Isotopic evidence for the origin of Proterozoic massif-type anorthosites and their relation to rapakivi granites in southern Finland and northern Brazil

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Cover photo: Outcrops of the Repartimento anorthosite of the Mucajaí complex along the Rio Repartimento, in Roraima, northern Brazil.

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Abstract

The purpose of this study was to apply novel isotopic methods to shed new light on the petrogenesis and magmatic evolution of Proterozoic massif-type anorthosites and related rapakivi granites in southern Finland and northern Brazil. Representative rock types from two AMCG (Anorthosite-Mangerite-Charnockite-Granite) complexes of similar age and composition, the 1640 Ma Ahvenisto complex in Finland and the 1530 Ma Mucajaí complex in northern Amazonia, were studied.

The methodological emphasis of the study was on high-precision conventional (ID-TIMS) U–Pb geochronology and the implementation of *in situ* analytics of Lu–Hf (LAM-ICP-MS) and oxygen (SIMS) isotopes of zircon. One of the aims was to create spatially coupled zircon Hf–O isotope datasets to trace the magmatic evolution of the parental melts of the anorthositic and granitic rocks. The results were compared to those from conventional isotopic methods (ID-TIMS, laser fluorination) that utilize bulk sampling (whole-rock and mineral) and routine isotope tracers such as the Sm–Nd, Rb–Sr, and Pb–Pb systems.

Absolute zircon U–Pb ages from 1642 Ma to 1636 Ma in the Ahvenisto complex and from 1527 Ma to 1519 Ma in the Mucajaí complex were measured. Crystallization age patterns in both complexes are similar. All major rock types (anorthositic and gabbroic rocks, granites, monzodiorites) have overlapping ages. Only the most evolved granitic rocks in both suites (late quartz-feldspar porphyries in Ahvenisto, biotite granite in Mucajaí) are marginally younger than the rest of the rock types.

Previous conventional tracer isotope methods (e.g., TIMS Sm–Nd and laser fluorination O) have provided overlapping compositions for all AMCG rock types, but results suggest that some source heterogeneity may exist. The whole-rock ϵ_{Nd} isotope values obtained in this study for AMCG rocks in both complexes follow similar patterns ranging from +0.4 to -2.1 (at 1640 Ma) in Ahvenisto and -1.9 to -2.8 (at 1525 Ma) in Mucajaí.

In situ analysis of zircon Lu–Hf isotopes conducted in the Finnish rapakivi suite rocks and especially in the Ahvenisto complex revealed a depleted mantle signature (initial $\varepsilon_{\rm Hf}$ values up to +9 at ~1640 Ma) in the most primitive, olivine-bearing, rock types; this could not be detected in the Repartimento anorthosite of the Mucajaí complex. Rapakivi granites from both study areas displayed rather homogeneous Hf isotope compositions (~±2 ε units at 2SD) with average $\varepsilon_{\rm Hf}$ values from +1.0 to -0.1 at ~1640 Ma and +0.1 to -2.2 at 1540 Ma in the Finnish suite and from -2.0 to -3.1 at 1525 Ma in the Mucajaí complex, which strengthens the hypothesis that the parental magmas of rapakivi granites were most likely derived from the Paleoproterozoic crust.

Zircon oxygen isotope studies in both study areas produced relatively low δ^{18} O value estimates for the primary magmas of the anorthositic rocks ($\delta^{18}O_{zir}$ approximately from 5.5‰ to 7.5‰ in Ahvenisto and ~6.5‰ in Mucajaí) and somewhat higher values for the rapakivi granites (~8.0‰ in Ahvenisto and from 6.0‰ to 7.5‰ in Mucajaí). The lowest $\delta^{18}O_{zir}$ values were observed in the Ahvenisto leucotroctolite ($\delta^{18}O_{zir}$ from 5.4‰ to 7.0‰), which also had the highest initial ϵ_{Hf} values (+0.3 to +5.2 at 1640 Ma) within the Ahvenisto complex. This correlation was taken as evidence of a depleted mantle source component in the Finnish AMCG suite, even though most of them display crustal isotope signatures ($\delta^{18}O_{zir}$ from 6.3‰ to 7.8‰; ϵ_{Hf} from -1.5 to +5.1; 1640 Ma). The somewhat higher $\delta^{18}O_{zir}$ values observed in the rapakivi granites from both complexes ($\delta^{18}O_{zir}$ ~8.0‰ in Ahvenisto and from 6.0‰ to 7.5‰ in Mucajaí) support the assumption of their crustal source.

Tiivistelmä (in Finnish)

Tämän tutkimuksen tavoitteena on ollut soveltaa uusia isotooppigeologisia menetelmiä proterotsooisten, massiivi-tyypin anortosiittien ja niihin liittyvien rapakivigraniittien synnyn ja magmaattisen evoluution tutkimukseen. Tutkimusalueina ovat olleet Etelä-Suomi ja Pohjois-Brasilia. Tarkempia tutkimuksia ja kenttätöitä tehtiin kahden samantyyppisen proterotsooisen AMCG (anortosiitti-mangeriitti-charnockiitti-graniitti) kompleksin alueella (1640 Ma Ahveniston kompleksi Etelä-Suomessa ja 1530 Ma Mucajaí Pohjois-Brasiliassa).

Tutkimuksen menetelmällinen painotus on ollut konventionaalisessa (ID-TIMS) U–Pb geokronologiassa ja zirkonin Lu–Hf (LAM-ICP-MS) ja O isotooppien (SIMS) *in situ* analytiikassa. Yhtenä tavoitteena on ollut saada aikaan spatiaalisesti kytkettyjä Hf–O isotooppiaineistoja, joiden avulla on pyritty tutkimaan anortosiittien ja graniittien kantasulien kehitystä. Tuloksia on vertailtu konventionaalisin, kokokivi- ja bulkkimineraalinäytteitä hyödyntävien menetelmin (ID-TIMS, laserfluorinaatio) mitattuihin Sm–Nd-, Rb–Sr-, Pb–Pb- ja O isotooppituloksiin.

Zirkoni U–Pb menetelmällä määritetyt kiteytymisiät vaihtelevat Ahvenistossa välillä 1642-1636 Ma ja Mucajaíssa 1527-1519 Ma. Molempien AMCG kompleksien sisäinen kivilajien ikäjakauma on samantyyppinen. Kaikki kompleksien pääkivilajit (anortosiitit ja gabrot, graniitit ja monzodioriittiset kivet) ovat saman ikäisiä. Vain kaikkein kehittyneimmät graniittiset kivilajit kummassakin kompleksissa ovat U–Pb-menetelmän määritysrajoissa muita kivilajeja nuorempia.

Konventionaalisten isotooppimenetelmillä (mm. TIMS Sm–Nd ja laserfluorinaatio 0) tehtyjen tutkimusten perusteella anortosiittien ja graniittien kantamagmojen lähteet ovat saattaneet olla peräisin kahdesta eri lähteestä, ylävaipasta ja alakuoresta. Päällekkäisyydet kivilajien isotooppikoostumuksissa ovat kuitenkin aiheuttaneet ongelmia lähdetulkintojen tekemisessä. Kivilajien initiaaliset Nd-isotooppikoostumukset noudattelevat samanlaisia linjoja molempien tutkittujen kompleksien osalta (kokokivi ε_{Nd} = +0.4--2.1; 1640 Ma Ahvenistossa ja -1.9--2.8; 1525 Ma Muca-jaíssa).

Etelä-Suomen rapakivialueen kivistä tehdyt zirkonin *in situ* Lu–Hf isotooppianalyysit paljastivat kaikkein primitiivisimmissä anortosiittisissa kivilajeissa köyhtyneen vaipan koostumusta muistuttavan isotooppisignaalin (initiaaliset $\varepsilon_{\rm Hf}$ arvot jopa +9; 1640 Ma), jota ei puolestaan havaittu Mucajaí kompleksin anortosiitissa. Molempien tutkimusalueiden rapakivigraniiteista mitattiin suhteellisen homogeeniset (~±2 ε yksikköä; 2SD) zirkonin Hf isotooppikoostumukset (keskiarvo $\varepsilon_{\rm Hf}$ arvot +1.0--0.1; 1640 Ma ja +0.1--2.2, 1540 Ma Etelä-Suomen graniiteissa ja –2.0--3.1; 1525 Ma Mucajaí kompleksin graniiteissa), joiden perusteella voidaan päätellä rapakivigraniittien kantamagmojen olleen peräisin paleoproterotsooisesta alakuoresta.

Molemmilla tutkimusalueilla tehtyjen zirkonin happi-isotooppitutkimusten perusteella anortosiittisten kivien kantamagmojen $\delta^{18}O$ arvot olivat suhteellisen matalat ($\delta^{18}O_{zir}$ ~5.5-7.5 ‰ Ahvenistossa ja ~6.5 ‰ Mucajaíssa). Matalimmat $\delta^{18}O_{zir}$ arvot mitattiin Ahveniston kompleksin leuktroktoliittisesta näytteestä ($\delta^{18}O_{zir}$ ~5.4-7.0 ‰), josta mitattiin myös Ahveniston kompleksin korkeimmat $\epsilon_{\rm Hf}$ arvot (+0.3-+5.2; 1640 Ma). Tämän korrelaation perusteella pääteltiin, että Etelä-Suomen anortosiittisten kivien kantamagmojen lähde sijaitsi köyhtyneessä vaipassa, vaikka niistä suurinta osaa luonnehtivatkin kuorelliset isotooppikoostumukset ($\delta^{18}O_{zir}$ arvot ($\delta^{18}O_{zir}$ ~8.0 ‰ Ahveniston graniiteille mitattiin jonkin verran korkeammat $\delta^{18}O_{zir}$ arvot ($\delta^{18}O_{zir}$ ~8.0 ‰ Ahveniston graniitissa, 6.0-7.5 ‰ Mucajaín graniiteissa), mikä tukee oletusta graniittien kuorellisesta lähteestä.

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"To see the world in a grain of sand..." -William Blake-

> "S = k ln W" -The Known Universe-

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List of original publications

This thesis is based on the four (I-IV) peer-reviewed articles listed below. The papers are referred to with their roman numerals in the text.

PAPER I: **Heinonen, A.**, Andersen, T., & Rämö, O.T. (2010). Re-evaluation of Rapakivi Petrogenesis: Source Constraints from the Hf Isotope Composition of Zircon in the Rapakivi Granites and Associated Mafic Rocks of Southern Finland. *Journal of Petrology* 51, 1687-1709.

PAPER II: **Heinonen, A.**, Rämö, O.T., Mänttäri, I., Johanson B.S., & Alviola R. (2010). Formation and fractionation of high-Al tholeiitic melts in the Ahvenisto complex southeastern Finland. *The Canadian Mineralogist* 48, 969-990.

PAPER III: **Heinonen, A.**, Andersen, T., Rämö, O.T., & Whitehouse, M. Combined *in situ* zircon Hf–O isotope analysis refines the role of depleted mantle in Proterozoic AMCG magmatism. *Manuscript submitted to Chemical Geology*.

PAPER IV: **Heinonen, A.**, Fraga, L.M., Rämö, O.T., Dall'Agnol, R., Mänttäri, I., & Andersen, T. (2012). Petrogenesis of the igneous Mucajaí AMG complex, northern Amazonian craton – geochemical, U–Pb geochronological, and Nd–Hf–O isotopic constraints. *Lithos*, in press.

Author's contributions

I: A.H. conducted the Hf isotope LAMS and trace element analysis with T.A. and O.T.R., processed the data, and prepared the manuscript with contributions from the other authors.

II: A.H. conducted the additional U–Pb ID-TIMS analysis and calculations with I.M., processed the geochemical and isotopic data, and prepared the manuscript with contributions from the other authors.

III: A.H. planned the study, conducted the isotopic analysis with T.A. and M.W., processed the data, and prepared the manuscript with contributions from the other authors.

IV: A.H. conducted the field-work with L.M.F, O.T.R., and R.D. and the isotopic analysis with I.M. and T.A., and prepared the manuscript with contributions from the other authors.

