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Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences

edited by Matti Vaasjoki

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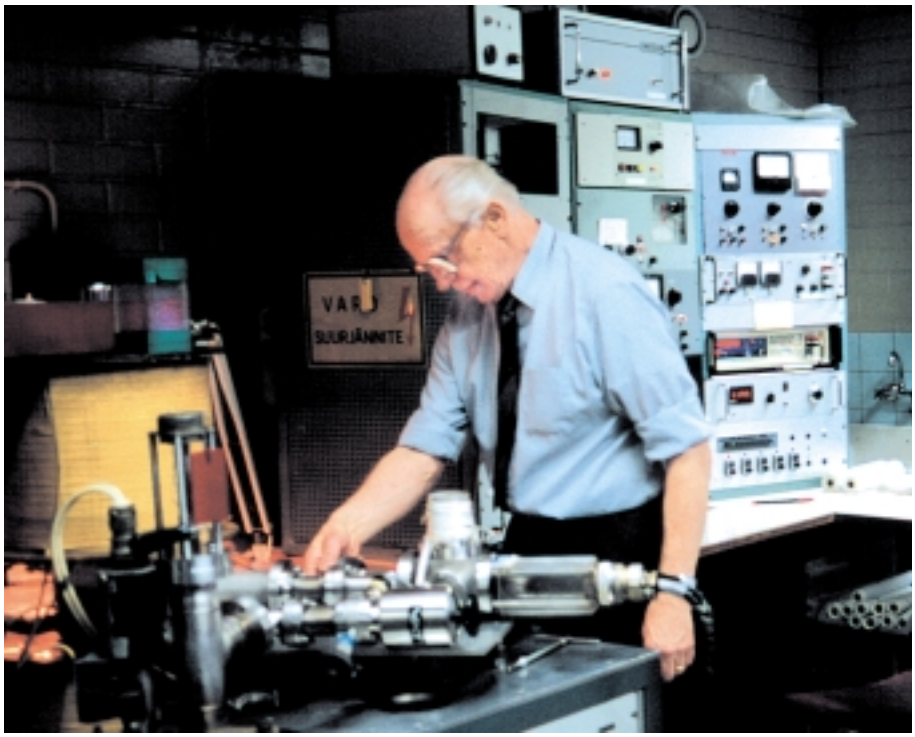

Recent pictures of Olavi Kouvo in the lab preparing the MS I for a run (top) and on a field excursion in central Lapland (bottom). Photos: Tapani Rämö and Pentti Rastas.

DEDICATION

Those aquainted with geochronological research in Finland will immediately spot a defeciency in the list of contributors for the papers in this volume. This is the absence of Dr. Olavi Kouvo's name among the authors.

Olavi Kouvo started his isotope geological career in the United States in the middle 1950's, where he joined George Tilton, George Wetherill, Paul Gast and others developing the methods and applications of the then infant science. His doctoral thesis in 1958 caused a fundamental change in the understanding of Finnish Precambrian geology. Sederholm, Eskola and other pioneers of Finnish geology had believed that there were two great Precambrian orogenies in Finland: the older Svecofennian and the younger Karelian. Olavi's results showed that the two were in fact coeval and that the granite-gneiss domain northeast of Karelides was more than a billion years older than the southwestern part of the country. In spite of Eskola's publicly expressed doubts, the existence of an ancient plate boundary along the Raahe-Ladoga zone was an accepted fact, not a mere working hypothesis even among Eskola's former students (of whom my father happened to be one), when I joined the laboratory in 1973.

In the early 1960s, Olavi was invited to establish an isotope laboratory at the Geological Survey of Finland. His choice of the principal method, U-Pb analyses on zircon, was not only farsighted, but enterprising, as that method was - from sample preparation to the final mass-spectrometric runs - both more demanding and tedious than other methods then employed.

Olavi has always utilized his thorough understanding of mineral chemistry for developing new analytical approaches. Two of these are directly related to this volume. As dating mafic rocks became important in the 1970s, he decided that gabbros should be sampled where they contained sufficient silica to form zircon - hence the dating of gabbro pegmatoids. Where silica was deficient, a primary phase accommodating sufficient uranium should be used. Thus the first baddeleyite analysis was completed at the Survey laboratory in 1976 - though the first in the world, Olavi neither sought nor received credit for it.

A basic tenet of Olavi's work, which he always has stressed to his students, is that isotope geology has to strive for understanding geological processes, not merely produce "ages" for rocks. This led to a long, fruitful and still ongoing co-operation between the field and laboratory staffs of the Survey, and in fact the emphasis on sample descriptions found in this volume is a direct consequence of Olavi's maxim that a result is only just as good as the original sampling.

Most of the analytical work and indeed much of the original sampling presented in this volume have been either personally carried out or initiated by Dr. Olavi Kouvo. Although still active at the Unit of Isotope Geology, from the leadership of which he retired in 1985, he has not wished to be included among the authors in this volume on grounds that we, if not understand, at least respect.

We therefore feel not only compelled to acknowledge Olavi's work in this connection, but in dedicating this volume to him, pay tribute to the founding father of practical isotope geology in Finland.

On the authors' behalf

Matti Vaasjoki

EDITOR´S PREFACE

The history of this volume is rather long and tedious, its general setting being the development of geological research in Finland in general and that at the Geological Survey in particular. As a result of a decentralization process that had been going on for several years, the Survey's Regional Office for North Finland was formally established in 1974, and this led to the stepping up of the 1:100,000 bedrock mapping program in Lapland. The bedrock survey of northern Finland was further intensified by the Lapland Volcanite Project in the latter half of the 1980s.

The isotope laboratory of the Survey had become operational in the middle 1960s, and it was natural that much of its work would be devoted to clarifying the often puzzling and at times controversial chronostratigraphic issues arising from the new field work in Lapland. Unfortunately, as is often the case, both the pace of gathering new evidence and the rate of evolution in interpretation were such that little time could be devoted to reporting of the results in their full scientific context.

This volume seeks to remedy the last mentioned deficiency. Its aim has been to properly document the geochronological data collected from Lapland over a period of almost 30 years. Many of the results have been publicized less formally, in conference papers and posters, but all too often the samples themselves, their geological context and the analytical data have been described only in passing, if at all. The focus is on the Central Lapland Greenstone and Peräpohja Schist belts (Fig. 1), which were the main concerns of the Lapland Volcanite Project, but minor contributions from the Kuusamo Schist belt and from the Finnish Caledonides have been added as well. Thus analytical data are reported in ten papers. To produce a well-rounded entity, summaries on the analytical techniques employed, on the history of stratigraphic research in Lapland and on the conclusions drawn from the geochronological studies have been added. The aim has been to produce a comprehensive volume, but also to address regional aspects in a manner to provide easy access to readers interested in specific problems or even in individual samples.

In this context, it is appropriate to acknowledge the contribution of the staffs of the laboratories of Isotope Geology and Mineralogy of the Geological Survey, without whose technical assistance the cumulation of the data would never have occurred. Though naming everyone involved is impossible, several key people have to be mentioned: Mr. Matti Sakko who assisted Dr. Olavi Kouvo in setting up the isotope laboratory and conducted innumerable mass-spectrometer runs, Mrs. Marita Niemelä and Mrs. Tuula Hokkanen who have done virtually all of the chemical purification of uranium and lead from the samples and much of the hand-picking of the analysed mineral fractions, Mrs. Mirja Saarinen and Mrs. Leena Järvinen who have done much of the heavy liquid separations and XRD mineral identifications and, finally, Mr. Matti Karhunen who has skilfully operated the shaking table and is also responsible for archiving the sample materials.

It is also my pleasure to thank those many colleagues who in various ways have assisted me in my editorial tasks either as referees or in a less official capacity. Any help and advice I asked for was willingly given. Thus the finalization of this volume should not be regarded as a feat of a restricted working group, but rather as an exemplary collective achievement of the geological community.

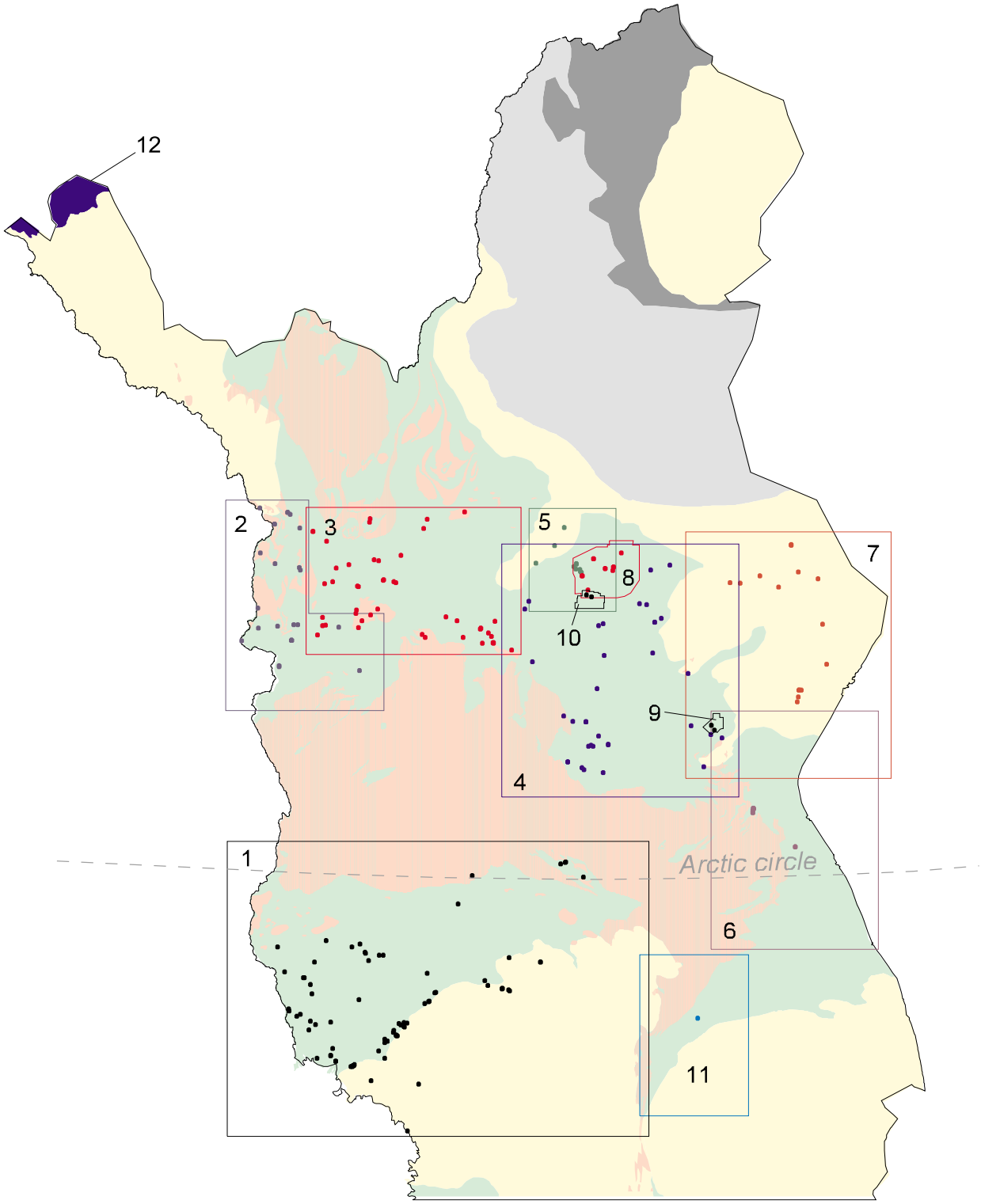

Fig. 1. The target areas of the papers in this volume: 1. Peräpohja (Perttunen & Vaasjoki), 2. Muonio-Kolari (Väänänen & Lehtonen), 3. Kittilä (Rastas et al.), 4. Sodankylä (Räsänen & Huhma), 5. Peurasuvanto (Manninen et al.), 6. Salla (Manninen & Huhma), 7. Eastern Lapland (Juopperi & Vaasjoki), 8. Koitelainen (Mutanen & Huhma), 9. Akanvaara (ibid.), 10. Keivitsa (ibid.), 11. Posio (Räsänen & Vaasjoki), 12. Halti-Ridnitsohkka (Vaasjoki & Sipilä).

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THREE DECADES OF U-Pb MINERAL ANALYSES AT THE GEOLOGICAL SURVEY OF FINLAND

by
Matti Vaasjoki

Vaasjoki, Matti 2001. Three decades of U-Pb mineral analyses at the Geological Survey of Finland. *Geological Survey of Finland, Special Paper 33,* 9-13.

An overview on the analytical methods employed for the U-Pb analyses of zircons and other mineral at Unit for Isotope Geology of the Geological Survey of Finland is given, and the methods used to present the data accumulated by the various methods over thirty years in a comparative manner are described.

Key words (GeoRef Thesaurus, AGI): absolute age, U/Pb, zircon, titanite, monazite, methods, review, Finland

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INTRODUCTION

The analytical work summarized in this volume has been carried out over a period of 30 years, which have seen a huge development in the methods employed. This applies to all stages of the work, starting from sample preparation and going on through the chemical purification of lead and uranium to the mass-spectrometric determinations of isotopic compositions. Even the final data reduction techniques have varied. Although individual methods will not be described in great detail, the development is traced through, and in the end of this section I will describe the measures that were taken in order to present the data set of this volume in as coherent a manner as possible. However, the development achieved over the past 30 years is best illustrated by an anecdote.

SAMPLE A305-LAURILA ALBITE DIABASE

As is related by Eero Hanski in his historical essay, early mappers of northern Finland had noticed curious mafic rocks, the albite diabases, of which there will be much talk of later on in this volume. They were known to penetrate the lower sedimentary sequences, but at least in the Peräpohja area, they did not enter the stratigraphically higher deposits. Thus their successful dating would clearly constrain the timing of the sedimentation.

In 1970, when I as a student was employed for the summer by the Mineralogical lab of the Department of Petrology of the Survey, a large (40 kg) sample from the Laurila diabase at Kemi had been collected. The reason was that titanite had been found in this rock in thin

sections, and it was hoped that the mineral could be dated by the U-Pb method. It was natural that the student, of whose geological prowess none could be too sure of, was given the task of mineral separation where he could not possibly cause any permanent damage.

When I had successfully separated the heavier than 3.6 g/cm^3 fraction, I spotted among the titanite occasional fragments of reddish-gray, turbid, stubby and apparently tetragonal crystals, which both Olavi Kouvo and Kai Hytönen deemed as potential zircon. A confirming x-ray powder photograph was promptly done, and I started hand-picking to obtain a clean fraction of the precious mineral. Borax fusion was then in use, at least 100 mg of the material was needed, and the mineral separate contained less than 10% of zircon. When I had to return to the university in September, the hand-picking was continued throughout the winter by Mirja Saarinen. The analysis of fraction A305A was done in June 1971, and the first results of albite diabase age determinations were published by Matti Sakko in the same year.

MINERAL SEPARATION

The usual sample size has been 10-30 kg, although especially drill core samples have been smaller. In some cases, it has been deemed necessary to resample the same locality. Generally, hand specimens of the samples are still retained by the laboratory. The samples were chopped with a hammer to about fist size, and then crushed to <1 cm by a jaw crusher. The final milling was done with a roller mill. Usually the grain size aimed at was 0.3 mm, but in the case of very fine-grained rocks was reduced to 0.2 mm.

A shaking table was used for initial coarse separation, with the aim of reducing the sample to about 2 kg by removing mainly feldspars, quartz and micas. Up to 1982, bromoform (density c. 2.8 g/cm^3) was used as the first heavy liquid, from then on the heavy liquid separation was started with methylene di-iodide (c. 3.3 g/cm^3). In the case of mafic rocks, this was followed by separation with a Carpco rotary magnet separator in order to remove magnetite and mafic silicates with magnetite inclusions. Further magnetic separation was done with a Frantz isodynamic magnetic separator.

The +3.3 non-magnetic fractions were treated further with Clerici's solution, which when heated in a water bath, can theoretically attain a density of 4.6 g/cm^3. In practice, standard solutions of 3.6, 3.8, 4.0, 4.3 and 4.5 g/cm^3 were used. In order to assist hand-picking, the samples were often sieved with 200 mesh (70 μm) and 100 mesh (160 μm) sieves.

As the discordance of particularly the early albite diabase data proved a problem, several measures were tried as a remedy. The most notable of these were slight crushing in a mortar and subsequent treatment with concentrated HF and/or HNO_3. When the air abrasion technique (Krogh 1982) became available, this was extensively used and though not always successful, proved more effective than the acid treatments.

Aliquots of the lighter fractions, including the overflow from the shaking table, and everything left over from the heavier fractions are still retained in the laboratory. It has thus been possible to analyze several samples during all four evolutionary phases (borax fusion, fraction work, air abrasion and minichemistry) of zircon work.

It should be emphasized that not all samples were treated in an analogous manner. Rather, the aim has always been to produce an as favorable a result as possible by varying routines according to the mineralogy of the particular rock studied.

EVOLUTION OF ZIRCON ANALYSIS: THE BATTLE OF BLANKS

The lead blank, i.e. the amount of lead inevitably added as laboratory contamination to the sample during its treatment has been the major factor determining the size of the zircon sample throughout the history of isotopic analyses. In the early days, when borax fusion was used for the combustion of zircons, most of the blank came from the borax. As this meant micrograms of external lead per analysis, preferably 100-300 milligrams of zircon were required. The horror story of early zircon analyses is the day-in day-out hand-picking of clean zircon fractions. As in the case of sample A305, even one analysis per sample required a tremendous amount of work.

With the advent of the hydrothermal combustion method (Krogh 1973), borax as a lead source was eliminated. The lead blanks dropped to the nanogram level, and the sample size could be cut down to 10-30 milligrams. This meant in effect that several analyses per sample could be carried out and calculating upper intercept ages became possible. The reduced sample

size also facilitated hand-picking of different morphological and color varieties. Reagents and the general laboratory environment became the principal worry.

Purification of the reagents is relatively straight forward: p.a.-grade reagents and deionized water are further purified by sub-boiling stills, as has been done already from the 1960s onwards. Thus the laboratory itself and the air it contained became the major concerns.

People building zircon laboratories have always been almost paranoid about building materials. Paints and mortars were analyzed, and any rusting components were avoided almost at any cost. Today, most laboratories are at least adequately build.

Sometimes during the 1970s the extent of the heavy metal, particularly lead, contamination of the environment by motor vehicles was realized. The pioneering work was done by Claire Paterson, who showed that the amount of lead even in the glaciers of Greenland had grown by an order of magnitude in the last century. In spite of high-quality particulate filters, air locks and laminar flow working stations, tetraethyl lead remained a major source for the blank.

We experienced this at the Survey lab in two stages. In 1986, the normal blank in the old laboratory located on ground level was 1.5 nanograms. When the laboratory was moved to a new facility to the fifth floor in 1987, the blank dropped to 0.8 ng, mainly by virtue of being moved upwards away from the parking lot. Today, as almost all cars in Finland operate with unleaded petrol, the blank in the conventional analyses has dropped again, to about 0.4 ng. This has allowed to reduce the sample size to 1-3 mg.

Another way of reducing the blank has been scaling down the volume of the chemical purification. This led to the so-called minichemistry, where the amounts of ion-exchange resin and the chemicals used was dropped by an order of magnitude when compared to the conventional technique. Routine blanks in this method are of the order 10-40 picograms, and this allows the analysis of even single (admittedly large) grains, as the sample size can be taken down to 0.1 mg. Naturally, new horror stories of spurious high blanks, often suspected to arise from variations in Teflon quality, have seen daylight.

The bulk of the data reported in this volume have been produced by the now "conventional" method of Krogh (1973). Some old borax fusion analyses are reported and defined as such in the Appendices. Minichemistry analyses are very few and can be recognized from their small sample sizes.

SPIKING TECHNIQUES

The isotope dilution technique is the most accurate way of to determine the concentration of any element. The procedure consists of adding an artificially enriched isotope, spike, to a concentration aliquot. When the isotopic compositions of the sample (IC) and the spike aliquot (ID) are measured, and the amounts of spike and sample in the aliquot are known from weighing, the concentration of the element can be accurately computed. The main asset of the technique is that, once the sample and the spike are properly mixed, sample recovery is not a factor as the measured variables are isotopic ratios.

As the isotopic composition of uranium is constant, a ^{235}U-spike is always used. For zircon and titanite analyses, where ^{206}Pb is the dominating lead species, a ^{208}Pb-spike is added. The thorogenic ^{208}Pb is the dominating lead isotope in monazites, and consequently the spiking is done with ^{206}Pb. Originally, the spikes were added as separate solutions. The idea was that the critical ratio for the ID measurement should be brought as a close to 1 as possible in order to minimize the effects of electronic bias. As electronics (or the user's confidence in them) improved, it was found that a range of 0.1-10 was quite sufficient, if the weaker isotopic signal exceeded a certain threshold, typically 10^{-13} A.

This allowed the use of mixed spikes to be considered. Thus e.g. for zircon analyses, a combined ^{208}Pb-^{235}Pb spike was produced. The use of a mixed spike has the advantage that weighing errors cancel out. In fact, accurate U/Pb ratios (and consequently analytical points for a concordia diagram) can be produced without any weighing. However, as concentration information is important for data assessment, the samples, IC and ID aliquots as well as spike amounts are still weighed.

An artificial lead spike, ^{205}Pb, has been available to isotope geologist since the middle 1980s. Its use facilitates the analyses of very small samples, as both IC and ID information can be collected during a single mass-spectrometric measurement. The problem with this spike is its limited accessibility (it has been successfully manufactured only once) and hence many laboratories, the GTK included, do not use it for routine analyses.

Most analyses reported in this volume have been carried out using separate lead and uranium spikes. A mixed spike has been used since about 1990, and the ^{205}Pb spike has not been used at all.

MASS-SPECTROMETRY

The basic configuration of the thermal ionization solid source mass-spectrometer (TIMS) has remained unaltered since its original construction in the 1930s. Though commercial instruments were available already in the 1960s, it was a feasible proposal for any laboratory to build its own machines, and this was the option selected at the Geological Survey of Finland.

Three home made instruments, MS I which is an original DTM design, MS II with a z-focusing property and MS III with an extended flight path were completed over a period of 20 years. All are single sample, single collector instruments, which have seen several modifications in their pumping and data acquisition systems. The MS II has been converted into negative ion mode for Re-Os work, which has not eventuated so far. A VG Sector 54 multicollector instrument with a 20 sample source was installed at the Geological Survey in 1990. All four instruments have been used for producing data published in this volume, but MS I has been the principal one, and therefore variations in its data treatment procedures have to be discussed.

Originally the mass-spectrometer output was monitored on a strip chart recorder, and measured by hand from the recorder charts. Interpolation by hand was a time-consuming procedure prone to errors in personal judgement, as signal instabilities had to be eliminated.

In 1974 a tailor-made electronic device which governed both the necessary peak-jumping procedure and punched the raw data on a paper tape was installed. The paper tape output was then fed via a teletype to the central computer of the Survey, and processed by a program written in the laboratory. This system was slow, as the computer tended to become sluggish whenever more than four time-sharing operators were hooked on. In addition, the punching apparatus produced errors as it sometimes punched additional holes, thus converting e.g. 0s to 4s, and these errors had to be weeded out at least from the first three digits of the five digit output. Obviously, there was no significant improvement in the speed of data processing.

The paper tape output was replaced in 1978 by output to a CD-cassette station on a TI Silent terminal, but the problem of the "mispunches" persisted. However, the data could be screened and weeded before feeding it to the central computer, which improved the result/effort ratio considerably. A change of the central computer shortened time-sharing operations into what we then considered as almost instantaneous.

From 1982 onwards, peak-jumping, data acquisition and data reduction have been done by a dedicated desk-top computer. This eliminated the "mispunches", but the inherent problem of a single collector machine remained: if there is a sudden change in the ion signal strength during a measuring sequence, two ratios within the data set are affected. Depending on the structure of the interpolation sequence and the instant of the intensity change during the data collection, one ratio is going to be eliminated by the 2σ-criterion, but the other one may remain in the data set even after a repeated 2σ-screening. Therefore, weeding of the raw data is still a necessity.

DATA REDUCTION AND AGE CALCULATIONS

In the early days, when only one analysis per sample was feasible and the analyses were often discordant, modeling was the only possibility to extract some sort of age estimates. At the GTK this was done using the radiation damage model of Wasserburg (1963). An alternative was to combine analyses from several presumably genetically related samples, and use a regression calculation to obtain an upper intercept age. Thus the advent of fraction work with its possibility of obtaining upper intercept estimates for individual samples was a vast improvement in zircon dating.

The use of the upper intercept of a linear regression as an age estimate is well founded. It has been shown that irrespective of the mechanism of lead loss (cf. Wetherill 1956, Tilton 1960, Wasserburg 1963), zircons form linear trends on the concordia diagram unless they are very discordant. Of course, this applies only to zircons which have been subjected to just one type of lead loss. If two or more processes have been operative at any time during the geological history of the sample, the results will scatter randomly, and the solution becomes a matter of geological interpretation. This aspect is discussed in several individual papers of this volume, particularly in the summary by Hanski et al.

Although the basic linear regression system at the laboratory has been based on the method of York (1969) since the middle 1970s, details particularly in the error estimates have varied. Moreover, calculating the mean squares of weighted deviates (MSWD) was not systematically done until the 1990s. It was therefore decided that all age calculations should be brought to the same footing by recalculating them with

the widely circulated and used Isoplot/Ex program (Ludwig 1998).

When Hannu Huhma and I started to scrutinize the data for this purpose, we ran into a vexing problem: in some cases, the error estimates for analyses were clearly overly confident, most likely due to excessive weeding of bad raw data. After some testing and much soul-searching we came to the conclusion to apply a minimum error to most of the analyses done with separate uranium and lead spikes, and use the calculated errors only when they exceeded this limit. The values we chose are: 0.65% for the Pb/U ratios, 0.15% for the $^{207}Pb/^{206}Pb$ ratio and 0.97 for the error correlation.

The linear regression calculations are very conservative. As the volume contains many samples with over ten analyzed zircon fractions, it would have been very tempting to enter a numbers game and fiddle with several discordia lines arrived at by rejecting various data points. We have striven to avoid this as much as possible. Usually, if data are rejected, there is a valid reason, such as differences in morphology and/or color, physical properties or extreme discordance. Often a reference line is reported meaning that the upper intercept is probably in the right ball park, but cannot be considered a valid age.

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Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
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HISTORY OF STRATIGRAPHICAL RESEARCH IN NORTHERN FINLAND

by
Eero Hanski

Hanski, Eero 2001. History of stratigraphical research in northern Finland. *Geological Survey of Finland, Special Paper 33*, 15–43. 6 figures.

An overview on the results of stratigraphical research carried out in northern Finland (Central Lapland, the Kuusamo and Peräpohja schist belts) since the early 20th century is presented. The traditional stratigraphical names, Lapponian and Kumpu and their derivatives, which have been widely used in northern Finland, were defined in the 1920s and 1930s when the stratigraphical interpretations were based on establishing cycles of sedimentation separated by diastrophisms and intrusions of granites with then unknown ages. Later results of isotopic and geological work have forced the investigators to revise their stratigraphical schemes and regional correlations, but the nomenclature rooted to the traditional names has continued their life until recently. Particularly, the chronostratigraphical position of the Lapponian and Kumpu rocks and their lithostratigraphical correlation with the traditional Karelian formations have been a controversial issue over the years. The review illuminates the reasons that led to this long-lasting lack of consensus and explains why the abandonment of the old names along with the adoption of the formal lithostratigraphical nomenclature was still necessary as late as in the 1990s.

Key words (GeoRef Thesaurus, AGI): stratigraphy, lithostratigraphy, chronostratigraphy, greenstone belts, schist belts, metamorphic rocks, Paleoproterozoic, Archean, research, history, Kuusamo, Peräpohja, central Lapland, northern Finland

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INTRODUCTION

We have all been hopelessly wrong at one time or another, and it would ill behove us to hold the great pioneers of the Precambrian in any the less esteem because we have the advantage of techniques undreamt of by them.

Arthur Holmes, 1963

Field studies are the cornerstone of all stratigraphical interpretations and models of geological evolution. However, in structurally complicated and poorly-exposed Archean and Paleoproterozoic terranes, field investigations alone have not always produced unambiguous results concerning the order of supracrustal sequences and their regional correlations. In the absence of fossil-bearing strata in the early Precambrian, radiometric datings have become invaluable in providing indirect or direct age information for supracrustal rocks. It is obvious that the first U-Pb zircon ages obtained in Finland quickly brought about fundamental changes in the conceptual framework of the Precambrian geology within the Fennoscandian

Shield (Kouvo 1958, Wetherill et al. 1962). However, despite the increasing number of radiometric data produced since then, their major impact on solving stratigraphical problems in northern Finland took place relatively late.

The Precambrian rocks in Finland are divided into three main units: 1) the Svecofennian terrane in the southwest representing a juvenile Paleoproterozoic arc complex, 2) the Archean granite-gneisses and greenstone belts in the northeast, and 3) the Paleoproterozoic supracrustal sequences, assigned commonly to the Karelian formations, which overlie and, at least for the most part, were originally deposited on the Archean sialic basement. The focus of this paper is on the third unit, particularly the Karelian metasedimentary and metavolcanic rocks occurring in Central Lapland and the Peräpohja and Kuusamo schist belts.

The Karelian supracrustal successions reach several kilometres in total thickness, are spread intermittently over an area occupying the northeastern half of the Fennoscandian Shield and record a geological evolution of several hundreds of millions of years. Due to their wide development in space and time, the stratigraphy of the Karelian formations and their correlation between different areas have drawn much attention since the beginning of the 20th century. The Karelian stratigraphy has often been described in terms of adjectival names such as Sariolian, Jatulian and Kalevian (e.g., Meriläinen 1980a,b). These poorly-defined, informal names, derived from the Karelian folklore, are deeply ingrained in the Finnish geological literature, but have been used in an inconsistent manner and have also turned out to be insufficient for accurate correlations (e.g., Laajoki 1986, 1988). The historical background of the definitions and subsequent development of these commonly used stratigraphical terms are not explained here, accounts of that can be found in Simonen (1955, 1986) and Laajoki (1986) and references in them. For the partly different nomenclature used in the eastern part of the Fennoscandian Shield, the reader is referred to Russian papers such as Kratts et al. (1984), Zagorodny et al. (1986) and Semikhatov et al. (1991). In northern Finland, additional terms, the Lapponian and Kumpu formations and their derivatives, have been widely employed, and on their relation to other Karelian formations, several differing opinions have been presented in the geological literature over the years.

In the 1980s, there was a tendency in Finland to combine the traditional (mythological) names with lithostratigraphical rank terms (e.g., Marmo et al. 1988, Luukkonen & Lukkarinen 1986, Silvennoinen 1985, Geological Map, Northern Fennoscandia 1987) but this was not regarded as satisfactory by everyone as reflected, for example, by the following statement by Marmo and Ojakangas (1984): "*Division of the rock sequences on the Baltic Shield into groups and formations apparently has not been done as rigorously as the Code of Stratigraphic Nomenclature prescribes for North American rock sequences.*" Nevertheless, this seemed to remain as a temporary practice before shifting to the more strict lithostratigraphical system based on international recommendations in which the stratigraphical names are formed by combining local geographic names of the type areas with the proper rank terms (Hedberg 1976, Salvador 1994). Applying this procedure, classification of the Karelian formations into formal lithostratigraphical groups and formations gradually took over in the Kainuu and northern Karelia schist belts in central and eastern Finland (Kontinen 1986, Gehör & Havola 1988, Strand 1988, Laajoki 1991, Pekkarinen & Lukkarinen 1991, Kohonen & Marmo 1992). Lately, the same classification system has also been applied to rock successions in northern Finland, both in the Peräpohja area and in Central Lapland (Perttunen et al. 1995, Räsänen et al. 1995, Lehtonen et al. 1998). Most geologists now prefer to restrict the use of the traditional names to broad, informal discussions only or, as done by Laajoki (1988, 1990), to the concepts of tectofacies.

Besides the lithological correlation with traditional Karelian stratigraphical units, the interpreted geochronological position of the supracrustal rocks in Lapland has varied much, from Archean to Proterozoic, in the articles by different researchers. This has evidently caused confusion among Finnish geologists as well as foreign investigators trying to make a synthesis on the geological evolution of northern part of the Fennoscandian Shield (e.g., Pharaoh & Brewer 1990, Brewer & Pharaoh 1990, Goodwin 1991). One of the reasons for differing views is the geological diversity of rocks in Lapland, making them partly similar to other Karelian formations, but on the other hand, there exist some rock units, for instance thick piles of komatiites, which hardly have any counterparts in Karelian formations in eastern or central Finland. Until recently, there has been a lack of reliable geochronological data to constrain the ages of certain sedimentary and volcanic rock units, but the situation is gradually improving as shown, for example, by the accompanying articles in this volume.

At this stage, when the new lithostratigraphical names have just been adopted and consequently more perplexity may still arise, it is prudent to review the development of stratigraphical concepts in northern Finland over the past century. It is hoped that this will

lead to a better understanding of the reasons for difficulties which were encountered by the previous investigators and hindered the achievement of endurable and uniformly accepted stratigraphical schemes in Lapland. In the following treatise, the Peräpohja schist belt is dealt with separately from the Central Lapland Greenstone Belt and its continuations in the Kuusamo schist belt because these two entities form geographically distinct areas and have usually been described separately using different stratigraphical terms. It is assumed that the reader has basic knowledge of the geology of northern Finland, as it is beyond the scope of this paper to present any detailed geological review.

HISTORY OF STRATIGRAPHICAL STUDIES IN CENTRAL LAPLAND AND KUUSAMO

The Paleoproterozoic greenstone belt considered here forms a volcano-sedimentary belt extending from Kuusamo through Salla, Sodankylä, Kittilä and Kolari to the western border of Finland and from Kittilä to the north across the Finnish-Norwegian border. This is the area where the terms 'Lapponian' and 'Kumpu' have generally been applied.

The 1920s to 1940s

Attempts at rationalization of the stratigraphy of Central Lapland were initiated by **Hackman (1927)** in his pioneering study of the Kittilä-Sodankylä area. His stratigraphical concepts are shown schematically in Figure 1. Hackman regarded the gneiss granites of Muonio, west of Kittilä, as the oldest rocks, on which two sequences of supracrustal rocks had been deposited. The older one of these sequences, which he called the "*older schists*", was divided into a lower and upper series with no apparent hiatus between them. The lower series comprised pelites, quartzites and carbonate rocks and also the iron formations of the Porkonen-Pahtavaara area occurring in the midst of the Kittilä greenstones, while the upper series consisted of quartzites, the "Sodankylä quartzites", and the spilitic metabasites (now known as mafic metavolcanics) of the Kittilä area. In the southern part of the area, the metasediments of the upper series are intersected by granites which belonged to the large "post-Kalevian" Central Lapland granite area. Hackman (1927) considered it likely that the metavolcanics are also older than these granites.

For the younger sedimentary sequence in the Kittilä-Sodankylä area Hackman (1927) coined the name "*Kumpu quartzites*", after the type occurrence at Kumputunturi. These metasediments, which form laterally discontinuous bodies extending from the Kaarestunturi area in Sodankylä to Pyhätunturi east of Kittilä, comprised coarse-clastic quartzites and polymictic conglomerates with brownish purple colour locally. One of the best representatives of the basal conglomerates was shown to be the Sirkka conglomerate on the northern flank of the fell Levitunturi. Hackman presented ample evidence for the existence of an unconformity between these sedimentary formations and the "older schists", including the presence of pebbles and cobbles in the basal conglomerates and conglomeratic quartzites, the rock types of which were unmistakably the same as in the older underlying rocks of the area. These pebbles were partly composed of quartzites and phyllites similar to those occurring in the "older schists", and also metabasites (metavolcanics and metadiabases) and iron ores and jaspers, evidently from the Porkonen-Pahtavaara area. Hackman also established that the Kumpu quartzites had different strikes and dips from the older rocks in places, indicating the presence of an angular unconformity.

As mentioned above, Hackman (1927) regarded

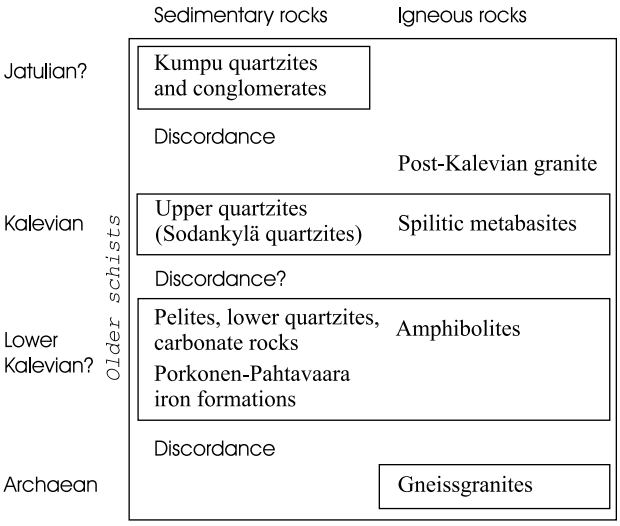

Fig. 1. Stratigraphical sketch of Hackman (1927) for the Kittilä-Sodankylä area in Central Lapland.

the granites cutting the "older schists" as similar to the "post-Kalevian" granites that were known to have penetrated Kalevian rocks in northern and eastern Finland. It was therefore logical to assign the "older schists" and the large volcanic region in Kittilä to the Kalevian. It is important to stress here that since the works of Frosterus (1902) and Ramsay (1902), the Kalevian schists were generally regarded as older than the Jatulian rocks, and consequently Hackman's "older schists" were considered pre-Jatulian.

Hackman (1927) could not draw any firm conclusion on the relationship between the "post-Kalevian" granites and the Kumpu quartzites. He had observed neither granitic dikes cutting into the Kumpu quartzites or their basal conglomerates nor any typical post-Kalevian granites as boulders in the conglomerates. The conglomerates do have granitic fragments, but according to Hackman, they could not be traced back to their source. Yet he was convinced that there was a great age difference between the Kumpu quartzites and the quartzites from the "older schists" and considered it probable that the former were younger than the post-Kalevian granites. In his stratigraphical scheme, Hackman correlated the Kumpu quartzites with the Jatulian formations, although with reservations on account of some important observations related to lithological differences and the occurrence of mafic dikes within the respective formations. For example, he pointed out that Kumpu-type polymictic conglomerates were absent from the Jatulian formations of the Rovaniemi-Tervola area in the Peräpohja schist belt, where the rocks were very similar to the quartzites of his older sedimentary unit. Hackman also drew attention to the fact that intrusive metadiabases and metagabbros were very common within Jatulian metasediments in the Kuusamo and Rovaniemi-Tervola areas but were practically non-existent within the Kumpu quartzites. He knew only one example, from Kaarestunturi. Instead, he maintained that such rocks occurred widely as boulders in conglomerates of the Kumpu metasediments. It is worth noting here that Hackman (1918) had already earlier wondered about the absence of intersecting metadiabases in the Pyhätunturi conglomerates of the Pelkosenniemi area.

The next contribution to the stratigraphy of Lapland was provided by **Sederholm (1932)** in his explanatory notes to the Geological Map of Fennoscandia. As stated earlier, Sederholm (1932) was the first to introduce the later frequently used term '*Lapponian*' for the widely distributed volcanic and sedimentary rocks of Central Lapland. To fully understand the terminological development, it is imperative to form an idea of the methods of stratigraphical classification employed at that time. As no radiometric age determinations were available, Precambrian sequences were classified on the basis of their relations to epochs of diastrophism, as manifested by intrusions of granites, for example. Using this method, Sederholm (1932) divided the Precambrian of Fennoscandia into four *cycles of sedimentation* (Fig. 2). The oldest, Katarchean rocks were represented by Svionian sedimentary and effusive rocks in the Svecofennian belt of southern Finland, which were cut by post-Svionian granites. Younger than these granites were the Younger Archean Bothnian sedimentary and volcanic rocks in southern and central Finland which were in turn intruded by post-Bothnian granites. The third cycle of sedimentation, the Karelidic cycle, comprised the Jatulian and Kalevian rocks, penetrated by post-Kalevian granites.

Sederholm (1932) assigned Hackman's (1927) Kumpu quartzites to the Jatulian rocks. According to him, the Kumpu formation was unconformably overlying a sequence of greenstones and associated schists which had a similar position to the Bothnian rocks between the post-Svionian and post-Bothnian granites. Hackman (1927) had correlated the same rocks with the Kalevian schists, but this correlation was no longer tenable since, as the result of the work of Wegmann (1929), Väyrynen (1928) and Hausen (1930), it had turned out that the Jatulian and Kalevian rocks of eastern Finland belonged to one Karelidic cycle of deposition and, furthermore, the Kalevian rocks were probably younger than the Jatulian ones. In contrast to the Karelidic cycle of deposition in eastern Finland, a great unconformity separated the different parts of the sedimentary rocks in Lapland, which had been collectively referred to as "Karelian". Sederholm thus considered that instead of one, at least two cycles of deposition and diastrophism had occurred in Lapland. This was supported by the interpretation that in addition to the post-Jatulian (post-Kalevian)

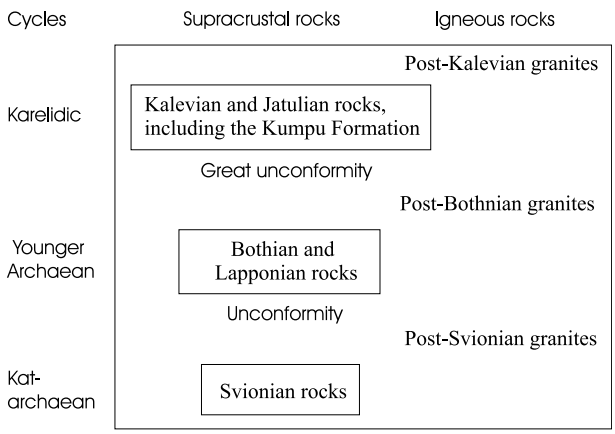

Fig. 2. Three oldest of the four cycles of sedimentation established by Sederholm (1932).

granites, the lower supracrustal rocks were also cut by an older plutonic suite, fragments from which are found in the basal conglomerates of the Kumpu metasediments.

This reasoning of Sederholm thus negated the possibility of correlating the Central Lapland greenstones and associated schists with either the Jatulian or Kalevian rocks. On the other hand, despite the inferred similar position to the Bothnian schists, Sederholm could not assign the Lappish sedimentary and volcanic rocks to the Bothnian schists because of differences in their primary characters. He concluded that these two formations were of different ages but, nevertheless, belonged to the same cycle of sedimentation. Therefore, Sederholm regarded it as necessary to adopt a special name, Lapponian, for these (pre-Jatulian) supracrustal rocks of Lapland.

It is important to note here that Sederholm (1932) did not exclude the possibility that some quartzites of northern Finland were even younger than the Jatulian proper. He was referring in this context to the Kumpu formation, the correlation of which with the Jatulian rocks of the Kemi and Kuusamo areas he regarded as not yet having been fully verified.

The general geological mapping of Finland on a scale of 1:400,000 by the Geological Commission (since 1945 the Geological Survey of Finland) was initiated in the 1890s, and was later also extended to Lapland. As the result of fieldwork extending from 1929 to 1935, **Erkki Mikkola's (1941)** posthumously published explanation to three map sheets (Tuntsajoki-Sodankylä-Muonio) covered the territory of Central Finnish Lapland in an E-W direction from the Soviet border to the Swedish border. For decades to come, this comprehensive treatise of Mikkola served as a basic source of lithological and stratigraphical information on Lapland. Many of the regional or local geological units recognized by him are still essentially valid.

The oldest geological unit distinguished by Mikkola was the Tuntsa-Savukoski supracrustal series, occurring in eastern Lapland, which was composed mainly of mica gneisses, quartzites and amphibolites. The next chronological group of rocks was the gneissose granites and granite gneisses, which occupy large areas of eastern Lapland but are encountered in other parts of the mapped territory as well. According to Mikkola, these gneisses definitely served as a basement for the supracrustal rocks, which he designated as the Lapponian Series, applying the term of Sederholm.

Mikkola (1941) divided the Lapponian metasediments into quartzites and schists, with the latter group being further divided into mica schists, highly aluminous schists and black schists. Mikkola considered the eruption of the associated volcanic rocks to be chronologically very close to the deposition of these sediments. He was able to recognize two main types of volcanic rocks, one with ordinary basicity, the so-called Kittilä greenstones, and the other with a more basic chemical nature, the amphibole-chlorite rocks, later to be known as komatiites (Mutanen 1976). Although they were areally fairly well separated, Mikkola thought that these two types of volcanic rocks were genetically related. The jasper quartzites and iron ores associated with the Kittilä greenstones in the Porkonen-Pahtavaara area were no longer regarded as epiclastic sediments but as chemical precipitates whose origin was closely linked with the volcanism. These metasediments were earlier documented as the oldest supracrustal rocks in Central Lapland (Hackman 1927, Sederholm 1932), but this view no longer fitted with Mikkola's interpretations.

Hackman (1927) had already demonstrated the existence of an unconformity beneath the Kumpu formations, and Mikkola (1941) agreed with him, regarding the evidence for the unconformity as convincing. In many areas, however, Mikkola had not found clear field evidence for an unconformity between Hackman's older quartzites and the quartzites of the Kumpu formations, and he therefore expanded the younger group to include more quartzites, especially in the southern and eastern part of Hackman's area, and even a considerable area of mica schists occurring around Oraniemi, south of Sodankylä. This extension of the geographical distribution of the younger group of sediments was accompanied by renaming the unit as the Kumpu-Oraniemi Series. Mikkola acknowledged a clear facies difference between the Kumpu quartzites and the other nearby quartzites, but still regarded both as post-Lapponian. It was shown later that the Oraniemi mica schists and many of the quartzites included by Mikkola in the younger sediments are not part of the Kumpu sediments, and hence Hackman's interpretation is closer to the scheme accepted by subsequent investigators (e.g., Rask 1978, Tyrväinen 1983).

As noted by Hackman (1927), the Sirkka conglomerates contain fragments of obviously intrusive granitic rocks, including gneissose granites and aplitic rocks of varying coarseness. Neither he nor Mikkola (1941) could confidently trace the source of these fragments, although Mikkola considered the aplites to be possibly derived from the 'syenite series' (later to be called the Haparanda series by Ödman et al. 1949 and the Haaparanta Suite by Finnish geologists, e.g., Lehtonen et al. 1998) which forms a characteristic igneous suite in western Lapland and the western part

of the Peräpohja schist belt.

The other issue concerning the granites is Mikkola's observation that the westernmost Kumpu quartzites in the Muonio area are traversed by granitic dikes and that these quartzites are associated with typical Sirkka conglomerates that have been strongly migmatized. Two decades later, Eskola (1963) emphasized this observation, which, according to him, suggests that the Kumpu-Oraniemi Series belongs to the same orogenic cycle as the Lapponian Series.

The correlation of the Lapponian Series with other stratigraphical units in Finland was regarded by Mikkola as an important problem, but he conceded that different conclusions may be arrived at depending on whether volcanic or sedimentary rocks are concerned in the first place. He tentatively correlated the Lapponian metavolcanics with the volcanic rocks of eastern Finland. Wilkman (1921) had described abundant basic metavolcanics and more basic chlorite-amphibole rocks in the Nurmes map sheet area (D4), and according to Väyrynen (1933), these rocks were similar to those occurring in the basement of the Jatulian system in the Pielisjärvi area (map sheet D3). Other counterparts of the Lapponian metavolcanics could be found in the Kuhmo-Suomussalmi greenstone belt, which in Mikkola's opinion could also precede the Jatulian.

When considering the sedimentary rocks of the Lapponian system, a different comparison had to be performed by Mikkola, since the above-mentioned greenstone areas in the basement do not contain many epiclastic sedimentary rocks. He admitted that the lithological similarity of the quartzites in the Jatulian and Lapponian systems and their widespread regional development as basal units would strongly suggest a common period of deposition, but because no such voluminous eruption of volcanic rocks as had taken place in Lapland as recorded in Jatulian successions elsewhere, the Jatulian-Lapponian correlation also seems to be excluded with regard to sedimentary rocks. A further indication of Mikkola's unwillingness to draw any firm conclusions on the regional correlations is revealed by the following passage: "*But as the author in any case considers the Lapponian formations, which have clearly a basement of gneissose granites etc. in E. Lapland, as joining more closely the Karelides in respect to the age of deposition and time of folding than other, more ancient orogenic cycle of the Svecofennides which occurred in Finland ...*" Mikkola considered it possible that the Kumpu-Oraniemi Series may correspond to the Jatulian or Kalevian rocks farther south, based on the assessment that the underlying Lapponian rocks are older than those stratigraphical units. However, this correlation was not without obvious difficulties due to the lithological deviations. Firstly, Mikkola stressed the lack of carbonate rocks and black schists in the Kumpu-Oraniemi Series, which are typical of a complete Jatulian succession. Secondly, the Kumpu-Oraniemi Series hardly bears more resemblance to characteristic flysch-type sediments of the Kalevian sequences, such as occur, for example, in the classic area southwest of Lake Pielinen (Northern Karelia).

Erkki Mikkola's geological research in Lapland was interrupted by the outbreak of the war between Finland and the Soviet Union in 1939. The commentaries to the map sheets (Mikkola 1941), completed by his colleagues Th. Sahama and K. Rankama, were published posthumously after Mikkola's premature death at the front in 1940 at the age of 35 years (see Eskola 1941a,b). Several important chapters of the explanation planned to be written by Mikkola, including that dealing with tectonic development of southern Lapland, were never realized (see Laitakari 1941).

The 1950s and 1960s

The next significant contribution to the discussion on the Lappish stratigraphy was made by **Väyrynen (1954)** in his book 'Suomen Kallioperä' (The Bedrock of Finland). This was largely based on the accounts of Hackman (1927) and Mikkola (1941), but he interpreted the available data in the light of the wide knowledge of the Karelian formations which he had achieved by studying these rocks in Kainuu and Northern Karelia (Väyrynen 1928, 1933).

Väyrynen (1954) pessimistically stated that igneous and sedimentogenic rocks lie in utter mutual disorder in Central Lapland, and therefore unravelling the detailed stratigraphical structure is a quite hopeless task in most parts of the area. Nevertheless, one clearly discernible stratigraphical unit, the Kumputunturi formation, can be shown to have been deposited discordantly on the Kittilä greenstones. These older greenstones and the associated metasediments did not show any signs of their mutual age relationships or stratigraphical positions. With regard to the sedimentological facies, however, the quartzites, phyllites and dolomites represented mainly the same formation types as occur in the Peräpohja schist belt, for example, and they thus probably correspond to the

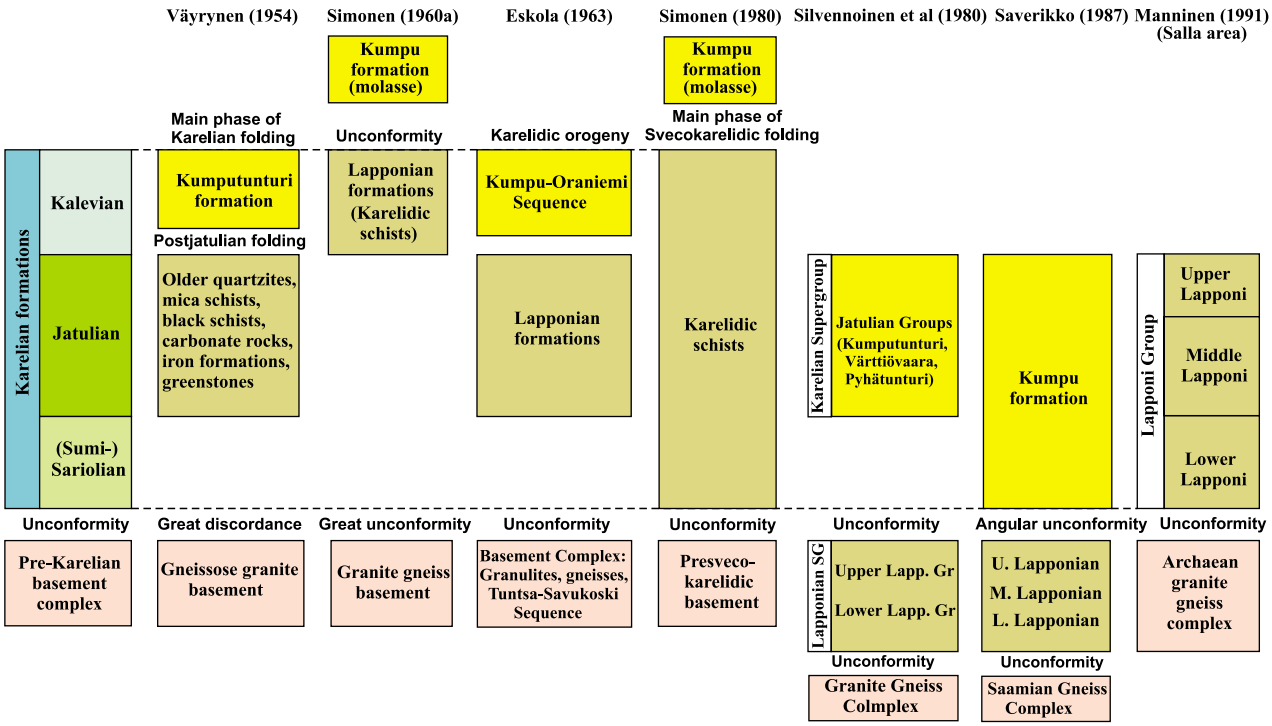

Fig. 3. Correlation of the Lapponian and Kumpu formations in Central Lapland with the traditional Karelian formations as presented by various authors. (See also the column of Silvennoinen, 1992, in Fig. 4).

Jatulian rocks. Väyrynen pointed out that only the Kumputunturi-Sirkka metasediments had earlier been included in the Jatulian because the youngest sedimentary formations had usually been called Jatulian irrespective of their depositional type.

Väyrynen (1954) hence regarded the older supracrustal group in Central Lapland as corresponding to the Jatulian (Fig.3), and consequently the Kumputunturi formation was bound to be post-Jatulian. This correlation with the Jatulian formations explains why he did not need to use the term Lapponian at all. He placed the Kumputunturi formation in the Kaleva in his stratigraphical scheme. Although the rock types of these two correlated units differ markedly in general, Väyrynen offered the Kalevian Jaurakka conglomerate in the Kainuu schist belt as an analogy for the Kumpu formation. Because of the relatively low metamorphic grade of the metasediments of the Kumputunturi formation, Väyrynen suggested that the Central Lapland granites do not intersect the Kumputunturi metasediments, and considered these rocks to represent molasse-type sediments deposited during folding. **T. Mikkola (1961)** also emphasized the molasse-like nature of the Kumpu formation. In his comprehensive overview of the Precambrian of Finland, **Eskola (1963)** shared Väyrynen's view on the correlation and held that it would seem natural to correlate the Lapponian supracrustals with the Jatulian rocks and the Kumpu-Oraniemi Series with the Kalevian rocks (Fig. 3). Referring to the Lapponian rocks, he stated: "*The sedimentary formations in Karelia, Kainuu and Lapland are everywhere similar in character, and no geological evidence of an age difference has been found.*"

Differing from Väyrynen (1954), **Simonen (1960a**, p. 26) correlated the Lapponian succession in Lapland with Kalevian micaceous schists (Fig. 3), which represent a geosynclinal association deposited after the Jatulian transgressive series. On the other hand, he regarded the arkoses and conglomerates of the Kumpu-Oraniemi formation as molasse-type sediments deposited at the end phases of the Karelidic orogenic cycle. With regard to the sedimentological facies, Simonen's opinion was compatible with that presented by Väyrynen (1954) a few years earlier, but he assessed the Kumpu-Oraniemi Series as being post-Kalevian rather than Kalevian, and hence maintained that it had no obvious counterparts in eastern Finland.

At this point, it is appropriate to review the revolution in geological thinking which took place in Finland in the late 1950s and early 1960s along with the appearance of the first radiometric dating results, especially by virtue of the pioneering work of O. Kouvo. Until then, the main means for placing the supracrustal belts in chronological order was the age relations of the schists to granitic plutons of supposedly different ages. As described earlier, this method led Sederholm (1932) to regard the Svecofennian

orogenic belt as considerably older than the Karelian orogenic belt, with both of them belonging to two independent orogenic cycles. In addition, he considered the Lapponian to be part of the same cycle of sedimentation as the Bothnian rocks of southern Finland.

When Kouvo (1958) and Wetherill et al. (1962) published their U-Pb zircon ages for orogenic plutonic rocks from both of the above-mentioned orogenic belts, it became immediately apparent that the prior assumptions on the markedly different ages of the post-Svionian, post-Bothnian and post-Kalevian granites were no longer valid, as the Svecofennian and Karelian plutonic rocks proved to be contemporaneous, with ages of c. 1.8 Ga. Additional geochronological data provided by Kouvo and Tilton (1966) confirmed the result further. The inescapable conclusion was that the Svecofennian and Karelian formations belonged to the same Svecofenno-Karelian orogenic belt (Simonen 1960b). The other significant geochronological result was that the ancient pre-Karelian gneissic basement and mantled gneiss domes in the Karelidic belt in eastern Finland yielded much older ages, in the range of 2.6-2.8 Ga (Wetherill et al. 1962, Kouvo & Tilton 1966).

The Svecofennian and Lapponian supracrustal rocks had previously been assigned to the Archean (e.g., Simonen 1953, Mikkola 1941), and when classifying the Karelian supracrustal sequences, Sederholm (1932) had stated that these rocks are generally similar to formations that have been regarded as Proterozoic in other regions, probably referring to the Huronian formations in Canada. The advent of radiometric age determinations gave new geochronological connotations to the words Archean and Proterozoic, although they continued to be used quite loosely until the general international agreement in the late 1970s to define the boundary at 2500 Ma (James 1978). The early results indicated that the Karelian formations were deposited between 2.6 and 1.8 Ga ago, and since then they have been generally regarded as Proterozoic. The Archean, pre-Karelian basement extends from southeastern Finland to eastern Lapland. Mikkola (1941) already regarded the Tuntsa-Savukoski Series in eastern Lapland as part of this basement complex, and the Archean age of that series has never been questioned since (see Juopperi & Veki 1988). The situation is different, however, where the Lapponian Series is concerned. This series has usually been considered Proterozoic, but in the late 1970s and the 1980s several authors expressed opinions favouring a late Archean age. This issue will be addressed in some detail later on.

Returning to our consideration of the stratigraphy of Central Lapland, geological research continued in the sixties and seventies mainly in the form of thematic studies carried out by university students and geologists employed by prospecting companies. Among the works touching upon wider stratigraphical aspects, the university thesis of **Mäkelä (1968)** concerning the Sirkka conglomerates and quartzites deserves mentioning. He regarded these sedimentary rocks together with the metasediments of Kumpututunturi proper as molasse-type accumulations. Apparently relying more on contemporary theoretical models of the geological evolution than on field evidence, he also outlined a model for the whole stratigraphy of Central Lapland. While the age relationship between the lower quartzitic sedimentary rocks and the Kittilä metavolcanics had been more or less obscure up to then, Mäkelä's evolutionary model placed the volcanites above the majority of these sediments in the stratigraphy. In addition, he distinguished certain aluminous mica schists and conglomerates from the other sediments and classified them as flysch-type accumulations deposited after the eruption of the Kittilä greenstones. These included the pelitic sediments of Oraniemi, which Mikkola (1941) had incorporated into his Kumpu-Oraniemi Series. Since then, the Oraniemi rocks, along with the geographical name, have usually been excluded from the youngest sedimentary group of Lapland.

Mäkelä (1968) surmised that the Central Lapland granite area had been uplifted and eroded and had acted as a source for the flysch-type deposits. After this stage, a long tranquil period had ensued before the deposition of the coarse-grained molasse-type Sirkka and Kumpu sediments. One of the interesting observations reported by Mäkelä is the presence of undeformed microcline granite pebbles in the Sirkka conglomerates. On the other hand, he described pegmatitic and aplitic dikes cutting quartzites on the hill of Aakenusvaara, for example.

The 1970s

Paakkola (1971) performed a detailed study of manganiferous iron ores and their environment in the Porkonen-Pahtavaara area, located within the Kittilä volcanic complex. He agreed with the explanation of Mikkola (1941) concerning the origin of the ores as related to submarine volcanic emanations. Paakkola's

stratigraphical scheme for Central Lapland was essentially the same as that proposed by Mäkelä (1968). He fixed the Kittilä volcanic complex and its iron formations in the pre-flysch stage of the Lapponian geosyncline development and the volcanites as products of the initial magmatism of the Karelidic orogeny.

The systematic geological mapping of Lapland to a scale of 1:400,000 was completed in Lapland in 1965, but in the meantime the Geological Survey (GTK) had initiated remapping of the bedrock of northern Finland to a scale of 1:100,000 in the early 1960s, with the Salla greenstone area and the eastern part of the Kuusamo schist belt as the first target areas. As a result of this mapping activity, **Silvennoinen (1972)** worked out a detailed stratigraphical subdivision for the Kuusamo schist belt. Earlier this region had been mapped by Hackman and Wilkman (1929) who distinguished older Kalevian mica schists, dolomites and metabasites in the north and younger Jatulian quartzites, dolomites and metabasites in the south, with a discordance between these two groups. The Jatulian rocks were thought to be separated from the oldest rocks, the pre-Kalevian gneisses and gneiss granites further south, by a tectonic contact.

The stratigraphical order of the main lithological units of the Kuusamo schist belt (older Kalevian, younger Jatulian), as presented by Hackman and Wilkman (1929), had already been rejected by Mikkola (1941). Silvennoinen (1972) then demonstrated the existence of a conglomerate at the base of the Jatulian succession, against the old (Saamian) granite gneiss complex, and he also divided the Karelian stratigraphical sequence into 11 formations, following the basal conglomerate, namely: the Greenstone Fm I, Sericite quartzite Fm, Sericite schist Fm, Quartzite schist Fm, Greenstone Fm II, Siltstone Fm, Greenstone Fm III, Rukatunturi quartzite Fm, Dolomite Fm and Amphibole schist Fm. The uppermost Amphibole schist Formation is composed of mafic tuffs intercalated with dolomites and black schist and also mica schists in the upper part of the formation. Silvennoinen could easily correlate this formation, coupled with the underlying Dolomite Fm, with the marine Jatulian formation of Väyrynen (1933) in southeastern and eastern Finland, implying that Kalevian rocks are absent in the Kuusamo area. The lower parts of the succession were more difficult to correlate, especially because of the presence of the lowermost volcanic unit resting immediately on the basal conglomerate, an unknown situation at that time elsewhere in the Finnish part of the Karelian belt. The correlation of the basal conglomerate with the Sariolian formations could not be confirmed because of the lack of a weathering crust above it.

A pronounced increase in interest in the Archean greenstone belts of the ancient Precambrian shields was manifested world-wide in the 1970s, and this trend also reached Finland during the second half of the decade. The recognition of komatiites, one of the most characteristic rock types of the Archean greenstone belts, in Finnish Lapland (Mutanen 1976) was among the reasons that apparently led **Gaál et al. (1978)** to suggest that the greenstone belts in Finnish Lapland are Archean in age and thus comparable with the Archean greenstone belts of eastern Finland (Kuhmo-Suomussalmi, Ilomantsi). Several separate volcanite-dominated greenstone belts were delineated by these authors on the geological map of Lapland, including those of Salla, Jauratsi and Kittilä. On the other hand, the Kumpu formation and most of the Lapponian epiclastic sedimentary rocks were regarded as younger than the greenstone belts and probably Proterozoic in age.

The 1980s

A three-year research project was undertaken at the University of Oulu in the late 1970s, one of the principal tasks of which was to elucidate the structure and stratigraphy of the bedrock in the Kittilä-Sodankylä area. After field studies conducted in several separate subareas, **Kallio (1980)** made a synthesis of the mapping profiles and formulated a general stratigraphical scheme for the supracrustal rocks. According to him, the evolution started with the deposition of an assembly of sedimentogenic rocks on the Archean basement, including orthoquartzites, greywackes and mica schists. These were followed by the eruption of volcanites, varying laterally from mafic (spilitic) to ultramafic. The transition from sedimentary to volcanic processes was a gradual one, as indicated by the presence of sedimentary interlayers within the metavolcanics. Following earlier investigators, the coarse-clastic Kumpu quartzites and conglomerates were distinguished as the youngest supracrustal rock unit.

Kallio (1980) also published an age determination of c. 2213 Ma for an albitite. This sample had been taken from a contact with the adjacent quartzite and was explained as having originated through a process of adinolization from a greenstone with purported volcanic structures. According to Kallio, this age demon-

strated that, contrary to the suggestions made not long before by Gaál et al. (1978), not all the volcanism in the Kittilä greenstone area could be of Archean age. However, due to the poor characterization of the dated rock, the involvement of secondary processes in its origin and the fact that similar ages have been obtained for mafic intrusive sills in the area (Tyrväinen 1983), it is difficult to assess how precisely the figure obtained is related to the actual age of the volcanism.

A sufficient number of radiometric age determinations, particularly on basic dike rocks, had accumulated in the 1970s for **Simonen (1980)** to be able to conclude in his explanation to the Geological Map of Finland (scale 1:1,000,000) that the Jatulian sedimentation took place approximately 2200-2000 Ma ago. The overlying Kalevian rocks were assessed as being younger than c. 2000 Ma but older than the synorogenic Svecokarelidic plutonic rocks of c.1900 Ma age. Simonen (1980) still regarded the Kumpu formation as a molasse-like deposit formed after the main Svecokarelian folding, but contrary to the situation 20 years earlier (Simonen 1960a), it sufficed for him to call the underlying volcanic and sedimentary rocks Karelidic schists, without any attempts at a more detailed correlation or reference to the term 'Lapponian' (Fig. 3).

A Finnish-Soviet symposium on Jatulian geology, held in Finland in 1979 as a final meeting of a collaborative endeavour that had started in 1973, yielded several contributions from northern Finland in its proceedings and stimulated discussion on the stratigraphy of Karelian formations in general. **Silvennoinen et al. (1980)** published a generalized stratigraphical classification of northern Finland expressed in lithostratigraphical terms. They distinguished Archean Lower and Upper Lapponian Groups, with the boundary between the Archean and Proterozoic aeons being set at 2600 Ma. These units were followed by the Karelian Supergroup consisting of the Lower, Middle and Upper Jatulian Groups representing a peneplane stage of the Svecokarelian orogeny. The overlying Kalevian Groups of the last flysch and molasse stages belonged to the Svecofennian Supergroup. The Lower and Upper Lapponian Groups were thought to have been deposited between 3100 and 2600 Ma ago, while the time boundary between the Lower and Middle Jatulian Groups was considered to be the same as the age of the 2.45-Ga mafic layered intrusions (Fig. 4). The Lower and Upper Lapponian Groups encompassed all the rocks in Central Lapland which had previously been called Lapponian, including the Kittilä and Salla greenstone complexes, and also the rocks of the Tuntsa-Savukoski area. On the other hand, the Jatulian rocks were represented in Central Lapland only by the Kumputunturi quartzite conglomerate formation and its equivalents at Värttiövaara and Pyhätunturi. Applied to the Kuusamo area, the scheme of Silvennoinen et al.

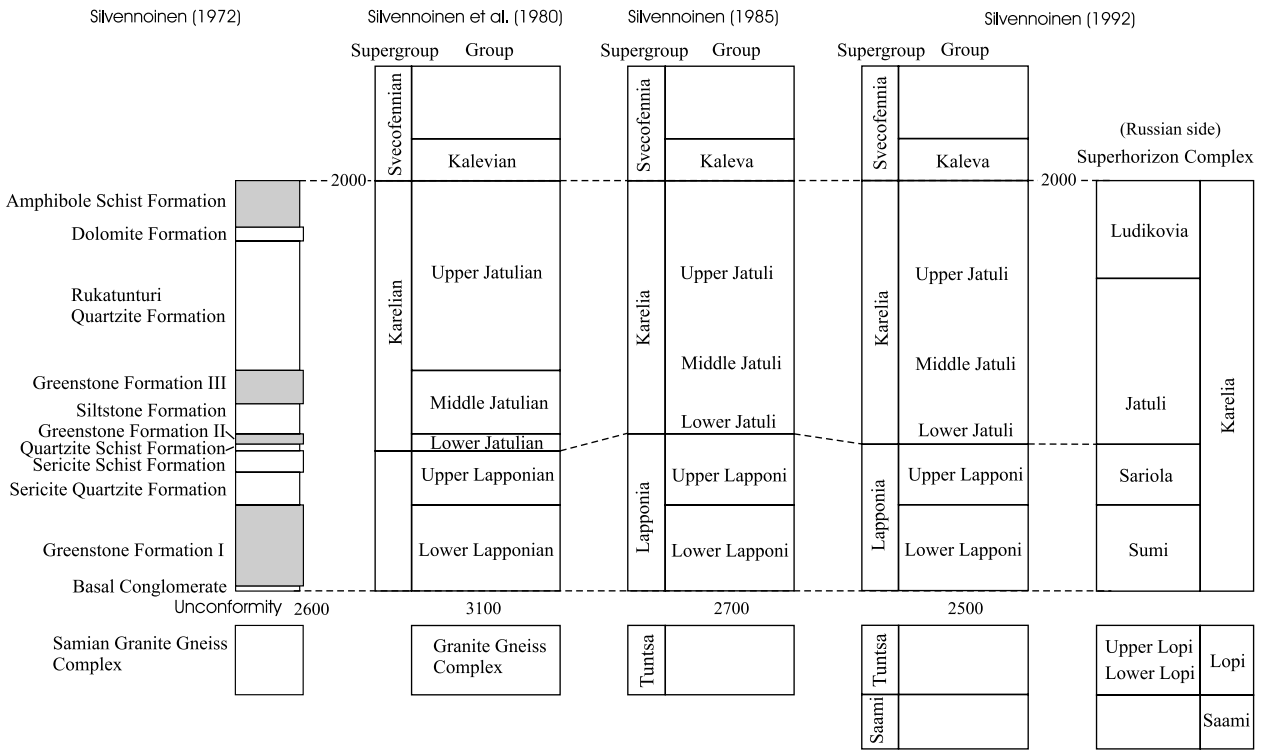

Fig. 4. Stratigraphical schemes presented for the Kuusamo schist belt since 1972 and correlation with rock formations on the Russian side of the border. Ages shown in million years.

(1980) resulted in the following classification: the basal conglomerate and the Greenstone Fm I belonged to the Lower Lapponian Group and the Sericite quartzite Fm and the Sericite schist Fm to the Upper Lapponian Group, while the other seven formations higher up in the sequence were distributed among the Lower, Middle and Upper Jatulian Groups (Fig. 4).

The stratigraphical classification of Silvennoinen et al. (1980) deviated in many respects from the views of other contemporary researchers (e.g., Simonen 1980, see Fig. 3). Firstly, the term Lapponian was extended from Central Lapland to the Kuusamo schist belt, where the rocks had previously been described with the terms Jatulian and Kalevian (Hackman & Wilkman 1929) or only Jatulian (Silvennoinen 1972). Secondly, the Lapponian metasediments and metavolcanics were excluded from the Karelian rocks (Karelian Supergroup), and thirdly, the Lapponian rocks were regarded as late Archean in age thus corresponding temporally to the Archean greenstone belts of eastern Finland, in analogy with the earlier suggestions made by Mikkola (1941) and Gaál et al. (1978). The last point gained support from geochronological data, as Silvennoinen et al. (1980) reported an age of 2790 Ma for an acid metavolcanic belonging to the Greenstone Fm I in the western part of the Kuusamo schist belt.

The stratigraphical column for the Kittilä area presented by **Rastas (1980)** at the above-mentioned symposium was in harmony with the general scheme of Silvennoinen et al. (1980). The Archean Lower Lapponian Group consisted of the amygdaloidal metavolcanics of Kaukonen, which were overlain by sericite quartzites and mica schists of the Archean Upper Lapponian Group. Higher up in the Upper Lapponian Group followed the metavolcanics and associated sedimentary rocks of the Kittilä greenstone complex. The oldest unit of the greenstone complex, according to Rastas (1980), was an acid volcanite at Jeesiö for which a zircon age of c. 2700 Ma had been obtained. In the Kittilä area, the Karelian Supergroup was represented by the lower Jatulian Luosujoki conglomerates and the middle Jatulian Kumpu formation. The Jatulian age for the Kumpu formation was corroborated by an age of c. 2200 Ma measured for an intersecting albite diabase at Värttiövaara (Kaarestunturi). Rastas (1980) also reported a U-Pb zircon age of c. 1885 Ma for the Latvajärvi volcanic rocks, an areally limited occurrence of acid and intermediate volcanic rocks discovered west of the village Kittilä at the beginning of the 1970s in connection with mapping by the GTK. The age of these rocks indicated that they were coeval with the nearby Kallo quartz monzonite massif belonging to the Haaparanta Suite. Rastas assigned the Latvajärvi volcanic rocks to the Kalevian Group, with a stratigraphical position above the Kumpu formation, while there still existed younger, molasse-stage conglomerates and siltstones of Vesikkovaara, resting on the Latvajärvi metavolcanics in his scheme. In contrast, based on the drillings and field investigations of a project on the iron ore mineralizations hosted by the Latvajärvi intermediate metavolcanics, **Puustinen et al. (1980)** maintained that the Kumpu-type quartzites were deposited after the eruption of the Latvajärvi volcanic rocks.

In a study aimed at explaining the genesis of the strata-bound skarn iron ores of Rautuvaara in western Lapland, **Hiltunen (1982)** extended his investigations to the stratigraphical and structural aspects of the whole Rautuvaara (Kolari) area. He divided it into eastern and western areas and constructed a separate lithostratigraphical column for each. The lowermost unit of the eastern section was the Niesakero-Kuertunturi quartzite complex. As the name implies, this is composed mainly of quartzites, but sillimanite-bearing and calc-silicate-bearing gneisses and amphibolites also occur. The Niesakero-Kuertunturi quartzite complex was overlain by the ore-bearing Rautuvaara Formation and higher up by the Kolari Greenstone Formation. The uppermost units were the Luosujoki conglomerate and the Yllästunturi quartzite.

In the western area, the Niesakero-Kuertunturi complex was correlated with the Mustijärvi quartzite, while the Kolari Greenstone Formation corresponded to the Siekkijoki Greenstone Formation. The Tapojärvi Quartzite Formation was in turn regarded as the counterpart to the highest Yllästunturi quartzite in the east.

Hiltunen (1982) did not extend his classification to the group or supergroup level but seemed to agree with Rastas' (1980) comparison of the sillimanite-bearing quartzites and mica schists of western Lapland with the sedimentary rocks of the Upper Lapponian Group in the Kittilä area. He also considered it possible to correlate his volcanic formations with the Upper Lapponian greenstone complex of Kittilä. An Archean age had been put forward for these Lapponian rocks by Rastas (1980) and Silvennoinen et al. (1980) two years earlier. Hiltunen (1982) did not express any decisive support for this view, but provided some apparently supportive evidence in terms of zircon dates. He had sampled two keratophyre layers in the upper part of the Siekkijoki Greenstone Formation, and although the analytical results were highly discordant, two zircon fractions suggested an Archean age for these rocks (the upper intercept of a two-point chord equivalent to 2738 Ma). (However, it is possible that the zircons are inherited in a similar manner as

reported by Perttunen and Vaasjoki, 2001, *this volume*).

The uppermost sedimentary rocks, the Luosujoki conglomerate, Yllästunturi quartzite and Tapojärvi Quartzite Formation, were readily correlated by Hiltunen with the Kumpu formation. If confirmed, this would be significant because, according to Hiltunen's map, a monzonite intrusion belonging to the 1.88 Ga Haaparanta Suite cuts into the Tapojärvi Quartzite Formation. This view is at variance with that of Mikkola (1941), who had earlier regarded the Tapojärvi quartzite as older than the Kumpu quartzites. It should also be mentioned in this context that when comparing the Rautuvaara rocks with those occurring on the Swedish side of the border, Hiltunen pointed out that even though sedimentary rocks younger than the Haaparanta Suite, belonging to Maattavaara Quarzite Group of Eriksson and Hallgren (1975), exist in the Vittangi area, no such rocks have been identified in the Rautuvaara area.

The base of the Niesakero-Kuertunturi quartzite complex was not known in the Rautuvaara area, but it was inferred from the gneissose plutonic rock fragments in the conglomerate of the lower part of the quartzite that the unit may have been deposited on a basement composed of granite gneisses. Such a basement was later recognized by **Väänänen (1989, 1992)** in the south, not far from the border of the area studied by Hiltunen (1982), and called the Venejärvi Gneiss Complex. Hiltunen (1982) similarly found no analogies to the lowermost Kaukonen amygdaloidal metavolcanics of the Kittilä area, assigned to the Lower Lapponian Group by Rastas (1980). The Venejärvi Gneiss Complex was instead shown to be overlain by the amygdaloidal, komatiitic Teuravuoma volcanite unit, which was considered to correspond to Lower Lapponian metavolcanics (Väänänen 1989, 1992).

Let us turn our attention back to the east. **Lauerma (1982)** was the first to criticize the inference of an Archean age for the lowermost supracrustal rocks of the Kuusamo schist belt, as proposed by Silvennoinen et al. (1980). He pointed out that ages between 2600 and 2800 Ma had been obtained for granitoids and pegmatites in the basement complex south of the Kuusamo schist belt and that no such granites had been seen to penetrate the schists. On the contrary, the schists overlie the basement complex unconformably, containing basal conglomerates with fragments from the basement in places. Lauerma (1982) also emphasized that the corresponding rocks on the Soviet side had been considered to be part of the Sumian-Sariolian group (Kulikov et al. 1980), implying that they cannot be much older than 2450 Ma.

It is necessary in this context to mention a Sm-Nd isotope study performed by **Krill et al. (1985)** on MgO-rich volcanic rocks from the Karasjok greenstone belt in northern Norway. Although the belt lies beyond the geographical borders of the area under discussion in this article, this isotopic study is of significance because the dated komatiites come from a belt regarded as a direct northerly continuation of the Kittilä greenstone complex, and also correspond structurally and geochemically to the komatiites occurring in Finnish Central Lapland. Krill et al. obtained a Sm-Nd age of 2085±85 Ma for the Karasjok komatiites, which argued strongly for an early Proterozoic age for the Karasjok greenstone belt, and for the Kittilä greenstone belt as well (cf. Silvennoinen et al. 1980).

In his stratigraphical classification of the Proterozoic rocks of northern Finland, **Silvennoinen (1985)** used essentially the same nomenclature as in Silvennoinen et al. (1980) with some purely cosmetic changes related to the spelling of the names; the Lapponian, Jatulian and Kalevian Groups being replaced by the Lapponia, Jatuli and Kaleva Groups and the Karelian and Svecofennian Supergroups by the Karelia and Svecofennia Supergroups, respectively (Fig. 4). More significant differences in the nomenclature included the introduction of the term 'Lapponia Supergroup', incorporating the Lower and Upper Lapponia Groups. Also, another previously unnamed supergroup was included in the stratigraphical column, the lowermost, Archean Tuntsa Supergroup. Silvennoinen (1985) still placed the Lower Lapponia Group entirely in the Archean (>2500 Ma) and the Upper Lapponia Group partly so, but commented that the chronostratigraphical correlation was not yet fixed for these units. Other changes included the lower time boundary of the Karelia Supergroup. Presumably influenced by the finding of a Jatulian basal conglomerate lying directly on c. 2.44 Ga mafic layered intrusions in the Peräpohja schist belt (Perttunen 1985), this boundary was regarded as younger than the age of the intrusions themselves, which were now taken as manifestations of Lapponian magmatic activity. Silvennoinen (1985) correlated the Lapponia Supergroup with the Sumi-Sariolian formations occurring on the Soviet side of the border. It is also worth of mentioning that Silvennoinen (1985, his Fig. 3) correlated the abundant mafic metavolcanics of the Salla area with the Greenstone Formation I of the Lower Lapponia Group. We will come back to this important issue below when discussing studies carried out by Manninen (1991) in the Salla area.

The stratigraphical scheme outlined by Silvennoinen (1985) was used as a base for classifying the rocks of northern Finland when the bedrock map of the

Nordkalott map series was completed a couple of years later as a joint effort between the Nordic geological surveys **(Geological Map, Northern Fennoscandia 1987)**. Both the Lapponia and Karelia Supergroups were defined in this map as being of early Proterozoic age **(Silvennoinen et al. 1986)**.

Following the terminology of Silvennoinen (1985), **Kortelainen et al. (1986)** considered the youngest quartzites and conglomerates in Lapland to belong to the Karelia Supergroup. The inner stratigraphy of these well-exposed deposits, which have been interpreted as fluviatile, was described in the west, at Sirkka, by **Kortelainen (1983)** and in the east, at Pyhätunturi, by **Räsänen and Mäkelä (1988)**. The latter distinguished four (Jatulian) formations in the Karelia Group in their study area. Subsequent studies have revealed that these do not form a continuous, northerly facing succession and only two of them, the Isokuru and Pyhätunturi Formations, belong to the sedimentary package regarded then as Jatulian (Räsänen & Huhma *2001, this volume*). This is worth mentioning here because the Pyhätunturi area has served as an important link when the rock sequences of the Kuusamo area have been traced towards northwest to Central Lapland.

One of the keenest advocates of an Archean age for the Lapland greenstone belts was **Saverikko (1987, 1988)**. While concurring with the earlier interpretation of Silvennoinen et al. (1980) that the greenstone belts running from Kolari via Kittilä and Sodankylä to Salla were Archean in age, he went still further in proposing that the whole Kuusamo schist belt belonged to the Archean, too. He placed the Kumpu formation in the Proterozoic, with its basal greywacke conglomerates correlated with the Sariolian and the overlying arkosic conglomerates and quartzites with the Jatulian, and subdivided the Lapponian rocks into three groups instead of the commonly used two-fold division (Fig. 3). The Lower Lapponian group was represented by volcanic rocks, e.g. those occurring in the Möykkelmä area of Central Lapland and in the Salla greenstone complex, and by quartzites and schists. The Middle Lapponian group comprised the Oraniemi association, an arkose-slate-quartzite sequence located between the Salla and Kittilä greenstone complexes, the transition to the Upper Lapponian group being marked by a graphitic slate unit overlain by pyroclastic komatiites in the Sattasvaara and Kummitsoiva areas. The Upper Lapponian group also included the large Kittilä greenstone complex.

A three-year joint Finnish-Soviet project (1980-1983) carried out by geologists from the North-Western Regional Geological Department (SEVZAPGEOLOGIA) and Rautaruukki Co. in order to evaluate the prospects for finding iron ores in Finland yielded various maps, including a regional stratigraphical sketch map, and several articles dealing with the stratigraphy of Lapland and other areas (Ivanov et al. 1986, Muradymov et al. 1986, 1988, Korsakova et al. 1988). **Ivanov et al. (1986)** incorporated the Lower Proterozoic formations in Finland into the Svecokarelian supergroup, consisting of the Sariolian, Jatulian and Kalevian-Svecofennian groups, and constructed stratigraphical columns with inferred mutual correlations between several areas in the Karelian and Lapland blocks. The basal conglomerate and the Greenstone Fm I of the Kuusamo schist belt were included in the Lower and Upper Sariolian subgroups, respectively, while all the overlying formations were assigned to the Lower and Upper Jatulian subgroups.

In contrast to the Kuusamo area, Ivanov et al. (1986) considered Central Lapland to be practically devoid of Sariolian rocks. Here the Lower Proterozoic formations begin with Lower Jatulian basal horizons, represented mainly by polymictic conglomerates and arkoses together with greenish fuchsite-bearing quartzites. It is worth mentioning at this point that Silvennoinen (1985) had emphasized the presence of green, chromium-bearing sericite in Lower Lapponian metasediments. Ivanov and his coworkers also included in the Lower Jatulian rocks the Kaarestunturi and Pyhätunturi conglomerates and arkoses which previously had usually been linked with the Kumpu formation (e.g., Silvennoinen 1985). In their opinion, the upper, volcanogenic section of the Lower Jatulian seemed to be missing in Central Lapland and the Lower Jatulian metasediments were directly overlain by Upper Jatulian volcanic and sedimentary rocks. They distinguished three Upper Jatulian type sections, essentially dominated by sedimentary (Pelkosenniemi structure), volcanogenic (Kittilä and Salla structures) and volcanogenic-sedimentary rocks (Kolari structure), respectively. The abundant volcanic rocks of the Kittilä area and the associated siliceous iron formations were thus regarded as Upper Jatulian (see also **Muradymov et al. 1986)**. As the uppermost formations, Ivanov et al. distinguished molasse-like Kumpu, Sirkka and Luosujoki conglomerates and quartzites, the overlying Latvajärvi metavolcanics and metasediments and finally the molasse-like Vesikkovaara conglomerates. They classified the Latvajärvi rocks as middle Kalevian, separating the Lower and Upper Kalevian metasedimentary units. The Kalevian sequence was obviously adopted from Rastas (1980) but elevated to a higher position in the stratigraphy.

It is clear from the preceding discussion that Ivanov et al. (1986) disagreed with those authors who had

argued for an Archean age for the greenstone belts in Kittilä and Salla (Gaál et al. 1978, Silvennoinen et al. 1980), but the project members did not totally exclude the presence of Archean supracrustal rocks in Central Lapland. In fact, **Muradymov et al. (1988)** claimed that Upper Archean formations are widespread in northern Finland. As indicated by the Schematic Tectono-Geologic Map of the Baltic (Fennoscandian) Shield, scale 1:2,500,000, accompanying **Korsakova et al. (1988),** these Archean supracrustal rocks were thought to occur partly in regions where komatiites are known to prevail. Muradymov et al. (1988) mentioned the Jauratsi and Sattasvaara structures, which they assigned to the Upper Lopian, though without providing any obvious evidence for their age.

Among the few age determinations that have so far been achieved directly from volcanic rocks in northern Finland, one from the Peurasuvanto area north of the Koitelainen layered intrusion deserves mentioning. Here **Peltonen et al. (1988)** described the volcaniclastic Rookkiaapa Fm, comprising conglomerates, tuffaceous arenites and interstratified pyroclastic layers which surround the Archean Tojottamaselkä granite gneiss dome (see also Pihlaja & Manninen 1988). Zircons from these basement gneisses were dated by Kröner et al. (1981) to c. 3.1 Ga, rendering these rocks some of the oldest in Finland. The overlying Rookkiaapa Formation contains various clasts, such as TTG gneisses, quartzites, acid to intermediate metavolcanics, diabases, and komatiites thought to have been derived from the basement gneisses or the Archean Tuntsa Supergroup, whereas the Lapponian aluminous schists and ultramafic metavolcanics and the c. 2.44 Ga Koitelainen intrusion are not represented among these clasts. According to Peltonen et al. (1988), an acid porphyry from an interbed of volcanic breccia yielded a U-Pb zircon age of 2526 Ma (published without errors), thus placing these rocks close to the Archean-Proterozoic boundary. As this rock represents the lowermost level of the Lapponian succession, the other Lapponian rocks can be assumed to be younger. This result is clearly in conflict with the postulated widespread occurrence of Archean greenstone belts in Central Lapland (e.g., Gaál et al. 1978), although given the heterogeneous nature of the host formation, there is a possibility that the sample may have contained a mixed zircon population (see Manninen et al. 2001, *this volume*). This means that the date represents a maximum age for the lowermost Lapponian rocks, which strengthens even further the conclusion of a Proterozoic age for the bulk of the Lapponian rocks.

The 5-year Lapland Volcanite Project (LVP) was launched at the Rovaniemi Office of the GTK in 1984 (Lehtonen 1989) with an aim of providing information on the generally poorly exposed volcanic rocks of northern Finland by means of effective sampling (including drilling and trenching) and geochemical methods. One of its goals was the chemical characterization of volcanic rocks at different stratigraphical levels. Before the actual commencement of the fieldwork for the project, a geological map of Central Lapland was compiled in collaboration with Lapin Malmi Co. on the basis of all the geological data available at that time **(Lehtonen et al. 1984)**. A partly overlapping stratigraphical map of western Lapland was also published the following year **(Lehtonen et al. 1985a)**. These maps and the explanation to the first of them **(Lehtonen et al. 1985b)** used the stratigraphical nomenclature of Silvennoinen (1985) but without any geochronological connotation.

The work of the LVP in 1986 included a field investigation into a 2.2 Ga diabase at Värttiövaara, an oval-shaped body 40x100 m in diameter at the contact between the quartzites and conglomerates that had traditionally been regarded as parts of the Kumpu formation. This diabase may be the one mentioned by Hackman (1927) as the only one that he knew to occur within the Kumpu quartzites. Later, the diabase has been used as evidence of a Jatulian age for the Kumpu formation (Rastas 1980). The excavations carried out by the LVP did not confirm an intersecting contact relation, as the contacts turned out to be strongly sheared. The age of c. 2.2 Ga obtained for the diabase has become an intriguing puzzle, as a younger age had been recorded for detrital zircons from the surrounding quartzite, and therefore a tectonic juxtaposition of the diabase and quartzite was judged to be a viable alternative **(Lehtonen 1987)**.

The small Möykkelmä granite gneiss dome within the Lapponian supracrustal rocks NNW of Sodankylä was selected as one of the first LVP target areas because **Tyrväinen (1983)** had just reported the existence of volcanic rocks directly on the granite gneiss basement in this area. **Räsänen et al. (1989)** documented a detailed study of the Möykkelmä metavolcanics which occur between the basement and the overlying Lapponian quartzites. A surface profile and a drill core section revealed the presence of two ultramafic and three mafic, subaerially erupted volcanic units with a total thickness of c. 250 m. Geochemically, the ultramafic rocks turned out to be komatiites and basaltic komatiites while the mafic rocks ranged from basalts to andesites. Räsänen et al. (1989) showed that all these volcanic rocks display a strong crustal signature in their chemical composition. In this respect, the Möykkelmä komatiites differed significantly from the megascopically similar,

volcaniclastic komatiites occurring higher up in the stratigraphy, e.g., in the Sattasvaara area. They also verified the existence of volcanic rocks with chemical characteristics of their own (enriched in LIL, depleted in HFSE), located below the Lapponian sedimentary rocks but younger in age than the Archean granite gneiss basement.

The 1990s

As part of the Lapland Volcanite Project, **Manninen (1991)** studied the stratigraphy and volcanite geochemistry of the Salla greenstone complex on the Finnish-Russian border. Various opinions had been offered earlier concerning the age and correlation of the Salla metavolcanics. The greenstones were candidates for counterparts of the Kittilä greenstones in the comparison made by Mikkola (1941), mainly due to the large volume of volcanic rocks in both areas. The same correlation was also favoured by Ivanov et al. (1986). Meanwhile, Gaál et al. (1978), Silvennoinen et al. (1980) and Saverikko (1987) had suggested an Archean age for the belt, whereas Lauerma (1982) regarded it definitely as Proterozoic. As noted before, Silvennoinen (1985) assigned it to his Lower Lapponia Group. In striking contrast, according to Kulikov et al. (1980), the corresponding metavolcanics on the eastern side of the border constitute the uppermost unit in the stratigraphy of the Paanajärvi-Kuolajärvi belt. These authors placed them in the Suisaarian group, which implies that they would be post-Jatulian in age.

Manninen (1991) divided the depositional sequence in the Salla area into seven sedimentary and volcanic formations and organized them into three groups, the Lower, Middle and Upper Lapponi Groups (Fig. 3). The majority of the supracrustal rocks in the Salla complex belong to the two lowermost volcanic Salla and Mäntyvaara Formations, forming the Lower Lapponi Group. The metavolcanics of the former formation range geochemically from basaltic andesites to rhyolites and those of the latter from komatiitic basalts to high-Mg basalts. They have a significant crustal signature in their composition, in the same manner as recorded by Räsänen et al. (1989) for the Möykkelmä metavolcanics. According to Manninen, the volcanic structures and geochemical composition of the Salla and Mäntyvaara Formations find their closest counterparts on the other side of the border in the Sumian and Sariolian metavolcanics, respectively. The Mäntyvaara Fm was also correlated with the Greenstone Fm I in the Kuusamo area, while the Salla Fm was considered to be older. Not far from Salla, the 2.44 Ga Akanvaara layered intrusion is known to intersect felsic metavolcanics that are comparable to those of the Salla Formation. Manninen (1991) concluded that this intrusion is roughly coeval with the Salla Formation.

The middle Lapponi quartzite-dominated rocks occur west of the metavolcanics described above and are separated from them by a tectonic contact. In the middle of these metasediments (Kelloselkä Fm, Matovaara Fm), there is a mafic volcanic formation (Tahkoselkä Fm) which can be traced farther in the south as merging with the Jatulian Greenstone Fm III of Silvennoinen et al. (1980). The supracrustal rocks classified as Upper Lapponian are encountered in the northern part of the Salla greenstone complex and are composed of ultramafic and mafic pyroclastic rocks similar to those found in the Kummitsoiva and Sattasvaara areas. They overlie the middle Lapponian dolomites, graphite-bearing mica schists and jaspilites of the Aatsinginhauta Formation.

The area investigated by Manninen (1991) is a key region because here the Central Lapland and Kuusamo schist belts join together. Several important aspects were reported by him. Firstly, the term Lapponi was not used in the sense defined e.g. by Silvennoinen (1985), that is as being pre-Jatulian (beneath the Karelian Supergroup), but instead the sequence of rocks forming the three Lapponi Groups was regarded as correlative with other Karelian units ranging from the Sumian and Sariolian through Jatulian to even Ludikovian. Secondly, most of the volcanic rocks of the Salla greenstone complex differed in their geochemical compositions drastically from the metavolcanics from the Kittilä greenstone complex, so that the correlation between these two complexes could not be substantiated. Finally, as the bulk of the Salla metavolcanics were correlated with the Sumian and Sariolian metavolcanics, the Suisaarian connection as advocated by Kulikov et al. (1980) was negated. Later, a dike cutting volcanic rocks in Salla has been dated, demonstrating a pre-Suisaarian age for these rocks (Manninen & Huhma 2001, *this volume*).

Silvennoinen (1972) had already recorded the presence of acid porphyry clasts in the basal conglomerate of the Kuusamo schist belt. In his explanation to the Kuusamo and Rukatunturi map sheets, **Silvennoinen (1991)** published U-Pb isotope results for zircons from three separate clasts, which taken together yielded an age of 2405±6 Ma. This can be regarded as

a maximum age for the lowermost Lapponian supracrustal rocks in the Kuusamo area, thus finally nullifying the proposed Archean age for the Lapponian rocks of the Kuusamo area (cf. Saverikko 1987).

The next account of the stratigraphy of the Kuusamo area was given by **Silvennoinen (1992)** in the proceedings of the Finnish-Soviet symposium on the Paanajärvi-Kuusamo-Kuolajärvi area. An attentative reader may note that in his description of the supergroups, Silvennoinen (1992) had shifted the lower boundary of the Karelia Supergroup from its original place at the base of the Quartzite schist Fm (Silvennoinen et al. 1980) to the base of the overlying Greenstone Fm II. In the correlation across the border, presented in the legend to the appendix map **(Silvennoinen et al. 1992)**, the Lapponia Supergroup was interpreted to be equivalent of the Sumi-Sariolian Superhorizon, while the Karelia Supergroup corresponded to the Jatuli and Ludikovia Superhorizons (Fig. 4).

Lehtonen et al. (1992) divided the Paleoproterozoic supracrustal rocks of Central Lapland into five lithostratigraphical groups: the Lower, Middle and Upper Lapponi Groups, overlain by the Lainio and Kumpu Groups. The type formations of the Lower Lapponi Group are the Madetkoski and Möykkelmä Formations, comprising crustally contaminated amygdaloidal and pyroclastic metavolcanics ranging from komatiites to dacites. The mainly sedimentogenic Middle Lapponi Group, with the Virttiövaara Fm as its type formation, includes traditional Lapponian quartzites, arkosites, mica schists etc., while the Upper Lapponi Group is composed mostly of volcanogenic rocks. The lowermost part contains pillowed and pyroclastic komatiitic and picritic volcanic rocks, represented by the Sattasvaara Formation, for example. In the Kittilä-Sodankylä area, the Sattasvaara Formation is overlain by a thick sequence of submarine basic volcanic rocks subdivided into the lower Fe-tholeiitic Kautoselkä Formation and the upper Mg-tholeiitic Vesmajärvi Formation, which are separated by the BIF-bearing Porkonen Formation.

The coarse-clastic sedimentary rocks deposited unconformably on the Lapponi Groups were split by Lehtonen et al. (1992) into two separate units, the Lainio and Kumpu Groups, on the basis of differences in their deformation histories. The Lainio Group comprises the sedimentogenic Ylläs and Vesikkovaara Formations and the intervening metavolcanics of the 1.9 Ga Latvajärvi Formation, while the overlying Kumpu Group is exemplified by the Levi Formation. The rocks representing the Kumpu Group had previously been included in the Middle Jatulian Group (Rastas 1980), but this assignment was now regarded as unwarranted because, as interpreted by Lehtonen et al. (1992), the Kumpu rocks are younger than the Latvajärvi metavolcanics. Lehtonen et al. (1992) provided further evidence for post-Jatulian deposition by reporting ages of 1913-2066 Ma for detrital zircons from the Kumpu quartzites. Rastas (1980) had based his Jatulian connection partly on an age of 2044 Ma obtained for a diabase which was assumed to have intersected the Kumpu quartzite at Sätkänävaara. Excavations carried out by the LVP had nevertheless revealed that cobbles from the diabase in question exist within the adjacent quartzite (Lehtonen et al. 1992).

In his investigation on the carbon isotope composition of sedimentary carbonate rocks from the Fennoscandian Shield, **Karhu (1993)** recognized a fairly systematic $^{13}C/^{12}C$ variation within individual stratigraphical sequences of Karelian formations and showed that this method can potentially be used as a new tool for stratigraphical correlation. When plotting the carbon isotope composition against time for the sedimentary rocks whose age could be constrained by radiometric dating of associated intrusive rocks, a distinct positive isotope anomaly in the ^{13}C range of +10±3 per mil could be observed in the time interval 2.2 to 2.1 Ga, that is when Jatulian carbonate rocks were precipitated. From the standpoint of the present review, it is interesting to note that Karhu also reported generally anomalous carbon isotope compositions for sedimentary carbonates within the Central Lapland quartzites that Lehtonen et al. (1992) had assigned to the Middle Lapponi Group. According to him, this result supports the correlation of the Middle Lapponi Group with the Jatulian formations in Karelia (cf. Manninen 1991). On the other hand, carbonates associated with the voluminous volcanic rocks of the Kittilä greenstone complex (north of the 'Sirkka tectonic line') displayed normal marine isotopic values, suggesting that they lie at a higher or alternatively lower stratigraphical level compared with the Jatulian carbonate rocks. Karhu favoured the former option. Among other interesting outcomes of Karhu's study was the finding that the excursion of the carbon isotope composition towards highly positive anomalies had already reached $\delta^{13}C$ values of up to +8‰ in carbonates from the Sericite Schist Formation in the Kuusamo area, representing part of the Lapponia Supergroup of Silvennoinen (1985, 1992). Even higher ratios were recorded for carbonates from the overlying Karelia Supergroup with the youngest carbonates belonging to the Upper Jatuli Group showing a tendency to decrease towards normal values.

A recent and fundamentally different approach to the stratigraphical classification of supracrustal rocks

in northern Finland was made in the final works produced within the Lapland Volcanite Project, including a new stratigraphical map of Central Lapland by **Räsänen et al. (1995)**, scale 1:200,000, and a report by **Lehtonen et al. (1998)** on the stratigraphy, petrology and geochemistry of the Kittilä greenstone area in Central Finnish Lapland. They renamed the rock units at the group and formation levels using geographical names together with lithostratigraphical rank terms and hence totally refrained from using names like Lapponian or Jatulian that had been common earlier. Although the term Lapponian and its derivatives have a long tradition behind them and are deeply ingrained in the Finnish geological literature, this new approach was chosen because it conforms better with the international recommendations for lithostratigraphical classification (Hedberg 1976, Salvador 1994) and avoids the ambivalent meanings coined by various authors with time for many loosely defined traditional terms.

For the Paleoproterozoic sedimentary and volcanic rocks, Räsänen et al. (1995) and Lehtonen et al. (1998) distinguished the following lithostratigraphical groups, from oldest to youngest: the Salla, Onkamo, Sodankylä, Savukoski, Kittilä, Lainio and Kumpu Groups. The two last mentioned ones, the Kumpu and Lainio Groups, represent the previous Kumpu formations or the Karelia Supergroup, while the other five groups are the heirs of the former Lapponian rocks or the Lapponi Supergroup. In addition to this correlation, Figure 5 shows the type formations for the groups and their approximate lithological compositions as taken from the legend to the Appendix map of Lehtonen et al. (1998). Field geological, lithological and geochemical characteristics of these units and their representative type formations were documented by Lehtonen et al. (1998), and the readers are referred to this work for more detailed information. It only suffices here to briefly summarize the most salient features of the five lowermost groups. The Salla and Onkamo Groups are composed mostly of volcanogenic rocks showing commonly structures indicative of subaerial eruptions. They vary in composition from intermediate to felsic in the Salla Group and from ultramafic to mafic in the Onkamo Group and can be correlated with the pre-Jatulian Sumi-Sariolian rocks in the eastern part of the Fennoscandian Shield. These units are overlain by the Sodankylä Group comprising the wide-spread "Lapponian" quartzites and spatially associated minor dolomites and mafic metavolcanics. These are the best candidates for counterparts of the traditional Jatulian quartzites in the Karelian schist belts. The

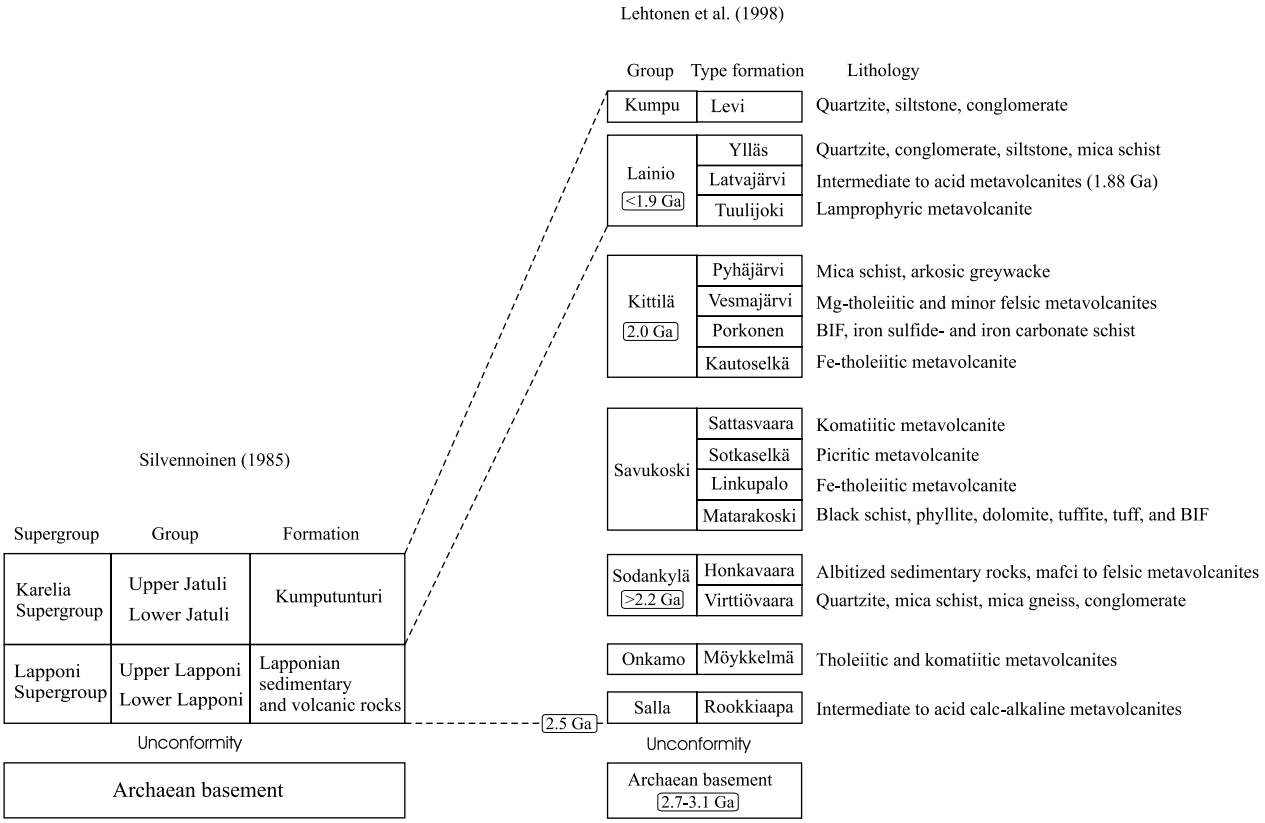

Fig. 5. Lithostratigraphical scheme of Lehtonen et al. (1998) for rocks in Central Lapland and its correlation with the Lapponian and Karelian Supergroups of Silvennoinen (1985, 1998).

next stratigraphical unit, lying conformably on the Sodankylä Group, is the Savukoski Group, which contains pelitic metasediments and minor mafic tuffites in its lower part and primitive volcanic rocks (komatiites and picrites) in its upper part. The volcanic rocks and associated chemical metasediments of the Kittilä greenstone area are assigned to the Kittilä Group. Hanski (1997) presented evidence for the presence of ophiolitic ultramafic rocks around Nuttio at the eastern margin of the Kittilä greenstone area. This, coupled with structural, geochemical and isotopic data on mafic and felsic rocks of the Kittilä Group, suggests that this rock complex forms an allochthonous unit representing a block of ancient oceanic crust (Hanski et al. 1998), and therefore its direct correlation with the Karelian cratonic volcanogenic-sedimentary sequences is not warranted.

Additional reasons, and perhaps the most compelling ones, for rejecting the traditional Lapponian-Jatulian nomenclature in Lapland were provided by U-Pb and Sm-Nd age determinations. In many stratigraphical schemes ever since the first definition by Sederholm (1932) to the system by Silvennoinen (1992, 1998), the Lapponian rocks have been regarded as *pre-Jatulian*, which was also implicit in the correlation of the Lapponian with the Sumi-Sariolian (Manninen 1991, Silvennoinen 1992). This implies that the radiometric age of the Lapponian rocks should be greater than that of the Jatulian rocks. Deposition of the Jatulian supracrustal rocks commenced well before 2.2 Ga, which is the age of the gabbro-wehrlite sills intruding into Jatulian quartzites (Hanski 1987). Against this background, it is significant that Hanski et al. (1997) reported a Sm-Nd isochron age of 1990±35 Ma for the mafic metavolcanics of the Vesmajärvi Formation belonging to the Kittilä Group. This age is compatible with the U-Pb zircon ages between 2012±5 and 2018±7 Ma obtained for felsic porphyries within the Kittilä Group. The field characteristics of these porphyries provide evidence for their coeval emplacement with the surrounding mafic metavolcanics (Hanski et al. 1998, Rastas et al. 2001, *this volume*). Although mafic volcanic rocks with an apparent pre-Jatulian age, such as the tholeiitic and komatiitic metavolcanics of the Onkamo Group lying directly on the Archean basement at Möykkelmä (Räsänen et al. 1989), exist in Central Lapland, the age determinations referred to above indicate that the majority of the mafic volcanic rocks of the Kittilä greenstone area, belonging to the Kittilä Group, are *post-Jatulian* in age, and are thus in striking contrast with the first definition and later usage of the term Lapponian.

The Jatulian-Kumpu correlation has become an impossible alternative as a result of U-Pb dating of felsic porphyry and granite fragments and detrital zircons from the Lainio and Kumpu Group metasediments. Hanski et al. (1997) obtained an age of 1927±6 Ma for a felsic porphyry cobble from the Mantovaara conglomerate located 20 km NE of Sirkka, demonstrating that these sediments were deposited later than this date. An even younger age of deposition is indicated by granitoid fragments from the Kellostapuli and Vesikkovaara conglomerates (Rastas et al. 2001, *this volume*) and detrital zircons from the Aakenustunturi and Värttiövaara quartzites (Hanski et al. 2000), as these contain rock fragments and zircons most likely derived from the 1.88 Ga Haaparanta Suite plutonic rocks and Latvajärvi Formation metavolcanics. These results are in harmony with the suggestion put forward already by Mikkola (1941) that there exist aplitic 'syenite series' clasts in the Sirkka conglomerates.

HISTORY OF STRATIGRAPHICAL STUDIES IN THE PERÄPOHJA SCHIST BELT

Introduction

With the exception of interpretations based on the first investigations carried out at the beginning of this century, the history of stratigraphical research in the Peräpohja schist belt does not involve such contradictory ideas on the general stratigraphical order of supracrustal rocks and their ages and regional correlation as are revealed by the foregoing pages on research into the rocks of Central Lapland. Instead, a consensus on the character of the main lithological units and their mutual relationships has prevailed for decades, with the rock associations being described as a typical Karelian succession composed of Jatulian and Kalevian rocks. The longest history of research, stretching back for about a century, concerns the SW part of the schist belt, where the stratigraphical outlines were also formulated. Later studies have shown that the geological structure of the belt becomes more complicated towards the north and east, where there are areas with higher degrees of deformation and metamorphism. These observations and the changes

in the general stratigraphical classification system in northern Finland, have necessitated recent modifications to the lithostratigraphical nomenclature (Perttunen et al. 1995).

From the 1910s to the 1950s

The bedrock mapping to a scale of 1:400,000 started at the end of the 19th century formed the basis for the first effort at evaluating the stratigraphy of the Peräpohja schist belt. In his explanation to the Rovaniemi, Tornio and Ylitornio map sheets, **Hackman (1914, 1918)** identified the gneissose granites and diorites in the southern part of the mapped area as the oldest, pre-Kalevian rocks, subdividing the overlying supracrustal rocks into older Kalevian and younger Jatulian rocks in accordance with the general opinion on the relation between the Kalevian and Jatulian formations in Finland at that time. The Kalevian rocks contained phyllites, mica schists and gneisses, quartzites, dolomites, amphibole schists and metabasites, while the Jatulian rocks were mainly composed of quartzites, metabasites and conglomerates, including the conglomerates of Pyhätunturi in the Pelkosenniemi area.

The large 'post-Kalevian' granite massif occurring on the northern side of the Peräpohja schist belt was described by Hackman (1914) as being fringed in many places by mica gneisses and veined gneisses that have clear indications of being affected by the granites. These gneisses were equated with the better-preserved phyllites occurring within the schist belt. Also, some schistose and glassy quartzites were observed to be intersected by granites and were therefore included in the Kalevian rocks. On the other hand, the quartzites with a more clastic appearance were not observed to be intersected by granites and were therefore regarded as younger and included in the Jatulian. Further evidence for a difference in metamorphic grade between the Kalevian and Jatulian rocks was provided by penetrating metabasites, those associated with Jatulian quartzites being argued to be less metamorphosed than those cutting only Kalevian metasediments. Hackman (1914) also mentioned conglomerates which could be regarded as Jatulian basal conglomerates and contained fragments of a probable Kalevian age. These stratigraphical notions arrived at by Hackman (1914) regarding the Peräpohja schist area were clearly influential in his subsequent treatise on the Kittilä-Sodankylä area as discussed earlier in this paper (Hackman 1927).

The distinction between the Kalevian and Jatulian rocks was thus partly based on metamorphic grade and, as admitted by Hackman (1914), was not always a straightforward matter, especially when dealing with quartzites. The other criteria included the relationship to the granites as it was asserted that the Jatulian rocks are nowhere observed to be cut by granites - which was obvious since this premise was included in the definition of Jatulian rocks. The tinge of circularity involved in this common definition was pointed out later by Sederholm (1932), for instance.

The commonly held view of older Kalevian and younger Jatulian rocks was challenged for the first time in the Peräpohja area by **Mäkinen (1916),** but his suggestions remained unacknowledged for many years. He studied only the extreme SW part of the Peräpohja schist belt, but was able to decipher a synclinal structure, revealing a consistent rock succession. The following quotation from Mäkinen (1916) illuminates his observations: "*Next to the basement there is a zone of quartzite, from 4 to 5 km broad, with a dip of 60°-70° inwards. The inner parts are occupied by phyllites and mica schists with intercalated layers of dolomite, all of which are conformable to the quartzite in their strike and having fairly steep, almost vertical dips. The quartzite forms the basal layers of the sedimentary series and contains in its lowest parts detritus from the subjacent basement*". Thus Mäkinen could not find any support for either the previously presented order of deposition or an unconformity between the phyllitic Kalevian and the quartzitic Jatulian rocks. He was inclined to regard the quartzites as basal layers of the same, Kalevian sedimentary series.

Hausen (1936), in his attempt to clarify the stratigraphical and tectonic evolution of the sedimentary series in the Peräpohja schist belt, established the following sequence of deposition of supracrustal rocks. A basal conglomerate (?) was deposited on the Archean basement of granites, granodiorites, and migmatites etc. and overlain by feldspar quartzites and sericite quartzites. These were followed by pure quartzites, with probably contemporaneous amygdaloidal lava rocks. The next large unit comprised argillites with intercalations containing a few quartzite conglomerates (Taivalkoski) and dolomites as well as "green schists" of tuffaceous origin. The last unit to precipitate contained dolomites with interlayers of marls and also psammites. It was possible for these dolomites to be still overlain by some argillites higher up in the sequence.

Much attention was paid later to the Taivalkoski

conglomerate (Mäkinen 1949, Härme 1949, Väyrynen 1954), which was described for the first time by Hausen (1936). This occurs within a phyllitic environment and contains mainly rounded pebbles of quartzite accompanied by lesser amounts of angular phyllite fragments. Hausen (1936) classified the conglomerate as intraformational and did not assign it any great tectonic significance in terms of stratigraphical breaks.

Besides working out the stratigraphical sequence, Hausen (1936) also provided it with a palaeosedimentological interpretation, based on beautifully preserved primary sedimentary structures. According to him, the whole series bears a transgressive character, the facies changing with stratigraphical height from epicontinental (deltaic-fluviatile) to littoral and shelf types. Hausen did not find any evidence for a stratigraphical break within the sedimentary series and therefore disagreed with the prior two-fold division into "Kalevian" and "Jatulian", which was based on the degree of metamorphism and the relation of the metasediments to younger granites. Instead, he considered the supracrustal rocks to be members of one and the same series, the North-Bothnian, which could have been formed during the "Karelidic" cycle of sedimentation of Sederholm (1932).

The first results of post-war investigations in the Peräpohja schist belt appeared in two academic dissertations based on fieldwork in partly overlapping areas in the western and southwestern parts of the belt (A. Mikkola 1949, Härme 1949). **Mikkola (1949)** shared the opinion of Mäkinen (1916) that quartzites form the lowermost known part of the sedimentary series, calling the well-preserved quartzites in the southern part of the belt the *Kivalo quartzites* and the more strongly metamorphosed, often fuchsite-bearing, anticlinal quartzites in the central part the *northern quartzites*. He regarded these quartzites as belonging to the same stratigraphical level, differing from each other only in their degree of metamorphism. Higher up in the sequence, he distinguished a separate black quartzite interstratified with slates and dolomites. In general, the slates were documented as lying directly on the quartzites without any unconformity.

Mikkola (1949) paid special attention to the distinguishing of volcanic (tuffitic greenstone, agglomerates, amygdaloidal lavas, pillow lavas) and hypabyssal rock types among those covered by the widely used collective term 'metabasite'. He also studied the contact relations of the Haaparanta Suite igneous rocks, varying from gabbros to granodiorites, and showed that these are definitely intrusive. Earlier they had been regarded as being older than the schist belt (Hackman 1914, Mäkinen 1916).

With the comprehensive description of the rocks of Central Lapland by E. Mikkola (1941) already published, A. Mikkola (1949) was able to make a good comparison between the supracrustal sequences of the Peräpohja and Central Lapland areas. According to him, the Lapponian Series contained the counterparts of all the rocks to be found in the Peräpohja schist belt. The greatest difference concerned the amount of carbonate rocks, which are not as abundant in Lapland as in Peräpohja. Correlation with the Kumpu-Oraniemi Series was not so straightforward. Mikkola (1949) pointed out the absence throughout the Peräpohja area of the polymictic conglomerates to be found in the basal formations of the Kumpu-Oraniemi Series. On the other hand, thick slate and mica schist piles typical of the Peräpohja area were not known in the Kumpu-Oraniemi Series.

Although the stratigraphical scheme of the Kuusamo-Salla area proposed by Hackman and Wilkman (1929) had become obsolete, a correlation between the Peräpohja schist belt and that area seemed obvious to Mikkola (1949), at least insofar as the occurrence of similar rock types in both areas was concerned. A more direct parallel could be made with the Kainuu schist belt, where counterparts for the Kivalo quartzites were found among the Kainuan-type quartzites of Väyrynen (1933), though Mikkola erroneously mentioned the absence of basal formations in the Peräpohja area as one difference (cf. Perttunen 1980). Furthermore, the transition from quartzites to varved slates and dolomites corresponded to the marine Jatulian in Väyrynen's nomenclature. Finally, a correlation could be drawn between the Kalevian schists and the fine-grained metasediments in the Peräpohja schist belt. A complicating factor, however, concerned the quartzites and conglomerates of the Jaurakka facies, marking an apparent unconformity at the base of the Kalevian formations. This did not conform with Mikkola's perception of the absence of such unconformities in the slates of the Kemi river basin.

Härme (1949) and Mikkola (1949) agreed on many aspects of the geology of the Peräpohja schist belt, such as the general order of deposition of the supracrustal rocks, the lack of major unconformities within the Karelian sequence and the intrusive nature of the Haaparanta Suite. Härme (1949) described mafic igneous bodies located between the Archean basement and the schist belt which were later to be called mafic layered intrusions, giving them the name 'anorthosite-serpentine series', but he could not ascertain whether they were older or younger than the schist belt.

The new discoveries made by Härme (1949) also included two occurrences of polymictic conglomerates at the southern margin of the belt which turned out

to be rather enigmatic. These conglomerates were found within the space of a few tens of metres on the Kivalo ridge, with rocks of the anorthite-serpentine series on one side and greenstones interpreted as hypabyssal diabases on the other. According to Härme (1949), the fragments were mostly composed of migmatitic granite and ossipite (uralite gabbro) supplemented with greenstone, anorthosite, quartzite and slate (see also Enkovaara et al. 1953). These fragments were recorded as being of exactly the same type as found in outcrops around the conglomerates. These features coupled with the position of the conglomerates between two mafic bodies of alleged intrusive origin and a low degree of deformation of the conglomerates were taken by Härme as evidence of the conglomerates being younger than the sedimentary cycle of the schist belt. Furthermore, he correlated the migmatitic granites in the SE side of the schist belt with granite veins cutting through rocks of the anorthosite-serpentine and Haaparanta Suite. This led him to infer that the migmatitic granite fragments in the conglomerates were also much younger than the Peräpohja sedimentary sequence. The conglomerates were thus assumed to have been deposited at a very late stage, when these migmatitic granites were subjected to erosion. The erroneous conclusion that Härme had reached about the nature of the granite fragments in the Kivalo conglomerates (cf. Perttunen 1980) should be seen against the background of the general belief held at the time that the migmatite-forming microcline granites in the basement area belonged to late-kinematic granites of the Karelidic orogenic cycle.

Since the work of Hackman (1914), **Väyrynen (1954)** was the first to propose a discordance between the Kalevian phyllitic and Jatulian quartzitic rocks in the Peräpohja schist belt, but this time the order of these rocks was reversed. Väyrynen suggested that certain quartzites, like that occurring at Kallinkangas, have features resembling those of the Jaurakka facies rocks in the Kainuu schist belt and could be basal parts of the phyllite formation. He also referred to the Ukonköngäs conglomerate, described by Mikkola (1949), which is situated in the SE corner of the Paakkola phyllite area and contains fragments of quartzite, dolomite and granite. The occurrence of phyllites in axial depressions and their close relation to tuffitic volcanic rocks which were thought to have erupted during the folding of the Jatulian rocks, and also the presence of the above conglomerates, led Väyrynen to imply some kind of discordance between the two major units in the Peräpohja schist belt. However, later workers have never been able to provide any unequivocal support for this suggestion.

The 1980s to 1990s

The Geological Survey of Finland started mapping the Peräpohja schist belt to a scale of 1:100,000 in the early 1960s under the leadership of V. Perttunen, and the first summary of these investigations was presented by **Perttunen (1980)** in connection with the Finnish-Soviet Jatulian symposium. He described a basal conglomerate which grades downwards to a disintegrated granite of the Archean basement complex. Further field observations have verified that this is a true basal conglomerate and not younger than the schist belt as argued previously by Härme (1949). The granite boulders in the conglomerate were reported by Perttunen (1980) to have an age of 2600 Ma, which was used as evidence for the Jatulian nature of the conglomerate as opposed to Lapponian (cf. Silvennoinen et al. 1980). Another discovery related to the lower part of the sequence was the recognition of a mafic volcanic formation above the basal conglomerate unit. This amygdaloidal metavolcanite, the Runkaus Volcanite Fm, is spatially closely associated with a sill-like differentiated diabase and was not distinguished from the latter by earlier investigators.

In his lithostratigraphical classification (see Fig. 6), Perttunen (1980) designated the basal conglomerate together with the overlying Runkaus Volcanic Formation as forming the Lower Jatulian Group. Higher up in the sequence lay the Kivalo Quartzite Formation and the Jouttiaapa Volcanic Formation of the Middle Jatulian Group. The Upper Jatulian Group was represented by the Kvartsimaa Quartzite, Tikanmaa Volcanic and Rantamaa Dolomite Formations. The uppermost formation of the Peräpohja schist belt was the Martimo Phyllite Formation assigned to the Kalevian Group. Following the general scheme of Silvennoinen et al. (1980), the Jatulian groups were united to the Karelian Supergroup while the Kalevian Martimo Phyllite Formation was taken being as part of the Svecofennian Supergroup.

The detailed stratigraphy of the upper part of the sequence had previously been somewhat obscure (Hausen 1936, Väyrynen 1954), but now Perttunen (1980) had definitely classified the phyllitic rocks of the Martimo formation as the youngest supracrustal rocks. The tuffaceous greenschists that earlier had often been associated with the slates (phyllites) constituted in Perttunen's scheme an independent forma-

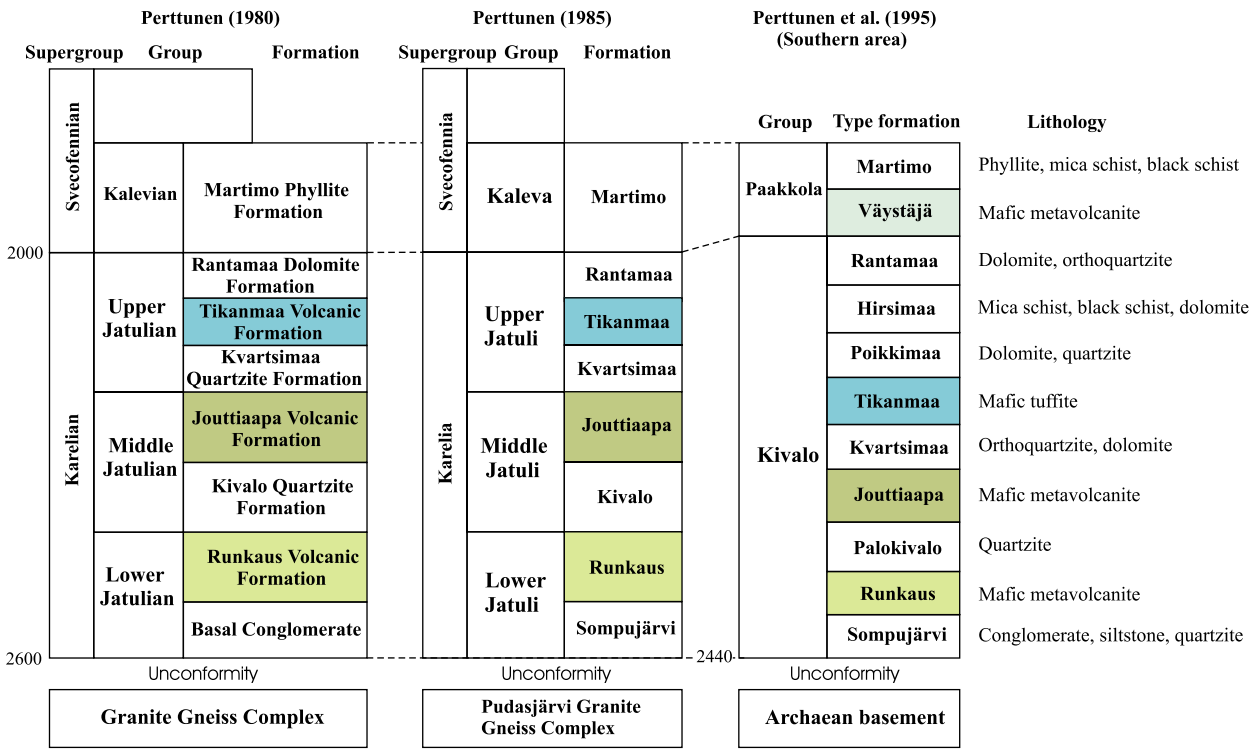

Fig. 6. Stratigraphical schemes presented for the Peräpohja schist belt since 1980. Ages shown in million years.

tion separated from the phyllites of the Martimo Formation by the Rantamaa Dolomite Formation.

According to Perttunen (1980), the quartzites of the Middle Jatulian Group were penetrated by differentiated albite diabase sills of an age c. 2200 Ma. This age together with that of 2600 Ma obtained for the granite boulders within the basal conglomerate thus constrained the time of deposition of the lower part of the Jatulian succession. On the other hand, the minimum age for the youngest supracrustal rocks of the schist belt was provided by the age of c. 1900 Ma determined for the intrusive rocks of the Haaparanta Suite.

The next account of **Perttunen (1985)** retained the same stratigraphical sequence but included superficial modifications in the nomenclature at supergroup to formation levels (see Fig. 6). The lowermost basal conglomerate and associated arkosic metasediments were now termed the Sompujärvi Formation. In the case of the other formation names, the attributes describing the rock types were left out, so that the former Runkaus Volcanic Formation was now called the Runkaus Formation, for example. One additional formal difference was the labelling of the Archean basement complex as the Pudasjärvi Granite Gneiss Complex.

Perttunen (1985) reported an important observation according to which the conglomerates of the Sompujärvi Formation were deposited unconformably on the 2.44 Ga mafic layered intrusions, even though no fragments from the latter were found within the conglomerates of the mapped area. Later, he provided evidence for this erosional contact in terms of outcrop and drill core observations **(Perttunen 1991),** while **Alapieti et al. (1989)** described the finding of a PGM-bearing phenoclast within a conglomerate, apparently derived from an underlying layered intrusion. These observations placed tighter time constraints on the beginning of Jatulian sedimentation in the area.

Huhma et al. (1990) reported an Sm-Nd isochron age of 2090±70 Ma for the metavolcanics of the Jouttiaapa Formation and applied the same method to the Runkaus Formation to obtain a somewhat imprecise age of 2330±80 Ma. Nonetheless, this was in good agreement with the dates obtained previously, 2.44 Ga for the underlying mafic layered intrusions and 2.2 Ga for the differentiated sills penetrating into the quartzites. Based on field geological evidence, it was known that the age of the Runkaus Formation should fall between these two figures.

In a study related to the Lapland Volcanite Project, **Perttunen (1989)** characterized geochemically the basic volcanic and sill-like units in the Peräpohja schist belt using the stratigraphical scheme of Perttunen (1985). The Runkaus Formation displayed an obvious influence of contamination with crustal material, whereas the Jouttiaapa and Tikanmaa Formations did not show any recognizable signs of sialic contamination. The Jouttiaapa Formation in particular has a very

depleted, MORB-like composition which renders it quite unique among volcanic rocks that have erupted in an environment of epiclastic sedimentary rocks (see also Huhma et al. 1990). In fact, it is difficult to find any geochemically analogous volcanic units in a similar setting elsewhere in the Fennoscandian Shield. The geochemical comparison further demonstrated that the 2.2 Ga differentiated sills did not have any genetically related counterparts among the overlying volcanic rocks, which was in contrast to some earlier suggestions (e.g., Härme 1949).

On the basis of long-term, detailed field studies carried out in the Peräpohja schist belt, as described above, carbonate rocks were known to occur at various, well-defined stratigraphical levels, and therefore this belt served as one of the key areas in **Karhu's (1993)** carbon isotope study of sedimentary carbonate rocks. The isotope analyses showed that the lowermost carbonates from the Sompujärvi Formation had already attained anomalous positive $\delta^{13}C$ values (+8.6‰) and this trend persisted higher up in the sequence until the uppermost Jatulian carbonate rocks of the Rantavaara Formation where they started to decline back to more normal values.

The most recent report on advances in stratigraphical studies in the Peräpohja area was published in a conference abstract by **Perttunen et al. (1995)**. This work is related to the stratigraphical map of the Peräpohja schist belt, scale 1:200,000. In their lithostratigraphical nomenclature, the authors abandoned the traditional Jatuli- and Kaleva-based names at the group level and replaced them with geographical names, in accordance with international stratigraphical guides (Hedberg 1976, Salvador 1994). This procedure resulted in a two-fold subdivision of the supracrustal sequence into the lower Kivalo Group and the upper Paakkola Group, corresponding to the former Jatulian and Kalevian rocks, respectively. Additional revisions were made at the formation level, including the distinction for the first time of a volcanic formation composed of pillow lavas (the Väystäjä Formation) in the former Kalevian rocks (now the Paakkola Group). The stratigraphy and lithology of the type formations in the southern part of the Peräpohja schist belt are shown in Figure 6.

SUMMARY AND DISCUSSION

The stratigraphical research in Finland was based in the first half of the 20th century on the distinction of cycles of sedimentation separated by events of diastrophism (Sederholm 1932). In the absence of radiometric data, this was the only possible way to proceed, and many historical accounts have shown that this method of deduction was not always devoid of *circulus in demonstrando*. Later age determinations and geological research have brought about revolutions in our understanding of the stratigraphical relations and the geological evolution of the Fennoscandian Shield. The rocks previously included in the Archean rocks have turned out to possess ages spanning a range of more than 1 billion of years and are now distributed between the Archean and Paleoproterozoic formations. It has been discovered that certain parts of the Finnish bedrock were generated in totally different geotectonic regimes (e.g., Karelian, Svecofennian), and therefore direct correlations of their rock units have become meaningless. Also in the course of research, the age relations of certain stratigraphical units have been reversed (e.g., Kalevian and Jatulian). All these developments have rendered previously presented stratigraphical schemes obsolete, but many traditional stratigraphical names have remained in common usage until the present day.

As the above historical overview reveals, there has been the general agreement since the 1920s that two major supracrustal rock compartments, traditionally called with the names Lapponian and Kumpu and their derivatives, occur in Finnish Lapland, but the agreement generally has not extended beyond that. Especially, their relationship to the Karelian formations elsewhere in the shield has been under dispute (Fig. 3). The Karelian formations were originally proposed by Eskola (1921, 1925) as a collective term for Jatulian, Kalevian and Ladogan deposits in eastern Finland and were later defined as comprising Sariolian, Jatulian and Kalevian rocks in Finland (Meriläinen 1980a,b, Simonen 1986). Since it had appeared that the Kalevian are probably younger than the Jatulian rocks without any notable stratigraphic breaks between them, Sederholm (1932) decided to unite the Jatulian and Kalevian rocks into the same cycle of sedimentation, the Karelidic cycle. The arguments for Sederholm to exclude the Lapponian rocks from the Karelidic cycle and give these rocks a special name included 1) their inferred position analogous to that of the rocks of the Bothnian cycle in southern Finland (though being lithologically different, both were interpreted as having been intruded by post-Bothnian granites) and 2) the presence of the great unconformity between the Lapponian formations and the overlying, Jatulian-correlated Kumpu

quartzites.

A common feature of all the Karelian formations in eastern Fennoscandia is that they were deposited on the deeply eroded Archean cratonic basement (e.g., Eskola 1963, Laajoki 1986). We now know that this is also true for many rock units assigned to the Lapponian by Sederholm (1932). The correlation of the Lapponian rocks with Karelian formations seems to have been evident for many prominent geologists from the 1950s to the 1970s in Finland (Fig. 3) and also in northern Sweden (Geijer 1963). For example, Väyrynen (1954) did not mention the term Lapponian at all and there were no difficulties for Eskola (1963) to include Mikkola's Lapponian and Kumpu-Oraniemi series in the 'karelides'. On the other hand, the molasse-like nature of the Kumpu formations, deposited after the main phase of Svecokarelidic folding, was advocated especially by Simonen (1960a,b, 1971, 1980). The Karelian-Lapponian and Kumpu-molasse connections were, however, broken at the turn to the 1980s (Fig. 3) when the Lapponian rocks were regarded as Archean in age and correlated with Archean greenstone belts, such as the Kuhmo-Suomussalmi belt in eastern Finland (Gaál et al. 1978, Silvennoinen et al. 1980). Three Supergroups, Lapponia, Karelia and Svecofennia, were established in Lapland, with only the Jatulian rocks (represented by the Kumpu formation in Lapland) included in the Karelia Supergroup (Silvennoinen 1985). Even though the post-Archean age for the Lapponian rocks has been widely accepted for some time (Silvennoinen 1986, Gaál & Gorbatchev 1987), the division into the above-mentioned supergroups has continued until very recently (Silvennoinen 1998).

In principle, placing of the Lapponia Supergroup beneath the Jatuli Groups of the Karelia Supergroup is in agreement with the original definition of Sederholm (1932) according to which the Lapponian rocks are pre-Jatulian in age. However, isotopic and geological studies carried out in the1980s and 1990s have provided ample evidence that Sederholm's "Jatulian" Kumpu quartzites are much younger (< 1.88 Ga) than the Jatulian supracrustal rocks elsewhere in the Fennoscandian Shield (Rastas et al. 2001, *this volume*, Hanski et al. 2000, 2001). On the other hand, the rocks previously called Lapponian seem to cover a very large range of ages. Most importantly, a large part of the Lapponian volcanic rocks have turned out to be younger than Jatulian, approaching an age of c. 2.0 Ga (Rastas et al. 2001, *this volume*, Hanski et al. 1998). Furthermore, typical Jatulian quartzites and hypabyssal layered sills within them as well as pre-Jatulian volcanic rocks correlated with Sumi-Sariolian rocks are found among the "Lapponian" rocks.

