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Autor(en): **Dürr, Sören B. / Ring, Uwe / Frisch, Wolfgang**

Objekttyp: **Article**

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

Band (Jahr): **73 (1993)**

Heft 3

PDF erstellt am: **21.07.2024**

Persistenter Link: https://doi.org/10.5169/seals-55584

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SCHWEIZ. MINERAL. PETROGR. MITT. 73, 407–419, 1993

Geochemistry and geodynamic significance of North Penninic ophiolites from the Central Alps

by *Sören B. Dürr*[1,2], *Uwe Ring*[1,3] and *Wolfgang Frisch*[1]

Abstract

In the Central Alps, underneath (i.e. west of) the main Alpine suture zone (Arosa Zone), ophiolitic rocks occur as isolated bodies and are considered to represent remnants of the North Penninic or Valais ocean. The chemistry of metabasaltic rocks from these bodies displays characteristics of transitional-type mid-ocean ridge basalts, slightly elevated in incompatible trace elements relative to N-type MORB. The rocks probably stem from a shallow asthenospheric source by ca. 15% partial melting. Ultramafic rocks associated with the metabasalts were formed from lherzolitic protoliths in a mid-ocean-ridge type environment. The association of peridotite, MORB-type basalt, and a pelagic sedimentary sequence testifies of oceanic crust at least in parts of the North Penninic trough. The ocean floor presumably formed at a slow-spreading ridge in a transtensional basin. We propose a tight genetic connection of the mantle sources of the North and South Penninic oceans. The basalts of the North Penninic ocean started to form when the South Penninic ocean already was in an advanced stage and displayed homogenization of its mantle source. The North Penninic ocean appears to have tapped this mantle source.

Keywords: ophiolite, North Penninic basin, Central Alps, geochemistry, Alpine orogeny.

Zusammenfassung

In den Zentralalpen unter (d.h. westlich) der alpinen Hauptsutur (Arosa-Zone) treten ophiolithische Fragmente als isolierte Gesteinskörper auf. Sie werden als Überreste des nordpenninischen oder Valais-Ozeans angesehen. Geochemische Daten der Metabasalte dieser Vorkommen zeigen Charakteristika von mittelozeanischen Rücken-Basalten, jedoch leicht angereichert an inkompatiblen Spurenelementen. Die Gesteine wurden wahrscheinlich durch ca. 15% partielle Aufschmelzung einer seichten Asthenosphärenquelle gebildet. Die Basalte bildeten sich wahrscheinlich an einem Rücken mit langsamer Spreading-Geschwindigkeit. Das Fehlen von weniger entwickelten, stärker angereicherten Basalten lässt vermuten, dass die Mantelquelle bereits vor der Bildung des nordpenninischen Ozeans verarmt war. Meta-Ultramafite, die mit den Metabasalten assoziiert vorkommen, bildeten sich aus lherzolithischen Protolithen in einem mittelozeanischen Rücken-Milieu. Die geochemischen Daten, das Auftreten der Meta-Ultramafite sowie auch von pelagischen Sedimenten belegen, dass ozeanische Kruste zumindest in Teilen des nordpenninischen Beckens auftrat. Die Basalte des Nordpenninischen Ozeans bildeten sich, als der südpenninische Ozean sich in einem fortgeschrittenen Entwicklungsstadium befand und eine Homogenisierung seiner Mantelquelle eingetreten war. Der nordpenninische Ozean scheint diese Mantelquelle angezapft zu haben.

Introduction

The Alpine orogeny may be described in terms of progradation of tectonic activity and suturing from internal to external parts of the orogen (Roeder and Bögel, 1978; Frisch, 1979). For the Eastern and Central Alps, two depositional troughs, the South and North Penninic oceans, opened and closed successively, thus showing a northwest- to northward trend in extensional and then compressional activity (Trümpy, 1973; Clar, 1973). In the Late Triassic to Early Jurassic a rift

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[1] Institut für Geologie, Universität Tübingen, Sigwartstrasse 10, D-72076 Tübingen, Germany.
[2] present address: Mineralogisches Institut, Universität Würzburg, Am Hubland, D-97074 Würzburg, Germany.
[3] present address: Institut für Geowissenschaften, Universität Mainz, Saarstrasse 21, D-55122 Mainz, Germany.

developed between the European continent and the Apulian plate (Channell and Horvath, 1976; Frisch, 1979; Dercourt et al., 1986). This rift evolved to a narrow oceanic basin, the South Penninic or Ligurian-Piemontais ocean, in the Middle Jurassic to Early Cretaceous. Ophiolitic remnants of this ocean floor are found in the Western and Central Alps (e.g., Bearth, 1967; Dietrich, 1969) and the eastern Alpine windows (e.g., Bickle and Pearce, 1975; Heugel, 1982). They comprise ultramafic rocks, metagabbro and metabasalt, and sedimentary sequences with radiolarian chert, shale, and pelagic limestone. High terrigenous input (Bündnerschiefer or schistes lustrés), starting in the Early Jurassic, points to a limited basin width (Frisch, 1980). The onset of subduction in the South Penninic ocean was in the Early Cretaceous (Frisch, 1979; Winkler and Bernoulli, 1986). Closure of the South Penninic ocean started in the Eastern Alps with the underplating of the Tauern terrane beneath the Apulian plate in the Late Cretaceous and prograded to the Western Alps in the Early Tertiary.

The North Penninic or Valais basin took up Bündner Schiefer from the Middle Jurassic on (Pantiç and Isler, 1978) but ocean floor was probably not generated before the Early Cretaceous (Frisch, 1979). The basin was filled with a thick sequence of hemipelagic calcareous sediments and flysch (e.g., Ziegler, 1956; Isler and Pantiç, 1980; Probst, 1980; Funk et al., 1987). The North Penninic basin was completely closed in the Eocene by collision with the Helvetic margin. As a consequence, depositional activity and tectonic contraction shifted northward into the molasse foredeep on the European margin.

The existence of oceanic crust has been widely accepted for the South Penninic ocean (e.g., Laubscher, 1969; Dewey and Bird, 1970; Oberhänsli and Dietrich, 1975; Frisch, 1974, 1976; Weissert and Bernoulli, 1985; Colombi and Pfeifer, 1986; Pfeifer et al., 1989), but is still a matter of debate for the North Penninic basin (e.g., Dietrich, 1976; Frisch, 1978; Trümpy, 1980; Schmid et al., 1990), for which very little geochemical data have been published yet (e.g, Dietrich and Oberhänsli, 1975; Schmutz, 1976; Colombi, 1989). On the basis of a new geochemical study, we argue in favour of existence of oceanic crust in the North Penninic basin.