Abbreviations

A	=?
AFC	=Assimilation Fractional Crystallization
AMCG	=Anorthosite-Mangerite-Charnockite-Granite
AMG	=Anorthosite-Mangerite(or Monzonite or Monzodiorite)-Granite
BSE	=Bulk Silicate Earth
cf.	=confer
CHUR	=CHondritic Uniform Reservoir
CIPW	=Cross, Iddings, Pirsson, and Washington normative calculation scheme
DM	=Depleted Mantle
e.g.	=exempli gratia
ESA	=ElectroStatic Analyzer
HFS(E)	=High Field Strength (Elements)
IAB	=Island Arc Basalt
ID-TIMS	=Isotope Dilution Thermal Ionization Mass Spectrometry
i.e.	=id est
LAM-ICP-MS	
(LAMS)	=Laser Ablation Multicollector Inductively Coupled Plasma Mass Spectrometry
LAQ-ICP-MS	=Laser Ablation Quadrupole Inductively Coupled Plasma Mass Spectrometry
LIL	=Large Ion Lithophile (Elements)
OIB	=Ocean Island Basalt
REE	=Rare Earth Element
SAMBA	=South AMerica Baltica
SD	=Standard Deviation
SEM	=Scanning Electron Microscope
SHRIMP	=Sensitive High Resolution Ion Micro Probe
SIMS	=Secondary Ion Mass Spectrometry
SMOW	=Standard Mean Ocean Water
T _{1/2}	=Half-life
T _{DM}	=Depleted mantle model age

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1. A as in Anorthosite: Brief introduction to A-type granites and massif-type anorthosites

1.1. A typical granite of A-type

The term A-type granite was originally coined by Loiselle & Wones (1979) in an abstract that since its publication has become one of the most quoted short communications in the history of petrological research. The original purpose of the paper was to extend the so-called "alphabetic granite classification" scheme of Chappell & White (1974) to include a fourth member with distinct geochemical features. Therefore in contrast to the original S- and I-type (Chappell & White, 1974) and the later added M-type (White, 1979) granites, which were defined by their source characteristics (S for sedimentary, I for Igneous, and M for mantle [as in oceanic slabs subducted to the upper mantle]) the A did not refer to the source of A-type granites. Instead it was used variably to refer to the Anorogenic nature of magmatism, or broadly Alkaline or Anhydrous geochemical features of the granites. In the recent debate on granite typology (Bonin, 2007) the letter A has mostly become to signify the Ambiguous (Whalen, 2005) nature of the group.

Despite of the vague original nature of the term, it has been eagerly adopted by several researchers to signify a granitic rock with high levels of HFS and LIL elements, total alkalis, iron, and Fe/Mg and K/Na ratios, low levels of elements compatible in mafic silicates (Co, Sc, Cr, Ni) and feldspars (Ca, Al, Ba, Sr, Eu) and low magmatic oxygen and water fugacities (e.g., Bonin, 2007). Often the term also carries a tectonic burden that places the rocks in an anorogenic, extensional, or rift tectonic setting.

The original definition of A-type granites (Loiselle & Wones, 1979) did indeed suggest an extensional tectonic environment and a mantle-derived alkali-basaltic composition for their primary magmas. Possible crustal contamination was deemed as less important than for the I- and S-types (Chappell & White, 1974) generated in convergent tectonic settings. Also, a residual crustal source (signified with the letter R) was considered by White (1979) for the "A-type-like" granites of southeastern Australia (Collins et al., 1982; Bonin, 2007) and generalized later by Creaser et al. (1991).

This early controversy has since allowed many different approaches and hypotheses to be set over the source of the granites and therefore, no consensus over the matter exists even to date (Bonin, 2007; Frost & Frost, 2011). Models for straight mantle derivation (e.g., Turner et al., 1992), crustal derivation (e.g., Rämö & Haapala, 2005), and several different combinations of sources and processes (e.g., Collins et al., 1982; Frost & Frost, 1997; Martin, 2006) have been set forward in pursuit of a unifying synthesis.

Unfortunately, but not unexpectedly, several attempts at classification, grouping (e.g., Pearce et al., 1984; Bonin, 2007; Dall'Agnol & Oliveira, 2007), and regrouping (Eby, 1990; 1992) of these rocks have met difficulties and inevitable critique as it has turned out that the original classification scheme most likely includes several different (but in crucial ways still similar) groups of granitic rocks (Frost & Frost, 2011) deriving from different origins in varying tectonic environments. Hence the ironic "there are A-type granites and A-type granites" anecdote very popular among certain professionals in the field.

Thus, the term A-type granite has become highly contested (Bonin, 2007) and is slowly being replaced by more descriptive terms based on geochemical classification (e.g., Frost et al., 2001; Frost & Frost, 2011) that recognize the variation within the group and allow a more distinctive classification. This progress will eventually also allow the solution of the prevailing source controversy by allowing different origins for different types of A-types.

1.1.1. Rapakivi granites

Consideration of granite typology is essential as this study deals with rapakivi granites, which are a special sub-group of A-type granites. Rapakivi granites are defined (Haapala & Rämö, 1992; Rämö & Haapala, 2005) as "A-type granites characterized by the presence, at least in the larger batholiths, of granite varieties showing the rapakivi texture."

Rapakivi granites are a prominent feature of the Precambrian bedrock of southern Finland recognized already in the late 19th century pioneering studies of Sederholm (1891). Rapakivi texture, which in addition to the specific geochemistry of the granites, defines the rapakivi granite clan consists of large, ovoid alkali-feldspar crystals that are mantled by plagioclase (oligoclaseandesine) rims (Vorma, 1976). Another diagnostic feature is the two-stage crystallization of alkali feldspar and quartz, which is manifested by the drop-like crystal forms of the earlier hightemperature generation (Vorma, 1976; Rämö & Haapala, 2005). A well-developed rapakivi texture with mantled alkali feldspar ovoids grants the name *"wiborgite"* (Fig. 1a) to the granite as a less well-developed porphyritic texture without the plagioclase rims corresponds to the *"pyterlitic"* (Fig. 1b) granite variety (Rämö & Haapala, 2005). As even-grained rapakivi granites are also very common, these type names are used in conjunction with normal descriptive rock type names.



Figure 1. Two main types of rapakivi texture from the Finnish rapakivi suite, (a) wiborgite showing the distinct plagioclase mantled ovoids and (b) pyterlite without the plagioclase rims. Figure is from Heinonen (2010a).

Rapakivi granites have, since their initial discovery in southern Finland, been described from all present continents and as spanning a considerable age distribution from the late Archean to the Phanerozoic (Fig. 2; Haapala & Rämö, 1999). Most known rapakivi granites are nevertheless Proterozoic (1.8 to 1.0 Ga) and typically intrude crustal terranes formed more than 50 m.y. earlier (Haapala & Rämö, 1999; Rämö & Haapala, 2005). Also the batholiths that comprise the early Mesoproterozoic rapakivi magmatism in the northern Amazonian craton (Venezuela and northern Brazil) have been recognized since the 1970s (Dall'Agnol et al., 1999 and references therein) and described principally as wiborgitic rapakivi granites (e.g., Gaudette et al., 1978).

1.2. Proterozoic massif-type anorthosites

The problem of A-type granite petrogenesis also converges with another major petrological dilemma of recent decades: the petrogenesis of the Proterozoic massif-type anorthosites (Ashwal, 1993).

Proterozoic massif-type anorthosites are igneous rocks that consist mainly of plagioclase (>90 vol.%; LeMaitre, 2002). Classical account of their classification was given by Ashwal (1993), who defined anorthositic rocks (leucogabbronorites, leuconorites, leucogabbros, and leucotroctolite) as containing more than 75 % of modal plagioclase, and less than 25 % of respective mafic minerals

(Px, Ol, Fe-Ti oxides, Ap). Pure massif-type anorthosite is essentially a cumulate rock, which contains more than 90 % of modal, usually intermediate $(An_{50\pm10})$ plagioclase.

Massif-type anorthositic rocks are commonly associated with coeval granitic rocks that have suggestive A-type (or charnockitic) features in the so-called AMCG (Anorthosite-Mangerite-Charnockite-Granite) suites (Emslie, 1978; Ashwal, 1993). Classical occurrences of AMCG rocks are found in the Grenville terrain of northeastern Canada (e.g., Emslie, 1978; Wiebe, 1992; Emslie et al., 1994; McLelland et al., 2010), where they comprise a major component of the continental crust along large areas (anorthosites comprise up to 8-9 % of areal extent; Ashwal, 2010).



Figure 2. Some known localities of Proterozoic AMCG magmatism with emphasis on occurrences of rapakivi granites. Numbers denote age in Ga. AMCG localities are divided into three different classes on the basis of the distribution of their dominant rock types. Occurrences dominated by massif-type anorthosites are most abundant in eastern Canada (Grenville & Labrador); elsewhere both of the components are equally represented or the granitic component dominates. These differences in distributions may represent a real difference in mutual abundances of the rock types but it is better explained by exposure bias. Seismic data from the southern Finnish rapakivi suite (Luosto et al., 1990; Elo & Korja, 1993; Rämö & Haapala, 2005) suggest that the rock type distribution in AMCG complexes is strongly affected by level of erosion (shallow exposures are dominated by granites, deeper ones by anorthositic rocks). Study areas of the present study are marked with reference to the pertaining papers and more detailed geological maps in Figs. 3 and 4. Figure is adapted from Haapala & Rämö (1999).

Petrogenesis of anorthosites has also been a long-standing point of debate in petrology (Bowen, 1917) as no volcanic rocks or inferred liquids of anorthositic composition are known in nature, and no common liquid would crystallize a rock containing as much plagioclase. Accumulation is therefore implied to have taken place during their genesis, but complementary ultramafic components are mostly absent from the associations.

This presupposition has produced a large number of competing theories on how these rocks were formed in the first place. The source, intrusion and emplacement mechanisms, amount of possible contamination, depth(s) of crystallization and emplacement, and tectonic setting have all been among major contestable points. As a result, a similar source controversy as in A-type granite petrology has ensued in anorthosite research.

Several models imply a significant mantle-contribution to the parental magmas of both the A-type granites and the anorthosites. These models include the so called *'tholeiitic model'* of Frost & Frost (1997) where mantle-derived magmas have intruded lower crustal levels and partially crystallized forming a mafic underplate. Continued mantle heat production has remelted the underplate and created partial melts that in turn react with (minimal amounts of) lower crustal components to

produce the granites and anorthositic rocks are formed as straight cumulates from the mantlederived melts. Other models (e.g., Turner et al., 1992; Frost et al., 2002; Anderson et al., 2003; Frost & Frost, 2011) suggest that at least some of the rock types in the associations are consanguineous *sensu stricto* via straight basaltic fractionation (with or without crustal contamination) from gabbroic rocks to monzodiorititic to granitic compositions.

At the other end of the spectrum are the *'all-crustal'* models (e.g., Duchesne et al., 1999; Longhi et al., 1999; Schiellerup et al., 2000), that explain all the rock types as consanguineous fractionates of a common crustal parental melt of jotunitic composition (Duchesne & Wilmart, 1997; Vander Auwera et al., 1998; Longhi et al., 1999) or as a result of two-stage lower-crustal partial melt extraction (Taylor et al., 1984). Most of these models emphasize the extensional tectonic control of petrogenesis that exposes lower crustal tongues along listric faults to astenospheric heat (Duchesne et al., 1999) causing partial melting.

Somewhere in the middle-ground lay the classical *'two source -models'* which distinguish a mantle source for the anorthosites and a dominant lower crustal source for the granitic rocks (e.g., Rämö, 1991; Emslie & Stirling, 1993; Emslie et al., 1994). These models have individually been applied to the anorthosite-dominated Grenvillean terranes (e.g., Emslie et al., 1994) as well as the *locus classicus* rapakivi granite area in southern Finland (Rämö & Haapala, 2005 and references therein; Papers I and II).

Tectonic environment of the massif-type anorthosites has also been a topic of active discussion (e.g., Ashwal, 2008, 2010; Corrigan, 2008; McLelland, 2008; McLelland et al., 2010) and a general trend of two broadly categorized tectonic settings, one of active margin (or post-collisional; e.g., Corrigan and Hanmer, 1997; McLelland et al., 2010) and the other of within-plate association, arises (e.g., Sharkov, 2010).

These several attempts at explaining both problems at once go to show that the petrogenesis of massif-type anorthosites and A-type granites are complicated processes and hold many common denominators. It is, however, very likely that not one of the above-mentioned schemes is correct for all studied associations and that some geodynamic contexts emphasize the importance of different theoretical facets than others (e.g., Bonin, 2007). Generalized inferences for A-type petrogenesis drawn from case studies therefore remain difficult to justify to the unforeseeable future.

2. Objectives of the study

The objective of this study was to shed light on the petrogenesis of Proterozoic A-type rapakivi granites and especially to decipher their petrogenetic relationship to the contemporaneous massiftype anorthositic rocks in two localities; southern Finland and northern Brazil. The use of elemental and especially major element geochemistry in the study of cumulate rocks, like massif-type anorthosites, can often be inconclusive. Therefore, most of the research presented in this thesis concentrates on petrologic tracer applications of isotopic methods, which are able to see through the cumulate effects, but as a tradeoff are susceptible to several other difficulties such as more complicated or unconventional methodology or lack of an agreed upon frame of interpretation.

Precambrian petrology and utilization of radiogenic tracer methods also heavily rely on U–Pb isotope geochronology capable of precise determination of igneous crystallization ages for magmatic rocks. Accordingly, the other focus in the study has been the application of suitable U-Pb dating methods to determine accurate ages for the AMCG rocks in both studied localities. Evidently AMCG complexes can be linked to at least two major geodynamic and tectonic environments, an extensional and a post-collisional one, that have been able to produce A-type granites and massif-type anorthosites through diverse petrogenetic processes (e.g., McLelland et al., 2010). The main purpose of this study has been to find out how the different models and processes suggested for AMCG petrogenesis test out and can be combined in the light of isotope data collected from the Ahvenisto and Mucajaí AMCG complexes. The 'two-source model' that has been traditionally applied to the Finnish rapakivi granite terrain (e.g., Rämö & Haapala, 2005) was taken as the working hypothesis from which the following spin-offs were derived:

Hypothesis 1: Massif-type anorthosites and rapakivi granites are genetically linked but derived from discrete sources.

