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Fast Track Article

Geochemical and isotopic constraints on the genesis of the Late Palaeozoic Deliktaş and Sivrikaya granites from the Kastamonu granitoid belt (Central Pontides, Turkey)

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With 8 figures and 2 tables

Abstract: The Late Palaeozoic Sivrikaya and Deliktaş granitoids of the Kastamonu granitoid belt (KGB) are of sub-alkaline affinity, belong to the high-K calc-alkaline series and display features of transitional to S-type granites. Sivrikaya granitoid is host to biotite-hornblende granodiorite-tonalite and minor two-mica granites. The rocks are 303–300 Ma old, have low initial $^{87}Sr/^{86}Sr$ ratios (0.7041–0.708), moderately low $\varepsilon Nd_{(t)}$ values (–1 to –3.8) and young T_{DM} model ages (0.75 to 1.08 Ga). All these characteristics, combined with low $Al_2O_3/(FeO + MgO + TiO_2)$ and $(Na_2O + K_2O)/(FeO + MgO + TiO_2)$ and $\delta^{18}O$ values of 10–11.6‰ point to dehydration melting of heterogeneous protoliths dominated by amphibolite and greywackes-type sources with mantle contribution. Chondrite-normalized REE patterns of the Sivrikaya rocks are characterized by concave-upward patterns suggesting that amphibole played a more significant role than garnet during magma segregation. The main portion of the Deliktaş granitoid consists of peraluminous muscovite-rich monzogranite. Compared to Sivrikaya, rocks from this pluton have higher initial Sr ratios (0.7109–0.7185), older Nd model ages (1.2 to 2.2 Ga) and similar $\varepsilon Nd_{(t)}$ values (–2.0 to –4.7). U–Pb zircon analyses give an age range of 295–275 Ma. The nearly constant $\delta^{18}O$ values (~ 11.5 to 11.7‰) in conjunction with the chemical characteristics indicate a predominantly pelitic source similar to the basement, which consists of felsic high-grade granulite-facies metasedimentary rocks, of continental origin.

Key words: NW Turkey, Central Pontides, Kastamonu granitoid belt, Palaeotethys; Hercynian magmatism, Genesis, VAG, volcanic-arc granitoids, Sr-, Nd-, O-isotopes.

Introduction

The numerous plutons of the Kastamonu granitoid belt (KGB) (Yilmaz & Boztuğ 1986) and associated felsic volcanic rocks (Peccerillo & Taylor 1976, Ustaömer & Robertson 1993) in the Central Pontides, NW Turkey, are interpreted by some authors as resulting from the late orogenic-collisional tectonics and crustal thickening stage, linked to the subduction of the Palaeotethys oceanic crust beneath south Eurasian margin (Şengör 1980, Robertson & Dixon 1984, Yilmaz & Boztuğ 1986, Boztuğ et al. 1995). Alternatively, the granitoids are seen as related to southward subduction and closure of the Küre marginal basin (Ustaömer & Robertson 1997).

The mineralogy and petrology of rocks from some of these granitoids have been studied (e. g. Boztuğ et al. 1984, 1995, Yilmaz & Boztuğ 1986, Boztuğ & Yilmaz 1991, 1995). Based on limited data, the granitoids were interpreted as hybrid products genetically related by extensive differentiation of mantle-derived parent magmas variably contaminated by continental crust (e. g. Boztuğ et al. 1984, 1995). The genesis of these granitoids is still enigmatic because key data such as crystallization ages, geochemical and isotope analyses necessary to identify the type and origin of a granitoid are not available. This study focuses on the origin of the Deliktaş (DLG) and Sivrikaya (SG) granitoids in the Central Pontides (NW Turkey, Fig. 1), using geochemical, Nd-, Sr- and O-isotopic data, to further constrain their genesis and discuss the tectonic setting. The granitoids were emplaced at about 303–275 Ma into the Devrekanı metamorphic unit, during and after the collision of the Pontides and the south Eurasian margin, following the northward subduction of the Palaeotethys.

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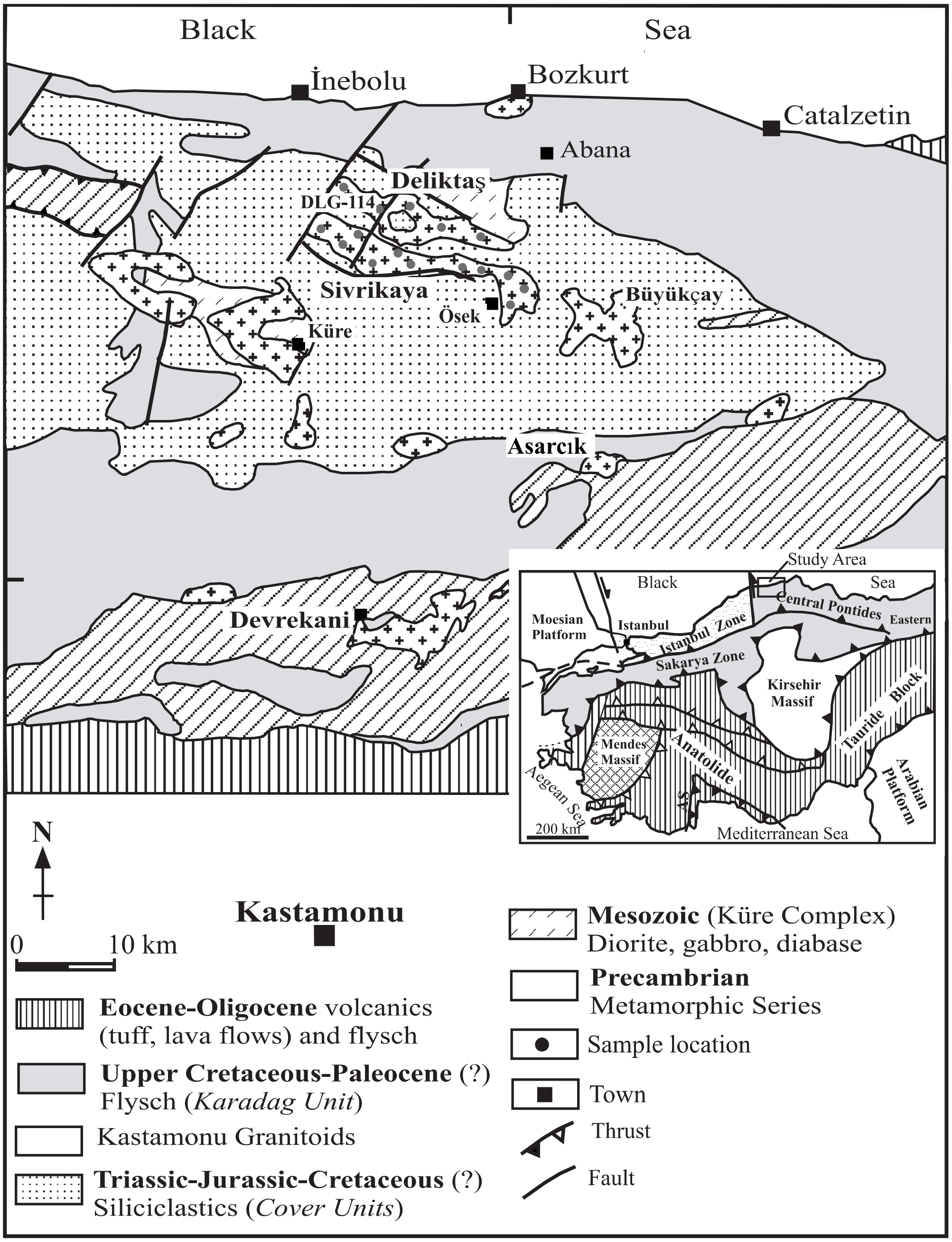

Fig. 1. Simplified map of study area (İnebolu-Sivrikaya) showing the Kastamonu granitoid belt (modified after Aydin et al. 1995). Inset shows main tectonic units of Turkey (modified after Okay & Tüysüz 1999). Sample localities are indicated.

