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Geochemical characterisation of Archaean granitoids in eastern Finland

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Abstract:

The oldest rocks in Finland are the Archaean grey gneisses of eastern and northern Finland. The Archaean of the Karelian craton spans about 1000 Ma of crustal growth and evolution and forms the core of the Fennoscandian shield. The Karelian province is a complex patchwork of different rock types. The individual formations are of small territorial extent in accordance with often postulated small Archaean plates. Overall, the Karelian craton is a granitoid-greenstone terrain with prevailing TTGs and younger granites, which show increasing level of potassium. The craton also includes a distinct sodic variety of granites that combines features of classical Archaean TTGs and late Archaean high-K granites. A minor number of Mg-rich lithologic units, including adakites and sanukitoids, are reported as well. A small number of A-type granites, syenites and S-type granites are widely distributed and of local nature only. Peculiarly, a large number of TTGs is peraluminous. The formation of Karelian craton may be explained by accretion of small plates, perhaps during the late Archaean supercraton event in a process that at least in later stages included active plate marginal processes.

Avainsanat – Nyckelord – Keywords
Archaean geochemistry, TTG, granitoid

Säilytyspaikka – Förvaringställe – Where deposited

Muita tietoja – Övriga uppgifter – Additional information

Geochemical characterisation of Archaean granitoids in eastern Finland

1. Introduction

Archaean nucleus of Fennoscandian shield covers wide area from northern Sweden and Norway across central and southeastern Finland and into northwestern Russia. It has become convenient to consider the Archaean of the Fennoscandian shield in terms of five crustal provinces: Murmansk, Kola, Karelia, Belomoria and Norrbotten (Hölttä et al., 2008). Karelian Province is subdivided into Western Karelian, Central Karelian and Vodlozero terrains. The western part of Karelian domain is the subject of this work. The domain exceeds 200 000 km^2 in Finland and adjacent Russia. It is flanked to the northeast by the Belomorian mobile belt. The southwestern contact is with the Svecofennian province, which is Proterozoic in age.

The Karelian domain comprises an extensive mosaic of tonalite-trondhjemite-granodiorite (TTG) areas, greenstone belts and Proterozoic rocks intruded into or deposited on the Archaean crust. The Archaean rocks crop out in the north-eastern part of Finland between Proterozoic formations (Fig. 1). The Archaean rocks have a complex pattern of ages from rare occurrences of >3.0 Ga to the major population of ~2.9 to 2.7 Ga (Sorjonen-Ward and Luukkonen, 2005).

Archaean granitoids form three major and a number of minor associations. Tonalite-trondhjemite-granodiorite (TTG) association dominate early Archaean. Granitoids diversify in the Neoarchaean (Moyen et al., 2003; Mikkola et al., 2012). The term TTG was introduced by Jahn et al. (1984). TTGs have become the defining feature of Archaean terranes and in general constitute the gneissic basement of all Archaean cratons. They volumetrically dominate Archaean crust and may represent up to 90 % of the overall juvenile continental crust of the Archaean cratons worldwide (e.g. Rapp et al., 1991; Springer, 1997; Martin, 2005; Smithies et al., 2009). Generation of large volumes of TTGs represents the first and essential step in generation of continental crust.

Archaean granites are biotite-bearing monzo- to syenogranites to granodiorites (Moyen et al., 2003). In most cases they represent the products of melting of older TTG associated with a highly enriched mantle-derived component (Watkins et al., 2007). Most Archaean granites were emplaced towards the end of Archaean and appear to be temporarily linked to craton stabilisation (Kusky et al., 1999). Archaean leucogranites form a minor family. Leucogranites are characterised by high K/Na ratios and moderately fractionated REE patterns with strongly negative Eu anomalies and typically high Th and Rb (Moyen et al., 2003).

Adakites have become somewhat controversial as the term has acquired ambiguous meaning (Hastie, 2020) as it has become used to describe a too large groups of rocks characterised only by high Sr/Y and La/Yb (Moyen, 2009). This work follows the definition of Martin et al. (2005) for adakites and includes the definition of Hastie et al. (2010).

High-Mg diorites and granodiorites termed Archaean sanukitoids (Smithies et al., 2000; Moyen, 2005) form a minor component of Archaean crust. They have been identified in the Archaean of Finland, and there is also a number of occurrences in neighbouring Russia (Lobach-Zhuchenko et al. 2004; Halla, 2008, Heilimo et al., 2010). They share similarities with and are often coeval with another group of mafic granitoids, the quartz diorites (Mikkola et al., 2012). Both are interpreted as products of variously metasomatized mantle.

The petrogenesis of TTGs has received considerable attention in recent years. Experimental petrology has demonstrated that sodic granitoid melts can be produced by partial fusion (up to 40 % melt) of hydrated metabasalts at pressures high enough to stabilize garnet (Rapp, 1995; Smithies, 2000). There is, indeed, wide agreement on metabasalt being the source of TTGs. The depth of melting, the mineralogy of the source geological and geotectonic scenarios of formation are still under debate.

Historically, the subduction (or hot subduction) model dominates the literature on TTG petrogenesis (Martin, 1986; 1987) subduction theory identifying TTGs as melts of basaltic slabs (Martin, 1998, 2005; Smithies et al., 2003; de Wit, 1997) but the debate is ongoing (Condie, 2003; Zegers et al., 2001). Another theory that is gaining acceptance is melting of thickened oceanic crust (Mikkola et al., 2012). Oceanic crust could have thickened through tectonic imbrications and stacking of oceanic crust (Smithies, 2000). Oceanic crust could have thickened through mafic underplating (Bédard, 2006; Champion and Smithies, 2007). Martin et al. (2014) advocated for TTG origin in subduction of plateau basalts. Condie (2005) favoured melting of lower portions of oceanic plateaus.

The geochemical diversity of TTG association points towards a great range of melting depths (and pressures). But this point has been questioned (Laurent et al., 2020). TTGs, adakites, sanukitoids, quartz diorites are product of mantle or oceanic crust and mark production of juvenile continental crust during which the recycling of continental crust was a subordinate process (Martin, 1993).

The changes observed from early to late Archaean TTG reflect the change in nature and depth of melting as well as progressively increasing degree of interaction between the subducting slab and mantle wedge peridotite (Martin, 1998; Smithies et al., 2009). The diversification of Archaean granitoids towards the end of Archaean eon stems from increased interactions between different reservoirs. Late Archaean hybrid granitoids (more than one source; mantle and crust) have been identified from all cratons; Karelia as well. They form a varied family that cannot be defined always by geochemical grounds alone due to myriad of sources and geochemical processes abound. Many types are present on all cratons but literature reports also unique (so far) types and local peculiarities (e.g. Bulai pluton, Limpopo belt, South Africa; Likamännikkö intrusion, Karelian province, Finland).

1.1 The aim of the study

The aim of this study is to describe the geochemical characteristics of Archaean granitoids in Karelia and to find and identify units that share similarities in composition. An attempt is also made to find geographically continuous areas of bedrock and explore their interrelations.

1.2. Samples and analytical methods

The samples were collected between 1991 and 1995 using a stratified procedure where number of samples per area depends on variation of lithology seen on geological maps. The sampling sites were pre-selected using available maps 1:100 000 and 1:400 000. Where available, maps 1:100 000 where preferred and maps 1:400 000 were used in areas lacking more detailed geologic maps. Sampling was directed to outcrops and they had effect on sampling density as areas devoid of outcrops have lower sampling density. The samples were taken by portable mini-drill. The drill cores measured 15-20 cm in length and 2.5 cm in diameter.

The samples were prepared and analysed between 1992 and 2001 at the geochemical laboratory of the Geological Survey of Finland. The samples were analysed by X-ray fluorescence spectrometry, inductively coupled plasma atomic emission spectrometry, atomic absorption spectrometry with electrochemical atomisation and inductively coupled plasma mass spectrometry.

The samples, sampling procedure and further treatments were described in greater detail by Rasilainen et al. (2007). In this work, a subset of 702 samples of Archaean magmatic rocks from the GTK rock geochemical database of Finland is used (Fig. 1).

2. Geographical division of the Karelian province

Finland is situated in the central part of the Fennoscandian shield that crops out between the Caledonian mountain chain to the west and younger sedimentary rocks to the east and south. The Karelian province forms a coherent late Archaean (3.2-2.7 Ga) cratonic nucleus that covers an area in excess of 200 000 km^2 (e. g. Sorjonen-Ward and Luukkonen, 2005). The Karelian province is bounded by Belomorian province in the northeast and by the Proterozoic Svecofennian province in the southwest (e. g. Bibikova, 1996; Gaál, 1990). The Karelian province is characterised by extensive TTG terrains and narrow N-S trending greenstone and metasedimentary belts. Greenstone belt volcanic rocks (not included in this study) fall to distinct age groups for each belt (Hölttä et al., 2012).

The understanding of crustal evolution is impeded by the lack of defined collisional systems and their sedimentary basins reflecting the scale of the craton which is too small to preserve the complete orogenic record similar to modern-day collisional orogens with their associated foreland basins, passive margins, possible accreted arcs and hinterlands.

The available U-Pb age determinations show that most of the magmatic rocks formed between 2.9 and 2.6 Ga and a minor proportion dates at 2.9-3.5 Ga (e.g. Slabunov et al., 2006; Mutanen and Huhma, 2003). The isotopic ages in Finland do not, however, delineate any regional spatial patterns (e. g. Sorjonen-Ward and Luukkonen, 2005).

Karelian craton has been often considered a part of larger Archaean supercraton or a craton cluster (Bleeker, 2003). It is thought to have formed by accretion of crustal fragments of different ages in a manner somewhat similar to modern continent-continent collision. It is postulated that Archaean plates were small: its modern analogue could be western Pacific domain. It occupies only 8% of modern Earth surface but is made of more than 20 microplates (Yamamoto, 2008). The boundaries of smaller units are often difficult to constrain with accuracy as they are reworked during the Proterozoic orogeny and also intruded by younger, post-Archaean rocks.

The division of the Karelian province is adopted from Sorjonen-Ward and Luukkonen (2005). The term terrain is used in a descriptive sense. Small Archaean windows and lenses within the schist zones bounding the Archaean formations are possibly fragments detached from major Archaean blocks during the Proterozoic tectonic movements (Luukkonen et al., 1997). The granitoid complexes were formed during several stages and are distinguished by scale and conditions of granitoid formation (Slabunov et al., 2006).

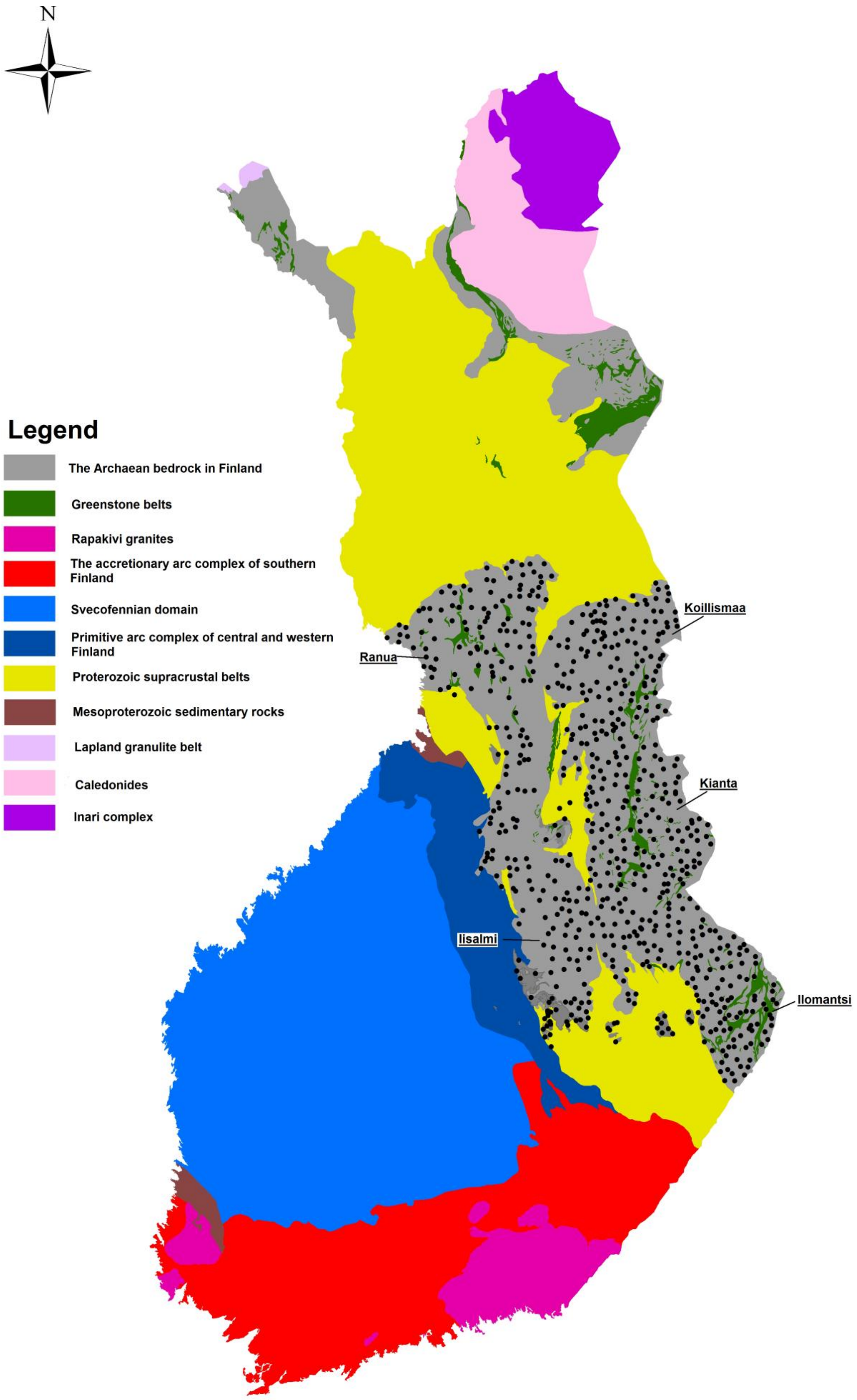

Figure 1: Simplified geological map of Finland with subdivision of Karelian province. Black symbols mark the locations of the samples of this study. Courtesy of Kalevi Rasilainen, GTK (2006).

