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1.8 Ga magmatism in the Fennoscandian Shield; lateral variations in subcontinental mantle enrichment

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Dedicated to the memory of Matti Vaasjoki

Abstract

The ca. 1.8 Ga post-collisional magmatism in the southern part of the 2.1–1.86 Ga Svecofennian domain of the Fennoscandian Shield have been studied with particular reference to the character and sources of the mafic rocks in a traverse from the Archaean craton margin in east, across the juvenile Svecofennian domain to its western margin. For this purpose three key areas were selected in the eastern, central and western parts of the domain: (i) in the western part of the domain, the Tjällmo-Vättern zone (southern Sweden) of the Transscandinavian Igneous Belt (TIB) consists of extensive areas of dominantly alkali-calcic granitoids associated with calc-alkaline to tholeiitic mafic rocks. Initial ε_{Nd} for the mafic rocks vary from around 0 to above +3. (ii) In the central part of the domain (SW Finland), the post-collisional rocks are represented by small intrusions consisting of calc-alkaline high-Ba–Sr granites associated with shoshonitic lamprophyres and their plutonic equivalents. Initial ε_{Nd} for the rocks in this series vary between 0 and +1. (iii) In the easternmost part of the Svecofennian domain (Russian Karelia and SE Finland) shoshonitic associations occur, comprising lamprophyres and their plutonic equivalents (apatite and magnetite-rich monzodiorites), which are related to syenites and high-Ba–Sr granites by fractional crystallization. All the rock types in this shoshonitic association have strongly elevated contents of P_2O_5, LREE and LILE. Initial ε_{Nd} for all rocks in Karelia fall between 0 and -1.

Geochemical and isotopic results indicate that the post-collisional rocks in the central and eastern part of the domain stem from lithospheric mantle sources that were enriched during the preceding Svecofennian orogeny. The HFSE depletion, combined with the strong Sr, LILE and LREE enrichment, recall signatures of increasingly carbonate-dominated metasomatism of the mantle eastwards towards the craton margin. In contrast, the mainly LILE enriched mafic rocks from TIB in the west signal sources subjected to H_2O-dominated metasomatism, that could in part be coeval with the magmatism. In all areas the rocks carry a subduction type chemistry with continental arc affinity, however, with strongly increasing enrichment levels eastwards.

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Rocks in the west are derived by relatively larger degrees of melting at shallower levels, from previously depleted spinel–phlogopite–amphibole lherzolites/harzburgites, while going eastwards successively smaller melt fractions were tapping deeper, more enriched mantle sections in the garnet stability field. The enrichment agents are interpreted to be LILE-bearing H_2O-dominated fluids from dewatering slabs in the west, changing to an increasing role for CO_2-dominated fluids/melts derived mainly from subducted Svecofennian metasediments eastwards.

A convergent continental margin setting with transpressional shearing was active during TIB formation in the west. Whether this shearing was instrumental in the formation of the 1.8 Ga magmatism further continentwards, or if the magmatism in the central and eastern areas was the result of extensional collapse or plume activity is presently not known.

Keywords: Mantle enrichment; Carbonate metasomatism; 1.8 Ga mafic magmatism; Fennoscandian Shield; Nd–Sr isotopes

1. Introduction

The Fennoscandian Shield consists of an Archaean cratonic nucleus in the northeast, in NE Finland and Russia (e.g. Gaál and Gorbatschev, 1987). In the Palaeoproterozoic large amounts of new, mostly juvenile, crust grew and successively accreted onto the SW margin of the old craton and the extensive Svecofennian (2.1–1.86 Ga) Domain was formed (e.g. Nironen, 1997). Continued westward growth added

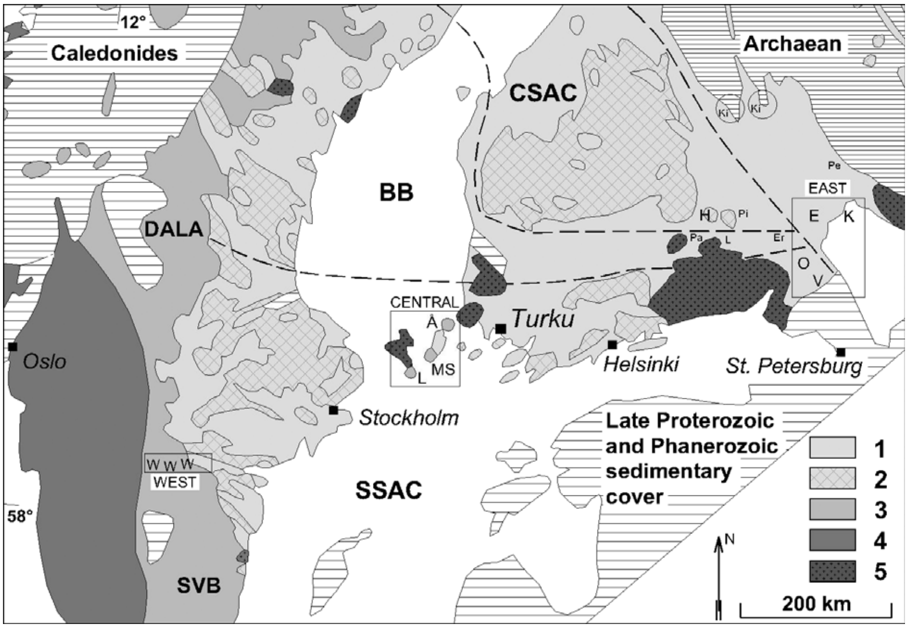

Fig. 1. The south-central part of the Fennoscandian Shield with the three areas of investigated ca. 1.8 Ga mafic rocks marked by frames. The western area is represented by the Tjällmo-Vättern zone of the Transscandinavian Igneous belt (cf. Andersson, 1997b). W = the principal sampling sites of the mafic rocks from this area. The central area is represented by the Åland islands with the investigated intrusions: L = Lemland, MS = Mosshaga and Seglinge, Å = Åva. The eastern area is represented by the post-collisional intrusions in Russian Karelia. V = Vuoksi, E = Elisenvaara, K = Kalto island. Other post-collisional intrusions are: O = Ojajärvi, Pe = Petravaara, Er = Eräjärvi, Pi = Pirilä, L = Luonteri, Pa = Parkkila, H = Halpanen carbonatite. Ki = kimberlite provinces. Inferred sutures are marked by broken lines separating the following terranes in the Svecofennian domain: CSAC = central Svecofennian arc complex; BB = Bothnian basin; SSAC = southern Svecofennian arc complex (after Nironen et al., 2000). Lithological units are: (1) roughly 1.9 Ga Svecofennian supracrustal rocks. (2) ca. 1.88 Ga syn- and 1.84–1.77 Ga late orogenic granitoids. (3) The Transscandinavian Igneous Belt (TIB). (4) Post-Svecofennian rocks of the Southwest Scandinavian Domain. (5) Rapakivi granites (1.6–1.5 Ga). SVB = Småland-Värmland belt in TIB. Dala = the Dala volcanic province (Dalarna).

the generally post-1.7 Ga terranes of SW Scandinavia (e.g. Lindh, 1987). Before that, however, almost the entire shield experienced extensive metamorphic and magmatic activity 1.85–1.75 Ga ago, with magmas ranging from S-type granites to various mantle-derived rocks (e.g. Korsman et al., 1999; Högdahl et al., 2004).

During and following the high-*T* and low-*P* metamorphism, the Svecofennian crust in the accretionary arc complex of southern Finland was punctured by at least 14 small P-, F-, Ba-, Sr-, and LREE-enriched bimodal mafic-felsic shoshonitic (cf. Morrison, 1980) intrusions (Fig. 1) into mid-crustal and hypabyssal levels (Eklund et al., 1998; Väisänen et al., 2000). These small intrusions occur in a 600 km long belt extending from Lake Ladoga in Russian Karelia in the east to the Åland archipelago in the west. Age determinations from different intrusions indicate that this particular shoshonitic event took place between 1815 and 1770 Ma ago, most ages clustering around 1800 Ma (Eklund et al., 1998). Simultaneously, extensive magmatic activity took place along the about 1500 km long Transscandinavian Igneous Belt (TIB) situated to the west of the Svecofennian domain (Högdahl et al., 2004).

In order to examine the lateral variation in type and origin of the ca. 1.8 Ga post-collisional or post-accretional magmatism and, in particular, the nature and mantle source character of the mafic rocks, across the Svecofennian domain of the Fennoscandian Shield, three key areas were selected (Fig. 1). These represent: (i) in the east, intrusions at the border between the old, Archaean craton and the juvenile Svecofennian (2.1–1.86 Ga) crust, (ii) in the center, ca. 1.8 Ga intrusions in the 'middle' of the juvenile Svecofennian block, and (iii) in the west, TIB intrusions at the western 'edge' of the Svecofennian Domain, where it borders the even younger domains of southwest Scandinavia. Hence, an objective is to make a first reconnaissance investigation of the differences in the character and origin of coeval mafic rocks formed across and immediately following the accretion of juvenile Palaeoproterozoic crust to a cratonic nucleus.

For this purpose we present geochemical data for mafic rocks from the three key areas, representing a 900 km cross section from the easternmost (Russian Karelia) to the westernmost (central southern Sweden) parts of the Svecofennian domain. In addition, we also give new Nd and Sr isotope data from the eastern and central key areas as well as a U–Pb zircon age determination for one intrusion in the east.

2. Geological framework

Rifting of the southeastern margin of the 3.1–2.6 Ga Archaean craton in the northeastern part of Fennoscandia was active in early Proterozoic times ca. 2.5–2.1 Ga, forming mafic layered intrusions and dyke swarms, and associated sediments in mainly NW–SE trending graben structures (e.g. Gaál and Gorbatschev, 1987; Huhma et al., 1990; Nykänen et al., 1994). During this period, most likely, part of the craton was rifted off and an intervening ocean formed (Peltonen et al., 1998).

Following this period, juvenile, early Svecofennian, crust was formed between 2.1 and 1.86 Ga (e.g. Wilson et al., 1985; Huhma, 1986; Patchett and Arndt, 1986; Patchett et al., 1987; Claesson et al., 1993; Lahtinen and Huhma, 1997), and is composed dominantly of calc-alkaline volcanic rocks and tonalite–trondhjemite–granodiorite rock series (including numerous mafic bodies). The crust formation occurred in several volcanic-arc systems with intervening sedimentary basins that by 1.86 Ga had accreted on to the Archaean continental nucleus in the northeast (Fig. 1) (e.g. Hietanen, 1975; Nurmi and Haapala, 1986; Gaál, 1986, 1990; Gaál and Gorbatschev, 1987; Front and Nurmi, 1987; Nironen, 1997). The pronounced mafic-felsic bimodality, particularly of the volcanic rocks, has led several workers to propose rifted continental arc settings for the Svecofennian volcanic belts (e.g. Claesson, 1985; Rickard, 1986; Vivallo and Claesson, 1987; Baker et al., 1988; Lagerblad, 1988; Vivallo and Willdén, 1988; Gaál, 1990; Allen et al., 1996). The earliest volcanic-arc crust formation is thought to have occurred 2.1–1.9 Ga, which was then subsequently partly rifted and obliterated during the later main early Svecofennian stage 1.9–1.86 Ga. The first phase is today only preserved as isolated units within the 1.9–1.86 Ga successions (e.g. Wasström, 1993, 1996; Skiöld et al., 1993; Welin et al., 1993; Lahtinen and Huhma, 1997; Vaasjoki et al., 2003), and as inherited zircons in metasediments and other rocks (Claesson et al., 1993; Kumpulainen et al., 1996; Andersson et al., 2000, 2004c; Lahtinen et al., 2002).

Crust formation and accretion were followed by continued convergence accompanied by episodes of metamorphism and crustal remelting 1.88–1.76 Ga in different terranes of the Svecofennian domain, generating migmatites and numerous smaller granitic intrusions of mainly S-type (sedimentary areas) and less abundant I-type granitoids (volcano-plutonic areas) (e.g. Suominen, 1991; van Duin, 1992; Ehlers et al., 1993; Romer and Öhlander, 1995; Claesson and Lundqvist, 1995; Öhlander and Romer, 1996; Andersson, 1997a; Andersson and Öhlander, 2004). In southern Finland this activity had ceased already at ca. 1.81 Ga (Lindroos et al., 1996; Korsman et al., 1999).