Geologic setting and previous work

The Central Pontides include two terranes. The Western Pontides (the Istanbul Zone) are characterized by an Ordovician-Carboniferous sedimentary succession overlain by Triassic and younger rocks (e. g. Görür et al. 1997). The Eastern Pontides (the Sakarya Zone) comprise a Permo-Triassic subduction-accretion basement complex overlain by Triassic and younger sediments (e. g. Okay & Tüysüz 1999, Okay et al. 1997 and Okay & Göncüoğlu 2004). These units were tectonically juxtaposed during the Early Cretaceous (e. g. Tüysüz 1999). Both the onset of the subduction and the timing of the collision between the Pontides and the south Eurasian margin have been a matter of debate (e. g. Robertson & Dixon 1984, Şengör 1980, 1984, Şengör et al. 1980, 1984, Okay et al. 1996, 1997, 2002, Tüysüz 1990, Ustaömer & Robertson 1994, 1997, Yilmaz et al. 1997 and Robertson et al. 2004). Okay et al. (1994) and Şengör et al. (1980, 1984) thought the Palaeotethys had been located to the north of the Istanbul Zone. On the other hand, ROBERTSON & DIXON (1984) and OKAY (1989) have placed the Palaeotethys between the Istanbul and Sakarya zones and have left the Sakarya Zone as part of the Gondwana margin, in continuity with the Anatolide-Tauride block. According to OKAY et al. (1994, 1996) the Palaeotethys had been located to the south of the Sakarya Zone, leaving the Palaeozoic basement of the Sakarya Zone in possible continuity with the Moesia/Istanbul and Strandja zones (Fig. 1, inset), all presenting a post-Hercynian continental block along the southern margin of Laurasia.

The Central pontides are divided into four tectonic units: The Küre Complex in the north is a Triassic to Early-Mid Jurassic subduction-accretion complex of southward polarity related to the closure of the Palaeotethyan marginal basin (USTAÖMER & ROBERTSON 1993, 1997). The Çangaldağ Complex, in the south, is interpreted as a Late Palaeozoic south-facing oceanic-arc-trench complex (USTAÖMER & ROBERTSON 1997, USTAÖMER 1993). The Devrekanı metamorphic unit which constitutes the basement of the Central Pontides, is exposed between the Küre and the Çangaldağ Complexes. This high-grade metamorphic continental assemblage comprises gneisses and amphibolites at the base, overlain by metacarbonates of the Samatlar epicontinental group (e. g. USTAÖMER & ROBERTSON 1997). The Devrekanı metamorphic unit is inferred to have rifted from the south Eurasian margin, thereby producing the Küre marginal basin in the Late Permian (AYDIN et al. 1995, USTAÖMER & ROBERTSON 1993, 1994, 1997). Based on stratigraphic correlation with the Palaeozoic rocks of the Istanbul Zone (OKAY 1989, 2000, OKAY et al. 1994), both Precambrian (YILMAZ 1980) and Lower Palaeozoic ages (TÜYSÜZ 1990) were suggested for the Devrekanı metamorphic unit. The Domuzdağ-Saraycıkadağ Complex, further south, is interpreted as a Palaeozoic-Mesozoic Palaeotethyan subduction-accretion complex of northward polarity (e. g. TÜYSÜZ & YIĞITBAS 1994, USTAÖMER & ROBERTSON 1993).

On the basis of geochemical and mineralogical studies, Sivrikaya and Deliktaş granitoid rocks were classified as intermediate granite and leucoadamellite, respectively (e. g. BOZTUĞ et al. 1984, 1995). Sivrikaya samples were collected at new road cuttings and quarries in the Sivrikaya/Elmalıçay town and SG–186 a was collected at the Dıkmentürbet village quarry, close to Ösek town. Rocks of the Sivrikaya granite (SG) are medium- to coarse-grained hornblende-biotite granodiorite-tonalites and two-mica granites. The major mineral assemblage of this granitoid is quartz, K-feldspar (orthoclase-microcline), dark-green hornblende, brown biotite in mafic members, quartz and euhedral plagioclase (anorthite in the basic samples). Zircon, calcite, hematite, apatite and allanite are accessory phases. In some samples the plagioclase grains are twinned and faintly zoned. Quartz forms aggregates and displays undulatory extinction under polarized light. Locally, biotite and hornblende are partially chloritized and epidotized, and usually contain zircon inclusions. The muscovite in the leucocratic members is almost completely altered. K-feldspar exhibits a microperthitic to perthitic texture wrapped around the plagioclase. Sivrikaya samples are relatively poor in silica and alkalis, and rarely orthopyroxene occurs in the mafic samples. Deliktaş samples were collected from quarries and new roads, at a distance of ~ 500 m around the Deliktaş/Ahiçay village which is located on this pluton. All of the Deliktaş rocks are leucocratic, predominantly coarse-grained displaying porphyritic texture with K-feldspar and muscovite phenocrysts. In addition to quartz, K-feldspar (albite-orthoclase-microcline), muscovite, plagioclase and zircons, minor cordierite, calcite, biotite and apatite occur. Quartz occurs as anhedral crystals partially recrystallized as graphic quartz and myrmekite. K-feldspar is cream-white and occurs as anhedral to subhedral crystals, often displays microcline twinning. Plagioclase is generally sub-euhedral, faintly zoned, twinned and mostly sericitized. Muscovite is coarse-grained, idiomorphic and "apparently primary" (SAAVEDRA 1978). Rare biotite occurs as minute flakes scattered throughout the rock.

Both plutons intruded the Precambrian-Palaeozoic Devrekanı metamorphic unit, the Çangaldağ and Küre Complexes and the flyschoid country-rocks (YILMAZ 1980). Intrusion caused low grade contact metamorphism. The basement rocks are Palaeozoic-Precambrian (YILMAZ 1980, TÜYSÜZ 1990) as indicated by zircon ages (NZEGGE in prep). Zircons separated from rocks of both granitoids were dated by the Pb-Pb evaporation and U-Pb isotope dilution methods and yielded 303–300 Ma for Sivrikaya and 295–275 Ma for Deliktaş granites (NZEGGE et al. 2002, subm. 2006).

Analytical procedures

Major and trace elements were determined by XRF (Bruker AXS S4 Pioneer spectrometer) using standard techniques at Tübingen. Lithium borate fusion disks were used for major and trace element analyses. Total iron concentration is expressed as Fe_2O_3. Analytical uncertainties range from 1–8 % and 5–13 % for major and trace elements, respectively. The trace elements Cs, Th, U, Ta, Hf, Sc and Pb and REE were determined by inductively

Table 1. Major (wt %) and trace (ppm) element abundances of samples from the Deliktaş (DLG) and Sivrikaya (SG).

