen isotope compositions of the rodingite minerals from the Bracco massif cover a wide range from 3.0 to 6.8‰ (Fig. 11). The garnet with the lowest $\delta^{18}O$-value of 3.0‰ is andradite-rich and overgrows the olivine domains, whereas a grossularite garnet associated with plagioclase domains from the same sample has a composition of 4.3‰. Obviously there was no isotopic equilibration on the scale of this specimen. The minerals in the contact zones of the serpentinites become continuously depleted in ^{18}O towards the rodingite (Fig. 11). The lightest value for garnet from a serpentinite (1.7‰) is lower than the lightest value from the rodingites.

5.2. DISCUSSION

The $^{18}O/^{16}O$ composition of a mineral depends on its oxygen isotope fractionation in relation to water. As there are no experimentally determined fractionation coefficients available for most of the minerals investigated, these coefficients were calculated as a function of their chemical compositions according to the method of SCHÜTZE (1980) as applied to silicate minerals by RICHTER and HOERNES (1988). Using this method mineral-water or mineral-mineral fractionations can be calculated for complicated solid-solutions, and even the influence of (OH)-substitution in garnet on the oxygen isotope fractionation can be estimated. The fractionation coefficients used in this study are listed in Tab. 5. For the oxygen isotope fractionation between magnetite and H_2O the experimental curve of BERTHENRATH et al. (1972) was used.

Temperatures. Because of the small fractionation (partly within the analytical measurement error) between most coexisting minerals from the Lanzo massif (Δ = 0.2–0.6) as well as from the Bracco massif (Δ = 1.1–3.1), the calculated temperatures scatter over a wide range and in most cases do not represent geologically significant conditions.

For the Lanzo massif the mineral pair zoisite-pumpellyite from sample LA5 (Δ = 0.9, Tab. 1) indicates a temperature of 420 °C. As pumpellyite was formed during greenschist-facies metamorphism, zoisite could have reequilibrated with pumpellyite during the overprinting at a temperature of about 420 °C.

For the Bracco massif the mineral pairs chrysotile-magnetite and vesuvianite-chlorite yield temperatures of 250, respectively 225 °C. Based on textural considerations, chrysotile and magnetite seem to have crystallized first during one of the

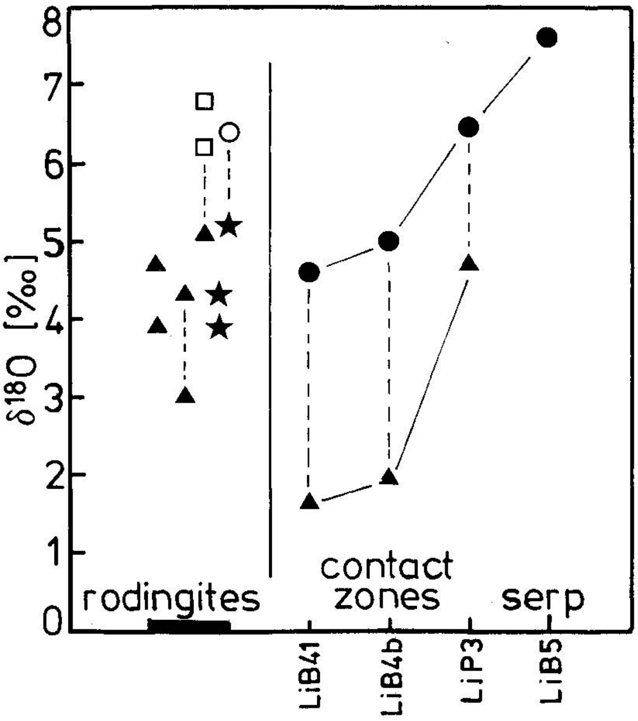

Fig. 11 The $\delta^{18}O$ mineral data in dependence of their position in the profiles from the Bracco massif. Symbols see Fig. 8, additionally: open square = prehnite.

Tab. 5 Mineral structural formula and calculated oxygen isotope fractionation coefficients for the equation $1000\ln\alpha_{(\mathrm{mineral\text{-}water})} = A + B \times 10^3 / T + C \times 10^6 / T^2$, T in [K] (RICHTER and HOERNES, 1988). Chemical data for the minerals may be requested from the authors (RÖSLI, 1988).