2.1. Ranua terrain

Ranua terrain is a roughly triangular crustal block that forms the north-western corner of Karelian Domain. It is bounded by Proterozoic formations to north and west and separated by Hirvaskoski shear zone from East Finland terrain and by Oulujärvi shear zone from Iisalmi terrain. It is mainly composed of tonalitic to granitic gneisses and migmatites that enclose several greenstone belts. It contains the oldest rocks in Finland and in the Fennoscandian Shield so far identified, the Siurua trondhjemitic gneiss, dated to 3.5 Ga. Signs of even older crust come from a core of one zircon dated to 3.73 Ga (Mutanen and Huhma, 2003; Slabunov et al., 2006).

2.2. Iisalmi terrain

The Iisalmi terrain is located in the southwestern corner of the craton, bounded by Proterozoic domain to the west and separated from Ranua terrain by N-NE trending Oulujärvi shear zone. Iisalmi terrain is heavily deformed and may be entirely displaced by Proterozoic events relative to other terrains of the Karelian craton. The Iisalmi terrain contains evidence of some of the oldest and youngest Archaean events recorded on the Karelian craton: the early paleosomes dated at 3.2 Ga and the 2.6 Ga Siilinjärvi carbonatite complex (e.g. Lukkarinen, 2000). Seismic reflection studies have revealed that the Moho is unusually deep beneath the Iisalmi terrain (55-60 km). The unusual thickness is probably the result of post-Archaean tectonic events (Korsman et al., 1999).

2.3. Ilomantsi terrain

Ilomantsi terrain is located at the southern corner of the Karelian craton. It is composed of several greenstone belts (e.g. Sorjonen-Ward et al., 1997) and granodioritic and tonalitic plutons. The western part (Lieksa complex) includes numerous porphyritic granitoids. Migmatitic gneisses and granulite facies supracrustal enclaves are widespread.

2.4. Kianta terrain

Kianta terrain is located north of Ilomantsi and east of Iisalmi terrain. It is composed of several greenstone belts that bisect it from the north to south, and surrounding granite-gneiss terrains. Much of the southern part of Kianta consists of Nurmes paragneisses (Kontinen et al., 2007). Similarities between granitoids of Kianta and Ilomantsi terrains suggest that they have developed as a single terrain since at least 2.74 Ga (Sorjonen-Ward and Luukkonen, 2005).

2.5. Koillismaa terrain

Koillismaa terrain is located in the north-eastern corner of the study area. It is bounded by the Kuusamo schist belt to the north and west, and a discontinuous belt of Proterozoic mafic intrusions to the south. It is composed of quartz dioritic to trondhjemitic gneisses and pelitic paragneisses.

Ilomantsi, Kianta and Koillismaa terrains form a continuous unit and are sometimes treated together as East Finland Archaean complex. It continues eastward across the border into Russia. It is bounded in the north by the Proterozoic Kuusamo schist belt, in the west by Hirvaskoski shear zone, Kainuu schist belt, Nunnalahti-Holinmäki shear zone and North Karelian schist zone (Luukkonen and Sorjonen-Ward, 1998).

3. Classification of rocks

3.1. General geochemical features of Karelian granitoids

The Karelian gneisses included in this study range from basic to acid (51-77.8 wt.% SiO_2). All the rocks follow a calc-alkaline trend on the AMF diagram. They show variable LILE and LREE enrichment and different degrees of alteration. The degree of LREE enrichment (normalized La/Sm) varies between 1.32 and 21.6 with the mean of 6.0. Typically to crustal rocks they display relative negative anomaly of Nb and Ta in variation diagrams. On tectonic discrimination diagrams of Pierce et al. (1984) majority of samples plot in the VAG field while only a small minority in the WPG and syn-collisional fields. The data used in this work do not include the paragneisses (e.g. Kontinen et al., 2007) or the greenstone belts (e.g. Sorjonen-Ward, 1997). Karelian domain records almost a billion years of magmatism, initially as crustal growth and in the late Archaean increasing crustal reworking.

Tonalite-trondhjemite-granodiorite association represents the majority of preserved early and middle Archaean crust. Juvenile crustal growth is defined as originating from the mantle or oceanic crust. Recently it has been noted that Archaean TTGs form a more heterogeneous group than previously thought. Divisions by REE characteristics (high- and low-HREE endmembers), interpreted formation pressure and general characteristics (transitional, enriched) have been used in literature. Their tectonic position of formation is disputed but there is general consensus of origin through melting of mafic protolith.

Adakites have become somewhat controversial after the term came to be used for rocks that differ significantly from original geochemical definition and are found in non-subduction settings and for magmas generated through processes other than melting of basaltic protoliths.

Sanukitoids are igneous rocks that share many geochemical characteristics with high-Mg andesites in Japan (sanukites). Their high content of both compatible and incompatible elements requires input from both mantle and crust. Quartz diorites and quartz monzonites resemble sanukitoids and share many elements in petrogenesis.

Late Archaean crustal and hybrid granites are identified widely. They formed by interaction (metasomatism, mixing, mingling) of different Archaean magmas or sources (TTG, sanukitoid s.l., granite). Patiño-Douce (1999) has claimed that only peraluminous leucogranites represent pure crustal melts. High-K granites (also biotite granites) are widespread on all Archaean cratons. After TTGs they form the second most abundant group of granitoids. Leucogranites are often included in this group as they share many aspects of origin and petrology. S-type and A-type granites are rare and less constrained on Archaean terrains.

The Archaean granites, gneissic granites examined in this study make up around 40% of the rocks. Granites are abundant in many orogenic and post-orogenic setting and models for origin and typology of granites abound (e.g. Barbarin, 1990; Sylvester, 1994). Late Archaean, typically high-K, granites intrude the earlier TTG terrains and greenstone belts. These granitoids present typically the last Archaean geological events in every craton worldwide (Laurent et al., 2014). This event is diachronous from craton to craton between 3.0 and 2.5 Ga and is apparently followed by a 250 Ma global magmatic shutdown. Late Archaean granites appear to be related to final stabilisation of Archaean cratonic lithosphere which testifies to geodynamic changes of the period. Late Archaean granites cover a wide range of compositions (Sylvester, 1994; Champion and Sheraton, 1997; Frost et al., 1998; Moyen et al., 2003; Käpyaho et al., 2006).

Probably most of the late Archaean high-K plutons belong to biotite-bearing granites. These are biotite (sometimes also hornblende) bearing monzo- to syenogranites to granodiorites with high K/Na (>1), moderately fractionated REE patterns with negative Eu anomaly. Most often they are interpreted as partial melts of pre-existing TTGs (Champion and Sheraton, 1997; Frost et al., 1998; Hölttä et al., 2012). Less common types include metaluminous to peralkaline syenites and A-type, two-mica peraluminous (A/CNK>1.1) leucogranites. Literature also reports granitoids that cannot be ascribed to any predefined group (Mikkola et al., 2011; Laurent et al., 2014).

Notably, the majority of the rocks are peraluminous, reflecting the proportion of high-potassium granites and the leucogranites. Also, a large proportion of TTGs are peraluminous. The samples are grouped using the O'Connor's (1965) diagram based on feldspar ratios for rocks with more than 10% normative silica. Streckeisen classification is used to determine rock types.

Archaean cratons follow typical evolution of granitoids from TTGs to sanukitoids to leucogranites. They are presented in the same order bellow.

3.2. Tonalite-Trondhjemite-Granodiorite (TTG) association

TTG rocks are most common rock suite on almost all Archaean granitoid terranes and they formed the embryos of the continental crust. In recent decades TTGs have attained considerable attention and it has become increasingly accepted that there is significant variation among the TTGs (e.g. Barker et al., 1976; Rapp et al., 1991; Smithies et al., 2000; Condie, 2004; Martin, 2005; Moyen, 2011). The TTGs formed the first and a major step in formation of Earth's continental crust and unravelling the genesis of TTGs is the crucial step in understanding Earth's crustal evolution.

There is no formally accepted definition of TTGs but the consensus is that TTGs are silica-rich and sodic (3 wt.%< Na_2O<7wt.%) with low K_2O/Na_2O (<0.5). TTGs are relatively leucocratic ($FeO+MgO+MnO+TiO_2 \leq 5$ wt. %) with a Mg# average of 43 (Moyen and Martin, 2012).

Historically, TTG is most often referred to high-Al TTGs (low-HREE TTG) and researchers agree that these TTG magmas were generated by partial melting of amphibolites or eclogites at high pressures (Martin, 1987; Rapp et al., 1990). All models for TTG generation postulate partial melting (5-30 %) of broadly (meta)basaltic precursor (Lopez et al., 2006; Rollinson, 1996; Martin, 1986; Moyen and Stevens, 2006; Moyen, 2011).

TTGs are traditionally regarded as a single type for which a single tectonic scenario should exist. However, many regional studies have identified subtypes or different geochemical series of TTGs (e.g. Halla et al., 2009), summarised neatly in Moyen (2011). The differences are primarily in pressure-sensitive elements (Sr, HREE) and interpreted to reflect differences in pressure and hence depth of melting. The mineralogy of the source is an associated question, i.e. eclogite or amphibolites (Foley et al., 2002). In the early days of TTG research, they were divided into high-Al and low-Al subtypes (Barker and Arth, 1976); at present preferred division is by HREE content into high-HREE and low-HREE TTGs (Halla et al., 2009).

While it is recognised that post-Archaean TTGs are formed in subduction-related tectonic environment, the question of Archaean TTGs' tectonic setting is still debated (e.g. Condie, 2009, Martin et al., 2005; Willbold et al., 2009; Halla et al., 2009; Moyen, 2011). The literature on TTG formation is dominated by subduction model but increasing numbers of alternative views have emerged.

Alternatives to slab melting include melting of roots of thick oceanic plateaus above long-lived mantle plumes or lower crust in arc systems (e.g. Condie, 2003; Bédard, 2006). Na-rich granitoids can also be produced at the base of thickened, perhaps stacked crust in a non-subduction setting by delamination and melting of thickened mafic crust at a range of crustal levels (Smithies, 2000) or at the base of thickened (>40 km) continental crust.

Halla et al. (2009) advocated subduction origin for high-HREE TTGs (low-Al TTG) and melting of deep parts of oceanic plateaus for low-HREE TTGs. Alternative mechanism for

Middle Archaean TTGs is through delamination of dense, eclogitic parts of oceanic plateaus (Zegers and van Keken, 2001) and melting of gabbroic and amphibolitic lower crust producing TTG melts following mantle decompression melting and uplift. Recently, Smit et al. (2019) have discovered B isotope evidence for 3.8-2.8 Ga TTGs that support plateau origin over subduction. The B isotopes in this study indicate that seawater did not contribute to the mafic source of TTGs.

TTGs are still an important constituent of juvenile crust in continental-margin arcs and they have contributed to crustal growth through the geologic time (Condie, 2003).