Sample	DLG-82	DLG-83	DLG-4	DLG-84	DLG-114*	SG-132	SG-124	SG-125	SG-276	SG-116	SG-115	SG-74	SG-88b	SG-72	SG-75a	SG-186a	SG-72a
SiO_2	77.08	77.38	77.64	78.51	79.77	68.70	69.00	69.40	69.80	69.90	71.20	74.90	75.00	75.10	75.90	76.00	76.07
TiO_2	0.04	0.04	0.04	0.03	0.07	0.39	0.39	0.34	0.15	0.36	0.23	0.34	0.07	0.19	0.34	0.04	0.58
Al_2O_3	14.30	13.72	13.18	13.94	11.40	16.17	16.13	15.81	14.47	15.89	15.77	14.52	14.33	011.8	13.68	14.46	15.15
Fe_2O_3	0.642	0.349	0.305	0.573	0.338	3.289	3.078	2.891	1.192	2.774	1.953	0.35	0.672	0.73	0.56	0.416	0.261
MnO	0.037	0.005	0.003	0.062	0.003	0.084	0.056	0.076	0.03	0.06	0.028	0.003	0.024	0.003	0.003	0.009	0.003
MgO	0.177	0.116	0.111	0.154	0.134	0.986	1.13	0.832	0.548	0.844	0.633	0.148	0.249	0.14	0.112	0.119	0.208
CaO	0.368	0.221	0.218	0.507	0.138	3.55	0.446	2.48	3.10	3.32	3.57	0.123	0.394	0.1	0.124	0.218	0.07
Na_2O	4.44	3.85	3.67	4.82	2.63	3.79	3.73	3.87	4.17	3.96	4.02	2.19	4.14	1.74	3.12	4.75	1.76
K_2O	3.67	5.16	5.16	2.14	4.66	1.54	3.36	3.01	3.04	1.94	1.80	5.48	3.98	4.55	5.28	3.43	4.87
P_2O_5	0.083	0.061	0.053	0.114	0.025	0.109	0.166	0.159	0.066	0.135	0.147	0.037	0.09	0.05	0.032	0.06	0.04
LOI	0.84	0.54	0.83	0.85	0.81	1.67	2.20	1.46	3.78	1.46	0.73	1.63	1.18	1.58	1.28	1.05	2.48
Sum	101.7	101.5	101.3	101.7	101.9	100.6	101.8	102.0	102.2	100.7	100.1	100.2	100.5	100.5	100.7	100.7	100.0
ASI	1.20	1.11	1.10	1.25	1.18	1.13	1.52	1.12	0.92	1.08	1.05	1.49	1.21	1.5	1.24	1.21	1.82
Cs	2.7	0.9	0.8	<0.1	1.4	3.2	<0.1	3.7	<0.1	<0.1	<0.1	<0.1	2.9	1.2	<0.1	<0.1	<0.1
Rb	82	124	132	138	148	49	124	100	87	69	66	254	133	25	272	118	207
Ba	306	430	341	56	418	457	629	872	990	488	491	287	448	558	318	185	334
Sr	75	71	66	53	68	393	164	363	254	345	330	52	100	161	69	55	74
Y	15	9	18	11	18	12	19	18	14	14	15	28	15	32.7	29	20.1	27
Zr	93	79	87	90	134	225	172	181	116	197	203	206	94	122	196	100	247
Pb	33	33	36	25	28	14	23	53	27	11	15	34	44	9.0	30	26.0	11
Zn	<0.1	<0.1	<0.1	<0.1	2.4	52	63	54	11	36	21	<0.1	0	18	<0.1	<0.1	<0.1
V	5	4	2	<0.1	4	29	37	31	16	28	16	19	2	50.9	14	4.2	39
Cr	2	6	2	6	2	46	7	6.3	52	33	58	12	5	24	29	17	27
Li	8	8	7	<0.1	13	25	<0.1	28	<0.1	<0.1	<0.1	<0.1	5.3	7.5	<0.1	<0.1	<0.1
Nb	15	19	11	13	15	10	16	21	<0.1	<0.1	<0.1	13	19	7.1	13	8.4	14
La	14	14	32	16	27	51	31	40	21	44	22	46	17	33	41	17.4	42
Ce	11.4	6.3	4.8	<0.1	42	64	62	71	34	62	0	85	13	43	80	<0.1	82
Pr	1.3	0.8	0.6	<0.1	5.2	6.2	<0.1	7.9	<0.1	<0.1	<0.1	<0.1	1.4	5.0	<0.1	<0.1	<0.1
Nd	14	23	4	14	25	21	29	30	10	30	14	30	5	20	42	9	37
Sm	3.0	4.9	0.9	3.9	5.2	1.3	5.5	5.6	3.1	4	<0.1	6	2.8	4.5	5	30	7
Eu	0.25	0.21	0.19	<0.1	0.27	1.15	0.70	1.20	0.70	1.00	0.30	0.70	0.31	0.78	0.20	<0.1	0.40
Yb	1.2	0.6	0.6	0.8	5.0	0.7	1.5	1.4	1.1	0.9	0.8	2.1	1.2	3.2	2.3	1.7	1.9
Th	3.7	0.9	1.4	15.1	21.8	11.9	13	14.6	10.9	12	5	23	6	6.3	23	8.1	18
U	5.2	1.3	2.1	1.3	1.2	4.8	<0.1	1.9	0.8	4	3	<0.1	15	0.9	<0.1	<0.1	<0.1
Co	<0.1	<0.1	<0.1	<0.1	<0.1	3.3	4	3.1	<0.1	3	1	<0.1	<0.1	7.5	<0.1	<0.1	<0.1
Ni	<0.1	<0.1	<0.1	<0.1	<0.1	15.6	<0.1	<0.1	10.7	2	16	<0.1	<0.1	6.4	<0.1	<0.1	<0.1
Gd	1.16	0.54	0.45	n.d.	4.13	1.66	n.d.	4.17	n.d.	n.d.	n.d.	n.d.	1.14	4.67	n.d.	n.d.	n.d.
Tb	0.22	0.10	0.10	n.d.	0.65	0.20	n.d.	0.55	n.d.	n.d.	n.d.	n.d.	0.22	0.75	n.d.	n.d.	n.d.
Dy	1.54	0.66	0.85	n.d.	3.95	1.02	n.d.	2.80	n.d.	n.d.	n.d.	n.d.	1.38	4.96	n.d.	n.d.	n.d.
Ho	0.30	0.14	0.15	n.d.	0.76	0.17	n.d.	0.45	n.d.	n.d.	n.d.	n.d.	0.27	1.08	n.d.	n.d.	n.d.
Er	1.00	0.50	0.48	n.d.	2.15	0.42	n.d.	1.12	n.d.	n.d.	n.d.	n.d.	0.79	3.30	n.d.	n.d.	n.d.
Tm	0.17	0.07	0.08	n.d.	0.32	0.06	n.d.	0.16	n.d.	n.d.	n.d.	n.d.	0.13	0.50	n.d.	n.d.	n.d.
Lu	0.18	0.06	0.08	n.d.	0.29	0.04	n.d.	0.12	n.d.	n.d.	n.d.	n.d.	0.15	0.49	n.d.	n.d.	n.d.
Hf	1.13	0.39	0.61	n.d.	1.16	0.70	n.d.	1.54	n.d.	n.d.	n.d.	n.d.	1.67	1.25	n.d.	n.d.	n.d.
Ta	3.34	0.43	0.21	n.d.	1.52	0.38	n.d.	1.73	n.d.	n.d.	n.d.	n.d.	4.89	0.37	n.d.	n.d.	n.d.
Eu/Eu*	0.31	1.19	0.92	n.d.	0.18	1.15	n.d.	0.75	n.d.	n.d.	n.d.	n.d.	0.42	0.53	n.d.	n.d.	n.d.
Nb/Ta	4.6	13.3	15.6	n.d.	13.4	19.4	n.d.	12.0	n.d.	n.d.	n.d.	n.d.	3.9	19.1	n.d.	n.d.	n.d.
$[La/Yb]_N$	7.6	15.9	35.5	n.d.	3.6	48.6	n.d.	19.1	n.d.	n.d.	n.d.	n.d.	9.5	6.8	n.d.	n.d.	n.d.
$[Gd/Yb]_N$	0.78	0.72	0.59	n.d.	0.66	1.9	n.d.	2.4	n.d.	n.d.	n.d.	n.d.	0.76	1.1	n.d.	n.d.	n.d.
Fe_2O_3/MgO	3.6	3.0	2.7	3.7	2.5	3.3	2.7	3.5	2.2	3.3	3.1	n.d.	2.7	5.2	5.0	3.5	1.3

* Pegmatite; n. d. not determines; ASI = Aluminium saturation Ondex (molar $Al_2O_3/(CaO+K_2O+Na_2O)$).

coupled plasma-mass spectrometry (ICP-MS) at the Memorial University of St. John's Newfoundland, using the sodium peroxide sinter technique. For detail analytical procedure, precision and accuracy of the data see Longerich et al. (1990) and Dostal et al. (1986, 1994).

For the determination of Sr and Nd isotopic ratios, whole-rock (WR) powders were spiked with mixed ^{84}Sr-^{87}Rb and ^{150}Nd-^{149}Sm tracer solution and digested in concentrated HF-HNO_3 in Teflon vials in PTFE bombs at 180 °C for 6 days. Digested samples were dried and redissolved in 2.5 N HCl. Rb, Sr, Sm, Nd and REE were separated by conventional cation exchange chromatography techniques. Rb, Sm and Nd were run on Re double filaments. Sr was run on W single filament. A detailed description of the analytical procedures is outlined in Hegner et al. (1995). Isotopic compositions were measured on a Finnigan MAT 262 multi-collector mass spectrometer at the University of Tübingen in data static collection mode. The Sr and Nd isotopic ratios were corrected for mass fractionation by normalizing to $^{86}Sr/^{88}Sr$

= 0.1194 and $^{146}Nd/^{144}Nd$ = 0.7219, respectively. Total procedural blanks were < 160 pg for Sr and < 80 pg for Nd. Replicate analyses of the NIST-987 Sr standard during the course of this study yielded a mean value of $^{87}Sr/^{86}Sr = 0.71026 \pm 10$ (2σ). Measurement of the Ames Nd standard preparation at Tübingen, within the same period, gave a mean $^{143}Nd/^{144}Nd$ ratio of 0.512129 ± 8. $^{87}Rb/^{86}Sr$ ratios for whole-rock samples were calculated based on the measured $^{87}Sr/^{86}Sr$ ratios and the Rb and Sr concentrations determined by XRF.

