

Petrochemistry of a Late Precambrian garnetiferous granite, pegmatite and aplite, southern Israel

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Abstract

Garnet is a widespread minor accessory mineral in the Late Proterozoic Elat-Quarry granite of southern Israel and is more abundant in the associated pegmatite and aplite. All garnets are dominated by almandine and spessartine end-members. Granite-hosted garnets are zoned with relative enrichment of Mn in the core and Fe in the rim. The chemistry of the garnet in the pegmatite and aplite are comparable to the rim compositions of garnets in the granite, but with a slight Fe-depletion at the rims. Geochemical parameters for the granite indicate fractional crystallization largely of an S-type source magma to a peraluminous composition.

In the highly evolved granite magma, Fe is relatively diminished, and only small amounts of biotite can crystallize. Manganese becomes a compatible element forming Mn-rich garnet (cores), and reducing the Mn content in the magma, subsequently leading to Fe-enriched rims. The greater abundance of garnet in the pegmatite and aplite (and its larger crystal-size in the former) relate to the enhanced presence of a hydrous fluid within the magma. The tendency for garnet crystals to concentrate in bands is much more developed in the pegmatite than in the granite, and is associated with the effects of hydrofracturing (fracture-filling), and the crystallization of coarse-grained alkaline feldspars.

KEYWORDS: garnet, granite, pegmatite, aplite, Israel.

Introduction

GARNETS, although not abundant, occur in a wide variety of granitoid rocks and their volcanic equivalents. Several different mechanisms have been proposed to explain their presence in these rocks. In certain cases, they have been related to allochthonous sources including restite, xenocrysts carried from depth of origin, and country rock xenocrysts (Abbot, 1981; Whitworth, 1992). Where garnet occurs in syntectonic granitoids, a metamorphic origin has also been proposed (Weigand *et al.*, 1981). However, it is widely recognized that garnet may crystallize directly at various stages from a granitoid melt ('magmatic garnet').

This study documents the nature of garnet in a late Proterozoic peraluminous granite and associated pegmatite and aplite in southern Israel.

Regional geology

In terms of its regional setting, the granite is considered part of an extensive calc-alkaline granitoid suite associated with the last stage of regional metamorphism in the northern part of the Arabo-Nubian Shield (Bentor, 1985; Garfunkel, 1980).

The orogenic cycle of the Late Proterozoic Arabo-Nubian Shield first involved different stages of accretion resulting from the juxtaposition of juvenile island arc belts along ophiolite-bearing suture zones (Stoesser and Camp, 1985; Stern, 1993). This was followed by large scale calc-alkaline plutonism, cratonization and ultimately continental collision leading to accretion with the Archaen to early Proterozoic African plate. The latest Precambrian magmatic phases are represented by post-accretion non-metamorphosed volcanics (mainly basalts) in

places with continental affinities (Stern, 1993) and at least in southern Israel with subduction affinities (Bogoch *et al.*, 1993).

Field description and petrography

The Elat-Quarry (E-Q) granite pluton occurs in the Nahal Shelomo area of southern Israel, approximately 3 km west of the town of Elat. It is a small intrusion of 2 km², however, the exposure is almost 100%. The intrusive contact with greenschist facies pelitic schists to the south and north is sharp and at least in the south is conformable with the structure of the local schistosity. The contact to the west is a fault zone and partially indurated Pliocene-Pleistocene conglomerates cover the Precambrian exposures to the east. Several N–W trending, K-feldspar-quartz porphyritic dykes and lesser dykes of intermediate composition cross the exposure.

The granite is pale grey and consists mainly of K-feldspar (~40%), quartz (~35%), albite–oligoclase (~20%) and biotite (~3–4%). It is medium- to coarse-grained (feldspars and quartz, 2–4 mm; biotite 0.5–1 mm) and non-porphyritic. Locally, very coarse K-feldspars (~1 cm and more) are present, mainly in clusters, which may have formed within a fluid-enriched portion of the intrusion.

The plagioclase and biotite have well-developed tabular forms whereas the quartz and microcline are generally anhedral. The K-feldspar contains between traces and 10% of irregularly distributed, parallel perthitic bands of albite. The biotite occurs either in elongated monomineralic, non-linear aggregates adjacent to feldspar or quartz (Fig. 1), or as single grains in apparent equilibrium contact with K-feldspar and quartz. However, both biotite and plagioclase also exhibit resorbed boundaries when in contact with the K-feldspar and quartz. The biotite is partly altered to hematite and 'limonite'. Sericite is irregularly present on the feldspars, and a few muscovite flakes were noted, generally in association with biotite, which are considered as late alterations. Accessory minerals include apparently late crystallized monazite (generally with biotite) and zircon, with lesser apatite, xenotime, and titanite.

Masking the primary texture of the E-Q granite, and particularly near grain boundaries, are several features related to the syntectonic intrusion of the granite, and to sub-solidus modifications. All quartz grains and most of the K-feldspar exhibit undulose extinction. Both feldspars and quartz contain microfractures, and many of these, as well as along grain boundaries, contain tiny, apparently recrystallized grains of quartz and minor K-feldspar. In places where they are slightly larger in grain size (~0.2 mm), they exhibit triple point junctions, and the undulose extinction is less pronounced. Certain

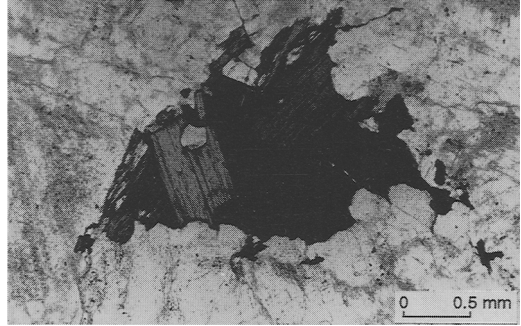


FIG. 1. Photomicrograph of biotite 'clot' within the E-Q granite (plain light). Note presence of some resorbed boundaries at edge of biotite crystals.

quartz–quartz boundaries are serrate, also suggesting solid phase recrystallization. Myrmekite is present in most thin sections, suggesting the presence of late-stage fluid circulation.

