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Geochemistry and petrogenesis of intrusive and extrusive ophiolitic plagiogranites, Central Anatolian Crystalline Complex, Turkey

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Abstract

Plagiogranites associated with the Sarikaraman ophiolite of the Central Anatolian Crystalline Complex, Turkey, closely resemble other plagiogranites from supra-subduction zone-type ophiolites of Neotethys. The ophiolite is remarkable in displaying a higher proportion of the plagiogranite suite (ca. 10% by volume) than is usually associated with such bodies. The Sarikaraman plagiogranites are represented by intrusive sheets and netvein trondhjemites largely developed at the top of the upper gabbros and as multiphase dykes within the sheeted dyke complex. The plagiogranite dykes are considered to feed extrusive silicified rhyolites associated with the basaltic lavas in the volcanic section of the ophiolite. Field relations suggest that the trondhjemites were probably generated from the roof section of a dynamic and evolving gabbroic magma chamber. Both the deep-seated trondhjemites and the volcanic rhyolites constitute the Sarikaraman plagiogranite suite. Geochemically there is complete overlap between the intrusive trondhjemites and extrusive rhyolites, which are characterised by (MORB-normalized) low HFS element contents with small negative Nb–Ta anomalies and variably enhanced LIL element abundances. Unlike other plagiogranites, however, the Sarikaraman suite is not characterized by consistently low K_2O contents; a feature that reflects the variable mobilization of the LIL elements under lower greenschist facies conditions. The REE are uniformly enriched relative to the basic components of the complex, but have similar normalized patterns exhibiting mild light REE depletion. In terms of their origin, the initial or most primitive plagiogranite melts could have been generated by either fractional crystallization (70–85% of clinopyroxene–feldspar ± amphibole) or partial melting (5–15% batch melting) of a gabbroic 'source material', although only the first process can produce most of the range of the plagiogranite compositions. As a group the plagiogranites exhibit some degree of internal variation which can be generated by further fractionation largely dominated by feldspar with minor apatite and amphibole. © 1998 Elsevier Science B.V.

Keywords: Plagiogranite; Petrogenesis; Rhyolite; Trondhjemite; Ophiolite; Central Anatolia

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1. Introduction

Small-volume acidic magmatic rocks associated with ophiolitic complexes have compositions ranging from alkali granite through trondhjemite and tonalite to diorite and are collectively referred to as oceanic plagiogranites (Coleman and Peterman, 1975). Plagiogranites typically occur as intrusive masses of varying size and form in the upper portions of the plutonic ophiolite sequence and are commonly associated with layered or massive gabbro. A number of processes have been suggested to account for their origin and development in the oceanic environment. These include: (a) the extreme differentiation of tholeiitic MOR-type magmas (e.g. Coleman and Peterman, 1975; Coleman and Donato, 1979; Pallister and Knight, 1981), (b) the partial melting of altered and hydrated basic rocks, including amphibolite (e.g. Gerlach et al., 1981; Spulber and Rutherford, 1983) and (c) under certain conditions, as immiscible liquids coexisting with mafic melts (e.g. Philpotts, 1976; Dixon and Rutherford, 1979). Broadly acidic/intermediate rocks with a similar petrography and chemistry to plagiogranites are also a relatively common occurrence in modern, and analogous, ancient oceanic settings such as mid-ocean ridges (e.g. Indian Ocean Ridge, Engel and Fisher, 1975), island arcs (e.g. Oman, Alabaster et al., 1982; Canyon Mountain, Gerlach et al., 1981) and back-arc basins (e.g. Bay of Islands, Malpas, 1979; S. Chile, Saunders et al., 1979; Kunar Massif, Vernikovskiy et al., 1993).

Apart from their variable petrogenesis, plagiogranitic rocks can play a significant part in understanding the tectonic setting of ophiolitic rocks which generally lack suitable exposures of the volcanic and sedimentary sections. For example, Pearce et al. (1984) have attempted to reconstruct the geodynamic environment of ophiolite formation based on the trace element content of the associated plagiogranites. As well as being associated with spreading centre generated oceanic crust, plagiogranites are common members of supra-subduction zone-type ophiolites (e.g. Alabaster et al., 1982; Pearce et al., 1984; Jenner et al., 1991; Jafri et al., 1995). This is a particular characteristic of Mesozoic Neotethyan ophiolites (e.g. Troodos, Hatay, Antalya, Vourinos, Oman) which display a significant proportion of plagiogranites throughout their magmatic pseudos-

tratigraphy (Aldiss, 1978; Pearce et al., 1984); this feature also includes the Sarikaraman ophiolite studied here.

2. Turkish ophiolites

Mesozoic ophiolites are ubiquitous in Turkey and the Middle East region and represent the variably disrupted fragments of strands of the Neotethyan ocean that was situated between Eurasia and Gondwanaland (Juteau, 1980; Sengor and Yilmaz, 1981; Robertson and Dixon, 1985). These oceanic strands comprise a northern branch, composed of two separate segments and a single southern branch, all of which crop out along a nearly east–west trending tectonic zone. The northernmost segment represents remnants of the Intra-Pontide Ocean, whereas the other segment is composed of allochthonous units derived from the Vardar–Izmir–Ankara–Erzincan Ocean just to the south. The southernmost branch separated the main body of Gondwanaland from the Tauride–Anatolide Platform to the north and is variably named the Peri-Arabic Belt (Ricou, 1971) or Southern Neotethyan Ocean (Sengor and Yilmaz, 1981. Despite these more-or-less well-defined oceanic strands, numerous poorly known and isolated ophiolitic bodies are exposed in the Central Anatolian Crystalline Complex (CACC) or Kirsehir Block, just to the south of the Vardar–Izmir–Ankara–Erzincan Ocean segment (Fig. 1). The CACC is a triangular area of metamorphic basement which is considered to represent the northern passive margin of the Mesozoic Tauride–Anatolide Platform (TAP), facing the Vardar–Izmir–Ankara–Erzincan Ocean (Ozgul, 1976). However, Gorur et al. (1984) suggest that the CACC/Kirsehir Block was separated from the TAP by a distinct oceanic strand called the Inner Tauride Belt that was unrelated to the northern branches of Neotethys. Isolated outcrops of ophiolitic rocks, termed the Central Anatolian Ophiolites (Goncuoglu et al., 1991, 1992) are found as allochtonous bodies in the CACC (Fig. 1), having been emplaced southwards via obduction from the Vardar–Izmir–Ankara–Erzincan Ocean (Ozgul, 1976).