Oxygen was extracted from dried whole-rock powder at 550 °C using BrF_5 as a reagent following the technique of Clayton & Mayeda (1963). Oxygen yields were between 95 and 100 %. The Oxygen was converted to CO_2 using graphite rod heated by a Pt-coil. CO_2 was analysed for its $^{18}O/^{16}O$ ratios with a Finnigan Mat 252 gas source mass spectrometer. The isotopic ratios are reported in δ-notation relative to Vienna Standard Mean Ocean Water (V-SMOW). All analyses have been duplicated with analytical precision better than ± 0.1 ‰. Replicate analyses of the NBS-28 standard quartz were $\delta^{18}O_{(V\text{-}SMOW)}$ of + 9.7 ± 0.1 ‰ (2σ). All data have been normalized to NBS-28 = +9.7‰.

Results

Major and trace element geochemistry

Representative chemical analyses of samples are listed in Table 1. The bulk-rock concentrations of the Sivrikaya and Deliktaş granitoids are characterized by high SiO_2 and low MgO, and low abundances of high-field strength elements (HFSE) (Nb, Th, Ta, Zr and Hf). These features are commonly considered as characteristics of subduction-related (Floyd & Winchester 1975, Keppler 1996). In terms of normative mineralogy, the Sivrikaya granite has tonalitic-granodioritic to granitic compositions (Fig. 2 a). In contrast, the Deliktaş granite melt approach minimum melt compositions. The molar A/CNK vs. A/NK diagram (Maniar & Piccolli 1989) defines the rocks as slightly metaluminous to strongly peraluminous, and of transitional to S-type character (Fig. 2 b) (e. g Chappell 1999). Major and trace elements variations are illustrated in Harker diagrams in Fig. 3. The samples exhibit a narrow range of SiO_2 content, 68 to 79 wt % for the Sivrikaya and 77 to 78.5 wt % for Deliktaş (except for one pegmatite sample DLG-114, 79.8 wt %). MgO, Al_2O_3, Fe_2O_3 and CaO decrease slightly with increasing SiO_2, whereas K_2O remain nearly constant. SG rocks exhibit a cluster of data in two groups, the intermediate and felsic samples (Fig. 3 a–e). All samples are of

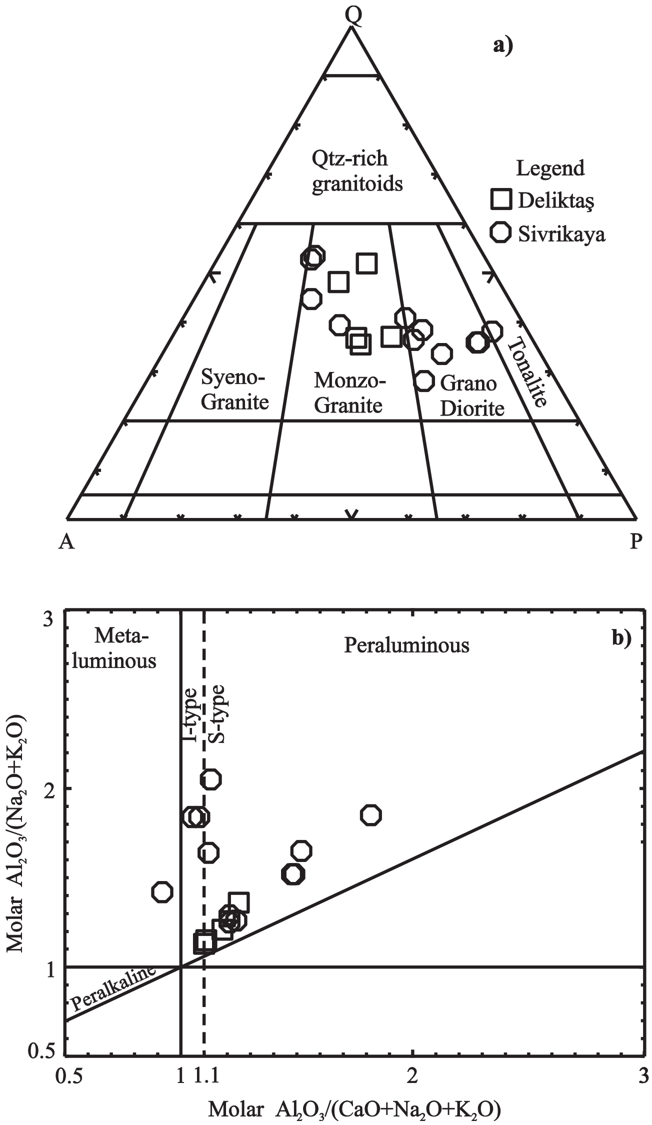

Fig. 2. **(a)** Ternary diagram illustrating the compositions of Deliktaş and Sivrikaya granitoids. Nomenclature taken from Le Maitre (1989): quartz (Q) – alkali feldspar (A) – plagioclase (P). **(b)** A plot of Shand's index for the Deliktaş and Sivrikaya granitoids. Discrimination for different types of granitoids (Maniar & Piccolli 1989), Shand (1927) are shown.

sub-alkaline affinity, belong to the calc-alkaline series and plot in the volcanic arc granitoids field of the Pearce et al. (1984) diagram (Fig. 4). The K_2O vs. SiO_2 plot shows Deliktaş samples to be dominantly of high-K affiliation (Fig. 3 b). Sr shows two groups of rocks for SG, but rather constant contents in the DLG (Fig. 3 f). Although the rocks of both granites exhibit typical high-K, calc-alkaline compositions, the variation diagrams reveal some differences between them. The Deliktaş rocks exhibit a higher and smaller range in SiO_2 content. Among the trace elements, samples of Sivrikaya show a

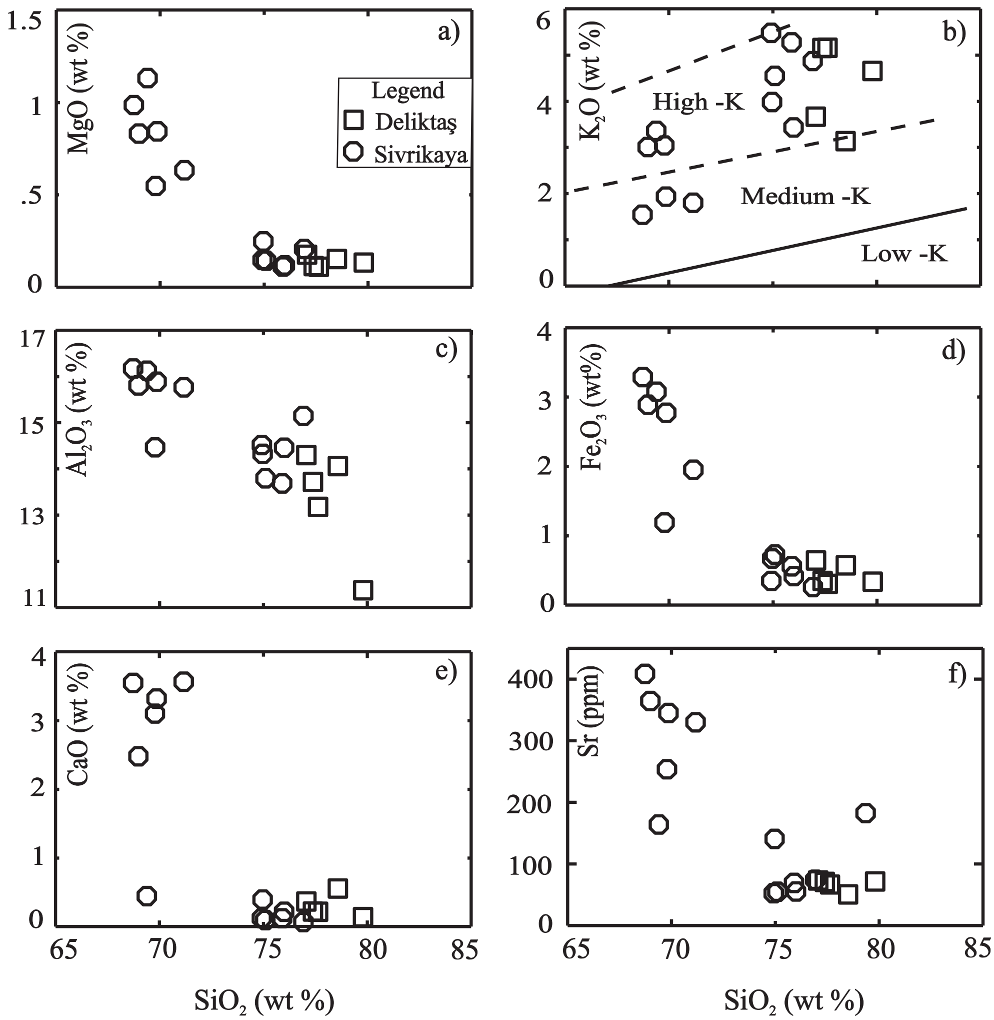

Fig. 3. Selected Harker variation diagrams of major **(a–e)** and trace elements **(f)** for Deliktaş and Sivrikaya Granitoids. The K_2O vs. SiO_2 diagram (Fig. 3b). shows field boundaries between medium-K (normal calc-alkaline), high-K series of Peccerillo & Taylor (1976).

much larger range in concentration than those from Deliktaş granitoid.

