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Isotopic (Rb–Sr and Sm–Nd) and geochemical studies of the metamorphosed volcanic rocks in the metamorphic series below the Semail ophiolite in the Dibba zone (United Arab Emirates)

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Abstract

Major and trace element geochemistry of the metavolcanic rocks occurring in a metavolcanic unit within the metamorphic series below the Semail ophiolite in the Dibba region of the United Arab Emirates (UAE) has been conducted. Analysis show that volcanism in this area was essentially bimodal alkalic to tholeiitic. The alkaline basaltic meta-pillow lavas, meta-hyaloclastites, meta-pillow breccias, and meta-tuffites show an intimate association with meta-carbonatites and deep sea sediments such as meta-radiolarites, and meta-siliceous carbonates. These are thought to be equivalents of the non-metamorphic Haybi/Hawasina volcanics underlying the Semail nappe in various areas along the Oman mountains. Sm–Nd and Rb–Sr isotopic investigations of the metavolcanics show that they are derived from a low ε_{Nd} mantle source, with an isotopic composition different from that of the mantle sequence rocks of the Semail ophiolite. A comparison with data from the meta-carbonatites shows that the Nd isotopic composition was not severly affected by post extrusive alteration processes. In contrast the Sr isotopic composition was probably slightly altered by such processes. Analysis of a serpentinite and a partly serpentinised harzburgite of the Semail sequence show that their Nd and Sr isotopic composition was strongly disturbed by the serpentinisation process.

Keywords: Oman mountains, Semail ophiolite, carbonatites, alkaline volcanics, Sm–Nd, Rb–Sr, metamorphic series.

1. Introduction

Geochemical investigations on the unmetamorphosed Haybi and Hawasina volcanics underlying the Semail ophiolite nappe in the Oman mountains have been conducted by SEARLE (1980) and BECHENNEC (1987). Their data indicate a succession of alkaline within plate basalts, tholeiitic mid ocean ridge basalts, island arc tholeiites and finally alkalic intraplate volcanics. K–Ar age data and biostratigraphic studies (SEARLE, 1980; BECHENNEC, 1987) on part of the Haybi and Hawasina volcanics and interbedded sediments suggest that initial alkalic volcanic activity occurred during the Permo-Triassic. Later, island arc type tholeiitic and alkaline intraplate magmatism took place during the Lower to Middle Cretaceous.

Earlier studies of ZIEGLER et al. (1990) identify the occurrence of carbonatites in association with alkalic basalts in the metamorphic series directly underlying the Semail ophiolite nappe in the Asimah window (UAE).

The purpose of the present study is to use major and trace element geochemistry to elucidate the connection between the Haybi and Hawasina volcanics and the volcanics of the meta-alkaline volcanic unit in the Asimah window. Rb–Sr and Sm–Nd isotopes of the metavolcanic rocks of the Asimah window could provide information on the possibility of a mantle source of the alkaline volcanic rocks, and their correlation with the carbonatites in the Asimah window and also the overlying mantle sequence rocks of the Semail nappe.

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2. Geological setting

The Oman mountains are located on the eastern margin of the Arabian Platform (Fig. 1) and consist of a series of allochthonous thrust sheets which are overlain by a large obducted ophiolite slab. This whole sequence was emplaced onto the passive continental margin during the late Cretaceous (GLENNIE et al., 1973, 1974). The allochthonous units underlying the sequence of Cretaceous oceanic lithosphere of the Semail ophiolite nappe (> 10 km thick) consist mostly of Mesozoic Tethyan oceanic sediments and volcanic rocks. From top to bottom they can be broadly subdivided into four major units:

1) the sub-ophiolitic metamorphic sheets,
2) the Permian to Triassic exotic limestones and alkalic to tholeiitic basalt melanges of the Haybi complex,
3) the distal slope and deep-sea oceanic sediments of the Hawasina complex,
4) the shelf edge and proximal slope sediments of the Sumeini Group.

Metamorphic rocks underly the Semail ophiolite complex of the Oman mountains on its whole length. The metamorphic series is separated by distinct thrust planes from the Semail ophiolite (Semail thrust plane) and the Haybi complex. The thickness of the metamorphic series which occurs as either imbricated slices in the Semail ophiolite masses (e.g. Bulaydah window, UAE) or as sheets that have been thrusted onto the Haybi complex (e.g. Haybi corridor, Oman) varies between a few meters and up to 7.5 km. However, the metamorphics often directly overlie Hawasina sediments (i.e. in the Wadi Kub area in the UAE). The sub-ophiolitic metamorphic series consists of several sheets of metamorphic rocks of both sedimentary and volcanic origin. Rocks similar to those of the Hawasina and Haybi complex are thought to have been the protoliths of the metamorphic series, consisting mainly of quartz-rich mica schists, metavolcanics and lesser amounts of carbonate rocks (ALLEMANN and PETERS, 1972; GLENNIE et al., 1974; SEARLE et al., 1980; ZIEGLER and STOESSEL, 1985). The maximum degree of metamorphism varies from granulite facies conditions (e.g. Bulaydah window, UAE; ZIEGLER and STOESSEL, 1985) to greenschist facies conditions (e.g. the northern part of the Asimah window, UAE) depending on the structural position of the individual thrust sheets.

Alkaline volcanic rocks are known from two units of the Oman mountains. First, SEARLE (1980) has reported such rocks from the Haybi complex which corresponds to parts of the Ha-

wasina series of BECHENNEC (1987). Second, alkaline volcanics are known from the metavolcanic unit in the eastern part of the Asimah window between Wadi Shimal and Wadi Uyaynah (ZIEGLER et al., 1990). Age and geochemistry of this Haybi/Hawasina volcanism have been investigated by SEARLE (1980) and BECHENNEC (1987). Their K–Ar and biostratigraphic dating reveals that the lower part of the Haybi/Hawasina volcanics is Upper Permian to Triassic. Radiolaria interbedded with the depleted tholeiites reveals a Cretaceous origin for the uppermost part of the volcanic sequence. ROBERTSON et al. (1990) propose an initial rifting phase during the Lower to Middle Permian for the Dibba zone. BECHENNEC (1987) concludes that initial rifting and formation of the Hamrat Duru basin (along the passive continental margin of Arabia) took place during the Upper Permian, accompanied by the extrusion of mainly alkalic and later MORB-type tholeiitic volcanics. Bechennec also concludes that the internal morphostructures of the Hawasina basin were formed by rifting processes and associated Horst-Graben structures during the Middle to Upper Triassic. This phase was accompanied by basaltic, andesitic and strongly alkaline volcanism in the Umar basin. In the view of BECHENNEC (1987) the existence of continental crust as substratum of the Hawasina ocean is supported by the limited occurrence of alkaline and hyperalkaline magmatism during the Jurassic and Cretaceous.