In terms of pseudostratigraphic relationships of magmatic units and chemical designation, the Cen-
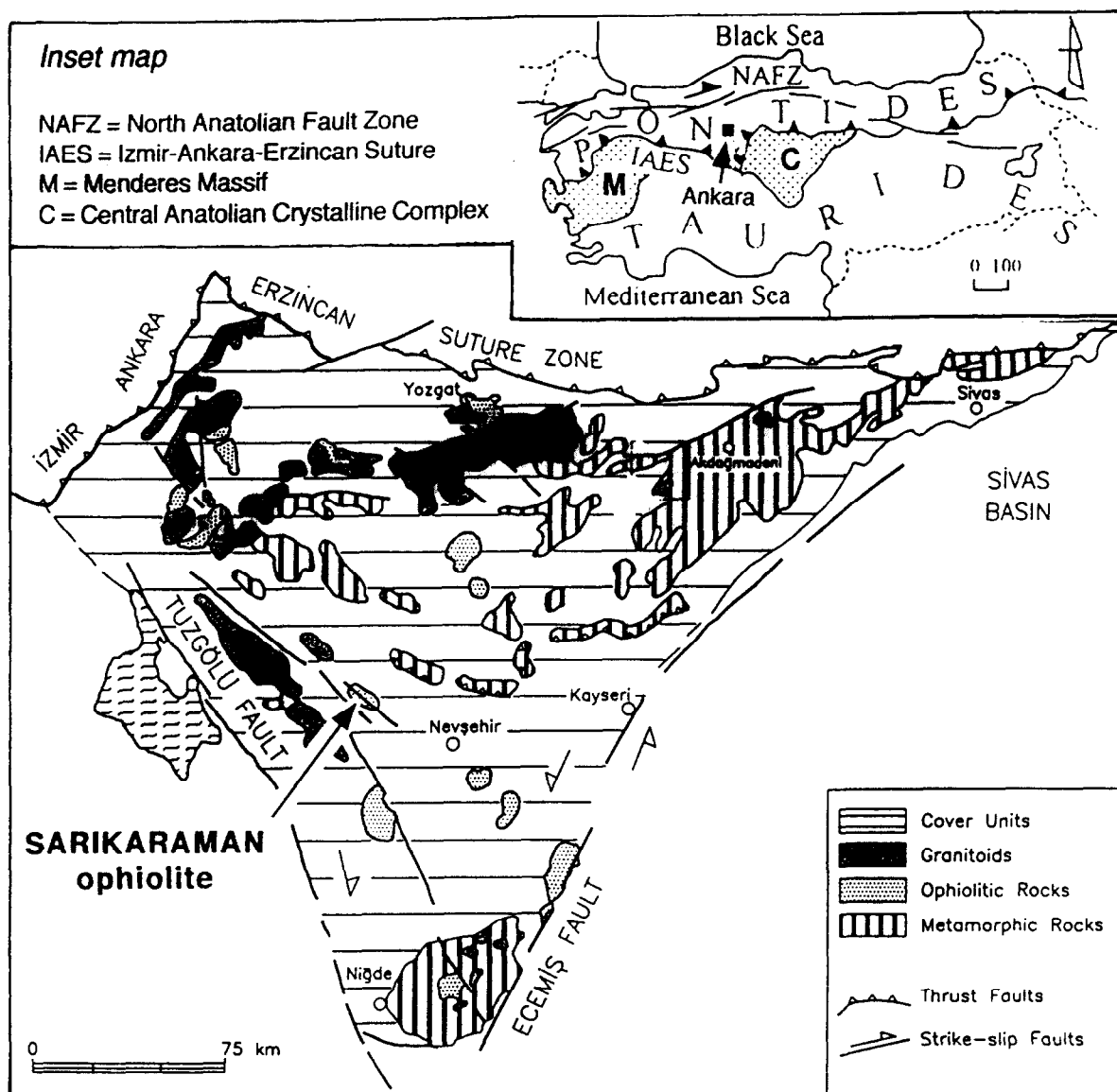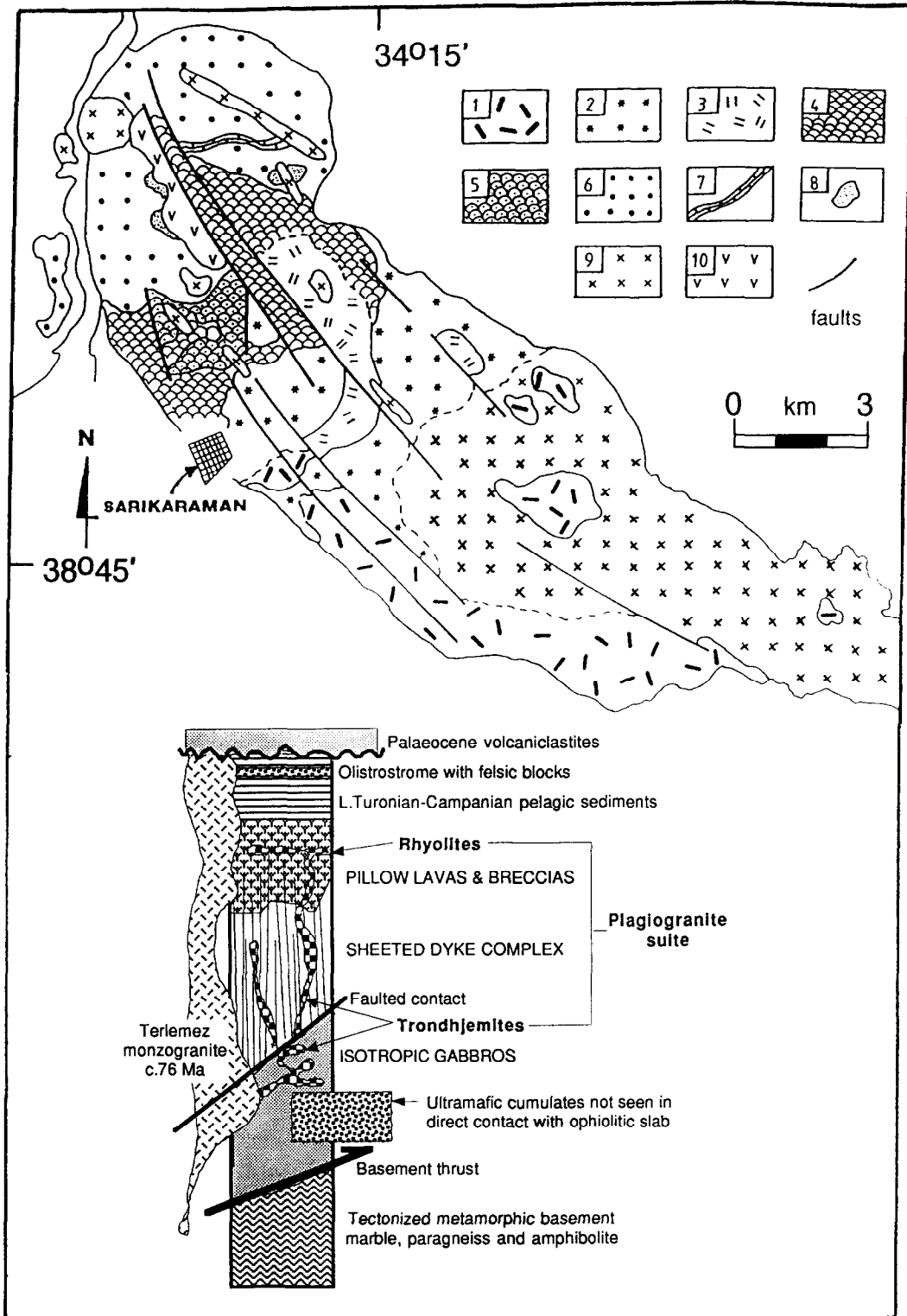
Fig. 1. Sketch map of the main lithologic units of the Central Anatolian Crystalline Complex (CACC), Turkey (after Bingol, 1989) showing the location of the Sarikaraman ophiolite body. Inset shows the position of the CACC directly to the south of the Izmir–Ankara–Erzincan Suture Zone.

tral Anatolian Ophiolites exhibit a supra-subduction zone setting and are characterized by a significant proportion of plagiogranites (Goncuoglu and Tureli, 1993; Yaliniz et al., 1996). These features are mirrored by the ophiolites derived from the southern branch of Neotethys, such as Troodos, Antalya, Hatay, Mersin and Oman. However, the plagiogranites of the Central Anatolian Ophiolites have not

been studied in detail and invite comparison with similar rocks in the southern branch of Neotethys in the eastern Mediterranean area.

