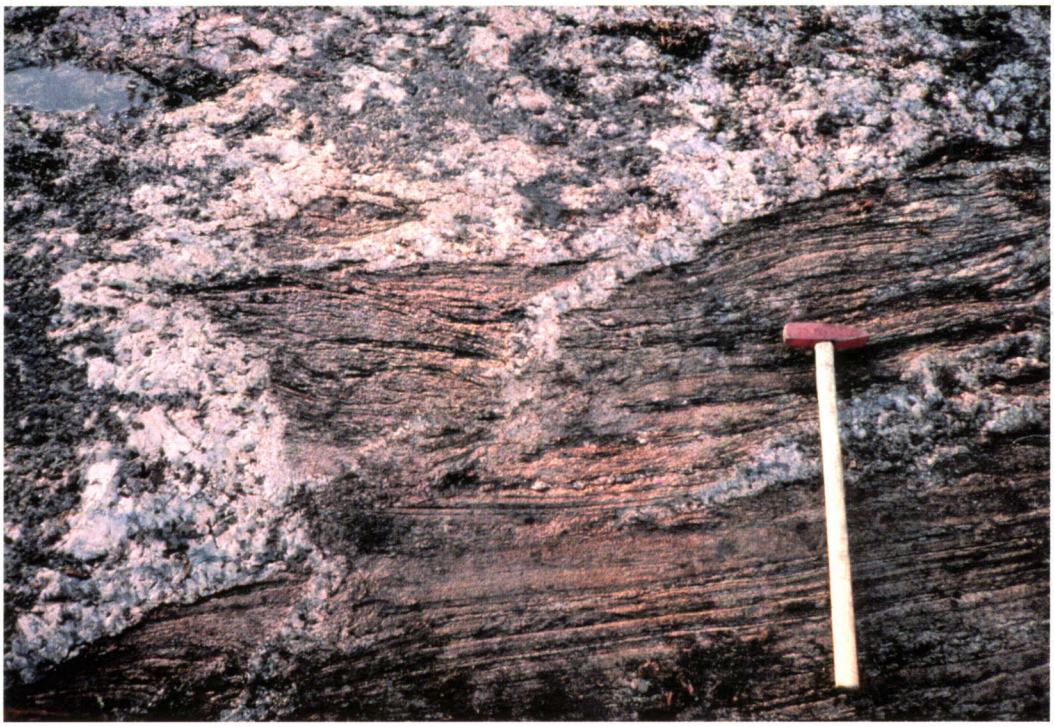
Geological Survey of Finland, Special Paper 30

Svecofennian granitic pegmatites (1.86-1.79 Ga) and quartz monzonite (1.87 Ga), and their metamorphic environment in the Seinäjoki region, western Finland

edited by Hannu Mäkitie



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Geological Survey of Finland
Espoo 2001

Front cover: Svecofennian mica schist cut by a pegmatitic granite resembling the c. 1.80 Ga rare-element pegmatites. Kyrkösvuori, Seinäjoki. Photo: Hannu Mäkitie. Farmland – a common view at Seinäjoki, western Finland. Photo: Erkki Halme. (background)

Back cover: Svecofennian mica gneiss and foliated tonalite cut by the 1.87 Ga Luopa quartz monzonite (on the left). Perämaa, Ilmajoki. Photo: Hannu Mäkitie.

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This volume describes Palaeoproterozoic Svecofennian granitic rocks which postdate the main stage (1.89-1.88 Ga) of crustal thickening of the Svecofennian orogeny (1.9-1.8 Ga) in the Seinäjoki region, western Finland. The studied rocks are rare-element (RE) pegmatite dykes (1.86-1.79 Ga), associated pegmatitic granites, and an older (1.87 Ga) quartz monzonite intrusion. They represent the youngest igneous rocks of the region. The petrography, chemical and mineralogical composition, structure, metamorphic environment and U-Pb ages of the rocks are described. The Seinäjoki region is located in the Bothnia Belt, with a southern part extending into the Central Finland Granitoid Complex.

About 250 RE-pegmatites were studied and most of these can be classed as either beryl pegmatites, complex pegmatites or albite-spodumene pegmatites, which are all of the LCT(lithium-caesium-tantalum) type. Beryl-columbite-cassiterite pegmatites (ca. 1.80 Ga, U-Pb tapiolite age) are most common near the town of Seinäjoki. Locally there are a few topaz-bearing pegmatites (ca. 1.80 Ga, U-Pb columbite age) that contain accessory columbite and fluorite, and form a unique subtype among the beryl pegmatites. Of the pegmatites studied, the complex pegmatites of Haapaluoma and Kaatiala (ca. 1.80 Ga, U-Pb columbite ages) are the richest in rare minerals such as columbite, cassiterite, spodumene and lepidolite. Albite-spodumene pegmatites (ca. 1.79 Ga, U-Pb columbite age) occur in the Kruunupyy-Ullava area, which lies 40 km NE of Seinäjoki. The columbites dated differ in chemical composition from ferrocolumbite to manganocolumbite, and there is minor variation in the contents of Ti and W. Older rare-earth-element (REE)-pegmatites (ca. 1.86 Ga, U-Pb monazite age) occur in the Alavus area, the southeastern part of the Seinäjoki region that is part of the Central Finland Granitoid Complex. Except for these REE pegmatites, all the other pegmatites are essentially in the rocks of the Bothnia Belt. There are about 12 RE-pegmatite groups in the study region. These are usually associated with pegmatite granites that are 10-100 m thick and 50-300 m long.

There is a clear relationship between the regional distribution of the granitic pegmatites and the metamorphic grade of the country rocks. Most of the RE-pegmatites occur in low-pressure areas containing low to medium amphibolite grade rocks, such as andalusite mica schists or sillimanite-muscovite-biotite gneisses. The characteristic rare minerals in the pegmatites disappear indicating that the metamorphic grade of the country rocks exceeded those P-T conditions that occurred between the first and second sillimanite isograd. The whole rock chemical composition of the pegmatites also changes relative to the metamorphic grade of the country rocks: the highest Li, Ta, Cs, Rb/Sr contents usually occur in those RE-pegmatites hosted by the andalusite mica schists while, in contrast, high Fe, Th and K/Rb characterise those RE-pegmatites that lie in rocks metamorphosed in the muscovite-sillimanite facies or at a higher metamorphic grade. Local minor variations in chemical composition are the result of differences in the degree of fractionation within the pegmatites. Additionally, the country rocks of the LCT-type pegmatites are enriched in Li (up to 500 ppm), Cs (up to 300 ppm) and Rb (up to 300 ppm). The topaz-bearing pegmatites are exceptional because they lie in high-grade metamorphic mica gneisses near the diatexitic Vaasa Migmatite Complex (1.89-1.88 Ga), which is part of the Bothnia Belt.

The LCT-type pegmatites are clearly younger than the peak of regional metamorphism (1.89-1.88 Ga). The studied pegmatites also cut the main schistosity (S_2) seen as a growth of metamorphic muscovite and biotite on the F_2 axial plane. Many of these pegmatites are, however, deformed by younger folding. Due to the

absence of granite plutons near most of the LCT-type pegmatites, it is suggested that the pegmatites represent either (1) a distinct tectonomagmatic episode that is notably younger than the culmination of the Svecofennian orogeny, or (2) the upward movement of low-solidus anatectic melts that initially formed in the neighbouring high-grade metamorphic migmatite terranes and migrated into the relatively low-grade metamorphic areas.

The porphyritic Luopa quartz monzonite (1.87 Ga, U-Pb zircon age) is located at the boundary of the Central Finland Granitoid Complex and the Bothnia Belt. The intrusion is hosted by migmatitic garnet-cordierite-sillimanite mica gneiss, pyroxene granite and foliated tonalite. The Luopa quartz monzonite is composed of three different rock varieties: a dark green fayalite-augite-hornblende quartz monzonite, a dark green augite-hornblende-biotite quartz monzonite and a light-coloured hornblende-biotite quartz monzonite. The REE distribution spectrum of the light-coloured variety is steeper than in the dark-green varieties: the aforementioned variety is more LREE enriched and HREE depleted. The chemical composition of the intrusion is characterized by low Rb/Sr (0.15-0.25), high X_{Fe} (0.9) and low A/CNK (0.91-0.96). The intrusion has the following specific features: an irregular form, a relatively high fayalite content and an intrusive relationship with the surrounding pyroxene granite. The quartz monzonite was emplaced at a relatively high crustal level and it cuts the main regional deformation phase (D_2).

The emplacement of the Luopa intrusion resulted in the formation of a contact metamorphic aureole, which overprints the slightly older upper amphibolite to lower granulite facies regional metamorphism. The contact aureole can be divided into a narrow granulite-grade inner aureole (2 m wide) and a wider outer aureole (30 m wide) of slightly lower metamorphic grade. Garnet-orthopyroxene and cordierite-orthopyroxene pairs, as well as symplectites containing hercynite, sillimanite and cordierite, occur in the inner aureole. Texturally, the metapelites in the contact aureole are either diatexites or strongly deformed, tightly banded mica gneisses (due to ballooning of the intrusion). The contact metamorphic temperature is estimated to have been about 750 °C.

One notable feature of the Seinäjoki bedrock is that there are two different high-grade metamorphic terranes. The southern one is composed of migmatitic garnet-cordierite-sillimanite mica gneisses, which contain minor kyanite. Foliated I-type tonalites and the Luopa quartz monzonite are located in this terrane. The other high-grade terrane is situated in the northwestern part of the study region and contains migmatitic garnet-cordierite mica gneisses (with minor andalusite) and granodioritic diatexites belonging to the Vaasa Migmatite Complex. There is a gradual transition from these diatexites, via metatexites, to the mica schists of the Bothnia Belt. One peculiarity of this terrane is the occurrence of topaz-bearing pegmatites.

The layering of the metapelites is more preserved in the southern terrane than in the northern terrane. This is mostly due to the intense melting and retrograde metamorphism that occurred in the northern gneisses. From an E-W trending andalusite zone containing RE-pegmatites (in the middle of Seinäjoki region), the regional metamorphic grade increases gradually towards the two high-grade metamorphic terranes, with a facies series of andalusite-sillimanite type. Northwards and southwards, there are four regional metamorphic zones, commencing with the andalusite zone. An E-W trending major deformation zone partly separates the high-grade metamorphic terranes.

Key words (GeoRef Thesaurus, AGI): granitoids, pegmatite, diatexites, quartz monzonite, absolute age, U-Pb, columbite, pyroxene, chemical composition, mineral composition, metamorphism, contact aureole, metamorphic zoning, metapelite, Bothnia Belt, Vaasa Migmatite Complex, Central Finland Granitoid Complex, Proterozoic, Svecofennian, Ostrobothnia, Seinäjoki, Finland

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PREFACE

The bedrock of southern and central Finland is characterized by large areas of Svecofennian granitic rocks (1.89-1.82 Ga) and a diverse network of metamorphosed supracrustal sequences (see Korsman et al. 1997). There are several reasons for studying the relationships between these Finnish intrusives and their supracrustal country rocks, and these include the following questions and considerations:

- there may be a genetic link between granitic rocks and neighbouring high-grade metamorphic migmatite terranes
- contact metamorphic studies allow clarification of a concept of regional(-scale) contact metamorphism in which intrusives have collectively increased the regional thermal gradient
- an understanding of the local relationship between granitoids and metamorphism is useful in making geotectonic comparisons at orogenic scale
- why does post-kinematic mangeritic magmatism usually occur in high-grade metamorphic terranes?
- why do rare-element pegmatites usually occur within supracrustal rocks that were metamorphosed under relatively low-grade conditions?

The Svecofennian granitoids in Finland have traditionally been divided into syn-, late- and post-kinematic groups (Eskola 1932, Saksela 1936, Nurmi and Haapala 1986). In the plate tectonic realm, the synkinematic granitoids are predominantly collision-related, foliated tonalites and granodiorites, while the post-kinematic granitoids (1.88-1.82 Ga) are compositionally granitic or monzonitic rocks that postdate the main stage of crustal thickening (e.g. Korsman et al. 1997). The emplacement of late- and post-kinematic rocks in Finland has been a long and episodic process between 1.88 and 1.77 Ga (Ehlers et al. 1993, Vaasjoki 1996, Eklund et al. 1998, Nironen et al. 2000). In addition to these complexities, the extension of the terms syn-, late- and post-kinematic over the entire Svecofennian Orogen faces severe problems due to overlapping ages (see Nironen et al. 2000). Metamorphically, the Svecofennian sequences may be divided into two parts (Korsman et al. 1984, Korja et al. 1994): the northern part (where metasediments are psammitic) is characterized by tonalite migmatites and the southern part (where metapelites with excess aluminium in relation to alkalies and calcium are dominant) by potassium granite migmatites. Both parts underwent intensive metamorphism 1.89-1.88 Ga ago, but only the southern part was overprinted by a strong thermal pulse 1.86-1.81 Ga ago (Korsman et al. 1984). Metamorphism in the Svecofennian sequences is generally characterized by low P/high T conditions and a change in metamorphic grade in narrow zones (e.g. Pajunen 1994). Consequently, the relationship between granitoids and their metamorphic environment probably varies within the Svecofennian of Finland.

In the Svecofennides, the relationships between granitoids, metamorphism and deformation have been studied locally in eastern, central and southwestern Finland (see Hölttä 1995, Korsman et al. 1999, Nironen et al. 2000). These studies have shown that there are several types of post-kinematic granitoids and that the time interval between intrusions, regional metamorphism and deformation varies locally. In western Finland, the relationship between granitic rocks and tectono-metamorphism is poorly known. Nevertheless, the region is promising for studies with the aim of clarifying this relationship because it contains, locally, several granitic rock types, while metamorphic zoning, polyphase deformation and shear zones are common (e.g. Mäkitie and Lahti 1991).

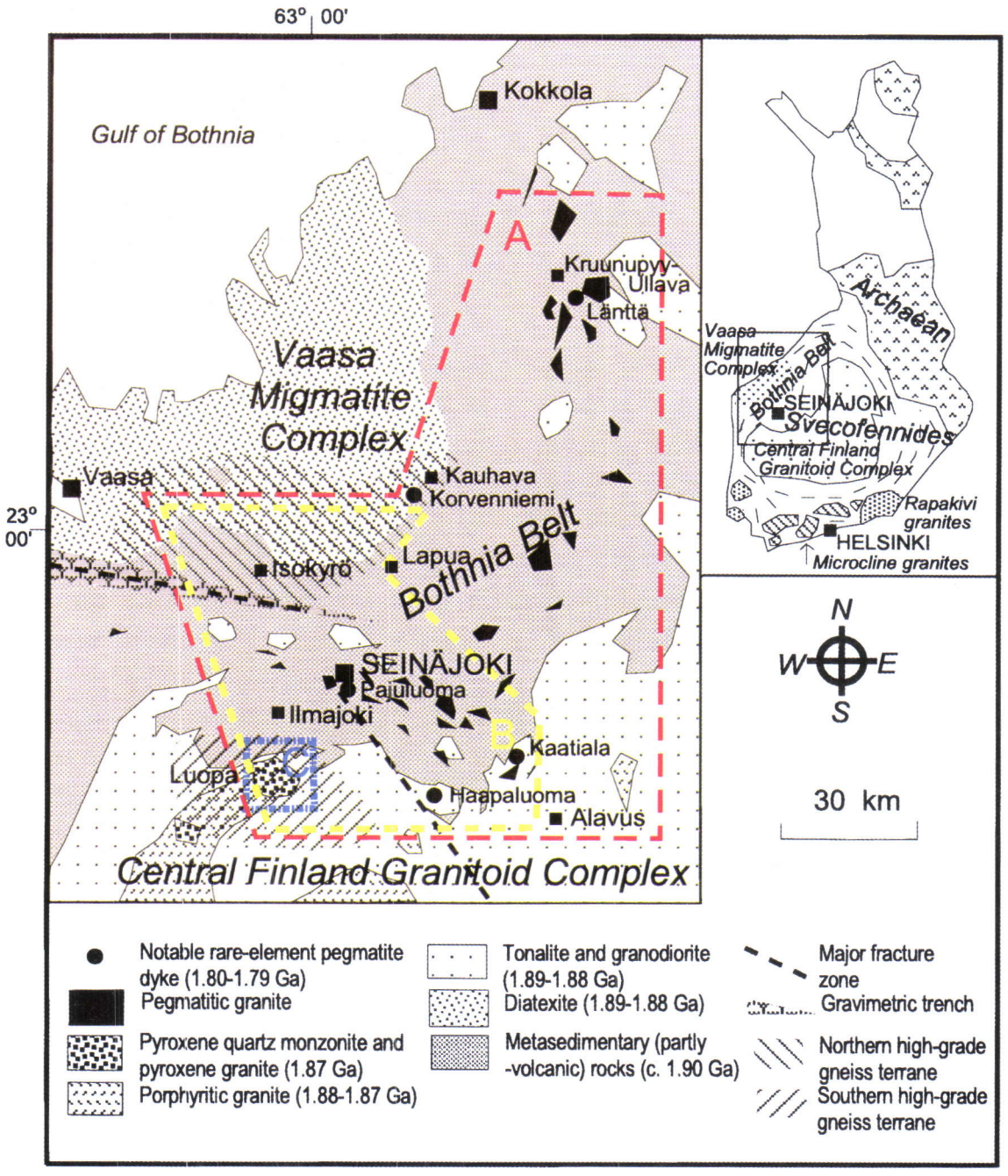

Fig. 1. Simplified geological map of the Seinäjoki region and its surroundings, western Finland. The areas delimited (A-C) are the study areas for the particular papers. The map is modified after Korsman et al. (1997).

This volume describes 1.87-1.79 Ga Svecofennian post-kinematic granitic rocks and the metamorphism of their slightly older country rocks in the Seinäjoki region, western Finland (Fig. 1). The intrusive rocks studied are a fayalite-augite quartz monzonite intrusion (age 1.87 Ga) and barren and rare-element pegmatite dykes (age 1.86-1.79 Ga) with associated pegmatitic granites. These rocks postdate the main stage (1.89-1.88 Ga, e.g. Vaasjoki 1996) of crustal thickening and are the youngest plutonic and dyke rocks in the region. The granitic rocks studied also cross-cut the main regional deformation phase (D_2) (Mäkitie 1999); however, many pegmatites are deformed by later folding.

The Seinäjoki region consists of an area of 6000 km^2 around the town of Seinäjoki (see Fig. 1). The region occupies part of the Central Finland Granitoid Complex and the Bothnia Belt, which also includes the Vaasa Migmatite Complex (formerly called the Vaasa granite) (Fig. 1). For comparison, the granitic pegmatites are examined locally over an area (Ostrobothnia) that is slightly larger than the Seinäjoki region. The Seinäjoki region belongs to the accretionary arc complex of the central and western Finland, which is part of the Palaeoproterozoic Svecofennian domain (cf. Korsman et al. 1997). The present volume includes three papers (see Fig. 1):

A) "Svecofennian rare-element granitic pegmatites of the Ostrobothnia region, western Finland; their metamorphic environment and time of intrusion" by R. Alviola, I. Mänttäri, H. Mäkitie and M. Vaasjoki, has some major themes. First, it presents an updated review of the rare-element (RE)-pegmatites and of their metamorphic environment in the Ostrobothnia region, locally with tectonic information. New U-Pb age data (ca. 1.80 Ga, mostly Nb-Ta mineral ages) from the characteristic RE-pegmatite types of the region are presented. As a conclusion, it is maintained that these RE-pegmatites (with other similar pegmatites elsewhere in Finland) may indicate a distinct tectonomagmatic episode separate from the Svecofennian orogenic culmination. Most of the new data given is from the Seinäjoki region.

B) "Compositional variation of granitic pegmatites in relation to regional metamorphism in the Seinäjoki region, western Finland", by H. Mäkitie, N. Kärkkäinen, S.I. Lahti and R. Alviola, describes the occurrence, mineralogy and chemical composition of the barren and rare-element pegmatites of the Seinäjoki region. The emphasis of the paper is on the whole rock chemistry and mineralogical composition of the pegmatites. The mutual relationship between the compositional variation of pegmatites and country rocks is discussed. In the region, there is a distinct positive correlation between the RE-pegmatites and the relatively low-grade metamorphic country rocks. Two slightly different high-grade metamorphic terranes (with various metamorphic zones) are described. The northern terrane contains diatexites and the southern terrane contains mangeritic magmatism. The RE-pegmatites are suggested to be derivates from the anatectic melts initially formed in the high-grade terranes. These melts intruded upwards into areas of lower metamorphic grade. Simultaneous fractionation of melts produced different types of pegmatite dykes.

C) "The fayalite-augite quartz monzonite (1.87 Ga) of Luopa, western Finland, and its contact aureole", by H. Mäkitie and S.I. Lahti, gives a description of the practically undeformed Luopa mangeritic intrusion that lies 20 km southwest of the town of Seinäjoki, at the boundary between the Central Finland Granitoid Complex and the Bothnia Belt. The structure, petrography, chemical and mineralogical composition of the intrusion, and its contact metamorphic aureole, are studied. The quartz monzonite crosscuts the neighbouring pyroxene granite and foliated tonalites, and deforms its metapelitic country rocks: it is post-kinematic relative to the main tectonic events in the region. The emplacement of the quartz monzonite resulted in the formation of a narrow zone of contact metamorphism, which overprints the effects of regional metamorphism. The inner part of the zone includes granulite-grade mineral assemblages not found elsewhere in the Seinäjoki region.

Espoo, October 27, 2000 *Hannu Mäkitie*

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SVECOFENNIAN RARE-ELEMENT GRANITIC PEGMATITES OF THE OSTROBOTHNIA REGION, WESTERN FINLAND; THEIR METAMORPHIC ENVIRONMENT AND TIME OF INTRUSION

by

Alviola, R., Mänttäri, I., Mäkitie, H. and Vaasjoki, M.

Alviola, Reijo, Mänttäri, Irmeli, Mäkitie, Hannu, & Vaasjoki, Matti 2001. Svecofennian rare-element granitic pegmatites of the Ostrobothnia region, western Finland; their metamorphic environment and time of intrusion. *Geological Survey of Finland, Special Paper 30,* 9-29, 10 figures, 5 tables and one appendix.

Ostrobothnia is situated in western Finland along the coast of the Gulf of Bothnia. It consists of the Ostrobothnia Schist Belt, the Vaasa Migmatite Complex and part of the Central Finland Granitoid Complex, and is part of the Paleoproterozoic 1.90-1.87 Ga Svecofennian accretionary arc complex of central and western Finland. The Ostrobothnia Schist Belt consists of turbiditic metapelites exhibiting Paleoproterozoic (-Archean) provenance, and local metavolcanic rocks as interlayers. The large Central Finland Granitoid Complex, which contains syn- and post-kinematic granitoids occurs east of the schist belt and to the west of the schists are the diatexites of the Vaasa Migmatite Complex. Most of the Ostrobothnian supracrustal rocks were metamorphosed under low-pressure amphibolite facies, locally granulite facies, conditions 1.89-1.88 Ga ago. The metamorphic grade is lowest in the central and eastern parts of the Ostrobothnia Schist Belt. The grade increases towards the Vaasa Migmatite Complex and also southwards in the southern part of the schist belt.

The rare-element (RE) granitic pegmatites of the region are divided into 12 groups. All pegmatites belong to the LCT (Li,Cs,Ta) family except the Alavus NYF (Nb,Y,F) pegmatites. Most pegmatites are beryl pegmatites of the beryl-columbite subtype, but three groups belong to the spodumene subtype and one group to the albite-spodumene subtype of complex pegmatites. Two groups of the beryl-columbite subtype are topaz/andalusite- and fluorite-bearing with biotite predominant over muscovite and without tourmaline. All the RE-pegmatites of the schist belt occur in the lower to medium amphibolite facies rocks with the exception of the topaz/andalusite- and fluorite-bearing pegmatites, which are in upper amphibolite facies mica gneisses surrounding the Vaasa Migmatite Complex. The pegmatites at Alavus lie within the Central Finland Granitoid Complex.

U-Pb ages have been determined from LCT pegmatite hosted Nb-Ta minerals. Columbites from the Kaatiala, Ullava, Haapaluoma, and Korvenniemi pegmatites and tapiolite from the Seinäjoki pegmatite yielded ages between 1.80-1.79 Ga. The dated columbites are ferro- and manganocolumbites. The Alavus NYF pegmatites do not contain Nb-Ta minerals applicable for U-Pb dating. The dated monazite from these pegmatites gives an age of 1.86 Ga and, thus, these dykes may be linked to the evolution of the 1.89-1.87 Ga Central Finland Granitoid Complex. The 1.80-1.79 Ga old RE-pegmatites indicate a tectono-magmatic episode distinct from the culmination (c. 1.88 Ga) of the Svecofennian orogeny.

Key words (GeoRef Thesaurus, AGI): pegmatite, absolute age, U-Pb, columbite, monazite, metamorphism, metapelites, Svecofennian, Proterozoic, Ostrobothnia, Bothnia Belt, Seinäjoki, Finland

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INTRODUCTION

Ostrobothnia in western Finland lies roughly at the centre of the Precambrian Fennoscandian (Baltic) Shield. The main geological units of the region are (1) the marginal areas of the Central Finland Granitoid Complex in the south and east, (2) the arcuate Ostrobothnia Schist Belt in the centre, and (3) the Vaasa Migmatite Complex in the northwest (Fig. 1). All these rocks belong to the Paleoproterozoic 1.90-1.87 Ga Svecofennian accretionary arc complex of central and western Finland (Korsman et al. 1997).

The prominent Kaatiala pegmatite was probably the first pegmatite known from Ostrobothnia. It was quarried for quartz on a small scale for a long time before industrial quartz production started in 1942. The production of feldspar for the domestic ceramic industry commenced soon afterwards.

The increase of rare element (RE) prices in the 1950s resulted in exploration for RE-pegmatites in southern Finland and in Ostrobothnia, by the Geological Survey of Finland (GTK) and private enterprises. Over the next 25 years pegmatite exploration covered the whole region. Spodumene pegmatites, containing a few millions of tons of Li-ore, were traced using boulder fans in the marshy Kruunupyy-Ullava area in the northern part of the region. The Haapaluoma complex pegmatite was discovered and opened for feldspar production in 1961. Several new minor RE-pegmatite populations were detected, and the Seinäjoki tin pegmatites were found in 1979. Only two RE-pegmatite groups, namely the Haapaluoma-Kaatiala and the Seinäjoki groups, have been studied in detail (Haapala 1966, Nieminen 1978, Oivanen 1983, Nurmela 1985, Alviola 1989a).

Granitic pegmatites are classified into (1) abyssal, (2) muscovite, (3) rare-element and (4) miarolitic according to their metamorphic environment, typical minor elements, nature of their parental granites and relation to their parental granites (see Ginsburg et al. 1979, Černý 1990). Pegmatites of the rare-element class are broadly divided into three families: LCT (Li,Cs,Ta), NYF (Nb,Y,F) and a rare mixed LCT+NYF family. These are further classified into 5 types and 9 subtypes according to their rock-forming mineralogy and geochemical signature (Černý 1991, Novák and Povondra 1995) (Table 1).

About 30 major RE-pegmatite populations occur in Finland (Alviola 1989a), many of them contain several groups. Two of the groups have Archean host rocks, while others occur in a Proterozoic environment. Most of the RE-pegmatites are Paleoproterozoic in age, but the pegmatites of the anorogenic rapakivi granites are Mesoproterozoic. Prior to the present study, isotopic ages had been determined for only three of the 30 known RE-pegmatite populations.

Table 1. Classification of granitic pegmatites of the rare-element class as modified by Černý (1998).

TYPE	SUBTYPES	FAMILY
Rare-earth	Allanite-monazite	NYF
	Gadolinite	NYF
Beryl	Beryl-columbite	LCT,NYF
	Beryl-columbite-phosphate	LCT
Complex	Spodumene	LCT
	Petalite	LCT
	Amblygonite	LCT
	Lepidolite	LCT
	Elbaite	LCT
Albite-spodumene		LCT
Albite		LCT

Most of the RE-pegmatites in Finland are members of the LCT family of pegmatites, but 6 groups belong to the NYF family of pegmatites. The Kemiö pegmatite group in southwestern Finland includes some NYF pegmatites, while the remainder belong to the LCT family.

Ostrobothnia is rich in RE-pegmatites, which occur as 12 groups (Fig. 1), separated by 10-25 km wide areas sparse in pegmatites. All groups contain beryl pegmatites of the LCT family, occasionally with some columbite±phosphates. Some pegmatites in the Haapaluoma-Kaatiala, Tarikko and Seinäjoki groups are complex pegmatites with spodumene±lepidolite and Li-tourmaline. The RE-pegmatites at Seinäjoki are particularly enriched in cassiterite. The albite-spodumene pegmatites of the Kruunupyy-Ullava group contain the largest Li-concentration in Finland. The only pegmatites of the NYF family in the region occur at Alavus. The columbite-bearing topaz pegmatites along the marginal zone between the Vaasa Migmatite Complex and the Ostrobothnia Schist Belt do not fit very well into the subtype of beryl-columbite pegmatites – particularly due to their topaz and fluorite contents.

In the following, the main geological units, metamorphic zones and tectonic setting of Ostrobothnia and its RE-pegmatite groups are briefly described, and some hitherto unpublished U-Pb data on zircons and monazites are presented. U-Pb analyses of columbite, tapiolite and monazite from six RE-pegmatites representing five different RE-pegmatite groups provide further isotopic constraints. The derivation of the pegmatite-forming magmas, the time of intrusion of the granitoids and the timing of RE-pegmatite formation are then discussed in a regional context.

LITHOLOGY OF OSTROBOTHNIA

Central Finland Granitoid Complex

The large Central Finland Granitoid Complex mostly comprises granodiorites and granites (Nurmi and Haapala 1986, Nironen et al. 2000) (Fig. 1). In the context of the Svecofennian orogeny, these granitoids can roughly be divided into the following two groups (Korsman et al. 1997, Nironen et al. 2000): collision-related, foliated synkinematic tonalites and granodiorites (1.89-1.88 Ga), and porphyritic or even-grained, locally pyroxene-bearing post-kinematic granites and quartz monzonites (1.88-1.87 Ga), post-dating the main stage of crustal thickening. Adjacent to the eastern side of the Ostrobothnia Schist Belt, synkinematic granodiorites and tonalites of the Central Finland Granitoid Complex dominate.

Ostrobothnia Schist Belt

The Ostrobothnia Schist Belt (practically the same area as the Bothnia Belt, see Nironen 1997) forms an arcuate belt about 350 km long and up to 70 km wide (Fig. 1). Bedrock maps and their explanations indicate that turbiditic metapelites, often with porphyroblastic texture, are the most common rocks within the belt (Nykänen 1960a,1960b, Tyrväinen 1970, 1971, Lonka 1971, Pipping 1979, Laitala 1981, Lahti and Mäkitie 1990, Mäkitie and Lahti 1991, Mäkitie et al. 1991, Pipping and Vaarma 1992, Vaarma and Pipping 1997, Lehtonen and Virransalo, in press). There are also metavolcanic rocks, such as metalavas at Evijärvi (Vaarma and Kähkönen 1994, Vaarma and Pipping 1997) in the northern part of the belt, and porphyrite sills hosting Sb mineralizations, particularly around Seinäjoki in the south (Mäkitie and Lahti 1991).

Sheets of pegmatite granite intrusions, barren pegmatite dykes and RE-pegmatites are common rocks in the Ostrobothnia Schist Belt (e.g. Alviola 1989b), and occur most frequently in the eastern and southern parts of the schist belt. Minor orogenic intrusive rocks, mainly foliated tonalites, granodiorites and granites, occur within the belt.

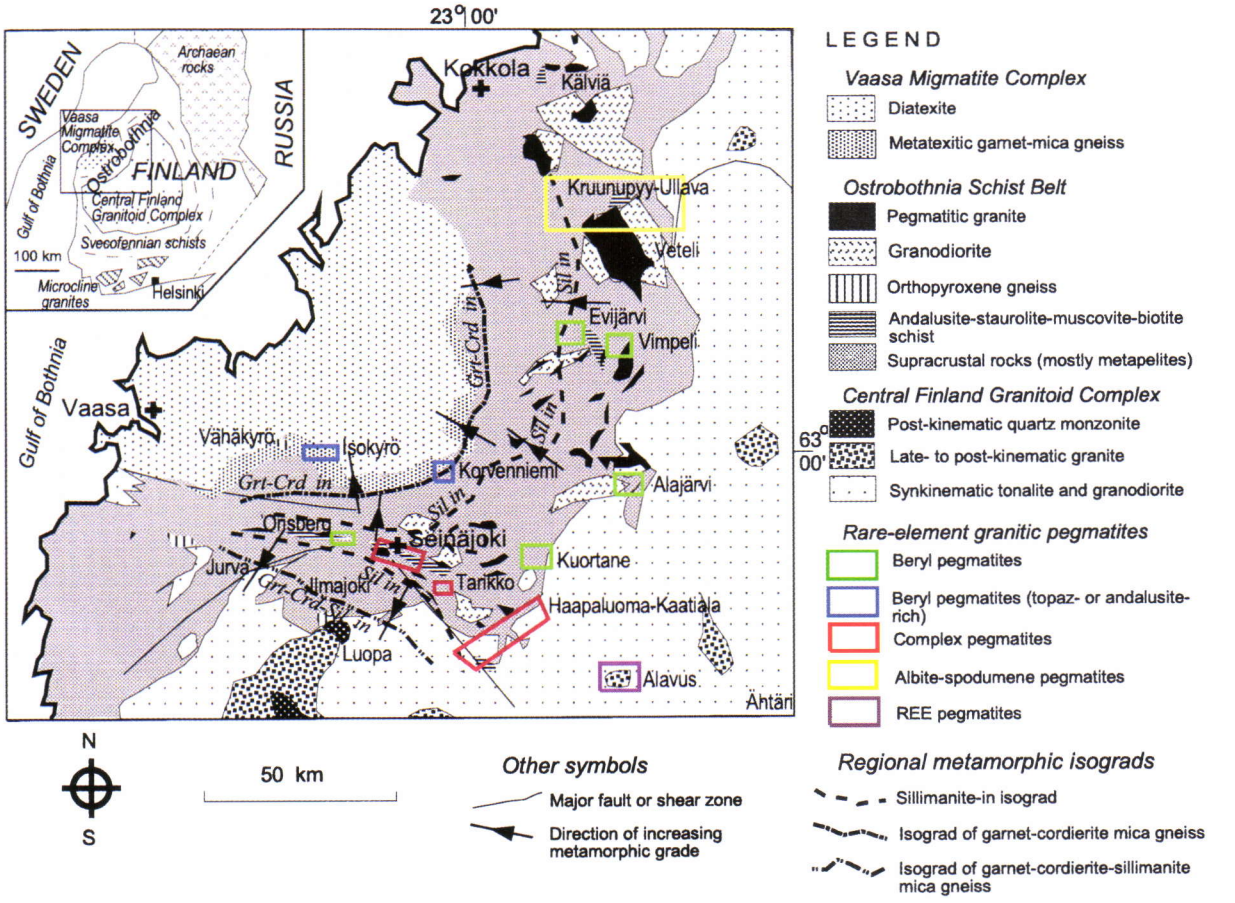

Fig. 1. Locations of the RE-pegmatite groups and the main metamorphic isograds in the Ostrobothnia Schist Belt. Abbreviations are after Kretz 1983 (map after Tyrväinen 1970, 1971, Pipping 1979, Västi 1986, Lahti and Mäkitie 1990, Mäkitie et al. 1991, Pipping and Vaarma 1992, Korsman et al. 1997, Lehtonen and Virransalo, in press).

Vaasa Migmatite Complex

A large area composed of mainly migmatitic rocks occurs north and west of the Ostrobothnia Schist Belt. These are locally granodioritic and contain leucotonalite-granodiorite veins (cf. Korsman et al. 1997). Traditionally, this area has been called the Vaasa granite (e.g. Laitakari 1942) and recently the Vaasa Granitoid Complex (Mäkitie et al. 1999). As these rocks clearly lie at the centre of a thermal dome surrounded by zones caused by regional metamorphism (Fig. 1 and the following paragraph), we have chosen to use the name Vaasa Migmatite Complex, even though some of the rocks resemble granitoids.

Biotite is usually the only mafic mineral in the rocks of the Vaasa Migmatite Complex. Garnet is a typical accessory mineral and supracrustal rock remnants are common. The southern part of the Vaasa Migmatite Complex is composed of granodioritic, peraluminous even-grained or porphyritic diatexites (Mäkitie et al. 1999). The rocks of this complex gradually change via metatexites to the mica schists of the Ostrobothnia Schist Belt.

METAMORPHISM AND TECTONICS

Metamorphism

The metamorphic grade is lowest in the central and eastern parts of the Ostrobothnia Schist Belt, where andalusite and/or staurolite-bearing muscovite-biotite schists dominate: e.g. in the Seinäjoki, Veteli and Kälviä areas (Fig. 1) (Västi 1986, Mäkitie and Lahti 1991, Vaarma and Pipping 1997). Most of the

RE-pegmatites occur within these mica schists. In metapelites, the mineral pair muscovite-quartz without sillimanite indicates low-pressure, lower amphibolite facies P-T conditions. The regional metamorphic grade increases from these mica schists towards the Vaasa Migmatite Complex and the schists change into gneisses as sillimanite becomes stable. Near Seinäjoki the regional metamorphic grade also increases towards the southwest. Thus, both on the western and southern side of the mica schist there is a sillimanite-in isograd (see Fig. 1). It is noteworthy that muscovite is usually stable in metapelites up to several kilometres on the higher temperature side of the sillimanite-in isograd.

Migmatitic mica gneisses without muscovite, but exhibiting equilibrated mineral pairs of sillimanite-K-feldspar or cordierite-K-feldspar occur on the western, northwestern and southern side of the aforementioned sillimanite-muscovite-biotite gneisses (see Mäkitie and Lahti 1991, Lundqvist et al. 1997, Vaarma and Pipping 1997). However, to the west and north these mica gneisses change over a short distance (1-3 km) into the garnet- and cordierite-bearing metatexites surrounding the Vaasa Migmatite Complex. In addition, the garnet/cordierite ratio of the gneisses increases towards these migmatites. Sillimanite is rare, probably because it was a component in the dehydration reactions that formed garnet, cordierite and K-feldspar. Hercynite occurs locally. The garnet-cordierite pair in the rocks indicates metamorphic P-T conditions of the upper amphibolite facies.

Fig. 2. Migmatitic garnet mica gneiss; the country rock of the Korvenniemi columbite-fluorite-topaz pegmatite. Korvenniemi RE-pegmatite group. x = 6994200, y = 2449420. Photo: H. Mäkitie.

The metatexitic mica gneisses in the boundary zone between the Vaasa migmatites and the Ostrobothnian schists are usually characterized by intense retrograde metamorphism. For example, cordierite is decomposed into andalusite and pinite; there has been muscovitization and chloritization in places; and garnet is partly decomposed into biotite. These rocks, usually migmatitic, are the country rocks of the Korvenniemi and Isokyrö RE-pegmatites (Fig. 2).

Southwest of Seinäjoki there are migmatitic garnet-cordierite-sillimanite mica gneisses (Mäkitie 1999). In addition to the abundance of sillimanite, these mica gneisses differ from the garnet-cordierite mica gneisses because they contain relatively well preserved bedding structures.

Compositionally intermediate, locally psammitic orthopyroxene-bearing gneisses indicating granulite facies metamorphism occur in the southern part of the Ostrobothnia Schist Belt, especially at Vähäkyrö, western Jurva and southern Ilmajoki (Fig. 1). In these areas the metapelites are migmatitic garnet-cordierite mica gneisses with or without sillimanite. The Luopa quartz monzonite, some 20 km south of Seinäjoki, is surrounded by a narrow granulite-grade contact aureole that overprints regional metamorphism (Mäkitie and Lahti 1991).

Tectonic setting and deformation

In the Svecofennian orogeny, the closure of the Bothnian basin, which, in fact, refers to the central part of the Svecofennian Domain, was associated with intense deformation and migmatization of their greywacke and argillaceous sedimentary fill (Gaál and Gorbatschev 1987). Furthermore, post-collisional crustal thickening in the Svecofennian Domain eventually resulted in the formation of crustal melts that were mainly emplaced in the sedimentary rocks of the Bothnian basin. Nironen (1997) recently modelled the evolution of the whole Svecofennian orogeny as a progressive accretion of two arc complexes to the Archean craton between 1.91 and 1.87 Ga.

Discordant U-Pb zircon data (Fig. 3, Table 2) with $^{207}Pb/^{206}Pb$ ages in excess of 2.1 Ga from a migmatitic, partially melted garnet-cordierite-sillimanite mica

gneiss at Mannilankallio, Ilmajoki, probably indicate the presence of both Archean and Paleoproterozoic grains (Huhma et al. 1991, Mäkitie and Lahti 1991). U-Pb zircon analyses from the Routakallio porphyrite sill at Seinäjoki (Table 2) indicate a Svecofennian age (c. 1.9 Ga) of emplacement into mica schists for this metavolcanic rock (see also Mäkitie and Lahti 1991).

Regional metamorphism was attained in Ostrobothnia 1.89-1.88 Ga ago (e.g. Vaasjoki 1996). A rather discordant monazite U-Pb determination from the aforementioned mica gneiss at Mannilankallio, 20 km SW of Seinäjoki, has a $^{207}Pb/^{206}Pb$ age of c. 1.86 Ga and sets a minimum age for the formation of the monazite either during pelite metamorphism or cooling of the crust below c. 520 °C (Mäkitie and Lahti 1991, see also Smith and Barreiro 1990) (Fig. 3, Table 2). The age estimate is in agreement with the age of the regional metamorphic peak.

Most of the plutonic rocks in Ostrobothnia are synkinematic relative to the main stage of deformation. The U-Pb ages for synkinematic granitoids within the Ostrobothnia region are as follows: c. 1.89 Ga for a

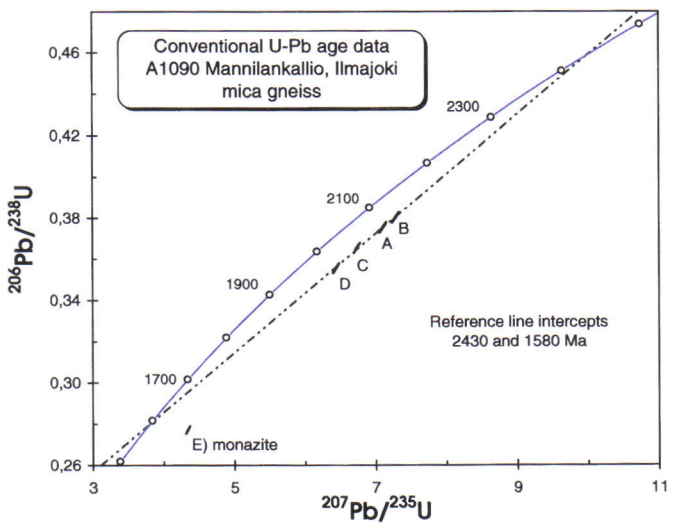

Fig. 3. Concordia diagram for analysed zircon and monazite fractions from a partially melted garnet-cordierite-sillimanite mica gneiss, Mannilankallio area.

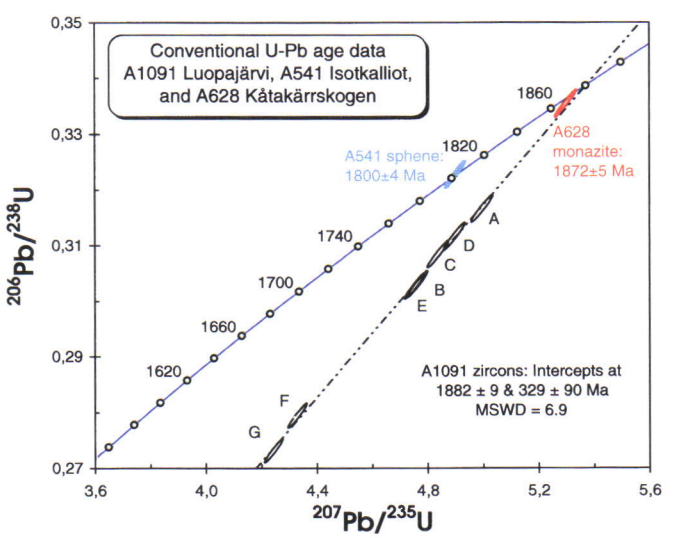

Fig. 4. Concordia diagram for analysed zircon fractions from the Luopajärvi tonalite and of sphene fractions from the Isotkalliot pegmatite, and monazite fractions of the granodioritic rock (diatexite) of Kåtakärrskogen.

gneissose granodiorite at Vimpeli (Kouvo and Tilton 1966); 1886±10 Ma for a synorogenic granite in the Ähtäri area c. 15 km southwest of the schist belt (Huhma 1986); 1882±9 Ma for the Luopajärvi tonalite at Jalasjärvi (Fig. 4, Table 2, see also Mäkitie and Lahti 1991) and 1883±6 Ma for a gneissose tonalite in the Evijärvi area, in the central part of the belt (Vaasjoki et al. 1996). However, the Evijärvi sample also contains older zircon populations (Vaasjoki et al. 1996) and ion-probe (NORDSIM) work has demonstrated the presence of Archean cores in some zircons (Vaasjoki et al. 1998).

The migmatites in the Vaasa Migmatite Complex are probably 1.89-1.88 Ga old because regional high-grade metamorphism and migmatization surrounding the complex was clearly associated with the culmination of the Svecofennian orogeny and the emplacement of the Svecofennian 1.89-1.88 Ga synkinematic granitoids (e.g. Vaasjoki 1996).

The quartz monzonite of Luopa (1872±2 Ma) in the southern part of the Ostrobothnia Schist Belt belongs to a suite of post-kinematic rocks in the Central Finland Granitoid Complex (Nironen et al. 2000, Mäkitie and Lahti, this volume). Pegmatites at Kälviä in the northern Ostrobothnia Schist Belt (see Wetherill et al. 1962) also clearly record a post-kinematic event.

