

Geochemistry of mafic rocks of the Karakaya complex, Turkey: evidence for plume-involvement in the Palaeotethyan extensional regime during the Middle and Late Triassic

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Abstract The Karakaya Complex within the Early Mesozoic Cimmerian Orogeny in northern Turkey represents the remnants of the Palaeotethys. It includes slivers and/or mega-blocks of slightly metamorphic basic volcanic rocks associated with fossiliferous sediments as well as hypabyssal and intrusive rocks with basaltic-andesitic to ultramafic compositions. They display two distinct compositional groups; namely alkaline and variably tholeiitic. The alkaline basalt samples are more akin to oceanic-island basalts (OIB) with relatively enriched trace element characteristics together with strong partitioning in HREE ($[La/Yb]_N = 5.8–16.2$), suggesting that garnet is present as a residual phase in the source of those basalts. The variably tholeiitic samples apart from diabases display E-MORB characteristics; being relatively depleted compared to the alkaline counterparts and less fractionated REE patterns ($[La/Yb]_N = 2.1–3.6$). The diabases, on the other hand, are distinctively different with a significant negative Nb anomaly ($Zr/Nb = 28.9–43.4$) and flat REE patterns ($[La/Yb]_N = 0.8–1.4$), suggesting their generation above a supra-subduction zone, probably a back-arc basin. These results may suggest that a mantle plume-related magmatism associated with extensional oceanic system should have been installed within Palaeotethys during Middle-Late Triassic time, which was then incorporated into subduction-accretion prism forming the final picture, that is, “the Karakaya Complex”.

Keywords Karakaya Complex · Mantle Plume · OIB · Triassic · Palaeotethys

Introduction

It is commonly accepted that during its evolutionary history Turkey has been characterized by a number of continental and oceanic microplates with different Palaeozoic and Mesozoic histories. The consecutive collision events between Laurasia and Gondwana as well as smaller microplates or terranes accounted for their association during the course of time. The “Tethyan oceans” separating these terranes include oceanic plates of Proto-, Palaeo-, Neo-, and Para-tethys (Sengör and Yılmaz 1981; Stampfli 2000; Robertson and Ustaömer 2002) that have been the subject of a long debate together with microplates adjacent to them [for a brief review see Stampfli (2000)].

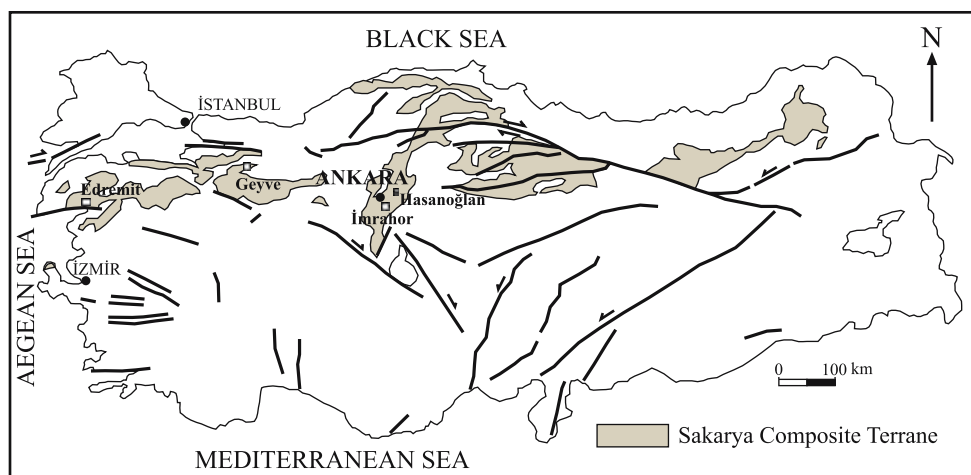
The closure of the Late Palaeozoic–Early Mesozoic “Palaeotethys” during Early Mesozoic led to accretion of the oceanic and continental units, eventually forming the “Karakaya Complex” (Tekeli 1981; Sengör et al. 1984). The tectonic unit produced is characterized by an east–west trending belt along northern Turkey (Fig. 1) and defined under several names, such as the Cimmerian Continent (Sengör et al. 1984), “pre-Jurassic basement” of the Sakarya Zone (Okay 1989) or the Sakarya Composite Terrane (Göncüoğlu et al. 1997).

The Karakaya Complex became a “puzzle” which is in dispute for many years, commencing actually with the name “Karakaya Formation” used by Bingöl et al. (1973), to define the Triassic low-grade metamorphic rock assemblage of “rift origin”. Tekeli (1981) was the first, who suggested a Palaeotethyan subduction-accretion origin for this rock unit. Many researchers have discussed this

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Fig. 1 Distribution of the Sakarya Composite Terrane in northern Turkey (after Göncüoğlu et al. 1997) and locations of the studied areas



controversial subject, offering an extensive scheme of tectonic environments comprising rift, forearc, seamount, etc. Recently Okay and Göncüoğlu (2004) attempted to reorganize the complex into several informal units including:

- Variably metamorphosed metaclastic–metabasic basement intruded by Palaeozoic granitoids (Göncüoğlu et al. 1987)
- Greywacke series (Orhanlar Greywacke of Okay et al. 1991)
- Arkosic sandstone series with basaltic lavas (Hodul Unit of Okay et al. 1991)
- Metabasite-phyllite-marble series (Nilüfer Unit of Okay et al. 1991)
- Basalt, limestone, grain flows, debris flows, and olistostrome series with limestone blocks (Cal Unit of Okay et al. 1991).

Except the greywacke Series, all units of the Karakaya Complex are characterized by the presence of variably metamorphosed mafic volcanic rocks (Capan and Floyd 1985; Floyd 1993; Ustaömer and Robertson 1994, 1999; Pickett and Robertson 1996, 2004; Yaliniz and Göncüoğlu 2002; Genc 2004) all along the complex. Nevertheless, they have not been satisfying either due to insufficient field data or lack of precise age data and/or reliable geochemical data including REE and isotope data. Hence, they resulted in very diverse interpretations or in over-generalizations.

The aim of this study is to determine the geological relationships, age and petrological features of these volcanic rocks in some selected areas for a better understanding of their tectono-magmatic setting and hence the geological evolution of the Karakaya Complex.

For this purpose, we selected four different areas in western and central parts of the complex (Fig. 1) that were previously studied by different authors. After a mapping campaign in these areas we sampled only those basic volcanic rocks, where we could obtain reliable age data

from the associated sediments. The geochemical data used for this purpose include major oxides, trace and rare earth element data; however, the evaluation of isotope data on the geochemical groups identified in this study is in progress.

Geological features

The first study area is located at Imrahor village, just 5 km southeast of Ankara (Fig. 1). The second one, namely Hasanoglan, is situated approximately 35 km northeast of Ankara (Fig. 1). The third study area is located at Ortaoba village, 4 km northwest of Edremit (Fig. 1). The last study area is located at the north of Kadirler village, 10 km southeast of Geyve (Fig. 1).

The Karakaya Complex in the Ankara region is mainly governed by three rock units, which are from oldest to youngest: (a) the Eymir Complex (Kocyigit 1992), (b) the Triassic Karakaya Group including the Bahcecik, and (c) Olukman Formations (Kocyigit et al. 1991; Altiner and Kocyigit 1993). It is unconformably overlain by Jurassic cover rocks.

The Eymir Complex, exposed in the region to the south Asagi Imrahor Village (Fig. 2), is composed of a regionally metamorphosed chaotic mixture of clastic rocks (mainly greywackes, slates, quartzo-feldspatic sandstones and conglomerates) with knockers of metabasic rocks, mainly Na-amphibole-bearing disrupted lava flows, which have been cross-cut by a number of diabase dykes (Sayit 2005).

The Bahcecik Formation comprises an alternation of spilitic basalts, shallow and deep marine limestones, mudstones and volcanoclastics as well. The syn-depositional nature of basaltic extrusions and limestones could be conceived by the occurrence of basalt clasts in limestone, and similarly the presence of limestone clasts in basalts (Fig. 3).

Fig. 2 Geological map of the Imrahor area (modified from Sayit 2005)

