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RADIOACTIVE AGE OF SOME FINNISH PRE-CAMBRIAN MINERALS

BY

OLAVI KOUVO

With 3 figures and 10 tables

HELSINKI 1958

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PREFACE

This research was supported by the United States National Science Foundation under Project No. NSF—G 1959, as part of a program of research into the application of radiation damage measurements in determining the age of geologic materials. The support was made available by Professor Heinrich D. Holland, to whom the author wishes to acknowledge his deep obligation, both for his valuable aid, and his continuous interest in the work.

In the course of this investigation the facilities of the Department of Geology, Princeton University, of the Geochemical Laboratory of the Lamont Geological Observatory, Columbia University, of the Geophysical Laboratory and Department of Terrestial Magnetism of the Carnegie Institution of Washington, and of the Bureau of Mineral Research, Rutgers University, were placed at the author's disposal. To all these institutions the author is greatly indebted.

The guidance of Drs. George Tilton, Geophysical Laboratory, Henry Faul, U. S. Geological Survey, and Mr. Donald Miller, Lamont Geological Observatory, has made the experimental work on lead, uranium, and thorium possible. The rubidium, strontium, and potassium analyses are part of a cooperative program with Dr. Paul Gast. It covers about twenty samples from the U. S. A., Canada, and Finland. The Finnish material is presented in this paper. The analytical work was done cooperatively, and the kind assistance of Dr. Gast with the masspectrometric techniques was essential in obtaining the rubidium-strontium data reported here. The author is greatly indebted to them all for their kind assistance.

The Geological Survey of Finland and the Department of Geological Exploration of Outokumpu Co., Finland, were most generous in providing specimens from Fennoscandia for study. The sincere thanks of the author go to all collaborators who have assisted him.

This investigation is an immediate continuation of the author's studies at Princeton University supported by an award from the Outokumpu Foundation and the Higgins Predoctoral Fellowship, Princeton University. It is a pleasure to acknowledge these awards.

To Professor Aarne Laitakari, Director of the Geological Survey of Finland, the author must extend his sincere gratitude for kindness in permitting the publication of this paper by the Geological Survey.

Finally the author wishes to thank Miss Thyra Åberg who has drawn the pictures, and Miss Anne Grinstead for correcting the English in the manuscript.

Outokumpu, October, 1957

Olavi Kouvo

INTRODUCTION

In spite of the fact that natural minerals used for radioactive age determinations may be parts of open chemical systems, that huge periods of time are involved, and that especially in the pre-Cambrian formations one is possibly faced with the problem of several periods of recrystallizations of material, studies over the last twenty years have shown that by careful selection of samples for age work a significant value for $t_{measured}$, which is based on several independent radioactive decay schemes, can be obtained. This $t_{measured}$ has been supposed to be the time elapsed since the last crystallization of the material concerned. If these ages are found to agree with one another within ten percent, the ages are said to be concordant. If they do not agree, they are said to be discordant.

If N_P is the number of atoms of a radioactive parent, and N_D is the number of atoms of a daughter element, then $N_P^o - N_P^t = N_D$. Since the rate of radioactive decay has been found to follow an exponential law, it follows that

$$t = \ln (N_D/N_P + 1)/\lambda$$

The quantity λ, the decay constant characteristic of the isotope responsible for the activity is independent of time. The values of lambda used for different decay schemes in this study are given in Table VIII (p. 41).

The wide range of the reported values for the half-life of Rb^{87} is the result of the unusual distribution of the energy of its electrons. In a recent paper Aldrich, Wetherill, Tilton, and Davis (1956) have presented a new measurement of the half-life for Rb^{87} based on Sr^{87}/Rb^{87} ratios in rubidium bearing minerals from rock units, for which concordant U—Pb ages have also been obtained. The proposed value lies in the range of values found by the more recent direct counting experiments. The value is generally accepted now and has been used in calculating the age of the present materials. The value of the half-life, like that determined for K^{40} by Wetherill, Wasserburg, Aldrich, Tilton, and Hayden (1956) is based on the value of the half-life of U^{238} given by Kovarik and Adams (1955) and that of U^{235}

given by Fleming, Ghiorso, and Cunningham (1952). The minerals used for comparison have been assumed to be cogenetic.

Isotope abundance measurements can now be made on samples weighing only a few micrograms as a result of advances in stable isotope dilution and thermionic emission techniques. These techniques reviewed by Inghram (1954) are now well established for use in analytical problems and are employed for the determinations of isotopic composition in the present investigation.

If the quantity λ is known and the ratio N_D/N_P is right within the stated limits of error, $t_{measured}$ equals t_{true}, if 1. chemical exchange with surrounding material has not occurred, 2. primary N_D equals zero or can be accurately estimated. This assumption includes the possibility of several generations of the mineral concerned.

Progress in the understanding of geochronometry during the past few years has shown that combined uncertainties in analytical techniques and nuclear decay constants are always very much smaller than discrepancies in results based on several decay schemes from one mineral or several cogenetic minerals. With the exception of a few cases, the majority of discordant ages published are under rather than overestimates. In other words, in the ratio N_D/N_P the amount of daughter element is too low or that of parent element is relatively too high. The first mentioned alternative, loss of daughter element, is generally supposed to be the reason for low age, but gain in the amount of parent element cannot be excluded.

The atoms or ions produced in radioactive minerals by the decay of radionuclides are often misfits in the mineral structure. Together the presence of strange ions and the continuous bombardment by nuclear particles tend to impair the structure of the host mineral.

The process of daughter-isotope diffusion as a possible cause of the discrepant age measurements is examined in a quantitative manner by Nicolaysen (1957). Interpretations of lead ages by acid washing experiments are given by Tilton in several papers (e. g., Tilton, 1956, Tilton and Nicolaysen, 1957). The recent »catastrophic» loss of lead is discussed by Aldrich et al. (1955). Several authors have demonstrated the loss of intermediate members of the radioactive decay scheme. This includes the theory of radon leakage presented first by Wickman (1942).

The purpose of this paper is to produce more comparative age material, calling special attention to the possible selection of material for isotope work. Within the limits of the mineral ages obtained some Fennoscandian geological problems are discussed.

MINERAL AND ROCK SAMPLES STUDIED

SOURCES

The primary use for the present material was to serve the project of research into the application of radiation damage in zircon to problems of age measurements. The major portion of the material was collected with Dr. H. D. Holland in the western part of the United States, particularly in the Boulder Batholith, Big Horn Mountain, and the Black Hills. The remainder of the samples came from Finland. The isotopically controlled ages for the Finnish material are given in this study.

The minerals studied consist of 21 zircons, one monazite, seven biotites, three muscovites, three potash feldspars, and 17 galenas. In selecting material, special attention was paid to geologically well-known rock units supposed to be rich enough in zircon and covering some of the most important intrusive groups of the Fennoscandian root zones of the Karelidic and of the Svecofennidic fold-chains. Figure 3 (p. 50) shows the geographic distribution of samples.

Aulanko granodiorite, Torikatu, Hämeenlinna, Finland.

Urjala granodiorite, Soukko, Urjalankylä, Urjala, Finland.

The Aulanko granodiorite and the Urjala granodiorite occur on the map sheet Hämeenlinna (Geological Map of Finland, 1:100 000, Sheet 2131, Hämeenlinna) described by Simonen (1949). According to him, the infracrustal rocks on the map sheet occur conformably with the Svecofennidic schists. They have split the schists during movement, and continued orogeny has caused the gneissose texture especially in the northern part of the map sheet which includes the Urjala granodiorite.

The exact geographic and geologic situation of Hämeenlinna city, where the Aulanko sample was taken, is given by Simonen (1948). A detailed petrographic description of the granodiorite is also given in the same investigation. According to him (Simonen, 1949, p. 27), the outlines of the rock area are not quite similar to the general, elongated shapes of granodiorites and gneissose granites in southwestern Finland, and gneissose texture has been observed only occasionally near the boundaries. As further stated by Simonen (1948), the hypidiomorphic texture is especially well pre-

served. Hornblende and biotite are found as mafic minerals, and the biotite groups occur as pseudomorphs after hornblende. Hornblende is met in the kernels of the crystals. The zoned plagioclase (An_{30-34}) is also presented in drawings by Simonen (1948, 1949). The sample for this work was kindly selected by Dr. Ahti Simonen. The accessory zircon from this rock was used for the Pb^{207}—Pb^{206} and the biotite fraction for the Rb^{87}—Sr^{87} and for the K^{40}—A^{40} age determinations.

Parts of the large Urjala granodiorite area occur on the Hämeenlinna map sheet and on the Forssa map sheet (Geological Map of Finland, 1:100 000, Sheet 2113, Forssa) described by Neuvonen (1956). These maps do not include the locality, where the sample studied in this investigation was found. As stated above, the rock is more gneissose, granulated, and the biotite occurs as pseudomorphs after hornblende. Modes of some members of this intrusive series are given by Neuvonen (1956). Neuvonen has suggested that the primary anorthite component was higher than the measured one. Olivine and pyroxene relics have been found. Besides zircon and apatite, pyrite was the most abundant accessory mineral. The pyrite fraction was used for the study of the sulphur isotopic ratio mentioned in a later section. The sample for the age work from the Urjala granodiorite area was chosen by Dr. K. J. Neuvonen. Zircon from this sample was used for the Pb^{207}—Pb^{206} age determination.

83-GSF, Kalajoki tonalite, Heikkilä, Kalajoki, Finland. (Museum sample, Geological Survey of Finland.)

The commercial name of this biotite-hornblende-quartz diorite is Kalajoki granite or »Russian blue». It is dark gray in color and greyish black if polished. The quartz-diorites occur transitionally between diorite and gabbro, and they show very different degrees of metamorphism. In some areas the primary hypidiomorphic structure is perfectly preserved, whereas in other regions the rock has become crystalloblastic, showing, at the same time, a more or less distinct parallel texture (Mäkinen, 1916, p. 145). The formation of infracrustal rocks of this area, the coastal region of western Finland, took place during the orogenic movements of the Bothnian supracrustal complex, which consists of mica gneisses, mica schists, quartzites, quartz-feldspar schists, limestones, basic intrusives, and effusive rocks. According to Saksela (1953, p. 37), »Die Intrusivzone von Kalajoki—Perho, ..., ist ihrem Charakter nach wohl in gewissem Masse diapiritisch, aber sie stellt keineswegs einen Antiklinalbatholith dar, wie aus Verfassers Untersuchungen deutlich hervorgehen dürfte (Saksela, 1932a, b, 1933b, 1934, 1935, 1936). Sie ist zunächst ein »Achsialkulminationsbatholith» und stellt demnach eine ganz andere tektonische Intrusionsform als ein Antiklinalbatholit dar. Der Kalajoki—Perho-Batholith liegt, im Kartenbilde das allgemeine Achsialstreichen überquerend, in der Gegend einer deutlich her-

vortretenden Achsialkulmination und hat offenbar die Bildung der letztgenannten verstärkt ....» Mäkinen (1916) has pointed out the analogy between these rocks and the rocks of the Skellefteå region in Sweden, on the opposite side of the Bothnian Gulf.

Chemical analyses, modes, and detailed petrographic description of this rock are given by Mäkinen (1916). A microphotograph of this rock is given by Saksela (1933b). The geographic location can be seen on the General Geological Map of Finland, Sheet B 4, Kokkola, described by Saksela (1933b) and on the map published by Mäkinen (1916).

The accessory zircon from this sample was used for the Pb^{207}—Pb^{206} age determination.

95-GSF, trondhjemite, Heinänen, Uusikaupunki, Finland. (Museum sample, Geological Survey of Finland.)

As suggested by Goldschmidt (1916), the early removal of potash within the crystallization of biotite at an early stage of a differentiation could produce trondhjemites. According to Goldschmidt, the magma masses have possibly intruded through sediments which have supplied the water needed for the crystallization of biotite. The main point of interest in this sample is the primary nature of the biotite flakes in comparison with granodiorites like Aulanko or Urjala, where they are, in part at least, pseudomorphs after hornblende.

The locality of this sample (Heinänen, Uusikaupunki) and a detailed petrographic description is given by Hietanen (1943). The genesis of these rocks is also thoroughly discussed by her. The Svionian formations in southwestern Finland are intruded by the infracrustal rocks of this series. According to Hietanen (1943):

»Die Uusikaupunkigranite (Nystadsgranite) wurden bisher allgemein als Vertreter der jüngeren Urgebirgsgranite oder der Granite der 2. Gruppe Sederholms aufgefasst, in erster Linie wohl mit Rücksicht auf das »granitische» Aussehen und die scheinbar richtungslose Struktur der als Bausteine verwandten Varietäten. Nach den vorliegenden Untersuchungen aber zeigen diese Gesteine, unsere Trondhjemite, vorwiegend solche Characterzüge, die den Graniten der 2. Gruppe durchaus fremd sind. Ihre Zusammensetzung ist weitgehend trondhjemitisch; sie gehören zu einer vollständigen Differentiationsserie an, die dioritischen bis gabbroiden Glieder dieser Serie weisen alle Merkmale der entsprechenden, den »Urgranite» verbundenen gneisartigen Dioriten oder Hornblendegabbros auf. Ferner sind alle diese Intrusivkörper, einschliesslich die Massen von hellfarbigen, quartzreichen Trondhjemiten, typisch synkinematische, das Gefüge der Gesteine ist granoblastisch, die Regelung hat sich überall an den Rahmen angepasst. Folglich waren die Gesteine der Trondhjemitserie eher zur I Gruppe Sederholms zuzuzählen. Nur die pegmatitischen Granite, die die Trondhjemitserie sowie die supracrustale Formation häufig schonungslos durchsetzen, sind echte Vertreter der 2. Gruppe.»

In Simonen's orogenic classification these rocks are classified as belonging to the synkinematic intrusions of the I cycle (Svecofennidic) (Simonen, 1953).

Biotite from this rock was used for the Rb^{87}—Sr^{87} age determination.

658a-MH/Kaavi-54, oligoclase granite, Vihtajärvi, Kaavi, Finland.

209-AH/Kaavi-54, oligoclase granite, Mäntyjärvi, Kaavi, Finland.

Kaavi granite is supposed to be a part of the large late-kinematic Maarianvaara massif intruding the Karelidic schist zone (III group in Sederholm's classification). The rock is homogeneous, even-grained, slightly deformed oligoclase granite, rich in monazite and zircon. The shape of the zircon crystals differs from those of Maarianvaara granite being unusually long. Mäkinen has analyzed one type of Kaavi granite from Meltusvirta, Kaavi. According to this analysis (Frosterus and Wilkman, 1920, p. 139), the percentages of CaO, Na_2O, and K_2O are 2.16, 2.98, and 3.95 respectively. The mode of the same rock shows slightly trondhjemitic nature: plagioclase (An_{15}) 29.2 %, microcline 2.0 %, biotite 23.9 %, chlorite 3.0 %, quartz 40.2 %. The same feature can be seen from the analyses of the present samples kindly placed at the author's disposal by Mr. Aarto Huhma, M. A.: 209-AH: CaO 2.73 %, Na_2O 3.59 %, K_2O 1.28 %, and SiO_2 68.15 %. 658a-MH: CaO 1.78 %, Na_2O 3.64 %, K_2O 1.49 %, and SiO_2 73.60 %. The sample 658a-MH contains only biotite, and the 209-AH is slightly hornblende bearing.

The locality of the samples is shown on The General Geological Map of Finland, 1:400 000, Sheet D 3, Joensuu described by Frosterus and Wilkman (1920). The samples were supplied to the author by Maija Huhma, M. A., and by Aarto Huhma, M. A., whose recent maps have been the sources to which the author has had access. According to them (personal communication) the rocks are only slightly deformed; they penetrate the schists and contain inclusions from the surrounding rocks. Some short and darker zircon grains were found among the very long zircons separated for age determinations. They are more like the zircons of the Maarianvaara granite but are most probably derived from the surrounding schists. Besides the shape of zircon, the abundance of monazite distinguishes the Kaavi granite from the proper Maarianvaara granite. The Kaavi granite was the richest in monazite of the approximately ninety samples, separated within the scope of this program for their accessory minerals (about 250 g/ton).

According to Sahama and Rankama (1938), the actual Maarianvaara region was found to be very poor in rare earths. A quantitative analysis of granite pegmatite from Luikonlahti and of monazite from the same rock to ascertain the amount of rare earths is given by Sahama and Vähätalo (1941).

The zircon and the monazite from these samples were used for the Pb^{207}—Pb^{206} age determinations and the biotite for the Rb^{87}—Sr^{87} age.

Maarianvaara granite, 25/27 km, E-side of the Outokumpu—Kaavi highway, Kaavi, Finland.

The locality of this sample can be seen on the map published by Vähätalo (1953, Appendix I). A short description of the Maarianvaara granite is also given by him (ibid., p. 15—16), and by Frosterus and Wilkman (1920). The Karelidic orogenic belt is intruded by this granite complex, and it is comparable to the Kaavi oligoclase granite mentioned above. The palingenic nature of this granite has been stressed by Eskola (1956a).

The treated sample was selected for age work by Mr. Erkki Viluksela, M. A. Zirzon from this rock was used for an approximate Pb^{207}—Pb^{206} age determination.

Granite pegmatite vein, Outokumpu mine, Kuusjärvi, Finland.

Some granite pegmatite dikes are found in the Outokumpu mine. According to Vähätalo (1953), the thickness of the dikes ranges from about 2 m to a few centimeters. They penetrate the ore and schist formation, and the contacts are sharp against the wall rock. They are supposed to be connected with the Maarianvaara granite mentioned above. These veins are thoroughly discussed by Vähätalo (1953). Only a few zircon grains were met in the heavy fraction separated from the sample. The heavies consisted almost entirely of garnet, rich in iron and manganese. Some biotite flakes were found among the muscovites separated from the rock.

Muscovite from this rock was used for the Rb^{87}—Sr^{87} age determination.

Kitee pegmatite granite, Kitee, Finland.