The metavolcanic sequence between Wadi Uyaynah and Wadi Shimal is separated from the rest of the metamorphics by a well defined slightly eastward dipping thrust plane which is marked by tectonic melange. Partly metamorphosed porphyritic pillow lavas, pillow breccias, hyaloclastites and tuffites with partially to well preserved original volcanic textures alternate from meters to a few tens of meters with red and green quartzites (metaradiolarites) and lesser amounts of marly metacarbonates. Some apatite rich carbonatite layers and lenses of both extrusive and intrusive origin occur in close association with the metavolcanic rocks of the sequence. Due to the strongly tuffaceous nature of a large part of the metavolcanic rock sequence the metavolcanic rocks represent a mechanical mixture of carbonatite and silicate components. Field relations within the metavolcanic series are difficult to determine because of the strong deformation which occurred during metamorphism and ophiolite obduction. Locally individual pillows were elongated by a factor of 10 along their a and b axes, thus top-bottom orientation of the volcanic sequence based on the orientation of pillows was not possi-
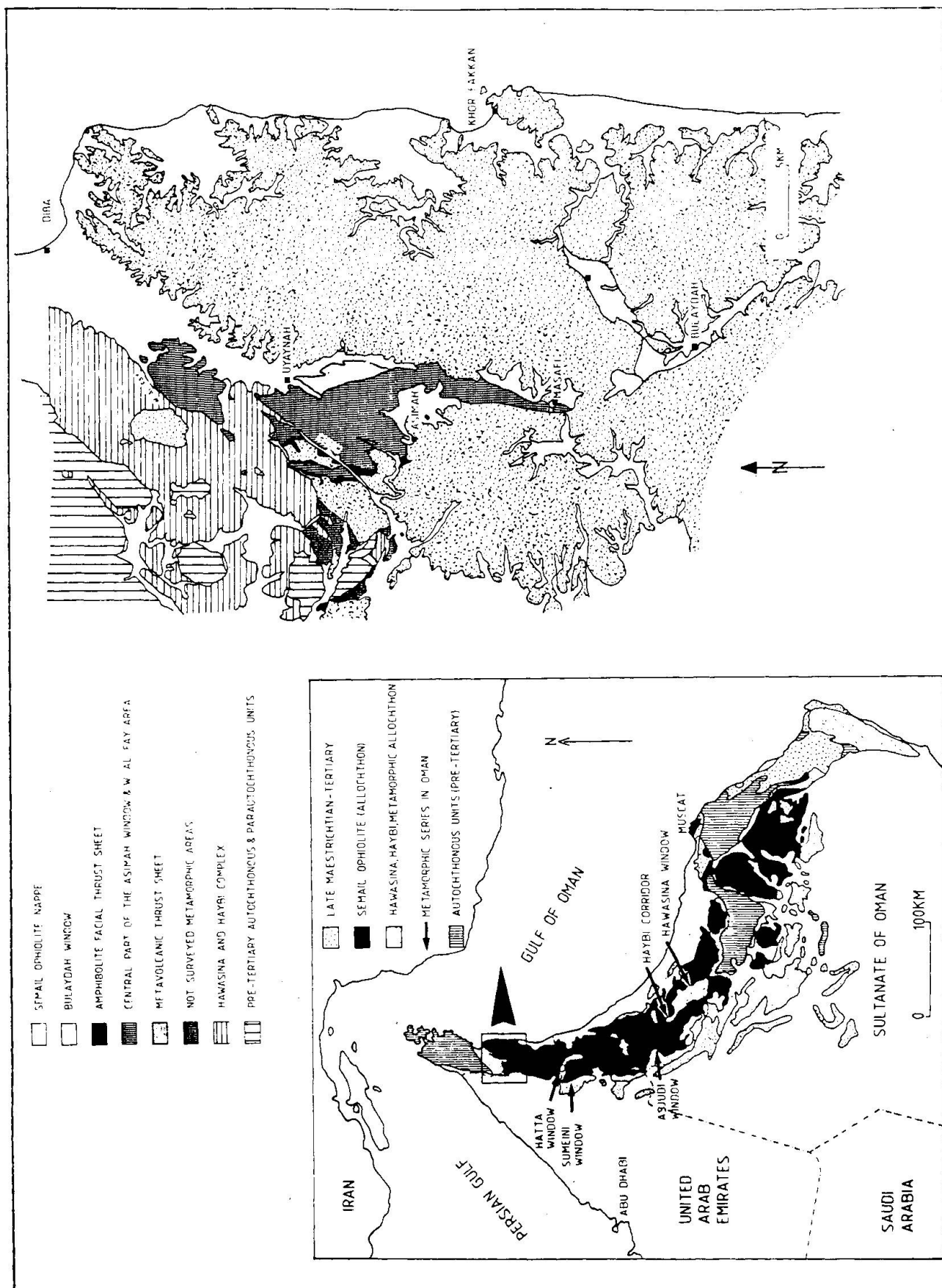
Fig. 1 Geological map of the Oman mountains and the area under investigation (after GLENNIE et al., 1974).

Tab. 1　Major (wt%) and trace element (ppm) analyses of metavolcanics from the meta-alkaline volcanic unit of the Asimah window (numbers in brackets denote detection limits due to count statistics, b.d. = below detection limit; samples labeled * have alkalic trace element characteristics while samples labeled ** are tholeiitic and *** denotes a carbonatitic sample).

Sample	** UAE4706	** UAE4707	** UAE4709	** UAE4710	* UAE4715	* UAE4719	** UAE4723	* UAE4724	* UAE4731	* UAE4732	* UAE4735	* UAE4737	* UAE4740
Major elements:													
SiO_2	38.24	45.87	46.38	41.59	36.48	42.91	42.39	44.64	39.99	43.73	42.95	30.89	48.56
TiO_2	1.15	1.06	1.09	0.95	2.89	3.01	1.11	3.17	5.54	2.57	4.93	2.44	3.61
Al_2O_3	13.43	15.39	15.99	14.60	8.68	11.69	14.98	11.27	15.72	12.20	16.15	8.27	11.27
Fe_2O_3	3.87	4.77	8.61	10.10	3.93	7.93	7.88	4.07	13.52	4.02	15.15	5.99	6.97
FeO	2.87	4.01	3.10	2.73	4.81	3.80	2.09	5.14	3.74	4.87	1.75	3.40	3.78
MnO	0.23	0.17	0.15	0.15	0.15	0.27	0.18	0.15	0.27	0.18	0.13	0.18	0.12
MgO	5.16	8.18	7.17	3.89	6.95	4.06	4.88	7.93	3.70	7.59	3.51	6.79	5.77
CaO	19.23	9.66	5.08	16.11	18.18	13.49	15.75	11.12	6.46	10.73	4.96	22.68	10.30
Na_2O	1.67	2.55	5.91	2.56	3.50	4.61	2.17	4.17	3.72	3.08	1.73	1.71	5.20
K_2O	0.45	0.06	0.07	0.05	0.63	0.74	0.04	0.66	2.02	0.07	5.25	1.09	0.40
P_2O_5	0.13	0.10	0.07	0.12	0.89	1.57	0.14	0.73	2.40	0.58	0.95	0.59	0.85
H_2O	3.36	4.74	3.41	2.73	2.72	1.60	3.22	3.03	3.20	4.35	2.95	2.61	1.07
CO_2	8.40	3.40	3.00	4.70	10.60	4.80	5.30	4.10	0.30	6.50	0.30	13.90	2.70
Cr_2O_3	0.05	0.06	0.08	0.06	0.02	0.02	0.06	0.03	0.02	0.03	0.03	0.04	0.02
NiO	0.02	0.02	0.04	0.04	0.02	0.01	0.02	0.02	0.02	0.02	0.02	0.02	0.01
Sum:	100.26	100.04	100.15	100.38	100.45	100.51	100.21	100.23	100.62	100.52	100.76	100.60	100.63
Trace elements:													
Nb (3)	8	7	b.d.	6	64	193	11	116	153	150	115	73	87
Zr (4)	90	83	79	72	226	623	94	279	489	513	426	228	317
Y (4)	40	32	39	37	31	72	36	36	90	41	59	36	41
Sr (4)	323	127	41	126	559	729	391	790	353	630	118	452	380
U (7)	7	b.d.	7	8	11	19	9	b.d.	8	13	13	16	10
Rb (4)	14	8	7	9	21	30	8	43	54	8	139	33	15
Th (6)	b.d.	b.d.	b.d.	b.d.	b.d.	19	b.d.	b.d.	b.d.	b.d.	9	b.d.	6
Pb (6)	b.d.	b.d.	b.d.	b.d.	7	10	b.d.	b.d.	8	17	8	b.d.	10
Ga (2)	13	13	13	11	11	18	15	15	27	19	29	12	14
Zn (5)	97	92	116	102	86	212	76	89	252	152	227	114	98
Cu (7)	59	220	216	90	45	8	56	147	62	30	63	55	44
Ni (7)	114	146	378	399	96	24	152	100	74	68	157	137	66
Co (6)	57	56	70	73	60	47	53	60	80	48	67	65	49
Cr (10)	292	411	623	483	94	22	417	118	45	145	127	285	31
V (3)	280	221	190	205	360	252	325	418	396	247	122	254	310
Ce (23)	b.d.	23	b.d.	28	161	489	28	186	468	292	274	127	147
Nd (12)	b.d.	b.d.	b.d.	b.d.	72	240	b.d.	64	284	113	123	46	65
Ba (2)	60	16	20	21	570	238	50	588	474	49	1086	308	193
La (12)	23	23	45	17	144	278	33	176	390	195	235	115	167
Sc (1)	32	33	29	30	25	16	43	27	45	17	35	29	22

ble. Large scale progressive interlayering of hyaloclastites, breccias, pillow lavas and tuffites could not be detected. Therefore the stratigraphy of the volcanic succession could not be determined.

