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Genesis and Emplacement of Carbonatites and Lamprophyres in the Svecofennian Domain

by

Jeremy Woodard

TURUN YLIOPISTO
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From the Department of Geology
University of Turku
FI-20014 Turku
Finland

Supervised by

Prof. Olav Eklund
Department of Geology
University of Turku
Turku, Finland

Dr. Hugh E. O'Brien
Finland Isotope Geology Laboratory (SIGL)
Geological Survey of Finland
Espoo, Finland

Reviewed by

Prof. Tom Andersen
Department of Geosciences
University of Oslo
Oslo, Norway

Dr. Axel Gerdes
Institut für Geowissenschaften
Johann-Wolfgang-Goethe University
Frankfurt am Main, Germany

Opponent

Prof. Dmitry Konopelko
Geology Faculty
St. Petersburg State University
St. Petersburg, Russian Federation

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ABSTRACT

A small carbonatite dyke swarm has been identified at Naantali, southwest Finland. Several swarms of shoshonitic lamprophyres are also known along the Archean-Proterozoic boundary in eastern Finland and northwest Russia. These intrusions, along with the carbonatite intrusion at Halpanen, eastern Finland, represent a stage of widespread low-volume mantle-sourced alkaline magmatism in the Svecofennian Domain. Using trace element and isotope geochemistry coupled with precise geochronology from these rocks, a model is presented for the Proterozoic metasomatic evolution of the Fennoscandian subcontinental lithospheric mantle. At ~2.2-2.06 Ga, increased biological production in shallow seas linked to continental rifting, resulted in increased burial rates of organic carbon. Subduction between ~1.93-1.88 Ga returned organic carbon-enriched sediments of mixed Archean and Proterozoic provenance to the mantle. Dehydration reactions supplied water to the mantle wedge, driving arc volcanism, while mica, amphibole and carbonate were brought deeper into the mantle with the subducting slab. The cold subducted slab was heated conductively from the surrounding warm mantle, while pressures continued to gradually increase as a result of crustal thickening. The sediments began to melt in a two stage process, first producing a hydrous alkaline silicate melt, which infiltrated the mantle wedge and crystallised as metasomatic veins. At higher temperatures, carbonatite melt was produced, which preferentially infiltrated the pre-existing metasomatic vein network. At the onset of post-collisional extension, deep fault structures formed, providing conduits for mantle melts to reach the upper crust. Low-volume partial melting of the enriched mantle at depths of at least 110 km led to the formation of first carbonatitic magma and subsequently lamprophyric magma. Carbonatite was emplaced in the upper crust at Naantali at 1795.7 ± 6.8 Ma; lamprophyres along the Archean-Proterozoic boundary were emplaced between 1790.1 ± 3.3 Ma and 1781 ± 20 Ma.

Keywords: carbonatite; deep carbon cycle; geochronology; lamprophyre; mantle metasomatism; shoshonitic; zircon

TABLE OF CONTENTS

ORIGINAL PUBLICATIONS

PREFACE

This thesis consists of a synopsis and four original publications, referred to with respect to their Roman numerals in the text, as designated below:

Paper I: **Woodard, J.** and Hölttä, P., 2005. The Naantali alvikite vein-dykes: a new carbonatite in southwestern Finland. *Geological Survey of Finland, Special Paper 38*, 5-10.

Paper II: **Woodard, J.** and Hetherington, C.J. Timing and conditions of carbonatite emplacement at Naantali, SW Finland. *Manuscript submitted to Precambrian Research*.

Paper III: **Woodard, J.**, Kietäväinen, R., Eklund, O. and Shebanov, A. Svecofennian post-collisional shoshonitic lamprophyres at the margin of the Karelia Craton: implications for mantle metasomatism. *Manuscript submitted to Lithos*.

Paper IV: **Woodard, J.** and Huhma, H. Paleoproterozoic mantle enrichment beneath the Fennoscandian Shield: isotopic insight from carbonatites and lamprophyres. *Manuscript submitted to Lithos.*

J. Woodard was responsible for all work pertaining to the original publications with the following exceptions: P. Hölttä made the original field discovery of the dykes in Paper I. C.J. Hetherington performed the apatite and monazite analyses, calculated the monazite age, wrote the EPMA methods section and contributed to the dissolution-reprecipitation discussion in Paper II. For Paper III, O. Eklund and A. Shebanov were responsible for the fieldwork and some of the zircon analyses. R. Kietäväinen contributed to the petrography. H. Huhma assisted with the radiogenic isotope data reduction in Paper IV.

1. INTRODUCTION

Alkaline igneous rocks, including volatile-rich varieties such as carbonatites, lamprophyres and kimberlites, have much greater petrological significance than their relative volumetric abundance would indicate. A common thread connecting these relatively diverse rock types is their generation via low-volume partial melting of mantle domains enriched in trace elements and volatiles. Owing to their deep source and rapid, often violent mode of emplacement, alkaline rocks are some of the most important carriers of mantle xenoliths. These xenoliths are vital to our understanding of the composition of the deep earth, as they provide a means for direct observations and analyses of material from the mantle. In addition to their scientific value, alkaline rocks are often of great economic value. Kimberlites and lamproites are important hard rock sources for diamonds (e.g. Mitchell, 1986), while some lamprophyres may be associated with gold deposits (e.g. Rock, 1991). Carbonatites have been mined for a multitude of purposes, including rare metals, phosphates and industrial minerals (e.g. Mariano, 1989).

Study of the alkaline rocks themselves is of equal importance. Combining observations on the geochemical characteristics of natural rocks with experimental results may provide insight into processes of mantle metasomatism and melt generation under high-pressure conditions. Metasomatism by carbonatite melts has been recognised as an important mechanism for enrichment of mantle domains. The effects of crustal contamination can complicate interpretation of geochemical data, and care must be taken to ensure that whole-rock analyses are representative of primary compositions. Isotope studies are useful in this regard. Owing to the high absolute concentrations of REE and Sr in alkaline rocks, the Sm-Nd and Rb-Sr isotopic systems are relatively insensitive to the effects of crustal contamination.

Although it has long been recognised that alkaline rocks occur in various tectonic settings worldwide (e.g. Woolley, 1989; Rock, 1991), study of these rocks, particularly carbonatites, has focused disproportionately on within-plate continental rift environments. Due to the typically low volume and high reactivity of these magmas, it may nonetheless be assumed that an extensional tectonic environment is prerequisite to their emplacement into the upper crust. In post-collisional extensional settings, alkaline rocks have the potential to provide information about the effects of convergent tectonic processes on the geochemical and isotopic evolution of the upper mantle. Determining the emplacement ages of these rocks may also be used as a proxy to date tectonic events.

1.1. Carbonatites

Carbonatites are defined by the IUGS as igneous rocks with > 50% modal carbonate minerals. They most commonly occur as small, hypabyssal bodies such as dykes, cone sheets, plugs and sills, or as composite plutonic ring complexes in association with alkaline silicate rocks (Barker, 1989). The most recent database lists 527 occurrences of carbonatite worldwide, roughly 24% of which have no associated silicate rocks (Woolley and Kjarsgaard, 2008a; 2008b). They occur on all continents as well as several oceanic islands, spanning geologic time from the oldest known intrusions at ~2.7 Ga to present day eruptions. Although the majority of known carbonatites are found in rift or near-rift settings, they may nonetheless occur in off-craton, orogenic or collision suture settings where extension may be localised in back-arc regimes or occur from widespread orogenic collapse (Woolley and Kjarsgaard, 2008a). Carbonatite melt is highly reactive, and as such there are physiochemical barriers to its emplacement in the upper crust. Simple thermal constraints on melt generation (outlined in Section 4.2.1) may also contribute to the relative dominance of rift or near-rift settings for carbonatite occurrences. It is estimated that the majority of primary carbonatite melt generated in the mantle is consumed by reaction with mantle peridotite and undergoes "chemical death" (Yaxley et al., 1991; Rudnick et al., 1993; Dalton and Wood, 1993; Barker, 1996; Bell et al., 1998). The implication is that the structural conduits linking the mantle source to the upper crust and allowing for rapid emplacement may only occur in an extensional environment. Recently reported examples of carbonatites from post-collisional tectonic settings include Maoniuping, Lizhuang and Dalucao, Sichuan, China (Hou et al., 2006) and Eden Lake, Manitoba, Canada (Chakmouradian, 2008).