The age of two samples from this massif was measured at the suggestion of Professor Pentti Eskola. The samples were kindly supplied to the author by Aarto Huhma, M. A., from a location four kilometers southwest of Kitee church. The locality of these samples can be found on the General Geological Map of Finland, 1 : 400 000, Sheet D 2, Savonlinna (Hackman, 1933), in the narrow neck on the southern shore of Lake Kiteenjärvi. Both samples, labelled I-Kitee and II-Kitee, were taken from the area between the shore of the lake and the highway running to the southeast. The locality of these samples is of special interest because of the Russian age determinations for rocks originating from the region north of Lake Ladoga.

The Kitee pegmatite granite pentrates the Karelidic orogenic belt. According to Hackman (1933), who has given a description of the rock, it belongs to the youngest pegmatites on the Savonlinna map sheet except the rapakivi group. This map sheet includes a typical Karelidic schist belt around Lake Jänisjärvi and on the eastern side of Lake Orivesi and the more gneissose belts striking from Lake Saimaa to the northeast. Hence its age has to be regarded as representing the lower age limit of all of these formations.

The Kitee heavies consisted almost completely of garnet as did the heavy fraction from the Outokumpu pegmatite. The x-ray fluorescence spectrum

showed that the ratio Mn/Fe is not as high as in the Outokumpu garnet. Only a few zircon grains were found.

Muscovite from the I-Kitee and II-Kitee samples was used for the Rb^{87}—Sr^{87} age determination. A K^{40}—A^{40} age measurement was made also for the II-Kitee muscovite.

Gneissose granite, Sotkuma, Polvijärvi, Finland.

As can be seen from The General Geological Map of Finland, 1:400 000, Sheet D 3, Joensuu (Frosterus and Wilkman, 1920), several openings in the Karelidic schist belt are shown to be composed of the basement complex. They represent the mantled gneiss domes described by Eskola (1949) in the Pitkäranta region north of Lake Ladoga and they are described in detail by Frosterus and Wilkman (1920). In all cases, pebbles of the underlying dome gneiss can be found in the surrounding sedimentary beds. The size of the pebbles varies from microscopically discernible pieces to 30 cm. The intrusive character of some southern gneissic domes is explained by Eskola (1949) as being the result of the reactivation of the old material by metasomatic granitization during a later folding. The same phenomenon has been described in several regions, for instance, in the Appalachian and Caledonian belts and in the Alps. No intrusive veins have been found around the Sotkuma dome.

The locality of this sample can be found on the map sheet Joensuu at the southern end of Reuhajärvi on the eastern side of the highway going to Joensuu. Biotite from this sample was used for the Rb^{87}—Sr^{87} age and a zircon fraction for the Pb^{207}—Pb^{206} age.

Onas granite, south of Porvoo city, about 1.3 km south of the junction Vassö.

The granite region of Onas is situated about 30 km east of Helsinki. The geology of the area is described by Sederholm (1923), and by Wegmann and Kranck (1931). The granite has been described by Borgström (1931), as well as a granite porphyry vein probably running from the Onas granite (Borgström, 1907).

The Onas granite forms a homogeneous body of about 270 km^2 and is a rather coarse-grained potash granite. According to an analysis published by Borgström, the SiO_2, CaO, Na_2O, and K_2O percentages are 73.78 %, 0.92 %, 2.50 %, and 6.49 % respectively. It is the youngest formation in this area and, as stated by Barth (1938), may be contemporaneous with the rapakivi formation. According to Sederholm (1926), the small coarse-grained porphyritic granite massives like Onas, Bodom, and Obbnäs (west of the big Viipuri rapakivi massif), Lemland and Mosshaga S and E of the Åland rapakivi massif, Åva SW of the Vehmaa rapakivi, and Kokemäki E of the large Laitila rapakivi area are between the Hanko granite and the rapakivi granite in age. Yet he was inclined to think that their age is nearer to that

of the rapakivi granite than that of the Hanko granite. The rock is rich in fluorite and zircon. Biotite and hornblende occur as colored constituents. Microcline is the typical feldspar.

Piggot (1938) has presented the radium content for several intrusive rocks of the four granite groups of Sederholm. An average of 1.53×10^{-12} grams Ra per gram of rock is given for the fourth group, including the rapakivi granite and some of the small potash granites mentioned above. The value for Onas granite is 1.03×10^{-12} grams Ra per gram of rock.

This sample was kindly selected for the age work by Dr. Ahti Simonen. Zircon was used for the isotopically controlled U—Pb and Pb^{207}—Pb^{206} ages.

29-GSF, Kokemäki granite, Kokemäki, Finland.

This sample, also labelled as 616-GSF, was submitted to the author by the museum of the Geological Survey of Finland.

The Kokemäki granite is a coarse-grained potash granite and is rich in fluorite and zircon. Biotite is the main colored mineral. The location can be seen on the General Geological Map of Finland, 1 : 400 000, Sheet B 2, Tampere (Sederholm, 1913). According to the description given by Sederholm (1913), the rock is very similar to rapakivi rocks, but it is slightly deformed, and oligoclase shells are not found around the potash feldspar grains. Yet it is less deformed than the other porphyritic granites on the map sheet and it resembles mostly the Onas and Mosshaga granites mentioned above. He has pointed out that they could be correlated with the post-Kalevian granites in North Finland and are possibly partly younger than those.

Zircon from this rock was used for the Pb^{207}—Pb^{206} isotope age.

Bodom granite, Espoo, Finland.

A petrographic description of the Bodom granite is given by Sederholm (1926).

Bodom granite occurs as a rounded, sharply-defined massif 16×5 km in size. The rock is mostly equigranular and only partly porphyritic. As stated by Sederholm, the biotite occasionally possesses good crystal form towards the microcline but is also in many cases — as shown by some photographs — decidedly xenomorphic. It has been seen also to encroach upon the plagioclase. In some varieties hornblende is the prevalent mafic mineral. Fluorite, zircon, orthite, and apatite are the predominant accessories. According to Sederholm, the quartz and biotite often show signs of having been crushed without having undergone any subsequent recrystallization.

According to a very detailed description of the contact phenomena of the Bodom and Obbnäs granites given by Sederholm, these granites are decidedly younger than the surrounding rocks. The observations in the Åland Islands made by Sederholm between the rapakivi formation and some

potash granites analogous with the Bodom suggest that the Bodom granite is also intermediate in age between the Hanko type of granite and the rapakivi granites.

The main difference between the accessory minerals in the Bodom and in the rapakivi granite is the abundance of orthite in the Bodom granite. Sahama and Vähätalo (1941) have given the rare earth content of the orthite of Bodom and that of the Bodom, Onas, and Obbnäs granites.

Halonen (1954) has more recently studied the Bodom granite. He also selected two samples for the present work. The first one labelled I-Bodom represents the even-grained hornblende bearing type and the second one, labelled II-Bodom, is the porphyritic type, in which biotite is the predominant mafic mineral. As stated later, a slight difference in age has been found for the biotites separated from these two types of rock.

Zircon from the I-Bodom sample was used for determining the U—Pb, Th—Pb, and the Pb^{207}—Pb^{206} isotope ages. Biotite from the I-Bodom and II-Bodom samples was used for determining the Rb^{87}—Sr^{87} age. A K^{40}—A^{40} age measurement was done for the I-Bodom biotite.

Rapakivi formations.

Nine samples were treated from the Finnish rapakivi areas. Eight of them were from different parts of the large Viipuri rapakivi area (35b-AS/Lemi-54, 805-ML/Iitti-54, 672-ML/Valkeala-54, 29-KM/Virojoki-55, 785-ML/Valkeala-54, 714-ML/Valkeala-54, 297-FP/Mentula-55, and 299-FP/Kiesilä-55). On the northern side of the main Viipuri rapakivi area there is a small separate Suomeniemi rapakivi massif. One sample, 292-FP/Suomenniemi-54 is from there. One sample is from the Åland Islands. All these samples have been obtained from the Geological Survey of Finland through the kindness of Dr. Ahti Simonen. A detailed description of the geology of the samples will be given by him.

Five zircon samples were separated by Dr. K. J. Neuvonen:

672-ML Pihlajasaari, Valkeala, Finland

805-ML About 2 km to the west of Hiidensaari island, Lyöttilä, Iitti, Finland

292-FP Suomenniemi, Finland

785-ML Valkeala, Finland

714-ML Valkeala, Finland

Only zircon fractions from these samples were available.

The sample 29-KM from Falin, Muurikkala, Virojoki, Finland is typical viborgite, in other words, a porphyritic rapakivi variety characterized by oligoclase shells around the potash feldspar grains. Biotite and hornblende are the main dark components. The rock is rich in fluorite, zircon, and apatite. Zircon was separated for determining the U—Pb and Pb^{207}—Pb^{206} ages and the biotite fraction was used for determining the Rb^{87}—Sr^{87} age.

35b-AS, dark, pyroxene bearing rapakivi variety, Värrö, Lemi, Finland.

This dark green rapakivi variety is characterized by quartz, microcline, oligoclase, hornblende, fayalite, biotite, and augite. The main accessories are fluorite, zircon, apatite, and magnetite. Several authors have described this type of rapakivi (Wahl, 1925; Laitakari, 1928; Hackman, 1934; Kranck, 1929; Savolahti, 1956).

Zircon from this rock was used for the U—Pb, Th—Pb, and Pb^{207}—Pb^{206} isotope ages.

297-FP, quartz-porphyry vein, Mentula, Suomenniemi, Finland.

299-FP, quartz-porphyry vein, Kiesilä, Suomenniemi, Finland.

The Mentula vein is about 100 m thick and can be followed from the Suomenniemi rapakivi massif. The Kiesilä quartz-porphyry vein cuts through the Suomenniemi rapakivi massif. The locality of the Suomenniemi rapakivi massif can be seen on the General Geological Map of Finland, 1:400 000, C 2, Mikkeli, described by Frosterus (1902).

Zircon from the Mentula vein was used for the lead:alpha age and zircons from both samples for measurements of radiation damage.

Zircon from the Åland quartz-porphyry was used for a radiation damage measurement. This sample originated from the museum of the Geological Survey of Finland.

Zircon from the 805-ML rapakivi mentioned above was used for the isotopically controlled U—Pb and Pb^{207}—Pb^{206} ages. The 292-FP, 785-ML, and 714-ML zircons served for the radiation damage measurements. The lead:alpha age measurement was made also for the 292-FP, 785-ML, and 672-ML zircons.

The isotopic composition of lead from some galenas and from a native lead sample was studied in an attempt to relate the age of mineralization to the broad metamorphic period of the host rock. The areas were exceptionally well-suited for this type of study, since the major rock units into which the lead minerals were emplaced had been dated by the Rb^{87}—Sr^{87}, K^{40}—A^{40}, and by the isotopically controlled U—Pb and Th—Pb methods.

The number of the samples is that used at the Lamont Geological Observatory, Columbia University. The analyses are reported in the data accumulated on Contract AT(30—1)—1114.

Pb-175, galena, Pakila, Finland.

The sample was obtained from the Geological Survey of Finland. It is from a small galena mineralization found at Pakila near Helsinki. According to Kulonpalo (1946), the country rock, biotite gneiss, is intruded by pegmatite veins, and includes some galena-quartz and galena-sphalerite veins. Some galena-sphalerite as well as pyrite-chalcopyrite-pyrrhotite impregnations are found also in the biotite gneiss.

Pb-238, galena, Korsnäs, Finland.

The galena mineralization of Korsnäs in western Finland was found recently and is not fully described. The occurrence is typical epigenetic mineralization taking place in a breccia and shear zone in the Svecofennidic zone. Lithologically the Korsnäs area is like the leptite areas of southern Finland, where there are carbonatite horizons bearing pyroxene gneisses. The ore is situated in a carbonatite horizon. The mineralization has taken place in a skarn rock formation.

Pb-228 and A-48, galena from the Outokumpu mine, Kuusjärvi, Finland.

The occurrence of galena in the Outokumpu ore is discussed by Vähätalo (1953). Galena is rare in this ore, and the author is indebted to Mr. Paavo Kupias for a sample found in the ore. The crystal was situated in a chalcopyrite-rich vein cutting through the ore.

A-58, galena, Outokumpu, Kuusjärvi, Finland.

The sample consists of a small galena crystal, associated with a pyrrhotite grain found by Mr. Yrjö Vuorelainen in a small fissure of a crushing zone about 20 m below the Outokumpu ore.

Pb-239, carbonate vein intruding the Karelidic schists, Panjavaara, Juuka, Finland.

Some galena was submitted to the author by Mr. Erkki Viluksela, M. A., from a carbonate-barite vein cutting through the Karelidic schists. Lead extracted from these samples was analyzed for its isotopes.

Pb-240, native lead, Keihäsjoki, Juuka, Finland.

Some grains of native lead were found by Mr. Erkki Viluksela, M. A., in soil in Keihäsjoki. Some of the grains were associated with carbonate minerals. The isotopic composition of this lead was studied.

Pb-247, galena, Nunnanlahti, Juuka, Finland.

Some galena was found in the amphibole skarns between the serpentine rocks and amphibolites in Nunnanlahti, Juuka. The sample was kindly submitted to the author by Mr. E. Penttilä.

Vaasjoki (1956) has compared by ore-microscopic and spectrographic methods some galenas from the occurrences in Svecofennian rock zones and from those within the rapakivi areas in Finland. The lead from these samples has been studied for their isotopes within the scope of this program. Those samples from Orijärvi, Aijala, Attu, Koskenkylä, Lemi, Sottunga, Luumäki, and Säkkijärvi are described by him (Vaasjoki, 1956) and not repeated in this paper.

SEPARATION OF MINERALS

The concentration of heavy accessories from rock samples was done by means of standard techniques. The rock samples were crushed and ground

in Denver laboratory jaw crushers and crushing rolls. A McCool laboratory pulverizer was used for grinding a part of the last biotite fraction. After every crushing, the material was sized by means of the Denver-Dillon vibrating screen, which was supplied with 30 and 60 mesh screens. Crushing was continued until most of the material was of -30 mesh size. Only the -60 mesh material was used for further work. The remaining +30 mesh fraction consisted mostly of mica. This was cleaned and used for the Rb^{87}—Sr^{87} and K^{40}—A^{40} age measurements.

A Denver-Wilfley laboratory concentration table was adapted for treating the fine fractions. The heavies were then cleaned by bromoform separation in a large separatory funnel. Biotite and other dark minerals, as well as monazite, were then removed in a Frantz electrodynamic separator set up for vertical feed. The zircon was left with the light-colored minerals for separation by centrifugation in Clerici solution. The Frantz electromagnetic separator set up for inclined feed was used to separate the zircon from the other constituents of higher magnetic susceptibility. Pyrite was removed either by an acid treatment or in a small flotation cell. Plenty of molybdenite was found in almost all zircon fractions. It was removed by a sheet of paper if it was coarse grained or by careful hand-picking. The zircon concentrate was then cleaned by means of the Frantz separator and by means of hand picking.

An x-ray fluorescence spectrum was run for all zircon samples with a view to checking the possible contamination of lead found in the Clerici solution. The Tl_{L_α} and Tl_{L_β} peaks were used for checking the presence of thallium.

To separate the biotite from the hornblende, an air table (Sutton, Steel and Steel) was supplied. In most cases sieving was found to be the most convenient way of obtaining a biotite fraction. A sheet of paper was usually used for the final cleaning of biotite. Remaining impurities were picked by hand from the sample.

DESCRIPTION OF SAMPLES

The morphological character of zircons indicates some applications of zircon studies to petrogenetic problems. The fundamental idea is that the form of the free-growing crystal depends on the physico-chemical conditions of the melt, if the grains are self-nucleated. Zircon is used, e. g., to distinguish connected or disconnected intrusives (Larsen and Poldervaart, 1956) and to establish mixing of magmas (Larsen et al., 1938). On the other hand several workers have been able to show that the overgrowth of zircons is characteristic of migmatic origin.

Material for elongation curves has not been collected from the present samples. However, even a cursory inspection of the zircon fractions shows certain well-defined features.

Zircons from the rapakivi granite and from the granites of the third group are surprisingly similar. Zircons are simple, uniformly elongated tetragonal prisms terminated by a dipyramid. They are strongly zoned and rich in inclusions. The zircons from three rapakivi quartz-porphyries are also similar but strongly pigmented. The first description of the rapakivi zircons was given by v. Ungern-Sternberg (1882).

The Maarianvaara granite complex consists of several small massives. The Kaavi granite is north of the actual Maarianvaara granite and is generally thought to be of the same age as Maarianvaara granite and to have a similar tectonic history. Zircons from two samples of the Kaavi oligoclase granite and from one sample of the Maarianvaara microcline granite show a distinct morphological difference. The Kaavi zircons are extremely long and simple tetragonal prisms in shape. The Maarianvaara zircons are short, dark, and richer in crystal faces.

Simonen (1949) has pointed out the different occurence of the Urjala and Aulanko granodiorites. The Urjala zircons are extremely poor in inclusions. The Aulanko zircons have numerous small inclusions.

Zircons from basic and from some intermediate rock types have been found to be rich in small cavities, which are explained as being the result of corrosion. Zircons from one fraction studied showed the same phenomenon. These zircons were from a hornblende gabbro.

The values of alpha activity for the Finnish samples are given in Table VII. The inhomogeneous distribution of uranium and thorium in the studied zircons is discussed elsewhere (p. 37). The total alpha activity of most of the samples studied lies between 150 and 385 α/mg/hr. The average activity of about 50 accessory zircons from different types of rocks were as follows (upper and lower limit in parenthesis): from hornblende gabbros 252 (169—324), from granodiorites 206 (187—260), from quartz monzonites 280 (135—655), from oligoclase granites 216 (131—300), from rapakivi granites 240 (120—544), from a dark pyroxene bearing rapakivi granite 109, from rapakivi quartz-porphyries 234 (206—266), from microcline granites 291 (144—405), from syenites 110 (86—160), and from dacites 343 (307—389) alphas/mg/hr.