Along with these results a terminological crisis was unavoidable. This was conveniently circumvented by shifting to a totally new, formal lithostratigraphical nomenclature (Räsänen et al. 1995, Lehtonen et al. 1998). As was stated in the Introduction, a similar stratigraphical reform had already been in process elsewhere in Finland for some years. The previous rocks of the Karelia Supergroup or Kumpu formations were assigned to the Lainio and Kumpu Groups, while the Lapponian rocks or the rocks of the Lapponia Supergroup were divided between the Salla, Onkamo, Sodankylä, Savukoski and Kittilä Groups (Lehtonen et al. 1998). Abandonment of the traditional stratigraphical names in formal classification of rock units has proceeded rather smoothly in northern and eastern Finland, but not everywhere on the Fennoscandian Shield. Particularly, there has been a major confrontation of the old and new systems in the Kola Peninsula as evidenced by the vigorous public debate carried out recently by Melezhik and Sturt (1998a,b), Smolkin (1998) and Sharkov and Smolkin (1998).

New research results from Finnish Lapland also include the proposal that the Kittilä Group represents an allochthonous piece of ancient oceanic crust (Hanski 1997). This interpretation and the recognition of Paleoproterozoic ophiolitic terranes elsewhere (e.g., Kontinen 1987, Scott et al. 1992) which may have originated hundreds or thousands of kilometres from their present sites, and whose geological history may have little or nothing in common with the continental mass to which they are now physically attached, has important implications for stratigraphical research. In the least, they provide a serious warning against the practise of extending a 'layer-cake' stratigraphy over a whole craton without caution. The Kittilä Group and the overlying coarse-clastic, molasse-like metasediments of the Lainio and Kumpu Groups possess lithological characteristics and geological histories which hardly find parallels in other Karelian formations which partly explains why previous workers have had difficulties in squeezing them into the standard Karelian stratigraphy.

In retrospect, it is interesting to note that there were obvious reasons for the cautiousness that several notable researchers had in taking the final stand in regard to the correlation of the Lapponian and Kumpu formations with other Karelian formations. Although the Kumpu-Jatulian connection was regarded as a viable alternative, many lithological observations were at odds with this option. Hackman (1927) pointed out the absence of intrusive metadiabases and metagabbros in the Kumpu formations. Mikkola (1941) in turn emphasized the obvious lack of carboniferous and carbonate-rich interbeds in these rocks and on the

other hand, drew attention to the general lithological similarity between the Lapponian and Jatulian quartzites. He also considered it possible that some granitoid fragments found in the Kumpu-Oraniemi series conglomerates were derived from the syenite series (Haaparanta Suite) plutonic rocks. Even Sederholm (1932) did not appear confident of the Kumpu-Jatulian correlation although much of his stratigraphical interpretation hinged on this assumption.

Not all the stratigraphical problems have yet been solved, but new tools, like NORDSIM in Stockholm, are now available for dating purposes. Although the radiometric datings are of paramount importance in stratigraphical research, they always need critical evaluation against the background of field evidence. The irony lies in the fact that some of the first zircon dates from northern Finland, namely the c. 2.2 Ga age for a diabase occurring among Kumpu-type quartzites at Värttiövaara (see Rastas et al. 2001, *this volume*) and the Archean age for a felsic rock from the western part of the Kuusamo schist belt (see Räsänen & Vaasjoki 2001, *this volume*) were among the evidence which was utilized to assign the Kumpu formations to the Middle Jatulian Group and all the Lapponian rocks to the Archean (Rastas 1980, Silvennoinen et al. 1980). Furthermore, should the first age determinations from the Peräpohja schist belt have come from acid dikes containing inherited zircons (see Perttunen & Vaasjoki 2001, *this volume,* Hanski et al. 2001*, this volume*), the interpretations of the geological evolution in that area would have been, at least temporarily, quite different. Datings can be utilized to solve problems, but they surely will sometimes create new ones and impose new challenges for further field geological research, as exemplified by the finding of unexpectedly young detrital zircons in arkosites from the Peräpohja schist belt (Perttunen & Vaasjoki 2001, *this volume*). Thus the stratigraphical schemes that we now espouse, may again be, at least partly, at stake.

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Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
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U-Pb GEOCHRONOLOGY OF THE PERÄPOHJA SCHIST BELT, NORTHWESTERN FINLAND

by
Vesa Perttunen and Matti Vaasjoki

Perttunen, V. & Vaasjoki, M. 2001. U-Pb geochronology of the Peräpohja Schist Belt, northwestern Finland. *Geological Survey of Finland, Special Paper 33,* 45-84. 14 figures, 2 tables and 6 appendices.

New U-Pb results as well as analytical data for some published preliminary age results are presented. The samples include basement rocks, layered intrusions, mafic and felsic sills and dykes, intersecting plutonic rocks and metasedimentary rocks. Zircon has been the principal mineral analyzed, but baddeleyite, monazite and titanite have also been used. U-Pb dating of mafic volcanic rocks has failed so far due to the lack of suitable uranium-bearing minerals.

Numerous mafic layered intrusions occur near the basement contact of the rocks of the Peräpohja Schist Belt. The results indicate that intrusion occurred near 2.43 Ga but much of the data show signs of later geological events.

The analyses reveal wide-spread magmatism at 2.2 Ga, when concordant, differentiated mafic sills intruded the quartzites of the Peräpohja area. These sills can be correlated with similar Haaskalehto-type rocks in Central Lapland and the gabbro-wehrlite association in eastern Finland. The existence of the 2.05 Ga old diabases, common in Central Finnish Lapland, has not been confirmed in the study area. Instead, there may be a set of about 2.1 Ga old diabases.

All analyzed felsic dykes contain xenocrystic, Archean zircons and their emplacement age is unknown. Also, especially felsic rocks from the plutons of the Haaparanta Suite contain inherited zircons, while some zircon as well as titanite analyses confirm the age of about 1.89 Ga, common elsewhere in Western Lapland and Northern Sweden.

The detrital zircons separated from all of the (Jatulian) quartzite samples show, irrespective of their stratigraphic setting, an Archean provenance, as does a sample from a quartzite interbed in an apparently Archean greenstone formation in the basement area. Zircons from the uppermost, (Kalevian) Martimo Formation give a Pb/Pb age result >2.58 Ga. Results from an arkosite, a little lower in stratigraphy, give an apparent age of 1985 Ma for the provenance of the material, so existence of the "Upper Kaleva" also in the Peräpohja area is plausible.

Key words (GeoRef Thesaurus, AGI): absolute age, U/Pb, zircon, baddeleyite, monazite, titanite, layered intrusions, dikes, diabase, plutonic rocks, metasedimentary rocks, Paleoproterozoic, Archean, Peräpohja, Lappi Province, Finland

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INTRODUCTION

The supracrustal rocks of the Peräpohja Schist Belt rest unconformably on the Archean Pudasjärvi Granite Gneiss Complex outcropping to the southeast of the belt. The basement rocks include Archean granitoids and remnants of greenstone belts as well as Paleoproterozoic mafic layered intrusions. In the north and west, the rocks of the Peräpohja Schist Belt are bounded by younger Paleoproterozoic granitic rocks of the Central Lapland Granite Complex. The Peräpohja Schist Belt itself consists of metasedimentary and mostly mafic metavolcanic rocks. They are cut by the plutonic rocks of the Haaparanta Suite as well as by mafic and minor felsic sills and dykes. The grade of metamorphism is low, mainly greenschist facies, reaching amphibolite facies in the north.

Regional geological mapping in the area has been carried out at several stages to the scales of 1:400,000 (Hackman 1910a and 1910b, Enkovaara et al. 1952 and 1953) and 1:100,000 (Perttunen 1971a, 1971b, 1972 and 1975). The layered mafic intrusions in the area host, apart from the huge Elijärvi Cr-deposit (Kahma et al. 1962), also several PGE-mineralizations (Halkoaho et al. 1990a and 1990b, Huhtelin et al. 1990). Minor sulfide vein occurrences (Rouhunkoski & Isokangas 1974) are also known to exist.

The first radiometric dates in the Peräpohja area derive from the early 1960's. These included a Rb-Sr age 1780 Ma for the biotite from the intersecting Nosa granodiorite (A154) which gave the minimum, Proterozoic age for the Peräpohja Schist Belt (Wetherill et al. 1962). The zircons from a granite (A52) belonging to the basement yielded an Archean U-Pb age of 2618 Ma (Kouvo & Tilton 1966), establishing the maximum age limit for the evolution of the schist belt. Since the pioneering work of Dr. O. Kouvo at the Geological Survey of Finland (GTK), the U-Pb method has remained the main method of dating in northern Finland.

Initially, felsic rocks were preferred for sampling as the zircon content in mafic rocks was considered too low. In the Peräpohja area, the U-Pb studies of diabases started with titanite in the late 1960's. Zircon was identified in coarse-grained, plagioclase-rich light-colored diabase types (often called albite diabases), which concentrate HFS-elements like Zr (Sakko 1971). Samples contained enough zircon (100-200 mg) for the borax fusion method then employed. After that discovery, dozens of "albite diabases" have been analyzed. Most of the zircon crystals are large and turbid, their uranium content is high, and thus the obtained results are rather discordant. Later, baddeleyite, often intergrown with zircon, was detected in the heavy mineral concentrates. When found, titanite was also analyzed.

The dating of the layered intrusions in the Peräpohja area started in the 1970's after the discovery of coarse-grained pegmatoids characterized by large zircon crystals in most of the intrusions. The heaviest zircon fraction is often low in uranium, resulting in negligible discordance while the large amount of zircon allows several fractions to be analyzed. Common lead is usually, but not always, low. Baddeleyite, often found along with zircon as separate crystals or as an intergrowth, may be rather abundant or even the principal zirconium mineral (Kouvo 1977).

In addition to albite diabases and layered intrusions, there are a few U-Pb age determinations from zircons of plutonic rocks and felsic dykes intersecting the supracrustal rocks of the Peräpohja Schist Belt. Some analyses have also been made on detrital zircons of quartzites and other sedimentary rocks. No suitable minerals for U-Pb dating have been obtained from mafic volcanic rocks.

Preliminary age results without analytical data from the Peräpohja rocks have been reported in many contexts. In this paper, we report in detail U-Pb isotope results on the various samples collected during the last decades. The samples represent (1) the rocks of the Pudasjärvi Basement Complex, (2) layered intrusions, (3) diabase sills and dykes, (4) felsic dykes, (5) plutonic intrusive rocks, and (6) detrital zircons from metasedimentary rocks. In addition, a short description is given of samples where U-Pb dating has been unsuccessful. Since the 1980's,Sm-Nd isotope analyses have shed new light on the evolution of the Peräpohja Schist Belt. Some pertinent age results will be discussed in conjunction with the assessment of the U-Pb data

Most of the samples have been collected during the mapping activities of the Geological Survey of Finland. Some samples were taken by Outokumpu Co. in connection with the mining and exploration activities of layered intrusions and plutonic rocks of the Haaparanta Suite. The samples from the Misi area in the northeastern corner of the schist belt have been obtained from Rautaruukki Co. The sampling sites are shown on a map in Figure 1 and the analytical data in Appendices I-VI. The results are summarized in Table 1, and a list of the samples found unsuitable for U-Pb work is given in Table 2.

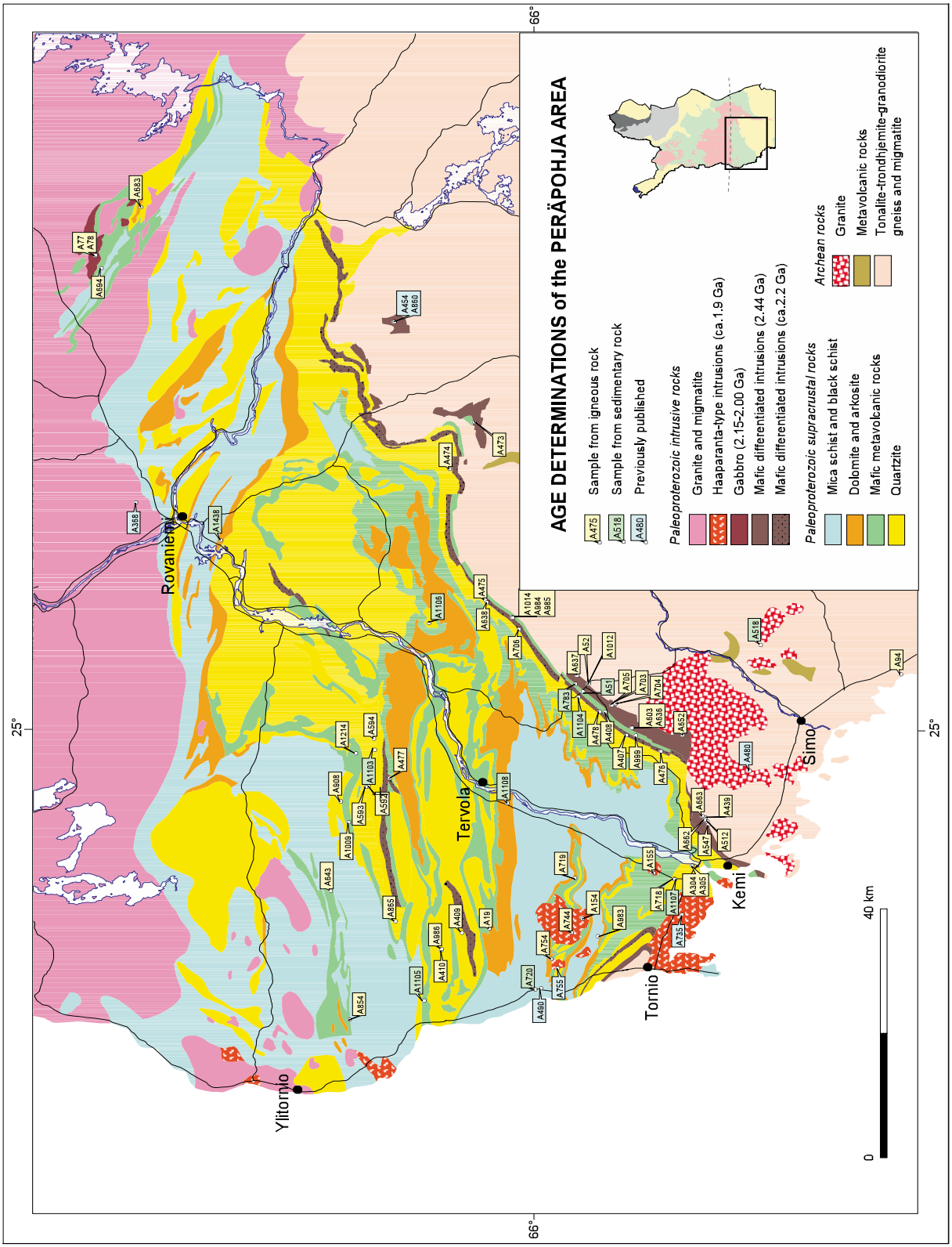

Fig. 1. Map of surface geology also showing radiometric sampling sites in the Peräpohja schist belt, Finland.

STRATIGRAPHY OF THE PERÄPOHJA SCHIST BELT

The stratigraphy of the well-preserved supracrustal rocks of the Peräpohja Schist Belt has been studied for more than a century. The first stratigraphic subdivision of the rocks into older Kalevian and younger Jatulian rocks was based on the metamorphic grade (Hackman 1914), but it was soon established that their age relationship was in fact opposite (Mäkinen 1916). The overall geology and most of the stratigraphic units have been established for a long time (Härme 1949). Later, in accordance to the division adopted for Northern Finland (Silvennoinen et al. 1980), the supracrustal rocks were divided into the Karelia Supergroup, comprising Lower, Middle and Upper Jatulian Groups, and the overlying Svecofennia Supergroup, represented by the litohologies of the previous Kalevian formation (Perttunen 1980, 1985).

On the stratigraphical map of the Peräpohja area, the supracrustal rocks were divided into two units: the lower Kivalo Group containing sedimentary rocks of the orthoquartzite-dolomite association as well as mafic volcanic rocks, and the upper Paakkola Group consisting of pelitic and black schists with minor mafic volcanic rocks (Perttunen et al. 1995; Fig. 2). The two groups correspond, respectively, to the Jatulian and Kalevian in the traditional classification of Karelian rocks. The Kivalo-Jatulian correlation in particular is supported by the pronounced positive $\delta^{13}C$ anomaly in carbonate rocks (Karhu 1993), interpreted as being related to a global increase of atmospheric oxygen c. 2.2 Ga ago (Karhu & Holland 1996).

The supracrustal rocks have been penetrated by many groups of intrusions. There are at least three sets of diabase sills and dykes, as well as the plutonic intrusive rocks of the Haaparanta Suite ranging petrographically from granodiorite and monzonite to gabbro. Two petrographically different granitic intrusions connected with the Central Lapland Granite Complex exist in the northern area. Practically all of the volcanic rocks of the area are mafic, and attempts of dating them by the U/Pb method have so far been unsuccessful.

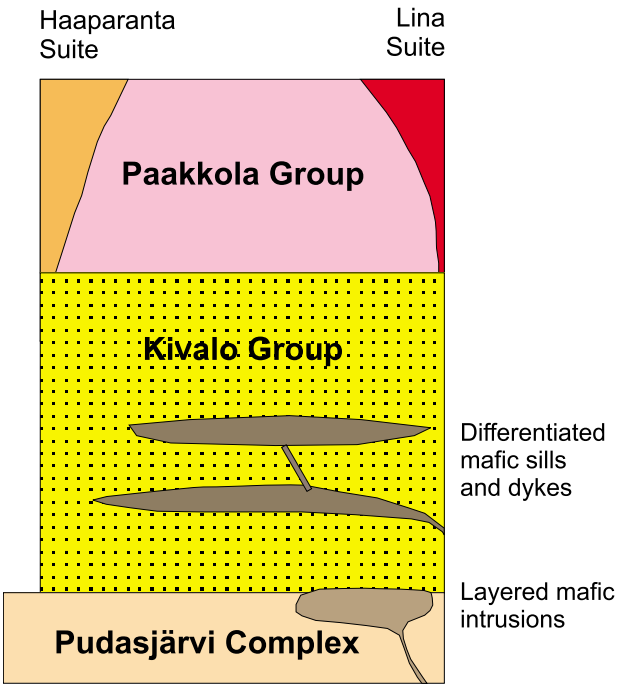

Fig. 2. General stratigraphy of the Peräpohja schist belt according to Perttunen et al. (1995).

ISOTOPIC STUDIES

The Pudasjärvi Gneiss Complex

The rocks of the Pudasjärvi Gneiss Complex include felsic para- and orthogneisses as well as diverse granitoids. The Oijärvi greenstone belt contains amphibolites and felsic gneisses. Due to sparse outcrops, the geological picture is not clear. There are only a few isotope analyses from the rocks of the Pudasjärvi Gneiss Complex.

Sample **A52 Sompujärvi** is a coarse-grained, slightly porphyritic microcline granite, taken some 50 m from the contact to the Penikat layered intrusion. This granite forms large areas in the Pudasjärvi Complex, and occurs as clasts in the basal conglomerates of the Peräpohja Schist Belt (Sompujärvi Formation). The six zircon fractions yield an Archean upper intercept age of 2620±21 Ma with a lower intercept at 515±97 Ma, but the MSWD is relatively high at 8.4. Omitting the old zircon analyses by borax fusion (Kouvo & Tilton 1966) in the age calculation does not significantly alter the result, as the new U-Pb data (4 fractions) define a reasonably good linear fit (MSWD=2.7) with intercepts at 2619±18 and 504±95 Ma. The sample contains nearly concordant Proterozoic monazite with a $^{207}Pb/^{206}Pb$ age of 1800±3 Ma (Fig. 3).

A94 Kuivaniemi is from a gray, banded, granodioritic and migmatized orthogneiss in the more

central part of the Pudasjärvi Gneiss Complex. Rather surprisingly, the abraded fraction F yields a slightly lower $^{207}Pb/^{206}Pb$ age than the other zircon fractions. Although the six zircon analyses exhibit considerable scatter (MSWD=15), the upper intercept age is relatively precise at 2667±11 Ma while the lower intercept passes within experimental error through the origin. Monazite in this sample is nearly concordant with a $^{207}Pb/^{206}Pb$ age of 2664±4 Ma (Fig. 3).

A439 Elijärvi is a granitic rock 100 m below the foot-wall contact of the Kemi layered intrusion. Most of the zircons in the sample occur as long, euhedral crystals. 110 is the dominating prism type. Some of the grains are subhedral and a little rounded. Five discordant zircon fractions define a reference line with intercepts at 2685 and 90 Ma (Fig. 3). As the granitic rock brecciates the rocks of the layered intrusion, the magma apparently results from the heating effect of the layered intrusion on the basement rocks. Old zircon has survived the melting and crystallizing processes, as is also suggested by the albitite dyke A512 (see page 56).

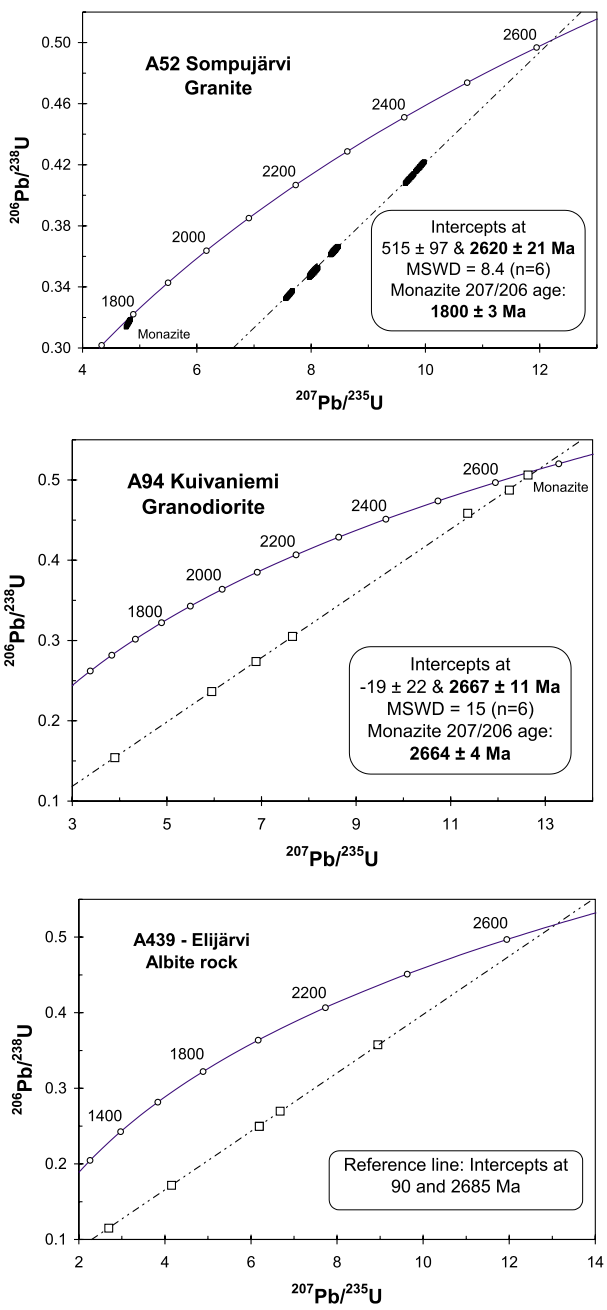

Fig. 3. Concordia diagrams presenting data for the Archean granitoids A52-Sompujärvi and A94-Kuivaniemi and the remobilized basement rock A439-Elijärvi. Note that in this and other figures the dimensions of the axes vary from one diagram to another.

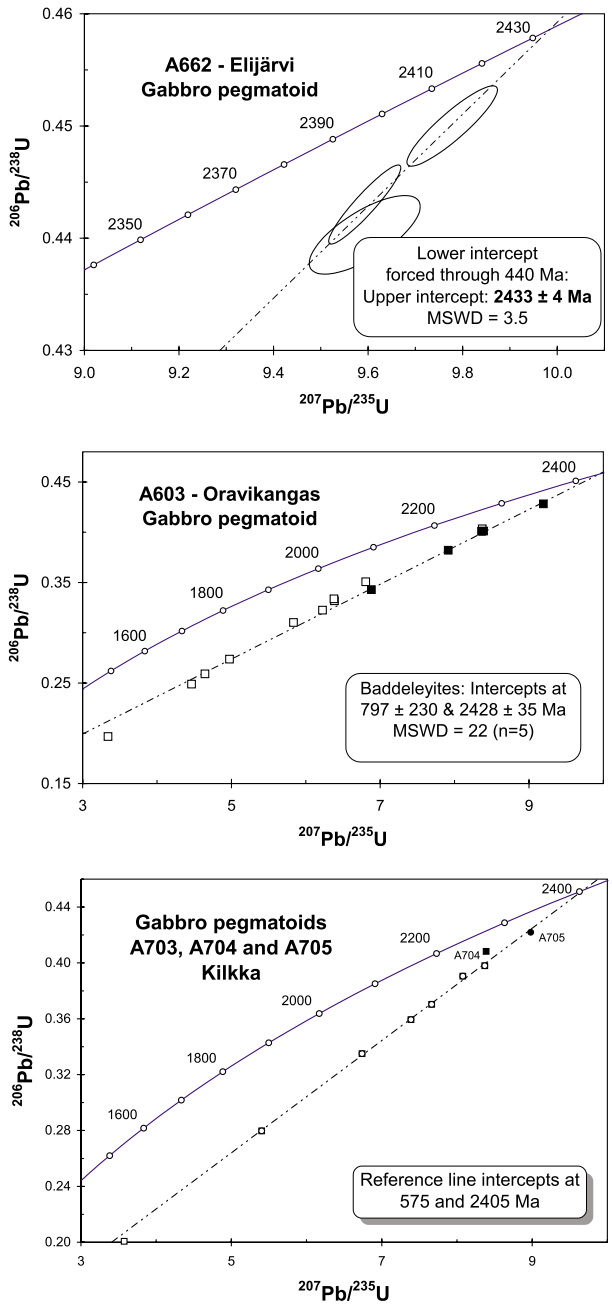

Fig. 4. Concordia diagrams presenting the data from the layered gabbro intrusions. The intercepts for A603-Oravikangas are calculated on the basis of the baddeleyite analyses shown as black squares.

LAYERED INTRUSIONS

There is a coherent east-west trending chain of mafic layered intrusions within the Archean Pudasjärvi Complex near the southern margin of the Peräpohja Schist Belt. The intrusions include three large ones, Tornio (Söderholm & Inkinen 1982), Kemi (Kujanpää 1986) and Penikat (Alapieti & Lahtinen 1986) in the west, and several small intrusions (Konttijärvi, Suhanko, Lihalampi, Kilvenjärvi, Nutturalampi, Kuohunki and Siikakämä; Alapieti et al. 1989 and 1990) in the Portimo area in the east. These layered gabbros clearly transgress the Archean basement area and are among the earliest Paleoproterozoic igneous rocks within the Fennoscandian Shield. The intrusions continue further to the east to Koillismaa (Alapieti 1982) and across the national border to Russia (Amelin et al. 1995).

Several U-Pb datings have been performed on the Kemi and Penikat intrusions. Zircons have been extracted mostly from coarse-grained pegmatoid gabbros. In addition to the traditional U-Pb method on zircons, Pb-Pb and Sm-Nd methods have been utilized (Alapieti 1982, Huhma et al. 1990, Manhes et al. 1980), and a Hf isotope study on zircons of sample A662 was carried out by Patchett et al. (1981). All different dating methods have yielded ages in the range from 2440 Ma to 2450 Ma. Some zircon samples of the Siikakämä (A454, A860; Mertanen et al. 1989) and the Penikat intrusions (this study) show a more complex history since their crystallization.

The Kemi layered intrusion

The Kemi intrusion is a lenticular c. 15 km long and up to 2 km wide ultramafic-mafic body and hosts the Kemi chrome mine of the Outokumpu Chrome Co. Most samples are from coarse-grained gabbro pegmatoids. The foot wall rocks of the intrusion consist of late Archean granitoids, often intensively albitized. Albite-rich granite (as sample A439 Elijärvi, see above) brecciates the rocks of the layered intrusion near the foot wall contact and fine-grained albitite dykes (as A512) penetrate the intrusion. A polymictic conglomerate of the Sompujärvi Formation lies unconformably on top of the intrusions (Perttunen 1991). Manhes et al. (1980) report common lead results for 5 whole rock analyses from basic-ultrabasic rocks from Elijärvi, which give an age estimate of 2440±160 Ma.

Sample **A662 Elijärvi** contains reddish brown, clear zircon, mostly as fragments after crushing and grinding. All three U-Pb analyses are very little discordant (Fig. 4), but do not coincide within experimental error, and one of them (fraction B) has a $^{207}Pb/^{206}$ Pb age slightly younger than the others. However, the Pb-Pb age 2430±4 Ma for the least discordant fraction (A) can be regarded as a relatively close estimate for the time of emplacement of the rock. A similar result, 2433±4 Ma is obtained when the lower intercept is forced through the 440 Ma point of the concordia curve. Zircons from the albitite (A512) and brecciating granite (A439) indicate Archean ages (see p. 49).

The Penikat layered intrusion

The Penikat intrusion is 2 to 3 km wide and 23 km long. Four samples have been analyzed from this intrusion. Sample **A603 Oravikangas** is a coarse-grained gabbro pegmatoid near the upper contact of the Penikat Intrusion. The primary minerals include plagioclase and clinopyroxene, the latter altered to albite, epidote, and amphibole.

The first eight zircon fractions (A-H) of A603 Oravikangas were analyzed in 1975 and 1976 while the other (I-O) fractions were done in the late 1980s. The early analyses have an upper intercept at 2288±38 Ma on the concordia diagram. The result is younger than expected, possibly indicating effects of later processes. This is also suggested by the later analyses on baddeleyite-zircon aggregates, which give intercepts at 2428±35 and 797±230 Ma (Fig. 4). The rather large MSWD (22) probably indicates some remaining sample heterogeneity, probably arising from the zircon overgrowths on the baddeleyites. The confirmation of this hypothesis would require spot analyses by an ion probe.

The **Kilkka** samples **A703, A704** and **A705** are from three gabbro pegmatoids in the middle of the layered Penikat intrusion. The sampling sites lie about 400 m apart from each other, with A703 as the lowest and A705 as the highest in the internal stratigraphy of the Penikat intrusion.

The zircons in sample A703 exhibit predominantly a simple prismatic-pyramidal morphology, but crystals with higher order index faces are not uncommon. The color is generally darkish brown, but lighter colored, platy grains do also occur. Sample A704 is similar, but contains also a large amount of rutile. In sample A705 there occur, in addition to the predominant dark brown euhedral zircon variety, longish lighter colored crystals with an L/B up to 5.

The seven analyzed fractions from A703 (Fig. 4) exhibit a usual discordancy pattern for magmatic rocks, as both the degree of discordance and uranium

contents increase with decreasing density. The analytical points do not, however, form a well defined linear trend (MSWD for all data 46). If the very discordant fraction F is omitted, a reference line with intercepts at 2405 and 575 Ma can be drawn through the other six data points. The relatively little discordant data from the other two samples plot off the reference line, A704 on the younger and A705 on the older side. The diffusion model (Wasserburg 1963) age for A705 is 2417 Ma. Thus the data from the Kilkka samples may be interpreted as being consistent with an emplacement age slightly in excess of 2.4 Ga, but the scatter of the data imply a secondary disturbance, possibly during the emplacement of the differentiated diabase sills and dykes.

Diabase sills and dykes

Several sets of diabase sills and dykes of probably more than one generation cut the Palokivalo Quartzite Formation. Most of the dated samples were taken from layered, 10-200 m thick concordant sills. They are frequently strongly differentiated, with ultramafic olivine (±clinopyroxene) cumulates in the lower part, clinopyroxene cumulates in the middle part, and gabbroic differentiates in the upper part. The ultramafic cumulates are characterized by brown amphibole as an intercumulus mineral. The main, gabbroic parts of the sills consist of plagioclase-clinopyroxene-magnetite cumulates. Both the proportion of plagioclase and the grain size of the rock increase upwards. In the uppermost part, quartz appears as an intercumulus mineral. The primary minerals have mostly been replaced by metamorphic assemblages: olivine by serpentine pseudomorphs, clinopyroxene by uralite, and plagioclase by albite. These rocks have often been called albite diabases and are similar to the sills assigned to the gabbro-wehrlite association in eastern Finland (Hanski 1987) and the Haaskalehto-type diabases in Central Finnish Lapland (Lehtonen et al. 1998).

The geology of the Misi area in the northeastern corner of the schist belt is not settled in detail because of poor exposure of the bedrock. Quartzitic and dolomitic sedimentary and mafic metavolcanic rocks predominate in the area. Gabbros and diabases intrude the quartzites. The metamorphic grade is rather high and scapolitization ubiquitous (Nuutilainen 1968). Preliminary age results (2.16 ± 0.015 Ga) of A694 were published by Patchett et al. (1981), but no analytical data were presented.

The basement rocks of the Peräpohja Schist Belt contain abundant diabase dykes, apparently representing swarms of different ages. So far, there is only one dating with an age of 2.15 Ga of the diabases within the basement area (A480 Sipojuntti, Perttunen 1987).

Samples and results

Samples **A304 Kallinkangas** and **A305 Laurila** were taken from a coarse-grained, albite-rich diabase sill in quartzite. Sulfide-bearing quartz-carbonate veins associated with the diabase have been targets for gold exploration (Tegengren 1951).

Of the 14 mineral fractions analyzed from sample A304 Kallinkangas, the relatively light (d<4.2) zircon fractions C-I are rather discordant, and also seem to exhibit a random scatter (Fig. 5). The heavier zircon fractions and the sole baddeleyite (A, B and K-M) define a relatively good (MSWD=1.8) linear trend with intercepts at 2216±10 and 241±38 Ma. The nearly concordant titanite is clearly younger and registers a $^{207}Pb/^{206}Pb$ age of 2173±34 Ma, the largish error arising from a rather poor lead isotopic composition run.

The three baddeleyite fractions from sample A305 Laurila plot close to the linear trend defined by the best analyses from sample A304 (Fig. 5), and form by themselves a good regression line (MSWD=0.67) with intercepts at 2221±5 and 311±30 Ma. Taken together, the data from samples A304 and A305 may be interpreted as indicating an intrusion age of c. 2220 Ma for the Kallinkangas albite diabase sill.

A differentiated diabase sill, tens of kilometers long and up to 200 m thick, occurs within the Palokivalo Quartzite Formation. Samples A407 Oravikangas, A408 Puukkokumpu I, A474 Konttikivalo, A475 Hattuselkä I, A476 Tornivaara and A706 Runkausvaara are all apparently from this sill.

Sample **A407 Oravikangas** is an albite-rich, coarse-grained variety near the top of the sill. The analyses of 8 zircon, two titanite, and two baddeleyite/zircon mixture fractions are shown on a concordia diagram in Figure 6. Especially the borax fusion analyses on the zircons are very discordant and the common lead in titanites is fairly high. If these are omitted in the calculation,the other data define a reference line with intercepts at 2225 and 275 Ma.

A408 Puukkokumpu I is near the upper contact of the sill, resembling petrographically and stratigraphically the other diabase samples from the sill. The amount of zircon is high. Most of the crystals are turbid and light-colored, while some crystals are clear and greenish yellow. Figure 6 shows that especially the turbid zircon fractions (B-D) scatter rather

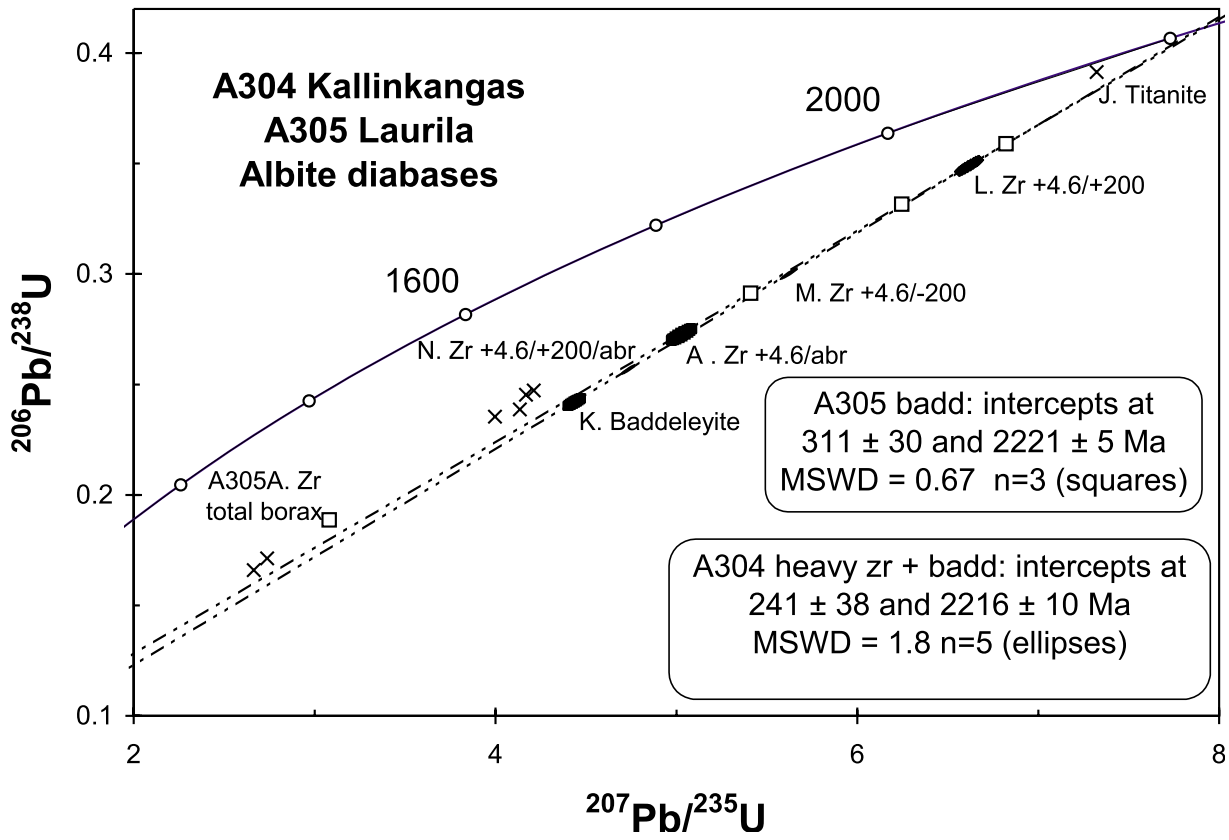

Fig. 5. Analytical results for the samples A304-Kallinkangas (ellipses indicating analytical uncertainty for heavy zircons and baddeleyite used for the calculation, the other analyses as crosses) and A305-Laurila (squares).

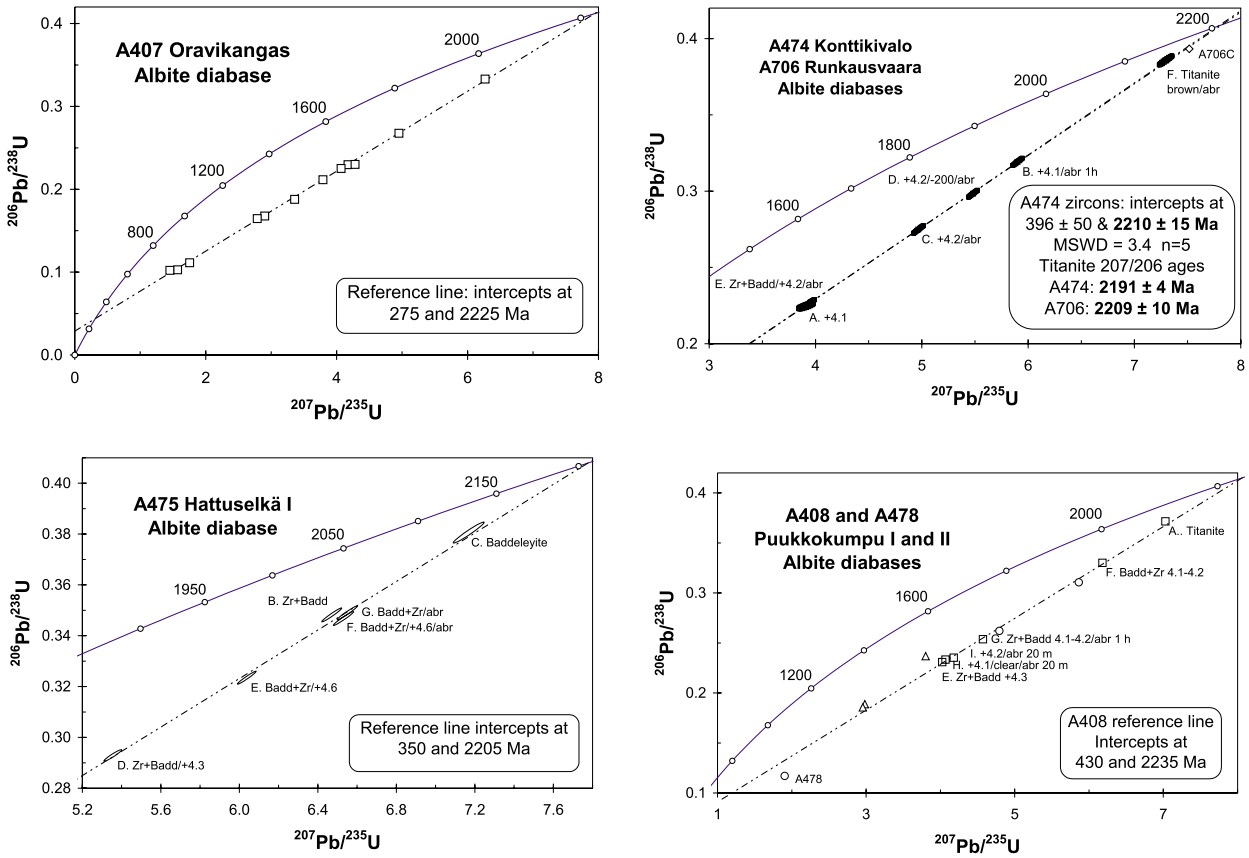

Fig. 6. Concordia diagrams for albite diabase sills within the Palokivalo quartzite formation. In the Puukkokumpu diagram squares indicate the clear zircon and zircon baddeleyite intergrowth fractions of A408, the triangles the turbid zircon fractions from the same sample and the circles analyses from A478.

randomly, nor do the clear zircons or the zircon-baddeleyite intergrowths (E-I) define an unequivocal linear trend, although they plot in the vicinity of a reference line with intercepts at 2235 and 430 Ma. The slightly discordant titanite has a $^{207}Pb/^{206}Pb$ age of 2192±15 Ma.

A478 Puukkokumpu II. The sill Puukkokumpu II together with A638 Hattuselkä II lies higher in the Palokivalo Quartzite Formation than the other dated diabases in the area. The contacts are not exposed, but according to magnetic surveys the thickness of the sill can be estimated as 30 m. The sample contains rather much zircon, mostly as turbid, brownish grains. Some zircon crystals are platy in shape. Another zircon population contains transparent, greenish yellow crystals, often intergrown with green amphibole and baddeleyite. The content of barite is high. Point A represents coarse-grained zircon (Ø<160 µm) with a density of >3.8 cm^3. Fraction B contains transparent zircon intergrown with baddeleyite. Fraction C is composed of transparent zircon only. The common lead content in all three fractions is high, which makes the results dependent on the common lead correction. Nevertheless, fractions B and C plot close to the reference line for sample A408 (Fig. 6). The concentration of common lead in the titanite is so high that calculation of U-Pb ratios was pointless.

Sample **A474 Konttikivalo** is a typical coarse-grained, pink, albite-rich diabase variety in the upper part of a sill within the Palokivalo quartzite. Most of the zircon crystals are clear and colorless, but turbid zircon crystals are also common. Baddeleyite is mostly surrounded by a rim of turbid zircon. The somewhat discordant zircon and zircon-baddeleyite intergrowth analyses define a relatively good (MSWD=3.4) linear trend with intercepts at 2210±15 and 396±50 Ma (Fig. 6). The slightly discordant titanite has a $^{207}Pb/^{206}Pb$ age of 2191±4 Ma.

A706 Runkausvaara. According to drillings, the Runkausvaara diabase sill is about 200 m thick. The sill is distinctly differentiated with ultramafic cumulates near the base. The sample is from the light-colored gabbroic rock rich in plagioclase (albite) and was taken from local, large boulders at the road. There is plenty of greenish grey zircon in the light fractions. The typical crystal faces are (100) and (001), the (110) prism has not been detected. Twin crystals are common. The titanite is clear, brown, and exists as separate grains. Only two zircon fractions and one of titanite were analyzed. Titanite is nearly concordant with a $^{207}Pb/^{206}Pb$ age of 2209±10 Ma, while the zircons are very discordant (Fig. 6).

A475 Hattuselkä I is a coarse-grained diabase variety with abundant baddeleyite, mostly intergrown

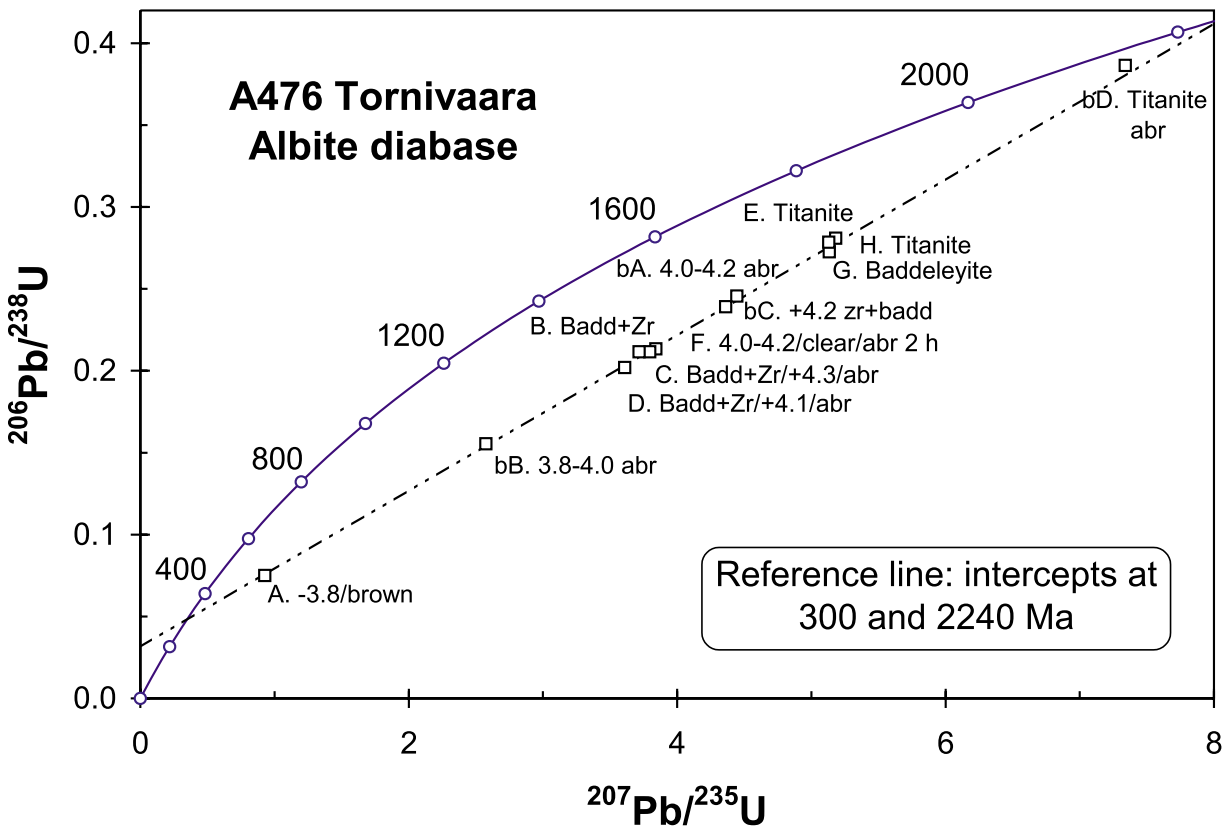

Fig. 7. Concordia diagrams for samples A476 Tornivaara.

with zircon, in the heavy fraction. Barite is also present. Zircon is mostly turbid, and white, pink or brownish in color. The crystals are mostly simple prisms with asymmetric pyramid faces. Analyses of the zircon, baddeleyite and zircon-baddeleyite intergrowths scatter in excess of analytical error (Fig. 6), but most data plot in the vicinity of a reference line with intercepts at 2205 and 350 Ma. Titanite cannot be used for U-Pb calculations because of its high common lead content, the $^{206}Pb/^{204}Pb$ ratio being 18.3.

A476 Tornivaara is a coarse-grained diabase near the upper contact of the sill against the quartzite of the Palokivalo Formation. The metamorphic mineral assemblage includes albite, actinolite, epidote, chlorite, and quartz. Barite, pyrite and chalcopyrite occur as accessory minerals with zircon, baddeleyite and titanite. Light zircon crystals (d<4.0 g/cm^3) are turbid and green in color. Some of the crystals are brown and translucent. Baddeleyite is abundant and is mostly intergrown with zircon. The rock was sampled on two occasions (samples A476 and A476b). The results exhibit an extreme range of discordancy on the concordia diagram (Fig. 7) and plot in the vicinity of a reference line with intercepts at 2240 and 300 Ma. The $^{207}Pb/^{206}Pb$ age of the least discordant titanite fraction is 2198±15 Ma.

A409 Virkkumaa. The Virkkumaa diabase occurs as a sill in the quartzite of the Palokivalo Formation. The sampling site is near the upper contact of the sill. Most of the zircon occurs as pale-colored, turbid crystal fragments in the processed samples. Also some transparent clear or brown crystals have been found. The clear crystals are long with L/B ratios sometimes >10 while the dark crystals are short. The results (Fig. 8) for the turbid fractions exhibit a random scatter, one HF-treated aliquot being slightly reversely discordant. However, the clear fractions A, B, P and R define a reasonably good (MSWD=1.6) linear trend with intercepts at 2213±5 and 304±40 Ma.

A592 Luppovaara I. The sample is from near the upper contact of a differentiated diabase sill in the Luppovaara Quartzite which belongs to the Palokivalo Formation. The amount of zircon was too low to allow analysis. Four titanite fractions give an upper intercept of 2217±10 Ma (Fig. 8).

A593 Luppovaara II is a 50-m thick diabase dyke distinctly intersecting the Luppovaara quartzite and also the diabase sill A592. The dyke is light-colored, mostly pink, medium-grained, only slightly differentiated, and is rich in plagioclase (albite). Many of the zircon crystals are highly idiomorphic, platy and water-clear; some are turbid. Long crystals are rare. The amount of zircon and baddeleyite in the sample was low, so only two fractions could be analyzed - turbid

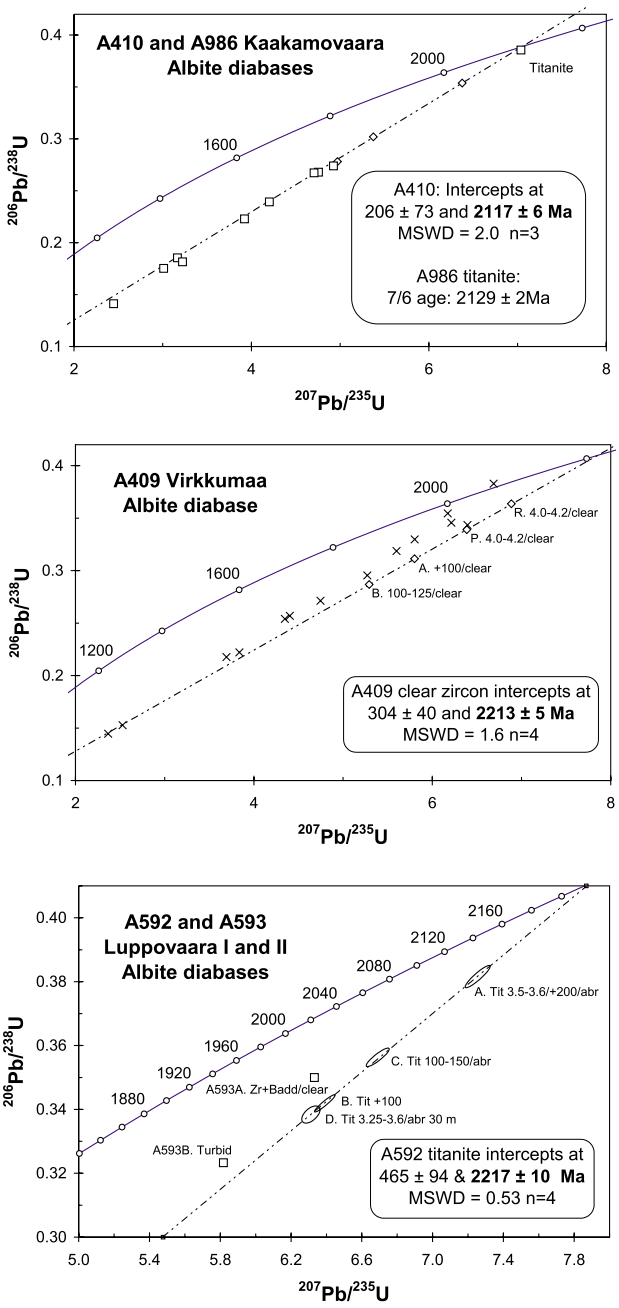

Fig. 8. Concordia diagrams for A. samples A410 Kaakamovaara (diamonds) and A986 Kaakamovaara (squares), B. A409 Virkkumaa with crosses indicating turbid and diamonds clear zircon fractions, and C. samples A592 and A593 from Luppovaara.

zircons, and clear zircons with a little baddeleyite. Both fractions have $^{207}Pb/^{206}Pb$ ages around 2100 Ma and plot on the younger side of the titanite regression for sample A592 (Fig. 8).

Sample **A410 Kaakamovaara I** was taken from a thin, coarse-grained sill in the Kaakamovaara quartzite of the Palokivalo Formation. As the amount of zircon was low, a new sample A986 was taken at a distance of some 100 m from the first locality. Zircon crystals in A410 are mostly simple tetragonal large prisms with a light brown, turbid color. Sometimes the core of the crystals consists of clear zircon. The

amount of titanite in the samples is fairly high. Three zircon analyses (Fig. 8) form a relatively good linear trend (MSWD=2) with intercepts at 2117±6 and 206±73 Ma.

A986 Kaakamovaara II is a differentiated diabase sill c. 100 m higher in the same quartzite as A410 Kaakamovaara I, which belongs to the undifferentiated diabase type. A general feature of the data (Appendix III) is the low $^{207}Pb/^{206}Pb$ age. Unfortunately, both the zircon and the baddeleyite data of this sample are discordant and exhibit such a scatter that no linear regression through the data is feasible. As the $^{207}Pb/^{206}Pb$ age of the nearly concordant titanite is 2129±2 Ma, and the zircon and baddeleyite data plot near the linear trend determined by A410 (Fig. 8), it seems that both types of diabase sills at Kaakamovaara are of similar age.

A477 Kivimaa. The sample is a reddish, medium-grained albite-rich diabase. The country rock is apparently a mafic volcanic rock of the Runkaus Formation. A small, now exhausted vein-type copper-gold deposit (Rouhunkoski & Isokangas 1974) is associated with pyroxenite within the same sill. Most of the zircon crystals are turbid and reddish brown in color. There are, however, translucent and transparent crystals, too. The zircon crystals are mostly short, even bipyramidic. Other accessory minerals include barite, titanite, and baddeleyite. All analyzed zircon and baddeleyite fractions are discordant (Fig. 9) and many of them are rather rich in common lead. The fractions with lower common lead contents ($^{206}Pb/^{204}Pb$>1000) plot close to a reference line with intercepts at 2215 and 215 Ma. The slightly reversely discordant titanite has a $^{207}Pb/^{206}Pb$ age of 2216±8 Ma.

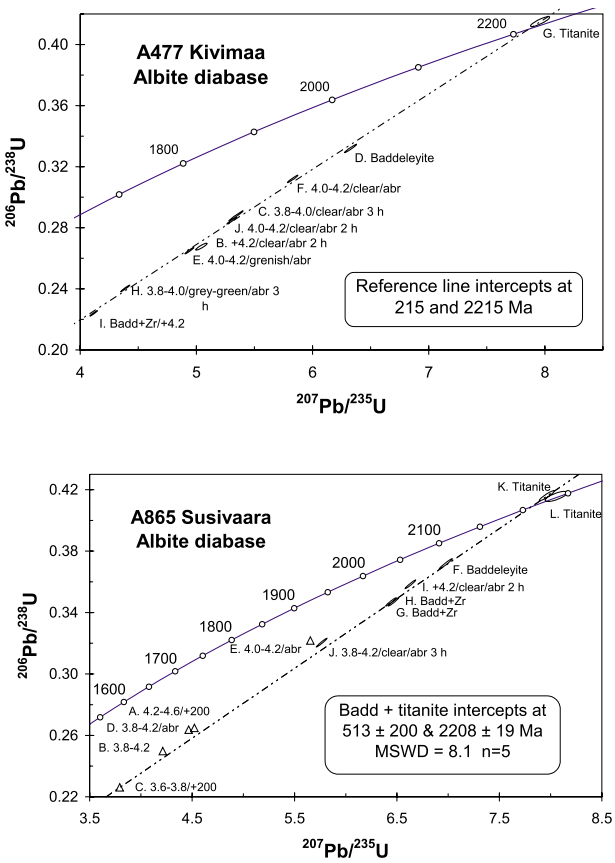

Fig. 9. Concordia diagrams for samples A477 Kivimaa and A865 Susivaara.

A865 Susivaara was taken from a 50 m thick differentiated sill in quartzites and siltstones of the Palokivalo Formation. The zircon is typical for the "albite diabases" and mostly fragmentary in the <75 mesh fractions. Unbroken crystals show simple, short, tetragonal bipyramidic forms. The color is brownish and turbid. Green amphibole inclusions are common in the zircon crystals. A small amount of clear, transparent zircon has also been detected. Baddeleyite is greyish brown and mostly surrounded by zircon. Titanite is common and occurs as brown or light-colored crystals. Five turbid zircon fractions gave an upper intercept of 2089±27 Ma but baddeleyite suggested an age 100 Ma older. Analyses from the clear zircon variety and the baddeleyite exhibit a rather large scatter (MSWD=8.1) from a linear trend with relatively imprecise intercepts at 2208±19 and 513±200 Ma (Fig. 9). The slightly reversely discordant titanites plot close to the trend defined by the zircon and baddeleyite analyses.

A480 Sipojuntti cross-cuts the Archean Pudasjärvi granite gneiss complex. The dyke is differentiated in a manner similar to the albite diabases. The darker varieties contain hornblende surrounded by albitic plagioclase. The light-colored portions consist of abundant albite and epidote and pale green amphibole. Zircons separated from the latter type of rock are euhedral with simple prismatic-pyramidal morphology. The 3.8-4.0 density fraction contains many crystals with a reddish tinge, while those in the 4.0-4.2 fraction are more often glassy and transparent. In the heavier fractions the zircons are either platy or needle-like.

The zircons form two groups according to their density (Fig. 10). The heavy fractions A, B, E and I plot off the main cluster, and possibly reflect the combined effects of a later geological episode (e.g. intrusion of Haaparanta granitoids) and subsequent continuous lead loss by diffusion. The lighter fractions define a linear trend (MSWD=4.3) with intercepts at 2118±14 and 257±190 Ma. It should be noted that the discordia passes within experimental error through the slightly reversely discordant titanite point, the large error of which is caused by a poor lead isotopic composition run.

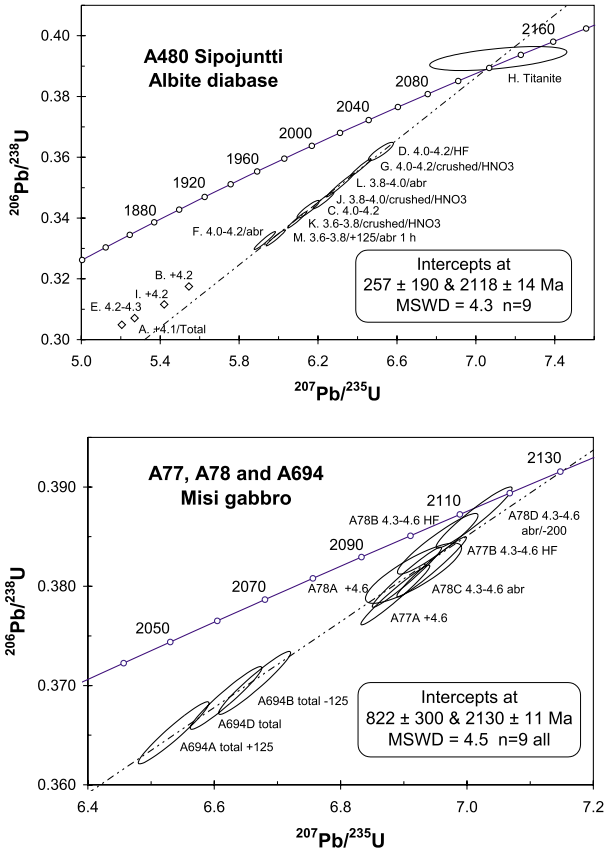

Fig. 10. Concordia diagrams for younger mafic rocks. A480-Sipojuntti cross-cuts the Archean Pudasjärvi complex, and A77, A78 and A694 represent the Misi gabbro in the eastern part of the study area.

A77 and **A78 Lapalionkangas** represent the Misi gabbro in the northeastern corner of the Peräpohja Schist Belt. Both samples contained plenty of clear, faintly brownish zircon. Their grain size has been fairly coarse, as most crystals have been broken during milling. Nevertheless, distinct crystal faces bear evidence of an original prismatic-pyramidal habit. A minority of platy crystals also occurs. **A694 Misi** is a pegmatoid gabbro block found close to the outcrops of A77 and A78. The zircon and titanite contents of the sample are high.

The data from the three samples vary from practically concordant (A78D) to very little discordant (Fig. 10). As the data points for individual samples are closely bunched, linear regression calculations become unreliable. If all analyses are pooled, the intercepts become 2130±11 and 822±300 Ma. However, the somewhat elevated MSWD (4.5) and lower intercept values may indicate different discordance mechanisms for the pegmatoid and even-grained rock types. Thus the $^{207}Pb/^{206}Pb$ age of the abraded fraction A78D, 2117±4 Ma, probably conveys the best estimate for the time of emplacement of the Misi gabbro.

Felsic dykes

The porphyry dykes of the Peräpohja Schist Belt range from 1 to 20 m in thickness. They intrude the Palokivalo quartzite, Tikanmaa tuffite, Rantamaa dolomite, and Martimo mica schist, as well as the layered intrusions. Some porphyry dykes are apparently genetically related to the intrusive rocks of the Haaparanta Suite (Mikkola 1949). In addition to dykes, a felsic, apparently effusive rock, A643 Keinokangas, is discussed in this chapter.

A19 Kapustavuoma. The Kapustavuoma felsic dyke cuts sharply a mafic tuffite of the Tikanmaa Formation. This E-W trending greyish pink dyke is 1-2 m thick, vertical and can be followed in outcrops for two kilometers. The rock is porphyritic with frequent idiomorphic plagioclase and rare quartz phenocrysts in a fine-grained groundmass. The primary plagioclase phenocrysts as well as the plagioclase in the groundmass have been altered to a mixture of albite and sericite. The heaviest (+4.6 g/cm^3) fraction contains water-clear crystals with well developed crystal faces (microgems). In the lighter fractions, turbid gray crystal are dominating, but there also occur dark brown and clear longish varieties, the latter with very sharp crystal face edges. The sample was the first one of its kind to be processed, and thus separating various kinds of zircons was attempted by all possible techniques including pre-treatment with HF. It is thus no wonder that the results for the six analyzed fractions scatter badly (Fig. 11), merely demonstrating a very heterogeneous zircon population.

A512 Elijärvi. Up to 3 m thick albitite dykes occur in the open pit of the Kemi chromite mine and clearly penetrate the ore and other rocks of the Kemi layered intrusion. The dykes are fine-grained with idiomorphic albite phenocrysts (Reino 1973). The zircon grains in the analyzed sample show mixed populations: metamict, rounded grains without any crystal faces, angular long prisms as well as long, somewhat rounded grains. Six U-Pb analyses from various kinds of zircons plot close to a reference line with intercepts at 2600 and 345 Ma (Fig. 11). As the host rock is certainly Proterozoic in age, the zircons of the albitite dykes are inherited from an older source, melted Archean basement rocks.

A643 Keinokangas is a felsic porphyritic rock

with microcline and rare quartz phenocrysts in a fine-grained groundmass. On the top of the porphyry there is a conglomerate with porphyry, siltstone, quartz, and quartzite pebbles. The porphyry crops out in an area of 2x0.5 km^2 and is the only known felsic effusive porphyry within the Peräpohja Schist Belt. The rock lies within mafic volcanic rocks of the Väystäjä Formation, but the stratigraphic setting is somewhat uncertain. The homogeneous zircon population consists of yellowish euhedral crystals with sharp edges, L/B ratio between 3 and 7, and the typical morphology (100). The zircon crystals exhibit many silicate overgrowths (biotite etc.), so it is not easy to achieve a satisfactory age result. Three HF-treated and air abraded fractions are concordant within experimental error, but the individual analyses barely overlap. A linear regression for all six analyses yields intercepts at 2050±8 and 289±150 Ma with an MSWD=5.5 (Fig. 11).

A718 Kallinkangas. A c. 4 m thick porphyry dyke occurs in the road cut on the Kallinkangas ridge. This dyke cuts the bedded quartzite and has fine-grained, chilled contacts. The phenocrysts include idiomorphic albitized plagioclase 1-5 mm in size and quartz. Thinner dark-colored, intermediate dykes cut the porphyry. Chemical analyses of the porphyry and the dark dykes have been published earlier (Perttunen 1991, Table 9). Similar felsic dykes also intrude the quartzite at Kitkiövaara (map sheet 2542 03). The sample contained abundant zircon with variable morphology. The principal type consists of clear prisms which often contain darkish cores. The other noteworthy type is "microgems", which are water-clear and bounded by higher index crystal faces. This and the occasional rounding of the crystal faces makes them outwardly stubby, even rounded in appearance. A good (MSWD=1.07) linear fit into the microgem-trend (fractions G, I, L and N, Fig. 11) defines concordia intercepts at 2679±19 and 1728±59 Ma, of which the former demonstrates an Archean origin and the latter is probably geologically meaningless. The abraded fractions (C-F), where analyses are most likely dominated by the brownish cores, are discordant, but nevertheless exhibit an even more pronounced Archean character than the microgems. The titanite (M) has a very high common lead content, which makes the result extremely dependent on the common lead correction used. Thus its plotting on the microgem-trend is fortuitous. Considering the stratigraphic position of the dyke, it is obvious that the zircons are

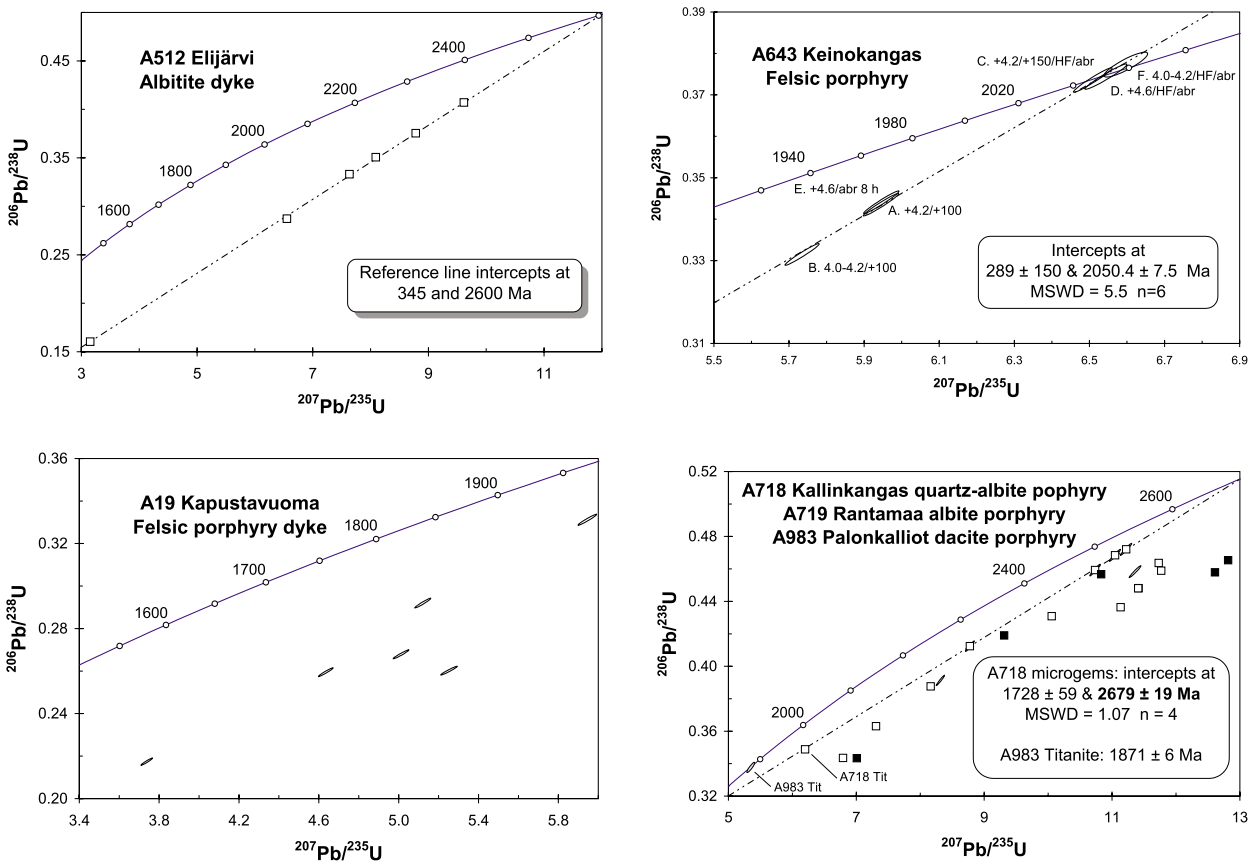

Fig. 11. Concordia diagrams for felsic dyke rocks in the Peräpohja area. Lower right diagram: squares - A718; ellipses with squares - A718 "microgems"; solid squares - A719; ellipses - A983.

xenocrystic Archean material.

A719 Rantamaa. A few 1-3 m thick, vertical porphyry dykes cut the dolomite of the Rantamaa Formation. Some dykes are composite, with no marked difference apart from color between the two types. The phenocrysts consist solely of albitized plagioclase. Petrographically, these dykes resemble A718 Kallinkangas and A19 Kapustavuoma. The zircons in this sample are very similar to those found in the Kallinkangas dyke, and are also of Archean origin (Fig. 11).

A983 Palonkalliot. The porphyry dyke is vertical and about 10 m thick and cuts the mica schist of the Martimo Formation (Härme & Siivola 1966). The phenocrysts include idiomorphic, zoned, fresh plagioclase with complex, multiple twinning as well as biotite and minor hornblende in a fine-grained matrix rich in plagioclase microlites. Dozens of dykes of the same type have been mapped. Petrograhically, these dykes can be correlated with the plutonic rocks of the Haaparanta Suite, as suggested e.g. by Mikkola (1949). The zircon population of this dyke is as mixed as the ones at Kallinkangas and Rantamaa, and only two analyses were carried out. Fraction A is similar to the microgems found in samples A718 and A719 and it plots on the concordia diagram (Fig. 11) close to the microgem-trend of sample A718 Kallinkangas. The other, a relatively dark variety, has a high uranium content (1100 ppm) and is clearly of Archean origin. The titanite from the sample is concordant at 1871±6 Ma.

Plutonic intrusive rocks

A set of plutonic rocks intrude the southwestern part of the Peräpohja Schist Belt. The suite is here called after the original spelling of the locality Haaparanta (instead of Haparanda) Suite (cf. Ödman et al. 1949). The largest of the intrusions are Kaakamo, Nosa and Ruottala (Härme 1949, Mikkola 1949), while smaller intrusions include Liakka, Kuttijänkkä and Poikkilahti. A not outcropping intrusion probably exists in Kuivanuoro in the estuary of Kemijoki. Its presence can be inferred from aeromagnetic low-altitude greytone maps. In addition, there are small, mostly unexposed bodies southwest of the Nosa intrusion.

The contacts of the intrusions are exposed in a few places and demonstrate that the rocks of the Haaparanta Suite are intrusive and younger than the supracrustal rocks of the Peräpohja Schist Belt (Härme 1949). Petrographically, the rocks mostly vary from granodiorite or quartz diorite to gabbro. A typical feature is the presence of plagioclase as idiomorphic, oscillatorily zoned grains (Härme & Siivola 1966). Microcline is common also in intermediate and basic rocks, monzonites and monzogabbros.

There is only one age determination (A368 Rovaniemi) of the plutonic rocks within the large Central Lapland Granite complex near the northern margin of the Peräpohja Schist Belt. That sample is a red, coarse-grained microcline granite (Lauerma 1982).

Samples

The Nosa stock is roundish, about 10 km in diameter. It consist mainly of granodiorite. At the contact zone against the country rock there is a more fine-grained quartz dioritic or dioritic variety (Härme 1949). Two samples, A154 Nosa and A744 Nahonmaa, about 3 km from each other, represent the Nosa intrusion. Petrographically they are coarse-grained, light gray granodiorites with plagioclase, microcline, quartz and biotite as the major minerals.

A154 Nosa contains three different zircon groups: "gem-like" roundish small crystals, long water-clear crystals and big dark grains. The gem-like crystals (fractions C, E and J-O) are transparent and almost colorless. A characteristic feature is the occurrence of numerous high-index crystal faces and the paucity of simple prismatic and pyramidal forms. In contrast, the long clear crystals (A, B, F and H) exhibit a simple morphology and have L/B ratios in excess of 5. Most crystals possess a darker core. The dark crystals (G and I) are rather large (>100 µm), turbid and euhedral, but exhibit also high index crystal faces. The sample also contained dark brown, translucent, anhedral titanite (fraction D).

As can be expected from the variation of the zircons, the analytical results are complex. The dark brown zircon fractions are very discordant, but clearly of Archean age (Fig. 12). The long clear crystals are also discordant and exhibit a large variation in their $^{207}Pb/^{206}Pb$ ages. With the exception of the unabraded fraction C, the gem-like crystals plot along a reference line with intercepts at 1775 and 2605 Ma. The titanite is concordant within experimental error and registers an age of 1879±3 Ma, which is consistent with the time of intrusion of the Haaparanta suite plutonic rocks obtained from other areas. Thus the results from the Nosa granodiorite seem to indicate that it contains a large amount of Archean crustal material remobilized

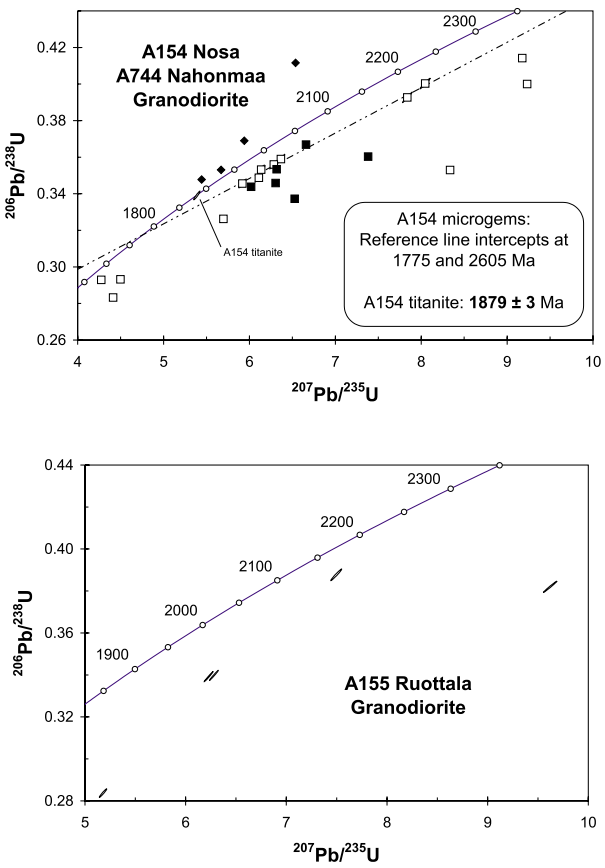

Fig. 12. Concordia diagrams for the felsic intrusions of the Haaparanta suite. Upper diagram: open squares - A154; solid squares - A744 zircons; solid diamonds - A744 titanites.

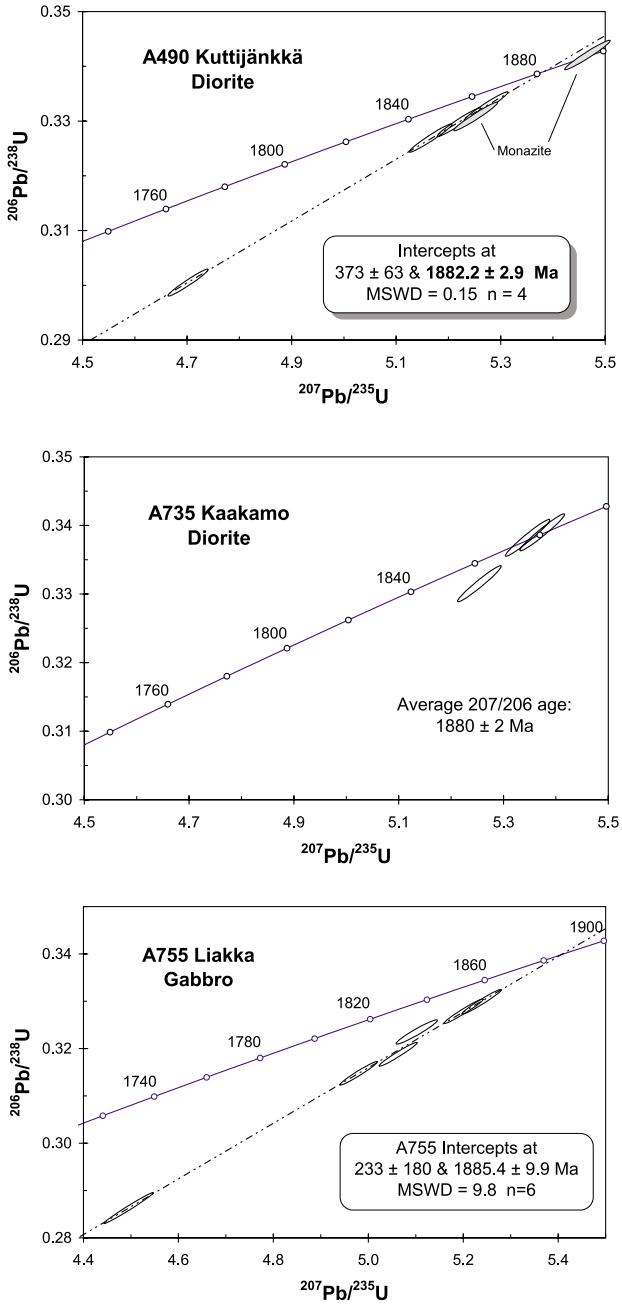

Fig. 13. Concordia diagrams for intermediate-mafic intrusions of the Haaparanta suite.

in Svecofennian times some 1880 Ma ago.

The zircons in sample **A744 Nahonmaa** are similar to those of sample A154 Nosa. Seven zircon and four titanite fractions were analyzed, and the results are consistent with those from A154 Nosa (Fig.12). All titanites are reversely discordant and give imprecise $^{207}Pb/^{206}Pb$ ages close to 1900 Ma. The zircons are discordant and the mixed fraction A indicates an Archean source.

A155 Ruottala. The roundish Ruottala stock, some 4 km in diameter, crops out only in a few places. The contacts against the country rock are unexposed. The shape and size are inferred from low-altitude aeromagnetic maps. The dated sample is a quartz-monzodiorite with plagioclase, hornblende, biotite, microcline and quartz as the main minerals. Most of the zircon crystals in the sample are water-clear and idiomorphic with sharp crystal edges. Many crystals are long 100-prisms without any inclusions. One special fraction (C) consists of small, dark, subhedral zircon crystals. Some zircons are colorless and rounded, reminiscent of detrital grains. Other accessory minerals are titanite and rare anatase. Five analyzed zircon fractions scatter without any linear trend. The short, dark-colored crystals are clearly Archean, and the other four fractions may also contain an older component (Fig. 12).

A490 Kuttijänkkä is from a dioritic portion of a mafic stock. Four zircon fractions form a perfect linear trend (MSWD = 0.15), which defines intercepts at 1882±3 and 373±63 Ma (Fig. 13). This is a typical age for the Haaparanta suite. Inexplicably, the concordant monazite registers an age of 1895±3 Ma, and the older result is confirmed by another, slightly discordant monazite analysis.

A735 Kaakamo. Two of the three zircons are

concordant and even the third lies very close to the concordia curve (Fig. 13). They have an average $^{207}Pb/^{206}Pb$ age of 1880±2 Ma, which is consistent with the results from the zircons of A490.

A755 Liakka. The only large outcrop of the Liakka stock consists of dark olivine gabbro. The main minerals include olivine, plagioclase, ortho- and clinopyroxene, and hornblende. The sample is a pegmatitic granite from a core drilled by Outokumpu Co. The zircons of the sample consist of very long, needle-like crystals (L/B c. 10) with a simple prismatic-pyramidal morphology. Results for three zircon fractions (A-C) have been published previously (Perttunen1987). The six analyses now available define a poor (MSWD=9.8) linear trend with intercepts at 1885±10 and 233±180 Ma (Fig. 13). In spite of its poor quality, the upper intercept age is consistent with the concept that the Liakka gabbro belongs to the Haaparanta suite.

Metasedimentary rocks

Sedimentary rocks in the Peräpohja Schist Belt occur in five sets, separated from each other by volcanic formations. The lowermost unit in the Kivalo Group, the Sompujärvi Formation, consists mainly of conglomerate with minor sandstone and siltstone. The next sedimentary formations contain quartzite and dolomite. The quartzite/dolomite ratio decreases upwards, and three volcanic formations are intercalated in the sequence. The uppermost Martimo Formation consists mainly of turbiditic greywacke with black schist and insignificant rare conglomerate interlayers. Detrital zircons from quartzites and from the turbiditic mica schist have been analyzed, as well as zircons from granitic pebbles from the basal Sompujärvi conglomerate. One sample is from a quartzite layer in a mafic volcanic rock of the Pudasjärvi Basement Complex. All of the analyzed samples contain apparently Archean zircons.

A51 Sompujärvi. The sample is from a granitic clast in a conglomerate which rests unconformably on the 2.45-Ga-old layered intrusions. A few kilometers to the north-east, the foot-wall rock of the conglomerate is a granite which petrographically resembles the pebbles. The zircon analyses exhibit some scatter, with the lighter (d 4.0-4.3 g/cm^3) fractions exhibiting younger $^{207}Pb/^{206}Pb$ ages. The analysis of the heaviest fraction I is suspect in view of its low $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{206}Pb$ ratios. The reference line through the rest of the data has intercepts at 2625 and 460 Ma, and unequivocally demonstrates the Archean provenance of this granite clast (Fig. 14).

A518 Lautakodanoja. Small areas of apparently volcanic amphibolite occur within the Pudasjärvi Gneiss Complex and contain occasional quartzite layers. The sample is from a 50 cm thick quartzite layer. Two zircon fractions, gray translucent and red prisms, were separated from a small quartzite sample. After air abrasion the zircons suggest an Archean age of about 2700 Ma.

A720 Napinmäki. The sample is from a mica schist of the Martimo Formation with distinct graded bedding. The thickness of the beds is 10-15 cm. In addition to the common minerals quartz, biotite, and chlorite the mica schist contains porphyroblasts of cordierite. The mica schist is cut by the plutonic rocks of the Haaparanta Suite dated at 1882±3 Ma (A490 Kuttijänkkä) and 1885±10 Ma (A755 Liakka). The distance from the quartz diorite sample A490 Kuttijänkkä is about 700 m. One zircon fraction is discordant with a $^{207}Pb/^{206}Pb$ age of 2575 Ma.

A783 Kirakka-aapa. The sample lies on the Penikat intrusion, but the contact cannot be seen. The rock was mapped as quartzite, but in thin section it exhibits a granophyric texture resembling the keratophyres roofing the layered intrusions at Koillismaa, some 200 km to the east. The sample contains three zircon populations. Fraction A represents brownish, well-rounded grains without crystal faces, while B and C consist of gray-colored zircon crystallized on A-type grains. According to the diffusion model, the age of fraction A is 2403 Ma suggesting derivation from the layered intrusions. The result 2048 Ma for fractions B and C represent zircon crystallized on older, rounded grains. In addition, there are water-clear, roundish crystals which have not been analyzed.

A1103 Kätkävaara. The sample is from the southern part of the Kätkävaara quartzite ridge. The quartzite overlies amygdaloidal and pyroclastic mafic effusive rocks of the Runkaus Formation and belongs to the Palokivalo Formation. Trough and thin tabular cross beds are common in the reddish to grey colored quartzite. The zircon grains are reddish and well rounded and demonstrate an Archean provenance, as in all of the six analyzed quartzite samples (A1103-A1108) from the Peräpohja area.

A1104 Palokivalo. The sample is from the more than 100 km long Kivalo quartzite ridge. Ripple marks and thick tabular cross beds are common; mud cracks are rare. The reference line shown for the detrital zircons of the Kivalo quartzites is drawn on the basis of this sample (Fig. 14).

A1105 Laitamaa. The Laitamaa quartzite overlies the mafic volcanic rocks of the Jouttiaapa Formation and apparently belongs to the Kvartsimaa Formation. Dolomite and conglomerate with orthoquartzite and dolomite clasts in a dolomitic arkose matrix lie on top of the orthoquartzite. As no rocks of the Tikanmaa Formation, next up in the stratigraphic order, occur near the sample locality, the Laitamaa quartzite sample may belong to the Rantamaa Formation.

A1106 Myllyköngäs. The quartzite sample is from an old mill site at the Vähäjoki River. Ripple marks and rare small-scale cross beds can be detected in the pink, very pure orthoquartzite. No carbonate-bearing rocks have been noticed in the neighborhood. The Myllyköngäs quartzite belongs to the Kvartsimaa Formation.

A1107 Kallinkangas. Kallinkangas is a 2 km long and 500 m wide quartzite ridge. There is a railway cut from the beginning of the 20th century. Primary sedimentary structures such as ripple marks, mud cracks, and cross bedding are common as described by Berghell and Hackman (1928). The sample is from the northern (upper) end of a new road cut through the ridge. The Kallinkangas quartzite is a geographic continuation of the Kivalo ridge and thus apparently belongs to the Palokivalo Formation.

A1108 Lehtosaari. The quartzite is very pure, bluish gray in color. The size of detrital quartz grains is 0.5 - 1.5 mm. The accessory detrital minerals include zircon and tourmaline. The matrix is composed of fine-grained quartz. The quartzite occurs as intercalations in the dolomite of the Rantamaa Formation. The zircons are well rounded, as in all of the quartzite samples of the Peräpohja Schist Belt.

A1438 Korkiavaara is a reddish, banded arkosite occurring in the upper part of the Kivalo Group near Rovaniemi. The content of microcline is fairly high. Chemically, this arkosite is marked by a high Nb content and a high Nb/TiO_2 ratio. These features suggest that the source area included A-type granites (Perttunen et al. 1996). So far, no such rocks are known to exist in the potential source area of the Archean gneiss complex. The amount of zircon is high. The grains are light pink and surprisingly slightly rounded.

Four conventional zircon analyses of sample A1438 Korkiavaara give an upper intercept of 1980+39/-26 Ma. The Pb/Pb age of the most concordant analysis 1962 Ma (A; Fig. 14) should be the minimum and 1975±7 Ma a good average age for the zircon population. SIMS analyses are needed to confirm that interesting age.

The provenance of the material of the quartzites of the Peräpohja Schist Belt is from Archean rocks as all of the analyzed detrital zircons of quartzites are distinctly Archean (Fig. 14). There is no systematic correlation between the age of detrital zircons and the stratigraphic order of quartzites. The apparent ages of zircons in quartzites are older than the zircon ages of granite pebbles in the basal conglomerate of the Sompujärvi Formation (A51), the granitic rocks (A52), and gneisses (A94) of the Pudasjärvi Gneiss Complex. In contrast to the quartzites, the arkosite from Korkiavaara seems to have a definitely younger, Paleoproterozoic provenance.

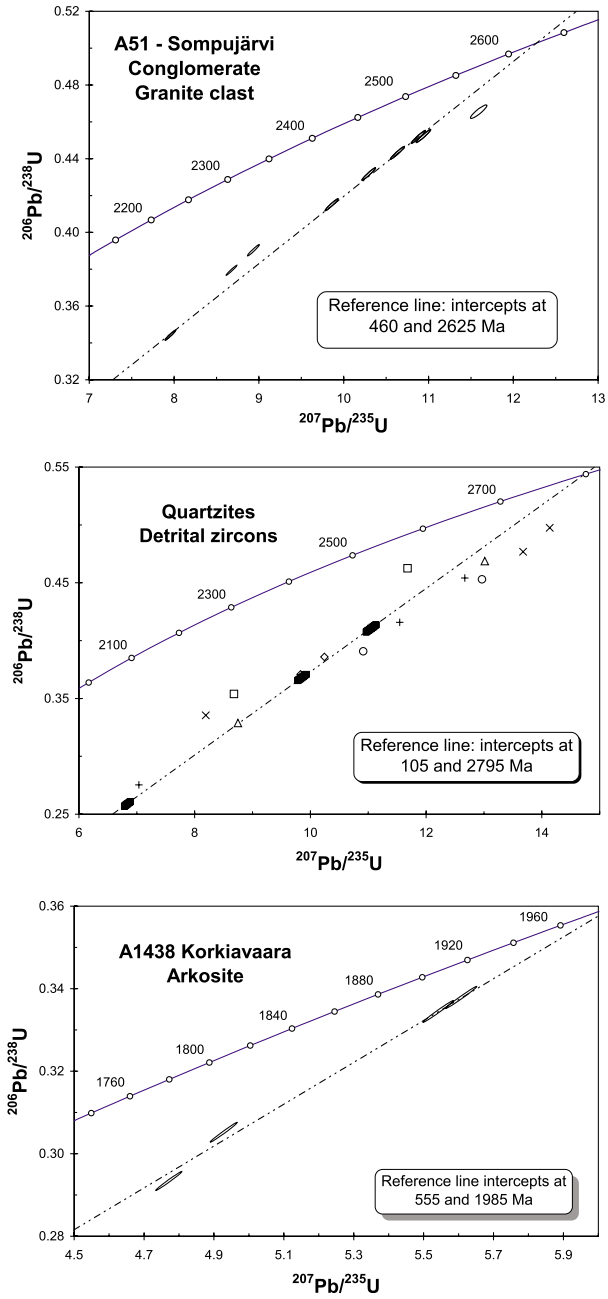

Fig. 14. Concordia diagram for detrital zircons of the metasedimentary rocks of the Peräpohja area. The reference line for the quartzite samples is drawn according the sample A1104 (ellipses). Other quartzite samples are: A518 (squares), A1103 (diamonds), A1105 (circles), A1106 (triangles), A1107 (plusses) and A1008 (crosses).

Table 1. Summary of U-Pb age data from the Peräpohja area.

Sample	Map	Northing	Easting	Location	Rocktype	Mineral	Age	Comment
Archean rocks								
A52	254404	7317.90	2554.45	SOMPUJÄRVI	GRANITE	Zircon	2620±21	Mz 1800±3 Ma;2 borax fusions on discordia
A94	253406	7271.00	2558.00	KUIVANIEMI	GRANODIORITE	Zircon	2667±11	Mz 2664±4 Ma;1 borax fusion near discordia
A439	254111	7299.26	2533.12	ELIJÄRVI	ALBITE ROCK	Zircon		Reference line, 2685 Ma
Layered mafic intrusions								
A603	254401	7311.00	2547.30	ORAVIKANGAS	GABBRO PEGMATOID	Baddeleyite	2428±35	MSWD=7.8; zircons heterogeneous
A662	254111	7299.36	2532.74	ELIJÄRVI	PERIDOTITE PEGMATOID	Zircon	2433±4	
A703	254404	7313.60	2555.50	KILKKA	GABBRO PEGMATOID	Zircon	-	Reference line, 2405 Ma
A704	254404	7313.70	2555.10	KILKKA	GABBRO PEGMATOID	Zircon	-	Reference line, 2405 Ma
A705	254404	7314.10	2550.80	KILKKA	GABBRO PEGMATOID	Zircon	-	Reference line, 2405 Ma
Diabase dykes and sills								
A304	254109	7300.87	2525.13	KALLINKANGAS	DIABASE (ALBITE-)	Zircon	2216±10	7 turbid zircons younger; Tit >2173 Ma
A305	254109	7300.87	2525.13	LAURILA	DIABASE	Baddeleyite	2221±5	zircon very discordant
A407	254401	7311.80	2546.10	ORAVIKANGAS	DIABASE (ALBITE-)	Zircon	-	Reference line, 2225 Ma
A408	254401	7315.20	2549.70	PUUKKOKUMPU	DIABASE (ALBITE)	Baddeleyite	-	Reference line, 2235 Ma
A409	254206	7335.00	2512.50	VIRKKUMAA	DIABASE (ALBITE-)	Zircon	2213±5	on clear zircons; turbid fractions scatter
A410	254203	7338.00	2509.40	KAAKAMOVAARA	DIABASE (ALBITE-)	Zircon	2117±6	
A474	352212	7339.50	3453.00	KONTTIKIVALO	DIABASE (ALBITE -)	Zircon	2210±15	Titanite 2191±4 Ma
A475	254409	7334,15	2567.20	HATTUSELKÄ	DIABASE (ALBITE -)	Baddeleyite	-	Reference line, 2205 Ma
A476	254303	7306.45	2543.65	TORNIVAARA	DIABASE (ALBITE -)	Baddeleyite	-	Reference line, 2240 Ma; Tit c. 2200 Ma
A477	263110	7347.07	2537.35	KIVIMAA	DIABASE (ALBITE -)	Titanite	2216±8	zircon reference line 2215 Ma
A478	254401	7316.00	2549.75	PUUKKOKUMPU	DIABASE (ALBITE-)	Zircon	-	Reference line, 2235 Ma
A480	254302	7292.96	2541.06	SIPOJUNTTI	DIABASE (ALBITE+F67-)	Zircon	2118±17	heterogeneous, imprecise titanite coeval
A592	263111	7350.20	2535.80	LUPPOVAARA	DIABASE (ALBITE-)	Titanite	2217±10	zircons heterogeneous
A593	263111	7350.60	2535.70	LUPPOVAARA	DIABASE (ALBITE -)	Zircon	-	two-point reference line, 2215 Ma
A706	254408	7328.80	2562.70	RUNKAUSVAARA	DIABASE (ALBITE -)	Titanite	2209±10	very discordant zircons
A865	263104	7345.14	2513.90	SUSIVAARA	DIABASE (ALBITE -)	Zircon	-	Reference line, 2215 Ma
A986	254203	7337.94	2509.67	KAAKAMOVAARA	DIABASE (ALBITE -)	Titanite	2129±2	Baddeleyite discordant; zircons heterogeneous
A77	361409	7392.76	3488.07	LAPALIONKANGAS	GABBRO (MISI)	Zircon	2130±11	best 207/206 estimate 2117±4 Ma
A78	361409	7392.85	3488.13	LAPALIONKANGAS	GABBRO (MISI)	Zircon	2130±11	best 207/206 estimate 2117±4 Ma
A694	361409	7392.00	3486.00	MISI	GABBRO PEGMATOID	Zircon	2130±11	best 207/206 estimate 2117±4 Ma

Table 1. continued

Sample	Map	Northing	Easting	Location	Rocktype	Mineral	Age	Comment
Felsic dykes								
A19	254206	7330.88	2513.35	KAPUSTAVUOMA	PORPHYRY	Zircon	-	heterogeneous
A512	254111	7298.95	2532.50	ELIJÄRVI	ALBITITE	Zircon	-	Reference line, 2600 Ma
A643	263105	7355.10	2518.40	KEINOKANGAS	PORPHYRY (QUARTZ -)	Zircon	2050±8	3 concordant fractions
A718	254109	7303.50	2522.85	KALLINKANGAS	PORPHYRY (ALBITE-QUARTZ -)	Zircon	2679±19	on microgems; discordant Tit. on same line
A719	254207	7318.40	2522.26	RANTAMAA	DYKE (ALBITE PHENOCRYSTS)	Zircon	-	xenocrystic, heterogeneous
A983	254204	7314.52	2512.88	PALONKALLIOT	PORPHYRY (DACITE - VEIN)	Titanite	1871±6	zircons xenocrystic
Proterozoic intrusive rocks								
A154	254204	7316.96	2515.70	NOSA	GRANODIORITE	Titanite	1879±3	zircons Archean; ref. line 2610 and 1780 Ma
A155	254109	7306.59	2523.67	RUOTTALA	GRANODIORITE	Zircon	-	xenocrystic, partly Archean
A490	254202	7322.70	2503.95	KUTTIJÄNKKÄ	DIORITE (QUARTZ-)	Zircon	1882±3	concordant monazite 1895±3 Ma
A735	254106	7302.00	2517.00	KAAKAMO	DIORITE	Zircon	1880±2	average 207/206 age
A744	254204	7318.50	2513.30	NAHONMAA	GRANODIORITE	Zircon	-	xenocrystic, partly Archean
A755	254202	7320.30	2507.33	LIAKKA	GABBRO	Zircon	1886±10	
Metasedimentary rocks								
A518	254308	7292.32	2561.80	LAUTAKADONOJA	QUARTZITE	Zircon	-	Archean provenance
A1103	263301	7349.55	2541.85	KÄTKÄVAARA	QUARTZITE	Zircon	-	Archean provenance
A1104	254404	7319.27	2552.00	PALOKIVALO	QUARTZITE	Zircon	-	Archean provenance
A1105	263101	7340.15	2501.10	LAITAMAA	QUARTZITE	Zircon	-	Archean provenance
A1106	363307	7342.44	3563.06	MYLLYKÖNGÄS	QUARTZITE	Zircon	-	Archean provenance
A1107	254109	7303.35	2522.82	KALLINKANGAS	QUARTZITE	Zircon	-	Archean provenance
A1108	254211	7329.20	2534.10	LEHTOSAARI	QUARTZITE	Zircon	-	Archean provenance
A1438	361207	7374.00	3441.30	KORKIAVAARA	ARKOSITE	Zircon	-	Reference line, 1985 Ma

Samples with no zircon

There are some samples which did not contain enough zircon or any other mineral amenable for U-Pb dating. Instead, some other dating methods may have been successful.

A594 Kätkävaara. This sample stems from a differentiated sill in the Kätkävaara Quartzite, Palokivalo Formation. Geographically, the Luppovaara and Kätkävaara hills form a practically continuous ridge, and the sample A592 Luppovaara I apparently represents the same sill.

A636 Keskipenikka is a gabbroic rock in the middle of the Penikat Layered intrusion near the pegmatoid sample A603 Oravikangas. The primary minerals, plagioclase and pyroxene, have been replaced by metamorphic minerals albite, clinozoisite and amphibole.

A637 Kirakkajuppura is a sample from a gabbro pegmatoid near the northeastern end of the Penikat intrusion and sample **A652 Paasivaara** represents a similar rock near the intrusion´s southwestern end.

A663 Elijärvi is a gabbro pegmatoid sample from a drill core in the Kemi layered intrusion.

A754 Liakka. The sample is from a drill core of a pegmatitic granite apparently connected with the intrusive rocks of the Haaparanta Suite.

A854 Rytijänkkä consists of coarse-grained, light-colored diabase. The sample lies in the northwest of the Peräpohja area. Its stratigraphic setting is not quite clear. About 600 m to the south turbiditic greywackes of the Martimo Formation crop out. To the north, there is an area of distinct mafic pillow lavas of the Väystäjä Formation. The amount of zircon in the sample was too small for analyses.

A908 Sivakkajärvi. The coarse-grained, dark colored Sivakkajärvi diabase sill is hosted by an orthoquartzite of the Kvartsimaa Formation. The dyke is vertical and about 20 m thick. This sill can be correlated with sample A1214 Koppakumpu, also with no zircon.

A985 Arppeenlampi. Some amygdaloidal lava flows of the Runkaus Formation crop out near Arppeenlampi, Tervola. Chemically, the lavas of the Runkaus Formation are tholeiitic basalts with relatively evolved compositions. No zircon has been detected. In addition to sample A985, samples A1014 and A999 were analyzed for Pb and Sm-Nd. The most precise Pb/Pb age (A999) of secondary titanite-leucoxene is c. 2250 Ma, which provides the minimum age for the Runkaus volcanics (Huhma et al. 1990). The regression line of all analyzed samples gives an age of 1972±80 Ma, apparently too young for an extrusion age. Whole rock analyses yield an imprecise Sm-Nd age of 2332±180 Ma (Huhma et al. 1990).

A999 Oravikangas. The sample is from a 20-m-thick tholeiitic basalt flow of the Runkaus Formation. The primary minerals include plagioclase and clinopyroxene. The rock was metamorphosed under low greenschist facies. The metamorphic minerals are albite, actinolite, epidote, and chlorite. Isotope results have been discussed in the previous sample A985.

A1009 Jyrövinsa. There are dozens of amygdaloidal lava flows in the Jouttiaapa Formation. Chemically, the lavas of the Jouttiaapa Formation are tholeiitic basalts characterized by depletion of incompatible trace elements. The best fit line through the data yields a Sm-Nd age of 2090±70 Ma (Huhma et al. 1990).

A1012 Sompuvaara. The sample is a pyroxenite

Table 2. Samples with no U-bearing minerals from the Peräpohja area.

Sample	Map	Northing	Easting	Location	Rock type
A594	263301	7349.80	2543.70	Kätkävaara	Albite diabase
A636	254401	7311.00	2547.30	Keskipenikka	Gabbro
A663	254111	7299.79	2533.33	Elijärvi	Gabbro pegmatoid
A754	2542	7321.40	2508.80	Liakka	Pegmatite granite
A854	261311	7351.22	2497.44	Rytijänkkä	Albite diabase
A908	263110	7354.29	2533.14	Sivakkajärvi	Albite diabase
A985	254408	7329.80	2564.50	Arppeenlampi	Basalt
A999	254401	7310.35	2546.05	Oravikangas	Basalt
A1009	263108	7352.90	2529.80	Jyrövinsa	Basalt
A1012	254404	7317.95	2554.18	Sompuvaara	Gabbro (pyroxenite)
A1014	2544	7329.90	2564.60	Runkausvaara	Basalt
A1214	263302	7352.40	2541.14	Koppakumpu	Diabase

low in the stratigraphy of the Penikat Layered Intrusion. It contains well-preserved cumulus orthopyroxene with some plagioclase (An_{50}) as an intercumulus phase. Whole rock, orthopyroxene, and plagioclase define a Sm-Nd isochron age of 2410±64 Ma (Huhma et al. 1990).

A1014 Runkausvaara. This basaltic sample of the Runkaus Formation is discussed in connection of samples A895 and A999.

A1214 Koppakumpu is from a coarse-grained diabase sill tens of meters thick. The host rock of the sill is a tuffite of the Tikanmaa Formation, so it is on a higher stratigraphic level than most of the diabases of the area. The main minerals include plagioclase and secondary amphibole.

DISCUSSION

The Pudasjärvi Basement Complex underlying the Peräpohja Schist Belt is of Archean age. This is confirmed by the analyses of samples A52 and A94 giving clearly Archean upper intercept ages. Also a Sm-Nd isotope analysis from a microcline granite just below the basal Sompujärvi conglomerate shows an ε_{Nd} value of -15.6 (Huhma 1986). On the other hand, the subsequent thermal histories of the Archean rocks may have varied considerably, as concordant monazites in the two samples analyzed during this study are totally different. While the grey banded gneiss A94 contains an Archean monazite with an age of 2662±6 Ma, the monazite in the granitic rock A52 near the contact of the Penikat layered intrusion is Proterozoic at 1799±3 Ma. However, this age, typical of e.g. the Revsund type granites occurring abundantly in central Sweden, is not recorded in other samples from the Peräpohja area.

The zircon data from sample A662 Elijärvi and the results for the zircon-baddeleyite composites from sample A603 Oravikangas indicate an intrusion age of c. 2430 Ma for the Kemi and Penikat layered intrusions. Previous whole rock analyses are consistent with this result, as a Pb-Pb age of 2.44±0.16 Ga has been reported for the Kemi intrusion (Manhes et al. 1980) and a Sm-Nd isochron age of 2410±64 Ma has been obtained from the Penikat intrusion (Huhma et al. 1990). However, in most cases zircons do not give precise ages.

Apparently, zirconium minerals formed originally about 2.43 Ga ago, and an overgrowth of original primary baddeleyite by secondary zircon is suggested by the observation from sample A603 Oravikangas from the Penikat intrusion, which shows a very complex zircon history. From the geological record it is obvious that the layered mafic intrusions were affected by many later events. Such were the intrusion of the diabase sills, the geochemical changes triggered by the regional albitization and the intrusion of the Haaparanta suite granitoids. Consequently, the results of the bulk zircon analyses are mixtures reflecting the effects of several geological events (see also Mertanen et al. 1989 and Hanski et al. 2001, *this volume*).

As demonstrated by the Sompujärvi conglomerate (A51), the supracrustal rocks of the Peräpohja Schist Belt lie unconformably on the basement gneisses and the layered intrusions, so the maximum age of the Schist Belt is about 2.43 Ga. Although the Runkaus volcanics in its lower part are clearly older than the overlying (Jatulian) Kivalo quartzites, there is, in contrast to central Lapland (cf. Manninen et al. 2001, *this volume*), no unequivocal evidence on the existence of volcanic rocks roughly coeval with the layered gabbro intrusions (Huhma et al. 1990). The minimum age for the whole Peräpohja Schist Belt, >1890 Ma, is obtained from the intersecting rocks of the Haaparanta Suite which by field evidence are younger than all of the supracrustals. Some further fine-tuning can be obtained from the various dykes and sills intruding the supracrustal sequence.

Most U-Pb data from the diabases in the Peräpohja area suggest emplacement ages near 2.2 Ga. By their age as well as by their chemical and mineralogical compositions, they can with certainty be correlated with the Haaskalehto type diabases in central Lapland (Rastas et al. 2001, *this volume*, Räsänen & Huhma 2001, *this volume*). This group of diabases occurs also in the Kuusamo area (Silvennoinen 1991), and in Karelia (gabbro-wehrlite association, Hanski et al. 1985, Vuollo et al. 1992). Some of the diabases (A410 and A986 Kaakamovaara, A480 Sipojuntti) may be about 100 Ma younger. The gabbros of the Misi area in the northeastern corner of the Peräpohja Schist Belt give ages near 2.15 Ga (A694 Misi; Patchett et al. 1981).

An interesting detail of the 2.2 Ga age group diabase sills is the frequent, if not ubiquitous occurrence of baddeleyite. This suggests that the magmas were originally undersaturated in silica, and zircon was not the primary zirconium phase to crystallize. As most of the zircons are turbid, and zircon overgrowths on baddeleyite are frequent, much of the

zircon in these rocks may well represent a secondary phase. The formation of secondary zircon may have been associated with the regional spilitization, which possibly has occurred at several stages, and, as much fluid was involved in the process, formed several generations of hydrozircons, i.e. those turbid zircons yielding apparently youngish and discordant U-Pb age data.

Diabases, dated to be about 2.05 Ga old, are common in Central Lapland (Rastas et al. 2001, *this volume*). The same age group possibly exists also in the Peräpohja area and could be represented by samples A908 Sivakkajärvi and A1214 Koppakumpu, which differ petrographically from and lie higher up in the stratigraphy than the coarse-grained (Haaskalehto type) albite diabases. Unfortunately, attempts at their dating have so far been unsuccessful as no zircon or other U-bearing minerals have been found.

Direct geochronological data on the mafic volcanic rocks of the Peräpohja Schist Belt are fragmentary. Radiometric dating of all mafic volcanic formations in the area has proven difficult due to the lack of suitable primary minerals, and thus evidence on their ages of emplacement are at best circumstantial. Whole rock and mineral Pb-Pb and Sm-Nd analyses have been carried out on three samples (A895, A999 and A1014) of the Runkaus Volcanic Formation. The most precise Pb/Pb age of secondary titanite-leucoxene of these samples is c. 2250 Ma, which provides the minimum age for the Runkaus volcanics (Huhma et al. 1990). The regression line of all analyzed samples gives 1972±80 Ma, an age apparently too young. Scattered whole rock data yield a Sm-Nd age of 2330±180 Ma (op. cit.). In any case, these volcanic rocks are younger than the layered intrusions, and their age is thus <2.45 Ga. The Sm-Nd age for the tholeiitic basalts of the Jouttiaapa Volcanic Formation is 2090±70 Ma (op. cit.). These volcanics have no chemical counterparts in northern Finland - they are depleted in LREE and have an ε_{Nd} of +4. This result is suggestive of derivation from a depleted mantle source and may imply a period of deep crustal rifting in the Peräpohja area at their time of extrusion.

The zircons of the felsic porphyry A643 Keinokangas give an age of 2050±7 Ma. The porphyry lies within the pillow lavas of the Väystäjä Formation in the Paakkola Group, but its stratigraphic position, unfortunately, is not quite clear.

The zircons in a thin quartzite layer in the apparently Archean volcanic rocks of the basement complex (A518 Lautakodanoja) show a definitely Archean (>2700 Ma) provenance. The quartzites of the Palokivalo, Kvartsimaa and Rantamaa formations (samples A1103 - A1108) are cut by the differentiated diabases, so the minimum age of their sedimentation is >2.2 Ga. U-Pb results on their detrital zircons do not give much information on the age of the source rocks of the sediments of the Peräpohja Schist Belt, as all of the analyses suggest an Archean provenance. Statistical analyses of the zircons of these rocks by means of e.g. an ICP-MS with a laser ablation source could yield significant information on the relative amounts of Archean continental crust of various ages available in their source areas.

The age of the sedimentation of the Martimo Formation is not precisely settled. The minimum age established by the Haaparanta Suite intrusions is <1900 Ma and the $^{207}Pb/^{206}Pb$ ratios of discordant zircons give a late Archean result at 2575 Ma. A mica schist sample from Taivalkoski registers an ε_{Nd} value of -3.0 (Huhma 1987). These rocks have been correlated with the Lower Kaleva assemblage in Eastern Finland.

Recent analyses of the arkosite sample A1438 Korkiavaara from the top of the Kivalo Group are interesting, because the zircons show that the sedimentation did not occur before 1980 Ma ago. As the zircon population is relatively homogeneous, and most zircons are not intensively abraded, it is easy to accept an age of less than 2000 Ma for the provenance. Chemically, the arkosite corresponds to an A-type granite (Perttunen et al. 1996), but no such rock or any other igneous rock with an age between 1.9 and 2.0 Ga is known to exist in northern Fennoscandia. The age of the mica schists lying stratigraphically above the Korkiavaara Formation must be between 1980 and 1890 Ma. This seemingly contradicts the result from A643 Keinokangas, which indicates an age of c. 2.05 Ga for the Väystäjä Formation, considered to be part of the Paakkola Group. In the Kainuu area, eastern Finland, similar mica schists have been divided into Lower and Upper Kaleva Formations (Kontinen 1997). The age limit between them is established by the age of the Jormua Ophiolite Complex at 1.95 Ga. In the Peräpohja area, solving the age problems within the Paakkola Group must await future SIMS analyses.

Three intermediate-mafic samples suggest ages of 1880-1890 Ma for the intrusion of the plutonic rocks of the Haaparanta Suite (A490 Kuttijänkkä, A735 Kaakamo and A755 Liakka). According to four zircon fractions, the age of A490 Kuttijänkkä is 1882±3 Ma. The analyses are little discordant and the scatter is small. However, the result of the monazite analysis (1898 Ma) from the same rock is problematic. Of course, the monazite may be partly xenocrystic and derive from the frequent mica schist inclusions encountered in the rock, but it is hard to conceive why this monazite would not have been thermally reset by

the dioritic magma. The results on sample A755 give an age of 1887±9 Ma for the Liakka gabbro. Two concordant analyses of the Kaakamo diorite (A735) barely overlap within experimental error and the third is somewhat discordant, but the average $^{207}Pb/^{206}Pb$ age of 1880±2 Ma is consistent with the apparent coevality of these rocks.

Two other analyzed felsic intrusions, Nosa and Ruottala, belong according to field evidence to the Haaparanta Suite. All three analyzed samples (A154 Nosa, A744 Nahonmaa and A155 Ruottala) contain mostly xenocrystic, Archean zircons which have survived melting and crystallizing processes. Also the granitic sample A754 from the Liakka intrusion contains Archean zircons while zircons in the gabbro A755 of the same intrusion are Proterozoic (Perttunen 1987). Similarly, the felsic porphyry dykes (A512 Elijärvi, A718 Kallinkangas and A719 Rantamaa) contain Archean zircons, though the host rocks - well-dated layered intrusions and overlying sedimentary rocks - are certainly Proterozoic. These results clearly indicate that the Archean basement was remobilized at least on two occasions, locally at the time of the emplacement of the layered mafic intrusions and on a regional scale during the culmination of the Svecofennian orogeny.

The intrusion of the plutonic rocks of the Haaparanta Suite in the Peräpohja area occurred near 1885 Ma. This finding is in concert with the ages obtained in Kolari-Muonio area (Väänänen & Lehtonen 2001, *this volume*) and also in northern Sweden where the Purissakero granodiorite registers an age of 1873±73 Ma (Skiöld 1981). Similar ages have also been obtained from the Degerberg granitoids in the Kalix area (Wikström et al. 1996). A noteworthy feature of these data is that one of the samples definitely contains inherited zircons, while the other one displays a very large MSWD of 20, also suggestive of heterogeneities within its zircon population. That practically all felsic rocks of the Haaparanta suite in the Peräpohja area (A154 Nosa, A744 Nahonmaa and A155 Ruottala) contain inherited zircons suggests a westerly rise in temperatures in the magma chambers where the rock melts were formed. A similar indication is also given by the Töre granodiorite northeast of Luleå, which contains a heterogeneous zircon population (Wikström & Persson 1997).The large influx of pre-existing Archean continental crust into at least the felsic members of the Haaparanta suite is also demonstrated by their markedly negative (-4 to -7) $\varepsilon_{Nd}(1900)$ values (Öhlander et al. 1993).

The age of the granitoids of the Central Lapland Granite Complex delimiting the Peräpohja Schist Belt in the north is based only on a few samples. The microcline granite A368 Rovaniemi contains both zircon and a coeval concordant titanite dated at 1770±8 Ma (Lauerma 1982) with an ε_{Nd} of -6.2 (Huhma 1986). The granite A145 Molkoköngäs is 1843±28 Ma old and has an ε_{Nd} of -6.4 (Huhma 1986). The granite A642 Pietariselkä at Kemijärvi contains both zircon and monazite with an age of c. 1800 Ma (Lauerma 1982).

CONCLUSIONS

The results presented in this paper and the foregoing discussion warrant the following conclusions:

1) The emplacement of the mafic layered intrusions near the basement contact of the rocks of the Peräpohja Schist Belt occurred c. 2.43 Ga ago, but much of the data show signs of later geological events.

2) The analyses reveal a wide-spread 2.2-Ga-old magmatism when concordant, differentiated mafic sills intruded the quartzites of the Peräpohja area. These sills can be correlated with similar, coeval Haaskalehto-type rocks in Central Lapland and the gabbro-wehrlite association of eastern Finland.

3) The existence of the 2.05-Ga-old diabases, common in Central Finnish Lapland, has not been confirmed in the study area. Instead, there may be a set of about 2.1 Ga old diabases.

4) All analyzed felsic dykes contain xenocrystic, Archean zircons and their emplacement age is unknown. Also many samples from the plutons of the Haaparanta Suite contain inherited zircons, while some zircon as well as titanite analyses confirm the age of about 1.89 Ga, common elsewhere in Western Lapland and Northern Sweden. This suggests that Archean crustal rocks were remelted twice in Paleoproterozoic times, first locally the during the emplacement of the layered mafic intrusions, and then on a regional scale when the magmas of the Haaparanta Suite were formed.

5) The detrital zircons separated from all of the Kivalo Group (Jatulian) quartzite samples show, irrespective of their stratigraphic setting, an Archean provenance, as does a sample from a quartzite interbed in an apparently Archean greenstone formation in the basement area.

6) A sample from the uppermost Martimo Formation gives a Pb/Pb age result >2.58 Ga. Results from an arkosite, at a slightly lower stratigraphic level give an apparent age of 1985 Ma for the provenance of the

material, so the existence of the "Upper Kaleva" is plausible also in the Peräpohja area.

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Appendix I. U-Pb mineral age data from Archean rocks of the Peräpohja area.

