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Origin of continental crust of 1.9–1.7 Ga age: Nd isotopes and U-Pb zircon ages in the Svecokarelian terrain of South Finland

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Abstract. Initial Nd isotopic ratios are reported for 23 samples representing magmatic crustal components in the Svecokarelian terrain of South Finland. U-Pb zircon ages are determined for all geologic units, involving 21 separate upper concordia intercept ages based on more than 100 U-Pb analyses. The ages range for all the rocks from 1.90 Ga for primitve plutonic rocks to 1.79 Ga for post-tectonic intrusions. The well-known gabbro-diorite-tonalite-trondhjemite association of the Kalanti district appears to consist of components with different ages: trondhjemites are probably 1.90 Ga or older, diorites/tonalites belong to the main Svecokarelian plutonic eposide at 1.89–1.87 Ga, and at least some gabbro has a post-tectonic age of 1.80 Ga.

ε_{Nd} (T) values range between $+2$ to $+3$ for meta-andesites, large gabbros and primitive granitoids to -0.5 for more evolved granitoids. A magma source with ε_{Nd} of at least $+2$ to $+3$ was available during 1.90 to 1.87 Ga, but evolved granitoids have ε_{Nd} close to zero. The preferred interpretation is that depleted mantle with $\varepsilon_{Nd} = +4$ to $+5$ was present beneath the Svecokarelian crust forming during 1.9–1.8 Ga, and that all rocks have been affected more or less by addition of an Archean crustal component with $\varepsilon_{Nd} = -9$ to -13. The primitive rocks with $\varepsilon_{Nd} = +2$ to $+3$ were only slightly affected, while granitoids with ε_{Nd} close to zero include a $\sim 10\%$ Archean component. The widespread nature of the Archean addition and the distance of up to 500 km to actual exposed Archean crust make it most realistic that the Archean component was added to the form of sediments delivered by subducting Proterozoic ocean crust. The plutonic rocks of the Finnish Svecokarelian crust in areas away from Archean cratons consist of $\sim 90\%$ newly mantle-derived material.

Introduction: 1.9–1.7 Ga crustal genesis and the Svecokarelian terrain

The 1.9–1.7 Ga cycle of tectonothermal events took place primarily in the present-day northern hemisphere of the Earth, and was the event responsible for establishing the "North Atlantic continents" in their cratonic form (Fig. 1, inset), welding together Archean continents and "new" 1.9 – 1.7 Ga crust. The crust stabilized during 1.9–1.7 Ga could have three fundamentally different types of origin: (1) simple reheating of crust that was separated from the mantle at a much earlier time; (2) crust produced from reprocessing of older continental materials through the sedimentary system; and (3) crust produced from the mantle without influence of older crust. Any single rock produced could have any of these origins or any combination of them. Sr, Pb, Nd and Hf isotopes can be used to distinguish newly mantle-derived material from recycled crust because of the large parent/daughter chemical fractionations involved in crust formation, and the resulting isotopic differences which develop with time between mantle and established crust. Nd and Hf (particulary Nd) are most useful in this respect because the mantle/crust fractionation effects are reasonably constant and well-understood, and because the isotopic compositions are insensitive to events on a less than 0.1 Ga time-scale. This means that the complexity of recycling and granitoid generation during development of a crustal terrain reduces to a single event as far as Nd and Hf isotopes are concerned, with isotopic values reflecting the contributing components rather than the details of the orogenic process.

The Svecokarelian terrain of Finland and Sweden (Fig. 1) is one of the largest regions of uninterrupted 1.9–1.7 Ga crust and is coherent in sedimentation, metamorphism and magmatism. This is so in spite of the fact that Svecokarelian geology has always been reviewed separately for Finland and for Sweden (e.g. Magnusson 1965; Hietanen 1975; Lundqvist 1979; Simonen 1980; Gaál 1982). For general factual background in Finland, the review of Simonen (1980) serves well. The Archean continental areas of Fennoscandia (Fig. 1) are overlain by platform sediments deposited from about 2.20 Ga onwards (e.g. Sakko 1971), and westwards this "Jatulian" shallow-water sedimentary facies gives way to a deeper-water turbidite-pelite association which is reasonably constant over the whole of southwestern Finland and northern and central Sweden. The earliest tectonic activity probably took place before 1.9 Ga, and there are certainly gabbros of 1.97 Ga age associated with serpentinites (e.g. Koistinen 1981). Many plutonic and volcanic rocks have U – Pb zircon ages

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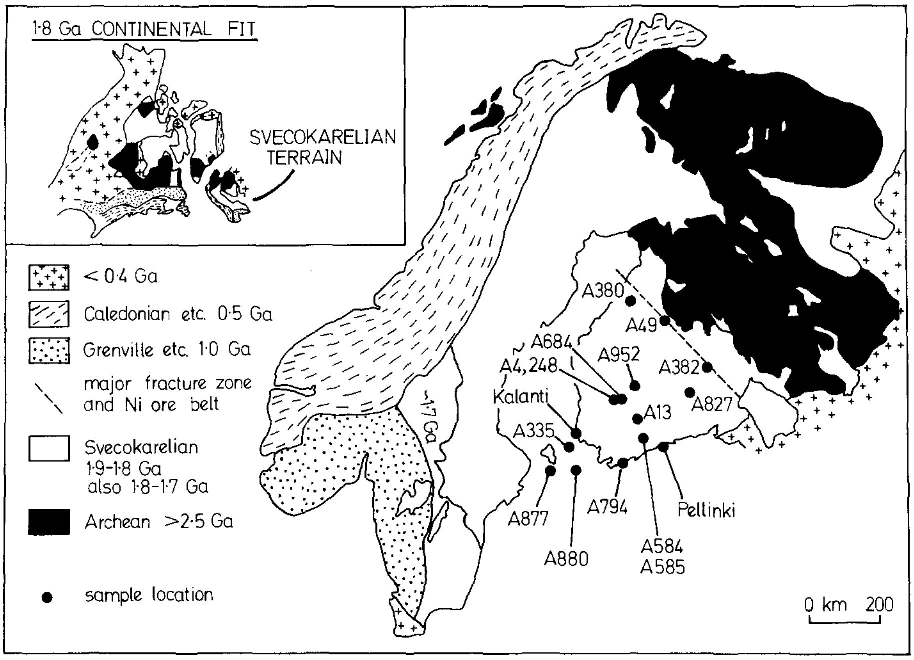

Fig. 1. The Svecokarelian terrain of Finland and Sweden, showing principal geologic features and sample locations. Inset: the North Atlantic region in its 1.8–1.2 Ga configuration

between 1.90 and 1.86 Ga (Kouvo and Tilton 1966; Gaál and Rauhamäki 1971; Skiöld 1979; Welin et al. 1980a, b; Aftalion et al. 1981; Salli 1983). These "main Svecokarelian events" are separated rather distinctly from late- and post-tectonic granitoids at 1.80 Ga (e.g. Table 2 of this paper), an apparent major belt (Fig. 1) of ~1.70 Ga volvanics and plutons in Sweden (Gorbatschev 1980), and the dispersed plutons of the 1.67–1.55 Ga "rapakivi episode" (Vaasjoki 1977).

There is no definite indication of where any possible plate-tectonic boundaries lay in the Svecokarelian terrain, although a major fault belt with considerable mafic and ultramafic rocks as well as transition-metal ore bodies (the Main Sulphide Ore Belt) does cut across Finland close to and parallel to the margin of the Archean continent (Fig. 1). The special character of this Main Sulphide Ore Belt is shown by a coincident deep negative gravity anomaly, and a distinct difference in isotopic composition of sulphide lead between the ore belt and areas to the west or east. Pb isotopes in the ore belt show pre-1.9 Ga systematics with a low μ-value which might well be consistent with a mantle origin. According to Vaasjoki (1981) and also Stacey et al. (1977), the ore belt Pb isotopic signatures vary from essentially mantle-type (Outokumpu) through those of more usual type with some continental influence (e.g. Pyhäsalmi), to those with "orogenic-type" Pb in the model of Doe and Zartman (1979). As a tectonic boundary, the Main Sulphide Ore Belt separates a region to the east with predominant shallow-water Proterozoic sediments and Archean gneiss domes from a region to the west with mainly flysch-type semipelitic micaceous schists and mafic-intermediate volcanics, and having no Archean relics. This zone has occasionally been suggested as a Svecokarelian suture zone (e.g. Bowes 1980). Otherwise, the Svecokarelian is conspicuous in lacking any convincing ophiolites or true pelagic sediments on any significant scale, and being characterized by high-temperature, low-pressure metamorphic environments (Lundqvist 1979; Simonen 1980). In terms of rock components, the Svecokarelian crust over the whole region can be summarized as ~75% granitoids and intermediate-felsic volcanics plus ~20% metaturbidites and metapelites, with only very small components of basalt, gabbro or other types of sediment.