Some original magmatic textures of the metavolcanics were observed and were not completely overprinted by metamorphic textures. Relic magmatic minerals such as clinopyroxene, plagioclase and amphiboles have been preserved in rare cases. A detailed description of the petrography of the metavolcanic rocks can be found in ZIEGLER and STOESSEL (1985). To summarize, the metavolcanic rocks show a large variety in mineralogical composition due to their different primary chemical composition and due to regional greenschist

Tab. 1   (cont.)

Sample	* UAE4741	* UAE4745	* UAE4749	* UAE4750	*** UAE4752	** UAE4753	** UAE4754	** UAE4755	* UAE4756	* UAE4760	* UAE4762	* UAE4775	** UAE4779
Major elements:													
SiO_2	36.78	42.29	42.01	23.46	8.48	49.59	45.02	42.72	30.00	39.85	23.78	32.57	39.51
TiO_2	3.45	3.31	3.77	3.01	1.14	1.57	2.22	1.14	1.98	3.11	2.18	2.23	1.05
Al_2O_3	13.60	13.56	15.80	7.90	2.30	17.33	13.32	18.49	6.24	10.04	7.28	7.72	13.73
Fe_2O_3	7.00	4.57	4.93	6.73	11.98	4.53	9.99	6.88	6.14	6.03	4.78	8.34	5.13
FeO	5.24	7.52	8.36	2.88	2.09	4.04	1.92	2.52	4.13	5.79	3.01	1.75	3.99
MnO	0.31	0.19	0.23	0.26	0.68	0.15	0.25	0.18	0.35	0.19	0.14	0.19	0.21
MgO	6.46	7.45	5.99	4.41	6.55	5.90	5.83	5.58	8.10	9.50	4.95	3.45	6.00
CaO	15.80	9.96	9.84	27.00	33.15	9.49	8.80	11.31	21.82	14.66	28.97	22.98	17.72
Na_2O	1.34	2.92	2.75	2.13	0.53	4.69	2.43	4.33	0.22	2.60	2.50	3.62	2.90
K_2O	0.34	2.32	2.49	0.21	0.20	0.13	3.81	0.24	0.00	1.27	0.10	0.31	0.07
P_2O_5	0.93	0.87	1.08	1.41	8.51	0.19	0.25	0.11	0.46	0.99	0.64	0.58	0.11
H_2O	4.13	3.45	2.84	2.70	2.72	1.47	2.70	3.26	4.46	2.90	2.02	1.84	3.09
CO_2	5.30	2.00	0.40	18.00	20.70	0.30	3.70	3.50	16.60	3.90	20.20	15.10	7.20
Cr_2O_3	0.04	0.05	0.03	0.01	0.01	0.04	0.04	0.05	0.10	0.05	0.02	0.03	0.04
NiO	0.02	0.02	0.02	0.00	0.01	0.02	0.01	0.02	0.05	0.04	0.01	0.01	0.02
Sum:	100.74	100.48	100.54	100.11	99.05	99.44	100.29	100.33	100.65	100.92	100.58	100.72	100.31

Trace elements:

		* UAE4741	* UAE4745	* UAE4749	* UAE4750	*** UAE4752	** UAE4753	** UAE4754	** UAE4755	* UAE4756	* UAE4760	* UAE4762	* UAE4775	** UAE4779
Nb	(3)	130	128	151	89	331	10	17	7	51	94	72	70	9
Zr	(4)	378	336	398	322	1296	135	174	91	173	402	233	246	71
Y	(4)	47	43	48	42	163	42	55	32	29	43	32	35	40
Sr	(4)	619	785	1157	942	1973	180	64	183	180	1064	716	389	121
U	(7)	12	12	12	9	22	9	8	8	12	14	12	11	b.d.
Rb	(4)	15	81	83	12	9	9	57	11	11	44	11	14	8
Th	(6)	10	7	8	b.d.	28	b.d.	b.d.	b.d.	b.d.	8	b.d.	b.d.	b.d.
Pb	(6)	10	10	10	b.d.	b.d.	b.d.	b.d.	b.d,	b.d.	b.d.	b.d.	b.d.	b.d.
Ga	(2)	20	20	21	12	8	16	19	18	10	18	10	10	14
Zn	(5)	163	133	153	153	353	80	106	97	113	142	106	97	88
Cu	(7)	155	91	113	29	b.d.	48	24	45	56	51	49	27	49
Ni	(7)	112	135	83	33	52	74	66	107	551	322	88	90	114
Co	(6)	72	70	71	57	72	50	54	52	90	75	61	56	60
Cr	(10)	314	361	141	b.d.	53	260	184	365	1086	377	91	199	283
V	(3)	426	400	446	207	117	318	484	204	285	300	196	159	291
Ce	(23)	203	191	320	179	1535	24	26	b.d.	161	160	85	125	b.d.
Nd	(12)	101	94	125	81	836	b.d.	b.d.	b.d.	42	71	16	46	b.d.
Ba	(2)	352	774	1763	187	211	67	118	40	41	432	113	226	32
La	(12)	206	193	233	158	566	55	95	22	102	162	106	114	22
Sc	(1)	26	36	21	22	21	41	53	17	32	20	23	24	47

metamorphism. However, they consist mostly of plagioclase (albite to oligoclase), chlorite, amphiboles, epidote minerals, ore grains and sphene. Relic brown kaersutitic hornblendes (which abundantly occur in the alkalic metabasalts) often show blue rims of sodic calcic to alkali amphibole composition. Greenish biotite becomes a major constituent in the southern part of the metavolcanic unit. The hyaloclastitic matrix is often found in association with pillows and consists of irregularly shaped brown glass shards (usually com-pletely recrystallised to very fine grained aggregates of chlorite, epidote minerals, sphene and ore grains) in a ground mass of chlorite, biotite, epidote minerals, albite, sphene, hornblende, calcite and ore grains. The spherulitic rims of pillows are characterized by radiate aggregates of fibrous plagioclase and/or hornblende needles. Most cores of the pillows have a porphyritic and often seriate texture. Plagioclase and less commonly enstatite rich diopsides are surrounded by a very fine grained groundmass. Larger calcite aggre-

Tab. 1 (cont.)

Sample	UAE*4781	UAE*4782	UAE*4784	UAE*4785	UAE**4787	UAE*4788	UAE*4789	UAE*4791
Major elements:								
SiO_2	41.57	36.82	45.68	39.48	58.38	43.50	41.18	31.34
TiO_2	3.91	3.18	3.84	4.16	0.70	1.91	3.73	3.35
Al_2O_3	14.57	11.29	16.75	12.46	15.04	11.17	15.15	9.85
Fe_2O_3	3.66	3.49	7.33	9.20	5.29	5.21	5.60	3.75
FeO	9.83	8.52	1.41	5.27	2.55	5.03	8.04	6.63
MnO	0.20	0.25	0.11	0.20	0.20	0.17	0.24	0.17
MgO	6.45	10.39	1.77	8.62	3.28	6.94	7.66	7.90
CaO	7.78	11.71	16.68	14.63	5.71	15.55	10.95	19.90
Na_2O	3.85	0.92	3.43	1.40	7.08	3.09	2.40	1.71
K_2O	1.34	0.74	0.25	1.08	0.64	0.47	0.07	1.20
P_2O_5	0.95	0.68	0.85	0.87	0.13	0.62	0.98	0.75
H_2O	4.05	5.93	1.07	3.03	0.56	2.07	4.30	3.69
CO_2	2.30	6.70	0.80	0.50	0.50	4.50	0.30	10.20
Cr_2O_3	0.03	0.08	0.05	0.09	0.03	0.05	0.05	0.05
NiO	0.02	0.03	0.01	0.02	0.02	0.03	0.02	0.02
Sum:	100.51	100.73	100.03	101.01	100.11	100.31	100.67	100.60

