



The Arosa Zone and Platta Nappe Ophiolites (Eastern Swiss Alps): Geochemical Characteristics and their Meaning for the Evolution of the Penninic Ocean

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11 Text-Figures and 3 Tables

*Ostalpen
Penninische Zone
Austroalpine Zone
Arosazone
Platta-Decke
Ophiolithe
Grünsteine
Ultramafite
Seltene Erden*

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Die Ophiolithe der Arosazone und der Platta-Decke (östliche Schweizer Alpen): Geochemische Charakteristik und ihre Bedeutung für die Entwicklung des Penninischen Ozeans

Zusammenfassung

Die Arosazone (Ostschweiz und Vorarlberg) bildet eine Mélange mit Anteilen der südpenninischen Ozeanbodenfolge und des ostalpinen Grund- und Deckgebirges in einer metapelitischen oder serpentinitischen Matrix. Die Plattadecke, die die Arosazone unterlagert, besteht aus intern weniger deformierten südpenninischen Ophiolithen. Die Metabasalte der Ophiolithe beider Zonen werden in zwei geochemisch unterschiedliche Gruppen mit unbekanntem Altersbezug unterteilt:

- I) Basalte, die an inkompatiblen Elementen angereichert sind und Charakteristika von Intraplattentholeiten besitzen, und
- II) Ozeanbodenbasalte des Normaltyps.

Zwischen beiden Gruppen bestehen Übergänge. Wir ziehen den Schluß, daß die angereicherten Basalte durch ca. 10 % Aufschmelzung einer Mantelquelle, die durch Diapirismus oder Metasomatose beeinflusst war, im Anfangsstadium der Bildung der südpenninischen ozeanischen Lithosphäre entstanden. Die Ozeanbodenbasalte des Normaltyps entstammen einer seichten Asthenosphärenquelle durch 15–20 % Aufschmelzung und wurden teilweise von der ersten, tieferen Quelle beeinflusst. Die ultramafischen Gesteine sind vorwiegend aus Lherzoliten, zu einem geringeren Teil auch aus Harzburgiten hervorgegangen. Das paläogeographische Milieu der Ophiolithe der Plattadecke und der Arosazone ist, entsprechend der chemischen Zusammensetzung der Gesteine und anderer geologischer Evidenz, das eines sich langsam ausbreitenden mittelozeanischen Rückens in einem schmalen ozeanischen Becken, das von zahlreichen Transformstörungen durchschnitten wurde.

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Abstract

The Arosa Zone of the eastern Swiss Alps is a *mélange* containing imbricates of the South Penninic ocean floor sequence as well as Austroalpine continental basement and cover rocks embedded in a metapelitic or serpentinitic matrix. The Platta Nappe underlying the Arosa Zone consists of internally less deformed South Penninic ophiolite. The metabasaltic rocks of the ophiolites of both zones are divided into two geochemically distinct groups of unknown age relationship:

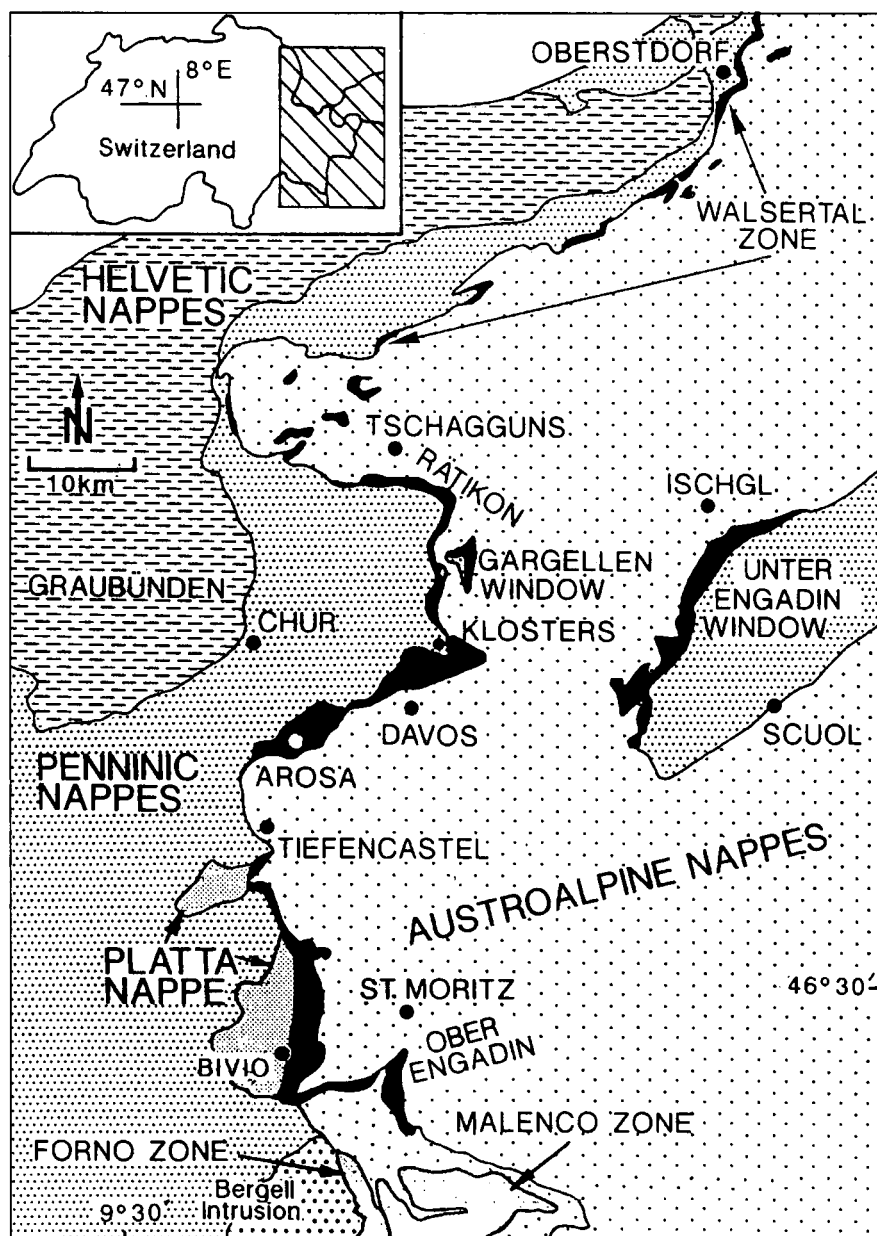
- I) basalts enriched in incompatible elements displaying characteristics of within-plate tholeiite, and
- II) normal-type mid-ocean ridge basalts.

Transitions between the two types occur. We propose that the enriched basalts were generated during the incipient stage of formation of South Penninic oceanic lithosphere by about 10 % partial melting of a mantle source, which was influenced by diapirism or metasomatism. The normal-type mid-ocean ridge basalts were derived from a shallow asthenospheric source by 15–20 % partial melting, and they are partly influenced by the first, deeper source. The ultramafic rocks are derived chiefly from lherzolitic and, to a lesser extent, from harzburgitic protoliths. The paleotectonic setting of the ophiolitic rocks of the Platta Nappe and the Arosa Zone as deduced from their chemical composition and other geologic evidence is a slowly spreading mid-ocean ridge in a narrow ocean cut by a series of transform faults.

1. Introduction

The ophiolites in the Alps are generally highly dismembered sections of ultramafic rocks, gabbro, dolerite, and pillow basalt associated with pelagic sedimentary rocks. They are interpreted as portions of a late Jurassic to early Cretaceous mid-ocean ridge system (LAUBSCHER, 1969; DEWEY & BIRD, 1970) cut by a series of transform faults (BERNOULLI & WEISSERT, 1985; WEISSERT & BERNOULLI, 1985). The South Penninic or Piemont Ocean separated the European continent in the northwest from the Adriatic promontory or microcontinent in the south-east (CHANNELL & HORVÁTH, 1977; FRISCH, 1979; DERCOURT et al., 1986). From the South Penninic and, in its continuation, the Ligurian ocean floor a number of ophiolite occurrences derived. They can be traced from southern Italy (LANZAFAME et al., 1979) across the northern Apennines (DECANDIA & ELTER, 1972) and the Western and Central Alps (BEARTH, 1967; DIETRICH, 1969; DE CAPITANI et al., 1981; PFEIFER et al., 1989) into the Penninic windows of the Eastern Alps (BICKLE & PEARCE, 1975; HEUGEL, 1982; VUICHARD, 1984; KOLLER & HÖCK, 1987). Many of these ophiolites underwent Cretaceous high-pressure metamorphism which was overprinted in the Lower Tertiary by regional medium-pressure metamorphism (BEARTH, 1967; DIETRICH & PETERS, 1972; FREY, 1974; DESMONS, 1986; GOFFÉ & CHOPIN, 1986; MILLER, 1986; FRY & BARNICOAT, 1987).

In eastern Switzerland, Penninic ophiolites mainly occur in the Platta Nappe, the Forno and Malenco Zones, the Arosa Zone and the Unterengadin Window (Text-Fig. 1). Pillow basalts associated with hyaloclastites and pelagic sedimentary rocks, and less abundant serpentinite and gabbro make up most of the Platta Nappe (DIETRICH,



Text-Fig. 1.
Geological sketch map of westernmost part of Eastern Alps.
The Arosa Zone and its equivalents in the Unterengadin Window are shown in black, Penninic ophiolite units (Platta Nappe, Forno and Malenco Zones) are in light shading. Note that the Arosa Zone is separated from the Platta nappe, which is in contrast to previous work (e.g., DIETRICH, 1970).

1970), which immediately underlies the Arosa Zone (RING et al., 1990). To the south, we consider the Forno and Malenco Zones as the continuation of the Platta Nappe. The Forno Zone comprises ultramafic and mafic rocks associated with pelagic sedimentary rocks and flysch (PERETTI, 1985). The Malenco Zone contains only ultramafic rocks (DE CAPITANI et al., 1981).