This paper is concerned with the petrological and geochemical features of plagiogranites that occur throughout the magmatic sequence of the Sarikaraman ophiolite (as representative of the Central Anatolian Ophiolites), together with extrusive flows in

34°15'

1 | 2 | 3 | 4

5 | 6 | 7 | 8

9 | 10

faults

0 km 3

N

SARIKARAMAN

38°45'

Palaeocene volcaniclastites

Olistrostrome with felsic blocks

L.Turonian-Campanian pelagic sediments

Rhyolites

PILLOW LAVAS & BRECCIAS

SHEETED DYKE COMPLEX

Plagiogranite suite

Faulted contact

Trondhjemites

ISOTROPIC GABBROS

Terlemez monzogranite c.76 Ma

Ultramafic cumulates not seen in direct contact with ophiolitic slab

Basement thrust

Tectonized metamorphic basement marble, paragneiss and amphibolite

the volcanic section and discusses their origin relative to the basic members of the complex.

3. Sarikaraman ophiolite

The Sarikaraman ophiolite is representative of a number of partly dismembered bodies within the CACC that retains a still recognizable pseudostratigraphy (Fig. 2). Whereas voluminous ultramafics are not exposed in direct contact with the rest of the ophiolitic slab, the lowest section is composed of isotropic gabbros, which are juxtaposed against a sheeted dyke complex that merges up-section into basalt lavas and breccias. The gabbros are cut by plagiogranites which are related to various high-level rhyolitic dykes and lavas that are found in the upper volcanic section of the ophiolite. All units are cut by a late set of isolated dolerite dykes. The ophiolite is conformably overlain by a sequence of Lower Turonian–Campanian pelagic sediments (Yaliniz et al., in press) intercalated with a volcanogenic olistrostromal sedimentary unit. Both the ophiolite and cover sediments are intruded by the late Cretaceous–Lower Palaeocene Terlemez monzogranite (ca. 76 Ma). This late granitic body is unrelated to the ophiolitic plagiogranites and has its counterpart in other areas of the CACC, representing the post-collisional product of the melting of thickened crust (Goncuoglu and Tureli, 1994; Yaliniz et al., 1996). Finally, the ophiolite and late granites are unconformably overlain by Palaeocene volcaniclastites. The field relationships and faunal age of the ophiolitic sediments indicate that the Sarikaraman ophiolite had a Turonian formation age before being rapidly obducted onto the CACC prior to late granite intrusion.

4. Field relationships

The dominant plutonic, together with the minor volcanic components of the plagiogranites occupy a significant volume (estimated at 8–12%) of the Sarikaraman ophiolite as a whole (Fig. 1). Although the plagiogranites mainly occur at the boundary between the isotropic gabbro and the sheeted dyke complex, as noted for other ophiolites (Coleman and Donato, 1979; Gerlach et al., 1981; Hopson et al., 1981), acid rocks of similar character are found spasmodically throughout the volcanic section of the ophiolite. At the gabbro boundary they are laterally discontinuous and intrude upwards into the dyke complex displaying both sharp and diffuse margins with no chilled zone.

Stratigraphically, within the ophiolite sequence, the plagiogranites exhibit a number of different emplacement relationships: (a) in the gabbros they occur as either simple narrow fracture-filling dykes or commonly as more wide-spread netveined or agmatitic complexes enclosing partly assimilated or resorbed mafic enclaves; (b) in the dolerite sheeted dyke complex they occur mainly as dykes or irregular pods; (c) in the upper basaltic massive and pillowed lava unit, plagiogranites occur both as small irregular dykes and felsic lava flows. Finally, similar felsic lava is also found as blocks in olistrostrome units within the sedimentary cover, implying that autobrecciated equivalents of the high-level volcanics were subjected to secondary redeposition via mass flow mechanisms. As with other ophiolites the greatest development of the plagiogranites are seen at the uppermost part of the gabbros and because of their intimate field relationships, the gabbros are considered the initial source or site of the plagiogranites. In the Sarikaraman ophiolite contact relationships within the deep-seated trondhjemite–gabbro netvein complex suggests that not only was there more than one development of acid melt, but that both acid and basic melts probably coexisted. Higher up in the ophiolite sequence (upper gabbros and sheeted dykes) the plagiogranites form brittle fracture infillings that both pre- and post-date dolerite dyke intrusion.

Fig. 2. Geological sketch map (Yaliniz et al., 1996) of the Sarikaraman ophiolite, Central Anatolia, Turkey, and a composite lithostratigraphic section showing location of the plagiogranite suite: 1, isotropic gabbros; 2, plagiogranites; 3, sheeted dyke complex; 4, pillow lavas; 5, pillow breccias; 6, Lower Turonian-Campanian sediments; 7, Campanian limestone unit; 8, felsic volcanic block unit; 9, Terlemez monzogranite (ca. 76 Ma) and 10, (?)Palaeocene volcaniclastic unit.

5. Petrography

Depending on their position in the ophiolite stratigraphy, the plagiogranites vary from medium/coarse grained plutonic varieties (deeper levels) to very fine-grained or vitric types (upper levels). Apart from their texturally heterogeneity, the main distinction is between the plutonic members at the gabbro boundary which are classified as trondhjemites (based on normative An–Ab–Or) or low mafic tonalites (Streckeisen, 1976) and the high-level dykes and lavas which have the appearance of altered silicic rocks (keratophyric). This later group have generally been referred to as rhyolites in the field, although their highly siliceous nature is probably a consequence of secondary alteration, as chemically they are similar to the trondhjemites (see below).

The trondhjemites are generally equigranular, leucocratic rocks consisting largely of plagioclase and quartz. Plagioclase is present as subhedral, sometimes zoned, but rarely twinned, albite (commonly An_{2-8}, or if zoned with a core of An_{20-25}), whereas quartz invariably forms a granular mosaic partly enclosing earlier feldspar. Textures range from typically hypidiomorphic granular to granophyric pro-

Table 1
Representative analyses of volcanic and plutonic plagiogranites from the Sarikaraman ophiolite, Turkey

| Sample No.: | KY-68 | KY-80 | KY-113 | KY-115 | M-28 | M-31 | M-32 | M-37 | M-50 | M-67 | M-68 | M-86 | ADAT |
Rock type:	Rhy	Rhy	Rhy	Rhy	Rhy	Rhy	Rhy	Rhy	Rhy	Rhy	Rhy	Rhy	Rhy
Major oxides (wt%): XRFS													
SiO_2	81.39	79.76	70.71	65.03	72.73	76.41	82.03	77.93	72.77	73.01	76.77	75.41	63.39
TiO_2	0.23	0.13	0.20	0.54	0.52	0.38	0.29	0.12	0.52	0.37	0.10	0.36	0.55
Al_2O_3	10.26	11.66	16.51	15.72	13.03	11.81	9.53	11.68	13.81	11.88	11.76	11.61	15.52
Fe_2O_3t	1.95	0.73	2.54	5.12	3.41	2.86	1.94	0.99	4.32	2.77	0.82	4.09	4.44
MnO	0.03	0.02	0.04	0.08	0.05	0.05	0.03	0.02	0.02	0.04	0.09	0.10	0.08
MgO	0.42	0.03	0.01	1.98	2.57	1.12	0.74	0.62	1.57	3.98	0.17	2.32	1.70
CaO	0.36	0.06	0.19	3.77	0.36	1.68	1.27	1.74	0.45	1.50	2.93	0.24	5.22
Na_2O	2.79	3.23	10.02	4.21	5.54	4.54	1.17	0.13	4.23	3.71	0.53	3.07	2.59
K_2O	1.28	3.65	0.05	2.09	0.06	0.26	0.84	3.39	0.35	0.23	2.85	0.81	4.04
P_2O_5	0.03	0.01	0.01	0.09	0.11	0.05	0.03	0.01	0.09	0.03	0.01	0.06	0.12
LOI	1.29	0.82	0.19	1.77	1.81	1.11	1.87	3.27	2.18	2.77	4.18	1.92	2.69
Total	100.03	100.10	100.47	100.40	100.19	100.27	99.74	99.90	100.31	100.29	100.21	99.99	100.34
Trace elements (ppm): XRFS													
Ba	103	2444	14	224	59	17	64	1687	12	24	59	52	715
Ce	21	78	7	43	11	13	7	71	12	10	24	5	91
Cl	57	42	25	43	66	46	36	30	51	39	33	49	33
Cr	23	29	17	48	18	20	23	15	15	59	18	33	26
Cu	2	1	2	5	4	1	1	1	1	1	1	6	12
Ga	14	9	16	17	12	11	9	10	13	10	16	13	21
La	6	42	1	16	2	5	2	39	1	1	6	1	45
Nb	1	4	1	3	1	2	1	7	1	1	4	1	17
Nd	29	33	9	25	14	14	6	24	12	7	18	25	31
Ni	5	4	5	10	3	5	4	5	5	28	2	4	9
Pb	13	21	1	13	3	7	2	10	1	1	9	1	77
Rb	35	100	4	62	5	11	29	92	17	10	127	26	156
S	71	72	63	79	65	60	70	107	63	64	68	70	67
Sr	36	150	49	148	49	96	70	23	101	82	40	40	616
Th	1	34	1	10	1	1	1	33	1	1	13	1	30
V	1	3	12	62	24	19	12	3	7	68	3	5	55
Y	44	16	64	27	44	55	38	16	49	24	25	35	22
Zn	68	42	12	39	84	15	38	19	18	29	22	84	79
Zr	137	95	176	168	179	191	71	96	131	143	55	102	202