3.3. Adakites

Adakites were originally defined by Defant and Drummond (1990) as suite of intermediate to felsic rocks in the range of andesite to dacite and rhyolite and corresponding plutonic equivalents. In this work definition of adakite is adopted from Martin et al. (2005). The defining characteristics of adakites are high Na_2O contents (>3wt. %) and Na_2O/K_2O values, silica >54wt%, high Mg# (>40), Ni (>24 ppm) and Cr (>36 ppm). Adakites have highly fractionated REE patterns with typically low HREE and Y (normalized La/Yb ≥10, Yb≤1.8 ppm and Y≤18 ppm). Amongst their defining characteristics are also high Sr (≥400 ppm) and Sr/Y values (>20). Martin et al. (2005) have divided adakites into high- and low-silica types (HSA and LSA, respectively) based on SiO_2 content (Table 1).

Table 1: Comparison of the main characteristics of adakites (Martin et al., 2005; Hastie, 2010).

	LSA	HSA	JTA
SiO_2	>56-60 wt.%	>60 wt.%	~68 wt.%
MgO	4-9 wt.%	0,5-4	<2 wt.%
$CaO+Na_2O$	>10 wt.%	<11	<10 wt.%
K_2O/Na_2O	up to 0,6	<0,5	<0,5
Sr	Sr>1100 ppm	<1100 ppm	mostly <400 ppm
Sr/Y	>40	>40	~20

Adakites and TTGs share a high degree of compositional overlap in major and trace elements that supports a widely held view that these rocks share many petrogenetic similarities (Martin, 2005). The genesis of adakites is explained by melting of subducting basalt slabs where the ensuing melts have interacted with mantle wedge peridotite during ascent through mantle wedge and incorporated Ni, Cr and MgO (Martin et al., 2005). Experimental results (Rapp et al., 2010) support the origin of adakites as slab melts that rise through mantle wedge where they assimilate mantle peridotite that increases their MgO, Ni and Cr content and decreases their SiO_2 content.

The subducted slab is metamorphosed into amphibolite or eclogite or garnet eclogite (Defant and Drummond, 1990; Rapp et al., 1991; Martin et al., 2005; Moyen, 2009). Controversy about adakites has arisen in early 2000s with the introduction of “continental” and even potassic “adakites” (Wang et al., 2005, Clemens and Xiao, 2003) with no links to subduction zones.

Moyen (2009) questioned the classification of these rocks as adakites on geochemical basis and emphasised the importance of all parts of adakite (or any other rock type) definition. Danyushevsky et al. (2008) found a range of adakitic compositions in inclusions in primitive adakite and argued against division by SiO_2 content and for magmatic suite that includes primitive high-Mg members and more evolved low-Mg members. Most of the adakites of this study are located in the Ilomantsi terrain.

3.4. Sanukitoid suite

Archaean sanukitoids, often called high-Mg diorites, were first described from Superior province by Shirey and Hanson (1984). In contrast to Archaean TTGs, they are characterised by lower silica, higher Mg#, Ni and Cr and large ion lithophile element (LILE) contents (Smithies et al., 2000). Stern et al. (1989) gave the first defined sanukitoids as plutonic rocks with SiO_2 55-60 wt.%, K_2O>1 wt.%, MgO>6 wt.%, Mg#>60, Ni>100 ppm, Cr> 200 ppm, Rb/Sr<0.1, Sr and Ba both over 500 ppm. Sanukitoids have highly fractionated REE pattern and only small Eu anomalies. Rocks of sanukitoid series have high compatible (K, Ba, Sr, P, LREE) and incompatible element (MgO, Ni, Cr) contents that does not depend on silica content.

Sanukitoids are currently considered a series of granitoid rocks (Lobach-Zhuchenko et al., 2005; Halla, 2005; Heilimo et al., 2010). Heilimo et al. (2010) proposed new geochemical definition for sanukitoid series: SiO_2=55–70 wt.%, Na_2O/ K_2O= 0.5–3, MgO= 1.5–9 wt.%, Mg#= 45–65, K_2O= 1.5–5.0 wt.%, Ba+ Sr>1400 ppm, and $(Gd/Er)_N$= 2–6. It gives wider compositional range than earlier definition and emphasizes the importance of LILE (Ba+Sr) and HREE contents. This definition of sanukitoids is adopted in this work.

Their genesis requires a source significantly more mafic than average Archaean basalt: a mantle source. Sanukitoids are among the first rock types that were derived from mantle and thus understanding their geotectonic position and petrogenesis carries important insights to crustal growth. Their sudden appearance at the end of Archaean also indicates a distinct tectonic process. Their formation has been explained by a two-stage process. In first stage fluids and possibly melts from subducting slab, which may include also sediments on the down going slab, induce metasomatism in the mantle wedge above the subducting slab. In the second stage, at the end or after the subduction the metasomatized mantle melts following a thermal event (Lobach-Zhuchenko et al., 2005; Halla, 2005).

The characteristics of sanukitoids are intermediate between the characteristics of high-Al TTGs and those of modern arc granitoids. They are typically metaluminous and moderately potassic and their features are interpreted as reflecting melting of mantle wedge peridotite that has been previously metasomatized by TTG melt from subducting slab. Heilimo et al. (2010) have argued that enrichment in Sr and Ba and their isotopic characteristics requires two events in the mantle: some enrichment results from basaltic slabs and sediments it carries. The other enrichment is related to upwelling of asthenosphere following slab break-off at the end of subduction (Halla et al., 2009; Heilimo et al., 2010).

3.5. Quartz diorites

The dataset provided by GTK includes a small number of intermediate rocks which differ from adakites and sanukitoids. At first approximation they are collectively referred to as quartz diorites. They have no apparent spatial relation to greenstone belts and in part have characteristics similar to TTGs. They are mostly calc-alkaline and metaluminous.

Mikkola et al. (2011) have interpreted quartz diorites from Suomussalmi complex of Kianta terrain as melts from metasomatized mantle or as felsic endmembers of a fractionation trend within a magma that fractionated an ultramafic phase rich in Mg, Cr and Ni.

They are characterised by lower (and variable) silica contents (53-66 % SiO_2) and wide range of compositions within a small number of rocks. In greater part they have higher transitional element and Mg# higher than TTGs. Quartz diorites include a subset that is characterised by peculiar, convex LREE patterns. These rocks are widely scattered and do not form larger groups but have greater concentration on the western half of the Karelian domain.

3.6. Quartz monzonites

Alkaline-rich rocks form a minor part of the current study as well as a minor part of Archaean geological record. There are many geotectonic models for the formation of alkaline-rich rocks. Typically proposed tectonic settings are often intracrustal but occasionally subduction and an extensional setting has been cited (Yang et al., 2005). They have some characteristics in common with A-type granites, as high total alkali content and some HFSE but mostly lower Nb and Y as well as FeO/MgO (this study) and more fractionated REE patterns. On tectonic diagrams of Pierce (1984) they plot into VAG field and on diagrams of Whalen et al. (1987) they straddle the boundary between the A-type and other granites. Further they differ from A-type granites by high Ba (>1000 ppm) and locally high Sr. Quartz monzonites also have higher Al_2O_3, MgO and Mg#.

Their REE patterns resemble those of leucogranites with enriched LREE, deep negative EU anomalies and, in cases, flat HREE end. They differ from leucogranites by lower silica and higher LILE. Quartz monzonites include two samples with peculiar, convex LREE patterns. This kind of pattern has been explained by Hölttä (1987) as a result of alteration by CO_2-rich fluids.

3.7. Transitional TTGs

Rocks similar to sodic TTGs, but potassic in composition, sometimes referred to as "enriched TTGs" or "transitional TTGs" abound in Archaean terrains. Compositionally they overlap with the felsic members of low-HREE TTGs and they follow similar fractionation trends. Transitional TTGs are slightly richer in SiO_2 than low-HREE TTGs.

Transitional TTGs are variably metaluminous to peraluminous with majority being peraluminous. K_2O concentrations are higher and Na_2O lower than in high-Al TTG. Eu anomalies vary from positive to negative. LREE enrichment is highly variable, on average higher than in high-Al low-HREE TTGs but include samples with low total REE content that display a positive Eu anomaly.

Transitional TTGs have mostly higher LILE content than high-Al low-HREE TTGs but lower Sr although there is significant overlap.

3.8. High-K granite

It is frequently assumed that the granites are products of intracrustal reworking of older TTGs though it has become increasingly obvious that there is considerable variation among the Archaean granites and in part granites are juvenile (Lopez et al., 2006). Clearly different granites, including high-Mg varieties, granites with distinct A-type affinities and syenites are also important in the Archaean (Champion et al., 1996; Jayananda et al., 2006). Many granites in Archaean terrains appear to be related to crustal thickening and anatexis during late stages of crustal collisions. In Karelia crustal thickening due to collisional tectonics is dated to 2.7 Ga. Despite being of significant volume and volumetrically dominant on some Archaean terrains the granites have been given relatively little attention in recent decades (Champion, 1997; Frost et al., 2005). Most Archaean granites were emplaced towards the end of Archaean and are temporally associated with craton stabilization (Kusky, 1999; Jayananda et al., 2006; Joshi et al., 2017). Sylvester (1994) assigned the geochemical diversity of high-K granites to differences in depth of melting. The unifying features of this group are high SiO_2 and LILE as well as low MgO concentration.

3.9. A-type granite

A-type granite was introduced in 1979 by Loiselle and Wones in an abstract. Originally A-type was defined as high K and K/Na, Fe/(Fe+Mg) and high incompatible element (Zr, REE, but Eu, Nb, and Ta) concentrations. Additionally, A-type granites have high halogen, especially F concentrations. Whalen et al. (1987) have devised a discrimination diagram for A-type granites. He has extended the composition of A-type granites to include calc-alkaline and peraluminous and metaluminous samples.

Eby (1990, 1992) has divided A-type granites into two geochemical subtypes by Y/Nb. A1 A-type granites have Y/Nb<1.2 and A2 A-type granites have Y/Nb>1.2. A-type granites of A1 subgroup occur on oceanic islands and continental extension zones. They are suggested to form from oceanic basalt source. The A2 subgroup of A-type granites has been suggested to arise from a number of different sources and by various mechanisms: island arc basalt, continental margin basalt or continental source than an earlier melt has been extracted from. Part of A-type granites has a rapakivi texture. Rapakivi-granites are iron-rich, mostly metaluminous and calc-alkaline.

A-type granites have low concentration of mafic mineral-compatible elements (Co, Sc, Cr, Ni) and elements compatible in feldspars (Ba, Sr, Eu). Martin (2006, 2012) has proposed a "fenitization" theory for the origin of A-type granites. According to the proposal, A-type granites form by melting of lower crust. Before melting lower crust is metasomatized by fluids in an extensional environment that follows a delamination event. Delamination lowers mantle pressure leading to decompression melting and release fluids rich in H_2O and CO_2. Fluids contain incompatible elements, Fe and alkalies, Si and heat. Fluids refertilize formerly sterile, unfertile lower crust and make thermally and chemically anomalous. Continuous rise of fluids causes alkali metasomatism and carbonisation. Melting of newly fertilized lower crust can give rise to anorogenic A-type magmas that can be oversaturated or undersaturated in silica. Silica saturation depends on the H_2O and CO_2 as only H_2O can transport silica.

A-type granites differ from leucogranites by their higher FeO, HFSE and mostly higher LREE content.

3.10. S-type granite

S-type granites were first proposed by Chappell and White (1974), along with I-type granite. They described granites in Australian Lachland Fold Belt that form two series with distinct petrographic and field characteristics and chemical composition.

S-type granites (and corresponding volcanic rocks) are defined as being over-saturated in Al and are peraluminous with A/CNK>1.1. This is a result of their lower Na and Ca that are presumed to be lost during a weathering of their source rocks. To the contrary they are rich in K which is sequestered in clays during weathering. S-type granites contain biotite and another alumina-rich mineral, often cordierite, muscovite or garnet and are postulated to form through melting of material that has undergone at least one weathering cycle, i.e. sediments.

Although division of granites into I- and S-types is an oversimplification, as admitted by the original authors (Chappell, 1999), the division has been widely accepted as it serves to distinguish rocks that have a significant weathered component in their genesis. It is still used (in 2020) and continues to be a subject of fruitful research and providing new insights into evolution of continental crust (Zhu et al., 2020).

4. Descriptions of rock suites: crustal growth

4.1. Karelian TTGs: this study

Karelian TTGs are grouped initially into high-Al and low-Al (low-HREE and high-HREE TTG) groups based on their major and trace element characteristics. The primary division follows the classification of Barker and Arth (1976), Martin et al. (2005) and Halla et al. (2009). There is little difference in major element compositions between different TTG subgroups. Then, groups deviating from initial division are identified. High-Al (low-HREE) TTGs form the majority of TTGs. High-HREE TTGs are very subordinate.

High-Al (low-HREE) TTGs as widely reported in literature are the largest group (157 samples).

High-Al (low-HREE) TTGs are further divided by Eu anomaly following Halla et al. (2009) example (80 samples). High-Al TTGs with positive Eu anomaly tend to have lower total REE concentrations than low-HREE TTGs (and other TTGs in the dataset).