Within this massive granite are a few irregular patches (up to ~2 m²) of granite with diffuse contacts with the host. The granite in these patches consists of the same mineralogical composition, but exhibits linear (non-penetrative) biotite grains within a 'normal' granite matrix. In places, this rock is bimodal in grain size with coarse K-feldspar and quartz whose long dimension parallels the linear biotite, in a fine- to medium-grained dense matrix of interlocked quartz, K-feldspar and plagioclase. Pitcher (1987) argued that simple mineral lineations in granites are largely the result of syn-tectonic intrusion.

Very rare enclaves of the country-rock schist are present in the granite mainly near its southern border.

Pegmatite and subordinate aplite are widely dispersed, and total some 5% of the rock. They occur as: (i) sub-horizontal dyke-like bodies (in places, 2–5 in parallel), reaching 50 cm in width, or as (ii) irregular-shaped patches up to ~1 m in maximum dimension. The sub-horizontal disposition suggests fracturing in response to horizontal pressure. No mineralogical zoning in either type of occurrence was observed. Aplite dykes are relatively rare and generally not in contact with the pegmatite. Temporal relationships between the pegmatite and aplite are thus unclear.

Feldspar is the dominant mineral in the pegmatite, attaining several cm in size as do the quartz and rare biotite flakes. The pegmatite is coarsely crystalline in several exposures, generally with a matrix similar in grain-size and composition to the granite. Biotite is present in both the pegmatite and aplite, in lesser amounts than in the granite. Many of the pegmatite

bodies contain abundant, generally sub-mm sized, reddish brown garnets.

The aplite is fine-grained and contains more quartz (~40% quartz, 30% K-feldspar, 30% plagioclase) than the granite. Grain boundaries are mainly straight, although curvilinear and more rarely serrated contacts are present.

Garnets

Within the pegmatite, garnet is universally present in concentrations of at least several crystals per hand specimen, and reaching 2–3 mm in diameter. Several exposures of pegmatite contain one or more discontinuous bands generally 2–3 mm in width, consisting of a dense concentration of dark red garnet crystals (Fig. 2). In most cases, the disposition of the garnet-rich bands in the pegmatite are structurally controlled within micro-fault zones, at the borders of large K-feldspar crystals and at aplite–pegmatite and pegmatite–granite contacts (Figs. 2, 3a–c).

Garnet is also sporadically visible to the naked eye within the granite. One to several crystals, generally with sub- to euhedral forms, were found in ~90% of the granite thin sections, and show equilibrium boundaries with both quartz and K-feldspar. Only one crystal was observed in contact with biotite, with no apparent reaction texture. Garnets are inclusion-free and vary in size from 0.1 to 1.0 mm (most 0.2–0.4 mm). In two samples of granite, three to six garnet crystals are aligned, with a spacing in the order of the 1–4 × the diameter of the garnet (Fig. 3d). A similar alignment of garnet was termed “train texture” by Macleod (1992) who considered that it may have resulted from magnetic properties of Fe-rich garnet rims.

Garnet is more abundant in the aplite and occurs in all samples (Fig. 3e). Its crystal-size is similar to that of the other minerals and contacts are sharp. The

pelitic schists are also garnetiferous, with crystals generally ranging from 1 to 5 mm in diameter.

Analytical methods

Powders of mineral concentrates were analysed using a Philips PW1730/1710 X-ray diffractometer with Cu-K radiation. Mineral analyses were made using a JEOL 840 SEM with an attached Link 10,000 EDS using the ZAF4 program for polished sections (comparison with standard block files). The accuracy is 5%. Major and trace elements were analysed by ICP-AES (Brenner *et al.*, 1980) and K₂O and Na₂O by AES. Accuracy for majors is better than 2%, and for trace elements, 10%. REE, Y, Cs, Hf, Ta, U, and Th were determined by INAA (Bergerioux *et al.*, 1979; Gibson and Jagan, 1980) using an ORTEC HPGe GMX-10180 coaxial photon detector wired to an ORTEC multichannel analyser for data collection and reduction. Accuracy and precision are 5% or better.

Mineral chemistry

Garnet. The spessartine component is significant in all analyses of garnet, consistent with the findings of previous studies of garnets from peraluminous granitoids. Most fall within, or close to the “igneous garnet field” of Barth and Ehlig (1988). Many of the larger garnet crystals within the granite are zoned (Table 1, Fig. 4). The smaller crystals and the rims of the larger crystals, which crystallized later and presumably more or less contemporaneously, are fairly homogeneous in their composition and average 59% almandine (57.5–62%), 34% spessartine (32–39%), 4% andradite (0.8–5.6%) and 3% pyrope (2–4%). The cores, which represent the first stage of garnet growth, are chemically more heterogeneous, with 46% almandine (37–57%), 51% spessartine (37–62%), 1.3% andradite (0.9–1.9%) and 1.7%