The Arosa Zone is a narrow, internally imbricated zone with an overall *mélange* structure sandwiched between the underlying Penninic and the overlying Austroalpine nappe systems (RING et al., 1990; Text-Fig. 1). It contains elements of both the Penninic and Austroalpine sequences. Based on detailed mapping, we separated the Arosa Zone from the Platta Nappe and traced it to the south into the Oberengadin. Its northern part is referred to as the Walsertal Zone by WINKLER (1988). Equivalents of the Arosa Zone are found in the Gargellen and Unterengadin windows (Text-Fig. 1; BERTLE, 1973; GÜRLER, 1982; RING et al., 1989) as well as in the Tauern Window (Matrei Zone; FRISCH et al., 1987; WAIBEL & FRISCH, 1989).

The Arosa Zone comprises ophiolitic material equivalent to that of the Platta Nappe as dismembered imbricates. The imbricates are made up of serpentinite as the most prominent member, as well as gabbro, rare rhyolite dikes, dolerite, pillow basalt, ophiocarbonate, and pelagic sedimentary rocks. Based on immobile trace elements, the pillow basalts of the Platta Nappe were determined by TROMMSDORFF & DIETRICH (1980) as ocean floor basalt. The serpentinites were interpreted as tectonized mantle lherzolites and harzburgites which were emplaced into the oceanic crust along fracture zones (WEISERT & BERNOULLI, 1985).

The rocks of the Arosa Zone are unmetamorphosed in the north (Walsertal Zone; Text-Fig. 1) and underwent low grade metamorphism in the south. The Platta Nappe also experienced low grade metamorphism.

The petrography and mineralogy of the ultrabasic and basic magmatic rocks of the Arosa Zone are described by PETERS (1963) and BURKHARD (1987). Only few chemical data are available (PETERS, 1963; TROMMSDORFF & DIETRICH, 1980). A regional study based on immobile trace elements and an overall interpretation of the geochemical results is lacking. In the present article we summarize the petrographic characteristics of the ophiolitic rocks and examine their chemical composition in order to constrain the geotectonic environment in which these rocks formed. X-ray fluorescence analysis was carried out in the frame of the doctoral thesis of RING (1989) and the diploma theses of BIEHLER (1988), BORCHERT (1990), and DÜRR (1989); additional sampling was done by DÖRRE (1989) and KNAUS (1990) during diploma mapping. A comparative study concerning the North Penninic ophiolites of eastern Switzerland is provided by DÜRR et al. (1993).

2. Geologic Setting

The Arosa Zone represents the main Alpine suture and is characterized by highly variable thickness (Text-Fig. 1; RING et al., 1990). Blocks and imbricates of any size are embedded in a matrix of Cretaceous shale and flysch or serpentinite. Components of South Penninic origin are the ophiolitic rocks, Upper Jurassic to Lower Cretaceous radiolarite and pelagic limestone, as well as Lower Cretaceous series with an increasing terrigenous input (Palombini and Lavagna Formations; LÜDIN, 1987). Austroalpine

members encompass Variscan basement rocks, Permo-Scythian arkose and quartzite, Middle to Upper Triassic dolomite, as well as Jurassic sandstone, shale and scarp breccia (Saluver Formation), radiolarian chert, and pelagic limestone. A paleogeographic position of the Arosa Zone close to the Austroalpine margin is indicated by numerous Austroalpine slabs which particularly contribute to the overall *mélange* character (LÜDIN, 1987; WAIBEL & FRISCH, 1989; RING et al., 1990).

The Platta Nappe is composed of a moderately tectonized ophiolite sequence that lacks exotic elements. We define the boundary between the Platta Nappe and the Arosa Zone as the boundary between the purely ophiolitic assemblage and the imbricated zone containing both Penninic and Austroalpine elements.

3. Greenstones (Metabasalt and Metadolerite)

Metabasalts in the Arosa Zone and Platta Nappe are fine grained and show evidence of an extrusive origin. Metadolerites are coarser grained (primary pyroxene and plagioclase generally > 1 mm), massive basaltic rocks. Metamorphism and deformation have only partly obliterated these primary features.

3.1. Petrography

The metabasalt frequently shows pillow structure. It occurs as fault-bounded, lens-shaped blocks and slices in flysch, or as flows in pelagic sedimentary rock. The best outcrop of pillow basalt is found at Hörnli west of Arosa. This pinnacle consists of pillow lava, pillow breccia, dolerite, and minor hyaloclastite, fault-bounded within pelagic sediment. The pillows show an increase in grain size from rim to core, which is interpreted in terms of inheritance of a primary zoning not destroyed by the metamorphic overprint. The originally glassy pillow rinds which are rich in chlorite, contain abundant pink and green varioles ($\varnothing$ 3–15 mm) composed of radiating albite fibres intergrown with prehnite and sphene, and show a mosaic texture typical of devitrified material (TROMMSDORFF & DIETRICH, 1980). Red slate, chert, and siliceous carbonate fill the interstices of the pillows.

Metadolerite occurs less frequently than metabasalt. Contacts between metadolerite and metabasalt or other rocks are generally tectonically overprinted. Metagabbro occurs as lenses within serpentinite and metabasalt; cumulate structures were not observed.

The mineralogical characteristics of the metabasalts and metadolerites near Arosa (Text-Fig. 1) are as follows: Augite and plagioclase originally formed an intersertal to subophitic texture in the basalt and a subophitic to ophitic texture in the dolerite. Augite is altered to sphene, pumpellyite or epidote, and chlorite, but augite relics occur in cores of pillows and in dolerite. Euhedral plagioclase shows bending of twin lamellae but no recrystallization. Although the shape of the primary plagioclase is preserved, its composition changed to albite. Part of the basalt and dolerite contained primary olivine which is altered to chlorite, serpentine, and opaque minerals.

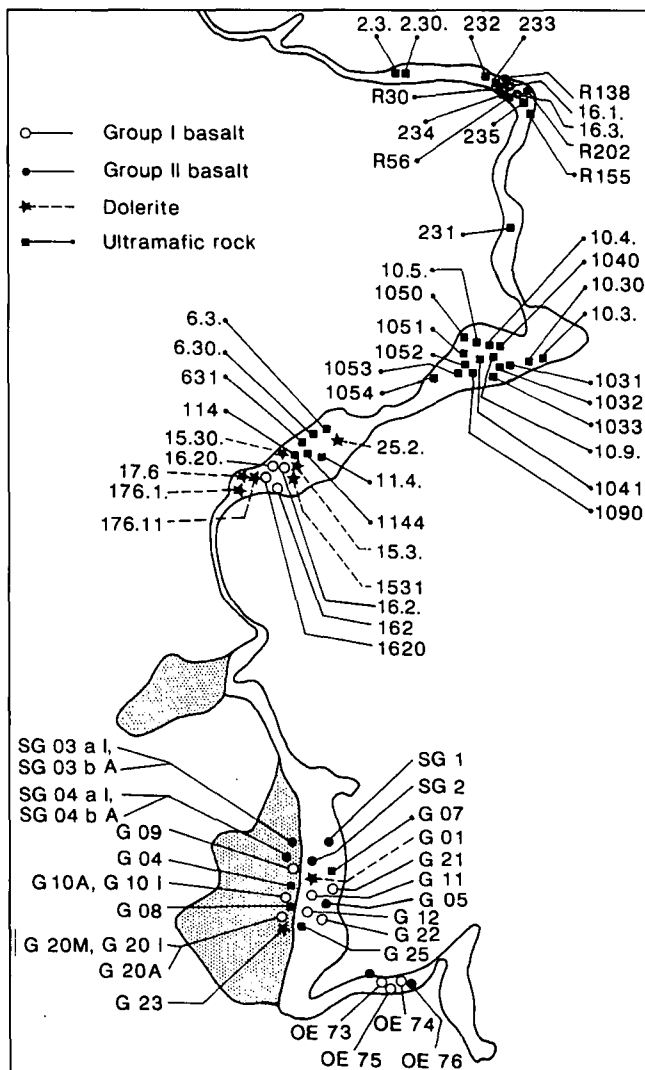
During regional metamorphism albite, chlorite, pumpellyite, prehnite, epidote, actinolite, calcite, quartz, hematite, and sphene formed in the area around Arosa

(TROMMSDORFF & DIETRICH, 1980). The occurrence of both pumpellyite and epidote indicates that the very-low grade/low grade boundary passed through this area. Mg-riebeckite, riebeckite, and richterite are reported from south of Bivio (e.g., PHILIPP, 1982).

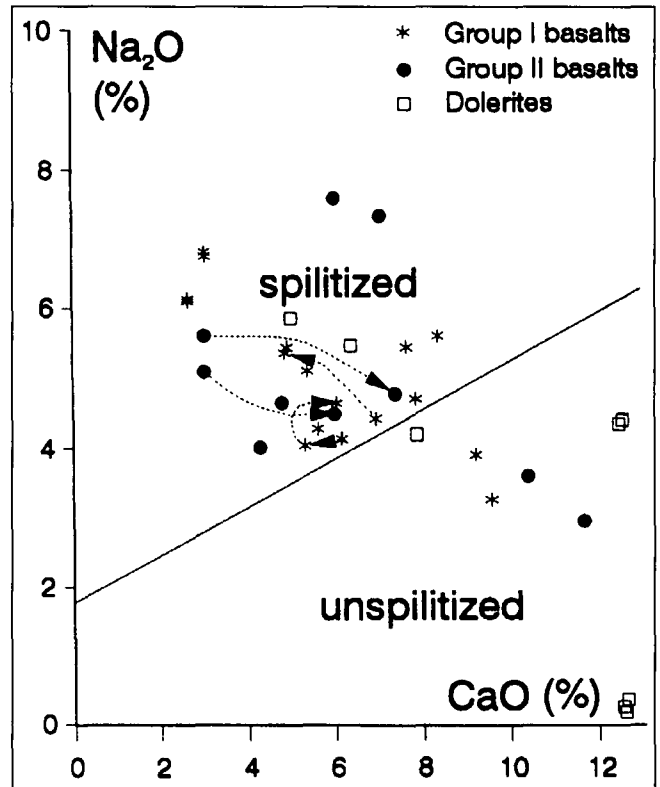
3.2. Geochemistry

The results of chemical analyses of 38 greenstone samples from the Arosa Zone between Rätikon and Engadin and from the Platta Nappe near Bivio are presented in Table 1. The sample localities are shown in Text-Fig. 2. For use in the diagrams (Text-Fig. 3–10), the chemical analyses were recalculated with 1 wt-% H₂O and no CO₂.

