Ancient and juvenile components in the continental crust and mantle: Hf isotopes in zircon from Svecofennian mafic rocks and rapakivi granites in Sweden

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ABSTRACT

The sources of igneous rocks in the continental crust are elusive, but they may be traced by radiogenic isotopes, which convey a message about the age and composition of the concealed parts of the continent. We investigated the Hf-isotope composition of zircon in ten rocks from central and southern Sweden. Two felsic metavolcanic rocks and two metagabbros (ca. 1.89 Ga) from Bergslagen, southern Sweden, show εHf(t) ranges of −1.8 to +5.1 and +2.6 to +6.8, respectively, suggesting that juvenile sources have contributed to both. A 1.85 Ga granite from southern Bergslagen shows a εHf(t) range of −2.6 to +4.8 for magmatic zircons, but both highly negative and positive values for inherited grains, providing evidence for both Archean and juvenile crustal sources. These and previous data confirm the existence of juvenile proto-Svecofennian crust (<2.2–1.9 Ga) with a minor Archean component, from which later crustal magmas were generated. The Hf-isotope evolution curve for this crust can be approximated by εHf(1.9) = 3 ± 3 and 176Lu/177Hf = 0.018. Similarly, the present data, together with data for younger mafic intrusions, can be used to infer the presence of a “mildly depleted” sub-Svecofennian mantle evolution curve with εHf(1.9) = 4.5 ± 2.5 and 176Lu/177Hf = 0.0315. Zircons from four out of five rapakivi intrusions (1.53–1.50 Ga) in central Sweden yield negative εHf(t) in the range −9.8 to −4.6, suggesting mixed Archean and juvenile Svecofennian sources. One intrusion farther south ranges between εHf(t) of −4.1 and −1.6, and has a larger contribution from Svecofennian crust. The data suggest that the crust in Bergslagen, southern Sweden, is dominantly Paleoproterozoic, while higher proportions of Archean material are present below central Sweden.

INTRODUCTION

The Hf-isotope ratios of individual zircon crystals record heterogeneities in the magmas of various igneous rock types at a much higher spatial resolution than do whole-rock isotopic data (e.g., Griffin et al., 2002; Kinny and Maas, 2003; Belousova et al., 2006; Hawkesworth and Kemp, 2006). Although whole-rock data yield information on the characteristics of the magma source(s) at the scale of whole-rock samples within a given volume of a rock suite, this approach averages out any evidence of mixed provenance within a single specimen. Source components of contrasting isotopic composition may be preserved in a magma on a very local scale (e.g., the size of a hand specimen) within early-crystallized minerals that are stable enough to have survived subsequent stirring and mixing (e.g., Waight et al., 2000; Charlier et al., 2007; Kemp et al., 2007; Qin et al., 2010). Initial Hf-isotope data in zircons separated from one sample may thus provide a wider spectrum of values than a suite of whole-rock samples (although with a somewhat lower analytical precision), even approaching the end-member source compositions (e.g., Griffin et al., 2002; Andersen et al., 2002, 2007; Yang et al., 2007; Rutanen et al., 2011). In addition, U-Pb and Lu-Hf spot analysis by laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) on cores and overgrowth zones in zircon from igneous rocks yields combined information on the spread in protolith and assimilant ages and the isotopic composition of the magmas from which they crystallized (e.g., Belousova et al., 2006; Andersen et al., 2009a; Kurhila et al., 2010). This provides a powerful tool for unraveling the history of crustal tracts. There exists a limited volume of whole-rock and multigrain-zircon Hf-isotope data from the Fennoscandian Shield, including various granitoid (Patchett et al., 1981; Vervoort and Patchett, 1996) and mafic suites (Patchett et al., 1981; Söderlund et al., 2005, 2006); these data essentially replicate the information provided by the more abundant Nd-isotope data (see compilations in, e.g., Lahtinen and Huhma, 1997; Andersson et al., 2002, 2004; Rutanen and Andersson, 2009; Rutanen et al., 2011; Appelquist et al., 2011). Recent in situ zircon-Hf data come mainly from the SW part of the shield, where 1.7–0.9 Ga granitoids yield initial εHf compositions ranging from depleted mantle to somewhat below chondritic uniform reservoir (CHUR) (Andersen and Griffin, 2004; Andersen et al., 2004, 2007; Pedersen et al., 2009), and even to highly negative values for some late Sveconorwegian granites (Andersen et al., 2002, 2009b). Only a few studies have reported such data from rocks of the Archean to Paleoproterozoic part of the shield (Andersen et al., 2009a; Heinonen et al., 2010; Kurhila et al., 2010; Rutanen et al., 2011; Lauri et al., 2011).

Here, we present new in situ zircon Hf-isotope data for 1.90–1.85 Ga mafic and felsic rocks from the Bergslagen region, southern Sweden; these data add further constraints on the origin of the continental crust in this part of the Fennoscandian Shield. In addition, data are reported for five 1.53–1.50 Ga rapakivi intrusions in central Sweden, yielding information pertaining to the sources for this magmatism. The results are combined with previous isotopic data to build an integrated crustal model for...
the western and southern parts of the Svecofennian Domain.

