

Magmatic and Postmagmatic Processes in Tin-mineralized Granites: Topaz-bearing Leucogranite in the Eurajoki Rapakivi Granite Stock, Finland

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The 1·57 Ga Eurajoki stock in southwestern Finland is composed of biotite–hornblende–fayalite granite, biotite granites and late-stage topaz-bearing granite, as well as related topaz-bearing rhyolite (ongonite) dykes. Miarolitic cavities indicate the presence of a separate fluid phase during late stages of crystallization of the topaz-bearing granite, and greisen-type Sn–Be–W–Zn mineralization is closely associated with it. Subsolvus reactions have modified the petrography and geochemistry of the topaz-bearing granite. The presence of magmatic topaz is indicated by petrographic features and crystallized melt inclusions entrapped in topaz. The magmatic origin of accessory cassiterite in the topaz-bearing granite (average 80 ppm Sn) is indicated by mode of occurrence and chemical composition of cassiterite: the Nb₂O₅ + Ta₂O₅ content of cassiterite is 4·8–8·7 wt % in granite, 2·3–3% in pegmatite and 0·0–0·6% in greisen. Some of the rhyolite dykes contain topaz as phenocrysts and as small prismatic crystals in the groundmass showing locally fluidal texture. These features indicate that the late-stage magma was highly enriched in F and Sn. Strong exsolution of alkali feldspar has produced intergranular albite rims and grain rows. De-anorthitization of primary plagioclase has produced turbid albite with inclusions of topaz, fluorite and quartz. Metasomatic albitization of K-feldspar and K-feldspathization of plagioclase have only locally been important. In greisenized granite, metasomatic addition of topaz and quartz is common. The topaz-bearing granite had originally anomalous geochemistry and mineralogy, and its anomalous character was further increased during postmagmatic fluid–rock interaction. The petrographic and geochemical peculiarities of the topaz-bearing granite of Eurajoki are characteristic of many other late-stage intrusive phases of rapakivi complexes and for many tin granites of different ages. In interpreting the origin of such highly evolved granites, the role of Na-metasomatic albitization has often

been overemphasized at the expense of exsolution and re-crystallization.

KEY WORDS: albitization; cassiterite; postmagmatic processes; rapakivi granite; topaz

INTRODUCTION

Primary tin mineralization is commonly associated with peraluminous leucocratic granites which represent late phases of late-orogenic to anorogenic granite complexes. These ‘tin granites’ or ‘stanniferous granites’ show petrographic and geochemical characteristics (e.g. occurrence of topaz, Li–Fe mica and albite; high F, Li, Rb, Ga, Sn and Nb; low Mg, Ti, Zr, Ba, Sr and Eu) which can be used to distinguish them from barren granites (Tischendorf, 1977). The origin of these geochemically specialized granites has been debated for decades. Several researchers have considered them to be products of intense postmagmatic metasomatism, and have especially emphasized albitization and topazization (e.g. Beus *et al.*, 1962; Stemprok, 1971), whereas others have interpreted them to be essentially magmatic rocks (Kovalenko *et al.*, 1971; Mackenzie *et al.*, 1988; Cuney *et al.*, 1992; Taylor, 1992), and some researchers have considered them to be ‘specialized’ magmatic granites whose petrography and

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geochemistry have been further modified by post-magmatic fluid–rock reactions (Haapala, 1977a, 1995). The discovery of subvolcanic and volcanic, in part glassy, equivalents of the topaz-bearing granites, such as the ongonites of Mongolia (Kovalenko *et al.*, 1971), topaz-bearing rhyolites (quartz porphyry) of Finland (Haapala, 1977a) and western USA (Christiansen *et al.*, 1986; Congdon & Nash, 1991), as well as the macusanites of Peru (Barnes *et al.*, 1970; Noble *et al.*, 1984; Pichavant *et al.*, 1988a, 1988b; London *et al.*, 1989), have given strong support to the magmatic model or its modifications.

The purpose of this paper is to present petrographic, mineralogical and geochemical observations from one typical locality, the Eurajoki rapakivi stock in south-western Finland, which can be used to evaluate the role of magmatic and metasomatic processes in the formation of late-stage tin-bearing granites. Similar topaz-bearing granites with tin–polymetallic mineralization have been described from other rapakivi complexes in southeastern Fennoscandia (Haapala & Ojanpera, 1972; Haapala, 1977b; Amelin *et al.*, 1991; Eden, 1991) and from Rondonia and Amazonas in Brazil (Bettencourt & Dall'Agnol, 1987; Horbe *et al.*, 1991).

THE EURAJOKI STOCK

The rapakivi granites of Finland form anorogenic 1.65–1.54 Ga batholiths and stocks, which generally show a bimodal magmatic association and A-type geochemical and mineralogical characteristics. Their emplacement is related to mantle upwelling and extensional tectonics (Haapala & Ramo, 1992; Korja *et al.*, 1993). Several rapakivi complexes contain topaz-bearing late-stage intrusive phases which are often associated with greisen-type Sn–W–Be–Zn mineralization.

The Eurajoki stock is a satellite of the Laitila rapakivi granite batholith. It is composed of several petrographically and geochemically distinct granite types (Fig. 1). The marginal parts are occupied by a horseshoe-shaped body of homogeneous, even- and medium-grained biotite–hornblende(–fayalite) granite called the Tarkki granite. The central parts consist of younger granitic rocks, which Laitakari (1928) called the 'Vakkara granite'. Haapala (1977a) showed that the 'Vakkara granite' itself is an intrusive complex composed of several granite types which may or may not be comagmatic: fine-grained, porphyritic, biotite granite ('contact type'); medium-grained biotite granite; and texturally inhomogeneous topaz-bearing granite. All these granites have sharply cross-cutting contacts with the Tarkki granite. The mutual contact relations between the younger granites are not known, but analogies with other similar rapakivi granite complexes suggest that the topaz-bearing granite represents the youngest intrusive phase. On the

basis of geochemical and mineralogical characteristics (Haapala, 1977a; this study) it is possible to distinguish three granite suites in the Eurajoki stock (Fig. 1): biotite–hornblende granite (Tarkki granite), biotite granites and topaz-bearing granite. These granite suites may represent three periods of emplacement.

The biotite–hornblende granite is cut by rhyolitic porphyry dykes, some of which are topaz bearing, as well as by intermediate dark porphyry dykes which represent hybrids between mafic (diabase) and felsic (rhyolite) magmas (Haapala, 1977a; Salonsaari & Haapala, 1994). Contact relations (Haapala, 1977a, p. 42) and similarities in petrography and geochemistry suggest that the rhyolite dykes were derived from the same magma chamber as the topaz-bearing granite. However, different rhyolite dykes may represent different stages of evolution of the fluorine-rich parent magma. Greisen-type Sn–Be–W–Zn mineralization is found as veins (with or without central quartz veinlet) in the biotite–hornblende granite and as veins, irregular lenses, and patches in the topaz-bearing granite.

The U–Pb zircon age of the biotite–hornblende (Tarkki) granite is 1570 Ma, which is also the age of the main phase of the Laitila batholith (Vaasjoki, 1977). The topaz-bearing granite gives a somewhat younger, discordant, zircon age (Vaasjoki, 1977). The contact relations as well as differences in petrographic characteristics indicate that the topaz-bearing granite crystallized markedly later and at a shallower depth than the biotite–hornblende granite, and the geochemical differences further indicate different parent magmas for these granites (Haapala, 1977a, p. 121).