Hypothesis 2: Mantle is the primary source of the massif-type anorthosites.

Hypothesis 3: Lower crust is the primary source of the A-type granites.

Hypothesis 4: The mantle component was variably depleted.

Hypothesis 5: Mafic and felsic magmas have interacted during their ascent and emplacement in an open system.

3. Methods

3.1. Isotopic methods in igneous petrology

The use of radiogenic isotope methods in igneous petrology is based on the predictable decay rate of radioactive parent isotopes to stable daughter nuclides and precise measurement of the mutual abundances of different isotopes of a certain element in rocks or different minerals separated from them (e.g., Faure & Mensing, 2005). The usefulness of a certain isotope system is based on the decay rate of the parent, which is defined by the decay constant (λ), and the differences in the affinities of the parent and daughter relative to each other in the magmatic system in question. The long half-lifes of radioactive isotopes of Sm, U, Th, Lu, and Rb make them especially practical in the study of Precambrian geology.

3.1.1. U-Pb geochronology

In this study U–Pb analysis of uranium-rich minerals (high initial U–Pb ratio), mainly zircon, has been the primary method of determining magmatic crystallization ages of the studied rocks. ²³⁵U and ²³⁸U decay at different rates ($T_{\frac{1}{2}}$ = 0.704 Ga and 4.468 Ga, respectively) to stable isotopes of Pb (²⁰⁷Pb and ²⁰⁶Pb, respectively) and by measuring the Pb isotopic composition of a uranium-rich mineral, the time of crystallization of that mineral from the magma can be calculated. Furthermore, due to the different rate of decay of the two uranium isotopes, the U–Pb method can be exploited to determine the crystallization age of a mineral even if the system has been disturbed by later Pb or U loss (Wetherill, 1956).

The U–Pb method requires a mineral that acquires a high U/Pb ratio at the time of crystallization and which is ideally resistant to later alteration. Zircon (ZrSiO₄) is a near-ubiquitous mineral in igneous rocks that fulfills these requirements and has therefore been used almost exclusively also in this study. Baddeleyite is another high-U/Pb mineral, which is especially useful in dating mafic rocks that often lack sufficient zircon altogether or require inconveniently large samples to produce adequate zircon fractions for isotopic work. Baddeleyite was extracted from some of the anorthositic rocks to provide additional age constraints to complement the zircon U-Pb data (Paper II).

U–Pb dating of rocks can be performed on a number of mass spectrometric instruments. The conventional ID-TIMS (Isotope-Dilution Thermal-Ionization-Mass-Spectrometry) method has been used to determine the ages of the rocks in Papers I-IV. The ID-TIMS method utilizes isotope spiking and chromatographic separation of U and Pb, most commonly, from a bulk mineral fraction (Krogh, 1973, 1982). The separated U and Pb are precipitated on either Re (Pb) or Ta (U) filaments and ionized thermally in a vacuum by leading a current through the filaments. The Pb and U ions, are then accelerated and their ratios measured by a multi-collector mass spectrometer.

Additional material reported in the Discussion section of this thesis also includes *in situ* zircon U–Pb data acquired by the SIMS (Secondary Ion Mass Spectrometry; e.g., Whitehouse & Kamber, 2005) and LAM-ICP-MS (Laser Ablation Multicollector Inductively Coupled Plasma Mass Spectrometry; e.g., Andersen et al., 2009; Rosa et al., 2009) methods. The LAM-ICP-MS or for short, the LAMS and SIMS are dating methods, where individual mineral crystals can be analyzed by ablating a solid sample with a high-energy laser beam under Ar or Ar + He flux (LAMS) or oxygen ion beam in a high vacuum (SIMS).

In LAMS the laser-ablated aerosol is led to an argon plasma torch burning at $\sim 10~000$ K, in which the sample is vaporized and ionized to be accelerated to a double-focusing mass spectrometer. In SIMS, the secondary ions released from the surface of the sample by the primary ion bombardment are directly accelerated through a double-focusing mass spectrometer.

LAMS and SIMS are referred to as *in situ* methods, because of the greater spatial resolution that can be achieved by solid sample introduction compared to bulk ID-TIMS analysis (see Section 2.1.3). Single crystals or mineral grains or, in some cases, even different portions of the same crystal can be analyzed and dated with relatively high precision.

As can be seen from Table 1, *in situ* and conventional methods are complementary and to obtain the best age result, both should ideally be used. First the zircon population should be examined by LAMS or SIMS to determine possible population heterogeneities. After the preliminary examination and identification of the most suitable and concordant mineral fractions, they should be analyzed with ID-TIMS to achieve the best possible precision. Often for homogeneous zircon populations, a high level of precision can be achieved also with *in situ* methods alone.

3.1.2. Isotopic tracers (Sm–Nd, Rb–Sr, Pb–Pb, Lu–Hf, δ^{18} O)

Isotopic tracer methods, like Sm–Nd, Lu–Hf or stable oxygen isotopes, are used in igneous petrology to trace the isotope composition of the magma that crystallized the studied rocks. The information about the initial magmatic isotope compositions of rocks can be used as petrogenetic indicators of their possible source components. Because heavy parent isotopes are fractionated from their stable daughters during magmatic differentiation, magmatic source reservoirs in the Earth are eventually characterized by different radiogenic isotope compositions. The isotopic composition of these reservoirs can be estimated through generalized models (e.g., DePaolo & Wasserburg, 1976; DePaolo, 1981; Griffin et al., 2000) that track the isotopic evolution through time after their initial segregation from the source. In contrast to the isotopic dating applications, which require high parent isotope abundances, in radiogenic tracer applications low parent/daughter ratios are beneficial as the correction to acquire initial values then becomes less significant.

Method	Pros	Cons
TIMS	 (+) High precision (2σ uncertainty in U/Pb ratios in the order of ~0.5 %). (+) Fairly simple and routine maintenance of equipment (compared to SIMS and LAMS). (+) Several different minerals can be routinely analyzed (e.g. zircon, baddeleyite, monazite, sphene, rutile). (+) Clean samples (single focusing MS instrument is sufficient, isobaric interference problems negligible). 	 (-) Large amount of sample needed (several μg:s). (-) Bulk analysis, no spatial control of data. (-) Long and meticulous chemical treatment of samples needed. (-) Single crystal analysis complicates chemical treatment. (-) Pre-treatment of crystals (physical or chemical abrasion) needed to enhance the concordance of results. (-) Slow measurement (Two or three different runs, separately for Pb and U, are needed for single analysis. In total several hours for a single measurement may be needed.)
LAMS	 (+) Small amount of sample needed (a single crystal or even a sufficiently large crystal domain is adequate for one analysis). (+) Fair spatial resolution (~30-50 μm spot). (+) Good spatial control of data (<i>In situ</i> analysis) can be used to resolve heterogeneous mineral populations. (+) Accurate documentation and control of analysis is possible (e.g. SEM imaging). (+) Easy and fast sample preparation (solid samples, no chemistry needed). (+) Fast measurement (~5 minutes for single run, where all species are measured simultaneously). (+) Spatial coupling to other isotope systems (e.g. Lu–Hf or O isotopes) or trace element compositions is possible. 	 (-) Semi-destructive, i.e. relatively high sample volume required to achieve best precision (larger than SIMS). (-) Low precision (2σ uncertainty in U/Pb ratios in the order of ~3.0 %). (-) Complicated maintenance of equipment. (-) Routine measurement of only zircon (other minerals, like monazite, are possible). (-) "Dirty samples" (double-focusing MS instrument with ESA is needed, interference problems might occur). (-) Complicated, matrix-dependent, standardization
SIMS	 (+) Small amount of sample needed (a single crystal or even a sufficiently large crystal domain is adequate). (+) Practically non-destructive (+) Good spatial resolution (~10 μm spot). (+) Good spatial control of data (<i>In situ</i> analysis) can be used to resolve heterogeneous mineral populations. (+) Accurate documentation and control of analysis is possible (e.g. SEM imaging). (+) Easy and fast sample preparation (solid samples, no chemistry needed). (+) Relatively fast measurement (~15 minutes for single run, where all species are measured simultaneously). (+) Spatial coupling to other isotope systems (e.g. Lu–Hf or O isotopes) or trace element compositions (by LAMS) is possible. (+) Routine analysis is possible, and enables chaining of analysis. 	 (-) Low precision (2σ uncertainty in U/Pb ratios in the order of ~2.0 %). (-) Complicated maintenance of equipment. (-) Routine measurement of only zircon (other minerals, like monazite, are possible). (-) "Dirty samples" (double-focusing MS instrument with ESA is needed, some interference problems might occur).

Table 1. Pros and cons of ID-TIMS, LAM-ICP-MS, and SIMS methods available in this study for U–Pb dating.

Sm–Nd and Lu–Hf systems are the main radiogenic tracer isotope tools utilized in this study. ¹⁴⁷Sm and ¹⁷⁶Lu are radioactive parents that decay to stable isotopes of ¹⁴³Nd (at $T_{\frac{1}{2}}$ = 106 Ga) and ¹⁷⁶Hf (at $T_{\frac{1}{2}}$ = 37.1 Ga), respectively. Sm–Nd isotopes are routinely analyzed from whole-rock powders by ID-TIMS following chromatographic separation of REE (Rare Earth Elements) and Sm and Nd aliquots from each other (Richard et al., 1976). Separation, chemical treatment, and thermal mass spectrometry of Lu–Hf isotopes is more complicated (e.g., Patchett et al., 1981) than for Sm–Nd, which has impeded routine analysis somewhat, but the recent advent of solid source ICP mass spectrometry has spurred a renewed interest in their petrologic applications. Lu–Hf isotopes are routinely analyzed by ICP-MS either with laser ablation or solution sample introduction.

Sm–Nd and Lu–Hf systems are analogous in the sense that the parent isotopes of both (¹⁴⁷Sm and ¹⁷⁶Lu) behave relatively more compatibly in partial melting of the mantle than their respective daughter isotopes (¹⁴³Nd and ¹⁷⁶Hf). Therefore the depleted upper mantle (DM; DePaolo, 1981), that has undergone repeated extractions of partial melts, is generally characterized by more radiogenic long-term daughter isotope compositions than contemporary crust with lower par-ent/daughter isotope ratios. This also leads to similar notions and applicability of the systems as for example the ε (epsilon) notion (Formulas 1A and 1B) is used for both systems to notify the divergence from a bulk earth value (BSE – the Bulk Silicate Earth, or CHUR – the CHondritic Uniform Reservoir).

(1A)
$$\varepsilon_{\text{Nd}}^{t} = \left[\frac{\left(\frac{1^{43}\text{Nd}}{1^{44}\text{Nd}}\right)_{\text{sample}}^{t}}{\left(\frac{1^{43}\text{Nd}}{1^{44}\text{Nd}}\right)_{\text{CHUR}}^{t}} - 1\right] \times 10^{4}$$

(1B)
$$\epsilon_{\rm Hf}^{\rm t} = \left[\frac{\left(\frac{176}{177}{\rm Hf}\right)_{\rm sample}^{\rm t}}{\left(\frac{176}{\rm Hf}\right)_{\rm CHUR}^{\rm t}} - 1\right] \times 10^4$$

where t is time. Values used in calculations of $\epsilon_{\rm Hf}$ values in this study are as follows: λ^{176} Lu = 1.867 x 10⁻¹¹ a⁻¹ (Scherer et al., 2001, 2007; Söderlund et al., 2004), present day CHUR composition 176 Hf/ 177 Hf = 0.282785, 176 Lu/ 177 Hf = 0.0336 (Bouvier et al., 2008) and for $\epsilon_{\rm Nd} \lambda^{147}$ Sm = 6.54 x 10⁻¹² a⁻¹ 143 Nd/ 144 Nd = 0.512638, 147 Sm/ 144 Nd = 0.1966 (DePaolo & Wasserburg, 1976).

Furthermore Lu is more compatible in relation to Hf than Sm is in relation to Nd and decay of ^{176}Lu is more rapid compared to ^{147}Sm ($\lambda^{176}Lu > \lambda^{147}Sm$). This leads to a systematic difference in resolution between the systems, which can also be utilized in petrogenetic inferences.

Initial Nd composition provided by bulk whole-rock ID-TIMS analysis is a robust indicator of source contribution and very useful in the study of magmatic systems, where internal sample heterogeneities are not implied. Also crustal residence times of the source, the T_{DM} (Nd) values, can be determined by extrapolating the Nd isotope evolution of the sample to the intersection point of DM curve (DePaolo, 1981). Whole-rock Sm–Nd isotopes were analyzed by ID-TIMS for Papers II and IV.