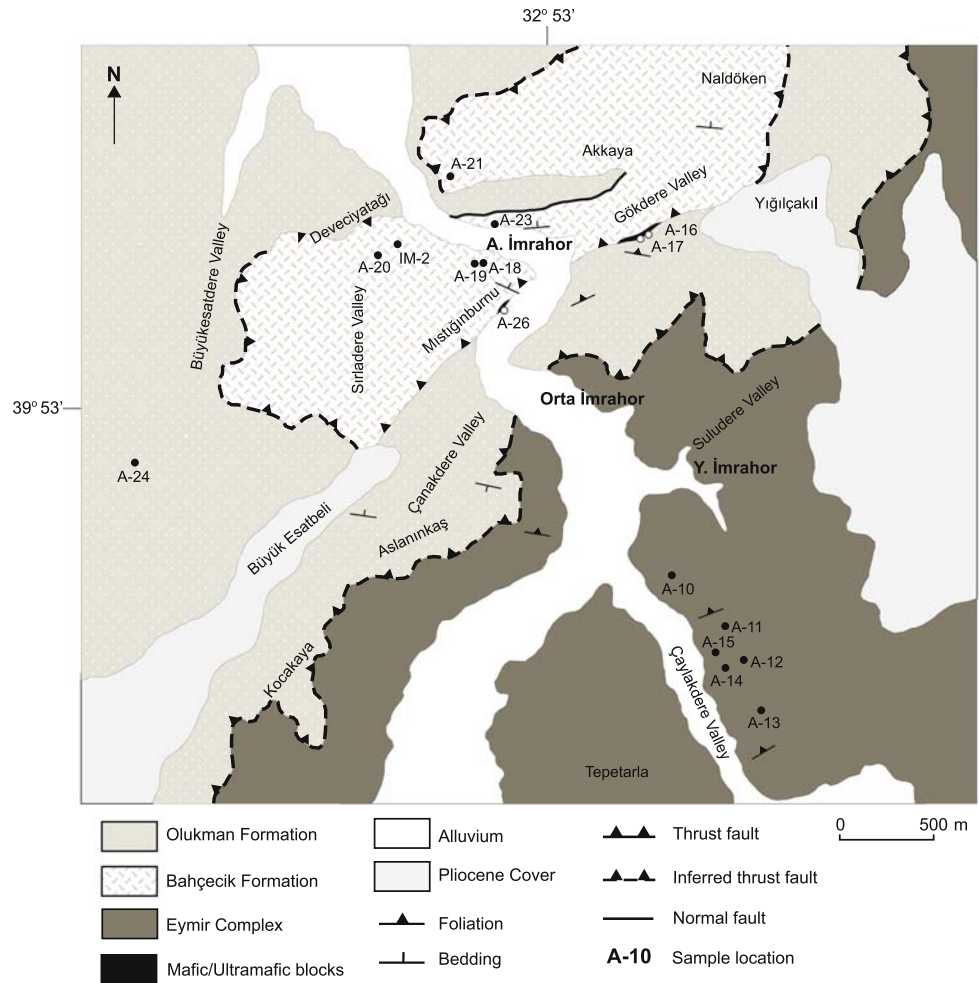


Fig. 3 The syn-depositional nature of basalt and limestone within the Bahçecik Formation. Note that they are interfingered with each other. While basalt clasts can be found in limestone, limestone clasts can also be found in basalt



In addition to the basaltic lava flows, the unit also comprises blocks of mafic/ultramafic intrusives in a small part of the study area. These intrusive rocks, including low-grade metamorphosed gabbros, together with minor amounts of serpentinized wherlitic ultramafics, occur mainly at the contact between the Olukman and Bahçecik Formations.

An Anisian age was assigned to the Bahçecik Formation by Özgül (1993) based on the foraminifers (*Glomospira densa*, *Glomospirella grandis* (?), *Endothrid* sp., *Meandrospira dinarica* (?)) found within the slightly recrystallized

limestones, which interfinger with the pillowed basalt flows. In the present study, the re-examination of these limestones in Mistığınburnu Hill to the west of Imrahor village, has revealed the fossil assemblage (*Pliammia densa*, PANTIC, 1965, *Meandrospira dinarica*, KOCHANSKY-DEVIDE and PANTIC, 1965, *Meandrospira deformata*, SALAJ, 1967, *Endoteba* sp., *Aulotortus* (?) sp., *Aulotortus* (?) *eotriasicus*, ZANINETTI, RETTORI and MARTINI, 1994 and *Paulbronnimannella whittakeri* RETTORI; determined by Dr. D. Altner) which characterizes Middle Anisian, thus placing constraints on the age of the basaltic volcanism

investigated. In the Hasanoglan area, the limestones also interfinger with the pillow lavas and include Anisian fossils (Altner and Kocyigit 1993; Fig. 4).

The Olukman Formation is composed of slightly metamorphic and intensely deformed clastic rocks, debris flows with volcanic clasts, greywackes, mica-rich sandstones, mudstones and huge blocks of recrystallized limestones of Permian age. An Upper–Middle Triassic is ascribed to the unit (Akyürek et al. 1984). It does not include syn-sedimentary basic volcanic lavas or phyllites, thus differing from the other Karakaya units.

In the Biga Peninsula, the Karakaya Complex includes the Ortaoba Unit of Pickett (1994) in addition to Cal and

Nilüfer Units. The unit comprises gray ribbon cherts alternating with slightly metamorphic greenish-gray siltstones with single chert bands and basaltic rocks (samples: Ortaoba Type-I, Fig. 5). A Ladinian (?)–Carnian age were assigned to these cherts of the Ortaoba Unit, based on the conodont finding (*Gladigondolella tethydis*) from a 60-cm thick gray chert band on the forest road (determination by Dr. H. Kozur, unpublished data). Further to the NW of this locality, spilitic basalts assigned to the “massive basalt flows” of the Nilüfer volcanics sensu Pickett and Robertson (1996) were sampled (Ortaoba Type-II) to check the REE geochemistry, not considered in the previous study of Pickett (1994).

Fig. 4 Geological map of the Hasanoglan area (after Altner and Kocyigit 1996)

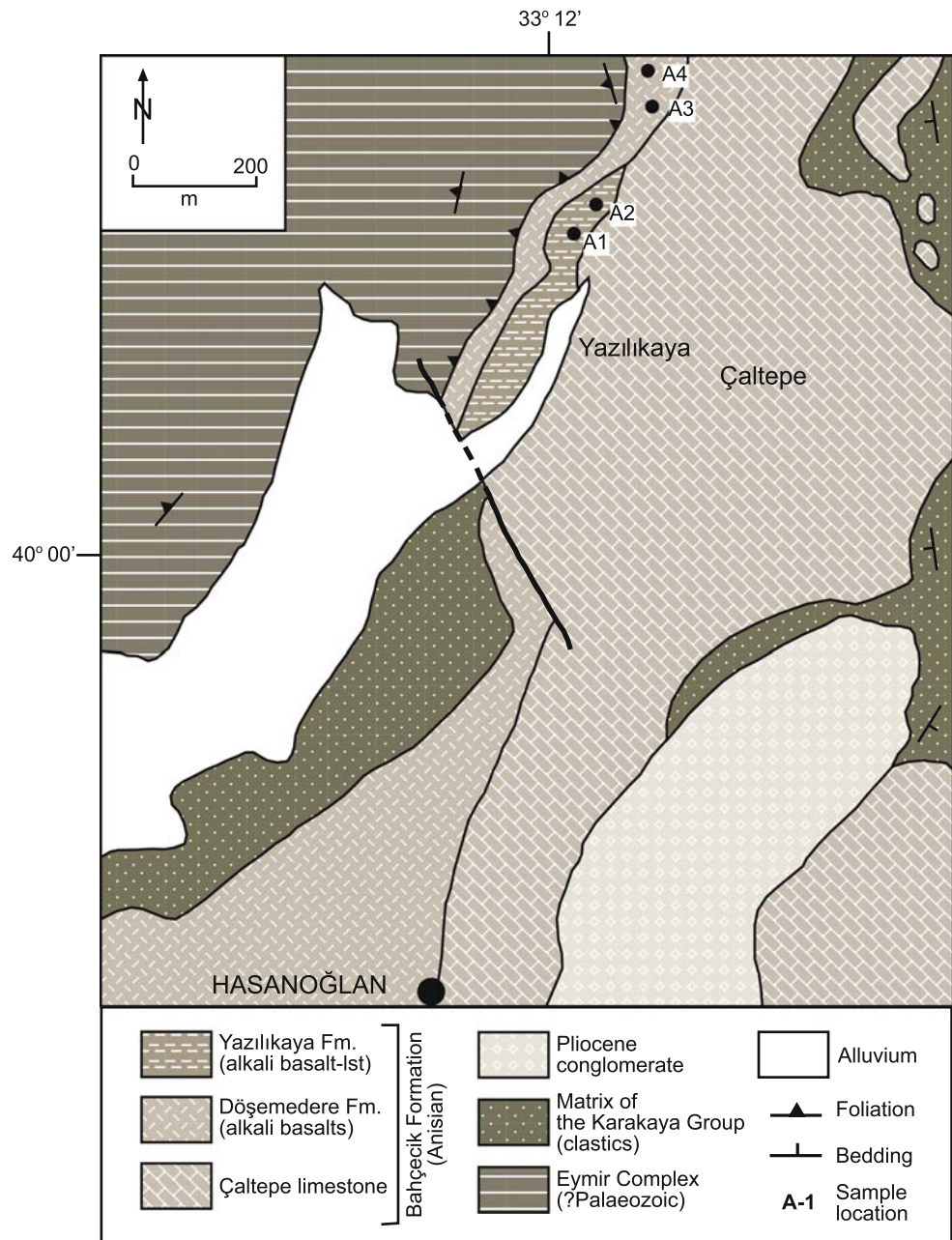
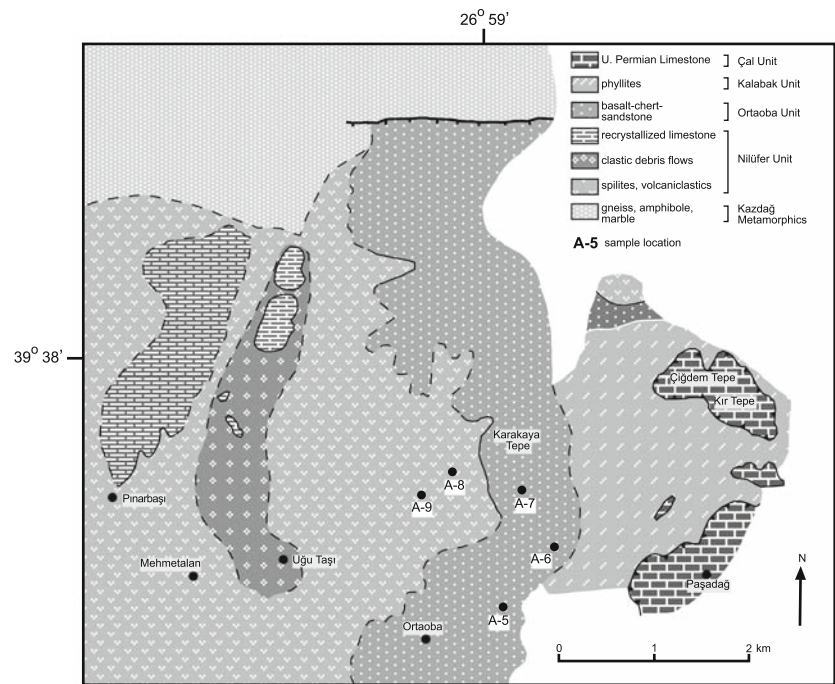