Rare earth elements (REE)

Chondrite-normalized REE patterns are plotted in Fig. 5. The patterns of all samples from both plutons are characterized by fractionation of the light and heavy REEs. The Sivrikaya samples are characterized by moderately fractionated REE patterns ($[La/Yb]_N$ = 6.8–49) and flat to concave-upwards HREE ($[Gd/Yb]_N$ = 0.76–2.37) patterns, and have mostly very small Eu-anomalies (Eu/Eu* = 0.42–1.15). The Deliktaş samples exhibit less fractionated REE patterns ($[La/Yb]_N$ = 3.6–35.5), flatter HREE ($[Gd/Yb]_N$ = 0.59–0.78), and have positive to negative Eu anomalies (Eu/Eu* = 0.18–1.19). The pegmatite sample, DLG–114 has the highest absolute REE abundances but the lowest $[La/Yb]_N$ ratio, which show slight enrichment of middle REEs (Gd to Er) relative to the other HREEs. Primitive mantle-normalized trace elements show enrichment in large ion lithophile elements (LILE) (e. g. Cs, Rb, Th, K and U) and SG samples exhibit troughs for HFSE (Nb and Ti) (Fig. 6). Whereas, DLG samples virtually have no negative Nb anomalies but more distinct Ti depletion. Negative Ti anomaly can be attributed to fractional crystallization of accessory phases, e. g. allanite and titanite. Some samples show decoupling of Ba and Sr from Rb and K as indicated by the troughs for Ba and Sr.

Sr-, Nd-, O-isotopic ratios

The Nd, Sr and O isotope data are presented in Table 2. Calculated initial Sr and Nd isotopic compositions are

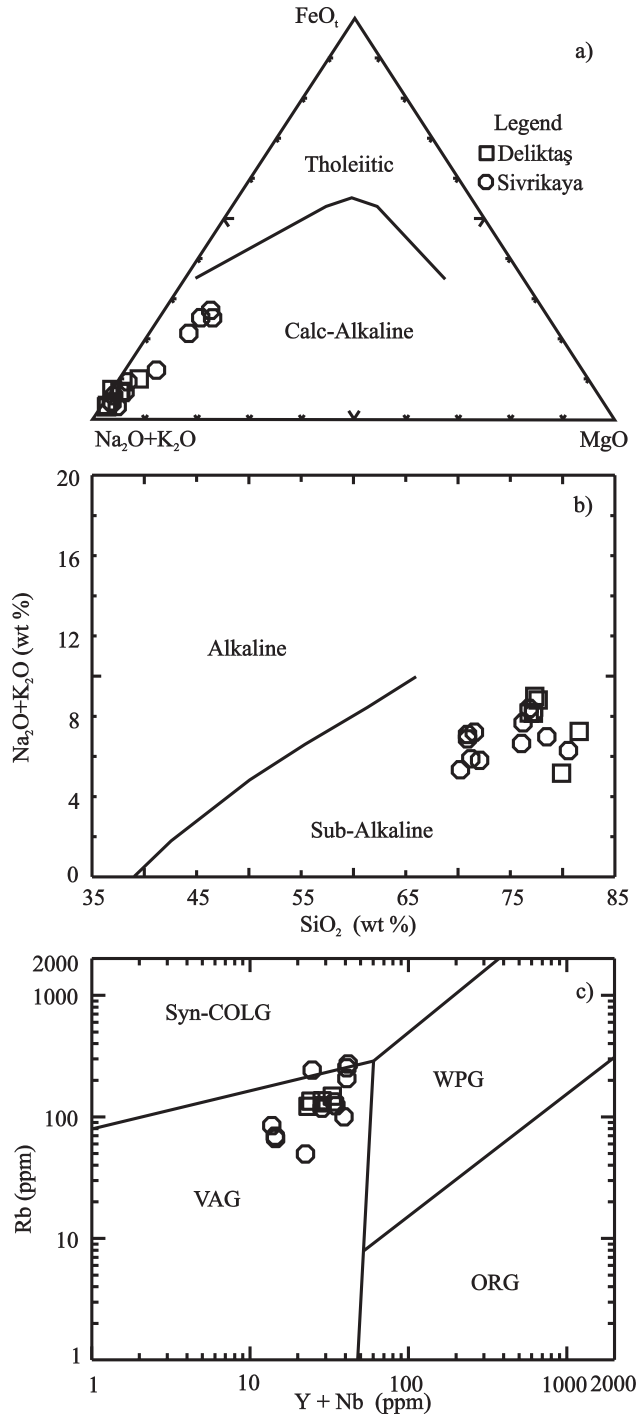

Fig. 4. Geochemical compositions of Deliktaş and Sivrikaya granitoids plotted in the Rb vs. Y + Nb discrimination diagram (PEARCE et al. 1984).

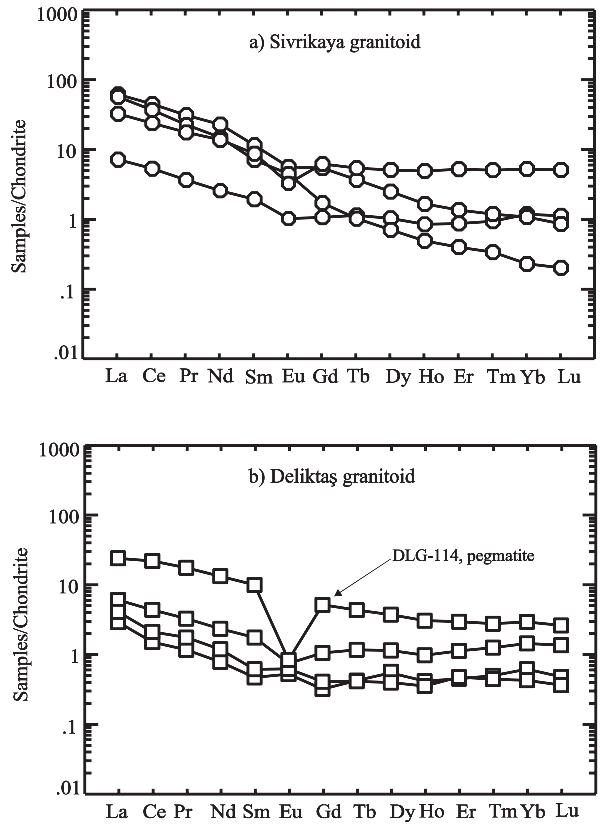

Fig. 5. Chondrite-normalized rare earth element abundances for Deliktaş and Sivrikaya granitoids. Normalizing values are from TAYLOR & MCLENNAN (1985).

based on U–Pb and Pb–Pb zircon ages of 300 Ma for Sivrikaya and 290 Ma for Deliktaş (NZEGGE et al. subm.). Fig. 7a shows the variation of initial $^{143}Nd/^{144}Nd$ expressed as $\varepsilon Nd_{(t)}$ values with initial $^{87}Sr/^{86}Sr$ ($Sr_{(i)}$) isotopic ratios. For Deliktaş samples $\varepsilon Nd_{(t)}$ values decrease with increasing $Sr_{(i)}$. Sivrikaya samples have similar $\varepsilon Nd_{(t)}$ but lower $Sr_{(i)}$ values (Fig. 7a, b).