**GEOLOGICAL SETTING AND SAMPLES**

The classical Bergslagen mining region in eastern southern Sweden (Stephens et al., 2009) consists of 1.91–1.87 Ga ore-bearing intercalated metavolcanic and metasedimentary rocks. The (mainly felsic) volcanic rocks dominate in the west and north, while the metasedimentary rocks are dominant in the SE part of the region (e.g., Allen et al., 1996; Lundström et al., 1998) (Fig. 1). The supracrustal rocks were intruded by an early Svecofennian granite-granodiorite-gabbro suite at 1.89–1.86 Ga, and a suite of late Svecofennian, migmatite-related granites and pegmatites at 1.85–1.75 Ga (summarized in, e.g., Andersson and Öhlander, 2004; Andersson et al., 2006a; Hermansson et al., 2001; Wikström and Andersson, 2004). The second, and dominant, episode (1.81–1.75 Ga) followed mainly farther away from the Svecofennian margin (e.g., Mansfeld, 1991; Andersson and Wikström, 2004), while the third episode (1.71–1.65 Ga) was emplaced even farther to the west and north (e.g., Ahl et al., 1999; Lundqvist and Persson, 1999; Söderlund et al., 1999; Brander and Persson, 2011).

Five samples from Bergslagen were studied: two mafic intrusions, two felsic volcanic rocks, and one sample of the earliest Transscandinavian Igneous Belt generation. The petrography and secondary ion mass spectrometry (SIMS) geochronology of these samples are described in Andersson et al. (2006a). All samples have been metamorphosed in middle to upper amphibolite facies. The two mafic intrusive rocks (R89-14: 1887 ± 5 and UB98-29: 1895

Figure 1. Geological overview of central Sweden (A) and the Bergslagen region (B), with sample locations indicated. Upper frame in inset map outlines position of A, and lower frame outlines position of B. BB—Bothnian Basin, LjD—Ljusdal domain, BS—Bergslagen region, TIB—Transscandinavian Igneous Belt. Figure is modified from Andersson (1997a), Andersson et al. (2006a), Högdahl et al. (2008), and references therein.
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± 5 Ma) were sampled from western Bergslagen, the two felsic volcanic rocks (U97-1: 1888 ± 12 and U97-3: 1892 ± 7 Ma) were sampled from the eastern part, and the early Transscandinavian Igneous Belt augen gneiss (Fin2: 1855 ± 6 Ma) came from the southern part (Fig. 1).

North of the Bergslagen region, there is a vast and relatively little-studied area (referred to as the Ljusdal domain; Högdahl et al., 2009) of mostly gneissic supracrustal rocks, and a suite of variably deformed 1.86–1.84 Ga plutonic rocks of the Ljusdal Batholith (e.g., Sjöström and Bergman, 1998; Högdahl et al., 2008). North of the Ljusdal domain, there lies the Bothnian Basin, an area strongly dominated by thick successions of metasedimentary rocks (Lundqvist, 1987), which are intruded by at least two generations of granitoid rocks, including pegmatites, at ca. 1.95–1.86 and 1.82–1.79 Ga (e.g., Claesson and Lundqvist, 1995; Romer and Smeds, 1997; Lundqvist et al., 1998; Rutland et al., 2001; Weihed et al., 2002; Högdahl et al., 2008; Sköld and Rutland, 2006). To the west and SW, the Bothnian Basin is intruded by ca. 1.87–1.85 Ga K-feldspar-megacryst-bearing granites and diatexites (Högdahl and Ahl, 2004; Högdahl et al., 2008) and, farther north, by the mainly 1.81–1.77 Ga Revsund granitoid suite (Claesson and Lundqvist, 1995; Weihed et al., 2002), which constitutes the northern continuation of the Transscandinavian Igneous Belt (e.g., Andersson, 1997b; Gorbatschev, 2004).

Composite rapakivi intrusions, including granites, syenites, gabbros, anorthosites, and numerous diverse dikes, intruded the southern part of the Bothnian Basin and the Revsund suite at 1.53–1.50 Ga (Andersson, 1997c; Andersson et al., 2002). One rapakivi intrusion was emplaced at Strömsbro farther south, right at the border between the Bergslagen region and the Ljusdal domain (Andersson, 1997a; Fig. 1B).

Descriptions and U-Pb zircon thermal ionization mass spectrometry (TIMS) age determinations of the five rapakivi intrusions can be found in Andersson (1997a, 1997b, 2001) and Andersson et al. (2002). These are from south to north (Fig. 1): the Strömsbro granite (1500 ± 20 Ma), the Rödön granite (1497 ± 6 Ma), the Mårdsjö granite (1524 ± 3 Ma), the Nordsjö syenite (1520 ± 3 Ma), and the Mullnäset granite (1526 ± 3 Ma).