A U-Pb monazite age of 1872±5 Ma (Fig. 4, Table 2) for a granodioritic rock from the Vaasa Migmatite Complex in the Kåtakärrskogen area, 10 km east of the town of Vaasa, is interpreted by the authors to indicate the cooling of crust below c. 600 °C. In addition, a U-Pb age of c. 1.84 Ga for sphene from a gneissose tonalite at Evijärvi indicates regional cooling of the crust below 500 °C (Vaasjoki et al. 1996), as probably does also the previously mentioned monazite age result for mica gneiss from Mannilankallio. The U-Pb sphene age 1800±4 Ma reported from the Isotkalliot quartz-rich pegmatite vein at Nurmo (Mäkitie and Lahti 1991) (Fig. 4, Table 2) is much younger than the aforementioned sphene from the gneissose tonalite in the Evijärvi area.

The Ostrobothnia region has undergone polyphase deformation: four deformation phases have been reported from the Evijärvi area (Vaarma and Pipping 1997) and five deformation phases at Seinäjoki (Mäkitie 1999). In both areas, the development of the S_2 schistosity has involved the synkinematic metamorphic growth of micas. Generally, the fold axes in the Ostrobothnia Schist Belt trend E-W and the region includes axial culmination and depression zones (Saksela 1935). Unfortunately, structural analysis remains to be carried out on large areas of Ostrobothnia.

In the Seinäjoki area, the pegmatite granites and associated pegmatite dykes cut the main foliation (S_2); however, some pegmatites also cut D_3 structures (Fig. 5a) (Mäkitie 1999). Many pegmatites are deformed (e.g. the cassiterite dyke of Pajuluoma in the town of Seinäjoki) and some are folded by the D_5 deformation

a)

b)

Fig. 5. Relationships between pegmatite dykes and deformation phases in the Seinäjoki RE-pegmatite group. a) Pegmatite dyke, one metre thick, sharply cutting the foliation (S_3) of mica gneiss. x = 6964300, y = 2435610. b) Crest of F_5 fold strongly deforming one metre thick pegmatite dyke. x = 6960330, y = 2443950. Photos: H. Mäkitie.

Table 2. U-Pb isotopic age data from western Finland.

Sample information	U	Pb	$^{206}Pb/^{204}Pb$	ISOTOPIC RATIOS[a,b]			APPARENT AGES / Ma (±2sigma)		
Analysed mineral / fraction	ppm		measured	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A1019 Routakallio, Seinäjoki, plagioclase porphyrite									
A) zircon d>4.5	175	56	6055	0,3208	5,089	0,1150	1793	1834	1880
B) zircon d: 4.3-4.5	289	91	7457	0,3153	5,033	0,1158	1766	1824	1892
C) zircon d>4.5, abraded	180	58	5228	0,3218	5,120	0,1154	1798	1839	1886
A1090 Mannilankallio, Ilmajoki, mica gneiss									
A) zircon d>4.5	261	105	14070	0,3754	7,101	0,1372	2054	2124	2192
B) zircon d>4.5, abraded	266	109	13980	0,3803	7,281	0,1389	2077	2146	2213
C) zircon d: 4.3-4.5	503	196	12400	0,3654	6,732	0,1336	2007	2076	2146
D) zircon d: 4.2-4.3	877	332	9276	0,3556	6,437	0,1313	1961	2037	2115
E) monazite	8612	5893	50170	0,2772	4,340	0,1135	1577	1700	1857±6
A1091 Luopajärvi, Jalasjärvi, tonalite									
A) zircon d: 4.3-4.5, +100mesh, abraded	669	224	8595	0,3167	4,996	0,1144	1773	1818	1870
B) zircon d:4.3-4.5, +100mesh	689	221	6821	0,3031	4,763	0,1140	1706	1778	1864
C) zircon d:4.2-4.3, +100mesh	831	271	6766	0,3085	4,837	0,1137	1733	1791	1859
D) zircon d:4.2-4.3, +100mesh, red	941	312	5008	0,3117	4,896	0,1139	1748	1801	1863
E) zircon d:4.2-4.3, +100mesh, white	808	258	8767	0,3028	4,754	0,1139	1705	1776	1862
F) zircon d:4.0-4.2, +100mesh, red	1190	352	5610	0,2795	4,329	0,1124	1588	1698	1838
G) zircon d:4.0-4.2, +100mesh, white	1031	298	6996	0,2732	4,243	0,1127	1556	1682	1843
A541 Isotkalliot, Seinäjoki, pegmatite[c]									
sphene	187	64	822	0,3227	4,896	0,1101	1802	1801	1800±4
A628 Kåtakärrskogen, Vaasa, granite[c]									
monazite	2218	644	53866	0,3353	5,295	0,1145	1864	1868	1872±5

a) Isotopic ratios corrected for fractionation, blank (500 pg) and age related common lead (Stacey & Kramers 1975).
b) Errors for Pb/U and $^{207}Pb/^{206}P$ ratios are less or equal than 0.8% and 0.25%, respectively.
c) Analyses by Dr. Olavi Kouvo, GTK.

(Fig. 5b). The ages of these relatively young deformations are not known.

At Seinäjoki, pegmatites in the mica schists have a greater tendency to be parallel to the foliation than those in more massive rocks such as plagioclase porphyrites (Nurmela 1985). According to Mäkitie (1999) the structure of the country rocks of the Seinäjoki RE-pegmatites is characterized by a large antiform in which S_2, S_3 and S_5 have almost parallel trends. It is probable that suitable extensional sites for pegmatite emplacement formed during the development of this antiform.

Haapala (1966, 1983) reported that the Haapaluoma pegmatite sharply intersects the lineation of its granodioritic country rock and that deformed pegmatites occur in the Haapaluoma-Kaatiala area. According to Vaarma and Pipping (1997), pegmatites at Evijärvi intersect local S_2 schistosity and their emplacement was probably synkinematic with D_3 and/or D_4.

Several faults and shear zones traverse Ostrobothnia (Fig. 1; see also Talvitie et al. 1975, Tyrväinen 1984, Korsman et al. 1997). Schollen migmatites occur in these shear zones. Locally, the foliation trends of rocks differ on different sides of the major shears which probably indicates that the zones are young (Tyrväinen 1984).

PEGMATITE GRANITES AND RE-PEGMATITE GROUPS

Most of the RE-pegmatites, particularly those containing spodumene, lie in the northern and southeastern parts of the Ostrobothnia Schist Belt (Fig. 1). A few beryl/beryl-columbite(± phosphate) pegmatite groups occur in the centre of the schist belt. Topaz/andalusite-bearing pegmatites are concentrated along the margin of the Vaasa Migmatite Complex, and the Alavus NYF-pegmatite group lies within the Central Finland Granitoid Complex. During the last decades, pegmatite exploration has been intensive in the southern and the northern parts of Ostrobothnia (Fig. 1). It is anticipated that more beryl pegmatites will be discovered especially from Evijärvi and Vimpeli around the Vimpeli pegmatite granites, in central Ostrobothnia.

Practically all of the RE-pegmatites within the Ostrobothnia Schist Belt are tourmaline-bearing with muscovite predominanting over biotite. The only exceptions are the topaz-bearing Isokyrö and Korvenniemi pegmatites where biotite clearly predominates over muscovite, and tourmaline is absent. The Alavus group pegmatites that occur outside the schist belt are also biotite pegmatites, but without tourmaline.

The Haapaluoma-Kaatiala group lies in the southeastern bend of the Ostrobothnia Schist Belt. The host rock of the pegmatites is usually a foliated tonalite/granodiorite. About 20 RE-pegmatites occur in the

Table 3. Classification (after Černý 1998, see also Table 1) of the Ostrobothnian RE-pegmatites. Characteristic rare minerals and country rocks of the RE-pegmatite groups are listed. Mineral abbreviations are after Kretz (1983) with the following additions: Amg = amblygonite, Clb = columbite, Tap = tapiolite, Trp = triphylite and Frg = fergusonite.

PEGMATITE TYPE	PEGMATITE SUBTYPE	PEGMATITE GROUP	CHARACTERISTIC RARE MINERALS IN THE PEGMATITES	COUNTRY ROCKS OF THE PEGMATITES
Rare-earth	Allanite-monazite	Alavus	Frg, Mnz, Aln	Granite & foliated tonalite
Beryl	Beryl-columbite	Vimpeli	Brl, Clb	St-Ms-Bt schist
		Orisberg	Brl, Clb	And-Ms-Bt schist
		Isokyrö	Toz, And, Clb	Grt(-Crd)-Bt gneiss
		Korvenniemi	Toz, Fl, Clb	Grt(-Crd-Hc)-Bt gneiss
	Beryl-columbite-phosphate	Evijärvi	Brl, Clb	St-Ms-Bt schist
		Kuortane	Brl, Clb, Trp	Foliated granodiorite & Ms-Bt gneiss
		Alajärvi	Brl, Clb, Trp	Foliated tonalite & Ms-Bt schist
Complex	Spodumene (with rare amblygonite pegmatites)	Haapaluoma-Kaatiala	Spd, Lpd, Elb, Brl	Foliated granodiorite & Sil-Ms-Bt gneiss
		Seinäjoki	Cst, Brl, Clb, Tap, Spd	And-Ms-Bt schist & metavolcanic rock
		Tarikko	Cst, Brl, Spd, Clb	Sil-Ms-Bt gneiss & And-Ms-Bt schist
Albite-spodumene		Kruunupyy-Ullava	Spd, Clb, Brl	And-Ms-Bt schist & metavolcanic rock

area and three of them have been studied in detail. These are the Hunnakko, Haapaluoma (Haapala 1966) and Kaatiala (Nieminen 1978) complex pegmatites. Of these, the large (30 m x > 200 m), almost horizontal (dip 15° NE) Kaatiala dyke is a spodumene pegmatite; the smaller (5 m x 70 m) Hunnakko dyke is an amblygonite pegmatite and the largest, the Haapaluoma dyke (10-30 m x 500 m, dip 50-60° N), is a spodumene-lepidolite subtype pegmatite. All other pegmatites are beryl pegmatites with or without columbite/tapiolite, except one dyke with red tourmaline and some lepidolite. The Haapaluoma dyke is a highly fractionated pegmatite with pollucite, but it is slightly similar to NYF pegmatites in character as indicated by accessory xenotime, monazite-cheralite and brockite. Altogether 48 minerals have been identified from the Haapaluoma-Kaatiala pegmatites. The Kaatiala pegmatite was extensively quarried during 1942-1968. It produced 160 000 tons of K-feldspar, 30 000 tons of quartz and a few tons of beryl and columbite. The Haapaluoma quarry was in operation during 1961-1998 and about 350 000 tons of K-feldspar were produced.

The pegmatite granites, i.e. pegmatite-generating granites, occur north of the Haapaluoma-Kuortane pegmatite population. Pegmatite granites contain coarse-grained pegmatitic patches or veins occasionally with a few beryl crystals.

The Seinäjoki, Tarikko and Orisberg groups occur at 10-15 km intervals in metasedimentary and metavolcanic rocks within the southern part of the schist belt, c. 20 km NW of the Haapaluoma-Kaatiala group (Fig.1). About 60 RE-pegmatites of the Seinäjoki group (Oivanen 1983, Nurmela 1985, Alviola 1986) lie between two sets of elongated pegmatite granite bodies. Barren and beryl pegmatites occur within and close to the pegmatite granites, whereas beryl-columbite or beryl-tapiolite pegmatites

a) b)

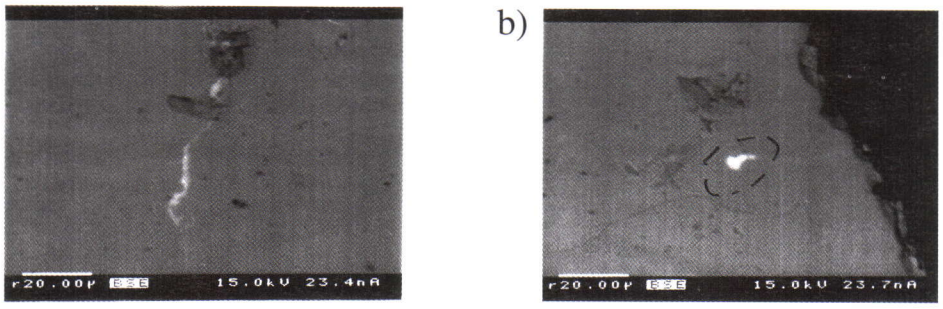

Fig. 6. Backscatter electron images of the dated columbites. a) Uranium-rich crack (light-coloured line in the centre of the Figure) in ferrocolumbite from the Kaatiala pegmatite. b) Uranium-, calcium- and titanium-bearing inclusion (marked by broken lie) in ferrocolumbite from the Korvenniemi pegmatite. Photos: B. Johanson.

with or without phosphates lie further away. Cassiterite-bearing beryl-columbite pegmatites occur in the centre of the pegmatite swarm, as do also the more south-easterly spodumene-bearing complex pegmatites. Altogether 47 minerals have been identified from the pegmatites at Seinäjoki. The Perälä cassiterite-bearing dykes are estimated to contain 100 000 tons of pegmatite with 0.3 % Sn to a depth of 100 m (Saltikoff 1982).

Long (0.1 km-3 km), narrow (20 m-200 m) pegmatite granite bodies at Seinäjoki contain numerous pegmatite veins running parallel, diagonally and at right angles to the contacts. Most of them are barren, but many are beryl-bearing and at least one of them is columbite-bearing. Usually the pegmatite granites are very coarse-grained with reddish or bluish phenocrysts of graphic feldspar up to 2 m in size. The matrix consists of K-feldspar, plagioclase, quartz, muscovite, crow-foot biotite, black tourmaline, garnet, apatite and arsenopyrite/löllingite. In places the pegmatite granite is fine- to medium-grained and banded with alternating aplitic (feldspar- and quartz-rich), and tourmaline- and garnet-rich bands.

There is a small group of RE-pegmatites at Tarikko (Alviola 1987). There are a few beryl-bearing pegmatites and one complex pegmatite of the spodumene subtype with accessory beryl, columbite, cassiterite, triphylite and alluaudite. The small Orisberg group consists of about half a dozen of beryl-pegmatites and two beryl-columbite pegmatites.

The Kuortane group RE-pegmatites are rather primitive beryl-columbite pegmatites with triphylite and some secondary phosphates e.g. strunzite.

The RE-pegmatites of the Kruunupyy-Ullava group are believed to belong to the albite-spodumene group of RE-pegmatites. These pegmatites and their exploitation were investigated for 30 years by a private company, but very little information has been published (Boström 1988). The dykes are 200-400 m long, some of them are very narrow, but others up to 10-20 m wide. In this marshy area the thickness of overburden is 2-10 m, and none of the dykes were originally exposed. However, the area is rich in erratic boulders of pegmatite, some of them barn-sized. The number of traced dykes is at least half a dozen. Li-ore

Table 4. Microprobe analyses of the dated columbite-tantalites from the RE-pegmatites.

PEGMATITE GROUP / host rock	columb.-tant. generation	FeOtot wt%	MnO wt%	MgO wt%	Nb_2O_5 wt%	Ta_2O_5 wt%	SnO_2 wt%	TiO_2 wt%	Sc_2O_5 wt%	U_2O_3 wt%	Total
ULLAVA R5 / gneiss	several										
average of 14 points in 3 grains		4,84	**14,46**	0,00	**64,17**	15,19	0,02	0,55	0,01	0,12	99,36
range		3.05-6.19	13.12-16.30	0.00-0.03	62.52-66.23	13.52-16.26	0.00-0.06	0.39-0.75	0.00-0.06	0.00-0.43	
KORVENNIEMI / mica gneiss	first										
average of 12 points in 4 grains		**11,83**	8,33	0,02	**71,11**	2,61	0,03	2,31	0,03	0,18	96,45
range		9.78-12.89	7.29-10.45	0.00-0.06	68.99-73.11	2.29-3.18	0.00-0.08	1.34-2.82	0.00-0.13	0.00-0.54	
SEINÄJOKI (Routakallio) * / pl. porphyrite	first										
average of 40 points in 6 grains		**13,18**	3,51	0,18	**42,72**	37,11	0,70	2,18	0,06	0,19	99,83
range		12.75-13.62	3.31-3.80	0.12-0.23	39.28-47.08	33.51-44.18	0.40-0.92	1.67-2.67	0.01-0.11	0.00-0.45	
SEINÄJOKI (Perälä W) * / mica schist	first										
average of 13 points in 2 grains		**9,22**	7,92	0,01	**43,51**	39,43	0.21	0,26	0,04	0,07	100,46
range		8.84-9.83	7.36-8.44	0.00-0.07	40.09-45.41	38.01-42.47	0.09-0.38	0.20-0.34	0.00-0.09	0.00-0.27	
KAATIALA / granodiorite	early										
average of 9 points in 3 grains		**13,56**	5,83	0,07	**69,47**	7,40	0,02	2,07	0,25	0,22	98,89
range		13.24-14.02	5.59-6.11	0.05-0.12	68.17-69.87	7.19-7.64	0.00-0.12	1.87-2.36	0.23-0.30	0.05-0.47	
HAAPALUOMA / granodiorite	early										
average of 15 points in 5 grains		5,30	**13,74**	0,00	**68,97**	8,34	0,28	1,53	0,84	0,28	99,28
range		2.70-5.98	12.75-16.55	0.00-0.01	67.97-71.05	8.20-8.54	0.00-0.46	1.27-1.60	0.77-0.90	0.11-0.47	

* These columbites were not dated (the dated Nb,Ta-mineral from the Seinäjoki pegmatites was TAPIOLITE).

reserves are a few million tons of spodumene-bearing pegmatite with about 1 % Li_2O. In addition to greenish and reddish spodumene, the dykes are known to contain feldspars, quartz, muscovite, accessory garnet, apatite, black tourmaline, triphylite, mangano-columbite, and trace beryl, amblygonite and bismuth.

The Kruunupyy-Ullava pegmatite dykes occur between three large granite-pegmatite granite intrusions. The southernmost of them, the Kaustinen pegmatite granite, is known to contain many pegmatitic pods. Some of them are beryl-bearing, but a few contain spodumene and a little triphylite.

The RE-pegmatites of the Alajärvi group lie south of the Alajärvi pegmatite granites. The pegmatite granites contain many pegmatitic pods, but all are barren. A few RE-pegmatites occur in three areas: 1) at Soini; a beryl pegmatite dyke rich in black tourmaline, 2) at Jyrkäys; a 10 m wide beryl-columbite pegmatite, and 3) at Kirkkokallio; a beryl-pegmatite with Fe,Mn-phosphates.

Almost twenty pegmatite granite bodies occur c. 30 km north and northwest of Alajärvi. At least some of the bodies are flat-lying, sheet-like intrusions. Barren pegmatites adjacent to them are rich in black tourmaline containing 10-25 % of schorl. A few beryl-columbite pegmatites, the Evijärvi group, with or without Li,Fe,Mn-phosphates were recently discovered unexpectedly northwest of the pegmatite granites, and southeast of the pegmatite granites, the Vimpeli group.

Near Isokyrö and Korvenniemi, in the boundary zone between the Vaasa Migmatite Complex and the Ostrobothnia Schist Belt, there are about half a dozen RE-pegmatites. These pegmatites are flat-lying, topaz- or andalusite- and fluorite-bearing, and, in particular, there is clearly more biotite than muscovite. The Isokyrö group pegmatites contain accessory beryl and columbite. There is light brown and diamond-shaped accessory muscovite and greenish yellow topaz crystals that are usually enveloped with brownish mica. West of Isokyrö, there are small andalusite-bearing pegmatites. These contain reddish gray, elongated crystals of andalusite up to 10 cm long, and some slightly altered cordierite crystals. Accessory minerals are muscovite, columbite and microlite. Large (100 x 40 m) pegmatites of the Korvenniemi group are more zoned than the Isokyrö dykes. Pinkish and fine- to medium-grained muscovite is present only in the coarse-grained K-feldspar cores, which are surrounded by quartz-topaz-microcline-plagioclase pegmatite with green and brown fluorite. Topaz crystals are subhedral, bluish-white in colour and 5-10 cm in diameter. In the presence of albitic 'sugary' plagioclase, the topaz crystals are surrounded by a greenish rim of fibrous mica. K-feldspar is in places amazonitic and graphic feldspar is absent. Accessory minerals are ferrocolumbite, apatite, garnet and pyrite.

The Alavus RE-pegmatite group occurs on either side of the border of the Alavus and Töysä parishes. About a dozen 1-10 m wide almost horizontal pegmatites occur in the vicinity of the Alavus microcline granite and one dyke lies within the granite. Most dykes are biotite pegmatites with accessory allanite crystals. Two of them have been quarried for feldspar in the 1980s. A few grains of fergusonite and monazite were found in the dumps of an abandoned quarry, located at the endocontact of the Alavus granite, and a few grains of allanite from another abandoned quarry.

The mineralogical characteristics and country rocks of the RE-pegmatite groups of Ostrobothnia are summarized in Table 3.

U-Pb DATING OF PEGMATITES

Radiometric dating of pegmatites in Precambrian areas has been problematic, as most traditional dating methods have often yielded results that, generally, have been of little value because of large analytical uncertainties. The main reason has been the nature and geological history of the available sample material. Thus, for example, zircon U-Pb data from pegmatites are very often discordant, which probably arises because the zircons were hydrated during crystallization and, as their lattices were originally distorted, were consequently more susceptible to lead loss by diffusion than ordinary zircons.

In a similar manner, precise U-Pb dating of sphene has often been hindered by the very high initial lead contents of the sample material, which has caused precise age estimates to depend on the isotopic composition chosen for the common lead correction. Determinations made with other isotopic systems, such as Rb-Sr and K-Ar, have proven unsatisfactory partly because of the uncertain closure temperatures of the minerals employed, and partly because of the effects of prolonged weathering on their isotopic systems.

In the 1990s, it became apparent that the columbite-tantalite series as well as Nb,Ta-minerals occurring in complex pegmatites may contain appreciable amounts

of uranium and are thus amenable for U-Pb dating (Romer and Wright 1992). Furthermore, experience (e.g. Romer and Smeds 1994, Romer et al. 1996) has shown that the results from these minerals are very often practically concordant, and thus the results obtained are rather precise.

Sample material

Various Nb,Ta-minerals were selected from different RE-pegmatite occurrences of Ostrobothnia for U-Pb dating. Monazite was used for the Alavus pegmatite, as fergusonite gave discordant ages. The Seinäjoki group is the only one where tapiolite was discovered. In the other groups columbite was used in age determination. The dated columbites have been analysed using a Cameca SX 50 electron microprobe at the Geological Survey of Finland and their chemical compositions are reported in Table 4.

The dated samples have usually been taken from large pegmatites. Manganocolumbite from the Kruunupyy-Ullava group is from the Länttä dyke at Ullava, which is the largest of the spodumene pegmatites. The sample is from a heavy mineral concentrate from drill hole 5/63. Backscatter electron (BSE) images indicate that the Ullava manganocolumbite is homogeneous in composition. The marginal parts of some grains contain inclusions of manganocolumbite with somewhat higher Ta-content.

The sample from Korvenniemi is a fragment of a single ferrocolumbite crystal from the largest pegmatite. The ferrocolumbite is inhomogeneous in composition, and contains small irregular patches with higher Mn-content and Ta-content. One U-Ca-Ti-bearing inclusion was detected.

The Haapaluoma columbite is a homogeneous manganocolumbite. The U-content is rather high (average 0.27 wt% U_2O_3), but no U-mineral inclusions were detected. Some cassiterite inclusions were found on grain margins. The Haapaluoma sample is in fact the Sc-bearing manganocolumbite$_I$ described by Haapala (1966). Columbite$_I$ crystallized with black tourmaline, cassiterite, albite and quartz in Na-replacement bodies. Needle-form Li-stage columbite gave discordant ages and was rejected.

The Kaatiala columbite represents also an early generation, which occurs as large platy crystals and is homogeneous ferrocolumbite in composition. The U-content (0.22 wt% U_2O_3) is roughly the same as that of the Haapaluoma mangano-columbite, but rare micron-sized U-mineral spots and fracture fillings were noted in BSE images (Fig. 6a).

The Alavus samples are from an abandoned feldspar quarry in the largest RE-pegmatite of the group. A U-rich fergusonite was rejected because it gave discordant ages and a monazite from the same pegmatite was used instead.

Analytical methods

For U-Pb dating, columbite and tapiolite grains were crushed and some fragments with fresh surfaces selected. Unleached and HF (5%, 10% and 20%) leached fractions were washed with a HNO_3-HCl-H_2O mixture in an ultrasonic bath. Sample dissolution was carried out in steel jacketed 0.3 ml Savillex® capsules or 2 ml Savillex® beakers. Aliquoting of the sample was done directly from the HF-solution to avoid the formation of some unsoluble complexes with HCl (methods described in Romer and Wright 1992). Monazite and fergusonite dissolution followed the procedure described by Krogh (1973).

A ^{235}U- ^{208}Pb-tracer was added to the aliquoted liquids. Uranium and lead were extracted using anion-exchange chromatography and uranium extracted from columbites and tapiolites were further purified with anion-exchange in HNO_3-form (Romer and Wright 1992). For isotopic composition measurement, lead was loaded on a single Re-filament (silica gel - phosphoric acid-mixture), and for concentration measurements lead and uranium (phosphoric acid) were loaded as a triple filament system (Pb on a centre Re-filament and U on a Ta-side filament). Based on multiple runs of SRM981 and U500 standards, the measured Pb and U ratios were corrected for mass fractionation of 0.12%/amu and 0.14%/amu, respectively.

U-Pb isotopic results

U-Pb data from the Kaatiala, Seinäjoki, Ullava and Haapaluoma columbites and tapiolites share some characteristics (Table 5): the U-Pb data from slightly (5% HF) leached and unleached fractions scatter on the

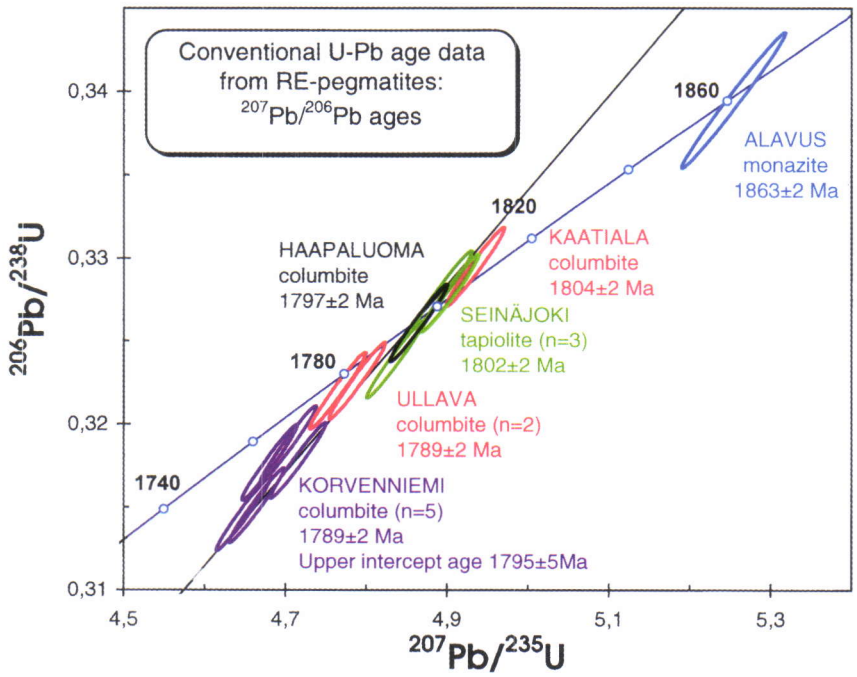

Fig. 7. Concordia diagram for analysed columbite, tapiolite and monazite fractions taken from the Ostrobothnia RE-pegmatites.

Table 5. U-Pb isotopic age data from RE-pegmatites, Ostrobothnia, western Finland. The data with bold letters are presented in Figure 7.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{206}Pb/^{208}Pb$	ISOTOPIC RATIOS[a,b]			APPARENT AGES / Ma (±2sigma)		
Analysed mineral / fraction	weight /mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
KAATIALA - Columbite											
A) Powdered grains	2.0	1731	478	2897	282	0,2828	4,251	0,1090	1605	1684	1783±2
B) Abraded grains	4.7	1795	524	1215	192	0,2901	4,329	0,1082	1642	1699	1769±2
C) 10% HF-leached grains	**2.4**	**1357**	**422**	**26113**	**443**	**0,3245**	**4,934**	**0,1103**	**1812**	**1808**	**1804±2**
SEINÄJOKI - Tapiolite											
36A) Untreated fragments	0.9	108	33	6586	1685	0,3189	4,835	0,1100	1785	1791	1799±2
36B) 5% HF-leached fragments	**2.1**	**113**	**35**	**9234**	**2068**	**0,3228**	**4,899**	**0,1101**	**1804**	**1802**	**1800±2**
36C) 10% HF-leached fragments	**1.8**	**110**	**34**	**2832**	**2954**	**0,3231**	**4,894**	**0,1099**	**1805**	**1801**	**1797±2**
36D) 20% HF-leached fragments	**1.6**	**127**	**39**	**16356**	**»10000**	**0,3229**	**4,903**	**0,1101**	**1804**	**1803**	**1802±2**
ALAVUS - Fergusonite											
106K) Translucent fragments	0.3	1.98%	5397	850	2.4	0,1973	2,877	0,1058	1161	1376	1728±2
106S) Turbid fragments	0.8	1.93%	6282	5038	2.3	0,2409	3,735	0,1125	1391	1579	1840±2
ALAVUS - Monazite											
112) Untreated fragments	0.2	449	1825	949	0.1	0,3271	5,141	0,1140	1825	1843	1864±2
112A) Abraded fragments	**0.9**	**507**	**2327**	**1157**	**0.1**	**0,3345**	**5,253**	**0,1139**	**1860**	**1861**	**1863±2**
KORVENNIEMI - Columbite											
Columbite plate; disordered lattice structure											
1) small fragments	1.7	1764	525	5354	102	0,3060	4,582	0,1086	1721	1746	1776±2
2) 20%HF-leached fragments	**0.6**	**1874**	**564**	**83414**	**99**	**0,3128**	**4,715**	**0,1094**	**1754**	**1770**	**1789±2**
3) abr. Fragments	2.3	1740	522	3474	113	0,3072	4,588	0,1083	1727	1747	1771±2
4) abr. 5%HF-leached fragments	**2.4**	**1658**	**495**	**14599**	**108**	**0,3096**	**4,648**	**0,1089**	**1739**	**1758**	**1781±2**
5) abr. 20%HF-leached fragments	**1.6**	**1705**	**514**	**12601**	**132**	**0,3127**	**4,680**	**0,1085**	**1754**	**1764**	**1775±2**
Columbite grain; ordered lattice structure											
6) abr. small fragments	2.4	1371	392	2470	84	0,2900	4,311	0,1078	1741	1695	1763±2
7) abr. fragments	**4.0**	**1492**	**447**	**10125**	**109**	**0,3101**	**4,664**	**0,1091**	**1741**	**1761**	**1784±2**
8) abr. 5%HF-leached fragments	**2.1**	**1289**	**392**	**6664**	**103**	**0,3138**	**4,703**	**0,1087**	**1759**	**1768**	**1778±2**
9) abr. small fragments	1.1	1332	377	1730	84	0,2847	4,216	0,1074	1615	1677	1756±3
ULLAVA (R5) - Columbite											
1d) abr. small fragments	0.3	2227	749	332	32	0,2909	3,974	0,0991	1646	1629	1607±5
2d) abr. larger fragments	2.4	1807	577	403	24	0,2816	4,049	0,1043	1599	1644	1702±5
3d) abr. 5%HF-leached fragments	0.5	1048	344	2388	77	0,3316	4,970	0,1087	1846	1814	1778±2
4d) abr. 10%HF-leached fragments	**0.6**	**1162**	**356**	**7089**	**128**	**0,3170**	**4,764**	**0,1090**	**1775**	**1779**	**1783±2**
5d) abr. 20%HF-leached fragments	**0.4**	**786**	**240**	**16300**	**135**	**0,3176**	**4,788**	**0,1094**	**1778**	**1783**	**1789±2**
HAAPALUOMA - Columbite I and II											
1/II) abr. 20%HF-leached fragments	3.6	869	283	2819	69	0,3296	5,043	0,1110	1836	1827	1816±2
2/II) abr. HNO3-leached fragments	0.9	2095	430	2119	65	0,2070	3,015	0,1057	1213	1412	1726±2
3/I) abr. 10%HF-leached fragments	**0.5**	**1622**	**502**	**12266**	**158**	**0,3211**	**4,864**	**0,1099**	**1795**	**1796**	**1797±2**

a) Isotopic ratios corrected for fractionation, blank (<50 pg) and age related common lead (Stacey & Kramers 1975).
b) Errors for Pb/U ratios are 0.6% and for $^{207}Pb/^{206}P$ ratios less or equal than 0.1%. Correlations for the $^{206}Pb/^{238}U$ vs. $^{207}Pb/^{235}U$ errors are 0.98.

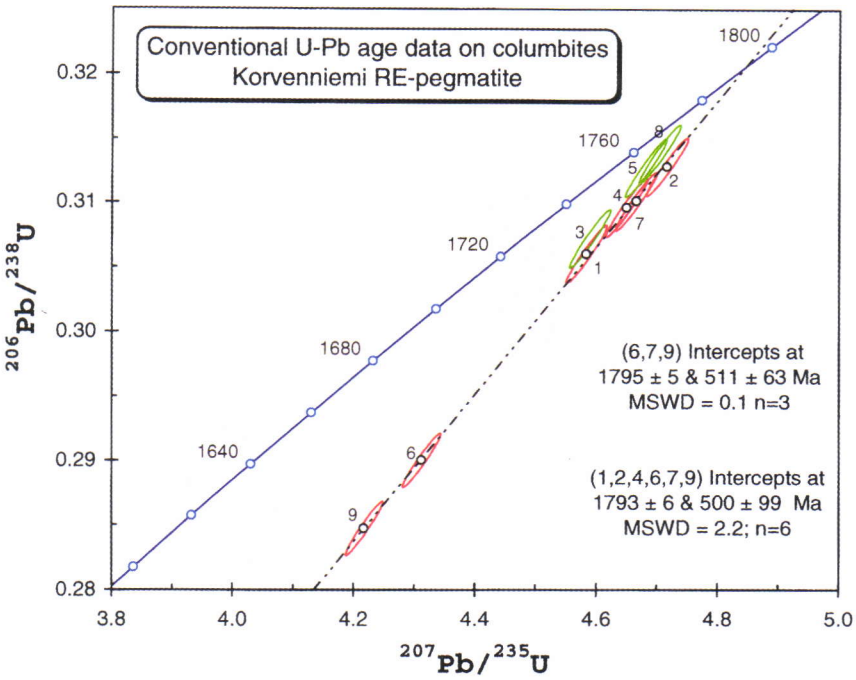

Fig. 8. Concordia diagram for analysed columbite fractions from the Korvenniemi RE-pegmatite.

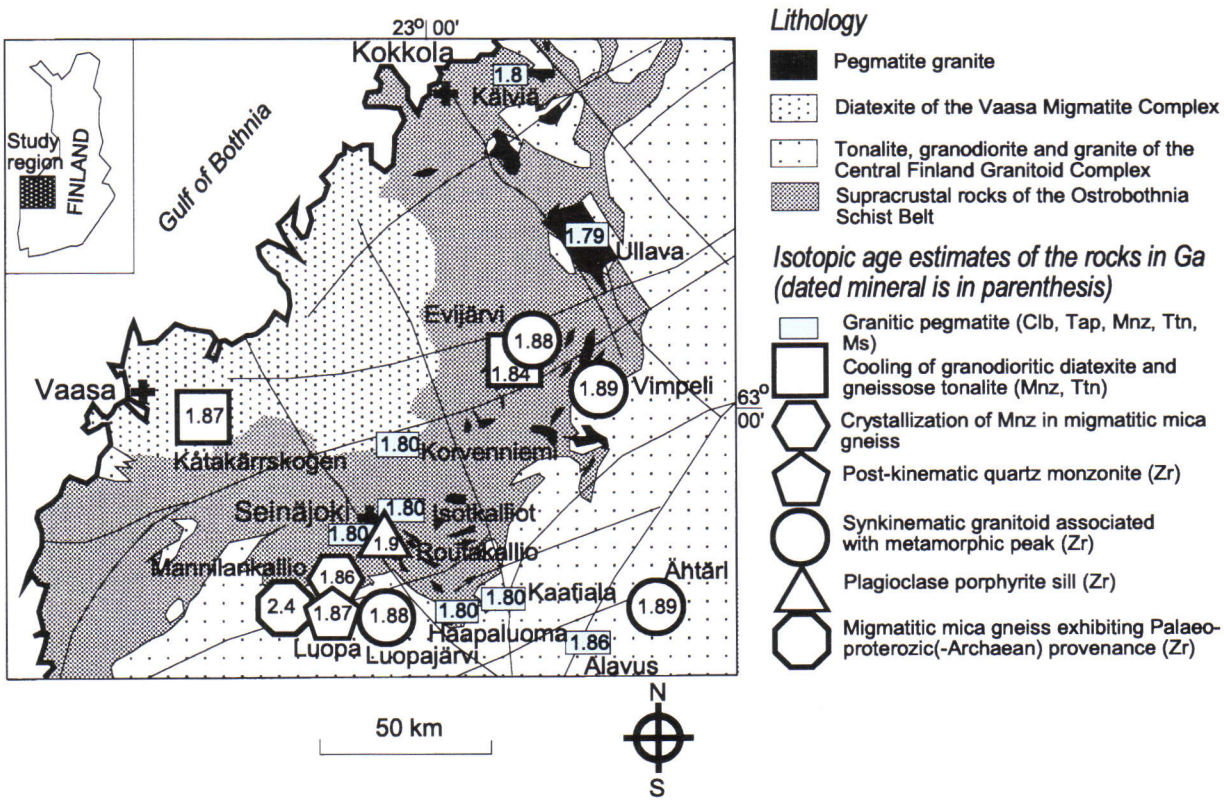

Fig. 9. Locations with age estimates of the isotopically dated rocks from Ostrobothnia. All age measurements, except that from the Kälviä area, are U-Pb measurements. For references, see Appendix 1 and for abbreviations, see Table 3. Lines indicate shear and fracture zones (map after Korsman et al. 1997).

concordia diagram and data points are normally or reversely discordant, but the 10% or 20% HF-leached fractions give concordant or nearly concordant ages. In addition, these leached fractions clearly have the most uranogenic $^{206}Pb/^{208}Pb$ and $^{206}Pb/^{204}Pb$ ratios. This can be explained by the existence of at least two separate U-Pb-systems in the mineral. When leached, the phase with more thorogenic lead (i.e. more ommon lead) is extracted and the $^{206}Pb/^{208}Pb$ and $^{206}Pb/^{204}Pb$ ratios increase significantly. These results agree with those reported by Romer and Wright (1992) and Romer et al. (1996), who have studied the U-Pb systematics of the columbite-tantalites and their potential use as a dating tool.

The concordant U-Pb age data for the Kaatiala, Seinäjoki, Ullava and Haapaluoma columbites and tapiolites are shown in Figure 7. The two unleached columbite fractions from the Kruunupyy-Ullava pegmatite give strongly discordant ages. However, the 10% HF-leached fraction (C) gives a concordant age of 1804±2Ma. All four analysed tapiolite fractions in the Seinäjoki pegmatite give concordant or nearly concordant ages. The 20% HF leached fraction (36D) gives the most concordant age of 1802±2Ma. Five fractions of columbite were analyzed from the albite-spodumene dyke from the Kaustinen area. In these, unleached fractions give discordant ages and the 5% HF-leached fraction gives a reversely discordant age. The 10% and 20% HF-leached fractions give nearly concordant ages. The highest $^{207}Pb/^{206}Pb$ age of the two concordant data points (4d and 5d) is 1789±2Ma, which may be considered as a good age estimate for the Ullava columbite. Two columbite grains were available from the Haapaluoma pegmatite. One gives discordant and reversely discordant ages while the other (3/I) is concordant at 1797±2Ma.

The uranium-rich fergusonite in the Alavus pegmatite is clearly metamict and gives strongly discordant ages (Table 5). However, the concordant monazite analysis (112A) gives an age of 1863±2 Ma for the Alavus pegmatite (Fig. 7, Table 5).

From the Korvenniemi pegmatite, nine leached and unleached fractions from two columbite grains were analyzed (Table 5). The data scatter on the concordia diagram (Fig. 8) and clearly do not share the common characteristics of the Kaatiala, Seinäjoki, Ullava and Haapaluoma columbites and tapiolites (cf. Table 5). However, all three unleached fractions from a columbite grain with an ordered lattice structure plot on a discordia line with an upper intercept age of 1795±3Ma (MSWD=0.1) (Fig. 8). Further, a six point discordia line gives a similar age of 1793±6 Ma with a slightly higher MSWD-value of 2.2. Three of the data points (3, 5 and 8) plot clearly on the younger side of these discordia lines, most probably indicating the existence of younger uranium-rich phases in microfractures, visible also in BSE images (Fig. 6b).

DISCUSSION

The data presented in this paper demonstrate that crustal evolution in Ostrobothnia followed the general pattern established elsewhere within the Svecofennian domain in Finland (cf. Vaasjoki 1996). Thus the zircons in the Mannilankallio metapelite exhibit the same pattern of bulk analyses as has been found in metapelites in which ion probe data has proven a bimodal Archean-Paleoproterozoic provenance. The age of the synkinematic Luopajärvi tonalite, 1882±9 Ma, is typical for the Svecofennian culmination in central Finland. However, both the monazite (although discordant) from Mannilankallio and the age of the undeformed Luopa quartz monzonite, 1872±2 Ma, indicate that orogenic deformation ceased soon after its culmination. The presence of a contact metamorphic aureole around the Luopa quartz monzonite constitutes particularly strong evidence that regional metamorphism had ceased when this pluton was emplaced. Inhomogeneous pegmatite granites and pegmatites are the youngest rocks in the region. Locations of the isotopically dated rocks of Ostrobothnia are shown in Figure 9.

The age of the Vaasa migmatitic gneisses (formerly the Vaasa granite) has been a major analytical problem for several decades, as all samples taken so far have contained heterogeneous zircon populations. However, the above constraints on the duration of regional metamorphism, the generally granodioritic composition of the neosome and, in particular, the nearly concordant monazite cooling age of 1872±5 Ma from the Kåtakärrskogen sample argue that migmatization in the Vaasa area was syndeformational with the tectono-metamorphic culmination (1.89-1.88 Ga) of the Svecofennian orogeny.

Almost all of the Ostrobothnian RE-pegmatites can be readily divided into the types and subtypes of Černý (1991, 1998). Exceptions are the pegmatites at Kruunupyy-Ullava, Isokyrö and Korvenniemi. Clas-

sification of the Kruunupyy-Ullava RE-pegmatites into albite-spodumene type is based on a somewhat scanty knowledge of these rocks and a future reassessment is possible. The RE-pegmatites at Isokyrö and Korvenniemi are primitive pegmatites of the beryl-columbite subtype, but are distinct – because biotite is the dominant mica – as they contain neither tourmaline nor Fe,Mn-phosphates, but topaz or andalusite, cordierite, and fluorite. In fact, these pegmatites are the only ones without tourmaline in the Ostrobothnia Schist Belt.

Every RE-pegmatite group occurs in the vicinity of pegmatite granite intrusions. It appears that these intrusions are often phacoliths, or flat-lying sheets, almost parallel to the schistosity of the adjacent rocks. Their size is very variable, from some hundreds of metres to a few kilometres, and only the major ones are shown on the lithological map (Fig. 1). The pegmatite granites contain lengthy pegmatitic bodies or veins which are mostly barren, but sometimes beryl±columbite±spodumene-bearing similar to the pegmatites in the vicinity.

The new age data from the pegmatites are particularly interesting against the regional background. According to radiometric dating, the pegmatites fall into two age groups: the NYF family pegmatites, such as the Alavus REE-pegmatites, crystallized at 1.86 Ga and other pegmatites, i.e. the pegmatites of the LCT family, much later at 1.80-1.79 Ga. A temperature-time diagram of the isotopically dated rocks from Ostrobothnia is given in Figure 10. The dated rocks with age results from the study area are summarized in Appendix 1.

Usually the youngest (c. 1.80 Ga) Svecofennian RE-pegmatites, for example the 1.79-1.80 Ga old Gruvdalen REE-pegmatites in Sweden and the Brändö REE-pegmatites in southwestern Finland, occur in the vicinity of granites of approximately the same age (Suominen 1991, Romer and Smeds 1994, Lindroos et al. 1996). In the Seinäjoki RE-pegmatite group, zircons in the rock suite called pegmatite granite appear to be metamictic. So far, no record of intrusive activity coeval with the c. 1.80 pegmatite age group is encountered in the Ostrobothnia region.

The Alavus NYF pegmatites are located both inside and outside of a microcline granite intrusion, 5 km in diameter, in the Central Finland Granitoid Complex. This intrusion contains accessory fluorite and belongs to a suite of post-kinematic rocks (Tyrväinen 1984, Nironen et al. 2000). The age difference between the granite pluton and the NYF pegmatites is about 10 Ma. The position of the pegmatites and the relatively small time gap indicate that the Alavus NYF pegmatites are derived from this granite intrusion.