The $\delta^{18}O$ values of the Deliktaş samples are high with a narrow range from 11.5 to 12.1‰ for whole-rocks, restricted 13‰ for quartz and 11.5‰ for K-feldspar. These values are typical of intracrustal melts or "S-type" granites (e. g. TAYLOR 1968, O'NEIL & CHAPPELL 1977). There is a slight positive correlation between $\delta^{18}O$ and $Sr_{(i)}$ and SiO_2 values indicated for the Deliktaş samples (Fig. 7 b, c). Sivrikaya whole-rock $\delta^{18}O$ values are slightly lower and more variable in the range of 10.0 to 11.6‰, ~12‰ for quartz and ~6‰ for biotite. The wider range in $\delta^{18}O$ values of the Sivrikaya whole-rock samples, reflect a heterogeneous source.

Discussion

Magma genesis and possible sources

WATSON & HARRISON (1982), WYLLIE (1984) and TEPPER et al. (1993) reported that partial melting of lower crustal metabasalt under variable H_2O conditions could

Table 2. Sm–Nd, Rb–Sr and O isotopic data of the Deliktaş and Sivrikaya granitoids.

Samples	Sm	Nd	$[^{143}Nd/^{144}Nd]m$	$[^{147}Sm/^{144}Nd]m$	$[^{143}Nd/^{144}Nd]i$	$\varepsilon Nd_{(t)}$	T_{DM} (Ga)	Rb	Sr	$[^{87}Rb/^{86}Sr]m$	$[^{87}Sr/^{86}Sr]m$	$[^{87}Sr/^{86}Sr]i$	$\delta^{18}O_{(VSMOW)}$ WR	Qtz	Kfs	Bt
Deliktaş																
DLG-4	0.9	4.1	0.512349±09	0.1285	0.512105	-2.8	1.23	136	67	5.889	0.73554 ±10	0.71124	11.6	13.3	11.5	n.d.
DLG-82	3.0	13.9	0.512394±09	0.1312	0.512394	-2.0	1.28	142	73	5.720	0.73261 ±12	0.71101	11.5	13.1	11.6	n.d.
DLG-83	4.9	23.3	0.512339±11	0.1262	0.512099	-2.9	1.22	127	71	5.779	0.73479 ±12	0.71095	11.7	13.1	11.5	n.d.
DLG-84	3.9	14.0	0.512312±10	0.1607	0.512007	-4.7	2.15	138	53	7.563	0.74231 ±10	0.71704	11.7	n.d.	n.d.	n.d.
DLG-114	15.2	41.9	0.512369±09	0.1663	0.512053	-3.8	2.08	150	75	5.775	0.74825 ±10	0.71848	12.1	13.2	n.d.	n.d.
Sivrikaya																
SG-88b	4.4	16.4	0.512492±09	0.2197	0.512058	-3.7	1.08	135	141	3.912	0.72460 ±11	0.70790	11.6	n.d.	n.d.	n.d.
SG-124	5.5	29.4	0.512300±09	0.0773	0.512147	-2.0	0.83	123	164	2.185	0.71423 ±10	0.70451	11.1	12.2	n.d.	6.2
SG-125	5.9	33.2	0.512282±09	0.1147	0.512056	-3.8	1.08	100	382	0.750	0.71103 ±11	0.70783	10.2	11.9	n.d.	n.d.
SG-132	5.7	31.5	0.512335±10	0.0689	0.512199	-1.0	0.75	50	409	0.342	0.70840 ±10	0.70694	10.8	12.3	n.d.	6.1
SG-186a	9.1	29.5	0.512529±10	0.1865	0.512162	-1.7	1.00	118	55	5.884	0.73246 ±10	0.70642	11.4	n.d.	n.d.	n.d.
SG-276	3.1	10.4	n.d.	n.d.	n.d.	n.d.	n.d.	87	254	1.545	0.71070 ±10	0.70408	12.4	n.d.	n.d.	n.d.

m = measured isotopic ratios; i = calculated initial isotopic ratios; $\varepsilon Nd_{(t)}$ values were calculated using present day $(^{143}Nd/^{144}Nd)_{CHUR}$ = 0.512638 and $(^{147}Sm/^{144}Nd)_{CHUR}$; CHUR = Chondrite uniform reservoir; $\lambda = 6.54.\ 10^{-12}\ a^{-1}$ (Steiger & Jager, 1977). The ages of 290 Ma (Delkitaş) and 300 Ma (Sivrikaya) are used for the $\varepsilon Nd_{(t)}$ and $Sr_{(i)}$ calculations. $[Sr_{(i)} = (^{87}Sr/^{86}Sr)i = (^{87}Sr/^{86}Sr)m - ^{87}Rb/^{86}Sr(e^{\lambda t}-1)$, $\lambda = 1.42.\ 10^{-11}\ a^{-1}]$. n.d., not determined.

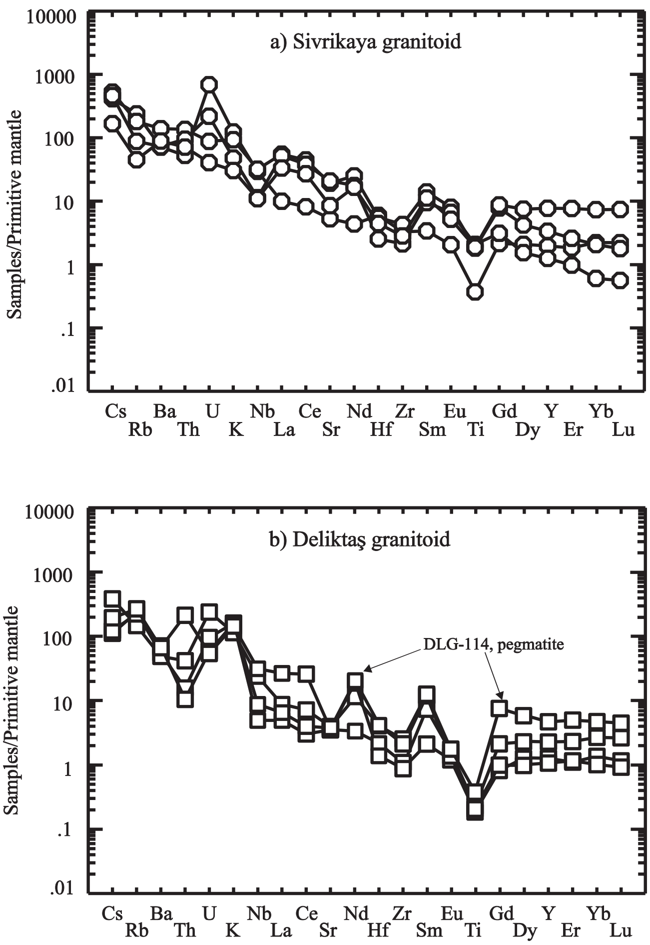

Fig. 6. Primitive mantle-normalized trace element abundances for the Deliktaş and Sivrikaya granitoids. Normalizing values are from Taylor & McLennan (1985).

yield a variety of granitoid rocks. Similarly, Roberts & Clemens (1993), based on the results of experiments on partial melting of crustal rocks stated that medium- to high-K, transitional- to S-type calc-alkaline granitoid magmas could be derived from partial melting of intermediate to felsic metamorphic rocks of the crust. The necessary heat for partial melting of the crust can be provided by repetitive underplating with basaltic magmas (e. g. Wolf & Wyllie 1994, Petford & Gallagher 2001, Annen & Sparks 2002, Borg & Clynne 1998, Bullen & Clynne 1990, Tepper et al. 1993, Guffanti et al. 1996).

Compositional diversity among crustal magmas may arise from melt fractionation, different source compositions, variable H_2O-activity, pressure, temperature and oxygen fugacity (e. g. Vielzeuf & Holloway 1988, Wolf & Wyllie 1994, Patiño Douce 1996/1999, Thompson 1996, Borg & Clynne 1998). Compositional differences of magmas produced by partial melting of different crustal source rocks such as amphibolites, gneisses, metagreywackes and metapelites, may be visualized in terms of major oxides ratios. Partial melts originating from mafic source rocks, for example, have lower $Al_2O_3/(FeO_t + MgO + TiO_2)$ and $(Na_2O + K_2O)/(FeO_t + MgO + TiO_2)$ than those derived from felsic metapelites.

