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Petrogenesis of Mesoproterozoic (Subjotnian) rapakivi complexes of central Sweden: implications from U–Pb zircon ages, Nd, Sr and Pb isotopes

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ABSTRACT: U–Pb zircon geochronology of Mesoproterozoic (Subjotnian) rapakivi complexes in central Sweden yields: 1526 ± 3 Ma (Mullnäset), 1524 ± 3 Ma (Mårdsjö), 1520 ± 3 Ma (Nordsjö) and 1497 ± 6 Ma (Rödön). Together with complexes further S in Sweden, they constitute the westernmost, youngest (1·53–1·47 Ga) belt of rapakivi magmatism in the Fennoscandian shield.

The low initial ε_{Nd} values (−8·9 to −4·8) of all studied Subjotnian basic, intermediate and silicic rocks, require an input from an old (Archaean) low-radiogenic source component, as evidence for Palaeoproterozoic protoliths in the age range 2·5–2·1 Ga is lacking in this region. Crustal, early Svecofennian + Archaean (roughly 30–40%) sources are suggested for the Subjotnian A-type granites and syenites, where the granites derive from undepleted, granodioritic, and the syenites from monzodioritic ($\pm$depleted crustal) protoliths. The basic rocks originate from a depleted mantle acquiring the enriched Nd isotopic signatures during interaction with an Archaean lower crust (20–40%), largely depleted after rapakivi melt extraction. Pb isotope data from feldspars ($^{207}Pb/^{204}Pb$ to 15·018–15·542) support the presence of Archaean components in the magmas.

The results indicate that an Archaean basement is underlying relatively wide areas of Svecofennian formations in central Sweden. This old basement section was most likely rifted off the Archaean craton in the NE in Palaeoproterozoic times.

KEY WORDS: Archaean basement, A-type, assimilation, contamination, initial ε_{Nd} ratios, LREE enrichment/depletion, mixing

Mesoproterozoic (Subjotnian) rapakivi intrusive complexes of the Fennoscandian shield extend from Russian Karelia in the E to central Sweden in the W (Fig. 1), including the classic Wiborg, Salmi and Åland batholiths, which contain granites with exceedingly well-developed rapakivi textures (cf. e.g. Rämö & Haapala 1995). Some Subjotnian complexes are hidden under the Palaeozoic cover of the E European platform in the Baltic states, including the huge Riga batholith (Soesoo & Niin, 1992; Rämö *et al.* 1996). Intrusive ages range from *c.* 1·65 Ga (1·67 Ga in associated mafic dykes) in southeastern Finland (Vaasjoki *et al.* 1991), to *c.* 1·50–1·47 Ga in central-southern Sweden (Claesson & Kresten 1997; Andersson 1997a, this paper).

The term Subjotnian was coined by Högbom (1909) because these rocks form the basement of the Jotnian (*c.* 1·50–1·27 Ga) sedimentary successions in several areas, particularly in the graben structures associated with the Salmi, Laitila and Strömsbro complexes (Asklund 1930; Amantov *et al.* 1996). In several complexes, however, the association with the Jotnian sediments is lost or only indicated by sandstone dykes. The term Subjotnian is used in this work to denote the specific intracratonic tectonomagmatic episode that produced the array of anorogenic intrusive complexes in the shield at 1·65–1·47 Ga. It is preferred over rapakivi, since it is a chronostratigraphic term rather than a petrographic–geochemical term. Moreover, granitoids of the Swedish complexes generally lack or have poorly developed rapakivi textures, except in the Rödö complex.

The Subjotnian complexes are characteristically bimodal, with a suite of gabbros (norites)–anorthosites–monzodiorites associated with the granites (Haapala & Rämö 1992; Eklund *et al.* 1994; Rämö & Haapala 1995). Numerous dykes are commonly associated with the plutonic complexes; dominantly dolerites and quartz-feldspar porphyries, sometimes comingled in composite dykes (e.g. Laitakari 1969; Lindberg *et al.* 1991; Rämö 1991). The associated mafic rocks constitute only a minor part of the complexes in Finland, but are much more abundant in central Sweden, particularly in the Nordingrå, Ragunda and Mårdsjö complexes (Lundqvist *et al.* 1997; Persson 1999). Syenites and peralkaline felsic rocks are also a significant part of the Swedish complexes (Kornfält 1976; Persson 1997; Andersson 2001).

The present study deals with five minor Subjotnian complexes in central Sweden, namely from the SE to NW (Fig. 2): the Rödö, Mårdsjö, Nordsjö, Mullnäset and Strömsund complexes. The Rödö complex comprises a small rapakivi-textured, shallow-level intrusion, containing abundant miarolitic cavities, filled with calcite, quartz, chlorite and sometimes fluorite. A large number of dykes is associated with the granite, including three main types: quartz-feldspar porphyries (QFPs), dolerites and hybrid porphyries. The hybrid porphyries are interpreted to represent magmatic mixtures of dolerites and QFPs in varying proportions and stages of fractionation (Andersson 1997d). Both the dolerites and the QFPs comprise a suite of dykes with strong chemical fractionation into very evolved compositions.

The other four complexes show some common geological features. They comprise gabbros, syenites and metaluminous to peralkaline granites, although not all rock types occur in each complex. The Mårdsjö complex is dominated by gabbros, showing extensive signs of magma mingling with the granites similarly to the Ragunda gabbros (Persson 1999). Syenites are

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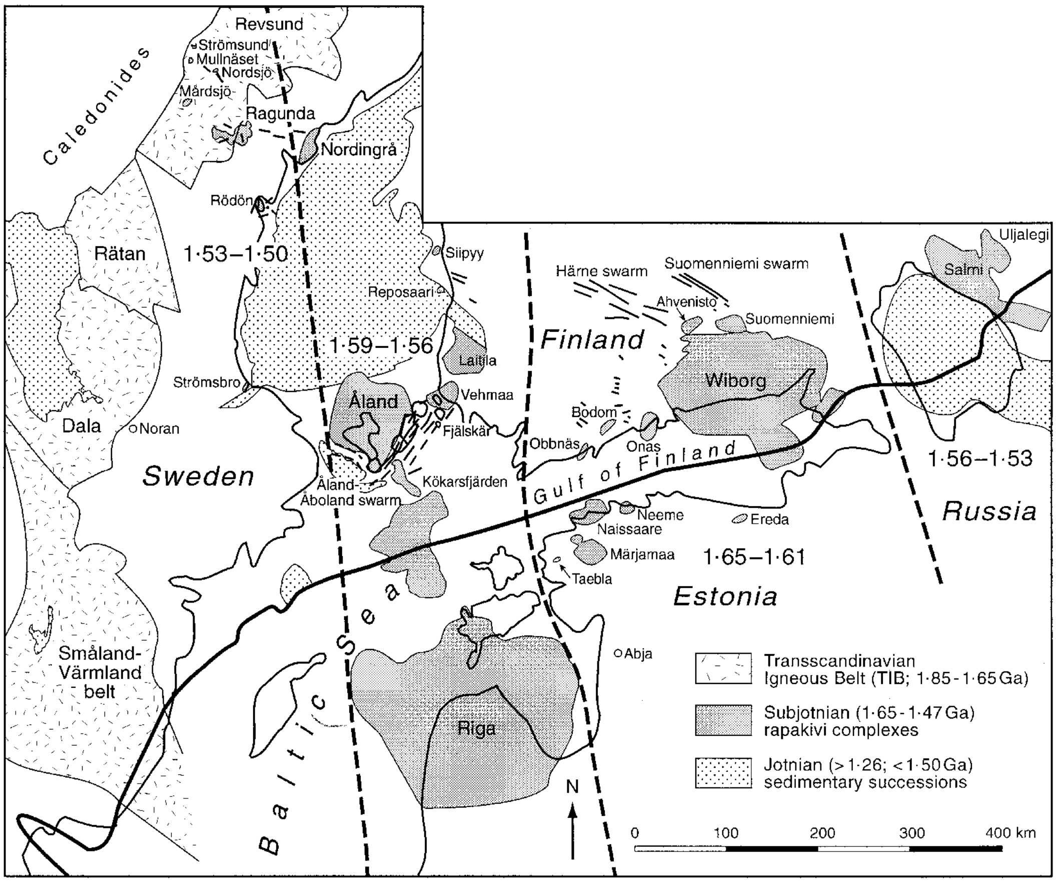

Figure 1 Distribution in space and time of the Fennoscandian rapakivi complexes. Thick slid line marks the northern limit of the Palaeozoic platform cover rocks. The TIB have been subdivided into the Småland–Värmland belt, the Dala province, the Rätan batholith and the Revsund intrusive suite. A systematic younging of the rapakivi complexes to the W is apparent, apart from Salmi in the E. The map is compiled from various sources including Lundegårth *et al.* (1984), Lundqivst *et la.* (1990), Gaál & Gorbatschev (1987); Rämö (1991), and Koistinen (1994). Sources of age data are given in the text.

also present. The Nordsjö complex consists entirely of syenites, but is associated with numerous dykes of plagoclase–porphyritic dolerites, trachytes and peralkaline aplites. The Mullnäset complex consists dominantly of syenites, minor gabbros and both metaluminous and peralkaline granites. The Strömsund complex comprises granites, both metaluminous and peralkaline, with minor quartz syenite. Details of the geology are given in Andersson (1997c).

In general, the granitoids of the Subjotnian complexes in central Sweden show geochemical characteristics (Fig. 3) of A-type magmatism, with relatively high contents of LIL and HFS elements, although commonly lower than in their Finnish counterparts (Andersson 1997b). Typically, the abundance of such elements as Ca, Mg, Al, P and Sr is low, while Fe/Mg, K/Na, and Ga/Al ratios are high relative to orogenic granites. The granitoids are mainly metaluminous, but evolve into the peraluminous field with fractionation. The associated basic rocks are generally quartz-normative tholeites. Contamination of the basic magmas with components of the coeval granitic magmas, particularly in the Rödö complex, produces a monzo-trend in the compositions. The syenitic magmas in the NW complexes fractionate into peralkaline silicic residual compositions. The petrochemical evolution is described in more detail by Andersson (1997c, 2001).

The purpose of the present paper is to put constraints on the variation in intrusive ages and source material for the various Subjotnian intrusive rocks in central Sweden by means of U–Pb zircon dating, Nd, Sr and Pb whole rock, and Pb–feldspar isotopic data, and to discuss the petrogenetic and regional implications of this.

1. Geological setting of the Subjotnian rapakivi complexes in central Sweden

The Proterozoic areas of central Sweden belong to the western part of the central Svecofennian subprovince of Gaál & Gorbatschev (1987), extending eastwards over central to eastern Finland. The Subjotnian intrusions of central Sweden are emplaced in dominantly early Svecofennian metasediments in the E (the Bothnian basin) and rocks of the Revsund granitoid suite in the W (Fig. 2).

The metasediments consist of a thick (up to 10,000 m) marine turbiditic sequence of metagreywackes, metargillites and subordinate mafic metavolcanics, variably metamorphosed from greenschist to upper amphibolite grade (Lundqvist 1987; Lundqvist *et al.* 1990; Gorbatschev 1997). The abundance of meta-arenites and felsic metavolcanics increases southwestwards, suggesting a transition to subaerial conditions of sedimentation (Lundqvist 1987). The age of deposition of the metasediments is Palaeoproterozoic pre-1·87 Ga (Welin 1987) and may have started early, even before 2·03 Ga (Lundqvist *et al.* 1998). The rocks contain detrital zircons mainly in the age range 2·1–1·9 Ga, but also *c.* 30% of Archaean grains are present with ages of 3·0–2·6 Ga (Claesson *et al.* 1993).

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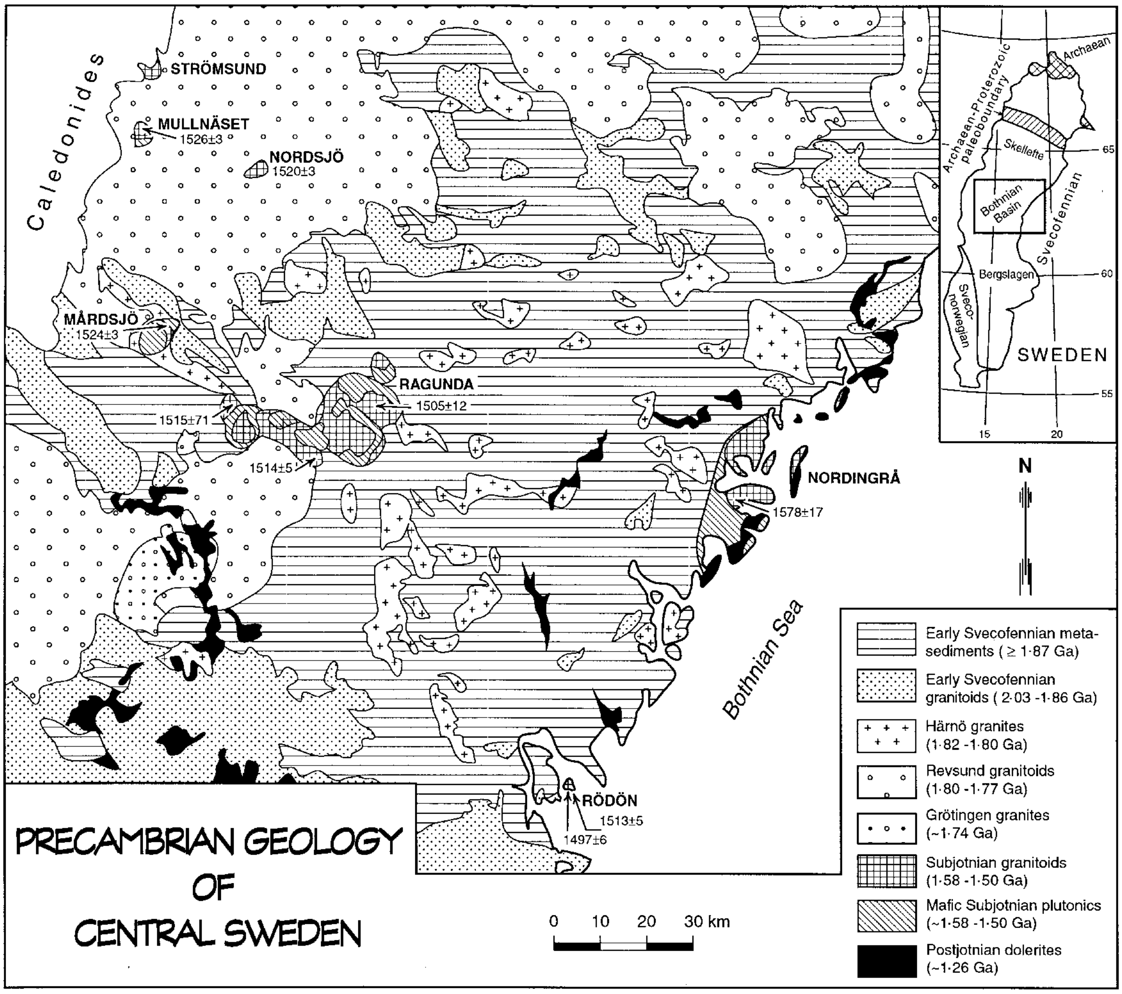


Figure 2 Precambrian geology of central Sweden; modified after Lundegårdh *et al.* (1984), and Lundqvist *et al.* (1990); Subjotnian ages (in Ma on the map) are after Welin & Lundqvist (1984), Welin (1994), Persson (1999), and this paper.

The metasediments are intruded by an early Svecofennian suite of calc-alkaline granitoids and gabbros yielding U–Pb zircon ages in the range 2·03–1·86 Ga (Waström 1993, 1996; Weihed & Vaasjoki 1993; Welin *et al.* 1993; Claesson & Lundqvist 1995; Lundqvist *et al.* 1998). Some of these zircons clearly contain inherited components (Welin *et al.* 1993; Lundqvist *et al.* 1998). Although the granitoids are usually described as intrusive into the sediments, similar rocks in the age range 2·1–1·87 Ga must have been in erosional positions as such ages predominate in the detritus (Claesson *et al.* 1993), and thus presumably also form the basement for the sediments.

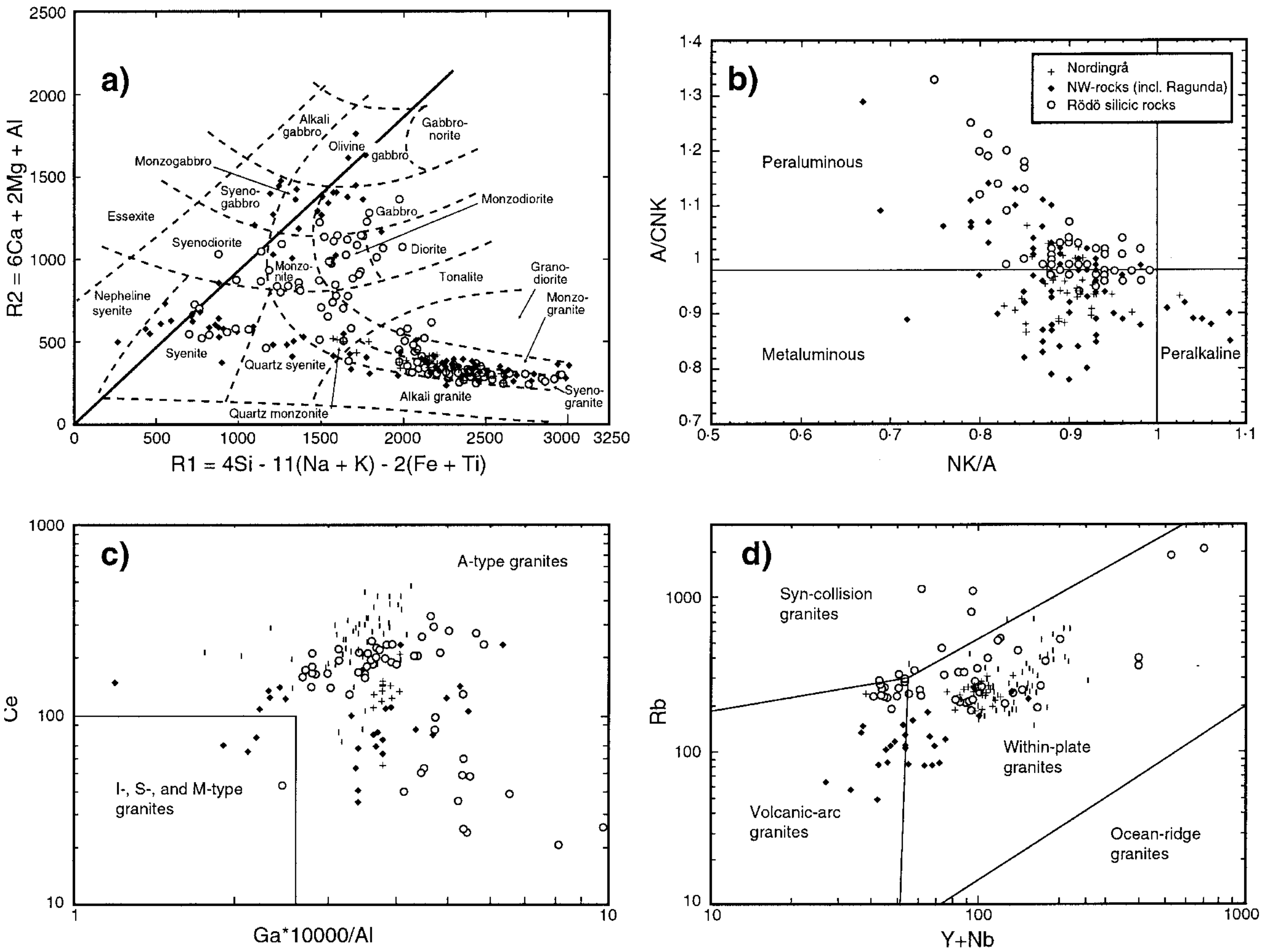
Late-orogenic, peraluminous granites, dominantly of S-type, form a suite of *c.* 1·82–1·79 Ga minor intrusive bodies (Härnö granites) , together with associated pegmatites, mainly in the sedimentary areas (Lundqvist *et al.* 1990; Romer & Smeds 1994, 1997; Claesson & Lundqvist 1995). Their geochemical and isotopic compositions are in concert with an origin as anatectic magmas derived from rocks similar to the surrounding metasediments, with a minor contribution from early Svecofennian granitoid sources (Claesson & Lundqvist 1995).