ANALYTICAL METHODS

The five zircon samples from the Bergslagen area were previously analyzed for U-Th-Pb using the SIMS technique (Andersson et al., 2006a). Our Lu-Hf analyses were obtained in the same analytical spots as were used for the previous U-Pb analyses (Fig. 2). Thus, the age of zircon growth was known for each of these spots. Crystallization ages for the five rapakivi intrusions were previously determined only by zircon U-Pb TIMS geochronology (Andersson, 1997c; Andersson et al., 2002). The great majority of crystals from these intrusions show uncomplicated magmatic growth zoning, and such crystals were selected both for the previous TIMS analysis and for the present Lu-Hf study.

Hf-isotope analyses were performed at National Key Centre for Geochemical Evolution and Metallogeny of Continents of the Australian Research Council (GEMOC ARC), Macquarie University, Australia. The 176Hf/177Hf ratios in zircon were measured with a New Wave Research 213 nm laser-ablation microprobe attached to a Nu Plasma multicollector ICP-MS. The analytical methods, including extensive data on the analysis of standard solutions and zircons, were discussed in detail by Griffin et al. (2000, 2002). Data were processed using the Nu Plasma

Figure 2. Cathodoluminescence images of analyzed zircons. The 207Pb/206Pb spot ages (Andersson et al., 2006a) are given, and the initial εHf data are from Table 1.
time-resolved analysis software, which allows selection of the most stable part of the ablative signal. Analyses were carried out at 5 Hz frequency with a beam diameter 55 μm and energies around 0.1 mJ per pulse. Background was measured for 60 s prior to ablation. The length of the analysis varied between 30 and 140 s, depending on the thickness of the grains. Repeated analyses of the Mud Tank zircon (long-term running average $^{176}$Hf/$^{177}$Hf = 0.28253 ± 43; Griffin et al., 2007) and the 91500 zircon (long-term running average $^{176}$Hf/$^{177}$Hf = 0.282307 ± 58; Griffin et al., 2006) were used to monitor data quality, translating to about ±2 μ units.

For the decay constant of $^{176}$Lu, a value of 1.867 × 10^{-11} yr^{-1} (Scherer et al., 2001, 2007) was applied. The present-day chondritic values used were $^{176}$Lu/$^{177}$Hf = 0.0336 and $^{176}$Hf/$^{177}$Hf = 0.282785 (Bouvier et al., 2008). For the depleted initial $^{176}$Hf/$^{177}$Hf = 0.28325 (εHf = +16.4), similar to mid-ocean-ridge basalt (MORB), and a $^{176}$Lu/$^{177}$HfDM value of 0.0388.

**RESULTS**

The results of all analyses are given in Table 1 and plotted in εHf versus time diagrams in Figure 3. Even though the $^{207}$Pb/$^{206}$Pb age was known for each individual spot for the Bergslagen zircons, the initial εHf value for each analysis from the magmatic parts of crystals is plotted at the calculated crystallization age of the rock. This is because of the relatively large error in the individual SIMS analyses, variable degrees of discordance, and possible nonzero lower intercepts. However, inherited cores and overgrowths are plotted at their individual $^{207}$Pb/$^{206}$Pb ages. The initial εHf values of the rapakivi zircons are plotted at the calculated TIMS age of crystallization. The homogenous zoning of the zircons and the coherency of the Hf isotopic results support a magmatic origin for the analyzed crystals. In reference to the data and discussion, one should bear in mind the analytical precision of ±2 μ units, which, however, will not alter the interpretations and conclusions.

The initial εHf values from the early Svecofennian mafic intrusive rocks range from +2.6 to +6.8, including a few slightly older cores and slightly younger overgrowths, while the data from the early Svecofennian felsic volcanic rocks fall in the range −1.8 to +5.1. The initial εHf results from the rapakivi zircons range between −9.8 and −1.6 and differ significantly among the intrusions. The zircons from the Strömsbro granite in the south show distinctly lower ones in Mullnäset. The range within each rapakivi intrusion does not exceed 3.5 μ units.

Including the previous analyses from the 1.85 Ga Finspång augen gneiss (Andersson et al., 2009a), magmatic crystals yielded initial εHf values between −2.6 and +4.6, while analyses of inherited cores show a considerable range in ages and initial εHf (Fig. 3). Analytical spots giving apparent $^{207}$Pb/$^{206}$Pb ages between 1.91 and 2.24 Ga range in εHf from −16.1 to +9.8, i.e., from the Archean crust to depleted mantle (DM). However, seven out of ten fall in the narrow range +0.6 to +3.6. Those with apparent ages in the range 2.35 to 2.95 Ga yielded εHf in the range