Trace elements:

		UAE*4781	UAE*4782	UAE*4784	UAE*4785	UAE**4787	UAE*4788	UAE*4789	UAE*4791
Nb	(3)	136	106	154	83	16	52	151	57
Zr	(4)	368	298	398	279	166	214	376	219
Y	(4)	47	39	46	42	41	31	47	31
Sr	(4)	385	453	1897	615	714	550	646	756
U	(7)	8	8	16	13	13	10	12	13
Rb	(4)	58	37	12	36	14	16	9	40
Th	(6)	6	b.d.	11	b.d.	6	b.d.	8	b.d.
Pb	(6)	b.d.	8	34	8	23	8	13	b.d.
Ga	(2)	22	21	27	22	16	16	25	15
Zn	(5)	148	144	36	128	99	109	154	114
Cu	(7)	116	78	11	74	204	61	111	83
Ni	(7)	97	225	44	199	108	227	127	207
Co	(6)	77	81	39	88	32	60	78	82
Cr	(10)	181	696	344	673	122	371	341	410
V	(3)	414	303	679	669	175	277	540	478
Ce	(23)	296	184	216	165	57	113	310	170
Nd	(12)	141	91	108	105	31	40	121	77
Ba	t2)	451	200	261	601	129	564	116	502
La	(12)	227	151	237	219	57	101	252	171
Sc	(1)	33	50	24	49	19	28	33	39

gates with straight grain boundaries are observed and represent calcite pseudomorphs after olivine. Original vesicules and amygdales of the pillows are now filled with either albite, calcite, greenish biotite, chlorite or an aggregate of epidote minerals.

3. Samples and analytical procedure

The eastern part of the Asimah window was mapped on a 1:25000 scale in 1983/84 and 1989/90.

A total of 53 samples of metavolcanic rocks (weights ranging from 2 to 25 kgs) were collected throughout the meta-alkaline volcanic rock unit of the Asimah window. In addition a sample of a serpentinite (OE666) was collected from a serpentinite lens (probably derived from the Semail ophiolite) within the greenschist facies metamorphics of the Asimah window. The partly serpentinised harzburgitic sample OE749 was collected in this part of the basal Semail nappe which directly overlies the metavolcanic thrust sheet. The geochemistry and isotope systematics of the car-

Tab. 2 Tabulation of the Rb and Sr results. Constants used are: ^{87}Rb decay constant = $1.42*^{-11}a^{-1}$ (Steiger and Jäger, 1977), ^{85}Rb/^{87}Rb = 2.59265, ^{86}Sr/^{88}Sr = 0.1194, ^{84}Sr/^{86}Sr = 0.056584 (Steiger and Jäger, 1977), ^{87}Sr/^{86}Sr$_{UR}$ = 0.7045 (e.g. De Paolo, 1988), ^{87}Rb/^{86}Sr$_{UR}^{87}$ = 0.0827 (e.g. De Paolo, 1988).

Sample	Rb ppm	Sr ppm	^{87}Sr/^{86}Sr	±	^{87}Rb/^{86}Sr	±	Epsilon Sr(0)	Epsilon Sr(90)	Epsilon Sr(250)	Epsilon Sr(340)
Metavolcanics										
UAE4707	0.71	123.31	0.70624	8	0.0166	0.0002	24.70	25.90	28.05	29.26
UAE4715	10.77	563.16	0.70473	5	0.0553	0.0006	3.26	3.80	4.69	5.15
UAE4723	0.97	385.45	0.70588	1	0.0073	0.0001	19.59	20.89	23.33	24.87
UAE4735	126.73	107.62	0.70902	1	0.0341	0.0003	64.16	3.84	-103.68	-164.32
UAE4737	24.75	442.60	0.70515	1	0.1618	0.0016	9.23	7.83	5.28	3.80
UAE4749	77.32	1177.32	0.70450	0.1	0.1900	0.0019	0.00	-1.96	-5.43	-7.37
UAE4750	4.30	958.86	0.70496	5	0.0130	0.0001	6.53	7.75	10.01	11.33
UAE4752	1.35	2085.51	0.70423	1	0.0019	0.0001	-3.83	-2.44	0.18	1.72
UAE4755	3.02	17.95	0.70561	0.1	0.4875	0.0049	15.76	8.41	-4.68	-12.06
UAE4762	1.50	713.61	0.70467	3	0.0061	0.0001	2.41	3.82	6.30	7.68
UAE4789	0.45	633.23	0.70573	0.1	0.0020	0.0001	17.46	18.87	21.48	23.01
UAE4791	32.52	766.11	0.70446	1	0.1228	0.0012	-0.57	-1.25	-2.55	-3.32
OE748	45.08	811.78	0.70414	3	0.1606	0.0016	-5.11	-6.35	-9.05	-10.47
Semail ophiolite										
OE666	0.30	5.32	0.70792	13	0.1623	0.0016	48.55	47.11		
OE749	0.91	3.96	0.71089	9	0.6646	0.0066	90.70	80.15		

Tab. 3 Tabulation of the Sm and Nd results. Constants used are: ^{147}Sm decay constant = 0.00654 b.y.$^{-1}$ (Lugmair and Marti, 1978), ^{143}Nd/^{144}Nd$_{CHUR(0)}$ = 0.51264, ^{143}Nd/^{144}Nd$_{DM(0)}$ = 0.51316, ^{147}Sm/^{144}Nd$_{CHUR}$ = 0.1967, ^{147}Sm/^{144}Nd$_{DM}$ = 0.2136.

Sample	Sm ppm	Nd ppm	^{143}Nd/^{144}Nd	±	^{147}Sm/^{144}Nd	±	Epsilon Nd(0)	Epsilon Nd(90)	Epsilon Nd(250)	Epsilon Nd(340)
Metavolcanics										
UAE4707	2.32	7.25	0.512901	8	0.1936	0.0019	5.01	5.05	5.11	5.15
UAE4715	10.19	59.08	0.512718	4	0.1043	0.0010	1.44	2.51	4.40	5.46
UAE4723	3.05	13.11	0.512818	6	0.1406	0.0014	3.39	4.04	5.19	5.83
UAE4735	14.61	93.17	0.512730	5	0.0948	0.0009	1.68	2.85	4.93	6.11
UAE4737	8.54	52.26	0.512737	8	0.0988	0.0010	1.81	2.94	4.94	6.07
UAE4749	14.71	86.09	0.512806	15	0.1033	0.0010	3.16	4.23	6.15	7.22
UAE4750	16.39	93.47	0.512691	3	0.1060	0.0011	0.92	1.96	3.81	4.86
UAE4752	95.70	556.22	0.512709	4	0.1040	0.0010	1.27	2.33	4.23	5.30
UAE4755	2.54	7.84	0.512941	4	0.1956	0.0020	5.79	5.81	5.83	5.85
UAE4762	7.53	42.73	0.512796	4	0.1066	0.0011	2.97	4.00	5.85	6.88
UAE4789	13.44	77.78	0.512731	5	0.1045	0.0010	1.70	2.76	4.64	5.71
UAE4791	10.64	61.78	0.512715	3	0.1041	0.0010	1.38	2.45	4.34	5.41
OE748	10.67	60.61	0.512738	3	0.1064	0.0011	1.83	2.87	4.72	5.76
Semail ophiolite										
OE666	0.38	0.50	0.512197	36	0.4587	0.0046	-8.72	-11.73		
OE749	0.05	1.65	0.512287	8	0.0181	0.0002	-6.96	-4.91		

bonatitic rocks of this area have been described in detail elsewhere (Ziegler et al., 1990).