Sample information	Weight	U	Pb	206Pb/204Pb	208Pb/206Pb							Rho**	APPARENT AGES (Ma)		
Analyzed mineral fraction	mg	ppm		measured	radiogenic	206Pb/238U	2SE%	207Pb/235U	2SE%	207Pb/206Pb	2SE%		206Pb/238U	207Pb/235U	207Pb/206Pb
A52-Sompujärvi granite															
A. Zr total, borax	469.7	658.4	276.3	871	0.1533	0.3515	0.83	8.020	0.87	0.1655	0.16	0.97	1941	2233	2512
B. Zr total, borax	474.0	831.4	339.5	1307	0.1469	0.3494	0.74	8.042	0.79	0.1668	0.24	0.95	1932	2235	2526
C. Monazite	3.5	1464.8	4691.5	6076	10.5692	0.3162	0.65	4.799	0.65	0.1101	0.15	0.97	1771	1784	1800
D. Zr +4.3/abr	6.3	601.5	283.3	2047	0.1313	0.4109	0.65	9.721	0.66	0.1716	0.15	0.97	2218	2408	2573
E. Zr 4.2-4.3/abr	7.6	640.6	313.6	1240	0.1387	0.4186	0.65	9.901	0.65	0.1715	0.15	0.97	2254	2425	2572
F. Zr +4.0/long	1.6	732.6	311.4	1039	0.1333	0.3638	0.65	8.406	0.65	0.1676	0.29	0.90	2000	2245	2533
G. Zr 3.8-4.0/abr	7.1	884.4	344.5	1141	0.1312	0.3349	0.65	7.615	0.65	0.1649	0.15	0.97	1862	2186	2506
A94-Kuivaniemi granodiorite															
A. Zr total, borax	357.5	594.6	276.6	775	0.8050	0.2738	0.93	6.883	1.17	0.1823	0.63	0.84	1569	2096	2674
B. Zr +4.6	18.7	399.4	237.3	1185	0.1817	0.4874	0.65	12.241	0.65	0.1822	0.15	0.97	2559	2623	2572
C. Zr 4.0-4.2	16.0	626.5	188.4	644	0.1825	0.2364	0.65	5.947	0.65	0.1825	0.15	0.97	1267	1968	2675
D. Zr -4.0	5.5	689.2	154.2	335	0.1834	0.1542	0.65	3.899	0.65	0.1834	0.15	0.97	924	1613	2684
E. Monazite	2.8	585.0	5031.0	1454	18.3921	0.5061	0.79	12.636	0.80	0.1811	0.23	0.96	2639	2652	2663
F. Zr +4.6/abr	6.9	384.2	200.9	2817	0.1252	0.4584	0.70	11.360	0.76	0.1798	0.32	0.91	2432	2553	2650
G. Zr 4.0-4.2/abr	13.2	602.5	251.7	876	0.3358	0.3051	0.66	7.655	0.67	0.1820	0.15	0.97	1716	2191	2671
A439-Elijärvi albite rock															
A. Zr total, borax	554.2	275.1	94.2	278	0.1771	0.2499	0.65	6.195	0.65	0.1798	0.29	0.90	1437	2003	2651
B. Zr +4.2/long	14.6	285.8	98.1	246	0.1553	0.2495	0.65	6.192	0.65	0.1800	0.16	0.97	1436	2003	2652
C. Zr +4.2/anhedral	21.2	247.2	92.1	307	0.2170	0.2698	0.65	6.677	0.65	0.1795	0.15	0.97	1539	2069	2648
D. Zr 4.0-4.2	7.6	870.0	208.5	70	0.1538	0.1149	0.65	2.696	0.65	0.1703	0.25	0.93	700	1327	2560
E. Zr +4.2	19.9	531.6	131.8	141	0.1192	0.1717	0.65	4.153	0.65	0.1754	0.15	0.97	1021	1664	2609
F. Zr +4.4/abr	11.3	181.2	84.5	430	0.1839	0.3577	0.67	8.946	0.68	0.1814	0.32	0.89	1971	2332	2665

*) Isotopic ratios corrected for fractionation, blank and age related common lead.
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = standard error in the mean

Appendix II. U-Pb mineral age data from layered gabbro intrusions.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A603-Oravikangas gabbro pegmatoid															
A. 4.0-4.2/124-200	14.0	769.4	334.7	1613	0.3789	0.3318	0.65	6.386	0.65	0.1396	0.20	0.95	1847	2030	2222
B. 3.6-3.8/-200	12.2	1168.2	313.3	770	0.4230	0.1968	0.65	3.341	0.65	0.1231	0.20	0.95	1158	1490	2001
C. 3.8-4.0/125-200	16.7	942.7	340.5	695	0.4418	0.2591	0.65	4.644	0.65	0.1300	0.20	0.95	1485	1757	2098
D. 3.8-4.0/-200	17.5	948.5	353.6	884	0.4074	0.2738	0.65	4.974	0.65	0.1318	0.15	0.97	1560	1814	2121
E. 4.0-4.2/+125	12.5	791.5	344.3	1363	0.4084	0.3225	0.65	6.224	0.65	0.1400	0.15	0.97	1802	2007	2226
F. 4.0-4.2/-200	23.4	781.2	347.8	950	0.3719	0.3337	0.65	6.378	0.65	0.1386	0.15	0.97	1856	2029	2210
G. -4.2	10.1	796.7	335.4	701	0.3876	0.3102	0.65	5.835	0.65	0.1365	0.15	0.97	1741	1951	2182
H. 4.0-3.8/+125	28.8	993.5	355.5	621	0.4763	0.2490	0.65	4.464	0.65	0.1300	0.15	0.97	1433	1724	2098
I. +4.2	4.0	519.9	272.4	1924	0.3393	0.4034	0.59	8.371	0.59	0.1505	0.15	0.97	2184	2272	2351
J. Badd+Zr	4.2	708.4	298.6	1047	0.1017	0.3821	0.59	7.915	0.62	0.1502	0.16	0.97	2086	2221	2348
K. Baddeleyite	3.1	686.9	298.2	1265	0.0815	0.4012	0.59	8.390	0.61	0.1517	0.15	0.97	2174	2274	2365
L. Baddeleyite	6.5	702.9	316.6	1942	0.0513	0.4283	0.58	9.193	0.58	0.1557	0.15	0.97	2297	2357	2409
M. Badd+Zr	6.1	701.7	299.2	2211	0.0713	0.4009	0.58	8.363	0.58	0.1513	0.15	0.97	2173	2271	2360
N. Zr+Badd	4.1	719.4	272.7	1678	0.1200	0.3429	0.60	6.882	0.60	0.1456	0.15	0.97	1900	2096	2294
O. 4.0-4.2/abr 3h	4.4	783.3	379.1	913	0.4127	0.3508	0.69	6.805	0.62	0.1407	0.18	0.96	1938	2086	2236
A662-Elijärvi gabbro pegmatoid															
A. Zr +4.2/+125	30.6	37.1	22.7	743	0.3490	0.4500	0.65	9.778	0.80	0.1576	0.36	0.90	2395	2414	2430
B. Zr +4.2/-125	54.0	36.4	21.7	863	0.3379	0.4430	0.65	9.593	0.65	0.1571	0.24	0.93	2363	2396	2424
C. Zr 3.8-4.2	17.2	37.0	22.4	603	0.3530	0.4403	0.65	9.593	1.00	0.1580	0.70	0.72	2351	2396	2434
A703-Kilkka gabbro pegmatoid															
A. +4.2/+100	31.6	512.8		470		0.3980	0.65	8.369	0.65	0.1525	0.15	0.97	2159	2271	2374
B. +4.2/100-200	31.9	520.2		542		0.3704	0.65	7.664	0.65	0.1501	0.15	0.97	2031	2192	2374
C. 4.0-4.2/+125	15.9	713.1		433		0.3594	0.65	7.388	0.65	0.1491	0.15	0.97	1979	2159	2335
D. 4.0-4.2/+150		744.9		565		0.3350	0.65	6.738	0.65	0.1459	0.15	0.97	1862	2077	2298
E. 3.8-4.0/+150		945.7		444		0.2797	0.65	5.403	0.65	0.1401	0.15	0.97	1589	1885	2228
F. 3.6-3.8/+100		1209.7		248		0.2007	0.65	3.574	0.65	0.1292	0.15	0.97	1178	1543	2086
G. 4.2/+200/abr 4 h		497.6		717		0.3905	0.65	8.078	0.65	0.1501	0.15	0.97	2124	2239	2346
A704-Kilkka gabbro pegmatoid															
A. +3.8		630.1		4415		0.4083	0.65	8.387	0.65	0.1490	0.15	0.97	2207	2273	2334
A705-Kilkka gabbo pegmatoid															
A. +3.8		488.5		7764		0.4220	0.65	8.981	0.65	0.1544	0.15	0.97	2269	2339	2395

*) Isotopic ratios corrected for fractionation, blank and age related common lead.
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = standard error in the mean

Appendix III. U-Pb mineral age data from diabase dykes and sills.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A304-Kallinkangas albite diabase															
A . Zr +4.6/abr	10.4	426.7	221.6	105.7	0.3547	0.2726	0.91	5.028	1.00	0.1338	0.49	0.87	1553	1824	2148
B. Zr 4.2-4.6/abr	4.4	338.5	227.1	123.3	1.0515	0.2788	2.00	5.224	11.00	0.1359	9.84	0.64	1585	1856	2175
C. Zr 4.0-4.2/abr	12.2	401.7	261.8	108.3	1.2523	0.2453	1.06	4.168	1.60	0.1232	1.30	0.59	1414	1667	2003
D. Zr 4.0-4.2/+100/abr	12.2	372.4	250.0	101.2	1.4119	0.2361	0.71	4.169	3.50	0.1213	3.04	0.71	1366	1623	1976
E. Zr 4.0-4.2/abr	15.3	410.2	237.7	138.5	1.1988	0.2354	0.65	3.997	0.70	0.1232	0.40	0.83	1362	1633	2003
F. Zr 4.0-4.2/abr	6.1	418.5	249.7	131.3	1.2088	0.2387	0.78	4.135	0.90	0.1257	0.49	0.84	1379	1661	2039
G. Zr 4.0-4.2/abr	15.7	397.5	261.7	114.7	1.3024	0.2473	0.70	4.212	0.80	0.1236	0.36	0.89	1424	1676	2008
H. Zr 3.8-4.0/abr	13.9	723.2	284.3	107.8	0.9226	0.1660	0.65	2.664	0.70	0.1164	0.42	0.81	990	1318	1901
I. Zr 3.8-4.0/+100/abr	12.7	747.6	302.1	113.1	0.9506	0.1713	0.71	2.738	0.76	0.1160	0.25	0.94	1019	1338	1894
J. Titanite	31.5	24.3	30.5	123.3	1.9687	0.3914	0.65	7.323	2.30	0.1357	1.90	0.70	2129	2151	2173
K. Baddeleyite	1.3	590.3	207.6	241.3	0.4010	0.2420	0.65	4.435	0.72	0.1329	0.30	0.91	1396	1718	2137
L. Zr +4.6/+200	2.1	448.2	174.0	816	0.0790	0.3490	0.65	6.618	0.70	0.1375	0.20	0.96	1929	2061	2196
M. Zr +4.6/-200	4.3	487.1	170.7	693	0.1347	0.3002	0.65	5.622	0.65	0.1358	0.15	0.97	1692	1919	2174
N. Zr +4.6/+200/abr	4.5	572.5	192.5	454	0.2396	0.2569	0.65	4.737	0.65	0.1338	0.15	0.97	1473	1773	2148
A305-Laurila albite diabase															
A. Zr total borax	122.6	615.4	286.7	106.4	1.0626	0.1887	0.85	3.081	2.10	0.1184	1.40	0.89	1114	1427	1933
B. Baddeleyite	4.6	454.5	173.3	606	0.0988	0.3317	0.65	6.246	0.65	0.1366	0.15	0.97	1846	2010	2184
C. Baddeleyite	6.3	426.8	177.5	520	0.0807	0.3590	0.65	6.822	0.65	0.1378	0.15	0.97	1977	2088	2200
D. Baddeleyite + zircon	4.0	548.2	219.1	323	0.2477	0.2914	0.65	5.410	0.65	0.1347	0.15	0.97	1648	1886	2160
A407-Oravikangas albite diabase															
A. +4.0/EM/+200	11.2	734.3	260.1	443	0.5071	0.2297	0.65	4.174	0.67	0.1318	0.16	0.97	1332	1668	2122
B. +4.0/M/+200	9.8	783.6	257.5	351	0.4811	0.2116	0.65	3.792	0.65	0.1300	0.15	0.97	1237	1596	2097
C. 3.8-4.0/green	10.1	918.4	265.2	217		0.1677	0.65	2.901	0.65	0.1254	0.15	0.97	999	1382	2034
D. green, borax	142.1	1189.1	231.7	213	0.6122	0.1114	0.72	1.755	1.00	0.1142	0.70	0.71	681	1028	1867
E. +100/crushed/borax	152.1	1289.3	225.3	280	0.6388	0.1029	0.70	1.571	1.00	0.1108	0.70	0.71	631	959	1812
F. 100-125/red/borax	115.9	1476.3	277.5	313	0.8433	0.1021	0.65	1.452	1.00	0.1032	0.70	0.72	626	910	1682
G. Titanite/borax	1995.6	17.2	18.3	81.1	3.2576	0.2301	0.82	4.282	2.00	0.1350	1.50	0.74	1334	1689	2163
I. Titanite/abr 15 m	24.8	50.3	43.2	66.3	2.8833	0.1879	0.65	3.358	2.00	0.1296	1.70	0.59	1110	1494	2092
J. 3.7-4.2/abr	12.4	1075.0	257.9	545	0.4455	0.1646	0.87	2.788	0.87	0.1229	0.16	0.98	982	1352	1998
K. Baddeleyite+zircon	2.3	424.8	154.4	1117	0.0774	0.3330	0.65	6.268	0.65	0.1365	0.15	0.97	1852	2014	2184
L. Baddeleyite+zircon	4.5	838.2	230.6	744	0.2007	0.2251	0.65	4.068	0.70	0.1311	0.23	0.94	1308	1647	2112
M. +4.0/abr 3 h	5.8	622.1	245.5	555	0.4576	0.2677	0.65	4.953	0.70	0.1342	0.21	0.95	1528	1811	2153

Appendix III continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A408-Puukkokumpu albite diabase															
A.. Titanite	1025.7	20.9	14.8	113	0.4525	0.3716	0.65	7.028	1.50	0.1372	1.00	0.86	2036	2114	2192
B. 100-125	217.3	830.4	274.9	241	0.6384	0.1887	0.66	2.982	1.00	0.1146	0.59	0.82	1114	1403	1873
C. 150-200	125.4	775.4	227.8	353	0.5445	0.1856	0.65	2.955	0.90	0.1154	0.49	0.85	1097	1396	1886
D. 4.-4.2/turbid	12.4	552.6	251.3	235	0.8116	0.2368	1.06	3.802	1.06	0.1164	0.25	0.97	1370	1593	1902
E. Zr+Badd +4.3	6.6	908.5	269.5	648	0.2629	0.2308	0.69	4.024	0.69	0.1264	0.15	0.96	1338	1639	2049
F. Badd+Zr 4.1-4.2	2.5	851.9	304.4	1353	0.0758	0.3301	0.65	6.181	0.70	0.1358	0.28	0.92	1838	2001	2174
G. Zr+Badd 4.1-4.2/abr 1 h	8.0	943.8	382.3	885	0.1643	0.2538	0.65	4.573	0.65	0.1307	0.15	0.97	1457	1744	2107
H. +4.1/clear/abr 20 m	1.3	887.2	274.8	751	0.3314	0.2334	0.65	4.068	0.70	0.1264	0.26	0.93	1352	1647	2049
I. +4.2/abr 20 m	1.4	1006.8	285.1	984	0.2065	0.2353	0.65	4.180	0.70	0.1288	0.24	0.94	1362	1670	2082
A409-Virkkumaa albite diabase															
A. +100/clear	38.8	1054.5	469.5	1299	0.4747	0.3114	0.65	5.800	0.90	0.1351	0.54	0.73	1747	1946	2165
B. 100-125/clear	31.4	1073.3	436.7	1063	0.4521	0.2868	0.65	5.292	0.75	0.1338	0.35	0.87	1625	1867	2148
C. 100-125/borax	105.1	693.5	238.9	428	0.5883	0.2177	0.65	3.692	1.00	0.1230	0.70	0.72	1269	1569	2000
D. +100/turbid/borax	133.0	2186.0	580.0	331	0.6996	0.1527	0.65	2.528	0.75	0.1201	0.27	0.94	915	1280	1957
E. 100-150/M	46.3	2272.6	598.6	273	0.7402	0.1446	0.65	2.367	0.65	0.1187	0.15	0.96	870	1232	1936
F. 3.8-4.0/-200/abr	8.6	1224.9	740.1	123	0.6343	0.2954	0.68	5.271	0.90	0.1294	0.62	0.73	1668	1864	2090
G. 4.0-4.2/-200/HF	11.0	632.1	507.5	93.4	0.7285	0.3456	0.77	6.213	3.50	0.1304	3.28	0.39	1913	2006	2103
H. 4.2-4.6/-200/HF	10.6	395.2	368.1	61.1	0.7231	0.3298	0.71	5.802	1.50	0.1276	1.03	0.79	1837	1946	2065
I. 3.8-4.0/-200/HF	6.4	2047.4	1412.6	104	0.6502	0.3187	1.57	5.599	1.58	0.1274	0.34	0.98	1738	1915	2062
J. 4.0-4.2/-200/HF	10.7	757.9	703.0	71.6	0.7504	0.3545	0.73	6.173	1.60	0.1263	1.13	0.78	1956	2000	2047
K. 4.2-4.6/-200/HF	10.6	352.5	399.4	56.9	0.7448	0.3829	0.70	6.687	1.00	0.1267	0.47	0.91	2089	2070	2052
L. 4.0-4.2/crushed/HNO3	13.3	526.3	215.5	803	0.5411	0.2713	0.67	4.747	0.68	0.1269	0.16	0.97	1547	1775	2055
M. +4.6/abr	1.6	722.6	259.1	425	0.5939	0.2221	0.70	3.838	0.90	0.1254	0.36	0.91	1292	1600	2034
N. 4.2-4.6/abr	3.6	511.2	242.9	1386	0.4250	0.3435	0.76	6.394	0.77	0.1350	0.15	0.97	1903	2031	2164
O. 4.0-4.2/turbid/crushed	15.2	487.0	203.4	376	0.6014	0.2540	0.70	4.347	0.90	0.1241	0.41	0.87	1459	1702	2016
P. 4.0-4.2/clear	5.4	714.2	320.1	2271	0.3704	0.3390	0.80	6.387	0.81	0.1367	0.15	0.97	1881	2030	2185
Q. 4.0-4.2/turbid/abr	17.1	492.2	201.1	595	0.6079	0.2569	0.87	4.402	0.90	0.1243	0.20	0.97	1473	1712	2018
R. 4.0-4.2/clear	13.7	694.2	332.9	3041	0.3721	0.3637	0.92	6.886	0.93	0.1373	0.15	0.97	1999	2096	2193
A410-Kaakamovaara albite diabase															
A. Total/borax	221.7	424.4	182.4	958	0.4725	0.3019	0.65	5.374	0.85	0.1291	0.44	0.86	1700	1880	2085
B. Total	6.9	392.4	153.9	639	0.4783	0.2781	0.67	4.970	1.80	0.1296	1.43	0.68	1581	1814	2092
C. +4.0/abr 2 h	4.7	254.9	119.5	2195	0.3818	0.3538	0.84	6.375	0.84	0.1397	0.15	0.97	1952	2028	2107

Appendix III continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A474-Konttikivalo albite diabase															
A. +4.1	25.4	900.9	249.7	713	0.2249	0.2247	0.75	3.909	1.50	0.1262	1.06	0.75	1306	1615	2045
B. +4.1/abr 1h	2.8	501.3	197.1	1367	0.2520	0.3192	0.65	5.899	0.65	0.1341	0.15	0.97	1785	1961	2151
C. +4.2/abr	7.2	860.6	265.1	938	0.1013	0.2749	0.77	4.967	0.77	0.1311	0.15	0.97	1565	1813	2112
D. +4.2/-200/abr	7.1	730.0	244.0	884	0.0942	0.2984	0.65	5.481	0.65	0.1332	0.15	0.97	1683	1897	2141
E. Zr+Badd/+4.2/abr	4.6	1005.9	269.4	560	0.1242	0.2273	0.65	3.956	0.75	0.1262	0.27	0.94	1320	1625	2046
F. Titanite/brown/abr 20 m	32.1	42.0	37.7	153	1.0908	0.3857	0.65	7.293	0.70	0.1371	0.23	0.94	2103	2148	2191
A475-Hattuselkä albite diabase															
A. +3.8	21.9	568.6	180.5	373	0.5384	0.2157	0.65	3.410	3.00	0.1146	2.78	0.43	1259	1506	1874
B. Zr+Badd	8.8	1022.5	388.8	2016	0.1023	0.3482	0.65	6.471	0.65	0.1348	0.15	0.97	1926	2041	2161
A475C. Baddeleyite	0.5	865.5	348.1	2388	0.0747	0.3806	0.89	7.169	0.90	0.1366	0.22	0.97	2079	2133	2184
D. Zr+Badd/+4.3	5.1	947.1	321.8	927	0.1448	0.2929	0.65	5.355	0.70	0.1326	0.20	0.96	1655	1877	2132
E. Badd+Zr/+4.6	4.0	1034.4	360.3	2376	0.0897	0.3235	0.65	6.038	0.65	0.1354	0.15	0.97	1806	1981	2168
F. Badd+Zr/+4.6/abr	2.0	969.6	355.1	3298	0.0708	0.3468	0.65	6.531	0.65	0.1362	0.15	0.97	1923	2050	2180
G. Badd+Zr/abr	2.2	977.0	360.9	3622	0.0785	0.3490	0.65	6.550	0.65	0.1361	0.15	0.97	1930	2052	2178
A476-Tornivaara albite diabase															
A. -3.8/brown	15.5	1873.0	173.2	570	0.2523	0.0750	0.66	0.928	1.30	0.0897	0.90	0.77	466	666	1420
B. Badd+Zr	3.4	1415.9	329.1	1343	0.0874	0.2135	0.65	3.841	0.65	0.1305	0.15	0.97	1247	1601	2104
C. Badd+Zr/+4.3/abr	9.4	1325.6	302.2	1530	0.0800	0.2116	0.65	3.796	0.65	0.1302	0.15	0.97	1237	1591	2100
D. Badd+Zr/+4.1/abr	4.1	1417.4	309.3	1502	0.0828	0.2021	0.65	3.609	0.65	0.1295	0.15	0.97	1186	1551	2091
E. Titanite	31.2	49.2	32.9	113	0.9605	0.2810	0.73	5.181	1.70	0.1337	1.25	0.75	1596	1849	2147
F. 4.0-4.2/clear/abr 2 h	1.2	1050.9	263.2	1113	0.1939	0.2116	0.65	3.717	0.70	0.1274	0.25	0.93	1237	1575	2062
G. Baddeleyite	0.7	1353.2	378.9	2994	0.0428	0.2726	0.65	5.133	0.65	0.1366	0.15	0.97	1554	1841	2184
H. Titanite	22.7	52.5	30.1	129	0.6870	0.2785	0.65	5.129	1.40	0.1336	1.07	0.68	1583	1840	2145
bA. 4.0-4.2 abr				1510		0.2456	0.65	4.444	0.65	0.1312	0.15	0.97	1415	1720	2114
bB. 3.8-4.0 abr				520		0.1555	0.65	2.574	0.90	0.1200	0.40	0.92	931	1293	1956
bC. +4.2 zr+badd				1952		0.2391	0.65	4.362	0.65	0.1323	0.15	0.97	1381	1705	2129
bD. Titanite abr				159		0.3865	0.65	7.338	0.90	0.1377	0.45	0.88	2106	2153	2198
A477-Kivimaa albite diabase															
A. 3.8-4.1	29.2	789.8	270.2	425	0.8413	0.1935	0.66	3.350	2.00	0.1255	1.53	0.91	1140	1492	2036
B. +4.2/clear/abr 2 h	4.4	536.2	159.1	1433	0.1085	0.2676	0.70	5.043	0.80	0.1367	0.33	0.91	1528	1826	2185
C. 3.8-4.0/clear/abr 3 h	7.3	1310.7	725.1	1956	1.0592	0.2878	0.94	5.342	0.94	0.1346	0.15	0.94	1630	1875	2159
D. Baddeleyite	7.1	533.2	185.6	2081	0.0504	0.3319	0.69	6.327	0.69	0.1382	0.15	0.97	1847	2022	2205
E. 4.0-4.2/grenish/abr	11.2	845.4	362.9	723	0.6515	0.2650	0.65	4.940	0.65	0.1352	0.15	0.97	1515	1809	2166
F. 4.0-4.2/clear/abr	11.2	992.3	544.2	1693	0.8633	0.3120	0.65	5.829	0.65	0.1355	0.15	0.97	1750	1950	2170

Appendix III continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A477G. Titanite	23.2	35.3	27.8	564	0.9423	0.4150	0.65	7.959	0.90	0.1391	0.43	0.84	2237	2226	2216
H. 3.8-4.0/grey-green/abr 3 h	5.6	1007.3	440.9	940	0.9089	0.2403	0.74	4.388	0.74	0.1325	0.15	0.97	1388	1710	2130
I. Badd+Zr/+4.2	4.1	781.6	210.7	1017	0.2019	0.2239	0.65	4.115	0.65	0.1333	0.15	0.97	1302	1657	2142
J. 4.0-4.2/clear/abr 2 h	3.6	605.0	274.0	1349	0.6640	0.2847	0.65	5.305	0.65	0.1352	0.15	0.97	1614	1869	2166
A478-Puukkokumpu albite diabase															
A. +3.8/+100	12.5	1428.7	323.9	89.5	0.2758	0.1171	0.65	1.904	1.20	0.1179	0.83	0.73	713	1082	1924
B. Badd+Zr/+4.6/abr	2.0	1013.3	362.6	721	0.1119	0.3106	0.65	5.863	0.65	0.1369	0.15	0.97	1743	1955	2188
C. 4.0-4.2/clear	6.7	984.7	353.7	301	0.2302	0.2621	0.65	4.794	0.70	0.1326	0.21	0.93	1500	1783	2133
A480-Sipojuntti albite diabase															
A. +4.1/Total	13.7	568.9	221.7	790	0.3566	0.3049	0.65	5.205	2.00	0.1238	1.60	0.72	1715	1853	2012
B. +4.2	4.8	615.4	249.3	1084	0.2969	0.3175	0.65	5.545	0.70	0.1267	0.26	0.93	1777	1907	2052
C. 4.0-4.2	10.7	830.0	381.4	1624	0.3856	0.3429	0.69	6.150	0.75	0.1301	0.31	0.91	1900	1997	2099
D. 4.0-4.2/HF	10.7	757.8	363.3	5889	0.3968	0.3619	0.68	6.519	0.80	0.1306	0.38	0.88	1991	2048	2106
E. 4.2-4.3	7.8	587.4	224.5	1562	0.2795	0.3071	0.68	5.270	0.70	0.1244	0.22	0.97	1726	1863	2021
F. 4.0-4.2/abr	8.9	763.9	331.8	2117	0.3563	0.3326	0.73	5.930	0.75	0.1293	0.21	0.96	1850	1965	2089
G. 4.0-4.2/crushed/HNO3	10.3	939.5	452.2	5174	0.4237	0.3567	0.65	6.417	0.66	0.1305	0.15	0.97	1966	2034	2104
H. Titanite	37.9	65.7	36.4	415	0.3450	0.3925	0.82	7.114	4.00	0.1315	3.62	0.54	2134	2125	2117
I. +4.2	2.9	560.1	215.1	1632	0.2680	0.3116	0.85	5.420	0.86	0.1262	0.17	0.97	1748	1888	2045
J. 3.8-4.0/crushed/HNO3	11.6	1037.5	500.7	1483	0.4446	0.3458	0.65	6.228	0.70	0.1306	0.22	0.94	1914	2008	2106
K. 3.6-3.8/crushed/HNO3	10.5	970.9	447.3	3781	0.4281	0.3394	0.65	6.094	0.65	0.1302	0.15	0.97	1883	1989	2101
L. 3.8-4.0/abr	1.0	1039.6	497.4	1892	0.4218	0.3509	0.92	6.314	0.92	0.1305	0.16	0.97	1939	2020	2104
M. 3.6-3.8/+125/abr 1 h	11.3	976.8	444.8	3363	0.4350	0.3335	0.65	5.986	0.67	0.1302	0.17	0.97	1855	1973	2100
A592-Luppovaara albite diabase															
A. Tit 3.5-3.6/+200/abr 30 m	31.3	37.2	28.1	507	1.0243	0.3816	0.75	7.255	0.80	0.1379	0.32	0.93	2083	2143	2200
B. Tit +100	29.0	42.9	29.5	557	1.0764	0.3419	0.65	6.391	0.75	0.1356	0.25	0.93	1895	2031	2171
C. Tit 100-150/abr	31.7	41.4	39.0	650	1.0414	0.3563	0.67	6.690	0.80	0.1362	0.36	0.89	1964	2071	2179
D. Tit 3.25-3.6/abr 30 m	31.9	45.5	29.9	630	1.0119	0.3384	0.65	6.312	0.67	0.1353	0.63	0.54	1879	2020	2167
A593-Luppovaara albite diabase															
A. Zr+Badd/clear/abr 1.5 h	3.2	636.7	305.9	1979	0.4306	0.3500	0.65	6.332	0.73	0.1312	0.23	0.93	1934	2022	2114
B. Turbid, abr 1.5 h	3.4	777.8	362.1	927	0.4702	0.3233	0.65	5.818	0.65	0.1305	0.15	0.97	1806	1949	2104

Appendix III continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A694-Misi gabbro															
A. +125/Total	47.1	244.9	105.5	7946	0.2336	0.3651	0.65	6.534	0.65	0.1298	0.15	0.97	2006	2050	2095
B. -125/Total	19.9	299.5	133.4	4233	0.2536	0.3703	0.65	6.664	0.65	0.1305	0.15	0.97	2030	2067	2105
C. Total	23.8	266.0	116.5	7944	0.2427	0.3687	0.65	6.618	0.65	0.1302	0.15	0.97	2023	2061	2100
A706-Runkausvaara albite diabase															
A. 3.8-4.0	8.2	1673.3	351.5	848	0.3735	0.1558	0.65	2.492	0.65	0.1071	0.15	0.97	933	1269	1896
B. 3.6-3.8/abr 3 h	2.0	1854.7	284.3	559	0.3886	0.1106	0.65	1.565	0.75	0.1026	0.30	0.90	676	956	1671
C. Titanite	24.1	23.2	21.1	210	1.2153	0.3933	0.65	7.515	1.00	0.1386	0.62	0.76	2137	2174	2209
A865-Susivaara albite diabase															
A. 4.2-4.6/+200	18.8	757.3	347.0	1734	0.8444	0.2649	0.67	4.523	0.80	0.1238	0.39	0.88	1515	1735	2012
B. 3.8-4.2	15.4	760.9	350.4	1349	0.9663	0.2498	0.71	4.214	0.71	0.1223	0.21	0.96	1437	1676	1990
C. 3.6-3.8/+200	11.1	1166.1	443.0	1250	0.7742	0.2262	0.69	3.794	0.90	0.1216	0.51	0.82	1314	1591	1980
D. 3.8-4.2/abr 3 h	11.6	882.3	419.4	1550	0.9235	0.2637	1.15	4.466	1.23	0.1228	0.49	0.92	1508	1724	1997
E. 4.0-4.2/abr 3 h	11.2	597.7	338.2	1929	0.8766	0.3218	0.70	5.654	0.71	0.1274	0.15	0.97	1798	1924	2063
F. Baddeleyite	1.3	643.9	256.8	4559	0.1037	0.3717	0.75	6.976	0.76	0.1361	0.16	0.97	2037	2108	2178
G. Badd+Zr	3.3	578.5	224.0	2791	0.1404	0.3470	0.65	6.467	0.65	0.1352	0.15	0.97	1920	2041	2166
A865H. Badd+Zr	3.1	565.4	218.5	4569	0.1483	0.3469	0.65	6.441	0.65	0.1347	0.15	0.97	1919	2037	2159
I. +4.2/clear/abr 2 h	4.1	819.7	437.4	4054	0.5780	0.3584	0.65	6.632	0.65	0.1342	0.15	0.97	1974	2063	2154
J. 3.8-4.2/clear/abr 3 h	7.4	1386.3	761.4	4343	0.8415	0.3204	0.74	5.765	0.74	0.1305	0.15	0.97	1791	1941	2105
K. Titanite	30.4	61.9	40.3	280	0.4289	0.4161	0.65	7.963	0.75	0.1388	0.27	0.90	2242	2226	2212
L. Titanite	32.1	56.6	36.1	376	0.4596	0.4156	0.65	8.055	1.10	0.1406	0.75	0.73	2240	2237	2234
A986-Kaakamovaara albite diabase															
A. +4.6	16.6	402.4	121.0	1640	0.3982	0.2230	0.65	3.921	0.65	0.1275	0.15	0.97	1297	1618	2064
B. +4.6/HF	20.6	303.4	108.1	6746	0.4066	0.2679	0.65	4.756	0.65	0.1288	0.15	0.97	1529	1777	2081
C. +4.6/abr 2 h	15.6	273.2	103.6	874	0.4375	0.2673	0.65	4.706	0.70	0.1277	0.27	0.92	1527	1768	2066
D. +4.6/abr 5 h	16.2	350.4	118.0	1130	0.4484	0.2393	0.96	4.202	1.00	0.1274	0.28	0.96	1383	1674	2062
E. 4.2-4.6/+100/abr 3 h	13.5	599.0	161.9	710	0.5700	0.1753	0.73	3.011	0.80	0.1245	0.26	0.94	1041	1410	2022
F. 4.2-4.6/+100/abr 5 h	13.3	567.6		853	0.5712	0.1855	0.65	3.166	0.65	0.1238	0.15	0.97	1097	1449	2011
G. 4.2-4.6/clear/abr 3 h	15.2	588.8	222.8	3784	0.4552	0.2740	0.65	4.924	0.68	0.1304	0.19	0.95	1561	1806	2102
H. Titanite	30.2	71.1	29.2	1038	0.0417	0.3856	0.65	7.037	0.65	0.1324	0.15	0.97	2102	2116	2129
I. Baddeleyite	3.8	347.7	67.9	1159	0.0712	0.1816	0.65	3.223	0.66	0.1287	0.17	0.97	1075	1462	2080
J. Badd+Zr	6.4	552.1	89.5	823	0.1292	0.1413	0.65	2.446	0.65	0.1255	0.15	0.97	852	1256	2036

Appendix III continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A77-Lapalionkangas gabbro															
A. +4.6	21.7	158	76	4998	0.32	0.37913	0.65	6.8868	0.65	0.13175	0.15	0.97	2072	2096	2121
B. 4.3-4.6/HF	20.6	218	105	9826	0.33	0.38196	0.65	6.9442	0.65	0.13186	0.15	0.97	2085	2104	2123
A78-Lapalionkangas gabbro															
A. +4.6	21.0	152	74	1841	0.31	0.38132	0.65	6.9154	0.90	0.13154	0.50	0.84	2082	2100	2118
B. 4.3-4.6/HF	13.3	242	115	5983	0.31	0.38428	0.65	6.9538	0.75	0.13125	0.25	0.95	2096	2105	2114
C. 4.3-4.6/abr 5 h	22.3	265	125	5122	0.30	0.38091	0.65	6.9049	0.65	0.13148	0.15	0.97	2080	2099	2117
D. 4.3-4.6/-200/abr 5 h	18.4	266	127	7996	0.29	0.38693	0.65	7.0114	0.70	0.13143	0.20	0.96	2108	2112	2117
A694-Misi gabbro															
A. total/+125	47	245	106	7946	0.23	0.36524	0.7	6.5359	0.7	0.12979	0.2	0.96	2006	2050	2095
B. total/-125	19.9	299	133	4233	0.25	0.37027	0.7	6.664	0.7	0.13054	0.2	0.96	2030	2067	2105
D. total	23.8	265	117	7944	0.24	0.36875	0.7	6.6187	0.7	0.13019	0.2	0.96	2023	2061	2100

*) Isotopic ratios corrected for fractionation, blank and age related common lead (see methods).
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = Standard error

Appendix IV. U-Pb mineral data from felsic dykes in the Peräpohja area.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A19-Kapustavuoma felsic porphyry dyke															
A. +4.6		135.4	30.4	954	0.1200	0.2595	0.65	4.635	0.65	0.1295	0.15	0.97	1487	1755	2091
B. 4.2-4.6/+100/HF		502.0	94.4	2588	0.0820	0.2174	0.65	3.738	0.65	0.1247	0.15	0.97	1267	1579	2025
C. 4.2-4.6/100-200		489.7	110.3	1849	0.1000	0.2602	0.65	5.252	0.65	0.1464	0.15	0.97	1490	1861	2304
D. +4.2/HF		302.4	70.1	3720	0.0830	0.2679	0.65	5.012	0.65	0.1357	0.15	0.97	1529	1821	2173
E. 4.2-4.6/+100/HF/abr 3 h		381.3	96.3	6016	0.0770	0.2920	0.65	5.121	0.65	0.1272	0.15	0.97	1651	1839	2059
F. +4.2/HF/abr 3 h		162.2	46.5	1055	0.1107	0.3312	0.65	5.945	0.65	0.1302	0.15	0.97	1844	1967	2100
A512-Elijärvi albitite															
A. Total	59.2	152.1	64.9	507	0.1969	0.3331	0.80	7.634	0.80	0.1662	0.20	0.97	1853	2188	2520
B. +4.6/+100/rounded	21.4	105.3	51.2	613	0.2378	0.3755	0.65	8.780	0.65	0.1696	0.15	0.97	2054	2315	2553
C. +4.6/subhedral	31.9	140.1	61.4	666	0.2040	0.3507	0.65	8.091	0.65	0.1674	0.15	0.97	1937	2241	2531
D. +4.6/long	13.7	246.2	90.1	437	0.1699	0.2873	0.65	6.554	0.65	0.1654	0.15	0.97	1628	2053	2512
E. -4.2/M	7.7	686.9	150.5	251	0.1829	0.1605	0.65	3.152	0.65	0.1424	0.15	0.97	959	1445	2257
F. +4.6/rounded/abr 3 h	9.3	121.1	62.0	711	0.2046	0.4071	0.65	9.614	0.65	0.1713	0.15	0.97	2201	2398	2570
A643-Keinokangas quartz porphyry															
A. +4.2/+100	52.2	125.0	58.0	302	0.2194	0.3435	0.65	5.945	0.65	0.1255	0.15	0.97	1903	1967	2036
B. 4.0-4.2/+100	21.1	159.7	70.8	350	0.2377	0.3315	0.65	5.735	0.65	0.1255	0.15	0.97	1845	1936	2035
C. +4.2/+150/HF/abr 2 h	4.2	87.8	37.8	1177	0.1621	0.3744	0.67	6.544	0.70	0.1268	0.18	0.97	2050	2051	2053
D. +4.6/HF/abr 3 h	2.8	89.6	38.3	1098	0.1564	0.3736	0.65	6.509	0.65	0.1264	0.16	0.96	2046	2047	2048
E. +4.6/abr 8 h	21.2	100.5	47.6	250	0.1954	0.3441	0.65	5.946	0.65	0.1253	0.15	0.97	1906	1968	2033
F. 4.0-4.2/HF/abr 3 h	9.8	83.4	43.8	221	0.1696	0.3773	0.65	6.597	0.72	0.1268	0.35	0.87	2063	2058	2054
A718-Kallinkangas albite-quartz porphyry															
A. +4.6/+50/large		181.9	67.8	1137	0.1756	0.4309	0.65	10.056	0.65	0.1693	0.15	0.97	2309	2439	2550
B. +4.6/-200		269.7	80.2	502	0.1480	0.3435	0.99	6.792	0.99	0.1434	0.31	0.95	1903	2084	2268
C. 4.2-4.6/+100/abr 1 h		611.2	245.2	720	0.1720	0.4637	0.71	11.729	0.71	0.1835	0.16	0.97	2455	2583	2684
D. 4.2-4.6/abr 3 h		680.8	264.0	609	0.1560	0.4482	0.67	11.416	0.67	0.1848	0.15	0.97	2387	2557	2696
E. 4.0-4.2/+100/abr 2 h/dark		1079.0	418.4	763	0.1550	0.4481	0.87	11.409	0.87	0.1847	0.15	0.98	2386	2557	2695
F. 4.0-4.2/100-200/abr 2 h		1061.0	400.7	556	0.1450	0.4364	0.69	11.134	0.69	0.1851	0.15	0.98	2334	2534	2698
G. +4.6/microgems		131.2	52.1	3018	0.2020	0.4593	0.65	10.734	0.63	0.1695	0.15	0.97	2436	2500	2552
H. +4.6/dark		438.0	173.9	752	0.1470	0.4589	0.65	11.769	0.70	0.1860	0.20	0.96	2434	2586	2707
I. +4.6/abr 3 h/clear		151.0	53.9	2541	0.1550	0.4124	0.65	8.777	0.65	0.1544	0.15	0.97	2225	2315	2395

Appendix IV continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A718J. +4.6/abr 2 h		248.9	83.5	2344	0.0980	0.3876	0.65	8.163	0.65	0.1528	0.15	0.97	2111	2249	2377
K. +4.6/powder		295.5	92.8	1878	0.0890	0.3631	0.65	7.308	0.65	0.1460	0.18	0.96	1996	2149	2299
L. 4.3-4.6/+100/microgems		102.8	42.0	5589	0.2360	0.4721	0.65	11.221	0.65	0.1724	0.15	0.97	2492	2541	2581
M. Titanite		13.1	3.9	34	1.0700	0.3488	0.65	6.198	1.80	0.1289	1.55	0.54	1928	2004	2082
N. 4.3-4.6/abr 1 h/microgems		98.5	39.9	2962	0.2400	0.4684	0.65	11.047	0.65	0.1711	0.15	0.97	2476	2527	2568
A719-Rantamaa albite porphyry dyke															
A. +4.2		411.2	165.6	1358	0.1480	0.4654	0.65	12.817	0.65	0.1997	0.15	0.97	2463	2666	2824
B. +4.2/long/abr 3 h		469.3	139.4	3161	0.0780	0.3434	0.65	7.008	0.65	0.1480	0.16	0.97	1902	2112	2323
C. +4.3/dark/abr 1 h		578.5	229.3	2903	0.1580	0.4579	0.65	12.610	0.65	0.1998	0.15	0.97	2430	2650	2824
D. +4.2/abr 2 h/microgems		160.0	58.0	4088	0.1300	0.4190	0.65	9.312	0.65	0.1612	0.15	0.97	2255	2369	2468
E. +4.2/abr 1 h/reddish		150.8	59.6	5021	0.1690	0.4567	0.65	10.830	0.65	0.1720	0.15	0.97	2425	2508	2577
A983-Palonkalliot dacitic porphyry															
A. +4.2/clear		178.2	60.3	3171	0.1460	0.3912	0.65	8.301	0.65	0.1539	0.15	0.97	2128	2264	2390
B. +3.8/abr 3 h/brown		1101.0	436.7	9861	0.0580	0.4584	0.65	11.359	0.65	0.1798	0.15	0.97	2432	2553	2650
C. Titanite		20.3	5.9	119	0.2900	0.3374	0.74	5.324	0.90	0.1145	0.42	0.89	1873	1872	1871

*) Isotopic ratios corrected for fractionation, blank and age related common lead (Stacey & Kramers 1975).
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = standard errorin the mean

Appendix V. U-Pb mineral data from Proterozoic intrusive rock in the Peräpohja area

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A154-Nosa granodiorite															
A. long clear/-100		247.1	62.7	1280	0.1090	0.2932	0.58	4.498	0.60	0.1113	0.22	0.90	1657	1730	1820
B. 4.2-4.6/-100/long turbid		371.0	90.9	564	0.1550	0.2833	0.60	4.412	0.61	0.1130	0.33	0.85	1607	1714	1847
C. microgems/-100		168.7	50.4	1098	0.1170	0.3456	0.58	5.918	0.59	0.1242	0.16	0.96	1913	1963	2017
D. Titanite		168.3	49.4	364	0.3790	0.3390	0.56	5.373	0.58	0.1149	0.17	0.95	1882	1880	1879
E. +4.3/abr 4 h/microgem		214.8	77.0	13551	0.1210	0.4142	0.58	9.177	0.58	0.1607	0.10	0.99	2233	2355	2463
F. long clear/abr 2 h		250.4	63.5	6218	0.1560	0.2929	0.58	4.275	0.58	0.1059	0.10	0.99	1655	1688	1729
G. dark/abr 2 h		620.7	189.6	1293	0.1000	0.3530	0.59	8.336	0.60	0.1713	0.10	0.99	1948	2268	2570
H. 4.2-4.6/+150/abr 3 h/long		234.2	66.1	9069	0.0840	0.3263	0.58	5.697	0.60	0.1266	0.10	0.99	1820	1930	2052
I. 4.2-4.6/+150/abr 3 h/dark		301.4	104.3	1670	0.0990	0.4000	0.60	9.234	0.60	0.1674	0.16	0.97	2169	2361	2532
J. +4.3/+100/microgems/abr		117.5	40.7	8524	0.1500	0.4003	0.60	8.046	0.61	0.1458	0.13	0.98	2170	2236	2297
K. 4.3-4.6/-100/microgems/abr		166.3	50.2	7535	0.0990	0.3488	0.57	6.109	0.57	0.1270	0.12	0.99	1928	1991	2057
L. 4.3-4.6/-100/microgems/abr		155.2	47.4	428	0.1770	0.3532	0.57	6.137	0.57	0.1260	0.22	0.90	1949	1995	2043
M. 4.2-4.6/-100/microgems/abr		170.6	53.0	1249	0.1210	0.3590	0.61	6.366	0.61	0.1286	0.10	0.99	1977	2027	2079
N. 4.2-4.6/-100/microgems/abr		167.4	51.6	9899	0.1040	0.3560	0.56	6.284	0.60	0.1284	0.10	0.98	1963	2016	2071
O. 4.3-4.6/microgems/abr		154.3	52.4	4273	0.1360	0.3927	0.61	7.838	0.61	0.1448	0.11	0.99	2135	2212	2285
A155-Ruottala granodiorite															
A. +4.6/abr 1 h/clear/sharp		174.4	58.5	22795	0.1230	0.3877	0.59	7.496	0.60	0.1402	0.12	0.98	2112	2172	2230
B. 4.2-4.6/abr 1 h		345.2	101.3	4533	0.1380	0.3392	0.59	6.229	0.59	0.1332	0.10	0.99	1882	2008	2140
C. +4.2/abr 1 h/dark/subhed		537.3	177.6	1812	0.1230	0.3820	0.58	9.621	0.58	0.1827	0.10	0.99	2085	2399	2677
E. 4.2-4.6/long		432.3	106.1	3298	0.1414	0.2837	0.60	5.179	0.60	0.1324	0.15	0.97	1609	1849	2130
F. 4.2-4.6		349.2	102.7	5221	0.1365	0.3398	0.59	6.281	0.59	0.1341	0.10	0.99	1885	2015	2151
A490-Kuttijänkkä diorite															
A. 4.2-4.3/+200		771.4	200.6	11755	0.0982	0.3005	0.67	4.702	0.67	0.1135	0.15	0.97	1693	1767	1855
B. 4.2-4.3/+200/abr 5 h		696.8	197.1	27257	0.0982	0.3269	0.65	5.166	0.65	0.1146	0.15	0.97	1823	1847	1874
C. 4.3-4-5/+200/abr 3 h		541.0	154.4	17509	0.0993	0.3298	0.65	5.220	0.65	0.1148	0.15	0.97	1837	1855	1876
D. 4.3-4.5/+200/abr 1 h		530.8	152.8	17302	0.1004	0.3327	0.65	5.272	0.65	0.1149	0.15	0.97	1851	1864	1878
E. Monazite		1042.0	307.7	19466	11.0200	0.3413	0.65	5.466	0.65	0.1162	0.15	0.97	1892	1895	1898
F. Monazite		765.1	218.4	11009	11.0700	0.3300	0.65	5.251	0.65	0.1154	0.15	0.97	1838	1860	1886
A735-Kaakamo diorite															
A. +4.2		448.6	128.7	17445	0.1937	0.3315	0.65	5.254	0.65	0.1150	0.15	0.97	1845	1861	1879
B. +4.2/abr 2 h		469.3	137.3	36209	0.1961	0.3382	0.65	5.346	0.65	0.1150	0.15	0.97	1878	1879	1880
C. +4.2/abr 2 h		456.1	133.8	28289	0.1955	0.3390	0.65	5.374	0.65	0.1150	0.15	0.97	1881	1880	1879

Appendix V continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A744-Nahonmaa granodiorite															
A. +4.6/+100/mixed		162.9	50.8	4056	0.1385	0.3603	0.61	7.382	0.61	0.1486	0.10	0.99	1983	2158	2329
B. +4.6/abr 1 h/long		210.9	63.1	3340	0.1024	0.3459	0.58	6.304	0.62	0.1322	0.20	0.95	1914	2019	2127
C. 4.2-4.6/abr 1 h/long		352.7	102.9	3136	0.1009	0.3373	0.74	6.526	0.75	0.1403	0.12	0.98	1873	2049	2231
D1. Titanite		74.6	22.8	167	0.7500	0.3531	0.76	5.670	0.76	0.1165	0.40	0.86	1949	1926	1902
D2. Tit/3.4-3.6/abr 20 m		64.9	20.7	139	0.8550	0.3690	2.70	5.940	2.86	0.1166	4.12	0.66	2024	1966	1905
D3. Tit/3.4-3.6/abr 20 m		73.8	22.2	208	0.7470	0.3478	0.81	5.444	1.01	0.1136	1.14	0.62	1923	1891	1857
D4. Tit/3.4-3.6/+100/abr 10 m		72.3	25.8	82	0.9424	0.4116	0.78	6.537	0.80	0.1152	1.74	0.61	2222	2050	1882
E. +4.6/+200/gems/abr 3 h		169.2	51.7	6621	0.1077	0.3534	0.59	6.316	0.60	0.1296	0.10	0.99	1950	2020	2093
F. 4.2-4.6/+200/gems/abr 3 h		239.6	71.3	2914	0.0996	0.3438	0.58	6.019	0.58	0.1270	0.17	0.95	1905	1978	2056
G. +4.3/gems/abr 4 h		148.2	47.0	8097	0.1200	0.3669	0.60	6.659	0.60	0.1316	0.10	0.99	2014	2067	2119
A755-Liakka gabbro															
A. +4.6		279.8	77.2	3544	0.3839	0.3188	0.65	5.063	0.65	0.1152	0.15	0.97	1783	1829	1883
B. 4.2-4.6		358.1	97.5	3275	0.2738	0.3148	0.65	4.980	0.65	0.1148	0.15	0.97	1764	1815	1876
C. 4.0-4.2		809.3	200.5	1839	0.2887	0.2863	0.94	4.495	0.95	0.1139	0.15	0.95	1622	1729	1862
D. 4.2-4.6/abr 1 h		339.7	95.1	3091	0.2751	0.3235	0.65	5.102	0.70	0.1144	0.20	0.90	1806	1836	1870
E. 4.2-4.6/abr 2 h		329.6	93.5	7175	0.2825	0.3279	0.65	5.201	0.65	0.1150	0.15	0.97	1828	1852	1880
F. 4.2-4.6/abr 3 h		315.8	90.2	5688	0.2944	0.3299	0.65	5.239	0.65	0.1152	0.15	0.97	1838	1858	1882

*) Isotopic ratios corrected for fractionation, blank and age related common lead (Stacey & Kramers 1975).
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = standard errorin the mean

Appendix VI. U-Pb mineral analyses from metasedimentary rocks in the Peräpohja area.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A51-Sompujärvi granite clast in conglomerate															
A. Total	21.0	663.1	249.7	2108	0.086	0.3442	0.65	7.956	0.65	0.1677	0.15	0.97	1906	2226	2534
B. +4.3/clear/abr 3 h	5.6	301.5	151.0	6803	0.105	0.4523	0.65	10.940	0.65	0.1755	0.15	0.97	2405	2518	2610
C. 4.2-4.6	10.5	481.2	219.9	3753	0.092	0.4150	0.65	9.857	0.65	0.1723	0.15	0.97	2237	2421	2580
D. 4.2-4.6/abr 5 h	10.8	441.8	221.0	7522	0.105	0.4518	0.65	10.880	0.65	0.1747	0.15	0.97	2403	2512	2603
E. 4.2-4.6/abr 14 h	10.7	453.8	222.7	4472	0.100	0.4432	0.65	10.634	0.65	0.1740	0.15	0.97	2365	2491	2596
F. 4.0-4,2/abr 3 h	10.9	799.9	346.9	2146	0.096	0.3903	0.65	8.935	0.65	0.1661	0.15	0.97	2124	2331	2518
G. 4.0-4.2/abr 2 h	3.3	793.3	333.5	2335	0.095	0.3794	0.60	8.676	0.61	0.1659	0.11	0.98	2073	2304	2516
H. +4.3/long/abr 3 h	4.6	488.9	231.7	4502	0.091	0.4317	0.65	10.295	0.65	0.1729	0.15	0.97	2313	2461	2586
I. +4.6/abr 3 h	1.4	64.6	37.0	609	0.041	0.4655	0.65	11.591	0.69	0.1806	0.29	0.91	2463	2571	2658
A518-Lautakodanoja quartzite															
A. Transparent grey	1.3	287.8	149.9	722	0.0460	0.4626	0.65	11.678	0.65	0.1893	0.15	0.97	2451	2578	2681
B. Red prisms	0.7	689.4	313.1	233	0.0374	0.3540	1.59	8.678	1.60	0.1778	0.45	0.96	1953	2304	2632
A720-Napinmäki mica schist															
A. 4.3-4.5/+300	6.1	306.1	133.6	2630	0.1551	0.3787	0.65	8.969	0.65	0.1718	0.15	0.97	2070	2334	2575
A783-Kirakka-aapa quartzite/keratophyre															
A. +3.8/brown/round/abr	2.0	806.3	342.9	7646	0.1532	0.3772	0.72	7.899	0.72	0.1519	0.15	0.97	2063	2219	2367
B. grey needles/abr	2.2	1051.5	306.4	1940	0.3150	0.2331	0.65	3.812	0.65	0.1186	0.15	0.97	1350	1595	1935
C. grey needles/abr	1.6	1173.1	242.7	2834	0.2997	0.1690	0.65	2.583	0.65	0.1109	0.15	0.97	1006	1295	1814
A1103-Kätkävaara quartzite															
A. +4.5	7.7	127.4	61.4	1164	0.2309	0.3859	0.65	10.241	0.65	0.1925	0.15	0.97	2103	2456	2763
B. 4.3-4.5	11.2	197.6	93.8	679	0.2475	0.3709	0.65	9.831	0.65	0.1923	0.15	0.97	2033	2419	2761
A1104-Palokivalo quartzite															
A. +4.5	6.5	126.3	62.4	2363	0.1991	0.4106	0.65	11.048	0.65	0.1952	0.15	0.97	2217	2527	2786
B. 4.3-4.5	7.8	220.0	97.6	1866	0.2000	0.3682	0.65	9.853	0.65	0.1941	0.15	0.97	2020	2421	2777
C. 4.2-4.3	6.4	413.1	135.0	665	0.2267	0.2589	0.65	6.836	0.65	0.1915	0.15	0.97	1484	2090	2755

Appendix VI continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A1105-Laitamaa quartzite															
A. +4.5/pale oval	11.1	124.1	73.7	930	0.2356	0.4688	0.80	13.013	0.81	0.2013	0.16	0.95	2478	2680	2837
B. 4.2-4.3/brown round	8.0	431.9	196.9	264	0.2563	0.3288	0.70	8.748	0.71	0.1930	0.15	0.96	1832	2312	2767
A1106-Myllyköngäs quartzite															
A. +4.5	9.4	123.4	69.7	1824	0.2248	0.4530	0.65	12.966	0.65	0.2076	0.15	0.97	2408	2677	2887
B. 4.2-4.3	10.5	541.2	258.9	968	0.1898	0.3908	0.65	10.913	0.65	0.1892	0.15	0.97	2126	2452	2735
A1107-Kallinkangas quartzite															
A. +4.2/rounded	3.7	137.3	78.7	2511	0.1794	0.4768	0.80	13.677	0.82	0.2081	0.16	0.95	2513	2727	2890
B. +4.2/irregular	9.9	154.0	93.4	2199	0.1977	0.4975	0.65	14.136	0.65	0.2061	0.15	0.97	2602	2758	2875
C. 4.0-4.2/euhedral	0.8	771.2	331.4	417	0.2033	0.3355	0.65	8.191	0.65	0.1771	0.15	0.97	1864	2252	2625
A1108-Lehtosaari quartzite															
A. +4.5	2.3	160.1	90.9	1446	0.2142	0.4541	0.65	12.669	0.65	0.2183	0.15	0.97	2413	2727	2968
B. 4.3-4.5/dark rounded	11.4	347.0	181.1	1018	0.2049	0.4158	0.65	11.544	0.65	0.2014	0.15	0.97	2241	2568	2837
C. 4.2-4.3/dark ronded	7.5	501.9	172.4	946	0.2096	0.2754	0.65	7.036	0.65	0.1853	0.15	0.97	1568	2116	2701
A1438-Korkiavaara arkosite															
A. +4.2/+200/abr 16 h	6.4	257.7	97.2	9884	0.173	0.3379	0.65	5.608	0.65	0.1204	0.1	0.99	1876	1917	1962
B. +4.2/+200/abr 6 h	5.9	269.6	100.1	12273	0.165	0.3345	0.65	5.542	0.65	0.1202	0.1	0.99	1860	1907	1958
C. +4.2/+200	5.8	340.9	110	10524	0.157	0.2934	0.65	4.771	0.65	0.1179	0.1	0.99	1658	1779	1925
D. +4.2/-200	7.2	305.4	101.2	12973	0.14	0.3052	0.65	4.928	0.65	0.1171	0.1	0.99	1717	1807	1912

*) Isotopic ratios corrected for fractionation, blank and age related common lead (Stacey & Kramers 1975).
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = standard errorin the mean

Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
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U-Pb ISOTOPIC AGE DETERMINATIONS FROM THE KOLARI-MUONIO AREA, WESTERN FINNISH LAPLAND

by
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Väänänen, J. & Lehtonen, M.I. 2001. U-Pb isotopic age determinations from the Kolari-Muonio area, western Finnish Lapland. *Geological Survey of Finland, Special Paper 33,* 85–93. 5 figures, 2 tables and one appendix.

Four zircon age groups of igneous rocks have been defined in the Kolari-Muonio area: 1) Archean (c. 2.6 Ga) for the granodioritic gneisses and granodiorites of the basement, 2) 2027±33 Ma for diabases intruding the volcanic rocks of the Kolari Formation, 3) 1860-1890 Ma for synorogenic granitoids of the Haaparanta Suite, and 4) c. 1.8 Ga for postorogenic microcline granites.

Several Archean and Paleoproterozoic zircon populations from the basement granodiorites in the Muonio area indicate apparently a Paleoproterozoic reactivation of the Archean basement. The titanite age, 1845 Ma, of the granodioritic gneisses and the titanite age distribution of 1885-1800 Ma of Paleoproterozoic rocks reflect metamorphic resetting of the Archean basement and the Paleoproterozoic rocks. Neither magmatic nor tectonometamorphic event are known to explain the exceptionally young zircon age of 1747 Ma and titanite age of 1748 Ma obtained for a mafic pegmatoid and a skarn rock, respectively, in the Rautuvaara mine.

Key words (GeoRef Thesaurus, AGI): absolute age, U/Pb, zircon, titanite, granites, diorites, gabbros, diabase, Paleoproterozoic, Archean, Kolari, Muonio, Lappi Province, Finland

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INTRODUCTION

The pioneering geological work in the Kolari-Muonio area was done during the first decades of the 20th century (Hackman 1927, Mikkola 1936, 1941). The geological picture of the area was updated in the 1970s and 1980s (Hiltunen & Tontti 1976, Rastas 1980, Lehtonen 1984, 1988, Hiltunen 1982, Lindroos & Henkel 1981, Lehtonen et al.1985, Väänänen 1989, Geological Map, Northern Fennoscandia, 1:1 mill. 1987). In conjunction with the 1:100,000 scale geological mapping of the Muonio, Kihlanki, Kittilä, Kolari and Kurtakko map sheets a number of radiometric age determinations were carried out (Lehtonen 1980, 1981, Rastas 1984, Väänänen 1984, 1992, 1998). The results from the Muonio area have been published and discussed already in detail by Lehtonen (1984). Hiltunen (1982) published and discussed data

from various rock types collected in connection with the mineral exploration by the Rautaruukki Oy in western Lapland. In this paper these previous results are briefly summarized and some new datings are reported. The datings were performed by O. Kouvo and discussed in his unpublished reports (archives of the GTK).

GENERAL GEOLOGY

The bedrock of the Kolari-Muonio area forms the western extension of the Paleoproterozoic Central Lapland Greenstone Belt (Fig. 1). The Kolari-Muonio area is predominantly occupied by supracrustal formations intruded by granitoid rocks and by minor dykes. In the Kolari area the supracrustal sequence lies on top of the Venejärvi Complex, which is mainly composed of migmatitic metasedimentary rocks and granites. From oldest to youngest, the Paleoproterozoic sequence comprises the following lithostratigraphic units: the Teuravuoma, Haisujupukka, Rautuvaara, Kolari and Tapojärvi Formations (Väänänen 1998). The Teuravuoma Formation consists of volcanic rocks corresponding with their stratigraphical position and geochemistry to the ultramafic and mafic volcanic rocks of the Onkamo Group (OnG c. 2.44 Ga) in the Central Lapland Greenstone Belt (see Räsänen et al. 1996, Lehtonen et al. 1998). The Haisujupukka Formation contains metasedimentary rocks, mainly quartzites, obviously correlative with the Virttiövaara Formation of the Sodankylä Group (SoG> 2.21 Ga). The amphibolites, graphite-sulphide schists, iron ores (Cu, Au) and skarn rocks of the Rautuvaara Formation and the volcanic rocks of the Kolari Formation can be assigned to the Savukoski Group (SaG> 2.05 Ga). The epiclastic metasedimentary rocks of the Tapojärvi Formation have been connected with the Lainio Group (LaG≤1.88 Ga). The stratigraphical sequence in the Muonio area matches well with that of the Kolari area (cf. Lehtonen 1984).

The most voluminous igneous rocks intruding the supracrustal sequence are synorogenic granitoids and gabbroic rocks as well as the postorogenic microcline granites. The synorogenic intrusions are variously deformed and have conformable contacts with their country rocks. These 1.89-1.86 Ga intrusions have been assigned to the Haaparanta Suite occurring widely in Western Lapland (called the Haparanda Series by Ödman et al. 1949 and Ödman 1957). These intrusions were called the Syenite Series by Mikkola (1941). Lehtonen (1988) divided the Haaparanta Suite into two lithological associations in the Muonio-Kihlanki area: the quartz monzodiorite association (QMD) and quartz diorite-tonalite-trondhjemite association (QTT). Microcline granites form plutons and migmatites cutting all the other rocks. They are comparable with the Lina granites in Northern Sweden (Ödman 1957).

PREVIOUS AGE DETERMINATIONS

Previously published age data (Hiltunen 1982, Lehtonen 1984) are summarized in Tables 1 and 2 and sample locations are shown in Figure 1. They have been recalculated using the IsoplotEx-program, and the results, including the MSWDs, are reported. If the data have been interpreted in a different manner than in the original papers, this is mentioned in the following paragraphs.

Monzonite samples A583 (1876±16 Ma) and A837 (1892±6 Ma) represent the quartz monzodiorite association (QMD), while the quartz diorite-tonalite-trondhjemite association (QTT) is represented by the granodiorite samples A967 (1850±41 Ma, combined fractions with A312), A523 (heterogenous zircon), A569 (heterogenous zircon), A838 (1832±7 Ma) and the sample A836 (heterogenous zircon, monazite age 1790 Ma). The granodiorite samples A835 (2591±16 Ma, titanite age 1845 Ma) and A378 (1977±46 Ma, 2444±96 Ma for two zircon fractions), and the pegmatitic microcline granite sample A966 (1778±2 Ma, MSWD=0.58) have been taken from the assumed Archean basement area in Muonio.

Samples A840 (1862±4 Ma) and A958 (1849±19 Ma) represent granitoids of the Haaparanta Suite in the vicinity of the Rautuvaara mine. A958 is an albititic contact variety taken from the foot wall of an ore body at Rautuvaara. The three zircon fractions analyzed are somewhat discordant and plot relatively close to each other. If the upper intercept for A958 is forced through the 1860 Ma point of the concordia curve, the MSWD increases from 0.57 to 2.1. Thus it may be argued that there is no significant difference between the results from the two samples. Samples A963 and A949 are skarn rocks associated with the Haaparanta Suite granitoids at the Hannukainen ore and Rautuvaara mine respectively. The three zircons from A963-

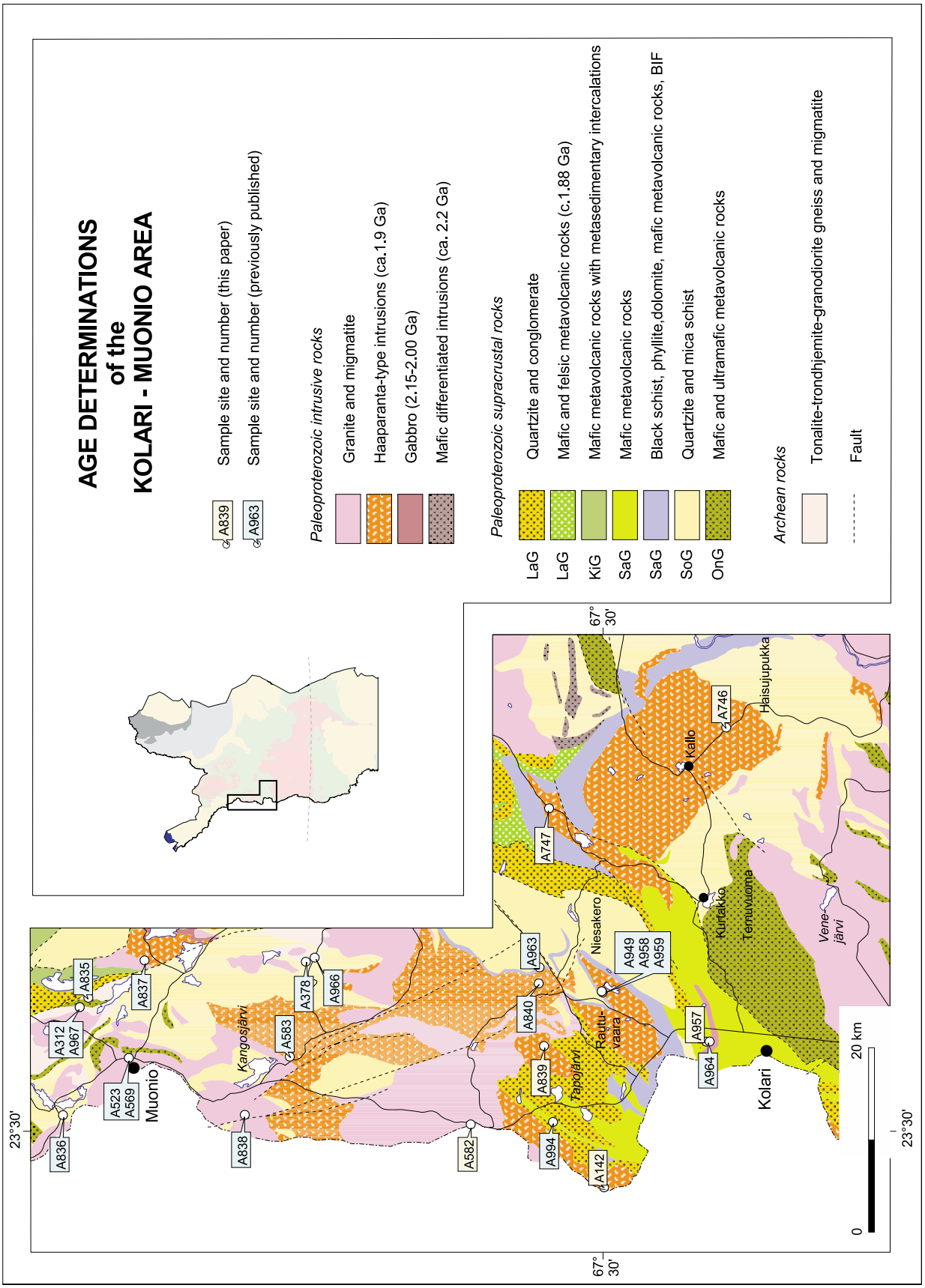

Fig. 1. Simplified surface geology and location of age determination samples from the Kolari-Muonio area, western Lapland. LaG = Lainio Group, KiG = Kittilä Group, SaG = Savukoski Group, SoG = Sodankylä Group, OnG = Onkamo Group.

Table 1. Previously published U-Pb age results from the Kolari-Kihlanki-Muonio area.

Sample	Map	Northing	Easting	Location	Lithology	Mineral	Age	Reference
A959	271312	7489.81	2496.95	Rautuvaara	Mafic pegmatoid	Zircon	1748±7	Hiltunen 1982
A963	271410	7496.79	2499.28	Kuervaara	Skarn	Zircon	1797±5	Hiltunen 1982
						Titanite	1800	Hiltunen 1982
A949	271312	7489.72	2496.82	Rautuvaara	Skarn	Titanite	1748	Hiltunen 1982
A840	271410	7496.80	2497.50	Kivivuopion-vaara	Monzonite	Zircon	1862±3	Hiltunen 1982
						Titanite	1783	Hiltunen 1982
A964	271311	7477.64	2491.93	Saarenpudas	Diabase	Zircon	2027±33	Hiltunen 1982
						Titanite	1820	Hiltunen 1982
A958	271312	7489.81	2496.95	Rautuvaara	Monzonite (albitite)	Zircon	1849±16	Hiltunen 1982
						Titanite	1783	Hiltunen 1982
A994	271407	7494.47	2482.14	Jaaravinsa	Keratophyre	Zircon	Archean	Hiltunen 1982
						Titanite	1793	Hiltunen 1982
A583	272307	7523.85	2488.00	Kangosjärvi	Monzonite	Zircon	1876±16	Lehtonen 1984
A838	272307	7528.50	2481.30	Kangosselkä	Paragneiss	Zircon	1832±7	Lehtonen 1984
A836	272309	7548.56	2480.27	Ylimuonio	Granodiorite (dyke)	Monazite	1790	Lehtonen 1984
						Zircon	heter.	Lehtonen 1984
A523	272309	7541.60	2487.00	Muonio	Granodiorite	Zircon	heter.	Lehtonen 1984
A569	272309	7541.60	2487.00	Muonio	Granodiorite	Zircon	heter.	Lehtonen 1984
A835	272312	7546.40	2493.48	Vuontisrova	Granodioritic gneiss	Zircon	2591±16	Lehtonen 1984
						Titanite	1845	Lehtonen 1984
A837	272312	7540.40	2497.95	Toras-Sieppi	Monzonite	Zircon	1892±6	Lehtonen 1984
						Titanite	1845	Lehtonen 1984
A967	272312	7547.30	2492.40	Kipparioja	Granodiorite	Zircon	1850±41	Lehtonen 1984
A312	272312	7547.30	2492.40	Kipparioja	Granodiorite	Zircon	heter.	Lehtonen 1984
A378	272310	7522.60	2498.87	Matala	Granodiorite	Zircon	1977±46	Lehtonen 1984
				Nivunkijärvi		Zircon	2444±96	Lehtonen 1984
A966	272310	7521.00	2499.17	Matalan-Nivunginselkä	Pegmatite granite	Zircon	1778±3	Lehtonen 1984

Kuervaara and the two titanites from A949-Rautuvaara have upper intercept ages which overlap within experimental error at c. 1795 Ma. One titanite fraction from A963 is only slightly discordant and exhibits a similar age, while another titanite with very large errors in the Pb/U ratios (arising from a poor U concentration run) is concordant at 1710 Ma. The mafic pegmatoid sample A959 (1747±14 Ma, MSWD=12) is from a coarse-grained vein cutting skarn rock and iron ore in the Rautuvaara mine.

Sample A964 (2025±19 Ma, MSWD=6.7) has been collected from a differentiated diabase dyke at Saarenpudas intruding concordantly volcanic rocks of the Kolari Formation and sample A957 represents a mafic rock occuring close by. Only the results for sample A964 were published. The unpublished preliminary results for sample A957 will be briefly discussed in the subsequent text. Sample A994 with an Archean zircon age (2738 Ma, titanite age 1793 Ma) is from a keratophyre dyke of the Siekkijoki Formation (correlates to the Kolari Formation).

ISOTOPIC STUDIES FROM THE KOLARI-MUONIO AREA

A142-Jalokoski. This sample (Table 2) represents a gabbroic intrusion (c. 35 km^2) outcropping on the banks of the Muonio river at the Finnish-Swedish border. The Jalokoski gabbro was regarded by Mikkola (1941, p. 109) as belonging to the Syenite Series (Haaparanta Suite), and is now considered part of the quartz monzodiorite association (QMD) of the Muonio-Kihlanki area (Lehtonen 1988). The composition of the igneous body varies from ultrabasic to intermediate due to differentiation (Lindroos & Henkel 1981, Hiltunen 1982). The sample was taken from a rather coarse-grained (0.2-5 mm) rock type consisting mainly of plagioclase (c. An_{45}), orthopyroxene and biotite. Accessory minerals include magnetite, apatite, epidote and locally ilmenite. Plagioclase contains secondary scapolite. Orthopyroxene is typically altered to amphibole and green biotite.

The analytical data for the zircon fractions of the Jalokoski gabbro are shown on a concordia plot in Figure 2, and define an upper intercept age of 1866±6 Ma. Only three zircon fractions have been used, as the fractions are nearly concordant and their $^{207}Pb/^{206}Pb$ ages are so similar that additional analyses were deemed unnecessary.

A746-Haisuvuoma and A747-Härkimännikkö. These samples represent a large granitoid body called

Table 2. New U-Pb age data from the Kolari-Kihlanki-Muonio area.

Sample	Map	Northing	Easting	Location	Lithology	Mineral	Age	Comment
A957	271311	7478.00	2491.90	Saarenpudas	Diabase	Zircon	1969	$^{207}Pb/^{206}Pb$ age
A582	271408	7503.54	2481.43	Kihlanki	Granite (Lina type)	Zircon	1805±10	
						Titanite	1792±5	$^{207}Pb/^{206}Pb$ age
A839	271410	7495.88	2490.56	Kiuaskero	Quartz diorite	Zircon	1860	reference line
A746	273108	7477.50	2527.00	Haisuvuoma	Quartz monzodiorite	Zircon	1882±1	$^{207}Pb/^{206}Pb$ age, average
						Titanite	1832±12	$^{207}Pb/^{206}Pb$ age
A747	273204	7496.60	2517.04	Härkimännikkö	Quartz monzonite	Zircon	1885	reference line
A142	271306	7489.15	2474.40	Jalokoski	Gabbro	Zircon	1866±6	nearly concordant

the Kallo massif (Mikkola 1941) occurring mainly in the Kurtakko and partly in the Kittilä map sheet areas. The intrusion is c. 300 km^2 in size. The granitoid intrudes concordantly the metasedimentary and volcanic rocks of the Sodankylä and Savukoski Groups. Geographically, the Kallo massif occupies a topographic low and its flat surface is only rarely exposed. Mikkola (1941) assumed that the intrusion was a differentiate of the Syenite Series. Rantataro (1988) regarded the massif as a synorogenic zonal intrusion and classified it geochemically as a calc-alkaline, I-type granitoid.

Field observations show that the various granitoid types within the Kallo massif were generated by several magma pulses. The oldest phase comprises mafic autoliths and quartz diorite in the southeastern part of the intrusive. During the main phase of intrusion, the tonalitic outer margin, felsic autoliths and the prevailing Haisuvuoma quartz monzodiorite were generated. The tonalitic margin grades into quartz monzodiorite. This phase of intrusion was followed by more felsic quartz monzonites and granodioritic pegmatites. There are two separate bodies of felsic quartz monzonite within the Kallo massif: Härkimännikkö on the northwestern and Poro-Heikinpalo on the eastern side of the intrusion. Their oblong shape and location around the massif indicate that their emplacement was possibly controlled by thrust or fault zones (cf. Rastas 1984, Väänänen 1992). Sample A746 was taken from the dominant Haisuvuoma quartz monzodiorite and sample A747 from the Härkimännikkö felsic quartz monzonite. The tonalitic margin consists mainly of plagioclase, alkali feldspar, quartz, biotite and hornblende. In the prevailing quartz monzodiorite clinopyroxene occasionally occurs in addition to the above mentioned main minerals. Titanite is the principal accessory mineral in the Kallo massif while apatite, magnetite and zircon are other accessories. The felsic quartz monzonites are porphyritic to some extent. Petrographically and geochemically, the Kallo massif resembles the granitoids of the quartz monzodiorite association (QMD) of the Haaparanta Suite in the Muonio-Kihlanki area (Väänänen1998).

The relatively large zircon grains extracted from sample A746 are mostly broken, brown, transparent fragments of originally larger crystals. Distinct growth zones and clear crystal surfaces are typical features. Zircon grains in sample A747 are relatively small and turbid with L/B c. 1.5-2. Crystal edges are blunt, and the surfaces contain many inclusions. The concordia plot for analytical data from the Kallo massif is presented in Figure 3. The age is given by three nearly

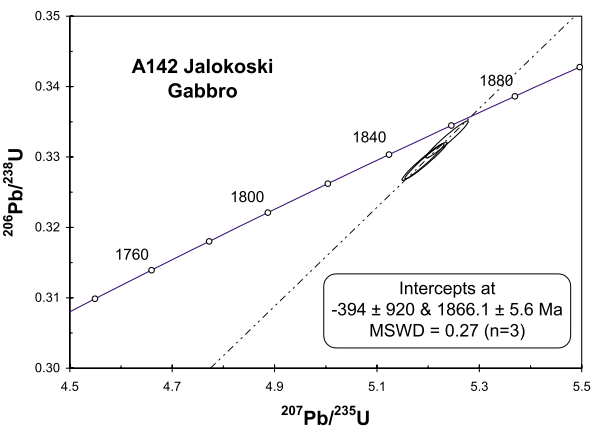

Fig. 2. Results for U-Pb determinations from zircon of the Jalokoski gabbro on a concordia diagram. The size of ellipses indicates the analytical error on a 2 -level.

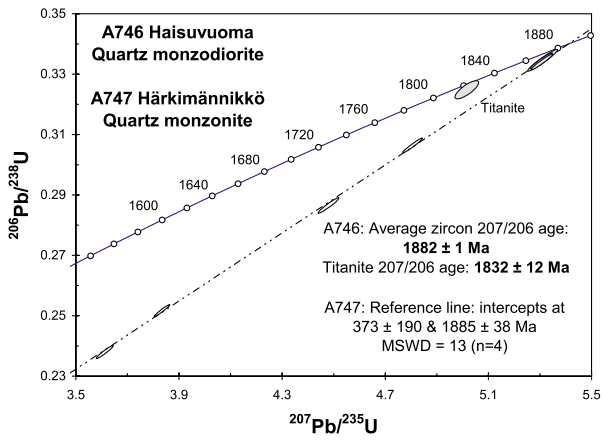

Fig. 3. Results for U-Pb mineral determinations from the Kallo massif on a concordia diagram. A746 closed symbols, A747 open symbols.

concordant zircon fractions from the Haisuvuoma quartz monzodiorite (A746) at 1882±1 Ma, while the titanite from the same rock yields an age of 1832±12 Ma. The turbid zircons from the Härkimännikkö quartz monzonite (A747) are rather discordant and exhibit a scatter in excess of analytical uncertainty (MSWD=13). Nevertheless, they define an upper intercept at 1885±38 Ma, which is in agreement with the result from A746 and supports a comagmatic origin for these rocks. The zircon and titanite ages of the Kallo massif are typical of the Haaparanta Suite in Western Lapland (cf. Tables 1 and 2).

A582-Kihlanki. The Kihlanki granite forms part of a large post-orogenic intrusion, which lies mainly on the Swedish side (so-called Lina-granite; Geijer 1931, Ödman 1957, Lindroos & Henkel 1981). In Finland, the dimensions of the Kihlanki granite are 10x15 km^2 (Lehtonen 1981). It belongs to the microcline granites of the Muonio-Kihlanki area (Lehtonen 1988), is medium-grained, massive and partly porphyritic (feldspar phenocrysts <1 cm). Country rock fragments and nebulitic mica-rich relics occur in the contact zones. The granite is clearly intrusive into its country rocks. Quartz, potassium feldspar and plagioclase are the main minerals. Biotite is the only dark mineral and forms 1-3 % of the total rock volume. Occasionally, the rock is sufficiently rich in magnetite (max. 2 %) to produce irregular anomaly patterns on the aeromagnetic map.

The zircon concentrates contained also rutile, which was removed by hand picking. The zircons from the Kihlanki granite often exhibit well developed 110-faces and are brownish in color. Large cores occur especially in big and broken crystals. The L/B ratio varies from two to more than ten. A few beautiful "micro-gem" specimens have also been seen. The data of six zircon and one titanite analyses are presented in Figure 4. The zircons define a linear trend

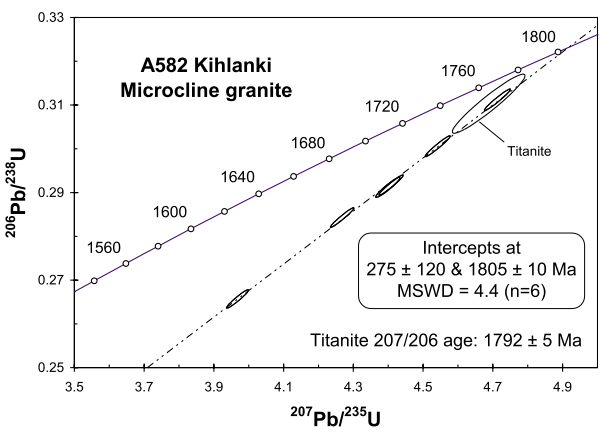

Fig. 4. Results for U-Pb mineral determinations from the Kihlanki granite on a concordia diagram.

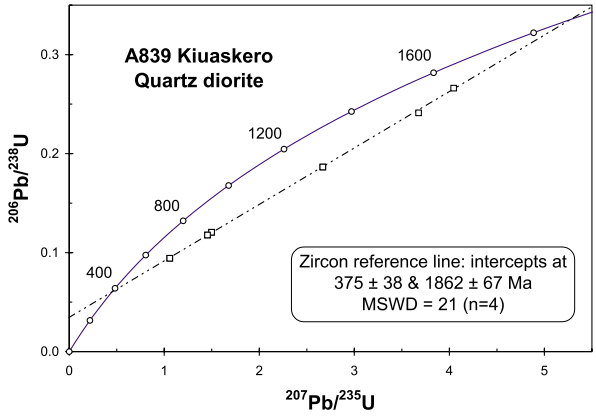

Fig. 5. Results for the Kiuaskero quartz diorite on a concordia diagram. Note that the scale is so large that displaying error ellipses is nonsensical.

with an upper intercept at 1805±10 Ma (MSWD=4.4). Although analytical errors overlap, the titanite is probably somewhat younger with a $^{207}Pb/^{206}Pb$ age of 1792±5 Ma.

A839-Kiuaskero. The dated quartz dioritic sample represents the Kiuaskero intrusion (c. 20 km^2) in the Kihlanki area c. 10 km northwest of Rautuvaara mine. The intrusion belongs to the quartz diorite-tonalite-trondhjemite association (QTT) of the Haaparanta Suite (Lehtonen 1988) and is intruded by the Kihlanki granite. The contact of the Kiuaskero intrusion with the Tapojärvi Formation in the south and west is deformed. Quartz monzodiorite inclusions occur in the intrusion, which is typically leucocratic and porphyritic; pink plagioclase phenocrysts (c. 1 mm) stand up in the finer-grained ground mass. The modal composition varies between quartz diorite and diorite.

All four zircon fractions from sample A839 have a high uranium concentration (1600-3900 ppm) and are consequently very discordant (Fig. 5). The radiation damage has probably also contributed to the apparent heterogeneity of the data (MSWD=21). The upper intercept of the line defined by the zircons is at 1862±67 Ma, which is consistent with the assumption that the Kiuaskero quartz diorite is part of the Haaparanta Suite. The two titanite analyses are rather imprecise, and the large common lead contents make them dependent on the value of the used for the common lead correction. If the two titanites are included in the linear regression, the result becomes 1865±31 Ma (MSWD=21).

A957-Saarenpudas. This sample has been collected from a road cut north of Kolari. The present authors are inclined to interpret it as a diabase similar to the rock represented by sample A964.

The sole zircon analysis from the rock is not

particularly discordant and the result of c. 1970 Ma suggests that this dyke is somewhat younger than most of the Jatulian diabases.

DISCUSSION

The oldest zircon ages obtained in the northern Kolari-Muonio area, 2591±16 and 2444±96 Ma, are from a granodioritic gneiss (A835) and a hypidiomorphic granodiorite (A378), respectively, representing the assumed Archean basement. As evidence of metamorphism appears on the U-Pb concordia plot in the form of a very high lower intercept (Lehtonen 1984), these ages could be taken as minima. Two zircon populations and heterogenous zircons with darker outer rims extracted from sample A378 also indicate metamorphic reactivation of the basement gneisses.

The zircon age of the Saarenpudas diabase (A964) is 2027±33 Ma (Hiltunen 1982) suggesting that it corresponds to the younger age group (2.0-2.05 Ga) of diabases in the Kittilä greenstone area rather than the older group having an age of c. 2.2 Ga (cf. Lehtonen et al. 1998). On the other hand, sample A957, taken close by from a diabase gives a markedly younger age, c. 1970 Ma. The meaning of this age is unknown for the present.

The zircon age distribution of 1860-1890 Ma obtained for the Haaparanta Suite (A840, A958, A746, A747, A583, A837, A836, A312, A967 A839, A142) indicates many magmatic pulses of different ages within the suite (cf. Lehtonen 1988, Rantataro 1988). This age distribution is typical of synorogenic igneous rocks (cf. Skiöld & Öhlander 1989, Hanski et al. 2001, *this volume*).

The zircon ages obtained from the microcline granites vary from 1778±3 to 1805±10 Ma (A966, A582) indicating postorogenic crystallization. This event belongs to the same stage of felsic magmatism that produced the Central Lapland Granite Complex (Lauerma 1982) and the microcline and migmatite-granites (Lina granites) in northern Sweden (Skiöld 1981, 1982, 1988, Skiöld & Öhlander 1989, Bergman & Skiöld 1998). These zircon ages are consistent with the field observations showing that the granites belong to the youngest rocks in the study area.

The youngest zircon age of 1748±7 Ma in the Kolari-Muonio area is from a mafic pegmatoid (A959) cutting skarn rock and iron ore in the Rautuvaara mine. No regional magmatic event explaining the origin of the mafic pegmatoid is presently known (see Hiltunen 1982).

The zircon ages of the keratophyre layers or dykes (A994) are interesting but problematic. Hiltunen (1982, p. 53) writes: "One limitation on the interpretation of the zircon data is that only one fraction from two samples was available (2 mg and 6 mg). A tentative interpretation would be that the two-point chord joining them has an upper intersection equivalent to an age of 2738 Ma, or else that they may not be cogenetic and thus fall on diffusion lines corresponding to ages of 2410 Ma and 2505 Ma. No detailed conclusions can be drawn, but their distribution suggests an Archean age of crystallization close to a minimum age of 2400-2500 Ma, or a maximum estimate of 2738 Ma". Possibly these results can be taken as evidence for inherited Archean zircons (cf. Perttunen & Vaasjoki, 2001, *this volume*). 2366 and 1968 Ma old zircons from samples A569 and A523 indicate assimilation of older crustal material into the tonalitic melts of the QTT association of the Haaparanta Suite (Lehtonen 1988).

The titanite age, 1845 Ma, of the granodioritic gneisses (A835) and the titanite age distribution of 1885-1800 Ma of diabases (A964), skarn rocks (A949, A963), keratophyre (A994) and synorogenic rocks (A840, A958, A837, A746) reflect metamorphic resetting of the Archean basement and the Paleoproterozoic rocks in the Kolari-Muonio area. The exceptionally young titanite age, 1748 Ma, for a single crystal of titanite from skarn rock (A949) in the Rautuvaara mine, is of the same order as that mentioned before for mafic pegmatoid (A959). No tectonometamorphic or magmatic event explaining this young titanite is known.

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Appendix I. U-Pb mineral ages from the Muonio area, western Lapland.

Fraction	Weigth mg	U conc. ppm	Pb conc. ppm	$^{206}Pb/^{204}Pb$ measured	$^{208}Pb/^{206}Pb$ radiogenic	$^{206}Pb/^{238}U$	±2 SE (%)	$^{207}Pb/^{235}U$	±2 SE (%)	$^{207}Pb/^{206}Pb$	±2 SE (%)	Corr.	$T_{206/238}$	$T_{207/235}$	$T_{207/206}$
A142-Jalokoski gabbro															
A. +4.6	20.5	138.1	53.8	3050	0.2380	0.3293	0.65	5.190	0.65	0.1143	0.15	0.97	1835	1851	1869
B. +4.6/HF	20.6	127.6	49.4	7943	0.2447	0.3295	0.65	5.196	0.65	0.1144	0.15	0.97	1836	1851	1870
C. +4.2/M/HF	15.2	124.3	48.7	7329	0.2476	0.3325	0.65	5.237	0.65	0.1142	0.15	0.97	1850	1858	1868
A746-Haisuvuoma quartz monzodiorite															
A. +4.6	25.7	444.9	161.7	19528	0.1451	0.3347	0.89	5.314	0.89	0.1152	0.15	0.96	1860	1871	1882
B. 4.2-4.6	21.0	718.6	263.2	14771	0.1535	0.3346	0.65	5.309	0.65	0.1151	0.15	0.97	1860	1870	1881
C. 4.0-4.2	21.8	1101.1	404.7	19793	0.1602	0.3342	0.65	5.313	0.65	0.1153	0.15	0.97	1858	1870	1884
D. Titanite	31.0	110.1	40.0	988	0.1231	0.3250	0.75	5.018	0.76	0.1120	0.65	0.79	1813	1822	1832
A747-Härkimännikkö quartz monzonite															
A. +4.6	20.7	635.8	198.4	5336	0.1156	0.2865	0.74	4.480	0.74	0.1134	0.15	0.96	1624	1727	1854
B. 4.2-4.6	21.9	983.0	252.4	6512	0.1351	0.2380	0.75	3.612	0.75	0.1101	0.15	0.96	1376	1552	1801
C. +4.6/abr 3 h	10.2	600.8	200.2	15156	0.1468	0.3064	0.65	4.806	0.65	0.1138	0.15	0.97	1723	1789	1860
D. 4.2-4.6/abr 3 h	7.9	957.3	259.5	10461	0.1363	0.2517	0.65	3.834	0.65	0.1105	0.15	0.97	1446	1599	1807
A582-Kihlanki microcline granite (Lina type)															
A. +4.2/+200	22.4	595.3	187.1	5625	0.1367	0.2917	0.65	4.408	0.65	0.1096	0.15	0.97	1650	1713	1793
B. Titanite	40.6	99.0	74.4	189	1.3852	0.3103	1.78	4.688	1.82	0.1096	0.35	0.95	1742	1765	1792
C. 4.0-4.2/+150/abr 2 h	4.4	875.2	277.8	2218	0.1733	0.2843	0.65	4.268	0.65	0.1089	0.15	0.97	1612	1687	1781
D. +4.2/+200/abr 3 h	6.2	620.3	206.2	2223	0.1599	0.3006	0.65	4.542	0.65	0.1096	0.15	0.97	1694	1738	1792
E. +4.2/-200/abr 3 h	3.2	626.2	206.0	1270	0.1762	0.2912	0.65	4.399	0.65	0.1096	0.21	0.95	1647	1712	1792
F. +4.3/+200/abr 2 h	5.3	582.8	199.6	2870	0.1573	0.3112	0.65	4.713	0.65	0.1099	0.15	0.97	1746	1769	1797
G. 4.0-4.2/+200/abr 2 h	4.1	889.2	271.2	859	0.1842	0.2657	0.65	3.968	0.65	0.1083	0.16	0.96	1519	1627	1771
A893-Kiuaskero quartz diorite (porphyritic)															
A. +4.2	3.9	1598.0	303.2	2696	0.0698	0.1865	0.65	2.668	0.65	0.1038	0.15	0.97	1102	1319	1692
B. 3.8-4.0/+100	6.7	2894.7	375.2	1081	0.1284	0.1206	0.65	1.499	0.65	0.0901	0.15	0.97	734	929	1428
C. 3.6-3.8/+100	3.6	3876.1	391.1	1319	0.1385	0.0943	0.65	1.058	0.65	0.0813	0.15	0.97	580	732	1229
D. Titanite	33.8	78.5	37.2	150	0.6473	0.2662	0.65	4.045	0.80	0.1102	0.42	0.85	1521	1643	1802
E. 3.8-4.0/+100/abr 1 h	2.1	2926.1	355.9	2934	0.1001	0.1178	0.65	1.455	0.65	0.0896	0.19	0.95	717	912	1416
F. Titanite	9.4	74.4	38.3	97	0.8164	0.2412	0.70	3.676	0.99	0.1105	0.61	0.79	1393	1566	1808
A957-Saarenpudas mafic volcanic rock															
A. 4.0-4.2	9.6	220.4	120.4	1522	0.7158	0.3405	0.65	5.666	0.65	0.1207	0.20	0.96	1889	1929	1966

Isotopic ratios corrected for fractionation, blank and age related common lead
Corr. = Error correlation for $^{207}Pb/^{235}U$ vs. $^{206}Pb/^{238}U$ ratios.
SE = Standard error in the mean

Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
Edited by Matti Vaasjoki
Geological Survey of Finland, Special Paper 33, 95–141, 2001.

U-Pb ISOTOPIC STUDIES ON THE KITTILÄ GREENSTONE AREA, CENTRAL LAPLAND, FINLAND

by
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Rastas, P, Huhma, H., Hanski, E., Lehtonen, M.I., Härkönen, I.†, Kortelainen, V., Mänttäri, I. & Paakkola, J. 2001. U-Pb isotopic studies on the Kittilä greenstone area, central Lapland, Finland. *Geological Survey of Finland, Special Paper 33,* 95-141. 31 figures, one table and 3 appendices.

This paper reports 227 previously unpublished U-Pb isotopic analyses on zircon, titanite and monazite from 51 samples in the Kittilä greenstone area, Central Finnish Lapland. These consist of metadiabases and metagabbros (17 samples), felsic volcanic rocks (11), granitoids (7), mainly quartzitic metasediments (10) and cobbles from conglomerates (6 samples).

The reported U-Pb age determinations on intrusive mafic rocks fall into three groups indicating ages of c. 2.2 Ga, 2.05 Ga and 2.0 Ga. The oldest group represents the Haaskalehto-type mafic differentiated intrusions, while the younger ones are collectively called metadiabases and gabbros. The 2.2 Ga Haaskalehto-type intrusions provide a minimum age for the metasediments of the *Sodankylä Group*. The 2.0-2.06 Ga metadiabases and gabbros in turn intrude rocks of the overlying *Savukoski Group.*

The felsic volcanic rocks in the Kittilä greenstone area yield both Archean and Proterozoic zircon ages. The rocks with Archean ages (2.75-2.72 Ga) are strongly altered dacitic and rhyolitic tuffs, but the Archean age is not fully supported by the geological interpretation of the volcanites and part of the zircons may be xenocrystic in origin. These rocks are older than the intruding 2.2 Ga old Haaskalehto-type dykes. According to the radiometric data, the Proterozoic porphyries can be divided into three main groups (2.01-2.02 Ga, c. 1.92 Ga and 1.88 Ga) having different geochemical and petrographical features. The porphyries of the oldest group are quartz- and plagioclase-phyric dacites and rhyolites. The field characteristics suggest they are coeval with the associated mafic volcanic rocks of the *Kittilä Group*. This is consistent with the age of a 1.92 Ga porphyry dyke, which cuts the metavolcanic rocks of the Kittilä Group. The 1.88 Ga porphyritic felsic metavolcanic rocks occur stratigraphically in the lower part of the *Lainio Group*. They are chemically calc-alkaline trachytes with a remarkably high content of K_2O (average 8.9 wt %). U-Pb ages show that they are roughly coeval with the Haaparanta Suite quartz monzonites in western Lapland.

The dated detrital minerals from quartzites belonging to the Proterozoic *Sodankylä Group* indicate an Archean provenance (~2.8 Ga). A much younger age of deposition is indicated for the *Lainio Group* conglomerates by fragments of felsic porphyry and deformed granitoid, which are 1873±11 Ma and 1888±22 Ma old, respectively. Relatively young sources are also indicated for the *Kumpu Group* quartzites by a 1928±6 Ma age obtained for a porphyry cobble from a conglomerate at Mantovaara. The dated monazites of the Kumpu Group are 1.89 Ga and 1.90 Ga old. The analyses show that a Jatulian or any older origin can be excluded for the Lainio and Kumpu Groups.

The age of the supracrustal rocks is further constrained by intruding granitoids. The metavolcanic rocks of the Kittilä Group are cut by the deformed Ruoppapalo granodiorite, which has been dated at 1914±3 Ma. The post-tectonic Tepasto granite within the Hetta Granite Complex is dated at 1802±10 Ma. Several analyses on titanite and monazite provide also ages of c. 1.8 Ga, which is considered a regional alteration phase with significant fluid activity. This has also affected the U-Pb systematics of zircon in some gabbroic rocks.

Key words (GeoRef Thesaurus, AGI): absolute age, U/Pb, zircon, monazite, titanite, metavolcanic rocks, metasedimentary rocks, plutonic rocks, Paleoproterozoic, Archean, Kittilä, Lappi Province, Finland

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INTRODUCTION

Since the beginning of the 20th century, Central Finnish Lapland and especially the Kittilä area has been a subject of active geological research and exploration. Although the detailed geological picture of this area has changed since the early works of Hackman (1927), Sederholm (1932) and Mikkola (1937, 1941), the overall geological interpretation has remained essentially the same as established in the first half of the 20th century. Although the relative ages of the major lithostratigraphical units have been mostly valid, opinions have differed on the chronostratigraphical position of these units (Mäkelä 1968, Gaál et al. 1978, Paakkola 1971, Saverikko 1987, Silvennoinen et al. 1980, Silvennoinen 1985, Silvennoinen et al. 1986, Lehtonen et al. 1992, 1998, Sorjonen-Ward et al. 1992). In a recent, comprehensive account on the rocks of Central Lapland, Lehtonen et al. (1998) described a lithostratigraphical scheme in which the Paleoproterozoic rocks lying on the Archean basement were divided into seven groups. These groups were shown to represent a supracrustal sequence deposited during a long period of time extending from the Archean-Proterozoic boundary to less than 1.9 Ga.

Due to the lack of suitable minerals for dating, the precise ages of mafic volcanic rocks in Central Lapland are still poorly known. Encouraging results have recently been obtained on these rocks by using the Sm-Nd method (Huhma et al. 1996, Hanski et al. 1997). Over the last 25 years, the zircon U-Pb method has been successfully applied to various intrusive rocks and felsic extrusive rocks by Dr. O. Kouvo and his colleagues at the Geological Survey of Finland (GSF Annual Reports 1972, 1975, GSF Petrological Department Annual Reports 1982-1985), and other workers including Kröner et al. (1981), Patchett et al. (1981), Hiltunen (1982), Kouvo et al. (1983), Jahn et al. (1984), and Huhma (1986). Preliminary results of the U-Pb isotopic work have been published in many other papers, but actual analytical data have so far been insufficiently reported (Rastas 1980, Lehtonen et al. 1992, 1998, Räsänen et al. 1995, Hanski et al. 1997). In this paper, we document U-Pb age results and isotopic data for zircons, titanites and monazites separated from various rock types from the Kittilä greenstone area, the number of samples totaling 51. The oldest measurements are from the 1970's, most data from the 1980's and the early 1990's, but some

recent analyses are also included. For many samples analyses have been made in several stages. Both magmatic, metamorphic and detrital minerals have been analyzed and the samples include mafic intrusive rocks (layered sills, gabbros, diabases), felsic porphyries, granitoids, quartzites and cobbles in conglomerates.

The analytical techniques have varied considerably over the years and range from the borax fusion technique to the present-day microanalyses. These have been dealt with in more detail in the summary of analytical techniques (Vaasjoki 2001, *this volume*). All age calculations and concordia diagrams have been made using the Isoplot/Ex program by Ludwig (1998).

The main lithostratigraphical features of the Kittilä greenstone area and the locations of the dated samples are shown in Figure 1. The sample locations, rock types and age results are summarized in Table 1 and detailed U-Pb analytical data on zircon, titanite and monazite are given in Appendices 1-3.

GENERAL GEOLOGICAL SETTING

The Kittilä greenstone area forms part of the Central Lapland Greenstone Belt and consists of Paleoproterozoic supracrustal rocks overlying the Archean basement. In the east, the area is bordered by the Sodankylä schist area, and elsewhere mainly by various granitoids including the Hetta Complex in the north, the Haaparanta Suite in the west, and the Central Lapland Granite Complex in the south (Fig. 1).

The Paleoproterozoic supracrustal rocks of Central Lapland have been divided into seven lithostratigraphical groups (Fig. 1), which are, from oldest to youngest, the Salla, Onkamo, Sodankylä, Savukoski, Kittilä, Lainio and Kumpu Groups (Lehtonen et al. 1998, see also Fig. 5 in Hanski 2001, *this volume*). The two oldest of these, the Salla and Onkamo Groups, comprise metavolcanic rocks representing products of an intracratonic rift environment. In the NE part of the study area of this report, the Salla Group consists of andesitic and dacitic lava flows as well as rhyolitic pyroclastic rocks, which are found deposited around the Archean Tojottomanselkä basement gneiss dome (Manninen et al. 2001, *this volume*). The Salla Group is also exposed in the Rajalompolo area at the border of the Kittilä and Sodankylä parishes. The Onkamo Group comprises pyroclastic ultramafic rocks and mafic to intermediate amygdaloidal lava flows. At Möykkelmä in western Sodankylä, these rocks form a 250 m thick unit deposited directly on the Archean basement (Räsänen et al. 1989).

Metasedimentary and minor metavolcanic rocks of the Sodankylä and Savukoski Groups were deposited in a deepening depositional basin in a cratonic or craton margin setting. The Sodankylä Group consists of arkosic quartzites, orthoquartzites and mica schists, which occur in the marginal parts of the Kittilä greenstone area. In the Sodankylä Group there are felsic metavolcanic rocks of uncertain ages. Progression from a shallow to a deeper-water sedimentational environment is manifested by the transition from the Sodankylä Group to the Savukoski Group. The latter comprises fine-grained metasedimentary and pyroclastic rocks, such as phyllites, black schists, tuffs and tuffites. Tholeiitic basalts as well as picritic and komatiitic volcanic rocks also erupted in this sedimentary basin.

The contact between the Savukoski and Kittilä Groups is characterized by thrust and fault zones. A number of serpentinite bodies occur within the thrust zones along the eastern edge of the Kittilä Group and have been interpreted as parts of dismembered ophiolites (Hanski 1997). The allochthonous Kittilä Group was obducted to its present position before the intrusion of the 1.91 Ga Ruoppapalo granodiorite (see p. 125 and Hanski et al. 1998). The Kittilä Group is mainly volcanic in origin and is regarded as representing ancient oceanic crust. It comprises mainly tholeiitic metavolcanic rocks, fine-grained marine metasediments and volcanoclastic rocks as well as chemically precipitated metasediments. Quartzose mica schists at Pyhäjärvi, in the northwestern part of the Kittilä greenstone area, represent the uppermost unit in the lithostratigraphy of the Kittilä Group.

The Lainio and Kumpu Groups lie disconformably on rocks of the Sodankylä, Savukoski and Kittilä Groups. They consist mainly of coarse-clastic metasediments. The common purplish color of the conglomerates is a diagnostic feature of the Kumpu Group metasediments. In addition to metasediments, there exist minor calc-alkaline metavolcanic rocks in the lower part of the Lainio Group. The rocks were deposited in a collisional margin environment after the obduction of the Kittilä Group.

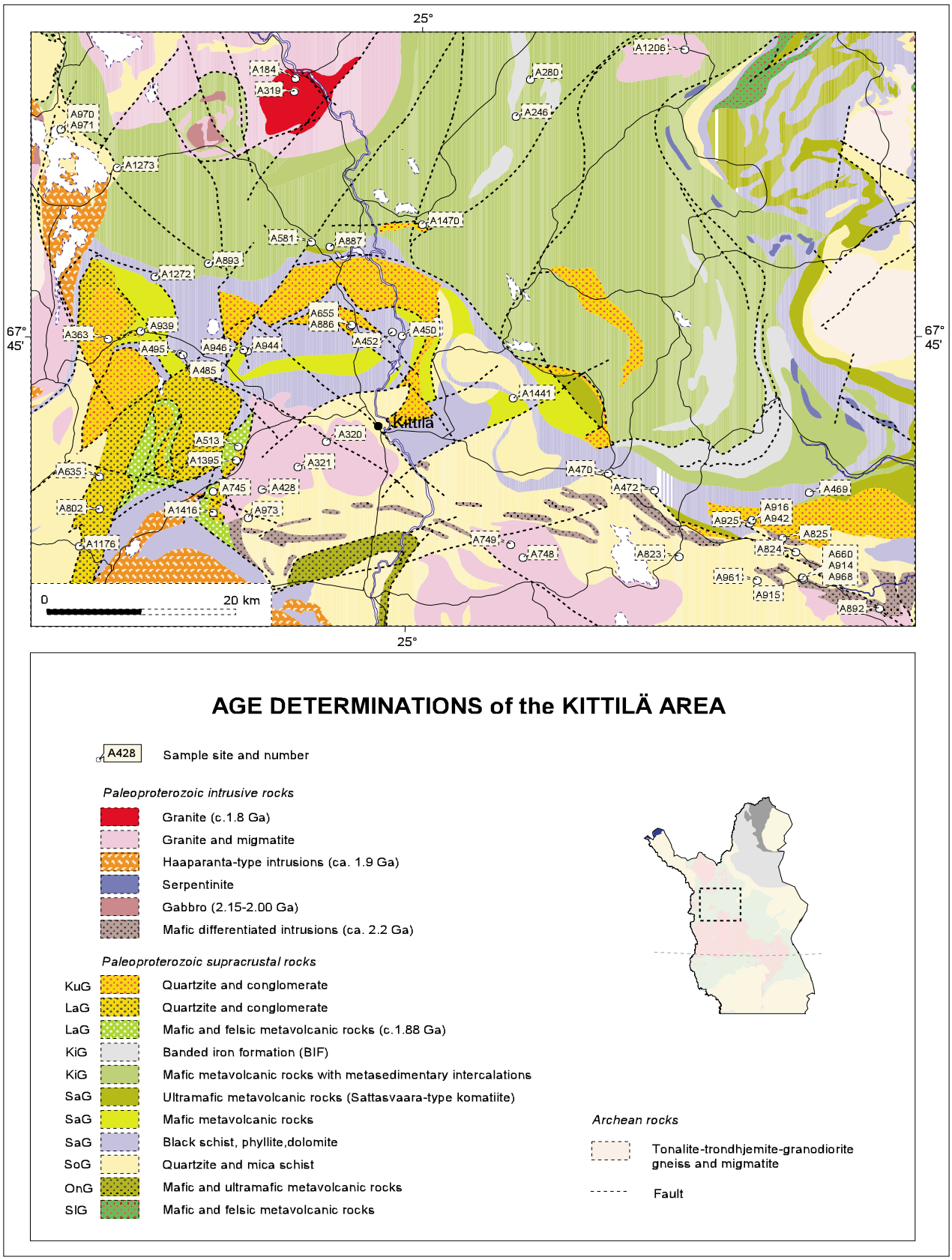

Fig. 1. Geological map of the Kittilä greenstone area with sampling sites. The lithostratigraphical groups from oldest to youngest are shown in the legend: SlG= Salla Group, OnG= Onkamo Group, SoG= Sodankylä Group, SaG= Savukoski Group, KiG= Kittilä Group, LaG= Lainio Group, KuG= Kumpu Group.

Table 1. U-Pb mineral age results for rocks from the Kittilä schist area.

Sample	Map	Northing	Easting	Location	Rock type	Mineral	Age	Comment
Igneous rocks								
A246	274305	7542.94	2551.91	Nyssäkoski Kittilä	Porphyry (Quartz-feldspar)	Zircon	1919±8	
A280	274306	7547.13	2553.22	Kapsajoki Kittilä	Porphyry (Quartz-feldspar; lava)	Zircon	2017±8	
A450	273403	7517.85	2541.06	Riikonkoski	Diabase (Meta-)	Zircon	2060±8	
A469	371211	7501.45	3457.30	Heinälammintievat Sodankylä	Diabase (Meta-)	Zircon	2052±7	
A470	371206	7503.55	3436.05	Eksymäselkä Kittilä	Gabbro (Carbonate-albite rock)	Zircon	2214±4	1 discordant fraction excluded
A472	371208	7501.71	3440.89	Palttina Kittilä	Gabbro	Zircon	2137±28	MSWD=29; 2 titanites close to discordia
A513	273208	7504.66	2524.35	Latvajärvi Kittilä	Rhyolite	Zircon	-	heterogeneous, cf. A745
A581	274110	7527.85	2530.97	Veikasenmaa Kittilä	Porphyry (Quartz-, dyke)	Zircon	2014±8	
A655	273212	7518.68	2535.52	Sätkänävaara Kittilä	Diabase (Albite-)	Zircon	2046±9	
A660	371210	7491.80	3456.70	Honkavaara Sodankylä	Volcanic rock (intermediate)	Zircon	2721±9	Titanite 7/6 c. 2200 Ma
A745	273207	7499.46	2521.86	Murtomaa Kittilä	Dacite (below Kumpu)	Zircon	1875±7	NC Monazite 7/6 age 1796±6 Ma
A823	371207	7494.25	3443.50	Lehtovaara Kittilä	Diabase (albite-, in quartzite)	Zircon	2210	discordant, plots on A825 discordia
A824	371210	7494.80	3455.85	Laitavaara Sodankylä	Albitite	Zircon	-	discordant, heterogeneous data
A825	371210	7496.35	3454.45	Sivakkavaara Sodankylä	Carbonate rock (Albite-)	Zircon	2213±14	
A887	274110	7527.42	2532.93	Yräjärvi Kittilä	Porphyry (Quartz-feldspar)	Zircon	2015±5	
A893	274107	7524.95	2520.18	Kiimarova Kittilä	Porphyry (felsic)	Zircon	2012±3	average 7/6 age, nearly concordant, bunched
A914	371210	7491.94	3456.54	Honkavaara I Sodankylä	Dacite	Zircon	2746±14	MSWD=23
A915	371210	7491.72	3456.70	Honkavaara II Sodankylä	Diabase (Meta-)	Zircon	2203±9	Tit 2204±5 Ma
A916	371210	7498.38	3451.26	Isolaki (Värttiövaara)	Diabase (Meta-)	Zircon	2210±4	
A944	273209	7515.36	2524.78	Home-Jolhikko Kittilä	Diabase (Gabbro-)	Zircon	2007±8	heterogeneous, also younger fractions
A946	273209	7515.49	2524.38	Home-Jolhikko Kittilä	Gabbro (Pegmatoid)	Zircon	2003±4	2 older fractions, 2007±6 Ma
A968	371210	7491.90	3456.60	Honkavaara III Sodankylä	Dacite	Zircon	2724±14	
A973	273207	7496.74	2525.75	Venejoki Kittilä	Rhyolite (Quartz keratophyre)	Titanite	1858±10	heterogeneous Archean zircons
A1272	274104	7523.20	2514.62	Pittarova Kittilä	Gabbro (Diabase)	Zircon	1995±9	
A1273	274105	7535.16	2510.03	Kulkujärvi Kittilä	Gabbro	Zircon	1986±10	fine fractions. 1888±30 Ma
A1395	273208	7503.12	2524.19	Latvajärvi Kittilä	Diabase (Albite-)	Zircon	1988	reference line
A1441	273406	7511.48	2553.12	Palovaara Kittilä	Lamprophyre	Monazite	1771±8	7/6 age, nearly concordanr
Metasedimentary rocks								
A363	273203	7516.00	2510.00	Outakero Kittilä/Muonio	Quartzite (Kumpu)	Monazite	1797±5	zircon heterogeneous, >2.4
A495	273206	7514.63	2517.77	Vasalampi	Quartzite	Monazite	1817±5	Archean zircon
A635a	273202	7500.60	2509.85	Kellostapuli	Conglomerate (felsic pebbles)	Zircon	-	discordant zircons
A635b	273202	7500.60	2509.85	Kellostapuli	Conglomerate (felsic pebbles)	Zircon	1888±22	MSWD=25
A802	273201	7497.00	2510.00	Ylläs	Quartzite	Zircon	-	hetrogeneous zircon, >2400 Ma
A886	273212	7518.82	2535.64	Sätkänävaara Kittilä	Quartzite (Kumpu)	Monazite	1905±5	3 zircons close to diabase A655 discordia
A925	371210	7498.00	3451.00	Värttiövaara Sodankylä	Quartzite	Monazite	1891±3	1 zircon <2200 Ma
A939a	273206	7517.02	2513.40	Linkupalo Kittilä	Conglomerate (Qtz-Fsp porphyry pebble)	Zircon	-	7/6 ages: 1950 and 2050 Ma
A939b	273206	7517.02	2513.40	Linkupalo Kittilä	Conglomerate (Qtz-Fsp porphyry pebble)	Zircon	1890	reference line, 1 fraction older
A942	371210	7498.33	3451.20	Värttiövaara Sodankylä	Quartzite (Kumpu)	Zircon	-	Nordsim: some Archean; sedimentation <1.89 Ga

Table 1. continued

Sample	Map	Northing	Easting	Location	Rock type	Mineral	Age	Comment
Metasedimentary rocks (continued)								
A961	371210	7491.60	3451.76	Virttiövaara Sodankylä	Quartzite (green)	Zircon	2801±10	all zircons Archean
A970	274102	7539.20	2503.90	Keimiöniemi Muonio	Schist	Zircon	-	heterogeneous Proterozoic >2150 Ma zircons
A971	274102	7539.16	2503.87	Keimiöniemi Muonio	Schist	Titanite	1792±5	2 other titanites deviate, 1 older zircon fraction
A1176	273201	7492.74	2508.10	Hyyverova Kittilä	Quartzite (Sillimanite)	Zircon	-	heterogeneous discordant zircons
A1416	273207	7497.13	2511.25	Vesikkovaara	Conglomerate (porphyry clasts)	Zircon	1873±11	MSWD=5
A1470	274302	7530.40	2542.60	Mantovaara Kittilä	Conglomerate (porphyry clast)	Zircon	1928±6	
Granitoid rocks								
A184	274109	7546.05	2528.38	Kotivaara Kittilä	Granite (Tepasto)	Zircon	1802±10	nearly coeval titanite
A319	274109	7544.55	2528.30	Hanhivaara Kittilä	Granite (Tepasto)	Zircon	-	discordant, close to A184 discordia
A320	273211	7505.65	2533.62	Isovaara Kittilä	Granite (Kittilä)	Zircon	-	2 discordant >2000 Ma zircon fractions
A321	273211	7502.72	2530.76	Hangasselkä Kittilä	Granite (Kittilä)	Zircon	-	zircon heterogeneous, discordant
A428	273208	7500.00	2527.10	Lappalaislampi Kittilä	Granite (Kittilä)	Titanite	1783	discordant; cf. A320 and A321, E-Nd=-10
A748	273404C	7493.75	2555.02	Iso-Nilipää Kittilä	Granite	Zircon	2136±5	1 discordant fraction excluded; E-Nd -2.5
A749	273404B	7495.09	2553.65	Pikku Nilipää Kittilä	Granite	Zircon	-	discordant; 7/6>2100 Ma
A1206	372207	7550.95	3444.10	Ruoppapalo Kittilä	Granodiorite	Zircon	1914±3	

U-Pb STUDIES ON MAFIC ROCKS

The mafic rocks of the Kittilä greenstone area form a broad range of rocks from gabbros to metadiabases and supracrustal rocks. Large c. 2.44 Ga mafic layered intrusions belong to the oldest major Paleoproterozoic magmatism in the Fennoscandian Shield (Alapieti et al. 1990, Amelin et al. 1995). In Finnish Lapland this age group includes the Koitelainen and Akanvaara intrusions (Huhma et al. 1996, Mutanen 1997, Mutanen & Huhma 2001, *this volume*). Isotopic information available from other mafic intrusive rocks dealt with in this section suggests younger ages. Here we report U-Pb ages for 17 mafic rock samples, which generally are coarse-grained varieties of metadiabases. Although light-colored, albite-rich varieties tend to contain significant amounts of zircon, it is generally noticed that zircon analyses in more dark-colored samples are less discordant.

The dated rocks are metadiabases and gabbros known earlier commonly as "albite diabases" (e.g., Meriläinen 1961), but we prefer to abandon this name because of its ambiguous nature. Due to their distinct character in terms of chemical composition, general appearance and age, the Haaskalehto-type mafic intrusions were recently separated from other diabases of the study area into a group of their own (Lehtonen et al. 1998). These c. 2.2 Ga intrusions are similar to the layered mafic sills assigned to the gabbro-wehrlite association in eastern Finland (Hanski 1986, 1987). Typical features of these intrusions include strong differentiation resulting in a rock series varying from amphibole-bearing peridotites through clinopyroxenites to magnetite gabbros. In this report, diabases younger than the Haaskalehto-type intrusions are grouped under the name "2.0-2.06 Ga metadiabases and gabbros".

2.2 Ga Haaskalehto-type mafic intrusions

The Haaskalehto-type mafic intrusions have their major occurrence in the Jeesiö River valley between the communal centers of Kittilä and Sodankylä. They intrude the sedimentary rocks of the Sodankylä Group.

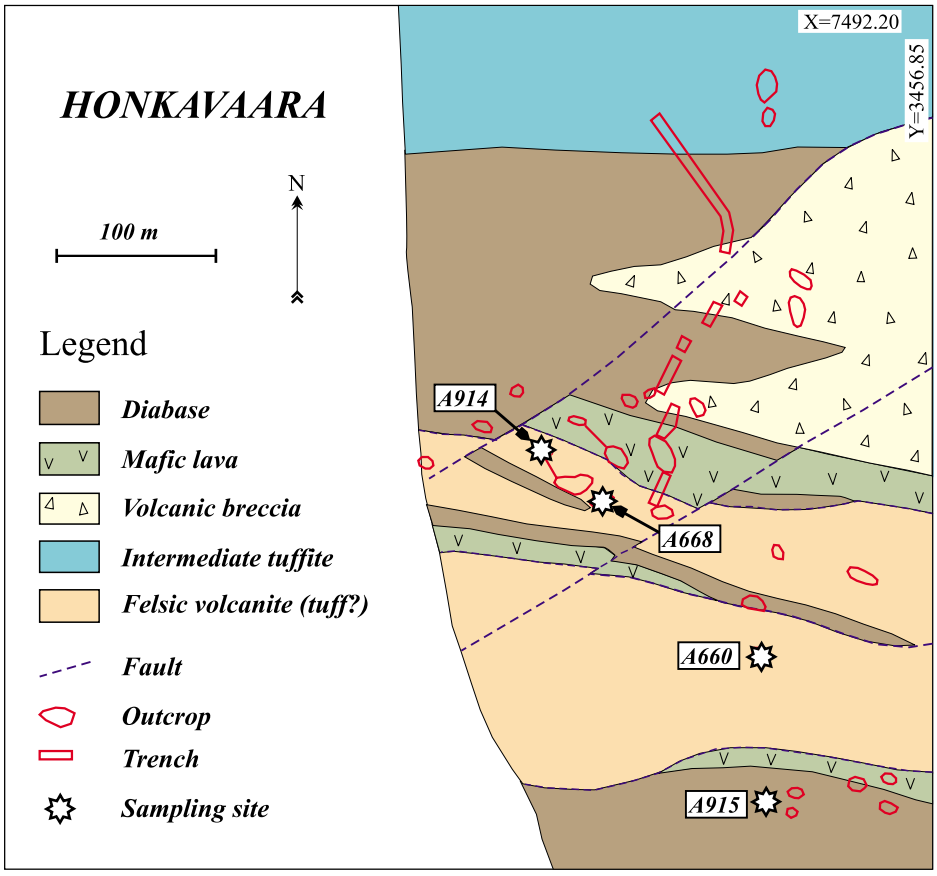

Fig. 2. Local geology of the Honkavaara area, Sodankylä.

The intrusions form typically a cumulate series varying from ultramafic to felsic in modal composition and have thicknesses of 50-1000 m. Due to their high content of magnetite, the Haaskalehto-type mafic intrusions are easily traced on the aeromagnetic maps.

A915 Honkavaara. The geology of the Honkavaara area has been described by Kallio et al. (1980), Eilu (1994) and Lehtonen et al. (1998). In this area, situated about 30 km west from Sodankylä, Haaskalehto-type sills and dykes intrude mafic and felsic supracrustal rocks of the Sodankylä Group (Fig. 2). Archean zircons have been found in these felsic rocks (see later). Most of the rocks at Honkavaara are strongly albitized.

Sample A915 was taken from an E-W-trending dyke in the southern part of the Honkavaara area. Due to poor exposure, the exact thickness of the dyke is unknown, but data from a magnetic ground survey suggest an estimate of c. 200 m. The dated metadiabase is massive, fairly coarse-grained and light greenish to pinkish in color. Roundish, light green epidote aggregates, 5-10 cm in diameter, are common. Albite and tremolite-actinolite are the main minerals together with epidote, magnetite and titanite.

A general characteristic of zircon crystals from these metadiabases is their short, stubby prismatic form and a turbid appearance. In most grains the translucent and red brown inner part tends to be surrounded by a turbid mantle. Five U-Pb zircon analyses form a linear trend, and if one slightly deviating fraction (C) is omitted the intercepts become 2203±9 Ma and 353±19 Ma with an MSWD of 1.9 (Fig. 3). The practically concordant titanite gives a $^{207}Pb/^{206}Pb$ age of 2204±3 Ma.

A916 Värttiövaara (Isolaki). At Värttiövaara, about 8 km NW of Honkavaara, a Haaskalehto-type metadiabase is enclosed by quartzites and conglomerates of the Kumpu Group (Lehtonen et al. 1998). The contact between the dyke and the quartzite is sheared and hence the age relationship remains unclear in the field (Fig. 4). Two samples (A916a and A916b) were taken for age determination from the same locality.

The Isolaki dyke is an elliptical, differentiated body of approximately 30x100 m in size with its northern part being ultramafic pyroxene cumulate (Eilu 1994), while the main body of the dyke is mafic. The dated sample was taken from a coarse pegmatoid pocket in the mafic part of the metadiabase. The dyke is subophitic in texture with albite, tremolite-actinolite, epidote, magnetite and titanite as its main constituents.

The heavy fraction contains zircon and abundant rutile. Zircon generally occurs as turbid, reddish brown fragments. The small-grained heavy fraction (d>4.3 g/cm^3), however, contains mainly transparent grains. Six zircon analyses define a line with an upper intercept at 2210±4 Ma and a lower intercept at 203±16

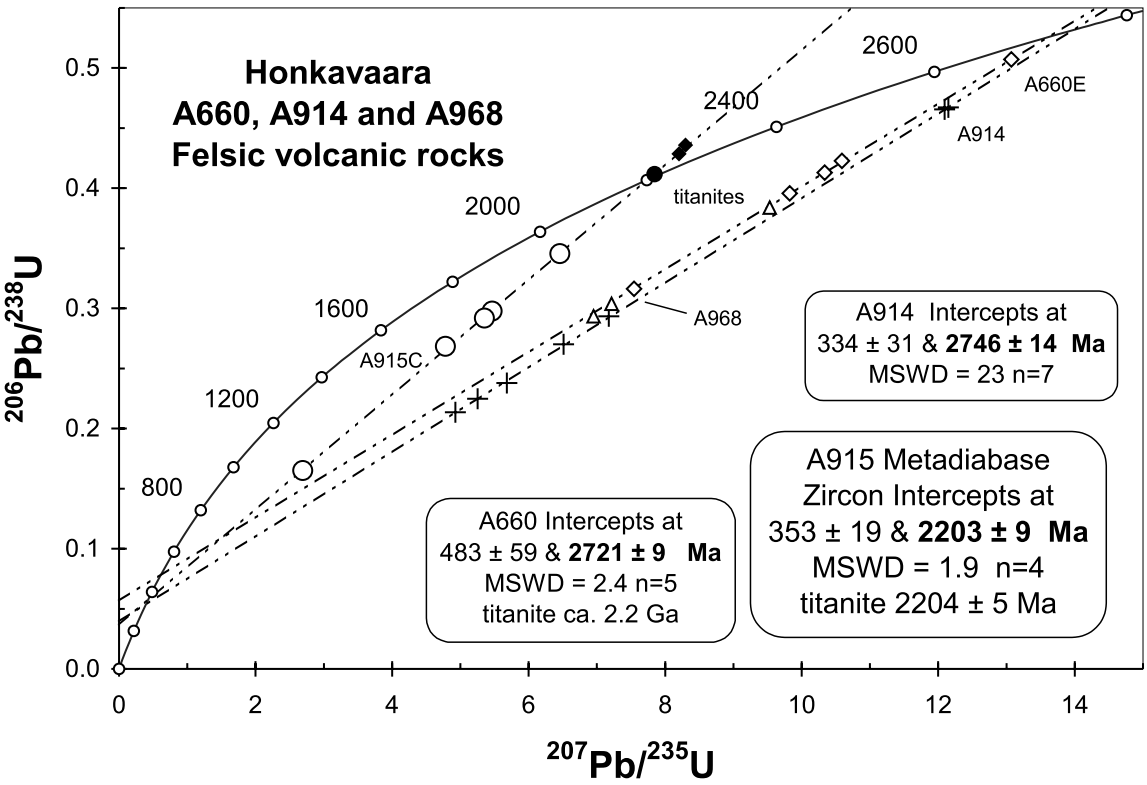

Fig. 3. U-Pb concordia diagram of zircons and titanites (solid symbols) from metadiabase and metavolcanic rocks in Honkavaara.

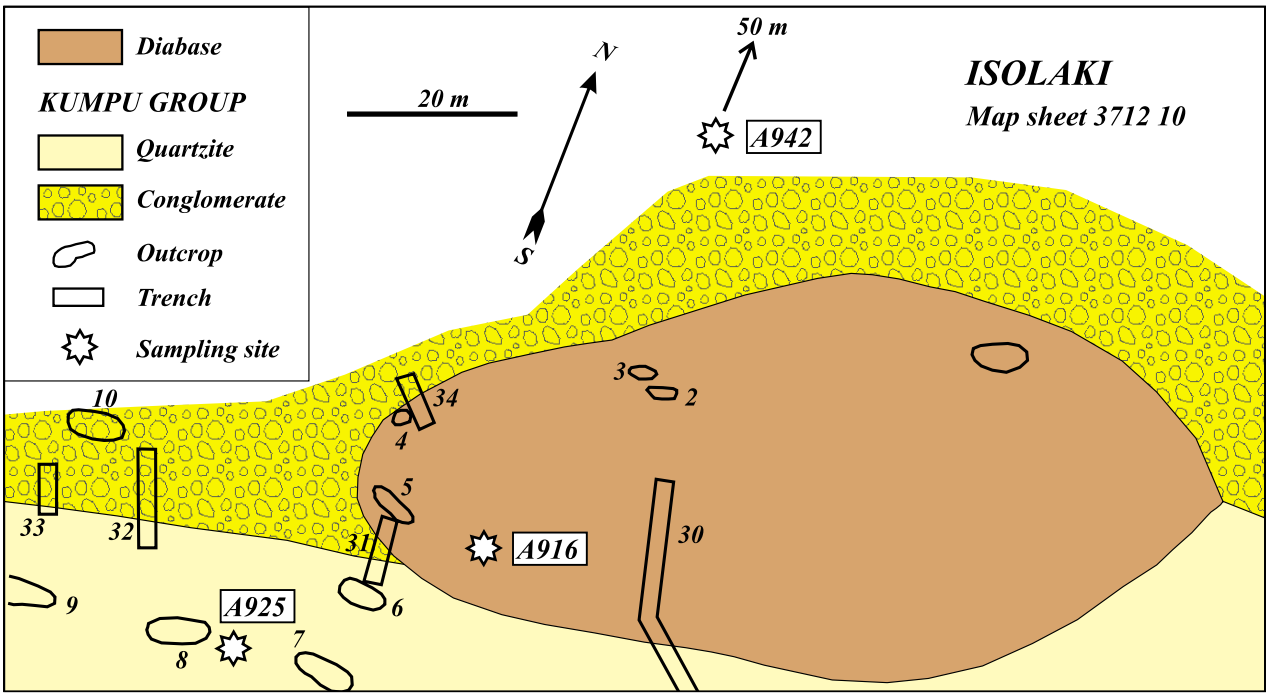

Fig. 4. Detailed geology of the Isolaki (Värttiövaara) area, Sodankylä.

Ma (MSWD 0.8). A discordant U-Pb analysis of titanite lies close to this line (Fig. 5). The result is interpreted as the emplacement age of this mafic rock and compares well with that of 2220±11 Ma (sample A892) reported previously for the Haaskalehto mafic intrusion by Tyrväinen (1983).

A470 Eksymäselkä. The sample (Fig. 1) was taken from the leucocratic part of a Haaskalehto-type intrusion cutting quartzites and metavolcanic rocks of the Honkavaara Formation at Eksymäselkä. The sampling site is close to the Kittilä-Sodankylä road c. 10 km west of the village of Tepsa. The upper contact of the intrusion with the quartzite is diffuse due to strong albitization affecting both the intrusion and its country rocks. The sample was taken from a pure, almost quartz-free albitite. The texture of the rock is fine-grained, allotriomorphic-granular and partly granoblastic. The main mineral, albite (c. 85% vol.), is accompanied by lesser amounts of carbonate, chlorite, sericite, epidote, quartz, titanite, zircon and opaque minerals. Zircon is relatively abundant and occurs as large, sharp-edged, clear, yellowish crystals. The zircon population seems to be homogeneous. No complete crystals are found among the separated fractions, but they contain broken grains with pyramid heads indicating that the crystal habit of zircon is rather simple. The uranium content varies between 400-1000 ppm and the common lead concentration is moderately high.

If the very discordant light fraction A is excluded from the calculation, the four other analyses of sample

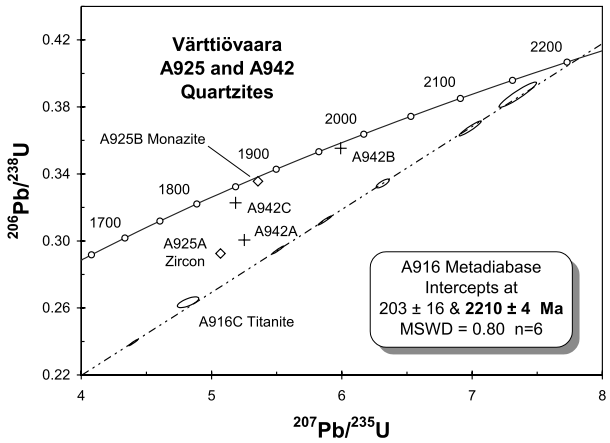

Fig. 5. Concordia diagram for zircon, titanite and monazite analyses from the Isolaki (Värttiövaara) area.

A470 define a linear trend with intercepts at 2214±4 Ma and 334±21 Ma (MSWD=0.2) (Fig. 6). If fraction A is included, the calculated upper intercept age would be 2209±7 Ma.

A472 Palttina. The sample for zircon dating was taken from a drill core No. 1 of Rautaruukki Co at Palttinavuoma, 1.5 km west of Lake Kuolajärvi and 3 km northwest of Tepsa (Fig. 1). The dated rock is an albite gabbro occurring at a depth interval of 45-88 m in the core. The gabbro intrudes albitized supracrustal rocks of the Sodankylä Group.

Separation yielded titanite and a small amount of zircon, which occurs as simple tetragonal prisms. The uranium content in zircon is high and the density

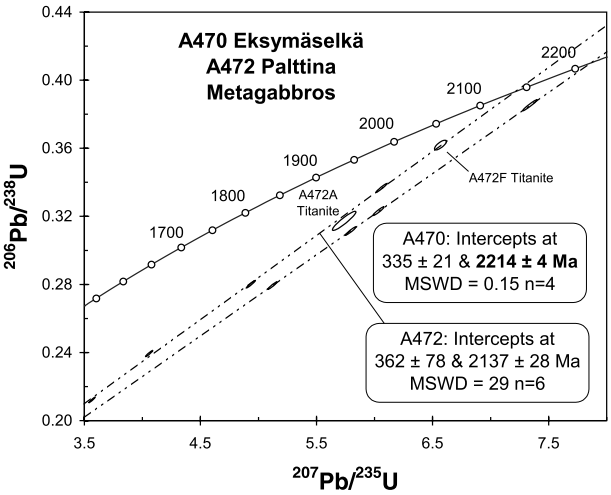

Fig. 6. Concordia diagram for zircon and titanite analyses from the Eksymäselkä and Palttina metagabbro samples.

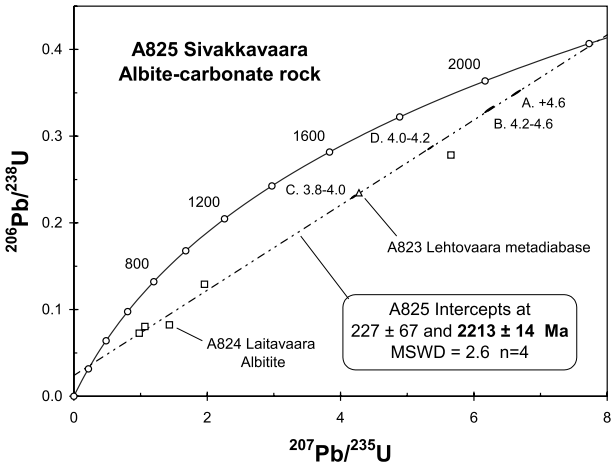

Fig. 7. Concordia diagram for zircon analyses from the Sivakkavaara albite-carbonate rock, the Laitavaara albitite and the Lehtovaara metadiabase.

accordingly low. Four discordant zircon and two discordant titanite fractions do not constrain a perfect chord, but nevertheless plot along a reference line with intercepts at 2137 and 362 Ma (Fig. 6).

A823 Lehtovaara. At Lehtovaara, 3 km south of Tepsa (Fig. 1) an arcuate, a few tens of meters thick metagabbro dyke penetrates quartzite belonging to the Virttiövaara Formation of the Sodankylä Group. Texturally the rock is hypidiomorphic and contains large (up to 1 cm), saussuritized albite grains enclosing ragged amphibole grains.

Zircon in sample A823 is small-grained and morphologically typical for zircons in igneous plutonic rocks; the length/breadth ratio (L/B) is around 3-4 and the prisms are simple. Only one fraction with a density of 3.8-4.2 g/cm^3 could be analyzed. The U concentration is relatively high (880 ppm), apparently having caused metamictization and loss of lead. One discordant analysis does not allow an age calculation, but the point plots close to the chord defined by sample A825 and is compatible with an age of c. 2.2 Ga (Fig. 7).

A824 Laitavaara. The albitite of Laitavaara (Fig. 1) is part of a differentiated Haaskalehto-type intrusion cutting metavolcanic rocks of the Honkavaara Formation. The sampling site lies close to the Sodankylä-Kittilä road, 5 km northwest of the village of Jeesiö.

The heaviest separated fraction (d>4.6) contains mainly barite, but also zircon with a normal, prismatic crystal habit and a brownish tint. In addition, some probably xenocrystic, clearly rounded zircon grains were discovered. As shown in Appendix I, the common lead contents are reasonably low. The uranium contents are very high (>2000 ppm) which has led to extensive metamictization and a high degree of discordance even in the case of the heaviest abraded fraction E representing probably xenocrystic zircon (Fig. 7). On the basis of the existing data, any age calculations are pointless.

A825 Sivakkavaara. Sample A825 comes from a little Haaskalehto-type intrusion (Fig. 1) occurring at Sivakkavaara, c. 7 km northwest of Jeesiö (Kallio 1980). The sample representing an albite-carbonate rock was taken near the contact with the greenish, fuchsite-bearing quartzite of the Virttiövaara Formation, and resembles the Eksymäselkä intrusion described above. The main minerals are albite, tremolite-actinolite, carbonate, quartz and chloritized biotite while accessory minerals commonly include epidote, magnetite and titanite.

Besides sulfides, the heaviest separated fraction (d>4.6 g/cm^3) contains barite and zircon. Zircon crystals are translucent and commonly reddish in color but some greyish ones were also found. L/B is usually about 2, in some cases 3. Morphologically the zircons are very similar to those occurring in plutonic rocks. The (100) faces are much larger than the (110) faces, and the (101) pyramid faces are larger than, for example, (211). In all fractions, the edges of the zircons are completely sharp. The freshest fractions contain a relatively small amount of uranium for an albitite, and the amount of common lead is low. The regression line on the concordia diagram defines an upper intercept age of 2213±14 Ma (Fig. 7) with an MSWD value of 2.6. The lower intercept falls into the usual range at 227±67 Ma.

2.0-2.06 Ga metadiabases and gabbros

In the southern part of the Kittilä greenstone area, mafic metadiabase and metagabbro dykes are known to occur as discordant and, more rarely, as concordant bodies in metavolcanic and metasedimentary rocks belonging mostly to the Savukoski Group. These intrusions are less differentiated than the older Haaskalehto-type intrusions dealt with in the previous section. The metadiabases belonging to this group are usually medium-grained and exhibit ophitic or subophitic textures. The main minerals are albitic plagioclase and various amounts of hornblende and biotite. Quartz, epidote, carbonate, scapolite, opaque minerals, chlorite, titanite, rutile and apatite are the most common accessory minerals. The gabbroic bodies form small intrusive lenses. They are generally more coarse-grained and contain more hornblende than the metadiabases and relicts of clinopyroxene may still be present in some samples.

Geochemically, the metadiabases are tholeiitic basalts, and they differ from the Haaskalehto-type intrusions, for example, in having flat chondrite-normalized REE patterns whereas those of the Haaskalehto-type intrusions are clearly LREE-enriched (Lehtonen et al. 1998). The metadiabases record ages around 2.05 Ga (Sätkänävaara, Riikonkoski and Heinälammintievat) while the gabbroic types are typically slightly younger with an age close to 2.0 Ga (Home-Jolhikko, Pittarova, Kulkujärvi, Latvajärvi). Similar age results have been reported for some metadiabases from Western Lapland (Väänänen & Lehtonen 2001, *this volume*). It is worth noting that ages falling in the range of 2.0-2.06 Ga have so far not been obtained for metadiabases in the Peräpohja schist belt (Perttunen & Vaasjoki 2001, *this volume*).

The age determinations of lamprophyre dykes at Palovaara are also treated in this connection.

A655 Sätkänävaara. At Sätkänävaara (Fig. 1), 10 km north of Kittilä, metadiabases occur together with mafic metalavas, metatuffs and metagreywackes of the Savukoski Group. This rock association is overlain by Kumpu Group metasediments. A 40 m thick metadiabase is in contact with quartzites and conglomerates (Fig. 8). Detailed mapping revealed no signs of thermal contact effects on the overlying metasediments and rounded metadiabase pebbles and cobbles ranging from 5 to 20 cm in size are encountered within the conglomerate. On the basis of these field observations, the metasediments of the Kumpu Group are considered younger than the Sätkänävaara metadiabase (Lehtonen et al. 1998).

The metadiabase was exposed with a trench across the contact zone against the overlying metasediments

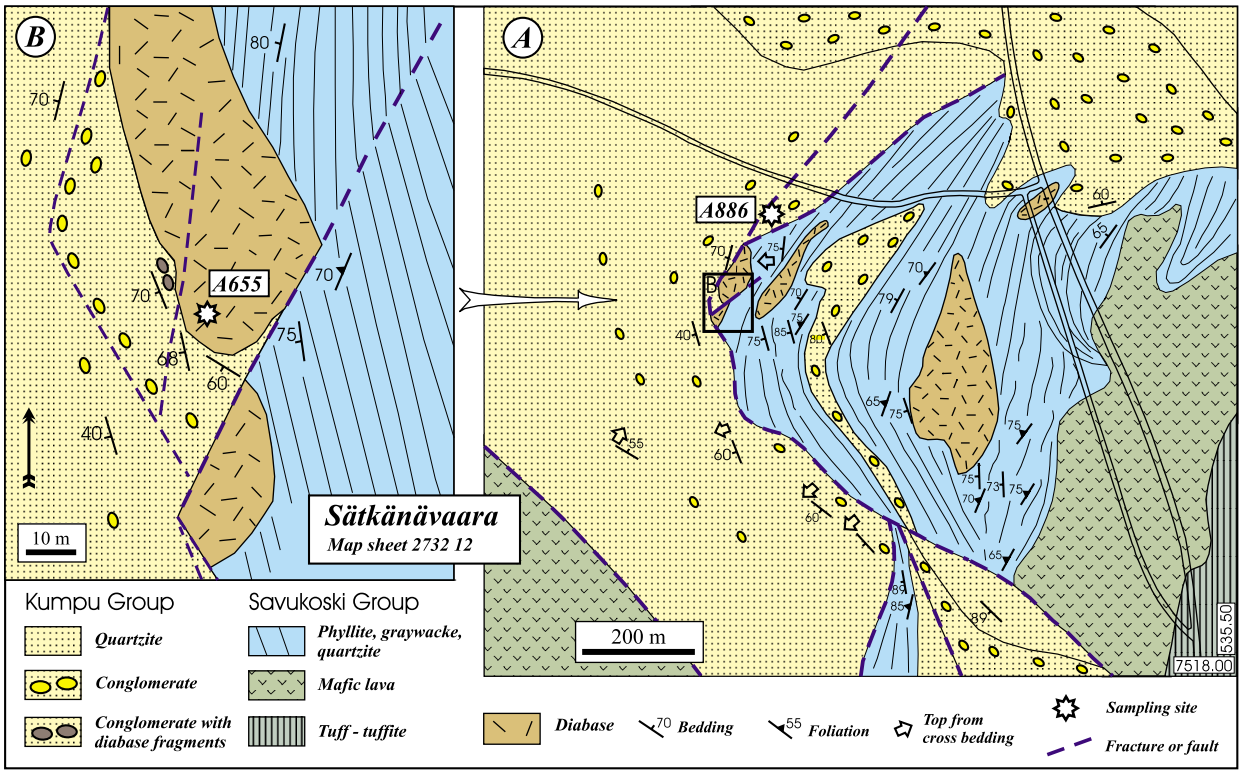

Fig. 8. Detailed geology of the Sätkänävaara area, Kittilä.

(Fig. 8B). The rock is a massive, coarse- to medium-grained, subophitic leucogabbro. It is composed of albite with lesser amounts of chlorite, carbonate, biotite, magnetite, titanite, quartz and sericite. The three metadiabase samples (A655a-c) picked at this locality differ slightly from each other in the abundance of the dark minerals, chlorite and biotite. A sample was also taken from the overlying Kumpu Group metasediments a few meters from the contact for the purpose of dating detrital zircons (sample A886, p. 119).

The separated zircon grains from the metadiabase samples are euhedral and transparent. In some grains, L/B is up to 5, and in sample A655c zircon occurs as grey or colorless, often tabular crystals. The zircons are thus morphologically different from those occurring in the typical Haaskalehto-type mafic intrusions. All eight U-Pb analyses from the three different samples define a discordia line with an upper intercept age of 2046±9 Ma, and a lower intercept at 312 Ma, but the MSWD is rather high at 7 (Fig. 9). Two of the analyses (E, F) plot slightly off this regression line and have larger analytical error than the other analyses. Omitting these analyses does not have a notable influence on the age (2048±8 Ma, MSWD 3). The obtained age is interpreted as representing the time of emplacement of the diabase and consequently demonstrates that the overlying Kumpu Group metasediments are younger than 2.05 Ga.

A450 and A452 Riikonkoski. In the Riikonkoski area west of Ounasjoki, 10 km north of Kittilä (Fig. 1), metadiabases occur as conformable or semi-conformable dyke-like intrusions and small lenses within tuffs, tuffites and phyllites of the Savukoski Group. Differentiation is observed in many places. Samples A450 and A452 come from two drill cores, R329 (depth 9.3-12.3 m) and R364 (200-203 m), respectively. The drill holes penetrate a metadiabase/gabbro which, according to geophysical data, reaches 1 km in length and 250 m in width (Yletyinen & Nenonen 1972, Nenonen 1975). The metadiabase is medium-grained and has an ophitic texture. Major minerals are albitic plagioclase and hornblende accompanied by lesser amounts of biotite and quartz. Accessories include titanomagnetite, sulfides, apatite and secondary epidote and chlorite.

Zircon grains in the metadiabase are mostly transparent and euhedral but occasionally also tabular crystals are encountered. L/B varies commonly between 2-3 but reaches 5 in some crystals. First U-Pb analyses were made in 1972 by using the borax fusion method (A450A, A452A). Later, four further analyses were done for A450, from which enough zircon was left over for fraction and air abrasion work. The five data points define intercepts at 2060±8 and 350±84 with an MSWD of 3 (Fig. 10). The fact that the borax fusion analyses for both A450 and A452, albeit discordant, plot on the discordia line defined by the four newer data points, is a fine tribute to the accuracy of the now outdated method.

A469 Heinälammintievat. Sample A469 represents an albitite in the middle of a 20 m thick metadiabase intruding metavolcanic rocks of the Savukoski Group at Heinälammintievat, 10 km southwest of the village of Rajala (Fig. 1). The sampled rock consists mainly of albite (c. 85%). Other minerals include carbonate, chlorite, sericite, epidote, titanite and opaque minerals. It is noteworthy that quartz is totally absent. The texture is hypidiomorphic-granular.

Zircon crystals are small-grained, tetragonal prisms. They are stained by a pigment but still transparent. Crystal edges are very sharp. L/B of the prismatic crystals is c. 5, particularly in the freshest fraction

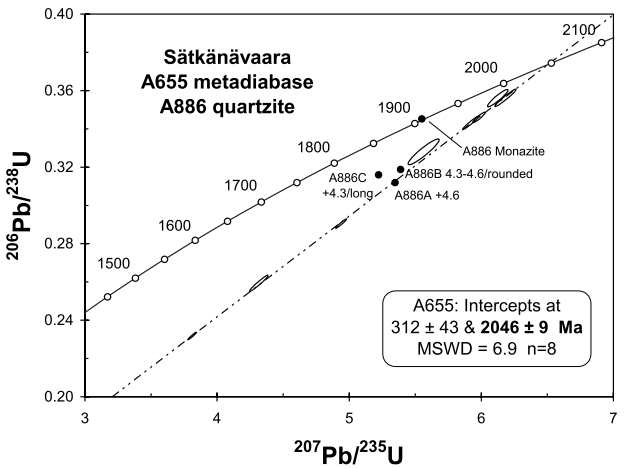

Fig. 9. Concordia diagram of zircon and monazite analyses from the Sätkänävaara metadiabase (error ellipses) and quartzite (dots).

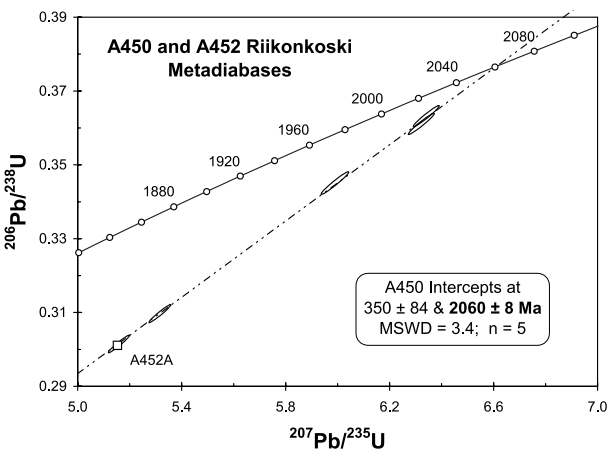

Fig. 10. Concordia diagram of zircon analyses from the Riikonkoski metadiabases.

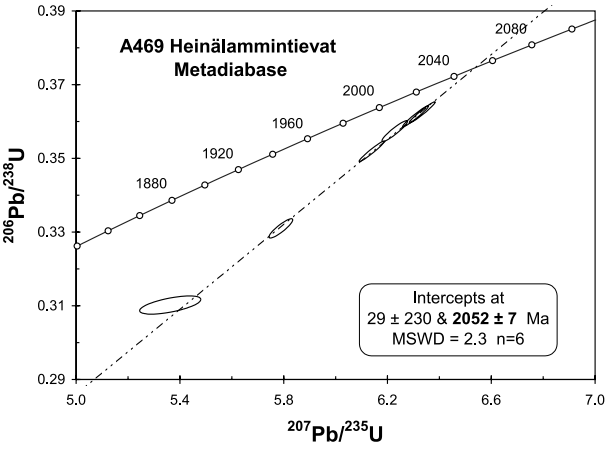

Fig. 11. Concordia diagram of zircon analyses from the Heinälammintievat metadiabase.

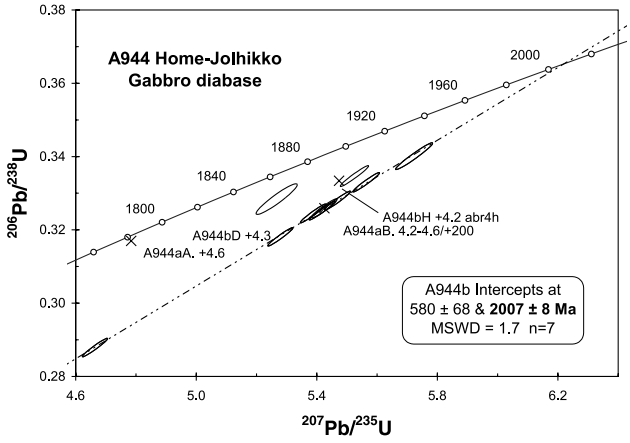

Fig. 12. Concordia diagram of zircon analyses from the Home-Jolhikko gabbro diabase. X – A944a, error ellipses – A944b.

(d>4.6 g/cm^3). The uranium content of zircon is relatively low. The analytical results of six zircon fractions are plotted on the concordia diagram of Figure 11, and define a chord with an upper intercept age of 2052±7 Ma and a lower intercept of 29±230 Ma (MSWD=2.3).