Unit cell dimensions for zircons are given in Table VII. The broad (200)-peak was found to indicate an inhomogenity in the metamict state of zircon. In an extreme case a double peak was obtained, when very coarse-grained accessory zircon was treated.

In general the fine-grained zircon fraction was found to be more suitable for age determinations, because the foreign inclusions were fewer in the smaller

crystals and because the smaller grains tended to be more homogeneous than the larger ones. Overgrowths which envelop the cores of the numerous large rapakivi zircons are completely absent in the fine-grained fraction.

One of the uncertainties in the use of radiation damage measurements in zircons as dosage indicators is the effect of the chemical composition of the zircons on their response to nuclear radiation. The concentration of uranium, thorium, hafnium, and yttrium in zircon samples was determined by means of an x-ray fluorescence technique. Preliminary data suggested that the effect of differences in composition is negligible in most zircons.

X-ray powder patterns were made for the identification of the three muscovites worked on for the Rb^{87}—Sr^{87} and K^{40}—A^{40} ages. It was concluded from these data that they have 2M structure (Yoder and Eugster, 1954 and 1955). The large flakes of muscovite in the I-Kitee sample are subhedral and show undulatory extinction like the plagioclase grains in the same rock. Plenty of sillimanite needles can be found. The muscovite in the II-Kitee sample contains only a few sillimanite needles and the rock is not deformed like the I-Kitee sample. The muscovite from the I-Outokumpu sample does not show alteration to sillimanite like the other ones. The flakes are subhedral, but not as well developed as the Kitee muscovites. All three samples of rock are rich in garnet.

The biotites from the even-grained, hornblende bearing I-Bodom granite and from the porphyritic II-Bodom biotite granite are xenomorphic and have narrow streaks of microcline and quartz along the cleavage planes. Microscopic study as well as the x-ray spectrograms indicate an alteration to chlorite. At least a part of the biotite in the I-Bodom sample is a product of secondary alteration of hornblende.

As stated above, the hornblende in the Aulanko sample shows relics of pyroxene and alteration to biotite. The biotite has become chloritized and is xenomorphic in shape. The hornblende and the biotite are very intimately associated.

The biotite from the Kaavi oligoclase granite and from the Uusikaupunki trondhjemite is supposed to have crystallized at an early stage. In both of these samples (658a-MH and 95-GSF respectively) the biotite flakes show a subhedral shape. Some alteration to chlorite can be found in the Kaavi biotite. The Uusikaupunki biotite does not show alteration but among the large biotite flakes some small chlorite grains can be found.

In the 29-KM rapakivi sample alteration of hornblende to biotite and of biotite to chlorite can be seen. Most of the biotite is still supposed to be a product of primary crystallization.

The Sotkuma gneiss granite is badly altered. The biotite also shows the process of replacement by chlorite. Numerous small separate chlorite grains can be found in the sample.

EXPERIMENTAL PROCEDURES

ANALYTICAL TECHNIQUES

The techniques developed at the Geophysical Laboratory and the Department of Terrestial Magnetism of the Garnegie Institution of Washington and at the Lamont Geological Observatory were used for the materials at hand. Most of the analytical work was carried out at the Geochemistry Section of the Lamont Geological Observatory, Columbia University.

The stable isotope dilution technique (Inghram, 1954) for the quantitative determination of each of the elements involved in the five decay schemes was used for all the analytical results reported in this paper, except for determinations on lead isolated from some feldspars and galenas. The methods have been discussed in some detail in many connections (Aldrich, Doak, and Davis, 1953; Aldrich, 1956; Aldrich, Davis, Tilton, and Wetherill, 1956; Bate, Miller, and Kulp, 1957; Gast, Kulp, and Long, 1957; Tilton, Aldrich, and Inghram, 1954; Tilton et al., 1955; Tilton and Nicolaysen, 1957; Tilton, Davis, Wetherill, and Aldrich, 1957) and only the broad outlines of most of the procedures used are given here.

L e a d. Zircon samples were cleaned by using conc. HNO_3 (if sulfides were visible; otherwise 6N HCl was used), washed in double-distilled water and dried. A 200—700 mg sample was fused in about four grams of borax to clear glass (about 10 minutes) in a 40 ml platinum crucible. This was digested in 3N HCl (usually overnight), the supernatant liquid was transferred to a 150—250 ml beaker, double-distilled water and a little 3N HCl were added to the crucible, warmed for an hour and the treatment repeated until all the solids had gone into solution. The silica was destroyed by hydrofluoric and perchloric acid digestion. The possible sediment was treated once more with one gram of borax. A 250 ml volumetric flask was used for the original and the resulting solution, and the content was diluted to the mark with double-distilled water.

A 100—150 ml aliquot was taken for the determination of lead isotopic composition. A 75 ml aliquot was »spiked» with Pb^{208} or Pb^{204} standard solution and used for the lead concentration analysis. The rest of the solution was used for uranium and thorium analyses.

Lead was isolated from the aliquots by extracting it into dithizone at a pH of 8.5—9; the first extraction was carried out in the presence of 30 % ammonium citrate. 15 ml of 2 % potassium cyanide was used in the second dithizone extraction to prevent the extraction of interfering elements. The dithizone was then shaken

with 2—3 ml of 2 % HNO_3; the lead bearing solution was poured into a 10 ml beaker and dried by evaporation. The lead was then transferred into a 3 ml centrifuge tube by using one ml of 2 % HNO_3.

The pH was adjusted to 4.5 with dil. ammonium hydroxide and the solution heated. Hydrogen sulfide was bubbled in, and the resulting lead sulfide precipitate was centrifuged out. The solution was then discarded, leaving about 1/10 ml for mass spectrometer filament.

Part of the lead sulfide was transferred to an outgassed tantalum ribbon filament using an acid-washed capillary tube. The importance of the presence of ammonium nitrate in the loading of lead sulfide is pointed out by Aldrich, Davis, Tilton, and Wetherill (1956), and by Tilton and Nicolaysen (1957).

The mass spectrometers used for the lead runs were the Department of Terrestial Magnetism 60°, 6 inch radius of curvature, Nier-type instrument, equipped with an electron multiplier, described by Aldrich, Davis, Tilton, and Wetherill (1956) and the same type of instrument at the Lamont Geological Observatory, described by Bate, Miller, and Kulp (1957).

The average of mean deviations calculated for the isotope abundance measurements is 1.55 percent. This number includes some anomalous poor runs.

The average of runs taken at several different intensities was usually used for the final result. The even peak height was supposed to be the most reliable. A small correction was made for the resistance of the voltage multiplier bridge following the vibrating reed electrometer.

Seventeen galenas were analyzed for their lead isotopes using leadtetramethyl introduction and $Pb(CH_3)_3^+$-spectrum. All samples were prepared for the mass spectrometer by Miss Elisabeth Hodges using the following technique:

Dissolve the galena in 3N HNO_3; filter and precipitate the $PbCl_2$ from the filtrate using HCl. Filter the crystals and dry overnight at 110°C. Pipette under nitrogen, anhydrous ethyl ether and commercial methyl magnesium bromide into the glass reaction flask, and after thorough stirring, slowly add to 200 mg of $PbCl_2$. Allow the ether to reflux for three hours, and stop the reaction by adding water dropwise until all the excess Grignard Reagent is destroyed. Decant the ether layer, and, using anhydrous calcium chloride, dry the solution overnight. Distill the ether from the sample using a small Vigreaux column equipped with a pouring spout. When the distillation is complete, pour the light yellow viscous tetramethyl lead into sample containers fitted with high vacuum stopcocks and ground glass joints. The sample is further purified on a low vacuum system using the difference in vapor pressure to separate the remaining ether from the tetramethyl.

The mass spectrometer II, described by Bate, Miller, and Kulp (1957), was employed in the course of the isotope analysis of prepared tetramethyllead.

Correction was made for the carbon-13-carbon-12 effect and for the hydrogen loss and hydride formation discussed by Bate and Kulp (1955)

and Bate, Miller, and Kulp (1957). All samples were run in duplicate or in triplicate. In routine work the average reproducibility was found to be better than 1 % for Pb^{204} and better than 0.5 % for Pb^{206}, Pb^{207}, and Pb^{208}.

U r a n i u m and t h o r i u m. The 25-ml aliquot taken for uranium and thorium analyses was »spiked» with 1.00 ml of known uranium standard solution and with 1.00 ml of known thorium standard solution, diluted to 100 ml and warmed with a small amount of carrier solution containing about one mg of iron. The hydroxides were precipitated with ammonia gas and centrifuged in a 50-ml tube. A few drops of concentrated nitric acid were used for dissolving the precipitation, 25 ml of nitric acid saturated $Al(NO_3)_3 \cdot 9H_2O$ (washed with hexone) was added, and uranium and thorium were isolated into hexone by shaking the solutions for about two minutes in a 60 or a 125 ml separatory funnel. Hexone was poured into a dry beaker, and then shaken with ten ml of water in a 60 ml separatory funnel and evaporated to dryness. Fifteen ml of saturated ammonium nitrate (washed with acid washed hexone) was added, the solution warmed, and uranium and thorium extracted from the cool solution with hexone. Hexone was removed into a dry beaker, the funnel rinsed and hexone shaken with 10 ml of water, and this extract was dried by evaporation.

Uranium and thorium were transferred to the mass spectrometer filament using a capillary tube and 1—2 drops of 10M HNO_3 or water. According to Tilton, Davis, Wetherill, and Aldrich (1957), the recovery of uranium is about 90 %, while that for thorium is only about 30 %. Thorium was determined only for two zircons.

All treated zircon fractions (200—700 mg) were rich enough for lead analysis. The alpha activity of samples varied from 109 to 379 alphas/mg/hr, and the lowest content of lead was about 50 ppm. All zircons were pre-Cambrian in age, most being about 1600—1800 m. y. old. The average lead content of the last precipitation was about 40—50 micrograms.

P o t a s s i u m, r u b i d i u m, and s t r o n t i u m.

The determinations of potassium, rubidium, and strontium are described by several authors mentioned above. The analytical methods used in different laboratories vary somewhat, but are the same in principle. The method followed and partly developed at the Lamont Geological Observatory (Gast, 1957, and personal communications) was used in the course of this work.

A sample of 1.0 to 2.5 grams was weighed into a 100 ml or 150 ml platinum dish. The sample was »spiked» for strontium by adding the known Sr^{86}-standard solution from a calibrated pipette. A known amount of Sr^{89}-tracer was added and used to check the yield in precipitations and ion exchange separations. The sample was dried by evaporation to drive off the HCl.

The dry sample was then treated with a mixture of 5—10 ml of 60 % hydrofluoric acid and 2—5 ml of $HClO_4$, the minerals decomposed, silicon tetrafluoride driven off, and the resulting solution taken to a moist »mush» of perchlorates. Water was then added, the sample warmed and dissolved. If not completely decomposed or dissolved, the sample was treated once more.

Both solutions were diluted to 100 ml in a volumetric flask, mixed well, and a known aliquot of 2—2.5 % removed with a calibrated pipette. In some cases the aliquot was checked gravimetrically.

The aliquot was »spiked» with known amounts of K^{41}- and Rb^{87}-standard solutions, evaporated to one ml and transferred to a small round bottom beaker. Some drops of $HClO_4$ were added; the sample was cooled, the solution removed, and the perchlorates dissolved in a small amount of water. From the clear solution sulfates were precipitated and a sulfate solution used for mass spectrometer filament.

The remaining sample was used for strontium analysis. The solution was evaporated to about 30 ml, cooled, and alkali perchlorates centrifuged off. In some cases the perchlorates were counted to check the possible strontium.

The strontium bearing solution was transferred into a 50-ml centrifuge tube. The hydroxides were precipitated with ammonia gas and centrifuged out. For some samples MgF_2 was used to carry down the strontium. Hydroxides were dissolved in some drops of 6N HCl and fluorides in about 10 ml of 20 % HNO_3. If HNO_3 was used, the solution was evaporated with 1 ml of perchloric acid to remove nitric acid. The perchlorates were dissolved in 10 ml of 1.0N HCl.

An ion exchange column (Analytical Grade Dowex 50, 8 % cross linked, 200—400 mesh resin) was used to separate strontium from calsium, magnesium, lead, and barium.

The use of strontium-89 tracer made possible the rapid identification of strontium and the estimation of yields. If hydroxides were used, the magnesium hydroxide was found to carry down the strontium more quantitatively. The average runs gave yields from about 70—80 %. In many cases a yield as high as 90—95 % was obtained.

The strontium-bearing solution was evaporated down in a 50-ml silica glass beaker. The remainder — about 1 ml — was transferred into a 2-ml round bottom beaker, and dried by evaporation with 50 λ of conc. $HClO_4$. The remaining Sr^{89} was counted to determine the total yield.

The strontium was transferred to the mass spectrometer filament in 25 λ of 0.2N $HClO_4$.

The Nier-type, first order direction focusing Lamont mass spectrometer with a 60° sector field and 6 inch radius of curvature was used for this work and is described by Gast (1957).

The uncertainty in the mass ratios was calculated from the equation $\varepsilon = \sqrt{[\triangle \cdot \triangle]/n-1}$, where $\triangle$ is the deviation of each number from the mean, and n is the number of scans. The average of uncertainties in different mass ratios was found to be as follows (the average number of scans in parenthesis): $K^{41}/K^{39} = 0.6$ % (17), $Rb^{85}/Rb^{87} = 0.7$ % (36), $Sr^{87}/Sr^{86} = 0.8$ % (33), and $Sr^{88}/Sr^{86} = 0.7$ % (21). The number of runs was 8, 15, 12, and 12 respectively.

As stated by Gast (1957), the problem of contamination in the analysis of potassium and rubidium can be avoided by using a filament previously used for strontium, because much higher temperatures are used for several hours to form thermionic strontium ions, and alkalis are evaporated in this way from the filament.

Contamination, apparatus, and reagents.

Work with quantities of material in the microgram range requires special care especially where lead is concerned. Contamination is introduced from reagents, apparatus, and the laboratory air. The methods of reagent purification, sample protection, and glass cleaning are similar to those described by Tilton et al. (1955). Pyrex glassware was used throughout.

All glassware was cleaned by boiling for several hours first in 10 % sodium hydroxide and then in 50 % nitric acid. The platinum dishes and crucibles were in addition kept several hours in 6.2N hydrochloric acid. Finally all containers were rinsed several times with double-distilled water and sealed with a semi-adhesive plastic film. The seal was broken only when material had to be transferred. A teflon dust shield or large watch glass was used, when long periods of evaporation or digestion took place. Pipettes were cleaned by pumping hot 50 % HNO_3 through them several times.

Distilled water from a Barnstead still was redistilled through a condenser using quartz glass system or running first successively through a cation (Dowex 50, 8 %, 200—400 mesh) and an anion (Dowex 1, 8 %, 100—200 mesh) exchange resin and then carrying out the final distillation through a quartz glass column. For Rb—Sr- and K-work only water run through the exchange resin was used. The water was stored in polyethylene bottles.

Table I. Standard solutions

Enriched isotope	Supplier of salt	Source of spike	Concentration of solutions	Amounts used
Pb^{208}	ORNL	GL	19 μg/ml	1.00 ml
Pb^{206}			0.303 »	
Pb^{204}	ORNL	LGO	4.69 μg/ml	9.88 ml
U^{235}	ORNL	LGO	3.89 μg/ml	10.00 ml
U^{235}	ORNL	GL	5.843 μg/ml	1.00 ml
U^{238}	ORNL	GL	0.004615 μg/ml	
Th^{230}	ORNL	GL	7.2028 μg/ml	1.00 ml
Th^{232}			0.7618 »	
$K^{41}I$	ORNL	LGO	0.8412 μm/ml	10.00 ml
$K^{41}II$	ORNL	LGO	0.8456 μm/ml	10.00 ml
$Rb^{87}I$	Harwell	LGO	0.2634 μm/ml	10.00 ml
$Rb^{87}II$	ORNL	LGO	0.2287 μm/ml	10.00 ml
$Sr^{86}I$	ORNL	LGO	0.07446 μm/ml	10.00 ml
$Sr^{86}II$	ORNL	LGO	0.05420 μm/ml	10.00 ml

ORNL—Oak Ridge National Laboratories
Harwell—British Atomic Energy Research Establishment, Harwell
LGO—Lamont Geological Observatory, Columbia University
GL—Geophysical Laboratory, Carnegie Institution of Washington

Vacuum distilled nitric and perchloric acid, and hydrofluoric acid made by bubbling the gas into purified water were used. Ammonium hydroxide was prepared by passing the gas through a filter into the purified water.

Solid reagents (ammonium citrate, potassium cyanide, borax, ammonium nitrate, aluminium nitrate ennehydrate) were extracted with dithizone in chloroform to remove lead or with washed hexone to remove uranium and thorium. The hexone was shaken with 2 % nitric acid and washed twice with double distilled water.

A complete blank run on all reagents used in normal zircon treatment showed a total lead contamination of 0.18 micrograms. A blank three times higher was used for three of the samples, because of the contamination found in borax used for the fusion of these zircons. The isotopic composition of the blank, communicated by Patterson (1951) in his early meteorite work, was used. The contamination of uranium was found to be negligible in all reagents used during a single uranium and thorium run.

S p i k e s o l u t i o n s. One of the most serious sources of error is the spike concentration. Standard solutions (Table I) prepared in the Geophysical Laboratory and in the Lamont Geological Observatory were used in the course of this work.