Geological Framework

Paleogeographic reconstructions (e.g., Frisch, 1979; Trümpy, 1980; Weissert and Bernoulli, 1985; Schmid et al., 1990) generally show two oceanic realms (South and North Penninic basins,

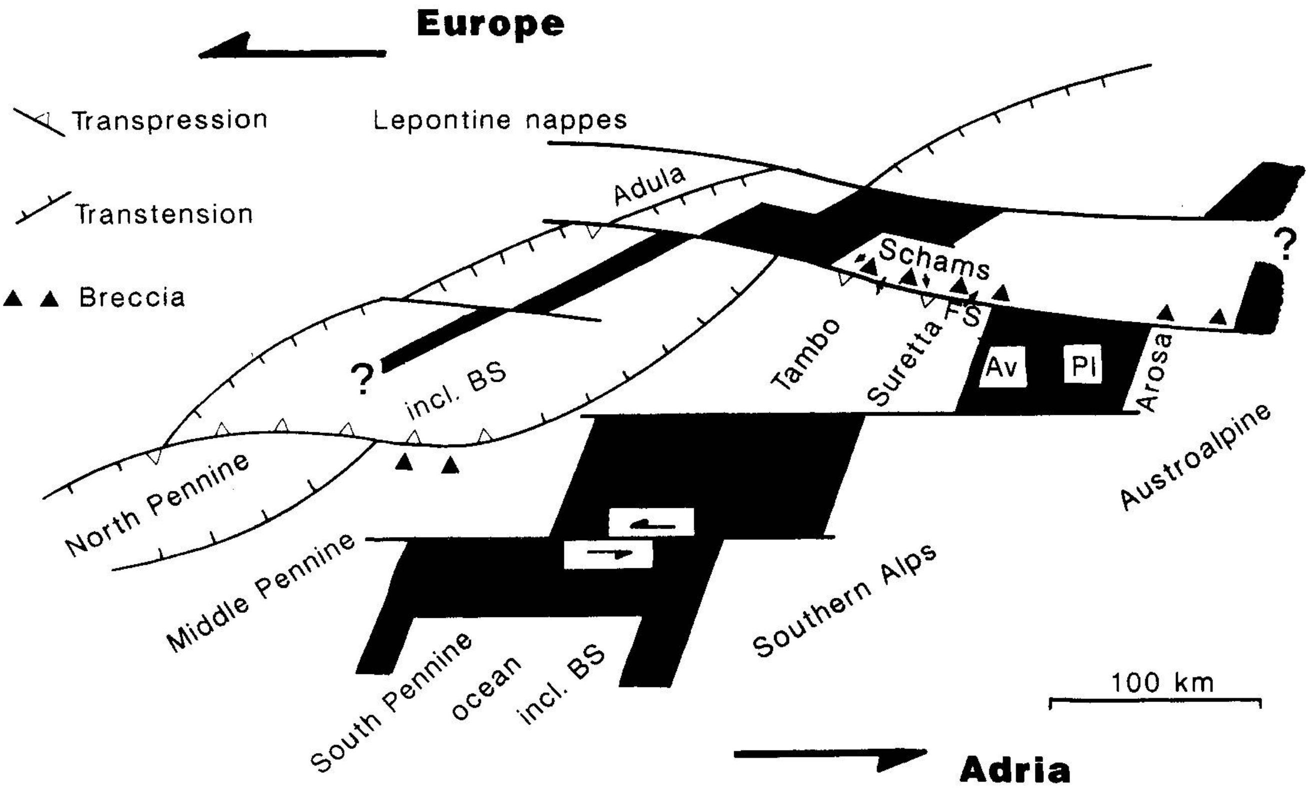

Fig. 1 Paleogeographic reconstruction of the Central Alps for the Jurassic (modified after Schmid et al., 1990). BS: Bündnerschiefer. FS: Falknis-Sulzfluh zone. Av: Avers Bündnerschiefer. Pl: Platta nappe. Black: oceanic crust.

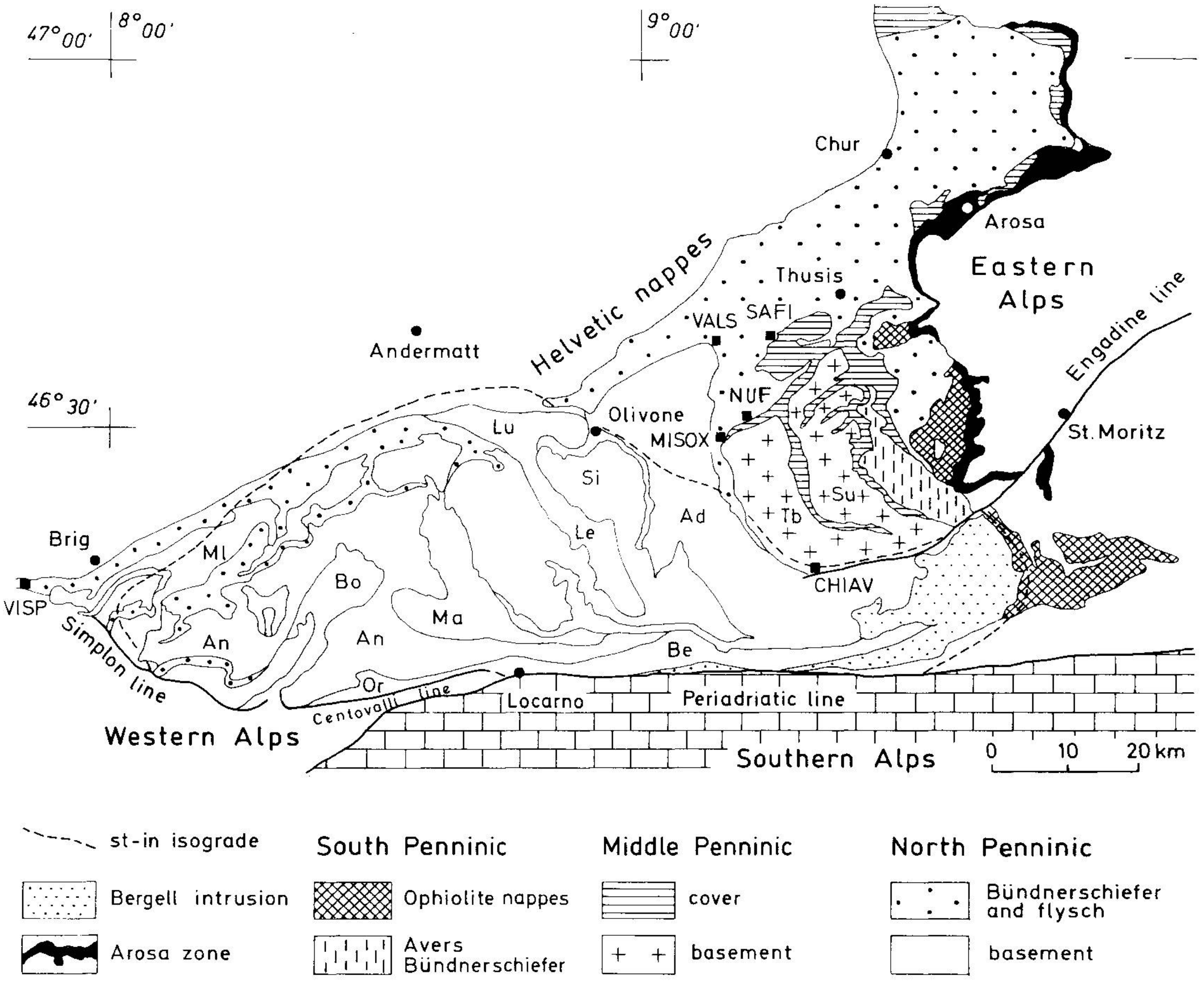

Fig. 2 (a) Tectonic map of the Central Alps. Locations used in text and sample localities. Staurolite-in isograd (dashed line; FREY et al., 1980) outlines Lepontine dome. Black squares: sampling localities. Abbreviations: Pl: Platta, MF: Malenco-Forno, Su: Suretta, Tb: Tambo, Ad: Adula (including the "Cima Lunga Lappen"), CHIAV: Chiavenna, SAFI: Safiental, Si: Simano, Le: Leventina, Lu: Lucomango, Be: Bellinzona, Ma: Maggia, An: Antigorio, ML: Monte Leone, NUF: Nufenen, Or: Orselina.

respectively) separated by a Middle Penninic rise (Fig. 1).