The overall geochemical pattern fits that of basaltic rocks but major element mobility is evident. No systematic difference in the chemistry between the Arosa Zone and the Platta Nappe greenstones exists so that the results of the samples from both tectonic units are treated together. Two groups of metabasalt are discerned by their different chemical patterns, referred to as Group I and Group II basalts. The metadolerites are similar to the Group II basalts



Text-Fig. 2. Location map (compare Text-Fig. 1) of analyzed samples from Arosa Zone (without shading) and Platta Nappe (light shading). Numbers are sample numbers as shown in Tables 1, 2, and 3. I, M, A refers to inner, intermediate, and outer zones of single basalt pillows.



Text-Fig. 3. Diagram after STILLMAN & WILLIAMS (1978) for discriminating between spilitized and unspilitized basalts. Arrows point from inner to outer zones of single pillows.

but in some respect transitional between the two groups (Text-Fig. 3–10).

3.2.1. Element Mobility

Spilitization affected most pillow basalts, but the dolerites to a lesser degree (Text-Fig. 3). This is interpreted in terms of ocean floor metasomatism. Pillows across which the chemical zoning was studied show no uniform trend between core and rim (Text-Fig. 3). Metasomatic and metamorphic alteration caused intense hydration and, in part, carbonatization of the rocks. This is reflected by H₂O contents of up to 8 wt-% in chlorite-rich samples, and by CO₂ contents of up to 5 wt-% in calcite bearing samples. Three highly deformed dolerite samples are strongly depleted in Na (Text-Fig. 3) and show low silica and elevated iron contents. These element changes are considered to have taken place during destruction of primary pyroxene and plagioclase and passive enrichment of iron oxide phases during regional metamorphism and deformation (see DOSTAL et al., 1980; FRISCH & RAAB, 1987).

Because of element mobility, the interpretation of the geotectonic environment in which the greenstone protoliths formed is mostly based on high field strength (HFS) trace elements. These elements are known to be largely immobile during sub-seafloor weathering and low-grade metamorphism (e.g., CANN, 1970; HUMPHRIS & THOMPSON, 1978). An immobility test are correlations with Zr as a reference (PEARCE, 1982). The correlation diagrams (Text-Fig. 4) infer stability of Zr, Ti, Y, P, and V, partial mobility of Mn, Ba, and Cr, and strong mobility of K. Values analyzed for Nb and Rb generally lie below the detection limit. A correlation of the incompatible trace elements with either H₂O or FeO/Fe₂O₃ as mobility indicators does not exist.

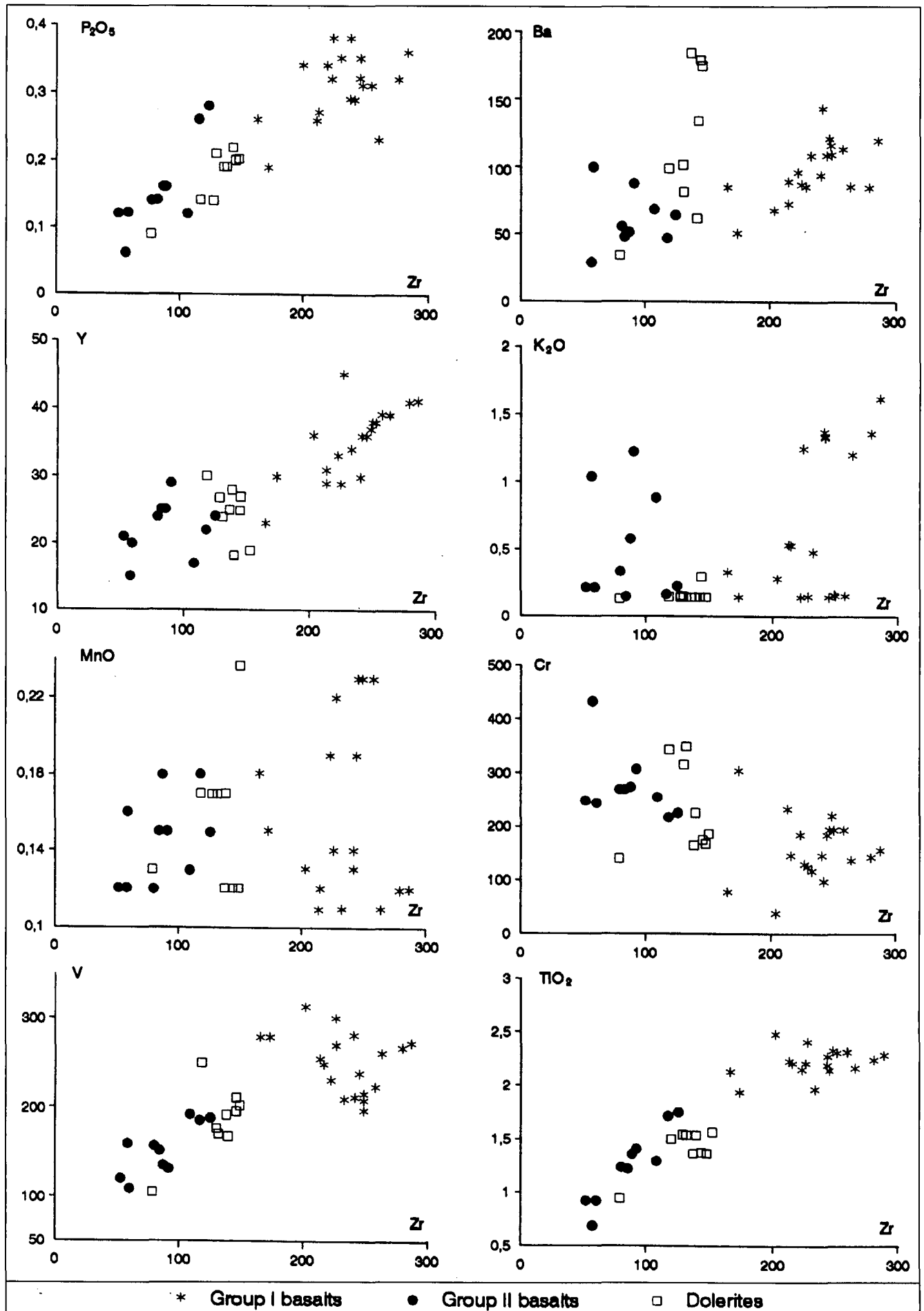
Table 1.

Geochemical results of Group I basalts, Group II basalts, and dolerites from Arosa Zone and Platta Nappe.

Main and minor elements in weight percent, trace elements in ppm. * with FeO refers to total iron. ** with H₂O refers to loss on ignition. Nb values are <10 ppm.

For sample locations, see Text-Fig. 2. Samples labeled „R“ are from BIEHLER (1988), „SG“ from BORCHERT (1990), „G“ from DÜRR (1989), „OE“ from KNAUS (1990), and without letter from RING (1989).