The Kemiö pegmatites of the LCT family, in southwestern Finland, are practically coeval with the Ostrobothnian LCT pegmatites (see Masuda et al. 1988, Lindroos et al. 1996). Thus, it appears that there is no relationship between the Finnish pegmatites of this age (c. 1.80 Ga, U-Pb age) and the important east-west trending terrane boundary in southern Finland as described by, for example, Vaasjoki and

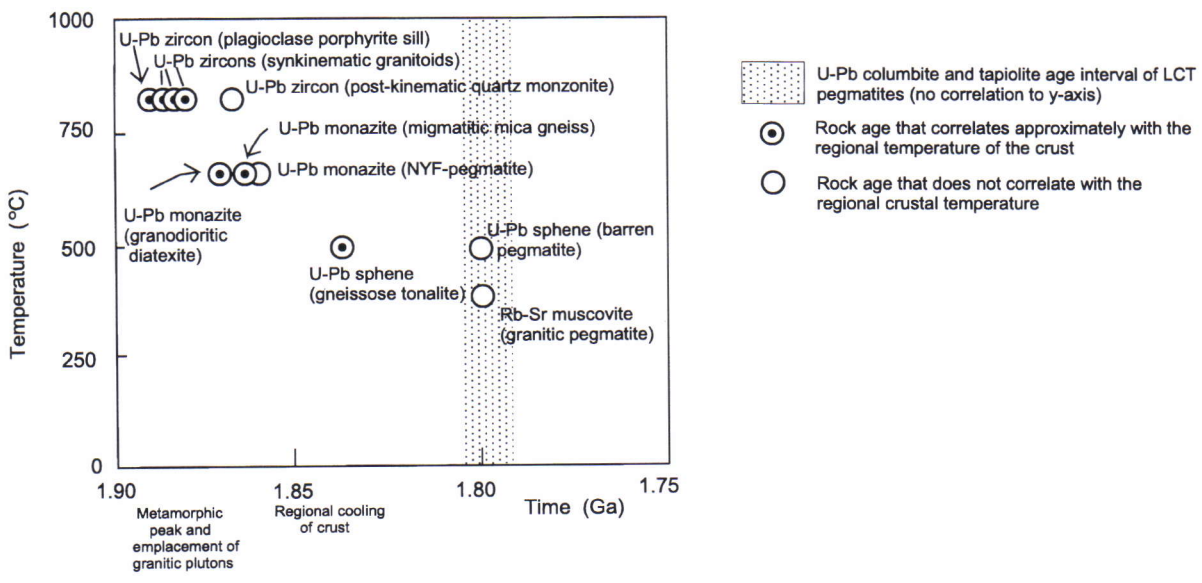

Fig. 10. Temperature-time diagram for isotopic rock ages, Ostrobothnia, western Finland.

Sakko (1988) and Kähkönen et al. (1994).

The closure temperatures of Nb-Ta minerals are not as well known as those for zircon, monazite, and sphene. In addition, comparison of zircon and monazite ages with columbite-tapiolite ages has been difficult, because related zircons and monazites in LCT pegmatites are either lacking or too metamict for accurate dating. However, if it is considered that pegmatite dykes are usually relatively narrow (<100 m) and intruded into cooled country rocks, we can assume that pegmatitic magma cools rapidly. Evidence for coeval crystallization ages for zircon, monazite and sphene have been reported from small granites and related dykes (e.g. Suominen 1991) and even monazites in the vast Wiborg batholith register identical ages with zircons of the same samples (Vaasjoki et al. 1991). More importantly, Romer and Wright (1992) report that the Vassijaure area, where U-Pb ages from columbite and zircon in pegmatite from the genetically related Vassijaure granite coincide at c. 1780 Ma, was subjected to lower amphibolite facies metamorphism during the Caledonian orogeny c. 400 Ma ago. We thus consider it safe to assume that columbite U-Pb ages from pegmatites date the emplacement of these rocks and have not been reset by subsequent geological processes.

In Central Finland, large scale ductile deformation ceased at the latest 1850 Ma ago (Vaasjoki 1996). However, many schist-hosted RE-pegmatites are deformed and folded in both the Seinäjoki and Haapaluoma-Kaatiala areas. Most of these pegmatites occur in prominent shear zones shown on large scale geological maps (e.g. Korsman et al. 1997). Clearly, the shear zones have been active later than 1.80 Ga ago.

Most of the LCT pegmatites occur in relatively low-grade metamorphic rocks such as andalusite mica schists, but a few, compositionally unique LCT pegmatites lie in high-grade metamorphic mica gneisses. Černý (1998) has shown that none of the metamorphogenic hypotheses, such as derivation from anatectic melts, can be used to explain the attributes of RE-pegmatites and, thus, magmatogenic derivation of RE-pegmatites should be the only model supported. However, in the study area, there is a strong correlation between the LCT type pegmatites and relatively low-grade metamorphic country rocks (see also Mäkitie et al., this volume) (Fig. 1). This indicates that the metamorphic zoning has had some control on the accumulation of the pegmatite melts, although the time gap between the metamorphic peak to the crystallization of the pegmatites is about 80 Ma.

All the 12 pegmatite groups studied have individual features separating them from each other. This indicates that each group of pegmatites is a product of the fractionation of local melts. It appears likely that the c. 1.80 Ga pegmatites occurring in the Ostrobothnia represent a distinct tectono-magmatic episode separate from the Svecofennian orogenic culmination (1.89-1.88 Ga) and from the migmatite forming microcline granite event (c. 1.83 Ga) in southern Finland.

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Appendix 1. Summary of the published isotopic age data from Ostrobothnia, western Finland. Abbreviations are explained in Table 3.

ROCK TYPE (dated mineral is in parenthesis)	AGE RESULT	LOCATION/ GEOLOGICAL INTERPRETATION	CLOSURE °C	REFERENCE
Albite-spodumene pegmatite (Clb)	1789 ± 2 Ma	Ullava (R5)/ crystallization		This paper
Columbite-topaz pegmatite (Clb)	1795 ± 5 Ma	Korvenniemi/ crystallization		This paper
Spodumene-lepidolite pegmatite (Clb)	1797 ± 2 Ma	Haapaluoma/ crystallization		This paper
Pegmatite granite (Ms)	c. 1.8 Ga	Kälviä/ crystallization	c. 350	Wetherill et al. 1962
Barren pegmatite (Ttn)	1800 ± 4 Ma	Isotkalliot/ crystallization	c. 500	This paper
Cassiterite-tapiolite pegmatite (Tap)	1802 ± 2 Ma	Seinäjoki/ crystallization		This paper
Spodumene-columbite pegmatite (Clb)	1804 ± 2 Ma	Kaatiala/ crystallization		This paper
Gneissose tonalite (Ttn)	1838 ± 5 Ma	Evijärvi/ cooling of crust	c. 500	Vaasjoki et al. 1996
Garnet-cordierite-sillimanite mica gneiss (Mnz)	≥1.86 Ga	Mannilankallio/ cooling of crust	c. 600	This paper
Fergusonite-monazite pegmatite (Mnz)	1863 ± 2 Ma	Alavus/ crystallization	c. 600	This paper
Granodioritic rock from the Vaasa Migmatite Complex (Mnz)	1872 ± 5 Ma	Kåtakärrskogen/ cooling of crust	c. 600	This paper
Quartz monzonite (Zr)	1872 ± 2 Ma	Luopa/ crystallization	c. 800	Mäkitie & Lahti, this volume
Tonalite (Zr)	1882 ± 9 Ma	Luopajärvi/ crystallization & metamorphic peak	c. 800	This paper
Gneissose tonalite (Zr)	1883 ± 6 Ma	Evijärvi/ crystallization & metamorphic peak	c. 800	Vaasjoki et al. 1996
Granite (Zr)	1886 ± 10 Ma	Ähtäri/ crystallization	c. 800	Huhma 1986
Gneissose granodiorite (Zr)	c. 1.89 Ga	Vimpeli/ crystallization	c. 800	Kouvo & Tilton 1966
Plagioclase porphyrite sill (Zr)	c. 1.9 Ga	Routakalliot/ emplacement	c. 800	This paper
Garnet-cordierite-sillimanite mica gneiss (Zr)	c. 2.4 Ga	Mannilankallio/ mixture of Archaean and Proterozoic grains	c. 800	This paper

Svecofennian granitic pegmatites (1.86-1.79 Ga) and quartz monzonite (1.87 Ga),
and their metamorphic environment in the Seinäjoki region, western Finland
Edited by Hannu Mäkitie
Geological Survey of Finland, Special Paper 30, 31–59.

COMPOSITIONAL VARIATION OF GRANITIC PEGMATITES IN RELATION TO REGIONAL METAMORPHISM IN THE SEINÄJOKI REGION, WESTERN FINLAND

by
Mäkitie, H., Kärkkäinen, N., Lahti, S.I. and Alviola, R.

Mäkitie, H., Kärkkäinen, N., Lahti, S. I. & Alviola, R. 2001. Compositional variation of granitic pegmatites in relation to regional metamorphism in the Seinäjoki region, western Finland. *Geological Survey of Finland, Special Paper 30*, 31–59, 12 figures, 5 tables and two appendices.

Granite pegmatite dykes and pegmatitic granite intrusions are common in the Palaeoproterozoic Svecofennian (1.9-1.8 Ga) schists of the Seinäjoki region, western Finland. The dykes often occur as swarms composed of barren and LCT (lithium-caesium-tantalum)-type rare-element pegmatites. Moderately fractionated beryl-columbite and beryl-columbite-phosphate pegmatites, and especially the well fractionated lithium mineral -bearing complex pegmatites, are found in the andalusite mica schists which form an east-west trending zone in the Seinäjoki region. The metamorphic grade gradually increases northwards and southwards from the andalusite zone trough andalusite-sillimanite grade up to granulite grade. Some complex pegmatites occur in the muscovite-sillimanite gneisses. The rare-element pegmatites rapidly disappear between the first and second sillimanite isograds, and are almost absent in the high-grade metamorphic zones.

In particular the concentrations of Li, Cs, Ta, Sn and P are highest in the pegmatites of the low-grade metamorphic zones. Supracrustal country rocks of the pegmatites are also enriched in Cs, Ta and Rb. Generally Sr/Rb, Ba/Rb and K/Rb decrease in the pegmatites as a function of the metamorphic grade of country rock. However, similarly fractionated pegmatites (in terms of their Sr/Rb and K/Rb) of different metamorphic environments and zones have different element concentrations: e.g. P and Ta are high in the andalusite zone, but Fe and Th are enriched in pegmatites of the high-grade metamorphic zones.

The pegmatites are post-kinematic in regard to the main stage of deformation in the region, but deformed dykes also occur. The pegmatites are clearly younger (ca. 1.80 Ga) than the regional metamorphic peak (1.89-1.88 Ga). The pegmatitic granites and pegmatites probably derived from anatectic melts – from northern and southern high-grade migmatite areas – intruded upwards in the crust. The rare-element pegmatites crystallized from the residual fractions enriched in volatiles and other elements at a higher crustal level where the grade of metamorphism was lower.

Key words (GeoRef Thesaurus, AGI): pegmatite, chemical composition, mineral composition, metamorphism, metapelite, zoning, Seinäjoki, Paleoproterozoic, Finland

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CONTENTS

INTRODUCTION

Numerous barren and rare-element pegmatite dykes of LCT (lithium-caesium-tantalum) type and pegmatitic granite intrusions are found in the Palaeoproterozoic Svecofennian (1.9-1.8 Ga) bedrock of the Seinäjoki region, South Pohjanmaa, western Finland (Fig. 1). The Haapaluoma and Kaatiala pegmatites, 30 km SE of the town of Seinäjoki, are richest in rare minerals such as beryl, columbite-tantalite, spodumene and lepidolite (Haapala 1966, Nieminen 1978, Lahti et al. 1982, Lahti and Saikkonen 1986, Teertstra et al. 1993). These pegmatites produced feldspar and quartz for the glass industry for many years (e.g. Alviola 1989).

The discovery of a cassiterite pegmatite at Pajuluoma, near Seinäjoki, prompted pegmatite studies by the Geological Survey of Finland in the 1980's and led to the discovery of numerous new rare-element pegmatites, some of them rich in tin but too small to warrant exploitation (Oivanen 1983, Alviola 1986a, Kärkkäinen 1993). Rare-element pegmatites differ from barren pegmatites by their trace element (Nb, Ta, Be, Sn, Li etc) and mineral (columbite, cassiterite, beryl, spodumene etc) contents (e.g. Černý 1991a).

The rare-element pegmatites of the Seinäjoki region are closely related to andalusite mica schists and muscovite-sillimanite gneisses of low-pressure, lower to middle amphibolite-facies metamorphism (Fig. 2). There is a correlation between the distribution of the granitic pegmatites and rare-element pegmatites in relation to the regional metamorphic grade of country rocks: rare minerals disappear from the pegmatites when the metamorphic grade of the country rock reaches P-T conditions between the first and second sillimanite isograd (Mäkitie et al. 1998). Only topaz pegmatites with columbite as a rare accessory mineral are encountered in high-grade mica gneisses, and are exceptions to the special relationship between the rare-element pegmatites and the relatively low-grade metamorphic areas.

This paper has emphasis on the compositional variation of pegmatites within the different metamorphic zones of the Seinäjoki region. Generally chemical composition correlates with the mineral composition (mode) of pegmatites. It is often difficult, however, to distinguish with the naked eye the poorly or moderately fractionated pegmatites (containing rare minerals with important trace elements such as Nb, Ta and Li) from barren pegmatites. Moreover, feldspars and micas in the less fractionated pegmatites contain trace elements such as Rb and Li. Chemical whole rock analysis of pegmatites is a useful method to reveal the concentration of these elements.

The metamorphic environment and mineralogy of the pegmatites is described in rather detail because the study area includes metamorphic zoning and high-grade metamorphic terranes, and because the mineralogy of the pegmatites not only indicates the chemical composition of the rock, but also it reflects the differences in crystallization conditions such as lithospheric pressure.

Compositional variations of pegmatites are usually explained by fractionation processes and by different sources of pegmatite melt (e.g. Černý 1991b). However, many other reasons have been reported to cause the compositional variation in pegmatites. For example, according to Jolliff et al. (1992) additional compositional variations among samples from wall zones of individual pegmatites in the Black Hills area, USA, were caused by local interaction with host metamorphic rocks and fractionation during internal evolution of the pegmatites as they crystallized. Moreover, Shearer et al. (1992) emphazise that in addition to fractional crystallization of pegmatites in the Black Hills area, partial melting appears to be important in controlling the composition of the volatile component and the incompatible element character of a distinct magma-type.

In terms of 'whole evolution' of pegmatites, Breaks and Moore (1992) discussed that the genesis of a peraluminous granite - rare-element pegmatite association in the Dryden area, Canada, can be explained by a complex interplay of petrogenetic processes that involves five stages. The first stage is clastic sedimentation into large basins followed by high-grade metamorphism and initial migmatization, and the last stage includes dispersion of rare-element-enriched melt/fluid masses along host rock anisotropies, and is associated by interaction of fluids derived from pegmatites with host rocks.

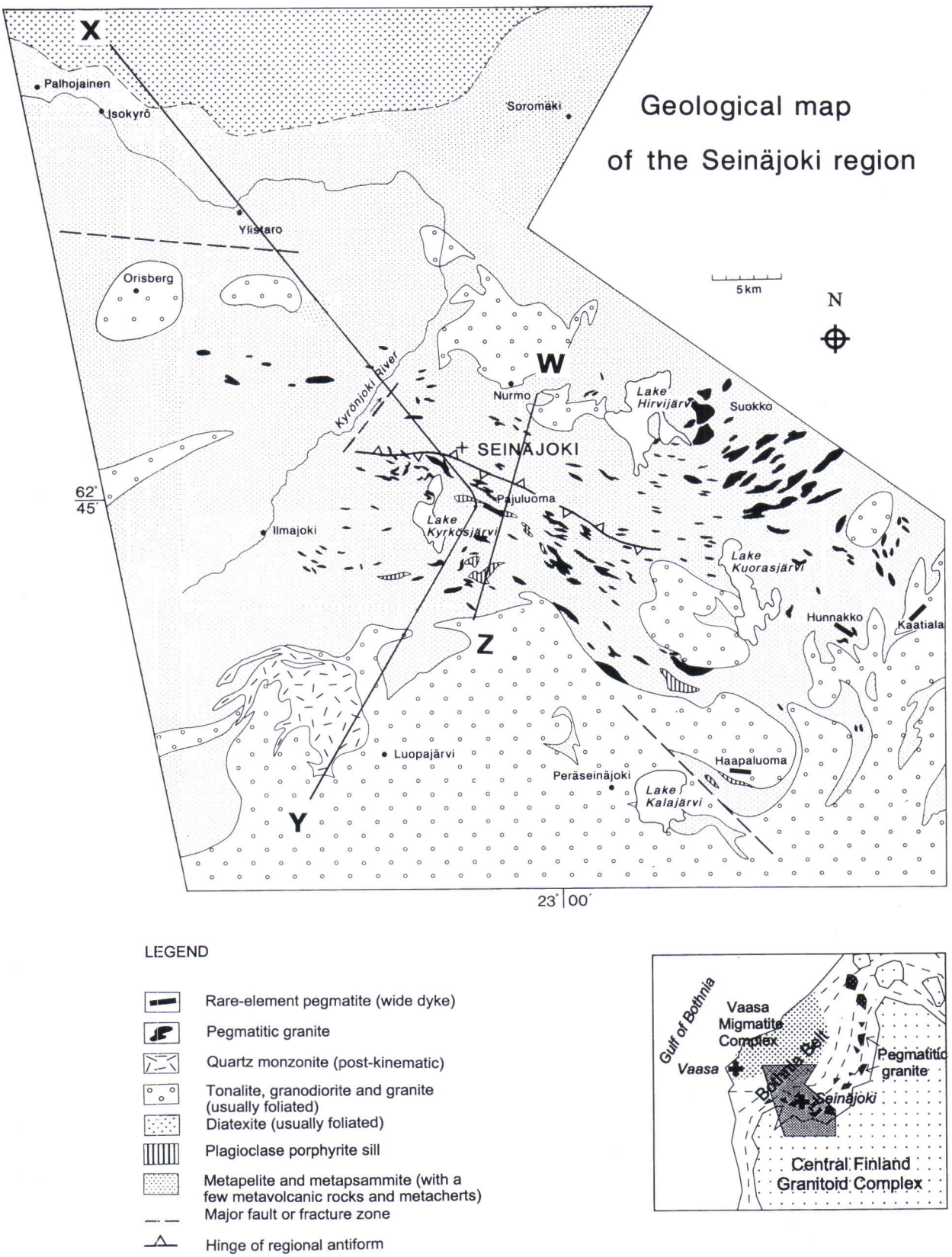


Fig. 1. Simplified geological map of the Seinäjoki region (modified after Tyrväinen 1970, 1971, Lahti and Mäkitie 1990 and Mäkitie et al. 1991). X-Y and Z-W lines on the map refer to cross-section profiles.

GENERAL GEOLOGICAL SETTING

The Seinäjoki region is a part of the Palaeoproterozoic Bothnia Belt (BB) of western Finland. This belt is about 80 km wide and curved in shape, and is located roughly in the middle of the Fennoscandian (Baltic) shield. The Bothnia Belt lies on the western side of the large Central Finland Granitoid Complex (CFGC), and includes the inhomogeneous Vaasa Migmatite Complex (VMC) (Fig. 1) (also known as the Vaasa granite or the Vaasa Granitoid Complex; Laitakari 1942, Mäkitie et al. 1999). All rocks of the study area belong to the Svecofennian accretionary arc complex (1.90-1.87 Ga) of central and western Finland, which is part of the Svecofennian domain (1.93-1.82 Ga) (Korsman et al. 1997, see also Nironen 1997). The rocks of the Seinäjoki region underwent regional metamorphic peak 1.89-1.88 Ga ago (Vaasjoki 1996).

The predominant rocks in the Seinäjoki region are porphyroblastic muscovite-biotite schists and biotite gneisses with psammitic intercalations (Mäkitie et al. 1991). A few tholeiitic metabasalts, metacherts with thin iron formations, ca. 1.9 Ga old plagioclase porphyrite sills and several Sb and Au mineralizations closely related to the metavolcanic rocks are also found in the region (Pääkkönen 1966, Borodaev et al. 1983, Mäkitie and Lahti 1991, Appelqvist 1993).

The most common plutonic rocks in the Seinäjoki region, especially in the south, are synkinematic foliated tonalites and granodiorites of the CFGC (Fig. 1; Tyrväinen 1970, 1971, 1984; Lahti and Mäkitie 1990; Mäkitie et al. 1991). Geochemically they are I-type rocks and have molar A/CNK in the range of 1.00-1.04 (Mäkitie et al. 1999). The U-Pb zircon age of tonalites is 1882±9 Ma (Mäkitie and Lahti 1991). The VMC is composed of porphyritic or even-grained granodioritic diatexites that typically contain remnants of supracrustal rocks and garnet. The A/CNK value of the VMC range from 1.11 to 1.32, and the rocks have also other features typical of S-type granites (Mäkitie et al. 1999). The contact between the VMC and the garnet-cordierite mica gneisses of the BB is gradual. The post-kinematic (1872±2 Ma), metaluminous Luopa quartz monzonite is located in the southwestern corner of the study area, at the contact between the BB and the CFGC (Mäkitie and Lahti, this volume).

Lensoid or irregular pegmatitic granite bodies and pegmatite dykes, usually following the main schistosity are common in the Seinäjoki region (Mäkitie et al. 1991). Here the term pegmatitic granite is used if it is an intrusive body tens to hundreds of metres wide and is composed of minerals with grain size > 3-5 cm. The pegmatitic granite intrusions are usually inhomogeneous and locally contain accessory beryl, fluorapatite and even tiny plates of columbite.

The pegmatites, in contrast, are zoned tabular dykes or lenses. Pegmatite dykes are found predominantly in schists, and their number is highest around the pegmatitic granite intrusions. Several generations of intraformational pegmatite dykes may also occur in the pegmatitic granite intrusions. The pegmatitic granites and pegmatite dykes cross-cut all the other rocks with sharp contacts. Previously, ca. 1.8 Ga U-Pb age has been reported for xenotime from the Haapaluoma pegmatite (Haapala 1966) and 1800±4 Ma for sphene from a pegmatite-rimmed quartz vein near the Seinäjoki town (Mäkitie and Lahti 1991). New U-Pb datings show that the LCT-type pegmatites of the Seinäjoki region are ca. 1.80 Ga old (Alviola et al., this volume).

TECTONIC AND METAMORPHIC ENVIRONMENT

Tectonic environment

The tectono-metamorphic evolution was complex in the Seinäjoki region. The structure of the central part of the region is dominated by an approximately east-west-trending antiform (Fig. 1) (Pääkkönen 1966): on the southern side of Seinäjoki the foliations of metapelites usually dip southwards, but on the northern side northwards. Major part of the granitic pegmatites lie in the hinge area of the antiform. The andalusite mica schists of this area form isoclinally folded (F_2) thick beds, with local kink folding (Fig. 3a) (Mäkitie 1999).

Field studies show that the pegmatites intersect synmetamorphic foliation (S_2) (Mäkitie 1999). Deformed pegmatites are common in the Seinäjoki region; folding in pegmatite dykes near the town of Seinäjoki is due to late deformation (D_5) (see also

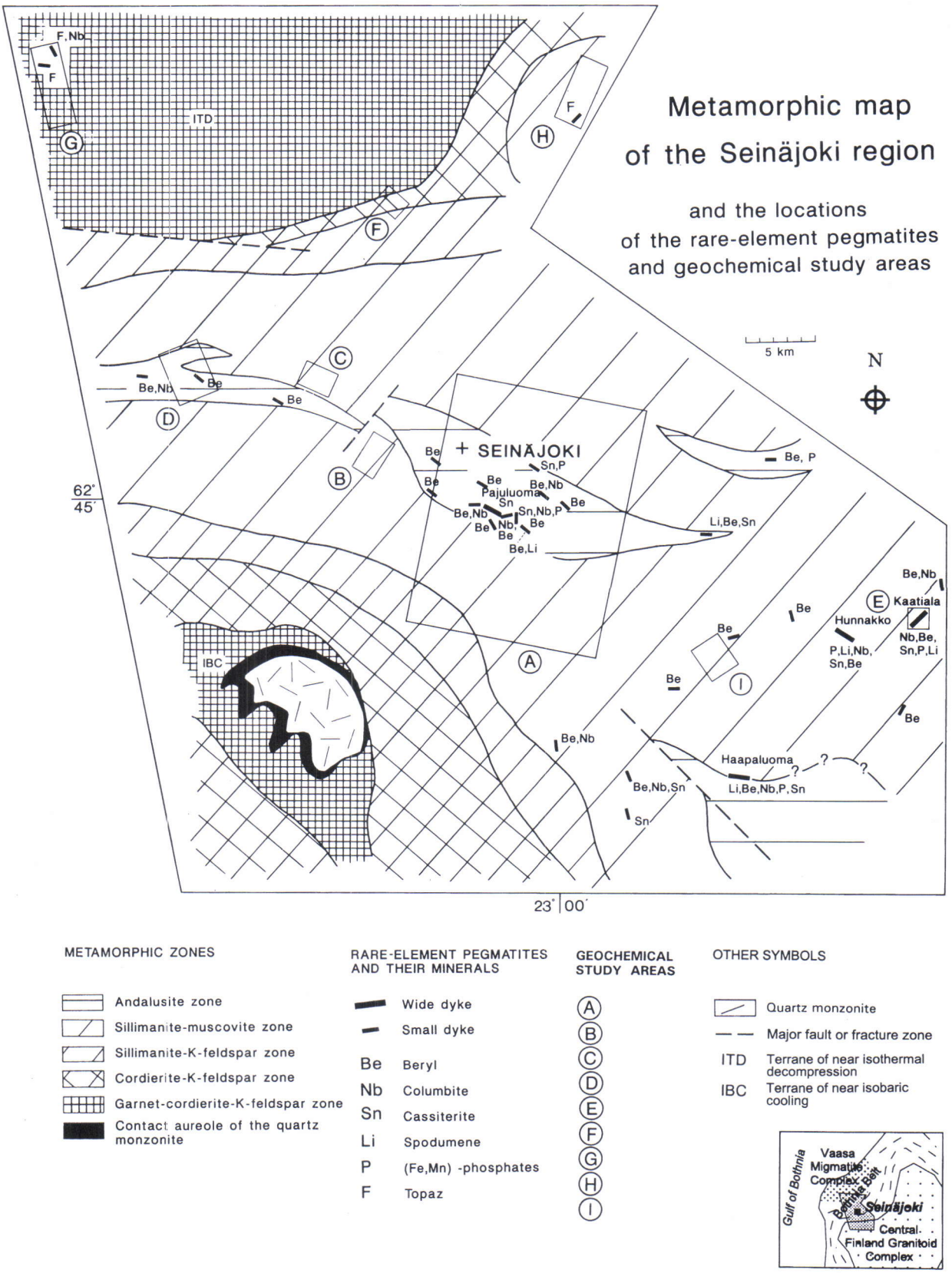


Fig. 2. Metamorphic map of the Seinäjoki region with locations of most important rare-element pegmatites. Rectangular areas marked with letters A-I refer to geochemical study areas of granitic pegmatites (modified after Tyrväinen 1970, 1971, 1984, Lahti and Mäkitie 1990, Mäkitie and Lahti 1991 and Mäkitie et al. 1991).

Mäkitie 1999, Alviola et al., this volume). The pegmatites in the mica schists are mainly oriented conformably with the general east-west-trending schistosity, but cross-cutting dykes are more common in the more competent plagioclase porphyrites. The Haapaluoma pegmatite sharply intersects the northwest-southeast-trending lineation in the country rocks (Haapala 1966).

Faults and shear zones, some of them several kilometres in length, are characteristic in the Seinäjoki region (Korsakova 1982, Mäkitie et al. 1991, Korsman et al. 1997); most significant faults are shown in Figure 1. The great fault trending northwest-southeast near Peräseinäjoki has caused a tectonic unconformity and dominates the strikes of the area. A notable shear zone trending north-south intersects the general strike of rocks in the southeastern corner of the study region (Tyrväinen 1984). The Haapaluoma, Pajuluoma and Kaatiala rare-element pegmatites occur close to or at the continuations of these lineaments. The deformed Pajuluoma cassiterite pegmatite lies roughly at a lithological boundary that may be partly tectonic (Fig. 4).

Ca. 10 kilometres northwest of Seinäjoki town, there is a notable geotectonic lineament striking 100° (Fig. 1; cf. Papunen and Gordunov 1985). The lineament contains schollen migmatites and coincides with a negative gravity anomaly. Moreover, it lies in relatively high-grade metamorphic rocks and no rare-element pegmatites are found near it.

Metamorphic zones

In the Seinäjoki region the metamorphic grade is lowest in the elongated east-west-trending andalusite schist zone where well-preserved sedimentary structures, rare-element pegmatites and pegmatitic granites are most common (Fig. 2). The metamorphic grade increases northwards and southwestwards to upper amphibolite and granulite facies (Mäkitie 1990, 1999).

The following regional metamorphic zones have been identified: andalusite, sillimanite-muscovite, sillimanite-K-feldspar, cordierite-K-feldspar and garnet-cordierite-K-feldspar (Fig. 2). The southern garnet-cordierite-K-feldspar zone is partly overprinted by a granulite-facies contact aureole of the Luopa quartz monzonite (Mäkitie and Lahti 1991). Mineral assemblages typical in metapelites of the zones are listed in Table 1. For references to the metamorphic reactions presented here, see Spear (1993).

Andalusite zone

The muscovite-biotite schists in the andalusite zone contain andalusite porphyroblasts 1-3 cm in diameter (Fig. 3a). Tourmaline is a common accessory mineral in the schists. Staurolite and garnet are found locally in well-preserved sedimentary layers. Chlorite-staurolite-muscovite-biotite schists occur sporadically. The schists also contain narrow discordant quartz veins and psammitic and tuffitic intercalations. The psammitic layers are usually fine grained, but there are also some beds of medium-grained metagreywackes with clasts.

Sillimanite-muscovite zone

The appearance of sillimanite and increase in the grain size of the metapelites mark the beginning of the sillimanite-muscovite zone. The main minerals of mica gneisses are quartz (30-40%), plagioclase (20-30%), biotite (10-30%), muscovite (2-15%), sillimanite (1-5%) and opaque (0-2%). Tourmaline is rare and K-feldspar is a sporadic component of the mica gneisses. Fibrolitic sillimanite is found in coarse-grained muscovite flakes and in their quartz inclusions. Knots of sillimanite-quartz-muscovite are locally characteristic features (Fig. 3b). The grain size of mica ranges from 1 to 3 mm and increases southwestwards. In places the mica gneisses are spotted by muscovite flakes and migmatitic due to the presence of narrow, intrusive pegmatite dykes. Tight folds with amplitudes of up to 1 m are common as are psammitic intercalations. Pegmatitic granite, granite dykes and veins in the zone contain visible tourmaline. These intrusive rocks cause that the mica gneisses locally are migmatitic in appearance.

Abbreviations for mineral names used in this article follow predominantly the usage of Kretz (1983): Ab (normative albite), Alb (albite), An (normative anorthite), And (andalusite), Bt (biotite), Crd (cordierite), Ecr (eucryptite), Grt (garnet), Kfs (potassium feldspar), Ms (muscovite), Or (normative orthoclase), Ptl (petalite), Q (normative quartz), Qtz (quartz), Sil (sillimanite), Spd (spodumene) and St (staurolite).

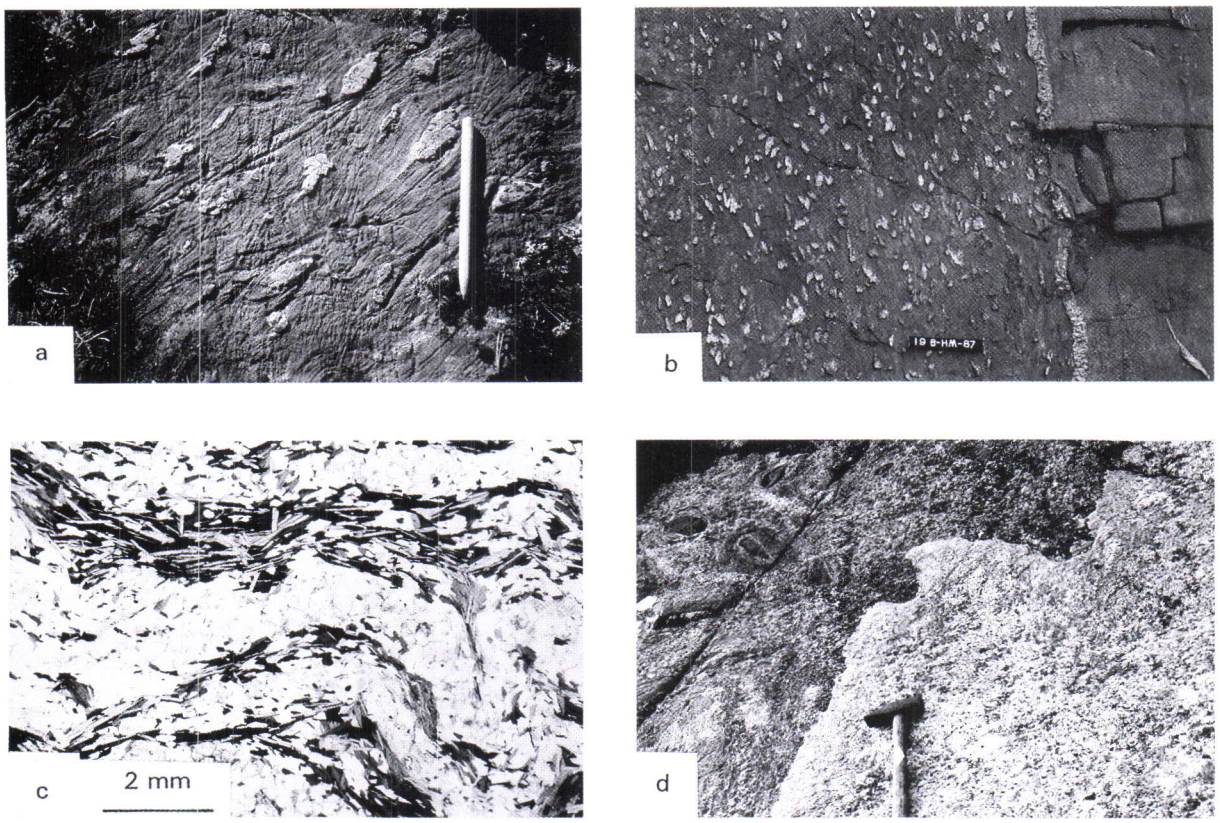


Fig. 3. Metapelites of the Seinäjoki region. **(a)** Porphyroblastic, folded (F_5 kinks seen as vertical stripes in the outcrop) andalusite mica schist, a common country rock of the rare-element pegmatites in the Pajuluoma pegmatite group. Andalusite zone. Length of pen 14 cm. **(b)** Layered mica gneiss containing knots of intergrown sillimanite, muscovite and quartz. Metapelite from the area (8 km northwest of the Haapaluoma complex pegmatite) where rare-element pegmatites disappear. Length of scale bar 12 cm. Southern Sil-Ms zone. **(c)** Microtexture of a veined muscovite-biotite gneiss 2 km southeast of the Kaatiala complex pegmatite. Muscovite (elongated light-coloured minerals) is a progressive mineral. Southern Sil-Ms zone. **(d)** Contact between the gently dipping topaz-bearing pegmatite at Palhojainen and a diatexitic garnet mica gneiss with remnants of psammitic layers including concretions. Length of hammer 30 cm. Northern Grt-Crd-Kfs zone.

$$\text{Ms} + \text{Alb} + \text{Qtz} = \text{K-richer Ms} + \text{K-richer Alb} + \text{Sil} + H_2O \quad (1)$$

$$\text{Ms} + H^+ = \text{Sil} + K^+ + H_2O + SiO_2 \quad (2)$$

$$\text{Bt} + H^+ = \text{Sil} + K^+ + (Mg,Fe)^{2+} + H_2O + SiO_2 \quad (3)$$

Sillimanite probably mostly formed by reaction (1), because fibrolitic sillimanite occurs in muscovite (see Thompson and Thompson 1976, Tracy 1978); only very rarely does sillimanite overgrow andalusite ideally in reaction And = Sil. Some sillimanite can also be formed by metasomatic reactions (2) and (3) (see Vernon 1979), because fibrolitic sillimanite is also detected in biotite. These three reactions refer to the first sillimanite isograd. A common reaction Ms + St = Bt + Sil + Qtz + H_2O is unlikely in the study area because staurolite is rare in the relative low grade biotite-muscovite schists.

In the eastern parts of the Seinäjoki region psammitic rocks dominate, but the metapelite

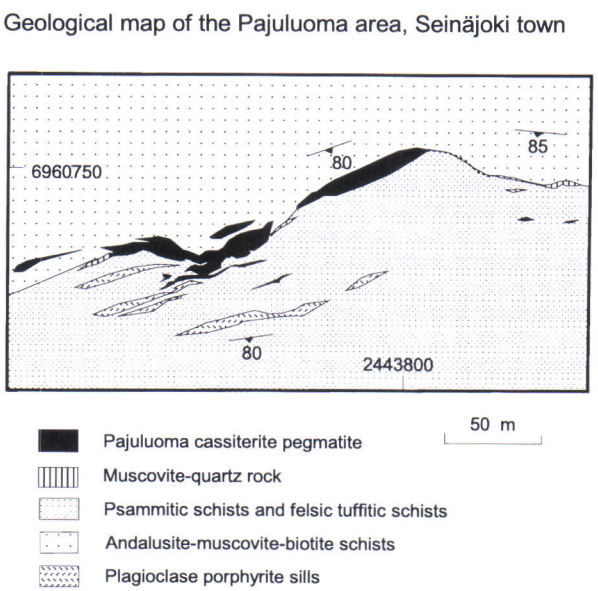


Fig. 4. The form of the cassiterite-rich pegmatite dyke of Pajuluoma (modified after Oivanen 1983).

Table 1. Mineral occurrences in metapelites of the southern and northern metamorphic zonations in the Seinäjoki region. Both zonations start in the same andalusite zone (see also Fig. 2). + = common mineral, (+) = rare mineral, * = retrogressive origin.

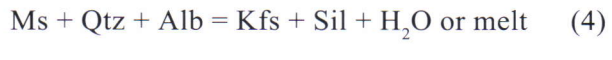
	Andalusite zone	Sillimanite-muscovite zone	Sillimanite-K-felspar zone	Cordierite-K-feldspar zone	Garnet-cordierite-K-feldspar zone	Contact aureole of the quartz monzonite
Southern zonation:						
Quartz	+	+	+	+	+	+
Plagioclase	+	+	+	+	+	+
K-feldspar		(+)	+	+	+	+
Tourmaline	+	(+)				
Muscovite	+	+				
Biotite	+	+	+	+	+	+
Andalusite	+					+*
Sillimanite		+	+	+	+	(+)
Kyanite					(+)*	
Staurolite	(+)					(+)*
Cordierite				+	+	+
Garnet	(+)		(+)	+	+	+
Hercynite						(+)
Orthopyroxene						(+)
Northern zonation:						
Quartz	+	+	+	+	+	
Plagioclase	+	+	+	+	+	
K-feldspar		(+)	+	+	+	
Tourmaline	+	(+)				
Muscovite	+	+			+*	
Biotite	+	+	+	+	+	
Andalusite	+				+*	
Sillimanite		+	+	+	(+)	
Staurolite	(+)					
Cordierite				+	+	
Garnet	(+)		+	+	+	
Hercynite					(+)	

intercalations contain discordant muscovite porphyroblasts (diameter 1-4 mm) with fibrolitic sillimanite. Muscovite-rich knots near the Hunnakko pegmatite are probably alteration products of andalusite (Haapala 1966). Near the Kaatiala dyke, the metapelites are veined sillimanite-muscovite-biotite gneisses in which muscovite is progressive (Fig. 3c).

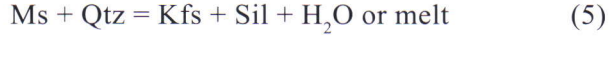
Sillimanite-K-feldspar zone

The mineral pair sillimanite-K-feldspar without muscovite is the index of the zone. The mineral assemblage of mica gneisses consists of quartz (20-30%), plagioclase (30-40%), biotite (20-30%), K-feldspar (1-10%) and sillimanite (1-10%). K-feldspar porphyroblasts occur as small (1-2 mm) white spots whereas the grain size of mica is 0.5-2.0 mm. Psammitic intercalations are common. The migmatitic texture in the gneisses is not as distinct as in the gneisses of the sillimanite-muscovite zone due to smaller amount of pegmatite dykes. Rare anatectic granite patches contain some garnet. In the northern zone some garnets have been decomposed into biotite, plagioclase and quartz.

K-feldspar was formed by reactions (4) and (5), with simultaneous disappearance of muscovite. The mineral modes in the metapelites are compatible with these reactions, which refer to the second sillimanite isograd.



$$Ms + Qtz + Alb = Kfs + Sil + H_2O \text{ or melt} \quad (4)$$



$$Ms + Qtz = Kfs + Sil + H_2O \text{ or melt} \quad (5)$$

Cordierite-K-feldspar zone

The main minerals of these mica gneisses are plagioclase, quartz, K-feldspar, biotite, cordierite, garnet and sillimanite. Cordierite and garnet have not been found in contact with each other. South of Seinäjoki the garnet/cordierite ratio and grade of migmatization increases southwestwards. Garnet and cordierite occur in the anatectic patches as

well in the paleosomes. However, the gneisses are not intensely migmatized. The zone contains some tourmaline free pegmatitic granites.

Most of the cordierite was formed by reaction (6), because the initial cordierite porphyroblasts which include small biotite grains occur only in the sillimanite-biotite gneisses. Garnet probably crystallized by reaction (7), because the modes of garnet and K-feldspar start to increase simultaneously. Reaction (7) needs a slightly higher temperature than reaction (6) (Miyashiro 1994). The boundary between the Sil-Kfs and Crd-Kfs zones, 15 km southwest of Seinäjoki, is rather sharp.

$$\text{Bt} + \text{Sil} + \text{Qtz} = \text{Crd} + \text{Kfs} + \text{melt} \quad (6)$$

$$\text{Bt} + \text{Sil} + \text{Qtz} = \text{Grt} + \text{Kfs} + \text{melt} \quad (7)$$

$$\text{Bt} + \text{Sil} + \text{Qtz} = \text{Crd} + \text{Grt} + \text{Kfs} + \text{melt} \quad (8)$$

Garnet-cordierite-K-feldspar zone

The pelitic rocks of the southern zone are migmatitic garnet-cordierite-sillimanite mica gneisses. Coexisting garnet-cordierite is the index of the zone but the minerals have seldom crystallized in equilibrium. A typical feature is the occurrence of garnet surrounding cordierite. Sillimanite is a common mineral and retrograde kyanite is occasionally encountered.

Generally the migmatization of metapelites is not intensive (Mäkitie 1999). Anatectic material occurs as patches, and more seldom as narrow elongated veins. The psammitic intercalations contain narrow granitic veins, but portions between psammitic intercalations are often composed of relatively homogeneous rocks showing only weak schistosity and small anatectic patches with garnet and cordierite. Pegmatites are rare in the zone. Coexisting garnet and cordierite – which do not rim each others – are formed by reaction (8).

The rocks have passed through granulite facies metamorphism because orthopyroxene has crystallized in the intermediate tuffitic schists, more than 1 km from the quartz monzonite. Orthopyroxenes are breakdown products of hornblende.

The main minerals in the pelitic rocks of the northern Grt-Crd-Kfs zone are garnet, cordierite, biotite, plagioclase, K-feldspar and quartz. Accessory minerals are sillimanite, hercynite, retrogressive andalusite, retrogressive muscovite, sericite and pinite. Generally the amount of cordierite decreases northwards due to decomposition. The migmatitic nature of mica gneisses also decreases northwards while the heterogeneous plutonic textures begins to dominate (Fig. 3d); the contact with the VMC is gradual and grades from metatexites to diatexites (Mäkitie et al. 1999, see also Sawyer 1995). The northern Grt-Crd-Kfs zone contains also a few orthopyroxene-bearing gneisses. These rocks refer to P-T conditions of granulite facies.

In the northern mica gneisses K-feldspar contains retrograde sericite and muscovite. Garnet is often surrounded by cordierite and sillimanite occurs as needles in the cordierite which is altered to pinite, sericite, andalusite and biotite. Cordierite pseudomorphs occasionally include hercynite. Garnet has often been decomposed into biotite, plagioclase and quartz. Retrogressive features become more common northwards. These microtextures imply that reactions (6) and (7) propagated in reverse directions.

Primary sedimentary structures in the northern Grt-Crd-Kfs zone have been destroyed (due to intense melting) more intensively than those in the mica gneisses of the southern high-grade terrane. The southern boundary of the high-grade terrane partly follows the east-west-trending shear zone (Fig. 2).

Contact aureole of the Luopa quartz monzonite

The post-kinematic (1872±2 Ma, Mäkitie and Lahti, this volume) fayalite-augite quartz monzonite of Luopa has produced an overprinting contact metamorphic aureole, which includes a narrow (~2 m) granulite grade inner zone and a wider (up to 30 m) outer zone characterized by upper amphibolite facies. The inner zone contains orthopyroxene-garnet pairs, retrogressive andalusite and symplectites composed of hercynite, sillimanite and cordierite. The metapelites in the contact aureole are locally diatexites due to intense melting, but locally tightly banded due to diapiric deformation.

Outlines of metamorphic evolution

The andalusite-muscovite-biotite schists and, locally, chlorite-staurolite mica schists indicate metamorphic P-T conditions of low pressure (1-3 kbar), lower amphibolite facies. Sillimanite-muscovite zone refers to bathozone 2 and cordierite-K-feldspar zone implies P-T conditions of ca. 600 °C, 3 kbar (see Bucher and Frey 1994). Thus the metamorphic zoning represents the andalusite-sillimanite series.

The metapelites of the southern Grt-Crd-Kfs zone have not been intensively migmatized. This with the fact that U-Pb isotopic ratios of zircon in the same rock have not been reset during regional metamorphism suggest that regional metamorphic temperature was below ca. 750 °C (Mäkitie and Lahti 1991). Unfortunately P_{H20} which has a notable role in anatectic reactions is not known. Thermometric determinations on garnet-cordierite pairs of the area suggest metamorphic temperature of at least 700 °C (Mäkitie 1999).