A944 and A946 Home-Jolhikko. The samples from Home-Jolhikko (Fig. 1) come from a metamorphosed gabbro diabase occurring as a semiconcordant body within tuffites of the Savukoski Group. Gabbro diabase occurs as a N-S trending, 20 m wide ridge of outcrops. On its western side, there exists another, parallel ridge of metadiabase outcrops. The outcrops in the area may belong to the same metadiabase dyke or form two parallel ones. Gabbro diabase is medium- to coarse-grained and consists mainly of euhedral plagioclase and light green hornblende with ragged margins. Plagioclase is partly altered to sericite and scapolite and contains numerous small amphibole inclusions. Opaque minerals, epidote, quartz, titanite and apatite are found as accessory phases.

Samples A944a and A944b represent the exposed gabbro diabase while the third sample, A946, was taken from a nearby gabbro pegmatoid boulder evidently of local provenance. The zircon crystals separated from the gabbro pegmatoid boulder are large, euhedral, glass-clear, simple tetragonal prisms with sharp edges. The crystals are long having an L/B of c. 5. In some grains, a brownish rim is seen. The zircons from the outcrop sample (A944) are morphologically similar but smaller in size and are not as beautifully clear as the zircons in sample A946.

Nine zircon fractions from the outcrop sample A944b form a scattered pattern on the concordia diagram (Fig. 12). Seven fractions of them plot roughly on a line, which indicates an age of 2007±8 Ma (lower intercept 580±68 Ma; MSWD 1.7). The two other fractions lie on its upper (younger) side, and represent heavy zircon. They also have lower Th/U (radiogenic ^{208}Pb/^{206}Pb in appendix), suggesting an origin different from the other zircon. This feature is more pronounced in the analyses of zircon from the second sample, A944a. The heavy zircon (A944aA) is close to concordance at 1790 Ga.

The 6 fractions separated from the boulder sample A946 seem to define two groups: both analyses from the +4.6 fractions A and C seem to be slightly older than the lighter fractions. The reference line for the heavy fractions has intercepts at 2007 Ma and 35 Ma, while the lighter fractions define a linear trend with the intercepts at 2003±4 Ma and 255±110 Ma (Fig. 13). Considering the evidence from the two samples, it seems likely that the original emplacement of the Home-Jolhikko dyke occurred 2007±8 Ma ago and that some zircon formed during a late hydrothermal stage c. at 1.8 Ga. Two fractions obviously contain composite grains including late overgrowth.

A1272 Pittarova. The Pittarova gabbro diabase occurs by the Saattopora-Pahtavuoma road, 3 km west of the closed Saattopora gold mine (Fig. 1). It slightly exceeds 100 m in thickness and cuts obliquely mafic tuffs and tuffites of the Savukoski Group. The sampled rock is coarse-grained and has a subophitic texture with euhedral albitic plagioclase and light green hornblende as the main minerals. In places, hornblende forms glomerophyric aggregates up to a couple of cm in size. Minor constituents include magnetite, quartz, apatite, epidote, carbonate and biotite.

The separated zircons are generally turbid and poorly defined morphologically, which may account for the fact that the three analyzed fractions are rather discordant. Nevertheless, as shown in Figure 14, they plot accurately on the same line indicating an age of

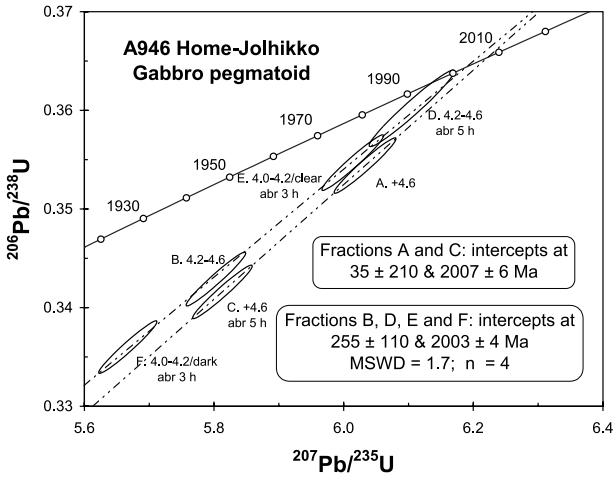

Fig. 13. Concordia diagram of zircon analyses from the Home-Jolhikko gabbro pegmatoid.

1995±8 Ma (lower intercept at 249±57 Ma, MSWD 0.06).

A1273 Kulkujärvi. The Kulkujärvi gabbro lies on the eastern side of Lake Jerisjärvi, by the Kittilä-Muonio road. The sample was taken from a 60 m long road cut. The rock is massive and medium-grained with some variation in the grain size. The major minerals are euhedral andesine, light green hornblende and augite. Accessories comprise opaque minerals, quartz, apatite, epidote and titanite. The texture of the rock is hypidiomorphic-granular.

Previously, a poorly exposed, 6 km long magnetic anomaly at Kulkujärvi was regarded as generated by a gabbro similar to that exposed at the road cut (Lehtonen et al. 1984, 1985). However, a drill hole profile made across the anomaly revealed that it is mainly caused by mafic volcanic rocks, now assigned to the Kittilä Group. Hence, the dated rock, situated at the edge of the magnetic anomaly, is a relatively small gabbroic body within the metavolcanic rocks of the Kittilä Group.

The separated zircons seem to form two groups in terms of their grain size. Most of the grains (95%) are larger than 70µm. This main population is represented by light-colored, bright and clearly translucent, mostly anhedral grains, probably being fragments of even larger crystals. The other population (5%) of smaller crystals (<70 µm) also consists of light-colored, bright and subhedral grains. Their small size appears to be an original feature and not the result of crushing during separation.

The nine analyzed zircon fractions do not form a linear array, making a straightforward interpretation of the analytical results impossible (see Fig. 14). It is noteworthy that the uranium concentration of all fractions is relatively low (52-257 ppm) which accounts for the low degree of metamictization and discordance. The heaviest zircons of the coarser main population are nearly concordant with $^{207}Pb/^{206}Pb$ ages from 1961 to 1965 Ma. Fraction H with the lowest U content (52 ppm) was analyzed after a 10 h abrasion and yielded a concordant age of 1986±2 Ma. A regression

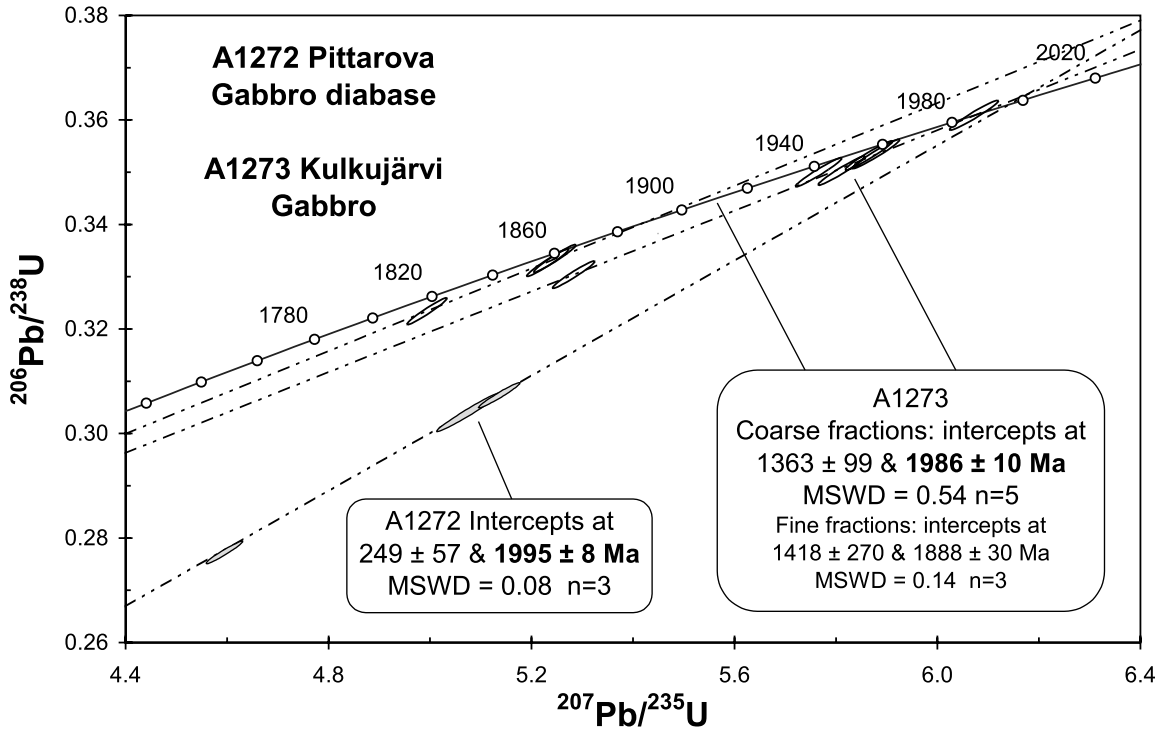

Fig. 14. Concordia diagram of zircon analyses from the Pittarova gabbro diabase and the Kulkujärvi gabbro.

line including 5 zircon fractions (A, B, C, G and H) of the main population produce an upper intercept age of 1986±10 Ma and a relatively high lower intercept age of 1363±99 Ma, the MSWD being 0.5. The minor fine-grained zircon population represented by fractions E, F and I seems to be younger than the main population. A three-point regression line yields intercepts at 1888±30 and 1418±270 Ma (MSWD 0.14), but it remains obscure whether these zircons represent a single population or composite grains between 2.0 Ga and e.g. 1.8 Ga, the age obtained for younger zircon population in the Home-Jolhikko sample A944 discussed above.

The age of 1986±2 Ma obtained for the concordant zircon fraction of the main population is thought to indicate accurately the time of emplacement of the Kulkujärvi gabbro. The apparent upper intercept age of 1888 Ma for the minor zircon population corresponds to that of the Haaparanta Suite plutons. This relation provides us with two interpretations. Firstly, the sampling site is not far from large plutons of the Haaparanta Suite in the Jerisjärvi and Äkäsjärvi areas and hence, the small younger zircons could represent metamorphic/hydrothermal crystals formed by a metamorphic effect coeval with the emplacement of these intrusions (Lehtonen 1984, Lehtonen et al. 1984). Alternatively, the younger zircons could have been derived from some dykes related to the Haaparanta suite magmatism. In fact, the road cut which was sampled, contains some a few cm wide felsic dykes.

A1395 Latvajärvi. A metadiabase has been penetrated by a drill hole (M52/2732/75/R307/166.30-172.45, x=7503.12, y=2524.187) at Latvajärvi c. 15 km from Kittilä (Fig. 1). It occurs in a tectonic contact zone between the c. 1.88 Ga Latvajärvi felsic porphyries and conglomerates and the 1.8 Ga Kittilä granite. The metadiabase is medium-grained (0.5-5 mm) and subophitic. The main minerals are euhedral albitic plagioclase and light green hornblende. Accessory phases are biotite, carbonate, titanite and opaque minerals.

Mineral separation produced a lot of titanite, but only a small amount of zircon. Some zircons are colorless, rather long-prismatic, others ill-defined chalky turbid. Four U-Pb microanalyses were made on zircon separates, three of which (A-C) are analytically normal, but the result of the fourth analysis (D) is uncertain and very discordant. As is seen in Figure 15, the fractions do not form a linear trend on a concordia diagram, which implies a heterogeneous population. In principle, the reason could be either contamination with an extraneous zircon or a metamorphic effect. On the basis of the zircon morphology, the turbid zircons (fraction C) seem younger and more discordant, which suggests a metamorphic effect. The apparently low Th/U in this fraction (as reflected by $^{208}Pb/^{206}Pb$, see Appendix I) is another difference when compared with the euhedral fractions.

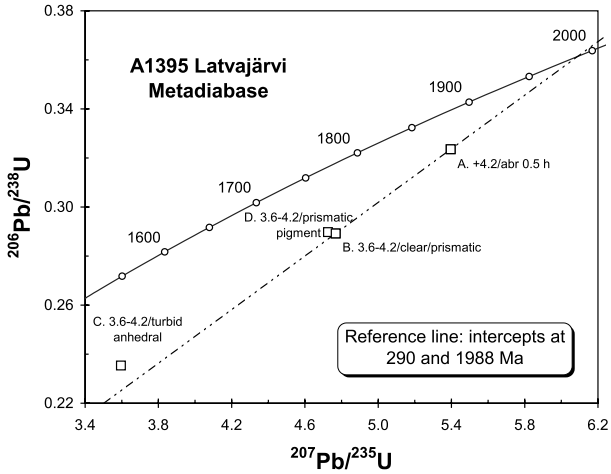

Fig. 15. Concordia diagram of zircon analyses from the Latvajärvi metadiabase.

The Pb-Pb age of 1970 Ma obtained for fraction A can be considered a minimum age of the zircon. A reference line drawn through fractions A and B cuts the concordia curve at c. 1990 Ma (Fig. 15), but this can be regarded only as a first order approximation. Nevertheless, as these fractions represent the same morphological type, the age estimate is hardly a consequence of mixing of Archean and 1.9 Ga zircons. Unfortunately, all zircon material was used up during the analytical work.

The titanite has a $^{207}Pb/^{206}Pb$ age of 1791±5 Ma, which is quite normal for titanites and monazites from this part of Lapland (e.g., Mänttäri 1995, Manninen et al. 2001, *this volume*).

It seems likely that the euhedral zircon (fractions A, B and D) belong to the magmatic phases of the gabbro diabase. In that case, the diabase can neither be assigned to the 1.88 Ga Latvajärvi Formation and nor can it cut this formation. Instead, the rock more probably belongs to the c. 2 Ga mafic magmatism, products of which occur widely in the Kittilä area. This interpretation is supported by a shear zone at the contact between the Latvajärvi Formation and the Kittilä granite.

A1441b Palovaara. Palovaara lies c. 15 km east of Kittilä (Fig. 1). Here a drilling profile was made during the studies of the Lapland Volcanite Project. The profile penetrated a supracrustal sequence containing mafic and ultramafic metavolcanic rocks of the Savukoski Group and phyllites, siltstones, quartzites,

tuff beds of the Sodankylä Group. In the drill cores, lamprophyre dykes 2-10 m in width cut obliquely the metasediment layers. According to Eilu (1994), the dykes are chemically calc-alkaline and the least altered types possess biotite-phyric, apatite-phyric and intergranular textures. The biotite phenocrysts 1-10 mm in size are bent and broken. The groundmass consists of albite, biotite, chlorite, quartz and magnetite. In the most mafic types, the groundmass consists mainly of talc. The dykes have undergone strong alteration. The aim of the isotopic study was to find out both the time of emplacement and alteration of the lamprophyres. For this purpose, three samples were picked from a drill core (R1-LVP-86, x=7511.48, y=2553.12). The samples and their depth intervals are as follows: A1439, 180.0-182.0 m, a c. 7 m thick dyke; A1441a, 189.60-192.50 m, a c. 8 m thick dyke, a sheared type; A1441b, 193.60-195.40 m, the same dyke as above, a non-sheared type.

A small amount of monazite was gained from sample A1441b, but otherwise the separation yield was unsuccessful for these samples. The monazite analysis is slightly discordant with a $^{207}Pb/^{206}Pb$ age of 1771±4 Ma, which is considered to be close to the age of monazite.

Another question is whether this date represents the time of crystallization or alteration of the lamprophyric rock. The age result is comparable with that obtained by Mänttäri (1995) for monazite(1781±18 Ma) from a carbonate-sulfide vein cutting a tuff and tuffite of the Savukoski Group at Saattopora. We regard the monazite age as recording the time of the latest major stage of alteration in Central Lapland.

U-Pb STUDIES ON FELSIC VOLCANIC ROCKS

Felsic volcanic rocks in the Kittilä area exhibit both extrusive and subvolcanic lithologies whose zircon ages range from Archean to early Proterozoic. In the following, they are discussed as two groups: rocks with Archean zircon and those with Proterozoic zircon. The second group includes both volcanic and intrusive rocks. It is important to emphasize here that the first group comprises altered felsic supracrustal rocks with a somewhat ambiguous origin. To some extent, the Archean ages obtained for some of these felsic supracrustal rocks are problematic as the age is inconsistent with the geological interpretation (see below).

Rocks with Archean zircon

A660, A914 and A968 Honkavaara. At some localities within the Central Lapland Greenstone Belt, felsic Na-rich rocks, interpreted as volcanic in origin, yield Archean U-Pb zircon ages. One of these is Honkavaara, where a crosscutting Haaskalehto-type metadiabase (A915) was dated at 2.2 Ga.

The geology of the supracrustal succession in the Honkavaara area has been described by Lehtonen et al. (1989) and Eilu (1994). The area forms the type locality for the Honkavaara Formation of the Sodankylä Group in the stratigraphical classification by Lehtonen et al. (1998) and consists of interlayered mafic to felsic supracrustal rocks, intruded by Haaskalehto-type mafic intrusions. The mafic rocks are interpreted as lava flows while the felsic varieties are considered tuffs and tuffites (Lehtonen et al. 1989). A chaotic breccia covers the central part of the study area (op. cit. Fig. 22) and is bordered by diabases in the west, mafic lavas in the south and tuffites in the north. The mostly angular fragments measure from a few millimeters to about half a meter in diameter. Sometimes the breccia is layered and graded, but usually its structure is chaotic. Lehtonen et al. (1989) regarded the breccia as a lahar deposit.

The petrography of the rocks at Honkavaara has been described by Eilu (1994). The rocks are highly altered; albitization, chloritization, carbonitization and sericitization have been the principal alteration processes. The porphyritic mafic lavas contain albite and actinolite phenocrysts in a groundmass of albite, actinolite and chlorite. The felsic tuffs/tuffites are banded or massive rocks with albite, quartz, tremolite-actinolite, magnetite and titanite as the main constituents. The tuffites are banded and layered with sedimentary interbeds, 1 to 2 m in thickness, with blastoclastic quartzite as the main component. Quartz, albite, hematite and rutile are the principal minerals of the tuffites, while quartz and albite dominate the mineralogy of the quartzites. The breccia fragments are highly altered and resemble lithologically the country rocks in the area. The main mineralogy of the breccia fragments is albite, quartz and carbonate and the composition of the matrix is albite, quartz and carbonate with occasional sericite and chlorite. The mafic

lavas in the Honkavaara area are subalkaline basalts and basaltic andesites with tholeiitic affinity while the felsic rocks are geochemically dacites and rhyolites, with most of them having a calc-alkaline character (Lehtonen et al. 1998).

Samples from three felsic rocks (A660, A914 and A968) were used for U-Pb studies. These rocks are fine-grained, granoblastic tuffs/tuffites with albite and tremolite-actinolite as the main minerals. Zircon grains in all samples are small and often full of pigment. Many grains are slightly rounded, but some have still sharp edges with L/B mostly at 1.5-2. Some variation can be observed in color. A HF-etched section of zircons from sample A660 shows that many grains have a well-developed euhedral zoning. Some grains have a distinct translucent overgrowth.

Five density fractions from sample A914 are very discordant, whereas two air-abraded fractions lie much closer to the concordia curve (Fig. 3). A seven point discordia yields an upper intercept age of 2746±14 Ma, the lower intercept being 334 Ma. However, the high MSWD (23) suggest scatter in excess of analytical error. Five analyses from sample A660 provide an age of 2721±9 Ma (lower intercept at 483 Ma, MSWD 2.4), and the three analyses from A968 lie close to the same chord. Two titanite analyses from sample A660 are slightly reversely discordant with $^{207}Pb/^{206}Pb$ ages at slightly over 2200 Ma, which corresponds to the age of the metadiabase (A915). Presumably the reverse discordance is due to slight inequilibrium between spike and sample solutions, but this does not affect the $^{207}Pb/^{206}Pb$ ages.

The calculated upper intercept ages above are marginally different. As the analyzed rocks should represent the same volcanic formation, one would expect the zircon ages to be the same. Evaluation of the ages is, however, complicated by the discordance and heterogeneity of the results. This difference becomes more pronounced when abraded zircon fractions are compared: analyses A660E, A914F and G are clearly different. If they ultimately represent coeval crystallization, either metamorphic effects or slight contamination would be required to explain the results.

A973 Venejoki. In the Venejoki area, c. 15 km SW from Kittilä, zircons have been extracted from a fine-grained albite-rich rock, which might originally represent volcanic or sedimentary material. The areal extent of this rock type is quite limited. The geology of the area is complex and the rock types are quartzites, mica schists, tuffs/tuffites, metadiabases, ultramafic rocks and granites (Rastas 1984).

The albitite rock at Venejoki belongs stratigraphically to the upper part of the Sodankylä Group and hence,

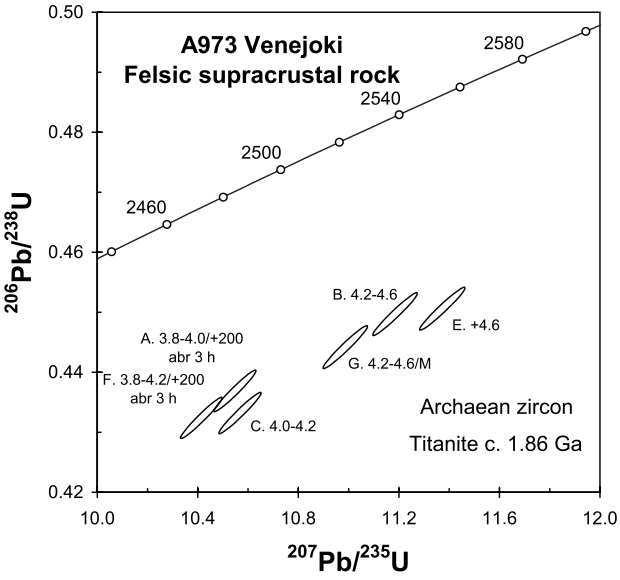

Fig. 16. Concordia diagram of zircon analyses from the Venejoki felsic supracrustal rock. Note that the analysis for the Paleoproterozoic titanite would plot outside of the diagram.

its position is comparable with that of the Honkavaara Formation described above. However, as the albite rock occurs close to the E-W-trending Venejoki fracture zone, its position could be tectonically disturbed, particularly as breccia structures are a striking feature of the albite rock. Locally pink, albite-rich fragments of varying size are abundant in a calc-silicate-bearing quartz-albite matrix. The form of the fragments varies from angular to rounded and more light-colored rims are often observed around the rounded fragments.

The sample for isotopic studies was blasted from a fine-grained, pink albite rock containing thin (1-2 mm) amphibole bands (tremolite/actinolite). The rock consists mainly of fine-grained albite with minor amounts of quartz, light-colored amphibole, titanite, biotite, opaque minerals and apatite. Geochemically it corresponds to rhyodacite.

Most zircons in the sample are rather fine-grained (<70 μm) and stubby (L/B<2). Some crystals are longer (L/B>3) and euhedral crystals occur occasionally. The short, fine-grained crystals vary from colorless to violet, mixed crystals of the two varieties being fairly common. Many of these zircons appear to be rounded, as only a few definite crystal faces have been observed.

Six zircon analyses have all Archean $^{207}Pb/^{206}Pb$ ages, but scatter rather randomly (Fig. 16), which is suggestive of an (possibly multiple) Archean provenance and/or complicated alteration history. The titanite from the sample is nearly concordant with a $^{207}Pb/^{206}Pb$ age of 1858±9 Ma, which may characterize the last time of cooling below c. 500°C.

Rocks with Proterozoic zircon

Proterozoic felsic porphyries have turned out to be good targets for zircon dating in the Kittilä area. They are volumetrically minor but areally wide-spread among the mafic rocks of the Kittilä Group and can be divided into two groups which differ from each other in terms of their age (2.01-2.02 and 1.92 Ga), petrographical features and chemical composition. The field characteristics of the *older group*, represented by the Kapsajoki, Veikasenmaa, Yräjärvi and Kiimarova porphyries, suggest their coeval nature with associated Kittilä Group mafic volcanic and subvolcanic rocks. These porphyries are normally quartz- and plagioclase-phyric and range geochemically from dacites to rhyolites, although some samples may plot in the andesite field due to mixing of felsic and mafic material (Lehtonen et al. 1998).

The *younger group,* represented by the Nyssäkoski porphyry dyke, contains quartz, plagioclase, amphibole and biotite as phenocrysts in a fine-grained felsic groundmass and has a more restricted chemical composition when compared to the older group, plotting in the dacitic field. The distinction between the two porphyry groups is obvious in their Nd isotopic and trace element compositions, the older group having higher ε_{Nd} values and higher concentrations of incompatible elements such as REE and HFSE (Hanski et al. 1997, Lehtonen et al. 1998).

In addition to the above mentioned felsic rocks within the Kittilä Group, a third type occurs in the Latvajärvi area, representing c. 1.88 Ga calc-alkaline metavolcanic rocks of the Lainio Group. In terms of their age and geochemical characteristics, they resemble the Kiruna porphyries in northern Sweden (Skiöld 1986, 1987). In the following, age determinations for the above described three porphyry generations are presented.

2.01-2.02 Ga porphyries

A280 Kapsajoki. At Kapsajoki, c. 40 km NNE of Kittilä, some felsic volcanic rocks are associated with voluminous mafic metavolcanics of the Kittilä Group (Fig. 1). Also chert-jaspilite units are abundant in this zone. The sampling site lies on the eastern side of River Kapsajoki, at Suasmuotkamaa, c. 200 m from the river. Sample A280 was taken from a lava-like quartz-feldspar porphyry. The mafic metavolcanic rocks in the vicinity are tuffs with tuffitic and black schist intercalations. Diamond drilling in the area has revealed that felsic porphyries occur both as extrusions and subvolcanic dykes coeval with the mafic volcanism (Kuronen 1983).

The felsic porphyry is exposed as a 2 m wide bed on the eastern margin of the outcrop. The contact with a mafic volcanic rock is almost vertical and seems somewhat sheared. Euhedral phenocrysts (0.5-4 mm) can be distinguished on the grey weathering surface of the porphyry. They comprise bluish quartz and whitish plagioclase crystals. In addition, 0.1-3 cm long amygdules, composed mostly of quartz and carbonate, are observed. Quartz phenocrysts contain abundant melt inclusions and deep embayments. The groundmass is very fine-grained and consists of quartz, albite and a small amount of carbonate, chlorite, opaque minerals and titanite. Chemically, the porphyry is rhyolitic with SiO_2 falling between 69-81% wt.

The zircon grains are generally pale-brown and translucent, although turbid yellowish crystals also occur. The grain size is rather small, 90% of the crystals being <70 µm. L/B is generally <3, but many longer crystals have apparently been broken during milling. Under oil immersion, most crystals exhibit sharp-edged, simple prismatic and pyramidal surfaces. Oscillatory zoning is common, but about 20% of the crystals are devoid of any internal texture. In some rare cases, unidentified transparent, rod-like inclusions were observed. The three analyzed zircon fractions are only slightly discordant and provide an upper intercept age of 2017±8 Ma (Fig. 17).

A581 Veikasenmaa. On the NW side of Sirkka village, the Veikasenmaa Formation belonging to the Kittilä Group mainly comprises mafic metavolcanic rocks including lavas, pyroclastic deposits, volcanic breccias and tuffites. The formation also contains

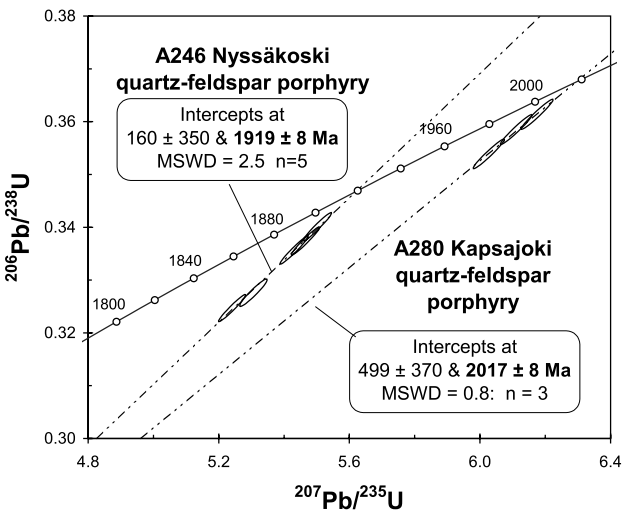

Fig. 17. Concordia diagram of zircon analyses from the Kapsajoki and Nyssäkoski quartz-feldspar porphyries.

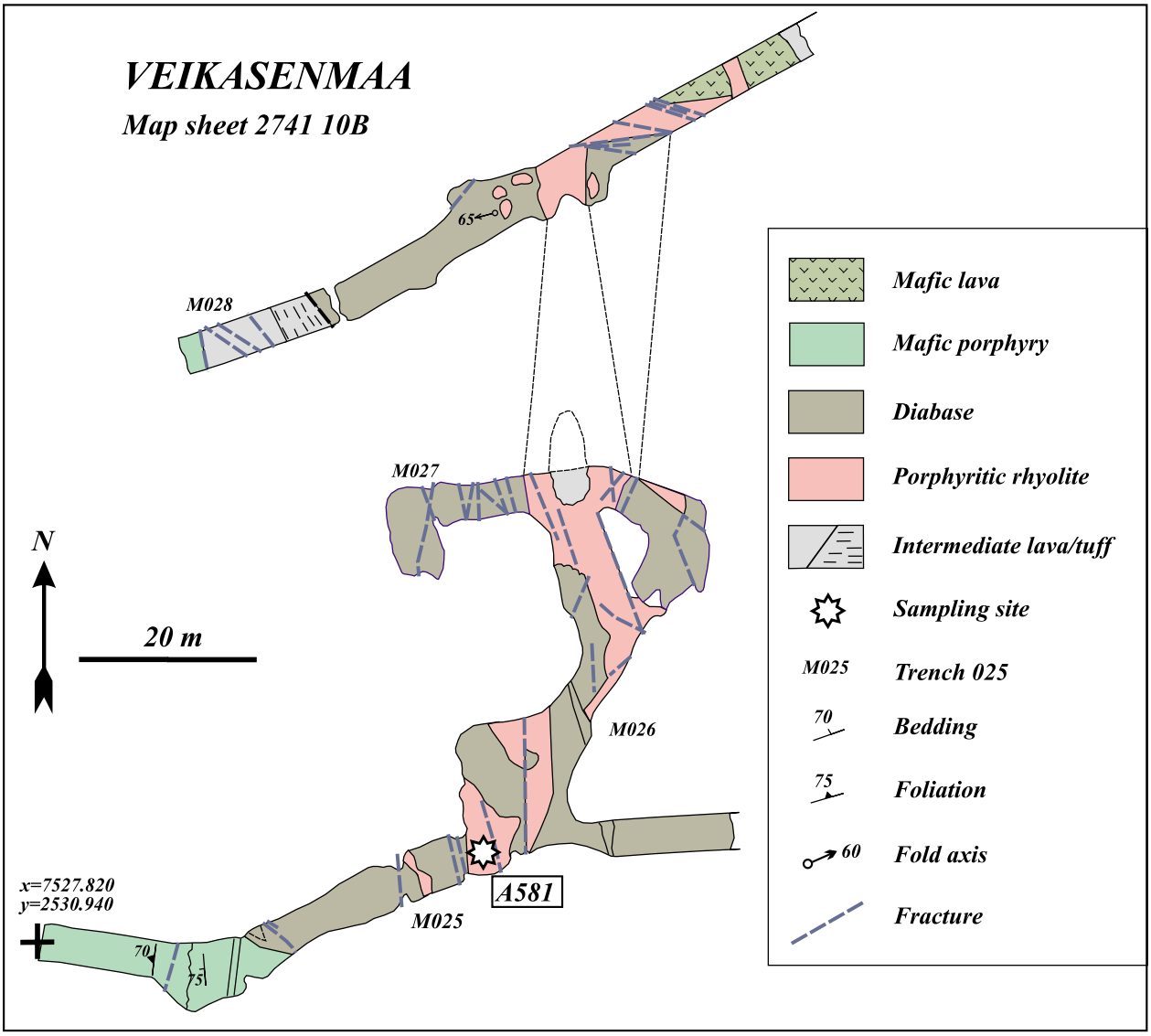

Fig. 18. Detailed geology from the trenches at Veikasenmaa, Kittilä.

conformable or crosscutting diabases, genetically related to the volcanic rocks, as well as felsic porphyries. The rock types are best exhibited at Veikasenmaa (Fig 1), about 25 km NNW of Kittilä. Here a porphyritic rhyolite occurs together with diabases and mafic volcanic rocks with massive or pillowed structures. Detailed field observations made in a trench suggest that the mafic volcanic rocks are cut by rhyolitic dykes, which in turn are intruded by a diabase dyke (Fig. 18). The felsic porphyry occurs commonly as large inclusions in diabase or smaller fragments in mafic lavas. It is noteworthy that mafic metavolcanite inclusions are observed in felsic porphyries, too, and thin apophyses from mafic rocks have injected into porphyries.

With regard to their major and trace element composition, the mafic metavolcanic and dyke rocks of the Veikasenmaa Formation are tholeiitic basalts very similar to each other. Accordingly, they are regarded as comagmatic and can be geochemically correlated with the metavolcanic rocks of the Vesmajärvi Formation, the type formation of the upper part of the Kittilä Group (Lehtonen et al. 1998). The obtained geochemical data coupled with the field observations made at Veikasenmaa suggest that the felsic porphyries are coeval with the mafic supracrustal and subvolcanic rocks and hence, the U-Pb zircon data on felsic rocks have important bearings on the age of the whole Kittilä Group.

On the light-colored or locally even white weathering surface of the felsic porphyries, dark quartz phenocrysts (1-2 mm) are distinguished together with somewhat larger plagioclase phenocrysts which locally form glomerophyric aggregates. The groundmass is composed of aphyric quartz, albite and chlorite mass.

The felsic rocks are calc-alkaline rhyolites and dacites with SiO_2 varying between 65 and 82% wt. A

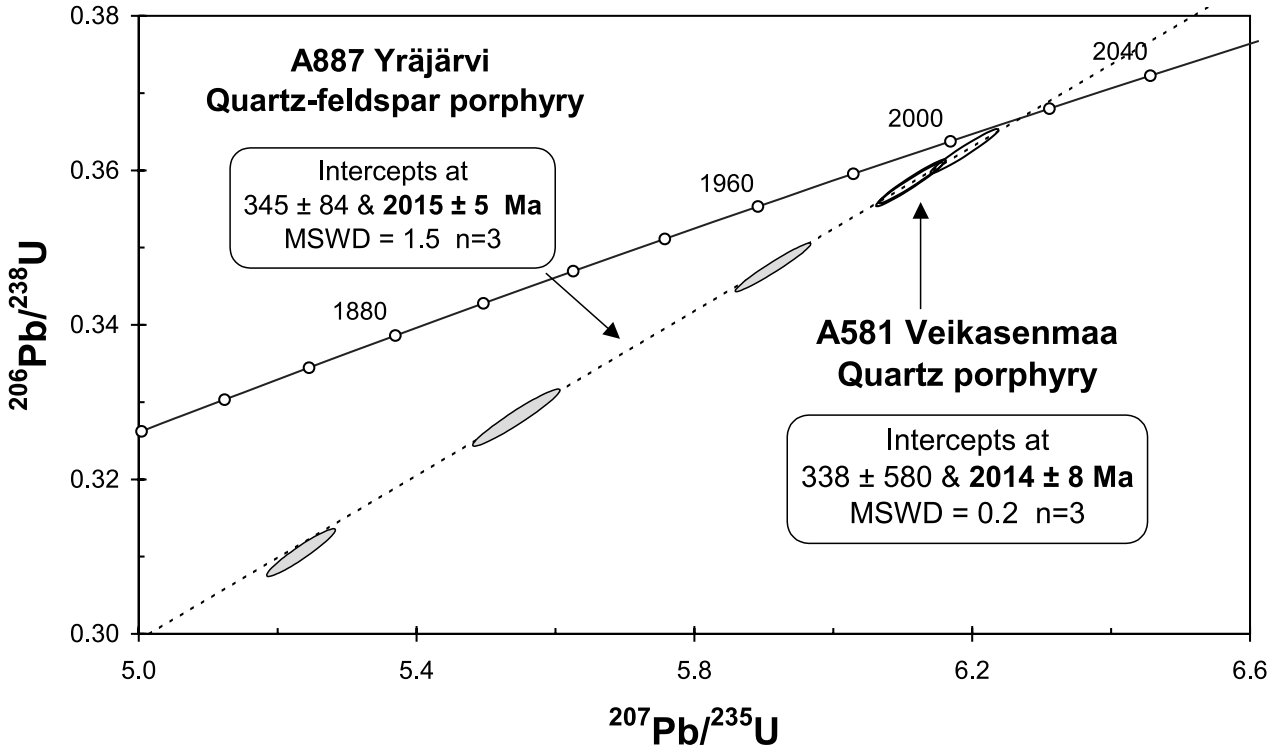

Fig. 19. Concordia diagram of zircon analyses from the Veikasenmaa quartz porphyry and the Yräjärvi quartz-feldspar porphyry.

typical feature of the Veikasenmaa porphyry as well as the other porphyries of this age group is a low K_2O/Na_2O value, mostly below 0.5 and only rarely exceeding 1.0. They are LREE-enriched ($La_N/Sm_N = 2.9$) but have non-fractionated HREE distributions and show well-defined negative Eu anomalies in chondrite-normalized REE diagrams. It is significant that Ta or Nb are not notably lower than La or Th in the MORB-normalized spidergrams which is contrary to what is commonly observed for felsic rocks derived from continental crustal material. A non-continental origin is also consistent with the Nd isotopic data obtained for the porphyries (initial-$\varepsilon = +3.8$, Hanski et al. 1997).

Zircon grains separated from a fine-grained quartz-feldspar porphyry sample (A581) occur generally as short prisms, though tabular morphologies are present as well. The three analyses are slightly discordant, very low in common Pb and define an upper intercept at 2014±8 Ma. Due to the close bunching of the analyses, the error limit is relatively high (Fig. 19).

A887 Yräjärvi. The rock types in the study area at Yräjärvi (Fig. 1), 2 km east of Veikasenmaa, are similar to those of the latter area. Fine-grained, felsic porphyritic rocks occur as interlayers and fragments in mafic volcanics. The contacts with mafic rocks can be gradational or sharp. Felsic porphyries are locally associated with intermediate crystal tuff layers. Part of the felsic rocks occur as narrow (30 cm) porphyry dykes and are younger than the mafic metavolcanic rocks. In addition, the exposed area contains fine-grained, narrow mafic dykes geochemically analogous to the associated mafic metavolcanic rocks. These observations suggest that the different types of mafic rocks are comagmatic and coeval with the felsic rocks.

The sample for isotope analysis (A887) is a very fine-grained quartz-feldspar porphyry. Albitic plagioclase is the major mineral in the groundmass, which contains also quartz and some chlorite, titanite and opaque minerals. A primary trachytoid texture as well as the presence of vesicular glass shards have been observed.

Zircons (fraction +4.3 g/cm^3) are small, euhedral, transparent and relatively short (L/B 2.5-3). Three fractions give an upper intercept age of 2015±5 Ma (lower intercept at 345 Ma, MSWD 1.5, Fig. 19).

A893 Kiimarova. At Kiimarova (Fig. 1), about 11 km SW of Veikasenmaa and in the vicinity of the now abandoned Saattopora gold mine, a felsic tuff unit occurs among mafic pillow lavas, which form a c. 200 m thick, SE-NW-trending unit. The primary structures of the pillows are well-preserved with their size varying between 10-120 cm. The rocks are fine-grained (0.1-0.8 mm) and consist of albite, tremolite-actinolite and chlorite with accessory titanite, epidote, quartz, opaque minerals, carbonate and biotite. Geochemically, the Kiimarova mafic metavolcanic rocks can be correlated with those occurring at

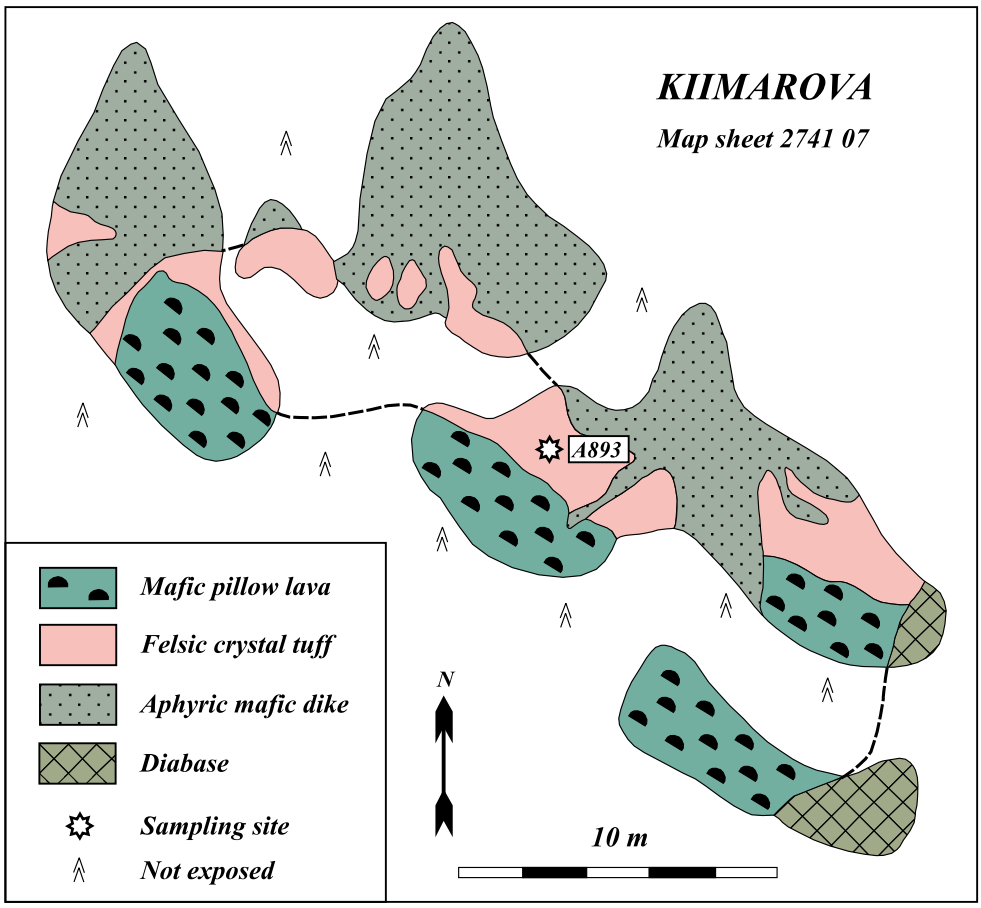

Fig. 20. Local geology from the Kiimarova area, Kittilä.

Veikasenmaa.

Mafic diabase dykes (0.05-8 m thick) have intruded the pillow lavas and some of them are concordant while others trend almost perpendicularly to the strike of the lavas. The latter ones are often fine-grained and can occur as a set of parallel dykes a couple of meters thick. In some places, fine-grained diabases are seen penetrating coarse-grained ones. The parallel dykes possibly represent incipient formation of a sheeted dyke complex. A more well-developed dyke complex was recently found at Selkäsenvuoma, 7 km west of Kiimarova (Lehtonen et al. 1998). On the basis of their geochemical composition, the mafic dykes are comagmatic with the associated mafic metavolcanic rocks.

In the eastern part of the outcrops at Kiimarova, a felsic porphyry is found as a 2-3 m thick layer on top of mafic pillow lava (Fig. 20). On its northern side, the porphyry is bounded by a massive, medium-grained mafic diabase-like rock, from which apophyses are seen to intrude through the felsic porphyry to the pillow lava. The porphyry also occurs in the diabase as inclusions of varying size . On the weathering surface, the felsic rock is usually massive and light-grey in color, but occasionally has a fragmental structure. The rock is interpreted as a crystal tuff layer.

Under the microscope, the felsic rock appears as porphyritic with albitic plagioclase (0.2-2.2 mm) and partly resorbed quartz (0.1-1.2 mm) as phenocrysts. The fine-grained groundmass is also composed mainly

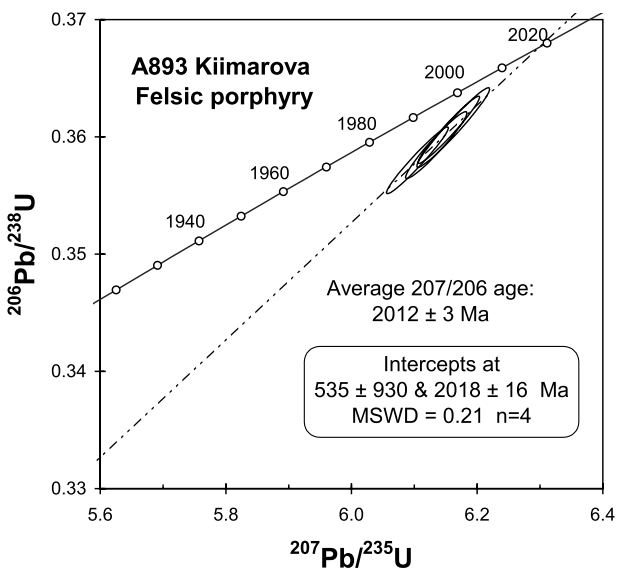

Fig. 21. Concordia diagram of zircon analyses from the Kiimarova felsic porphyry.

of quartz and albite. Additional minerals are tremolite-actinolite, titanite, chlorite, tourmaline and zircon. The chemical composition of the Kiimarova felsic porphyry corresponds to calc-alkaline dacite and is similar to that of the Veikasenmaa porphyries discussed previously.

Zircon generally occurs as small, euhedral, clear and short (L/B c. 2) prisms. The four analyzed density fractions (Fig. 21) are nearly concordant and bunch so close together that any statistical calculation is futile. The average $^{207}Pb/^{206}Pb$ age, 2012±3 Ma, should be a minimum and close to the age of the emplacement of the rock.

Altogether, the four felsic porphyries (A280, A581, A887 and A893) discussed above provide reliable zircon ages demonstrating that this felsic magmatism occurred about 2015 Ma ago. Geological and geochemical criteria indicate an intimate relationship in space and time with the mafic volcanic rocks observed to occur with them.

1.92 Ga porphyry dykes

A246 Nyssäkoski. The sample was taken from the western shore of River Kapsajoki at Nyssäkoski c. 4 km SSW of the sampling site of A280 (Fig 1). A 20-m-thick quartz-feldspar porphyry dyke crosscuts the volcanic rocks of the Kittilä Group. The SW-NE-trending contact with the mafic country rock is sharp and vertical. Abundant, light-colored plagioclase phenocrysts (0.3-7 mm) are distinguished on the weathering surface. Compositionally they are albitic and commonly sericitized and stained with small opaque inclusions. Quartz phenocrysts (0.1-3 mm) are bluish and mostly euhedral. Light green, prismatic amphibole (0.1-3 mm) and minor biotite occur among the phenocrysts. The groundmass is very fine-grained and consists mainly of albite, quartz and lesser amounts of amphibole, biotite, opaque minerals, titanite and carbonate.

The zircons are pale-brown and transparent, with simple prismatic and pyramidal crystal faces dominating. Although some of the larger crystals have been broken during milling, L/B seems to vary from 2.5 to 4. Oil immersion reveals a faint oscillatory zoning typical of magmatic zircons, but no significant inclusions were detected.

The five zircon fractions analyzed (Fig. 17) are slightly discordant and define a linear trend with intercepts at 1919±8 Ma and 160±350 Ma (MSWD 2.5). A peculiar feature of this magmatic zircon is the very low Th/U ratio, which can be seen from the radiogenic $^{208}Pb/^{206}Pb$ ratio listed in the table (Appendix I).

Similar porphyry dykes are found in many other places among the mafic metavolcanic rocks of the Kittilä Group and their crosscutting nature is even more obvious than at Nyssäkoski (Fig. 1). So far, only the Nyssäkoski dykes and an analogous cobble in the Mantovaara conglomerate (A1470) have been dated.

1.88 Ga felsic volcanic rock

Among the Proterozoic felsic porphyries in the Kittilä area, the Latvajärvi Formation belonging to the lower part of the Lainio Group stands out as a type of its own (Lehtonen et al. 1998). This formation occurs as a 2x12 km, N-S-trending unit c. 15 km SSW of Kittilä (Fig. 1). It is dominated by felsic volcanic rocks, although mafic variants are also met with. Small tuffitic and volcanic conglomerate intercalations are common as well as thin beds of fragmented and brecciated rocks. A small iron ore deposit has been discovered in the Latvajärvi Formation (Puustinen et al. 1980). All rocks belonging to the formation are strongly schistose and a W-trending lineation is clear.

On the weathering surface, the felsic porphyries are greyish or light brick red. Ignimbrite-like features, i.e. small fiamme-lenses (flame structure), are seen locally. The rocks contain potassium and sodic feldspar phenocrysts up to 8 mm in size in a fine-grained groundmass composed of potassic feldspar, quartz, plagioclase, biotite and muscovite. Rare amphibole and quartz phenocrysts are also present. Accessory phases are chlorite, tourmaline, leucoxene, apatite, zircon and opaque minerals. In addition, the rock often

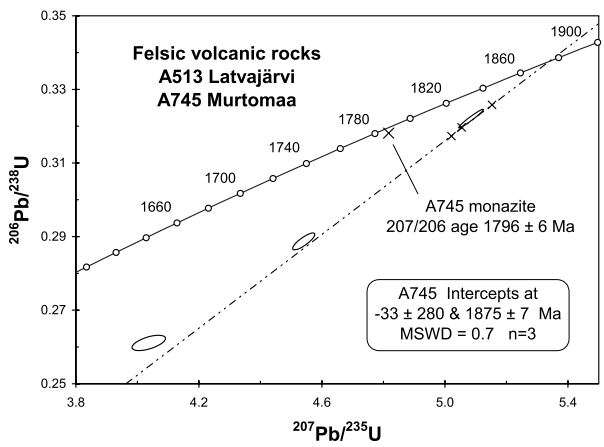

Fig. 22. Concordia diagram of zircon and monazite analyses from the Latvajärvi (error ellipses) and Murtomaa (crosses) felsic volcanic rocks.

contains scapolite porphyroblasts. Chemically, the Latvajärvi porphyries are calc-alkaline trachytes with average SiO_2 and K_2O contents of 62.1 and 8.9 %, respectively. Chondrite-normalized REE patterns are LREE enriched with an average La_N/Sm_N of 4.2 (Lehtonen et al. 1998).

A513 Latvajärvi. The sample for age determination was taken from the NE side of Lake Latvajärvi (Fig. 1). The rock is a typical feldspar porphyry, with abundant plagioclase and potassium feldspar phenocrysts in a fine-grained groundmass. Zircons separated from the sample are euhedral with sharp crystal edges, and often long prismatic (L/B can be >5). Some grains have a tabular appearance. Due to an inferior Pbic-run, data for one of the fractions (B) exhibits large errors. The other analyses define a reference line with intercepts at 1875 and 110 Ma. In view of the relatively small degree of discordancy of fraction C, the upper intercept appears to be a plausible age estimate (Fig. 22).

A745 Murtomaa. Another porphyry sample from the Latvajärvi Formation was taken at Murtomaa (Fig. 1), 6 km south of Latvajärvi. This rock contains a smaller proportion of phenocrysts in comparison to A513 and is strongly foliated. The sampling site is less than 10 m from the contact with an overlying conglomerate assigned to the Vesikkovaara Formation (Lainio Group). Porphyries derived evidently from the Latvajärvi Formation (A1416, p. 121) are found among the rounded fragments in this polymictic conglomerate. The conglomerate matrix contains detrital felsic volcanic material.

Zircons separated from the porphyry sample are small, clear, euhedral prisms with L/B of c. 2. The concentration of common lead is low and the three analyses are technically very good. However, due to the "bunching effect", the error limit is unduly large. The concordia intercepts are at 1875±7 Ma and -33 Ma (MSWD = 0.7). The $^{207}Pb/^{206}Pb$ ages for the three fractions (1874-1876 Ma) can be taken as a minimum age for zircon (Fig. 22). As the analyses are nearly concordant, these should be close to the real age. This interpretation is also consistent with the data obtained for sample A513. When the six analyses on these two samples are regressed together, the intercepts are at 1880±8 Ma and 204 Ma (MSWD=3). The marginally discordant monazite has a $^{207}Pb/^{206}Pb$ age of 1796±6 Ma, which is consistent with titanite and monazite data from elsewhere in western Lapland.

U-Pb STUDIES ON CLASTIC METASEDIMENTS

Epiclastic metasediments in the Kittilä greenstone area occur principally at two stratigraphical levels separated by the extensive volcanic formations of the Savukoski and Kittilä Groups. The older metasediments, mostly meta-arkoses and quartzites, belong to the Sodankylä Group and are often characterized by a greenish color due to the presence of chromian mica (fuchsite). The younger metasediments are coarse-clastic meta-arkoses and conglomerates of the Lainio and Kumpu Groups. In order to get some constraints on the average provenance and age of deposition, radiometric datings were performed on multi-grain zircon fractions separated from quartzites as well as from felsic cobbles in conglomerates.

Detrital zircons in quartzites

A961 Virttiövaara. The Virttiövaara quartzite of the Sodankylä Group (Fig. 1) has been studied in detail by Nikula (1985), who divided the quartzite formation into three members: the sericite schist-orthoquartzite member, the fuchsite quartzite member and the sericite quartzite-orthoquartzite member. Sedimentary facies analysis indicates that the deposition occurred in a tidal environment. The analyzed sample belongs to the sericite quartzite-orthoquartzite member. The texture of the sericite quartzite is blastoclastic and medium- to coarse-grained (0.2-3 mm). It is composed of monocrystalline quartz clasts in a sericite-rich matrix. Opaque minerals and zircon are accessory components. Thin, brown bands in the quartzite are due to concentrations of rounded rutile.

Zircons are generally rounded, but some grains still have good crystal surfaces with only slightly eroded edges. The seven analyzed fractions plot roughly on a line which gives an upper intercept at c. 2.8 Ga. The large MSWD (58) shows, however, some scatter in excess of analytical error, which is naturally expected of a detrital zircon population (Fig. 23). In addition to standard density fractions, four different morphological populations were handpicked and abraded from the heavy fraction. Two of these (F colorless, very clear; G pale-violet) are very low in uranium and concordant at 2.8 Ga (Appendix II).

A1176 Hyyverova. The Hyyverova sample was

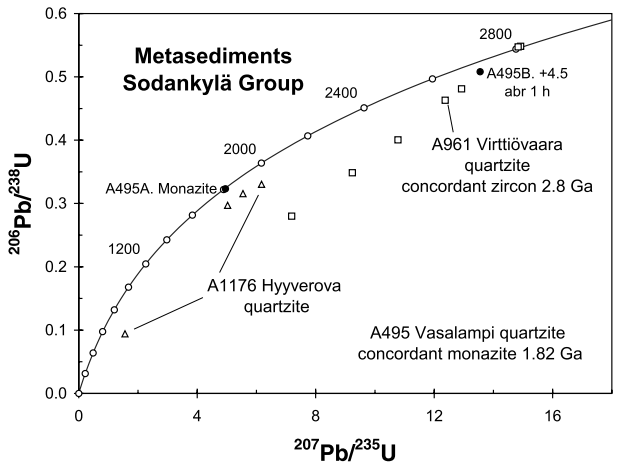

Fig. 23. Concordia diagram of zircon and monazite analyses from metasedimentary rocks of the Sodankylä Group.

taken from a sillimanite-bearing quartzitic gneiss (Fig.1). A stratigraphical position in the Sodankylä Group has been suggested (Lehtonen et al. 1998). The sampling site lies 2 km SE of Lake Luosujärvi, by River Luosujoki in the south end of the Yllästunturi fell. Sillimanite-bearing gneiss of high metamorphic grade occurs widely to the southwest of Hyyverova (Väänänen 1992, 1998). The rock is strongly lineated and banded with mica- and quartz-rich layers. The sample is dominated by quartz (0.1-2 mm, with strong undulatory extinction), accompanied by lesser amounts of biotite and its alteration product, fibrous sillimanite. Minor phases include potassium feldspar, plagioclase, muscovite, tourmaline, garnet, opaque minerals, titanite, apatite and zircon.

The zircon in the rock is relatively fine-grained (90%<70 µm), mainly pale-brown. Simple prismatic and pyramidal crystal faces are common, but also fairly rounded grains occur. Generally crystal edges are more or less rounded, but in contrast to pitted surfaces encountered in other quartzite zircons in this study, the surfaces are shiny. The effects of transportational abrasion are not conspicuous.

The four analyzed fractions are discordant and form no coherent pattern on the concordia diagram. The $^{207}Pb/^{206}Pb$ ages range from 1.96 to 2.17 Ga (Fig. 23). Although metamorphic effects on the U-Pb systematics remain obscure, it is likely that the zircon population primarily represents mixed Paleoproterozoic and Archean sources. Consequently, the average provenance and abrasion pattern of zircons are different from the Virttiövaara quartzite described above.

A495 Vasalampi. The sample for isotopic work was collected from a narrow group of outcrops between River Aakenusjoki and Lake Vasalampi in the Aakenusjoki-Housuoja shear zone (Fig. 1; Rastas 1984). Stratigraphically, the Vasalampi arkosite is thought to belong to the Sodankylä Group and has been thrust tectonically onto tuffites of the Savukoski Group, which form the principal rock type in the area. The metamorphic grade in the region is relatively low. The sampled rock, an arkosic quartzite, is nearly white on the weathered surface, but looks pinkish in a fresh cut. The rock is even- and fine-grained with a granoblastic texture and shows distinct cross-bedding and vague ripple-mark structures. Quartz, albitic plagioclase and carbonate are the main constituents, while biotite, chlorite and zircon form the accessory minerals. The proportion of quartz and feldspar varies, but the sampled rock has a relatively high content of feldspar.

Zircon separated from the rock is well-rounded, whereas monazite is euhedral and transparent, and characterized by very sharp crystal edges and clear surfaces. U-Pb analyses show that the detrital zircon is on an average Archean, like zircon from the Virttiövaara quartzite (A961, Fig. 23). The age of the metamorphic monazite is 1817±6 Ma.

A802 Ylläs. The Ylläs Formation, which has been assigned to the Lainio Group, forms a sedimentary unit reaching almost 2 km in thickness. Siltstone, sericite quartzite, arkosic quartzite and orthoquartzite, often as alternating beds, were deposited on a polymictic basal conglomerate. The metamorphism has been high enough to destroy the original clastic texture of the quartzites. However, local cross bedding is observed. Discrete rock fragments composed of vein quartz, quartzite, schists, and most strikingly red jasper occur in the Ylläs Formation quartzite. In addition, some conglomerate interlayers are known, the most notable being the Kellostapuli conglomerate (see p. 121).

The quartzitic sample for dating detrital zircon from the Ylläs Formation was picked at the summit of the Yllästunturi fell (Fig. 1). The rock is a white, massive,

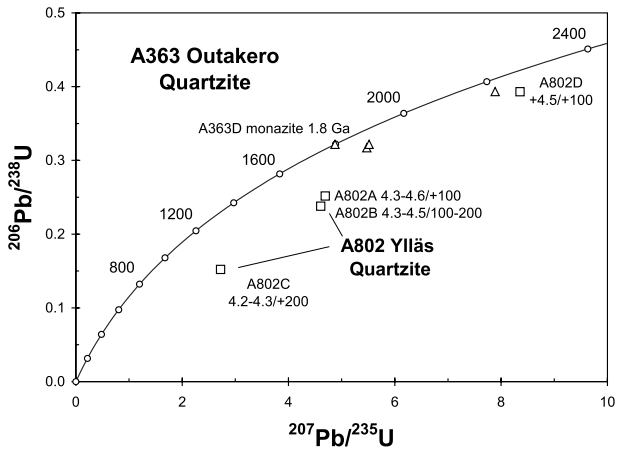

Fig. 24. Concordia diagram of zircon and monazite analyses from the Outakero and Ylläs quartzites.

granoblastic and small-grained (0.1-1 mm) orthoquartzite. Besides quartz, it contains orientated muscovite flakes and sporadic titanite, zircon and opaque minerals. In places, the primary bedding is feebly visible.

The four analyzed zircon fractions have Paleoproterozoic $^{207}Pb/^{206}Pb$ ages, and the zircon population shows heterogeneity on the concordia diagram (Fig. 24). The zircon is rounded, and it is likely that both Archean and Proterozoic detrital material contributed to these sediments.

A925 and A942 Värttiövaara. In the Värttiövaara area (Fig. 4), about 7 km N of Virttiövaara and 11 km NW from Jeesiö, quartzites and conglomerates belong to the Kumpu Group, which is the youngest quartzite sequence in Lapland. A 100-200 m thick, approximately E-W-trending conglomerate unit is bordered by quartzites both north- and southwards. Sample A942 represents the northern, and sample A925 the southern quartzite. The Värttiövaara metadiabase (A916) discussed previously (see p. 103) occurs at the contact between the conglomerate and the southern quartzite. The sites of the two quartzite samples are close to this metadiabase, sample A925 10 m and sample A942 70 m from it.

Nikula (1985) studied the sedimentology and depositional environment of the metasediments in the Värttiövaara area. The quartzite north of the conglomerate (quartzite II, sample A942), is a light-grey, fairly homogeneous, medium-grained sericite quartzite. Most clasts are composed of well-rounded quartz. Granoblastic quartzose rock fragments are common. Accessory clast material includes opaque minerals, K-feldspar and zircon. The matrix consists almost entirely of sericite. The quartzites south of the conglomerate (quartzite I, sample A925) are also medium-grained, blastoclastic sericite quartzites. The clasts are rounded quartz, plagioclase and a variety of rock fragments. Opaque minerals, zircon, titanite and muscovite are the main accessories and the matrix is composed of sericite, fine-grained quartz, plagioclase and occasional carbonate. Nikula (1985) interpreted the Värttiövaara metasediments as braided river deposits.

Zircons separated from these quartzites are variably rounded. The three analyzed fractions from sample A942 indicate that the zircons were derived from several sources (Fig. 5). One of the analyses (A942B) is nearly concordant at c. 1990 Ma, whereas clear zircons (C) suggest an even younger age ($^{207}Pb/^{206}Pb$ age of 1903 Ma). An analysis on monazite (A925B) is concordant within experimental error at 1891 Ma. The data clearly show the difference between the average provenance of quartzites in the Kumpu and Sodankylä Groups (A961 discussed above). Some monazite grains are rounded, but generally the effects of abrasion are inconspicuous. If monazite originates from rock fragments, it is conceivable that no signs of abrasion are visible, and the judgement whether the analyzed material is detrital or authigenic becomes difficult. Monazite in this sample appears to be quite different from the authigenic monazite in the Vasalampi quartzite, which is euhedral and transparent.

Provided that monazite and clear zircon (C) are detrital, the deposition of the Kumpu Group quartzites at Värttiövaara occurred after 1.9 Ga. This interpretation is supported by recently obtained NORDSIM data on detrital zircons from sample A942, where both Archean and Proterozoic grains, some younger than 1.9 Ga, were encountered (Hanski et al. 2000).

It is important to point out here that the U-Pb zircon age obtained for the Värttiövaara metadiabase, 2210±4 Ma, is older than the ages obtained for the detrital minerals from the Värttiövaara quartzite. Hence, it is unlikely that the diabase intruded into the sediments but was tectonically juxtaposed against the Kumpu Group quartzites.

A886 Sätkänävaara. At Sätkänävaara, 10 km N of Kittilä (Fig. 1), Kumpu Group quartzites and conglomerates of the Sirkka type (Kortelainen 1983) are in contact with an older sequence of metavolcanic and metasedimentary rocks of the Savukoski Group (Fig. 8). On the basis of field observations, the 2046 Ma metadiabase (A655, p. 105) is also considered older than the deposition of the Kumpu Group metasediments. As the youngest formation of the Kittilä area, the Kumpu Group rocks did not suffer the earliest deformation phases observed in Central Lapland (Lehtonen et al. 1998).

Sample A886 was taken from the coarse-clastic, cross-bedded quartzite about 10 m from the contact with the underlying schists. Quartz clasts (1-15 mm) occur in a matrix of sericite, muscovite and carbonate. The magnetite content is rather high both in the quartzite and the intervening conglomerate and the Sätkänävaara Formation is clearly visible on magnetic maps.

The three analyzed zircon fractions scatter randomly with $^{207}Pb/^{206}Pb$ ages of 1954-2019 Ma. Zircons of fraction B are large and rounded. The fraction with the youngest Pb-Pb age consists of long, handpicked crystals. A U-Pb analysis on monazite is concordant at 1905 Ma. In contrast to the monazite from Värttiövaara (A925B), monazite crystals in sample A886 have often sharp edges and good faces. If the monazite is detrital, its age puts strict constraints on the age of deposition of the sediment. The analyses

on zircon strongly suggest that the quartzite contains material younger than the adjacent 2046 Ma metadiabase (Fig. 9).

A363 Outakero. The sample was collected at the southern end of the Linkukero fell (Outakero) close to the border of the Muonio and Kittilä communes (Fig. 1). The white, medium-grained and clastic quartzite belongs to the Linkukero Formation of the Kumpu Group (Fig. 1; Rastas 1984). Its clasts in a sericite matrix are mostly quartz but also some angular plagioclase is found. Fine-grained, grey silt interlayers and sporadic rock fragments also occur. Cross-bedding is a common feature.

Zircons separated from the Outakero quartzite are clearly less rounded than those from the Ylläs quartzite (A802). In the density fraction of 4.2-4.3 g/cm^3, some zircons are fairly sharp-edged. Two analyses are discordant with the $^{207}Pb/^{206}Pb$ ages exceeding slightly 2.0 Ga. The third analysis was made on heavy and rounded zircons and gives older age indications (Fig. 24). The monazite is concordant at c. 1800 Ma, and may reflect the same thermal episode that is found in the Kolari-Muonio area (Väänänen & Lehtonen 2001, *this volume*).

Keimiöniemi supracrustals

A970 Keimiöniemi. Southwest of the Keimiöniemi fell (Fig 1), there occurs a belt of supracrustal rocks which has been considered to form a part of the Matarakoski Formation of the Savukoski Group. In addition to mica and graphite-sulphide schists, this poorly exposed belt contains a fine-grained, schistose rock of pale-grey, occasionally greenish or purplish color. A sample was taken from this rock type, which is probably epiclastic in origin, but a pyroclastic derivation cannot be totally excluded. The rock is mostly massive with occasional thin laminar structures. A microscopic study revealed that the granoblastic rock is composed of quartz, albite and green biotite as well as minor opaque minerals.

The zircon population as a whole is heterogeneous and most grains are rounded with corroded and pitted surfaces suggesting a detrital origin. Occasionally relics of crystal faces are observed. The seven analyzed zircon fractions scatter on the concordia diagram (Fig. 25), and have $^{207}Pb/^{206}Pb$ ages from 2.13 to 2.19 Ga. This may indicate an average provenance, as multi-grain analyses do not provide detailed information.

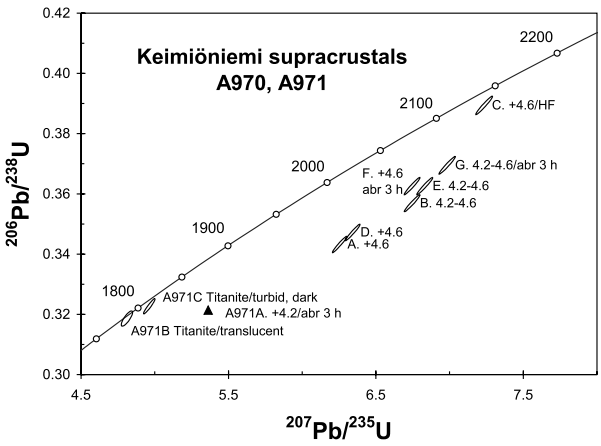

Fig. 25. Concordia diagram of zircon and titanite analyses from the Keimiöniemi supracrustal rocks.

A971 Keimiöniemi. This sample is a dark grey rock with porphyritic appearance. The matrix of the rock is small-grained and schistose. It contains quartz, albite and biotite as the main, titanite, epidote and carbonate as accessory minerals. The phenocrysts form aggregates of quartz and albite as well as uralite (0.5-1.5 mm). The rock can be pyroclastic in origin and, hence, may belong to the same sequence as the tuffitic sample A970 above. However, a tectonic origin cannot be totally excluded in this case (microbreccia).

Sample A971 is characterized by a high content of titanite whereas the amount of zircon is low. In contrast to A970, zircons are not rounded. Both simple prisms and multi-faceted grains occur, and the population looks heterogeneous also due to a variety of colors from dark to limpid. An analysis on abraded heavy zircon is discordant and gives a $^{207}Pb/^{206}Pb$ age of 1.97 Ga, which should be considered as a minimum age for the analyzed material. Two varieties of titanite have been picked for analysis. A nearly concordant analysis on dark, turbid grains gives an age of 1825 ±5 Ma whereas translucent crystals provide a slightly younger age of c. 1792±10 Ma.

Felsic cobbles in conglomerates

A939 Linkupalo. The Linkupalo volcanic park, opened by the Geological Survey of Finland and the Forest and Park Service in 1989, lies 30 km W of Kittilä by the road leading to Äkäslompolo. At this site

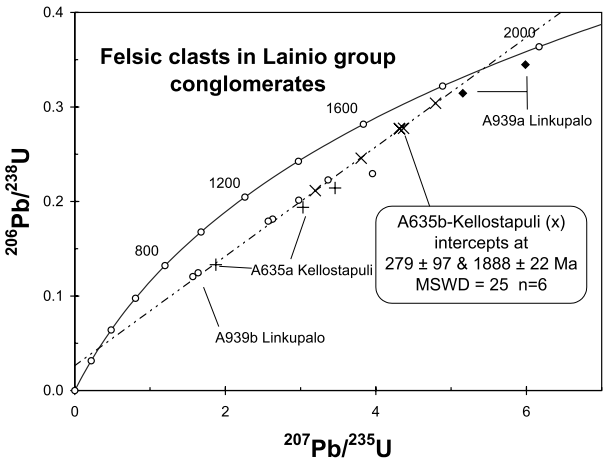

Fig. 26. Concordia diagram of zircon analyses from felsic clasts of the Linkupalo and Kellostapuli conglomerates of the Lainio Group.

there occur conglomerates and various mafic metavolcanic rocks including pillow lavas, variolitic and massive lavas, subvolcanic mafic sills, volcanoclastic rocks, and tuffites. Just by the road, a breccia has been exposed which is interpreted as the eruption pipe of an ancient volcano.

The mafic metavolcanic rocks at Linkupalo belong to the Savukoski Group and are overlain by a 50 m thick, polymictic conglomerate assigned to the Lainio Group (Fig. 1; Rastas 1984). It contains rounded, felsic porphyry fragments ranging from 10 to 30 cm in diameter. Phenocrysts in these porphyries are mainly formed by euhedral, zoned plagioclase 2-15 mm in size. Quartz, biotite and amphibole can also occur as phenocrysts. In addition to these minerals, the groundmass consists of carbonate, epidote, chlorite, scapolite and opaque minerals, mainly magnetite.

Two porphyry samples were collected for age determination. Sample A939a contains relatively little stubby, turbid zircon, which is Paleoproterozoic in origin, apparently older than the $^{207}Pb/^{206}Pb$ age of 2.04 Ga measured for the heavy fraction. Sample A939b contains more zircon, and an effort was made to hand-pick transparent, euhedral fractions (analyses A, B and D). Six of the seven fractions analyzed from this sample are from very low-density, high uranium zircon, and are discordant plotting close to a reference line defined by zircons from the Kellostapuli sample below (Fig. 26). The deviating fraction E from slightly heavier zircon is also very discordant with a $^{207}Pb/^{206}Pb$ age in excess of 2 Ga (Fig. 26). As the data on the fragments are heterogeneous, the age of deposition cannot be precisely constrained.

A635a, A635b Kellostapuli. The Kellostapuli conglomerate unit, described already by Hackman (1927), occurs in the quartzitic Ylläs mountain range (Lainio Group) at Äkäslompolo. Stratigraphically it belongs to the upper part of the Ylläs Formation (Fig. 1; Rastas 1984). The N-S-trending unit has been exposed as a strip 40 m in width. The fragments of this polymictic conglomerate consist mainly of rounded granite with the diameter varying between 5-30 cm. In addition, occasional angular schist fragments are found. The major minerals of the granitic fragments are albitic plagioclase, quartz and varying amounts of K-feldspar while the accessory minerals include biotite, titanite, muscovite, apatite, opaque minerals and sometimes amphibole.

Two types of fragments were collected from the Kellostapuli conglomerate for age determination. Sample A635a consists of red granitoids (monzosyenites), whereas fragments of sample A635b are slightly more mafic, coarse-grained and slightly deformed in hand specimen. Zircon crystals are different in these two rock types. Zircon in A635a occurs as translucent, yellowish grains in the +4.0 g/cm^3 fraction, but is more greyish in lighter fractions. Many crystals in the heaviest fraction have sharp edges with L/B up to 3. The density fraction 3.8-4.2 g/cm^3 is heterogeneous containing rounded grains, which may have originated from the matrix of the conglomerate. Zircon in sample A635b occurs as translucent, brownish and short prisms (L/B 1-2) with sharp crystal edges.

The three analyzed fractions from A635a are very discordant (Fig. 26). Although no linear fit can be calculated, the $^{207}Pb/^{206}Pb$ age in fraction A (1913 Ma) suggests an origin older than 1900 Ma. However, if some zircon grains in this sample originated from the matrix, as suggested above, this result does not necessarily constrain the age of the granite fragments.

Six analyses from sample A635b are less discordant and define a reference line with an upper intercept at 1888±22 Ma and a lower intercept at 279 Ma. The large MSWD (25) suggests a scatter in excess of analytical error. Still, the upper intercept may be interpreted as indicating the magmatic crystallization age of the granitoid fragments, meaning that their age is coeval with the Haaparanta suite intrusions. This interpretation is also consistent with the petrographical and geochemical data, as the fragments are similar to the quartz monzonite of the 1.89 Ga Härkimännikkö felsic intrusion (see Väänänen & Lehtonen 2001, *this volume*) c. 10 km SE of Kellostapuli. Furthermore, the age 1888±22 Ma obtained for the granitoid cobbles in the Kellostapuli conglomerate demonstrate that the age of deposition of the conglomerate (and the whole Lainio Group) must be younger than this. Deformation of the granitoid took place before its exposure by erosion.

A1416b Vesikkovaara. Sample A1416 is from a conglomerate interlayer in the quartzites of the

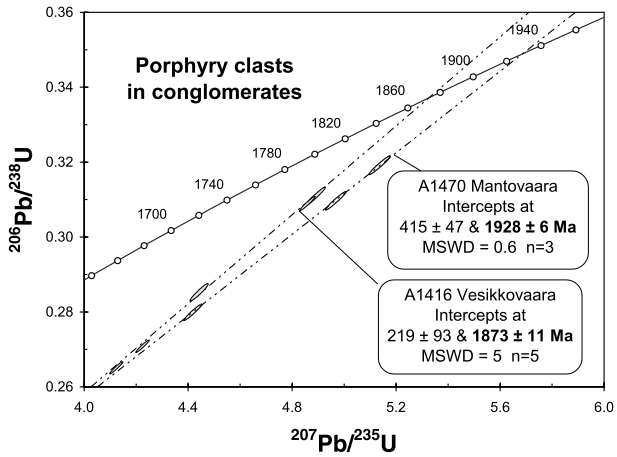

Fig. 27. Concordia diagram of zircon analyses from porphyry clasts in the Vesikkovaara and Mantovaara conglomerates.

Vesikkovaara Formation belonging to the Lainio Group (Fig. 1; Rastas 1984). Two porphyry cobbles (A1416a, 1416b) were extracted from the conglomerate. Phenocrysts in the porphyries are mainly formed by euhedral, albitic plagioclase and slightly rounded quartz. A sufficient amount of zircon for isotopic analysis was acquired only from sample 1416b. Zircon is mostly euhedral, stubby (L/B 2-3), with simple morphology and often translucent. In most grains, the color is normal reddish brown, but light-colored crystals are also present.

Two analyses were made on the most abundant density fraction 4.0-4.2 g/cm^3 by conventional techniques and a third one as well as two analyses on the density fraction >4.2 g/cm^3 by the "microtechnique", for which the blank level is below 30 pg Pb. A regression line drawn through all five analytical points gives an upper intercept at 1873±11 Ma and a lower intercept at 219 Ma (MSWD=5, Fig. 27). One of the older analyses is slightly off the line, and omitting this, the upper intercept would be 1873±3 Ma and MSWD = 0.4. In contrasts to a normal situation, analyses on the heavy fraction (d>4.2 g/cm^3) are more discordant than the lighter ones. The age is comparable with that of the U-Pb zircon age of the Latvajärvi felsic porphyry and a similar age range is also indicated by the granitoid fragments from the Kellostapuli conglomerate. These isotopic data strongly imply a relatively young age (<1.88 Ga) for deposition of the Lainio Group metasediments.

A1470 Mantovaara. The Mantovaara conglomerate (Fig. 1) occurs as an E-W-trending, 1-km-wide and 5-km-long rock unit, situated about 10 km NE of Sirkka. It was recognized already by Hackman (1927) as being similar to the Kumpututunturi conglomerates and thus belonging to the "Kumpu quartzites". The conglomerate lies discordantly on Kittilä Group mafic metavolcanic rocks and contains a variety of rock fragments in an arkosic matrix. The fragments range up to 0.5 m in size and include mafic volcanic rocks and jasper obviously derived from the underlying Kittilä Group. A special feature is the presence of numerous fragments of felsic porphyry and non-schistose, clastic quartzite with a purplish tint. In addition, a few aplite granite cobbles have been found.

In order to constrain the depositional age of the Kumpu Group, we sampled a rounded felsic porphyry cobble with a diameter of 20 cm and extracted a zircon fraction from it. Major and trace element as well as Nd isotopic analyses revealed that the rock is identical to the younger porphyry type cutting the mafic metavolcanic rocks of the Kittilä Group (Hanski et al. 1997) which allowed us to expect an age slightly higher than 1.9 Ga as recorded for this porphyry type at Nyssäkoski (A246, p. 116)

The zircons from the porphyry fragment display mainly euhedral, stubby forms and are translucent-reddish in color. Although there is some variation in their grain size, the zircons give an impression of a homogeneous, magmatic population. Three fractions were analyzed (Appendix II). As is seen in Figure 27, they are somewhat discordant (the degree of concordance 90-76%) but are well-aligned along a line crossing the concordia curve at 1928±6 Ma (lower intercept at 415 Ma). This can be regarded as a good estimate of the crystallization age for the felsic porphyry. From this result we can draw two conclusions. Firstly, as expected, the age is very similar to that of the Nyssäkoski dyke (1919±8 Ma) and thus further corroborates the genetic link between the cobble in the conglomerate and the felsic dyke type represented by the Nyssäkoski porphyry. Secondly and more importantly, the age of the cobble demonstrates that the Kumpu Group sediments were deposited later than 1927 Ma ago.

U-Pb STUDIES ON GRANITOIDS

Granitic rocks in the Kittilä area are mainly limited to the northern and southern sides of the volcanic belt, with the exception of the Ruoppapalo granodiorite belonging to the Hetta Complex and occurring within the Kittilä Group metavolcanic rocks in the northern part of the study area. The Kallo quartz monzonite

massif in the southwest partly occupies the Kittilä area, but the radiometric data for this Haaparanta Suite granitoid (A747 Härkimännikkö, 1.89 Ga) are reported in connection of other results from Western Lapland (Väänänen & Lehtonen 2001, *this volume*). In the south, the volcanic belt is bounded by the Central Lapland Granite Complex and in the north by the Hetta Complex (Hetta granites), both being very heterogeneous regions of granitoid rocks. Earlier studies dealing with zircon morphology and radiometric dating from the Central Lapland Granite Complex indicate a multi-phase crustal history with zircon ages from Archean to 1770 Ma (Lauerma 1982, Huhma 1986). Neither is the age of the Hetta Complex granitoids a simple matter: earlier studies indicate both mixed zircon populations (A891-Kappera and A355-Peltovuoma, Mänttäri 1995 and Lehtonen 1984) and a well-defined age of 1810±14 Ma (A181-Peltovuoma, Huhma 1986). One new sample for the Hetta Complex proper (Ruoppapalo) is included in this report along with two samples from a separate massif within the Hetta Complex granitoids (Tepasto granite; Kouvo et al. 1983, Huhma 1986). West of Kittilä there is another granite massif, the Kittilä granite (previously known as the Central Kittilä granite, Huhma 1986), from which samples for isotope dating have been taken at three localities.

Kittilä granite

The Kittilä granite covers an area of 10x20 km^2 just WSW of Kittilä (Fig. 1; Rastas 1984). It contains many inclusions of country rocks such as quartzite, mica schist, amphibolite, gabbro and diabase. Two different phases can be distinguished on magnetic maps: the magnetic eastern block and the weakly magnetic western block. Three samples were taken for isotope studies. Sample A320 Isovaara comes from the magnetic block while samples A321 Hangasselkä and A428 Lappalaislampi represent the less magnetic part of the massif.

A320 Isovaara. Isovaara lies 5 km SW of Kittilä. The rock is a mostly red, medium-grained (0.5-4 mm) granite with local pegmatitic portions. Quartzite and mica schist inclusions have been found. The main minerals are plagioclase ($An_{<30}$), microcline and quartz occurring in equal amounts. Accessory minerals include biotite, chlorite, muscovite, opaque minerals, titanite, apatite and epidote. The texture is hypidiomorphic-granular and biotite is orientated giving the rock a slightly gneissose appearance.

The two analyzed zircon fractions are fairly discordant and cannot be used for precise dating (Fig. 28). A line through these points has intercepts with the concordia curve at 2090 and 555 Ma.

A321 Hangasselkä. The Hangasselkä sample was taken from a wide group of outcrops c. 4 km SW of Isovaara described above. The rock is mineralogically and texturally similar to the granite at Isovaara, but its color is somewhat greyer.

The three analyzed zircon fractions do not form a good linear array. The $^{207}Pb/^{206}Pb$ ages vary from 1824 to 1925 Ma, of which the oldest age was yielded by a HF-leached fraction (Fig. 28).

A428 Lappalaislampi. Sample A428 Lappalaislampi comes from a 50x100 m outcrop occurring on the same profile as A320 Isovaara and A321 Hangasselkä with the sampling site lying 4 km SW of the latter (Fig. 1). The rock is a red-colored, medium-grained granite. Pegmatitic portions and dykes are

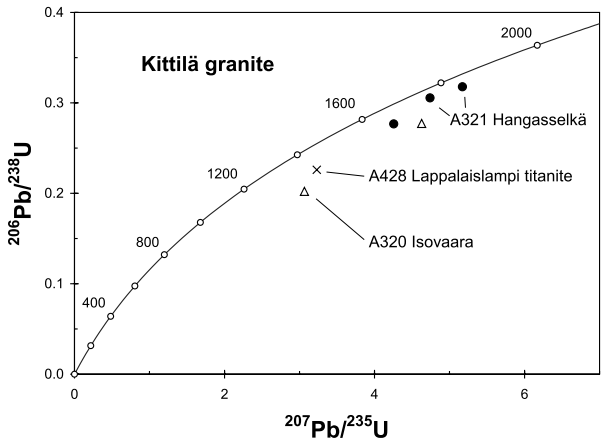

Fig. 28. Concordia diagram of zircon and titanite analyses from the Kittilä granite.

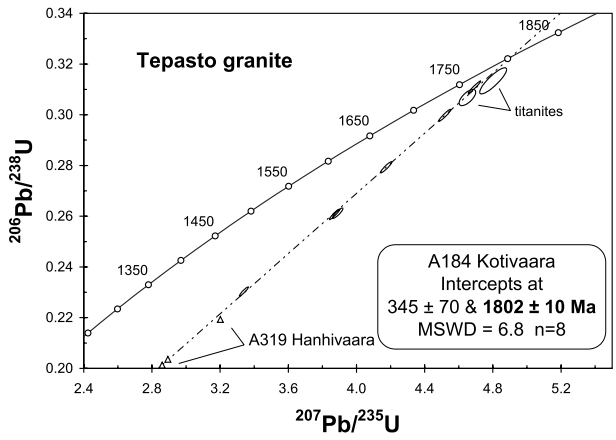

Fig. 29. Concordia diagram of zircon and titanite analyses from the Tepasto granite.

found locally.

The amount of zircon in this sample was very small and only titanite was analyzed. The result is, however, fairly discordant (Fig. 29) and is only suggestive of an age of about 1.8 Ga. The origin of this granite has been studied by Huhma (1986) by using the Sm-Nd method. The Sm-Nd ratio of sample A428 is nearly chondritic and the ε_{Nd} of c. -10 suggests an origin from Archean crust. This result together with the scattered U-Pb data on zircon indicate that zircon may be partly inherited from the older (Archean) crust. The exact age of this granite remains unknown.

Tepasto granite

The Tepasto granite (Fig. 1) on the western side of the village of Tepasto in Kittilä belongs to the *Nattanen-type granites* as defined by Mikkola (1928, 1941). Other representatives of this type are the granites of Nattastunturit, Riestovaara, Salmurinvaara, Pomovaara in the Sodankylä parish, the Vainospää granite NE of Lake Inari and a chain of granites on the Kola Peninsula (Juvoaivi, Litsa-Araguba; Haapala et al. 1987). These massive, homogeneous granites are medium- to coarse-grained. Microcline and plagioclase commonly occur in equal proportions and the amount of quartz is variable. Accessories include biotite, muscovite, titanite, orthite, apatite and zircon.

Aarnisalo et al. (1982), Front et al. (1989), Haapala et al. (1987), Lehtiö (1993), Pakkanen (1993) and Summanen (1993) have investigated the Tepasto granite and its environment. According to these studies, the post-orogenic, zoned Tepasto stock comprises several, practically undeformed granitic intrusion phases. The oldest of them is coarse-porphyritic granite (Kotivaara) which was followed by biotite-bearing, medium-grained granite (Hanhivaara). The youngest intrusion is the Kokonpesävaara aplitic granite, which has generated a small molybdenum mineralization (Eeronheimo 1981, Pakkanen 1993). Samples for zircon dating were taken at two localities, Kotivaara and Hanhivaara (Fig. 1).

A184 Kotivaara. Sample A184 Kotivaara was taken from an outcrop immediately south of the Tepasto school. The rock is a light-grey, coarse-grained, porphyritic granite composed mainly of potassium feldspar, plagioclase ($An_{<30}$) and quartz. Perthitic microcline phenocrysts may reach 20 mm in length. Apatite, titanite, biotite, hornblende, epidote, magnetite and zircon are found as accessory phases. Myrmekitic texture is common. According to Front et al. (1989), coarse-grained, porphyritic granite forms the outermost zone of the Tepasto stock.

Separation yielded abundant zircon, which is stubby and transparent especially in the heavy fraction. Six zircon fractions from the Kotivaara sample exhibit a rather large variation in their degree of discordancy. Although there is probably some scatter in excess of analytical error (MSWD=7), the upper intercept age estimate is relatively precise at 1802±10 Ma, which is consistent with other results from the Nattanen-type granites (Huhma 1986). The two titanite analyses support this result (Fig. 29).

A319 Hanhivaara. The Hanhivaara sample A319 represents a very light-colored, medium-grained biotite granite occurring in the middle part of the Tepasto massif (Front et al. 1989). The rock is mainly composed of quartz, microcline and plagioclase ($An_{<30}$), which is slightly sericitized and in some cases zoned. Apart from biotite, other accessory phases are rare.

Analyses of the three zircon fractions from the Hanhivaara sample are so discordant that an age calculation is meaningless. They plot, nevertheless, in the vicinity of the extension of the linear regression for the zircons from sample A184, demonstrating that the two varieties of the Tepasto intrusion can be roughly coeval (Fig. 29).

Other granites and granitoids in the Kittilä area

Other granitic plutons in the Kittilä area, which were sampled for age determination, include two samples (Iso Nilipää and Pikku Nilipää) from the northern margin of the Central Lapland Granite Complex and a granodiorite at Ruoppapalo located in the northern part of the Kittilä greenstone area (Fig. 1).

A748 Iso Nilipää. The Iso Nilipää sample was taken at the northern margin of the Central Lapland Granite Complex c. 20 km southeast of Kittilä. The rock is a light-grey, coarse-grained, porphyritic granite with perthite-bearing microcline, plagioclase ($An_{<30}$) and quartz as major minerals. Euhedral microcline and less abundant plagioclase phenocrysts may reach 3 cm in length. Accessories include biotite, hornblende, titanite, epidote, apatite and zircon. The dark minerals and phenocrysts have often a preferred orientation

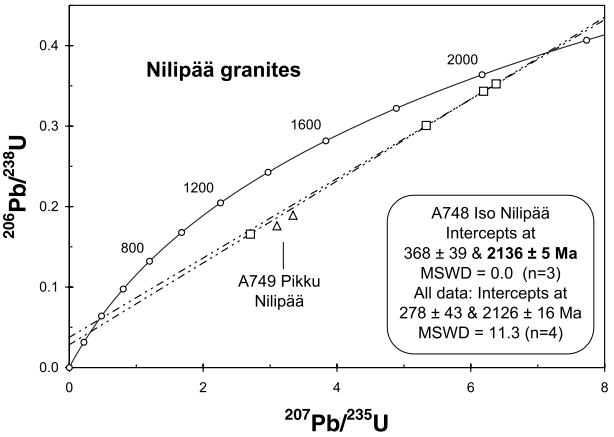

Fig. 30. Concordia diagram of zircon analyses from the Nilipää granites.

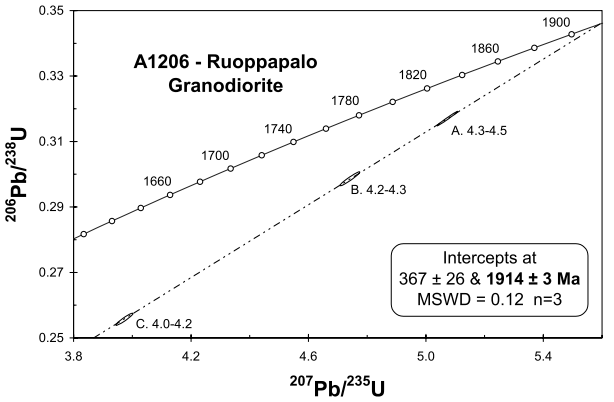

Fig. 31. Concordia diagram of zircon analyses from the Ruoppapalo granodiorite.

and the appearance is somewhat cataclastic.

Earlier isotopic data by Huhma (1986) provide an age of 2136±5 Ma (n=3, MSWD=0.0), and if the fourth very discordant analysis is included the upper intercept would be 2126±16 Ma (MSWD=11). Titanite from this sample has yielded an age indication of the same range. Zircon in this sample consists of euhedral zoned crystals without distinct cores and provides quite unusual age information for granite magmatism.

A749 Pikku Nilipää. The site of this sample is 3 km NW of Iso Nilipää. The rock is coarse-grained, red granite. Locally one gets an impression that the red granite phase might be younger than the porphyritic granite. The mineralogy is similar to that of sample A748 Iso Nilipää, with the exception that amphibole is not present.

A relatively small amount of zircon was gained from this sample. Zircon occurs as euhedral, red-brown, often turbid grains. The results of two analyses are very discordant with $^{207}Pb/^{206}Pb$ ages of c. 2.07 Ga. They plot slightly off the line defined by sample A748 (Fig. 30).

A1206 Ruoppapalo. The Ruoppapalo pluton lies in northern Kittilä. It is an inhomogeneous, oval body covering an area of ~10x5 km^2. According to gravimetric profiles, the contacts with the surrounding metavolcanic rocks of the Kittilä Group are relatively steep and the intrusive nature is visible also in the general appearance of the rock. The pluton consists of a variety of felsic intrusive rocks. In the exposed areas, the main types are granodiorites, tonalites, quartz diorites, grey granites, minor potassium granites and monzonites. They are generally medium- to fine-grained although coarser varieties also occur, especially among granodiorites and granites. Intersecting felsic dykes are frequently observed. They are mainly porphyritic albitites with brecciating quartz-carbonate vein networks. The granitic rocks are weakly to distinctly foliated and in some places also folded. Their syn- or late-orogenic origin seems obvious. Intersecting dykes are not folded but tectonic dislocations are common. The surrounding mafic volcanic rocks are metamorphosed to amphibolites within a distance of 1-3 km from the contact. Outside of this peripheral zone the metavolcanic rocks possess greenschist facies mineral parageneses. This phenomenon is interpreted to be a result of contact metamorphism.

Zircons for the age determination were separated from a typical medium-grained and weakly foliated granodiorite in the eastern part of the massif. Zircons are generally euhedral, translucent, reddish brown, simple prisms with L/B typically around 3. Only three analyses have been made and they form a good linear trend (MSWD=0.12) with intercepts at 1914±3 and 367±26 Ma (Fig. 31).

DISCUSSION AND SUMMARY

Depositional ages of the major supracrustal rocks in northern Finland, and, particularly, in the Kittilä greenstone area, are known only approximately as the prevailing mafic metavolcanic rocks generally do not contain primary igneous minerals for dating. Recently, however, Sm-Nd analyses on primary clinopyroxenes and whole rocks have given promising age results on mafic metavolcanic rocks. These include the ages of 1990±35 Ma for the Vesmajärvi Formation, Kittilä Group (Hanski et al. 1997), and 2063±34 Ma for the Jeesiörova komatiites in the Savukoski Group (Hanski et al, 2001). The chronostratigraphic scheme for the

Central Lapland Greenstone Belt is mainly based on dating of intrusive rocks, as well as analyses of detrital materials. Felsic volcanic rocks encountered in some formations have also been employed. Most age determinations utilize conventional multi-grain analyses on zircon, and problems with the homogeneity of the analyzed materials are often obvious. Nevertheless, the data produced over the decades provide several reliable age assessments.

The large Koitelainen and Akanvaara mafic layered intrusions (Mutanen & Huhma 2001, *this volume*), east of the Kittilä greenstone area, are products of the extensive mafic magmatism that occurred c. 2.44 Ga ago. They are known to have intruded metavolcanic rocks of the Salla Group. Age determinations (Räsänen & Huhma 2001, *this volume*, Manninen et al. 2001, *this volume*) suggest that at least some of these metavolcanic rocks are roughly coeval with the layered intrusions. All these magmatic rocks manifest a breakup of the Archean craton and form a violent prelude to the ensuing, long Paleoproterozoic supracrustal evolution.

The Sodankylä Group is dominated by epiclastic sedimentary rocks. These are accompanied by less abundant mafic metavolcanites and are intersected by the Haaskalehto-type mafic sills and dykes. Generally, coarse-grained gabbroic differentiates in these intrusions contain abundant zircon, which is turbid and provides discordant U-Pb analyses (e.g., A824, see also Perttunen & Vaasjoki 2001, *this volume*). In spite of this problem, several samples within the Jeesiö River valley between the communal centers of Kittilä and Sodankylä have yielded reliable upper intercept ages of c. 2.2 Ga, and thus give a minimum age for the supracrustal rocks of the Sodankylä Group. Some of these results are supported by U-Pb ages on titanite. In terms of their geochemistry, mode of occurrence as usually conformable bodies and age, these rocks perfectly correspond to the differentiated mafic-ultramafic sills occurring in eastern Finland (gabbro-wehrlite association, Hanski 1986) and the Peräpohja schist belt (Perttunen & Vaasjoki 2001, this volume), providing thus indirect evidence for potential correlation of their quartzitic country rocks (cf. Hanski 1987).

Komatiitic and Fe-tholeiitic metavolcanics and fine-grained metasediments are the principal rock types of the Savukoski Group. The sedimentary rocks have been intruded by metadiabases and gabbros which have yielded U-Pb zircon ages in the range of 2.06-2.0 Ga. In addition to those reported here, mafic intrusions within this age range have also been recognized elsewhere in Lapland including, e.g. the Keivitsa layered intrusion northeast of Sodankylä (Mutanen & Huhma 2001, *this volume*) and the Saarenpudas dyke in the Kolari area (Väänänen & Lehtonen 2001, *this volume*). These metadiabases and gabbros are considered to give a minimum age for the supracrustal rocks of the Savukoski Group (Lehtonen et al. 1998).

Large areas in Central Lapland consist of mafic volcanic rocks of the Kittilä Group. Mafic dykes associated with these metavolcanics are fine-grained and undifferentiated and generally yield no zircon for dating. On the basis of field observations, it is concluded that the generation of the mafic and felsic metavolcanic rocks of the Veikasenmaa Formation of the Kittilä Group were very closely related in space and time. The U-Pb ages of c. 2015 Ma obtained for the felsic porphyries at four different locations are regarded as representing the age of the upper part of the Kittilä Group (Veikasenmaa and corresponding Fms.). This result is consistent with the Sm-Nd isochron age of 1990±35 Ma reported by Hanski et al. (1997) for mafic metavolcanic rocks of the Vesmajärvi Formation which, on geological grounds, has been correlated with the Veikasenmaa Formation (Lehtonen et al. 1998). The correlation of these two formations is also supported by similar ε_{Nd} results (+3.7/+3.8, Hanski et al. 1998), indicative of a depleted mantle source for both formations. These ages are also consistent with the U-Pb zircon age of c. 1986 Ma obtained for the Kulkujärvi gabbro, which appears to be intruding the metavolcanic rocks of the Kittilä Group. The radiometric ages of the Kittilä Group are also in harmony with the results of carbon isotope data (a mean $\delta^{13}C_{Tot}$ value of +1.5±1.3‰) as the dolomites of the Kautoselkä Formation (Kittilä Group) are interpreted as post-Jatulian (Karhu 1993). According to Karhu (1993), all Jatulian sedimentary carbonates deposited c. 2.1-2.2 Ga are highly enriched in ^{13}C ($\delta^{13}C_{Tot}$ c. +10‰).

The current interpretation regards the Kittilä Group as an allochthonous complex representing a block of ancient oceanic lithosphere, which was obducted to its present position at c. 1.92 Ga (Hanski et al. 1998). This age is derived from the Nyssäkoski felsic dyke and the Ruoppapalo granodiorite, which intrude the metavolcanic rocks of the Kittilä Group.

The major stage of Svecofennian magmatic activity at c. 1.88 Ga is manifested in our study area by the Latvajärvi felsic metavolcanic rocks assigned to the Lainio Group, and the quartz monzonites of the Haaparanta Suite in Western Lapland (see also Väänänen & Lehtonen 2001, *this volume*). High-K metavolcanic rocks are also found in northern Sweden, where the so-called Kiruna porphyries fall in the same age range as the Latvajärvi trachytes (Skiöld 1986, 1987).

The U-Pb age determinations on detrital zircons

from quartzites of the Sodankylä and Kumpu Groups provide contrasting results. The apparent age of c. 2.8 Ga on detrital zircons from quartzites of the Sodankylä Group suggest at least a predominant, if not exclusive, Archean zircon provenance. The multi-grain detrital zircon $^{207}Pb/^{206}Pb$ ages from the Kumpu Group range from 1.91 to 2.02 Ga. Further constraints for a relatively young age of deposition are provided by rock fragments in conglomerates. A felsic porphyry cobble at Mantovaara, Kittilä (Kumpu Group) has been dated to 1928±6 Ma, and cobbles of felsic porphyry and deformed granitoid from the Lainio Group conglomerates give ages of 1873±11 Ma and 1888±22 Ma, respectively. Recent SIMS data also suggest that many detrital zircons are c. 1.88 Ga in age (Hanski et al. 2000). Consequently, both Kumpu and Lainio Groups are distinctly younger than the Jatulian quartzites in eastern Finland. Instead, possible Jatulian counterparts most probably are the quartzites assigned to the Sodankylä Group.

At Värttiövaara, a 2.2-Ga diabase is surrounded by quartzites and conglomerates containing detrital zircons with younger ages (p. 119). At the time of sampling, the diabase was considered to be intrusive with respect to the Kumpu-type quartzite, but trenching of the originally buried contacts showed that the contacts between the diabase and quartzite are strongly sheared. Tectonic emplacement of the diabase has been evoked to explain the apparently contradictory age results (Lehtonen et al., 1998). Alternatively, the diabase outcrops may represent a window in the metasediments exposing underlying older rocks.

The picture of a long Proterozoic supracrustal evolution of more than 500 Ma, described briefly above, is well-based on isotopic results. However, there still exist cases where the apparent age results are not easily reconciled with field geological observations. These include the felsic supracrustal rocks from Honkavaara and Venejoki. The two dated rocks from the Honkavaara area, corresponding chemically to metadacite or andesite, yielded ages of c. 2.72 Ga and 2.75 Ga. These Archean ages are at variance with the geological interpretation of the Honkavaara Formation as being the youngest member of the Paleoproterozoic Sodankylä Group. As the rocks are strongly altered, the volcanic origin can be questioned. The zircon population does not show distinct transportational abrasion, and despite some scatter, the analyses plot relatively well on a chord suggesting that zircons ultimately should represent a relatively homogeneous age population. The Sm-Nd systematics on whole rocks further show that the Honkavaara felsic supracrustal rocks as well as the sample from the Venejoki area have an Archean source (Geological Survey of Finland, unpublished data).

Conceivable explanations would include either tectonic (exotic) or sedimentary origins for these rocks. Lehtonen et al. (1998) considered the possibility of the Honkavaara Formation representing an uplifted Archean block. The sampled rock from Venejoki is a quartz-albite rock and resembles those at Honkavaara. All U-Pb analyses on zircon from Venejoki have Archean $^{207}Pb/^{206}Pb$ ages, but the data are clearly heterogeneous. Many zircon grains have a slightly rounded appearance and a sedimentary origin for this rock may be viable. Similar rocks with an Archean age have also been found elsewhere in Finnish Lapland (Räsänen & Vaasjoki 2001, *this volume*). Like the rocks at Honkavaara and Venejoki, they all seem to lie more or less around the Central Lapland Granite Complex. This could refer to an Archean tectonic wedge within the Proterozoic environment. More detailed studies and analyses on zircon would be needed to solve this problem.

In addition to the 1.9 Ga granitoids mentioned above, granites are the principal rock type in large areas of Central Lapland. The isotopic studies reported here and previously have shown that these granites often provide heterogeneous data due to inherited zircon (Lauerma 1982, Lehtonen 1984, Mänttäri 1995). Major involvement of Archean crustal material in the generation of most Proterozoic granites in Finnish as well as Swedish Lapland is also evident through the Sm-Nd isotope geochemistry (Huhma 1986, Skiöld et al. 1988, Öhlander et al. 1993).

The three samples from the Kittilä granite yielded relatively little zircon, which turned out to give discordant and scattered results. In contrast, the sample studied (A184) from the Tepasto granite contained abundant zircon and provided a good age of 1802±10 Ma. The Tepasto granite is a zoned, multi-phase intrusion and is classified as one of the post-tectonic, dominantly I-type granitoids of the Nattanen-type (Haapala et al. 1987, Front et al. 1989). Analogously with the Kittilä granite, it is considered as late-tectonic in origin. Archean crustal involvement in the origin of both the Kittilä and Tepasto granites is demonstrated by isotopic data (ε_{Nd} -5.7 for Tepasto and ε_{Nd} c. –10 for Kittilä, Huhma 1986).

The available isotopic data suggest that many granites in Lapland formed at about 1.8 Ga ago. These include the microcline granites of western Lapland (the Hanhipalo and Kihlanki granites, Väänänen & Lehtonen 2001, *this volume*) as well as the Lina granites in northern Sweden (the Kihlanki granite and the Lina granite join at the state boundary). According to Öhlander and Skiöld (1994), the Lina-type granitoids are c. 1.8 Ga old, minimum melt rocks derived from

continental crust. On the basis of U-Pb data on monazite, titanite and zircon, the bulk of the granites in Central Lapland Granite Complex, south of the Kittilä greenstone area, are also considered to have been formed at c. 1.8 Ga (Lauerma 1982, Huhma 1986). A significant exception to this concept is the Iso Nilipää granite from the northern edge of CLGC, which has provided an age of c. 2.14 Ga.

The age of c. 1.8 Ga is also typical for metamorphic titanites and monazites, which show that roughly contemporaneously with the formation of granites, vast areas were affected by strong thermal and fluid activity. Hydrothermal fluids have also affected the U-Pb systematics of zircon, which is clearly seen in gabbros from Kulkujärvi and Home-Jolhikko, for example.

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Appendix I. U-Pb mineral age data from igneous rocks in the Kittilä area

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A246-Nyssäkoski quartz-feldspar porphyry															
A. +4.6	13.2	565.8	193.9	4262	0.0278	0.3401	0.65	5.502	0.65	0.1174	0.15	0.97	1886	1900	1916
B. +4.6/HF	3.1	407.0	131.5	8265	0.0245	0.3276	0.65	5.306	0.70	0.1175	0.20	0.96	1826	1869	1918
C. 4.2-4.6/+100	15.1	639.7	215.6	3974	0.0280	0.3374	0.65	5.466	0.65	0.1175	0.15	0.97	1874	1895	1918
D. 4.2-4.6/100-200	15.6	680.0	228.1	4565	0.0305	0.3357	0.65	5.431	0.65	0.1173	0.15	0.97	1866	1889	1916
E. 4.0-4.2	15.6	793.6	259.3	3536	0.0348	0.3247	0.65	5.241	0.65	0.1171	0.15	0.97	1812	1859	1912
A280-Kapsajoki quartz-feldspar porphyry (lava)															
A. +4.6/100-200	20.3	210.6	80.7	23357	0.1327	0.3539	0.65	6.027	0.65	0.1235	0.15	0.97	1953	1979	2007
B. +4.6/100-200/HF	19.8	183.7	71.9	27772	0.1325	0.3614	0.65	6.175	0.65	0.1239	0.15	0.97	1988	2000	2013
C. 4.2-4.6	17.9	401.6	155.3	19649	0.1268	0.3585	0.65	6.111	0.65	0.1236	0.15	0.97	1975	1991	2009
A450-Riikonkoski metadiabase															
A. Total, borax	50.0	629.0	244.5	2937	0.3162	0.3097	0.65	5.316	0.69	0.1245	0.19	0.96	1793	1871	2021
B. +4.2/abr 1.5 h	2.1	448.2	199.7	3458	0.2917	0.3611	0.65	6.322	0.65	0.1270	0.15	0.97	1987	2021	2057
C. -3.8/abr 3 h	8.2	516.0	233.6	4839	0.3082	0.3632	0.65	6.341	0.65	0.1266	0.15	0.97	1997	2024	2051
D. 3.6-4.0/abr 3 h	3.4	1003.5	390.7	3524	0.3578	0.3015	0.65	5.159	0.65	0.1241	0.15	0.97	1698	1845	2015
E. 4.0-4.2/abr 2 h	6.3	679.5	300.3	3341	0.3423	0.3451	0.73	5.990	0.73	0.1259	0.15	0.97	1911	1974	2041
A452-Riikonkoski metadiabase															
A. Total, borax	42.0	603.0	218.3	2467	0.2535	0.3011	0.65	5.153	0.72	0.1241	0.24	0.94	1696	1844	2016
A469-Heinälammintievat metadiabase/ albitite															
A. +4.6	20.1	166.6	82.7	1211	0.6818	0.3102	0.65	5.363	1.80	0.1254	1.49	0.62	1741	1879	2034
B. 4.2-4.6	20.7	216.6	116.3	1260	0.7053	0.3310	0.65	5.789	0.75	0.1268	0.30	0.92	1843	1944	2054
C. +4.6/abr 5 h	14.4	152.9	88.4	3008	0.7085	0.3614	0.65	6.308	0.65	0.1266	0.15	0.97	1988	2019	2051
D. +4.6/abr 2 h	3.9	147.6	89.3	479	0.7024	0.3574	0.65	6.228	0.72	0.1264	0.24	0.94	1969	2008	2048
E. 4.2-4.6/abr 2 h	4.3	241.5	141.0	5117	0.7317	0.3625	0.65	6.334	0.65	0.1267	0.15	0.97	1993	2023	2053
F. +4.6/abr 6 h	20.4	155.6	88.2	1404	0.6953	0.3519	0.65	6.139	0.65	0.1265	0.15	0.97	1943	1995	2050
A470-Eksymäselkä metagabbro/ albitite															
A. 3.3-3.8	33.5	1052.5	182.0	1822	0.2695	0.1413	0.65	2.283	0.75	0.1172	0.31	0.91	851	1207	1914
B. +3.8	46.0	587.4	193.0	2080	0.2027	0.2793	0.65	5.120	0.69	0.1330	0.18	0.97	1587	1839	2137

Appendix I continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A470C. +4.6	23.1	455.8	171.6	773	0.1848	0.3114	0.65	5.795	0.69	0.1350	0.18	0.97	1747	1945	2163
D. 4.2-4.6	24.3	480.3	190.0	631	0.1822	0.3231	0.65	6.035	0.70	0.1355	0.19	0.96	1805	1980	2170
E. 4.2-4.6/abr 5 h	9.6	409.3	179.5	2917	0.1629	0.3853	0.70	7.339	0.71	0.1381	0.15	0.97	2101	2153	2204
A472-Palttina metagabbro															
A. Titanite	30.9	40.0	29.6	178	1.1761	0.3174	1.45	5.739	1.60	0.1311	0.56	0.94	1777	1937	2113
B. 3.8-4,3/clear/abr 3 h	2.5	1635.1	825.4	4754	0.5962	0.3367	0.65	6.060	0.65	0.1306	0.15	0.97	1870	1984	2105
C. 3.8-4.3/turbid/abr 2 h	1.7	1160.0	399.3	1434	0.5030	0.2394	0.65	4.058	0.69	0.1230	0.19	0.96	1383	1645	1999
D. 3.6-3.8/clear/abr 1 h	2.9	2198.6	990.5	3046	0.7160	0.2801	0.65	4.934	0.65	0.1278	0.15	0.97	1592	1808	2067
E. 3.6-3.8/turbid/abr 1 h	4.3	1796.6	600.8	1386	0.6595	0.2121	0.65	3.565	0.65	0.1219	0.15	0.97	1240	1541	1984
F. Titanite	26.3	75.7	73.3	98	1.1746	0.3619	0.65	6.568	0.90	0.1316	0.45	0.88	1991	2055	2119
A513-Latvajärvi felsic volcanic rock															
A. +4.1	21.5	509.5	172.4	3462	0.2464	0.2887	0.65	4.541	0.80	0.1141	0.38	0.88	1634	1738	1865
B. 3.8-4.1	15.6	743.2	222.0	1962	0.2066	0.2611	0.65	4.037	1.10	0.1121	0.89	0.59	1495	1641	1834
C. +4.1/abr 3 h	8.4	470.8	169.1	16531	0.1780	0.3221	0.65	5.085	0.65	0.1145	0.15	0.97	1800	1833	1872
A581-Veikaisenmaa quartz porphyry dyke															
A. 4.3-4.6	14.4	325.4	126.9	17974	0.1381	0.3582	0.65	6.110	0.65	0.1237	0.15	0.97	1973	1991	2010
B. 4.2-4.3	15.0	685.1	272.9	22777	0.1650	0.3586	0.65	6.114	0.65	0.1237	0.15	0.97	1975	1992	2010
C. 4.2-4.3/hf	15.0	724.6	292.4	20648	0.1679	0.3625	0.65	6.189	0.65	0.1238	0.15	0.97	1994	2001	2012
A655-Sätkänävaara metadiabase															
aA. +4.0/M	4.2	766.6	272.1	2819	0.6350	0.2318	0.70	3.812	0.71	0.1193	0.16	0.97	1343	1595	1945
aB. +4.0	5.0	619.4	288.6	5084	0.7244	0.2905	0.69	4.939	0.70	0.1233	0.15	0.97	1644	1808	2004
aC. 3.6-4.0	3.2	755.6	315.0	1586	0.7036	0.2593	1.35	4.316	1.35	0.1207	0.24	0.97	1468	1696	1967
bD. 4.0-4.2/+200	5.3	531.8	310.7	3870	0.8327	0.3432	0.85	5.921	0.86	0.1251	0.18	0.97	1901	1964	2031
bE. 4.2-4.3/+200	0.8	396.5	207.6	3103	0.7254	0.3280	1.68	5.563	1.80	0.1230	0.59	0.93	1828	1910	2000
bF. +4.3/+200	1.2	277.0	164.9	1700	0.7890	0.3561	1.04	6.125	1.04	0.1247	0.39	0.93	1963	1993	2025
cG. +4.2	2.4	424.7	212.1	4236	0.4900	0.3562	1.06	6.183	1.07	0.1259	0.18	0.97	1964	2002	2041
cH. 4.0-4.2	7.4	653.5	345.5	2982	0.6256	0.3460	0.65	5.983	0.72	0.1254	0.24	0.94	1915	1973	2035
A660-Honkavaara intermediate volcanic rock															
A. +4.6	7.3	178.4	92.6	3413	0.2728	0.4130	0.65	10.335	0.70	0.1815	0.20	0.96	2228	2465	2666
B. 4.2-4.6	20.8	363.2	185.5	6800	0.2147	0.4228	0.65	10.590	0.65	0.1817	0.15	0.97	2273	2487	2668
C. 4.0-4.2	17.8	647.8	302.5	9460	0.1855	0.3957	0.70	9.826	0.71	0.1801	0.15	0.97	2149	2418	2653

Appendix I continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A660D. 3.8-4.0	12.1	972.7	360.6	4587	0.1764	0.3163	1.38	7.544	1.38	0.1730	0.15	0.97	1771	2178	2586
E. 4.2-4.6/abr 3 h	11.5	296.7	175.5	19100	0.1654	0.5071	0.76	13.072	0.76	0.1870	0.15	0.97	2644	2684	2715
F. Titanite	24.6	55.7	48.0	704	1.0985	0.4285	0.65	8.201	0.77	0.1388	0.35	0.89	2298	2253	2212
G. Titanite	30.2	61.9	53.2	704	1.0544	0.4359	0.65	8.298	0.68	0.1381	0.18	0.96	2332	2264	2203
A745-Murtomaa felsic volcanic rock															
A. +4.6	22.9	396.5	139.7	13852	0.1732	0.3173	0.65	5.021	0.65	0.1148	0.15	0.97	1776	1822	1876
B. 4.3-4.6/grey/abr 3 h	10.5	468.6	168.0	22413	0.1864	0.3197	0.65	5.054	0.65	0.1147	0.15	0.97	1788	1828	1874
C. 4.3-4.6/red/abr 3 h	10.3	539.2	200.5	28171	0.2102	0.3258	0.65	5.153	0.65	0.1147	0.15	0.97	1817	1844	1875
D. Monazite	2.4	278.0	5181.5	6813	66.4626	0.3181	0.68	4.817	0.75	0.1098	0.33	0.90	1780	1787	1796
A823-Lehtovaara metadiabase															
A. 3.8-4.2	6.7	886.8	308.1	1017	0.5240	0.2347	0.74	4.276	0.76	0.1321	0.23	0.96	1359	1688	2126
A824-Laitavaara albitite															
A. 3.8-4.0	5.8	1750.3	246.7	3135	0.1412	0.1292	0.65	1.959	0.75	0.1100	0.40	0.85	783	1101	1799
B. 3.6-3.8	1.1	2294.8	189.3	755	0.1685	0.0727	0.65	0.981	1.40	0.0979	1.00	0.76	452	694	1585
C. 3.6-3.8	4.1	2062.4	184.0	3015	0.1797	0.0804	0.66	1.066	0.72	0.0961	0.32	0.90	498	736	1549
D. 4.0-4.2/abr 3 h	1.1	2764.2	248.0	2180	0.1180	0.0823	0.65	1.433	0.68	0.1262	0.18	0.96	510	902	2045
E. 4.2-4.6/abr 3 h	1.0	669.2	209.0	1187	0.1113	0.2782	0.65	5.655	0.77	0.1475	0.41	0.85	1582	1924	2316
A825-Sivakkavaara albite-carbonate rock															
A. +4.6	24.5	244.5	111.8	10390	0.3760	0.3498	0.77	6.636	0.79	0.1376	0.19	0.97	1933	2064	2197
B. 4.2-4.6	21.6	352.9	154.2	4574	0.3851	0.3309	0.89	6.231	0.89	0.1366	0.15	0.97	1842	2008	2184
C. 3.8-4.0	21.9	684.7	206.0	3971	0.3711	0.2301	0.65	4.196	0.72	0.1323	0.31	0.90	1335	1673	2128
D. 4.0-4.2	21.2	461.0	171.9	3836	0.3593	0.2867	0.65	5.350	0.65	0.1354	0.15	0.97	1624	1876	2168
A887-Yräjärvi quartz-feldspar porphyry															
A. +4.3	12.3	239.1	88.0	3514	0.1413	0.3340	0.79	5.649	0.82	0.1227	0.21	0.96	1857	1923	1995
B. +4.3/crushed/HF	15.5	213.6	81.7	11071	0.1373	0.3512	0.66	5.982	0.67	0.1236	0.15	0.97	1940	1973	2008
C. 4.2-4.3	10.9	589.7	211.9	2738	0.1647	0.3186	0.69	5.370	0.70	0.1222	0.21	0.95	1782	1880	1989

Appendix I continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A893-Kiimarova felsic porphyry															
A. +4.6	3.5	247.3	96.7	9034	0.1309	0.3606	0.65	6.155	0.67	0.1238	0.17	0.97	1985	1998	2011
B. 4.3-4.6	18.3	424.5	165.8	40941	0.1387	0.3593	0.65	6.135	0.68	0.1239	0.18	0.96	1978	1995	2012
C. 4.2-4.3	7.2	575.0	228.2	14578	0.1508	0.3608	0.77	6.162	0.77	0.1239	0.15	0.97	1985	1999	2013
D. 4.0-4.2	3.3	681.2	269.8	16178	0.1597	0.3580	0.65	6.105	0.66	0.1237	0.15	0.97	1972	1990	2010
A914-Honkavaara felsic volcanic rock															
A. +4.6	12.7	383.2	130.7	3011	0.1609	0.2934	0.65	7.173	0.65	0.1774	0.15	0.97	1658	2133	2628
B. 4.2-4.6	21.8	548.2	171.0	3000	0.1476	0.2701	0.65	6.511	0.65	0.1749	0.15	0.97	1541	2047	2604
C. 4.0-4.2	15.4	914.5	247.3	5546	0.1383	0.2379	0.65	5.679	0.65	0.1732	0.15	0.97	1375	1928	2588
D. 3.8-4.0	9.8	1250.9	300.4	2996	0.1208	0.2136	0.65	4.928	0.70	0.1674	0.21	0.95	1247	1807	2531
E. 3.8-4.0	10.6	1126.9	283.0	7108	0.1222	0.2247	0.82	5.252	0.82	0.1695	0.15	0.97	1306	1861	2552
F. 4.2-4.6/abr 4 h	12.1	299.5	160.9	15095	0.1494	0.4656	0.65	12.094	0.65	0.1884	0.15	0.97	2464	2611	2728
G. 4.2-4.6/abr 5 h	7.8	300.8	162.3	12728	0.1497	0.4672	0.65	12.147	0.65	0.1886	0.15	0.97	2471	2615	2729
A915-Honkavaara metadiabase															
A. Titanite	31.2	73.8	54.9	1929	0.9270	0.4118	1.03	7.844	1.03	0.1382	0.18	0.97	2223	2213	2204
B. 4.2-4.6	22.3	436.0	171.2	3160	0.3769	0.2978	0.66	5.465	0.66	0.1331	0.19	0.96	1680	1895	2139
C. 4.0-4.2	13.9	714.8	270.8	4327	0.4933	0.2684	0.68	4.781	0.68	0.1292	0.19	0.96	1532	1781	2087
D. +4.6	9.9	386.9	150.0	1455	0.3642	0.2919	0.65	5.351	0.65	0.1330	0.19	0.95	1650	1877	2137
E. 3.8-4.0	15.1	1409.0	347.8	3841	0.5985	0.1652	0.65	2.691	0.65	0.1182	0.15	0.97	985	1325	1928
F. 4.2-4.6/abr 3 h	10.3	387.8	180.8	6583	0.4209	0.3456	0.66	6.457	0.67	0.1355	0.15	0.97	1913	2040	2170
A916-Isolaki metadiabase															
A. 3.8-4.0	13.8	1120.0	455.3	2498	0.4406	0.2946	0.65	5.510	0.65	0.1356	0.15	0.97	1664	1902	2172
B. 3.6-3.8	11.2	1388.6	486.0	1169	0.5079	0.2394	0.65	4.401	0.65	0.1334	0.15	0.97	1383	1712	2142
C. Titanite	30.5	75.7	31.9	220	0.4155	0.2634	1.00	4.820	1.40	0.1327	1.00	0.70	1507	1788	2134
D. +4.3	0.6	459.0	230.6	1209	0.3281	0.3873	1.60	7.351	1.70	0.1377	0.49	0.96	2110	2155	2198
E. 4.2-4.3	1.0	475.3	222.3	2485	0.3276	0.3672	1.00	6.986	1.00	0.1380	0.30	0.96	2016	2109	2202
F. 4.0-4.2	7.8	758.3	327.9	1985	0.3325	0.3343	0.65	6.314	0.85	0.1370	0.43	0.87	1858	2020	2190
G. 3.8-4.0	8.2	1199.4	444.6	3061	0.3463	0.3122	0.71	5.869	0.72	0.1363	0.15	0.97	1751	1956	2181
A944-Home-Jolhikko gabbro diabase															
bA. 4.0-4.2/abr 3 h	8.0	1175.4	556.9	2891	0.5513	0.3245	0.65	5.388	0.65	0.1204	0.15	0.97	1811	1882	1962
bB. 4.0-4.2/abr 5 h	7.3	1186.0	564.4	3289	0.5568	0.3253	0.65	5.418	0.65	0.1208	0.15	0.97	1815	1887	1968
bC. 4.0-4.2/abr 8 h	5.1	1156.1	558.0	4528	0.5467	0.3330	0.65	5.564	0.65	0.1212	0.15	0.97	1852	1910	1974
bD. +4.3	9.8	388.7	156.7	1930	0.2755	0.3284	1.04	5.267	1.06	0.1163	0.37	0.94	1830	1863	1900

Appendix I continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A944bE. 4.2-4.3	13.8	684.1	320.1	3392	0.4552	0.3401	0.88	5.723	0.88	0.1220	0.15	0.97	1887	1934	1986
bF. 4.0-4.2/abr 10 h	6.0	1169.5	558.1	2764	0.5433	0.3280	0.65	5.469	0.65	0.1209	0.16	0.97	1828	1895	1970
bG. 3.8-4.0/abr 8 h	9.7	1325.6	558.9	1810	0.5453	0.2879	0.72	4.664	0.74	0.1175	0.17	0.97	1630	1760	1918
bH. +4.2/abr 4 h	5.7	595.3	260.9	3771	0.3844	0.3347	0.70	5.525	0.70	0.1197	0.15	0.97	1861	1904	1952
bI. 4.0-4.2/abr 3 h	8.2	979.8	456.5	1570	0.5375	0.3181	0.65	5.279	0.65	0.1204	0.15	0.97	1780	1865	1962
aA. +4.6	2.8	309.6	111.5	1218	0.1750	0.3170	1.10	4.784	1.30	0.1095	0.70	0.84	1775	1782	1790
aB. 4.2-4.6/+200	7.6	525.2	225.2	2988	0.3500	0.3334	1.10	5.474	1.20	0.1191	0.50	0.91	1854	1896	1942
aC. 4.0-4.2	21.8	887.6	409.0	2533	0.4940	0.3259	0.65	5.425	0.70	0.1207	0.20	0.95	1818	1888	1967
A946-Home-Jolhikko gabbro pegmatoid															
A. +4.6	26.8	144.6	56.3	6199	0.1491	0.3544	0.65	6.033	0.65	0.1235	0.15	0.97	1955	1980	2007
B. 4.2-4.6	26.6	366.5	140.0	5299	0.1607	0.3429	0.65	5.803	0.65	0.1228	0.15	0.97	1900	1946	1996
C. +4.6/abr 5 h	9.4	133.0	49,8	15076	0.1500	0.3416	0.65	5.813	0.65	0.1234	0.15	0.97	1894	1948	2006
D. 4.2-4.6/abr 5 h	5.0	381.8	152.4	10462	0.1613	0.3602	0.88	6.105	0.88	0.1230	0.15	0.98	1982	1991	1999
E. 4.0-4.2/clear/abr 3 h	8.0	208.4	84.8	1778	0.1761	0.3547	0.65	6.014	0.65	0.1230	0.16	0.97	1956	1977	2000
F. 4.0-4.2/dark/abr 3 h	9.2	827.1	314.4	2342	0.1650	0.3360	0.65	5.667	0.65	0.1223	0.15	0.97	1867	1926	1990
A968-Honkavaara felsic volcanic rock															
A. +4.6	7.0	416.6	146.4	3184	0.1582	0.3038	0.72	7.215	0.80	0.1723	0.34	0.91	1710	2138	2579
B. 4.2-4.6	21.4	608.5	205.2	4113	0.1498	0.2935	1.41	6.952	1.41	0.1718	0.27	0.97	1658	2105	2575
D. 4.0-4.2/abr 3 h	10.1	660.9	288.1	12224	0.1353	0.3838	0.65	9.531	0.65	0.1801	0.15	0.97	2094	2390	2654
A973-Venejoki felsic volcanic rock (?)															
A. 3.8-4.0/+200/abr 3 h	6.5	912.5	443.5	19017	0.1154	0.4358	0.65	10.521	0.65	0.1751	0.15	0.97	2331	2481	2607
B. 4.2-4.6	20.4	611.2	310.3	12763	0.1279	0.4497	0.65	11.185	0.65	0.1804	0.15	0.97	2393	2538	2656
C. 4.0-4.2	15.2	856.4	416.3	10362	0.1223	0.4332	0.65	10.568	0.65	0.1770	0.15	0.97	2319	2485	2624
D. Titanite	30.5	71.6	37.0	4425	0.6830	0.3308	1.64	5.182	1.69	0.1136	0.45	0.95	1842	1849	1858
E. +4.6	20.5	314.7	162.2	10333	0.1426	0.4506	0.65	11.373	0.65	0.1831	0.15	0.97	2397	2554	2680
F. 3.8-4.2/+200/abr 3 h	5.4	931.3	448.7	17778	0.1747	0.4324	0.65	10.414	0.65	0.1747	0.15	0.97	2316	2472	2603
G. 4.2-4.6/M	10.7	624.6	315.7	18863	0.1401	0.4442	0.66	10.987	0.66	0.1794	0.15	0.97	2369	2522	2647
A1272-Pittarova gabbro diabase															
A. 4.0-4.2/abr 5 h	5.8	1062.0	382.8	2808	0.3652	0.2774	0.65	4.590	0.65	0.1200	0.15	0.97	1578	1747	1956
B. 4.2-4.3/abr 3 h	5.2	616.6	233.6	1958	0.2918	0.3042	1.00	5.074	1.01	0.1210	0.15	0.98	1712	1831	1970
C. 4.2-4.3/abr 5 h	6.5	684.6	260.3	2931	0.2929	0.3074	0.65	5.134	0.65	0.1211	0.15	0.97	1727	1841	1973

Appendix I continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A1273-Kulkujärvi gabbro															
A. +4.5/+200	10.7	65.8	24.8	4106	0.1177	0.3503	0.65	5.811	0.65	0.1203	0.15	0.97	1936	1948	1960
B. +4.5/+200/abr 3 h	10.4	61.9	23.4	4857	0.1141	0.3532	0.65	5.865	0.65	0.1204	0.15	0.97	1949	1956	1963
C. 4.3-4.5/+200	7.2	179.6	64.6	2777	0.1273	0.3304	0.65	5.283	0.65	0.1160	0.15	0.97	1840	1866	1894
D. +4.5/+200/abr 3 h	11.1	66.5	24.9	7000	0.1187	0.3500	0.65	5.766	0.65	0.1195	0.15	0.97	1934	1941	1948
E. +4.5/-200/abr 5 h	7.0	103.5	36.5	6137	0.1086	0.3335	0.65	5.246	0.65	0.1141	0.15	0.97	1855	1860	1865
F. 4.3-4.5/-200	7.1	256.5	87.6	19506	0.1134	0.3234	0.65	4.994	0.65	0.1120	0.15	0.97	1806	1818	1832
G. +4.5/+200/abr 6 h	9.8	58.8	22.3	6436	0.1181	0.3535	0.65	5.879	0.65	0.1206	0.15	0.97	1951	1958	1965
H. +4.5/+200/abr 10 h	12.2	51.8	20.0	13254	0.1225	0.3609	0.65	6.072	0.65	0.1220	0.15	0.97	1986	1986	1986
I. 4.3-4.5/-200/abr 5 h	8.1	227.5	80.8	12888	0.1225	0.3327	0.65	5.232	0.65	0.1141	0.15	0.97	1851	1857	1865
A1395-Latvajärvi metadiabase															
A. +4.2/abr 0.5 h	0.2	503.0	189.8	6687	0.2294	0.3236	0.50	5.396	0.50	0.1210	0.10	0.98	1807	1884	1971
B. 3.6-4.2/clear/prismatic	0.2	1023.3	358.0	3487	0.2759	0.2892	0.50	4.768	0.50	0.1196	0.10	0.98	1637	1779	1950
C. 3.6-4.2/turbid/anhedral	0.2	1218.6	307.6	4662	0.1288	0.2354	0.50	3.595	0.50	0.1108	0.10	0.98	1363	1548	1812
D. 3.6-4.2/prismatic/pigment	<0.1			156	0.2750	0.2898	0.60	4.726	0.87	0.1183	0.55	0.78	1640	1772	1930
E. Titanite	20.4			249						0.1095					1791
A1441b-Palovaara lamprophyre dyke															
C.Monazite	0.24	129.0	399.0	4228	10.3000	0.3120	0.65	4.659	0.65	0.1083	0.15	0.97	1750	1760	1771

*) Isotopic ratios corrected for fractionation, blank and age related common lead (Stacey & Kramers 1975).
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = standard error in the mean

Appendix II. U-Pb mineral age data from (meta)sedimentary rocks in the Kittilä area.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$							Rho**	APPARENT AGES / Ma		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A363-Outakero quartzite (KUG)															
A. 4.2-4.3	15.5	632.5	214.7	8170	0.1117	0.3171	0.65	5.473	0.65	0.1252	0.15	0.97	1775	1896	2031
B. 4.2-4.3/-200	15.6	628.7	222.7	6951	0.1460	0.3218	0.65	5.516	0.65	0.1243	0.15	0.97	1798	1903	2019
C. +4.5/round/abr 3 h	6.4	183.4	81.0	5416	0.1501	0.3934	0.65	7.887	0.65	0.1453	0.15	0.97	2139	2218	2291
D. Monazite	2.8	716.5	3147.7	8528	14.6290	0.3218	0.65	4.874	0.70	0.1098	0.28	0.92	1798	1797	1797
A495-Vasalampi quartzite (SOG)															
A. Monazite	3.2	266.3	591.8	3603	6.8187	0.3232	0.65	4.951	0.70	0.1111	0.25	0.93	1805	1811	1817
B. +4.5/abr 1 h	11.7	153.5	91.6	10598	0.1670	0.5081	0.81	13.550	0.81	0.1934	0.15	0.97	2648	2718	2771
A635a-Kellostapuli conglomerate, reddish granite clast (LAG)															
A. +4.6	11.9	561.9	143.9	2203	0.2458	0.2142	0.65	3.459	0.65	0.1172	0.15	0.97	1250	1517	1913
B. +4.2	8.3	825.8	188.0	1998	0.2227	0.1938	0.65	3.032	0.65	0.1135	0.15	0.97	1141	1415	1855
C. 3.8-4.3	11.8	1523.4	232.3	2141	0.1966	0.1331	0.65	1.874	0.65	0.1021	0.15	0.97	805	1071	1663
A635b-Kellostapuli conglomerate, darkish granite clast (LAG)															
A. +4.2/abr 3 h	10.0	636.1	212.7	22884	0.1625	0.3039	0.65	4.792	0.65	0.1144	0.15	0.97	1710	1783	1870
B. +4.2	9.1	566.1	173.4	5523	0.1609	0.2775	0.65	4.368	0.69	0.1142	0.20	0.95	1578	1706	1867
C. 4.0-4.2/+200	9.2	1101.1	299.7	4936	0.1642	0.2457	0.65	3.805	0.65	0.1123	0.15	0.97	1416	1593	1837
D. 4.0-4.2/+200/abr 3 h	7.0	1068.3	325.3	17330	0.1607	0.2772	0.65	4.311	0.65	0.1128	0.15	0.97	1577	1695	1845
E. 4.0-4.2/+200/HF	10.7	1041.3	316.2	22746	0.1609	0.2768	0.65	4.315	0.65	0.1131	0.15	0.97	1575	1696	1849
F. 3.8-4.0/+200	7.9	1452.1	334.4	8530	0.1508	0.2115	0.65	3.195	0.65	0.1096	0.15	0.97	1236	1456	1792
A802-Ylläs quartzite (LAG)															
A. 4.3-4.6/+100	4.7	321.9	93.7	2056	0.1810	0.2517	1.03	4.691	1.05	0.1351	0.18	0.97	1447	1765	2166
B. 4.3-4.5/100-200	20.5	338.3	92.5	1934	0.1612	0.2380	0.78	4.607	0.78	0.1404	0.15	0.97	1376	1750	2232
C. 4.2-4.3/+200	15.4	571.6	97.9	1407	0.1326	0.1521	0.73	2.721	0.82	0.1298	0.44	0.86	912	1334	2095
D. +4.5/+100	10.3	152.8	69.3	5088	0.1754	0.3931	0.65	8.356	0.65	0.1542	0.15	0.97	2137	2270	2392
A886-Sätkänävaara quartzite (KUG)															
A. +4.6	5.1	439.2	153.8	3802	0.1678	0.3119	0.86	5.347	0.87	0.1243	0.15	0.98	1749	1876	2019
B. 4.3-4.6/rounded	13.3	257.5	91.1	7771	0.1617	0.3188	0.76	5.389	0.76	0.1226	0.15	0.98	1783	1883	1994

Appendix II continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$							Rho**	APPARENT AGES / Ma		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
C. +4.3/long	6.3	458.3	151.0	6431	0.0844	0.3160	0.85	5.222	0.86	0.1199	0.18	0.97	1769	1856	1954
Monazite	4.2	3250.2	5314.2	39468	4.3539	0.3452	0.65	5.550	0.67	0.1166	0.15	0.97	1911	1908	1905
A925-Värttiövaara quartzite (KUG)															
A. 4.3-4.5/-200	18.4	400.7	128.5	3397	0.1296	0.2925	0.65	5.069	0.65	0.1257	0.15	0.97	1654	1830	2038
B. Monazite	2.3	2070.7	3610.7	18426	4.8821	0.3356	0.65	5.356	0.70	0.1157	0.21	0.95	1865	1877	1891
A939a-Linkupalo conglomerate, felsic porphyry clast (LAG)															
A. +4.2	2.1	634.8	222.5	11834	0.0530	0.3448	0.65	5.987	0.71	0.1260	0.23	0.95	1909	1974	2041
B. 4.0-4.2	4.9	938.2	295.6	17554	0.0432	0.3145	0.65	5.155	0.65	0.1189	0.15	0.97	1762	1845	1939
A939b-Linkupalo conglomerate, felsic porphyry clast (LAG)															
A. 3.6-4.0/+200/abr 3 h/transp	5.0	1938.2	382.9	4617	0.1472	0.1814	0.65	2.629	0.70	0.1051	0.22	0.95	1074	1308	1716
B. -3.6/+200/abr 3 h/transp	4.5	2180.6	288.7	2274	0.1552	0.1205	0.65	1.568	0.65	0.0944	0.15	0.97	733	957	1516
C. 3.6-4.0/-200/abr 3 h	1.8	2121.2	465.9	5128	0.1497	0.2014	0.65	2.975	0.65	0.1071	0.15	0.97	1182	1401	1751
D. -3.6/abr 6 h/transp	4.0	2212.4	302.5	2929	0.1584	0.1246	0.65	1.639	0.65	0.0954	0.15	0.97	756	985	1535
E. +4.0/abr 2 h	1.5	800.6	199.0	5796	0.1291	0.2295	0.65	3.954	0.65	0.1250	0.15	0.97	1331	1624	2028
F. 3.8-4.0/abr 2 h	0.8	1768.2	434.8	3178	0.1577	0.2228	0.65	3.364	0.69	0.1095	0.19	0.96	1296	1496	1791
G. 3.6-3.8	1.6	2215.5	470.2	1364	0.2248	0.1792	0.65	2.570	0.69	0.1040	0.19	0.96	1062	1292	1696
A942-Värttiövaara quartzite (KUG)															
A. 4.3-4.6/-200	16.8	352.4	112.9	4897	0.0998	0.3006	0.71	5.252	0.80	0.1267	0.38	0.88	1694	1861	2053
B. 4.3-4.6/turbid	16.2	267.8	106.7	4962	0.1513	0.3553	0.83	5.991	0.84	0.1223	0.19	0.97	1959	1974	1990
C. 4.3-4.6/+100/clear	0.9	242.9	84.8	3504	0.1405	0.3228	1.34	5.185	1.50	0.1165	1.00	0.76	1803	1850	1903
A961-Virttiövaara quartzite (SOG)															
A. +4.6/+100	28.8	170.0	84.7	902	0.1834	0.4006	0.65	10.777	0.65	0.1951	0.15	0.97	2171	2504	2786
B. 4.2-4.6	20.9	424.5	178.7	890	0.1418	0.3485	0.65	9.235	0.65	0.1922	0.15	0.97	1927	2361	2760
C. 4.0-4.2	20.1	791.5	262.6	916	0.1222	0.2801	0.65	7.187	0.65	0.1861	0.15	0.97	1591	2134	2708
D. +4.6/dark,turbid/abr 3 h	1.8	361.5	193.3	2571	0.1288	0.4631	0.75	12.371	0.76	0.1937	0.18	0.97	2453	2632	2774
E. +4.6/dark purple/abr 3 h	3.9	213.1	120.4	3457	0.1551	0.4812	0.70	12.927	0.71	0.1949	0.15	0.97	1532	2674	2783
F. +4.6/colorless/abr 3 h	3.0	74.4	48.9	2745	0.1885	0.5482	0.86	14.921	0.87	0.1974	0.22	0.97	2817	2810	2805
G. +4.6/pale purple/abr 3 h	10.1	90.3	59.9	2654	0.1904	0.5473	0.68	14.836	0.69	0.1966	0.15	0.97	2814	2804	2798

Appendix II continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$							Rho**	APPARENT AGES / Ma		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A970-Keimiöniemi schist (SKG?)															
A. +4.6	20.0	479.1	177.4	3050	0.1124	0.3432	0.65	6.254	0.65	0.1322	0.15	0.97	1901	2012	2127
B. 4.2-4.6	16.7	304.1	117.7	3177	0.1151	0.3569	0.65	6.746	0.65	0.1371	0.15	0.97	1967	2078	2190
C. +4.6/HF	19.7	382.1	161.8	2559	0.1166	0.3894	0.65	7.235	0.65	0.1348	0.15	0.97	2120	2140	2161
D. +4.6	20.1	469.7	174.8	4597	0.1067	0.3472	0.65	6.343	0.65	0.1325	0.15	0.97	1921	2024	2131
E. 4.2-4.6	20.4	303.4	118.3	7082	0.1095	0.3626	0.65	6.832	0.65	0.1267	0.15	0.97	1994	2089	2185
F. +4.6/abr 3 h	4.8	491.1	190.9	6924	0.1073	0.3626	0.65	6.746	0.65	0.1349	0.15	0.97	1994	2078	2163
G. 4.2-4.6/abr 3 h	4.2	285.1	113.3	11510	0.1077	0.3695	0.65	6.984	0.65	0.1371	0.15	0.97	2026	2109	2191
A971-Keimiöniemi schist (SKG?)															
A. +4.2/abr 3 h	3.6	536.1	185.3	13355	0.1251	0.3214	0.65	5.362	0.65	0.1210	0.15	0.97	1796	1878	1971
B. Titanite/translucent	11.5	111.4	50.0	306	0.2960	0.3186	0.65	4.811	0.80	0.1095	0.40	0.87	1782	1786	1792
C. Titanite/turbid, dark	15.1	81.0	33.6	558	0.2650	0.3227	0.65	4.963	0.70	0.1116	0.20	0.95	1802	1813	1825
D. Titanite/translucent a1h	10.1			711	0.2110					0.1096	0.24				1785
A1176-Hyyverova sillimanite quartzite (SOG)															
A. +4.5	10.9	309.8	111.2	2167	0.0976	0.3304	0.70	6.172	0.71	0.1355	0.15	0.97	1840	2000	2170
B. 4.3-4.5	8.3	563.0	188.2	2968	0.0824	0.3155	0.65	5.543	0.65	0.1274	0.15	0.97	1767	1907	2062
C. 4.2-4.3/+200	7.0	881.6	271.7	4005	0.0674	0.2971	0.65	5.025	0.65	0.1227	0.15	0.97	1676	1823	1995
D. 4.0-4.2/+200	7.5	1291.2	125.8	3779	0.0663	0.0941	0.73	1.560	0.73	0.1202	0.15	0.97	580	954	1959
A1416-Vesikkovaara conglomerate, felsic porphyry clast (LAG)															
A. 4.0-4.2	7.0	1307.4	421.1	7892	0.1938	0.2855	0.65	4.439	0.70	0.1128	0.20	0.95	1619	1719	1844
B. 4.0-4.2/abr 6 h	5.1	1299.8	457.8	14604	0.1997	0.3110	0.65	4.886	0.65	0.1140	0.15	0.97	1745	1799	1863
C. +4.2/abr 3 h	0.4	1313.9	402.6	3825	0.2100	0.2654	0.50	4.119	0.50	0.1126	0.12	0.97	1517	1658	1841
D. +4.2	0.5	912.0	278.7	8021	0.1900	0.2712	0.50	4.220	0.50	0.1129	0.12	0.97	1547	1678	1846
E. 4.0-4.2/abr 18 h	0.5	1236.5	435.2	10731	0.2030	0.3091	0.50	4.855	0.50	0.1139	0.12	0.97	1736	1795	1863
A1470-Mantovaara conglomerate, felsic porphyry clast (KUG)															
A. +4.2	5.5	655.0	210.9	16834	0.0892	0.3097	0.65	4.968	0.65	0.1163	0.15	0.97	1739	1813	1900
B. +4.2/abr 3 h	4.2	656.1	218.2	27419	0.0928	0.3192	0.65	5.137	0.65	0.1167	0.15	0.97	1785	1842	1906
C. 4.0-4.2	5.7	989.5	292.9	9259	0.1082	0.2799	0.65	4.418	0.65	0.1145	0.15	0.97	1590	1715	1872

*) Isotopic ratios corrected for fractionation, blank and age related common lead.
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = Staqndard error in the mean

Appendix III. U-Pb mineral data from granitoid rocks in the Kittilä area.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$							Rho**	APPARENT AGES / Ma		
Analysed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A184-Kotivaara granite (Tepasto)															
A. 4.4-4.6	20.4	211.8	74.6	3348	0.2373	0.2998	0.65	4.518	0.70	0.1093	0.20	0.96	1690	1734	1788
B. 4.4-4.6/HF/uncrushed	20.3	205.2	74.6	5169	0.2372	0.3109	0.65	4.692	0.65	0.1095	0.17	0.97	1745	1765	1790
C. 4.3-4.4	16.3	525.4	158.5	2073	0.2018	0.2612	0.65	3.890	0.75	0.1080	0.30	0.92	1496	1611	1766
D. 4.3-4.3/HF/uncrushed	15.1	485.0	153.4	5226	0.1950	0.2793	0.65	4.174	0.72	0.1084	0.24	0.94	1588	1668	1772
E. 4.2-4.3	10.4	945.7	250.0	1539	0.1849	0.2303	0.65	3.339	0.72	0.1051	0.23	0.95	1336	1490	1716
F. 4.2-4.3/HF/uncrushed	10.1	823.4	242.5	3648	0.1869	0.2607	0.65	3.871	0.70	0.1077	0.20	0.96	1493	1607	1760
G. Titanite	30.7	144.0	96.0	1280	1.2985	0.3134	1.35	4.802	1.50	0.1111	0.69	0.89	1757	1785	1818
H. Titanite/HF	28.8	143.6	93.4	928	1.2732	0.3068	0.85	4.651	1.00	0.1100	0.69	0.73	1724	1758	1800
A319-Hanhivaara granite (Tepasto)															
A. +4.2	5.0	504.2	126.8	1661	0.1886	0.2193	0.65	3.201	0.65	0.1059	0.19	0.96	1278	1457	1730
B. 4.0-4.2/+100	2.3	1308.3	320.0	1280	0.2420	0.2036	0.65	2.893	0.70	0.1030	0.20	0.95	1194	1380	1679
C. 4.0-4.2	8.7	1296.2	310.3	1065	0.2153	0.2013	0.68	2.860	0.95	0.1031	0.70	0.68	1182	1371	1680
A320-Isovaara granite (Kittilä)															
A. +4.2	2.1	488.2	148.2	4095	0.1412	0.2772	1.00	4.628	1.00	0.1211	0.17	0.99	1577	1754	1972
B. 3.3-4.2	2.6	1399.7	315.4	1529	0.1435	0.2022	0.65	3.066	0.69	0.1100	0.19	0.96	1186	1424	1799
A321-Hangasselkä granite (Kittilä)															
A. +4.3	10.7	442.5	154.2	4535	0.1992	0.3055	0.65	4.739	0.65	0.1125	0.15	0.97	1718	1774	1840
B. +4.2/HF	10.1	394.3	141.4	10857	0.1876	0.3179	0.65	5.171	0.65	0.1180	0.15	0.97	1779	1847	1925
C. 4.0-4.2	9.4	663.0	206.2	4126	0.1793	0.2768	0.80	4.256	0.85	0.1115	0.30	0.94	1575	1684	1824
A428- Lappalaislampi granite (Kittilä)															
A titanite	17.7	278.0	86.1	168	0.4800	0.2260	1.00	3.230	1.00	0.1036	0.30	0.96	1313	1464	1690
A748-Iso Nilipää granite (Huhma 1986)															
A. +4.6	25.2	539.0	217.0	3109	0.1760	0.3524	0.65	6.378	0.65	0.1313	0.15	0.97	1946	2029	2115
B. 4.2-4.6/+100	15.4	826.0	318.0	1867	0.1380	0.3432	0.65	6.192	0.65	0.1309	0.15	0.97	1901	2003	2109
C. 4.0-4.2	10.8	1123.0	380.0	1440	0.1320	0.3008	0.65	5.334	0.65	0.1286	0.15	0.97	1695	1874	2079
D. 3.6-3.8/+200	8.7	2490.0	475.0	891	0.1440	0.1658	0.65	2.706	0.65	0.1184	0.15	0.97	988	1330	1931
E. titanite	32.0	193.0	242.0	60.5	0.5700					0.1310	1.00				2110

Appendix III continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$							Rho**	APPARENT AGES / Ma		
Analysed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A749-Pikku Nilipää granite															
A.+4.2	6.2	1251.3	297.1	559	0.2080	0.1893	0.65	3.343	0.72	0.1281	0.30	0.90	1117	1491	2072
B. 4.0-4.2	10.6	1701.1	352.1	676	0.1387	0.1761	0.65	3.103	0.65	0.1278	0.15	0.97	1045	1433	2068
A1206-Ruoppapalo granodiorite															
A. 4.3-4.5	5.4	673.4	220.2	3258	0.0637	0.3170	0.65	5.073	0.65	0.1161	0.15	0.97	1774	1831	1896
B. 4.2-4.3	5.3	878.9	269.7	3541	0.0616	0.2986	0.65	4.741	0.65	0.1152	0.15	0.97	1684	1774	1882
C. 4.0-4.2	5.8	1247.7	331.7	1993	0.0629	0.2557	0.65	3.973	0.65	0.1127	0.15	0.97	1468	1628	1843

*) Isotopic ratios corrected for fractionation, blank and age related common lead (Stacey & Kramers 1975).
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = standard error in the mean

Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
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THE U-Pb AGE OF A FELSIC GNEISS IN THE KUUSAMO SCHIST AREA: REAPPRAISAL OF LOCAL LITHOSTRATIGRAPHY AND POSSIBLE REGIONAL CORRELATIONS

by
Jorma Räsänen and Matti Vaasjoki

Räsänen, J. & Vaasjoki, M. 2001. The U-Pb age of a felsic gneiss in the Kuusamo schist area: reappraisal of local lithostratigraphy and possible regional correlations. *Geological Survey of Finland, Special Paper 33,* 143-152. 4 figures and one table.