Fig. 5 Geological map of the Ortaoba (after Pickett and Robertson 1996)



In the Geyve region, the sampling was carried out on the “pillow basalt–limestone association” of Göncüoğlu et al. (2004), representing the Bahcecik Formation in Ankara or the Cal Unit in the Biga Peninsula (Fig. 6). In this area, the pillow basalts are relatively fresh and interfinger with pinkish-white, thick-bedded oolitic limestones (Fig. 7). Basalts are sometimes observed as lava-flows with baked zones on red mudstones with volcanoclastic detritus between thick layers of white limestone. The lowermost part of the overlying limestone layers includes clasts of amygdaloidal basalt. Vice versa, there are reddish white limestone fillings between the pillows obviously evidencing the coeval formation of the pillow basalts and limestones. Overall, an Anisian age is accepted for this unit, based on a correlation with the fossil-bearing outcrops in Middle Sakarya (Göncüoğlu et al. 2004) and E of Bursa (conodont finding, Genc et al. 1986).

Petrography

An overall consideration of the investigated basic volcanic rocks indicates that primary mineralogy in the alkali basalts is composed of clinopyroxene (pinkish Ti-augite) + plagioclase + olivine + sphene ± kaersutite + opaque oxides, whereas the secondary mineral phases are represented by serpentine + epidote + chlorite + calcite ± actinolite ± prehnite ± pumpellyite ± leucoxene. The tholeiitic basalts are also represented by a similar mineral assemblage

including clinopyroxene (brownish Ti-augite) + plagioclase + olivine + sphene + opaque oxides + serpentine + epidote + chlorite + prehnite ± pumpellyite. Most of the basalts are highly porphyritic, including phenocrysts of clinopyroxene, plagioclase and pseudomorphs after olivine. Clinopyroxene appear to be the dominant phenocryst phase, generally retaining its original structure. Sub-ophitic texture is sometimes observed, including clinopyroxenes associated with smaller plagioclase laths. Glomeroporphyritic aggregates of clinopyroxene and plagioclase were also found in the studied basalts. The olivine phenocrysts, on the other hand, have been completely replaced mainly by serpentine minerals. Pinkish Ti-augite is the typical clinopyroxene phase especially for the alkaline basalts sampled. Brown amphibole (probably kaersutite) is a distinctive mineral phase for the alkali İmrahor lavas, observed generally at the rims of Ti-augite as needle-like, prismatic minerals. It must be noted that alkaline nature of the basalts is clearly reflected by their Ti-rich mineralogy (Ti-augite ± kaersutite ± leucoxene). Ti-rich tholeiitic gabbros/microgabbros in İmrahor are characterized by a similar mineral assemblage as the basalts such that clinopyroxene (brownish Ti-augite) + plagioclase + olivine + sphene + opaque oxides are primary mineral phases, while secondary mineralogy includes serpentine + epidote + chlorite + prehnite + pumpellyite. The ultramafic sample is somewhat similar to the basalts and gabbros in terms of its Ti-rich composition as indicated by the abundant presence of Ti-augites and rarely kaersutite. Olivine is another

Fig. 6 Geological map of the Kadirler area (after Gönçüoğlu et al. 2004)

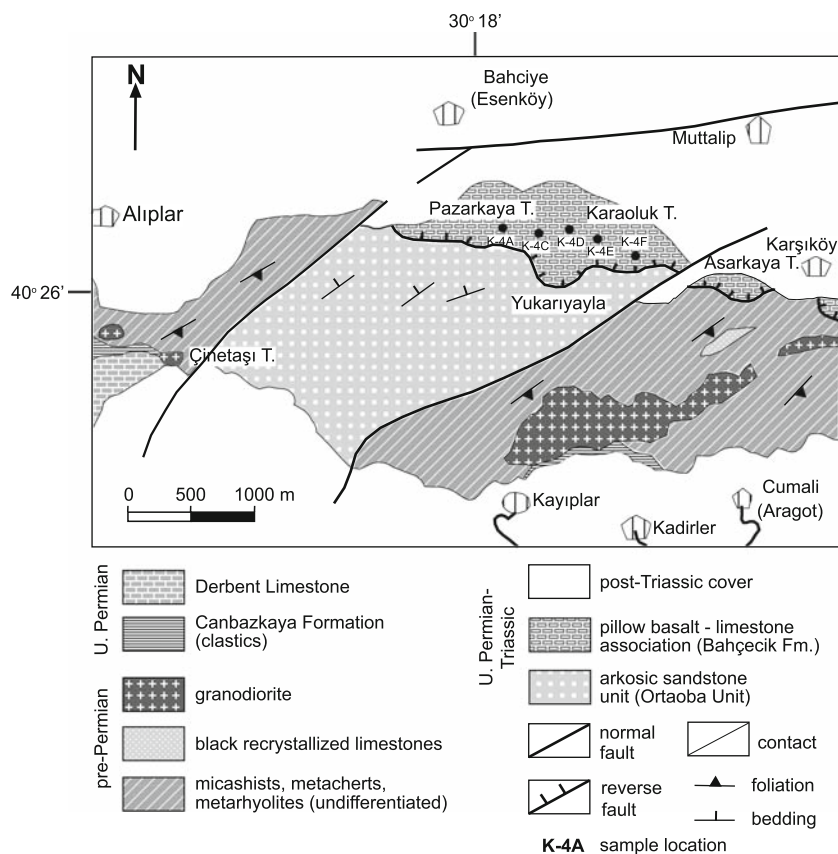


Fig. 7 The relationship between basalt and limestone from the Kadirler area. The clasts of basalt can be observed within limestone. Similarly, limestone fillings are present between the pillow lobes, suggesting their syn-depositional character

major mineral phase; however, it appears to be entirely serpentinized, displaying well-developed mesh-texture. The diabbases, compared with the rest of samples, are somewhat different in such a way that hornblende and plagioclase are the major primary phases while clinopyroxene is found in minor amounts as well as accessory sphene. However, the secondary mineral assemblage is very similar, including chlorite + epidote + prehnite.

Geochemistry

For the geochemical investigation, 34 relatively less-altered samples were selected under microscope. The major elements together with Sc, Ba and Ni were analyzed by inductively coupled plasma-atomic emission spectrometry (ICP-AES), while the rest of trace elements and rare earth elements (REE) were determined by inductively coupled plasma-mass spectrometry (ICP-MS). All the analyses were conducted by ACME Analytical Laboratories Ltd., Canada.

Alteration

Intensely influenced by low-grade hydrothermal alteration, the presence of selected element mobility is expected to occur in most of the samples, especially comprising the large ion lithophile (LIL) elements (e.g. Pearce and Cann 1973; Thompson 1991; Staudigel et al. 1996). The loss on ignition (LOI) values are also highly variable (up to 13.7%, Table 1), indicating some degree of alteration and the presence of secondary volatile and carbonate phases.

Highly inconsistent abundances exhibited by most major elements (except Ti and P) together with LIL elements

Table 1 Chemical composition of the investigated igneous rocks from the Karakaya Complex