In the central part (SW Finland) and eastern part (Russian Karelia) of the Svecofennian domain, shoshonitic (transitional to alkaline) magmatism took place directly after the end of metamorphism and migmatisation at 1.81–1.77 Ga (Konopelko, 1997; Eklund et al., 1998; Väisänen et al., 2000).

The voluminous N–S trending magmatism of the Transscandinavian Igneous Belt (TIB) intruded the western margin of the Svecofennian domain 1.85–1.65 Ga ago (Högdahl et al., 2004). In a southern Sweden NE–SW transect, rocks become successively younger from ca. 1.85 Ga at the Svecofennian margin and progressively further away across the TIB, down to ca. 1.65 Ga in the west where Sveconorwegian (1.1–0.9 Ga) overprinting becomes penetrative (e.g. Åhäll and Larsson, 2000; Andersson and Wikström, 2001; Wahlgren and Stephens, 2004, and ref. therein). The earlier stages of TIB magmatism (1.85–1.75 Ga) overlap late Svecofennian metamorphism, anatexis, and granite formation in time (Andersson, 1991, 1997a,b; Andersson and Öhlander, 2004, and ref. therein). The TIB extends for about 1500 km from southernmost Sweden northward beneath the Caledonides to the Atlantic coast in northern Norway (Gaál and Gorbatschev, 1987; Gorbatschev, 1985; Skår, 2002; Corfu, 2004). TIB is dominated by granites to quartz monzodiorites in large areas, usually with a coarse porphyritic texture (Gorbatschev, 2004). Associated volcanic rocks are abundant in the south (Småland) and central (Dalarna) areas. In many areas mafic to intermediate plutonic rocks are common, and in Dalarna also basalts to andesites (Nyström, 1999, 2004). Mafic magmatic enclaves, and evidences for magma mingling and mixing have been reported from both plutonic and volcanic areas showing the contemporaneity of the mafic and felsic magmatism (Andersson, 1991, 1997b; Nyström, 1999).

The 1.8–1.7 Ga supracrustal and plutonic rocks of the Dala province (Dalarna) in central Sweden, traditionally included in the TIB, appear to represent a transition from post-collisional to anorogenic magmatism (Ahl et al., 1999, 2004; Lundqvist and Persson, 1999). However, according to Nyström (1982, 1999, 2004) the volcanic rocks in the Dala province have been formed in an Andino-type continental margin setting.

The last major episode of reworking of the Svecofennian crust was the emplacement of the rapakivi granite complexes with associated gabbro-anorthosites at 1.65–1.47 Ga (Rämö and Haapala, 1995; Andersson, 1997c; Andersson et al., 2002, and ref. therein).

3. Key areas

3.1. West

The rocks from the TIB in the west come from the Tjällmo-Vättern zone (TVZ), which is situated within the southern TIB province of the Småland-Värmland Belt (SVB), close to the southwestern margin of the

Notes to Table 1:

* = Not analyzed, 0 = below detection limit, significant number of digits reflects reported analytical precision.

Localities: 1 = Tjällmo-Vättern zone; 2 = Lemland intrusion; 3 = Åva ring-intrusion; 4 = Elisenvaara intrusion; 5 = Vuoksi intrusion; 6 = Halpanen carbonatite dyke.

West: major elements, Ba, Rb, and Zr by XRF. Cs, Hf, and REE by NA, and Sr, Pb, Cr, and Ni by ICP-AES. Analyses by X-ray Assay Laboratories, Canada (see Andersson, 1997a,b,c for details), except Th, U, Y, Nb and Ta by ICP-MS: Genalyses, Australia.

Central: 11–16; major elements by XRF, trace elements by ICP-MS provided by Allan Kolker. 17–28: analyzed at X-ray Assay Laboratories, Canada; major elements, Ba, Rb, Sr, Zr, Y, Nb, and Cr by XRF. REE, Cs, Hf, Ta, Th, and U by NA, while Ni and Pb were analyzed by ICP-AES. East: 29–31; major elements by ICP-AES, Ecole des Mines de Saint Etienne, France. Trace elements by ICP-MS, EU geochemical facility, Bristol (analyst S. Fröjdö); 32; major elements by ICP-AES Ecole de Mines de Saint Etienne, France, trace elements by ICP-MS, Actlabs, Canada. 33; ICP-MS, CO_2 by infrared, Actlabs, Canada.

Table 1
Major and selected trace elements for samples from West, Central and East

Number	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Sample	8828B	8802	8829	8846	8848	8822	8860	8842A	8839	8850	49NOE85	86.11 lant	86.2 lant	pumx2 lant	pumx4 lant	pumx5 lant
Area	west	west	west	west	west	west	west	west	west	west	central	central	central	central	central	central
Locality	1	1	1	1	1	1	1	1	1	1	2	2	2	2	2	2
Rock type	mafic dyke	gabbro	medium grained gabbro	coarse hbl gabbro	ultra-basite	fine-grained basite	medium-grained gabbro	medium-grained gabbro	coarse gabbro	mafic dyke	px-hbl gabbro	monzonite	monzonite	monzonite	monzonite	monzonite
Major elements																
SiO_2	50.00	46.10	51.50	44.30	44.90	47.00	49.50	52.20	48.40	50.70	46.62	54.42	51.05	50.14	56.71	55.05
TiO_2	1.53	1.13	0.92	1.18	0.27	1.62	0.99	1.14	1.37	0.71	0.77	1.34	1.90	1.66	1.18	1.44
Al_2O_3	15.50	14.30	16.90	16.00	18.40	15.80	15.90	15.80	15.40	11.70	11, 51	14.30	14.96	15.61	16.42	15.53
Fe_2O_3	12.20	13.70	10.60	14.80	9.49	13.50	10.80	11.00	13.10	10.70	14.74	9.66	10.08	9.65	7.14	8.22
MnO	0.18	0.17	0.14	0.15	0.14	0.21	0.17	0.18	0.18	0.18	0.25	0.17	0.14	0.14	0.10	0.12
MgO	5.54	7.17	4.71	6.72	10.40	6.88	5.82	5.05	5.87	9.73	11, 83	5.08	5.27	6.05	3.49	4.24
CaO	8.02	10.60	8.73	12.10	9.59	8.38	9.38	7.78	9.35	10.10	8.01	7.21	6.62	6.77	4.39	5.64
Na_2O	2.24	2.08	2.12	1.35	1.45	2.58	2.47	2.95	2.21	2.55	2.08	3.12	2.49	2.70	3.05	2.59
K_2O	2.49	1.36	1.70	0.71	0.84	1.30	1.46	1.48	1.41	1.09	3.05	2.79	4.56	4.23	5.39	4.41
P_2O_5	0.57	0.13	0.13	0.14	0.07	0.15	0.26	0.27	0.49	0.17	0.1	1.32	1.93	1.85	1.13	1.47
CO_2	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*
LOI	1.39	1.93	1.93	1.16	3.16	1.47	1.77	1.08	0.47	0.70	*	*	*	*	*	*
Sum	99.66	98.67	99.38	98.61	98.71	98.89	98.52	98.93	98.25	98.33	98.96	98.60	98.14	97.98	98.39	98.01
Trace elements																
Ba	581	308	522	185	191	221	332	458	426	171	1410	1459	4414	4138	4492	4016
Rb	108	38	55	32	21	55	33	60	52	80	78	104	103	108	112	106
Sr	270	336	367	439	490	258	452	231	393	332	464	1219	2240	2342	1783	1926
Cs	4	2	3	2	4	4	3	3	1	2	2	3	2	2	2	2
Th	2.88	1.56	4.03	1.90	1.23	1.83	2.96	1.81	2.57	2.99	22	9	14	12	14	13
U	0.92	0.82	1.60	0.56	0.45	0.49	1.07	0.65	1.22	0.78	1	4	4	3	5	3
Y	27.65	17.04	16.56	14.95	6.36	25.27	23.41	31.70	26.96	15.25	26	29	35	31	27	33
Zr	178	90	65	42	22	114	85	169	162	76	174	290	305	337	307	364
Hf	4	2	3	2	1	4	3.8	4.5	4.1	2.0	4	8	8	9	9	11
Nb	11.53	5.15	4.65	3.56	2.15	4.08	5.98	10.26	8.03	4.88	15	20	23	21	20	25
Ta	0.59	0.36	0.39	0.26	0.20	0.36	0.35	0.76	0.44	0.43	2	1.1	1.3	1	0.9	1
Pb	0	0	0	0	0	0	0	6	0	4	8	26	22	24	26	28
Cr	112	63	22	70	33	55	54	69	74	654	963	109	30	70	28	33
Ni	48	8	5	19	109	62	10	35	30	156	78	0	0	0	0	0
REEs																
La	36.4	15.3	17.1	14.3	7.9	10.1	24.2	34.2	29.0	18.4	298	107	242	227	163	210
Ce	76	33	37	28	16	25	50	68	58	38	393	209	444	411	305	383
Pr	*	*	*	*	*	*	*	*	*	*	28	*	*	*	*	*
Nd	33	16	19	15	8	15	23	33	31	17	90	82	163	162	119	147
Sm	6.5	3.7	4.3	3.1	1.5	4.1	5.5	6.9	7.1	3.6	10	12.1	21.2	19.9	15.5	18.2
Eu	1.9	1	1.3	0.9	0.4	1.5	1.5	1.4	2.0	1.1	1, 9	2.9	5.7	4.7	3.9	4
Gd	*	*	*	*	*	*	*	*	*	*	6.9	9	14.1	14.1	11.3	13.2
Tb	0.8	0.5	0.6	0	0	0.8	0	0.9	0.7	0.5	0.86	*	*	*	*	*
Dy	*	*	*	*	*	*	*	*	*	*	4.7	5.7	8.4	8	7.2	8.2
Ho	*	*	*	*	*	*	*	*	*	*	0.9	*	*	*	*	*
Er	*	*	*	*	*	*	*	*	*	*	2.7	2.5	3.2	3.5	3	3.2
Tm	*	*	*	*	*	*	*	*	*	*	0.36	*	*	*	*	*
Yb	2.9	1.6	1.6	1.4	0.7	2.4	2.4	3.6	2.7	1.4	2.6	2.2	2.4	2.8	2.9	2.9