For this study the major and trace elements of 34 metavolcanic rocks have been determined by X-ray fluorescence analyses. Major elements were determined from glass beads which were fused from ignited powders plus $Li_2B_4O_7$ (1:5 ratio) in gold-platinum crucibles while trace ele-
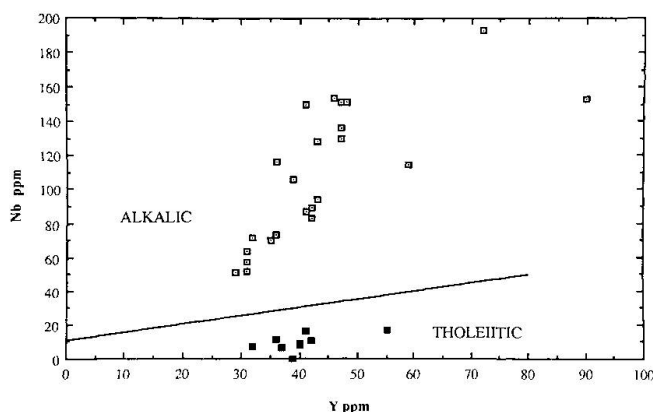
Fig. 2 Plot of the Y–Nb ratios after PEARCE and CANN (1973), and FIELDS and ELLIOTT (1973). (Open squares represent alkalic samples while solid squares represent tholeiitic samples.)
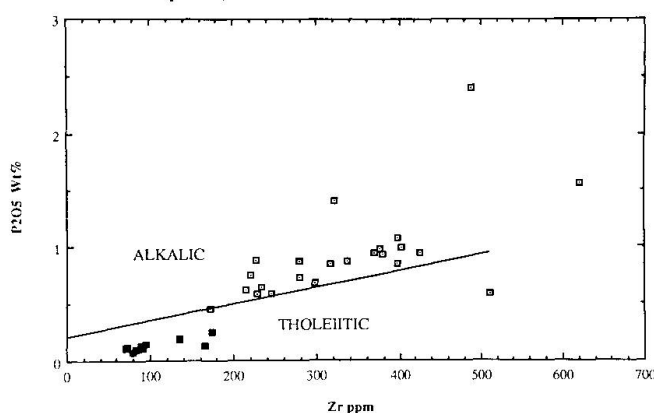

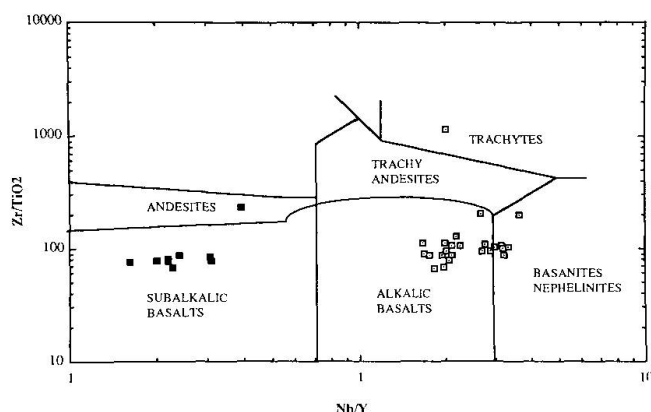*Fig. 3* P_2O_5 versus Zr diagram after ELLIOTT (1973). (Same notation as in Fig. 2.)

Fig. 4 Nb/Y versus Zr/TiO_2 diagram of WINCHESTER and FLOYD (1977) for the analysed metavolcanics. (Same notation as in Fig. 2).

Fig. 5 Ti/100, Zr, 3Y diagram after PEARCE and CANN (1973). A = Ocean Floor Basalts, B = Low-K Tholeiites, C = Calc-Alkaline Basalts, D = Within Plate Basalts. (Same notation as in Fig. 2.)
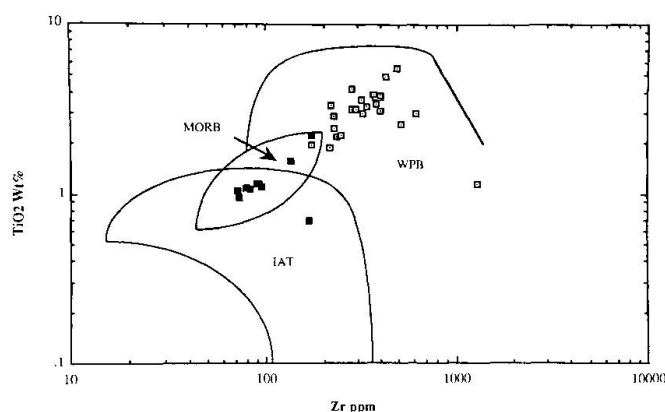
Fig. 6 TiO_2 versus Zr diagram after PEARCE and NORRY (1979). WPB = Within Plate Basalts, MORB = Mid Ocean Ridge Basalts, IAT = Island Arc Tholeiites. (Same notation as in Fig. 2.)
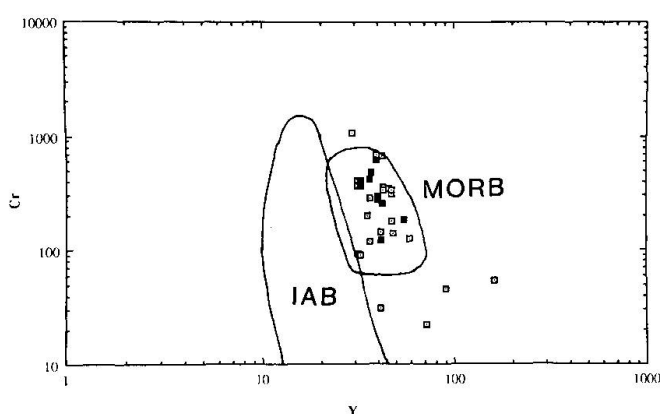

Fig. 7 Y versus Cr diagram after PEARCE (1980). (Same notation as in Fig. 2.)

ment concentrations were determined on pressed powder tablets. The XRF measurements were performed on an automated Philips PW1400 spectrometer (University of Fribourg). The resulting data were corrected for drift, background and matrix. Twelve USGS rock samples were used as standards for calibration. The FeO content of the samples was determined colorimetrically. CO_2 contents were determined on a Coulo-

mat by coulometric alkalimetric titration (Geology Department of the University of Bern). H_2O contents were calculated from loss on ignition in conjunction with the previously determined CO_2 concentrations. Trace element determinations are

based on the synthetic background method in which major element concentrations are known. The trace element concentrations were then calculated according to the method described by Nisbet et al. (1979) (including the calculation of background, interference and mass absorption effects). For an estimate of the resulting accuracies we refer to the considerations of Reusser (1987). Major and trace element data are compiled in table 1.

Rb–Sr and Sm–Nd studies were carried out on 15 whole rock samples of which an aliquot was completely dissolved in a mixture of hydrofluoric and perchloric acid. The Rb–Sr concentrations and the Sr isotopic compositions were determined using the isotope dilution method as described by Jäger (1979). The Sm–Nd concentrations and the Nd isotopic compositions were determined according to the method described by Stoessel and Ziegler (1989) using a two column element separation procedure. The Rb analyses were carried out on an "Ion Instruments" solid source mass spectrometer Elemental Sr, Sm and Nd measurements were carried out on a "VG Sector" thermal ionisation mass spectrometer by single and triple filament modes, respectively. The $^{87}Sr/^{86}Sr$ ratios are normalized to $^{88}Sr/^{86}Sr = 0.1194$. The mean $^{87}Sr/^{86}Sr$ ratio for 7 measurements of the NBS SRM-987 standard is 0.71022 ± 3 (σm). The $^{143}Nd/^{144}Nd$ ratios are normalized to a $^{146}Nd/^{144}Nd$ ratio of 0.72190. The mean value for 14 analyses of the La Jolla isotopic standard was 0.511843 ± 15 (σm). Total blanks were < 1 ng for Sm, 0.01–0.1 ng for Nd and < 10 ng for Sr. The results of the Rb–Sr analyses are listed in table 2 while those of the Sm–Nd analyses are given in table 3.

