Geological Survey of Finland

Bulletin 337

Sm-Nd, U-Pb and Pb-Pb isotopic evidence for the origin of the Early Proterozoic Svecokarelian crust in Finland

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Geologian tutkimuskeskus
Espoo 1986

1886 1986

Geological Survey of Finland, Bulletin 337

Sm-Nd, U-Pb AND Pb-Pb ISOTOPIC EVIDENCE FOR THE ORIGIN OF THE EARLY PROTEROZOIC SVECOKARELIAN CRUST IN FINLAND

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with 22 figures, 4 tables and 2 appendices

GEOLOGIAN TUTKIMUSKESKUS
ESPOO

Huhma, H. 1986. Sm-Nd, U-Pb and Pb-Pb isotopic evidence for the origin of the Early Proterozoic Svecokarelian crust in Finland. *Geological Survey of Finland, Bulletin 337.* 48 pages, 22 figures and 2 appendices.

The isotopic composition of Nd together with the U-Pb isotopic ages of zircons are reported for about 30 samples from different geotectonic settings in Finland, and have been used to deduce the origin of the Early Proterozoic Svecokarelian crust. Analyses from primitive mafic rocks indicate that a time-integrated light REE depleted mantle reservoir with $\epsilon_{Nd}(T)$ from +4 to +5 existed beneath the Baltic Shield during Early Proterozoic times, and was the source of some 2.1 Ga old Jatulian and 1.9 Ga old Svecofennian basalts. Material from such sources also contributed to the generation of the Svecokarelian felsic crust about 1.9 Ga ago.

The U-Pb zircon age of the majority of the granitoids in southern and central Finland is about 1.88 Ga, whereas in this work the zircons from the Proterozoic granites in northern Finland are mostly about 1.8 Ga old. All the granitoids investigated from the Svecofennian domain in southern and central Finland, and therefore not close to the Archaean craton, have $\epsilon_{Nd}(1.9)$ from −1 to +3. This indicates that they consist largely of newly mantle-derived material, with only minor admixture of older continental crust. There is no isotopic evidence so far for the existence of any Archaean crust in the Svecofennian domain.

In contrast, almost all the Svecokarelian granitoids investigated from northern and eastern Finland, and therefore close to the Archaean craton, have $\epsilon_{Nd}(T)$ from −3 to −10. This indicates that the contribution of recycled Archaean crustal material was significant in the genesis of the Svecokarelian granites in the Karelian domain. Even so, many of these granites do not contain old inherited zircons. The division of the Baltic Shield into the continental Karelian and orogenic Svecofennian blocks is strongly supported by isotopic data.

The formation of the Svecokarelian crust involved processes in which large quantities of new mantle-derived material was added to the continent and mixed effectively with a smaller amount of older crustal material. A modern analogue would be processes related to subduction in plate boundaries. Evidence for the mixing is observed not only in granitoids but also in some mafic orogenic rocks. A few analyses from the Svecofennian metagraywackes also suggest only a minor input from much older crust. The Kalevian metasediments relatively close to the Archaean in eastern Finland, in contrast, have older crustal residence age, although purely Archaean provenance is excluded. A model based on mixing during a 1.9 Ga orogeny is presented to explain the lead isotopic data on Finnish galenas.

During the later stages of the orogeny, input from the mantle diminished and crustal recycling was the dominant process. In the Karelian domain, granites formed largely from reworked Archaean crust, whereas partial fusion of the newly formed crust was involved in the generation of late-orogenic granites in the Svecofennian domain.

Key words: crust, genesis, absolute age, Sm/Nd, U/Pb, mantle, granites, gabbros, metabasalt, Proterozoic, Finland

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ISBN 951-690-232-4
ISSN 0367-522X

Helsinki 1986. Valtion painatuskeskus

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INTRODUCTION

The primary sources of material in the generation of new continental crust are the upper mantle and the existing upper and lower crust. The involvement of old recycled crustal material in this process has been an important issue in recent isotope geological studies. Much of this research has focused on granitoids, since they frequently form the bulk of the crust. Studies on the granite-generating processes themselves are not much aided by the radiogenic isotopes, which, rather, give unique information about the materials involved in the genesis of the granitoids, and help constrain the geotectonic evolution models.

Discussion on the origin of the granites was prompted by early observations of migmatites (e.g., Sederholm, 1926) and the experimental results obtained by Bowen and Tuttle (1950), which showed that the partial melting of crustal rocks (anatexis) offers a mechanism for producing large amounts of granitic rocks. On the basis of experimental data, Wyllie (1977) suggests that the normal product of high grade regional metamorphism is an H_2O-undersaturated granitic liquid. With increasing temperature or pressure the composition of the liquids trends towards granodiorite. The existing crust could thus be a major source of the granitic rocks. Batholiths dominated by tonalites probably include heat and magma transported from the mantle sources with subsequent fractionation and crustal assimilation (Wyllie, 1984).

The generation of Phanerozoic calc-alkaline magmas (e.g., batholiths) has often been connected with the subduction process at convergent plate boundaries (e.g., Gill, 1981). The source rocks of the magmas in subduction zones include subducted oceanic crust (altered basalts and ocean-floor sediments), mantle peridotite above the subduction zone and metamorphosed continental crust. The dehydration of subducted oceanic crust may initiate partial melting of the overlying mantle and subducted oceanic crust (Ringwood, 1975), producing magmas that may rise, fractionate and hybridize with anatectic crustal melts to form the parental material for granitoid batholiths (e.g., Wyllie, 1984).

Recent isotope studies on the involvement of older crustal material in the genesis of new crust are largely based on the Sm-Nd method, but information about the genesis of the granitoids has also been obtained using Sr, Pb, Hf and oxygen isotopes alone or combined with each other and Nd (e.g., Kistler and Peterman, 1973; Moorbath *et al.*, 1981; Patchett *et al.*, 1981; Taylor, 1980). The contribution of the old crust to the genesis of Phanerozoic granitoids is highly variable (e.g., Farmer and DePaolo, 1983). Many of these granitoids have a substantial old recycled component, suggesting that recent juvenile additions to the crust have been minor (Allègre and Ben Othman, 1980; Hamilton *et al.*, 1980; DePaolo, 1981; McCulloch and Chappell, 1982; Halliday, 1984). In contrast, the majority of the reported isotope data from the Proterozoic and Archaean granitoids suggest relatively short mean crustal residence time for their source materials (e.g., Moorbath and Taylor, 1981; Patchett *et al.*, 1981, Patchett and Bridgwater, 1984).

The generation of the Svecokarelian crust in the Baltic Shield over a period of 1.9—1.8 Ga (e.g., Kouvo and Tilton, 1966) was an important crust-forming event in Early Proterozoic times. Granitoids make up the major

part (70%) of this crust. The ages of many of these rocks are known as a result of U-Pb zircon work done at the Geological Survey of Finland. Although dating has constituted the major task of isotope work in this laboratory, lead isotopes have also been used for petrogenetic studies. In the early seventies a large variation was observed in the lead isotopic composition of K-feldspars from various Svecokarelian granitoid massifs, showing that Archaean material was probably involved in the genesis of some of the granites (Rouhunkoski, 1968; Meriläinen, 1976 and Kouvo, pers. comm.). A variation in source characteristics has also been observed using lead isotopes of Svecokarelian galenas in Finland (Kouvo and Kulp, 1961) and has been attributed to mixing of mantle- and crustal-derived leads (Vaasjoki, 1981). Hafnium isotopes, however, suggest that the relative amount of recycled Archaean crustal material was small during the formation of the Svecokarelian crust (Patchett *et al.*, 1981).

The aim of this study was to establish the initial $^{143}Nd/^{144}Nd$ ratios of some Finnish granitoids from different geotectonic environments using U-Pb zircon ages determined during this work or zircon ages reported in the literature. These data have been used to examine the relative involvement of mantle and old crustal materials in the genesis of these rocks. The emphasis is on the granitoid areas of central Lapland and central Finland, and on the change, if any, in the initial Nd ratio across the Raahe—Ladoga zone from the Archaean boundary in the east to the granitoid area of central Finland in the west (Fig. 1). Since the upper mantle is an important source in the formation of new crust, the evolution of the Proterozoic mantle is also discussed.

As this study is a survey of Nd-initial ratios from various granitoids, less attention has been paid to the variation within a single granitoid body. Variations may occur in the initial ratios as a result of diverse levels of contamination, but the data accumulated in this study and from the literature suggest that the qualitative conclusions reached are not dependent on the small variations in ϵ_{Nd} that may exist within a single granitoid massif.

GEOLOGICAL SETTING

The major Precambrian units in Finland are the granitoids and greenstone/schist belts of late Archaean (about 2.7 Ga) age, and the Early Proterozoic Svecokarelian terrains (Fig. 1). The Svecokarelian supracrustal rocks in eastern and northern Finland have been referred to as the Karelides and the schists in southern and western Finland as the Svecofennides (e.g., Eskola, 1963, Simonen, 1980). This division was based mainly on differences in the character of the sedimentary rocks, brought about by different geotectonic environments. The Karelian domain is characterized by epicontinental sediments (e.g., quartzites and dolomites) that deposited on the Archaean crust 2.45—1.9 Ga ago. The Svecofennian supracrustal rocks, in contrast, constitute 1.9 Ga old orogenic terrains, and consist mainly of micaceous greywacke-schists and volcanic rocks deposited in marine environments. The limestones and iron formations in southernmost Finland indicate relatively shallow marine conditions (Simonen, 1953).

The boundary zone between the Karelian and Svecofennian domains coincides with the major NW-trending lineament system (Gaál, 1972) known as the Main Sulphide Ore Belt (Kahma, 1973) or Raahe—Ladoga lineament

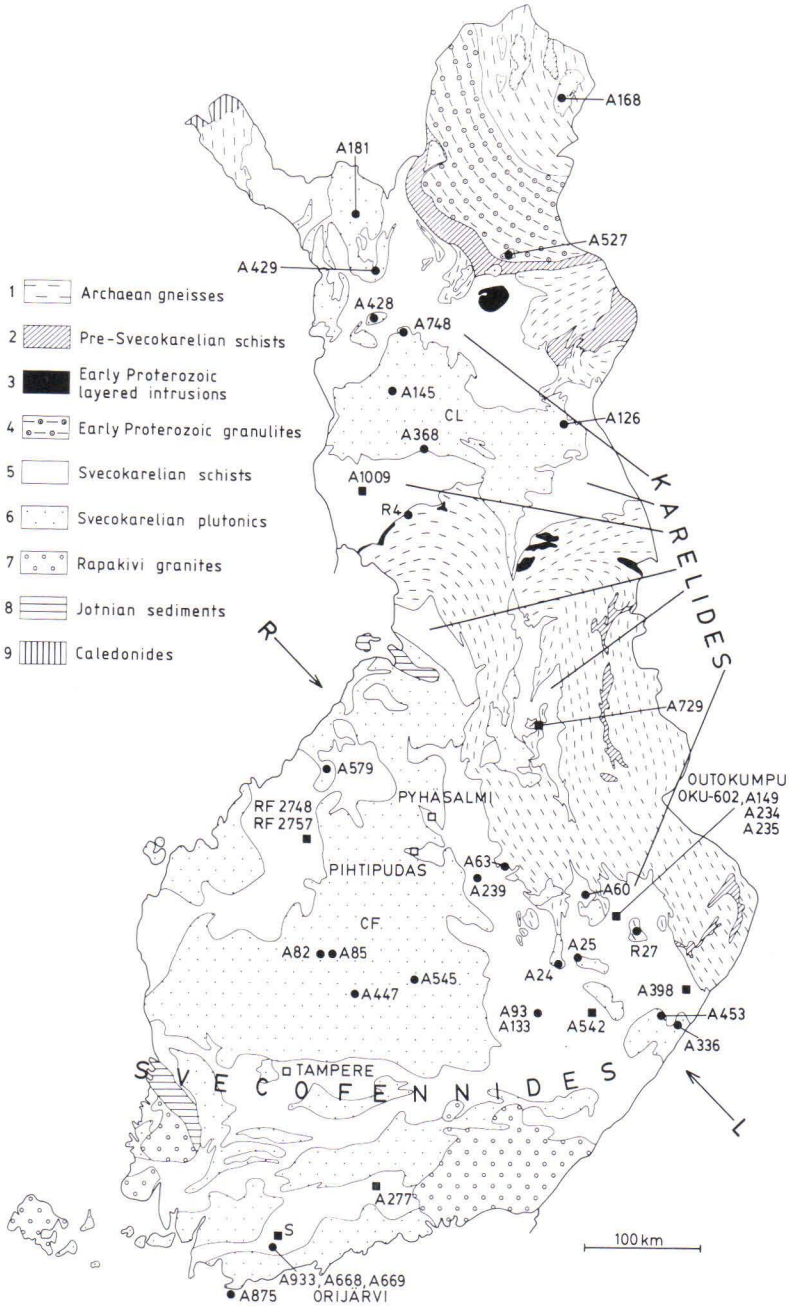

Fig. 1. Geological map of Finland, showing the Raahe—Ladoga lineament zone (R-L) and sample locations. Modified from Simonen (1980). 1. Archaean gneisses, 2. Archaean greenstone and schist belts, and the Tana belt of unknown age, SW of the granulites in northern Finland. 3. 2.45 Ga old layered mafic intrusions, 4. Early Proterozoic granulite belt in Lapland, 5. Svecokarelian supracrustal rocks, 6. Svecokarelian plutonic rocks (predominantly granitoids), 7. Proterozoic Rapakivi granites, 8. Mid-Proterozoic Jotnian sediments, 9. Caledonides. CL = granitoid area of central Lapland, CF = granitoid area of central Finland

(R-L in Fig. 1). A negative gravity anomaly and a deep-seated fracture zone revealed by deep seismic sounding (Luosto *et al.*, 1982, see also Korsman *et al.* 1984) are associated with this zone, which is also characterized by the distinct Pb-isotopic composition of galenas (Kouvo and Kulp, 1961; Vaasjoki, 1981).

Lower Karelian assemblages of clastic sediments (Sariolian and Jatulian) are associated with mafic volcanic rocks and sills 2.45—2.05 Ga in age (Sakko, 1971; Kratz *et al.*, 1976), which indicate intracratonic rifting and incipient basin development, as suggested by Hoffman (1980) for the roughly coeval Wopmay orogen in Canada. The upper part of the Karelian schists contains marine flysch-type sediments (Kalevian) intruded by the 1.87 Ga old Svecokarelian granitoids, and which deposited 2.05—1.9 Ga ago at the edge of the Archaean craton (Koistinen, 1981, Bowes *et al.*, 1984).

Plutonic rocks ranging from peridotite to granite intrude the Svecofennian volcanic-sedimentary terrain. This synorogenic plutonism and volcanism took place about 1.9—1.87 Ga ago (e.g., Neuvonen *et al.*, 1981). The 1.89 Ga granitic pebbles in Svecofennian conglomerates from the Tampere schist belt imply that part of the sedimentation was roughly contemporaneous with igneous activity (Kouvo and Tilton, 1966). Presumably, most of the Svecofennian geosynclinal turbiditic sediments deposited shortly before that (Simonen, 1953, 1980). No basement for the Svecofennian supracustal rocks has been demonstrated.

In the light of plate tectonic models, the Svecofennides have been considered ancient island arcs and the Raahe—Ladoga zone a palaeosuture zone between continental Karelian and oceanic Svecofennian blocks (e.g., Hietanen, 1975). An island-arc environment without a plate tectonic framework had in fact already been postulated by Simonen (1953, p. 35). The plate tectonic setting is corroborated by the chemical data on volcanic rocks from southern Finland (e.g., Orijärvi, Tampere) and the Raahe—Ladoga zone (e.g., Pyhäsalmi, Pihtipudas). The volcanic rocks often have calc-alcaline affinities (e.g., Latvalahti, 1979; Kähkönen and Laitakari, 1983), but tholeiitic basalts also occur (e.g., Mäkelä, 1980; Kousa, 1985). A large proportion of the Svecofennian metasediments have been considered products of turbidity currents. It has been suggested that the Outokumpu ophiolite assemblage (Koistinen, 1981), including the 1.97 Ga old gabbro (Oku 602 in Fig. 1), and some of the Kalevian flysch-type deposits in eastern Finland are an allochtonous thrust nappe (Koistinen, 1981). In spite of many similarities there are differences from recent orogenies, such as the very limited age interval within the synorogenic magmatism in large areas of the Svecofennian domain (zircon ages are frequently the same within error). The roughly 1.9 Ga old Wopmay orogen in Canada an its plate tectonic interpretations (Hoffman and Bowring, 1984) provide many analogies with the Svecokarelian orogeny.

The Svecokarelian orogeny is characterized by low pressure-high temperature metamorphic facies (e.g., Hietanen, 1975) and by the abundance of granitoids and migmatites (e.g. Sederholm, 1926). Sederholm (1932) divided the Precambrian granites in Fennoscandia into four groups. The first group was composed of gneissose granitoids, and the second group consisted of migmatite-forming microcline granites that cut the gneissose granitoids. In the magmatectonic classification these correspond to the synorogenic and late-orogenic phases. The third group comprises post-orogenic intrusions, and the fourth group post-Svecokarelian rapakivi granites (Simonen, 1980). Using mineralogical and geochemical data, Simonen (1960) divided the Svecofennian plutonic rocks into five provinces: granodiorite, trondjhemite, charnockite, granite and migmatite-forming microcline granite provinces. He considered rocks of the first four

provinces synorogenic but the migmatite-forming microcline granite late-orogenic and a product of granitization.

Most synorogenic granitoids contain 1.9—1.88 Ga old zircons (Kouvo and Tilton, 1966; Neuvonen *et al.*, 1981). Migmatite-forming late-orogenic microcline granites in southern Finland have zircon ages of about 1.83 Ga, and some postorogenic phases are about 1.80 Ga old (e.g., Vaasjoki, 1977). The youngest granites are the rapakivi granites with U-Pb zircon ages of 1.7 to 1.54 Ga (Kouvo, 1958; Vaasjoki, 1977). Areas in central Finland (CF) and central Lapland (CL) are dominated by granitic rocks (Fig. 1).

Although there is no evidence for Archaean crust in the Svecofennian domain, assumptions have been made concerning the existence of Archaean crust or reworked material in the granitoid areas of central Finland and central Lapland (e.g., Salop, 1983 and discussion in Lauerma, 1982). One of the purposes of the present study is to throw more light on this problem.

BASIC PRINCIPLES

Sm-Nd method

Since the work by Lugmair et al. (1975) the Sm-Nd isotopic method has become one of the most important tools for recording planetary evolution. The method, which is based on the alpha decay of samarium –147 (half life 106 Ga) to neodymium –143, has been used to determine the ages of rocks and minerals, and especially for petrogenetic studies. The abundance of ^{143}Nd is reported as the $^{143}Nd/^{144}Nd$ ratio. The measurable differences in this ratio in natural reservoirs are produced by the decay of ^{147}Sm during a considerable period of time or, if open systems are considered, by mixing of materials from different sources with varying $^{143}Nd/^{144}Nd$ ratios. The assumption of local equilibrium during partial melting, which is fundamental to the interpretation of isotope and trace element data, is generally held to be valid (cf., Hofmann and Hart, 1978). Isotope fractionation, which is the basis of the variation in light stable isotopes, can be considered negligible for Nd, because differences in the chemical properties of isotopic species diminish rapidly with increasing atomic weight (Urey, 1947). The fractionation correction calculated to eliminate the mass fractionation during mass spectrometric determination in fact corrects all fractionation.