Probe	Min	Structural Formula	$I-{}^{18}O$	A	B	C
LA5	zoi	$Ca_2^{[VIII]}Al_3^{[VI]}Si_3O_{12}(OH)$	0.625	-2.01	-1.02	0.79
	pum	$Ca_4^{[VII]}Al_4^{[VI]}(Al_{0.6}Fe^{3+}_{0.3}Mg_{1.1})^{[VI]}Si_6O_{21}(OH)_7$	0.569	-1.83	-1.17	0.36
LA7/8	gra	$Ca_3^{[VIII]}Al_2^{[VI]}Si_3O_{12}$	0.617	-1.98	-1.04	0.73
	chl	$(Mg_{8.40}Fe^{2+}_{1.37}Al_{2.23})^{[VI]}Al^{[IV]}_{2.17}Si_{5.83}O_{20}(OH)_{16}$	0.610	-1.95	-1.06	0.67
LA9	ves$^{(1)}$	$Ca_{19}^{[VIII]}(Mg_{1.9}Al_{9.9}Fe^{3+}_{1.2})^{[VI]}Si_{18}O_{69}(OH)_9$	0.577	-1.85	-1.15	0.42
	chl	$(Mg_{9.5}Fe^{2+}_{0.6}Al_{1.9})^{[VI]}Al^{[IV]}_{1.9}Si_{6.1}O_{20}(OH)_{16}$	0.611	-1.96	-1.05	0.68
LA10c	chl	$(Mg_{9.2}Fe^{2+}_{1.0}Al_{1.8})^{[VI]}Al^{[IV]}_{1.8}Si_{6.2}O_{20}(OH)_{16}$	0.612	-1.96	-1.05	0.69
LA10c1	dio	$(Ca_{0.91}Na_{0.06})^{[VIII]}(Mg_{0.96}Fe^{2+}_{0.02}Fe^{3+}_{0.05})^{[VI]}Si_2O_6$	0.682	-2.19	-0.86	1.23
	chl	$(Mg_{9.55}Fe^{2+}_{0.80}Al_{1.65})^{[VI]}Al^{[IV]}_{1.65}Si_{6.35}O_{20}(OH)_{16}$	0.613	-1.97	-1.05	0.70
LA10c2	dio	$(Ca_{0.84}Na_{0.04})^{[VIII]}(Mg_{1.05}Fe^{2+}_{0.04}Fe^{3+}_{0.03})^{[VI]}Si_2O_6$	0.679	-2.18	-0.87	1.21
	ant$^{(2)}$	$(Mg_{44.5}Fe^{2+}_{3.1}Al_{0.4})^{[VI]}Al^{[IV]}_{0.4}Si_{33.6}O_{85}(OH)_{62}$	0.631	-2.03	-1.00	0.84
LiB5	chr	$(Mg_{11.1}Fe^{2+}_{0.6}Fe^{3+}_{0.2}Al_{0.1})^{[VI]}Al^{[IV]}_{0.3}Si_{7.7}O_{20}(OH)_{16}$	0.625	-2.01	-1.02	0.79
LiP3	gar	$Ca_3^{[VIII]}(Al_{0.06}Fe^{3+}_{1.94})^{[VI]}Si_{2.82}O_{11.28}(OH)_{0.72}$	0.609	-1.95	-1.06	0.67
		$Ca_3^{[VIII]}(Al_{0.06}Fe^{3+}_{1.94})^{[VI]}Si_{2.975}O_{11.90}(OH)_{0.10}$	0.626	-2.01	-1.01	0.80
	chr	$(Mg_{11.4}Fe^{2+}_{0.6})^{[VI]}Si_{8.0}O_{20}(OH)_{16}$	0.627	-2.01	-1.01	0.80
LiB4b	gar	$Ca_3^{[VIII]}(Al_{0.28}Fe^{3+}_{1.72})^{[VI]}Si_{2.85}O_{11.40}(OH)_{0.60}$	0.612	-1.96	-1.05	0.69
		$Ca_3^{[VIII]}(Al_{0.28}Fe^{3+}_{1.72})^{[VI]}Si_{2.63}O_{11.52}(OH)_{1.48}$	0.580	-1.86	-1.14	0.44