There is a group of low-HREE TTGs with low LREE and thence unfractionated REE pattern (High-Al low HREE unfrac: 8 samples). They have HREE in the low-HREE TTG range but LREE are low. A subset of low-HREE TTGs is characterised by low LREE (La 8-19 ppm) and LILE contents and moderately fractionated REE pattern with low chondrite-normalized La/Yb (5-9) and small negative Eu anomaly (Eu/Eu* 0.67-1). It overlaps in composition with low-HREE TTGs in major and trace element content, though on a given silica content it has low Na_2O and high CaO concentration. It has LREE content similar to low-HREE TTG subgroup with positive Eu anomaly but HREE higher and lower than high-HREE TTGs.

There is a small group of mafic low-Al (low-HREE) TTGs with low LREE similar to previous group. It is similar to adakites and might represent altered adakites. REE pattern is unfractionated (chondrite-normalized La/Yb 5-<10). TTG suite here include a subset of low-HREE TTGs (and low Y<13 ppm) with low content of LREE and Sr resulting in a unfractionated (normalized La/Yb<10) patterns and low Sr/Y. Their transitional element contents (MgO, Ni, Cr) are well in adakite compositional level. If they have lost LREE and Sr post-emplacement they could be considered HSA. They are recorded as gneisses and have all the features of HSA but low LREE. There are 5 samples.

There are two samples with unfractionated REE pattern that is very irregular that might be heavily altered samples. They are so outlying that are discarded.

The low-Al TTGs (high-HREE TTGs) have less fractionated REE patterns with flat HREE ends that is taken to indicate the absence of garnet in the residue of melting. The negative Eu anomaly indicates presence of plagioclase in residue or plagioclase fractionation, as in low-HREE group.

Low-Al TTGs have low abundances of Rb and Sr, REE patterns characterised by light LREE enrichment, negative europium anomalies and flat heavy REE (Fig. 2) (Barker et al., 1976). They are widely scattered, mostly on southern Kianta terrain, and appear as lone samples. They can be divided into two groups by their HFSE content (high-HFSE and low-HFSE). The high-HFSE group is more enriched in HFSE, MREE and is more metaluminous.

They have narrow silica range (66.4-69.6 wt.% SiO_2). They are widely scattered as individual samples. In the classification of Frost et al. (2001) they are calcic, magnesian and metaluminous, on discrimination diagram of Martin (1986) they plot on the overlapping field of adakite/TTG and ordinary arc lavas (Fig. 3). Low LREE may be a result of fractionation of monazite? Their transitional element content equals the content in both high- and low-HREE TTGs suggesting that their formation was similar in general terms and did not include interaction with mantle wedge. REE pattern is shown in Fig. 2.

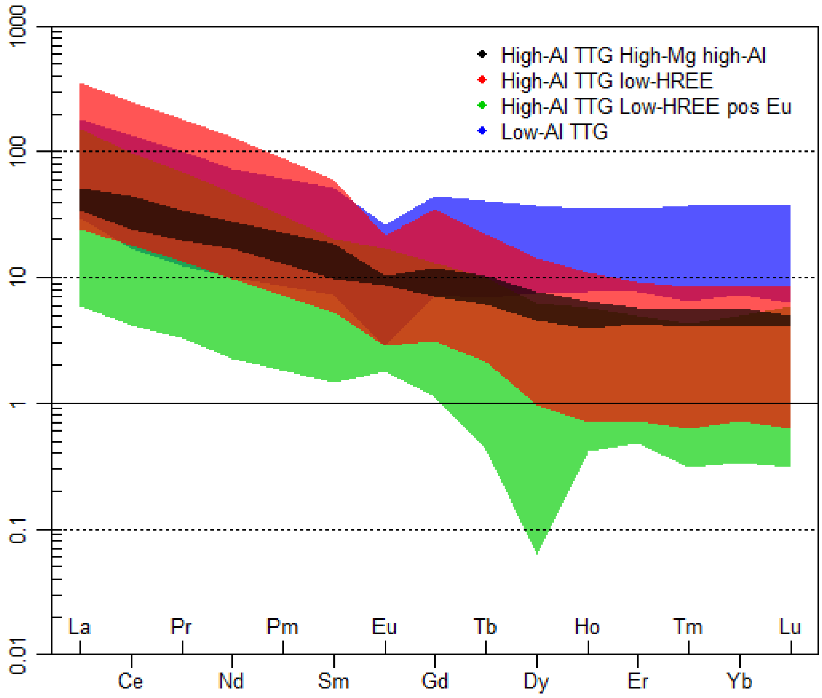

Fig.2. REE patterns of Karelian TTGs of this study demonstrate their division into high and low-HREE groups but also show divergent groups; the mafic TTGs with minor REE fractionation resulting from low-LREE with the same HREE content as low-HREE TTGs. Normalization values after Boynton (1984).

4.2. Formation of TTGs

High-Al TTGs (low-HREE TTGs) correspond to Archaean TTGs widely reported in literature. They have steeply fractionated REE patterns with significant depletion of HREE which is taken to reflect significant garnet in the residue of melting retaining the HREE (Martin, 1987; Moyen and Stevens, 2006; Halla et al., 2009; Moyen, 2011; Mikkola et al., 2011; Laurent et al., 2014).

Other processes can lead to high La/Yb: melting of high La/Yb source, fractional crystallization or interaction of felsic melts with mantle. Low mantle-compatible element (Mg, Ni, Cr) concentrations preclude mantle interaction in these samples. Eu anomalies are explained by preferential incorporation of Eu (as Eu^{2+}) in plagioclase. The negative Eu anomalies indicate plagioclase in the source or plagioclase fractionation. The positive Eu anomaly suggests accumulation of plagioclase.

Moyen and Stevens (2006) gave experimental constraints on TTG petrogenesis. They suggested that TTGs form a continuum of compositions, forming at a range of pressures, as indicated by pressure-sensitive elements, from low to high pressures. Low-pressure TTGs (ca. 10 kbar) are have low Sr (<400 ppm) are relatively undepleted in HREE and have negative Eu anomaly. REE systematics are coupled to Sr and Y. Moyen (2009) has related high Sr/Y TTGs (low-HREE TTG) to deep melting of basaltic source and low Sr/Y TTGs to shallow melting of similar, basaltic source. He has further developed the theory and envisaged generation of TTG in similar tectonic scenario as modern S-type granites but dominated by mafic, basaltic composition producing TTG melts at a range of pressures.

Halla et al. (2009) have suggested a possible tectonic setting where both high and low HREE groups may exist: an incipient subduction under an oceanic plateau where the mantle is unusually hot. Based on the example from Aruba, Caribbean (White et al., 1999), Halla et al. proposed an origin for high-HREE and low-HREE TTGs with contributions from subducting slab, mantle wedge and overlying oceanic plateau. In this scenario low-HREE TTGs are formed in deep root parts of oceanic plateau at high pressures where garnet is stable and sequesters HREE. High-HREE TTGs were formed at lower pressures and shallower depths by interactions of subducting slab and mantle wedge. The dataset used in this study includes 13 samples of high-HREE TTGs that overlap in Ni, Cr and MgO contents with equally low content in low-HREE TTGs. As HREE and Y are garnet-controlled elements melting of amphibolites at different pressures provides an explanation for HREE variation in TTGs.

The group with positive Eu anomalies has on average lower total REE and Y. There is also tendency for higher silica and Al_2O_3 that makes majority of them peraluminous but there significant overlap. Terekhov and Shcherbakova (2006) proposed that the origin of positive Eu anomalies is in deep-seated origin of fluids that participated in their formation. The proposed zone is the boundary in the crust between the brittle and ductile behaviour. The periods of horizontal extension were characterised by inflow of reduced fluids into the fractures and partial melting of host rocks. This suggestion is partially supported by lower Rb and higher Sr and Ba and consequently higher Sr/Rb in TTGs with positive Eu anomalies.

Episodic generation of TTG crust has been known for long time (e.g. Condie, 1998; Stein and Hofmann, 1994; Condie et al., 2018). Martin et al. (2014) have integrated the episodic generation of TTGs with subduction of oceanic plateaus in analogy with a modern process, subduction of Carnegie ridge beneath South American continent. This process today produces adakites, often considered modern analogues of Archaean TTGs (Martin et al., 2000), although the analogy is disputed (Smithies, 2000; Condie, 2005). In the light of dataset used in this study that includes low Ni and Cr TTGs and adakites that origin is not probable.

Halla (2020) proposed another explanation for coupled formation of high- and low-HREE TTGs: partial melting of granulite facies amphibolite can produce low-HREE. As low-TTG melts leave lower levels of amphibolite they leave garnet retaining the HREE behind. As the melts arrive to higher levels and lower pressures they induce water-fluxed melting in the amphibolite-facies yielding high-HREE TTGs because garnet is not stable in water-fluxed melting conditions. At equal mantle-compatible element concentration levels this explanation is more credible. It does not explain why the high-HREE TTGs are so subordinate.

Recently Smit et al. (2020) has found that boron isotopes of TTGs do not suggest interaction of parental material of TTGs with seawater arguing against TTG origin though subduction and in favour of an intraplate (crustal thickening, volcanic resurfacing, delamination or stagnant lid), as argued e.g. by Bédard (2006) or Hoffmann et al. (2011).

The mafic TTG group has notably higher transitional element concentrations and higher Mg# (>45) which indicate (more) mantle contribution in their formation if they are not altered adakites as their composition, apart from LREE, suggests (Fig. 2).

4.3. Karelian adakites: this study

Adakites in this dataset include samples with silica range from 54.8-70.7 wt.% SiO_2. This range covers both the low-silica adakites (LSA), high-silica adakites (HSA) and Jamaica-type adakites (JTA) but the adakites here have mostly the characteristics of HSA (Fig. 3.). They lack the extreme enrichment in Sr (400-1215 ppm), with only one sample reaching typical LSA values (>1100 ppm Sr) and it has lower MgO content and $CaO+Na_2O<11$ wt.%, typical of HSA. The dataset includes only two samples of the low-silica variety and these are only marginally low-silica (58.4-59% SiO_2) and their other characteristics point to HSA. The adakites have high $(La/Yb)_{Ch}$, low Y and Yb concentrations and negative Nb-TA-Ti anomalies on normal primitive mantle- normalized multielement diagrams (Fig. 4.).

Adakites form three compositional groups: high-silica adakites (HSA), low-silica adakites (LSA) and Jamaica-type adakites (JTA). They are widely scattered, mostly as lone samples but on Ilomantsi terrain JTA appear to form a short belt. Adakites are most often explained as products of slab melting in subduction environment (Defant and Drummond, 1990; Martin et al., 2005; Defant and Kepezhinskas, 2001; Hastie et al., 2010). As in the case of TTGs low HREE and highly fractionated REE pattern indicate garnet in residue. Negative Nb-Ta-Ti anomalies (Fig. 4.) suggest residual rutile and/or amphibole in residue (Moyen, 2009). Alternatively they can result from enrichment of LILE (K, Ba, Sr) and LREE (La, Ce) relative to Nb and Ta as a result of slab melt mixing with slab-derived aqueous fluids.

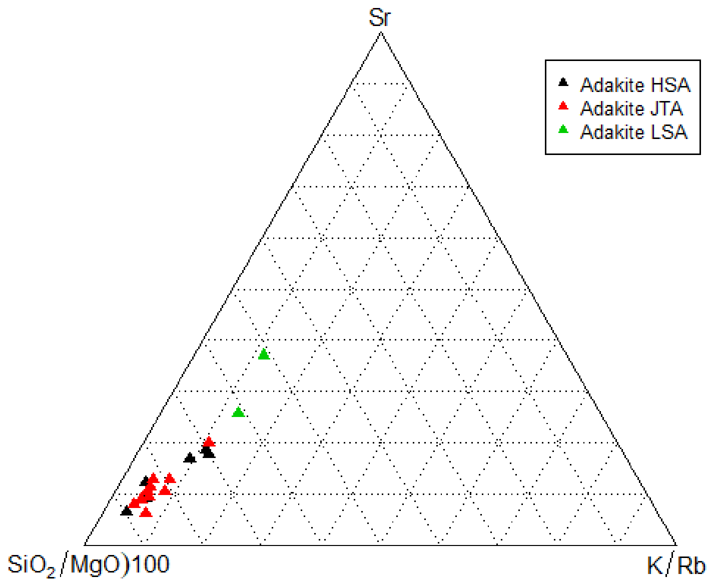

Fig. 3. Adakites in this study plot to HSA corner of the diagram. Martin et al., 2005.

Major elements mostly correlate with silica but Na and K show scatter as do most LILE (Fig. 4.