T h e e q u a t i o n s u s e d. The derivation of the mathematical relationships used for the calculation of lead, uranium, and thorium is as follows:

Let: A = isotope A
B = isotope B
a = micrograms of isotope A per milliliter of spike solution

$$\left(\frac{B}{A}\right)_{\text{spiked run}} = \frac{B_{\text{spike}} + B_{\text{sample}}}{A_{\text{spike}} + A_{\text{sample}}} \tag{1}$$

Substituting $\left(\frac{B}{A}\right)_{\text{spike}}$ for R_{spike} and $\left(\frac{B}{A}\right)_{\text{sample}}$ for R_{sample}:

$$R_{\text{spiked run}} = \frac{[R_{\text{spike}} \times (A_{\text{spike}})] + [R_{\text{sample}} \times (A_{\text{sample}})]}{A_{\text{spike}} + A_{\text{sample}}}$$

If 1 ml of spike solution was used, one can write

$$R_{\text{spiked run}} = \frac{a\,(R_{\text{spike}}) + X\,(R_{\text{sample}})}{a + X}$$

Solving for X = micrograms of isotope A in sample aliquot:

$$X = \frac{a\,(R_{\text{spike}} - R_{\text{spiked run}})}{R_{\text{spiked run}} - R_{\text{sample}}} \tag{2}$$

If w = the weight in grams in aliquot of the sample being analyzed, and $R_{sample} = R_{unspiked\ run}$

$$ppm_A \text{ in sample} = \frac{a}{w}\left(\frac{R_{spike} - R_{spiked\ run}}{R_{spiked\ run} - R_{unspiked\ run}}\right) \tag{3}$$

The concentration of other lead isotopes can be calculated in the following way:

$$ppm_{208} = ppm_{206} \times (208/206) \times (\text{wt. ratio})$$

if the concentration of isotope 206 and the ratios (208/206), (207/206), and (204/206) are known respectively.

In the case of rubidium, potassium, and strontium the first equation was used as follows:

$$\left(\frac{B}{A}\right)_{spiked\ run} = \frac{c'_B\, M_{spike} + c_B\, M_{sample}}{c'_A\, M_{spike} + c_A\, M_{sample}} \tag{4}$$

where B/A is Rb^{85}/Rb^{87}, K^{41}/K^{39}, or Sr^{88}/Sr^{86}; M is micromoles of spike or sample (nonradiogenic); c′ is the atom percent of the isotopes of a spike material, and c is the natural isotope abundance. Solving the equation (4) for M_{sample} the micromoles of rubidium, potassium, and normal strontium can be obtained.

If the atomic ratio Sr^{87}/Sr^{86} in the spiked sample is known, the micromoles of the radiogenic strontium can be found by solving the same equation and supposing that the total Sr^{87} consists of three members: added spike-Sr^{87}, primary Sr^{87}, and the radiogenic Sr^{87}:

$$\frac{Sr^{87}}{Sr^{86}} = \frac{c'_{87}\, Sr^{87}_{spike} + c_{87}\, Sr^{87}_{primary} + Sr^{87}_{radiogenic}}{c'_{86}\, Sr^{86}_{spike} + c_{86}\, Sr^{86}_{sample}}$$

The present day ratios used are given in Table VIII (p. 41). The influence of contamination on the lead concentrations was eliminated by a blank correction.

A r g o n. The argon concentrations in the three mica samples mentioned in this paper were determined by Mr. Leon Long (Lamont Geological Observatory, Columbia University) by using the standard isotope dilution technique described, e.g., by Inghram (1954) and by Damon (1957).

ALPHA COUNTING

In making age determinations on zircon samples by the lead-alpha activity method described by Larsen and others (1952) and by the radiation damage method (Holland and Kulp, 1950; Hurley and Fairbairn, 1953;

Holland and Gottfried, 1955), the alpha-particle emission of each sample was determined in thick-source alpha scintillation counters, using an R. C. A. 5819 photomultiplier tube and housing, described by Kulp, Holland, and Volchok (1952). The counter was supplied with a register which recorded the accumulated counts every fifteen minutes.

The alpha activity of most samples was determined in two ways: (1) using natural grain size and (2) in pulverized form. If available, one hundred milligrams of sample were always introduced into a sample well of known size. The surface of the sample was checked in all cases under a binocular microscope. The fine-grained sample powder was settled in methyl alcohol. This method produced an even distribution of sample grains, but had the disadvantage that the sample dish was hard to clean. Several measurements were made on the same sample by using a more or less rough sample surface. Differences were not found outside the statistical counting error. Small spots of alcohol suspension of sample outside the sample well, as well as the small uncovered points in the well, were found to introduce a small error.

Twelve to eighteen hours counting time was used for all samples. The average of the probable statistical counting error of 150 measurements was 0.9 %, the average activity being 260 alphas/mg/hr. For some low level samples (about 8 alphas/mg/hr) the error was 7 %, and for some high activity materials (800—7000 alphas/mg/hr), such as monazites and uranium-thorium rich zircons, it was found to be 0.5 %.

The background count rate was usually measured once a week. Counters were checked every day by running a small samarskite sample, counted relatively to standard zircons.

Because of the relative nature of results obtained, the zircons were counted relative to two standard zircons: R-1932 from Iredell county, N. C., U. S. A. and Zr2—11 from Ceylon. The uranium and thorium content of the first of these zircons was determined by the fluorimetric method (Waring, U. S. G. S.) and of the second by the isotope dilution technique (Tilton, Geophysical Laboratory).

The traffic counter type automatic register was found to be invaluable for evaluating counting statistics and for detecting instrument failure.

MEASUREMENT OF THE UNIT-CELL DIMENSIONS

In order to study the relationship between radiation dosage (alpha-disintegrations per mg of sample) and disorder in zircon, it is desirable to measure the changes in c_0 (unit cell vertical axis) to approximately 1.5 parts per 10 000.

For preparing mounts, the sedimentation cell technique, based on the principle of simultaneous sedimentation of sample and silicon standard (Holland et al., 1955) was used. The quantity of silicon and sample was approximately 4 mg and 10 mg respectively. The zircon peaks commonly measured were those from the (200) and (112) planes, the reference peak in all measurements being Si-(111). Unfiltered copper radiation and a Norelco high-angle x-ray spectrometer were used.

The average deviation of calculated c_0 in 74 samples, each containing two slides and each slide 5—10 scans, was $\pm$.0005 Å, the value of c_0 ranging from 5.9802 Å to 6.0390 Å.

RESULTS

ANALYTICAL RESULTS

Isotopic composition of lead. As mentioned above, the distinctly anomalous character of the isotopic composition of lead extracted from the galenas in the veins within the rapakivi formation raised the problem of primary lead correction for the zircon leads from all the postkinematic potash granites of southern Finland. For this purpose three potash feldspars were treated, and the isotopic abundance data for these samples are shown in Table II. For each analysis the isotopic composition is reported in two ways: with the normal percentage composition in the first line and with $Pb^{204} = 1.000$ in the second line.

Table II. Isotopic abundance of common lead from three potash feldspars from rapakivi granite, Finland

No.	Locality	Source	204	206	207	208	
A—31	Tulisenlampi	Pegmatite vein	1.445	23.68	22.43	52.62	s
	Lemi		1.000	16.39	15.52	36.42	
A—42	Matalaluoto	»	1.446	24.04	22.25	52.43	s
	Lemi		1.000	16.63	15.39	36.26	
A—45	Pyhtää	Rapakivi ovoid	1.404	24.00	22.23	52.60	s
			1.000	17.09	15.83	37.46	

The results represent a potash feldspar from typical viborgite and potash feldspars from two rapakivi pegmatite veins. The three treated samples suggest an age approximately consistent with the age determined by the Rb^{87}—Sr^{87}, K^{40}—A^{40}, and by the isotopically controlled Th—Pb and U—Pb methods. This is in sharp contrast to the galena ore-lead veins in the rapakivi formation, which were found to have a modern lead isotope composition. Again there is evidence either of laboratory contamination (A-45) or of a real difference in the local environment of feldspar formation. It is worthy of note that the leads from pegmatite feldspar are similar but that the third lead from ovoid feldspar from ordinary rapakivi granite has a slightly different composition.

All the isotopic abundances of common lead extracted from seventeen Finnish galenas are listed in Table III.

Table III. Isotopic lead data for some Finnish galenas

No.	Locality	Mineral	204	206	207	208	
			I. Svecofennidic range, Finland				
Pb—175*	Pakila	Galena	1.484	23.35	22.79	52.38	g
			1.000	15.73	15.36	35.30	
Pb—232	Lemi	Galena	1.449	23.37	22.60	52.77	g
			1.000	16.13	15.60	36.42	
Pb—233	Attu	Galena	1.468	23.22	22.71	52.69	g
			1.000	15.82	15.47	35.89	
Pb—234	Aijala	Galena	1.453	23.18	22.72	52.65	g
			1.000	15.95	15.64	36.24	
Pb—235	Pernaja	Galena	1.463	23.19	22.69	52.66	g
			1.000	15.85	15.51	36.00	
Pb—238	Korsnäs	Galena	1.454	23.19	22.64	52.72	g
			1.000	15.95	15.57	36.26	
A—67	Orijärvi	Galena	1.471	23.24	22.74	52.78	s
			1.000	15.80	15.46	35.88	
Pb—237	Orijärvi	Galena	1.453	23.38	22.61	52.55	g
			1.000	16.09	15.56	36.17	
			II. Karelidic range, Finland				
Pb—228	Outokumpu	Galena	1.507	22.46	22.95	53.06	g
			1.000	14.90	15.23	35.21	
A—48	Outokumpu	Galena	1.510	22.41	22.97	53.29	s
			1.000	14.84	15.21	35.29	
A—58	Outokumpu	Galena	0.542	65.28	15.35	18.23	s
			1.000	120.44	28.32	33.64	
Pb—239	Panjavaara	Galena	1.467	23.09	22.66	52.78	g
			1.000	15.74	15.45	35.98	
Pb—240	Keihäsjoki	Metallic lead	1.456	23.18	22.86	52.50	g
			1.000	15.92	15.70	36.06	
Pb—247	Nunnanlahti	Galena	1.573	21.46	23.29	53.70	g
			1.000	13.64	14.81	34.14	
			III. Galena-bearing veins within the rapakivi formation				
Pb—174*	Säkkijärvi	Galena	1.332	25.51	21.04	52.12	g
			1.000	19.15	15.80	39.13	
Pb—230	Säkkijärvi	Galena	1.308	25.50	21.04	52.15	g
			1.000	19.50	16.09	39.87	
Pb—229	Luumäki	Galena	1.317	25.46	21.48	52.17	g
			1.000	19.33	16.31	39.61	
Pb—231	Sottunga	Galena	1.246	27.78	20.46	50.51	g
			1.000	22.30	16.42	40.54	

*Determined 1955 by George Bate
g gas run
s solid run

Ten of these samples were suggested by Vaasjoki, based on his comparison of the minor base metal contents of the same samples (Vaasjoki, 1956). Most of the samples were analyzed by the tetramethyl technique (g), although in some the surface emission method (s) was used.

The galenas from the Svecofennidic range are surprisingly coherent: in fact, they would appear to vary only slightly, if we discount the experimental error. The maximum variation is essentially represented in the same locality (i. e. Orijärvi). In sharp contrast to the simultaneity in the Svecofennidic province, the rapakivi granite veins are essentially modern lead with considerable radiogenic contamination. The situation with the galenas in the Karelidic province is much less simple.

Thirteen zircons and one monazite from Finnish rocks have been analyzed for their lead isotopic composition. In five cases the thorium and/or uranium determinations were completed using the isotope dilution technique. The isotopic composition of lead from these accessory minerals is given in Table IV.

The zircon analyses, completed for their uranium and thorium determinations, are given in the following chapter together with the calculated age data.

Table IV. Isotopic composition of lead from some radioactive minerals from Finland

No.	Source	Mineral	204	206	207	208
A—34	Oligoclase granite 658a-MH/Kaavi-54	Zircon	0.220 1.000	67.82 308.27	10.52 47.82	21.10 95.91
A—38	Oligoclase granite 209-AH/Kaavi-54	Zircon	0.333 1.000	67.96 204.08	11.68 35.08	19.68 59.10
A—95	Maarianvaara granite	Zircon	0.413 1.000	65.25 158.00	12.56 30.41	21.63 52.37
A—50	Oligoclase granite 658a-MH/Kaavi-54	Monazite	0.001 1.000	17.64 17640	2.13 2130	80.66 80660
A—61	Aulanko granodiorite	Zircon	0.194 1.000	74.90 386.08	11.00 56.70	13.46 69.38
A—71	Urjala granodiorite	Zircon	0.0535 1.000	79.87 1 492.90	9.65 180.37	10.18 190.28
A—84	Kalajoki granodiorite	Zircon	0.125 1.000	80.42 643.36	10.36 82.88	9.13 73.04
A—97	Sotkuma granite II-Sma	Zircon	0.063 1.000	79.44 1 261.0	13.76 218.4	6.23 98.89
A—46	Kokemäki granite	Zircon	0.119 1.000	74.28 624.20	8.98 75.46	16.20 136.13

Rubidium, strontium, and potassium. The data obtained in stable isotope dilution analyses of Finnish samples for rubidium, strontium, and potassium are shown in Table V.

Table V. Analytical data for micas from some intrusive rocks in southern Finland

Sample	Sample weight (gram)	Parts per million by weight			Sr^{87}/Rb^{87} (molar)	K(percent)
		Rb	SrN	SrR		
I—Bodom biotite	0.7095	953±21	8.6	5.90±.10	0.0219 ±.0006	6.07±.12
	1.2997	966±16	7.5	5.90±.10	0.0216 ±.0005	—
	0.050	957±20	—	—	—	—
II-Bodom biotite	0.7012	1 247±26	6.9	7.94±.13	0.0225 ±.0006	6.66±.14
29—KM biotite	1.2199	1 260±21	4.9	7.72±.14	0.0217 ±.0005	7.39±.25
	0.0546	1 205±25	—	—	—	—
I—Kitee muscovite	1.2997	—	1.7	5.29±.08	0.0250 ±.0006	—
	0.05005	747±13	—	—	—	8.72±.26
II—Kitee muscovite	1.4000	—	1.6	5.78±.07	0.0254 ±.0008	—
	0.0500	806±19	—	—	—	8.25±.25
658a—MH biotite	2.4999	350± 6	10.0	2.44±.05	0.0247 ±.0007	—
	1.2618	341± 7	—	—	—	—
95—GSF biotite	2.5000	632±13	9.0	4.45±.08	0.0249 ±.0007	—
	2.5334	642±15	—	—	—	—
Aulanko biotite	1.1011	696±14	5.54	5.03±.08	0.02554±.0008	7.17±.25

Some of these samples were analyzed in duplicate and triplicate and these separate determinations are shown in the table. For several samples 50 mg of sample were analyzed separately for potassium and rubidium.

Argon analyses. Three mica samples from Finland were run for argon by Mr. Leon Long, who kindly made them available for this paper. The A^{40}/K^{40} ratios for the Aulanko biotite, I-Bodom biotite, and II-Kitee muscovite are 0.1649, 0.1405, and 0.173 respectively.

Alpha activity data and lead determinations by using the spectrographic method for zircon samples are given in the following chapter together with the ages calculated for them.

COMPARISON OF AGE DATA

As pointed out by Larsen et al. (1952), the almost equal ionic radii of potassium and lead (1.33 Å and 1.32 Å) permit lead to concentrate in the

potassium minerals. The close packing of zircon and the great difference in ionic radii of zirconium (0.82 Å) and lead (1.32 Å) do not encourage the lead to intrude the zircon.

The possibility of determining the age of a mineral by the so called lead:alpha method has been studied by Larsen and his coworkers. This method yields an age based on measurements of the total lead content of a mineral and its alpha activity, assuming that all of the lead in the mineral is of radiogenic origin, that no lead has been lost, and the amount of radioactive elements has not been changed by processes other than radioactive decay. Hence if the mineral has been a subject of gain or loss of lead or radioactive elements, an anomalous lead:alpha age should indicate a discordant lead:uranium isotopic age.

The lead:alpha ages shown in Table VI were used as a basis for comparison of the ages determined on the basis of the radiation dosage-alpha activity ratio measurements.

Table VI. Lead:alpha ages for zircons

Sample	Activity α/mg/hr	Lead (ppm)	Age (m. y.)	Remarks
29—KM, viborgite	150±2	72, 70, 69	1 045	+ 200 mesh
29—KM, viborgite	194±2	72, 71, 68	830	— 200 mesh
292—FP, rapakivi granite	108±2	22, 19	444	
805—ML, rapakivi granite	150±2	100, 108	1 463	
35b—AS, dark rapakivi	109±2	54, 53, 54	1 100	
785—ML, rapakivi granite	124±2	60, 60	1 080	
672—ML, rapakivi granite	620±3	220, 215	805	
Aulanko granodiorite	195±2	75, 72, 71	857	
Urjala granodiorite	250±2	110, 112, 114	1 007	
658a—MH, oligoclase granite	300±2	110, 112, 114	855	
297—FP, quartz-porphyry	223±2	106, 104, 102	1 045	

The alpha activity was measured in the Department of Geology, Princeton University, by the equipment described above. The spectrographic lead determinations were kindly submitted to the author by C. L. Waring from the U. S. Geological Survey. Ages were computed according to the equation $t_1 = c\,Pb/\alpha$ and corrected for radioactive parent decay according to the formula, $t = t_1 - \frac{1}{2}(t_1^2 \times k)$, where α = alpha disintegrations per

milligram-hour, Pb is the lead concentration in parts per million, and t_1 and t are ages in millions of years. The constant c is taken as 2500, according to two uranium-thorium analyses made on the samples concerned. The correction made for parent decay is similar to that described by Keevil (1939). The constant k was chosen as 1.8×10^{-4}, according to uranium-thorium analyses mentioned above.

These ages were used when samples were selected for lead:uranium isotopic age work on the assumption that samples having high lead:alpha age would most probably yield a concordant isotopically controlled lead:uranium age. Inspection of the data shows that two of the eight zircon samples separated from the postkinematic potash granite group in southern Finland having a probable age of 1620 m. y. have given a lead:alpha age of 1460 m. y.: 805-ML, rapakivi granite (Table VI) and Bodom granite (Jaffe, Gottfried, and Waring, 1955). The probable true ages are based on intercomparisons of the isotopically controlled lead:uranium, potassium:argon, and rubidium:strontium ages listed in Table IX and X. Although too few samples are available, it is worth noting that zircon from Bodom granite having a high lead:alpha age yielded a concordant isotopic age.