Number	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33
Sample	ava-1	ava-12	ava-13	ava-15	ava-17A+B	ava-18	ava-19	ava-33A	ava-38	ava-5	ava-6	ava-8	elislamp	elne	elmela	vuoksi 2686	halp-white
Area	central	central	cent ral	central	central	central	central	central	central	central	central	central	east	east	east	east	east
Locality	3	3	3	3	3	3	3	3	3	3	3	3	4	4	4	5	6
Rock type	lamprophyre	lamprophyre	lamprophyre	lamprophyre	lamprophyre	lamprophyre	lamprophyre	lamprophyre	lamprophyre	lamprophyre	lamprophyre	lamprophyre	lamprophyre	nevoite	ladogite	lamprophyre	carbonatite
Major elements																	
SiO_2	55.40	49.00	47.70	48.90	46.90	49.40	50.50	47.50	46.90	49.00	48.20	45.90	46.68	42.93	52.02	40.39	0.19
TiO_2	1.70	1.74	1.62	1.75	1.93	1.88	2.21	1.66	1.92	1.77	1.90	2.32	1.85	1.72	2.05	2.08	0.49
Al_2O_3	13.80	13.90	14.10	13.90	14.70	13.60	13.80	15.20	14.40	14.50	13.30	13.40	14.95	11.21	15.96	12.44	0.04
Fe_2O_3	9.46	10.20	11.70	10.10	10.99	10.70	11.40	10.80	11.59	10.60	11.10	13.40	10.40	9.46	9.55	11.12	0.42
MnO	0.13	0.15	0.17	0.15	0.16	0.20	0.19	0.16	0.17	0.16	0.15	0.21	0.16	0.16	0.12	0.16	0.15
MgO	3.61	4.55	5.71	4.51	5.52	5.23	4.06	5.01	5.49	4.62	4.56	4.67	4.71	11.11	3.77	9.65	1.06
CaO	5.44	7.34	7.56	7.12	6.84	6.20	6.28	6.57	6.81	6.39	7.20	7.77	6.79	13.23	5.57	14.48	49.30
Na_2O	2.62	3.15	2.99	3.10	2.43	1.59	2.66	2.72	2.47	2.77	2.20	2.38	3.37	1.78	4.59	1.01	0.07
K_2O	3.78	3.80	3.13	4.21	4.27	3.19	3.24	4.29	4.43	4.14	4.66	2.99	5.39	3.70	4.25	3.39	0.07
P_2O_5	1.04	1.77	1.44	1.69	1.96	1.89	1.33	1.53	1.85	1.64	1.77	2.08	1.95	3.21	1.65	3.64	0.02
CO_2	0.13	0.02	0.22	0.03	0.04	0.04	0.48	0.05	0.10	0.06	0.06	0.06	*	*	*	*	43.00
LOI	1.25	1.25	1.30	1.15	1.30	3.20	1.70	1.65	1.15	1.00	1.10	1.65	0.85	0.87	0.55	1.42	1.13
Sum	98.90	97.70	98.20	97.40	97.90	97.90	98.10	98.00	98.10	97.50	96.90	97.50	99.10	98.51	99.53	99.78	95.94
Trace elements																	
Ba	3470	4260	3870	4300	4690	5300	3680	4740	4680	4580	4130	4010	6360	9400	6753	10607	4110
Rb	103	68	77	96	86	75	51	77	63	84	94	82	72	66	57	52	0
Sr	1340	2680	2150	2470	2640	1310	1610	2640	2690	2460	2260	1830	6289	5190	4415	3123	29400
Cs	3	3	4	1	3	1	1	1	1	1	1	1	0	1	0	0.3	0
Th	9	8	5	8	11	11	40	6	8	9	9	60	16	32	8	67	3
U	5	3	2	3	3	4	3	3	3	3	2	2	5	6	2	8	0
Y	38	30	24	28	19	33	27	21	29	29	39	35	29	42	29	24	109
Zr	607	444	285	451	448	499	578	421	456	463	367	427	304	226	257	292	5
Hf	13	10	6	12	10	9	13	10	10	11	10	9	0	2	0	0	0
Nb	47	20	5	21	33	38	41	30	30	28	17	31	29	12	28	21	0
Ta	1	1	1	1	12	1	3	1	1	1	1	1	1.5	0.6	0.4	0.5	2.1
Pb	1	17	16	19	1	1	9	17	9	1	32	1	78	50	73	24	103
Cr	0	0	0	0	0	0	0	0	0	0	0	0	35	192	39	46	480
Ni	28	33	68	37	55	48	32	24	33	34	35	19	22	119	22	66	0
REEs																	
La	188	216	178	233	264	196	225	203	224	300	300	278	365	588	449	487	1060
Ce	322	355	295	392	441	345	385	341	370	483	465	458	750	1092	819	1058	2079
Pr	*	*	*	*	*	*	*	*	*	*	*	*	92	121	92	118	245
Nd	112	114	98	130	141	125	127	110	123	150	137	144	288	433	299	414	914
Sm	15.9	15.6	13.2	16.4	18.8	17.7	18.4	14.6	16.2	19.9	17.6	18.9	36	38	38	48	127
Eu	4.4	5.1	3.9	5.4	6	5.4	5.3	5.2	5.6	5.4	6.2	5.8	12.3	12.6	11.7	16	41
Gd	*	*	*	*	*	*	*	*	*	*	*	*	0	33	0	0	82
Tb	1.3	1	1	1.2	1.3	1.4	1.4	1	1.1	1.4	1.2	1.1	2.8	2.8	3	2.7	7.6
Dy	*	*	*	*	*	*	*	*	*	*	*	*	13	9.7	2.3	14.7	25
Ho	*	*	*	*	*	*	*	*	*	*	*	*	0	1.4	0	0	3.3
Er	*	*	*	*	*	*	*	*	*	*	*	*	3.5	3.7	3.8	3.6	8
Tm	*	*	*	*	*	*	*	*	*	*	*	*	0	0.4	0	0.24	0.8
Yb	3.5	2.7	2.5	2.6	2.8	2.9	3.6	2.4	2.7	3.7	3.3	4.1	1.7	2.4	2.1	1.6	4.6

Svecofennian Domain (Fig. 1). These rocks were selected as representatives of mafic rocks from SVB because of their central location and that they are relatively well studied (Andersson, 1997b). Furthermore, all mafic SVB rocks studied so far essentially show similar continental-arc geochemical signatures (Andersson et al., 2004a, and ref. therein). The TVZ consists of an east–west trending array of mafic to intermediate rocks within coarse porphyritic SVB granitoids, showing frequent evidence of mixing and mingling relationships, e.g. abundant mafic-hybrid magmatic enclaves (Andersson, 1991, 1997b; Wikström and Andersson, 2004). No direct age determination is available for the TVZ rocks, but the zone is surrounded by similar SVB granitoids to the west and east giving ages of ca. 1855 and 1815 Ma, respectively (Wikström, 1996; Andersson, 1997b). The mixing–mingling relations support a coeval nature for the mafic and granitoid magmatism. The SVB intrusions of this area were generally emplaced at ca. 5.5 ± 1 kbar as determined from contact metamorphic assemblages (Andersson, 1997a). The rocks in the TVZ comprise alkali-calcic granite–monzonite suites, calc-alkaline to tholeiitic gabbro, gabbronorites and dykes, as well as some ultramafic rocks (spinel–phlogopite–amphibole lherzolites/harzburgites). The rocks reported here represent variably evolved gabbroic magmas, of which the fine-grained mafic dyke (anal. 10) has the most primitive composition. They have evolved through fractionation of primarily clino- and orthopyroxene, olivine and minor plagioclase (Andersson, 1997b). Petrographically, the mafic rocks are mostly amphibolitic, with locally preserved clino- and orthopyroxene, hydrated late- to postmagmatically.

3.2. Central

The 1.8 Ga post-collisional intrusions in the central part of the Svecofennian domain are situated in the archipelago of SW Finland (the Åland Islands, Fig. 1) and close to the town of Turku. The oldest shoshonitic post-collisional rock in this area has been dated to 1815 Ma (Väisänen et al., 2000) and the youngest to 1770 Ma, but the intrusions center around 1800 Ma (see Eklund et al., 1998 and references therein). The oldest intrusions are elongated and concordant with the regional E–W orientation of the orogenic rocks in southern Finland. The emplacement level of these intrusions was determined to be 4.5 kbar based on the mineral assemblage in the contact aureole (Väisänen et al., 2000), this corresponds to the pressure of the regional metamorphism in the area (Väisänen and Hölttä, 1999). The younger intrusions form small upper crustal intrusions and cone sheets consisting of calc-alkaline high-Ba–Sr granites associated with shoshonitic lamprophyres and their plutonic equivalents. Petrographically, the monzonites and lamprophyres are characterized by panidiomorphic texture with glomeroporphyritic aggregates of biotite, hornblende and sometimes clinopyroxene in a fine-grained matrix of plagioclase, K-feldspar and sometimes quartz. In places the mafic aggregates contain abundant apatite. Typical accessory minerals are titanite, apatite, magnetite, and allanite.

3.3. East

In the easternmost part of the Svecofennian domain (NW of Lake Ladoga and SE Finland; Fig. 1) there are shoshonitic lamprophyres and their plutonic equivalents (apatite and magnetite-rich monzodiorites) intruding an area 100×50 km in size. These plutonic varieties fractionate into syenites and high-Ba–Sr granites. All rock types in this shoshonitic to slightly alkaline association are strongly enriched in P_2O_5, LREE and LILE; the degree of enrichment and volume of enriched rocks in the east exceed those described in the central part of the Svecofennian domain (Konopelko, 1997; Eklund et al., 1998; Konopelko et al., 1998; Ivashchenko, 1999).

One of the key areas in the east is the Elisenvaara group of intrusions. This group includes dykes and several small pipe-like bodies with a complex inner structure, comprised of a rock sequence ranging from ultramafic to leucosyenitic rocks. Rocks with SiO_2 contents close to 40% are considered to represent the parental magma composition, while rocks with lower SiO_2 contents represent apatite rich cumulates, based on geochemical modelling and the composition of the associated lamprophyric dykes which are considered equivalents of the ‘primary’ magmas of the plutonic rocks suite, containing more than 4% P_2O_5 and skeletal apatite (Konopelko et al., 1998; Eklund et al., 1998). From this extremely enriched (cf. Table 1) lamprophyric composition the magmas in the eastern complexes evolved by fractionation of early biotite, clinopyroxene, apatite, magnetite, sphene and allanite, and in later stages amphibole and feldspars. This is supported by the presence of phenocrysts of clinopyroxene, biotite, apatite and magnetite in both the plutonic rocks and the lamprophyres, which also occur accumulated in low SiO_2-rocks. The rocks are relatively little altered; feldspars are slightly sericitized and clinopyroxene is partly altered to amphibole. For a more detailed petrographic description, see Konopelko (1997), Eklund et al. (1998), and Konopelko et al. (1998).

4. Results

4.1. Geochemistry

Major and trace element data of mafic rocks from the three key areas are presented in Table 1. Most of these rocks are considered to represent near-liquid compositions in variable stages of magmatic evolution. Samples no. 5, 11 and 30 contain minor cumulus components. For the purpose of this regional comparison of source characteristics, however, this is not significantly affecting the interpretations. The data set was obtained from different laboratories and analytical techniques over time, but repeated analyses of several samples by variable techniques essentially show good reproducibility (cf. data in Rutanen et al., 1997; Eklund et al., 1998). Note that the Nb values (XRF) originally reported in Andersson (1997b) have been replaced by ICP-MS values of the same samples. The latter are preferred since abundances of Nb are close to the XRF detection limits. In addition to the data reported here, some data from Andersson (1997b), Eklund et al. (1998) and Rutanen et al. (1997) are used in the plots.

4.1.1. Major elements

K_2O contents in mafic rocks from the west are relatively low, and the samples plot mainly in the calc-alkaline field (Fig. 2a). With increasing SiO_2 contents, K_2O increase and they trend into the fields of high-K calc-alkaline and shoshonitic rock series. Mafic rocks from central and east are invariably rich in K_2O and plot within the field of shoshonitic rock series over the whole SiO_2 range. Many of the eastern rocks are characterized by low SiO_2 contents (Fig. 2), which is coupled with a generally higher abundance of biotite, apatite and magnetite. As noted above, rocks with ca. 40% SiO_2 are close to the parental magma composition, while those with $<40\%$ SiO_2 are cumulate-enriched (cf. Konopelko, 1997; Eklund et al., 1998).

Although the rocks from the west show the lowest K_2O/Na_2O ratios, they still mainly plot in the field for shoshonitic series (Fig. 3a). Rocks from the central and eastern areas have distinctly higher K_2O/Na_2O ratios, and the central rocks are concentrated almost entirely within the shoshonitic field (Fig. 3b). At corresponding K_2O contents, the rocks from the east show a range in Na_2O contents from 1% to 5%, where those lowest in Na_2O trend into the ultrapotassic field

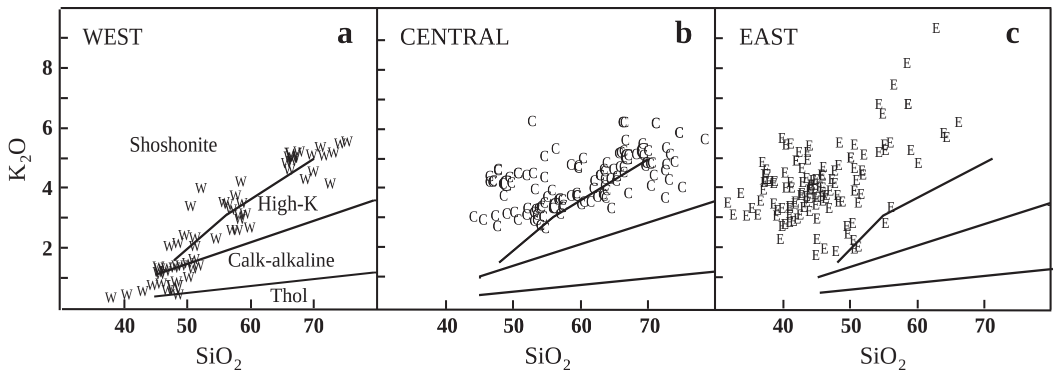

Fig. 2. (a–c) K_2O vs. SiO_2 variation diagrams of rock series from the studied key areas. Boundaries from Peccerillo and Taylor (1976). Data from this paper, Andersson (1997b), Konopelko (1997), Rutanen et al. (1997), Eklund et al. (1998). (a) WEST, shows rocks from the ca. 1.8 Ga Tjällmo-Vättern zone in TIB; (b) CENTRAL, shows rocks from the ca. 1.8 Ga coeval lamprophyre-high Ba–Sr granite intrusions in SW Finland; (c) EAST, shows lamprophyre–granitoid intrusions from the Lake Ladoga district in Russia.