The $^{143}Nd/^{144}Nd$ ratio measured today for a reservoir with a constant Sm/Nd ratio is given by:

$$(1)\quad \frac{^{143}Nd}{^{144}Nd)_{measured}} = \frac{^{143}Nd}{^{144}Nd)_{init}} + \frac{^{147}Sm}{^{144}Nd)_{measured}} \times (\exp(\lambda T) - 1),$$

where T is the age, λ the decay constant of $^{147}Sm = 6.54 \times 10^{-12}\ y^{-1}$, and $^{143}Nd/^{144}Nd_{)init}$ the initial ratio T years ago. Because the decay constant of ^{147}Sm is small the graph of this equation is very close to a straight line, and the rate of change of $^{143}Nd/^{144}Nd$ is directly proportional to $^{147}Sm/^{144}Nd$ (Fig. 2). Since the condensation and accretion of the solar system 4.5 Ga ago, the $^{143}Nd/^{144}Nd$ ratio in average chondrites has increased from 0.5068 to the present-day value of 0.5126 (Fig. 2). Because the rare earth elements (REE) have similar chemical properties, the fractionation between

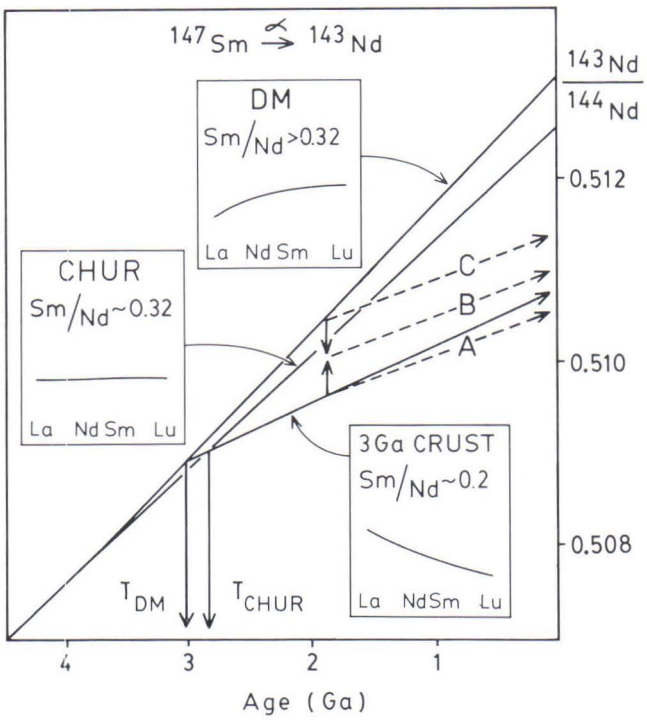

Fig. 2. Evolution of $^{143}Nd/^{144}Nd$ in the Earth's mantle and continental crust. The evolution lines are shown for the bulk Earth (=CHUR), for LREE depleted mantle (=DM), for 3 Ga old crust and for 1.9 Ga old crust: A= 1.9 Ga old granitoid generated by anatexis from 3 Ga old crust, B= 1.9 Ga old granitoid generated from material that was a mixture of juvenile and Archaean crustal material, C= 1.9 Ga old granitoid generated from juvenile mantle material. Model ages are shown for a sample from 3 Ga old crust: T_{DM} = model age calculated relative to the depleted mantle, T_{CHUR} = model age relative to the chondritic reservoir (see text). The insets show the conventional chondrite-normalized REE patterns for the reservoirs involved.

Sm and Nd is relatively small in natural processes compared with that between parent and daughter elements for other widely used isotope systems (U-Pb, Rb-Sr). Consequently, the variation in $^{143}Nd/^{144}Nd$ in natural systems is small (Fig. 2), and the use of Nd in dating or geochemistry requires routine measurement of $^{143}Nd/^{144}Nd$ to a precision of 4 to 5 parts per 10^5 or better.

For Sm-Nd dating we need a set of cogenetic samples that have enough variation in the Sm/Nd ratio, and which had the same initial Nd ratio during the time of their crystallization. A further requirement is that the Sm/Nd ratio for each sample should not have changed since crystallization. Because of the chemical similarity of the REE, the latter assumption is likely to be valid in most cases, but for the same reason, it is very difficult to find a set of whole rock samples that fulfil all the requirements. The age determination is more promising, if minerals of the primary crystallization, e.g. plagioclase and pyroxene, are available. The basis for the age calculation is equation 1, which has two unknowns (age and initial ratio), and can be written for all minerals and whole rocks. In practice, this leads to the least squares fit to the data points in the plot of $^{143}Nd/^{144}Nd$ vs. $^{147}Sm/^{144}Nd$, where the age T is proportional to the slope m of the straight line (e.g., Fig. 18): $T = 1/\lambda \times \ln(m+1)$.

The evolution of the bulk Earth, which has been inferred from analysing chondritic meteorites, is referred to as the CHUR (''chondritic uniform reservoir'') evolution (DePaolo and Wasserburg, 1976), with $^{147}Sm/^{144}Nd$ and $^{143}Nd/^{144}Nd$ ratios the same as in average chondrites (Jacobsen and Wasserburg, 1980). Using normalization based on $^{146}Nd/^{144}Nd = 0.7219$, the present-day values are 0.1966 and 0.51264, respectively. Some laboratories use another basis for the normalization and obtain

totally different Nd ratios (op.cit.). The $\epsilon_{Nd}(T)$ notation eliminates the discrepancy, whereafter the measurements from all laboratories are comparable. (Without normalization, which corrects the fractionation during the mass spectrometric run, the Sm-Nd isotope method would be absolutely useless.)

The $^{143}Nd/^{144}Nd$ ratios and especially the initial ratios (the value at the time of crystallization) are often reported relative to the CHUR by epsilon notation:

$$(2)\ \epsilon_{Nd}(T) = \left(\frac{^{143}Nd/^{144}Nd)_{sample}(T)}{^{143}Nd/^{144}Nd)_{CHUR}(T)} - 1 \right) \times 10^4$$

The formation of the continental crust involves large chemical fractionation, and the Sm/Nd ratio of the crust is generally about 40% lower than that of CHUR. For any sample the model age T_{CHUR} may be calculated relative to the CHUR curve:

$$3)\ T_{CHUR} = 1/\lambda \times \ln\left(1 + \frac{^{143}Nd/^{144}Nd)_{meas} - 0.51264}{^{147}Sm/^{144}Nd)_{meas} - 0.1966}\right)$$

Most of the upper mantle is depleted in incompatible and, especially, in light rare earth elements (LREE). This depletion can be seen in the REE patterns for present mid-ocean ridge basalts (MORB). The high ϵ_{Nd} values from MORB (+10) further indicate a time-integrated LREE depletion in the MORB sources (Fig. 2). The accumulating data show that this depletion took place already during Proterozoic and Archaean times as a consequence of the formation of large amounts of LREE-enriched continental crust. Thus, the upper mantle does not generally follow the chondritic evolution, and the model ages relative to some depleted mantle reservoir (DM) may be more useful. In this paper, the model ages, T_{DM}, are calculated relative to the model by DePaolo (1981), which is

$$(4)\ \epsilon_{Nd}(T) = 0.25 \times T^2 - 3 \times T + 8.5,$$

where T is the age in Ga.

As a consequence of elapsed time, it is obvious that Archaean crustal material with a low Sm/Nd ratio, and also a 1.9 Ga-old granite generated by remelting of that crust, had a much lower $^{143}Nd/^{144}Nd$ ratio (negative ϵ_{Nd}) during the Svecokarelian orogeny about 1.9 Ga ago than did the CHUR or a melt from the depleted mantle reservoir (Fig. 2). This systematic variation in ϵ_{Nd} in the mantle and crustal reservoirs can be used for inferring the sources of the granitic magmas, and is also the basis for this study. The initial Nd-isotopic composition for each whole rock sample was calculated using equation (1) from the measured Sm and Nd concentrations and the Nd-isotopic composition, assuming no fractionation between Sm and Nd after the crystallization of the sample. Because the Sm/Nd ratio of the granitoids is close to the average Sm/Nd ratio of the crust, the model age, T_{DM}, of a granitoid is reasonable and can be considered a crust-formation age or a crustal residence age, as defined by O'Nions *et al.* (1983).

U-Pb analyses on the concordia diagram

The U,Th-Pb method is based on the decay of ^{238}U to ^{206}Pb, ^{235}U to ^{207}Pb and ^{232}Th to ^{208}Pb. The number of atoms of Pb that accumulate in a time T are given by:

$$(5)\ ^{206}Pb = ^{238}U \times (\exp(\lambda_{238} \times T) - 1)$$

$$(6)\ ^{207}Pb = ^{235}U \times (\exp(\lambda_{235} \times T) - 1),$$

where the λ's are the decay constants. Combining these equations and using the constant present-day value of 137.88 for $^{238}U/^{235}U$, we get the lead/lead age, which can be obtained from the isotopic analysis of lead alone:

$$(7)\quad \frac{^{207}Pb}{^{206}Pb} = \frac{1}{137{,}88} \times \left(\frac{\exp(\lambda_{235} \times T) - 1}{\exp(\lambda_{238} \times T) - 1} \right)$$

The concordia diagram (Wetherill, 1956) is widely used for presenting U-Pb zircon analyses. The concordia curve is the locus of all the points for which the $^{206}Pb/^{238}U$ ages equal the $^{207}Pb/^{235}U$ ages. The great advantage of the U-Pb system is that there are two coupled decay systems, which thus offer a method for testing the assumption of a closed U-Pb system with a single analysis. In practice, closed systems are rare, and normally zircons are discordant. For a single analysis this usually means that

$T_{205Pb/238U} < T_{207Pb/235U} < T_{207Pb/206Pb}$.

In a normal situation, if the analysis is not very discordant, the $^{207}Pb/^{206}Pb$ age is only slightly younger than the true zircon age.

The analyses of fractions of cogenetic zircon populations usually plot in a straight line on the concordia diagram, and according to the episodic lead loss (Wetherill, 1956) or continuous diffusion models (e.g., Tilton, 1960), the upper intercept of this line with the concordia curve can be considered the time of crystallization of the zircons. In many Phanerozoic examples, the lower intercept is taken as the crystallization age of the rock because of the presence of older zircon populations together with younger magmatic zircon. Old zircons surviving in rocks have been called inherited zircons (Pidgeon and Aftalion, 1978). More complex models have frequently been invoked e.g. due to multiphase zircons. For an introduction to U-Pb dating reference is made to Gebauer and Grünenfelder (1979).

SAMPLING

An effort was made to analyse synorogenic and late-orogenic granitoids from different Svecokarelian tectonic environments in the Svecofennian and Karelian domains. One criterion for the sampling was existing isotopic data. Where U-Pb zircon results were available, a new sample (5 kg) was, in most cases, collected from the same locality for Sm-Nd analysis. Some old zircon data (by borax fusion) were confirmed by analysing zircon fractions separated from the original sample. Four new sampling sites were selected in the granitoid areas of central Finland (CF) and central Lapland (CL).

Mafic rocks

For the study of evolution of the Proterozoic mantle, primitive mafic rocks were analysed from four localities representing the evolutionary stage of the Svecokarelian orogeny. These include the c. 2.1 Ga Jatulian basalts from the Jouttiaapa formation (A1009; Perttunen, 1985) and Oravaara (A398; Nykänen, 1968) and two gabbros (A729, OKU—602) associated with serpentinites and which presumably belong to the ophiolite systems 1.97 Ga in age (Koistinen, 1981; Kontinen, in press). The synorogenic mafic samples about 1.9 Ga in age include a hornblende gabbro from Mäntsälä (A277, A278-Soukkio; Härme, 1978), a Ni-sulphide-bearing norite from Laukunkangas (A542; Grundström, 1980), a peridotite from the Salittu massif (S, Eskola, 1914) and two tholeiitic basalts from Evijärvi (RF2748, RF2757; Vaarma, 1984). These basalts are from the Pohjanmaa schist belt, which contains Svecofennian metagraywackes and metavolcanic rocks.

Granites from northern Finland

The geology and granites of northern Finland have been described by Mikkola (1941). Samples from central Lapland and Nattanen type granites are included in the present study. The Nattanen-type granites are multiphase, relatively uniform intrusions compared with the large granitoid area in central Lapland (Fig. 1.), which is composed of various types of granitoid, schist inclusions and migmatitic rocks (Mikkola, 1941; Lauerma, 1982). The rocks included in the present work are true granites (Streckeisen, 1974), probably close to the minimum melting composition and often very poor in mafic minerals. The geology of the large Vainospää granite (A168), which intrudes Archaean gneisses and schists, has been described by Meriläinen (1976). The Nattanen-type granites form a group of intrusions cutting granulites (A527-Nattanen) and gneissose granitoids (A429-Kuivavaara). The granite of central Kittilä (A428-Lappalaislampi) intrudes the Lapponian supracrustal rocks (Rastas, 1984). The granitoid area of central Lapland is not very well known, but pink granite, similar to sample A145, occurs in large areas along the river Ounasjoki (Rastas, pers. comm.). The granite from Salla in the present study is a fine-grained rock (A126) and has been described by Lauerma (1982), whereas the granite of Iso-Nilipää (A748) is an amphibole-bearing, porphyritic and coarse-grained rock (Rastas, pers. comm.).

Granitoids from southern and central Finland

The granitoids from southern and central Finland sampled for this study range from granite to quartz diorite. The samples from the Svecofennian domain (southwest of the Raahe—Ladoga zone) comprise intrusive granodiorites from the Rautio batholith (A579-Susineva; Gaál and Isohanni, 1979; Nurmi *et al.*, 1984) and Orijärvi (A933; Eskola, 1914); late-orogenic migmatite-forming Hanko granite (A875-Märaskär; Sederholm, 1926); a gneissose trondjhemite/ tonalite (A93-Saunakangas) from closer to the Raahe—Ladoga lineament zone and somewhat similar to the 1.88 Ga old Tuusmäki gneissose tonalite studied by Korsman *et al.* (1984); and four representative rocks from the granitoid area of central Finland.

The granitoid area of central Finland (CF in Fig. 1) is composed largely of various granitoids, in which supracrustal rocks occur as inclusions. The samples include a coarse-grained microcline granite from Jyväskylä (A545-Lehesvuori), where six granitoid types from diorite to granite have been encountered (Nurmi *et al.*, 1984). The microcline-rich granite included in this study is the youngest phase observed in the field. The other samples are a slightly foliated granodiorite from Keuruu (A447), which is the most common rock type in the area (Marmo, 1963), and two samples from the Ähtäri map sheet area (Sjöblom, 1984); a coarse-grained amphibole-bearing granite (A82-Hankavesi) and a coarse-porphyritic granodiorite /granite (A85-Kiukaanniemi). The latter granitoid is somewhat similar to those described by Aho (1979) from Pihtipudas and by Nurmi *et al.* (1984) from Viitasaari.

The samples from the intrusions in the Karelian domain, eastern Finland, are granites, granodiorites and quartz diorites, including the lateorogenic Puruvesi granite (A453-Syrjäsalmi, Nykänen, 1975), and grey granodioritic rocks from Kermajärvi (A25-Viitalahti), Suvasvesi (A24-Parkkolansaari), Onkivesi (A63-Jussilansaari) and Maarianvaara (A60-Vihtajärvi). The ''Maarianvaara granite'' is composed of diorite to granite, mainly granodiorite (A.

Huhma, 1975, 1976), and intrudes the Kalevian metasediments and Archaean gneisses. The Onkivesi intrusion is situated in a somewhat similar position. The Kermajärvi intrusion is composed mainly of granodiorite and tonalite with inclusions of schists. Both Kermajärvi and Suvasvesi granitoids intrude Kalevian metasediments, which are slightly migmatized.

The samples are described in Appendices 1 and 2. The localities are shown in Fig. 2. The geological references are given in Table 2.

ANALYTICAL PROCEDURES

The chemical preparation of samarium and neodymium was based essentially on the method described by O'Nions et al. (1976). The uranium- lead analyses followed procedures established by Krogh (1973). The isotopic measurements were made on two 9''-radius, 60° sector Nier-type mass spectrometers. Built at the laboratory of the Geological Survey of Finland, they are similar to the one constructed at the DTM, Carnegie Institution, Washington D.C.

U-Pb method

The analytical methods used for U-Pb zircon dating in this laboratory have been described by Vaasjoki (1977). The final separation of zircons was done using a Frantz magnetic separator, heavy liquids and handpicking under a binocular microscope. The sample, consisting of 5—20 mg of zircon, was washed in hot 7N HNO_3 and rinsed several times with H_2O in an ultrasonic bath to remove any surface contamination. Before washing, some fractions were treated in an ultrasonic cleaner with 5% HF for 4 minutes. This procedure removes altered outer margins and gives more concordant analyses (Krogh *et al.*, 1982). The analyses were carried out in the way described by Krogh (1973) using a Teflon capsule with a steel jacket to dissolve the zircon, and anion exchange chromatography to separate uranium and lead. The sample was spiked with ^{235}U tracer and, after aliquoting, with ^{208}Pb- or ^{206}Pb tracer.

Isotopic measurements of lead were made using the silica-gel technique, Ta_2O_5 being used as an activator for uranium. For most analyses, the $^{206}Pb/^{204}Pb$ ratio was measured on the chart. Total Pb blank levels ranged from 0.1 to 1.3 ng. The very high $^{206}Pb/^{204}Pb$ ratios (up to 40,000) measured for some zircons in our laboratory show that, in these cases, the blank was significantly less than 1 ng, since, assuming that all ^{204}Pb was derived from the blank, we get Pb blank levels of about 0.6 ng.

The 2-sigma error estimates based on standard and replicate sample analyses are about 0.7% for U/Pb ratios and 0.2—0.3% for $^{207}Pb/^{206}Pb$ ratios. York's (1969) method was used to fit the regression line to the U-Pb data, and the age uncertainties of the concordia intersections are given at the 2-sigma level. Individual data points are weighted according to their precision and a correlation coefficient of 0.92 to 0.96 between $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ errors is used according to the treatment of Ludwig (1980). The statistical parameter MSWD (=Mean Square of Weighted Deviates) was calculated to illustrate the quality of the regression (Table 2). If MSWD is less than or equal to 1, the data

points fit the chord well, whereas numbers exceeding 1 indicate some extra scatter in addition to the experimental error. All the ages were calculated using the decay constants recommended by Steiger and Jäger (1977).

Sm-Nd method

Sample preparation and chemical separation

Samarium and neodymium were separated chemically using a method adopted from the University of Leeds, Department of Earth Sciences (Thirlwall, 1982, Cliff *et al.*, 1983), and based on the procedure described by O'Nions *et al.* (1976). The samples collected for the Nd work usually weighed a few kilograms, though an experiment showed that the result was the same when a rock sample weighing only 10 g was used (sample A24 vs. A24-small in Table 4.). The fresh sample was split into small pieces, and about 200 g of fresh rock was washed with clean water in an ultrasonic bath and crushed in an iron pan in a swing-mill.

All the reagents used were distilled in vessels made of quartz or teflon. The ion exchange columns, beakers, etc., were quartz or teflon. Between 100 and 200 mg of sample powder was placed in a teflon "bomb" (a teflon capsule sealed in a steel jacket) and dissolved for a few days in a HF/HNO_3-mixture at 200°C. After the fluorides had been evaporated with HNO_3, the residue was dissolved overnight in $HNO_3 + H_2O$ at 160°C using a teflon "bomb". After cooling, the clear solution was poured into a clean teflon beaker. This dissolution procedure was repeated until a clear solution was obtained, which in some cases involved total volumes of 30 ml. The clear solution was poured several times back and forth between two teflon beakers to ensure that the solution was well mixed and homogeneous. This solution was aliquoted; about one-third was spiked with a mixed ^{149}Sm-^{145}Nd spike and the remainder was used for determining the Nd isotopic composition.

The separation of Sm and Nd involved two ion exchange columns made of quartz. In the first step, the REE were separated as a group, and in the second step, Sm and Nd were separated from other REE and each other. The sample was dissolved in 5 ml of a mixture of 75% CH_3COOH + 25% 5M HNO_3. The first column, filled with 3 ml of Bio-Rad AG 1x8 (200—400 mesh) anion exchange resin, was prepared with a mixture of 90% CH_3COOH + 10% 5M HNO_3 (90/10). After centrifuging, the clear solution was loaded onto the prepared column with a disposable pipette. The sample solution was allowed to pass through the column, and the column was washed twice with liquid 90/10. The column was eluted with 35 ml of 90/10 liquid. The REE group was collected with 10 ml of 0.05M HNO_3. The second temperature-controlled ion exchange column was filled with about 4 ml of the same anion-exchange resin. The column was calibrated using CH_3OH + CH_3COOH + HNO_3 + H_2O mixtures, Sm-Nd solution and a drop counter. The dry residue was dissolved in 1 ml of a solution composed of 75% MeOH, 10% 8M HOAc, 10% 5M HNO_3 and 5% H_2O (liquid A). The sample was loaded at 25°C onto the prepared column. After washing with 2 x 1 ml of liquid A and about 20 ml elution of liquid A, Sm was collected with about 25 ml of liquid A. The temperature was increased to 35°C and about 20 ml of liquid B was eluted (B: 75% MeOH, 10% 8M HOAc, 5% 5M HNO_3 and 10% H_2O). Nd was collected with about 30 ml of liquid B.