The dominant carbonate mineral can be calcite, dolomite, or ankerite; these may also constitute an evolutionary series related to fractional crystallisation processes (e.g. Le Bas, 1989). Aside from the main carbonate minerals, the most common accessory minerals are fluorapatite, phlogopite, magnetite, hematite, titanite, pyrochlore, pyroxene and amphibole (Hogarth, 1989). Chemically, carbonatites are characterised by extreme trace element enrichment, including having the highest concentration of lanthanides (REE) of any known rock type (e.g. Woolley and Kempe, 1989). In addition, characteristic alkaline metasomatic alteration (fenitisation) typically surrounds carbonatite complexes. Enrichment in high field strength elements (HFSE) such as niobium and tantalum, once considered an essential characteristic of carbonatite, is conspicuously absent from carbonatites in post-collisional tectonic settings (Hou et al., 2006; Chakmouradian, 2008; Paper I). Three main models have been proposed for the petrogenesis of carbonatite magma: partial melting of carbonated mantle to generate primary carbonatite magma (Gittins, 1989; Harmer & Gittins, 1998; Gittins & Harmer, 2003); derivation in the crust from a carbonated silicate parent melt via immiscibility (Le Bas, 1977; Kjarsgaard & Hamilton 1989; Lee & Wyllie 1998); and as "carbothermal residua" resulting from extensive crystal fractionation (Veksler et al., 1998; Mitchell, 2005).

1.2. Lamprophyres

Lamprophyres form a diverse group of volatile-rich, peralkaline to alkaline, mafic to ultramafic igneous rocks, typically occurring as dyke swarms. Rock (1991) described the "lamprophyre clan" as consisting of five different groups: shoshonitic (or calc-alkaline) lamprophyres, alkaline lamprophyres, ultramafic lamprophyres, lamproites and kimberlites. Under the current IUGS guidelines however, lamproites and kimberlites are distinct and should not be considered as lamprophyres (Woolley et al., 1996). Mineralogical classification is based upon a two-tier system involving the modal abundance of light-coloured minerals (feldspars and felspathoids) and predominant mafic minerals. Ultramafic lamprophyres contain > 90% mafic minerals, melilite and/or primary carbonate, as well as macrocrysts of olivine and phlogopite (Tappe et al., 2005). Shoshonitic lamprophyres contain feldspar as a matrix phase, while alkaline lamprophyres may contain either feldspars or felspathoids (Le Maitre et al., 1989). Shoshonitic lamprophyres with biotite as the dominant mafic mineral are termed minettes (alkali feldspar dominant) or kersantites (plagioclase dominant). Corresponding names for amphibole-dominant varieties are vogesite and spessartite (Le Maitre et al., 1989; Rock, 1991).

Shoshonitic lamprophyres are typically found in convergent or passive margin settings, often associated with other shoshonitic or calc-alkaline igneous rocks (Rock, 1991). Alkaline lamprophyres are found in all tectonic settings, typically alone or associated with mildly alkaline igneous rocks. Ultramafic lamprophyres, on the other hand, are found only in divergent margin and intraplate settings and are most commonly associated with carbonatite-ijolite-nephelinite complexes. This reinforces the notion that ultramafic lamprophyres are not related to the other lamprophyres, with distinct differences in tectonic setting, source and genetic characteristics (cf. Tappe et al., 2005).

1.3. Carbonatites and Lamprophyres in Fennoscandia

A wide spatial and temporal distribution of carbonatites and lamprophyres is known throughout Fennoscandia (Figure 1). Much of the classic work on alkaline rocks was done in Fennoscandia, as is expressed in type-locality rock names including alnöite, alvikite and beforsite (from Alnö, Alvik and Bergeforsen, Alnö complex, north-central Sweden; Rosenbusch, 1887; von Eckermann, 1928a, von Eckermann, 1942), ijolite (from Iivaara,

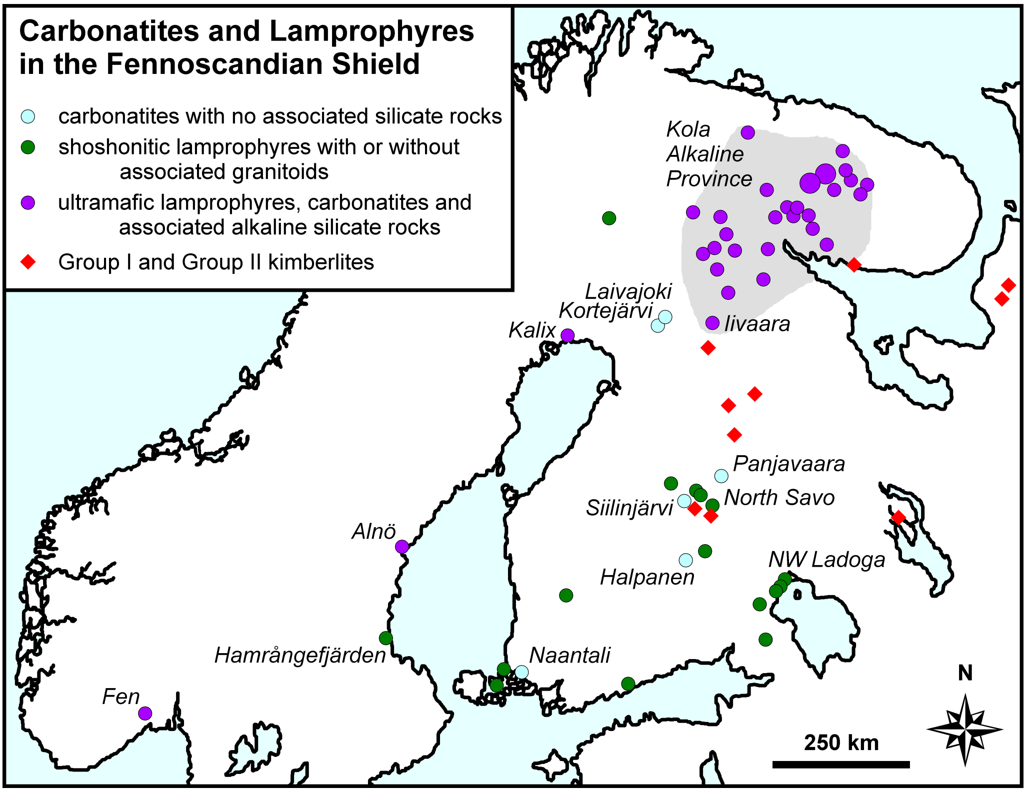


Figure 1 Map showing the spatial distribution of carbonatites, lamprophyres and kimberlites in the Fennoscandian Shield.

northern Finland; Ramsay and Berghell, 1891) as well as damtjernite, sövite, sannaite and fenite (from Damtjern, Søve, Sannavand and Fen, Fen complex, southern Norway; Brøgger, 1921; Sæther, 1957).

The ~2.6 Ga calciocarbonatite at Siilinjärvi, eastern Finland, is one of the oldest known carbonatite intrusions in the world (O'Brien et al., 2005). Two intrusions at Laivajoki and Kortejärvi, identified using aeromagnetics and sampled only by drilling, have an estimated age of ~2.02 Ga (Vartiainen and Woolley, 1974). The newly discovered dykes at Naantali, southwest Finland (Paper I; Woodard, 2005) have many characteristics in common with those at Halpanen, southeast Finland (Puustinen and Karhu, 1999) and Panjavaara, eastern Finland (Torppa and Karhu, 2007). At Kalix, northern Sweden, silicocarbonatite dykes have been reported with an age of 1142 Ma (Kresten and Brunfelt, 1980). The classic carbonatite occurrences at Fen and Alnö are coeval, with ages determined at 583 Ma (Meert et al., 1998) and 589 Ma (Rukhlov and Bell, 2009) respectively. The Devonian (360-380 Ma; Kramm et al., 1993) Kola Alkaline Province of northwest Russia and northeast Finland is the largest alkaline magmatic province in the world.