Sivrikaya rocks have moderate $Al_2O_3/(FeO_t + MgO + TiO_2)$, $(Na_2O + K_2O)/(FeO_t + MgO + TiO_2)$ and a rather wide range of $CaO/(FeO_t + MgO + TiO_2)$ ratios (Fig. 8). These features suggest a derivation from a wide range of rocks, which include felsic pelite, greywackes and am-

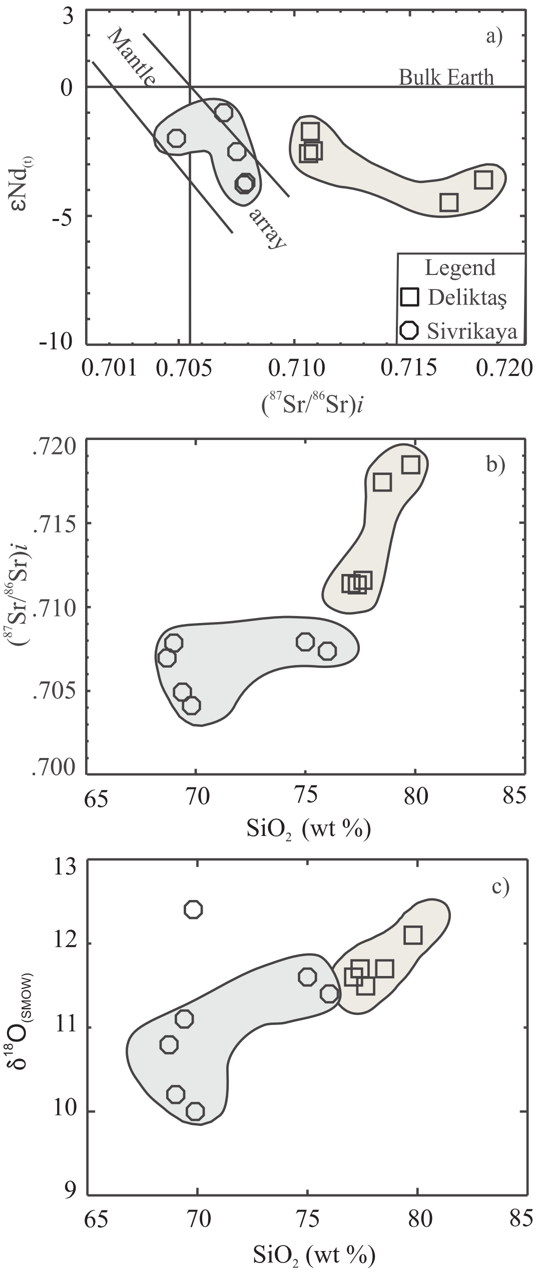

Fig. 7. Isotopic compositions of samples from Deliktaş and Sivrikaya granitoids. **(a)** Initial εNd(t) values vs. initial Sr isotopic ratios; **(b)** and **(c)** initial Sr isotopic ratios and $\delta^{18}O$ values vs. SiO_2 respectively.

phibolite. Evidence for the melting of heterogeneous and old (amphibolitic) sources is further provided by missing to very small negative Eu-anomalies in the REE patterns, and the variable CaO and Sr contents of the samples. Some of the Sivrikaya rocks plot in the peraluminous leucogranites field (Fig. 8), as these samples contain secondary muscovite. Muscovitization of feldspar and chloritization of biotite in these samples suggest hydrothermal alteration. Partial melting models (Clemens & Vielzeuf 1987, Vielzeuf & Holloway 1988, Patiño Douce & Johnston 1991), suggest partial melting of metabasic rocks might generate rocks which exhibit transitional- to S-type geochemical characteristics. On the Na_2O vs. K_2O diagram (Fig. not shown) only the Sivrikaya transitional samples (SG–115, –116, –125, –132) plot in the field outlined for typical I-type granite of the Lachlan Fold Belt (Chappell & White 1974, 1992). The fact that some Sivrikaya rocks are compositionally transitional between the I- and S-type granites implies that the Sivrikaya granitoid might have been derived from partial melting of acid to intermediate igneous rocks or immature sediments. Sivrikaya rocks show relatively large variations in isotopic composition ($\varepsilon Nd_{(t)}$ = –1 to –4; $Sr_{(i)}$ = 0.7040–0.7079) further suggesting their derivation from heterogeneous and old sources. The slightly to strongly negative $\varepsilon Nd_{(t)}$ values, low $Sr_{(i)}$ ratios, and young Nd model ages (T_{DM} = 0.75–1.08 Ga) indicate significant input of mantle-derived component during magma generation. The Sivrikaya basement consists of orthogneisses and amphibolites with similar $Sr_{(i)}$ (~ 0.704–0.706), higher $\varepsilon Nd_{(t)}$ values (–1.4 to 9.4) and lower $\delta^{18}O_{whole\text{-}rock}$ values (8.2 to 10.3 ‰) compared to the granitoid samples. However, the young model Nd ages (0.75–1.08 Ga) of the SG samples indicate that the petrogenesis of this granitoid involved the input of juvenile mantle component (e. g. DePaolo et al. 1992). Therefore, a mixture of juvenile material and old continental crust may characterize the petrogenesis of this granitoid. Furthermore, some samples show the concave-upward REE patterns and are depleted in MREE relative to HREE (Fig. 5 a), indicating that amphibole played a more significant role than garnet during magma segregation. Given the existing experimental constrains and the overall geochemical and isotopic characteristics of the rocks, we think the origin of the Sivrikaya granitoids involves partial melting of crustal protolith having heterogeneous composition, under variable H_2O activities, leaving restite with variable proportions of amphibole. This process combined with fractional crystallization of the melts en route to higher crustal levels can generate the whole spectrum of rock types represented in the Sivrikaya pluton, e. g. the transitional types of the Moldanubian granites (e. g. Liew & Hofmann 1988).

All plots in Fig. 8 indicate an origin of the Deliktaş magmas by dehydration melting of felsic pelite-type source rocks. High and nearly constant Rb/Sr ratios (5.6–5.7) and $\delta^{18}O_{whole\text{-}rock}$ values (~ 11.5–11.7‰) of the Deliktaş rocks might suggest derivation from a homogeneous source. However, the samples have variably high initial $Sr_{(i)}$ ratios and low $\varepsilon Nd_{(t)}$ values and Nd model ages of 1.2–2.2 Ga. This and the positive to nega-

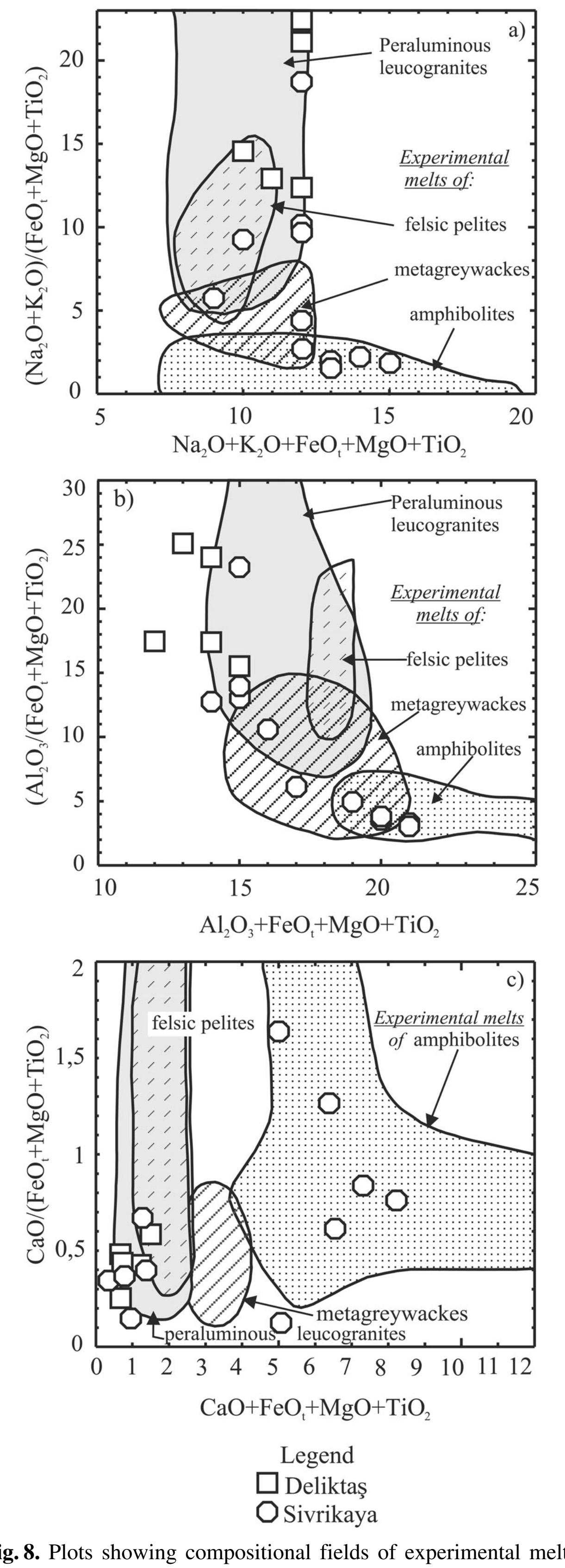

Fig. 8. Plots showing compositional fields of experimental melts derived from partial melting of felsic pelites, metagreywackes and amphibolites (Patiño Douce 1999) and composition of studied samples from Deliktaş and Sivrikaya granitoids.