The Revsund granitoid suite (RGS) covers extensive areas in central Sweden, particularly in the W (Figs 1, 2), and consists of reddish-grey granites to monzodiorites with very restricted abundance of associated basic rocks (Persson 1978; Lundqvist *et al.* 1990; Claesson & Lundqvist 1995; Gorbatschev 1997). No associated supracrustal rocks are known. A group of more evolved, red, coarse prophyritic granites, termed Grötingen granites, and some associated pegmatites, are associated with the RGS (Gorbatschev 1997). Ages of RGS fall in the range 1·80–1·77 Ga (Wilson *et al.* 1985; Patchett *et al.* 1987; Skiöld 1988; Claesson & Lundqvist 1995). Geochemically, the RGS rocks are dominantly alkali-calcic, met- to peraluminous, with an average molecular $Al_2O_3/CaO + Na_2O + K_2O = 1{\cdot}0$, and transitional between I- and A-type affinities (Claesson & Lundqvist 1995). In certain areas, however, the RGS rocks are more peraluminous and S-type in character (Wilson 1980; Armands & Xefteris 1987), presumably due to more extensive assimilation of metasediments. The RGS is considered post-orogenic in relation to the Svecofennian orogeny (Lundqvist *et al.* 1990; Claesson & Lundqvist 1995), and by some authors included in the TIB (Gorbatschev & Bogdanova 1993; Gorbatschev 1997), based on petrographic, geochemical and age similarities.

Postjotnian dolerite dikes and sills of the central Scandinavian dolerite group (Gorbatschev *et al.* 1979, 1987) constitute the youngest areally important rock group in the investigated area (Fig. 2). Rb–Sr whole rock-mineral ages for these intrusions are 1·25–1·16 Ga (Patchett 1978), while U–Pb zircon ages from equivalent lithologies in Finland are 1·27–1·26 Ga (Suominen 1991).

2. Analytical techniques

The studied zircon fractions were recovered using standard Wilfey table, magnetic, heavy liquid and hand-picking



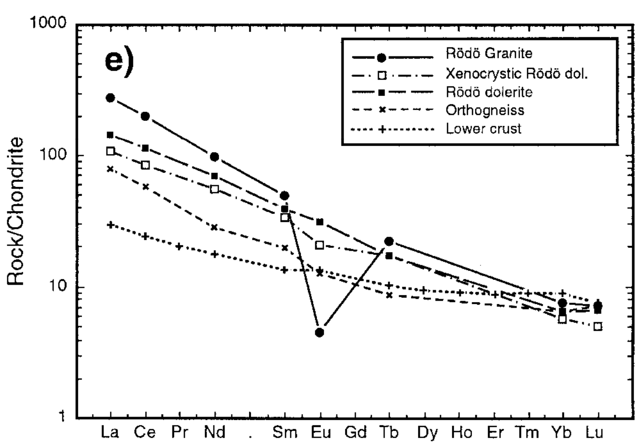


Figure 3 Geochemical plots for central Swedish Subjotnian rocks. Sources of data are Kornfält (1976), Lundqvist *et al.* (1990), Lindh & Johansson (1996), and Andersson (1997c, 2001). In addition to the legend in (b), small vertical bars indicate rocks from the Suomenniemi Subjotnian complex in SE Finland (Rämö 1991). (a) Diagram after De la Roche *et al.* (1980) for rocks of all compositions of the complexes; line with 45° slope is line of silica saturation; the rocks from the NW complexes display a distinct syenitic trend, partly undersaturated; nearly all rocks from the Rödö complex are saturated, more calcic and intermediate; only granitoids are plotted from the Nordingrå complex, falling generally in the syenogranite field. (b) Alumina–alkali saturation diagram for the granitoids; the NW rocks range from metaluminous (syenites or biotite–hornblende granites) to strongly peraluminous, or peralkaline, granitic rocks; no peralkaline rocks occur associated with the Rödö complex, but some quartz–feldspar porphyries (QFPs) are strongly peraluminous; the Nordingrå granites are met- to peraluminous. (c) Ce vs Ga/Al diagram after Whalen *et al.* (1987), showing that essentially all Subjotnian granitoids plot in the field of anorogenic granites. (d) Rb vs Y + Nb diagram after Pearce *et al.* (1984); the Nordingrå, as well as the Suomenniemi, granitoids fall in the within-plate field, while the Rödö granites and NW syenites and granites plot transitional between the fields. (e) REE plot of some rocks of the Rödö complex, an example of early Svecofennian orthogneiss, and the lower crustal average proposed by Taylor & McLennan (1985).

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Table 1 U–Pb zircon results, Rödö granite and Nordsjö syenite

Fraction	Weight (μg)	Number of crystals	U (ppm)	Pb_{tot} (ppm)	Pb_{com} (ppm)	$^{206}Pb/^{204}Pb$ (1)	rad. Pb atomic % $^{206-207-208}Pb$	$^{206}Pb/^{238}U$ (2)	$^{207}Pb/^{235}U$ (2)	$^{207}Pb/^{206}Pb$ (2)	$^{207}Pb/^{206}Pb$ (age)
Rödö granite											
R1 >210 μm LP brown	138·6	2	746	183·8	–	2107	90·2–8·4–1·4	0·2236 ± 9	2·879 ± 14	0·09336 ± 28	1495 ± 4
R2 150–210 μm SP clear	104·0	8	97	17·0	0·526	801	76·7–7·1–16·2	0·1507 ± 31	1·922 ± 42	0·09252 ± 77	1478 ± 12
R3 150–210 μm LP brown	215·5	9	455	94·6	5·48	835	78·4–7·1–14·5	0·1786 ± 2	2·227 ± 10	0·09044 ± 35	1435 ± 6
R4 100–150 μm SP clear	132·9	16	79	25·0	0·381	1578	76·7–7·2–16·1	0·2789 ± 8	3·590 ± 23	0·09335 ± 52	1495 ± 8
R5 100–150 μm LP brown	241·1	23	601	118·7	5·40	1058	77·1–6·9–16·0	0·1687 ± 2	2·083 ± 7	0·08951 ± 28	1415 ± 4
R6 74–100 μm SP clear	180·1	50	200	45·0	2·89	710	78·8–7·2–14·0	0·1927 ± 3	2·436 ± 13	0·09168 ± 42	1461± 7
R7 74–100 μm LP brown	444·3	94	486	99·0	3·37	1456	78·3–7·1–14·6	0·1789 ± 2	2·230 ± 11	0·09043 ± 42	1435 ± 7
R9 >100 μm abraded	60·1	17	237	65·3	1·20	1540	80·0–7·4–12·6	0·2515 ± 5	3·228 ± 10	0·09310 ± 21	1490 ± 3
R10 >100 μm abraded	48·6	11	64	19·8	1·22	413	77·4–7·3–15·3	0·2605 ± 16	3·367 ± 36	0·09375 ± 81	1503 ± 13
Nordsjö syenite											
N1 >210 μm LP	741·7	5	286	68·2	0·002	19909	78·9–7·5–13·6	0·2189 ± 4	2·853 ± 6	0·09452 ± 7	1519 ± 1
N2 >210 μm SP clear	276·9	7	267	78·6	0·190	6611	80·1–7·6–12·3	0·2735 ± 5	3·567 ± 8	0·09460 ± 13	1520 ± 2
N3 150–210 μm LP	187·5	8	440	124·3	0·340	6245	76·5–7·2–16·3	0·2505 ± 3	3·261 ± 6	0·09442 ± 13	1516 ± 2
N4 150–210 μm clear	258·4	15	173	44·8	0·119	6325	81·7–7·7–10·6	0·2446 ± 4	3·192 ± 7	0·09464 ± 14	1521 ± 2
N5 45–74 μm SP clear	101·8	205	95	21·4	0·116	811	82·2–7·7–10·1	0·2137 ± 10	2·745 ± 32	0·09313 ± 90	1491 ± 14
N6 45–74 μm SP dark	52·5	84	612	171·3	14·0	508	76·6–11·8–11·6	0·2285 ± 5	4·849 ± 50	0·15393 ± 132	2390 ± 21

LP = long prismatic, SP = short prismatic, Pb_{tot} = total lead, Pb_{com} = common lead. (1) $^{206}Pb/^{204}Pb$ only corrected for mass discrimination.
(2) Ratios corrected for mass discrimination, blank (0·01 ng U and 0·06 ng Pb),and common Pb ($^{206}Pb/^{204}Pb = 15{\cdot}8$, $^{207}Pb/^{204}Pb = 15{\cdot}2$, $^{208}Pb/^{204}Pb = 35{\cdot}3$). Errors are 2σ.

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Table 2 U–Th–Pb isotopic data for air-abraded single zircon grains from rapakivi rocks, central Sweden

Weight (μg)	U (ppm)	Pb_{tot} (ppm)	Th (ppm)	$^{206}Pb/^{204}Pb$	Measured ratios $^{206}Pb/^{207}Pb$	$^{206}Pb/^{208}Pb$	$^{206}Pb/^{238}U$	Ages (Ma)(1) $^{207}Pb/^{235}U$	$^{208}Pb/^{232}Th$	$^{207}Pb/^{206}Pb$
Mårdsjö granite (sample M14a)										
3·4	89·8	22·4	61·5	325·3	7·491	3·226	1510	1516	1575	1525 ± 8
3·9	95·9	28·6	72·4	512·6	8·496	3·511	1447	1472	1437	1509 ± 6
1·7	160·1	47·9	116·5	905·1	9·462	3·975	1490	1504	1516	1524 ± 4
Mullnäset granite (sample Mu7)										
4·5	58·8	18·2	26·9	385·1	7·919	4·399	1496	1507	1574	1524 ± 6
4·5	67·9	20	32·7	816·8	9·255	5·702	1535	1532	1465	1527 ± 4
4·6	66·6	20	29·6	431·2	8·121	4·731	1530	1529	1550	1526 ± 7

(1) Ages based on radiogenic Pb corrected for mass-fractionation, spike composition, total blank, plus initial common Pb at 1525 Ma according to the Stacey & Kramers (1975) model.

techniques. The multigrain analyses in Table 1 were performed by the first author (UBA) at the Laboratory for Isotope Geology (LIG), Swedish Museum of Natural History in Stockholm, following standard routines (cf. Krogh 1973). The analyses were performed on a multicollector Finnegan MAT-261 mass spectrometer in a static mode. Measured U and Pb blanks during the course of this study were about 0·01 ng and 0·04–0·12 ng, respectively. The lead isotopic composition of feldspars from the Nordsjö syenite (sample N12; Table 5; $^{206}Pb/^{204}Pb = 15{\cdot}8$, $^{207}Pb/^{204}Pb = 15{\cdot}2$, $^{208}Pb/^{204}Pb = 35{\cdot}3$) was used for common lead correction in the rapakivi age calculations. The isotopic composition of the laboratory blank was: $^{206}Pb/^{204}Pb = 18{\cdot}5 \pm 2$, $^{207}Pb/^{204}Pb = 15{\cdot}6 \pm 0{\cdot}2$, $^{208}Pb/^{204}Pb = 38{\cdot}5 \pm 1{\cdot}5$. A mass discrimination of $0{\cdot}1 \pm 0{\cdot}04$ per atomic mass unit was used, based on replicate measurements of the NIST SRM981 lead standard. An average value for 24 parallel analyses of the SRM981 standard during the course of data collection was: $^{206}Pb/^{204}Pb = 16{\cdot}893 \pm 0{\cdot}015$ (2σ), $^{207}Pb/^{204}Pb = 15{\cdot}434 \pm 0{\cdot}016$, $^{208}Pb/^{204}Pb = 36{\cdot}519 \pm 0{\cdot}054$. The calculation of the corrected isotopic ratios, intercept ages and errors reported in this paper was performed using the programs by Ludwig (1993, 1995), applying the decay constants recommended by Steiger & Jäger (1997). Stated errors are 2σ.

Single-grain analyses reported in Table 2 were performed by the second author (LAN) at the US Geological Survey Laboratory in Denver. Based on selection criteria such as crystal completeness, surface reflectance, colour and freedom from inclusions and cracks, approximately 100 of the best crystals were chosen from the 75–150 μm-sized fraction and abraded with pyrite in a pneumatic steel abrader (Krogh 1982). The three cleanest of the rounded and polished crystals from each of the two samples (Mårdsjö and Mullnäset) were chosen for analysis. The small (1·7–4·6 μg) zircon grains were leached in 1N HNO_3 and rinsed in H_2O on a hot plate. Each crystal was transferred by micropipette onto a microbalance for weighing, and finally into a teflon capsule where *c.* 300 μl HF, *c.* 50 μl HNO_3, and 4 μl of ^{205}Pb-^{235}U-^{230}Th spike were added. Six sample and one blank capsules were loaded for simultaneous digestion into a steel-jacketed teflon bomb containing an additional reservoir of HF. Complete digestion was accomplished in a week in an oven maintained at 210°C, after which the solutions were evaporated to dryness, refilled with HNO_3, and re-evaporated to dryness to ensure equilibration with the spike. Without further chemistry the sample was picked up in a drop of silica gel-H_3PO_4 and loaded onto the source filament of a VG Isomass 54E mass spectrometer equipped with Daly multiplier. Pb, U and Th blanks for the total process over the duration of this study averaged 4 ± 1 pg, 1 pg and 2 pg, respectively.

The WR Sm–Nd and Rb–Sr analyses in Table 3 were performed by UBA at LIG. The analyses were performed on a Finnegan MAT 261 multicollector mass spectrometer in a static mode. The measurements were normalised to $^{146}Nd/^{144}Nd = 0{\cdot}7219$ and $^{88}Sr/^{86}Sr = 8{\cdot}375209$. Analytical stability of the instrument was monitored by repeated measurements of the La Jolla Nd standard. The measured $^{143}Nd/^{144}Nd$ ratios of the samples was corrected according to the average measured value for the La Jolla standard during discrete time periods of relative homogeneity, and normalised to its 'accepted' value of 0·511854. Similarly, the SRM987 Sr standard was monitored during the time of study. Average $^{87}Sr/^{86}Sr$ values were calculated and the sample measurements normalised according to an 'accepted' value of 0·710240 (Gladney *et al.* 1990). Errors in the measured $^{147}Sm/^{144}Nd$ values are estimated to be lower than 0·5% (2σ). Repeated measurements of the BCR-1 basalt standard during the course of analysis gave results close to 'accepted' values (cf. Gladney *et al.* 1990), e.g. $^{143}Nd/^{144}Nd = 0{\cdot}512636 \pm 40$ and $^{87}Sr/^{86}Sr = 0{\cdot}705066 \pm 55$ (1s, n = 6). The Rb/Sr ratios were obtained from XRF measurements, and the $^{87}Rb/^{86}Sr$ ratio calculated from the isotopic abundances of Rb and the mass spectrometric measurements of Sr. The reproducibility ranges from 0·12 to 6·29% with an average of 0·75%. The number of significant figures given in Tables 3 and 4 is indicative of the calculated errors for each sample.

The Sm–Nd data reported in Table 4 were obtained by LAN at the Institute of Precambrian Geology and Geochronology in St Petersburg. The samples were totally spiked with a mixed $^{149}Sm/^{146}Nd$ spike solution, measurements performed with a Finnegan MAT 261 in a multicollector static mode, and normalised to $^{148}Nd/^{144}Nd = 0{\cdot}24157$. Analysed blanks are between 0·1 and 0·4 ng for Sm and from 0·1 to 0·5 ng for Nd. 18 measurements on the La Jolla Nd standard during the same period as the samples gave 0·511860 ± 9, close to the 'accepted' value of 0·511854. Duplicate analyses from separate dissolutions of the same samples show good reproducibility (Table 4), and one duplicate sample, M15a, analysed both in St Petersburg and Stockholm shows excellent interlaboratory reproducibility.

The Rb–Sr analytical data in Table 4 were obtained by LAN at the USGS laboratory in Denver. Rb and Sr were determined on separate aliquots of the samples by conventional isotope dilution methods (^{87}Rb and ^{84}Sr spikes), using HF and H_2SO_4 dissolution. Sr isotopic composition was measured using a fully automated multicollector MAT-262 mass-spectrometer in a static mode. Data collected on the USGS Sr standard EN-1 during the course of this study give 0·709060 ± 0·000008. Replicate analyses for granitic rocks performed in the

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Table 3 Nd and Sr isotopic results, Rödö area

	SiO_2 (XRF)	MgO (XRF)	Sm (MS)	Nd (MS)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	$\varepsilon_{Nd}(1{\cdot}50)$	$\varepsilon_{Nd}(t)$	T(DM)	T(CHUR)	Rb (XRF)	Sr (XRF)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	$^{87}Sr/^{86}Sr(1{\cdot}50)$	$^{87}Sr/^{86}Sr(i)$
Rödö granites																
R126a	73·4	0·36	13·54	88·18	0·0928	0·511274 (6)	−6·65	→	2·23	1·99	223	58	10·74	0·913555 (15)	0·6823	→
R127:1 + 2	73·2	0·38	12·86	80·98	0·0960	0·511295 (12)	−6·86	→	2·27	2·03	226	96	8·59	0·881124 (13)	0·6962	→
R133:2	72·2	0·36	12·06	77·73	0·0937	0·511296 (6)	−6·40	→	2·22	1·98	208	151	5·80	0·820179 (21)	0·6953	→
R138	70·7	0·37	11·61	74·13	0·0947	0·511254 (13)	−7·41	→	2·30	2·06	227	131	5·429	0·811473 (27)	0·6946	→
Quartz-feldspar porphyries																
R73a	74·4	0·12	12·82	77·46	0·1001	0·511336 (8)	−6·85	→	2·30	2·05	281	41	30·0	1·235583 (20)	0·5897	→
R96a + b	69·9	0·20	13·96	77·77	0·1076	0·511384 (8)	−7·36	→	2·37	2·11	208	56	14·5	0·978438 (16)	0·6663	→
R117g1	72·8	0·17	4·10	11·24	0·2204	0·512606 (15)	−5·20	→	—	—	1870	35	218	4·728422 (156)	0·0352	→
R160	75·9	0·08	24·47	134·45	0·1100	0·511405 (4)	−7·41	→	2·42	2·16	237	< 10	41·0	1·447649 (211)	0·5650	→
R167:2	73·7	0·17	11·95	33·24	0·2174	0·512598 (8)	−4·78	→	–	–	451	41	43·86	1·563732 (89)	0·6194	→
R144a + b	77·2	0·01	11·55	48·05	0·1454	0·511800 (8)	−6·51	→	2·81	2·48	251	23	30·57	1·092798 (148)	0·4346	→
Aplite																
R126f	75·9	0·08	10·65	34·18	0·1883	0·512214 (11)	−6·68	→	—	—	540	< 10	239	5·198968 (15521)	0·0537	→
Rödö dolerites																
R101a + b	59·6	0·61	14·38	82·42	0·1055	0·511346 (6)	−7·69	→	2·40	2·15	237	89	6·4582	0·845026 (22)	0·7060	→
R111	53·4	6·91	5·56	30·29	0·1110	0·511424 (12)	−7·23	→	2·41	2·15	41	564	0·2421	0·708770 (9)	0·7036	→
R134a	51·7	2·69	9·31	50·89	0·1106	0·511353 (3)	−8·54	→	2·51	2·27	185	265	2·218	0·758789 (28)	0·7110	→
R14c	53·5	3·32	7·61	41·41	0·1110	0·511340 (18)	−8·87	→	2·54	2·30	48	522	0·1398	0·708316 (14)	0·7053	→
Hybrid porphyries																
R86a	61·5	0·45	13·49	73·38	0·1040	0·511367 (10)	−6·99	→	2·33	2·08	178	63	8·60	0·869521 (38)	0·6844	→
R133b1	56·2	2·43	7·70	42·89	0·1085	0·511414 (19)	−6·94	→	2·37	2·11	129	455	0·9920	0·727760 (20)	0·7064	→
R68(1)	68·8	1·23	12·07	75·13	0·0971	0·511262 (5)	−7·72	→	2·33	2·10	172	270	2·4541	0·760926 (14)	0·7081	→
R68(2)			12·12	75·31	0·0972	0·511305 (8)	−6·89	→	2·28	2·03			2·4542	0·761140 (13)	0·7083	→
R115a1	59·1	3·12	8·24	47·50	0·1054	0·511358 (8)	−7·44	→	2·38	2·13	147	446	2·408	0·770593 (15)	0·7188	→
R151h	62·3	0·78	10·39	56·69	0·1108	0·511370 (5)	−8·25	→	2·49	2·24	225	243	1·012	0·733373 (15)	0·7116	→
Hybrid enclave																
R18Enk1	57·6	1·64	8·88	37·88	0·1416	0·511709 (23)	−7·55	→	2·85	2·56	538	65	20·74	1·10547 (31)	0·6590	→
Sedimentary gneisses																
R91	68·7	2·15	5·29	28·88	0·1106	0·511418 (6)	−7·27	−2·92	2·41	2·15	142	254	1·7041	0·752605 (29)	0·7159	0·7063
R89	68·0	2·29	5·24	28·51	0·1110	0·511397 (5)	−7·76	−3·43	2·45	2·20	105	253	1·505	0·745135 (26)	0·7127	0·7042
Palingenic granites																
R22	67·3	1·95	5·34	29·53	0·1092	0·511435 (13)	−6·67	−3·04	2·35	2·09	130	259	1·708	0·751064 (18)	0·7143	0·7063
R38b	69·4	1·86	6·07	31·73	0·1156	0·511501 (13)	−6·61	−3·25	2·40	2·13	146	215	2·206	0·766978 (35)	0·7195	0·7092
Orthogneisses																
R94a + b	69·1	1·14	4·07	23·79	0·1035	0·511560 (13)	−3·12	1·59	2·05	1·76	101	287	1·026	0·729691 (19)	0·7076	0·7018
R15	70·0	1·23	10·06	53·09	0·1145	0·511558 (12)	−5·28	−1·13	2·29	2·00	164	198	2·947	0·782908 (13)	0·7195	0·7027

Initial values calculated at 1·89 Ga for the sedimentary gneisses and orthogneisses and at 1·82 Ga for the palingenic granites. Measured $^{143}Nd/^{144}Nd$ is corrected according to average La Jolla, and measured $^{87}Sr/^{86}Sr$ ratios to average SRM987 standard measurements, at the time of analysis. $^{147}Sm/^{144}Nd$: relative std dev. estimated to be better than 0·5%. Rb/Sr measured by XRF. Normally 4 measurements on 2 briquettes. Average relative 'spread' 0·012%. Significant number of digits in relation to 'spread' of data.