Lamprophyres have an equally wide distribution throughout Fennoscandia, and are often found in direct association with carbonatites. Ultramafic lamprophyres are found at Kalix (Geijer, 1928; Kresten and Brunfelt, 1980), Alnö (Rosenbusch, 1887; von Eckermann, 1948), Fen (Brøgger, 1921; Dahlgren, 1994) and the Kola Alkaline Province (Ramsay and Hackman, 1894; Vartiainen et al., 1978). Shoshonitic or alkaline lamprophyres occur in several locations including North Savo (Hackman, 1914; Koistinen, 1965), Viljakkala (Stigzelius, 1944), Helsinki (Eskola, 1954), Haukivesi, Pielavesi (Laukkanen, 1983) and Palovaara (Rastas et al., 2001) in Finland and in the NW Ladoga region (Ivashchenko and Lavrov, 1993; Eklund, 2003) in Russia. The "hamrongite" dykes of von Eckermann (1928b) from

Hamrångefjärden, near Gävle in central Sweden, should be classified as kersantite under the modern classification guidelines.

For this study, samples from three separate regions within the Fennoscandian Shield were examined in detail. Carbonatite and fenite samples were collected from Naantali, southwest Finland. Shoshonitic lamprophyres were collected from three separate areas in the North Savo region, eastern Finland and four locations in the NW Ladoga region, northwest Russia.

1.4. Objectives of this Study

Based on the trace element characteristics of post-collisional shoshonitic (monzonite - granite ± lamprophyre) intrusions, Eklund et al. (1998) suggested that the lithospheric mantle beneath the Svecofennian Domain had been affected by carbonatite metasomatism. When carbonatite dykes were discovered at Naantali, southwest Finland (Paper I; Woodard, 2005), together with the carbonatite at Halpanen, southeast Finland (Puustinen, 1986), it was speculated that there could be a relation. These carbonatites both have island arc trace element signatures, while a preliminary age determination from Halpanen (Puustinen and Karhu, 1999) suggested emplacement coeval with the post-collisional shoshonitic intrusions. Potassic fenitisation around the Naantali dykes further suggested an association with shoshonitic or ultrapotassic magmatism (Woodard, 2005). Several swarms of shoshonitic to ultrapotassic lamprophyres are also known in Fennoscandia, implying that all of these intrusions could be related to a single, shield-scale event. These mantle-sourced rocks offer a means to examine processes of enrichment in the subcontinental lithospheric mantle. Essential to any model of the Svecofennian mantle evolution is precise geochronological data, and obtaining this from the carbonatites and lamprophyres was an important goal of this study. This precise age data, combined with trace element and isotopic data, also facilitates correlation of the intrusions despite their wide geographic distribution. The trace element and isotopic data were subsequently used to develop a more complete model for mantle metasomatism beneath the Fennoscandian Shield.

2. SUMMARY OF THE ORIGINAL PUBLICATIONS

Paper I

Paper I uses field relationships, mineralogy and geochemistry to verify carbonate dykes at Naantali, Finland as a *bona fide* occurrence of carbonatite. A small swarm of narrow dykes (mostly 3-20 cm wide) are found to intrude Svecofennian pyroxene tonalite in the town of Naantali, southwest Finland. At least fifteen straight, sub-parallel dykes were identified in an area roughly 1 km wide and 2 km long. The dykes cut the regional SW-NE schistosity in a NW-SE orientation, dipping ~45° to the NE. An aureole of potassic fenitisation extends up to 1 km outward from the dyke swarm. The dykes contain 90-95% calcite, minor fluorapatite and allanite and accessory titanite, fluorite, chlorite and quartz. Monazite and bastnäsite were identified as inclusions in the fluorapatite. Chemically, the dykes are enriched in Sr, Y and REE, with very high relative LREE/HREE enrichment, while the fenites are also enriched in K and Ba. Based on the combined evidence of trace element enrichment, intrusive nature of the veins and the fenite alteration halo, it was concluded that the Naantali dykes formed by intrusion and crystallisation of carbonatite magma.

Paper II

In Paper II, two independent methods were used on two different mineral phases in order to determine the age of the carbonatite dykes at

Naantali, southwest Finland. Fluorapatite macrocrysts displaying irregular zoning patterns and alteration zones containing abundant inclusions, including monazite, were analysed by electron probe microanalysis (EPMA) in order to better understand their conditions of formation. The unaltered fluorapatite domains are Th-Si-LREE enriched and have distinct positive europium anomalies. Altered domains have lower concentrations of Th, Si and LREE, no Eu anomalies, and contain abundant inclusions of monazite, quartz, allanite and bastnäsite. Based on experimental partitioning data, positive europium anomalies will not form in fluorapatite without a pre-existing anomaly in the crystallising environment, either in bulk or as a local anomaly created by a co-crystallising phase. The Naantali dykes contain neither Eu anomalies in the whole rocks nor any phase known to exclude Eu relative to the other REE; therefore the fluorapatite is interpreted as xenocrystic. Most likely, fluorapatite crystallised in the presence of significant clinopyroxene and in the absence of plagioclase, for instance during metasomatism of the upper mantle. Textures within the altered domains are characteristic of dissolution-reprecipitation reactions, further indicating disequilibrium between the fluorapatite and the carbonatite. The observed inclusion assemblage can be explained by a fluid catalysed *in situ* mineral reaction such as:

$$\text{(Th-Si-LREE)-Fluorapatite} + CO_2 \rightarrow \text{Fluorapatite} + \text{Monazite} + \text{Quartz} \pm \text{Bastnäsite}$$

Furthermore, the influx of Si, Al, Fe and Mg from the wall rock alteration, together with Ca in the carbonatite promoted epidote-group mineral stability, which in the presence of REE released by fluorapatite dissolution, led to the growth of allanite.

A U-Th-Pb chemical (EPMA) date of 1797 ± 34 Ma (2σ) was determined from the monazite inclusions within fluorapatite macrocrysts, placing an age on the monazite-forming mineral reaction. In addition, inclusion free, light pink, gemmy zircon grains, interpreted as mantle xenocrysts, were analysed with the ion microprobe, resulting in an age of 1795.7 ± 8.5 Ma (2σ). The correlation in the age results suggests that mantle zircons, entrained by the carbonatite magma, became closed to diffusive lead-loss at the time of carbonatite emplacement, while the monazite inclusions formed via dissolution-reprecipitation triggered by the magmatic fluid.

Paper III

Paper III describes the petrology, geochemistry and geochronology of the post-collisional shoshonitic lamprophyres in Fennoscandia. Specifically, lamprophyre dykes were investigated from two regions in close proximity to the Archean-Proterozoic boundary: North Savo, Finland and NW Ladoga, Russia. The dykes (minettes and kersantites) contain abundant mica and apatite macrocrysts set in a matrix dominated by feldspars, mica, clinopyroxene and apatite. The magmas were produced by low-volume partial melting in the lithospheric mantle, which experienced two stages of metasomatic enrichment. First, a hydrous alkaline silicate melt enriched the mantle wedge in Al, Fe, K, Ba, Rb, P, Sr, Th, U, F, LREE and H_2O, probably crystallising as veins. This was followed by preferential infiltration of carbonatite melt along the metasomatic veins, enriching the source area in Ca, LREE, Y and CO_2. Geochemistry indicates a destructive-margin setting, with clear negative spikes for Ti, Nb and Ta in multi-element plots, suggesting melting of subducted sediment as the source for the metasomatising melts. U-Pb analyses of mantle zircons by ion microprobe resulted in ages of 1790.1 ± 3.3 Ma, 1784.1 ± 4.0 Ma and 1783.7 ± 5.4 Ma from North Savo and 1781 ± 20 Ma from NW Ladoga. The close correlation in ages (identical within error limits) indicates a shield-scale shift to an extensional tectonic regime. Furthermore, the lamprophyres, along with other coeval shoshonitic intrusions in the

Svecofennian Domain (including the carbonatites investigated in Papers I and II) may be considered to belong to a single, large shoshonitic magmatic province. In addition, inherited zircons in these rocks provide direct evidence of Proterozoic crust underlying the western margin of the Archean Karelian Province as well as Archean crust beneath the Proterozoic rocks south of the Meijeri Thrust in northwest Russia.