Two lead:alpha ages made on the synkinematic intrusive group of Svecofennidic orogeny showed an average age of 900 m. y. Because of these probable low ages, only Pb^{207}—Pb^{206} isotopic age measurements were made. On the basis of these ages, as well as the ages based on Rb^{87}—Sr^{87} and K^{40}—A^{40} results, the most probable true age for this intrusive group is 1800 m. y.

Holland and Kulp (1950) suggested that the degree of damage in minerals caused by alpha-particle bombardment could be used as an independent age method, if the alpha activity were measured. Hurley and Fairbairn (1953) made a preliminary study, using x-ray diffraction and alpha activity measurements to establish the relation between radiation dosage and damage measured by changes in the (112) lattice spacing. Holland and Gottfried (1955) were able to determine the changes in c_0 (unit cell vertical axis) to changes in dosage (alphas/mg of sample) as a function of time by using gem quality zircons from Ceylon, which proved to have a concordant isotopic age as well as a wide range of alpha activity. Recently Fairbairn and Hurley (1957) have studied radiation damage in zircon from 42 rocks. The ratio dosage/activity has been found to correspond to other ages in some cases, but some have been obviously too low. These »ages» have been suggested as representing the date of a later metamorphism.

Age measurements based on this technique have been made on about 70 samples with an age of 70—2700 m. y. The data on the Finnish samples will be presented in this connection, together with the other ages based on isotopically controlled results.

The clean zircon samples were washed in an acid mixture of HCl + HNO_3 + H_2O (1 : 1 : 1). The material was kept in a beaker on a hot plate at a moderate temperature for about one hour to eliminate surface concentrations of radioactive material. Where available, 100 mg of sample was used. The alpha activity was measured first in the sample of natural grain size as far as accessory zircon was concerned, and afterwards in the pulverized sample. One interesting feature was found: in almost all cases the activity of zircon from the granitic or intermediate intrusive rocks was lower when the sample was crushed. For a hornblende gabbro from the Boulder batholith and for a dark, pyroxene-bearing variety of rapakivi rock the alpha activity of the pulverized zircon sample was found to be higher (2 % and 9 % respectively). Only in three cases was it found to be higher, where granitic rock was concerned: zircon from two rapakivi granites and Bodom granite. In all these cases the difference was within the limits of error: 1 % and 1½ % for rapakivi zircons and 1 % and 2 % for Bodom granite (-230 mesh and +100 mesh respectively). In 33 cases the activity was found to be lower, the average being about 10 %. For one quartz monzonite from Boulder batholith no difference was found. Several grain sizes were used for some big zircon crystals without any difference in activity. As far as this difference is real (and it is not supposed to be due to the escape of any radioactive material), this is supposed to indicate the geochemical differentiation of uranium, thorium, and zirconium during the crystallization process.

If possible, as sharply defined a fraction of zircon as possible was used. Most measured samples were between 200—325 mesh or minus 325 mesh fractions (mostly natural grain size). The activity measurements indicate in all cases more extensive difference between uncrushed and pulverized samples, when coarse-grained fractions were treated. Such inhomogeneity of the sample will usually be apparent in a broadening and often a skewing of the x-ray reflection peaks. A more homogeneous fraction will always give a sharper peak shape.

The results are listed in Table VII. The uncertainty arises from the error directly introduced with the statistical counting error and from that contributed by the radiation dosage measurement. A small correction to the radiation dosage/alpha activity ratio ($D_c/\bar{\alpha}$) is due to the U/Th ratio (Holland, 1955), and the final ratio indicating the age of the mineral is given in column $D_c/\bar{\alpha} \times \psi$. All calculations are based on the assumption that a given dosage has produced the same damage in specimens 1600—1800 million years in age, as shown by Ceylon zircons of 550 million years in age.

Inspection of the data shows three groups of ratios: 1. most of the zircons separated from the postkinematic potash granites suggest an age of about 800 m. y.; 2. the syntectonic intrusives are between 600 and 800 m. y.

Table VII. Radiation damage in zircons from some Finnish intrusive rocks. Dosage-damage relation the same as that shown by Ceylon zircon (550 m. y.)

Sample	a_o (Å)	c_o (Å)	Activity α/mg/hr	Radiation damage ratio		Comments
				$D_c/\bar{\alpha}$	$D_c/\bar{\alpha} \times \psi$	
29—KM, viborgite	6.6302±.0010	6.0231±.0012	150±2	1 113	910±26	60—200 mesh
29—KM, viborgite	6.6300±.0012	6.0225±.0014	194±2	849	790±32	-200 mesh
292—FP, rapakivi granite	6.6195±.0003	6.0077±.0008	117±2	921	851±37	
805—ML, rapakivi granite	6.6235±.0005	6.0132±.0005	150±1	859	799±18	carefully picked by hand
805—ML, rapakivi granite	6.6273±.0005	6.0187±.0007	244±2	615	584±15	cleaned only by Frantz
35b—AS, dark rapakivi ..	6.6167±.0004	6.0039±.0006	109±2	854	793±32	
785—ML, rapakivi granite	6.6294±.0005	6.0204±.0008	195±2	803	749±26	
Onas granite	6.6196±.0003	5.0099±.0003	120±2	968	890±20	-200 mesh
I—Bodom, Bodom granite	6.6400±.0014	6.0382±.0020	261±2	863	800±33	+100 mesh
II—Bodom, Bodom granite	6.6222±.0002	6.0151±.0004	281±2	485	466± 9	200—325 mesh
714—ML, rapakivi granite	6.6230±.0005	6.0289±.0012	384±2	493	473±14	
Quartz-porphyry, Åland ..	6.6196±.0008	6.0104±.0010	266±2	444	427±18	
297—FP, quartz-porphyry.	6.6196±.0016	6.0065±.0018	232±2	444	427±32	
299—FP, quartz-porphyry .	6.6228±.0016	6.0028±.0009	206±2	432	417±20	
Aulanko granodiorite	6.6306±.0003	6.0242±.0005	195±2	878	817±15	
Urjala granodiorite	6.6405±.0025	6.0390±.0017	250±2	912	844±30	
658a—MH, oligoclase granite	6.6328±.0005	6.0297±.0010	300±2	641	606±17	-230 mesh

old; 3. four samples belonging to the first-mentioned group indicate an age of about 450 m. y. The probable age of these two intrusive groups is 1620 m. y. and 1800 m. y. as mentioned above.

It can be seen that the radiation damage ratio — generally supposed to be the minimum age — equals about 800 m. y. for most of the samples from the rapakivi group. For all quartz-porphyry samples the measured activity is relatively higher and the damage ratio lower. For sample No. 714 activity as high as 384 alphas/mg/hr was obtained. In this case the measured age is also lower. Generally speaking, the impure fractions were found to be more radioactive, the impurities being unidentified inclusions, dark pigment, or inhomogenities such as more metamict grains etc. For sample No. 805

the »primary» fraction was treated and the age 584 m. y. obtained, while pure, carefully handpicked material indicated an age of 799 m. y.

Two zircon fractions from the Svecofennidic granodiorites were selected for radiation damage work: the Aulanko and Urjala granodiorites. Both intrusions occur on the same map sheet, Hämeenlinna, described by Simonen (1949). Both zircons gave an lead isotope age (Pb^{207}—Pb^{206}) of about 1850 m. y. The Rb^{87}—Sr^{87} age for Aulanko granodiorite is 1820 m. y. and the A^{40}—K^{40} age for the same rock 1725 m. y. The radiation damage age for Aulanko zircon is 817 m. y., whereas that for Urjala zircon is 844 m. y. The lead:alpha ages are 857 and 1007 m. y. respectively.

The discrepancies between the ages measured by the radiation damage method and the probable true age can be explained in several ways: inhomogeneity in the distribution of the radioelements, thermal annealing of the radiation damage, a later gain of radioactive elements, and the fact that two or more generations of zircon were measured. Holland (1956) has found that dry heating most zircons at a temperature of over 500°C produces appreciable annealing of radiation damage within one day. According to Ingerson (1955), Frondel has studied zircons hydrothermally. He has found that zircon recrystallizes at a temperature at least as low as 500°C under hydrothermal conditions without excess pressure, beyond that developed by heating. As pointed out by Holland (1956), the radiation damage measurements for determining the age of the last period of metamorphism of a region must be applied with caution. Only if all of the previously produced damage is removed, may the radiation damage age be considered to be a measure of the time which has elapsed since the last thermal metamorphism.

In Fig. 1 c_0 has been plotted as a function of the present alpha activity and of the radiation dosage calculated for 1 600 million years. It can be seen that the points fall on two straight lines. The line A goes through the point $c_0 = 5.9797$ (the length of the c-axis of the undamaged Ceylon zircon), and it is the locus of points of most of the common rapakivi zircons. The line B does not go through the point $c_0 = 5.9797$ and it consists of three zircons from the rapakivi quartz-porphyries and of two other zircons. Two points fall outside the lines. They belong to the samples 29-KM and II/805-ML. The sample 29-KM represents a coarse-grained (60—100 mesh) accessory zircon from rapakivi granite. The material is inhomogenous, as described above. The sample II/805-ML represents a rather impure fraction, which consists yet almost completely of zircon.

The recurrence of the age of 800 million years in several samples from different places (line A) suggests a comprehensive action. The uranium-thorium isotopic analyses exclude the possibility of relatively too high activity due to some secondary processes. If the assumption that a given dosage has produced the same damage in specimens of 1600 m. y. in age as shown by

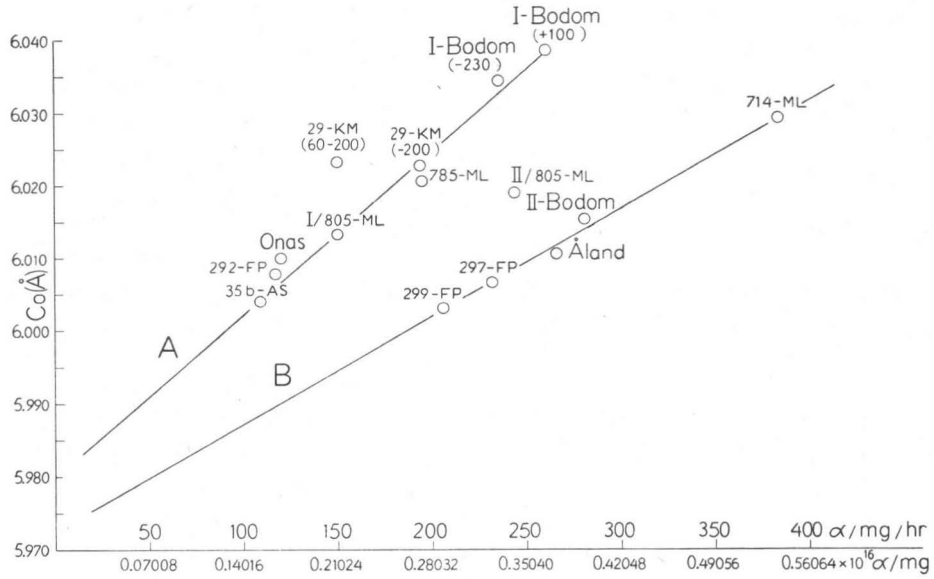

Fig. 1. c_0 for zircons from the postkinematic granites as a function of the present α-activity and of the total α-dosage calculated for 1 600 million years.

Ceylon zircons of 550 million years in age is valid, the thermal annealing of radiation damage could be suspected as the cause of the low age.

Worthy of note is the high activity of the samples falling on the line B and the impure nature of the samples (especially the pigment described above).

Thirteen zircons and one monazite from Finnish rocks have been analyzed for their lead isotope composition. In five cases the uranium and thorium determinations have also been completed, using the isotope dilution technique.

The isotopic compositions of lead used for the primary lead correction are as follows:

1. Potash granite group of the age of about 1 600 m. y.

The average isotopic composition of lead extracted from feldspars from one viborgite ovoid and from pegmatite veins in rapakivi granite given in Table II:

204	206	207	208
1.000	16.70	15.75	36.68

2. Intrusive group in Svecofennidic range of the age of about 1 800 m. y.

The average isotopic composition of galena lead from six samples from southern Finland mentioned in Table III (Attu, Aijala, Orijärvi (A-67), Pernaja, Pakila, Korsnäs):

204	206	207	208
1.000	15.88	15.51	35.96

3. Intrusives in Karelidic range of the age of about 1 800 m. y.

The isotopic composition of Panjavaara galena mentioned in Table III.

The isotopic abundances and decay constants used in the age calculations are given in Table VIII.

Table VIII. Decay constants and isotopic abundances of radioactive isotopes used in this study

Isotope	Decay constants	Isotopic abundance
U^{238}	$\lambda_\alpha = 1.54 \times 10^{-10}yr^{-1}$	0.9929 g/g U
U^{235}	$\lambda_\alpha = 9.72 \times 10^{-10}yr^{-1}$	0.0071 g/g U
Th^{232}	$\lambda_\alpha = 4.99 \times 10^{-11}yr^{-1}$	1 g/g Th
Rb^{87}	$\lambda_\beta = 1.39 \times 10^{-11}yr^{-1}$	0.283 g/g Rb
K^{40}	$*\lambda_\beta = 4.8 \times 10^{-10}yr^{-1}$	1.22×10^{-4} g/g K
	$**\lambda_e = 0.581 \times 10^{-10}yr^{-1}$	

* Based on 28 β/g/sec. (Gast et al., 1957)
** —»— 3.39±.15 γ/g/sec. (Wetherill, 1957)

The results of the isotopically controlled zircon analyses are given in Table IX. Uncertainties in isotopic composition are mean deviations of 15—25 sets of ratios.

Table IX. Age data for zircons from postkinematic potash granites from southern Finland

Bodom zircon

I—Bodom

U (p. p. m.): 557.16 ± 5.57, Th (p. p. m.): 174.0 ± 1.74, Pb (p. p. m.): 163.4

Lead data	204	206	207	208
Isotopic composition (moles)	0.057±0.003	100	10.90±0.10	11.29±0.11
Concentration (p. p. m.)	0.0756	133.5	14.62	15.22
Primary lead (measured)	1.00	16.70	15.57	36.68
Radiogenic lead (p. p. m.)	0	132.2	13.43	12.34

	U^{238}—Pb^{206}	U^{235}—Pb^{207}	Pb^{207}—Pb^{206}	Th^{232}—Pb^{208}
Age in million years	1 590±30	1 620±20	1 675±30	1 530±35

Onas zircon

U (p. p. m.): 291.7 ± 6.1, Pb (p. p. m.): 88.64

Lead data	204	206	207	208
Isotopic composition (moles)	0.260±0.003	100	13.66±0.12	19.82±0.44
Concentration (p. p. m.)	0.170	66.15	9.08	13.24
Primary lead (measured)	1.00	16.70	15.57	36.68
Radiogenic lead (p. p. m.)	0	63.28	6.39	6.88

	U^{238}—Pb^{206}	U^{235}—Pb^{207}	Pb^{207}—Pb^{206}
Age in million years	1 460±60	1 545±35	1 670±42

Rapakivi zircon

805—ML/Iitti—54

U (p. p. m.): 492.04 ± 36.9, Pb (p. p. m.): 143.5

Lead data	204	206	207	208
Isotopic composition (moles)	0.1609±0.0064	100	12.31±0.10	17.82±0.15
Concentration (p. p. m.)	0.175	109.91	13.60	19.78
Primary lead (measured)	1.00	16.70	15.57	36.68
Radiogenic lead (p. p. m.)	0	106.95	10.83	13.23

	U^{238}—Pb^{206}	U^{235}—Pb^{207}	Pb^{207}—Pb^{206}
Age in million years	1 470±140	1 550±95	1 670±30

Rapakivi zircon

29—KM/Virojoki—55

U (p. p. m.): 383.36 ± 4.98, Pb (p. p. m.): 114.05

Lead data	204	206	207	208
Isotopic composition (moles)	0.1769±0.0076	100	12.36±0.32	17.77±0.45
Concentration (p. p. m.)	0.1532	87.37	10.85	15.68
Primary lead (measured)	1.00	16.70	15.57	36.68
Radiogenic lead (p. p. m.)	0	84.79	8.43	9.95

	U^{238}—Pb^{206}	U^{235}—Pb^{207}	Pb^{207}—Pb^{206}
Age in million years	1 490±90	1 550±55	1 637±68

Rapakivi zircon
35b—AS/Lemi—54
U (p. p. m.): 228.4 ± 2.3, Pb (p. p. m.): 50.95

Lead data	204	206	207	208
Isotopic composition (moles)	0.0604±0.003	100	10.82±0.11	17.93±0.18
Concentration (p. p. m.)	0.02348	39.31	4.274	7.12
Primary lead (measured)	1.00	16.70	15.57	36.68
Radiogenic lead (p. p. m.)	0	38.91	3.903	6.24

	U^{238}—Pb^{206}	U^{235}—Pb^{207}	Pb^{207}—Pb^{206}	Th^{232}—Pb^{208}
Age in million years	1 165±30	1 350±20	1 650±30	1 050±100

It may be seen from Table IX that the results of these determinations fit into the general pattern $\frac{208}{232} < \frac{206}{238} < \frac{207}{235} < \frac{207}{206}$, which is developed from measurements on various geological units. In Fig. 2 the mole ratios Pb^{206}/U^{238} (D_1/P_1) and Pb^{207}/U^{235} (D_2/P_2) are plotted on a »concordia» diagram (Wetherill, 1956a, 1956b). The ratios lie on a straight line going through the points 120 and 1 670 on the curve marked »concordia», i. e. the locus