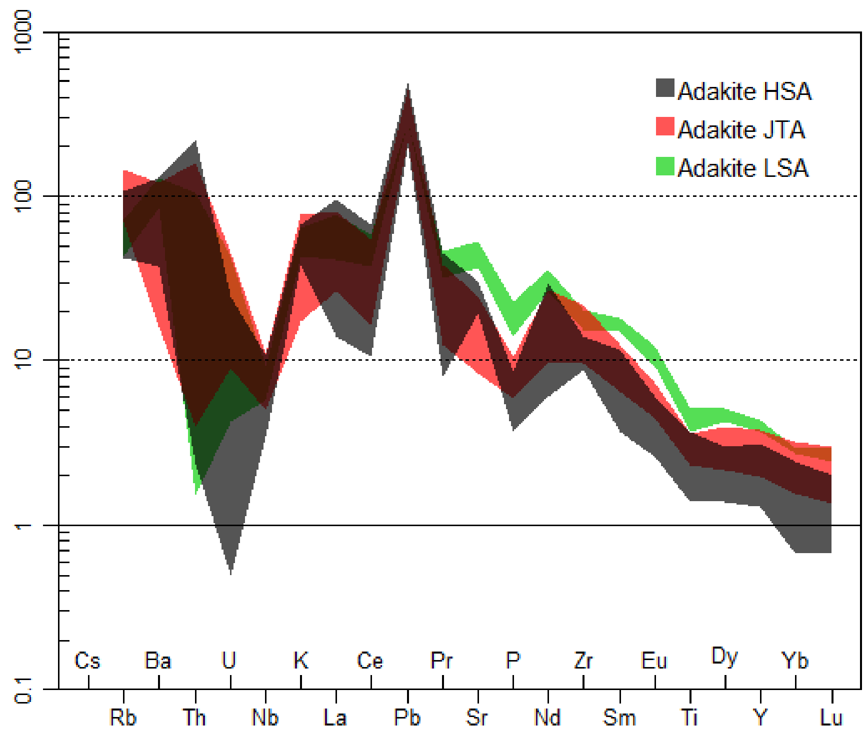

Fig. 4. Primitive mantle normalized element variation diagram for adakites. Normalization values after Sun and McDonough, 1989. Positive Sr anomaly for LSA is visible as are negative Nb-Ta-Ti anomalies in all samples.

4.4. Formation of adakites

Adakites are pre- to syntectonic granitoids and volcanics found in modern subduction settings. There are some suggestions of modern adakites being possible analogues to Archaean sanukitoids, at least at lower silica range (Martin et al., 2005). However, sanukitoids have higher K_2O content and distinctly higher Sr and Ba. Furthermore, sanukitoids are 10-100 Ma younger than TTG magmatism in a given region in Karelia (Käpyaho et al., 2006) and thus post tectonic.

Sanukitoids in Karelia are postulated (Heilimo et al., 2010, 2013), as well as quartz diorites (Mikkola et al., 2011) to form from magmas originally from subduction related metasomatized mantle. The subduction that metasomatized the mantle region may have produced also the adakites.

Higher MgO, Ni and Cr in adakites compared with TTGs are interpreted as result of melts from subducting slab rising through a “slice” of mantle wedge peridotite. During the ascent slab melts assimilate peridotite and increase their MgO, Cr and Ni as well as Mg# and decrease silica contents (Defant and Drummond, 1990). Such adakites form at present in South America (Martin et al., 2014). There is also evidence for melt-mantle interactions in form of adakitic veins in mantle xenoliths (Kepezhinskas et al., 1995). Experimental evidence and geochemical modelling support this possibility (Rapp et al., 2010; Prouteau et al., 2001; Moyen, 2009).

Interestingly the largest group of adakites present here is of the Jamaica-type. Archaean oceanic plates are postulated to have been thicker (e.g. Foley et al., 2003) and perhaps more akin to oceanic plateaus. Using this analogy there are two possibilities to form JTA: relatively shallow partial melting of subducting slab aided by higher mantle temperatures without interaction with mantle wedge. Second possibility is melting of plateau-like oceanic slab and interaction of melts with a thin or discontinuous (boudinaged) mantle wedge. Hastie et al. (2010) suggested that Jamaican adakites form as melts from subducting Caribbean oceanic plateau.

4.5. Karelian sanukitoid suite: this study

Sanukitoid suite displays significant uniformity in its composition. They follow same trends on Harker diagrams. Martin et al. (2005; 2010) have attempted division of sanukitoid suite into high-Ti and low-Ti sanukitoids but the division has not gained wide acceptance.

Sanukitoids follow the calc-alkaline trend. They have remarkably uniform REE patterns with small negative Eu anomalies. Even the most primitive members of sanukitoid suite are remarkably enriched in LILE. This point implies that enrichment in LILE is a primary characteristic of sanukitoids and not a consequence of fractional crystallization.

Most sanukitoid studies explained the petrogenesis of sanukitoids and their transitional characteristics as results from interaction (effectively hybridisation) of mantle peridotite with felsic, TTG melt (Heilimo et al., 2010). Experimental studies support this conclusion (Rapp et al., 2010). The source of sanukitoids has been suggested from the beginning of sanukitoid studies to be enriched mantle wedge (Shirey and Hanson, 1984; Rapp et al., 2010).

4.6. Formation of sanukitoids

Formation of sanukitoids has been explained most often by melting of previously metasomatized mantle. Metasomatism of the mantle source of sanukitoids is achieved by addition of fluids and/or melts in subduction environment. This explanation is adopted by Heilimo et al. (2010, 2013) for sanukitoids in Karelia. They have postulated two metasomatic events: in the first stage mantle is metasomatized by melts and fluids from subducting slab.

The second metasomatic event follows after the end subduction and continent-continent collision as the subducting slab breaks-off. Slab break-off and sinking into mantle induces upwelling in of the asthenospheric mantle that carries alkaline fluids that further enriched and metasomatized the mantle wedge. Partial melting of the twice metasomatized mantle produced the sanukitoid magmas with adequate element and isotopic composition.

Heilimo et al. (2010) claimed that fractionation in Karelian sanukitoids was rather weak and unclear at the series level although locally significant. They ascribe the difference between sanukitoid intrusions to source heterogeneity and differences in depth of melting.

Support for this interpretation come from Silurian granites of Caledonian orogeny in Scotland. Caledonian high Sr-Ba granites are similar in occurrence and geochemical characteristics to Karelian sanukitoids (Atherton and Ghani, 2002; Fowler and Rollinson, 2012; Bruand et al., 2014). These were formed in narrow time interval as were sanukitoids and form narrow, linear belt as expected from a single slab breakoff event. Karelian sanukitoids form narrow belts along the margins of Archaean domains of Karelian province.

4.7. Quartz diorites: this study

Quartz diorites form minor but diverse and widespread groups of rocks on Archaean cratons. These rocks share similar characteristics with sanukitoids but differ in compositional detail. They are often grouped together in broad grouping as sanukitoids s.l. (Laurent et al., 2014). Both have moderately fractionated REE patterns and higher transitional element content than TTGs. Quartz diorites can be differentiated from sanukitoids by higher Ba, K_2O and LREE although there is overlap between the groups. Quartz diorites in this study range from 51-63.5 wt.% SiO_2, Mg# 39-63. They include high-HREE and low-HREE subgroups in an analogous manner to TTGs. There are two samples with positive Eu anomalies and low total REE content. Both high-HREE and low-HREE subgroups include a subset with unfractionated REE patterns that have peculiar, almost flat to convex LREE (La-Sm) pattern (Figs. 5. and 6.).

Hölttä (1997) offered possible explanations for similar pattern in Varpaisjärvi enderbites: a LREE rich phase selectively removes LREE or a CO_2-rich fluid preferably fractionates LREE. Such fluid metasomatism offers an explanation for the rocks studied here. Low LREE contents correlate with low Th and K_2O contents suggesting too granulite facies metamorphism might have caused the depletion of these elements in samples with peculiar LREE patterns. The rocks in this dataset are mostly metaluminous to the contrary to enderbites in Hölttä.

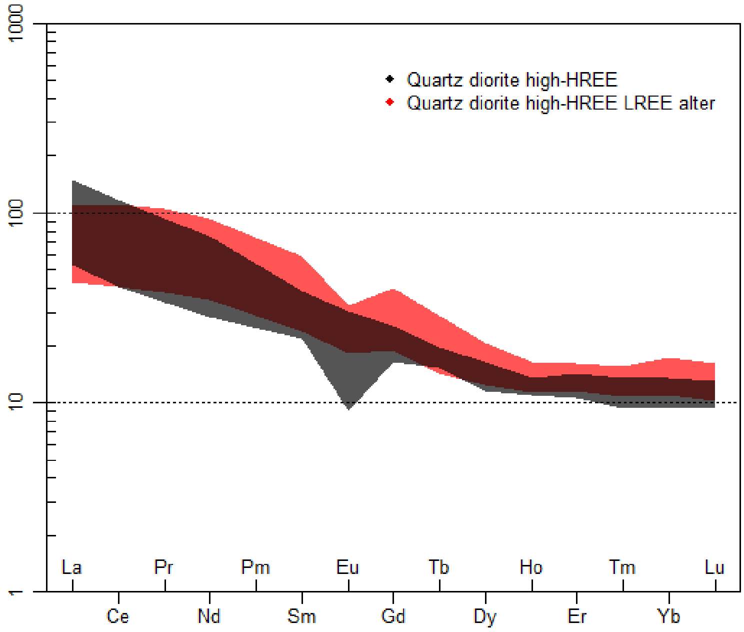

Fig. 5. High HREE quartz diorites. Black: apparently unaltered; red: high-HREE quartz diorites with peculiar, convex LREE patterns.

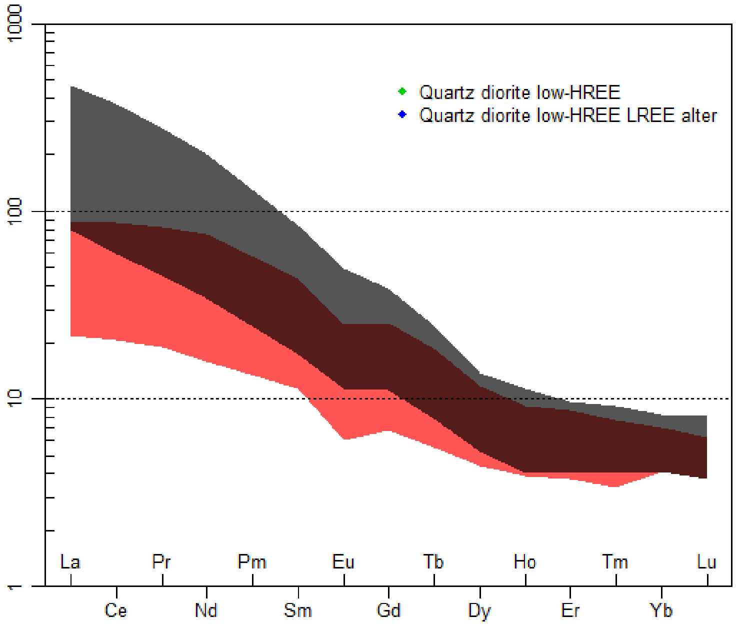

Fig. 6. REE patterns of low-HREE quartz diorites. Low-HREE quartz diorites have also two subsets with contrasting REE patterns.

Apparently unaltered high-HREE samples have moderately fractionated REE patterns (chondrite-normalized) La/Yb 11.1-15.25 whereas altered samples have La/Yb<10. Both altered and unaltered subsets have negative or (one sample) negligible Eu anomalies (0.48-1.01). Apparently unaltered low-HREE samples have fractionated REE patterns (ch. normalized) La/Yb 10-70, though mostly 20-30 whereas altered samples have La/Yb<10 the difference coming from lower LREE enrichment (or resulting from loss after formation) (Fig. 6.).

4.8. Formation of quartz diorites

Both sanukitoids and quartz diorites share similar petrogenesis as melts from variously metasomatized mantle and an incompatible element rich component (Mikkola et al., 2011; Laurent et al., 2014). Mikkola et al. (2011) have studied quartz diorites in Suomussalmi and found them similar to sanukitoids in many aspects. The differences between sanukitoids and quartz diorites may be caused by different mantle metasomatism or higher degree of partial melting in the mantle. The difficulty is explaining the lower Cr, Ni and MgO in quartz diorites compared to sanukitoids. Another possibility is that quartz diorites represent a felsic endmembers of a magma that fractionated ultramafic phase rich in Ni, Cr and MgO. Here, the difficulty is finding a suitable rock, more mafic than quartz diorite. The differences in HREE content may have resulted from different depth of melting. The division by REE characteristics reveals striking similarity to similar division in TTGs.