Paper IV

The Rb-Sr, Sm-Nd, C and O isotope geochemistry of both the carbonatites and the lamprophyres are presented in Paper IV. All of the samples plot to the right of the mantle array in the Nd-Sr correlation diagram, with compositions trending toward EMII (high radiogenic Sr; chondritic Nd), resulting from recycling of terrigeneous sediments back into the mantle. A mathematical mixing model for both the Rb-Sr and Sm-Nd isotope systems was developed to quantify mantle enrichment by recycling of a subducted mixture of Archean sediments and juvenile Proterozoic material. The results of this expanded on the two-stage metasomatic model presented in Paper III. It was shown that Rb, Sr and importantly $^{87}Sr/^{86}Sr$ were enriched in the first (hydrous alkaline silicate) stage of metasomatism, while the REE were confined to the second (carbonatitic) stage. Assuming a mixture of 45% Archean and 55% Proterozoic material, addition of 2.4% silicate melt to the depleted mantle in the southwest, increasing to 2.7% in the northeast and up to 3.0% in the southeast would produce the observed $^{87}Sr/^{86}Sr$ ratios. Subsequent addition of 4.0-4.6% carbonatite melt (in all areas) would produce the observed $^{143}Nd/^{144}Nd$ ratios. If the subducted material only contained 30% Archean sediments, the predicted percentages rise to 3.9%, 4.5%, 5.1% and 6.4-7.4% respectively. Carbon and oxygen isotopes were also analysed from calcite concentrates. Oxygen isotope ratios are slightly higher than the normal range for mantle carbonate, most likely due to minor fractionation effects during crystallisation or post-magmatic interaction with meteoric water. Carbon isotope ratios, however, are anomalously light with respect to normal mantle carbonate, which may result from large amounts of organic carbon in subducted sediments having been recycled back into the mantle. It was estimated that the metasomatising carbonatite melt in close proximity to the Archean continent had maximum $\delta^{13}C$ ratios between -17 and -19‰, similar to early Proterozoic black shales in Fennoscandia. This supports a model in which recycled organic carbon was the source of this metasomatising melt. With increasing distance from the Archean continent, $\delta^{13}C$ ratios become less negative, indicating mixing of increasing amounts of inorganic carbon with the recycled sediments.

3. TECTONIC EVOLUTION OF THE SVECOFENNIAN DOMAIN

The Fennoscandian Shield may be divided into the Archean Karelian Craton in the north, which may be further divided into the Kola, Belomorian and Karelian Provinces, the Paleoproterozoic Svecofennian Domain and the Transscandinavian Igneous Belt in the centre and the Neoproterozoic Southwest Scandinavian Domain in the southwest (Figure 2, e.g. Gaál and Gorbatschev, 1987; Nironen, 1997). The tectonic evolution of the Svecofennian Domain has been discussed in detail by many other researchers (e.g. Gaál and Gorbatschev, 1987; Nironen, 1997; Väisänen, 2002; Lahtinen et al., 2005; Korja et al., 2006), and there is no need to review all aspects here. The purpose of this section is to review the important tectonic events and how they affected the evolution of the Fennoscandian subcontinental lithospheric mantle. It should also be noted that in the following section, all compass directions refer to the modern orientation of crustal blocks.

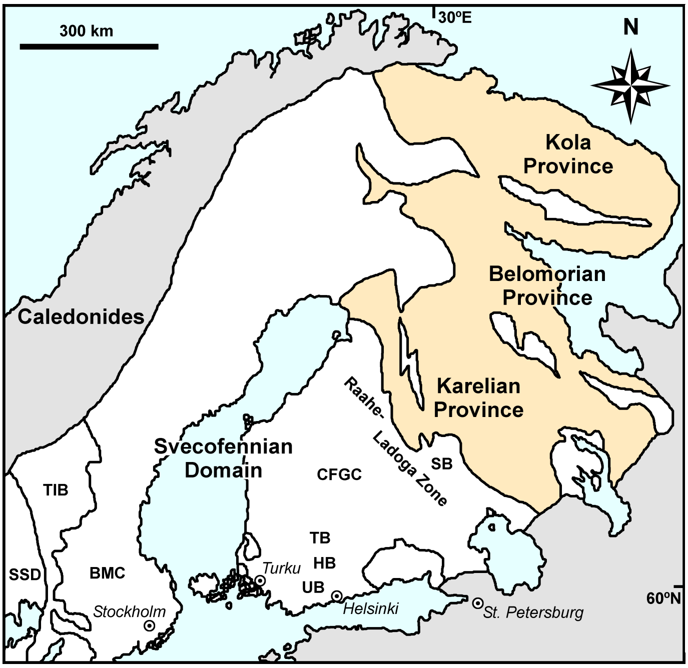


Figure 2 Map of Fennoscandia after Gaàl and Gorbatchev (1987) and Nironen (1997). Abbreviations are: SB = Savo Belt; CFGC = Central Finland Granitoid Complex; TB = Tampere Belt; HB = Häme Belt; UB = Uusimaa Belt; BMC = Bergslagen Microcontinent; TIB = Transscandinavian Igneous Belt; SSD = Southwest Scandinavian Domain.

3.1. Rifting of the Archean Continent

The first stage of rifting of the Karelian Craton is marked by the intrusion of mafic dyke swarms and large layered mafic intrusions at around 2.45 Ga, while several generations of younger, mainly tholeiitic dyke swarms intruded at 2.32 Ga, 2.2 Ga, ~2.1 Ga, ~2.05 and 1.98 Ga (Vuollo and Huhma, 2005). Continental break-up occurred beginning at ~2.2 Ga, resulting in the opening of an ocean to the southwest of the Karelian Province (Korja et al., 2006). Age determinations of 1.97-1.95 Ga from gabbros, clinopyroxenite dykes and hornblendite dykes from the Jormua and Outokumpu ophiolite complexes constrain the age of seafloor magmatism (Huhma, 1986; Kontinen, 1987; Peltonen et al. 1998).

The break-up of the Archean continent correlates with the Lomagundi-Jatuli carbon isotope event, a worldwide positive excursion in carbon isotope ratios in sedimentary carbonate from 2.22-2.06 Ga (Karhu and Holland, 1996; Melezhik et al., 2007). The formation of shallow seas during the first stages of ocean opening led to an overall increase in continental shelf area. This may have resulted in increased biological production and, consequently, increased burial rates of organic carbon (Karhu, 1993; Karhu and Holland, 1996). Evidence of further ocean opening is provided by a transition at 2.06-1.96 Ga from shallow water to turbidite and then deep-water sedimentation (Laajoki, 2005). The ensuing decrease in continental shelf area and, correspondingly, stromatolite abundance, coincided with and may have been causal to the end of the Lomagundi-Jatuli carbon isotope event (Melezhik et al., 2007).

3.2. Subduction and Accretion of Arcs

Active subduction of the newly formed oceanic crust beneath the margin of the Karelian Craton began at about 1.93 Ga, resulting in the formation of the island arc volcanics of the Savo Belt (Korsman et al., 1984; Vaasjoki and Sakko, 1988; Kousa et al., 1994). These tonalites and associated felsic volcanics have distinctly positive $\varepsilon_{Nd}(T)$ values, indicative of a juvenile depleted mantle source (Lahtinen and Huhma, 1997). Subduction reversal occurred at least twice, resulting in a gradual shift of the subduction zones to the south and southwest (Korja et al., 2006). This resulted in the formation of a succession of island arcs (north to south, the Tampere, Häme and Uusimaa Belts) from 1.90-1.88 Ga (e.g. Patchett and Kouvo, 1986; Kähkönen et al., 1989; Väisänen and Mänttäri, 2002). These newly formed arcs collided with each other and accreted onto the Archean Karelian Craton (Lahtinen et al., 2005). The maximum age for the obduction of the Jormua and Outokumpu ophiolites, and thus the closure of the intervening ocean, is 1.92 Ga, given by the youngest detrital zircons in the Kaleva metasediments (Claesson et al., 1993). Along the Archean-Proterozoic boundary, granulite facies metamorphism at 1.89-1.875 Ga (Hölttä, 1988; Vaasjoki and Sakko, 1988) coincides with the accretion of the Bergslagen microcontinent and subsequent shift of subduction zones to the southwest (Lahtinen et al., 2005; Korja et al., 2006). Intrusion of the mildly shoshonitic granitoids at Puutsaari, northwest Russia, indicates a midly enriched mantle beneath the Archean-Proterozoic boundary already at at ~1.87 Ga and that some tectonic extension may have occurred (Konopelko and Eklund, 2003).