	Kadirler (basalt)					Hasanoglan (basalt)			
	4A	4C	4D	4E	4F	A-1	A-2	A-3	A-4
SiO ₂	43.24	38.77	37.42	43.25	40.67	44.19	49.4	43.71	42.9
Al ₂ O ₃	13.79	12.1	11.91	14.06	11.73	12.98	11.52	9.8	14.66
Fe ₂ O ₃	10.88	8.77	9.54	7.56	9.25	14.11	10.32	9.16	10.64
MgO	5.31	4.07	3.6	3.62	3.92	9.34	5.01	2.35	3.66
CaO	10.82	16.25	16.2	11.4	15.17	7.36	8.13	14.94	8.83
Na ₂ O	3.39	2.95	3.22	4.64	4.27	4.26	4.32	3.84	5.63
K ₂ O	1.14	1.44	1.42	0.99	0.33	0.22	0.07	0.08	0.05
TiO ₂	2.44	2.11	2.08	2.45	2.01	2.67	2.78	2.38	3.56
P ₂ O ₅	0.36	0.33	0.32	0.37	0.31	0.32	0.35	0.52	0.53
MnO	0.11	0.11	0.1	0.14	0.08	0.13	0.07	0.09	0.09
Cr ₂ O ₃	0.056	0.045	0.044	0.039	0.062	0.092	0.097	0.055	0.065
LOI	8.2	12.5	13.7	11.1	11.5	4	7.7	13.1	9.1
Total	99.77	99.47	99.58	99.66	99.32	99.7	99.81	100.04	99.74
Sc	22	20	20	29	23	29	24	15	26
Ni	216.4	186.9	140.8	273.4	128.7	223.8	355	116.7	143.4
Pb	1.8	2.2	2.2	1.1	2.4	0.5	1.4	1.9	1.1
Ba	320.3	280.7	220.7	312.8	194.9	147	47	53.8	78
Co	63.1	50	45.1	82.2	46.3	61.4	62.1	28.7	45.4
Cs	0.6	0.8	1.1	0.5	0.4	1.3	<.1	0.3	<.1
Ga	20.4	17.8	17.7	19.6	15.5	20	17.4	11.5	21.4
Hf	4.3	4	3.7	3.5	4.2	3.4	4.7	4.7	6.7
Nb	40.8	36.4	34.4	26.4	34.7	34.2	28.9	32.5	49.5
Rb	21.7	31	32.8	16.9	4	3.1	0.7	1.1	1.5
Sr	404.3	380.3	332.6	416.7	403.6	234.6	190.7	336.7	365.2
Ta	2.1	2.1	2	1.5	2	2.2	1.7	2.2	3.3
Th	4.5	4.2	2.9	2.6	3.4	2	2.5	2.3	4.2
U	0.9	1	1.1	1	0.7	0.6	0.8	0.7	0.9
V	225	197	197	291	220	301	251	180	276
W	1	0.3	0.1	0.1	0.7	0.2	0.2	0.4	0.9
Zr	178.1	156.1	153.8	133.7	152	118.5	175.8	192.2	260.4
Y	21.6	19.7	19.4	23.3	22	21.9	22.3	21.2	25.7
La	24.5	25.2	23.8	14.9	22.6	22.5	24.2	29.1	33.9
Ce	50.4	49.7	47.2	34.7	45	45.8	54.4	58.9	70.8
Pr	6.23	6.02	5.75	4.63	5.51	5.51	6.8	7.34	8.31
Nd	27.2	24.9	24.7	24	23.9	23.9	30.4	31.2	35.6
Sm	5.5	5.6	4.8	5.6	5.4	5.8	7.4	7.3	8
Eu	1.8	1.62	1.47	1.82	1.68	1.95	2.26	2.17	2.46
Gd	5.3	4.65	4.72	5.15	4.89	5.24	5.92	5.76	6.22
Tb	0.73	0.63	0.67	0.8	0.7	0.82	0.93	0.86	1.01
Dy	4.33	3.42	3.43	4.33	3.59	4.36	4.48	4.18	5.19
Ho	0.76	0.65	0.64	0.8	0.75	0.89	0.87	0.81	1.04
Er	1.76	1.61	1.67	2.01	1.68	2.05	1.93	1.77	2.34
Tm	0.24	0.24	0.22	0.27	0.25	0.28	0.27	0.24	0.32
Yb	1.57	1.64	1.28	1.85	1.58	1.52	1.56	1.3	1.81
Lu	0.23	0.23	0.18	0.24	0.24	0.22	0.21	0.18	0.3

(except Th) makes them unreliable indicators to infer petrogenesis. Thus, the geochemical discussion given below is based on the elements thought to be relatively immobile under conditions of low-grade metamorphism (e.g. Pearce 1975; Floyd and Winchester 1978; Bienvenu et al. 1990; Kerrich et al. 1998).

Discrimination of the volcanic rocks

The Zr/TiO₂ vs. Nb/Y diagram (Winchester and Floyd 1977) indicates that the studied samples are categorized into two groups, namely, alkaline basalts and variably tholeiitic basalts (Fig. 8). The Hasanoglan and Kadirler lavas fall into

Table 1 continued

	Ortaoba (basalt)				
	A-5	A-6	A-7	A-8	A-9
SiO ₂	46.94	47.22	49.02	46.54	48.88
Al ₂ O ₃	16.74	14.09	12.8	12.64	14.19
Fe ₂ O ₃	7.92	12.93	11.53	12.09	11.7
MgO	8.13	7.55	6.38	8.65	6.43
CaO	10.77	7.76	8.73	7.03	8.49
Na ₂ O	2.96	3.34	3.15	3.79	4.05
K ₂ O	0.9	0.63	0.04	0.3	0.44
TiO ₂	0.97	1.61	1.25	2.53	2
P ₂ O ₅	0.12	0.21	0.17	0.3	0.29
MnO	0.23	0.2	0.19	0.19	0.18
Cr ₂ O ₃	0.045	0.033	0.016	0.068	0.024
LOI	4.3	4.2	6.6	5.6	3.1
Total	100.04	99.79	99.89	99.75	99.78
Sc	47	43	41	31	27
Ni	84.9	33.9	32.7	135.4	42
Pb	1.2	1.0	0.8	1.7	0.5
Ba	224.1	389.5	76.3	118.4	140.4
Co	47.7	47.1	41.9	46.3	39.9
Cs	3.9	1.9	3.4	1.8	0.7
Ga	17	17.4	16	19.9	18.9
Hf	1.6	2.8	2.1	4.4	3.7
Nb	6.8	12.3	9	27.7	19.3
Rb	22.6	15.7	2	9	5
Sr	148.3	254	243.5	169.3	546.7
Ta	0.4	0.8	0.5	1.8	1.2
Th	0.5	0.7	0.9	3.1	1.7
U	0.2	0.3	0.1	0.6	0.5
V	249	363	280	300	241
W	0.9	0.5	0.8	0.8	0.4
Zr	54.1	101.2	70.8	163	151
Y	17.6	32.3	23.2	22.5	26.9
La	5.4	9.9	7.7	22.7	17
Ce	12.5	22.4	17	50.4	39
Pr	1.61	2.95	2.12	6.1	5
Nd	7.6	13.8	10.5	28.9	23
Sm	2.4	4.4	3.1	6.1	5.9
Eu	0.91	1.34	1.03	1.99	1.87
Gd	2.65	4.51	3.3	5.67	5.04
Tb	0.52	0.93	0.68	0.9	0.98
Dy	3.22	5.18	3.81	4.65	5.03
Ho	0.69	1.25	0.86	0.89	1.08
Er	1.8	3.37	2.32	2.11	2.5
Tm	0.23	0.49	0.34	0.3	0.36
Yb	1.48	2.91	2.06	1.71	2.11
Lu	0.23	0.48	0.33	0.25	0.3

the alkaline field, while the Eymir diabase dykes, which are markedly confined in a different place relative to the rest, are plotted in the sub-alkaline field. The Ortaoba and Imrahor samples are characterized by two different types of igneous rocks such that three Ortaoba basalts are of sub-alkaline character, whereas the other two samples from the same

locality reflect alkaline character. If the Imrahor samples are considered, it is observed that gabbros are clearly gathered in sub-alkaline field. Most of the basalts from Imrahor, however, fall into alkaline field. The two basalt samples and an ultramafic sample are plotted just on the boundary separating sub-alkaline basalts from alkaline ones.

Table 1 continued

Imrahor (Eymir) Diabase Dykes										
	A-10	A-11	A-12	A-13	A-14	A-15	A-25	J-19	J-25	IM-11
SiO₂	53.91	52.83	53.73	50.35	52.92	53.05	52.97	50.29	50.18	50.24
Al₂O₃	14.45	14.95	15.13	14.68	14.95	14.32	15.72	14.99	15.43	14.96
Fe₂O₃	12.53	13.4	12.21	15.34	13.22	13.28	11.32	11.69	12.03	11.83
MgO	3.46	4.01	3.68	4.49	3.99	4.02	5.14	5.63	4.1	6.06
CaO	8.08	4.92	5.47	3.58	5.33	6.96	2.6	8.94	7.87	6.15
Na₂O	2.59	3.15	2.69	3.29	2.73	3.11	4.3	3.58	3.62	4.75
K₂O	0.04	0.08	0.04	0.26	0.04	0.05	0.33	0.51	0.23	0.25
TiO₂	1.35	1.5	1.51	1.75	1.53	1.57	1.53	1.42	1.48	1.45
P₂O₅	0.15	0.13	0.11	0.11	0.13	0.13	0.14	0.12	0.14	0.13
MnO	0.18	0.15	0.2	0.18	0.2	0.18	0.09	0.18	0.14	0.2
Cr₂O₃	0.044	0.026	0.029	0.027	0.03	0.042	0.018	0.022	0.012	0.016
LOI	3.1	4.6	5	5.6	4.6	3.1	5.4	2.5	4.7	3.8
Total	99.88	99.75	99.81	99.66	99.65	99.81	99.56	99.89	99.94	99.83
Sc	33	39	36	43	38	39	40	42	42	43
Ni	23.7	14.5	14.2	17.2	15.2	20.4	24.7	13.5	13.4	15
Pb	1.0	1.6	1.0	1.5	1.4	1.0	1.0	0.7	0.7	0.7
Ba	35.8	82.6	127.7	132.2	35	41.7	439.8	446.3	171.6	685.7
Co	31.5	32.9	29.4	38.6	30.6	31.8	41.7	39.4	35	39.3
Cs	0.1	0.3	0.2	0.4	<1	0.1	1.6	1.1	0.3	0.6
Ga	20.1	19.2	20.4	20.8	19.4	19.2	20.4	18	18.9	19.1
Hf	2.7	2	2	1.7	1.5	1.9	2.5	2.6	2.9	2.2
Nb	2.5	1.9	1.6	1.7	1.7	1.9	1.9	2	2.2	2.3
Rb	0.6	3.1	0.6	10.2	<5	1	12.5	10.5	4.5	5.7
Sr	340	298.9	263.9	269	328.5	369.7	854.4	407.9	310.9	648.4
Ta	0.1	0.1	0.1	0.1	<1	0.1	0.1	0.1	0.2	0.2
Th	1.4	0.6	0.7	0.5	0.8	0.7	0.5	0.2	0.4	0.4
U	0.3	0.4	1.2	0.3	0.3	0.2	0.1	<1	0.1	0.1
V	406	473	446	623	435	421	327	353	348	333
W	0.4	0.1	0.4	0.2	1.3	0.6	0.2	0.4	0.3	0.6
Zr	79.9	64.5	57.2	51.8	56	54.9	82.5	85.9	93.3	81.5
Y	33.8	23.2	25.2	23.6	25	26.2	27.6	30.1	32.6	28.1
La	5.9	4.9	4	3.6	3.8	4.1	3	4.2	5.3	4.4
Ce	14.3	11.8	10	9	9.4	9.7	9.3	13.7	14	12.8
Pr	2.04	1.7	1.53	1.4	1.37	1.47	1.48	2.03	2.15	1.95
Nd	10.3	9.3	8.7	7.3	7.7	8	9.7	9.9	12.1	11
Sm	3.4	2.8	3.1	2.5	2.8	2.8	3.1	3.5	3.8	3.6
Eu	1.34	1.07	1.1	1.01	1.08	1.18	1.19	1.53	1.55	1.25
Gd	4.17	3.26	3.48	3.08	3.4	3.53	3.97	4.44	4.41	4.35
Tb	0.89	0.6	0.68	0.63	0.69	0.65	0.8	0.92	0.81	0.79
Dy	5.49	4.12	3.77	4.01	4.09	4.18	4.65	4.95	5.38	4.99
Ho	1.3	0.95	1	0.95	0.95	0.96	1.05	1.06	1.15	1
Er	3.58	2.63	2.66	2.55	2.5	2.7	2.84	3.13	3.64	3.17
Tm	0.49	0.39	0.43	0.4	0.39	0.43	0.48	0.45	0.53	0.44
Yb	3.07	2.67	2.63	2.33	2.35	2.31	2.85	3.01	3.09	2.7
Lu	0.47	0.39	0.43	0.36	0.39	0.42	0.44	0.45	0.49	0.46