THE TOPAZ-BEARING GRANITE

Petrography

The texture of the topaz-bearing granite varies markedly. In general it is slightly porphyritic, with 1–3 cm long alkali feldspar and 0.5–1 cm quartz megacrysts occurring in a medium-grained or fine-grained matrix (Fig. 3a, below); locally sub-equigranular fine-grained or coarse-grained varieties are seen. Granophyric texture is locally present. The alkali feldspar is usually perthitic (string, vein and patch perthite), whereas in strongly recrystallized granite it may be nearly homogeneous K-feldspar. Quartz occurs in two generations, as euhedral early crystals and as anhedral late grains. Sometimes the quartz crystals have a euhedral core marked by zonally arranged mineral inclusions, and an anhedral marginal zone. Albitic plagioclase (An_{0–5}) occurs in the matrix, sometimes as small megacrysts. Biotite occurs in the matrix as brown to pale greenish brown, anhedral flakes filling the spaces between quartz and feldspar. Topaz occurs in the matrix usually as subhedral grains, up to 1–2 mm in diameter, and

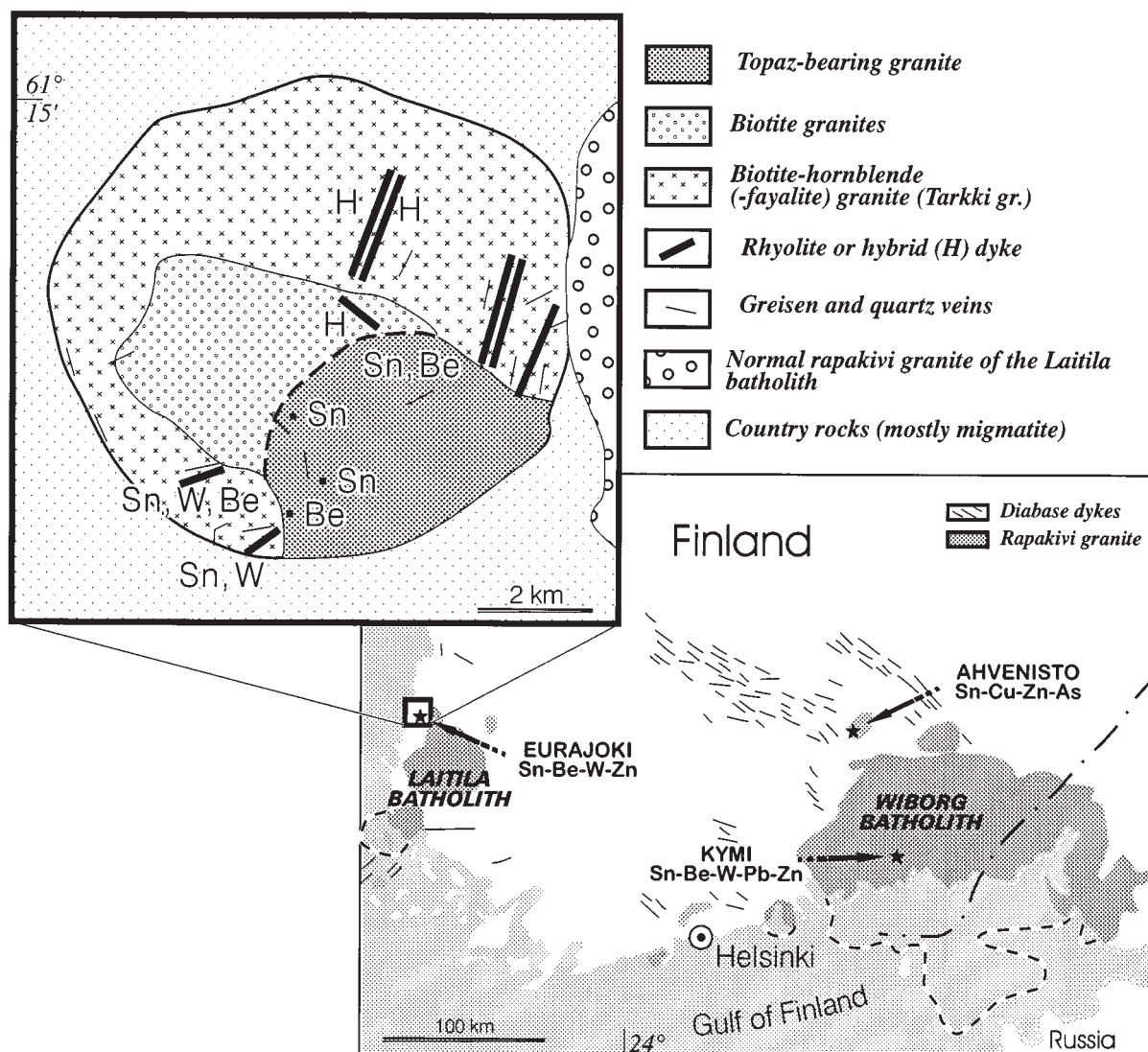


Fig. 1. The Eurajoki rapakivi granite stock, modified from Haapala (1977a). The index map shows the rapakivi granites and coeval diabase dykes cutting the Palaeoproterozoic crust of southern Finland, as well as greisen-type mineralization associated with the topaz-bearing granites.

commonly shows crystal faces against late quartz. Topaz thus crystallized generally after feldspars (especially plagioclase) and early quartz, but before late quartz. In chemical composition the biotite is lithian siderophyllite (protolithionite) (Haapala, 1977a, 1988). The volume percentages of the major and minor minerals are: alkali feldspar 25.1–44.4 (mean 31.9), quartz 24.0–47.2 (25.3), albitic plagioclase 16.6–36.8 (23.9), biotite \pm chlorite \pm muscovite 2.2–8.2 (4.8), topaz 0.0–4.3 (2.1), fluorite 0.0–3.0 (1.4). Typical accessory minerals, identified from the heavy mineral fractions of several granite specimens, are monazite, bastnasite, xenotime, zircon, ilmenite, cassiterite, columbite and thorite (Haapala, 1977a).

Miarolitic cavities are common and range in size from microscopic intergranular vugs (Fig. 3b, below), filled by clay minerals, to pegmatite-lined druses several centimetres in diameter. Locally, an interconnected miarolitic texture (Candela & Blevin, 1995) is visible. The miarolitic cavities indicate the existence of a separate fluid phase obviously separated from the fluid-saturated residual magma in the shallow level (epizonal) magma chamber.

The topaz-bearing granites are characterized by textures indicating subsolidus reactions (exsolution, recrystallization, mineral alterations). Recrystallization phenomena are often visible at the boundaries of feldspar grains; weak recrystallization has produced sutured

Table 1: Chemical analyses of granites and rhyolite dykes of the Eurajoki stock

Analysis:	Bt-hbl	Biotite granite		Topaz-bearing granite					Rhyolite	
	granite	Porph.	Even gr.							
	8/IH/85	309/PL/IH	11/IH/96	5/IH/85	1/IH/85	2/IH/96	98/MK/ER	10/PL/IH	97/IH/67	229B/IH/68
SiO ₂	66.0	72.92	74.5	75.4	73.9	74.0	74.94	74.83	73.1	75.46
TiO ₂	0.739	0.33	0.204	0.027	0.025	0.042	0.02	0.02	0.016	0.02
Al ₂ O ₃	13.1	12.93	11.8	13.1	14.3	13.5	13.77	14.06	14.9	13.77
FeO ^W	5.2		0.8	0.1	0.6	0.4				
Fe ₂ O ₃	1.87	3.03*	0.95	0.37	0.44	0.62	1.07*	1.17*	0.61*	1.21*
MnO	0.08	0.04	0.01	<0.01	0.05	0.01	0.05	0.04	0.03	0.02
MgO	0.43	0.38	0.06	<0.01	0.01	<0.01	0.09	0.06	<0.01	0.12
CaO	2.21	0.74	0.75	0.82	0.70	0.76	0.68	0.64	0.02	0.60
Na ₂ O	2.37	2.64	1.81	3.45	3.31	3.44	3.65	3.67	5.20	4.19
K ₂ O	5.17	5.60	6.84	5.61	4.66	4.97	4.78	4.34	4.54	4.41
P ₂ O ₅	0.26	0.13	0.02	<0.01	<0.01	<0.01	0.01	0.01	0.02	0.01
F ^W	0.18	n.a.	0.40	0.60	1.40	0.96	1.06	0.90	0.33	n.a.
S ^L	0.060	0.01	0.009	0.010	0.08	0.009	0.01	0.0	0.009	n.a.
H ₂ O ^{+W}	1.1	n.a.	0.7	0.4	0.5	0.5	0.4	0.5	0.4	n.a.
—O = 2F	—0.08	n.a.	—0.17	—0.25	—0.59	—0.40	—0.45	—0.38	—0.14	
Total	98.689	98.75	98.683	99.667	99.395	98.83	100.08	99.86	99.045	
Be ¹	4	n.a.	7	4	4	13	n.a.	n.a.	4	n.a.
Li ¹	22	n.a.	18	42	375	153	240	480	235	n.a.
Rb	205	336	365	894	1050	756	1058	963	1050	738
Cs ²	4	3.4	2	3	7	5	7.6	5.5	6	4.6
Sr	178	86	80	17	13	12	28	<5	12	28
Ba	1290	521	356	60	50	56	155	129	104	174
Ga ³	27	n.a.	26	45	55	48	n.a.	n.a.	79	n.a.
Y ³	90	65 ¹	97	130	76	132	71 ¹	63 ¹	4	81
Sc ¹	19	8.4 ²	8	5	8	9	10 ²	8.9	8	9.9 ²
Zr	619	260	328	467	93	113	58	30	76	52
Sn ³	8	n.a.	32	28	93	41	36	100	182	n.a.
Nb	26	11	22	53	57	42	60	61	70	53
Ta ²	2.5	2.7	1	9.8	14	8.4	23	28	49	14.6
W ²	3		<1	15	22	5	n.a.	n.a.	7	6.6
Th ²	17	32	51	24	21	28	28	19	13.0	28.0
U ²	3.8	3.4	8.8	6.4	8.4	7.5	7.6	3.6	7.8	12
Hf ²	14	9.2	11	7.0	6.2	7.1	6.2	5.0	6.8	6.4
Ni ¹	7	n.a.	6	6	3	6	n.a.	n.a.	6	n.a.
Cu ¹	13	n.a.	2.5	8.6	3.3	1.1	n.a.	n.a.	1.4	n.a.
Zn ¹	184	85 ²	138	34.6	184	102	159 ²	175 ²	65.2	65 ²
Cd ³	0.6	n.a.	1.5	1.0	2.9	1.5	n.a.	n.a.	5.1	n.a.
Cr ²	22	n.a.	18	44	24	12	n.a.	n.a.	20	n.a.
V ¹	13	n.a.	2	<2	<2	<2	n.a.	n.a.	<2	n.a.
Mo ¹	15	n.a.	1	<1	<1	4	n.a.	n.a.	2	n.a.
Pb ¹	27	n.a.	68	31	13	65	n.a.	n.a.	27	n.a.
Cl ^W	468	n.a.	255	<100	<100	55	n.a.	n.a.	<100	n.a.