3.3. Continental Collision

Continent-continent collision began at about 1.84 Ga with the convergence of Laurentia from the north, Amazonia from the west and Sarmatia from the southeast (Korja et al., 2006). Subduction continued in an Andean-type active margin to the southwest (Lahtinen et al., 2005; Korja et al., 2006). Transpressional shear zones developed as a result of oblique collision, and large-scale fold and thrust belts indicate crustal stacking (e.g. Ehlers et al., 1993). As a result, the Fennoscandian lithosphere is abnormally thick, up to 240 km (Kukkonen and Peltonen, 1999). In southern Finland, a second metamorphic peak (upper amphibolite to granulite facies) occurred at 1.84-1.815 Ga, manifested in extensive migmatisation and crustal anatexis forming Late Svecofennian "S-type" granites (Vaasjoki and Sakko, 1988; Suominen, 1991; Ehlers et al., 1993; Väisänen et al., 2002).

3.4. Post-collisional Extension (orogenic collapse)

Orogenic collapse is the isostatic stabilisation of the crust after a period of thickening and compression. Whether or not orogenic collapse occurred in the Svecofennian Domain is still a matter of controversy. For example, Cagnard et al. (2007) argue that regionally homogeneous metamorphic conditions and lack of major jumps in P-T conditions across shear zones are evidence against thrusting or localised detachments. According to Chardon et al. (2009), ultra-hot orogens such as the Svecofennian do not collapse, which is evidenced by the preservation of thick lithosphere. The latest of the Late Svecofennian granites in southwest Finland (Runosmäki, Turku, 1814.3 ± 2.7 Ma) was emplaced at a minimum depth of 14-15 km (~4.1 kbar; Väisänen et al., 2000). Ductile deformation was still active at ~1.79 Ga at pressures in excess of 4 kbar in shear zones in southwest Finland (Torvela et al., 2008). Carbonatite emplacement at 1795.7 ± 8.5 Ma at Naantali, southwest Finland indicates an extensional tectonic regime (Paper II). Furthermore, the presence of a calcite-prehnite-epidote-actinolite assemblage in the fenites implies pressures of ~2 kbar (Liou, 1971; Paper I). The Naantali and Runosmäki intrusions are also located within 15 km of each other, with no intervening metamorphic jumps or tectonic boundaries. Shoshonitic lamprophyres in eastern Finland and northwest Russia intruded between 1.790-1.781 Ga (Paper III). Small intrusions of post-collisional shoshonitic granitoids, indicative of a transition from compressional to extensional tectonic regime, are found in an east-west trending belt across the Svecofennian domain (e.g. Eklund et al., 1998; Andersson et al., 2006). In southeast Finland, the post-collisional Ruokolahti granite (1795 Ma; Nykänen, 1988) was emplaced at 2.5 kbar and 650ºC (Niiranen, 2000). Cagnard et al. (2007) propose that lateral ductile flow and erosion compensate for crustal thickening, and that isostatic equilibrium was likely reached before the end of regional compression. According to Harrison (1994), normal erosion rates in mountainous regions are 0.235-0.212 km/My, and significantly faster denudation rates require tectonic extension or lithospheric delamination. Assuming the crust in the Turku area was uplifted from 14 km depth at 1814.3 Ma to 6.8 km at 1795.7 Ma, an erosion rate of 0.387 km/My (> 65% faster than normal rates) would be required. In addition, Niiranen (2000) estimated a similar uplift rate (0.37 km/My) from 1810-1795 Ma in the eastern part of the Svecofennian Domain. These indications of rapid uplift and tectonic extension are taken as evidence of orogenic collapse.

4. DISCUSSION

4.1. Mantle Metasomatism

4.1.1. Types of metasomatism

Metasomatism, by definition, is the alteration of the chemical composition of a solid by interaction with a fluid phase. This very broad definition can encompass a variety of fluid compositions and processes within the crust and mantle. Fenitisation, for example, is a metasomatic process in which Na- or K-rich fluids evolved from intruding alkaline magma cause characteristic alteration in the host rocks (Brøgger, 1921). At mantle pressures and temperatures, the distinction between aqueous fluid and hydrous melt becomes blurred, such that in addition to water- or CO_2-rich liquids, the fluids responsible for mantle metasomatism can be silicate or carbonatite melts.

It is not possible to define a set of general characteristics for silicate melt metasomatism due to the wide compositional variation of such melts. Experimental evidence and observations on ultrahigh pressure rocks show that the elements enriched in arc lavas (LILE, LREE, Th, U) can be transported in significant quantities from the slab to the mantle wedge by granitic melts (Rapp et al., 1999; Hermann et al., 2006). Metasomatism by a silica-rich melt may result in the formation of orthopyroxene at the expense of olivine, increasing bulk SiO_2 while maintaining compatible trace element characteristics (e.g. Beccaluva et al., 2004; Rehfeldt et al., 2008). It has been proposed that the MARID assemblage found in some mantle xenoliths formed via high-pressure crystallisation of a melt resembling lamproite (Waters, 1987). Metasomatism by hydrous alkaline silicate melt has been considered in many areas as a necessary precursor to the generation of lamprophyric, lamproitic or kimberlitic magmas (e.g. Bergman, 1987, Mitchell, 1995; O'Brien et al., 1995; Tappe et al., 2008; Paper III).

Carbonatite metasomatism is characterised by enrichment in CaO, particularly relative to TiO_2 and Al_2O_3 (Yaxley et al., 1991; Rudnick et al., 1993). Trace element characteristics of carbonatite metasomatism include very low Ti/Eu, low Hf/Sm, high LREE/HREE, high Zr/Hf and a correlation of Zr/Hf with Ca/Sc (Yaxley et al., 1991; Dupuy et al., 1992; Rudnick et al., 1993). Enrichment in LILE and LREE without corresponding HFSE enrichment is also characteristic of carbonatite metasomatism (Yaxley et al., 1991). Recent experimental work has shown that some or all of these characteristics may be produced by silicate metasomatism of varying compositions, however the combination of several of these indicators is nonetheless probably a result of carbonatite metasomatism (Foley et al., 2009).

4.1.2. Source of the metasomatising melts

It was proposed in Paper III that the source of enrichment in the Fennoscandian subcontinental lithospheric mantle was subducted carbon-rich sediments. During subduction, metamorphism of carbon-rich pelitic sediments releases water driving arc magmatism, leaving a remnant mineral assemblage of garnet, clinopyroxene, biotite (or phengite), feldspars, kyanite, quartz and calcite (e.g. Spear, 1995). Carbonates and K-micas have a stability range high enough to bypass the volcanic arc region such that in most subduction zones CO_2, H_2O and K_2O may be carried to depths greater than 120 km (Schmidt et al., 2004; Thomsen and Schmidt, 2008). Experiments on Fe-rich carbonate-saturated pelite compositions show that at upper mantle pressures and temperatures ($P > 2.5$ GPa; $T > 900$°C), the breakdown of phengite + quartz/coesite controls silicate melting, producing hydrous alkaline silicate melt with K_2O/Na_2O wt-ratios of 5.8-8.6 (Thomsen and Schmidt, 2008). Such a melt would be an ideal metasomatising agent for the source of ultrapotassic magmas (e.g. Bergman, 1987; Ulmer and Sweeney, 2002). Carbonates continue to be stable to still higher pressures and temperatures, disappearing only through the formation of calciocarbonatite melt at 3.7-5.0 GPa and T > 1100°C (Thomsen and Schmidt, 2008). The implication of these experiments is that in an open system, subducted carbon-rich pelitic sediment may melt in two stages to first produce hydrous alkaline silicate melt and subsequently produce carbonatite melt (Paper III).