Table 4 Nd–Sr isotopic results, Jämtland-Ångermanland

Sample no.	Description	SiO_2 wt%	Sm ppm	Nd ppm	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	$\varepsilon_{Nd}(1{\cdot}52)$	$\varepsilon_{Nd}(t)$	T_{DM}	T_{CHUR}	Rb	Sr	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Sr_i (1·52)	Sr_i	$\varepsilon_{Sr}(1{\cdot}52)$
M18	Mårdsjo gabbro	55·2	5·80	33·32	0·1052	0·511317 (4)	−7·96	←	2·43	2·19							
M15a(1)	Mårdsjö gabbro	55·0	6·87	40·61	0·1026	0·511286 (3)	−8·06	←	2·42	2·18	65·63	715·13	0·2656	0·70884 (07)	0·7030	→	5
M15a(2S)	Mårdsjö gabbro	55·0	6·90	40·40	0·1033	0·511295 (8)	−8·03	←	2·42	2·18			0·269	0·70894 (17)	0·7031	→	5
M20	Monzordiorite, Mårdsjö	58·4	6·68	43·02	0·0942	0·511180 (10)	−8·50	←	2·38	2·16							
M11(1)	Mårdsjö syenite	59·7	11·32	67·21	0·1022	0·511298 (6)	−7·75	←	2·39	2·15							
M11(2)	Mårdsjö syenite	59·7	9·87	58·48	0·1024	0·511306 (7)	−7·63	←	2·39	2·14							
M14a	Märdsjö granite	65·5	11·90	76·34	0·0946	0·511257 (6)	−7·06	←	2·29	2·05	153·57	195·03	2·2882	0·75275 (08)	0·7027	→	1
M15b(1)	Mårdsjö granite	67·7	9·37	62·45	0·0910	0·511196 (7)	−7·55	←	2·30	2·07							
M15b(2)	Mårdsjö granite	67·7	9·15	65·15	0·0852	0·511140 (5)	−7·52	←	2·26	2·04							
N7	Nordsjö dolerite	52·1	7·87	39·39	0·1211	0·511551 (6)	−6·50	←	2·47	2·18							
N9(1)	Nordsjö syenite	61·2	10·19	70·09	0·0811	0·511264 (6)	−5·67	←	2·16	1·92							
N9(2)	Nordsjö syenite	61·2	12·02	79·24	0·0920	0·511287 (7)	−5·97	←	2·20	1·96							
N11	Nordsjö syenite	58·9	10·60	57·05	0·1127	0·511496 (6)	−5·93	←	2·34	2·07							
Mu11a(1)	Mullnäset gabbro	57·1	6·68	30·01	0·1350	0·511688 (6)	−6·53	←	2·64	2·34	43·77	456·39	0·22775	0·70943 (08)	0·7034	→	9
Mu11a(2)	Mullnäset gabbro	57·1	6·72	30·24	0·1347	0·511684 (5)	−6·55	←	2·64	2·33							
Mu5	Mullnäset syenite	63·0	8·32	42·69	0·1183	0·511536 (6)	−6·23	←	2·42	2·13							
Mu12	Mullnäset syenite	60·8	5·95	28·78	0·1254	0·511578 (4)	−6·80	←	2·54	2·26							
Mu8	Mullnäset granite	69·9	8·06	43·08	0·1135	0·511460 (5)	−6·79	←	2·42	2·15	133·04	115·39	3·3571	0·77290 (10)	0·6994	→	−44
S7	Strömsund granite	73·4	5·99	36·66	0·0991	0·511263 (5)	−7·83	←	2·37	2·14							
S6	Strömsund granite	70·3	8·71	47·46	0·1113	0·511367 (7)	−8·18	←	2·51	2·26							
S4	Strömsund granite	71·8	9·31	50·12	0·1127	0·511416 (14)	−7·49	←	2·47	2·21							
M1	Svecof. metasediment	66·3	3·84	21·72	0·1072	0·511363 (8)	−7·45	−3·16	2·41	2·16	150·43	244·52	1·7878	0·75308 (08)	0·7139	0·7044	162
M7(S)	Early Svecof. granitoid	63·2	6·02	33·44	0·1088	0·511463 (7)	−5·81	−1·60	2·30	2·03			2·180	0·76092 (22)	0·7134	0·7016	151
M5(1)	Early Svecof. granitoid	69·6	4·46	23·21	0·1166	0·511616 (9)	−4·34	−0·51	2·24	1·94	290·50	101·27	8·4646	0·91111 (12)	0·7258	0·6809	337
M5(2)	Early Svecof. granitoid	69·6	4·57	24·00	0·1154	0·511596 (7)	−4·51	−0·61	2·25	1·95							
Mu2	Revsund granitoid	64·9	9·13	46·69	0·1186	0·511682 (7)	−3·45	−0·72	2·18	1·86	213·25	197·30	3·1500	0·78241 (10)	0·7135	0·7013	156
Mu4	Dark Revsund granitoid	57·3	10·74	52·52	0·1240	0·511762 (6)	−2·93	−0·39	2·18	1·83							
N2	Revsund granitoid	67·6	8·13	43·11	0·1143	0·511630 (6)	−3·62	−0·74	2·17	1·86							
N8	Pink Revsund granitoid	64·6	7·02	35·36	0·1203	0·511707 (5)	−3·29	−0·62	2·18	1·85							
BCR-1	USGS Basalt standard		6·40	27·96	0·1388	0·512646 (6)											
BCR-1	USGS Basalt standard		6·44	28·26	0·1383	0·512640 (8)											
BCR-1	USGS Basalt standard		6·67	28·56	0·1390	0·512643 (9)											
Literature values (Gladney *et al.* 1990)			*6·59*	*28·8*	*0·1371*	*0·512640*								*0·70501*			

(1) and (2) are duplicate analyses. (S) are analysed at LIG-NRM in Stockholm, the rest at IPGG St Petersburg. SiO_2 is extracted from Andersson (1997c). Rb and Sr were determined by IDMS. Numbers in parenthesis denote within run 2σ(m) in $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ ratios in the fifth and sixth decimals. T_{DM} after DePaolo (1981); $\varepsilon_{DM}(t) = 0{\cdot}25t^2 - 3t + 8{\cdot}5$. $f_{Sm/Nd} = [^{147}Sm/^{144}Nd(sample)/^{147}Sm/^{144}Nd(CHUR\text{-}1)] - 1$. Measurements on the WR-standard BCR-1 in St Petersburg and Stockholm are included. $\varepsilon_{Nd}{}^{t}$ and Sr_i at 1·89 Ga for M1, M5 and M7, and at 1·79 Ga for Mu2, Mu4, N2 and N8.

Table 5 Pb-feldspar isotopic results, Jämtland-Ångermanland, and Rödö whole-rock Pb results

Sample	Mineral	Rocktype	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
Jämtland-Ångermanland					
M15a	plagioclase	Subjotnian gabbro	15·308	15·019	35·037
M11	mesopherthite	Subjotnian syenite	15·758	15·130	35·057
M14a	perthite	Subjotnian granite	15·941	15·215	35·662
Mu11a	plagioclase	Subjotnian gabbro	15·809	15·217	35·231
Mu12	mesoperthite	Subjotnian syenite	15·286	15·018	34·971
Mu8	perthite	Subjotnian granite	16·610	15·191	35·878
N12(1)	mesoperthite	Subjotnian syenite	15·856	15·201	35·359
N12(2)	mesoperthite	Subjotnian syenite	15·790	15·187	35·296
M1	plagioclase	Early Svec. metased.	16·462	15·452	36·295
M5	feldspar	Early Svec. granitoid	16·879	15·500	35·932
Mu2(1)	feldspar	Revsund granitoid	16·893	15·521	35·706
Mu2(2)	feldspar	Revsund granitoid	17·220	15·542	35·835

(1) and (2) denote replicate analyses.

Rödö

Sample	Rocktype	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
R138	Rödö granite	19·47	15·49	40·58
R126a	Rödö granite	20·617	15·673	42·610
R127:1 + 2	Rödö granite	20·169	15·649	41·468
R133:2	Rödö granite	18·856	15·558	40·345
R73a	Qz-fsp porphyry	21·540	15·786	42·912
R96a + b	Qz-fsp porphyry	18·88	15·45	38·38
R117g1	Qz-fsp porphyry	34·554	16·520	55·166
R126f	Rödö aplite	29·753	16·587	45·993

Samples with two decimals have somewhat lower precision.

laboratory indicate an uncertainty of 1·5% in the $^{87}Rb/^{86}Sr$ ratios. Rb and Sr total chemistry blanks were lower than 5 ng.

Pb isotopic analyses reported in Table 5 were also performed by LAN in Denver on feldspar fractions from the four Subjotnian complexes in the NW and selected country rocks. 50 mg to 100 mg of the cleanest grains were hand-picked from each fraction, leached in HNO_3 and HCl, and digested overnight in a mixture of HF and HNO_3 in tightly capped teflon vials. The measured Pb blank for the total procedure during this study was 100 pg. Therefore, assuming that the Pb content of the feldspars is higher than 1 ppm, no correction for lead blanks was necessary. The analyses were performed on the MAT-262 mass-spectrometer in a static mode using silica-gel emitter. Mass discrimination of $0·05 \pm 0·02$% per amu, determined by 30 runs of the NIST Pb standard SRM982, was used to correct the data. Average ratios for SRM982 measured during the course of these analyses were: $^{206}Pb/^{204}Pb = 36·720 \pm 0·024$ (2σ), $^{207}Pb/^{204}Pb = 17·143 \pm 0·018$, and $^{208}Pb/^{204}Pb = 36·692 \pm 0·047$. Within-run errors for all the sample measurements were better than 0·005% (2σ). An external 2σ error of 0·05% per amu was accepted for the samples, based on the SRM982 measurements.

Some of the Pb fractions collected during the chemical preparation of the Sr–Nd whole rock samples (Table 3), were measured by the third author (KB) at LIG (Table 5). After standard ion exchange procedures, the evaporated fractions were loaded onto Re single filaments with a mixture of Si-gel and 0·25N H_3PO_4. Isotopic measurements were made with a Finnegan MAT-261 mass spectrometer in a static mode. Corrections for mass fractionation were made relative to the NIST standard SRM981 measured at different temperatures and amounts according to a specific scheme developed by Hans Schöberg (H. Schöberg pers. comm. 1995).

3. Results

3.1. U–Pb zircon age determinations

3.1.1. The Rödö granite. Nine fractions of zircons of different types and sizes were handpicked. Two of these were air-abraded. The zircons were of two different morphological types, one long prismatic with a length-to-breadth ratio of around 4:1, and one short prismatic with a length-to-breadth ratio of around 1:1. The former were generally brown and turbid, but with well-developed crystal forms. The latter were clear, slightly yellow, also with sharp crystal forms. The analytical results are shown in Table 1. All nine fractions give a poor discordia with an upper intercept age of 1500 ± 26 Ma (MSWD = 22) (Fig. 4a). The data for the most discordant fraction (R2) are of lower analytical quality and a little way off the discordia line to the right. If this fraction is omitted from the discordia calculation, the upper intercept age becomes 1504 ± 19 Ma (MSWD = 10). The relatively large MSWD value indicates excess scatter among the fractions that cannot be accounted for by analytical errors. If all fractions with positions slightly to the right of the discordia (R1, R2, R6 and R10) are omitted from the age calculation, the age obtained is, within errors, the same as before (1493 ± 12 Ma, MSWD = 1·8). The two abraded fractions (R9 and R10) are most concordant (3 and 0·3% discordance, respectively), and yield $^{207}Pb/^{206}Pb$ ages of 1490 and 1503 Ma, respectively. An age calculation using only these two abraded fractions gives 1507 ± 47 Ma. One fraction is above the concordia curve (R4) (Fig. 4a). If this fraction together with those that are slightly displaced to the right (R1, R2, R6) are omitted the statistically most well constrained age is calculated (1497 ± 6 Ma, MSWD = 1·3; lower intercept is 229 ± 21).

3.1.2. The Nordsjö syenite. Six fractions were handpicked

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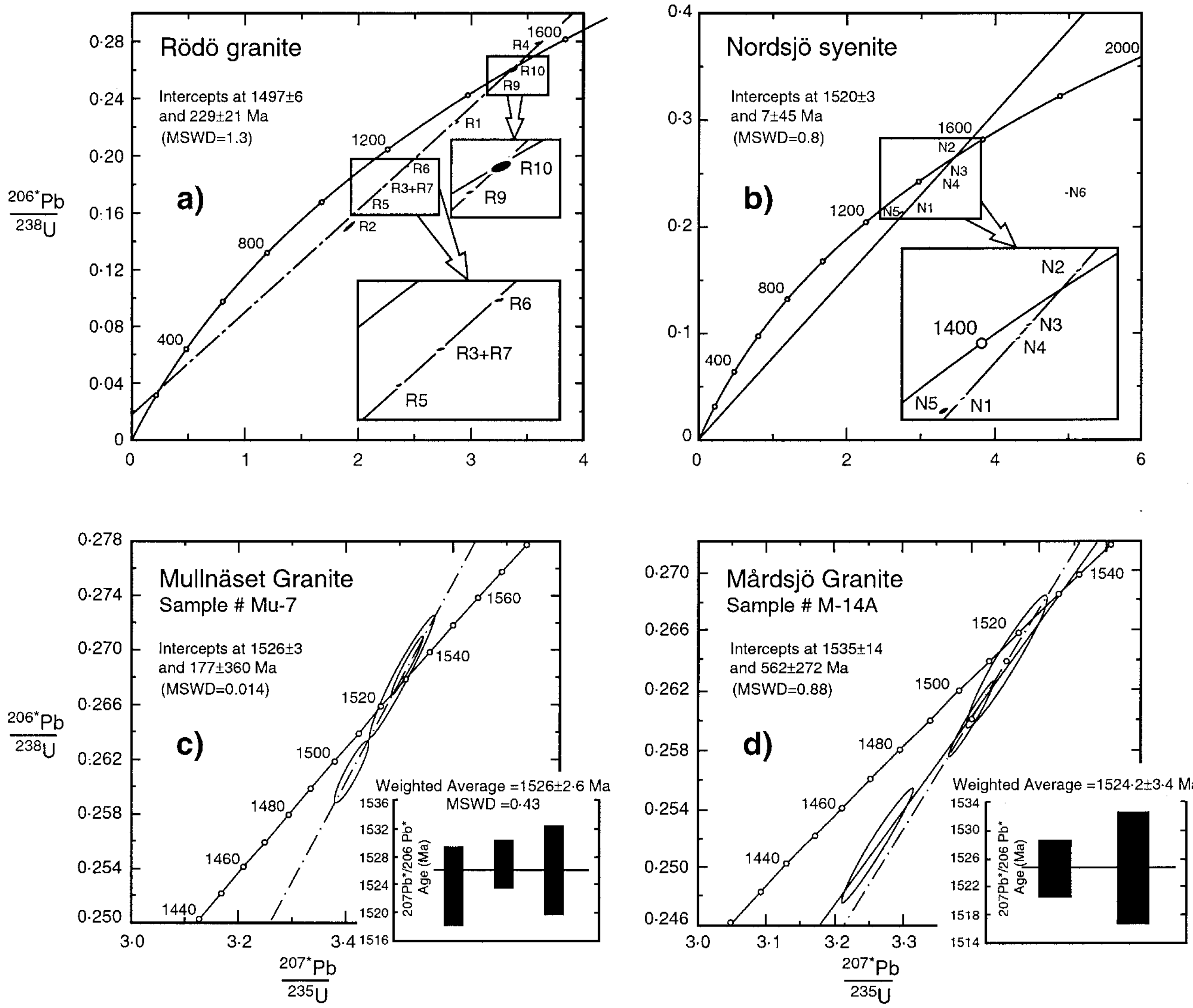

Figure 4 U–Pb zircon concordia diagrams; preferred ages are indicated. (a) The Rödö granite; nine fractions were analysed; two air-abraded (R9 and R10); the analyses show some scatter around the discordia; the preferred age is obtained by omitting fractions R1, R2, R4 and R6. (b) The Nordsjö syenite; six fractions were analysed; the discordia were calculated omitting fractions N5 and N6; fraction N6 is strongly offset to the right of the rest of the analysed fractions, with a $^{207}Pb/^{206}Pb$ age of 2390 Ma. (c, d) Concordia diagrams for Mullnäset and Mårdsjö granite, respectively; analyses are performed on air-abraded single grains. Mullnäset granite (c): all three grains follow a discordia line with an upper intercept of 1526 ± 3 Ma. Mårdsjö granite (d): two analyses are near concordant and one slightly less so; regression through all three analyses gives an upper intercept of 1535 ± 14 Ma; the weighted average of the $^{207}Pb/^{206}Pb$ age of the two most concordant grains (1524 ± 3 Ma) is, however, considered a more reliable estimate of the crystallisation age; this age is, within errors, the same as the regressed age.

from different size groups (Table 1). The majority of crystals were clear and short-prismatic, with a length-to-breadth ratio of about 2:1. Less abundant were long-prismatic (*c.* 5:1) crystals clear to brownish in colour. Fraction N6 is especially rich in U and Pb. The fractions N1–N5 adhere relatively well to a discordia line with an upper intercept age of 1522 ± 22 Ma (MSWD = 5·2). The fraction N5 has a large error due to low U and Pb contents and low $^{206}Pb/^{204}Pb$ (Table 1). A discordia calculated omitting N5 (Fig. 4b), gives a more precise age of 1520 ± 3 Ma (MSWD = 0·8). This is the preferred value for the crystallisation age of the Nordsjö syenite.