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<tr>
<th>Sample no. Coordinates, RT90 2.5 gon V</th>
<th>Age (Ma)</th>
<th>$^{176}$Hf/$^{177}$Hf</th>
<th>2se</th>
<th>$^{176}$Lu/$^{177}$Hf</th>
<th>2se</th>
<th>$^{176}$Yb/$^{177}$Hf</th>
<th>2se</th>
<th>$^{176}$Hf/$^{177}$Hf initial</th>
<th>εHf initial</th>
<th>2se</th>
<th>T(DM)* (Ga)</th>
<th>T(DM)* crustal (Ga)</th>
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<td>Amphibolitic gabbro (663060/144590)</td>
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<td>R8914.n746-02a 1920 0.281772 0.000022 0.001658 0.049513 0.281711 5.4 0.8 2.09 2.23</td>
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(continued)
### TABLE 1. Lu-Hf DATA ON ZIRCONS FROM SVECOFENNIAN AND RAPAKIVI ROCKS IN SWEDEN (continued)

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Coordinates, RT90 2.5 gon V</th>
<th>Age (Ma)</th>
<th>(^{176})Hf/(^{177})Hf</th>
<th>2se</th>
<th>(^{176})Lu/(^{177})Hf</th>
<th>(^{176})Yb/(^{177})Hf</th>
<th>initial</th>
<th>(\varepsilon_{Hf}) initial</th>
<th>2se</th>
<th>T(DM)* (Ga)</th>
<th>T(DM)* crustal (Ga)</th>
</tr>
</thead>
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**Svecofennian rocks**

Felitic supracrustal rock (668180/163190)

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<th>Sample no.</th>
<th>Coordinates, RT90 2.5 gon V</th>
<th>Age (Ma)</th>
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<th>2se</th>
<th>(^{176})Lu/(^{177})Hf</th>
<th>(^{176})Yb/(^{177})Hf</th>
<th>initial</th>
<th>(\varepsilon_{Hf}) initial</th>
<th>2se</th>
<th>T(DM)* (Ga)</th>
<th>T(DM)* crustal (Ga)</th>
</tr>
</thead>
</table>

**Finspång augen gneiss (650975/149720)**

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<th>Age (Ma)</th>
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<th>2se</th>
<th>(^{176})Lu/(^{177})Hf</th>
<th>(^{176})Yb/(^{177})Hf</th>
<th>initial</th>
<th>(\varepsilon_{Hf}) initial</th>
<th>2se</th>
<th>T(DM)* (Ga)</th>
<th>T(DM)* crustal (Ga)</th>
</tr>
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</table>

**Rapakivi complexes**

**Strömsbro granite (673225/157457)**

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<th>Sample no.</th>
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<th>2se</th>
<th>(^{176})Lu/(^{177})Hf</th>
<th>(^{176})Yb/(^{177})Hf</th>
<th>initial</th>
<th>(\varepsilon_{Hf}) initial</th>
<th>2se</th>
<th>T(DM)* (Ga)</th>
<th>T(DM)* crustal (Ga)</th>
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**Rödö granite (692031/159275)**

<table>
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<th>Sample no.</th>
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<th>2se</th>
<th>(^{176})Lu/(^{177})Hf</th>
<th>(^{176})Yb/(^{177})Hf</th>
<th>initial</th>
<th>(\varepsilon_{Hf}) initial</th>
<th>2se</th>
<th>T(DM)* (Ga)</th>
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**Nordjöy syenite (706214/151693)**

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<th>(^{176})Lu/(^{177})Hf</th>
<th>(^{176})Yb/(^{177})Hf</th>
<th>initial</th>
<th>(\varepsilon_{Hf}) initial</th>
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**Mårdsjö granite (702589/149466)**

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<th>(^{176})Yb/(^{177})Hf</th>
<th>initial</th>
<th>(\varepsilon_{Hf}) initial</th>
<th>2se</th>
<th>T(DM)* (Ga)</th>
<th>T(DM)* crustal (Ga)</th>
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**Mullnäset granite (707200/149208)**

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<th>Sample no.</th>
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<th>2se</th>
<th>(^{176})Lu/(^{177})Hf</th>
<th>(^{176})Yb/(^{177})Hf</th>
<th>initial</th>
<th>(\varepsilon_{Hf}) initial</th>
<th>2se</th>
<th>T(DM)* (Ga)</th>
<th>T(DM)* crustal (Ga)</th>
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\(^{1}\)Minimum depleted mantle (DM) age, calculated using the measured \(^{176}\)Lu/\(^{177}\)Hf of the zircon.