Fractional crystallization and magma source

The compatible elements (e.g. Ni, Cr and V) tend to retain in the residual solids during magma crystallization (e.g. Wilson 1989). They are useful indicators as their concentrations are controlled by fractionation of Fe–Mg minerals such as olivine and clinopyroxene. Therefore, the

compatible trace element contents of the studied rocks reflect fractional crystallization of observed mafic phenocryst phases.

When Zr/Y ratios of the Hasanoglan, Kadirler and Imrahor basalts are taken into consideration, it is observed that Zr/Y values increase with an increase in Zr at the same time (Fig. 9b). The Zr/Y ratios, if small degrees of melting

Table 1 continued

	Imrahor (gabbro & ultramafic)					Imrahor (basalt)				
	A-16	A-17	A-26	A-18	A-19	A-20	A-21	A-22	A-23	A-24
SiO₂	44.82	44.24	38.74	44.93	42.95	40.64	47.37	41.28	43.11	44.66
Al₂O₃	14.8	16.04	7.45	16.65	14.61	12	15.48	11.59	15.62	14.99
Fe₂O₃	14.96	11.34	13.07	10.35	11.69	12.01	13.18	12.11	11.95	9.67
MgO	6.69	9.17	23.66	8.03	10.84	8.11	5.4	9.13	8.46	4.74
CaO	7.82	9.18	4.17	8.55	8.98	13.5	5.05	12.61	6.59	12.63
Na₂O	3.57	2.91	0.11	2.89	2.51	2.51	3.37	2.99	3.39	3.94
K₂O	0.2	0.13	0.09	1.07	0.43	0.94	0.54	0.06	0.96	0.29
TiO₂	1.84	1.24	0.66	2.19	1.9	3.11	3.65	3.78	1.7	1.53
P₂O₅	0.12	0.08	0.05	0.24	0.2	0.6	0.45	0.67	0.14	0.14
MnO	0.16	0.13	0.15	0.14	0.14	0.12	0.13	0.1	0.14	0.11
Cr₂O₃	0.03	0.043	0.223	0.03	0.058	0.059	0.017	0.054	0.034	0.042
LOI	4.7	5.4	11.4	4.8	5.6	6.3	5	5.2	7.7	7.1
Total	99.72	99.92	99.90	99.89	99.93	99.93	99.64	99.59	99.81	99.86
Sc	28	31	19	32	34	22	31	23	35	31
Ni	79.8	131.6	933.2	59.7	155.6	196.2	45.1	191.7	95.8	85.5
Pb	0.6	0.2	0.2	0.7	0.8	1.4	1.6	1.3	0.5	0.5
Ba	119.3	76	18.8	182.9	91.5	722	413.2	1879.3	74.6	74.1
Co	54.5	51.7	110.7	37.6	47.2	50.7	46	56	48.5	40.6
Cs	0.4	1.1	2	0.4	0.4	0.7	0.6	0.4	0.3	0.1
Ga	20	17.6	8.7	22.5	18.8	20.4	26.4	20	20	19.7
Hf	2.1	1.7	0.8	3.6	2.8	4.2	6.1	4.9	2.1	2.2
Nb	8.4	5.6	4.9	28.3	22	49.8	31.9	63.8	12.2	11.8
Rb	2.8	2.2	5.5	18.3	7	13	15.2	1.6	8.7	3.4
Sr	360.2	292	131.8	308.2	313.4	526	418.3	452.3	171	230.2
Ta	0.5	0.3	0.3	1.7	1.3	2.9	1.9	3.9	0.7	0.7
Th	0.5	<.1	0.2	1.4	1.5	3.3	2.8	2.8	0.7	0.9
U	0.1	<.1	<.1	0.5	0.4	0.8	0.7	0.9	0.4	0.2
V	293	245	134	287	260	261	336	287	265	276
W	0.3	0.1	<.1	0.1	0.1	0.4	0.4	0.5	0.1	0.1
Zr	76.4	51.6	25.8	125.3	100.7	155.1	236.5	182.4	80.1	75
Y	24.1	17.9	7.6	25	21.8	25.5	36.5	28.3	20.5	19.5
La	6.2	4.3	3.5	21	17.3	32.1	25.6	35.8	7.8	7.5
Ce	15.3	10.5	7.6	42.1	34.2	65	58.4	73.9	19	16.5
Pr	2.23	1.59	0.99	5.11	4.05	8.11	7.62	9.24	2.46	2.25
Nd	11	8.2	4.4	22.8	19.2	36.1	36.1	40.6	11.6	10.6
Sm	3.6	2.6	1.2	5.5	4.5	8.4	8.6	9.8	3.8	3.1
Eu	1.33	1.02	0.45	1.9	1.59	2.86	2.87	3.17	1.21	1.19
Gd	3.93	2.87	1.4	4.97	4.2	7.44	8.04	8.48	3.65	3.43
Tb	0.76	0.56	0.22	0.86	0.76	1.09	1.42	1.3	0.64	0.65
Dy	4.25	3.05	1.44	4.34	3.84	5.48	7.24	6.16	3.64	3.44
Ho	0.93	0.64	0.27	0.89	0.79	0.91	1.44	1.03	0.82	0.71
Er	2.31	1.75	0.71	2.37	1.9	1.99	3.23	2.27	2.04	1.88
Tm	0.34	0.24	0.12	0.32	0.28	0.25	0.45	0.27	0.29	0.28
Yb	2.15	1.39	0.69	1.8	1.62	1.49	2.68	1.59	1.81	1.5
Lu	0.29	0.23	0.1	0.26	0.26	0.19	0.42	0.2	0.23	0.24

are considered, reflect relatively high values owing to higher incompatibility of Zr compared to Y in mantle phases (e.g. Pearce and Norry 1979; Pearce 1980; Nicholson and Latin 1992). Therefore, the Hasanoglan, Kadirler and Imrahor basalts could have been generated by varying degrees of melting. It must also be noted that amphibole fractionation may have accounted for a small increase in Zr/Y ratios.

Relatively high Zr/Y values of the samples (Zr/Y = 2.88–10.13) apart from the Imrahor dykes (Zr/Y = 2.10–2.99) may suggest their derivation from enriched mantle sources or possible contribution of a mantle plume (OIB, Zr/Y = 9.66, E-MORB = 3.32; Sun and McDonough 1989). The lower Zr/Nb values (Zr/Nb = 2.9–9.2) also support this suggestion. On the other hand, considerable high Zr/Nb values (Zr/Nb = 28.9–43.4) together with

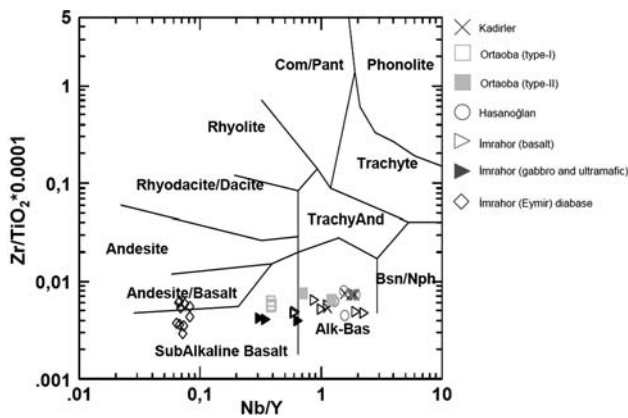


Fig. 8 Classification of the samples on the diagram based on relatively immobile elements (after Winchester and Floyd 1977)

relatively low Zr/Y ratios of the diabases may provide the contribution of depleted mantle during their generation (MORB, Zr/Y = 2.64, Zr/Nb = 31.8; Sun and McDonough 1989).

Y/Nb ratios can be useful at this point to further clarify that the generation of the Eymir dykes may have been associated with a depleted mantle source and the other samples under question are representative of more enriched sources. The resulting picture is that the alkaline samples are of the lowest Y/Nb values (Y/Nb = 0.44–1.39) among the investigated samples; however, variably tholeiitic samples (dykes excluded) are characterized by relatively high values of Y/Nb (1.55–3.20). The Eymir diabases, on the other hand, exhibit the highest ratios (Y/Nb = 12.21–

15.75). Thus, these results support the idea that the samples, except the diabases, may be related to the involvement of enriched mantle sources, but the case is opposite for the latter.