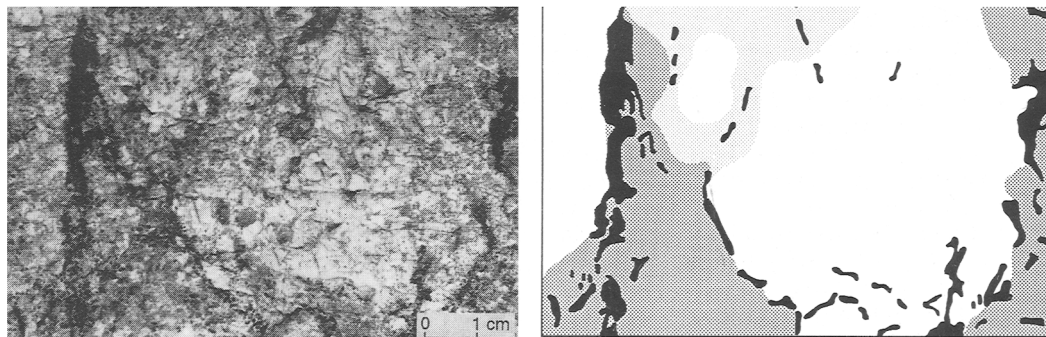


FIG. 2. Photograph of hand specimen of pegmatite exhibiting discontinuous garnet bands. Drawing: black bands are garnet-rich; medium-grey shading represents granitic ‘phase’ within pegmatite; light-grey shading is medium-grained; white represents the coarse-grained pegmatitic material.

TABLE I. Representative analyses of garnet

A. Garnet in granite								
	c - - - - r		c - - - - 1/3 - - - - r					
SiO ₂	34.46	34.71	34.47	34.45	33.93	34.28	34.60	
Al ₂ O ₃	20.14	19.62	19.67	19.77	19.92	19.95	20.48	
FeO	26.05	27.05	18.62	21.45	26.91	27.87	25.93	
MnO	16.73	14.79	25.35	22.48	14.91	15.29	17.18	
CaO	0.66	1.89	0.38	0.41	1.64	0.83	0.27	
MgO	1.11	0.66	0.42	0.44	1.02	0.53	0.82	
Total	99.15	98.72	98.92	99.00	98.33	98.75	99.28	
alm	56.9	59.7	39.7	47.1	58.6	61.4	57.4	
spes	37.0	33.1	57.7	50.0	32.9	34.2	38.6	
andr	1.9	5.3	1.0	1.2	4.5	2.3	3.2	
pyr	4.2	1.9	1.6	1.7	4.0	2.1	0.8	

	c - - - - r		c - - - - r		c - - - - r - - - - r - - - - r			
SiO ₂	34.56	34.63	34.27	34.62	34.54	34.45	34.35	34.76
Al ₂ O ₃	19.92	20.46	20.28	19.86	19.85	19.62	19.66	20.27
FeO	16.59	26.03	23.42	27.25	18.52	26.64	26.49	26.41
MnO	27.43	16.66	19.90	15.10	25.43	15.08	15.01	15.29
CaO	0.34	0.78	0.33	1.34	0.66	2.01	1.93	1.68
MgO	<0.1	0.56	0.60	0.97	nd	0.82	0.88	0.76
Total	98.84	99.12	98.80	99.14	99.00	99.63	99.32	99.17
alm	37.4	58.0	52.0	59.2	41.1	58.0	57.9	58.2
spes	62.4	37.6	44.7	33.3	57.1	33.2	33.3	34.1
andr	0.2	2.2	0.9	3.7	2.8	5.6	5.4	4.7
pyr	—	2.2	2.4	3.8	—	3.2	3.4	3.0

B. Garnet in pegmatite								
	c - - - - r		c - - - - r		c - - - - r		c - - - - r	
SiO ₂	34.25	34.73	35.12	34.82	35.18	34.84	34.42	34.62
Al ₂ O ₃	19.54	19.33	19.61	19.96	19.55	19.53	19.26	19.32
FeO	28.96	27.73	28.82	27.73	27.87	27.73	29.73	28.28
MnO	14.49	14.54	14.47	14.70	14.31	14.56	14.72	15.29
CaO	0.35	1.62	0.74	2.03	1.50	1.78	0.59	1.42
MgO	0.64	0.58	0.44	0.57	0.57	0.50	0.55	0.61
Total	98.77	98.59	99.20	99.83	99.00	98.98	99.27	99.74
alm	64.0	60.8	63.7	60.0	61.5	60.7	64.1	60.6
spes	32.5	32.3	32.5	32.2	32.0	32.3	32.2	33.2
andr	1.0	4.6	2.1	5.6	4.2	5.0	1.6	3.9
pyr	2.5	2.3	1.7	2.2	2.3	2.0	2.1	2.3

(contd. on next page)

pyrope (traces—4.5%). In all zoned crystals in the granite, the rims are consistently enriched in almandine, and andradite (and in most cases in pyrope), and depleted in spessartine, relative to the cores.

In general, the values in the pegmatite and aplite are comparable to the rim compositions of garnets in the granite (~60% alm; ~30% spess.). Zoning is not conspicuous although there is a slight tendency for Fe-poor rim compositions (Fig. 4).

TABLE 1. (contd.)