Rhy = volcanic rhyolites; Tron = plutonic trondhjemites.

duced by late plagioclase and quartz intergrowths. Primary mafic phases (5–10%), such as biotite and amphibole, are rarely preserved, commonly being replaced by chlorite and epidote group minerals. Titanite and rare apatite occur as accessory phases.

Rhyolite dykes are fine-grained with a microgranophyric texture composed of albite and quartz. Originally the groundmass was probably glassy, but is now devitrified as shown by the development of occasional relict spherulites within a granular recrystallized mosaic. The high-level sills and lava flows are typically aphyric, very fine-grained or cryptocrystalline and often exhibit flow banding. Originally holohyaline they have undergone considerable devitrification with the growth of radial quartzfeldspar spherulites, together with the variable development of secondary K feldspar in the matrix.

Secondary minerals occur not only in the matrix, but are common in veinlets throughout the plagiogranites. Secondary mineralization comprises mainly epidote group, chlorite and quartz with subordinate calcite and traces of titanite. Zoned turbid feldspar often has an epidotized core or may be largely kaolinitized. Primary and secondary quartz frequently occur in a wide variety of textures, sometimes with no clear distinction between them, espe-

16 Rhy	20 Rhy	116-B Rhy	128 Rhy	184 Rhy	TL-4 Tron	M-7 Tron	144 Tron	147 Tron	149 Tron	157-T Tron	163 Tron	165 Tron	166 Tron
77.69	79.04	79.39	77.77	81.50	72.63	80.42	77.01	77.14	77.91	72.82	73.07	77.15	79.38
0.22	0.24	0.27	0.08	0.13	0.52	0.19	0.29	0.45	0.33	0.41	0.30	0.37	0.17
12.45	10.41	10.57	12.06	9.93	13.84	10.94	12.06	12.89	12.21	13.44	13.07	12.14	12.02
1.17	2.18	1.98	0.72	1.37	1.64	0.79	2.18	0.95	0.77	3.48	2.80	1.59	1.03
0.02	0.08	0.04	0.07	0.01	0.05	0.01	0.03	0.02	0.03	0.09	0.04	0.03	0.02
0.27	1.24	0.55	0.19	0.01	1.81	0.56	0.50	0.28	1.08	1.24	0.48	0.79	0.30
0.97	2.11	0.98	1.84	0.66	2.27	0.99	1.67	1.65	0.87	1.50	4.36	1.14	0.13
6.70	2.27	5.03	2.35	5.67	4.92	4.45	5.30	4.68	5.19	4.90	5.13	5.68	6.45
0.02	1.42	0.25	2.12	0.03	0.90	0.60	0.41	1.31	0.42	0.92	0.03	0.41	0.17
0.01	0.02	0.02	0.01	0.01	0.09	0.02	0.02	0.06	0.05	0.06	0.03	0.03	0.01
0.36	1.24	0.75	2.80	0.91	1.39	0.89	0.90	0.87	1.11	1.23	0.63	0.89	0.51
99.88	100.25	99.83	100.01	100.23	100.06	99.86	100.37	100.30	99.97	100.09	99.94	100.22	100.19
5	77	27	49	26	16	149	45	43	14	94	32	24	31
12	13	12	21	14	12	15	6	34	16	23	2	20	42
71	27	40	52	34	74	66	65	36	59	43	38	115	29
20	25	26	12	21	29	26	23	20	17	25	31	19	25
1	1	1	1	1	2	1	2	1	1	2	7	1	1
10	11	11	13	7	11	11	11	11	10	12	11	13	12
4	3	2	5	2	1	2	1	3	2	2	1	4	2
1	1	1	3	1	1	1	1	1	1	1	1	1	1
16	12	16	16	26	17	19	10	19	12	21	10	25	30
4	5	5	1	3	6	3	4	3	3	3	9	4	4
1	7	1	11	1	3	1	4	3	1	4	6	1	1
5	34	9	119	5	30	22	15	35	15	30	4	14	6
62	71	62	67	70	67	65	62	63	64	64	64	63	65
62	94	63	110	46	123	70	116	124	102	104	130	66	31
1	1	1	13	1	1	1	1	1	1	1	1	1	1
6	21	24	11	6	54	10	30	24	8	29	54	26	2
54	25	29	22	42	26	33	26	30	40	29	22	33	52
9	64	21	16	24	26	15	16	11	21	38	17	18	13
114	145	146	54	104	171	165	170	179	115	136	164	173	195

cially where large quartz grains partially replacing plagioclase are in optical continuity with interstitial quartz. The highly siliceous nature of the volcanic 'rhyolites' in the field is largely a consequence of replacement by secondary quartz, where total quartz can comprise up to 75% of the mode. Epidote group minerals are invariably associated with turbid plagioclase and secondary quartz; clinozoisite typically replaces the more calcic cores or zones of plagioclase crystals, whereas later epidote is more pistacitic and is developed throughout the matrix. In extreme cases progressive epidotization of the rhyolites culminates in the formation of epidosite pods. In the deeper seated trondhjemites, epidote frequently forms within or adjacent to joints and veinlets.

6. Geochemistry

Representative samples of the Sarikaraman plagiogranite suite (9 trondhjemites, 18 rhyolites) were analysed for major oxides and selected trace elements (data in Table 1) on an ARL8420 X-ray Fluorescence Spectrometer (Department of Earth Sciences, Keele University, UK) calibrated against > 30 international and internal Keele standards of appropriate composition. Details of method, accuracy and precision are given in Floyd and Castillo (1992). Three samples were also determined for Hf, Ta, Th, U and the rare earth elements (REE) by Instrumental Neutron Activation Analysis (Activation Laboratories, Canada) and presented in Table 2.