The peak of regional metamorphism in the southern high-grade terrane was obviously associated with emplacement of the abundant ca. 1.88 Ga tonalites, but about 20 km south of Seinäjoki the crust was also locally heated during the emplacement of the 1.87 Ga Luopa quartz monzonite. Monazite separated from partially melted metapelite of the southern Grt-Crd-Kfs zone gave a U-Pb age of ca. 1.86 Ga which refers to cooling of the crust to below ca. 650 °C (Mäkitie and Lahti 1991). The dated metapelite is relatively close (~1.5 km) to the quartz monzonite, and therefore the age may refer to cooling of the quartz monzonite contact aureole.

The southern and northern high-grade terranes probably proceeded along different P-T paths and metapelites of both areas contain different textures. In the southern high-grade terrane, garnet rims cordierite and rare retrograde kyanite is detected, and the absence of retrogressive andalusite and muscovitization is typical. On the contrary, the garnet-cordierite mica gneisses of the northern high-grade terrane include retrogressive andalusite and muscovitization. Moreover, cordierite surrounds garnet, and marginal parts of single garnet crystals contain biotite, plagioclase and quartz. Regional scale homogenizing and gradual changes to diatexites are characteristic. The southern boundary of the northern Grt-Crd-Kfs zone is located in a major east-west-trending shear zone, in an elongated negative gravimetric anomaly (Fig. 2; Kiviniemi 1980). We conclude that the metamorphic path in the northern terrane has been closer to an isothermal decompression (ITD) path than in the southern terrane, in which almost isobaric cooling (IBC) path is probable after peak regional metamorphism (Fig. 5). The post-kinematic Luopa quartz monzonite in the southern terrane may refer to heating of the lower crust during crustal thickening – resulting a steady state geotherm and that environment will undergo isobaric cooling (see, e.g. Ellis 1987).

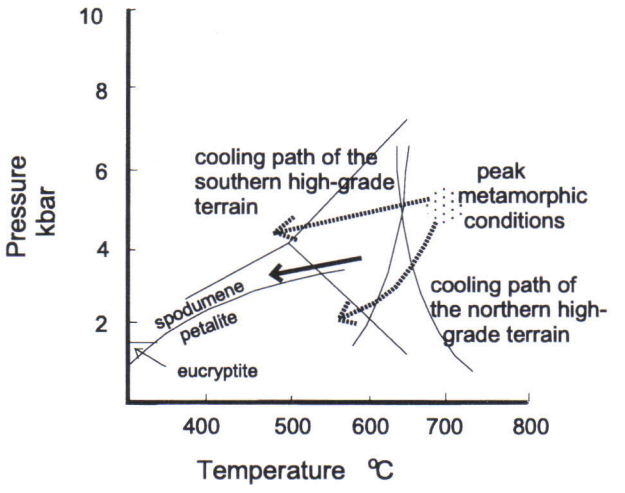


Fig. 5. P-T diagram of the Seinäjoki region with probable cooling paths of the high-grade metamorphic terranes. The black arrow indicates the direction of fractionation of LCT-type pegmatites containing spodumene but not petalite. Ecr-Spd-Ptl-Qtz invariant point by London (1984). Al_2SiO_5 triple point by Salje (1986). Muscovite breakdown reaction after Chatterjee and Johannes (1974). Solidus curve of granite at P(fluid) = P(load) after Jahns (1982)

GRANITIC PEGMATITES

Regional distribution and classification

The rare-element pegmatites of the Seinäjoki region are enriched in B, Be, Nb, Ta, Li, Rb, Cs, Li (see also Virkkunen 1963, Haapala 1966, Oivanen 1983, Nurmela 1985, Kärkkäinen 1993) and can be regarded as typical LCT-type pegmatites in the modern classification by Černý (1990, 1991a, 1991b). The rare-element pegmatite dykes usually occur as geographically separate swarms. The rare-element pegmatites have been divided into nine groups according to their location in the study area: Pajuluoma, Orisberg, Tarikko, Haapaluoma, Hunnakko, Kaatiala, Suokko, Palhojainen and Soromäki (Figs. 2 and 6). The individual dykes of each group contain variable amounts of rare minerals representing various differentiation stages of a pegmatite magma. Mineralogical classification of the pegmatite groups is given in Table 2.

A large proportion of the rare-element pegmatites are beryl-columbite pegmatites characterized by schorl, beryl and Nb-Ta minerals (mostly columbite-tantalite), and beryl-columbite-phosphate pegmatites. Complex pegmatites crystallized from the most fractionated pegmatite magma are found locally. The bulk of the complex pegmatites can be included in the group of spodumene pegmatites, but a few lepidolite pegmatites and one amblygonite (montebrasite) pegmatite are also known. Some pegmatites (Table 2) are difficult to classify because they tend to be intermediate in mineralogical composition and contain various amounts of type minerals such as beryl, columbite-tantalite, lithium silicates and phosphates.

Pegmatite dykes of the Pajuluoma group are

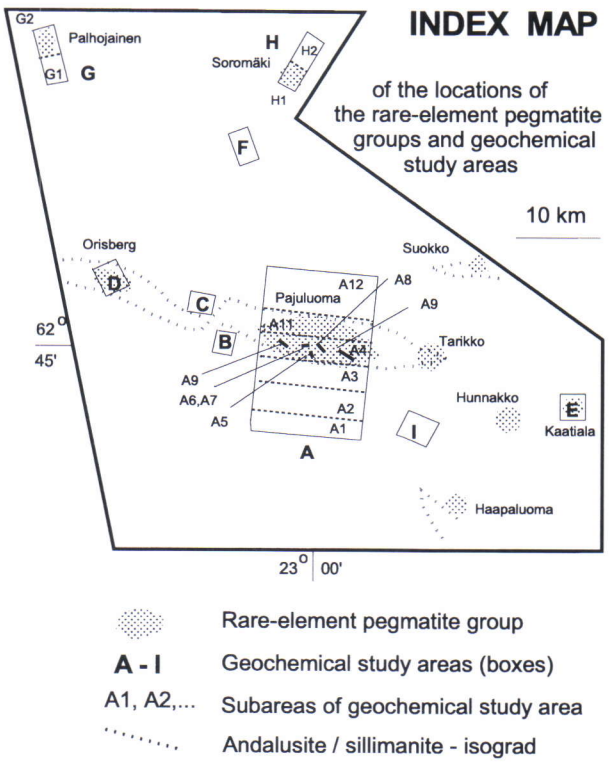


Fig. 6. Index map of the locations of rare-element pegmatite groups and geochemical study areas (A-I). The size of the geochemical study areas is shown by rectangles, and the subareas are indicated by dashed lines (cf. Figs. 1 and 2).

Table 2. Classification of rare-element pegmatite groups in the Seinäjoki region.

Pegmatite group	Moderately-well fractionated granitic pegmatites					
	Beryl pegmatites		Complex pegmatites			Topaz pegmatites
	Beryl-columbite pegmatites	Beryl-columbite-phosphate pegmatites	Spodumene pegmatites	Lepidolite pegmatites	Amblygonite pegmatites	Topaz pegmatites
Pajuluoma	+	+	+			
Orisberg	+					
Tarikko		+	+			
Haapaluoma	+		+	+		
Hunnakko					+	
Kaatiala	+		+			
Suokko		+				
Palhojainen						+
Soromäki						+

Table 3. Minerals identified in pegmatites of the Pajuluoma group and in the dykes of Haapaluoma, Kaatiala and Hunnakko. ++ = common mineral, + = trace mineral, (+) = sporadic mineral.

	Pajuluoma group	Haapaluoma dyke	Kaatiala dyke	Hunnakko dyke
Silicates:				
orthoclase	++			
microcline	++	++	++	++
plagioclase	++	++	++	++
spodumene	+	++	++	
eucryptite		(+)	(+)	
pollucite		(+)		
chabazite	(+)			
analcime		(+)		
biotite	++	++	+	+
muscovite	++	++	++	+
lepidolite		++	+	
chlorite	(+)	(+)		
cookeite	+	+	+	
montmorillonite	+	+		
serpentine	(+)			
kaolinite	+	+	+	
schorl	++	++	++	+
elbaite	+	++	+	+
beryl	++	++	++	++
bertrandite	+	+	+	
garnet	++	++	++	+
zircon	+	+	+	
thorite		(+)		
titanite	(+)			
topaz			+	(+)
bityite			(+)	
dumortierite			(+)	
dravite			(+)	
Phosphates:				
apatite	++	+	+	+
triphylite-lithiophilite	+		+	
ferrisicklerite-sicklerite	+			
heterosite-purpurite	+			
alluaudite	+		+	+
vivianite	+		+	
phosphosiderite	(+)			
lipscombite	(+)			
monazite		(+)		
brockite		(+)		
xenotime		(+)		
libethenite	(+)			
amblygonite-montebrasite			(+)	+
triploidite				+
childrenite				+
varulite			(+)	
Oxides, sulphides and other minerals:				
quartz	++	++	++	++
columbite-tantalite	+	+	+	+
tapiolite	+			
microlite	(+)	(+)		
cassiterite	++	+	+	+
uraninite	(+)			
arsenolite			(+)	
pyrite	(+)			(+)
marcasite	(+)			
sphalerite	(+)			
chalcopyrite			(+)	
löllingite	+	(+)	+	+
arsenopyrite	(+)	(+)	(+)	
tetraedrite			(+)	
covelite			(+)	
herzenbergite	(+)			
antimony	(+)			
stibnite	(+)			
graphite	(+)			
fluorite	(+)	(+)	(+)	
calcite	(+)			
siderite	(+)			
scorodite	+		+	
parasymplesite			(+)	
symplesite			(+)	
kaatialaite			(+)	

closely associated with the pegmatitic granite bodies abundant in this area (Fig. 7a). Pegmatite dykes near or in the pegmatitic granite bodies are relatively simple in mineralogy, whereas the dykes farther away may contain spodumene, beryl, cassiterite, columbite, triphylite and other rare accessories. Apatite, triphylite and varulite are characteristic of the phosphate-rich dykes. About 50 rare-element pegmatites occur in the Pajuluoma pegmatite group. The minerals identified in them are listed in Table 3.

Cassiterite is often more common than Nb-Ta minerals and thus the beryl-columbite pegmatites of this dyke swarm might preferably be called beryl-cassiterite pegmatites (Fig. 7b). For example, according to chemical analyses, the 10-m-wide Pajuluoma pegmatite (also known as the Perälä dyke) contains about 100 000 tonnes of tin ore with an average grade of 0.3% Sn (Saltikoff 1982, Oivanen 1983, Alviola 1989).

Typical examples of complex pegmatites are the Haapaluoma, Kaatiala and Hunnakko dykes (Haapala 1966) in the southeastern part of the study area where the pegmatites usually occur in granitoids of the sillimanite-muscovite zone. The pegmatites have been described by, among others, Volborth (1952), Haapala (1966), Nieminen (1978), Lahti et al. (1982). The minerals identified in these three dykes are listed in Table 3.

The Haapaluoma pegmatite is located some 30 km southeast of Seinäjoki (Fig. 1). The dyke is at least 500 m long and 30 m wide (Haapala 1966). Close to the southern contact there is a smaller parallel dyke of the same type. From 1961 to 1997 some 300 000 tonnes of K-feldspar and quartz were quarried from the pegmatite (see Fig. 7c). In addition, small amounts of gem minerals were discovered (Vilpas 1996).

The pegmatite is of an intermediate type between spodumene and lepidolite pegmatites. Typical minerals in the albite-rich replacement bodies and fracture fillings in the dyke are blue fluorapatite, yellowish or reddish beryl, cassiterite, columbite-tantalite, grey or lilac spodumene, red or green elbaitic tourmaline and purple lepidolite. During regional bedrock mapping several small beryl-columbite pegmatites with abundant cassiterite were discovered in the surroundings of the Haapaluoma quarry.

The Kaatiala dyke lies about 20 km northeast of the Haapaluoma quarry. The dyke is a subhorizontal, 40-m-wide zoned lens. The pegmatite, quarried during 1942-1969, produced some 160 000 tonnes of K-feldspar and 30 000 tonnes of quartz (Nieminen 1978). Also the Kaatiala pegmatite may be included in the group of spodumene pegmatites, although abundant lepidolite, lithiophilite, varulite and even minor amounts of amblygonite-montebrasite have been found in addition to spodumene in the strongly albitized zones close to the quartz core.

The Hunnakko amblygonite pegmatite is located about 5 km west of Kaatiala. The dyke is 5 m wide and

intersects the schistosity of the country rock, a migmatitic granodiorite (Haapala 1966). The zoned texture of the Hunnakko dyke is complex. There are quartz cores in the central parts and albite-rich parts with beryl, columbite-tantalite and phosphate minerals including apatite, montebrasite, varulite, triploidite and childrenite. The schists, gneisses and plutonic rocks surrounding the Hunnakko and Kaatiala pegmatites in places contain beryl-columbite pegmatites (e.g. Volborth 1952).

The Suokko pegmatite is located some 15 km and the Tarikko pegmatite some 20 km east of Seinäjoki. Because of the thick overburden the distribution of the adjacent pegmatite dykes is poorly known. Spodumene, beryl, cassiterite, columbite and triphylite are characteristic minerals of the Tarikko dyke (Alviola 1987) which occurs in a separate continuation of the Seinäjoki andalusite schists. The Suokko pegmatite contains beryl, columbite, triphylite and ferrisicklerite as characteristic accessories (Alviola 1989). The Suokko group is part of the rare-element pegmatites of the Kuortane area located outside the study region.

In the village of Orisberg, ca. 20 km west of Seinäjoki town, a few rare-element pegmatites containing beryl and some columbite have been discovered. Most of the pegmatite dykes occur in an andalusite mica schist (Fig. 2).

Tabular pegmatite dykes and larger pegmatitic granite intrusions are almost totally absent from the high-grade metamorphic mica gneisses in the Crd-Kfs and Grt-Crd-Kfs zones, but there are narrow irregular biotite-pegmatite bodies that were formed during anatexis of the rocks. These are very simple in mineralogy but may contain garnet and cordierite.

A few small topaz-bearing pegmatite dykes with rare columbite were discovered in the northern part of study region, in high-grade garnet-cordierite mica gneisses (Fig. 3d) (Alviola 1989, Vilpas 1996). Three dykes occur at Palhojainen, Isokyrö, and two at Soromäki, Lapua (Fig. 2); all are close to the

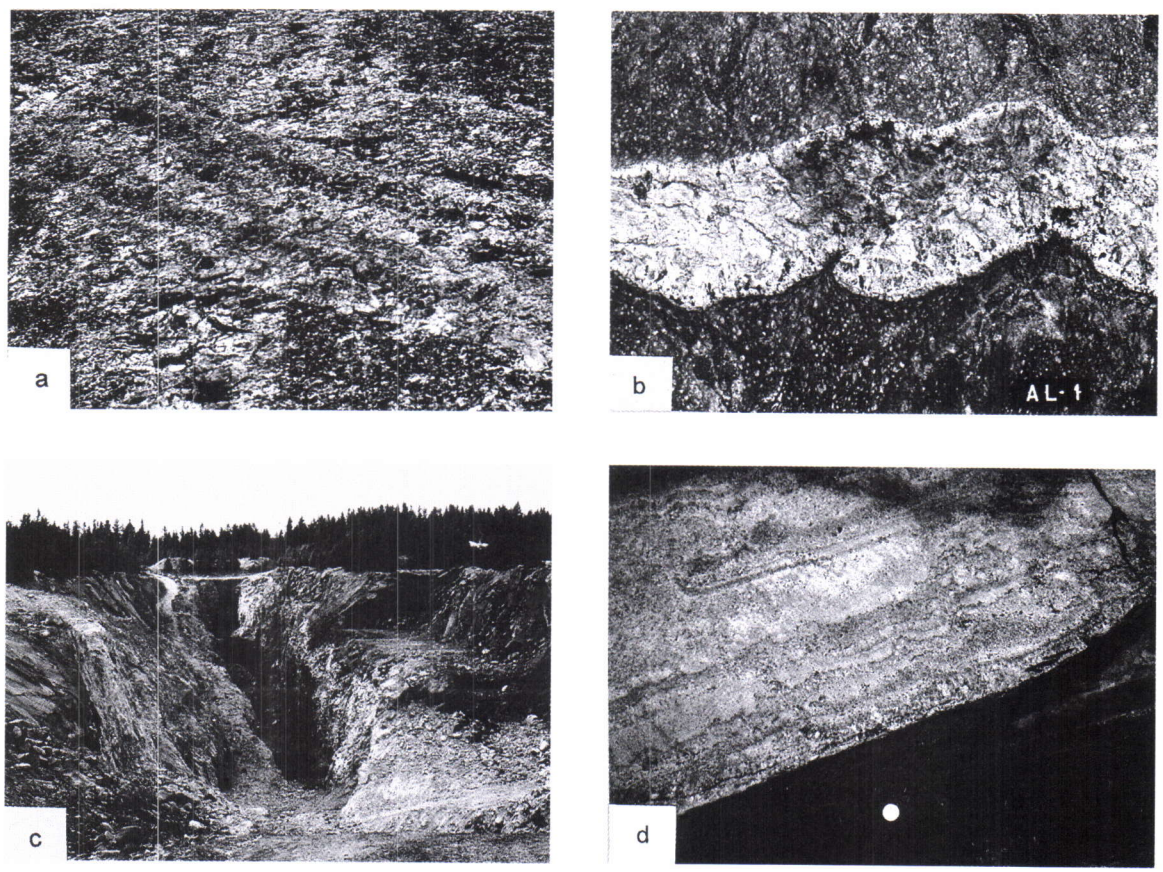


Fig. 7. Granitic pegmatites of the Seinäjoki region. **(a)** Pegmatitic granite with two intersecting pegmatite dykes. Length of scale bar 12 cm. Pajuluoma pegmatite group. **(b)** Deformed (probably D_5, see Mäkitie 1999) cassiterite-bearing pegmatite dyke in a foliated plagioclase porphyrite. The dyke has an unusual structure, cassiterite (black grains) having concentrated near the boundaries. The dyke also contains beryl and columbite. Length of scale bar 4 cm. Pajuluoma pegmatite group. **(c)** The Haapaluoma complex pegmatite quarry in the 1980's. Haapaluoma pegmatite group. **(d)** Magmatic layering in the wall zone of pegmatitic granite intrusion. Diameter of coin 2 cm. Pajuluoma pegmatite group.

VMC. Geochemically these pegmatites exhibit also features of the NYF (niobium-yttrium-fluorine)-type pegmatites (cf. Černý 1990, 1991a, 1991b), and they form a separate group in the present pegmatite classification (Table 2). These pegmatites may also be classified as a subtype of beryl pegmatites due to their columbite content but absence of lithium minerals.

Zoning

The pegmatites are poorly to well zoned, rarely banded, and their mineralogical composition may vary largely from one zone to another. Three separate zones can usually be recognized: a border zone, a wall zone and an intermediate zone. The intermediate zone together with the quartz cores constitutes the interior of the dykes. Banded or layered textures caused by several successive magma flows are locally found in the wall zones (Fig. 7d). In places deformation has destroyed the zoning and, in addition, the pegmatite often exhibits an irregular form (Fig. 4).

The border zone is seldom more than 1-10 cm wide, actually no more than a fine-grained reaction rim between pegmatite and the wall rock. Albite, quartz and muscovite, or locally biotite, are dominant in the border zone of the rare-element pegmatites, but the mineralogy also varies from dyke to dyke. The wall zone, which is composed of coarse albite-quartz-microcline-mica pegmatite, usually contains equal amounts of albite and K-feldspar. Graphic feldspar is characteristic of the pegmatitic granites and the border and wall zones of many pegmatite dykes.

Main mineralogy and characteristic accessory minerals

The occurrence of the main and most important accessory minerals has been described in several of the papers referred to above and is therefore only reviewed here. The minerals encountered in the Pajuluoma group and in the Haapaluoma, Kaatiala and Hunnakko dykes are listed in Table 3.

The main minerals of the pegmatites are K-feldspar, quartz, albitic plagioclase, muscovite and/or biotite. K-feldspar is usually perthitic microcline and grey, white or sometimes pinkish in colour. The grain size and the abundance of K-feldspar increase from the border zone inwards. The intermediate zone is composed of coarse-grained (∅ 10-30 cm) or very coarse-grained (>30 cm) pegmatite. The microcline crystals or crystal accumulations in the largest dykes may be huge, even several metres long.

Yellowish, greenish or greyish muscovite is the most common, and indeed the only mica in most of the rare-element pegmatites. Biotite is characteristic of the pegmatitic granites and the pegmatites of the high-grade metamorphic zones. Biotite is often intergrown with muscovite in the two-mica pegmatites and pegmatitic granites. The mica sheets may measure several centimetres or, in larger rare-element pegmatites, even many decimetres.

The distinctly zoned pegmatite dykes always have an oval or irregular quartz core or several separate cores of various size randomly distributed in the intermediate zone. In some dykes the quartz cores are connected with each other and form a continuous quartz-vein type of core, the width of which varies largely and often depends on the size of the pegmatite dyke. In the largest dykes the cores may be several metres wide. The mineral is usually grey, brownish, white or colourless, but large rose quartz cores have also been observed in some dykes.

Late- or postmagmatic replacement bodies and fracture fillings are characteristic of the furthest fractionated rare-element pegmatites (e.g. Haapala 1966). Their main minerals are sugary or platy albite, quartz and muscovite.

Locally, the albite-rich fracture fillings, which may follow the direction of the main dyke, have garnet-, schorl-, mica- or apatite-rich bands. Schorl is common in various parts of the dykes and beryl, apatite, columbite and cassiterite, which are typical accessories, may also occur locally in the border and wall zones. Spodumene, lepidolite, triphylite-lithiophilite, coloured elbaitic tourmalines, Nb-Ta minerals, cassiterite, topaz and other rare minerals are always concentrated in the last-crystallized central parts of the dykes, in the intermediate zone, and in albite-rich replacement bodies and fracture fillings.

Common accessory minerals in the barren pegmatites and rare-element pegmatites are black tourmaline (schorl), greenish to greyish fluorapatite,

Table 4. Average chemical compositions of pegmatites in the geochemical study areas and subareas. The amount of chemical analyses in each study area and subarea is shown in an own row.

Geochemical study area	A												B	C	D	E	F	G		H		I
Subarea	A1	A2	A3	A4	A5	A6	A7	A8	A9	A10	A11	A12						G1	G2	H1	H2	
Pegmatite type	Dykes	Dykes & pegm. granites	Dykes & pegm. granites	Dykes	Sn-dyke (Paju-luoma)	Be,Sn,Nb-dykes	Be,Sn-dyke	Pegm. granites (small)	Pegm. granite (Hallilanvuori)	Pegm. granites (large)	Dykes	Dykes	Dykes	Dykes	Dykes	Li,Nb,Be-dyke (Kaatiala)	Dykes	Topaz-dyke (Palhojainen)	Dykes	Topaz-dyke (Soromäki)	Dyke	Pegm. granite
Amount of analyses	2	5	16	4	6	2	2	9	6	9	5	9	2	3	4	average	4	2	2	2	1	4
Amount of analysed pegmatites	2	5	16	4	1	2	1	9	1	2	4	8	2	3	4	1	4	1	2	1	1	1
SiO_2 (wt%)	68,95	72,82	72,21	72,27	68,92	71,73	73,88	71,83	74,09	73,23	73,15	71,13	75,05	74,16	74,36	71,41	72,59	70,24	70,56	70,77	70,66	74,24
TiO_2	0,10	0,10	0,03	0,03	0,02	0,06	0,03	0,06	0,07	0,06	0,09	0,07	0,06	0,05	0,03	0,08	0,13	0,13	0,21	0,04	0,25	0,06
Al_2O_3	14,59	14,66	15,22	14,83	16,17	15,64	15,79	14,71	14,56	15,26	14,05	14,60	14,62	14,97	15,20	15,41	14,65	14,36	13,91	13,91	13,18	15,02
Fe_2O_3tot	0,99	0,87	0,74	0,66	0,39	0,52	0,31	0,49	0,65	0,69	0,32	0,74	0,60	0,66	0,52	0,81	1,06	1,75	2,11	0,59	1,14	0,85
MnO	0,03	0,01	0,01	0,09	0,02	0,08	0,02	0,04	0,02	0,04	0,02	0,02	0,00	0,05	0,02	0,03	0,04	0,08	0,05	0,01	0,01	0,02
MgO	0,29	0,40	0,29	0,09	0,04	0,03	0,03	0,14	0,31	0,34	0,07	0,25	0,36	0,30	0,10	0,28	0,24	0,29	0,24	0,06	0,35	0,43
CaO	0,43	0,93	0,88	0,53	0,67	0,39	0,54	0,43	0,66	0,76	0,66	0,63	0,90	0,66	0,56	0,52	0,50	0,88	0,51	0,60	0,66	0,78
Na_2O	2,57	2,77	3,30	4,64	4,99	5,64	6,52	4,06	3,92	3,88	4,47	3,23	3,96	4,28	3,91	4,60	2,22	3,74	2,74	3,37	2,30	3,58
K_2O	7,83	5,37	5,15	2,81	3,49	2,88	1,05	4,37	3,89	4,38	3,08	6,08	3,10	3,32	3,80	5,40	7,11	3,73	6,06	5,27	6,89	3,89
P_2O_5	0,08	0,18	0,29	0,21	0,76	0,46	0,48	0,26	0,12	0,16	0,20	0,18	0,13	0,17	0,17	0,06	0,15	0,01	0,01	0,02	0,07	0,17
Total	95,86	98,11	98,12	96,16	95,47	97,43	98,65	96,39	98,29	98,80	96,11	96,93	98,78	98,62	98,67	98,60	98,69	95,21	96,40	94,64	95,51	99,04
Rb ppm	249	147	152	238	300	289	166	283	136	146	117	188	116	-	124	300	326	485	559	455	173	142
Ba	919	843	758	45	90	46	71	61	70	404	200	449	491	172	293	30	690	84	207	293	1210	-
Sr	189	170	149	31	31	30	44	18	44	-	54	134	120	-	67	30	221	30	56	89	121	70
Li	23	-	34	58	37	8	86	22	30	-	11	12	-	-	17	150	13	73	49	14	11	-
Cs	19	10	14	21	25	24	16	38	16	16	10	7	12	13	8	30	9	4	13	10	1	12
Ta	1	-	1	5	83	54	82	35	31	-	1	1	-	-	2	4	7	4	4	3	0,2	-
U	18	5	8	8	13	32	18	14	7	9	8	5	4	11	8	3	6	9	10	6	2	6
Th	5	2	1	1	0,4	1	1	1	1	1	1	1	-	1	0,6	3	29	28	42	24	46	1
La	7	5	4	2	1	2	2	3	-	-	3	5	-	-	2	1,5	66	29	43	24	76	-
Sm	3	2	1	0,4	0,1	0,0	0,2	0,5	-	-	1	1,4	-	-	0,5	2	8	10	12	4	9	-
Y	3	16	13	4	1	1	1	3	-	-	5	7	-	-	3	5	9	27	34	9	6	-
Sc	3	5	5	2	1	1	1	1	3	5	1	3	6	4	2	4	4	17	8	4	3	5
Sn	7	8	12	33	2569	34	50	42	12	16	8	10	-	19	9	12	-	11	0	0	0,2	15
As	5	15	22	36	4	9	15	48	181	91	28	2	6	418	4	30	2	0	1	1	0	21
B	-	79	172	-	-	-	-	1040	1238	746	-	-	31	973	-	1200	-	-	-	-	-	176
Be	-	-	-	-	-	-	-	49	7	10	-	-	-	16	-	6	-	-	-	-	-	-
Zr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	25	13	-	45	81	-	142	-
A/ CNK	1,08	1,22	1,21	1,27	1,22	1,19	1,23	1,21	1,23	1,22	1,18	1,12	1,27	1,26	1,31	1,07	1,20	1,22	1,16	1,13	1,06	1,30
K/Rb	261	303	281	98	97	83	53	128	237	249	219	268	222	-	254	149	181	64	90	96	331	227
Sr/Rb	0,76	1,16	0,98	0,13	0,10	0,10	0,27	0,06	0,32	-	0,46	0,71	1,03	-	0,54	0,10	0,68	0,06	0,10	0,20	0,70	0,49
Ba/Rb	3,69	5,73	4,99	0,19	0,30	0,16	0,43	0,22	0,51	2,77	1,71	2,39	4,23	-	2,36	0,10	2,12	0,17	0,37	0,64	6,99	-
K/Na	3,4	2,2	1,7	0,7	0,8	0,6	0,2	1,2	1,1	1,3	0,8	2,1	0,9	0,9	1,1	1,3	3,6	1,1	2,5	1,7	3,4	1,2
K/Cs	3421	4458	3054	1111	1159	996	545	955	2018	2273	2557	7211	2145	2120	3943	1494	6558	7741	3870	4375	57198	2691
Fe/Cs	364	608	370	220	109	152	136	90	284	302	224	739	350	355	455	189	824	3060	1135	413	7973	495
Mg/Rb	7,0	16,4	11,5	2,3	0,8	0,6	1,1	3,0	13,7	14,0	3,6	8,0	18,7	-	4,9	5,6	4,4	3,6	2,6	0,8	12,2	18,3
Ba/La	131	169	190	23	90	23	36	20	-	-	67	90	-	-	147	20	10	3	5	12	16	-
La+Sm+Y	13	23	18	6	2	3	3	7	-	-	9	13	-	-	6	7	83	66	89	37	91	-
Ti/Sn	86	75	15	5	0,05	11	4	9	35	22	67	-	-	16	20	40	-	71	-	-	7494	24
Th/U	0,28	0,4	0,13	0,13	0,03	0,03	0,03	0,07	0,14	0,11	0,13	0,20	-	0,09	0,08	1,00	4,83	3,11	4,20	4,00	23,00	0,17
Host rock	Tonalite	Mica gn.	Mica gn.	Mica sch.	Mica sch.	Mica sch.	Mica sch.	Graywac.	Mica sch.	Mica sch.	Graywac.	Mica gn.	Mica gn.	Mica gn.	Graywac.	Granod.	Mica gn.	Mica gn.	Mica gn.	Mica gn.	Mica gn.	Mica gn.
Metamorphic zone	Sil-Kfs (southern)	Sil-Ms (S-part of southern)	Sil-Ms (N-part of southern)	And	And	And	And	And	And	And	And	Sil-Ms (northern)	Sil-Ms (southern)	Sil-Ms (northern)	And	Sil-Ms	Crd-Kfs (northern)	Grt-Crd-Kfs (northern)	Grt-Crd-Kfs (northern)	Sil-Kfs (northern)	Crd-Kfs (northern)	Sil-Ms (southern)

red almandine-spessartine garnet and small dark crystals of zircon. Schorl, which may be locally very abundant, occurs as huge prismatic crystals in the dykes. Iron sulphides, arsenopyrite or löllingite are locally common. Löllingite also occurs as large aggregates and veinlets in the Kaatiala pegmatite. In places, graphite coats the shear-planes and occurs as small (1-5 mm) poorly crystallized spherules (thucholites).

The minerals characteristic of the more fractionated beryl-columbite and beryl-columbite-phosphate pegmatites are beryl, columbite-tantalite, cassiterite and triphylite-lithiophilite, which is often replaced by varulite or other secondary phosphates (e.g. ferrisiclerite-siclerite, heterosite-purpurite, vivianite, phosphosiderite, lipscombite). In addition to columbite-tantalite, tapiolite and microlite are characteristic Nb-Ta minerals in some pegmatites. Topaz has been encountered in the Kaatiala pegmatite, but the mineral occurs more commonly as small (1-5 cm) prismatic crystals in the pegmatite dykes in the northern part of the study region.

Beryl is yellowish or greyish, rarely pink, in colour. The crystals are prismatic, usually some centimetres in diameter. Bertrandite is a common alteration product of beryl. Columbite is the most common Nb-Ta mineral in the pegmatites, but tapiolite was also discovered. Columbite-tantalite usually occurs as black platy, rarely columnar crystals some millimetres or rarely some centimetres long. Cassiterite is common and usually occurs as dark brown circular grains measuring 5-15 mm (Fig. 7b). Euhedral crystals are rare, but short prismatic crystals with some simple pyramidal faces were locally observed.

Characteristic accessory minerals in the most fractionated complex pegmatites are spodumene, lepidolite and coloured elbaitic tourmalines. In addition, the pegmatites may contain large quantities of common beryl, columbite-tantalite, cassiterite, löllingite, amblygonite-montebrasite, triphylite-lithiophilite, varulite and numerous rarer minerals. Schorl, beryl and columbite may occur as huge crystals in the complex pegmatites. For example, in the Kaatiala pegmatite the biggest beryl and schorl crystals were 3-4 decimetres in diameter and several metres long; the columbite crystals could weigh as much as several kilograms.

Spodumene occurs in pegmatites as long (5-30 cm) crystal laths. The mineral is grey, but lilac and rarely greenish varieties occur in the Haapaluoma pegmatite. Spodumene may be associated with purple lepidolite and with green and/or red elbaite. Kaolinite, smectite-group clay minerals and cookeite are common alteration products of spodumene. Eucryptite has been discovered in the Haapaluoma and Kaatiala pegmatites with altered spodumene.

Elbaitic tourmaline usually occurs as small green or red prismatic (diameter 1-50 mm) crystals in the central parts of pegmatites. Purple cookeite locally replaces altered red tourmaline crystals or occurs as fine-grained light-green aggregates in the late-stage mineral associations. Lepidolite, usually purple or pink in colour, occurs as small-scale accumulations or as larger separate sheets some centimetres in diameter. In the Haapaluoma pegmatite lilac spodumene, purple lepidolite, reddish beryl and red tourmaline were exceptionally common in the cleavelandite-rich replacement bodies and fracture fillings.

Grey, greenish or bluish triphylite-lithiophilite occurs as circular crystals or aggregates some 10 cm in diameter. The mineral is often strongly altered and replaced by varulite-group minerals, ferrisiclerite-siclerite, heterosite-purpurite or by water-bearing (Fe,Mn)phosphates. In the phosphate-rich parts of the Kaatiala pegmatite, lithiophilite and varulite crystals were large, up to 30 cm in diameter. Lithiophilite is partly or totally replaced by dark-green varulite, grey or black apatite, vivianite and a brown mixture of various low-temperature phosphates.

The nearby Hunnakko pegmatite is also locally rich in phosphate minerals. Montebrasite, alluaudite, triploidite, childrenite, vivianite and fluorapatite have been described from the locality. Montebrasite and varulite crystals or crystal aggregates may be up to 30 cm in diameter. Two generations of montebrasite have been described from the pegmatite. As in the Kaatiala pegmatite, varulite is very fine-grained and occurs in dark green or brown pseudomorphs after lithiophilite (Haapala 1966).

GEOCHEMISTRY OF THE PEGMATITES

Introduction

Much of what is known about the chemical composition of pegmatites is based on the chemical composition of single minerals in pegmatites. The concentrations of certain elements in feldspars and micas are therefore used as a fractionation index (e.g. Černý et al. 1985). The occurrence of specific rare minerals, e.g. lithium-minerals, columbite-tantalite and beryl, is used to classify the pegmatites. The crystals of the indicator minerals are often so large that the minerals are easily visible. On the other hand, sampling for chemical whole rock analyses is much more difficult from very coarse-grained, zoned dyke rocks. Chemical whole rock analyses are useful in poorly or moderately fractionated pegmatites, where indicator minerals are rare or difficult to observe.

The ratios of alkali and alkali earth metals are useful in whole rock chemical studies of pegmatites, because they are of the same range in both feldspar and mica due to the similarity of the distribution coefficients (e.g. Cesbron 1989). The ratio is independent of the absolute abundances of these elements, and of the volume of quartz causing dilution by decreasing the absolute proportions of elements other than 'inert' silica. However, the ratios do not guard against sampling a zone that is richer in K-feldspar in one dyke compared to another sample containing more biotite. This can be controlled by element concentrations and ratios of Fe, Mg carried by biotite. Whole-rock samples can be analysed, even for small abundances of trace elements such as Sn, Ta, Li, which first concentrate into mica and only in extreme concentrations precipitate as their own minerals.

The geochemistry of whole rock samples was used in Sn exploration in the research area, with encouraging preliminary results (Kärkkäinen 1986, 1993). Even a coarse sample grid revealed a potential area, especially when alkali and alkali earth element ratios were used. In a 20-km-long section trending north-

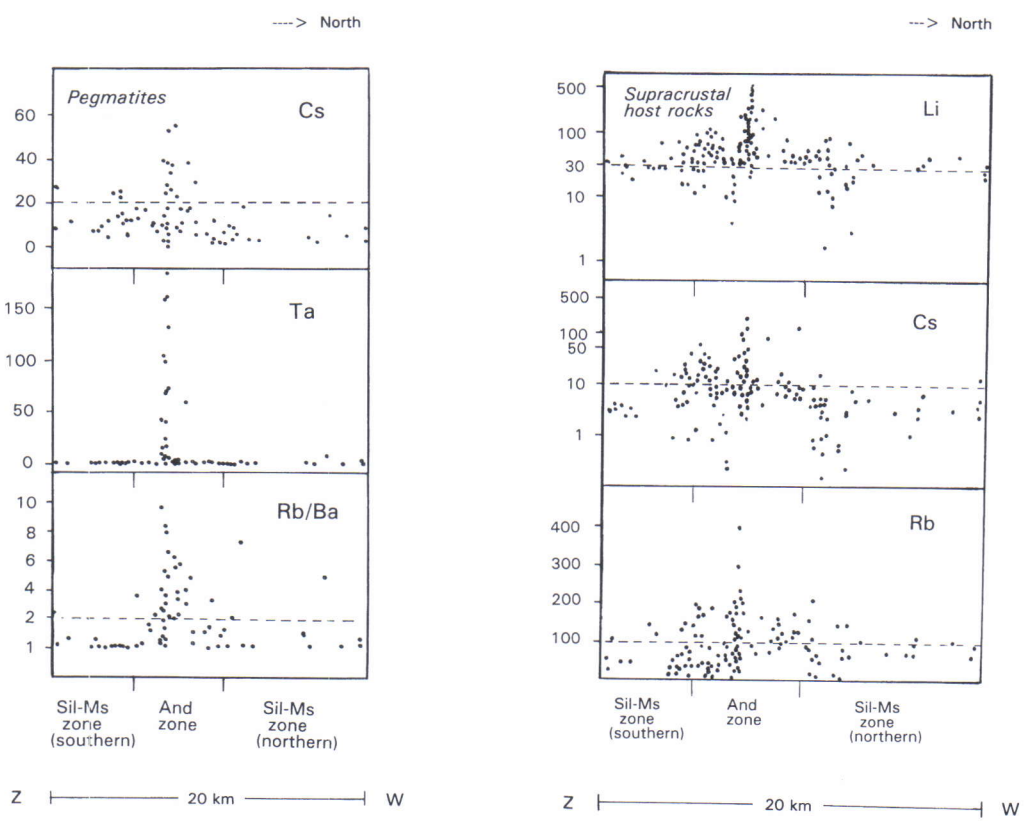


Fig. 8. Trace element concentrations and ratios in the pegmatites (left) and their country rocks from a 20-km-long, N-S-trending cross-section (W-Z) passing Seinäjoki town and shown in Figure 1 (after Kärkkäinen 1993).

south across the schist belt through Seinäjoki the chemical compositions of the pegmatites vary, showing some distinct features such as clustering of the Rb/Ba ratios (Fig. 8; see Kärkkäinen 1986). Ta and Cs of pegmatites are anomalous in a restricted area and the supracrustal country rocks of pegmatites are enriched in Li (Fig. 8).

Sampling and analytical procedures

Several samples were taken from each pegmatite for the chemical determinations, because the dykes tend to be very coarse-grained, zoned or otherwise heterogeneous. One to three samples weighing 0.5-3 kg were drilled from narrow dykes (width < 0.5 m), but from the wide dykes one sample was taken from the core and one from the wall zone close to the contact. For the larger pegmatitic granite intrusions, several samples were drilled from sites across the body. The samples were taken from the relatively homogeneous parts of the dykes, and the samples of the very coarse-grained parts were combined; the analytical results are representative.

Major elements and Zr were analysed by X-ray fluorescence (XRF) and inductively coupled plasma mass-spectrometry (ICP-MS) at the Chemistry Laboratory of the Geological Survey of Finland (GTK). The total sums (%) (wt) of elements in analyses are often relatively low (95-98%) (Table 4). The reason for this is not clear. However, the low sums probably are to some extent due to OH-groups of micas. Li and Sr were determined partly by ICP-MS and partly by neutron activation analysis (INAA). As, Ba, Cs, Rb, Sc, Ta, La, Sm, U and Th were determined by INAA at the Technical Research Centre of Finland. Tin, yttrium and beryllium were mainly determined by optical emission spectrometry (OES) at the GTK.

Geochemical study areas

The granitic pegmatite samples from the Seinäjoki region were taken from separate restricted localities, called here geochemical study areas and referred to by the letters A-I (for sites and dimensions see Figs. 2 and 6). The areas extend across different metamorphic zones, and represent barren pegmatites and a few of the present rare-element pegmatite groups. Some of them, however, include different pegmatites, and were therefore divided into subareas, referred to here by the Arabic numerals A1, A2,..(Table 4). A brief introduction to each geochemical study area is in Appendix 1.

Chemical analyses of pegmatites are available for each study area and subarea (Table 4). The chemical compositions of each areas and subareas are shown as average value. Standard deviations of the analysis are in Appendix 2.

Table 5. Grouping of the geochemical study areas by the metamorphic grade of their country rocks and 'degree' of fractionation, with the symbols referring to them in scattergrams.

Metamorphic grade of host rocks	'Degree' of the fractionation of the pegmatites	Geochemical study area or subarea	Symbol of the areas in the scattergrams
And zone	Poorly fractionated	A4, A8, A9, A10, A11, D	○
	Moderately-well fractionated	A5, A6, A7	●
Sil-Ms zones or Sil-Kfs zones	Poorly fractionated	A1, A2, A3, A12, B, C, I	◇
	Moderately-well fractionated	E, H1	◆
Crd-Kfs zones or Grt-Crd-Kfs zones	Poorly fractionated	F, G2, H2	□
	Moderately fractionated	G1	■

General geochemical features of the pegmatites

The reliability of the chemical analyses that here represent the dyke composition is based on the number of sampling points. Pegmatitic granite intrusions and dykes were analysed separately (see Table 4). The following general observations were made concerning the composition of the major and

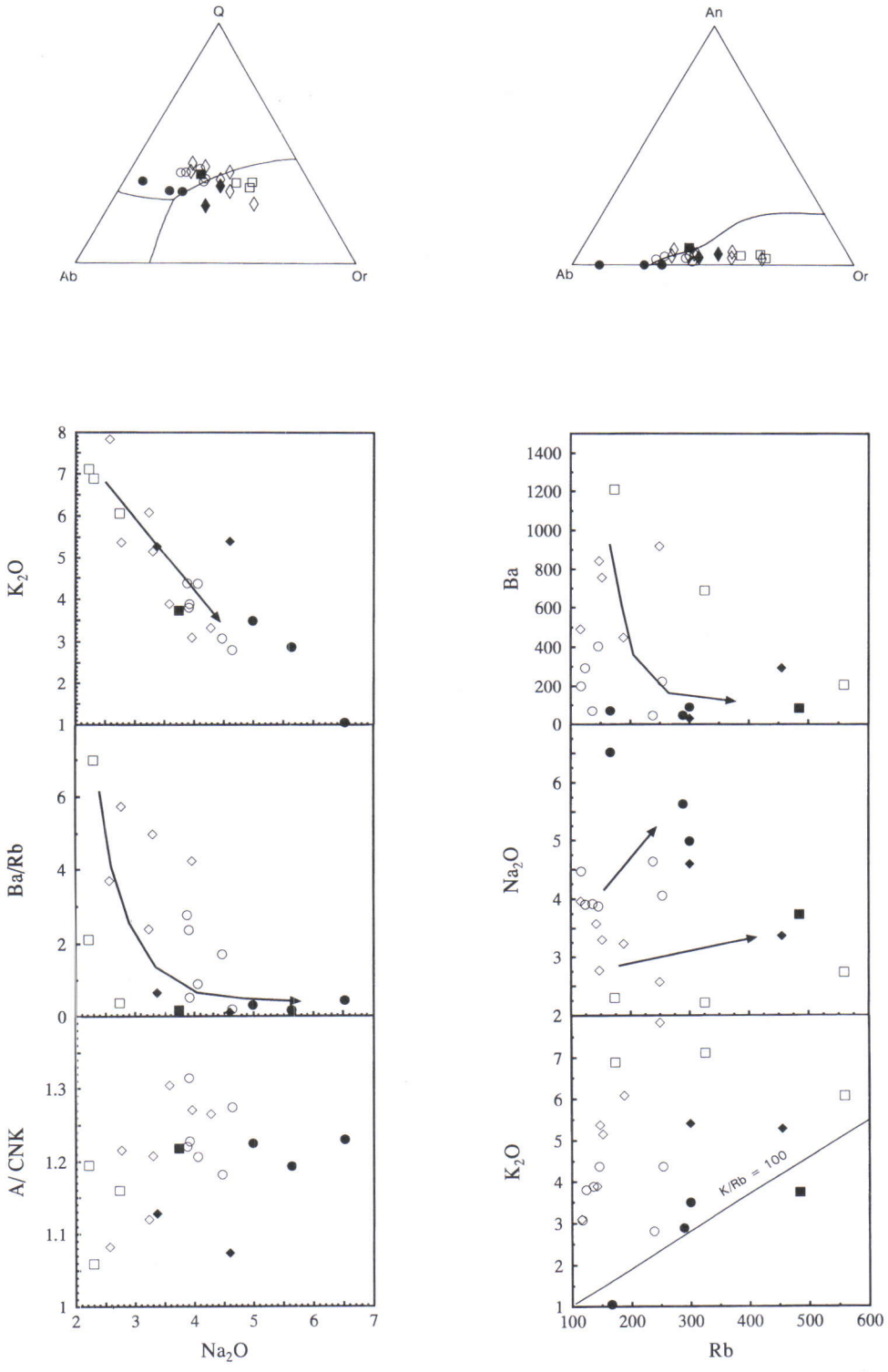


Fig. 9. Normative Q-Ab-Or and An-Ab-Or diagrams (with isobaric cotectic lines at 5 kbar water pressure, cf. Winkler 1979) and scattergrams of Na, K, Rb, Ba, A/CNK concentrations of the pegmatites of the Seinäjoki region. General trends of increasing fractionation are shown by arrows. For symbols, see Table 5.

trace elements in the granitic pegmatites of the various metamorphic zones. The concentrations given in parenthesis in the following text refer to averaged analysis (see Table 4) and, thus, the concentrations in individual samples may be notably higher.