4. Major and trace element geochemistry

As discussed earlier the metavolcanic rock sequence has undergone greenschist metamorphism and the original magmatic rocks are altered to low grade metamorphic rocks. Post-magmatic bulk chemical changes must be considered when attempting to define the original tectonic setting of the volcanic sequence. Chemical changes can occur by reaction with seawater after submarine extrusion or by hydrothermal alteration (e.g. Hart, 1970 and Hart et al., 1974). Also, tectonism and metamorphism related recrystallisation and chemical exchange with country rock are possible (e.g. Wood, 1976; Condie et al., 1977). The possible mobilisation of elements such as Si, Al, Fe, Mg, Ca, Na, K, Rb, Sr, Ba and the L.O.I. during post-magmatic alteration of basic rocks has been shown by various studies (e.g. Humphries and

Thompson, 1978a and 1978b; Mottl and Holland, 1978; Pearce, 1975; Gill and Bridgewater, 1979). In contrast, elements such as Ti, P, Zr, Nb, Y, V, Cr, and Ni are considered to be refractory and relatively immobile under hydrothermal ocean ridge metamorphism up to epidote amphibolite facies conditions (Pearce, 1975; Humphries and Thompson, 1978a and 1978b; Winchester and Floyd, 1976, 1977; Floyd and Winchester, 1978). Numerous studies during the
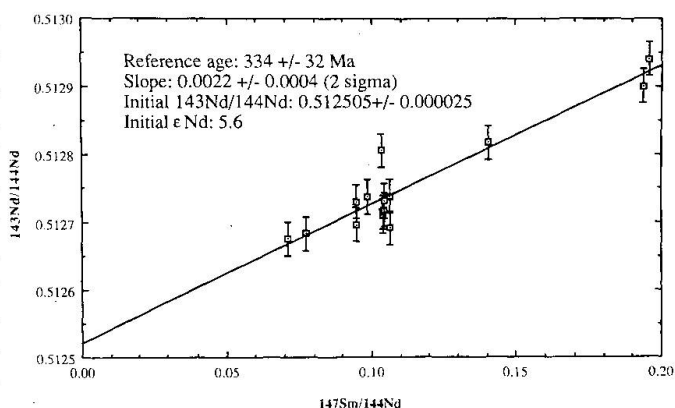

Fig. 8 Sm–Nd isochron plot for the analysed metavolcanic rocks, including the data of Ziegler et al. (1990) for the meta-carbonatites from the same thrust sheet. (See text for discussion.)

past two decades have shown that the Rare Earth Elements (REE) have to be considered as elements with limited mobility during post-magmatic alteration processes of basic rocks (e.g. Humphries et al., 1978c; Ludden and Thompson, 1979; Menzies and Seyfried, 1979).

The metavolcanics show strong variations in their major element composition. For example, the SiO_2, MgO and Na_2O contents range between 23–59 wt%, 1.7–10.4 wt% and 0.2–7.1 wt%, respectively. A comparison of the major element variation (not shown) with the trace element characteristics of the metavolcanics clearly indicates that post magmatic alteration has affected the major element chemistry of the partly strongly tuffaceous samples. The refractory trace elements mentioned above yield consistent and therefore interpretable characteristics. The following geochemical discussion is thus restricted to the refractory elements Ti, P, Zr, Nb, Y, V, Cr, and Ni. A discussion of the Sm–Nd systematics is given in chapter 5.

The Nb contents of the specimens were plotted versus their Y contents (Fig. 2) because of the relative stability of the Nb/Y ratios during metamorphism (Pearce and Cann, 1973; Fields et al., 1974). The result is a clear subdivision of the metavolcanics into an alkalic (high Nb contents)
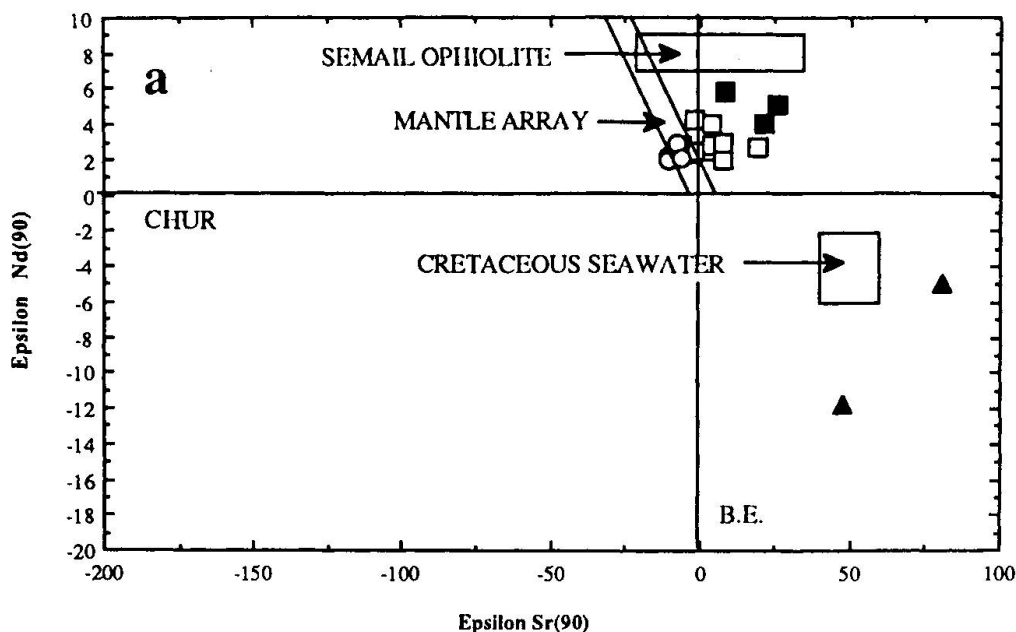
Fig. 9a   Plot of ε_{Sr} versus ε_{Nd} for the analysed metavolcanics (open squares represent alkalic samples, solid squares tholeiitic samples, circles carbonatites) in comparison an assumed minimum formation age of 90 Ma. The ε_{Nd} of a model depleted mantle would be 9.9. The solid triangles represent the analysed harzburgite and serpentinite samples. The data for the metacarbonatites was taken from ZIEGLER et al. (1990) while those for the Semail ophiolite and the Cretaceous seawater was taken from McCULLOCH et al. (1980).
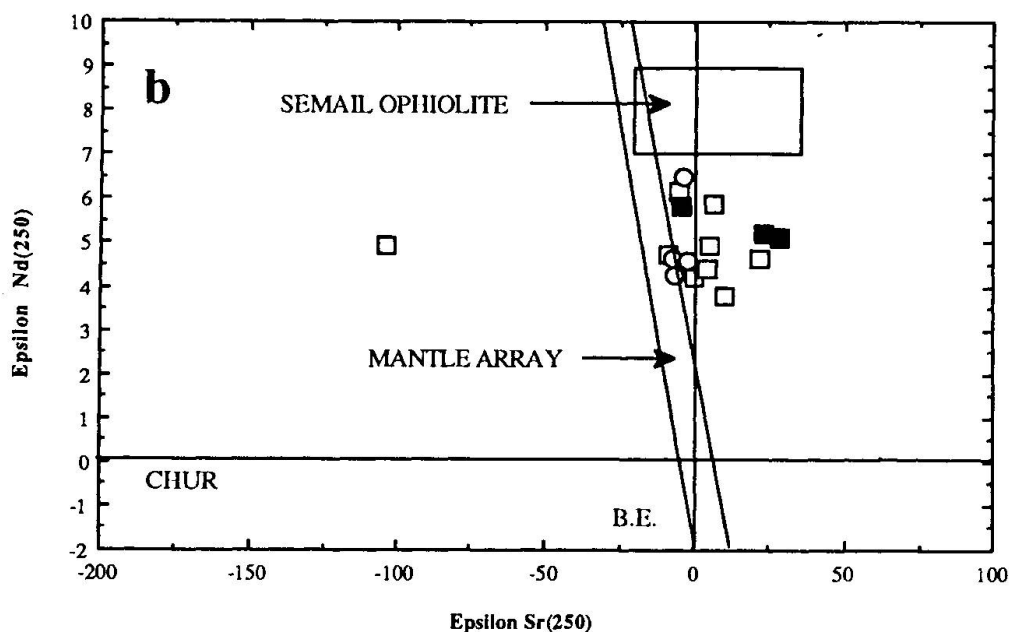
Fig. 9b ε_{Sr} versus ε_{Nd} diagram for an assumed age of the analysed volcanics of 250 Ma. (Same notation as in Fig. 9a.) The ε_{Nd} of a model depleted mantle would be 9.5.