In this study the LAM-ICP-MS method has been used to analyze Lu–Hf isotopes from zircon (Papers I, III, IV). Zircon is the ideal medium for the Hf isotope tracer method as it has a very high affinity for tetravalent Hf ([VIII]Hf⁴⁺ = 0.83 Å ~ [VIII]Zr⁴⁺ = 0.84 Å) and incorporates relatively small amounts of trivalent Lu (e.g., Hoskin & Schaltegger, 2003). Therefore only small corrections for radiogenic growth after crystallization are needed to acquire magmatic initial isotope ratios. As a tradeoff, extrapolation of Hf isotope compositions from the zircon initial value has to utilize an assumed Lu/Hf ratio for growth in reservoir, which introduces additional uncertainty to the inferred T_{DM}(Hf) values and crustal residence times.

In contrast to elemental fractionation, fractionation of light isotopes such as ¹⁸O from ¹⁶O is nearly negligible in magmatic processes at high temperatures (e.g., Allègre, 2008). In the case of oxygen isotopes, enrichment in ¹⁸O results mainly from low-temperature processes and interaction with meteoric water at surface conditions (e.g., Valley, 2003; Valley et al., 2005; Bindeman, 2008). Therefore magmatic $\delta^{18}O$ (‰ difference relative to VSMOW, Vienna Standard Mean Ocean Water; ¹⁸O/¹⁶O = 2005.20 ± 0.43; Formula 2) signature preserves petrogenetic information and can be used as a proxy for supracrustal or recycled material in the source of igneous rocks or, alternatively, the influence of such a contaminant to a primary magmatic composition. In this study (Papers III and IV) O isotopes of zircon were analyzed by SIMS (Whitehouse & Nemchin, 2009).

(2)
$$\delta^{18}0 = \left[\frac{\binom{^{18}0}{^{16}0}}{\binom{^{18}0}{^{16}0}}_{_{VSMOW}}\right] \times 10^3$$

3.1.3. In situ vs. the incredible bulk

The majority of conventional methods in isotope analytics have for the past fifty years or so relied heavily on the ID-TIMS method, which has been perfected for routine analysis in, for example, U–Pb geochronology and Pb–Pb, Rb–Sr, and Sm–Nd tracer systems. The advantage of the TIMS method relies in high level of precision, unattainable by any other method to date (c.f. Table 1). The major setback of the method is, however, the loss of spatial information. Most of the TIMS analyses are performed on bulk samples, which are in the case of whole-rock samples essentially mixtures of the different rock-forming constituents - the individual minerals, or in the case of U–Pb dating, monomineralic, but potentially heterogeneous mixtures of high U/Pb minerals like zircon or baddeleyite.

In the course of its evolution, magma can experience several different stages or igneous processes, like assimilation or mixing, which might alter the isotope composition of the system and create isotopic heterogeneities or inherited domains in the resulting rocks. The minerals that crystallize during these processes are in some cases able to record the changes and information about their compositions could be used to track the processes. The averaging bulk samples utilized in ID-TIMS analytics are prone to lose all this information. Analogously the method of choice in 0 isotope studies has been laser fluorination analysis (e.g., Valley, 2003), which utilizes bulk mineral fractions (e.g., zircon) to determine δ^{18} O values.

Development of *in situ* isotope microanalytics during the past decades has opened new possibilities to overcome these problems and decipher possible heterogeneities in the composition of minerals. SIMS and SHRIMP (U–Pb and O) and, most recently, the LAMS (U–Pb and Hf) methods have provided powerful tools to overcome the challenges in spatial resolution (c.f., Section 2.1.1.).

The most obvious application of *in situ* methods in igneous petrology is to unravel heterogeneous zircon populations and resolve geochronological problems arising from mixed U–Pb ages (e.g., Huhma et al., 1991). It also seems that zircon Hf and O isotope compositions within single samples can be quite variable and that this may have petrogenetic significance (Papers I and IV).

With *in situ* methods it is also possible to combine isotopic information from two or even three different systems to create spatially coupled datasets. In theory it would be possible to analyze for example oxygen isotopes from zircon by SIMS (\sim 10-20 µm spot size, negligible pit depth), and after light repolishing of the sample mounts analyze U–Pb by LAM-ICP-MS (\sim 30 µm laser spot, \sim 30 µm pit depth) from the same spots to provide a crystallization age, and then, material permitting, also Hf isotopes from the same spots or adjacent domains in the same grain by LAM-ICP-MS (\sim 50 µm laser spot, >30 µm pit depth). As LAM-ICP-MS is effectively a destructive method, spatial

resolution in depth is understandably lost to some extent and requirements for the sample material are high (large grains are needed).

Ideally, coupled datasets could be acquired simultaneously from a single laser ablation by dividing the introduced sample volume between different ICP-MS instruments. This method has been utilized for example to acquire trace element data, U–Pb ages (by LAQ-ICP-MS instrument, Laser Ablation Quadrupole Inductively Coupled Plasma Mass Spectrometry), and Hf isotope (LAM-ICP-MS) data from a simultaneous ablation (e.g., Yuan et al., 2008). However promising the *in situ* methodology might be, the applications still suffer from, among other problems, poor precision compared to TIMS. The tradeoffs have to, at least for the moment, be compensated by using the methods in concert to complement the acquired information.

In this study coupled zircon O–Hf isotope datasets for Papers III and IV, and partially coupled zircon Hf isotope-trace element dataset for Paper I were collected. In the Discussion, partially coupled U–Pb and Hf data (LAM-ICP-MS) on selected samples was used.

4. Geological background

A-type granites and massif-type anorthosites are widespread and known from almost all Precambrian cratons (Bonin, 2007; Ashwal, 2010). The temporal occurrence of AMCG complexes and A-type granites is almost as widespread, from Neoarchean (Bonin, 2007) to Miocene (Calzia & Rämö, 2005), but 'true' massif-type anorthosites are restricted to the Proterozoic (Wiebe, 1992; Ashwal, 1993; Ashwal, 2010). Early theorization on massif-type anorthosite petrology also saw suggestions of a special global anorthosite event that took place at approximately 1.5 Ga (e.g., Hertz, 1969; Anderson, 1975) but the large age span (2491 – 530 Ma; Ashwal, 2010) revealed by later geochronologic studies has refuted the existence of such an event (Ashwal, 1993). This study was conducted in two AMCG localities of roughly the same age (1.65-1.5 Ga), but located on different continents nearly on the opposite sides of the world (Fig. 2), in southern Finland, the type locality of A-type rapakivi granite magmatism (Papers I-III) and in northern Brazil (Paper IV).

4.1. The rapakivi suite of southern Finland

The Proterozoic (1.67-1.54 Ga) rapakivi granites of southern Finland (Rämö & Haapala, 2005, and references therein) comprise four major batholiths (Wiborg, Ahvenanmaa, Laitila, and Vehmaa) and a group of smaller intrusions (Suomenniemi, Ahvenisto, Onas, Bodom, Obbnäs, Peipohja, Mynämäki, Eurajoki, Reposaari, Siipyy, Fjälskär and Kökarsfjärden) (Fig. 3), which sharply crosscut the Paleoproterozoic (1.9-1.8 Ga) Svecofennian country rocks of the Arc Complex of southern Finland (Lahtinen et al., 2005). The Finnish province is associated with the Salmi rapakivi intrusion in Russian Karelia (Neymark et al., 1994) and the Baltic rapakivi intrusions of Riga, Märjamaa, and Naissaare (Rämö et al., 1996). Also the rapakivi occurrences in central Sweden (e.g., Andersson, 1997; Persson, 1999; Andersson et al., 2001) are considered to belong to the province.

The Finnish rapakivi granites have most likely been emplaced in an extensional tectonic environment and are regarded as anorogenic (Rämö & Haapala, 2005) or distally orogenic (Åhäll et al., 2000) relative to the contemporaneous Gothian orogenesis. They can be divided into two geographic age-groups, the older, 1.67-1.62 Ga, southeastern and the younger, 1.59-1.54 Ga, southwestern group.

Compositionally the Finnish rapakivi granites span from primitive fayalite-hornblende granite, through hornblende granite, biotite-hornblende granite, and biotite granite to evolved biotite granite and (locally topaz-bearing) alkali-feldspar granite (Lukkari, 2007).



Figure 3. (a) The location and (b) generalized geologic map of the Finnish rapakivi province in the Fennoscandian shield. TIB-Transcandinavian Igneous Belt. Map is adapted from Paper I.

The Finnish rapakivi suite is effectively bimodal (felsic – mafic) with only minor intermediate rock-types (monzodiorites and quartz monzodiorites; e.g., Alviola et al., 1999). The associated mafic rocks are coeval massif-type anorthosites (Ashwal, 1993), which are poorly exposed and mostly inferred from geophysical interpretations (Elo & Korja, 1993). The known occurrences are located in Ahvenisto (Savolahti, 1956,1966; Alviola et al., 1999; Heinonen, 2010b; Paper II), Kolinummi (Norokallio, 2003), and Ylämaa (Arponen et al., 2009; Rämö & Arponen, 2010) areas. There is also a host of diverse hypabyssal rocks associated with the rapakivi granites: diabase dike swarms (e.g., Luttinen et al., 2010) and quartz-feldspar dikes (Rämö, 1991), as well as several localities of hybrid rocks (e.g., Rämö, 1991; Salonsaari, 1995; Alviola et al., 1999; Kosunen, 2004; Arponen et al., 2009). Associated supracrustal rocks are restricted to two localities: the Taalikkala area in eastern Finland (Harju et al., 2010), and the island of Suursaari in the Gulf of Finland, Russia (Rämö et al., 2009, 2010).

Paper I deals with the petrogenesis of the whole Finnish rapakivi suite in general and Papers II and III offer a more detailed look on characteristics of the the prominent AMCG locality in the association, the Ahvenisto complex.

4.2. The Mucajaí AMG complex, northern Amazonian craton

The Mesoproterozoic AMCG magmatism in the Guyanan Shield in northern Amazonian Craton is manifested by a NW-SE trending ~900-km-long rapakivi granite belt (Dall'Agnol et al., 1999). The belt comprises three known localities of rapakivi granites and associated anorthositic rocks: the Parguaza batholith in Venezuela (Gaudette et al., 1978; Mendoza, 1975), the Surucucus suite that straddles the border of Brazil and Venezuela (Dall'Agnol et al., 1975; Pinheiro et al., 1981) and the Mucajaí AMG complex in the northern Brazilian state of Roraima (Paper III, Fraga, 2002; Fraga et al., 2009a). The rapakivi magmatism occurred between 1.55 and 1.52 Ga (Fraga et al., 2009a; Gaudette et al., 1976; Santos et al., 1999, 2003; Paper IV) and is most likely generated by a single within-plate igneous event (Dall'Agnol et al., 1999), termed the Parguaza

event by Gaudette et al. (1978). The rock types in the belt are predominantly wiborgitic and pyterlitic (Rämö & Haapala, 2005) rapakivi granites (Dall'Agnol et al., 1999) and gabbroanorthositic rocks are rare. The best-known locality of massif-type anorthosite, the Repartimento anorthosite, is associated with the Mucajaí complex rapakivi granites (Paper IV, Fraga, 2003; Fraga et al., 2009a).



Figure 4. (a) Geological overview map of the Mucajaí complex and (b) locations of the Mucajaí complex and the other Mesoproterozoic northern Brazilian rapakivi batholiths. Map is adapted from Paper IV.

The Mucajaí AMG complex (Paper IV, Fraga, 2003; Fraga et al., 2009a; Fig. 4) is a ~4000 km² multiphase rapakivi granite – massif-type anorthosite complex that contains pyterlitic rapakivi granites, fayalite-bearing monzonites and syenites, subordinate monzodiorites, and the Repartimento anorthosite. The ~1525 Ma (Paper IV) Mucajaí complex intruded the Paleoproterozoic basement that has been characterized somewhat differently by different authors. The local division of Mucajaí country rocks south of the Cauarane Coeroeni Belt by Fraga et al. (2009b) into the ~1.94 Ga Igarape Branco (allanite-biotite-hornblende granitic gneiss) and Igarape Miracelha (titanite-hornblende-biotite gneiss) units (A-type) and the charnockitic Serra da Prata suite is preferred in this study over regional generalizations (Tassinari & Macambira, 2004; Santos et al., 2006).

4.3. Geochemical overview of the studied AMCG complexes

Overall geochemical patterns in both studied localities, Ahvenisto and Mucajaí, are characteristic for AMCG complexes. The distribution is essentially bimodal (SiO₂<~57 wt.% or SiO₂>~65 wt.%) with only minor intermediate (monzonitic; SiO₂ = 50-65 wt.%) compositions in Mucajaí and none in Ahvenisto (Fig. 5). Plagioclase accumulation, and to some extent modal olivine and pyroxenes, control the whole-rock compositions of the anorthositic rocks and a similar fractionation controlled trend can be observed for granites in both complexes (Papers II and IV; Fig. 5). Monzodioritic rocks sampled from Ahvenisto (Johanson, 1984) define a clear compositional array

spanning SiO₂ values of ~40-55 wt.% (Fig. 5) that has been interpreted as a residual liquid trend after anorthosite fractionation with some cumulus effect from plagioclase accumulation (Paper II). Rudimentary plagioclase-equilibria model-melts calculated after Morse (2006) for plagioclase from the Ahvenisto anorthositic rocks (Johanson, 1984) overlap with the monzodioritic array (Fig. 5), which lends further support to the assumption that the rock groups are consanguineous.