The fraction N6 shows a completely different isotopic composition, and gives a $^{207}Pb/^{206}Pb$ age of about 2390 Ma. The zircons of this fraction are small (45–74 μm), grey-black, turbid and elongated, with a length-to-breadth ratio of about 3:1. They have surprisingly well-preserved crystal faces, although usually somewhat rounded. The data point is highly discordant and the calculated $^{207}Pb/^{206}Pb$ age gives a minimum age of an older component present. These crystals thus show that inherited material, at least *c.* 870 Ma older than the rock itself, was involved in the syenite formation.

3.1.3. The Mårdsjö and Mullnäset granites. The Mårdsjö granite (M14a) and Mullnäset granites (Mu7), were selected for single grain zircon analysis (Table 2), to minimise the risk of including inherited components. The analyses plot nearly concordant (Fig 4c–d). For the Mullnäset granite, all three analysed zircons follow the calculated discordia line closely, giving an upper intercept age of 1526 ± 3 Ma (MSWD = 0·014), regarded as the crystallisation age of the intrusion.

A discordia through all three of the analysed Mårdsjö zircons gives an upper intercept age of 1535 ± 14 Ma (MSWD = 0·88). One grain is, however, more discordant than the other two, and a weighted average of the $^{207}Pb/^{206}Pb$ ages of the two most concordant points gives 1524 ± 3 Ma. This is regarded as the best estimate of the true crystallisation age of the Mårdsjö granite.

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3.2. Nd isotopes

Whole-rock Sm–Nd analyses of Subjotnian and Svecofennian rocks (gneisses) from the Rödö area are reported in Table 3, and from an area in the NW in Table 4. The measured $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ ratios are low for the Rödö granites (0·511254–0·511296 and 0·0928–0·0960, respectively), while ranging up to very high values for quartz-feldspar porphyries (0·512606 and 0·2204, respectively). Mafic and hybrid rocks are intermediate. For the NW complexes there are considerable overlaps in measured values between rock types and complexes.

Furthermore, there is a considerable overlap in contents of Sm and Nd between the different rock types, although the granites and felsic QFPs tend to be higher (< 134 ppm Nd) and basic-intermediate rocks lower (> 30 ppm Nd). The low REE contents and high Sm/Nd ratios in some Rödö QFPs (evolved) are due to strong fractionation of LREE-phases (Andersson 1997d, 2001). The REE contents of mafic Subjotnian rocks are high (cf. also Rämö 1990, 1991; Andersson 1997b; Neymark *et al.* 1994; Lindh *et al.* 2001), as compared with mafic rocks in general (Faure 1986; cf. e.g. the Postjotnian and other dolerites, Rämö 1990; Johansson & Johansson 1990; Patchett *et al.* 1994).

The initial ε_{Nd} values of the Rödö rocks are very low, irrespective of chemical composition, and range in the silicic rocks between −7·4 and −4·8, and in the dolerites and hybrid porphyries between −8·9 and −6·9. The basic rocks of the suite thus actually tend to have somewhat lower ε_{Nd} values than the silicic. For all rocks of the NW complexes, initial ε_{Nd} values are also consistently very low, ranging between −8·5 and −5·7 (Table 4). Irrespective of chemistry, rocks from the same complex have similar initial ε-values. In the Mårdsjö and Norsjö samples, the basic rocks have even slightly lower ε_{Nd} (1·52) values than the silicic-intermediate, whereas for the Mullnäset complex the opposite relation is observed. There are also some differences between the complexes. The most negative ε_{Nd} (1·52) values are found in the Strömsund and Mårdsjö complexes: −8·5 to −7·1, whereas the Mullnäset and Nordsjö complex has the least negative values: −6·8 to −5·7. Thus, the observed differences seem to be related more to geography of the complexes, than to rock composition. Calculated depleted mantle ages (T_{DM}; DePaolo 1981), as well as CHUR-ages, are much older than crystallisation ages of these rocks. T_{DM} values vary between 2·81 and 2·16 Ga (Tables 3 and 4).

The range of initial ε_{Nd} values, together with data from other Fennoscandian rapakivi complexes and other relevant rocks (at t = 1·51 Ga; the mean age of Subjotnian rocks in central Sweden), is shown in Figure 5. The evolution of ε_{Nd} with time for the different rock groups is shown in Figure 6.

The initial ε_{Nd} values of 1·53–1·50 Ga Subjotnian rocks from central Sweden essentially range between −10 and −5 (including one gabbro as low as −10·1 in the Ragunda complex; Persson 1997), while those for the 1·58 Ga Nordingrå complex in the same region are more radiogenic (−3·5 to −1; Fröjdö *et al.* 1996; Lindh & Johansson 1996; Lindh *et al.* 2001). The central Swedish Subjotnian rocks have completely overlapping Nd isotope composition with the 1·56–1·53 Ga Salmi rocks emplaced at the Archaean–Svecofennian border in the E (Neymark *et al.* 1994; Rämö 1991). The Nordingrå rocks have overlapping evolution with the 1·59–1·56 Ga SW Finnish and Riga, and the 1·65–1·62 Ga SE Finnish and Estonian Subjotnian complexes (Figs 5, 6).

The ε_{Nd} (1·52–1·50) values of the sedimentary gneisses and the associated palingenic granites (local melts from the sediments) are all similar to those of the Subjotnian rocks, whereas the early Svecofennian granitoids tend to be higher (Tables 3 and 4). The sedimentary gneisses analysed in this study (Tables 3 and 4) fall in the interval of previously analysed corresponding rocks of the Bothnian basin (Claesson 1987; Patchett *et al.* 1987; Claesson & Lundqvist 1995). Similarly, the orthogneisses also fall in the interval of previously analysed early Svecofennian granitoids (cf. Wilson *et al.* 1985, 1987; Huhma 1986; Patchett & Kuovo 1986; Patchett *et al.* 1987; Valbracht 1991; Welin *et al.* 1993; Claesson & Lundqvist 1995; Billström & Weihed 1996; Lahtinen & Huhma 1997), even if sample R15 is on the lower edge. In the Bothnian basin of central Sweden there is a slight tendency towards lower ε_{Nd} values in the early Svecofennian granitoids (Welin *et al.* 1993; Claesson & Lundqvist 1995; this paper), compared with other parts of the Svecofennian (cf. Fig. 5, Table 6).

Four analysed rocks belonging to the Revsund granitoid suite (Table 4), show ε_{Nd} (1·52) values significantly higher than those recorded for the Subjotnian rocks, and exhibit very coherent initials (−0·7 to −0·4 at 1·79 Ga). Generally, initial ε_{Nd} values for RGS rocks vary between −3·4 and +0·5 (Wilson *et al.* 1985; Patchett *et al.* 1987; Claesson & Lundqvist 1995) and follow a more radiogenic time-integrated evolution than both the metasediments of the Bothnian basin and Subjotnian complexes, but similar to that of the early Svecofennian granitoids (Fig. 6).

The Sm–Nd data of the Rödö rocks are plotted in an isochron diagram in Figure 7a. Only the silicic rocks (unfilled squares) are used in the calculation, which gives an age of 1608 ± 45 Ma (MSWD = 0·53). This low MSWD indicates that the isochron is of acceptable quality. All samples but one are within analytical error on the isochron line. The age obtained is significantly (beyond error limits) higher than the 1497 ± 6 Ma age derived from the U–Pb zircon data. The data from the basic and intermediate rocks do not give any meaningful Sm–Nd age results.

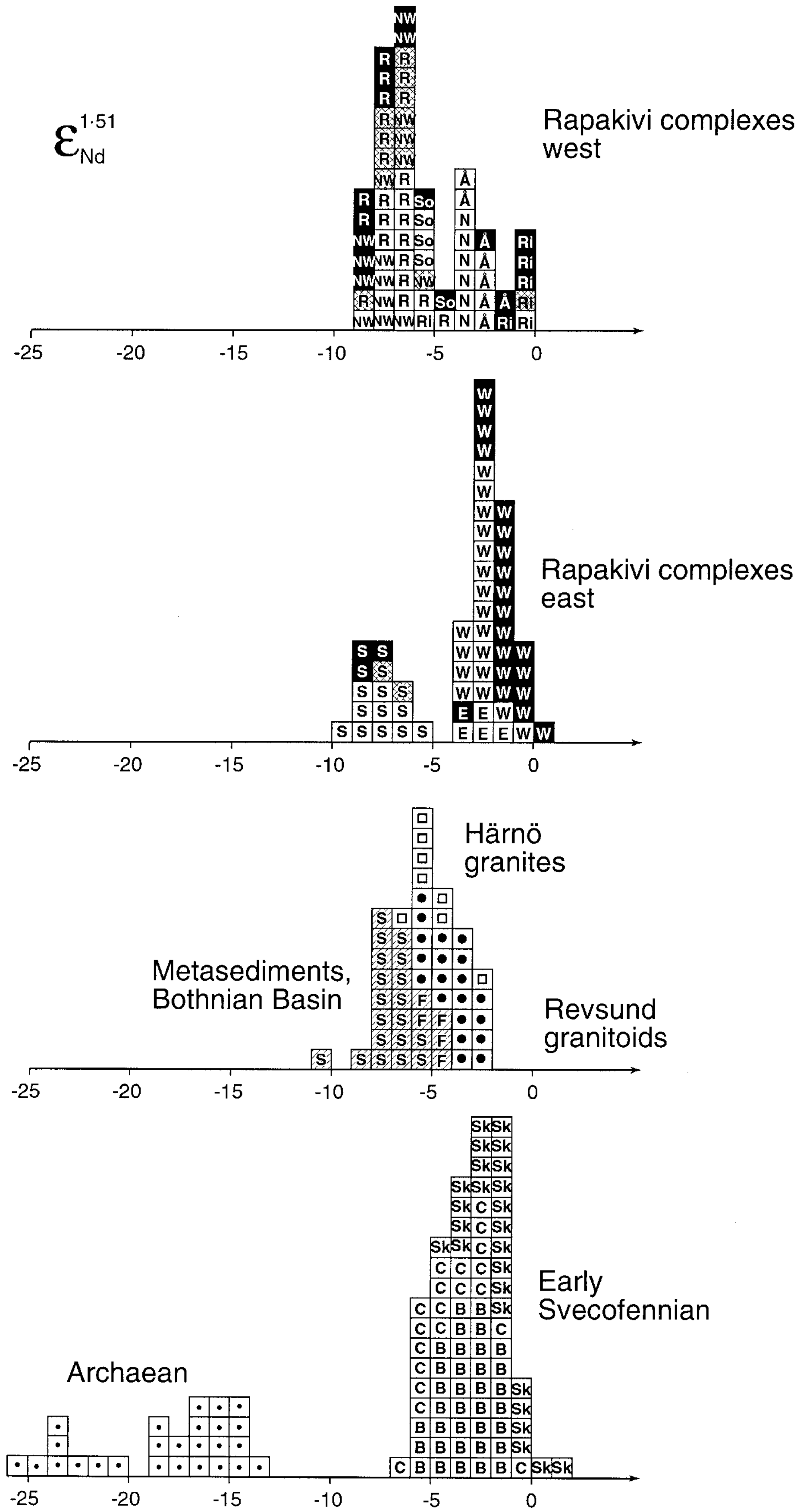
3.3. Sr isotopes

All samples from the Rödö and seven from the area in the NW were also analysed for Sr isotopes (Tables 3 and 4). Some of the Rödö QFPs and the aplite are strongly radiogenic with measured $^{87}Sr/^{86}Sr$ values up to 5·2 and $^{87}Rb/^{86}Sr$ ratios of up to 239, while the $^{87}Rb/^{86}Sr$ ratio of the granites reaches 10·7.

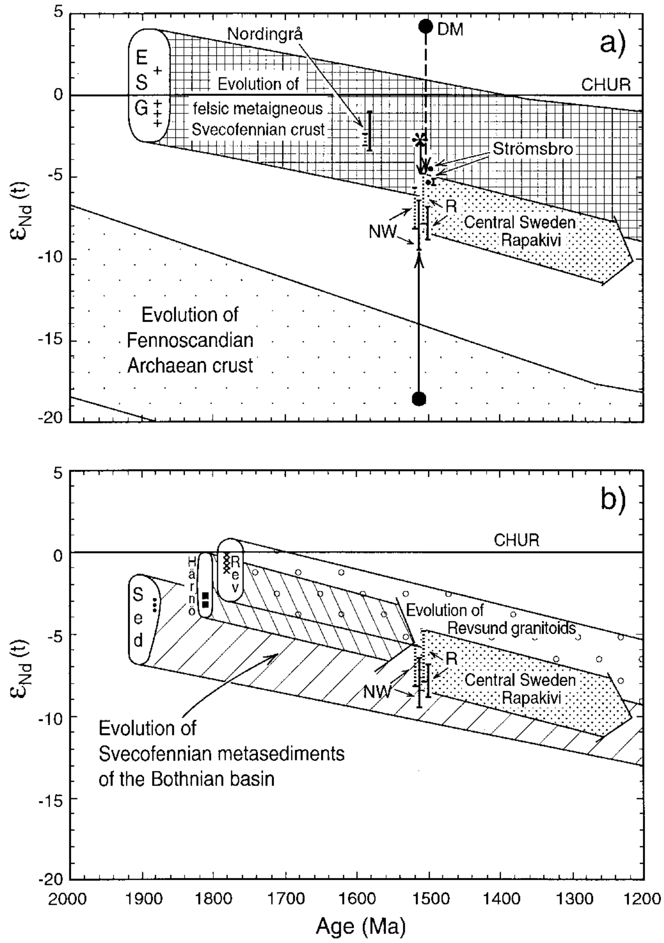
All apparent initial Sr values (except one; M14a, Table 4) of the granite Subjotnian rocks are very to extremely low, which is unreal. They are, for example, considerably lower than the value for an assumed bulk earth (UR = uniform reservoir) at this time (0·7027). The initial ratios for the basic and intermediate rocks are mostly realistic, and may be subdivided in three groups: one fairly primitive (0·7030–0·7053), a second intermediate (0·7060–0·7064), and a third with distinctly higher (0·7110–0·7188) values.

The Svecofennian metasediments and the palingenic granites have initial Sr values between 0·7042 and 0·7092, which at *c.* 1·5 Ga have evolved to 0·7127–0·7195. The Svecofennian orthogneisses were initially (at 1·89 Ga) rather primitive (0·7016–0·7027; excluding one unrealistically low at 0·6809), but have at *c.* 1·5 Ga evolved quite variably (0·7076–0·7195), partially overlapping with the metasediments. The orthogneisses show a correlation between low ε_{Nd} and high $^{87}Sr/^{86}Sr$ values at *c.* 1·5 Ga.

The Rb–Sr WR data of the Rödö rocks are plotted in an isochron diagram in Figure 7b. Only data for silicic rocks were used in the calculation. The calculated 'age' is 1340 ± 39 Ma (MSWD = 154); 160 Ma less than the U–Pb zircon age. The very high MSWD value indicates strong scatter around the line, which cannot be regarded as an isochron. This is more clearly shown in Figure 7c using the diagram of Provost (1990), which is a transformation of the axes of the conventional isochron plot, where the fit of individual analyses to



$\varepsilon_{Nd}^{1.51}$
Rapakivi complexes west
Rapakivi complexes east
Härnö granites
Metasediments, Bothnian Basin
Revsund granitoids
Archaean
Early Svecofennian
-25
-20
-15
-10
-5
0



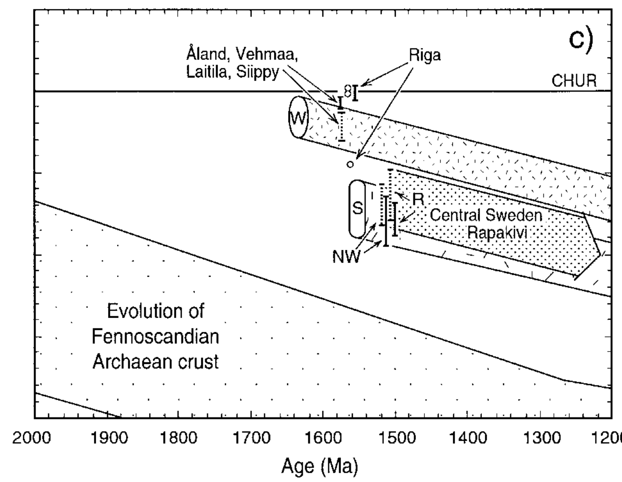


Figure 6 ε_{Nd} evolution vs time for rocks from Fennoscandian rapakivi complexes: ESG, early Svecofennian metaigneous rocks (crosses are those measured in this study); Sed, sedimentary rocks in the Bothnian basin (dots represent those measured in this study); Härnö, Härnö suite granites (*c.* 1·82 Ga) (solid boxes represent palingenic granites from the Rödö area included here); Rev, 1·80–1·77 Ga Revsund granitoid suite (inclined crosses are those samples reported here); W, evolution of the Wiborg batholith granitoids and its satellites (particularly Suomenniemi); S, rocks of the Salmi batholith; central Sweden rapakivi denotes the evolution of silicic–intermediate rocks of this study; solid bars denote basic Subjotnian rocks and broken bars silicic–intermediate; Strömsbro dots are dolerites, and bars granites, whereas unfilled circles represent granitoids from Riga batholith; DM, approximate composition of the depleted mantle at 1·51 Ga (DePaolo 1981); black dot in the Archaean field denotes the average composition of Fennoscandian Archaean rocks, and the star similarly denotes the early Svecofennian metaigneous rocks (cf. Table 6); sources of data as in Figure 5.

the regression line is more clearly seen. Calculations omitting certain samples do not improve the results. If, for example, the two samples with the most elevated Rb/Sr ratios are omitted, this gives an almost identical age result (1346 ± 28 Ma, MSWD = 48).

3.4. Pb isotopes

3.4.1. Feldspar Pb isotopes from the NW complexes. Feldspars from seven rocks of the rapakivi complexes and three country rocks were analysed for lead isotopes (Table 5). The same rocks were also analysed for Sr and Nd WR isotopes (Table 4). $^{206}Pb/^{204}Pb$ values for the leached rapakivi feldspars vary between 15·286 and 15·941, except one value as high as 16·610. The latter feldspar may have included some small, undetected U-rich inclusion, adding some radiogenic Pb. Pb in the other rapakivi feldspars is regarded as essentially unradiogenic, showing the isotopic composition at the time of crystallisation.