The North Penninic crystalline nappes represent distal European basement. The North Penninic Bündnerschiefer are sandwiched between the basement nappes and are composed of large masses of monotonous calcareous schists which are sandy at the base. Hemipelagic sedimentation grades into flysch deposits from the Mid-Cretaceous (ISLER and PANTIÇ, 1980) to the Eocene (ZIEGLER, 1956). The paleogeographic positions of the Chiavenna ophiolite (e.g., SCHMUTZ, 1976) and some other scattered basic and ultrabasic intercalations in the highly disrupted Bündnerschiefer succession are not accurately known. In general, however, the ophiolitic intercalations are thought to be restricted to the higher portions of the North Penninic Bündnerschiefer, i.e., to the vicinity of the Middle Penninic rise (PROBST, 1980). The tectonic contact of the overlying Middle Penninic nappes and the North Penninic Bündnerschiefer is decorated by ophiolitic melange units, the Areua-Bruschghorn (e.g., SCHREURS, 1990) and the Martegnas zone (EIERMANN, 1988). The Middle Penninic basement is made up of the Tambo and Suretta nappes which are overlain by a complex nappe structure of Mesozoic shelf sediments (Schams, Falknis and Sulzfluh nappes). The Avers or South Penninic Bündnerschiefer (including the Lizun ophiolites), as well as the ophiolitic Forno, Malenco and Platta nappes are remnants of the South Penninic ocean and are separated from the overlying Austroalpine units by the Arosa zone, which forms a tectonic melange (e.g., WEISSERT and BERNOULLI, 1985; RING et al., 1990).

The Adula, Tambo and Suretta nappes as well as parts of the North and South Penninic Bündnerschiefer suffered Cretaceous and/or Lower Tertiary high pressure metamorphism (e.g., HEINRICH, 1986; RING, 1992). In Mid-Tertiary times, the entire edifice of the Central Alps was affected by a Barrovian-type regional metamorphism. It reached greenschist-facies conditions in the studied region (FREY et al., 1980).

Metabasalts

30 metabasalt samples were collected from different North Penninic ophiolite occurrences. Figure 2 shows the sampling localities, table 1 presents the geochemical analyses by X-ray fluorescence as well as the geographic coordinates of the sampling locations. Despite differing Alpine metamorphism and deformation of the sampled occurrences, the geochemical patterns of the metabasalts show far-reaching similarities. Therefore these rocks are treated as a group. The sampling localities are distinguished by different symbols in the diagrams (Figs 3 to 10).

PETROGRAPHY AND MINERALOGY

The primary textures of the metabasalts are nearly completely obliterated in favour of a pronounced cleavage, recrystallization, blastesis, and mineral reaction. However, deformed pillows are discerned in places (e.g., DIETRICH and OBERHÄNSLI, 1975; EIERMANN, 1988). Most of the samples display mineral parageneses typical of low-grade metamorphism: albite, in places with oligoclase rims, actinolite to actinolitic hornblende, clinozoisite-epidote, chlorite, sphene and opaque minerals. Albite is dynamically recrystallized or shows xenoblastic to poikiloblastic growth with inclusions of amphibole needles. Sporadic corroded shape relics of magmatic plagioclase occur. Few samples contain accessory white mica which we interpret as a product of K-metasomatism from the surrounding metasedimentary rocks (see below). Lack of quartz speaks against sedimentary mixing with pelitic-psammitic material. Strongly deformed rocks are rich in carbonate and chlorite but poor in amphibole. Growth of carbonate and chlorite at the expense of amphibole in strongly sheared rock was also observed in a low-grade amphibolite of the Tauern window (FRISCH, 1984).

The studied metabasalts from the Chiavenna ophiolite are metamorphosed in the uppermost low grade. They are chraracterized by large, poikilitic green amphibole (magnesio-hornblende or tschermakitic hornblende), xenoblastic plagioclase with albite cores and oligoclase rims, clinozoisite-epidote, accessory sphene and opaque minerals (detailed petrography, see SCHMUTZ, 1976).

GEOCHEMISTRY

To get a sound base for comparison with unaltered rocks, element concentrations recalculated with 1 wt% H_2O and no CO_2 were used in the data plots (Figs 3 to 10). These values are close to the average of fresh basalts (e.g. MANSON, 1967; PRINZ, 1967).

Element mobility

Intense hydration and carbonatization of the rocks are recorded by H_2O, CO_2, or loss on ignition contents of several wt% and elevated CaO contents of some samples. The carbonatized samples are accordingly poor in SiO_2. The alteration pattern is different from spilitization (Fig. 3). Therefore, metamorphism and rock deformation rather than sea floor metasomatism were responsible for geochemical alteration. Exceeding Al_2O_3 contents correlate with low CaO contents. This is interpreted in terms of destruction of the anorthite component during sea-floor metasomatism. K-metasomatism, evident from metamorphic white mica (see above), was caused by a high geochemical gradient between the metabasalts (primarily poor in K_2O) and surrounding metapelitic sediments and occurred during regional metamorphism.

To avoid misinterpretation due to alteration effects, our interpretation of the protolith geotectonic environment mainly bases on abundances and ratios of trace elements generally considered as being immobile during weathering, low-grade metamorphism and metasomatism. Immobility up to and including low-grade metamorphism has been shown for Ti, Zr, Nb, Y, Cr, and, with limitations, for P (e.g., CANN, 1970; HUMPHRIS and THOMPSON, 1978). Since these elements, except for Cr, behave incompatibly during basalt fractionation, a method of testing their immobility is to plot their concentrations against each other (CANN, 1970). A good positive correlation in our

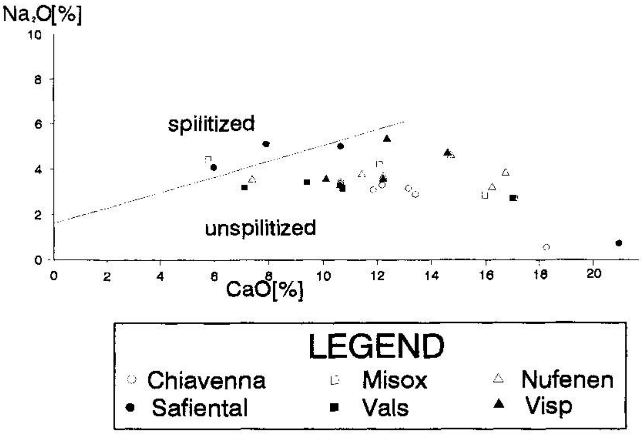

Fig. 3 Diagram after Stillman and Williams (1978) for the discrimination of spilites.