The Svecokarelian samples studied

The nature of the present sampling was dictated by Nd isotopic criteria rather than systematic U–Pb geochronology. The general aim was to sample most major igneous units of the Svecokarelian 1.90–1.80 Ga crustal development while avoiding areas close to Archean crustal relics. Thus the main part of the Finnish Svecokarelian terrain in the southwestern half of Finland was sampled for volcanics, gabbros, trondhjemites, diorites/tonalites, granodiorites and granites (Tables 2, 3 and Fig. 1).

The Pellinki district (Fig. 1) has well-preserved arkosic and other sediments overlain by pyroclastics and basaltic to andesitic lavas (Laitala 1973, 1984). From the descriptions and statements in Laitala (1973, 1984), we take the rock association to correspond to a present-day island-arc environment.

Intrusive gabbros were taken from three intrusions formerly studied for Hf isotopes in zircons (A684, A794, A877; Patchett et al. 1981), for the purposes of $\varepsilon_{Nd}-\varepsilon_{Hf}$ comparison. These three samples were the late-crystallized mafic pegmatoid phase used for zircon separation. More important geologically is that two major gabbro bodies, with sizes of several km, were included. Hyvinkää is situated in southern Finland and was described by Härme (1954, 1978), who considered it and neighboring gabbro bodies to be the subvolcanic magma chambers of the mafic and intermediate volcanics with which they are surrounded. For this gabbro a pegmatitic facies (A585) is used for U–Pb zircon, and a normal gabbro (A584) for Sm–Nd analysis. The Ylivieska gabbro from central Finland was described by Salli

(1961), and is a large body of noritic gabbro; sample A380 is a normal gabbro used both for Sm – Nd and U – Pb zircon analysis.

The Kalanti disdtrict of southwesternmost mainland Finland became classic for the occurrence of trondhjemites and related rocks after the work of Hietanen (1943). Gabbros, diorites, tonalites and trondhjemites occur together and have appeared to be related in an evolution series. Hietanen's work was extended by Arth et al. (1978), who reported considerable petrographic, major-element and REE data and presented a detailed evolutionary scheme whereby trondhjemite was the ultimate differentiation product of gabbro. The samples all lay close to a Rb – Sr regression line of 1,934 ± 104 Ma (2σ error, Arth et al. 1978). Our Kalanti samples are all from the southern part of the district in the Uusikaupunki and Kalanti areas together with nearby islands. The samples include: trondhjemites A493 (A493b is equivalent), A553, A955 and A174; diorites/tonalites A552, A554 from our collection and 103/1939 from Hietanen (1943); gabbros A568a, a pegmatitic facies, and A568b, a normal type. Locality names for the Kalanti district are given in Table 1 and the appendix. Several of the samples are from the same localities used by Hietanen (1943) and Arth et al. (1978), and details are available on request.

Syn-tectonic granitoids with ages close to 1.88 Ga are sampled from right across the Svecokarelian in Finland (Fig. 1). The Aulanko granodiorite (A13) is described by Simonen (1948), Vaaraslahti hypersthene-granite (A49) by Salli (1983), who also gives a U – Pb zircon age. Hämeenkyrö granodiorite (A4, A248) is described by Simonen (1952), Voinsalmi hypersthene-granite (A382, A383) by Gaál and Rauhamäki (1971) who also give a U – Pb zircon age, Mörskär by Suominen (1985, in preparation) and Kopelojärvi (A952, A952b) by Laitakari (1985, in preparation) and Laiti (1985, in preparation).

The ~1.80 Ga post-organic granitoids were previously analyzed for Hf in zircons by Patchett et al. (1981), and those authors give geological and age references.

In general, where precise U – Pb zircon age data for the samples, or a summary thereof, is not available in the literature, those data are reported here. Some reported U – Pb ages have improved with new analyses for this work. The samples used for Sm – Nd work are identical with those used for U – Pb zircon dating wherever possible. Where a different sample was used, this was either because (a) for Sm – Nd normal gabbros were preferred to the late pegmatic facies used to separate zircon, or (b) sufficiently large pieces of coarse-grained granitoids were not available for making statistically meaningful rock powders, necessitating a new sample from the same locality or at least the same geologic unit.

Analytical methods

U – Pb zircon and monazite chemistry and mass spectrometry followed techniques of Krogh (1973). For titanites strong HCl was used to separate uranium and lead in the presence of titanium and iron in anion-exchange columns. All isotopic analyses were performed on a 60°, 9″ Nier-type mass spectrometer of the Geological Survey of Finland. Corrections for common Pb in analyses follow the California Institute of Technology composition. Uranium and lead spikes were calibrated against shelf solutions checked through interlaboratory calibration.

Sm – Nd analyses were carried out at MPI Mainz. Analytical techniques for Sm – Nd of old samples have been described in detail by Patchett and Bridgwater (1984), and only the following salient points need to be repeated here:

(1) Sample dissolution was performed in sealed bombs at 180° C over 7 days, after preliminary dissolution of major minerals in the open bomb on a hotplate. This two-step procedure ensures zircon dissolution.

(2) Our ^{149}Sm-^{145}Nd isotopic spike solutions have been calibrated in triplicate against standard solutions prepared from high-purity Sm and Nd metals. A calibration check using the Sm – Nd solution distributed by the California Institute of Technology yielded agreement at the 0.02% level for Sm/Nd.

(3) Although no duplicate analyses are reported here, Patchett and Bridgwater (1984) report eight duplicates which show that for a rock powder, ^{147}Sm/^{144}Nd can be reproduced to within 0.4%, and ε_{Nd} (T) to within 0.5 ε-units. We regard these as conservative and reliable error estimates.

(4) This study was concurrent with that on South Greenland, and the mean value for the La Jolla Nd isotopic standard was ^{143}Nd/^{144}Nd = 0.511848 where 0.000013 was the standard deviation of a population of 22 runs. ^{146}Nd/^{144}Nd = 0.7219 was used to correct all Nd isotopic ratios for fractionation.

U – Pb zircon ages

Available age determinations on samples studied for neodymium isotopes are summarized in Table 2 and Figures 2–3. Aside from the high and dispersed ages on silicic trondhjemites from the Kalanti region (Fig. 1), the dates all fall in the 1.872–1.893 Ga range common for the synorogenic intrusions in S.W. Finland (or 1.872–1.887 Ga if isochrons with high lower intercept are omitted), and 1.794–1.803 Ga on two post-tectonic intrusions. These ages are simple in interpretation, very consistent, and need no further comment (see Table 2).

The Kalanti trondhjemite complex is approximately correlative with the other synorogenic intrusions as far as the diorite (Hietanen 1943; melatonalite by Arth et al. 1978) of the Uusikaupunki area (A522, A554), normal gabbro type of Tynki (A568b, Fig. 2) and pyroxene trondhjemite of Putsaari (A174) are concerned. Arth et al. (1978) were able to show that the concentrations of all rare earths decrease, and europium anomalies become increasingly positive with increasing SiO_2 content in the supposed Kalanti evolution trend. During the course of this study it became evident that zircons from the silicic members of the trondhjemite suite show higher and diverging ages. Data for the nine zircon fractions from sample A493-Lepäinen show uranium contents from 700–1,200 ppm. On the concordia diagram in Fig. 2 the U – Pb analyses are not colinear but define (probably accidentally) two four point chords which have upper intercepts corresponding to ages of 1.910 ± 0.023 Ga and 1.959 ± 0.014 Ga. Instead of representing two different ages, the analyses should probably be regarded as covering a wedge-shaped area, because two groups cannot be distinguished on any criteria. All fractions are clear, euhedral long needles, with length/breadth (L/B) ratios being mainly 2 to 3 but reaching as high as 7. Among density fractions less than 4.1 g cm^{-3} some crystals were found having relict zircon cores and a shell resembling the overwhelming majority of other zircons. It is conceivable that the spread of data points is induced by certain amounts of inheritance. Treatment with cold 4% HF for 10 min in an ultrasonic bath produced a zircon fraction that is con-