The separation of Sm and Nd was good, since Sm interference during the Nd run was usually negligible. At the beginning of the work, the total Nd blank was 0.4—1 ng; later

it was 0.2 ng. The last blank measurement, which includes all the steps after the bomb dissolution, was 0.09 ng.

Mass spectrometry

Isotopic measurements of Sm and Nd were made using a triple or double rhenium filament technique. The sample was loaded in a drop of H_2O onto a filament previously outgassed in vacuum at 4A for two hours. Automated operation was achieved using a Commodore 4000 computer linked up to the mass spectrometer with an IEEE—488 bus. Nd^+ and Sm^+ beams were measured using a 10^{11} Ohm input resistor, a VG CA—3 ampere meter, a Fluke 8502A Digital Multimeter and a Kepco SN 488—121 programmer for peak jumping. A stable beam intensity of 10^{-11} A in mass ^{144}Nd was often achieved. For the Nd analyses, the mass spectrum was scanned by jumping the magnetic field sequentially to the following mass positions: 146, 144, 143, 146.6, 147 and 145. The data were collected in sets of 10 ratios each; 20—30 sets were normally measured. During the data collection, the signal was allowed to settle for 5 seconds and integrated for 2 seconds. The slight Sm interference sometimes observed at the beginning of the Nd run was seldom detectable later on. The Nd isotope ratios were normalized to $^{146}Nd/^{144}Nd = 0.7219$. In the isotope dilution analyses, $^{145}Nd/^{144}Nd$, $^{145}Nd/^{143}Nd$ and $^{149}Sm/^{147}Sm$ ratios were used for isotope dilution calculations. The fractionation was corrected with $^{146}Nd/^{144}Nd$ and $^{147}Sm/^{152}Sm$ ratios with the method described by Hofmann (1971) using the samarium isotopic composition of Wasserburg et al. (1981).

Quality of the data

For good-quality mass spectrometer measurements, the 2-sigma-mean (2x standard error of mean =2SEM) uncertainty in the $^{143}Nd/^{144}Nd$ ratio was about 2—3 x 10^{-5}. The reproducibility of the data was estimated by measuring the La Jolla Nd standard received from Dr. G.W. Lugmair. The mean of 18 runs since the beginning of this work was $^{143}Nd/^{144}Nd = 0.511872 \pm 17$ (1 std. dev.) and $^{145}Nd/^{144}Nd = 0.348422 \pm 10$ (1 std. dev.). During the sample measurement, the $^{145}Nd/^{144}Nd$ ratio was also measured to monitor data quality (see Tables 3 and 4). The mean of the last 20 samples during this work was $^{145}Nd/^{144}Nd = 0.348418 \pm 11$ (1 std. dev.). The $^{143}Nd/^{144}Nd$ ratio was also measured from the spiked runs; although usually less precise, it was equal to the ratio obtained from the isotopic composition runs. Analysis of standard BCR—1 yielded $^{143}Nd/^{144}Nd = 0.512664 \pm 28$ and $^{148}Nd/^{144}Nd = 0.241576 \pm 26$ (one measurement, 2SEM). These values are in close agreement with those obtained by other workers (e.g., Hamilton *et al.*, 1983).

The reproducibility of the $^{147}Sm/^{144}Nd$ ratio was estimated from duplicate analyses to be within 0.4%. It seems to be better for basalts than for granites, probably owing to powder heterogeneity or, in some cases, to problems in dissolution. The Sm/Nd ratio of the spike was calibrated against the gravimetric standard solution prepared from Sm_2O_3 and Nd_2O_3 powders. Stoichiometry was improved by heating the oxides and weighing them repeatedly until a constant weight was obtained. The loss of weight was about 12%. The Sm/Nd ratio of the spike was subsequently calibrated relative to the CalTech mixed Sm/Nd standard (Wasserburg et al., 1981), which resulted in a very slight correction to the Sm/Nd ratio of our standard. The Sm/Nd ratios reported here should now be compatible with the Caltech standard. The Sm/Nd ratio measured for BCR—1 was 0.2285. The model ages were calculated applying $^{147}Sm/^{144}Nd)_{CHUR} = 0.1966$, $^{143}Nd/^{144}Nd_{CHUR} = 0.51264$ and $\lambda^{147}Sm = 6.54 \times 10^{-12}\ y^{-1}$.

RESULTS

Age data

The U-Pb isotopic data produced in this study are given in Table 1 and shown on the concordia diagrams in Figs. 3—16. For other samples included in the neodymium studies, old U-Pb data were utilized. These analyses were made earlier at the isotope laboratory of the Geological Survey of Finland. All the U-Pb results are summarized in Table 2.

The U-Pb ages are based on analyses on zircon density fractions. As has commonly been observed, the heavier fractions contain less uranium and tend to be most concordant (Table 1). Fractions pretreated with dilute HF are more concordant than the corresponding untreated fractions, and usually plot on the same discordia chord as the other fractions.

Table 1. U-Pb analytical results

Sample[1]	Concentrations (ppm)		Measured	Atomic abundance,[2] ^{206}Pb = 1000			Atomic ratios[3]			Age (Ma)[4]
d = density Ø = size in μm HF = preleached in HF	^{238}U	$^{206}Pb_{rad}$	$\frac{^{206}Pb}{^{204}Pb}$	^{204}Pb	^{207}Pb	^{208}Pb	$\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{207}Pb}{^{206}Pb}$	$\frac{^{207}Pb}{^{206}Pb}$
A398-Oravaara, metabasalt										
A d>4.6	111.5	35.05	501	1.781	153.96	1891.3	0.3633	6.481	0.12938	2089 ±18
B 4.2<d<4.6	150.5	48.10	335.4	2.897	169.45	2176.2	0.3693	6.614	0.12989	2096 ±11
C 3.8<d<4.2	181,1	57.39	286.3	3.382	176.33	2365.9	0.3663	6.579	0.13028	2101 ± 9
Horsmanaho, gabbro pegmatoid										
A149A total	387.4	105.4	555.4	1.640	139.18	284.23	0.3144	5.075	0.11706	1912 ±12
A234A d>4.6	256.7	77.1	4334	0.1550	122.11	193.58	0.3471	5.743	0.12002	1956
A234B 4.2<d<4.6	840.6	231.4	1665	0.5151	125.07	267.98	0.3182	5.183	0.11813	1928
A235A d>4.2	282.9	83.1	7507	0.0383	120.55	168.81	0.3394	5.616	0.12004	1956
A542-Laukunkangas, norite										
A d>4.65	285.0	82.78	9144	0.0782	115.79	113.13	0.3357	5.310	0.11473	1875
B 4.6<d<4.65	299.7	87.15	12639	0.0710	115.85	117.18	0.3361	5.324	0.11489	1878
C 4.6<d<4.65/HF	281.3	81.59	16500	0.0503	115.59	113.70	0.3352	5.310	0.11491	1878
A277-Soukkio A, Mäntsälä, gabbro pegmatoid										
A d>4.6, Ø>160	153.0	44.23	9310	0.0906	116.11	176.46	0.3340	5.263	0.11431	1869
B 4.2.<d<4.6, Ø>160	363.5	101.1	8500	0.1093	116.26	166.84	0.3214	5.061	0.11420	1867
A278-Soukkio B, Mäntsälä, gabbro pegmatoid										
A d>4.6, Ø>160	114.0	32.74	13850	0.0509	115.42	177.08	0.3319	5.224	0.11417	1866
B d>4.6, Ø<160	121.3	33.79	8556	0.0934	115.39	157.98	0.3219	5.040	0.11355	1857
C 4.2<d<4.6,Ø>70	291.3	83.03	6927	0.1346	116.56	210.64	0.3295	5.186	0.11415	1866

Table 1. (continued)

A145-Molkoköngäs, granite										
A 4.4<d<4.55, Ø<160	372.4	88.92	2110	0.4524	119.15	401.55	0.2759	4.299	0.1130	1848 ±11
B 4.2<d<4.3, Ø>160	697.5	121.1	819.3	1.210	126.75	253.02	0.2007	3.052	0.1103	1804
C 4.2<d<4.3, Ø>160/HF	660.9	119.6	922.6	1.071	125.46	199.96	0.2091	3.197	0.1109	1814
D 4.0<d<4.2, Ø>160	895.9	128.5	507.4	1.965	137.00	269.69	0.1657	2.520	0.1103	1804
E 4.4<d<4.5, Ø<110	349.0	85.08	385.9	2.527	148.10	737.73	0.2817	4.424	0.1139	1862 ±30
F 4.3<d<4.4, Ø>160	514.7	110.0	738.7	1.344	128.99	208.20	0.2470	3.773	0.1108	1812
G 4.0<d<4.2, Ø>130	895.2	112.9	815.7	1.209	125.13	166.17	0.1458	2.185	0.1087	1777
A748-Iso Nilipää, granite										
A d>4.6	535.6	163.3	3109	0.3140	136.11	188.09	0.3515	6.361	0.13126	2115
B 4.2<d<4.6, Ø>160	820.1	243.5	1867	0.5258	138.57	157.45	0.3427	6.184	0.13088	2109
C 4.0<d<4.2, Ø>160	1115	290.3	1440	0.6821	138.44	155.58	0.3005	5.330	0.12863	2079
D 3.6<d<3.8, Ø>70	2472	354.7	961.2	1.106	134.01	182.07	0.1654	2.699	0.11837	1931
E titanite	192.2	76.37	60.49	16.37	351.26	1012.1	0.4585	8.2785	0.13096	2110 ±30
A24-Parkkolansaari, granodiorite										
A total, borax[5]	678.9	160.7	2019	0.3583	117.61	72.59	0.2735	4.253	0.11277	1844 ±10
B 4.2<d<4.6	759.1	200.7	3656	0.2681	117.33	69.26	0.3055	4.789	0.11371	1859
C 4.2<d<4.6/HF	623.0	176.2	4852	0.1994	116.89	75.00	0.3269	5.147	0.11420	1867
D d<4.2	932.6	210.7	3368	0.2905	116.49	67.02	0.2611	4.052	0.11255	1841
A25-Viitalahti, granodiorite										
A total, borax[5]	902.3	171.9	1031	0.8489	123.97	85.36	0.2202	3.417	0.11259	1841 ±10
B 4.2<d<4.6	839.9	192.9	3787	0.2585	116.98	61.78	0.2656	4.156	0.11351	1856
C 4.2<d<4.6/HF	641.1	173.6	14116	0.0649	115.04	59.16	0.3130	4.926	0.11417	1867
A60-Vihtajärvi, granodiorite										
A total, borax[5]	778.1	201.4	5850	0.0325	113.33	204.48	0.2991	4.656	0.11288	1846
B 4.2<d<4.6	556.6	152.9	7879	0.1217	114.98	193.10	0.3175	4.962	0.11334	1853
C 4.0<d<4.2	770.7	205.6	5388	0.1790	115.36	207.76	0.3082	4.799	0.11294	1847
D 4.0<d<4.2/HF	659.5	183.5	30311	0.0241	113.55	213.47	0.3215	5.019	0.11322	1851
E monazite	1043.0	300.0	20761	0.0293	114.24	4905.5	0.3324	5.218	0.11385	1861 ± 7
A875-Märaskär, granite										
A d>4.0	1063.7	220.2	1584	0.5973	118.88	359.54	0.2386	3.621	0.11009	1801
B 3.8<d<4.0	1786.2	238.1	1152	0.8278	118.68	106.44	0.1536	2.259	0.10668	1743
C monazite	2228.9	619.1	4918	0.1628	114.61	5724.4	0.3205	4.941	0.11182	1829
A668-Orijärvi, granodiorite										
A d>4.0, Ø>70	663.5	135.3	1532	0.3923	116.37	124.30	0.2357	3.587	0.11040	1806
B d>4.0, Ø<70	728.3	141.9	1396	0.5624	118.28	132.62	0.2252	3.414	0.10994	1798
C 3.8<d<4.0	1052	175.5	1248	0.5178	114.26	139.81	0.1928	2.832	0.10654	1741
A669-Nyckeln, quartz diorite										
A 70<Ø<130	711.5	159.0	1722	0.4115	117.98	84.59	0.2585	3.983	0.11174	1828

Table 1. (continued)

A993-Orijärvi, granodiorite											
A d>4.6	383.9	87.68	1281	0.6118	122.83	122.64	0.2639	4.143	0.11387	1862	±20
B 4.2<d<4.6	507.7	109.2	847	1.136	127.62	156.27	0.2484	3.816	0.11141	1822	± 9
C 4.0<d<4.2	824.6	159.6	356	2.731	148.43	289.90	0.2240	3.404	0.11022	1803	±30
A82-Hankavesi, granite											
B 4.3<d<4.6	330.9	86.69	3700	0.2627	118.1	118.36	0.3027	4.782	0.11455	1873	
C 4.2<d<4.3	589.1	152.1	3082	0.3074	118.41	124.56	0.2985	4.701	0.11425	1868	
D 4.0<d<4.2	868.0	213.4	2442	0.4017	118.88	152.38	0.2842	4.445	0.11344	1855	
E 4.3<d<4.6/HF	309.6	88.74	4805	0.1992	117.70	119.94	0.3313	5.253	0.11500	1880	
F titanite	181.6	52.98	1589	0.6165	123.39	845.71	0.3372	5.349	0.11506	1880	±10
A85-Kiukaanniemi, porphyritic granite											
A d>4.3	542.0	102.5	1410	0.6964	123.57	117.08	0.2186	3.440	0.11415	1866	± 7
B 4.2<d<4.3	876.4	158.2	1449	0.6799	121.73	107.41	0.2086	3.235	0.11251	1840	
C 4.0<d<4.2	1361	245.3	1849	0.5331	119.18	100.66	0.2083	3.214	0.11194	1831	
D 4.2<d<4.3/HF	588.2	128.9	2315	0.4185	119.70	98.16	0.2533	3.982	0.11404	1864	
E 4.0<d<4.2/HF	885.3	216.2	4807	0.1975	116.63	87.45	0.2822	4.434	0.11396	1863	
A447-Keuruu, granodiorite											
A Ø>130, borax	668.3	158.4	1839	0.4219	120.16	127.32	0.2739	4.321	0.11443	1871	±10
B d>4.6	374.6	96.86	2742	0.3529	119.63	124.53	0.2988	4.732	0.11486	1877	
C 4.3<d<4.6, 70<Ø<160	522.8	127.9	1533	0.6426	122.72	135.50	0.2827	4.445	0.11403	1864	
D 4.3<d<4.6, 70<Ø<160 HF	422.6	116.6	3450	0.2816	118.65	121.34	0.3190	5.051	0.11484	1877	
A545-Lehesvuori, granite											
A d>4.3	407.7	101.65	3389	0.2756	118.59	109.61	0.2881	4.563	0.1149	1877	
B 4.2<d<4.3/HF	536.0	138.9	4496	0.2150	117.81	106.18	0.2994	4.744	0.1149	1878	
C 4.2<d<4.3	619.4	149.5	3553	0.2740	118.21	108.57	0.2790	4.405	0.1145	1872	
A93-Saunakangas, Joroinen, trondhjemite/tonalite											
A d>4.5	248	69.4	10129	0.0864	117.02	95.22	0.32385	5.1729	0.11585	1893	± 8
B 4.35<d<4.5/HF	376	105	13403	0.0665	117.22	104.05	0.32336	5.1859	0.11632	1900	
C 4.35<d<4.5	396	107	5836	0.1640	118.30	108.40	0.31188	4.9902	0.11605	1896	
A133-Saunakangas, Joroinen, granodiorite											
A d>4.5	253	68.6	2553	0.3150	120.08	103.73	0.31337	5.0046	0.11583	1893	
B 4.35<d<4.5/Ø>160/HF	385	106	5365	0.1783	118.34	102.30	0.31952	5.1074	0.11594	1894	
C 4.35<d<4.5/Ø>160	413	109	4184	0.2310	118.73	103.50	0.30642	4.8845	0.11562	1889	
A579-Susineva I, granodiorite											
A d>4.3, Ø>160	463.1	115.0	4430	0.2110	117.39	100.58	0.2870	4.532	0.1145	1872	
B 4.2<d<4.3	757.8	169.8	2440	0.4013	119.52	105.41	0.2590	4.074	0.1141	1865	
C 4.0<d<4.2	1022	208.1	1668	0.5918	121.63	116.57	0.2352	3.684	0.1136	1858	

1) All fractions are zircon unless otherwice indicated. Borax = old analyses, borax-fusion method.
2) Corrected for blank, U-blank ca. 0.5 ng, Pb-blank ca. 1 ng.
3) Corrected for blank and common lead according to the age.
4) 2-sigma error is 0.15—0.3 %, unless otherwise indicated.
5) recalculated from A. Huhma (1976).

Table 2. Summary of U-Pb zircon ages with references to the data.

Sample		Rock type	Age ± 2s (Ma)	Lower intercept (Ma)	N	MSWD	References Age	Geologic	Comments
A527	Nattanen	granite	1772 ± 10	129	19		1	4	concordant monazite
A168	Vainospää	granite	1790 ± 22	350	16		2,1	2	concordant titanite
A126	Hatajavaara	granite	x				3	3	
A181	Peltovuoma	granite (Hetta)	1810 ± 14	345	7		1	4	
A368	Rovaniemi mlk	granite	1770 ± 8	251	3		3	3	2 zircons + titanite
A145	Molkoköngäs	granite	1843 ± 23	92	7	10	5	4	Age = 1870 ± 30 using five most concordant fractions
A184	Kotivaara	granite (Tepasto)	1798 ± 6	288	6		1	4	(Nd from A429)
A428	Lappalaislampi	granite (C.Kittilä)						4	titanite 1760 Ma
A748	Iso Nilipää	granite	2136 ± 16	365	3	0.002	5	4	
A453	Syrjäsalmi	granite (Puruvesi)	1797 ± 19	229	4		7	18	concordant monazite
A336	Närsäkkälä	quartz diorite	1884 ± 19	141	4		7	18	
A24	Parkkolansaari	granite/granodiorite (Suvasvesi)	1870 ± 4	193	4	0.2	8,5	8	
A25	Viitalahti	trondhjemite/ granodiorite (Kermajärvi)	1871 ± 5	107	3	0.004	8,5	8	
A60	Vihtajärvi	granodiorite (Maarianvaara)	1857 ± 8	205	4	0.8	8,5	8	concordant monazite
A63	Jussilansaari	granite/granodiorite (Onkivesi)	x						
A239	Molkanjärvi	quartz diorite	1882 ± 4		3				close to concordia
A875	Märaskär	granite (Hanko)	1830 ± 10	142	3	0.2	5	17	monazite + 2 zircons
A933⁺	Orijärvi	granodiorite	1891 ± 13	340	6	1	5	6	
A82	Hankavesi	granite	1886 ± 10	263	4	2	5	9	
A85	Kiukaanniemi	granodiorite	1883 ± 17	140	5	6	5	9	
A447	Keuruu	granodiorite	1883 ± 14	150	4	3	5	19	
A545	Lehesvuori	granite	1889 ± 15	145	3	0.7	5	20	
A93	Saunakangas	trondhjemite/tonalite	1903 ± 10	192	6	2	5	21	
A579	Susineva	granodiorite (Rautio batholith)	1883 ± 10	108	3	0.002	5	11,20	
A398	Oravaara	basalt	2105 ± 15		3		5	12	close to concordia
A235*	Horsmanaho	gabbro	1972 ± 18	657	4	2	13,5	13	(Nd from Oku602)
A729	Jormua	gabbro	1970 ± 20	564	2		22	22	
A542	Laukunkangas	norite	1880 ± 3		3		5	14	close to concordia
A277°	Soukkio	gabbro (Mäntsälä)	1870 ± 8	292	5	3	5	15	

Ages are upper intercept ages on the concordia diagram, exept A239, A398 and A542. x, scattered zircon data.*, includes samples A149, A234, A235.⁺, includes A933 and A668.°, includes A277 and A278. N = number of fractions. MSWD = Mean Square of Weighted Deviates, see text on p. 14. References: 1, Kouvo *et al.* (in prep.); 2, Meriläinen (1976); 3, Lauerma (1982); 4, Mikkola (1941) 5, Huhma (this study); 6, Eskola (1914); 7, Nykänen (1983); 8, Huhma, A. (1976); 9, Sjöblom (1984; 11, Gaál and Isohanni (1979); 12, Nykänen 1968); 13, Koistinen (1981); 14, Grundström (1980); 15, Härme (1978); 17, Sederholm (1926); 18, Nykänen (1975); 19, Marmo (1963); 20, Nurmi *et al.* (1984); 21, Pekkarinen and Hyvärinen (1984); 22, Kontinen (in press).