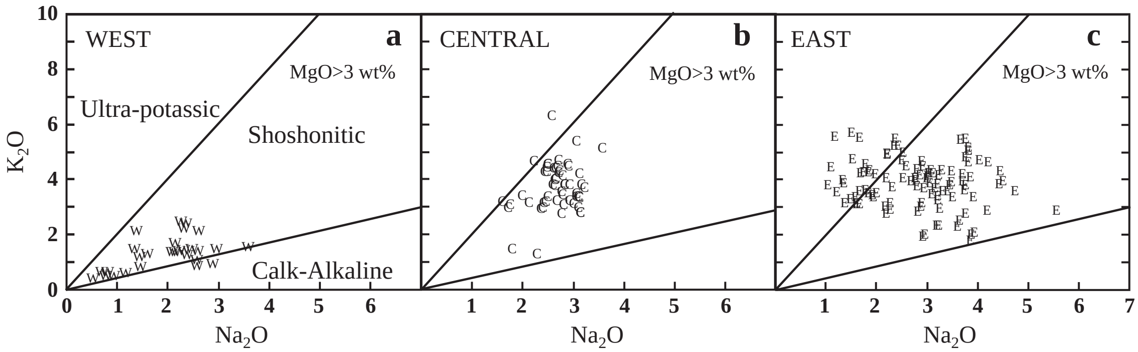

Fig. 3. (a–c) Plots of Na_2O vs. K_2O for the western, central, and eastern rocks discriminating calc-alkaline, shoshonitic and ultrapotassic rock series. The upper line represent $K_2O/Na_2O=2$ and the lower line $K_2O/Na_2O=0.5$ (Turner et al., 1996).

(Fig. 3c). In addition to fractionation–cumulation trends in each intrusion, the K_2O/Na_2O ratios are related to individual primary differences between the intrusions (cf. Konopelko, 1997; Eklund et al., 1998). According to the classification for ultrapotassic rocks used by Foley et al. (1987) ($K_2O>3$ wt.%, $MgO>3$ wt.%, and $K_2O/Na_2O>2$ wt.%), these rocks also classify as ultrapotassic. However, they do not contain typical alkaline mineral assemblages like alkaline mafic silicates or feldspatoids.

4.1.2. Trace elements

The trace element geochemical patterns for rocks of the three investigated areas are plotted in Fig. 4. In rocks from the west (Fig. 4a), the abundance of the least incompatible elements Zr, Ti and Y are mainly lower and equal to that of N-MORB. Some samples are somewhat enriched in Zr. In contrast, the more pronouncedly incompatible elements Th, Nb and Ce are enriched compared to N-MORB, however, with a distinct negative anomaly for Nb. In rocks from the central area abundances of Ti and Y spread around N-MORB, while Zr is enriched between 4 and 7 times relative to N-MORB. While Th and Ce are enriched up to 80 times compared to N-MORB, Nb is much less enriched, which produces a deep anomaly in the diagram (Fig. 4b). Rocks from the eastern area have abundances of Zr, Ti and Y that plot slightly above N-MORB. In some samples, Th and Ce are enriched to more than 100 times relative to N-MORB, with an even more pronounced negative Nb anomaly (Fig. 4c).

Oceanic tholeiitic volcanic-arc basalts (VAB) are characterized by a depletion of Nb, Zr, Ti and Y relative to N-MORB. Calc-alkaline and high-K calc-alkaline VAB (continental) contains higher concentrations of these elements, whereas Ti and Y remain below N-MORB in abundance (cf. Pearce, 1996). The trace element spider diagrams (Fig. 4) highlights the overall

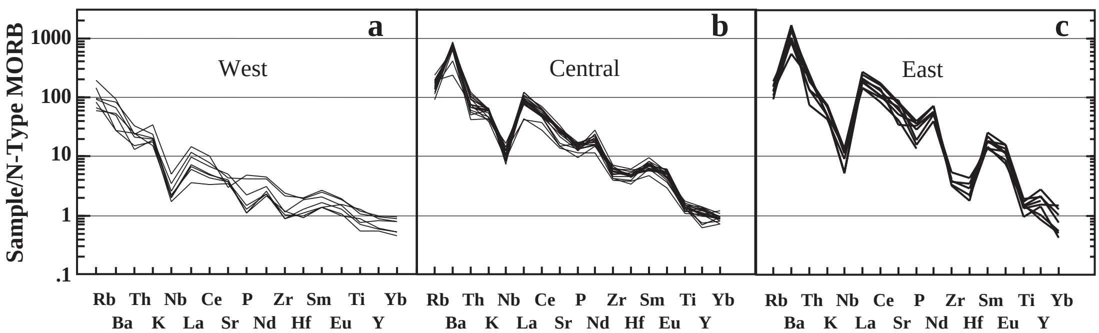

Fig. 4. (a–c) Primitive mantle-normalized (norm: Sun and McDonough, 1989) trace element diagram indicating a trough at Nb for rocks from all areas, a positive spike for Ba in the central and eastern, but negative in the western rocks. Note also the high Th concentration and the high Sm/Zr ratio in the east.

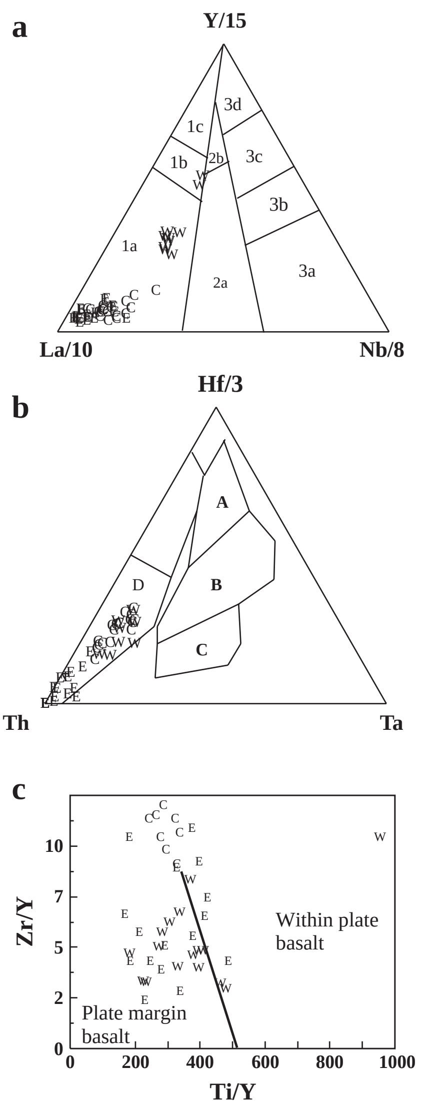

Fig. 5. Trace element discrimination diagrams indicating the principal differences between the sources in the three areas. (a) La/10–Y/15–Nb/8 by Cabanis and Lecolle (1989). Field 1 contains volcanic-arc basalts, field 2 continental basalts and field 3 oceanic basalts. The subdivision of the fields are as follows: 1A, calc-alkali basalts; 1C, volcanic-arc tholeiites; 1B is an area of overlap between 1A and 1C; 2A, continental basalts; 2B, back-arc basin basalts; 3A, alkali basalts from intercontinental rifts; 3B, 3C, E-type MORB (3B enriched, 3C weakly enriched), 3D, N-type MORB. (b) Th–Hf/3–Ta diagram after Wood (1980). The fields are: A, N-type MORB; B, E-type MORB and tholeiitic within plate basalts (WPB) and differentiates; C, alkaline WPB, WPB and differentiates; D, destructive plate-margin basalts and differentiates. (c) Zr/Y vs. Ti/Y diagram after Pearce and Gale (1977).

variation in enrichment level between the areas. In addition to the trough for Nb there are strong positive spikes for Ba in rocks from the central and eastern areas, but negative in west. Note also the high concentration for Th in the east. The eastern rocks are characterized by a considerable relative drop in Zr and Hf abundances, lacking in the other areas. This produces particularly high Sm/Zr ratios. In diagrams designed to unravel tectonic setting for basaltic rocks (Cabanis and Lecolle, 1989; Wood, 1980) using trace elements such as Zr, Nb, Y, La, Th, and Hf, the rocks from all three areas tend to plot in the fields for destructive margins (Fig. 5a, b), but in the Ti/Y–Zr/Y diagram (Fig. 5c) (Pearce and Gale, 1977) there is some overlap into the within-plate field, partly due to the elevated Zr/Y ratios of the rocks from the central area.

The extremely LREE-enriched character (more than 1000 times chondritic values) of the rocks from the central and eastern areas is displayed in Fig. 6. Rocks in the west only show LREE values 50–100 times chondrite. The abundance of the HREEs varies much less, between 10 and 40 times chondrite, but also increasing eastwards.

In order to characterize different types of sources for the mafic rocks of the three regions, some trace element ratios were utilized. In the Nb/Ba vs. Nb/Zr diagram (Hooper and Hawkesworth, 1993) in Fig. 7a, rocks from all areas plot close to the field for subcontinental lithospheric mantle and far away from oceanic island basalt (OIB) and E-MORB sources.

To illustrate different types of mantle enrichment and the effects of melting such sources, Sun and Stern

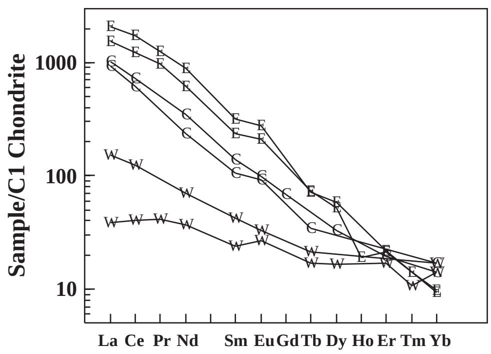

Fig. 6. Normalized rare earth element concentration diagram indicating the difference in enrichment for rocks from the west (w), center (c) and east (e). The samples are the same as in Fig. 4 (norm: chondrite, Sun and McDonough, 1989).

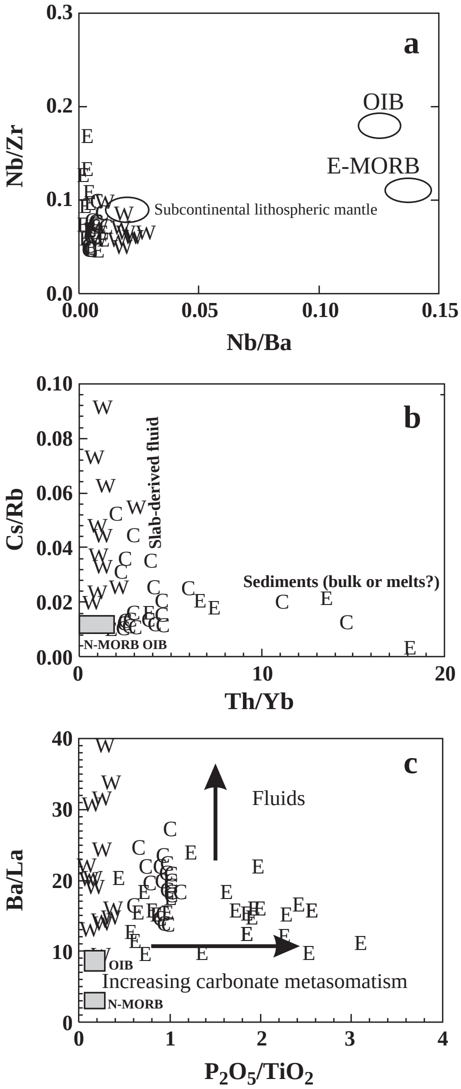

Fig. 7. (a) Nb/Zr vs. Nb/Ba diagram after Hooper and Hawkesworth (1993) indicating that rocks from all areas stem from a subcontinental, lithospheric, mantle source. (b) Cs/Rb vs. Th/Yb diagram after Sun and Stern (2001) discriminating magmas derived from a source enriched by slab-derived fluids or by sediments. (c) Ba/La vs. P_2O_5/TiO_2 diagram discriminating magmas derived from fluid enriched sources and sources affected by carbonate metasomatism.