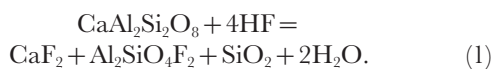
Table 1: continued

Analysis:	Bt-hbl	Biotite granite		Topaz-bearing granite					Rhyolite	
	granite	Porph.	Even gr.							
	8/IH/85	309/PL/IH	11/IH/96	5/IH/85	1/IH/85	2/IH/96	98/MK/ER	10/PL/IH	97/IH/67	229B/IH/68
La ³	109	77.1 ¹	158	49.2	32.6	59.0	39 ¹	27.8 ¹	4.7	41.6 ¹
Ce ³	227	161.2 ¹	303	104	76.1	126	90 ¹	60.0 ¹	18.8	100.0 ¹
Pr ³	27.8	20.3 ¹	34.3	13.9	9.1	15.8	11 ¹	6.6 ¹	1.9	13.0 ¹
Nd ³	111	71.1 ¹	119	52.1	29.0	59.1	36 ¹	20.1 ¹	3.5	38.6 ¹
Sm ³	23.7	13.5 ¹	23.9	18.1	8.5	18.2	9.6 ¹	5.1 ¹	0.8	11.5 ¹
Eu ³	2.64	0.92 ¹	0.64	0.10	0.10	0.23	0.20 ¹	0.05 ¹	<0.05	0.23 ¹
Gd ³	17.8	11.1 ¹	17.8	15.3	6.4	15.3	9.0 ¹	5.0 ¹	0.5	9.6 ¹
Tb ³	2.8	1.7 ¹	2.9	3.6	1.5	3.3	2.0 ¹	1.4 ¹	0.1	2.4 ¹
Dy ³	16.4	9.0 ¹	17.3	27.5	12.2	23.8	15 ¹	11.2 ¹	0.9	18.9 ¹
Ho ³	3.23	1.7 ¹	3.64	6.05	2.83	5.09	3.4 ¹	2.9 ¹	0.18	4.3 ¹
Er ³	9.0	4.5 ¹	10.9	20.8	10.5	17.4	12 ¹	11.2 ¹	0.6	15.4 ¹
Tm ³	1.3	0.60 ¹	1.7	4.2	2.2	3.4	2.1 ¹	2.2 ¹	0.2	3.0 ¹
Yb ³	8.2	3.8 ¹	11.6	34.5	19.2	28.6	17 ¹	18.3 ¹	2.3	23.4 ¹
Lu ³	1.25	0.58 ¹	1.72	5.42	3.11	4.67	2.8 ¹	3.0 ¹	0.35	3.5 ¹
La _N /Yb _N	9.5	14.6	9.8	0.89	1.22	1.48	1.65	1.09	1.47	1.28
Eu _N /Eu* _N	0.39	0.23	0.09	0.02	0.04	0.04	0.07	0.03		0.05

Analyses 98/MK/ER (256/73) and 10/PL/IH(149/70) are from Taylor (1992), analyses 309/PL/IH and 229B/IH/68 are provided by R. P. Taylor. Other analyses are by X-Ray Assay Laboratories Ltd (Canada). Trace elements in ppm. Analytical methods: XRF-F, if not otherwise mentioned; ¹ICP; ²NA; ³ICP-MS; ^wwet chemical; ^LLeco.

*Total iron as Fe³⁺; n.a., not analysed.

boundaries, whereas strong recrystallization has produced smooth, more regular boundaries between feldspar grains. Primary plagioclase is nearly always altered to turbid albite which contains very small inclusions of topaz and fluorite as well as some quartz and sericite (Fig. 3c and d, below). These are formed by reactions between fluorine-bearing fluids and the anorthite molecule of plagioclase:



The time of de-anorthization is not known. It probably started under supersolidus (liquid + crystals) conditions, when HF or F dissolved into the melt reacted with the plagioclase crystals (see Price *et al.*, 1996), and continued under subsolidus conditions.

Other characteristic subsolidus alterations include partial sericitization of plagioclase and topaz, chloritization and muscovitization of biotite, and weak kaolinization of alkali feldspar. Na-metasomatic albitization of K-feldspar and K-feldspathization of plagioclase are encountered only locally (Fig. 5a, below; see also Haapala, 1977a). In greisenized granite, feldspars are partially replaced by quartz, white mica and topaz (Fig. 5, below).

Geochemistry

The topaz-bearing granite deviates in chemical composition markedly from the other granites of the Eurajoki stock and of the whole Laitila batholith (Vorma, 1976; Haapala, 1977a). This peraluminous high-SiO₂ granite is characterized by high F (0.60–1.40%), Li (42–590 ppm), Rb (590–1250 ppm), Ga (45–55 ppm; Table 1), Sn (19–240 ppm) and Nb (40–100 ppm), as well as by low MgO (<0.01–0.06 wt %), TiO₂ (0.02–0.07 wt %), Sr (<5–28 ppm; Table 1), Ba (50–155 ppm; Table 1) and Zr (30–467 ppm). The REE pattern is characterized by relative enrichment in heavy rare earth elements (HREE) (La_N/Yb_N=0.89–1.65; Table 1 and Fig. 2) and a marked negative Eu anomaly (Eu_N/Eu*_N=0.02–0.07), which are typical features for highly evolved granites. The rhyolite dykes show geochemical characteristics similar to those of the topaz-bearing granite; the dyke richest in topaz (specimen 97/IH/67) is even more anomalous than the typical topaz-bearing granite. These are characteristic features for many tin granites, and are similar to those of ongonites (Kovalenko *et al.*, 1971; Stempok, 1991) and topaz rhyolites from the USA (Christiansen *et al.*, 1986; Congdon & Nash, 1991) (Table 1 and Fig. 2c).