and a tholeiitic group (low Nb contents). The Y concentrations are basically identical in both the alkalic and tholeiitic groups. The Zr/P_2O_5 ratio remains approximately stable with an increase in metamorphic grade (although Zr and P_2O_5 may show a slight increase in content). ELLIOTT (1973), ENGEL and ENGEL (1962) and STILLE (1980) showed that Zr and P_2O_5 have limited mobility

during metamorphism. Figure 3 is a plot of P_2O_5 versus Zr and also subdivides the metavolcanic rocks in an alkalic group (with high P_2O_5 and Zr contents) and a tholeiitic group (with low contents of the respective elements). A plot of Nb/Y versus Zr/TiO_2 (Fig. 4) (as discussed by WINCHESTER and FLOYD, 1977) confirms the distribution of both alkalic and tholeiitic rocks. This diagram
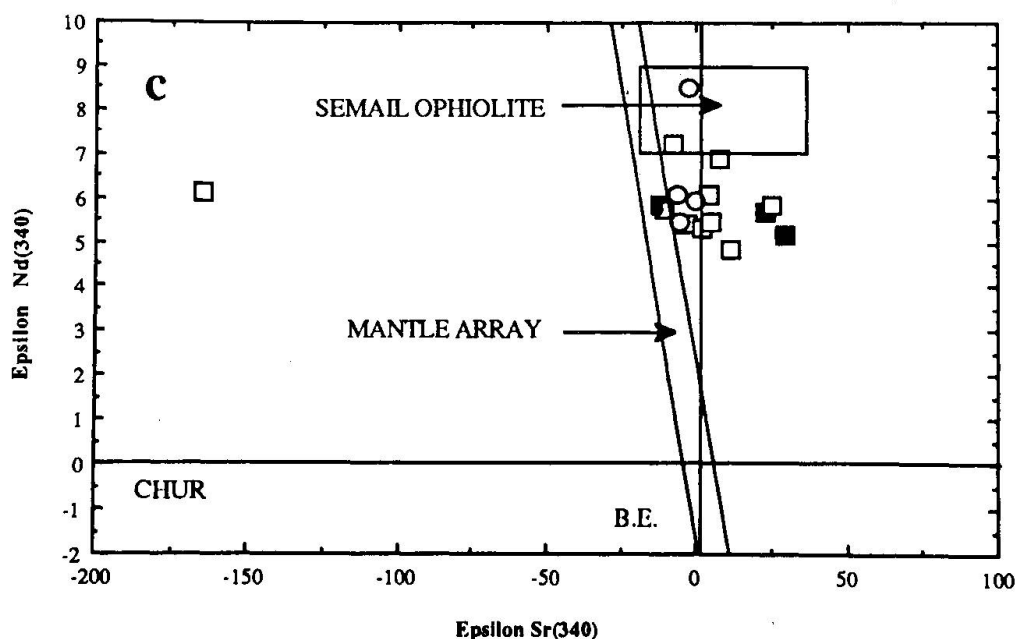
Fig. 9c ε_{Sr} versus ε_{Nd} diagram for an assumed maximum age of the analysed volcanics of 340 Ma. (Same notation as in Fig. 9a.) The ε_{Nd} of a model depleted mantle would be 9.3.

also shows the bimodal alkaline and subalkaline-tholeiitic type distribution of the metavolcanic rocks. The tectonic discriminant Ti/100, Zr, 3Y triangular diagram (Fig. 5) as suggested by PEARCE and CANN (1973) shows that the tholeiitic basalts of the metavolcanic unit represent ocean floor basalts while the alkaline basaltic rocks formed in an intraplate environment. This is confirmed by a plot of TiO_2 versus Zr (Fig. 6) as suggested by PEARCE and NORRY (1979). The alkalic samples plot in the field of within plate basalts with a trend towards a mid ocean ridge-type TiO_2–Zr distribution. The tholeiitic samples, plot in the fields of mid ocean ridge basalts and island arc tholeiites. Figure 7 is a plot of the Cr versus Y diagram (PEARCE, 1980) and shows that the samples cannot be classified as island arc tholeiite. This suggests that the metavolcanic thrust sheet consists entirely of tholeiitic mid ocean ridge basalts along with the alkalic within plate basalts.

The Ni contents of the metavolcanics are in the range which is typical for basic magmas. An evaluation of the Ti–V correlation according to SHERVAIS (1982, not shown) supports the subdivision of the metavolcanics into mid ocean ridge and within plate basalts. The tholeiites as compared to the alkaline volcanics have generally lower Ti contents and generally lower V values. The metamorphic overprint of the anlysed volcanics destroyed the major part of the primary mineralogical composition and it is not possible to make more detailed interpretations. Thus a discussion of the fractionation and differentiation processes using the trace element data is not justified.

5. Isotope geochemistry

According to the considerations of SEARLE (1980) and BECHENNEC (1987) concerning the age of the Haybi/Hawasina volcanism the ε_{Sr} and ε_{Nd} values have been calculated for an assumed approximate age of the metavolcanic rocks of 250 Ma (Permo-Triassic boundary).

The $^{87}Rb/^{86}Sr$ ratios of the specimens are (with the exception of UAE4735 and UAE4755 which yielded ratios of 3.4 and 0.48, respectively) below a value of 0.2 and thus do not allow the calculation of Rb–Sr isochron ages. With the exception of UAE4735 the initial Sr isotopic compositions (calculated for an assumed age of 250 Ma) range between 0.7038 and 0.7062 and thus extend from bulk earth values towards slightly elevated $^{87}Sr/$ ^{86}Sr ratios. The calculated (primordial) initial Sr composition of UAE4735 for t = 250 Ma of 0.6969, indicates the limitations of the assumption that the metavolcanic series formed approximately 250 Ma ago.

The $^{147}Sm/^{144}Nd$ ratios of the metavolcanic samples range between 0.09 and 0.20. A Sm–Nd reference line calculated through the data points (including the carbonatite data of ZIEGLER et al., 1990; disregarding the carbonatitic sample UAE4020 and the alkalic sample UAE4762) following the method of YORK (1969) yields an age of 334 Ma ± 32 Ma with a correlation coefficient of 0.94 and an intercept at $^{143}Nd/^{144}Nd = 0.512505$ ± 0.000025 (Fig. 8). This Carboniferous age is in conflict with the assumed Permian to Triassic stratigraphic age of the Haybi/Hawasina volca-

nism as determined by SEARLE (1980) and BECHENNEC (1987).

The initial ε_{Nd} and ε_{Sr} values of the metavolcanic rocks have been plotted in Fig. 9 a, b and c. Available data on the Upper Cretaceous Semail ophiolite (McCULLOCH et al., 1980) and the metacarbonatites associated with the metavolcanic rocks (ZIEGLER et al., 1990) have been included for comparison. Fig. 9 a shows the ε_{Nd}-ε_{Sr} values for an assumed minimum age of 90 Ma; Fig. 9b for an assumed age of 250 Ma and Fig. 9c for a maximum age of 340 Ma.

In spite of the large assumed age difference, the resulting patterns of Figs 9 a, b and c are quite surprising. In all the diagrams the metavolcanic rocks plot above ε_{NdCHUR} and well below the ε_{Nd} values obtained for the Semail ophiolite (McCULLOCH et al., 1980) and a model depleted mantle at t = 340 Ma (ε_{Nd} = 9.34), 250 Ma (ε_{Nd} = 9.53) and 90 Ma (ε_{Nd} = 9.87), respectively. The ε_{Nd} values of the metavolcanic rocks are in the same range as those reported for the metacarbonatites (ZIEGLER et al., 1990). A trend towards slightly elevated ε_{Nd} values can be observed in the tholeiitic samples. In Fig. 9 a, b, c the ε_{Sr} values of the specimens plot to the right of the mantle array and thus indicate an alteration of the Sr isotopic composition after the separation of the magmas from their mantle source by processes like crustal contamination, ophiolite obduction related metamorphism, hydrothermal and/or seawater alteration.