Mafic rocks

The ages of mafic rocks range from 2.1 Ga to 1.87 Ga. The U-Pb age estimates for the mafic rocks are reliable since the zircon analyses are highly concordant (Figs. 3 and 4). The Jatulian basalts from Jouttiaapa (A1009) and Oravaara (A398, Fig. 3) represent mantle-derived material extruded during the rifting of the Archaean continent about 2.1 Ga ago. The age of the Jouttiaapa basalts was determined from the whole rock Sm-Nd isochron (Huhma, 1984).

Granitoids

The majority of the granitoids from southern and central Finland are synorogenic intrusions (1.87—1.89 Ga) , whereas the ages of the granites from northern Finland reported here are usually about 1.8 Ga. The majority of the zircons yield good chords, and no major inherited component can be found. The zircon crystals seldom have either rounded cores or distinct overgrowths. There are only two granites (A126-Hatajavaara, A63-Jussilansaari) from which the zircon data are scattered (Lauerma, 1982; Kouvo, pers.comm.), implying an inherited zircon component. Two granites from Lapland, i.e. Central Kittilä (A428) and Tepasto (A429), are very poor in zircon, but the U-Pb analyses on titanites and the intrusive nature of the granites (Mikkola, 1941) suggest an age of about 1.8 Ga. New analyses on the porphyritic Tepasto granite (A184-Kotivaara) confirm this age (Kouvo *et al.*, in prep.)

The zircon data on the granites in the *central Lapland granitoid area* are variable (Table 2). Lauerma (1982) reports rather scattered zircon data on fairly homogeneous fine-grained granites in the Salla area (including A126) and concludes that the majority of the zircons have a minimum age of about 2.2 Ga. Most of the zircon crystals in sample A126 are rounded, and some have a rounded core and a zoned

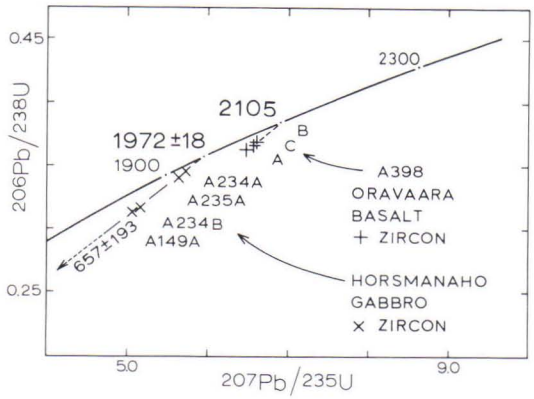

Fig. 3. Concordia plot of U-Pb zircon data on coarse Oravaara basalt (2.1 Ga) and Horsmanaho gabbro pegmatoid (1.97 Ga).

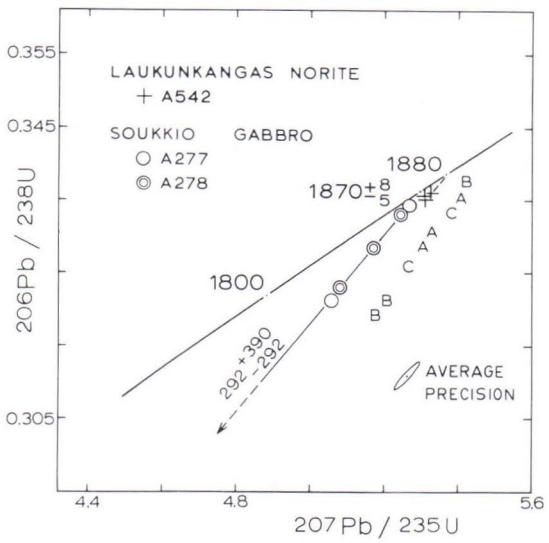

Fig. 4. Concordia plot of U-Pb zircon data on Laukunkangas norite (1.88 Ga) and Soukkio gabbro pegmatoid (1.87 Ga).

shell, indicating a multiphase crustal history. In fact three fractions from A126 give an apparent upper intercept age of about 2.53 Ga, whereas the lower intercept is about 1.4 Ga.

Seven zircon fractions from a typical pink granite at Molkoköngäs (A145, Fig. 5) in the

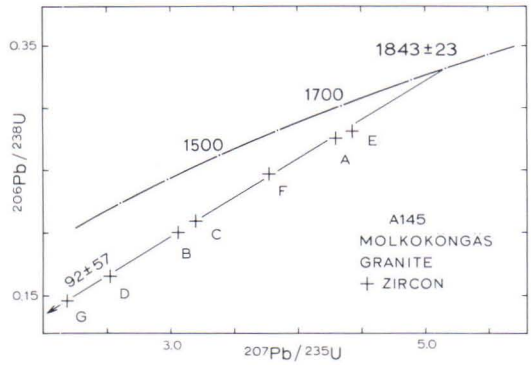

Fig. 5. Concordia plot of U-Pb zircon data on Molkokongäs granite, central Lapland granitoid area.

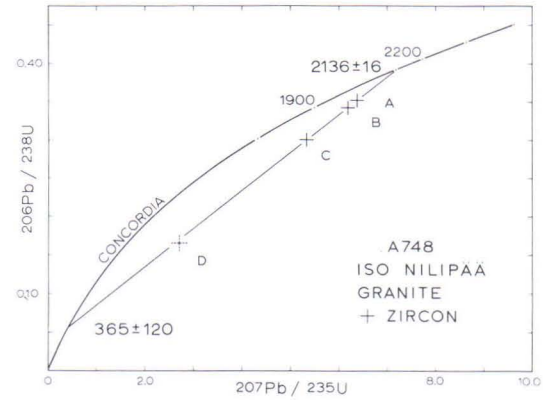

Fig. 6. Concordia plot of U-Pb zircon data on Iso Nilipää granite, central Lapland granitoid area.

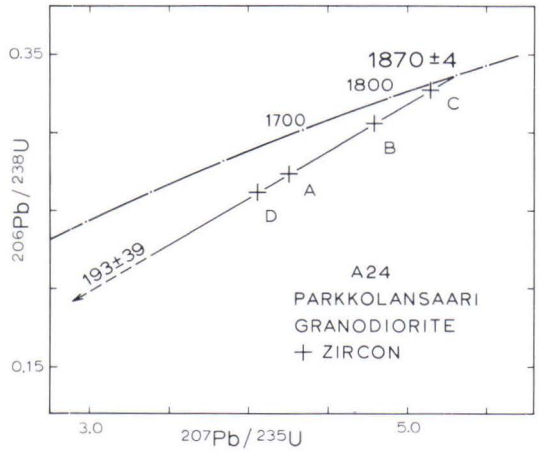

Fig. 7. Concordia plot of U-Pb zircon data on Parkkolansaari (Suvasvesi) granodiorite, Karelian domain.

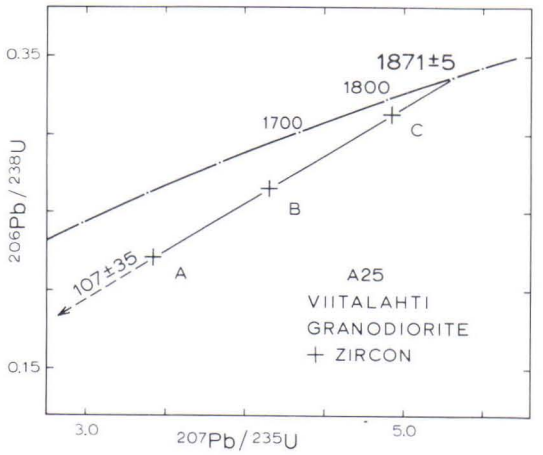

Fig. 8. Concordia plot of U-Pb zircon data on Viitalahti (Kermajärvi) granodiorite, Karelian domain.

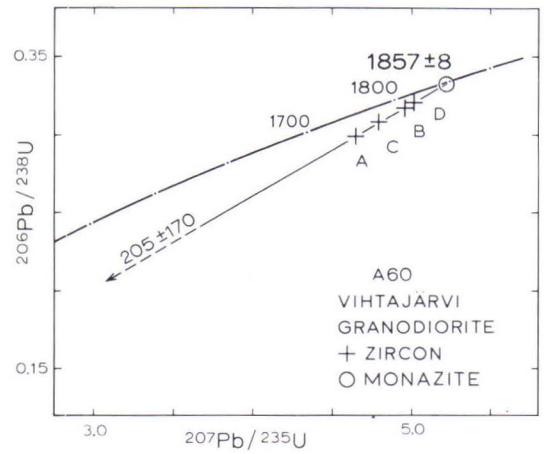

Fig. 9. Concordia plot of U-Pb zircon data on Vihtajärvi (Maarianvaara) granodiorite, Karelian domain.

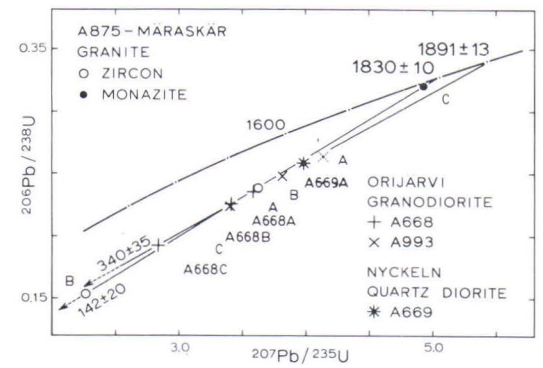

Fig. 10. Concordia plot of U-Pb zircon and monazite data on Märaskär (Hanko) granite (1.83 Ga) and Orijärvi granodiorite (1.89 Ga) including one fraction from Nyckeln quartz diorite, southern Finland.

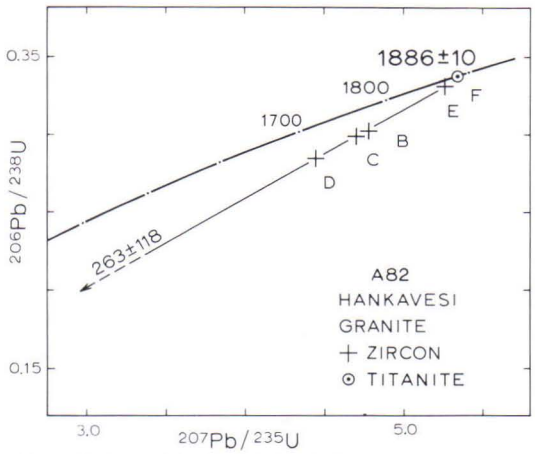

Fig. 11. Concordia plot of U-Pb zircon and titanite data on Hankavesi granite, central Finland granitoid area.

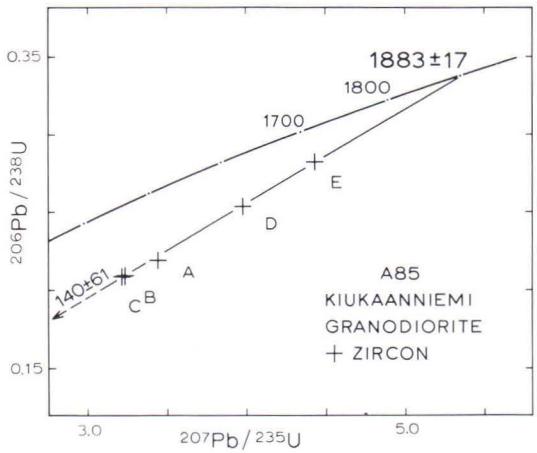

Fig. 12. Concordia plot of U-Pb zircon data on Kiukaanniemi granodiorite, central Finland granitoid area.

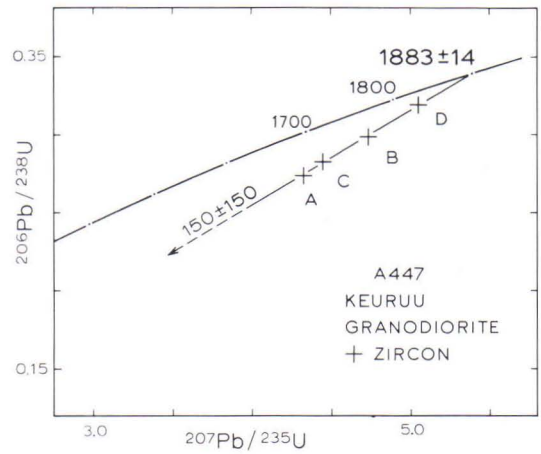

Fig. 13. Concordia plot of U-Pb zircon data on Keuruu granodiorite, central Finland granitoid area.

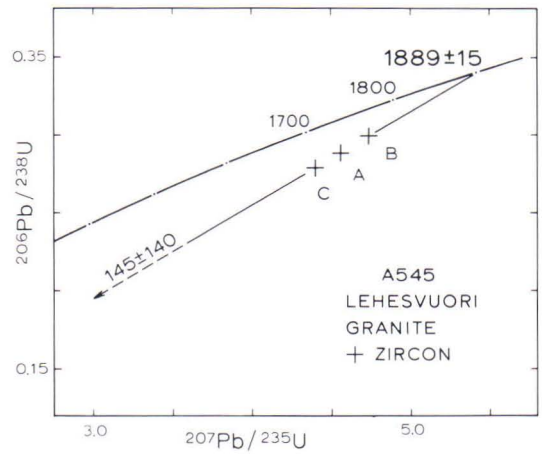

Fig. 14. Concordia plot of U-Pb zircon data on Lehesvuori granite, central Finland granitoid area.

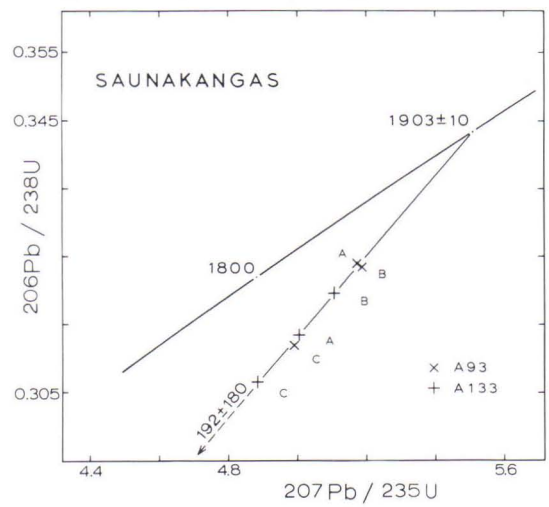

Fig. 15. Concordia plot of U-Pb zircon data on Saunakangas trondjhemite (A93) and granodiorite (A133).

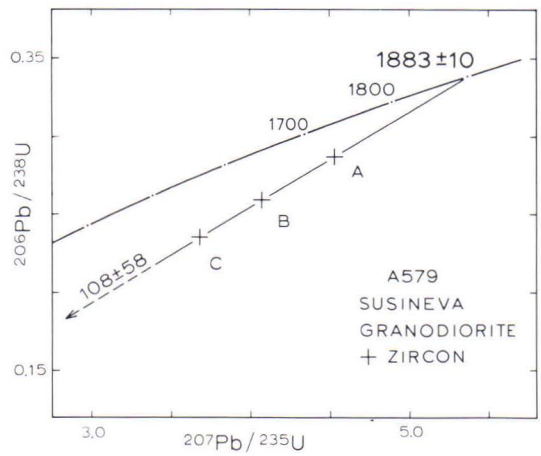

Fig. 16. Concordia plot of U-Pb zircon data on Susineva granodiorite, Rautio batholith.

central Lapland granitoid area define an upper intercept age of 1843 ± 23 Ma with a high MSWD (= 10, Mean Square of Weighted Deviates), indicating some extra scatter in addition to the experimental error, probably due to a small inherited zircon component. Regression of the five most concordant fractions yields an age of 1870 ± 30 Ma with MSWD = 11. The two zircon fractions and a concordant titanite from a pink microcline granite in Rovaniemi (A368) give a slightly younger age of 1.77 Ga (Lauerma, 1982).

An exceptional age of 2.14 Ga was obtained for a granite from Iso Nilipää (A748, Fig.6). The age of the zircons from A748, which are euhedral zoned crystals without any visible rounded cores, is consistent with the U-Pb age of titanite. The sample is microscopically somewhat cataclastic and contains a small amount of strongly coloured amphibole. If the zircon really dates the crystallization of the granite, it would imply that granites could have been formed during Jatulian magmatism, before the Svecokarelian orogenic stage. Hence, the central Lapland granitoid area is geologically and geochronologically heterogeneous (Lauerma, 1982), but the widespread, pink microcline granites presumably crystallized finally during the late-orogenic stage of the Svecokarelian orogeny.

The age of 1.8 Ga for the Hetta granite reported here (A181-Peltovuoma) is well defined by seven zircon fractions. Other samples from the Hetta-type granite, however, contain mixed zircon populations with inherited zircons, suggesting a, partially at least, older crustal source lithology (Kouvo *et al.*, in prep., see also Lehtonen, 1984). This is to be expected since the granite is often migmatitic and contains assimilated fragments of schists and gneisses (Mikkola, 1941).

The migmatizing Hanko granite (A875-Märaskär, Fig. 10) in *southern Finland* is about 1.83 Ga old and clearly younger than the roughly 1.9 Ga synorogenic phase. This age is consistent with other zircon analyses from related microcline granites (e.g., Vaasjoki, 1977; Hopgood *et al.*, 1983; Kouvo, unpublished data). The development and ages of a similar migmatite complex have been described by Hopgood *et al.* (1983), who emphasize the relatively short time-span during the major tectonic and metamorphic evolution of the migmatites. Another younger granite in the present study is the 1.8 Ga old Puruvesi granite (A453-Syrjäsalmi, Nykänen, 1983), which is situated close to the Raahe—Ladoga lineament zone.

Four granitoids (A82-Hankavesi, A85-Kiukaanniemi, A447-Keuruu, A545-Lehesvuori) from the granitoid area of central Finland have zircon ages of about 1.885 Ga (Figs. 11—14), which is close to the average U-Pb zircon age for the Svecokarelian synorogenic rocks (e.g. Neuvonen *et al.*, 1981). One of these zircon ages, for the granite from Hankavesi (A82), is confirmed by the concordant 1.88 Ga titanite (Fig. 11). This is contrary to the situation at Pihtipudas (Fig. 1) and Viitasaari, some 120 km northeast, close to the Raahe—Ladoga zone, where granitoids with 1.88 Ga zircons contain 1.8 Ga titanites and monazites (Aho, 1979). This suggests different tectonothermal histories for these bodies, and probably for larger blocks of the crust as well.

The three granodiorites from the Karelian domain in *eastern Finland* are significantly younger. They are also characterized by long, light-coloured zircon crystals (Kouvo, 1958). Based on one zircon analysis per sample, A. Huhma (1976) reported an age of about 1.9 Ga for these rocks, one of which (A60) was included in the pioneering study by Kouvo (1958). The analyses of zircon fractions yield an age of 1.87 Ga for the Suvasvesi (A24-Parkkolansaari, Fig. 7) and Kermajärvi (A25-Viitalahti, Fig. 8) granodiorites, while the granodiorite at Maarianvaara (A60-Vihtajärvi, Fig.9) is 1.857 ± 8 Ga old. The zircon age for the granodiorite at Vihtajärvi is confirmed by

concordant monazite. This is close to the roughly 1.84 Ga Rb-Sr muscovite age for a pegmatite at Sivakkavaara, about 5 km from sample A60 (van Breemen and Bowes, 1977; recalculated using the decay constant of Steiger and Jäger, 1977). Data suggest that these granodiorites were emplaced slightly later than most Svecokarelian synorogenic intrusions in southern and central Finland, e.g. the "late" microcline-rich granite (A545-Lehesvuori, T = 1889 ± 15 Ga, Fig. 14) in the granitoid area of central Finland.

The data available support the observation that the 1.80 —1.83 Ga postorogenic and late-orogenic granites may be absent from large areas of central Finland, apparently only occurring in a belt in southern Finland and in Lapland. An example of this late thermal activity in SE Finland is the Sulkava area, where zircon and monazite from granitic neosome and paleosome yield U-Pb ages of about 1.8—1.83 Ga (Korsman *et al.*, 1984). Small 1.8 Ga old post-orogenic granodiorite intrusions also exist in the same area (op.cit.).