	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	H ₂ O	CO ₂	Total	Ba	Cr	Ni	Rb	Sr	V	Y	Zn	Zr
Group I basalts																							
16.1.	49.72	2.08	16.44	7.29	2.40	0.18	7.67	2.84	6.61	< 0.14	0.33	1.55	2.66	99.91	95	181	88	< 12	98	232	32	93	216
16.2.	48.64	2.25	17.59	5.91	2.36	0.22	9.32	2.45	5.93	< 0.14	0.30	1.71	2.82	99.64	110	187	103	< 12	56	214	38	135	249
16.20.	48.66	2.24	17.47	5.71	2.57	0.22	9.29	2.45	5.89	< 0.14	0.31	1.71	2.78	99.44	118	215	109	< 12	86	199	37	134	240
162	48.56	2.23	17.46	5.87	2.41	0.22	9.34	2.46	5.93	< 0.14	0.30	1.76	2.75	99.43	106	190	120	< 12	90	190	37	131	241
1620	48.57	2.24	17.56	5.63	2.66	0.22	9.34	2.46	5.93	< 0.14	0.34	1.69	2.80	99.58	113	189	118	< 12	51	209	36	136	229
16.3.	50.26	2.10	16.55	7.23	2.54	0.18	7.70	2.86	6.58	< 0.14	0.28	1.46	2.29	100.17	107	179	92	12	110	231	35	88	238
OE 73	50.54	1.92	16.46	4.49	3.60	0.11	4.33	8.12	5.47	0.47	0.34	3.50 **	—	99.35	107	114	91	15	301	204	33	84	227
OE 74	46.83	2.29	14.53	3.52	7.67	0.21	5.20	9.05	3.09	< 0.14	0.36	6.35 **	—	99.24	81	118	58	< 12	177	283	43	113	216
OE 75	49.12	2.47	14.52	2.21	7.20	0.13	7.39	9.13	3.88	0.28	0.34	1.75 **	—	98.42	69	37	< 40	< 12	261	311	36	27	201
G 09	51.97	2.18	15.02	1.62	7.05	0.11	7.37	4.75	5.35	0.53	0.27	3.02	0.46	99.70	73	145	74	< 12	173	244	28	74	211
G 10 A	51.61	2.19	15.04	1.25	7.12	0.12	7.43	4.70	5.27	0.53	0.26	3.05	0.51	99.08	89	230	51	< 12	175	250	30	74	210
G 10 I	47.63	2.12	15.38	1.46	7.80	0.13	7.38	6.63	4.26	1.20	0.31	3.54	2.04	99.88	85	125	< 40	13	274	259	28	78	217
G 11	47.76	2.21	16.20	2.70	5.67	0.13	7.42	5.39	4.17	1.29	0.28	3.56	0.84	97.82	93	141	65	12	449	271	29	78	234
G 12	49.46	2.11	14.50	2.13	7.71	0.18	7.05	7.70	4.64	0.32	0.26	2.82	0.51	99.39	85	78	49	< 12	48	274	23	70	163
G 20 A	50.24	2.11	15.28	1.73	6.78	0.11	6.92	5.84	4.53	1.18	0.22	2.99	1.28	99.11	85	135	64	< 12	272	254	38	67	257
G 20 I	48.32	2.18	15.58	1.90	7.40	0.12	7.36	5.92	4.01	1.32	0.31	3.32	1.49	99.23	83	141	67	< 12	285	260	40	75	271
G 20 M	48.20	2.23	16.17	2.26	7.61	0.12	7.30	5.11	3.94	1.57	0.35	3.57	0.87	99.30	118	154	80	< 12	351	265	40	77	279
G 21	46.93	2.10	15.71	6.50	2.96	0.13	4.75	7.26	5.20	1.30	0.36	4.81	0.99	99.00	139	93	213	41	103	202	34	142	231
G 22	49.38	1.90	14.93	2.78	6.46	0.15	9.24	5.20	5.01	< 0.14	0.19	3.57	0.28	99.23	51	299	192	< 12	69	272	29	80	170
Group II basalts																							
OE 76	48.85	0.68	15.97	1.13	4.78	0.12	8.82	11.34	2.86	1.01	0.06	3.98 **	—	99.60	29	421	93	< 12	108	155	15	23	56
SG 1	49.09	1.66	17.74	—	9.01 *	0.17	9.03	4.04	3.86	0.15	0.25	5.04 **	—	100.04	46	210	131	< 12	96	179	21	78	113
SG 2	49.53	1.70	18.23	—	9.01 *	0.14	7.90	4.57	4.50	0.22	0.27	4.54 **	—	100.61	64	217	134	< 12	122	182	22	83	122
SG 03 a I	47.14	1.37	20.19	—	8.97 *	0.14	8.20	2.81	4.91	1.18	0.15	4.74 **	—	99.80	85	296	222	30	70	128	28	96	88
SG 03 b A	45.19	1.20	18.23	—	13.20 *	0.12	5.35	5.72	4.33	0.33	0.13	4.82 **	—	98.62	54	260	191	< 12	96	151	23	74	76
SG 04 a I	47.61	1.33	19.78	—	9.33 *	0.17	8.56	2.82	5.42	0.55	0.15	4.61 **	—	100.33	50	265	213	16	78	131	24	85	85
SG 04 b A	48.13	1.18	17.39	—	11.20 *	0.14	4.30	7.05	4.59	0.14	0.13	5.08 **	—	99.33	48	260	173	< 12	134	144	24	51	82
G 05	48.01	1.26	16.61	2.00	5.25	0.13	7.47	10.12	3.50	0.87	0.12	3.42	0.83	99.59	68	249	110	17	230	187	17	47	106
R138	51.69	0.92	16.34	—	8.26 *	0.12	3.64	5.89	7.52	0.31	0.12	1.92	3.16	99.89	69	246	221	< 12	167	119	21	55	52
R202	50.39	0.93	15.84	—	7.03 *	0.16	3.74	6.91	7.25	0.22	0.12	2.11	4.39	99.09	100	242	249	13	147	107	20	55	59
Dolerites																							
15.3.	44.56	1.29	16.17	4.99	2.47	0.11	5.50	11.66	4.09	< 0.14	0.20	3.16	5.06	99.40	167	164	86	< 12	183	183	23	48	134
15.30.	44.47	1.29	16.10	4.85	2.61	0.11	5.49	11.65	4.08	< 0.14	0.19	3.11	5.08	99.17	162	157	79	< 12	183	195	25	46	136
1531	44.44	1.28	16.17	4.97	2.50	0.11	5.49	11.59	3.99	< 0.14	0.21	3.07	5.20	99.16	164	146	92	< 12	103	183	22	50	130
17.6	43.47	1.46	16.23	7.21	2.91	0.16	8.75	11.81	0.22	< 0.14	0.13	2.66	4.35	99.50	97	297	165	< 12	36	165	25	63	122
176.1.	43.55	1.45	16.18	7.44	2.74	0.16	8.61	11.91	0.30	< 0.14	0.20	2.71	4.36	99.75	78	326	185	< 12	37	160	23	66	124
176.11	43.59	1.45	16.23	7.34	2.81	0.16	8.61	11.86	< 0.14	< 0.14	0.18	2.65	4.34	99.50	60	215	182	< 12	41	158	26	65	132
25.2.	48.59	1.45	15.98	5.77	2.88	0.16	8.04	6.07	5.25	< 0.14	0.13	1.89	3.12	99.47	96	330	117	< 12	455	239	29	64	114
G 01	50.44	1.64	17.09	1.34	6.77	0.23	6.31	4.89	5.87	0.42	0.21	3.46	0.78	99.45	150	176	< 40	19	159	195	19	91	155
G 23	47.51	0.92	17.75	1.16	7.02	0.13	8.13	7.60	4.07	< 0.14	0.09	3.96	0.82	99.30	< 40	139	109	< 12	175	102	13	55	77



Text-Fig. 4.

Correlation diagrams using Zr as a reference (PEARCE, 1982).
Zr, Y, V, Ba, Cr in ppm, TiO_2 , P_2O_5 , MnO, K_2O in weight percent.

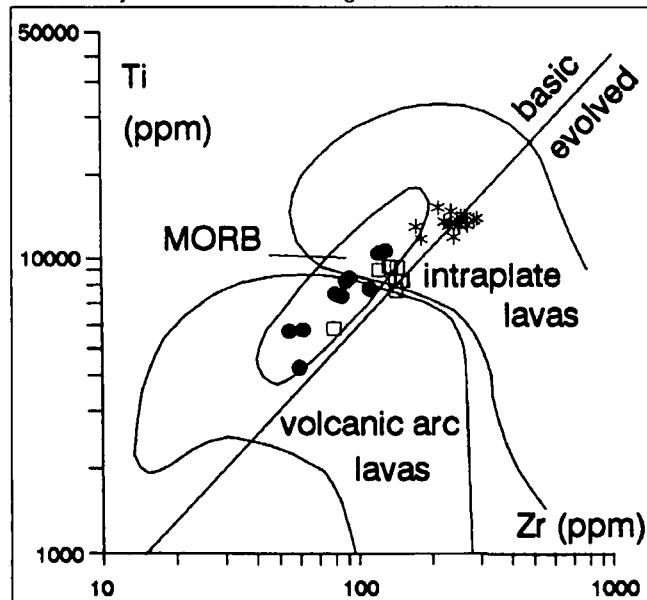
3.2.2. Basaltic Nature of Protoliths

A prerequisite for the use of basalt discrimination diagrams is that the original composition of the rock was in fact basaltic. According to BECCALUVA et al. (1983) the Al_2O_3 and TiO_2 contents of our samples are generally those of primary basaltic melts and exclude a cumulate origin. Moreover, among the major elements the MgO and FeO_{tot} contents of most samples are in the range of basaltic rocks. We therefore suggest that these elements did not substantially change their abundances with the exception of those 20 % of samples which obviously lost MgO by alteration processes (values around or below 5 wt-%). Omitting these samples, the average Mg number $[100 \text{ Mg}/(\text{Mg} + \text{Fe}_{\text{tot}})]$ is 62, and the average M number $[100 \text{ Mg}/(\text{Mg} + \text{Fe}^{+2})]$ – assuming around 15 % Fe^{3+} of total Fe in the protolith – is around 65. These values indicate high-level fractionation after segregation from the mantle source, since an M value of 70 defines basaltic magma in equilibrium with mantle olivine (WILSON, 1989). Ni contents <250 ppm ($\bar{\varnothing}$ 120 ppm) and Cr contents <400 ppm ($\bar{\varnothing}$ 200 ppm) underline the evolved nature of the basalts, with olivine and chrome spinel fractionation playing an important role.

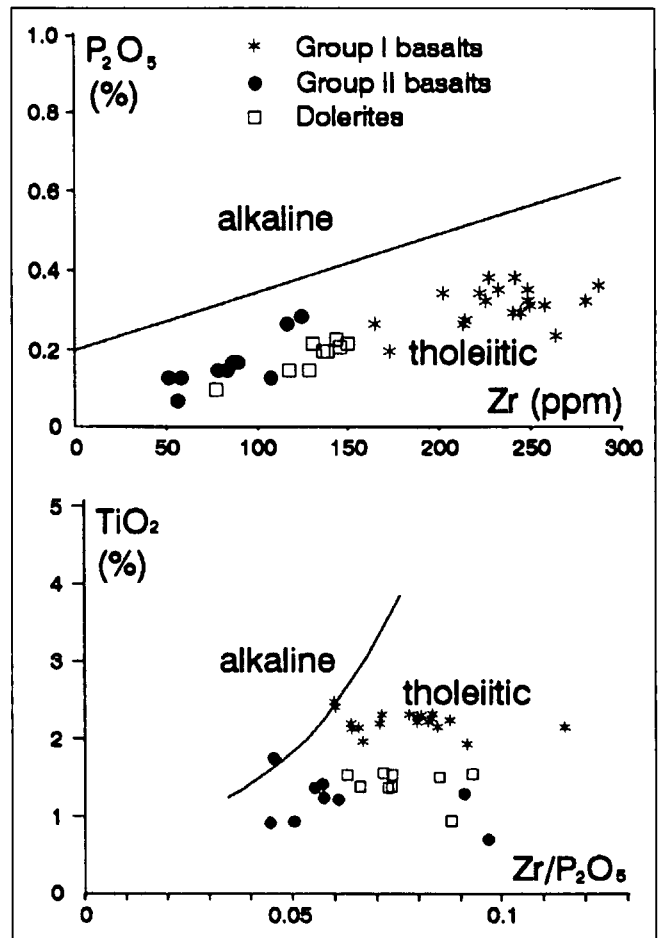
The Ti-Zr diagram (Text-Fig. 5; PEARCE et al., 1981) allows screening between basaltic and more acidic protoliths. Since the precipitation of a Ti bearing oxide causes or accompanies an increase of silica in the melt, the transition from a basic to a more acidic magma is generally accompanied by a decrease in the Ti/Zr ratio. The studied greenstone samples cluster within the basic field or around its boundary.

3.2.3. Discrimination of Protolith Environment

A selection of diagrams is used to discriminate the basaltic rocks in terms of the geotectonic environment in which they were formed. Two geochemical trends are ob-



Text-Fig. 5.
Ti-Zr diagram after PEARCE et al. (1981) showing fields of mid-ocean ridge basalts (MORB), volcanic arc and intraplate lavas. Straight line separates basaltic rocks from their differentiates.



Text-Fig. 6.
Diagrams after FLOYD & WINCHESTER (1975) for discriminating between alkaline and tholeiitic basalts.