The Kuusamo schist area in the SE part of the Paleoproterozoic Central Lapland Greenstone Belt is a geological key area in northern Finland. The lithostratigraphy observed in eastern Kuusamo in 1970s was later applied also to the western part of the area and to the Central Lapland Greenstone Belt. New geologic data collected from the greenstone belt gives more information of its lithostratigraphy and sheds new light on its evolution.

This paper reviews briefly the geology of the Kuusamo Schist area and some new aspects of the lithology are discussed in the light of current investigations. Based on U-Pb data obtained from zircons of a felsic volcanic rock (A906-Niskavaara) with an age of 2800±8 Ma (MSWD=4.9), the stratigraphic scheme applied previously to the western part of the Kuusamo Schist area is deemed questionable. It appears that the earlier classification derived from Proterozoic supracrustal rocks in eastern Kuusamo does not apply to the entire Kuusamo schist area.

The concept of the presence of remnants of a late Archean schist belt extending from Sodankylä to Puolanka is supported by field evidence and isotopic data.

Key words (GeoRef Thesaurus, AGI): absolute age, U/Pb, zircon, metavolcanic rocks, lithostratigraphy, Central Lapland Greenstone Belt, Proterozoic, Archean, Kuusamo, Posio, Finland.

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INTRODUCTION

The Kuusamo schist area forms the SE corner of the Paleoproterozoic Central Lapland Greenstone Belt (Fig. 1). This triangular, mainly NE-SW trending belt is one of the most intensely investigated regions in northern Finland and has been one of the geological key areas. About two thirds of the bedrock consists of metasediments while the remainder are mainly mafic volcanic rocks.

The Kuusamo schist area was first investigated by Hackman and Wilkman (1929), who divided the supracrustal rocks into the older Kalevian and younger Jatulian cycles. This classification, adopted from south-

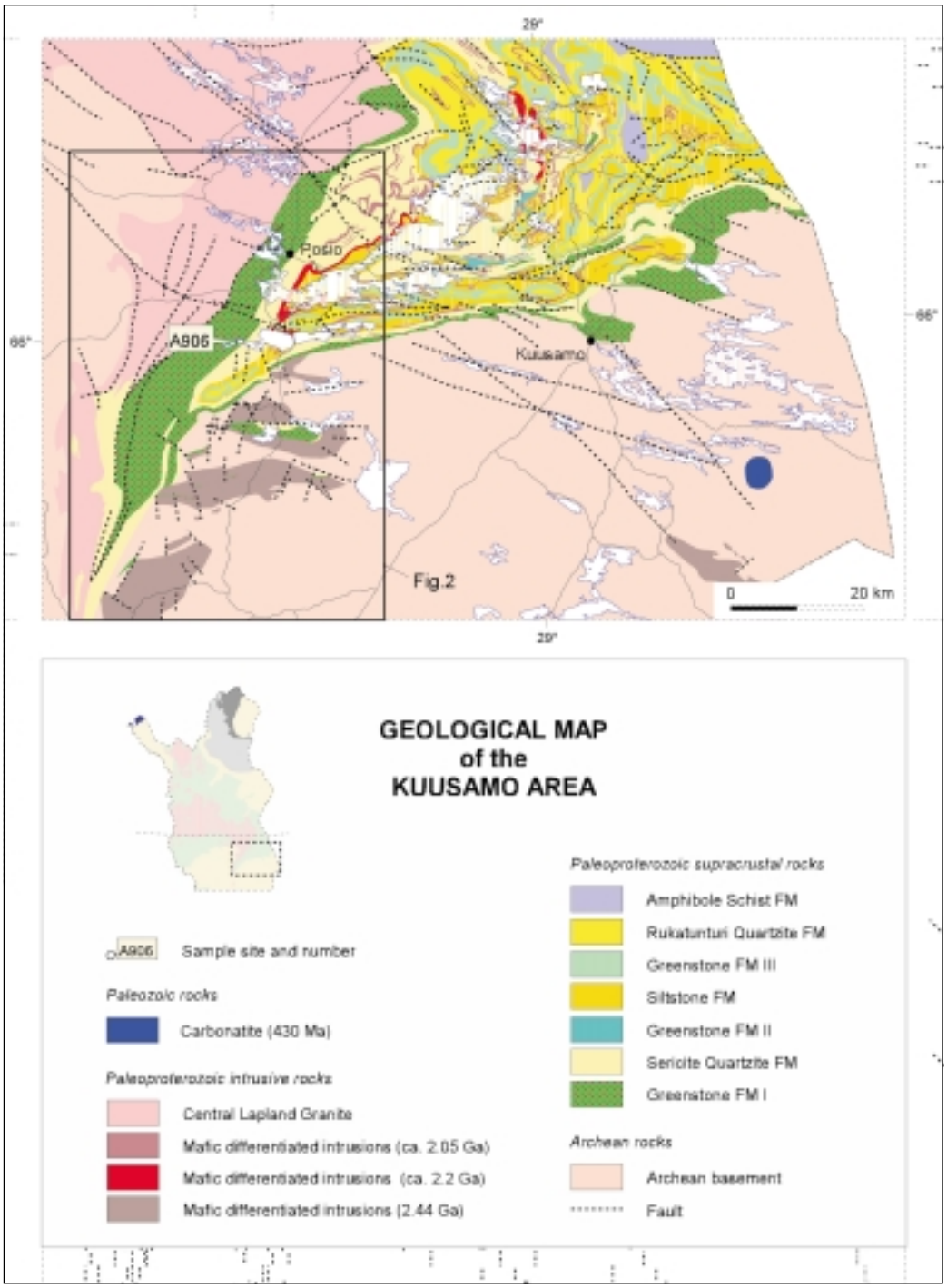

Fig. 1. Location and general geology of the Kuusamo schist area.

eastern Finland, was used by Hackman (1918) in spite of Mäkinen´s (1916) opinion that the age relations were exactly the opposite. Since the 1950s the area has been studied extensively during prospecting activities and many subareas have been described in university theses. The western part of the schist area was investigated by Enkovaara et al. (1953) and Honkamo (1979), while the eastern part was studied by Piispanen (1972), Silvennoinen (1972, 1989, 1991) and Pekkala (1985). Silvennoinen (1972) observed that the lithostratigraphy does not correspond to that found in eastern Finland and could not be classified using the traditional terms. The stratigraphic scheme used by Silvennoinen (op. cit.) was later applied over the whole Central Lapland Greenstone Belt, interpreted then as containing both Archean and Proterozoic supracrustal formations (Silvennoinen et al. 1980). Afterwards all the supracrustal rocks have been considered as Paleoproterozoic in origin (Silvennoinen 1992, Korsman et al. 1997).

The ongoing bedrock mapping of the Koillismaa area adjoining the Kuusamo schist area commenced in the late 1990s. The new data revealed that the bedrock in the western part of the Kuusamo schist

area differs considerably from the better known eastern part. The western part consists of almost NS-striking rocks comprising mica and quartz-feldspar schists, felsic metavolcanic rocks, amphibolites and ultramafic rocks, none of which are met within the eastern part of the Kuusamo schist area. Silvennoinen et al. (1980) have published a U-Pb zircon age of c. 2790 Ma for a metavolcanic rock from Niskavaara in the western part. As felsic volcanics also occur as clasts in a conglomerate lying low in the stratigraphy in eastern Kuusamo, these and the age in the west were correlated and it was concluded that Archean volcanic rocks formed the basal unit of the Kuusamo schist area. Later, the felsic clasts of the conglomerate in eastern Kuusamo were dated to be 2405±6 Ma old (Silvennoinen 1991), which demonstrated that the basal unit in the east was not Archean, but Proterozoic. However, the Archean age result from Niskavaara was overlooked and the entire Kuusamo Schist Belt (as well as the Central Lapland Schist Belt) was interpreted as Proterozoic (Silvennoinen 1992). In this paper, the Archean result is reconsidered and the probability of widespread Archean schists in the western part of the Kuusamo Schist Belt is demonstrated on the basis of geological and geophysical data.

GEOLOGICAL SETTING OF THE KUUSAMO SCHIST AREA

The area is bounded in the south by the Archean Kuhmo Complex and in the southwest by the 2.44 Ga old Koillismaa layered intrusions. In the west, it is delimited by the Archean Pudasjärvi Complex but comes also into contact with the Proterozoic Lapland Granitoid Complex. In the north, the contact with the Archean Suomujärvi Complex is granitized. In contrast, the contact with the Salla greenstones in the northeast is tectonic. In the east, the Kuusamo schists continue into Russia.

According to Silvennoinen (1972, 1991), the Paleoproterozoic supracrustal sequence in the eastern part of the schist area starts locally with a breccia-like conglomerate. The conglomerate lies on Archean granite gneisses and, in addition to these gneisses, contains also fragments of quartz porphyry which register a U-Pb age of 2405±6 Ma. The overlying Greenstone Formation I lies mostly directly on Archean granitoids and exhibits volcanic breccias at its base, but the bulk is composed of massive and amygdaloidal mafic flows. These are covered by tuffitic schists with some dolomitic beds overlain by a transgressive sequence of sedimentary rocks. The latter commence with the Sericite Quartzite Formation, which is mainly arkosic in composition and contains conglomeratic interlayers. Its uppermost part grades into the Sericite Schist Formation, which includes mainly fine-grained carbonate-bearing schists. Upwards it changes further into phyllites containing dolomite interlayers on their top. Some dolomites display stromatolitic structures forming up to 35 m thick deposits (Pekkala 1985). The overlying Quartzite Schist Formation, about 50 m thick, consists mainly of phyllitic, usually chlorite-bearing sericite schists. Silvennoinen (1992) noted that this volcano-sedimentary succession, called *the Lapponia Supergroup,* is comparable to *the Sumi-Sariola superhorizon* in Russian Karelia.

Pillowed mafic flows and breccias are characteristic of Greenstone Formation II, which overlies the transgressive sequence. The age of the metavolcanics is unknown but they are cut by a mafic dyke with a U-Pb age of 2206±9 Ma (Silvennoinen 1991). The next sedimentary sequence is regressive and starts with the Siltstone Formation, which is an inhomogeneous deposit characterized by fine-grained and commonly reddish, hematite-bearing graded rocks. It also contains arkose, orthoquartzite, phyllite and dolomite members, but their internal stratigraphy is not known in detail. The top of the formation, however, displays ripple-stratified beds with mud cracks indicative of shoreline sediments. The overlying Greenstone Formation III consists of homogeneous massive flows with some amygdaloidal parts. The extensive occurrence of the formation has been interpreted to indicate a plateau basalt. The extrusives are highly magnetic (Elo 1992, Ruotoistenmäki 1992) and form on low-altitude geophysical maps banded magnetic patterns, which can be traced as a folded stratigraphic key horizon. The overlying Rukatunturi Formation contains minor graded schists, dolomites and sericite quartzites at base, but the bulk is formed by orthoquartzites. An age of 2078±8 Ma has been reported for a mafic dyke penetrating the lowermost part of the Rukatunturi Formation. This quartzite-dominated rock pile with two volcanic formations is taken to represent the *Jatuli succession* in eastern Finland (Silvennoinen 1992). It is covered by the Dolomite Formation, considered partially organic in origin. The overlying Amphibole Schist Formation contains black schist, some dolomite interbeds and undifferentiated hornblendite sills and is regarded to be mainly volcanogenic. On the top of the formation there occur also mica schists, which represent normal clastic sediments. This volcano-sedimentary pile is

interpreted to indicate a deep-water environment correlating with the *Marine Jatuli* described by Väyrynen (1933) in eastern Finland. These six conformably overlying formations constitute *the Karelia Supergroup* of Silvennoinen (1992).

South of Kuusamo the basement granite gneiss is intruded by a tonalitic granite with a U-Pb age of c. 2.74 Ga (Lauerma 1982). In addition, the lowermost breccia-like conglomerates in eastern Kuusamo have been recently recognized as mainly felsic volcanic breccias, which contain fragments of underlying quartz porphyries (Räsänen 1999). Chemically they are rhyolitic in composition. Similar acidic extrusive rocks have been found over a large area and in places they resemble porphyritic granites containing cm-size euhedral K-feldspar phenocrysts. Current investigations have shown that south of Kuusamo there exist also faintly porphyritic A-type granites yielding a U-Pb age of 2442±3 Ma (Landén & Mänttäri, in prep.). This age coincides with the 2443±5 Ma obtained for layered mafic intrusions (Alapieti 1982), which indicates their coeval emplacement and implies an extensional crustal setting.

OUTLINE OF THE WESTERN PART OF THE KUUSAMO SCHIST AREA

The bedrock mapping at Posio in the western part of the Kuusamo schist area was started by GTK in the 1970s but was later discontinued and little has been done since the 1980s and the area is not yet known in detail. However, non-magnetic mafic volcanic rocks closely resembling those of the Greenstone Formation I in Kuusamo were observed at the time. Quartzites similar to those in Kuusamo were found at Posio. More similarities in lithology were observed in the southernmost end of the schist area where non-magnetic mafic volcanic rocks and quartzites exist (Honkamo 1979). Although abundant mica gneisses were also found, the observed similarities gave reason to extrapolate the lithostratigraphy used in eastern Kuusamo (Silvennoinen 1972) to the western part of the schist area and also to the Central Lapland Greenstone Belt (Silvennoinen et al. 1980). An Archean age of c. 2.79 Ga for a felsic volcanic rock at Posio was reported and, as the age was connected with the felsic fragments of the lowermost conglomerate in Kuusamo, rocks of the Lapponi(an) Supergroup were interpreted to be of Archean age (op. cit.). It should be noted, however, that the bedrock mapping had just started in Lapland, and only a few chemical analyses and no low-altitude geophysical maps were available. Today all these supracrustal rocks are interpreted to be Proterozoic.

Some years ago the Geological Survey of Finland and the University of Oulu started jointly to investigate the structure and lithology in the western part of the Kuusamo schist area (Airo & Laajoki 1996). Based on sedimentological studies, Laajoki (1997) demonstrated that the mafic metavolcanics at Posio (Karkuvaara Formation) underlie turbiditic schists (Ahola Formation), which in turn are overlain by quartzites (Kirintökangas Formation). In addition, seeking a larger context for the old and questionable lithostratigraphy, Laajoki (1997) compared these quartzites with some quartzites at Puolanka, about 100 km south of Posio,and expressed that those occurrences "*are considered to represent the same storm-dominated muddy shelf*" and, "*their fluvial and near-shore counterparts should be found in the eastern part of the Kuusamo belt*". Later Airo (1999), on the basis of a geophysical structural model over the western part of this schist area, argued that the mafic volcanics, called the Posio greenstone, dip eastwards under the Kuusamo schist area. The highly magnetic schists and gneisses on either sides of the greenstones were thought of as underlying on the western side while those east of them were considered overlying, both the gneisses and the greenstones being part of the same structural unit.

Recent low-altitude geophysical maps contain much information on the bedrock and its structures. As seen in Figure 2, a long, mostly NS-trending non-magnetic zone extends from Posio to Jukua. Magnetic bands of varying intensity frame this magnetically quiet zone, come in contact south of Jukua and fuse to each other further southwards. This might be an image of a master fold-structure. Similarly, a large fold-like structure can be supposed to exist south of Niskavaara where the southernmost edge of the Kuusamo schist area is framed by highly magnetic bands. A conspicuous feature on the eastern side of the non-magnetic zone is that it terminates abruptly against the long banded magnetic highs, and could indicate a tens of km's long tectonic contact. The outlines of lithology and main structural features based on current bedrock mapping of the western part of the Kuusamo schist area are shown in Figure 3.

The Posio greenstones comprise principally lava-born volcanic rocks showing regularly very low ($<10^2$ Si) susceptibility values. These rocks are well exposed and consist of several almost non-magnetic massive and amygdaloidal flows with minor tuffitic

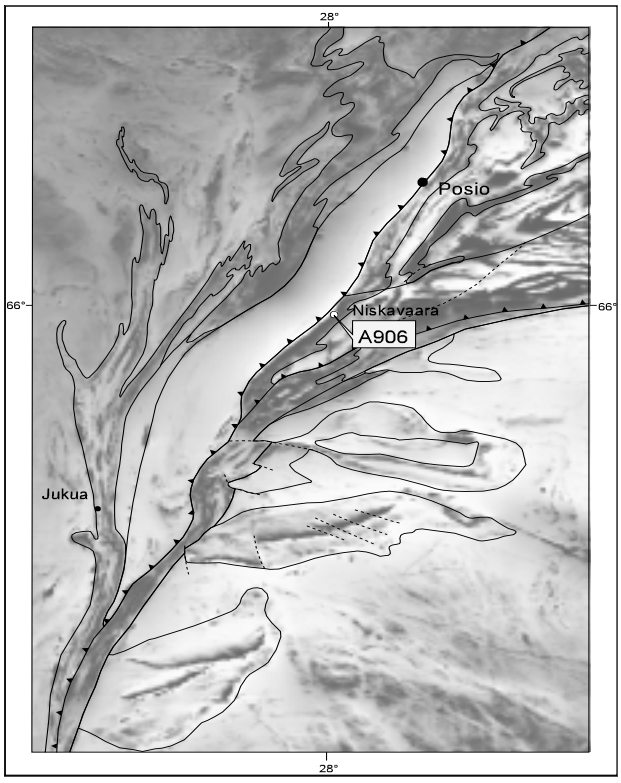

Fig. 2. Low-altitude aeromagnetic map of the western part of the Kuusamo schist area.

interbeds e.g. at Karkuvaara, north of Posio. They are basaltic andesites and andesites resembling chemically the metavolcanics of the Greenstone Formation I at Kuusamo. In the east they come in contact with highly magnetic mica gneisses. The contact is covered, but probably tectonic as the rocks on both sides are strongly sheared. Similar volcanic rocks occupy a large area west of Niskavaara and according to diamond drilling their eastern contact with the mica gneisses is tectonic. Also the almost non-magnetic zone east of Jukua consists of volcanic rocks comparable to the ones west of Niskavaara and those of the Greenstone Formation I at Kuusamo. This southernmost part of the Posio greenstones is intensely deformed, the rocks being strongly sheared and intruded by late granites. There also occur mafic and ultramafic intrusive rocks, as observed earlier by Honkamo (1979).

The Posio greenstones are framed by a rock succession of forming a belt with mica and quartz-feldspar gneisses, mafic amphibolites and felsic tuffitic rocks. The mica gneisses in particular are magnetite-bearing and show high (10^3-10^4 Si) susceptibility values and commonly produce banded magnetic anomalies. The gneisses are multiply deformed exhibiting early isoclinal folds with transposed and redeformed parallel layers. Owing to later shear folding they contain mylonitic and sheared zones probably associated with a long N-S trending structure called the Hirvaskoski Shear Zone (Kärki et al., 1993). However, graded beds up to some cm in thickness are often observed, and form at times meters of thick layers with or without porphyroblasts. Garnet and staurolite are most frequent, but also staurolite and andalusite form alternating layers. Sillimanite-bearing quartzites occur as well. The structures suggest a turbidite origin and the Ahola occurrence at Posio with small-scale graded beds has been interpreted to represent distal turbidites (Evins 1997). The latest mapping results suggest that also felsic volcanic rocks occur close to Ahola.

Around Jukua on the western side of the Posio greenstone the gneiss belt consists mainly of fine-grained quartz-feldspar rocks with varying mica contents and alternating mica-rich beds, but also of minor mica gneisses. Banded amphibolites containing garnet-, calc-silicate- and quartz-rich layers are not uncommon and there is also a long stratiform horizon of sulfide-bearing felsic gneisses. Graded structures suggest that the rocks are probably proximal turbidites with closely associated mafic and felsic metavolcanics. However, they are so strongly deformed that their internal stratigraphy is hard to resolve. Quartzites cover them and these gneisses have been previously interpreted as quartzites containing mica-rich, mafic and calc-silicate intercalations (Honkamo 1979). In places the gneisses are highly sheared and intruded by granites, but they can be traced as a belt far southwards. At Puolanka in the Kainuu schist belt quite similar multiply deformed gneisses are interpreted as mainly turbidites in origin (Laajoki 1991).

The gneiss belt extends northwards from Posio for tens of kilometers into the Archean Suomujärvi Complex, where porphyroblast-bearing mica gneisses are still recognized while more felsic gneisses are thoroughly granitized. The gneisses are magnetic and they and their granitized remnants can be traced on low-altitude geophysical maps. Most probably, the same turbiditic gneisses with felsic volcanic rocks are found on the southwestern side of the Sodankylä schist area (Räsänen & Huhma 2001, *this volume*). However, alternating layers rich in porphyroblasts are not found through the long gneissic belt. Gneisses without porphyroblasts or with some garnet are more common south of Posio and, at Niskavaara, the belt includes also felsic volcanic rocks. The gneisses also come into contact with the 2.44 Ga old layered intrusions, which seem to penetrate them, but so far their primary contact is unknown.

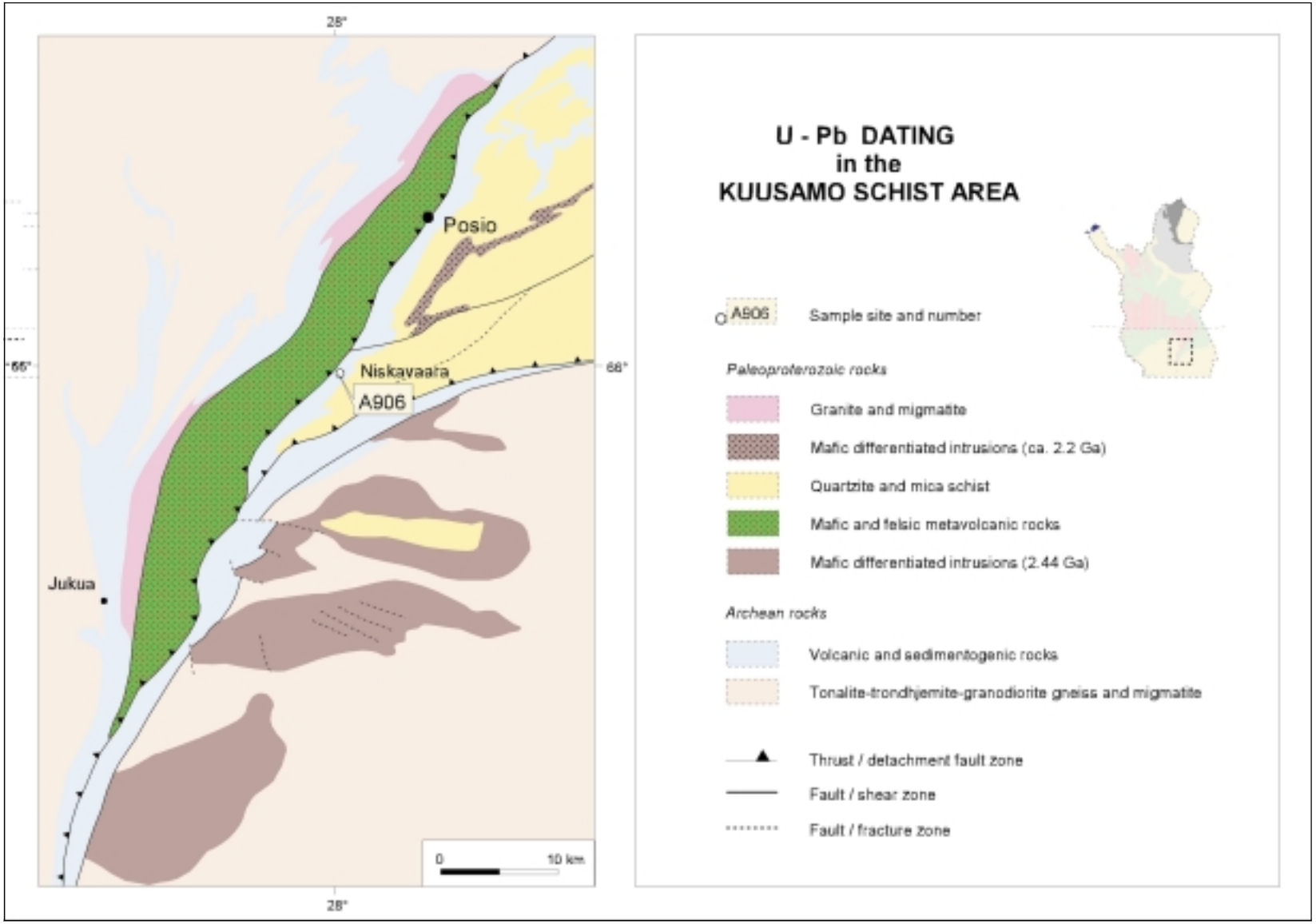

Fig. 3. Lithological outline and main structural features of the western part of the Kuusamo schist area.

THE NISKAVAARA FELSIC VOLCANIC ROCK: SAMPLE A906

Niskavaara is a hill about 15 km southwest of Posio on the southern shore of lake Livojärvi west of the Kuusamo Schist area (Figs. 2 and 3) and, the acidic volcanic rocks on and around it are called the Niskavaara Formation. They are intensely folded and highly schistose, and almost all primary structures are destroyed. For example, bedding can be observed only where compositional differences are recognized. The formation includes at least one massive unit, which may be lava-born in origin, but as no structures demonstrating primary lava flows are found, it is interpreted to represent a tuff. It occurs together with graded schists, which probably are reworked tuffites as they contain fine-grained layers rich in quartz and feldspar and very fine-grained layers rich in sericite. These also contain fine-grained magnetite, which may form up to some mm thick stratified laminae. Some of the tuffites contain up to 1 mm large garnet porphyroblasts and are weakly sulfide-bearing.

Sample A906 (map sheet 3544 02, northing 7322.31, easting 3545.72) was taken from a massive, homogeneous and about 10 m thick unit occurring with stratified tuffites. Chemically the rock is a rhyolite. It is weakly microphyric containing subhedral and angular to oval-shaped crystals of plagioclase and quartz up to 1 mm in size in a very fine-grained groundmass. The plagioclase is twinned but sericitized and mainly albite in composition. Quartz occurs commonly as elongate crystals with undulatory extinction. The groundmass around the phenocrysts is composed of sericite and chlorite with minute quartz and plagioclase. The most common accessory mineral is magnetite. It occurs as disseminated grains usually less than 0.5 mm in size and forms minute streaks parallel to the schistosity. Also some pyrite and pyrrhotite exist. The texture is foliated and exhibits a gneiss-like fabric due to deformation.

The sample contained abundantly relatively small-grained (70% <70 μm) zircons, but there are also larger, partly broken (>150 μm) crystals, possibly derived from a more coarse-grained source. In contrast to most igneous zircons, which exhibit simple prismatic-pyramidal morphology, zircons with pyramidal faces having higher order indexes are common in the Niskavaara sample. The crystal edges of some grains are rather sharp, while others appear more rounded, reminiscent of abrasion during transportation. The finer crystals occur as two color varieties, translucent pale brown and transparent colorless. There is no obvious morphological difference between the two types, but this may be due to the smallish grain size. The larger crystals are invariably brown.

The analytical results (Table 1) show that the

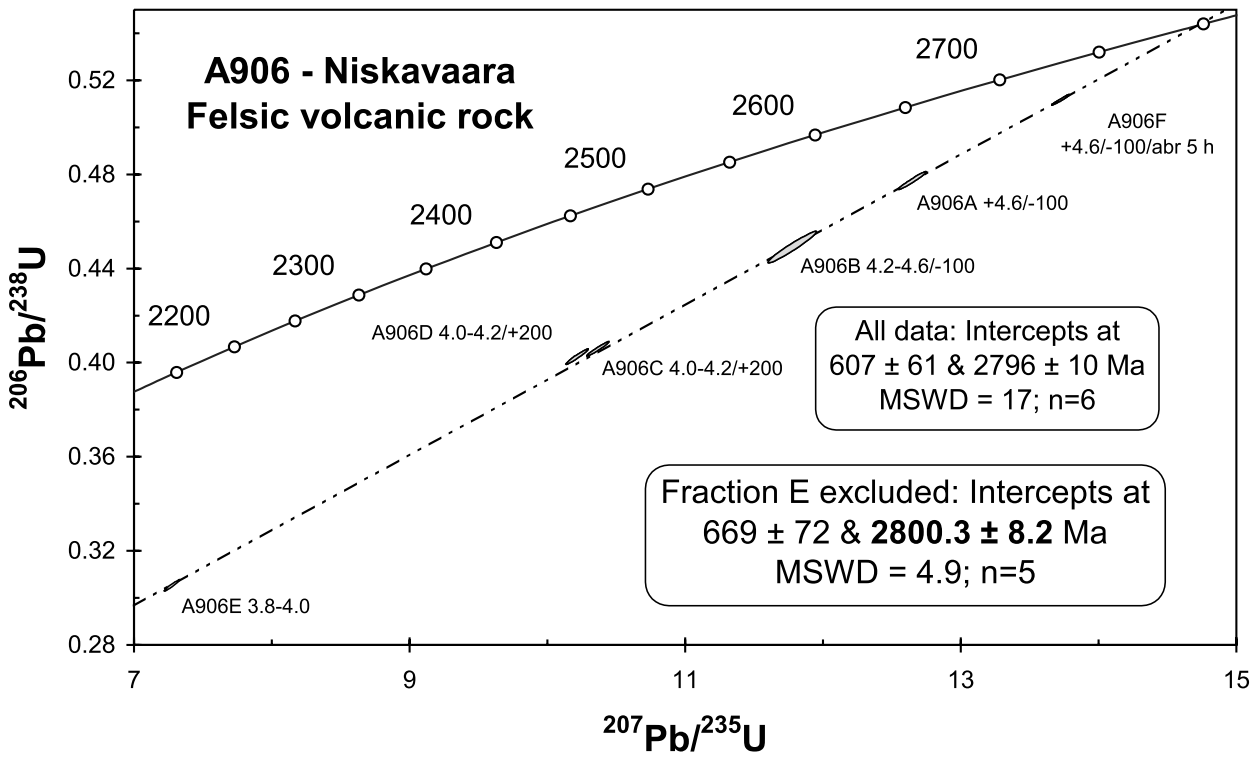

Fig. 4. Concordia diagram of the U-Pb analyses of zircons from sample A906-Niskavaara.

Table 1. U-Pb age data on zircons from the volcanogenic felsic gneiss A906 - Niskavaara.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES / Ma		
Fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A906A +4.6/-100	27.8	211.4	114.9	9908	0.126	0.4768	0.65	12.645	0.65	0.1923	0.15	0.97	2513	2653	2762
A906B 4.2-4.6/-100	24.6	400.1	203.9	6232	0.116	0.4486	1.23	11.777	1.23	0.1904	0.15	0.97	2389	2586	2745
A906C 4.0-4.2/+200	16.9	671.9	303.7	6559	0.103	0.4052	0.65	10.368	0.65	0.1856	0.15	0.97	2193	2468	2703
A906D 4.0-4.2/+200	18.4	682.9	305.9	9768	0.106	0.4020	0.65	10.217	0.65	0.1844	0.15	0.97	2178	2454	2692
A906E 3.8-4.0	12.3	971.9	325.5	6030	0.100	0.3045	0.65	7.284	0.65	0.1735	0.15	0.97	1713	2146	2591
A906F +4.6/-100/abr 5 h	0.6	64.6	38.2	18169	0.149	0.5113	0.38	13.719	0.39	0.1946	0.07	0.98	2662	2731	2782

*Isotopic ratios corrected for blank and common lead
** error correlation

uranium contents in the zircons are usual, ranging from 65 to 970 ppm. The discordancy pattern of the analyses is normal, i.e. the abraded fraction F is least discordant and the degree of discordance increases with increasing uranium contents and decreasing density. As demonstrated by the high $^{206}Pb/^{204}Pb$ ratios, the initial lead contents are low, and consequently the exact lead isotopic composition of the common lead correction has no bearing on the various isotope ratios determined. The largish errors in the U/Pb ratios of fraction B arise from an unusually poor uranium concentration run.

On a concordia diagram (Fig. 5), the data form a sublinear trend with intercepts at 2796±10 and 607±61 Ma. However, there is some scatter in excess of analytical error, demonstrated by the largish MSWD of 17. If the very discordant analysis of the lightest fraction E is excluded, the results become 2800±8 and 669±72 Ma with an MSWD of 4.9.

As all of the analyses from sample A906 are bulk fractions, the MSWD of 4.9 might indicate some variation in the isotopic properties of the different color and morphological zircon varieties. The only proper way to check this would be carrying out an ionprobe investigation. However, the scatter on the concordia diagram is relatively small and arises mainly from the 4.0-4.2 density fractions C and D. Even if the Niskavaara rock should consist of redeposited volcanic material, the time interval between the metamorphic and volcanic episodes has apparently been small, and the new analysis of the abraded fraction F in particular proves the Archean origin of the rock beyond any doubt.

DISCUSSION AND CONCLUSIONS

Correlations within the Kuusamo schist area

Niskavaara-type gneisses have not been found in eastern Kuusamo, where the supracrustal succession commences with Proterozoic volcanic rocks. In the western Kuusamo schist area the gneisses enclose the Posio greenstone which contains volcanogenic mafic rocks showing many similarities with mafic lavas of the Greenstone Formation I at Kuusamo. Using sedimentary criteria, Laajoki (1997) concluded that the Posio greenstone is overlain by turbiditic rocks. Later Airo (1999), based on geophysical parameters, interpreted that these rocks belong to a single structural unit and that the Posio greenstone, lying between gneisses, dips to the east and continues under the Kuusamo schist area.

However, field observations during current bedrock mapping indicate that the gneisses east and west of the Posio greenstone do not significantly differ from each other. They contain different lithologies but both of them display sedimentary structures suggesting similar turbiditic environments of deposition and probably represent different facies of turbidity fans associated with volcanism. The western contact of the Posio greenstone is so far unknown, but the eastern contact seems to be tectonic throughout, as is also indicated by the low-altitude magnetic data. The asymmetric mode of occurrence of the Posio greenstone supports a large half-graben environment for these mainly subaerial flows, which probably extruded on folded Archean supracrustal rocks.

The Posio greenstone is still correlated with Greenstone Formation I at Kuusamo. However, because of the definite Archean age of sample A906-Niskavaara, the rocks of the gneiss belt must be considered as a stratigraphic unit in its own right. This invalidates the previous lithostratigraphic correlation with the Proterozoic rocks in eastern Kuusamo. We emphasize that this does not disprove the present lithostratigraphic classification in eastern Kuusamo, but simply shows that it cannot be applied as such to the western end of the Kuusamo schist area.

Regional considerations

The felsic metavolcanics at Niskavaara yielding an Archean age occur together with turbiditic, multiply deformed gneisses, which exist as a long, mostly NS-trending belt on the western side of the Kuusamo schist area. Most probably the same volcano-sedimentary gneiss belt continues in the north into the Sodankylä schist area, where similar gneisses with Archean (2775±25 Ma) tuffitic (Keski-Loviselkä, Räsänen & Huhma 2001, *this volume*) rocks are found. In the south the gneiss belt continues into the highly deformed and tectonized Kainuu schist belt characterized by shear zones, several faults and thrusts, but also there felsic volcanic rocks have been recently recognized (Kontinen et al. 1996).

The Central Puolanka Group in the Kainuu schist belt has been considered Paleoproterozoic and is often correlated stratigraphically with the Lapponia Supergroup (e.g., Laajoki 1986a and b, 1988, 1991) explaining that they belong in the same Kainuu-Lapponi tectofacies (e.g., Laajoki 1990, 1991). The correlation is no more valid because the gneisses of the Central Puolanka Group contain Archean volcanic rocks. Zircons from a felsic tuff yield ionprobe ages of 2.72 Ga and some close to 2.8 Ga (Huhma et al. 2000). Likewise, neodymium T_{DM} model ages in the order of 2.9-3.0 Ga have been reported for three volcanic rocks in the same area (op. cit.).

We thus conclude that the reinterpretation of the stratigraphical position of the acidic volcanic rock at Niskavaara suggests that remnants of a late Archean schist belt extend from Sodankylä via Kuusamo to Puolanka. Their geological and temporal similarity is supported by field evidence and isotopic data, and aeromagnetic data also seem to favor the concept.

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Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
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U-Pb DATINGS IN THE SODANKYLÄ SCHIST AREA, CENTRAL FINNISH LAPLAND

by
Jorma Räsänen and Hannu Huhma

Räsänen, J. & Huhma, H. 2001. U-Pb datings in the Sodankylä schist area, central Finnish Lapland. *Geological Survey of Finland, Special Paper 33,* 153-188. 19 figures, 2 tables and one appendix.

The bedrock of the Sodankylä schist area consists mainly of Paleoproterozoic volcanic and sedimentary rocks penetrated by mafic and felsic intrusives. This paper reports c. 200 conventional U-Pb analyses mainly on zircon. A total of 34 samples were studied from 12 mafic intrusions, 3 granitoids, 5 metavolcanic rocks and 4 metasediments. Many samples yielded discordant and heterogeneous data, either due to inheritance or effects of metamorphism.

The age results obtained suggest that a supracrustal belt running on the southwestern side of the Sodankylä schist area and continuing inside the Central Lapland Granite Complex is late Archean in origin. Besides amphibolites, it includes mainly different kinds of gneisses associated with 2775±25 Ma old felsic volcanic rocks. The Proterozoic volcanic succession starts with felsic extrusives, which have yielded an age of 2438±11 Ma and seem to be coeval with the emplacement of the layered mafic intrusions which indicate the break-up of the Archean continent at the beginning of the Proterozoic eon.

The overlying volcano-sedimentary pile is penetrated by intrusive rocks of various ages. The most reliable results from mafic intrusions include ages such as 2222±6 Ma (Harjunoja), 2148±11 Ma (Rantavaara), 2070±5 Ma (Ahvenselkä) and 2055±5 Ma (Rovasvaara). Accordingly, the age of deposition of the Sodankylä and Savukoski Group supracrustal rocks is considered older than 2.22 Ga and 2.05 Ga, respectively.

A granodiorite associated with a gabbroic rock inside the schist area yields an age of 1891±5 Ma indicating that intrusions of the Haaparanta Suite are not restricted only to western Lapland. The 1.8 Ga thermal episode in the area is manifested by an age of 1808±8 Ma for a granite pegmatite dyke, as well as a few U-Pb ages on titanites. Several granites formed at this stage contain zircons inherited from older, presumably Archean sources.

Key words (GeoRef Thesaurus AGI): absolute age, U/Pb, zircon, metavolcanic rocks, intrusions, gabbros, metadiabase, granites, Paleoproterozoic, Archean, Sodankylä, Lappi Province, Finland

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INTRODUCTION

The Central Lapland Greenstone Belt in northern Finland is part of a widespread, c. 1000 km long NW-SE trending Paleoproterozoic supracrustal belt, extending from Russian Karelia to northern Norway and Sweden. On the basis of lithology, the Central Lapland Greenstone Belt can be divided into subareas, which are from southeast to northwest the Kuusamo schist area, the Salla greenstone area, the Sodankylä schist area and the Kittilä greenstone area. The Peräpohja schist belt further southwest records the same early Proterozoic regime, but is separated from the Central Lapland Greenstone Belt by a large granitic terrain, about 20,000 km^2 in extent, called the Central Lapland Granite Complex and generally considered to be principally Proterozoic in age.

The Sodankylä Schist area occupies the middle part of the Central Lapland Greenstone Belt. The bedrock consists mainly of Proterozoic supracrustals containing volcanic and sedimentary rocks penetrated by mafic and felsic intrusives. Direct age determinations on volcanic rocks have generally been difficult and only few age data have been available. Instead, dating of several intrusive rocks has been successful. The data considered in this paper are based on c. 200 conventional U-Pb analyses of zircons and represent 34 rock samples taken by several researchers during the last decades. Most of the dated rocks are mafic intrusions and some of the ages have been published earlier. However, as their analytical data are still unpublished and as additional analyses and even resampling has been done in some cases, all the data will be published here.

Because of the fact that, probably due to later metamorphism, many samples contain heterogeneous zircon populations and yield discordant data, no precise ages could be determined for many of the rocks. Dating rocks containing inherited zircons, such as S-type granites is hard by conventional methods and requires ionprobe techniques. The same applies to detrital zircons in sedimentary rocks. In spite of these problems, several reliable ages for mafic intrusions have been obtained and these give information of separate magmatic phases, some of which have produced also extrusive rocks.

It must be stressed that all the described lithological units have undergone metamorphism and consequently all rock names should also carry the prefix meta-. However, for simplicity and to avoid tautology it is not used in this paper. It should be also mentioned that the map in Figure 1, based on the 1:1,000,000 bedrock map of Finland, does not show all lithological variations in the study area.

GEOLOGICAL OUTLINE

The Central Lapland Greenstone Belt consists of areas where either volcanic or sedimentary rocks are prevailing. The middle part of the belt, roughly the area between Sodankylä and Savukoski, is dominated by sedimentary rocks and is called the Sodankylä schist area (Räsänen 1991). It is bounded in the southeast with a gradational contact to the Salla greenstone area but in the northwest its contact with the Kittilä greenstone area is tectonic. The area is bounded in the northeast by the Archean Basement Complex and comes in contact with the Central Lapland Granite Complex in the southwest (Fig. 1).

The geological outline of Finnish Lapland was drawn by Mikkola (1941) and the bedrock mapping at the scale 1:100,000 initiated in the Sodankylä district in the beginning of the 1970s (Tyrväinen 1983). Based on studies of the Lapland Volcanite Project, Räsänen et al. (1996) presented a formal stratigraphic classification scheme, where the Paleoproterozoic supracrustal rocks of the Central Lapland Greenstone Belt were divided into seven groups. From oldest to youngest, they are the Salla, Onkamo, Sodankylä, Savukoski, Kittilä, Lainio and Kumpu Groups. The same classification together with the type formations was also used by Lehtonen et al. (1998).

Well-preserved sedimentary rocks occur in the Sodankylä schist area and it is easy to observe that the volcanic rocks of the Salla and Onkamo Groups form anticlinal structures while the rocks of the Savukoski Group are placed in synclinal structures. However, the bedrock is polyfolded and all supracrustal rocks have been thrust eastwards, appearing now as different kinds of bulging longitudinal lithological units which can be traced on low-altitude geophysical maps. In places the folds have been overturned which, coupled with faults and overthrusts of varying intensity, has caused both discontinuities and repetitions of the supracrustal sequences.

The Proterozoic supracrustal lithology of the Sodankylä schist area starts with felsic volcanic rocks which belong to the Salla Group. However, they are not ubiquitous, and e.g. at Möykkelmä mafic and ultramafic volcanic rocks of the Onkamo Group lie directly on basement granitoids (Räsänen et al., 1989).

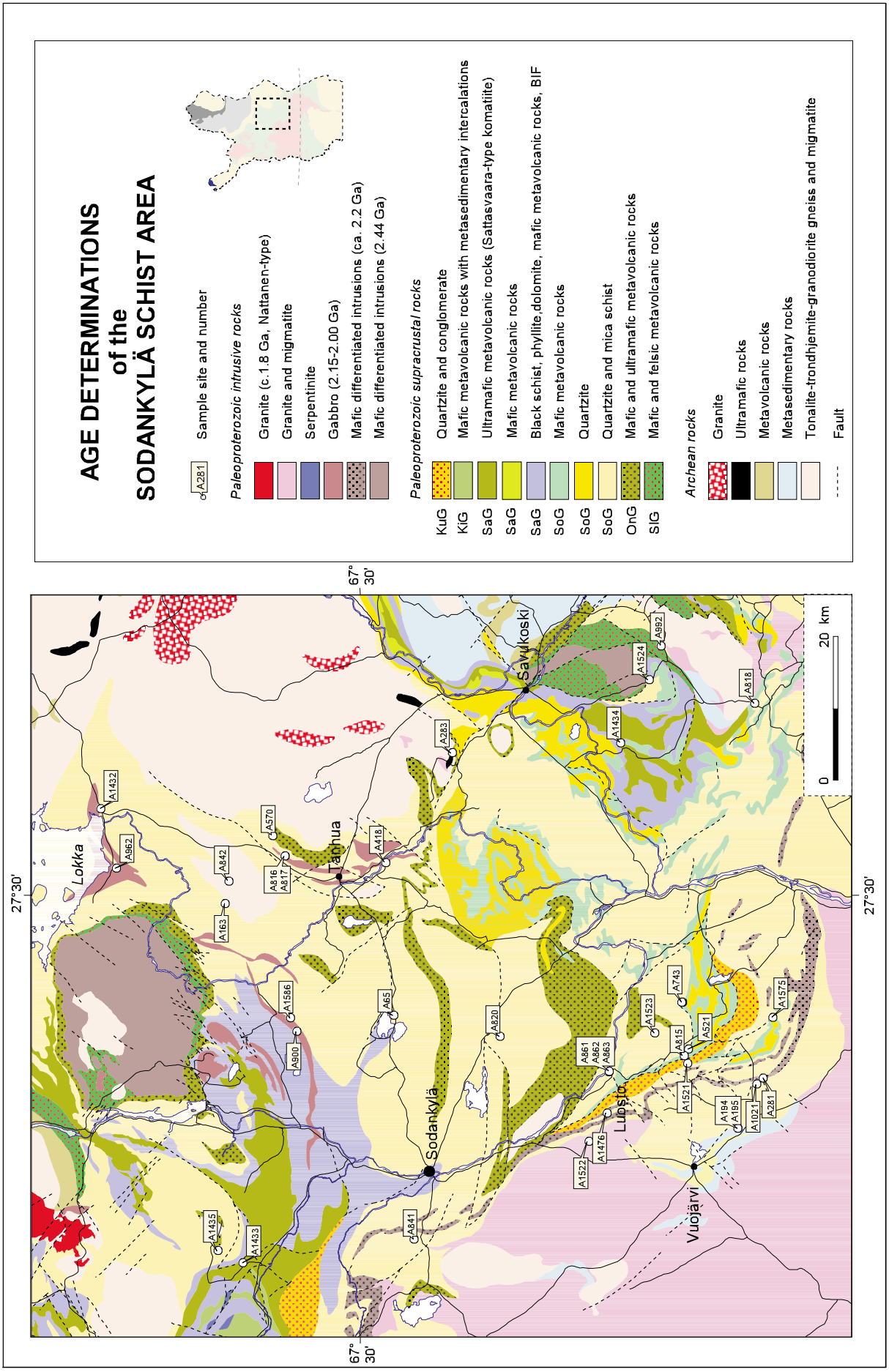

Fig. 1. Geological map of the Sodankylä Schist area. KuG = Kumpu Group, KiG = Kittilä Group, SaG = Savukoski Group, SoG = Sodankylä Group, OnG = Onkamo Group, SIG = Salla Group.

In the Central Lapland Greenstone Belt these rocks are volcanic products of the earliest Proterozoic continental break-up, and seem to be spatially closely related to 2.4 Ga old layered intrusions supporting an intracontinental, rift-related origin as interpreted in the Salla (Manninen 1991) and Kuusamo areas (Räsänen 1999). In the Sodankylä schist area, they are overlain by conglomerates, epicontinental clastic sediments and associated volcanic rocks of the Sodankylä Group, which represents a later stage of rifting with sedimentation and coeval volcanism.

On the whole the supracrustal pile thickens laterally westwards and the rocks of the Savukoski Group appear on the top of the Sodankylä Group. They start with graded schists and contain minor dolomites, black quartzites and sulfide-bearing black schist. These are intruded by the Keivitsa mafic intrusion dated at 2.05 Ga (Mutanen & Huhma 2001, *this volume*). The schists are overlain by mafic and ultramafic volcanic rocks and several mafic and ultramafic dykes, called diabases and olivine gabbro-diabases, which penetrate the Keivitsa intrusion (Mutanen 1997). These dykes are Mg-basalts and komatiites in chemical composition and resemble the komatiites of the Sattasvaara Formation, ranging from Mg-basalts to basaltic and peridotitic komatiites (Räsänen 1996). Chemical data suggest that these dykes may form part of the Sattasvaara Formation. However, reliable age data of these rocks are still missing.

In the west, a major tectonic zone marked by a chain of ophiolitic serpentinites (Hanski 1997) occurs on the eastern border of the Kittilä greenstone area (Lehtonen et al. 1998) and separates the rocks of the Savukoski Group from the volcanic rocks of the Kittilä Group. U-Pb ages of 2.01 Ga for felsic porphyries and a Sm-Nd age of 1.99 Ga for tholeiitic basalts of the Vesmäjärvi Formation on the top of the Kittilä Group have been reported (Hanski et al. 1997). Also a U-Pb age 1.92 Ga for a felsic dyke has been reported (Lehtonen et al. 1992).

Polymictic conglomerates and associated quartzites of the Kumpu Group overlie unconformably the folded supracrustal pile described above. Most of these coarse clastic deposits occur in the western part of the Central Lapland Greenstone Belt and based on U-Pb ages of 1.88 Ga for granitoid pebbles of conglomerates (Hanski et al. 1997) they are post-collisional molasse deposits. They are found also in the Sodankylä Schist area (Haimi 1977, Räsänen 1977, Nikula 1985). A typical feature of these conglomerate-quartzite deposits is that they are not more than a few kilometers wide, but may reach thicknesses of more than two kilometers. At Pyhätunturi, in the southern part of the area, occurs a similar thick sequence of well-exposed conglomerates and quartzites, which unconformably overlie supracrustals of the Sodankylä Group. Based on sedimentary studies they are interpreted to be mainly a river-born deposit (Räsänen & Mäkelä 1988) and are called the Pyhätunturi Formation assigned to the Kumpu Group (Räsänen et al. 1995).

The Central Lapland Granite Complex is heterogeneous and composed of various Proterozoic granites post-dating the crustal thickening during the Svecofennian orogeny, as discussed by Gaál and Gorbatschev (1987). This granitic complex, close to 20 000 km^2 in size and so far poorly known, includes remnants of Proterozoic schists and a variety of Archean rocks. For example, the Suonivaara granodiorite about 15 km SW of Sodankylä, which is one of the Archean basement rocks (Tyrväinen 1979, 1983), is an inhomogeneous gneissose granite associated with migmatitic cordierite-, garnet-, staurolite- and andalusite-bearing mica gneisses together with quartz-feldspar gneisses. These gneisses, though they are folded and granitized to variable degrees, can be followed for tens of kilometers southeastwards through the granitoid terrain. In the Vuojärvi area, they are much better preserved and associated with felsic volcanic rocks (Räsänen 1986a).

ARCHEAN ROCKS

A194 and A195 Keski-Loviselkä felsic volcanic rock. Keski-Loviselkä lies on the southwestern border of the Sodankylä schist area, close to Vuojärvi, about 40 km south of Sodankylä (Fig. 1). In the key area, felsic volcanic rocks occur in association with cordierite-, garnet-, staurolite- and andalusite-bearing mica gneisses called the Loviselkä Formation. All rocks are isoclinally folded and the felsic volcanic rocks have been metamorphosed to quartz-feldspar gneisses. In spite of folding, sedimentary structures are well-preserved and show that the porphyroblast-bearing mica gneisses with quartzitic lower parts, containing almost complete Bouma-cycles, are proximal turbidites in origin. The rocks can be traced southeastwards into the granitoid complex and the latest mapping results indicate that similar turbiditic gneisses together with Archean felsic volcanics can be found as a long NS-trending belt on the western side of the Kuusamo Schist area (Räsänen & Vaasjoki 2001, *this volume*).

The felsic volcanic rocks at Keski-Loviselkä are dacites or quartz andesites and differ from the associated mica gneisses in chemical composition for example by having higher SiO_2- and lower Al_2O_3 and $Fe_2O_3^{tot}$ contents. They display a cyclic appearance suggesting a partly reworked tuff/tuffitic (ash-flow) deposit. The lower part of a unit, about 10 m in thickness, is massive but contains some mica-rich intraclasts. It is overlain by a 1-2 m thick ripple-cross laminated part and capped by parallelly laminated calcareous layers. Diamond drilling conducted by Rautaruukki Co. in the late 1970s has shown that they also include layers of felsic volcaniclastic rocks, which may be pyroclastites in origin. However, the stratified upper part suggests that they are redeposited and probably originate from a mass flow of volcanic debris carried by a turbidity current further down slope into a basin, while the associated turbiditic greywackes indicate the contemporaneity of sedimentation and volcanism.

Sample A194 is taken from the massive lower part and under microscope the rock is granoblastic and very fine grained (0.01-0.4 mm). It consists of oligoclase, quartz and biotite with muscovite, chlorite and K-feldspar. Magnetite, apatite, rutile and zircon, up to 0.1 mm, are common accessory minerals. The zircon occurs as yellowish short prisms, which often have slightly rounded edges. In the fine-grained (-200 mesh) heavy fraction crystals are generally clear with sharp edges. Sample A195 was taken from the graded tuffites, which include some brecciated parts rich in scapolite. It contains more biotite and also some titanite, but otherwise it does not significantly differ from the massive unit in mineral composition. The appearance of the zircon population does not much differ from zircon crystals of the massive part.

Analytical results are given in Appendix I and a concordia diagram is shown in Figure 2. The four discordant U-Pb analyses from zircons in the massive flow A194 define a trend which gives an upper intercept age of 2775±25 Ma, and a lower intercept of 430±150 Ma. The large MSWD of 16 is due to incompatibility of the two most discordant analyses. It is quite unusual that heavy zircons (A) do not yield analysis closer to concordia compared to the less heavy fraction C. As the old analyses (both A and C) have been made using separate Pb and U spike solutions, it may be conceivable that e.g. occasional weighing error could explain the problem. We have to point out, however, that several duplicated analyses in the laboratory have yielded consistent results.

If the analytically somewhat inferior fraction A is omitted from the calculation, the upper intercept age for the massive flow A194 becomes 2777±3 Ma. The overlying tuffites (A195) register an upper intercept age of 2780±72 Ma with an MSWD of 9, which may reflect the incorporation of external detrital material into this rock. The titanite in this rock (not shown on the diagram) is slightly discordant, but nevertheless demonstrates a later, Paleoproterozoic thermal event, most likely during the Svecofennian orogeny.

A283 Kivivuotsonselkä ultramafic volcanic rock. Kivivuotsonselkä lies on the northeastern border of the Sodankylä Schist area about 10 km northwest of Savukoski (Fig. 1). The bedrock consists of granitic gneisses and ultramafic rocks, which are overlain by supracrustal rocks (Juopperi 1986). The ultramafic rocks are komatiites and display well-preserved lava structures with cumulate bases overlain by spinifex-textured zones and flow top breccias (Räsänen 1984). They occur in conjunction with granitic gneisses and, in addition to orthogneisses, there occur also plenty of cordierite-, garnet- and

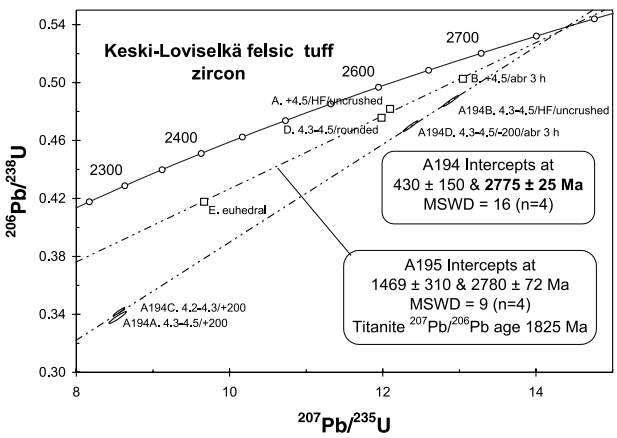

Fig. 2. Results for U-Pb analyses on zircons from the Keski-Loviselkä felsic rocks. The ellipses denote the analytical error on the 95% confidence level.

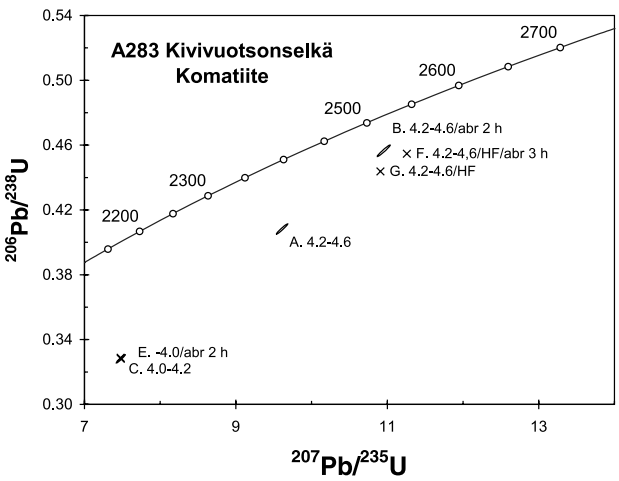

Fig. 3. Results for U-Pb analyses on zircons from the Kivivuotsonselkä komatiite.

staurolite-bearing paragneisses as layers reminiscent of primary bedding.

Between the granitic gneisses and the komatiites there is a 1-3 m thick ultramafic chlorite-rich hybrid rock, which in places contains partly melted and at their rims corroded xenoliths of granitic gneisses.In addition to chlorite, amphibole and biotite are the major minerals, while magnetite, apatite, ilmenite and zircon are common accessories. Zircon, up to 0.3 mm in size, is mainly euhedral but turbid with rounded edges.

Sample A283, taken by H. Juopperi in the 1980s, is a chlorite-rich hybrid rock. Chemically it is a komatiite, but it is contaminated and strongly enriched in LREE indicating that the rock originated from a hybrid melt, which was formed by thermal erosion of country rock by hot komatiite flows. This is a common phenomenon and often occurs during the ascent of an extremely hot komatiitic magma or when it flows over a felsic base rock (e.g. Groves et al. 1986, Juteau et al. 1988).

The analytical data are given in Appendix 1 and a U-Pb concordia diagram is presented in Figure 3. From the seven fractions treated, one (D) was analyzed only for its lead isotopic composition, and is consequently not shown on the diagram. The two most discordant fractions (C and E) are almost identical and overlap in Figure3. There is considerable heterogeneity in the other four fractions, and no proper age result can be obtained apart from the bulk zircons being of a definitely Archean origin. Given the komatiitic nature of the rock and its established contamination, it seems likely that the zircons are xenocrystic and reflect rather a minimum age for the country rocks than the time of lava extrusion. In fact, the Kivivuotsonselkä komatiite flows can be Paleoproterozoic in origin.

PROTEROZOIC ROCKS

Felsic volcanic rocks

A992 Purkkivaara felsic tuff. Purkkivaara is a hill 20 km south of Savukoski (Fig. 1). In the area there occur extensive felsic volcanic rocks consisting of amygdaloidal lavas and tuffs, called the Purkkivaara formation. In the southeast, it is associated with quartzites and minor garnet-staurolite gneisses, while 5 km to the northwest, at Akanvaara, the felsic volcanics are in contact with a southeastward-facing layered gabbro intrusion capped by a granophyre (Räsänen 1983). TheAkanvaara gabbro is c. 2430 Ma old (Mutanen & Huhma 2001, *this volume*) and was investigated in detail during prospecting activity by the GTK in the 1990s. It is well-documented by Mutanen (1997), according to whom the magma intruded into the supracrustal rocks described above and *"all the acid volcanites intersected by diamond drill holes below the intrusion are hornfelses"* and *"the hornfelses analyzed so far are compositionally affected by the intrusion and hence resemble the ferrogranophyres of the roof"*. Thus the age of the Purkkivaara felsic volcanic rocks, dacites and rhyolites in chemical composition, is of considerable interest.

Sample A992, taken in the 1980's, is a crystal tuff from a large outcrop on the southeastern slope of the Purkkivaara hill. The rock is very fine-grained, its weathered surface is whitish gray with thin biotite-streaks while the fresh surface is dark gray in color. All the outcrops are schistose due to strong axial plane cleavage associated with folding which has generally destroyed the primary structures. However, it is occasionally obvious that the rocks were originally stratified. There exist layers up to few tens of cm's thick containing very fine-grained lithic fragments and plagioclase crystals, up to 1-2 mm in size, and alternating homogeneous layers with obscure contacts without plagioclase. Also amphibole-bearing interlayers containing abundant biotite exist among them.

Microscopically plagioclase ($An_{\sim 10}$) forms eu- and subhedral or roundish crystals up to 2.5 mm in size and the crystals are commonly twinned. Some crystal-rich layers contain broken crystals indicative of pyroclastic fallout deposits. The very fine-grained matrix is principally composed of quartz and plagioclase with biotite, while amphibole is a common accessory mineral. The amount of secondary sericite is highly variable and usually magnetite is also present.

Four samples, both flows and tuffs, were taken in the 1980s from these volcanic rocks of the Purkkivaara formation. However, only a small amount of very fine-grained zircon was extracted. They are commonly brownish in color and often have a crushed appearance. They are not rounded and a quarter of the zircons exhibit tetragonal prismatic crystal forms with L/B 1.5-2. Some transparent crystals occur as well.

The zircon has a high uranium content, and is consequently of very low density, providing discordant analyses. The four analyses give an upper intercept age of 2412±40 Ma, but the high MSWD (24) shows scatter in excess of analytical error (Fig 4.). It should be emphasized that the two most

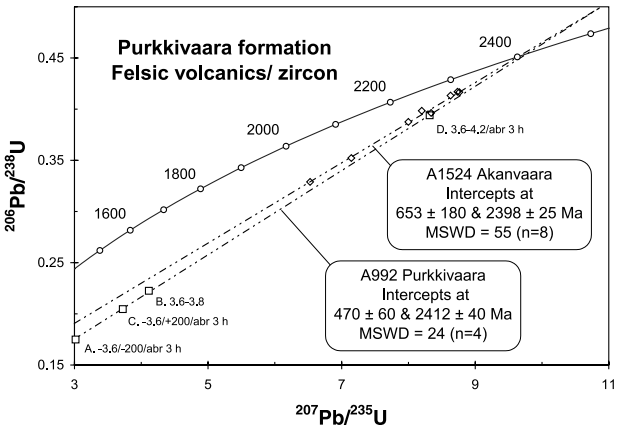

Fig. 4. Results for U-Pb analyses on zircons from the Purkkivaara felsic tuff and the Akanvaara quartz-feldspar porphyry.

discordant analyses were made on unusually low-density zircon and should be treated with caution. It is likely that the felsic volcanic rocks are Proterozoic, but the exact age remains obscure. The obtained age estimate is close to the age of the Akanvaara layered mafic intrusion, 2.43 Ga, (Mutanen & Huhma 2001, *this volume*), suggesting that felsic volcanism may have been associated with this enormous mafic magmatic event.

A1524 Akanvaara quartz-feldspar porphyry. Akanvaara is a large hill about 15 km south of Savukoski (Fig. 1.). The bedrock comprises felsic volcanics of the Purkkivaara formation and plagioclase-phyric amygdaloidal flows associated with quartz-feldspar porphyries. They are overlain by quartzites and schists of the Ritaselkä formation in the east and south. In the north, they are closely associated with granophyric rocks lying on the top of the Akanvaara layered mafic intrusion without any observed sharp contact.

Sample A1524 was taken from a rhyolitic quartz-feldspar porphyric rock representing a massive flow with quartz-filled vesicles on top. It contains 1-2 mm phenocrysts of quartz and feldspar in a very fine-grained granular groundmass composed of quartz and feldspar with biotite and sericite. Magnetite, some carbonate mineral and also zircon are present. Albitic plagioclase exhibit lamellar twinned euhedral phenocrysts up to 2.5 mm in size, which contain minute inclusions of quartz and potassium feldspar. Some phenocrysts of quartz show angular hexagonal outlines obscured by embayments and minute inclusions of feldspar. Also granophyric crystallization of quartz and potassium feldspar exists and spherulitic structures of plagioclase are not uncommon. The structures indicate high-temperature devitrification of natural glass found in volcanic rocks.

Sample A1524 yielded abundant zircon, which occurs as euhedral, relatively short, pale-brown crystals. In contrast to the previous sample (A992) much of the zircons are heavy (+4.3), but somewhat turbid. Eight U-Pb analyses made on sub-milligram quantities are technically good, but give scattered data on the concordia diagram (Fig. 4.). The data points define a line which has an upper intercept at 2398±25 Ma, and a lower intercept at 653±180 Ma (MSWD=55!). The most deviating points represent clear zircon (D) and abraded long crystals (F), and without these points the intercepts would be 2394±11 Ma and 629±79 Ma (MSWD=7). The heterogeneity observed is considered to be due to metamorphic effects presumably during 1.8-1.9 Ga tectonothermal events. Similar, more pronounced effects have been observed e.g. at the Siikakämä intrusion (Mertanen et al., 1989). The data available do not allow reliable dating, but it is very likely that the age of volcanism is fairly close to the age of the Akanvaara gabbro, which is considered intruding these volcanics.

A1432 and A1498 Sakiamaa volcanics. Sakiamaa is in the northeastern corner of the Sodankylä Schist area about 60 km north of Savukoski on the southern shore of the Lokka reservoir (Fig. 1). The lithological information on the area is mainly based on works by Mikkola (1937, 1941), according to whom the bedrock consist of quartzites and mica schists together with leptitic schists. At the time the term leptite was used for most felsic schists of unknown origin composed principally of quartz, feldspar and mica, often with gneissose fabric. Some targets of these poorly known rocks have been reinvestigated and the rocks described formerly as quartz-feldspar schists or mica arkoses are, in fact, felsic volcanics in origin. They include stratified tuffs and plagioclase crystal tuffs containing also felsic interflows. These are associated with garnet-bearing quartz-feldspar schists and garnet-staurolite gneisses and quartzites. Chemically the volcanics are dacites and rhyolites.

Samples A1432 and A1498 were taken from a massive part of the volcanic unit, which is about 80 m thick and contains also stratified tuffites. The massive rock, about 5 m thick, is a very fine-grained slightly orientated plagioclase-phyric lava with granoblastic groundmass containing also vesicles, up to 3 mm in diameter, filled with quartz, epidote and chlorite. Quartz, saussuritized plagioclase exhibiting eu- to subhedral phenocrysts up to 1 mm in size and deep green hornblende are the main minerals, while chlorite, magnetite, biotite, epidote and titanite are common accessories. Also apatite and some hematite are present.

Zircon in sample A1432 occurs as small subhedral grains, which range from light-colored turbid to dark-

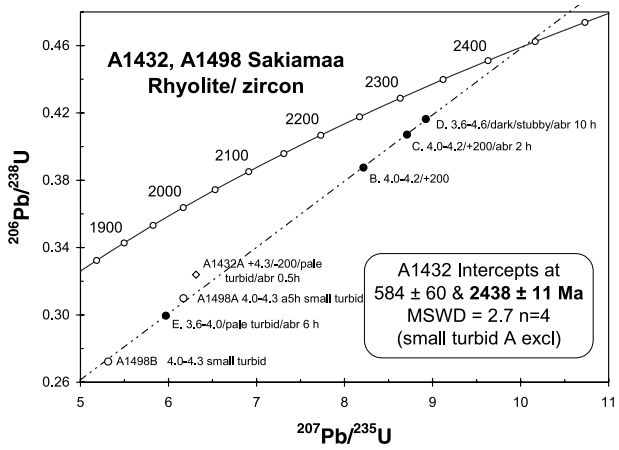

Fig. 5. Results for U-Pb analyses on zircons from the Sakiamaa felsic volcanics.

ish and fairly clear. Results of analyzed zircon fractions are given in Appendix I and the concordia diagram is presented in Figure 5. In spite of a high contents of uranium, the analyses on the dark zircon are not particularly discordant. Instead, an analysis on the heaviest fine-grained (-200 mesh), though turbid zircon (A) is badly discordant, although its uranium content is lower. This analysis lies clearly off the line defined by the others, and suggests disturbances due to metamorphism. The four analyses (on +200 mesh) yield an upper intercept age of 2438±11 Ma and a lower intercept at 584±60 Ma (MSWD=3, Fig. 5).

The zircon age estimate of 2438±11 Ma demonstrates volcanism coeval with the plutonic phase that generated numerous layered mafic intrusions in northeastern Fennoscandian Shield generally, and in particular the local Koitelainen and Akanvaara layered intrusions, dated at 2.43 Ga (Mutanen & Huhma 2001, *this volume*), lying on either side of the Sodankylä Schist area.

Mafic intrusions

A281 Harjunoja mafic sill. Harjunoja lies in the southwestern part of the schist area about 45 km south of Sodankylä (Fig. 1). The bedrock consists of quartzites, dolomites and siltstones associated with minor felsic volcanic rocks together with phyllites and carbonaceous schists. This volcano-sedimentary rock association, called the Harjunoja Formation, is penetrated by a dyke swarm composed of highly magnetic differentiated mafic sills. The magnetic susceptibility of the sills is about 100 times higher than that of the surrounding rocks and, consequently, they can be traced on aeromagnetic low-altitude maps for more than 50 km along strike.

The thicknesses of individual sills vary from tens to hundreds of meters, and they may reach one kilometer in total. The sills are differentiated and the thickest contain serpentinized olivine cumulates at the base, but generally the lowermost rocks are olivine-bearing pyroxene cumulates composed of variably uralitized pyroxene and epidotized labradoritic plagioclase with magnetite. The upper differentiates are gabbroic cumulates composed of plagioclase (An_{20-30}), amphibole, biotite, magnetite and epidote. Also some magnetite-rich plagioclase cumulates are present.

Sample A281 was taken from a coarse-grained gabbroic differentiate of the upper part of a sill and it contained abundant zircon. The zircon population is quite homogeneous and euhedral crystals possess well-developed (100) and (101) crystals faces, indicating a high temperature during crystallization (e.g. Pupin 1981). There also exist needle-like (L/B>5) crystals of zircon, which could indicate a fast crystallization rate. In addition, there are also glassy zircons, which exhibit numerous crystal faces.

Results of analyses are listed in Appendix I. The high $^{208}Pb/^{206}Pb$ ratio of suggests that when the rock crystallized Th and U were not yet fractionated in the source melt. The concordia plot for zircons is presented in Figure 6. The glassy zircon fractions of (F,G,H) are least discordant and a linear fit through all 12 zircon fractions has an upper intercept at 2222±6 Ma and a lower intercept at 383±58 Ma with an MSWD of 7. Exclusion of the most discordant fraction J from the calculation does not significantly alter these figures.

The obtained U-Pb age of 2222±6 Ma for the

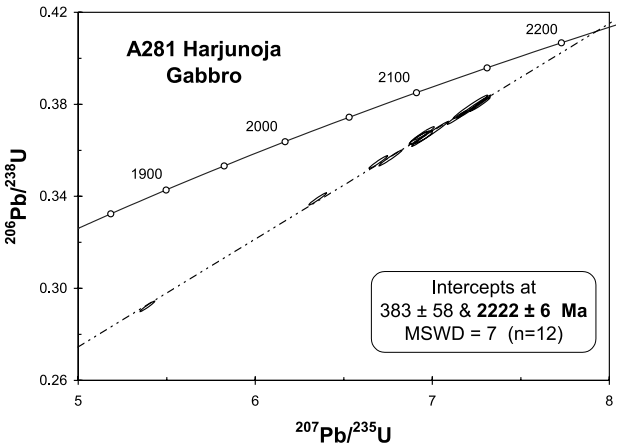

Fig. 6. Results for U-Pb analyses on zircons from the Harjunoja mafic intrusion.

Harjunoja layered sill is in good agreement with the Sm-Nd mineral age of 2.26±0.04 Ga for the same intrusion at Ahvenvaara (Huhma et al. 1996). A differentiated sill-like intrusion at Haaskalehto, which occurs about 20 km NW of Sodankylä, yields an U-Pb zircon age of 2220±11 Ma (Tyrväinen 1983; A892) and a similar Sm-Nd age of 2.19 Ga. From Silmäsvaara, 10 km further NW, a Sm-Nd age at 2.22 ±0.05 Ga is reported by Huhma et al. (1996). All these rocks have initial Nd-ratios close to the chondritic value. Similar differentiated dykes in the Salla area at Jokinenä and Pirtinkehäkukkura yield U-Pb zircon ages of 2209±10 Ma and 2202±2 Ma (Lauerma 1995; A379, A420+A830), respectively. Silvennoinen (1991; A847 Jäkäläniemi) reported a U-Pb age at 2206±9 Ma for a long dyke in the Kuusamo area and, there are abundantly quite similar differentiated mafic dykes in the Peräpohja area. E.g. at Kallinkangas, a U-Pb age of 2216 Ma has been obtained (Perttunen & Vaasjoki 2001, *this volume*).

A900 Rantavaara gabbro. Rantavaara is a hill about 20 km NNE from Sodankylä (Fig. 1.) where a mafic intrusion called the Rantavaara gabbro is exposed. It is a part of a more extensive, 200- 500 m thick sill-like intrusion, which can be traced for almost 25 km along strike roughly in a NE-direction. At Rantavaara the NW dipping (40-50°) intrusion is slightly differentiated containing minor ultramafic rocks at base, but the bulk consists of gabbroic rocks characterized by alternating plagioclase- and amphibole-rich magmatic layers with varying thickness. Epidotized plagioclase and green hornblende are the main minerals, often with chlorite, while titanite and magnetite are common accessories with some pyrite.

The intrusion penetrates quartzites and graded Al-bearing mica schists, which have been found on its southern side and are interpreted to represent the uppermost sedimentary rocks of the Sodankylä Group (Räsänen et al. 1996). Later it has been observed that on the northern side the intrusion is in contact with black schists and it has been presumed (Lehtonen et al. 1998) that the intrusion penetrates also the lowermost rocks of the Savukoski Group. The contact between the rocks is covered and, according to diamond drillings, it is sheared. No signs of a thermal effect caused by the intrusion have been observed in black schists; in contrast, they are highly folded against the intrusion. Based on low-altitude geophysical maps and ground measurements the black schists run for kilometers along the northern side of the intrusion. However, minor black schists are found also on the southern side of the intrusion, but also their contact with the intrusion consists of highly sheared and broken rocks as observed in drill cores. All these observations indicate a tectonic contact and support an interpretation of a thrust. Actually, this black schist may be one of those several thrust sheets observed in the Rajala area, which accompany the overthrust of the Kittilä greenstones.

Earlier an age of 2130±30 Ma was obtained for the Rantavaara intrusion and, based on its high ε_{Hf} value of +9.8, Patchett and Kouvo (1986) concluded that the rock originated from a depleted mantle source. This is supported by Nd-Sm data from a pyroxenite at base of the intrusion, which shows an ε_{Nd} value of +3 also compatible with a depleted mantle source (Huhma et al., 1996). Sample A900 is a very coarse-grained hornblende-rich gabbro pegmatoid, which yielded abundant zircon. The separated zircons are generally fragments of larger grains, exhibiting still sharp crystal edges and shiny surfaces. The transparency is generally moderate. In addition to the old data we have picked clear zircon for three new U-Pb analyses. All seven analyses provide a chord which gives an upper intercept age of 2148±11 Ma and a lower intercept of 652±190 Ma (MSWD=4, Fig. 7).

A1586 Metsävaara gabbro. Metsävaara is a hill about 2.5 km northeast from Rantavaara (Fig. 1.)and consists of the same intrusion, which here displays excellent magmatic layering rich in plagioclase. Sample A1586 was taken from a light-colored and coarse-grained gabbro. Separation yielded a small amount of clear zircon. A U-Pb analysis is slightly discordant and plots on the chord defined by sample A900 (Fig. 7).

The data presented here confirms the previous age results for the Rantavaara gabbro and the obtained U-Pb age of 2148±11 Ma can be regarded as the age of this extensive mafic intrusion. However, it is highly doubtful to argue that the intrusion penetrates the black schists at the base of the Savukoski Group.

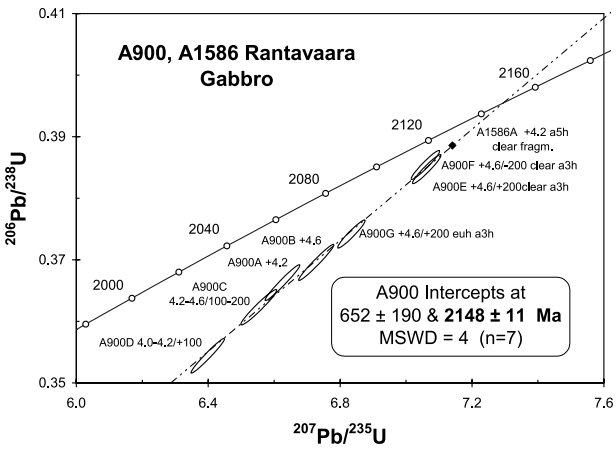

Fig. 7. Results for U-Pb analyses on zircons from the Rantavaara mafic intrusion.

A841 Pittiövaara albitite. Pittiövaara hill lies 10 km west of Sodankylä (Fig. 1). The bedrock is dominated by quartzites comparable to the Virttiövaara Formation in lithology. Arkose quartzites with mica schist interlayers exist but greenish sericite quartzites are dominating and they contain fuchsite-bearing layers. These quartzites are cut by a dyke, called "albite diabase" and represented by sample A841, for which an upper intercept of 2132±30 Ma has been reported by Tyrväinen (1983). The sample A841 shows a high ε_{Hf} value of +4.5 (Patchett et al. 1981).

The rock represented by sample A841 is a fine-grained carbonate-bearing albitite, composed mainly of albite and quartz with carbonate, magnetite, mica and other accessory minerals. Gabbroic rocks occur close to the place where the sample A841 was taken and more albitites are found about 8 km southwards. On reinvestigation, it has turned out that most albitites are associated with gabbroic rocks and it is evident that the gabbro is more extensive than previously thought.

Zircon in sample A841 occurs as turbid stubby prisms, which are quite typical for many "albite diabases". The analytical results for sample A841 are shown in Appendix 1 and presented on a concordia diagram in Figure 8. The six old analyses are fairly discordant and scatter slightly along the line, which gives intercepts at 362±190 and 2135±31 Ma (MSWD=13). The upper intercept provides the minimum age, but as the zircon is turbid and the data discordant and slightly scattered, the exact age of this sample remains unknown.

A962 Povivaara gabbro pegmatoid. The Povivaara mafic intrusion lies about 60 km northeast of Sodankylä, on the southern shore of the Lokka reservoir (Fig. 1). It is the middle part of a larger intrusion, up to several kilometers long, which has been brittle during folding and broken into pieces. It consists of two or three parts separated by schists and cut by mafic dykes. The investigated portion is composed of slightly differentiated gabbroic rocks, which include alternating plagioclase- and amphibole-rich magmatic layers, up to some meters thick, resembling the Rantavaara intrusion (A900 and A1586) discussed before. The intrusion has chilled margins and penetrates quartzites and mica schists. Close to the contact of the gabbro there exists a hornfels-like schist rich in biotite, which indicates that the schists are thermally affected by the intrusion.

Sample A962 was taken by A. Karvinen in order to find out whether or not the intrusion is a satellite of the 2.43 Ga old Koitelainen layered intrusion, which lies only a few kilometers westwards. The rock is a slightly scapolitized coarse-grained gabbro-pegmatoid with abundant zircon and rutile. The zircon is euhedral and long (L/B>5) prismatic crystals are common. Some zircons contain unidentified inclusions. Although only three fractions were analyzed, the upper intercept age of 2142±5 Ma seems quite reliable (MSWD=0.33). The lower intercept is at 438±48 Ma. A U-Pb concordia diagram is shown in Figure9 and the analytical results are given in Appendix I.

The age result the Povivaara intrusion shows clearly that it is not a satellite of the Koitelainen intrusion (2.43 Ga), but belongs to another magmatic phase which intruded about 300 Ma later.

A418 Koskenkangas gabbro. Koskenkangas is about 40 km east of Sodankylä in the vicinity of Tanhua. This intrusion is one of poorly exposed gabbro intrusions continuing as a chain for at least 10 km but reaching a thickness of no more than 400 m. Sample A418 is taken from a 2-3 km long intrusion consisting mainly of medium-grained and homogeneous gabbro showing only minor differentiation. However, some hornblende-plagioclase bands indicate primary magmatic layers. The gabbro is surrounded by quartzites

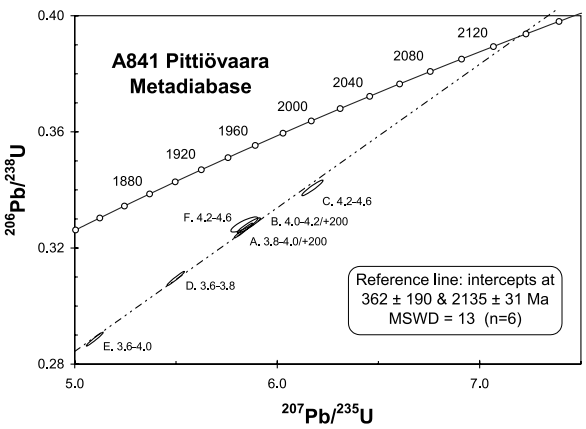

Fig. 8. Results for U-Pb analyses on zircons from the Pittiövaara metadiabase.

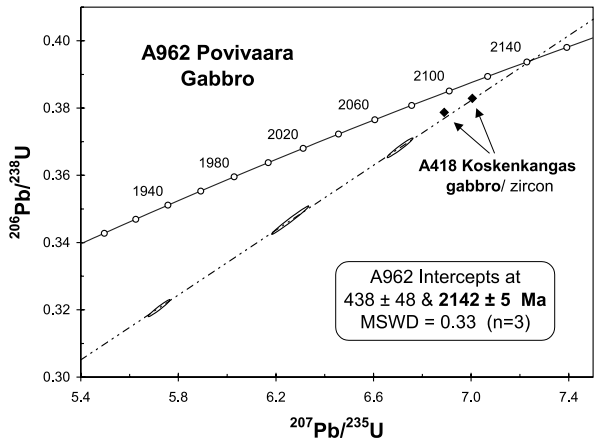

Fig. 9. Results for U-Pb analyses on zircons from the Povivaara and Koskenkangas gabbros.

and mica schists of the Sodankylä Group, but the contacts are covered.

The gabbro is quite even-grained and composed of hornblende and andesine as the main minerals,while zircon and rutile are common accessory minerals. As mentioned earlier, there are many similar gabbro intrusions in the Tanhua area and the sample was taken to determine the age group of these longish gabbros. Only two analyses are available and results of them are listed in Appendix I. A concordia diagram with plotted data is shown in Figure 9. The data plot close to the concordia and, as the less discordant heavy fraction (A) plots close to the chord defined by the previous sample A962, it is likely that all these gabbros are part of the same magmatic event as the intrusions of Povivaara (A962) and Rantavaara (A900+A1586) described before.

A816 and A817 Kylälampi gabbro. The Kylälampi intrusion occurs also in the Tanhua area about 45 km east of Sodankylä, close to the eastern border of the Sodankylä Schist area (Fig. 1). This highly magnetic intrusion is unexposed and was investigated by diamond drilling by Rautaruukki Co. in the 1970s. The intrusion faces to the west and consists of layered gabbroic rocks, which penetrate amphibolites and associated felsic volcanics. All of the intrusive rocks are amphibole- and albite-altered and contain abundantly magnetite. Two samples, A817 and A816, were taken by Rautaruukki Co.

Drilling did not reach the base of the intrusion, but the lowermost rock, more than 150 m thick in drill core, is a mafic biotite-bearing gabbro containing variable amounts of magnetite and quartz and carrying enclaves of amphibolite and volcanic rocks. The middle part, about 40 m thick in drill core, is more felsic and composed of an albite-magnetite rock with a sheared upper contact. The upper part, 17 m thick in drill core, consists of light-colored albitite with a varying amount of magnetite. The roof of the Kylälampi intrusion consists of volcanic rocks and, in drill core, they are penetrated by several albitite- and gabbro-dykes. Actually, this highly magnetic body could be the northernmost end of a more extensive gabbroic massif continuing over 10 km southwards and reaching a thickness about half a kilometer.

Sample A817 is taken from albite-magnetite rock, called albitite, at a depth between 78.70-93.00 m in the middle part of the intrusion. The rock is fine-grained, slightly orientated and dark gray in color consisting mainly of albite and magnetite with quartz. Biotite and hornblende are common accessory minerals and diopside, apatite and carbonate are also present together with titanite and leucoxene.The texture is granoblastic, but there are relics of saussuritized subhedral plagioclase crystals, up to 3 mm in size, indicating the primary texture. Zircon is transparent and exhibits needle-like crystal forms with well-developed crystal faces. Some platy and short prismatic, transparent crystals are present.

Data, listed in Appendix I, exist for six zircon fractions and titanite. As shown on the diagram in Figure10, a linear fit through the zircon analyses has an upper intercept at 2114±6 Ma. The least discordant points represent abraded fractions (F and G), but also an analysis of the HF-pretreated zircons (E) provides similar ratios. The nearly concordant titanite with a $^{207}Pb/^{206}Pb$ age of 1802±10 Ma, indicates a later metamorphic event.

Sample A816 is also from the Kylälampi intrusion and was taken from the same drill core as sample A817. The rock, taken from the upper part of the intrusion from the depth interval 50.00-64.00 m, represents a light gray, slightly reddish-colored albitite. In drill core its upper contact is characterized by a layer

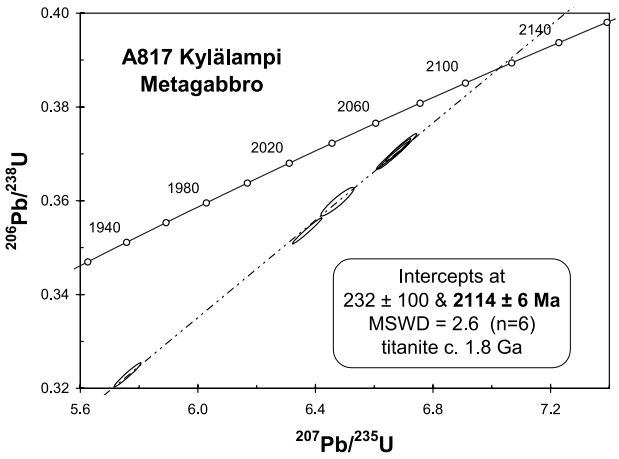

Fig. 10. Results for U-Pb analyses on zircons from the Kylälampi metagabbro A817.

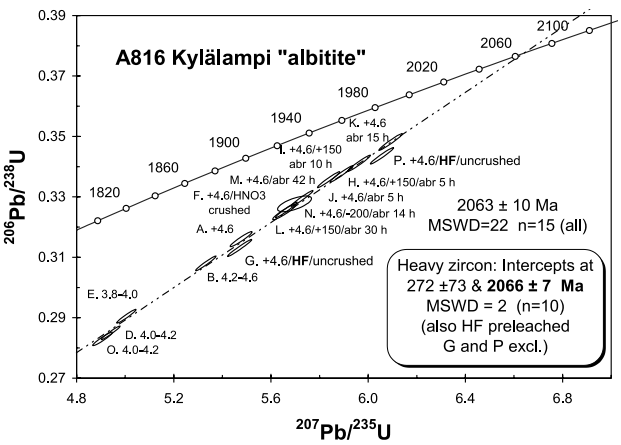

Fig. 11. Results for U-Pb analyses on zircons from the Kylälampi "albitite" A816.

of calc-silicate rock, while the lower contact with albite-magnetite rock is highly sheared. As said above, the roof of the intrusion consists of volcanic rocks penetrated by albite dykes and, on that ground, this rock could also be interpreted as a dyke. However, no proper chilled margins have been observed.

The rock is hypidiomorphic or weakly ophitic in texture. Albite, as 1-3 mm sub- and euhedral crystals and as large crystal-aggregates, and quartz are the main rock-forming minerals, while light green amphibole and magnetite are common accessory minerals with some titanite. Zircon crystals are euhedral and transparent, and do not seem to differ from the zircons in sample A817. However, their analyses give a different result.

Results of 15 analyses are given in Appendix I, and a concordia diagram with plotted data is presented in Figure11. The regression of all points provides a chord with intercepts at 2063±10 Ma and 193±97 Ma. However, the high MSWD (22) shows that the data are scattered. It is particularly obvious that the HF treated heavy fractions G and P plot off the line (Fig. 11). As there is a potential danger of U/Pb fractionation during HF-leaching, these analyses should be considered with caution.Rejecting these two and the analyses on less heavy (<4.2 g/cm^3) zircon, the intercepts would be at 2066±7 and 272±73 Ma (MSWD=2, n=10).

The conventional interpretation of the upper intercept age would suggest that in the Kylälampi drill core there exists intrusive material of two different ages. It should be noted, however, that all data are discordant and scatter exists as discussed above. An alternative interpretation would involve significant metamorphic effects on zircon at 1.8-1.9 Ga, in which case the primary magmatic age would be much older than the apparent upper intercept. In such case, however, one might expect even more scatter than observed.

A818 Ahvenselkä mafic dyke. Ahvenselkä lies about 30 km to the south of Savukoski (Fig. 1). The bedrock consists mainly of quartzites and mafic volcanic rocks, but felsic volcanic rocks occur as well (Räsänen 1983). However, the local bedrock is almost entirely covered and observations are based mainly on samples obtained by percussion drilling and on some diamond drill cores of Rautaruukki Co. drilled in the 1970s during their activity in the area.

Sample A818, called then an albite diabase, was also taken from a mafic dyke by Rautaruukki Co. in the 1970s. The outcrop is quite small but according to the aeromagnetic low-altitude map the dyke is several tens of meters thick and about 500 m long. It penetrates reddish quartzites, which are mainly arkosic in composition and contain calcareous and mica-bearing

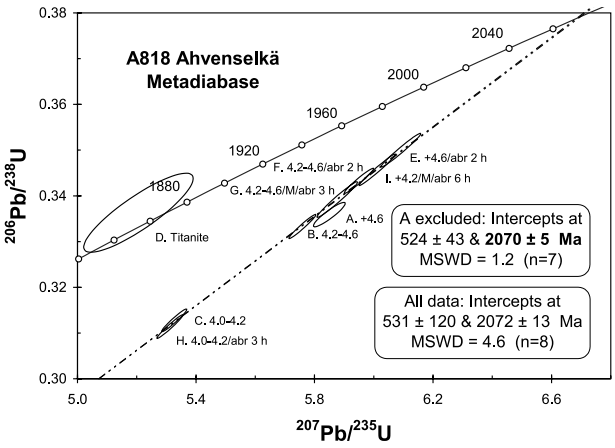

Fig. 12. Results for U-Pb analyses on zircons from the Ahvenselkä metadiabase.

interbeds. The quartzites are overlain by mafic volcanic rocks.

The dyke is medium-grained and massive in appearance. Hornblende and plagioclase are the main minerals, but the rock is altered. Plagioclase (An_{25-35}) displays albitized euhedral crystals with saussuritized cores rich in minute epidote and sericite. Hornblende is uralitic, slightly fibrous or flaky in appearance and contains relics of pyroxene. Quartz is also present and magnetite, titanite and scapolite are common accessories. Zircon is euhedral as prismatic crystals with well developed (100) and (101) crystal faces.

Results of zircon and titanite U-Pb analyses are listed in Appendix I. The $^{208}Pb/^{206}Pb$ ratios are relatively high (c. 0.6), which may indicate that Th and U were not fractionated in the melt. A regression of all 8 discordant zircon analyses gives intercepts at 531±120 and 2072±13 Ma (MSWD=4.6). Compared to other data, analysis A is of poor quality and without it the upper intercept age would be at 2070±5 Ma and the lower 524±43 Ma (MSWD=1.2, Fig. 12). The analysis for titanite is concordant but rather imprecise. Nevertheless, it indicates a later thermal event in Svecofennian times at 1836±30 Ma.

A820 Rovasvaara gabbro. Rovasvaara lies about 20 km southeast of Sodankylä (Fig. 1). Sample A820 was taken in the 1970s in connection with investigations made by Rautaruukki Co. The first age estimate (2060±2 Ma) was published by Patchett et al. (1981).

The Rovasvaara gabbro is part of an intensely folded mafic intrusion, tens of kilometers in length with a thickness varying from 100 m to almost 500 m. In the Oraniemi area it penetrates the rocks, which were previously known as the classical Kumpu-Oraniemi sedimentary series of Mikkola (1941). Later Saverikko (1977) stated that, in the Oraniemi area, the supracrustal succession is intensely folded and lies on gneissic granitoids. At the same time it was also observed that

the Kaarestunturi Formation, which is one of those typical 'Kumpu rocks', lies unconformably on highly folded rocks resembling the Oraniemi supracrustals. In addition, Räsänen (1977) found that mafic diabases do not penetrate this coarse clastic formation as argued by Hackman (1927). In contrast, the highly polymictic conglomerate at the base of the formation contains a lot of pebbles derived from mafic diabases and gabbros. These two university theses demonstrated that no connected Kumpu-Oraniemi sedimentary series exists. Based on two misinterpreted ages of volcanic rocks, Saverikko (1988) assumed the Oraniemi supracrustals to be Archean in origin. According to the current interpretation the Rovasvaara gabbro penetrates the Paleoproterozoic quartzites and aluminous mica schists of the Sodankylä Group.

The gabbro is mainly small- to medium-grained and quite homogeneous in appearance but, locally, it also contains more differentiated and coarse-grained, plagioclase-rich cumulates and exhibits ophitic texture. The most common rocks are hypersthene-gabbros, composed of pink pyroxene and plagioclase (An_{40-50}) with some hornblende and magnetite. Uralitized hornblende-rich parts contain biotite while epidote is a common product of altered plagioclase.

Sample A820, originally a pyroxene-plagioclase cumulate, is a very coarse-grained, pegmatoid variety of the gabbro. It is amphibolitized and epidotized and contains plenty of magnetite and accessory titanite with zircon. Initial $^{176}Hf/^{177}Hf$ ratio of the zircons shows an ε_{Hf} value of -3.2 (Patchett et al. 1981) indicating that the basaltic source material was crustally contaminated when compared to a depleted mantle source (Patchett & Kouvo 1986).

Since then, two additional fractions of zircons were analyzed at the GTK. The zircons are euhedral and exhibit splintery and platy crystal forms. Long needle-like crystals with L/B>10 are quite characteristic. Results of all analyses are listed in Appendix I and a concordia diagram is presented in Figure 13. A regression through 4 slightly discordant zircon fractions has an upper intercept at 2055±5 Ma and a lower intercept at 347±58 Ma (MSWD=0.02). An analysis on zircon with red pigment (E) is off the line, and omitted from the calculation.

The U-Pb age of this gabbro, 2055 Ma, is the same as that of the Keivitsa mafic intrusion, 2057±8 Ma (Mutanen & Huhma 2001, *this volume*) and also in good agreement with the U-Pb age of 2060±8 Ma from an albite gabbro at Riikonkoski in the Kittilä area (Rastas et al. 2001, *this volume*). Furthermore, the Sm-Nd data of the Keivitsa intrusion, which lies within the schists of the Matarakoski Formation at base of the Savukoski Group, yields an initial ε_{Nd} value of -3.5

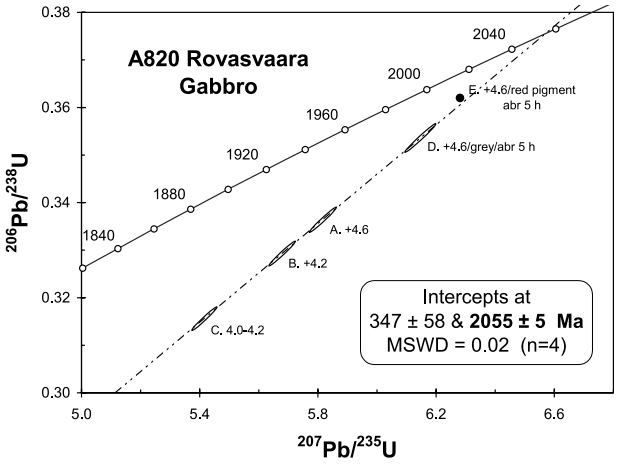

Fig. 13. Results for U-Pb analyses on zircons from the Rovasvaara metagabbro.

(Huhma et al. 1995), which shows that it is crustally contaminated as demonstrated by Mutanen (1997).

A861, A862 and A863 Törmänen mafic dyke. Törmänen is about 30 km southeast of Sodankylä by the river Kitinen (Fig. 1). The local bedrock is characterized by fine-grained but inhomogeneous sedimentary rocks, which are penetrated by mafic dykes. The sedimentary rocks, called the Akankoski Formation, include plenty of graded and commonly reddish, hematite-bearing siltstones associated with pinkish orthoquartzites together with bedded dolomites and laminated schists rich in calcite. These sedimentogenic carbonate rocks are highly enriched in ^{13}C, calcite rocks show positive $\delta^{13}C$ values between 11.1–12.5 ‰ and dolomites 10.1–10.8 ‰ (Karhu 1993). The Akankoski Formation can be compared lithologically with the Siltstone Formation in Kuusamo, which also is an inhomogeneous sedimentary package characterized by fine-grained, hematite-bearing sedimentary rocks together with quartzites, schists and sedimentary carbonate rocks (Silvennoinen 1972, 1991). According to Karhu (op.cit.) a dolomite sample of the Siltstone Formation also yielded a highly positive $\delta^{13}C$ value of 12.1 ‰ and a calcite rock from the Kelloselkä Formation in the Salla greenstone area, which can be compared with the former in stratigraphy, has also yielded a $\delta^{13}C$ value of +12.7 ‰.

The mafic dykes reach a thickness from some meters to tens of meters and exhibit chilled margins. However, all observed contacts are schistose or sheared indicating post-emplacement movements. Commonly the dykes are small- to medium-grained, exhibit ophitic texture and contain some layer-like, coarse-grained parts rich in plagioclase. All dykes are similar in texture as well as in mineral composition indicating that they are of the same generation.

Samples A861a, A862 and A863 were taken by P. Mielikäinen in 1970s. They represent coarse-grained plagioclase-rich parts of a single dyke, called then albite diabase, composed mainly of oligoclase and hornblende. Also some biotite and quartz are present. Magnetite, titanite and epidote occur frequently as accessory minerals. Sample A861b was taken afterwards from an other outcrop of the same dyke and does not significantly differ from the others. It also consists of hornblende and plagioclase (An_{20-30}) but contains hematite in addition to magnetite. Hematite is also present in zircons, which contain zones colored reddish by hematite dust. These zones coincide with original crystal forms, which are commonly short prismatic with sharp edges.

Results of all 22 analyses are listed in Appendix I and plotted on the concordia diagram in Figure14. Analyses on the original samples are all very discordant and scattered. Nevertheless, the five analyses on sample A863 give intercepts at 2061±47 and 518±130 Ma (MSWD=9). In contrast, sample A861b yielded less discordant analyses, and particularly the clear heavy zircon handpicked from this sample gives points fairly close to concordia curve (Fig. 14). The six analyses from A816b provide a trend with an upper intercept age of 2050±14 Ma and a lower intercept at 361±130 Ma. The large MSWD of 17 suggests, however, some scatter in excess of analytical error. If the most discordant analysis on the low-density turbid zircons is rejected the intercepts would be 2054±14 and 459±170 Ma (MSWD=8). Following a conventional interpretation this upper intercept age should be considered as the best estimate for the age of magmatic zircon and intrusion. The scatter observed particularly in sample A862 would suggest significant metamorphic effects probably during the 1.8-1.9 Ga tectonothermal stage, i.e. lead loss and possibly some recrystallization of zircon. If such domains existed in zircons which yielded the least discordant analyses (A861b), the magmatic age should be older than the apparent upper intercept.

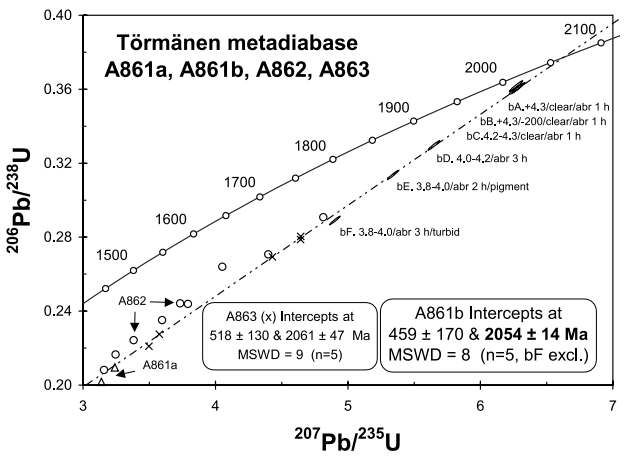

Fig. 14. Results for U-Pb analyses on zircons from the Törmänen metadiabase.

A401, A741, A742 and A743 Mairivaara mafic dyke. Mairivaara lies about 40 km southeast of Sodankylä, 10 km north of Pyhätunturi (Fig. 1). The bedrock consist of quartzites, which are capped by mafic tuffitic schist and penetrated by a mafic dyke, called albite diabase by Mielikäinen (1979). It was then assumed that the quartzites are the same as the quartzites at Pyhätunturi and comparable to the Kumpu quartzites, which both Hackman (1927) and Mikkola (1941) regarded as youngest sedimentary rocks in Finnish Lapland, lying probably unconformably on Jatulian rocks classified as Proterozoic. In order to get an age constraint for these quartzites three samples were taken from the mafic dyke. However, owing to the correlation of the Mairivaara and Pyhätunturi quartzites, the first results yielding an age c. 2.0 Ga for the dyke together with another age c. 2.2 Ga at Värttiövaara led to a confusion in lithostratigraphy as Silvennoinen et al. (1980) concluded that the Kumpu formation (presently the Kumpu Group) represents Jatulian rocks. The underlying supracrustal succession was interpreted as Archean and called Lapponian after the term 'Lapponium' (Eskola 1932). In fact, Hackman (1927) previously presented the same conclusion, supposing that as the Kumpu quartzites at Kaarestunturi seem to be penetrated by a mafic dyke, they belong to Jatulian rocks.

The Mairivaara quartzite is reddish and mainly arkosic in composition with thin orthoquartzitic interbeds and contains typically only a little micas. Crossbeds associated with some conglomeratic interbeds containing highly weathered gneissose granitoid pebbles are common in the lower part of the quartzite, but the main part is dominated by laminar beds with some cross-bedded interlayers. The uppermost part is amphibole-bearing and includes mica-rich siltstones with laminar and ripple cross-laminated interlayers, and changes to stratified amphibole-rich tuffite. Observed sedimentary structures are indicative of deposits formed in shallow water not far from shoreline, suggesting shelf-like prograde conditions during sedimentation followed by volcanism. Later investigations have shown that the Mairivaara quartzite (Mairivaara Formation) lies close the base of the Sodankylä Group.

The Pyhätunturi quartzite was also reinvestigated and described in detail by Räsänen and Mäkelä (1988). It lies on a thick pile of highly polymictic conglomerates. The bulk of the quartzites are homogeneous ortho- and sericite quartzites containing only a little feldspars. Well-preserved primary structures exhibit

alternating tabular beds from 10 cm to 160 cm in thickness, which include angular and tangential cross-bedded, horizontal-bedded and wavy-laminated sheets. Based on sedimentary structures this formation has been interpreted as a braided river deposit. Although no penetrating mafic dyke was found, Räsänen and Mäkelä (op. cit.) still assumed that the mafic lavas of the Kiimaselkä Formation overlie the Pyhätunturi Formation, which led to another confusion in general lithostratigraphy. Based on conglomerate pebbles and observations from sedimentary rocks, Laajoki (1988 p.103) explained the Pyhätunturi quartzites to be typically Jatulian rocks and as such correlative with "*the mature Jatuli-type quartzites of the third cycle*".

It is important to point out in this connection that in the 1970s there was no geophysical low-altitude map of the area and during 1970s and 1980s the role of tectonics such as faults and thrusts was almost entirely ignored in northern Finland as seen e.g. on the Geological Map of Northern Fennoscandia (1987). Later studies, based on lithological and geophysical data together with new field-observations, have suggested that both in north and east, the contact of the Pyhätunturi Formation is tectonic and the formation is separated by a fault, probably a thrust, from the mafic volcanics at Kiimaselkä and the quartzites at Mairivaara. Though the contacts of the Pyhätunturi Formation are covered, diamond drillings show that the bedrock including the Mairivaara quartzite was folded before the deposition of the Pyhätunturi Formation. Some results of these studies are also indicated on the Bedrock map of Finland (Korsman et al., 1997).

The mafic dyke is well exposed, 350-400 m thick and can be followed on outcrops for about 1 km in length. It is not differentiated, but exhibits alternating amphibole- and plagioclase-rich layers of magmatic origin. The margins of the dyke are chilled and glassy, elsewhere the rock is small- to medium-grained and ophitic in texture. Coarse-grained varieties are rare. The rock was originallypyroxene-bearing but is altered and now the main minerals are hornblende and plagioclase (An_{25-30}). The cores of plagioclase crystals are filled with minute epidote crystals. Secondary albite with subordinate quartz as well as scapolite are common. Titanite, often associated with biotite, displays large anhedral crystals or crystal-aggregates and magnetite is commonly present. Samples A741, A742 and A743 were taken P. Mielikäinen in 1970s during the first mapping activities in the area, while sample A401 was taken more recently. All samples are close to each other. Zircon occurs as small, euhedral, prismatic crystals and also some twinned crystals are found. The appearance of the zircon population is somewhat heterogeneous due to a range of colors and transparency from clear to very dark. Some baddeleyite occurs in the heavy fractions of samples A401 and A742.

There are 38 analyses of zircon fractions listed in Appendix I, and corresponding concordia diagrams are presented in Figures 15a and b. As a whole the analyses are discordant and do not provide well-defined chords on the concordia diagram. Nevertheless, excluding the two most discordant points, the data from these four closely spaced samples yield a regression line which has intercepts at 2029±7 Ma and 389±69 Ma (MSWD=38). The data for individual samples are also heterogeneous. An analysis on red opaque zircon (L) in sample A401 is discordant and clearly off the line defined by other eleven analyses, which give intercepts at 371±38 and 2038±5 Ma (MSWD=7, n=11). As some of the analyses represent very low-density material (<3.8 g/cm^3) and give highly discordant points, it should be appropriate to consider only good analyses on heavy zircon. Seven points yield a chord, which gives an upper intercept age of 2046±18 Ma, and a lower intercept of 530±260 Ma (MSWD=3, Fig. 15a). Most analyses on sample A741 represent

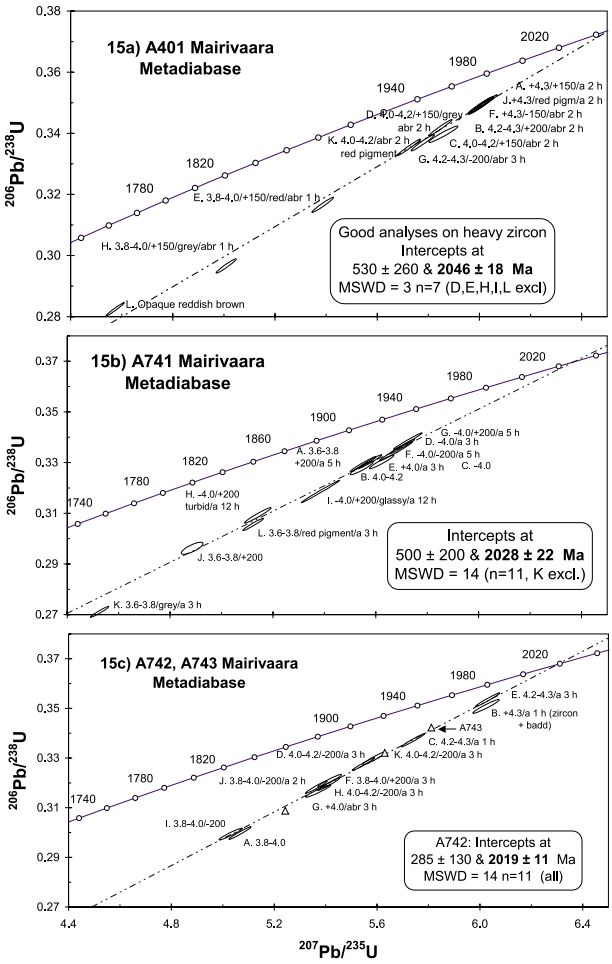

Fig. 15a) - c). Results for U-Pb analyses on zircons from the Mairivaara metadiabase.

low-density zircon. Apart of the most discordant data point (fraction K) the analyses from A741 form an ill-defined cluster which plots along a 2028-500 Ma reference line. The data from sample A742 are also heterogeneous plotting roughly along a 2019-285 Ma reference line. The three analyses on sample A743 plot close to this line.

The Mairivaara data are not unequivocal, but suggest intrusion not later than 2.03 Ga ago. The upper intercept age of 2046±18 Ma calculated for the best material of sample A401 may be considered with caution as an age estimate. However, as discussed above (Törmänen) metamorphic effects on zircon at 1.9-1.8 Ga are the likely reason for the heterogeneity observed. If such an effect was very pronounced, the age of intrusion should be older than the result of this simple calculation.

A815 Rykimäkuru mafic dyke. Rykimäkuru lies in the Luosto area about 40 km southeast from Sodankylä, 10 km west from Mairivaara and 12 km northwest from Pyhätunturi (Fig. 1.). The bedrock was mapped by Haimi as a student practice carried out for Rautaruukki Co. in the 1970s. The lithology consists principally of quartzites but there occur also schists and dolomites, as well as mafic dykes and mafic volcanic rocks, both flows and tuffs. Haimi (1977) considered that the quartzites were of same formation and, together with the quartzites at Mairivaara and Pyhätunturi, formed a syncline penetrated by mafic dykes and capped by the mafic volcanics. The idea for sample A815, taken 1977 by M. Haimi, was to produce an age constraint for the quartzite deposition.

Based on geological and geophysical information the Rykimäkuru dyke resembles that at Mairivaara, but is more deformed due to its proximity to a tectonized zone. It penetrates the fine-grained arkose quartzite of the Mairivaara Formation. It should be noted, that no dykes penetrating the Pyhätunturi Formation have been observed during careful reinvestigations in the area.

The sample A815 is ophitic and reddish in color, and rich in plagioclase containing subhedral crystals of albite up to 3 mm in size and abundant carbonate minerals. Hornblende and leucoxene after titanomagnetite are common and also minor quartz exists. Zircon occurs as short, reddish brown or pale, completely turbid crystals. The seven U-Pb analyses on zircon are very discordant and plot along a line which gives an upper intercept at 1912±18 Ma and an unusually low lower intercept at 22±49 Ma (Fig. 16). As the data are heterogeneous (MSWD=22) and very discordant, the significance of the age remains obscure.

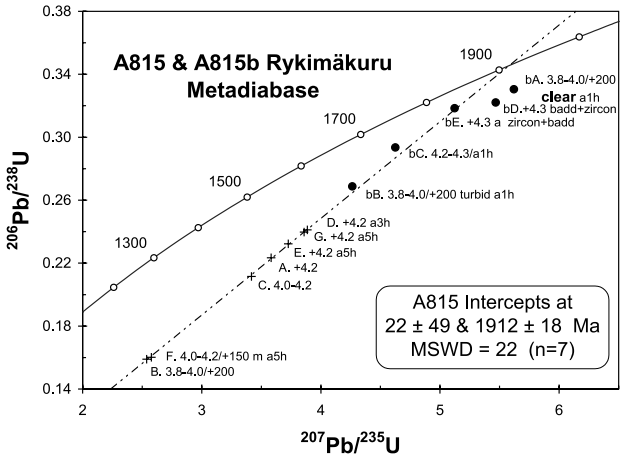

Fig. 16. Results for U-Pb analyses on zircons from the Rykimäkuru metadiabase.

In order to get better material for dating, a new sample (A815b) was taken from the Rykimäkuru mafic dyke. Sample A815b is ophitic in texture and mainly blackish gray in color containing some light-colored magmatic layers rich in plagioclase. Metamorphic hornblende containing some remnants of pyroxene and albitized plagioclase together with secondary biotite are the main minerals, while scapolite and leucoxene-altered titanomagnetite are common accessories. In addition to turbid zircon the heavy minerals separated contain transparent crystals of zircon and baddeleyite as well as zircon-baddeleyite aggregates. The five U-Pb analyses are less discordant than the data from the original sample, but do not provide any chord. Three analyses on turbid zircon are roughly on the line defined by sample A815, but an analysis on clear zircon (A) as well as the baddeleyite-dominated fraction (D) give clearly older age indications exceeding 2 Ga (Fig. 16). It should be noted that the clear zircons were picked from relatively low-density material (<4.0 g/cm^3).

There are cases where analyses on transparent zircons, baddeleyite or titanite have provided reliable age estimates for magmatic zircon, while turbid zircons in the same rock have yielded discordant data with relatively low $^{207}Pb/^{206}Pb$ ages. For example, the Jäkäläniemi mafic dyke in Kuusamo containing turbid and transparent zircons yielded originally a poor age estimate of about 2.0 Ga. Later, after baddeleyite and titanite were analyzed, their results together with transparent zircons yielded a reliable age of 2206±9 Ma as reported by Silvennoinen (1991). The same phenomenon is also observed at Susivaara, in the Peräpohja Schist belt (Perttunen & Vaasjoki 2001, *this volume*).

The results from the Rykimäkuru dyke do not allow a confident age estimate. The two analyses on the

clear zircon and baddeleyite plot on a line defined by the previous sample A401 from Mairivaara, which was considered geologically analogous. Natural interpretation for the data available is, again, 1) magmatic crystallization more than 2 Ga ago, 2) followed by significant disturbance of U-Pb in zircon at 1.8-1.9 Ga, and 3) finally more recent lead loss.

Felsic intrusions

A65 Kelujärvi granodiorite. Kelujärvi lies about 20 km to the east of Sodankylä (Fig. 1). The local bedrock is composed of quartzites and associated sericite and Al-bearing schists of the Sodankylä Group, which are penetrated by a gabbroic intrusion. Sample A65, taken in the 1980s, stems from the proximity of a gabbroic massif close to a contact with sedimentary rocks composed of fine-grained, graded biotite-sericite schists, rich in albite. The same granodioritic rock occurs with a gabbro close the place where the sample was taken.

Sample A65 is a medium-grained slightly reddish gray rock with faint orientation. It is patchy in appearance, due to amphibole aggregates. The texture is hypidiomorphic and the main minerals are plagioclase, quartz and hornblende with some potassium feldspar and biotite. The plagioclase (An_{20-30}) is weakly zoned and contains inclusions of quartz and microcline. Some thin sections show abundant sericitized potassium feldspar which contains inclusions of plagioclase and forms intergrowths with quartz. Hornblende is uralitic, bluish green in color and always associated with biotite. Abundant zircon occurs as small, short, euhedral and almost even-grained translucent crystals, exhibiting well developed (100) faces. Crystal edges are sharp and no rounded grains are found.

Results of the analyses are listed in Appendix I and plotted in Figure 17. Fractions D and E are preleached in HF. All analyses are very discordant and there is a positive correlation between increasing uranium content and increasing discordance, a feature shared by many igneous rocks. The data provide a chord which gives intercepts at 1891±5 and 106±12 Ma (MSWD=0.8).

The upper intercept age coincides with the ages of collision-related granitic to gabbroic intrusions of the Svecofennian orogeny, which are common in Western-Lapland and known as intrusive rocks of the Haaparanta Suite. All the dated intrusives yield U-Pb ages close to 1.89 Ga (e.g. Lehtonen 1984, Perttunen 1987, 1991 and Väänänen 1998). However, rocks belonging to the Haaparanta Suite lying as far in the east as A65 Kelujärvi have not been recognized before.

A570 Mustatseljät granite. Mustatseljät is a hill area on the eastern side of the schist area, about 50 km northeast from Sodankylä (Fig. 1). The bedrock consists of sulfide- and garnet-bearing, mainly banded amphibolites overlain by garnet-cordierite-sillimanite gneisses and quartzites. This volcano-sedimentary package, lying on Archean granitic gneisses with a brecciated tectonic contact, is intruded by a reddish granite with biotite streaks. Sample A 570 was taken from a granite dyke, about 2 m in thickness, in order to get some age constrains for the supracrustals. The rock is a foliated, small- to medium-grained granitic apophysis, which sharply penetrates the gneisses and amphibolites.

The rock is hypidiomorphic in texture and the main minerals are plagioclase, microcline and quartz, though

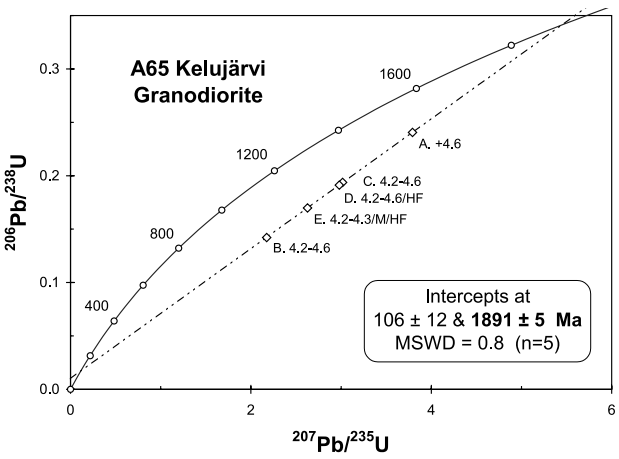

Fig. 17. Results for U-Pb analyses on zircons from the Kelujärvi granodiorite.

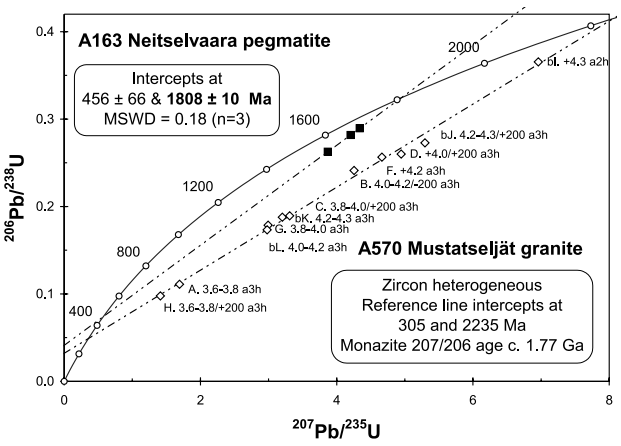

Fig. 18. Results for U-Pb analyses on zircons from the granite dykes.

microcline dominates some thin sections. Plagioclase (An_{25}) is rimmed with secondary albite and sometimes it includes a sericite-altered core exhibiting remnants of original, euhedral crystals, up to 2 mm in size, with weakly zoned textures. Biotite and muscovite together with monazite and apatite are common accessories and, in addition to zircon, also garnet, up to 2.5 mm in size, is frequent as an accessory mineral. Based on mineralogy the rock can be classified as an S-type granite.

The zircon population is somewhat heterogeneous in size and color. Turbid, red-brown grains are common, but also transparent crystals occur. The forms range from elongated euhedral simple prismatic to short anhedral. The U-Pb data are presented in the Appendix and Figure18. Most analyses on zircon are very discordant, particularly those made on very high-uranium, low-density material. The content of common lead is also relatively high. The regression provides intercepts at 2236±49 and 305±58 Ma, but the data are badly scattered (MSWD=475!). The U-Pb analyses on monazite give clearly younger age indications, the $^{207}Pb/^{206}Pb$ age being c. 1.77 Ga. This age registers either the intrusion of this granitic dyke or the metamorphism in the area. It is very likely that the upper intercept age above does not have any geological meaning due to inherited material. S-type granites like A570 may contain zircons derived from different sources including both magmatic and sedimentary rocks and, in addition, may have been mantled by a new generation of zircon afterwards. Lauerma (1982) reported a discordant "age" at 2216±70 Ma for the Hatajavaara granite (A126+A640+A641) in Salla and discussed the mixed zircon population and the origin of this inhomogeneous granite, which also contains supracrustal inclusions. Monazites in the Salla granites give U-Pb ages of c. 1.8 Ga and can be considered the date the crystallization of these granites (Lauerma 1982, Huhma 1986).

A163 Neitselvaara granite pegmatite. Neitselvaara lies in the Tanhua area about 45 km northeast from Sodankylä (Fig. 1.). The bedrock, mainly based on Mikkola (1937, 1941), is composed of weakly foliated and light-colored pinkish to grayish granite, which dominates the lithology over a large area. At Neitselvaara the granite is quite homogeneous, pinkish gray in color and weakly orientated, but it also contains inclusions rich in biotite and is penetrated by granitic dykes with sharp contacts.