6.1. Chemical classification and alteration effects

In terms of standard chemical classification of the plagiogranites, such as the SiO_2–total alkalis diagram (Le Bas et al., 1986), both the trondhjemites and the rhyolites have broadly similar compositions and are grouped as 'granites' and 'rhyolites' respectively. However, this is in part a consequence of element mobilization during low-grade alteration (the alkali contents are highly variable) and secondary silification. Stable element ratios, such as Zr/TiO_2 (Winchester and Floyd, 1977) which again group both the trondhjemites and rhyolites together (ratio ranges from 0.02–0.10) indicate that there is variation from andesitic to rhyodacitic (and plutonic

Table 2
Rare earth and other element abundances in representative Sarikaraman plagiogranites

Sample No.:	M-31	M-7	163
Rock type	Rhy	Tron	Tron
Trace elements (ppm): INAA			
Hf	4.40	4.60	4.20
Sc	12.2	4.9	6.3
Ta	0.08	0.11	0.13
Th	0.60	1.00	0.50
U	0.30	0.40	0.20
La	7.10	4.30	3.70
Ce	23.00	14.00	12.00
Nd	17.00	9.00	7.00
Sm	4.49	2.47	2.02
Eu	1.43	0.56	0.82
Gd	6.20	4.10	2.80
Tb	1.20	0.70	0.60
Yb	5.70	3.92	2.83
Lu	0.76	0.57	0.45

equivalent) compositions. In view of the effects of alteration, the petrographic and field designation of the plagiogranites as trondhjemites and rhyolites will be maintained, although it is recognized that they both originally had similar, broadly 'intermediate' compositions, typical of many oceanic plagiogranites (e.g. Coleman and Donato, 1979).

In common with the basaltic members of the Sarikaraman ophiolite, the plagiogranites have undergone secondary alteration in the lower greenschist facies. Chemically this has largely effected the large-ion-lithophile (LIL) element contents, such that K_2O is highly variable (ranging from 0.02 to 3.65 wt%) and not characteristically low and uniform as in most plagiogranites (e.g. Canyon Mountain ophiolite, Gerlach et al., 1981; Mingora ophiolite, Barbieri et al., 1994). However, high field strength (HFS) element contents and REE patterns (see below) are more stable and comparable to other oceanic plagiogranites such that they are considered to maintain a compositional range and coherence typical of magmatic processes (e.g. Herrmann et al., 1974; Floyd and Winchester, 1978; Pallister and Knight, 1981).

6.2. Chemical comparison with late granitoids

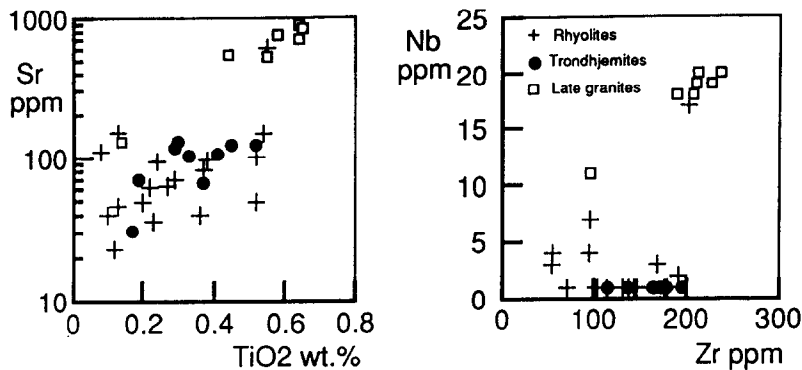Although field evidence demonstrates that the Terlemez monzogranite is later than all the members

Fig. 3. Chemical distinctions between the late cross-cutting granites and the plagiogranites (trondhjemites and rhyolites) of the Sarikaraman ophiolitic complex. The rhyolite sample plotting with the late granites suggests that it is related to the granites rather than being associated with the Sarikaraman plagiogranite suite.

of the ophiolite, it is worth pointing out that there is also a marked chemical distinction between the late granitoids and the ophiolitic plagiogranites. The compositional differences reflect the eruptive setting of the two groups, the granites being generated in thickened crust after continental collision (Goncuoglu and Tureli, 1994; Erler and Goncuoglu, 1996), whereas the plagiogranites were developed in an incipient arc/oceanic environment (Yaliniz et al., 1996). As seen in Fig. 3 the granitoids typically have higher incompatible element contents (e.g. especially high Nb values) or develop different trends in binary diagrams. On this basis the one rhyolite sample plotting with the late granites has been misclassified in the field and probably represents an 'aplitic' dyke from the main granite. Note also the marked overlap and association of the trondhjemite and rhyolite samples collected from the ophiolite assemblage. Although the occurrence of high-level 'keratophyric' lavas, broadly similar to the altered silicic lavas seen here, have been noted in other ophiolite complexes they are not always considered part of the plagiogranite suite (e.g. Brown et al., 1979; Gerlach et al., 1981).

6.3. Characteristic chemical features

Ocean ridge granite (ORG) normalized patterns (values from Pearce et al., 1984) are shown in Fig. 4 and demonstrate that the Sarikaraman plagiogranites are characterized by relatively low HFS element contents and highly variable LIL element values. The HFS elements are particularly depleted relative to hypothetical ORG, and exhibit small negative Nb–Ta anomalies. The slight difference in the normalized values for Ta and Nb are a consequence of analytical error at these low concentrations (Table 2). Geochemically they do not differ markedly from other arc or supra-subduction zone related ophiolitic plagiogranites, although the normalized LIL variability for the Sarikaraman plagiogranites probably reflects alteration rather than subduction enrichment. Similarly the spread of plagiogranite data across the volcanic arc-ocean ridge fields in tectonic discrimination diagrams for acid rocks (Fig. 5) is similar to other Tethyan supra-subduction zone ophiolitic plagiogranites (cf. Pearce et al., 1984). A supra-subduction zone environment for the Sarikaraman plagiogranites is also borne out by a similar chemical designation for the basaltic members (Yaliniz et al., 1996).

Chondrite-normalized REE patterns (Fig. 6a) for the three samples analysed are essentially sub-parallel, but with varying Eu anomalies indicative of feldspar involvement during fractionation and/or melting. The absence of a negative Eu anomaly in one sample can be attributed to elevated Sr and Rb contents and may reflect an element of feldspar accumulation (e.g. Birk et al., 1979). The Sarikaraman REE distribution is similar to the Samail plagiogranites (Pallister and Knight, 1981) and does not show any overlap with the associated basalts and
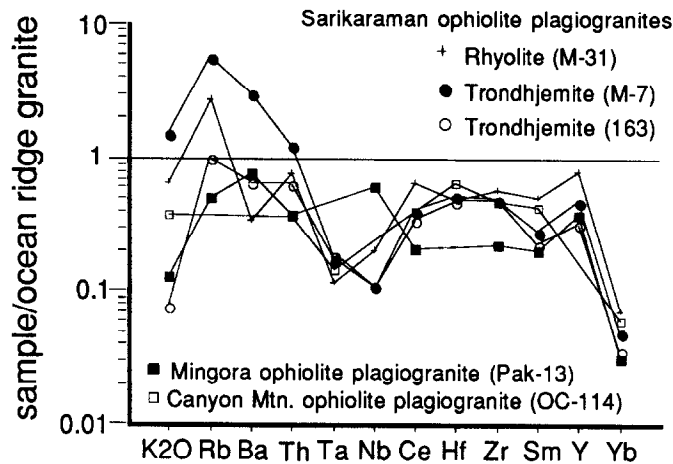
Fig. 4. Depleted incompatible element characteristic of the Sarikaraman ophiolite plagiogranites and other ophiolitic plagiogranites relative to an ocean ridge granitic differentiate (normalizing values from Pearce, 1983). Mingora plagiogranite from Barbieri et al. (1994); Canyon Mountain plagiogranite from Gerlach et al. (1981).

gabbros, unlike the plagiogranites from the Canyon Mountain ophiolite (Gerlach et al., 1981) (Fig. 6b). The largely flat patterns of the Sarikaraman plagiogranites with minor light REE depletion (La_N/Yb_N ca. 0.7–0.9) are typical of plagiogranites generally (Coleman and Peterman, 1975).