Table 1. U – Pb isotopic data

Sample	Analysis Code	Fraction[a]	Concentration			Atom Ratios					
			µg/g		measured	Blank-corrected			Corrected for blank and common lead		
			^{238}U	^{206}Pb radiogenic	$\frac{^{206}Pb}{^{204}Pb}$	$\frac{^{206}Pb}{^{204}Pb}$	$\frac{^{207}Pb}{^{206}Pb}$	$\frac{^{208}Pb}{^{206}Pb}$	$\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{207}Pb}{^{206}Pb}$ age, Ga
Pellinki mafic-intermediate volcanics[b]											
A616b-Västaholm, meta-andesite											
A616b	D	$d>4.2$; HF	334.0	90.62	8,043	10,722	0.11617	0.08606	0.3136±21	4.968±33	1.878± 4
Intrusive gabbros											
A380-Ylivieska											
A380	A	total; borax	536.1	155.93	1,619	2,072	0.12174	0.15360	0.3361±24	5.338±46	1.883± 8
A585-Hyvinkää											
A585a	A	total	725.6	205.75	4,411	4,925	0.11728	0.13070	0.3277±17	5.175±28	1.872± 4
A585b	B	$d>4.2$; HFcr	361.3	104.48	24,231	57,057	0.11521	0.12738	0.3343±17	5.299±28	1.879± 3
A585b	C	$d>4.2$; HF	386.9	110.54	10,237	11,521	0.11618	0.12047	0.3303±17	5.237±28	1.880± 3
A585b	D	$d>4.2$; HF repeat	360.4	103.9	21,598	26,770	0.11557	0.12489	0.3333±17	5.288±28	1.881± 3
A585b	E	$3.8<d<4.2$; HF	1,205.5	314.61	19,047	21,629	0.11547	0.11260	0.3016±16	4.776±25	1.877± 3
A585b	F	$3.8<d<4.2$	1,316.6	343.13	5,255	5,538	0.11654	0.10831	0.3012±19	4.738±30	1.865± 3
A684-Varissaari											
A684	A	$d>4.2$	425.4	121.13	2,232	2,352	0.12090	0.09074	0.3291±17	5.225±35	1.882± 9
A684	B	$4.0<d<4.2$	985.6	282.74	22,950	35,379	0.11521	0.08362	0.3316±17	5.249±28	1.877± 3
A684	C	$3.8<d<4.0$	1,666	448.76	19,780	25,483	0.11473	0.09072	0.3113±16	4.902±26	1.867± 3
A684	D	titanite	83.4	23.72	571	597	0.13690	0.78822	0.3288±17	5.178±50	1.868±15
A684	E	$4.3<d<4.6$; HF	307.2	88.55	26,517	48,307	0.11533	0.07240	0.3331±23	5.284±38	1.880± 4
A684	F	$4.2<d<4.3$; HF	588.3	167.18	37,029	61,811	0.11520	0.07728	0.3285±21	5.207±35	1.879± 6
Kalanti district											
A493-Lepäinen, trondhjemite											
A493a	A	$d>4.1$	825.6	195.43	1,420	1,819	0.12466	0.07032	0.2736±17	4.423±36	1.914± 9
A493a	B	$d<4.1$; $\varnothing>70$	1,115	237.40	1,351	2,373	0.12100	0.05772	0.2460±16	3.910±49	1.884±16
A493a	C	$d<4.1$; $\varnothing<70$	1,083	216.03	1,652	1,993	0.11956	0.05564	0.2305±15	3.583±42	1.844±18
A493a	D	monazite; $\varnothing>160$	2,544	723.79	6,295	–	0.11282	382.93	0.3288±21	5.114±43	1.845±10
A493a	E	$d>4.1$	719.3	176.96	1,592	2.726	0.12250	0.07507	0.2844±17	4.608±47	1.919±13
A493b	F	$d>4.2$	688.5	175.02	1,312	1,848	0.12506	0.06097	0.2938±18	4.771±46	1.922±12
A493b	G	$d>4.1$	1,038	236.72	986	1,165	0.12600	0.06506	0.2636±16	4.159±34	1.870± 9
A493b	H	$4.1<d<4.2$	939.4	227.59	1,280	1,564	0.12350	0.05749	0.2800±18	4.434±36	1.878± 9
A493b	I	$3.8<d<4.1$	1,215	262.13	1,181	1,784	0.12063	0.05283	0.2493±16	3.886±44	1.849±15
A493b	J	$d>4.1$; HF	717.5	214.31	2,333	2,360	0.12162	0.06306	0.3452±18	5.516±30	1.894± 4
A553-Arvassalo, trondhjemite											
A553a	A	$3.8<d<4.2$	1,642	286.13	431.7	450.1	0.14134	0.12325	0.2014±12	3.084±25	1.816±10
A553a	B	$3.8<d<4.2$	1,510	294.55	713.5	717.2	0.13085	0.09266	0.2255±12	3.480±19	1.831± 4
A553a	C	$3.8<d<4.2$; HFcr	1,239	313.97	4,588	4,736	0.11738	0.05876	0.2928±15	4.623±24	1.872± 3
A553b	D	$4.0<d<4.2$; HF	1,102	266.00	2,337	2,379	0.12288	0.09110	0.2789±15	4.507±24	1.914± 3
A553b	E	$3.8<d<4.0$; HF	1,391	279.58	1,557	1,574	0.12125	0.06718	0.2324±12	3.609±20	1.842± 4
A553c	F	$d>4.2$; $\varnothing>70$	692.8	176.37	1,470	1,542	0.13015	0.17865	0.2942±29	4.927±50	1.977± 4
A553c	G	$3.8<d<4.0$; $\varnothing>70$; HF	1,445	279.97	1,489	1,504	0.12124	0.06795	0.2240±17	3.465±27	1.835± 5
A553a	H	monazite; $d>4.2$	15,975	4,539	61,191	84,009	0.11249	58.424	0.3284±41	5.086±65	1.837± 6
A553b	I	$4.0<d<4.2$; $\varnothing>160$	1,114	250.13	1,074	1,093	0.13079	0.08136	0.2595±19	4.239±38	1.933±10
A553b	J	$4.0<d<4.2$; $\varnothing<160$	1,137	262.04	1,566	1,587	0.12576	0.07275	0.2663±18	4.305±33	1.915± 8
A553b	K	$3.6<d<3.8$	1,925	249.07	405.6	407.0	0.14187	0.12714	0.1495±10	2.235±20	1.772±11
A552-Kaleva, diorite/tonalite											
A552	A	$d>4.6$	305.6	86.58	6,118	6,468	0.11622	0.10350	0.3275±17	5.153±27	1.866± 3
A552	B	$d>4.6$; HFcr	279.1	80.82	25,154	34,303	0.11491	0.10727	0.3347±18	5.285±28	1.872± 3
A552	C	$4.3<d<4.6$; HFcr	392.5	113.15	27,721	34,862	0.11492	0.11065	0.3332±17	5.261±28	1.872± 3
A552	D	$4.3<d<4.6$; HF	421.7	120.66	8,984	9,596	0.11570	0.10809	0.3307±17	5.212±28	1.868± 2
A552	E	$4.2<d<4.3$; HF	584.5	163.97	3,855	3,960	0.11747	0.11293	0.3242±17	5.098±27	1.865± 3

Table 1 (continued)