1) The concentrations of K_2O, CaO, Fe_2O_3tot, MgO, TiO_2, Ba, Sr, Th, La, Y, Sm and Sc are high in the pegmatites of metamorphic areas above the sillimanite isograd, but low in the pegmatites of the andalusite zone.

2) The concentrations of Na_2O, P_2O_5, Ta, Cs, Sn, As and B are high in the pegmatites of the andalusite zone, but low in those of the higher metamorphic areas.

3) Li and Rb have concentrated in pegmatites of

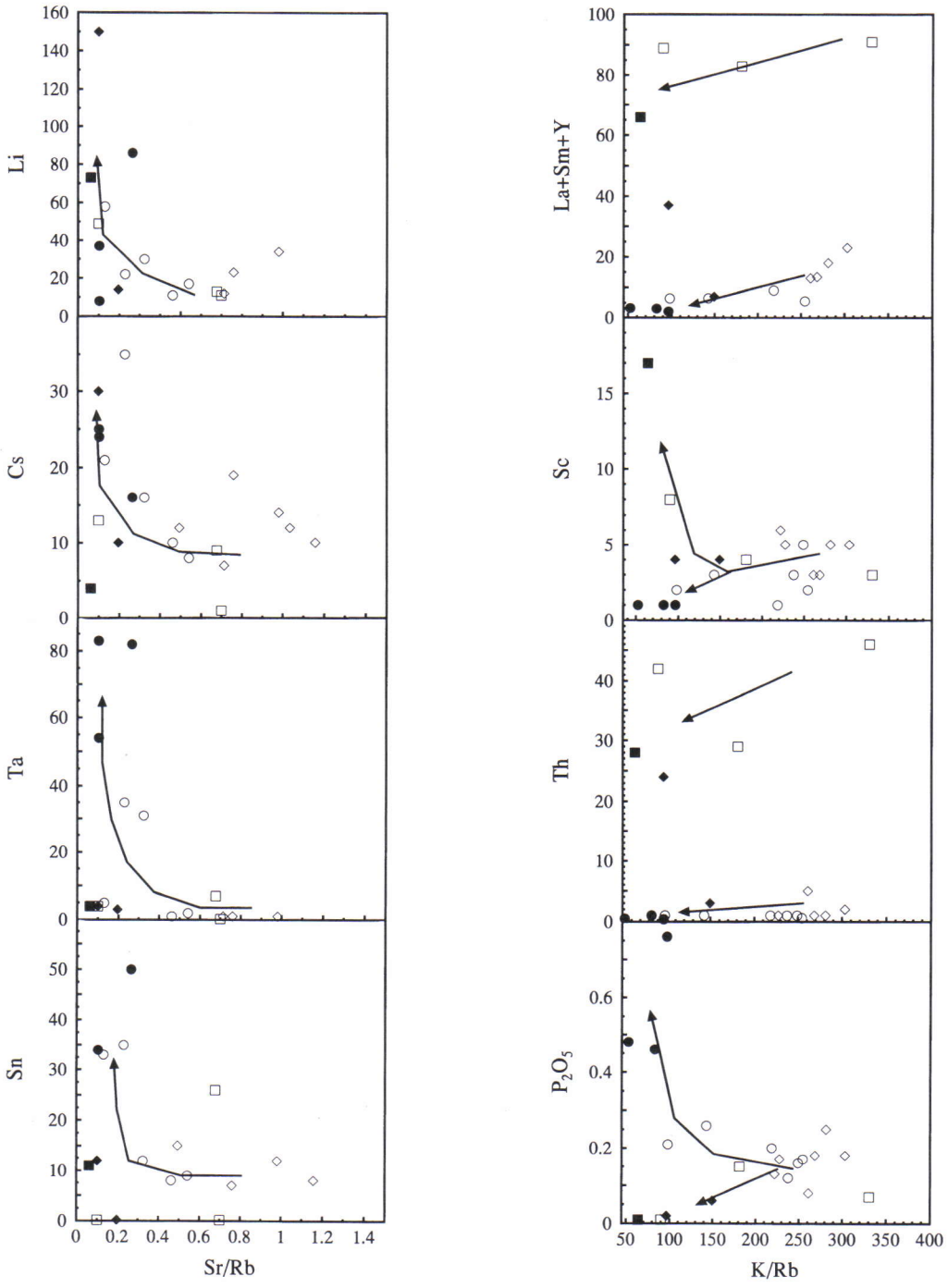


Fig. 10. Scattergrams of concentrations of typical elements in the LCT-type (on the left) and NYF-type pegmatites against indicators of fractionation of the pegmatites of the Seinäjoki region. For symbols, see Table 5. General trends of increasing fractionation are shown by arrows.

low-grade metamorphic areas and in some moderately fractionated pegmatites of the high-grade metamorphic zone.

4) B is concentrated in pegmatitic granites of the andalusite zone, but also in some dykes of the Sil-Ms zone.

There are, however, exceptions to these variation trends. Barium is low (30 ppm) in the complex pegmatite of the Kaatiala dyke in the muscovite-sillimanite zone. Phosphorus, boron and lithium are common fluxing elements in pegmatites and they lower the solidus temperature of magma. The Kaatiala dyke has a gentle dip (Nieminen 1978), which possibly caused that the B- and Li-rich melt could not move upwards from the country environment of relatively high-grade metamorphism.

The compositional variation in geochemical study area A is interesting, because the area intersects the first sillimanite isograd. The overcrossing of the isograd is best marked by the increase in MgO (up to 0.43 wt%), Ba (up to 1210 ppm), Sr (up to 189 ppm), Sc (up to 5 ppm) and K/Na (over 1) in granitic pegmatites (Table 4). The increase in Ba/Rb (over 2) and decrease Sn (less than 10 ppm) in pegmatites is also sharp at the first sillimanite isograd.

The enrichment in K_2O (up to 7.8 wt%) and CaO (up to 0.9 wt%) of pegmatites in the relatively high-grade metamorphic areas indicates that the pegmatites contain more K-feldspar and biotite than do those in the andalusite zone and that plagioclase is richer in anorthite component. Sr and Ba concentrations are also generally higher, because Sr is concentrated in anorthite and Ba in K-feldspar.

Because biotite is more common than muscovite in pegmatites of the high-grade metamorphic areas, these pegmatites have elevated Fe, Mg and Ti concentrations. Some fractionated pegmatites have high Fe contents (> 0.8 wt% Fe_2O_3tot), but their country rocks are of a relatively high metamorphic grade (cf. the Kaatiala complex pegmatite and the topaz pegmatites). However, in the Kaatiala complex pegmatite, in which biotite is rare, most of the iron is concentrated in columbite, löllingite and schorl (Nieminen 1978).

Lithium (up to 86 ppm) and caesium (up to 35 ppm) are enriched both in fractioned pegmatites and surrounding host rocks in the Seinäjoki region (Fig. 6). Thus, routine analyses of Li and Cs from bulk rock samples of all rock types is a good method in exploring fractioned pegmatites (Kärkkäinen 1986). Feldspar and micas were enriched in Cs and Rb in the fractionation process. Saturation of Cs in the Haapaluoma dyke has resulted in the precipitation of Cs-silicate pollucite.

In general, Li is dispersed among the many silicates and phosphates often randomly located in pegmatites. In barren pegmatites Li content is 8-35 ppm that is of same grade than in the supracrustal country rocks. The general Li content is not very high (30-86 ppm) in the fractioned pegmatites of the Seinäjoki region (Table 4). The average Li concentration of the Kaatiala spodumene pegmatite is only 150 ppm (Nieminen 1978), which is still many times higher than that in the moderately and well fractionated cassiterite-columbite pegmatites of the Pajuluoma group (cf. Table 4).

The topaz pegmatites and some barren pegmatites of the northern high-grade terrane show slightly increased concentrations of Li (up to 73 ppm) and Rb (up to 559 ppm), which are probably mainly in muscovite and feldspars, because no Li silicates or phosphates have been discovered.

Tantalum content is usually 0.2-7 ppm in barren pegmatites and clearly increased (30-83 ppm) in fractioned pegmatites and pegmatitic granites in the Pajuluoma group (Table 4). Tantalum is the best pathfinder element of bulk rock samples from pegmatites in exploration of tin (Kärkkäinen 1986). Ta and Nb are the main constituents of columbite-tantalite, tapiolite and microlite, minerals characteristic of beryl pegmatites. Ta and Nb contents can be very high, 100 ppm and 240 ppm, respectively, in some spodumene-bearing dykes of the Pajuluoma group (Alviola 1986a). For comparison, the Kaatiala dyke averages 60 ppm Nb (Nieminen 1978). Tin is the main constituent of cassiterite, but the muscovites of pegmatites may contain several hundred ppm of Sn, even if the dyke has no cassiterite (Alviola 1986b).

Apatite is the most common phosphate mineral in pegmatites, but other phosphates, too, occur in the complex pegmatites (especially Li phosphates if P is available) (Table 3). The phosphorus content tends to increase in pegmatites during fractionation from 0.01 wt% P_2O_5 in barren pegmatites up to 0.73 wt% P_2O_5 in fractioned pegmatites (see subarea A5 in Table 4). An exception is the low P content (0.06 wt% P_2O_5) of the Kaatiala complex pegmatite (Nieminen 1978). This is probably due to the relatively high-grade metamorphic environment, Sil-Ms zone, of the surrounding bedrock. Because of the low phosphorous, lithium formed spodumene instead of Li phosphates in the Kaatiala pegmatite. Fractionated topaz dykes in the north also contain only minor amounts of phosphorous, due to either high-grade metamorphic regime at the time of emplacement or the NYF-type shade of pegmatites.

Beryllium, which mainly occurs in beryl, and bo-

ron, the main element in tourmaline, are common constituents of rare-element pegmatites. In this study beryllium is analysed only from few pegmatites and pegmatitic granites, and concentrations vary from 3 ppm to 36 ppm (Table 4). The Be content in some single Be-enriched dykes in the Seinäjoki group can be as high as 190 ppm (Alviola 1986a). The average contents of boron in the pegmatites of the study region vary over a large range (Table 4).

Interpretation of scattergrams

For scattergrams (X-Y diagrams) and other diagrams the pegmatites (actually the geochemical study areas and subareas) were divided by the metamorphic grade of their country rock into three classes (Table 5):

(1) the pegmatites in the andalusite zone

(2) the pegmatites in the Sil-Ms or Sil-Kfs zones

(3) the pegmatites in an environment of a higher metamorphic grade than the Sil-Kfs zone

The pegmatites were further divided on the basis of their fractionation degree as indicated by the ratio of certain elements into (Table 5) (see Černý 1991):

1) poorly fractionated pegmatites, with high values of Ba/Rb (> 0.5), Sr/Rb (> 0.2) and K/Rb (> 150)

2) moderately to well fractionated pegmatites, with relatively low values of Ba/Rb (0.1-0.5), Sr/Rb (0.05-0.2) and K/Rb (50-150)

Thus, the geochemical study areas and subareas can be divided into six groups, each with its own symbols in scattergrams (Table 5). The aim of this classification is to establish whether there are any regular changes (correlations) in the metamorphic grade of the country rocks or in the fractionation of the pegmatites and the element concentrations of the pegmatites. The study areas can also be divided into those (F, G and H) in a probable ITD-type terrane and those (A-E, I) near Seinäjoki town.

Plots of normative compositions (weigth norm, Fe_2O_3 and FeO after Irvine and Baragar 1971) for the

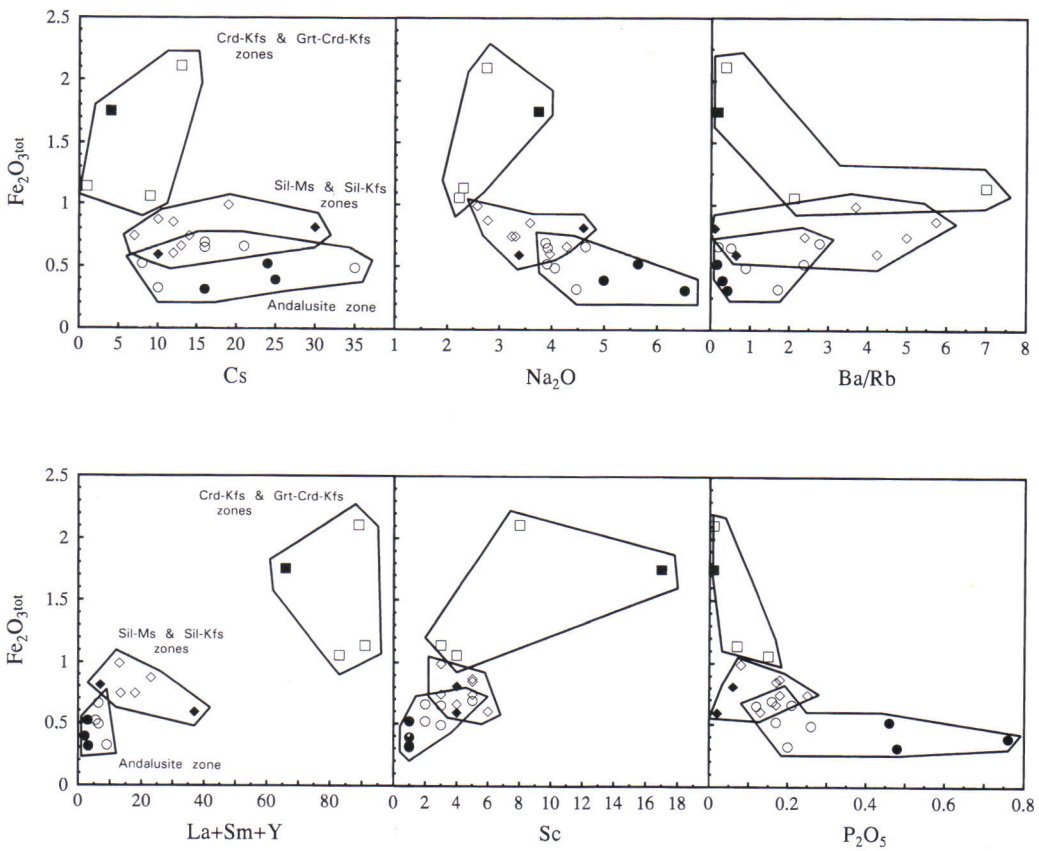


Fig. 11. Scattergrams of various elements vs. Fe_2O_3tot for the pegmatites of the Seinäjoki region. For symbols see Table 5. Pegmatites of the andalusite zone, Sil-Ms and Sil-Kfs zones, and Crd-Kfs and Grt-Crd-Kfs zones are separated by lines.

studied pegmatites within the Q-Ab-Or-An tetrahedron for conditions where $P_{H2O} = P_{total} = 5$ kbar show that they fall close to the cotectic line formed by intersection of cotectic surfaces (Fig. 9). There is a trend that pegmatites of the relatively low grade country rocks are rich in normative albite.

Generally the K/Na ratio of pegmatites decreases as a function of the decrease in metamorphic grade of the country rocks (see K_2O-Na_2O diagram, Fig. 9). Overlapping is due to local differences in fractionation. The Rb content in pegmatites increases during fractionation, but the A/CNK value does not (Fig. 9). The K/Rb ratio of pegmatites is >100.

Many geochemical features of the fractionated pegmatites are seen in the diagrams in which typical minor elements of pegmatites are plotted against the indicators of fractionation.

1) Li, Cs and Ta concentrations, typical of the LCT-type pegmatites, start to increase in the pegmatites when the fractionation indicator, Sr/Rb, of the rock reaches 0.2-0.3 (Fig. 10). Tin, a characteristic element of geochemical study area A, has a similar trend.

2) The pegmatites of the northern areas have elevated Th, Sc, Y+La+Sm, but lower P contents (Fig. 10), as is more or less characteristic of NYF-type pegmatites (cf. Černý 1991b, 1992). The high Th/U and K/P ratios of the northern pegmatites also distinguish them from the other pegmatites (Table 4).

3) P contents have decreased in the fractionated pegmatites of the relatively high-grade metamorphic areas, but increased in the fractionated pegmatites of the andalusite zone (Figs. 10 and 11).

The pegmatites of the andalusite zone, sillimanite zones and higher-grade zones form distinct groups in the plot diagrams of Fe to Cs and Fe to Na (Fig. 11). Iron is high in the pegmatites of the northern areas, and the Ba/Rb ratio (and K/Rb and Sr/Rb, see Table 5) of pegmatites shows moderate fractionation (Fig. 11).

RESULTS AND DISCUSSION

The LCT-type pegmatites of the Seinäjoki region occur within a zone characterized by an andalusite mica schist, but some dykes occur in mica gneisses and granitoids in areas of the sillimanite-muscovite grade. The most fractionated pegmatites in the andalusite zone are generally in the central part of the zone, implying the occurrence of metamorphic subzones (Fig. 8). The characteristic rare-element pegmatite minerals, e.g. beryl, columbite, cassiterite and spodumene, disappear from these pegmatites when the metamorphic grade of the country rocks reaches the sillimanite isograd. Pegmatites in the areas metamorphosed above the second sillimanite isograd are barren. Thus, there is a clear correlation between the distribution of LCT-type rare-element pegmatites and the metamorphic grade of the country rocks.

A pegmatite distribution similar to that in the Seinäjoki region has been reported from many pegmatite fields (e.g. Magrygina 1979, Lahti and Korsman 1990, see also Norton and Redden 1990). However, the Seinäjoki region is exceptional in

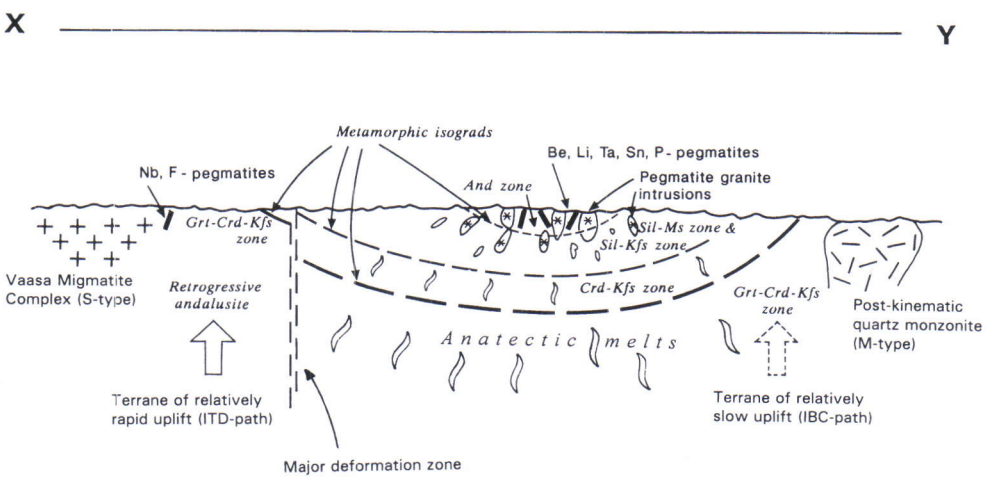


Fig. 12. Schematic north-south-trending cross-section (X-Y, shown in Figure 1) of the upper crust of the Seinäjoki region.

that the topaz pegmatites including columbite are located in high-grade supracrustal rocks. These pegmatites are unique; normally, the volatile components essential for the development of rare-element pegmatites are driven upwards earlier from the dry and hot environment of high-grade metamorphism. The geological environment of the topaz pegmatites differs from that of other rare-element pegmatites in the study region; the northern and southern high-grade terranes have somewhat different cooling paths (Figs. 5 and 12). The topaz pegmatites which slightly resemble NYF (niobium-yttrium-fluorine)-type pegmatites may represent the latest low-solidus melts and volatiles of the anatectic processes that generated the S-type Vaasa Migmatite Complex. The interpretation is consistent with the fact that the concentration of fluorine in the Vaasa Migmatite Complex is much higher than in other parts of the Bothnia Belt or in the Central Finland Granitoid Complex (see Lahermo et al. 1990).

The LCT(lithium-caesium-tantalum)-pegmatite dykes of the Seinäjoki region often occur around the pegmatitic granite intrusions as swarms, indicating that the pegmatites derived from the volatile-rich residual magmas of pegmatitic granites. The scarcity of major granite plutons near these pegmatitic granites and pegmatites implies that these granitic pegmatites have no magmatic derivations. Baker (1998) calculated that pegmatites should only rarely be found more than ca. 10 km from their host pluton (dimensions 10 x 10 x 10 km) and that pegmatites should not be associated with small plutons. Our interpretation of the genesis of the studied LCT-type pegmatites is that they represent low-solidus melts that moved upwards and were formed by anatexis in the adjacent high-grade terrane which continues below the present surface (Fig. 12). Some small granodiorites, including the rare-element pegmatites of the study area, are subhorizontal formations intersected by the pegmatites.

There is a time difference of several tens of millions of years between the peak of regional metamorphism (1.89-1.88 Ga) and cooling of pegmatites (ca. 1.80 Ga) (see also Alviola et al., this volume). The solidus of the pegmatite magma may be lowered by volatiles and other fluxing elements, with the consequence that melts can keep on intruding for long periods in the crust and concentrate upwards to the relatively low metamorphic areas. Regional cooling rate of the Seinäjoki region is not known. Slow regional cooling (1-4 °C/ Ma), often characteristic of Precambrian terranes (e.g. Spear 1993), can contribute to the length of the pegmatite intrusion period. According to Alviola et al. (this volume) pegmatites of the Seinäjoki region belong to a tectonomagmatic episode distinct from the culmination of Svecofennian magmatism.

Generally the whole rock chemical composition of pegmatites, e.g. Fe/Na, correlates with changes in metamorphic grade of the country rock. Overlappings in the correlation, e.g. in Ba/Rb and K/Rb ratios, in the pegmatites are mostly due to local differences in fractionation (see Figs. 9-11). Important elements in the LCT-type pegmatites, such as Li and Ta, show most extensive enrichment when Sr/Rb is smaller than 0.3 (Fig. 10). However, it seems that all element variations cannot be explained solely by fractionation. For example, fractionated pegmatites of the andalusite zone are clearly enriched in P, but similarly fractionated (in terms of Sr/Rb, K/Rb and Ba/Rb) Kaatiala dyke and topaz-bearing pegmatites in areas of relatively high-grade metamorphism are low in P (Fig. 10). In addition, the iron content is low in the fractionated LCT-type pegmatites but high in the fractionated topaz pegmatites and the Kaatiala dyke (Table 4, Fig. 11).

The supracrustal country rocks of the pegmatites studied have elevated Li and Cs (Fig. 8). This indicates that fluids have emanated from the pegmatites during crystallization. The lack of petalite and SQI (spodumene-quartz-intergrowth) in the pegmatites of the Seinäjoki region indicates that lithostatic pressure was moderately high, 2-4 kbar during pegmatite crystallization (Fig. 5) (see also London 1984). The metamorphic grade of the country rocks only roughly equals to the P-T conditions during pegmatite emplacement, because the pegmatites are late- to postmagmatic rocks, postdating the peak of regional metamorphic evolution.

The relationships between pegmatites and their country rocks indicate that in the Seinäjoki region the whole rock chemical and mineral composition in granitic pegmatites is a complex function, of metamorphic grade (e.g. lithostatic pressure) of country rock, source of magma and local differences in magma fractionation.

Our study implies that metapelite -dominated rocks metamorphosed in lower to middle amphibolite facies of the andalusite-sillimanite series are promising for rare-element pegmatite prospecting. The existence of a metamorphic map is a very important tool in choosing the exploration areas, and the bulk rock or mineral chemistry of pegmatites is the best tool in efforts to trace the mineralization process – particularly in areas containing poorly or moderately fractionated pegmatites.

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APPENDIX 1. Brief description of the geochemical study areas and subareas.

Geochemical study area A. A large, slightly N-S trending study area that goes through Seinäjoki town and includes rare-element pegmatites and barren pegmatites of the Pajuluoma group. It is divided into 12 subareas (see Fig. 6):

- *Subarea A1* has two barren pegmatites situated in deformed tonalites in the southern Sil-Kfs zone.
- *Subarea A2* represents barren pegmatites and small pegmatitic granites in the southern part of the southern Sil-Ms zone.
- *Subarea A3* refers to pegmatite dykes and small pegmatitic granites in the northern part of the southern Sil-Ms zone.
- *Subarea A4* consists of beryl pegmatite dykes in the andalusite zone. The subarea resembles subarea A6.
- *Subarea A5* is the Pajuluoma cassiterite pegmatite located in the andalusite zone.
- *Subarea A6* refers to two narrow beryl-columbite-cassiterite pegmatites lying near the Pajuluoma cassiterite dyke in the andalusite zone
- *Subarea A7* is a narrow beryl-cassiterite pegmatite, about 150 m west of the Pajuluoma Sn-dyke.
- *Subarea A8* represents relatively small pegmatitic granite intrusions in the andalusite zone.
- *Subarea A9* is the large Hallilanvuori pegmatitic granite intrusion in the central part of the andalusite zone.
- *Subarea A10* contains pegmatites of the large pegmatitic granite intrusions at Jouppilanvuori and Mykänristinkallio, both in the andalusite zone.
- *Subarea A11* represents pegmatites in the northern part of the andalusite zone.
- *Subarea A12* includes the barren pegmatites in the northern Sil-Ms zone.

Geochemical study area B. This small study area lies 5 km west of Seinäjoki town, and includes barren pegmatites in the southern Sil-Ms zone.

Geochemical study area C is small and lies in the Sil-Ms zone, close to the andalusite zone (Fig. 2).

Geochemical study area D contains barren pegmatites of the andalusite zone 20 km west of study area A.

Geochemical study area E is composed of the Kaatiala pegmatite dyke (Nieminen 1978). The dyke occurs in granodiorite, but near muscovite-bearing mica gneisses (cf. Fig. 3c).

Geochemical study area F lies about 20 km north of Seinäjoki and includes four barren pegmatites across the Crd-Kfs zone. The country rocks of the southernmost pegmatite probably represent the Sil-Kfs zone; the northern dyke is in the Grt-Crd-Kfs zone (Fig. 2).

Geochemical study area G lies in the northwest corner of the study region, at Isokyrö. The country rocks are composed of mica gneisses of the northern Grt-Crd-Kfs zone (Fig. 2). The area is divided into two subareas:

- *Subarea G1* contains topaz pegmatites of Palhojainen, where the country rocks are diatexitic garnet mica gneisses (Fig. 3d).
- *Subarea G2*, about 2 km south of Palhojainen, includes barren pegmatites in mica gneisses.

Geochemical study area H lies in the northeast corner of the study region (Fig. 2), and it, too, is divided into two subareas:

- *Subarea H1* is the topaz-bearing pegmatite of Soromäki surrounded by garnet mica gneisses.
- *Subarea H2* is a narrow, barren pegmatite dyke about 3 km northeast of Soromäki. The country rocks are migmatitic garnet mica gneisses of the Crd-Kfs zone.

Geochemical study area I refers to a large pegmatitic granite intrusion, about 20 km southeast of Seinäjoki town. Country rocks are deformed tonalites and mica gneisses of the southern Sil-Ms zone.

APPENDIX 2. Standard deviations of the chemical analyses from the geochemical study areas and subareas. 0 = standard deviation is very small.

Geochemical study area	A												B	C	D	E	F	G		H		I
Subarea	A1	A2	A3	A4	A5	A6	A7	A8	A9	A10	A11	A12						G1	G2	H1	H2	
Pegmatite type	Dykes	Dykes & pegm. granites	Dykes & pegm. granites	Dykes	Sn-dyke (Paju-luoma)	Be,Sn,Nb-dykes	Be,Sn-dyke	Pegm. granites (small)	Pegm. granite (Hallilanvuori)	Pegm. granites (large)	Dykes	Dykes	Dykes	Dykes	Dykes	Li,Nb,Be-dyke (Kaatiala)	Dykes	Topaz-dyke (Palhojainen)	Dykes	Topaz-dyke (Soromäki)	Dyke	Pegm. granite
Amount of analyses	2	5	16	4	6	2	2	9	6	9	5	9	2	3	4	average	4	2	2	2	1	4
Amount of analysed pegmatites	2	5	16	4	1	2	1	9	1	2	4	8	2	3	4		4	1	2	1	1	1
SiO_2 wt%	0,43	2,98	3,16	1,55	2,91	1,50	0,21	1,53	1,07	2,38	1,57	1,61	0,65	1,11	2,41	-	3,20	0,43	0,11	0,96	-	0,73
TiO_2	0,07	0,02	0,02	0	0,01	0	0,01	0,06	0,04	0,02	0,02	0,05	0	0	0,01	-	0,11	0,08	0,03	0,02	-	0
Al_2O_3	0,15	0,59	1,23	0,99	1,12	0,52	0,80	1,07	0,68	1,21	0,31	0,40	0,15	0,75	0,83	-	0,31	0,39	0,93	0,57	-	0,24
Fe_2O_3tot	0,75	0,14	0,30	0,1	0,13	0,02	0,02	0,09	0,28	0,27	0,07	0,42	0,10	0,12	0,18	-	0,69	0,52	0,41	0,17	-	0,05
MnO	0,02	0,01	0,02	0,08	0,01	0,03	0,01	0,05	0,02	0,07	0,01	0,03	0	0,06	0,01	-	0,04	0,02	0,01	0	-	0,02
MgO	0,22	0,06	0,11	0,06	0,02	0,01	0,01	0,12	0,27	0,10	0,04	0,14	0	0,06	0,05	-	0,15	0,22	0,02	0,01	-	0,05
CaO	0,13	0,47	0,44	0,26	0,16	0,16	0,41	0,14	0,10	0,25	0,24	0,34	0,05	0,18	0,12	-	0,34	0,06	0,23	0,34	-	0,08
Na_2O	0,11	0,62	0,92	1,08	2,35	0,14	0,22	0,59	0,34	1,20	1,31	1,07	0,81	1,05	1,51	-	0,25	0,22	0,24	1,11	-	0,94
K_2O	0,36	1,56	2,10	1,25	2,82	1,57	0,15	0,94	0,93	2,60	0,76	2,13	0,40	1,50	1,70	-	0,64	0,11	1,83	1,63	-	0,73
P_2O_5	0,03	0,12	0,02	0,13	0,17	0,09	0,32	0,18	0,03	0,08	0,12	0,04	0,02	0,03	0,06	-	0,04	0	0	0	-	0,03
Rb ppm	1	37	48	186	183	103	77	149	30	55	51	72	14	-	35	-	103	31	121	165	-	18
Ba	501	673	580	10	41	1	24	27	26	304	262	507	220	34	425	-	407	21	36	145	-	-
Sr	62	123	79	24	7	7	31	7	19	-	39	78	30	-	80	-	57	7	7	1	-	10
Li	17	-	33	27	10	6	4	12	17	-	7	6	-	-	13	-	8	15	9	2	-	-
Cs	7	3	4	13	16	14	8	26	11	8	10	5	5	7	3	-	9	0	0	5	-	4
Ta	0	-	1	3	66	46	78	51	29	-	0	1	-	-	2	-	11	1	1	1	-	-
U	15	2	9	6	6	23	12	21	4	7	8	5	1	1	9	-	4	2	1	0	-	2
Th	3	1	1	1	0	0	0	1	0	1	1	1	-	0	0	-	34	9	8	3	-	0
La	0	0	2	2	0	0	0	3	-	-	2	3	-	-	1	-	76	8	10	5	-	-
Sm	2	0	2	0	0	0	0	0	-	-	1	1	-	-	0	-	6	3	1	1	-	-
Y	1	11	6	3	0	0	0	1	-	-	3	5	-	-	1	-	5	8	1	4	-	-
Sc	2	3	3	1	1	0	0	1	1	3	1	2	1	1	0	-	2	5	0	2	-	1
Sn	7	2	4	13	3197	4	-	66	4	5	3	5	-	6	5	-	-	1	-	-	-	2
As	0	13	46	34	1	2	11	58	114	172	30	2	3	215	4	-	1	-	0	0	-	12
B	-	17	230	-	-	-	-	60	1299	658	-	-	-	803	-	-	-	-	-	-	-	130
Be	-	-	-	-	-	-	-	43	1	1	-	-	-	10	-	-	-	-	-	-	-	-
Zr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	10	-	-	45	81	-	-	-
Host rock	Tonalite	Mica gn.	Mica gn.	Mica sch.	Mica sch.	Mica sch.	Mica sch.	Graywac.	Mica sch.	Mica sch.	Graywac.	Mica gn.	Mica gn.	Mica gn.	Graywac.	Granod.	Mica gn.	Mica gn.	Mica gn.	Mica gn.	Mica gn.	Mica gn.
Metamorphic zone	Sil-Kfs (southern)	Sil-Ms (S-part of southern)	Sil-Ms (N-part of southern)	And	And	And	And	And	And	And	And	Sil-Ms (northern)	Sil-Ms (southern)	Sil-Ms (southern)	And	Sil-Ms	Crd-Kfs (northern)	Grt-Crd-Kfs (northern)	Grt-Crd Kfs (northern)	Sil-Kfs (northern)	Crd-Kfs (northern)	Sil-Ms (southern)

Svecofennian granitic pegmatites (1.86-1.79 Ga) and quartz monzonite (1.87 Ga),
and their metamorphic environment in the Seinäjoki region, western Finland
Edited by Hannu Mäkitie
Geological Survey of Finland, Special Paper 30, 61-98.

THE FAYALITE-AUGITE QUARTZ MONZONITE (1.87 GA) OF LUOPA, WESTERN FINLAND, AND ITS CONTACT AUREOLE

by

Hannu Mäkitie & Seppo I. Lahti

Mäkitie, Hannu & Lahti, Seppo I. 2001. The fayalite-augite quartz monzonite (1.87 Ga) of Luopa, western Finland, and its contact aureole. *Geological Survey of Finland, Special Paper 30*, 61–98, 19 figures, 8 tables and 5 appendices.

The Svecofennian quartz monzonite intrusion (1872±2 Ma) of Luopa, western Finland, forms an almost undeformed irregular stock which occupies an area of about 35 km^2. The majority (95%) of the intrusion is composed of porphyritic, dark green quartz monzonite but, also, porphyritic light-coloured quartz monzonite varieties are locally observed. There are no intrusive contacts between the two quartz monzonite varieties. The dark green variety is often strongly weathered. The main minerals of the quartz monzonite are K-feldspar (as megacrysts), andesine, hornblende, quartz and locally biotite. Two mineral assemblages are found in the dark green variety, one is fayalite-augite-hornblende and the other augite-hornblende-biotite. In the light-coloured quartz monzonite, pyroxene and fayalite are practically absent.

Ferromagnesian silicates have high Fe/Mg. The SiO_2 content of the Luopa quartz monzonite is 57-63 wt%. Fe content (6-13 wt% Fe_2O_3tot) as well as X_{Fe} value (0.9) are high. Rb/Sr is low, 0.15-0.25. The REE distribution curve of the light-coloured quartz monzonite variety is slightly steeper than that of the dark green quartz monzonites. The quartz monzonites are metaluminous, have I-type characteristics and trace element contents similar to within plate granites. The quartz monzonite crystallized from iron-rich magma and was emplaced at a relatively shallow crustal level. The intrusion cross-cuts foliated tonalites and brecciates pyroxene granite; therefore it was emplaced after main stage of crustal thickening and is thus post-kinematic.

The emplacement of the quartz monzonite caused a narrow high-grade contact metamorphic aureole which may be divided into an inner granulite-grade aureole (width 2 m) and wider outer aureole (width 2-30 m) of slightly lower metamorphic grade. The contact aureole overprints 1.89-1.88 Ga upper-amphibolite facies regional metamorphism. Orthopyroxene-bearing mineral assemblages and sillimanite-hercynite-cordierite symplectites were formed in the diatexitic metapelites and garnet-orthopyroxene pairs in the nebulitic intermediate rocks of the inner aureole. The outer contact aureole is composed of diatexitic metapelites containing garnet, cordierite and/or sillimanite, and some mica gneisses that became tightly banded during ballooning of the intrusion. The tonalites of the inner aureole are characterized by the presence of orthopyroxene and the total dehydration of hornblende.

The X_{Mg} value of garnet in the metapelites of the inner aureole is 0.24-0.28 and the average TiO_2 content of biotite is 4.2 wt%. Thermometric temperature calculations imply contact metamorphism of at least 710 °C. Retrogressive andalusite and staurolite as well as garnet rims around orthopyroxene were formed in the contact aureole during cooling.

Key words (GeoRef Thesaurus, AGI): Quartz monzonite, granites, contact metamorphism, metapelite, fayalite, pyroxene, chemical composition, electron probe data, Proterozoic, Luopa, Ilmajoki, Finland

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CONTENTS

INTRODUCTION

Pyroxene-bearing granitoid intrusions form a noteworthy and exceptional part of the Paleoproterozoic, Svecofennian rocks of Finland. Pyroxene granitoids occur in the tectonically complex Raahe-Ladoga zone and within the Central Finland Granitoid Complex (CFGC) (Wahl 1963, Lahti 1995, Elliott et al. 1998, Nironen et al. 2000). They are often situated in areas of high-grade regional metamorphism and it seems that their intrusion has caused significant contact metamorphic aureoles (Hölttä 1995). The pyroxene-bearing granitoids are mostly post-kinematic and they vary in composition from tonalitic to granitic and syenitic (e.g. Nurmi and Haapala 1986, Rämö et al. 1999).

Many interesting questions arise on the origin and source of these intrusions, and on their relationship to the adjacent high-grade metamorphic areas. For example, do the protoliths of the intrusions represent parts of the deeper continental crust, subduction zones, or mantle, and can their contact aureoles provide a method for tracking the tectonothermal evolution of metamorphic belts? The sites of the Finnish Svecofennian pyroxene granitoids are relatively well known (e.g. Rämö et al. 1999), however, detailed studies of individual intrusions are mostly lacking – except for the well studied pyroxene granitoid of Vaaraslahti in eastern Finland (cf. Hölttä 1995, Lahti 1995).

The fayalite-augite quartz monzonite of Luopa (Figs. 1 and 2a) is located at the village of Luopa in the municipality of Ilmajoki, western Finland, about 70 km southeast of the town of Vaasa. The rock has been poorly studied: short descriptions have been given by Väyrynen (1923, 1936), Saksela (1934), Laitakari (1942), Alviola (1989), Lahti and Mäkitie (1990), Mäkitie (1999), Mäkitie and Lahti (1991), Mäkitie et al. (1991, 1999), Elliott et al. (1998) and Nironen et al. (2000).

For this study the rocks of the Luopa intrusion and of its surroundings were studied in detail. Moreover, the contact metamorphism of the intrusion was examined.

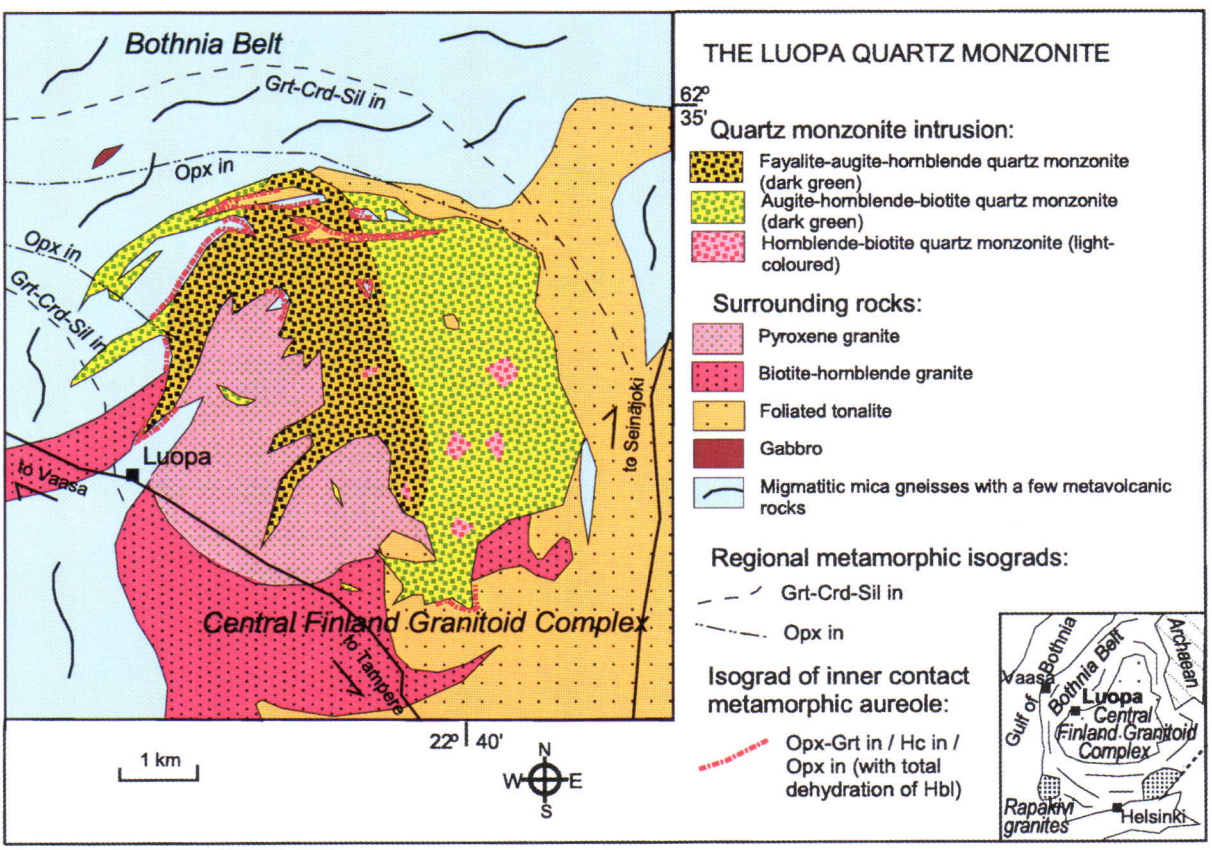


Fig. 1. Simplified geological map of the Luopa quartz monzonite. Map modified after Lahti and Mäkitie (1990) and Mäkitie et al. (1991).

GEOLOGICAL SETTING

The Luopa quartz monzonite occurs at the boundary of the CFGC and the Bothnia Belt (BB) within the accretionary arc complex of central and western Finland (Lahti and Mäkitie 1990, Korsman et al. 1997) (Fig. 1). The intrusion is composed of two types of quartz monzonite: a dark green type (dominant) and a light-coloured type (Fig. 2b).

The country rocks around the stock comprise migmatitic garnet-cordierite-sillimanite mica gneisses of the BB, equigranular foliated tonalites and a porphyritic, locally pyroxene-bearing granite of the CFGC. The boundary between the pyroxene granite and a biotite-hornblende granite is not exposed.

The regional metamorphic grade in the region is

a)

b)

Fig. 2. Samples of the Luopa quartz monzonite. **(a)** An outcrop surface of the porphyritic quartz monzonite; **(b)** a hand specimen of a partly dark green (right) and partly light-coloured (left) quartz monzonite. Note that both types have similar texture and that ideal intrusive contact between them is questionable.

transitional between upper amphibolite facies and granulite facies. Metamorphic grade increases to the southwest and the intrusion lies in the area of highest grade metamorphism (Mäkitie and Lahti 1991). Supracrustal rocks in the Luopa area are polyphasically deformed (Mäkitie 1999).

The Luopa quartz monzonite has previously been dated by the U-Pb zircon method at 1871±1 Ma (Mäkitie and Lahti 1991). A U-Pb zircon age of a tonalite located 4.5 km east of the Luopa quartz monzonite is 1882±9 Ma, and this tonalite intruded during the regional metamorphic peak (Mäkitie and Lahti 1991, see also Vaasjoki 1996). A U-Pb monazite age of 1857 Ma reported from migmatitic garnet-cordierite-sillimanite mica gneiss about 1.5 km northwest of the Luopa intrusion indicates either regional cooling of the crust below 600-650 °C or the age of contact metamorphism (see Mäkitie and Lahti 1991).

On aeromagnetic maps published by the Geological Survey of Finland (GTK), the stock generally appears as a negative anomaly whereas on a gravimetric map (Kiviniemi 1980) the stock and its surroundings are indicated by a weak positive anomaly.

QUARTZ MONZONITE

Modal composition

Because of the large variation in the proportions of the felsic minerals (determined by point-counting thin sections from coarse-grained samples) the composition points of the Luopa quartz monzonite are widely-scattered in granitoid classification diagrams, but the average composition of the rock plots on the quartz monzonite field (Fig. 3 and Table 1).

Quartz monzonites do not belong to the granite family in the classification of plutonic rocks by Streckeisen (1976) because quartz monzonites do not contain enough quartz (> 20 % of felsic minerals). Moreover, they do not belong to the granitoids, because quartz is not an essential (>20 %) component (Bates and Jackson 1980). Strict terminology and classification is important when comparing, for example, numerical data of various granitoids or granitic rocks. However, terms 'granitoid' and 'granite' are often used inaccurately and in this study the rock is classed as a granitoid.

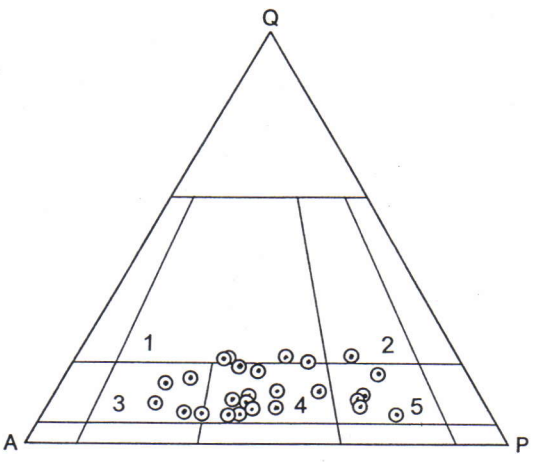


Fig. 3. Proportion of plagioclase, quartz and K-feldspar of the quartz monzonite plotted on the A-Q-P diagram by Streckeisen (1976). The fields in the diagram are as following: 1=granite, 2=granodiorite, 3=quartz syenite, 4=quartz monzonite, 5=quartz monzodiorite and quartz monzogabbro. ⊙= dark green or light-coloured quartz monzonite.