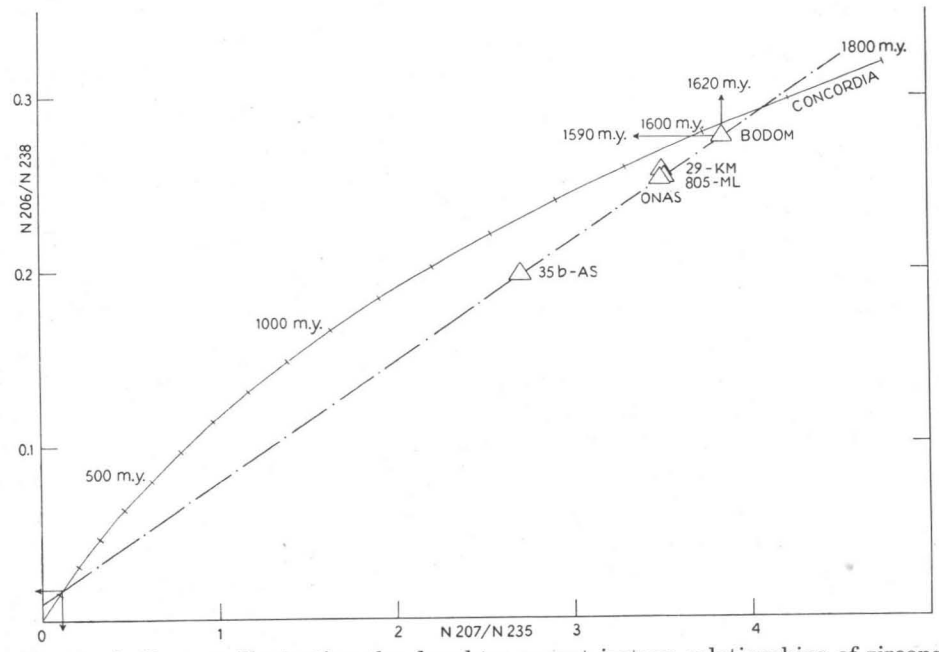

Fig. 2. A diagram illustrating the daughter-parent isotope relationships of zircons from the postkinematic granites in southern Finland.

of all points, such that $t_{U\text{-}235} = t_{U\text{-}238}$. Wetherill has demonstrated this kind of regularity and the possible use of discordant uranium:lead ages to determine some aspects of the history of known geological units. He has been able to give a mathematical proof of the graphical procedure for calculation of discordant uranium:lead ages, resulting from multiple episodes of uranium-lead fractionation or from the presence of primary radiogenic lead or a constant percentage loss of intermediate daughter products. This graphical procedure is found useful in giving information on the possible histories of a given geological area. It is not possible, however, to infer the history uniquely, if only the discordant ages are given. According to him, this kind of linearity could be explained by a chemical removal of lead or uranium. An alternate explanation has been given by Ahrens (1955a and b) and by Russell and Ahrens (1957). Instead of chemical removal of lead, these authors suggest that physical control of lead escape rather than chemical may be one of the possible mechanisms for discordance. According to them, the ejection of members of the decay series into microfissures by the recoil energy from alpha-particle emission, followed by the removal of these members or their descendants by diffusion, leaching or any other processes, provides a possible mechanism for the loss of lead and could explain the fractionation of lead isotopes in the direction mentioned above. As pointed out by former authors, more detailed studies of single deposits are needed to explain the most proper mechanism. Worth noting is that Tilton (1956) has been able to show that radiogenic Pb^{208} in zircon and sphene is more susceptible to removal than are radiogenic Pb^{206} and Pb^{207}. In a later paper Tilton, Davis, Wetherill, and Aldrich (1957) have thoroughly discussed the present state of explanations for the age discordances in zircons.

Aldrich, Davis, Tilton, and Wetherill (1956) have pointed out that an agreement of rubidium-strontium and potassium-argon ages is potentially a better measure of absolute age than are the two U—Pb ages. This belief depends on the fact that potassium and rubidium decay to daughters which are different elements, while U^{235} and U^{238} decay to two isotopes of the same element which might be expected to behave in a more similar manner during processes affecting their host mineral. On the other hand, the half-lives of U^{235} and U^{238} differ so greatly from each other that any measurable change of lead or uranium will destroy the agreement. Actually the best laboratories nowadays are equipped for Pb—U and Pb—Th as well as for A—K and Rb—Sr work. This makes possible an internal checking of the ages of a sample, if the necessary minerals are available.

Some facts can be given for the postkinematic potash granite group in southern Finland: 1. the average of six Pb^{207}—Pb^{206} ages indicates an age of 1 655 $\pm$ 15 m. y.; 2. the solid straight line along the experimental points on the D/P diagram cuts the curve »concordia» at the T_1-point 1 670 m. y.;

3. the only slightly discordant I-Bodom age pattern gives a most probable age of 1 605 m. y. (average of U^{235}—Pb^{207} and U^{238}—Pb^{206} ages); 4. Rb^{87}—Sr^{87} ages for rapakivi biotite 29-KM/Virojoki-55, II-Bodom, and I-Bodom biotites (Table X) are $1\,610 \pm 40$ m. y., $1\,605 \pm 40$ m. y., and $1\,555 \pm 20$ m. y. respectively; 5. a K^{40}—A^{40} age for I-Bodom biotite is $1\,550 \pm 60$ m. y. (Table X).

The studied zircons have given discordant ages so frequently that a question arises as to which, if any, of these ages are correct. However, the agreement of Rb^{87}—Sr^{87} and U—Pb ages excludes the possibility of an accidental agreement of the zircon isotope ages of the Bodom sample. The Rb^{87}—Sr^{87} age of 29-KM rapakivi as well as the excellent agreement of six Pb^{207}—Pb^{206} ages unite the other samples in a coherent intrusive group of the same age, within the indicated limits of experimental error. From the slight discordance of I-Bodom zircon ages, as well as the about 3 % higher Pb^{207}—Pb^{206} ages and the location of the point T_1 on the D/P diagram, the question arises of whether the agreement of Rb^{87}—Sr^{87} and both Pb—U ages is not rather accidental. Since the I-Bodom zircon isotope ages have to be regarded as slightly discordant, 1 605 m. y. is to be regarded as the probable minimum age, the probable maximum age being the Pb^{207}—Pb^{206} age = 1 655 m. y. Instrumental considerations alone limit the accuracies of the ages to about three percent. The U^{235}—Pb^{207} age of Bodom granite, 1 620 m. y., is probably the most nearly correct age.

One fact still stands out regarding these age comparisons. Zircon from the I-Bodom sample gave the most concordant isotope age. Biotite separated from the same rock gave a definitely low age of $1\,555 \pm 20$ m. y. This age was checked twice afterwards, and the low result is not due to analytical error. As noted before, three samples treated were selected because of the primary nature of biotite: II-Bodom, Uusikaupunki trondhjemite, and Kaavi oligoclase granite. In this case the biotite separated from the biotite granite (II-Bodom) gave the more correct age (1 605 m. y.).

Daughter product diffusion as a possible cause of the discrepant age measurements has been examined in detail and developed in graphical form by Nicolaysen (1957). As can be seen, the age obtained from the graphs is always higher than the Pb^{207}—Pb^{206} age, usually found to be slightly higher than the most probable true age. The five intersections obtained for the age patterns given in Table IX suggest numbers such as (parameters D/a^2 in parenthesis) 1 686 m. y. (0.5×10^{-12} year^{-1}), 1 710 m. y. (3.3×10^{-12} year^{-1}), 1 715 m. y. (3.3×10^{-12} year^{-1}), 1 668 m. y. (1.7×10^{-12} year^{-1}), and 1 790 m. y. (19.8×10^{-12} year^{-1}) respectively. The average of these numbers is 6 % higher than the most probable true age for that granite group.

The fore-going discussion has not considered the other Finnish intrusive rocks studied. Table X shows the comparative Pb^{207}—Pb^{206}, Rb^{87}—Sr^{87},

Table X. Comparative age results for zircons and associated micas

Sample	Location	Rock	Mineral	Age in millions of years			Remarks
				Pb^{207}—Pb^{206}	Rb^{87}—Sr^{87}	K^{40}—A^{40}	
			Postkinematic granites				
35b—AS	Lemi	Dark rapakivi	Zircon	1 650±30	—	—	U—Pb and Th—Pb ages are given in Table IX
805—ML	Iitti	Rapakivi granite	Zircon	1 670±30	—	—	
29—KM	Virojoki	Viborgite	Zircon/ biotite	1 637±68	1 610±40	—	
Onas	Porvoo	Potash granite	Zircon	1 670±42	—	—	»
I—Bodom	Espoo	Potash granite (even grained)	Zircon/ biotite	1 675±30	1 555±20	1 550±60	»
II—Bodom	Espoo	Potash granite (porphyritic)	Biotite	—	1 605±40	—	
29—GSF	Kokemäki	Potash granite	Zircon	1 630±37	—	—	
			Svecofennidic range				
Aulanko	Hämeenlinna	Granodiorite	Zircon/ biotite	1 845±30	1 820±45	1 725±70	
Urjala	Urjala	Granodiorite	Zircon	1 860±34	—	—	
83—GSF	Kalajoki	Qtz. diorite	Zircon	1 790±40	—	—	
95—GSF	Uusikaupunki	Trondhjemite	Biotite	—	1 760±45	—	
			Karelidic range				
658a—MH	Kaavi	Oligoclase granite	Zircon/ biotite	1 845±45	1 780±47	—	
			Monazite	<2 000			
209—AH	Kaavi	Oligoclase granite	Zircon	1 735±80	—	—	
Maarian-vaara	Kaavi	Microcline granite	Zircon	1 750±75	—	—	
I—Kitee	Kitee	Pegmatite granite	Muscovite	—	1 780±42	—	
II—Kitee	Kitee	Pegmatite granite	Muscovite	—	1 810±53	1 775±50	
Outokumpu	Kuusjärvi	Pegmatite vein	Muscovite	—	1 845±45	—	
II—Sotkuma	Polvijärvi	Gneissose granite	Zircon/ biotite	2 530±38	1 804±40	—	

and K^{40}—A^{40} ages of intrusive rocks from southern Finland. As can be seen for the Svecofennidic intrusives, the two Rb^{87}—Sr^{87} ages imply an age of 1 790 $\pm$ 45 million years. The average of the three Pb^{207}—Pb^{206} ages is 1 832 $\pm$ 35 million years or about 2.4 % higher.

The excellent agreement of the Rb^{87}—Sr^{87} and K^{40}—A^{40} ages of the Karelidic intrusive group indicates a most probable true age of 1 800 $\pm$ 45 million years. The average of the Pb^{207}—Pb^{206} ages is about 1 800 $\pm$ 60 million years This number does not include the high monazite age and anomalous low zircon age of 1 735 m. y.

As can be seen from Table IV, the 658a-monazite sample is extremely poor in primary lead. Unfortunately the mass spectrometry run was not good enough to separate the small 207-peak from the big 208- and 206-peaks because of the poor resolution due to the preparation of the sample. Only an estimation of less than 2 000 m. y. can be given. The age of the sample 209—AH is presumed to be too low because of analytical reasons due to the extremely small sample size.

Because of the low Larsen-ages, only Pb^{207}—Pb^{206} ages were made for most of the zircons. Results have shown that Pb^{207}—Pb^{206} ages are slightly high but good enough to give reasonable information for most geological problems, especially as it is possible to get a statistical result easily in most cases.

The accuracies stated do not include errors arising from incorrect decay-constants or from the geological history. As stated by Wetherill et al. (1956), the half-life for U^{235} (7.13×10^8 years) with a probable error $\pm 0.16 \times 10^8$ years introduces additional errors of $\pm$ 35 m.y. in the $U^{235} - Pb^{207}$ age and $\pm$ 90 m.y. in the $Pb^{207} - Pb^{206}$ age of a 1 600 m.y. mineral.

GEOLOGIC INTERPRETATION OF RESULTS

SOME GENERAL ASPECTS OF THE GEOLOGICAL PROBLEMS INVOLVED

One task of geological endeavour is to go downward in the stratigraphical sequence. Together with petrogenesis, the chronological subvision of rocks of lithological complexes has always been one of the final purposes of geological survey in all continents.

Many of the recent efforts to obtain more information regarding conditions in pre-Cambrian time have been concerned with the cycles of sedimentation defined by their relation to tectonic movements. The classification is mainly based on the relations of the supracrustal rocks to epochs of diastrophism and to the intrusive rocks which have been intruded. The rule of »tracing from place to place», if continuous schist belts or certain granite types are available, is the commonest way used for working out the stratigraphic succession.

Contact phenomena play the most important role in the correlation of rocks lying near together. However, many difficulties arise especially in pre-Cambrian geology, because often uncertainties are multiplied and in any case only relative ages can be obtained. For instance, in the Fennoscandian shield a plastic Cambrian clay lies on the undeformed rapakivi granite, and Paleozoic fossiles are found in sandstone veins cutting it. These observations show that rapakivi granite is pre-Cambrian. Thus the relative age is known, but still the space of time between these two geological formations is not cleared up. Numbers like 200 million years as well as 1 000 million years have been presented.

All field geologists are agreed on an interesting and confusing phenomenon: a crystallizing intrusive massif can cut its own already solidified outer part by sending pegmatitic or aplitic veins from the inner part through its outer part. As far as relative age is concerned, the vein is younger than the host rock, but the absolute age difference is practically zero. In such a case the value of the measured absolute age of the penetrating vein may be higher than that of the intrusive host rock, the result being quite correct within the specified limits of experimental error notified.

One major difficulty arises from the correlation of large scale orogenic belts which do not come into contact.

In one of his last publications Sederholm (1934) says (p. 96): »The classification of pre-Cambric rocks will also need to be greatly changed. On the basis of the evidence which has hitherto been obtained in the field, we are already able to assert that the pre-Cambric far from being an inconsiderable appendix to the succession of fossiliferous rocks, must contain the greatest part of the geological record. It contains more cycles of sedimentation, separated by cycles of diastrophism than all later stratal successions, and, therefore also is thicker and more varying in sedimentation than those, although we do not yet know more than a part of them.» In the same paper he announces (p. 95): »The cooperation of geo-physical research is also very necessary for the age determinations among the oldest rocks. . . . But there ought to be collaboration. Especially it is necessary that the material investigated by a geo-chemist, or geo-physicist is collected by geologists possessing a thorough knowledge of the region in question.»

The fact that the Earth is the scene of processes that continually change its isotope composition has given us a geological clock. New methods developed in the last fifteen years have extended the possibilities to the more common rock-forming minerals. The result of this evolution is that neither analytical requirements nor the limited amounts of useful minerals are restricting factors any more. A confining factor is the material itself.

If the age of synkinematic intrusive groups belonging to two orogenic cycles is known, the space of time between the climaxes of these two orogenies should be measurable within the limits of error indicated. However, there is still the problem of the true nature of intrusives selected for investigation. As shown, for example, by the well studied Appalachian mountain chain, the orogenic belt may be a composite one, including two or more sub-cycles, each with its own plutonic activity. The rejuvenation of material by a new orogenic activity is perhaps the most confusing factor within the old pre-Cambrian formations. To what extent the old schist belts are intruded by later orogenic intrusives is of the greatest importance to the interpretation of all age results. The close connection of orogenic and epeirogenetic movements will obscure the boundaries of two orogenic belts, and more difficulties will arise as far as age results are concerned.

The minerals mostly used for age determinations are from acidic or intermediate rocks. A single crystallization of material is provided. However, the current hypotheses as to the genesis of these groups of rock are the most confusing ones. One school of thought postulates that most granite is formed from magma. Another school of thought argues that most granite is of metasomatic origin. Even though these problems have to be regarded as unsolved, and the interpretation of age determinations made on minerals separated from granitic rocks may be considered to be uncertain, the comparative studies of some geochronological data so far obtained have shown

that it is possible, to some extent, to assess the degree of confidence that may be placed in the respective methods employed. The question is still open as to what is the amount of primary daughter element included in the material granitized during later events.

The area where samples for this research were collected is presented in Figure 3. The necessary background of general geological knowledge was provided by the studies of several Finnish geologists. Some excellent summaries are given by Sederholm (1927, 1932, 1934), Eskola (1927, 1941), Wahl (1936), Saksela (1953), Simonen (1953), and Väyrynen (1954). Additional information on the detailed geological settings of each sample was provided by the various geologists who actively mapped each area.

The pre-Cambrian formations in Finland are separated into two systems as stated above — an upper Karelidic system, which includes undoubtedly original sediments lying on an old base, and a lower Svecofennidic system, which consists of original sediments, represented by graded bedded graywacke slates, arkose sandstones, conglomerates, tuffs, and lavas, and of granitic, gneissic, and schistose rocks, among which are seldom found beds of quartzite and ophiolite like intrusives. The stratigraphy and sedimentation of the Svecofennidic supracrustal rocks in southwestern Finland have recently been described by Simonen (1953). The tectonic characteristics of the Svecofennian mountain chain have been demonstrated, e. g., by Saksela (1953). The sedimentary characteristics and the stratigraphy of the Karelidic formations in Finland have been summarized in several papers by Väyrynen (1933, 1954).

Both cycles of sedimentation exhibit close similarities on one hand and some remarkable differences on the other. Each schist belt has its own graywacke slates. In certain places the graded bedding and other primary sedimentary features are well preserved. Simonen (1953) has described an eugeosynclinal accumulation in the Tampere schist belt at least 8 km thick. Many graphite-bearing schists similar to recent muds occur in various sections of both schist belts. Volcanic activity has produced numerous lava and tuff beds in both areas. The limestones and calcareous sediments in the Karelidic mountain chain are, however, dolomites, whereas in the Svecofennidic area they are calcites. The rarity of true quartzites in the Svecofennidic belt is remarkable. The huge quartzite beds in the Karelidic chain are the most characteristic formations indicating a tectonically stable platform.

In the Karelidic area the underlying old basement can be easily found, but it has not yet been traced definitely in the Svecofennidic area. Some geologists (Saksela and Heiskanen, 1952; Huhma, Salli, and Matisto, 1952) have given, however, evidences of possible pre-Bothnian rocks in southern Finland.

