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Rapakivi granites and other postorogenic rocks in Finland: their age and the lead isotopic composition of certain associated galena mineralizations

by Matti Vaasjoki

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RAPAKIVI GRANITES AND OTHER POSTOROGENIC ROCKS IN FINLAND: THEIR AGE AND THE LEAD ISOTOPIC COMPOSITION OF CERTAIN ASSOCIATED GALENA MINERALIZATIONS

BY
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With 10 figures and 12 tables in the text and two appendixes

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The ages of the postorogenic rocks of southern Finland have been investigated by utilizing the U—Pb method on zircons. There exists one group of postorgenic intrusions in the area of the Åland Islands with ages ranging from 1840 to 1800 Ma. The rapakivi granites are younger, the Wiborg massif being 1700—1650, the Åland massif 1670, the Vehmaa massif 1590 and the Laitila massif 1570 Ma old. In the cases of the Wiborg, Åland and Laitila massives, younger intrusive phases of 1640 Ma, 1620 Ma and 1540 Ma, respectively, have been met with. The porphyry dikes are in every case of the same age as the main parts of the massifs. Anorthosites spatially associated with the Laitila and Wiborg massifs register a temporal relationship as well.

Within the Wiborg massif there seem to have occurred three major magmatic phases: at 1700—1660 Ma, at 1650 Ma and at 1640 Ma.

The results of the 76 U—Pb determinations suggest that zircons formed from residual magmatic solutions are liable to produce unusually discordant age patterns. This property is attributed to an abnormally high initial lead content, which may have resulted in an initial distortion of the zircon lattice.

The Pb—Pb determinations made from galena occurrences in the rapakivi massifs suggest that they are generated by their host rock. The results also suggest that the radiogeneity of the leads contained in the vein deposits increases as the temperature and the pressure prevailing during ore formation decrease.

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INTRODUCTION

Throughout the works of J. J. Sederholm, a fourfold classification of the granites of southern Finland appears. The gneissose granites are the oldest and they are penetrated by the migmatite-forming microcline granites. The migmatite formation in its turn is brecciated by small, so-called third-group granites, and one of these is cut by veins derived from one of the large rapakivi massifs.

While Sederholm (1923, 1926, 1934) was advancing his four granite groups, different ideas — much inspired by the alpinotectonic principles introduced in Finland by C. E. Wegmann — were evolving. P. Eskola (1932) called Sederholm's gneissose granites synkinematic and the migmatite-forming microcline granites late kinematic, and he incorporated the third-group granites and rapakivi massifs in the postkinematic formations. M. Saksela (1936) preferred the terms synorogenic and late orogenic for the first two groups, and W. Wahl (1936) was the first to refer to the third-group and rapakivi granites as postorogenic.

In 1960, Eskola differentiated the orogenic syn-, sero- and postkinematic granites as well as the epirogenic rapakivi massifs. A. Simonen (1960, 1970) classified the South-Finnish granites into the syn-, late- and postorogenic rocks of the Svecokarelidic cycle and the anorogenic rapakivi granites.

Though the terms anorogenic and postorogenic both imply that the rocks so designated had not been subjected to orogenic movements during their emplacement, they differ in a decisive manner: the former does not allow for a causal relationship with a preceding orogeny, whereas the latter takes no stand in the matter. Vorma (1976) has expressed the opinion that there may be a causal relationship between the Svecokarelian orogeny and the origin of the rapakivi magma. Since the present author knows of no objections to this thesis, and certain results arrived at during this work seem to support it, the term postorogenic will be used exclusively in the following text.

The rapakivi granites have always been considered to be one of the youngest intrusive rock groups in Finland, but their age as related to that of the orogenic rocks was obscure for a long time. Eskola (1932, p. 477) thought originally that their age should be intermediate between the Karelian and Caledonian orogenies. Väyrynen (1954, p. 94) cited a chemical U, Th-Pb determination by Lokka (1950) from the Åva intrusive and remarked of its result, 1116 Ma, that »... it is, however, unexpectedly high for the age of a rapakivi» (translation by the present author). On the basis of isotopic U, Th-Pb determinations made from zircons, Kouvo (1958) concluded that the age of the Wiborg rapakivi massif and of the smaller Kokemäki, Onas and Bodom granites was 1700 Ma. Sederholm's concept of a third group of granites was revived when Neuvonen (1970) reported U-Pb determinations from zircons and titanite occurring in the Åva ring complex that indicated an age of 1830 Ma.

A few years ago, age determinations carried out in conjunction with mapping activities in the Åland area and tin prospecting in the Wiborg, Åland and Laitila rapakivi massifs revealed that there may be significant, though small, age differences within various postorogenic rock groups encountered in these provinces. This led to a full-scale survey of the rapakivi massifs utilizing U-Pb determinations made from zircons. The extent of the rediscovered 1800 Ma age group of postorogenic granites had to be investigated as well. This also seemed to be a proper occasion to reanalyze certain galenas from the Wiborg, massif, the old analyses (Kouvo 1958) having been carried out without standardization by present-day methods. New galena material was also incorporated in this work, and the results obtained with zircons contained in the Laitila rapakivi massif necessitated lead isotopic analyses carried out with whole rock samples and feldspars.

The outcome of the survey is presented in this work. In addition to the results of the age determinations and the comments on them, a brief contribution to the discussion of the discordancy problem of zircons has seemed proper, since the material includes several suites of cogenetic zircons and some abnormally discordant samples. The results of the common lead work carried out with the galenas of the Wiborg massif also necessitate a longer discussion than a mere presentation of analytic results.

It is perhaps proper to stress in this connection that though the author alone is responsible for the conclusions arrived at during this work, his views have been greatly enriched and improved by discussions with several friends and colleagues mentioned in the acknowledgments. Without their help and cooperation, this paper would have remained more incomplete than it is now.

SOME THEORETICAL BACKGROUND FOR THE U-Pb AND Pb-Pb METHODS

This chapter is meant for users of the results obtained in experimental geochronology. Its sole purpose is to introduce the reasoning behind the geological interpretations in this paper. Thus its scope has been limited to the methods used in the course of this work. Also, many of the derivations of the formulae have been omitted, since the mathematical work has been elucidated in the papers of, for example, Russel and Farquhar (1960), Catanzaro (1968), Kanasewich (1968), Doe (1970) and Gale and Mussett (1973). On the other hand, some basic corrections well known to any geochronologist, but only briefly mentioned in the literature, are discussed at somewhat greater length.

The U-Pb method

Uranium occurs in nature as two long-lived, radioactive isotopes, ^{238}U and ^{235}U. These, as well as the similarly radioactive ^{232}Th, produce over lengthy decay series the stable lead isotopes ^{206}Pb, ^{207}Pb and ^{208}Pb, respectively. A fourth isotope of lead, ^{204}Pb, is not radiogenic, and it has not changed in amount during the evolution of our planet. Neither uranium nor lead is known to be subject to isotopic fractionation during normal geochemical processes. In naturally occurring uranium, the molar ratio of the two isotopes is a constant within a certain margin of error.

In the ^{238}U, ^{235}U and ^{232}Th decay series, the half-lives of the parent isotopes are of the order of 10^9 years and much longer than the half-lives of any intermediate members of the series (10^4—10^{-11} years). It is thus obvious that after a certain time of equilibration, the cumulation of the radiogenic lead isotopes is solely governed by the decay of uranium and thorium.

In geochronology, it is practical to measure time as positive backwards. By such a convention, the law of radioactive decay may be written as:

$$\frac{dp}{dt} = \lambda P$$

where λ is the decay constant of the nucleus and dp denotes the quantity of nuclei that disintegrate from P nuclei during the time dt. If P nuclei are observed today, then their quantity P_0 a time t ago would have been:

$$P_0 = Pe^{\lambda t}$$

Daughter nuclei are produced at the same rate as the parent nuclei disintegrate. Thus the number of daughter nuclei D generated during a time t from originally P_0 parent nuclei is:

$$D = P_0 - P = P(e^{\lambda t} - 1)$$

The ratio of daughter and parent nuclei observed today may be written as:

$$R = \frac{D}{P} = e^{\lambda t} - 1$$

This formula may justly be called the basic equation of geochronology.

In praxis, as, e.g., in the analysis of a population of zircons, the ratio observed is not the one on which the age calculation is based. First of all, the system may have contained some daughter nuclei at its formation, and second, both the daughter and parent isotopes may have received additions from laboratory contamination during the analytic procedure. Thus the ratio actually observed is:

$$R_{obs} = \frac{D_{rad} + D_{init} + D_{cont}}{P_{pres} + P_{cont}}$$

The laboratory contamination is simply determined by running a blank analysis. If sufficient control is kept over the working procedures and the purification of reagents, the amounts of daughter and parent nuclei in the blank remain rather constant. The estimation of the amount of initial daughter nuclei, especially in the case of the U-Pb system, is a more difficult problem to solve.

The isotopic composition of lead has changed throughout geologic time as a result of the production of radiogenic lead. Thus the composition of initial lead incorporated in the zircon depends on the time of formation of the mineral and on the geochemical history of the lead. Moreover, the amount of common lead in zircons may vary greatly. This is best demonstrated by comparing the total ^{206}Pb with the nonradiogenic ^{204}Pb. The $^{206}Pb/^{204}Pb$ ratios encountered during this work range from 70 to 29,000. It is obvious that in the latter case the correction for common lead is insignificant, while in the case of a small $^{206}Pb/^{204}Pb$ ratio errors of the order of 10 % in the estimated composition of the common lead may induce marked differences in the results of the age calculations.

Most geochronologists agree that the composition of the common lead contained in zircons is the least radiogenic lead of the geological environment considered. To estimate this, analyses of galenas, feldspars and other minerals containing little or no uranium and contemporaneous with the zircons are carried out. As a rough rule of thumb, it may be stated that if, after the correction for the blank, the initial lead composition is erroneous by 10 % and the $^{206}Pb/^{204}Pb$ is

1) less than 200, the erroneous composition of initial lead may induce significant errors in the age determination,
2) 200—2 000, the results may become somewhat erroneous, but the uncertainties usually bear no significance on the geologic interpretation of the analytic results,
3) larger than 2 000, the error induced by the correction for initial lead is even in the grossest cases less than 1 % of the total analytic error and has no significance.

The Pb-Pb or common lead method

Since the amount of ^{204}Pb has remained a constant during the geologic age of the earth, it has become a common practice to compare the other three, partly radiogenic, lead isotopes with this internal standard provided by nature. For the $^{206}Pb/^{204}Pb$ ratio developed in a sample during a time interval $t = t_1 - t_2$, we may write:

$$\frac{^{206}Pb}{^{204}Pb} = \frac{^{206}Pb_{rad} + {}^{206}Pb_{init}}{^{204}Pb} = \frac{^{206}Pb_{rad}}{^{204}Pb} + \frac{^{206}Pb_{init}}{^{204}Pb}$$

The amount of radiogenic ^{206}Pb depends on two factors: the abundance of ^{238}U in the system and the length of the time interval during which lead is produced. The amount of uranium is usually indexed by the $^{238}U/^{204}Pb$ ratio, conventionally marked as μ. Since the amount of uranium may vary during the development of the system for reasons other than radioactive decay, we must consider μ as a function of time. We may thus write:

$$\frac{^{206}Pb}{^{204}Pb} = \lambda_1 \int_{t_2}^{t_1} \mu(t) e^{\lambda_1 t} dt + a$$

where a is the initial ratio of $^{206}Pb/^{204}Pb$.

If the final $^{206}Pb/^{204}Pb$ ratio is designated X, and it is assumed that μ has been changed only by radioactive decay, we obtain by integration:

$$X = \mu(e^{\lambda_1 t_1} - e^{\lambda_1 t_2}) + a$$

and similarly for the $^{207}Pb/^{204}Pb$ ratio

$$Y = \frac{\mu}{k}(e^{\lambda_2 t_1} - e^{\lambda_2 t_2}) + b$$

the factor k arising from the natural ratio of the two uranium isotopes and λ_1 and λ_2 being the decay constants of ^{238}U and ^{235}U, respectively.

By combining the last two equations, it can be seen that in $^{206}Pb/^{204}Pb$ vs. $^{207}Pb/^{204}Pb$ coordinate system samples of the same age (t_1, t_2) and having the same initial composition of lead (a, b) can be plotted on a straight line, the slope of which is a function of t_1 and t_2. The formula for the slope is known as the *Holmes-Houtermans equation*

$$m = \frac{dy}{dx} = \frac{e^{\lambda_2 t_1} - e^{\lambda_2 t_2}}{k(e^{\lambda_1 t_1} - e^{\lambda_1 t_2})}$$

It is obvious that, having measured the slope, we may solve for either t_1 or t_2 if the other one is independently known. Alternatively, we may get an upper estimate for t_1 by assuming $t_2 = 0$. In this case, the equation becomes:

$$m = \frac{e^{\lambda_2 t_1} - 1}{137.88\,(e^{\lambda_1 t_1} - 1)}$$

where the factor 137.88 reflects the present-day molar ratio of $^{238}U/^{235}U$. Obviously, for systems where lead is still evolving from uranium (eg., whole rock samples), the last equation is applicable as such. The constants currently used in the U-Th,Pb and Pb—Pb age calculations are given in Table 1.

Table 1. The current constants used in the U, Th—Pb and Pb—Ph age calculations.

Symbol	Meaning	Value	Source
μ_1	dec.const. for ^{238}U	1.155125×10^{-10}	Jaffey *et al.* (1971)
μ_2	» ^{235}U	9.8485×10^{-10}	»
μ_3	» ^{232}Th	4.9475×10^{-11}	LeRoux & Glendenin (1963)
k	present $^{238}U/^{235}U$	137.88	Jaffey *et al.* (1971)
a_0	prim. $^{206}Pb/^{204}Pb$	9.307	Tatsumoto *et al.* (1973)
b_0	prim. $^{207}Pb/^{204}Pb$	10.294	»
c_0	prim. $^{208}Pb/^{204}Pb$	29.476	»
t_0	earth's age	4570 Ma	»
μ	$^{238}U/^{204}Pb$	variable	———
W	$^{232}Th/^{204}Pb$	»	———

The discordancy problem of zircon age determinations

The two U—Pb ages and the Pb—Pb age derived during an analysis usually deviate from each other. The deviation is generally much larger than the analytic uncertainty and follows the pattern

$$t_{206Pb/238U} < t_{207Pb/235U} < t_{207Pb/206Pb}$$

Ahrens (1955) showed that contemporaneous samples plot close to a straight line, and that at one point in this line all three ages coincide, yielding a »convergent» age.

Wetherill (1956) demonstrated that an episode occurring at a time t_2 and causing lead loss in a suite of samples formed previously at a time t_1 would produce a linear array on the $^{207}Pb/^{235}U$ vs. $^{206}Pb/^{238}U$ diagram and intersect the concordia curve, which is the locus of the »convergent» ages, at the points defined by t_1 and t_2. This is called the episodic lead loss model and the diagram the concordia diagram.

On the basis of an earlier work of Nicolaysen (1957), Tilton (1960) and Wasserburg (1963) showed that continuous loss of lead by diffusion would result in an almost linear distribution of discordant sample material on the concordia diagram in the area where analytic results are normally plotted. Noting that discordancy usually seems to increase as a function of a growing uranium content, Wasserburg (1963) correlated the rate of diffusion with increasing lattice damage by alpha bombardment of the mineral as it became older.

Both the episodic lead loss and continuous diffusion models have been utilized extensively in the interpretation of discordant analytic data. It has been shown repeatedly (e.g., Silver and Deutsch 1963, Catanzaro 1968, Meriläinen 1976) that all analytic data do not fit a continuous diffusion model. It has also been demonstrated experimentally that conditions similar to pneumatolytic ore formation and metamorphism in the greenschist facies are sufficient to leach lead rapidly from zircon (Pidgeon *et al.* 1966). Recent results of investigations of low-grade metasediments (Gebauer and Grünenfelder 1976) support this finding.

On the other hand, the age indicated by the lower intersect of the isochron with the concordia curve cannot always be accounted for or tied up with a geologic event. This is especially true of the Precambrian of Finland, where the range of the lower intersect ages (t_2) is from 0 to 500 Ma. Mechanisms like recent leaching by weathering may be envisaged, but in the present author's opinion there is very little evidence of large-scale episodic events during the Phanerozoic time in southern Finland if one excludes sedimentation on top of the Precambrian craton and its subsequent denudation. Moreover, it has been demonstrated repeatedly that combined continuous and episodic losses of lead may result in sublinear distributions on the concordia diagram (Shukolyukov 1964, Allégre *et al.* 1974). In these cases, neither the upper nor the lower intersects yield ages of any geologic significance.

Problematic as the discordant age data of zircons may seem, it should be remembered that this is due to the mineral's unique resistance in later geologic processes. For instance, titanite and uraninite, both of which often produce concordant U—Pb age data, are far more readily reset by metamorphism than is zircon, which almost invariably retains some »memory» of its original formation. Thus zircon, in spite of its metamictization and tendency to produce discordant age data, is one of the best recorders of ancient events.

LABORATORY PROCEDURES

Sample preparation

Zircons. The zircon fractions were purified from crushed rock samples (2—80 kg) by separations on a shaking table, subsequent employment of heavy liquids (bromoform, dimethyliodide, clerici's solution) and by magnetic means. As a final step, the samples were purified by handpicking under a stereoscopic microscope.

Galenas. The samples were picked directly from hand specimens. The material was cleaned by hand-picking from other sulphides and silicates under a stereo-microscope.

Whole rock samples. The whole rock samples (about 500 g) were prepared according to the procedure described by Sakko and Laajoki (1975).

Feldspars. The feldspar was separated by means of bromoform diluted with ethyl alcohol from the light fraction of the bromoform separation.

Chemical treatment

Reagents. All the reagents used were clean commercial products, which had been purified further by distillation in quartz stills and subsequent evaporation at sub-boiling temperatures in teflon vessels.

Zircons. The samples were washed in HNO_3conc. and rinsed several times with H_2O. Before 1973, the sample size was 200—600 mg. Most of the recalculated samples were decomposed by borax fusion. Since the adoption of the technique described by Krogh (1973), the sample size has been 5—50 mg. All the analyses done by the present author were carried out utilizing this technique.

Galenas. The galena samples were washed for one hour in 1—N HCl, rinsed several times with sub-boiled H_2O, and then dissolved in 1 ml of HNO_3conc. The purification of the lead was accomplished by anion exchange techniques in HCl solutions.

Whole rock samples and feldspars. The samples were washed for 20 minutes in 1—N HCl, rinsed several times with sub-boiled H_2O, and decomposed according to the procedure described by Sakko and Laajoki (1975). The purification of the lead was accomplished by anion exchange techniques in HCl solutions.

Mass-spectrometry

The mass-spectrometer utilized is a Nier-type, 9″—60°, single-focussing, single-filament machine built at the Geological Survey according to the construction worked out at the Carnegie Institution of Washingston. A Cary 31 vibrating reed electrometer is used to measure the ion current detected by a Faraday cup-collector. Before 1974, the measurements were done by hand from strip-chart recordings. Since then, the electrometer reading is digitalized with a Fluke 8200A DVM. A device built at the Finnprospecting Ky calculates the mean of 400 measurements and punches it on a paper tape. An automatic five-channel magnet control unit allows switching from peak to peak every 8 sec and triggers the measuring event after a 7 sec relaxation time. The paper tape output is processed by the time-sharing HP 3000 computer of the Geological Survey.

Some of the earlier runs were conducted from tantalum filaments. At present, all samples are loaded on rhenium filaments employing H_3PO_4-silica gel emitters for Pb and H_3PO_4-Ta_2O_5 emitters for U runs.

A mass-spectrometer run involves 2—4 sets of 15 measurements from each of the 5 channels. This results in 14 measurements per set of the $^{206}Pb/^{204}Pb$, $^{206}Pb/^{207}Pb$ and $^{206}Pb/^{208}Pb$ ratios when determining the isotopic composition of lead. When the concenteration of Pb or U is involved, one set contains 42 measurements of either the $^{206}Pb/^{208}Pb$ or the $^{235}U/^{238}U$ ratio. The mean and its standard deviation are derived from the measured ratios, after which values deviating from the mean by more than $1.5 \times$ STD are eliminated and the means and the STDs are recalculated foı the remaining data. The final result for each isotopic ratio is the weighted mean of the results of the sets. In the case of Pb, a correction based on calibration with the CIT lead is made. The isotopic composition of the CIT lead is assumed to be that reported by Catanzaro (1967).

Age calculations

All the age calculations presented in this work are based on the latest decay constants for U (Jaffey *et al.* 1971) and Th (LeRoux and Glendenin 1963). Thus the U-Pb ages for some of the recalculated zircon analyses are about 30 Ma lower than those reported in previous publications.

The model ages of the galenas were calculated by utilizing the two-stage model of Stacey and Kramers (1975). The age calculation programs used allow also the derivation of the $^{238}U/^{204}Pb$ ratio (μ_2) for any given sample, and these data are repoıted along with the model ages.