Figure 5. Samples from Ahvenisto and Mucajaí complexes plotted on (a) SiO₂ vs. Al₂O₃ and (b) SiO₂ vs. FeO_{tot} diagrams. Data are compiled from Johanson (1984), Fraga et al. (2009a), and Papers II and IV. Also shown are rudimentary compositions of melts in equilibrium with plagioclase compositions from the anorthositic rocks in Ahvenisto complex calculated after D values of Morse (2006) from equation $C_L = C_s/D$. Plagioclase compositions used in calculations are from Johanson (1984).

Monzodioritic rocks (2 samples) from the Mucajaí complex plot outside the Ahvenisto monzodioritic array at relatively higher Al_2O_3 and lower FeO^{tot} owing most likely to higher abundance of cumulus plagioclase at an earlier stage of fractionation (Paper IV). Normative mineral compositions of the anorthositic rocks (Fig. 6) further illustrate their leucocratic nature.



Figure 6. Anorthositic and monzodioritic rocks from Ahvenisto and Mucajaí plotted on CIPW-based classification diagrams for gabbroic rocks (LeMaitre, 2002). Data are compiled from Johanson (1984), Fraga et al. (2009a), and Papers II and IV.

Major element geochemistry of the studied granites in both localities displays classical A-type characteristics. All studied granites are clearly ferroan (Frost et al., 2001; Fig 7a) but define some-what different alkali enrichment trends (Frost et al., 2001). The Mucajaí granites are alkali-calcic to alkalic but Ahvenisto rocks define a broader variation from calc-alkaline to alkalic values (Fig. 7b) suggesting a more complex fractionation pattern, possibly involving mafic silicates at an earlier stage of crystallization (Frost & Frost, 2011). General trace element patterns define the within-plate and A-type nature of the granites in both complexes (Fig. 7c and 7d). Lower Fe/Mg ratios (Fig. 7e) suggest that the magmatic conditions might have been slightly more oxidizing in Ahvenisto compared to Mucajaí, but all studied granites nevertheless fall broadly to the reduced A-type field of Dall'Agnol & Oliveira (2007).

Trace element discrimination diagrams of Eby (1992) show the granites from both complexes straddling the A1-A2 -type boundary (Fig. 8). Despite overlap, A2-type tendency seems to be more prevalent suggesting crustal influence to the granite chemistry in both localities. This fits well the prevailing hypotheses on the within-plate tectonic setting and dominant crustal source of the granites from both localities (Rämö, 1991; Rämö & Haapala, 2005; Fraga, 2009a). It is, however, recognized that crustal contamination may influence the composition of originally A1-type magmas that are essentially implied to be mantle-derived (Eby, 1992) so that their Y/Nb ratios shift to higher values (>1.2) and plot on the A2 fields on the discrimination diagrams (Eby, 1992; Frost & Frost, 2011). The classification of marginal cases on these diagrams, like Ahvenisto and Mucajaí granites (Fig. 8), may therefore be contestable.



Figure 7. Rapakivi granites of the Ahvenisto and Mucajaí complexes plotted on classification diagrams for A-type granites from (a, b) Frost et al. (2001), (c) Whalen et al. (1987), (d) Pearce et al. (1984), and (e) Dall'Agnol & Oliveira (2007). Data are compiled from Johanson (1984), Fraga et al. (2009a), and Papers II and IV.



Figure 8. Rapakivi granites of the Ahvenisto and Mucajaí complexes plotted on the A1-A2 -type discrimination diagrams of Eby (1992). Data are compiled from Johanson (1984), Fraga et al. (2009a), and Papers II and IV.

5. Summaries of the original papers

5.1. Paper I: Re-evaluation of Rapakivi Petrogenesis: Source Constraints from the Hf isotope Composition of Zircon in the Rapakivi Granites and Associated Mafic Rocks of Southern Finland

The aim of the research was to attempt a first stage application of the *in situ* LAM-ICP-MS method to study the magmatic Lu–Hf isotopes of zircon in the rapakivi granites and associated anorthositic rocks and diabases from the Proterozoic southern Finnish rapakivi association. The working hypothesis postulated that the shorter half-life of ¹⁷⁶Lu to ¹⁷⁶Hf compared to the popular whole-rock Sm–Nd system would provide better compositional resolution to address the source question, and that by studying robust zircon isotope compositions instead of whole-rock samples, possible later alteration and equilibration could also be ruled out. A brief examination of zircon trace elements on two representative samples was also included to support the Hf isotope findings.

According to the measurements, the Finnish rapakivi granites have a chondritic and relatively homogeneous initial zircon Hf isotope composition (~0±4 $\epsilon_{\rm Hf}$ units) that overlaps significantly with the composition of the Paleoproterozoic country-rocks and that the mafic rocks (the anorthositic rocks and a diabase) of the association reveal a clearly more juvenile Hf isotope signal ($\epsilon_{\rm Hf}$ up to +9). In addition to more radiogenic initial Hf isotope values, the mafic rocks also exhibit a wider within-sample range of initial $\epsilon_{\rm Hf}$ values (from ~0 to +9), which was interpreted as an isotope signature from a significantly depleted primary mantle source. The uniform and homogeneous initial Hf isotope composition of the granites was taken as an indication of the lack of mantle component in their source and used to argue for an essentially crustal origin in accordance to the prevailing two-source hypothesis for the association.

5.2. Paper II: Formation and fractionation of High-Al tholeiitic melts in the Ahvenisto complex, southeastern Finland

The second paper is a regional study that summarizes geochronological, geochemical and isotopic data collected from the \sim 1640 Ma Ahvenisto AMCG complex in southeastern Finland and proposes a petrogenetic model for the anorthositic rocks associated with the rapakivi granites of southern Finland.

The aim of the study was to refine the existing ID-TIMS U–Pb geochronology of the Ahvenisto complex (Alviola et al., 1999) and analyze additional heavy mineral fractions from the already dated samples. A new age determination from a previously undated anorthosite enclave was also included in the study. In addition to the mineral U–Pb studies, an extensive whole-rock Nd-Sr-Pb (ID-TIMS) isotopic dataset backed up by elemental geochemical analysis was collected to compare the compositions of different rock types in the complex and to propose petrogenetic considerations and comparisons.

The U–Pb geochronology supports the earlier finding (Alviola et al., 1999) that all the rock types of the complex are coeval. Furthermore the age of the anorthosite enclave (1641 ± 2 Ma) shows that the anorthosite fragments found in the complex are autolithic and do not represent an earlier magmatic phase. All the rock types of the complex, save for the late quartz-feldspar porphyry (1636 ± 2 Ma), are coeval at ~ 1641 ± 2 Ma.

Comagmatic relationship between the anorthositic (ϵ_{Nd} –0.9 to –0.5, Sr_i 0.7037 to 0.7041) and monzodioritic rocks (ϵ_{Nd} –1.1 to –0.2, Sr_i 0.7028 to 0.7040) of the complex was determined from their similar Nd-Sr isotope compositions. The slightly more juvenile Nd isotope composition of the olivine-bearing gabbroic rocks (ϵ_{Nd} +0.2 to +0.4, Sr_i 0.7034 to 0.7035) was taken as evidence of an

open system magmatic evolution of the primary anorthositic magma, most likely caused by crustal contamination.

Supported by major element and REE geochemistry, the isotope results were used to constrain a two-stage petrogenetic model where the primary juvenile magma assimilated lower crust during the first stage before the main phase of anorthosite fractionation. Monzodioritic rocks were deemed to represent residual liquids after anorthosite fractionation. The Ahvenisto rapakivi granites (ϵ_{Nd} –2.1 to –0.1) were, in contrast to the mafic rocks, concluded to represent crystallization products of crustally derived magmas.

5.3. Paper III: Combined *In situ* zircon Hf–O isotope analysis refines the role of depleted mantle in Proterozoic AMCG magmatism

Paper III elaborates the observations of Papers I and II by introducing a coupled zircon O–Hf isotope dataset from the AMCG rocks of the Ahvenisto complex.

A hypothesis was formed on the basis of the heterogeneous initial zircon Hf isotope composition observed in the mafic rocks of the Finnish rapakivi association (Paper I) that the same effect should be observable in the zircon oxygen isotope compositions. A coupled, comprehensive O-Hf (SIMS - LAM-ICP-MS) isotope dataset was collected on the previously U-Pb dated samples (Paper II) of all the major rock types of the Ahvenisto complex.

The zircon O–Hf compositions of the most juvenile rock type of the complex, a leucotroctolite, record a correlating trend that suggest interaction of two source components possibly through an AFC (Assimilation Fractional Crystallization) process acting on a mafic magma derived from the depleted mantle as predicted by the results of Paper I. Homogeneous isotope signatures in the zircons from the anorthositic and monzodioritic rocks show they crystallized only after major assimilation had taken place. The zircon from the Ahvenisto rapakivi granite displays the least radiogenic Hf (low ε_{hf}) and most supracrustal O isotope compositions (high δ^{18} O) suggesting a possible crustal primary source, but the cogenetic relationship with the anorthositic rocks remains a possibility.

5.4. Paper IV: Petrogenesis of the igneous Mucajaí complex, northern Amazonian craton – geochemical, U–Pb geochronological, and Nd–Hf–O isotopic constraints

The fourth paper concentrates on the isotope systematics of the \sim 1525 Ma Mucajaí AMG complex in northern Brazil and compares the observed features to the Ahvenisto complex and Laramie complex in Wyoming, USA.

Zircon U–Pb ages of the major rock types of the Mucajaí complex, including a biotite-hornblende granite, a biotite granite, a monzonite, and an anorthosite were determined by ID-TIMS method. The results imply that the biotite granite is marginally younger (~1520 Ma) than the rocks of rest of the complex that intruded the Paleoproterozoic (~1.94 Ga) country-rocks at ~1526 Ma. This indicates that AMCG magmatism in the area was rather short-lived and lasted at maximum only about 12 million years.

Whole-rock Sm–Nd (ID-TIMS) and zircon Lu–Hf (LAM-ICP-MS) and oxygen (SIMS) isotope compositions were also analyzed to determine the possible source of the Mucajaí complex rocks. Considerable isotopic overlap between the rock types was observed (whole-rock ID–TIMS ϵ_{Nd} from –1.9 to –2.8, zircon ϵ_{Hf} from –2.0 to –3.1, and zircon δ^{18} O from 6.1 to 7.0‰). Therefore no conclusions on discrete source component compositions could be drawn. The influence of Paleoproterozoic crust in the petrogenesis was, however, found to be considerable.

It was concluded that the petrogenetic processes that have acted on the compared three complexes (Mucajaí, Ahvenisto, and Laramie) seem similar, but the lack of comparable rock types preclude their full comparison.

6. Discussion

6.1. Source and tectonic setting of AMCG magmatism

To a first degree, the chemical composition of any igneous rock reflects the source from which it was derived. Tectonic setting, on the other hand, dictates what sources are available and are able to interact with each other during generation and emplacement of the magmas. If the rocks in AMCG associations have been generated in several different tectonic environments (i.e., McLelland et al., 2010), it is therefore also viable to assume that their geochemical and isotopic compositions can reveal differences accordingly (Eby, 1990, 1992).

Compositional bimodality (felsic-mafic) that is characteristic of AMCG complexes (graniteanorthosite) is often quoted as ample evidence of extensional tectonic setting, where mantlederived magmas are able to intrude in and interact with the crust (Bonin, 2004). Felsic and mafic parental magmas are suggested to have formed by partial melting of two different sources, the lower crust and the upper mantle, respectively. Limited interaction between the melts from different sources is reflected in discontinuous elemental geochemical trends leading to the paucity of intermediate rock types in the associations.

Occurrence (though limited) of A-type granites in oceanic settings confirms that their genesis does not necessarily require continental crust at all (Bonin, 2007), but isotopic evidence (e.g., Papers I-IV) suggests that in continental settings, regardless of tectonic environment, crustal influence on the granite chemistry is often considerable. Consanguinity of the felsic and mafic rocks in some Atype associations has, however, been suggested based on Fe-Ti oxide -controlled rapid decline in SiO₂ during crystallization (i.e., Turner & Rushmer, 2010, and references therein) and an upper mantle origin has been invoked for many A-type granite magmas (e.g., Frost & Frost 1997; Frost et al., 2002).