B. Garnet in pegmatite (contd.)						
	c - - - - r		c - - - - r			
SiO ₂	35.12	34.83	35.31	34.94	34.91	35.72
Al ₂ O ₃	19.56	19.08	20.45	19.81	19.88	20.85
FeO	28.09	27.74	28.44	28.30	28.01	27.81
MnO	14.95	14.93	14.51	14.10	14.28	14.30
CaO	1.42	1.66	0.76	1.32	2.49	0.70
MgO	0.58	0.10	0.88	0.55	0.71	0.81
Total	99.72	98.36	100.35	99.02	100.28	100.14
alm	61.0	61.4	62.3	62.5	59.7	62.3
spes	32.9	33.5	32.2	31.6	30.8	32.5
andr	3.9	4.7	2.1	3.7	6.8	2.0
pyr	2.2	0.4	3.4	2.2	2.7	3.2

C. Garnet in aplite										
	c - - - - r		c - - - - r		c - - - - r		c - - - - r		c - - - - r	
SiO ₂	35.40	36.01	35.50	35.55	34.89	34.81	34.84	36.12	34.70	35.13
Al ₂ O ₃	20.34	20.73	20.16	20.97	19.96	19.75	19.94	20.56	19.72	20.44
FeO	29.42	26.48	27.89	27.04	27.96	27.58	27.94	27.12	28.22	27.91
MnO	13.60	14.91	13.58	14.80	13.78	14.76	13.96	13.94	13.45	13.43
CaO	2.02	1.91	2.15	1.95	1.86	2.01	1.99	1.93	2.18	1.88
MgO	0.73	0.42	0.55	0.57	0.75	0.61	0.65	0.52	0.79	0.78
Total	101.51	100.46	98.83	100.88	99.20	99.52	99.32	100.21	99.06	99.57
alm	62.5	59.1	61.5	59.3	61.3	59.7	61.1	60.7	61.2	61.6
spes	29.3	33.7	30.3	33.0	30.5	32.4	30.9	31.7	29.6	30.0
andr	5.4	5.5	6.1	5.5	5.2	5.6	5.5	5.5	6.1	5.4
pyr	2.8	1.7	2.1	2.2	3.0	2.3	2.5	2.1	3.1	3.0

c - core; r - rim

Garnets in the pelitic schist contain considerably more of the almandine (~72%) and pyrope (~10%), and less of the spessartine (~15%) molecules.

Other minerals. Analyses of the other minerals in the suite show that the Na₂O content of the K-feldspar ranges between 0.7 and 2% (Ab 7–17), with most grains measuring between 1 and 1.5%. The plagioclase is fairly homogeneous in composition, mainly a low-An oligoclase, and clear-cut zoning was not observed. The biotites (both in clots and as single grains) plot in the peraluminous magmatic biotite fields of Abdel-Rahman (1994), with no apparent enrichment in MnO (average ~0.8%). Their aluminum saturation index (ASI = A/CNK) is generally >1.4, a value for igneous biotite which does not co-exist with hornblende (Zen, 1988). The Mg/(Mg+Fe) ratio is typical of biotite from peraluminous granites (see discussion in France-Lenord and Le Fort, 1988). Monazite has a high La:Nd ratio (>1.5), unlike those found in highly

differentiated granites (Forster and Rhede, 1995), where the LREE pattern tends to be more horizontal with La:Nd ratios less than 1. Based on two analyses of monazite crystals, its approximate formula is: (La_{0.20}Ce_{0.35}Nd_{0.13}Sm_{0.02}Th_{0.04}Si_{0.07})_{0.81}PO₄.

Whole rock geochemistry

Representative whole rock and trace element analyses of the granite, pegmatite and aplite are presented in Table 2. The chemical data for the granite indicate a rather restricted composition with SiO₂ values between 73.9 and 76.7%, and low contents of MgO, CaO, TiO₂ and P₂O₅. In these aspects, it is similar to that of a minimum melt granite (Harrison, 1988). The iron (1.0–2.5% Fe₂O₃) can largely be accounted for in the biotite and its alteration products. The generally higher MnO values in the pegmatite and aplite relate to higher concentrations of garnet.