tive Eu anomalies, emphasizing that young and old crustal material contributed in their petrogenesis. The Deliktaş basement rocks (gneisses and granites) have similar range in $\varepsilon Nd_{(t)}$ (−6.6 to −3.4), $\delta^{18}O_{whole\text{-}rock}$ values (10 to 11.3‰) and $Sr_{(i)}$ ratios (0.707–0.712), and Nd model ages (1.1–1.3 Ga), compared to the granitoid samples (Nzegge in prep). The more evolved and less variable geochemical and isotopic characteristics of the Deliktaş granitoid rocks are similar to plutons that are entirely derived from crustal sources (e. g. Chappell & White 1974, 1992, Liew & Hofmann 1988). These characteristics further reflect low water activity and lower melt fraction of a muscovite-rich metapelite (e. g. Thompson 1996) (e. g. Fig. 8), with sufficient volume to generate the observed proportion of the Deliktaş leucogranite. An origin by pure remagmatization (dehydration melting) of felsic crust is regarded as the most viable model for generating the chemical signatures of the Deliktaş leucogranites.

In conclusion, we suggests that the Deliktaş magmas were generated by dehydration melting of metapelitic material and the Sivrikaya magmas originated by dehydration melting of a heterogeneous mafic crust (amphibolite-greywacke-type).

Fractional crystallization

Small or no negative anomalies of Eu, Ba and Sr indicate that fractionation of plagioclase has not played an important role in the petrogenesis of Sivrikaya granite (Figs. 5 a and 6 a). The fractionation of accessory phases such as zircon, allanite and titanite can account for depletion in zirconium and titanium (e. g. Weaver 1990). The Sivrikaya samples display moderate concave-upward REE patterns and relative depletion of middle REEs with respect to HREE (Fig. 5 a), which can be attributed to fractionation of amphibole and/or titanite (e. g. Romick et al. 1992, Hoskin et al. 2000). High SiO_2-contents as well as high values of Fe_2O_3/MgO ratios of some Sivrikaya samples (Table 1) might also indicate that the parent magma experienced magmatic differentiation (e. g. Whalen et al. 1987). Increase in $\delta^{18}O_{whole\text{-}rock}$ values of ~0.6‰ in the Sivrikaya samples can be attributed to fractional crystallization without significant contamination by continental crust (e. g. Taylor 1978, Harmon & Gerbe 1992).

In contrast, the high and narrow range in $\delta^{18}O_{whole\text{-}rock}$ values of the Deliktaş samples and the restricted $\delta^{18}O_{qtz}$ and $\delta^{18}O_{K\text{-}feldspar}$ values could be attributed to significant contribution of continental crust (e. g. Taylor 1978, O'Neil & Chappell 1977) (Table 2, Fig. 7b, c). Furthermore, Deliktaş granitoid rocks are all leucocratic and exhibit porphyritic textures, with "apparently" primary muscovite megacrysts, myrmekite and graphic quartz, which are not observed for the Sivrikaya rocks.

The initial Sr isotopic compositions of Deliktaş samples lie outside the range of the Sivrikaya rocks (Fig. 7 a, b). The Deliktaş granitoid has older Nd model ages (T_{DM} = 1.2–2.2 Ga) than Sivrikaya (T_{DM} = 0.75–1.08 Ga), suggesting that they were derived from different sources. Discontinuous chemical variations illustrated in Harker diagrams (Fig. 3), the high Silica-content ($SiO_2 > 60$ wt %), the considerable 10 Ma difference in emplacement age (Nzegge et al. subm.) and the volume of the Sivrikaya and Deliktaş granitoids makes it unlikely that the magmas were derived from a basaltic parent magma through fractional crystallization processes (e. g. Grove & Donnelly-Nolan 1986, Bacon & Druitt 1988).

Geodynamic implications

On the Pearce et al. (1984) diagram the Deliktaş and Sivrikaya plutons plot in the volcanic arc granitoids (VAG) field (Fig. 4 c). The high silica-content samples from the Sivrikaya granitoid plot on the VAG and syn-COLG boundary. Crossing of the VAG and syn-COLG boundary as observed for the Sivrikaya samples is likely due to progressive magmatic differentiation (e. g. Förster et al. 1997). Most chemical signatures of Deliktaş samples indicate they were anorogenic, implying the VAG characteristics were inherited from the arc environment in which they were generated (e. g. Pearce et al. 1984, Harris et al. 1986, Rogers & Hawkesworth 1989, Sajona et al. 1996). Pitcher (1982, 1983), Pearce et al. 1984 and Harris et al. 1986 showed that granitoid types broadly correlate with geological environment and tectonic regimes. Using the Pitcher-Pearce-Harris scheme, the peraluminous (ASI ≥ 1.1) assemblage of Deliktaş leucogranite rocks are classified as post-collisional and S-type, emplaced after the peak of crustal thickening (e. g. Sylvester 1998). Increase in K- and Sr-contents, high Rb/Sr ratios and initial Sr values of the Deliktaş leucogranite samples, are clearly related to a thickened crust and the involvement of more crustal material in their genesis (e. g. Bennett & De Paolo 1987). In addition, Deliktaş monzogranites have K-feldspar megacrysts and apparently "primary" muscovite, common in the S-type plutons of Hercynian belts of Western and central Europe (e. g. Pagel & Laterrior 1980, Lameyer et al. 1980, Liew et al. 1989, Barbarin 1999, Fernandez-Saúrez et al. 2000). The geochemical data of the Deliktaş and Sivrikaya granitoids can be summarized as follows: the subduction-related Sivrikaya granitoid is ultimately derived from mixed mafic-felsic material (e. g. Hawkesworth et al. 1993), and crustally derived high-K Deliktaş granites resulting from partial melting of crustal material, indicating a major crustal thickening episode (e. g. Boztuğ et al. 1995) that followed subduction and collision of micro-continental plates with South Eurasian active margin (e. g. Şengör 1984, Robertson & Dixon 1984, Ustaömer & Robertson 1993). The activity of the Eurasia-Pontides convergent zone probably consisted of successive periods of compression and piling of the crustal fragments on one hand and tension on the other hand, to permit respectively genesis of magmas and alternating emplacement of (1) mixed (crust and mantle) origin (e. g. Sivrikaya granite), the result of subduction, transitional regimes and continental collision; and (2) peraluminous granitoids (muscovite- and cordierite-bearing, K-rich peraluminous granites, associated with the climax of orogenesis) dominantly of crustal origin (e. g. Deliktaş granite). Deliktaş rocks were probably emplaced during the relaxation phases (e. g. Barbarin 1999) where anatectic complexes formed are commonly porphyritic-K-feldspars-rich (e. g. Wyllie 1977).

Conclusions

Geochemical signatures, a wide range of Sr–O isotopic compositions and geochronological data (e. g. ~ 10 Ma age difference) preclude a comagmatic origin of the Sivrikaya and Deliktaş magmas. An origin by dehydration melting of homogeneous, felsic upper crust is regarded as the most viable model for generating the Deliktaş leucogranite. In contrast, the Sivrikaya magmas originated by partial melting of heterogeneous pelite–amphibolite–greywackes-type sources. The lowermost crust must have been hot, implying extensive heating by mantle-derived mafic magmas.

These types of granitoids are widespread in the other Pontides and the Hercynian belt of Central and Western Europe.

The mineralogical and geochemical data and isotopic characteristics of the Sivrikaya and Deliktaş granitoids are predominantly the consequences of melting, mixing and assimilation of crustal rocks, and the compositions of the source regions rather than large-scale, post-emplacement differentiation. Major and trace element compositions suggest the Sivrikaya granitoid to be subduction-related, while the Deliktaş granite may be post-orogenic, with inherited arc-related signatures.

The Late Carboniferous-Early Permian ages for the Sivrikaya and Deliktaş granitoids and the geochemical and isotopic data of the granitoids and basement, suggest that they were components of the Hercynian orogenic belt.

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