The use of the two-stage model may seem to be a handicap in the calculation of »primary» ages from the slopes of »secondary» isochrons. Admittedly, the model employed adds one more stage to any multistage model. However, as can be seen from the Holmes-Houtermans equation, the slope of the isochron is still the function of the same interval of time no matter whether it is termed the 2nd, 3rd or *n*th stage. Thus the geologic event to be dated is the same, and the slope of the »secondary» isochron yields the same age for both the single-stage and two-stage models.

Estimation of analytic errors

It should be emphasized that error estimates for individual deteminations given in this work as well as in most other publications on geochıonology are analytic errors. They represent only the measurable uncertainties of laboratory work. The geologic accuracy remains a question of interpretation and cannot be measured by means of mathematical formulae and statistical methods.

Zircons. In the case of the pre-1975 analyses, no separate error estimates have been made. The introduction of computer techniques has greatly facilitated such calculations, which are now a routine part of every analysis. The error estimate is based on the mean weighing error in the laboratory, deviations obseıved in the mass-spectrometeı runs and uncertainties of the spike compositions and concentrations. From these, the maximum error is calculated. Since the occurrence of the maximum error is highly improbable, the margins of error are reported on the 95.6 % (2σ) confidence level.

Common lead. Common lead analyses are extremely vulnerable to errors in the $^{206}Pb/^{204}Pb$ ratio. This becomes especially crucial if age determinations are based on a single sample, as is the case in the lead model age calculations. Therefore, from each ore specimen several batches of galena underwent chemical treatment and 3—4 runs were performed on each chemistry. Thus the number of runs per galena varied from 6 to 12. This procedure also served to establish the isotopic inhomogeneity of the lead in some of the samples.

Since the whole rock and feldspar samples were homogenized as a result of thorough grinding, only three runs from the same chemistry were performed.

The errors reported for the common lead determinations are the between-run STDs in the cases of the measured ratios $^{206}Pb/^{204}Pb$, $^{206}Pb/^{207}Pb$ and $^{206}Pb/^{208}Pb$. Compound errors aie reported for the calculated ratios $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$.

Computer programming

The computer programming was carried out in the HP 3000 versions of Basic and Fortran. The author is responsible for most of the programming. Ms. Marjatta Asa, M.Sc., is to be credited with programming the Wasserburg model. In this particular program, the author has made only some minor revisions, which were necessitated by a change of computer.

SIGNIFICANCE OF THE ZIRCON AGES

Zircons occur in the rapakivi granites usually as subhedral or euhedral grains. The grain size is quite variable even within a single sample, ranging from 50—200 mesh. Many of the larger grains, especially, exhibit euhedral zoning, which is commonly thought to reflect quite uninhibited crystal growth in a more or less liquid environment (e.g., Silver 1969). Inclusions of ilmenite are common, and at times semiopaque cores of a rounded, partly resorbed and apprently older zircon generation may be observed. With this in view, it is necessary to consider the significance of the zircon data that will be presented in the course of this work. It is especially important, since in one case (p. 30) an age of 2750 Ma was obtained from rounded, partially resorbed zircons.

The purification of the analyzed material was carried out with utmost care. Most of the grains containing inclusions of ilmenite could be removed with a Frantz isodynamic magnetic separator, but the final step was invariably hand-picking, the remaining grains that contained impurities and exhibited older cores being removed at this stage. Nothing could be done about the zoning, since it is a characteristic feature of the rapakivi zircons. Thus there is a possibility that zones of different ages could be incorporated in the analyzed material.

Zircon is an early crystallizing phase in granitic rocks, and the zoning, as already stated, reflects uninhibited growth in a magmatic environment. If age differences between the zones were significant, a set of mixed ages, with an upper limit determined by the beginning of the formation of the mineral and a lower limit determined by the consolidation of the magma, would be formed. This would result in a distribution of the analytic material along the concordia curve, and, after loss of lead by continuous or episodic diffusion processes, in a random spread on the concordia diagram. Since such distributions have not been observed in the several zircon suites obtained from one sample, it must be concluded that each of these suites was formed by a discrete episode, i.e., age differences larger than the analytic error cannot be detected for zircon fractions separated from the same rock.

The nature of this discrete episode is, however, not revealed by analytic techniques. It is unfortunate that very little experimental phase work has been carried out on zircons, and that none known by the present author relates to the uranium and lead equilibria between zircons and magmatic melts. With no such data available, circumstantial evidence must be utilized.

Various investigators have come to the conclusion that rapakivi magma has been generated at great depths in conditions corresponding to the granulite facies (e.g., Kranck 1968, Barkeı *et al.* 1975, Vorma 1976). The conditions in the anatectic magma seem to have been on the borderline of the stability of zircon, as demonstrated by the residual cores. The slow corrosion of the primary zircon generation seems to indicate that the crystals and the magma never reached a state of equilibrium. Consequently, it must be assumed that, since the magma was able to corrode the old zircons, it must have been undersaturated in respect of zirconium, and thus unable to produce a new generation of zircon as long as it remained in the infıastructure of the earth's crust. Only as the distensional phase of the crustal movements began and the magma started its ascent to the upper regions of the crust, a new generation of zircon was able to form. Thus there are grounds for assuming that the zircon ages of the rapakivis represent maximum estimates concerning the emplacement of these rocks. Since in the mafic rocks the major phases had crystallized prior to the formation of zircon, the zircon ages are minimum estimates for the gabbro-anorthosites analyzed.

MASSIFS AND STOCKS STUDIED AND THEIR AGES

The Wiborg massif

Outline

Of the postorogenic granite bodies of souther Finland, the Wiborg rapakivi massif is by far the largest. Like the other rapakivi massifs of the Fennoscandian (or Baltic) shield, it cuts sharply across the surrounding metasediments, the synorogenic and late orogenic granites and the set of the Subjotnian diabase dikes described by I. Laitakari (1969). A thermal aureole has been established around it by investigation of the ordering of potassium feldspar in the surrounding granitoids (Vorma 1972). Previous age determinations (Kouvo 1958) indicate that it is definitely younger than the Svecokarelian diastrophism.

According to the classic definition (cf. Vorma 1976), rapakivi is a potassium granite exhibiting ovoid crystals of potassium feldspar covered by a mantle of oligoclase. The ground mass is a medium- or coarse-grained granite, in which biotite and hornblende are the usual mafic constituents. Fluorite is the characteristic accessory mineral, and zircon occurs more abundantly than in granites in general. Usually, the accessory titanium minerals are ilmenite and anatase, titanite being more of a rarity. The name of the rock is derived from its conspicuous mode of crumpling (e.g., Eskola 1930).

Most of the Wiborg massif consists of this typical rapakivi, which is called wiborgite. However, varieties lacking the oligoclase mantle around the ovoid potassium feldspar (pyterlites) are relatively widespread. In addition, there occur fayalite-bearing, darkish varieties (tirilites) and porphyritic and even-grained granites exhibiting the chemical characteristics of rapakivi (Wahl 1925, Hackman 1934, Lehijärvi and Lonka 1964, Haapala 1974). At least around the massif of Ahvenisto (Savolahti 1956 and 1966) and at Ylämaa in the Wiborg massif gabbros and anorthosites are met with. A number of granite porphyry and quartz porphyry dikes are known from the massif (Wahl 1925, Savolahti 1956, Vorma 1971). In the Gulf of Finland, on the islands of Hogland and Sommarö, there occur acid and mafic effusive rocks, which have been equated with the more deep-seated rocks of the Wiborg massif (Wahl 1947). Recently, another acid effusive occurrence of rapakivitic affinity has been described from the Ruoholampi roof pendant (Vorma 1975). The area of the Wiborg massif also exhibits a number of small-scale galena mineralizations.

The existing age determinations from the Wiborg massif are few and scattered in many publications. Kouvo (1958, p. 43), on the basis of U-Pb determinations of zircons, estimated 1670 Ma to be the age of the rapakivi and the 3rd group granites of Sederholm. If the current decay constants (Jaffey *et al.* 1971) are used, this would equal an age of between 1630 and 1650 Ma, depending on the discordancy of the sample material. The K-Ar and Rb-Sr ages from the micas and feldspars of these rocks were slightly lower, ranging from 1560 to 1630 Ma (op.cit.). The radiation damage ages (op.cit.) were of the order of 800 Ma, except in the case of the porphyry dikes, which yielded estimates of around 450 Ma. Kouvo (op.cit.) found that the lead in several galena veins of the Wiborg massif was »J-type anomalous.» In the same paper, analyses of the Hyvärilä and Koskenkylä deposits, situated in the contact zone of the Wiborg massif, were published. Kouvo and Kulp (1961) also reported analyses of some galenas contained in the Wiborg massif. Later on, some age determinations were published in the Annual Reports of the Geological Survey (1970, 1973). Vorma (1975) reported the ages of an acid effusive rock from the Ruoholampi roof pendant and of a Subjotnian diabase occurring in the Hyvärilä roof pendant. Recently, three fossil fission track ages for apatites, with a mean of 570 Ma, were published by Lehtovaara (1976).

Age determinations and discussion

Brief descriptions of the sampled rocks as well as pertinent references are given in Appendix 1. The sampling sites are also indicated on the index map (Fig. 1). The results of the analyses are presented in Table 2 together with the ages calculated

Fig. 1. The postorogenic formations of southern Finland. The encircled numbers represent the sampling sites. The numbers refer to the index numbers used in Tables 2—11.

Table 2. U—Pb analyses on zircons

Sample No.	Index No.	Locality	Rock type	Analyst & year	^{238}U ppm	Radiogenic lead (ppm) ^{206}Pb	^{207}Pb	^{208}Pb
A 18	1	Värtö	Tirilite	MVa, 75	227.7	55.48	5.63	8.93
A 21	6	Mentula	Granite porphyry	MVa, 76	1 175.3	209.63	20.87	48.49
A 29	3	Muurikkala	Wiborgite	MVa, 75	530.1	104.21	10.49	14.37
				MVa, 75	952.3	182.31	18.22	24.23
A 69	5	Verla	Porphyritic granite	MVa, 76	634.9	129.14	13.07	51.06
				MVa, 76	2 161.1	402.23	39.71	84.63
A 96	4	Hiidensaari	Porphyritic granite	OKo, 56[x]	488	106.4	10.7	12.6
				MVa, 76	536.5	111.81	11.30	13.66
A 98	2	Suomenniemi	Hornblende granite	MVa, 76	305.4	67.41	6.84	9.61
A118	7	Nurmaa	Gabbro-anorthosite	OKo, 73[y]	281.9	69.58	7.07	13.11
				OKo, 73[y]	768.4	190.73	19.36	41.37
A323	8	Hamina	Quartz porphyry	OKo, 70[z]	410.8	68.2	6.8	10.8
				MVa, 76	370.7	61.76	6.17	10.11
				MVa, 76	542.3	89.34	8.84	9.94
A422	9	Ruoholampi	Rapakivi porphyry	OKo, 72[wv]	131.5	31.24	3.18	7.02
A423	17	Hyvärilä	Subjotnian diabase	OKo, 73[yv]	233.1	57.25	5.87	7.81
A524	10	Kymi	Porphyritic granite	OKo, 74	1 551.3	257.5	25.04	33.20
A627	11	Hogland, USSR	Rapakivi porphyry	MVa, 75	113.3	24.33	2.48	5.34
A629	12	Parola	Contact variety	MVa, 76	1 370.5	326.23	32.80	41.11
				MVa, 76	2 622.3	581.49	57.93	81.76
				MVa, 76	1 992.8	384.05	37.75	62.03
A630	12	Parola	Pyterlite	MVa, 76	1 049.0	210.54	21.21	24.96
				MVa, 76	1 509.4	317.88	32.14	43.74
A631	13	Pesäntäjärvi	Wiborgite	MVa, 76	453.1	79.46	7.94	9.52
				MVa, 76	913.5	155.29	15.52	20.09

[x] Previously published in Kouvo (1958), recalculated.
[y] Previously published in Annual Report of Activities (1973), recalculated.
[z] Previously published in Annual Report of Activities (1970, recalculated.
[w] Previously published in Annual Report of Activities (1972), recalculated.
[v] Previously published in Vorma (1975), recalculated.

according to the ^{206}Pb/^{238}U, ^{207}Pb/^{235}U and ^{207}Pb/^{206}Pb ratios and the continuous diffusion model of Wasserburg (1963). A concordia plot derived from the analytic data is presented in Fig. 2.

The mean age of the analyzed material as a whole worked out from the ages accoiding to the Wasserburg model is 1 664$\pm$21 Ma. If the somewhat younger samples A524 and A629 are omitted, the figure becomes 1670$\pm$17 Ma. Considering the systematic differences induced by the new decay constants, the result is in excellent agreement with earlier work. A reanalysis of some of the samples has cut down the range of the individual determinations from 100 to 70 Ma. The more abundant material has also caused a definite decrease in the standard deviation of the mean of the ages.

As far as the areal distribution of the analytic results is concerned, the highest age, 1 693$\pm$12 Ma (mean of the analyses of sample A323), occurs in the southern part of the massif. Excepting sample A631, the ages are slightly lower in the northern part of the massif, being as low as 1 652 Ma in the Ahvenisto satellite. The age of the Suomenniemi massif is about 1 670 Ma.

from the Wiborg rapakivi massif.

Total Pb ppm	$^{206}Pb/^{204}Pb$	Radiometric ages $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{206}Pb/^{207}Pb$	Wasserb	Fraction density/mesh size
70.40	5 870	1 599±20	1 617±12	1 641± 2	1 650±10	+4.2/—100
299.98	703	1 208±51	1 359±39	1 607± 6	1 672±12	3.8—4.1
135.52	1 127	1 319±26	1 442±18	1 628±10	1 674±15	+4.1/50—150
238.33	939	1 288±45	1 416±32	1 614±18	1 664±20	3.8—4.1/50—150
286.24	203.4	1 360± 8	1 472± 6	1 637± 5	1 678±10	+4.1/100—200
644.64	250.6	1 255±27	1 385±19	1 591± 5	1 645±10	3.6—3.8/+100
143.47	628	1 449	1 524	1 630	1 652	Total
136.57	5 700	1 391±34	1 490±21	1 634± 5	1 668±10	Total
85.09	3.710	1 464±15	1 538± 8	1 641± 2	1 664±10	+200
90.24	7 400	1 617	1 629	1 645	1 654	+4.1
253.33	9 200	1 625	1 633	1 642	1 650	3.8—4.1
120.4	150.0	1 131	1 309	1 612	1 702	Total
107.95	157.9	1 135±14	1 311±14	1 612± 5	1 698±10	+4.2
135.46	240.8	1 123±21	1 296±14	1 596± 5	1 680±11	3.8—4.1
44.1	861	1 563	1 600	1 650	1 662	Total
71.04	1 226	1 610	1 632	1 661	1 674	+4.0
317.76	139	1 131	1 289	1 562	1 638	Total
34.16	845	1 429± 9	1 519± 5	1 648± 7	1 682±10	Total
401.28	28 800	1 566±82	1 591±49	1 625± 6	1 633±10	4.0—4.2/+150
724.13	13 600	1 470±73	1 527±45	1 608± 5	1 626±10	3.8—4.0
494.08	2 630	1 296±16	1 408±11	1 583± 5	1 628±10	3.6—3.8
259.67	4 900	1 344±64	1 458±42	1 628± 5	1 675±10	+4.0/—100
401.36	2 890	1 404±29	1 498±19	1 635±10	1 669±13	—4.0/+150
99.41	2 220	1 189±26	1 349±19	1 613± 7	1 685±10	+4.2/—100
203.77	845	1 156±15	1 326±11	1 614± 6	1 697±12	3.8—4.2/—200

Calculations with the decay constants of Jaffey *et al.* (1971).
All analyses corrected for blank. Common lead corrected according to G272 (Table 10)
Analysts: Matti Vaasjoki (MVa) and Olavi Kouvo (OKo).

Sample A627, representing the red porphyry of Hogland (Wahl 1947), is slightly older than the mean age of the massif. The porphyry A422 in the Ruoholampi roof pendant (Vorma 1975) is also somewhat older than the oldest intrusive rapakivi in the vicinity, the tirilite (Vorma 1971, p. 6). Although the evidence drawn from only two samples is anything but conclusive, the results seem to substantiate the view that these porphyries represent eruptions belonging to the early intrusive phases of the Wiborg massif (Wahl 1947, Vorma 1975).

The Ahvenisto satellite massif was studied thoroughly by Savolahti (1956, 1966), who, in contrast to an earlier opinion of Frosterus (1903), concluded that the gabbro-anorthosite arc surrounding the rapakivi is much older than the rapakivi and the Subjotnian diabase dikes met with at Lovasjärvi and Vuohijärvi. He stated, however, that the diabase-like portions observed within the gabbro-anorthosite arc could be interpreted as marginal variants and apophyses of several intrusive pulses (Savolahti 1956, p. 46), and later reported the existence of seven different mafic intrusions in the arc surrounding the Ahvenisto massif (Savolahti 1966, p. 175). Buddington (1959, p. 685) classified the Ahvenisto massif and the surrounding mafic rocks as

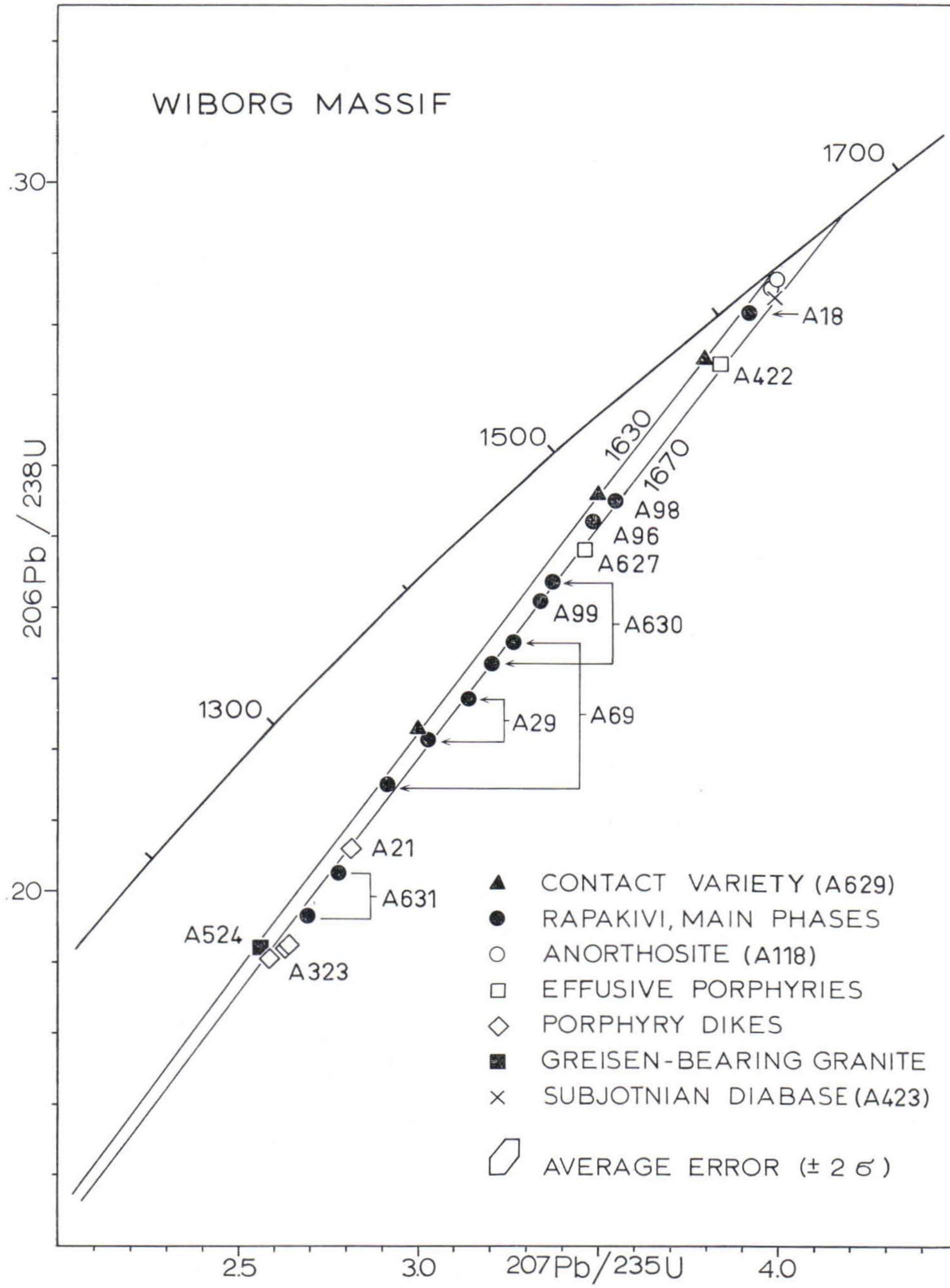

Fig. 2. The concordia diagram of analyses from the Wiborg massif.

a ring-dike complex. The two analyses of sample A118 suggest that the gabbro-anorthosite is contemporaneous with the rapakivi. It should be noted that the surprisingly large amount of zircon found in the sample may indicate a genetic relationship with the rapakivi.