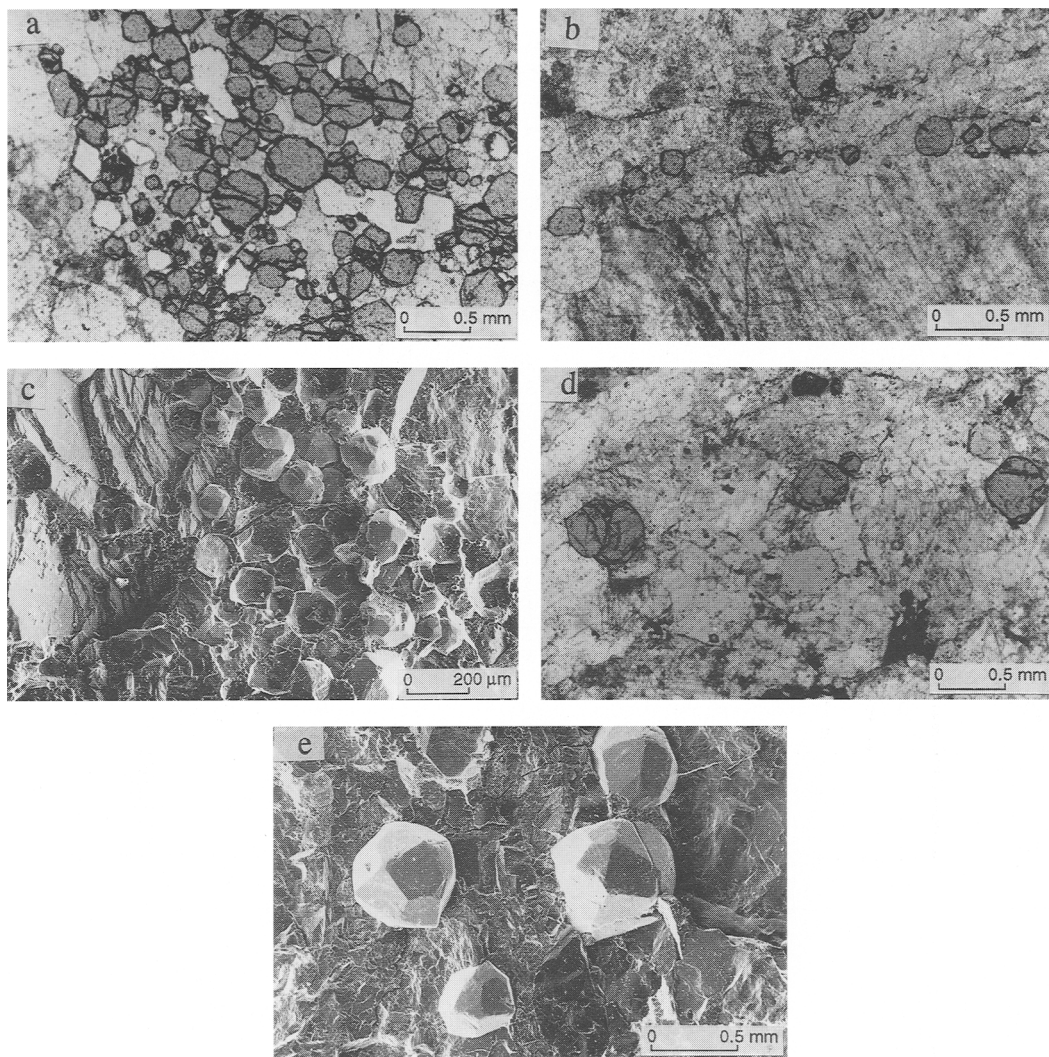


FIG. 3. (a) Photomicrograph of band of garnet crystals in micro-fault zone within pegmatite, plain light; (b) Photomicrograph of 'train' of garnet crystals within pegmatite at contact with coarse K-feldspar and within micro-crack, plain light; (c) SEM photograph of band of garnet crystals within pegmatite body adjacent to large K-feldspar crystal; (d) Photomicrograph of aligned garnet crystals within granite, plain light. This type of alignment is rare in the granite; (e) SEM photograph of unusually abundant garnet euhedra within aplite.

There is an apparent increase in the $\text{Na}_2\text{O}:\text{K}_2\text{O}$ ratio as a function of increasing SiO_2 . Sodium and potassium contents are similar although the latter exhibits greater variations, which could relate in part to slightly varying biotite contents.

The chondrite-normalized REE patterns for the granite (Fig. 5) feature a moderate slope (LaN:YbN generally >10) with LaN values typically between

50–150 × chondrite. The La concentrations decline as a function of increasing silica. This is probably due to decreasing amounts of monazite in the more evolved rocks. The monazite shows a combined $(\text{La}_2\text{O}_3 + \text{Ce}_2\text{O}_3)$ concentration of approximately 45 wt%; therefore a significant proportion of the LREE in the rock must be stored in this mineral. There is a prominent negative

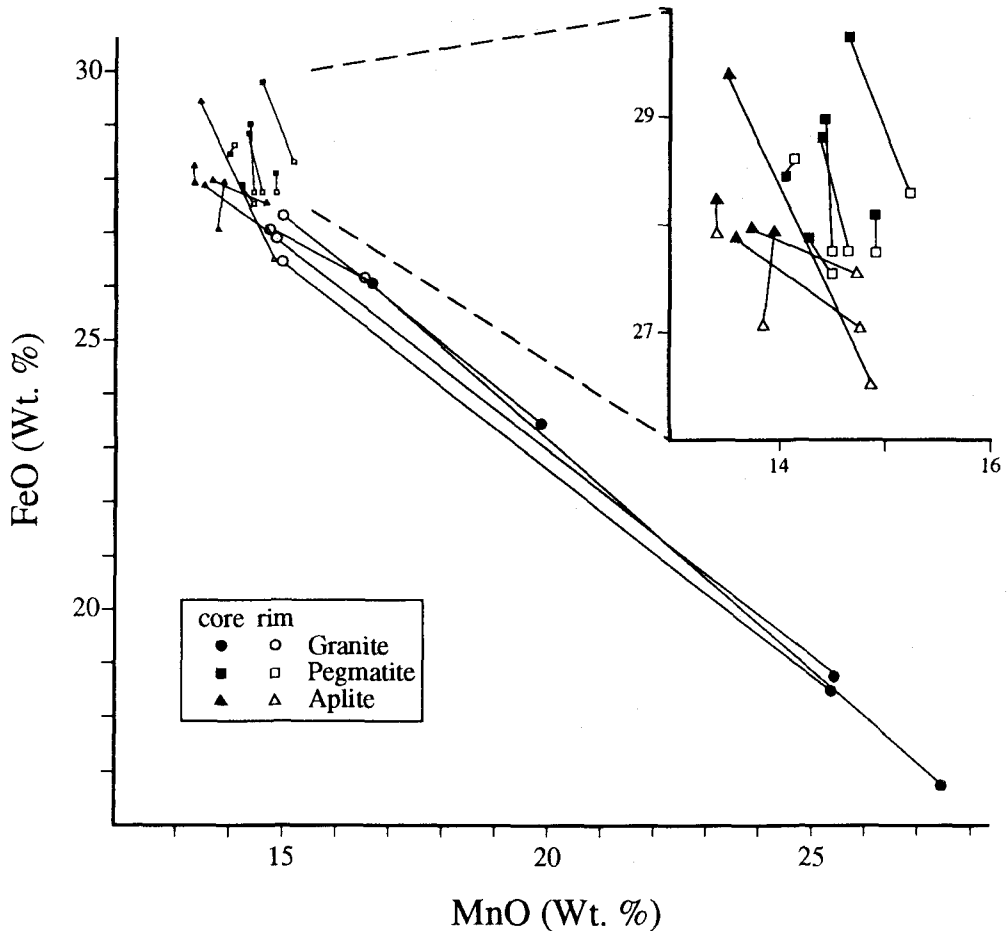


FIG. 4. FeO vs. MnO plot of zoned garnets.

europium anomaly in most samples ($Eh:Eu^* 0.25-0.55$), which is most likely unrelated to the monazite chemistry, and suggests fractional crystallization of plagioclase. This is supported by the low Sr values (Chappel *et al.*, 1987).