The ε_{Nd} and ε_{Sr} values of the analysed serpentinite and harzburgite samples from the Semail nappe (with a determined Cretaceous age; McCULLOCH et al., 1980) are displayed in Fig. 9a. This data plot outside the range of the fields defined by data of McCULLOCH et al. (1980) for the Semail ophiolite and the data for the metavolcanics of this study. Both samples plot close to the field representing Cretaceous seawater in an area which is typical of crustal material. These values can be interpreted as serpentinisation related alteration of the primary ultramafic mantle rocks by fluids with crustal Sr and Nd characteristics, leading to their present Sr–Nd isotopic composition. Such an alteration of the isotopic composition of the harzburgite and serpentinite samples is also indicated by their $^{147}Sm/^{144}Nd$ ratios which are outside the range commonly observed in oceanic crust material (DE PAOLO, 1989).

6. Discussion

The trace element data obtained for the metavolcanic unit between W. Shimal and W. Uyaynah in the UAE could (in spite of the tuffaceous nature

of many of the samples and greenschist facies metamorphism) show the protolithic nature of alkalic and tholeiitic volcanics within this subophiolitic thrust sheet. A comparison of the geochemical results of this study with those obtained by SEARLE et al. (1980) and BECHENNEC (1987) on unmetamorphosed Haybi/Hawasina volcanics underlying the Semail ophiolite nappe shows many similarities. It can be interpreted that the analysed metavolcanic rocks from the Asimah window represent a metamorphosed equivalent to the unmetamorphosed Haybi/Hawasina volcanics found throughout the Oman mountains.

The initial Nd and Sr isotopic compositions of the metavolcanic rocks are in close association to those of the metacarbonatites (ZIEGLER et al., 1990). It is of interest that the observed ε_{Nd} values are within the same range as those of the metacarbonatites while the ε_{Sr} values are usually slightly higher than those of the metacarbonatites. The $^{147}Sm/^{144}Nd$ ratios of the alkalic samples scatter around a value of 0.1 while those of the tholeiitic samples range between 0.16 and 0.2. In combination with the corresponding ε_{Nd} values varying between 3.5 and 6.5 this indicates that the alkalic metavolcanics (as well as the metacarbonatites) have a Sm/Nd fractionation factor α with values < 0.7 which according to DE PAOLO (1988) are typical for alkali basalts and kimberlites. While the tholeiitic samples have α-values between 0.7 and 1.0 and according to De Paolo are typically observed in tholeiitic basalts. These results thus confirm the geochemical classification of the samples as discussed above.

The ε_{Nd} values are slightly lower than the PREMA mantle component of ZINDLER and HART (1986) while the ε_{Sr} values are slightly higher than PREMA. The observed ε_{Nd} and ε_{Sr} values are within the range of data observed for oceanic islands such as Hawaii, the Azores, and Bouvet (ZINDLER and HART, 1986). Data on the Semail ophiolite (McCULLOCH et al., 1980) and the Bay of Islands Ophiolite Complex (JACOBSEN and WASSERBURG, 1979), yield ε_{Nd} values for ophiolitic rocks of approximately +8. This shows that the metavolcanic rocks have ε_{Nd} values well below those found in ophiolite sequences.

It is thus clear that the metavolcanic rocks do not represent an undisturbed partial melt fraction of a depleted mantle source similar to the Oman ophiolite sequence. The present Nd and Sr isotopic composition of the analysed samples can thus be explained by the following two models:

A) If the isotopic composition of the metavolcanic rocks has not changed since their fractionation from their mantle source then they are either derived from a low ε_{Nd} subcontinental

mantle (this mantle type is described e.g. by MENZIES et al., 1983; SEMKEN, 1984; and PERRY et al., 1987) or they are derived from an upwelling asthenospheric low ε_{Nd} mantle (similar to the ideas of NELSON et al., 1988 on the formation of carbonatite magmas) with a slightly enriched composition in comparison with the PREMA and HIMU mantle components of ZINDLER and HART (1986).

B) If, the isotopic composition of the metavolcanics has been affected by crustal contamination (causing a shift of the ε_{Nd} values towards lower values and a shift of the ε_{Sr} values towards higher values) or by post formation alteration processes they might still be derived from a depleted mantle source.

A comparison of the obtained results with data from crustal contaminated provinces such as the Scottish Hebrides (CARTER et al., 1978) or southern Africa (HAWKESWORTH et al., 1982), shows that they do not form a similar mixing pattern. Therefore if contaminated they do not represent the result of pure crustal contamination. The observation that the obtained ε_{Sr} values plot to the right of the mantle array points to the possibility that the Sr isotopic composition of the metavolcanic rocks was altered by exchange with hydrothermal fluids derived from ocean water or from metamorphic fluids. This idea is supported by the observation that the ε_{Sr} values of the metacarbonatites (with their very high Sr contents > 2800 ppm; ZIEGLER et al., 1990) are in general below those of the metavolcanic rocks with considerably lower Sr contents. The metabasalts due to their lower Sr content the Sr isotopic composition can be more easily altered by circulating fluids in comparison with the metacarbonatites (which plot closer to mantle array).

The positive ε_{Nd} values are well within values typical of mantle magmas and are scattered around the values observed for the metacarbonatites. The observed deviation of the ε_{Nd} values from those of the metacarbonatites could be explained by slight alterations of the Nd isotopic composition by processes such as mantle-crust interaction, wall rock assimilation and seafloor – or regional metamorphism. The difference between the calculated Sm–Nd reference age of 334 Ma ± 32 Ma and the assumed age of the Hawasina/Haybi volcanism according to BECHENNEC (1987) and SEARLE (1980) can be interpreted in two ways. First, by the idea of a slight disturbance of the Nd isotopic system, and secondly it might indicate that Hawasina/Haybi magmas have already been generated in the mantle during the Lower Carboniferous (as opposed to only during the Upper Permian as it was previously assumed).

In spite of the differences mentioned in the uncertainties of the degree of disturbance of the Sr and Nd isotopic systems, the authors favour the idea that the metavolcanic rocks initially had Nd and Sr isotopic compositions similar to those of the metacarbonatites of the same area. Thus they are derived from a mantle source with relatively low ε_{Nd} characteristics.

7. Conclusions

The study of the metavolcanic rocks occurring beneath the Semail ophiolite nappe in the Asimah window in the UAE has shown that volcanism can be classified into two different magmatic series: alkaline and tholeiitic. Such a distribution of volcanism has also been described for the unmetamorphosed Haybi/Hawasina volcanism. It can thus be postulated that rocks similar to the Haybi/Hawasina volcanics represent the unmetamorphosed protoliths of the metavolcanic rocks of the metamorphic thrust sheets below the Semail nappe. According to their initial Nd isotopic composition the metavolcanic rocks either are derived from a subcontinental lithospheric or from an asthenospheric, low ε_{Nd} mantle and not from a "Semail type" depleted mantle source. The Sr isotopic composition of the metavolcanic rocks (and to a lesser degree also the Nd isotopic composition) indicate alteration by oceanic or metamorphic hydrothermal fluids. The Lower Carboniferous Sm–Nd reference age for the metavolcanics of 334 Ma ± 32 Ma for the onset of the Haybi/Hawasina volcanism is considerably higher than the Upper Permian age which was postulated by BECHENNEC (1987) and SEARLE (1980). The intimate association of the metavolcanic rocks (including the partly subaerial meta-carbonatites; ZIEGLER et al., 1990) with deep sea sediments such as metaradiolarites strongly suggests that they formed in an ocean island tectonic setting along the passive margin of the Arabian continent. The occurrence of alkaline volcanics in association with carbonatites on oceanic islands has only been observed on the Canary Islands and the Cape Verde islands (see e.g. LE BAS, 1984; SILVA, 1981; ALLÈGRE et al., 1971 and DAVIES et al., 1989). It is thus postulated that the metavolcanic series formed in a tectonic situation similar to that of the mentioned islands at the margin of the African continent

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