	chr	$(Mg_{11.5}Fe^{2+}_{0.3}Al_{0.2})^{[VI]}Al^{[IV]}_{0.2}Si_{7.8}O_{20}(OH)_{16}$	0.625	-2.01	-1.02	0.79
LiB41	gar	$Ca_3^{[VIII]}(Al_{0.48}Fe^{3+}_{1.52})^{[VI]}Si_{2.90}O_{11.60}(OH)_{0.40}$	0.617	-1.98	-1.04	0.73
		$Ca_3^{[VIII]}(Al_{0.48}Fe^{3+}_{1.52})^{[VI]}Si_{2.67}O_{10.68}(OH)_{1.32}$	0.584	-1.87	-1.13	0.47
	chr	$(Mg_{10.6}Fe^{2+}_{0.4}Fe^{3+}_{0.1}Al_{0.8})^{[VI]}Al^{[IV]}_{0.9}Si_{7.1}O_{20}(OH)_{16}$	0.619	-1.99	-1.03	0.74
LiB1	gar$^{(3)}$	$Ca_3^{[VIII]}(Al_{1.84}Fe^{3+}_{0.16})^{[VI]}Si_{2.75}O_{11.00}(OH)_{1.00}$	0.583	-1.87	-1.13	0.46
	gar$^{(4)}$	$Ca_3^{[VIII]}(Al_{0.54}Fe^{3+}_{1.46})^{[VI]}Si_{2.75}O_{11.00}(OH)_{1.00}$	0.595	-1.91	-1.10	0.56
		$Ca_3^{[VIII]}(Al_{0.54}Fe^{3+}_{1.46})^{[VI]}Si_{2.52}O_{10.08}(OH)_{1.92}$	0.562	-1.80	-1.19	0.30
LiB2	gar	$Ca_3^{[VIII]}(Al_{1.96}Fe^{3+}_{0.04})^{[VI]}Si_{2.75}O_{11.00}(OH)_{1.00}$	0.581	-1.87	-1.14	0.45
		$Ca_3^{[VIII]}(Al_{1.96}Fe^{3+}_{0.04})^{[VI]}Si_{2.64}O_{10.56}(OH)_{1.44}$	0.566	-1.82	-1.18	0.33
LiB18	gar	$Ca_3^{[VIII]}(Al_{1.96}Fe^{3+}_{0.04})^{[VI]}Si_{2.80}O_{11.20}(OH)_{0.80}$	0.589	-1.89	-1.12	0.51
LiB19	ves$^{(1)}$	$Ca_{19}^{[VIII]}(Mg_{2.1}Al_{9.9}Fe^{3+}_{1.0})^{[VI]}Si_{18}O_{69}(OH)_9$	0.570	-1.83	-1.17	0.36
LiB203	ves$^{(1)}$	$Ca_{19}^{[VIII]}(Mg_{1.7}Al_{10.6}Fe^{3+}_{0.7})^{[VI]}Si_{18}O_{69}(OH)_9$	0.573	-1.84	-1.16	0.39
LiB24	ves$^{(1)}$	$Ca_{19}^{[VIII]}(Mg_{1.8}Al_{9.5}Fe^{3+}_{1.6}Cr_{0.1})^{[VI]}Si_{18}O_{69}(OH)_9$	0.574	-1.84	-1.15	0.39
	chl	$(Mg_{9.7}Fe^{2+}_{0.6}Al_{1.7})^{[VI]}Al^{[IV]}_{1.7}Si_{6.3}O_{20}(OH)_{16}$	0.612	-1.96	-1.05	0.69
LiP1	gar	$Ca_3^{[VIII]}(Al_{1.96}Fe^{3+}_{0.04})^{[VI]}Si_{2.80}O_{11.20}(OH)_{0.80}$	0.589	-1.89	-1.12	0.51
		$Ca_3^{[VIII]}(Al_{1.96}Fe^{3+}_{0.04})^{[VI]}Si_{2.67}O_{10.68}(OH)_{1.32}$	0.570	-1.83	-1.17	0.36
LiP1/4e	pre$^{(5)}$	$Ca_2^{[VII]}Al^{[VI]}Al^{[IV]}Si_3O_{10}(OH)_2$	0.659	-2.12	-0.92	1.05