Table 1. Average mineral modal composition of the dark green and light-coloured quartz monzonite varieties. 'Others' are sericite, chlorite, zircon, apatite, orthopyroxene, allanite and tremolite-actinolite.

	Dark green type	Light-coloured type
Plagioclase	34,2	35,6
Quartz	7,1	15,5
K-feldspar	37,2	26,7
Biotite	2,1	8,5
Hornblende	13,6	12,8
Fayalite	1,6	0,1
Augite	1,5	0,1
Serpentine	0,3	0,1
Opaque	2,0	0,3
Others	0,4	0,3
Total	100,0	100,0

Structure and emplacement

Form

The Luopa quartz monzonite forms a sub-circular, irregular stock occupying an area of ca. 35 km^2, and has a diameter of about 6 km (Fig. 1). Wide (5-100 m), locally apophysic dykes branch from the main body in the northwestern part of the intrusion. These dykes are considered to be indicative of relatively shallow level of emplacement into relatively cool and rigid crustal rocks. Narrow aplitic dykes, representing the latest magma phases, are locally characteristic of the contact areas (Fig. 4a).

The quartz monzonite appears to be undeformed (Figs. 2a and 2b) although there are sporadic narrow (1-3 cm) ductile shear zones. In places, the marginal part (up to some 15 m) of the intrusion is weakly deformed parallel to contacts, especially against the metapelites (Fig. 4a). This deformation is due to compression during 'ballooning' of the magma: the earlier crystallized outer part of the stock was pushed against the rigid country rocks during the emplacement and crystallization of the final magma phases. Elsewhere, the observed orientation of K-feldspar megacrysts indicates magmatic flow.

The dips of the contacts are usually gentle in the central part of the stock. In other parts, the dips are steep.

Contact relationships and structural setting

The porphyritic pyroxene granite (where clinopyroxene is more common than orthopyroxene near quartz monzonite) country rock is brecciated by the quartz monzonite into angular fragments (Fig. 4b). Between the fragments, the magmatic foliation of quartz monzonite is well-preserved. The quartz monzonite also intersects the trend of the foliation and elongation of enclaves in the surrounding tonalites (Fig. 4c). This foliation in tonalites was formed during the main deformation phase (D_2) of the region (Mäkitie 1999) and, thus, the quartz monzonite is post-kinematic in regard to the main phase of deformation in the region.

The contact between the quartz monzonite and supracrustal country rocks is sharp, and characterized by a narrow (5-50 cm) zone of garnet- and biotite-rich coarse-grained granitic rock. Texturally, there are two types of metapelite at the contacts of the quartz monzonite, diatexitic metapelites and tightly banded mica gneisses (Figs. 4d, 5a and 5b). The diatexitic metapelites were formed during intense melting caused by thermal metamorphism. However, these diatexites still contain some remnants of supracrustal intermediate rocks. The deformation textures in the tightly banded mica gneisses are due to compression caused by ballooning of the stock. Foliation of these mica gneisses is parallel to the contact. Most commonly the tightly banded mica gneisses occur between parallel quartz monzonite dykes in the outermost part of the stock.

Some supracrustal rocks at contacts are nebulitic migmatites (Fig. 5c) containing inclusions from metavolcanic rocks and metapelites in the partially melted, compositionally intermediate matrix. Moreover, the quartz monzonite sharply intersects some layered, intermediate metavolcanic rocks, which are relatively resistant to metamorphism.

There is a tectonic contact trending 340° between quartz monzonite and adjacent tonalite on the northeastern side of the intrusion that is visible also in aeromagnetic map. This deformation zone is about one metre wide and is composed of intensively deformed, locally mylonitic rocks (Fig. 4e).

The different types of the contacts between the quartz monzonite and the adjacent rocks result mostly from differences in the competency of the country rocks; incompetent, plastic metapelites are more easily deformed than the competent granitoids and metavolcanites. Typically, the latter rocks are either broken into angular fragments or cut sharply by the stock.

Petrography and mineralogy

Dark green quartz monzonite

The predominant rock type (95 %) of the Luopa stock consists of homogeneous, coarse-grained, porphyritic quartz monzonite. The colour of the rock is dark green on a fresh surface but on weathered surfaces light brown (Figs. 2a and 2b). The grain size in the intrusion is fairly constant, but some medium-grained varieties are found in the northern part of the intrusion. Enclaves are rare and, where observed, composed of plagioclase and augite. The quartz monzonite is locally strongly weathered and covered by saprolitic layers.

The main minerals of the rock are andesine, K-feldspar, quartz, hornblende, augite, fayalite and biotite (Table 1). There are two different varieties of dark green quartz monzonite (cf. Fig. 1): one is characterized by fayalite-augite-hornblende and the other by augite-hornblende-biotite. No separation between these two varie-

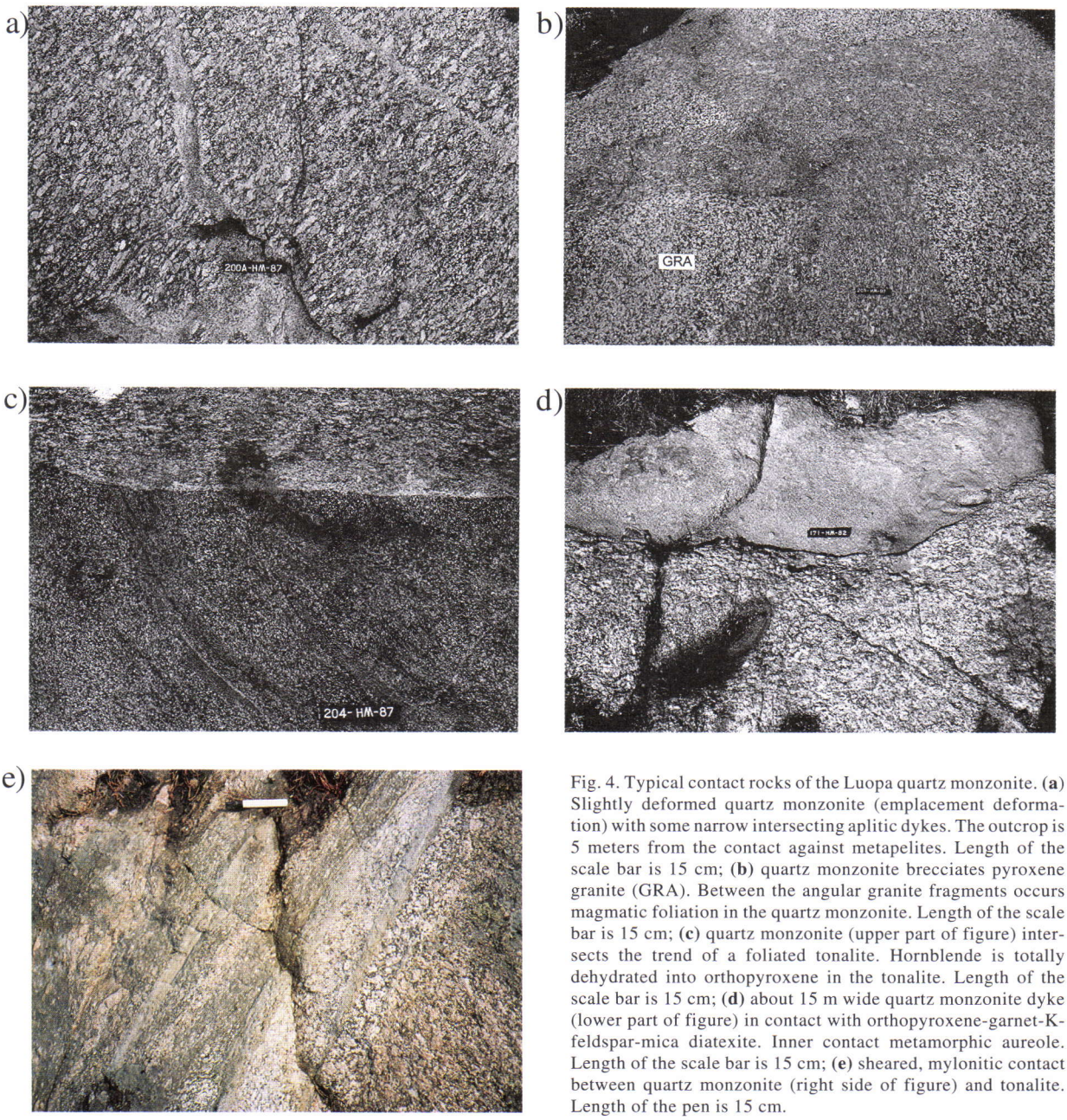


Fig. 4. Typical contact rocks of the Luopa quartz monzonite. **(a)** Slightly deformed quartz monzonite (emplacement deformation) with some narrow intersecting aplitic dykes. The outcrop is 5 meters from the contact against metapelites. Length of the scale bar is 15 cm; **(b)** quartz monzonite brecciates pyroxene granite (GRA). Between the angular granite fragments occurs magmatic foliation in the quartz monzonite. Length of the scale bar is 15 cm; **(c)** quartz monzonite (upper part of figure) intersects the trend of a foliated tonalite. Hornblende is totally dehydrated into orthopyroxene in the tonalite. Length of the scale bar is 15 cm; **(d)** about 15 m wide quartz monzonite dyke (lower part of figure) in contact with orthopyroxene-garnet-K-feldspar-mica diatexite. Inner contact metamorphic aureole. Length of the scale bar is 15 cm; **(e)** sheared, mylonitic contact between quartz monzonite (right side of figure) and tonalite. Length of the pen is 15 cm.

ties is made in the following petrographic and chemical descriptions. Common accessory minerals are allanite, zircon, apatite, ilmenite and black fayalite. Minor amounts of orthopyroxene, sulphides, retrogressive garnet, serpentine and chlorite have been detected.

Microscopically the Luopa quartz monzonite is inequigranular, porphyritic and hypidiomorphic (Fig. 6a). Straight grain boundaries and triple junctions between even-grained plagioclase grains are common. Poikilitic texture is rare, but a few quartz inclusions occur in mafic minerals.

Feldspars and quartz. K-feldspar occurs as coarse-grained (Ø 1-5 cm) megacrysts, which may be macroscopically zoned. X-ray diffraction studies indicate that alkali feldspars from the dark green quartz monzonite are transitional types between intermediate microcline and orthoclase (Fig. 7). The crystals are cryptoperthitic, but may contain microscopic plagioclase laths accreted on growth surfaces, and narrow microperthite strings (Fig. 6b). Often K-feldspar does not contain inclusions, which probably indicates that the mineral crystallized early. K-feldspar megacrysts are mostly untwinned. Myrmekite between K-feldspar and plagioclase is rare.

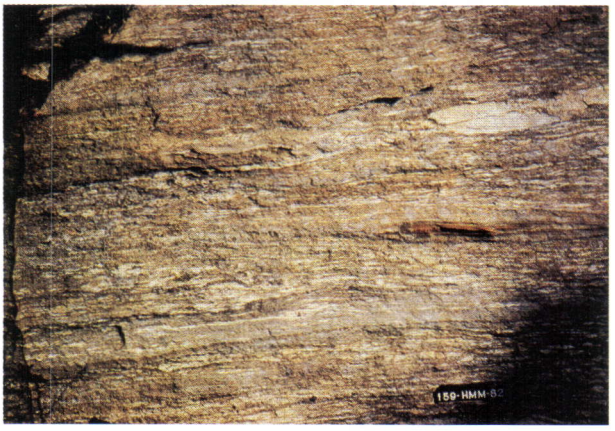
Plagioclase (Ø 0.1-3.0 mm) is polysynthetically twinned andesine (Table 2). The crystals are often optically zoned, and thus their composition probably slightly varies. In places, the crystal cores contain fine-grained saussurite or sericite as an alteration product. The molecular An % for plagioclase, calculated from whole rock chemical analyses, is about 32.

Quartz occurs as small interstitial, xenomorphic grey crystals (Ø 0.01-0.5 mm) and as irregular inclusions in ferromagnesian minerals, indicating that it partly crystallized in the reaction processes of mafic silicate minerals.

Olivine and allanite. Olivine occurs as rounded xenomorphic grains (Ø 0.1-0.3 mm) (Fig. 6a). They are often surrounded by hornblende resembling a corona texture, and occasionally by augite or biotite when

a)

b)



c)

Fig. 5. Rocks in the contact aureole of the Luopa quartz monzonite. **(a)** Garnet-cordierite-mica diatexite in the outer contact metamorphic aureole, about 5 m from the quartz monzonite. Length of the scale bar is 15 cm; **(b)** tightly banded garnet-cordierite-sillimanite mica gneiss near the contact. Compression induced deformation, about 3 m from the contact, is due to the ballooning of the quartz monzonite magma. Outer contact metamorphic aureole. The length of the scale bar is 15 cm; **(c)** nebulitic intermediate orthopyroxene-garnet migmatite with remnants of metavolcanic rocks, located in the inner contact metamorphic aureole about 1 m from the quartz monzonite. Coin 3 cm in diameter.

Table 2. Microprobe analyses of minerals in the Luopa quartz monzonite. Mineral abbreviations after Kretz (1983).

Mineral	Fa	Fa	Fa(opaq.)	(Fa(opaq.)	Opx	Opx	Aug	Aug	Kfs	Pl	Ilm	Hbl	Hbl
Sample	167-HMM	167-HMM	901B	901B	167-HMM	164-HMM	167-HMM	164-HMM	167-HMM	167-HMM	901B	167-HMM	901B
SiO_2 wt%	30,9	31,0	29,8	29,8	48,3	46,7	49,9	49,0	64,8	60,8	0,0	41,0	38,7
TiO_2	0,0	0,0	0,0	0,0	0,1	0,1	0,1	0,2	0,0	0,0	48,6	1,8	2,1
Al_2O_3	0,0	0,0	0,0	0,0	0,3	0,3	0,8	0,8	18,8	25,2	0,0	10,9	11,2
FeOtot	66,0	65,9	61,5	63,1	45,7	46,0	24,8	25,5	0,1	0,1	46,3	28,4	28,1
MnO	1,7	1,7	1,4	1,4	1,4	1,4	0,7	0,7	0,0	0,0	0,8	0,4	0,3
MgO	1,8	2,0	1,4	1,4	4,5	4,4	4,1	3,5	0,0	0,0	0,1	3,0	2,2
CaO	0,0	0,1	0,0	0,0	0,8	0,8	19,8	20,0	0,1	6,6	0,0	9,9	10,4
Na_2O	0,0	0,1	0,0	0,0	0,1	0,1	0,3	0,3	1,9	7,9	0,0	1,8	1,5
K_2O	0,0	0,0	0,0	0,0	0,0	0,0	0,0	0,0	13,4	0,3	0,0	1,5	1,7
Total	100,5	100,8	94,2	95,8	101,2	99,8	100,5	100,0	99,1	100,9	95,8	98,7	96,2
Si	1,02	1,02	1,04	1,03	2,01	1,98	1,99	1,98	2,99	2,69	0,00	6,47	6,31
Al	0,00	0,00	0,00	0,00	0,02	0,01	0,04	0,02	1,02	1,31	0,00	2,03	2,15
Ti	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,01	0,00	0,00	0,97	0,21	0,26
Mg	0,09	0,10	0,07	0,07	0,28	0,28	0,24	0,21	0,00	0,00	0,00	0,71	0,54
Fe	1,82	1,81	1,80	1,82	1,59	1,63	0,83	0,86	0,00	0,00	1,03	3,75	3,83
Mn	0,05	0,05	0,04	0,04	0,05	0,04	0,02	0,02	0,00	0,00	0,02	0,05	0,04
Ca	0,00	0,00	0,00	0,00	0,04	0,04	0,85	0,87	0,01	0,31	0,00	1,67	1,82
Na	0,00	0,01	0,00	0,00	0,01	0,01	0,02	0,02	0,17	0,68	0,00	0,55	0,47
K	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,79	0,02	0,00	0,30	0,35

Table 3. Microprobe analyses across (width175 μm) an elongated allanite crystal.

Mineral	Aln	Aln	Aln	Aln	Aln	Aln	Aln	Aln	Aln	Aln
Sample	2	4	6	8	10	12	14	16	18	20
SiO_2 wt%	28,3	28,3	28,4	28,6	28,3	28,6	28,6	28,6	28,7	28,5
TiO_2	1,8	1,8	1,5	1,7	1,7	1,7	1,7	1,7	1,6	1,5
Al_2O_3	11,1	11,1	11,3	11,5	11,1	11,3	11,2	11,5	11,6	11,4
FeOtot	11,4	11,2	11,1	11,0	11,2	11,5	11,6	11,3	11,0	11,1
MnO	1,0	1,0	1,1	1,0	1,1	1,0	1,0	1,0	1,1	1,2
MgO	0,1	0,1	0,2	0,2	0,1	0,1	0,2	0,1	0,2	0,1
CaO	8,4	8,0	8,2	8,5	8,3	8,6	8,5	8,4	8,3	8,5
Na_2O	0,2	0,2	0,2	0,2	0,2	0,2	0,2	0,2	0,2	0,2
K_2O	0,0	0,0	0,0	0,0	0,0	0,1	0,1	0,0	0,0	0,0
P_2O_5	0,2	0,1	0,2	0,2	0,2	0,2	0,1	0,2	0,2	0,2
ThO_2	0,5	0,6	0,5	0,3	0,4	0,5	0,4	0,3	0,4	0,4
U_2O_3	0,2	0,0	0,1	0,0	0,0	0,0	0,0	0,0	0,0	0,1
La_2O_3	6,5	6,6	6,5	6,5	6,5	6,6	6,6	6,4	6,5	6,3
Ce_2O_3	15,5	16,1	16	15,5	15,9	15,6	15,5	15,6	16,1	16,1
Nd_2O_3	4,5	4,4	4,6	4,5	4,4	4,3	4,5	4,4	4,4	4,4
Total	89,7	89,5	89,9	89,7	89,4	90,3	90,2	89,7	90,3	90,0

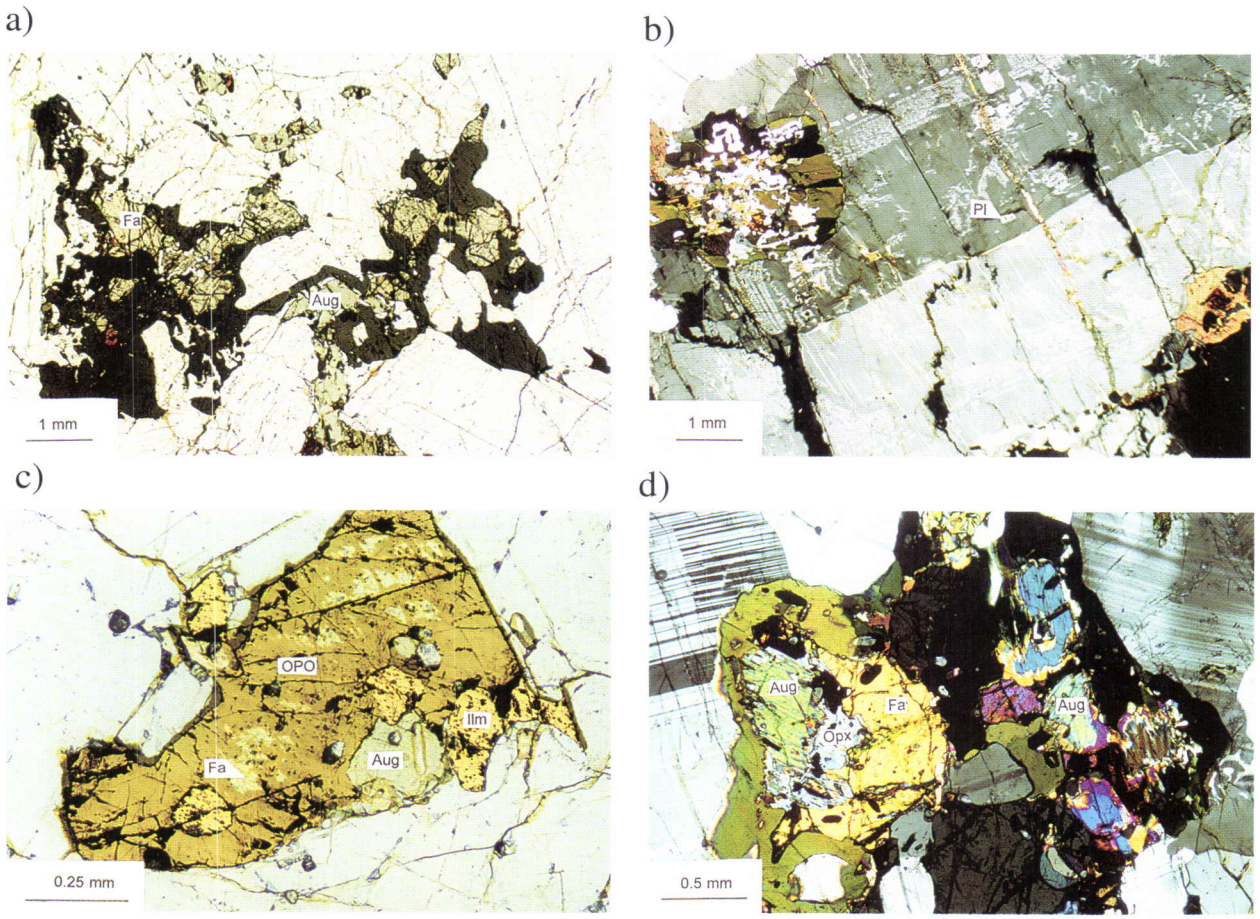


Fig. 6. Microtextures of the Luopa quartz monzonite. **(a)** Undeformed quartz monzonite, where fayalite (Fa) and augite (Aug) are surrounded by hornblende; **(b)** zoned megacryst of K-feldspar with some laths of plagioclase (Pl) accreted on grown surfaces. Light linear lines are perthite exsolutions. Fayalite reacted to hornblende, pyroxene, and quartz intergrowths (left side of photomicrograph); **(c)** grain of black altered fayalite (OPO), 'opaque' in appearance, containing light relicts of unaltered fayalite (Fa). Other minerals are ilmenite (Ilm) and augite (Aug); **(d)** zoned augite (Aug) crystals (zones distinguished by variation in birefringence). Orthopyroxene (Opx) occurs between fayalite (Fa) and augite (Aug). Hornblende surrounds these minerals.

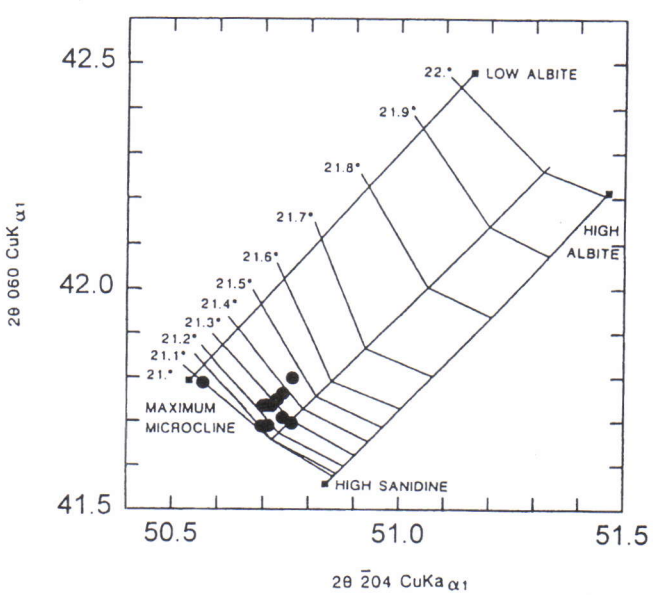


Fig. 7. The structural state of alkali feldspar from dark green quartz monzonite. The reflections (2θ204 and 060, CuKa$_1$ - peaks) were measured from the X-ray powder diffractograms and plotted on the diagram by Wright (1968). ● = K-feldspar from dark green monzonite.

the mineral is in contact with quartz. These corona-like textures are mostly due to reactions between fayalite, plagioclase, quartz and H_2O. Some olivine has altered to yellowish serpentine and to brownish, unidentified mineral aggregates. Olivine contains inclusions of zircon, apatite, allanite and opaque minerals.

Olivine is fayalite (Fo_0-Fo_{10}, see Table 2), and often altered to black 'opaque' material (Fig. 6c). The chemical composition of this material resembles olivine, but the totals of the main elements are only 94-96 % (Table 2), indicating that the mineral must contain, in addition to olivine, some hydrous or hydrated alteration products.

Allanite-Ce occurs as small prismatic crystals. The mineral is often metamict and optically almost isotropic. Allanite crystals are often surrounded by a fan of microfractures. The mode of allanite in the quartz monzonite varies within the intrusion from 0-2 %.

The allanite studied contains 15.5-16.1 wt % of Ce_2O_3, about 6.5 wt % of La_2O_3 and about 4.5 wt % of Nd_2O_3 (Table 3). The mineral is weakly zoned in terms of e.g. Ce_2O_3 and ThO_2.

Pyroxenes. Clinopyroxene (Ø 0.1-0.3 mm) is the most common pyroxene in the rock. It occurs as anhedral crystals often coexisting with olivine, orthopyroxene and hornblende. Locally, clinopyroxene and hornblende occur as worm-like intergrowths. Some clinopyroxene exhibit narrow twin lamellae. The pleochroic colour of clinopyroxene is pale green and, in places, variations in its birefringence indicate weak zoning (Fig. 6d). Narrow tremolitic zones between clinopyroxene and plagioclase occur sporadically.

Chemical analyses of clinopyroxene in the Luopa intrusion, given in Table 2, indicate that the mineral is augite or ferroaugite in composition.

Rare orthopyroxene occurs as reaction rims either between fayalite and quartz or around those augites of greater birefringence. The mineral has high Fe/Mg and is compositionally equivalent to orthoferrosilite (Table 2). The orthopyroxene ($FeSiO_3$) in the aforementioned rim is a reaction product after olivine (gave Fe) and quartz (SiO_2).

Amphibole and biotite. Hornblende grains, 0.2-1 mm in diameter, are irregular and their surface is often ragged. They often surround olivine and augite apparently in part replacing these minerals (Fig. 6a). The pleochroic colour (γ) of hornblende is brownish green to dark green.

Hornblende has high Fe/Mg -ratios (Table 2). The Na_2O content of the mineral is 1.5-1.8 wt%, less than that in typical alkali amphiboles. The microprobe analyses are consistent with an old wet chemical hornblende analysis reported by Väyrynen (1923, p. 33) from the Luopa intrusion. The analytical data indicate that hornblende is barkevikitic and rich in iron.

The amount of biotite is variable in the two dark green quartz monzonite varieties and, locally, may be absent. The biotite flakes are 0.2-2 mm long and their pleochroic colour (γ) is reddish brown.

Oxides and sulphides. The mode of opaque minerals is relatively high, from 2% to 3%, and grain size varies between 0.1 and 1 mm (Fig. 6c). The most common oxide is rounded or irregular grains of ilmenite (TiO_2 content of 48.5 wt %, Table 2), and magnetite is extremely rare. Chalcopyrite and pyrrhotite were observed in a few samples, and molybdenite occurs sporadically at the contacts between pyroxene granite and quartz monzonite.

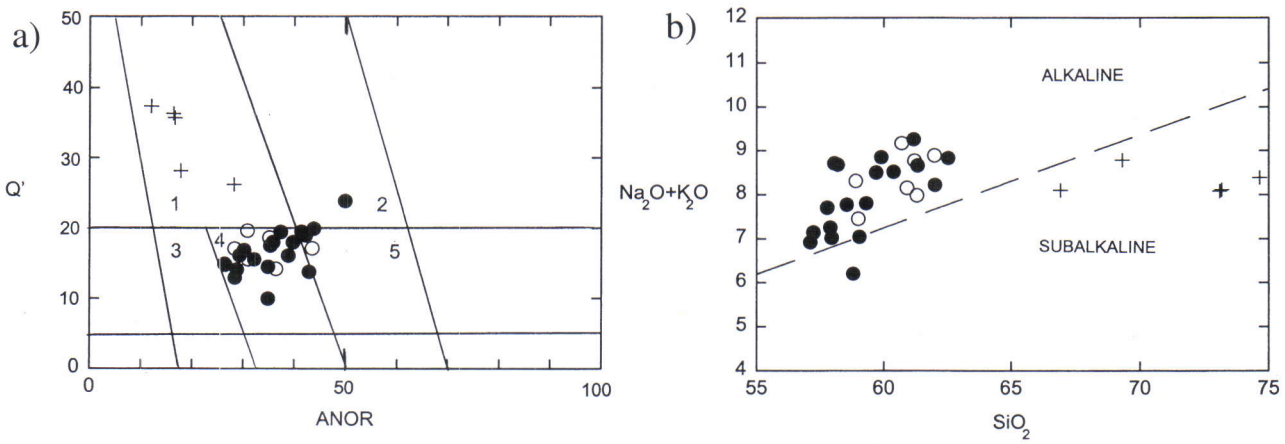
Other minerals. Zircon crystals are idio- to subidiomorphic, with cross section aspect ratio of about 1:4. Apatite is a common accessory mineral. Garnet, sericite and chlorite were detected in those rock samples that underwent retrogression and ductile deformation. Biotite sometimes occurs as a reaction product around garnet.

Light-coloured quartz monzonite

The dark green quartz monzonite occasionally (about 5% of the stock area) has a patchy appearance, chancing into a light-coloured variety (Fig. 2b). The textures of these quartz monzonite varieties are similar, and do not show much textural variation at the contact between the two types.

The main minerals of the light-coloured quartz monzonite are plagioclase, K-feldspar, quartz, biotite and hornblende. The quartz content in the rock is usually slightly higher than in the dark green rock, and thus it is more granitic in composition than the dark green rock variety (Table 1). The accessory minerals are augite, allanite, zircon, apatite, garnet and ilmenite. Fayalite is very rare. The main chemical compositions of both colour varieties of the quartz monzonite are rather similar (Table 4). The light-coloured quartz monzonite does not display as intense weathering as the dark green variety. Colour differences between the dark green and light-coloured quartz monzonites in the Luopa intrusion are predominantly due to the change in the colour of feldspars. Variations of trace Fe^{2+} and Fe^{3+} in feldspars have notable effects to the colour

of K-feldspar and plagioclase (Deer et al. 1965). In general, the pyroxene granite has a higher quartz mode and a smaller grain size than the quartz monzonite, and, locally, it can be difficult to recognize from the light-coloured quartz monzonite.



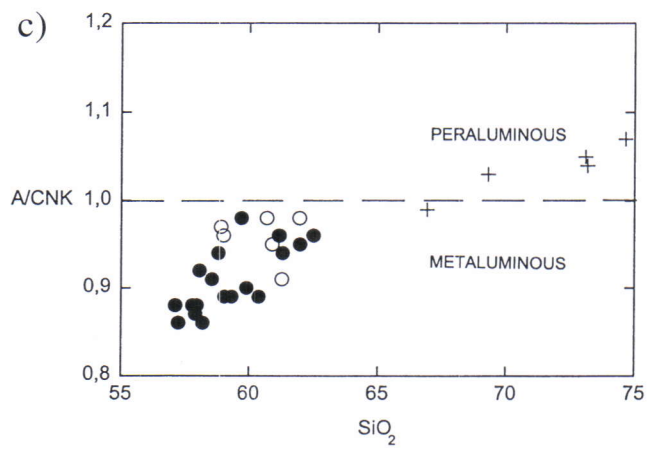


Fig. 8. Chemical classification diagrams of quartz monzonite and pyroxene granite. **(a)** Chemical composition of quartz monzonite and pyroxene granite samples plotted on normative Q'- ANOR diagram containing rock classification fields by Streckeisen and LeMaitre (1979). The fields are as following: 1=granite, 2=granodiorite, 3=quartz syenite, 4=quartz monzonite, 5= quartz monzodiorite and quartz monzogabbro; **(b)** alkalies ($Na_2O + K_2O$) vs. silica (SiO_2) and, **(c)** molar A/CNK vs. SiO_2 diagram. Lines separate alkaline and subalkaline fields (after Irvine and Baragar 1971) in (b) and peraluminous and metaluminous fields (after Shand 1947) in (c).

● = dark green quartz monzonite
○ = light-coloured quartz monzonite
+ = pyroxene granite

Table 4. Average chemical composition of 18 dark green quartz monzonite samples and 7 light-coloured quartz monzonite samples from the Luopa intrusion.

	Dark green type	Light-coloured type
number of analyses	18	7
SiO_2 wt%	59,27	60,57
TiO_2	0,80	0,68
Al_2O_3	16,17	16,61
Fe_2O_3tot	9,69	8,09
MnO	0,17	0,13
MgO	0,64	0,62
CaO	4,03	3,56
Na_2O	3,29	3,13
K_2O	4,68	5,26
P_2O_5	0,26	0,27
Sum	99,01	98,91
A/CNK	0,91	0,96
NK/A	0,65	0,65
Rb ppm	69	80
Ba	2816	3007
Sr	423	406
Zr	767	734

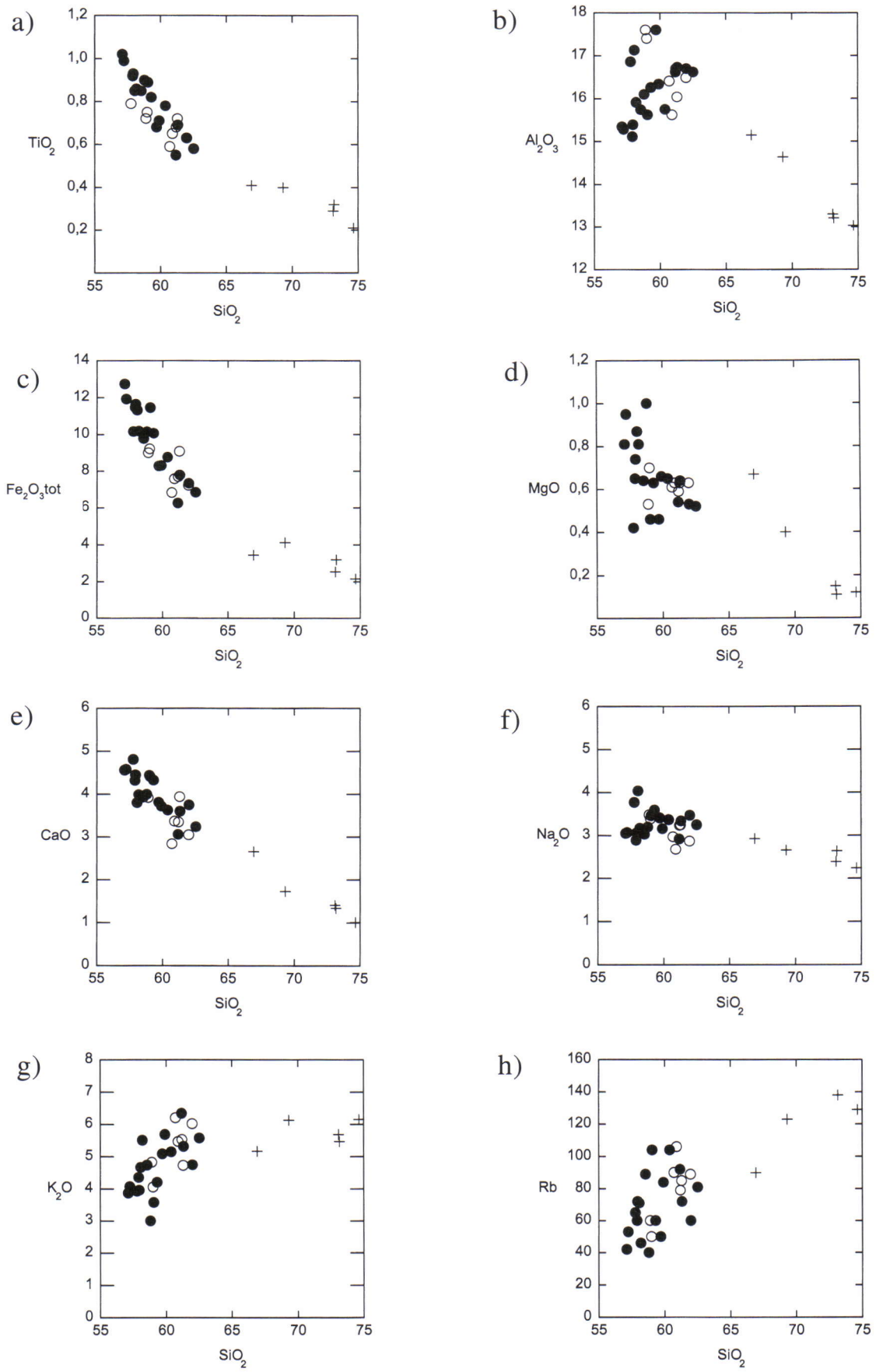


Fig. 9. Harker and trace element variation diagrams of the Luopa quartz monzonite and pyroxene granite. Symbols as in Figure 8.

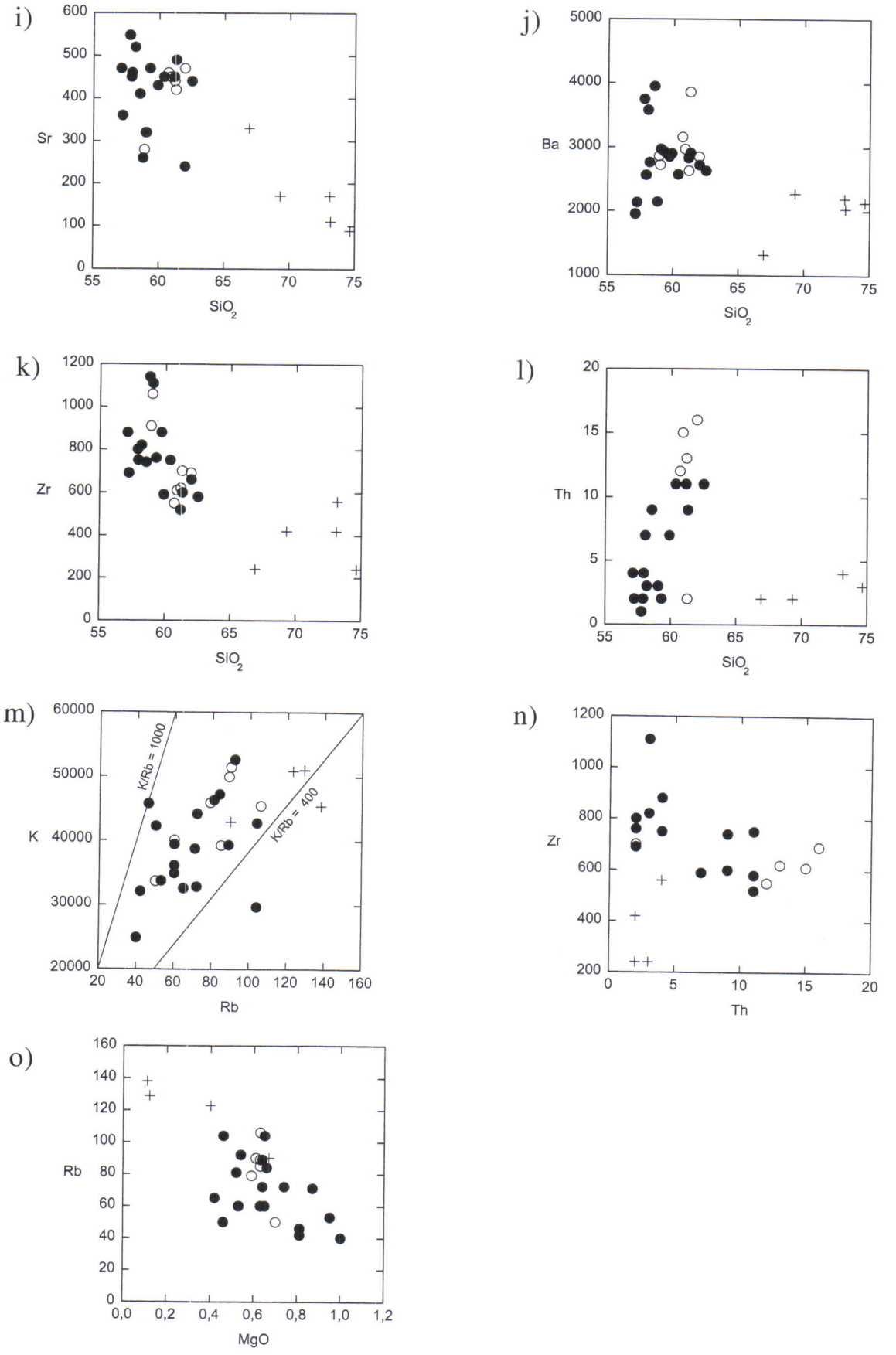


Fig. 9. Continued

Aplitic dykes

Aplitic, locally pegmatitic, dykes 0.05-0.5 m wide occur in the quartz monzonite near the contacts of the stock (Fig. 4a). The largest number of dykes were observed at the eastern margin of the Luopa intrusion. The dykes are considered to be the latest magmatic phases that penetrated into fractures of the earlier crystallized, outer part of the intrusion.

Granitic margin

There is a 0.05-0.5 m wide margin composed of a garnet-biotite granite with a grain size of 0.3-2.5 cm between the quartz monzonite and country rocks, best developed against metapelites. In the margin garnet and biotite contain worm-like quartz inclusions. Garnet is partly altered to biotite, alkali feldspar and Fe-oxides.

Whole rock chemistry

Outlines and analytical procedures

In total, 25 rock samples from the Luopa intrusion were analysed. 18 analyses are from the dark green quartz monzonite and seven from to the light-coloured variety (Table 4). In addition to major elements, a number of trace elements were analysed, however, not from every sample. A few analyses were also made from the surrounding pyroxene granite and biotite-hornblende granite for comparison. Geographical locations of the analysed samples are in Appendix 1.

Major elements were determined by X-ray fluorescence method (XRF) at the GTK and at the Laboratory of the Rautaruukki Company (RR). Most of the trace elements were assayed by neutron activation analysis (NAA) at the Technical Research Centre of Finland (VTT). Trace elements of a few samples were analysed by XRF at the GTK and RR. All the chemical analyses are given in Appendices 2 and 3.

Geochemistry

For comparative purposes, both of the quartz monzonite types and the adjacent pyroxene granite are plotted on the same diagrams. On the classification diagrams based on the whole rock chemistry, the Luopa rocks mainly plot in the field of quartz monzonite, but the pyroxene granites plot in the granite field (Fig. 8a).

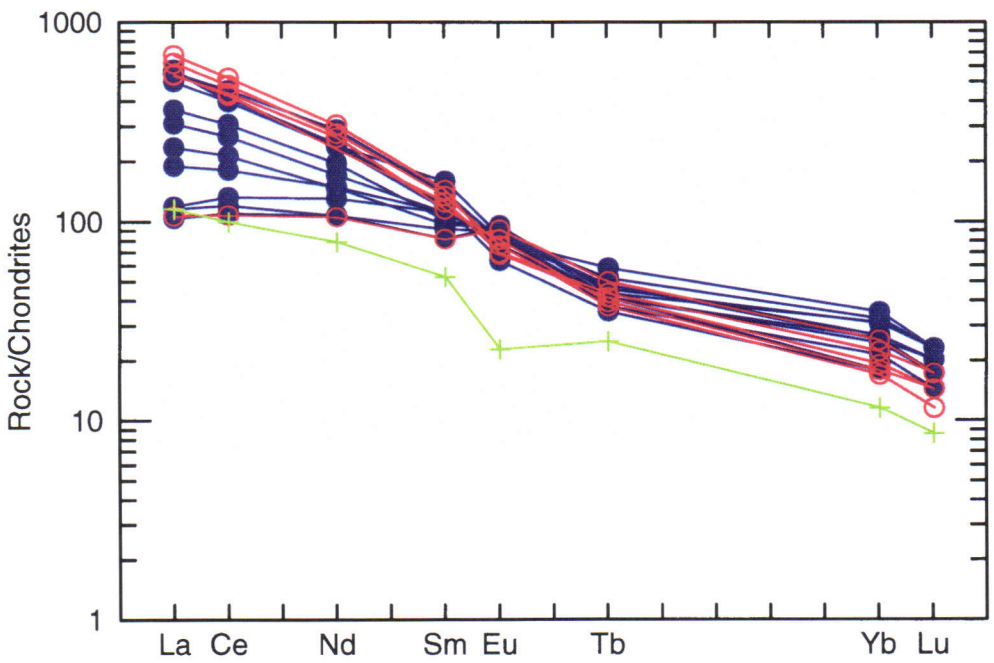


Fig. 10. REE distribution curve (chondrite normalized values) of 8 dark green quartz monzonite, of 5 light-coloured quartz monzonites and of one pyroxene granite sample. Symbols: red open circles = light-coloured quartz monzonite, blue filled circles = dark green quartz monzonite, green crosses = pyroxene granite.

The alkalies (Na_2O+K_2O) vs. silica (SiO_2) -diagram indicates that the quartz monzonite is relatively more alkaline than the pyroxene granite (Fig. 8b).

The quartz monzonites are metaluminous and the light-coloured quartz monzonites have slightly higher A/CNK values than the dark green variety (Fig. 8c). The adjacent pyroxene granite has slightly higher molar A/CNK, extending into the field for peraluminous granites (Fig. 8c). Both intrusions belong to I-type granitoids because their A/CNK values are less than 1.1. Generally the average A/CNK value of the pyroxene granites is slightly less than that of the biotite-hornblende granites (Mäkitie et al. 1999). This is expected because the biotite-hornblende granite has a higher mode of quartz compared to the pyroxene granite – in the Luopa area and in its vicinities the A/CNK value correlates with the mode of quartz in granitoids (Mäkitie et al. 1999).