Compatible element (Mg, Cr and Ni) abundances are all typically low (Table 1), although coupled with the variation displayed by stable incompatible ele-

ments (including the REE) suggest some degree of chemical differentiation within the plagiogranite suite as a whole. As stated above, no chemical distinction was noted between the trondhjemites and the rhyolites and suggests that the latter are the extrusive equivalents of the plutonic members associated with the deeper-level gabbros. There is no suggestion that the volcanic members are more chemically evolved than the plutonic plagiogranites.
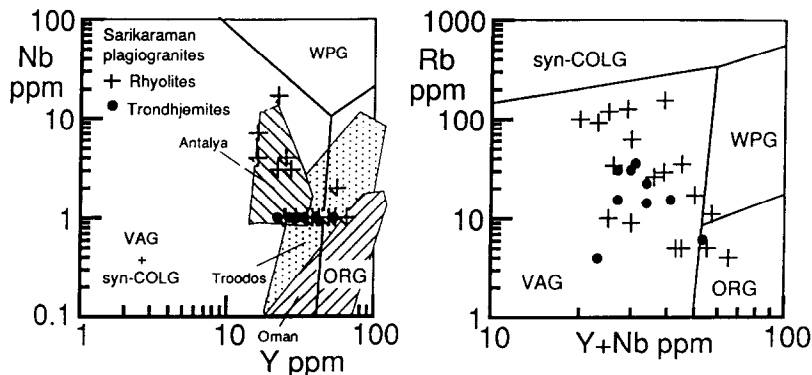
Fig. 5. Sarikaraman ophiolite plagiogranites plotted in tectonic environment discrimination diagrams for granitic rocks (after Pearce et al., 1984). In common with other Neotethyan plagiogranites they have features similar to both volcanic arc and ocean ridge granites: Troodos (Aldiss, 1978); Semail (Alabaster et al., 1982); Antalya, (Cocherie, 1978). ORG, ocean ridge granite field; VAG, volcanic arc granite field; WPG, within-plate granite field and syn-COLG, syn-collisional granite field.

7. Plagiogranite petrogenesis

Field evidence shows the close association of the plagiogranites with the upper levels of the isotropic gabbros in a variety of emplacement relationships from coexisting melts to intrusive contacts and netvein complexes. Acidic dykes also traverse the sheeted dyke complex and are present as extrusive

Fig. 6. Rare earth element patterns for ophiolitic plagiogranites (normalized values from Sun and McDonough, 1989). (a) Normalized patterns for typical Sarikaraman ophiolite basaltic rocks with associated trondhjemites and rhyolites. (b) Distinction between plagiogranites formed in arc-type environments by different postulated mechanisms: Canyon Mountain, hydrous melting of a basic source (Gerlach et al., 1981), and Samail, fractional crystallization (Pallister and Knight, 1981). (c) Model generation of acid melts by variable fractionation and partial melting of a Sarikaraman gabbro 'parent' and comparison with the range of Sarikaraman plagiogranite compositions observed. Distribution coefficients from compilation by Rollinson (1993).

lavas in the uppermost volcanic section of the ophiolite. The roof of the magma chamber is probably represented at or near the level of these upper gabbros and the site from which plagiogranite melts were generated. Acidic melts evolved at various times during the dynamic evolution of the magma chamber roof during which stopping and melting of altered (hydrated) gabbro also probably occurred (cf. Hopson et al., 1981). At the same time crystallization of minor magmatic amphibole and Fe-rich opaques in the associated gabbros would drive any fractionating liquid towards silicic compositions. Both these processes, partial melting of hydrated gabbro and extreme fractionation of a gabbroic liquid, could produce melts of plagiogranite composition in this environment. These processes will be evaluated in the light of the chemical variation displayed by the plagiogranites and in terms of the initial generation of the acidic melts, and their subsequent evolution.
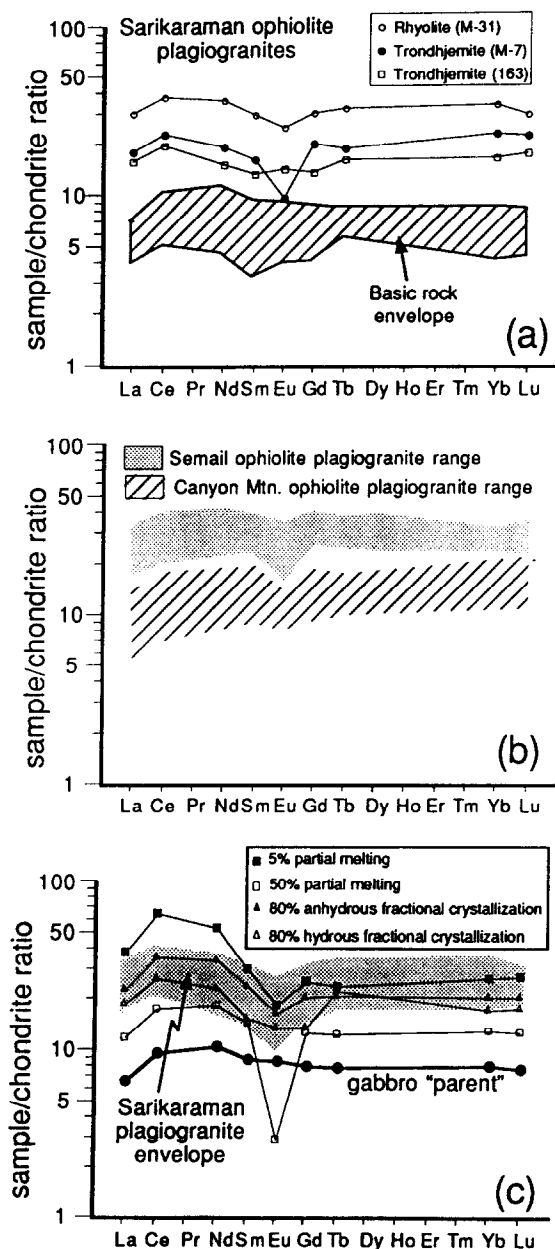
7.1. Extreme fractionation of basic parents