Mineral abbreviations see Tab. 3 and 4. From most garnets two formula have been calculated, with a low and a high OH-content (see text).

(1) Structure of vesuvianite after HOISCH (1985) and VALLEY et al. (1985).

(2) Structure of antigorite after KUNZE (1956) and MELLINI et al. (1987).

(3) Garnet grown over plagioclase domains.

(4) Garnet grown over olivine domains.

(5) Structure of prehnite after PREISINGER (1965).

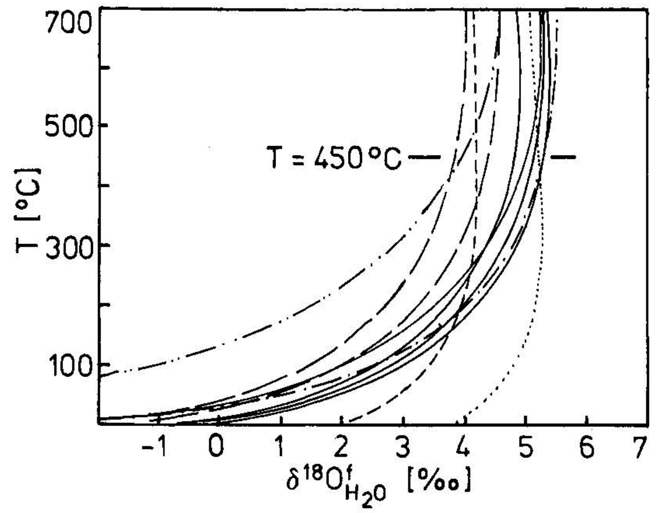

Fig. 12 The calculated $\delta^{18}O^{f}_{H_2O}$ for the minerals of the Lanzo massif in dependence of the equilibrium temperature. Lined = chlorite and antigorite, dash-dot-dotted = diopside, long dashed = garnet, short dashed = vesuvianite, dash-dotted = zoisite, dotted = pumpellyite.

oceanic metamorphic cycles. During a later, probably Alpine stage, garnet partially replaced magnetite, therefore the 250 °C could represent the temperature of the oceanic serpentinization and rodingitization.

Water composition. To determine the oxygen isotope composition of the water coexisting with the minerals, the equilibrium $\delta^{18}O$-values of water were calculated for various temperatures (Figs 12 and 13), using the measured $\delta^{18}O$ values of the mineral phases and the mineral-water fractionation factors of Tab. 5.

For the Lanzo rodingites and contact zones the curves in Fig. 12 meet at temperatures above 420 °C. The $\delta^{18}O$ of equilibrium water at this temperature varies between 3.8 and 5.2‰ (Rösli, 1989). In comparison to oceanic water ($\delta^{18}O < 0.0$‰) the water composition is enriched in ^{18}O but is lighter than magmatic water (5.5 to 9.5‰, Sheppard et al., 1969). This suggests that the fluids involved in rodingitization could have been dominated by oceanic water that had exchanged with magmatic rocks.

A further hint to the origin of the water involved in the metasomatic processes in the Lanzo massif is given by combined $\delta^{18}O$ and δD data for six serpentine minerals from the Lanzo massif (Javoy, 1980) (Fig. 14). The $\delta^{18}O$-value of 5.1‰ for one antigorite is comparable to the massive serpentinite sample LA20, whereas two other antigorite samples show values of 3.74 and 3.4‰, similar to the other serpentinites from the Lanzo massif. The δD values are all similar (–33.8 to –48.3‰). The antigorite data plot in the field for oceanic serpentinites formed by interaction with sea water whereas the lizardite data plot in the field for chrysotile-lizardite serpentinites from continental ophiolites which probably have exchanged with meteoric water at low temperatures (Sheppard, 1980).