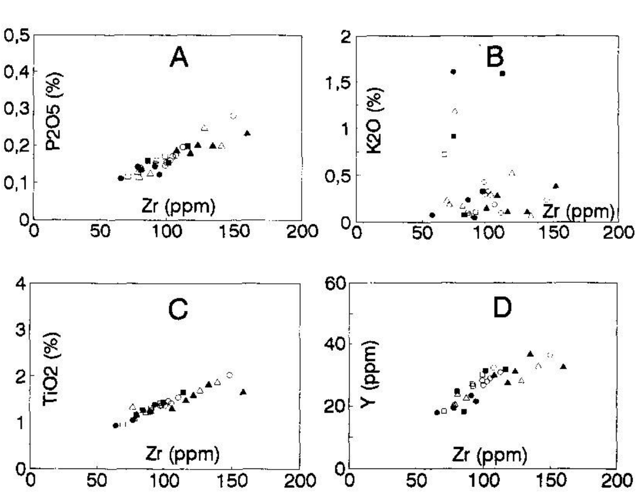

Fig. 4 Correlation diagrams to test the immobility of incompatible trace elements. Good correlation of P, Ti and Y with Zr testifies to the immobility of these elements, whereas K shows mobility. Symbols as in figure 3.

samples exists between Ti, Y, P and Zr (Fig. 4) and indicates the immobility of these elements. This applies as well for the samples in which CaO, K_2O or Al_2O_3 mobility is evident (see above). Nb has been excluded from interpretation, because the values are generally below the detection limit of 6 ppm.

Tholeiite basaltic nature of the metabasalts

The Ti–Zr systematics are able to discern between basaltic and evolved volcanic rocks (Fig. 5). Fractionation of a Ti-bearing oxide, and therefore depletion in Ti, generally correlates with, and enhances, SiO_2 increase in the magma, while Zr becomes passively enriched. In tholeiite basaltic melts, however, both Ti and Zr become increasingly enriched in early stages of magma evolution, since a low oxygen fugacity prevents early precipitation of FeTi-oxides. This trend is displayed by our samples (Fig. 5). Their subalkalic or tholeiitic character is supported by additionally taking P into consideration (Fig. 6).

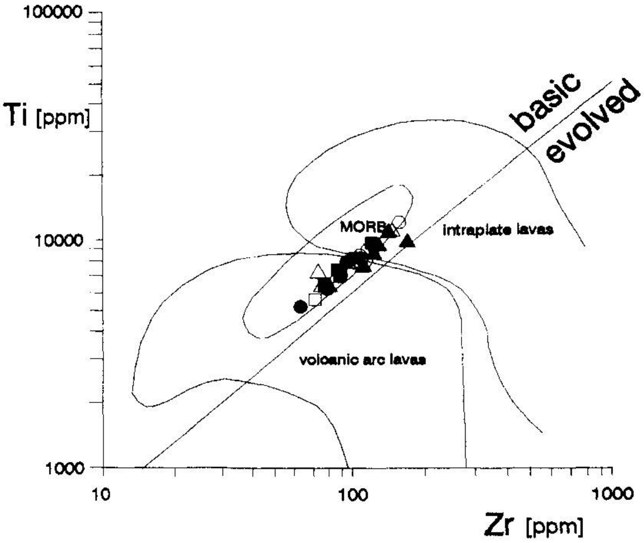

Fig. 5 Diagram after Pearce et al. (1981) showing fields of mid-ocean ridge basalts (MORB), volcanic arc and intraplate lavas. The straight line separates basic from evolved rocks. Symbols as in figure 3.

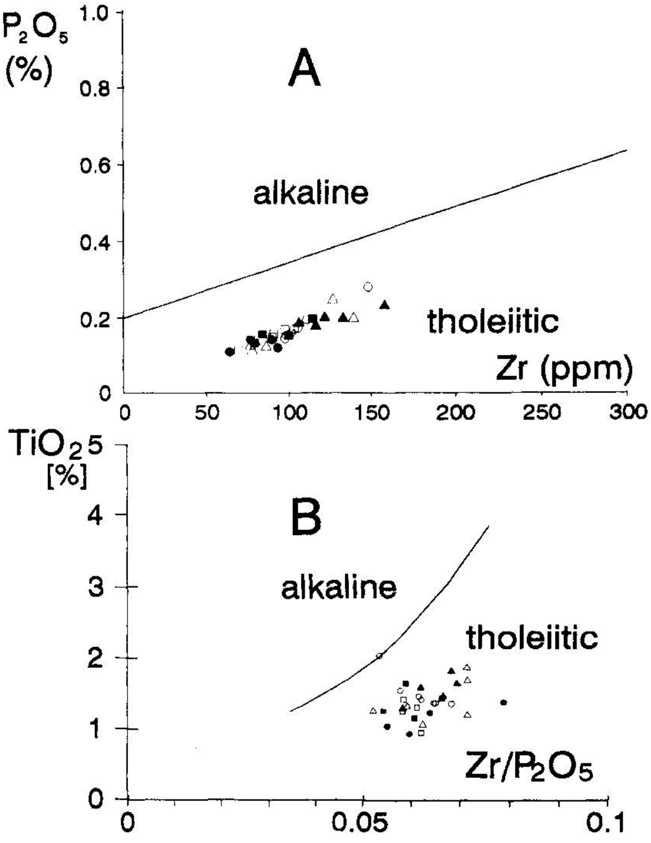

Fig. 6 Diagrams after Winchester and Floyd (1976), discriminating alkaline from subalkaline/tholeiitic basalts. Symbols as in figure 3.

Tab. 1 Chemical results and coordinates of sampling locations of basaltic rocks. det.l.: detection limit. Nb values were under the detection limit of 6 ppm, except for Misox 4 (10 ppm), Safi3, Safi4 and Visp4 (all 7 ppm). Asterisk in CO_2 column: not determined, then H_2O = loss on ignition.