(2001) used Ba/La, Pb/Ce, Cs/Rb, and U/Th ratios to identify fluid enriched sources. In these ratios the numerator has higher fluid mobility than the denominator, which means that magmas derived from fluid enriched sources are enriched in Ba, Pb, Cs, and U, relative to La, Ce, Rb and Th, respectively. Woodhead et al.

Table 2
U–Pb mineral analyses from the Elisenvaara syenite, Russian Karelia

Fraction	Weight (mg)	U conc. (ppm)	Pb conc. (ppm)	$^{206}Pb/^{204}Pb$ (meas.)	$^{208}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	±2 S.E. (%)	$^{207}Pb/^{235}U$	±2 S.E. (%)	$^{207}Pb/^{206}Pb$	±2 S.E. (%)	Corr.	$T_{206/238}$	$T_{207/235}$	$T_{207/206}$
(A) +4.3/abr 12 h	0.48	253.7	87.9	1659	0.1708	0.3049	0.43	4.627	0.48	0.1101	0.22	0.89	1715	1754	1801
(B) +4.3/abr 6 h	0.46	318.4	110.2	1002	0.1571	0.3011	0.41	4.575	0.48	0.1102	0.23	0.87	1697	1745	1803
(C) +4.3/abr 3 h	0.52	348.9	123.0	1273	0.1665	0.3076	0.40	4.669	0.44	0.1101	0.19	0.90	1729	1762	1801
(D) +4.3	0.64	382.8	137.9	964	0.2594	0.2904	0.47	4.411	0.52	0.1102	0.21	0.91	1644	1715	1802

(1998) suggested that if the mantle is enriched by melts from subducted sediments, it is reflected in the Th/Yb ratio in mafic magmas. In the Th/Yb vs. Cs/Rb diagram in Fig. 7b rocks from the western area show a trend of increasing relative Cs, while the rocks from the central area show a variable relative enrichment in Cs or Th. The rocks from east are least enriched in Cs but develop a trend towards Th enrichment.

Melts generated from mantle sources subjected to carbonate metasomatism are believed to be characterized by high P_2O_5/TiO_2 ratios (Wyllie, 1995). In Fig. 7c, the rocks with ≥ 40% SiO_2 from the east form a trend towards high P_2O_5/TiO_2 ratios, while the rocks from the west do not show any particular P_2O_5 enrichment compared to OIB and N-MORB. Relatively high Ba/La ratios characterize all three areas, but tend to increase westwards.

4.2. Isotopes

4.2.1. U–Pb data

Four multi-grain zircon fractions were separated from a syenite of the Elisenvaara complex in Russian Karelia. The mineral separation techniques and chemical purification techniques were those described by Vaasjoki et al. (1991). The density fraction >4.3 g/cm^3 consists mainly of euhedral-subhedral, clear, fine-grained (<70 μm) zircons which exhibit primarily simple prismatic and pyramidic faces; however, higher order crystal faces are also present. There is also a coarser, turbid, brownish zircon variety in the separate. From the clear zircon variety, crystals without visible inclusions or cores were handpicked into four fractions. The turbid brownish variety was not analyzed. Three of the fractions were subjected to air-abrasion for 3, 6, and 12 h, respectively. The results are given in Table 2.

The four fractions plot on a discordia with an upper intercept of 1800 ± 6 Ma (MSWD = 0.24), and with a lower intercept of −47 Ma. The least discordant fraction is the one abraded only for 3 h, suggesting that the crystals of this fraction were of the best quality, while the unabraded fraction is most discordant, as expected. The negative lower intercept, however, indicates that the lead loss was recent. If the discordia is constrained through the origin (Fig. 8), an age estimate of 1801 ± 4 Ma (MSWD = 0.28) is obtained. This is the preferred age.

4.2.2. Nd and Sr data

Four samples from the Åva complex and one from the Lemland complex in the central area (Fig. 1) were analyzed for Sr and Nd isotopic compositions in the Laboratory for Isotope Geology (LIG) at the Swedish Museum of Natural History in Stockholm. Nd and Sm concentrations were determined by isotope dilution,

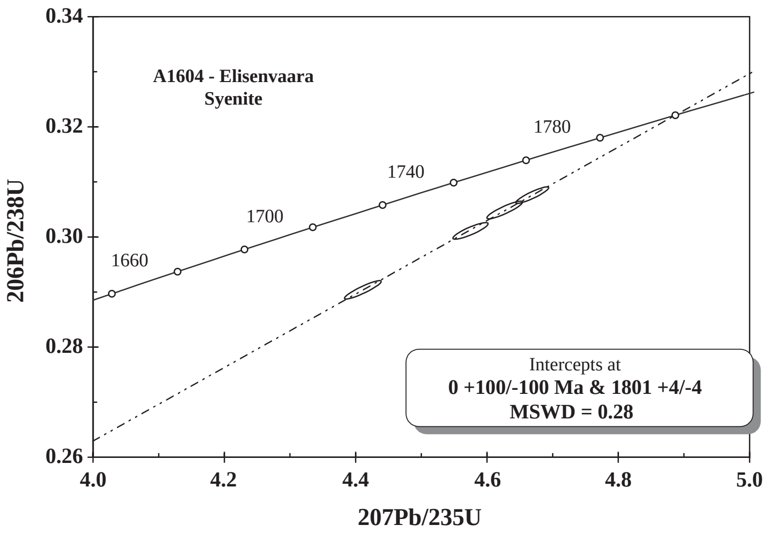

Fig. 8. U–Pb zircon concordia diagram for the Elisenvaara intrusion. A discordia forced through the origin results in an upper intercept age of 1801 ± 4 Ma; the preferred age of intrusion.

Table 3

Nd and Sr isotopic data for some ca 1.8 Ga plutonic shoshonitic complexes in the Fennoscandian Shield

Region/sample	Rock type	Age	Sm	Nd	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	$\pm 2\sigma M$	ε_{Nd} (T)	ε_{Nd} (1800)	T_{CHUR}	T_{DM}	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	$\pm 2\sigma M$	$^{87}Sr/^{86}Sr$ (T)	$^{87}Sr/^{86}Sr$ (1800)	Lab.
Central																	
Åva																	
AvaGr1	Granite	1804	19.7	150	0.0795	0.511274	5	0.5	0.4	1.80	2.00	0.241	0.708658	7	0.7024	0.7024	LIG
AvaLp1	Lamprophyre	1804	23.8	204	0.0704	0.511176	4	0.7	0.6	1.79	1.97	0.076	0.704867	8	0.7029	0.7029	LIG
AvaLp2	Lamprophyre	1804	21.5	186	0.0705	0.511164	4	0.4	0.4	1.80	1.99	0.113	0.705652	8	0.7027	0.7027	LIG
AvaMz2	Monzonite	1804	17.6	135	0.0786	0.511257	4	0.4	0.3	1.81	2.00	0.067	0.704503	9	0.7028	0.7028	LIG
Lemland																	
49NOE	Mafic fragment	1775	13.0	73.8	0.1065	0.511582	3	−0.1	0.2	1.82	2.06	0.491	0.715694	9	0.7032	0.7030	LIG
02NOE85	Mafic Enclave	1775	15.4	109	0.0854	0.511360	5	0.4	0.8	1.78	1.99	0.182	0.707051	8	0.7024	0.7024	GTK
PUMX86.6	Granite	1775	7.12	53.2	0.0809	0.511256	4	−0.6	−0.2	1.85	2.04	0.837	0.722495	9	0.7011	0.7008	GTK
PUMX86.7	Granite	1775	3.22	25	0.0765	0.511222	4	−0.3	0.1	1.82	2.01	0.263	0.709556	10	0.7028	0.7027	GTK
PUMX86.9	Granite	1775	4.59	35.1	0.0789	0.511271	4	0.1	0.5	1.80	1.99	2.696	0.762994	9	0.6942	0.6932	GTK
PUMX86.11	Mafic Enclave	1775	9.26	65.2	0.0858	0.511344	4	0.0	0.4	1.81	2.01	0.247	0.709152	10	0.7029	0.7028	GTK
East																	
Elisenvaara																	
ElLeLa	Leucoladogite	1800	45.6	386	0.0714	0.511135	5	−0.4	−0.4	1.85	2.03	0.031	0.704162	15	0.7034	0.7034	GTK
ElMeLa	Mesoladogite	1800	36.6	321	0.0688	0.511085	6	−0.8	−0.8	1.88	2.05	0.037	0.704266	15	0.7033	0.7033	GTK
ElNe	Nevoite	1800	51.1	437	0.0707	0.511118	5	−0.6	−0.6	1.86	2.04	0.038	0.704284	15	0.7033	0.7033	GTK
ElSy	Syenite	1800	14.2	128	0.0674	0.511089	5	−0.4	−0.4	1.85	2.03	0.029	0.704147	13	0.7034	0.7034	GTK
Kalto																	
KaLaCont	Lamprophyre	1800	64.9	513	0.0764	0.511192	5	−0.4	−0.4	1.86	2.05	0.035	0.704382	9	0.7035	0.7035	GTK
KaLaWide	Lamprophyre	1800	69.5	557	0.0754	0.511189	5	−0.3	−0.3	1.85	2.03	0.046	0.704684	23	0.7035	0.7035	GTK

Samples analyzed at: LIG = Laboratory for Isotope Geology at the Museum of Natural History in Stockholm, Sweden; GTK = Geological Survey of Finland, Espoo, Finland. All analyses are corrected for fractionation to $^{146}Nd/^{144}Nd = 0.7219$ and $^{86}Sr/^{88}Sr = 0.1194$. The $^{87}Rb/^{86}Sr$ ratios for Åva and Lemland samples were calculated using ICP-MS concentration data; Elisenvaara and Kalto from ID concentrations.

Errors are given as 2× the standard error of the mean (=$2\sigma_M$) in last significant digits. Estimated errors in $^{147}Sm/^{144}Nd$ are <0.4% (GTK) and <0.5% (LIG), and for $^{87}Rb/^{86}Sr$: 0.5% (GTK) and <1% (LIG). ε_{Nd} is calculated relative to a model chondritic mantle with present day values for $^{143}Nd/^{144}Nd = 0.512638$ and $^{147}Sm/^{144}Nd = 0.1966$. T_{DM} model ages are calculated relative to the Depleted Mantle model of DePaolo (1981).

using a mixed ^{150}Nd–^{147}Sm spike. The Nd and Sr isotopic compositions were measured on a Finnegan MAT 261 in static mode and were corrected for fractionation to ^{146}Nd/^{144}Nd = 0.7219. Estimated errors for ^{147}Sm/^{144}Nd are < 0.5%. The Sr isotopic compositions were measured on unspiked samples and reported ^{87}Rb/^{86}Sr ratios are calculated from ICP-MS concentration data. The precision of ^{87}Rb/^{86}Sr is estimated to be better than 1%. The Sr isotope ratios were normalized to ^{86}Sr/^{88}Sr = 0.1194. The average of measured ^{87}Sr/^{86}Sr ratios on the SRM 987 Sr standard during the course of sample analyses was 0.710213 ± 15 (n = 18; 1σ), which is 0.000027 lower than the 'preferred' value (0.710240; Gladney et al. 1990), and the sample Sr measurements were normalized to this 'preferred' standard value. The average measured ^{143}Nd/^{144}Nd ratio of the LaJolla Nd-standard during the same period was 0.511764 ± 12 (n = 15; 1σ). The sample Nd measurements were normalized to the average standard value obtained at the GTK laboratory (0.511851; see below) to allow direct comparisons (cf. the similar procedure applied by e.g. Mitchell et al., 1995).

Five additional samples from the Lemland complex, all samples from the Elisenvaara complex and the Kalto dyke swarm were analyzed at the Geological Survey of Finland (GTK), Espoo. The samples were spiked with a ^{150}Nd–^{149}Sm tracer and Sm and Nd concentrations were determined by isotope dilution. The error in ^{147}Sm/^{144}Nd is estimated to be less than 0.4%. The isotopic ratios were measured on a VG Sector 54 mass spectrometer in dynamic mode and were normalized to ^{86}Sr/^{88}Sr = 0.1194 and ^{146}Nd/^{144}Nd = 0.7219. For a more detailed description of the methods, see e.g. Rämö et al. (1996). ICP-MS data for Rb and Sr concentrations were used to calculate the ^{87}Rb/^{86}Sr ratio. A long-term average for the LaJolla standard was ^{143}Nd/^{144}Nd = 0.511851 ± 6 (1 S.D.; n = 40), which is close to general averages obtained for this standard (e.g. Raczek et al., 2003), and no correction was applied. For the SRM987 standard ^{87}Sr/^{86}Sr = 0.710229 ± 13 (1 S.D.; n = 24) was obtained and the sample Sr measurements were normalized to the 'preferred' value for this standard (see above). Measurements on standards and the consistency of the results confirm that the corrections made are justified.