Sample	Analysis Code	Fraction[a]	Concentration µg/g ^{238}U	Concentration µg/g ^{206}Pb radiogenic	Atom Ratios, measured $^{206}Pb/^{204}Pb$	Blank-corrected $^{206}Pb/^{204}Pb$	Blank-corrected $^{207}Pb/^{206}Pb$	Blank-corrected $^{208}Pb/^{206}Pb$	Corrected for blank and common lead $^{206}Pb/^{238}U$	Corrected for blank and common lead $^{207}Pb/^{235}U$	Corrected for blank and common lead $^{207}Pb/^{206}Pb$ age, Ga
A568-Tynki; a = pegmatoid, b = normal gabbro											
A568a	A	$d>4.6$; HF	354.5	96.79	4,440	5,136	0.11379	0.03889	0.3156 ± 16	4.836 ± 26	1.818 ± 4
A568a	B	$4.3<d<4.6$; HF	621.1	164.28	3,597	3,933	0.11316	0.08535	0.3057 ± 16	4.624 ± 25	1.794 ± 3
A568a	C	$4.3<d<4.6$	831.2	197.17	790	798	0.12514	0.13273	0.2742 ± 14	4.084 ± 23	1.766 ± 5
A568a	D	$4.0<d<4.2$; HFcr	864.0	223.21	4,300	4,724	0.11205	0.10394	0.2986 ± 16	4.494 ± 25	1.785 ± 4
A568a	E	$4.0<d<4.3$; HF	420.7	99.98	8,394	17,346	0.11231	0.02843	0.2747 ± 14	4.224 ± 24	1.824 ± 4
A568a	F	$4.0<d<4.2$; $\varnothing<50$	1,586	330.10	616.7	620.5	0.12803	0.17707	0.2405 ± 13	3.515 ± 43	1.731 ± 21
A568a	G	titanite; $\varnothing>160$	66.24	18.62	495.7	504.1	0.13814	0.10537	0.3249 ± 61	4.980 ± 95	1.818 ± 5
A568a	H	$4.3<d<4.6$	882.7	200.52	712.3	722.0	0.12671	0.17864	0.2625 ± 16	3.902 ± 29	1.762 ± 9
A568b	I	$d>4.3$; $\varnothing<160$	416.7	114.76	8,594	10,442	0.11584	0.10377	0.3183 ± 21	5.027 ± 36	1.872 ± 6
A568b	J	$d>4.3$; $\varnothing<160$; HF	409.4	115.44	7,797	9,261	0.11610	0.10443	0.3259 ± 28	5.150 ± 45	1.874 ± 4
A568b	K	$4.2<d<4.3$; $\varnothing<160$; HF	567.7	160.50	25,362	56,285	0.11474	0.11400	0.3268 ± 20	5.159 ± 32	1.872 ± 5
A568b	L	$4.0<d<4.2$; HF	732.1	203.01	18,179	26,662	0.11505	0.12749	0.3205 ± 21	5.061 ± 34	1.872 ± 4
A955-Hiukalliot, trondhjemite											
A955	A	$d>4.2$	715.3	181.96	4,344	4,581	0.11733	0.09062	0.2940 ± 30	4.637 ± 49	1.870 ± 7
A955	B	$4.0<d<4.2$	1,045	256.36	3,359	3,424	0.11756	0.04057	0.2836 ± 20	4.442 ± 34	1.857 ± 6
A955	C	$3.6<d<4.0$	1,425	316.77	1,579	1,604	0.11805	0.04132	0.2570 ± 23	3.880 ± 11	1.792 ± 44
A955	D	monazite	2,705	760.43	33,563	41,917	0.11295	345.82	0.3250 ± 30	5.046 ± 48	1.842 ± 6
A554-Uusikaupunki, diorite/tonalite											
A554	A	$d>4.1$; $\varnothing>160$; dark	996.2	250.98	851.5	934.6	0.12569	0.13959	0.2912 ± 18	4.460 ± 36	1.817 ± 10
A554	B	$d>4.1$; $\varnothing>160$; light	512.4	139.49	2,909	8,751	0.11607	0.13148	0.3146 ± 19	4.968 ± 47	1.872 ± 12
A554	D	$d>4.6$; $\varnothing>160$; dark	614.2	161.15	3,236	3,328	0.11820	0.12394	0.3032 ± 16	4.772 ± 26	1.866 ± 3
A554	E	$d>4.6$; $\varnothing>160$; light	455.7	121.14	2,357	2,394	0.11941	0.14028	0.3072 ± 16	4.818 ± 26	1.860 ± 2
A554	F	$4.2<d<4.6$; $\varnothing>160$; dark	860.5	237.55	3,955	4,065	0.11732	0.13460	0.3191 ± 17	5.014 ± 27	1.864 ± 3
A554	G	$d>4.6$; $70<\varnothing<160$; HF	460.2	128.09	11,284	12,166	0.11552	0.11531	0.3217 ± 17	5.074 ± 27	1.870 ± 2
A554	H	$d>4.6$; $70<\varnothing<160$; HFcr	430.1	119.61	41,151	57,127	0.11454	0.11003	0.3214 ± 18	5.065 ± 29	1.869 ± 3
A174-Putsaari, trondhjemite											
A174a	A	$d>4.3$; HF	524.7	144.01	5,771	6,465	0.11612	0.05288	0.3172 ± 17	4.987 ± 27	1.864 ± 3
A174a	B	$4.2<d<4.3$; HF	665.1	179.75	6,292	6,682	0.11549	0.04255	0.3124 ± 17	4.887 ± 27	1.855 ± 3
A174a	C	$4.0<d<4.2$; $\varnothing>70$; HF	1,101	281.53	2,552	2,595	0.11727	0.04960	0.2956 ± 15	4.566 ± 24	1.832 ± 2
A174a	D	$4.0<d<4.2$; $\varnothing>70$	1,191	296.98	1,196	1,204	0.12271	0.07660	0.2883 ± 15	4.430 ± 25	1.823 ± 6
A174b	F	titanite; $\varnothing>160$	166.7	46.96	1,085	1,092	0.12315	0.16525	0.3256 ± 35	4.970 ± 11	1.810 ± 35
Syn-tectonic granitoids											
A13-Aulanko, granodiorite											
A13	A	$d>4.6$	304.0	82.37	4,083	4,235	0.11734	0.09276	0.3132 ± 16	4.928 ± 26	1.866 ± 3
A13	B	$d>4.6$; HFcr	261.1	75.31	9,938	11,081	0.11614	0.08420	0.3333 ± 17	5.281 ± 28	1.878 ± 3
A13	C	$4.2<d<4.6$	556.0	128.79	1,220	1,232	0.12297	0.10662	0.2677 ± 14	4.133 ± 22	1.831 ± 4
A13	D	$4.2<d<4.6$; HFcr	403.5	107.33	5,826	6,219	0.11693	0.07806	0.3075 ± 16	4.865 ± 26	1.876 ± 3
A4-Vastamäki, Hämeenkyrö granodiorite											
A4	A	total; borax[c]	406.8	112.45	3,370	7,189	0.11514	0.08800	0.3195 ± 28	4.988 ± 75	1.852 ± 23
A4	B	$d>4.2$	420.9	109.76	4,736	5,364	0.11639	0.09303	0.3014 ± 16	4.731 ± 25	1.861 ± 3
A4	C	$d>4.6$	302.4	84.08	4,949	6,187	0.11718	0.08905	0.3213 ± 18	5.094 ± 30	1.879 ± 4
A4	D	$4.2<d<4.6$	502.9	127.42	3,272	3,627	0.11717	0.09378	0.2928 ± 16	4.580 ± 26	1.855 ± 5
A248-Äkönmaa, Hämeenkyrö granodiorite											
A248	A	$4.2<d<4.6$	435.2	112.29	2,202	2,251	0.11997	0.09591	0.2982 ± 15	4.685 ± 28	1.863 ± 6
A248	B	$4.0<d<4.2$	698.9	161.53	1,661	1,676	0.12090	0.09755	0.2671 ± 14	4.155 ± 25	1.845 ± 6
A248	C	titanite; $3,5<d<3.6$	41.8	12.19	596	632	0.13650	0.13641	0.3372 ± 19	5.351 ± 46	1.881 ± 11
A382-Voinsalmi, hypersthene granite											
A382	A	$\varnothing>110$; borax[d]	356.2	100.57	10,619	–	0.11460	0.09047	0.3264 ± 20	5.155 ± 36	1.873 ± 6
A382	B	$d>4.6$	221.6	64.27	23,447	36,242	0.11605	0.09004	0.3352 ± 18	5.345 ± 30	1.890 ± 3
A382	C	$4.2<d<4.6$	832.6	230.93	10,286	11,345	0.11569	0.09888	0.3206 ± 19	5.061 ± 31	1.872 ± 3
A382	D	$4.0<d<4.2$	1,958	502.24	5,374	5,497	0.11517	0.10361	0.2964 ± 18	4.606 ± 29	1.843 ± 3
A382	E	$d>4.6$; $\varnothing>160$	224.1	64.90	22,135	58,801	0.11563	0.08921	0.3347 ± 18	5.325 ± 29	1.886 ± 3

Table 1 (continued)