Nd results

The Sm-Nd isotopic data are given in Tables 3 and 4, and the initial $^{143}Nd/^{144}Nd$ ratios are plotted as $\epsilon_{Nd}(T)$ in Fig. 17, which also shows the CHUR (= chondritic uniform reservoir, $^{147}Sm/^{144}Nd = 0.1966$) and a depleted mantle evolution line according to DePaolo (1981).

Mafic rocks

A well defined initial value has been observed for the Jatulian basalts from the Jouttiaapa formation (A1009). Ten samples define a Sm-Nd isochron age of 2.09 ± 0.07 Ga, with $\epsilon_{Nd} = +4.2 \pm 0.5$ (Huhma, 1984).

Table 3. Sm-Nd isotopic data for the mafic rocks

Sample		Rock type/ Mineral	Age (Ga)	Sm (ppm)	Nd (ppm)	(a $^{147}Sm/^{144}Nd$	(b $^{143}Nd/^{144}Nd$	(c ϵ_{Nd}	(b $^{145}Nd/^{144}Nd$
A398	Oravaara	basalt	2.10	1.89	5.20	0.2193	0.513085 ± 35	+2.6 ± 0.8	0.348418 ± 34
Oku602	Horsmanaho	gabbro	1.97	1.21	3.24	0.2255	0.513171 ± 30	+3.1 ± 0.7	0.348432 ± 30
A729b	Jormua	gabbro	1.97	0.76	2.00	0.2274	0.513200 ± 46	+3.2 ± 1.0	0.348407 ± 35
A542	Laukunkangas	norite	1.88	6.39	31.55	0.1224	0.511729 ± 20	+0.2 ± 0.5	0.348418 ± 13
S	Salittu	peridotite	(d 1.89	2.44	9.38	0.1571	0.512307 ± 32	+3.1 ± 0.7	0.348418 ± 14
A1009	Jouttiaapa	basalt	2.09	± 0.07	(Sm-Nd	whole	rock isochron)	+4.2 ± 0.5	
A277	Soukkio	gabbro	1.87						
		hornblende		10.37	51.74	0.1211	0.511820 ± 40	+2.2 ± 0.9	
		apatite		142.0	719.4	0.1193	0.511807 ± 30	+2.3 ± 0.7	
		plagioclase		0.2198	1.971	0.06738	0.511150 ± 40	+2.0 ± 0.9	
		zircon		6.25	9.20	0.4107	—		
RF2748	Evijärvi	basalt	(d 1.9	2.46	6.91	0.2151	0.513020 ± 96	+2.9 ± 2.0	0.348430 ± 60
RF2757	Evijärvi	basalt	(d 1.9	2.06	5.87	0.2123	0.513066 ± 44	+4.5 ± 1.0	0.348412 ± 31

(a, Estimated error is 0.4%.
(b, Errors are 2 SEM in last significant digits. Values are normalized to $^{146}Nd/^{144}Nd = 0.7219$.
(c, $\epsilon_{Nd}(T) = 10^4 \times [^{143}Nd/^{144}Nd_{)Sample\ at\ T}/^{143}Nd/^{144}Nd_{)CHUR\ at\ T} - 1]$.
(d, ages inferred from the geological data.

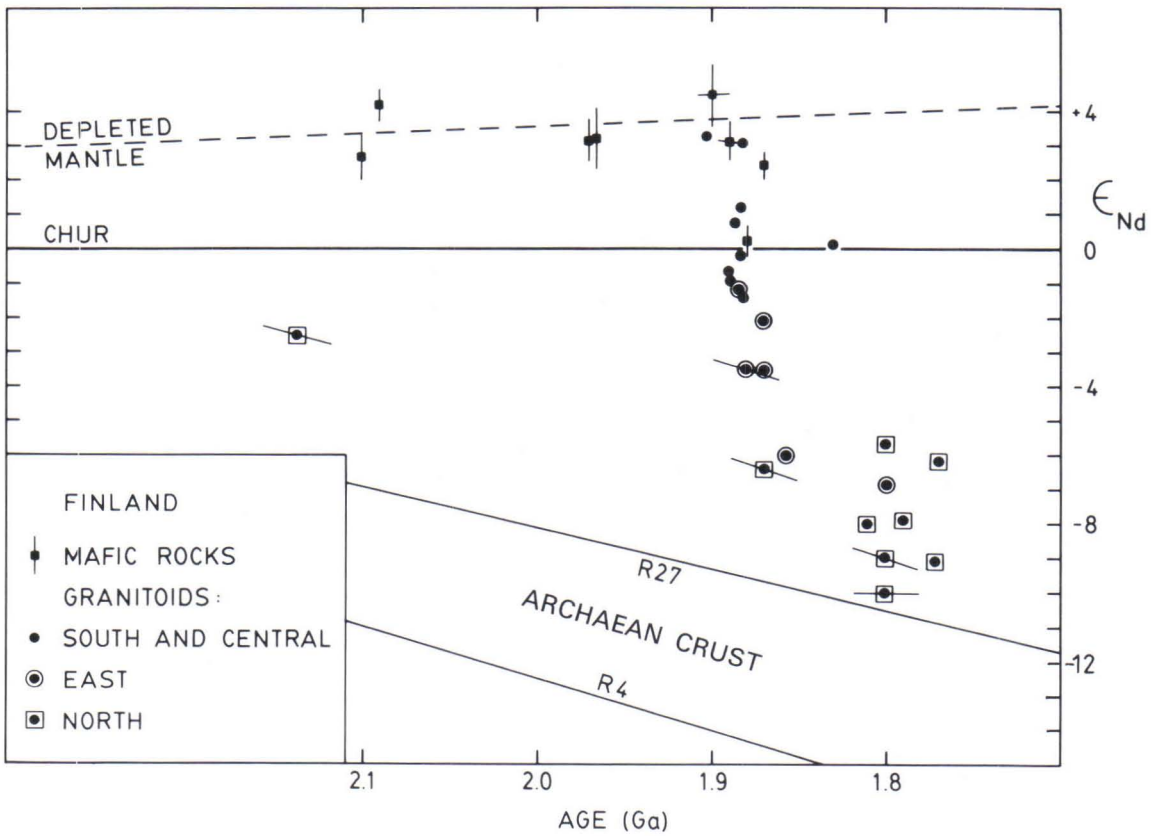

Fig. 17. ϵ_{Nd} vs. age plot showing the data of this study. Different symbols are used for mafic rocks and for granitoids from the Svecofennian domain (south and central), and from the Karelian domain in eastern and northern Finland. CHUR = evolution line of undifferentiated Earth. Depleted mantle is according to DePaolo (1981): $\epsilon_{Nd}(T) = 0.25xT^2 - 3xT + 8.5$. A part of the evolution line calculated for granitoids of somewhat uncertain zircon age is also shown. 2-sigma error bars are shown for analyses of mafic rocks. The evolution lines of two Archaean granitoids are also displayed (R4, R27).

The two gabbros (Oku602, A729) associated with the ophiolites are slightly depleted in LREE (Table 3), and give $\epsilon_{Nd} = +3.2 \pm 0.7$ 1.97 Ga ago. The two tholeiitic basalts from Evijärvi (RF2748, RF2757) also have a relatively primitive chemical composition, the Sm/Nd ratio implying a pattern slightly depleted in LREE. The age of these rocks has not been defined by radiometric methods, but they belong to the Svecofennian schist belt of Pohjanmaa, which contains abundant turbiditic metagreywackes and metavolcanic rocks presumably about 1.9 Ga old. The $\epsilon_{Nd}(1.9) = +4.5 \pm 1.0$ is not much dependent on the age used in the calculation. Other synorogenic mafic rocks in this study are chemically more evolved and the initial ratios are variable. No age data are available from the Salittu (sample S) peridotite massif, but geological evidence supports a normal synorogenic age (Eskola, 1914). An initial value of $\epsilon_{Nd}(1.89) = +3.1$ was observed, but the Sm/Nd ratio shows a pattern unexpectedly enriched in LREE. The Sm-Nd mineral

isochron age (apatite, hornblende, plagioclase) for the Soukkio gabbro (A277) is 1.91 ± 0.13 Ga (Fig. 18), which is equal to the zircon age of 1.87 Ga (Fig. 4) and gives $\epsilon_{Nd}(1.87) = +2.3$.

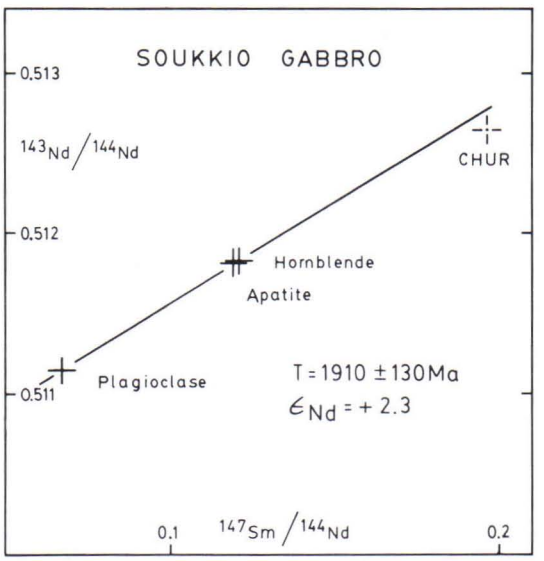

Fig. 18. $^{143}Nd/^{144}Nd$ versus $^{147}Sm/^{144}Nd$ diagram for minerals from the Soukkio gabbro (A277). CHUR = undifferentiated earth.

Granitoids

The mafic samples described above represent mantle-derived material that was involved in the generation of the Svecokarelian crust. Another potential source of the Svecokarelian magmas could have been Archaean crustal material. Two analyses on Archaean granitoids, a Runkaus granite from the Peräpohja area (R4, Perttunen, 1985) and a granodioritic gneiss from the Sotkuma dome (R27; Kouvo, 1958; A. Huhma, 1975), eastern Finland (Fig. 2), are given in Table 4. Figure 17 displays the neodymium evolution paths of these samples. The calculated $\epsilon_{Nd}(1.8)$ values are −15.6 and −10.5, which are typical of the Archaean material with an LREE-enriched pattern (e.g., McCulloch and Wasserburg, 1978).

The Sm-Nd isotopic data on the granitoids are listed in Table 4 and displayed on a diagram of ϵ_{Nd} vs age in Fig. 17. The granitoids with a somewhat uncertain zircon age (except A748) also show part of their evolution according to the $^{147}Sm/^{144}Nd$ ratio measured (Fig. 17).

All nine granites from northern Finland have relatively unradiogenic initial Nd-isotopic compositions, the $\epsilon_{Nd}(1.8)$ being from −5.7 to −11. Two of these analyses (A527, A168) have been reported by Kouvo et al. (1983). The T_{DM} model ages are about 2.4—2.5 Ga. The mean $^{147}Sm/^{144}Nd$ ratio of eight samples is 0.083, which is slightly lower than the average for Phanerozoic granites (e.g., McCulloch and Chappell, 1982). Sample A428 (Lappalaislampi) from the Central Kittilä granite has a roughly chondritic Sm/Nd ratio (Fig. 19) and a low zircon content. This may be attributed to the compatible behaviour of LREE in the generation of the felsic magma (Miller and Mittlefehldt, 1982), an explanation that is supported by the low concentration of Ti and P (analyses of Rastas, pers.comm.). The low REE and zircon abundances in the even-grained Tepasto granite (A429b-Kuivavaara) may also be related to this phenomenon. In other words, the residue in the melting contained accessory phases, such as allanite, monazite and zircon or, alternatively the fractionation of these phases took place before the crystallization of the granite.

Seven granitoids from the Karelian domain in eastern Finland, to the east of the Raahe—Ladoga geosuture zone, also have negative epsilon values, quartz diorites (A336b- Närsäkkälä, A239b-Molkanjärvi) having the highest $\epsilon_{Nd} = -1.2$. The mean $^{147}Sm/^{144}Nd$ ratio is 0.091 and the T_{DM} model ages range from 2.5 to 2.2 Ga.

Six granitoids from the Svecofennian domain have $\epsilon_{Nd}(T)$ close to zero. $\epsilon_{Nd}(T) = +3$ have been obtained for the trondhjemite (A93) situated a few tens of kilometres west of

Table 4. Sm-Nd isotopic data for the granitoids.

Sample		Age (Ga)	Sm (ppm)	Nd (ppm)	a $^{147}Sm/^{144}Nd$	b $^{143}Nd/^{144}Nd$	c $\epsilon_{Nd}(T)$	T_{CHUR} (Ga)	T_{DM} (Ga)	a,d Control ratio
Northern Finland (Karelian domain)										
A527/12	Nattanen	1.77	9.36	68.61	0.08249	0.510842 ± 24	−9.1 ± 0.5	2.39 ± 0.04	2.55	0.24157 ± 2
A168	Vainospää	1.79	6.73	52.60	0.07727	0.510830 ± 14	−7.9 ± 0.4	2.30 ± 0.03	2.47	0.24157 ± 2
A126	Hatajavaara		3.43	27.99	0.0740	0.510731 ± 34 °	−9.0 ± 0.7	2.36 ± 0.05	2.52	0.241566 ± 27
A181	Peltovuoma	1.81	3.47	25.73	0.08153	0.510860 ± 46	−8.0 ± 1.0	2.35 ± 0.06	2.51	0.24158 ± 5
A368	Rovaniemi ltk	1.77	3.95	26.69	0.08942	0.511075 ± 96	−6.2 ± 2.0	2.22 ± 0.14	2.41	0.24157 ± 8
A145	Molkokongäs	1.87	3.25	29.61	0.06636	0.510710 ± 31	−6.4 ± 0.7	2.25 ± 0.05	2.41	0.348417 ± 16
A429b	Kuivavaara	1.8	1.28	7.54	0.1024	0.511230 ± 28	−5.7 ± 0.6	2.27 ± 0.05	2.48	0.348391 ± 17
A428b	Lappalaislampi	1.8	5.71	17.49	0.1974	0.512131 ± 46	−9.9 ± 1.0			0.348510 ± 32
*			5.77	17.67	0.1974	0.512076 ± 56	−11 ± 1.2			0.348435 ± 28
A748b	Iso Nilipää	2.14	3.96	26.40	0.09075	0.511021 ± 20	−2.5 ± 0.5	2.32 ± 0.04	2.50	0.348413 ± 17
Eastern Finland (Karelian domain)										
A453b	Syrjäsalmi	1.80	6.28	42.24	0.08987	0.511021 ± 38	−6.9 ± 0.8	2.30 ± 0.06	2.50	0.241596 ± 30
A336b	Närsäkkälä	1.88	3.19	19.63	0.09811	0.511357 ± 24	−1.2 ± 0.5	1.98 ± 0.04	2.22	0.241556 ± 30
A24b	Parkkolansaari	1.87	3.66	22.75	0.09710	0.511243 ± 34	−3.4 ± 0.7	2.13 ± 0.06	2.35	0.241592 ± 27
A24b	'' (small)	1.87	4.08	27.89	0.08835	0.511125 ± 26	−3.6 ± 0.6	2.13 ± 0.04	2.33	0.241609 ± 24
A25b	Viitalahti	1.87	5.12	40.44	0.07674	0.511056 ± 22	−2.1 ± 0.5	2.01 ± 0.04	2.20	0.348407 ± 16
*			5.18	41.06	0.07625					
A63	Jussilansaari		3.70	29.15	0.07672	0.510978 ± 42 °	−3.6 ± 0.9	2.11 ± 0.06	2.29	0.348429 ± 21
A239b	Molkanjärvi	1.88	4.27	22.00	0.1173	0.511586 ± 39	−1.4 ± 0.9	2.02 ± 0.07	2.30	0.241586 ± 40
A60	Vihtajärvi	1.86	5.88	39.62	0.08964	0.511023 ± 35	−6.0 ± 0.8	2.29 ± 0.05	2.48	0.348412 ± 26
Southern and central Finland (Svecofennian domain)										
A875	Märaskär	1.83	9.29	57.47	0.09778	0.511453 ± 22	+0.1 ± 0.5	1.83 ± 0.04	2.08	0.241576 ± 25
A933	Orijärvi	1.89	4.39	22.22	0.1175	0.511618 ± 50	−0.7 ± 1.1	1.96 ± 0.09	2.25	0.348423 ± 42
A82	Hankavesi	1.89	4.92	27.82	0.1068	0.511559 ± 18	+0.7 ± 0.5	1.83 ± 0.04	2.11	0.348412 ± 11
A85	Kiukaanniemi	1.88	8.50	55.40	0.09278	0.511412 ± 21	+1.2 ± 0.5	1.80 ± 0.04	2.05	0.348421 ± 22
A447b	Keuruu	1.88	3.72	19.00	0.11817	0.511658 ± 20	−0.2 ± 0.5	1.90 ± 0.04	2.20	0.348415 ± 20
A545	Lehesvuori	1.89	7.82	32.91	0.14367	0.511934 ± 34	−0.9 ± 0.7	2.02 ± 0.10	2.41	0.348407 ± 30
*			7.79	32.78	0.14356					
A93	Saunakangas	1.90	6.81	40.84	0.1007	0.511609 ± 20	+3.3 ± 0.5	1.64 ± 0.04	1.93	0.348416 ± 20
A579	Susineva	1.88	2.54	17.09	0.08986	0.511468 ± 33	+3.0 ± 0.7	1.67 ± 0.05	1.93	0.348434 ± 41
*			2.45	16.56	0.08930		+3.2			
Archaean										
R4	Runkaus, granite		4.24	34.08	0.0752	0.510404 ± 28 °	−15.6 ± 0.5	2.79 ± 0.04	2.91	0.241601 ± 21
R27	Sotkuma, gneiss		1.98	11.19	0.1067	0.511039 ∓ 60 °	−10.5 ± 1.2	2.70 ± 0.10	2.86	0.348457 ± 39

a, b, c refer to Table 2. d, control ratio either $^{148}Nd/^{144}Nd \approx 0.24157$ or $^{145}Nd/^{144}Nd \approx 0.348418$.
°, ϵ calculated at 1.8 Ga
*, duplicates
(small), small 10 g sample from the same locality. Model ages T_{CHUR} and T_{DM} were calculated according to evolution of chondrites (Jacobsen and Wasserburg, 1980) and depleted mantle by DePaolo (1981), see text.

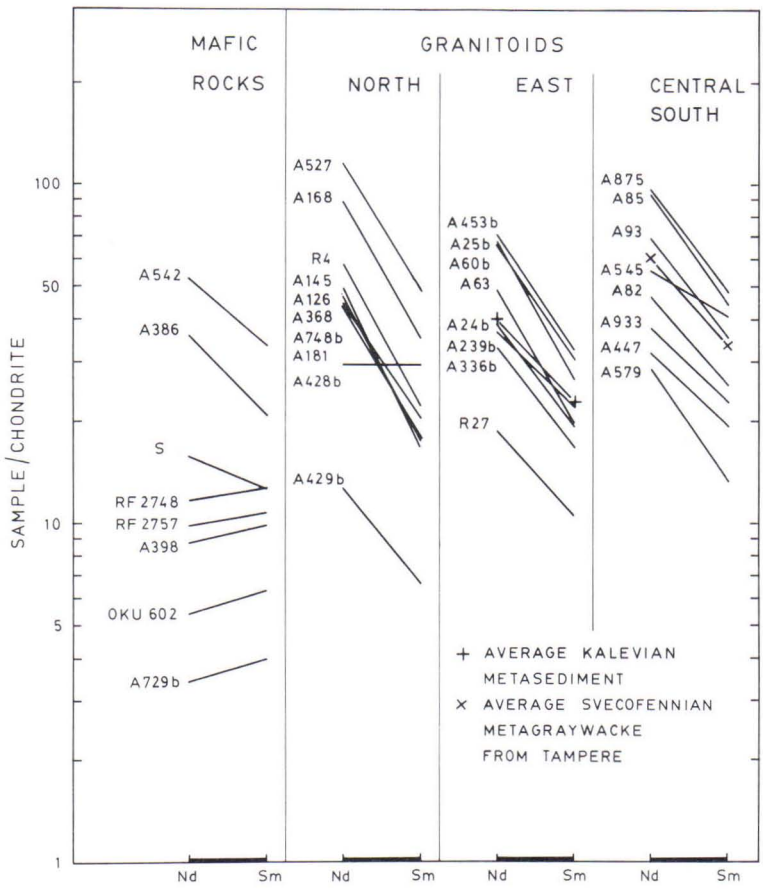

Fig. 19. Chondrite-normalized Sm-Nd concentration patterns for the samples in this study. The average concentrations for the metasediments and the data on A386, which is a metadacite from the Tampere schist belt, are from Huhma (in press). Values used for normalization are: Nd = 0.597 ppm, Sm = 0.192 ppm.

the Raahe—Ladoga zone, and for the granodiorite from the I-type Rautio batholith (A579). The mean $^{147}Sm/^{144}Nd$ for the samples in the Svecofennian domain is 0.108. The $^{147}Sm/^{144}Nd$ ratio of 0.143 of the Lehesvuori microcline granite is high (A545, Fig. 19), and results in the high T_{DM} model age of 2.41 Ga. Other T_{DM} ages range from 2.25 to 1.93 Ga.