All samples are strongly LREE-enriched and have very low ^{147}Sm/^{144}Nd ratios (all below 0.0854), except a mafic fragment (0.1065) from the Lemland complex (Table 3), yielding steep and subparallel slopes in the time-integrated evolution (Fig. 9). The calculated initial ε_{Nd} values show narrow ranges, tending to be lowest in the east (−0.8 to −0.3), with the central intrusions showing slightly but significantly higher values (−0.6 to +0.8).

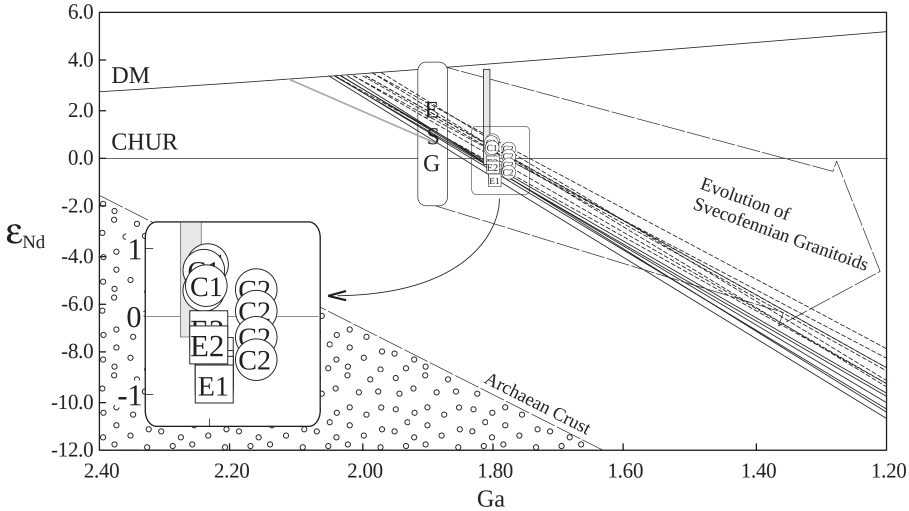

Fig. 9. Evolution of ε_{Nd} with time for the measured samples of this study. The ε_{Nd} range of the mafic rocks in the Tjällmo-Vättern zone (Andersson, 1997b) in the west is given as a bar. The evolutionary field for early Svecofennian (1.95–1.86 Ga) and Archaean granitoid crust is given for comparison (taken from Andersson et al., 2002).

The $^{87}Rb/^{86}Sr$ ratios are low (below 0.9), with the exception of one of the Lemland granites (2.696). Initial $^{87}Sr/^{86}Sr$ (I_{Sr}) ratios are in the narrow range 0.7033–0.7035 for the east, but significantly lower for the central rocks, below 0.7032. Calculated I_{Sr} for two Lemland granites are very low (0.7011 and 0.6942) and unreal. Unrealistically low I_{Sr} values in the Åland rocks may result from thermal and/or fluid impact of the 1575 Ma old Åland rapakivi batholith.

The Nd and Sr data from the central and eastern areas, presented here (Fig. 10), are compared with mafic rocks from a section of the SVB in the southern TIB in the west (Figs. 10 and 12; Andersson, 1997b). The initial ε_{Nd} ratios of these rocks are more variable and trend to significantly more depleted compositions (−0.3 to +3.5), with less pronounced LREE-enrichments ($^{147}Sm/^{144}Nd$ ratios between 0.1014 and 0.1635). Kononova et al. (1999, 2000) presented similar Sr and Nd isotopic results from rocks in the east.

4.2.3. O and C isotopic data

To obtain information on the origin of the carbonates encountered in the rocks from the east, O and C isotopes were analyzed from carbonate vesicles in the marginal zone and in the central part of a 50 cm wide lamprophyric dyke located at Kalto in Lake Ladoga. Measurements of oxygen yielded $\delta^{18}O$ values of 11.35 and 16.29 (relative to SMOW), and of carbon $\delta^{13}C$ values of −9.11 and −8.24 (relative to PDB), respectively.

5. Discussion

5.1. Age

The obtained age estimate for the crystallization of the Elisenvaara leucosyenite (1801 ± 4 Ma) is within the range of other post-collisional intrusions in southern Finland and Russian Karelia (1815–1770 Ma;

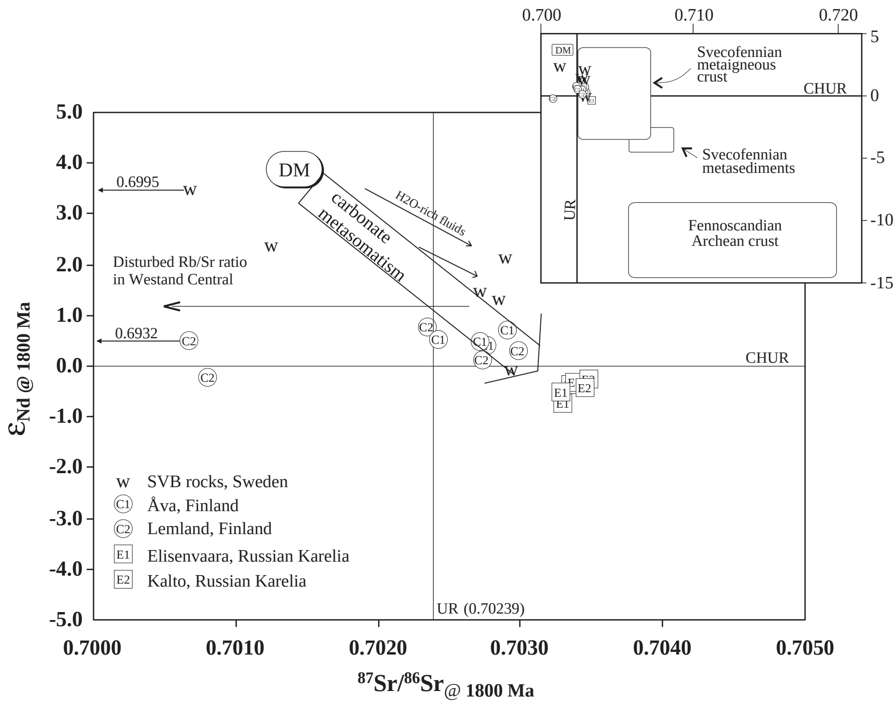

Fig. 10. The initial isotopic composition of the samples in ε_{Nd} (1800 Ma) vs. $^{87}Sr/^{86}Sr$ (1800 Ma). Disturbance of the Rb/Sr system has affected several samples from the western and central areas, lowering their calculated initial values. The inset figure also gives the compositional fields for three important Fennoscandian crustal components (data as compiled in Andersson et al., 2002, and unpublished data by the authors). Arrows schematically give the proposed direction of isotopic fluid/melt enrichment in the mantle source regions in the western, and central + eastern areas, respectively.

Eklund et al., 1998; Väisänen et al., 2000, and ref. therein), and thus supports its assignment to this group with respect to timing. It is also within error of the previously estimated 1775 ± 65 Ma Pb–Pb age (Ivanikov et al., 1996). U–Pb zircon ages determined of genetically related intrusions in the east are: Vuoksi 1802 ± 17 Ma (Konopelko and Ivanikov, 1996), Luonteri 1802 ± 22 Ma (Korsman et al., 1984), Parkkila 1794 ± 5 Ma (Simonen, 1982), Pirilä 1815 ± 7 Ma (Vaasjoki and Sakko, 1988) and Eräjärvi 1795 ± 5 (Nykänen, 1988). Precise geochronological data are lacking from the Petravaara (poor resolution Pb–Pb evaporation age around 1750 Ma; Geologinen tutkimuslaitos, 1966), Ojajärvi intrusion (poor resolution K–Ar amphibole and Pb–Pb zircon evaporation ages center around 1800 Ma; D. Konopelko, unpubl.), and the Kalto lamprophyric dyke swarm, which on petrological grounds are included in the post-collisional suite (Eklund et al., 1998). For orientation, see Fig. 1.

5.2. Nature of the mantle source regions

5.2.1. Geochemical evidence

The trace element spider diagram patterns for all rocks in the studied transect mimic patterns typical for continental arc rocks, with increasing, even extreme, enrichments eastwards. The limited relative enrichment of particularly Ti and partly Zr, combined with the pronouncedly negative Nb–Ta anomalies do not lend support for a within-plate, or constructive margin, character of these rocks.

The geochemistry of rocks from collision zones range from volcanic-arc to within-plate in character, depending on if the source experienced a previous subduction enrichment, or not (e.g. Pearce et al., 1990). A diagnostic feature of collision-related, mantle-derived rocks is an enrichment of Zr relative to Ti and Y (Pearce, 1996). Particularly the rocks from the central area display high Zr/Y, while the Ti/Y ratios indicate characteristics transitional between plate-margin and within-plate basalts with dominance for the former in the rocks from the east and west. However, the variable but low Ti/Y ratios in the east can in part be related to oxide fractionation (Eklund et al., 1998).

The distinctive enrichment levels of LREE and LILE for each area, irrespective of fractionation degree (e.g. Mg#), increase eastwards (Fig. 11a, b) and should reflect variable source enrichment. Accordingly, the mantle source regions for the mafic magmas are mildly enriched in the west, strongly enriched in the center and extremely enriched in the east.

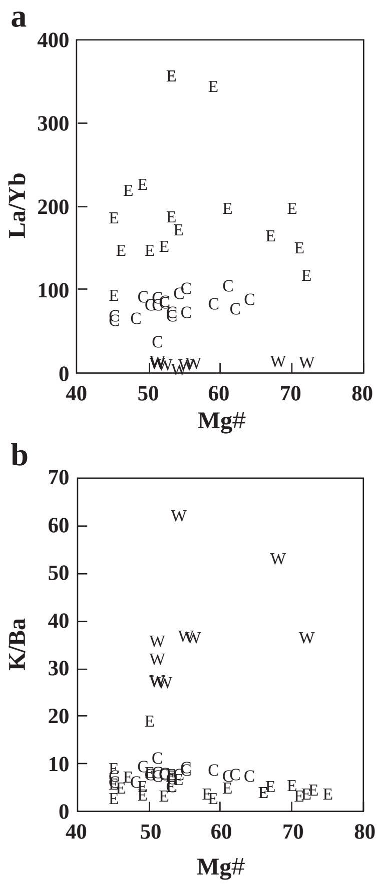

Fig. 11. (a) La/Yb vs. MgO/(MgO + FeO) showing that the LREE enrichment is unrelated to the degree of fractionation of the samples. (b) K/Ba vs. MgO/(MgO + FeO) showing that the LILE enrichment levels (here represented by Ba) are not related to fractionation level.

Additionally, the extreme enrichment in LREE/(HREE-Y) in rocks from the central and eastern areas (Figs. 4, 6 and 7) suggests relatively deep mantle sources in the garnet lherzolite stability field (cf. e.g. Doe, 2002). In contrast, the rocks from the west have moderately enriched LREE and flat HREE patterns, suggestive of shallower melting within a spinel lherzolite mantle.

The individual, but subparallel, variation in enrichment level in the eastern rocks is partly due to some amounts of fractionation of particularly clinopyroxene, apatite (note variable P in Fig. 4), and magnetite (note variable Ti; cf. Eklund et al., 1998), but most probably also some variations in mantle enrichment and degree of melting (note variable Y and REE). In the west, the spread in enrichment relative to MORB can also be correlated with variable fractionation of olivine, clinopyroxene (± opx), and plagioclase (note variable Sr, Fig. 4) (cf. Andersson, 1997b), but most likely some variations are also inherited from the mantle source.