Sample	Analysis Code	Fraction[a]	Concentration µg/g ^{238}U	µg/g ^{206}Pb radiogenic	measured $\frac{^{206}Pb}{^{204}Pb}$	Atom Ratios, Blank-corrected $\frac{^{206}Pb}{^{204}Pb}$	$\frac{^{207}Pb}{^{206}Pb}$	$\frac{^{208}Pb}{^{206}Pb}$	Corrected for blank and common lead $\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{207}Pb}{^{206}Pb}$ age, Ga
A383-Voinsalmi, hypersthene quartz diorite											
A383	A	total; borax[d]	957.3	260.3	5,704	10,427	0.11569	0.07054	0.3143±21	4.956±39	1.870± 8
A383	B	$d>4.3$	417.2	118.80	30,590	–	0.11527	0.09681	0.3291±20	5.226±32	1.882± 3
A383	C	$4.2<d<4.3$	730.7	187.05	12,085	14,533	0.11577	0.08129	0.2959±16	4.684±26	1.877± 4
A383	D	$4.2<d<4.3$; ∅>70; abr	744.1	208.92	14,357	15,350	0.11552	0.08383	0.3245±17	5.129±27	1.874± 3
A952a-Koppelojärvi, granite											
A952a	A	$d>4.6$	282.0	71.77	2,275	2,315	0.12004	0.11558	0.2941±18	4.631±34	1.867± 8
A952a	B	$4.2<d<4.6$	419.4	98.13	1,409	1,426	0.12221	0.11292	0.2705±15	4.203±28	1.843± 7
A952a	C	$d>4.6$; $70<d<160$; HFcr	216.8	58.75	6,808	7,848	0.11639	0.11025	0.3131±18	4.950±29	1.874± 3
A952a	D	$4.2<d<4.6$; HFcr	311.8	86.25	6,747	7,214	0.11602	0.10062	0.3197±17	5.031±27	1.866± 4
Post-tectonic granitoids											
A334-Åva, monzonite											
A334	A	total; borax[e]	216.7	55.31	390.9	423.2	0.14158	0.21911	0.2949±21	4.448±44	1.789±11
A334	B	titanite; borax	104.6	27.00	173.3	178.0	0.18364	0.65463	0.2983±21	4.403±57	1.750±19
A334	C	$d>4.6$	177.2	44.47	1,808	2,014	0.11581	0.16253	0.2900±15	4.360±28	1.783± 8
A334	D	$4.2<d<4.6$	295.1	71.40	2,141	2,392	0.11437	0.13231	0.2796±15	4.189±25	1.777± 6
A334	E	$d>4.6$; HF	171.6	44.32	3,473	3,951	0.11254	0.15645	0.2986±29	4.491±44	1.784± 5
A335-Åva, granite											
A335	A	total[e]	428.5	88.78	568.1	613.7	0.12955	0.21072	0.2395±18	3.543±35	1.754±13
A335	B	titanite	103.0	27.75	210.8	211.9	0.17149	0.64537	0.3114±31	4.595±60	1.749±16
A335	C	$d>4.6$	323.7	71.94	1,978	2,189	0.11461	0.17399	0.2568±14	3.838±22	1.772± 4
A335	D	$4.2<d<4.6$	880.7	159.41	1,050	1,080	0.11868	0.18012	0.2092±11	3.058±19	1.732± 7
A335	E	$d>4.6$; HF	280.1	66.84	4,093	4,512	0.11178	0.16554	0.2758±23	4.136±36	1.778± 4
A826-Parkkila, granodiorite[f]											
A826	A	$d>4.6$; ∅>160[f]	348.5	90.89	1,620	1,714	0.11759	0.11692	0.3014±16	4.559±25	1.794± 4
A826	B	$4.2<d<4.6$[f]	509.0	128.07	2,272	2,395	0.11516	0.13740	0.2908±16	4.390±26	1.791± 5
A827-Parkkila, granodiorite[f]											
A827	A	$d>4.6$[f]	290.2	77.06	2,293	2,522	0.11494	0.15832	0.3069±18	4.636±29	1.791± 5
A827	B	$4.2<d<4.6$[f]	407.0	100.23	4,510	5,326	0.11213	0.18546	0.2846±15	4.300±24	1.792± 4

[a] All fractions are zircon unless otherwise indicated. d=density g cm^{-3}; ∅=size in microns; HF=grains HF-leached; HFcr=crushed material HF-leached; abr=grains abraded; borax=old analyses, borax-fusion method
[b] For other Pellinki data see Laitala 1984
[c] From Kouvo and Tilton 1966
[d] Recalculated from Gaál and Rauhamäki 1971
[e] Recalculated from Vaasjoki 1977
[f] Recalculated from Simonen 1982

cordant just above 1.89 Ga (Fig. 2; the analysis J of Table 1).

It has been found in other studies that if further divided, the density fractions define separate chords of their own, the heavy fractions having the highest intercept at concordia. In the case of sample A553-Arvassalo (Fig. 2) density fractions 3.8–4.2 g cm^{-3} lie on a well-defined chord having upper and lower intercepts corresponding to ages of 1.890 ± 0.007 Ga and 0.202 ± 0.026 Ga, respectively. All five fractions are clear long needles with L/B ratios up to 7. The two fractions 3.8–4.0/+70 µm and 3.6–3.8 g cm^{-3} include about 1% of short prisms. The coarse fraction 4.0–4.2/+160 µm consists of clear long prisms (L/B higher than 5), and some platy crystals can be found. The offset of the fraction +4.2/+70 µm appears to be a real effect, because the zircon is morphologically different; the crystals are clear and short rather than simple tetragonal long prisms like other fractions. Visible cores were not found. The fractions analyzed from the third trondhjemite sample, A955-Hiukalliot, behave like the samples from Lepäinen and Arvassalo.

Zircons from diorite/tonalite (A522-Kaleva and A554-Uusikaupunki) show recrystallization effects. In sample A552 aggregate crystals occur. This is also the case for

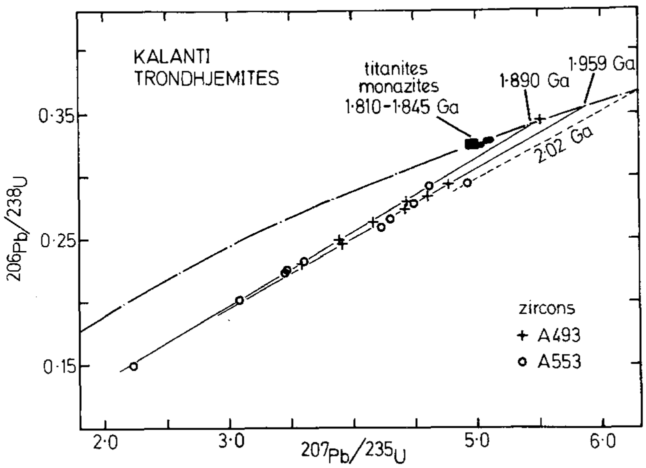

Fig. 2. U – Pb age data for Kalanti trondhjemites

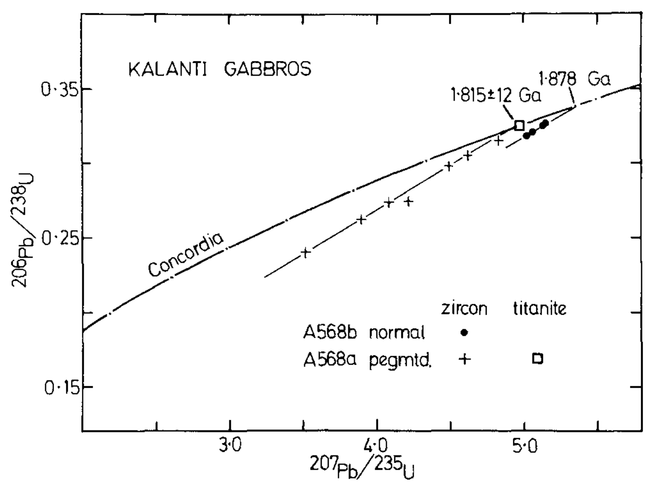

Fig. 3. U – Pb age data for Kalanti gabbros

zircons separated from sample A554, and here dark-colored shells around light-colored cores are common. However, no age difference between the light and dark zircons was found.