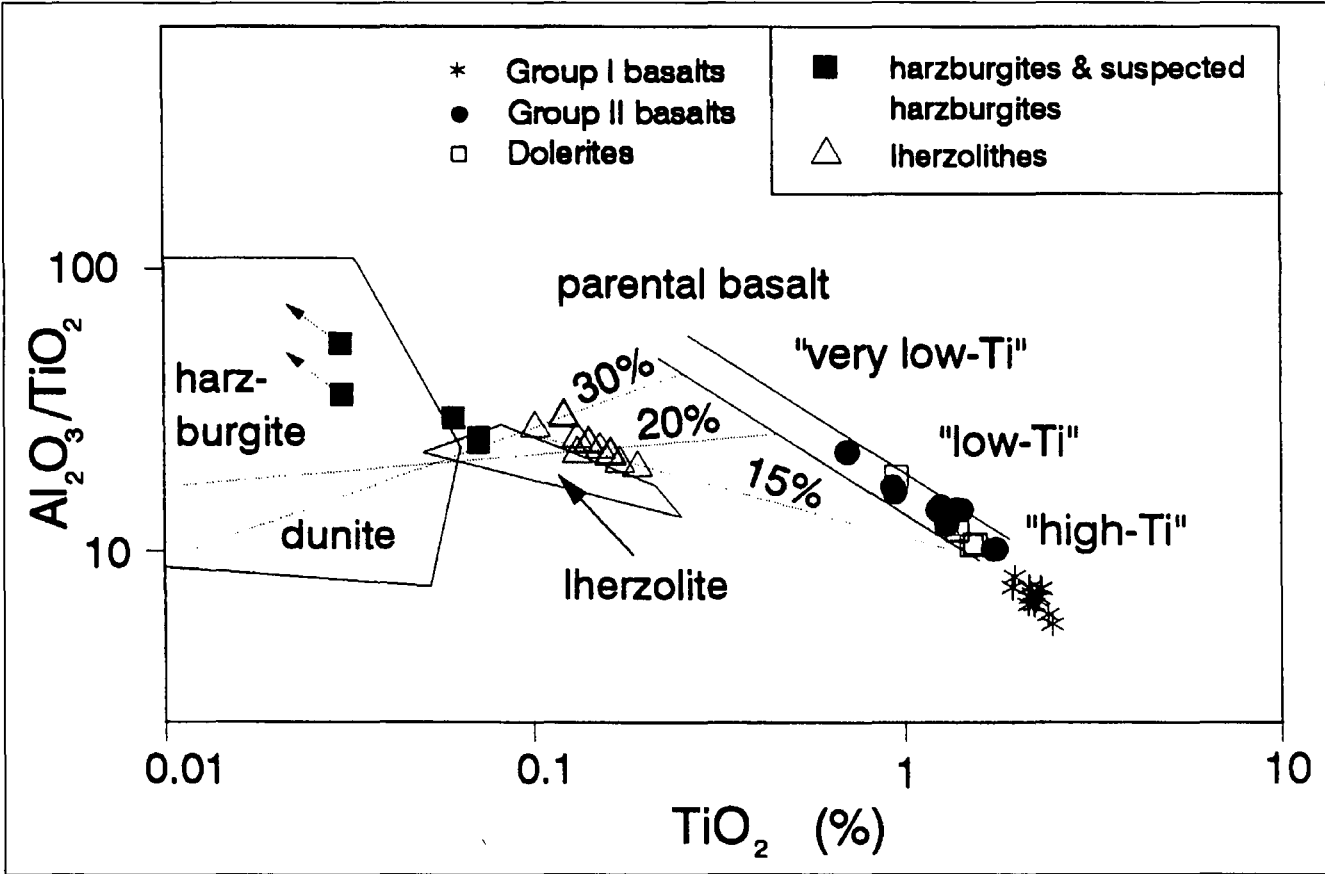
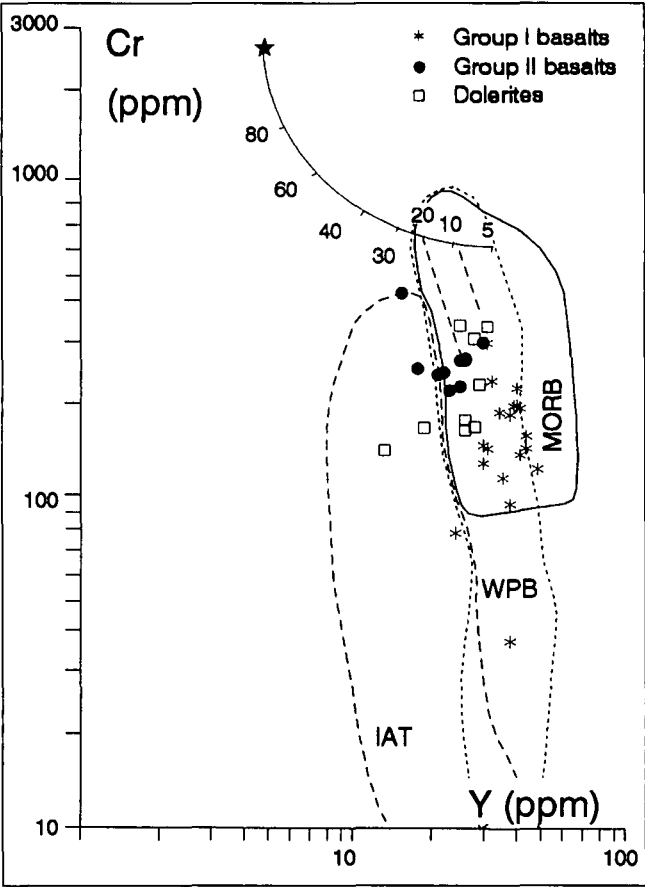
served for the pillow basalts. The Group I basalts display higher concentrations in P, Zr, Ti, and Y and high Y/Cr ratios relative to the Group II basalts and dolerites. The Group II basalts and dolerites show concentrations in P, Zr, Ti, Y, and Cr similar to those of normal-type mid-ocean ridge basalt (N-type MORB; see Text-Fig. 10).

In the Ti-Zr diagram after PEARCE et al. (1981) the Group II basalts and dolerites cluster in the mid-ocean ridge basalt (MORB) field (Text-Fig. 5). The Group I basalts form a separate cluster which reveals a within-plate component. A generally subalkalic character is displayed in the diagrams of FLOYD & WINCHESTER (1975; Text-Fig. 6). The higher contents in the incompatible elements Ti, Zr, and P of the Group I basalts suggest an enriched mantle source and/or a lower degree of partial melting than the Group II basalts and dolerites. The Cr-Y pathways of the magma as modelled by PEARCE (1980) indicate, despite some scatter of the data, a lower amount of partial melting of the mantle source for the Group I basalts (Text-Fig. 7). This is corroborated by the $\text{Al}_2\text{O}_3/\text{TiO}_2$ - TiO_2 systematics according to BECCALUVA et al. (1983). There the Group II basalts and dolerites classify as high- to low-Ti basalts with partial melting between 15 and 20 % whereas the Group I basalts are shown to have been generated by partial melting of less than 15 % (Text-Fig. 8).

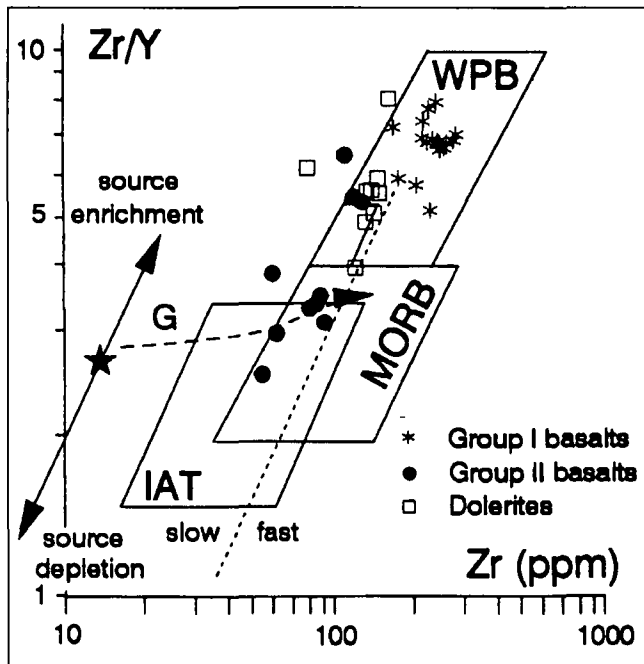
Text-Fig. 7.
Diagram after PEARCE (1980) and PEARCE et al. (1984) for modelling genetic pathways of basaltic magmas. Asterisk shows element concentration for mantle source C₃ chondrite. Numbers on melting curve refer to percent of partial melting. Straight dashed lines indicate pathways of magma evolution during fractionation of mafic minerals. MORB = mid-ocean ridge basalt; WPB = within-plate basalt; IAT = island arc tholeiite.

The magma pathways in the Zr/Y-Zr system for partial melting in the order of 15 % and subsequent fractional crystallization indicate that the great part of the Group II basalts, which show a MORB character, can be derived from a mantle source approximately matching primordial composition (Text-Fig. 9). This is in agreement with PEARCE's (1980) results from the Grossglockner ophiolites in the Tauern Window (path G in Text-Fig. 9). Group I basalts clearly plot in the within-plate field and derive from an enriched mantle source. A part of the Group I and II basalts and the dolerites take an intermediate position between the two groupings and show transitions between the basalts displaying within-plate and mid-ocean ridge characteristics.