The age of the quartz porphyry dike at Hamina (A323) is the highest in the analyzed material. No analyses have been made of its immediate environment; but a comparison between the ages of the dominant rock type of the Suomenniemi massif (A98) and the granite porphyry of Mentula (A21) belonging to a dike set traversing it indicates that, in this case, the ages of the country rock and the intersecting dike coincide within their limits of analytic error. There is some basis, therefore, for regarding the age of the Hamina dike as a reasonable approximation of the age of its country rock, too.

Sample A524, representing the cupola formation at Kymi, is somewhat younger than the mean age of the Wiborg massif. Considering the geologic field relationships (Haapala and Ojanperä 1973, Haapala 1974), the difference appears to be real and not induced by analytic errors.

Sample A629, taken from the northern contact of the Wiborg massif, is definitely the youngest in the analyzed material. The composition of the rock is granitic, potassium feldspar, plagioclase and quartz appearing as the major minerals. Biotite and olivine, slightly serpentinized, appear as minor constituents. Zircon and opaques form the most conspicuous accessory minerals. At least two generation of both feldspars and quartz occur. Analyses of three different fractions of zircons cover a fairly wide range of discordancy, and the results are in perfect agreement, yielding a mean age of 1 629$\pm$4 Ma.

Sample A630, taken at a distance of only 200 m from sample A629, yields a mean age of 1 672 Ma. Thus there is a 43 Ma age discrepancy within a single outcrop, something that cannot be attributed to analytic errors. The conclusion must therefore be that the ages obtained represent two different geologic events. There are at least two alternative interpretations to consider.

Since the contact variety (A629) exhibits a sharp border on the side of the rapakivi, too, it could be argued that we are dealing with a dike injected into a natural zone of weakness. Its material could have been derived from residual phases or remobilizates of the rapakivi proper. However, younger contact injections encountered, e.g., in the Lappeenranta area (Vorma 1965) are normal granite porphyries, which do not resemble the rock represented by sample A629.

On the other hand, within the road cut of Parola there are several successive layers, which represent a transition from the contact variety (A629) to the pyterlite (A630), all exhibiting sharp borders against each other and containing an increasing amount of feldspar phenocrysts in the direction of the rapakivi proper. This could be interpreted as a chilled margin with a laminar flow structure. The age data could be accounted for by assuming that the rapidly chilled contact rock was for a considerable time thermally metamorphosed by its host pluton, resulting in lead loss from

Table 3. U—Pb analyses on zircons from

Sample No.	Index No.	Locality	Rock type	Analyst & year	^{238}U ppm	Radiogenic lead (ppm) ^{206}Pb	^{207}Pb	^{208}Pb
A 97	23	Onas	Potassium granite	OKo, 56[x]	290.5	62.78	6.28	6.28
A101	24	Bodom	Even-grained granite	OKo, 56[x]	552.9	132.08	13.40	12.19
				MVa, 76	568.9	136.00	13.73	13.04
A102	24	Bodom	Porphyritic granite	MVa, 76	824.7	169.34	17.01	19.48
A602	25	Obbnäs	Porphyritic granite	MVa, 75	284.2	60.73	6.09	9.10
				MVa, 75	1 072.8	181.97	17.76	23.13
				MVa, 75	1 093.4	170.68	16.35	22.17
				MVa, 75	1 811.1	220.44	20.50	32.34

Analysts: Matti Vaasjoki (MVa) and Olavi Kouvo (OKo).
[x]Previously published in Kouvo (1958), recalculated.

its zircons at early stages of development. It is remarkable, indeed, that all the fractions of sample A629 exhibit abnormally high $^{206}Pb/^{204}Pb$ ratios, which indicate an unsually low content of common lead in the sample. A lead loss at an early stage would, of course, result in the loss of initial lead as well, leading to high $^{206}Pb/^{204}Pb$ ratios at a later stage of development.

The Onas, Bodom and Obbnäs granites

Outline

The small granite bodies of Onas, Bodom and Obbnäs have been dealt with in the literature mostly in conjunction with other problems, and have therefore received only a limited amount of attention (Sederholm 1923 and 1926, Borgström 1931, Laitala 1961 and 1973). The rocks are fairly uniform throughout the intrusive bodies, the main parts being coarse-grained porphyritic in texture. Potassium feldspar is the dominant mineral, forming usually angular crystals 2—3 cm long and 1—2 cm wide. The other main minerals are quartz, oligoclase, biotite and hornblende. Fluorite and zircon are the usual accessory minerals. These rocks differ from the rapakivi granites inasmuch as their titanium-bearing mineral is titanite.

Structurally, the Obbnäs and Bodom granites exhibit an orientation of the potassium feldspar crystals, which runs parallel to the strike of the Porkkala—Mäntsälä fault, passing from SW to NE in between these granites. Also, according to the present author's own investigations and earlier studies (Sederholm 1926), the quartz in both rocks is in some places crushed in a cataclastic manner, and, at least in one case (see Appendix 1), mylonites have been met with.

Kouvo (1958) reported one U-Pb age for zircons contained in each of the Onas and Bodom granites, and he concluded that these rocks are contemporaneous with the Wiborg rapakivi massif.

the Onas, Bodom and Obbnäs granites

Total Pb ppm	$^{206}Pb/^{204}Pb$	Radiometric ages $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{206}Pb/^{207}Pb$	Wasserb	Fraction density/mesh size
86.6	386	1 437	1 511	1 618	1 642	Total
163.4	1 770	1 571	1 601	1 642	1 650	Total
178.87	813	1 572±23	1 598±14	1 632± 8	1 642±10	Total
228.81	522	1 372±29	1 474±28	1 623±10	1 659±14	Total
91.78	300.8	1 422±16	1 503±12	1 619± 7	1 647±10	+4.2/100—200
266.80	279.2	1 153±35	1 307±20	1 570± 7	1 644±13	4.0—4.2/100—200
262.28	237.1	1 069±23	1 233±14	1 534±10	1 620±13	4.0—4.2/+100
367.18	177.4	848±22	1 043±17	1 478±12	1 626±15	3.8—4.0/+150

Calculations with the decay constants of Jaffey *et al.* (1971).
All analyses corrected for blank.
Common lead correction according to G272 (Table 10).

Age determinations and discussion

Brief descriptions of the rock samples as well as pertinent references are given in Appendix 1. The sampling sites are also indicated on the index map (Fig. 1, p. 17). The results of the analyses are presented in Table 3. A concordia diagram is shown in Fig. 3.

The arithmetic mean of the ages according to the Wasserburg model for the eight analyses carried out is 1 641±13 Ma. A least squares fitted line yields an age of 1 645±5 Ma for the upper intersect with the concordia curve. Since the two most discordant samples from specimen A602 (Obbnäs) would be plotted above the continuous diffusion trajectory, the author has chosen to use the upper intersect of the least squares fit as the best age estimate. From the geologic point of view, it is immaterial which interpretation is used, since the upper intersect and the mean of the Wasserburg model ages coincide within their margins of error. As for the lower intersect, yielding an age of 253 Ma, it is impossible to assign any known geologic event from Finland to this date. It should be pointed out that the lower intersect age in the case of the Wiborg massif would be 176 Ma.

The orientation of the potassium feldspar crystals in the Bodom and Obbnäs granites parallels both the long axes of these intrusive bodies and the Porkkala—Mäntsälä fault. This orientation may be due to either magmatic flow or to tectonic movements during the emplacement of these granites. The present field observations are insufficient for deciding this matter. The mylonites observed in the Obbnäs granite, which also parallel the Porkkala—Mäntsälä fault, suggest that tectonic activity took place at some time after the emplacement of the Obbnäs granite.

The analyses of the minor granite massifs of southern Finland demonstrate that these rocks are definitely younger than most of the Wiborg massif, and are very likely contemporaneous with the rocks of the Ahvenisto massif and, possibly, of the greisen-bearing Kymi cupola formation. The best age estimate for the Onas, Bodom and Obbnäs granites in 1 645 Ma.

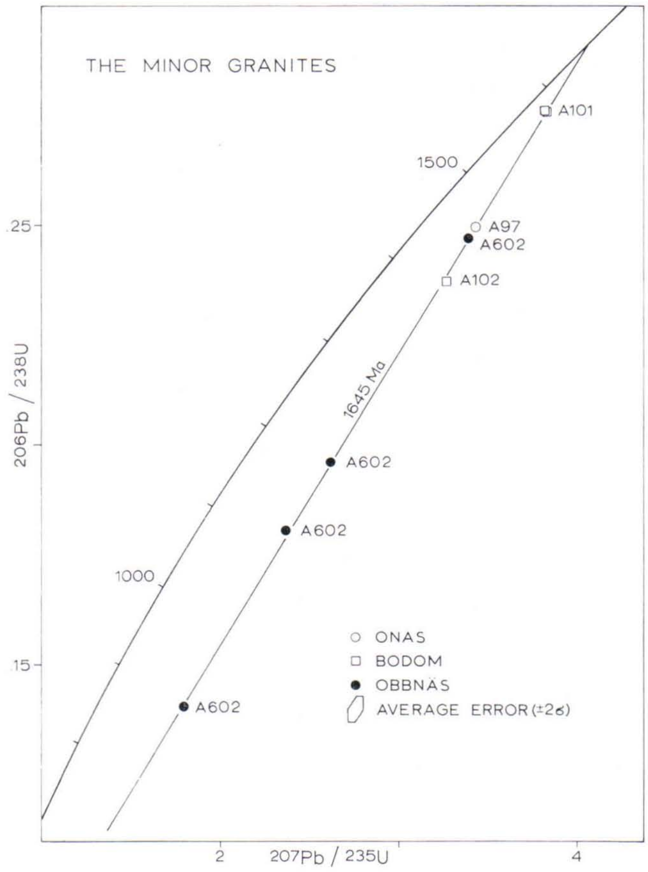

Fig. 3. The concordia diaram of analyses from the Onas, Bodom and Obbnäs granite stocks.

The Åland area

Outline

The postorogenic rocks of the Åland area have not been studied as closely as the rapakivi granites of the Wiborg massif. In the papers of Sederholm (1934) and Hausen (1964), the rocks are described in a general manner, but only the ring-dike complexes of Åva (Kaitaro 1953, Bergman 1973a) and Seglinge (Björk 1976) and the circular even-grained rapakivi granite of Fjälskär (Beıgman 1973a) have received detailed petrological attention.

The main rock type within the Åland rapakivi massif deviates from the classic definition of rapakivi (p. 15) inasmuch as the pyterlitic ovoid crystals of potassium feldspar prevail over the wiborgitic ones. Even-grained and porphyritic varieties are occasionally met with, and these granites contain several greisen formations (Bergman 1973b). Quartz porphyry dikes also occur, and in one case, on the islet of Blåklobb, such a dike cuts a mafic rock designated by Sederholm (1934, p. 60) as ossipite. A general description furnished by him (*op.cit.*, p. 55) makes it evident that the rocks called ossipites in the Åland area are gabbro-anorthosites similar to those encountered in the Wiborg and Laitila massifs.

Sederholm (1934, p. 46) also noted that the potassium-rich porphyritic granites of Lemland and Mosshaga brecciate the Svecofennian formation, but that the Lemland granite is cut by dikes of rapakivitic affiliation. He called these granites as well as the monzonite-potassium granite ring complex of Åva third-group granites, a name still current in colloquial usage among Finnish geologists.

In a thorough study of the Åva complex, Kaitaro (1953, p. 66) concluded that the geologic relations in the area indicated separate monzonitic and granitic intrusive phases, the potassium granite being the younger one. Neuvonen (1970) reported, in connection with a paleomagnetic study, that U—Pb analyses of the titanite and zircon contained in the Åva intrusives yielded an age of 1 830 Ma. Other radiometric ages from the third-group granites have been published in the Annual Report of the Geological Survey of Finland (1972). Using criteria derived from tectonic observations, Bergman (1971) concluded that the Fjälskär granite is of younger origin than the Åva intrusives and probably of the same ages as the rapakivi granites. This thesis was verified in 1972, when age determinations of the Fjälskär granite and the slightly younger rapakivi of Kökar fjärd (e.g., Sederholm 1934) were published in the Annual Report of the Geological Survey.

Age determinations and discussion

The analytic results are presented in Table 4, and a concordia diagram is given in Fig. 4. The sampling sites may be found on the index map (Fig. 1), and brief descriptions of the rock samples are given in Appendix 1.

As in the cases of the Wiborg massif and the minor granite bodies, there is by and large no significant difference between the results of the continuous diffusion and episodic lead loss models. Only in the case of the porphyry dike (A715) can a difference be noted. The material presented suggests clearly that there are granitic rocks of at least three different ages among the post-orogenic rocks of the Åland area.

The oldest category consists of the third-group granites of Sederholm, for which the mean of the ages calculated according to the Wasserburg model is 1 813±14 Ma. An episodic lead loss interpretation would yield the same upper intersect, the lower one being about 250 Ma. The range of the ages calculated according to the Wasserburg model is 1 800—1 836 Ma, and no values have been recorded in the 1 820—1 830 Ma range. Thus one could envisage two subgroups, one of 1 830—1 840 Ma

Table 4. U—Pb analyses on

Sample No.	Index No.	Locality	Rock type	Analyst & year	^{238}U ppm	Radiogenic lead (ppm) ^{206}Pb	^{207}Pb	^{208}Pb
A334	30	Åva	Monzonite	OKo, 70^{xy}	216.7	55.27	6.07	7.86
A335	30	Åva	Granite	OKo, 70^{xy}	428.5	88.81	9.58	14.12
A440	38	Ytterklobb	Lemland granite	OKo, 72^{z}	557.3	115.82	12.63	17.32
A441	32	Söderharn	Kökar rapakivi	OKo, 72^{z}	1 855.5	224.07	20.77	22.29
				OKo, 72^{z}	585.3	88.71	8.45	10.06
A442	34	Sälskär	Mosshaga granite	OKo, 72^{z}	392.2	79.46	8.63	12.15
A443	34	Vänskär	Mosshaga granite	OKo, 72^{z}	378.7	84.62	9.16	12.53
A444	35	Fjälskär	Rapakivi granite	OKo, 72^{z}	613.4	120.79	12.15	33.55
A467	36	Eckerö	Rapakivi	MVa, 75	370.5	58.07	5.66	7.32
A468	37	Lumparland	Even-grained granite	MVa, 75	376.6	70.27	6.79	9.51
A508	31	Kummelskär	Seglinge granite	MVa, 65	835.4	145.37	15.36	15.69
A509	31	Kummelskär	Seglinge granite	MVa, 75	225.3	57.13	6.27	9.96
A510	31	Kummelskär	Seglinge granite	MVa, 75	939.3	158.32	16.60	16.34
A715	49	Blåklobb	Quatrz porphyry	MVa, 76	673.8	133.37	12.91	20.57
				MVa, 76	1 211.3	226.13	21.63	37.34
				MVa, 76	983.7	176.46	16.61	30.07

Analysts: Matti Vaasjoki (MVa) and Olavi Kouvo (OKo).
^{x}Previously published in the Annual Report of Activities (1970), recalculated.
^{y}Previously published in Neuvonen (1970), recalculated.
^{z}Previously published in the Annual Report of Activities (1972), recalculated.

and another of 1 800—1 820 Ma. On the other hand, the discrepancies may as well be ascribed to analytic errors. This must be accepted as the more likely explanation, since there are no geologic grounds known to the present author to support an assumption that the granite of Lemland (A440) and part of the Mosshaga granite (A442) would be definitely older than the other third-group granites in the Åland area.

The main type of Åland rapakivi (A467) and the rapakivi granite of Fjälskär (A444) are of the same age as the main phases of the Wiborg rapakivi massif, In the Åland area, especially in view of the paucity of analyses, the author considers 1 670$\pm$ 20 Ma to be a reasonable age estimate for the main parts of the rapakivi massif.

The Kökar massif and the greisen-bearing, even-grained rapakivi granite of Lumpaıland form another group within the analytic results. They can be plotted within their margin of analytic uncertainty on the 1 620 Ma continuous diffusion trajectory derived from the Wasserburg model.

The three analyses performed on sample A715 (Blåklobb) do not conform to the Wasserburg model. The model ages become increasingly younger as the discordancy of the analyses increase. A least-squares fitted line yields an upper intresect of 1 662$\pm$ 7 Ma, and a lower one of 568$\pm$17 Ma. As in the Wiborg massif, the porphyry dikes seem to be as old as the main phases of the rapakivi proper.

The data presented here demonstrate that in the Åland area the third-group granites of Sederholm are no doubt about 150 Ma older than the rapakivi massifs. As far as the Åva intrusives are concerned, the analyses of the granite and the monzonite yield zircons from the Åland area.

Total Pb ppm	$^{206}Pb/^{204}Pb$	Radiometric ages $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{206}Pb/^{207}Pb$	Wasserb	Fraction density/mesh size
78.16	429	1 664	1 720	1 788	1 803	Total
122.10	623	1 383	1 537	1 755	1 815	Total
170.50	335	1 387	1 548	1 775	1 836	+3.8/+150
280.80	1 160	841	1 036	1 472	1 623	3.8—4.0/+100
116.92	648	1 040	1 208	1 523	1 617	+4.0/+100
104.99	1 155	1 355	1 524	1 766	1 831	+4.1/+150
110.13	1 515	1 480	1 599	1 760	1 800	+4.1/+150
302.63	77.1	1 321	1 442	1 625	1 671	Total
77.44	635	1 072	1 248	1 566	1 659	+4.2
111.92	206.5	1 258	1 370	1 550	1 594	+4.2
184.03	1 305	1 181	1 385	1 716	1 813	+4.2
73.37	3 935	1 656	1 714	1 786	1 801	+4.1/—100
210.69	565	1 146	1 355	1 702	1 804	+4.1
176.74	943	1 328±31	1 417±21	1 554± 9	1 588±12	+4.1/—70
318.46	482	1 259±32	1 364±22	1 532±10	1 574±14	3.8—4.1/—70
258.68	357	1 214±18	1 322±15	1 502± 8	1 546±10	3.6—3.8/—70

Calculations with the decay constants of Jaffey *et al.* (1971).
All analyses corrected for blank.
Common lead corrected according to G 272 (Table 10).

the same age within the limits of analytic error. Since field evidence indicates beyond any doubt that the monzonite is the older of the two, it must be concluded that their emplacement occurred during a relatively short period of time.

The Vehmaa massif

Outline

The Vehmaa massif is situated between the Åland and Laitila rapakivi massifs. One special study on the area has been published (Kanerva 1928), and it is briefly mentioned in several other papers dealing mainly with the surrounding migmatites of the Svecofennian formations (e.g., Sederholm 1934, Härme 1960).

The massif consists mostly of a biotite-hornblende rapakivi similar to the main type occurring in the Laitila massif. The ovoid crystals of potassium feldspar are predominantly pyterlitic, somewhat larger and less densely packed than in the Åland area. Kanerva (1928, p. 12) classifies also the coarse, porphyritic biotite rapakivi of Lokala as belonging to the main type contained in the massif, although he describes a sharp contact between the two major varieties and assumes the contacts to be sharp also in the areas where they are not exposed.

The third major variety found in the Vehmaa massif is the even-grained topaz-bearing biotite granite occurring around the villages of Uhlu and Helsinki (not to be confused with the capital of Finland.) In the eastern paıt of the massif, at Uhlu, the

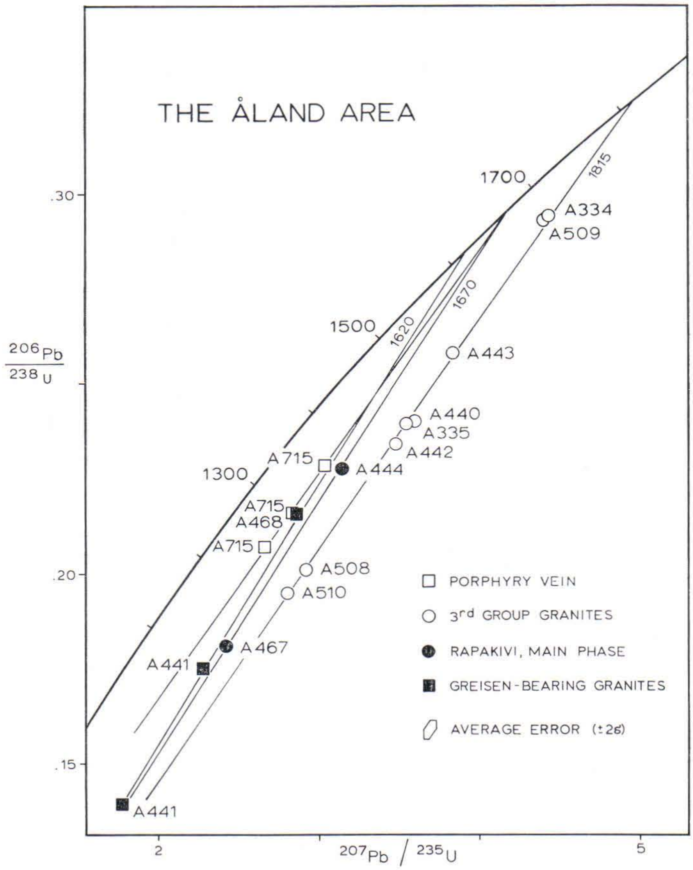

Fig. 4. The concordia diagram of analyses from the Åland area.

rock forms a roundish bulge in the Svecofennian formation. In the proximity of the contact against the migmatites in a road cut along the Raisio—Kustavi highway, it exhibits autoliths of a more mafic nature. The contact relationships between the even-grained biotite granite and the normal rapakivi of the Vehmaa massif are not known to the present author. Kanerva (1928, p. 4) states, apparently for structural reasons, that the even-grained biotite granite is the younger of the two.