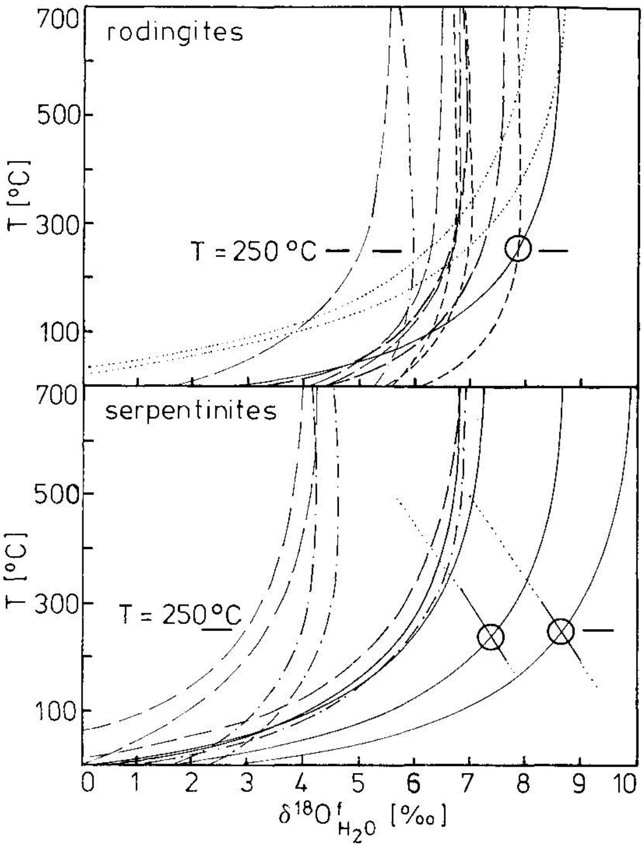

Fig. 13 The calculated $\delta^{18}O^{f}_{H_2O}$ for the minerals from rodingites and serpentinites of the Bracco massif in dependence of the equilibrium temperature. Lined: chlorite and chrysotile/lizardite; dash dotted: magnetite; long dashed: garnet; short dashed: vesuvianite; dotted: prehnite.

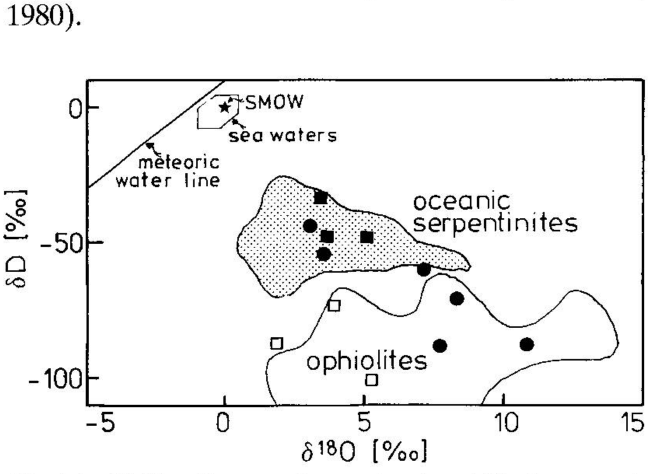

Fig. 14 $\delta^{18}O^{f}_{H_2O}$ diagram for antigorites (filled squares) and lizardites (open squares) from the Lanzo massif (Javoy, 1980) and chrysotile lizardites (filled dots) from the Bracco massif (Barrett and Friedrichsen, 1989). Fields for oceanic and ophiolitic serpentinites after Sheppard (1980).

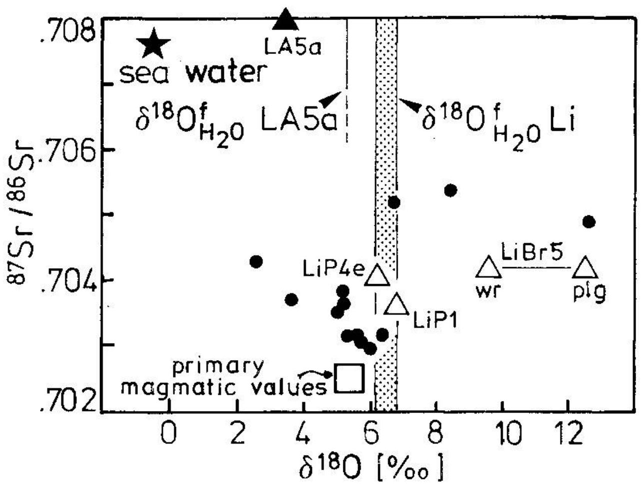

Fig. 15 $\delta^{18}O^f_{H_2O}$ $^{87}Sr/^{86}Sr$ diagram. For sample LiBr5 the whole rock (wr) as well as the calculated plagioclase (plg) composition is given. Medium Cretaceous sea water with a $\delta^{18}O$ of 0.5 ‰ (Gregory and Taylor, 1981) and $^{87}Sr/^{86}Sr$ of 0.7076 (McCulloch et al., 1981). Dots are data of gabbros, basalts and sheeted dikes from the Semail Opholite, Oman (McCulloch et al., 1981) $\delta^{18}O^f_{H_2O}$ for LA5a calculated at 250 °C (Tab. 2).