Sample A163, taken in the early 1980s, represents a red-colored and coarse-grained pegmatitic granite dyke. The zircon population is relatively uniform consisting of elongated, euhedral, yellowish prisms. The three analyses shown in the Appendix and Figure18 give intercepts at 1808±10 and 456±66 Ma (MSWD=0.18). Following a conventional interpretation the upper intercept obtained should be considered as the age for the Neitselvaara granite pegmatite. This age of c. 1.8 Ga is well within the range obtained for late granites in Lapland. Isotopic studies indicate that the source of these granites originated from or at least involved Archean rocks (Patchett et al. 1981, Kouvo et al. 1983, Huhma 1986).

Detrital zircons in sedimentary rocks

A1021 Haikaraselkä quartzite. Haikaraselkä lies in the Vuojärvi district 40 km south of Sodankylä (Fig. 1.). The area is dominated by greenish sericite quartzites, called the Haikaraselkä quartzite, comparable to the Virttiövaara Formation as both of them are characterized by greenish fuchsite-bearing layers and exhibiting well-preserved ancient tidal deposits (Nikula 1985, Räsänen 1986b). The Haikaraselkä quartzite reaches a probable thickness of close to 1000 m and can be followed for tens of kilometers. It overlies the turbiditic mica gneisses associated with the Archean felsic volcanics (A194) at Keski-Loviselkä described before.

Typically, the quartzite is fine-grained and contains rounded to well-rounded quartz clasts in a sericite-rich matrix. A conspicuous feature is the bright green layers, owing to great abundance of fuchsite (1% Cr), which also reflects in the main chemistry of the quartzite (Cr>650ppm). In this association there exist brownish yellow laminae, up to 2-4 mm thick, which are rich in rutile and zircon. Sample A1021 was taken from these laminae which contain rounded detrital zircons.

Analytical results are given in Appendix I and shown in Figure19. The six analyses on detrital zircon are discordant and give $^{207}Pb/^{206}Pb$ ages from 2.70 to 2.75 Ga, which should be considered a minimum for the average zircon provenance. They do not form a precise chord, as the regression gives a MSWD of 32 and intercepts at 2856±180 and 1295±680 Ma. The data are comparable with the results from the Virttiövaara quartzite (Rastas et al. 2001, *this volume*), c.75 km NW, which exhibits the same sedimentologic features and also corresponds to the Haikaraselkä quartzite in stratigraphy. Similarly, ionprobe analyses on detrital zircons from quartzites in

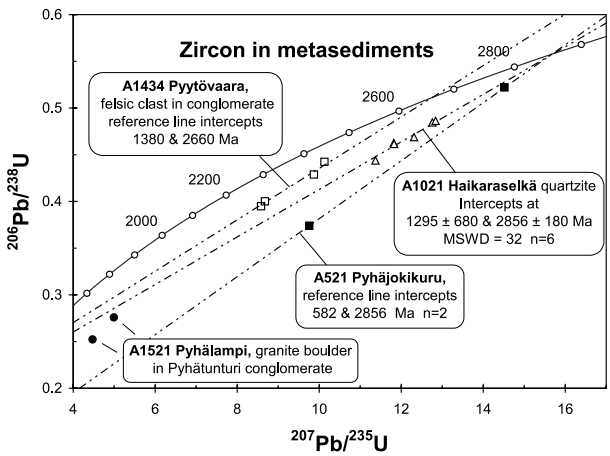

Fig. 19. Results for U-Pb analyses on zircons from the sedimentary rocks.

the Suomujärvi Complex, c. 70 km SE, are predominantly 2727±13 Ma old but also indicate another, older source (Evins 2001). So far no Proterozoic zircon population has been obtained from these quartzites which indicates that they derive from Archean rocks. However, this does not exclude the Paleoproterozoic deposition of sands.

A521 Pyhäjokikuru conglomerate. Pyhäjokikuru is in the Luosto area about 40 km southeast of Sodankylä and 10 km west from Mairivaara (Fig. 1). It is a river canyon lying in a fracture zone which cuts the bedrock and displays a lithostratigraphic section of it. The bedrock is dominated by reddish arkose quartzites belonging to the Mairivaara Formation of the Sodankylä Group and contain some conglomeratic and mica-rich interlayers. Several mafic dykes, often with albite-rich contacts, penetrate the quartzite which has been interpreted as part of the Pyhätunturi Formation (Haimi 1977, Mielikäinen 1979). The quartzite is conformably overlain by mafic volcanic rocks of the Kiimaselkä Formation displaying massive flows with amygdaloidal flow tops and tuffaceous beds. Based on lithology and lithological similarities Silvennoinen (1985) considered the Kiimaselkä volcanics part of the Greenstone Formation III, which represents a plateau-type basalt and is the best stratigraphic key horizon in the Kuusamo schist area (Silvennoinen 1972, 1991). On low-altitude geophysical maps it is characterized by magnetic highs and can be traced over very large areas in the Central Lapland Greenstone Belt. Based on similarities in lithological associations as well as geophysical and geochemical parameters, recent studies strongly suggest that this correlation is still valid and the volcanics of the Kiimaselkä and Greenstone Formation III can be compared to each other in stratigraphy.

Sample A521 was taken from the Mairivaara Formation and it represents a slightly conglomeratic interlayer of the quartzite. The rock is albitized and mainly granoblastic in texture exhibiting only weak clastic textures, however, it contains plenty of well-rounded zircons of undoubted detrital origin. Furthermore, there are two kinds of zircons, some having red pigment, while others are gray in color and do not show any pigment at all.

Only two analyses of zircons with red pigment were made, and both indicate an Archean age. The $^{207}Pb/^{206}Pb$ age of 2.84 Ga for the less discordant analysis suggests a relatively old average age for the provenance. This apparent age is quite close to the U-Pb age of 2832±10 Ma of a uniform detrital zircon population of the quartzite at Virttiövaara (Rastas et al. 2001, *this volume*) and gives more evidence for the existence of Archean rocks older than 2.8 Ga, which have been exposed and denuded while the Paleoproterozoic sands were deposited.

A1434 Pyytövaara volcanic conglomerate. Pyytövaara lies about 15 km southwest of Savukoski (Fig. 1). The local bedrock consists of mafic and felsic volcanics with a BIF, which are overlain by volcanic conglomerates and komatiites (Räsänen 1983). The volcanics were formerly known as the Jauratsi greenstone belt and, as age determinations were lacking, regarded as Archean in age (Gaál et al. 1978 and Silvennoinen et al. 1980). Later Saverikko (1983) observed that the komatiites are the uppermost volcanic rocks in the area but afterwards still considered them of an Archean age (Saverikko 1987).

During the mapping by GTK in the 1980s many samples from felsic volcanic rocks were taken to get some age constraints for the overlying komatiites, called today the Kummitsoiva Formation, but no zircon was found. Sample A1434 was taken later and it is a light-colored, fine-grained felsic boulder (about 1 m^3 in size) at the base of a volcanic conglomerate, which also contains phenoclasts of a banded iron formation.

The felsic boulder consists mainly of quartz and plagioclase (An_{20-30}) but also contains abundant tremolite and some microcline. Almost euhedral plagioclase crystals, up to 0.4 mm in size, exist but there are also roundish quartz crystals, up to 0.5 mm in size, indicating that the rock is reworked. Zircon and apatite are common accessories. The zircons are transparent and reddish in color. However, all zircon crystals seem to be slightly rounded and some of them are clearly rounded and evidently detrital in origin. This conflicts with the original attempt to constrain the age of deposition of these supracrustal rocks by dating a felsic igneous clast.

The four analyses listed in the Appendix are technically very good, but provide discordant U-Pb results

(Fig. 19). The $^{207}Pb/^{206}Pb$ ages range from 2.43 to 2.53 Ga and represent the minimum age for the analyzed material. These results do not help much in constraining the age of the Kummitsoiva Formation. However, it is known today, that the Kummitsoiva komatiites are correlative to the komatiites of the Sattasvaara Formation, in Sodankylä, and on well proven lithostratigraphic grounds can be regarded as Proterozoic in age (cf. Räsänen 1990, 1996 and Lehtonen et al. 1998). Recently Hanski et al. (2001) have obtained a Sm-Nd age of 2063±34 Ma for the Jeesiörova komatiites at Kittilä. This is an extensive komatiite-picrite association which can be traced for hundreds of km through the Finnish Lapland to Norway, where Krill et al. (1985) have reported a Sm-Nd age at 2085±85 Ma for the Karasjoki komatiites.

A1521 Pyhälampi polymictic conglomerate. Pyhälampi lies about 40 km southeast from Sodankylä on the western side of Luostotunturi (Fig. 1). The local bedrock consists of conglomerates and quartzites of the Pyhätunturi Formation, which belong to the Kumpu Group. Hematite- and carbonate-bearing fine-grained schists together with graded siltstones and pinkish quartzites of the Harjunoja Formation and mafic volcanics of the Kiimaselkä Formation belong to the Sodankylä Group. Diamond drilling has shown that the mafic volcanics of the Kiimaselkä Formation overlie the supracrustals of the Harjunoja Formation, which also includes some felsic volcanic rocks but is comparable in lithology and in stratigraphy with the Akankoski Formation on the eastern side of the Pyhätunturi Formation. The Pyhälampi conglomerates representing the coarse-clastic molasse-type deposits are the same as described by Räsänen and Mäkelä (1988) at base of the Pyhätunturi Formation and called then the Isokuru Formation. At Pyhälampi this highly polymictic conglomerate exhibits meters of thick alternating sandy and pebbly layers, some of them containing abundant boulders and cobbles of homogeneous red granite with bluish gray quartz. Because similar granites occur nearby at Vaiskonselkä in the Central Lapland Granite Complex and are generally regarded as post-orogenic, c. 1.8 Ga old (see Korsman et al. 1997), a sample was taken to get maximum age constraints for the deposition of the coarse-clastic sediments.

Sample A1521 was a well-rounded boulder, about 35 cm in diameter taken from a layer containing abundantly pebbles and cobbles of similar granite. The rock is a medium-grained and unorientated red granite. Its texture is hypidiomorphic and the main minerals are microcline, plagioclase and quartz. Both feldspars are turbid and reddish in color due to abundant pigment of dust-like hematite, and also contain hematite-filled fissures. Quartz is transparent and exhibits almost euhedral crystals, 2-4 mm in size. It contains commonly abundant linearly aligned minute inclusions of rutile, which probably cause the bluish color of quartz. The most common accessory mineral is hematite, which exists as platy crystals up to 3 mm in size. The amount of magnetite and biotite is insignificant.

The sample yielded a small amount of zircon, which is relatively large in size and dark red-brown due to pigment. Crystals are generally euhedral exhibiting often sharp edges. The two U-Pb analyses are very discordant providing $^{207}Pb/^{206}Pb$ ages of c. 2.1 Ga (Fig. 19). This should be considered as a minimum age for the material analyzed, but the age of this granite pebble remains obscure. The results obtained do not give essential information on the age of sedimentation, but on sedimentogenic grounds the Pyhätunturi Formation can be compared to similar deposits of the Kumpu Group also characterized by molasse-like, several hundreds of meters thick coarse-clastic deposits. They all are well-preserved, and typically longitudinal in shape but areally very restricted in width. Recently a U-Pb age of 1888±16 Ma has been obtained for a felsic pebble picked from conglomerates of the Kumpu Group in Kittilä (Hanski et al., 2000, Rastas et al. 2001, *this volume*). The age supports their deposition after collision-related intrusion of the Haaparanta Suite during the Svecofennian orogeny, which indicates that they are comparable to the molasse sediments proper in a tectonic sense.

Other samples

Attempts to determine the geochronology of Central Lapland have involved the collection of many samples during the last decades. Some of them have yielded no suitable material for dating, and are listed below (Table 1).

A1433 Visakuppura. At the base of the Sattasvaara Formation at Visakuppura, about 30 km NNW of Sodankylä (Fig. 1), there exist fragments and angular boulders of fine-grained felsic rocks of probably volcanic origin with highly corroded contacts because they are affected by the komatiite. Sample A1433 was taken from a felsic boulder to get some age constrains for the Sattasvaara Formation but separation yielded no zircon.

A1435 Möykkelmä. The Möykkelmä Formation about 35 km NNW of Sodankylä (Fig. 1) lies on

Table 1. Samples with insufficient material for U-Pb dating from the Sodankylä area.

Sample	Map	Northing	Easting	Locality	Rock type
A1433	371405	7507.04	3470.38	Visakuppura	Andesite (ejecta in komatiite)
A1435	371406	7510.50	3472.10	Möykkelmä	Andesite (basaltic)
A1476	362412	7456.40	3491.25	Keski-Luosto	Dacite (felsic volcanics)
A1522	362409	7458.90	3487.30	Vaiskonselkä	Granite
A1573	362411	7448.92	3493.32	Luosto	Arkosic quartzite
A1575	364201	7433.32	3504.54	Sienioja	Dacite (felsic volcanics)

Archean granitoids and consists of two ultramafic and three mafic volcanic units of unknown age. Sample A1435 was taken from the basaltic andesite member representing the middle mafic unit but no zircon was recovered.

A1476 Keski-Luosto. The Harjunoja Formation, penetrated by 2.2 Ga old differentiated sills (A281 Harjunoja, described above), consists of sedimentary and volcanic rocks. The supracrustals are poorly exposed and a part of them has been intersected by diamond drilling made by Rautaruukki Co. in 1970s. The drill cores have been reinvestigated and sample A1476, taken recently from these drill cores, represents a plagioclase-phyric felsic flow, andesite in chemical composition, associated with felsic tuffs and breccias. No zircon sufficient for conventional U-Pb analyses was recovered.

A1522 Vaiskonselkä. Vaiskonselkä is 25 km SSE from Sodankylä (Fig. 1). The area consists of medium grained red granites characterized by blue quartz and closely resembling boulders and pebbles found in conglomerates at base of the Pyhätunturi Formation (A1521 Pyhälampi, described above). Sample A1522 was taken from the granite and it contains zircon and a lot of rutile, but as the zircon population is obviously heterogeneous conventional analyses have been not done.

A1573 Luosto. Sample A1573 taken from the Luosto area about 30 km SSW of Sodankylä (Fig. 1) represents a quartzite layer of the Pyhätunturi Formation. A small amount of zircon was found among abundant rutile, but no conventional analyses have been made.

A1575 Sienioja. Sienioja lies roughly 50 km SSE from Sodankylä on the southern side of Pyhätunturi (Fig. 1). Sample A1575 was taken from a drill core made by GTK some years ago. It is a very fine-grained felsic volcanic rock, chemically a andesite, associated with fragmental and graded tuffites resembling the sample A1476 Keski-Luosto. Only a few zircon grains were recovered.

Jauratsi is about 20 km southwest of Savukoski (Fig. 1). In this area, there exist mafic and felsic volcanic rocks associated with a banded iron formation. They lie on quartzites and the whole sequence is overlain by komatiites of the Kummitsoiva Formation comparable to the Sattasvaara Formation. Three samples were taken from felsic volcanic rocks during regional mapping in the 1980s, but no zircon was found in these amygdaloidal lavas and breccias.

DISCUSSION

Dating problems

The most reliable U-Pb dating often involves heavy, clear zircon, and nearly concordant analyses. In this study such material and data are not generally available, and the U-Pb dating always involves interpretation of the analytical results. As the data are often very discordant and heterogeneous, the drawing of reliable conclusions becomes difficult if not impossible. Occasional analytical problems are conceivable, as the old analyses have been made using separate Pb and U spike solutions. However, several duplicated analyses in the laboratory have yielded consistent results, and the principal source for heterogeneity should be in zircon. Possible reasons for scatter in multigrain conventional data include primarily heterogeneous zircon populations (detrital or inherited zircon) and metamorphic effects.

It is known that zircon will stay closed at high temperatures (e.g. 800°C), but may lose Pb if it experienced extended time periods at low temperatures especially when abundant fluids were available (Mezger & Krogstad 1997). Apart of clearly detrital or inherited material encountered in few samples, we consider this as the principal cause for the heterogeneity observed in this study as (1) magmatic crystallization occurred before 2 Ga and was (2) followed by disturbance of U-Pb system at 1.8-1.9 Ga, which may have involved dissolution and recrystallization of metamict zircon. Finally, (3) more recent lead loss may

have occurred.

A sample subjected to this sequence should generally provide very scattered data (A862, Fig.14), but occasionally good-looking chords with geologically meaningless upper intercept ages are formed (A815, Fig.16). Generally, such chords exhibit elevated MSWD-values and apparent upper intercept ages too young for magmatic crystallization in the particular geological environment. This may be the case with the metarhyolites from the Purkkivaara formation (A992 and A1524), as well as gabbroic rocks from Pittiövaara (A841), Mairivaara (A741-A734) and Luosto (A815). Typically zircons in these gabbroic rocks are very turbid and of low-density. Examples show that transparent zircon, baddeleyite or even titanite have provided more reliable age estimates, while turbid zircons in the same rock have yielded discordant data with relatively low $^{207}Pb/^{206}Pb$ ages (Silvennoinen, 1991, Lauerma, 1995; Perttunen & Vaasjoki 2001, *this volume*; sample A861b).

Apart from zircon dating, there are occasionally problems in determining the primary rock types in highly metamorphosed and tectonized lithologies, i.e. whether a sample in question is truly magmatic with magmatic zircon or contains a detrital component.

Geological implications

The results of the isotopic analyses presented previously are summarized in Table 2.

The extent of Archean rocks

In the northeast, the Sodankylä schist area is bounded by an Archean terrain, which – excluding the gneisses of Tuntsa Suite northeast of Savukoski - is thought to consist mainly of orthogneisses. However, during this study both ortho- and paragneisses have been observed. Similar gneisses are also found on the southern side of the area, and at Keski-Loviselkä they occur with a felsic volcanic rock yielding an age of c. 2.77 Ga. In addition, associated turbiditic greywackes strongly suggest the contemporaneity of sedimentation and volcanism at that time, which is well-known in the Canadian Shield (e.g. Ojakangas 1985) and also reported from the Fennoscandian Shield (Rundquist & Mitrofanov 1993, Vaasjoki et al. 1993). These gneisses can be followed northwestwards into the Central Lapland Granite Complex and, Archean ages of c. 2.72 - 2.74 Ga are obtained from felsic volcanic rocks at Honkavaara (Rastas et al. 2001, *this volume*), about 35 km W of Sodankylä.Likewise the gneisses can be traced far southeastwards into the Archean Suomujärvi Complex, where they are highly granitized but still contain porphyroblast-bearing parts, traceable for tens of km southwards into the Posio area. At Niskavaara, Posio, on the western side of the Kuusamo Schist area, a felsic volcanic rock associated with turbiditic gneisses has a U-Pb zircon age of 2800±8 Ma, and Räsänen and Vaasjoki, (2001, *this volume*) argue that all these rocks belong to the same volcanic-sedimentary unit.

These findings suggest the presence of an extensive late Archean volcanic-sedimentary belt on the southeastern side of the Central Lapland Greenstone Belt and continuing within the Central Lapland Granite Complex. The gneisses are magnetic and, though granitized, can be seen as banded magnetic anomalies forming images of large scale fold structures on low-altitude geophysical maps. This gives reason to presume that the Central Lapland Granite Complex includes much more Archean rocks and their granitized relicts than previously thought. So far this huge granitoid terrain, about 20 000 km^2, is poorly known, but it seems likely that the Suomujärvi Complex, separated from the rest of the Central Lapland Granite Complex by a fault, actually comprises mostly Archean rocks. The basement consists of 2.82 Ga old orthogneisses, which are penetrated by 2.75 Ga granitic rocks and also includes paragneisses and quartzites containing zircons older than 2.72 Ga (Evins 2001). Felsic volcanics are yet to be recognized as the rocks were highly granitized and partly remobilized during the Svecofennian orogeny. Migmatites and penetrating granites yield heterogeneous and discordant zircon data with $^{207}Pb/^{206}Pb$ ages from 1.77 to 2.22 Ga (Lauerma 1982) comparable with the granitoids in the Tanhua area (A163, A570), on the northern side of the Central Lapland Greenstone Belt, indicating their similar geological history.

The age of the komatiites at Kivivuotsonselkä can not be compared to that of the komatiites of the Archean Kuhmo-Suomussalmi belt, which are penetrated by intrusive rocks dated at 2.74 Ga (Hyppönen 1983, Vaasjoki et al. 1999). Likewise, the ultramafic volcanic rocks of the Tulppio Suite in the Archean terrain north of Savukoski seem to be older (Juopperi 1994, Juopperi & Vaasjoki 2001, *this volume*). The Paleoproterozoic komatiites of the Vetreny Belt, in Russia, yield Sm-Nd ages of 2.44 Ga and 2.41 Ga and an associated felsic volcanic rock yields a U-Pb age at 2.43 Ga (Puchtel et al. 1996, 1997). However, the

Table 2. Summary of results of U-Pb mineral analyses from rocks of the Sodankylä schist area.

Sample	Map	Northing	Easting	Locality	Rock type	Age (Ma)	N	MSWD	Comment
A194	362407	7438.20	3489.10	Keski-Loviselkä	Felsic volcanic rock	2775±25	4	16	
A195	362407	7438.20	3489.10	Keski-Loviselkä	Felsic volcanic rock	2780±72	4	9	Titanite >1825 Ma
A283	373302	7477.88	3541.40	Kivivuotsonselkä	Komatiite	-	7	-	Archean, heterogeneous, inherited?
A992	364405	7448.85	3556.23	Purkkivaara	Felsic volcanic rock	2412±40	4	24	
A1524	364406	7450.50	3551.50	Akanvaara	Quartz-feldspar porphyry	2398±25	8	55	two deviating analyses
A1432	374110	7526.80	3533.60	Sakiamaa	Rhyolite	2438±11	4	2.7	
A1498	374110	7526.80	3533.60	Sakiamaa	Rhyolite	-	2	-	compatible with A1432
A281	362410	7434.70	3497.75	Harjunoja	Gabbro	2222±6	12	7	
A900	373201	7499.53	3502.61	Rantavaara	Gabbro	2148±11	7	4	E-Hf=+9.8
A1586	373202	7500.40	3504.50	Metsävaara	Gabbro	-	1	-	compatible with A900
A841	371306	7483.21	3473.71	Pittiövaara	Albitite	2135±31	6	13	discordant, suspect; E-Hf=+4.5
A962	374107	7524.67	3525.30	Povivaara	Gabbro pegmatoid	2142±5	3	0.3	
A418	373109	7487.16	3526.08	Koskenkangas	Gabbro	-	2	-	compatible with A962
A816	373208	7501.19	3526.98	Kylälampi	Albitite	2066±7	15	2.0	10 anal. on heavy zircon; Tit 1890±9 Ma
A817	373208	7501.19	3526.98	Kylälampi	Metagabbro	2114±6	6	2.6	Titanite 1801±13
A818	364401	7435.85	3548.25	Ahvenselkä	Metadiabase	2070±5	8	1.2	1 fraction excluded; titanite c. 1.84 Ga
A820	373102	7471.25	3501.90	Rovasvaara	Gabbro	2055±5	4	0.02	1 fraction excluded; E-Hf=-3.2
A861	362412	7456.16	3496.98	Törmänen	Metadiabase	2054±14	6	8	turbid crystals discordant, excluded
A862	362412	7456.14	3497.00	Törmänen	Metadiabase	-	9	-	heterogeneous, metamorphic?
A863	362412	7456.10	3497.06	Törmänen	Metadiabase	2061±47	5	9	discordant data
A401	364202	7445.86	3506.66	Mairivaara	Metadiabase	2046±18	12	3	7 good analyses on heavy zircon
A741	364202	7445.86	3506.56	Mairivaara	Metadiabase	2028±22	12	14	heterogeneous, 1 fraction excluded
A742	364202	7445.96	3506.66	Mairivaara	Metadiabase	2033±20	11	8	heterogeneous, 3 fractions excluded
A815	362411	7445.62	3499.23	Rykimäkuru	Metadiabase	-	12	-	heterogeneous, baddeleyite >2032 Ma
A65	373103	7486.04	3504.84	Kelujärvi	Granodiorite	1891±5	5	0.8	
A163	373208	7509.50	3520.40	Neitselvaara	Granite pegmatite	1808±10	3	0.2	
A570	373208	7502.90	3529.80	Mustatseljät	Granite	-	11	-	heterog. Inherited?, monazite >1.77 Ga
A521	364202	7445.08	3500.12	Pyhäjokikuru	Quartzite	-	2	-	Archean, apparently >2.8 Ga
A1021	362410	7435.55	3495.25	Haikaraselkä	Quartzite	-	6	-	Archean, apparently >2.8 Ga
A1434	364403	7454.48	3542.74	Pyytövaara	Felsic clast in conglomerate	-	4	-	discordant, apparently >2.5 Ga
A1521	362411	7445.20	3498.12	Pyhälampi	Granite clast in conglomerate	-	2	-	discordant, >2.1 Ga

N = number of analyses
MSWD = Mean square of weighted deviates

REE-patterns of the Kivivuotsonselkä spinifex-textured komatiites show no signs of LREE-enrichment characteristic of the crustally contaminated rocks of the Vetreny Belt. Likewise, the Kivivuotsonselkä rocks differ chemically from the Möykkelmä komatiites (Räsänen et al. 1989, Manninen 1991). Also a comparison to the Sattasvaara-type Kummitsoiva komatiites is unwarranted as the komatiite flows at Kivivuotsonselkä lie on Archean banded gneisses and are probably overlain by a pile of supracrustals which form the base of the Kummitsoiva komatiites. As the rocks do not contain any primary magmatic minerals, successful Sm-Nd dating is unlikely, and the true age of these komatiites remains unknown.

Felsic volcanism and layered mafic intrusions

An important finding concerning the Proterozoic supracrustal rocks within the Sodankylä Schist area is that the lowermost felsic flows at Sakiamaa yield an age of 2.43 Ga, comparable to ages obtained for the Purkkivaara and Akanvaara felsic volcanics. These are convincing evidence for volcanism at the start of the rifting of the Archean continent at the onset of the Paleoproterozoic coinciding with the emplacement of layered mafic intrusions at 2.43-2.44 Ga. Although volcanic rocks of this age have been previously identified elsewhere in the Fennoscandian Shield (Kratz et al. 1976, Gorbunov et al. 1985, Amelin et al. 1995, Puchtel et al. 1996, 1997), such volcanism has not been conclusively demonstrated in northern Finland before, even if several plausible candidates have been identified. These studies (see also Manninen et al. 2001, *this volume*) indicate that the volcanic rocks forming part of the 2.44 Ga igneous episode occur throughout the Fennoscandian Shield.

Periods of mafic magmatism

In spite of the dating problems discussed above, several samples from mafic intrusions have provided good results sufficient for constraining the age of mafic magmatism and associated supracrustals in the Sodankylä schist area. The data obtained show clearly that, in addition to the 2.44 Ga old Koitelainen and Akanvaara intrusions, there exist several younger generations of mafic magmatism, with U-Pb ages of 2222±6 Ma, 2148±11 Ma, 2114±6 Ma, 2070±5 Ma and 2055±5 Ma. In addition to these, the 2058±4 Ma Keivitsa intrusion is cut by komatiite dykes.

The differentiated sill-like mafic intrusions, dated to be 2222±6 Ma old at Harjunoja, penetrate quartzites with associated dolomites and schists of the Harjunoja Formation, which lies at the base of the Sodankylä Group and will be compared here to the Jatuli-type rocks in eastern Finland. The intrusions are older than the volcanics of the Kiimaselkä Formation as they do not penetrate those. On the basis of geochemistry they can not be feeder channels for the Kiimaselkä volcanic rocks. In addition to the Sodankylä schist area, similar intrusions occur also at Salla, where U-Pb ages of c. 2.21 Ga are observed both at Pirtinkehäkukkura and at Jokinenä (Lauerma 1995). The Jäkäläniemi mafic dyke in Kuusamo also yields an age of c. 2.21 Ga (Silvennoinen 1991) and in the Peräpohja schist area there exist several differentiated mafic intrusions, which belong to this age group (Perttunen & Vaasjoki 2001, *this volume*). These intrusions concentrate on the borders of Proterozoic volcano-sedimentary belts and penetrate their lowermost supracrustal rocks forming a belt, which can be followed on outcrops and as magnetic highs for hundreds of kilometers. Evidently these rocks are genetically associated with a rifting phase of the Archean craton, which occurred 2.22-2.20 Ga ago and gave rise to these enormous magmas. Their initial Nd-ratios are close to chondritic values (Huhma et al, 1990, 1996). Volcanic counterparts for them have not so far been recognized in northern Finland.

The mafic intrusions yielding U-Pb ages between 2.15 and 2.11 Ga are mainly gabbroic in composition and commonly slightly differentiated but include also albitites and albite-magnetite rocks. These intrusions, often characterized by alternating plagioclase- and amphibole-rich magmatic layers, are common in the study area. The Hf and Nd isotopic data (Patchett et al. 1981) of the Rantavaara intrusion indicate that it stems from a depleted mantle source, but no isotopic source rock studies have been done on the other intrusives of this group. Most probably the intrusives have volcanic counterparts in the area, but the relationships between the various volcanic rocks and sets of dykes are still obscure. In any case, these intrusions penetrate the graded Al-rich mica schists on top of the Sodankylä Group, but so far there is no evidence that they would penetrate the black schists at the base of the Savukoski Group.

The third group of mafic intrusions consists of gabbros and mafic dykes, which yield ages in the 2.07-2.05 Ma range, and penetrate the Matarakoski Formation of the Savukoski Group. Most of the intrusions are almost undifferentiated dykes with ophitic texture but they also include gabbroic rocks, which exhibit magmatic layering and contain both pyroxene- and albite-bearing derivatives. It has been confirmed recently that this magmatic phase also includes large ore-bearing layered intrusions like the 2.06 Ga old Keivitsa

mafic-ultramafic intrusion (Mutanen 1997, Mutanen & Huhma 2001, *this volume*), which lies in the northern part of the Sodankylä Schist area. Both Hf and Nd isotopic data (Patchett et al. 1981; Huhma et al. 1995, 1996), indicate that the source material of these intrusions was crustally contaminated when compared to a depleted mantle source. Recently Hanski et al. (2001) have obtained a Sm-Nd mineral age of 2.06±0.04 Ga for the komatiites in Kittilä and demonstrated that also the extensive Sattasvaara-type komatiites belong to this magmatic episode. There is no reason to be suspicious of that age result, especially as field observations indisputably show that several basaltic and peridotitic komatiite dykes also penetrate the Keivitsa intrusion. Mutanen (1997) called them olivine diabases and based on later investigations they are chemically similar to the komatiites of the Sattasvaara Formation and will be described in another paper. The dating of these dykes is yet to be completed.

Younger igneous rocks

With regard to the younger intrusive rocks in the study area, it is worth to note that the U-Pb zircon age of 1891±5 Ma from the granodiorite at Kelujärvi represents the age of the Haaparanta Suite of intrusions. These relatively homogenous rocks have been found previously in northern Finland only in western Lapland, but according to this study, the occurrence of the Haaparanta Suite rocks is areally not so restricted. Actually, it is logical to assume that they occur in eastern Lapland as well, as the extensive granitic rocks of the Central Lapland Granite Complex continue eastwards for more than 200 km up to the Kuusamo schist area. In contrast, the granites of the Tanhua district on the eastern side of the Sodankylä schist area are heterogeneous S-type granites, which is also apparent from their discordant zircon population. The granite at Mustatseljät is garnet-bearing and, like the Hetta-type granites, contains in places supracrustal inclusions. At Neitselvaara they are cut by a pegmatitic granite dyke, which gives an age of 1808±10 Ma. This age is within the range obtained for late granites in Lapland, and together with U-Pb ages on titanites and monazites (e.g. Lauerma 1982) manifest the widespread 1.8 Ga thermal episode in Northern Finland.

Concluding remarks

The conventional U-Pb data reported in this paper have provided a geochronological framework for the evolution in Central Lapland. Several stages of magmatic activity spanning the period 2.43 - 1.80 Ga have been defined, which in early stages are related to continental rifting and later to orogenesis and reworking. However, the results for many samples have remained inconclusive, and more sophisticated methods are needed in order to unravel these problems. The contribution from Archean crust is evident in many Paleoproterozoic geological processes in Central Lapland, but the extent of Archean supracrustal rocks remains controversial. These problems are met especially in the Vuojärvi area (Fig.1) and at Honkavaara c. 30 km west of Sodankylä, where fine-grained felsic rocks interpreted as volcanic in origin have provided Archean age estimates (A194, A195 and Rastas et al. 2001, *this volume*). Several lines of evidence suggest that the bulk of the supracrustal rocks in the Honkavaara area are Paleoproterozoic in origin. These include the moderately enriched $\delta^{13}C$ value at +5.9±0.7 of the Kuoninkivaara dolomites suggesting a probable age under 2.2 Ga for these sedimentary deposits (Karhu 1993). Honkavaara lies in a highly tectonized and altered zone, where the bedrock is cut by a major thrust (the Sirkka Line) and is characterized by several faults and small thrusts associated with tectonic breccias. Accordingly, Lehtonen et al (1988) have interpreted the felsic volcanics as an Archean tectonic wedge enclosed by Paleoproterozoic rocks.

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Appendix I. U-Pb mineral age data from rocks in the Sodankylä area

Sample information[1)]	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS[2)]						Rho[3)]	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A65-Kelujärvi granodiorite															
A. +4.6	18.9	481.4	133.8	1827	0.191	0.2406	0.65	3.792	0.65	0.1143	0.15	0.97	1389	1591	1869
B. 4.2-4.6	15.2	894.8	174.0	283	0.228	0.1421	0.65	2.175	0.80	0.1110	0.48	0.80	856	1173	1815
C. 4.2-4.6	12.5	716.7	171.6	584	0.206	0.1937	0.65	3.016	0.68	0.1129	0.20	0.96	1141	1411	1846
D. 4.2-4.6/HF	13.2	581.5	129.9	1288	0.192	0.1912	0.65	2.980	0.65	0.1131	0.15	0.97	1128	1402	1849
E. 4.2-4.3/M/HF	12.5	745.1	165.8	366	0.219	0.1698	0.65	2.628	0.75	0.1122	0.25	0.95	1011	1308	1835
A163 Neitselvaara granite pegmatite															
A. +4.2	16.6	723.8	220.5	1072	0.176	0.2627	0.65	3.869	0.65	0.1068	0.15	0.97	1503	1607	1746
B. 4.0-4.2 abr	3.5	933.3	295.5	1311	0.145	0.2818	0.65	4.204	0.70	0.1082	0.20	0.96	1600	1674	1769
C. +4.0 abr	2.5	611.0	201.6	2000	0.180	0.2896	0.65	4.337	0.70	0.1086	0.20	0.96	1639	1700	1776
A194-Keski-Loviselkä felsic lava (base)															
A. 4.3-4.5/+200	11.9	325.2	126.3	5056	0.147	0.3378	1.01	8.539	1.10	0.1834	0.33	0.95	1875	2290	2683
B. 4.3-4.5/HF/uncrushed	20.0	210.3	119.6	3634	0.156	0.4873	0.65	12.889	0.66	0.1918	0.15	0.97	2559	2671	2758
C. 4.2-4.3/+200	18.6	428.9	170.7	2257	0.158	0.3419	0.68	8.558	0.72	0.1816	0.20	0.96	1895	2292	2667
D. 4.3-4.5/-200/abr 3 h	7.0	235.2	128.6	4740	0.155	0.4699	0.65	12.362	0.65	0.1908	0.15	0.97	2483	2632	2749
A195-Keski-Loviselkä felsic lava (top)															
A. +4.5/HF/uncrushed	20.5	148.2	81.7	5078	0.141	0.4818	0.65	12.094	0.65	0.1821	0.15	0.97	2534	2611	2672
B. +4.5/abr 3 h	10.5	131.8	77.0	4422	0.156	0.5026	0.65	13.046	0.65	0.1883	0.15	0.97	2624	2683	2727
C. Titanite 3.4-3.5	23.9	100.4	31.9	3720	0.047	0.3171	0.92	4.879	0.97	0.1116	0.30	0.95	1775	1798	1825
D. 4.3-4.5/rounded	7.2	252.0	141.2	3649	0.177	0.4757	0.65	11.983	0.65	0.1827	0.17	0.97	2508	2603	2677
E. euhedral	1.9	376.1	173.5	6140	0.112	0.4177	0.65	9.668	0.65	0.1679	0.15	0.97	2250	2403	2536
A281-Harjunoja diabase															
A. +4.3/abr 13 h	14.2	352.5	194.4	1576	0.588	0.3654	0.65	6.923	0.65	0.1374	0.15	0.97	2007	2101	2194
B. 4.2-4.3/clear/abr 7 h	13.0	515.7	285.0	1979	0.639	0.3567	0.80	6.764	0.80	0.1376	0.15	0.98	1966	2081	2196
C. 4.3-4.5/clear/+150/abr 9 h	16.2	352.9	195.2	1892	0.589	0.3671	1.20	6.986	1.20	0.1380	0.15	0.99	2015	2109	2202
D. +4.3/clear/abr 3 h	2.5	284.6	151.9	2527	0.545	0.3656	0.65	6.937	0.75	0.1377	0.36	0.88	2008	2103	2198
E. 4.0-4.2/+100/clear/abr 3 h	8.3	701.0	393.3	1479	0.669	0.3549	0.65	6.695	0.65	0.1368	0.15	0.97	1958	2071	2187
F. 4.0-4.2/clear/abr 3 h	8.5	699.7	413.8	2193	0.653	0.3789	1.10	7.216	1.10	0.1381	0.22	0.98	2071	2138	2203
G. 4.3-4.5/+200/clear/abr 3 h	16.1	337.4	192.5	3112	0.590	0.3800	0.65	7.269	0.65	0.1387	0.15	0.97	2076	2145	2211
H. 4.2-4.3/clear/+150/abr 3 h	13.1	511.1	298.8	2898	0.636	0.3790	1.10	7.236	1.10	0.1385	0.15	0.99	2071	2140	2208
I. 3.6-4.0/+150/clear/abr 3 h	5.9	1201.7	659.4	3379	0.729	0.3389	0.65	6.353	0.65	0.1360	0.15	0.97	1881	2025	2176

Appendix I continued

Sample information[1)] Analyzed mineral/fraction	Sample weight / mg	U ppm	Pb ppm	$^{206}Pb/^{204}Pb$ measured	$^{208}Pb/^{206}Pb$ radiogenic	ISOTOPIC RATIOS[2)] $^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%	Rho[3)]	APPARENT AGES (Ma) $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A281J. 3.6-4.0/+150/clear/abr 5 h	6.8	1125.2	519.0	2306	0.679	0.2922	0.65	5.391	0.65	0.1338	0.15	0.97	1652	1883	2148
K. 4.3-4.5/-200/clear/abr 4 h	12.8	313.0	176.3	2161	0.589	0.3743	0.65	7.141	0.65	0.1384	0.15	0.97	2049	2129	2207
L. +4.3/abr 4 h (separated from a portion rich in epidote)	3.4	339.1	192.5	1466	0.629	0.3674	0.65	6.954	0.68	0.1373	0.18	0.96	2017	2105	2193
A283-Kivivuotsonselkä mafic lava															
A. 4.2-4.6	20.0	251.5	114.9	5496	0.125	0.4082	0.65	9.609	0.65	0.1707	0.15	0.97	2206	2398	2565
B. 4.2-4.6/abr 2 h	4.3	204.5	104.2	23338	0.123	0.4565	0.65	10.952	0.65	0.1740	0.15	0.97	2424	2519	2596
C. 4.0-4.2	17.9	481.6	178.1	7257	0.137	0.3288	0.65	7.477	0.65	0.1649	0.15	0.97	1832	2170	2506
D. +4.6	2.4			3792						0.1715	0.15				2572
E. -4.0/abr 2 h	1.2	482.4	181.0	1629	0.147	0.3281	0.65	7.479	0.65	0.1653	0.15	0.97	1829	2170	2511
F. 4.2-4,6/HF/abr 3 h	3.9	141.2	72.6	10505	0.133	0.4547	0.65	11.257	0.65	0.1796	0.15	0.97	2416	2544	2649
G. 4.2-4.6/HF	21.0	177.3	88.6	51684	0.130	0.4438	0.65	10.910	0.65	0.1783	0.15	0.97	2367	2515	2637
A401-Mairivaara albite diabase															
A. +4.3/+150/abr 2 h	17.8	454.2	182.4	11613	0.205	0.3496	0.71	6.018	0.71	0.1248	0.15	0.98	1932	1978	2026
B. 4.2-4.3/+200/abr 2 h	13.0	552.0	224.6	11370	0.224	0.3491	0.68	6.006	0.68	0.1248	0.15	0.98	1930	1976	2025
C. 4.0-4.2/+150/abr 2 h	4.3	630.1	253.0	14626	0.238	0.3418	0.65	5.847	0.65	0.1241	0.15	0.97	1895	1953	2016
D. 4.0-4.2/+150/grey/abr 2 h	1.5	577.3	233.5	3010	0.241	0.3397	0.65	5.859	0.80	0.1251	0.25	0.96	1885	1955	2030
E. 3.8-4.0/+150/red/abr 1 h	4.3	940.8	367.7	9373	0.304	0.3166	0.65	5.387	0.65	0.1234	0.15	0.97	1773	1882	2006
F. +4.3/-150/abr 2 h	15.0	439.5	175.7	9135	0.199	0.3491	0.67	6.008	0.68	0.1248	0.15	0.98	1930	1976	2026
G. 4.2-4.3/-200/abr 3 h	4.1	631.0	245.2	4395	0.203	0.3369	0.65	5.778	0.65	0.1244	0.15	0.97	1871	1943	2020
H. 3.8-4.0/+150/grey/abr 1 h	4.1	1116.5	425.9	5120	0.359	0.2966	0.65	5.009	0.65	0.1225	0.16	0.97	1674	1820	1992
I. 3.6-3.8/red pigment/abr 3 h	4.8	1400.1	434.9	3656	0.350	0.2426	0.65	3.958	0.65	0.1184	0.15	0.97	1399	1625	1931
J. +4.3/red pigment/abr 2 h	4.2	449.2	179.2	7141	0.196	0.3491	0.65	6.001	0.65	0.1247	0.15	0.97	1930	1975	2024
K. 4.0-4.2/red pigment/abr 2 h	6.1	652.7	255.0	9752	0.224	0.3354	0.65	5.724	0.65	0.1238	0.15	0.97	1864	1934	2011
L. Opaque reddish brown	5.3	881.1	280.7	5781	0.183	0.2824	0.65	4.574	0.65	0.1175	0.15	0.97	1603	1744	1918
A418-Koskenkangas gabbro															
A. 4.3-4.6	21.6	263.5	116.7	8081	0.203	0.3829	0.65	7.005	0.65	0.1327	0.15	0.97	2089	2112	2134
B. 4.2-4.3/+200	20.8	582.3	268.5	5408	0.270	0.3787	0.65	6.890	0.65	0.1319	0.15	0.97	2070	2097	2124
A521 Pyhäjokikuru albite rock															
A. +4.5 rounded red pigm a2h		121.0		1430		0.5221	0.65	14.519	0.65	0.2017	0.18	0.96	2708	2784	2839
B. 4.3-4.5 rounded red p. a2h		315.0		3308		0.3740	0.65	9.758	0.65	0.1893	0.15	0.97	2047	2412	2735

Appendix I continued

Sample information[1)]	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS[2)]						Rho[3)]	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A570 Mustatseljät granite															
A. 3.6-3.8 a3h	4.0	2547.0	323.0	548	0.095	0.1111	0.65	1.692	0.67	0.1105	0.17	0.97	678	1005	1807
B. 4.0-4.2/-200 a3h	5.3	807.0	234.0	639	0.166	0.2412	0.76	4.255	0.78	0.1280	0.17	0.98	1392	1684	2070
C. 3.8-4.0/+200 a3h	7.0	1349.0	296.0	616	0.113	0.1893	0.65	3.310	0.65	0.1268	0.15	0.97	1117	1483	2054
D. +4.0/+200 a3h	3.5	835.0	262.0	519	0.130	0.2600	0.65	4.946	0.65	0.1380	0.15	0.97	1489	1810	2201
F. +4.2 a3h	2.0	760.0	225.0	1060	0.147	0.2564	0.65	4.663	0.68	0.1319	0.18	0.96	1471	1760	2123
G. 3.8-4.0 a3h	2.9	1380.0	279.0	874	0.119	0.1786	0.65	2.993	0.68	0.1215	0.18	0.96	1059	1405	1979
H. 3.6-3.8/+200 a3h	2.6	2588.0	294.0	821	0.160	0.0978	0.65	1.413	0.70	0.1047	0.26	0.93	601	894	1709
bI. +4.3 a2h	2.9	468.0	203.0	1225	0.195	0.3654	0.65	6.959	0.65	0.1381	0.15	0.97	2007	2106	2204
bJ. 4.2-4.3/+200 a3h	2.0	742.0	231.0	778	0.103	0.2729	0.65	5.296	0.70	0.1408	0.28	0.92	1555	1868	2236
bK. 4.2-4.3 a3h	3.3	1071.0	243.0	515	0.150	0.1877	0.65	3.202	0.73	0.1237	0.33	0.89	1108	1457	2011
bL. 4.0-4.2 a3h	4.1	1442.0	292.0	533	0.108	0.1734	0.65	2.980	0.68	0.1246	0.18	0.96	1030	1402	2023
E1. Monazite				1274						0.1083	0.15				1771
A741-Mairivaara albite diabase															
A. 3.6-3.8/+200/abr 5 h	5.9	924.3	376.1	1771	0.276	0.3294	0.74	5.551	0.74	0.1223	0.15	0.98	1835	1908	1989
B. 4.0-4.2	3.6	1041.2	412.3	4243	0.253	0.3279	0.66	5.546	0.67	0.1227	0.25	0.93	1828	1907	1995
C. -4.0	15.4	935.4	371.7	2743	0.256	0.3303	0.68	5.623	0.70	0.1235	0.20	0.96	1839	1919	2007
D. -4.0/abr 3 h	7.7	973.5	391.4	5930	0.255	0.3363	0.71	5.702	0.71	0.1230	0.15	0.98	1868	1931	1999
E. +4.0/abr 3 h	4.0	945.0	376.3	2230	0.249	0.3305	0.65	5.579	0.65	0.1224	0.15	0.97	1840	1912	1992
F. -4.0/-200/abr 5 h	6.2	988.8	404.8	2050	0.266	0.3345	0.93	5.679	0.93	0.1231	0.15	0.99	1860	1928	2002
G. -4.0/+200/abr 5 h	7.8	905.8	377.0	1906	0.271	0.3385	0.79	5.725	0.80	0.1227	0.15	0.98	1879	1935	1995
H. -4.0/+200/turbid/abr 12 h	6.0	1050.8	390.3	4066	0.260	0.3092	0.81	5.141	0.83	0.1206	0.20	0.97	1736	1842	1965
I. -4.0/+200/glassy/abr 12 h	3.5	1291.0	520.9	3441	0.330	0.3186	1.14	5.385	1.14	0.1226	0.15	0.99	1782	1882	1994
J. 3.6-3.8/+200	13.3	1086.7	403.0	1770	0.298	0.2962	0.66	4.886	0.71	0.1196	0.45	0.79	1672	1799	1950
K. 3.6-3.8/grey/abr 3 h	1.1	1018.7	334.0	1890	0.254	0.2711	0.65	4.526	0.65	0.1211	0.16	0.97	1546	1735	1972
L. 3.6-3.8/red pigment/abr 3 h	2.8	1015.5	381.3	3139	0.286	0.3056	0.65	5.123	0.65	0.1216	0.15	0.97	1719	1839	1979
A742-Mairivaara albite diabase															
A. 3.8-4.0	7.3	972.3	342.9	2225	0.216	0.2998	0.70	5.067	0.70	0.1226	0.15	0.98	1690	1830	1994
B. +4.3/abr 1 h (zircon + badd)	1.6	500.0	195.4	4443	0.163	0.3508	0.65	6.024	0.70	0.1245	0.20	0.96	1938	1979	2022
C. 4.2-4.3/abr 1 h	4.4	729.5	276.8	4867	0.172	0.3372	0.65	5.740	0.65	0.1235	0.15	0.97	1872	1937	2007
D. 4.0-4.2/-200/abr 3 h	6.4	844.0	314.1	3551	0.180	0.3276	0.65	5.547	0.65	0.1228	0.15	0.97	1826	1907	1997
E. 4.2-4.3/abr 3 h	3.4	761.4	306.6	2602	0.175	0.3535	0.65	6.030	0.65	0.1237	0.15	0.97	1951	1980	2010
F. 3.8-4.0/+200/abr 3 h	3.3	1098.6	414.0	2302	0.214	0.3209	0.65	5.416	0.70	0.1224	0.18	0.97	1794	1887	1992
G. +4.0/abr 3 h	1.9	855.0	311.9	2002	0.186	0.3167	0.65	5.371	0.75	0.1230	0.22	0.96	1773	1880	2000
H. 4.0-4.2/-200/abr 3 h	4.2	863.0	318.9	1998	0.189	0.3195	0.65	5.412	0.65	0.1228	0.15	0.97	1787	1886	1998
I. 3.8-4.0/-200	4.1	996.0	369.0	684	0.218	0.2994	0.65	5.031	0.75	0.1219	0.22	0.96	1688	1824	1984
J. 3.8-4.0/-200/abr 2 h	2.2	1036.7	392.6	2000	0.229	0.3183	0.65	5.364	0.65	0.1222	0.15	0.97	1781	1879	1988
K. 4.0-4.2/-200/abr 3 h	4.9	867.4	327.7	3531	0.190	0.3299	0.65	5.589	0.65	0.1229	0.19	0.96	1837	1914	1999

Appendix I continued

Sample information[1)]	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS[2)]						Rho[3)]	APPARENT AGES (Ma)		
Analyzed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A743-Mairivaara albite diabase															
A. 3.8-4.0	5.2	888.6	322.8	1828	0.212	0.3088	0.70	5.243	0.70	0.1232	0.15	0.98	1734	1859	2002
B. 4.0-4.2/abr 2 h	5.4	860.7	341.5	3967	0.208	0.3423	0.70	5.811	0.70	0.1231	0.15	0.98	1897	1948	2002
C. 3.8-4.0/abr 3 h	5.5	1098.5	438.7	2400	0.247	0.3321	0.75	5.630	0.75	0.1229	0.15	0.98	1848	1920	1999
A815, A815b Luosto diabase															
A. +4.2	20.7	531.0	177.0	1003	0.544	0.2234	0.65	3.582	0.65	0.1163	0.15	0.97	1299	1545	1899
B. 3.8-4.0/+200	12.5	729.0	171.0	1030	0.534	0.1590	0.65	2.537	0.68	0.1157	0.18	0.96	951	1282	1891
C. 4.0-4.2	12.1	598.0	185.0	1312	0.528	0.2115	0.65	3.417	0.65	0.1172	0.15	0.97	1236	1508	1913
D. +4.2 abr3h	4.7	539.0	189.0	2247	0.548	0.2397	0.65	3.859	0.65	0.1168	0.15	0.97	1385	1605	1907
E. +4.2 abr5h	7.4	522.0	187.0	529	0.555	0.2323	0.65	3.726	0.68	0.1163	0.20	0.96	1346	1577	1900
F. 4.0-4.2/+150 m abr5h	10.2	653.0	156.0	934	0.539	0.1602	0.65	2.577	0.68	0.1166	0.18	0.96	958	1294	1905
G. +4.2 abr5h	4.1	541.0	201.0	553	0.552	0.2411	0.65	3.886	0.70	0.1169	0.22	0.95	1392	1610	1909
bA. 3.8-4.0/+200 clear abr1h	3.0	2316.0	1029.0	26387	0.345	0.3305	1.00	5.620	1.00	0.1233	0.15	0.99	1840	1919	2005
bB. 3.8-4.0/+200 turbid abr1h	4.0	1426.0	505.0	5059	0.404	0.2688	0.65	4.264	0.65	0.1151	0.15	0.97	1534	1686	1881
bC. 4.2-4.3/abr1h	2.0	505.0	187.0	567	0.237	0.2935	0.65	4.625	0.73	0.1143	0.28	0.92	1659	1753	1868
bD. +4.3 badd (+zircon)	3.6	858.0	295.0	2808	0.096	0.3221	0.65	5.467	0.65	0.1231	0.15	0.97	1800	1895	2001
bE. +4.3 a zircon+badd	6.2	497.0	182.0	1945	0.187	0.3185	0.65	5.122	0.65	0.1166	0.15	0.97	1782	1839	1905
A816-Kylälampi albitite															
A. +4.6	27.9	227.1	81.8	849	0.124	0.3160	0.65	5.477	0.68	0.1257	0.20	0.96	1769	1897	2039
B. 4.2-4.6	31.0	444.6	156.9	1025	0.142	0.3080	0.65	5.332	0.65	0.1255	0.15	0.97	1730	1873	2036
D. 4.0-4.2	17.1	609.8	195.7	2282	0.155	0.2850	0.65	4.940	0.65	0.1257	0.15	0.97	1616	1809	2038
E. 3.8-4.0	7.4	408.0	131.5	2400	0.143	0.2905	0.65	5.005	0.66	0.1249	0.17	0.97	1644	1820	2028
F. +4.6/HNO_3/crushed	20.6	218.0	83.0	664	0.124	0.3277	0.65	5.697	1.00	0.1261	0.74	0.67	1826	1930	2044
G. +4.6/HF/uncrushed	20.7	196.3	66.1	20757	0.121	0.3131	0.74	5.471	0.75	0.1267	0.15	0.98	1756	1895	2053
H. +4.6/+150/abr 5 h	16.5	218.4	82.9	1150	0.124	0.3376	0.66	5.890	0.66	0.1265	0.15	0.97	1874	1959	2050
I. +4.6/+150/abr 10 h	22.3	215.4	80.4	2986	0.123	0.3411	0.65	5.962	0.65	0.1268	0.15	0.97	1891	1970	2053
J. +4.6/abr 5 h	20.4	214.1	81.0	781	0.122	0.3295	0.65	5.728	0.65	0.1261	0.15	0.97	1835	1935	2044
K. +4.6/abr 15 h	7.3	190.2	72.0	3935	0.130	0.3482	0.65	6.093	0.65	0.1269	0.15	0.97	1925	1989	2056
L. +4.6/+150/abr 30 h	20.5	216.0	77.7	1569	0.137	0.3257	0.65	5.663	0.65	0.1261	0.15	0.97	1817	1925	2044
M. +4.6/abr 42 h	20.8	209.7	77.0	2234	0.131	0.3357	0.65	5.837	0.65	0.1261	0.15	0.97	1865	1951	2044
N. +4.6/-200/abr 14 h	20.0	218.1	81.6	758	0.167	0.3250	0.65	5.655	0.70	0.1262	0.24	0.94	1814	1924	2046
O. 4.0-4.2	11.6	702.4	233.6	835	0.198	0.2830	0.65	4.907	0.68	0.1258	0.20	0.96	1606	1803	2040
P. +4.6/HF/uncrushed	20.8	153.1	56.8	8897	0.125	0.3436	0.65	6.057	0.65	0.1279	0.15	0.97	1903	1984	2069

Appendix I continued

Sample information[1)] Analyzed mineral/fraction	Sample weight / mg	U ppm	Pb ppm	$^{206}Pb/^{204}Pb$ measured	$^{208}Pb/^{206}Pb$ radiogenic	ISOTOPIC RATIOS[2)] $^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%	Rho[3)]	APPARENT AGES (Ma) $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A817-Kylälampi albitite															
A. +4.6	31.5	169.0	65.9	1967	0.121	0.3537	0.65	6.372	0.65	0.1307	0.15	0.97	1951	2028	2107
B. 4.2-4.6	25.6	473.1	172.3	1545	0.139	0.3229	0.65	5.761	0.65	0.1294	0.16	0.97	1803	1940	2090
C. Titanite	28.5	45.1	19.8	1348	0.454	0.3176	0.65	4.822	0.90	0.1101	0.52	0.82	1777	1788	1801
D. +4.6/HNO_3/crushed	20.3	149.8	60.3	1482	0.126	0.3598	0.69	6.475	0.71	0.1305	0.27	0.93	1981	2042	2105
E. +4.6/HF/uncrushed	22.2	140.4	56.1	28441	0.125	0.3700	0.72	6.666	0.72	0.1307	0.15	0.98	2029	2068	2107
F. +4.6/abr 10 h	19.9	143.4	57.3	9023	0.120	0.3697	0.65	6.667	0.65	0.1308	0.15	0.97	2028	2068	2108
G. +4.6/abr 5 h	20.1	150.9	62.0	1737	0.119	0.3714	0.65	6.692	0.65	0.1307	0.16	0.97	2035	2071	2107
A818-Ahvenselkä diabase															
A. +4.6	21.0	363.7	189.0	1342	0.620	0.3359	0.65	5.850	0.75	0.1263	0.30	0.92	1866	1953	2047
B. 4.2-4.6	16.0	467.8	241.0	1742	0.632	0.3333	0.65	5.759	0.65	0.1253	0.15	0.97	1854	1940	2033
C. 4.0-4.2	15.8	552.7	270.5	1698	0.654	0.3128	0.65	5.327	0.65	0.1235	0.17	0.97	1754	1873	2007
D. Titanite	30.9	14.6	16.9	573	2.780	0.3363	2.10	5.206	2.90	0.1123	1.50	0.87	1869	1853	1836
E. +4.6/abr 2 h	3.2	429.3	236.1	1066	0.637	0.3500	0.86	6.092	0.90	0.1262	0.18	0.98	1934	1989	2046
F. 4.2-4.6/abr 2 h	2.9	553.9	295.7	1491	0.636	0.3433	0.65	5.950	0.68	0.1257	0.17	0.97	1902	1968	2039
G. 4.2-4.6/M/abr 3 h	1.7	378.0	218.5	1169	0.800	0.3392	0.94	5.877	0.95	0.1256	0.20	0.98	1882	1957	2038
H. 4.0-4.2/abr 3 h	15.2	628.6	306.5	1473	0.647	0.3114	0.68	5.312	0.68	0.1237	0.15	0.98	1747	1870	2010
I. +4.2/M/abr 6 h	10.2	506.6	280.4	886	0.655	0.3457	0.82	6.012	0.84	0.1262	0.15	0.98	1914	1977	2045
A820-Rovasvaara gabbro pegmatoid															
A. +4.6	27.8	342.2	123.5	4096	0.108	0.3364	0.65	5.817	0.65	0.1254	0.15	0.97	1869	1948	2034
B. +4.2	27.1	492.5	177.0	3925	0.128	0.3293	0.65	5.679	0.65	0.1251	0.15	0.97	1835	1928	2030
C. 4.0-4.2	16.2	703.5	246.2	3360	0.146	0.3156	0.65	5.416	0.65	0.1245	0.15	0.97	1768	1887	2021
D. +4.6/grey/abr 5 h	7.7	343.6	129.9	7358	0.110	0.3536	0.70	6.147	0.70	0.1261	0.15	0.98	1951	1996	2044
E. +4.6/red pigment/abr 5 h	9.3	361.5	139.2	9349	0.105	0.3620	0.73	6.281	0.73	0.1258	0.15	0.98	1991	2015	2040
A841-Pittiövaara albite diabase															
A. 3.8-4.0/+200	18.2	1036.7	448.1	5251	0.398	0.3260	0.65	5.833	0.65	0.1298	0.15	0.97	1818	1951	2095
B. 4.0-4.2/+200	12.0	819.9	343.9	1891	0.320	0.3278	0.65	5.860	0.80	0.1297	0.30	0.94	1827	1955	2093
C. 4.2-4.6	5.9	600.9	246.6	4097	0.256	0.3407	0.65	6.174	0.70	0.1314	0.22	0.95	1890	2000	2117
D. 3.6-3.8	14.7	890.2	357.8	4437	0.367	0.3095	0.65	5.499	0.65	0.1289	0.15	0.97	1738	1900	2083
E. 3.6-4.0	6.7	951.8	355.8	4342	0.364	0.2886	0.68	5.099	0.68	0.1282	0.15	0.98	1634	1835	2072
F. 4.2-4.6	2.3	635.2	257.7	1865	0.255	0.3281	0.65	5.838	0.94	0.1291	0.55	0.82	1828	1952	2085

Appendix I continued

Sample information[1)] Analyzed mineral/fraction	Sample weight / mg	U ppm	Pb ppm	$^{206}Pb/^{204}Pb$ measured	$^{208}Pb/^{206}Pb$ radiogenic	ISOTOPIC RATIOS[2)] $^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%	Rho[3)]	APPARENT AGES (Ma) $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A861a, A861b-Törmänen albite diabase															
aA. 3.8-4.0/abr 3 h	1.6	833.4	272.0	692	0.603	0.2095	0.65	3.239	0.70	0.1121	0.22	0.95	1226	1466	1834
aB. 3.8-4.0	12.1	791.1	250.8	811	0.630	0.2017	0.85	3.138	0.85	0.1129	0.19	0.98	1184	1442	1846
bA. +4.3/clear/abr 1 h	1.3	373.3	161.3	8335	0.267	0.3601	0.65	6.267	0.70	0.1262	0.20	0.96	1982	2013	2046
bB. +4.3/-200/clear/abr 1 h	3.6	392.3	171.6	8880	0.278	0.3608	0.65	6.288	0.65	0.1264	0.15	0.97	1986	2016	2048
bC. 4.2-4.3/clear/abr 1 h	4.4	609.6	264.7	11213	0.265	0.3618	0.65	6.270	0.65	0.1257	0.15	0.97	1990	2014	2038
bD. 4.0-4.2/abr 3 h	3.2	903.9	366.9	1390	0.260	0.3296	0.65	5.648	0.65	0.1243	0.15	0.97	1836	1923	2018
bE. 3.8-4.0/abr 2 h/pigment	4.4	1117.5	430.4	4343	0.289	0.3136	0.65	5.339	0.65	0.1235	0.15	0.97	1758	1875	2007
bF. 3.8-4.0/abr 3 h/turbid	3.0	1083.8	386.1	1697	0.271	0.2890	0.66	4.901	0.67	0.1230	0.15	0.97	1636	1802	2000
A862-Törmänen albite diabase															
A. 4.0-4.2	15.4	718.6	254.0	1860	0.535	0.2439	0.96	3.792	0.96	0.1128	0.15	0.99	1406	1590	1844
B. 3.8-4.0	10.4	843.3	250.5	1729	0.510	0.2082	0.72	3.158	0.90	0.1100	0.50	0.83	1219	1446	1798
C. 3.8-4.0/abr 3 h	1.1	876.8	280.8	1559	0.513	0.2244	0.65	3.382	0.70	0.1093	0.22	0.95	1304	1500	1788
D. +4.2/abr 1 h	2.9	614.3	253.2	2778	0.504	0.2908	0.65	4.813	0.65	0.1201	0.15	0.97	1645	1787	1957
E. +4.0/+100/abr 2 h	5.2	610.4	257.2	2026	0.657	0.2707	0.70	4.398	0.72	0.1178	0.20	0.96	1543	1711	1924
F. +4.0/-100/abr 8 h	9.0	728.7	250.7	1867	0.550	0.2352	0.71	3.596	0.71	0.1109	0.15	0.98	1361	1548	1813
G. 4.0-4.2/abr 12 h	4.7	737.5	260.9	2096	0.540	0.2442	0.73	3.733	0.73	0.1109	0.15	0.98	1408	1578	1813
H. 4.0-4.2/abr 5 h	2.6	713.0	283.1	860	0.558	0.2640	0.73	4.051	0.75	0.1113	0.22	0.96	1510	1644	1820
I. 3.8-4.0/+150/abr 4 h	6.1	804.4	259.0	765	0.528	0.2167	0.69	3.245	0.70	0.1086	0.18	0.97	1264	1467	1776
A863-Törmänen albite diabase															
A. 4.0-4.2	12.9	702.5	282.2	2603	0.529	0.2787	0.66	4.642	0.67	0.1208	0.20	0.95	1584	1756	1968
B. 3.8-4.0	12.7	790.8	262.9	1412	0.586	0.2210	1.00	3.497	1.04	0.1148	0.66	0.79	1287	1526	1876
C. 3.8-4.0/-200/abr 3 h	5.1	939.5	306.1	647	0.445	0.2275	0.85	3.576	0.86	0.1140	0.22	0.97	1321	1543	1864
D. +4.0/abr 2 h	1.7	559.5	266.4	639	0.748	0.2804	0.65	4.643	0.73	0.1201	0.24	0.95	1593	1756	1957
E. +4.0/M/abr 1 h	2.3	786.2	286.2	1319	0.403	0.2693	0.76	4.430	0.77	0.1193	0.15	0.98	1537	1717	1946
A900 Rantavaara gabbro															
A +4.2	20.08	260	113	6718	0.240	0.3663	0.65	6.625	0.65	0.1312	0.15	0.97	2012	2062	2113
B +4.6	28.76	180	78.6	2189	0.210	0.3696	0.65	6.728	0.65	0.1320	0.15	0.97	2027	2076	2125
C 4.2-4.6/100-200	25.19	401	179	1558	0.250	0.3623	0.65	6.553	0.68	0.1312	0.18	0.96	1993	2053	2114
D 4.0-4.2/+100	24.00	649	297	3458	0.350	0.3544	0.65	6.399	0.65	0.1309	0.15	0.97	1955	2032	2110
E +4.6/+200clear abr3h	0.53	211	102	3890	0.310	0.3847	0.50	7.063	0.50	0.1331	0.15	0.96	2098	2119	2140
F +4.6/-200 clear abr3h	0.22	187	86.6	2918	0.240	0.3854	0.50	7.059	0.50	0.1328	0.15	0.96	2102	2119	2136
G +4.6/+200 euh abr3h	0.29	216	97	1893	0.230	0.3742	0.50	6.835	0.50	0.1325	0.15	0.96	2049	2090	2131

Appendix I continued

Sample information[1)] Analyzed mineral/fraction	Sample weight / mg	U ppm	Pb ppm	$^{206}Pb/^{204}Pb$ measured	$^{208}Pb/^{206}Pb$ radiogenic	ISOTOPIC RATIOS[2)] $^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%	Rho[3)]	APPARENT AGES (Ma) $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A962-Povivaara gabbro pegmatoid															
A. +4.6	25.4	299.3	121.5	3592	0.133	0.3682	0.65	6.709	0.65	0.1322	0.15	0.97	2020	2073	2127
B. 4.2-4.6	25.9	465.8	180.5	3405	0.150	0.3467	1.00	6.259	1.00	0.1309	0.15	0.99	1919	2012	2110
C. 4.0-4.2	13.2	888.4	332.4	2366	0.200	0.3205	0.65	5.721	0.65	0.1295	0.15	0.97	1792	1934	2090
A992-Purkkivaara felsic volcanic rock															
A. -3.6/-200/abr 3 h	1.2	1077.7	237.6	883	0.300	0.1750	0.65	3.016	0.70	0.1250	0.22	0.95	1039	1411	2028
B. 3.6-3.8	5.5	1923.9	517.7	1882	0.253	0.2226	0.82	4.113	0.83	0.1340	0.18	0.97	1295	1656	2151
C. -3.6/+200/abr 3 h	2.6	2101.0	538.5	1206	0.294	0.2046	0.65	3.724	0.70	0.1320	0.22	0.95	1200	1576	2124
D. 3.6-4.2/abr 3 h	1.2	1572.0	715.5	6539	0.185	0.3940	0.65	8.317	0.65	0.1531	0.15	0.97	2141	2266	2381
A1021-Haikaraselkä fuchsite quartzite															
A. 4.2-4.3/-200	15.4	568.0	301.5	6189	0.116	0.4689	0.61	12.313	0.65	0.1905	0.23	0.94	2478	2628	2746
B. 4.0-4.2/-200	15.5	765.0	381.6	6363	0.111	0.4439	0.63	11.372	0.64	0.1858	0.15	0.97	2368	2554	2705
C. 4.2-4.3/+200	7.6	559.2	305.4	14400	0.118	0.4846	0.64	12.762	0.76	0.1910	0.41	0.84	2547	2662	2750
D. 4.0-4.2/+200	8.8	727.8	377.4	7837	0.110	0.4626	0.63	11.819	0.63	0.1853	0.15	0.97	2451	2590	2701
E. 4.2-4.3/abr 3 h	8.3	617.3	337.7	17284	0.113	0.4862	0.65	12.839	0.66	0.1915	0.15	0.97	2554	2667	2755
F. 4.0-4.2/abr 3 h	6.1	735.2	379.5	13122	0.110	0.4615	0.63	11.824	0.63	0.1858	0.15	0.97	2446	2590	2705
A1432-Sakiamaa felsic volcanic rock															
A. +4.3/-200/pale turbid/abr 0.5 h	0.3	602.1	232.3	1693	0.200	0.3239	0.50	6.313	0.50	0.1414	0.15	0.96	1809	2020	2244
B. 4.0-4.2/+200	5.0	1147.6	544.0	3815	0.250	0.3876	0.50	8.215	0.50	0.1537	0.15	0.96	2111	2254	2388
C. 4.0-4.2/+200/abr 2 h	0.4	1215.9	601.3	4848	0.250	0.4071	0.50	8.710	0.50	0.1552	0.15	0.96	2202	2308	2404
D. 3.6-4.6/dark/stubby/abr 10 h	0.8	2021.6	1030.3	6599	0.260	0.4164	0.50	8.923	0.50	0.1554	0.15	0.96	2244	2330	2406
E. 3.6-4.0/pale turbid/abr 6 h	0.4	2590.0	994.4	4165	0.330	0.2995	0.50	5.970	0.50	0.1446	0.15	0.96	1689	1972	2283
A1434-Pyytövaara felsic clast in volcanic conglomerate															
A. +4.5/abr 4 h	8.2	146.5	75.3	19894	0.184	0.4424	0.50	10.129	0.50	0.1661	0.15	0.96	2361	2446	2518
B. +4.5	6.9	155.6	77.2	16045	0.177	0.4289	0.50	9.872	0.50	0.1670	0.15	0.96	2300	2422	2527
4.3-4.5	8.4	382.2	171.6	18589	0.145	0.4002	0.50	8.676	0.50	0.1572	0.15	0.96	2169	2304	2426
4.2-4.3	8.6	539.5	279.9	23877	0.130	0.3949	0.50	8.581	0.50	0.1576	0.15	0.96	2145	2294	2430
A1498 Sakiamaa rhyolite															
A 4.0-4.3 abr5h small turbid	0.42	881	330	2240	0.230	0.3099	0.50	6.169	0.60	0.1444	0.30	0.87	1740	2000	2280
B 4.0-4.3 small turbid	0.64			2222	0.230	0.2723	0.50	5.314	0.50	0.1416	0.15	0.96	1552	1871	2246

Appendix I continued

Sample information[1)] Analyzed mineral/fraction	Sample weight / mg	U ppm	Pb ppm	$^{206}Pb/^{204}Pb$ measured	$^{208}Pb/^{206}Pb$ radiogenic	ISOTOPIC RATIOS[2)] $^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%	Rho[3)]	APPARENT AGES (Ma) $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A1521 Pyhälampi granite pebble in conglomerate															
A +4.3 abr3h sharp prisms	0.24	404	142	2011	0.320	0.27590	0.5	4.9931	0.5	0.13126	0.15	0.96	1571	1818	2115
B +4.2 abr3h large prisms	0.19	385	130	1169	0.370	0.25228	0.5	4.4700	0.5	0.12851	0.15	0.96	1450	1725	2078
A1524 Akanvaara rhyolite															
A +4.3 abr6h	0.45	450	209	13255	0.140	0.41692	0.5	8.7351	0.5	0.15195	0.15	0.96	2247	2311	2368
B +4.3	0.51	522	225	11121	0.140	0.38750	0.5	7.9989	0.5	0.14971	0.15	0.96	2111	2231	2343
C 4.2-4.3	0.88	607	222	8843	0.140	0.32896	0.5	6.5274	0.5	0.14391	0.15	0.96	1833	2050	2275
D +4.3 clear	0.25	499	224	9203	0.150	0.39623	0.5	8.3409	0.5	0.15267	0.15	0.96	2152	2269	2376
E +4.3 +200 long pale	0.51	477	189	4798	0.150	0.35221	0.5	7.1444	0.5	0.14712	0.15	0.96	1945	2130	2313
F +4.3 +200 long pale abr5h	0.58	436	194	10157	0.140	0.39837	0.5	8.1992	0.5	0.14928	0.15	0.96	2162	2253	2338
G +4.3 +200 abr16h	0.54	499	230	12599	0.140	0.41320	0.5	8.6315	0.5	0.15151	0.15	0.96	2230	2300	2363
H +4.3 -200 abr16h	0.48	462	217	7708	0.150	0.41659	0.5	8.7629	0.5	0.15256	0.15	0.96	2245	2314	2375
A1586 Metsävaara (Rantavaara gabbro)															
A +4.2 abr5h clear fragm.	0.56	430	234	14126	0.480	0.38858	0.5	7.1407	0.5	0.13328	0.15	0.96	2116	2129	2142

[1)] Zircon unless otherwise stated. Density (in g cm^{-3}), size (in mesh), duration of abrasion (abr, in hours), preleached in HF (HF)
[2)] Isotopic ratios corrected for fractionation, blank and age related common lead.
[3)] Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
Pb blank 20-50 pg for small samples (< mg), for others 200-500 pg.
SE = standard error in the mean

Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
Edited by Matti Vaasjoki
Geological Survey of Finland, Special Paper 33, 189–200, 2001.

U-Pb GEOCHRONOLOGY OF THE PEURASUVANTO AREA, NORTHERN FINLAND

by
Tuomo Manninen, Pekka Pihlaja and Hannu Huhma

Manninen. T., Pihlaja, P. & Huhma, H. 2001. U-Pb geochronology of the Peurasuvanto area, northern Finland. *Geological Survey of Finland, Special Paper 33*, 189-200. 5 figures, one table and one appendix.

U-Pb zircon age determinations are presented for ten samples from the Peurasuvanto area, northern Finland. The area consists of both Archean and early Paleoproterozoic rocks northwest of the large 2.44 Ga Koitelainen layered mafic intrusion. Most of the data were collected during prospecting and mapping of the area in 1970-80, and some age estimates, without analytical data, have been published earlier.

Archean rocks older than 3 Ga are exposed in the Tojottamanselkä dome, whereas heavy zircons from a quartzofeldspathic gneiss within the Pomokaira basement complex give an age of 2486±4 Ma, thus actually representing the Paleoproterozoic cover of the basement. A felsic tuff from the Rookkiaapa Formation of the Salla Group has been dated at 2438±8 Ma, providing further evidence for the felsic volcanism associated with the c. 2.44 Ga old layered intrusions. A granitic clast from a volcanic conglomerate belonging to the Rookkiaapa Formation does not constrain the age of deposition, as the only, very discordant analysis suggests a clear Archean age. U-Pb zircon data on a felsic breccia interbed of this conglomerate are slightly heterogeneous suggesting involvement of xenocrystic zircon. A hornblende gabbro that intrudes the volcanic conglomerate is 2124±5 Ma old.

Key words (GeoRef Thesaurus, AGI): absolute age, U-Pb, zircon, gneisses, metavolcanic rocks, quartzites, gabbros, Paleoproterozoic, Archean, Peurasuvanto, Lappi Province, Finland

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INTRODUCTION

The Peurasuvanto area lies on the northwestern margin of the Paleoproterozoic Sodankylä schist belt, in the northern part of the Fennoscandian Shield. The geology of the area was outlined by Erkki Mikkola (1941), and since then more detailed investigations in the region have been carried out (e.g. Puustinen 1977,

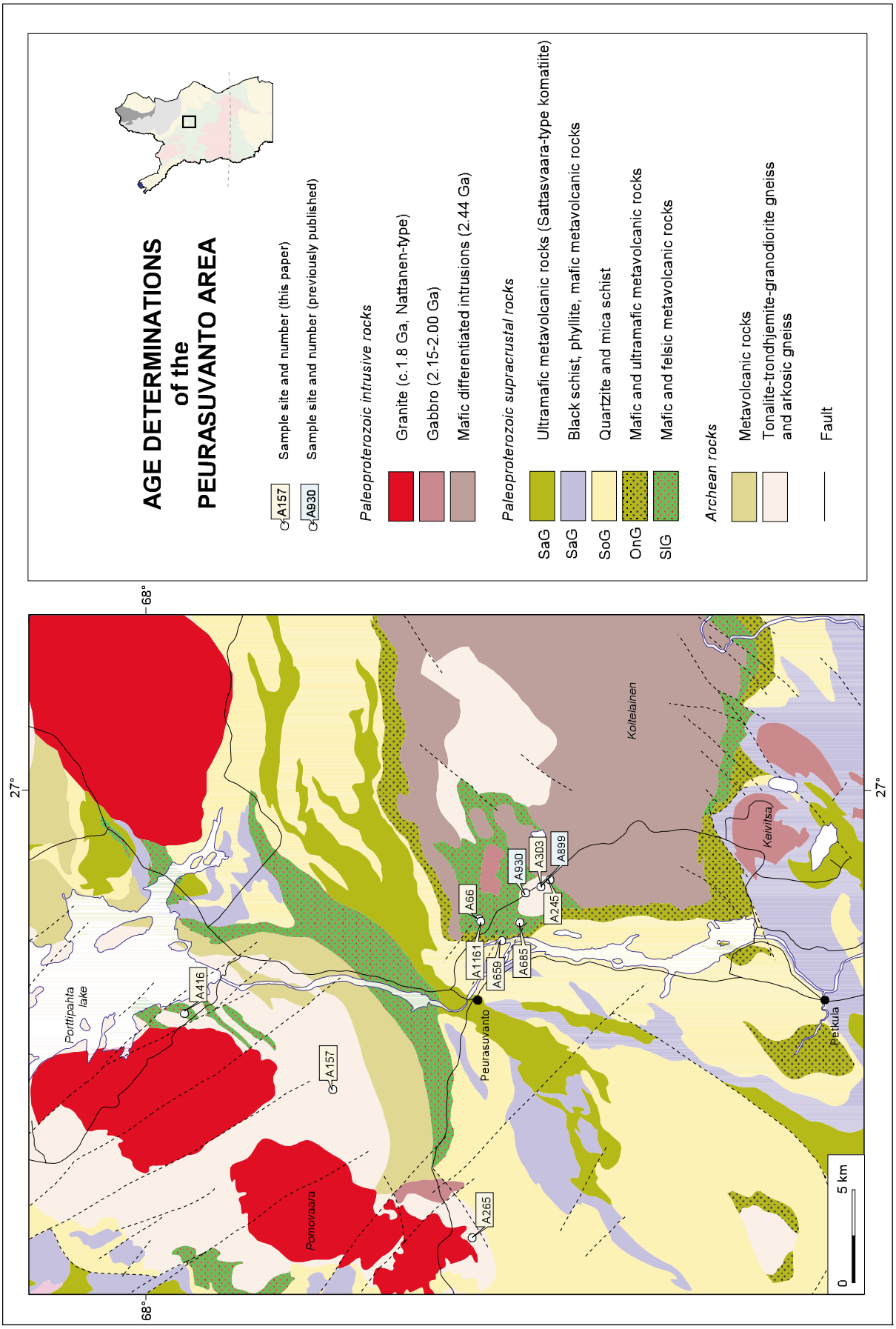

Fig. 1. Geological map of the Peurasuvanto area. SlG = Salla Group, OnG = Onkamo Group, SoG = Sodankylä Group, SaG = Savukoski Group.

Kröner et al. 1981, Jahn et al. 1984, Peltonen 1986, Peltonen et al. 1988, Pihlaja & Manninen 1988). The lithological map of the area in a scale 1:100,000 was compiled by Pihlaja and Manninen in 1993. Recently, a revised lithostratigraphical division, affecting also the supracrustal rocks of the study area, has been presented in the new stratigraphical map of Finnish Central Lapland (Räsänen et al. 1996, Lehtonen et al. 1998).

The bedrock in the Peurasuvanto area is composed of gneisses of the Archean basement complex, overlain by the Paleoproterozoic volcanic-sedimentary sequence of the Sodankylä schist belt. As a part of the Central Lapland Greenstone Belt (cf. Lehtonen et al. 1998), the supracrustal rocks of the study area represent the c. 2.5-2.0 Ga old fillings of an intracratonic rift zone (e.g. Gaál & Gorbatschev 1987, Gorbatschev & Bogdanova 1993, Lehtonen et al. 1998). The Koitelainen and Keivitsa layered mafic intrusions as well as the post-tectonic Nattanen-type granites were emplaced in the supracrustal rocks of the study area about 2.44 Ga, 2.05 Ga and 1.77 Ga ago, respectively (Front et al 1989, Huhma et al. 1996, Mutanen 1997, Mutanen & Huhma 2001, *this volume*).

The rocks of the Archean basement complex and especially the lowermost lithological units of its Paleoproterozoic supracrustal cover are poorly dated, hampering lithostratigraphic correlations between different greenstone belts of the Fennoscandian Shield. Also the basement-cover relationships of the Central Lapland greenstone belt area are still largely unknown. This paper presents U-Pb zircon analyses for 10 samples from the quartzofeldspathic gneisses of the Archean basement complex, from the overlying Paleoproterozoic supracrustal rocks, and from a 2.1 Ga old hornblende gabbro from the Peurasuvanto area, northern Finland. The data were collected during the prospecting and mapping of the area in 1970-80, and some of these ages have been published earlier in various papers without analytical data (see e.g. Pihlaja & Manninen 1988). General geology and sample localities are shown in Figure 1, and the U-Pb analytical data in Appendix I.

REGIONAL GEOLOGY

The Archean Pomokaira Basement Complex (Pomokaira basement complex; Lehtonen et al. 1998) belongs into the Karelian Domain of the Fennoscandian Shield (Gaál & Gorbatschev 1987). The ortho- and paragneisses of the complex, exposed in the cores of the domal structures include grey, granodioritic orthogneisses, fine- to medium-grained quartz-feldspar gneisses, variably granitized arkose gneisses and minor arkose quartzites. A reddish tonalitic gneiss from the small Tojottamaselkä dome in the center of the study area, dated at 3.1 Ga, belongs to the most ancient rocks of the Fennoscandian Shield (Kröner et al. 1981, Kröner & Compston 1990).

Applying the revised lithostratigraphical division of the Central Lapland greenstone belt, the Paleoproterozoic supracrustal sequence of the Peurasuvanto area has been divided into the Salla, Onkamo, Savukoski and Sodankylä Groups. Intermediate and felsic metavolcanites, a volcanic conglomerate and minor volcanoclastic rocks of the Rookkiaapa Formation (the former Madetkoski Formation of Lehtonen et al. 1992), belonging into the Salla Group (see Lehtonen et al. 1998) represent the lowermost lithostratigraphic unit of the Sodankylä schist belt. Due to faulting and poor exposure, neither the basal relations to the Archean basement gneisses below, nor the internal stratigraphy of the Formation are well established. Intermediate metavolcanic rocks, probably situated in the lower part of the sequence, are mostly andesitic amygdaloidal lava flows, with the amount of small, albite and epidote-filled amygdules exceeding locally 30 vol.% of the rock. Texturally these rocks are nematoblastic, consisting mainly of tremolite-actinolite needles and strongly altered, albitic plagioclase. Along the Kitinen river in the western part of the Peurasuvanto area, dacitic lavas and rhyolitic crystal tuffs generally exist only as variably thick interlayers within the andesitic flows, but predominate the area around the Tojottamaselkä basement gneiss dome in the east. The uppermost part of the Rookkiaapa Formation comprise a thin but extensive sheet of arkosic metasediment, probably also volcanoclastic in origin.

A unit of high-Mg basalts belonging to the Vajukoski Formation of the Onkamo Group (see Manninen et al. 1995), occurs only sporadically along the strongly tectonized western margin of the Koitelainen layered intrusion. Psammitic and pelitic metasediments of the Postojoki Formation, belonging to the Sodankylä Group, dominate the area west of the intrusion. The predominantly psammitic rocks include light or brownish arkose quartzites, greyish to greenish sericite quartzites and minor orthoquartzites, which generally exhibit a distinct laminar bedding and a small-scale cross bedding. Quartzites turn gradually upwards into sericite schists and aluminous schists with abundant porphyroblasts of various Al-silicates. The uppermost section of the

metasedimentary sequence, the Matarakoski Formation of the Savukoski Group, comprises variably thick beds of phyllites and black schists with tuffitic interlayers.

The uppermost lithology of the area is represented by the Peuralampi and Peurasuvanto Formations of the Savukoski Group, composed entirely of Fe-tholeiites and basaltic to peridotitic komatiites, respectively.

Based on similarities in geotectonic environment, geochemical composition, and some earlier age determinations, the Salla and Onkamo Groups (the former Lower Lapponi of Lehtonen et al. 1992) representing the lowermost metavolcanic units of the Central Lapland greenstone belt have provisionally been correlated to the Sumian and Sariolan Groups of northern Karelia and Kola Peninsula in Russia (Manninen 1991, Lehtonen et al. 1992, Manninen & Huhma 2001, *this volume*). However, the limited number of datings from these rocks has prevented the correlation of lower-rank stratigraphical units between separate greenstone belts of the inferred intracratonic rift zone.

U-PB GEOCHRONOLOGY

Gneisses

A **tonalitic gneiss (A303)** was taken from the only known outcrop of the Tojottamanselkä gneiss dome, and represents the same gneiss unit from which the 3.1 Ga zircon ages have been reported by Kröner et al (1981). The distance to the large 2.44 Ga Koitelainen layered intrusion is less than 1 km. The strong deformation and granitization have metamorphosed the tonalitic protolith into a reddish-brown augen gneiss with microcline porphyroblasts in a fine to medium-grained, foliated groundmass.

Zircons are euhedral, pale brown and elongate with somewhat rounded edges. Eight zircon fractions have been analyzed, but they do not plot on a chord in the concordia diagram, suggesting heterogeneity due to effects of metamorphism and/or inheritance (Fig. 2). Nevertheless, the data are compatible with the previous estimates of ages above 3 Ga. The oldest $^{207}Pb/^{206}Pb$ age of 3044 Ma for an individual analysis should be considered as a minimum. The Th/U in the zircons, inferred from the radiogenic $^{208}Pb/^{206}Pb$, is relatively low in this sample.

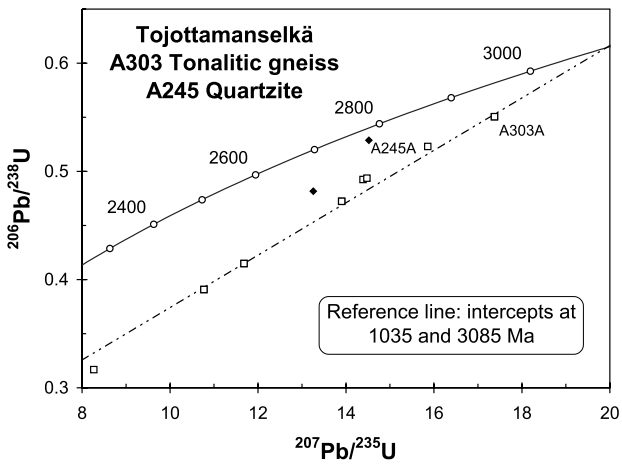

Fig. 2. U-Pb concordia diagram for the tonalitic gneiss (A303, open squares) and quartzite (A245, solid diamonds) from Tojottamanselkä.

The **quartz-feldspar gneiss (A157)** has been taken from a large, in situ boulder field at Kaunismännikkö c. 15 km NW of the Koitelainen intrusion. It is composed entirely of reddish or buff quartz-feldspar gneiss, a rock type which generally has been associated with the lithology of the Pomokaira Basement Complex (Pihlaja & Manninen 1988, Pihlaja & Manninen 1993). Although faint streaks are visible in this fine to medium-grained, only moderately foliated rock, the distinct banding, which generally appears in arkose gneisses of the Pomokaira basement complex, is absent. Instead, the geochemical composition of the rock closely coincides with the tuffaceous metasediments of the Rookkiaapa Formation, described around the Tojottamanselkä gneiss dome by Peltonen et al. (1988). Texturally the gneiss is granoblastic and consists mainly of quartz and microcline, with minor amounts of dark green hornblende and albitic plagioclase. Biotite, magnetite, sphene, zircon and apatite are accessory minerals.

The zircon population of the sample appears to be bimodal. The heavy zircon (d>4.6) consists of small (<70 m), short, pale brown, often very clear and euhedral simple tetragonal prisms, whereas in the low-density fraction (d<4.0) crystals are dark brown and cloudy, some grains considerably larger than in the heavy fraction. The nine fractions analyzed follow the general trend: The low density material has high U content and the U-Pb data are very discordant, whereas heavy zircons especially after air-abrasion provide analyses fairly close to concordia (Fig. 3). If all data are regressed together the upper intercept would be 2472±17 Ma, but the high MSWD (92) shows considerable scatter in excess of analytical error. Four analyses of the heavy zircon yield an upper

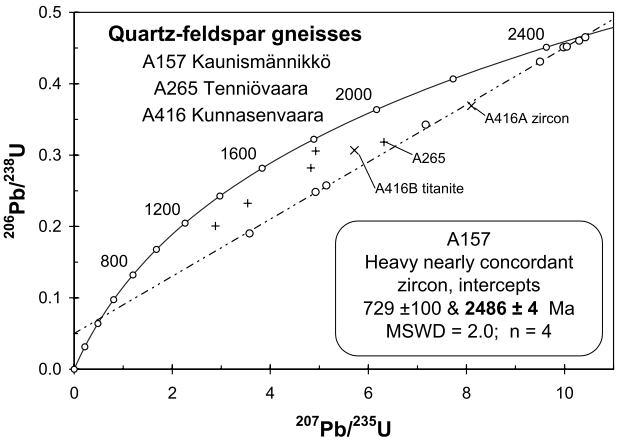

Fig. 3. U-Pb concordia diagram for the quartz-feldspar gneisses A157 (circles), A265 (crosses) and A416 (Xs).

intercept age of 2486±4 Ma and a lower intercept of 729±110 Ma (MSWD=2, the HF-pretreated fraction B has been omitted). Considering the quality of zircon and the near concordance of the analyses, this result should be taken as an age of the crystallization of magmatic zircon and the rock itself.

Sample **A265** is from a light brown, moderately foliated **quartz-feldspar gneiss**, known as the Tenniövaara gneiss, which crops out near the southern margin of, or as inclusions in, the ~1.77 Ga old Nattanen-type Pomovaara granite (Front et al. 1989). The sampled rock is fine to medium-grained, composed of quartz, microcline, plagioclase and small amounts of biotite, sphene, epidote, allanite, apatite and zircon. The cataclastic deformation of the Tenniövaara gneiss is manifested by feldspar and quartz megacrysts, 5-10 mm in size, while the prevalent homogeneous, granoblastic texture of the rock suggests an orthogneiss origin.

The sample yielded abundant subhedral to euhedral, brown and fairly turbid zircon. The grain size varies considerably and also elongate crystals exist. The general appearance does not suggest a detrital origin. Nevertheless, the five analyses scatter and no real age estimates are possible. The $^{207}Pb/^{206}Pb$ ages for the discordant analyses range from 1.70 to 2.27 Ga (Fig. 3).

Quartz-feldspar gneiss (A416). The sample has been taken from a reddish or buff quartzofeldspathic gneiss at the eastern summit of the Kunnasenvaara hill, about 9 km NE of sample A157. The fine to medium-grained, foliated gneiss exhibits locally a faint banding, and consists mainly of quartz, microcline, plagioclase and hornblende.

Zircon in this sample is euhedral, stubby and generally brown and turbid. One old borax-fusion analysis (1972) is fairly discordant and has a $^{207}Pb/^{206}Pb$ age of c. 2.45 Ga (Fig. 3). No real age can be determined, but on the concordia diagram the analysis plots close to the data array of the other gneiss sample A157.

An old borax-fusion analysis on titanite is also available on this sample. Although the result is discordant the $^{207}Pb/^{206}Pb$ age of c. 2.17 Ga should be considered as a minimum. The titanite still available from this sample appears to be quite "bad-looking", which makes the evaluation of the result somewhat uncertain.

Supracrustal rocks

Felsic volcanic breccia (A685, A206). Both samples were taken from a felsic volcanic breccia interlayer of the volcanic conglomerate member of the Rookkiaapa Formation at Sadinoja (see Peltonen et al. 1988, Fig. 4) close to the contact of the Archean Tojottamanselkä dome. The observed length of the breccia layer is c. 3 km, while its thickness is unknown. The pyroclasts consist mainly of dacitic to rhyolitic quartz porphyry and andesitic amygdaloidal lava fragments ranging from lapilli size to angular or subangular blocks up to 10-15 cm in diameter. The matrix of the breccia is composed predominantly of a rhyolitic crystal tuff, but intermingled epiclastic material may also occur.

Zircon separated from sample A685 consists mainly of anhedral to subhedral, small, grey, stubby crystals, which contain dark inclusions mainly on the surfaces. Three U-Pb analyses made in 1985 on this sample are relatively little discordant, and provide $^{207}Pb/^{206}Pb$ ages of 2.51 to 2.53 Ga, but show heterogeneity in excess of analytical error (Fig. 4).

The U-Pb zircon analyses on the other Sadinoja

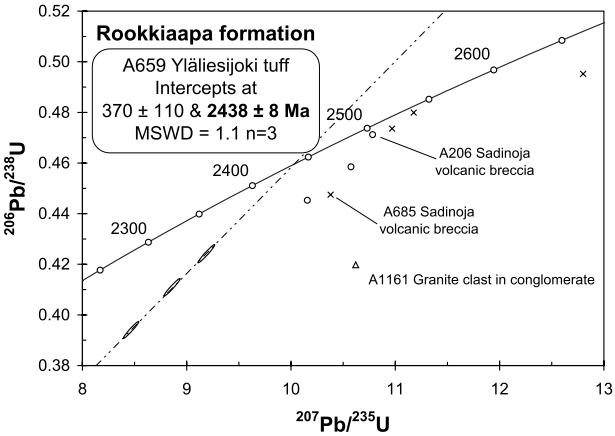

Fig. 4. U-Pb concordia diagram for samples A206 (circles), A685 (Xs), A659 (ellipses) and A1161 (triangle) from the Rookkiaapa formation.

sample, A206, yield also heterogeneous data with slightly older $^{207}Pb/^{206}Pb$ ages. Dark zircon crystals give an age estimate >2.7 Ga, and it is questionable whether any of the analyzed fractions represent homogeneous zircon populations. Although some of the data lie very close to the concordia, reliable age determination is difficult from these conventional multigrain analyses.

The felsic tuff sample A659-Yläliesijoki was taken from a small outcrop of a light grey, foliated felsic metavolcanic rock, which corresponds texturally and geochemically to the felsic tuffs and lavas of the Rookkiaapa Formation. The sampled outcrop occurs within a strongly tectonized zone close to the western (upper) contact of the Koitelainen layered intrusion, the assumed contact between the intrusion and the country rocks lying about 200 m to the east of the sample site. The felsic tuff is fine-grained, consisting of small albite phenocrysts or crystal fragments in a well-orientated quartz-feldspar groundmass. Subeuhedral phenocrysts of plagioclase are 0.5-2.5 mm in diameter, varying from resorbed to fractured and angular in shape. The matrix is composed of recrystallized grains of quartz and plagioclase with epidote, magnetite, chlorite, biotite, titanite and zircon as accessory minerals.

The sample yielded abundant zircon, which is anhedral to subhedral, brown, turbid and variable in grain size. The heavy fraction contains a lot of rutile. Three fractions analyzed in 1984 define a chord with an upper intercept age of 2438±8 Ma, and lower intercept at 370 Ma (Fig. 4). This is similar to the age of 2439±3 Ma determined for the Koitelainen layered intrusion (Mutanen & Huhma 2001, *this volume*).

Volcanic conglomerate (A1161). Clasts of conglomerate have been successfully used for constraining the age of deposition (e.g. Rastas et al. 2001, *this volume*). Sample A1161 represents selected microcline granite clasts from a large outcrop of the volcanic conglomerate member of the Rookkiaapa Formation (see Peltonen 1986, Peltonen et al. 1988). In addition to the sampled granitic clasts, the polymictic clast-supported conglomerate also locally contains boulder-size clasts of granite gneisses and TTG gneisses as well as various types of the metavolcanic and metasedimentary rocks set in an intermediate tuff or tuffitic matrix. The stratigraphic position of the probably thin but areally extensive (>10 km^2) conglomerate is unclear. However, the pebbles of intermediate lavas and tuffs, existing as subrounded or rounded clasts within the conglomerate correspond chemically to the andesitic rocks of the Rookkiaapa Formation, and suggest a rather high intraformational position (Peltonen 1986). The sampled clasts are well-rounded and consist of reddish, medium grained, massive or only weakly foliated microcline granite.

Zircons from the granite clasts are red-brown euhedral crystals. Only one analysis has been made, which is discordant and provides Archean $^{207}Pb/^{206}Pb$ age (Fig. 4). As the aim of dating was to constrain the age of deposition, such an old age does not provide much help.

The **quartzite sample A245** was taken from a local block field of the arkosic quartzite, which lies south of the Tojottamanselkä gneiss dome. The lithostratigraphic position of the light brown, distinctly laminated quartzite is unclear; its occurrence at the contact of the Archean gneiss dome suggests it belongs to the metasedimentary basement cover.

Zircons from this sample are short and variably rounded. The two analyses made have $^{207}Pb/^{206}Pb$ ages above 2.8 Ga (Fig. 2). Analysis on the rounded clear crystals has a fairly low U-content.

Intrusive rocks

Hornblende gabbro (A66). In the Pitkännuottionpulju area a dark green, medium-grained hornblende gabbro intrudes the volcanic conglomerate member of the Rookkiaapa Formation (see Peltonen 1986, 1988). The undeformed gabbro occurs as rather small, separate sills and bodies, consisting mainly of hornblende, plagioclase (An_{55}), scapolite, biotite and magnetite, while sericite, epidote and zircon are accessory minerals. Locally, a narrow chilled margin against the conglomeratic host rock can be noticed. Sample A66 was taken from a pegmatoid portion of

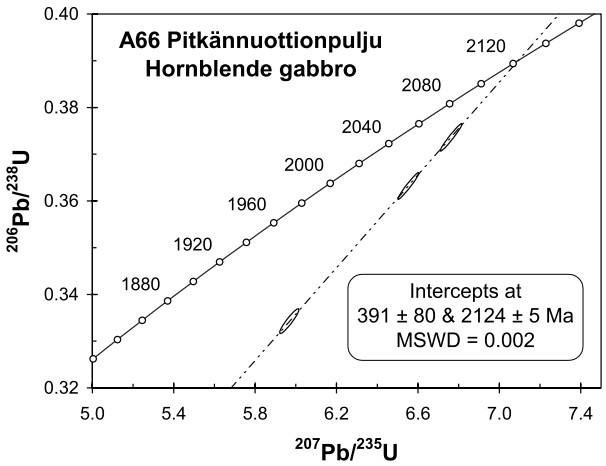

Fig. 5. U-Pb concordia diagram for Pitkännuottionpulju gabbro.

the gabbro.

The three fractions analyzed yield an upper intercept of 2124±5 Ma, which can be considered as an age of crystallization of the gabbro (Fig. 5).

DISCUSSION

This paper is a compilation of conventional multigrain U-Pb data analyzed over a long period of time (results summarized in Table 1). Although the approach provides often heterogeneous data, and more sophisticated methods would be needed for high-precision dating, the results available are useful in constraining the geological history of the Peurasuvanto area.

The Archean basement gneiss domes of the Peurasuvanto area comprise a structurally and lithologically diverse rock sequence ranging from variably granitized ortho- and paragneisses to mafic and felsic metavolcanic rocks. Except of the small Tojottamanselkä dome in the middle of the study area, where an age of c. 3.1 Ga has been obtained for the tonalitic gneisses with multi-stage geological history (Kröner et al. 1981, Jahn et al. 1984, Kröner & Compston 1990, sample A303, this study), the geochronology of the rocks belonging to the basement complex is poorly known. Because all features which may indicate the nature of their protolith have been obliterated by granitization, migmatization and penetrative deformation, only a thorough geochemical study might separate e.g. psammitic metasediments from felsic metavolcanic rocks. No attempts for a detailed stratigraphic subdivision for the rocks of the Pomokaira basement complex have been made so far.

Light reddish or brownish, distinctly foliated, fine- to medium-grained quartzofeldspathic gneisses, called granite gneisses by Mikkola (1941), dominate the Pomokaira basement complex around the Nattanen-type granites to the west of the study area. Similar gneisses have been described by Meriläinen (1976) from the West Inari schist zone, which is a direct continuation for the supracrustal rocks of the Pomokaira basement complex along the Lapland Granulite Complex to the northwest. Based on the dacitic to rhyolitic chemical composition of the quartz-feldspar gneisses, Hörmann et al. (1980) interpreted these rocks as magmatic in origin.

On the geological maps published recently from the Peurasuvanto area (Pihlaja & Manninen 1993, Lehtonen et al. 1998), the gneisses from Kaunismännikkö (A157) and Kunnasenvaara (A416) are assigned to the basement gneiss complex. The U-Pb data obtained suggest, however, that these rocks might equally well be part of the Paleoproterozoic supracrustal sequence covering the Archaean basement. This also contradicts the assumed Archean age of some other lithological units of the Pomokaira basement complex, including the mafic metavolcanic rocks and amphibolites of the Tankajoki Suite (cf. Lehtonen et al. 1998).

The emplacement of the mafic layered intrusions of the Fennoscandian Shield is reasonably well dated using U-Pb on zircon (Alapieti 1982, Amelin et al. 1995, Mutanen 1997, Mutanen & Huhma 2001, *this volume*). Although closely associated Paleoproterozoic supracrustal rocks are known on the Kola Peninsula (Bayanova & Balashov 1995) and Karelia (Amelin et al. 1995), isotopic data from the coeval volcanic sequence are scarce. The U-Pb zircon ages obtained from the felsic metavolcanic rocks of the Peurasuvanto area help to estimate the time and, indirectly, also the volume of the volcanism on the Archean-Proterozoic boundary. They also provide a fairly reliable tool for the correlation of metavolcanic rocks of the Rookkiaapa Formation with other lithostratigraphical units in other Paleoproterozoic greenstone belts of the Fennoscandian Shield.

According to the new stratigraphic division and nomenclature of Lehtonen et al. (1998), the Rookkiaapa Formation represents the lowermost lithologic unit within the Sodankylä schist belt of the Paleoproterozoic Central Lapland greenstone belt. Apart of the study area, metavolcanic rocks with similar geochemistry and lithostratigraphic position have been described from the Salla schist belt, northeastern Finland (Manninen 1991, Manninen & Huhma 2001, *this volume*).

The U-Pb data from Sadinoja are heterogenous, reflecting partly xenocrystic zircon from pre-existing older crust either as detrital component or from the depth. Thus the best age estimate for the Rookkiaapa Formation in the Peurasuvanto area is the 2438±8 Ma provided by the Yläliesijoki sample. Consequently, the volcanism would be roughly coeval with the emplacement of the Koitelainen and Akanvaara layered intrusions (Mutanen & Huhma 2001, *this volume*). This conclusion is also supported by the U-Pb zircon result for felsic extrusive rocks occurring at Sakiamaa some 45 km east from Peurasuvanto, where an age of 2438±11 Ma has been recorded (Räsänen & Huhma 2001, *this volume*).

A similar close temporal correlation of c. 2.44 Ga old layered intrusion and rift-related volcanism has

Table 1. Summary of U-Pb determinations from the Peurasuvanto area.

Sample	Map	Northing	Easting	Location	Mineral	Rock type	Age (Ma)	N	Comment
A66	372310	7527.45	3492.88	Pitkännuottionpulju	zircon	Hornblende gabbro	2124±5	3	
A157	372308	7535.60	3483.35	Kaunismännikkö	zircon	Quartz-feldspar gneiss	2486±4	9	intercept for 4 heaviest fractions
A206	372310	7525.80	3492.70	Sadinoja	zircon	Volcanic breccia	Archean	4	heterogeneous; may be xenocrystic
A245	372310	7523.80	3494.80	Tojottamanselkä	zircon	Quartzite	Archean	2	zircons c. 2.8 Ga
A265	372304	7527.83	3475.08	Tenniövaara	zircon	Quartz-feldspar gneiss	Proterozoic	5	heterogeneous, 207/206 ages 1.8-2.3 Ga
A303	372310	7523.87	3494.78	Tojottamanselkä	zircon	Tonalitic gneiss	Archean	8	heterogeneous, discordant, >3 Ga
A416	372309	7543.86	3487.60	Kunnasenvaara	titanite	Quartz-feldspar gneiss	Proterozoic	2	discordant, 207/206 age c. 2.17 Ga; zircon 207/206 c. 2.45 Ga
A659	372310	7526.18	3491.67	Yläliesijoki	zircon	Felsic tuff	2438±8	3	
A685	372310	7525.18	3492.68	Sadinoja	zircon	Volcanic breccia	Archean	3	heterogeneous; may be xenocrystic
A1161	372310	7527.35	3492.75	Rookkiaapa	zircon	Granite clasts in conglomerate	Archean	1	discordant

been documented in various greenstone belts in the Fennoscandian Shield (Amelin et al. 1995, Bayanova & Balashov 1995). In Russia, the lowermost Paleoproterozoic supracrustal rocks have been traditionally assigned to the Sumian Group (e.g. Sharkov & Smolkin 1997). In the Pechenga-Varzuga Belt of the Kola Peninsula, a subvolcanic granophyre of the Seidorechka Formation in the upper part of the Sumian Group yields U-Pb baddeleyite ages of 2434±15 Ma and 2442±1.4 Ma (Amelin et al. 1995, Bayanova & Balashov 1995), consistent with a concept of coeval volcanism within the separate schist belts of the Fennoscandian Shield. The short time span of the igneous episodes, indicated by the emplacement of layered intrusions and coeval extrusive rocks, has been explained by mantle-plume activity in an extensional geotectonic setting (e.g. Amelin et al. 1995).

CONCLUSIONS

Although frequently heterogeneous, conventional multigrain U-Pb zircon data for gneisses from the Archean basement complex, felsic metavolcanic rocks from its Paleoproterozoic cover, and from a mafic intrusive rock in the Peurasuvanto area, northern Finland, help to clarify the geology of the Archean basement and constrain the history of the volcanism on the Archean-Paleoproterozoic boundary. The nearly concordant zircon age of c. 2.48 Ga for the quartz-feldspar gneiss at Kaunismännikkö and a consistent analysis from Kunnasenvaara indicate that these rocks, interpreted earlier as Archean in age, form rather a part of the Paleoproterozoic basement cover and, concluding from their chemical composition, probably are products of felsic volcanism. These rocks together with the felsic crystal tuff from the Rookkiaapa Formation, dated at 2438±8 Ma, would suggest that volcanism occurred over a time interval of c. 40 million years at the Archean-Proterozoic boundary. Nevertheless, the bulk of volcanism is considered roughly coeval with the emplacement of the 2.44 Ga Koitelainen and Akanvaara layered intrusions. Judging from the age data, it seems likely that the felsic metavolcanic rocks of the Rookkiaapa Formation of the Salla Group are coeval with the Seidorechka Formation of the Sumian Group on the Kola Peninsula, Russia.

The hornblende gabbro, which intrudes the volcanic conglomerate of the Rookkiaapa Formation was emplaced 2124±5 Ma, and thus belongs to the group of ~2.1 Ga old mafic intrusions of Central Lapland.

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Appendix I. Mineral U-Pb data from the Peurasuvanto area.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES / Ma		
Analysed mineral/fraction	weight / mg	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A66-Pikännuottionpulju hornblende gabbro															
A. +4.6	19.8	158.9	67.7	1255	0.1409	0.3736	0.65	6.763	0.65	0.1323	0.15	0.97	2046	2080	2115
B. 4.2-4.6	20.4	511.7	184.8	2186	0.0967	0.3343	0.65	5.968	0.70	0.1295	0.23	0.94	2859	1971	2091
C. 4.2-4.6/HF	21.1	403.9	159.2	5025	0.1174	0.3632	0.65	6.553	0.65	0.1309	0.15	0.97	1997	2053	2109
A157-Kaunismännikkö quartz-feldspar gneiss															
A. +4.6	20.4	161.8	84.1	1516	0.2187	0.4310	0.65	9.501	0.65	0.1599	0.15	0.97	2310	2387	2454
B. +4.6/HF	20.8	145.0	77.6	3264	0.2070	0.4511	0.65	9.978	0.65	0.1604	0.15	0.97	2400	2432	2460
C. 4.0-4.2	17.7	983.0	393.1	2077	0.1883	0.3427	0.68	7.169	0.70	0.1517	0.15	0.97	1899	2132	2365
D. 3.8-4.0/+100	16.6	1352.2	395.5	1237	0.1969	0.2484	0.65	4.921	0.65	0.1437	0.15	0.97	1430	1805	2271
E. 3.8-4.0/-100	15.7	1344.1	412.6	1058	0.2080	0.2575	0.65	5.143	0.65	0.1449	0.16	0.97	1476	1843	2286
F. 3.6-3.8/+200	10.7	1548.6	368.5	753	0.2693	0.1903	0.65	3.575	0.65	0.1362	0.18	0.96	1123	1543	2179
G. +4.6/abr 3.5 h	20.5	141.9	80.3	745	0.2467	0.4521	0.65	10.047	0.65	0.1612	0.15	0.97	2404	2439	2468
H. +4.6/abr 7 h	20.3	138.9	77.4	2016	0.2271	0.4604	0.65	10.300	0.65	0.1623	0.15	0.97	2441	2462	2479
I. +4.6/abr 12 h	12.5	125.7	71.0	1902	0.2308	0.4652	0.65	10.425	0.65	0.1625	0.15	0.97	2462	2473	2482
A206-Sadinoja volcanic breccia															
A. +4.5/abr 2 h	22.0	171.9	94.7	10996	0.1791	0.4736	0.65	10.969	0.65	0.1680	0.18	0.96	2499	2520	2537
B. 4.3-4.6/abr 3 h	19.3	257.9	133.5	11354	0.1710	0.4475	0.65	10.380	0.66	0.1683	0.15	0.97	2383	2469	2540
C. +4.6/abr 2 h	8.6	171.1	95.9	5860	0.1791	0.4799	0.65	11.176	0.65	0.1689	0.15	0.97	2526	2537	2546
D. dark crystals	3.7	663.2	363.7	10274	0.0953	0.4952	0.72	12.801	0.73	0.1875	0.15	0.97	2593	2665	2720
A245-Tojottamanselkä quartzite															
A. 4.3-4.6/clear/rounded	11.5	95.7	61.1	2680	0.1826	0.5287	1.86	14.522	1.86	0.1992	0.17	0.99	2735	2784	2819
B. 4.3-4.6/dark/rounded	6.3	285.2	166.5	1287	0.1612	0.4817	0.91	13.258	0.91	0.1996	0.17	0.97	2534	2698	2823
A265-Tenniövaara quartz-feldspar gneiss															
A. 4.3-4.6/100-200	11.5	353.9	118.0	2317	0.2222	0.2821	0.65	4.826	0.65	0.1241	0.15	0.97	1601	1789	2015
B. 4.3-4.6/-200/HF	10.1	250.5	92.6	4206	0.2743	0.3057	0.65	4.925	0.65	0.1169	0.15	0.97	1719	1806	1908
C. 4.2-4.3/+200	10.4	985.7	273.5	1590	0.2347	0.2325	0.65	3.537	0.73	0.1104	0.42	0.82	1347	1535	1805
D. 4.0-4.2/+200	11.4	1501.6	373.9	1083	0.2762	0.2006	0.65	2.881	0.65	0.1042	0.19	0.95	1178	1376	1699
E. 4.3-4.6/+100	4.4	305.3	116.2	968	0.1782	0.3182	0.65	6.319	0.65	0.1440	0.15	0.97	1780	2021	2276

Appendix I. continued

Sample information / Analysed mineral/fraction	Sample weight / mg	U ppm	Pb ppm	$^{206}Pb/^{204}Pb$ measured	$^{208}Pb/^{206}Pb$ radiogenic	ISOTOPIC RATIOS* $^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%	Rho**	APPARENT AGES / Ma $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A303-Tojottamanselkä tonalitic gneiss															
A. 4.3-4.3/+125/abr 2 h	7.1	419.0	262.7	16088	0.0919	0.5506	0.79	17.378	0.80	0.2289	0.16	0.97	2827	2955	3044
B. <4.3/+200/abr 4 h	15.3	587.8	347.2	6963	0.0838	0.5231	0.65	15.859	0.65	0.2199	0.15	0.97	2712	2868	2979
C. 4.2-4.3/+125	10.5	741.0	391.2	10008	0.0800	0.4723	0.65	13.901	0.65	0.2135	0.15	0.97	2493	2743	2932
D. 4.0-4.2/+100/abr 3 h	6.5	952.6	411.2	6335	0.0749	0.3908	0.65	10.769	0.65	0.1999	0.15	0.97	2126	2503	2825
E. 4.0-4.2/large/abr 1.5 h	1.2	1235.9	431.3	2833	0.0733	0.3168	0.95	8.267	0.95	0.1893	0.15	0.98	1773	2260	2736
F. 4.2-4.3/125-200/abr 8 h	10.5	737.8	406.8	10362	0.0840	0.4925	0.81	14.389	0.82	0.2119	0.15	0.98	2581	2775	2920
G. 4.2-4.3/125-200/abr 10 h	8.8	780.6	432.3	9892	0.0855	0.4937	0.65	14.477	0.65	0.2127	0.15	0.97	2586	2781	2926
H. 4.0-4.3/short/abr 2 h	2.0	816.5	377.1	3562	0.0748	0.4149	0.80	11.682	0.80	0.2042	0.15	0.97	2237	2579	2860
A416-Kunnasenvaara quartz-feldspar gneiss															
A. Total/borax	355.5	600.2	240.1	3875	0.1434	0.3691	0.80	8.101	0.80	0.1592	0.20	0.96	2028	2243	2448
B. Titanite (borax)	1510.3	78.1	44.7	135	0.4935	0.3067	0.65	5.717	0.95	0.1352	0.87	0.50	1724	1933	2166
A659-Yläiesijoki felsic tuff															
A. 4.0-4.2/+100/HF	9.3	1487.5	742.9	6777	0.2011	0.4242	0.70	9.189	0.72	0.1571	0.15	0.97	2279	2357	2424
B. 3.8-4.0/+100	9.4	1762.8	822.1	5764	0.2078	0.3941	0.70	8.467	0.72	0.1558	0.15	0.97	2141	2282	2410
C. 3.8-4.0/+100/HF	8.1	1779.6	864.1	7866	0.2084	0.4109	0.70	8.859	0.72	0.1564	0.15	0.97	2218	2323	2417
A685-Sadinoja volcanic breccia															
A. +4.5	21.3	199.1	105.6	9325	0.1713	0.4585	0.65	10.575	0.67	0.1673	0.15	0.97	2432	2486	2530
B. 4.3-4.5	16.1	275.0	143.8	8287	0.1918	0.4453	0.65	10.157	0.65	0.1654	0.15	0.97	2374	2449	2512
C. 4.3-4.5/HF	21.0	246.7	137.6	14076	0.2052	0.4712	0.65	10.782	0.65	0.1659	0.15	0.97	2488	2504	2517
A1161-Rookkiaapa granite clast in conglomerate															
A. +4.2/-100	5.2	586.0	281.8	737	0.0624	0.4198	0.95	10.620	0.96	0.1835	0.24	0.97	2259	2490	2684

*) Isotopic ratios corrected for fractionation, blank and age related common lead (Stacey & Kramers 1975).
**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = standard error in the mean

Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
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A NEW U-Pb ZIRCON CONSTRAINT FROM THE SALLA SCHIST BELT, NORTHERN FINLAND

by
Tuomo Manninen and Hannu Huhma

Manninen, T. & Huhma, H. 2001. A new U-Pb zircon constraint from the Salla schist belt, northern Finland. *Geological Survey of Finland, Special Paper 33,* 201-208. 2 figures and one table.

A mafic dyke that cuts the volcanic formations in the center of the Paleoproterozoic Salla schist belt in the northeastern part of the Fennoscandian Shield yields a minimum U-Pb zircon age of 2325 Ma. Due to metamorphic effects the available data do not allow an exact dating of the magmatic zircons, but the upper intercept age for the less turbid zircons, 2383±33 Ma, can be considered a best estimate. This result demonstrates that the intermediate to felsic volcanic rocks of the Salla Formation, as well as the siliceous high-Mg basalts of the overlying Mäntyvaara Formation represent the lowermost volcanic units of the belt. On the base of their inferred stratigraphical position, lithology, and chemical composition, these rocks are correlated to the volcanic rocks of the Sumi and Sariola Groups, respectively.

Key words (GeoRef Thesaurus, AGI): absolute age, U/Pb, zircon, dikes, diabase, schist belts, metavolcanic rocks, Paleoproterozoic, Onkamonlehto, Salla, Finland

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INTRODUCTION

The Salla schist belt forms the southeastern extension of the Paleoproterozoic Central Lapland Greenstone Belt (see Lehtonen et al. 1998), comprising a volcanic-sedimentary rock succession in the northeastern part of the Fennoscandian Shield. Divided by the Finnish-Russian borderline, this approximately 100 km long and 40 km wide, NNW-SSE trending belt is compressed between the Central Lapland granite complex in the west and the Archean rocks in the east.

The only geological map covering the entire belt has been compiled by Hackman and Wilkman (1925) in the scale 1:400,000. In their explanation to the map sheet, they also describe ”Kalevian metabasites”, volcanic rocks which form the extensive Salla greenstone area proper in the center of the schist belt (Hackman & Wilkman 1929). The Finnish side of the belt has been partly mapped in a scale 1:100,000 by Lauerma (1967a,b) and Silvennoinen (1982). During the Lapland Volcanite Project in the 1980s, a number of key areas of the belt were mapped in detail, and particular attention was paid to the geochemical properties of the volcanic rocks (Manninen 1991). The lithology and lithostratigraphy of the Russian part

of the Salla(-Kuolajärvi) schist belt has been dealt with by various authors (e.g. Korosov 1979, Kulikov et al. 1980, Zhuravlev et al. 1980, Polehovskij 1985, Voinov and Polehovskij 1985, Kulikov 1988, Korosov 1989, Melezhik 1992, Radchenko et al. 1994). The Mesoproterozoic Salla dyke swarm has been described by Hackman (1914), Väänänen (1965) and Lauerma (1987), who has also reported the age, 1122±5 Ma, for the dyke in his explanatory text for the map sheet (Lauerma 1995, Fig. 1).

Some U-Pb zircon ages for 2200-2210 Ma old metadiabases, occurring in the western part of the Salla schist belt have been published by Lauerma (1982, 1995, Fig. 1). Due to lack of appropriate lithologies, only a few attempts at determining ages of supracrustal rocks in the Salla schist belt have been made, and so far all have failed because of difficulties in separating zircon from the felsic volcanic rocks. In this paper, we report new zircon data on a mafic dyke intruding the volcanic formations in the center part of the belt.

GENERAL GEOLOGY

The Archean basement rocks beneath the Salla schist belt are exposed only on the eastern (Russian) side of the border. Due to it's remote location in the Finnish-Russian boundary zone, the lithology and lithostratigraphy of the area are still rather poorly known and particularly the tectonic setting of the greenstone area proper in the middle of the belt is controversial. According to some Russian geologists (e.g. Kulikov et al. 1980), the Salla(-Kuolajärvi) schist belt forms the northern limb of a large Paanajärvi-Kuolajärvi synclinorium, with the volcanic formations in the central part of the area (the Kuolajarvi Group of Radchenko et al. 1994) representing the uppermost lithostratigraphical unit of the sequence.

Based primarily on the geochemical and geophysical properties of the volcanic formations of the Salla schist belt, Finnish researchers (Manninen 1991, Ruotoistenmäki 1992) prefer a geotectonic model in which the volcanic rocks of the Salla greenstone area proper form an uplifted, rigid block of older (Sumian-Sariolian) metavolcanic rocks in a metasediment-dominated belt, overthrusting and folding a pile of younger (Jatulian) metasedimentary and metavolcanic formations against the basement rocks in the east and southeast. The geochemistry of the carbonaceous rocks found within these volcanic rocks in the Salla and Kuolajärvi area is also consistent with this model (cf. Melezhik 1992).

In his work, Manninen (1991) divided the volcanic rocks of the Salla schist belt into four formations: the Salla Formation (lowermost), the Mäntyvaara Formation, the Tahkoselkä Formation, and the Tuohivaara Formation (uppermost), all separated by metasedi-mentary units.

The Salla Formation consists of a co-magmatic volcanic series ranging from calc-alkaline basaltic andesites and andesites to dacites and rhyolites of tholeiitic affinity. While dominated by intermediate lava flows of variable thicknesses, the apparently subaerial volcanic sequence also includes dacitic and rhyolitic lavas, ash-flow tuffs and ignimbrites. The Salla Formation has been tentatively correlated with the upper part of the Seidorechka Formation of the Sumi Group (e.g. Melezhik & Sturt 1994) in the Imandra-Varzuga belt of the Kola Peninsula, Russia.

The overlying Mäntyvaara Formation is separated from the rhyolitic crystal tuff of the Salla Formation by a weathering crust and a thin layer of metasediments. The subaqueous volcanic succession comprises mainly siliceous high-Mg basalts, basaltic andesites and komatiitic basalts. Based on similarities in lithology, geochemistry, and stratigraphical position above the Salla and Seidorechka Formations, respectively, the Mäntyvaara Formation was compared with the Polisarka Formation (Melezhik & Sturt 1994) of the Sarioli Group in the central part of the Imandra-Varzuga belt.

The volcanic rocks of the Salla and Mäntyvaara Formations are strongly enriched in LREE and other incompatible elements, probably due to crustal contamination. This view is possibly corroborated by Sm-Nd whole rock data, as the $\varepsilon_{Nd}(2400)$ values of the siliceous high-Mg basalts are clearly negative and the T_{DM} consequently range from 2.8 to 3.0 Ga (Huhma et al. 1996).

The tholeiitic basalts of the Tahkoselkä Formation above lie between the clastic metasediments of the Kelloselkä and Matovaara Formations, the former being separated from the underlying Salla and Mäntyvaara Formations by a fault zone. The REE and trace-element distributions of the massive or porphyritic lavas of the Tahkoselkä Formation have MORB-like characteristics, indicating a lack of interaction with sialic crust. The geochemistry and the stratigraphical position of the Tahkoselkä Formation is analogous with the Greenstone Formation III of the Kuusamo schist belt (cf. Silvennoinen 1972)

The ultramafic and mafic pyroclastic rocks of the Tuohivaara Formation represent the uppermost unit in the lithostratigraphical column of the Salla schist belt. The subaqueously erupted peridotiitic komatiites of the formation resemble structurally and geochemically the LREE-depleted komatiites of the Sattasvaara Formation in Sodankylä, Central Lapland (cf. Lehtonen et al.

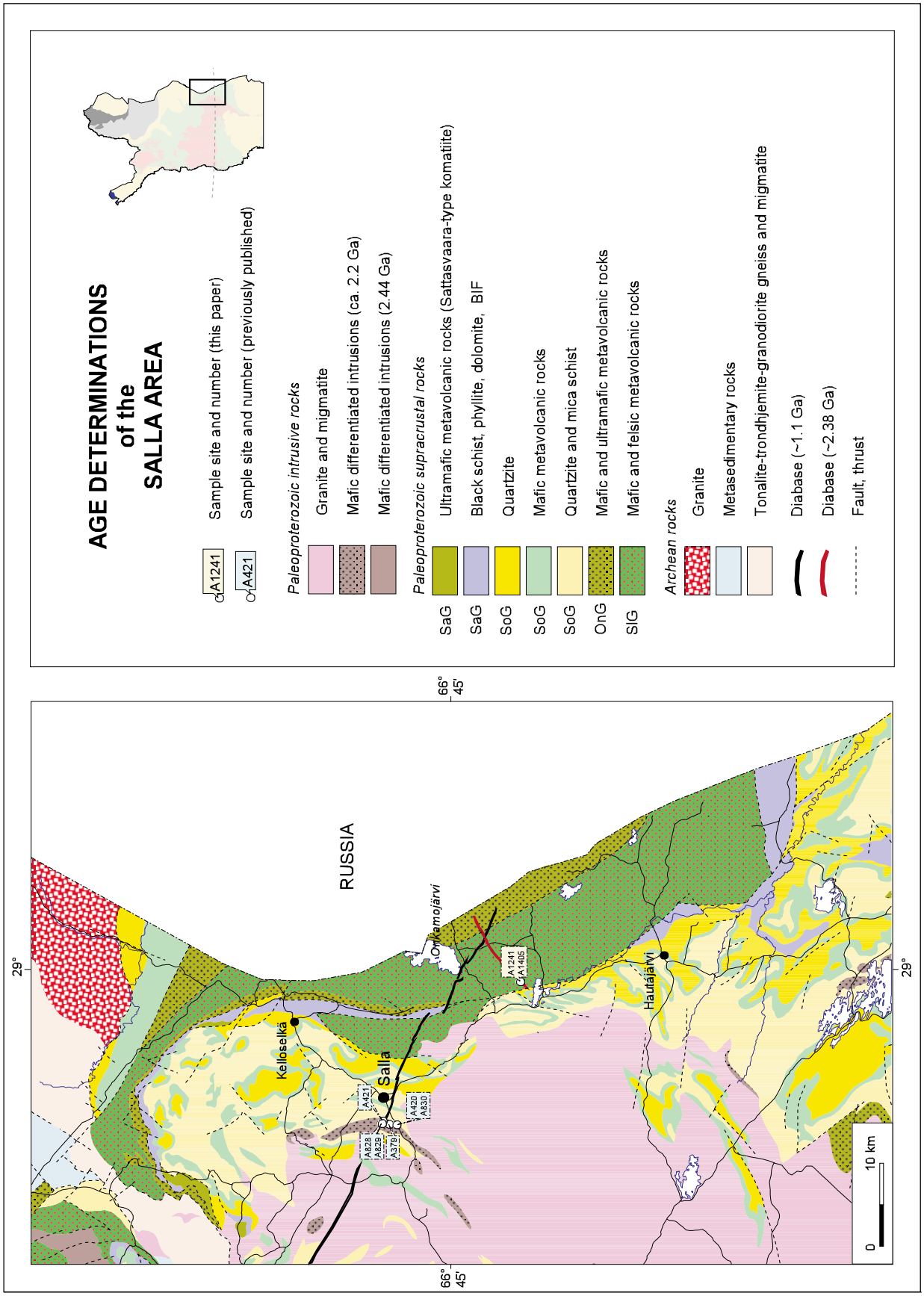

Fig. 1. General geology and age determination sites from the Salla area. For previously published determinations, see Lauerma (1995) and Silvennoinen (1972, 1982 and 1991).

1998).

Except of the strongly deformed marginal zones and some internal thrust fault zones, the volcanic rocks of the Salla greenstone area proper have suffered only moderate pervasive deformation, and exhibit therefore generally well-preserved volcanic textures. The observed fold axes are generally subhorizontal or plunge gently to the north. The prevailing mineral assemblages indicate that the rocks of the study area were metamorphosed under low to medium-grade conditions.

THE ONKAMONLEHTO MAFIC DYKE

A mafic dyke intrudes the andesitic lava flows of the Salla Formation at Onkamonlehto, about two kilometres north of the lake Kallunkijärvi. The eastern portion of the dyke also cuts the siliceous high-Mg basalts of the Mäntyvaara Formation in the Esikkovaara area near the Russian border. The maximum thickness of the subvertical, NE-SW trending dyke is over 150 m, the minimum length being over twelve kilometers. The lowermost part of the distinctly differentiated dyke consists of a bright green, hornblenditic cumulate layer, while the middle part comprises a medium-grained hornblende gabbro. The upper part of the dyke is composed of grey, medium-grained granophyric rock with a magnetite dissemination resulting in a distinct positive anomaly on the aeromagnetic map of the area. The exposed lower contact against the andesitic lavas of the Salla Formation is sharp; the upper contact of the dyke is not exposed. The samples for the age determination A1241 (map sheet 461212, northing 7398.75, easting 4455.60) and A1405 (461212, 7398.70, 4455.60) were taken from the massive, gabbroic part of the dyke (Fig. 1).

RESULTS

Zircon extracted from the original sample A1241 was very turbid, small and anhedral. As turbid zircons tend to yield discordant data and unreliable age estimates, a new sample A1405 was taken. Zircons from this sample are less turbid, pale greyish brown, mostly stubby but also often flaky in appearance. The results for the U-Pb analyses are given in Table 1 and summarized in Figure 2.

The three fractions analyzed from sample A1241 are discordant but form a well-defined chord with intercepts at 2324±10 and 695±25 Ma (MSWD=1, n=3). Seven zircon fractions have been analysed from sample A1405, three of them using the low-blank microchemistry. They are less discordant and yield an upper intercept age of 2383±33 Ma. The lower intercept is highish at 1206±180 Ma. The high MSWD (9) shows that the data points do not plot exactly on a chord. In principle heterogeneous zircon populations in mafic rocks can be produced by either contamination or metamorphism. Zircon morphology do not support crustal contamination and it is considered that the heterogeneity is due to metamorphic (hydrothermal) effects during the Svecofennian tectonothermal pulse at 1.9-1.8 Ga. This is supported by similar results on turbid zircons from other mafic rocks, particularly from 2.2. Ga mafic sills (Perttunen 1985, Silvennoinen 1991, Räsänen & Huhma 2001, *this volume*). The zircon data from the c. 2.44 Ga layered intrusions also occasionally show significant metamorphic effects (Mertanen et al. 1989, Perttunen & Vaasjoki 2001, *this volume*). Consequently, the upper intercept age of 2324 Ma from very turbid zircons of the sample A1241 can also be questioned.

We conclude that the highest $^{207}Pb/^{206}Pb$ age of an individual analysis, 2325 Ma, provides the absolute minimum age for the zircons and the dyke. The upper intercept age of 2383±33 Ma from the less turbid zircons can be considered as the best age estimate at the moment.

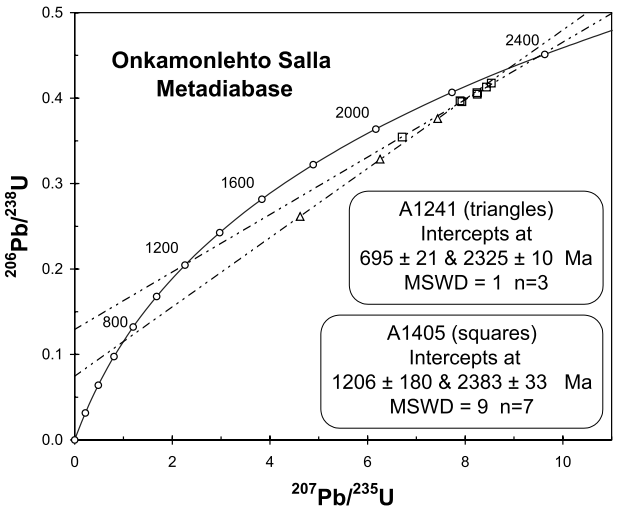

Fig. 2. Concordia diagram for the U-Pb analyses from samples A1241 and A1405 representing the Onkamonlehto metadiabase dyke from Salla.

Table 1. U-Pb data on zircons from the Onkamonlehto dyke, Salla schist belt, Finland.

Sample	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES (Ma)		
Analyzed fraction	Size (mg)	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A1241 Onkamonlehto, metadiabase (461212/7398.75/4455.60):															
A. 3.8-4.0/+200 a2h	0.6	866	273	3410	0.26	0.2616	0.65	4.620	0.66	0.1281	0.3	0.90	1497	1752	2072
B. 4.0-4.2	4.7	695	276	4001	0.26	0.3285	0.65	6.257	0.65	0.1382	0.15	0.97	1831	2012	2204
C. +4.2 a2h	1.2	495	225	9848	0.26	0.3759	0.65	7.437	0.65	0.1435	0.15	0.97	2057	2165	2269
A1405 Onkamonlehto, metadiabase (461212/7398.77/4455.60):															
A. 4.3-4.5	6.2	486	233	8013	0.25	0.3959	0.65	7.934	0.65	0.1454	0.15	0.97	2150	2223	2292
B. 4.2-4.3 a2h	5.1	576	280	8030	0.26	0.3969	0.65	7.894	0.65	0.1443	0.15	0.97	2155	2218	2279
C. 4.3-4.5 a3h	6.2	442	222	15166	0.25	0.4176	0.65	8.535	0.65	0.1482	0.15	0.97	2249	2289	2325
D. 4.0-4.2	7.1	698	302	5426	0.26	0.3544	0.65	6.709	0.65	0.1373	0.15	0.97	1955	2073	2193
E. +4.5	0.18	380	181	15727	0.22	0.4046	0.65	8.249	0.65	0.1479	0.15	0.97	2190	2259	2322
F. 4.3-4.5 a2h	0.36	459	225	13636	0.25	0.4063	0.65	8.241	0.65	0.1471	0.15	0.97	2198	2258	2313
G. 4.3-4.5 a3h	0.64	469	234	17402	0.25	0.4130	0.65	8.436	0.65	0.1482	0.15	0.97	2229	2279	2325

* Isotopic ratios corrected for fractionation (0.15%/amu), blank and age related common lead.
** error correlation between Pb/U ratios
Analyses from small samples (< 1 mg) are made by I.Mänttäri with blanks 5 pg for U and <50 pg for Pb.
The Pb blank for the older analyses (wt>1 mg) is 0.5 ng.
Sample locations in parentheses: (Map sheet/Northing/Easting) Finnish basic map grid.

DISCUSSION AND CORRELATION

Although not revealing the absolute age for the the volcanic rocks of the Salla and Mäntyvaara Formations, the minimum age for the intruding Onkamonlehto mafic dyke, 2325 Ma, sets the lower age limit for the rocks under discussion. The obtained age also conflicts with the proposed synformal structure of the Salla(-Kuolajärvi) schist belt. Instead, a domal model, with the lowermost units of the sequence exposed in the center of the belt, is supported by several facts:

1) The Salla schist belt forms a part of the Karasjok-Kittilä Greenstone Belt (Gaál et al. 1989) or Lapland Greenstone Belt (Sorjonen-Ward et al. 1997), an intracontinental rift-zone with 2.5-2.4 Ga layered intrusions and associated flood basalts in the lowest part of the sequence (see Amelin et al. 1995, Heaman 1997). The overall lithology and observed primary textures within the volcanic rocks of the Salla Formation (see Manninen 1991) are characteristic for the initial stage of the continental, rift-related volcanism (e.g. Green 1983, 1989).