The isotopic data on monazites and titanites from five samples (Fig. 2) yield lower ages (1.80 Ga) which are consistent with each other and probably indicate a time of metamorphism. These ages are in general agreement with results from nearby areas. There is also other compelling evidence for advocating an important thermal event at or about 1.80 Ga. This is illustrated in Fig. 3. Two different zircon isochrons were obtained on samples from the same outcrop: 1.878 Ga for normal gabbro and 1.815 Ga for pegmatoidal variety representing the matrix between large brecciated fragments of normal gabbro. This gabbro at Tynki is called by Hietanen (1943) "schiefriger Gabbro". During the course of this work it was found that the brecciated character is not restricted to the contacts only but the gabbro itself is fragmentary in nature. At least some of the pieces of normal gabbro are round and covered by thin shell of plagioclase. Zircons and titanite were separated both from the normal gabbro and from the pegmatoidal gabbro occurring between the fragments. The almost concordant age of 1.818 Ga for titanite from the pegmatoidal gabbro is shown also in Fig. 3.

A Rb – Sr age on biotite from the Heinänen trondhjemite ca. 3 km SE of Lepäinen (A493) is low, 1.718 Ga (recalculated from Kouvo 1958). This late event recorded by Rb – Sr and K – Ar systems on biotites is manifested in numerous rock units in Finland, e.g. Aulanko granodiorite (A13): Rb – Sr 1.776 ± 0.045 Ga and K – Ar 1.728 ± 0.070 Ga (recalculated from Kouvo 1958) and Vaaraslahti hypersthene granite (A49): Rb – Sr 1.723 Ga and K – Ar 1.733 Ga (Salli 1983; recalculated from Wetherill et al. 1962).

Clearly the upper concordia intercepts listed in Table 2 give unambiguous emplacement ages for the majority of the igneous rocks. The frequency of 1.89–1.87 Ga ages reflects the intensity of the main Svecokarelian igneous events. Even the one sample with only a single zircon analysis, A380-Ylivieska, is so consistent with this age that there can be no uncertainty. We do feel that we are justified in assigning older ages, around 1.90 Ga, to at least some of the Kalanti trondhjemites, and so 1.90 Ga is used for calculated Nd initial isotopic ratios for A493 and A553.

Nd isotopic results

Sm – Nd analyses are listed in Table 3 and initial $^{143}Nd/^{144}Nd$ ratios plotted as $\varepsilon_{Nd}(T)$ in Fig. 4. Fig. 4 shows the CHUR evolution line for bulk Earth ($^{147}Sm/^{144}Nd = 0.1966$), a suggested depleted mantle evolution similar to DePaolo (1981) as well as a generalized band for Archean crustal evolution taken from Patchett and Bridgwater (1984). As in previous studies such as DePaolo (1981) and Patchett and Bridgwater (1984), we assume that the Sm/Nd ratio of single rock samples is undisturbed since rock consolidation, in keeping with the widely-known stability of rare-earth elements with respect to low-temperature events. No major thermal events occurred in Finland after ~1.6 Ga. Our Nd isotopic initials are thus calculated from single well-chosen and dated samples.

The Pellinki volcanics define $\varepsilon_{Nd}(T)$ values from $+0.9$ to $+2.4$. Except for one hornblende-phyric basalt, the $^{147}Sm/^{144}Nd$ ratios are generally low and similar to average continental crust (0.125). The meta-andesites show the higher $\varepsilon_{Nd}(T)$ values. Some of the intrusive gabbros also exhibit crustal-type $^{147}Sm/^{144}Nd$ and have $\varepsilon_{Nd}(T)$ around $+1$ or less. These are mafic pegmatoid phases of smaller gabbro bodies. The two large gabbros Hyvinkää and Ylivieska, however, show $\varepsilon_{Nd}(T)$ of $+2.6$.

The Kalanti diorites/tonalites and the normal gabbro, with 1.88–1.87 Ga ages, all have $\varepsilon_{Nd}(T) = +3.2$, supporting a magmatic relationship between these rocks. The trondhjemites, however, appear to be different, with $\varepsilon_{Nd}(1.90) = +1.7$ and $+2.4$; if a 1.87 Ga age were used, the ε_{Nd} value would be even lower. This Nd isotopic difference supports the U – Pb zircon age evidence in suggesting that the classic trondhjemites and "quartz-drop trondhjemites" are unrelated to and older than the main plutonic rocks of the Kalanti district. If this is so, the trondhjemites cannot have been derived from the gabbros and diorites as proposed by Arth et al. (1978).

Most of the synorogenic and post-orogenic granitoids have $\varepsilon_{Nd}(T)$ close to zero except for A13, the Aulanko hornblende-granodiorite, with $+1.9$. This would seem to be a

Table 2. Summary of U – Pb ages and geologic references

Sample	Geologic Unit/Location	Number of Fractions	Concordia intercepts[a]		Geologic Reference
			Upper	Lower	
Pellinki mafic-intermediate volcanics					
A615, A616, A617		5	1.887±14	0.13± 6	Laitala 1973, 1984
Intrusive gabbros					
A380	Ylivieska	1	1.883[b]		Salli 1961
A585	Hyvinkää	6	1.880± 5	0.13±16	Härme 1954, 1978
A684	Varissari	5	1.885± 5	0.35±15	Simonen 1952
A794	Skåldö, four samples	8	1.885± 7	0.26± 5	Hopgood et al. 1983
A877	Svartgrund	7	1.891±11	0.66± 2	Suominen, in preparation
Kalanti district					
A493	trondhjemite, Lepäinen	9	(1.90)[c]		Hietanen 1943; Arth et al. 1978
A553	trondhjemite, Arvassalo	11	(1.90)[c]		
A174	trondhjemite, Putsaari	4	1.892± 5	0.64± 5	
A552	diorite, Kaleva	5	1.874± 2	0.45±11	
A554	diorite, Uusikaupunki	6	1.872±11	0.16±24	
A568a	gabbro pegmatoid, Tynki	6	1.812±11	0.41±10	
A568b	normal gabbro, Tynki	4	1.878[d]		
Syntectonic granitoids					
A13	Aulanko granodiorite	4	1.886±14	0.35±14	Simonen 1948
A49	Vaaraslahti hy-granite	6	1.884± 5	0.11± 8	Salli 1983
A4, A248	Hämeenkyrö granodiorite	6	1.882± 6	0.27± 7	Simonen 1952
A382, A383	Voinsalmi hy-granodiorite	9	1.887±11	0.34±23	Gaál and Rauhamäki 1971
A880	Mörskär granite	3	1.881± 9	0.27± 5	Suominen, in preparation
A952a	Koppelojärvi granite	4	1.879±14	0.24±17	Laitakari, in preparation
Post-tectonic granitoids					
A334, A335	Åva ring-complex	8	1.797± 4	0.22± 2	Kaitaro 1953; Ehlers and Ehlers 1981
A826, A827	Parkkila granodiorite	4	1.794± 5	0.02±10	Simonen 1982

[a] Errors on concordia intercepts are 2σ, and correspond to last figures shown
[b] Age from a single zircon analysis lying close to concordia; age is $^{207}Pb/^{206}Pb$ age
[c] Preferred interpretation of Kalanti trondhjemite age; see Fig. 2
[d] Points all lie close to concordia; age is computed assuming continuous Pb diffusive loss; see Fig. 3

plutonic rock of rather primitive, mantle-influenced chemistry. Among the remaining granitoids, it may or may not be significant that the samples closest to the Archean continent, A49b and A382 have the most negative $\varepsilon_{Nd}(T)$ of -0.6. The uncertainty on ε_{Nd} of 0.5 ε-units does not allow a definite statement.

Six of the samples in Table 3 have previously been analyzed for $\varepsilon_{Nd}(T)$ using zircons (Patchett et al. 1981). Comparison of the Nd and Hf initials (Table 4) shows that indeed ε_{Hf} exceeds ε_{Nd} by a factor probably around 2–3, as observed in Colorado (Patchett et al. 1981). The samples with the lowest ε_{Nd}, A794 and A382, have also the lowest ε_{Hf}, but the relationship is clearly not simple. Even allowing for uncertainties, the ε_{Hf} of the A335 Åva zircon seems too high, or the whole-rock ε_{Nd} too low (Table 4). When the mixed origin of the granitoids and the contaminated nature of the gabbros (as inferred in a later section) is considered, a regular relationship would not be expected between Nd and Hf isotopes, because of differences in concentrations and behavior of the elements.