No previous radiometric age determinations from the Vehmaa massif are known to the present author.

Age determinations and discussion

The results of the U-Pb analyses are presented in Table 5, and a concordia diagram deduced from the analytic data is shown in Fig. 5. The sampling sites appear on the index map (Fig. 1, p. 17) and short descriptions of the samples collected are given in Appendix 1.

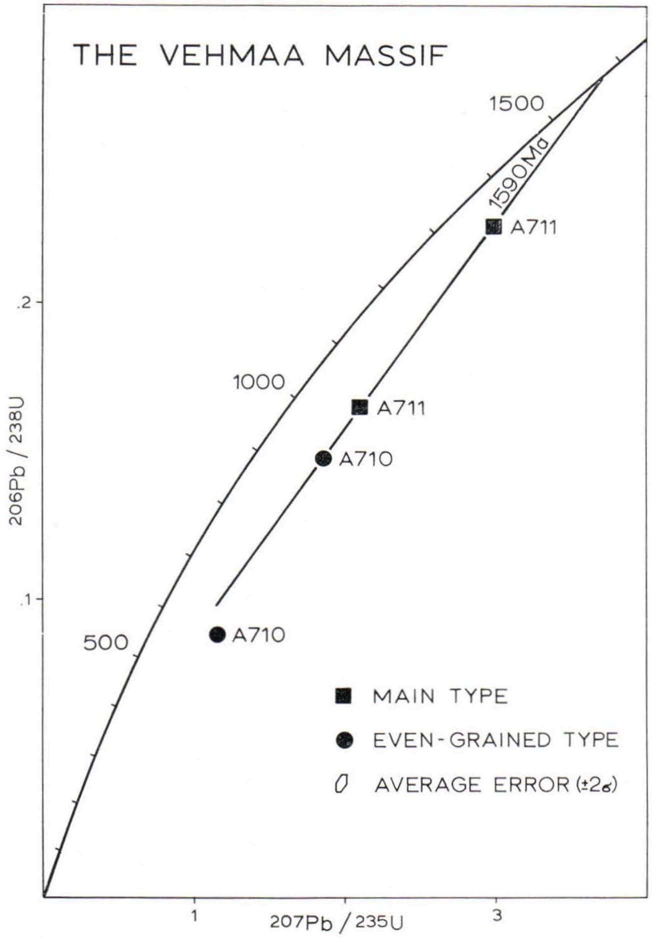

Fig. 5. The concordia diagram of analyses from the Vehmaa massif.

The analyses made of the normal Vehmaa rapakivi (A711) are in close agreement, the ages according to the Wasserburg model being virtually the same. The lighter fraction from the even-grained type (A710) can likewise be plotted on the 1 590 Ma continuous diffusion trajectory. The heavier fraction of A710 does not fall on this curve. A least squares fitted line for all four analyses would produce upper and lower intersects with the concordia curve suggesting episodic events at 1 531 and 64 Ma, respectively.

Sample A710 is the only case in the entire analytic material presented in this study where the heavier and less uraniferous fraction is significantly more discordant than the lighter fraction. The analytic results show that the heavy fraction is also the sole

Table 5. U—Pb analyses on zircons

Sample No.	Index No.	Locality	Rock type	Analyst & year	^{238}U ppm	Radiogenic lead (ppm) ^{206}Pb	^{207}Pb	^{208}Pb
A710	46	Uhlu	Even-grained granite	MVa, 76	868.7	66.05	6.37	112.14
				MVa, 76	2 425.4	308.32	28.69	97.81
A711	45	Vehmaa	Rapakivi	MVa, 76	291.3	56.84	5.50	9.75
				MVa, 76	888.3	126.85	11.85	16.83

Calculations with the decay constants of Jaffey *et al.* (1971).
All analyses corrected for blank.

case in this study where the thorogenic ^{208}Pb is the most abundant lead isotope. Since separate grains of xenotime have been found in this rock as well (determination by X-ray diffraction), it seems reasonable to assume that in the case of the heavy fraction of A710, the material analyzed consisted of crystal intergrowths of the isostructural minerals zircon and xenotime in the same manner as is reported by Grünenfelder *et al.* (1968). The high ^{208}Pb-content of the mixture is a result of the tendency of the latter mineral to admit Th^{4+} into its Y-sites.

The discordancy characteristics of mixed mineral systems are known to a lesser extent than the behaviour and interpretation of discordant zircon data. Considering the analysis of the heavy fraction of A710 as of equal importance with the three other analyses from the Vehmaa massif would invalidate an interpretation resulting in an age of around 1 590 Ma (it is immaterial whether the continuous diffusion or episodic lead loss model is applied to the three points.) In the author's judgment, it cannot be feasible to allow an analysis made from uncertain material to affect the interpretation of three other analyses of material whose discordancy characteristics are much better established. On the other hand, deviating data should not be totally ignored simply because of the deviation.

A third zircon population was also separated from sample A710, which represents the even-grained granite of the Vehmaa massif. While most rapakivi zircons are subhedral or euhdral, this one exhibits a rounded morphology resembling the suites of oval zircons described from the Lewisian granulites of Scotland by Pidgeon and Bowes (1972). The crystal forms are only occasionally barely discernible and the mineral bears a strong resemblance to older cores encountered at times in the rapakivi zircons (p. 14). The yield from a sample of 20 kg was minimal; only 0.8 mg could be extracted by careful handpicking.[1]) The small amount of the sample material prevented a complete analysis, and therefore only the isotopic composition of the lead was determined. The $^{207}Pb/^{206}Pb$ age thus obtained is 2 750$\pm$50 Ma, suggesting that we are dealing with a remnant of a much older rock, which barely escaped total resorption during the development of the rapakivi magma.

[1]) Since the crushing procedures involve cleaning of the mills by quartz and no samples from the Presvecokarelian basement areas (age 2600—2900 Ma) were handled in the laboratory during the treatment of A710, laboratory contamination is an unlikely possibility.

from the Vehmaa rapakivi massif.

Total Pb ppm	$^{206}Pb/^{204}Pb$	Radiometric ages $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{206}Pb/^{207}Pb$	Wasserb	Fraction density/mesh size
200.39	302.7	542±15	783±16	1 549±10	NC	+4.2
477.06	520	883±18	1 072±15	1 480± 9	1 618±16	3.8—4.0/—100
72.91	4 750	1 310±17	1 405±12	1 552± 6	1 588±10	+4.2
162.96	1 194	984±20	1 153±16	1 487±10	1 587±12	4.0—4.2

Common lead correction according to G63 (Table 11).
Analyst: Matti Vaasjoki (MVa).
NC = not calculated.

In the light of the foregoing discussion, it may be concluded that at least a portion of the Vehmaa massif originated from assimilated and remelted material, parts of which have been 2 750 Ma old. The emplacement of the massif occurred most likely 1590±15 Ma and definitely not later than 1530 Ma ago.

The Laitila massif

Outline

The Laitila massif has been incorporated by various authors (Sederholm 1911, Laitakari 1925, Kahma 1951) in the Jotnian formation of Satakunta. This formation includes, besides the rapakivi granites, an arkosic sandstone (Sederholm 1911, A. Laitakari 1925, Kahma 1951, Simonen and Kouvo 1955), and an olivine diabase (A. Laitakari 1928, Kahma 1951), which cuts across the two other rocks. No visible contacts exist between the rapakivi and the sandstone, but observations from analogous formation in Sweden (e.g., Geijer 1963) and in the Soviet Union (e.g., Polkanov 1956) demonstrate that the sandstone is the younger of the two. Therefore, the Laitila rapakivi massif (as well as other rapakivi massifs in southern Finland) is nowadays often referred to as a Subjotnian formation.

Texturally the normal rapakivi of the Laitila massif is pyterlitic, ovoid crystals of potassium feldspar without an oligoclase mantle being more abundant than mantled ones. Biotite and hornblende usually occur together as the mafic minerals, but varieties containing only biotite are occasionally met with. In the eastern part of the massif, even-grained biotite granites (the Lellainen granite, Vorma 1976) are met with, and in the southern part there occur two bodies of a mainly porphyritic biotite granite (the Ytö granite, Vorma 1976), exhibiting at its contacts large phenocrysts of potassium feldspar. In the northern part of the massif, northeast of Eurajoki, there occurs a composite stock containing two granites, even-grained Tarkki granite and topaz-bearing, in many cases porphyritic Väkkärä granite (Laitakari 1928, Kahma 1951, Haapala 1974 and 1977). Both rocks of the Euıajoki stock contain tin-bearing greisen formations similar to those encountered within the Kymi cupola of the

Wiborg massif (Haapala and Laajoki 1969, Haapala and Ojanperä 1973, Haapala 1974 and 1977). On the southeastern border of the Laitila massif, there occurs a small, elongate anorthosite body striking in a northeasterly direction (Hietanen 1947).

The contact relationships between the various granites of the Laitila massif are, thanks to recent mapping, quite well established. Contacts between the normal Laitila rapakivi and the Lellainen granite are at times quite sharp, but often a gradation from normal rapakivi over a hornblende-bearing variety into the Lellainen granite may be observed. Vorma (1976), on the basis of extensive petrochemical material, has suggested that the normal rapakivi and the Lellainen granite belong to the same intrusive phase.

The contacts between the Ytö granite and the Laitila rapakivi are quite sharp in all instances. The contact seems to dip gently below the normal rapakivi, and there is a marked orientation of the feldspar laths of the Ytö granite parallelling the contact. Because of the contact relationships and petrochemical differences, Vorma (1976) has postulated that the Ytö granite and related rocks represent a younger intrusive phase than the normal rapakivi of the Laitila massif.

The age relationships within the Eurajoki stock are clear: the Väkkärä granite brecciates the Tarkki granite, being thus indisputably the younger. When visible, the contact between these rocks dips gently below the Tarkki granite, the Väkkärä granite thus forming an evidently cupola-shaped intrusion in it. In the Väkkärä granite, the contact varieties are usually even-grained, the centre of the intrusion being mostly porphyritic (Haapala 1974 and 1977). Two sets of porphyry dikes, perpendicular to each other, traverse the Tarkki granite. On one occasion, a dark porphyry dike (Haapala 1977) was seen to cut the marginal part of the Väkkärä granite. The porphyry dikes are evidently older than the Postjotnian diabases encountered in the area (Haapala and Ojanperä 1973, Fig. 2).

The age relationship between the Tarkki granite and normal rapakivi cannot be deduced from contact phenomena. At the only outcrop where these two rocks may be seen together, the contact is visible for only a few metres and no fragments or gradations may be observed in either rock. Chemical analyses (Vorma 1976, Haapala 1977) establish that the Tarkki granite is the most basic granite within the Laitila massif, and its REE distribution suggests that it was formed at a relatively high temperature (Vorma 1976).

Previous work thus indicates that the Tarkki granite is the oldest intrusive within the Eurajoki stock, and may well be older than the normal rapakivi of the Laitila massif. A previous age determination from the Tarkki granite indicates an age of about 1 600 Ma (Geological Survey, Annual Report 1969). Within the Väkkärä granite, the porphyritic varieties seem to constitute the last differentiates considerably enriched in a hydrous phase (Haapala 1977).

Concerning the general geologic position of the Laitila massif, two comments should be made. First, the interpretation of gravimetric data by Laurén (1970) indicates that the Laitila massif extends several kilometres below the present level of erosion. Second, the presence of a breccia in the middle of the massif (Vorma 1976) suggests that the present level of erosion is quite close to the roof of the intrusive. Apparently, the contrary opinion of A. Laitakari (1928, p. 7) was based on differences in mapping (compare *op. cit.*, Fig. 1, and Haapala and Ojanperä 1973, Fig. 2).

The Jotnian sandstone of Satakunta is situated in a graben striking northwest in the direction of Kokemäenjoki. The existence of the graben is well established by gravimetric and seismic data as well as by deep drilling (cf., Laurén 1970). In a recent report (Elo 1976), the thickness of the sandstone formation at Pori was calculated to be 1 300—1 800 m. This figure also represents the minimum vertical displacement along the graben of Satakunta. Simonen (1960) has reported K-Ar ages from micas of the Jotnian sandstone ranging from 1 150 to 1 300 Ma.

Age determinations and discussion

The results of the U-Pb analyses of zircons are given in Table 6, and a concordia diagram deduced from the results is presented in Fig. 6. The sampling sites may be found on the index map (Fig. 1, p. 17), and short descriptions of the samples collected are given in Appendix 1.

Compared with the results from the other rapakivi massifs, the Laitila massif deviates markedly in one respect: the discordancy pattern cannot be accounted for by the continuous diffusion model of Wasserburg. The majority of the analyses are plotted on a straight line intersecting the concordia at 1 573$\pm$8 Ma and at 70$\pm$89 Ma. No measurable age difference can be detected between the Tarkki granite (A255 and A606) and the main rapakivi type of the Laitila massif (A608 and A689). The gabbro-anorthosite at Kolinnummi (A691) is also contemporaneous with the main part of the massif.

The two analyses from the Ytö granite (A690) are likewise plotted on a straight line parallel to the main trend but intersecting the concordia curve at 1 540 and 105 Ma. Since the heavy zircon fraction is concordant within its limits of analytic error, and the geologic observations suggest a younger origin of the Ytö granite, the age difference must be regarded as real.

The age of the porphyritic potassium granite of Kokemäki (Peipohja) coincides with the age of the major parts of the Laitila massif, 1 573$\pm$8 Ma. This result clearly demonstrates that the granite, situated on the northeastern border of the Satakunta graben, belongs to the post-orogenic suite of southern Finland. The existence of rocks of similar age on both sides of the Jotnian sandstone also sets an upper time limit for the formation of the Satakunta graben.

The zircons of the topaz-bearing variety of the Väkkärä granite register an unusually high content of common lead, and thus its composition may affect the results of the age calculations (see p. 8). It was thought proper not to carry out the correction for initial lead by the standard method of employing a nearby least

Table 6. U—Pb analyses on zircons

Sample No.	Index No.	Locality	Rock type	Analyst & year	^{238}U ppm	Radiogenic lead (ppm) ^{206}Pb	^{207}Pb	^{208}Pb
A129	26	Kokemäki	Porphyritic granite	OKo, 70[x]	310.4	67.90	6.59	13.14
				MVa, 75	499.2	77.77	7.46	11.47
A255	27	Tarkki	Tarkki granite	MSa, 68[y]	122.9	28.16	2.76	6.12
				MVa, 75	119.8	26.67	2.62	5.68
				MVa, 75	126.6	28.17	2.79	5.89
A515	27	Linnamaa	Väkkärä granite	MVa, 74	215.4	29.46	2.85	6.94
A606	27	Kylämaa	Tarkki granite	MVa, 75	224.3	52.66	5.11	11.04
				MVa, 75	197.0	39.08	3.80	7.60
A607	27	Koivuniemi	Väkkärä granite	MVa, 75	671.4	103.08	9.56	30.90
A608	28	Untamala	Rapakivi	MVa, 75	214.3	48.42	4.74	9.93
				MVa, 75	698.5	144.52	14.03	30.71
				MVa, 75	617.2	131.91	12.79	27.35
A609	27	Heikkilä	Väkkärä granite	MVa, 76	2 264.5	97.83	8.00	35.87
A686	27	Metsäranta	Väkkärä granite	MVa, 76	468.5	79.48	7.77	46.97
				MVa, 76	964.1	144.74	13.77	40.88
A687	27	Koivuniemi	Väkkärä granite	MVa, 76	1 615.1	177.26	16.05	37.47
				MVa, 76	2 326.6	130.27	10.87	34.00
A689	42	Hinnerjoki	Rapakivi	MVa, 76	335.2	71.68	7.01	13.34
				MVa, 76	900.4	159.66	15.23	24.53
A690	43	Kusni	Ytö granite	MVa, 76	307.9	71.94	6.90	38.65
				MVa, 76	1 124.9	161.00	15.18	41.88
A691	44	Kolinnummi	Gabbro-anorthosite	MVa, 76	380.2	88.66	8.72	24.98
A707	27	Pistola	Dark porphyry	MVa, 76	220.6	48.15	4.66	8.97

All calculations with the decay constants of Jaffey *et al.* (1971).
All analyses corrected for blank.
Common lead corrections with either the mean of the samples G259, G260 and G274 (Table 11) or with the lead of the respective potassium feldspar (Table 7).
NC = not calculated.

radiogenic lead (in this case, the mean of the galena compositions of the local greisen mineralizations), but by using the lead of the potassium feldspars corrected for their uranium-derived portions. The results of the analyses are given in Table 7. Since no separate analysis for thorium was carried out, and the Th content of feldspars is known to be low, no correction was applied to the ^{208}Pb-data.

With the composition of common lead ascertained as closely as possible, it must be concluded that the analyses obtained from the Väkkärä granite yield two kinds of results: 1) the even-grained and porphyritic varieties situated mostly at the contacts of the Tarkki granite are of an age similar to that of most of the other rocks of the Laitila massif, 2) the topaz-bearing porphyritic varieties of the Väkkärä granite are plotted on a straight line intersecting the concordia curve at 1 501$\pm$11 Ma and at 109$\pm$10 Ma. Thus the petrographic differences between the various parts of the Väkkärä granite seem to be reflected in the zircon age data as well.

The marginal and topaz-bearing varieties of the Väkkärä granite differ, however, both mineralogically and chemically less from each other than from the Tarkki granite or from the normal rapakivi of the Laitila massif (Haapala 1974 and 1977). Among the rocks sampled, there exist complete chemical analyses of the even-grained (A515), porphyritic (A607) and coarse-grained (A687) types of the Väkkärä granite (Haapala from the Laitila rapakivi massif.

Total Pb ppm	$^{206}Pb/^{204}Pb$	Radiometric ages $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{206}Pb/^{207}Pb$	Wasserb	Fraction density/mesh size
99.1	413.6	1 452	1 497	1 560	1 570	Total
105.99	591	1 066	1 233	1 536	1 629	+4.2/+100
52.31	418.3	1 514	1 539	1 574	1 581	Total
36.08	2 275	1 475±15	1 518±18	1 579± 2	1 595±10	+4.2
37.39	2 254	1 475±17	1 525±11	1 594± 3	1 610±12	3.8—4.1
47.97	378.5	945	1 147	1 549	1 680	Total
72.08	1 135	1 547±39	1 552±19	1 559± 9	1 560±15	+4.2/—200
52.25	1 545	1 330±17	1 422±12	1 563±10	1 596±12	+4.2/+100
224.18	105.3	1 053±31	1 199±31	1 474± 9	1 552±14	Total
66.94	3 490	1 495±25	1 528±16	1 574±20	1 584±25	+4.2/100—200
191.06	5 555	1 382±21	1 453±15	1 560±10	1 595±13	4.0—4.2/—200
174.65	2 430	1 423±24	1 477±18	1 557±13	1 575±15	4.0—4.2/+100
284.22	64.05	314±35	452±25	1 230± 5	NC	Total
136.20	2 810	1 154±23	1 308±17	1 571± 4	1 645±10	+4.2
205.99	1 533	1 031±25	1 201±19	1 521± 5	1 614±10	4.0—4.2
441.83	75.01	770±29	961±26	1 437± 5	1 602±12	4.0—4.2
307.60	84.98	404±14	564±13	1 270± 5	NC	3.8—4.0/—150
92.63	8 250	1 423±11	1 484± 8	1 572± 7	1 593±10	+4.2
209.51	1 107	1 201±24	1 323±18	1 526±15	1 579±17	3.8—4.0
125.25	758	1 541±15	1 539±10	1 536± 5	NC	+4.2/—150
235.14	519	986±17	1 161±14	1 504± 6	1 608±25	3.8—4.0/—150
122.08	14 380	1 538±32	1 557±18	1 584± 2	1 603±10	+4.1
65.23	984	1 450±14	1 493± 9	1 555± 5	1 576±10	Total

Analysts: Matti Vaasjoki (MVa), Olavi Kouvo (OKo) and Matti Sakko (MSa).
[x]Previously published in the Annual Report of Activities (1970).
[y]Previously published in the Annual Report of Activities (1968).

Table 7. Feldspar lead compositions from the Laitila rapakivi massif.

Sample	Index No.	$^{206}Pb/^{208}Pb$	$^{206}Pb/^{204}Pb$	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
A686	27	.45909±.00005	17.095±.001	1.1029±.0002	15.500±.004	37.24±.01
A687	27	.45643±.00005	16.397±.008	1.0641±.0003	15.409±.012	35.92±.02
A688	27	.55600±.00030	20.501±.005	1.3217±.0004	15.511±.008	36.87±.03
A689	42	.45331±.00007	16.346±.002	1.0593±.0007	15.430±.012	36.06±.01

All ratios normalized using the CIT-standard (Catanzaro 1967).