An important issue to be considered is whether the garnet is a residual phase in the source or not. The existence of residual garnet results in depletion of HREE relative to LREE because of the strong retention of HREE in garnet (e.g. Wilson 1989; Spath et al. 1996), as also indicated by high partition coefficients of those elements (e.g. Green et al. 1989; Jenner et al. 1994). The alkaline basalts display considerable depletion HREE compared to LREE ($[La/Yb]_N = 5.8–16.2$; Fig. 10), suggesting the presence of garnet as a residual phase in the source of these basalts. It must also be noted that $[Tb/Yb]_N$ values of the Kadirler and alkaline Ortaoba basalts ($[Tb/Yb]_N = 1.75–2.38$; $2.11–2.39$, respectively) are somewhat in agreement with those of alkaline basalts from Hawaii ($[Tb/Yb]_N = 1.89–2.45$; Frey et al. 1991), which are also thought to be derived from a garnet-bearing source. However, the Hasanoglan and alkaline Imrahor basalts ($[Tb/Yb]_N = 2.45–3.01$; $2.13–3.72$, respectively) do not seem to be consistent with those of Hawaii basalts. The relatively high Dy/Yb ratios displayed by alkaline basalts ($Dy/Yb = 2.09–3.87$) with respect to that of chondrite ($Dy/Yb = 1.49$; Sun and McDonough 1989), on the other hand, may suggest a garnet-bearing lherzolite source for the derivation of those alkaline basalts.

The variably tholeiitic samples (dykes excluded) are characterized by less fractionated patterns as reflected by

Fig. 9 Tectonomagmatic discrimination plots of the studied samples **a** after Shervais 1982, **b** after Pearce and Norry 1979, **c** after Meshede 1986. Fields: *A* and *AII*: WPB; *B*: E-MORB; *C*: WPT and VAB; *D*: N-MORB and VAB. **d** after Wood 1980. Fields: *A*: N-MORB; *B*: E-MORB; *C*: WPB; *D*: destructive plate margin basalts). For symbols see Fig. 8

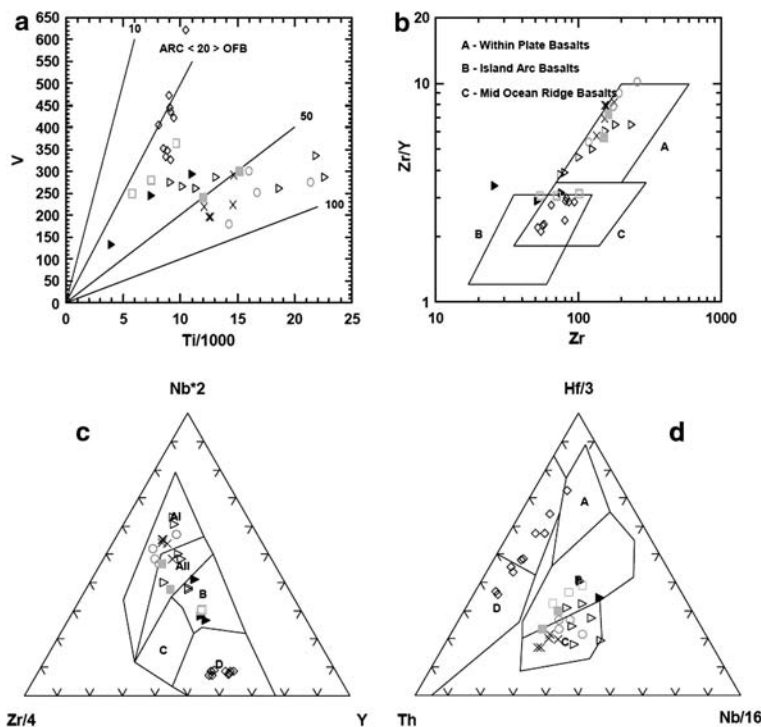
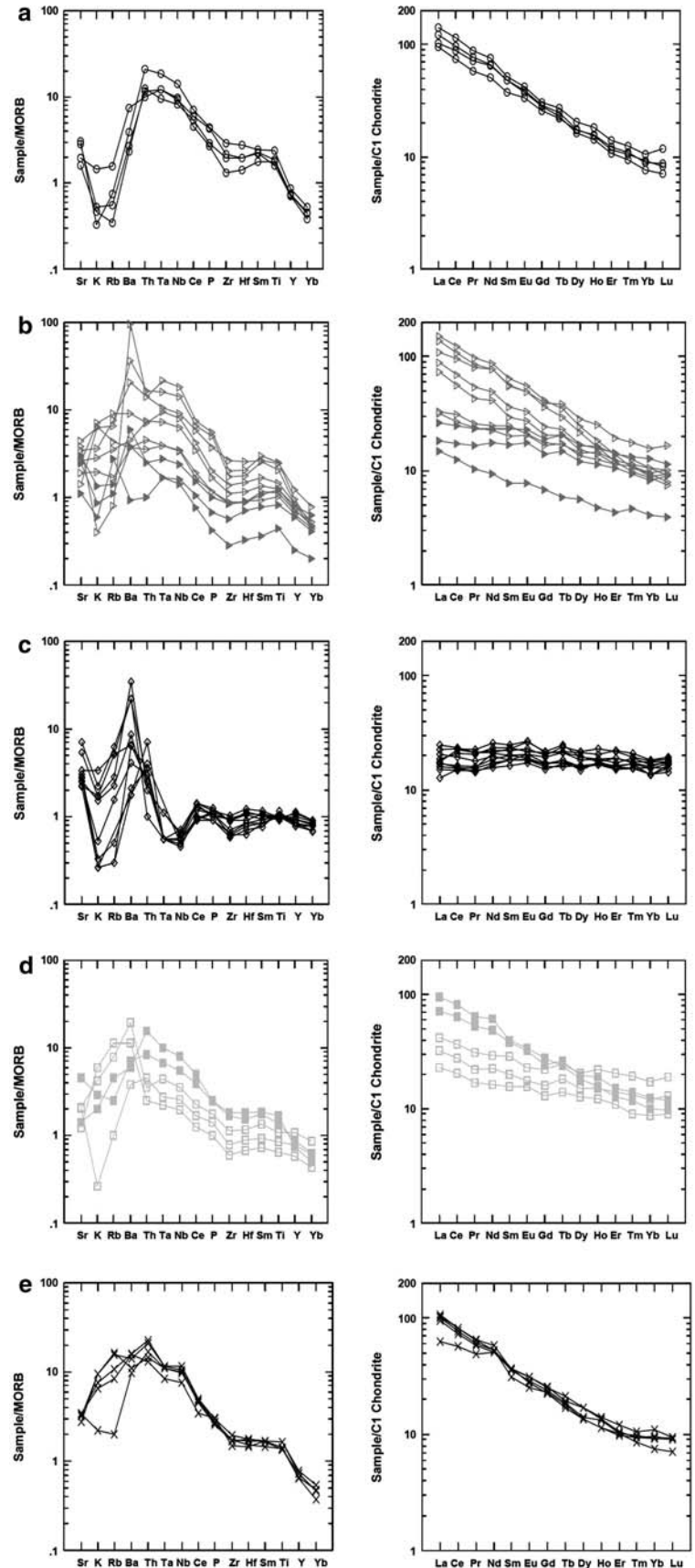


Fig. 10 MORB-normalized trace element patterns (normalization values from Pearce 1983) and C1 chondrite-normalized REE patterns (normalization values from Sun and McDonough 1989) for the studied samples. For symbols see Fig. 8



lower $[La/Yb]_N$ ratios (2.1–3.6) compared to the alkaline samples (Fig. 10). There are two options which can be considered probable at this point in order to explain the source of these relatively enriched tholeiitic samples; (a) an LREE-enriched garnet-bearing source where garnet is consumed in response to large degrees of melting, or (b) a source represented by relatively low LREE abundances with no garnet present. Consequently, the latter possibility, a spinel lherzolite source, seems more suitable to enlighten the genesis of variable tholeiitic samples other than the diabases.

Another subject to be discussed is whether the samples have experienced crustal contamination or not. At this point, attention should be paid to the possibility that some chemical parameters may have the same signature as supra-subduction zone (SSZ) effects. Basaltic rocks not influenced by crustal contamination display ratios; $La/Ta < 22$, $La/Nb < 1.5$ (e.g. Fitton et al. 1988; Leat et al. 1988; Hart et al. 1989). All the investigated samples apart from the diabases have low elemental ratios ($La/Ta = 9.18–15.40$; $La/Nb = 0.56–0.90$; $Th/Nb = 0.04–0.11$), indicating that crustal contamination might be insignificant for those samples. The diabases, on the other hand, exhibit considerably higher values ($La/Ta = 22.00–59.00$; $La/Nb = 1.58–2.58$; $Th/Nb = 0.10–0.56$), which could arise from crustal contamination or SSZ signature. High La/Nb values are also widely recognizable in arc mafic rocks (e.g. Gill 1981). The comparatively low Th/La ratios of the diabases ($Th/La = 0.05–0.24$), however, are consistent with the latter scheme, since those values are in harmony with those observed in intra-oceanic arcs ($Th/La = 0.09$, Mariana Arc; McCullough and Gamble 1991) which display no major crustal influence in contrast to continental margins ($Th/La > 45$, Aeolian Arc; Ellam et al. 1988).