Taylor (1992) has divided the topaz-bearing granites into two subtypes, 'low-P subtype' (P₂O₅<0.1 wt %,

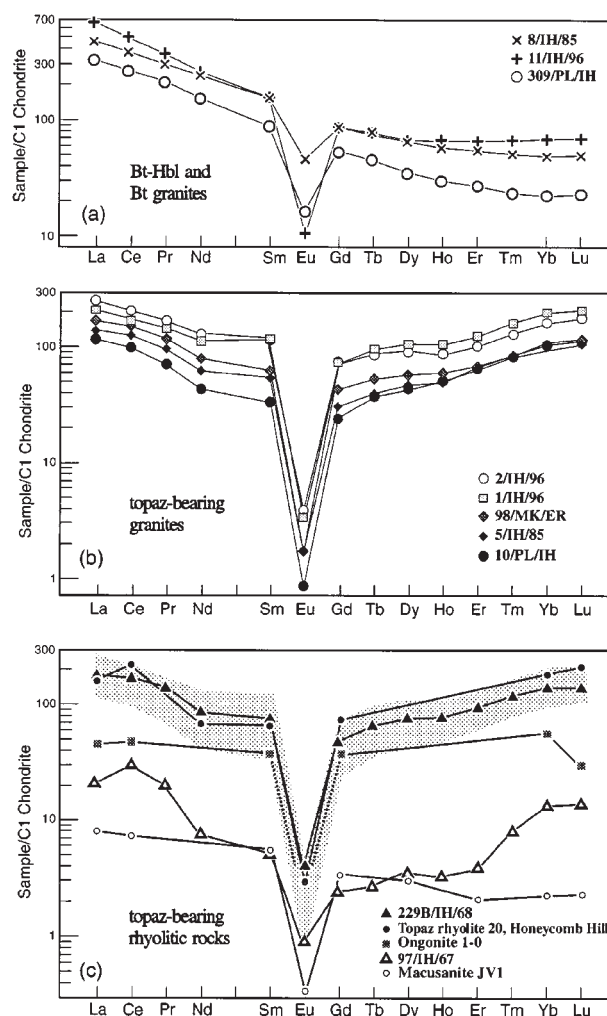


Fig. 2. Chondrite-normalized REE patterns of the granites and rhyolite dykes of the Eurajoki stock. The analyses are presented in Table 1. (a) The biotite–hornblende granite (Tarkki granite, specimen 8/IH/85), fine-grained, porphyritic biotite granite (specimen 309/PL/IH) and medium-grained biotite granite (specimen 11/IH/96); (b) topaz-bearing granites; (c) topaz-bearing rhyolite dyke samples from the Eurajoki stock compared with the topaz-bearing granites (shaded) and some other topaz-bearing rhyolitic rocks.

$\text{Al}_2\text{O}_3 < 14.5$ wt %, $\delta^{18}\text{O} < 10\text{‰}$) and ‘high-P subtype’ ($\text{P}_2\text{O}_5 > 0.4$ wt %, $\text{Al}_2\text{O}_3 > 14.5$ wt %, $\delta^{18}\text{O} > 10\text{‰}$). The low-P granites have a meta-igneous protolith, whereas the high-P granites have a significant pelitic metasedimentary source component. The topaz-bearing granite from Eurajoki with $\delta^{18}\text{O}$ of 9.6‰ and 9.8‰ belongs to the low-P subtype (Taylor, 1992).

Intergranular albite

One characteristic feature of the topaz-bearing late-stage granite is the occurrence of water-clear albite rims between turbid plagioclase and alkali feldspar or between adjacent alkali feldspar grains. Euhedral plagioclase inclusions in alkali feldspar grains usually have continuous

albite rims in crystallographic continuity with the turbid plagioclase. The rims are thinner on (010) faces than on faces perpendicular to (010) (Fig. 3c and d). The turbid plagioclase cores are euhedral, but the water-clear albite rims often have irregular outer margins. The albite twinning usually continues from core to rim, but sometimes the rim has more irregular twinning. In the K-feldspar, the amount of perthite albite is often smaller near the randomly oriented, albite-mantled, plagioclase crystals than farther away. Similarly, turbid plagioclase grains adjacent to alkali feldspar grains have water-clear albite rims against alkali feldspar. Between two differently oriented alkali feldspar grains, the water-clear albite forms swapped albite rims or grain rows, each albite grain having the same optical orientation as perthite albite on

the opposite side of the grain boundary (Fig. 3e and f). Sometimes, the albite rim has as remnants inclusions of perthite albite (different orientation), which indicates that the formation of some perthite albite preceded the formation of rim albite in that zone.

The albite rims could have been formed by metasomatic albitization or exsolution. In both cases, the albite would have nucleated on a plagioclase or alkali feldspar grain, adopting its crystallographic orientation, and grown into the adjacent alkali feldspar. In the case of the Eurajoki granites, the exsolution model is preferred because a large number of electron microprobe analyses with large beam diameter have shown that the total amount of albite component in perthite decreases with coarsening of the perthite texture (albite lamellae and patches) and with simultaneous widening of the albite rims (Haapala, 1977a; see Fig. 6, below). Also, the common decrease of the amount of the perthite lamellae near the albite rims fits the exsolution model: albite has migrated to the grain boundaries to form the intergranular albite. The exsolution is catalysed by postmagmatic fluids, and strong recrystallization has led to coarsening of the exsolved albite grains of the swapped albite rims. These exsolution–recrystallization–ordering processes probably took place at temperatures around or below 500°C (see Brown & Parsons, 1989, 1994). In greisenized and chloritized granite, practically all the albite has exsolved out and migrated to the grain boundaries, where it has recrystallized to larger subhedral or even euhedral albite grains. The grains still have the same optical orientation as occasional perthite albite patches in the K-feldspar grains (last stage in Fig. 6, below).

Topaz

Topaz is a common minor constituent in the peraluminous, late-stage, intrusive phases of the rapakivi granite complexes. In Eurajoki the amount of topaz is usually 1–3%. Topaz has clearly formed in different stages, being in part magmatic, and in part metasomatic. The magmatic topaz mainly occurs in the matrix of the granite, usually as subhedral grains showing crystal faces against late quartz (Fig. 4a and b). Topaz crystals sometimes contain small crystallized melt inclusions (Haapala, 1977a), which during heating experiments give a solidus temperature of ~570°C (R. Thomas, personal communication, 1996). Naumov *et al.* (1971) have described solidified (glass + gas bubbles) melt inclusions in topaz phenocrysts of Mongolian ongonites. These inclusions began to melt usually at 550–600°C, i.e. in the same range as the crystallized melt inclusions in topaz crystals from Eurajoki. In some specimens, topaz occurs with feldspars and biotite as euhedral primary inclusions along former growth zones in marginal parts of quartz megacrysts (Fig. 4c). In some cases, radial microcracks have

developed around such topaz inclusions, because of stronger thermal contraction of quartz during cooling (Haapala, 1977a, p. 37).

Some of the rhyolitic porphyry dykes also contain topaz. It occurs in part as phenocrysts (Fig. 4d), but more commonly it occurs as small prisms in the matrix, sometimes showing preferred orientation (flow texture) (Fig. 4f). The magmatic origin of such very small ($\leq 5 \mu\text{m}$) topaz crystals in the matrix and some larger ragged topaz insets is not undisputable, because they may be formed also as a result of devitrification and vapour-phase alteration. However, it is extremely difficult to challenge the magmatic origin of topaz occurring as euhedral inclusions in quartz phenocrysts (Fig. 4e).

Secondary topaz occurs, together with fluorite, quartz and sericite, as very small inclusions in plagioclase throughout the topaz-bearing granite. These small inclusions give a turbid appearance to the plagioclase. They are obviously formed by reactions between plagioclase and fluorine-bearing late magmatic to postmagmatic fluids [see reaction (1)]. In greisenized granite, topaz often replaces plagioclase and more rarely K-feldspar as irregular grains, and sometimes topaz and quartz appear to have replaced the albite veins or strings in perthite leaving the K-feldspar undisturbed. Topaz grains are often altered along their margins and fractures to sericite and sometimes to clay minerals or chlorite.

Cassiterite

Cassiterite occurs in the Eurajoki stock in three different associations: (1) in the topaz-bearing granite as a ubiquitous accessory mineral; (2) in pegmatite dykes and pockets; (3) in greisen and quartz veins. The cassiterite of different associations shows differences in crystal form, colour and composition. In the topaz-bearing granite, the accessory cassiterite occurs in the matrix as small elongated prisms which are black or dark brown under stereomicroscope and homogeneous dark brown in thin section. The heavy mineral fractions of the topaz-bearing granite contain also a little columbite, and, rarely, niobian rutile (Haapala, 1977a; Taylor, 1992). In pegmatite, and especially in greisen, cassiterite is lighter in colour and exhibits well-developed colour zoning. Niobian rutile is also found as inclusions in cassiterite of greisen veins in the biotite–hornblende granite.