Figures 9a to 9o show Harker-diagrams and trace element scatter diagrams of the quartz monzonite and pyroxene granite. These diagrams show that the compositions of the two quartz monzonite types overlap, and K/Rb is very high, 400-800 (see Appendix 2). High K/Rb is typical for unaltered magmatic rocks, contrary to low ratios that would suggest fluid interaction (Clarke 1992). Rb/Sr in the quartz monzonite is low, even less than in the pyroxene granite, indicating that the magma was relatively undifferentiated. Zr and Th do not decrease together in the quartz monzonite as is usual in many granitoids, indicating that these elements were not together removed by zircon. Thorium was probably retained by allanite instead of zircon.

The light-coloured quartz monzonite has a slightly steeper REE distribution curve than in the dark green variety (Fig. 10). REE concentrations are lower in the pyroxene granite than in the quartz monzonites. The REE distribution pattern shows that the concentration of europium remains rather constant in the quartz monzonite samples.

Granite typology

In the Zr+Nb+Ce+Y vs. (K_2O+Na_2O)/CaO discrimination diagram of Whalen et al. (1987), the

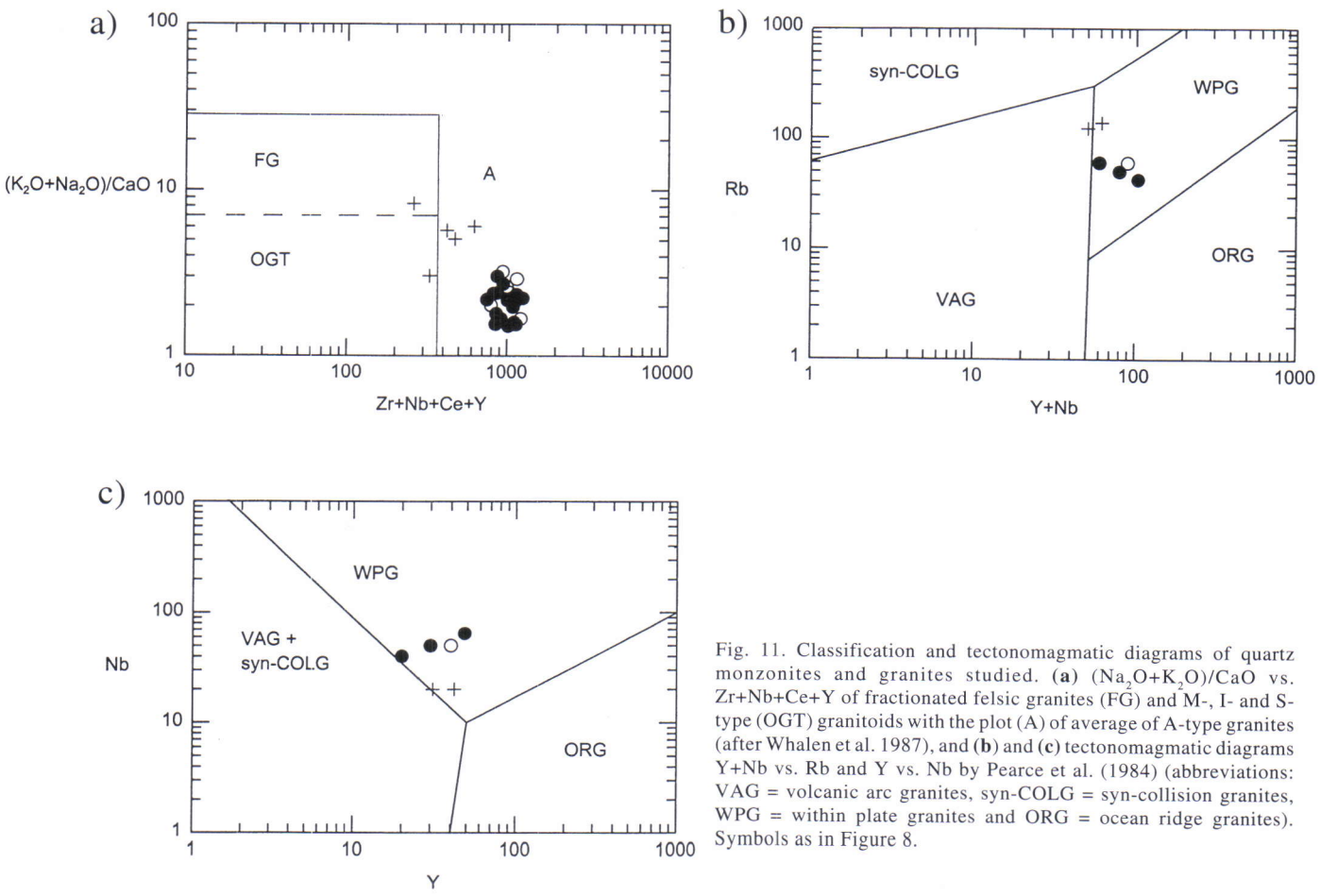


Fig. 11. Classification and tectonomagmatic diagrams of quartz monzonites and granites studied. (**a**) (Na_2O+K_2O)/CaO vs. Zr+Nb+Ce+Y of fractionated felsic granites (FG) and M-, I- and S-type (OGT) granitoids with the plot (A) of average of A-type granites (after Whalen et al. 1987), and (**b**) and (**c**) tectonomagmatic diagrams Y+Nb vs. Rb and Y vs. Nb by Pearce et al. (1984) (abbreviations: VAG = volcanic arc granites, syn-COLG = syn-collision granites, WPG = within plate granites and ORG = ocean ridge granites). Symbols as in Figure 8.

quartz monzonite compositionally plots in the field of A-type granites and outside the fields of fractionated felsic granites and M-, I-, S-type granites (Fig. 11a). According to these diagrams, the pyroxene granites have geochemical affinities to A-type granites. However, many geochemical features in the quartz monzonite, such as a high Sr/Rb, CaO, and Ba, and low SiO_2 are in contrast with typical A-type granitoids, for example, rapakivi granites described by Eby (1990) and Rämö (1991). The given sums of Zr+Nb+Ce+Y in the aformentioned diagrams may indicate to a 'minimum value', because niobium, cerium and yttrium were not analysed from every rock sample (see Appendix 2).

In the tectonomagmatic Nb vs. Y and Rb vs. Y+Nb diagrams by Pearce et al. (1984) the Luopa quartz monzonite and the neighbouring pyroxene granite plot in the fields of within plate granites (WPG) and volcanic arc granites (VAG) (Figs. 11b and 11c). Noteworthy is that the Y/Nb ratio in rocks is robust and remains relatively constant throughout the evolution of a particular suite (Eby 1990).

Magnetic susceptibility and zoning

The average magnetic susceptibility of the intrusion is 2400 μSI, but the individual values vary largely (400-14000 μSI). In the central parts of the stock, the magnetic susceptibilities are higher than in the marginal parts. In the eastern part of the intrusion, magnetic susceptibilities are generally low, ranging from 500 to 950 μSI (Fig. 12).

The rocks with fayalite- and hornblende-bearing assemblages, in which biotite is often absent, are most common in the centre of the intrusion (Fig. 12). The Thornton-Tuttle (T-T) differentiation index (Thornton and Tuttle 1960), calculated from the CIPW normative minerals (weight norm, FeO and Fe_2O_3 after Irvine and Baragar 1971), range from 64 to 76 in the quartz monzonite intrusion. The light-coloured quartz monzonites have slightly higher T-T indexes than the dark green quartz monzonites (Fig. 12). The fayalite and hornblende-bearing quartz monzonites have generally lower (< 70) T-T indexes than in the marginal parts of the stock, indicating that the stock is slightly zoned.

Significance of fayalite and ilmenite

Fayalite-bearing igneous mineral assemblages are usually observed in highly evolved magmatic rocks associated with anorthositic, gabbroic and alkalic intrusions and in anorogenic granitoids and rhyolites (e.g. Simonen 1961, Frost et al. 1988, Fuhrman et al. 1988, Nordgulen and Mitchell 1988, Rämö 1991). However, olivine very rarely occurs in quartz monzonites (see, however, Oyawoye and Makanjuola 1972, Emslie et al. 1984, Dada and Respaut 1989, Elliott et al. 1998).

Fayalite typically crystallizes in rocks with a very high FeO/(FeO + MgO) ratio, usually higher than 0.9 (e.g. Malm and Ormaasen 1978). Excess of Fe^{3+} inhibits crystallization of fayalite, whereas reducing conditions promote its formation (Wyderko 1969). The stability of Fe-Ti oxides in the presence of fayalite is a function of both silica activity and oxygen fugacity (Frost et al. 1988). The relationships of fayalite and coexisting quartz with Fe-Ti oxides are governed by equation (1) abbreviated QUIlF (Frost et al. 1988):

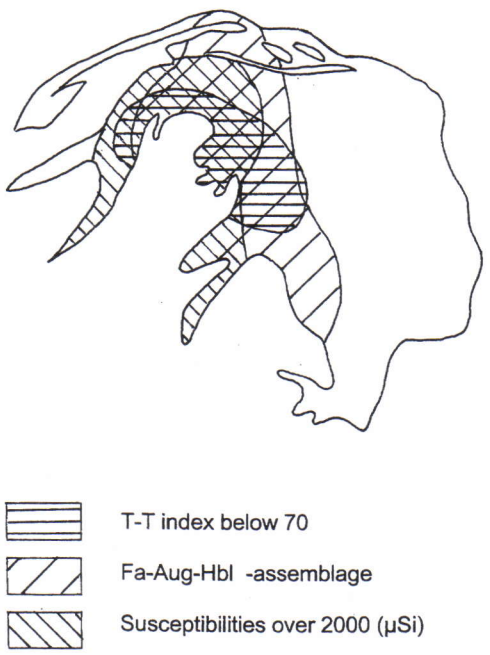


Fig. 12. Variation of Thornton-Tuttle differentiation index and magnetic susceptibility within the Luopa quartz monzonite. Distribution of fayalite-bearing rock types are also indicated.

$$SiO_2 + 2Fe_2TiO_4 = 2FeTiO_3 + Fe_2SiO_4 \quad (1)$$

quartz(Q) ulvospinel(U) ilmenite(Il) fayalite(F)

Granitoids which contain fayalite and quartz usually also contain ilmenite, and suggest that these rocks have been crystallized at or below the QUIlF surface, but those which also contain titanomagnetite may be either saturated or undersaturated with respect to silica (Frost et al. 1988). For example, the relatively magnesian portions in monzosyenitic plutons have the assemblage fayalite (0.73-0.88)-pigeonite-augite plus two Fe-Ti oxides (Fuhrman et al. 1988). With increasing Fe-enrichment, both pigeonite and magnetite disappear while quartz becomes crystallized. This indicates that the conditions of crystallization crossed the QUIlF surface at about the same stage that pigeonite became unstable with respect to augite-fayalite-quartz at an inferred pressure of 3 kbar (Fuhrman et al. 1988). Oxide-silicate re-equilibration persists in plutonic rocks to temperatures well below the solidus, but Fe-Ti oxides much more readily re-equilibrate during cooling than silicates (Frost et al. 1988).

The chemical and mineral composition of the well studied fayalite-, quartz- and ilmenite-bearing Sybille and Wolf River monzonites in USA (e.g., Frost et al. 1988, Fuhrman et al. 1988) greatly resemble those of the Luopa intrusion. High FeO/(FeO + MgO) and the predominance of ilmenite over magnetite (> 99%) are characteristic in these intrusions. Thus, it can be suggested that the crystallization of the Luopa intrusion took place in reducing conditions at or below the QUIlF-equilibrium line shown by Frost et al. (1988, figure 8).

Age determination

The U-Pb age (1871±1 Ma) of the Luopa intrusion (earlier known as the Ilmajoki intrusion) reported by Mäkitie and Lahti (1991) has been recently recalculated and corrected to 1872±2 Ma (Matti Vaasjoki, pers.comm. 1997). The isotopic data is on Table 5 and the concordia diagram is given in Figure 13.

Both light and dark zircons, all idiomorphic and partly zoned, were used in U-Pb age determinations. The dark varieties are richer in U than the light varieties. The zircons are very slightly discordant (Fig. 13), and thus the resulting age is a good estimate for crystallization of the Luopa quartz monzonite.

Table 5. U-Pb isotopic data on the Luopa quartz monzonite.

Fraction	Weigth (mg)	U conc (ppm)	Pb conc (ppm)	$^{206}Pb/^{204}Pb$ (meas.)	$^{08}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	±2 SE (%)	$^{207}Pb/^{235}U$	±2 SE (%)	$^{207}Pb/^{206}Pb$	±2 SE (%)	Corr.	$T_{206/238}$	$T_{207/235}$	$T_{207/206}$
A. +4.5/+100/clear	4,0	207,4	70,8	8736	0,0888	0,3285	0,70	5,181	0,72	0,1144	0,21	0,96	1830	1849	1870
B. +4.5/+100/dark	10,5	229,8	80,2	9246	0,0877	0,3359	0,84	5,300	0,85	0,1144	0,10	0,99	1866	1868	1871
C. 4.3-4.5/+100/dark	5,8	846,7	277,7	17495	0,0723	0,3205	0,79	5,054	0,79	0,1144	0,10	0,99	1792	1828	1870
D. 4.3-4.5/+100/clear	6,7	345,3	122,3	19916	0,0862	0,3336	0,60	5,265	0,60	0,1145	0,10	0,99	1855	1863	1871

Isotopic ratios corrected for fractionation, blank and common lead (6/4: 15.7; 7/4: 15.3; 8/4: 35.2)

Model of magma evolution in space and time

The mineral composition of the Luopa quartz monzonite is rather similar to that of the calc-alkaline monzonitic series described by Lameyre and Bowden (1982), and Lameyre and Bonin (1991). These authors suggested that it may be possible to relate the major trends of the calc-alkaline trondhjemitic (low K), calc-alkaline granodioritic (medium K) and calc-alkaline monzonitic series (high K) to both their position within the orogen and their chronology trends: calc-alkaline trondhjemitic rocks intruded first, followed by granodioritic magmatism and the latest intrusives are calc-alkaline monzonitic by composition.

The Luopa region contains intrusions resembling the aforementioned calc-alkaline trondhjemite series (foliated tonalites east of Luopa), calc-alkaline granodiorite series (various granites south of Luopa)

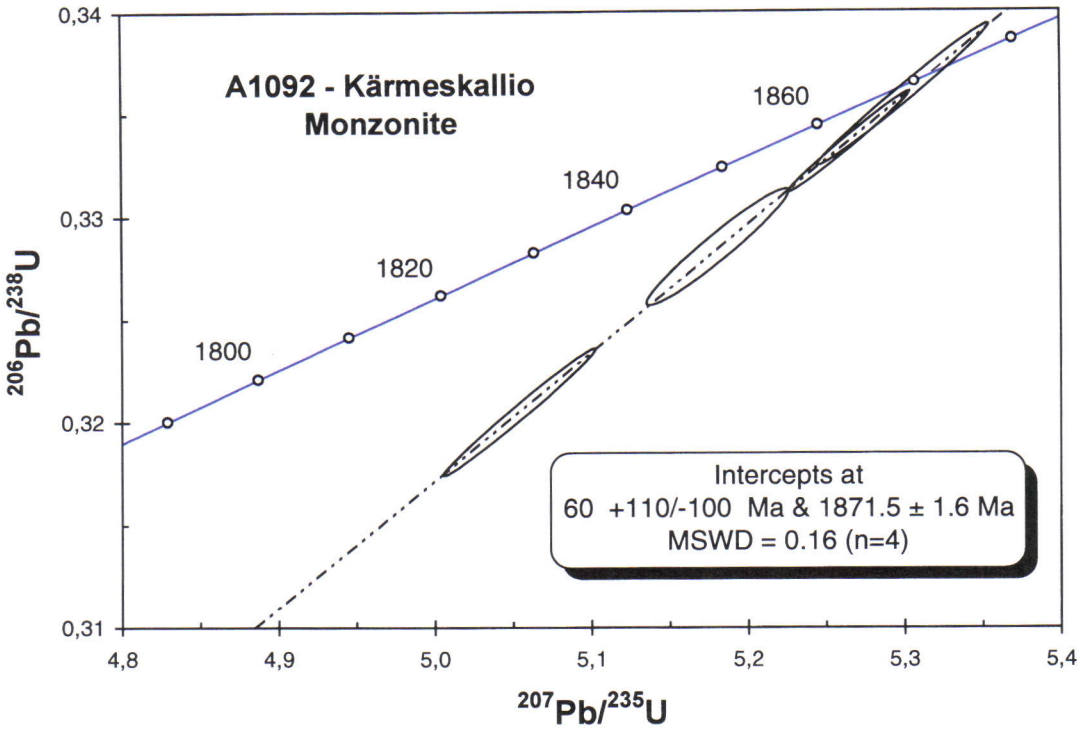


Fig. 13. The concordia diagram for the U-Pb isotope ratios of zircon fractions from the Luopa quartz monzonite, Kärmeskallio area.

and calc-alkaline monzonitic series (the Luopa quartz monzonite). The tonalites are cut by granites, which are themselves cut by the Luopa quartz monzonite. Modal compositions (seen in APQ diagram) of the Luopa granitoids show similar trends to the rocks presented in the model by Lameyre and Bowden (1982). However, the modes of K-feldspar and concentrations of K_2O are fairly similar and high within the granites and quartz monzonites of the Luopa region. The granite cut by quartz monzonite, therefore, does not well represent a calc-alkaline 'medium K' granodiorite. The magmatic evolution model shown in Figure 14 and discussed by Lameyre and Bowden (1982) and Lameyre and Bonin (1991) is one interpretation for the unusual intersection relationship between granitoids and quartz monzonite in the Luopa region.

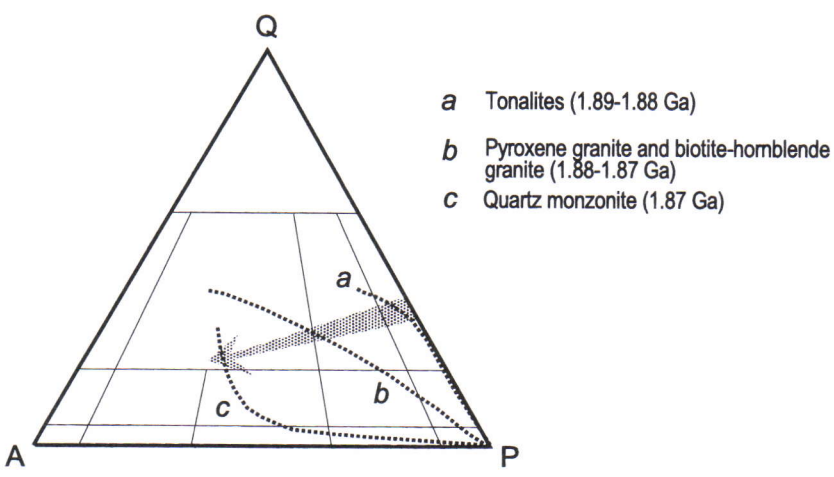


Fig. 14. Schematic representation (in A-P-Q diagram of Streckeisen 1976) of the evolution in space and time (shaded arrow) of the three major trends in the calc-alkaline spectrum after Lameyre and Bowden (1982, figure 6): a = calc-alkaline trondhjemitic (low K), b = calc-alkaline granodioritic (medium K); c = calc-alkaline monzonitic (high K).

THE CONTACT AUREOLE

Introduction

The metapelites at the contacts of the quartz monzonite contain mineral assemblages and textures, which do not occur in the surrounding regional metamorphic high-grade terrane. This indicates that contact metamorphism is clearly the result of the Luopa quartz monzonite intrusion. The contact metamorphic aureole of the Luopa quartz monzonite can be divided into an inner and an outer aureole.

The inner contact aureole extends up to two metres from the contact of the stock, although its width also depends on the regional position of outcrops. The inner aureole is best recognizable in the parts that are closer to the centre of the irregular stock. Mineral assemblages in the inner aureole indicate P-T conditions of granulite facies. The outer contact aureole extends to about 30 m from the quartz monzonite contact and includes mineral assemblages indicative of a slightly lower metamorphic grade, in the upper-amphibolite facies.

The inner aureole is directly adjacent to the fayalite-bearing quartz monzonite. The contact metamorphic aureoles are difficult to present on geological maps because of the irregular form of the stock and because of variation in the dip of the contacts. Mineral assemblages encountered in various rock types within the contact metamorphic aureoles are listed in Table 6.

Pelitic rocks in the contact aureole can be divided into three groups: (1) diatexitic metapelites (observed from the inner and outer aureole), (2) tightly banded mica gneisses (observed in the outer aureole) and (3) psammitic metapelites (observed from the inner and outer aureole). The metamorphic reactions presented are in simplified forms. For references to the reactions, see Spear (1993) and Bucher and Frey (1994). Abbreviations used are explained in Appendix 4.

Inner contact aureole: petrography and metamorphic reactions

Pelitic rocks

Diatexitic metapelites. Intensely melted diatexitic metapelites occur particularly at those contacts which are situated closer to the central part of the stock (fayalite-bearing quartz monzonite). The texture of the rock resembles intrusive rocks, but the rock often contains dark ghost-like gneiss remnants (Fig. 5a).

Those diatexitic metapelites situated very close (< 0.5 m) to the contacts with quartz monzonite are orthopyroxene-bearing (Fig. 4d). The main minerals in the rocks are quartz (20-30 %), biotite (20-25 %), untwinned K-feldspar (15-20 %), plagioclase (5-30 %), orthopyroxene (1-5 %) and garnet (1 %). Orthopyroxene and garnet were found in the same thin sections, but coexisting pairs were not. Orthopyroxene and K-feldspar crystallized by the reaction (2), which is a common reaction in the formation of orthopyroxene-quartz assemblages in granulite-grade metamorphism. This reaction requires a temperature above 700 °C to initiate (e.g. Bucher and Frey 1994).

Diatexitic cordierite-orthopyroxene-K-feldspar mica gneisses rarely occur in the inner contact aureole (Fig. 15a). This mineral assemblage probably could be formed by reactions (3) or (4), which require temperatures of about 750 °C to begin (e.g. Bucher and Frey 1994). Formation of the cordierite-orthopyroxene-melt assemblage is favoured by the relative high magnesium content, 4.28 (wt%), of the rock (cf. Appendix 5, sample 166d).

Sillimanite is rare in the metapelites of the inner contact aureole and it has a fibrolitic appearance. Sillimanite is rare because the reactions (5) and (6) inhibit the crystallization of sillimanite; all sillimanite has been used in the formation of cordierite- and garnet-bearing melt. Opaques in the mica gneisses are usually pyrites surrounded by ilmenite.

$$\text{Bt} + \text{Qtz} = \text{Opx} + \text{Grt} + \text{Kfs} + \text{melt} \quad (2)$$

$$\text{Bt} + \text{Sil} + \text{Qtz} = \text{Crd} + \text{Opx} + \text{Kfs} + \text{melt} \quad (3)$$

$$\text{Bt} + \text{Grt} + \text{Qtz} = \text{Crd} + \text{Opx} + \text{Kfs} + \text{melt} \quad (4)$$

$$\text{Bt} + \text{Sil} + \text{Qtz} = \text{Crd} + \text{Kfs} + \text{melt} \quad (5)$$

$$\text{Bt} + \text{Sil} + \text{Qtz} = \text{Grt} + \text{Crd} + \text{Kfs} + \text{melt} \quad (6)$$

Some moderately melted mica gneisses include symplectites and macroscopic (5-10 mm) corona textures composed of hercynite, sillimanite, cordierite, biotite, quartz and plagioclase (Figs. 15b and 15c).

There are two kinds of coronas and symplectites: 1) biotite corona surrounding Hc-Pl symplectites on large sillimanites (Fig. 15c), 2) plagioclase surrounding cordierite, which surrounds Hc-Sil symplectite and sillimanite on small hercynite grains (Fig. 15b). Hercynite in the second coronas may be formed by reaction (7) and (8). Similar intergrowths of hercynite, sillimanite and cordierite are reported in many contact metamorphic aureoles (e.g. Grant and Frost 1990, Kuznetsova et al. 1994). Retrogressive andalusite, usually intergrown with biotite, is a fairly common accessory mineral in the hercynite-bearing mica gneisses around the Luopa quartz monzonite.

Minor garnet crystals may be surrounded by cordierite, indicative of decreasing pressure during cooling, and the reversal of reaction (9). Occasionally, there is fibrolitic sillimanite at the contacts of these garnet and cordierite grains. Garnet and cordierite

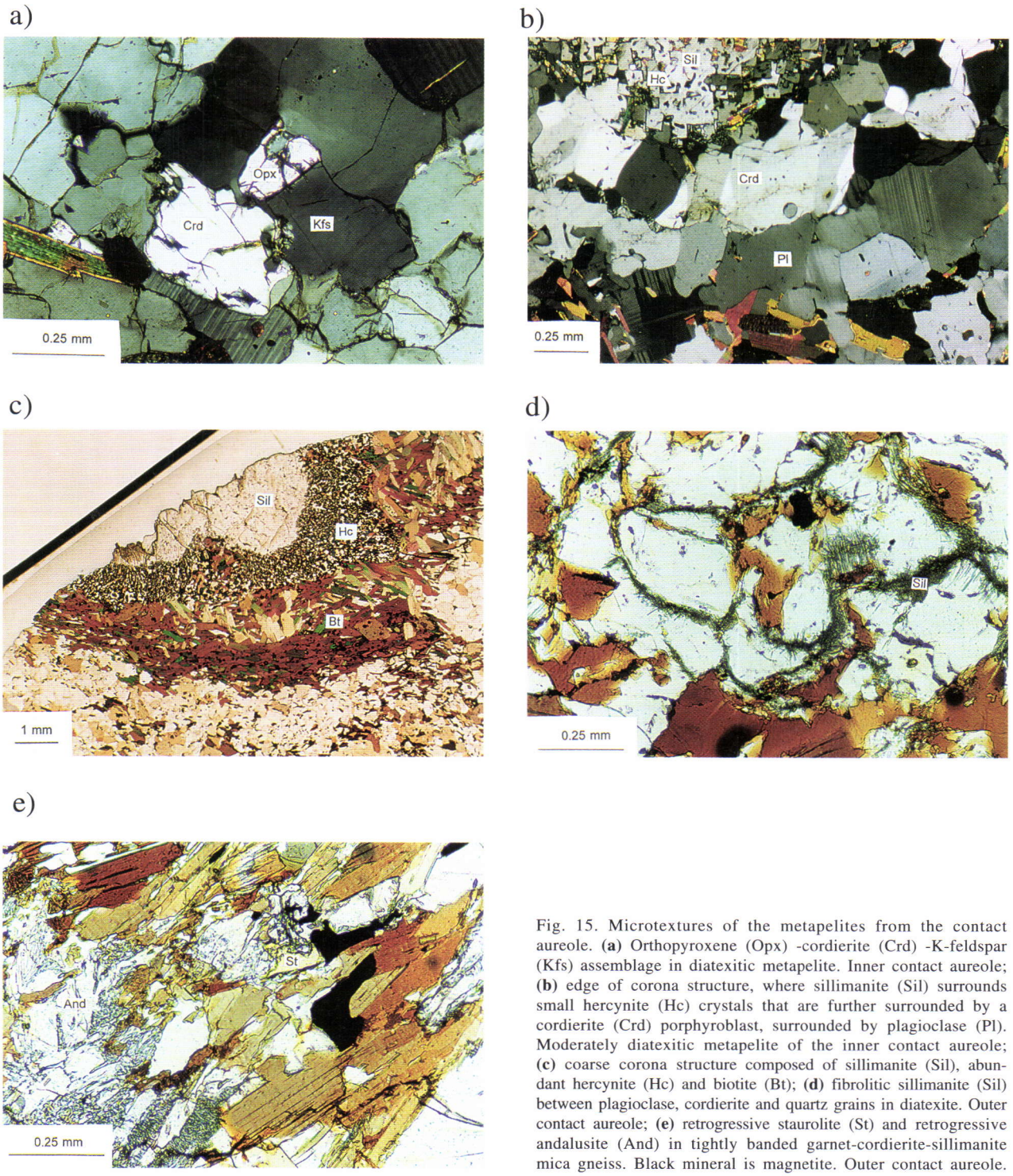


Fig. 15. Microtextures of the metapelites from the contact aureole. **(a)** Orthopyroxene (Opx) -cordierite (Crd) -K-feldspar (Kfs) assemblage in diatexitic metapelite. Inner contact aureole; **(b)** edge of corona structure, where sillimanite (Sil) surrounds small hercynite (Hc) crystals that are further surrounded by a cordierite (Crd) porphyroblast, surrounded by plagioclase (Pl). Moderately diatexitic metapelite of the inner contact aureole; **(c)** coarse corona structure composed of sillimanite (Sil), abundant hercynite (Hc) and biotite (Bt); **(d)** fibrolitic sillimanite (Sil) between plagioclase, cordierite and quartz grains in diatexite. Outer contact aureole; **(e)** retrogressive staurolite (St) and retrogressive andalusite (And) in tightly banded garnet-cordierite-sillimanite mica gneiss. Black mineral is magnetite. Outer contact aureole.

Table 6. Mineral assemblages of rocks in the contact metamorphic aureole of the Luopa quartz monzonite. All assemblages include plagioclase and quartz. Abbreviations are explained in Appendix 4.

ROCK TYPE	MINERAL ASSEMBLAGE	
	inner aureole	outer aureole
Diatexitic metapelite	Hc-Sil-Crd-Bt Hc-Grt-Bt Opx-Grt-Bt Opx-Crd-Bt-Kfs Opx-Bt-Kfs(-Grt) Grt-Crd-Bt-Kfs	Grt-Crd-Bt-Kfs(-Sil)
Tightly banded mica gneiss		Grt-Crd-Kfs-Bt-Sil
Psammitic metapelite	Opx-Grt-Bt	Grt-Bt
Intermediate nebulite	Opx-Grt-Bt Opx-Cpx-Bt	
Intermediate metavolcanic rock	Opx-Bt	
Foliated tonalite	Opx-Bt Opx-Kfs-Bt	Cpx-Bt-Hbl(-Opx-Kfs)

are also retrogress to biotite. Locally, the felsic minerals occur as inclusions in poikiloblasts of garnets.

$$\text{Sil} + \text{Bt} + \text{Qtz} \pm \text{Pl} = \text{Crd} + \text{Hc} + \text{Kfs} \quad (7)$$

$$\text{Sil} + \text{Grt} = \text{Hc} + \text{Crd} \quad (8)$$

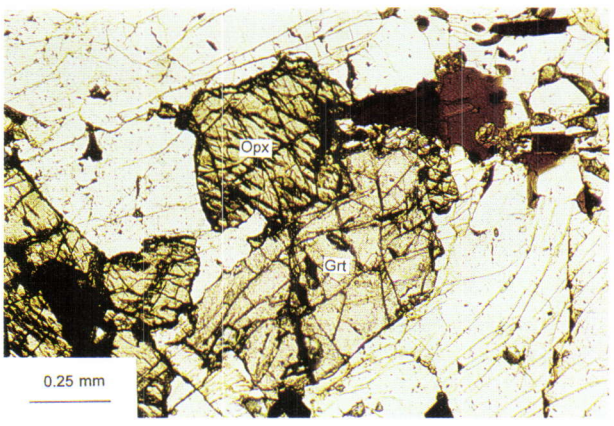
$$\text{Crd} = \text{Grt} + \text{Sil} + \text{Qtz} \quad (9)$$

Psammitic metapelites. A few psammitic metapelites form wide intercalations (0.2-2 m) between the mica gneisses. The mineral composition of these rocks are: quartz (30-50 %), plagioclase (25-40 %), biotite (4-10 %), garnet (2-5 %), orthopyroxene (1-2 %), and opaques (1 %). Orthopyroxene is often altered to a brownish phyllosilicate mineral. Locally, the orthopyroxene is surrounded by garnet (see Mäkitie 1990), which is probably formed by the reversal of reaction (10).

Intermediate rocks

Intermediate nebulites. The inner aureole contains nebulitic intermediate portions, which were formed by mixing components of metapelites and intermediate metavolcanic rocks during contact metamorphism (Fig. 5c). These portions include remnants of supracrustal rocks that are rich in biotite and hornblende. The main minerals are quartz, plagioclase, biotite and orthopyroxene. Accessory minerals are

a)



b)

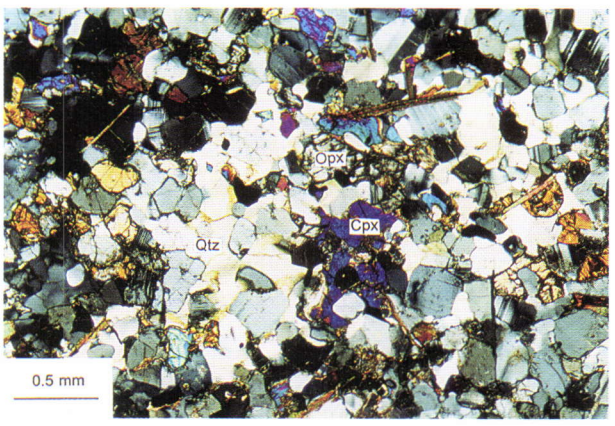


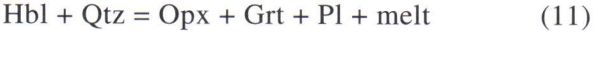
Fig.16. Microtextures in intermediate nebulites. **(a)** Orthopyroxene (Opx) -garnet (Grt) pair in intermediate nebulite. Inner contact aureole; **(b)** orthopyroxene (Opx) -clinopyroxene (Cpx) pair in intermediate nebulite. Quartz (Qtz) occurs as interstitial crystals between other minerals.

hercynite, clinopyroxene, garnet, hornblende, ilmenite and magnetite. The intermediate nebulites have gradual contacts with the neighbouring diatexitic garnet-cordierite mica gneisses.

The garnet and pyroxene assemblages (Figs. 16a and 16b) in the nebulites were probably formed by reactions (10), (11) and (12). A favourable chemical composition, with a relatively high amount of FeOtot and CaO, allowed equilibration of the garnet and orthopyroxene in the rocks. Orthopyroxene inclusions in the garnet indicate that orthopyroxene is locally transformed to garnet, probable by reaction (10). In places, quartz was crystallized in the interstices between plagioclase and pyroxene (Fig. 16b).

Hornblende only occurs as retrogressive rims around pyroxene. Locally, there are hercynite- and plagioclase-bearing inclusions and embayments in the garnets. Ilmenite is often surrounded by magnetite.

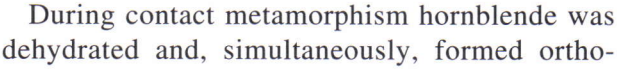
Grt + Qtz = Opx + Pl (10)



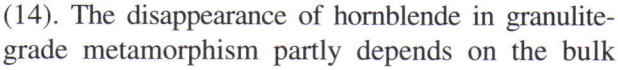
Hbl + Qtz = Opx + Grt + Pl + melt (11)

Hbl + Qtz = Opx + Cpx + Pl + melt (12)

Intermediate metavolcanic rocks. A few, rather well preserved, metavolcanic rocks occur at the contacts of the Luopa intrusion. The main minerals in the rocks are plagioclase, quartz, orthopyroxene, biotite, and, locally, minor amounts of hornblende and clinopyroxene. Orthopyroxene is formed by reaction (13) at the expense of hornblende.



Hbl + Qtz = Opx + Pl + melt (13)



Hbl + Bt + Qtz = Opx + Pl + Kfs + melt (14)

Tonalites

Foliated tonalites at the contacts with the fayalite-bearing quartz monzonite typically contain orthopyroxene, but not hornblende (Fig. 4c). The tonalites have a greenish brown colour on their fresh surface, are slightly weathered, and grain boundaries are 'muddy'. In places, the tonalites contain untwinned metamorphic K-feldspar (mode 4-15 %) and grade to granodiorite.

During contact metamorphism hornblende was dehydrated and, simultaneously, formed orthopyroxene and K-feldspar by reactions (13) and (14). The disappearance of hornblende in granulite-grade metamorphism partly depends on the bulk chemical composition of the rock (e.g. Bucher and Frey 1994). Typically, the metamorphic temperature estimations for reactions (13) and (14) are about 700-750 °C (e.g. Bucher and Frey 1994).

Outer contact aureole: petrography and metamorphic reactions

Pelitic rocks

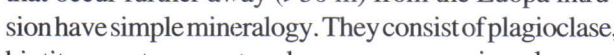
Diatexitic metapelites. The outer contact aureole also includes diatexitic metapelites. Unlike the metapelites of the inner contact aureole, these do not contain orthopyroxene or hercynite. The main minerals of this rock type are quartz (20-40 %), plagioclase (20-30 %), K-feldspar (10-20 %), biotite (10-15 %), garnet (5-10 %) and cordierite (10-15 %). Sillimanite is rare, but some fibrolitic sillimanite may occur between feldspars and quartz, or within cordierite (Fig. 15d). Garnet and cordierite are texturally crystallized in equilibrium.

Tightly banded mica gneisses. These rocks are garnet-cordierite-sillimanite mica gneisses, strongly deformed and tightly banded, and only associated with the outer contact aureole (Fig. 5b). Moreover, they are located, approximately, in the outermost parts of the quartz monzonite intrusion. The main minerals include quartz (25-40%), plagioclase (20-30%), K-feldspar (5-15%), garnet (10-15 %), cordierite (5-10 %) and sillimanite (2-6 %). Sillimanite often has a relatively coarse grain size and may occur as intergrowths with biotite.

Retrogressive staurolite and andalusite have been detected in some relatively magnesium-rich (MgO ~ 5.1 wt%) mica gneisses (Fig. 15e), where staurolite was probably formed by the reversal of reactions (15) and (16). These staurolite-bearing rocks do not contain muscovite, probably because staurolite remains stable in high-grade rocks in the absence of muscovite

(Bucher and Frey 1994). Locally, in the tightly banded mica gneisses, garnet has a narrow inclusion-free rim, but the core of the crystal contains quartz inclusions (see Mäkitie 1990).

$$St + Qtz = Grt + Sil + H_2O \quad (15)$$



$$St + Qtz = Grt + Crd + Al_2SiO_5 + H_2O \quad (16)$$

Psammitic metapelites. The psammitic metapelites that occur further away (>30 m) from the Luopa intrusion have simple mineralogy. They consist of plagioclase, biotite, quartz, garnet and some opaque minerals.

Intermediate rocks

The limit of the outer contact aureole is not clear, because orthopyroxene coexisting with hornblende, biotite, plagioclase and quartz is found in the intermediate metavolcanic rocks of the regional metamorphic terrane, within several kilometers of the Luopa intrusion (see Fig. 1).

Tonalites

In the tonalites, hornblende becomes common at distances of 2-3 meters from the quartz monzonite. Simultaneously, the amount of orthopyroxene decreases and clinopyroxene begins to occur. Locally, orthopyroxene and clinopyroxene are surrounded by retrogressive hornblende. The core of plagioclase crystals typically includes saussurite and sericite. The rock gradually changes into hornblende-biotite tonalite, at about 40 m from the Luopa intrusion contact. The regional metamorphic 'opx in' isograd lies in the same area (Fig 1.).

Pyroxene is more common in the foliated tonalites on the northern side of the Luopa intrusion than on eastern side, where apophyses of the intrusion are rare. The reason for the small amount of pyroxene is not known. It may be due to a lower thermal effect in the area, because fayalite-bearing assemblages (probably indicating a higher temperature magma) are absent in the eastern part of the Luopa intrusion (cf. Fig. 1).

Whole rock chemistry

Analytical procedures

The main elements of rock samples (from the contact aureole) were analysed by XRF at the Chemistry Laboratory of the Rautaruukki Company (RR). Trace elements have been assayed by XRF in RR, except yttrium which was analysed by OES in GTK. Most samples were taken with a portable diamond drill. All analyses are in Appendix 5. Locations of samples are shown in Appendix 1.

Pelitic rocks

Three whole rock analyses are available from the inner contact metamorphic aureole: two from the diatexitic metapelites and one from a psammitic metapelite (see Appendix 5). In addition, two rock samples were analysed form the outer contact metamorphic aureole, one from diatexitic metapelite and one from tightly banded mica gneiss.

The metapelites analysed typically contain lower SiO_2 but higher FeOtot and MgO than mica gneisses of the surrounding, regional metamorphic high-grade terrane (see Mäkitie 1999). The decrease of SiO_2 in the contact aureoles is relatively common and probably caused by the segregation and extraction of a silicate melt due to either granulite-grade metamorphism or the outward flux of low pH magmatic fluids dissolving feldspars and quartz (e.g. Compton 1960, Munksgaard 1988). The decrease of silica also indicates that the relative proportions of iron and magnesium have increased in the rest of the pelitic rocks.

Barium is slightly higher in the metapelites of the contact aureole than in the mica gneisses of the surrounding regional high-grade metamorphic terrane (Mäkitie 1999). The authors suggest that barium is derived from quartz monzonite (Ba content ~ 3000 ppm) and transported by aqueous fluids during the crystallization of the intrusion.

Tonalites

The chemical composition of four pyroxene tonalites from the contact aureole of the Luopa intrusion does not differ notably from the sphene- and epidote-bearing hornblende-biotite tonalites of the surrounding regional metamorphic terrane (cf. Appendix 5, Mäkitie et al. 1999). The macroscopic textures of the rocks support this observation because granitic veins, which should indicate segregation of light elements, are practically absent.

Geothermobarometry and mineral chemistry

Introduction

The microprobe analyses of minerals from pelitic rocks, intermediate nebulites and tonalites from the inner and outer contact metamorphic aureoles were made from polished thin sections by computerised electron microprobes (JEOL-JCXA 733) at GTK and at the University of Oulu (Table 7). Both cores and rims of coexisting garnet and cordierite crystals were analysed. The thermometers used were those of Holdaway and Lee (1977), Perchuk (1989), Harley (1984), Bhattacharya et al. (1991) and Fonarev and Grapchikov (1982). The barometers used were those of Perkins and Newton (1981) and Perkins and Chipera (1985).

Pelitic rocks

Garnet-cordierite thermometry. Cores of garnets in the metapelites of the contact aureole seem to have higher X_{Mg} (up to some 0.29) than their rims, where X_{Mg} values are about 0.19. Locally, the Mn and Ca content of garnet is relatively high, probably indicating retrogression (Table 7). Moreover, Mn content is higher in the rims than in the cores of garnet.

The thermometer of Holdaway and Lee (1977) gives temperature estimates of 590-670°C for the contact metamorphism (Table 7). The estimated pressure is 3.0 kbar and $P_{H2O} = 0.4 \times P_{tot}$. Slightly lower temperatures are obtained by using analyses from the rims of minerals rather than points from the cores. The thermometer of Perchuk (1989) gives higher estimates for metamorphic temperature than that of Holdaway and Lee (1977) (Table 7).

Garnet-orthopyroxene thermometry. The Al_2O_3 content of orthopyroxene in the pelitic rocks of the inner contact aureole range 0.9-3.8 wt%. The garnet and orthopyroxene (almost coexisting minerals in thin section) in the psammitic metapelites (sample 167L) from the inner contact aureole give unexpectedly low temperatures, 577 °C, using the thermometer of Harley (1984) and a higher estimate using the thermometer of Bhattacharya et al. (1991) (Table 7). The temperature calculations assume a pressure of 3 kbar.

Cordierite-orthopyroxene assemblage. Chemical compositions of coexisting cordierite and orthopyroxene are in Table 7. In terms of experimental metamorphic petrology (e.g. Spear 1993), this cordierite-orthopyroxene pair indicates higher metamorphic temperatures than the estimates by the aforementioned thermometers. Al_2O_3 content of orthopyroxene, a classical indicator of metamorphic grade, is relative high (3-4 wt%).

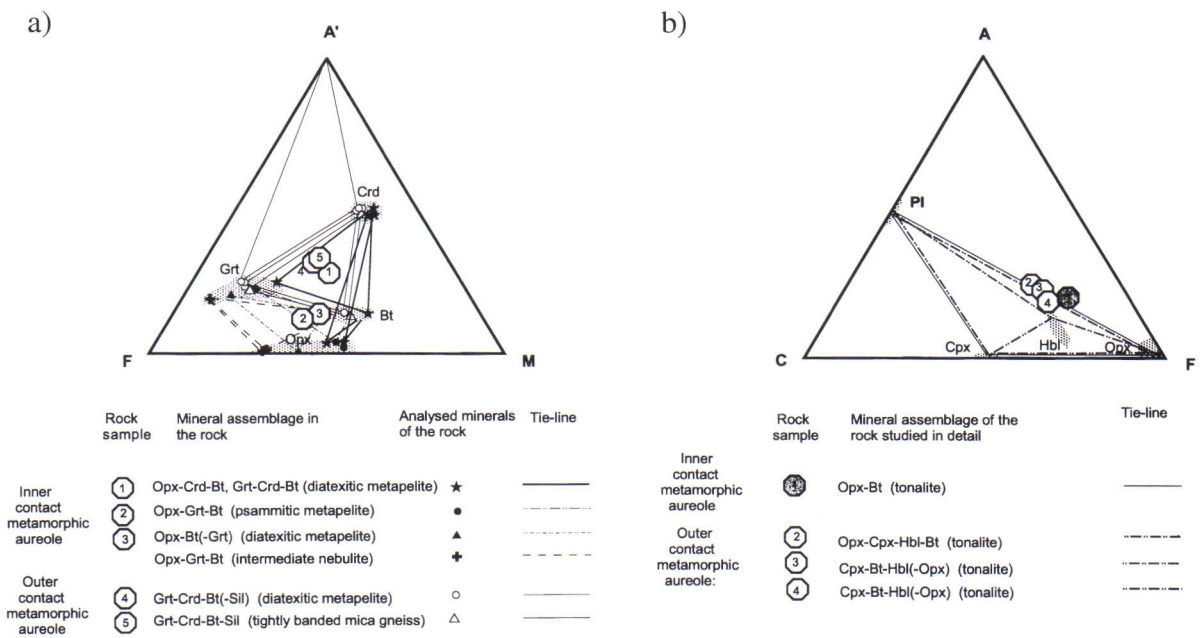


Fig. 17. Mineral assemblages of the rocks studied in detail. **(a)** A'FM-diagram (Reinhardt 1968) of pelitic rocks and intermediate nebulites for the inner and outer contact aureoles; **(b)** ACF-diagram (Eskola 1915) of tonalites for the inner and outer contact aureoles. All assemblages include plagioclase and quartz and, in the metapelites, K-feldspar.