EARLY PROTEROZOIC MANTLE IN THE BALTIC SHIELD

Mafic rocks before the synorogenic stage

Much of present-day continental crustal growth takes place near the zones of subduction, where mantle material contributes significantly to the formation of new crust. It is therefore important to know the chemical characteristics of the contributing mantle in order to study the origin and evolution of the continents. The results given here from relatively primivite mantle-derived material show that a LREE- depleted asthenospheric mantle reservoir, similar to the source of mid-ocean ridge basalts today, existed beneath the Baltic

Shield during Early Proterozoic and Archaean times (Fig. 17). The highest and most precise initial $^{143}Nd/^{144}Nd$ ratio ($\epsilon_{Nd}(2.09) = +4.2$) is above the depleted mantle evolution curve by DePaolo (1981). The new model by Nelson & DePaolo (1984) may be more appropriate, if any globally uniform asthenospheric mantle reservoir, existed at all. Other mafic samples older than 1.9 Ga plot slightly below (but are equal within error to) the DM model curve. Gabbros Oku602 and A729 have a very low REE concentration with a slight LREE depletion (Fig. 19) and $\epsilon_{Nd}(1.97) = +3.2$, supporting the suggestion that they could well belong to the ophiolite assemblages (Koistinen, 1981; Kontinen and Meriläinen, in prep.).

On the other hand, there are some Jatulian mafic rocks, 2.2—2.4 Ga old, with much lower $\epsilon_{Nd}(T)$ (Huhma, in press; and in prep.), but they are all enriched in LREE. They may have been contaminated by the Archaean lower or upper crust, or they may have a different upper mantle source with $\epsilon_{Nd}(T)$ about 0. The distinction between mantle heterogeneity and crustal contamination is a controversial issue and difficult to verify (e.g., Carlson *et al.*, 1981; DePaolo, 1983; Chauvel *et al.*, 1985). Mantle xenoliths, which have been considered to derive from the continental lithospheric mantle, are chemically heterogeneous. Their isotopic composition often indicates long-term LREE enrichment, exhibiting negative initial $\epsilon_{Nd}(T)$ ratios (e.g., Cohen *et al.*, 1984). Whatever the explanation for this enrichment, the consequence of a significant contribution of such old lithospheric mantle material to the genesis of volcanic rocks, and of new crust in general, would be similar to that of crustal contamination. Because Jatulian igneous rocks occur in the proximity of Archaean crust, old lithospheric mantle (and crust) was available. If it did contribute as intimated above, $\epsilon_{Nd}(2.2—2.4) = 0$ for the Jatulian LREE-enriched basalts can be readily explained.

In addition to the high initial neodymium ratios and slightly LREE depleted chemistry, the 1.97 Ga ophiolitic gabbros (Oku602-Horsmanaho and A729-Jormua) and the 2.1 Ga old basalt of Oravaara (A398) are characterized by the high thorogenic lead abundances of the zircon. The radiogenic $^{208}Pb/^{206}Pb$ ratio of zircon is about 2 for the basalt of Oravaara and about 0.7 for the gabbros, in contrast to a ratio of about 0.1 for most of the samples included in this study and zircon in general. For closed zircon systems, these ratios correspond to Th/U ratios of 7, 2.5 and 0.4, respectively. Both Th and U are highly incompatible elements in the generation of basaltic magmas, and the Th/U ratio in the melt should not differ much from that in the source. The Th/U ratio in basalts and in many igneous rocks is about 3.5 to 4 (Sun, 1980) in contrast to less than 1 for the zircon in a great variety of rocks (e.g., Ahrens et al., 1967; this study). The zircon in the basic magma has usually crystallized during the late stage, after the saturation of the liquid with Zr and SiO_2. Since it is unlikely that a rather primitive magma had a high Th/U ratio (see discussion in Sun, 1980), the very high thorite component in zircon is probably a consequence of conditions during crystallization. (Thorite was not detected by X-ray diffraction).

Synorogenic 1.9 Ga old mafic rocks

The Nd-isotopic compositions for 1.9 Ga old synorogenic mafic rocks vary, the $\epsilon_{Nd}(1.9)$ being from 0 to +4.5 (Fig. 20). The highest value ($+4.5 \pm 1.0$), which was obtained from LREE-depleted basalts, is similar to values reported from the USA (DePaolo, 1981,

Nelson and DePaolo, 1984), South Greenland (Patchett and Bridgwater, 1984) and Kiruna, Sweden (Skiöld and Cliff, 1984) for Proterozoic tholeiites. Lower $\epsilon_{Nd}(T)$ ratios, which are generally associated with LREE enriched rocks, could indicate long-term heterogeneity in the mantle or, what is more likely, admixture of an LREE-enriched crustal component with mafic mantle-derived magma during the orogeny. As was discussed above, it is difficult to distinguish between these alternatives. The existing isotopic data would, however, suggest that very old lithospheric mantle was probably not available beneath the Svecofennian domain. On the other hand, isotopic and trace element data from ocean island basalts indicate that reservoirs in which Nd is less radiogenic than in MORB sources have existed in areas where old lithosphere was unlikely present (e.g., White, 1985).

The Sm/Nd ratios of many synorogenic mafic rocks indicate a chemical composition enriched in incompatible elements (e.g., Fig. 19). In the case of gabbroic rocks, this is partly due to crystal fractionation and cumulation, but partly also due to a source effect, as suggested also by the high P concentrations (17 times chondritic) and high P/Ti ratio in the peridotite at Laukunkangas (Grundström, 1980). In terms of the partitioning of Sm and Nd alone, the LREE-enriched pattern with relatively high REE abundances could be produced from the mantle by a small degree of partial melting combined with the fractionation processes (Gast, 1968). In the case of A542-Laukunkangas, derivation from spinel lherzolite with unfractionated 1.5 x chondritic REE abundances would involve a small amount of partial melting (5%) with clinopyroxene in the residue, associated with subsequent large amounts (70%) of olivine, clinopyroxene and plagioclase fractionation (using batch melting and Rayleigh fractionation trace element modelling and parameters by Hanson, 1978; see also DePaolo, 1983). Direct derivation from an LREE-depleted source by a single- stage melting process requires less than 1% partial melting and would probably be almost impossible.

Enrichment might be easier to explain by multistage melting processes, and if all these processes took place shortly before the eruption, the original Nd isotopic composition of the mantle would be recorded. Although fractionated mafic rocks could thus be pure derivatives from mantle sources, lithospheric contamination would be more likely than for the more primitive samples. Consequently, rocks with strongly LREE-enriched patterns are probably not good candidates for recording the Nd isotopic composition of major convecting mantle reservoirs. If local disequilibrium melting is significant, the most primitive samples might indicate the Nd isotopic ratio of the source more successfully than the samples that represent only a small degree of partial melting. This should not be a problem for partially molten mantle, however since the mantle would tend to equilibrate locally in less than 10^6 years (Hofmann and Hart, 1978).

Although the contribution of mantle sources with $\epsilon_{Nd}(1.9)$ lower than $+4$ cannot be excluded, it is most probable that the material from a LREE-depleted asthenospheric mantle with $\epsilon_{Nd}(1.9)$ from $+4$ to $+5$ played a major role in the formation of the Svecokarelian crust. This is supported by the $\epsilon_{Nd}(T) = +3$ from some Svecokarelian granitoids (this study; Patchett and Kouvo, 1986; Wilson *et al.,* 1985) and by data from the Ketilidian (Patchett and Bridgwater, 1984), which was formed in the proximity of the Svecokarelian crust. Rocks with LREE-enriched patterns and $\epsilon_{Nd}(1.9)$ lower than $+4$ most probably contain admixtures of continental crust, as has been proposed for recent island-arc volcanics (e.g., White and Patchett, 1984). The enriched mantle sources probably fed only insignificant amounts of exotic magmas such as some postorogenic granitoids and lamprophyres.

REWORKED VERSUS NEW COMPONENTS IN THE SVECOKARELIAN GRANITOIDS

The large range in ϵ_{Nd} values (from −10 to +3) in the Svecokarelian granitoids is attributed to the mixing of Archaean ($\epsilon_{Nd}(1.9)$ about −12) and juvenile mantle-derived material ($\epsilon_{Nd}(1.9)$ about +5) during the orogeny (Figs. 17 and 20). From the isotopic point of view, the mantle-derived material does not here necessarily mean a mantle origin at the time of granitoid formation; it only refers to the material that has REE and isotopic characteristics similar to the mantle. The orogenic rock associations with mantle-derived mafic lithologies, and the large quantity of heat required for producing granitic rocks, however, implies direct transfer of heat and magma from below the continental crust (Wyllie, 1984).

Karelian domain

A striking feature of all the granites in northern Finland included in this study is the major reworked Archaean crustal component in them. This was noted in the Nattanen (A527), Vainospää (A168) and Hetta (A181) granites back in the early seventies, when Pb isotopes were analysed from whole rocks and K-feldspar (Meriläinen, 1976; Kouvo *et al.*, 1983). The Hf isotopes also indicate a marked contribution of Archaean material to the Nattanen and Vainospää granites (Patchett *et al.*, 1981). The relatively unradiogenic Pb-isotopic composition of K-feldspars indicates that the lead was derived from a reservoir with a time-integrated low U/Pb-ratio, such as the lower crust (e.g. Doe and Zartman, 1979; Moorbath *et al.*, 1981). The high proportion of Nd of Archaean crustal derivation in the Nattanen granite is interesting when compared with the $\epsilon_{Nd}(1.9) = 0$ for the mafic granulites (Bernard-Griffiths *et al.*, 1984) that surround the Nattanen granite. Thus, the material from the (mafic) granulites did not contribute much to the formation of the Nattanen granite, and Archaean crust probably exists beneath these Early Proterozoic granulites.

Despite their old crustal precursors, obvious inherited zircons were found only from sample A126 in this study, implying that the provenance of the granites was at least partly Archaean (Lauerma, 1982). Watson and Harrison (1983) found that zircon solubility in crustal melts increases with increasing temperature, and that basic and peralkaline felsic compositions show higher zircon solubilities. The lack of inherited zircons could thus indicate that all the zircons of the source were dissolved by the granitic melts at relatively high temperature, probably at deep crustal levels (or, alternatively, that the source was poor in zircon). If melting takes place at lower temperatures, it is possible that only a fraction of the recycled zircon would dissolve in the melt to reach saturation, and that some of the zircon would be preserved. Zircons much damaged by radiation would not survive since recrystallization is plausible even at low temperatures (e.g., Gebauer and Grünenfelder, 1976; Krogh and Davis, 1975).

The Svecokarelian granitoids from the Karelian domain in eastern Finland also contain a substantial reworked component of LREE-enriched Archaean crust. The Puruvesi (A453) and Maarianvaara (A60) granites seem to have been formed largely from anatectic Archaean crustal material as were the granites in the north. The zircons from these two samples define well-fitting chords without any inherited component (see Fig. 9, A60-Vihtajärvi). Another sample from the Maarianvaara

granite complex, however, also contained small amounts of Archaean rounded zircons (A. Huhma, 1976). The possible influence of old crustal material in the Puruvesi (A453) , Kermajärvi (A25) and Suvasvesi (A24) granitoids was observed earlier in the lead isotopic compositions of the K-feldspars, which are less radiogenic than those in the Svecofennian domain (Kouvo, pers.comm.). The contrasting Pb isotopic evolution in the lower and upper crust, however, complicates the use of Pb isotopes. Another characteristic feature of these granitoids is the low radiogenic $^{208}Pb/^{206}Pb$ ratio (0.06) in zircon, implying a low Th/U ratio in the zircon and probably in the melt as well. The radiogenic $^{208}Pb/^{206}Pb$ ratio of most granitoids is 0.1, although much higher values also exist. These variations are probably connected with the generation and evolution of granite magmas, when Th and U might be compatible elements, i.e. the bulk distribution coefficient between solid and liquid phases is higher than 1. Zircons from the Kermajärvi, Suvasvesi and Maarianvaara intrusions are typically long, light-coloured prisms (Kouvo, 1958), distinct from other zircons in this study.

Svecofennian domain

In contrast, the granitoids investigated from the Svecofennian domain, including samples from the granitoid area of central Finland, are not primarily reworked Archaean crust. In this respect they are very similar to the majority of granitoids in the Svecofennian domain in Finland (Fig. 20) and Sweden (Patchett *et al.*, 1984), and also similar to the Ketilidian granitoids in South Greenland (Patchett and Bridgwater, 1984). Including those analysed by Patchett and Kouvo (1986), twelve of the granitoid samples analysed from large areas of the Finnish Svecofennian domain have $\epsilon_{Nd}(T)$ from −0.9 to +1.2. Three pyroxene-bearing granitoids from the Raahe—Ladoga zone (this study and metasediments from the Tampere

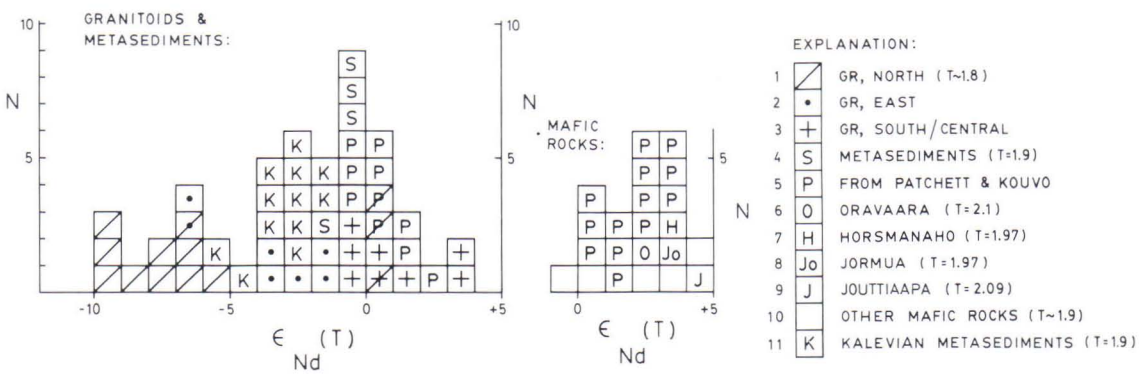

Fig. 20. Histogram of $\epsilon_{Nd}(T)$ for Svecokarelian rocks of Finland. Explanation: 1. $\epsilon_{Nd}(T)$ for granites from northern Finland, 2. granitoids from the Karelian domain in eastern Finland, 3. granitoids from the Svecofennian domain in central and southern Finland, 4. $\epsilon_{Nd}(1.9)$ for the Svecofennian metasediments from Tampere area (Huhma, in press), 5. Data on igneous rocks in the Svecofennian domain from Patchett and Kouvo (1986), 6. $\epsilon_{Nd}(2.1)$ for Oravaara basalt (A398), 7 and 8. $\epsilon_{Nd}(1.97)$ for Horsmanaho (Oku602) and Jormua (A729) gabbros, 9. $\epsilon_{Nd}(2.09)$ for Jouttiaapa basalts (A1009), 10. other mafic rocks, 11, $\epsilon_{Nd}(1.9)$ for the Kalevian metasediments in eastern Finland. For samples with a diagonal line, $\epsilon_{Nd}(T)$ was calculated using T ca. 1.8 Ga. N= number of samples.

schist belt (Huhma, in press) and some mafic rocks also have $\epsilon_{Nd}(1.9)$ close to zero, suggesting that the mean crustal residence time before their final formation was relatively short for all these rocks.

It is impressive that the two most "primitive-type" granitoids from the Svecofennian domain in this study, the I-type calc alkaline Rautio batholith (A579-Susineva) and the gneissose trondjhemite/tonalite at Joroinen (A93-Saunakangas), seem to contain the smallest component (if any) of older LREE- enriched crust ($\epsilon_{Nd} = +3$). Similar positive $\epsilon_{Nd}(T)$ values from +1.7 to +3.2 have recently been reported by Patchett and Kouvo (1986) from trondjhemites and diorites in the Kalanti district, southwestern Finland and by Wilson et al. (1985) from the granitoids in the Skellefte district in Sweden. The $\epsilon_{Nd}(T) = +3$ indicates that a mantle with positive $\epsilon_{Nd}(T)$ rather than some other mantle source contributed significantly to the genesis of these granitoids, and that the crustal residence time of their source was very short. Tonalites may have been formed by partial melting of short-lived garnet-bearing basaltic crustal material, which was a product of partial melting of the LREE-depleted upper mantle (Hanson, 1978). Arth *et al.* (1978) suggested that hornblende fractionation could account for the pattern of REE distributions in the trondhjemites in SW Finland. If the mantle with positive $\epsilon_{Nd}(T)$ was the major contributory primary source, Nd data further suggest that the involvement of older crustal material in other granitoids in the Svecofennian domain (with $\epsilon_{Nd}(T) = 0$) was slightly greater.

MODEL AGES

The model ages for the granitoid samples are given in Table 4. From the preceeding results it is clear that a depleted mantle with $\epsilon_{Nd}(T)$ from +4 to +5 contributed to the formation of the Svecokarelian crust, and that the model ages calculated relative to the chondritic uniform reservoir (T_{CHUR}) are probably not useful. Since the most primitive samples approach the depleted mantle by DePaolo (1981) in their $\epsilon_{Nd}(T)$ (see Fig. 17), the D_{DM} model ages could well be used to evalueate the mean crustal residence ages, the period of time during which the rare earth elements resided within the continental crust (O'Nions et al., 1983). On the other hand substantial REE fractionation could take place during the generation of granitic magma and, in an extreme case, it would not be possible to calculate any model ages (e.g., sample A428). The Sm/Nd ratios for most granitoids are roughly similar to the ratios in the average continental crust, implying only minor fractionation (e.g., Farmer and DePaolo, 1983). If granite was formed by partial melting in the lower crust, with pyroxene ±garnet in the residue, the Sm/Nd ratio in the melt would be lower than in the source (Hanson, 1978; Farmer and DePaolo, 1983), resulting in model ages "too young" for the crustal residence of the material. This is what may have happened with many of the Svecokarelian granites in northern Finland, since their Sm/Nd ratios are relatively low.

Since partial melting is not necessarily an equilibrium process, the melt formed by anatexis may possibly have a slightly different isotopic ratio from that in the source (Fourcade and Allègre, 1981). If old LREE-enriched phases, important for the REE budget, were contained in the unmelted residue without equilibration, the Nd ratio in the anatectic melt would be more radiogenic than the source, resulting in the "too young" model ages (and

too high $\epsilon_{Nd}(T)$). This effect is probably very small, because the Sm/Nd ratios for granites are close to the ratios in the average crust.

Model ages do not date geological events, mainly because of the mixed sources. In this study the most primitive Svecokarelian granitoids have a mean crustal residence age of T_{DM} = 1.93 Ga, which is close to the U-Pb zircon ages. For the granites in northern Finland this age is about 2.5 Ga, which is generally 0.7 Ga in excess of their U-Pb zircon age.

THE CHEMICAL COMPOSITION OF THE GRANITES

A discussion of whether or not the genetic differences shown by the $\epsilon_{Nd}(T)$ of the granites from different geotectonic environments are also reflected in their chemical composition is beyond the scope of this paper. Some general observations can be made from the few analyses published, but the data are too limited for conclusions to be drawn. The granites from northern Finland included in this study are very poor in mafic minerals and approach the minimum melt composition (SiO_2 76%). They should be compared with the true granites from the Svecofennian domain in southern Finland. The few chemical analyses available (see Appendix 2, with references) suggest similar major element compositions for late-orogenic and postorogenic granites at Hanko (A875), Puruvesi (A453), Vainospää (A168), Nattanen (A527) and Hetta (A181).