The strong relative enrichment in Cs over Rb in the rocks from the west (Fig. 7b) is suggestive of sources enriched by slab-derived H_2O-rich fluids (cf. e.g. Altherr et al., 1999; Sun and Stern, 2001; Melzer and Wunder, 2000). In contrast, the enrichment of Th over Yb in rocks from the east supports mantle wedge enrichment by melting of subducted sediments (cf. e.g. Woodhead et al., 1998, Becker et al., 2000). Rocks from the center trend in both directions indicating derivation from partly fluid- and partly melt-enriched mantle sections.

Barium is extremely enriched in the central and eastern rocks and relatively high Ba/La ratios characterizes all three areas, but tend to increase westwards. Since Ba has a higher fluid mobility than La, it is interpreted that the Ba enrichment over La indicates that fluids enriched the sources. Fluids expelled from pelagic sediments returned to the mantle by subduction may enrich the mantle wedge in Ba, radiogenic Sr and unradiogenic Nd (Fitton et al., 1991; Thirlwall et al., 1996; Becker et al., 2000). Based on mass balance considerations, Hawkesworth et al. (1995) suggested that the elevated Ba/La and La/Nb in basaltic rocks from the Basin and Range province cannot realistically be attributed to crustal contamination of MORB-type magmas, and that such rocks are connected to source regions in the subcontinental lithospheric mantle. Because of the overall high Ba/La ratio in Fig. 7c, it is here interpreted that the high Ba-content in the central and eastern rocks may be a result of enrichment by slab-derived fluids or melts. High Ba together with Sr and LREE is also typical of mantle carbonates and carbonated mantle xenoliths (e.g. Ionov, 1998). Melts generated from such sources would yield high Ba, as well as high Ba/Rb ratios. This contrasts with the rocks in the west, which have lower Ba/Rb related to dominantly H_2O-metasomatized mantle (cf. Becker et al., 2000).

Doe (2002) surveyed the variations in Ba/Ce and Ce/Yb ratios in mantle-derived rocks and concluded that strongly Ce/Yb enriched rocks derive from CO_2-metasomatized garnet lherzolites, while Ba/Ce enriched rocks typically should derive from amphibole–spinel harzburgites (with some phlogopite), mainly metasomatized by H_2O-dominated fluids (cf. Hartmann and Wedepohl, 1993). The latter is applicable to the rocks in the west and supported by some occurrences of associated spinel–phlogopite–amphibole lherzolites/harzburgites. The rather high Ba/La ratios also in the central and eastern rocks indicate the presence of some hydrous mafic silicates also in the deeper source regions of these magmas. In general, O'Brien et al. (1995) related high Ba contents with the presence of phlogopite in the mantle source lithologies. Conceição and Green (2004) showed experimentally that shoshonitic magmas can be derived by decompression melting of phlogopite + pargasite lherzolitic mantle sources.

Carbonatitic mantle melts are able to dissolve and transport major amounts of P (Green and Wallace, 1988; Barker and Wyllie, 1992; Rudnick et al., 1993; Wyllie, 1995) and the high P_2O_5/TiO_2 ratios in the rocks from the center, and particularly from the east, suggest that their mantle source region was subjected to strong carbonate metasomatism (Fig. 7c). Such a signature is lacking in the west. According to Dupuy et al. (1992), small amounts of CO_2 (< 1%) in the mantle source may reduce the Hf/Sm and Zr/Sm ratios and increase the Zr/Hf ratio for subsequently generated melts. The pronouncedly low Zr/Sm ratios in rocks from the east, as well as increasing Zr/Hf ratios eastwards (Fig. 4c), are thus also in agreement with mantle sources overprinted by strong carbonate metasomatism (cf. Eklund et al., 1998).

The latter is corroborated by the results of Blundy and Dalton (2000), who showed that the LREE are strongly partitioned into melts generated from mantle lherzolites affected by carbonate metasomatism, while Si, Al, Ga, Ti, Zr are retained in the solid phases, and further supports carbonate metasomatism in the source regions of the rocks from the central and eastern areas (cf. Figs. 4 and 6).

The importance of carbonates in the formation of post-collisional magmas is further supported by the presence of the coeval ca. 1.7–1.8 Ga Halpanen carbonatite dyke in SE Finland (Fig. 1) (Puustinen and

Karhu, 1999). This dyke consists of almost pure calcite carbonatite (alvikite) in its center but becomes more apatite-rich towards its contacts. The carbonatite is strongly enriched in Y, Sr, Ba and REE (Table 1). Stable isotope measurements from the alvikite have yielded $\delta^{18}O$ (SMOW) +15.5 and $\delta^{13}C$ (PDB) -12.3 (weighted average of 6 analyses: Puustinen and Karhu, 1999). Puustinen and Karhu (1999) concluded that the $\delta^{13}C$ value for the carbonatite is outside the range for primary mantle melts, why they suggested either a crustal source or extensive crustal differentiation of mantle derived material.

The measured stable isotopic composition of carbonates in the vesicles of the Kalto lamprophyre show $\delta^{13}C$ values in the lower end of the compositional range considered compatible with the primary mantle compositions (cf. e.g. Deines and Gold, 1973; Mitchell, 1986), while the $\delta^{18}O$ values are considerably enriched in ^{18}O in comparison with the field for primary mantle composition. The latter is, however, a rather common phenomenon, e.g. in kimberlites (Mitchell, 1986), and may be related to several processes, e.g. loss of isotopically light water during emplacement, exchange with magmatic or hydrothermal water with high ^{18}O contents, or influx of meteoric water at low temperature (Deines, 1989).

The diversity of rock types found in the central and eastern intrusions has been modelled to result mainly from crystal fractionation (Konopelko, 1997; Eklund et al., 1998). Variations in the ratios of alkalis, e.g. K_2O/Na_2O, may also partly derive from variations in, e.g. the amphibole/phlogopite ratios in the mantle source regions of the primary magmas. The extreme initial enrichment in Ba, Sr, LREE and moderate in some HFSE compared with average crustal composition tend to efficiently mask any traces of possible contamination with crustal material, even up to 30–40 wt.% of the bulk rock chemistry (cf. e.g. Hegner et al., 1998, Wenzel et al., 2000).

5.2.2. Isotopic evidences

5.2.2.1. Nd isotopes. The initial ε_{Nd} (at 1.8 Ga) of the individual samples reported here from central and east are plotted in Fig. 9 together with their respective individual time-integrated evolution lines. The evolution lines are all rather steeply inclined, due to their exceptionally low $^{147}Sm/^{144}Nd$ ratios (below 0.085, except for one mafic autolith from Lemland 0.1065; Table 3). Noticeable is that all analyzed rocks are encompassed by the evolutionary field of the early Svecofennian (1.95–1.86 Ga) calc-alkaline metaigneous crust. The western rocks are characterized by positive initial $\varepsilon_{Nd}(t)$ values within the range 0 to +3.4 and model ages T(DM) of 1.88–2.22 Ga (Andersson, 1997b), while samples from the center and east have initial ε_{Nd} values between -1 and +1, and model ages T(DM) of 1.97–2.06 Ga.

The absolute time for the enrichment of the mantle sources of the mafic magmas is not known, but is not likely to be very much older than the magmatism itself. The very strongly LREE-enriched mafic rocks from the central and eastern areas, with T(DM) ages essentially younger than 2.0 Ga, should derive from mantle sources with correspondingly strong LREE enrichment, i.e. low Sm/Nd ratios. However, the enrichment in the source is presumably somewhat less strong than the partial melts derived from it. Amphibole–phlogopite-bearing spinel lherzolites (-harzburgites) found in association with the mafic rocks in the west have $^{147}Sm/^{144}Nd$ values of ca. 0.135, and may represent enriched mantle pieces (Andersson, 1997b). Because the rocks in the center and east stem from more strongly enriched sources, this can be taken as a maximum value for the $^{147}Sm/^{144}Nd$ ratio in the source. Using this value and assuming an ε_{Nd} (1.80) of 0 results in a maximum age of depleted mantle enrichment of ca. 2.19 Ga. Kononova et al. (2000) calculated that the mantle enrichment in the east took place at ca. 2 Ga.

More likely, the mantle enrichment occurred during the preceding early Svecofennian ((2.1–)1.95–1.86 Ga) arc-magmatism, when systems of arcs and intervening basins were assembled and accreted to the Archaean craton nucleus in the NE (cf. Nironen, 1997, and ref. therein), including slab-enriched mantle sections. Such enrichment is shown by the ca. 1.87 Ga Puutsaari intrusion in the East (Konopelko and Eklund, 2003). When recalculated to 1.90 Ga, using their measured $^{147}Sm/^{144}Nd$ ratios, the rocks from the center and east yield ε_{Nd} values between +0.9 and +2.4, while the rocks from the west reach the DM curve. Even if the sources would have had somewhat higher Sm/Nd ratios, they would already 100 m.y. earlier have been mildly depleted in the central and eastern areas, ranging up to strongly depleted in the west. Assuming that the mantle was essentially of DM-type before enrichments at ca. 1.9 Ga, fluids and melts carrying unradiogenic Nd, in addition to low Sm/Nd (and more radiogenic Sr; see below), must have been added to the sources in order to decrease their ε_{Nd} (1.9) ratios. However, it is also possible that additional low-ε_{Nd} source enrichment occurred in immediate connection with the magmatism itself at ca. 1.8 Ga. Thus, this enrichment should have consisted of fluids and/or melts derived from newly formed slab-segments probably containing a minor component of Archaean material, presumably mostly in the form of continental detritus in Svecofennian sediments.

5.2.2.2. Sr isotopes and combined Nd–Sr considerations. Disturbance of the Sr isotopic system in some of the samples is clearly indicated by very low calculated I_{Sr} (1.80) (Fig. 10). A similar disturbance is also noted for some of the mafic rocks in the west (cf. Andersson, 1997b). However, most of the data plot at I_{Sr} values that are realistic, although minor disturbances cannot be ruled out.

The rocks from the east show a very little spread in I_{Sr} around 0.7035, while rocks from the central and western areas plot at slightly, but distinctly, lower I_{Sr} (0.7024–0.7030). The values are all at or somewhat above that of the Uniform Reservoir (UR) at 1.80 Ga (Fig. 9) (cf. Faure, 1986). In the east, the relatively low I_{Sr} values is a unique feature found in the shoshonitic intrusions in the area, while surrounding Svecofennian anatectic granites have much higher I_{Sr} (1850) values above 0.7070 (Konopelko, 1997).

On the I_{Sr} vs. ε_{Nd} diagram (Fig. 10) the analyses of rocks from the east form a distinct cluster at higher I_{Sr} and lower ε_{Nd}, while rocks from the center and west overlap. The latter, however, trend towards more depleted ε_{Nd} compositions. Assuming that the isotopic compositions are due primarily to enrichment levels in the mantle source regions, as indicated by the very high abundances of LILE and LREE in the central and eastern rocks (higher than can be accounted for by any known crustal rocks available as contaminants (see above; cf. Hegner et al., 1998; Wenzel et al., 2000)), the sources of rocks from the east are also most enriched in terms of Sr and Nd isotopic compositions. For comparison, Beccaluva et al. (2004) reported amphibole–phologopite-bearing harzburgitic mantle xenoliths with strongly enriched Nd and Sr isotopic signatures from subduction-related metasomatism. During fluid/melt carbonate metasomatism the mantle would become progressively enriched in more radiogenic Sr and less radiogenic Nd (cf. Xu et al., 2003), away from an assumed original DM composition (big arrow in Fig. 10). In the west, where strongly depleted ε_{Nd} compositions partly prevail, enrichments by H_2O-dominated fluids (cf. geochemistry above) will cause a shift in Sr isotopic compositions, but little change in ε_{Nd}. The metasomatizing agents, melts and/or fluids, apparently were derived from subducting oceanic crust and sediments (cf. Peccerillo, 1990; Beccaluva et al., 1991). Thirlwall et al. (1996) proposed that H_2O-dominated slab-derived fluids could explain mantle enrichment in mainly LILE (and lesser LREE), where Sr isotopic enrichment derive from oceanic crust and Nd isotopic enrichment from subducted sediments. As inset in Fig. 10, boxes with the compositional ranges for the Archaean Fennoscandian continental crust, the newly formed early Svecofennian metaigneous crust, and Svecofennian metasediments are shown. These represent possible sources of mantle enrichment or contamination of the magmas upon emplacement, in addition to the subducted oceanic crust which presumably would represent a relatively depleted component. Of these components, the Svecofennian metasedimentary component is expected to be the most important; as such sediments are likely to have descended into the mantle together with the oceanic crust during the early Svecofennian subduction. The provenance of these sediments have generally been shown to consist of a mixture of early Svecofennian juvenile and old Archaean continental components, mostly in proportions of ca. 70% juvenile and 30% Archaean material, based on detrital zircon populations in metasediments (Huhma et al., 1991; Claesson et al., 1993; Andersson et al., 2000, 2004c; Väisänen et al., 2002, Lahtinen et al., 2002). This is also supported by the Nd–Sr isotopic data of the metasediments plotting in between that of the juvenile metaigneous and the Archaean crust (Fig. 10).