1974, p. 164, 165, analyses 3, 6 and 8). From these, it seems that 1) iron in the topaz-bearing types is moie readily oxidized to the trivalent state, 2) the topaz-bearing types contain about twice as much rubidium and half as much stronitum as the even-grained Väkkärä granite, and 3) the zirconium contents of the porphyritic variety range from very low to nil — certainly not a typical feature of rapakivi granites.

The last-mentioned difference is reflected in the yields of zircon from the samples. The porphyritic varieties invariably proved to be quite poor in zircon as compared to the even-grained varieties. In one case, a 60 kg sample designated A688 (Haapala 1974, analysis No. 7, p. 165) did not yield enough zircon to facilitate an age determina-

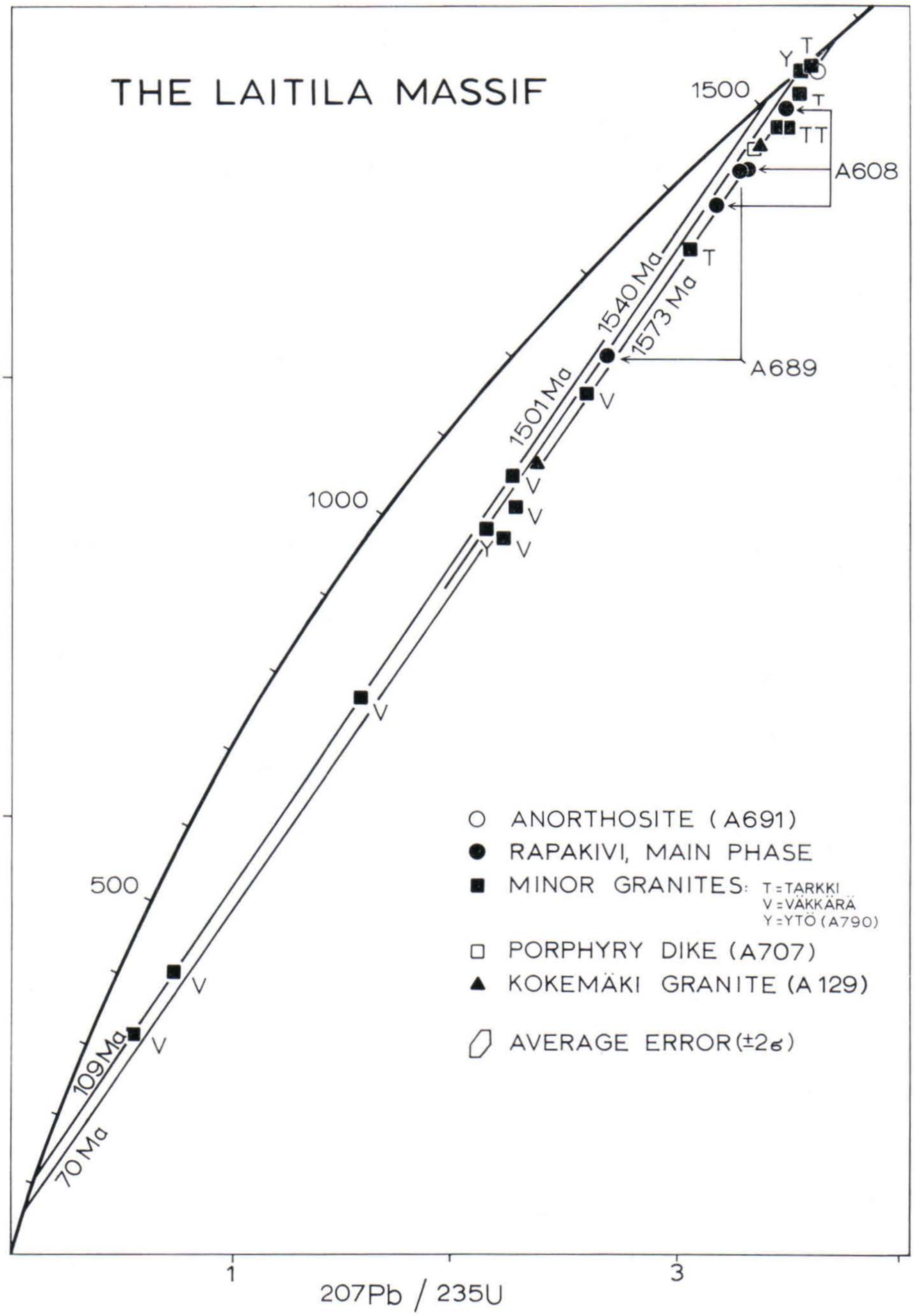

Fig. 6. The concordia diagram of analyses from the Laitila massif.

tion. The zirconium content of this rock has been reported as zero. The tin granite nature of the Väkkärä granite (*op.cit.*) was reflected in the fact that the heavy (d > 4.2) non-magnetic fractions of samples from the topaz-bearing types always contained cassiterite, which had to be removed by hand-picking before analysis of the zircons.

As far as age determinations are concerned, it should be noted that though petrographic and chemical data suggest that the topaz-bearing types of the Väkkärä granite are the last differentiates in the intrusive (Haapala 1977), it is hard to visualize a 70 Ma difference within the stock when there is no measurable age difference between its mariginal part and the surrounding rocks. Sample A707 from a dark porphyry dike, which intersects both the marginal part of the Väkkärä granite and the Tarkki granite, registers an age comparable to the majority of the analyses from the Laitila massif, thus indicating that the ages obtained from the marginal part of the Väkkärä granite cannot be erroneous. On the other hand, no dikes have been observed to cut the topaz-bearing types of the Väkkärä granite, so the result obtained from A707 has no direct bearing on the conclusions concerning the U-Pb analyses yielding an age of 1 501 Ma.

To clarify the results obtained from the U-Pb deteminations of zircons from the Laitila massif, one sample from the Tarkki granite, four samples of normal rapakivi and five samples of Väkkärä granite weie analyzed for their whole rock lead isotopic compositions. The analytic results are presented in Table 8 and the data, along with the feldspar and galena determinations, are plotted on a $^{206}Pb/^{204}Pb$ vs. $^{207}Pb/^{204}Pb$ diagram in Fig. 7.

Of the whole rock samples, those representing the normal rapakivi of the Laitila massif form a straight line whose slope (method of York 1966) yields an age of 1 628 ± 60 Ma. In view of the considerable geographic distances separating the samples, the formation of an isochron demonstrates that the lead isotopic composition within the initial material of the Laitila massif had been quite homogeneous, possibly as the result of the formation of a rapakivi magma in the infrastructure of the earth's crust. With some speculation, the slightly higher age as compared to the zircon age determinations could be taken as the beginning of the ascent of the magma to higher levels, which resulted in the formation of several subsystems with differing $^{238}U/^{204}Pb$ (μ) values.

The single analysis of the Tarkki granite does not fall into the same array with either the normal rapakivi or the Väkkärä granite, indicating thus that it may have a different parent magma or that its material is an extreme differentiate of the main rapakivi magma. A single analysis does not warrant any conclusions as to whether a possible differentiate has been an early or a late one.

The five whole rock analyses from the various types of Väkkärä granite can be plotted on a straight line, the slope of which yields either an age of 115 Ma (least-squares fit) or a modern age (method of York 1966). At first, the result seems striking, since the discordances of the zircon data of the Laitila massif cannot be accounted for by a continuous diffusion model, but rather suggests an episodic loss of lead at 95 ± 21 Ma (mean of the ages indicated by the lower intersects). A closer scrutiny

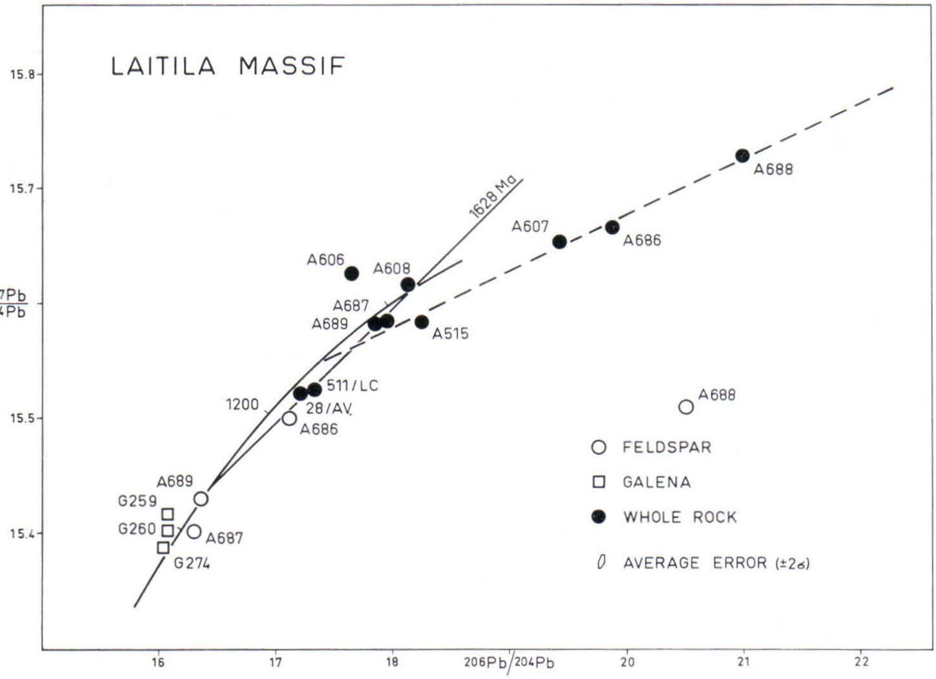

Fig. 7. The common lead analyses from the Laitila massif. The growth curve is drawn according to the two-stage model of Stacey and Kramers (1975) with $\mu_2 = 9.8$.

of the data reveals, however, that the »isochron» produced by the whole rock lead compositions of the Väkkärä granite is merely an accidental distribution.

From the theoretical point of view, whole rock samples should represent closed systems, the lead isotopic composition of which has changed only as the result of radioactive decay of uranium and thorium. The last resetting of the system should be indicated by the slope of an eventual isochron. As can be seen from Table 8, the uranium contents of the various types of Väkkärä granite are so small that the amount of lead required for the formation of the present isotopic ratios of sample A688 from the lead composition presented by sample A687 during the last 115 Ma is impossible. Moreover, the feldspar lead compositions demonstrate that there had been a large variation of the initial lead isotopic ratios, probably arising from differentiation processes (Haapala 1977) within the Väkkärä granite at the time of its formation. The lead now present has evolved from the original lead represented by the feldspars during some 1 500—1 600 Ma, and the formation of the phantom »isochron» can be ascribed to the different μ values of the various types of Väkkärä granite.

It is evident that the common lead work carried out does not help in the matter of the contradicting zircon age data obtained from the Väkkärä granite. With no other facts available, the following interpretation is a purely theoretical one, and must be regarded as a tentative solution.

Table 8. U—Pb determinations of whole rock samples from the Laitila rapakivi massif.

Sample	Index No.	$^{206}Pb/^{208}Pb$	$^{206}Pb/^{204}Pb$	$^{206}Pb/^{207}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Pbtot*	Utot*	μ
A515	27	.44755 ± .00001	18.249 ± .021	1.1710 ± .0004	15.584 ± .023	40.78 ± .05	—	—	—
A607	27	.48089 ± .00026	18.429 ± .017	1.2412 ± .0009	15.653 ± .025	40.40 ± .06	29.66 ± .09	2.93 ± .02	6.53
A686	27	.43789 ± .00005	19.872 ± .004	1.2687 ± .0002	15.663 ± .006	45.38 ± .01	15.77 ± .12	1.85 ± .01	8.34
A687	27	.48092 ± .00007	17.941 ± .002	1.1512 ± .0001	15.585 ± .003	37.31 ± .01	51.43 ± .05	5.78 ± .04	6.98
A688	27	.53683 ± .00003	20.991 ± .003	1.3346 ± .0003	15.728 ± .006	39.10 ± .01	13.16 ± .01	3.24 ± .02	16.35
A606	27	.46057 ± .00004	17.637 ± .010	1.1287 ± .0001	15.626 ± .010	38.29 ± .03	—	—	—
A608	28	.45375 ± .00025	18.118 ± .038	1.1601 ± .0001	15.618 ± .034	39.93 ± .10	—	—	—
A689	42	.46534 ± .00015	17.839 ± .012	1.1449 ± .0001	15.581 ± .012	38.34 ± .04	—	—	—
511/LC	47	.45736 ± .00001	17.321 ± .006	1.1156 ± .0002	15.526 ± .008	37.87 ± .01	—	—	—
28/AV	48	.44996 ± .00007	17.203 ± .006	1.1083 ± .0003	15.522 ± .010	38.23 ± .02	—	—	—

*ppm
Lead ratios normalized using the CIT-standard (Catanzaro 1967).

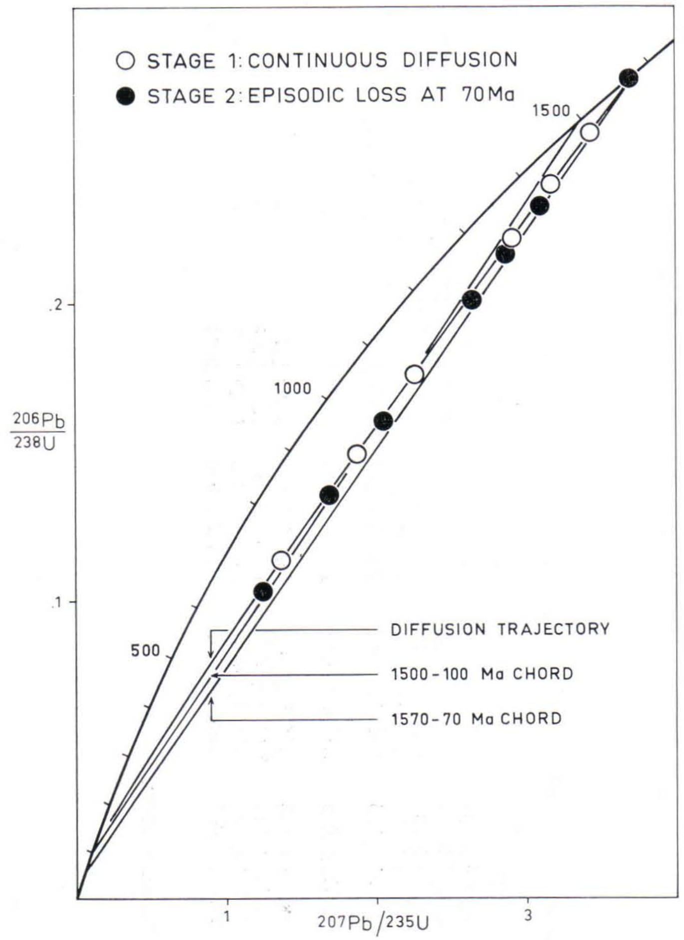

Fig. 8. Schematic diagram of the effects of combined continuous and episodic lead losses on a relatively concordant and a very discordant zircon population.

It has been pointed out repeatedly (Shukolyukov 1964, Allègre *et al.* 1974) that lead loss by continuous diffusion combined with a super-imposed episodic event will result in a sublinear distribution wheie neither of the intersects of a best-fitting line with the concordia will have any chronologic significance. The drawback of the model is the need for so many unmeasurable parameters that practical calculations result in guesswork. It is helpful, however, when trying to establish possible processes qualitatively.

In Fig. 8 there is an outline of what will happen to a set of zircons formed some 1 570 Ma ago when subjected to various degrees of lead loss by continuous diffusion and a 10 % loss of the remaining lead by an episodic event 70 Ma ago. The distribution thus obtained corresponds strikingly to the analytic results presented in Fig. 6, and it would be even more pronounced if a more recent loss of lead be assumed.

The final conclusion is that most of the Laitila massif was emplaced 1 573±8 Ma ago. The discordancy pattern of the sample material, including the Väkkärä granite, is a result of the combined effects of lead loss by continuous diffusion with a superimposed episodic event 0—70 Ma ago. The age of the Ytö granite and related rocks is 1 540±10 Ma.

ON FACTORS AFFECTING THE DEGREE OF DISCORDANCY OF ZIRCON AGES

The material accumulated during this work supports the old findings that the discordancy of a sample is generally a function of its uranium content (e.g., Wasserburg 1963, Kouvo and Tilton 1966), particularly in the case of cogenetic zircon suites (Silver and Deutsch 1963, Meriläinen 1976), i.e., zircon fractions with differing physical properties (density, magnetic susceptibility, grain-size) separated from the same rock sample. The present material suggests, however, that there may also be some petrologic control superimposed on this universal trend.

Rocks derived from the more basic (and also probably dryer) magmas seem to contain the more concordant zircons. In the sample material of this work, this is well demonstrated by the anorthosite samples A118 and A691 (see also Gulson and Krogh 1975), the fayalite-bearing darkish rapakivi variety tirilite (A18), the Tarkki granite (A255 and A606) and the Subjotnian diabase dike A423. On the other hand, some minor late derivatives of the rapakivi material seem to produce zircons especially susceptible to lead loss (A21, A323, A515, A607, A609 and A687). In the present material, very discordant ages occur in zircons exhibiting abnormally high uranium contents in conjunction with exceptionally abundant common lead. The topaz-bearing type of the Väkkärä granite is a perfect example of this. That a high uranium content does not necessarily lead to extreme age discordancy is best demonstrated by the contact variety of the Wiborg massif (A629), the zircons of which also exhibit very low amounts of common lead.

The crystal structure of zircon and especially the coordination of the Zr^{4+}-ion within it has been known to be of a singular nature for a long time (e.g., Machatski 1941). It is established that the Zr-ion is coordinated with eight oxygen atoms, four of which lie closer to it (2.15 Å) and four of which lie farther away (2.29 Å) (Deer *et al.* 1962, Vol. 1, p. 59). It is also known that metamict zircons contain larger amounts of water than fresher (and less discordant) ones (Grünefelder 1963, Grünefelder *et al.* 1968). Shukolyukov (1964) has suggested that at least part of the lead loss of zircons is induced by chemical reactions occurring along the micropores of damaged crystals.

In view of the fact that the most discordant samples of the material presented come from the Väkkärä granite — a rock representing the last phases of the rapakivitic bulk material and with an exceptionally high water and metal content, manifested in the numerous greisen mineralizations of the area (Haapala 1974 and 1977) —, the structural characteristics of zircon suggest that the mineral may at its formation absorb lead from residual phases, which results in an initial distortion of its lattice. Thus, to begin with, conditions are favourable for rapid metamictization, which would also result in a larger than average loss of lead. The process is obviously intensified by a greater than average uranium content in the zircon.

COMMON LEAD STUDIES FROM GALENA MINERALIZATIONS

The galena mineralizations of the Wiborg massif and their lead isotopic composition

General description and ore microscopic studies

A number of small sulfide mineralizations occur within the Wiborg rapakivi massif and its immediate surroundings. Although a number of them have been quarried as either silver oı iron ores during modern history, none is at the present of any economic impoıtance. Only three of them (Inkeroinen, Virojoki and Pernaja) are briefly mentioned by Kahma (1973) in his treatise on Finnish mineral deposits. The latest investigations with an economic aim were carried out in the late 1960s, when the Geological Survey prospected for tin in the greisen bodies associated with the even-grained and porphyritic rapakivi granites of Kymi. These investigations did not result in any mining operations, and their negative outcome may have been partly the reason why Isokangas (1976) listed the rapakivi massifs as non-metalliferous areas within the Fennoscandian (or Baltic) shield.

By their mode of occurrence, the galena mineralizations connected with the Wiborg massif may be divided into three broad groups:

1) Ore mineralizations in Svecofennian country rocks in the immediate vicinity of the massif
2) Ore mineralizations associated with greisen bodies
3) Ore veins within the massif, epigenetic in their mode of occurrence.

The sampling localities have been described in a somewhat greater detail in in Appendix 2.

To obtain a more detailed picture of the nature of the ore mineralizations studied, an ore microscopic survey was carried out. The results are summarized in Table 9.

The chief minerals of the ores emplaced in the Svecofennian country rocks are galena, sphalerite and chalcopyrite. In the case of Hyvärilä (G7) and Koskenkylä (G266), the sphalerite contains blebs of chalcopyrite and pyırhotite. At Hogland (G272), the sphalerite is void of inclusions on the microscale, but it is often rimmed by either chalcopyrite or galena.

Table 9. Ore microscopic results from the Wiborg rapakivi massif.

Sample	GAL	SPHA	CP	PY	ASPY	COV	PYRH	FeOx	Location	
G 2	●	●	+	○		+		○	Säkkijärvi	3b
G 3	●	●	+	●		+		+	Luumäki	3b
G 62	●	○	○	+		+			Anjalankoski	3a
G141	●	○	○	+					»	3a
G151	○	●	●					+	Kymi	2
G 7	●	●	●	○			○		Lemi	1
G266	●	●	●	○	+		○		Pernaja	1
G272	●	●	●						Hogland, USSR	1

● Major ○ Minor + Accessory

GAL = Galena; SPHA = Sphalerite; CP = Chalcopyrite; PY = Pyrite;
ASPY = Arsenopyrite; COV = Covellite; PYRH = Pyrrhotite; FeOx = Iron oxides

In the greisen formations of the Kymi cupola, the findings of Haapala and Laajoki (1969) were confirmed. Galena is always associated with sphalerite, which often exhibits pseudoanisotropic zoning and lamellae caused by minute inclusions of chalcopyrite. The latter mineral occurs also as larger blebs in sphalerite and as independent grains.