In Paper IV, isotope data was used to show that the subducted sediments were of mixed Archean and Proterozoic provenance, similar to the Svecofennian metasediments examined by Lahtinen et al. (2002). Radiogenic isotope characteristics are an important way of tracing mantle metasomatic processes. Mid-ocean ridge basalts have the lowest $^{87}Sr/^{86}Sr$ and the highest $^{143}Nd/^{144}Nd$, defining a depleted mantle isotopic reservoir relative to the chondritic bulk earth composition (DePaolo and Wasserburg, 1976; Richard et al., 1976; O'Nions et al., 1977). Ocean island basalts have slightly higher $^{87}Sr/^{86}Sr$ and lower $^{143}Nd/^{144}Nd$, while continental basalts have the highest $^{87}Sr/^{86}Sr$ and the lowest $^{143}Nd/^{144}Nd$. These characteristics define a normal "mantle array" for mantle derived igneous rocks (DePaolo and Wasserburg, 1979). A mantle reservoir defined by lower $^{87}Sr/^{86}Sr$ and lower $^{143}Nd/^{144}Nd$ relative to the mantle array, recognised as derived from recycled pelagic sediment, is defined as EMI (enriched mantle 1; Weaver, 1991; Dickin, 2005). The EMII (enriched mantle 2) component, which forms a trend toward extremely high $^{87}Sr/^{86}Sr$, is attributed to terrigeneous sediment recycling (Zindler and Hart, 1986; Dickin, 2005).

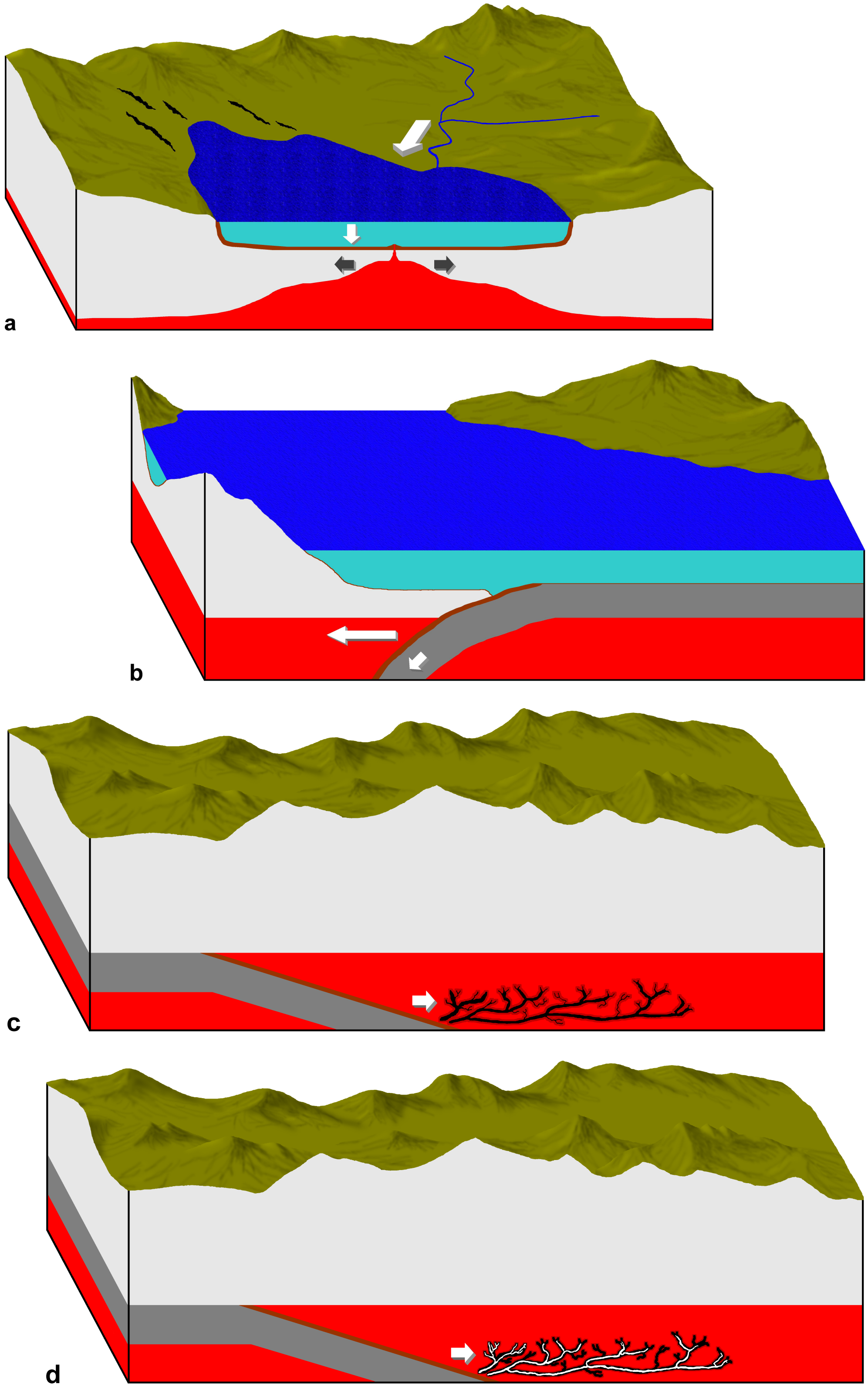
Recycling of carbon back to the mantle may also be traceable using stable isotope ratios. The carbon isotope composition of peridotitic diamonds shows a distinct mode at $\delta^{13}C = -5 \pm 1‰$ (Deines, 1980), and this is widely accepted as representing the primitive mantle carbon isotope composition.

Overall, > 90% of carbonatites worldwide have -8‰ < $\delta^{13}C$ < -2‰ (Deines and Gold, 1973; Deines, 1989). Abnormally heavy carbon isotope ratios in some primary carbonatites could result from recycled inorganic carbon in the mantle source (Ray et al., 1999; Manthilake et al., 2008). In addition, Torppa and Karhu (2007) suggested that the ^{13}C depletion in Fennoscandian carbonatites could be the result of recycled organic carbon.

4.1.3. Model for Paleoproterozoic mantle metasomatism in Fennoscandia

The first processes related to Paleoproterozoic mantle enrichment in Fennoscandia began in connection to rifting of the Archean continent (Figure 3a). At 2.2-2.06 Ga, a worldwide large positive carbon isotope excursion known as the Lomagundi-Jatuli carbon isotope event occurred (e.g. Karhu, 1993; Karhu and Holland, 1996). During this time, marine carbonates deposited in Fennoscandia have $\delta^{13}C = 10 \pm 3$‰, while those deposited after 2.06 Ga have $\delta^{13}C = 0 \pm 3$‰ (Karhu, 1993). Although it has been suggested that the Lomagundi-Jatuli carbon isotope event was caused by an increase in organic carbon burial rates, remarkably few stratigraphic units of this age containing significant organic matter have been preserved, and it has been suggested that they may have been subducted (Karhu, 1993; Karhu and Holland, 1996). Most black shales deposited between 2.2-2.06 Ga in Fennoscandia have $\delta^{13}C$ between -21 and -17‰, although values as low as -43‰ have been reported (Karhu, 1993).

With the transition to a convergent tectonic setting between ~1.96-1.93 Ga, most of the material in the sediments was derived from the juvenile volcanics from the island arcs. Metasediments deposited between 1.93-1.88 Ga in the Svecofennian Domain are comprised of mostly juvenile Paleoproterozoic material with addition of 30-45% terrigeneous material from the Archean continent (Lahtinen et al., 2002). Some of these sediments were subsequently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returned to the mantle along with the subducting slab (Figure 3b). It is reasonable to assume that the sediments that were subducted had the same mixed age provenance as those that were preserved. In Paper IV, it was estimated that the subducted sediments had maximum $\delta^{13}C$ ratios between -17 and -19‰ in the northeast and about -12 to -14‰ in other areas. This implies that in northeast, the carbon content was dominated by organic material of similar isotopic composition to the black shales, while with increasing distance from the Archean continent the proportion of inorganic material increased (Paper IV).