\(^{2}\)Depleted mantle age, calculated using a bulk crust \(^{176}\)Lu/\(^{177}\)Hf value of 0.018.
Figure 3. (A–C) $\varepsilon_{\text{Hf}}$ versus time diagram for the present samples and other relevant data. Chondritic uniform reservoir (CHUR) is after Bouvier et al. (2008). Depleted mantle (DM) is after Griffin et al. (2000), modified by the CHUR data of Bouvier et al. (2008). Upper limit of Fennoscandian Archean crust is as defined by Andersen et al. (2009a). Green ellipse in A encompasses the zircon data for early Svecofennian mafic rocks of this study and includes the data of Patchett et al. (1981). The ellipse for early Svecofennian granites includes data from Bergslagen and south-central Finland (Andersen et al., 2009a; Heinonen et al., 2010). The ellipse for late Svecofennian granites includes data from Vervoort and Patchett (1996), Andersen et al. (2009a), and Kurhila et al. (2010). Violet diamonds—Transscandinavian Igneous Belt (TIB) granitoids (Andersen et al., 2009a). Various red and green squares—Finnish rapakivi granites and their associated mafic rocks, respectively (Patchett et al., 1981; Heinonen et al., 2010). Crosses—various suites of younger mafic intrusions in the central part of the Fennoscandian Shield (Söderlund et al., 2005). The latter data are broadly encompassed by the evolutionary field for the Paleoproterozoic Fennoscandian “mildly depleted” subcontinental mantle (MDM), approximated by: $\varepsilon_{\text{Hf}}(1.90) = 4.5 \pm 2.5$ and $^{176}\text{Lu}^{177}\text{Hf} = 0.0315$. SJPC—Svecofennian juvenile protocrust, representing new crust formed after 2.2 Ga in the Svecofennian Domain, partly by mixing with Archean material (indicated by large arrows). Post–2.0 Ga felsic crust in the southern Svecofennian province is largely formed from this juvenile crust and lies within an evolutionary field defined by: $\varepsilon_{\text{Hf}}(1.90) = 3 \pm 3$ and $^{176}\text{Lu}^{177}\text{Hf} = 0.018$, including Transscandinavian Igneous Belt granitoids and Finnish rapakivi granites. Frames in A outline areas enlarged in B and C. Blue arrow in B indicates minor Archean contributions to Svecofennian granitoids plotting below the juvenile crust, supplied mainly through melting of metasediments. Arrows in C indicate Svecofennian and Archean contributions to the rapakivi granitoids from various local mixtures of meta-igneous sources.
Svecofennian metasediments typically contain with its inferred character as a redeposited volcanic rocks. The true ages of the analyses older than 2.2 Ga are poorly constrained, particularly those from Andersen et al. (2009a), because of discordance and large analytical errors. However, the presence of zircon that crystallized in early Paleoproterozoic and Archean time is clear.

DISCUSSION

Svecofennian Juvenile Crust

Following rifting of the Archean craton margin in the NE, new crust started to form in several arc systems in successive stages outboard of the craton from ca. 2.1 Ga (Gädl and Gorbatschev, 1987; Lahtinen and Huhma, 1997; Nironen, 1997), creating early “microcontinents” in the period 2.1–1.93 Ga (Lahtinen et al., 2005, 2009a); little of this older crust is exposed at the present erosion level. This proto-Svecofennian crust was strongly reworked from 1.91 to 1.86 Ga, when most of the presently exposed rocks in the domain were formed. The preexistence of the proto-Svecofennian crust is substantiated by a few outcropping 1.95–1.91 Ga igneous rocks, and especially by the dominance of juvenile (<2.1 Ga) depleted-mantle Nd model ages in the 1.91–1.86 Ga felsic rocks, as well as in the felsic Transscandinavian Igneous Belt rocks (e.g., Patchett et al., 1987; Valbracht et al., 1994; Andersson, 1997b; Lahtinen and Huhma, 1997). Even stronger evidence comes from the zircon record. Large numbers of Svecofennian metasedimentary rocks carry detrital zircons (60%–70%) in the age range 2.1 to 1.86 Ga; the remainder mostly are Archean, and a very few grains fall in the age range 2.45 to 2.1 Ga (e.g., Claesson et al., 1993; Lahtinen et al., 2002, 2009b, 2010; Sultan et al., 2005; Andersson et al., 2006a; Bergman et al., 2008; Skjöld and Rutland, 2006; Williams et al., 2008). Inherited grains and cores in igneous rocks also follow this distribution (e.g., Kumulainen et al., 1996; Ehlers et al., 2004). These data clearly suggest major formation of crust ca. 2.1–1.91 Ga; the available Nd-isotope data and zircon-Hf isotopic data indicate that much of this crust was juvenile.

Except for one grain, the present data for zircons from the felsic metavolcanic rocks, and most of the igneous zircons of the Transscandinavian Igneous Belt granite, show positive initial \(^{176}\text{Lu}/^{177}\text{Hf} \) values. Sample U97-3 shows slightly lower \(^{176}\text{Lu}/^{177}\text{Hf} \) values, compared with U97-1, including one slightly below CHUR. This may be correlated with its inferred character as a redeposited volcanogenic sediment (Andersson et al., 2006a); Svecofennian metasediments typically contain higher proportions of older material compared with the meta-igneous rocks (see summary in Andersson et al., 2002). The total range of values is \(7.7 \varepsilon \) units of the \(16 \varepsilon \) units between DM and the upper limit of the evolution for the Fennoscandian Archean crust, and they all fall entirely on the depleted (high-\( \varepsilon \)) side of CHUR.