Table 7. Microprobe analyses of orthopyroxene-cordierite, garnet-cordierite, orthopyroxene-garnet and orthopyroxene-clinopyroxene pairs as well as biotites in the pelitic rocks and intermediate nebulites from the contact metamorphic aureole of the Luopa quartz monzonite. The pairs refer to coexisting minerals, if not indicated otherwise. Metamorphic temperatures and pressures are calculated using the following thermo- and barometers (assuming a pressure of 3 kbar or temperature of 710 °C): T_1 (Holdaway and Lee 1977), T_2 (Perchuk 1989), T_3 (Harley 1984), T_4 (Bhattacharya et al. (1991), T_5 (Fonarev and Grapchikov 1982), P_1 (Perkins and Newton 1981), P_2 (Perkins and Chipera 1985). Oxygens: Grt 12, Crd 18, Opx 6, Cpx 6, Bt 22, Pl 8.

Rock	Diatexitic metapelite		Diatexitic metapelite		Diatexitic metapelite		Diatexitic metapelite		Diatexitic metapelite		Diatexitic metapelite		Diatexitic metapelite	
Sample	166B-HMM-82/2		166-HMM-82/1		42-HM(166)		43-HMM(9-10)		151AF		43-HMM(7-8)		38-HM(22-23)	
Location	inner zone		inner zone		inner zone		outer zone		outer zone		outer zone		outer zone	
Minerals	Opx	Crd	Opx	Crd	Grt(core)	Crd(core)	Grt(core)	Crd(core)	Grt(core)	Crd(core)	Grt(rim)	Crd(rim)	Grt(rim)	Crd(rim)
SiO_2	49,51	48,89	49,37	48,51	39,20	48,20	37,87	48,52	37,67	48,62	37,32	48,37	38,14	48,69
TiO_2	0,09	0,05	0,00	0,00	0,05	0,01	0,07	0,00	0,00	0,03	0,05	0,00	0,10	0,00
Al_2O_3	3,54	32,56	3,76	32,58	22,00	33,40	21,60	32,67	21,07	32,50	21,35	32,64	21,31	32,29
FeOtot	27,55	6,40	29,34	7,40	29,70	5,70	34,55	6,90	33,66	7,25	34,72	7,37	32,78	7,49
MnO	0,33	0,10	0,34	0,06	0,90	0,05	0,98	0,07	0,76	0,05	1,05	0,06	1,01	0,01
MgO	18,46	9,93	16,74	9,41	6,90	9,50	4,63	8,98	5,57	9,13	4,24	8,71	4,44	8,39
CaO	0,10	0,06	0,07	0,02	1,10	0,00	1,15	0,01	1,13	0,04	1,19	0,01	1,21	0,04
Na_2O	0,12	0,16	0,13	0,11	0,00	0,05	0,02	0,11	0,12	0,21	0,01	0,12	0,07	0,20
K_2O	0,02	0,02	0,03	0,02	0,00	0,00	0,01	0,02	0,03	0,03	0,00	0,00	0,03	0,00
Total	99,72	98,17	99,78	98,11	99,85	96,91	100,88	97,28	100,01	97,86	99,93	97,28	99,09	97,11
T_1 (°C)					672		593		655		594		633	
T_2 (°C)					710		624		693		624		665	
Si	1,898	5,003	1,906	4,989	3,046	4,972	2,991	5,015	2,993	5,009	2,986	5,010	3,043	5,052
Al	0,160	3,927	0,171	3,949	2,015	4,061	2,011	3,980	1,973	3,946	2,013	3,985	2,004	3,949
Ti	0,030	0,004	0,000	0,000	0,003	0,001	0,004	0,000	0,000	0,002	0,003	0,000	0,006	0,000
Mg	1,055	1,515	0,964	1,443	0,799	1,461	0,545	1,384	0,660	1,402	0,506	1,345	0,528	1,298
Fe	0,883	0,548	0,947	0,636	1,930	0,492	2,282	0,596	2,237	0,625	2,323	0,638	2,187	0,650
Mn	0,011	0,009	0,011	0,005	0,059	0,004	0,066	0,006	0,051	0,004	0,071	0,005	0,068	0,001
Ca	0,004	0,007	0,003	0,002	0,092	0,000	0,097	0,001	0,096	0,004	0,102	0,001	0,103	0,004
Na	0,009	0,032	0,010	0,022	0,000	0,010	0,003	0,022	0,018	0,042	0,002	0,024	0,010	0,040
K	0,001	0,003	0,001	0,003	0,000	0,000	0,001	0,003	0,003	0,004	0,000	0,000	0,001	0,000

Rock	Psammitic metapelite		Diatexitic metapelite		Intermediate nebulite			Intermediate nebulite			Intermediate nebulite		Intermediate nebulite	
Sample	167L		140-HM		922-HMM			923-HMM			141-HM-82		141-HM-87	
Location	inner zone		inner zone		inner zone			inner zone			inner zone		inner zone	
Minerals	Opx	Grt (almost coexisting)	Opx	Grt (almost coexisting)	Opx	Grt	Pl	Opx	Grt	Pl	Opx	Cpx	Opx	Cpx
SiO_2	51,34	38,40	50,80	39,10	48,71	36,86	59,84	48,88	36,71	58,71	51,17	52,54	51,18	52,47
TiO_2	0,08	0,02	0,05	0,01	0,12	0,07	0,05	0,08	0,09	0,04	0,05	0,13	0,05	0,06
Al_2O_3	2,33	19,32	2,80	21,80	0,37	20,27	24,93	0,65	20,15	25,80	0,47	0,78	0,41	0,77
FeOtot	26,91	30,66	27,80	29,80	38,72	33,95	0,07	38,43	35,37	0,04	30,12	11,34	30,21	11,35
MnO	0,31	1,30	0,40	1,30	0,53	2,02	0,00	0,68	2,02	0,00	0,79	0,19	0,67	0,32
MgO	17,98	5,51	16,80	6,50	9,99	2,02	0,00	10,64	2,61	0,00	17,35	12,56	16,48	12,50
CaO	0,13	1,74	0,20	1,30	0,72	4,25	7,04	0,59	2,41	7,81	0,59	21,94	0,60	21,03
Na_2O	0,02	0,05	0,10	0,01	0,11	0,12	6,84	0,14	0,11	6,35	0,11	0,31	0,00	0,15
K_2O	0,02	0,00	0,01	0,01	0,00	0,03	0,24	0,02	0,02	0,33	0,03	0,00	0,01	0,01
Total	99,12	97,00	98,96	99,83	99,27	99,59	99,01	100,11	99,49	99,08	100,68	99,79	99,61	98,66
T_3 (°C)	577		696		545			579						
T_4 (°C)	646		744		607			649						
T_5 (°C)											679		675	
P_1 (kbar)					5,7			3,6						
P_2 (kbar)					3,7			2,6						
Si	1,966	3,116	1,959	3,049	1,990	3,000	2,686	1,976	2,995	2,641	1,972	1,985	1,991	1,999
Al	0,105	1,848	0,127	2,004	0,018	1,944	1,319	0,031	1,938	1,368	0,021	0,035	0,019	0,035
Ti	0,002	0,001	0,001	0,001	0,004	0,004	0,002	0,002	0,006	0,001	0,001	0,004	0,001	0,002
Mg	1,026	0,667	0,966	0,756	0,608	0,245	0,000	0,641	0,317	0,000	0,997	0,707	0,956	0,710
Fe	0,862	2,081	0,897	1,944	1,323	2,310	0,003	1,300	2,414	0,002	0,971	0,358	0,983	0,362
Mn	0,010	0,089	0,013	0,086	0,018	0,139	0,000	0,023	0,140	0,000	0,026	0,006	0,022	0,010
Ca	0,005	0,151	0,008	0,109	0,032	0,371	0,339	0,026	0,211	0,376	0,024	0,888	0,025	0,859
Na	0,001	0,008	0,007	0,002	0,009	0,019	0,595	0,011	0,017	0,554	0,008	0,023	0,000	0,011
K	0,001	0,000	0,000	0,001	0,000	0,003	0,014	0,001	0,002	0,019	0,001	0,000	0,000	0,000

Table 7. Continued

Rock	Diatexitic metapelite		Diatexitic metapelite		Diatexitic metapelite	Diatexitic metapelite	Tightly banded mica gneiss	Tightly banded mica gneiss	Diatexitic metapelite	Diatexitic metapelite
Sample	166D		171-K		166D	171E/34	722	719	151AF	91-150CF
Location	inner zone		inner zone		inner zone	inner zone	outer zone	outer zone	outer zone	outer zone
Minerals	Opx	Grt	Opx	Grt	Bt	Bt	Bt	Bt	Bt	Bt
	(not coexisting)		(not coexisting)							
SiO_2	50,79	39,08	50,52	37,40	36,9	34,7	36,1	36,0	35,5	36,1
TiO_2	0,05	0,03	0,05	0,00	4,9	3,7	3,7	3,8	4,4	4,7
Al_2O_3	2,79	21,82	0,93	19,46	17,1	17,9	18,2	18,8	17,6	18,2
FeOtot	27,76	29,84	33,62	32,71	16,0	18,5	16,7	17,0	17,0	17,9
MnO	0,40	1,25	0,92	1,82	0,0	0,0	0,0	0,0	0,0	0,1
MgO	16,76	6,53	13,31	3,08	10,7	8,4	10,3	10,9	10,3	10,5
CaO	0,15	1,32	0,21	2,22	0,0	0,0	0,0	0,0	0,1	0,0
Na_2O	0,08	0,05	0,01	0,08	0,0	0,1	0,2	0,2	0,2	0,2
K_2O	0,03	0,02	0,00	0,03	8,9	9,1	8,8	9,7	9,7	9,7
Total	98,81	99,94	99,57	96,80	94,6	92,3	94,0	96,4	94,8	97,2
Si	1,961	3,045	1,997	3,091	2,77	2,71	2,74	2,68	2,70	2,68
Al	0,127	2,004	0,043	1,896	1,51	1,65	1,62	1,65	1,58	1,59
Ti	0,001	0,002	0,001	0,000	0,28	0,22	0,21	0,21	0,25	0,26
Mg	0,964	0,758	0,785	0,379	1,20	0,98	1,16	1,21	1,17	1,16
Fe	0,896	1,944	1,112	2,261	1,00	1,21	1,06	1,06	1,08	1,11
Mn	0,013	0,082	0,031	0,127	0,00	0,00	0,00	0,00	0,00	0,01
Ca	0,006	0,110	0,009	0,197	0,00	0,00	0,00	0,00	0,01	0,00
Na	0,006	0,008	0,001	0,013	0,00	0,01	0,03	0,03	0,03	0,02
K	0,001	0,002	0,000	0,003	0,85	0,91	0,85	0,92	0,94	0,91

Biotite. Biotite in pelitic rocks is characterized by high TiO_2 (3.7-4.9 wt%) (Table 7). The X_{Mg} value of the mineral varies from 0.45 to 0.53, with an average value of 0.51. Ti contents correspond to those of granulite-grade terranes (cf. Bailey 1984).

Intermediate nebulites

Garnet-orthopyroxene-plagioclase-quartz thermobarometry. Orthopyroxene in the intermediate nebulites has a lower aluminium content than in the pelitic rocks (Table 7). Moreover, Mg/Fe in orthopyroxene is less in the nebulites than in the pelitic rocks. Garnet in contact with orthopyroxene in the intermediate nebulites has higher Ca and Mn concentrations than garnet in the pelitic rocks (Table 7). These variations reflect differences in the whole rock chemical compositions between pelitic rocks and intermediate nebulites. The garnet-orthopyroxene pairs in the inner contact aureole give unexpectedly low temperature estimates (550-650 °C; assuming a pressure of 3 kbar) using the thermometer of Harley (1984) and Bhattacharya et al. (1991) (Table 7). Assuming a temperature of 710°C, geobarometers give pressure estimates of 2.6 - 5.7 kbar.

Orthopyroxene-clinopyroxene assemblage. In the intermediate nebulites, where the clinopyroxene is diopside, the Al_2O_3 content of orthopyroxene is relatively low, about 0.45 wt% (Table 7). Temperatures of slightly below 700 °C were obtained from the nebulites using the geothermobarometer of Fonarev and Grapchikov (1984) (Table 7).

Tonalites

The chemical composition of plagioclase, pyroxene and biotite from the foliated tonalite (for comparison, also from the pyroxene granite) are shown in Table 8. The high titanium content (TiO_2 ~ 4 wt%) of biotite is probably the result of intense contact metamorphism in the area.

Table 8. Microprobe analyses of minerals from pyroxene granite and foliated tonalite adjacent to the Luopa quartz monzonite. Rock types: 1 = pyroxene granite, 2 = foliated pyroxene tonalite. Oxygen: Opx 6, Bt 11, Pl 8, Cpx 6.

Rock type	1	1	1	2	2	2
Sample	140-C	140-C	140-C	258	258	258
Distance between sample and the quartz monzonite	50 m	50 m	50 m	25 m	25 m	25 m
Mineral	Opx	Bt	Pl	Cpx	Bt	Pl
SiO_2	49,8	34,7	59,6	54,5	36,4	60,7
TiO_2	0,1	4,4	0,0	0,1	4,0	0,0
Al_2O_3	0,3	13,6	26,2	0,7	14,5	26,0
FeOtot	41,4	26,6	0,1	15,7	22,8	0,1
MnO	0,9	0,1	0,0	0,6	0,2	0,0
MgO	7,8	5,6	0,0	10,2	9,1	0,0
CaO	0,8	0,1	7,6	20,3	0,0	7,2
Na_2O	0,1	0,5	7,5	0,3	0,2	7,5
K_2O	0,0	8,7	0,3	0,0	9,1	0,3
Total	101,2	94,3	101,3	102,4	96,3	101,8
Si	2,02	2,79	2,63	2,03	2,80	2,66
Al	0,01	1,29	1,36	0,03	1,31	1,34
Ti	0,00	0,27	0,00	0,00	0,23	0,00
Mg	0,47	0,67	0,00	0,57	1,04	0,00
Fe	1,40	1,79	0,00	0,49	1,46	0,00
Mn	0,03	0,01	0,00	0,02	0,01	0,00
Ca	0,04	0,01	0,36	0,81	0,00	0,34
Na	0,01	0,08	0,64	0,02	0,03	0,64
K	0,00	0,89	0,02	0,00	0,89	0,02

Triangular phase diagrams

Pelitic rocks and intermediate nebulites

The A'FM-diagram of Reinhardt (1968) and ACF-diagram of Eskola (1915) for granulite-grade metamorphism are used to illustrate relations between chemical composition and mineral assemblages of the rocks from the contact aureole. The present chemical data does not separate the Fe^{2+} and Fe^{3+}: iron is assumed to be Fe^{2+}.

Inner contact aureole. The A'FM diagram shows that the chemical composition of both diatexites and psammitic metapelites containing orthopyroxene (samples coded as 2 and 3 in Figure 17a) plot in the field of Grt-Bt-Opx and, moreover, relatively close to orthopyroxene. In addition, the crystallization of orthopyroxene in the orthopyroxene-cordierite-garnet diatexite (sample 1), which chemical composition is pelitic, is indicated by tie-lines Grt-Opx-Crd-Bt (Fig. 17a). In the Figure, compositional plots of orthopyroxene-garnet pairs from intermediate nebulite are also shown, even though the chemical composition of this rock is not available. Relations between rock compositions and mineral assemblage indicate that the inner contact aureole underwent granulite-grade metamorphism.

Outer contact aureole. On the A'FM diagram, the chemical data for garnet-cordierite-mica diatexite and tightly banded garnet-cordierite-sillimanite mica gneiss (samples 4 and 5 in Figure 17a) plot inside tie-lines between Grt-Crd-Bt. This is in agreement with observed mineral assemblage (Grt-Crd-Bt) in the rocks. However, these rocks indicate to a slightly lower metamorphic grade than the orthopyroxene-bearing diatexite (sample 1) of the inner contact aureole, because the rocks (samples 1, 4 and 5) have approximately similar chemical composition, but different mineral assemblages (Fig. 17a).

Tonalites

Inner and outer contact aureole. On the ACF diagram by Eskola (1915) for the pyroxene-hornfels facies (i.e. granulite facies, see Miyashiro 1994), all chemical compositions (samples 1-4 in Figure 17b) for pyroxene-bearing tonalites plot near to each other and approximately on Pl-Opx tie-line. However, sample 1 did not contain hornblende and clinopyroxene, and it is situated at a distance of one meter from quartz monzonite. This indicates a higher metamorphic conditions (in the inner contact aureole) for sample 1 than the other rock samples.

The other specimens (2-4), which are hornblende- and clinopyroxene-bearing, were taken from the outcrops about 20-30 meters from the quartz monzonite. Thus, these specimens refer to the outer contact aureole. Many assemblages such as Pl-Cpx-Hbl, Cpx-Hbl-Opx and Pl-Hbl-Opx are observed in the tonalites. This is expected because the chemical composition of the rocks overlap tie-lines of these assemblages (see Fig. 17b).

P-T path and width of the reheating of crust

A P-T path suggested for the contact metamorphism of the Luopa quartz monzonite is shown in Figure 18, which also includes a schematic petrogenetic grid modified after Pattison (1989, figure 5). The progressive part of the path is drawn with regard to the occurrence of garnet-cordierite-K-feldspar and cordierite-orthopyroxene-K-feldspar assemblages in the diatexitic metapelites of the contact aureole. The irregular form of the Luopa intrusion indicates a relatively high level of emplacement in the crust and, contemporaneously, a rather low lithospheric pressure of 2-4 kbar. Retrogressive andalusite in the contact aureole indicates cooling in the field of andalusite stability (Fig. 18). In the contact metamorphic facies series classification by Pattison and Tracy (1991), the present contact aureole resembles the type 2, in which high-grade assemblages typically contain biotite, garnet, sillimanite, K-feldspar and cordierite.

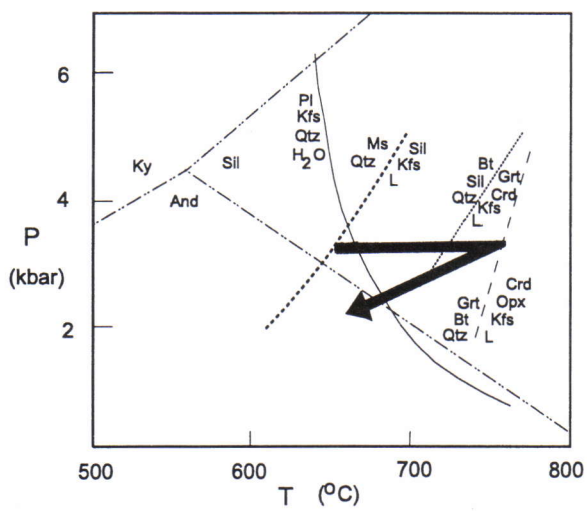


Fig. 18. Shematic, partial petrogenetic grid for a portion of KFMASH (modified after Pattison 1989). The solid arrow represents the approximate P-T path of the contact metamorphism around the Luopa intrusion.

The width of a thermal aureole around an intrusion is a complex function of the size and shape of the intrusion, its temperature, the composition of host rocks, and whether heat loss occurred by conduction or mass transfer (Parmentier and Schedl 1981). Granulite-facies contact metamorphism does not generally extend far from an intrusion (e.g. Tilley 1924, Miyashiro 1994). The temperature of contact metamorphism, at a distance of 1 kilometre from granitic intrusions with a ca. 10 km diameter, is about 500 °C (see Delaney 1988, Bowers et al. 1990, Voll et al. 1991). For example, in the Ardara aureole (Donegal, Ireland), the metamorphic temperature was less than 500 °C 0.4 km from the contact (Kerrick 1987). Moreover, if the country rocks have a metamorphic grade similar to that of the contact aureole, the outer limit of the aureole cannot be clearly defined (Dickerson and Holdaway 1989, Kerrick 1991). Thus, at the contacts of the Luopa intrusion – even if the intrusion had a high temperature, as indicated by the presence of fayalite – the effects of the highest grade contact metamorphism should not extend far because the contacts are often the edges of relatively narrow (5-100 m) quartz monzonite dykes.

Haudenschild (1995) reported that the cooling of contact aureole of the Vaaraslahti pyroxene intrusion, eastern Finland, was relative slow. However, it seems that the regional metamorphism and mangeritic magmatism was coevel in the Vaaraslahti area (Hölttä 1995) and a time difference of these shorter than at the Luopa area. Thus, the cooling rate of the Vaaraslahti intrusion area cannot be applied to the Luopa area, even though the two intrusions are relatively similar in petrography and metamorphic environment.

The regional high-grade metamorphic terrane of the Ilmajoki region includes migmatitic garnet-cordierite-sillimanite mica gneisses and pyroxene-bearing intermediate metavolcanic rocks several kilometres from the studied intrusion (Mäkitie et al. 1991, Lehtonen and Virransalo, in press). These areas do not contain intrusive pyroxene granitoids and, thus, the authors suggest that the high-grade regional metamorphic isograds of the region are not caused by the Luopa intrusion.

DISCUSSION

Some interesting differences between the areas of contact metamorphism and regional metamorphism are observed. Open to tight folds, common in the high-grade regional metamorphic metapelites, are absent in the vicinity (<30 m) of the Luopa intrusion. Metapelites near (at distances less than 20 m) the quartz monzonite are characterized by diatexitic textures or tightly banded mica gneisses (Figs. 4d, 5a and 5b) – both these features are not found elsewhere. The microtextures observed in thin sections differ between the regional metamorphic terrane and the inner contact aureole (Fig. 19); microscopic foliation is practically absent in the diatexites of the inner contact aureole. In particular, the orthopyroxene-garnet assemblages of the contact aureole indicate, without doubt, that different metamorphic conditions prevailed in the area. Some chemical properties in minerals, such as titanium content in biotite, differ between the metapelites of the contact aureole and metapelites of regional metamorphic terrane (Mäkitie 1999). This also indicates differences in metamorphic evolution for the areas.

The genesis of pyroxenes in the neighbouring pyroxene granite is problematic, because orthopyroxene occurs in the granite as far as 1.5 km from the Luopa quartz monzonite. The pyroxenes seem to have a magmatic origin, because crystallization of metamorphic orthopyroxene in granitoids requires temperatures of at least 750 °C (cf. Bucher and Fry 1994). This is supported by the fact that intense contact metamorphic textures are absent in the pelitic rocks occurring at a distance of 1.5 km from the quartz monzonite of Luopa.

Currently, three main types of post-kinematic granitoids within the CFGC are distinguished (Rämö et al. 1999). According to Nironen et al. (2000, see also Rämö et al. 1999) the Luopa intrusion belongs to Type 3. The Luopa intrusion has unusually high TiO_2 and Fe/Mg compared to most of the post-kinematic pyroxene granitoids. Typically, the trace element concentrations often correlate with changes in the SiO_2 content of pyroxene granitoids. For example, Sr in the SiO_2-rich pyroxene granite at Vaaraslahti, eastern Finland, is ca. 100 ppm. The corresponding value in quartz syenites (SiO_2~ 64 wt%) of the same area is ca. 150 ppm (Lahti 1995), but the Sr concentration in the SiO_2-poor Luopa quartz monzonite is ca. 400 ppm. However, the Luopa intrusion has lower Rb/Sr than Central Finland Granitoid Complex post-kinematic granitoids, in general, with the same SiO_2 content (cf. Nironen et al. 2000). In addition to the Luopa

a) b)

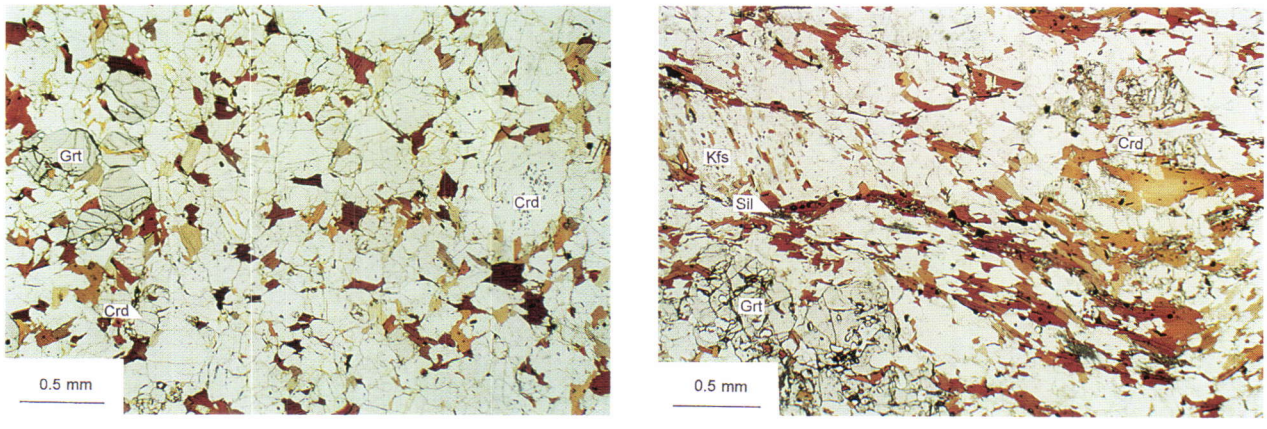


Fig. 19. Comparison of the microtextures between contact and regional metamorphic metapelites. **(a)** Garnet (Grt) -cordierite (Crd) -mica diatexite. Outer contact aureole; **(b)** migmatitic garnet (Grt) - cordierite (Crd) - sillimanite (Sil) mica gneiss from the regional metamorphic terrane, about 4 km northwest of the Luopa quartz monzonite. K-feldspar (Kfs) includes parallel small biotite flakes.

intrusion, fayalite is detected in the Jämsä intrusion (Elliott et al. 1998).

The evolution of the high-grade regional metamorphic terrane surrounding the Luopa quartz monzonite differs slightly from another high-grade terrane (with no mangeritic magmatism) located some 50 km north (Mäkitie et al., this volume). This northern terrane show features of isothermal decompression (ITD), compared to the areas surrounding the Luopa quartz monzonite, where features of an isobaric cooling (IBC) are encountered (Mäkitie 1999, Mäkitie et al., this volume). It is suggested that the Luopa quartz monzonite is derived from a relatively dense parental magma body. Due to this density difference, the local uplift of the crust was slowed and thus, partly caused the local IBC-path.

The authors suggest the following magmatic history for the Luopa quartz monzonite. During the evolution and formation of granitoid magmas in the base of the crust, olivine and pyroxene crystallized and accumulated at the bottom of the magma chamber. This unusual Fe-rich material (crystal mush) remained in the lower crust for a few million years. Finally it was emplaced to higher levels forming the Luopa fayalite-augite quartz monzonite. The average density of the dark green quartz monzonite variety is 2.76. During crystallization of the Luopa intrusion, it is likely that oxygen fugacities varied in the magma chamber, because there are light-coloured and dark green types (without clear cutting relations) of quartz monzonite in the intrusion.

The magma evolution model, including the relative position of granitoids in the orogen shown in Figure 14, corroborated by the fact that the Luopa quartz monzonite intersects pyroxene granite and foliated tonalites in the study region. The Luopa quartz monzonite and the pyroxene granite form a circular plutonic complex, suggesting that the intrusions are comagmatic. There are, however, some arguments against the possibility that the pyroxene granite and quartz monzonite intrusions have a common source. First, the unusual contact relation (SiO_2 -poor quartz monzonite brecciates SiO_2 -rich granite) is peculiar for order of emplacement and evolution of granitoid magmas. Secondly, the scattering of some elements (e.g. Mg, Zr and Th vs. SiO_2, Figure 9) in the pyroxene granite, when compared to the same elements in the quartz monzonite, does not support derivation from a common parental magma.

By age, the Luopa quartz monzonite is associated with the post-collisional magmatism of the Svecofennian evolution as described by Lahtinen (collisional magmatism at 1.89-1.88 Ga; 1994). The ages of the Svecofennian 'magmatic groups', however, overlap in space and time and pose problems when applied to plate tectonic models (e.g. Nironen et al. 2000). For example, the post-kinematic intrusions within CFGC probably show an overall decrease of crystallization age from east to west (Nironen et al. 2000). The Luopa stock is ca. 15 Ma younger than the Vaaraslahti pyroxene intrusion (see Hölttä 1995) in the tectonically complex Raahe-Ladoga zone, but the peak of regional metamorphism in the areas has been contemporaneous, 1.89-1.88 Ga. This greater time difference between the metamorphic peak and mangeritic magmatism at Luopa indicates cooler country rocks (and higher crustal level) during emplacement.

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Appendix 1. Geographical locations of the analysed samples taken from the Luopa region.

Sample	Rock type	Map	Northing	Easting
H-202	Quartz monzonite	222106	6947340	2432780
H-11A	Quartz monzonite	222106	6946800	2431800
H-205B	Quartz monzonite	222103	6943620	2427750
H-11C	Quartz monzonite	222106	6946900	2431800
P-238B	Quartz monzonite	222106	6947200	2432500
H-167	Quartz monzonite	222103	6948120	2427640
H-199	Quartz monzonite	222103	6946630	2429400
H-196	Quartz monzonite	222103	6946930	2429920
P-233	Quartz monzonite	222103	6947400	2433100
S-652	Quartz monzonite	222103	6947900	2425800
H-18	Quartz monzonite	222106	6946750	2430700
P-290	Quartz monzonite	222106	6947700	2430600
P-234	Quartz monzonite	222106	6947300	2432800
H-254	Quartz monzonite	222103	6946420	2429230
H-1	Quartz monzonite	222106	6947050	2430080
141B	Quartz monzonite	222103	6944000	2427850
H157	Quartz monzonite	222103	6949840	2425710
NKK178	Quartz monzonite	222103	6947500	4260000
V-256	Quartz monzonite (light-coloured)	222106	6949900	2431500
H200B	Quartz monzonite (light-coloured)	222201	6950730	2428180
H201	Quartz monzonite (light-coloured)	222106	6944760	2432050
H-11D	Quartz monzonite (light-coloured)	222106	6946800	2431900
H-11E	Quartz monzonite (light-coloured)	222106	6946700	2431800
P-328A	Quartz monzonite (light-coloured)	222106	6947200	2432500
H-11B	Quartz monzonite (light-coloured)	222106	6946800	2431800

Sample	Rock type	Map	Northing	Easting
H-141A	Pyroxene granite	222103	6944000	2427850
H-233	Pyroxene granite	222103	6946100	2426640
H-266	Pyroxene granite	222103	6945000	2427750
H-205A	Pyroxene granite	222103	6943620	2427750
H-140	Pyroxene granite	222103	6944480	2427820
H-171	Biotite-hornblende granite	222103	6945100	2424450
H-153	Biotite-hornblende granite	222103	6941600	2425700
600-HM	Foliated pyroxene tonalite	222103	6949990	2426010
52-LIS	Foliated pyroxene tonalite	222103	6949970	2428220
53-LIS	Foliated pyroxene tonalite	222103	6949670	2428420
54-LIS	Foliated pyroxene tonalite	222103	6949590	2429310
166D	Diatexitic metapelite	222103	6948030	2427770
171-E	Diatexitic metapelite	222103	6945120	2424840
171K	Diatexitic metapelite	222103	6945090	2424820
722	Tightly banded mica gneiss	222204	6950560	2428420
167-L	Psammitic metapelite	222103	6948020	2427780

Appendix 2. Chemical compositions of 18 dark green quartz monzonite samples (coded as 1) and of 7 light-coloured quartz monzonite samples (coded as 2) from the Luopa intrusion. * = trace elements analysed at GTK, ¤ = trace elements analysed at VTT, # = trace elements analysed at RR, ** = analysed by OES at GTK.

Rock type	1	1	1	1	1	1	1	1	1	1	1	1	1
Laboratory	#	¤	#	¤	¤	¤	¤	¤	¤	¤	¤	¤	¤
Sample	H-202	H-11A	H-205B	H-11C	P-238B	H-167	H-199	H-196	P-233	S-652	H-18	P-290	P-234
SiO_2 wt%	59,70	62,52	62,00	61,17	60,38	57,12	57,95	57,24	59,90	59,31	61,32	58,19	58,55
TiO_2	0,68	0,58	0,63	0,55	0,78	1,02	0,93	0,99	0,71	0,82	0,69	0,86	0,85
Al_2O_3	17,60	16,62	16,70	16,62	15,75	15,34	15,39	15,28	16,34	16,26	16,73	15,91	15,74
Fe_2O_3tot	8,28	6,84	7,33	6,25	8,76	12,73	11,64	11,92	8,31	10,07	7,79	10,18	9,77
MnO	0,15	0,11	0,15	0,10	0,14	0,22	0,19	0,20	0,13	0,17	0,13	0,17	0,15
MgO	0,46	0,52	0,53	0,54	0,65	0,81	0,74	0,95	0,66	0,63	0,64	0,81	0,64
CaO	3,81	3,24	3,75	3,06	3,63	4,56	4,45	4,58	3,72	4,33	3,60	3,99	3,93
Na_2O	3,41	3,25	3,47	2,92	3,37	3,05	3,06	3,07	3,16	3,59	3,34	3,17	3,03
K_2O	5,09	5,58	4,75	6,34	5,15	3,87	3,96	4,07	5,69	4,21	5,32	5,51	4,74
P_2O_5	0,42	0,15	0,44	0,14	0,18	0,31	0,27	0,29	0,21	0,22	0,18	0,25	0,24
S	0,03	-	0,04	-	-	-	-	-	-	-	-	-	-
Total	99,63	99,41	99,79	97,69	98,79	99,03	98,58	98,59	98,83	99,61	99,74	99,04	97,64
T-T index	71,65	75,62	73,48	75,29	72,68	63,98	65,22	64,49	72,5	68,89	73,45	70,13	68,29
A/CNK	0,98	0,96	0,95	0,96	0,89	0,88	0,88	0,86	0,90	0,89	0,94	0,86	0,91
NK/A	0,63	0,69	0,65	0,70	0,71	0,60	0,61	0,62	0,70	0,64	0,67	0,70	0,64
Rb/Sr	-	0,18	0,25	0,20	0,23	0,09	0,16	0,15	0,20	0,13	0,15	0,09	0,22
Rb/Ba	0,02	0,03	0,02	0,03	0,04	0,02	0,03	0,02	0,03	0,02	0,02	0,02	0,02
K/Rb	845	572	657	572	411	765	457	637	562	582	613	994	442
Ba ppm	2850	2630	2720	2830	2570	1950	2560	2130	2900	2930	2910	2760	3950
Rb	50	81	60	92	104	42	72	53	84	60	72	46	89
Sr	-	440	240	450	450	470	460	360	430	470	490	520	410
Th	-	11	-	11	11	4	4	2	7	2	9	3	9
Zn	190	134	170	-	-	-	-	339	286	93	331	-	161
Zr	880	580	660	520	750	880	750	690	590	760	600	820	740
Nb	50	-	40	-	-	-	-	-	-	-	-	-	-
La	130	193	-	169	187	48	39	64	104	35	122	79	191
Ce	290	367	30	350	400	140	106	160	235	97	270	188	352
Nd	-	158	-	153	186	99	69	96	110	69	126	95	155
Sm	-	25	-	24	29	24	19	23	23	17	21	20	33
Eu	-	5	-	6	6	1	7	7	7	8	7	7	7
Tb	-	2	-	2	2	3	2	3	2	2	2	2	2
Yb	-	4	-	4	6	8	7	7	6	7	5	6	6
Lu	-	0,5	-	0,5	0,7	0,9	0,8	0,8	0,6	0,8	0,5	0,6	0,7
Y	30	-	20	-	-	-	-	-	-	-	-	-	-
Sn	10	-	10	-	-	-	-	-	-	-	133	81	55

Appendix 2. Continued

Rock type	1	1	1	1	1	2	2	2	2	2	2	2
Laboratory	¤	¤	¤	#	*	¤	#	#	¤	¤	¤	¤
Sample	H-254	H-1	141B	H157	NKK178	V-256	H200B	H-201	H-11D	H-11E	P-238A	H-11B
SiO_2 wt%	57,90	58,06	59,05	58,80	57,78	61,29	59,00	58,90	61,98	60,91	61,20	60,70
TiO_2	0,92	0,85	0,89	0,90	0,79	0,72	0,75	0,72	0,63	0,65	0,68	0,59
Al_2O_3	15,11	17,13	15,62	16,10	16,86	16,04	17,40	17,60	16,49	15,62	16,70	16,41
Fe_2O_3tot	11,47	11,32	11,45	10,15	10,16	9,08	9,21	9,00	7,22	7,58	7,69	6,83
MnO	0,20	0,19	0,20	0,21	0,19	0,15	0,14	0,16	0,12	0,13	0,13	0,10
MgO	0,65	0,87	0,46	0,62	0,42	0,63	0,70	0,53	0,63	0,63	0,59	0,61
CaO	4,32	3,80	4,41	4,49	4,81	3,94	4,44	3,92	3,05	3,37	3,35	2,84
Na_2O	2,89	4,04	3,46	3,20	3,77	3,25	3,39	3,48	2,87	2,68	3,24	2,97
K_2O	4,36	4,67	3,58	3,44	3,93	4,73	4,06	4,83	6,02	5,47	5,53	6,20
P_2O_5	0,25	0,24	0,29	0,33	0,24	0,24	0,54	0,47	0,16	0,16	0,17	0,15
S	-	-	0,06	0,07	-	-	0,05	0,04	-	-	-	-
Total	98,07	101,17	99,47	98,31	98,95	100,07	99,68	99,65	99,17	97,20	99,28	97,40
T-T index	66,06	70,24	67,09	64,97	66,73	71,63	67,39	70,3	75,08	71,95	73,88	74,89
A/CNK	0,87	0,92	0,89	0,94	0,88	0,91	0,96	0,97	0,98	0,95	0,96	0,98
NK/A	0,63	0,68	0,61	0,56	0,62	0,65	0,57	0,62	0,68	0,66	0,68	0,71
Rb/Sr	0,13	-	0,33	0,15	0,12	0,20	0,16	0,21	0,19	0,24	0,18	0,20
Rb/Ba	0,02	0,02	0,04	0,02	0,02	0,02	0,02	0,02	0,03	0,04	0,03	0,03
K/Rb	603	546	286	714	502	462	674	668	561	428	581	572
Ba ppm	2560	3580	2970	2140	3750	3860	2720	2860	2850	2970	2630	3160
Rb	60	71	104	40	65	85	50	60	89	106	79	90
Sr	450	-	320	260	548	420	320	280	470	450	440	460
Th	2	7	3	-	1	2	-	-	16	15	13	12
Zn	214	-	253	260	-	-	220	190	-	-	-	-
Zr	800	-	1110	1140	-	700	1060	910	690	610	620	550
Nb	-	-	-	60	-	-	50	50	-	-	-	-
La	40	-	59	10	34	36	10	50	229	209	184	182
Ce	117	-	-	60	-	95	80	120	462	423	378	390
Nd	84	-	-	-	-	68	-	-	196	178	-	170
Sm	23	-	-	-	-	17	-	-	30	28	26	24
Eu	6	-	-	-	-	7	-	-	5	6	6	6
Tb	3	-	-	-	-	2	-	-	2	2	2	2
Yb	8	-	-	-	-	6	-	-	4	4	5	4
Lu	0,8	-	-	-	-	0,6	-	-	0,5	0,4	0,6	0,5
Y	-	50	94**	50	36	-	30	40	-	-	-	-
Sn	-	-	10	20	-	-	10	10	-	-	-	90

Appendix 3. Chemical compositions of pyroxene granites and biotite-hornblende granites surrounding the Luopa quartz monzonite. Pyroxene granite (coded as 1) and biotite-hornblende granite (coded as 2). ¤ = trace elements analysed at the VTT, # = trace elements analysed at RR. ** = analysed by OES at GTK.

Rock type	1	1	1	1	1	2	2
Laboratory	¤	¤	¤	#	¤	¤	¤
Sample	H-141-A	H-233	H-266	H-205-A	H-140	H-171	H-153
Distance between sample and the quartz monzonite	10 m	100 m	50 m	1 m	100 m	500 m	2 km
Main mafic minerals	Cpx-Hbl-Bt	Cpx-Hbl-Bt	Cpx-Hbl-Bt	Cpx-Hbl-Bt	Cpx-Hbl-Bt	Bt-Hbl	Bt-Hbl
SiO_2 wt%	73,17	74,65	69,31	73,10	66,93	74,76	72,80
TiO_2	0,32	0,21	0,40	0,29	0,41	0,14	0,28
Al_2O_3	13,21	13,03	14,64	13,30	15,15	13,55	13,49
Fe_2O_3tot	3,20	2,16	4,12	3,13	3,45	1,74	3,29
MnO	0,05	0,04	0,06	0,06	0,03	0,03	0,03
MgO	0,11	0,12	0,40	0,15	0,67	0,08	0,28
CaO	1,34	1,01	1,73	1,41	2,66	0,76	1,36
Na_2O	2,64	2,24	2,66	2,39	2,93	2,57	2,35
K_2O	5,47	6,15	6,13	5,69	5,17	6,11	5,73
P_2O_5	0,05	0,03	0,09	0,09	0,18	0,03	0,10
S	0,01	-	-	0,01	-	-	-
Total	99,57	99,64	99,54	99,62	97,58	99,77	99,71
T-T index	88,65	91,12	85,06	88,28	79,25	92,53	87,90
A/CNK	1,04	1,07	1,03	1,05	0,99	1,11	1,08
NK/A	0,78	0,79	0,75	0,72	0,69	0,80	0,75
Rb/Sr	1,25	1,45	0,72	-	0,27	3,93	1,19
Rb/Ba	0,07	0,06	0,05	-	0,07	1,26	0,14
K/Rb	329	396	414	-	477	215	307
Rb ppm	138	129	123	-	90	236	155
Sr	110	89	170	170	330	60	130
Ba	2030	2130	2270	2190	1320	187	1070
Th	4	3	2	-	2	12	18
Zn	-	-	-	80	-	-	-
Zr	560	240	420	420	240	130	260
Nb	20	10	20	-	-	10	10
La	49	52	37	-	39	-	103
Ce	-	-	-	-	88	-	-
Nd	-	-	-	-	51	-	-
Sm	-	-	-	-	11	-	-
Eu	-	-	-	-	2	-	-
Tb	-	-	-	-	1	-	-
Yb	-	-	-	-	3	-	-
Lu	-	-	-	-	0,3	-	-
Y	42**	-	31**	-	-	-	-

Appendix 4. Abbreviations used in the text. The abbreviations follow the usage of Kretz (1983).

Abbreviation	Mineral
Aln	allanite
And	andalusite
Aug	augite
Bt	biotite
Cpx	clinopyroxene
Crd	cordierite
Fa	fayalite
Grt	garnet
Hbl	hornblende
Hc	hercynite
Ilm	ilmenite
Kfs	K-feldspar
Ky	kyanite
Ms	muscovite
Opx	orthopyroxene
Pl	plagioclase
Qtz	quartz
Sil	sillimanite
St	staurolite

Appendix 5. Chemical composition of rocks from the contact aureole around the Luopa quartz monzonite. Diatexitic metapelite (coded as 1), tightly banded mica gneiss (coded as 2), psammitic metapelite (coded as 3) and foliated pyroxene tonalite (coded as 4). # = trace elements analysed at RR, ** = analysed by OES at GTK.

Rock type	1	1	1	2	3	4	4	4	4
Laboratory	#	#	#	#	#	#	#	#	#
Sample	166d	171-E	171K	722	167-L	600-HMM	52-Lis	53-Lis	54-Lis
Distance between sample and the quartz monzonite	2 m	10 m	0.2 m	20 m	2 m	1 m	25 m	30 m	25 m
Main mafic minerals	Grt-Crd-Bt(-Opx)	Grt-Crd-Bt	Opx-Kfs-Bt(-Grt)	Grt-Crd-Sil-Bt	Opx-Grt-Bt	Opx-Bt	Cpx-Hbl-Bt	Cpx-Hbl-Bt	Cpx-Hbl-Bt
SiO_2 (wt-%)	56,30	51,05	60,10	48,10	64,2	58,62	64,97	68,28	63,4
TiO_2	0,98	1,06	0,79	1,08	0,81	1,26	0,52	0,33	0,52
Al_2O_3	20,20	22,79	16,70	22,50	16,80	17,48	16,13	15,23	16,29
FeOtot	8,75	11,18	6,73	12,83	5,57	8,37	5,43	4,23	5,55
MnO	0,12	0,12	0,04	0,10	0,10	0,10	0,08	0,08	0,10
MgO	4,28	4,96	2,27	5,13	2,06	2,72	1,82	0,95	2,04
CaO	2,45	1,45	2,20	1,30	4,60	4,72	4,43	3,36	4,43
Na_2O	2,48	1,79	2,83	1,53	3,20	3,46	3,21	2,97	3,27
K_2O	2,91	3,60	5,52	3,55	1,24	1,79	2,47	3,72	3,45
P_2O_5	0,11	0,10	0,05	0,26	0,22	0,60	0,27	0,19	0,29
S	0,11	0,29	0,11	1,76	0,52	0,17	0,04	0,04	0,03
Total	98,69	98,39	97,34	98,14	99,32	99,29	99,37	99,38	99,37
A/CNK	-	-	-	-	-	1,08	1,01	1,01	0,95
NK/A	-	-	-	-	-	0,44	0,49	0,59	0,56
Sr	240	140	170	100	250	190	280	240	290
Ba	1090	1020	1020	890	640	350	600	1170	680
Th	-	20	-	-	-	-	-	-	-
Zn	-	-	-		-	140	50	100	70
Zr	180	210	210	180	160	130	140	200	140
Nb	-	20	-	-	-	-	-	-	-
V	280	230	130	270	190	-	-	-	-
La	-	30	-	-	-	-	-	-	-
Ce	-	80	-	-	-	-	-	-	-
Y	-	34**	-	-	-	-	-	-	-

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