The arc volcanics in the Savo Belt (1.93-1.91 Ga) show little to no isotopic evidence of an Archean component, which has been taken as evidence that the island arcs were situated at some distance from the Archean continent (Huhma, 1986; Lahtinen and Huhma, 1997). However, metasediments throughout the Svecofennian Domain show evidence of a significant amount of Archean terrigeneous material (Lahtinen et al., 2002). An equally plausible solution, therefore, is that while slab dehydration beneath the island arcs triggered melting in the mantle wedge, the fluids were not able to carry the trace elements necessary to significantly alter its isotopic composition (e.g. Hermann et al., 2006). Trace element transfer (particularly the REE) to the

Figure 3 Model for mantle metasomatism beneath the Fennoscandian Shield. a) Rifting of the Archean continent led to the opening of a shallow sea at ~2.2-2.06 Ga. In this environment, increased biological production, combined with sediment influx from the Archean continent, resulted in increased burial rates of organic carbon. b) Convergent tectonics led to the formation of multiple island arcs between ~1.93-1.88 Ga. Subduction brought the organic carbon-enriched sediments of mixed (Archean and Proterozoic) provenance back into the mantle. Dehydration reactions, primarily in the blueschist facies, supplied the water to the mantle wedge, allowing for melting and driving the arc volcanism. Mica, amphibole and carbonate-bearing sediments survived this process and brought with them K_2O, H_2O and CO_2 into the mantle with the subducting slab. c) After the migration of the subduction front to the southwest, the cold slab was heated by the surrounding warm mantle. Pressure increase was more gradual, driven by crustal stacking and thickening during arc accretion (~1.88-1.87 Ga) and continent-continent collision (~1.84-1.815 Ga). Melting of the sediments produced a hydrous alkaline silicate melt, which infiltrated the mantle wedge and crystallised as veins. d) After temperatures reached 1100°C (at ~1.8 Ga), residual carbonates in the sediments began to melt. This carbonatite melt preferentially infiltrated the pre-existing metasomatic vein network.

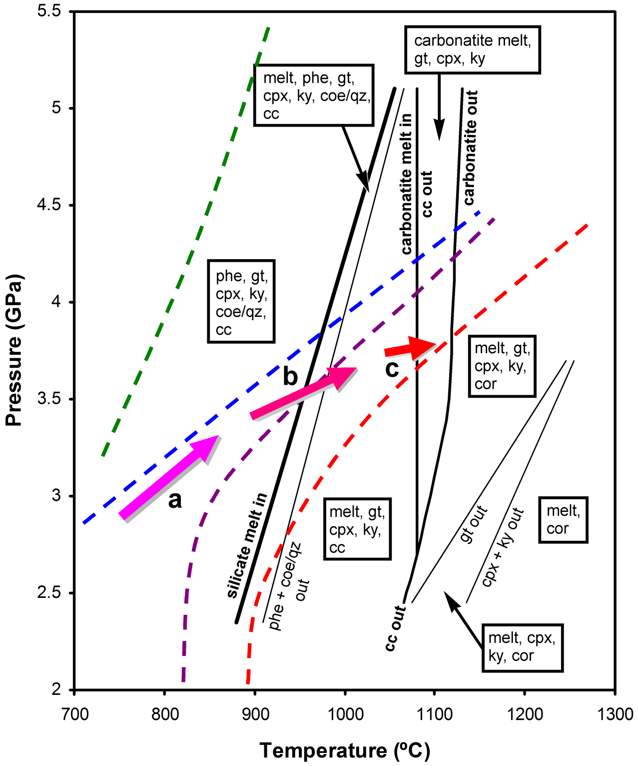

Figure 4 Phase diagram depicting a potential pressure-temperature-time path for the subducted sediments. Arrows represent: a) deep subduction of sediments prior to ~1.85 Ga. b) stagnation of subduction and gradual conductive heating (~1.85-1.82 Ga); generation of hydrous alkaline silicate melt at ~950ºC. c) Heating above 1100ºC at ~1.8 Ga; generation of carbonatite melt. Black phase boundaries are from Thomsen and Schmidt (2008); green dashed line is the H_2O-saturated solidus curve from Schmidt et al. (2004). The other three dashed lines are the modelled geotherms from Kukkonen and Lauri (2008) for 1.85 Ga (blue), 1.82 Ga (purple) and 1.80 Ga (red). Abbreviations are: phe = phengite; gt = garnet; cpx = clinopyroxene; ky = kyanite; coe = coesite; qz = quartz; cc = calcite; cor = corundum.

mantle wedge did not occur until higher pressures and temperatures were attained and the sediments themselves began to melt.

The relative downward motion of the slab likely stagnated after the closure of the ocean and migration of the active subduction zone to the southwest at ~1.87 Ga. However, pressures continued to slowly increase due to thickening of the overlying crust during continent-continent collision at ~1.84-1.815 Ga. Melting of the high-pressure carbon-rich metapelite assemblage of garnet, clinopyroxene, biotite (or phengite), feldspars, kyanite, quartz, calcite and apatite was most likely triggered by the breakdown of mica, which occurs at a minimum temperature of 900°C at ~2.5 GPa (Figure 4; Thomsen and Schmidt, 2008). Thermal modelling of southern Finland suggests that subduction and crustal stacking would have depressed temperatures such that prior to 1.86 Ga, temperatures above 900°C would only be attained at depths greater than 120 km (Kukkonen and Lauri, 2009). These depths correspond to pressures in excess of 3.5 GPa (e.g. Spear, 1995). This melting would produce a hydrous alkaline silicate melt, which could have enriched the mantle wedge in Al, Fe, K, Ba, Rb and H_2O, probably as metasomatic veins (Figure 3c). Despite its high thermal stability, apatite is highly soluble in such a melt and likely enriched it in P, F, Sr, LREE, Th and U (e.g. Watson, 1980). The mildly shoshonitic character of the granitoids at Puutsaari, northwest Russia, indicate that this process had already begun by ~1.87 Ga (Konopelko and Eklund, 2003). By about 1.80 Ga, radiogenic heat production and conductive heat transfer could have raised temperatures at this depth to ~1100°C (Kukkonen and Lauri, 2009). The residual carbonates in the subducted sediments would melt under these conditions, producing calciocarbonatite melt (Thomsen and Schmidt, 2008). Such a melt would move preferentially along zones of weakness, likely infiltrating the mantle wedge along the pre-existing metasomatic veins (Figure 3d; cf. Tappe et al., 2008). This would cause enrichment of the mantle wedge in Ca, LREE, Y and CO_2 (Paper III). As apatite is also highly soluble is carbonatite melt (e.g. Baker and Wyllie, 1992), any apatite formed during the first phase of metasomatism was likely dissolved, transported and recrystallised. Under such conditions, the Th-Si-LREE enriched fluorapatite grains bearing positive Eu anomalies found as xenocrysts in the carbonatites and lamprophyres could have formed (Paper II). The end result of this metasomatic process would be garnet lherzolite containing metasomatic veins of diopside, biotite, amphibole, apatite, zircon and carbonates (Paper III). From Sr and Nd isotopes in the carbonatites and lamprophyres, it can be seen that a small and regionally variable amount (2.4-3.0%) of silicate melt infiltration into depleted mantle, followed by metasomatism by 4.0-4.6% carbonatite melt would produce the required source characteristics given a 45/55% (Archean/Proterozoic) mixture in the subducted sediments (Paper IV).