sample	Swiss coordinates		SiO_2	TiO_2	Al_2O_3	$FeO^{(tot)}$	MnO	MgO	CaO	Na_2O
	longitude	latitude	%	%	%	%	%	%	%	%
Chiav01	752 000	130 900	49.62	1.33	15.49	8.92	0.14	7.84	10.47	3.33
Chiav02	750 900	132 600	49.84	1.34	15.52	9.02	0.15	7.85	10.55	3.43
Chiav03	751 100	132 300	47.54	1.39	16.96	9.0	0.15	6.39	13.29	2.86
Chiav04	753 400	132 200	48.38	1.42	15.86	9.30	0.14	6.47	11.95	3.25
Chiav05	751 700	131 700	42.68	1.97	16.04	11.18	0.12	5.77	17.84	0.53
Chiav06	752 500	132 000	48.41	1.51	15.70	10.34	0.17	5.3	13.0	3.13
Chiav11	751 300	132 500	48.23	1.33	15.79	9.32	0.15	7.87	11.74	3.06
Misox1	735 100	153 500	42.37	1.31	14.30	9.41	0.16	6.53	14.81	2.64
Misox2	735 200	152 700	49.23	1.22	13.20	8.16	0.12	5.82	11.40	3.98
Misox3	735 800	154 600	50.28	0.9	20.57	6.67	0.12	5.87	5.52	4.31
Misox4	736 000	154 500	43.44	1.15	13.44	8.6	0.16	6.24	15.8	2.51
Nuf1	738 500	155 500	45.5	1.55	14.66	8.3	0.12	3.05	13.69	4.26
Nuf2	738 400	155 600	43.2	1.11	17.52	9.26	0.13	6.35	11.46	3.45
Nuf3	737 750	156 100	46.06	1.76	15.66	10.13	0.15	6.18	10.92	3.61
Nuf4	737 700	156 050	49.03	1.01	18.81	8.08	0.12	6.44	7.11	3.42
Nuf5	737 800	156 200	40.38	1.17	12.63	8.1	0.15	7.36	14.99	3.44
Nuf6	737 800	156 150	40.79	1.02	14.34	8.64	0.16	5.42	14.43	2.83
Safi1	742 500	165 200	46.43	1.15	16.58	8.53	0.14	5.45	10.05	4.75
Safi2	742 100	165 000	43.53	0.82	14.60	8.57	0.14	2.83	19.2	0.67
Safi3	741 700	166 600	47.63	0.99	21.7	7.63	0.13	5.65	5.73	3.98
Safi4	742 800	165 900	48.5	1.31	16.4	8.86	0.13	6.86	7.56	4.94
Vals1	733 000	165 200	47.53	1.2	15.47	9.63	0.14	8.98	9.08	3.34
Vals2	733 300	163 100	47.2	1.36	15.13	9.71	0.14	7.4	10.25	3.03
Vals3	735 100	168 300	46.52	1.57	17.72	10.97	0.14	6.61	6.81	3.09
Vals4	734 400	167 400	42.68	1.06	15.78	8.72	0.13	4.37	15.87	2.54
Visp1	632 700	125 800	44.84	1.33	15.58	7.6	0.12	4.45	11.31	4.89
Visp2	634 100	125 000	45.16	1.18	15.12	7.69	0.14	3.96	13.51	4.33
Visp3	633 900	124 800	48.41	1.75	15.52	9.36	0.15	7.35	9.8	3.47
Visp4	634 000	125 100	44.08	1.47	14.84	9.49	0.16	7.59	11.43	3.33
Visp5	632 600	126 000	50.59	1.56	12.96	10.07	0.2	5.61	10.19	3.17
det.l.			1.19	0.03	0.33	0.55	0.01	0.25	0.2	0.18

Discrimination of the geotectonic environment

Concentrations and ratios of a series of minor and trace elements are characteristic features for certain geotectonic environments in which basalts formed. A number of basalt discrimination diagrams has been developed on an empirical base to enable a genetic discrimination. Since the interpretation based on these diagrams alone is often ambivalent, other geological information is necessary for a sound conclusion.

Ti, Zr, Y and Cr are present in our samples in concentrations typical for mid-ocean ridge basalt (MORB) (Figs 5, 7, 9). Multi-element patterns, normalized to average N-type MORB (Fig. 8), show enrichment, relative to MORB, of the elements with low ionic potential (Sr, K, Rb, Ba) and concentrations close to unity of the elements with intermediate ionic potential (P, Zr, Ti, Y, Cr). The first group of elements tends to be mobile during processes such as sea-floor weathering and metamorphic or metasomatic alteration (Hart, 1969; Pearce, 1975, 1976). As an example, K does not correlate with Zr as a representative of the immobile elements (Fig. 3).

The flat pattern of the elements from P to Cr in figure 8 is also typical of a tholeiitic MORB (for comparison, the bottom diagram in figure 8 depicts typical basalt patterns). The general slight negative slope between P and Y signifies Zr/Y ratios > 3 and TiO_2/P_2O_5 ratios < 12. These ratios slightly deviate from typical N-type MORB values (Beccaluva et al., 1983) and indicate a mantle source enriched in incompatible elements. Such source enrichment, relative to C_3 chondrite composition, is also evident from the ZrY systematics (Fig. 9). Elevated Zr/Y ratios are typical for enriched, transitional (T-type) MORB and within-plate basalt (Pearce and Norry, 1979). The fact that partial melting as well as subsequent fractional crystallization affects Zr

K_2O %	P_2O_5 %	Ba ppm	Cr ppm	Nb ppm	Ni ppm	Rb ppm	Sr ppm	V ppm	Y ppm	Zn ppm	Zr ppm	CO_2 %	H_2O %	Sum %
0.33	0.16	< 46	341	< 6	102	13	138	253	28	66	100	*	0.69	98.42
0.33	0.15	< 46	338	< 6	104	10	140	245	27	59	98	0.76	0.23	99.26
0.19	0.31	< 46	322	< 6	106	< 9	215	250	33	60	105	1.05	0.45	99.66
0.29	0.17	< 46	372	< 6	111	12	134	234	29	83	102	0.61	0.39	98.33
0.24	0.27	72	279	6	98	13	727	366	36	68	145	0.91	0.8	98.53
0.1	0.19	< 46	201	< 6	92	< 9	146	264	31	71	110	0.63	0.38	98.98
0.43	0.14	86	246	< 6	99	23	228	225	29	69	97	0.55	0.33	99.11
0.11	0.16	< 46	313	< 6	159	9	320	249	29	68	91	4.89	2.34	99.16
0.08	0.29	< 46	314	< 6	163	< 9	377	230	26	63	86	4.01	2.15	99.78
0.72	0.11	52	409	< 6	161	26	197	113	18	55	67	1.19	3.19	99.53
0.1	0.15	< 46	267	< 6	136	9	303	222	25	101	85	5.48	1.88	99.04
0.52	0.39	81	159	10	50	13	154	196	27	49	118	5.95	1.66	99.73
0.17	0.26	< 46	228	< 6	159	9	187	156	22	68	81	3.28	3.33	99.62
0.07	0.19	61	197	< 6	105	< 9	180	240	32	69	133	1.06	2.99	98.87
1.18	0.21	66	219	< 6	77	21	263	161	20	54	75	*	3.2	98.73
0.23	0.27	< 46	351	6	219	10	211	117	19	64	69	7.7	3.82	100.33
0.19	0.12	< 46	298	< 6	152	12	235	163	22	65	71	8.0	3.39	99.44
0.24	0.28	< 46	262	< 6	132	9	70	155	23	70	85	3.41	2.52	99.61
0.08	0.1	< 46	339	< 6	159	< 9	281	132	16	43	58	4.11	3.6	98.36
1.61	0.13	146	271	< 6	93	40	195	147	19	54	74	*	3.71	99.09
0.05	0.31	80	254	7	134	< 9	122	164	21	67	90	1.66	2.84	99.48
0.08	0.30	< 46	397	7	214	< 9	213	174	18	69	82	*	3.57	99.28
0.33	0.14	< 46	385	< 6	171	13	147	242	31	66	96	1.4	2.88	99.07
1.59	0.19	76	234	6	116	29	94	236	31	82	111	0.6	3	98.98
0.91	0.28	52	262	< 6	95	21	177	218	24	79	74	5.42	1.48	99.34
0.28	0.16	< 46	270	< 6	130	5	259	155	26	58	107	*	8.55	99.29
0.15	0.32	< 46	267	< 6	104	12	345	195	28	49	99	*	7.81	99.48
0.11	0.19	65	289	< 6	91	9	304	297	36	70	130	*	2.62	98.85
0.11	0.18	49	323	< 6	166	< 9	200	260	30	74	115	*	5.84	98.64
0.38	0.38	143	147	7	22	< 9	193	204	32	97	152	*	4.09	99.29
0.05	0.03	46	15	6	9	9	14	9	4	12	19			