The present data corroborate previous juvenile bulk-rock Hf-isotope data from Svecofennian granites in Bergslagen and southern Finland (Patchett et al., 1981; Vervoort and Patchett, 1996) and zircon Hf-isotope data from Svecofennian and Transscandinavian Igneous Belt granites in Sweden (Andersen et al., 2009a; Fig. 3). This restricted range of positive initial ratios does not lend support to a model of mixing between Archean and DM sources, which would tend to generate a larger spread in initial ratios, but rather suggests a dominance of sources separated from the mantle in early Paleoproterozoic time (cf. Andersen et al., 2009a). Similarly, Hf data for zircons from Svecofennian granitoids in southern Finland tend to be relatively juvenile in the west but carry higher proportions of Archean signatures toward the craton margin in the east (Kurhila et al., 2010; Heinonen et al., 2010; Rutanen et al., 2011).

Inherited cores of zircons from the Transscandinavian Igneous Belt augen gneiss span large ranges in both age and initial \(^{176}\text{Lu}/^{177}\text{Hf} \) (Fig. 3). However, most of those in the age range 2.1 to 1.9 Ga have initial \(^{176}\text{Lu}/^{177}\text{Hf} \) values on the juvenile side (+0.6 to +3.6). Only one originally crystallized at ca. 1.97 Ga from a magma derived from Archean crust (\(\varepsilon_{\text{Hf}} = -16.1\)). The five cores with \(^{206}\text{Pb}/^{208}\text{Pb} \) ages in the range 2.39 to 2.95 Ga show initial \(^{176}\text{Lu}/^{177}\text{Hf} \) values in the range \(-2.4 \) to \(-4.6 \), and these values are typical for Fennoscandian Archean crust. Even if the exact ages of these grains are uncertain, due to discordance, they most certainly contain dominantly Archean Hf. Similarly, the ages are uncertain for the two juvenile analyses at 2.24 and 2.35 Ga, and they may also in reality be Archean. One of these has a very radiogenic Hf-isotope composition. Due to the low Lu/Hf ratio in zircon, this depleted character will be present in this sample irrespective of age, and it thus shows the existence of juvenile material in the crustal mixture at the time of its formation.

Andersen et al. (2009a) suggested that the limited spread in initial \(^{176}\text{Lu}/^{177}\text{Hf} \) data for zircons from Svecofennian and Transscandinavian Igneous Belt granitoids could be explained by derivation from preexisting juvenile Svecofennian crustal Hf isotope evolution. This is approximately defined by the evolution \(^{176}\text{Lu}/^{177}\text{Hf} = 2.3 \pm 3 \) and \(^{176}\text{Lu}/^{177}\text{Hf} = 0.015 \). The presence of a juvenile Svecofennian protocrust is supported by the positive \(^{176}\text{Lu}/^{177}\text{Hf} \) values of 2.2–1.9 Ga inherited zircons in granites (Fig. 3; Andersen et al., 2009a; Kurhila et al., 2010). The coexistence of early Paleoproterozoic inherited grains with both DM and Archean initial Hf isotopic compositions suggests that the earliest Svecofennian protocrust formed from a mixture of these components after 2.2 Ga (Fig. 3). This protocrust probably was created in early arc systems of variable maturity from a depleted to increasingly enriched mantle, mixed with subordinate amounts of components derived from the Archean craton (cf. Patchett and Bridgwater, 1984; Patchett et al., 1987; Andersson, 1991). The development of this protocrust involved the creation and accretion of several arc systems, as well as reworking of early arcs into mature arcs (microcontinents), and collisions with the pre-accreted continent over such extended time (see, e.g., Nironen, 1997; Lahtinen et al., 2005, 2009a, 2009b).

During the proto-Svecofennian (ca. 2.1–1.91 Ga; Andersen et al., 2006a), and the following early Svecofennian (1.91–1.86 Ga) period of major juvenile crust formation (a major worldwide crustal growth episode; e.g., Condé, 2000), Fennoscandia became included in an assembled Paleoproterozoic supercontinent (e.g., Rogers and Santosh, 2002; Zhao et al., 2004, 2006). The 1.85–1.65 Ga Transscandinavian Igneous Belt formed part of a long-lived active continental margin along one side of the supercontinent, stretching from SW Baltica all through southern Laurentia (e.g., Gower et al., 1990; Karlstrom et al., 2001; Johansson, 2009).

The Hf-isotopic evolution of the Svecofennian protocrust can be broadly constrained to \(^{176}\text{Lu}/^{177}\text{Hf} = 3 \pm 3 \) and \(^{176}\text{Lu}/^{177}\text{Hf} = 0.018 \), based on the presently available Svecofennian data, and data from Transscandinavian Igneous Belt granitoids that are considered to be derived from the Svecofennian crust (Fig. 3). This \(^{176}\text{Lu}/^{177}\text{Hf} \) ratio is higher than that of estimates for the average continental crust (0.010–0.015), but it is in the range estimated for the lower crust (0.015–0.020) (e.g., Taylor and McLennan, 1995; Wedepohl, 1995; Rudnick and Gao, 2003). A relatively mafic composition, similar to that of the lower crust, would be anticipated for such a juvenile mantle–derived protocrust. This evolution trend is slightly revised from that of Andersen et al. (2009a), taking into account the additional data.