4.9. Quartz monzonites: alkaline-rich intrusions, this study

Quartz monzonites subtly differ from similar sanukitoids and quartz diorites by their higher K_2O and lower MgO, Ni, Cr, FeO and mostly TiO_2. Partial melting of quartz diorites and sanukitoids could explain overall similarity, the shape of REE pattern, lower transitional element contents and Mg#. P_2O_5 poses a problem in a fraction of cases as its concentration is same in many samples across the rock types. Quartz monzonites have similarities with A-type granites, as high alkalies and FeO but differ by lower concentration of Nb and Y and higher of Ba and locally of Sr. They are also more fractionated in REE. They are shoshonitic with one sample high calk-alkaline with silica range from 63-67 wt. % SiO_2. MgO is mostly low (max. 2.25 wt.%), K_2O and Na_2O are high (4.25-6.39 and 2.44-5.53 wt.%, respectively). Ba and Sr are high, up to 3474 and 1897 ppm, respectively. Their REE patterns are variable but highly fractionated and include two samples with similar, convex LREE pattern as in quartz diorites.

4.10. Formation of quartz monzonites

Petrogenetic models for alkaline-rich rocks invoke melting of lower or middle crust with possible contribution from mantle and possible assimilation of upper crust or partial melting of mantle with assimilation of crustal material and fractionation of magma (Smithies and Champion, 1999). They clearly differ from TTGs that are generally accepted as partial melts of basaltic precursors. Leucogranites are clearly more silica-rich and in most cases interpreted as melts of TTGs. Heilimo et al. (2017) have speculated on the possibility that quartz monzonites are partial melts of sanukitoids, quartz diorites or amphibolites. Amphibolites are an unlikely candidate as they have relatively flat REE pattern in Karelia (Heilimo et al.,

2017) and formation of fractionated REE patterns would require melting in the presence of significant garnet. Moreover high Ba and K_2O cannot be produced by melting of amphibolites. Partial melting of sanukitoids or quartz diorites could explain most of their characteristics though in most cases P content is evidently problematic. Their isotopic compositions point towards complex mixture of sources including mantle, continental and oceanic crust and sediments (Heilimo et al., 2017).

5. Descriptions of rock suites: Crustal Progeny and hybrid granites

Late Archaean crustal and hybrid granites are identified widely. They formed by interaction (metasomatism, mixing, mingling) of different Archaean magmas or sources (TTG, sanukitoid s.l., granite). Patiño-Douce (1999) has claimed that only peraluminous leucogranites represent pure crustal melts. High-K granites (also biotite granites) are widespread on all Archaean cratons. After TTGs they form the second most abundant group of granitoids. Leucogranites are often included in this group as they share many aspects of origin and petrology. S-type and A-type granites are rare and less constrained on Archaean terrains.

5.1. Biotite granite (Archaean high-K granite), this study

Typical high-K granites are high silica (68-75 wt.% SiO_2), moderately peraluminous (A/CNK≥1.0-1.1). They have low content of ferromagnesian elements ($MgO+FeO+MnO+TiO_2 \leq 4$ wt.%) and are clearly potassic ($K_2O>4$ wt.%) with high K_2O/Na_2O (>1). Their trace element diagrams are similar to those of TTGs but have higher content of incompatible elements (Rb, Th). Transitional element and HFSE are low. Their REE patterns are moderately fractionated and have significant negative Eu anomalies but with higher Y-HREE than TTGs. The high-K granites are interpreted as partial melts of older crust, i.e. TTGs (Sylvester, 1994; Moyen et al., 2003; Moyen, 2011; Mikkola et al., 2012). In this dataset high-K granites form such a group but there are also some deviating samples:

A subset is characterised by positive Eu anomalies. These have highly fractionated REE patterns (all but one) with variable but high positive Eu anomalies. They are moderately peraluminous (A/CNK 1.01-1.11), depleted Y and Yb, variable Sr (161-854 ppm) and high Sr/Y in the extreme case reaching over 1000. It is limited to high silica content (>73 wt%). These can be Eu positive counterpart to Eu negative samples, possibly plagioclase cumulates. Negative Eu anomaly is a typical characteristic of granites and is taken to reflect plagioclase effect.

In their major characteristics they are typical late Archaean high-K granites as reported widely in literature. They are calc alkaline and in the extreme cases reach shoshonitic values. On the source diagram of Laurent et al. (2014) they plot along the tonalite-metasediment boundary with three samples in high-K mafic source rock source field. This is in accordance with the often cited origin. Shoshonitic rocks are associated with late stage of subduction in post-Archaean systems (Morrison, 1980).

5.2. Formation of high-K granites

High-K granites are interpreted as crustal melts of pre-existing TTGs (with or without sedimentary component). In Karelia they are interpreted to have formed in response to crustal thickening at 2.7 Ga (Mikkola et al., 2011). The heat for crustal melting may have come from underplating or intrusions of mafic magma, i.e. sanukitoids and quartz diorites that provided K-rich aqueous fluids (Patiño-Douce, 1990; Castro, 2020).

Debate has arisen from claims that most TTGs are too sodic to produce large amounts of potassic granites (Watkins et al., 2007). This has led the authors to suggestions of highly metasomatized, high-K mantle source. However, lack of affinity to sanukitoids (s.l.) and absence of intermediate and mafic phases are at odds with this proposal. Further, experimental partial melts of TTGs have K_2O content within range of TTGs (Patiño-Douce et Beard, 1995). Their compositional variation is attributed to source heterogeneity and various magmatic processes. There are, however, new proposal relating sanukitoid formation with high-K granite formation (Käpyaho et al., 2006, and references therein). The mantle upwelling and intrusion of sanukitoid magmas may have heated the crust sufficiently to generate crustal anatexis and the formation of various granites. The same explanation has been suggested before for Karelian potassic granites (Kovalenko et al., 2005).

The group with positive Eu anomalies may have similar origin in deep crust as the low-HREE TTG group or it may be inherited feature from the source. Condie (1986) has offered an explanation for similar granites and quartz monzonites in India. The Indian granites were interpreted as melts of older tonalite and complement to granites with negative Eu anomalies higher in the crust. The proposal suggests that in response to continent collision upper mantle devolatilize. Devolatilization of upper mantle drives water from lower crust. The fluids rich in CO_2 and water move upward and carry K, Rb, Ba and LREE. Partial melting of tonalites is possible in water-rich environment. Magmas rise and undergo fractional crystallization. Residual liquids form eventually form high level granites that are depleted in Sr and Eu as feldspars are removed. Condie (1986) suggested a continental margin arc setting for their formation.

Additional source of K and water can be melting of sediments (graywackes and metapelites). Their inherited zircons serve to further strengthen the crustal origin of these granites. The samples with shoshonitic affinity may represent highly fractionated samples. On the source diagram of Laurent et al. (2014) high-K granites plot along the tonalitic source-metasedimentary source boundary with s small minority into high-K mafic source field in accordance with often assumed origin from melting of older TTGs with addition of sediments.

High-K granites correspond to low-HREE monzogranites of Joshi et al. (2017). They have crustal signature but as crust is more geochemically variable more variable granites can form. Here granites diverse but form compositional continuum (all are granites s.s.). Progressive melting of the older crust explains the enrichment in alkalies and incompatible elements. Negative Eu anomalies are universally attributed to fractionation of plagioclase. Low HREE content indicates garnet signature in these rocks at some point of their evolution. It may be inherited from low HREE source, as suggested by source diagram, as the majority of TTGs are of low-HREE group (Fig. 7).

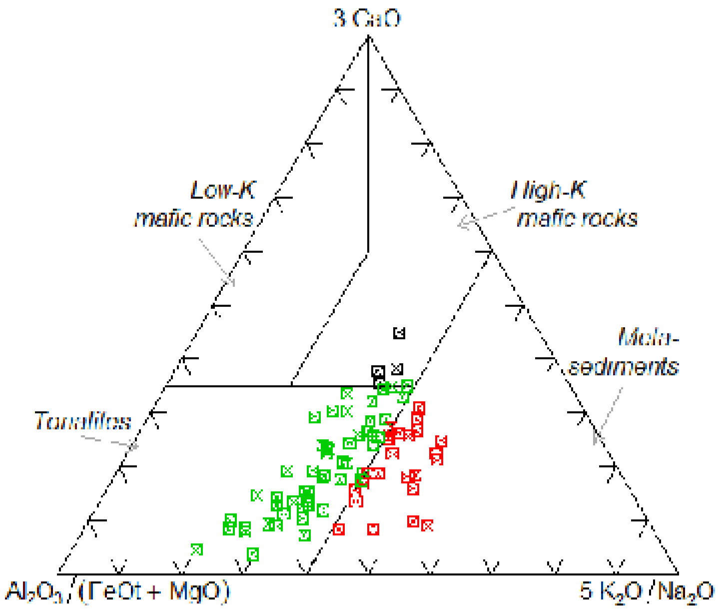

Fig. 7. Source diagram of the high-K granites of this study shows granites plotting along tonalite-metasedimentary boundary reflecting their origin as melts from older granitoids and sediments as universally postulated on all reports on Archaean high-K granites.

5.3. Karelian leucogranites: this study

A subset with variously fractionated LREE patterns, unfractionated, parallel HREE with deep negative Eu anomalies, although with a range of HREE and Y (Yb 0.9-3 ppm and 8.46-3.01 ppm, respectively). Further they have low transitional element and total ferromagnesian element content ($MgO+FeO+MnO+TiO_2<1.7$), MgO 0.02-0.1 wt.%), characteristic of leucogranites. They are variably marginally metaluminous (A/CNK 0.97-0.99; two samples) to peraluminous (A/CNK 1.01-1.08). Mikkola et al. (2012) interpreted similar granites as partial melts of both TTGs (s.s.) and transitional TTGs (TTG s.l.). They include two compositional groups: high-K granites (granite s.s.) and trondhjemite-granodiorite (transitional TTG on first assumption). The high-K granites in this group correspond to low-Eu monzogranites of Joshi et al. (2017). The grading of the series is observable in their highly variable CaO content (0.17-1.62 wt.%). Included in this subset are two samples with nearly strait, linear REE patterns with negligible negative Eu anomalies (0.97 and 0.98) that do not necessarily belong here.

5.4. Formation of leucogranites

Mikkola et al. (2012) interpreted similar granites as partial melts of both TTGs (s.s.) and transitional TTGs (TTG s.l.). They contain on average higher Th concentration than TTGs, which is considered defining feature of Archaean granitoids than include input from older granitoids (Moyen, 2011). On the source diagram of Laurent et al. (2014) they plot mostly in the field of tonalite source but include also metasedimentary source and mafic sources. Multitude of sources is in accordance with the numerous experimental results that indicate that granitoid magmas are products of virtually any crustal rock type while the exact composition depends on many factors (source composition, mineralogy, degree of melting, amount of volatiles). Progressive melting of the older crust and sediments may explain enrichment of alkalies and incompatible elements. Their flat HREE patterns indicate lack of garnet control in their formation further affirming their purely crustal origin. Their low Mg#, Ni and Cr indicate lack of mantle signature. Low Sr and negative Eu anomalies on the REE fractionation diagram indicate significant plagioclase control in their formation. As Mikkola et al. (2012) suggested they probably represent local crustal melts at low pressure.

5.5. Transitional TTG: this study

Low-HREE transitional TTG overlap with the felsic members of the low-HREE TTG subgroup and follow parallel fractionation trends albeit at higher SiO_2 content. Transitional TTG are more peraluminous and have distinctly higher K_2O and LILE content. Three broad groups can be defined: Transitional low-HREE TTGs very similar to low-HREE TTGs, transitional TTGs with positive Eu anomalies similar to low-HREE TTGs with positive Eu anomalies and a group characterised by high Y and Yb, less fractionated REE patterns and lower Sr/Y. Similar to corresponding group in high-K granite. The HREE pattern is convex indicating garnet at some point in its petrogenetic history (Fig. 8.).

Similarly to low-HREE TTGs, transitional TTGs with positive Eu anomalies are limited to high silica compositions. They are limited to high SiO_2 values (>70 wt%) and at the given silica range have higher Sr than the samples displaying negative Eu anomalies.

There is a small group that combine characteristics of high-HREE TTGs and transitional TTGs (Fig. 8). As in the case of TTGs, transitional high-HREE TTGs are very subordinate (11 samples). They have lower range of HREE and concentrate on the lower range of HREE content (~2-4 ppm Yb) compared to low-HREE TTGs.

One sample shares the shape of LREE pattern with CO_2-altered quartz diorites with low-HREE and negative Eu anomaly.