The $^{207}Pb/^{204}Pb$ ratios vary between 15·018 and 15·217, plotting between Rämö's (1991) approximated evolution curves for the depleted mantle ($\mu_2 = 9{\cdot}17$) and lower crust ($\mu_2 = 8{\cdot}00$) which are growth curves commencing at stage two (t = 3·7 Ga) according to the two-stage evolutionary model of Stacey & Kramers (1975) (Fig. 8a). A large number of Pb isotope data of feldspar and galena from the Fennoscandian shield added for reference in Figure 8. The range in Pb isotopic composition is as large within the complexes as between them,

Figure 5 Histogram showing distributions of the ε_{Nd} values for different Fennoscandian rapakivi complexes and pertinent rocks calculated at 1·51 Ga. The lettering refers to the following. Early Svecofennian (metaigneous): B is Bergslagen region, C is central Sweden (Bothnian basin) and Sk is the Skellefte district. Metasediments in the Bothnian basin: S is in the Swedish part and F in the Finnish part. Rapakivi complexes: S is Salmi, W is Wiborg batholith and its satellites (particularly Suomenniemi), E is the Estonian complexes, Ri is Riga batholith, Å is Åland and other SW Finnish complexes, N is Nordingrå, R is Rödön. So is Strömsbro, and NW is the northwestern complexes in central Sweden. Filled boxes are basic rocks, unfilled boxes silicic rocks and checkered boxes intermediate rocks. There is a 'bimodal' distribution of Nd isotopic characteristics, where the Salmi and the central Swedish (excluding Nordingrå) complexes have lower initial ε_{Nd} compared to Nordingrå and the Finnish–Estonian equivalents. At 1·51 Ga (age of rapakivi magmatism in central Sweden) the central Swedish complexes have lower ε_{Nd} than the early Svecofennian metaigneous rocks of the region, but are similar to the metasediments of the Bothnian basin, while the Finnish complexes and Nordingrå compare well with the metaigneous Svecofennian rocks. Sources of data, for the Presubjotnian rocks: this paper; Martin *et al.* 1983; Jahn *et al.* 1984; Wilson *et al.* 1985, 1987; Miller *et al.* 1986; Patchett & Kuovo 1986; Öhlander *et al.* 1987; Patchett *et al.* 1987, 1994; Claesson 1987; Huhma 1986, 1987; Valbracht 1991; Welin *et al.* 1993; Claesson & Lundqvist 1995; Billström & Weihed 1996; Kumpulainen *et al.* 1996; Lahtinen & Huhma 1997; Mellqvist 1997), and for the Subjotnian complexes: (this paper; Rämö 1991; Rämö *et al.* 1996; Neymark *et al.* 1994; Lindh & Johansson 1996; Andersson 1997a; Lindh *et al.* 2001).

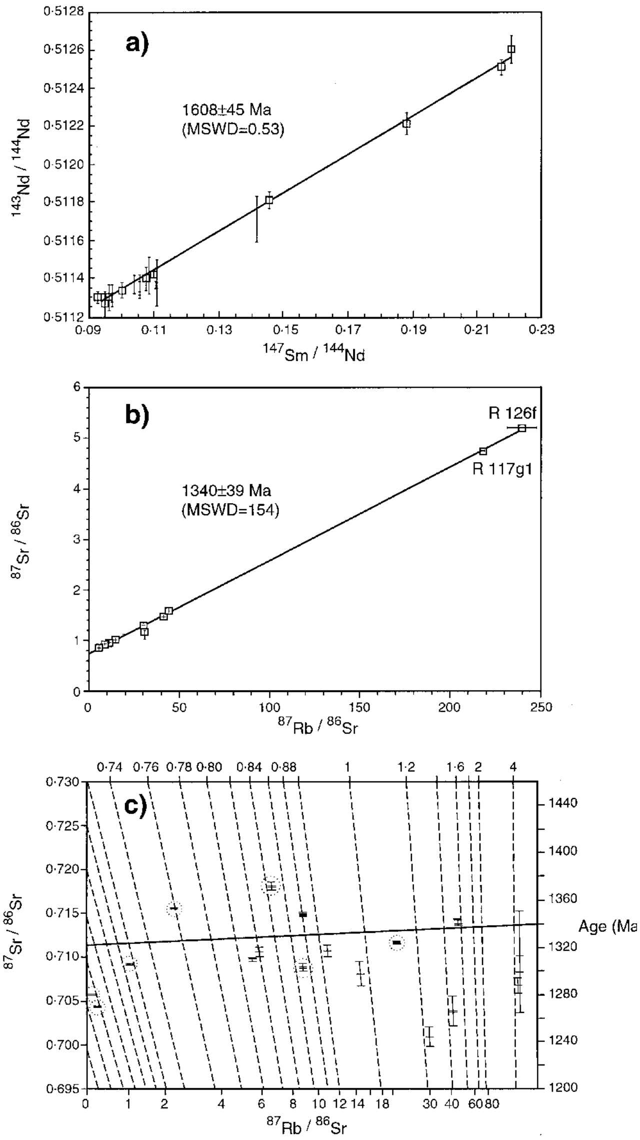


Figure 7 (a) $^{147}Sm/^{144}Nd$–$^{143}Nd/^{144}Nd$ diagram for the rocks of the Rödö complex; silicic rocks are marked with squares; an isochron of acceptable quality was calculated for the silicic rocks (1608 ± 45 Ma); the most evolved rocks have extremely high Sm/Nd ratios and control the isochron to a large extent. (b) $^{87}Rb/^{86}Sr$–$^{87}Sr/^{86}Sr$ diagram for the rocks of the Rödö complex; the calculated errorchron give an 'age' of *c*. 160 Ma younger than the U–Pb zircon age (cf. Fig. 4); two samples, one aplite (R126f) and one QFP (R117g1), have very high Rb/Sr ratios and totally control the slope of the errorchron. (c) 'Improved isochron' diagram of Provost (1990) for the Rödö rocks; in this diagram the axes have been transposed in such a way that the adherence to the calculated line of individual analyses is more clearly displayed irrespective of their Rb/Sr ratio; the left vertical axis gives the initial Sr ratio and the right vertical axis the age; only one of the silicic Rödö rocks overlaps within error the errorchron, explaining the high MSWD value.

and cannot be related to the rock chemistry in any straightforward way.

The $^{208}Pb/^{204}Pb$ values are between 34·971 and 35·878, and spread out in a way similar to the $^{207}Pb/^{204}Pb$ values. The two samples having the least radiogenic Pb isotope ratios (indicating the lowest U/Pb and Th/Pb ratios in their source; Mu12 and M15a) are located slightly above the mantle curve, near the feldspar data from the Salmi complex (Neymark *et al.* 1994). The Salmi feldspars and galenas, however, generally have higher $^{208}Pb/^{204}Pb$. Other data points are distributed near the mantle and upper crustal curves in the model of Zartman & Doe (1981) (Fig. 8b).

The three feldspars from the country rocks show markedly higher Pb isotopic ratios, especially $^{206}Pb/^{204}Pb$, suggesting an addition of uranogenic lead. The $^{207}Pb/^{204}Pb$ values conform relatively well with the evolutionary curve for 'Svecofennian juvenile crust' of Rämö (1991). Most of the analysed feldspars and galenas from Finnish rapakivi complexes (Vaasjoki 1981; Rämö 1991) fall along this trend (Fig. 8a). In contrast, feldspars from the Subjotnian complexes analysed here have substantially more primitive (i.e. less radiogenic) compositions.

3.4.2. Pb whole rock isotopes from the Rödö complex. Four granites, three QFPs and one aplite from the Rödö complex were analysed for WR–Pb (Table 5; Fig. 9). Because of the relatively high WR–U content, the Pb isotopic signatures are strongly radiogenic, particularly samples R126f and R117g1; i.e. the same samples that are also the most evolved in the Nd and Sr systems. The data cannot be fitted together in a single secondary isochron. An errorchron fitted through the six least radiogenic samples, which appear to form a linear array, gives an 'age' of 1762 ± 1100 Ma (Fig. 9).

In comparison with corresponding analyses from granitoids in Finnish rapakivi complexes (Rämö 1991), the Rödö samples generally have lower $^{207}Pb/^{204}Pb$ ratios. Lines with slopes corresponding to the crystallisation age of the Rödö granite (*c*. 1·50 Ga), extrapolate back from the WR-compositions to the composition measured for the more radiogenic of the feldspars of the NW complexes (cf. Fig. 8a), except for sample R117g1.

4. Discussion

4.1. Ages

Welin (1994) has previously determined a U–Pb zircon age of the Rödö granite, from four zircon fractions, to 1513 ± 5 Ma; however, the analytical details were not reported. The scatter obtained in the U–Pb data of the present study significantly exceeds the analytical errors (MSWD values $\gg$1), suggesting involvement of geological factors, such as e.g. remnants of older material. The analysed fractions plotting slightly to the right of the discordia may contain an older component (Fig. 4a). This component is either minor in volume and much older, or larger in volume but only slightly older, since the points are still close to the discordia. The former is more likely in the light of the low ε_{Nd} values for the Rödö granites and associated rocks, suggesting a major Archaean input to the magmas, even though no specific inherited types of zircon have been identified. Even the most concordant fraction (R10) seems to be slightly displaced to the right of the discordia and may also contain small amounts of an older component.

The fraction plotting above the concordia (R4) has most likely experienced apparent U loss due to incomplete digestion of the crystals in the analytical process, where some of the bound U (but not the more loosely bound Pb) was lost. If this is the case, then the R4 point has moved along a line away from the origin in the diagram. Since the calculated discordia line does not run through the origin, but has a lower intercept of 229 Ma, a more recent U loss in fraction R4 should have moved this point somewhat off the discordia to the left. If this fraction, and those that are slightly displaced to the right (R1, R2, R6) are omitted, the statistically best constrained age is 1497 ± 6 Ma (MSWD = 1·3). This age is the preferred age estimate, as it probably reflects only a minimum influence of older components (Fig. 4a).

Within errors, all discordia calculations given the same age around 1500 Ma, which is slightly but probably significantly lower than the 1513 ± 5 Ma age obtained by Welin (1994).

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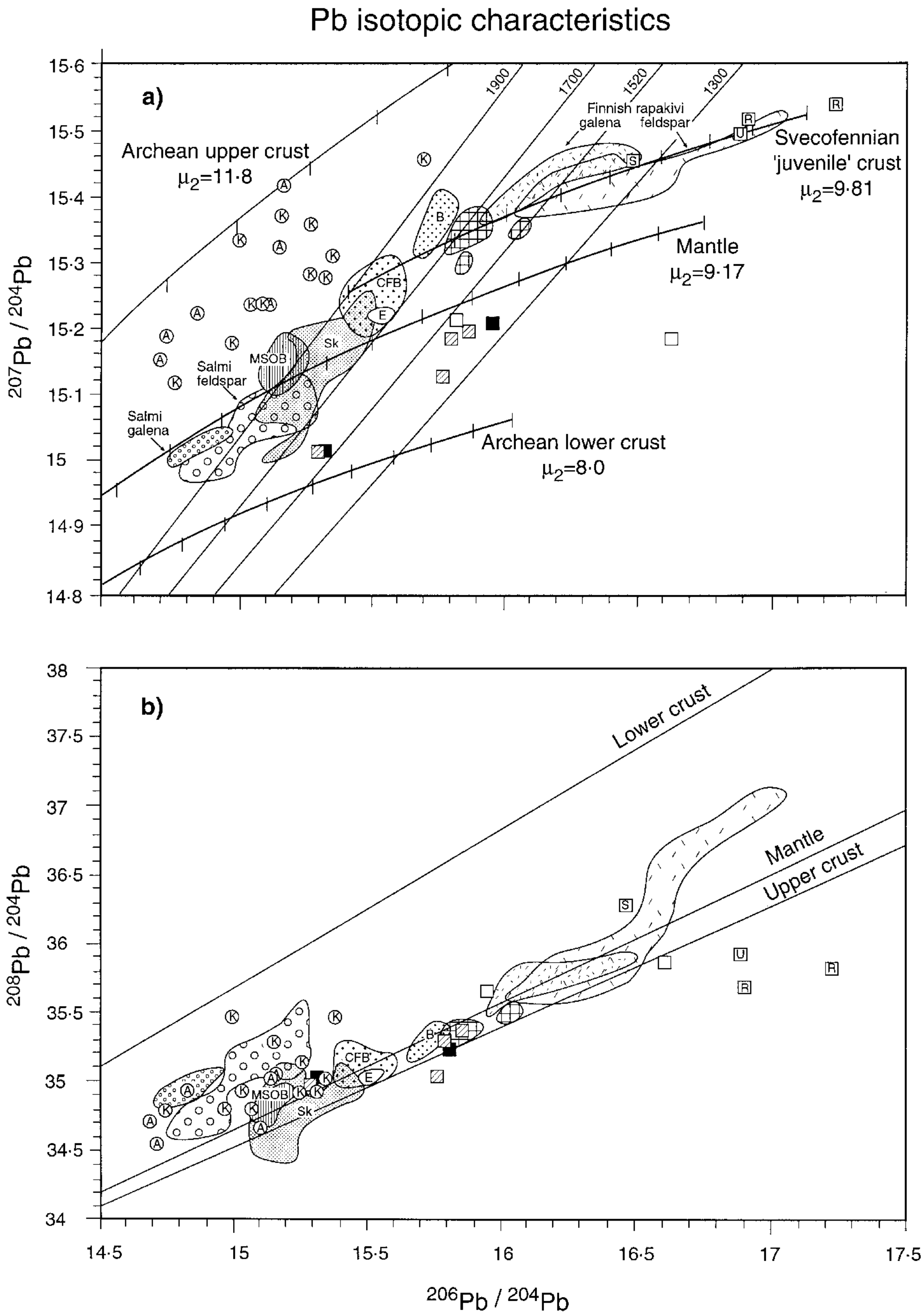

Figure 8 Pb isotopic composition of the feldspar fractions of this study (NW area), in the frame of other Fennoscandian Pb data compiled from the literature. Differently stippled and ruled fields are early Svecofennian galenas from different geographic/geological areas: B, Bergslagen, CFB, central Finnish batholith area, E, Enåsen deposit in the Bothnian basin, SK, Skellefte district, MSOB, main sulphide ore belt in Finland; fields with circles are Salmi galenas and feldspars, and fields with small dashes galenas and feldspars from Finnish rapakivi complexes; checkered fields are galenas from Postsvecofennian granite-related galenas; circles with an included A represent galenas from Archaean rocks and circles with a K are galenas from Karelian cover rocks on the Archaean craton; samples of this study (feldspars, Table 5), are plotted as squares: unfilled = granites, filled = basic rocks, striped = syenites from the Subjotnian complexes, filled with an S = from an early Svecofennian metasediment, filled with a U = from an early Svecofennian granitoid, and those filled with R = from a Revsund granitoid. (a) $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram; evolutionary curves are stage two curves after Stacey & Kramers (1975) for different postulated reservoirs with different μ_2 ($^{238}U/^{204}Pb$) values: Archaean upper crust, $\mu_2 = 11{\cdot}8$, Archaean lower crust, $\mu_2 = 8{\cdot}0$, depleted mantle, $\mu_2 = 9{\cdot}17$, and Svecofennian 'juvenile' crust, $\mu_2 = 9{\cdot}18$ (Rämö 1991); stage two model isochrons (Stacey & Kramers 1975) for 1300, 1520, 1700, and 1900 Ma have also been included. (b) $^{208}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$ diagram; evolutionary curves are after Zartman & Doe (1981); data sources for the compiled data are: Vaasjoki (1981), Rickard & Svensson (1984), Billström (1985), Johannson & Rickard (1985), Åberg & Charalampides (1986), Vaasjoki & Sakko (1988), Hallberg (1989), Vaasjoki & Vivallo (1990), Rämö (1991), Sundblad (1991), Sundblad *et al.* (1993), Billström & Vivallo (1994), Neymark *et al.* (1994) and Billström *et al.* (1997).

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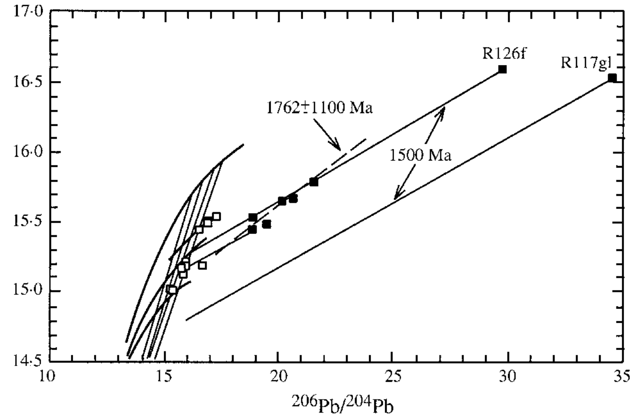

Figure 9 $^{206}Pb/^{204}Pb$–$^{207}Pb/^{204}Pb$ diagram for the silicic WR-samples (filled squares) from Rödön (Table 5); the diagram also includes the feldspar data from the NW complexes (unfilled squares; cf. Table 5), as well as the second-stage growth curves and isochrons plotted in Figure 8; a secondary errorchron fitted through the six least radiogenic WR-samples gives an 'age' of 1·76 Ga, which has no geological meaning and probably is a function of the variation in initial compositions and μ-values; lines with slopes representing the crystallisation age of the Rödö complex (1·50 Ga), have been indicated; all except one of the Rödö WR-samples extrapolate back (at 1·5 Ga) approximately to the isotopic composition (initial) of the NW feldspars.

Based on these results, and the low initial ε_{Nd}, one may argue that the result of Welin (1994), calculated from only four fractions, may also contain unidentified older components, giving a slightly too old intrusion age. Thus, the age preferred from the present data may more accurately define the intrusion age.

The Sm–Nd WR isochron age of 1608 ± 45 Ma (MSWD = 0·53; Fig. 7a) of the silicic rocks is the Rödö suite, is about 110 Ma older than the U–Pb age. There is a large spread in the $^{147}Sm/^{144}Nd$ ratios, but only a few samples have very high ratios exerting a strong impact on the isochron calculation. Small biases in their determined values will have large effects on the determined age. On the other hand, all but one analytical point fall within error on the isochron (Fig. 7a), which makes it a relatively well-determined WR isochron.

Any difference in the initial $^{143}Nd/^{144}Nd$ ratios in the system would make the isochron unreliable. If one accepts the 1497 Ma U–Pb age as the true intrusion age, the calculated initial ε_{Nd} values show a significant spread, where the samples with the highest Sm/Nd ratios have the highest ε_{Nd} values (Table 3). This means that it may be a case of an isochron with a primary positive slope, resulting in an overestimation of the real age. The reason for this may be sought in the strong fluid control exerted on the magmatic evolution, and especially on the LREEs (Andersson 1997c, 2001). The fluid extraction of the LREEs from the magmas has caused fractionation between Sm and Nd, leading to a strong increase in the Sm/Nd ratio in the most evolved silicic magmas. The strong partitioning of LREEs into the F- and CO_2-rich fluid phase may also have caused fractionation of the initial $^{143}Nd/^{144}Nd$ ratios between variably evolved magmas, which now biases the isochron age.

The very low and unreal calculated initial Sr ratios of the silicic rocks and the strong scatter around the errorchron (Table 3, Fig. 7c) indicate that the primary magmatic Rb–Sr system has suffered disturbance some unconstrained time after crystallisation, such that the original Rb/Sr ratio has increased or radiogenic Sr has been lost. Possible later events include (i) *c.* 1·27 Ga Postjotnian sill intrusions (e.g. one big sill can be inferred from aeromagnetic data to cross-cut the Rödö intrusion; Geological Survey of Sweden 1997, unpublished map), and (ii) the *c.* 0·58 Ga Alnö alkaline intrusion which has affected parts of the intrusion (including carbonatite dykes). Palaeomagnetic data seem to support a disturbance in Postjotnian times (Moakhar & Elming 2000).

4.2. Age distribution

The spatial distribution of intrusive ages among the Fennoscandian Subjotnian rapakivi complexes has been poorly constrained until recently, when new geochronological data have become available. The well-constrained ages from the three NW complexes reported here fall in the narrow range 1529–1517 Ma including errors, *c.* 60 Ma younger than the U–Pb zircon age of the Nordingrå complex (1578 ± 19 Ma; Welin & Lundqvist 1984). Recent reported U–Pb zircon ages of the Subjotnian Ragunda and Strömsbro complexes give, together with Rödön, slightly younger ages in the range 1513–1497 Ma (Andersson 1997a; Persson 1999). Claesson & Kresten (1997) also included the small 1·47 Ga anorogenic Noran intrusion, in southern central Sweden, among the rapakivi complexes of the shield.

The Swedish rapakivi complexes, which occupy the westernmost part of the Fennoscandian suite of rapakivi granite complexes, are thus the youngest (cf. Fig. 1), except Nordingrå which is similar in age to the SW Finnish and the Riga complexes (1·59–1·56 Ga; Vaasjoki 1997; Vaasjoki *et al.* 1988; Idman 1980; Suominen 1991; Rämö *et al.* 1996). The otherwise youngest complex in the suite is the easternmost Salmi massif in Russian Karelia (1·56–1·53 Ga; Suominen 1991; Neymark *et al.* 1994; Amelin *et al.* 1997). The oldest rapakivi complexes in Fennoscandia are the Wiborg complex together with associated smaller plutons in SE Finland (1·65–1·61 Ga, from 1·67 Ga including mafic dykes; Suominen 1991; Vaasjoki *et al.* 1991; Alviola *et al.* 1999), as well as small intrusions in northern Estonia (Kirs & Petersell 1994; Rämö *et al.* 1996) and on the island of Hogland in the Gulf of Finland (Belyaev *et al.* 1998).