4.2. Post-collisional Shoshonitic Magmatism (carbonatites and lamprophyres)

4.2.1. Genesis

A commonly accepted model is that primary carbonatites form by extremely low-volume partial melting of carbonated mantle peridotite (e.g. Gittins, 1989; Harmer & Gittins, 1998; Gittins & Harmer, 2003; Mitchell, 2005). Although the carbonatite solidus temperature in simple experimental systems (such as CaO-MgO-Al_2O_3-SiO_2-CO_2) is noticeably pressure-dependent (Gudfinnsson and Presnall, 2005), recent experiments using synthesised natural compositions show no discernable change in solidus temperature with changing pressure (Thomsen and Schmidt, 2008). This implies that the generation of carbonatite by decompression melting is not feasible in nature (see Figure 4). Therefore, an increase in temperature seems to be prerequisite to carbonatite melt generation. Models involving lithospheric delamination (Kukkonen et al., 2008) or slab break-off (Väisänen et al., 2000; Eklund and Shebanov, 2002), resulting in asthenospheric upwelling, have been proposed for the Svecofennian Domain, however such models are not without controversy (Cagnard et al., 2007; Chardon et al., 2009). Rb-Sr and Sm-Nd isotopic characteristics suggest derivation from a previously depleted mantle source enriched by subducted sediments (Paper IV), although isotopes alone can neither confirm nor rule out a fresh asthenospheric component. However, corresponding enrichment in compatible trace elements (e.g. Ni, Cr, V) and HFSE (Ti, Nb, Ta, Zr, Hf) would be expected if melt generation involved fresh asthenospheric input. Paleoproterozoic carbonatites and lamprophyres in Fennoscandia have arc signature trace element patterns, with relative to extreme depletion in compatible elements and HFSE (Paper I, Paper III). This indicates that no significant mass transfer from the asthenosphere was involved. Sufficient heat may be generated *in situ* by the radioactive decay of U, Th and K, particularly in apatite or mica bearing metasomatised areas (O'Reilly and Griffin, 2000). Alternatively, asthenospheric upwelling may provide the required thermal energy by conductive heat transfer without any accompanying mass transfer (e.g. Turner et al., 1992).

Generation of the lamprophyric magmas, and subsequently also the shoshonitic monzonites, requires more extensive melting and more absolute heat energy. Contrary to the carbonatites however, these melts may also be produced via decompression melting. Furthermore, it has been shown that the presence of CO_2 destabilises hydrous minerals at high pressures (Ulmer and Sweeney, 2002). As the stage of silicate metasomatism resulted in crystallisation of mica and amphibole in the source region, subsequent infiltration of carbonatite melt could have fluxed melting, producing the lamprophyric magma (Paper III). Xenocrystic mantle zircons found in both the carbonatites and lamprophyres give intrusion ages, meaning that prior to generation and emplacement, ambient temperatures in the source region were in excess of 1000ºC (Cherniak and Watson, 2001; Paper II, Paper III). This implies source depths of at least 110 km and up to 150 km (~3.2-4.4 GPa) depending on the geotherm used (Kukkonen and Peltonen, 1999; Kukkonen and Lauri, 2008). Preliminary thermobarometric data from xenocrystic biotite in the lamprophyres also imply source conditions of ~4 GPa (135 km) and ~1200ºC (Woodard et al., 2008).

4.2.2. Emplacement

Emplacement of carbonatite occurred at Naantali, southwest Finland at 1795.7 ± 8.5 Ma (Paper II) and at Halpanen, southeast Finland at 1792.0 ± 0.9 Ma (Rukhlov and Bell, 2009). The lamprophyres in eastern Finland and northwest Russia were emplaced at 1790-1781 Ma (Paper III). This suggests a very short time interval between the carbonatite metasomatism (see section 4.1.3) and the generation of these shoshonitic magmas. The stability range of the assemblage of calcite-prehnite-epidote-actinolite in the fenites implies emplacement pressures of ~2 kbar (Liou, 1971; Paper I). Preliminary thermobarometric data from matrix biotite in the lamprophyres also implies an emplacement depth less than 5 km (1-2 kbar; Woodard et al., 2008), and suggests that significant uplift had already occurred by ~1.79 Ga. This is further support that over a wide geographical area at ~1.79 Ga, the onset of orogenic collapse provided the necessary extensional environment and structural conduits that allowed primary carbonatitic and lamprophyric magmas to survive transport through > 105 km of lithosphere.

Crustal zircon xenocrysts with Proterozoic ages in the North Savo lamprophyres provide direct evidence of Proterozoic crust underlying the western margin of the Archean Karelian Province in this region (Paper III), as has been proposed by Plomerová et al. (2006). Similarly, the NW Ladoga lamprophyres, which intrude Proterozoic rocks, have inherited Archean crustal zircons (Paper III). The possibility must be considered that these zircons were inherited from metasediments of mixed age provenance. However, this is considered highly unlikely because for each of the four sampling area, the ages of xenocrysts are either all Archean or all Proterozoic (Paper III), whereas if the zircons were inherited from metasediments, a mix of ages similar to those within the metasediments themselves would be expected. This is therefore taken as evidence of buried Archean crust in this area, the first such evidence of Archean south of the Meijeri Thrust in northwest Russia.

5. CONCLUSIONS

The main conclusions that may be drawn as a result of this study are:

1: The carbonate dykes at Naantali are in fact carbonatite, with distinct crosscutting intrusive structures and magmatic textures. An aureole of potassic fenitisation surrounds the dykes. Their carbonatitic nature is confirmed by their trace element characteristics, having strong enrichment in both LILE and REE, with an extremely high LREE/HREE ratio.

2: The combination of trace element geochemistry, isotope geochemistry and geochronology confirms that the Paleoproterozoic carbonatites and lamprophyres in Fennoscandia can be correlated and are the result of prolonged mantle enrichment, followed by low-volume partial mantle melting and rapid emplacement in the upper crust.

3: Mantle metasomatism occurred as a two-stage process. First, melting of subducted sediments produced a hydrous alkaline silicate melt that crystallised veins of diopside, biotite, amphibole, apatite and zircon in the garnet lherzolite stability field. This process enriched the mantle wedge in Al, Fe, K, Ba, Rb, P, Sr, Th, U, F, LREE and H_2O. A second stage of metasomatism occurred with the infiltration of carbonatite melt, causing enrichment in Ca, LREE, Y and CO_2.

4: Isotopic mixing models predict that the first stage of metasomatism involved addition of ~2.4% hydrous alkaline silicate melt to the depleted mantle in the western Svecofennian Domain, increasing to ~2.7% in the north and ~3.0% in the east while assuming the recycled material was a mixture of 45% Archean and 55% Proterozoic material. The second stage of metasomatism involved addition of ~4.0-4.6% carbonatite melt to the metasomatised mantle. These percentages rise with decreasing proportions of Archean material.

5: Emplacement of the carbonatite dykes at Naantali occurred at 1795.7 ± 8.5 Ma. Lamprophyres were emplaced along the Archean-Proterozoic boundary in eastern Finland and northwest Russia in a ~10 My interval, the earliest in North Savo at 1790.1 ± 3.3 Ma and the latest in the NW Ladoga region at 1781 ± 20 Ma.

6: Due to the high reactivity of these magmas, emplacement must have been rapid and an extensional tectonic environment was required. Preliminary thermobarometric data indicate emplacement in the upper crust; evidence for significant uplift throughout the Svecofennian Domain within a ~15 million year period. These indications of extension and uplift are taken as evidence of orogenic collapse in the Svecofennian Domain beginning shortly before 1.795 Ga.

7: Inherited zircons in the lamprophyres provide direct evidence of Proterozoic crust beneath the margins of the Archean Karelian Province. Likewise, inherited zircons in the NW Ladoga lamprophyres provide the first evidence of buried Archean crust south of the Meijeri Thrust.

8: Further study of these rocks could provide insight into the role of these metasomatic events in the formation of Finnish diamonds. The diamond potential of these rocks is still somewhat of a mystery, however preliminary data indicate the source may have been within the diamond stability field.

However, it must be noted that diamond indicator minerals such as Cr-pyrope, Cr-diopside or picroilmenite have never been found in these rocks. Additional thermobarometry on these and other known lamprophyres will better constrain both emplacement and source depths, which combined with geochronology may also assist with determining crustal uplift rates. Xenocrysts and xenoliths from these rocks provide direct samples from the lower crust and upper mantle. In short, these exotic rocks offer intriguing opportunities to expand our knowledge about the evolution of the crust and mantle in Fennoscandia.

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