Sub-Svecofennian Mantle

The data for the two mafic samples are the first reported for zircons in early Svecofennian mafic rocks. Even if the overlap with zircons from the felsic rocks is substantial, their \(^{176}\text{Lu}/^{177}\text{Hf} \) in general ranges to higher values (Fig. 3). The range indicates mildly to relatively strongly depleted sources, though not as depleted as the DM. Initial \(^{176}\text{Lu}/^{177}\text{Hf} \) data for early Svecofennian...
mafic rocks range from DM to values slightly below CHUR, but most are “mildly depleted” (see compilations in Andersson et al., 2004; Rutanen and Andersson, 2009). These data, and particularly the dominant initial εNd (≈1 to +2) of mafic Transscandinavian Igneous Belt rocks in southern Sweden (e.g., Andersson, 1997b; Claeson, 2001; Andersson et al., 2007; Rutanen and Andersson, 2009), suggest the widespread occurrence of “mildly depleted” mantle sources in this part of the shield. Similarly, in southern Finland, mafic 1.9–1.8 Ga rocks have chondritic to mildly positive εNd values (e.g., Lahtinen and Huhma, 1997; Andersson et al., 2006b; Rutanen et al., 2011). This, together with MORB-like, or lower, contents of high field strength elements and enrichments in light rare earth and large ion lithophile elements relative to MORB, led Andersson et al. (2006b, 2007) to suggest that the sources for this magmatism represent previously depleted mantle that became enriched during Svecofennian (<2.1 Ga) development (cf. refertilization of the subcontinental lithospheric mantle discussed by, e.g., Griffin et al., 2000, 2009, and references therein).

Rutanen et al. (2011) suggested a heterogeneous distribution of such mantle sections to explain the variation from slightly subchondritic to strongly depleted initial εNd values in the early Svecofennian mafic rocks. An evolution of the sub-Svecofennian mantle defined by 147Sm/144Nd = 0.155 ± 0.015 would encompass the initial εNd ranges for most post-Svecofennian mafic suites in the Svecofennian Domain, and it is compatible with the 147Sm/144Nd ratios measured in enriched mantle xenoliths from elsewhere (e.g., McDonough and McCulloch, 1987; Voshage et al., 1987; Goring and Kay, 2000; Schmidberger et al., 2001).

Although the initial εNd values of various 1.6–0.28 Ga mafic rock suites in southern Fennoscandia show variations in detail, compatible with variable degrees of source enrichment (Söderlund et al., 2005), they broadly follow a common evolution that was defined by Andersson et al. (2009a) as εNd (1.60) = 3 ± 3 and 176Lu/177Hf = 0.033 (excluding the more depleted Central Svecofennian Dolerite Group), i.e., essentially parallel to CHUR. However, the present early Svecofennian data show slightly more depleted initial values and constrain, together with the younger intrusions from southern Fennoscandia, an evolution of approximately: εNd (1.89) = 4.5 ± 2.5 and 176Lu/177Hf = 0.0315 (Fig. 3), representing an early Svecofennian “mildly depleted mantle,” similar to the one suggested by Nd isotopes for the Transscandinavian Igneous Belt and many early Svecofennian rocks (e.g., Andersson et al., 2004, 2007; Rutanen and Andersson, 2009). Thus, the early Svecofennian mafic rocks were derived from depleted mantle sections that acquired their variably mildly depleted characters through additions of fluids and melts carrying an Archean isotopic imprint into the mantle wedge by subduction during arc and microcontinent assembly and reworking in proto-Svecofennian time (<2.3–1.9 Ga). The associated felsic crustal rocks derived their overlapping isotopic signature through reworking of the juvenile mafic crust and further minor additions of Archean crustal components (Fig. 3).

With the accumulation of more data on early Svecofennian mafic rocks, initial εHf values are likely to show an increased spread, reflecting variously enriched sections of the sub-Svecofennian mantle. This is tentatively demonstrated by Hf-isotope data from 2.15 to 1.97 Ga mafic intrusions in Finland (Patchett et al., 1981), some of which reflect increased contributions of Archean crustal components toward the craton in the east. However, if the Hf-isotope data remain similar to the Nd-isotope data, most of the data are expected to be “mildly depleted,” roughly coinciding with the evolution sketched in Figure 3.

**Source Variations in Fennoscandian Rapakivi Complexes**

The zircon data for the central Swedish rapakivi granitoids show an overall range in initial εHf values of −9 εHf units (−10 to −1.5), and a within-sample variation of less than 3.5 εHf units, while the total spread between DM and the Svecofennian Archean crust at 1.5 Ga is ~22 units (Fig. 3). Thus, the Hf data allow an interpretation that the zircons crystallized from thoroughly mixed magmas with variable proportions of crustal- and mantle-derived material, or from magmas with an origin entirely within a strongly enriched mantle. However, the dominantly granitic compositions (cf. Andersson, 1997a, 1997c, 2001), relatively unradiogenic Hf isotopes, and small and characteristic within-sample Hf isotopic variations favor distinct crustal sources for the granitoids of each complex.