2) The primary volcanic textures of the Salla and Mäntyvaara Formations are generally well-preserved, suggesting a coherent block with the rigid Archaean basement below. Instead, the boundary against the intensely folded (Jatulian) volcanic-sedimentary formations in the west is a strongly tectonized fault zone, clearly recognizable both on the outcrops and on the aerogeophysical maps.

3) Felsic metavolcanic rocks, analogous to those of the Salla Formation, form the floor and roof rock for the 2.44 Ga old Akanvaara layered intrusion in the Savukoski area, about 20 km to the northwest of the Salla schist belt (Mutanen 1997).

4) The volcanic rocks of the Salla and Mäntyvaara Formations exhibit strong enrichment in LREE and LIL elements, probably indicating a notable crustal contamination during their ascent trough the sialic crust (Manninen 1991). This is a general phenomenon within all the lowermost volcanic units of the Central Lapland Greenstone Belt (Räsänen et al. 1989, Lehtonen et al. 1998).

5) In northeastern Karelia, the layered intrusions of the Olanga Complex bracket the age of the Sumi-Sariolian volcanic sequence at ~2.45-2.44 Ga (Amelin et al. 1995). A U-Pb age of 2434±15 Ma has been determined for subvolcanic rocks of the Seidorechka Formation, which form the upper part of the Sumi Group in the Imandra-Varzuga belt (Bayanova & Balashov 1995). The geochemical data available from the volcanic rocks of the Sumi and Sarioli Groups (see e.g. Melezhik & Sturt 1994, Sharkov & Smolkin 1995, 1997) is notably alike the geochemistry of the rocks of the Salla and Mäntyvaara Formations, respectively.

According to these data presented, the volcanic rocks of the Salla and Mäntyvaara Formations of the Salla schist belt probably correspond to the volcanic rocks of the Sumi and Sarioli Groups, respectively.

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Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
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U-Pb MINERAL AGE DETERMINATIONS FROM ARCHEAN ROCKS IN EASTERN LAPLAND

by
Heikki Juopperi and Matti Vaasjoki

Juopperi, H. & Vaasjoki, M. 2001. U-Pb mineral age determinations from Archean rocks in eastern Lapland. *Geological Survey of Finland, Special Paper 33*, 209-227. 12 figures, one table and one appendix.

The bedrock in eastern Lapland is divided into the Naruska, Ahmatunturi and Vintilänkaira-Kemihaara granitoid complexes and the supracrustal Tuntsa metasedimentary and Tulppio metavolcanic belts. Traditionally the bedrock in the area has been regarded as Archean, but previous radiometric age determinations are few. New U-Pb data from 16 samples taken mainly from granitoid rocks within eastern Lapland are presented in this paper.

The Lomperovaara tonalite within the Vintilänkaira-Kemihaara complex and syenitic rocks cross-cutting the Tulppio metavolcanic belt have minimum zircon ages of c. 2.80 Ga, and a hornblende-gneiss inclusion within the Lomperovaara tonalite gives the same 2.83 Ga as a tonalite of the Ahmatunturi complex. The tonalites of the Naruska granitoid complex register an age of 2.70-2.75 Ga. The granitic rocks within the same area give ages at 2.70-2.72 Ga and are thus coeval with the tonalites and possibly represent their migmatized counterparts. The 2.57 Ga registered by zircons of a granitic vein cross-cutting the Lomperovaara tonalite cannot be considered a true age.

A quartz-feldspar schist from the Tulppio metavolcanic belt, interpreted as a felsic volcanic rock, contains zircons of Archean origin, which may indicate an age in the 2.8-2.9 Ga bracket.

The minimum ages of the Takatunturi mafic dike cutting the Naruska granitoid complex and the Sulkarinoja granite within the Vintilänkaira-Kemihaara granitoid complex are 1.94-1.90 Ga and demonstrate that magmatic activity occurred in Paleoproterozoic times as well. However, as Archean titanites have been preserved in the granitic gneisses found at Kairijoki and Naruskajoki it seems that the Paleoproterozoic magmatism was of rather local nature.

The results from eastern Lapland support the findings of earlier studies indicating that the continental crust in northeastern Fennoscandian Shield was formed mainly in late Archean times and consists of rock associations generated at slightly different times in different geotectonic environments.

Key words (GeoRef Thesaurus, AGI): absolute age, U/Pb, zircon, titanite, granites, tonalite, syenites, gneisses, diabase, Paleoproterozoic, Archean, Savukoski, Salla, Lappi Province, Finland

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INTRODUCTION

The Archean rocks of eastern Lapland are delimited in the south and in the west by Paleoproterozoic schists and in the north by the rocks of the Lapland granulite belt (Fig. 1). In the east, the Archean formations continue across the Finnish-Russian border to the Kola Peninsula and Russian Karelia. South of the granulite belt, the Archean rocks continue as a narrow belt north-westwards to the Norwegian border.

The first descriptions of the bedrock in eastern Lapland were published by Erkki Mikkola (1941) in the explanatory text of his 1:400,000 maps (Mikkola 1936 and 1937). Mikkola addressed the problems of the age and mutual age relationships of the rocks, and considered the supracrustal Tuntsa-Savukoski formation to be Archean in age and intruded by the surrounding gneissose granites. An actual proof for the Archean age for the rocks in eastern Lapland was obtained in the 1970s, when a granitoid sample from Naruskajärvi (A216), collected by Kauko Meriläinen and dated by Olavi Kouvo, was found to be c. 2.7 Ga old. A sample from the Kairijoki granite (A215, Meriläinen 1976) gave a very discordant result, but nevertheless demonstrated an Archean age also for this granite.

In the 1970s, Rautaruukki Oy carried out an extensive exploration program in eastern Lapland, but isotopic investigations were focused on the Paleozoic Sokli carbonatite intrusion. The 1980s were a period of intensive geological research in eastern Lapland. The Geological Survey commenced the 1:100,000 mapping of eastern Lapland in 1978, Lapin Malmi Oy continued the mineral exploration started by Rautaruukki Oy, the project on Archean ore deposits by the University of Oulu investigated the surroundings of Sokli, and the Lapland Volcanite Project by the Geological Survey carried out detailed studies of the volcanic rocks. Isotopic studies were, however, almost non-existent as the only sample taken for U-Pb determinations was from a mafic dike (A1001) cross-cutting the Archean rocks in the southern part of the area. In 1992 research activities in the North Finland Regional Office of the Geological Survey were reorganized on a project basis, and a two-year project called "Archean bedrock in eastern Lapland" commenced (Juopperi 1994). In 1994 the project "Archean schist belts in eastern Lapland and their exploration potential" continued the research. During these two projects 14 samples for U-Pb age determinations were collected from Archean rocks in eastern Lapland. This paper presents results from 13 of these new samples as well as earlier, yet unpublished data for a cross-cutting mafic dike (A1001) and the samples A215 and A216.

GEOLOGICAL SETTING

In the final report of the project "Archean bedrock in eastern Lapland" Juopperi (1994) divided the Archean area of eastern Lapland into five geological units: the igneous Naruska, Ahmatunturi and Vintilänkaira-Kemihaara granitoid complexes and the supracrustal Tuntsa metasedimentary and the Tulppio metavolcanic belts.

Archean supracrustal rocks

The Tuntsa metasedimentary belt extends from the Savukoski area into Russia, and consists mainly of metasedimentary gneisses called by Mikkola (1941) the Tuntsa-Savukoski formation. Typical rocks within the belt are quartz-feldspar and mica gneisses with their migmatized counterparts. Strong deformation and recrystallization in amphibolite facies conditions have obliterated the primary textures of the rocks which are locally altered into augen gneisses or even-grained granitic gneisses. With increasing metamorphic grade the metasedimentary rocks have also remelted and become heterogeneous gneisses containing diffuse relics of metasediments. Sporadically there occur up to several meters thick felsic dykes which migmatize the gneisses at their contacts. Within the Tuntsa metasedimentary belt there occur also amphibolites and amphibole-chlorite schists which correspond chemically to the rocks of the Tulppio metavolcanic belt. Thin mafic dykes are found in the southern and western parts of the Tuntsa belt. In the northern part of the Tuntsa belt ultramafic bodies and gabbro dykes, possibly representing feeder channels of mafic layered intrusions, are abundant. In the central and northern parts of the Tuntsa belt there are numerous pegmatite dykes typified by an occasionally rather high content of tourmaline. No observations are available from the northern, probably tectonic, contact between the Tuntsa belt and the Ahmatunturi granitoid

complex.

The Tulppio metavolcanic belt consists principally of amphibolites and ultramafic rocks and strikes in an E-W direction north of the Ahmatunturi granitoid complex. In the west and north it is in contact with the Vintilänkaira-Kemihaara granitoid complex. Although primary textures indicating the origin of the rocks are missing, the chemical composition of the amphibolites suggests that they represent Mg- and locally Fe-tholeiitic lavas erupted in a marine environment. They are associated with unmagnetized and strongly deformed amphibole-chlorite schists. These rocks can be interpreted as komatiitic lavas or tuffs. In contrast, the origin of some strongly magnetized ultramafic rocks is debatable. They can be interpreted as cumulates or feeder-cumulates for Archean komatiitic lavas, but may equally well be Proterozoic in age. It is also possible that they are feeder-cumulates of layered gabbros. Occasional amphibole and/or graphite-bearing Al-rich schists and quartz-rich cherty interlayers within the volcanic rocks may be sedimentary or tuffitic deposits between lava beds. Within the Tulppio belt there occur also quartz-feldspar schists (felsic volcanic rocks) and mica schists. The entire rock association has been called the "Tulppio group" in the reports of the project on Archean ore deposits (Virransalo 1985, Piirainen 1985, Kauniskangas 1987) and was divided into several formations. However, reliable lithostratigraphic classification is hampered by poor exposure and lack of primary textures in the strongly folded rocks. The contact between the Tulppio metavolcanic belt and the Vintilänkaira-Kemihaara granitoid complex is apparently intrusive and partially tectonized. The Paleozoic (Kramm et al. 1993) Sokli carbonatite complex has intruded the contact zone.

Archean granitoid complexes

The Naruska granitoid complex lies in the southeastern part of the study area and is bordered in the south by overthrusted Paleoproterozoic schists and in the north by the metasedimentary rocks of the Tuntsa belt. It comprises tonalitic and granitic gneisses, which locally contain metasedimentary and metavolcanic relics. The granitic gneisses may be migmatized tonalites. Mafic dykes cross-cutting the granitoids occur in the northern part of the complex. The contact towards the Tuntsa belt is ambiguous, as gradational, intrusive and tectonic features have all been observed. The Tuntsa metasediments and tonalitic gneisses are migmatized near their contact, which also obscures its nature. As an entity, the contact probably does not represent a boundary between two geotectonic units, but rather demonstrates a transition towards a deeper erosional section within the same unit.

The Ahmatunturi granitoid complex lies north of the Tuntsa metasedimentary belt and is delimited by Paleoproterozoic schists in the west and by the Tulppio metavolcanic belt in the north. In the east, the complex extends across the border into Russia. The principal rocks are moderately but unevenly magnetized tonalitic gneisses. In the southern part of the complex they contain abundant mafic and intermediate rock fragments and relics. Westwards there occur small granite intrusions along the center of the complex and larger massifs occur in the northwest at the Sotatunturi and Lipakka fells. The contact between the Ahmatunturi complex and the Tulppio metavolcanic belt on its northern side is presumably intrusive and partially tectonized.

In the northern part, near the Sauoiva hill close to the Russian border there occurs a granite, which according to Mikkola (1941) differs from other granites and granitic gneisses in eastern Lapland. On the basis of chemical compositions, the Sotatunturi granites within the Ahmatunturi complex and the Kairijoki granite within the Vintilänkaira-Kemihaara complex could represent the same phase of intrusion, as their REE-patterns are similar to that of the Sauoiva granite. In the eastern part, in a rather large area at the Värriötunturit fells, a Sauoiva-type granite migmatizes a tonalite containing amphibolite relics.

The western and northern part of the Archean in eastern Lapland is formed by the *Vintilänkaira-Kemihaara granitoid complex*, which in the north is terminated by the Lapland granulite belt and is elsewhere bordered by the Tulppio metavolcanic belt and Paleoproterozoic schists. The complex is the least known Archean area within eastern Lapland, and the geological interpretation is largely based on Mikkola´s (1936, 1937 and 1941) works. The complex can be divided into two parts, of which the southern part, Vintilänkaira, consists of tonalitic gneisses and migmatizing or cross-cutting granites and pegmatites. Massive and homogeneous granites in the southern part of this area have been interpreted as Proterozoic in age, as they intrude also Paleoproterozoic schists. In contrast, the more northerly, generally foliated and heterogeneous granites have been considered Archean, which is supported by the age determination from the Kairijoki granite (Meriläinen 1976). Some faint relics of metasediments have also been observed in the Vintilänkaira area. The northern, Kemihaara, part of the complex is, according to Mikkola (1941),

dominated by small-grained, schistose and banded gneissic granites, which he called "mylonitic". These are apparently migmatized tonalitic gneisses. According to new observations there are also more coarse-grained and less well foliated tonalites with locally abundant amphibolite inclusions in the Kemihaara area. On the basis of some field observations and low-altitude airborne magnetic mapping, larger remnants of schists resembling rocks of the Tulppio metavolcanic belt also occur. A small felsic intrusive body, which has been classified as a syenite (Vartiainen & Woolley 1974) occurs west of Sokli in the contact of the Tulppio belt at the Marjavaara hill. The same rock occurs probably also elsewhere in the contact zone of the granitoid complex and the metavolcanic rocks between the Marjavaara and Kuttusvaarat hills.

SAMPLE MATERIAL AND ANALYTICAL RESULTS

The sites of the samples are given in Figure 1 and location information as well a summary of the age results are presented in Table 1. The analytical results are given in Appendix I and summarized on concordia diagrams (Figs. 2-12).

The Naruska granitoid complex

A216-Naruskajoki. This sample represents a granitic gneiss on the border zone of the Naruska granitoid complex and the Tuntsa metasedimentary belt. The rock is medium-grained and with clearly orientated potassium feldspar, quartz and partly sericitized plagioclase as felsic constituents. The principal mafic mineral is partly chloritized biotite. Apatite, opaque minerals, zircon and titanite occur as accessories.

Unfortunately, the original borax fusion analyses devoured all zircon and titanite separates, so no optical description is possible. However, the $^{207}Pb/^{206}Pb$ ages for both the zircon and titanite, 2636±11 and 2647±3 Ma (Appendix 1), demonstrate both the Archean nature of the rock and that no significant metamorphic event influenced it at later times.

A1342-Jauratustunturi. The sample was collected from the Naruska granitoid complex and represents a fairly coarse-grained and foliated tonalite, which is pale grey on the weathered surfaces. The rock is characterized by relatively abundant, unevenly distributed biotite. Its zircon is relatively fine-grained (c. 80%<100 µm). The pale brown crystals are somewhat rounded at their edges, and the length to breadth ratio (L/B) has a median of 3. About 10% of the zircons are reddish in color, and these grains were removed by hand-picking.

The analyses for four zircon fractions exhibit rather normal uranium concentrations, although the abraded fraction A is slightly more discordant than the corresponding unabraded fraction B. The results define a reasonably linear trend (MSWD=5.3) which gives an upper intercept age estimate of 2744±25 Ma (Fig. 2). The lower intercept is highish at 1090±170 Ma. If the analytically somewhat dubious (poor Pbic run) fraction D is excluded, the result becomes 2744±5 and 1070±40 Ma (MSWD=1.5).

A1343-Keppervaarat 1. This sample represents tonalites from the migmatized contact zone of the Naruska granitoid complex with the Tuntsa metasediments. The rock is a medium grained, foliated tonalite with a light-coloured weathering surface. In thin section about 0.1 mm wide shearing seams and late hydrothermal alteration are evident. According to field observations, the tonalite brecciates a rock type consisting of biotite, epidote, plagioclase, hornblende and quartz. Both rocks are cut by granitic veins (A1344-Keppervaarat 2).

The majority of the zircon crystals in the tonalite are euhedral, with simple prismatic-pyramidal morphology. The L/B ratio varies from 2 to 4, the median being c. 2.5. The crystals are light brown in color and do not contain any significant internal features apart of the oscillatory zoning typical for magmatic zircons.

The four analyzed fractions exhibit a normal discordancy pattern, i.e. the abraded fraction is the least discordant one and the degree of discordancy increases with increasing uranium contents and decreasing density. The trend defined is linear (MSWD=0.35). The upper intercept age estimate is 2702±5 Ma while the lower intercept is fairly high at 1276±21 Ma (Fig. 3).

A1344-Keppervaarat 2. The granite is from the migmatized contact zone between the Naruska granitoid complex and the Tuntsa metasedimentary belt. The sample is medium-grained, foliated and reddish on the weathered surfaces. It is clearly granitic in composition, but thin section observations suggest it may be a migmatized tonalite. Most zircons in the heavy (d>4.5 g/cm^3) fraction are roundish, while about 30% of the crystals are euhedral. Both varieties contain ill-defined reddish patches which cannot be removed by hand-picking. The lighter fractions consist almost

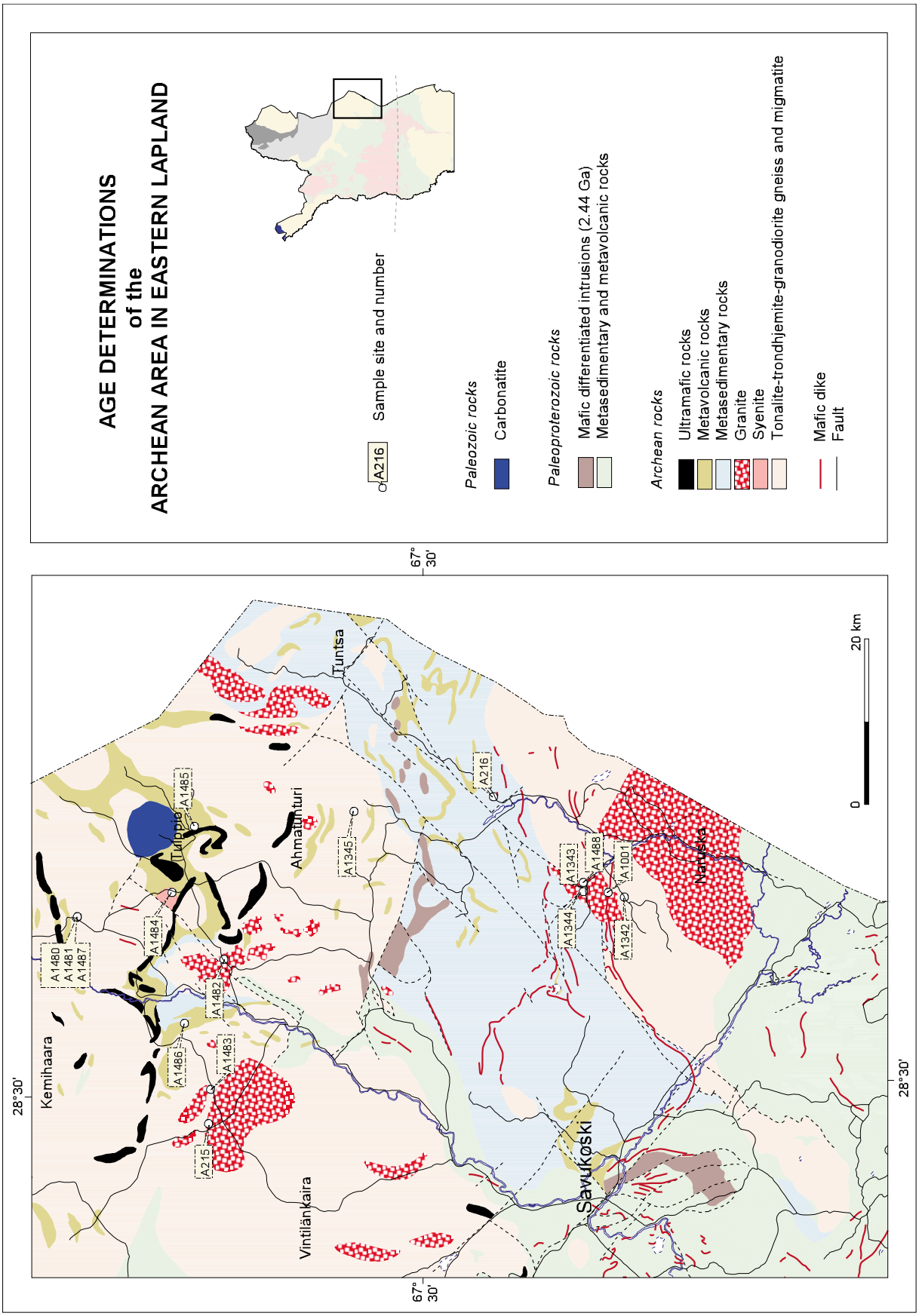

Fig. 1. General geology and sample locations in the Archean part of eastern Lapland.

Table 1. Synopsis of U-Pb mineral age determinations from eastern Finnish Lapland.

Sample	Map	Northing	Easting	Locality	Mineral	Rock type	UI	Err	LI	Err	N	R	Comment
The Naruska granitoid complex													
A216	471306	7480300	4473300	NARUSKAJOKI	Titanite	GNEISS (GRANITIC)	2647	3			1	?	7/6 age; zircon discordant, Archean
A1001	471301	7466080	4460300	TAKATUNTURI SALLA	Titanite	DIABASE (META-)	1888	23			1	?	7/6 age; 4 zircons discordant, Proterozoic
A1342	471110	7464030	4459580	JAURATUSTUNTURI	Zircon	TONALITE	2744	5	1070	40	4	I	1 fraction excluded
A1343	471301	7469250	4460830	KEPPERVAARAT 1	Zircon	TONALITE	2702	5	1276	21	4	I	
A1344	471301	7469310	4460550	KEPPERVAARAT 2	Zircon	GRANITE		-			5	?	Reference line; UI c. 2.70 Ga
A1488	471301	7469220	4461740	SUOLTIJOKI	Zircon	GRANITE	2721	15	1127	56	4	I	MSWD=2.0
The Ahmatunturi granitoid complex													
A1345	471404	7498420	4472280	MUJUVAARA	Zircon	TONALITE	2833	22	1220	230	5	I	MSWD=8.3
A1482	471212	7516160	4454100	RANNINMMAINEN SOTATUNTURI	Zircon	GRANITE	2896	8	911	11	4	?	MSWD=1.7; discordant data
The Kemihaara-Vintilänkaira granitoid complex													
A215	373406	7519000	3559900	KAIRIJOKI SAVUKOSKI	Titanite	GRANITE (PORPHYRITIC)	2541	11			1	NC	4 younger discordant hetrogeneous zircons
A1480	472302	7534940	4460400	LOMPEROVAARA 1	Zircon	TONALITE	2805	4	1299	75	7	I	two younger zircon fractions excluded
A1481	472302	7534850	4460500	LOMPEROVAARA 2	Zircon	GRANITE	2571	59	814	70	6	?	MSWD=8.3; 2 discordant fractions excluded
A1487	472302	7534820	4460490	LOMPEROVAARA 3	Zircon	MAFIC ENCLAVES	2832	6	1291	46	4	I	1 younger fraction excluded
A1483	471206	7518660	4437480	SULKARINOJA	Monazite	GRANITE	1897	7			1	?	7/6 age; 3 zircons heterogeneous, Proterozoic
The Tulppio metavolcanic zone													
A1484	472301	7522430	4463010	MARJAVAARA	Zircon	SYENITE	-				4	?	Reference line; UI c .2.80 Ga
A1485	471406	7519120	4471370	SUURKOVANSELKÄ	Titanite	ALKALI SYENITE	2683	1			1	?	4 discordant zircons; reference line 2.8 Ga
A1486	472107	7521620	4446160	ROVAUKONSELKÄ	Zircon	QUARTZ FELDSPAR SCHIST	-				4	?	Reference line; UI c. 2.85 Ga

I = linear fit; NC = nearly concordant; ? = uncertain estimate

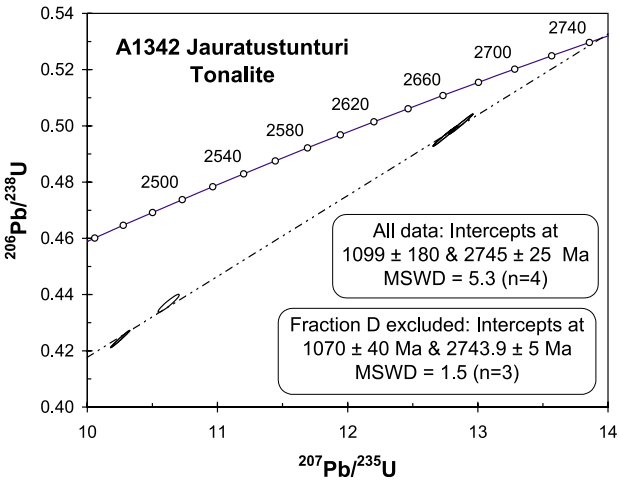

Fig. 2. Results for isotopic U-Pb mineral analyses on from the Jauratustunturi tonalite A1342, Naruska granitoid complex. The ellipses denote the analytical error on the 95% confidence level. Other symbols are used in cases where indicating the error is pointless because of the scale of the figure.

entirely of the euhedral crystal variety.

The analytical results provide a slight surprise: the euhedral crystals (fraction A) contain more uranium and are consequently more discordant than the anhedral zircons (fraction B). The lightest fraction (D) is very discordant, and if it is included in the age calculation, the MSWD becomes very high at 108. If this clearly anomalous fraction is eliminated, the MSWD reduces to a still very high 33 and the intercepts become 2705±48 and 911±290 Ma (Fig. 3).

A1488-Suoltijoki. The sample represents the northern part of the Naruska granitoid complex and was taken in order to cross-check the result from A1344-Keppervaarat. The sample is from a red, medium-grained and strongly deformed granite which consists principally of potassium feldspar, plagioclase and quartz. Accessory constituents are partly chloritized biotite, sericite, titanite, apatite, opaque minerals and zircon. Also carbonate and epidote occur as alteration products of the plagioclase. Microscopically the rock is somewhat fresher than A1344.

The zircons in the sample are generally euhedral with predominating simple prismatic and pyramidal faces. They often appear rather stubby on account of their relatively low (1.5-3) L/B. In the lighter fractions, longer crystals become more prevalent and the L/B may be up to 5 with 3 as a median. Under oil immersion, numerous microcracks are evident. This hampers the observation of inner textures, but occasional oscillatory zoning can be observed.

The four zircon analyses from the Suoltijoki granite are rather discordant, but define a relatively good linear trend (MSWD=2) with intercepts at 2721±15 and 1127±56 Ma (Fig. 4). The discordancy pattern is usual, and rejections of either the most discordant (D) or the coarse (A) crystal fractions do not markedly change the result, which - within error limits - is similar to that from the Keppervaarat granite A1344.

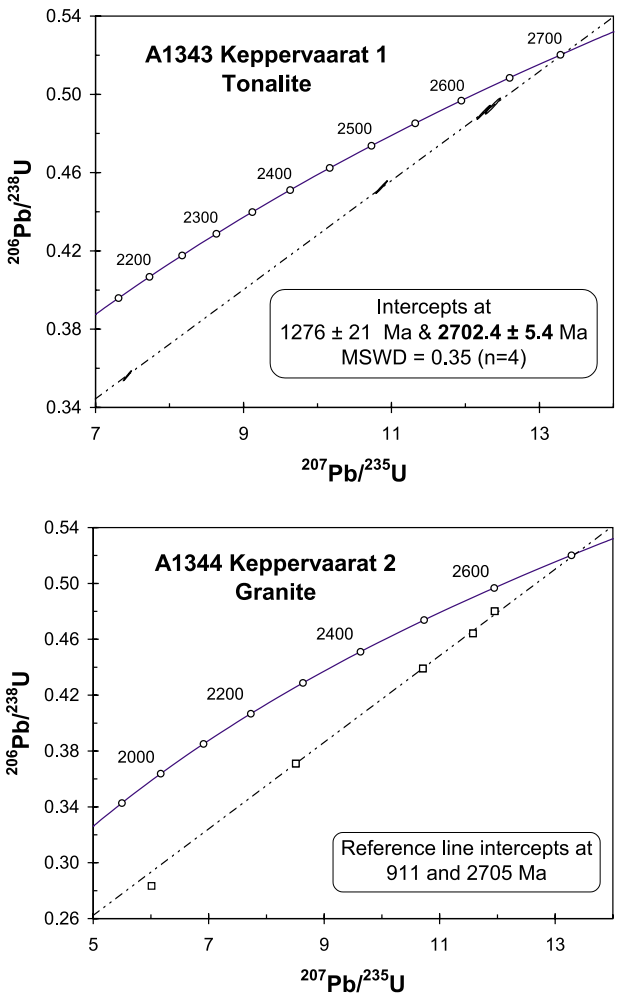

Fig. 3. Results for U-Pb zircon analyses from Keppervaarat, Naruska granitoid complex.

A1001-Takatunturi. On the low-altitude aeromagnetic maps of eastern Lapland there occurs a c. 30 km long, narrow magnetic anomaly caused by

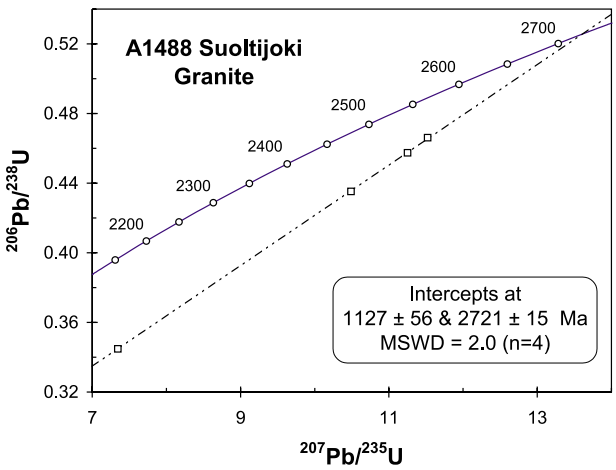

Fig. 4. Results for U-Pb zircon analyses from the Suoltijoki granite, Naruska granitoid complex.

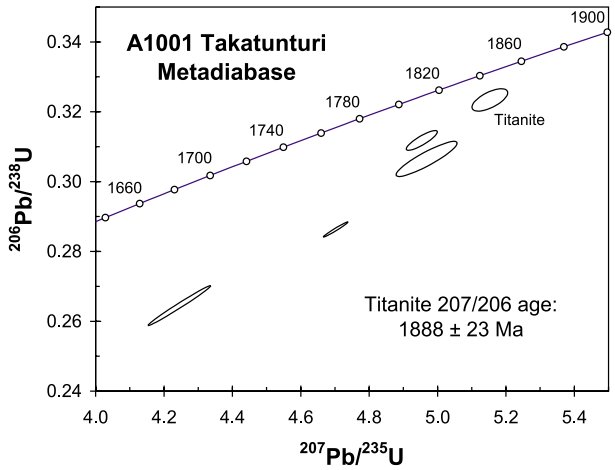

Fig. 5. Results for U-Pb mineral analyses from the Takatunturi mafic dyke intersecting the Naruska granitoid complex.

a 20-50 m wide, foliated and locally schistose dyke intersecting the Naruska granitoid complex.

In thin section, the main mineral is plagioclase, which contains numerous quartz inclusions. Quartz occurs also as larger, evidently primary grains. The principal mafic mineral is hornblende. Accessory minerals are apatite, titanite, zircon and titanomagnetite, which has occasionally been altered into leucoxene.

The zircons in sample A1001 are generally euhedral and only some larger individuals exhibit occasional high order crystal surfaces. The L/B ratio varies from 1.5 to 3, the median being ~2.5. The crystals are turbid and pale brown in color, and larger ones contain erratic whitish patches. The zircon also contains inclusions of a dark, undefined mineral. Both the size and the frequency of the inclusions increase as the density decreases. The translucent, brown titanite is anhedral.

Although the analytical results demonstrate the usual dependence of the degree of discordancy from the uranium contents, the analytical points do not form a linear array on the concordia diagram (Fig. 5). This reflects most probably a combined effect of several geological events, original emplacement and subsequent deformation being the two most likely ones. Thus the highest $^{207}Pb/^{206}Pb$ age observed, 1943±4 Ma for fraction B, must be interpreted as a minimum estimate for the emplacement of the Takatunturi dyke, while the $^{207}Pb/^{206}Pb$ age of the titanite, 1888±23 Ma, indicates the rock's last cooling below c. 500°C.

The Ahmatunturi granitoid complex

A1345-Mujuvaara. The Mujuvaara tonalite lies in the Ahmatunturi granitoid complex north of the Tuntsa metasedimentary belt. The rock is medium-grained, foliated and only slightly altered hornblende-biotite tonalite. The relatively abundant zircon is very fine-grained (90%<70 µm). Most of the crystals are euhedral, pale brown and translucent prisms with a L/B ratio of 3-4. About 30% are more stubby prisms with a L/B ratio <2.5.

The five analyzed fractions exhibit some scatter (MSWD=8) from a linear trend defining intercepts at 2833±22 and 1220±230 Ma (Fig. 6). A closer examination of the data reveals that there may be a slight difference between the two morphological varieties, as the stubby crystals (fraction E) contain less uranium and are less discordant than the longish crystals (fraction D). However, there is no valid reason for eliminating any of the data points from the calculation.

A1482-Rannimmainen Sotatunturi. The sample was collected in order to establish an age for the western part of the Ahmatunturi granitoid complex and to provide simultaneously a minimum age estimate for the supracrustal rocks of the Tulppio suite. It represents a reddish, medium-grained and orien-

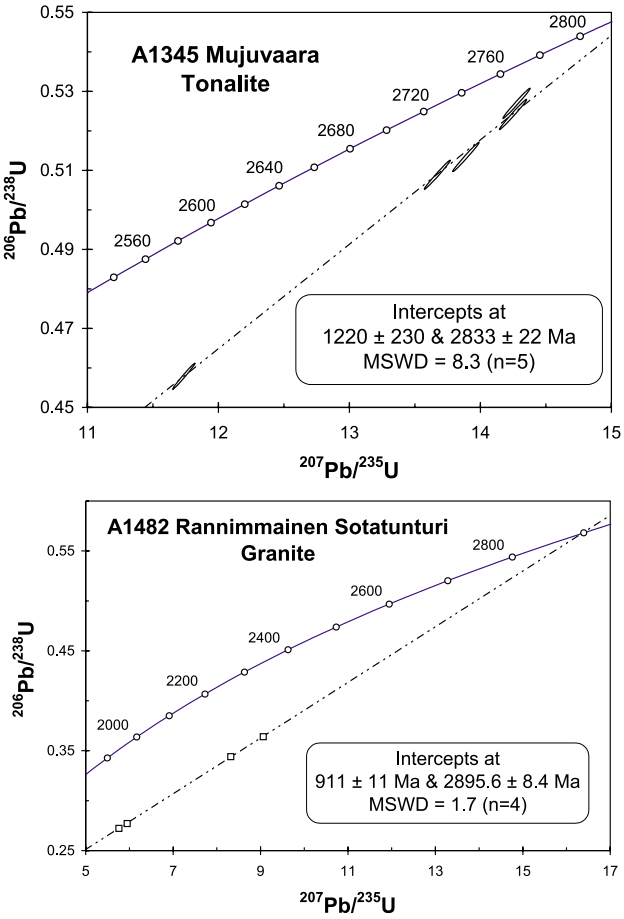

Fig. 6. Results for U-Pb zircon analyses on from the Mujuvaara tonalite and the Rannimmainen Sotatunturi granite, Ahmatunturi granitoid complex.

tated granite. Its major minerals are potassium feldspar, quartz, plagioclase and biotite. Opaque minerals, apatite, epidote and zircon occur as accessory components.

The zircons from the sample exhibit a low density, no fraction heavier than 4 g/cm^3 was found. The crystals are either euhedral or subhedral with simple prismatic and pyramidal faces dominating. The median of the L/B is 2-3. There occur two color varieties: a dominating pale greyish-brownish and a subordinate reddish. Both varieties are turbid to the extent that no internal crystal structures could be observed under oil immersion.

As can be expected from the low density, the zircons have a high uranium contents (800-2000 ppm) and are consequently very discordant. Nevertheless, the four analyses define a relatively good (MSWD=1.7) linear trend with intercepts at 2896±8 and 911±11 Ma (Fig. 6). These fairly good numeric values should be, however, treated with some caution because of the high degree of discordancy of the analytical results.

The Vintilänkaira-Kemihaara granitoid complex

A1480-Lomperovaara 1. This sample is from the northern (Kemihaara) part of the Vintilänkaira-Kemihaara granitoid complex and represents a medium-grained tonalite containing locally abundant amphibolitic inclusions. The major minerals are plagioclase, quartz and biotite. Accessories are potassium feldspar, apatite, titanite, zircon and opaque phases.

The sample contained abundant fine-grained (90%<70 µm), pale brown zircon. In the heaviest (d>4.5 g/cm^3) fraction the crystals are transparent, but the amount of translucent and turbid crystals increases with decreasing density. Under oil immersion two morphological types become evident: a needle-like (L/B c. 5) type with simple prismatic-pyramidal morphology and a stubby type (L/B 1.5-3) which exhibits also higher index crystal surfaces. No significant internal structures such as zonation or inclusions were observed. The crystal edges of both types are somewhat rounded, and consequently the stubby type has an ill-defined appearance under air.

The analytical results demonstrate a normal discordancy pattern in as much as the abraded fractions (A, E) are the least discordant ones. The lightest fraction (D) is abnormally discordant. A closer scrutiny reveals that fraction F representing the stubby zircon variety plots on the "younger" side of the linear trend defined by the other less discordant fractions (Fig. 7). The linear fit for these five fractions (A-C, E, G) is good (MSWD=0.6) and the intercepts are 2805±4 and 1299±75 Ma. In view of the near concordance of the abraded fractions, this may be regarded as a real age.

A1481-Lomperovaara 2. The granite at Lomperovaara occurs as a 10-20 m wide dyke cross-cutting the tonalite (A1480). The rock is somewhat foliated and contains potassium feldspar, quartz and plagioclase as major minerals. Partially chloritized

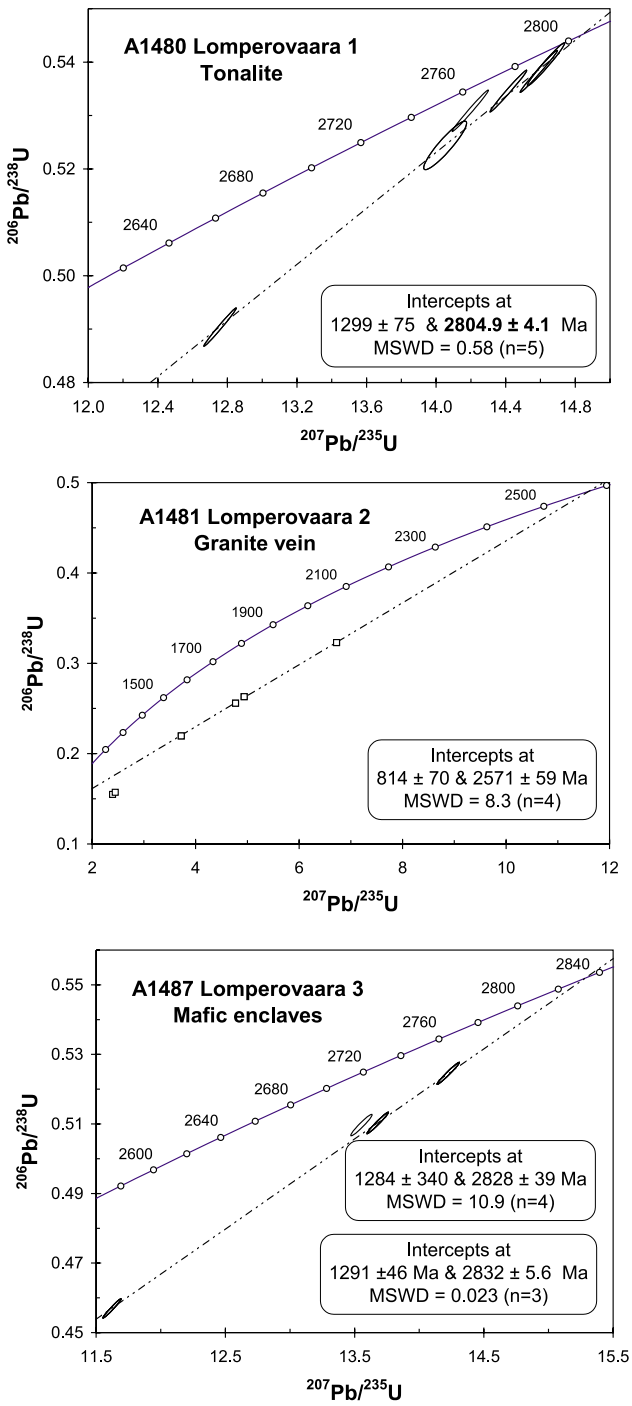

Fig. 7. Results for U-Pb zircon analyses from the Lomperovaara area, Kemihaara-Vintilänkaira granitoid complex.

biotite, muscovite and opaque minerals are present.

The density of the zircon is abnormally low. Because experience shows that light zircons are generally discordant, analyses were carried out on six zircon fractions in order to minimize the error estimate of the age calculation. Under oil immersion, the zircons appeared generally rounded, while characteristic prismatic and pyramidal surfaces are scarce. The crystals were so turbid that internal structures (zonation, inclusions) can not be discerned.

The analytical results demonstrate the correctness of the original assumption. All fractions exhibit high uranium contents, and discordancy increases with increasing uranium contents and decreasing density. Using all six points the upper intercept lies at 2460 Ma, but the MSWD is an extremely high 680. If the two most discordant data points (the 3.6-3.8 density fractions) are omitted from the calculation, the MSWD reduces to 8.3, and the upper intercept becomes 2571±59 Ma (Fig. 7). Even this cannot be regarded as a proper age result, but nevertheless demonstrates an Archean origin for the zircons.

A1487-Lomperovaara 3. The sample represents a greyish, medium-grained and foliated rock, which occurs as inclusions in the Lomperovaara tonalite A1480. Major minerals are plagioclase, quartz, biotite and hornblende, with potassium feldspar as a minor phase. Titanite, apatite, epidote, opaque minerals and zircon form the accessary minerals. Initially these inclusions were interpreted as intermediate volcanic rocks, but microscopic observations suggest that they could as well be partly assimilated amphibolite, or represent an earlier intrusion phase (autolith) of the Lomperovaara tonalite.

The relatively fine grained zircon (85%<70 µm) is under air generally euhedral and longish with L/B at 1.5-4, the median being at 3. In the heavy fraction the crystals are colorless and nearly transparent, but in the lighter fractions brownish hues become more evident and the optical character changes to translucent. In all fractions there occur some anhedral turbid crystals, which most likely arise from very coarse zircon grains broken during milling. Under oil immersion it is evident that most crystal edges of the euhedral longish zircons are somewhat rounded. The heavy fraction does not exhibit any noteworthy internal textures, but the inner parts of some crystals in the lighter fractions are characterized by opaque dust.

Four fractions representing different crystal morphologies were analyzed. The three analyses of long, nearly euhedral zircons (B-D) exhibit a discordancy pattern typical for magmatic rocks and define a precise linear trend (MSWD=0.02), while fraction A representing the coarse anhedral crystal plots off this line. The intercepts for the long zircons lie at 2832±6 and 1291±46 Ma (Fig. 7). The high age for the upper intercept may indicate that the material for A1487 represents xenoliths rather than autoliths, as it is hard to envisage that any part of a magma would have partly solidified and remained geochemically closed for over 30 Ma before its final emplacement.

A215-Kairijoki. Meriläinen (1976) reported a very discordant zircon analysis giving an Archean age estimate for the so-called Kairijoki granite at the Vintilänkaira part of the Vintilänkaira-Kemihaara granitoid complex. In a hand specimen from the collection of the Unit for Isotope Geology, the rock appears to be slightly foliated and greyish in color. Principal minerals are potassium feldspar, plagioclase, quartz an biotite. This sample is problematic as the first author has not been able to locate a corresponding rock type in the field area indicated by the available coordinates. A possible cause for this discrepancy is that when the sample was originally collected in the 1960s, orienteering was done on aerial photographs, and the basic 1:20,000 topographic maps became available only later on.

Three further zircon analyses from the sample are very discordant and scatter in such a manner that attempting a regression analysis would be nonsensical (Fig. 8). An important feature is, however, the titanite analysis, which is nearly concordant and exhibits a $^{207}Pb/^{206}Pb$ age of 2541±11 Ma, which may be interpreted as indicating that no significant thermal event affected the sample since late Archean times.

A1483-Sulkarioja. The present sample is a red, medium-grained and almost unfoliated granite collected less than 10 km east of the location of the Kairijoki granite. Potassium feldspar, plagioclase and quartz are major minerals, accessory phases being

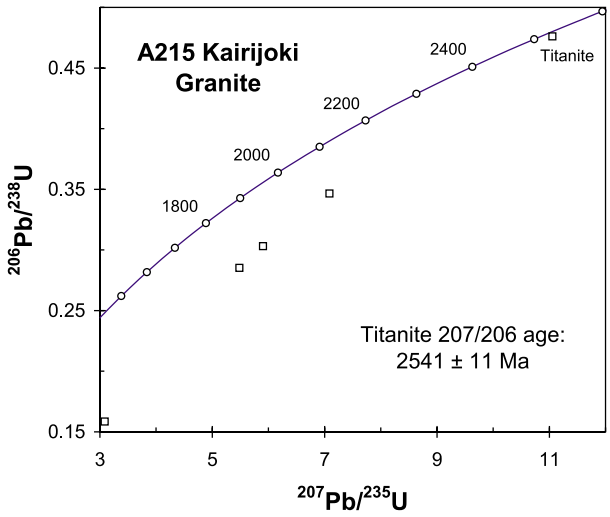

Fig. 8. Results for U-Pb mineral analyses from the Kairijoki granite, Kemihaara-Vintilänkaira granitoid complex.

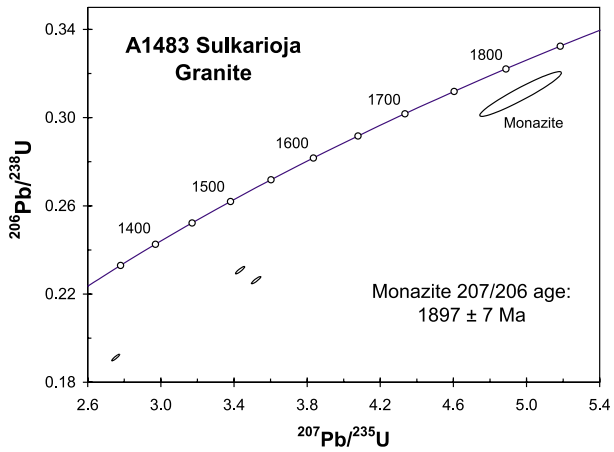

Fig. 9. Results for U-Pb mineral analyses from the Sulkarijoki granite, Kemihaara-Vintilänkaira granitoid complex.

biotite, iron sulfides, muscovite, zircon and monazite. It should be noted that in hand specimen samples A215 and A1483 look totally different.

Sample A1483-Sulkarioja contains two different zircon varieties: about 70% of the zircons are pale brown, translucent, longish (L/B 3-5) and euhedral with simple prismatic-pyramidal morphology. About 30% of the separates are turbid brown and anhedral. The euhedral variety exhibits strong oscillatory zoning, and about 10% contain older, somewhat rounded cores. The monazite is yellowish brown and anhedral.

The analytical results indicate a high uranium contents for all analyzed fractions. Also, the low $^{206}Pb/^{204}Pb$ values demonstrate a relatively high concentration of initial common lead. Apparently the combined effects (initial distortion of the zircon lattice and subsequent metamictization, cf. Vaasjoki 1977, pp. 41-42) of these factors have resulted in a high degree of discordancy. The analyses do not define a linear trend, but all three fractions are clearly Proterozoic in age (Fig. 9). The monazite is slightly discordant, and its $^{207}Pb/^{206}Pb$ age, 1897±7 Ma, gives a minimum age for a cooling through the 520°C isotherm.

The Tulppio metavolcanic belt

A1484-Marjavaara. About 6 km west of the Sokli carbonatite at the contact zone between the Tulppio metavolcanic belt and the Vintilänkaira-Kemihaara granitoid complex there occurs a small syenite stock. The rock is unfoliated with potassium feldspar, hornblende and plagioclase as major minerals. Accessories are titanite, quartz, iron sulfides, apatite, carbonate, epidote and zircon. The hornblende has been dated with the K-Ar method and the result, 1740±35 Ma, was regarded as an intrusion age by Vartiainen and Woolley (1974).

The zircon in sample A1484-Marjavaara is brown and turbid. The zircon crystals have been originally rather large, as they have been crushed during milling, and actual crystal faces (simple prisms and pyramids) are scarce. For these reasons, no observations on internal structures could be done under oil immersion. The radioactivity (corresponding to the uranium contents) of the titanite separated from the sample was so low that there was no point in attempting a U-Pb analysis.

The discordancy pattern of the analyzed fractions is usual, with clear correlations between degree of discordance, uranium contents and density. However, the analyses do not form a proper linear array, which is demonstrated by the high MSWD value of 22. The relatively narrow error estimate (±20 Ma) is a mathematical artefact caused by the wide dispersion of the analytical points on the concordia diagram (Fig. 10).

The data can be interpreted in two ways. Assuming

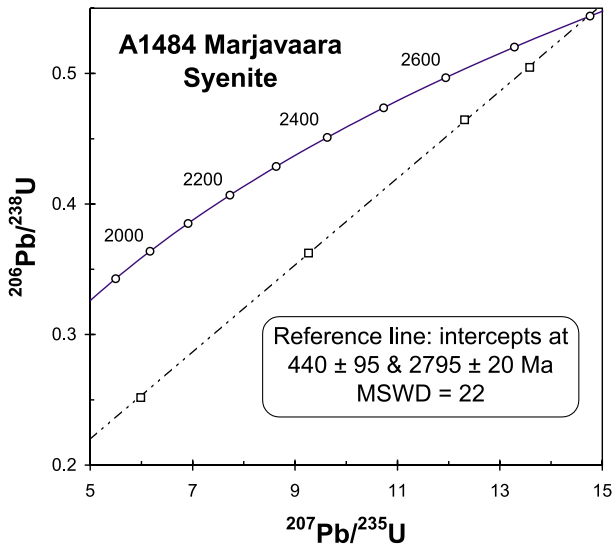

Fig. 10. Results for U-Pb zircon analyses from the Marjavaara syenite, Tulppio metavolcanic belt.

that the K-Ar system of the rock has been reset in Paleoproterozoic times, the upper intercept of the reference line, 2795 Ma, may be considered a minimum estimate for the intrusion of the syenite stock. Alternatively, considering the unfoliated nature of the rock, it may be argued that it is of Paleoproterozoic origin, and the zircons are xenocrystic, reflecting an Archean protolith. As the heavy abraded fraction (A) is relatively little discordant, we favor an Archean age of intrusion.

A1485-Suurkovanselkä. Small-grained, reddish

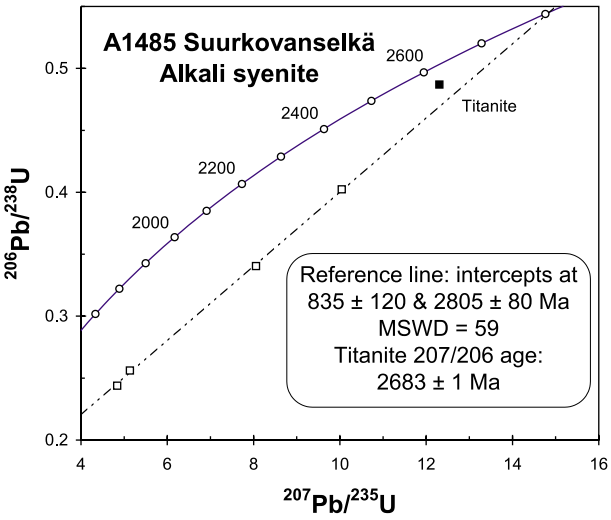

Fig. 11. Results for U-Pb mineral analyses from the Suurkovanselkä alkali syenite, Tulppio metavolcanic belt.

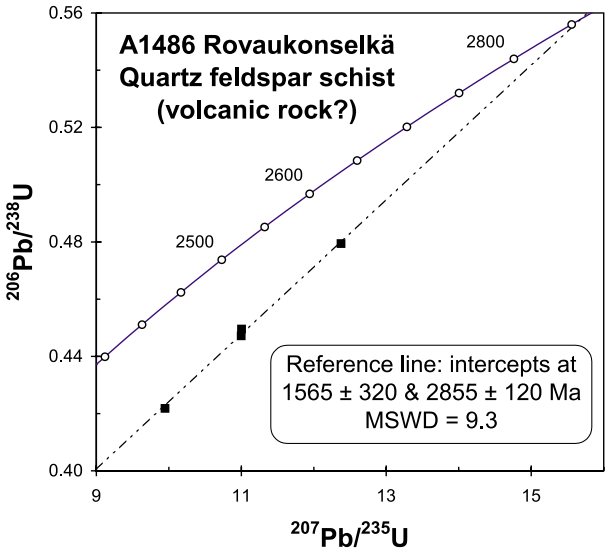

Fig. 12. Results for U-Pb zircon analyses from the Rovaukonselkä quartz feldspar schist, Tulppio metavolcanic belt.

or greyish and slightly foliated rocks occur as occasionally cross-cutting bodies within the amphibolites and ultramafic rocks of the Tulppio suite, and have been regarded as intrusive rocks forming part of the Ahmatunturi granitoid complex. However, according to its mineral composition (potassium feldspar, plagioclase and a pyroxene with greenish pleochroic properties as main constituents and accessory quartz, titanite, epidote, apatite, opaque minerals and zircon) the rock is rather an alkali syenite and could possibly be related to the Marjavaara syenite discussed above.

The zircons from the rock are pale brown, euhedral with simple prismatic faces dominating and rather long, the L/B being typically about 4. Because of the small yield of zircon, no oil immersion study was made, as this always consumes some material. The rock contained also abundantly titanite.

The four zircon fractions are very discordant and exhibit some scatter and thus the intercepts of the reference line (Fig. 11), 2805 and 835 Ma, should be treated with caution. According to weighing, fraction C was only 0.06 mg, but the very high U and Pb contents of the analysis when compared to fraction D suggest a weighing error. However, because of the mixed U-Pb spike employed, this does not affect the Pb/U ratios. The titanite is somewhat discordant, and thus its $^{207}Pb/^{206}Pb$ age, 2683±1 Ma, should be considered a minimum estimate.

A1486-Rovaukonselkä. The rock represents quartz-feldspar schists within the volcanic rocks of the Tulppio Suite, and has been provisionally interpreted as a felsic volcanic rock. The sample is a light-colored schist with plagioclase, potassium feldspar and quartz as the major minerals. Accessory constituents are sericite, biotite, carbonate, apatite, zircon and opaque minerals. The plagioclase occurs partly as somewhat larger, porphyroblast-like grains, and quartz forms in places ball-shaped clusters.

The zircons from this rock are light brown and translucent, generally subhedral grains with a fairly large amount of simple prismatic and pyramidal crystal faces. Under oil immersion, numerous cracks can be discerned and this hampers the detection of other inner textures. However, faint oscillatory zoning and a few euhedral cores have been observed. Fractions lighter than 4.3 consist principally of rutile.

The four zircon fractions analyzed from this rock are rather discordant and exhibit some scatter (MSWD=9), and thus only a reference line with intercepts at 2855 and 1565 Ma is shown on the concordia diagram (Fig. 12). If analysis D, made of the fine-grained zircons is omitted, the other three points lie on a line with intercepts at 2857±18 and 1547±47 Ma (MSWD=0.4). However, this cannot be regarded as a permissible age, as there is no tangible mineralogical reason to exclude the deviating analytical point.

DISCUSSION

The bedrock in eastern Lapland forms part of the Archean crust within the Fennoscandian Shield. Traditionally, Russian geologists have subdivided the evolution in this area into two orogenic phases: the early Archean Saamian and the late Archean Lopian orogenies. Almost all high-grade metamorphic gneiss

and granitoid areas have been regarded as Saamian, on which the younger Lopian formations were supposed to have been deposited (Stenar 1988). The type area for the Saamian formations has been the Belomorian belt extending from the west coast of the White Sea towards north west and encompassing at least part of the Archean bedrock of eastern Lapland.

Gaál and Gorbatschev (1987) subdivided the Archean domain of the Fennoscandian Shield into the Karelian, Belomorian and Kola provinces. They also envisaged two periods of crustal accretion, the Saamian (3.2-2.9 Ga) and the Lopian (2.9-2.6 Ga). According to them, the continental crust formed by the Saamian granitoids originated from pre-Saamian, mantle-derived tholeiitic crust in the course of primitive plate-tectonic processes. The Lopian orogeny (op. cit.) was explained by late Archean plate-tectonic events, which gave rise to the greenstone belts within the Karelia and Kola provinces and the volcanic and sedimentary sequences within the Belomorian province. The latter were interpreted as a sedimentary wedge accreted in a subduction zone, where later metamorphic and magmatic events are related to the collision of the Karelia and Kola provinces.

Geological observations and age determinations in Russia during the last two decades have profoundly changed the concepts on the age of the Saamian formations previously regarded as early Archean. Based on U-Pb determinations on zircons, Bogdanova and Bibikova (1993) concluded that the oldest orthogneisses within the Belomorides were 2.8-2.7 Ga old, and regarded even xenocrystic zircons no more than 3 Ga in age. According to them, metasedimentary gneisses are not significantly older than the orthogneisses. Also, based on the findings of Daly and Mitrofanov (1990), relic volcanic belts within the Belomorides can be correlated with the Lopian greenstone belts within the Karelia province. Bogdanova and Bibikova (1993) interpret the Belomorian belt as a fringe of the Karelian granite-greenstone terrain, where strong tectonic movements caused folding, high pressure metamorphism, formation of anatectic granites and migmatization about 2.65-2.60 Ga ago. According to them, at the onset of Paleoproterozoic times there occurred rifting which facilitated the extrusion of 2.5-2.4 Ga old volcanic rocks and the emplacement of 2.44-2.40 mafic and felsic intrusions. The 1.9-1.8 Ga old overthrust associated with the Svecofennian orogeny then caused structural and metamorphic overprints in parts of the Belomorides.

A regional granulite event, probably resulting from a collision of the Kola and Karelia blocks (e.g. Barbey et al. 1984) has affected a large area in the northern part of the Fennoscandian Shield. Its intensity is demonstrated by recent results from an undeformed garnet-rich quatrzo-feldspathic leucosome at Kolvitsa within the Umba granulite block at the White Sea coast of the Kola Peninsula, where zircons register an age of 1912±2 Ma (Kislitsyn et al. 1999). Bogdanova and Bibikova (1993) also associate the intrusion of granite pegmatites into the Belomorian Belt with this event.

The Sm-Nd isotopic investigations by Daly et al. (1993) and Timmerman and Daly (1995) from the Kola Peninsula and the Belomorides support the concept of a late Archean crustal accretion in these areas. According to Timmerman and Daly (1995) the Sm-Nd T_{DM} ages for Archean granitoids are 2.68-2.96 Ga, 2.86-2.93 Ga for metasediments and 2.77 Ga for felsic metavolcanic rocks. No indication of older crustal material was encountered during these studies. These conclusions are further supported by the results of Bibikova et al. (1996), who concluded on the basis of U-Pb zircon determinations and whole rock Sm-Nd data that the Belomorides had a short crustal prehistory before the intrusion of the tonalitic-trondhjemitic igneous rocks which occurred in two generations at c. 2.8 and 2.74-2.72 Ga ago. It should be noted that the latter finding corresponds to results from Archean rocks in eastern Finland (Vaasjoki et al. 1993 and 1999) where zircons from intrusive granitoid rocks record two major magmatic events at 2.83 and 2.74-2.72 Ga.

The dating results from the Archean area of eastern Lapland presented in this study agree rather well with the results from other Archean areas of the Fennoscandian Shield outlined above. The ages of the granitoids from Jauratustunturi and Keppervaarat (2.70-2.75 Ga) show that the Naruska granitoid complex belongs to the same age group as the orthogneisses of the Belomorian belt on the Russian side and the younger intrusive rocks of the TTG association in eastern Finland. Unpublished Sm-Nd results from mica schists of the Tuntsa metasedimentary belt indicate a late Archean crustal provenance for these rocks.

The age of the Mujuvaara tonalite from the Ahmatunturi granitoid complex is at 2.83 Ga slightly older than zircon U-Pb data from the Belomorides (cf. Bogdanova and Bibikova 1993) and of the same age as the older tonalites in the Puukari area (Vaasjoki et al. 1999). On the other hand, the apparently older result from Rannimmainen Sotatunturi, representing a proper granite of the Ahmatunturi complex, is rather suspect because of the very discordant analytical results. We emphasize that this is a case of two samples yielding rather poor analytical data, and thus the age of the Ahmatunturi complex, apart of falling

into the 2.8-2.9 Ga bracket, must be regarded as unresolved.

In this context the result for the long zircons from A1487-Lomperovaara from the Kemihaara area, c. 2.83 Ga, is also interesting, as it coincides with the result for the Mujuvaara tonalite. It should also be noted that 2.83 Ga ages for granitoid rocks have been recorded on several occasions in rocks surrounding the Kuhmo-Suomussalmi greenstone belt in eastern Finland (Luukkonen 1985, Vaasjoki et al. 1999). As this rock occurs as enclaves in the Lomperovaara tonalite (A1480), which gives a good age at 2805±4 Ma, the results from both the Ahmatunturi and Kemihaara-Vintilänkaira complexes may indicate the existence of somewhat older continental crust in present-day eastern Lapland before the formation of the main late Archean granitoid complexes 2.75-2.70 Ga ago.

Although a Paleoproterozoic age cannot be completely ruled out, a minimum age of c. 2.8 Ga for the Marjavaara syenite is supported by the similar reference line intercept obtained from the admittedly very discordant data representing the Suurkovanselkä alkali syenite. This may indicate that the rocks of the Ahmatunturi granitoid complex are slightly older than those of either the Naruska granitoid complex or the Belomorian belt. Although neither of these results can be regarded as conclusive, they indicate together an age in excess of 2.8 Ga for the Tulppio metavolcanic belt, and this is supported by the data from the Rovaukonselkä quartz feldspar schist, which alone would also be rather inconclusive.

There are also strong indications of early Paleoproterozoic rifting in eastern Lapland. In the northern part of the Tuntsa metasedimentary belt there occur olivine cumulates and gabbros which display Sm-Nd characteristic similar to those of the c. 2.44 Ga layered gabbro intrusions (Huhma, GTK internal report 1997). Felsic intrusive rocks belonging to this age group have not been found so far in eastern Lapland, but volcanic rocks presumably of this age abound in the lower units of the Paleoproterozoic schist belts fringing the Archean area.

The minimum age for the Takatunturi mafic dike, c. 1.94 Ga demonstrates that magmatic activity has taken place in the Tuntsa area also later. The minimum age of the Sulkarinoja granite within the Vintilänkaira granitoid complex (c. 1.9 Ga) corresponds temporally to the overthrusting of the Lapland granulite belt and the emplacement of associated granite pegmatites as suggested by Bogdanova and Bibikova (1993). However, the indisputably Archean titanites found in samples A215-Kairijoki and A216-Naruskajoki suggest that this magmatic activity has not affected the adjacent granitoid rocks on a regional scale.

Thus the present study from eastern Lapland as well as earlier work on the Kola peninsula and the Belomorides show that the continental crust in northeastern Fennoscandian Shield was generated in late Archean times. Together these results demonstrate that the Archean crust consists of a variety of rock associations formed, metamorphosed and deformed at slightly different times in varying geotectonic settings. Gaál and Gorbatschew (1987) suggested that plate-tectonic processes have been involved in the formation of the late Archean crust within the Fennoscandian Shield, and a similar sequence of crustal development has been explained in plate-tectonic terms also in the Superior province of the Canadian Shield (Card 1990). The results of this study, or those of other newer studies do not contradict this concept, but rather support it. However, more work on rock associations, their contact relationships and radiometric ages as well as a better understanding of magmatic, tectonic and metamorphic events are needed in order to establish the exact nature (e.g. plate size and thickness, rate of subduction) of these late Archean plate tectonic processes. Also the impact and intensity of the Svecofennian orogeny at 1.9-1.8 Ga on the then cratonized Archean crust remain to be more thoroughly investigated.

CONCLUSIONS

In summary, it must be noted that much of the data from the present sample material are inconclusive (only 5 of the 16 analyzed samples give definite age information) and are thus subject to a high degree of interpretation. With the reservations discussed above, we conclude that:

1) The ages of the granitoids from Jauratustunturi and Keppervaarat (2.70-2.75 Ga) show that the Naruska granitoid complex belongs to the same age group as the orthogneisses of the Belomorian belt on the Russian side and most other Archean granitoid rocks elsewhere in Finland.

2) The results from both the Ahmatunturi and Kemihaara-Vintilänkaira complexes indicate the existence of 2.80-2.85 Ga old continental crust in present-day eastern Lapland while the formation of the main late Archean granitoid complexes occurred 2.75-2.70 Ga ago.

3) The Marjavaara syenite and the Suurkovanselkä alkali syenite belong most likely to the same intrusive

phase and their most probable age is about 2.8 Ga.

4) As the syenitic granitoids cross-cut the Tulppio metavolcanic belt, they set a minimum age for the latter. The data from the Rovaukonselkä quartz-feldspar schist demonstrate, however, that the metavolcanic complex is still late Archean in age.

5) The minimum ages of the Takatunturi mafic dike cutting the Naruska granitoid complex and the Sulkarinoja granite within the Vintilänkaira part of the Vintilänkaira-Kemihaara granitoid complex are ~1.94 and ~1.9 Ga, respectively, and demonstrate that magmatic activity occurred within Paleoproterozoic times as well. However, as Archean titanites have been preserved in the granitic gneisses found at Kairijoki and Naruskajoki it seems that the Paleoproterozoic magmatism was rather local.

Acknowledgements

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Appendix 1. U-Pb mineral analyses from eastern Finnish Lapland.

Fraction	Weigth (mg)	U conc (ppm)	Pb conc (ppm)	$^{206}Pb/^{204}Pb$ (meas.)	$^{208}Pb/^{206}Pb$ (radiogenic)	$^{206}Pb/^{238}U$	±2 SE (%)	$^{207}Pb/^{235}U$	±2 SE (%)	$^{207}Pb/^{206}Pb$	±2 SE (%)	Corr.	Apparent ages $T_{206/238}$	$T_{207/235}$	$T_{207/206}$
A215-Kairijoki granite															
A. Total/borax	402.8	857.0	150.9	827	0.1165	0.1586	0.60	3.086	0.61	0.1412	0.15	0.97	948	1429	2241
B. +4.2	15.5	685.1	261.2	1346	0.1073	0.3466	0.60	7.085	0.64	0.1483	0.22	0.94	1918	2122	2326
C. Titanite	513.2	144.8	113.8	205	0.5583	0.4761	0.94	11.055	1.20	0.1684	0.64	0.85	2510	2527	2541
D. -4.2	14.3	938.6	300.2	744	0.1214	0.2852	0.60	5.482	0.60	0.1394	0.10	0.99	1617	1897	2219
E. -4.2/abr 2 h	4.6	589.2	196.8	1056	0.1100	0.3031	0.60	5.904	0.61	0.1413	0.10	0.99	1706	1961	2243
A216-Naruskajoki															
A. Titanite/borax	1992.5	57.5	54.0	134	0.8720	0.4550	1.03	11.258	1.05	0.1794	0.24	0.97	2417	2544	2547
B. Total/borax	177.0	254.3	126.4	1026	0.1850	0.4361	0.77	10.717	1.00	0.1783	0.59	0.81	2332	2498	2636
A1001-Takatunturi diabase															
A. 4.3-4.6	15.6	108.7	62.8	1422	0.9865	0.3118	0.74	4.954	0.75	0.1152	0.47	0.86	1749	1811	1883
B. 42.-4.3	3.3	331.6	138.9	2975	0.5634	0.2863	0.62	4.702	0.63	0.1191	0.21	0.98	1623	1767	1943
C. 4.0-4.2	6.5	604.3	257.6	2827	0.4745	0.3065	1.34	4.968	1.47	0.1175	0.64	0.90	1723	183	1919
D. 3.6-4.0	5.2	1026.2	363.3	5891	0.4270	0.2645	1.77	4.245	1.77	0.1164	0.10	0.99	1512	1682	1902
E. Titanite	31.3	12.5	5.9	308	0.3715	0.3234	0.81	5.153	0.83	0.1156	1.28	0.65	1806	1844	1888
A1342-Jauratustunturi tonalite															
A. +4.5/abr 5 h	9.9	157.1	91.1	9249	0.1683	0.4963	0.60	12.747	0.60	0.1863	0.10	0.99	2597	2661	2709
B. +4.5	7.1	154.3	90.4	5773	0.1696	0.5006	0.60	12.870	0.60	0.1865	0.10	0.99	2616	2670	2711
C. 4.3-4.5/+200	7.3	283.5	141.4	4264	0.1844	0.4242	0.60	10.253	0.60	0.1753	0.10	0.99	2279	2457	2609
D. 4.3-4.5/+200/abr 3h	7.1	281.6	146.7	2935	0.1982	0.4369	0.60	10.625	0.63	0.1764	0.22	0.94	2336	2490	2619
A1343-Keppervaarat tonalite															
A. +4.5/abr 3 h	7.2	208.9	119.8	5546	0.1622	0.4940	0.68	12.370	0.68	0.1816	0.10	0.98	2588	2632	2667
B. +4.5	6.3	202.8	115.1	5519	0.1583	0.4908	0.61	12.245	0.61	0.1810	0.10	0.99	2573	2623	2661
C. 4.3-4.5	7.4	295.5	153.8	3172	0.1530	0.4526	0.60	10.867	0.60	0.1741	0.10	0.99	2406	2511	2598
D. 4.2-4.3/+200	7.9	622.9	241.2	1442	0.0917	0.3562	0.60	7.431	0.60	0.1513	0.10	0.99	1964	2164	2361

Appendix 1. continued

Fraction	Weigth (mg)	U conc (ppm)	Pb conc (ppm)	$^{206}Pb/^{204}Pb$ (meas.)	$^{208}Pb/^{206}Pb$ (radiogenic)	$^{206}Pb/^{238}U$	±2 SE (%)	$^{207}Pb/^{235}U$	±2 SE (%)	$^{207}Pb/^{206}Pb$	±2 SE (%)	Corr.	Apparent ages $T_{206/238}$	$T_{207/235}$	$T_{207/206}$
A1344-Keppervaarat granite															
A.+4.5/euhedral	7.3	229.5	114.5	3055	0.1340	0.4391	0.60	10.710	0.60	0.1769	0.10	0.99	2346	2498	2624
B. +4.5/rounded	9.1	165.4	89.6	3975	0.1684	0.4643	0.60	11.576	0.60	0.1808	0.10	0.99	2458	2570	2660
C. 4.3-4.5/100-200	8.0	448.5	188.6	1350	0.1281	0.3710	0.60	8.508	0.60	0.1663	0.10	0.99	2034	2286	2521
D. 4.2-4.3/100-200	6.8	674.5	219.0	943	0.1449	0.2834	0.60	6.011	0.60	0.1538	0.10	0.99	1608	1977	2389
E. +4.5/abr 3 h	7.2	187.2	104.4	5090	0.1641	0.4801	0.60	11.954	0.60	0.1806	0.10	0.99	2527	2600	2658
A1345-Mujuvaara tonalite															
A. +4.5/+200/abr 6 h	7.2	118.0	68.6	10056	0.0908	0.5242	0.60	14.250	0.60	0.1972	0.10	0.99	2716	2766	2803
B. +4.5/100-200	9.0	123.0	70.3	4434	0.0916	0.5134	0.60	13.889	0.60	0.1962	0.10	0.99	2671	2742	2795
C. 4.3-4.5/+200	7.2	302.2	152.4	4483	0.0875	0.4579	0.60	11.736	0.60	0.1859	0.10	0.99	2430	2583	2706
D. +4.5/long	6.2	144.4	81.4	5619	0.0885	0.5089	0.60	13.671	0.60	0.1948	0.10	0.99	2652	2727	2783
E. +4.5/rounded	6.4	107.2	63.0	4205	0.0933	0.5270	0.60	14.276	0.60	0.1965	0.10	0.99	2728	2768	2797
A1480-Lomperovaara tonalite															
A. +4.5/abr 3 h	6.2	125.8	78.0	18849	0.1407	0.5383	0.60	14.591	0.60	0.1966	0.10	0.99	2776	2788	2798
B. +4.5	6.7	126.6	76.9	52299	0.1245	0.5346	0.60	14.416	0.60	0.1956	0.10	0.99	2760	2777	2789
C. 4.3-4.5/anhedral	6.9	217.3	118.2	14534	0.0994	0.4904	0.60	12.758	0.60	0.1887	0.10	0.99	2572	2661	2731
D. 4.2-4.3/needly	0.8	874.0	223.8	2764	0.0322	0.2497	0.60	5.246	0.60	0.1524	0.12	0.98	1436	1860	2373
E. +4.5/abr 6 h	5.7	120.6	74.7	26215	0.1376	0.5396	0.61	14.632	0.60	0.1967	0.11	0.98	2781	2791	2798
F. +4.5/stubby	2.1	111.1	66.9	8416	0.1232	0.5309	0.60	14.197	0.60	0.1939	0.14	0.99	2745	2762	2776
G. +4.5/needly	1.1	123.6	73.4	15303	0.1215	0.5244	0.71	14.052	0.72	0.1944	0.34	0.92	2717	2753	2779
A1481-Lomperovaara granite vein															
A. +4.2	4.1	1034.1	351.6	3769	0.0653	0.3230	0.60	6.723	0.60	0.1510	0.10	0.99	1804	2075	2357
B. 4.0-4.2/abr 3 h	6.6	1758.2	483.3	2100	0.0994	0.2558	0.66	4.771	0.67	0.1353	0.25	0.94	1468	1779	2167
C. 4.0-4.2	6.0	1704.8	474.9	3205	0.0860	0.2630	0.61	4.937	0.62	0.1362	0.13	0.98	1505	1808	2178
D. 3.8-4.0/+200	6.5	2635.5	621.5	3083	0.1150	0.2197	0.62	3.721	0.65	0.1229	0.19	0.95	1280	1575	1998
E. 3.6-3.8/abr 3 h	6.3	3433.8	580.1	1945	0.1396	0.1549	0.70	2.395	0.71	0.1121	0.15	0.98	928	1241	1834
F. 3.6-3.8	6.6	3497.3	598.6	1967	0.1360	0.1573	0.63	2.447	0.64	0.1128	0.12	0.98	941	1256	1845

Appendix 1. continued

Fraction	Weigth (mg)	U conc (ppm)	Pb conc (ppm)	$^{206}Pb/^{204}Pb$ (meas.)	$^{208}Pb/^{206}Pb$ (radiogenic)	$^{206}Pb/^{238}U$	±2 SE (%)	$^{207}Pb/^{235}U$	±2 SE (%)	$^{207}Pb/^{206}Pb$	±2 SE (%)	Corr.	Apparent ages $T_{206/238}$	$T_{207/235}$	$T_{207/206}$
A1482 Rannimmainen Sotatunturi granite															
A. +4.0/large/abr 16 h	0.5	815	339	2983	0.1200	0.3641	0.50	9.062	0.50	0.1805	0.10	0.98	2002	2344	2658
B. +4.0/euh/abr 16 h	0.5	838	321	4607	0.1100	0.3440	0.50	8.320	0.50	0.1754	0.10	0.98	1906	2267	2610
C. 3.8-4.0/+200/euh/abr 16 h	0.6	2188	701	1775	0.1500	0.2771	0.50	5.948	0.50	0.1557	0.10	0.98	1577	1968	2409
D. 3.8-4.0/-200/abr 16 h	0.5	2243	707	1675	0.1500	0.2723	0.50	5.758	0.50	0.1534	0.10	0.98	1552	1940	2384
A1483-Sulkarinoja granite															
A. +4.3/long	1.3	1150.0	310.1	1161	0.2413	0.2263	0.60	3.521	0.60	0.1128	0.27	0.91	1315	1532	1845
B. 4.2-4.3/long	3.7	1114.8	297.6	968	0.1996	0.2308	0.60	3.433	0.60	0.1079	0.29	0.90	1338	1511	1764
C. 4.0-4.2/long	3.8	1916.6	412.2	1057	0.1696	0.1912	0.65	2.753	0.65	0.1044	0.28	0.92	1127	1342	1704
D. Monazite	1.0	3555.1	9154.0	1450	8.3389	0.3107	2.67	4.968	3.68	0.1159	0.28	0.95	1744	1813	1894
A1484-Marjavaara syenite															
A. +4.5/abr	5.3	220.2	141.6	6670	0.2802	0.5047	0.74	13.583	0.74	0.1952	0.10	0.99	2634	2721	2786
B. +4.5	5.5	250.0	146.4	3943	0.2643	0.4645	0.60	12.310	0.60	0.1922	0.10	0.99	2459	2628	2761
C. 4.3-4.5	5.1	443.0	203.0	1511	0.2616	0.3624	0.60	9.264	0.60	0.1854	0.10	0.99	1993	2364	2701
D. 4.2-4.3	5.7	786.0	253.8	728	0.2720	0.2517	0.63	5.986	0.63	0.1725	0.29	0.91	1447	1973	2581
A1485 Suurkovanselkä alkali syenite															
A. +4.3/-200/abr 5 h	0.5	543	243	4023	0.0900	0.4024	0.50	10.0410	0.50	0.1810	0.10	0.98	2180	2439	2662
B. +4.3/-200	0.5	596	224	2515	0.0800	0.3405	0.50	8.0573	0.50	0.1716	0.10	0.98	1889	2238	2574
C. 4.2-4.3/abr 16 h	0.1	2311	640	1501	0.0700	0.2562	0.50	5.1327	0.50	0.1453	0.10	0.98	1470	1842	2292
D. 4.2-4.3	0.4	1297	342	1456	0.0700	0.2440	0.50	4.8399	0.50	0.1439	0.10	0.98	1407	1792	2275
E. Titanite	9.2	66.5	38.8	702	0.1200	0.4869	0.50	12.3070	0.50	0.1833	0.12	0.97	2557	2628	2683
A1486-Rovaukonselkä quartz feldspar schist (volcanic rock?)															
A. 4.3-4.5/large/pale/turbid	0.6	340	158	4430	0.1000	0.4219	0.50	9.947	0.50	0.1710	0.10	0.98	2269	2430	2568
B. +4.5/abr 5 h	0.4	231	126	9804	0.1300	0.4794	0.50	12.375	0.50	0.1872	0.10	0.98	2525	2633	2718
C. 4.3-4.5/+200/abr 16 h	0.4	306	153	6165	0.1100	0.4472	0.50	10.998	0.50	0.1784	0.10	0.98	2383	2523	2638
D. 4.3-4.5/-200/abr 16 h	0.4	358	181	7742	0.1200	0.4495	0.50	11.004	0.50	0.1776	0.10	0.98	2393	2524	2630

Appendix 1. continued

Fraction	Weigth (mg)	U conc (ppm)	Pb conc (ppm)	$^{206}Pb/^{204}Pb$ (meas.)	$^{208}Pb/^{206}Pb$ (radiogenic)	$^{206}Pb/^{238}U$	±2 SE (%)	$^{207}Pb/^{235}U$	±2 SE (%)	$^{207}Pb/^{206}Pb$	±2 SE (%)	Corr.	Apparent ages		
													$T_{206/238}$	$T_{207/235}$	$T_{207/206}$
A1487-Lomperovaara															
A. +4.5/large	0.4	167	95	7784	0.1000	0.5097	0.50	13.553	0.50	0.19285	0.14	0.96	2655	2719	2767
B. +4.5 long/clear/abr 16 h	0.3	156	94	5496	0.1300	0.5246	0.50	14.224	0.50	0.19667	0.10	0.98	2719	2765	2799
C. +4.5/long/clear	0.3	180	104	15792	0.1100	0.5103	0.50	13.679	0.50	0.19441	0.10	0.98	2658	2728	2780
D. 4.3-4.5/+200/long	0.5	286	147	7917	0.1200	0.4570	0.50	11.623	0.50	0.18447	0.10	0.98	2426	2575	2694
A1488-Suoltijoki granite															
A. +4.5/+100/large	0.6	227	118	2350	0.1200	0.4574	0.50	11.251	0.50	0.17839	0.10	0.98	2428	2544	2638
B. +4.5/-100/abr 16 h	0.5	263	138	3597	0.1100	0.4661	0.50	11.522	0.50	0.17928	0.10	0.98	2467	2566	2646
C. +4.5/-100	0.6	250	123	1896	0.1000	0.4353	0.50	10.489	0.50	0.17478	0.10	0.98	2329	2479	2604
D. 4.3-4.5/+100/abr 16 h	0.5	487	195	580.6	0.0900	0.3448	0.50	7.3458	0.50	0.15452	0.10	0.98	1910	2154	2397

Isotopic ratios corrected for fractionation, blank and age related common lead.
Corr. = Error correlation for 207Pb/235U vs. 206Pb/238U ratios.
SE = Standard error in the mean

Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
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Geological Survey of Finland, Special Paper 33, 229–246, 2001.

U-Pb GEOCHRONOLOGY OF THE KOITELAINEN, AKANVAARA AND KEIVITSA LAYERED INTRUSIONS AND RELATED ROCKS

by
Tapani Mutanen and Hannu Huhma

Mutanen, T. & Huhma, H. 2001. U-Pb geochronology of the Koitelainen, Akanvaara and Keivitsa layered intrusions and related rocks. *Geological Survey of Finland, Special Paper 33,* 229-246. 13 figures, one table and one appendix.

Conventional multi-grain U-Pb zircon results from three mafic layered intrusions in Finnish Lapland are summarized. Gabbroic rocks from the Koitelainen and Akanvaara intrusions give ages of 2439±3 Ma and 2436±6 Ma, respectively. Granophyres from both intrusions as well as quartz monzonites intruding the lower parts of the Koitelainen intrusion provide U-Pb ages of the same order. Archean gneisses form two domes in the Koitelainen area. The U-Pb ages from the Tojottamanselkä dome are c. 3.1 Ga, whereas zircon data on the Kiviaapa gneisses are heterogeneous and discordant providing Pb/Pb ages in the range of 2.74-2.82 Ga. Granophyre results provide the minimum age for the overlying metasedimentary and metavolcanic rocks.

The contact of the Keivitsa intrusion lies only about 1 km south from the edge of the Koitelainen intrusion. Nearly concordant zircon from an olivine pyroxenite gives an age of 2058±4 Ma, which is considered to date the intrusion. This is supported by the age of 2054±5 Ga from a diorite dyke crosscutting the mafic body. Further evidence of this age range is provided by a quartz porphyry at Matarakoski, 10 km southwest from the Keivitsa intrusion, which gives an U-Pb age of 2048±5 Ma. These results provide the lower age limits for the deposition of surrounding metasediments and the extrusion of the komatiites of the Sattasvaara formation.

Key words (GeoRef Thesaurus, AGI): absolute age, U/Pb, zircon, titanite, layered intrusions, gabbros, granophyre, quartz monzonite, pyroxenite, gneisses, Paleoproterozoic, Archean, Koitelainen, Akanvaara, Keivitsa, Finland

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INTRODUCTION

Paleoproterozoic mafic layered intrusions in Northern Finland have been a subject to exploration and research for a long time (see Mutanen 1997). In the 1970s the work was focused on the large Koitelainen intrusion, which was also dated at 2.44 Ga using U-Pb zircon (Kouvo 1977). Soon it became evident that the mafic intrusions further south (Kemi, Penikat, Koillismaa) are roughly coeval with Koitelainen

(Alapieti 1982). In contrast, the age correlation of the mafic intrusions in Central Lapland has been more difficult. It is obvious that intrusions of various ages exist, but zircon data have been frequently discordant and heterogeneous presumably due to extensive lead loss during metamorphic fluid activity.

A large low-grade Cu-Ni sulfide deposit was discovered in the Keivitsa intrusion in 1987, just about 1 km south of Koitelainen. A natural suggestion was that Keivitsa might be a higher level satellite of the Koitelainen intrusion (Mutanen 1989).

The conventional multi-grain U-Pb zircon data produced at the GSF laboratory since 1974 from the mafic layered intrusions of Finnish Lapland are summarized in this paper. These data provide the background for research on Paleoproterozoic mafic magmatism, including Sm-Nd isotopic studies, which aim to constrain the age and genesis of these rock associations (Hanski et al. 2001). This should help in modeling the geological evolution of Central Lapland and northern parts of the Fennoscandian Shield in general.

GEOLOGICAL SETTING

Paleoproterozoic supracrustal rocks cover large areas in Central Lapland, and range from c. 2.5 Ga to c. 1.9 Ga in age. Archean rocks are exposed especially in the east (Tuntsa), but also elsewhere in places, and the involvement of Archean crustal material in Proterozoic granitoids is often significant (Huhma 1986). Following the lithostratigraphic classification by Räsänen et al. (1996) the supracrustal rocks, which formerly were called Lapponian, have been divided into five groups (Salla, Onkamo, Sodankylä, Savukoski, and Kittilä Groups).These supracrustal rocks are associated with several mafic-ultramafic intrusions of various ages. The youngest sedimentary formations consist of the Lainio and Kumpu Groups (Lehtonen et al. 1998), which contain detrital zircons younger than 1.9 Ga (Hanski et al. 2000).

The Koitelainen and Akanvaara mafic magmas both intruded through Archean basement as well as the lowermost Proterozoic supracrustal rocks. Archean gneisses form two domes in the middle of Koitelainen, but the extent of Proterozoic supracrustal rocks clearly older than the intrusions is not well known, and no reliable isotopic ages are available on these rocks. The geology of the Koitelainen, Akanvaara and Keivitsa intrusions has been described in detail by Mutanen (1997).

U-Pb GEOCHRONOLOGY

This paper gives 83 conventional multi-grain U-Pb zircon and titanite analyses on 17 samples from the mafic layered intrusions and some related rocks in Finnish Lapland. Most data have been analyzed during the 1990s, but initial work on some of the Koitelainen samples was done already in the 1970s. Except of a few recent analyses, the sample size has generally been several milligrams of handpicked zircon. The methods follow conventional techniques (Krogh 1973, 1982). The older analyses were made using separate ^{208}Pb and ^{235}U spike solutions and a non-commercial mass-spectrometer. The Pb blank levels for older analyses have been on the order of 0.5 ng, and recently c. 30 pg or less. The input errors used in Isoplot/Ex calculation (Ludwig 1998) for a normal good analysis are 0.65% for Pb/U-ratios and 0.15% for Pb/Pb ratio. The age results are summarized in Table 1 and the figures, where the ages are quoted on a 2σ confidence level. The analytical data are summarized in Appendix I.

The Koitelainen intrusion

The Koitelainen intrusion (Fig. 1) is a flat, oval brachyanticline structure, 26 km x 29 km in size (Mutanen 1997). The magma intruded through the Archean basement and its lowermost Proterozoic cover rocks, and differentiated to form a layered structure of c. 3 km in thickness as measured from the base of the deepest cumulate basin in the west to the top of the granophyre cap.

The intrusion thins at the northern and western limb away from the thickest ultramafic cumulate basin in the west and the ultramafic cumulates are lacking. We presume that this is the direction of transgression of the basal contact across the sedimentary country rocks, a feature commonly found among well-exposed layered intrusions. Tectonic thinning is drastic in all of the western limb and in the uppermost granophyre in

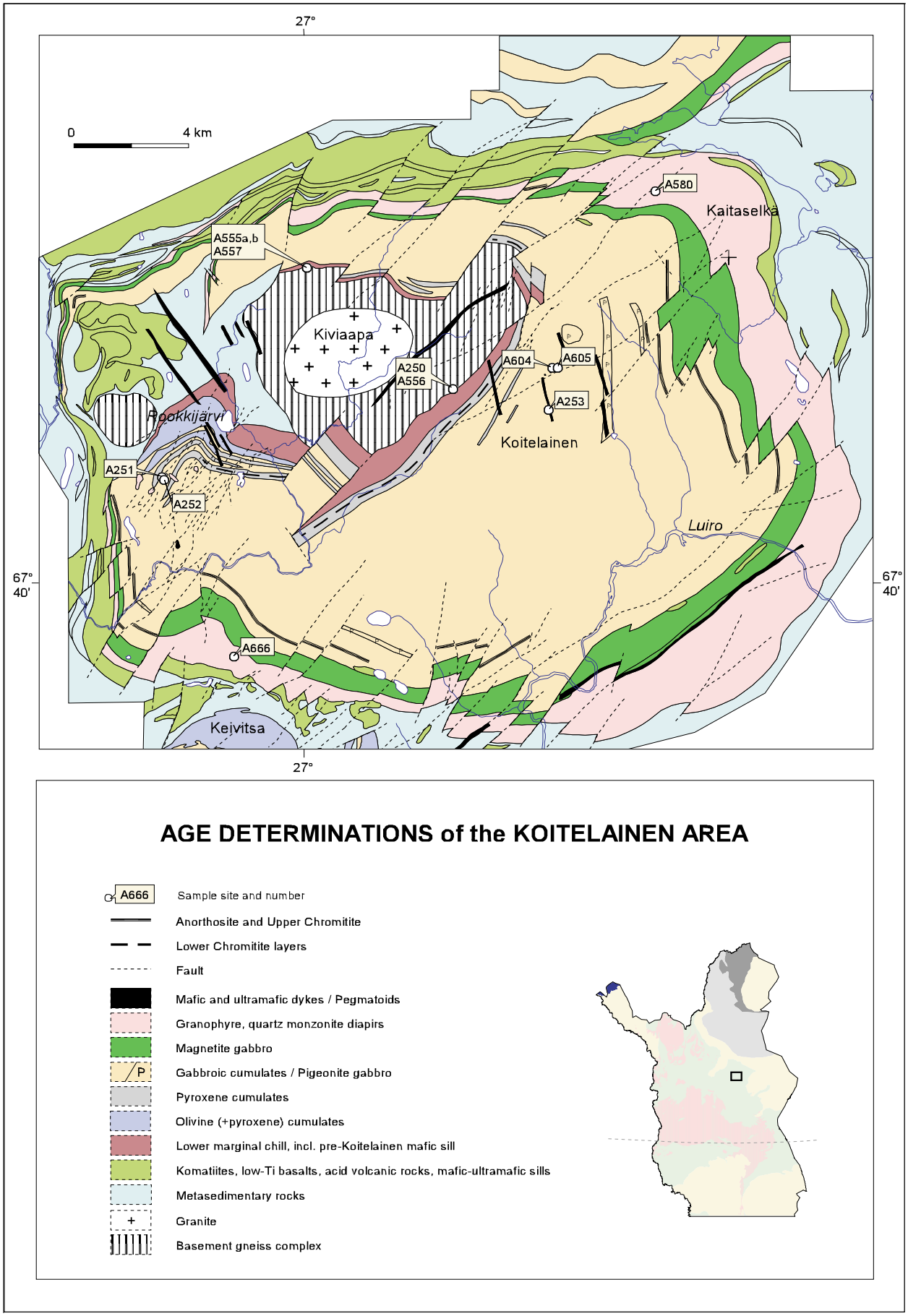

Fig. 1. Geological map of the Koitelainen layered mafic intrusion and the sampling sites.

the northern and northeastern part of the intrusion.

The intrusion is divided into an ultramafic Lower Zone, a gabbroic Main Zone and a gabbroic Upper Zone with anorthosite and magnetite gabbro. Granophyre forms a continuous cap above the mafic succession. Quartz monzonite diapirs intruded the cumulates of the Lower Zone. Several chromitite layers were intersected by diamond drilling in the orthopyroxene cumulates of the uppermost Lower Zone. A thin chromitite layer occurs at the base of a pyroxene cumulate layer among gabbroic cumulates in the lower part of the Main Zone. The peculiar Upper Chromitite layer occurs at the base of the Upper Zone.

Gabbro pegmatoids A604 and A605, Lakijänkä. The samples A604 (129-TM/73) and A605 (130-TM/73) were taken from the Koitelainen hill and represent ultramafic gabbro pegmatoids located in the Main Zone of the Koitelainen intrusion. The pegmatoids are pipes and dykes with sharp boundaries against the surrounding gabbros. The rocks were originally clinopyroxene rocks with interstitial plagioclase, ilmenite and magnetite. The deepest parts of the pipes are rich in ilmenomagnetite and ilmenite. The original Ca-clinopyroxene in these pegmatoids is mostly altered to amphiboles. Primary quartz occurs as blue grains and its amount increases upwards in the pipes. Other primary minerals are zircon, apatite and allanite. The pegmatoid material represents part of the intercumulus melt drawn into contraction voids from surrounding gabbroic cumulates. The pegmatoid melt evolved along a cotectic path of crystallization of primary titanomagnetite, Ca-clinopyroxene, a silica phase, zircon, apatite and plagioclase. For more details of the petrology of pegmatoids see Mutanen (1997). As for the interpretation of the zircon ages, it is important to emphasize that the pegmatoids are not

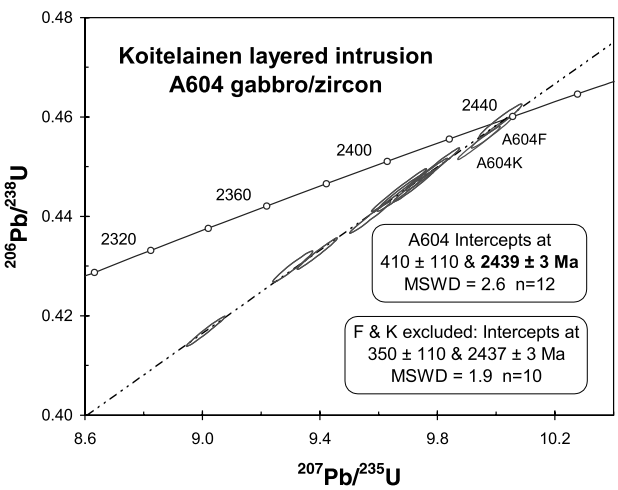

Fig. 2. Results for U-Pb analyses on zircons from the Koitelainen gabbro A604. The ellipses indicate the analytical error on the 95% confidence level.

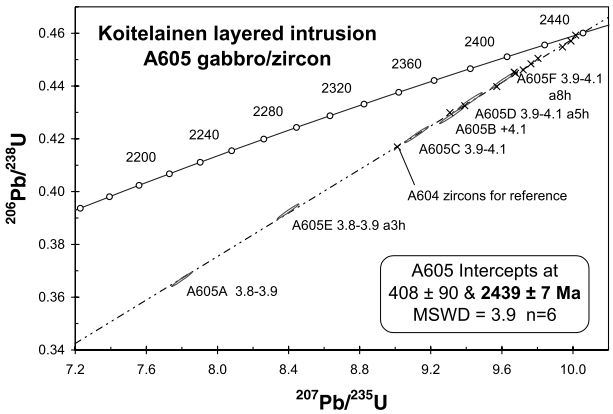

Fig. 3. Results for U-Pb analyses on zircons from the Koitelainen gabbro A605. The analytical points of sample A604 are shown as crosses for reference.

younger hydrothermal-metasomatic rocks but represent local intercumulus melt, and zircon crystallized from this melt.

The extraction of zircon and first analyses were performed in 1975. Zircon is dark brown and mostly fairly large (>160 μm). Simple prisms or fragments with sharp edges are typical, but tabular crystals are also common. Some grains have cores of baddeleyite, some are transparent and contain reddish resorbed cores, which were broken out from the mantles during crushing. Tabular crystals may have inherited their shape from baddeleyite seed crystals, through growth by reaction with silica in melt.

From the sample A604 U-Pb analyses have been made on twelve fractions, one of which is concordant at 2436 Ma. The regression of all analyses provides an upper intercept age of 2439±3 Ma and a lower intercept of 410±110 Ma (MSWD=2.6, Fig. 2). Fractions representing different morphological types yield practically coeval results, although the handpicked zircon cores (A604F, A604K) tend to give marginally older age indications. The cores have lower U than the rest of the zircons and should represent early stages of magmatic crystallization of the pegmatoid melt. Without these two analyses the upper intercept age is 2437±3 Ma (MSWD=1.9).

The six zircon fractions analyzed from sample A605 provide an upper intercept age of 2439±7 Ma and a lower intercept at 408±90 Ma (Fig. 3). The 18 analyses on samples A604 and A605 give an intercept of 2439±2 Ma (MSWD=2.6).

Quartz monzonites A251 and A252, Koitelainen. Quartz monzonite stocks (diapirs) intrude through the lower zone cumulates. They represent large-scale melting of the floor rocks at the early stage of the consolidation history of the intrusion. The original magmatic minerals are plagioclase, perthitic potassium feldspar, iron-rich pigeonite and fayalite.

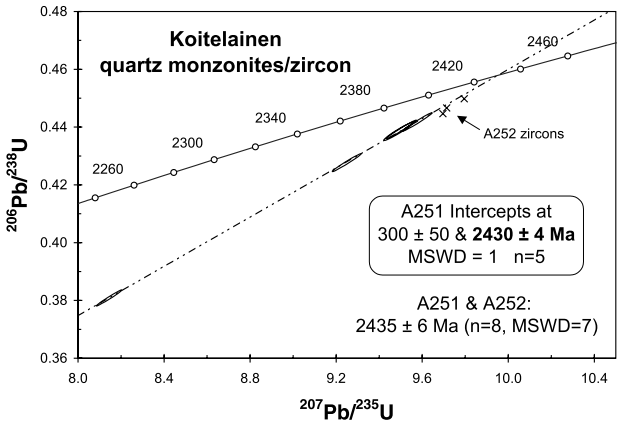

Fig. 4. Results for U-Pb analyses on zircons from the Koitelainen quartz monzonites. Ellipses for A251, crosses for A252.

Details of these "intra-intrusions" are given by Mutanen (1989, 1997). Similar "rheomorphic" melts are found among the lower parts of other layered intrusions, e.g. Bushveld, Stillwater and Insizwa (for references, see Mutanen 1997, p.73).

The samples taken for U-Pb dating are medium-grained homogeneous amphibole-plagioclase rocks. Sample A252 (TM-80-160.3) is characterized by unorientated laths of plagioclase (1 cm), whereas the other sample A251 (TM-80-143.3) is even-grained.

Separation of sample A251 yielded abundant brown, large zircon, which often has been broken during crushing. Sharp edges and simple tetragonal forms as well as mafic inclusions on the surfaces are characteristic. The five analyzed zircon fractions yield a chord with intercepts at 2430±4 and 300±50 Ma (MSWD=1, Fig. 4).

Zircons from the other sample A252 are quite different. They are colorless, water clear, small euhedral crystals with sharp edges. The three analyses plot relatively close to the concordia curve, marginally to the "older side" from the discordia defined by the sample A251 (Fig. 4.). No chord can be calculated from these data, but the Pb/Pb ages of individual analyses (2431-2436 Ma) should be considered minima for the zircons. If regressed together, the eight analyses from samples A251 and A252 give an upper intercept age of 2435±6 Ma. The fairly large MSWD of 7 suggests, however, some scatter in excess of analytical error.

Granophyre A580, Kaitaselkä. Granophyre forms the uppermost, up to 400 m thick unit between the mafic cumulates and supracrustal rocks on the roof all around the Koitelainen intrusion. The main minerals of the granophyres are albite+epidote (originally euhedral to subhedral plagioclase), dark hornblende and biotite (both secondary). Sample A580 was taken from Kaitaselkä on the NE edge of the Koitelainen intrusion. The rock is relatively fine-grained, pinkish and shows slight lineation in hand specimen.

The sample yielded abundant brown, often turbid zircon, which occurs as long simple prisms. The edges are sharp. The first analyses were made already in 1975, and are quite discordant. Since then analyses were made also using air-abrasion, and some improvement in concordance was achieved. The nine fractions available provide intercepts at 2405±9 and 240±34 Ma (MSWD=12, Fig. 5). The large MSWD shows that the data are not perfectly linear on the concordia diagram. Two old analyses are from very low-density (3.6-3.8) fractions and quite discordant. By omitting these two points the result becomes 2410±11 and 278±60 Ma (MSWD=8), and the four abraded fractions give intercepts at 2413±14 and 305±100 Ma (MSWD=2.4). Due to the heterogeneous and discordant data the age estimate derived for this rock should be taken with caution.

Granophyre A666, Vaiskonselkä. Sample A666 (30-KP/75) was taken from Vaiskonselkä at the southern margin of the Koitelainen intrusion some 30 km from the Kaitaselkä granophyre A580. The rock is medium-grained, slightly foliated in hand specimen and contains more dark minerals than A580. Plagioclase, now altered to albite and clinozoizite-epidote, crystallized as subhedral primocrysts. The matrix consists of quartz, potassium feldspar (partly microcline), dark blue-green hornblende and biotite. Pyrite, apatite and zircon are accessory minerals. Apatite occurs as big crystals in association with hornblende. The rock contains salic, quartz-rich eyes.

Zircon from the sample A666 is dark brown. Three old and two new, less discordant analyses lie fairly well on a chord which gives intercepts at 2432±4 and 331±50 Ma (MSWD=1.4, Fig. 5). Compared to the previous sample the data from this granophyre are more concordant, and the age is more reliable.

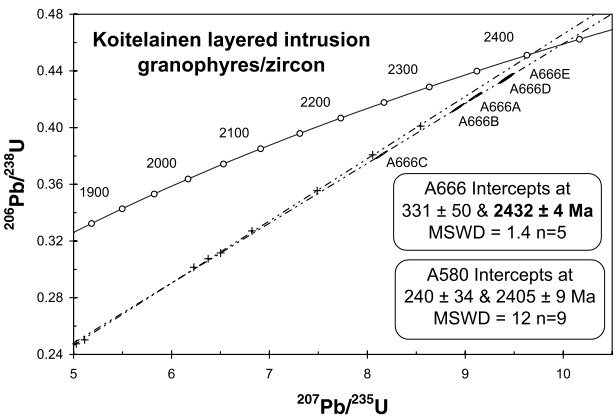

Fig. 5. Results for U-Pb analyses on zircons from the Koitelainen granophyres. Ellipses for A666, crosses for A580.

Archean rocks from the Kiviaapa dome

Mapping and research on the Koitelainen intrusion has involved also country rocks. Isotopic dating results on the western margin of Koitelainen are reported by Manninen et al. (2001, *this volume*), whereas Räsänen and Huhma (2001, *this volume*) discuss geochronology elsewhere in the Sodankylä area. Archean rocks occur in two domal structures, the well-known Tojottamanselkä area (Kröner et al. 1981) and the large Kiviaapa structure further east.

Samples for U-Pb dating have been taken from two localities of the Kiviaapa dome. Samples A250 and A556 are from the eastern side of the dome c. 500 m from the intrusion contact, whereas A555a, A555b and A557 come from the northern contact area. All rocks have been considered contact hornfelses of the Koitelainen intrusion, altered in a later regional metamorphism (Mutanen, 1997).

Metatuffite A250 and paragneiss A556, Kiviaapa. The samples were taken from a group of frost-shattered outcrops ("rakka") located between the low-gravity core of the Kiviaapa dome and the base contact of the Koitelainen intrusion.

Sample A250 (TM-80-87.2) is a fine-grained, homogeneous and dark rock in hand specimen. It is considered a tuffitic rock in a gneissic environment. The main minerals are quartz, albite, epidote and amphibole; titanite is a characteristic accessory mineral. The zircon population is heterogeneous. Many grains are somewhat rounded and have rough surfaces, whereas some have fairly sharp edges. Few grains are relatively large for a fine-grained volcanic rock. An analysis has been made only from the rounded zircons, and the result lies relatively close to the concordia curve providing a $^{207}Pb/^{206}Pb$ age of 2.79 Ga (Fig.6).

Titanite is mostly dark brown and clear. A U-Pb analysis shows a relatively low content of common lead, providing a reliable result fairly close to the concordia at about 2.52 Ga. The exact meaning of this age remains unclear, but certainly the titanite was formed before the emplacement of the Koitelainen intrusion, and the heat of the mafic magma was not able to reset the U-Pb clock. Neither did resetting occur during the later regional metamorphism.

The felsic gneiss sample A556 (TM-80-87.1) was collected from the same outcrop group, but the origin of this rock remains somewhat unclear. The sample is a pink, medium to fine-grained rock in hand specimen. The main minerals are quartz, cloudy feldspar, K-feldspar and epidote.

Separation yielded abundant zircon, which ranges from light-colored clear to brown turbid in color. Both euhedral and rounded grains occur, and the population is heterogeneous. The three analyses are slightly discordant and do not plot on a straight line on the concordia diagram. Two fractions from the euhedral grains yield $^{207}Pb/^{206}Pb$ ages of c. 2.74 Ga, and one from the rounded zircon gives an older Pb/Pb age of 2.82 Ga (Fig. 6).

Metasedimentary rocks A555a and b and meta-arkose A557, Kiviaapa. Sample A555a (TM-80-86.1a) contains plagioclase and potassium feldspar granules (clasts?) in a dark matrix which consists of quartz, feldspar and epidote. In field the rock was considered volcanogenic in origin, but microscopic study suggests that it is a coarse clastic metasediment. The zircon population in this sample is bimodal. One population consists of generally small, transparent light brown and euhedral crystals, whereas others are larger, turbid dark brown and often slightly rounded.

The U-Pb data are also heterogeneous. Two discordant analyses from the small, euhedral grains

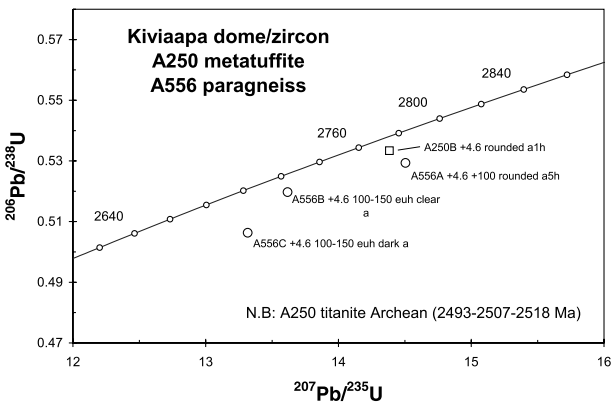

Fig. 6. Results for U-Pb analyses on zircons from the southeastern rim of the Kiviaapa dome.

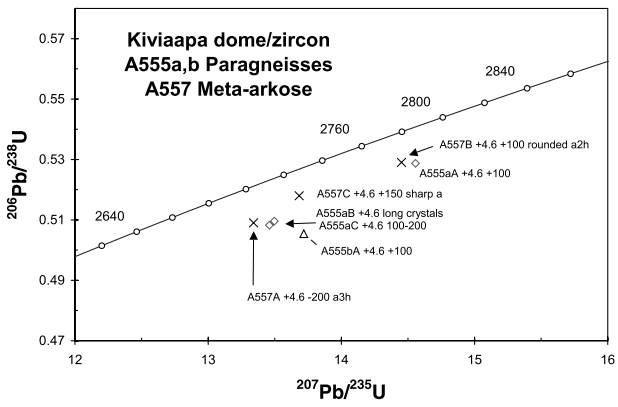

Fig. 7. Results for U-Pb analyses on zircons from the northern contact area of the Kiviaapa dome.

provide a $^{207}Pb/^{206}Pb$ age of 2.76 Ga, whereas the analysis on the large grains gives an older age indication (Fig. 7).

The other sample A555b is a medium-grained, pink, felsic rock with occasional faint banding. It is spatially closely related to A555a. The main minerals are quartz, albite, potassium feldspar and epidote. The zircon in this sample consists of yellow-brown short, often slightly rounded crystals. Only one analysis from the large grains has been made, which is discordant and gives a $^{207}Pb/^{206}Pb$ age of 2.8 Ga (Fig. 7.).

The third sample from this locality, A557 (TM-80-86.2), is a very fine-grained pink rock. The texture is banded in this section, which is caused by alternating quartz-rich and albite+K-feldspar+epidote dominated layers. The rock is quite similar to meta-arkosic (sub-arkosic) rocks encountered in diamond drill holes and in outcrops below the Koitelainen intrusion in several places near the western part of the intrusion.

Separation yielded abundant zircon, which is pale brown and clear. Many grains are rounded, and have a detrital appearance. The three U-Pb analyses resemble the data from the other Kiviaapa samples. Rounded large grains tend to give older age indications than the euhedral small crystals.

Metadiabase A253, Keskilaki. Diabase dykes with chilled margins cross-cut the Koitelainen gabbro. Sample A253 (TM-80-91.1) represents a coarse-grained magnetite-rich portion of a 90 m thick dyke. Its mineralogy is partly altered due to regional metamorphism, which presumably occurred 1.8-1.9 Ga ago.

Separation of the sample yielded a small amount of zircon with baddeleyite intergrowths. Crystals are brown and often slightly rounded. Three analyses have been made, but the data are discordant and heterogeneous, with $^{207}Pb/^{206}Pb$ ages ranging from 2016 to 2226 Ma. No age dating is possible from these data.

Akanvaara layered intrusion

The geological setting and the internal structure of the Akanvaara layered intrusion (Fig. 8) resemble those of Koitelainen (Mutanen 1997). However, due to the lack of outcrops its detailed contact relationships are poorly known. The lower contact has been intersected by several diamond drill holes. The intrusion is a block-faulted monocline with a general dip towards southeast. The present surface area is c. 50 km^2. The immediate floor consists of acid volcanic rocks (metalavas?). The immediate roof is not exposed nor are there any drill intersections. The lowermost exposed roof rocks are felsic volcanic rocks. Xenoliths of fine-grained porphyritic felsic rocks and basaltic rocks are known to occur among the granophyre cap. Mafic to ultramafic dykes, representing at least three different magma types, cut the intrusion and the country rocks.

Gabbro A1310, Tuorelehto. A vanadium-rich chromitite layer similar to the Upper Chromitite of the Koitelainen intrusion occurs at the base of the Upper Zone (Mutanen 1997). This chromitite is well exposed in a trench and an excavated area at the Tuorelehto hill. The sample A1310 for isotopic studies was taken 6 m above the Upper Chromitite layer, and represents a coarse leucocratic gabbro with a crescumulate texture.

Zircons are clear, brown, subhedral prismatic crystals or fragments. Tabular forms are common. U-Pb analyses have been made on six fractions. Regression on five analyses gives intercepts at 2436±6 and 731±130 Ma (MSWD=1.2, Fig. 9). An analysis on the fine-grained heavy zircon lies slightly off the line and is not included in this calculation. In contrast to the normal situation this fraction as well as the other analyses on the heavy zircon are most discordant. The heavy zircon also seems to have slightly lower Th/U than the lighter fractions (see radiogenic $^{208}Pb/^{206}Pb$ in Appendix).

This type of heterogeneity is not uncommon in the mafic rocks from Lapland. Previous attempts to date the Akanvaara intrusion have provided discordant data with rough estimates at about 2.3 Ga, and the analyses on heavy transparent zircons from the Siikakämä layered intrusion suggested hydrothermal zircon growth during a 1.9-1.8 Ga thermal pulse (Mertanen et al 1989).

Granophyre A1311, Akanvaara. The sample A1311 was taken from the Akanvaara hill. It represents an iron-rich granophyre, which belongs to the uppermost part of the granophyre cap of the intrusion. The granophyre is a medium-grained, granoblastic rock, the major minerals being plagioclase (now albite and muscovite), quartz, biotite and chlorite. Microcline is totally albitized. Accessory minerals are apatite, zircon, titanite, ilmenite and magnetite. The presence of ilmenite and magnetite is not typical for the Akanvaara granophyre.

The appearance of the zircon population is similar to the gabbro A1310, although the crystals are more turbid. The five zircon fractions analyzed provide

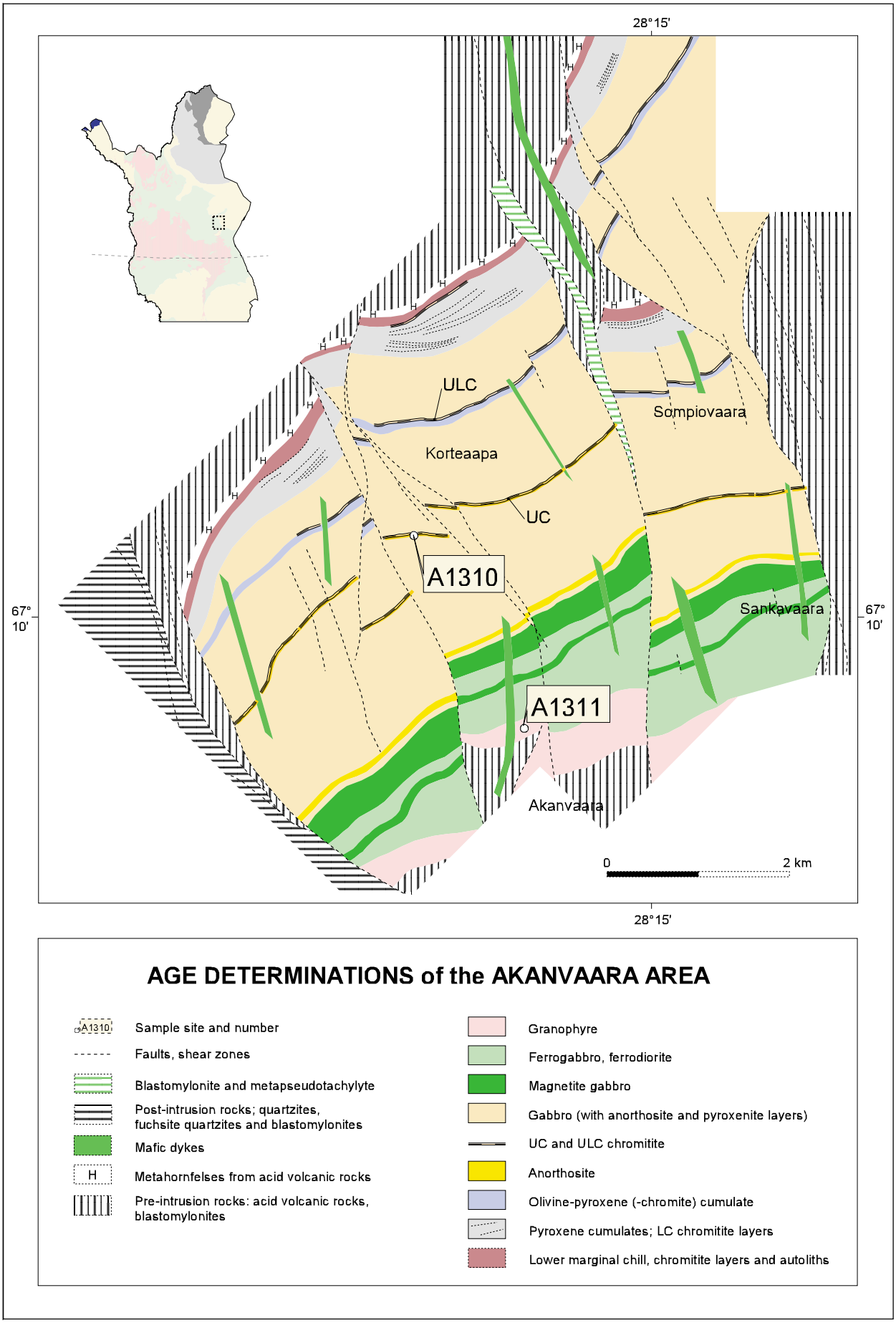

Fig. 8. Geology of the Akanvaara mafic layered intrusion and the age determination sampling sites.

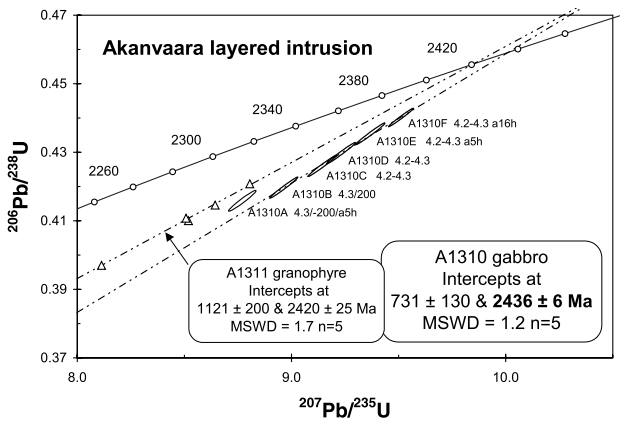

Fig. 9. Results for U-Pb analyses on zircons from both gabbro (A1310, ellipses) and granophyre (A1311, triangles) from the Akanvaara layered intrusion.

intercepts at 2420±25 and 1122±200 Ma (MSWD=1.7, Fig. 9). The large error estimate is mainly due to the limited dispersion of the discordant data points. Anyway, the result is compatible with the age derived from the gabbro A1310, but clearly the discordance pattern (lower intercept) is quite different.

Keivitsa mafic intrusion and related rocks

The Keivitsa magma intruded supracrustal rocks consisting mainly of pelitic and sapropelic metasediments (Fig. 10). The magma differentiated to produce a chilled microgabbro at the base, thick olivine-pyroxene cumulates in the lower part, gabbros, ferrogabbros and magnetite gabbros in the upper part and a sodic granophyre at the top. A disseminated Cu-Ni sulphide deposit is located in the upper part of the ultramafic zone. This zone also contains compositionally and texturally variable ultramafic xenoliths of komatiitic parentage, big pelitic hornfels xenoliths and "rotten" pelitic xenoliths. Lumps of graphite are what remain of large amounts of carbonaceous sapropelic material digested by the magma. Disseminated graphite is common among olivine pyroxenites, and euhedral cumulus and intercumulus graphite occurs in a pigeonite gabbro.

The deposit consists of compositionally distinct ore types, which occur in close proximity to each other and are intermingled. Ore types typically grade into each other. The most common type is 1) Cu-Ni-PGE-Au-type. This often grades into sulphide-rich 2) false ore type, which is low in sulphidic Ni and noble metals. The 3) Ni-PGE type is poor in sulphides and contains the highest PGE concentrations but has low Au values. The 4) intermediate type between the main type and Ni-PGE-type contains rather high PGE-Au values and the bulk sulphides are rich in Ni (see Mutanen 1997). Primary igneous minerals are well preserved in most places in the Keivitsa intrusion.

Pyroxenite A1390, Keivitsa. A fresh olivine pyroxenite (sample A1390, TM-93-6.1) near the base contact contains skeletal zircon crystals (see Mutanen 1997, Fig. 46f, p. 148-149). Electron probe microanalyses show U concentrations of 2600-5300 ppm for these crystals. The anomalous U concentrations are evidently due to diffusion of U from adjacent pelitic floor rocks. Contamination is further attested by the occurrence of chlorapatite and plentiful primary biotite in the sample. The main minerals are orthopyroxene, calcic clinopyroxene and olivine, and other minerals include intercumulus plagioclase, primary biotite, brown primary hornblende, apatite, small amounts of pyrrhotite, magnetite, chalcopyrite, pentlandite and ilmenite. Secondary alteration (local rodingitization of plagioclase, secondary amphiboles) is slight.

The separated zircon consists of dark brown, very transparent, well-faceted and equant crystals and fragments. The U-Pb analyses have been made at two stages. The first two analyses were made on hand-picked bulk material using the conventional method. The four other analyses were made on carefully selected high-quality crystals using new low-blank techniques. These four analyses contain a very low amount of common lead (Appendix I). In spite of the high U concentration and consequent metamictization of zircon, the U-Pb analyses are nearly concordant. This could be related to the preservation of magmatic mineralogy.

The regression of all six analyses provides intercepts at 2057±8 and 476±430 Ma (MSWD=2.6, Fig. 11). As the four new analyses are made on freshest zircons and yield analytically good results, the age based on these data should be considered as a best estimate. These four points give an age of 2058±4 Ma (MSWD=0.67).

This age is compatible with the U-Pb zircon age from the associated diorite dyke (below) and with the Sm-Nd age measured from the primary igneous min-

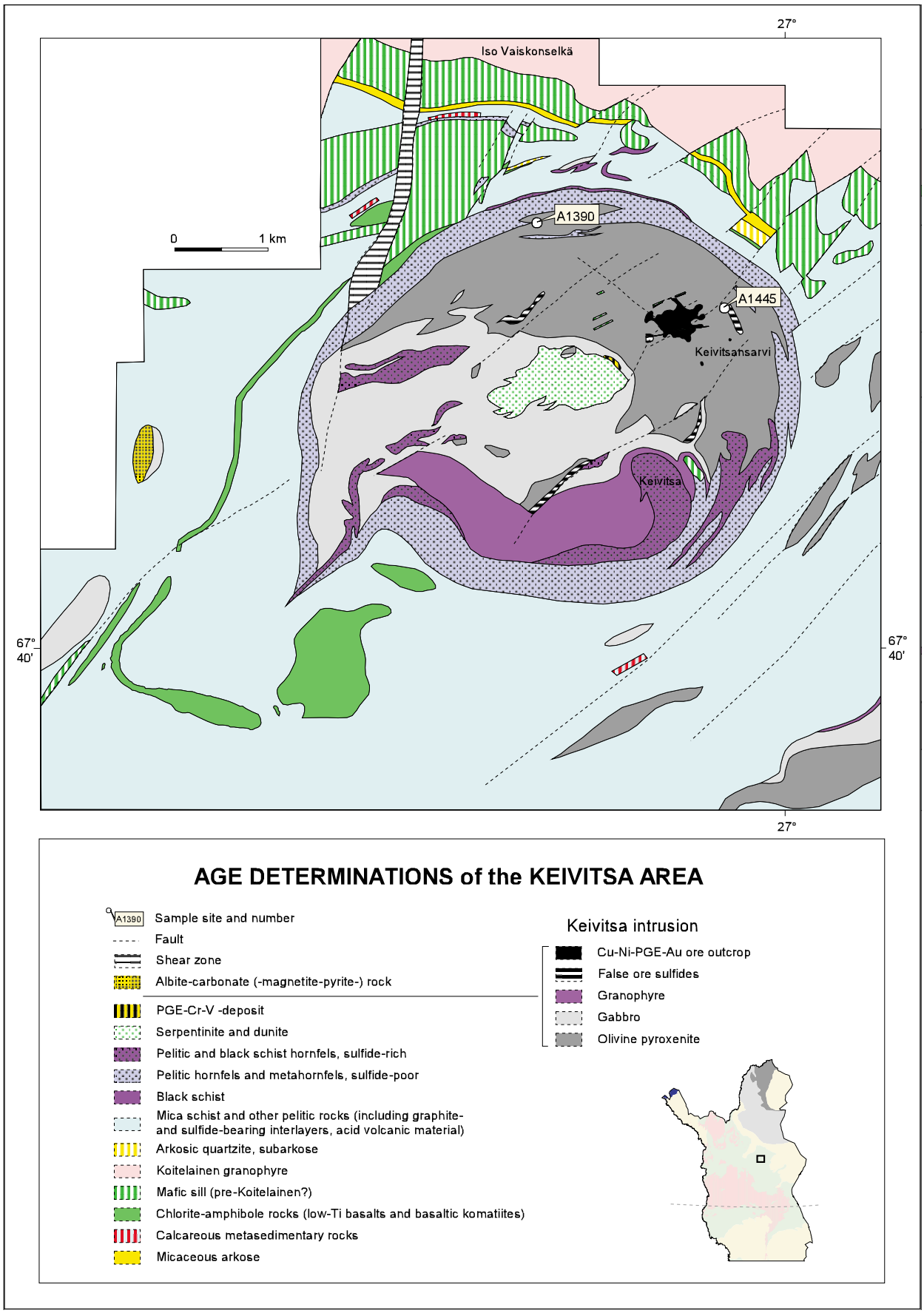

Fig. 10. Geology of the Keivitsa layered intrusion and the age determination sampling sites. The site of sample A566-Matarakoski plots just west of the map area.

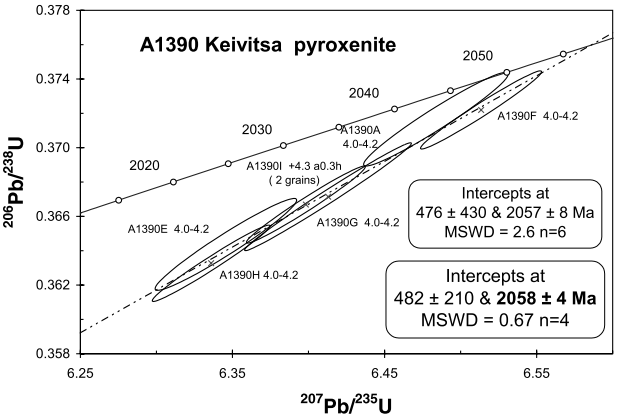

Fig. 11. Results for U-Pb analyses on zircons from the Keivitsa pyroxenite A1390.

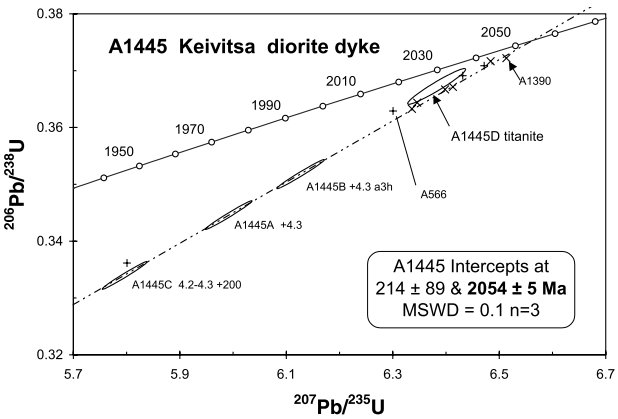

Fig. 12. Results for U-Pb analyses on zircons from the Keivitsa diorite dyke A1445.

erals of the Keivitsa intrusion (Huhma et al. 1996), and should thus be considered close to the age of the Keivitsa intrusion. One may, however, argue the concept of dating an ultramafic cumulate by U-Pb zircon. The natural suggestion would be that zircon must have been derived from external sources. As discussed above, contamination by felsic material is evident also here (Mutanen 1997), and presumably zirconium, uranium, along with potassium and chlorine, were derived from country rocks. The skeletal zircon, also with high U, without doubt crystallized from a melt, and thus there is no need to interpret the zircons as inherited (xenocrystic) material.

Diorite dyke A1445, Keivitsa. Mafic and felsic dykes have been found in trenches and are often intersected in diamond drill holes in the Keivitsa intrusion. Sample A1445 (R923/35.90-38.75) represents a medium-grained diorite dyke, and consists of plagioclase (partly altered to scapolite, epidote-clinozoisite and pistacite), calcic clinopyroxene, primary brown hornblende, secondary green hornblende and cummingtonite as well as small amounts of biotite. Titanite and large crystals of apatite are common. Rutile, ilmenite, carbonate, pyrite and pyrrhotite occur in small amounts.

The diorite dykes have sharp, winding contacts. Contact effects in the surrounding olivine pyroxenite are often very slight. The dykes typically have coarse marginal zones with large zoned plagioclase crystals. The inner part consists mainly of plagioclase, hornblende and quartz. Hedenbergite and brown magmatic hornblende are found in some felsic varieties (see Mutanen 1997, p. 163-164). The dykes have a high Na/K ratio. Evidently, the diorite veins were formed by anatectic roof melting; these magmas intruded downwards (back intrusion) into contraction cracks opened during late stages of the consolidation of the intrusion.

Separation has yielded a small amount of zircon and much titanite. Zircon occurs mainly as light coloured anhedral to subhedral and turbid grains. The three U-Pb analyses made give intercepts at 2054±5 and 214±89 Ma (Fig. 12). Titanite is dark brown and transparent. The lead is fairly radiogenic and gives a $^{207}Pb/^{206}Pb$ age of 2044±4 Ma. The analysis is slightly discordant and plots marginally to the "younger" side of the chord defined by three zircon analyses. This provides a minimum age for the dyke and the Keivitsa intrusion.

Quartz porphyry A566, Matarakoski. At Matarakoski by the river Kitinen, ten kilometers southwest from the Keivitsa intrusion, there occurs a felsic porphyry closely associated with komatiitic metavolcanics. A several meters thick unit of porphyry can be traced from the old riverside outcrop (marked as "leptitic schist" by Mikkola 1941, Fig. 7, p. 51) at least a hundred meters towards the west. The porphyry occurs among metapelites, black schists and komatiitic metavolcanic rocks, and is considered an eruptive rock rather than a sill. Euhedral plagioclase phenocrysts and a few lenses of granulated quartz

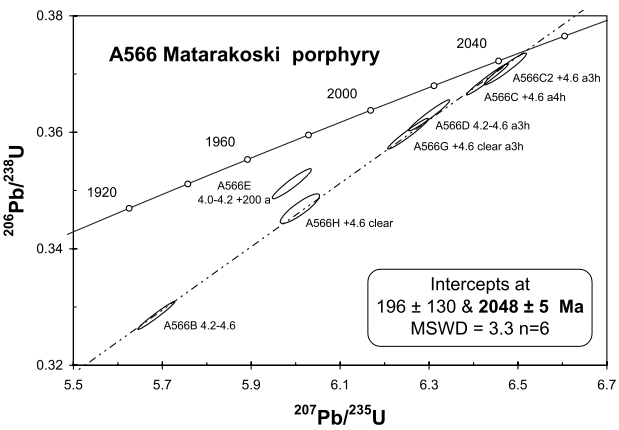

Fig. 13. Results for U-Pb analyses on zircons from the Matarakoski quartz porphyry A566.

phenocrysts occur in a fluidal-textured felsic groundmass containing chlorite and muscovite (See Fig 40b in Mutanen 1997, p. 132).

Separation in the early 1980's yielded zircon, which occurs as small, nearly colorless, clear crystals. Simple euhedral prisms are common, but many grains are subhedral. The appearance of the zircon population is homogeneous.

The U-Pb analyses have been made at various stages, early analyses in 1982 and the air-abraded analyses in the 1990's. Six of the seven analyses plot on a chord, which gives intercepts at 2048±5 and 196±130 Ma (MSWD=3.3, Fig. 13). An analysis of light zircon plots off the line, presumably because of metamorphic effects. Two analyses on the abraded heavy (+4.6) zircon fraction give nearly concordant ages and regression on the four points of heavy zircon gives an upper intercept age of 2051±3 Ma (MSWD=0.4).

DISCUSSION AND CONCLUSIONS

The interpretations of the data outlined in the preceding text are summarized in Table 1 together with location data for the analyzed samples.

Several well known problems are involved in interpreting conventional multi-grain U-Pb zircon results. Except of the few newer data, analyses generally involved several milligrams of handpicked zircon. Even though careful selection was maintained in handpicking, there certainly are difficulties with the homogeneity of the zircon population used for analysis. This becomes critical especially when the quality of zircon is less favorable e.g. due to turbidity. The analyses are then easily discordant, and the age determination always involves interpretation. This is the case with many samples in this paper.

An evaluation of the results shows that there exists quite often a scatter in excess of estimated analytical error in the data set. This means that the errors are either underestimated or the zircon populations of the igneous rocks in question are heterogeneous. There are several examples in the laboratory, where old analyses (made since seventies) are well on the chord defined by new data. Generally duplicated analyses are also well compatible. Thus the error estimates used here for individual data points should be generally appropriate, and in any case the extra scatter (MSWD>1) has been taken into account in the final upper intercept age calculation.

Consequently, the scatter in the data must be due to multistage geological history. One of the extreme cases was found in the Siikakämä intrusion south of Rovaniemi where slightly discordant analyses on the heavy transparent zircon yield the youngest $^{207}Pb/^{206}Pb$ ages (c. 2 Ga), whereas other fractions give older indications (Mertanen et al. 1989). Similar, less pronounced cases are common in the 2.44 Ga Penikat layered intrusion (Perttunen & Vaasjoki 2001, *this volume*), and suggest that metamorphic (1.8-1.9 Ga?)

Table 1. Summary of U-Pb mineral results from the layered mafic intrusions in central Lapland.

Sample	Map	Northing	Easting	Location	Mineral	Rock type	Age	N	Comment
A604	374101	7526.05	3508.88	Lakijänkä	zircon	Gabbro pegmatoid	2439±3	12	MSWD=2.6
A605	374101	7526.06	3509.07	Lakijänkä	zircon	Gabbro pegmatoid	2439±7	6	MSWD=3.9
A251	372310	7522.13	3495.30	Koitelainen	zircon	Monzonite	2430±4	5	MSWD=1
A252	372310	7522.02	3495.38	Koitelainen	zircon	Monzonite	–	3	nearly concordant, bunched
A580	374105	7532.40	3512.47	Kaitaselkä	zircon	Granophyre	2405±9	9	MSWD=12
A666	371412	7515.68	3497.81	Vaiskonselkä	zircon	Granophyre	2432±4	5	MSWD=1.4
A250	374101	7525.30	3505.42	Kiviaapa	zircon	Metatuffite	–	1	207/206 ages: zircon 2789, titanite 2518
A556	374101	7525.30	3505.42	Kiviaapa	zircon	Paragneiss (hornfels)	–	3	heterogeneous Archean, c. 2.8 Ga
A555	374101	7529.66	3500.34	Kiviaapa	zircon	Metasediment (basement gneiss)	–	4	heterogeneous Archean, c. 2.8 Ga
A557	374101	7529.66	3500.34	Kiviaapa	zircon	Mata-arkose (basement gneiss)	–	3	heterogeneous Archean, c. 2.8 Ga
A253	374101	7524.54	3508.75	Keskilaki	zircon	Metadiabase	–	3	heterogeneous, 207/206 ages 2.02-2.23 Ga
A1310	364406	7454.50	3551.70	Tuorelehto	zircon	Gabbro	2436±6	5	MSWD=1.2
A1311	364406	7452.40	3552.90	Akanvaara	zircon	Granophyre	2420±25	5	MSWD=1.7
A1390	371412	7513.52	3497.33	Keivitsa	zircon	Pyroxenite	2058±4	6	2 fractions excluded; MSWD=0.67
A1445	371412	7512.60	3499.35	Keivitsa	zircon	Diorite dyke	2054±5	3	MSWD=0.1; titanite slightly younger
A0566	371411	7505.34	3491.13	Matarakoski	zircon	Quartz porphyry	2048±5	7	one fraction excluded; MSWD=3.3

fluids have played a major role in disturbing the U-Pb chronometer in zircon. In these cases, zircon inheritance is not considered to be the cause for heterogeneity.