A synoptical presentation of the geochemical pattern is facilitated by the MORB normalized multi-element diagrams after PEARCE et al. (1981) (Text-Fig. 10). The meaning of the values of the large ion lithophile (LIL) elements Sr, K, Rb, and Ba is limited since these elements are mobile during alteration processes (see Text-Fig. 4). The flat

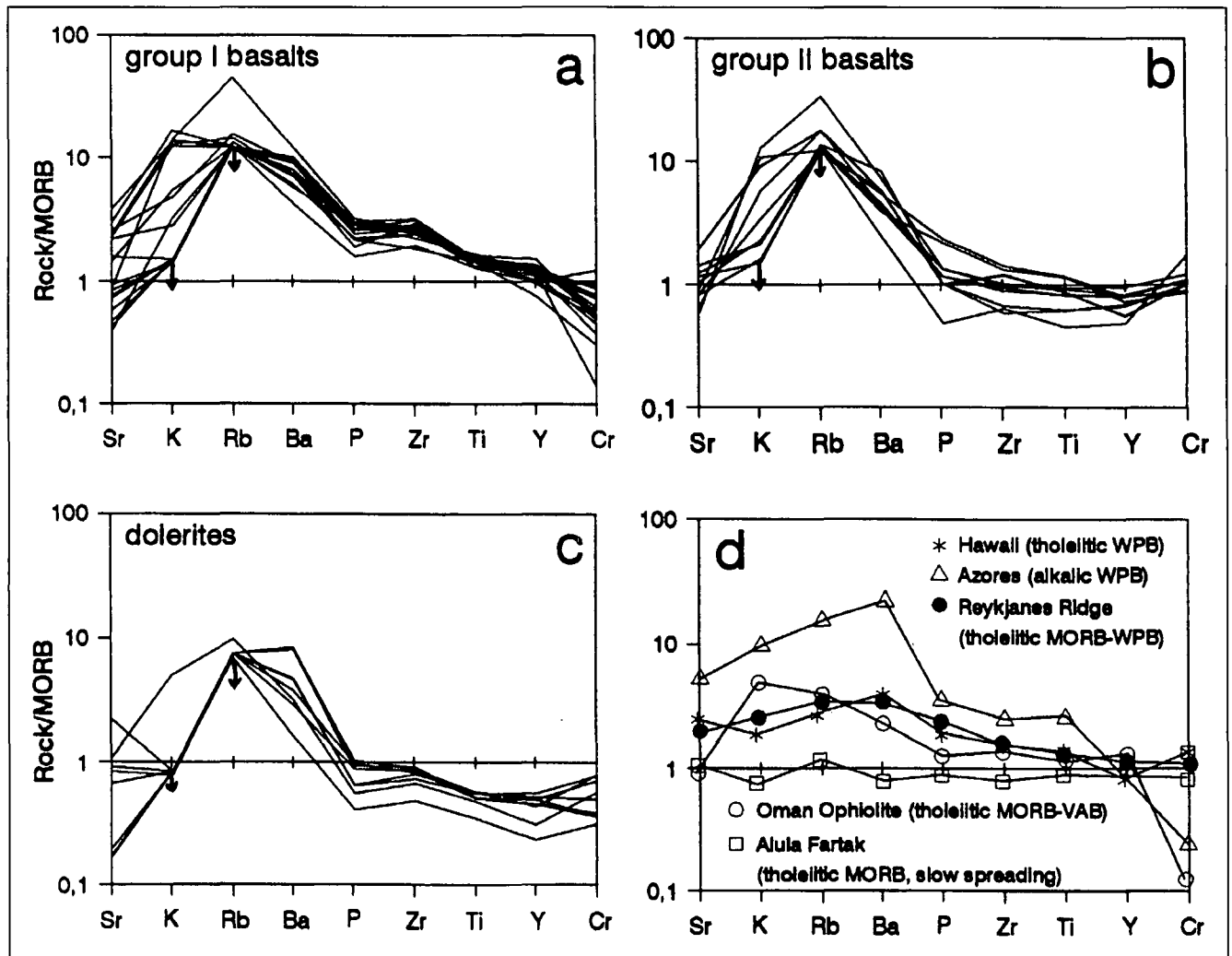


Text-Fig. 8.
Diagram after BECCALUVA et al. (1983) showing genetic relationships (dotted lines) between basaltic melts and ultramafic residua derived from a hypothetical pyrolite source. Percent numbers refer to amount of partial melting.



Text-Fig. 9.
Diagram after PEARCE (1980) for modelling genetic pathways of basaltic magmas.
Arrowed line with asterisk shows values for depleted and enriched mantle sources relative to C_3 chondrite (asterisk). Dashed arrow labeled „G“ indicates magma evolution path during partial melting and fractional crystallization for Grossglockner ophiolites (Tauern Window) after PEARCE (1980). Dotted line separates MORB from slow and fast spreading ridges (‘‘2 cm/a’’ unilateral spreading rate). Abbreviations as in Text-Fig. 7.

HFS element pattern of most Group II basalts and some dolerites with values close to unity (Text-Fig. 10b,c) reflect a mid-ocean ridge origin of the basalts. The Cr values also agree with the average of MORB. We interpret the negative slope from P to Y in several samples, especially the dolerites, in terms of an influence of a within-plate component. A pronounced within-plate or E-(enriched-)type MORB character is evident in the Group I basalts (Text-Fig. 10a). The curves harmonize with tholeiitic within-plate basalts (WPB) as from Hawaii or with transitional MORB-WPB as from the Reykjanes ridge (Text-Fig. 10d). According to MESCHÉDE (1986) the high Zr/Y ratio of the



Text-Fig. 10.
MORB-normalized multi-element diagrams after PEARCE et al. (1981) for basaltic rocks.

a: Group I basalts; b: Group II basalts; c: dolerites; d: for comparison, basalts from different geotectonic settings after PEARCE et al. (1981, 1984) and PEARCE (1982).

Nb concentrations in a-c are all below detection limit (10 ppm) and have therefore been omitted in the graphs. Downheaded arrows (K, Rb) indicate concentrations below detection limit.

Group I basalts support the within-plate rather than an E-type MORB character. The Cr values are generally below the MORB average and partly show a negative correlation with Y. This indicates a low degree of partial melting or differentiation and early fractionation of Cr in the magma.

A depletion in HFS elements relative to MORB in some Group II basalts (Text-Fig. 10b) puts the question of a subduction component in the mantle source. The high Ti/V ratios of around 50 do not support this view but are consistent with a MORB or MORB/WPB-transitional environment (SHERVAIS, 1982).

The Zr/Y-Zr diagram (Text-Fig. 9) is capable of discriminating between slow-spreading and fast-spreading ridges (PEARCE, 1980). However, the derivation of spreading rates from metamorphic basalts presents problems due to possible small-scale element mobility. Therefore, conclusions should only be drawn in connection with other supporting observation. The data (Text-Fig. 9) suggest that the studied samples with MORB characteristics derive from a slow-spreading ridge (unilateral spreading rate <2 cm/a). Such a ridge is characterized by small magma chambers giving rise to primitive, less differentiated melts. Therefore, incompatible elements are less concentrated compared with basalts from fast-spreading ridges. The slightly depleted HFS element pattern of some Group II basalts (Text-Fig. 10b) is similar to that of slow-spreading ridges such as the Alula Fartak ridge (Text-Fig. 10d; PEARCE et al., 1981). Small magma chambers produce limited amounts of cumulate rocks according to thermal models (e.g., SLEEP, 1975) and geophysical observation (e.g., NISBET & FOWLER, 1978). The scarcity of cumulate rocks therefore supports the assumption of a slowly spreading ridge. There are additional arguments in favour of a low spreading rate (see below).

3.2.4. Rare Earth Element Patterns

The rare earth element (REE) concentrations of the studied greenstones (Table 2) yield flat patterns in the chondrite normalized diagram (Text-Fig. 11). However, the Group I and Group II basalts show distinct patterns, and the dolerites take a transitional position between these two. The characteristics of each group are reflected by the normalized La/Sm and La/Lu ratios (Table 2).

The positive slope in light REE of the Group II basalts (Text-Fig. 11), although usually more pronounced than in our samples, is typical of N-type MORB. (La/Sm)_{cn} and (La/Lu)_{cn} ratios <1 reflect a mantle source depleted in light REE

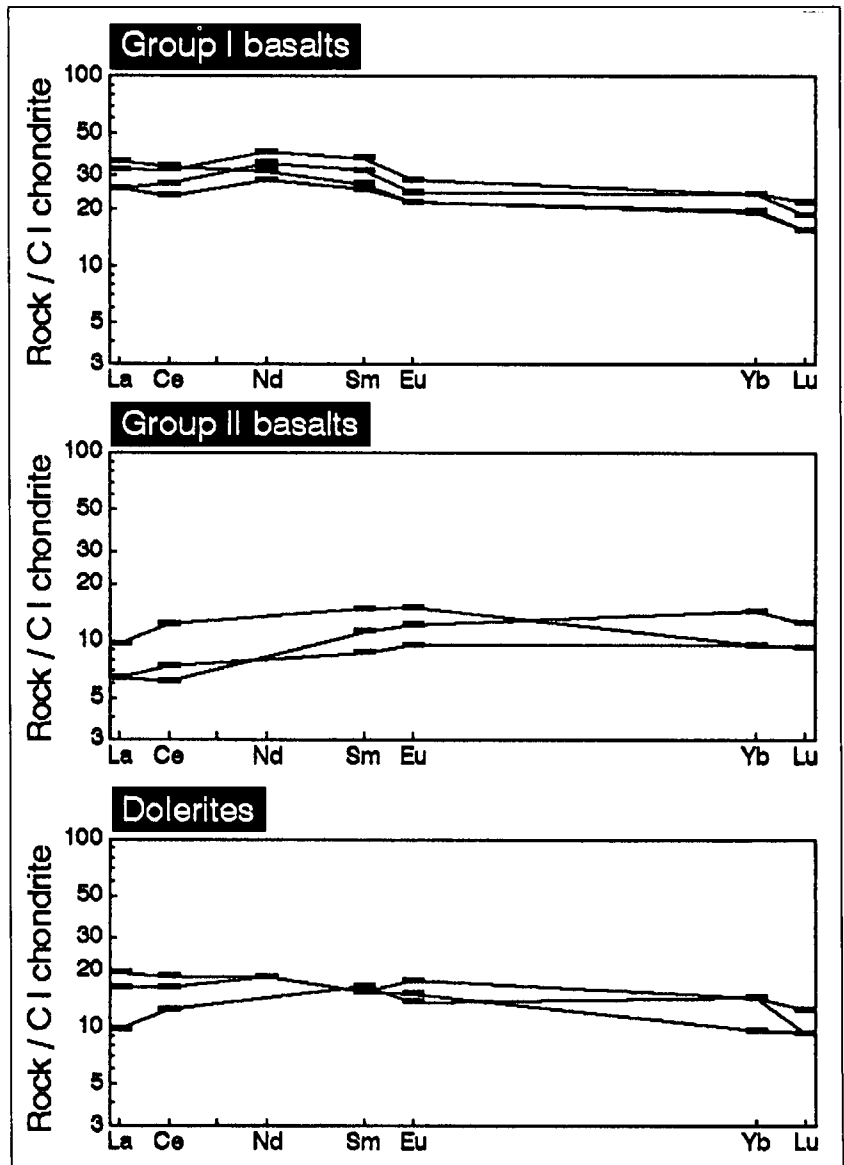
(SAUNDERS, 1984). The curves are compatible with about 20 % melting of a peridotite source (HASKIN, 1984).