Discussion: make-up of the Svecokarelian crust

Figure 4 shows the initial ε_{Nd} values of all the samples, a generalized evolution curve for Archean crustal material, as well as the more specific PAAS (post-Archean average Australian shale) evolution used by Patchett and Bridgwater (1984) to model typical sedimentary material derived from Archean continents. The Sm/Nd of PAAS was defined by Nance and Taylor (1976).

The first general point to be taken from the Svecokarelian data is that unlike Colorado (DePaolo 1981) and South Greenland (Patchett and Bridgwater 1984) where depleted mantle ε_{Nd} values of $+4$ could be identified in tholeiites, the results of this study only show $\varepsilon_{Nd}(T)$ values up to $+3.2$. $\varepsilon_{Nd}(T)$ greater than $+2$ is obtained from Kalanti

Table 3. Sm – Nd isotopic data

Sample	Rock Type/ Locality	Age (T)	Sm ppm	Nd ppm	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ [a]	$(^{143}Nd/^{144}Nd)_0$	$\varepsilon_{Nd}(T)$ [b]
Pellinki mafic-intermediate volcanics								
A615a	meta-andesite	1.89	5.21	25.42	0.1239	0.511820 ± 13	0.510279	+1.7
A615e	metabasalt	1.89	3.30	15.93	0.1251	0.511833 ± 17	0.510277	+1.7
A615f	metabasalt	1.89	1.77	6.90	0.1554	0.512170 ± 13	0.510237	+0.9
A616	meta-andesite	1.89	5.17	25.08	0.1246	0.511855 ± 19	0.510305	+2.2
A617	meta-andesite	1.89	5.21	25.58	0.1231	0.511848 ± 19	0.510317	+2.4
Intrusive gabbros								
A380	Ylivieska	1.88	2.05	6.74	0.1844	0.512617 ± 17	0.510336	+2.6
A584	Hyvinkää	1.88	2.65	11.19	0.1430	0.512113 ± 15	0.510353	+2.6
A684	Varissaari	1.88	11.78	56.21	0.1267	0.511836 ± 21	0.510260	+1.3
A794	Skåldö	1.88	3.39	12.72	0.1610	0.512200 ± 18	0.510208	+0.1
A877	Svartgrund	1.89	7.30	37.29	0.1183	0.511680 ± 13	0.510201	+0.4
Kalanti district								
A493b	trondhjemite	1.90	1.08	8.57	0.0758	0.511253 ± 18	0.510305	+2.4
A553	trondhjemite	1.90	0.99	4.78	0.1247	0.511825 ± 13	0.510266	+1.7
A552	diorite	1.87	9.22	50.77	0.1097	0.511726 ± 15	0.510376	+3.1
103/1939	diorite	1.87	9.59	50.56	0.1147	0.511795 ± 20	0.510384	+3.2
A568b	gabbro	1.88	2.98	14.04	0.1285	0.511955 ± 15	0.510365	+3.1
Syn-tectonic granitoids								
A13	Aulanko	1.89	2.18	10.40	0.1267	0.511868 ± 18	0.510292	+1.9
A49b	Vaaraslahti	1.88	4.75	27.79	0.1033	0.511455 ± 18	0.510177	−0.6
A248	Hämeenkyrö	1.88	4.69	22.23	0.1276	0.511791 ± 15	0.510212	+0.1
A382	Voinsalmi	1.89	7.83	36.01	0.1314	0.511796 ± 14	0.510162	−0.6
A880	Mörskär	1.88	6.61	41.19	0.0969	0.511416 ± 13	0.510217	+0.2
A952b	Koppelojärvi	1.88	9.63	50.61	0.1150	0.511608 ± 10	0.510185	−0.4
Post-tectonic granitoids								
A335	Åva	1.80	26.18	195.44	0.0810	0.511277 ± 14	0.510318	+0.2
A827	Parkkila	1.79	22.98	180.89	0.0768	0.511245 ± 15	0.510341	+0.3

[a] Errors on $^{143}Nd/^{144}Nd$ are $2\sigma_m$

[b] ε_{Nd} uncertainty is 0.5 ε-units

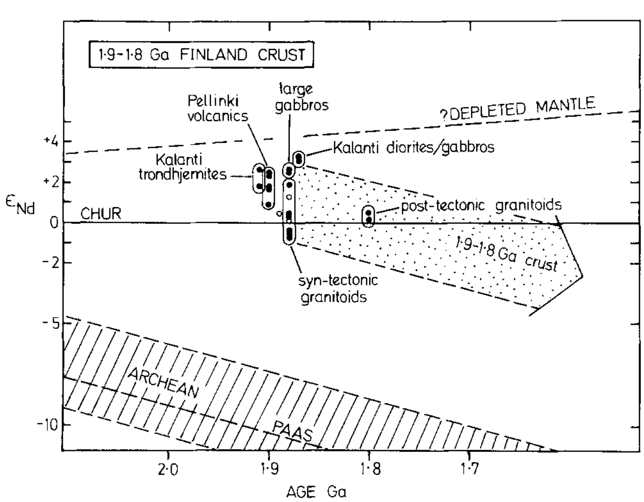

Fig. 4. $\varepsilon_{Nd}(T)$ plotted against age known from U – Pb zircon dating. CHUR = evolution of undifferentiated Earth. PAAS = post-Archean average Australian Shale, taken as typical Archean crust available as sediments 1.9–1.7 Ga ago. Open symbols are small gabbro intrusions considered to be crustally-contaminated compared to the large gabbros. For clarity, Pellinki volcanics and Kalanti trondhjemites are plotted at 0.01 Ga older than their true age

Table 4. Comparison of $\varepsilon_{Nd}(T)$ for total rock and $\varepsilon_{Hf}(T)$ for zircon

Sample	Unit	$\varepsilon_{Nd}(+0.5)$	$\varepsilon_{Hf}(\pm 0.8)$ [a]
A684	Varissaari gabbro	+1.3	+3.0
A794	Skåldö gabbro	+0.1	+2.1
A877	Svartgrund gabbro	+0.4	+3.1
A382	Voinsalmi granitoid	−0.6	+1.6
A335	Åva granite	+0.2	+3.9
A827	Parkkila granitoid	+0.3	+1.2

[a] Data from Patchett et al. (1981)

trondhjemites (1.90 Ga), from Pellinki andesites (1.89 Ga) and from Kalanti diorites/tonalites (1.87 Ga). There is thus clear evidence that depleted mantle material was repeatedly involved in magma genesis in the Finnish Svecokarelian, but either the mantle was less depleted than in Colorado/Greenland, or the $\varepsilon_{Nd} = +4$ to $+5$ was always reduced to some extent by contamination with material of Archean crustal origin (Fig. 4).

The results for Pellinki volcanics and the intrusive gabbros with $\varepsilon_{Nd}(T)$ less than $+2$ support a mixing hypothesis.

At Pellinki the volcanics overlie an extensive pile of sediments (Laitala 1973), and while the andesites with 25 ppm Nd show $\varepsilon_{Nd}(T) = +1.7$ to $+2.4$, and plagioclase-phyric basalt with 16 ppm Nd shows $+1.7$, it is the basalt with the lowest Nd = 7 ppm which has the lowest $\varepsilon_{Nd}(T) = +0.9$ (Table 3). Likewise for the gabbros, it is the very large intrusions Ylivieska and Hyvinkää which show $\varepsilon_{Nd}(T) = +2.6$, while the mafic pegmatoids of three smaller bodies have $^{147}Sm/^{144}Nd$ similar to continental crust and lower $\varepsilon_{Nd}(T)$ in the range 0 to $+1.3$.