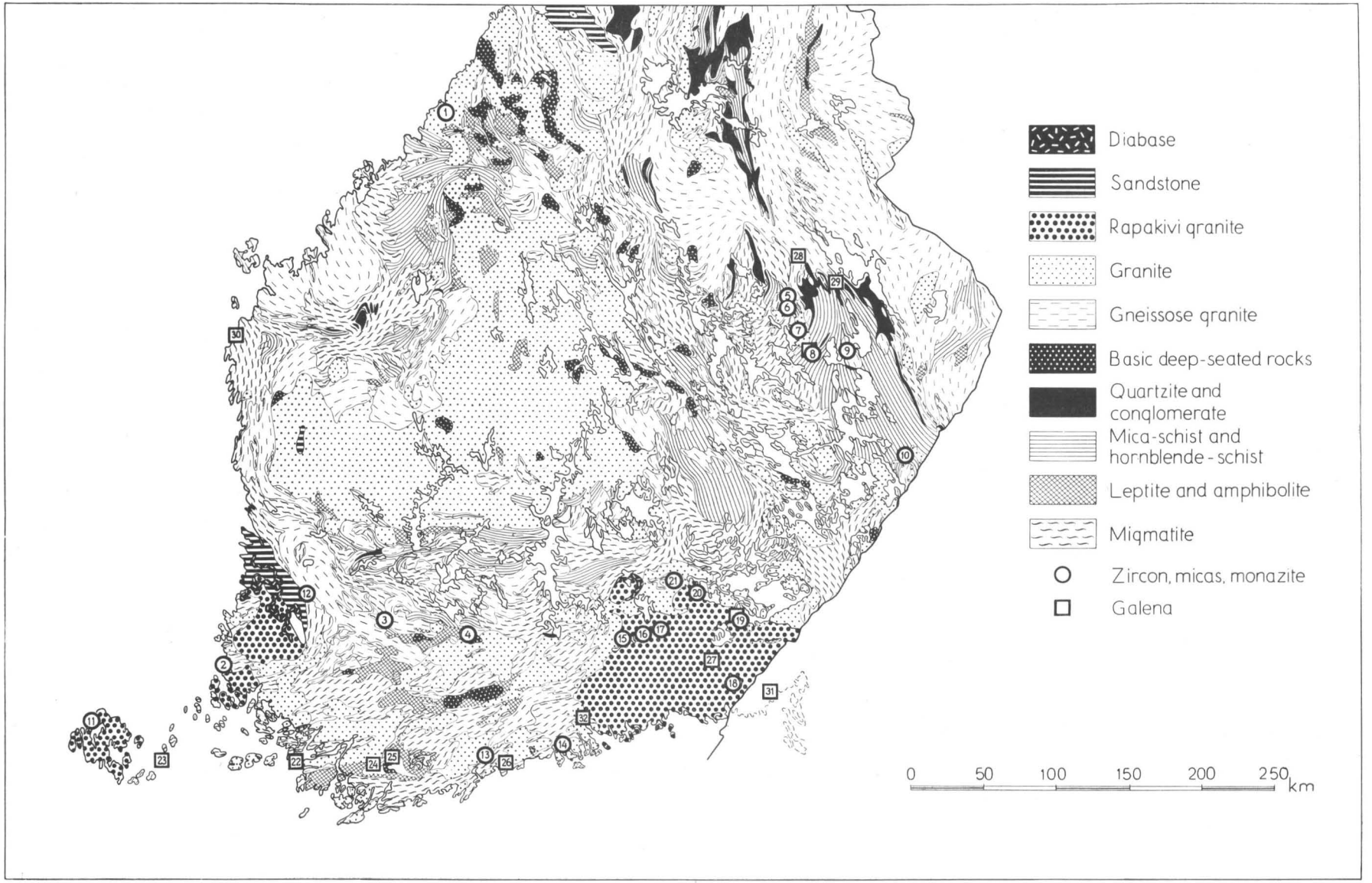

Fig. 3. Location of the samples investigated. (Map by Erkki Mikkola.)

1. Kalajoki quartz-diorite (83-GSF)
2. Uusikaupunki trondhjemite (95-GSF)
3. Urjala granodiorite
4. Aulanko granodiorite
5. Kaavi granite (209-AH)
6. Kaavi granite (658a-MH)
7. Maarianvaara granite
8. Outokumpu
9. Sotkuma granite
10. Kitee granite
11. Quartz-porphyry, Åland
12. Kokemäki granite (29-GSF)
13. Bodom granite
14. Onas granite
15. Rapakivi granite (805-ML)
16. » » (785-ML)
» » (714-ML)
17. » » (672-ML)
18. » » (29-KM)
19. » » (35b-AS)
Pb-232, Lemi
20. Rapakivi granite (292-FP)
Quartz-porphyry (299-FP)
21. » » (297-FP)
22. Pb-233, Attu
23. Pb-231, Sottunga
24. Pb-234, Aijala
25. Pb-237, Orijärvi
A-67, Orijärvi
26. Pb-175, Pakila
27. Pb-229, Luumäki
28. Pb-239, Panjavaara
29. Pb-247, Nunnanlahti
30. Pb-238, Korsnäs
31. Pb-174, Säkkijärvi
Pb-230, Säkkijärvi
32. Pb-235, Pernaja

As already pointed out by Hess about twenty years ago, the relationship of serpentines to island arcs adds weight to the hypothesis that island arcs represent an early stage in alpine-type mountain building. According to him (Hess, 1955), the serpentinized peridotites probably are intruded during the first great deformation of a mountain belt and do not recur in subsequent deformations of the same belt. Hence they enable one to date orogenies by dating serpentines and to follow the axis of an ancient orogenic belt in some cases for thousands of miles. In world wide research it has been found (Hess, 1955) that in the oldest rocks serpentinized peridotites do not occur in belts but are ubiquitous throughout the whole terrain. These ultramafic ophiolites are rare in the Svecofennidic schist belt but can be traced throughout the Karelidic orogenic belt from Lake Ladoga to Lapland.

The present correlations between distant pre-Cambrian successions and the local stratigraphic problems have been recently discussed, e. g., by Saksela (1953) and Simonen (1953). The orogenic position of the intrusive rocks where present samples are from has been discussed in the chapter »Sources». The consequences of the assignment of the ages given in the previous chapters to the orogenic classification will be valid if the following assumptions are fulfilled:

a) Proper corrections have been made for the initial concentration of daughter elements in each radioactive decay scheme used for age calculations. This premise includes the possibility of a rejuvenation of old rock massives as well as the granitization of older material.

b) There have been no gains or losses of parent or daughter or intermediate members in the radioactive decay scheme during the time since the formation of the system. As far as practical age determinations for the pre-Cambrian minerals are concerned, the limits of error of 30—50 million years allow alterations which will not change the final age by more than that amount.

c) Only one generation of each mineral is used and the age of the mineral is the age of the rock. The suitability of biotite from trondhjemitic rocks has been pointed out before, provided that the biotite is rich enough in the parent radio element.

d) The intrusives of each orogenic belt belong to that orogenic cycle. The age of the post-tectonic intrusives very closely defines the end of the cycle or sub-cycle of activity to which they belong, but they may be very much younger than the deep-seated surroundings where they are found.

If these assumptions are fulfilled, some conclusions can be drawn from the results obtained.

1. The differences in the ages of the minerals from the rocks intruding the Svecofennidic orogenic belt and from the rocks intruding the Karelidic orogenic belt are within the limits of the indicated error of the age measurements.

2. As far as the geologic interpretation of the nature of these intrusive masses is correct, the ages of the synkinematic intrusives suggest a climax of the orogenic movements about 1 800 million years ago.

3. The close agreement in the age of the members of the rapakivi granite and that of the Bodom, Onas, and Kokemäki granite confirms the assumption that all these potash granites (III and IV granite group in Sederholm's classification) might represent a coherent intrusive group. According to the age results gathered within this material, they intruded 1 620 $\pm$ 20 million years ago.

4. The galenas from the Svecofennidic range are surprisingly coherent; in fact they would appear to vary only slightly outside experimental error. The maximum variation is chiefly represented in the same locality i. e. Orijärvi. The Lemi sample is slightly more radiogenic and comes near the contact with the rapakivi granite. The isotopic composition of these samples would suggest an average model age of 1 740 million years, assuming the lead was derived from average source rock in the crust. This is essentially only slightly lower than the age of the Svecofennidic rocks determined by Rb^{87}—Sr^{87}, K^{40}—A^{40}, and U—Pb methods.

This relationship presents another case such as that observed in the Keewatin rocks of the Canadian shield and the pre-Cambrian ones of the Black Hills.

5. In sharp contrast with the simultaneity in the Svecofennidic province and with the isotopic composition of lead in the rapakivi feldspars, the veins in the rapakivi granite contain essentially modern lead with considerable radiogenic contamination. The sample from Sottunga has the highest radiogenic component but does not exceed in radio-lead many samples of lead in the Missisippi Valley deposit. Of the greatest interest are the determinations on the sulfur S^{32}/S^{34} ratios from these same samples made by Ault (1957). Except Sottunga the S^{32}/S^{34} ratio is 21.59.

6. The situation with the galenas in the Karelidic province is much less simple. The samples from Panjavaara (1 800 million years) and from Keihäsjoki (1 900 million years) appear to have the isotopic composition of lead corresponding to that found in the Svecofennian rocks. The Outokumpu and Nunnanlahti galenas (2 260 and 2 750 million years respectively) show distinctly different and less radiogenic leads. In addition, A-58, which was found in a shear zone outside the Outokumpu ore deposit, appears to represent a remobilized older radiogenic material. Both the 207/206 ratio and the Pb^{208} content suggest a source in the range of 2 100 million years in age. The other older samples (e. g., A-247) may also represent mobilization of older lead which was subsequently in a low U/Pb environment. The only old age obtained for zircon was that found for a sample from the Sotkuma gneiss cupole. This suggests an age of 2 530 $\pm$ 38 million years

for the old basement underlying the material sedimented and folded during the mountain forming processes about 1 800 million years ago. The Rb^{87}—Sr^{87} age for mica from the same sample (Gast, personal communication) and K^{40}—A^{40} age for potash feldspar from the same outcrop (Kahma, 1956) suggest ages of 1 804 and 1 860 million years respectively.

Sotkuma sample represents a case where the age for zircon is higher than that for mica. This phenomenon has been found also in some other localities.

The history in the Karelidic province appears considerably more complicated and is more analogous to the situation in Southern Rhodesia and South Africa as far as galenas are concerned.

COMPARISON OF DATA WITH PUBLISHED DETERMINATIONS

The Archean formations of southern Finland belong to Fennoscandia, that is, to the territory which comprises great parts of Scandinavia, all of Finland, Russian Karelia, and the Kola Peninsula. The region where pre-Cambrian rocks prevail is usually called the Fennoscandian or Baltic shield. This area is a part of the Fennosarmatia.

As pointed out by v. Bubnoff (1952), the pre-Cambrian formations of the Fennoscandian shield have their continuation further to the south, 187 m deep near Leningrad, about 700 m deep near Daugavpils, and 1 570 m deep near Moscow, where cordierite-sillimanite-garnet gneisses are met. According to magnetometric measurements east of Lake Peipus, the Svecofennidic belt turns towards the south or south-east. As delineated by Archangelski (Bubnoff, 1952), the regional N-S strikes of the magnetic anomalies of the West Sarmatian provinces indicate that the Karelidic mountain range extends further to the south and has its continuation in the N-S striking Ukrainian block and in the block of Woronesh as well.

With the object of correlating the ages of these formations with each other and the age determinations presented in this research with other ages made for formations in southern Finland and for corresponding formations in Scandinavia and Russian Karelia, a comparison with the other published data is here set out.

Lokka (1928, 1950) has worked on the chemistry of radioactive minerals found in Finland. He has presented several chemical ages from 1 081 up to 1 207 million years for the minerals from southern Finland and 1 192 million years for a monazite from sand from Ivalojoki River in North Lapland, Finland. Worth noting are the distinctly higher ages of 1 505 to 1 600 million years for the wiikites from Salla, Karelidic mountain range.

Holmes, Shillibeer, and Wilson (1955) have presented a K^{40}—A^{40} age of 1 810 $\pm$ 120 million years for a potash feldspar from the Varala pegma-

tite, about 30 km southeast of Tampere in southwestern Finland. This age agrees with the age of the intrusives presented in this paper from the same Bothnian schist belt belonging to the Svecofennidic orogenic belt.

The most remarkable work on the ages for Finnish rocks has been done by Kahma in collaboration with the age work group of the Department of Physics, University of Toronto. According to a short abstract (Kahma, 1956), the results of 28 K^{40}—A^{40} age determinations made for ten Finnish granites and gneisses indicate an existence of two intrusive groups of the ages of 1 400—1 450 ± 100 million years and of 1 770—1 860 ± 120 million years. In the second Nordic Geological Winter Meeting in Oslo, January 5—7, 1956, Kahma discussed the relation of the intrusive groups to the Karelidic and Svecofennidic orogenic belts. The younger age group includes Bodom granite, a potash granite of Pernaja, Aavasaksa granite, and a quartz-porphyry from Aitolahti. An interesting result is the average of 1 830 million years from seven measurements for Maarianvaara granite. The corresponding average age of Kaavi and Maarianvaara granites presented in this paper is 1 790 million years. The age of the Sotkuma gneiss granite discussed above is 1 860 ± 120 million years. Worth noting is that potash feldspar was used for all these measurements and, in order to compensate for some loss in argon, a smaller value for the branching ratio was used (Wilson, personal communication). These data also indicate an age of slightly more than 1 800 million years for the Karelidic orogeny. The age of 1 400—1 450 million years for Bodom granite has to be considered low, as well as the age of 1 500 ± 120 million years presented for Bodom granite by Stevens and Shillibeer (1956). The concordant isotope age for Bodom zircon presented in this paper indicates an age of 1 620 million years.

Some interesting chemical ages for Svecofennian pegmatites in the neighborhood of Stockholm are given by Parwell and Wickman (1954): uraninite from Digelskär (Pb/U—Th) = 1 770 million years, fergusonite from Ytterby (Pb/U—Th) = 1 790 million years, and hjelmite from Nya Karvet (Pb/U—Th) = 1 810 million years. These ages agree with the ages presented in this paper for the Svecofennian intrusives.

One remarkable age study done on the Fennoscandian minerals was the careful dating of lepidolite from Varuträsk, Sweden by Aldrich, Wetherill, and Davis (1956) using the Rb^{87}—Sr^{87} method. According to this work, the age of this mineral and hence the age of a Svecofennian pegmatite is 1 790 ± 40 million years. In 1939 Wahl made a lead isotope analysis on a uraninite from the Varuträsk pegmatite. The age calculated from these results is about 1 700 m. y. (Quensel, 1957).

Most of the work on the dating of formations of the Fennoscandian shield area has been done in Russia. These measurements cover the eastern part of the shield. Seventeen old chemical ages for fresh uraninites from

N. E. Karelia give a most probable age of 1 765 ± 10 million years for the Marealbian pegmatites (Holmes, 1948). These samples are as follows: Sinaja Pala: 1 760 million years (Nenadkevich, 1926): and five ages for uraninites from Khito Island: 1 760, 1 795, 1 790, 1 795, and 1 780 million years (Khlopin and Vladimirova, 1938, analysts: V. M. Permjakov, I. E. Starik, and L. V. Komlev). Eight ages were made by the same authors for the uraninites from the Chornaya Salma: 1 710, 1 700, 1 715, 1 750, 1 800, 1 805, 1 730, and 1 790 million years, and for the Samoilovich Vein: 1 805, 1 800, and 1 775 million years.

Further, two monazite ages are known from the Chornaya Salma and from the Samoilovich Vein pegmatites (Labuntsov, 1935). Two different techniques were used, and the ages are 1 800 and 1 860 million years, and 2 090 and 2 015 million years. One monazite age for the Chornaya Salma pegmatite gives 1 720 million years (Labuntsov, 1935). All these data indicate an age only 30—40 million years lower than the average age of 1 800 ± 40 million years presented in this paper for the Karelidic pegmatite minerals.

Much valuable work on Fennoscandian minerals has been done by Gerling and his co-workers. Several papers have appeared, dealing with the application of the decay of K^{40} to A^{40}. Polkanov (1954) has presented some interesting correlations between the K^{40}—A^{40} ages obtained and the geological formations in Russian Karelia. Table 8 summarizes some comparisons as follows (only ages for micas notified):

Jotnium-hoglandium	Granites of IV group	Upper proterozoicum	1 500 million years
Granites in Karelia	Granites of III group	Lower proterozoicum	1 560 million years
Bothnian granites	Granites of II group	Upper Archean	1 800 million years
Svionian granites	Granites of I group	Lower Archean	1 850 million years

Of special interest are the single age determinations for rapakivi granites: rapakivi from Valamo = 1 560 million years, rapakivi from Pitkäranta = 1 485 million years, and rapakivi from Viipuri = 1 570 million years. All these determinations were made on potash feldspar, and 20 % was added to the ages to correct for argon leakage. For the Viipuri rapakivi one biotite age is given: 1 500 million years. These ages presented for the rapakivi formation are about 100 million years lower than the 1 620 million years given in this paper. This difference is probably partly due to error in the value of the decay constant used by Gerling and others.

In a later paper Gerling, Sastsenko, and Jermolin (1957) have given several K^{40}—A^{40} ages for the pre-Cambrian minerals from Russian Karelia. Values $\lambda_K = 6.02 \cdot 10^{-11}$ year $^{-1}$ and $\lambda_\beta = 4.9 \cdot 10^{-10}$ year $^{-1}$ are used and for the calculation of the concentration of K^{40} $1.2 \cdot 10^{-4}$ is given. About 100 ages are given in that publication. The pre-Cambrian ages are summarized in a table (Table 4, p. 26). As pointed out by the authors, all the ages for feldspar are too low. The average ages made on micas are as follows:

I magmatic series	Oligoclase granites, biotite gneisses, and micas from the White Sea region	1 860 million years
II magmatic series	Micas from the pegmatite veins, White Sea region	1 800 million years
III magmatic series	Micas from the Ladoga formation	1 680 million years
	Micas from the pegmatites around Lake Ladoga.	1 700 million years
	Micas from the schists, Central Karelia	1 730 million years
	Micas from the schists, South Karelia	1 560 million years
IV magmatic series	Micas from pegmatites	1 510 million years

The authors have stated that, according to these results, the age difference between the oldest formations (White Sea region) and the youngest granites of IV group is 350 million years.