Clear differences occur in the chemical composition of cassiterite of different associations. Electron microprobe analyses show that the contents of $\text{Nb}_2\text{O}_5 + \text{Ta}_2\text{O}_5$ and FeO in cassiterite decrease from granite through pegmatite to greisen, obviously in order of decreasing temperature of crystallization (Table 2; Fig. 7). Because columbite and the rare niobian rutile occur together with cassiterite in the topaz-bearing granite, it is obvious that

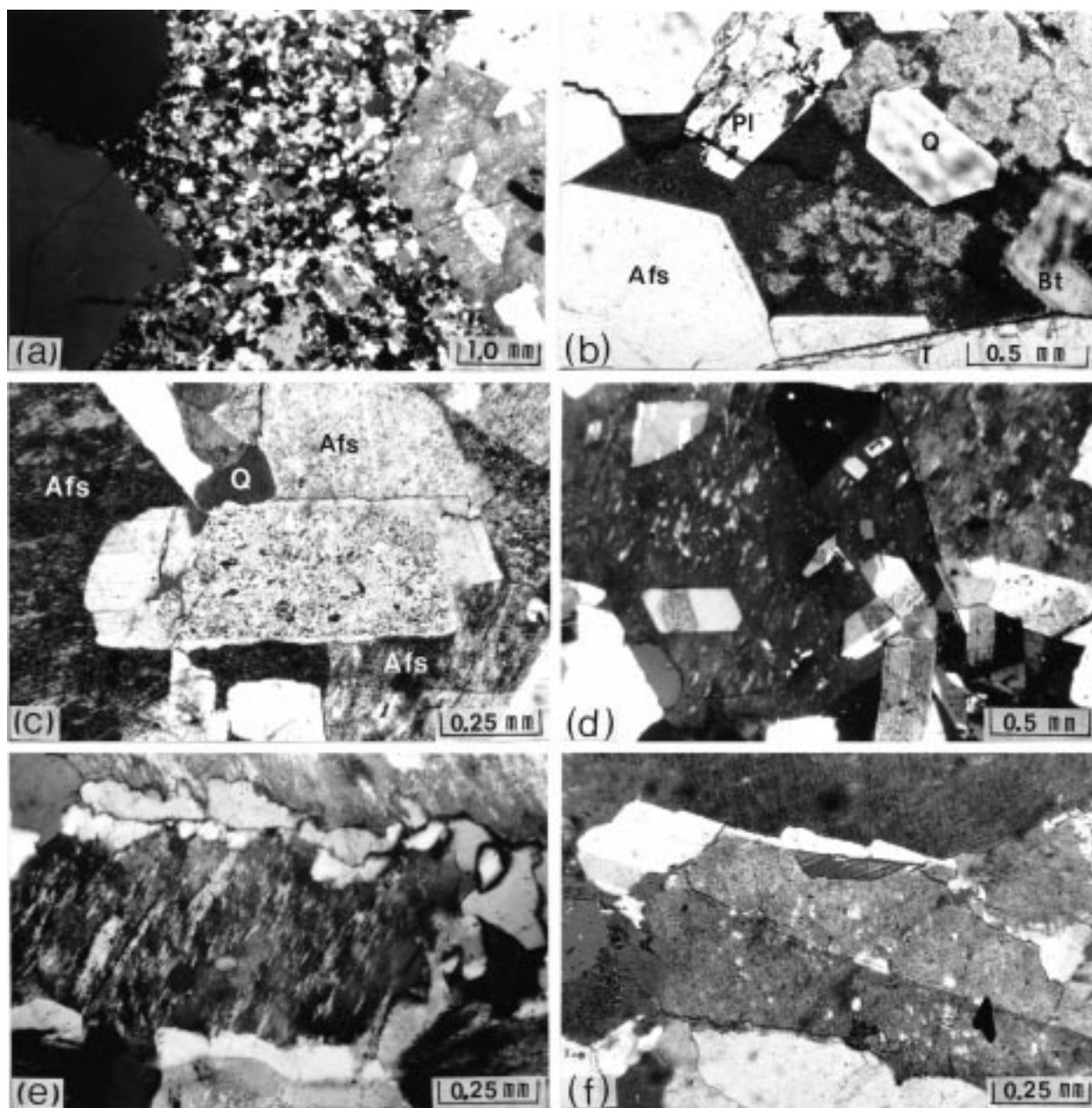


Fig. 3. Photomicrographs, taken with partly crossed Nicols, of the topaz-bearing granite of the Eurajoki stock. (a) Porphyritic topaz-bearing granite with alkali feldspar and quartz megacrysts. Sample 98/MK/67. (b) Miarolitic cavity filled with clay minerals and bordered by alkali feldspar (Afs) albite (Pl), quartz (Q), topaz (T) and altered biotite (Bt). Sample 25/MK/67. (c) Euhedral turbid albite plagioclase surrounded by alkali feldspar grains. Water-clear albite has grown on the turbid plagioclase against alkali feldspar. An anhedral quartz grain follows the former boundary of the euhedral plagioclase crystal. Sample 37/MK/67. (d) Water-clear albite grown on euhedral turbid albite plagioclase inclusions in alkali feldspar. Perthite texture is weakly developed near the plagioclase inclusions. Sample 301/PL/68. (e) Swapped albite rims between perthitic alkali feldspar grains. The alkali feldspar grain in the middle of the photo is surrounded on three sides by other alkali feldspar grains, with swapped albite rims at the boundaries. The intergranular albite grains have the same optical orientation as the perthite albite on the opposite side of the rim. Sample 16/PL/68. (f) Swapped albite rims between adjacent alkali feldspar grains; sample 109/MK/67.

the system was, at least temporarily, (Nb, Ta) and Ti saturated. This systematic change in composition of cassiterite shows that cassiterite did not crystallize in one

stage (e.g. greisenization) only, but was deposited in different generations. It is obvious that the accessory cassiterite of the granite is a magmatic mineral which