High levels of fractional crystallization (up to 80%) of ophiolitic basalt magmas have been suggested to generate plagiogranitic melts (e.g. Stern, 1979; Pallister and Knight, 1981; Jafri et al., 1995), either from gabbroic or sheeted dyke parental compositions. In a similar manner, rhyolites of oceanic islands such as Iceland where no continental crust is involved, have been modelled by the extreme fractionation of tholeiitic parental liquids (e.g. Furman et al., 1992). In the case of the Sarikaraman plagiogranites, the upper gabbros are considered to approximate the initial composition that undergoes fractionation.
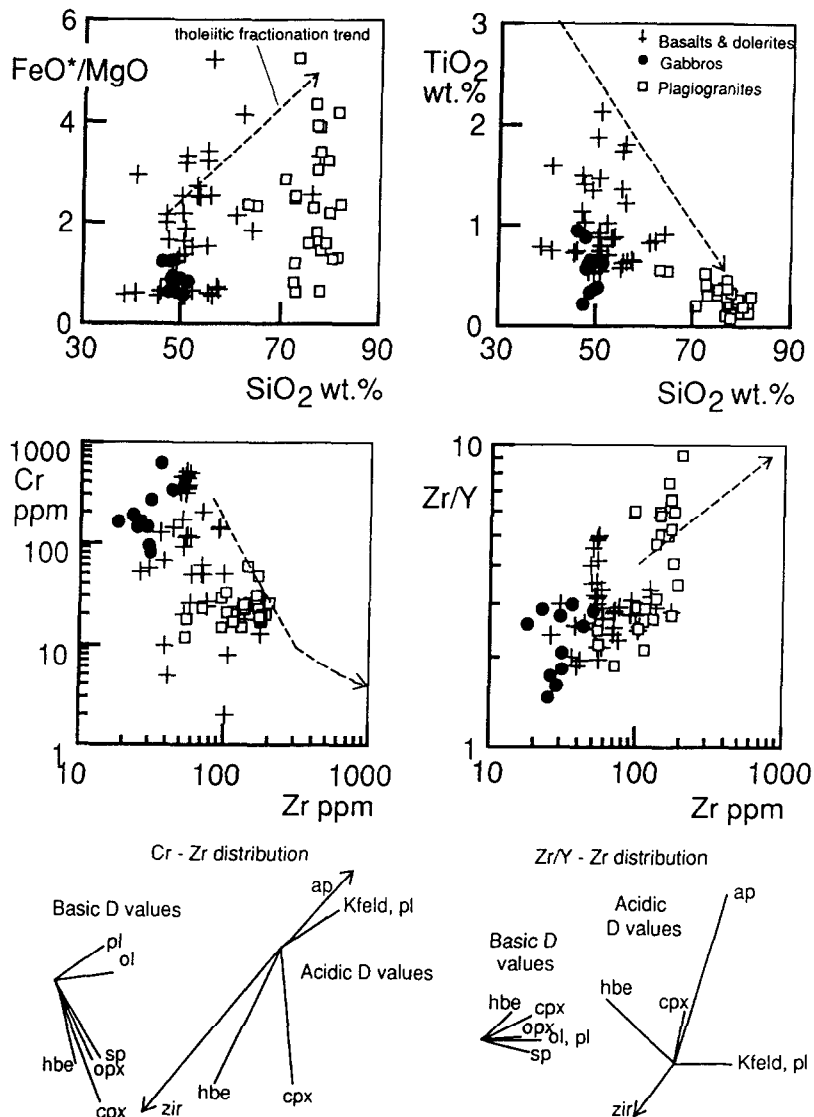
Fig. 7. Comparison of selected major oxide and trace element characteristics of Sarikaraman ophiolite basaltic components (gabbros, dolerite dykes, basalt lavas) and associated plagiogranites relative to progressive fractionation, as measured by increasing SiO_2 and Zr contents. Fractionation mineral vectors (for $F = 0.4$) for basic and acidic compositions are illustrated for Cr–Zr and Zr/Y–Zr relationships (distribution coefficients from compilation by Rollinson, 1993). The dashed line with arrow represents a generalized fractionation trend for Icelandic tholeiite suites ranging from basalt to rhyolite (data from Carmichael, 1964; Furman et al., 1992; Jonasson et al., 1992).

As seen in Fig. 7, plagiogranite variation is compared with the ophiolitic gabbros and dolerites/basalts from the dykes and lavas, relative to fractionation indices, SiO_2 and Zr. An averaged trend for an Icelandic basalt-to-rhyolite suite is used as a comparison for chemical variation due to fractionation. Apart from some of the variation being

due to alteration (scatter for FeO^*/MgO and maybe Cr) and differences in parental compositions (Zr), the ophiolitic suite does not appear to define typical linear trends indicative of fractionation from basaltic parents. Furthermore, there could be a compositional gap between the gabbro 'parents' and the 'differentiated' plagiogranites, although such features have been

recorded for comagmatic fractionated suites (e.g. Jonasson et al., 1992). However, the stable element plot, utilizing Zr/Y ratio versus Zr, is probably the most reliable in Fig. 7 and suggests there may be two groups of gabbros with variable Zr/Y ratio. The group with the lowest Zr/Y ratio (< 2) might represent a 'parental' composition to some of the plagiogranites that all lie on a similar trend of increasing Zr/Y with Zr. With reference to the mineral fractionation vectors, initial clinopyroxene $\pm$ amphibole fractionation (basic compositions) and then clinopyroxene–feldspar (acidic compositions), could generate the more primitive plagiogranites on this trend. However, the plagiogranites with a progressively increasing Zr/Y and decreasing Cr and TiO_2 at essentially constant Zr could not have been generated in this way (Fig. 7). These more evolved plagiogranite compositions probably reflect internal fractionation of the acidic melts. The additional influence of apatite (to produce high Zr/Y ratios) and amphibole fractionation (decreasing Cr, Ti) together with feldspar are probably responsible. However, although both hornblende and apatite are observed petrographically in these rocks they are only present in relative minor proportions. Further evidence for some degree of internal fractionation is shown by the relatively parallel REE patterns for the trondhjemites (Fig. 6a) that with a progressive increase in total REE develop a negative Eu anomaly indicative of feldspar fractionation.

On the basis that some of the plagiogranites could be generated by fractional crystallization of the gabbros, normalized REE patterns were modelled for 80% fractionation from both an anhydrous gabbro 'parent' (fractionating in the proportions $ol_{10}cpx_{40}pl_{50}$) and a hydrous gabbro 'parent' with amphibole and minor apatite (fractionating in the proportions $ol_5cpx_{35}pl_{50}hbe_{9.5}ap_{0.5}$). As seen in Fig. 6(c) both modelled patterns fall within the plagiogranite envelop. Too much amphibole and/or apatite fractionation suppresses REE enrichment (moving the patterns below the plagiogranite envelope) so these phases probably only played a minor role, as suggested by their paucity in the actual rocks. Of the two, fractionation of gabbro by largely anhydrous phases (between about 70 to 85%) covers much of the plagiogranite range, including the development of a small negative Eu anomaly.

In summary, largely anhydrous phase fractionation of gabbro could produce some of the variation shown by the plagiogranite, although internal fractionation by the acid melt (involving minor amphibole and apatite) is required to cover the full range.

7.2. Partial melting of basic source rocks

Experimental data for the hydrous partial melting of basalts and the dehydration melting of amphibolite (where water is only supplied by the breakdown of amphibole) has shown that both processes can
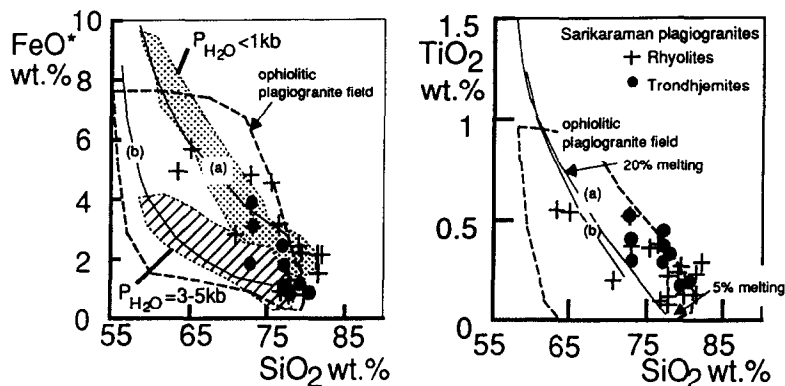