Excluding the Salmi batholith, there is thus a westerly younging age trend among the Fennoscandian rapakivi complexes. This trend, however, does not seem to record a smooth geographical age distribution. Instead, rocks of similar ages appear to form discrete, roughly N–S trending belts (cf. Fig. 1). The youngest belt (1·53–1·47 Ga) trends from Noran and Strömsbro in the S to Mullnäset and Strömsund (no age determined) in the N, followed to the E of an older belt (1·59–1·56 Ga) from Riga in the S over SW Finland to Nordingrå in the N. Next, to the E is the oldest belt (1·67–1·61 Ga) trending from northern Estonia to SE Finland (and SW Russian Karelia) with the huge Wiborg batholith as its major expression, and in the eastern end the Salmi complex, *c.* 100 Ma younger than the Wiborg complex (*c.* 1·56–1·53 Ga).

The reason for this age distribution is unclear, but the belt-like geometry argues for a relation to large-scale tectonic processes. Åhäll *et al.* (2000) have, for example, proposed that eastward subduction and accretion of successive belts of calc-alkaline rocks onto a penecontemporaneous evolving continental margin of Baltica in the W, was instrumental for the initiation of within-craton extension and successive rapakivi magmatism. Although possible, the position of these subduction systems in relation to Baltica, and their time of docking with the craton is still uncertain and controversial (Andersson 2001, and references therein). Previous models have emphasised crustal extension, thinning and anorogenic magmatism along the Gulf of Finland, where the crust is thinner than the surroundings, possibly in response to extended extensional collapse of overthickened Svecofennian crust (e.g. Korja *et al.* 1993; Windley 1993; Korja & Heikkinen 1995; Rämö & Haapala 1995, and references therein). The large Riga complex does not follow this pattern, as it is intruded into thicker crust.

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However, recent interpretations of geophysical data suggest that the Rödö and Nordingiå complexes are small on-land expressions of much larger batholiths underlying the northern Baltic Sea palaeorift, which extends in a N–S direction (Korja *et al.* 2001), conformable with the apparent age belts. Puura & Flodén (1999, 2000) preferred to group the Fennoscandian rapakivi complexes into rounded subprovinces of discrete ages, though this subdivision is not consistent with the reported ages in the W (cf. Figs 1, 2). They also included the *c.* 1·55–1·50 Ga (Rb–Sr ages; Patchett 1978) Breven–Hällefors dyke province in eastern southern Sweden, which lacks associated felsic plutonics, as a certain rapakivi subprovince. While lacking a definite plate tectonic model, it is clear that Fennoscandian rapakivi magmatism was activated in a stepwise fashion westwards: 1·67–1·61 Ga → 1·59–1·56 Ga → 1·53–1·47 Ga (with an intervening pulse, 1·56–1·53 Ga, in the E), including associated dyke swarms at right angles to each other (Fig. 1), in response to extensional stress fields changing with time.

4.3. Nd isotopic relations

4.3.1. Origin of the Subjotnian granitoids of central Sweden. The very low calculated initial ε_{Nd} values for the Subjotnian rocks suggest the involvement of an older unradiogenic component in the magmas. This is true irrespective of the chemistry of the rocks. In fact, the basic-intermediate rocks tend to have even slightly lower initial ratios (Tables 3 and 4), e.g. in the Rödö complex: −8·9 to −6·9 (av. −7·6), as compared to the silicic rocks: −7·4 to −4·8 (av. −6·6).

In order to try to put some quantitative constraints on possible sources for the Subjotnian rapakivi rocks, we have compiled published Nd isotopic data for rocks in the Svecofennian and Archaean provinces of the Fennoscandian shield. For comparative purposes this compilation is summarised in Figure 5, as a set of histograms of the ε_{Nd} (1·51) of all rocks. The age of 1·51 Ga was chosen as it represents an average age of the Swedish rapakivi complexes (excluding Nordingrå). In comparison with rocks from SE and SW Finnish and Estonian rapakivi complexes (Rämö 1991; Rämö *et al.* 1996) the central Swedish rocks have markedly lower ε_{Nd} values (Fig. 5). The rocks of the Salmi complex in Russian Karelia (Rämö 1991; Neymark *et al.* 1994), on the other hand, show complete overlap with the Swedish ones, while the Strömsbro rocks (Andersson 1997a) plot in the upper part of the range for the other Swedish complexes. The Salmi complex is emplaced on the border between the exposed Archaean craton and the Svecofennian domain, and it slow initial ε_{Nd} values were explained by a mixture of those components (Rämö 1991; Neymark *et al.* 1994).

The temporal Nd isotope evolution of different Fennoscandian crustal lithologies is shown in Figure 6. The Finnish–Estonian and Nordingrå rapakivi granites are clearly included within the evolution of the early Svecofennian metaigneous rocks, and can thus, with regard to the Nd isotopes, be interpreted as deriving solely from crustal melting of such sources (Rämö 1991; Lindh & Johansson 1996). This explanation is not feasible for the rocks of the Rödö and NW complexes. These occur more than 300 km S and SW of the nearest exposed Archaean crust in the Luleå area (Lundqvist *et al.* 1996; Wilkström *et al.* 1996), with a juvenile arc-like terrane (the Shellefte district) in between, but still show low ε_{Nd} values similar to those of the Salmi complex.

The calculated depleted mantle ages, T_{DM} (DePaolo 1981), and chondritic ages, T_{CHUR}, listed in Tables 3 and 4 are in the range 2·5–2·0 Ga for the rapakivi granitoids. These mantle model ages imply that possible juvenile mantle-derived protoliths of the rapakivi granitoid magmas should have been extracted from the mantle in this time span. Possible material with ages between 2·5 and 2·0 Ga, suitable as protoliths for the rapakivi magmatism, have not been recorded in this region. In the shield as a whole, rocks formed in this time span are dominated by metasediments and mafic to ultramafic volcanics and intrusions, belonging to the Lapponian and Karelian supergroups, covering the eroded Archaean basement in the NE (e.g. Gaál & Gorbatschev 1987; Huhma *et al.* 1990), with only minor intercalations of felsic igeneous rocks. One single U–Pb zircon dating of early Svecofennian rocks in the region gives an age older than 2·0 Ga (2·03 Ga; Welin *et al.* 1993), otherwise the early Svecofennian ages are constrained to the period 1·95–1·86 Ga. Ion-probe U–Pb work on early Svecofennian metasediments has identified essentially no zircons in the age range 2·5–2·1 Ga (Claesson *et al.* 1993). Most of the zircons are in the range 2·1–1·9 Ga and a limited number (30%) are Archaean (>2·6 Ga). These data do not support the existence of a major crustal segment of age 2·5–2·1 Ga in the Svecofennian part of the Fennoscandian shield, although the existence of such rocks at depth cannot be ruled out.

The alternative is mixing of two or more components, with different isotopic characteristics. As can be seen in Figures 5 and 6, only an Archaean crustal component has distinctly lower ε_{Nd} isotopic composition than that of the rapakivi granitoids. Thus, this is the only known potential source capable of lowering the ε_{Nd} of the rapakivi magmas significantly. A Svecofennian metaigneous derivation needs to be complemented by an input of Archaean material. Table 6 lists the average concentration of Nd (C_{Nd}) and ε_{Nd} (1·51) for different potential source rocks, compiled from literature data supplemented by the data of this paper.

Table 6 Concentration of Nd (*ppm*) and ε_{Nd} at the time of rapakivi magmatism for potential source rocks.

	C_{Nd}	$\varepsilon_{Nd}(1{\cdot}51)$
Depleted mantle[1]	5–15	+4·5
Non-depleted mantle[2]	30–47	±0
Archean[3]	24	−18·5
Metasediments, Bothnian basin[4]	30	−7·1
Svecofennian, metaigneous		
Bothnian basin area only (Swedish parts)[5]	36	−3·9
Skellefte field and southwards (incl. s. Finland)[6]	32	−2·7
Revsund granitoids[7]	41	−4·1
Rapakivi granitoids and syenitoids[8]	60	−6·8
Rapakivi-associated basic rocks[9]	34	−7·7

Notes:
1. Expected range in C_{Nd} estimated from data of Huhma (1986), Patchett & Kuovo (1986), DePaolo (1988); Björklund & Claesson (1992) and Kumpulainen *et al.* (1996). $\varepsilon_{Nd}(1{\cdot}51)$ after DePaolo (1981); $\varepsilon = 0{\cdot}25T^2 - 3T + 8{\cdot}5$.
2. C_{Nd} based on the Suomenniemi and Häme swarms (Rämö 1991), being a possible non-depleted source. $C_{Nd} = 30$, assumed initial diabase magma of Rämö (1991). $C_{Nd} = 47$, average of the Häme and Suomenniemi diabase analyses.
3. Average of analyses from Martin *et al.* (1983); Jahn *et al.* (1984); Huhma (1986); Öhlander *et al.* (1987) and Mellqvist (1997). C_{Nd}, n = 16. ε_{Nd}, n = 26.
4. Averages of 19 analyses from this paper, Miller *et al.* (1986), Claesson (1987), Patchett *et al.* (1987), Welin *et al.* (1993) and Claesson & Lundqvist (1995).
5. Averages of 19 analyses from this paper, Welin *et al.* (1993) and Claesson & Lundqvist (1995).
6. Averages of 113 analyses from those in (5) and Wilson *et al.* (1985, 1987), Huhma (1986), Patchett & Kouvo (1986), Patchett *et al.* (1987), Valbracht (1991), Billström & Weihed (1996), Kumpulainen *et al.* (1996), Lahtinen & Huhma (1997) and Andersson (1997a,c).
7. Averages of 19 analyses from this paper, Wilson *et al.* (1985), Patchett *et al.* (1987) and Claesson & Lundqvist (1995).
8. This paper. n = 22 (Tables 3 and 4).
9. This paper. n = 8 (Tables 3 and 4).

The Svecofennian metasediments of the Bothnian basin show a Nd evolution completely overlapping that of the silicic Rödö rocks, with a spread in $\varepsilon_{Nd}(1{\cdot}51)$ between $-10{\cdot}2$ and $-5{\cdot}8$, and an almost identical average ($-7{\cdot}1$; Table 6) (Miller *et al.* 1986; Claesson 1987; Patchett *et al.* 1987; Welin *et al.* 1993; Claesson & Lundqvist 1995; this paper). Data of Huhma (1987) for the Tampere schist belt in southern Finland give slightly less negative ($-5{\cdot}3$ to $-4{\cdot}4$) values and have not been included. Metaigneous rocks of the early Svecofennian episode are generally more radiogenic, with only a little overlap with the rapakivis (Fig. 6). The spread in $\varepsilon_{Nd}(1{\cdot}51)$ is wide ($-6{\cdot}0$ to $+1{\cdot}1$), but the average ($-2{\cdot}7$) is almost four ε units higher than for the rapakivi granitoids. The Svecofennian data represent rocks occurring over a large area, from the Skellefte district in the N (Wilson *et al.* 1985, 1987; Billström & Weihed 1996) to the relatively juvenile Bergslagen district in the S (Patchett *et al.* 1987; Valbracht 1991; Kumpulainen *et al.* (1996), including recent data from central Sweden (Welin *et al.* 1993; Claesson & Lundqvist 1995; this paper) as well as data from southern Finland (Huhma 1986; Patchett & Kuovo 1986; Lahtinen & Huhma 1997). The average $\varepsilon_{Nd}(1{\cdot}51)$ value of the Revsund suite granitoids is somewhat lower compared with the early Svecofennian metaigneous rocks ($-4{\cdot}1$), and the range is more restricted ($-5{\cdot}9$ to $-2{\cdot}4$; Wilson *et al.* 1985; Patchett *et al.* 1987; Claesson & Lundqvist 1995; this paper). However, the Revsund rocks also generally display a more radiogenic Nd evolution compared with the studied rapakivis (Fig. 6b). The Revsund suite rocks are restricted geographically to central Sweden and northwards to the Skellefte district. The $\varepsilon_{Nd}(1{\cdot}51)$ values in Archaean rocks of the Fennoscandian shield spread between $-25{\cdot}2$ and $-13{\cdot}8$ (Martin *et al.* 1983; Jahn *et al.* 1984; Huhma 1986; Öhlander *et al.* 1987; Mellqvist 1997), and have an average $\varepsilon_{Nd}(1{\cdot}51)$ of $-18{\cdot}5$.

A likely range in Nd concentration for a possible depleted mantle component in the Svecofennian is 5–15 ppm (cf. e.g. Huhma 1986; Patchett & Kuovo 1986; DePaolo 1988; Björklund & Claesson 1992; Kumpulainen *et al.* 1996), and the $\varepsilon_{Nd}(1{\cdot}51)$ is $+4{\cdot}5$ (DePaolo 1981). A possible undepleted mantle source of this particular age could be similar to the source of the Suomenniemi and Häme dolerite dyke swarms, which are related to the SE Finnish rapakivi complexes and have initial ε_{Nd} values close to zero (Rämö 1991). The average Nd concentration of these dolerites is 47 ppm. However, Rämö (1991) assumed in his calculations a C_{Nd} of 30 ppm in an initial mafic magia in these dykes, based on the most depleted rocks with the highest initial ε_{Nd}. Thus, a reasonable range in Nd concentration for mafic magmas derived from an undepleted mantle source associated with the rapakivi magmatism may be 30–47 ppm. The Nd concentration in Archaean rocks from the NE parts of the Fennoscandian shield is quite variable, 7–69 ppm (Jahn *et al.* 1984; Huhma 1986; Öhlander *et al.* 1987; Mellqvist 1987), with an average of 24 ppm. C_{Nd} for Svecofennian metasediments of the Bothnian basin varies between 12 and 44 ppm with an average of 30 ppm (Miller *et al.* 1986; Patchett *et al.* 1987; Claesson 1987; Welin *et al.* 1993; Claesson & Lundqvist 1995; this paper). Early Svecofennian metaigneous rocks are also quite variable in C_{Nd} (4–102 ppm), but a majority are in the range 20–40 ppm with an average of 32 ppm (Wilson *et al.* 1985, 1987; Huhma 1986; Patchett & Kuovo 1986; Patchett *et al.* 1987; Valbracht 1991; Welin *et al.* 1993; Claesson & Lundqvist 1995; Billström & Weihed 1996; Kukmpulainen *et al.* 1996; Lahtinen & Huhma 1997; this paper). Revsund granitoids range in C_{Nd} between 13 and 68 ppm, but 64% are between 30 and 50 ppm with an average of 41 ppm (Wilson *et al.* 1985; Patchett *et al.* 1987; Claesson & Lundqvist 1995; this paper).

As far as the Nd isotopic composition is concerned, the Subjotnian granitoids could be entirely derived from melting of metasedimentary rocks (Fig. 6). Metasediments, however, are a very unlikely source for A-type granitoids. The sediments are in part muscovite-bearing, Al-rich rocks (with Al_2O_3 up to 21%; Lundqvist *et al.* 1990), and melts from such a source would crystallise to peraluminous, two-mica S-type granites (e.g. Chappell & White 1974), like the Härnö granites, occurring exclusively within the sedimentary basin question (Fig. 2) (Lundqvist *et al.* 1990; Claesson & Lundqvist 1995). Moreover, the feldspar Pb and whole-rock Sr isotope data preclude metasedimentary sources (see below). Another possibility would be an early Svecofennian metasedimentary source depleted by a previous extraction of granitic melts, e.g. the Härnö granitic suite. Such a depleted crustal source is, however, not likely to contain an appropriate mineralogical composition. A previous, presumably H_2O-undersaturated, melt extraction event from semipeletic (greywacke) sources would proceed by the dehydration melting of micas, depleting the protolith in, for example, K and LIL elements, leaving a residue dominated by quartz, plagioclase, garnet, orthopyroxene, cordierite and sillimanite (e.g. Vielzeuf & Holloway 1988; Clemens 1992; Patiño Douce 1996; Thompson 1996 and references therein). Such a depleted source composition will not be capable of producing A-type granitic magmas rich in K, LIL and other characteristic elements (e.g. Creaser *et al.* 1991; Emslie 1991).

Hence, mixing of at least two components appears to be required to explain the origin of the rapakivi magmas. Mantle sources are possible, but require substantial crustal additions, unless the Subsvecofennian mantle at the time was highly enriched in LREE. There are data supporting mantle sources for silicic rocks, both for Tertiary to Recent rhyolites (DePaolo *et al.* 1992; Perry *et al.* 1993), and Proterozoic A-type granitoids (Emslie & Hegner 1993; Kerr & Fryer 1993). In these cases, however, the rocks exhibit positive ε_{Nd} (t) values, suggesting depleted mantle sources or sources with short crustal residence times, which is not applicable in the case of Swedish rapakivis. Furthermore, the ε_{Nd} (t) of the Subjotnian granitoids is lower than in any known mantle source from the shield.

Substituting relevant data into the general two-component mixing equation (cf. Faure 1986), the mixing of Archaean material with a mantle or crustal component can be described as:

$$x = \frac{C_{Nd}(S)[\varepsilon_{Nd}1{\cdot}51(S) - \varepsilon_{Nd}1{\cdot}51(Rap)]}{C_{Nd}(S)[\varepsilon_{Nd}1{\cdot}51(S) - \varepsilon_{Nd}1{\cdot}51(Rap)] + C_{Nd}(Arch)[\varepsilon_{Nd}1{\cdot}51(Rap) - \varepsilon_{Nd}1{\cdot}51(Arch)]}$$

where (Rap) denotes the rapakivi magma, (Arch) the Archaean component, (S) source material other than the Archaean component, and x is the fraction of Archaean material in the mixture.

In order to obtain some approximate quantitative constraints on the mixing of possible source components, averages in Table 6 were used in the calculations. A derivation of the rapakivi granitoids from a depleted mantle source requires the input of 17–38% of Archaean material, depending on the mantle C_{Nd} (5–15 ppm). An undepleted mantle source requires even more Archaean input (42–53%). To obtain the rapakivi granitic magmas from such mantle sources, an extreme fractionation (>90%; cf. e.g. Hess 1989, p. 231) is required, in addition to the large degrees of assimilation implied from the calculations. If assimilation in the lower crust is the major process, this lower crust has to be made up almost exclusively of Archaean material, otherwise an even larger degree of assimilation is necessary. Such large degrees of assimilation cause substantial energy loss and promote crystallisation of the magma (e.g. McBirney 1979; Hess 1989). The magma would tend to crystallise before reaching upper crustal levels. No field evidence for massive assimilation has been observed,

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such as, for example, restite enclaves. On the other hand, an enriched mantle source will require very little crustal assimilation, and may thus be more realistic. Archaean material might have been added to the subcontinental mantle at some previous orogenic stage (e.g. during the Svecofennian), for example by percolating slab-derived fluids and/or melts (cf. e.g. Zindler & Hart 1986).

Crustal sources are, however, more petrologically tenable. Most workers suggest non-depleted metaigneous lower crustal protoliths for A-type granitoids (Anderson 1983; Creaser *et al.* 1991; Emslie 1991; Skjerlie & Johnston 1993). In this case, a magma derived from an early Svecofennian metaigneous source (Table 6), requires an Archaean contribution of about 32% to arrive at $\varepsilon_{Nd}(1{\cdot}51) = -6{\cdot}8$. The Archaean contribution decreases to 27% if we consider sources similar to early Svecofennian rocks from the Bothnian basin area of central Sweden ($C_{Nd} = 36$, $\varepsilon_{Nd} = -3{\cdot}9$; Table 6). For a Revsund source, 28% of Archaean contribution is necessary. These calculations are strongly dependent on the Nd concentration in the averages. The reported Nd content in the Archaean rocks is very variable (Jahn *et al.* 1984; Huhma 1986; Öhlander *et al.* 1987; Mellqvist 1997), with only a few analyses showing more than 50 ppm and most below 20 ppm. An average of 14 ppm, calculated omitting the few data with high contents, gives an Archaean contribution to the rapakivi magmas of 44%. The spatial distribution of Revsund suite rocks (Figs 1, 2) suggests that this potential source is available primarily for the western intrusions and not for the Rödö granite. In the western area a combination of early Svecofennian and Revsund protoliths is permissible.