Slight differences are observed in the aluminous character. Granites at Puruvesi (A453 with $\epsilon_{Nd}(1.8) = -6.9$) and Maarianvaara (A60 with $\epsilon_{Nd}(1.86) = -6$) have higher molar ratios of $Al_2O_3/(CaO + K_2O + Na_2O)$ (= A/CNK) than the other granites (Appendix 2). Peraluminosity is considered to be characteristic of S-type (sedimentary provenance) granites (Chappell and White, 1974), which also tend to have lower $\epsilon_{Nd}(T)$ than the I-type granites (McCulloch and Chappell, 1982). Both Puruvesi and Maarianvaara granites contain accessory monazite, which is also commonly present in the S-type granites. On the other hand, the Hanko ($\epsilon_{Nd}(1.83) = 0$) and Nattanen granites ($\epsilon_{Nd}(1.77) = -9$) also contain accessory monazite.

The migmatizing late-orogenic granites in southern Finland (e.g. the Hanko granite) generally have high K/Na ratios and contain varieties in which almandine and cordierite are common (e.g., Sederholm, 1923), pointing to S-type characteristics. The calculated A/CNK ratios of late-orogenic granites are variable, often exhibiting values around 1.1 (Simonen, 1960; Härme, 1965; Nurmi *et al.*, 1984).

Consequently, most granites included in the present study are not highly peraluminous, suggesting that their source was not strongly weathered highly aluminous sedimentary one, regardless of the unradiogenic neodymium encountered in many of them. The unradiogenic Pb of the K-feldspars in some granitoids from the Karelian domain also indicates a cratonized rather than a sedimentary source.

SVECOKARELIAN METASEDIMENTS, EVIDENCE FOR RAPID RECYCLING

Clastic sediments form an important part of the upper continental crust. The Sm-Nd method has made it possible to determine the mean age of sediment provenances (e.g., McCulloch and Wasserburg, 1978). Sm-Nd isotope analyses of the Svecofennian and

Kalevian metasediments have recently been carried out in order to study crustal recycling (Huhma, 1985 and in press). The $\epsilon_{Nd}(1.9)$ data are summarized in Fig. 20 together with other Nd data on the Finnish Svecokarelian.

The Svecofennian metagraywackes in the Tampere schist belt constitute the lowest major stratigraphic unit (Simonen, 1953), and were most probably deposited from turbidity currents shortly before 1.9 Ga. The mean T_{DM} = 2.22 Ga and $\epsilon_{Nd}(1.9)$ = –0.6 for the metagraywackes are similar to those of many intrusive granitoids in the Svecofennian domain, and indicate that the bulk of the sediments were eroded from recently formed crust. Thus the Archaean crust could not have been a major source of these turbidites. Neither have the neighbouring igneous rocks been postulated as the source, since the metagraywackes are considered the lowest formation, although contrary views have also been expressed (e.g., Mäkelä, 1980). In fact, the metasediments with $\epsilon_{Nd}(1.9)$ = 0 could have been the source for some granites (Fig. 20). The Nd data by Huhma (in press) suggest that metagraywackes contain early orogenic Svecofennian detritus that, according to the sedimentological study by Ojakangas (1986), is largely of volcanogenic origin. The relatively coarse detrital zircons with an apparent U-Pb age of 2.3 Ga (Kouvo and Tilton, 1966) imply the presence of a small detrital component at least 2.3 Ga old. Hence, these metasediments are products of rapid recycling and suggest short crustal residence times for their major precursors, supporting conclusions derived from the granitoids in the Svecofennian domain.

The Kalevian metasediments, which were deposited about 2.1—1.9 Ga ago on the edge of the Archaean craton in eastern and northern Finland, have a mean crustal residence age of T_{DM} = 2.41 Ga (Huhma, in press), which is about 0.4—0.5 Ga in excess of the inferred stratigraphic age. This indicates that their provenance was, on average, older than that of the Svecofennian metasediments investigated. Their source could have been neither purely Archaean nor similar to the bulk of the 1.9 Ga old igneous rocks in the Svecofennian domain. The average $\epsilon_{Nd}(1.9)$ = –2.8 (Fig. 20) is close to that of some Svecokarelian granitoids in the same area (A24, A25); such sediments could thus be candidates for the source of these granitoids. The K-feldspars from the granitoids, however, have relatively unradiogenic Pb isotopic compositions (Kouvo, pers.comm.), pointing to the influence of a lower crustal source rather than that of a sedimentary source with more radiogenic Pb. The sediments reflect a crustal averaging process like that achieved by a granitic magma evolving within the crust by assimilation and melting (cf. Halliday and Stephens, 1984). Similar ϵ_{Nd} values can obviously be reached by different mixing processes.

MIXING MODEL FOR COMMON LEAD DATA FROM FINLAND

The variable initial Nd isotopic composition of the Svecokarelian synorogenic mafic rocks in the Svecofennian domain can be compared with the differences in the initial Pb isotopic composition. Whole rock lead isochrons from the volcanic suites at Pyhäsalmi and Pihtipudas (Fig. 1), close to the Raahe—Ladoga zone, both yielded an age of about 1.9 Ga (Helovuori, 1979; Aho, 1979). The distinct initial ratios of these isochrons are confirmed by lead isotopic data on galenas in associated sulphide ores (Kouvo and Kulp, 1961; Vaasjoki, 1981). These data indicate different sources for approximately coeval Svecofennian volcanic rocks and lead in ores.

The lead isotope ratios of the Svecofennian

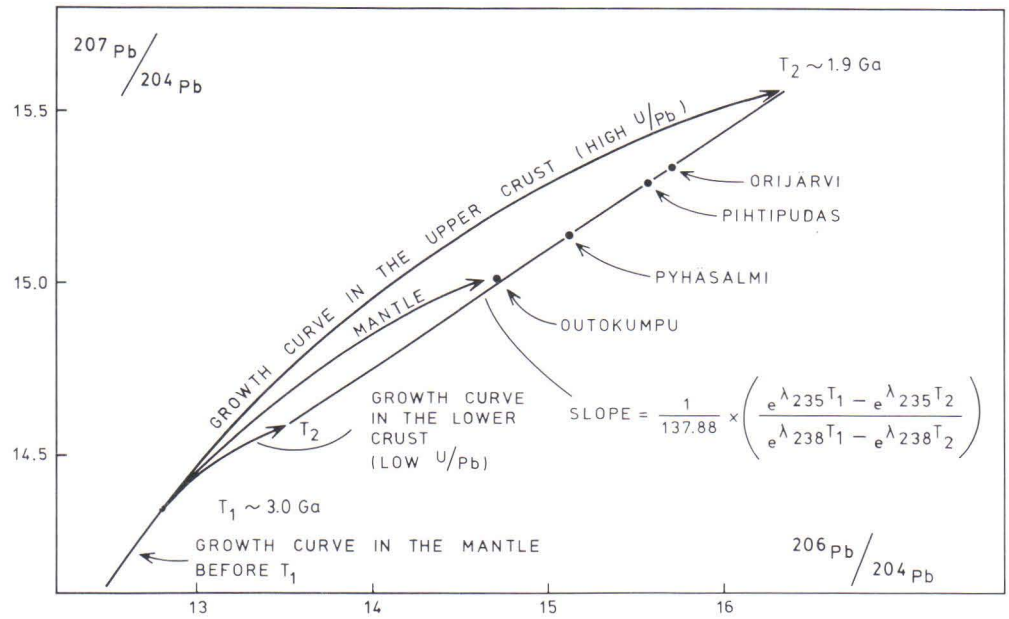

Fig. 21. $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram illustrating the lead isotopic evolution in the mantle and continental crust, starting from a common mantle isotopic composition 3 Ga ago and extending to 1.9 Ga. The lead isotope composition of Finnish galenas (Vaasjoki, 1981) are roughly compatible with the mixing model (see text).

galenas in Finland form groups that are geographically and geologically definitive, and are attributed to a varying degree of mixing between mantle and Archaean crustal materials (Vaasjoki, 1981; see also Rickard, 1978). The isotopic data on galenas define an orogenic trend in the $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram (Vaasjoki, 1981), which has a slope of about 0.37. The data and trend can be interpreted by a two-stage model using the equation given in Fig. 21, which has been derived from the growth equations for the lead isotopic ratios between T1 and T2 (see e.g., Köppel and Grünenfelder, 1979). According to this model, mixing took place 1.9 Ga ago (T2). It involved a mantle component and an Archaean upper crustal component that differentiated from the mantle about 3.0 Ga ago (T1), and in which a high U/Pb ratio generated relatively radiogenic lead during the time between T1 and T2. Material from Archaean lower crustal sources, where relatively unradiogenic lead was produced in low U/Pb environments, probably also contributed.

This model has been illustrated in Fig. 21, which shows the inferred growth curves for the mantle and for the upper and lower crust, starting from a uniform lead isotopic composition at 3 Ga and extending to 1.9 Ga. Since the Pb isotopic data are scattered around the given line, the value reported here for T1 (3 Ga) and for the initial Pb isotopic ratio T1 Ga ago is no more than approximate. However, using the line given by Vaasjoki (1981) the calculated initial Pb isotopic composition 3 Ga ago ($^{206}Pb/^{204}Pb$ = 12.9, $^{207}Pb/^{204}Pb$ = 14.34) is very close to the average terrestrial lead growth curve 3 Ga ago (Stacey and Kramers, 1975). Variable mixing during the 1.9 Ga orogeny is now exhibited as lead isotope provinces (e.g. Outokumpu, Pyhäsalmi, Pihtipudas, Orijärvi) on a line in the $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$

diagram. An infinite number of T1 and T2 pairs fulfil the equation, but the model given above is geologically the most reasonable one.

The consistency of the Pb/Pb model ages of the galenas with the inferred stratigraphic age is obviously better with this model than with the global models, such as that by Stacey and Kramers (1975). The model ages, in general, are, however, very sensitive to the parameters, and should obviously not be used as ages.

Because the galenas often occur in stratiform deposits associated with sediments, their Pb isotopic composition cannot be correlated directly with the Nd isotopic composition of the igneous rocks. Nevertheless, the diversity in both Pb and Nd isotopic data could be due to mixing of materials from two distinct sources. Current data do not allow more detailed comparison. Although the data on the galenas could also be attributed to the heterogeneity of the mantle, the mixing hypothesis seems to be more plausible, since many of the galenas may have been formed on the sea floor from migrating ore-bearing fluids that probably extracted material from various sources, e.g. Archaean crust (cf. Plumbotectonic model by Doe and Zartman, 1979). The differences in the initial Pb isotope ratios between volcanic suites suggest that mixing did not only take place on the surface but also before extrusion, during the orogeny. Sediments with more radiogenic lead were probably mixed effectively with mantle-derived magmas in the subduction zone.

According to the model (as shown in Fig. 21), much of the lead in galenas from Orijärvi (and from all the Svecofennian supracrustal formations given in Vaasjoki, 1981) and Pihtipudas had Archaean upper crustal sources, which appears to be contrary to the conclusions derived from the Nd isotope data. This is, however, probably a consequence of the low Pb concentration ratio between mantle- and crustal-derived materials, which can be shown by mass balance calculations. Following Faure (1977, p. 98), we get the equation:

$$f = \frac{C^{MA} \times (R^{M} - R^{MA})}{C^{C} \times (R^{C} - R^{M}) + C^{MA} \times (R^{M} - R^{MA})}$$

where f is the weight fraction of crustal component in the mixture, C^{MA}, C^{C}, R^{MA}, R^{C} are the Pb concentrations and $^{206}Pb/^{204}Pb$ isotope ratios of mantle- and crustal-derived materials, respectively, and R^{M} is the $^{206}Pb/^{204}Pb$ ratio in the mixture.

The concentration ratio of $C^{C}/C^{MA} = 15$ (e.g., $C^{C} = 15$ ppm, $C^{MA} = 1$ ppm) is considered quite reasonable (see e.g., Weaver and Tarney, 1980 and Sun, 1980). The $^{206}Pb/^{204}Pb$ ratio of $R^{C} = 16.3$ is assumed for the Archaean crust 1.9 Ga ago, which corresponds to $\mu = 13$ between T1 and T2. The mantle end member 1.9 Ga ago is assumed to be similar to the source of lead in Outokumpu ore ($R^{MA} = 14.8$), which is associated with the Outokumpu ophiolite (Koistinen, 1981). Galenas from Orijärvi have $R^{M} = 15.7$, and using the numbers given above, we get f = 9% for Orijärvi. This is close to the estimate derived from the Nd data for $\epsilon_{Nd} = 0$. At Pihtipudas ($R^{M} = 15.5$), the crustal component accounted for 6% of the mixture and at Pyhäsalmi ($R^{M} = 15.1$), for 1.5%. This could be an indication of a positive $\epsilon_{Nd}(1.9)$ for the volcanic rocks at Pyhäsalmi and of $\epsilon_{Nd}(1.9)$ close to 0 at Orijärvi and Pihtipudas; no Nd data are yet available on these rocks, however. If the concentration ratio between mantle- and crustal-derived materials was lower in the mixing during the orogeny, even smaller amounts of Archaean crustal material would have been needed to produce the large spread in the lead isotope ratios. Owing to the uncertainties in the end member values the results of these calculations should be considered tentative.

DISCUSSION

Jatulian mantle-derived mafic rocks 2.1—2.2 Ga old (with ϵ_{Nd} up to +4.2) indicate extensive rifting of the Archaean craton. This led to the development of the Svecofennian geosyncline and the subsequent rapid crustal growth about 1.9 Ga ago. Slices of the 1.97 Ga old oceanic crust yield $\epsilon_{Nd}(1.97) = +3.2$ (Fig. 22).

During the orogeny, variable amounts of Archaean crustal material were mixed with the juvenile material deep in the crust or below it. This is manifested as the large spread in $\epsilon_{Nd}(T)$ from roughly coeval felsic and mafic igneous rocks (Figs. 20 and 22). Similar evidence is supplied by the common lead data on galenas and some volcanic rocks (Fig. 21). The Nd-isotopic results with $\epsilon_{Nd}(T) = +3$ from some granitoids in the Svecofennian domain suggest that crust-formation processes, which transfer material from the mantle to tonalitic-granodioritic-granitic crust, took place within a relatively short time span (e.g., 0.1 Ga). Experimental and trace element data indicate that granites do not derive from primary magmas from the mantle or subducted oceanic crust (Green, 1980; Wyllie, 1984; Hanson, 1978) and that the crust-generating processes probably include many stages of partial melting, crystal fractionation and solidification. Granitoids may have been formed by partial melting of newly mantle-derived basaltic lower crustal material. The generation of syenogranites, such as sample A545-Lehesvuori, probably involved a more felsic, short lived crustal precursor.

Data on the Svecofennian domain further suggest that the generation of synorogenic granitoids was not aided significantly by

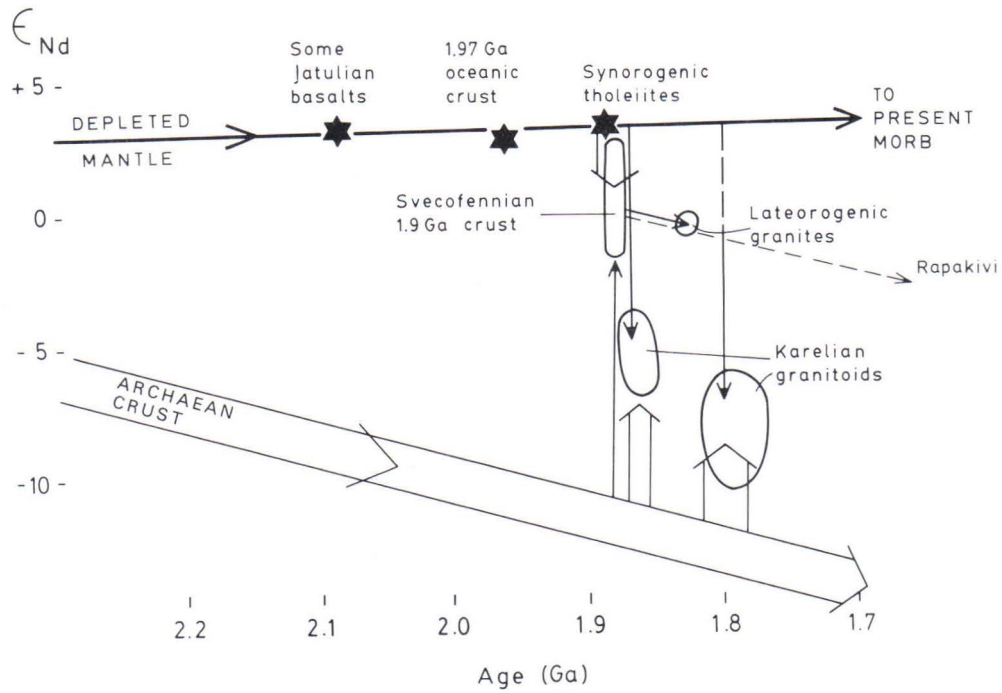

Fig. 22. Model for the origin of the Svecokarelian crust in ϵ_{Nd} vs. age diagram. Changes in ϵ_{Nd} are due to: 1) radioactive decay - gently sloping arrows, 2) mixing - vertical arrows, heavy arrows illustrate large input.

anatectic melts from much older crust and that this process most likely occurred in a region where Archaean crust was not present. In the light of recent orogenic analogues, this could involve subduction zones, where crustal-derived sediments on the ocean floor have been subducted and mixed, resulting in abundant orogenic terrains (cf., Patchett and Bridgwater, 1984). Some tholeiitic magmas (such as the samples from Evijärvi in this study) were formed by partial melting of LREE depleted upper mantle without any mixing with subducted crustal material.

After the major orogenic phase about 1.89 Ga ago, which created most of the Svecokarelian crust largely from newly mantle-derived material, crustal evolution continued in many places with the formation of late-orogenic microcline granites about 1.8 Ga ago, which migmatized the existing rocks. Although the ϵ_{Nd} data on late-orogenic granites in the Svecofennian domain are few, they suggest that these granites were formed largely from anatectic melts from the 1.9 Ga old Svecokarelian crust. An anatectic crustal origin for microcline granites had already been proposed by Sederholm (1926), Simonen (1960) and Härme (1965). The Nd-data exclude a major contribution of Archaean crustal material in the genesis of these late granites of the Svecofennian domain, whereas the late-orogenic and postorogenic granites in the Karelian domain formed largely from anatectic melts of the Archaean crust.

The 1.7—1.5 Ga old rapakivi granites, too, could have been formed from reworked 1.9 Ga crust, but the associated anorthosites, gabbros and diabases indicate input from the mantle as well (Vorma, 1976; Haapala, 1985; Laitakari, 1969). Only two Nd analyses are available of Finnish rapakivi granites, which have initial Nd ratios very similar to those of granitoids of the Svecofennian domain (T_{CHUR} about 1.9 Ga; T. Sando, pers.comm.). The Pb isotope data on the galenas associated with the rapakivi granites are also consistent with this interpretation. They have μ-values similar to those from the galenas in the Svecofennian supracrustal formations (Vaasjoki, 1981), which are higher than the μ-values for the average terrestrial lead in Stacey and Kramers (1975). However, this is not in agreement with the Hf isotopic data ($\epsilon_{Hf}(T) = 0 — +3$) on rapakivi zircon reported by Patchett *et al.* (1981). An exotic result was obtained by Vaasjoki (1977), who reports a very small amount of Archaean zircon in one sample of 1.59 Ga old rapakivi granite in SW Finland.

The proportions of the components in mixing can be estimated by the REE concentrations and Nd isotopic ratios of the end members. Patchett & Bridgwater (1984) and Patchett and Kouvo (1986) estimated the Archaean component to be about 10% to produce material with $\epsilon_{Nd}(T) = 0$. This was based on concentrations of Nd = 32 ppm and $\epsilon_{Nd}(1.9) = -12$ in the crustal component and Nd = 10 ppm and $\epsilon_{Nd}(1.9) = +4$ in the mantle-derived component. Assuming the same Nd concentrations, a simple mass balance to explain $\epsilon_{Nd}(1.9) = -8$ requires a 50% Archaean component (= 75% of Archaean Nd) to be mixed with the mantle material. If the Archaean gneiss from Sotkuma (R27) with Nd = 11 ppm and $\epsilon_{Nd}(1.8) = -10.5$ is used, the crustal component in the granites is even higher. If the REE rich phases are contained in the unmelted residue during anatexis without equilibrating with the melt, calculations would be even more tentative. It is obvious, however, that only a few percent of Archaean material will shift the $\epsilon_{Nd}(1.9)$ down from the mantle evolution line (Fig. 22).