concentrations in the melt but leaves Zr/Y ratios essentially unchanged, provides a key to source enrichment or depletion (Fig. 9). In the Zr/Y–Zr diagram (Fig. 9), modelled genetic magma pathways for partial melting in the order of 15% (see below) lie subparallel to the x-axis. The North Penninic samples plot in the top sector of the MORB field, thus revealing an enriched character of the mantle source.

Modelled petrogenetic magma pathways enable to infer the degree of melting necessary to produce basaltic rocks. The Cr–Y relations (Fig. 7) are particularly appropriate, because neither Cr nor Y appear to be significantly affected by mantle heterogeneities (Pearce et al., 1982). Three subsequent compositional trends are distinguished (Fig. 7). During progressive partial melting in the range between 5 and 30%, Y decreases in the melt but Cr concentrations experience little change (dotted line in Fig. 7). Subsequent crystallization of olivine, Cr-spinel and clinopyroxene reduces Cr concentrations but only insignificantly raises Y values in the melt (solid line, steep trend in figure 7; line shown for 15% partial melting). The appearance of plagioclase as a crystallizing phase results in enhanced Y enrichment (solid line, flat trend in Fig. 7). Our samples are in agreement with the differentiation trend modelled for a melt produced by 10 to 15% partial melting.

Around 15% of partial melting are indicated by the Al_2O_3/TiO_2–TiO_2 systematics introduced by Beccaluva et al. (1983; Fig. 10). In contrast to Pearce and Norry (1979) and Pearce et al. (1982), who, for their modelling, assumed trace element concentrations for a mantle source similar to C_3 chondrite, Beccaluva et al. (1983) calculated their model with a spinel lherzolite mantle source, slightly enriched in incompatible elements relative to chondrite.

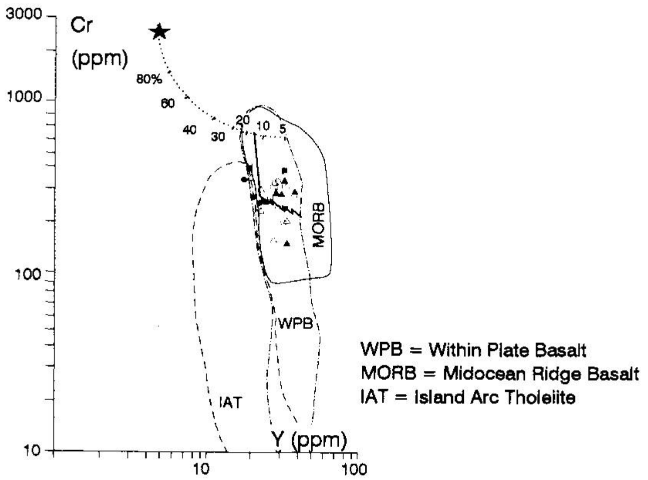

Fig. 7 Diagram after Pearce (1980) for discrimination of island arc tholleiites (IAT), within-plate basalts (WPB) and mid-ocean ridge basalts (MORB), and modelling of genetic magma pathways. Asterisk gives mantle source composition (C_3 chondrite). Dotted line shows the melting curve for equilibrium partial melting (percentages indicated by numbers). Solid line shows the closed sytem magma fractionation trend for magma formed by ca. 15% partial melting. Symbols as in figure 3.

For a number of samples the concentrations of elements with intermediate ionic potential are slightly lower as compared to N-type MORB (Fig. 8). This is indicative of basalts derived from slow spreading ridges (Pearce and Norry, 1979). In accordance, the Zr/Y–Zr diagram (Fig. 9), capable to roughly distinguish between MORB from slow (bilateral spreading rate < 2 cm/a) and faster spreading ridges, tends to reveal slow spreading rates. The distinction between ridges with different spreading rates is thought to be due to a generally larger and more stable magma chamber under fast spreading ridges, allowing mixing of the primitive, newly generated magma with magma in the chamber enriched by fractionation (Nisbet and Fowler, 1978). The resulting displacement towards higher Zr values is therefore roughly correlative with the spreading rate (Pearce, 1980). However, the derivation of spreading rates from the geochemistry of deformed and metamorphosed basalts alone is problematic, but additional arguments in favour of a slow spreading rate will be given below.

Ultramafic rocks

Ultramafic rocks are rare in North Penninic ophiolite complexes. Six samples from the Chiavenna ophiolite and from Visp were studied. Chemical analyses and coordinates of the sampling locations are given in table 2.

PETROGRAPHY

Thorough serpentinization affected all studied ultramafic rocks. The predominating serpentine mineral in the Chiavenna samples is antigorite (see also Schmutz, 1976). Olivine is frequent and appears to be a product of metamorphism since it

Tab. 2 Chemical results and sampling locations of ultramafic rocks. Asterisks as in table 1.

sample	Swiss coordinates longitude	Swiss coordinates latitude	SiO_2 %	TiO_2 %	Al_2O_3 %	$FeO^{(tot)}$ %	MnO %	MgO %	CaO %	Na_2O %	K_2O %	P_2O_5 %	Ba ppm	Cr ppm
Chiav08	752 100	132 400	40.14	0.1	2.92	9.16	0.13	37.57	0.23	0.21	0.0	0.18	< 46	2554
Chiav09	751 400	132 600	41.02	0.11	2.79	9.23	0.13	35.93	1.33	0.21	0.0	0.18	< 46	2596
Chiav10	751 700	132 100	41.54	0.09	2.97	7.96	0.13	37.91	< 0.2	0.21	0.0	0.18	< 46	2098
Chiav13	751 900	132 300	40.61	0.1	2.83	8.5	0.12	37.82	0.46	0.21	0.0	0.19	< 46	2584
Visp7	632 800	126 500	39.93	0.08	2.28	8.48	0.09	36.15	< 0.2	0.18	0.0	0.18	< 46	2504
Visp8	633 000	126 500	40.07	0.06	2.05	7.71	0.1	37.36	< 0.2	< 0.18	0.0	0.03	< 46	2435
det.l.			1.19	0.03	0.33	0.55	0.01	0.25	0.2	0.18	0.05	0.03	46	15