Tectonomagmatic evaluation

Largely varying major element compositions reflected by the Imrahor and Ortaoba mafic rocks may suggest a non-uniform origin in terms of their genesis. This picture is especially apparent from the TiO_2 contents such that the Imrahor samples have TiO_2 (wt%) contents ranging between 0.66 and 3.78%, and the Ortaoba basalts between 0.97 and 2.53%. The alkaline Imrahor and Ortaoba basalts are represented by the contents varying between 1.90 and 3.78%, thus being comparable to OIB (>2 wt% TiO_2 ; e.g. Harris 1983). This suggestion is further supported by the tectonomagmatic discrimination diagrams, since these alkaline basalts are plotted in the within-plate field (Fig. 9). The humped trace element patterns ($Zr/Nb = 3.1–7.8$; Fig. 10) as well as fractionated REE variations ($[La/Yb]_N = 5.8–16.2$) may be further indicators in regard to their within-plate generation

(OIB-like). In the plot of Th/Yb vs. Nb/Yb , this feature can be clearly observed (Fig. 12).

Such OIB-like extrusives within the Karakaya Complex were also documented in the Ankara Group (Capan and Floyd 1985; Floyd 1993). In NW Anatolia the Nilüfer Unit also includes OIB-like basalts (Pickett and Robertson 1996, 2004; Yaliniz and Göncüoğlu 2002; Genç 2004). However, in contrast to Genç (2004), who considered all basic volcanic rocks within the Karakaya Complex as members of the Nilüfer Unit, there are several basalts, which are more akin to N-MORB and E-MORB rather than OIB in his samples.

The sub-alkaline samples from the Imrahor (apart from diabases) and Ortaoba are relatively enriched in terms of TiO_2 (wt%) contents varying between 0.97 and 1.84% compared to N-MORB ($TiO_2 = 1.15–1.31$; Schilling et al. 1983). These samples appear to be E-MORB tholeiites on the Nb–Zr–Y (Meshede 1986) and Hf–Th–Nb (Wood 1980) tectonomagmatic discrimination plots (Fig. 9). Their variably tholeiitic compositions together with E-MORB characteristics may suggest the existence of a spreading-center system with involvement of enriched mantle sources, such as a mantle plume, during their magmatic evolution (Figs. 9, 10). Indeed, the multi-element and REE variations show that they are not similar to N-MORB (Fig. 10). Their relative enrichment compared to N-MORB ($Zr/Yb = 6.4–9.2$) and LREE-enriched natures ($[La/Yb]_N = 2.1–3.6$) indicate that a non-depleted source should be concerned for their genesis. The presence of E-MORB-type

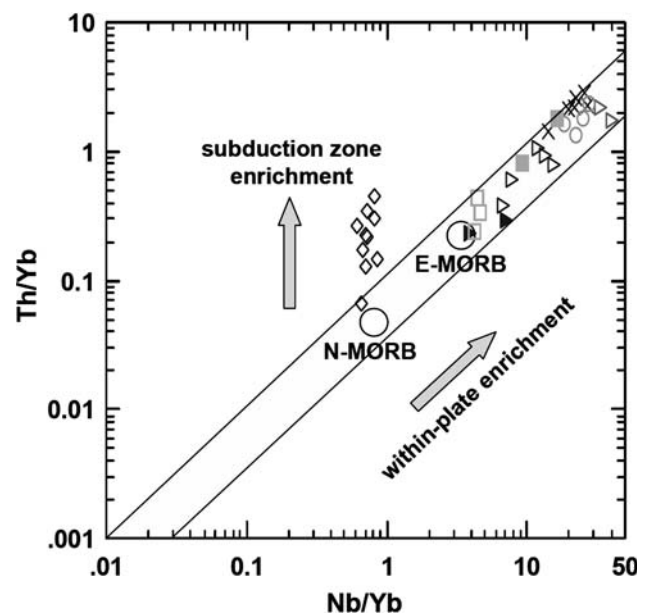


Fig. 11 Th/Yb vs. Nb/Yb diagram signifying mantle enrichment for OIB- and E-MORB-like samples and effect of subduction enrichment on the diabases (after Pearce and Peate 1995; N-MORB and E-MORB values from Sun and McDonough 1989). For symbols see Fig. 8

basalts within the Karakaya Complex in the Ankara region was also mentioned by Floyd (1993). Our study, however, reveals that the Ortaoba basalts are represented by both E-MORB-type (Type-I) rocks featuring the Ortaoba Unit and OIB-type basic volcanics (Type-II) characterizing the Nilüfer Unit (Sayit 2005). The same basalts from the Ortaoba Unit were regarded to have N-MORB characteristics by Pickett and Robertson (1996).

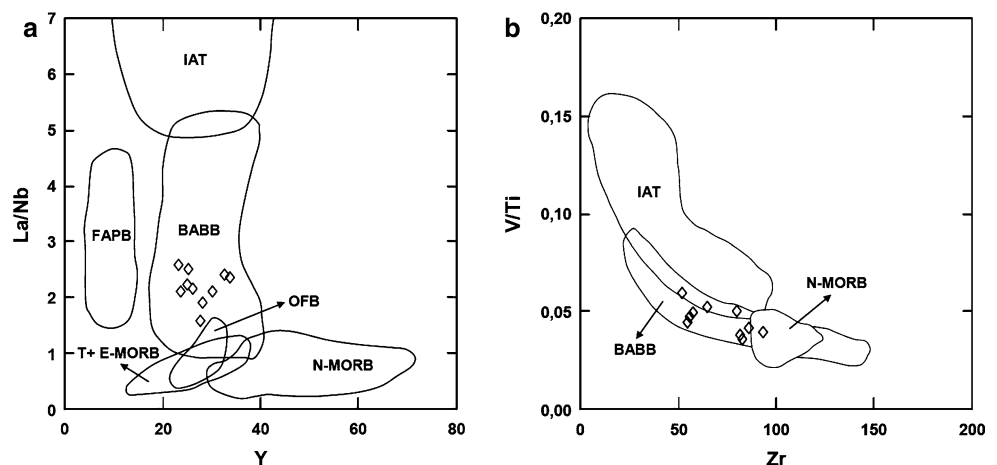
In comparison to the Imrahor and Ortaoba extrusives, those from Kadirler and Hasanoglan lavas, which have been first reported in this study, appear to have been derived from single tectonic setting, as indicated by the restricted TiO_2 (wt%) contents with 2.38–3.56% and 2.01–2.79%, respectively. These TiO_2 values are similar to those of OIB at the first sight. Indeed, the discrimination diagrams and multi-element variations suggest that they are akin to alkaline within-plate basalts with OIB signature (Figs. 9–11). This is also evidenced by the REE patterns with varying degrees of enrichment of LREE relative to HREE ($[\text{La}/\text{Yb}]_N = 5.8\text{--}16.1$).

The diabase dykes of Eymir Complex, in contrast to the rest of samples, are very unlike in terms of their depleted source characteristics. Their TiO_2 (wt%) contents seem to be confined to a small field (1.35–1.75), thus suggesting a uniform origin. Even though this compositional interval is somewhat comparable to that of E-MORB, in the tectonomagmatic discrimination diagrams the diabases fall into both MORB and IAB fields (Fig. 10). On the other hand, the multi-element patterns with apparent negative Nb anomaly ($\text{Zr}/\text{Nb} = 28.9\text{--}43.4$; Fig. 10) indicate that they might have been generated above a supra-subduction zone. The geochemical features of the diabase dykes within the Eymir Complex have been documented for the first time in this study and they are similar to the Küre basalts, which are also of SSZ-related character (Ustaömer and Robertson 1994). The important issue to be considered at this point is

whether the dyke samples are of IAB, CAB or BABB origin.

When compared with MORB and OIB, island-arc basalts (IAB) are characterized by higher abundances of LREE and LIL elements relative to HFS elements (e.g. Saunders et al. 1980; Gamble et al. 1993; Thirlwall et al. 1994; Pearce and Peate 1995), which is considered to be attributable to re-fertilization of the mantle wedge by slab-derived fluids (e.g. Gill 1981; Arculus and Powell 1986; Pearce and Parkinson 1993; Arculus 1994). The effect of subduction enrichment on the diabases is clearly reflected by Th/Yb and Nb/Yb ratios (Fig. 11). There is a general consensus that high LIL/HFSE ratio is a signature of subduction-related events (Arculus and Powell 1986; Hawkesworth et al. 1993; Brenan et al. 1994). This island-arc signature, indeed, is very evident on the multi-element variation plot with variably enriched LIL element abundances followed by a nearly flat pattern (Fig. 10). The relatively flat REE patterns may also be encouraging with respect to their subduction-related origin (Fig. 10). On the other hand, their transitional nature between MORB and IAB may suggest that they have been generated in supra-subduction extensional regimes, that is, back-arc basin (e.g. Stern et al. 1990; Ewart et al. 1994; Pearce et al. 1995; Petford and Atherton 1995). The geochemical compositions of back-arc basins are widely variable such that it is sometimes impossible to distinguish them from arc basalts with distinctive subduction signatures or from basalts with little or no evidence of subduction (e.g. Hochstaedter et al. 1990; Ewart et al. 1998; Gribble et al. 1998). It must be noted that flat REE patterns are recognizable in some back-arc basin basalts (e.g. Leat et al. 2000; Corcoran and Dostal 2001; Fretzdorff et al. 2002). These flat patterns may further signify the absence of residual garnet from the source, thus indicating melting depths shallower than those dominated by garnet-lherzolite (e.g. Arndt et al. 1993; Dostal

Fig. 12 La/Nb vs. Y plot (after Floyd et al. 1991) and V/Ti vs. Zr (after Woodhead et al. 1993) that implies the BABB signature of the Eymir diabase dykes



et al. 2001). The TiO_2 (wt%) compositions are also consistent with those of BABB (generally $\text{TiO}_2 > 1\%$; e.g. Pearce et al. 1995; Gribble et al. 1998; Leat et al. 2000; Fretzdorff et al. 2002; Leat et al. 2004) rather than those of IAB (generally $\text{TiO}_2 < 1$; e.g. Peate et al. 1997; Woodhead et al. 1998).