The higher (La/Sm)_{cn} and (La/Lu)_{cn} values of the Group I basalts reflect a different, slightly REE enriched mantle source. However, the partly positive slope in the light REE pattern (Text-Fig. 11) indicate a magma relationship to the Group II basalts. The higher REE concentrations in the Group I basalts testify lower amounts of partial melting which may be in the range of 10 per cent, and/or fractional crystallization (HASKIN, 1984). A slight negative Eu anomaly is due to plagioclase fractionation in limited depth. There is no pronounced fractionation of heavy REE.

The rare earth element data indicate heterogeneous mantle composition, but also common features of the mantle sources. This is stressed by the dolerites which take a transitional position. The REE data are in line with the results from the other trace elements.

4. Ultramafic Rocks

The ultramafic rocks of the Platta Nappe and the Arosa Zone include harzburgites and lherzolites serpentinized to various degrees. The abundance of opicalcite has been



Text-Fig. 11.
Chondrite normalized rare earth element diagrams for selected Group I basalt, Group II basalt, and dolerite samples.
Normalization values after HENDERSON (1984).

Table 2.

Sc and rare earth element (REE) concentrations as well as chondrite-normalized (cn) La/Sm and La/Lu ratios, of selected Group I basalt, Group II basalt, and dolerite samples from Arosa Zone and Platta Nappe. Values in ppm.

	Sc	La	Ce	Nd	Sm	Eu	Yb	Lu	(La/Sm) _{cn}	(La/Lu) _{cn}
Group I basalts										
16.1	35.5	8	19	17	5.0	1.6	4	0.5	1.01	1.66
OE 73	26.3	11	27	19	5.3	1.6	4	0.5	1.31	2.28
OE 74	33.5	10	26	24	7.2	2.1	5	0.7	0.87	1.48
G 20 I	32.0	8	22	21	6.2	1.8	5	0.6	0.81	1.39
Group II basalts										
SG 04 aI	39.9	2	5	<10	2.2	0.9	3	0.4	0.57	0.52
G 05	28.7	3	10	<10	2.9	1.1	2	0.3	0.65	1.04
R 202	29.1	2	6	<10	1.7	0.7	2	0.3	0.74	0.69
Dolerites										
15.3	25.7	6	15	11	3.0	1.1	2	0.3	1.26	2.08
17.6	29.6	5	13	11	3.0	1.3	3	0.3	1.05	1.73
25.2	35.9	3	10	<10	3.2	1.0	3	0.4	0.59	0.78

stressed by SCHIDLOWSKI & STAHL (1971), BERNOULLI & WEISSERT (1985), and WEISSERT & BERNOULLI (1985).

4.1. Petrography

The ultramafic rocks are generally serpentinized to a high degree but in some large bodies (e.g., the Todtalp complex near Davos) the original mineralogy is largely preserved. According to GEES (1954), PETERS (1963), and our analyses the original rocks consisted of 50–70 % olivine, 25–30 % orthopyroxene, and 5–20 % clinopyroxene and are therefore classified after STRECKEISEN (1976) as orthopyroxene-rich lherzolites and harzburgites.

Completely serpentinized peridotite is common in highly tectonized zones, for instance where serpentinite forms the matrix of the tectonic mélange (RING et al., 1990). It is generally associated with ophicalcite. X-ray diffraction studies show that north of Bivio the serpentine minerals are chrysotile and lizardite. Antigorite is stable south of Bivio (DIETRICH & PETERS, 1972). The serpentine minerals show a mesh texture (PETERS, 1963) where the rock escaped internal deformation. Finely dispersed magnetite grains derived from the fayalite component of olivine are concentrated in the partings of the mesh cells. Rare augite, enstatite, chromite, and picotite occur as relictic phases.

Around Bivio and in the Todtalp complex decimeter-thick lenses of medium- to coarse-grained peridotite occur within the serpentinites. Partly serpentinized orthopyroxene is arranged in layers and thus reveals a cumulate origin of the rock. PETERS (1963) reported pyrope-rich garnet (76 % pyrope) from these lenses. PETERS & STETTLER (1987) calculated formation pressures for the garnets between 1.0 and 1.6 GPa.

4.2. Geochemistry

The geochemical results of the ultramafic rocks are presented in Table 3, sample localities are shown in Text-Fig. 2. The chemical analyses were recalculated on a water and carbon dioxide free base. Normative corundum in some

samples indicates alteration which is likely to have occurred during sea-floor metasomatism and regional metamorphism. It led to partial removal of Ca from the clinopyroxenes. Loss of Ca was also recognized in the basaltic rocks (see above). Therefore, classification and genetic interpretation of the ultramafic rocks are based on the largely immobile elements rather than on normative composition.

Al and Ti are considered to behave largely immobile during alteration processes (PEARCE, 1976). The $\text{Al}_2\text{O}_3/\text{TiO}_2 - \text{TiO}_2$ relation (BECCALUVA et al., 1983) shows a lherzolitic composition for most of the studied ultramafic rocks (Text-Fig. 8). Although the classification of some samples is ambiguous, two groups are discernible. Those labeled as lherzolite in Text-Fig. 8 are characterized by higher Al_2O_3 (usually >3 wt-%) and TiO_2 (usually >0.10 wt-%) contents than those labeled as harzburgite or suspected harzburgite (see Table 3).

In the diagram of Text-Fig. 8 parental basalts and their ultramafic counterparts are interpreted in terms of generated melts and refractory residua. Transitional mid-ocean ridge/within-plate basalts like the Group I basalts may be derived from less than 15 % equilibrium partial melting of a spinel lherzolite source (BECCALUVA et al., 1983). We interpret the harzburgites and ultramafic cumulates to represent uppermost mantle and lowermost crust and consider the Group II basalts and dolerites to correspond with these counterparts (Text-Fig. 8).

To get a clue for the geotectonic position of the ultramafic rocks, PEARCE et al. (1984) considered their TiO_2 contents, which enables discrimination between ophiolites of supra-subduction zone and mid-ocean ridge environments. The TiO_2 contents of >0.05 wt-% in nearly all samples display a mid-ocean ridge position of the ultrabasic rocks of the Platta Nappe and Arosa Zone and confirm the geochemical results of the basaltic rocks.

5. Discussion and Conclusions

The geochemical patterns discriminate the dolerites and part of pillow basalts (Group II basalts) as mid-ocean

Table 3.
Geochemical results of ultramafic rocks.
Explanation as in Table 1. Nb and Rb <10 ppm, Sr <30 ppm, Zr <20 ppm.