	Ni ppm	Rb ppm	Sr ppm	V ppm	Y ppm	Zn ppm	Zr ppm	CO_2 %	H_2O %	sum %
Chiav08	2256	< 9	< 14	72	< 4	49	< 19	*	9.01	100.14
Chiav09	2003	< 9	< 14	75	< 4	47	< 19	*	6.94	99.99
Chiav10	1890	< 9	< 14	66	< 4	51	< 19	*	8.16	99.78
Chiav13	2129	< 9	< 14	63	< 4	53	< 19	*	8.81	100.16
Visp7	2145	< 9	< 14	57	< 4	41	< 19	*	11.89	99.89
Visp8	2112	< 9	< 14	50	< 4	44	< 19	*	11.88	99.92
det.l.	9	9	14	9	4	12				

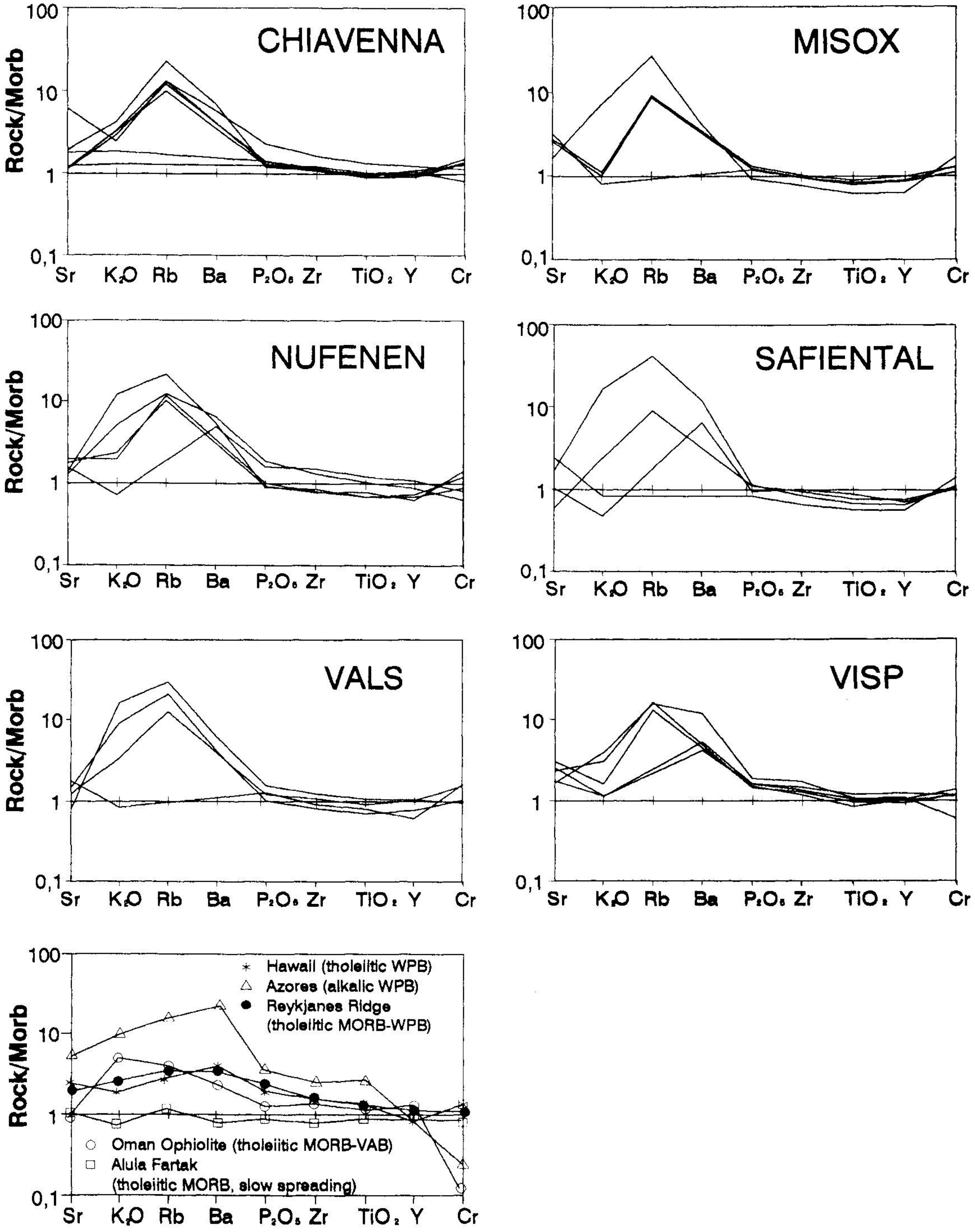

Fig. 8 Chemical patterns in MORB-normalized multi-element diagrams after PEARCE (1980). The bottom diagram shows patterns typical for specific geotectonic settings, for comparison (after PEARCE, 1980, 1982; PEARCE et al., 1981, 1984). Rb and Ba have been omitted for those samples with values under the respective detection limits.

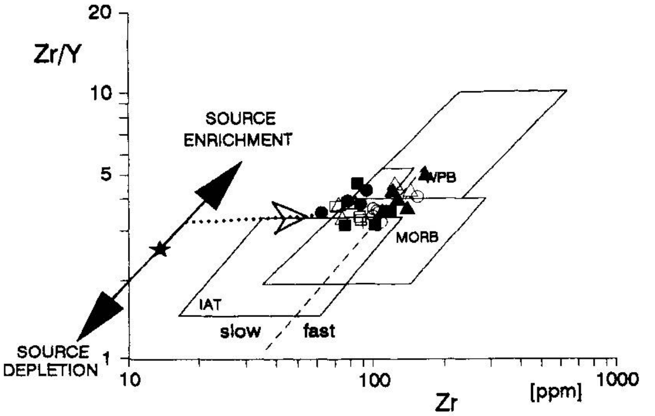

Fig. 9 Diagram after PEARCE (1980) for modelling genetic pathways of basaltic magmas. Asterisk gives mantle source composition (C_3 chondrite). Horizontal dotted line with open arrow head shows partial melting trend for a slightly enriched mantle source, with subsequent closed system fractional crystallization. Abbreviations as in figure 7. Symbols as in figure 3.

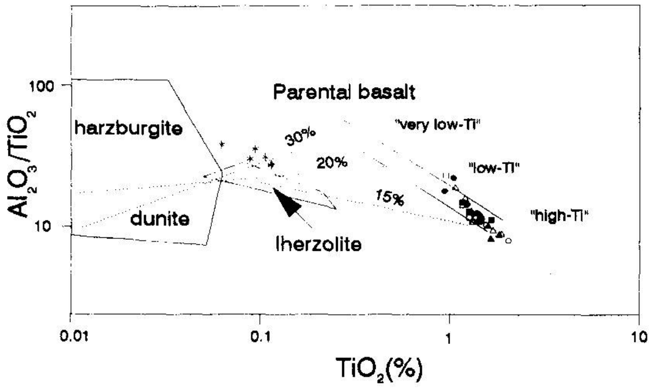

Fig. 10 Diagram after BECCALUVA et al. (1983) showing the modelled genetic relationships (dotted tie-lines) between basaltic melts and their ultramafic residua, derived from a (hypothetical) spinel lherzolite source. Percent numbers refer to degrees of equilibrium partial melting. Symbols as in figure 3; snow flakes: ultramafic rocks.

overgrows serpentine minerals. Elongate olivine grains are parallel to the cleavage. Colourless Mg-amphibole crosscuts the cleavage planes and is randomly oriented. Secondary carbonate aggregates occur in the serpentine matrix. Frequent opaque minerals are mainly magnetite and minor ilmenite (SCHMUTZ, 1976). Due to strong alteration of the primary mineral assemblages, the ultramafic rocks are classified according to their chemical composition rather than on petrographical grounds.