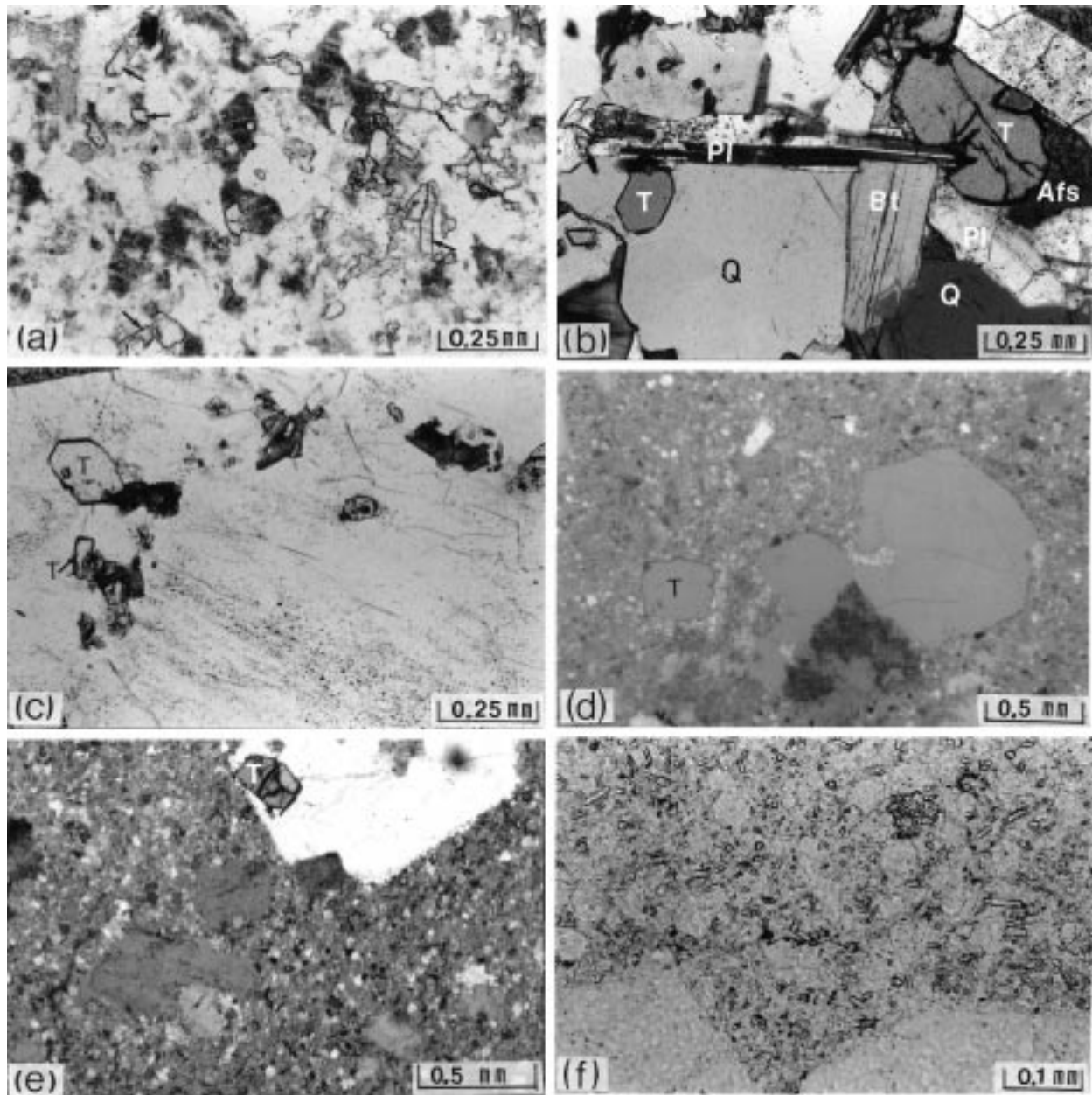


Fig. 4. Photomicrographs showing the occurrence of topaz in the topaz-bearing granite (a–c) and in a rhyolite dyke (sample 97/IH/67; d–f). (a) Groundmass of a porphyritic granite showing subhedral topaz crystals (some marked with an arrow) together with quartz, alkali feldspar (turbid), albite plagioclase and biotite. The photo is from the centre of Fig. 3a (sample 98/MK/67). (b) A detail of the groundmass of a porphyritic granite (sample 1/IH/85). Plagioclase (Pl) is euhedral against topaz (T) and biotite (Bt), but topaz shows crystal faces against alkali feldspar (Afs), late quartz (Q) and biotite. (c) Zonally arranged mineral inclusions (biotite, topaz, plagioclase) in marginal parts of a quartz megacrysts. Sample 1/IH/85. (d) Rhyolite (sample 97/IH/67) with phenocrysts of quartz, alkali feldspar and topaz. (e) Quartz and alkali feldspar phenocrysts in rhyolite (same sample as in d and e). A euhedral topaz inclusion in the quartz phenocryst. (f) Groundmass of rhyolite, showing small prismatic topaz crystals. Two quartz phenocrysts in the lower part of the photo.

Table 2: Mean contents (wt %) of minor constituents in cassiterite from different types of occurrence in the Eurajoki stock

	Cassiterite from granite					Cassiterite from pegmatite		Cassiterite from greisen				
Nb ₂ O ₅	3.9	4.9	3.1	3.0	2.8	2.7	1.6	0.0	0.2	0.2	0.5	0.6
Ta ₂ O ₅	2.5	2.9	2.7	2.4	2.0	0.3	0.7	0.0	0.1	0.0	0.0	0.0
FeO	1.7	2.4	1.6	1.8	1.4	1.4	0.9	0.3	0.1	0.1	0.1	0.2
TiO ₂	0.4	0.3	0.1	0.3	0.2	0.6	0.5	0.8	0.6	0.5	0.2	0.1
SnO ₂	91.5	89.5	98.5	92.5	93.6	95.0	96.3	98.9	99.0	99.2	99.2	99.1

The results are based on means of microprobe analyses of cassiterite fractions of five specimens of topaz-bearing granite, two specimens of pegmatite, and five specimens of greisen. SnO₂ is calculated to make the sum of oxides 100. Summarized from Haapala (1977a).

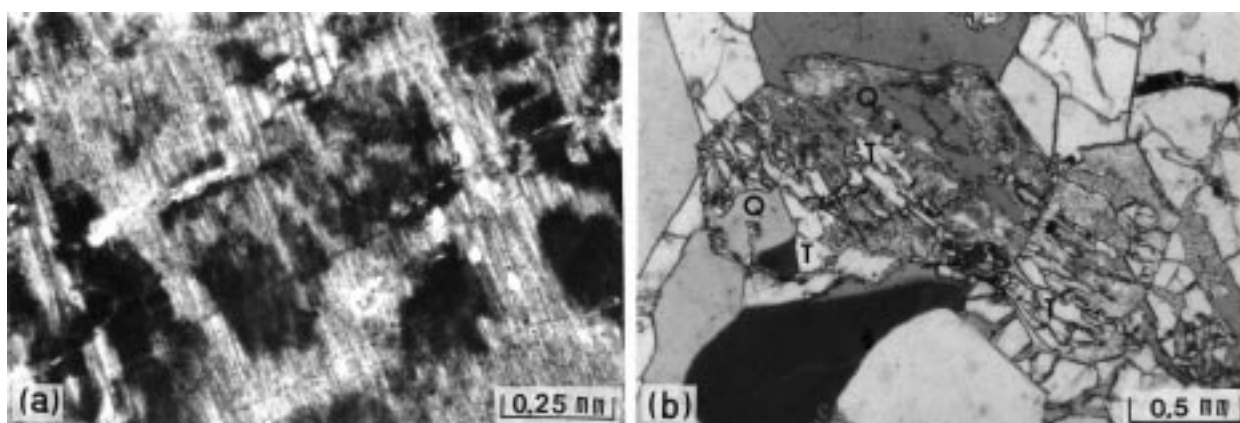


Fig. 5. Photomicrographs showing Na-metasomatic albitization (a) and topazization (b) in granites of the Eurajoki stock. (a) Uniformly oriented remnants of a K-feldspar grain (dark grey) in discontinuously twinned albite (chessboard albite). The replacement has advanced along grain margins and fractures. Even-grained biotite granite, sample 207/IH/67. (b) Replacement of K-feldspar crystals by vein-like strings and lenses of topaz (T) and quartz (Q). Greisenized granite, sample 304/PL/68.

crystallized from the residual interstitial melt (or melt–fluid system) before complete solidification of the magma (Haapala, 1977a).

DISCUSSION

Albitization of granites

Albitization is often regarded as one of the most typical petrographic features of the mineralized granites (e.g. Beus *et al.*, 1962; Stempok, 1971). However, the exact meaning of the word ‘albitization’ is usually not given, but from the context it often becomes apparent that the word is used in the sense of Na-metasomatic addition of albite (e.g. by albitization of K-feldspar, or by albitization of plagioclase which involves the coupled exchange reaction Na^+ for Ca^{2+} and Si^{4+} for Al^{3+}). Albite in granitic rocks may be formed by several processes: (1) magmatic

crystallization; (2) exsolution of alkali feldspar; (3) de-anorthitization of plagioclase; (4) Na-metasomatic albitization. Although magmatic albite is very rare, experimental studies indicate that it may crystallize from highly evolved magmas (e.g. Weidner & Martin, 1987; London *et al.*, 1989). Although albitization formed by processes (2)–(4) are postmagmatic, only case (4) involves inevitably sodium metasomatism. It is probable that formation of intergranular albite rims and grains by exsolution and minor recrystallization (the growing of differently oriented albite into the host alkali feldspar requires reorientation of the Si–Al–O framework and hence recrystallization) is more common than generally thought. Obviously, much of the albite described in the literature as a product of albitization is actually intergranular albite formed by exsolution and minor recrystallization. De-anorthitization of plagioclase involves destruction of the anorthite molecule of plagioclase, not necessarily sodium