Fig. 8. Distribution of total Fe (as FeO) and TiO_2 within Sarikaraman plagiogranites relative to progressive fractionation as measured by increasing SiO_2 content. The distribution can be compared with experimentally produced melts at variable water pressures (patterned envelopes; data compiled by Thy et al., 1990) and progressive melting of different basaltic compositions: line (a) = MORB (Spulber and Rutherford, 1983) with compositions for 20 and 5% melting marked for TiO_2 and line (b) = Kilauea tholeiite (Helz, 1976). Ophiolitic plagiogranite field from Gerlach et al. (1981) and references therein.
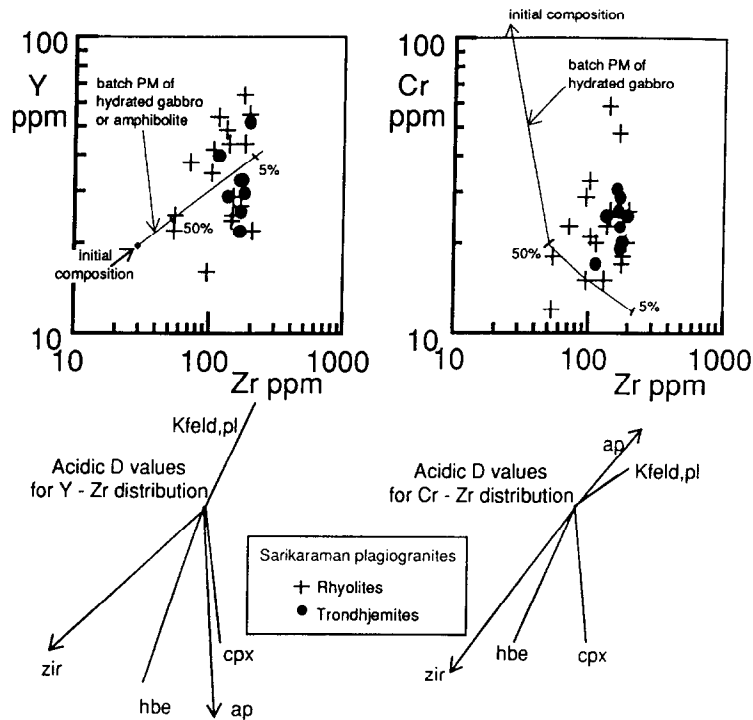
Fig. 9. Distribution of Y and Cr within Sarikaraman plagiogranites relative to increasing Zr content. Also shown are modelled partial melting trends for a Sarikaraman gabbro source containing minor amphibole (hydrated gabbro). An amphibolitic mineralogy for the same initial composition produces similar melting trends. Composition points for 50 and 5% melting are indicated. Fractionation mineral vectors (for $F = 0.4$) for acidic compositions are illustrated for Y–Zr and Cr–Zr relationships (distribution coefficients from compilation by Rollinson, 1993).

produce melts of broadly acidic composition (e.g. Helz, 1976; Spulber and Rutherford, 1983; Beard and Lofgren, 1991; Wyllie and Wolf, 1993). On this basis, the melting of hydrated or hydrothermally altered ocean crust has been suggested as a mechanism for the generation of Icelandic rhyolites (Spulber and Rutherford, 1983; Thy et al., 1990) and the partial melting of amphibolites to produce Archean trondhjemites (Rapp et al., 1991). The experimentally produced silicic melt compositions also compare favourably with oceanic plagiogranites and this origin has been suggested for some ophiolitic trondhjemites (e.g. Gerlach et al., 1981). As demonstrated by Beard and Lofgren (1991) water has a marked effect on melt compositions. Water-saturated melting produces peraluminous melts (with amphibole-rich restite) that are generally unlike natural silicic rocks, whereas dehydration melting of altered basalts can simulate typical oceanic low-K acid melts.

In this context the Sarikaraman plagiogranites were compared with experimentally generated compositions, although the gabbro 'source material' has only a small proportion of amphibole (1–5%) available for dehydration melting. However, as hornblende is the main mafic mineral (up to 10%) in the plagiogranites, it is assumed that melting (and crystallization) took place under hydrous conditions. Fig. 7 shows the distribution of the Sarikaraman plagiogranites relative to a general ophiolitic plagiogranite field and experimental melts produced at low ($P_{H_2O} < 1$ kb, $P_{total} = 1$–3 kb) and higher ($P_{H_2O} = 3$–5 kb, $P_{total} = 3$–5 kb) water pressures. Curves representing progressive melting compositions are also shown with low degrees of melting producing the most silicic compositions (Fig. 8, TiO_2–SiO_2 diagram). Allowing for secondary silification, which places some compositions outside the normal field, the Sarikaraman plagiogranites define a trend ap-

proximating that of low water pressure generated acid melts derived from MORB compositions. The major oxide data suggest that the plagiogranites could form by variable, but small degrees of melting of a hydrated gabbro. This was tested by trace element modelling for REE (Fig. 6c) and Y, Cr and Zr distributions (Fig. 9). The initial starting composition was taken as a typical high-level gabbro with a mineralogical composition of $cpx_{40}pl_{55}hbe_5$, normalized REE distribution as in Fig. 6(c) and 148 ppm Cr, 18 ppm Y and 29 ppm Zr. As seen in Fig. 9 simple modal batch melting of between 5 and 50% cannot provide the full range of plagiogranite compositions observed. However, if the initial ('parental') plagiogranite melts were produced in this way (by about 5–15% melting), then subsequent fractionation of feldspar, minor apatite and/or hornblende would be required to produce the internal variation (compare with mineral fractionation vectors). The REE patterns generated by 5 and 50% batch melting are also shown in Fig. 6(c), but the degree of enrichment calculated for the light REE at low degrees is too high and clearly outside the natural plagiogranite envelope.

In summary, both fractional crystallization and dehydration partial melting of gabbroic 'parent material' could provide the *initial* generation of plagiogranite melts. However, the former process (with variable amphibole fractionation) is more likely to produce most of the range of plagiogranite compositions observed. Any internal variation represented by more evolved compositions is a consequence of fractional crystallization involving feldspar, with minor apatite and amphibole. Intimate field relationships (suggesting co-existing liquids, resorption) and lithological gradation between the deep-seated plagiogranites and the associated gabbros, coupled with the broad similarity of their REE patterns generally favour the fractionation model. However, in a dynamic spreading centre the periodic intrusion of further basic melt batches would provide a supply of heat to readily cause dehydration melting of existing altered gabbro.

8. Conclusions

(1) Petrographic and geochemical data for plagiogranites associated with the Sarikaraman ophiolite of the Central Anatolian Crystalline Complex closely resemble other ophiolitic plagiogranites, especially those from supra-subduction zone-type ophiolites of the northern branch of Neotethys.

(2) The Sarikaraman plagiogranite suite is composed of intrusive trondhjemites and extrusive rhyolites, although the latter have been modified by secondary silification. Various episodes and modes of emplacement are seen from netvein complexes between plagiogranite and high-level gabbros, to intrusive dykes that presumably fed the acid lavas in the volcanic section of the ophiolite. As the trondhjemites are largely found concentrated near the top of the gabbro section they were probably generated from the roof section of a dynamic and evolving magma chamber.

(3) Geochemically the plagiogranites are characterised by (MORB-normalized) low HFS contents including small negative Nb–Ta anomalies and variable LIL element contents, the latter of which (in part) reflect low-grade alteration. Normalized REE patterns are uniformly enhanced relative to the basic components of the complex, but have similar patterns exhibiting mild light REE depletion. Overall a degree of internal chemical variation is present, although there is complete overlap in composition between the trondhjemites and rhyolites and no variation with stratigraphic height was noted.

(4) In terms of their origin, the Sarikaraman plagiogranites were compared with the acidic members of basalt–rhyolite suites generated by progressive fractional crystallization and experimentally-derived silicic melts produced by dehydration melting of altered basalts. The initial or most primitive plagiogranite melts could have been generated by either fractional crystallization (70–85%) or partial melting (5–15%) of a gabbroic 'source material', although only the first process can produce most of the range of the plagiogranite compositions. As a group the plagiogranites exhibit some degree of internal variation which can be generated by further fractionation largely dominated by feldspar with minor apatite and amphibole.

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