GEOCHEMISTRY

The chemical analyses were recalculated on a H_2O- and CO_2-free base. Normative corundum in all samples (CIPW norm) points to alteration processes causing enrichment in Al_2O_3, a feature becoming evident from the Al–Ti systematics (Fig. 10).

The Al_2O_3/TiO_2-TiO_2 diagram (Fig. 10) allows classification of ultramafic rocks. TiO_2 contents around 0.1% are typical for lherzolite. Modelling by BECCALUVA et al (1983) shows that the refractory residua of a spinel lherzolite mantle source retain their lherzolitic character after extraction of a MOR-type basalt ("high-Ti") by ca. 15% partial melting (dotted tie-line in Fig. 10) and display the observed TiO_2 values. Considering secondary enrichment of the ultramafic rocks in Al_2O_3, a generated melt-refractory residua relationship between the greenstones and the ultramafic rocks therefore seems possible (tie-line in Fig. 10). Moreover, the TiO_2 values of the ultramafic rocks are typical for a MOR origin as opposed to a supra-subduction origin (Fig. 11), thus corroborating the results of the greenstone chemistry.

Discussion

The geochemical data discriminate the North Penninic metabasalts as mid-ocean ridge basalts (MORB). Considering the chemical diagrams and their genetic implications derived from modelled petrogenetic pathways, the metabasalts formed by 10–15% partial melting of a slightly enriched mantle source along a slowly spreading ridge. Trace element relations favour a T-type MORB rather than an N-type MORB protolith. The accompanying serpentinized ultramafic rocks are classified as residual lherzolites formed due to

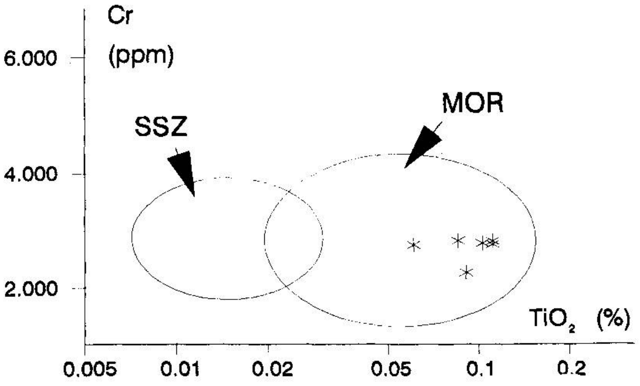

Fig. 11 Diagram after PEARCE et al. (1984) discriminating ultramafic rocks into mid-ocean ridge (MOR) and supra-subduction zone (SSZ) derivates.

magma extraction from a mantle source in a mid-ocean ridge environment.

The existence of oceanic crust in the North Penninic basin was proposed on grounds of paleogeographic reconstructions and the occurrence of ultramafic-basic igneous rock associations (e.g., FRISCH, 1978, 1979; DERCOURT et al., 1986; SCHMID et al., 1990). Our geochemical evidence fully confirms this concept. We like to stress, however, that large parts of the North Penninic sediments were deposited on the attenuated continental margins as evidenced by the geological relations in the Central Alps.

In paleogeographic reconstructions, the North Penninic basin only acquires a limited width (FRISCH, 1979; DERCOURT et al., 1986; SCHMID et al., 1990). The narrowness of the basin is also indicated by a thick sequence of predominantly detrital and turbiditic sediments rich in siliciclastic material. Ocean floor existed at least in parts of the North Penninic basin over a period of 80 Ma: this is the period of deposition of the Rhenodanubian flysch, which lacks a continental basement in favour of ophiolite fragments. Estimating the ocean floor as not more than a few hundred kilometers wide (see FRISCH, 1979), an average spreading rate in the order a few millimeters per year results. However, it seems likely that the opening of the basin occurred somewhat faster in the early stages of formation since there is no indication of synsedimentary widening of the basin. These observations are in accordance with a low spreading velocity of the North Penninic basin as may be suggested from the metabasalt geochemistry (Fig. 9; see above).

PEARCE (1980) investigated the chemistry of Mesozoic ophiolites in the Mediterranean region, including occurrences from the South Penninic realm in the Alps. He found rocks with geochemical signatures of MORB, intraplate basalts, and transitional types. This is in accordance with a recent comprehensive study of South Penninic ophiolites from the Arosa zone (FRISCH et al., in press.). Ophiolitic rocks of the Antrona and Zermatt-Saas zones of the western Central Alps were investigated by PFEIFER et al. (1989) and COLOMBI (1989). The metabasites from these zones also show a considerable chemical variety, thought to result partially from rock alteration (PFEIFER et al., 1989).

PEARCE (1980), however, interprets the chemical variety as the result of a geotectonic setting characterized by several small, Red-sea type basins related to the fragmentation of continental crust during initial opening of the Atlantic ocean. Such varied basalt types are today found in diffuse spreading centres as Iceland, Afar, or the Azores. In such environments an episode of mantle enrichment in incompatible elements prior to continental fragmentation, for instance by plume acitivity, still has its imprint on mantle source composition.

A genetic connection of the North and South Penninic oceans has been proposed by TRÜMPY (e.g., 1992) and SCHMID et al. (1990) for structural and sedimentological reasons. We therefore propose that the North Penninic mantle source for the formation of the ocean floor was either identical with, or strongly influenced by, the South Penninic mantle source. Despite the close neighbourhood of the two oceans (Fig. 1), the North Penninic metabasalts do not show the chemical variation of the South Penninic metabasalts. We interpret this in terms of mantle source depletion by magma extraction during formation of the South Penninic ocean floor, and homogenization of the mantle source. The homogenization between enriched and secondarily depleted mantle material with time gave rise to the only slightly enriched MOR-type basalts of the North Penninic realm. From geological reasoning the formation of ocean floor in the the North Penninic basin occurred later than in the South Penninic basin (e.g., FRISCH, 1979).

Acknowledgements

Thanks go to Roland Oberhänsli (Mainz), whose constructive criticism greatly improved an earlier version of the manuscript. The Geochemisches Zentrallabor in Tübingen carried out the XRF analyses. Ferdinand Thieme drafted many versions of the figures, including the final ones. The thoughtful review of an anonymous referee further added to the quality of the text. All this is gratefully acknowledged.

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Manuscript received April 4, 1993; revised manuscript accepted August 24, 1993.