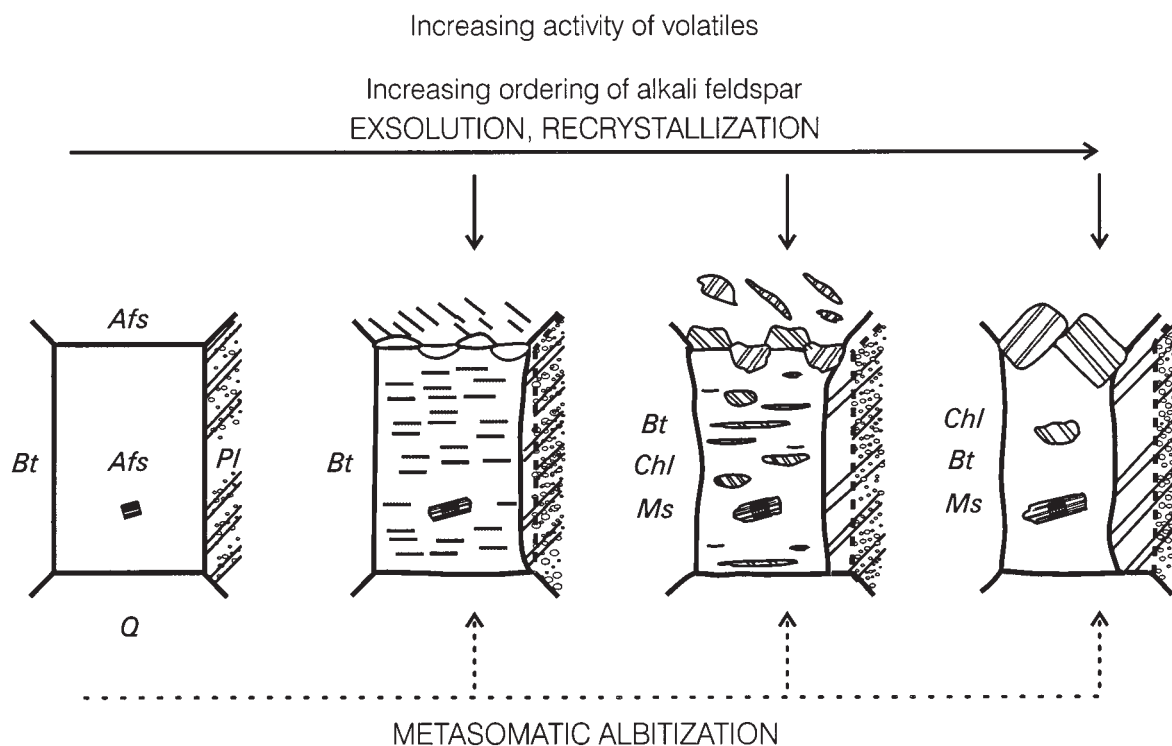


Fig. 6. A schematic model of the progressive evolution of the perthite textures and the intergranular albite in the topaz-bearing granite of the Eurajoki stock. The picture shows an alkali feldspar grain (Afs), surrounded by another alkali feldspar, plagioclase (Pl), quartz (Q) and biotite (Bt). Volatile-catalysed exsolution, accompanied by partial recrystallization, is the major mechanism in producing an aggregate of gradually coarsening perthite from an originally homogeneous alkali feldspar grain (first film and string perthite, then vein and patch perthite, and finally nearly pure K-feldspar microcline with accidental albite patches), water-clear albite rims on turbid plagioclase (adjacent grain and a small primary inclusion in alkali feldspar) against alkali feldspar, and swapped albite rims or grain rows between adjacent alkali feldspar grains. The postmagmatic fluids may cause alteration of biotite to chlorite (Chl) and muscovite (Ms), de-anorthitization of plagioclase, and other alteration reactions. Simplified from Haapala (1977a).

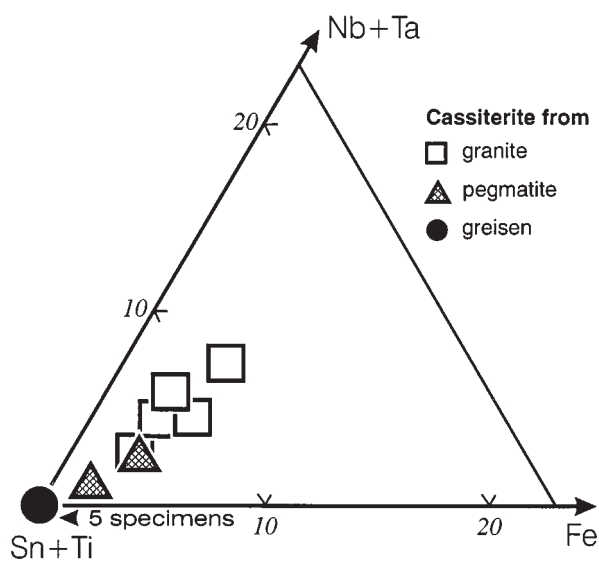


Fig. 7. Composition of cassiterite in terms of Nb+Ta, Sn+Ti and Fe, calculated from Table 2. The diagram separates cassiterite from the topaz-bearing granite, pegmatite and greisen.

metasomatism; the cations are used to form small mineral inclusions in the plagioclase. Similar de-anorthitization of plagioclase takes place in greenschist facies metamorphism and propylitic alteration.

In the case of the Eurajoki stock, secondary albite has been formed in the topaz-bearing granite by all the processes (2)–(4), but exsolution and de-anorthitization have been the major mechanisms. Locally, however, metasomatic albitization has affected the granites of the Eurajoki stock. Such albitization has often advanced along fissures and cracks, and changed original alkali feldspar grains partly or totally to chessboard albite (Fig. 5a).

Origin of topaz

Various petrographic features, described in detail by Haapala (1977a), indicate that the late-stage granite of the Eurajoki stock contains both magmatic and metasomatic topaz. Similar observations have been made from several other areas. In the topaz rhyolites of southwestern USA, topaz generally occurs in vugs and has precipitated from a vapour phase (Christiansen *et al.*, 1986), but some rhyolitic lavas contain magmatic topaz as megacrysts and phenocrysts, and in the groundmass as abundant needles (Congdon & Nash, 1991). Also, the ongonites of Mongolia contain topaz as 1 mm phenocrysts and in the groundmass as small acicular crystals which often form fluidal textures (Kovalenko *et al.*, 1971; Stempok, 1991). The topaz-bearing rhyolite (quartz porphyry) of Eurajoki is, in terms of its petrography and geochemistry, very similar to these topaz-bearing volcanic–subvolcanic rocks. They all contain phenocrysts of quartz, alkali feldspar, albite, topaz and mica in a glassy or fine-grained groundmass.

The occurrence of magmatic topaz in Eurajoki in two or more generations (euhedral topaz inclusions within quartz phenocrysts in rhyolite and quartz megacrysts in granite; topaz phenocrysts in rhyolite; subhedral topaz grains in the matrix of granite and rhyolite) is probably related to crystallization at different pressures. The experimental studies of Weidner & Martin (1987) on the topaz-bearing leucogranite (1.15 wt % F) from the St Austell pluton in Cornwall, UK, suggest that in H₂O-saturated magmas the relative order of crystallization of topaz, feldspar, quartz and muscovite depends on the pressure. Increasing pressure causes stabilization of topaz at higher temperatures. Thus, at 4 kbar pressure topaz is on the liquidus, but at 1 kbar topaz crystallizes after plagioclase, quartz and alkali feldspar. This may explain the occurrence of euhedral topaz inclusions in quartz phenocrysts in the rhyolite and in quartz megacrysts in the granite: these minerals have crystallized deep in the crust, where topaz was one of the early stabilized minerals. At higher levels, topaz usually crystallized after feldspars and early quartz in the matrix.

The common occurrence of topaz as both magmatic and metasomatic mineral is in agreement with experimental studies which show that fluorine partitions in substantial amounts between both granitic melt and coexisting aqueous fluid phase (see London, 1987; Webster & Holloway, 1990; Keppler & Wyllie, 1991). Experimental studies also indicate that topaz (and albite) may crystallize in supersolidus conditions from fluorine-enriched granitic magmas (Anfilogov *et al.*, 1973; Webster *et al.*, 1987; Weidner & Martin, 1987).

Solubility of cassiterite in granitic melts

Experimental studies (see Stempok, 1990; Taylor & Wall, 1992; Linnen *et al.*, 1995, 1996) show that the solubility of cassiterite in granitic melts varies markedly, decreasing with decreasing temperature and increasing $f(\text{O}_2)$, as well as with decreasing (Na + K)/Al and Na/K of the melt. In their experiments, Taylor & Wall (1992) obtained solubilities of 400–2500 ppm Sn in peraluminous granitic melt, but these results may be affected by reactions between tin and the gold capsule during the runs (Linnen *et al.*, 1995). Utilizing the diffusion profile method, Linnen *et al.* (1996) showed experimentally that at 850°C and oxygen fugacities around 2 log₁₀ units or more above the FMQ (fayalite–magnetite–quartz) buffer, the solubility of cassiterite in subaluminous or weakly peraluminous [Al/(Na + K) 1.02–1.22] water-saturated haplogranitic melts is ~400 ppm SnO₂. Because the solidus temperature of highly evolved fluorine-rich granites is around 600°C or even lower (Manning, 1981; Pichavant *et al.*, 1988a; London *et al.*, 1989), the saturation concentration of cassiterite in such granites may be well below 400 ppm Sn at the above-mentioned $f(\text{O}_2)$ conditions. Under more reducing conditions, the solubility of cassiterite increases rapidly.