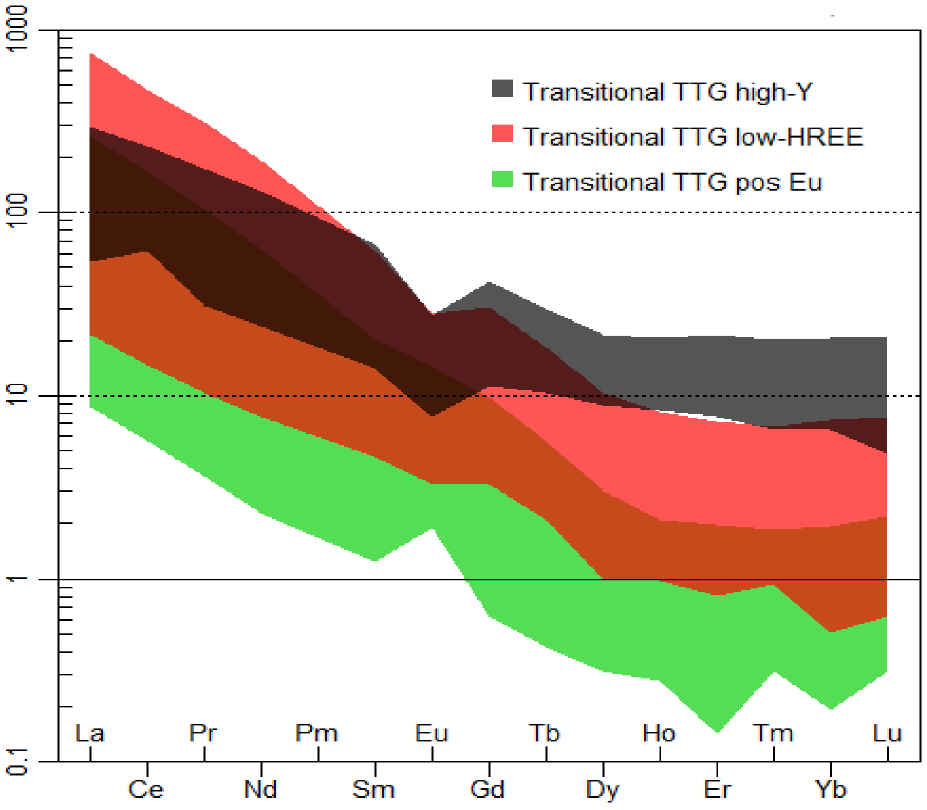

Fig. 8. Transitional TTGs show similar range of REE characteristics as sodic TTGs. Normalization values after Boynton (1984).

5.6. Formation of transitional TTGs

Transitional TTGs are very similar to TTGs and share many similarities and extensively overlap compositionally. They are typically more potassic and less sodic. Their combined range of Sr, Y, HFSE and HREE is greater than that of TTGs and indicate variety of processes that generated them. Champion and Sheraton (1997) have advocated generation of transitional TTGs over a range of pressures for similar rocks on the Yilgarn craton. Individual groups do not show an obvious geographical distribution but composite units might be seen as separating TTG blocks. Similar rocks have been reported recently from Wyoming Province (Frost et al., 1998, 2005).

In both Australian and Wyoming suites the authors have concluded, based on geochemical and isotopic data and presence of inherited zircons, that the petrogenesis of transitional TTGs requires input from pre-existing crust, but whether it is in the form of subducted sediments, or they represent purely crustal melts is unclear. Champion and Sheraton (1997) have also indicated a possibility that lower degree of melting and higher degree of crystal fractionation may give rise to granitoid rocks with higher potassium content. Champion and Sheraton (1997) have argued that their higher LILE require a source more felsic than MORB and the variation of Sr and Y argue against the source solely in melting slab. Champion et al. (1997) have advocated "remagmatisation" theory earlier proposed by Wyborn (1992). The term was introduced by Chappell and Stephens (1988) to refer to a process where granites are produces by partial melting with no significant change in composition, i.e. they inherit older compositions.

Remelting of a TTG protolith with minor restite separation could explain the overall similarity to TTGs, higher LILE and more felsic compositions of transitional TTGs. This process could provide some insight for the presence of isotopic evidence for older felsic crust and the presence of inherited zircons on the Karelian craton (Sorjonen-Ward et al., 2005). Negative Eu anomalies in many samples indicate residual or fractionating plagioclase whereas predominantly low Y and HREE indicate residual garnet in some stage of their development.

Joshi et al. (2017) have suggested that transitional TTGs with high HREE characteristics probably represent local *in situ* of amphibolite enclaves of typical TTGs. Enrichment of HREE may be result of garnet, allanite and apatite accumulation or fractionation from the restite or from the melt.

In Karelia transitional TTGs form two major age groups: 2.95 Ga and 2.83-2.78 Ga. Mikkola et al. (2011) have suggested the formation of transitional TTGs by melting of lower crust. Their formation is attributed to thickening of continental crust following continent collision, heating provided by mantle upwelling and intrusion of sanukitoid magmas. The continent collision may have been amalgamation of Superia supercraton, as suggested by Bleeker (2003).

5.7. A-type granite in this study

This study includes three A-type granites. They plot into WPG field of Pierce et al. (1984). They have slightly fractionated REE patterns (Fig. 9) with high HREE and negative Eu anomalies. In the classification of Frost et al. (2011) they straddle the calcic and calc-alkalic boundary. They have high silica (SiO_2 71-77 wt.%).

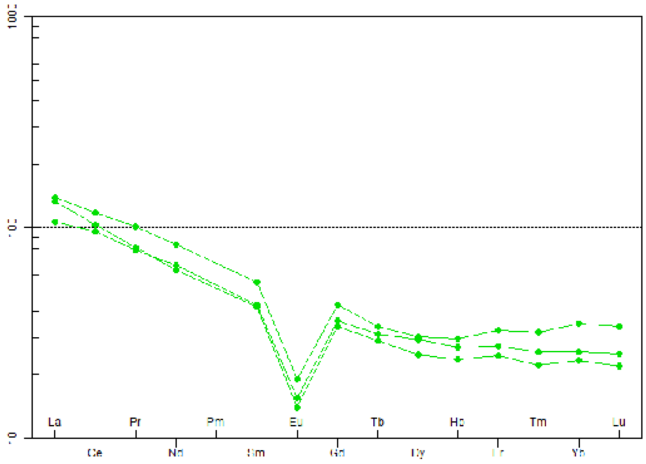

Fig. 9. The REE patterns of A-type granites display characteristic deep negative Eu anomalies and flat HREE end. Normalization after Boynton (1984).

Eby (1990, 1992) subdivided A-type granites in two groups by their Y/Nb. Here the samples are distinctly A2 type of Eby. A2 A-type granites include greater diversity of compositions: metaluminous, peraluminous and peralkaline. Here, they are peraluminous. They are widely scattered and appear as lone samples. Calc-alkalic A-type granites have typically high silica and lack low silica members. They have low Al_2O_3 relative to I-type granites (10-12 wt.% SiO_2). They have been postulated to result from melting of felsic granulite (Collins et al., 1982) or granodiorite and tonalite (Patiño-Douce, 1997). Calcic A-type granites are rare, high silica and lack low-silica members. These have low Al_2O_3 (10-13 wt.%) and very similar to experimental melts of granodiorite by Patiño-Douce (1997). Partial melting as a way of formation can help explain the lack of intermediate and mafic members.

Metaluminous members can form from low-pressure differentiation of tholeitic magmas; these are, however, associated with mafic intrusions and likely to have extremely high Fe. Calcic A-type granites form as a result of fractionation of basalt.

5.8. Formation of A-type granites

Many origins are proposed for A-type granites. Original proposition suggested formation by fractionation of mantle derived alkali basalt with possible crustal interaction. Often postulated crustal component or crustal source (Collins et al., 1982) is lower crustal granulite left after previous melting episode. Other proposed sources are quartz diorite, tonalite and granodiorite (Creaser et al., 1991). They are also similar to experimental melts by Patiño-Douce (1997). Martin (2006, 2012) has proposed a model for generation of A-type granites and their compositional variability. The model is inspired by the nonexistence of leucosome of A-type composition in migmatites terrains (Bonin, 2008) and apparent lack of crustal debris and refractory restitic material. It suggests a nearly complete melting of the source. In the model infertile lower crust that has been subjected to earlier melt removal is refertilized by fluid mix of H_2O and CO_2 from the mantle that carries alkalies, Fe, Si and mantle incompatible elements. Mantle upwelling follows crustal extension or delamination of lower crust following crustal thickening and formation of collisional granites. The model is supported by experimental data (Martin, 2012). The models support the formation of A-type granites in zones of crustal extension, which are at the same time zones of mantle upwelling and loci of mantle degassing. Fluids and gases bear silica and alkalies and can transport wide variety of elements and cause fenitization ion the lower crust that can subsequently melt completely to produce A-type magmas.

5.9. S-type and peraluminous granites in this study

S-type granites and leucogranites form a heterogeneous group unified by their strongly peraluminous nature (in this dataset A/CNK 1.11-1.37), normative corundum (>1%) and mostly high silica. Compositionally S-type granites range from true granites (s.s.) to trondhjemites. They are characterised by expanded silica range (60-76 wt.% SiO_2), variable major and trace element contents, K_2O/Na_2O, low contents of Ni, Cr, MgO and Mg# and variably enriched/depleted REE patterns. The only consistent characteristic is their highly peraluminous nature. Their peraluminous nature points towards origin by melting of graywackes and metapelites. Four groups can be defined. They appear as lone samples, widely scattered, apparently at internal and external boundaries of the Province.

Largest group spans narrow silica range (SiO_2 68-73.8 wt.%), low MgO (<1.75 wt.%) and other ferromagnesian oxides (≤5.26 wt.%). They have moderately fractionated REE patterns with negative Eu anomalies and flat HREE end of the pattern. Three samples have distinctly lower SiO_2 (58-63 wt.%) They have high Al_2O_3 (>15 wt.%), FeO (6-6.58 wt.%) and TiO_2. They have variably fractionated REE patterns with high HREE. Ni, Cr and MgO are high as well as Fe. Two samples have high Y and HREE and moderately fractionated REE patterns with deep negative Eu anomalies. One sample has distinct combination of high SiO_2 (75.2 wt.%) and MgO (2.44 wt.%), high Na_2O and low K_2O (4.32 and 0.22 wt.%, respectively) with low alumina (10 wt.%) and high A/CNK (1.37). In feldspar triangle of O'Connor (1965) it plots into trondhjemite corner.

5.10. Formation of S-type and peraluminous granites

S-type granites are results of interaction of metasedimentary rocks with basaltic magma (Patiño-Douce and Beard, 1995; Healy et al., 2004; Kemp et al., 2006). They form during the latter stages of supercontinent amalgamation (Zhu et al., 2020). In this case the supercraton amalgamation may have been Superia supercraton (Bleeker, 2003). Their diverse characteristics point to local origin and nature of formation. There is also a possibility that they are erroneously included into this dataset and that they are Proterozoic instead.

6. Discussion

Karelian province records almost a billion years of magmatism that includes both crustal growth and crustal reworking. It records a long period of TTG-only magmatism starting with rare occurrences of 3.5 Ga TTGs with major phases at 2.7 Ga and show evolution from TTGs to sanukitoids to granites (as major phases). Sanukitoids and granites appear in the Neoarchaean and mark the change from TTG to higher K calc-alkaline magmatism.

TTGs are commonly thought to be dominant rock type in Archaean terranes. Their dominance decreases towards the end of Archaean with the appearance of more mafic lithologies and start of intensive crustal reworking that produced late Archaean high-K granites. The characteristics of classical Archaean TTGs suggest little, if any, input from mantle in their genesis and as such they contrast the more mafic lithologies. The mantle contribution becomes established with the generation of adakites, sanukitoids and quartz diorites at the end of the Archaean, although it may be suggested by small number of TTGs and dioritic rocks with more mafic characteristics. The change in lithological characteristics is indicative of a change in the dominant crust-producing processes. Granites generally appear in the late Archaean and some of them are spatially and temporally associated with high-Mg rock types (adakites, sanukitoids and Closepet-type granites). These are often cited to be related to the final stabilisation of the craton at the transition from Archaean to Proterozoic (Kusky, 1999; Martin and Moyen, 2005). The origin of these granites is one of most important questions in relation to crustal evolution as they mark a change in crustal composition as well as in tectonic process. This broadly established sequence often ignores minor granitoid types (transitional TTGs, high-HFSE granites with A-type affinities, syenites and mafic granites) that make up the diversity of Archaean magmatism.

Large variation in major and trace element content is indicative of a variety of magma-producing processes operating during the Archaean. The lateral heterogeneity of Archaean cratons, craton-scale deformation patterns, strike-slip faults make some sort of plate-tectonic process an appealing mechanism to explain the differences between various crustal blocks (Bleeker, 2002). It has became apparent that Neoarchean crustal growth and evolution involved both the lateral accretion of juvenile terranes and the intrusion of arc magmas formed from mantle-derived and juvenile crustal sources as well as crustal reworking (Bleeker, 1998). In the early Archaean the crustal growth may have proceeded by low-angle underthrusting or crustal stacking (Smithies, 2000). Melting of oceanic plateaus has been also often cited as the source of Archaean TTGs (e.g. Condie, 2003). The small mafic component, most probably derived from subduction-modified mantle, strongly suggests that modern convergent margin processes operated at least at the end of the Archaean.