As pointed out previously the fluid composition in the Bracco massif during Alpine metamorphism is controlled by mineral reactions. Accordingly, the diagrams $\delta^{18}O^f_{H_2O}$ vs. T for the minerals from the Bracco massif exhibit rather complicated patterns (Fig. 13). At 250 °C, the temperature determined from the chrysotile-magnetite pairs, the calculated $\delta^{18}O^f_{H_2O}$ for the rodingite minerals ranges from 5.0 to 7.9‰ (Fig. 13). The lowest values (5.0, 5.9‰) pertain to garnets from an olivine domain of sample LiB1. These are clearly not in equilibrium with the other minerals from the same sample. The lowest values are similar to the equilibrium water calculated for the garnet from the serpentinitic contact zones (2.9 to 6.4‰) (Fig. 13) and depend on the OH-content and on the distance from the rodingite. Garnet 20 cm away from a rodingite has equilibrated with water of an isotope composition similar to that of water calculated from the rodingite data. The same, but less evolved trend is also visible for chrysotile/lizardite, the $\delta^{18}O^f_{H_2O}$-values of which vary between 6.4 and 8.6‰. The highest value was observed in a massive serpentinite (LiB5), the lowest one in a serpentinitic contact zone. Water/rock-calculations, for an open and for a closed system (Taylor, 1977) show that the chrysotile sample LiB5 (and also the chlorite sample LiB24) equilibrated with a water at $\delta^{18}O >$ 8‰, in contrast to the contact chrysotile samples which equilibrated with $\delta^{18}O_{Fluid} <$ 7‰ (Rösli, 1988).

These patterns and the mineralogical evolution of the rocks from the Bracco massif indicate at least two phases of fluid-rock interaction. First the rocks were serpentinized and rodingitized under influence of sea water enriched in ^{18}O which could have resulted from a previous exchange with magmatic rocks or mixing with magmatic water. In a second stage, fluid flow took place at the contact between serpentinites and rodingites. The contact zones were partially sheared and chlorite was formed. Garnet replaced magnetite and clinopyroxene. The $\delta^{18}O$-composition of the fluid was ≤ 2.5‰.

A further indication of the water composition in the Bracco massif is obtained from the gabbro samples (LiBr4, LiBr5). With the whole rock data (Tab. 4) and mineral modes, the oxygen isotope composition of plagioclase can be estimated: The gabbro LiBr5 consists of 60 vol.% plagioclase, 29 vol.% unaltered magmatic clinopyroxene and 11 vol.% olivine. The influence of olivine might be neglected for a rough estimate. By assuming a composition of 60 vol.% plagioclase and 40 vol.% clinopyroxene and a $\delta^{18}O$ of 5.2‰ for clinopyroxene we calculate $\delta^{18}O$-values of plagioclase of 10.0‰ in sample LiBr4 and 12.4‰ in sample LiBr5; both of which are comparable to plagioclases from other gabbroic and basaltic rocks from Liguria (Barrett and Friedrichsen, 1989). With the oxygen isotope data and the chemical composition of the plagioclases the equilibrium water $\delta^{18}O$-values can now be calculated. Plagioclase from LiBr4 has a mean X_{An} of 0.62. Plagioclase from LiBr5 is partially altered to albite. 14 measurements give a mean X_{An} of 0.26. With the fractionation curves of Richter and Hoernes (1988) the equilibrium $\delta^{18}O$ of water at 250 °C is 4.5 and 6.7‰ respectively. This water is similar to the first stage water in equilibrium with most rodingites and serpentinites, as discussed above.

Another assessment of the original composition of the water involved in the rodingitization and serpentinization processes in the Bracco massif can be obtained from the $\delta^{18}O$–δD data for chrysotile (Barrett and Friedrichsen, 1989). Two of their chrysotiles samples plot in the field for oceanic serpentinites (Fig. 14, Sheppard, 1980). The other samples are enriched in ^{18}O (with $\delta^{18}O$ from 7.0 to 10.8‰) with low δD (–60 to –88‰). These samples probably reequilibrated to varying degrees with a water containing a meteoric component (Barrett and Friedrichsen, 1989).