Available experimental and geochemical studies indicate that, under relatively reducing conditions, tin (as Sn²⁺) is a highly incompatible element in granitic systems and fractionates in the residual melt. Thus, the tin content is lower in feldspar and quartz-rich granitic ‘cumulate’ than in the melt from which it crystallized (Linnen *et al.*, 1992). In the case of the Eurajoki stock, apparently the best estimates of parent melt of topaz-bearing rocks can be obtained from the aphanitic chilled margins of the topaz-bearing rhyolite (quartz porphyry) dykes. Analyses of aphanitic contacts of three dykes (samples 97B/IH/67, 228A/IH/68 and 229A/IH/68) gave values of 100, 82 and 125 ppm Sn, respectively (Haapala, 1977a, p. 106). Thus, 100 ppm Sn can be used to approximate the tin content of the parent rhyolite melt, and obviously it may be used as a rough approximation of the tin content in the topaz-bearing granite melt. The average tin content of the topaz granite is 80 ppm.

Assuming a bulk solid–melt partition coefficient K^{Sn} of 0.1–0.3 (see Linnen *et al.*, 1992), 90% fractionation of a melt with an original tin content of 100 ppm would produce, according to the Rayleigh fractionation equation, a residual melt with between 800 and 500 ppm Sn, respectively. These values are of the same magnitude as (actually somewhat higher than) the saturation concentration of cassiterite in low-temperature granitic melts under oxidizing conditions (Linnen *et al.*, 1996). It is interesting to note that the chemical composition of biotites suggests that in the rapakivi complexes $f(\text{O}_2)$ varies from about FMQ buffer values in some of the early (biotite + hornblende \pm fayalite and magnetite) granites to values between NNO (nickel–nickel oxide) and HM (haematite–magnetite) buffers in the topaz-bearing late-stage granites of Eurajoki and Kymi (Rieder *et al.*, 1996). Under reducing conditions during the main stage of crystallization, tin (largely as Sn^{2+}) has fractionated into the residual melt, but increased $f(\text{O}_2)$ in the latest stages of magmatic evolution (tin largely as Sn^{4+}) has effectively decreased the saturation concentration of cassiterite and caused its crystallization as an accessory mineral. Cassiterite is occasionally found in the granite matrix as euhedral crystals embedded in biotite, which indicates that cassiterite crystallized earlier than, or simultaneously with biotite.

The distribution of tin between granitic melt and coexisting aqueous fluid is poorly known. Available experimental studies indicate that tin favours the melt phase, but that its partition into the coexisting fluid increases with increasing HCl (and less with HF) concentrations (Keppler & Wyllie, 1991).

Magmatic cassiterite has recently been described from the highly evolved ‘high-P subtype’ topaz–lepidolite–albite granite of Beauvoir, French Massif Central (Cuney *et al.*, 1992), and from the Nong Sua aplite–pegmatite complex in Thailand (Linnen *et al.*, 1992). In the Younger Granites of the Erzgebirge, granites with tin contents >30 ppm have cassiterite as an accessory mineral (Satran & Klominsky, 1968). Taking into account the fact that the tin content of granite is obviously much less than the tin content of the residual highly fractionated melt, it is possible that the accessory cassiterite of the unaltered Erzgebirge granite is also of late magmatic origin.

CONCLUSIONS

Proterozoic anorogenic granite complexes often contain leucocratic topaz-bearing granites as the latest intrusive phases. These granites, of which the topaz-bearing granite of the Eurajoki stock is a good example, commonly show the petrographic and geochemical characteristics of tin granites, and in many cases (southeastern Fennoscandia,

Amazonas and Rondonia in Brazil) Sn(–W–Be–Zn–Cu) mineralization is associated with them.

On the basis of petrographic, mineralogical and geochemical observations presented here and by Haapala (1977a), the following conclusions can be reached with respect to the origin of the petrographic and mineralogical characteristics of the topaz-bearing granites and rhyolite (ongonite) dykes of the Eurajoki rapakivi granite stock:

(1) The topaz-bearing granite and related rhyolite dykes of the Eurajoki stock were emplaced after crystallization, cooling and fracturing of the biotite–hornblende–fayalite granite (Tarkki granite). Textural and geochemical characteristics suggest that the two granite suites crystallized under different P – T conditions and from different parent magmas. The biotite granites are temporally and compositionally intermediate between the two suites.

(2) The presence of miarolitic cavities in the more or less porphyritic topaz-bearing granite indicates separation of an aqueous fluid phase during late stages of crystallization of the shallow level (epizonal) magma chamber.

(3) The topaz-bearing granites and associated topaz-bearing rhyolite dykes show geochemical peculiarities typical for tin-bearing granites of different ages: high F, Li, Rb, Ga, Sn and Nb, and low Mg, Ti, Ba, Sr, Zn and Eu. Similarities with subvolcanic–volcanic equivalents, ongonites and topaz rhyolites, are obvious.

(4) The topaz-bearing granite is characterized by subsolidus reactions, including exsolution of alkali feldspar, recrystallization and mineral alterations. Exsolution of alkali feldspar, accompanied by minor recrystallization, has produced albite rims between primary, turbid plagioclase and alkali feldspar grains, as well as swapped albite rims or grain rows between adjacent alkali feldspar grains. These exsolution–recrystallization reactions are accompanied by ordering of alkali feldspar and coarsening of perthite texture, and have been promoted by intergranular fluids. Interaction between fluorine-bearing fluids and primary plagioclase has caused de-anorthitization of plagioclase, visible as development of very small fluorite, topaz and quartz inclusions in the plagioclase. Other common mineral alterations include chloritization and muscovitization of biotite. Albitization of K-feldspar and K-feldspathization of plagioclase are local metasomatic replacement reactions, as is greisenization of the granite. It is possible that ion exchange reactions have caused some petrographically indistinguishable chemical changes (e.g. increase in Rb) in feldspars and micas.

(5) Topaz has crystallized during different stages of magmatic evolution, probably resulting from changes in pressure, but it is mainly late magmatic (subhedral grains in the matrix of the granite). In part, topaz is metasomatic, replacing feldspars, especially plagioclase.

(6) Cassiterite is a characteristic accessory mineral in the topaz-bearing granite. Its mode of occurrence and

chemical composition (high Nb + Ta) indicate a primary, late magmatic origin, which is in agreement with available experimental studies.

(7) The occurrence of topaz and cassiterite as primary, magmatic minerals indicates that the late-stage granite magmas of the Eurajoki complex were strongly enriched in fluorine and tin. The topaz-bearing granite was originally geochemically anomalous, and this 'specialization' was increased by subsolidus fluid-rock reactions. It is natural that the crystallization of highly evolved, volatile-saturated melt is accompanied by such postmagmatic reactions.

The observations and interpretations presented above for the topaz-bearing granite from Eurajoki are valid for several other late-stage intrusive phases of the Proterozoic rapakivi granite complexes as well for many tin granites in general. The role of Na-metasomatic albitization has often been overemphasized owing to misinterpretation of the textures. First, the intergranular albite rims and coarse perthite texture are in many cases formed by exsolution and minor recrystallization rather than albitization of K-feldspar. Second, de-anorthitization of plagioclase should not be confused with Na-metasomatic albitization; it is a question of destruction of the anorthite molecule of plagioclase, not necessarily Na-metasomatism, although Na-metasomatism may be accompanied by de-anorthitization. De-anorthitization of plagioclase can be identified in thin section from the occurrence of numerous very small mineral inclusions in albitic plagioclase. These inclusions (fluorite, topaz, quartz, sericite, calcite, epidote, etc.) contain cations of the anorthite molecule, and are formed by plagioclase-fluid interaction. Greenschist facies metamorphism and propylitic alteration of granitoids are well-known examples involving such de-anorthitization. However, true Na-metasomatic albitization of K-feldspar is common in many late-stage granites and pegmatites, and cassiterite often occurs in such albitized rocks (e.g. Haapala, 1966). This albitization is often controlled by fractures, and has locally led to complete albitization of K-feldspar. In granite such albite commonly exhibits chessboard texture; in pegmatites it may be fine-grained sugar albite or platy cleavelandite.

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