A further evidence for the back-arc origin of the Eymir dykes comes from La/Nb vs. Y and V/Ti vs. Zr plots (Fig. 12). The diabase samples are confined well within the area, which characterizes BABB and they have clearly lower La/Nb and V/Ti values (at constant Zr) than those displayed by IAT (Floyd et al. 1991; Woodhead et al. 1993). The back-arc basin signature of the diabase dykes, therefore, suggests that they should have been generated during or after the accretion event (syn- or post accretion), thus representing initiation of back-arc spreading.

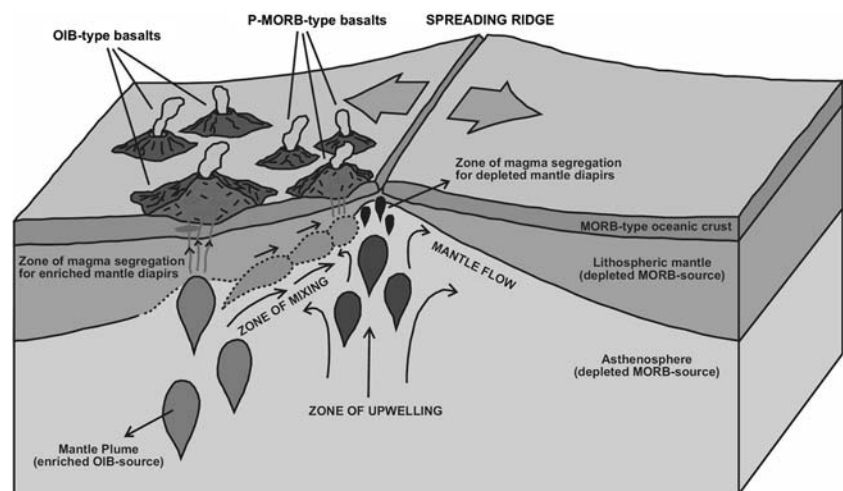
To summarize, the studied samples from the Karakaya Complex constitute two compositional groups, namely alkaline basalts and variably tholeiitic basalts. The Kadirler and Hasanoglan basalts are characterized by OIB-like signatures with substantial depletion in terms of HREE relative to LREE, suggesting their derivation from a garnet-bearing source. Similarly, the alkaline Imrahor and Ortaoba samples are also akin to OIB generated from a garnet-lherzolite source as indicated by variable enrichment with respect to most of the elements, while variably tholeiitic ones are of enriched nature but more depleted relative to the former, suggesting E-MORB affinity, probably derived from a shallower spinel-lherzolite source devoid of garnet phase. The diabase dykes, however, are distinctively different from the rest as revealed at the first glance by the transitional characteristics between MORB and IAB. Their transitional nature together with considerable negative Nb anomaly leads to the suggestion that they should have been generated in supra-subduction zone, probably in a back-arc basin with the involvement of a depleted mantle source where no residual garnet is present.

Constraints for geological evolution

The Olukman Formation, which is characterized by considerable amount of clastic rocks of turbiditic origin, probably reflects a tectonically unstable environment as deduced from rapid changes in lithology and the presence of olistostromal parts. Neritic limestone clasts embedded in a sandy matrix should possibly have been associated with gravity flows along a slope. Furthermore, the presence of a small lens of Bahcecik Formation consisting of pillow basalts within the Olukman Formation may suggest that pillow basalts together with syn-sedimentary limestones have been incorporated into the clastics by mass flows. Indeed, if the field relationships between two units in Imrahor and Hasanoglan are taken into consideration, it can be proposed that the Bahcecik Formation represents huge megablocks within the Olukman Formation. On the other hand, the intense deformation and shearing should be linked with a later tectonic event.

In contrast to the significant content of clastic rocks within the Olukman Unit, the Bahcecik Formation is dominated by basalt flows interlayered with both pelagic and neritic limestones. The presence of both OIB- and E-MORB-type rocks within the unit indicates the change in type of volcanism during the evolutionary history. As there are no indications for the geochemical involvement of a continental crust in the Bahcecik lavas we suggest that the magmatic evolution is mainly within an oceanic plate. The suggested picture is that within a pre-existing MORB-type oceanic crust the initiation of an extensional regime has resulted in the intrusion of asthenospheric magmas probably associated with a mantle plume (or hot spot) nearby the ridge axis. Pen-contemporaneous involvement of the deeply placed mantle plume most likely led to the MORB-type magmas becoming relatively enriched, producing enriched-MORB (E-MORB) magmas. Thus, the most enriched melts (OIB-type) should likely have been derived from the

Fig. 13 Middle–Upper Triassic reconstruction of the Palaeoethys influenced by a fixed hot-spot induced volcanism which gives way to generation of OIB- and E-MORB-type melts



mantle plume itself, whereas E-MORB magmas have been produced at the shallower levels where mixing takes place between OIB- and N-MORB-type melts, leading to generation of relatively enriched E-MORBs (e.g. Hekinian et al. 1999; Fig. 13). The neritic limestones associated with the basalts, on the other hand, should have been deposited during local pauses in volcanism in the fringes of the oceanic islands. The pelagic sediments and turbidites, on the other hand, were formed on the slopes of the seamounts during the Middle Anisian.

The Eymir Complex probably characterizes a subduction/accretion complex as indicated by the presence of olistostromal sequences (greywackes, black slates and sandstones) and metabasic rocks with HP/LT minerals (Na-amphibole). The clasts in the metaclastic rocks include igneous rocks of felsic-intermediate compositions (granite, rhyolite, dacite, andesite) and detrital white mica and may have been derived from a volcanic arc. The intense shearing and deformation seem to have resulted from the accretion process during subduction. The diabase dykes cutting across the unit; however, appear to be related to a magnetic event after the subduction/accretion process, as they do not bear any HP/LT imprint. The geochemical signatures of the diabbases suggest that they have been generated above supra-subduction zone, presumably a back-arc basin. The presence of subduction-related volcanism within the Eymir Complex contrasts with the suggestion of Pickett and Robertson (2004), who considered the lack of arc-related volcanism as a contra-evidence for the back-arc setting of the Karakaya Basin as suggested by Sengör et al. (1984) and Göncüoğlu et al. (2000).

The Ortaoba Unit, which comprises deep-marine deposits and radiolarian cherts but no carbonates, most likely reflects a slope setting as deduced from the presence of volcanoclastics associated with E-MORB-type magmatism. This plume-related volcanism may be closely related with the magmatic events, which were processing during the tectonic evolution of the Bahcecik Formation.

Conclusions

The Anisian OIB- and E-MORB-type volcanism described from the Bahcecik Formation together with the Ladinian (?)-Carnian E-MORB-type magmatism found in the Ortaoba Unit suggests the presence of a mantle plume and an oceanic extensional system within the Palaeotethyan oceanic crust. Obviously, a considerable part of the Middle Triassic magmatism was influenced by this mantle plume. This led Okay (2000) to suggest a huge oceanic plateau (Nilüfer Volcanics) within the Palaeotethys. Based on a geochemical comparison with the better-known oceanic plateaus, Yaliniz and Göncüoğlu (2002) preferred a

number of seamounts. The recent finding of fossiliferous limestones and clastics associated with the OIB-type volcanics (Bahcecik Formation) indicate that these Middle Triassic seamounts had their own platform and slope deposits.

The Ortaoba Unit with volcanic rocks, deep-marine sediments (i.e. chert bands) with the absence of carbonates has most likely been formed within the Palaeotethyan basin mentioned above. Therefore, the recent Late Triassic age ascribed to the Ortaoba Unit as well as additional geochemical evidence for co-existing E- and N-MORB may support the existence of this basin.

The Eymir Complex represents a subduction/accretion prism with HP/LT metabasic rocks. The geochemical features and ages of these metabasic rocks are not yet known. Regarding the petrography of the associated clastic sediments, the complex includes arc-derived igneous material. On the other hand, an oceanic subduction system should have been installed during intrusion of the diabase dykes into the Eymir Complex, as indicated by the geochemical characteristics of the dykes. The dykes are BABB-type and definitely younger than the metamorphism of the subduction–accretion complex but older than the pre-Jurassic juxtaposition of the Karakaya units.

To conclude, the authors strongly emphasize that the Karakaya Complex is the product generated by involvement of a number of units from different tectonic settings (seamount, plume-related MOR and back-arc basin). The presence of all these units in close association with each other suggest that they should have been accreted and integrated within a tectonic melange probably during Latest Triassic (e.g. Tekeli 1981). By this, the present study does not support the Early Triassic continental-rift basin model (e.g. Bingöl et al. 1973; Sengör and Yılmaz 1981; Altınır and Kocyigit 1993; Genc and Yılmaz 1995) at least, for the studied tectonic units concerned.

The new working-model here involves the presence of the Palaeotethyan oceanic crust prior to middle Anisian. Whether it was opened in relation with the Eymir back-arc volcanism (this study) within Palaeotethys or as a marginal basin within a Permian/Lower Triassic platform (Sengör et al. 1984; Göncüoğlu et al. 2000; 2004) is not yet clear. During Anisian, the system was affected by a mantle plume with the development of an oceanic crust associated with seamounts. The influence of the mantle plume was also dominant during Late Triassic, and ridge spreading was also still active. However, the latest Triassic was the time when all the pieces of the huge oceanic realm, the Palaeotethys, were joined and incorporated within a subduction/accretion system, which is known as the Karakaya Complex.

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