On the Rb-(Nb+Y) plot proposed by Pearce *et al.* (1984) to distinguish between syn-collision granites and volcanic arc granites, the E-Q granite falls just within the VAG field, possibly indicative of late loss of alkaline elements. The relative depletion in Ta, Nb, P and Ti, and enrichment in Rb, Th, K, and LREE suggest a calc-alkaline chemical affiliation. Rubidium and Th enrichments (relative to Nb and Ta), and Ce enrichment (relative to Nb and Hf), suggest crustal involvement (Pearce *et al.*, 1984), as possibly do the high Pb (and Zn) values. The chemical analyses of the granite are very similar to

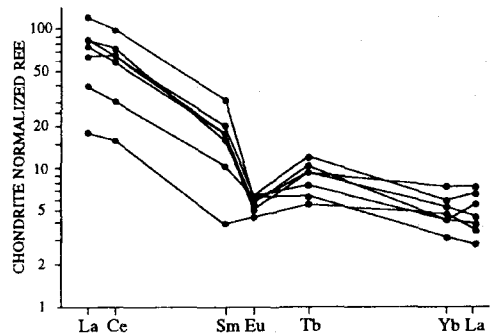


FIG. 5. Chondrite-normalized REE patterns for the granite.

TABLE 2. Whole rock analyses of the E-Q granite, pegmatite and aplite (wt.%)

	2	3	106	111	113	121	131	108	109	110b
SiO ₂	76.7	75.0	75.0	76.0	74.7	76.5	76.0	77.9	74.5	78.2
TiO ₂	0.2	0.1	0.13	0.07	0.14	0.1	0.1	0.03	0.03	0.03
Al ₂ O ₃	13.4	13.5	13.7	14.0	14.1	12.4	13.2	14.3	14.1	12.6
Fe ₂ O ₃	1.8	1.7	2.0	1.3	2.0	2.5	1.9	0.7	1.5	1.0
MnO	0.01	0.01	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.5	0.47	0.13
MgO	0.3	0.2	0.3	0.1	0.3	0.3	0.3	0.2	< 0.1	< 0.1
CaO	1.0	0.9	0.9	0.7	0.8	0.8	0.4	1.0	0.3	0.6
Na ₂ O	3.8	3.6	4.1	3.9	3.7	4.2	4.0	3.8	2.4	3.1
K ₂ O	3.2	4.1	3.0	3.9	3.7	2.7	3.4	1.8	7.4	4.3
P ₂ O ₅	< 0.1	< 0.1	0.1	0.1	—	< 0.1	< 0.1	—	—	—
LOI	0.3	0.4	0.6	0.5	0.4	0.4	0.5	0.9	< 0.1	0.2
Total	100.61	99.51	99.83	100.57	99.84	99.9	99.8	100.63	100.7	100.16
ASI	1.16	1.12	1.18	1.18	1.22	1.10	1.19	1.42	1.13	1.16
Qtz	38.79	35.23	36.55	36.32	35.83	38.23	37.41	45.92	30.31	41.73
Cor	1.81	1.44	2.25	2.27	2.49	1.04	2.15	4.13	1.52	1.69
Or	18.91	24.23	17.73	23.05	21.87	15.96	20.10	10.64	43.74	25.42
Ab	32.15	30.46	34.69	33.00	31.31	35.54	33.84	32.15	20.31	26.23
An	5.18	4.69	4.02	3.02	4.21	4.24	2.21	5.50	1.77	3.22
Hyp	2.69	2.32	2.85	1.64	2.83	3.48	2.78	1.26	2.56	1.35
Mag	0.52	0.49	0.58	0.38	0.58	0.72	0.55	0.20	0.43	0.29
Ilm	0.19	0.19	0.25	0.13	0.27	0.19	0.19	0.06	0.06	0.06
Apat			0.24	0.25						0.24
Zir	0.03	0.02				0.03				
Rb	101	134	112	116	133	103	96	42	208	131
Sr	100	9	90	85	85	81	94	370	70	75
Ba	654	820	680	670	755	891	705	1480	950	805
Zr	160	120					170	170		
Be			2.0	2.5	3.1			1.6	0.9	1.4
Y	14	11	8	13	16	13	10	3	27	50
La	27	13	27	6	40	26	21	4.2	3.1	10
Ce	56	27	61	14	86	54	57	8.1	8.4	23
Sm	3.6	2.2	3.4	0.81	6.5	3.6	4.1	0.53	0.71	2.34
Eu	0.42	0.5	0.45	0.36	0.5	0.4	0.5	0.79	0.30	0.33
Tb	0.51	0.35	0.55	0.30	0.64	0.5	0.42	0.06	0.46	0.89
Yb	1.20	0.75	1.00	1.10	1.30	1.0	1.00	0.41	5.5	8.3
Lu	0.16	0.10	0.14	0.13	0.24	0.26	0.20	0.06	0.74	1.20
Cz	2.9	2.0	3.1	2.0	2.5	0.8	2.9	2.7	2.0	1.4
Hf	3.8	3.2	4.2	2.1	5.0	4.2	4.1	1.1	4.8	5.8
Ta	0.5	0.5	0.5	< 0.5	1.0	< 0.5	< 0.5	< 0.5	< 0.5	< 0.5
U	1.6	1.0	1.3	1.9	3.4	1.6	2.3	0.7	2.2	5.1
Th	10	4.6	11	2.4	18	12	11	1.1	2.1	5.9
F	253	367	407	223			305			
Nb	0.3	0.3				< 0.3	0.5			