A combination of three sources (Archaean, early Svecofennian and mantle) is also possible. The association of the rapakivi granites with intermediate and mafic rocks points to this option. In such a case it is impossible to separate the relative importance of mantle and Svecofennian sources. It is evident, however, that a major Archaean component is present in all the magmas of the Subjotnian complexes of central Sweden.

In summary, we suggest that the source of the central Swedish rapakivi granitoid magmatism is a mixture of early Svecofennian metaigneous (possibly including also Revsund) and Archaean material, where the Archaean proportion is *c.* 30–45%. This Archaean material is presumably of tonalitic, trondhjemitic or granodioritic (TTG) composition, similar to the majority of both the early Svecofennian granitoids, and the Archaean gneisses in the NE parts of the shield (cf. e.g. Gaál & Gorbatschev 1987). The existence of Archaean material in the lower crust of central Sweden was verified by Claesson *et al.* (1997), who found 2·7 Ga zircons in the Nordsjö syenite.

4.3.2. Origin of the Subjotnian basic rocks in central Sweden. The similarly low ε_{Nd} (t) values of the associated basic rocks cannot be explained by a model involving crustal sources only. The basic magmas should have their ultimate origin in the mantle. Most likely the heat supplied by these basic magmas as they were emplaced in the lower crust started the melting and generation of the rapakivi magmas (Huppert & Sparks 1988; Andersson 1991; Haapala & Rämö 1992; Korja *et al.* 1993). This is supported by the bimodal nature of almost all Fennoscandian rapakivi complexes. However, the Nd isotopic nature of such mantle-derived basic magmas would be expected to be quite different from a crustal magma, especially when the crustal magma has such low initial ε_{Nd}. Two possible explanations of the low initial ε_{Nd} of the basic to intermediate rocks are: (i) these magmas derive from a strongly enriched mantle source, or (ii) they were originally chondritic or depleted in character but assimilated sufficient lower crustal material to change completely their Nd isotope characteristics.

No basic rocks with such strongly negative ε_{Nd} (t) have previously been reported for the part of the Svecofennian domain that lies outside areas known to be underlain by Archaean basement, e.g. in the early Svecofennian, 1·95–1·86 Ga (e.g. Huhma 1986; Patchett *et al.* 1987; Valbracht 1991; Björklund & Claesson 1992; Billström & Weihed 1996; Kumpulainen *et al.* 1996), the Transscandinavian igneous belt (TIB) and related rocks, 1·850–1·65 Ga (Mansfeld 1995; Andersson 1997c), or the Postjotnian, 1·26 Ga (Claesson 1987; Rämö 1990; Patchett *et al.* 1994; Andersson 1997a). Thus, the reason why the mantle should be enriched at 1·52–1·50 Ga and not before or after this time interval remains obscure. Moreover, the Svecofennian province in central Sweden was stabilised at *c.* 1·77 Ga after the regional metamorphism and magmatism culminated with the Revsund granitoid suite. Further orogenic activities that could enrich the mantle with crustal components should have been minimal after this time. If the mantle was enriched at the TIB stage, this should also be evident in the mafic rocks of this suite, which is not the case in the southern Svecofennian province. The basic TIB rocks show in general a mildly depleted character (Andersson 1997c). No basic rocks associated with the Revsund granitoids have been analysed to date.

Nevertheless, LREE-enriched mantle sources cannot be ruled out. It is possible that pieces of an Archaean subcontinental mantle are preserved beneath sections of a lower crustal Archaean basement, and that this old uppermost mantle was enriched in pre-Subjotnian times (for example, due to Svecofennian subduction episodes). Such strong enrichment then appears to be restricted to mantle sections localised only beneath central Sweden, and not in the rest of the shield (except Salmi; Neymark *et al.* 1994). Basic magmas associated with the rapakivi complexes in Finland and Estonia have near chondritic ε_{Nd}(t) values (Rämö 1991; Rämö *et al.* 1996). Furthermore, the Postjotnian dolerites occurring in the same general areas in central Sweden as the rapakivi complexes derive from relatively depleted sources (Claesson 1987; Patchett *et al.* 1994; Andersson 1997a), implying that they would need to have been extracted from completely different (deeper?) mantle sections compared with the rapakivi-related basic rocks. Such depleted mantle sources could be envisaged also for the Subjotnian basic rocks, but then supplemented with relatively higher degrees of crustal assimilation (Rämö 1990, see below).

However, it seems more probable that the low ε_{Nd}-initials of the basic rocks are predominantly caused by crustal assimilation. This requires a very high proportion of an Archaean component, compared with an early Svecofennian component, in the assimilated material. Otherwise the original basic character of the magmas would be changed too much to be consistent with the observed chemical compositions. A basic magma derived from a depleted mantle source (ε_{Nd} $(1{\cdot}51) = +4{\cdot}5$) having an assumed Nd content of 5–15 ppm, requires an Archaean addition of between 20% and 50%, depending on the Nd content of the Archaen source. If Svecofennian material is also present in the assimilant, the proportion of crustal component would be even higher. Another possible magma source may be an undepleted mantle with $\varepsilon_{Nd}(1{\cdot}51) \approx 0$, giving a primary LREE-enriched mantle-derived magma with Nd content of 30–50 ppm. Such a basic component was suggested by Rämö (1991) based on the most primitive dolerites in the Häme and Suomenniemi swarms in SE Finland, which are associated with SE Finnish rapakivi complexes. Despite its lower ε_{Nd} (t) values, this sort of mantle source requires even larger amounts of Archaean input (>50%) to the mantle magma to reach the low ε_{Nd} of the Subjotnian basic rocks, because of its much higher Nd content. Bulk assimilation of 20–50% or more of crustal rocks of tonalite–granodiorite composition is probably not reasonable, as it would strongly change the original basic composition of the magmas, which is not observed in the rocks (cf. Rämö 1990; Neymark *et al.* 1994; this paper).

A typical feature of the central Swedish complexes is the strongly overlapping initial ε_{Nd} values of the associated basic rocks with the rapakivi granitoids. This is also the case for complexes with less negative initial ε_{Nd} values, Nordingrå (Fröjdö *et al.* 1996; Lindh & Johansson 1996; Lindh *et al.* 2001) and Strömsbro (Andersson 1997a). In fact, in the central Swedish complexes with the lowest initial ε_{Nd}, the basic rocks trend to even lower values than the granitoids (Tables 3, 4). In the Ragunda complex a gabbro with ε_{Nd} (1·51) as low as −10·1 has been recorded (Persson 1997). The overlapping initial ε_{Nd} values of the basic and silicic rocks in the central Swedish complexes could be taken as an indication of a common origin for all magmas in an enriched upper mantle. However, from the discussion above it is suggested that the granitoid and basic rocks originate from different sources.

As regards rhyolitic volcanics and granitoids associated with basic rocks elsewhere, common sources in the mantle have been invoked (e.g. Kerr & Fryer 1993; Perry *et al.* 1993). In these cases, however, all the rock types (including the silicic ones) have depleted Nd isotopic signatures. These authors suggest that all rocks belong to a common line of differentiation from an original basic mantle-derived magma. A differentiation origin from a common primary magma is unlikely in the present cases, because the relative amount of silicic magma is large in most complexes, much larger than can be derived by fractional crystallisation of a basic parental magma (< 10%; cf. e.g. Hess 1989). Crustal assimilation could contribute to an increase in the relative volume of silicic to basic rocks, but most probably not account for the apparent broadly equal amounts observed in the complexes. The latter is a crude estimate because the different levels of exposure (and absence of detailed geophysical studies) makes it hard to estimate reliably the relative volumes of rocks in each case. In addition, a dominantly intermediate rather than bimodal compositional character of the complexes would be expected. Another major problem with this model is that an extended fractional crystallisation would yield huge amounts of mafic cumulates to account for this amount of silicic rocks, of which there is no sign. Anorthositic cumulates of considerable volume exist in some complexes, but there is no field evidence allowing us to relate these cumulates to fractionation yielding a rapakivi-type residual magma. Magma-mingling structures are common between basic and silicic magmas in the complexes, suggesting that the basic and silicic magmas were actually active contemporaneously, rather than the silicic magma being a late derivative from the basic magma. Hence, separate sources of the associated granitoid and basic magmas in the rapakivi complexes are most likely.

Emslie & Hegner (1993), and Emslie *et al.* (1994) proposed a model to explain crustal-like Nd compositions in anorthosite massifs, associated with rapakivi granites. They suggested that as the mantle-derived basic magmas come into the lower crust and induce partial melting, forming the rapakivi magmas, the restite material left behind after the crustal melting is largely dissolved in the mantle magmas below. This restite would supply the basic magmas with material contributing a crustal Nd isotopic signature (Archaean-dominated in this case). The restite would be dominated by refractory phases such as pyroxene, plagioclase, and probably some accessory phases such as apatite and zircon, left after granite extraction (cf. the experiments of Skjerlie & Johnston 1993; Singh & Johannes 1996). If this restitic assemblage is dominated by these minerals with only insignificant zircon and/or garnet, it would be somewhat LREE enriched (cf. compilation in Grauch 1989), which would be appropriate for an assimilant.

However, this model also encounters some major difficulties, such as low contents of REEs in lower crustal granulites (cf. e.g. compilation in Table 6 of Emslie *et al.* 1994; Kempton *et al.* 1990). These contents are on the same level as REE contents in the most primitive basic rocks in the Subjotnian complexes, but are not high enough to explain those in the more evolved basic rocks, e.g. in the Rödö complex that have substantially higher LREEs (cf. Andersson 1997c). A depleted lower crustal contaminant would thus not be able to increase the LREE content to the observed level. The addition of some undepeleted material to the magmas is therefore likely. Accompanying fractionation of olivine and pyroxenes could also contribute significantly to the increase in REE, and especially the LREE contents of the basic magmas (cf. Emslie 1985; Mitchell *et al.* 1995).

The following scenario is proposed to account for the enriched Nd isotopic character of the Subjotnian basic rocks in central Sweden: (1) mantle-derived magmas associated with the rapakivi magmatism underplate the crust. These magmas are mildly depleted to slightly enriched (similar to those in Finland) and initially have relatively low REE contents (perhaps similar to the least enriched Postjotnian dolerites, $C_{Nd} \approx$ 5–15 ppm, Rämö 1990; Patchett *et al.* 1994; Andersson 1997a); (2) melting of the undepleted lower crust is induced. This crust is dominantly tonalitic to granodioritic in composition, and is composed of both Archaean and early Svecofennian components. The Archaean components are suggested to be dominant in the lowermost parts closest to the lithospheric mantle, being remnants of an old basement in this part of the shield. This Archaean basement was intruded and overlain by calcalkaline rocks in the early Svecofennian. The melting of this composite material results in the strongly LREE-enriched rapakivi granite melts and a pyroxene and plagioclase-dominated residue.

In the lowermost crustal levels, the percentage of Archaean material available for assimilation in the basic magmas is higher. Both the crustal residue and the newly generated rapakivi magmas, as well as undepleted crustal material, should be potentially available as contaminants in this environment. The residue contains plagioclase with a substantial positive Eu peak, which is counteracted by a negative one in pyroxene (and also apatite, if present; cf. Grauch 1989). The net effect of mixing is suggested to be a minor positive anomaly (cf. Taylor & McLennan 1985, p. 81). Undepleted Archaean–Svecofennian metaigneous material is generally expected to lack or show a minor positive Eu anomaly and be strongly LREE-enriched (cf. Jahn *et al.* 1984; Öhlander *et al.* 1987; Andersson 1997c). The rapakivi magmas are more LREE-enriched, but with a substantial negative Eu anomaly (Fig. 3; cf. e.g. Vorma 1976; Rämö 1991; Andersson 1997a,b). Due to the much higher REE concentration in the rapakivi magma, only a minor assimilation of this component would erase a positive anomaly induced by the other two.

It is suggested that the main part of the assimilated material, causing the low initial ε_{Nd} values in the Subjotnian basic rocks, is composed of the depleted residue after extraction of the rapakivi magma. This is supplemented by a relatively minor component of undepleted Archaean-dominated lowermost crust and considerable high-P fractionation of mainly pyroxenes and olivine to arrive at the LREE-enriched composition of the basic rocks. If the mantle magma is mildly depleted and the assimilant is Archaean-dominated (depleted or undepleted), 15–40% assimilation is calculated, which is realistic because most of the Subjotnian basic rocks are quartz tholeiites showing abundant chemical and petrographic signs of crustal involvement (Andersson 1997c). The lowermost crust may also contain substantial amounts of previously underplated more basic rocks, which appears likely in light of the numerous basic granulite xenoliths recovered from volcanic pipes and kimberlites world-wide (e.g. Taylor & McLennan 1985; Rudnick *et al.* 1986; Kempton *et al.* 1990). In this case, more lower-crustal assimilation can be accommodated without significant changes in the composition of the basic magma. A model of this kind has been entertained by Neymark *et al.* (1994) for the Salmi batholith. They reported, for example, similar Nd and Sr isotope characters as in the Swedish Subjotnian rocks, in lower crustal mafic granulite xenoliths recovered from kimberlites in the easternmost part of the shield. Small additions of the newly extracted rapakivi magma may also be incorporated in parts of the suite, which is corroborated by an abundance of felsic xenocrysts and negative Eu anomalies (Fig. 3e). This might have occurred during transport in conduits to the upper crustal emplacement levels.

4.3.3. Features of the northwestern complexes. The differences in initial ε_{Nd} are generally larger between the four NW complexes than between rocks of different composition within each complex (Table 4), suggesting derivation from geographically slightly heterogeneous sources in the lower crust (and upper mantle). If the granitoids (including the intermediate rocks) are derived from melting of a lower crust composed of a heterogeneous mixture of early Svecofennian and Archaean metaigneous sources, the proportions of these two components varied in the sources for the different complexes in accordance with their respective initial ε_{Nd} values. The Strömsund and Mårdsjö complexes, with their lower ε_{Nd} (1·52) compared with Nordsjö and Mullnäset, should thus contain slightly more of an Archaean source component. A spatial variation in the concentration of Nd within each source component could also give such a variation. This is perhaps less likely since such differences would probably be of a small-scale and would tend to average out during magma generation in a source region.

If generation of the complexes from an upper mantle source is favoured, this mantle should also have had a spatial variation in the degree of LREE enrichment. In the Mårdsjö complex, the basic rocks have systematically lower initial ε_{Nd} values than the granitoids. This seems to be the case also for the Nordsjö rocks, but not for those from Mullnäset where the values overlap (Table 4). This might be an indication that the basic and silicic/intermediate rocks derive from different sources, mantle and crustal, respectively. The mantle-derived magmas do not necessarily have, in this case, a higher percentage of the Archaean component, because a small input of the Archaean component which is much richer in Nd than the basic magma may have a large effect on the ε_{Nd}. However, the relatively high contents of Nd (30–40 ppm) in the studied basic rocks (Table 4) favour substantial additions of crustal material to the magmas if a depleted original nature is assumed. High-pressure fractionation of olivine and orthopyroxene can also contribute significantly to LREE enrichment in the residual magma (cf. Emslie 1985; Mitchell *et al.* 1995). On the other hand, if the mantle-derived magmas initially had *c.* 30 ppm Nd (as proposed by Rämö 1991, for the Suomenniemi dykes), the low ε_{Nd} values must be reflecting an enriched mantle source.

The Nd-data of the NW intrusions are very similar to those obtained from the Salmi rapakivi complex in Russian Karelia (Rämö 1991; Neymark *et al.* 1994), which is located at the margin of the exposed Archaean craton; therefore a large Archaean component in its magmas is expected (Rämö 1991; Neymark *et al.* 1994). The Nd-evolution with time for this batholith strongly overlaps that of the NW and Rödö complexes (Fig. 5). This circumstance supports the presence of an Archaean basement also beneath central Sweden.

4.4. Sr isotopic relations

The Rb–Sr system in the silicic Subjotnian rocks has been severely disturbed, obscuring the primary igneous Sr isotopic compositions. All samples, except one, have unrealistically low initial $^{86}Sr/^{87}Sr$ (I_{Sr}) ratios (Tables 3, 4), lower than, for example, the estimated bulk earth (UR, I_{Sr} (1·51) = 0·7027), which is due to a later closure of the Rb–Sr system, postcrystallisation increase in the Rb/Sr ratio and/or loss of radiogenic Sr (cf. e.g. Romer *et al.* 1992). The reason and timing for this disturbance remain unclear. In the case of Rödön it may be related to fluid circulation connected to the *c.* 550 Ma (Kresten 1990) Alnö intrusion 7 km to the NW. The large sills of Postjotnian dolerites (*c.* 1·26 Ga; Suominen 1991) which penetrate the granite area including the granite (Lundqvist *et al.* 1990; aeromagnetic map, Geological Survey of Sweden 1997) could also have induced a redistribution of these elements. A Postjotnian influence on the Rödö granite is supported by its palaeomagnetic pole position (Moakhar & Elming 2000). Some of the basic to intermediate Rödö rocks also have disturbed Rb–Sr systems, with high Rb/Sr ratios and low apparent initial ratios. A few, however, have realistic I_{Sr} values, and can be divided into two groups: one with relatively primitive initial Sr ratios (0·7036–0·7064) and one with more evolved values (0·7081–0·7188).

The $^{86}Sr/^{87}Sr$ (1·51)-ratio is plotted against ε_{Nd} (1·51) in Figure 10 for the Rödö and NW rocks, together with other rocks in the Bothnian basin area (Tables 3, 4; Claesson & Lundqvist 1995), and a few Archaean rocks from the NE part of the shield (Martin *et al.* 1983). Very little information on the primary initial Sr composition of the silicic rocks can

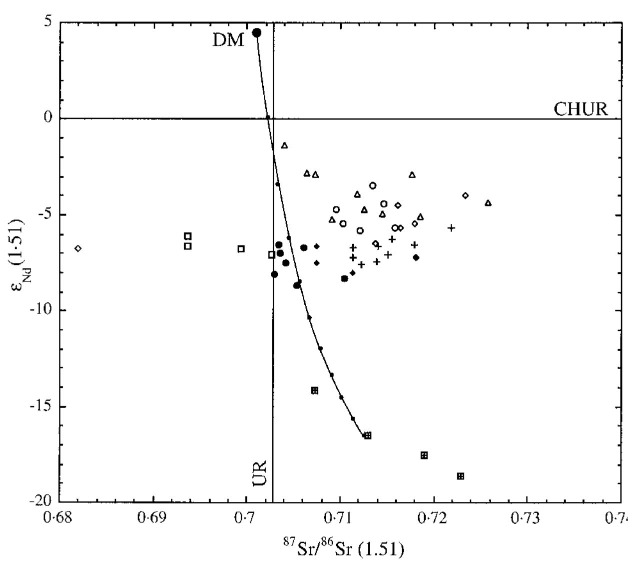

Figure 10 ε_{Nd} vs $^{87}Sr/^{86}Sr$ diagram at 1·51 Ga for the rapakivi rocks of this paper (Tables 3 and 4) and relevant country rocks in the Bothnian basin (Claesson & Lundqvist 1995; this paper), including a few Archaean rocks from the NE part of the shield (Martin *et al.* 1983), representing a possible Archaean component. Unfilled squares = rapakivi granites, filled circles = basic rapakivi-related rocks, filled diamonds = hybrid porphyries in the Rödö complex, crosses = metasediments in the Bothnian basin, triangles = early Svecofennian granitoids, unfilled circles = Revsund granitoids, unfilled diamonds = Härnö granites. CHUR is the chrondritic uniform reservoir (Nd) and UR is the hypothetical uniform reservoir (Sr). Many samples plot off the diagram to the far left (cf. Tables 3 and 4); these have strongly disturbed (not preserved magmatic) Rb–Sr systematics. A calculated, hypothetical mixing curve between a depleted mantle and Archaean lower crustal sources has been included; dots on the curve represent 10% increments. Concentrations of Nd and Sr, assumed in the calculation were, for DM: 10 ppm Nd and 200 ppm Sr, and for the Archaean source: 24 ppm Nd and 200 ppm Sr.

be extracted, due to the strong disturbance of the Rb–Sr systems. The strongly disturbed rocks plot to the left of the UR vertical line (Fig. 10) and partly far off the diagram to the left. Only one granite (M14a) from the NW shows an undisturbed (?) composition, which is identical to the value from the uniform reservoir (UR; Table 4; Fig. 10). This meagre evidence seems to suggest that the silicic rapakivi magmas may be characterised by a primary low-radiogenic Sr composition, close to that of the basic rocks, which would preclude Revsund and metasedimentary source components in the magmas.