The age of 2439±3 Ma for the Koitelainen layered intrusion is well defined since the best data points from the ultramafic pegmatoid A604 are nearly concordant. The calculated upper intercept ages for other rock types from the Koitelainen intrusion seem to be marginally younger. These ages are based on discordant data points, and sometimes very discordant analyses play a major role in line fitting. This simple approach is, however, not valid if e.g. post-Koitelainen burial metamorphism or the 1.9-1.8 Ga event has influenced the U-Pb system in zircon: The line-fitting gets worse and the chord could then define an upper intercept, which gives an age too young for magmatic zircon. We think this is the major reason for some young apparent upper intercept ages at both Koitelainen and Akanvaara.

The Kaitaselkä granophyre A580 provides the most pronounced example. The calculated upper intercept age is 2405±9 Ma, and would as such suggest that the granophyre is much younger than the gabbro, and also younger than the other granophyre A666, which has an upper intercept age of 2432±4 Ma. However, the age for A580 is based on nine discordant data points which give a MSWD of 12 for the line fitting. Based on geological evidence, this rock should be roughly coeval with the main Koitelainen body, and the apparent upper intercept age must be considered too young.

Also the data from the Akanvaara granophyre (A1311) seem to be disturbed by later processes and have a significantly different discordia pattern when compared to the zircon from the Akanvaara gabbro (A1310), but the error limits for the ages are fairly large and overlapping.

This paper has presented U-Pb isotopic data warranting the following conclusions.

1) The large layered intrusions in Koitelainen and Akanvaara are c. 2.44 Ga old and coeval with many other similar intrusions in the Fennoscandian Shield (Alapieti 1982, Amelin et al. 1995). Granophyre ages represent the minimum age for the overlying metasedimentary and metavolcanic rocks.

2) The Keivitsa magma intruded the surrounding sedimentary and volcanic rocks c. 2.06 Ga ago.

3) The age of the Matarakoski porphyry is c. 2.05 Ga, which is the lower age limit for the komatiites of the Sattasvaara formation.

These results are consistent with the Sm-Nd mineral ages of 2435±41 Ma, 2430±26 Ma and 2052±25 Ma obtained for the Koitelainen, Akanvaara and Keivitsa intrusions, respectively (Hanski et al. in press, Huhma et al. 1996). All these intrusions are characterized by low initial epsilon values (-2.0, -2.2, -3.5), which suggests major contribution from enriched Archean lithosphere in their genesis.

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Appendix I. U-Pb mineral age data from layered mafic intrusions in central Lapland.

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES / Ma		
density, mesh, abrasion	(mg)	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A250 Kiviaapa metatuffite															
A. titanite 3.4-3.5 a1h	29.8	117.6	96.9	753.1	0.772	0.4723	0.80	10.816	0.80	0.1661	0.17	0.98	2493	2507	2518
B. +4.6 rounded a1h	8.3	151.3	94.0	10811	0.153	0.5334	0.65	14.384	0.65	0.1956	0.17	0.97	2755	2775	2789
A251 Koitelainen quartz monzonite															
A. +4.6 a3h	20.3	291.5	143.3	13042	0.141	0.4386	0.65	9.490	0.65	0.1570	0.17	0.97	2344	2386	2423
B. 4.2-4.6 +200 a3h	18.1	418.3	209.1	8991	0.151	0.4418	0.65	9.577	0.65	0.1572	0.15	0.97	2358	2394	2426
C. 4.0-4.2 +150 a5h	12.3	576.5	286.5	17921	0.155	0.4392	0.65	9.503	0.65	0.1569	0.15	0.97	2347	2387	2423
D. 3.8-4.0 +150 a5h	11.1	927.3	452.9	13790	0.166	0.4277	0.65	9.249	0.65	0.1569	0.15	0.97	2295	2362	2422
E. 3.6-3.8 a3h	6.0	1770.0	767.0	12529	0.162	0.3809	0.65	8.144	0.65	0.1551	0.15	0.97	2080	2247	2403
A252 Koitelainen quartz monzonite															
A. +4.6 a3h	22.7	97.1	51.3	2645	0.184	0.4498	0.65	9.795	0.65	0.1580	0.15	0.97	2394	2415	2433
B. 4.2-4.6 a3h	20.3	198.9	100.9	5511	0.156	0.4447	0.65	9.697	0.65	0.1582	0.16	0.97	2371	2406	2436
C. 4.0-4.2 +200 a4h brown	4.0	662.1	333.2	9598	0.145	0.4466	0.65	9.714	0.65	0.1578	0.15	0.97	2380	2408	2431
A253 Keskilaki metadiabase															
A. zr+badd 3.8-4.0 a	1.7	1907.0	829.0	3827	0.33	0.3434	0.70	5.875	0.70	0.1241	0.15	0.98	1902	1957	2016
B. total a	2.6	622.2	260.8	4090	0.29	0.3382	0.65	6.338	0.65	0.1359	0.16	0.97	1878	2023	2176
C. zr+badd a	3.0	433.6	198.6	4210	0.218	0.3881	0.65	7.489	0.65	0.1400	0.16	0.97	2113	2171	2226
A555a Kiviaapa paragneiss															
A. +4.6 +100	11.0	159.0	96.1	8973	0.128	0.5287	1.00	14.557	1.00	0.1997	0.15	0.99	2736	2786	2823
B. +4.6 long crystals	6.7	229.6	131.5	4377	0.101	0.5096	0.65	13.497	0.65	0.1921	0.20	0.95	2654	2715	2760
C. +4.6 100-200	20.8	204.7	117.1	20409	0.115	0.5083	0.65	13.460	0.65	0.1921	0.15	0.97	2649	2712	2760
A555b Kiviaapa paragneiss															
A. +4.6 +100	9.2	117.3	67.6	7932	0.123	0.5054	0.65	13.718	0.65	0.1969	0.15	0.97	2637	2730	2800

Appendix I. continued

Sample information density, mesh, abrasion	Sample (mg)	U ppm	Pb ppm	$^{206}Pb/^{204}Pb$ measured	$^{208}Pb/^{206}Pb$ radiogenic	ISOTOPIC RATIOS* $^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%	Rho**	APPARENT AGES / Ma $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A556 Kiviaapa paragneiss															
A. +4.6 +100 rounded a5h	10.1	138.8	86.3	8959	0.16	0.5293	0.65	14.504	0.65	0.1987	0.15	0.97	2738	2783	2815
B. +4.6 100-150 euh clear a	6.0	120.6	72.8	9514	0.155	0.5197	0.65	13.616	0.65	0.1900	0.15	0.97	2698	2723	2742
C. +4.6 100-150 euh dark a	3.7	227.0	130.3	6923	0.118	0.5064	0.65	13.316	0.65	0.1907	0.16	0.97	2641	2702	2748
A557 Kiviaapa meta-arkose															
A. +4.6 -200 a3h	10.5	226.5	132.5	12931	0.141	0.5090	0.65	13.341	0.65	0.1901	0.15	0.97	2652	2704	2743
B. +4.6 +100 rounded a2h	4.9	150.1	92.4	8450	0.15	0.5290	0.65	14.452	0.65	0.1981	0.15	0.97	2737	2779	2811
C. +4.6 +150 sharp a	6.8	163.6	97.7	6575	0.139	0.5180	0.65	13.683	0.65	0.1916	0.15	0.97	2690	2728	2756
A566 Matarakoski porphyry															
B. 4.2-4.6	21.6	199.0	77.4	11045	0.245	0.3286	0.65	5.687	0.65	0.1255	0.15	0.97	1831	1929	2036
C. +4.6 a4h	11.5	131.8	57.4	7639	0.237	0.3691	0.65	6.431	0.65	0.1264	0.15	0.97	2025	2036	2048
C.2. +4.6 a3h	12.2	128.9	56.3	31381	0.241	0.3709	0.65	6.472	0.65	0.1266	0.17	0.97	2033	2042	2050
D. 4.2-4.6 a3h	10.6	174.0	75.5	36613	0.263	0.3629	0.65	6.300	0.65	0.1259	0.16	0.97	1995	2018	2041
E. 4.0-4.2 +200 a	1.0	489.7	207.0	9217	0.274	0.3512	0.65	5.991	0.65	0.1238	0.20	0.95	1940	1974	2011
G. +4.6 clear a3h	0.1	138.0	58.0	7498	0.235	0.3598	0.65	6.253	0.65	0.1261	0.16	0.97	1981	2012	2044
H. +4.6 clear	0.6	152.0	62.0	35792	0.243	0.3469	0.65	6.010	0.65	0.1257	0.30	0.89	1920	1977	2038
A580 Kaitaselkä granophyre															
A. 3.8-4.1 -100	16.8	1317.0	501.0	2971	0.275	0.3075	0.65	6.375	0.65	0.1504	0.15	0.97	1728	2028	2350
B. 3.6-3.8	19.6	1646.0	500.0	2740	0.243	0.2503	0.65	5.113	0.65	0.1482	0.15	0.97	1440	1838	2324
C. 3.8-4.1 +100	35.0	1284.0	500.0	4232	0.288	0.3116	0.65	6.502	0.65	0.1514	0.15	0.97	1748	2046	2361
D. 3.8-4.1 -100	10.7	1371.0	518.0	1499	0.276	0.3015	0.65	6.229	0.65	0.1498	0.20	0.95	1698	2008	2344
E. 3.6-3.8 magn	9.6	1719.0	513.0	2521	0.241	0.2475	0.65	5.028	0.65	0.1473	0.15	0.97	1425	1823	2315
F. +4.1 +200 transp a2h	4.1	753.7	383.0	2069	0.284	0.4010	0.65	8.544	0.65	0.1545	0.15	0.97	2173	2290	2396
G. 3.8-4.1 -100 a5h	5.0	1315.0	630.0	10007	0.299	0.3808	0.70	8.055	0.70	0.1534	0.15	0.98	2080	2237	2384
H. 3.8-4.1 -100 magn a5h	4.9	1359.0	605.0	6283	0.294	0.3556	0.70	7.487	0.70	0.1527	0.15	0.98	1961	2171	2376
I .4.0-4.2 +100 transp a3h	3.1	1428.0	589.0	2670	0.289	0.3272	0.65	6.822	0.65	0.1513	0.15	0.97	1824	2088	2360
A604 Lakijänkä gabbro															
A. +4.2 +100 a1h	8.3	601.0	310.0	7856	0.179	0.4461	0.65	9.720	0.65	0.1580	0.15	0.97	2377	2408	2434
B, 3.9-4.2 +100	11.0	1513.0	728.0	5742	0.175	0.4169	0.65	9.012	0.65	0.1568	0.15	0.97	2246	2339	2421

Appendix I. continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES / Ma		
density, mesh, abrasion	(mg)	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
C. +4.2 -100	16.9	650.0	322.0	3575	0.158	0.4325	0.65	9.391	0.65	0.1575	0.15	0.97	2317	2376	2428
D. +4.2 +100 a5h	6.4	622.0	331.0	6885	0.182	0.4593	0.65	10.015	0.65	0.1582	0.16	0.97	2436	2436	2436
E. 3.9-4.2 +100 a5h	4.4	1431.0	707.0	10724	0.173	0.4299	0.65	9.308	0.65	0.1570	0.16	0.97	2305	2368	2424
F. +4.2 +150 red cores a2h	6.1	326.0	175.0	6187	0.193	0.4571	0.65	9.989	0.65	0.1585	0.15	0.97	2426	2433	2439
G. +4.2 100-150 long a2h	8.4	648.0	333.0	4568	0.168	0.4453	0.80	9.669	0.80	0.1575	0.15	0.98	2374	2403	2429
H. +4.2 +100 thin plates a2h	11.3	703.0	373.0	3287	0.21	0.4448	0.70	9.677	0.70	0.1578	0.16	0.97	2372	2404	2432
I. +4.2 +100 short dark a5h	6.9	670.0	350.0	6566	0.184	0.4505	0.65	9.804	0.65	0.1580	0.15	0.97	2395	2416	2434
J. +4.2 +100 long a3h	10.4	638.0	327.0	4524	0.18	0.4398	0.65	9.572	0.65	0.1579	0.15	0.97	2349	2394	2432
K. +4.2 +200 red cores a1h	3.9	354.0	190.0	5996	0.201	0.4548	0.65	9.942	0.65	0.1586	0.15	0.97	2416	2429	2440
L. +4.2 100-150 long a3h	8.9	630.0	321.0	9535	0.159	0.4484	0.65	9.764	0.65	0.1579	0.15	0.97	2388	2412	2433
A605 Lakijänkä gabbro															
A. 3.8-3.9	7.6	2305.0	945.0	9558	0.142	0.3667	0.65	7.799	0.65	0.1543	0.15	0.97	2013	2208	2393
B. +4.1	10.4	927.0	455.0	8665	0.169	0.4289	0.65	9.318	0.65	0.1576	0.15	0.97	2300	2369	2430
C. 3.9-4.1	9.2	1723.0	817.0	13418	0.148	0.4216	0.65	9.121	0.65	0.1569	0.15	0.97	2267	2350	2422
D. 3.9-4.1 a5h	5.5	1723.0	843.0	26268	0.149	0.4343	0.65	9.421	0.65	0.1573	0.15	0.97	2325	2379	2427
E. 3.8-3.9 a3h	3.0	2321.0	1012.0	22862	0.133	0.3925	0.65	8.398	0.65	0.1552	0.15	0.97	2134	2274	2403
F. 3.9-4.1 a8h	4.0	1782.0	889.0	41994	0.152	0.4422	0.65	9.608	0.65	0.1576	0.15	0.97	2360	2397	2430
A666 Vaiskonselkä granophyre															
A. 4.0-4.2 +200	15.9	1192.0	599.0	10763	0.222	0.4219	0.65	9.102	0.65	0.1565	0.15	0.97	2269	2348	2418
B. 3.8-4.0 +200	8.7	1539.0	755.0	15125	0.215	0.4140	0.65	8.926	0.65	0.1564	0.15	0.97	2233	2330	2416
C. 3.6-3.8 +100	11.4	1744.0	789.0	10700	0.221	0.3808	0.65	8.144	0.65	0.1551	0.15	0.97	2080	2247	2403
D. 4.0-4.2 +200 a5h	5.1	1176.0	612.0	9954	0.225	0.4351	0.65	9.421	0.65	0.1571	0.15	0.97	2328	2379	2424
E. 3.8-4.0 +200 a5h	4.4	1610.0	834.0	17596	0.226	0.4344	0.65	9.419	0.65	0.1573	0.15	0.97	2325	2379	2426
A1310 Tuorelehto, Savukoski, gabbro															
A. +4.3/-200/a5h	5.3	346.0	124.0	7414	0.18	0.4159	0.60	8.771	0.60	0.1530	0.15	0.97	2241	2314	2379
B. +4.3/+200	8.1	372.0	135.0	10652	0.18	0.4197	0.60	8.964	0.60	0.1549	0.12	0.98	2259	2334	2401
C. 4.2-4.3	8.2	500.0	184.0	14265	0.21	0.4261	0.60	9.144	0.60	0.1557	0.12	0.98	2287	2352	2409
D. 4.2-4.3	5.9	483.0	180.0	9292	0.21	0.4298	0.60	9.232	0.60	0.1558	0.12	0.98	2304	2361	2410
E. 4.2-4.3 a5h	6.0	481.0	181.0	15093	0.21	0.4354	0.60	9.368	0.60	0.1561	0.12	0.98	2329	2374	2413
F. 4.2-4.3 a16h (M)	0.6	525.0	200.0	18424	0.22	0.4402	0.50	9.511	0.50	0.1567	0.12	0.97	2351	2389	2421

Appendix I. continued

Sample information	Sample	U	Pb	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	ISOTOPIC RATIOS*						Rho**	APPARENT AGES / Ma		
density, mesh, abrasion	(mg)	ppm		measured	radiogenic	$^{206}Pb/^{238}U$	2SE%	$^{207}Pb/^{235}U$	2SE%	$^{207}Pb/^{206}Pb$	2SE%		$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A1311 Akanvaara, Savukoski, granophyre															
A. 4.3-4.5 a3h	9.2	414.0	147.0	9841	0.13	0.4100	0.60	8.519	0.60	0.1507	0.12	0.98	2215	2287	2353
B. 4.3-4.5	7.7	415.0	147.0	7926	0.13	0.4108	0.60	8.508	0.60	0.1502	0.13	0.98	2218	2286	2348
C. 4.2-4.3	7.6	512.0	176.0	10090	0.14	0.3970	0.60	8.113	0.60	0.1482	0.12	0.98	2155	2243	2325
D. +4.5	1.3	344.0	123.0	3838	0.14	0.4146	0.70	8.643	0.70	0.1512	0.13	0.98	2235	2301	2359
E. 4.3-4.5 a16h	5.3	431.0	157.0	16499	0.14	0.4208	0.60	8.805	0.60	0.1518	0.12	0.98	2264	2318	2366
A1390 Keivitsa olivine pyroxenite (M = carefully selected crystals analysed by I.Mänttäri):															
A. 4.0-4.2	1.5	1911.0	615.0	2454	0.32	0.3717	0.60	6.484	0.60	0.1265	0.16	0.96	2037	2043	2050
E. 4.0-4.2	0.7	1561.0	490.0	2872	0.36	0.3644	0.60	6.346	0.60	0.1263	0.16	0.96	2002	2024	2047
F. 4.0-4.2 (M)	0.1	1711.0	551.0	47743	0.34	0.3722	0.50	6.514	0.50	0.1269	0.10	0.98	2040	2047	2056
G. 4.0-4.2 (M)	0.1	2156.0	684.0	36016	0.29	0.3667	0.50	6.398	0.50	0.1266	0.10	0.98	2014	2032	2051
H. 4.0-4.2 (M)	0.1	1524.0	480.0	54741	0.3	0.3633	0.50	6.336	0.50	0.1265	0.10	0.98	1998	2024	2050
I. +4.3 a0.3h (M, 2 grains)				10466	0.33	0.3672	0.70	6.413	0.70	0.1267	0.10	0.99	2016	2034	2052
A1445 Keivitsa diorite dyke															
A. +4.3	5.2	415.0	124.0	7961	0.32	0.3446	0.60	5.992	0.60	0.1261	0.12	0.98	1908	1974	2044
B. +4.3 a3h	5.1	388.0	118.0	11666	0.31	0.3518	0.60	6.127	0.60	0.1263	0.12	0.98	1943	1994	2047
C. 4.2-4.3 +200	6.6	584.0	169.0	3065	0.37	0.3340	0.60	5.797	0.60	0.1259	0.12	0.98	1857	1945	2041
D. titanite	8.8	41.4	32.8	1450	1.33	0.3672	0.70	6.383	0.70	0.1261	0.20	0.96	2016	2030	2044

*) Isotopic ratios corrected for fractionation, blank (0.5 ng Pb for old method, 0.03 ng Pb for new method) and age related common lead.

**) Error correlation for 207Pb/235U vs. 206Pb/238U ratios.

Radiometric age determinations from Finnish Lapland and their bearing on the timing of Precambrian volcano-sedimentary sequences
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U-Pb ISOTOPIC DETERMINATIONS ON BADDELEYITE AND ZIRCON FROM THE HALTI-RIDNITSOHKKA INTRUSION IN FINNISH LAPLAND: A FURTHER CONSTRAINT ON CALEDONIDE EVOLUTION

by
Matti Vaasjoki and Pekka Sipilä

Vaasjoki, M. & Sipilä, P. 2001. U-Pb isotopic determinations on baddeleyite and zircon from the Halti-Ridnitsohkka intrusion in Finnish Lapland: a further constraint on Caledonide evolution. *Geological Survey of Finland, Special Paper 33,* 247-253. 3 figures and one table.

A concordant baddeleyite from gabbro sills intersecting metasediments in the Halti-Ridnitsohkka area in NW Finland yields a U-Pb age of 434±5 Ma. Probably xenocrystic, but thermally reset zircons from the same rock give $^{207}Pb/^{206}Pb$ ages in the range of 441-451 Ma. A zircon fraction from an anatectic granite within the contact metamorphic aureole of the mafic intrusion appears younger on the concordia diagram, but has a $^{207}Pb/^{206}Pb$ age of 457±17 Ma, which indicates at least at partial derivation from the metasedimentary host rocks. The baddeleyite age estimate (434 Ma) is a minimum for the deposition of the country rocks, which may possibly be correlated with the Guolasjav'ri sequence, dated by fossil evidence to be Silurian-Middle Ordovician in age. The age of the gabbro sills also sets a maximum limit for the thrust which bounds the Halti-Ridnitsohkka allochton. On the basis of the present radiometric data, it is apparent that mafic rocks within the Seiland province are considerably older than the Halti gabbros and therefore also the correlation of sedimentary rocks between the Halti and Seiland areas is questionable. Thus the stratigraphic position of the Halti-Ridnitsohkka sequence is interpreted to be in the Upper Allochton of the Caledonides instead of the Middle Allochton as formerly thought.

Key words (GeoRef Thesaurus, AGI): absolute age, U/Pb, baddeleyite, zircon, gabbros, sills, granites, dikes, Caledonides, Paleozoic, Halti, Ridnitsohkka, Finland

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GEOLOGICAL SETTING

The magmatic complex at the Halti and Ridnitsohkka fells in Finland forms both topographically and tectonostratigraphically the highest unit within the Finnish Caledonides, and occurs on either side of the Finnish-Norwegian border (Fig. 1). The complex consists in the east of gabbro sills at Ridnitsohkka and in the west of dunite-troctolite-olivine gabbro cumulates at Halti (Fig. 2). The two rock units are considered comagmatic and belong to a subhorizontal allochthonous overthrust plate (Sipilä 1992). Rocks

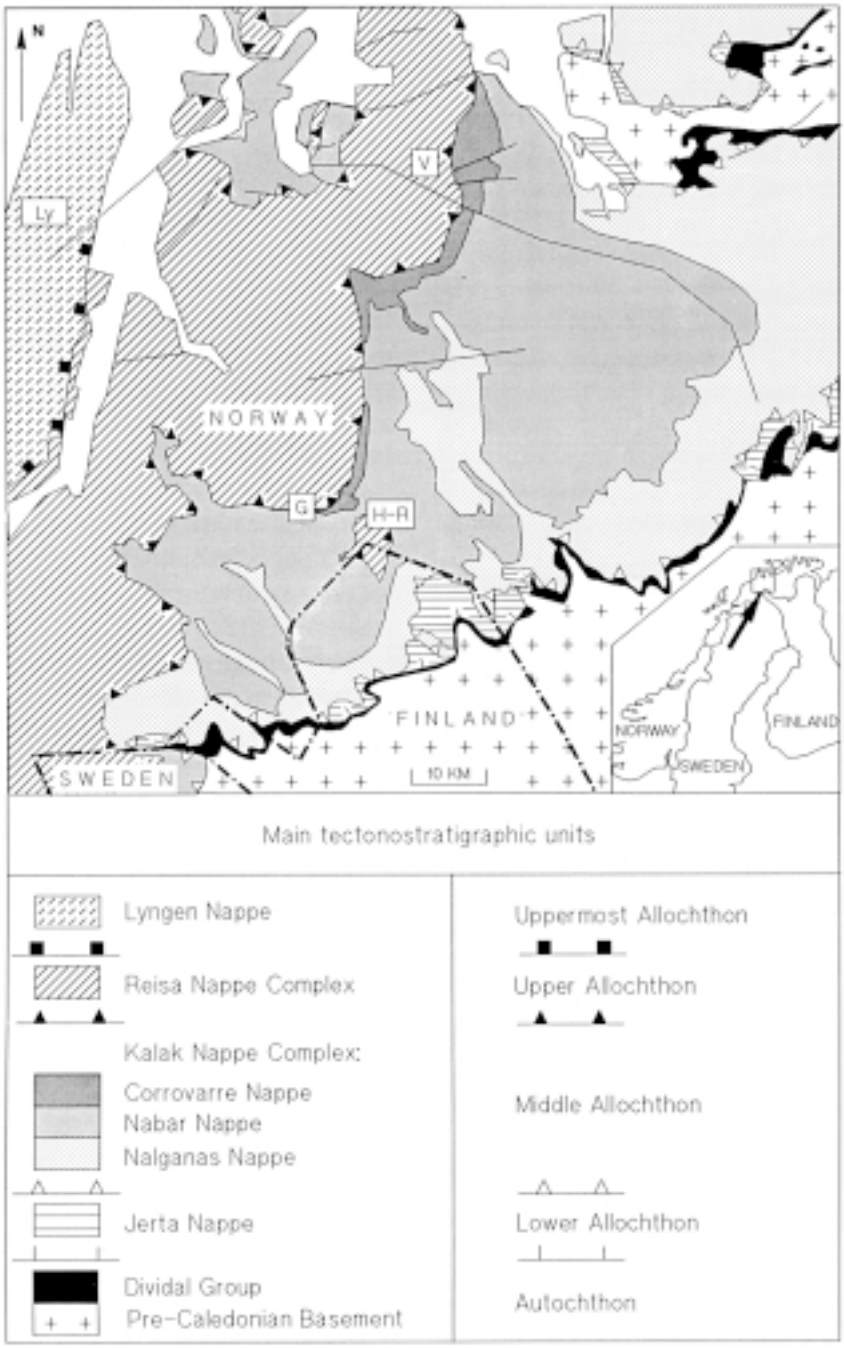

Fig. 1. Tectonostratigraphic map of part of northern Norway (after Zwaan 1988 and Zwaan & Roemund 1990) and of northwestern Finland (after Lehtovaara 1989). The stratigraphic interpretation of the Halti-Ridnitsohkka area is based on the present study and Sipilä (1992). H-R = Halti-Ridnitsohkka, G = Guolasjav'ri, Ly = Lyngen, V = Vaddas.

above the plate have been eroded but the lower contact is clearly tectonic and transgresses the bedding of the overlying rocks. In the sill area, the lower contact is less distinct as mylonitization occurs in a wider zone. East of the sill area, there is a 500 m wide contact metamorphic aureole in the adjacent metasedimentary rocks. In the innermost part of the aureole there occur clearly anatectic granite veins.

The Halti-Ridnitsohkka mafic and ultramafic complex has been investigated by Hausen (1941, 1942), Böe (1976), Lehtovaara (1984, 1987, 1989), Lehtovaara and Sipilä (1987) and Sipilä (1987, 1989, 1991 and 1992) and has usually been assigned to the Middle Allochton of the Scandinavian Caledonides. It was deformed before the last phase of overthrusting, during which the narrower mica-gneiss enclaves were boudinaged and the sills, especially those close to the cumulate massif, developed a lineation parallel to the gneiss enclaves. The magmatic activity occurred apparently in two phases: 1) the intrusion of the

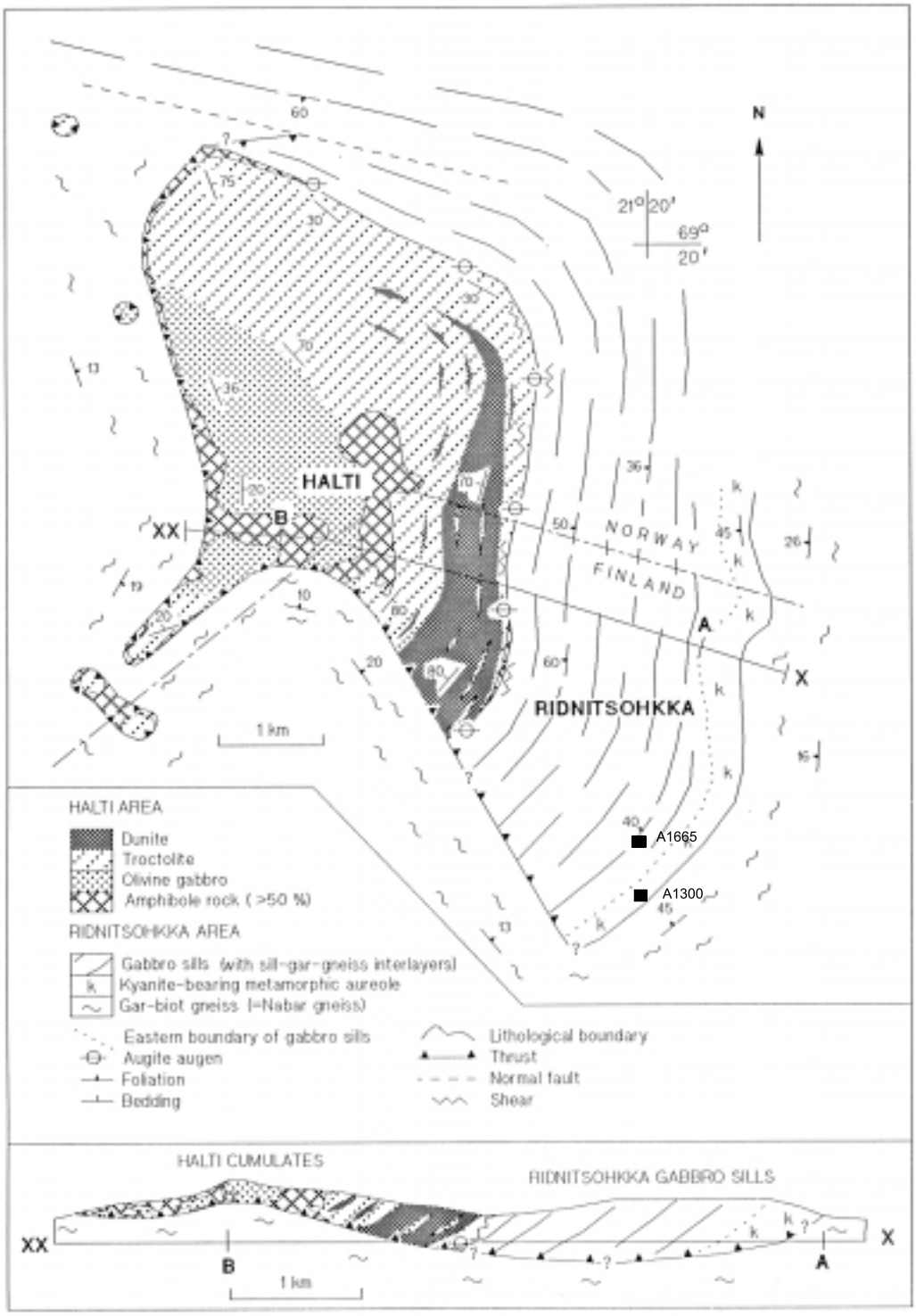

Fig. 2. Simplified geology in the Halti-Ridnitsohkka area. The sample locations are indicated by squares. Based on Sipilä, 1992.

tholeiitic sills at Ridnitsohkka and 2) the emplacement of the layered dunite-troctolite-olivine gabbro cumulates at Halti.

The equigranular microscopic texture of the gabbro sills, the paucity of chilled margins, the partial melting of gneiss enclaves and the contact metamorphic aureole demonstrate that the intrusion occurred fairly deep within the crust. The kyanite-bearing aureole just outside of the gabbro sill area and the sillimanite-bearing gneiss enclaves in the sill area show that pressure during the intrusion of the sills was above the stability range of andalusite (4-5 kb; e.g. Salje 1986). The cumulates at Halti were formed at rather low pressures, because formation of troctolitic rocks at

higher pressures than 5 kb would be impossible as olivine and Ca-rich plagioclase would not equilibrate with a Ca- and Mg-enriched melt (Presnall et al. 1978).

A large number of mafic and ultramafic rocks have been recorded in the North-Scandinavian Caledonides. Diabase dykes occur widely in the Kalak and Laksefjord Nappes thought to belong to the Middle and Lower Allochtons. In eastern Finnmark they have been dated at 640 Ma (Beckinsale et al. 1975) and a Sm-Nd age of 580±30 Ma has been reported from the Corrovarri-nappe north of Halti (Zwaan & van Roemund 1989). These dykes, often occurring in wide-spread swarms, are thought to reflect a crustal extension at that time, and they have been equated with the Ottfjället diabases in the South-Scandinavian Särv Nappe (Townsend & Gayer 1989), whose Ar-Ar age of 665±10 Ma (Claesson & Roddick 1983) has been linked with the opening of the Iapetus Ocean.

The complex Seiland magmatic province begins only 50 km north of Halti, and apparently formed deep in the Earth's crust during an extensional phase (Townsend & Gayer 1989). Its mafic-alkaline rocks cross cut the sediments of the Kalak Nappe complex (ibid.). The Seiland province has generally been considered a synorogenic product of the Finnmarkian orogeny (e.g., Robins & Gardener 1975) but some researchers regard it as pre-orogenic in respect to the Scandian orogeny (e.g. Krill and Zwaan, 1987) and question the orogenic status of the Finnmarkian extensional episode (cf. Binns 1989, Townsend & Gayer 1989).

The magmatism in the Seiland province commenced with the intrusion of subalkaline tholeiitic gabbros and continued by fractional crystallization resulting in the emplacement of alkali-olivine basaltic and picritic complexes at pressures of 6-10 kb (Robins & Gardener 1975, Bennet et al. 1986). These metamorphosed and deformed rocks are cut by alkaline nepheline-syenite dykes, whose zircons on Seiland yield concordant U/Pb ages of 530±2 Ma but are about 10 Ma younger on the nearby island of Söröy (Pedersen et al. 1989).

Mattson and Lindqvist (1990) reported a tentative Sm-Nd age of 650±115 Ma and an ε_{Nd} of +4.4 from four whole rock samples of the Halti cumulates. From Ridnitsohkka, two mineral separates and two whole rock samples gave a similar linear trend, indicating a slightly lower ε_{Nd} value.

SAMPLE MATERIAL AND ANALYTICAL PROCEDURES

The samples for the U-Pb determinations were collected from a gabbro sill and an apparently anatectic granite dyke, both cross-cutting the metasediments at Ridnitsohkka.

The sills consist of a sub-ophitic gabbro where plagioclase (An_{50-60}) and clinopyroxene, partly altered to amphibole and biotite, are the major minerals. The gabbro sample (Finnish map grid Northing = 7688.90; Easting = 1513.50) consisted of medium and fine-grained portions which grade into each other over a distance of 1 cm. The 60 kg sample contained baddeleyite and zircon. The latter consisted mainly of large crystals as crystal faces are rare in the coarse (+150 m) fractions indicating that the mineral grains were broken during sample processing.

The anatectic granite (7688.42; 1513.67) consists mainly of potassium feldspar, plagioclase (An_{20-30}), quartz and biotite. Its zircons are fine-grained (<100 μm), euhedral and without any visible zonation.

Both zircon and baddeleyite were separated using a shaking table, heavy liquids and magnetic means. The final purification was done by hand-picking under a stereoscopic microscope. The analyzed samples, ranging from 1-5 mg in weight, were prepared for analysis using the methods developed by Krogh (1973). The U and Pb concentrations were determined by isotope dilution. The errors in the age estimates are reported on the 2σ level.

The results are summarized in Table1 and presented graphically on a concordia diagram in Figure 3. Both the baddeleyite and the zircon fractions from the gabbro sill yield concordant U/Pb ages ranging from 434 to 443 Ma. As zircon should not form in a silica-deficient magma, and because of the large grain-size of the original crystals, we consider that the zircon may be xenocrystic, particularly as the $^{207}Pb/^{206}Pb$ ages are between 441 and 451 Ma. The baddeleyite, however, is a mineral characteristic of silica-deficient melts, and in this case it also exhibits the euhedral form expected of a primary mineral. On these grounds, we argue that the baddeleyite data from Ridnitsohkka especially is worth serious geological consideration. The fact that the zircon results are almost identical to the baddeleyite data is a result of the high temperature of the basaltic magma, which resulted in an almost complete loss of previously formed radiogenic lead from the probably xenocrystic zircons.

The single zircon fraction from the granite A1300 is marginally discordant and, consequently, appears younger on the concordia diagram than the analyses

Table 1. U-Pb mineral analyses from Ridnitshohkka.

Fraction	Weigth (mg)	U conc (ppm)	Pb conc (ppm)	$^{206}Pb/^{204}Pb$ (meas.)	$^{208}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	±2 SE (%)	$^{207}Pb/^{235}U$	±2 SE (%)	$^{207}Pb/^{206}Pb$	±2 SE (%)	Corr.	$T_{206/238}$	$T_{207/235}$	$T_{207/206}$
A1165-Ridnitshohkka gabbro sill															
A. Baddeleyite	3.0	815.5	57.5	661	0.018	0.0698	0.89	0.5335	0.89	0.05554	0.25	0.96	434	434	434
B. +4.5/+200	10.8	567.1	47.2	2802	0.280	0.0712	1.74	0.5477	1.72	0.05581	0.32	0.98	443	443	445
C. 4.3-4.5	6.3	777.4	66.4	2626	0.316	0.0710	0.65	0.5478	1.00	0.05596	0.46	0.93	442	443	451
D. 4.2-4.3	6.8	2000.1	184.9	5764	0.453	0.0706	0.65	0.5422	0.66	0.05571	0.16	0.97	439	439	441
A1300-Ridnitshohkka anatectic granite vein															
A. +4.3	1.8	1646.6	229.5	742	1.236	0.0677	1.62	0.5235	1.66	0.05612	0.75	0.89	422	427	457

Isotopic ratios corrected for blank and age related common lead

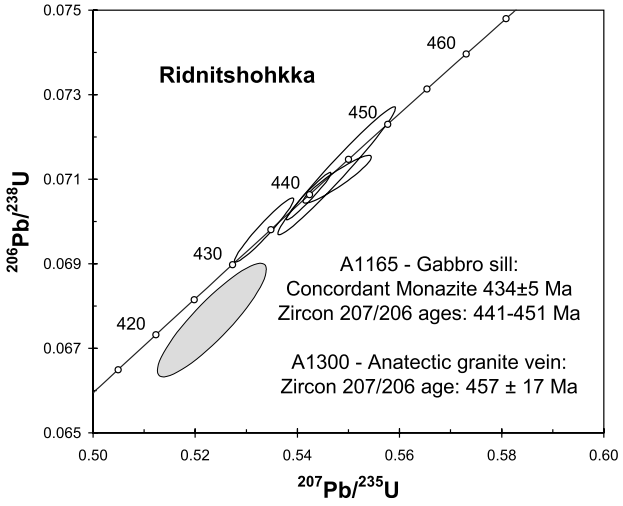

Fig. 3. A concordia diagram depicting the results of the baddeleyite and zircon analyses from the Ridnitsohkka area. The size of the ellipses indicates experimental error. The analysis from the granite vein A1300 is shown by a filled ellips.

from the gabbro sill. Its $^{207}Pb/^{206}Pb$ age is, however, 457±17 Ma which may indicate an older inherited lead component within the mineral. The zircon result from the granite supports nevertheless the results obtained from the gabbro sill.

DISCUSSION

Ages similar to those reported in this paper have been recorded from Caledonian mafic intrusions on several occasions: Dunning and Pedersen (1988) reported a 443±3 Ma zircon result from the Solund ophiolite in southern Norway. They associated the emplacement of the Solund complex with a near-continental basin environment which formed just before the closure of the Iapetus Ocean. In central Scandinavia Wilson et al. (1983) give an age of 426±5 Ma for felsic differentiates of the synorogenic Fongen-Hyllingen layered basic complex. Senior and Andriassen (1990) determined a zircon age of 434±5 Ma from the syndeformational Artfjället gabbro. In northernmost Norway the layered mafic/ultramafic Honningsvåg intrusive suite was emplaced into early Silurian flysch-type metasediments within the Mageröy Nappe during the Scandian orogeny (Robins et al. 1986). Tucker et al. (1990) measured a U-Pb zircon age of 437±1 Ma from a gabbro-norite from the Råna intrusion and drew attention to chronological similarities between the Caledonides and the Appalachians. Dunning et al. (1990) recorded ages in the range of 440 to 410 Ma in Newfoundland, and argued the Scandian-aged event was important along much of the strike length of the Appalachian-Caledonian Orogen, and that this event is demonstrable separable from Ordovician and Devonian orogenic pulses.

The present geological observations and radiometric dating from the Halti-Ridnitsohkka intrusion in Finland suggest that mafic magmatism at 425-445 Ma was more wide-spread in the Caledonides than previously thought. Our result on the baddeleyite from sample A1165 demonstrates that the intrusion age of the Halti-Ridnitsohkka mafic-ultramafic complex is most likely 434±5 Ma. As the intrusion must have occurred at a considerable depth and as a definite contact metamorphic aureole containing anatectic granites of a similar age exists in the mica gneisses surrounding the gabbro sill area, the age of the gabbro sills also gives a minimum age for the mica gneisses and a maximum age for the overthrust which resulted in the final emplacement of the Halti-Ridnitsohkka allochton. This implies that the mica gneisses at Ridnitsohkka could be of similar age as the Guolas series, which has been dated by fossil evidence to be Silurian or Middle Ordovician in age (Binns & Gayer 1980) and is separated from the Middle Allochton by the Stordalen thrust. As the type locality of the Guolas series (Guolasjav'ri) lies less than 10 km from Halti (Fig. 1), the overthrust surface observed below the Halti complex (Fig. 2) is an extension of the Stordalen thrust underlying the Guolas sequence (Binns 1989) and the Halti complex is part of the Upper Allochton.

In contrast, concordant zircon data from the nearby Seiland magmatic province (Pedersen et al. 1989) demonstrates that cross-cutting mafic rocks in this area are at least 90-100 Ma older, and consequently set a minimum for those metasediments. Moreover, the prolonged Stjernöy stage of magmatism lasted from 700 to 520 Ma (Daly et al. 1990), and thus supracrustal rocks in the Seiland province may be considerably older than the rocks of the Halti-Ridnitsohkka allochton.

CONCLUSIONS

The results presented in this paper show that the gabbro sills of the Halti-Ridnitsohkka complex were emplaced 434±5 Ma ago. This is also the minimum age for the sedimentation of the adjacent supracrustal rocks and a maximum age for the overthrusting of the Halti-Ridnitsohkka plate. Consequently, these rocks belong to the Upper Allochton of the Caledonides and may be tectonostratigraphically equated with the Guolas series, which, on the basis of fossil evidence, has been dated to be of an Silurian-Upper Ordovician age. As mafic rocks in the Seiland region have been reliably dated to be at least 520 Ma old, a correlation between metasediments in the Halti and Seiland areas is not feasible.

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GEOCHRONOLOGY OF NORTHERN FINLAND: A SUMMARY AND DISCUSSION

by
Eero Hanski, Hannu Huhma and Matti Vaasjoki

Hanski, E., Huhma, H. & Vaasjoki, M. 2001. Geochronology of northern Finland: a summary and discussion. *Geological Survey of Finland, Special Paper* 33, 255-279. 8 figures and one table.

The isotopic data presented in this volume from northern Finland are summarized. The results show that ages from the Archean basement complexes fall mostly in the range of 2.68 to 2.83 Ga with ages higher than 3.0 Ga being found only at one location (Tojottamanselkä gneiss dome). The Paleoproterozoic magmatism occurred as distinct episodes which in this review are divided into the following age groups: c. 2.43-2.50 Ga intrusions and metavolcanics, 2.22 Ga differentiated mafic sills, 2.14 Ga mafic intrusions, c. 2.05 Ga intrusions and dykes, c. 2.0 Ga mafic and felsic dyke rocks, 1.91-1.92 Ga felsic plutons and dykes, c. 1.88 Ga synorogenic plutonism and volcanism, and c. 1.80 Ga (postorogenic) granitoids. Isotopic data on bulk detrital zircons can be used to distinguish quartzites from different stratigraphical levels: zircons in the Sodankylä and Kivalo Groups quartzites were derived exclusively from Neoarchean sources while the younger Lainio and Kumpu Group quartzites contain a significant zircon component from c. 1.9 Ga synorogenic plutonic rocks. Comparison and correlation of certain magmatic and sedimentological phases is made beyond the western and eastern borders of northern Finland and some problems in the available geochronological data and in their relation to geological observations are discussed.

Key words (GeoRef Thesaurus, AGI): geochronology, stratigraphy, Precambrian, Archean, Paleoproterozoic, Peräpohja, Lapland, northern Finland.

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INTRODUCTION

This paper attempts to summarize age determinations from northern Finland presented in this volume together with those published earlier on. The focus of the review is on the segment of northern Finland extending from the Peräpohja Schist Belt to the Lapland Granulite Belt, that is the area from where the

dated rocks dealt with in this volume have been collected.

The bulk of the available age determinations from northern Finland have been made over a period of tens of years using the conventional multi-grain U-Pb method on zircon. With the exception of a few newer analyses, the sample size has been several milligrams of handpicked zircon. Major problems encountered in these datings are not purely analytical, but concern the discordancy and heterogeneity of the isotopic data. Reasons for scatter in multigrain conventional data include primarily heterogeneous zircon populations (detrital or inherited zircons) and metamorphic effects. It is known that zircon will stay as a closed system at high temperatures (e.g., 800°C), but may lose Pb if it experiences extended time periods at low temperatures, especially when abundant hydrothermal fluid is available (Mezger & Krogstad 1997). In addition to the above mentioned problems related to zircon dating, there are occasionally difficulties in determining the primary rock types in highly metamorphosed and tectonized lithologies, i.e. whether a sample in question is truly magmatic with magmatic zircon or contains a detrital component.

A compilation of zircon and baddeleyite ages on igneous rocks from northern Finland is shown as histograms in Figure 1 together with comparative data from northern Sweden. It should be emphasized that the errors of these age determinations have not been taken into account in the histograms, but the ages are simply divided into ten-million-year categories on the face value basis. Thus several problematic age results, discussed in more detail later on, are included in the data. Nevertheless, the histograms give a useful overview on the general age distribution of the dated igneous rocks and reveal notable peaks of magmatic activity in northern Finland. In the following, the most salient features of these data are discussed in a chronological order starting from the oldest. Particular attention is paid to their implications for regional correlation and to unsolved geochronological problems or discrepancies between the geological and geochronological knowledge. It is namely true with geochronological studies, as with any scientific research, that apart from solving earlier problems new discoveries often create new puzzles or alternative interpretations not fitting perfectly, if at all, with the previous concepts.

GEOLOGICAL BACKGROUND

Before plunging into the geochronology of northern Finland, the geological background and especially the stratigraphical nomenclature employed in two areas, the Peräpohja Schist Belt and the Central Lapland Greenstone Belt, should be shortly made familiar to readers less versed in the geology of northern Finland. The general stratigraphy of the Paleoproterozoic Peräpohja Schist Belt is well-established (Perttunen et al. 1995). The supracrustal rocks are divided into two major units: the lower Kivalo Group containing sedimentary rocks of the orthoquartzite-dolomite association and a few intervening mafic volcanic units, and the upper Paakkola Group consisting of pelitic schists, black schists and minor mafic metavolcanics. These two groups correspond, respectively, to the Jatulian and Kalevian rocks of the traditional classification of the Karelian rocks. They all are younger than the c. 2.44 Ga mafic layered intrusions which were exposed and partly eroded before the deposition of the lowermost sedimentary rocks of the schist belt. On the other hand, the minimum age for the deposition of the Paakkola Group is provided by intersecting Haaparanta Suite plutonic rocks dated at c. 1.89 Ga (Perttunen 1991).

The Paleoproterozoic Central Lapland Greenstone Belt extending from the Salla-Kuusamo area at the eastern border of Finland over Central Lapland to the Finnish-Norwegian border records a sedimentary and volcanic evolution spanning hundreds of millions of years. The general geology of the belt was recently presented by Lehtonen et al. (1998) and the reader is referred to this work for more detailed information. The lower parts of the supracrustal sequence were deposited unconformably on Archean basement gneisses with radiometric ages varying from 3.1 Ga (Kröner et al. 1981) to 2.6 Ga (Meriläinen 1976). The Paleoproterozoic stratigraphy of the belt is shown schematically in Figure 3 of Lehtonen et al. (1998) and in Figure 3 of Hanski (2001, *this volume*). It is divided into seven lithostratigraphic groups which are, from oldest to youngest, the Salla, Onkamo, Sodankylä, Savukoski, Kittilä, Lainio and Kumpu Groups.

The stratigraphic position of the volcanic rocks directly upon the Archean granite gneisses is taken to indicate that the volcanism of the Salla Group was related to an initial stage of rifting of the Archean craton. The volcanic rocks of this stage are mostly subaerially erupted andesites to rhyolites. Metavolcanics rocks forming the next unit, the Onkamo Group, have been discovered in a wide area extending from Salla at the Russian border to near Kolari close to Sweden. In the type area (Onkamojärvi, Salla), the

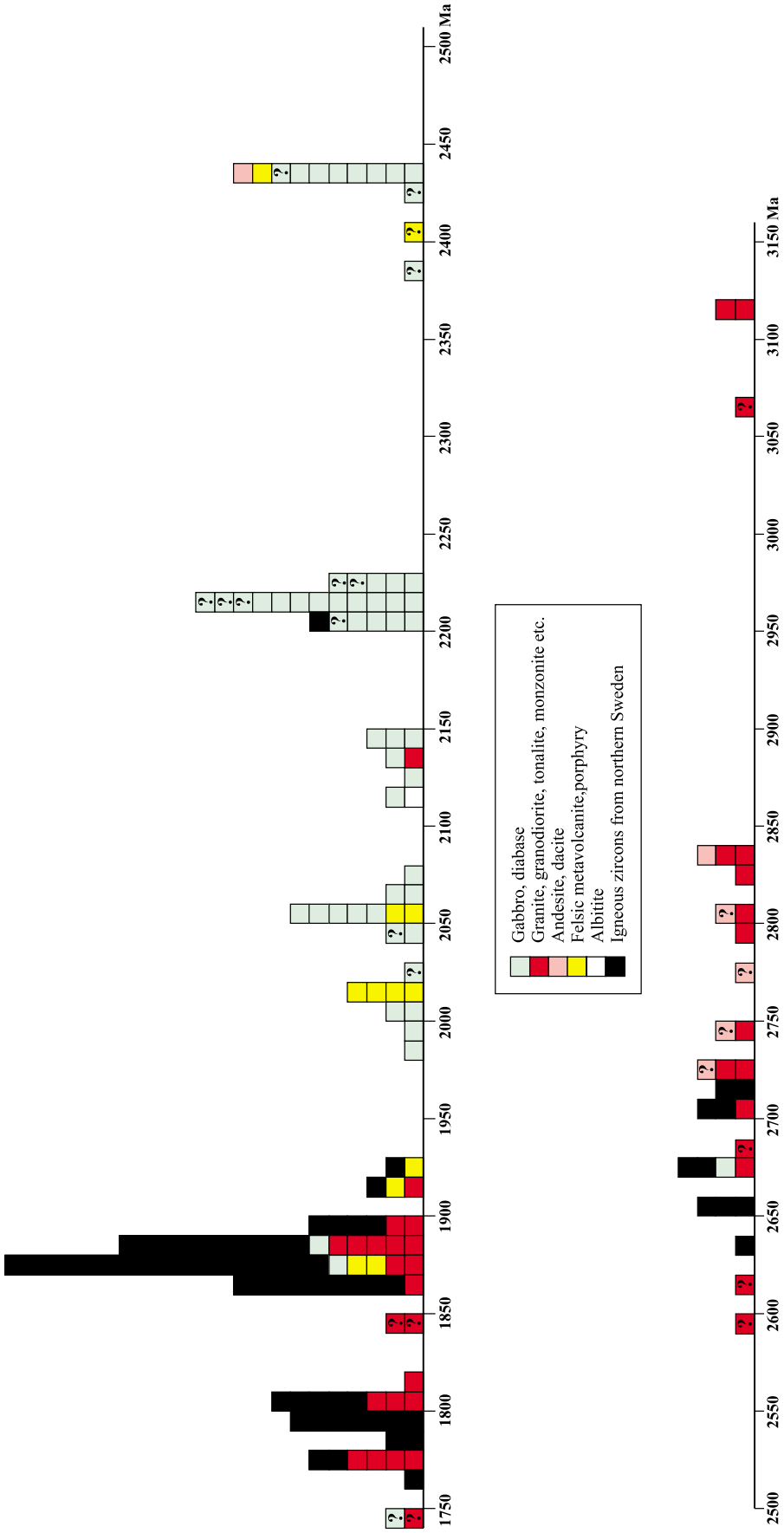

Fig. 1. Histograms of zircon and baddeleyite ages obtained for igneous rocks from northern Finland coupled with zircon ages for igneous rocks from Norrbotten County, northern Sweden (literature source available from the first author). Some ages are labeled with a question mark due to large analytical errors or problems with the interpretation of these ages.

Onkamo Group metavolcanics are represented by high-Mg basalts and andesites. At Möykkelmä, komatiites of the Onkamo Group were erupted discordantly on Archean granite gneisses (Räsänen et al. 1989) and were succeeded by andesitic and basaltic, highly amygdaloidal lavas.

The volcanic activity manifested as the Salla and Onkamo Groups was followed by a more tranquil period during which thick, wide-spread quartz-rich epiclastic sediments were deposited together with lesser amounts of carbonate rocks, pelitic sediments and mafic volcanic rocks. This phase of sedimentation is represented by the supracrustal rocks of the Sodankylä Group. The sediments were deposited either directly on the Archean granite gneisses or on the volcanic rocks of the Salla and Onkamo Groups. The metasediments are penetrated by 2.2 Ga, differentiated gabbro-wehrlite sills in many places. Dolomitic rocks possess positive $\delta^{13}C$ values typical of Jatulian carbonates which is compatible with a sedimentation age of 2.1-2.2 Ga (Karhu 1993).

The Savukoski Group contains graphite- and sulfide-bearing pelitic schists which act as important marker horizons and are best represented by the Matarakoski Formation in Sodankylä. Volcanic activity resumed when minor mafic tuffs deposited together with pelites, while voluminous ultramafic volcanics (komatiites and picrites) erupted at a later stage, the Sattasvaara Formation being the type formation. Komatiites form part of a long ultramafic belt extending from northern Norway through Sodankylä and Salla into Russia. A minimum age for the deposition of the supracrustal rocks of the Savukoski Group, 2.05 Ga, is obtained from the crosscutting Keivitsa intrusion (Mutanen 1997).

The Kittilä Group forms a large, volcanite-dominated rock complex in Central Lapland which, according to geophysical measurements, has a great present-day thickness of up to 6 km. The lower contact of the Kittilä Group is tectonic, as a pronounced shear zone runs close to the western margin of the Sovasjoki gneiss dome. Ophiolitic serpentinite lenses are found along this thrust zone (Hanski 1997). The Kittilä Group includes massive and pillowed mafic lavas, hyalotuffs, breccias and volcanic conglomerates as well as pelitic and chemical metasediments including the Porkonen banded iron formation, indicating deposition in a marine environment. In terms of geochronology, minor 2.0 Ga felsic subvolcanic and volcanic magmatism is an important constituent.

Largely because of different inferred deformation histories, the youngest arenaceous metasediments in northern Finland, constituting the Lainio and Kumpu Groups, were distinguished as two separate units by Lehtonen et al. (1998). Both of them comprise coarse, molasse-like metasediments which unconformably lie on folded formations of older groups. The Lainio Group also contains minor calc-alkaline volcanic rocks with an age of c. 1.88 Ga. As the historical account of Hanski (2001, *this volume*) illustrates, the correlation of these rocks with Karelian formations has long been controversial, but presently they are regarded as post-Jatulian and even as younger than synorogenic plutonic rocks in northern Finland (Rastas et al. 2001, *this volume*).

Paleoproterozoic felsic plutonic rocks cutting supracrustal rocks in northern Finland are divided into two broad groups, the c. 1.9 Ga synorogenic plutons, including the Haaparanta Suite, and postorogenic, c. 1.8 Ga granites. Minor Neoproterozoic magmatism is represented by the Salla and Laanila diabase swarms dated at 1.1-1.0 Ga (Lauerma 1995, Mertanen et al. 1996). The latest, Phanerozoic magmatic events are recorded by the Sokli carbonatite and the Iivaara alkaline intrusion in eastern Finnish Lapland and the Caledonian magmatic rocks in the most northwestern part of Finnish Lapland (Kramm et al. 1993, Vaasjoki & Sipilä 2001, *this volume*).

ARCHEAN ROCKS

As a whole, the presently available, still scarce Archean dates do not seem to indicate a strong magmatic activity at any particular time interval but most of the ages spread evenly between 2.57 and 2.84 Ga. Three deviating age determinations at c. 3.07 and 3.11 Ga come from the small Tojottamanselkä gneiss dome on the northwestern side of the Koitelainen layered intrusion in Central Lapland, the only area which has so far yielded ages older than 3.0 Ga in northern Finland (Kröner et al. 1981). Kröner and Compston (1990) re-examined one of the tonalite samples using single grain SHRIMP analysis. Part of the zircon grains formed a discordia line intersecting Concordia at 3115±29 Ma, which is close to the conventional zircon age of 3110±17 Ma obtained by Kröner et al. (1981) and was interpreted as the time of crystallization of the tonalitic magma. Other zircon grains defined a younger age of 2836±30 Ma reflecting isotopic resetting related to a later metamorphic recrystallization and partial melting event. The younger age, 3069±16 Ma, reported by Kröner et al. (1981) falls between these two events and its meaning re-

mains so far unclear.

Most of the dated Neoarchean rocks are different kinds of granitoids and gneisses. Archean ages for such rocks have been reported from the Tuntsa Basement Complex by Juopperi and Vaasjoki (2001, *this volume*), from the Muonio Basement Complex by Lehtonen (1984) and Väänänen and Lehtonen (2001, *this volume*), from the Pudasjärvi Basement Complex by Perttunen and Vaasjoki (2001, *this volume*) and from the basement rocks south of the Kuusamo schist belt by Lauerma (1982).

Evins et al. (1997, 2000) have delineated an Archean gneiss region tens of kilometers in extent (Suomujärvi Complex) in the SE part of the Central Lapland Granite Complex (Kemijärvi Complex). This was based on the U-Pb zircon ages of c. 2815 and 2754±16 Ma obtained for tonalitic intrusions. Detrital zircons from adjacent quartzites record ages of 2727±13 Ma and c. 3.4 Ga, which are among the oldest zircons dated in the Fennoscandian Shield. The studies carried out by Evins et al. (1997, 2000) suggest that the occurrence of Archean gneisses within the Central Lapland Granite Complex is probably much more wide-spread than previously thought.

There exist five age determinations on rocks thought to be of volcanic origin with a dacitic to rhyolitic chemical composition (Lehtonen et al. 1998, Räsänen & Huhma 2001, *this volume*). Four of them occur close to the northern or eastern margin of the Central Lapland Granite Complex. The locations and ages are as follows: Honkavaara in Sodankylä, 2721±9 Ma, 2746±14 Ma (Rastas et al. 2001, *this volume*); Keski-Loviselkä in Sodankylä, 2775±25 Ma (Räsänen & Huhma 2001, *this volume*); Niskavaara in Posio, 2800±8 Ma (Räsänen & Vaasjoki 2001, *this volume*). All these ages are both technically and geologically problematic. Firstly, zircons in the studied samples represent heterogeneous populations. Secondly, the large age range suggests that the dated rocks do not belong to the same volcanic stage or the ages do not precisely bracket the eruption times of the rocks. Thirdly, at least in the Honkavaara area, the rocks exhibit effects of strong Paleoproterozoic hydrothermal processes, including carbonatization and albitization, masking their original features (Eilu 1994). Fourthly, the volcanic nature of the rocks is not unequivocal in all cases, particularly at Keski-Loviselkä where the felsic rocks, from which zircons were separated, occur as intercalations in greywacke-like metasediments (Räsänen & Huhma 2001, *this volume*). Finally and most importantly, an interpretation that the rocks are Archean metavolcanics is not easy to reconcile with the general geology of the schist belts. For example, at Honkavaara, the dated rocks are surrounded by quartzites of the Sodankylä Group and a suggestion of the presence of an Archean tectonic wedge in the area has been made (Lehtonen et al. 1998, Rastas et al. 2001, *this volume).*

PROTEROZOIC ROCKS

U-Pb zircon and baddeleyite ages determined for both intrusive and volcanic Paleoproterozoic igneous rocks from northern Finland are shown as the upper histogram in Figure 1. Among these rocks, the older part is dominated by mafic rocks while the younger ones are mostly various granitoids. The mafic rocks seem to show distinct groupings at c. 2.44, 2.21, 2.14 and 2.05-2.0 Ga with the last interval probably containing two or more geologically meaningful subgroups. Felsic rocks also appear to display two or three age concentrations at the younger end of the histogram.

C. 2.43-2.50 Ga intrusions and metavolcanics

Large Paleoproterozoic mafic-ultramafic layered intrusions are widespread in the eastern part of the Fennoscandian Shield (Alapieti et al. 1990). Together with outpourings of voluminous subaerial volcanic rocks, ranging in composition from komatiites to rhyolites, they manifest the first major episode of magmatism related to rifting of an Archean craton. In the Russian part of the shield, supracrustal rocks of this pre-Jatulian magmatism are generally assigned to Sumian, Sariolian or Sumi-Sariolian formations (e.g., Zagorodny et al. 1986, Gaskelberg et al. 1986), while in Finnish Lapland they were previously part of the Lapponian rocks but are now incorporated into the Salla and Onkamo Groups (Lehtonen et al. 1998).

Mutanen and Huhma (2001, *this volume*) have dated rocks from the Koitelainen and Akanvaara intrusions occurring in Central Lapland (see Mutanen 1997). A gabbroic pegmatoid and coarse leucocratic gabbro from the Koitelainen and Akanvaara intrusions yielded ages of 2439±3 Ma and 2436±6 Ma, respectively. Comparable results were also obtained for a granophyre and monzodiorite from Koitelainen

(see Table 1). These ages are similar to that (2432±6 Ma) reported for the Kemi intrusion in the Peräpohja area by Perttunen and Vaasjoki (2001, *this volume*) and those published earlier for the Koillismaa intrusions by Alapieti (1982).

Similar ages have been obtained for metavolcanics of the Salla Group in the vicinity of the Koitelainen intrusion. Manninen et al. (2001, *this volume*) determined ages of 2438±8 Ma for a felsic tuff of the Rookkiaapa Formation and 2486±4 Ma for an arkosic gneiss in the Pomokaira area, which they regard as volcanic in origin. (However, due to the uncertainty regarding the origin of the rock, this date is excluded from Fig. 1). Räsänen and Huhma (2001, *this volume*) published an age of 2438±11 Ma for a dacitic metavolcanite at Sakiamaa, being identical with the age of the felsic tuff from the Rookkiaapa Formation.

Geochronological data from Finland and also from the Russian side of the shield (Table 1) thus show that large volumes of continental volcanic rocks were emplaced contemporaneously with the 2.44-2.45 Ga layered intrusions.

Somewhat enigmatic is the age of 2405±6 Ma reported by Silvennoinen (1991) for acid porphyry clasts in a basal conglomerate of the Kuusamo schist belt. This age was obtained by combining results from three separate samples, and the chord and upper intercept is strongly controlled by discordant data points. The four zircon fractions from sample A646 are the least discordant, but plot in a cluster with poorly constrained intercepts. Nevertheless, recalling the dating problems discussed in this volume, an age slightly older than reported is conceivable. An age of 2432±22 Ma has been determined for quartz porphyries in the Paanajärvi area on the Russian side of the border (Buiko et al. 1995).

Another erratic date is the age of 2383±33 reported by Manninen and Huhma (2001, *this volume*) for the Onkamonlehto mafic dyke cutting intermediate to felsic volcanic rocks of the Koutoiva Formation (Salla Group) and high-Mg basalts of the Mäntyvaara Formation (Onkamo Group) in the Salla schist belt. The authors emphasize that the age is technically poor and only provides a minimum age for the intrusion. Nevertheless, the result is highly significant as it also gives a minimum age for the country rocks of the intrusion. It demonstrates that the volcanic rocks adjacent to the border are pre-Jatulian on the Finnish side, whereas on the Russian side, the analogous formations have commonly been considered to be post-Jatulian in age and belong to the Suisaarian rocks (Kulikov et al. 1980, Radchenko et al. 1994).

A great number of age determinations have been performed on Sumian-Sariolian rocks in the Russian part of the shield. These are compared with geochronological data from Finland in Table 1 which calls for several comments. While the Finnish mafic layered intrusions provide ages of c. 2.43-2.44 Ga, data from Russian intrusions indicate at least two distinct episodes of magmatism. In the Karelian block, the U-Pb zircon ages of the intrusions converge to an age of about 2.44-2.45 Ga (Burakovka, Lukkulaisvaara, Tsipringa, Kivakka) (Amelin et al. 1995, Balashov et al. 1993). On the other hand, on the Kola Peninsula, the Imandra lopolith belongs to the 2.44 Ga age group, whereas three others, the Mt. Generalskaya and Fedorovo-Pansky intrusions and the Monche pluton, have yielded higher ages close to 2.5 Ga (Amelin et al. 1995). The dated Sumi-Sariolian volcanic and subvolcanic rocks have in turn produced U-Pb ages in the range of 2.43-2.44 Ga (see Table 1) and are in good agreement with the zircon ages obtained for Salla Group metavolcanics (Manninen et al. 2001, *this volume*, Räsänen & Huhma 2001, *this volume*) and the Sm-Nd isochron ages determined for komatiitic rocks from the Vetreny belt, southern Russian Karelia (Puchtel et al. 1997).

The tectonic regime as well as geological, geochemical and isotopic characteristics of the Sumi-Sariolian and corresponding magmatism in Finland are compatible with an origin related to upwelling of a mantle plume beneath the Archean craton (Amelin et al. 1995, Amelin & Semenov 1996, Puchtel et al. 1997). However, the large spread of U-Pb zircon ages recorded by the layered intrusions is difficult to reconcile with this hypothesis as the magmatism related to a starting mantle plume is generally considered to be very short-lived lasting only a few million years. This question was discussed by Amelin et al. (1995) who concluded that, though the time estimates both for the c. 2.45 Ga and 2.5 Ga magmatic episodes are similar or slightly longer than the duration of younger flood basalt volcanisms, two independent plumes affecting the Fennoscandian Shield are unlikely within a time period of c. 60 Ma.

Since Amelin et al. (1995) published their results, the geochronological ramifications concerning the layered intrusions in the Kola Peninsula has become more obscure, as recently two discrete U-Pb zircon ages were reported from different levels of the same layered intrusion at two locations. The Mt. Generalskaya intrusion was documented by Bayanova et al. (1999a) to contain gabbro-norites with an age of 2496±10 Ma and anorthosites with an age of 2447±10 Ma higher up in the cumulate sequence. Similarly, Mitrofanov and Bayanova (1999) published an age of 2491±1.5 Ma for a gabbro-norite and an age of 2447±12 Ma for an anorthosite in the Fedorovo-

Table 1. U-Pb zircon and baddeleyite ages together with Sm-Nd ages obtained for igneous rocks in the northern and eastern part of the Fennoscandian Shield, which have been dated at ca. 2.43-2.50 Ga.

Locality	Rock type	Age	Method	Reference
Mafic intrusions in Russia				
Burakovka	Gabbro-norite	2449 ± 1.1 Ma	Zr	Amelin et al. (1995)
Lukkulaisvaara	Lherzolite, pyroxenite	2439 ± 11 Ma	Zr+Bd	Balashov et al. (1993)
Lukkulaisvaara	Gabbro-norite	2442.1 ± 1.4 Ma	Zr+Bd	Amelin et al. (1995)
Tsipringa	Gabbro	2441.3 ± 1.2 Ma	Zr+Bd	Amelin et al. (1995)
Kivakka	Gabbro-norite	2445 ± 2 Ma	Zr	Balashov et al. (1993)
Imandra	Gabbro-norite	2396 ± 7 Ma	Zr	Balashov et al. (1993)
Imandra	Gabbro-norite	2446 ± 39 Ma	Zr	Bayanova & Mitrofanov (1999)
Imandra	Gabbro-norite	2441.0 ± 1.6 Ma	Zr	Amelin et al. (1995)
Imandra	Gabbro-norite	2444 ± 77 Ma	Sm-Nd	Balashov et al. (1993)
Monchegorsk	Gabbro-norite	2504.4 ± 1.5 Ma	Zr	Amelin et al. (1995)
Monchegorsk	Gabbro-norite	2493 ± 7 Ma	Zr	Balashov et al. (1993)
Monchegorsk	Gabbro-norite	2482 ± 48 Ma	Sm-Nd	Tolstikhin et al. (1992)
Monchegorsk	Pyroxenite	2492 ± 31 Ma	Sm-Nd	Tolstikhin et al. (1992)
Fedorovo-Pansky	Gabbro-norite	2501.5 ± 1.7 Ma	Zr	Amelin et al. (1995)
Fedorovo-Pansky	Gabbro-norite	2470 ± 9 Ma	Zr	Balashov et al. (1993)
Fedorovo-Pansky	Gabbro-norite	2487 ± 51 Ma	Sm-Nd	Balashov et al. (1993)
Fedorovo-Pansky	Gabbro-norite	2491 ± 1.5 Ma	Zr	Bayanova et al. (1994)
Fedorovo-Pansky	Gabbro-norite	2491 ± 1.5 Ma	Zr	Bayanova & Balashov (1995)
Fedorovo-Pansky	Gabbro-norite	2491 ± 1.5 Ma	Zr	Mitrofanov & Bayanova (1999)
Fedorovo-Pansky	Gabbro pegmatite	2470 ± 9 Ma	Zr	Mitrofanov & Bayanova (1999)
Fedorovo-Pansky	Anorthosite	2447 ± 12 Ma		Mitrofanov & Bayanova (1999)
Mt. Generalskaya	Gabbro-norite	2505.1 ± 1.6 Ma	Zr	Amelin et al. (1995)
Mt. Generalskaya	Gabbro-norite	2496 ± 10 Ma	Zr	Bayanova et al. (1999a)
Mt. Generalskaya	Anorthosite	2447 ± 10 Ma	Zr	Bayanova et al. (1999a)
Mt. Generalskaya	Anorthosite	2446 ± 10 Ma	Zr	Bayanova et al. (1999b)
Mt. Generalskaya	Anorthosite	2585 ± 10 Ma	Zr (xen.)	Bayanova et al. (1999b)
Mt. Generalskaya	Gabbro-norite	2453 ± 42 Ma	Sm-Nd	Balashov et al. (1993)
Mt. Generalskaya	Gabbro-norite	2496 ± 10 Ma	Zr	Mitrofanov & Bayanova (1999)
Mt. Generalskaya	Anorthosite	2447 ± 10 Ma	Zr	Mitrofanov & Bayanova (1999)
Lake Pääjärvi	Gabbro-norite dike	2446 ± 5 Ma	Bd	Vuollo et al. (1995)
Mafic intrusions in northern Finland				
Koillismaa	Various rocks	2436 ± 5 Ma	Zr	Alapieti (1982)
Penikat	Various rocks	2410 ± 64 Ma	Sm-Nd	Huhma et al. (1990)
Kemi	Pegmatitic gabbro	2432 ± 6 Ma	Zr	Perttunen & Vaasjoki (2001, *this volume*)
Koitelainen	Gabbro	2439 ± 3 Ma	Zr	Mutanen & Huhma (2001, *this volume*)
Koitelainen	Monzodiorite	2434 ± 5 Ma	Zr	Mutanen & Huhma (2001, *this volume*)
Koitelainen	Granophyre	2432.5 ± 6 Ma	Zr	Mutanen & Huhma (2001, *this volume*)
Akanvaara	Gabbro	2436 ± 6 Ma	Zr	Mutanen & Huhma (2001, *this volume*)
Onkamonlehto, Salla	Mafic dike	2383 ± 33 Ma	Zr	Manninen & Huhma (2001, *this volume*)
Volcanic and subvolcanic rocks in Russia				
Seidorechka	Felsic volcanics	2434 ± 15 Ma	Zr+Bd	Bayanova & Balashov (1995)
Imandra	Felsic subvolcanic rock	2442.2 ± 1.7 Ma	Bd	Amelin et al. (1995)
Paanajärvi	Quartz porphyry	2432 ± 22 Ma	Zr	Buiko et al. (1995)
Lekhta	Quartz porphyry	2442.8 ± 4.8 Ma	Zr	Levchenkov et al. (1994)
Vetreny Belt	Dacite	2437 ± 3 Ma	Zr	Puchtel et al. (1997)
Vetreny Belt	Komatiitic volcanite	2410 ± 34 Ma	Sm-Nd	Puchtel et al. (1997)
Vetreny Belt	Komatiitic volcanite	2449 ± 35 Ma	Sm-Nd	Puchtel et al. (1997)
Volcanic and subvolcanic rocks in northern Finland				
Rookkiaapa	Felsic tuff	2438 ± 14 Ma	Zr	Manninen et al. (2001, *this volume*)
Sakiamaa	Dacite	2438 ± 11 Ma	Zr	Räsänen & Huhma (2001, *this volume*)
Kuusamo	Porphyry clast in congl.	2405 ± 6 Ma	Zr	Silvennoinen (1991)

Pansky intrusion. The time interval indicated by these ages exceeds by far estimated crystallization times of mafic layered intrusions (Cawthorn & Walraven 1998). The ages were interpreted by Mitrofanov and Bayanova (1999) as indicating a long duration of the plumbing systems, which produced the composite layered intrusions, feeding from a long-lived mantle plume through repeated magma pulses. However, it is very unlikely that plume-related magmas find the same channels and are injected into the same chambers in the continental crust within a time period of 50±20 Ma as suggested by Mitrofanov and Bayanova (1999) and Bayanova and Mitrofanov (1999), especially when the products of the earlier magmatic pulses would likely have been completely crystallized and cooled down before the injection of the suggested later pulse(s). A further complicating factor is the additional discovery of still older zircons (2585±10 Ma) in some layered intrusions. These zircons have been interpreted as xenocrystic (Bayanova et al. 1999a,b).

2.2 Ga differentiated mafic sills (GWA)

As shown in Figure 1, there is a large time gap in the age determinations from northern Finland after the cessation of the magmatism at c. 2.43 Ga and before the next peak occurring at c. 2.2 Ga. However, this does not mean a period of complete absence of igneous activity in northern Finland. For example, on geological grounds we can deduce that the mafic lavas of the Runkaus Formation in the Peräpohja Schist Belt are younger than the 2.43 Ga layered intrusions but older than 2.2 Ga which is the minimum age for the overlying quartzites (Perttunen & Vaasjoki 2001, *this volume*). In fact, secondary titanite from a Runkaus Formation metalava has provided a U-Pb age of c. 2.25 Ga (Huhma et al. 1990). In addition, in the Archean Pudasjärvi and Kuhmo basement complexes, concrete geochronological evidence demonstrates the presence of mafic dykes with an age of c. 2.33 Ga (Vuollo et al. 2000), but these rocks are beyond the scope of this review.

Perttunen and Vaasjoki (2001, *this volume*), Rastas et al. (2001, *this volume*) and Räsänen and Huhma (2001, *this volume*) report numerous ages of c. 2.2 Ga for ultramafic to mafic, differentiated sills occurring within Kivalo Group quartzites in the Peräpohja Schist Belt and Sodankylä Group quartzites in Central Lapland. These intrusions are characterized by a rock sequence varying, from bottom to top, from amphibole-bearing peridotites through clinopyroxenites to magnetite-rich gabbros. Due to a low Al_2O_3 content of the parental magma of these intrusions, plagioclase appears relatively late as a cumulus mineral allowing ultramafic cumulates rich in olivine and clinopyroxene to form at the bottom of the sills. In the literature, these rocks have traditionally been included in a rock class loosely termed as "albite diabases ", while more recently they were classified as the "Haaskalehto-type intrusions " in Central Lapland (Räsänen et al. 1995, Lehtonen et al. 1998). Hanski (1982, 1986, 1987) pointed out the wide-spread occurrence of differentiated sills of distinct chemical and mineralogical characteristics in eastern and northern Finland and assigned them to the "gabbro-wehrlite association" (GWA) according to the major rock types found in these sills. The same rocks have also been termed "karjalites" in eastern Finland (Vuollo & Piirainen 1992).

It is now well-established that the sills of the GWA form a magmatic phase which has a restricted age distinguishable from other magmatic events in the Fennoscandian Shield, but possesses a large regional extension, being found in Karelian schist belts (Northern Karelia, Kainuu, Kuusamo, Peräpohja, Central Lapland) in eastern and northern Finland. They were intruded concordantly into sediments which have traditionally been regarded as Jatulian or occur immediately beneath them in the Archean basement. In addition, differentiated bodies of this magmatic event have been encountered in the Archean Kuhmo greenstone belt forming an enigmatic occurrence as they are not met with in the surrounding gneiss complex. In general, products of this magmatic event are absent or very scarce in the Archean granitoid basement with the exception of the immediate contact zone to the overlying Paleoproterozoic schist belts (Vuollo 1994). The highest stratigraphical level where rocks of the GWA intrusions have been found is the Sodankylä and Kivalo Groups and other units traditionally assigned to the Jatulian and therefore these intrusions provide a minimum age for these supracrustal rocks. It is noteworthy that they have never been encountered within the Savukoski Group or stratigraphic units above it in Central Lapland or within the Paakkola Group in the Peräpohja Schist Belt. Neither have they been observed to intrude the Kalevian metasediments of the Karelian schist belts in eastern Finland. It is also important to note that so far no genetic counterparts of this magmatic phase have been recognized among volcanic rocks.

The GWA intrusions have become one of the most frequently dated magmatic events in the Finnish Precambrian with the ages converging at 2.2. Ga (Fig.

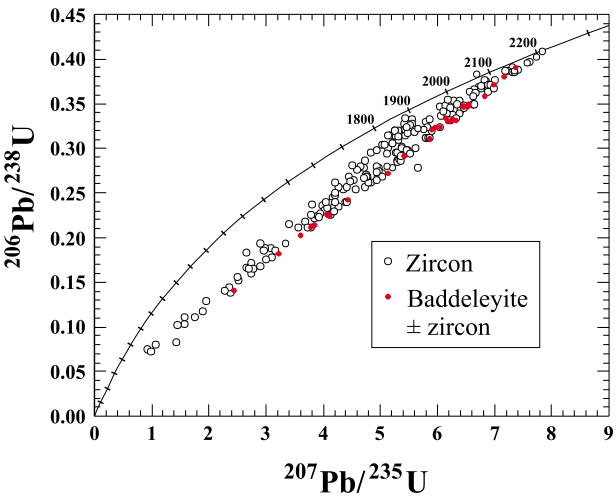

Fig. 2. U-Pb concordia diagram for zircons and baddeleyites from 2.2 Ga GWA intrusions (data from *this volume*; Tyrväinen 1983, Hyppönen 1983, Paavola 1984, Silvennoinen 1991, Lauerma 1995, unpubl. analyses of GSF).

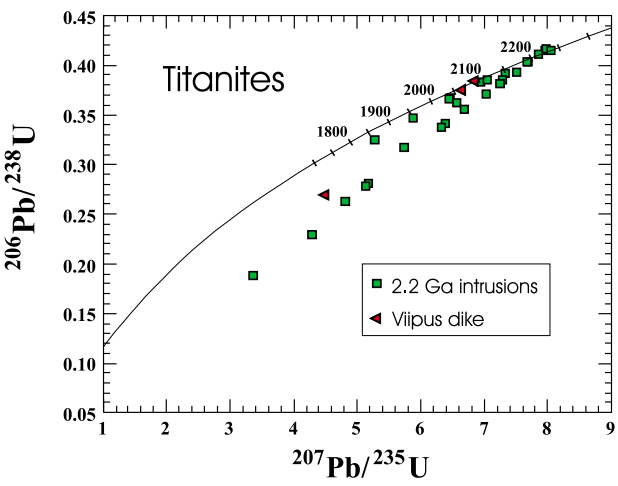

Fig. 3. U-Pb concordia diagram for titanites separated from 2.2 Ga gabbros (data from Rastas et al. 2001, *this volume,* Perttunen & Vaasjoki 2001, *this volume*, Silvennoinen 1991, Lauerma 1995). Also shown are two titanite fractions from the Viipus dyke, Kuusamo (data from Silvennoinen 1991).

1). Characteristic features of zircons from the GWA intrusions are their turbid color and porous texture and a high degree of metamictization resulting usually in a high degree of variation in the discordance on concordia diagrams. Among the most well-preserved intrusions having nearly concordant zircons is the Haaskalehto intrusion in the Sodankylä parish, Finnish Lapland which has yielded an age of 2220±11 Ma (Tyrväinen 1983). Good linear arrays on concordia diagrams are formed by zircons from some other Lappish intrusions, such as those at Eksymäselkä and Isolaki, defining ages of 2214±4 Ma and 2210±4 Ma, respectively (Rastas et al. 2001, *this volume*). Lauerma (1995) obtained an age of 2210±4 Ma for a concordant zircon fraction separated from metadiabase (Jokinenä A379) occurring in the Salla schist belt. Among the age determinations published by Perttunen and Vaasjoki (2001, *this volume*), the result of 2221±5 Ma for the Laurila sill, based on zircon and baddeleyite analyses, deserves mentioning. We regard the above listed ages as reliable estimates of the timing of the emplacement of the GWA intrusions. The age determination on zircons from Isolaki, made already in the late 1970s, brought about confusion in the stratigraphical correlation of the associated sedimentary rocks of the Värttiövaara Formation which are now regarded as younger that the mafic body and part of the Kumpu Group (see Rastas 1980, Silvennoinen et al. 1980, Lehtonen et al. 1998, Hanski 2001, *this volume*).

In addition to the conventional U-Pb zircon technique, the 2.2 Ga magmatism has been studied using single zircons and NORDSIM. Evins and Laajoki (in press) obtained an age of 2216.2±3.8 Ma for the Tokkalehto gabbro in the western part of the Kuusamo schist belt. Zircon grains analyzed by these authors were almost without exception nearly concordant, which is rarely the case when the 2.2 Ga intrusions are concerned, as discussed below.

The 2.2 Ga. intrusions often contain disturbed zircon fractions displaying a high degree of discordance and deviating from a single chord, and thus it has been unavoidable in several cases to exclude some of the zircon fractions from age calculation (e.g. sample A304-Kallinkangas in Figure 6 in Perttunen & Vaasjoki 2001, *this volume*). Therefore, it is useful to consider the total spread of analytical points on a concordia diagram, produced by those intrusions which, based on geological, petrological and isotope data, are confidently known to belong to the GWA and thus are c. 2.2 Ga old. In Figure 2, isotopic analyses are shown for zircons separated from GWA intrusions occurring in eastern and northern Finland. Huge variation in the $^{207}Pb/^{235}U$ and $^{206}Pb/^{238}U$ ratios generates a very long array with the lower intercept of the whole trend projecting to c. 300 Ma. The lower boundary of the "cloud" is rather sharp but there is a significant vertical scatter, especially with zircons having $^{207}Pb/^{235}U$ between 4.0 and 6.5. Also plotted in Figure 2 are analyses for baddeleyites and zircon-bearing baddeleyite fractions as red dots. The baddeleyites also display variable discordance, but in contrast to the zircons, form a reasonably good linear array on the lower edge of the zircon field. The scatter of the titanite trend is smaller than that of the zircons and many titanites are nearly concordant, but with a wider scatter of ages than the least discordant zircons (Fig. 3). The scatter observed in the zircon U-Pb data suggests multiphase histories: magmatic crystalliza-

tion 2.2 Ga ago, followed by disturbance of the U-Pb system at 1.8-1.9 Ga, which may have involved dissolution and recrystallization of metamict zircons, and finally more recent loss of lead either through continuous diffusion or by an episodic event.

Given the wide-spread regional distribution of the 2.2 Ga dykes and sills in Finland, it is surprising that this age group has not been reported to occur in the Russian part of the Fennoscandian Shield. There are no geological reasons which would suggest their absence within the Jatulian metasediments in Russian Karelia or on the Kola Peninsula. Neither have they been reported from Precambrian terranes in northern Norway. Only one such GWA-type diabase has been dated from northern Sweden. It is from north of Kiruna and contains complex zircon populations due to later metamorphic crystallizations, yielding a minimum age of c. 2.18 Ga (assumed to fall between 2.20 to 2.21 Ga in Fig. 1) for the dyke intrusion (Skiöld 1986).

2.14 Ga mafic intrusions

Mafic intrusions with an age approaching 2.14 Ga have so far been recognized mainly in a relatively restricted area in Finnish Central Lapland. Räsänen and Huhma (2001, *this volume*) published data on three intrusions occurring within Sodankylä Group quartzites, i.e. Rantavaara (2148±11 Ma), Povivaara (2142±5 Ma) and Koskenkangas (c. 2.14 Ga). The Rantavaara intrusion has been regarded as the type occurrence of this age group. It occurs as a conformable sill with a thickness of c. 0.5 km and a length of more than 20 km. Locally it has an olivine-bearing ultramafic base but the bulk of the body is composed of metagabbroic rock. Tyrväinen (1983) published an age of 2132±30 Ma for the Pittiövaara dyke, which is only 10 m thick and called an "albite diabase" by Tyrväinen. As zircons are turbid and the data discordant and heterogeneous, the reported upper intercept age should be considered very suspect and this rock may well belong to the 2.2 Ga age group (see below).

The majority of the zircon dates for mafic rocks from the Peräpohja Schist Belt are c. 2.2 Ga or older but there are some exceptions. Perttunen and Vaasjoki (2001, *this volume*) published results of three samples from the Misi gabbro (A694, A77, A78) located in the Misi area in the NE corner of the Peräpohja Schist Belt. Taken together these samples yield an age of 2131±12 Ma and thus may warrant classification of this occurrence together with the 2.14-2.13 Ga intrusions in Cental Lapland.

The second, more problematic case is provided by sample A480 (Sipojuntti) taken from a dyke cutting the Archean basement complex. This sample contains a heterogeneous zircon population. The low-density material yields an age of 2111±6 Ma, whereas analyses on the heavy zircon are more discordant and plot on the "younger side" of the chord (Perttunen & Vaasjoki 2001, *this volume*). The third example is a 50-m thick diabase dyke intersecting a differentiated sill of the 2.2 Ga age group at Luppovaara (A593). Unfortunately only two zircon fractions with a minor amount of baddeleyite were available to Perttunen and Vaasjoki (2001, *this volume*), both yielding $^{207}Pb/^{206}Pb$ ages at around 2.1 Ga.

All the c. 2.14 Ga rocks described above are gabbros or diabases. One exceptional representative of this age group is the Iso Nilipää porphyritic granite from the northern edge of Central Lapland Granite Complex, which has provided an age of 2136±5 Ga (Huhma 1986, Rastas et al. 2001, *this volume*). The geological significance of this age is unknown.

C. 2.05 Ga intrusions and dykes

This age group of mafic dykes and intrusions is incoherent as shown by Figure 1. Nevertheless, at least some of them are found to reach higher stratigraphic levels (Savukoski Group) than the older intrusions discussed above. The oldest apparent age for the rocks dealt with in this chapter is 2078±8 Ma which was reported by Silvennoinen (1991) for a mafic dyke (Viipus) in the Kuusamo schist belt. Before moving further to other ages approaching 2.05 Ga, we discuss a problem encountered with zircon geochronology of mafic rocks, which can be well illustrated using the Viipus dyke as an example.

When only discordant analytical results for zircon from a mafic intrusive body are available, a large error in the age estimate is inevitable and even assignment of the dated rock to a wrong age group is possible, as discussed in this volume. Given the wide scatter on concordia diagrams of the analytical points for zircons from 2.2 Ga intrusions (see Fig. 2), a question arises if some of the dyke rocks, giving superficially a younger age, are in fact members of the 2.2 Ga group containing disturbed zircons. In Figure 4, shown as red circles

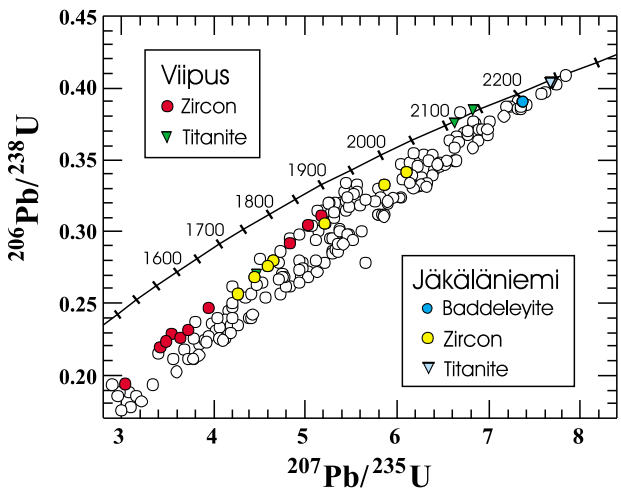

Fig. 4. U-Pb concordia diagram comparing analytical results for different minerals from from the Viipus dyke and the 2.2 Ga Jäkäläniemi intrusion with zircon data from other 2.2 Ga intrusions. Data for Viipus and Jäkäläniemi from Silvennoinen (1991), other reference data same as in Figure 2.

and green triangles are zircons and titanites, respectively, from two diabase samples, A412 and A487, corresponding to locations Viipus I and II in the Kuusamo area. The age of 2078±8 Ma given by Silvennoinen (1991) for sample A412 was based on two transparent but discordant zircon fractions and three titanite fractions of which two are concordant.

It is obvious from Figure 4 that the Viipus zircons fall within the field defined by the zircons from 2.2 Ga GWA intrusions. Silvennoinen (1991) also published isotopic data for zircon, baddeleyite and titanite fractions from sample A847 representing a diabase at Jäkäläniemi. Regressed together, transparent zircons, baddeleyite and titanite yield an age of 2206±9 Ma. This date together with the near-concordance of baddeleyite demonstrate that this diabase belongs to the 2.2 Ga GWA. However, as revealed by Figure 4, several of the turbid zircons from the Jäkäläniemi dyke plot close to the trend formed by the zircons from the Viipus dyke and hence, zircons cannot be utilized to distinguish between the two age groups of mafic rocks in this particular case. The supposedly younger age of the Viipus dyke relies heavily on the two titanite analyses with $^{207}Pb/^{207}Pb$ ages of 2076±30 and 2080±23 Ma. As shown in Figure 3, among the titanites analysed from 2.2 Ga intrusions, there exist near-concordant ones with $^{207}Pb/^{206}Pb$ ages significantly less that the magmatic age of their host rocks. The large range of ages shows that even concordant titanites are not reliable indicators of magmatic ages. Petrographic observations attest that titanites in mafic dykes are hardly primary magmatic phases but were formed by later autometasomatic or metamorphic processes. Furthermore, in the Kuusamo area, concordant titanite (A414) extracted from a mafic lava belonging to the Greenstone Formation III yields an age of 1809±18 Ma (Silvennoinen 1991) which is evidently too young for an eruption age of this lava. It seems that, without the help of more concordant zircons or baddeleyites, the age of the younger dyke group remains somewhat speculative. It should also be noted here that the age of c. 2.08 Ga reported for the Viipus dyke is uncommon among the U-Pb ages obtained for mafic dykes and intrusions in northern Finland (see Fig. 1). It is also noteworthy that Silvennoinen (1991) published analyses of three zircon fractions of the Matosuo diabase and commented that two of them plot close to the Viipus dyke chord and one on a 2.2 Ga chord, thus further stressing the uncertainty in dividing the Kuusamo dykes into different age groups. The phenomenon discussed above may also concern some of the 2.13-2.14 Ga intrusions dealt with in the previous chapter, particularly the Pittiövaara dyke. All zircon fractions from this intrusion are moderately to highly discordant and plot totally within the field of zircons from the GWA.

Returning to the zircon ages of c. 2.05 Ga which have been obtained for several mafic dykes, one layered intrusion and two felsic porphyries, we note that the Keivitsa mafic layered intrusion, in Central Lapland and hosting a large, low-grade Ni-Cu-PGE deposit, was dated by Mutanen and Huhma (2001, *this volume*) using the U-Pb zircon method. A concordant zircon from an olivine pyroxenite gave an age of 2058±4 Ma. A diorite dyke crosscutting the mafic body revealed a very similar age, 2054±5 Ma. These ages are also compatible with the Sm-Nd and Re-Os data on the intrusion (Hanski et al. 1997).

Several mafic dykes occurring in Central Lapland have been dated at c. 2.05 Ga. Räsänen and Huhma (2001, *this volume)* give the following results: Kylälampi 2066±7 Ma, Rovasvaara 2055±5 Ma, Törmänen 2054±14 Ma, Mairivaara 2046±18 Ma and Ahvenselkä 2070±5 Ma, while Rastas et al. (2001, *this volume*) report the following: Sätkänävaara 2046±9 Ma, Riikonkoski 2060±8 Ma and Heinälammintievat 2052±7 Ma. All of them are relatively narrow dykes reaching usually only some tens of meters in thickness and displaying no significant magmatic differentiation. They occur within metavolcanics and metasediments corresponding to the stratigraphic level of the Matarakoski Formation, the type formation of the lower part of the Savukoski Group. According to Räsänen and Huhma (2001, *this volume*), equivalent dykes have not been met with in the Sattasvaara or Sotkaselkä Formations higher up in the Savukoski Group.

Mutanen and Huhma (2001, *this volume*) have

studied a quartz porphyry at Matarakoski (A566), 10 km southwest from the Keivitsa intrusion, which gives a U-Pb age of 2048±5 Ma. The authors regard the porphyry as an extrusive rock, in which case its age would also date the time of deposition of the associated pelitic metasediments. Lehtonen et al. (1998) in turn considered the Matarakoski porphyry a hypabyssal intrusion. In any case, the age of the porphyry and all the dates for mafic dykes mentioned above attest that the minimum age for the pelitic metasediments of the Savukoski Group is 2.05 Ga which is in accord with the Sm-Nd age of 2056±25 Ma obtained for the overlying komatiites of the same group (Hanski et al. 2001).

It is noteworthy that this age group is lacking among the dated mafic rocks in the Peräpohja Schist Belt as all the concordant mafic bodies within the Kivalo quartzites have turned out to belong to the 2.2 Ga age group. However, according to geological and geophysical observations, there exist probably younger discordant mafic dykes in the belt, but so far no geochronological data are available for these rocks. Anyway, there is some evidence for the presence of magmatism of this age group in the form of felsic dykes. A felsic porphyritic rock (Keinokangas) lying within mafic volcanic rocks of the Väystäjä Formation was dated by Perttunen and Vaasjoki (2001, *this volume*) at 2051±7 Ma.

C. 2.0 Ga mafic and felsic dyke rocks

An age determination connecting the c. 2.0 and 2.05 Ga age groups in Figure 1 is the result of 2027±19 Ma for a mafic dyke in the Kolari area (Hiltunen 1982, Väänänen et al. 2001, *this volume*). Due to the large error in the age, this case could have been considered in the previous chapter as well as here. The dyke intrudes concordantly mafic metavolcanics of the Kolari Formation which is correlated with the Matarakoski Formation of the Savukoski Group.

Three other gabbroic dykes record ages of c. 2.0 Ga in the western part of the Central Lapland Greenstone Belt. Two of them occur within tuffites of the Savukoski Group and yield ages of 2007±8 Ma (Home-Jolhikko) and 1995±8 Ma (Pittarova), while the Kulkujärvi dyke cutting metavolcanics of the Kittilä Group gives an age of 1986±2 Ma.

Because of the lack of reliable direct or indirect age determinations of the voluminous mafic metavolcanics of the Kittilä Group in Central Lapland, its correlation with Karelian stratigraphic units has long been contentious (see Hanski 2001, *this volume*). A solution to this problem was provided by felsic porphyries occurring as dykes and crystal tuffs and forming a minor, but significant component of the bimodal magmatism in the Kittilä greenstone complex. Though the zircon ages of these rocks have been known for many years, their true significance as one of the results of the studies of the Lapland Volcanite Project was appreciated only quite recently (Lehtonen et al. 1998). Detailed field investigations including drilling and trenching showed that these felsic rocks and associated mafic lavas and dykes exhibit undisputable magma mingling and mixing structures and are thus coeval. Hence the U-Pb zircon dates of the felsic dykes also indicate the age of the mafic magmatism of the Kittilä Group. The following coherent zircon ages have been obtained for felsic porphyries at four localities north of Kittilä (Rastas et al. 2001, *this volume*): Kapsajoki 2017±8, Veikasenmaa 2014±8, Yräjärvi 2015±5 and Kiimarova 2012±3 Ma. These results are consistent with a Sm-Nd isochron age of 1990±35 Ma determined for mafic metavolcanics of the Vesmajärvi Formation of the Kittilä Group (Hanski et al. 1998). The initial ε_{Nd}-value is +3.7±0.2, and samples from other localities exhibit only a small variation of ε_{Nd}(2 Ga) from +3.3 to +4.4, which is consistent with a depleted mantle source.

The relatively young age of 2.0 Ga obtained for the Kittilä Group along with its lithological and geochemical characteristics partly explains why, over the years, it has been so difficult to find suitable counterparts for these rocks among the traditional Karelian formations. In fact, assuming that the interpretation of its allochthonous, ophiolitic nature is correct (Hanski et al. 1998), it is debatable whether the Kittilä Group can be included in the Karelian formations at all, as the latter are generally thought to involve rocks deposited or extruded on the Archean basement complex (e.g., Laajoki 1986).

1.91-1.92 Ga felsic plutons and dykes

Rastas et al. (2001, *this volume*) give an age of 1919±8 Ma for a felsic dyke crosscutting mafic metavolcanics of the Kittilä Group at Nyssäkoski. An identical rock type with a similar age (1928±6 Ma, see

below) has been discovered as abundant pebbles in the Mantovaara conglomerate (Kumpu Group) which is indicative of a wide-spread occurrence of this kind of dykes within the underlying Kittilä Group metavolcanics. A comparable age of 1914±3 Ma has also been obtained for a larger body of granodiorite occurring at Ruoppapalo with Kittilä Group metavolcanics as its country rocks (*ibid.*). These ages confirm that the emplacement of the Kittilä Group rocks took place at or before c. 1.92. Hanski et al. (1998) suggested that the genesis of the dykes and intrusions of this age group was genetically related to the tectonic emplacement of a slab of former oceanic lithosphere, now represented by the Kittilä Group. This interpretation is consistent with the fact that dykes and intrusions of this age have so far not been found outside of the Kittilä Group rocks in Central Lapland.

C. 1.88 Ga synorogenic plutonism and volcanism

Väänänen and Lehtonen (2001, *this volume*) summarize age determinations made for rocks occurring in the Kolari-Muonio area, western Finnish Lapland. Among the dated rocks is the Kallo massif, a characteristic representative of synorogenic plutonic rocks belonging to the Haaparanta Suite (called also the Haparanda Suite). The age obtained by Väänänen and Lehtonen (2001, *this volume*) for this tonalitic to monzonitic massif is 1881±1 Ma, while a gabbroic body at Jalokoski yielded an age of 1866±6 Ma. Lehtonen (1984) had earlier published zircon analyses from several plutonic rocks in the Muonio area, but many of the samples revealed a heterogeneous zircon population containing old zircons apparently inherited from the Archean basement. Nevertheless, two quartz monzonite occurrences produced reasonably good arrays on concordia diagrams allowing these intrusions to be dated at 1876±16 Ma and 1892±6 Ma. Hiltunen (1982) also published geochronological data from western Lapland (Kolari) including ages of 1849±16 Ma and 1862±3 Ma for two monzonites.

Very similar ages are reported by Perttunen and Vaasjoki (2001, *this volume*) for plutons cutting supracrustal rocks in the Peräpohja Schist Belt: 1887±9 Ma, 1882±3 Ma and 1880±2 Ma for an olivine gabbro (Liakka), dioritic portion of a mafic stock (Kuttijänkä) and quartz monzonite (Kaakamo), respectively. Also noteworthy is that zircons measured from more felsic Haaparanta Suite intrusions and supposedly cogenetic dykes are generally xenocrystic with Archean ages (ibid.).

Previously plutonic rocks of the Haaparanta Suite in Finland have been described in a north-south trending zone occurring close to the Finnish-Swedish border, as also indicated by the dated examples listed above. Räsänen and Huhma (2001, *this volume*) report a granodiorite occurrence at Kelujärvi (A65) with an age of 1891±5 Ma, situated about 150 km west of the border and thus expanding the aerial distribution of this magmatic phase significantly to the east.

Coeval calc-alkaline volcanic rocks occur at Latvajärvi, c. 15 km west of village Kittilä, for which Rastas et al. (2001, *this volume*) report an age of 1880±8 Ma. In northern Sweden, there exist a large, c. 1.9 Ga (Svecofennian) volcanic terrain, extending from the Skellefte to Kiruna area, which contains subaerial felsic to mafic lavas. Perdahl and Frietsch (1993) assigned this felsic-dominated lava suite to the Kiruna-Arvidsjaur Porphyry Group. The Latvajärvi Formation in the Kittilä area can be regarded as the only equivalent of the Kiruna-Arvidsjaur Porphyry Group so far recognized in northern Finland.

Most of the ages obtained for intrusions belonging to the age group under consideration fall within the range of 1.87 to 1.89 Ga which is generally considered to represent a peak of the Svecofennian orogenic magmatism (e.g., Skiöld et al. 1993). In addition to age determinations from northern Finland, Figure 1 exhibits U-Pb zircon ages published for Paleoproterozoic felsic igneous rocks occurring in northern Sweden, north of the Skellefte area, highlighting the strong magmatic activity within the above mentioned time period. The diagram indicates not only the amount of geochronological work done on the c. 1.88 Ga rocks by Swedish workers but also their wide-spread occurrence in northern Sweden. This age group encompasses the Haparanda and Perthite Suite rocks and such plutons as the Jörn complex and Degerberg massif in the Norrbotten County, northern Sweden (e.g., Wilson et al. 1987, Skiöld 1988, Wikström et al. 1996, Wikström & Persson 1997, Martinsson et al. 1999).

The youngest members of the orogenic magmatism in northern Sweden have been reported to be c. 1.86 Ga old (Skiöld 1988, Wikström & Persson 1997, Persson & Lundqvist 1997). After the termination of this later phase of Svecofennian orogenic magmatism, a quiescence of igneous activity lasting c. 60 Ma ensued (see Fig. 1), which led Skiöld and Öhlander (1989) to suggest that the next significant magmatic stage at c. 1.80 Ga should not be considered as part of the Svecofennian orogeny. Age determinations from

the Finnish side of the border, however, contain some examples which seem to fall in this gap. These are the Molkoköngäs granite and Rautuvaara monzonite for which Huhma (1986) and Hiltunen (1982) obtained ages of 1843±23 Ma and 1849±16 Ma, respectively. If the errors are taken into account, these figures overlap with the lower limit of 1.86 Ga for the orogenic magmatism. Based on Sm-Nd analysis, the granite at Molkoköngäs contains a major contribution from Archean crust (Huhma 1986), and a minor xenocrystic zircon component would easily result in an apparent upper intercept age too old for the granite emplacement. This is compatible with the slight heterogeneity encountered in the zircon U-Pb data. Whether there exist any granitoids with ages between 1.82-1.85 Ga in northern Finland has some geotectonic significance as this "late orogenic" age group has been recognized in southern Finland and south-central Sweden (Suominen 1991, Nironen 1997). The available geochronological data suggest their absence or scarcity in the north.

C. 1.8 Ga granitoids

The youngest rocks dealt with in this review are potassic granitoids which, as shown by Figure 1, seem to form two subgroups in northern Finland with ages of c. 1.80 Ga and 1.77 Ga. The Tepasto granite occurring c. 40 km north of Kittilä belongs to the *Nattanen-type granites* as originally defined by Mikkola (1928, 1941). It forms the southwestern end of a chain of post-orogenic, generally undeformed granites of this type, extending to NE and including the granites of Nattastunturit, Riestovaara, Salmurinvaara and Pomovaara in the Sodankylä parish, the Vainospää granite NE of Lake Inari and Juvoaivi and Litsa-Araguba on the Kola Peninsula (see Haapala et al. 1987). Rastas et al. (2001, *this volume*) report an age of 1802±10 Ma for the Tepasto granite (Kotivaara). Huhma (1986) published U-Pb zircon ages for two Nattanen-type granites: 1772±10 Ma for Nattastunturit and 1790±22 Ma for Vainospää. Chen et al. (1998) recently reported concordant U-Pb zircon ages averaging 1765±2 Ma for three porphyritic granites, which cut Archean rocks in the superdeep borehole in the Pechenga area (Kola Peninsula). These dates together with ages from northern Finland suggest that the emplacement of the Nattanen-type granites took place over a time period of c. 30 Ma.

Another, more heterogeneous granite type defined by Mikkola (1941) is the *Hetta-type* occurring on the northern and northwestern side of the Kittilä greenstone complex. These granites differ from the Nattanen-type in being more sheared or gneissose and Mikkola (1941) thought they had intruded contemporaneously with the Lapland granulites. Dating the Hetta-type granites has turned out to be difficult. Mänttäri (1995) published isotopic data for two granites, one from Kappera and the other from Peltovuoma. Both of them contained heterogeneous zircon, probably inherited populations. In contrast, an old sample from Peltovuoma collected by Meriläinen has provided an age of 1810±14 Ma (Huhma 1986), which is fairly close to the U-Pb age of 1770±22 Ma for titanite in the Kappera granite (Mänttäri 1995). On the other hand, the Ruoppapalo granodiorite with a U-Pb zircon age of 1914±3 Ma (Rastas et al. 2001, *this volume*) could be correlated with the Hetta-type granites (T. Manninen, *pers. comm. 2000*).

A microcline granite straddling the Finnish-Swedish border at Kihlanki and representing an easterly extention of the so-called Lina granites of northern Sweden, gives an age of 1805±15 Ma (Väänänen and Lehtonen 2001, *this volume*). Another example of this age group in the Muonio area is a pegmatite granite at Matalan-Nivunginselkä for which Lehtonen (1984) published an age of 1778±2 Ma.

The Kittilä granite, which intrudes the CLGB and is on geological grounds generally thought to belong to the c. 1.8 a age group, has not been amenable to zircon dating as the samples collected from it have proven to be very poor in zircon (Huhma 1986). In addition, the separated zircons have yielded discordant and scattered results. Rastas et al. (2001, *this volume*) attempted to solve the age problem using samples from three different locations, but the results indicated the presence of inherited zircons from the Archean basement and the precise emplacement age remains unknown.

The Central Lapland Granite Complex (also called the Kemijärvi Complex), comprising large amounts of granitoids between the Peräpohja Schist Belt and the Central Lapland Greenstone Belt, is so far poorly characterized in terms of lithology and geochronology. Some datings have been done mainly in its peripherial parts. In general, this complex is thought to have been generated c. 1.8 Ga ago with a significant contribution from Archean crustal sources (Huhma 1986). However, the geochronological situation seems to be more complicated. Earlier we already presented the results on the Molkoköngäs and Nilipää granites recording apparent ages higher than 1.8 Ga. In the Salla area,

granitic dykes have provided heterogeneous U-Pb zircon data, which do not allow dating with any confidence (Lauerma 1982). However, monazites in these granites yield U-Pb ages of c. 1.8 Ga, which can be considered to date the emplacement of the granites. A slightly younger age was published by Lauerma (1982) for a granite exposed at the Rovaniemi Airport close to the northern margin of the Peräpohja Schist Belt. The result, 1770±8 Ma, was based on one concordant titanite and two discordant zircon fractions. In addition to revealing the presence of Archean gneisses within the Kemijärvi Complex, recent studies carried out by the University of Oulu in the southeastern part of the complex have resulted in some age data on postkinematic granites. The undeformed Pernu granite has yielded an age of 1746±2 Ma (Evins et al. 1997). This is the youngest U-Pb zircon age determined for post-orogenic granites in northern Finland and outside the usual range of 1.80 to 1.77 Ga.

The last comment on the individual ages presented in the histograms of Figure 1 concerns an age documented by Hiltunen (1982) for a mafic pegmatoid intersecting skarn rock and skarn iron ore at Rautuvaara (Kolari area). This age, plotting at the younger end of the histogram, is 1748±7 Ma. The result is ambiguous because firstly, the mineralogy of the rock including plagioclase, hornblende and diopside possibly represents a metamorphic assemblage and secondly, manifestations of mafic magmatism of this age are unknown elsewhere in northern Finland.

The 1.80 Ga granites are known to penetrate rocks of the Kittilä and Savukoski Groups. No reliable field observations have been reported demonstrating that sizeable bodies of granite cut metasediments of the Kumpu or Lainio Groups. Pegmatitic and aplitic dykes crosscut quartzites at Aakenusvaara (Mäkelä 1968), but no geochronological data are available to confirm the genetic or coeval relationship of these dykes with the 1.80 Ga magmatism. As the datings of detrital minerals and granitoid pebbles from the Kumpu and Lainio Group metasediments definitely show, these rocks were deposited later than c. 1.88 Ga ago (see below). To constrain their minimum age, data on their field relationship with the post-orogenic granites would be important. Based on information on analogical molasse-like metasediments elsewhere, and particularly in northern Sweden (see below), it seems natural to presume that the 1.80 Ga granites postdate the Kumpu and Lainio Group rocks.

In northern Sweden, post-orogenic magmatism is represented by porphyritic and migmatite-associated granites divided into two main categories, the Lina- and Edefors-types, which range from 1.77 to 1.80 Ga in age with a dominance closer to the latter value (e.g., Öhlander & Skiöld 1994). The name Revsund-type has also been attached to some granites of this age group (Skiöld 1988). In addition, some syenitic intrusions and Perthite granites have yielded ages within the range of 1.80-1.77 Ga (Romer et al. 1994, Martinsson et al. 1999). The age determinations show that different kinds of granitoids were formed almost contemporaneously in northern Sweden. On the other hand, combining the above mentioned names with the Nattanen- and Hetta-types in northern Finland results in a jungle of nomenclature which needs to be clarified. More importantly, the genetic relationship between the above mentioned granitoid categories is presently not well-established.

Granitoid pebbles in conglomerates

To constrain the age of deposition of the uppermost sedimentary units in Central Lapland, pebbles have been collected for U-Pb dating from conglomerates. Felsic porphyry pebbles in a Kumpu Group conglomerate at Mantovaara have an age of 1928±6 Ma (Rastas et al. 2001, *this volume*), which is close to that recorded by the Nyssäkoski felsic dyke occurring within the underlying Kittilä Group metavolcanics (see above). A Haaparanta Series provenance for some of the conglomerates is corroborated by the U-Pb zircon age of 1888±22 Ma obtained for a granitoid pebble in the Kellostapuli conglomerate of the Lainio Group (*ibid.*). A similar age (1873±11 Ma) has been determined for felsic porphyry pebbles in the Vesikkovaara conglomerate (Lainio Group), suggesting that the porphyry pebbles were derived from the Latvajärvi Formation.

Detrital zircons in metasediments

A concordia diagram for analyses of bulk detrital zircons from various metasedimentary formations from northern Finland is presented in Figure 5. There are two major groupings on the diagram. The lower, longer trend contains zircons from the Kivalo and Sodankylä Group quartzites (Peräpohja Schist Belt

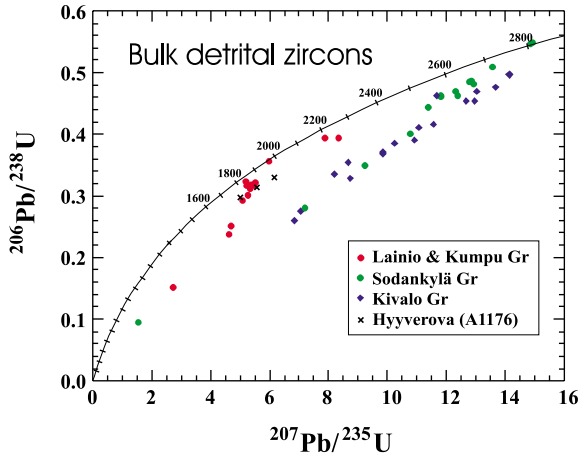

Fig. 5. U-Pb concordia diagram showing analyses of detrital zircons from the Kivalo and Sodankylä Groups and from the Lainio and Kumpu Groups. Also shown as crosses are detrital zircons from Hyyverova assigned to the Sodankylä Group (for sampling site, see sample A1176 in Figure 1 of Rastas et al. 2001, *this volume*).

and Central Lapland Greenstone Belt, respectively). The array indicates a dominance of Neoarchean gneisses and granites of the basement complex in their provenance. The other main group is formed by zircons from the Lainio and Kumpu Group quartzites located in Central Lapland. These analyses seem to be mixtures with a significant zircon component from obviously younger sources than the sources that furnished detrital minerals to the Sodankylä and Kivalo Group quartzites. Rastas et al. (2001, *this volume*) deduced that this relatively young component was derived from 1.88 Ga Haaparanta plutons and coeval volcanic rocks which is strongly supported by the ages of granitoid and porphyry pebbles (see above) and ion probe (NORDSIM) studies of detrital zircons (Hanski et al. 2000). The two main stratigraphical levels in Lapland composed of epiclastic metasediments are thus clearly distinguishable on the basis of conventional analyses of bulk detrital zircons. The results also demonstrate that the older one of these levels corresponds to traditional Jatulian quartzites as represented by the Kivalo Group. In addition, the available geochronological data cannot be used to show that the deposition of the Kumpu Group took place later than that of the Lainio Group. The Lainio and Kumpu Group rocks were separated by Lehtonen et al. (1998) on the basis of structural observations, according to which effects of the " main deformation stage ", seen in the Lainio Group metasediments, have not been found in the Kumpu Group rocks.

Figure 5 shows as crosses analyses of detrital zircons from quartzites occurring at Hyyverova (A1176), west of Kittilä. This is the only location of quartzites assigned to the Sodankylä Group which did not produce detrital zircons plotting in the general field of "Jatulian" quartzites in Figure 5. We do not know the exact reason for this exceptional phenomenon but some comments can be made. Firstly, the Hyyverova quartzites lie geographically very close to the Lainio Group quartzites as shown by figure 1 of Rastas et al. (2001, *this volume*). Secondly, the host metasediment of the analyzed zircons is a highly metamorphosed, sillimanite-bearing quartzitic gneiss. Thirdly, the zircons in sample A1176 are fine-grained and turbid with simple prismatic and pyramidal crystal faces devoid of effects of transportational abrasion (*ibid.*) and thus their morphology is not typical of detrital zircons and may indicate effects of metamorphic recrystallization. Finally, the quartzites in the area are cut by numerous granite pegmatite veins and dykes varying in thickness from a few centimeters to several meters.

A bulk zircon analysis from a turbiditic mica schist (A720, Napinmäki) of the Martimo Formation (Paakkola Group), occurring in the western part of the Peräpohja Schist Belt, yielded a Pb-Pb age of about 2575 Ma (Perttunen & Vaasjoki 2001, *this volume*). This result and data in Figure 5 show that detrital minerals from metasediments in the Peräpohja Schist Belt have mostly provided Archean ages, with one conspicuous exception. Perttunen and Vaasjoki (2001, *this volume*) obtained an age of 1975±7 Ma for a homogeneous, detrital zircon population separated from an arkosite (A1438) belonging to the Korkiavaara Formation, the sampling site being situated at Korkiavaara, 6 km south of Rovaniemi. Earlier Perttunen et al. (1996) had pointed out that these arkosites have a distinct chemistry in terms of elevated high-field-strength element (HFSE) contents which suggests an occurrence of A-type granites in their source area. This finding gave an impetus to apply the U-Pb method to detrital zircons from this rock as the provenance seemed to differ from typical TTG gneisses or granites occurring in the Archean basement. The obtained relatively young age for bulk zircon demonstrates that the zircons are really dominated by those derived from a non-Archean source. Furthermore, the deposition of the sediment must have taken place later than c. 1.98 Ga ago.

The Korkiavaara Formation consists of arkosites, amphibolites and calc-silicate- and carbonate-bearing arkosites. It lies stratigraphically in the upper part of the Kivalo Group and is overlain by graphite- and sulphide-bearing schists of the Pöyliövaara Formation belonging to the Paakkola Group (Perttunen et al. 1996), meaning that the deposition of the Paakkola Group, the former Kaleva Group of Perttunen (1985), would have taken place later than c. 1.98 Ga ago. This interpretation has a bearing on the general question of the age of Kalevian rocks. Geological evidence sug-

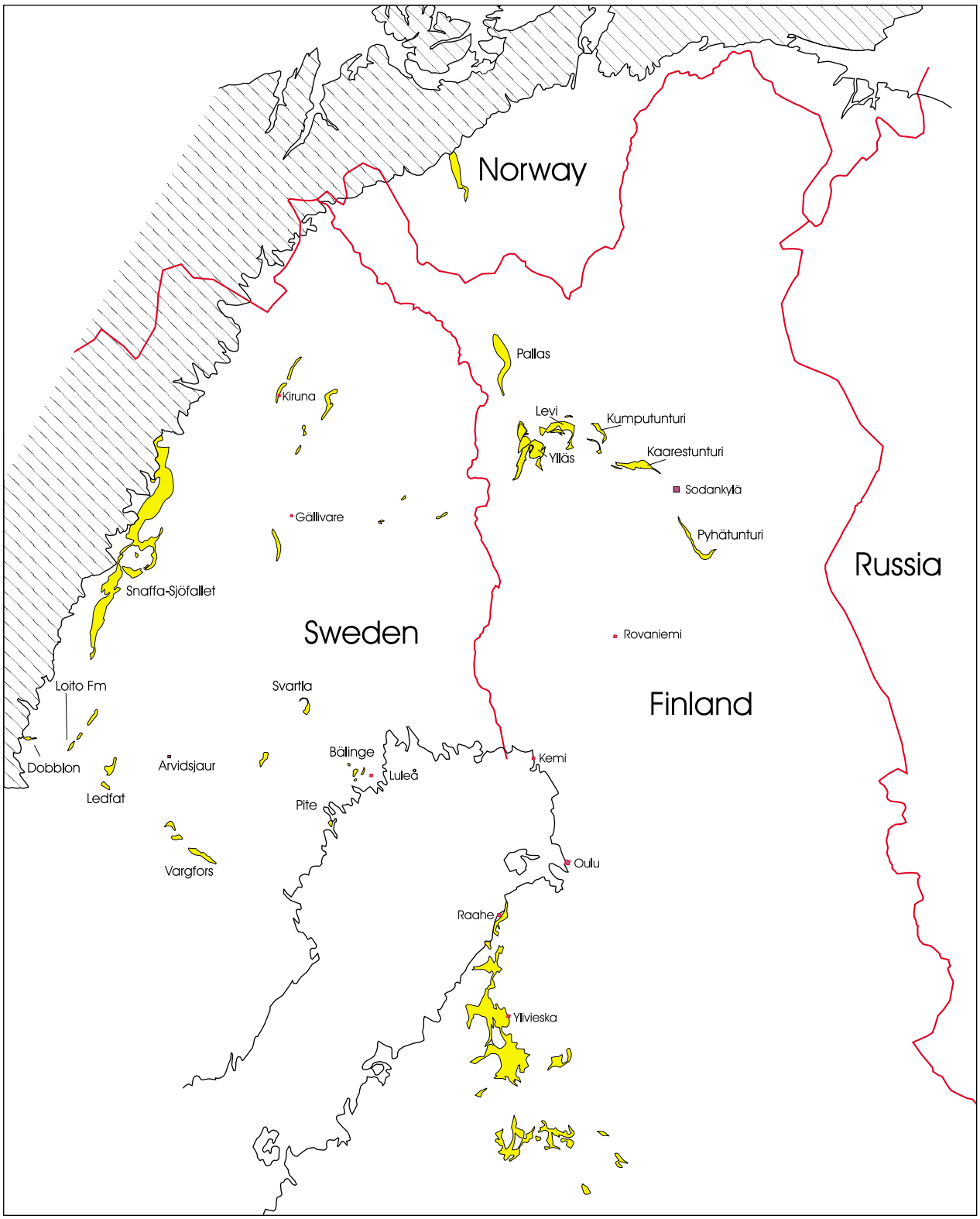

Fig. 6. Map showing potential sedimentary formations in northern part of the Fennoscandian Shield, deposited contemporaneously with the Lainio and Kumpu Groups.

gests that the Paakkola Group can be correlated with the "lower Kaleva" assemblage whose age has generally been bracketed in the interval of 2.1 to 1.95 Ga with the minimum age coming from the age of the Jormua ophiolite complex (Peltonen et al. 1996). The "upper Kaleva" was deposited on the ophiolitic rocks, being thus younger than 1.95 Ga. The youngest detrital zircons with an age of 1.92±0.04 Ga yield the maximum age limit for the upper Kalevian rocks (Claesson et al. 1993). The age obtained for the

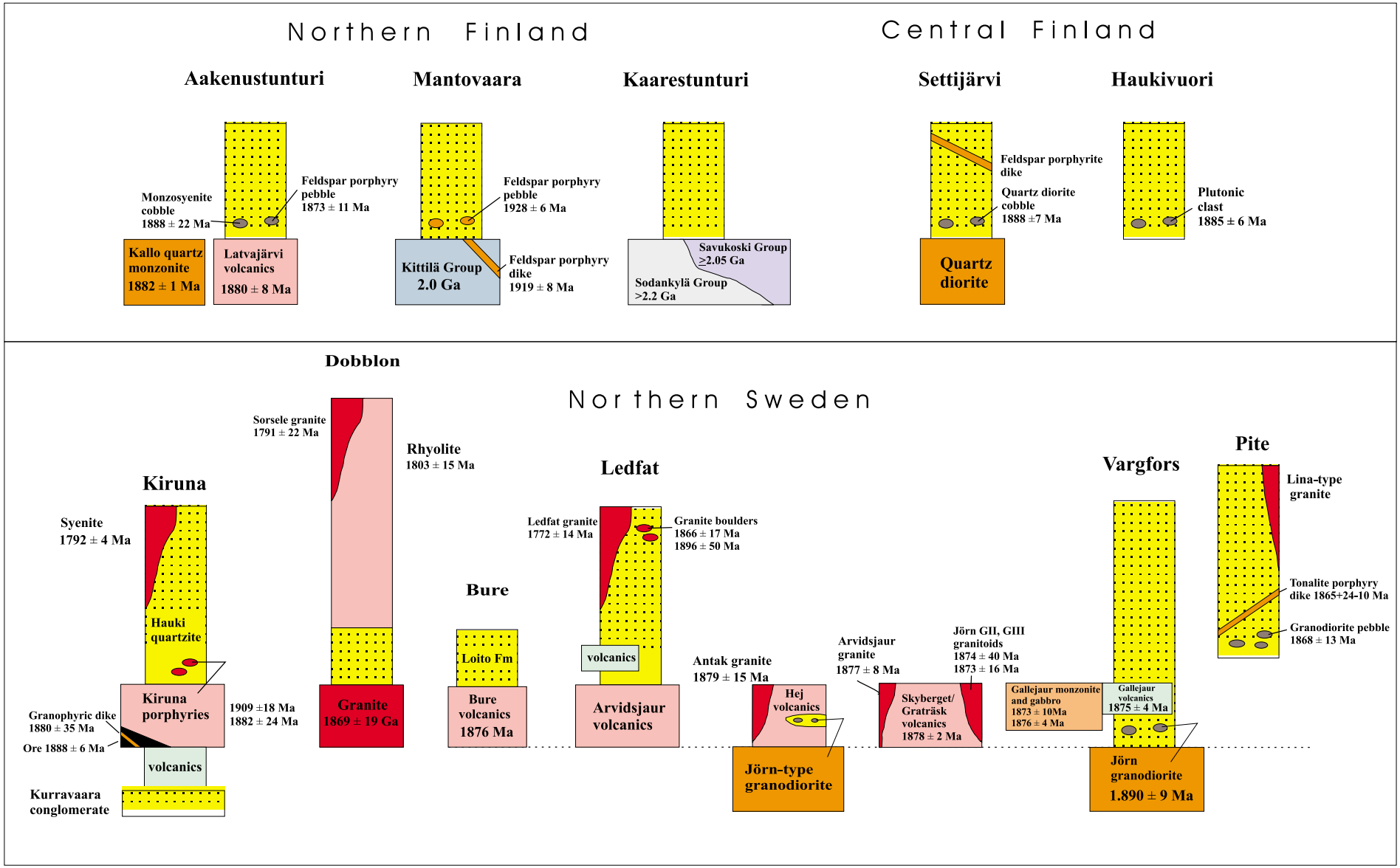

Fig. 7. A sketch showing simplified stratigraphic columns from northern part of the Fennoscandian Shield, containing sedimentary units with analogous age and setting as possessed by the Lainio and Kumpu Groups.

Korkiavaara sample would constrain the deposition of the "lower Kaleva" in the Peräpohja area to a very tight time interval between 1.98 Ga and 1.95 Ga. This contradicts the age of c. 2.05 Ga obtained for a felsic porphyry within the Paakkola Group rocks (Perttunen & Vaasjoki 2001, *this volume*). One solution could be that the Korkiavaara arkose represents a tectonic wedge of younger metasediments belonging to a so far unrecognized sedimentation cycle but no geological evidence for this interpretation is presently available.

The age determination of the Korkiavaara arkosite also causes contradictions in the stratigraphical correlation of rock sequences between the Peräpohja Schist Belt and the Central Lapland Greenstone Belt. The first appearance of significant black schist deposits in the stratigraphical sequence is found in the Savukoski Group lying conformably on the Sodankylä Group in Central Lapland and in the Paakkola Group lying similarly conformably on the Kivalo Group at Peräpohja. As was stated previously, the 2.05 Ga intrusions in Central Lapland determine the minimum depositional age for the Savukoski Group. On the other hand, detrital zircons separated from the Korkiavaara Formation arkosites suggest that the deposition of the Pöyliövaara Formation and correlative formations of the Paakkola Group took place later than 1.98 Ga. Solving this question clearly demands more geological and geochronological work.

The last issue handled shortly in this chapter is the possible correlation of the Lainio and Kumpu Group rocks with other rock units elsewhere in the Fennoscandian Shield, having a similar petrological character and tectonic position. Figure 6 shows examples of sedimentary units in the northern part of the Fennoscandian Shield which were deposited unconformably either on felsic metavolcanics or granodioritic plutonic rocks with ages of c. 1.88 Ga and thus are likely temporal counterparts of the Lainio and Kumpu Groups.

Starting from the north, the uppermost stratigraphic unit of the Kautokeino greenstone belt, the Çaravarri Formation, is composed of molasse-like quartzites and conglomerates (Siedlecka et al. 1985, Torske 1997) which are very similar to the Kumpu Group rocks and indeed have been correlated with them in the past (e.g., Geological Map, Northern Fennoscandia 1987). In northern Sweden, the Kiruna-Arvidsjaur Porphyry Group metavolcanics (Skiöld & Cliff 1984, Skiöld 1987, 1988) are known in many places to be overlain by coarse-clastic metasediments including conglomerates with pebbles and cobbles from the underlying felsic metavolcanics. These have been found, for example, in the following areas: Kiruna (the Hauki quartzite, Martinsson 1997; the Vakko quartzite, Ödman 1957), Vittangi (the Mattavaara quartzite, Eriksson & Hallgren 1975), Tärendö (the Haukkalaki conglomerate, Padget 1970), Pajala (the Palovaara conglomerate and Kuusivaara quartzite, Padget 1977). Moving further south to the Arvidsjaur region, analogous metasediments include the Loito Formation (Perdahl & Frietsch 1994) and the Ledfat conglomerates and quartzites (Offerberg 1959, Skiöld 1988). On the other hand, part of the Svecofennian terrestrial metasediments in northern Sweden have been demonstrated to lie directly on eroded, deformed felsic intrusions which are generally regarded as synorogenic plutonic rocks. In other cases there is indirect evidence for this kind of relationship in the form of granodioritic pebbles in conglomerates. Examples include the Dobblon conglomerate (Einarsson 1979, Skiöld 1988), the Vargfors Group metasediments (Kautsky 1957, Skiöld 1988) and the Pite conglomerate (Persson & Lundqvist 1997).

Crossing the Bothnian Bay to the Raahe-Ladoga zone in Finland, potential temporal counterparts are the Settijärvi and Haukivuori conglomerates containing syntectonic granitoid fragments with U-Pb zircon ages of 1888±7 Ma and 1885±6 Ma, respectively (Vaasjoki & Sakko 1988, Korsman et al. 1988). Further southeast, Vepsian red sandstones and quartzites in the western Lake Onega region, southern Karelia may also belong to this stage of sedimentation (Zagorodny et al. 1986, Heiskanen 1992) though the timing of deposition of these rock is so far poorly constrained. Figure 7 summarizes the radiometric age data and shows highly schematic stratigraphic sketches of the discussed post-1.88 Ga supracrustal formations in Finland and Sweden. It can be concluded that the temporal correlation of the Lainio and Kumpu Groups with Svecofennian metasediments (the informal Upper Svecofennian assemblage of Kousa et al. 2000) undermines the validity of the earlier stratigraphic schemes in which these rocks were assigned to the Karelian formations.

Lapland granulite

Although no new isotopic data are reported in this volume from the Lapland granulite belt in northern Finland, it is pertinent in this context to review the progress made in recent years in geochronological studies of the granulite belt. This progress has been achieved not only utilizing conventional U-Pb datings,

but also Sm-Nd isotopic and ion probe studies. Based on whole-rock Sm-Nd isotopic analyses, Huhma and Meriläinen (1991) concluded that the average provenance for metasedimentary granulites is Paleoproterozoic (T_{DM} c. 2.3 Ga), differing not much from the age of the provenance for Svecofennian and upper Kalevian metasediments. This was confirmed by single zircon ion probe analyses (both SHRIMP and NORDSIM), which provided a range of ages from 1.9 to 3.0 Ga (Sorjonen-Ward et al. 1994, Huhma 1996, Tuisku & Huhma 1998). The major provenance is about 2 Ga old and part of the detritus is younger than 2 Ga constraining the age of deposition. These studies and the U-Pb results on monazites (Meriläinen 1976) further suggest that granulite facies metamorphism took place at c. 1.9 Ga ago.

Summary

A summary of geochronological data from northern Finland is shown in Figure 8 as a cartoon containing the group-level lithostratigraphic units in the Central Lapland Greenstone Belt and Peräpohja Schist Belt together with approximate ages in billion years for some intrusive and volcanic rocks and pebbles in conglomerates. The obtained isotopic data record a long geological history in northern Finland: the Archean basement contains rocks with ages ranging between 3.1 and 2.6 Ga, while the deposition of the overlying Paleoproterozoic supracrustal sequence took place in the time interval of 2.45 Ga to less than 1.88 Ga, thus lasting for more than 550 Ma. The Paleoproterozoic evolution ended with the intrusion of post-orogenic granites c. 1.8 Ga ago.

The majority of the geochronological information come from mafic and felsic plutonic rocks which have been utilized to constrain the age of deposition of different lithostratigraphic units, though detrital zircons have also played a role in distinguishing quartzites between different stratigraphic levels. The Paleoproterozoic magmatic evolution shows an episodic character with ages of both mafic and felsic rocks being concentrated into several distinct groups. However, there are still uncertainties as to how strictly each episode is constrained in time, particularly when comparisons beyond the geographical boundaries of northern Finland are made. The available isotopic data appear to indicate tens of millions of years for some magmatic stages while some other

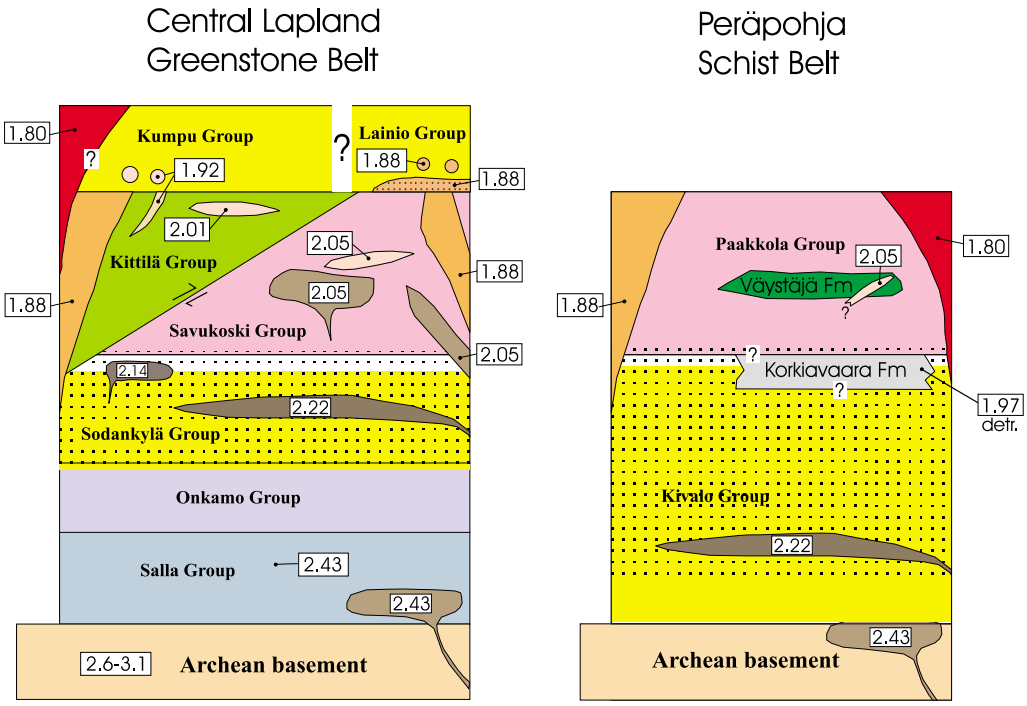

Fig. 8. Stratigraphical sketches the Central Lapland Greenstone Belt and Peräpohja Schist Belt with pertinent geochronological data for intrusive and volcanic rocks and pebbles in conglomerates. Ages are approximate and given in billion years.

intrusive events, such the 2.2 Ga mafic magmatism, seem to have occurred within a short time period and probably almost instantly over a large area. The scatter and large errors of the reported ages are in most cases not due to analytical reasons, but primarily result from the effects of inheritance or post-magmatic disturbances. Most of the dates are in accordance with the general geological information, but there are some rock units both in the Central Lapland Greenstone and Peräpohja Schist Belts exhibiting sharp discrepancies between their inferred geological position and isotopic dating results.

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