Our preferred interpretation of the Finnish data is that indeed depleted mantle with $\varepsilon_{Nd} = +4$ to $+5$ was available, but that none of the sampled rocks completely escaped mixing effects with material of Archean crustal origin. The samples with $\varepsilon_{Nd}(T)$ greater than $+2$ are merely those where this effect is minimal. At this point we note that Patchett et al. (1981), on the basis of an initial ε_{Hf} study of several Finnish Svecokarelian granitoids and the smaller gabbro intrusions, concluded that the mantle giving rise to the 1.9–1.8 Ga crust was characterized by $\varepsilon_{Hf} = +3$, corresponding to $\varepsilon_{Nd} = +1$. This light degree of old mantle depletion was in contrast to Colorado, with $\varepsilon_{Nd} = +4$ (DePaolo 1981) and $\varepsilon_{Hf} = +10$. Patchett et al. (1981) argued that the gabbros must have been mantle-derived, and the fact that four of these from widely-separated locations gave $\varepsilon_{Hf} = +3$, and in general agreed with granitoids, then the results could not be due to mixing, which could not have generated such uniformity. Therefore a rather undepleted mantle was invoked, and implied gross global heterogeneity of the "depleted mantle" 1.9–1.7 Ga ago. Although there is still no reason why the latter point should not be true, the detailed interpretation in Finland advocated by Patchett et al. (1981) is made impossible by the new results. Volcanics, gabbros and primitive plutonics with $\varepsilon_{Nd}(T)$ of $+2$ to $+3$, corresponding to $\varepsilon_{Hf}(T) = +5$ to $+9$ make clear that the smaller gabbro intrusions with $\varepsilon_{Hf}(T) = +3$, upon which so much significance was placed earlier, have most probably been crustally contaminated. There is further no reason why the larger gabbros should not have been contaminated to some extent, as would be the case for the Pellinki andesites which are most probably of island-arc affinity where such effects are common (McCulloch and Perfit 1981; White and Patchett 1984). The Kalanti trondhjemites and diorites/tonalites are plutonic rocks typical of crustal batholiths, although of primitive type, and could also easily have a mixed origin. Indeed the range of $\varepsilon_{Nd}(T)$ for all rock types and groups in Fig. 4 strongly suggests mixing and contamination processes. A small inherited zircon component in the Kalanti trondhjemites would support such mixing, and is consistent with the reduced ε_{Nd} in those rocks compared to the Kalanti diorites and tonalites. Therefore we now revoke the interpretation of Patchett et al. (1981) which is negated by the new results, and assume instead that depleted mantle of the conventional type with $\varepsilon_{Nd} = +4$ to $+5$ was present beneath Finland during Svecokarelian crustal genesis.

The granitoids of this study range between two types: (1) rather primitive plutonics, the Kalanti trondjemites and diorites/tonalites with $\varepsilon_{Nd}(T) = +2$ to $+3$. The Aulanko granodiorite is a compositionally primitive type with low K content (Simonen 1948), only 10 ppm Nd, and $\varepsilon_{Nd}(T) = +1.9$ (Table 2), and seems to belong to this group; (2) normal synorogenic granitoids with higher K content and Nd = 20 to 50 ppm, which have $\varepsilon_{Nd}(T)$ values close to zero. Although the Kalanti diorites/tonalites also have 50 ppm Nd, and may suggest some geographical differences, it *is* clear that there is a range of ε_{Nd} between $+3$ and -0.5, and that this is correlated with other chemical parameters such as K content. The post-tectonic granitoids have almost 200 ppm Nd, and $\varepsilon_{Nd}(T) = 0$ to $+0.5$, and may or may not be derived from ~1.88 Ga crust by remelting (Fig. 4).

The T_{DM} model ages of the granitoids naturally reflect the compositional trend as well. The Kalanti samples have T_{DM} ages less than 2.0 Ga, while the synorogenic more-evolved granitoids give 2.0 to 2.1 Ga. This general T_{DM} age distribution is clear from Fig. 4.

In a geochemically extremely similar situation in South Greenland, Patchett and Bridgwater (1984) interpret $\varepsilon_{Nd}(T)$ values grouped around zero as due to mixing between a magmatic component of $\varepsilon_{Nd} = +4$ to $+5$ depleted-mantle origin and a component derived from Archean crust with $\varepsilon_{Nd} = -9$ to -13. The samples in South Greenland, and indeed in Finland, were deliberately chosen far from any actual Archean crust, and the Archean mixed component appeared there, as here, to be present irrespective of location in the terrain. This led to the inference that the Archean crustal component was delivered in the form of sediments lying on oceanic crust, which were mixed tectonically and by melting into the terrain, in a reasonably uniform way so long as subduction continued. This interpretation is in line with plate-tectonic models for the origin of the Finnish Svecokarelian (Hietanen 1975; Gaál 1982). Huhma (1985) presents Nd isotopic evidence for the origin of Svecokarelian terrain flysch sediments. His T_{DM} model ages (for example, 2.2 Ga for schist central to the Finnish part of the Svecokarelian) show only a slightly higher component (10–20%) of Archean crust than our granitoids, and are therefore also a mixture of a small proportion of Archean with new Proterozoic materials (Huhma 1985). Patchett and Bridgwater (1984) emphasized that probably no observed rocks in South Greenland were first-generation inputs to the orogenic zone, and this may well be the case for components of the Svecokarelian terrain as well.

The crustal component can be taken as PAAS with 32 ppm Nd, which agrees with other sedimentary averages and with Canadian Shield crustal composite samples (Patchett and Bridgwater 1984). The sedimentary ε_{Nd} was -11 at 1.85 Ga, while the depleted mantle was $+4$. Taking the mantle-derived component as an island-arc one leads to a range of possible Nd contents 5–20 ppm, though this must be extended to 25 ppm if the Pellinki andesites are considered. A reasonable bulk Nd content for present-day island arcs is 10 ppm (White and Patchett 1984). A mass balance for the $\varepsilon_{Nd} \sim 0$ granitoids performed using this value leads, as in South Greenland, to a 10% Archean component being needed to produce the Svecokarelian granitoids. Thus the igneous part of the 1.9–1.8 Ga crust of Finland, at least in the areas far from Archean cratons, consists of around 90% (actually 85–95%) new, mantle-derived material, and represents a substantial mantle-to-crust differentiation between 1.9 and 1.8 Ga ago.

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Appendix. Location of samples

No.	Location	Map Sheet	Grid Coordinates
A4	Vastamäki, Ylöjärvi	2124 04 Viljakkala-Teisko	6834.47–472.52
A13	Louhimokatu, Hämeenlinna	2131 09 Hämeenlinna	6764.65–524.60
A49	Mäkikylä, Vaaraslahti	3314 09 Pielavesi	7030.85–486.36
A174	Putsaari, Uusikaupunki	1131 01 Uusikaupunki	6742.45–508.74
A248	Riihioja, Äkönmaa	2124 01 Viljakkala-Teisko	6834.96–469.16
A334	Getören, Brändö	1041 03 Iniö	6758.0 –502.7
A335	Åva, Brändö	1041 03 Iniö	6706.4 –503.3
A380	Someronperä, Ylivieska	2431 07 Ylivieska	7108.62–521.75
A382	Voinsalmi, Rantasalmi	3234 07 Varkaus	6896.70–561.95
A383	Voinsalmi, Rantasalmi	3234 07 Varkaus	6894.85–563.55
A493	Lepäinen, Uusikaupunki	1131 04 Uusikaupunki	6743.77–514.95
A552	Kaleva, Kalanti	1131 07 Uusikaupunki	6745.93–526.73
A553	Arvassalo, Uusikaupunki	1042 09 Vehmaa	6739.36–524.80
A554	Ruokola, Uusikaupunki	1131 07 Uusikaupunki	6744.05–524.41
A568	Tynki, Kalanti	1131 07 Uusikaupunki	6747.77–525.79
A584	Hyvinkää	2044 02 Riihimäki	6626.07–543.38
A585	Hirvijärvi, Riihimäki	2042 11 Karkkila	6629.47–533.76
A615	Långhällen	3012 09 Porvoo	6678.64–444.42
A616	Västaholm	3012 09 Porvoo	6678.16–444.65
A617	Trutklobben	3012 09 Porvoo	6677.70–443.70
A684	Varissaari, Kohmalanlahti, Ylöjärvi	2124 07 Viljakkala-Teisko	6833.88–483.22
A794	Norrsida, Skåldö	2013 03 Jussarö	6644.52–469.72
A826	Parkkila	3142 07 Mikkeli	6835.84–523.70
A827	Parkkila	3142 07 Mikkeli	6835.04–524.06
A877	Svartgrund, Åland	1011 09 Lågskär	6645.02–446.41
A880	Käringen, Mörskär	1031 01 Utö	6628.67–408.61
A952a	Koppelojärvi, Juupajoki	2142 09 Orivesi	6858.50–525.80
A952b	Lylyjärvi, Juupajoki	2231 07 Mänttä	6864.00–526.70
A955	Hiukalliot, Kalanti	1131 07 Uusikaupunki	6747.80–522.35

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