	SiO2	TiO2	Al2O3	Fe2O3	FeO	MnO	MgO	CaO	Na2O	K2O	P2O5	H2O	CO2	Total	Ba	Cr	Ni	V	Y	Zn
Lherzolites																				
2.3.	39,28	0,11	3,33	4,33	3,18	0,14	37,43	< 0,24	< 0,14	< 0,14	< 0,03	12,13	0,30	100,78	< 40	3649	1880	80	8	57
2.30.	39,34	0,11	3,31	4,36	3,20	0,14	37,45	< 0,24	< 0,14	< 0,14	< 0,03	12,19	0,22	100,87	< 40	3576	1890	82	7	59
232	39,55	0,11	3,33	4,44	3,12	0,14	37,58	< 0,24	< 0,14	< 0,14	< 0,03	12,20	0,30	101,32	< 40	3642	1882	81	6	55
239	39,48	0,11	3,35	4,53	3,01	0,15	37,54	< 0,24	< 0,14	< 0,14	< 0,03	12,19	0,22	101,13	< 40	3633	1906	81	6	59
234	39,20	0,11	3,30	4,46	3,09	0,14	37,51	< 0,24	< 0,14	< 0,14	< 0,03	12,20	0,19	100,75	< 40	3631	1880	76	7	58
235	39,29	0,11	3,30	4,60	2,94	0,14	37,46	< 0,24	< 0,14	< 0,14	0,05	12,30	0,14	100,85	< 40	3576	1874	69	6	56
6.3.	41,50	0,15	3,31	4,33	3,51	0,13	33,09	4,48	0,18	< 0,14	< 0,03	10,30	0,17	101,32	< 40	2436	1858	64	9	48
6.30.	41,31	0,15	3,30	4,36	3,47	0,13	32,86	4,48	0,18	< 0,14	0,04	10,37	0,24	101,03	< 40	2441	1850	75	7	42
631	41,58	0,15	3,32	4,35	3,49	0,13	33,09	4,51	0,21	< 0,14	< 0,03	10,19	0,25	101,44	< 40	2423	1848	74	7	43
10.3.	39,40	0,13	3,17	5,22	3,21	0,13	33,56	4,05	0,19	< 0,14	0,07	10,80	0,22	100,29	< 40	2076	1999	71	9	50
10.30.	39,44	0,12	3,18	5,30	3,11	0,13	33,38	4,08	< 0,14	< 0,14	< 0,03	10,71	0,26	100,02	< 40	2078	2033	63	8	45
1031	39,45	0,13	3,16	5,25	3,19	0,13	33,39	4,07	0,14	< 0,14	< 0,03	10,87	0,23	100,18	< 40	2047	2045	66	9	39
1032	39,32	0,14	3,18	5,19	3,17	0,13	33,19	4,00	< 0,14	< 0,14	< 0,03	10,79	0,22	99,64	< 40	2157	2030	78	8	41
1033	39,35	0,13	3,17	5,15	3,20	0,13	33,29	4,04	< 0,14	< 0,14	< 0,03	10,61	0,31	99,69	< 40	2082	2050	66	8	44
10.5.	39,42	0,14	3,26	5,61	2,90	0,13	34,79	1,73	< 0,14	< 0,14	< 0,03	11,51	0,24	100,06	< 40	2470	2132	72	7	51
1050	39,29	0,14	3,27	5,29	3,17	0,13	34,72	1,70	< 0,14	< 0,14	0,06	11,45	0,24	99,74	< 40	2410	2098	59	7	51
1051	39,38	0,14	3,29	5,45	3,07	0,13	34,84	1,71	< 0,14	< 0,14	< 0,03	11,49	0,30	100,11	< 40	2454	2127	73	7	56
1052	39,43	0,14	3,26	5,24	3,22	0,13	34,62	1,71	< 0,14	< 0,14	0,03	11,51	0,41	100,00	< 40	2442	2103	80	8	53
1053	39,15	0,14	3,26	5,58	2,86	0,13	34,82	1,72	< 0,14	< 0,14	< 0,03	11,37	0,28	99,82	< 40	2431	2108	85	8	51
1054	39,26	0,14	3,32	5,48	3,04	0,13	35,26	1,71	< 0,14	< 0,14	< 0,03	11,40	0,30	100,35	< 40	2511	2147	89	7	54
10.9.	39,04	0,12	3,06	4,65	3,61	0,12	34,35	2,68	0,26	< 0,14	0,06	11,12	0,41	99,82	< 40	2265	1971	67	7	49
231	39,08	0,11	3,29	4,43	3,16	0,14	37,15	< 0,24	< 0,14	< 0,14	< 0,03	12,22	0,30	100,43	< 40	3642	1888	84	7	55
G 08	38,60	0,17	3,55	7,13	3,11	0,16	34,03	2,09	< 0,14	< 0,14	< 0,03	10,87	0,57	100,59	< 40	2791	2015	78	6	43
G 25	39,07	0,13	3,22	3,35	4,35	0,14	35,75	1,20	< 0,14	< 0,14	< 0,03	11,83	0,44	99,79	< 40	2674	2091	60	< 5	59
R30	39,11	0,12	2,69	—	7,26 *	0,11	35,73	2,27	< 0,14	< 0,14	0,04	11,98 **	—	99,59	< 40	2175	2075	84	< 5	42
R56	39,05	0,09	2,53	—	7,53 *	0,14	37,05	0,39	< 0,14	< 0,14	< 0,03	12,38 **	—	99,47	63	2742	2271	76	< 5	48
Harzburgites (suspected)																				
114	39,23	< 0,03	1,65	5,13	3,01	0,10	37,86	< 0,24	< 0,14	< 0,14	< 0,03	12,17	0,28	100,01	< 40	2830	2518	51	6	64
1144	39,08	0,06	1,64	3,99	4,04	0,11	37,77	< 0,24	< 0,14	< 0,14	0,05	12,31	0,22	99,79	< 40	2818	2516	57	6	62
11.4.	39,29	< 0,03	1,65	4,57	4,24	0,10	37,90	< 0,24	< 0,14	< 0,14	< 0,03	12,27	0,28	100,88	< 40	2826	2543	46	7	62
G 07	39,50	0,05	1,76	5,30	2,58	0,12	37,62	1,10	< 0,14	< 0,14	< 0,03	12,11	0,44	100,89	< 40	2731	2334	51	< 5	41
R155	38,58	0,06	1,57	—	4,77 *	0,12	38,15	0,37	< 0,14	< 0,14	< 0,03	12,76 *	—	96,69	< 40	1921	2438	33	5	39
Harzburgites																				
10.4.	38,39	< 0,03	0,98	5,27	2,96	0,12	36,74	2,54	< 0,14	< 0,14	< 0,03	13,11	0,31	100,76	< 40	2400	2206	31	7	55
40.1.	38,08	< 0,03	0,98	5,30	2,88	0,12	36,10	2,52	< 0,14	< 0,14	< 0,04	13,21	0,26	99,80	< 40	2393	2184	23	6	55

ridge basalts (MORB) derived from a slow-spreading ridge. A component characteristic of intraplate basalts occurs in several samples, especially the dolerites. The greater part of the studied basalts (Group I basalts) are tholeiites enriched in incompatible elements and correlates with within-plate tholeiites (WPT), E-type MORB, or transitional types. Similar basalts occur at diffuse, slow-spreading ridges like Afar, Iceland, or other anomalous ridge segments of the Atlantic ocean (WOOD, 1976; PEARCE, 1980).

Low spreading rates for the South Penninic ocean are also deduced from geological constraints:

- a) Reconstructions of plate motions (FRISCH, 1979; DERCOURT et al., 1986) establish a limited width of the South Penninic Ocean, from which the Platta and Arosa ophiolites were derived, and allow only bilateral spreading rates of <2 cm/a.
- b) The bilateral spreading rates in the Central Atlantic Ocean between Newfoundland and the Canary Islands were about 2 cm/a in the Jurassic and Early Cretaceous (LE PICHON & FOX, 1971). Since the South Penninic Ocean was part of this early Atlantic spreading system (FRISCH, 1979) and its spreading axis was positioned closer to the pole of rotation, a spreading rate of less than 2 cm/a results.
- c) A limited width of the oceanic basin is corroborated by a high sedimentation rate of clastic sediments (Bündner Schiefer; FRISCH, 1980).
- d) As discussed above, the scarcity of cumulate rocks may also reflect slow spreading.

The enriched tholeiitic basalts may be derived from a subcontinental mantle beneath attenuated crust, where metasomatism was responsible for enrichment of incompatible elements. In view of lack of age constraints for the basaltic protoliths we speculate that the Group I basalts have been produced either in an incipient stage of ocean formation in a transitional within-plate/mid-ocean ridge setting (oceanic/continental lithosphere interface), or at a later stage tapping mantle heterogeneities inherited from the subcontinental era. The N-type MORB was produced during the period of South Penninic ocean floor formation. The mantle source of the dolerites and some of the Group II basalts was influenced by the enriched mantle source by heritage (in time) or neighbourhood (in space). Variations in oceanic tholeiites with different light/heavy REE ratios are generally explained by mantle heterogeneities and mixing. From trace element modelling including REE we conclude that there was also a difference in the amount of partial melting. The elevated contents in REE of the Group I basalts relative to the Group II basalts indicate a lower degree of partial melting, as do the higher Y contents at a given Cr concentration. The estimated amount of partial melting of the mantle source is about 10 % for the Group I basalts and about 15–20 % for the Group II basalts. From all this, we speculate, as a likely scenario, that the Group I basalts are generally older and originated during incipient drifting from a still deeper source. The N-type MORB Group II basalts formed in an advanced stage of ocean floor formation at a shallow level. The enriched magma source contributed to part of these melts which is evidenced by slightly but significantly elevated Mn, P, and Zr concentrations and higher Zr/Y ratios relative to N-type MORB. This points to gradual transitions with decreasing influence of the deeper-seated source to the N-type MORB source.

Mantle heterogeneities are frequent in the Atlantic spreading system and produce wide variations in the composition of mid-ocean ridge basalts (WILSON, 1989). As a general explanation it is suggested that mantle metasomatism was a significant effect in the subcontinental setting prior to continental splitting (SAUNDERS, 1984).

From the types of basaltic rocks we conclude that the South Penninic ocean floor in the region of the Platta Nappe and the Arosa Zone consisted of partly abnormal oceanic crust. Basaltic rocks from other South Penninic ophiolite bodies in the Eastern Alps suggest a slightly enriched (PEARCE, 1980) or slightly depleted mantle source (KOLLER & HÖCK, 1987). The stronger enrichment in incompatible elements of a part of the basalts in the Platta Nappe and Arosa Zone is therefore considered to be due to localized mantle diapirism and metasomatism.

Similar situations are inferred to be responsible for the formation of MORB/WPB transitional basalts that occur in ophiolite complexes elsewhere in the Late Jurassic/Early Cretaceous spreading system of the Alpine Mediterranean orogenic belt (PEARCE, 1980).

Phlogopite from ultramafic rock yielded a $^{39}\text{Ar}/^{40}\text{Ar}$ age of 160 ± 8 Ma, which is interpreted as the time of protolith formation (PETERS & STETTLER, 1987). This Middle/Late Jurassic age approximately coincides with other estimates of continental breakup for the South Penninic ocean (e.g., DERCOURT et al., 1986). Therefore, we infer that the age of the Group I basalts, which are considered to have formed shortly after the breakup event of the Penninic ocean, is around early Late Jurassic. The harzburgitic ultramafic rocks, the cumulate rocks, and the gabbros are considered to represent oceanic uppermost mantle and lower crust and to correlate with the N-type MORB (Group II basalts) erupted in an advanced stage of spreading.

Lherzolitic protoliths of Penninic ultramafic rocks are reported by PETERS (1963) from the Arosa Zone and VUICHARD (1984) from the nearby Unterengadin Window. KOLLER (1985) described cumulate gabbro, harzburgite, and lherzolite from the Rechnitz Window, HÖCK (1983) probable cumulate gabbro and ultramafic rocks from the Tauern Window, and HÖCK & KOLLER (1987) harzburgite and gabbro from the Unterengadin Window.

North Penninic ophiolitic greenstones are similar to our Group II basalts and dolerites (DÜRR et al., 1993). We interpret this in terms of a common mantle source. The North Penninic ocean is considered to have formed later than the South Penninic ocean so that the early, enriched types do not occur there. A direct connection between North and South Penninic oceans in the meridian of eastern Switzerland has been proposed by several authors (see TRÜMPY, 1992).

The paleogeographic situation of the South Penninic ocean in the Late Jurassic and Early Cretaceous is that of a narrow appendix of the early Central Atlantic ocean, transferred by transform faults and forming a connection to the Tethys ocean in the east (FRISCH, 1979; DERCOURT et al., 1986). The topography of the oceanic basement is suggested to have been rugged and complex in that it was produced by slow spreading and dominated by transform offsets. BERNOULLI & WEISSERT (1985) and WEISSERT & BERNOULLI (1985) interpreted the frequent occurrence of ophicalcite by the presence of a transform fracture system. Serpentinized mantle rocks extruded as diapirs using the fractures as pathways to the surface and were topped by and mixed with calcareous and siliceous ooze. The

narrow basin was quickly filled by clastic sediment shed through localized channels and distributed by longitudinal currents (FRISCH, 1980).

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