The crustal component that has lowered the initial ε_{Nd} in the basic rocks should have been relatively depleted in Rb (low Rb/Sr ratio) for those rocks which plot close to UR (Fig. 10). Even though all basic-intermediate Subjotnian rocks have low initial ε_{Nd} values, their I_{Sr} values vary from primitive to evolved. This is consistent with primitive mantle sources for the basic magmatism combined with different types and degrees of crustal assimilation. A calculated mixing curve, between a depleted mantle and an Archaean component, has been included in Figure 10. This calculation demonstrates that with respect to the Sr isotopic composition, as with the Nd data, a mixing of a mantle-derived Subjotnian magma with 30–40% Archaean lower crust is compatible with those samples having the lowest initial Sr-ratio. The group of basic-hybrid rocks having higher initial Sr should have assimilated material with high Rb/Sr ratios and thus more radiogenic $^{86}Sr/^{87}Sr$ ratios. The trend towards the metasediments in Figure 10 suggests that these could be a substantial assimilant in some of these magmas. However, since $^{86}Sr/^{87}Sr$ (1·51) ratios for the early Svecofennian granitoids also range up to high values, mixtures between granitoids and Archaean sources may account for the spread in initial Sr ratios in the basic-hybrid rocks. In any case, undepleted components have been assimilated in these rocks, presumably in the upper crust. In the hybrid porphyries there is abundant evidence of magma mixing with crystal-bearing rapakivi magmas, but in some cases there is petrographic evidence also for upper crustal assimilation. For example, sample R134a contains large scattered pink Kfsp megacrysts dissimilar to those in the rapakivi rocks, as well as cellular megacrysts of albite, of unknown, presumably xenocrystic, origin (cf. Andersson 1997c).

The Sr isotopic composition of a combination of early Svecofennian and Archaean metaigneous more or less depleted lower crustal material, residual after extraction of the rapakivi granite magmas, will be close to that of the basic rapakivi-related rocks (curve). The residue will be dominated by pyroxene and plagioclase and be rich in Sr and poor in Rb (cf. Emslie *et al*. 1994). Because $^{86}Sr/^{87}Sr$ (1·51) ratios of the residue will probably be close to that of the basic magmas, assimilation will have only a negligible effect on the Sr isotopic ratio of the basic magmas. Therefore, a model of assimilation of dominantly residual lower crustal components in the basic rocks with lowest initial Sr values is supported by the Sr data.

4.5. Pb isotopic relations

4.5.1. Feldspar Pb isotope relations in the NW complexes. The results of the Pb feldspar analyses of rocks from the NW complexes (Table 5) spread out between the model curves for the mantle and Archaean lower crust (Fig. 8) (cf. Rämö 1991). Sample Mu8 has an elevated $^{206}Pb/^{204}Pb$-ratio compared with the others that can be ascribed either to some remaining U, which has resulted in excess radiogenic Pb, or to a disturbance of the WR U–Pb system in similarity with its Rb–Sr system. The low Pb isotopic ratios indicate a marked but apparently more heterogeneous Archaean component in the magmas, compared with the Nd isotopes. This Archaean component is indicated to be U-depleted (low-radiogenic) and, therefore, no indications of U-rich Archaean upper crustal components are present in these magmas.

Two of the samples (M15a, gabbro, and Mu12, syenite), from different complexes, plot at very primitive (low-radiogenic) compositions near the lower crustal curve in Figure 8a, whereas the others spread toward higher $^{207}Pb/^{204}Pb$ values closer to the mantle curve. Most of the samples plot relatively close to the 1520 Ma second-stage model isochron after Stacey & Kramers (1975), which indicates that the samples essentially have retained their values since the time of crystallisation. The Pb isotopic composition of these samples is significantly more primitive than those of the corresponding rapakivi feldspars in Finland (Fig. 8a; Rämö 1991). The latter can be correlated with a Pb evolution from relatively evolved early Svecofennian sources (CFB or B in Fig. 8a) in agreement with the Nd isotopes (Rämö 1991).

The Pb results of the central Swedish Subjotnian feldspars can be interpreted similarly to the Nd isotopes above, i.e. as being formed by a mixture of Archaean and Svecofennian source components. In this case, however, the relative importance of these components appears to differ substantially from sample to sample even within the same complex, while the Nd isotope systematics suggested a more homogeneous mixture of components (cf. Table 4). One explanation for this is the large spread in possible Pb isotopic compositions for the crustal components.

Measured Pb isotopic compositions in galena from a large number of Svecofennian mineralisations have been plotted in Figure 8, and the spread of the data is substantial. There appears to be a geographic control on the Pb isotopic composition of the early Svecofennian galenas. The most primitive, low U/Pb sources are present in the Skellefte district in the N and the most evolved high U/Pb ones in the Bergslagen district in the S, while the Bothnian basin samples (E and CFB) fall in between. This trend also correlates with the distance from the exposed Archaean craton and the most primitive galenas in the Skellefte district may thus contain a significant component of U-depleted Archaean lower crust (not, however, recorded in Nd-data). At the same time, the Pb-ores in Bergslagen seem to be derived entirely from early Svecofennian sources. The role of the mantle component is, however, difficult to constrain from the Pb isotope data alone. The implication of this distribution regarding mixed sources has been discussed by several authors (e.g. Vaasjoki 1981; Huhma 1986; Billström 1989; Sundblad 1991; Sundblad *et al.* 1993; Billström & Vivallo 1994), and means that it is hard to constrain the isotopic composition of an early Svecofennian source component.

The most probable composition is, however, represented by the areas labelled E or CFB in Figure 8. E is the Enåsen mineralisation (Hallberg 1989) located in the western part of the Bothnian basin, and CFB is the central Finnish batholith area (Vaasjoki 1981), corresponding to galena Pb compositions from the eastern segment of the Bothnian basin. A Svecofennian component with this composition can be mixed in different proportions with an Archaean component plotting on the hypothetical evolution curve of Archaean lower crust ($\mu_2 = 8{\cdot}00$; Rämö 1991), to yield the range in the analysed silicic-intermediate rapakivi samples. However, since the compositions of the components in the mixture cannot be unequivocally determined, it is impossible to quantify their relative proportions. If the compositions of the components are the same for all the granitoid samples, the proportions will not be the same as calculated from the Nd isotopes. It is, however, probable that the Pb isotopes are more sensitive tracers of small-scale variations in the U/Pb ratio in the lower crustal source regions, and that this has influenced the data. Also, U/Pb and Th/Pb ratios might have varied in the Archaean lower crust due to different degrees of depletion, resulting in Pb isotope compositions deviating from the $\mu_2 = 8{\cdot}00$-curve. Sample Mu12 appears to contain a more depleted Archaean component, compared with the other syenites. This may, however, be due to a more primitive composition of the Svecofennian component, similar to galenas from the Skellefte district (Sk in Fig. 8). The Pb in feldspar of the gabbro of the same complex (Mu11a) is substantially more radiogenic. This is possibly related to a mantle source with $\mu_2 = 9{\cdot}17$ or slightly higher for the parent magma, which has assimilated relatively undepleted Archaean material. Such interpretation is consistent with the quartz-rich, high-silica character of this rock. The other granites and syenites require more radiogenic Svecofennian components, similar to those of CFB or Enåsen, and the Archaean component may be less depleted. The low-radiogenic Pb isotope composition of the Mårdsjö gabbro (M15a) should be related to an assimilated strongly U-depleted Archaean component (cf. ε_{Nd} values, Table 4), which is consistent with its more basic character, compared with Mu11a.

The feldspars from Svecofennian country rocks show markedly evolved (radiogenic) Pb compositions, which suggest that they contain U giving *in situ* radiogenic Pb (Table 5). They conform relatively well to the evolution curve for 'juvenile Svecofennian' crust ($\mu_2 = 9{\cdot}81$; Rämö 1991), which has its starting point in the CFB-field (Fig. 8a).

With regard to ^{208}Pb, the samples spread between primitive compositions in the area of Archaean Pb and evolved compositions near those of the Finnish rapakivis. Sample Mu8 and the Svecofennian feldspars show evidence of excess ^{208}Pb. Curves for the evolution of the mantle and upper and lower crust have been added in Figure 8b (after Zartman & Doe 1981). The ^{208}Pb isotopic contents are also consistent with different degrees of mixing of an Archaean component with a Svecofennian and/or mantle component, but quantification of their relative contributions is impossible. Evidently, samples M5a and Mu12 have more primitive compositions and appear to contain a higher degree of older material than the others, which is not generally supported by the Nd isotopic systematics. In comparison with feldspars from the Salmi complex, the central Swedish ones are more radiogenic with respect to U (^{206}Pb/^{204}Pb), but mainly less radiogenic with respect to Th (^{208}Pb/^{204}Pb), implying lower source Th/U ratios. The old lower crustal protoliths of the Salmi complex are thus more strongly depleted in U than in Th (cf. Larin *et al.* 1990; Neymark *et al.* 1994), compared with the protoliths of the central Swedish complexes.

Summarising this section: Pb isotopes alone do not give conclusive evidence for the relative proportions of the different source components, but still clearly indicate a major contribution of a much older lower crust in this magmatism. Ashwal *et al.* (1986) and Ashwal (1993) have also interpreted similar Pb isotopic variations in Mesoproterozoic anorthosite–granite complexes in Labrador as resulting from interactions with a depleted, low-radiogenic Archaean crust.

4.5.2. Pb–Nd relations. Relations between samples where both WR–Nd and fsp–Pb have been measured are presented in Figure 11, together with the estimated isotopic composition of some potential sources (from Table 6 and Fig. 8). Because variations in initial ε_{Nd} between the rocks are limited, while the ^{207}Pb/^{204}Pb ratios vary more strongly, this diagram provides further insights into possible mixing processes. Different mixing curves have been calculated, using primarily different relative contents of Pb, and those deemed more realistic have been plotted on the diagram. The gabbro (M15a) having the most primitive (low-radiogenic) Pb isotopic composition may have formed as a result of mixing of a depleted mantle component with *c.* 45% of an Archaean lower crustal component with an approximately five times higher Pb content. Such mantle magma–lower crustal mixing has been proposed to explain the variation in isotopic composition among suites of mafic lower crustal xenoliths from alkali basalts (Rudnick & Goldstein 1990). The other gabbro (Mu11a) seems to require an additional *c.* 20% input of an early Svecofennian component to this mixed mantle–lower crust (MLC) composition. The Subjotnian granites, which have the same Nd isotopic composition as the gabbros, are not likely to contain any mantle component (see previous discussion). Instead, they can be modelled (in agreement with the Nd discussion above) as a mixture of Svecofennian 'juvenile' crust and 30–40% Archaean lower crust with similar Pb contents (Fig. 11). The syenites, on the other hand, show a more complicated pattern, with ^{207}Pb/^{204}Pb ratios ranging from the most primitive to more evolved values. This can perhaps also be understood in terms of mixing of several sources. Regarding the most primitive sample (Mu12), it seems to have a mixed mantle–Archaean crust origin with little Svecofennian input. In this case, the mixing most likely occurred in pre-Subjotnian times (during the early Svecofennian or Revsund events), thus forming a mixed mantle–Archaean lower crust (MLC in Fig. 11) of diorite–menzodiorite composition, suitable for generating syenite magmas. The other syenite samples apparently have a considerable input of Svecofennian material. In the figure, a mixing curve has also been sketched between the Svecofennian 'juvenile' crust and an Archaean upper crust to show that the metasediment sample (M1) fits such a model, in accordance with the detrital zircon ages reported by Claesson *et al.* (1993).

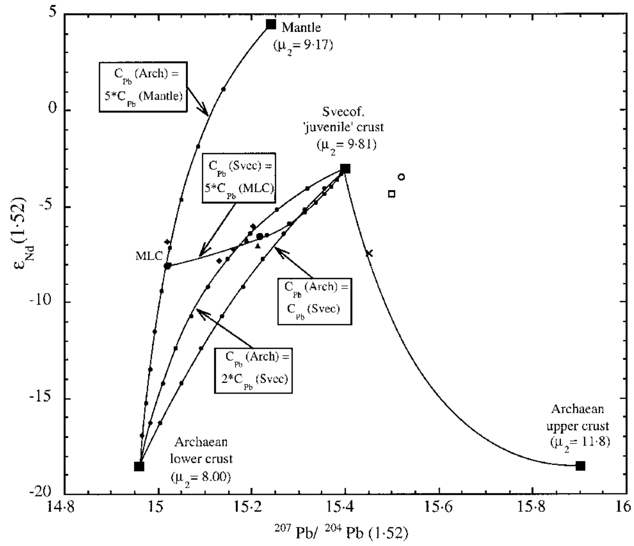

Figure 11 Plot of isotopic composition of whole rock–Nd and ^{207}Pb/^{204}Pb-fsp for the NW rocks, and some potential source rocks at 1·52 Ga. Solid triangles = granites, solid diamonds = syenites, solid circles = gabbros, open circle = Revsund granite, open square = early Svecofennian granite and inclined cross = metasediment. Curves with small dots are calculated mixing lines (dots meaning 10% increments), assuming certain ratios between Pb contents in the different sources (stated in the frames), which reproduced the most realistic results. C_{Nd} used was, for mantle: 15 ppm, Archaean lower crust: 24 ppm, Svecofennian crust: 35 ppm (from Table 6), and mixed mantle–lower crust (MLC): 20 ppm. A mixing curve has also been sketched between the Svecofennian 'juvenile' crust and Archaean upper crust, to show that such a relationship is indicated for the metasediment sample (M1).

4.5.3. Pb–whole rock relations in some Rödö rocks. The six least radiogenic samples appear to form a linear array (Fig. 9). The very imprecise 'age' of 1·76 ± 1·10 Ga corresponding to the errorchron has no geological meaning. The linear distribution of the data points is probably only an artifact of varying initial Pb compositions, where higher $^{207}Pb/^{204}Pb$ ratios are accompanied by higher μ-values ($^{238}U/^{204}Pb$).

Extrapolation of secondary 1500 Ma reference lines from the range in WR-compositions backwards to possible initial compositions, coincides with the compositions of the more radiogenic feldspars of the NW complexes, between the evolution curves for the mantle ($\mu_2 = 9{\cdot}17$) and Archaean lower crust ($\mu_2 = 8{\cdot}00$) (Fig. 9, cf. Fig. 8a). This supports the involvement of Archaean lower crustal components of similar Pb isotopic compositions in the Rödö magmas as in the NW complexes. This contrasts with the Pb data for the SE Finnish rapakivi granitoids (Rämö 1991), which have distinctly higher WR$-^{207}Pb/^{204}Pb$ ratios and trend to initial compositions of relatively evolved (radiogenic) Svecofennian galenas (cf. Fig. 8a), showing no signs of lower crustal contribution. The Pb–WR data thus support the conclusion from the Nd isotopes of a major Archaean lower crustal input to the Rödö rapakivi magma. The sample R117g1 extrapolates to an unreasonably low initial $^{207}Pb/^{204}Pb$ ratio. This sample is extremely rich in U (52·8 ppm) and probably has not remained a closed system after crystallisation (may have experienced U gain).

5. Conclusions

(1) Ages of intermediate–silicic rocks from four different Mesoproterozoic (Subjotnian) rapakivi complexes from central Sweden have been determined to lie in the range 1529–1491 Ma (including errors). This, together with data from other complexes extending southwards (Andersson 1997a; Claesson & Kresten 1997; Persson 1999), shows that the Swedish Subjotnian magmatism belongs to the youngest episode (1·53–1·47 Ga) of rapakivi magmatism in the Fennoscandian shield. The spatial distribution of the Subjotnian complexes appears to take the form of discrete N–S-trending belts, with a stepwise eastwards increase in age (1·53–1·47, 1·59–1·56, 1·65–1·61 Ga), except for the easternmost part which is again younger (1·56–1·53 Ga) (Fig. 1). The reason for this age distribution is at present not clear, but could be related to shifting extended post-collisional extensional collapse after the Svecofennian orogeny (Korja *et al.* 1993; Windley 1993; Korja & Heikkinen 1995), or to continent-side extension in response to eastward subduction under the continental margin in the W (Åhäll *et al.* 2000).

(2) Isotopic results of Nd, Sr and Pb strongly suggest that silicic, intermediate and basic magmas in all studied complexes contain a considerable amount (generally *c.* 20–40%) of an old Archaean source component. In the case of the basic rocks, this could be due to derivation from a mantle that was LREE-enriched, e.g. during a previous subduction event. More likely, the basic Subjotnian magmas were generated in a non-enriched mantle and then mixed with Archaean lower crustal material. Lower crustal Archaean source material is suggested for the associated silicic–intermediate rocks. These can be isotopically modelled as derived from a mixture of dominantly early Svecofennian and *c.* 30–40% Archaean metaigneous protoliths, rather than by fractionation from basic parental magmas. The latter is unlikely primarily because of lack of field evidence for extended crystal fractionation, evidence for coeval emplacement of magmas of different composition, and volume relationship between basic and silicic rocks. The nature of these crustal protoliths, however, may have varied from relatively mafic, containing depleted crustal components, formed during a previous episode of underplating (sources for syenite magmas), to essentially undepleted crustal mixed Archaean–early Svecofennian (sources for the rapakivi granites).

(3) A major section of previously undetected Archaean basement rocks are thus indicated to be preserved, buried beneath the Svecofennian successions in central Sweden. Direct evidence for this has recently been provided by the discovery of *c.* 2·7 Ga zircons recovered from the Nordsjö syenite (Claesson *et al.* 1997). The 'provenance' of this Archaean basement is unclear; it may be exotic, for example, have an origin in Laurentia, or be an early crustal block rifted off the craton to the NE in pre-Svecofennian, early Palaeoproterozoic times.

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7. Appendix. National grid coordinates of samples

Sample number	National grid coordinates
R126a	692031/159275
R127:1 + 2	692077/159238
R133:2	692130/159243
R138	692222/159610
R126f	692039/159286
R96a + b	691838/158520
R73a	692458/159520
R144a + b	692026/159122
R160	692178/159144
R167:2	692168/158953
R117g1	691511/159039
R133b1	692129/159244
R115a1	691989/159153
R68	691882/159037
R18Enk1	691772/158958
R151h	692102/159153
R86a	691780/158667
R101a + b	691667/158441
R134a	691700/158385
R111	691862/159039
R14c	692015/159123
R89	691829/157824
R91	692224/158190
R22	692032/159035
R38b	692017/158864
R15	691756/158930
R94a + b	692412/158001
M14a	702589/149466
M15b	702248/149593
M11	702265/149736
M20	702292/149404
M18	702300/149462
M15a	702248/149593
M5	703407/148803
M7	704115/148207
M1	700580/149005

continued on facing page

Sample number	National grid coordinates
N2	706375/151745
N8	706240/151415
N9	706320/151810
N11	706260/151648
N7	706270/152128
S4	708740/149535
S6	708580/149476
S7	708708/149449
Mu8	707144/149270
Mu5	707197/148955
Mu12	706974/149315
Mu11a	706800/149205
Mu2	707309/149000
Mu4	706812/148752

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