Also several experimental results (e.g., Skjerlie & Johnston, 1993; Patiño Douce, 1997; Bogaerts et al., 2006) suggest it is highly unlikely that partial melting of a (tonalitic to granodioritic) quartzofeldspathic source in the lower crust could alone be responsible for the whole spectrum of observed A-type compositions (peralkaline to peraluminous, alkali to calc-alkalic). Especially high-P environments at lower crustal levels, often quoted as the most important source region of for example the rapakivi granites, are prone to produce magnesian, calc-alkalic, peraluminous partial melts (Frost & Frost, 2011), which are not suitable as the parental magmas of the ferroan, alkalic to alkali-calcic, dominantly metaluminous compositions observed in most of the granites in this study (Papers II and IV; Fraga et al., 2009a). In contrast to these rebuttals, widely quoted experimental results by Wyllie (1977), however, suggest that it is possible to produce potassium-enriched, silica-poor magmas via anhydrous melting of continental crust at high-pressure, which would thus be a viable source for A-type granites.

Based on major element compositions, Frost & Frost (2011) distinguished at least eight different types of A-type granites and quoting the experimental evidence, postulated that most of them are generated by fractionation from mantle-derived basaltic magmas, some of which are potentially influenced by crustal assimilation.

Isotope compositions of the different rock types in the AMCG suites would ideally be able to resolve the source controversy. Discerning definite mantle and crustal isotopic signals from A-type granites and massif-type anorthosites has, however, proved to be a strenuous task (i.e., Ashwal, 1993; Paper I). Strongly overlapping isotope compositions between the different rock groups allow ambiguous conclusions to be drawn and the final answer still eludes the researchers in the field.

The crustal isotopic signatures of the Finnish rapakivi granites have until now been interpreted to originate by straightforward partial melting of a lower crustal Paleoproterozoic source (Rämö, 1991; Rämö & Haapala, 2005 and references therein). Fraga et al. (2009a) adopted a similar model also for the Mucajaí complex and suggested the two-source model postulated by Emslie (1978) and Emslie et al. (1994) to be the best explanation for the genesis of the Repartimento anorthosite. Overlapping isotope compositions, however, still cause ambiguous interpretations of the origin of the rocks (Papers II and IV).

Also the tectonic setting of AMCG magmatism is elusive. Formerly, Proterozoic anorthosite magmatism was thought to be characteristic for extensional or strictly "anorogenic" tectonic environments but recent consensus has shifted towards regional convergent tectonic settings (e.g., Ashwal, 2010). Also the rapakivi granites of southern Finland have been suggested to have formed in a distal orogenic, rather than in extensional, within-plate setting (Åhäll et al., 2000).

Evidence for the post-collisional setting comes mainly from detailed geochronology, which has revealed that the waning stages of orogenic events have been coeval with the onset of AMCG magmatism (Ashwal, 2010; McLelland et al., 2010). This model invokes delamination of lower lithosphere followed by astenospheric upwelling that would be a direct response to collisional stacking during a preceding orogeny (Corrigan & Hanmer, 1997; Ashwal, 2010; McLelland et al., 2010).

For some AMCG associations such a model is harder to reconcile as no coeval orogenic event from the immediate environment is known. These include, for example, the Nain complex in Labrador (e.g., Myers et al., 2008), the Finnish AMCG association (e.g., Rämö & Haapala, 2005; Papers I-III), and the Parguaza event AMCG rocks in northern Brazil and Venezuela (Dall'Agnol et al., 1999; Fraga et al., 2009a; Paper IV). To explain these AMCG occurrences, effects of contemporaneous but distal orogenic events (e.g., Åhäll et al., 2000), reactivation of earlier lithospheric structures (e.g., Myers et al., 2008), delayed heat production through crustal stacking (Kukkonen & Lauri, 2009) or truly "anorogenic" causes, like possible plume activity, have to be considered.

McLelland et al. (2010) proposed a generalized model for AMCG magmatism based on Grenvillian and Labradorian AMCG localities that span almost 600 million years of the Proterozoic. They divided AMCG occurrences in two groups based on their temporal relationship with broadly coeval orogenic events. The post-collisional (most of the Grenvillian) occurrences are a direct result of lithospheric delamination in response to lithospheric stacking during the preceding orogenesis. The second group, that have no immediate orogenic predecessor, are viewed as a result of transtensional reactivation of lithospheric structures (suture zones etc.), which can take place also after a protracted period of time after an orogenic event.

Geochronological results obtained in this study (Papers II and IV) place the Fennoscandian rapakivi association and the northern Brazilian AMCG rocks most definitely into the latter ('intraplate') group of McLelland et al. (2010), as no immediate preceding orogenic event is known from these regions. The main Paleoproterozoic crustal generation event in southern Finland took place during the multi-phase Svecofennian orogeny at 1.92-1.79 Ga (Lahtinen et al., 2005), which was followed by the 1.73-1.55 Ga Gothian orogeny on the southwestern margin of the craton

(Åhäll & Larson, 2000). Åhäll et al. (2000) suggested that the \sim 1.65-1.54 Ga Fennoscandian rapakivi magmatism (Papers I-III) has been a direct distal result of the Gothian orogeny, but this view has not been unequivocally accepted (e.g., Rämö & Haapala, 2005), as the large-scale age pattern within the rapakivi province does not strictly correspond to a southwestward retreating collisional environment.

The tectonic evolution of the northern Amazonian craton is somewhat more obscure and contested but no interpretation of a proximal orogenic event coeval with the 1.55-1.50 Ga Paraguaza rapakivi event (Dall'Agnol et al., 1999) exists in the literature (e.g., Tassinari & Macambira, 2004; Santos et al., 2006; Fraga et al., 2009a; Paper IV). The immediate country rocks of the Mucajaí complex featured in this study (Paper IV) are igneous gneisses and granitoids ~1.94 Ga in age (Fraga et al., 2009b). Therefore, the age difference between the regional orogenic crustal formation event and the following Parguaza AMCG magmatism in northern Amazonia is in the order of ~380-440 million years.

All the rock types in the studied suites are strictly speaking coeval. Almost all of the crystallization ages obtained for granites and associated massif-type anorthosites in the individual complexes (Papers II and IV; Ahvenisto at ~1640 Ma and Mucajaí at ~1525 Ma) are the same within analytical errors. Only a marginal difference between the last granitic phases (1636 Ma quartz feldspar porphyry in Ahvenisto and the 1520 Ma biotite granite in Mucajaí) and the rest of the complexes can be discerned.

Whatever the initial cause of magmatism, most of the models set in both tectonic settings (McLelland et al., 2010) invoke mantle influence, either by lower lithospheric delamination or direct underplating, allowing astenospheric material and heat to access higher lithospheric levels. A combined experimental-modeling approach, however, offers an alternative explanation, that does not require primary influence of mantle material in AMCG petrogenesis. High-P experimental work and equilibrium calculations (Longhi, 2005) suggest that suitable anorthosite parental magma compositions cannot be produced by partial melting of (peridotitic) upper mantle material. Primitive partial melts have problems reaching sufficient plagioclase saturation at high-P fractionation conditions (>11 kbar) that are indicated by the presence of high-Aluminum orthopyroxene megacrysts in many AMCG occurrences (i.e., Ashwal, 1993). Purely crustal origin of AMCG magmatism has been postulated to have taken place via introducing crustal slivers, or "tongues" in to the mantle along low-angle listric faults (e.g., Duchesne et al., 1999) to produce partial melts of mafic lower crustal material parental to the massif-type anorthosites. Applying this model to Ahvenisto is problematic, since the most primitive rock types of the complex record isotopic signatures indicative of a depleted mantle source (Paper III). In principle, the isotope results from Mucajaí are permissible of sole crustal origin of the AMCG magmatism, but the mafic rock types that are the key to resolve the mantle source (as in Ahvenisto; Paper III) have not been detected and sampled there.

Zircon Hf and O signatures reported in the present study, however, establish the depleted mantle origin of the massif-type anorthosites in Finland (Papers I and III) that has already been suggested by previous Nd isotope studies (Rämö, 1991; Paper II) but at the same time pose the question as to whether the prevailing crustal source model (e.g., Rämö, 1991; Rämö & Haapala, 2005; Paper II) is plausible for the Finnish rapakivi granites. Significant overlap in the isotope compositions of the Mucajaí rocks does not warrant an unambiguous answer to the anorthosite source question but clearly shows that a major Paleoproterozoic crustal source was involved in the petrogenesis of all the rocks in the Mucajaí complex (Paper IV). Considering the similar geotectonic setting (Fraga et al., 2009a), geochemistry (Paper IV), and isotope compositions of the respective rock types in Mucajaí and Ahvenisto (Papers II and IV), similar petrogenetic processes (partial melting of two

discrete sources and contamination of a mantle-derived magma by crustal material) may have been active in both complexes.

6.2. Comparison of Hf and Nd isotopes in the studied AMCG rocks

Routine whole-rock Nd isotope studies conducted in the Finnish rapakivi terrane (e.g., Rämö, 1991) have given systematic indications of the influence of more juvenile material in the genesis of the anorthositic rocks than in the related rapakivi granites. Theoretically, Lu-Hf and Sm-Nd systems behave in an analogous manner in partial melting and in igneous processes, which should lead to comparative initial epsilon values in the resulting igneous rocks. Relatively more incompatible nature of Hf as compared to Nd, however, should lead to stronger depletion of mantle source reservoirs. Coupled with higher rate of decay of ¹⁷⁶Lu compared to ¹⁴⁷Sm, a higher contrast in the isotopic composition of Hf than in Nd in source reservoirs should ensue. Resulting initial epsilon values have generally been estimated to roughly follow the formula (e.g., Patchett et al., 1981):

(3)
$$\varepsilon^{t}_{Hf} = 2\varepsilon^{t}_{Nd}$$

where t denotes time of crystallization.

Based on a large amount of combined whole-rock Nd-Hf data, Veervort et al. (1999) found that majority of terrestrial samples follow remarkably closely the relationship:

(4)
$$\epsilon^{t}_{Hf} = 1.36\epsilon^{t}_{Nd} + 2.95$$

and that local deviations from this array most likely reflect fractionation of the Hf and Nd isotopic systems by some petrogenetic processes.

Basically, the Hf isotope results obtained by *in situ* methods should therefore provide the same results as the whole-rock Nd isotope studies but with better theoretical resolution, which has been one of the motivations to utilize the method in this study to detect possible differences in rocks that have overlapping Nd isotope compositions and are potentially derived from different sources with similar isotopic compositions.



Figure 9. Comparison of zircon $\epsilon_{\rm Hf}$ (LAMS) and whole-rock $\epsilon_{\rm Hd}$ (TIMS) values obtained for the studied AMCG rocks. Linear fit to all data points, the "terrestrial array" ($\epsilon^t_{\rm Hf}$ = 1.36 $\epsilon^t_{\rm Nd}$ +2.95) from Veervort et al. (1999), and a 1:1 line are also shown for comparison. Error bars denote 2SD. Estimated uncertainty in $\epsilon_{\rm Nd}$ values is 0.4 units. Data is from Papers I-IV.

In Fig. 9 the results from whole-rock Nd ID-TIMS analyses (Papers II and IV and references therein) are plotted against *in situ* zircon Hf LAMS analyses (Papers I, III, and IV) to show the relation-

ship of these two isotopic systems in the studied rocks. A linear fit for un-weighted means for all the samples in both systems produces a robust relation of:

(5)
$$\varepsilon^{t}_{Hf} = 2\varepsilon^{t}_{Nd} + 3.4$$

The predicted average initial ϵ_{Hf} values within this sample set can then be calculated from this formula by inserting corresponding ϵ_{Nd} values. Fig. 10 illustrates the deviation of the measured ϵ_{Hf} values from these predicted values. Majority of the anorthositic rocks deviate to relatively more positive ϵ_{Hf} values and the rocks of the Mucajaí complex, regardless of rock-type, deviate to unexpectedly low ϵ_{Hf} even lower than corresponding ϵ_{Nd} values.



Figure 10. Comparison of measured zircon ε_{Hf} (LAMS) and whole-rock ε_{Nd} (ID-TIMS) values to ε_{Hf} values calculated from formula $\varepsilon_{Hf}^{t} = 2\varepsilon_{Nd}^{t} + 3.4$. Error bars on measured ε_{Hf} values denote within-sample 2SD. All samples in Papers I-IV with both Hf and Nd data available are plotted. Sample numbers are listed on the x-axis, granitic rocks in white and anorthositic and other mafic rocks in grey boxes. Brasilian samples are marked with Br. All other samples are from southern Finland. See Papers I-IV for sample descriptions and original data.

Both of these deviations may be caused by the stronger resolution of the Lu-Hf system relative to Sm-Nd. In case of the Finnish anorthositic rocks from the predicted values illustrates the same point that has been made from the absolute ϵ_{Hf} values (Paper I). Stronger depletion of the upper mantle in Hf relative to Nd causes greater differences in the absolute ϵ_{Hf} values of rocks derived from different source regions. In northern Brazil, the deviation to lower values may reflect the composition of the relatively older crustal source component than in southern Finland.

However, two problems arise from this approach. Firstly, if the within-sample variation of Hf isotopes in the anorthositic rocks is caused by analytical problems, the precision of the measurements may be too poor to make any justified conclusions based on the data. Secondly, even if the variation is real, it is admittedly contestable to propose that such a variation could be seen in single samples, which supposedly crystallize zircon at a quite late stage in their magmatic evolution.