Samples 2 to 131 — granite; 108,109 — pegmatite; 110b — aplite

those of the average Himalayan Manaslu leucogranite (France-Lanord and Le Fort, 1988) in terms of the major element concentrations; however, the U and F values are distinctly lower and the Ba and LREE are higher in the E-Q granite.

Discussion

Of the plethora of petrogenetic classifications of granitoid rocks (Barbarin, 1990, lists twenty one), the S-, I-, M- and A-types are most frequently referred to

in the literature. This is perhaps because they provide an insight both to the tectonic framework and the nature of the dominant source. There are limitations in applying these classifications because numerous factors can affect the 'original' composition of a granitoid source (assuming that the bulk chemistry reflects a relatively large sample of the source, and that the rocks are not in part cumulates, Zen, 1988). Thus even for 'classic' S-type locations such as the Lachlan Fold belt in Australia, the data can yield quite different interpretations (e.g. Chappel, 1995 – restite model; Collins, 1995 – three component mixing). In addition, there are possibly limitations on the use of the 'popular' Pearce *et al.* (1984) discrimination diagrams (Tischendorf *et al.*, 1995). However, the high peraluminosity (ASI up to 1.22) and the presence of garnet in the E-Q granite appear to point to an S-type source.

In the restite model (Chappel and White, 1974) the composition of the granitoid is determined by a combination of the melt and trapped restite material (Chappel *et al.*, 1987). However, restite components may be difficult to recognize (Pitcher, 1987; Wall *et al.*, 1987; Zen, 1988). In the E-Q granite, there are three possible 'candidates' for restite: plagioclase, biotite clots, and garnet. The well-formed crystals and apparent lack of calcium-rich cores appears to rule out the plagioclase. Mineral clots (or aggregates or glomerocrysts) are the most commonly cited restite material (Chappel, 1987; Castro *et al.*, 1991; Wall *et al.*, 1987). Although most biotite in the E-Q granite occurs in early-formed monomineralic clots, they also do not represent restite since their chemistry is identical to individual biotite crystals, and the crystals are euhedral. The most likely explanation for the origin of the biotite clots is that of synneusis, i.e. the drifting together and attachment of crystals suspended in a melt (Hogan, 1993), possibly in order to minimize surface energy (Wall *et al.*, 1987 quoting E.B. Watson, 1987, pers. comm.).

There are several examples of granitoid-hosted garnets which have been interpreted as restite (e.g. Birch and Gleadow, 1974; Hamer and Moyes, 1982), and the restite garnets described by Pattison *et al.* (1982) have features considered typical of melt-derived garnets such as subhedrality and an inclusion-free (or inclusion-poor) nature. A restite (or other autochthonous source) for the E-Q garnets is unlikely if only for the reason that their formation was after the early formed plagioclase and biotite.

Peraluminous granitoids do not imply a specific source or mode of genesis. They may derive from I-, S- or A-type sources (Brown *et al.*, 1984; Anderson and Thomas, 1985), through partial melting (anatexis), especially of chemically similar sedimentary or igneous rocks, by differentiation of metaluminous magma, whatever its origin, with or without

assimilation, particularly involving amphibole fractionation, and by late stage loss of alkalis through vapor phase transport (Pitcher, 1987; Zen, 1988). The E-Q granite lacks two minerals commonly considered as definitive for strongly peraluminous rocks. Cordierite (White *et al.*, 1986) is absent, although Zen (1988) noted that cordierite-bearing granites are not necessarily more peraluminous, and (magmatic) cordierite can appear over wide ranges of bulk aluminosity. A lack (or rarity?) of primary muscovite may relate to the relatively low fluorine-values since high F tends to lower the solidus allowing the crystallization of muscovite (Zen, 1988).

There is a hint of subsolidus removal of alkaline elements since the rocks become less peraluminous as a function of increasing silica, although the lack of strong alteration suggests that this is most likely a minor effect.

Based on various discrimination diagrams and on multielement plots, much of the trace element data from the E-Q granite shows affinity with syntectonic (generally peraluminous, S-type granites, Pearce *et al.*, 1984) and mature continental arcs (S-type, Brown *et al.*, 1984). They differ from these mainly in their slightly lower Rb, Th, U, Ta, and very low Nb values, suggesting a contribution from a source (I-type?, calc-alkaline, Harrison, 1988) depleted in these elements.

Fractionation in the E-Q granite is indicated by relatively high Na-values (Thompson and Tracy, 1979), by the REE pattern, by the negative Eu-anomaly, and low Sr-values suggesting the early removal of plagioclase. This fractionation also contributes to chemical variations (Wall *et al.*, 1987), although an expected enrichment in Rb related to fractionation is not in evidence, which also suggests a Rb-depleted source. The low Sr values suggest either plagioclase fractionation or a crustal source.