The distinctly more radiogenic zircons of the Strömsbro granite, compared with those from the complexes farther north, strengthen the case made from whole-rock Nd isotopes (Andersson, 1997c) that this complex contains significantly less of a much older crustal component. In fact, the initial Nd-isotope composition of the Strömsbro granite is encompassed by the evolutionary trend of the early Svecofennian metaigneous crust (even if within its lowest range), while the compositions of those farther north range to lower values, suggesting involvement of up to 40% Archean components (Andersson et al., 2002). Similarly, the proposed Hf isotopic evolution of the Svecofennian crust encompasses the composition of the Strömsbro zircons, but not that of the northern complexes (Fig. 3). Thus, the Lu-Hf isotopic system also supports the presence of significant Archean components in these rocks.

As with the Nd isotopes (e.g., Rämö, 1991), the εNd evolution of the Svecofennian crust encompasses the compositional range of zircons from rapakivi granites from SE and SW Finland (Fig. 3), suggesting juvenile crustal sources for this magmatism. However, the variable presence of small amounts of Archean components in the Svecofennian meta-igneous crust (e.g., Andersen et al., 2009a; Kurhila et al., 2010) should also have affected the younger igneous rocks derived from it. In the Strömsbro complex, the presence of Archean components is indicated by the equally low εNd values (~5) for the associated mafic rocks, suggesting either Archean enriched-mantle sources or contamination by Archean material in the crust (Andersson, 1997c). The variation in the ranges of εHf among the individual rapakivi complexes in central Sweden is most likely related to local variations of the Archean/Svecofennian ratio in their respective source areas. However, the previously reported presence of Archean zircons in the 1.52 Ga Nordsjö syenite (Claesson et al., 1997) has not been possible to reaffirm as yet (Andersson and Claesson, personal commun.). In any case, the zircon Hf-isotope data provide additional independent evidence for the presence of a buried Archean basement beneath central Sweden (cf. Andersson et al., 2002).

This basement has been interpreted to represent the core of the Bothnian microcontinent that docked with the craton around 1.90 Ga in the orogenic collage of Lahtinen et al. (2005). The appearance of anorthosite-mangerite-charnockite-granite suites, rapakivi granites, and associated dike swarms at ca. 1.6–1.5 Ga has been related to the initiation of breakup of the Paleoproterozoic supercontinent (e.g., Rogers and Santosh, 2002; Zhao et al., 2004).

Initial εHf data for zircons from the rapakivi-associated mafic rocks in southern Finland (Heinonen et al., 2010) are largely encompassed by the evolution of the “mildly depleted” Fennoscandian lithospheric mantle (Fig. 3), suggesting that this represents a possible major source also for these magmas. However, excursions to both more and less radiogenic compositions indicate magma components from the crust as well as a more depleted mantle.

**CONCLUSIONS**

1. Initial εNd values in zircon from early Svecofennian (ca. 1.89 Ga) felsic volcanic rocks range from −1.8 to +5.1, similar to the range in
early Svecofennian granitoids, supporting a dominance of juvenile Proterozoic sources for such magmas in the southern Svecofennian province. This implies a reworking of material separated from the mantle <2.2 Ga ago and is further supported by the presence of inherited 2.2–1.9 Ga zircons with positive εHf values in granitoids from Bergslagen and southern Finland.

2. Zircons from a ca. 1.85 Ga granite from southern Bergslagen show εHf values ranging from −2.6 to +4.6, compatible with crystallization from a heterogeneous magma derived from a mix of <2.2 Ga and Archean crustal sources. The presence of inherited juvenile and Archean zircons supports this notion. The peraluminous composition of this granite suggests derivation from, or major assimilation of, Svecofennian sediments that contained Archean debris.

3. Initial εHf values in zircon from early Svecofennian (ca. 1.89 Ga) mafic intrusions overlap the values of the felsic rocks but extend to higher values (+2.6 to +6.8), though not as high as the depleted mantle. This is compatible with a main source region in a “mildly depleted” Svecofennian mantle, i.e., a previously depleted mantle that was enriched during proto-Svecofennian subduction. Components of a more strongly depleted mantle cannot be ruled out.

4. Zircons from granitoids of the central Scandinavian rapakivi complexes show small but individually distinct ranges in their initial εHf values (<3 ± 1 units). The initial εHf values are lower than expected for magmas with juvenile (<2.2 Ga) Svecofennian sources. Thus, these granitoids appear to be derived from sources with characteristic Archean/Svecofennian source mixtures.

5. The conclusion, drawn earlier from Nd-isotope data, that central Sweden is underlain by lithosphere that contains Archean components, is further supported by the new HF data. In contrast, the Svecofennian in Bergslagen is dominated by juvenile (<2.2 Ga) crustal additions with only minor Archean components, presumably mostly derived from the metasediments.

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