The analytical problems of LAM-ICP-MS zircon Hf method have been considered thoroughly in this case and the exhaustive method descriptions are given in Paper I, but the mechanism of how such

isotopically heterogeneous zircon populations are introduced into massif-type anorthositic rocks, however, still remains an open question and possibly opens a future avenue of research. A further consideration of age-determination related errors in the issue of heterogeneity, is given in the following section.

6.3. Hf isotope heterogeneity of zircon populations, true or false?

Geological interpretation of heterogeneity in initial zircon Hf isotope composition of igneous rocks implied by several studies (e.g., Papers I and III) has inherent problems relating to possible inheritance of zircon crystals from older sources. It is possible that without strict age control on single spot Hf isotope analyses heterogeneity may be interpreted where none actually exists. Heterogeneity of zircon Hf isotope compositions in the anorthositic rocks of the Finnish rapakivi suite is integral for the interpretation of the depleted mantle source of massif-type anorthosites (Papers I and III) and therefore requires rigorous verification.

Ideally U–Pb and Hf isotope data should be collected from the same laser ablation with two ICP-MS settings enabling strict analysis of the same volume for both systems. This option has not, however, been available in the present study. Instead, to carry out a rudimentary test of temporal homogeneity of the studied zircon populations, preliminary U–Pb (LAM-ICP-MS) data (Appendix A.) partially coupled to Hf isotope data from the same grain domains were collected from the Ahvenisto leucotroctolite (A1271). In addition, uncoupled zircon SIMS U–Pb data from the Ahvenisto anorthosite (A1933; Appendix A.) were also assessed for possible inheritance effects.



Figure 11. The Hf–O compositions of zircons from the main rock types of the Ahvenisto complex show variation in the leucotroctolite sample (A1271) interpreted as a depleted mantle source signal (Paper III). Figure is adapted from Paper III.

Zircon from the leucotroctolite A1271 displays correlating initial Hf and O isotope signals (Fig. 11), which were used to sketch a possible model for the interaction of a mantle-derived parental melt of the Finnish massif-type anorthosites with the continental crust (Paper III). Due to the lack of coupled U–Pb data at the time, the initial Hf isotope compositions were calculated assuming homogeneous crystallization ages for all the grains at the U–Pb TIMS age (1636 Ma) of the leucotroctolite (Paper II). At the implied crystallization age, the range in initial Hf isotope values is considerable (from ~0 to +8 ϵ_{Hf} units). Thirteen crystals and crystal domains were analyzed by LAM-ICP-MS first for U–Pb and then for Hf isotopes creating a coupled (²⁰⁷Pb/²⁰⁶Pb)-Hf dataset. Eight of the analyzed points proved nearly concordant (Fig. 12a) and nine have ²⁰⁷Pb/²⁰⁶Pb ages clustering around 1654 Ma (Fig. 12a). Three grains have apparent Paleoproterozoic ²⁰⁷Pb/²⁰⁶Pb ages (1870, 1806, and 1723 Ma) and one yields a clearly younger ²⁰⁷Pb/²⁰⁶Pb age of 1495 Ma. At

face value, the ages provided by these deviant points could be taken as evidence that heterogeneity of the zircon population may have contributed to the observed Hf isotope variation (Papers I and III). However, as none of these ages are provided by concordant analyses and their ²⁰⁶Pb/²⁰⁴Pb ratios are relatively low, the apparent discordance of the grains is most likely due to analytical errors. Therefore, considering these preliminary U-Pb ages, the zircon population in A1271 most likely is homogeneous.



Figure 12. In situ zircon U–Pb data for (a) the Ahvenisto leucotroctolite (A1271; LAM-ICP-MS) and (b) anorthosite (A1933; SIMS)

If the age of some of the zircons is, however, hypothetically assumed to be higher (i.e., grains might be inherited), and the earlier reported values calculated at \sim 1640 Ma (Paper IV) are considered in

effect to be without geological meaning, their recalculated $\epsilon_{\rm Hf}$ values would still plot near the mean of the main cluster (Fig. 13) and do not considerably increase the observed Hf isotope range at 1640 Ma. This naturally does not conclusively prove that all possibly inherited grains in the original Hf isotope variation behave in a similar manner and some of the higher $\epsilon_{\rm Hf}$ values could still be caused by inherited grains. The correlating behavior of O and Hf isotopes in the zircon population (Paper III) provides independent indication, however, of possible magmatic variation caused by contamination of a mantle-derived magma with crustal material.



Figure 13. Coupled U–Pb – Hf data for the Ahvenisto leucotroctolite (A1271). ²⁰⁷Pb/²⁰⁶Pb ages and initial Hf isotope compositions of individual grains are plotted with 2σ error estimates against CHUR and depleted mantle evolution curves. Dashed gray lines connect the "outlier" coupled data points to the initial Hf isotope values recalculated to 1635 Ma showing that calculation of Hf isotope values from the possibly older grains at ~1640 Ma does not produce additional variation in the Hf isotope compositions.

Ten grains from the Ahvenisto anorthosite (A1933) analyzed for U–Pb by SIMS are coherently near-concordant and the mean ${}^{207}Pb/{}^{206}Pb$ age coincides with the TIMS age (1640 ± 2 Ma; Paper II) at 1640 ± 5 Ma (Fig. 12b) which proves that the zircon population in the sample is temporally homogeneous. No Hf isotope heterogeneity is, therefore, implied by the *in situ* U–Pb data in sample A1933.

6.4. Mesoproterozoic AMCG magmatism in the context of supercontinent Columbia

In several recent Proterozoic supercontinent reconstructions the East European and Amazonian cratons have been proposed to have formed a continuous crustal aggregate in the hypothetical Paleo- to Mesoproterozoic supercontinent Columbia (e.g., Zhao et al., 2002; Vigneresse, 2005; Cordani et al., 2009; Johansson, 2009). Continuous Paleo- to Neoproterozoic accretionary orogens extending from Fennoscandia to southeastern Amazonia (1.9-1.8 Ga Svecofennian/Ventuari – Tapajos, 1.7-1.5 Ga Gothian/Rio Negro Juruena) have been outlined to support these reconstructions (e.g., Zhao et al., 2002; Johansson, 2009). Paleomagnetic studies also place Amazonia and Baltica on similar paleolatitudes at 1.6 to 1.5 Ga (Mertanen & Pesonen, 2005). These suggestions are congruent with the prevailing Neoproterozoic Rodinian reconstructions and further supported by the subsequent 1.65 to 1.5 Ga AMCG magmatism evident in the 1.95 to 1.8 Ga Svecofennian and

Ventuari-Tapajos domains (Dall'Agnol, 1999; Rämö & Haapala, 2005; Papers I-IV). Geochemical evidence (Fraga et al., 2009a), geochronology, and the Nd isotopic similarities of the AMCG rock types of Mucajaí and Fennoscandia reported in this study (Papers II and IV), fit the hypothesis that the Paleoproterozoic crustal domains hosting these rocks may have been in close proximity during Paleo- to Mesoproterozoic times.

Fig. 14 shows a tentative reconstruction of the relative positions of Baltica and Amazonia within the supercontinent Columbia as proposed by Johansson (2009). The geographic distribution of 1.67–1.47 Ga AMCG magmatism in both cratons would define a ~2000-2500 km long magmatic belt parallel to the proposed accretionary fronts in the present west-southwest. The similarities and concurrence of the Fennoscandian and Amazonian AMCG suites alone naturally cannot prove the Proterozoic proximity of the regions but nevertheless suggest that the assemblage of Johansson (2009) may be more likely than the original Columbian reconstruction of Rogers & Santosh (2002) or other tectonic models that place the Amazonian craton in a closer proximity to the present south-western margin of North America (Mesoproterozoic Laurentia, e.g., Santos et al., 2008).



Figure 14. A schematic chronotectonic reconstruction of a possible Amazonia-Baltica continental assembly (SAMBA, South America Baltica; Johansson, 2009) in the Proterozoic supercontinent Columbia showing the rapakivi suites of Fennoscandia and northern Brazil. The figure is modified after Johansson (2009).

7. Concluding remarks

Different components (i.e., granites and anorthosites) of intraplate AMCG magmatism are strictly speaking coeval and often intrude simultaneously at similar crustal levels. In the studied AMCG complexes (Ahvenisto and Mucajaí) the anorthositic magmatism is essentially of the same age as the main-phase granitic magmatism. In both cases the most evolved magmatic phases (the biotite granite in Mucajaí and late granitic porphyries in Ahvenisto) are marginally younger than the other rock types. Considering the uncertainties in the zircon U–Pb ages, the duration of AMCG magmatism within individual complexes was fairly limited (10-15 m.y.) and it can be concluded that all rock types were emplaced during a single event.

Conventional isotopic methods that utilize bulk samples do not necessarily see the limited isotopic contrast between the possible source end-members of AMCG complexes. *In situ* methods (LAM-ICP-MS, SIMS), in conjunction with stable high-temperature isotope tracers (O isotopes), are more suited to probe the subtle variations in magmatic isotopic compositions recorded by refractory mineral phases like zircon. Coupled Hf–O isotope studies show internal variation within AMCG suite samples that can be used (to certain limits) to decipher primary source components unseen by conventional bulk methods.

The set hypotheses tested out in the studied localities as follows:

Hypothesis 1: Massif-type anorthosites and rapakivi granites are genetically linked but derived from discrete sources – Pass/Fail.

The massif-type anorthosites and granites in the studied localities are most definitely linked, both temporally and spatially, very closely. Isotopically discrete source components contributed at least in the Ahvenisto complex but the primary source of the rapakivi granites remains an open question.

Hypothesis 2: Mantle is the primary source of massif-type anorthosites - Pass.

Hf–O compositions of zircon from AMCG rocks of southern Finland have shown that the depleted mantle is the most likely primary source of massif-type anorthosite in the within-plate AMCG suite of southeastern Fennoscandia. Despite significant petrological similarities, extrapolating the model to northern Brazil or occurrences elsewhere is, however, not permissible based on the data presented here alone.

Hypothesis 3: Lower crust is the primary source of the A-type granites - Pass/Fail.

The source of the rapakivi and A-type granites remains contestable. Straightforward interpretation of minute Hf isotope variations and chondritic average compositions in the rapakivi granites of southern Finland could be used to argue for a homogeneous crustal source. However, coupled zircon Hf–O signatures cannot rule out mantle derivation with considerable crustal contamination as the mode of petrogenesis for either of the studied complexes.

Hypothesis 4: The mantle component was variably depleted - Pass.

A depleted mantle component is indicated by the initial Hf isotope compositions of the most primitive, olivine-bearing anorthositic rocks in Fennoscandia. Similarly depleted values (nor olivinebearing units) were not observed in Mucajaí.

Hypothesis 5: Mafic and felsic magmas have interacted during their accent and emplacement in an open system – Pass.

The open system evolution of a mantle-derived primary melt is demonstrated by almost all isotopic evidence presented in the study. The isotopic crustal overprint on the mafic magmas is considerable and also the main reason for the crustal Nd and Pb signatures of the bulk of the anorthositic rocks. Zircon Hf–O systematic of the most juvenile anorthositic rocks in Ahvenisto suggests that the process was possibly one of combined assimilation-fractional crystallization.

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Spot		Ages		Measured			lsot	ope ratios					
(n=13)	²⁰⁷ Pb/ ²⁰⁶ Pb (1σ)	²⁰⁷ Pb/ ²³⁵ U (1σ)	²⁰⁶ Pb/ ²³⁸ U (1σ)	²⁰⁶ Pb/ ²⁰⁴ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb	1SE	²⁰⁷ Pb/ ²³⁵ U	1SE	²⁰⁶ Pb/ ²³⁸ U	1 S	Е	ho E _{Hf}	(ini)
01a	1656 ± 13	1635 ± 14	1618 ± 22	3278	0.10174	± 0.00075	4.00518	± 0.0712	2 0.28524	700 ⁰ Ŧ	445 0.	52 3	8.
02a	1643 ± 13	1630 ± 14	1619 ± 22	9168	0.10101	± 0.00072	3.98109	± 0.0702	5 0.28551	∓ 0.00	448 0.	62 3	<u>.</u>
02b	1655 ± 13	1633 ± 12	1615 ± 21	10045	0.10166	± 0.00071	3.99527	± 0.0602	2 0.28478	+ 0.00	420 0.	72 3	4.
02c	1670 ± 12	1640 ± 16	1616 ± 24	2442	0.10250	± 0.00070	4.02966	± 0.0783	3 0.28488	± 0.00	469 0.	73 4	ن.
03a	1661 ± 16	1513 ± 14	1409 ± 20	398	0.10202	± 0.00086	3.43742	± 0.06290	0.24436	± 0.00	380 0.	73 3	0.
03b	1495 ± 20	1455 ± 16	1426 ± 20	155	0.09335	± 0.00102	3.19026	± 0.0647	l 0.24753	± 0.00	394 0.	63	2.4
03c	1806 ± 15	1667 ± 17	1559 ± 23	1206	0.11039	± 0.00099	4.16555	± 0.0846!	5 0.27367	± 0.00	448 0.	50 8	.1
04a	1651 ± 14	1643 ± 15	1634 ± 23	8370	0.10147	± 0.00081	4.04291	± 0.0734;	2 0.28847	+ 0.00	462 0.	34 3	2.7
04b	1641 ± 16	1626 ± 14	1613 ± 22	658	0.10090	± 0.00091	3.96107	± 0.0665	3 0.28435	+ 0.00	438 0.	50 4	.2
04c	1636 ± 12	1666 ± 13	1687 ± 23	30620	0.10063	± 0.00067	4.15729	± 0.0646	2 0.29906	± 0.00	462 0.	84 4	9.1
13a	1679 ± 15	1636 ± 15	1600 ± 23	1805	0.10299	± 0.00092	4.00745	± 0.0718	5 0.28167	± 0.00	466 0.	61 3	5.
14b	1870 ± 18	1727 ± 17	1610 ± 24	698	0.11438	± 0.00118	4.47808	± 0.0913	1 0.28379	± 0.00	485 0.	38 6	5.5
17d	1723 ± 15	1656 ± 18	1603 ± 26	3403	0.10551	± 0.00085	4.10885	± 0.0922	3 0.28236	+ 0.00	523 0.	70 6	.2

Appendix A.1 LAM-ICP-MS U-Pb data and corresponding initial $\epsilon_{\rm H}$ values for A1271 Suolampi leucotroctolite.