Garnet. There is some degree of consensus regarding certain petrographic features of magmatic garnets which crystallized in granitoids, in that, like those in the E-Q granite, they tend to be eu- to subhedral, free or relatively free of inclusions, and are generally finely-crystalline (~<2 mm) (Allan and Clarke, 1981; Clemens and Wall, 1984; Wall *et al.*, 1987; Harrison, 1988; Speer and Becker, 1992; Macleod, 1992; Whitworth, 1992). It appears that the more mafic granitoids (e.g. granodiorites; monzodiorites) contain Fe-rich garnet (almandine-dominated; Green and Ringwood, 1968; Clemens and Wall, 1984; Barth and Ehlig, 1988), whereas the garnet within more evolved hosts, particularly peraluminous granites, contains an important proportion of spessartine (Allan and Clarke, 1981; Speer and Becker, 1992; Whitworth, 1992). Experimental studies of Green (1976, 1977) and Clemens and Wall (1981) tend to confirm a relationship between higher

P-T (almandine-rich) and lower *P-T* (spessartine-rich) garnet in rocks of granitoid composition.

Magmatic garnets are commonly zoned, although the chemical variations are not consistent, nor is there a relation between the nature of the zoning and the host rock (i.e. granite, pegmatite, aplite). Thus, there is frequently an antithetic relationship between Fe and Mn with relative enrichment of Mn in the core (Leake, 1967; Macleod, 1992), in the rim (du Bray, 1988), in both core and rim (even in the same pluton, Speer and Becker, 1992), or with no apparent Fe-Mn zoning (Vennum and Meyer, 1979). Similar relationships are also evident for Mg and Ca, relative to each other or to Fe and/or Mn. These contradictory data contrast with the view of Mn-enrichment in these evolved granitoids resulting in the formation of increasingly larger amounts of the spessartine molecule (Miyashiro, 1955), quoted in Deer *et al.*, 1982; Whitworth, 1992) which suggests that Mn-rim enrichment should be the norm. du Bray (1988), Macleod (1992) and others attributed zoning in garnets to fractional crystallization. In the E-Q granites, the garnets are consistently zoned with Fe > Mn in the rims, whereas in the pegmatite and aplite, there is a slight Fe-depletion in the rims.

Experimental studies have shown that almandine-pyrope-enriched garnets are normal products of (early magmatic) medium-pressure crystallization of peraluminous magmas (Clemens and Wall, 1981); with increasing addition of Mn, spessartine forms at relatively low pressures (~3 kbar) in granitic liquids (Green, 1977), and the garnet is zoned with >Mn (vs. Fe) on the rims. Since 'normal' zoning in garnets (Mn-cores, Fe-rims) is considered characteristic of increasing pressure and temperature associated with progressive metamorphism (Kretz, 1973; Green, 1977), a decrease in temperature and pressure would be expected to promote the incorporation of Mn relative to Fe. Abbott's (1981) AFM liquidus projections for late granite intrusions essentially predict an early Fe-rich garnet that evolves to a more Mn-rich composition. However, based on rim-core relations in the E-Q granite, the last phases of garnet growth involved an increase in Fe (and Ca) and a decrease in Mn.

The following possible scenario for the formation of the garnets in the E-Q granite may be envisaged:

(1) Mn behaves as an incompatible element increasing in concentration in the magma as the temperature falls and fractional crystallization continues. This is largely because biotite and ilmenite, which are the predominate mafic phases at this stage, incorporate Fe (and reject Mn);

(2) The magma evolves to peraluminous compositions and garnet crystallization commences. At first, a smaller amount of garnet is co-crystallizing with a much larger amount of biotite. Since biotite still

dominates, the Mn is still increasing in the magma, and the garnets forming ('early garnets') are Mn-rich. This stage could also produce reverse zoning in garnet as noted by du Bray (1988);

(3) As Fe diminishes, smaller amounts of biotite can crystallize. Mn becomes a compatible element forming Mn-rich garnet (cores), and reducing the Mn content in the magma, subsequently leading to Fe-enriched rims.

Reverse zoning results when the volume of biotite is much greater than that of garnet, whereas the normal pattern in evolved granites develops when that ratio drops considerably. Thus, the type of pattern depends on where you are in the crystallization sequence: early garnets have reversed patterns, later garnets have irregular or no zoning, and the latest garnets (e.g. the E-Q granite) have normal zoning.

The differences between garnets in the granite and those in the pegmatite and aplite are the composition, amount present and crystal size. The composition of the garnets in the pegmatite and aplite are similar to rim compositions of garnets in the granite although the subtle zoning in the former indicates a slight drop in the Fe-content of the fluid during garnet growth. Since relatively little biotite crystallized in the pegmatite and aplite, this may have resulted in greater Fe-compatibility in the early stages (cores) of garnet growth. If the aplite/pegmatite crystallized from a magma formed of melt and co-existing hydrous fluid as shown by Jahns and Burnham (1969), it appears likely that the enhanced volume of garnet in these rocks relates to the presence of the aqueous phase. One explanation for the upward mobility of Mn in rhyolite magma was attributed by Hildreth (1981) to hydroxyl complexing. In the aplite, the garnet crystal size is controlled by rapid growth of all phases in a confined space; with increasing water (+volatile) content in the pegmatite (Jahns and Burnham, 1969; Stern *et al.*, 1986), the crystal-size of the garnet is also increased.

The tendency for garnet crystals to concentrate in 'trains' or bands is much more developed in the pegmatite than in the granite. These garnets formed within cracks as well as along contacts between pegmatitic and granitic material, suggesting that the fluid phase had a hydrofracturing effect as well as assisting in the development of large feldspar and quartz crystals, and playing a role in late garnet crystallization. Possibly the rapid development of large feldspars in the pegmatite had the effect of pushing aside available garnet nuclei, concentrating them along margins as well as within the fractures.

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