An alternative explanation for $\epsilon_{Nd} = 0$ would be to assume differentiation of depleted mantle about 2.2 Ga ago, which would have produced material with LREE enriched chemistry and should thus have $\epsilon_{Nd}(1.9) \approx 0$. Some Jatulian volcanic rocks and sills are enriched in LREE with $\epsilon_{Nd}(1.9)$ about –2 (Huhma, in

press). There is not much evidence, however, that large amounts of such material was available to contribute to the formation of the Svecokarelian crust 1.9 Ga ago, even though the detrital zircons from the Svecofennian metagraywackes in the Tampere area have apparent U-Pb ages of about 2.3 Ga (Kouvo and Tilton, 1966). T_{DM} ages of about 1.9 Ga from some granites are not consistent with this model either.

The role of the granulite belt in Lapland, which is about 2 Ga old (Meriläinen, 1976; Fig. 2), has not been established. It has been suggested that it might be possible to find the molasses of this uplifted belt in the Svecofennian geosyncline (Barbey *et al.*, 1984). $\epsilon_{Nd}(1.9)$ about 0 has been measured for LREE-enriched mafic granulites (Bernard-Griffiths *et al.*, 1984), showing an initial Nd-isotopic composition similar to that of the bulk of the Svecokarelian domain. The common lead data on K-feldspars of the granulites (Kouvo, pers.comm.; Meriläinen, 1976) are also consistent with a Proterozoic origin. Thus, the initial isotopic compositions available of the granulite material are similar to those of the Svecofennian crust. Whether or not they are genetically related is unknown.

The formation of 1.8—1.9 Ga crust from dominantly juvenile mantle material has recently been established using initial Nd ratios in Colorado (DePaolo, 1981), in Greenland (Patchett and Bridgwater, 1984), in Sweden (Patchett *et al.*, 1984; Wilson *et al.*, 1985) and in Finland (Patchett *et al.*, 1981; Patchett and Kouvo, 1986; present study). Many Phanerozic I-type batholiths formed at the continental margins also contain a significant, albeit variable, component of depleted mantle (e.g., DePaolo, 1980; Hensel *et al.*, 1985).

Only a few Nd data exist on 1.8—1.9 Ga granites generated largely by partial fusion of Archaean crust (Kouvo *et al.*, 1983; Karlsbeek and Taylor, 1985). In contrast, many Phanerozoic granites, e.g. Caledonian and Hercynian, seem to have a significant old component (e.g., Hamilton *et al.* 1980; Allègre and Ben Othman, 1980; McCulloch and Chappell, 1982; Farmer and DePaolo, 1983; Halliday, 1984). This is comparable to the granites of the Karelian domain in the present study. In the granites from the Mesozoic and Tertiary western USA, the proportion of crustal component increases from 10% in the west to 100 % in the east (Farmer and DePaolo, 1983; Kistler and Peterman, 1973). According to Farmer and DePaolo (op.cit.), the granites formed through the interaction of magma derived from a LREE-depleted mantle reservoir with an inland-increasing proportion of assimilated continental-derived pelagic sedimentary rock. The trend is similar in the Svecokarelian terrain as a whole, though no systematic trend can be found within the Svecofennian domain. A significant increase in the proportion of assimilated Archaean rocks is observed on the Karelian side of the Raahe—Ladoga zone.

Somewhat similar to the data on the Svecofennian domain are the recent data on the Phanerozoic New England Batholith, where a narrow range ($+1.0 \pm 1.5$ ϵ units) of different granitoids of the New England Super-Suite was observed. The source rock of the granodiorites and tonalites of the Nundle suite, which have ϵ_{Nd} values from $+3.3$ to $+6.1$, was probably an intermediate volcanite of island-arc character (Hensel *et al.*, 1985).

The generation of new continental crust during the Phanerozoic is closely associated with subduction processes, and a similar mechanism has been postulated for the formation of Proterozoic orogenic belts (e.g., Hoffman, 1980). The Svecokarelian orogeny involved a process in which large quantities of continental crust were formed from dominantly new mantle-derived material at the margin of the Archaean craton. Although the rapid crustal growth during the Svecokarelian orogeny differs from modern orogenic belts,

the isotopic data are compatible with a plate tectonic model in which mantle-derived material was mixed at the continental margin with Archaean-derived sediments by a process related to subduction. The division of the Baltic Shield into the continental Karelian block and the oceanic Svecofennian block by Hietanen (1975) is strongly supported by isotopic data, which also suggest that the western edge of the Archaean craton coincides roughly with the Raahe—Ladoga geosuture zone. The $\epsilon_{Nd}(T)$ values from the synorogenic igneous rocks in the Svecofennian domain are comparable to the variation encountered in modern island arc magmas with LREE-enriched chemistry, where a small recycled continental component is required to explain the radiogenic isotope ratios and the abundances of incompatible trace elements (e.g., Hawkesworth *et al.*, 1977; Cohen and O'Nions, 1982 ; White and Patchett, 1984).

The heat required for the generation of late-orogenic and postorogenic granites was probably a consequence of some other process, e.g. underplating of the crust by mantle-derived magmas (mantle plumes) or process that created tectonically thickened crust (cf., Korsman *et al., 1984),* which would increase the temperature of the crust above the H_2O-saturated granite solidus (Wyllie, 1977).

A large number of U-Pb zircon datings of Svecokarelian orogenic rocks exhibit ages around 1.89 Ga (e.g., Neuvonen *et al.,* 1981), and much of the present Nd isotopic study was concentrated on these rocks. However, there are some older rocks in the Svecofennian domain, such as the foliated tonalites with 1.93 Ga zircons (Helovuori, 1979) close to the Raahe—Ladoga zone in the proximity of the Archaean crust, which seem to be associated with certain metamorphic blocks as recently described by Korsman *et al.* (1984). If the tectonic models involving the accretion of continental fragments or microplates (Nur and Ben-Avraham, 1982) turn out to be appropriate for the Svecokarelian terrain, it might be possible to find even older crustal fragments in the Svecofennian domain. Existing isotopic data, however, give no evidence of that.

CONCLUSIONS

The main conclusions of this study are:

1) A depleted mantle with $\epsilon_{Nd}(T)$ about +4 — +5 existed beneath the Baltic Shield during Early Proterozoic times, and contributed to the formation of the Svecokarelian crust.

2) During the main orogenic stage 1.9—1.87 Ga ago large quantities of new continental crust were formed from predominantly newly mantle-derived material at the margin of the Archaean craton. The granitoids investigated from the Svecofennian domain have ϵ_{Nd} from −1 to +3, suggesting a small admixture of Archaean continental crust. The mixing is further supported by the neodymium isotopic data on synorogenic mafic rocks (ϵ_{Nd} from 0 to +4.5) and by the lead isotopic data on Svecofennian galenas and some metavolcanic rocks.

3) At the moment there is no evidence for predominantly reworked Archaean crust in the Svecofennian domain, including the granitoid area of central Finland, where granitoids are mainly about 1.89 Ga old.

4) There was considerable involvement of Archaean crustal material during the generation of the Svecokarelian granitoids in the Karelian domain in northern and eastern Finland, where the majority of the granites investigated have ϵ_{Nd} from −9 to −6. Isotopic data strongly support the division of the Baltic

Shield into continental Karelian and orogenic Svecofennian terrains.

5) The late-orogenic 1.83 Ga old granites in the Svecofennian domain are reworked 1.9 Ga old crust, whereas the late-orogenic and post-orogenic granites in the Karelian domain derive largely from reworked Archaean crust.

6) The isotopic data are consistent with a plate tectonic model in which Archaean-derived sediments were mixed with juvenil mantle material via subduction-related processes at the continental margin. At the synorogenic stage, 1.9—1.87 Ga ago, the Svecofennian domain originated from this mixed material in an island arc environment. At later stages, granites formed from anatectic crustal melts.

ACKNOWLEDGEMENTS

First and foremost I thank Dr. Olavi Kouvo for his continuous help and interest in this study. His enthusiasm for geology prompted numerous stimulating discussions, from which I derived much benefit in the course of my work. He introduced me to the realm of isotope geology and encouraged and helped me to start the Sm-Nd method in our laboratory. All the zircon analyses, which play such an important role in this work, were made under his supervision. My coworkers at the Isotope laboratory deserve special thanks: Tuula Hokkanen and Marita Niemelä for careful handpicking and zircon chemistry; Matti Sakko, Arto Pulkkinen, Timo Lahti and Matti Ryynänen for keeping the apparatus going and for running Pb and U; and Juha Karhu and Matti Vaasjoki for our many discussions.

My visit to the Department of Earth Sciences, University of Leeds, was very important in launching the Nd isotope work. I am deeply grateful to the staff of the Isotope Laboratory at Leeds, especially Drs. R.A. Cliff and M. Dodson.

I wish to thank those who contributed samples or data for this study, particularly Pentti Rastas, Jorma Paavola, Antti Pääjärvi, Veli Suominen, Ulla Mäkelä, Ossi Ikävalko, Lauri Pekkarinen, Osmo Nykänen, Pertti Hakanen, Asko Kontinen, Leo Grundström and Markus Vaarma. I am indepted to Maija Huhma, my mother, for helping me to clarify my ideas and to collect the samples from eastern Finland. I also thank Matti Karhunen, Mirja Saarinen, Anneli Rinne and Leena Järvinen for carrying out the crushing and mineral separation, and Liisa Siren and Ritva Forsman for the final drawings. My special thanks go to Pertti Hautala for innumerable discussions during the past few years. Professors G.W. Lugmair and G.J. Wasserburg are acknowledged for providing standard material.

Useful comments on an early version of the manuscript were received from Ilmari Haapala, Atso Vorma, Olavi Kouvo, Matti Vaasjoki, Juha Karhu, Pertti Hautala and Maija Huhma. I also appreciate comments by Jonathan Patchett and Yrjö Kähkönen. The English was corrected by Mrs Gillian Häkli. I am happy to record my appreciation to these persons and to all those not mentioned who contributed to the study in many ways.

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Appendix 1.

Sampling sites and short descriptions of samples.

Sample		Location Map sheet	Grid Coordinates		Description	Acc.	Sample for Nd collected by
Granites from northern Finland:							
A527	Nattanen, Sodankylä	3742	7557.95	516.80	red even-grained Nattanen granite Zircons: brown, euhedral, transparent, no cores, l:w = 1:1 —1:3	Mon	Drilling
A168	Vainospää, Inari	4911	7716	445,7	grey slightly foliated Vainospää granite Zircons: light, strongly zoned, subhedral, no cores, l:w = 2:1	Ti	Meriläinen
A126	Hatajavaara, Salla	4621	7412.50	437.05	fine-grained pale granite of CL Zircons: mainly rounded, some crystals have a core and zoned shell, l:w = 2:1		Lauerma
A181	Peltovuoma, Enontekiö	2831	7591.45	505.60	homogeneous Hetta granite Zircons: strongly zoned, euhedral, no cores l:w = 2:1 —4:1	Ti	Matisto
A368	Rovaniemi ltk	3612	7386.80	447.50	red granite of CL Zircons: brown, not transparent, often zoned, partly euhedral - partly irregular, l:w = 2:1	Ti	Nuutilainen
A145	Molkokoöngäs, Rovaniemen mlk	2644	7440.55	547.35	slightly foliated red granite of CL Zircons: brown, poorly transparent, strongly zoned, subhedral-partly irregular, some crystals have a core, l:w variable	Ru	Rastas
A429	Kuivavaara, Kittilä	2741	7546.35	525.45	even-grained granite of Tepasto, poor in zircon. New zircon age from A184-Kotivaara, porphyritic granite of Tepasto	Ti	Rastas
A428	Lappalaislampi, Kittilä	2732	7500.00	527.10	even grained granite of central Kittilä, poor in zircon	Ti	Rastas
A748	Iso Nilipää, Kittilä	2734	7493.75	555.02	pale porphyritic coarse-grained, amphibole-bearing granite of CL. Zircons: brown, euhedral, transparent, strongly zoned with cracks, l:w = 3:1	Ti	Rastas
Granitoids from eastern Finland:							
A453	Syrjäsalmi, Kitee	4213	6879.20	494.78	red Puruvesi granite Zircons: red-brown subhedral-irregular	Mon	Huhma
A336	Närsäkkälä, Kitee	4231	6868.85	506.20	quartz diorite Zircons: light-coloured, clear, euhedral, l:w = 2:1		,,
A24	Parkkolansaari, Leppävirta	3243	6929.30	559.80	grey Suvasvesi granite/granodiorite with orthoclase porphyroblasts. Zircons: light-coloured, euhedral, zoned, no cores, l:w = 4—5:1		,,

A25	Viitalahti, Heinävesi	4221	6932.90	421.10	grey Kermajärvi trondjhemite (Nd sample)/ granodiorite with orthoclase porphyroblasts (zircon sample)		Huhma
		3243	6936.80	577.00	coordinates for the zircon sample. Zircons: similar to A24		
A60	Vihtajärvi, Kaavi	4311	6991.10	434.18	grey fine-grained Maarianvaara granodiorite Zircons: very light-coloured long euhedral l:w = 6:1	Mon	''
A63	Jussilansaari, Lapinlahti	3332	7017.70	516.34	grey Onkivesi granite/granodiorite with K-feldspar porphyroblasts	Ti	Paavola
A239	Molkanjärvi, Pielavesi	3313	7005.28	490.04	pyroxene-bearing quartz diorite		Pääjärvi/ Huhma
Granitoids from southern and central Finland:							
A875	Märaskär, Hanko	2011	6634.33	442.55	red, homogeneous Hanko granite Zircons: mainly light-coloured (some red), long, euhedral, l:w = 4:1	Mon	Suominen
A933	Orijärvi, Kisko	2014	6678.80	474.70	granodiorite with inclusions of country rock Zircons: light-coloured with inclusions, poorly developed crystal faces		U. Mäkelä
A668	Orijärvi	2014	6678.79	474.60	granodiorite (cf. A933)		
A669	Nyckeln, Kisko	2014	6670.85	468.88	quartz diorite		
A82	Hankavesi, Ähtäri	2241	6936.80	509.60	coarse amphibole bearing granite of CF Zircons: very transparent, strongly zoned, poorly developed crystal faces, l:w = 2:1	Ti	Huhma
A85	Kiukaanniemi, Ähtäri	2241	6937.10	519.00	coarse porphyritic granodiorite of CF K-feldspar porphyroblasts up to 3 cm long Zircons: light-coloured to reddish		''
A447	Keuruu	2232	6905.14	539.20	slightly foliated granodiorite of CF Zircons: light-coloured, euhedral		''
A545	Lehesvuori, Jyväskylä	3212	6915.27	435.73	coarse microcline granite of CF (51% Mi in mode) Zircons: light-coloured to reddish, mainly euhedral, strongly zoned, l:w = 2:1		Ikävalko
A93	Saunakangas, Joroinen	3231	6889.45	539.00	slightly foliated trondhjemite/tonalite Zircons: very transparent, light-coloured, euhedral, l:w = 2:1		Pekkarinen
A133	Saunakangas	3231	6889.33	539.16	granodiorite		
A579	Susineva, Rautio	2342	7098.59	506.06	granodiorite of Rautio batholith Zircons: brownish, short, subhedral		Huhma
Mafic rocks:							
A398	Oravaara, Tohmajärvi	4232	6900.40	517.30	coarse basalt, rich in amphibole		Nykänen
Oku 602	Horsmanaho, Polvijärvi	4224	6968.93	461.83	coarse amphibole-rich mafic gabbro pegmatoid A149, A234, A235 from the same locality		Hakanen
A729	Jormua, Kajaani	3434	7136.80	550.21	coarse gabbro with saussuritic plagioclase		Kontinen
A542	Laukunkangas, Enonkoski	4211	6882.70	435.00	Ni sulphide-bearing norite		Grundström
A277	Soukkio, Mäntsälä	2044	6731.45	566.43	Mäntsälä hornblende-gabbro		Huhma
A1009	Jouttiaapa, Tervola	2631	7352.9	529.8	basalts (amphibole + albite + epidote)		''

S	Salittu, Suomusjärvi	2023	6688.0	480.0	peridotite (amphibole-rich)	Vokurka/ Huhma
RF 2748	Evijärvi	2314	7034.08	485.88	basalt, pillow lava	Vaarma
RF 2757	Evijärvi	2314	7027.45	492.14	basalt	"
Archaean:						
R4	Runkaus, Tervola	2544	7329.7	564.9	coarse pink granite	Huhma
R27	Sotkuma, Polvijärvi	4224	6957.42	474.30	granodiorite gneiss	"

l:w, average length/width ratio of zircon crystals
Acc., characteristic accessory mineral, Mon = monazite, Ti = titanite, Ru = rutile
CL, central Lapland granitoid area
CF, central Finland granitoid area

Appendix 2.

Modes and brief chemical characteristics of the granitoids.

Intrusion	Mode Plag	Kspar	Qu	Bt	Mu	Af	SiO_2 (wt%)	K_2O (wt%)	Al_2O_3/ CaO + Na_2O + K_2O (moles)	References for modes and analyses, comments
Nattanen (cf. A527)	33.5	24.7	30.7	3.2	4.6		72.82	5.00	1.04	Mikkola, 1941, p. 279
Vainospää (cf. A168)	33.9	35.0	27.4	2.9			73.67	5.01	1.05	Meriläinen, 1976, p. 38
Hetta (cf. A181)	38	25	29	3.8	2.5		72.01	4.75	1.03	Mikkola, 1941, p. 278
Tepasto, A429	*	*	*	2			78.46	4.98	1.00	Analyses from Rastas, pers.comm.
Central Kittilä (A428)	37.7	21.8	31.5	3.7	2.5		73.3	4.38	1.06	Mikkola, 1941, p. 280
Iso Nilipää, A748	*	*	*	3		1	72.2	4.42	1.06	Analyses from Rastas, pers.comm.
Puruvesi (cf. A453)	*	*	*	2	3		73.2	3.99	1.18	Nykänen, 1983, p. 58
Närsäkkälä (cf. A336)	57	1	22	10		5	67.7	2.14	0.99	Nykänen, 1975, p.24 (in average)
Kermajärvi, A25-Zr	46	12.7	27	14.7						sample for zircon analysis, HH
Kermajärvi, A25-Nd	57	0.8	29	9.8						sample for neodymium analysis, HH
Vihtajärvi, A60	39.1	22.8	32.3	5.7			72.63	3.75	1.18	Huhma, A., 1976, p. 14
Hanko (cf. A875)	26	37	32	3			74.76	4.62	1.03	Sederholm, 1926, p. 64
Orijärvi (cf. A933)	35.7	9.0	36.1	7.4		4.7	71.36	2.26	0.98	Eskola, 1914, p. 41
Hankavesi, A82	35.1	24.7	26.0	7.4		4.4				HH
Ähtäri (cf. A82)							70.72	4.29	0.98	Sjöblom, 1984, p. 25
Keuruu, A447	45.1	10.4	21.8	14.9		7.6				Pekkarinen, pers.comm.
Palokka (cf. A545)	7	51	37	2			78	5	1.05	Nurmi et al., 1984, p. 25
Saunakangas, A93	45.1	0.7	46.9	5.6						HH
Rautio (cf. A579)	48	11	24	8		2.7	68	4	1.08	Nurmi et al., 1984, p. 25

Plag = plagioclase, Kspar = K-feldspar, Qu = quartz, Bt = biotite, Mu = muscovite, Af = amphibole.
* = major minerals in approximate equal proportions.
HH = mode by the author.

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Puh. (90) 662 801

Denna publikation säljs av

STATENS TRYCKERICENTRAL

POSTFÖRSÄLJNINGEN
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GOVERNMENT PRINTING CENTRE

MAIL ORDERS
P.O.B. 516
SF-00101 Helsinki Finland
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ISBN 951-690-232-4
ISSN 0367-522X