Spot			Mea	sured isotop	e ra	itios				Ē	itial	isotope rati	os*		
(n=13)	¹⁷⁶ Hf/ ¹⁷⁷ Hf		1SE	¹⁷⁶ Lu/ ¹⁷⁷ Hf		1SE	¹⁷⁶ Υb/ ¹⁷⁷ Hf		1SE	¹⁷⁶ Hf/ ¹⁷⁷ Hf		2SE	ε _{Hf} (t)		2SE
01a	0.281870	+1	0.000014	0.001052	+1	0.000007	0.0639	+1	0.0012	0.281837	+I	0.000028	3.8	+1	1.0
02a	0.281891	+I	0.000010	0.001453	+1	0.000004	0.0941	+1	0.0012	0.281846	+1	0.000020	3.8	+1	0.7
02b	0.281873	+I	0.000010	0.001473	+I	0.000005	0.0938	+1	0.0010	0.281827	+I	0.000020	3.4	+I	0.7
02c	0.281893	+I	0.000014	0.001421	+1	0.000004	0.0908	+1	0.0007	0.281848	+I	0.000028	4.5	+1	1.0
03a	0.281829	+I	0.000015	0.000601	+1	0.000010	0.0350	+1	0.0003	0.281810	+1	0.000030	3.0	+1	1.0
03b	0.281787	+I	0.000013	0.000697	+1	0.000021	0.0420	+1	0.0010	0.281767	+1	0.000026	-2.4	+1	0.9
03c	0.281906	+I	0.000013	0.001331	+I	0.000008	0.0808	+1	0.0004	0.281860	+I	0.000026	8.1	+I	0.9
04a	0.281868	+I	0.000015	0.000987	+I	0.000035	0.0641	+1	0.0028	0.281837	+I	0.000030	3.7	+I	1.0
04b	0.281893	+I	0.000015	0.001151	+I	0.000024	0.0804	+1	0.0015	0.281857	+I	0.000030	4.2	+I	1.0
04c	0.281902	+1	0.000012	0.000957	+1	0.000035	0.0604	+1	0.0019	0.281872	+I	0.000024	4.6	+I	0.8
13a	0.281827	+I	0.000016	0.000460	+I	0.000005	0.0273	+1	0.0001	0.281812	+I	0.000032	3.5	+I	1.1
14b	0.281805	+1	0.000016	0.000892	+1	0.000037	0.0532	+1	0.0023	0.281773	+I	0.000032	6.5	+I	1.0
17d	0.281912	+1	0.000015	0.001590	+1	0.000000	0.0952	+1	0.0015	0.281860	+1	0.000030	6.2	+1	1.0

Appendix A.2. U-Pb coupled LAM-ICP-MS zircon Lu-Hf data for sample A1271 Suolampi leucotroctolite.

*Initial isotope ratios calculated at corresponding ²⁰⁷Pb/²⁰⁶Pb ages given in Appendix A.1.

(n=10) 207 pb/ 206 pb 207 pb/ 206 pb 10 pb/ 206 pb 10 pb/ 206 pb 10 pb/ 205 pb/ 238 U 10 pb/ 238 U	Spot		Ages		Measured				Isot	ope rat	tios				
01a 1642 8 1615 9 1594 14 60128 0.100991 ± 0.459021 3.90579 ± 1.09859 0.280495 ± 0.99805 02a 1642 ± 10 1626 9 1614 ± 14 51311 0.100956 ± 0.532577 3.95945 ± 1.09160 0.285337 ± 0.99561 03a 1647 ± 8 1618 ± 14 51649 0.101239 ± 0.447596 3.98298 ± 1.09160 0.285337 ± 0.99561 04a 1644 ± 9 16641 0.1001084 ± 0.407109 ± 1.01925 0.286805 ± 0.99561 04b 1634 ± 1640 1.638 ± 1647 ± 0.447596 3.99735 ± 1.099561 ± 0.99561 04b 1637 ± 165641 0.100705 ± 0.233544 4.01	(n=10)	²⁰⁷ Pb/ ²⁰⁶ Pb (1σ)	²⁰⁷ Pb/ ²³⁵ U (1σ)	²⁰⁶ Pb/ ²³⁸ U (1σ)	²⁰⁶ Pb/ ²⁰⁴ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb		1σ %	²⁰⁷ Pb/ ²³⁵ U	1	a %	²⁰⁶ Pb/ ²³⁸ U		1σ %	Rho
02a 1642 10 1626 9 1614 14 51311 0.100956 ± 0.532577 3.95945 ± 1.12629 0.284448 ± 0.99561 03a 1647 8 1631 9 1618 14 51649 0.101239 ± 0.447596 3.95938 ± 1.09160 0.285337 ± 0.99561 04a 1644 4 9 1626 ± 14 51649 0.1001084 ± 0.488667 3.99735 ± 1.09160 0.286805 ± 0.99561 04b 1637 ± 9 1626 ± 14 1600705 ± 0.233544 4.01109 ± 1.01889 0.288876 ± 0.99517 05a 1640 ± 1634 ± 165041 0.100705 ± 0.237555 ± 0.99117 ± 1.01889 0.28876 ± 0.99117 05a 1640 ± 1634 1 142004 0.100775 ± 0.277655 3.99912 ± 1.02733 0.28	01a	1642 ± 8	1615 ± 9	1594 ± 14	60128	0.100991	+1	0.459021	3.90579	+ 1.0	9859	0.280495	+	0.998097	0.91
03a 1647 ± 8 1631 ± 9 1618 ± 14 51649 0.101239 ± 0.447596 3.98298 ± 1.09160 0.285337 ± 0.99561 04a 1644 ± 9 1634 ± 9 1626 ± 14 6940 0.101084 ± 0.488667 3.99735 ± 1.09160 0.286805 ± 0.9581 04b 1637 ± 4 1636 ± 8 1636 ± 14 169641 0.100705 ± 0.233544 4.01109 ± 1.01889 0.288876 ± 0.99176 05a 1640 ± 5 1634 ± 8 1629 ± 14 142004 0.100872 ± 0.237555 ± 1.02933 0.287536 ± 0.99117 05a 1631 ± 12 1612 ± 10 1597 ± 14 142004 0.100872 ± 0.277655 3.99912 ± 1.02933 0.287536 ± 0.99117 06a 1631 ± 12 1612 ± 10 1597 ± 14 20473 0.100379 ± 0.646293 3.89022 ± 1.05078 0.299117 07a 1634 ± 6 1622 ± 14 106613 0.100557 ± <t< td=""><td>02a</td><td>1642 ± 10</td><td>1626 ± 9</td><td>1614 ± 14</td><td>51311</td><td>0.100956</td><td>+1</td><td>0.532577</td><td>3.95945</td><td>+ 1.1</td><td>12629</td><td>0.284448</td><td>+1</td><td>0.992412</td><td>0.88</td></t<>	02a	1642 ± 10	1626 ± 9	1614 ± 14	51311	0.100956	+1	0.532577	3.95945	+ 1.1	12629	0.284448	+1	0.992412	0.88
04a 1644 ± 9 1634 ± 9 1626 ± 14 6940 0.101084 ± 0.488667 3:99735 ± 1.10925 0.288876 ± 0.99581 04b 1637 ± 4 1636 ± 8 1636 ± 14 169641 0.100705 ± 0.233544 4.01109 ± 1.01889 0.288876 ± 0.99176 05a 1640 ± 5 1634 ± 8 1629 ± 14 142004 0.100872 ± 0.277655 3:99912 ± 1.02933 0.287536 ± 0.99117 06a 1631 ± 12 1612 ± 10 1597 ± 14 20473 0.100379 ± 0.646293 3:89022 ± 1.18825 0.281080 ± 0.99712 07a 1634 ± 6 1627 ± 9 16613 0.100357 ± 0.344280 3:96691 ± 1.05078 0.286115 ± 0.99277	03a	1647 ± 8	1631 ± 9	1618 ± 14	51649	0.101239	+1	0.447596	3.98298	± 1.0	<u>9160</u>	0.285337	+1	0.995618	0.91
04b 1637 ± 4 1636 ± 8 1636 ± 14 169641 0.100705 ± 0.233544 4.01109 ± 1.01889 0.288876 ± 0.99176 05a 1640 ± 5 1634 ± 8 1629 ± 14 142004 0.100872 ± 0.277655 3.99912 ± 1.02933 0.287536 ± 0.99176 06a 1631 ± 12 1612 ± 10 1597 ± 14 20473 0.100379 ± 0.646293 3.899022 ± 1.18825 0.281080 ± 0.99717 07a 1634 ± 6 1627 ± 9 1652 ± 14 106613 0.100357 ± 0.344280 3.96691 ± 1.05078 0.286115 ± 0.99727	04a	1644 ± 9	1634 ± 9	1626 ± 14	6940	0.101084	+1	0.488667	3.99735	+ 1.1	10925	0.286805	+1	0.995811	0.90
05a 1640 ± 5 1634 ± 8 1629 ± 14 142004 0.100872 ± 0.277655 3.99912 ± 1.02933 0.287536 ± 0.99117 06a 1631 ± 12 1612 ± 10 1597 ± 14 20473 0.100379 ± 0.646293 3.89022 ± 1.18825 0.281080 ± 0.99712 07a 1634 ± 6 1627 ± 9 1622 ± 14 106613 0.100557 ± 0.344280 3.96691 ± 1.05078 0.286115 ± 0.99277	04b	1637 ± 4	1636 ± 8	1636 ± 14	169641	0.100705	+1	0.233544	4.01109	± 1.0	11889	0.288876	+1	0.991761	0.97
06a 1631 ± 12 1612 ± 10 1597 ± 14 20473 0.100379 ± 0.646293 3.89022 ± 1.18825 0.281080 ± 0.99712 07a 1634 ± 6 1627 ± 9 1622 ± 14 106613 0.100557 ± 0.344280 3.96691 ± 1.05078 0.286115 ± 0.99277	05a	1640 ± 5	1634 ± 8	1629 ± 14	142004	0.100872	+1	0.277655	3.99912	± 1.0	12933	0.287536	+1	0.991170	0.96
07a 1634 ± 6 1627 ± 9 1622 ± 14 106613 0.100557 ± 0.344280 3.96691 ± 1.05078 0.286115 ± 0.99277	06a	1631 ± 12	1612 ± 10	1597 ± 14	20473	0.100379	+1	0.646293	3.89022	± 1.1	18825	0.281080	+1	0.997121	0.84
	07a	1634 ± 6	1627 ± 9	1622 ± 14	106613	0.100557	+1	0.344280	3.96691	± 1.0	05078	0.286115	+1	0.992775	0.94
08a 1645 ± 6 1628 ± 9 1614 ± 14 132402 0.101112 ± 0.344765 3.96751 ± 1.05083 0.284587 ± 0.99266	08a	1645 ± 6	1628 ± 9	1614 ± 14	132402	0.101112	+1	0.344765	3.96751	± 1.0)5083	0.284587	+I	0.992665	0.94
09a 1643 ± 16 1623 ± 11 1608 ± 14 25990 0.101007 ± 0.840616 3.94595 ± 1.30472 0.283335 ± 0.99782	09a	1643 ± 16	1623 ± 11	1608 ± 14	25990	0.101007	+1	0.840616	3.94595	+ 1.5	30472	0.283335	+1	0.997826	0.76

anorthosite.
Pitkäjärvi
for A1933
U-Pb data
A.3. SIMS
pendix