6.1. TTG-series

Classical TTGs are regarded products of partial melts of basaltic precursors though the processes that generated them remain debated (e.g. Rapp et al., 1990; Smithies, 2000; Smithies et al., 2003; Condie, 2004; Martin, 2004; Boily et al., 2004). Often postulated settings include melting of thick oceanic plateaus, shallow subduction of thick, hot plates (Martin, 1998), and in-situ crustal differentiation and delamination (Zegers, 2001). Many Archaean granite-greenstone terranes are interpreted as juvenile island arc sequences that grew up above subduction zones and later amalgamated during collisional orogenesis to form new continental crust (e.g. Martin, 1994).

Negative Nb-Ta-Ti-P anomalies are characteristic features of many continental crustal rocks, including the Archaean TTGs (Martin, 1994). Negative Nb-Ta and P anomalies are present in all Karelian TTGs but a small number of Karelian TTGs lacks a Ti anomaly. The negative Nb-Ta-Ti anomalies are frequently cited argument for island arc setting of Archaean TTGs (island arc signature) (Martin, 2005; Condie, 2008). Small number of TTGs with positive Ti anomalies has TiO_2 (~0.55 wt %) up to twice the average of Archaean TTGs (Martin, 1994).

Small Eu anomalies are often cited characteristic of Archaean high-Al TTGs. High-HFSE TTGs plot to the A-type granite field of Whalen (1987) but to VAG field of Pierce (1984) as do all Karelian TTGs. High-HFSE sum is a consequence of high Zr and Ce, as Nb and Y are well within typical Archaean TTG values. They may have different source, more enriched than that of classical TTGs, or higher temperature of melting, as postulated for true A-type granites (Whalen, 1987). They differ from true A-type granites by higher Eu and Sr and lack of significant negative Eu anomalies.

The intermediate rocks with TTG characteristics may represent higher degrees of melting of similar source (basaltic in composition) under similar conditions that did not involve contribution from mantle. The transitional rocks have lower silica and higher MgO, TiO_2 and CaO than true TTGs but are characterised by same transitional element (Ni, Cr, Co, V) contents, Mg# and REE characteristics as classical TTGs. Greater degree of melting may have contributed higher MgO, CaO, FeO and P transfer from the source. Martin does not uniformly use his classification in all his work and the silica range is frequently somewhat flexible.

Low-Al TTGs are located on the margins of high-Al classical TTG blocks, widely distributed as lone samples. Halla (2009) has argued that incipient hot subduction is a possible tectonic scenario for both high-Al and low Al TTGs where high-Al TTGs could be formed in the lower part of subducting crust and melting at shallow depths and low pressures could generate low-Al TTGs. Their locations on the margins of larger TTG blocks may lend some credence to this theory.

Transitional TTGs are not true TTGs according to definition used in this work (Martin, 1994) and many classifications used and accepted in literature (Lopez et al., 2006; Martin et al., 2005, Moyen et al., 2003). They belong to upper half of calc-alkaline and high-K calc-alkaline series on SiO_2 vs. K_2O plot of Peccerillo and Taylor (1976), similarly to high-K granites. They differ from high-K granites by lower silica, higher CaO and they follow different (subhorizontal) trend on Al_2O_3 vs. SiO_2 diagram in the case of high-Al high-K TTGs. Yet they are very similar in many aspects to sodic (classical Archaean) TTGs corresponding to high-Al or low-Al types, respectively. The major differences are higher K (>2 wt. % K_2O), parallel but not superimposed trends on Al_2O_3 vs. silica on Harker diagrams (low Al_2O_3 at given silica relative to TTGs), lack of most silica-rich compositions (>73 wt. % SiO_2) and lower Na_2O in transitional TTGs. HFSE, HREE and Sr are well within values of classical TTGs although they are concentrated on higher averages.

High-Y transitional TTG samples plot towards WPG field on tectonic discrimination diagrams and are located geographically adjacent to classical TTGs whereas low-HREE transitional TTGs are mostly located adjacent to high-K granites. The characteristics (HREE, HFSE and Sr) of high-Y TTGs point towards low pressure derivation from a source more enriched than the source of classical TTGs (Moyen, 2010).

Subduction setting may also provide explanation for transitional TTGs that are almost invariably located on margins of TTG blocks. Interaction with previously formed crust or subducted sediments may have provided for higher potassium and K_2O/Na_2O values. Thicker crust is typically associated with more felsic nature and higher K_2O and K_2O/Na_2O values (Best, 2003) at given silica values. Thicker crust induces a longer path and arrests or slows down rising magmas allowing time for fractionation.

Another rock that lack very convincing evidence for mantle interaction is quartz monzonite group. It is similarly depleted as low-HREE TTGs in HREE and Y that suggest of presence of garnet in restite as in the case of TTGs and high Sr content and minor negative Eu anomaly indicate high pressure origin. Low MgO, Ni and Cr contradict postulated mantle origin

Broadly linear and often parallel distribution of TTGs in belts on the Karelian craton is suggestive of tectonic stress during their genesis. Tectonic stress indicates lateral movement. The belts made up of a TTG subgroup are more frequently intersected by other rocks types than by other TTG groups suggesting that TTG belts may have originally been more extensive as may be suggested by earlier isotopic studies (Luukkonen et al., 1998). Where present in greater numbers the TTGs form belts that are nearly parallel to one another, curved from slightly to almost to 90 degrees, suggestive of some sort of plate tectonic process and induced stress. As TTGs are characterised by low content of transitional elements and low Mg# the tectonic setting probably did not include a development of mantle wedge and significant, if any, contribution from mantle (Smithies, 2009). A comfortable explanation that is often invoked to explain genesis of Archaean TTGs involves plate tectonic processes that differ from modern ones by lower angle of subduction that has prohibited formation of extensive, if any, mantle wedge (Smithies 2003, 2009; Rapp et al., 1991). However, alternatives to plate tectonics are gaining more acceptance at the expense of plate tectonic theory despite variations of. Evidence in favour of an origin in partial melting of oceanic plateaus is mounting

6.2. Mantle contribution: Adakites, sanukitoids, quartz diorites and high-Mg TTGs

Adakites and sanukitoids are often postulated to represent the earliest evidence of mantle interaction with crust and oldest hint of modern style plate tectonic and plate marginal processes (Martin, 1998; Martin et al., 2005; Condie, 2005; Samsonov et al., 2004). The theories of adakite, sanukitoid and Closepet-type granite genesis, although debated, are well established (Martin, 1994, 2005; Smithies et al., 2000; Richards et al., 2007; Thorkelson et al., 2005; Condie, 2004).

High-Mg TTGs resemble adakites closely, differing from them in detail. Major and the less mobile elements (HFSE, HREE, Ni and Cr) are within the range of adakites. This may suggest that these rocks are altered adakites.

High-Mg TTGs contrast the classical TTGs with their higher transitional element content. The generation of Archaean TTGs is often postulated to have formed via flat subduction (Smithies et al., 2003; Foley, 2008) or delamination of thickened lower crust after tectonic shortening (Condie, 2003) in both cases without significant interaction with mantle and consequently classical TTGs have low Mg#, Ni and Cr. If high-Mg TTGs are not altered adakites they may have also resulted from development of mantle wedge in subduction zones. Low angle subduction may result from high plate convergence rates (Currie and Beaumont, 2011; Wyman et al., 2008) and higher spreading rates are postulated for Archaean (Smithies, 2003). High-Mg TTGs appear as lone samples that indicate that the processes that led to their formation operated on a very limited scale.

Sanukitoids are frequently explained as results of melting of mantle wedge that has previously been metasomatized by TTG or adakite melt (e.g. Lobach-Zhuchenko et al., 2005; Kovalenko et al., 2005; Martin et al., 2010) and mark the first instance of detection of enriched mantle bellow Archaean crust (Mikkola et al., 2011). The linear distribution of adakites and sanukitoids is perhaps the best argument within the scope of this work in favour of plate tectonic process for the generation of adakites and sanukitoids. Lobach-Zhuchenko et al. (2005) have also argued in favour of subduction-related setting for Karelian sanukitoids. Sanukitoids are spatially related to syenites similarly to description of sanukitoid-syenite association in Russian Karelia. There are interesting parallels between low-HREE TTGs and low-HREE quartz diorites reflected in high-HREE TTGs and high-HREE quartz diorites. Both indicate pressure control during their generation.

6.3. Crustal progeny granites

Evolution towards higher K_2O contents is well attested (Moyen, 2011). Input from older granitoids could explain the differences between TTGs and transitional TTG. Maybe it can explain the difference between transitional TTGs and high-K granites. Are transitional TTGs an intermediate step in the generation of high-K granites? Granites of Karelian craton can be divided into two large groups: transitional TTGs and high-K granites and three minor groups (A-type granites, S-type granites and leucogranites). Transitional TTGs are second most abundant group (after TTGs) in this study.

Tectonic settings are out of the scope of this work and cannot be constrained on chemical data alone but the characteristics of deep-level granites are consistent with granites from continental-margin arcs (Condie, 1986). High silica, Sr and Zr, and positive Eu, Sr and Zr anomalies on element variation diagrams are also characteristics of felsic cumulates (Whalen et al., 2002). Fractionation of minor phases, as in the case of some TTGs with similar trace element characteristics, may be another possibility.

High-K granites are similar to late Archaean high-K granites as reported widely in literature. They have characteristics of I-type granites produced by intracrustal melting (López et al., 2006; Moyen et al., 2003; Ilbeyly et al., 2003; Boily et al., 2004; Castro, 2020). The generation of late Archaean granites is best explained by intracrustal melting of pre-existing TTGs (e.g. Springer et al., 1992; Boily et al., 2004; López et al., 2005; Moyen et al., 2003; Rapp et al., 1992).

High-K granites and leucogranites reflect further evolution and increased reworking of Archaean crust and the tendency of crustal rocks to become more potassic (LILE-enriched; Cassidy, 2006) and also more variable in time (Champion, 2001). High-K granites (biotite granite) and leucogranites are given separate treatment in this work as they were discovered separately. S-type and peraluminous granites have wide range of compositions and represent extreme crustal reworking.

Granites were, along with sanukitoids, typically final additions to continental crust that are related to stabilisation of Archaean cratons that have provided buoyancy to protect the cratons from recycling back into mantle, that is cratonize a province (Cassidy, 2006). The Karelian craton records about 1 000 Ma of crustal evolution from rare occurrences of 3.5 Ga TTGs to the end of the Archaean (late Archaean granites (2.7 Ga in Karelia) and sanukitoids). The beginning of evolution of Karelian craton may have started by 3.6-3.3 Ga and the oldest preserved rocks are dated to 3.5 Ga the time interval between about 2800 and 2600 Ma in the late Archaean was a period of intensive magmatism and crustal growth and in the end also of crustal reworking (Sorjonen-Ward et al., 2005; Heilimo et al., 2017). It also carries the record of transition from Archaean-like plate tectonics towards modern-like plate tectonics though the nature of Archaean tectonics remains controversial in spite of intensive research put into Archaean tectonic and crustal evolution. In part it is a time of simultaneous mantle and crust derived magmatism that has resulted in compositional diversification of forming rocks. As TTG-greenstone formation was largely confined to this period in Earth's history the processes involved may have differed significantly from those operating today although they may persist on a limited scale today.

7. Conclusions

In geochemical terms Archaean evolution of Karelia can be considered typical of Archaean cratons worldwide.

There are numerous geochemical hints towards subduction although subduction cannot be deduced on geochemical grounds alone. TTGs have an "arc signature" and literature on TTGs is dominated by subduction-related theories. However, evidence against subduction-origin for TTGs is mounting (Smit et al., 2019, Halla, 2020). Temporal evolution of Archaean cratons follows same pattern albeit at different pace at each craton.

TTG-intrusions have been suggested to represent magma chambers of ancient, long gone volcanoes (Laurent et al., 2020), a suggestion also speculated by Mikkola et al. (2011). In parallel to this idea a search for volcanic products of sanukitoid and granite magmatism could be attempted to further the understanding of Archaean geology.

Adakites, sanukitoids and TTG form compositional continuum where endmembers are clearly differentiated but gradation is continuous. It is no surprise as the metasomatizing fluid in sanukitoids is postulated TTG melt. Adakites are TTG melts contaminated by mantle wedge peridotite. On the other hand it confirms Moyen's thesis (2011) of ubiquitous mafic source in the early to middle Archaean and significant presence till the end of Archaean eon.

An effort was made to use the dataset to the maximum. This resulted in a description of Karelian granitoids "as is", including altered samples normally filtered out. In this case only the most extreme samples were filtered.

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