The epigenetic veins may be divided into two subgroups, differing greatly in their ore paragenesis:

3a: Galena is the predominant ore mineral. Sphalerite and chalcopyrite are the principal accessory ore minerals. Sample G62 (Inkeroinen, Anjalankoski) exhibits no inclusions in the sphalerite, and chalcopyrite in some instances forms a rim around it. At some places the chalcopyrite is partly altered into covellite.

3b: Galena is the chief ore mineral, but the wide occurrence of pyrite and honey-coloured inner reflections of sphalerite are the most conspicuous common features of the samples belonging to this group. Chalcopyrite occurs as fillings in fissures and is invariably associated with secondary covellite. In places, hematite was observed to occur together with spherulitic limonite.

The isotopic compositon of lead in the galenas

The results of the common lead isotopic determinations from the galenas are summarized in Table 10. A $^{206}Pb/^{204}Pb$ vs. $^{207}Pb/^{204}Pb$ diagram is shown in Fig. 9. In Fig. 10, a trilinear ^{206}Pb- ^{207}Pb- ^{208}Pb plot is presented. Both diagrams illustrate that similarities observed between the samples in their mode of occurrence and ore mineralogic properties are reflected in their lead isotopic compsition as well:

— The lead in the samples belonging to group 1 is less radiogenic than in the rest of the sample populations.

Table 10. Lead isotopic compositions from galena mineralizations in the Wiborg rapakivi massif.

Sample	Index	Locality	$^{206}Pb/^{208}Pb$	$^{206}Pb/^{207}Pb$	$^{206}Pb/^{207}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	T	μ_2
G 7	17	Lemi	.44560 ± .00022	1.0225 ± .0005	15.719 ± .009	15.373 ± .008	35.28 ± .02	1813	10.14
G263	16	Pernaja	.44555 ± .00023	1.0241 ± .0009	15.705 ± .008	15.336 ± .021	35.25 ± .04	1767	9.90
G266	16	Pernaja	.44683 ± .00033	1.0377 ± .0006	15.802 ± .012	15.388 ± .021	35.36 ± .05	1771	10.13
G272	15	Hogland, USSR	.44959 ± .00011	1.0352 ± .0001	15.908 ± .004	15.367 ± .005	35.38 ± .02	1653	9.85
G151	10	Kymi	.45136 ± .00020	1.0431 ± .0003	16.094 ± .008	15.429 ± .012	35.66 ± .03	1606	10.04
G261	10	Kymi	.44951 ± .00064	1.0424 ± .0009	16.073 ± .015	15.420 ± .028	35.76 ± .09	1610	10.01
G 62	19	Anjalankoski	.45575 ± .00059	1.0558 ± .0013	16.334 ± .015	15.471 ± .033	35.84 ± .08	1488	10.05
G141	19	Anjalankoski	.45468 ± .00019	1.0525 ± .0004	16.251 ± .007	15.441 ± .013	35.74 ± .03	1504	9.95
G277	16	Pernaja	.45535 ± .00032	1.0523 ± .0002	16.273 ± .006	15.464 ± .008	35.74 ± .04	1524	10.07
G 2	22	Säkkijärvi, USSR	.49414 ± .00012	1.2146 ± .0003	19.171 ± .012	15.783 ± .014	38.80 ± .03	—	—
G 3	21	Luumäki	.49135 ± .00021	1.2085 ± .0007	19.052 ± .015	15.766 ± .022	38.78 ± .05	—	—
G 61a	20	Virojoki	.49835 ± .00084	1.2250 ± .0007	19.346 ± .015	15.793 ± .021	38.82 ± .10	—	—
G 61b	20	Virojoki	.50045 ± .00035	1.2343 ± .0005	19.521 ± .015	15.816 ± .019	39.01 ± .06	—	—
G265a	20	Virojoki	.50386 ± .00025	1.2431 ± .0002	19.686 ± .011	15.835 ± .001	39.07 ± .04	—	—
G265b	20	Virojoki	.50067 ± .00014	1.2355 ± .0009	19.540 ± .013	15.817 ± .022	39.03 ± .04	—	

T_2 and μ_2 calculated according to the two-stage model of Stacey and Kramers (1975).
All results normalized against the CIT-standard (Catanzaro 1967).

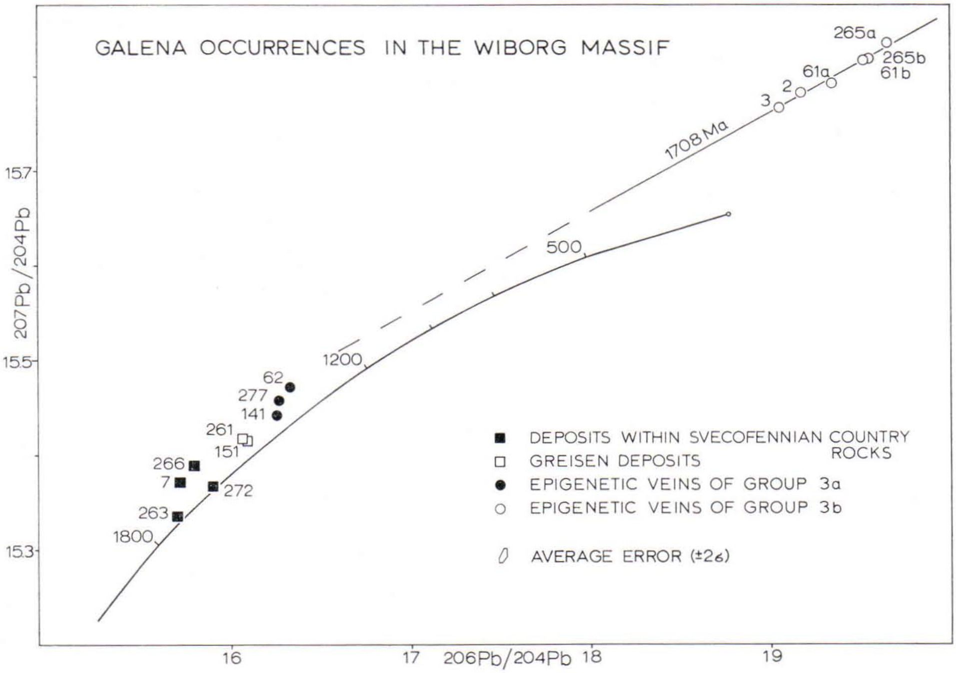

Fig. 9. The common lead analyses from the galena occurrences in the Wiborg rapakivi massif and its immediate environment. The growth curve according to Stacey and Kramers (1975), $\mu_2 = 9.85$.

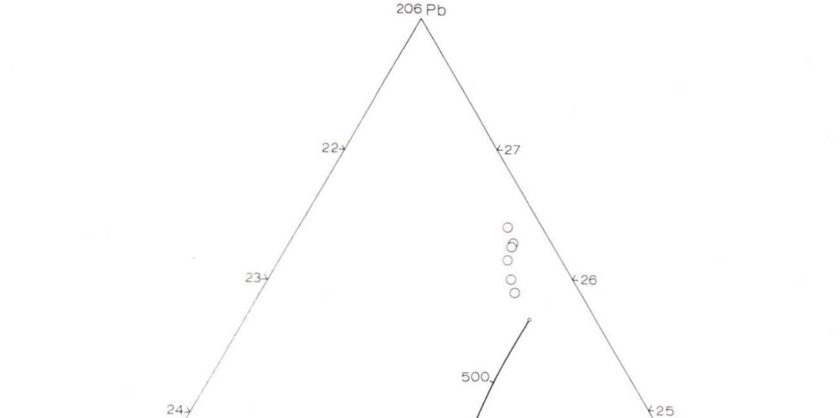

Fig. 10. The relative lead abundances in the galena occurrences of the Wiborg rapakivi massif and its immediate environment. The growth curve according to Stacey and Kramers (1975), $\mu_2 = 9.85$, $W_2 = 36.84$.

- The greisen deposits seem to be somewhat youngei than the mineralizations belonging to group 1.
- The samples belonging to group 3a contain more radiogenic lead than the samples belonging to groups 1 and 2, but are not »anomalous» in the respect that model ages can be computed for them.
- As far as model ages are concerned, the samples of group 3b are »anomalous.» Their radiogenic component consists mostly of the uranogenic isotopes ^{206}Pb and ^{207}Pb. The addition of the thorogenic ^{208}Pb is, as often is the case with radiogenic ore lead, relatively small (e.g., Stanton and Russell 1959).
- Samples G61 and G265 are inhomogeneous with respect to their lead isotopic composition.

The evolution of the lead in the ores of the Wiborg massif

Ore genetic considerations

The mineralizations at Hyvärilä, Lemi (G7) and at Koskenkylä, Pernaja (G263 and G266), are clearly older than the rapakivi. On the other hand, their lead isotopic composition is not typical of the Svecofennian formations (Kouvo 1958, Kouvo and Kulp 1961, Whittles 1962, Aho 1975), and their model ages are not sufficiently high to allow an interpretation involving an origin about 1 850—1 900 Ma ago. Considering the varied mineralogy of especially the Koskenkylä deposits (0. Vaasjoki 1953), it seems probable that the ores were originally formed during the late orogenic phase of the Svecokarelidic cycle, like, for example, the Telkkälä deposit (Häkli *et al.* 1975). Situated in the contact metamorphic aureole of the Wiborg massif, the existence of which has been established by studying the inversion of microcline to orthoclase in the surrounding rocks (Vorma 1972), both the Hyvärilä and Koskenkylä deposits were heated at least up to 350 °C. At the same time, hypogenic lead derived from the rapakivi was added to the ore material, as suggested by O. Vaasjoki and Kouvo (1959). The present lead isotopic compositions reflect this mixing.

Sample G272 from Hogland, USSR, comes from a quartzitic tectonic breccia in Svecofennian migmatites (Ramsay 1892), situated close to effusive rapakivitic rocks (Ramsay 1890, Kranck 1929), and it was probably formed during the period of volcanism represented by the effusive beds. The model age of the sample, 1 653 Ma, is in good agreement with the results obtained from the U-Pb age determinations carried out on zircons from the Wiborg massif. It is therefore feasible to propose that the lead of sample G272 derives from the rapakivitic volcanic rocks and was quenched in the described breccia. Thus this lead should represent the least radiogenic lead derived directly from the rapakivi.

Apperently, the greisen deposits of group 2 were formed from hypogenic solutions derived from the porphyritic and even-grained rapakivi granites of Kymi during the pneumatolytic stage.

In all the cases represented by groups 2 and 3, the ore mineralization took place after the bulk of the granitic material had consolidated. Thus it is obvious that the ores as we observe them today are younger than their host rocks and especially their zircons. On the other hand, the cooling of the rapakivi can hardly have taken 150 Ma, as suggested by the difference of the model ages of samples G272 (Hogland) and G141 (Anjalankoski). The vein mineralizations could, of course, be ascribed to a »deep-seated source.» However, the present minimum thickness of the Wiborg massif has been estimated, on the basis of gravimetric observations, to be 5 km (Laurén 1970). An eventual outside source must therefore be deepseated indeed. The most probable and feasible hypothesis is that the mineralizations found at Anjalankoski derive from their country rock, the rapakivi. Also the slope of the best-fitting line (method of York 1966) for the samples of group 3b yields a maximum age of 1 708 Ma, clearly supporting this assumption. It is thus obvious that the two-stage model used for the model age calculations is inadequate for explaining the lead isotopic composition of the galenas belonging to group 3.

The evolution of the lead in the sample material

When the ore-microscopic observations and the analyses of lead-isotopic composition are considered together, it becomes evident that the ıadiogenic component of the leads increases as the thermal state of the formation of the mineralizations becomes less intense.

With the exception of davidite-bearing uranium ores, most uranium mineralizations are considered to be telethermal or sedimentary in origin. Both of these interpretations presuppose low p,T-conditions during the ore formation. The geochemical behaviour of uranium also favours enrichement into residual solutions (McKelvey *et al.* 1955). A perfect example of this is provided by the leachable uranium of many granitic rocks, which most likely is present as an intergranular film (Larsen *et al.* 1955). It should also be noted that the sulfide mineralizations belonging to group 3b exhibit mineralogical features similar to those of uranium ores belonging to the felsic association (McKelvey *et al.* 1955). No uıanium minerals, however, have been found in the vein deposits of the Wiborg massif.

The two-stage model depicting the evolution of lead under ordinary conditions seems to work well up to the start of the emplacement of the main phases of the Wiborg massif. In the light of the foregoing discussion, the following conclusions seem to be realistic:

— The mineralizations belonging to group 3 crystallized in cracks formed during the cooling of the pluton, possibly as the result of cooling contraction of the solid phases. The sulfide minerals were formed from hypogene solutions deıived from the country rock itself.

— The crystallization temperature of the ore veins decreased in the order G141, G277, G62, group 3b.
— It is highly likely that the uranium, at least to a certain extent, was enriched in the liquid phase responsible for the formation of the sulfide mineralizations. The process has probably also involved the oxydation of U^{4+} into UO_2^{2+}, which remains in solution more readily than U^{4+} or Th^{4+} (McKelvey *et al.* 1955, Hostetler and Garrels 1962). This caused a geochemical separation of the uranium and thorium, which is reflected in the relatively low content of ^{208}Pb found in the galenas of the group 3b veins.

It is therefore reasonable to suppose that the vein deposits were formed during a third stage, which is characterized by a steadily increasing $^{238}U/^{204}Pb$ ratio (μ_3). As a result, the isotopic composition of the lead deposited in the galenas during this stage becomes a function of the conditions of ore formation so that the amount of radiogenic lead increases as the temperature and pressure decrease. The third stage was terminated when the residual phase was so enriched in the uranium that the uranium compounds — most likely taking the form of an intergranular film of UO_2 — began to crystallize together with the sulfides in the veins of group 3b. For these mineralizations, a fourth stage, during which the $^{238}U/^{204}Pb$ ratio (μ_4) was locally a constant, then began. It is interesting to note that the mineralogically transitional sample G62 (occurrence of covellite) lies quite close to the extension of the best-fitting straight line calculated for the group 3b veins.

It has already been stressed that no uranium minerals have been found in the samples belonging to group 3b in the course of the ore-microscopic survey. It is, of course, possible that minerals occurring in small quantities may have been overlooked, but to the author it seems far more probable that the uranium minerals were removed from the veins at a later stage of development. Analytically, this is supported by the fact that the age calculated from the slope of the best-fitting straight line, 1 708 Ma, is a maximum estimate, since t_4 has been assumed to be the present time (see p. 9). To obtain an age closer to the emplacement of the Wiborg massif, t_4 should be more than zero, indicating that the evolution of the radiogenic lead had stopped before the present time. Practically, this means, of course, the ejection of uranium out of the system.

It is well established that the denudation of the Precambrian in southern Finland had reached approximately the present level of erosion by the onset of the Cambrian. This is best demonstrated by the occurrence of Cambrian sandstone — probably of shelf facies — in cracks of the Åland rapakivi and elsewhere in southwestern Finland (e.g., Tanner 1911, Simonen and Kouvo 1955, Bergman 1976), as well as by the occurrence of Ordovician limestone in the graben of the bay of Lumparn in the Åland archipelago (e.g., Sederholm 1934). The Wiborg massif, for its part, plunges at its southeastern border under sand and claystone formations of Cambrian origin (Frosterus 1925). Lately, Soviet drilling operations have brought to light granites identical and contemporaneous with the Onas, Bodom and Obbnäs granites (Puura *et al.* 1974) under the thin (150—400 m, Vaher 1973) Phanerozoic strata of northern Estonia.

Thus it seems evident that all the mineralizations belonging to group 3 weie within the reach of supergene solutions at the onset of the Cambrian.[1]) The veins of group 3b were radically affected by this inasmuch as their uranium content — possibly present only as an intergranular film — was leached away. At the same time, the uranogenic lead evolved in the course of 900 Ma was incorporated in the galenas. During the diagenesis of the Cambrosilurian sediments, diagenetic p,T-conditions prevailed in the crakcs of the rapakivi massif, resulting in partial (Ravijoki) or total (Säkkijärvi) homogenization of the lead isotopes of the ore-bearing veins.

Galena occurrences in the Åland area and the Vehmaa and Laitila rapakivi massifs

There are several minor galena occurrences in the Åland area as well as in the Vehmaa and Laitila rapakivi massifs. Short descriptions of them are furnished in Appendix 2, and the analytic results of their lead isotopic analyses are presented in Table 11. The sampling sites are marked on the index map (Fig. 1, p. 17).

Sample G1 from Södö, Sottunga, is included in this study mainly for the sake of completeness, since its isotopic composition was reported by 0. Vaasjoki and Kouvo (1959) in connection with several of the epigenetic veins of the Wiborg massif. In contrast to the group 1 deposits of the Wiborg massif, its lead is the most radiogenic common lead encountered in the course of this work, and its mineralogy is comparable to the veins of group 3b in the Wiborg massif. Geologic observations have shown that the ore is definitely of Svecofennian origin (Laurén 1974). This deposit thus demonstrates that the formation of radiogenic lead in an ore deposit is related primarily to the conditions of its formation, and only in a secondary manner to its geologic age.

The grouping of the deposits as derived for the Wiborg massif can be applied to the ore deposits of the rapakivi massifs of western Finland as well. Of group 2 deposits, there are the greisen formations at Eurajoki, which were a target of intensive reconnaissance prospecting in the late 1960s. These deposits are mineralogically in every respect similar to those encountered in the Kymi cupola formation of the Wiborg massif (Haapala and Ojanperä 1973, Haapala 1974 and 1977). Even their lead isotopic compositions are extremely similar to the extent that the range of the determinations from the Eurajoki deposits is laiger than the difference between the Eurajoki and Kymi averages.

[1]) Lehtovaara (1976) interprets fission track data from apatites (570 Ma) as indicating an uplift of 8 km at a mean rate of 14 mm/ka since the beginning of the Cambrian. Such a model cannot, however, account for the geologic field observations.

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Table 11. Lead isotopic composition of galenas in the Åland area, and the Vehmaa and Laitila massif.

Sample	Index	Locality	$^{206}Pb/^{208}Pb$	$^{206}Pb/^{207}Pb$	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	T_2	μ_2
G 1a	33	Sottunga	.56951±.00066	1.3726±.0008	22.204±.010	16.176±.016	38.988±.063	—	—
G 1b	33	Sottunga	.56344±.00028	1.3633±.0005	22.016±.011	16.149±.014	39.075±.039	—	—
G259	27	Eurajoki	.44820±.00025	1.0424±.0006	16.071±.012	15.417±.020	35.856±.047	1608	10.00
G260	27	Eurajoki	.44976±.00078	1.0431±.0005	16.067±.005	15.403±.012	35.724±.073	1585	9.90
G274	27	Eurajoki	.44842±.00007	1.0427±.0003	16.037±.003	15.381±.008	35.764±.014	1575	9.80
G 63	29	Taivassalo	.44929±.00038	1.0377±.0006	15.968±.010	15.388±.019	35.540±.052	1640	9.92
G258	39	Finnström	.45846±.00064	1.0654±.0016	16.470±.012	15.458±.034	35.924±.076	1362	9.85
G250	40	Vårdö	.52852±.00034	1.3233±.0008	21.222±.021	16.037±.026	40.153±.066	—	—
G257	41	Saltvik	.53673±.00022	1.3453±.0009	21.676±.018	16.112±.024	40.385±.050	—	—
G275	40	Vårdö	.53730±.00020	1.3427±.0007	21.625±.003	16.110±.010	40.240±.020	—	—
G276	40	Vårdö	.52834±.00005	1.3201±.0006	21.090±.020	15.980±.020	39.920±.040	—	—

T_2 and μ_2 calculated according to the two stage model of Stacey and Kramers (1975).
All results normalized against the CIT-standard (Catanzaro 1967).

Of the representatives of group 3a, sample G63 is the least radiogenic of the population. Its model age compares favourably with the age of the Vehmaa massif. The other representative of the group, G258, derives from gangue material, which may have caused a contamination of the lead in the galena.

The epigenetic veins in group 3b are more radiogenic than the corresponding veins in the Wiborg massif. Also, they cannot be plotted on a straight line, even within a single occurrence (Vårdö, G250, G275, G276). It appears, therefore, that within the Åland area the residual phases of the rapakivi massif have been more heterogeneous than in the Wiborg massif.

The results obtained from western Finland seem to support the findings from the Wiborg massif. Especially the hypothesis that the increasing radiogenity of the lead in the galenas is a function of the lowering of temperature (and possibly of pressure) during ore formation receives additional support from these data.