Alteration path. Figure 15 compares the oxygen isotope compositions with the strontium isotope data of the rocks (see Tab. 2). The primary magmatic composition of these rocks may have been altered in two phases. During a first stage, the rocks interacted with more or less pure sea water, which caused a slight ^{87}Sr enrichment in rocks of the Bracco massif and a strong ^{87}Sr enrichment in rocks of the Lanzo massif and probably a depletion in ^{18}O which is no longer visible. During the second phase, water which had equilibrated with mafic and ultramafic rocks caused an enrichment in ^{18}O. A similar but clearer trend has been found for gabbros, basalts and sheeted dikes from the Semail Ophiolite, Oman (Fig. 15) (McCulloch et al., 1981).

6. Summary and conclusions

Rodingites occur in most ophiolite massifs of the Alps and the Apennines. Often these rodingites are intensively overprinted by Alpine metamorphism so that their pre-Alpine evolution is difficult to understand. Two ophiolite complexes with different degrees of Alpine recrystallizations, the eclogitized Lanzo massif from the Western Alps and the very low-grade Bracco massif from the Apennines have been investigated. All the studied samples are derived from intrusive magnesiogabbro dikes. Lead, strontium and oxygen isotope ratios have been measured on whole rocks and minerals. These data, together with a careful mineralogical investigation, give clear indications for an oceanic and an Alpine evolution of these two massifs.

In both ophiolite complexes rodingitization is clearly associated with the serpentinization process. The magmatic lead isotope ratios and the content in total lead indicate that the investigated samples never exchanged with water in equilibrium with crustal or sedimentary rocks, but with oceanic water. The first events of hydrothermal alteration took place in the oceanic environment, predating the Alpine history. The peridotites and the gabbro dikes interacted with almost pure sea water producing chrysotile-lizardite serpentinites with δD-values in equilibrium with the oceanic water. A first rodingitization event took place. The calcium necessary for this hydrothermal alteration could have been derived directly from the sea water or from the serpentinized pyroxenes. The newly formed calcium-silicate minerals were enriched in ^{87}Sr, the zoisite from Lanzo even shows the same composition (0.70794) as Cretaceous sea water (0.7076, McCulloch et al., 1981). A second exchange took place at T of 250 °C with sea water enriched in ^{18}O up to 5.2‰ for the Lanzo massif and up to 8.6‰ for the Bracco massif (Rösli, 1989).

In the present day oceanic environment two regimes of water circulation can be distinguished. The first is the near surface area consisting of pillow lavas and the sheeted dikes complex, where oxidizing sea water circulates at low temperatures (Spooner et al., 1977 a, b). A second convection cell of water is situated under the Mid-ocean Ridge magma chamber (Alt et al., 1986). There, the circulating water has reacted with the mafic-ultramafic rocks and thus has become enriched in ^{18}O at elevated temperatures. By analogy, the first event of serpentinization and rodingitization of the Lanzo and Bracco massifs could have occurred in an environment similar to the first regime. The second event occurred under the influence of the second convection cell, perhaps at some distance from the ocean ridge. A participation of magmatic water, however, cannot be excluded. A better developed two phase evolution has been found for gabbros, basalts and sheeted dikes from the Semail Ophiolite, Oman (McCulloch et al., 1981), whereas for the rodingites of the same region only the second event is documented (Gregory and Taylor, 1981). The relationship of these hydrothermal events with the oceanic metamorphism (e.g. described by Cortesogno and Lucchetti, 1982; Compagnoni et al., 1984) is not clear.

During the Alpine orogeny the two ophiolite massifs sustained different metamorphic histories. The rocks from Lanzo massif were subducted and recrystallized under eclogite-facies conditions at 12–18 kbar and 450 ± 50 °C (Pognante and Kienast, 1987; Kienast and Pognante, 1988). The minerals recrystallized, e.g. antigorite was formed and oxygen isotope fractionation occurred at the higher temperatures. The greenschist-facies overprint is only partly developed. In shear zones and veins pumpellyite crystallized at about 420 °C and the peridotites were partly serpentinized. But as the water involved in these processes was in equilibrium with the surrounding rocks, no alteration in isotope composition occurred.

During a late Alpine phase the Bracco rocks were recrystallized under prehnite-pumpellyite-facies conditions. At the contact between rodingites and serpentinite, diffusion zones developed with the formation of chlorite in the rodingite and garnet in the adjoining serpentinite. Here, water with low $\delta^{18}O$-composition (down to 2.9‰) circulated. This late stage water could have had a meteoric component, which can be postulated from the isotope composition of some serpentinites (Barrett and Friedrichsen, 1989).

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