Some interesting correlations can be made between these results and those given in this paper. The age of the first and second group given by Gerling and others is 1 800—1 860 million years. The number 1860 is an average age derived from ages which vary from 1 660 to 2 030 million years. Hence the ages of these series agree with the ages given in this paper for the Svecofennidic and Karelidic intrusives using several methods. The age limits of 1 560—1 700 million years for the third series include ages for micas from the Ladoga formation and from the pegmatites around Lake Ladoga. According to Hackman (1933), the pegmatite granite massif of Kitee is the youngest intrusive found in the area north of Ladoga except the rapakivi formation. The two Rb^{87}—Sr^{87} ages and one K^{40}—A^{40} age for this intrusive presented in this paper indicate ages of $1\,780 \pm 42$, $1\,810 \pm 53$, and $1\,775 \pm 50$ million years respectively. The age 1 510 million years given for rapakivi formation is about 100 million years lower than that given in this paper.

Wahl (1936) has suggested four periods of mountain building in Fennoscandia: 1) the Svecofennian in Sweden and Finland; 2) the Gothic in SW Sweden and SE Norway; 3) the Karelian in E and N Finland and in

Russian Karelia; 4) the Caledonian in Norway and N Sweden. Backlund has united the Gothic and the Karelian foldings to a single one, the Gotokarelide, and has added a Norwegosamidic folding which comprises the Archean of the Kola peninsula, the North Finnish granulite formation, and the South Varanger formation. Gerling et al. (1957) have stated that, according to K^{40}—A^{40} data presented by them, a time space of at least 700—800 million years has to be considered between the Gothic and the Karelidic orogenic cycles. If the ages presented in this paper are correlated with the age results (about 1 000 million years) given by Wickman (personal communication) for the pegmatites near Gothenburg and Karlstad (Kil) which intrude the Gothic gneiss belt, the existence of two separate orogenic belts seems more probable.

The age of 1 860 $\pm$ 130 million years for the oldest rocks in the White Sea region is considered to represent the time elapsed since the metamorphism of these formations (Gerling et al., 1957). The present material includes one sample from the old basement underlying the Karelidic sediments. As stated before (p. 53), three ages are available for this sample from the Sotkuma dome: 2 530 m. y. (zircon), 1 804 m. y. (biotite), and 1 860 m. y. (potash feldspar). The discordance between the high Pb^{207}—Pb^{206} age for zircon and the low Rb^{87}—Sr^{87} age for mica and K^{40}—A^{40} age for feldspar could be explained by the regeneration hypothesis presented by Eskola (1949). Some support for the high zircon age is furnished by the common lead determinations on some galenas from the Karelidic occurences.

According to the age model presented by Bate, Gast, Kulp, and Miller (1957), the galenas from the Svecofennidic region analyzed within the limits of this study give an average model age of 1 740 million years with a spread from 1 650 to 1 820 million years. Two galenas from the Karelidic range give ages of 1 800 and 1 900 million years. The Outokumpu galenas are 2 260 million years old and the age of one galena from a skarn rock suggests that the material is as old as 2 750 million years. Some other common lead ages are presented from the Fennoscandian shield. Bate, Gast, Kulp, and Miller (1957) have presented two similar ages suggesting rather close agreement with the data presented in this paper: Falun, Sweden = 1 850 million years and Långban, Sweden = 1 820 million years. Vinogradov (1955) has reviewed a large amount of isotope data from various parts of the globe. These data include some common lead ages from Fennoscandia. As pointed out by Bate et al. (1957), in a number of cases there appears to be some evidence that Russian data are not strictly comparable in that the 204 abundance is high, relative to that obtained by Lamont and Toronto. However, some correlations can be made. Two galena analyses given for material from Orijärvi, Finland, suggest ages of about 1 800 and 1 510 million years. Data given in this paper for Orijärvi galenas are 1 770 and 1 650

million years. Two galenas from Viena (Kemi, Karelia and the White Sea region, Karelia) suggest ages of 1 500 and 2 040 million years respectively. Two ages are given for galenas from Kinon-Koznala, Finland: 2 180 and 2 030 million years. Finally one age as high as 2 280 million years is given for a galena from Karelo-Finnish S.S.R. The geologic setting of those samples is not available, and the interpretation is difficult. One analysis for lead from galena from Sala, Sweden, is given, suggesting an age of approximately 1 615 million years. Nier's earlier work included an isotopic analysis of native lead from Långban, Sweden. This analysis suggests an age of about 1 740 million years.

Some interesting correlations can be made between the K^{40}—A^{40} ages given for the Ukrainian shield. Gerling (1955) has published 33 ages for these formations. Ten of them suggest an age of 1 985—1 915 million years. These were made on muscovite, and on micas from granodiorite, from biotite gneiss, and from graywackes. Sixteen ages of 1 890—1 800 million years are given. Most of these ages are for muscovites, but the data include measurements on mica from red granite, and green mica, and biotite from granite. One age of 1 890 million years is given for a chrome mica. The rest of the ages are between 1 470—1 760 million years. Only two of the numbers are less than 1 500 million years.

POSSIBLE IMPLICATIONS OF THE AGE MEASUREMENTS

Besides areal stratigraphic correlations, several other problems are closely connected with age determinations. Dating has thrown new light on puzzles like the mutual relationship of orogenic belts in different continents, the origin of various ore-deposits, the basement subdivision of old shield areas, the age of very ancient signs of the beginning of life, and the age of the Earth.

Von Bubnoff (1956, p. 73) has reviewed the present knowledge of signs of ancient life. In a recent paper Goldich, Nier, and Baadgaad (1957) have discussed the widespread graphitic beds in northern Wisconsin and in Michigan. They have stated that these beds indicate that plant life developed prior to the 1 700 million years orogeny, and that correlations with the Biwabik and Gunflint formations, on geologic grounds, date the algal and other fossil remains of these formations as older than 1.7 billion years.

One interesting correlation can be drawn also with the indications of past life that have been detected in the pre-Cambrian rocks of Africa. As summarized by Holmes (1951), those remains can hardly have an age of less than 1 500 million years.

The analogy of the Karelidic schists to those of the Huronian series has been pointed out in many connections (e. g., Sederholm, 1927, and Holmes,

1951). The age determinations for the intrusives penetrating these two formations suggest a similar age of 1 700—1 800 million years. Probable signs of ancient life have been seen also in the Karelidic and in the Svecofennidic schist belts. These have been discussed in several papers (Sederholm, 1899, 1913, 1934; Eskola, 1931, 1932, 1954, 1956b; Rankama, 1948a, 1948b, 1950, 1954a, 1954b; Metzger, 1924; Marmo, 1956).

Metamorphism has done much to obliterate the oldest fossils. Evidence of two forms has been presented in the Finnish Archean: Corycium enigmaticum and Carelozoon jatulicum. The nature of both of these is most uncertain, and their identification as fossils is still open to criticism. The most remarkable carbon-bearing formation in Fennoscandian Archean is the so-called shungite. Metzger (1924, p. 65) has explained it as follows: »Die Schungitbildung ist also durchaus durch organische Entstehungsweise erklärt. Sie bildet einen faziellen Faktor in der gesamten Pelitformation, die einen Sedimentabsatz in einem flachen, offenen Meere darstellt. Diese Gesteinsserie bildet den Abschluss der jatulischen Sedimentbildungen nach oben.» Some boulders indicate shungite beds up to 80 cm thick. The shungite was first described by Inostranzeff in Shunga near Onega lake. The most common carbon-bearing schist in the Karelidic and in the Svecofennidic schist formation is the so-called black-schist. These well known sediments are also rich in sulfides.

Wickman (1952) first suggested the possibility of investigating the early history of life by means of the C^{12}/C^{13} ratios. The work done on Fennoscandian material is summarized by Rankama (1954b).

The mineral ages presented in this paper show a minimum age of 1 800 million years for the shungites as well as for the other possible signs of ancient life in the Karelidic and in the Svecofennidic orogenic belts.

Thode and his co-workers first studied sulfur isotopes and found that no fractionation occurred between sulfates and sulfides until 800 million years ago, the isotopic content S^{32}/S^{34} being the same value as that for meteoritic and igneous rock sulfur. They have suggested that life involving the sulfur isotopes became abundant about 800 million years ago, when isotopic exchange between sulfate and sulfide began and became more and more rapid. Kulp, Ault, and Feely (1956) found that larger differences existed in the S^{32}/S^{34} ratios of sulfate and sulfide of sedimentary rocks at least 1 000 million years ago than had been previously concluded.

Within the compass of this research, pyrite samples were separated from some well known black schists, from the Outokumpu ore, and from some intrusive rocks from the U. S. A. and from Finland. All black schists were from the Karelidic schist belt, and they represent sapropelic beds rich in sulfur and carbon in a sedimentary column. The samples were analyzed for sulfur isotopes partly at Hamilton College, McMaster University by

Thode and his co-workers, and partly at the Lamont Geological Observatory, Columbia University by Ault. The results obtained for five pyrites from the Karelidic black schists and for one pyrite from a pyrite ore, Karhunsaari, Liperi (Saksela, 1933a) were as follows (Thode, 1956, personal communication, and Ault, 1957):

Sample	Geographic location	S^{32}/S^{34}	Analyst
Black schist	Kulkevainen, Polvijärvi, Finland	22.454	Thode
»	Kultakallio, » »	22.389	Thode
»	Marjakuiva, » »	22.398	Thode
Black schist within the ore	Outokumpu, Kuusjärvi, »	22.32	Ault
Black schist 100 m from the ore	Outokumpu, » »	22.38	Ault
Pyrite ore	Karhunsaari, Liperi, »	22.65	Ault

As can be seen from the results, an average fractionation of one percent can be found from the meteoritic value of 22.21 given by Ault (1957). The mineral age of 1 800 million years for the intrusives which penetrate these schists has to be considered as the lower limit for the fractionation of sulfur isotopes, assuming that no fractionation had taken place during the later metamorphic processes.

As stated before, the galenas from the Svecofennidic range form a surprisingly coherent age group of about 1 740 million years (Attu 1 880 m. y., Pakila 1 730 m. y., Lemi 1 650 m. y., Aijala 1 820 m. y., Korsnäs 1 772 m. y., Orijärvi 1 770 and 1 650 m. y., Pernaja 1 790 m. y.), assuming that the lead was derived from average source rock in the crust. The average S^{32}/S^{34} ratio (Ault, 1957) for six of those galenas was 22.21 (Attu 22.22, Pakila 22.18, Lemi 21.94, Aijala 22.27, Korsnäs 22.26, Orijärvi 22.21, Pernaja 22.14). The average does not include the anomalous low value for the Lemi sample from a place 500 m from the contact of the rapakivi granite. The lead data presented in this paper and the sulfur data (S^{32}/S^{34} for the rapakivi galenas: 21.59) presented by Ault (1957) combine to suggest that the galena mineralization in the rapakivi granite represents a case like S. E. Missouri district, Ellenville, N. Y., and Bellevue, Idaho, where lead is anomalous but the sulfur is very heavy.

The two common lead ages made on Outokumpu galena suggest an age of 2 260 million years, consequently about 450 million years higher than that for the intrusive rocks around the ore body. The S^{32}/S^{34} ratio of the ore minerals of this ore body shows only little fractionation from the meteoritic value:

Sample description	S^{32}/S^{34}	Analyst
Pyrite, coarse grained	22.237	Thode
Same, more magnetic fraction	22.235	Thode
Pyrite, large cobalt bearing crystals	22.181	Thode
Pyrite, fine grained, chalcopyrite bearing	22.212	Thode
Pyrite, associated with anhydrite, 2 m above footwall	22.17	Ault
Chalcopyrite from the same sample	22.09	Ault
Pyrite, cubanite ore	22.21	Ault
Pyrite, disseminated ore	22.14	Ault
Galena, chalcopyrite vein	22.23	Ault
Average	22.19	

The small amount of data available suggests 1. that the Outokumpu ore is older than 1 845 million years, i. e. older than the granite-pegmatite intrusion; 2. that it is characterized by a S^{32}/S^{34} ratio close to that of meteoritic and magmatic sulfur; 3. that the isotopic composition of lead isolated from the galena differs from that of the galenas in the Svecofennidic range, suggesting that the ore material was separated from its source about 2 200 million years ago. More evidence could be collected as regards the history of the ore by dating the cogenetic mica using the K^{40}—A^{40} method, by dating the magnetite by the He-method, and by investigating the isotopic composition of lead from the sulfides like pyrite within the ore body.

Most of the ages presented in this paper fall near 1 600—1 800 million years. Similar ages have been found in various other parts of the Earth, e. g. from the Satpura cycle (1 625 $\pm$ 75 million years) in India (Aswathanarayana, 1956), Athabasca province (1 800—1 900 million years) in Canada (Wilson et al., 1956), granitic intrusions in central Minnesota (late Huronian-early Keweenawan, 1 700 million years) in the U. S. A. (Goldich et al., 1957), the Ukrainian shield (1 800—1 900 million years; Gerling, 1955), and various ages from widely separated localities in Africa (Holmes and Cahen, 1955).

The material studied here includes one sample from the basement underlying the Karelidic schists and some galenas indicating a long interval between 2 300—2 700 million years. These ages might be derived from an older material existing before the orogenic cycle which includes the intrusives of 1 800 million years ago. This time interval includes two orogenic cycles seen in other continents, e. g., Yellowknife province (2 200—2 400 million years) and Keewatin (2 600 million years) in Canada (Wilson et al., 1956), the eastern Ghats cycle (2 300 million years) and Dharwar cycle in India (Aswathanarayana, 1956), 2 000—2 300 million years and 2 650 million years presented from various parts of Africa (Holmes and Cahen, 1955), Beartooth Range, Montana (2 750 million years), Bighorn Mountains, Wyom-

ing (2 460—2 800 million years), N. W. Angle region of Minnesota (2 510 million years) in the U. S. A. (Gast et al., 1957), Winnipeg River-Johnston Lake area of Manitoba (2 650 million years) in Canada (e. g., Gast et al., 1957), the Ukrainian shield (2 600—2 700 million years; Vinogradov, 1956), and four localities in western Australia (2 700 million years; Jeffery, 1956). Ages of 1 000 million years, such as southern Scandinavia in Fennoscandia, the Delhi cycle in India (955 million years), parts of Cape province and parts of South-West Africa (1 025 million years) have not been found in the present work.

SUMMARY

1. A set of samples of zircon, monazite, muscovite, biotite, feldspar, and galena from various intrusive groups of the orogenic belts in Finland has been subjected to isotopic and chemical as well as to radiation damage measurements in order to obtain apparent ages from the isotopic ratios Pb^{206}/U^{238}, Pb^{207}/U^{235}, Pb^{208}/Th^{232}, Pb^{207}/Pb^{206}, Rb^{87}/Sr^{87}, and K^{40}/A^{40}.

2. Comparisons of Rb^{87}—Sr^{87}, K^{40}—A^{40}, and Pb^{207}/Pb^{206} ages with concordant, isotopically controlled U—Pb and Th—Pb ages made on zircons from the postkinematic potash granite group in southern Finland indicate a most probable true age of 1 620 million years. The average of six Pb^{207}—Pb^{206} ages is 2.2 % higher. The Rb^{87}—Sr^{87} ages for biotite from biotite granite (II-Bodom) and from biotite-rich viborgite agree at 1 605—1 610 million years, but for biotite from an even grained hornblende-biotite granite (I-Bodom) it is certainly lower (1 555 million years).

3. The zircon results for four out of five samples completed for the thorium and/or uranium determination are internally discordant and indicate that these zircons have been disturbed since their crystallization. The samples are from the undeformed postkinematic potash granites including the rapakivi granite. No relation between the disagreement and the amount of common lead has been found. The radiation damage measurements made on these samples as well as for the concordant Bodom zircon indicate »ages» from 700 to 900 million years.

4. The results obtained by the Pb^{207}—Pb^{206} method on six zircons and one monazite from the intrusives of the Svecofennidic and the Karelidic orogenic belts show close agreement with age determinations by the Rb^{87}—Sr^{87} and K^{40}—A^{40} methods on micas from the same rocks, suggesting an average age slightly higher than the most probable age of 1 790 million years in the Svecofennidic and Karelidic zones.

5. One of the currently important problems in geochronometry is the disagreement of the zircon ages with those obtained for micas and potash feldspar from some gneisses. The data recorded for zircon, biotite, and potash feldspar from the Sotkuma gneiss cupole suggest that the Pb^{207}—Pb^{206} age for zircon is 700 million years higher than the Rb^{87}—Sr^{87} age for biotite and the K^{40}—A^{40} age for potash feldspar.

6. As now dated, the Pb^{207}—Pb^{206} age for the Sotkuma zircon suggests an age of about 2 500 million years for the old basement underlying the Karelidic sediments. Some support for this result is furnished by the common lead determinations on some galenas from the Karelidic occurences.

7. The average common lead age of eight galenas from the Svecofennidic region was found to be 1 740 million years. The spread of these measurements is from 1 650 to 1 820 million years. These data differ greatly from those connected with rapakivi galenas. The history in the Karelidic province appears considerably more complicated.

8. The results obtained place a lower limit of 1 800 million years on the age of formations where probable signs of ancient life have been found.

9. According to present material, the fractionation of the sulfur isotopes found in the Karelidic schist belt must have taken place at least 1 800 million years ago, providing that no fractionation has occurred during later metamorphic events.

10. A noteworthy agreement exists between the geochronological data presented in this paper and those recorded in the literature for several parts of Fennoscandia as well as for those presented for the Ukrainian pre-Cambrian shield.

11. The presented age data do not seem to substantiate all descriptive geologic interpretations. No final conclusions as to the geologic significance of the present material can be reached from the presented age data alone, but they are contributory evidence for the mineral ages collected by other workers.

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