DISCUSSION AND CONCLUSIONS

In the course of the preceding chapters, it has been established, that

— there exists in the Åland area a group of granites exhibiting postorogenic features and registering ages between 1 800 and 1 840 Ma;
— the emplacement of the Finnish rapakivi granites occurred during a time interval between 1 700 and 1 540 Ma ago;
— there are marked age differences between intrusives, and, in addition, within each major intrusive, measurable age differences supported by geologic evidence;
— at least in the Wiborg massif, the residual phases of the rapakivi proper seem to have generated epigenetic galena mineralizations, the lead of which becomes progressively more radiogenic as the conditions of ore formation become less intense.

In the following discussion, each of these results will be treated separately in the order listed in the foregoing.

The 1 840—1 800 Ma age group in the Åland area is quite well established by the analytic material presented in chapter 4. Most of the rocks bear the characteristics of Sederholm's 3rd group of granites: with the exception of the Åva monzonite and the Seglinge granodiorite, they are rich in potassium and none of them forms veined migmatites, but brecciates the adjoining Svecofennian rock. They differ from the rapakivi granites mainly in age and texture. Around the composite ring intrusions of Åva (Kaitaro 1953, Bergman 1971 and 1973a) and Seglinge (Björk 1976), the tectonic features of the country rock have changed locally to parallel the contact of the stock. In the sense of Buddington (1959), these rocks might well be called stocks transitional between the epizone and mesozone. As related to the Svecokarelian orogeny, they are rather post- than late orogenic, mainly because they exhibit only a few signs of deformation.

The Nattanen and Vainospää granites in Lapland (Meriläinen 1976) and most Swedish granites (Welin *et al.* 1971) exhibiting a similar mode of occurrence register an age between 1 730 and 1 780 Ma, while others in Finland (e.g., the Hirvensalo granodiorite, Korsman and Lehijärvi 1973) are of an age corresponding to that of the Åland intrusives. Most of the rocks of the 1 840—1 800 Ma age group in Sweden and some in Finland (e.g., a pegmatite granite at Kitee, Kouvo 1958) must be classified as late orogenic, or, in the sense of Sederholm, migmatite-forming. In the classification of Buddington (1959), they are meso-catazonal batholiths.

One possible source of the age discrepancy observed between the Finnish U-Pb data on zircons and the Swedish Rb-Sr whole rock determinations could be systematic analytic differences. Though this cannot be ruled out completely, it seems unlikely, since age determinations of synorogenic granites and rapakivi granites agree remarkably well. Furthermore, as was already noted, the zircons of some Finnish late orogenic granites register U-Pb ages in the 1 840—1 800 Ma time interval. Also, in the cases where the Finnish U-Pb determinations of zircons and the Swedish Rb-Sr determination of whole rock samples exhibit similar ages, the results relating to the Rb-Sr mineral ages agree as well.[1])

In the present author's view, the problem of the 1 840—1 800 Ma age group is very much a verbal one, stemming from the fourfold granite classification of Sederholm, which is still widely used in the jargon of Finnish geologists. The Åland area is the type location of this classification, extracted from the contact relationships of the various granites. Thus Sederholm's grouping refers to the relative ages of the intrusive bodies as characterized by their mode of occurrence in the Åland area. As in the case of many other Precambrian classifications, an extrapolation to another area may be fallacious. This happened, in fact, to Sederholm himself, when he tentatively assumed by analogy that the Onas and Obbnäs granites belong to his third group (Sederholm 1934, p. 54). Obviously, the same reasoning as that applied to Sederholm's granite groups is equally applicable to the likewise relative nomenclature, synorogenic, lateorogenic and postorogenic.

Buddington has stated that his zones of emplacement refer »... in substantial part to *intensity zones* rather than to strictly depth zones» (1959, p. 676). Considering the discussion in the foregoing paragraphs, the following conclusion can be made: the 1 840—1 800 Ma group of intrusive stocks in the Åland area demonstrates that the intensity of the orogenic processes had greatly diminished in this region at the level represented by today's erosion surface some 1 840—1 800 Ma ago. In other parts of the Fennoscandian shield, representing different zones of intensity, the cessation of the orogenic movements probably occurred at other times, thus resulting in a multitude of overlapping ages for the late orogenic and early postorogenic granites within the whole craton.

[1]) It is assumed that the »geologic» decay constant of Rb, 1.39×10^{-11}/a is employed.

The emplacement of the Finnish rapakivi granites occurred during the time 1 700—1 540 Ma. Within the Fennoscandian shield, the result compares favourably with the ages obtained from the Salmi massif (Polkanov and Gerling 1960), the Dala (Welin and Lundquist 1970) and the Småland porphyries (Åberg 1972), as well as with the postorogenic granites in northern Sweden (Welin *et al.* 1971, Gulson 1972). The rapakivi granites of Ragunda and Nordingrå in Sweden register Rb-Sr whole rock ages of 1 320 and 1 445 Ma, respectively, and the gabbro-anorthosite associated with the latter one is 1 585 Ma old (Kornfält 1976, p. 102). Within other Precambrian areas, postorogenic rocks corresponding in age to those in Finland have been recorded at least in the Ukrainian shield (Semenenko *et al.* 1968) and in North America (van Schmus *et al.* 1975). All these rocks are either rapakivi granites proper or chemically akin to them. The rapakivi granites of Greenland, probably representing a deeper level of intrusion, are somewhat older, being of the same age as the early postorogenic granites of Sweden (Gulson and Krogh 1975).

The fact that, in the light of abundant analytic material, the range of the ages of the Finnish rapakivi granites may be safely extended to 1 540 Ma makes certain correlations more credible than ever. The correlation between the Fennoscandian shield and the Precambrian of Scotland (Bowes 1975) is impressive in itself, but the existence of epizonal granites in Finland exhibiting an age similar to that encountered in pebbles of acidic hypabyssal rocks in the Torridonian sediments (Moorbath *et al.* 1967) adds another finishing touch to the general picture.

The contemporaneity of the Penokean and Svecokarelian orogenies has often been referred to (e.g., Sederholm 1896, Kouvo 1958, Goldich *et al.* 1962, Welin 1970); and it has been attempted to draw a correlation between the American and European Precambrian terrains, especially in the context of the anorthosite problem (Kranck 1968, Herz 1969, Bridgewater and Windley 1973). Van Schums *et al.* (1975) were forced to conclude, since they lacked the present data, that no correlation between the postorogenic (in their view anorogenic) granites of the Fennoscandian shield and those of North America could be attempted because of excessive differences in age. The new data, collected since 1973, remove this obstacle to an intercontinental temporal comparison.

As far as the anorthosites are concerned, at least two, and most likely all the Finnish anorthosites spatially associated with rapakivi granites register a temporal association as well. Considering the heating effect of the Wiborg massif on its country rocks (Vorma 1972), the contemporaneity cannot be assigned to intense regional metamorphism, as in the case of the Adirondacks (Silver 1968), but must be regarded as real. The temporal and spatial relationship between the rapakivi massifs and the associated anorthosites, along with the fact that the latter contain abundant zircon evidencing uranium contents quite normal for rapakivi (see also Gulson and Krogh 1975), suggest their genetic association with the rapakivi granites.[1]) The hypothesis

[1]) It should be noted that, in view of the work carried out on rocks of the granulite facies in Scotland (Pidgeon and Bowes 1972, Moorbath and Park 1972) and in Finland (ages reported by Meriläinen 1976), the extremely low amounts of uranium encountered in the zircons of the Adirondacks by Silver (1968) may reflect either the metamorphic conditions or origin from a basic starting material.

advanced by Kranck (1968) seems therefore to be substantiated at least in the cases of the Wiborg and Laitila massifs.

In the Wiborg massif, the ages of the main rapakivi phases range from 1 700 to 1 650 Ma. From the distribution of the age data, it seems that most of the old ages are found in the main parts of the massif, while the youngest ones occur on the fringes, in the tirilite and the Ahvenisto satellite. The ages registered in the northwestern part of the massif and the Suomenniemi satellite are intermediate between these extremes. The present age data rather favour Wahl's (1947) concept of a sheet-like intrusion expanding from north to south, although some local discrepancies — which also apply to Vorma's (1971) succession of intrusive phases — do arise.

As depicted by Vorma (1971, p. 65), the intrusive sequence would have occurred as follows: 1) the tirilite and its associates, 2) the normal rapakivi varieties, 3) the Jaala-Iitti dike and 4) the porphyry dikes. The zircon age data from the Wiborg massif indicate that the tirilite is distinctly younger than either of the porphyry dikes and that the Jaala-Iitti dike is intermediate between them. These results disagree with the intrusive sequence outlined.

On the other hand, the tirilite is evidently the oldest intrusive rock of the rapakivi suite in the Lappeenranta area. It grades into the Lappee granite, which in its turn has a sharp contact against the wiborgitic main type of the massif. At the contact, there occur ovoids of metasomatic origin in the ortherwise even-granied Lappee granite (Hackman 1934, p. 22). Also, the chemical characteristics of the tirilite (Vorma 1971 and 1976) seem to indicate an early differentiation from the rapakivitic bulk magma.

The contacts of the Jaala-Iitti dike against the adjoining rapakivi varieties prove beyond any doubt that this rock is younger than the main phases of the Wiborg massif encountered in the northwestern part of the massif (Lehijärvi and Lonka 1964).

The very largness of the Wiborg massif suggests that it did not originate by a single pulse of magmatism. The age data obtained from the zircons also exhibit a wide range of values, many apparently contradictory to the geologic picture developed by previous workers in the area. The geologic and geochronologic data may, however, be brought to an acceptable agreement by assuming that several separate phases of intrusion occurred at relatively short intervals. Though the data presented in this study are too few to allow any sweeping statements, the author tentatively suggests the following time-scale for the emplacement of the Wiborg massif: 1) 1 700—1 660 Ma: intrusion in the southern, northern and northwestern parts of the Wiborg massif and in the Suomenniemi statellite, 2) 1 660—1 640 Ma: intrusion in the Lappeenranta area and in the Ahvenisto massif, 3) 1 640±15 Ma: emplacement of the Kymi cupola.

Only the existence of the last-mentioned group as a welldefined unit is supported by field evidence (Haapala and Ojanperä 1973, Haapala 1974). Elsewhere in the Wiborg massif, sharp contacts between various coarse-grained rapakivis have been occasionally met with, but no actual borders between the intrusive pulses as outlined in the previous paragraph can be drawn. In the author's opinion, this is no serious flaw when considering a tentative scheme combining the results of the age determinations and the established field relationships in a relatively simple manner.

Recently, on the basis of extensive petrochemical material, Vorma (1976) concluded that no chemical factors preclude the possible existence of separate pyterlitic and wiborgitic magmas within the Wiborg massif, as was suggested by Wahl (1925, p. 81). With the regional interpretation just offered, it would be tempting to suggest that the various intrusive phases may have originated from different magmas, which have been emplaced at somewhat different times. The analytic data presented, however, warrant no such view. It may be stated in this connection that since two different types of magma would possibly exhibit different initial lead and strontium isotopic compositions, an extensive analytic work utilizing the common lead and Rb-Sr whole rock mehtods might be helpful in solving this problem. [1])

In the rapakivi massifs of southwestern Finland several ages seem to occur. The two analyses of the rapakivi granites in the Åland area demonstrate that this massif is of the same age as the Wiborg massif. As in southeastern Finland, so within the Åland area younger rapakivitic intrusions also occur, notable examples being the greisen-bearing even-grained granites and the massif of Kökar fjärd. The analyses made of the porphyry dike of Blåklobb demonstrate that, as in the case of the Wiborg and Laitila massifs, the dike rocks in the Åland area are of the same age as the main intrusive phases.

The most probable ages of the Vehmaa and Laitila massifs are 1 590 and 1 570 Ma, respectively, the difference being just about one-half of the age range encountered within the main phases of the Wiborg massif. Also, the rock types encountered in the Laitila and Vehmaa massifs are quite similar, as shown by the descriptions given by Kanerva (1928) and Vorma (1976). Especially the mineralogic resemblance between the even-grained granite at Uhlu and the porphyritic type of Väkkärä granite (Haapala 1974 and 1977) is striking. The fact that, in spite of its extreme uranium content, the zircon of the Uhlu granite exhibits neither a high initial common lead nor an extraordinary degree of discordancy indicates that this rock may have been somewhat dryeı than the Väkkärä granite.

Considering the geographic proximity of the Vehmaa and Laitila massifs, the similarity of their rocks and ages and the fact that they both crop out above the same negative gravity anomaly, there would seem to be eveıy reason to repeat the often expıessed thought that these two massifs are actually one and the same, with the Svecofennian migmatites between them filling only a shallow undulation in the surface of a huge intrusive body. If this is true, then also the hypothesis about the proximity of the present level of erosion in the Laitila and Vehmaa massifs to the ancient roof of the pluton could be considered as proven.

[1]) It should be noted that the whole rock work in the Laitila area indicates different initial Pb composition for at least the main type of Laitila rapakivi and the Tarkki granite. Initial Sr compositions from Sweden (Welin and Lundqvist 1970, Welin et al. 1971, Åberg 1972, Gulson 1972) exhibit quite uniform values for all the postorogenic granites contemporaneous with the Finnish ones, whereas the younger Nordingrå and Ragunda rapakivis (Kornfält 1976) register much higher initial ^{87}Sr.

As was pointed out during the discussion of the isotopic evolution of ore lead, the lead in ores seems to become more radiogenic as the thermodynamic conditions characterizing the environment of mineral deposition become less intense. This finding accords with the obsetvations of Doe and Stacey (1974) and Loveless (1975). Further, the trace lead study of the chalcopyrites contained in the minor Telkkälä Ni-Cu-depotis (Häkli *et al.* 1975) brings to light a similar trend. The study carried out in the Wiborg massif makes it possible to extend the relationship discovered by other investigators to lower thermal states. From the ore-genetic point of view, this means that a continuous evolution of the lead isotopic composition can be proven to exist in ore deposits situated in a magmatic environment. In these deposits, the composition of the radiogenic addition in the lead reflects the conditions of ore formation rather than the age of the deposit. Only if lead and uranium undergo the same early development within a whole deposit or set of deposits is a »secondary» isochron formed. For the magmatic environment, the isotopic composition of the lead in galenas and other sulfides may be developed into a useful ore-genetic indicator.

SUMMARY OF THE RESULTS

The geochronologic conclusions arrived at in the foregoing pages may be summed up as follows:

— The early postorogenic granites of the Åland area are 1 840—1 800 Ma old. Their postorogenic features result from emplacement in a zone of lesser orogenic intensity than in the case of most other granites within the Fennoscandian shield exhibiting a similar age.
— Of the Finnish rapakivi granites, the Wiborg massif is the oldest, its intrusion having occurred 1 700—1 640 Ma ago. It is tentatively suggested that there occurred three major magamatic phases: at 1 700—1 660 Ma in the southern, northern and northwestern parts of the massif and in the Suomenniemi satellite; at 1 660—1 640 Ma in the northeastern part of the massif and in the Ahvenisto satellite; and at 1 640±15 Ma in the Kymi cupola.
— The Onas, Bodom and Obbnäs granites are 1 650—1 640 Ma old and contemporaneous with the Ahvenisto massif.
— The main phases of the Åland rapakivi massif are 1 670±20 Ma old, being thus almost of the same age as the Wiborg massif. The Kökar massif and the greisen-bearing granites are 1 620±20 Ma old.
— The most probable age of the Vehmaa massif is 1 590 Ma, but igneous activity may have continued in the area until 1 530 Ma ago.
— The most probable age of the Laitila massif is 1 570 Ma. The Ytö granite is definitely younger, registering an age of 1 540 Ma. The younger ages of the topaz-bearing variety of the Väkkärä granite are probably due to the extreme discor-

dancy of the samples, arising from the combined effects of continuous diffusion and episodic lead loss processes.

— The epigenetic galena occurrences observed within the rapakivi massifs are of the age of the rapakivi and derived from it.

Conclusions not strictly geochronologic are the following:

— The Satakunta graben was formed after the emplacement of the Laitila massif and before the intrusion of the Jotnian diabases.

— At the onset of the Cambrian, the rapakivi granites in Finland had reached practically their present level of erosion and became exposed to weathering processes.

— In addition to the spatial relationship of anorthosites and rapakivi granites in southern Finland, there is a temporal and possibly also a genetic association.

The methodologic results relating to the analytic procedures used during this work are:

— Zircons formed from residual magmatic solutions seem to be more liable to produce discordant age patterns. This is probably the result of initial defects in the lattice caused by the abnormally high initial contents of common lead.

— In a magmatic environment, the amount of the radiogenic component of lead in the ore minerals is a function of the conditions of ore formation, and it might serve as a useful ore-genetic indicator.

The geochronology of southern Finland since the Svecokarelidic orogeny, along with the amendments necessitated by these results, is presented in Table 12.

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Table 12. Geochronology of southern Finland since 1900 Ma.

PHANEROZOIC				Uplift (Quaternary)
		Erosion		
				Subsidence?
		Phanerozoic sedimentation	?—600 Ma	
PRECAMBRIAN		Peneplanation		
		Postjotnian diabases	1250—1300 Ma	
		Jotnian sedimentation	1300—1400 Ma	
	Platform	Erosion		Uplift (2—3 km) Formation of the Satakunta graben Faulting
		Rapakivi granites Gabbro-anorthosites	1540—1700 Ma	Crustal distension
		Subjotnian diabases	1680—1700 Ma	
		Erosion		Uplift (5—10 km)
		Early postorogenic granites Late orogenic granites	1800—1860 Ma	Cessation of compression
	Svecokarelian orogeny			Crustal compression
		Synorogenic granites	1860—1900 Ma	

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The sample material used for U-Pb analyses of zircons

The sample designations are in the order: laboratory number, locality, index map number and field number of the Geological Survey of Finland if referred to in other publications.

The Wiborg massif

A 18 Värtö, Lemi. (Index map No. 1). Geological Survey of Finland field No. 35b/AS/54. A darkish, even-grained, hornblende-bearing rapakivi granite (tirilite). Main minerals are orthoclase, oligoclase, quartz and hornblende. The accessory constituents are olivine ((Fa_{80}), zircon, apatite and magnetite. Previous age determination by Kouvo (1958). Wet-chemical analysis of amphibole in Simonen and Vorma 1969. Intergrating stage analysis and chemical analyses of total rock and olivine in Simonen (1961). Ordering of the potassium feldspar by Vorma (1971).

A 21 Mentula, Suomenniemi (6). Field No. 297/FP/55. Granite porphyry dike. Angular potassium feldspar and idiomorphic quartz phenocrysts in a phaneritic groundmass. Biotite and hornblende occur as the mafic minerals. Accessory constituents: zircon, apatite and fluorite. Lattice constants and radiation-damage age of zircon by Kouvo (1958).

A 29 Falin, Muurikkala, Virojoki (3). Field No. 29/KM/55. A darkish wirborgite. The ovoid potassium feldspar phenocrysts are formed of a mixture of orthoclase and microcline. Quartz occurs as smoke-coloured idiomorphic grains. Biotite and hornblende form the main mafic constituents. Previous age determination by Kouvo (1958). Ordering of the potassium feldspar by Vorma (1971).

A 69 Verla, Iitti (5). Field No. 714/ML/54. A porphyritic biotite granite with a coarse-grained groundmass. Wet-chemical analysis in Simonen and Vorma (1969). Lattice constants and radiation damage age of zircon by Kouvo (1958). Ordering of the potassium feldspar by Vorma (1971).

A 98 Suomenniemi (2). Field No. 292/FP/55. The main rock type of the Suomenniemi satellite massif. A coarse-grained biotite-hornblende granite exhibiting erratic ovoid crystals of potassium feldspar. The lattice constans, radiation damage age and lead-alpha age of zircon by Kouvo (1958). Ordering of the potassium feldspar by Vorma (1971). Analyses of amphiboles and biotite in Simonen and Vorma (1969).

A 118 Nurmaa, border of the parishes of Mäntyharju and Jaala (7). The gabbro-anorthosite arc surrounding the Ahvenisto satellite massif. Main minerals: plagioclase (An_{60}) and hornblende. Most conspicuous accessory minerals: biotite, chlorite, zircon and opaques. A dark, medium-grained rock, exhibiting labradorizing surfaces in some places. Described in detail by Savolahti (1956, 1966). Age determination recalculated from the Annual Report of the Geological Survey (1973).

A 323 Hamina (8). Field No. 18/AS/56. A quartz porphyry vein with a dark red aphanitic groundmass. Bright red, angular phenocrysts of potassium feldspar and idiomorphic grains of quartz. Hornblende and biotite occur in the groundmass as the other main constituents. The accessory minerals include at least zircon, fluorite and ore minerals. The length of the dike is at least 50 m and the minimum width 3 m. A 30 cm broad chilling margin void of phenocrysts may be observed. The vein is embedded in a pyterlitic rapakivi. Wet-chemical analysis and ordering of potassium feldspar reported by Vorma (1971). Previous age determination in the Annual Report of Activities (1970), recalculated.

A 422 Ruoholampi, Lappee (9). Field No. 795. A massive, light-grey granite porphyry. Light potassium feldspar and dark quartz phenocrysts densely scattered in a grey phaneritic groundmass. In places, slight alignment of the phenocrysts may be observed. Close to the contacts, flow-texture-like banding is visible. Wet-chemical and spectrographic analysis in Vorma (1975). Age determination in Annual Report of Activities (1973) and Vorma (1975), recalculated.