Crustal contamination of the 1.8 magmas during ascent through Svecofennian crust would not alter the isotopic composition of the magmas significantly, especially not those most strongly enriched in the east. Furthermore, the compositional range of the Svecofennian crust overlaps that of the newly formed 1.8 Ga rocks (Fig. 10). However, the granitic rocks, particularly those from Lemland that are less LREE-enriched than their associated mafic rocks (cf. Lindberg and Eklund, 1988), may in fact be entirely derived within and from the early Svecofennian metaigneous crust. Their initial isotopic ratios are compatible with such a model. However, the isotopic values of the granites are also indistinguishable from those of the mafic, enriched mantle-derived rocks (Table 3). In contrast, the Åva granites are more LILE and LREE enriched, but still lower than the associated lamprophyres, and could have suffered considerable crustal additions. In both complexes, widespread occurrences of mingling and mixing attest that the granites and mafic magmas intruded coevally (Lindberg and Eklund, 1988).

Fig. 12 shows that rocks from central and eastern areas generally have lower ε_{Nd} values but higher Th/Yb values compared to the rocks from the west. This negative correlation suggests that the eastwards increasing LILE and LREE enrichment was caused by melts/fluids carrying partly older crustal components into the mantle source regions, where the 1.8 Ga magmas were subsequently generated. The relatively low value of sample 8840 from the west is corroborated by field evidence of assimilation of crustal granitoids (cf. Andersson, 1997b).

5.3. Tectonic and regional implications

The degree and type of ca. 1.8 Ga mafic magmatism across the Svecofennian Domain, from close to the Archaean craton in the east, over the accreted calc-alkaline crust in the center, to the 1.8 Ga continental margin in the west, apparently represents a shift in the type and character of the respective mantle source enrichment, as shown above. The reason for this across-shield variation may be found in the tectonic evolution.

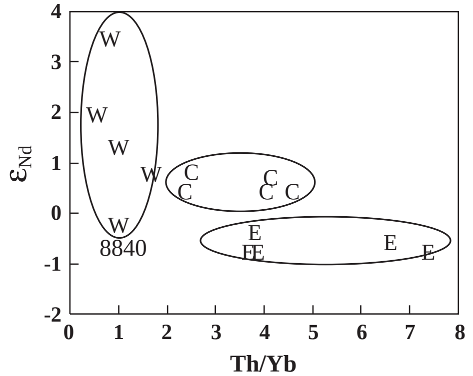

Fig. 12. ε_{Nd} vs. Th/Yb showing the decrease in ε_{Nd} with increasing amounts of slab-derived crustal material in the source.

During the period (2.1–)1.95–1.86 Ga essentially all crust in the Svecofennian Domain formed from mainly juvenile sources in island-arc systems and intervening sedimentary basins and was added to the Archaean cratonic nucleus by consecutive accretion in a today's southerly or southwesterly direction (Nironen, 1997). This continuing arc-accretion has been mostly interpreted to be associated with north-directed subduction (Gaál and Gorbatschev, 1987), supplying the metasomatizing melts/fluids to the overlying mantle wedges.

5.3.1. West

After accretion the juvenile crust was stabilized, and post-accretion extensional or transpressional forces ensued (Korja et al., 1993; Korja and Heikkinen, 1995; Högdahl and Sjöström, 2001; Andersson et al., 2004b) leading to the 'post-collisional' ca. 1.8 Ga magmatism studied here. Simultaneously in the west, an extensive magmatism was developed which reorganized the newly formed Svecofennian crust into the vast chains of batholiths of the Transscandinavian Igneous Belt (TIB), consisting dominantly of alkali-calcic granitoid magmatism (cf. Andersson, 1997b; Högdahl et al., 2004). The granitoid magmatism was generated by extensive underplating of the Svecofennian crust by mafic magmas derived from the uppermost spinel lherzolitic mantle, of which the here studied rocks of TVZ (west) is one expression. Essentially all mafic rocks in TIB studied so far have yielded calc-alkaline to tholeiitic continental arc geochemical signatures (Nyström, 1999; Claeson and Andersson, 2000; Claeson, 2000; Andersson et al., 2004a; Rutanen et al., 2005). A substantial part of the mantle enrichment shown by the western rocks may, in fact, be penecontemporaneous with the subduction and magmatism at ca. 1.8 Ga. The mafic rocks in the west are thus part of a voluminous suite of mafic magmatism, which is spread out over a large geographical area and most probably occupying substantial volumes of unexposed crustal sections. This is consistent with relatively large percentages of mantle melting, ca. 20–30% has been suggested (Andersson, 1997b), which diluted the enrichment levels in the resultant melts.

5.3.2. Central and east

It has been suggested that LRE- and LIL-element enriched shoshonites and alkaline rocks that clearly postdate active subduction but still show trace element patterns typical of subduction zones (negative spikes for Nb, Ta, Ti) originate in a layer in the lithospheric mantle that is enriched in incompatible elements. Examples have been described from the shoshonites in the Tibetan plateau (Turner et al., 1996), the Pliocene potassic magmas in Sierra Nevada (Feldstein and Lange, 1999), the Tertiary volcanic rocks in Anatolia (Pearce et al., 1990; Wilson et al., 1997), Caledonian minettes (Canning et al., 1996) and the Kilmerford shoshonites in Northern Britain (Zhou, 1987) and the Basin and range province, USA (Hawkesworth et al., 1995). However, there are contrasting models on how these layers may have originated: (1) enrichment by fluids from the asthenosphere to the lithosphere (e.g. McKenzie, 1989; Menzies et al., 1991), or (2) enrichment during a previous subduction stage (e.g. Fitton et al., 1991; Feldstein and Lange, 1999; Hawkesworth et al., 1995; Canning et al., 1996; Wilson et al., 1997). Pearce et al. (1990) describe the difference in composition of volcanoes in Eastern Anatolia as a consequence of both processes.

The combination of high Ba, Sr, LREE, CO_2, P_2O_5 and F encountered in the central, and particularly, in the eastern rocks, clearly favour mantle sources enriched by carbonate-dominated melts/fluids. The combined trace element signatures suggest that these melts/fluids were supplied to the mantle in a subduction zone setting, most probably derived from sedimentary components of subducted slabs.

The substantially lower enrichment levels of these elements in the magmas of the west signal a smaller role for carbonate melt-enriched mantle sections, while the equally high contents of other LILE (Rb, Cs, K, U) indicate a major role for an H_2O-dominated fluid enrichment process.

According to McKenzie (1989) metasomatic layers in the lithosphere have solidus temperatures considerably lower (900 °C) than the dry solidus for peridotite (1300 °C). Such layers could produce extensive volumes of high-potassium melts when regenerated by adiabatic decompression or by a plume (McKenzie, 1989), or by convective removal of the thermal boundary layer (Platt and England, 1994).

Studies of zircon xenocrysts from Palaeozoic kimberlites (Fig. 1) and their lower crustal xenoliths (Hölttä et al., 2000; Peltonen and Mänttäri, 2001) have shown that significant portions of the mafic lower crust in eastern Finland formed around 1.8 Ga. This suggests a relation between emplacement of the ca. 1.8 Ga upper crustal intrusions and widespread mafic underplating at this time. Peltonen et al. (2000) suggest that this magmatism may have been caused by a mantle plume that impinged on the base of the old lithosphere. However, already in early Svecofennian time (1.87 Ga) enriched mantle sources were available in the east (Konopelko and Eklund, 2003).

Erosion of the lithospheric mantle by asthenospheric upwelling may have caused the small but significant uplift of the crust between 1.83 and 1.80 Ga identified in southern Finland. The regional metamorphic peak in the accretionary arc complex of southern Finland took place roughly at 1.83 Ga, at a pressure around 5 kbar. The emplacement level of a ca. 1.80 Ga shoshonitic intrusion in the center has been determined to less than 2 kbar (Eklund and Shebanov, 2005) and in eastern Finland the contact aureole around a 1.8 Ga granite dyke was determined to around 2 kbar (Niiranen, 2000) in the same areas as the ca. 5 kbar, 1.83 Ga regional metamorphism.

The generation of the 1.8 Ga mafic rocks in the central and eastern areas may have been related to upwelling of hot asthenospheric material due to plume activity (Peltonen et al., 2000) or the break-off of an earlier subduction slab (Väisänen et al., 2000; Eklund and Shebanov, 2002), that caused fusion of sections of the strongly enriched lithospheric mantle by heating from below. However, no traces of 1.8 Ga magmas derived from such asthenospheric sources have been found.

Alternatively, complexes in the center and east, emplaced further inward the newly established Svecofennian continent, may have been generated as an effect of tectonic processes related to the contemporaneous active subduction beneath the continental margin in the southwest. The mafic rocks in the central and eastern areas obviously derive from smaller melt fractions of deeper mantle sections (in the garnet stability field), where carbonate metasomatism was strong. The more limited magmatism eastwards may be an effect of localized continentward extension and associated deep melting in response to the extensive continental margin processes of the TIB in the west. An Andean-type convergent margin setting has been proposed for the TIB (Nyström, 1982, 1999; Andersson, 1991; Åhäll and Larsson, 2000), associated with an overall transpressional regime in the overriding plate (Högdahl and Sjöström, 2001; Andersson et al., 2004b). This transpressional shearing may have had lithospheric-scale effects connected to magmatism reaching as far east as Russian Karelia. Above an N(E)-directed subduction zone at ca. 1.85–1.75 Ga, metasomatism and melting would also affect continuously deeper mantle sections in the same direction, but if this can be taken to affect areas as far east as Russian Karelia is uncertain.

6. Conclusions

During the build-up of the early Svecofennian ((2.1–)1.95–1.86 Ga) island-arc crust, subduction dehydration and melting caused extensive fluid and/or melt additions to the sub-Svecofennian mantle. The enriched mantle was accreted together with the overlying calc-alkaline crust to the Archaean cratonic nucleus in today's NE, thus forming the juvenile Svecofennian Domain which was accomplished by 1.86 Ga. High degrees of mantle melting in a converging continental margin in the west at ca. 1.8 Ga resulted in voluminous mafic magmatism that caused extensive melting and reworking of the juvenile Svecofennian crust (cf. Andersson et al., 2004b). The mantle enrichment could also partly have taken place in immediate connection with the 1.8 Ga magmatism. Geochemical and isotopic data suggest that the mantle sources in the west (TIB) were previously depleted sections of the uppermost mantle that was overprinted by slab-derived H_2O-dominated fluids, causing enrichment mainly in the LILE.

Eastwards, in the central and particularly in the eastern part of the domain, the level of enrichment in the mantle source regions of the ca. 1.8 Ga magmatism were considerably higher than in the west, as clearly shown by the geochemical and isotopic data. The nature of the enrichment is also different, as shown by a strong enrichment in LREE, P and F, in addition to the LILE, which is interpreted as the result of infiltration of carbonatitic melts. The isotopic data suggest that a significant amount of radiogenic Sr and low-radiogenic Nd was added to the mantle by the carbonic melts. The very strong enrichment in these elements in the derived melts also gives evidence for low melt fractions, much lower than in the west. The source regions are also situated deeper than in the west, in the garnet stability field as evidenced by the strong relative depletion of HREE and Y (garnet lherzolite).

The palaeotectonic setting and cause of the mantle magma generation and mafic underplating further continentwards, in the areas from the center to the east, is enigmatic. Models of mantle plume activity or the break-off of an inactivated subducted slab have been proposed. Both result in asthenospheric upwelling and fusion of the enriched lithospheric mantle by heating from below. On the other hand, transpressional shearing related to the coeval continental-margin convergence associated with the voluminous TIB magmatism in the west has affected the juvenile Svecofennian continent and may have reached through the whole lithosphere and triggered deep melting far continentward of the active continental margin. Future research may resolve the tectonic mechanisms behind this magmatism.

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