A 423 Hyvärilä, Lemi (17). Field No. 740. A Subjotnian diabase in the roof pendant of Hyvärilä (Vorma 1975). Age determination in Annual Report of Activities (1973) and Vorma (1975), recalculated.

A 524 Kymi (10). A porphyritic rapakivi granite containing greisen formations. Dominant minerals: potassium feldspar, plagioclase, quartz, biotite and hornblende. Accessory minerals: topaz, fluorite, zircon and ore minerals. For a detailed description and wet-chemical analysis, see Haapala and Ojanperä (1973) and Haapala (1974).

A 627 Hogland, USSR (11). A sample from the collections of the Geological Survey of Finland. A bright red quartz porphyry containing phenocrysts of red, angular potassium feldspar and idiomorphic quartz in an aphanitic groundmass. Biotite occurs as the mafic mineral. Accessory constituents: zircon and ore minerals. According to the labelling of the sample, the red porphyry of Wahl (1947). Other references: Ramsay (1890 and 1892), Kranck (1929).

A 629 Parola, Valkeala (12). A road cut along the Kouvola—Mikkeli highway. A rock immediately adjacent to the contact of the Wiborg massif, about 1 m from the contact surface on the side of the rapakivi. A dark, fine-grained rock, the major minerals being potassium feldspar, quartz and plagioclase (An_{10}). Olivine (Fa_{80}) and biotite occur as minor constituents. Accessory minerals: zircon and ore minerals. The texture is hypidiomorphic. A few scattered potassium feldspar phenocrysts occur. The rock belongs clearly to the rapakivi massif, forming a layer about 2 m wide parallel to the contact. On the side of the rapakivi, it exhibits a sharp border against the next layer of similar composition, which contains a larger amount of feldspar phenocrysts. A third layer of this nature grades into a pyterlite.

A 630 Parola, Valkeala (12). From the same road cut as sample A 629. A pyterlitic rapakivi about 200 m from the contact of the rapakivi. The ovoid potassium feldspar crystals are smallish, 1—2 cm Ø. Biotite is the chief dark mineral. In places, about 1 dm^2 -sized segregations of mafic minerals (biotite and hornblende) may be seen.

A 631 Pesäntäjärvi, Valkeala (13). A typical wiborgite in a road cut along the Kouvola—Mikkeli highway about 2 km south of the northern contact of the Wiborg massif.

The Onas, Bodom and Obbnäs granites

A 97 Porvoo parish (23). Typical Onas granite. Large grains of densely packed potassium feldspar occur in a medium-grained groundmass consisting of plagioclase ($An_{15—55}$), quartz, biotite and hornblende. Accessory minerals: fluorite, zircon, titanite and ore minerals. Age determination by Kouvo (1958), recalculated.

A 101 Bodom, Vantaa (24). Even-grained hornblende-bearing potassium granite. The main constituents are microcline, plagioclase ($An_{15—55}$), quartz and hornblende. Accessory minerals: biotite, fluorite, apatite, titanite and zircon. Age determinations by Kouvo (1958), using U—Pb, Rb—Sr and K—Ar methods.

A 102 Bodom, Vantaa (24). Red coarse-grained potassium granite. The major minerals are potassium feldspar, plagioclase ($An_{10—50}$), quartz and biotite. Hornblende occurs as a minor constituent. The potassium feldspar occurs as densely packed phenocrysts in a medium-grained ground-mass. The angular phenocrysts (1×3) are somewhat oriented, the maximum orientation occurring in the direction of the elongate stock (N50°E) and parallel to the Porkkala—Mäntsälä fracture. A Rb—Sr age of 1605 Ma for biotite is reported by Kouvo (1958).

A 602 Vitträsk, Kirkkonummi (25). The typical Obbnäs granite, a red coarse-grained potassium granite. Major minerals: microcline, plagioclase (An_{10}—$_{50}$), quartz and biotite. Hornblende occurs as a minor constituent. Accessory minerals: apatite, titanite, zircon, fluorite and molybdenite. The microcline forms angular phenocrysts (1×3 cm) embedded in a medium-grained ground-mass. The phenocrysts are oriented along the long axis of the oval stock, parallelling the strike of the Porkkala—Mäntsälä fracture. Some erratic ovoid crystals of potassium feldspar occur. At the outcrop where the sample was taken, several mylonites, also parallelling the strike of the said fracture can be observed. References: Sederholm (1926) and Laitala (1961).

The Åland area

A 334 Åva, Brändö (30). Monzonite. Plagioclase (An_{50}—$_{30}$), orthoclase, biotite and hornblende form the major minerals. Accesory constituents include quartz, titanite, apatite, zircon and fluorite. Phenocrysts of potassium feldspar up to 2 cm in length occur sparsely. A detailed treatise by Kaitaro (1953). Age determination in Neuvonen (1970), recalculated.

A 335 Åva, Brändö (30). Coarse-grained porphyritic biotite granite. Potassium feldspar, plagioclase (An_{10}—$_{50}$), quartz and biotite are the major minerals. Hornblende occurs as a minor constituent. Apatite, titanite, zircon and fluorite occur as the accessory minerals. The potassium feldspar forms rectangular phenocrysts (1×2) cm). Field observations (e.g., Kaitaro 1953) demonstrate that this rock is younger than the monzonite encountered in the area. Wet chemical analysis in Kaitaro (1953). Age determination in Neuvonen (1970), recalculated.

A 440 Lemland (38). Red coarse-grained potassium granite. Potassium feldspar, oligoclase, quartz and biotite are the main minerals. Titanite and zircon occur as accessory constituents. The potassium feldspar forms angular grains of ca. 1×2 cm. Briefly described by Sederholm (1934, p. 45), who based his division between the rapakivi and the third-group granites on the contact relations of the Lemland granite. Age determination in the Annual Report of Activities (1972), recalculated.

A 441 Söderharn, Kökar (32). A red coarse-grained potassium granite with 1—3 cm laths of potassium feldspar. Major minerals: potassium feldspar, quartz, plagioclase (An_{10}—$_{50}$) and biotite. Hornblende occurs as a minor constituent. Accessory minerals: fluorite and zircon. Ovoid crystals of potassium feldspar occur sporadically. Briefly mentioned by Sederholm (1934, p. 66). Age determination in Annual Report of Activities (1972), recalculated.

A 442 Sälskär, Sottunga (34). Mosshaga granite. Coarse-grained red potassium granite exhibiting 1—2 cm long laths of potassium feldspar. Other major minerals: quartz, plagioclase (An_{10}—$_{20}$) and biotite. Briefly mentioned by Sederholm (1934, p. 46). Age determination in the Annual Report of Activities (1972), recalculated.

A 443 Vänskär, Sottunga (34). Mosshaga granite. Medium-grained potassium granite occurring as ill-defined patches in the coarse granite represented by sample A 442. Major minerals: potassium feldspar, quartz, biotite and plagioclase (An_{18}—$_{50}$). Geologically, the age relation of the two varieties is difficult to determine, but the medium-grained type may be somewhat older, possibly an autolith. Age determination in the Annual Report of Activities (1972), recalculated.

A 444 The Fjälskär granite (35). A medium- to fine-grained biotite-muscovite-potassium granite Major minerals: microcline, quartz, biotite and muscovite. Plagioclase (An_{5}—$_{15}$) occurs as a minor constituent. Accessory minerals: fluorite, apatite and zircon. The circular intrusion at Mossala fjärd, in contrast to the Åva intrusions, has not had an influence on the lineation and fracturing of the surrounding Svecofennian rocks (Bergman 1971, 1973a). Age determination in the Annual Report of Activities (1972), recalculated.

A 467 Eckerö (36). Typical rapakivi of the Åland massif. Potassium feldspar forms ovoid crystals of 2—3 cm Ø, about 2/3 of which are surrounded by a mantle of oligoclase. Quartz and biotite are the other major minerals; hornblende occurs as a minor constituent. Fluorite, apatite, zircon and anatase are the accessory minerals.

A 468 Lumparland (37). Even-grained greisen-bearing rapakivi granite. A medium-grained potassium granite exhibiting hypidiomorphic texture. Occasional porphyritic phenocrysts and pyterlitic ovoid crystals of potassium feldspar. Major minerals: Potassium feldspar, plagioclase (An_{10}—$_{50}$), quartz and biotite. Minor constituent: hornblende. Accessory minerals: fluorite, zircon, ore minerals.

A 508 Kummelskär, Kumlinge (31). Part of the Seglinge granite exhibiting long laths of feldspar and quartz. The rock contains numerous fragments of Svecofennian rocks. Reference: Björk (1976).

A 509 Kummelskär, Kumlinge (31). Part of the Seglinge granite. The potassium feldspar is reddish brown in colour. The rock deviates structurally from A 508 and A 510 by being more even and finer in grain. It contains a large amount of fragments of Svecofennian rocks, but appears itself as xenoliths in the granite represented by A 508. Reference: Björk (1976).

A 510 Kummelskär, Kumlinge (31). Part of the Seglinge granite. Potassium feldspar occurs as large, broken grains, smaller grains occurring around the larger ones. The weathering surface is darkish red and white grains of quartz may be readily observed. The rock resembles A 508, but is not so uniform in structure. Also, the contacts against the country rocks are not quite so sharp as in the case of A 508.

A 715 Blåklobb, Eckerö (49). Quartz porphyry. Angular phenocrysts of potassium feldspar (15 × 8 mm) and idiomorphic quartz (Ø 5 mm) in an aphanitic groundmass. The rock exhibits a flow structure, and forms a rather flat dike with an irregular contact against an »ossipite» in the sense of Sederholm. Map of the island and chemical analysis in Sederholm (1934).

The Vehmaa massif

A 710 Uhlu, Vehmaa (46). An even-grained biotite rapakivi. Potassium feldspar, oligoclase and biotite are the major minerals. Quartz and hornblende occur as minor constituents. Accessory minerals: fluorite, topaz, monazite, zircon and xenotime. The rock forms a bulge into the Svecofennian formations at the eastern contact of the Vehmaa massif. It is quarried for commercial purposes and sold as the Balmoral Red. Reference: Kanerva (1928).

A 711 Vehmaa (45). The main type of the Vehmaa massif. A rapakivi, where pyterlitic and wiborgitic ovoid crystals of potassium feldspar occur in approximately 2 :1 ratio. Biotite is the dominant mafic mineral, oligoclase, quartz and hornblende occur as major constituents as well. Accessory minerals: zircon and fluorite. Reference: Kanerva (1928).

The Laitila massif

A 129 Peipohja, Kokemäki (26). This rock is separated from the Laitila massif by the Jotnian sandstone of Satakunta, but the age determination allows it to be classified togerher with the Laitila massif. Sederholm (1911) took this porphyritic potassium granite as similar to those of Onas and Mosshaga. The rock exhibits some tectonic features probably produced during the formation of the Satakunta graben. Previous age determination by Kouvo (1958). Wet chemical analysis in Vorma (1976).

A 255 Tarkki, Eurajoki (27). The Tarkki granite. The marginal rock of the polyphasic Eurajoki stock. The contact relations against the main type of the Laitila massif are unknown, but the Tarkki granite is definitely brecciated by the Väkkärä granite. The major minerals are orthoclase, quartz, plagioclase (An_{15}—$_{25}$) and biotite. Hornblende occurs a as minor constituent. Accessory minerals: Olivine (Fa_{85}), apatite, zircon and iron oxides. A few wiborgitic ovoid crystals of potassium feldspar occur sporadically. The rock has been dealt with by A. Laitakari (1928), Haapala and Ojanperä (1972), and Haapala (1974 and 1977). Kahma (1951) paid special attention to the palingenesis of the Tarkki granite during the intrusion of the Jotnian diabases. Chemical analyses in Laitakari (1928) and Haapala (1974). Age determination in the Annual Report of Activities (1968), recalculated.

A 515 Linnamaa, Eurajoki (27). Contact type of Väkkärä granite. Surrounded by the Tarkki granite, the Väkkärä granite is cut only by a dark porphyry vein (A 707). Potassium feldspar, quartz and plagioclase (An_5—$_{15}$) are the major constituents. Biotite and chlorite occur as the minor minerals. Accessory constituents are fluorite, apatite, iron oxides and zircon. The rock is fine-grained at the contact and grades into a medium-grained variety. Chemical analysis in Haapala (1974).

A 606 Kylämaa, Eurajoki (27). Tarkki granite. Similar to A 255.

A 607 Koivuniemi, Eurajoki (27). Porphyritic type of the Väkkärä granite. Microcline, albite and quartz from the major minerals. Minor constituents include biotite and topaz. Accessory minerals: fluorite, monazite, cassiterite and zircon. Occasionally, a sharpish contact between the porphyritic and even-grained varieties of the Väkkärä granite may be observed.

A 608 Untamala, Laitila (28). Field No. 327/LC/68—73. Normal type of Laitila rapakivi. A hornblende-biotite granite exhibiting both pyterlitic and wiborgitic ovoid crystals of potassium feldspar in a 2: 1 ratio. A chemical analysis in Vorma (1976).

A 609 Heikkilä, Eurajoki (27). Porphyritic type of Väkkärä granite. For description, see sample A 607. Chemical analysis in Haapala (1974).

A 686 Metsäranta, Eurajoki (27). Contact type of Väkkärä granite. For description, see sample A 515. The sample has been taken about 3 m from the contact of the Tarkki and Väkkärä granites. Chemical analysis in Laitakari (1928).

A 687 Koivuniemi, Eurajoki (27). Coarse-grained type of Väkkärä granite. Highly similar in composition to the porphyritic type of the same granite, but the coarser grain-size (microcline phenocrysts of 1×2 cm) makes it hard to observe the porphyritic texture. The coarse-grained and porphyritic varieties grade into each other at several outcrops. Chemical analysis in Haapala (1974).

A 688 Koivuniemi, Eurajoki (27). Even-grained, topaz-bearing type of the Väkkärä granite. Chemical analysis in Haapala (1974). No zircon could be separated from this sample.

A 689 Hinnerjoki (42). Field No. 30—2/AV/67—73. Typical Laitila rapakivi. For description, see sample A 608. Chemical analysis in Vorma (1976).

A 690 Kusni, Laitila (43). Field No. 1—1/AV/ 66—73. The Ytö granite. A stock of porphyritic granite in the southern part of the Laitila massif. Phenocrysts of potassium feldspar (1×2 cm) occur in a medium-grained porphyritic groundmass consisting of potassium feldspar, plagioclase and biotite. Zircon, apatite and monazite occur as the accessory minerals with iron oxides. Very subordinate molybdenite has been observed. Chemical analysis in Vorma (1976).

A 691 Kolinnummi, Yläne (44). Gabbro-anorthosite. Situated on the southeastern border of the Laitila massif, not in direct contact with the rapakivi. An elongate mafic intrusion, the long axis parallelling the strike of the surrounding kinzigitic migmatites. Studied to some extent by Hietanen (1947).

A 707 Pistola, Eurajoki (27). A dark porphyry dike. Large phenocrysts of plagioclase (An_{55}—$_{65}$) and smaller ones of potassium feldspar as well as rounded grains of quartz occur in an aphanitic groundmass of dacitic composition. The rock cuts the marginal type of Väkkärä granite, and other dikes are known to cut the Tarkki granite as well (Haapala 1977).

511/LC/68—73. Kylmäkorpi, Kalanti (47). Normal Laitila rapakivi. Chemical analysis in Vorma (1976).

28/AV/66—73, Lammi, Mynämäki (48). Normal Laitila rapakivi. Chemical analysis in Vorma (1976).

The galena material analyzed from the sulfide deposits of the Finnish rapakivi massifs

The Wiborg massif

G 7 Hyvärilä, Lemi (17). The sulfide ore of Hyvärilä lies in Svecofennian country rocks of a root pendant located at the northeast fringe of the Wiborg massif. The ore was sporadically quarried during the 17th and 18th centuries. It is formed by a set of veins and pockets in gneisses about 500 m from the contact of the rapakivi. The predominant ore minerals are pyrrhotite and pyrite, but chalcopyrite, sphalerite and galena occur in lesser amounts as well. Field report M/17/Le46/1 by Pääkkönen in the files of the Geological Survey.

G 263 Koskenkylä, Pernaja (16). The former silver quarry of Koskenkylä lies about 3 km west of
G 266 the western contact of the Wiborg massif, but still within its contact aureole and possibly also under the influence of the Onas granite. The ore is formed by two parallel fractures filled by galena veins. Pyrrhotite, chalcopyrite, arsenopyrite, sphalerite and loellingite also exist in the deposit. A detailed description by O. Vaasjoki (1953).

G 272 Suursomerikko, Hogland, USSR (15). On the western shore of Hogland, there are several tectonic breccias in the Svecofennian migmatites (»rifningsbreksior», Ramsay 1890, 1892), some of which contain galena. The sample belongs to the collections of the Geological Survey of Finland.

G 151 Peippola, Kymi (10). At Kymi, an oval cupola formation containing even-grained and
G 261 porphyritic types of rapakivi intersects a typical wiborgite. Several greisen-formations are found in the area. Some of them contain sphalerite, galena and other ore minerals. These occurrences have been decribed more closely by Haapala and Laajoki (1969), Haapala and Ojanperä (1973) and Haapala (1974).

G 62 Inkeroinen, Anjalankoski (19). At Anjalankoski, in the railroad yard of the Tampella paper mill, there occurs an epidote vein containing both galena and chalcopyrite. The country rock is a typical wiborgite. Field report M/17/Spl47/1 by Kulonpalo in the files of the Geological Survey.

G 141 Anjala, Anjalankoski (19). Close to the site of G62, but on the other side of the Kymi river, there occurs a small galena-quartz vein in a wiborgitic rapakivi.

G 277 Pernaja (50). In the first road cut exhibiting rapakivi on the eastern side of the Porvoo—Loviisa highway, there occurs a galena-bearing epidote vein. The country rock is a typical wiborgite, in which the oligoclase mantles of the potassium feldspar crystals have been epidotized in the surroundings of the vein.

G 2 Muhutniemi, Säkkijärvi, USSR (22). Southeast of the church of Säkkijärvi, on the shore of the Gulf of Finland, the wiborgite contains a swarm of galena-quartz veins. The sample analyzed belongs to the collections of the Geological Survey of Finland. A brief mention by Berghell and Frosterus (1897).

G 3 Hopiala, Luumäki (21). A galena-bearing quartz vein, intersecting wiborgite, was quarried for silver in the 16th and 17th centuries. Large mineralizations are no longer found in the outcrops. The analyzed sample was taken from gangue material containing specks of sulfides about 5—1 cm Ø. Field report M/17/Lä47/1 by Kulonpalo in the files of the Geological Survey.

G 61 Ravijoki, Virojoki (20). In a road cut close to the entrance of an army field camp, there
G 265 occurs a small galena-bearing quartz vein. About 2 km north of this occurrence, a similar vein is to be found in a bottom outcrop of the Ravijoki river. In both cases, the host rock is wiborgite.

The Åland area and the Vehmaa and Laitila massifs

G 1 Södö, Sottunga (33). A skarn deposit in Svecofennian rocks. The deposit is described in close detail by Laurén (1974). Galena is the major mineral in a small vein about 2 cm wide; otherwise, the deposit consists of magnetite and iron sulfides. Pyrite and chalcopyrite also occur in the vein, the latter as crack fillings. Sample obtained from the collections of the Geological Survey. Previous determination of lead isotopic composition in O. Vaasjoki and Kouvo (1959).

G 259 Eurajoki (27). In the greisen-mineralizations occurring in the Tarkki and Väkkärä granites
G 260 (Appendix 1.), sphalerite and cassiterite occur as the major minerals, but galena and chalco-
G 274 pyrite are also met with. The sphalerite contains inclusions of chalcopyrite and in some cases exhibits pseudoanisotropic halos and lamellae (Haapala and Laajoki 1969).

G 63 Helsinginranta, Taivassalo (29). An epigenetic vein of galena, epidote and quartz occurs in an outcrop on the shoreline. The country rock is the normal type of Vehmaa massif (see sample A 711). Field report M17/Ts—61/1 by Kulonpalo in the files of the Geological Survey.

G 258 Grelsby, Finnström (39). A mineralization consisting mostly of galena and sphalerite, with strongly sericitized wall rocks in the normal type of Åland rapakivi. The occurrence has been quarried for silver during the 16^{th} and 17^{th} centuries and no ore proper is left in the deposit. The sample analyzed stems from strongly sericitized gangue material, which also contains some erratic grains of chalcopyrite.

G 250 Löören, Vårdö (40). Several epigenetic galena-quartz veins in the normal type of Åland
G 275 rapakivi. Pyrite and sphalerite occur in places as minor ore minerals. Sometimes a narrow
G 276 seam of hematite is visible between the veins and the host rock. On the same island, there also occur some sandstones of both clastic and cataclastic texture. The latter is also mineralized.

G 257 Silverskär, Saltvik (41). A galena and sphalerite vein in normal Åland rapakivi. In the middle of the vein, there occurs a clastic sandstone filled fissure about 10 mm wide.

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