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METAMORPHISM AS AN INDICATOR OF EVOLUTION AND STRUCTURE OF THE CRUST IN EASTERN FINLAND

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with 17 figures and 8 tables in the text

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Studies on metamorphism play a key role in establishing the structure and evolution of the crust. It is particularly important in these studies to work out the development of the geothermal gradient, to define the position and continuity of the isograd surfaces and to establish the relationship between metamorphism, magmatism and structures.

The metapelite area of southeastern Finland is characterised by highly developed and extensive metamorphic zonig. The metamorphism is of the low-pressure type. A rise in the grade of metamorphism is not closely related to magmatism; only the volume of migmatising potassium granite increases towards the area of the most intensely metamorphosed thermal dome. Crystallisation took place in this area at a substantially higher temperature but at a lower pressure than in the environment. The metamorphic evolution of the area took place over a relatively long period. Metamorphism began more than 1880 Ma ago, culminated between 1850 and 1810 Ma ago and terminated before the emplacement of the granitoids of the 1800 Ma age group. The zonal metamorphism of the metapelite area of southeastern Finland is reminiscent of the metamorphism of tectonically thickened crust.

Extensive metamorphic zoning has not been encountered in the northern part of the Savo schist belt, with its characteristic metamorphic blocks. The metamorphic features within the blocks are uniform, but between blocks crystallisation conditions change suddenly at the same time as the nature of the metamorphism changes. In some blocks the metamorphism of the granulite facies is associated in both time and space with the hypersthene-granitoid intrusions of the 1880 Ma age group. The variations in the crystallization conditions of the blocks of the granulite facies are related to the depth at which the hypersthene granites were emplaced.

Key words: Metamorphism, zoning, isograds, metamorphic rocks, granites, absolute age, PT-conditions, crust, evolution, Proterozoic, eastern Finland.

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INTRODUCTION

In his concept of metamorphic facies published in 1915, Eskola compared the facies classification of metamorphic rocks with the depth zone classification of Becke and Grubenmann. Eskola stated that the classification of depth zones also was based on mineral composition controlled by pressure and temperature. On page 116 Eskola continues: »The temperature increased downwards, it is true, but the rate of the increase is very different and in a high degree modified by the neighbourhood of magma masses which play such and important part in the rock metamorphism.»

It is only since Miyashiro (1961) perceived the regularity of metamorphic zoning in relation to pressure and temperature that light has been thrown on the problem outlined by Eskola. Miyashiro's study indicated that metamorphism imprints the bedrock with specific structural features and it is closely connected with the evolution of the crust.

Even though pressure and temperature were to increase in a zone undergoing metamorphism as a function of depth, the rise in the metamorphic grade as a function of depth is not straightforward. In their theoretical observations on metamorphic zoning, England and Richardson (1977) have shown that erosion is crucial to the development of metamorphic zoning. A high temperature is not reached in tectonically thickened crust until long after overthrust. Pressure is then diminishing under the impact of erosion, and so the maximum temperature is attained at what is, in a sense, relatively low pressure.

If we assume that the metamorphic isograd surfaces of a tectonically thickened crust are formed horizontally, then even minor tilting of the postmetamorphic crust, by as little as a few degrees, might expose intense metamorphic zoning after erosion. If, however, the metamorphism took place before strong movements, the metamorphic zoning might be completely destroyed in the movements.

If, on the other hand, the temperature of the crust undergoing metamorphism has risen exceptionally high as a result of magmatic activity, then metamorphic zoning associated with time and space would be expected. Temperature rises in areas of magmatic activity more rapidly relative to pressure than in areas in which the geothermal gradient increases in a normal heat flow as a consequence of tectonically thickened crust. Thus the difference in development of the metamorphic zones would be displayed in the character of the isograd surfaces and in the metamorphic zoning in general (cf. England and Richardson op.cit.).

The foregoing implies that study of isograd surfaces and metamorphic zoning in general will enable us to delineate the development of metamorphism and to establish the relationship of metamorphism to the geological events intrinsic to the development of the crust. Figure 1 illustrates an imaginary relationship of the isograd surfaces to magmatism and the structures of the crust. The purpose of the figure is to show the potential of studies on metamorphism for delineating the development of the crust. Wide-ranging structural studies on the target are required though in order to establish the dip and position of the isograd surfaces (hot side up or

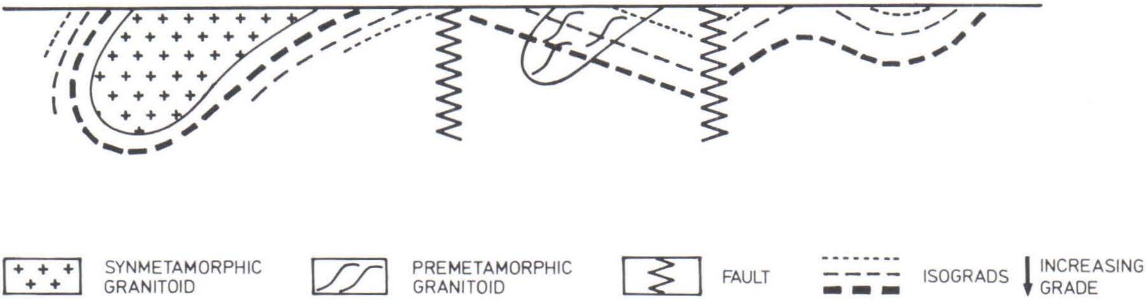

Fig. 1. Schematic presentation of the position of the isograd surfaces in the metamorphic zones.

hot side down); something which has seldom been feasible (cf. St. Onge and Hoffman 1980).

Some studies on the orogenic model of the Finnish bedrock have endeavoured to link metamorphism with the development of the crust. Gaál (1982) has presented a plate tectonic model, chiefly of the Cordillera type, for the Svecokarelian orogeny. In his model he associates the linear areas of granulite facies with island arc formations. Campbell (1980) has come to the conclusion in his studies on metamorphism that the evolution of the Svecokarelides represents continent-continent collision of the Tibetan type, in which the granulite facies metamorphism of the suture belt would have been caused by tectonically thickened crust. Neuvonen et al. (1980) have also favoured the continent-continent collision orogenic model, but in their opinion magmatic activity would have brought about metamorphism of the granulite facies. Korsman (1978, 1982) has paid attention to the variation in the type of metamorphism in the Finnish bedrock, which he attributes to differences in the age of the metamorphism or in the structure of the crust.

The aim of the present study was to clarify the extent to which metamorphism reflects the evolution and structure of early Proterozoic bedrock. An answer has been sought to the problem by investigating the relationship between the areas that underwent metamorphism under conditions of granulite facies and their environment. The study areas are in the suture zone of the Savo schist belt and in the metapelite zone of southeastern Finland, which can be considered as an inferred extension of the suture belt (Fig. 2).

Kalevi Korsman directed the investigation and compiled the summary. Tuula Hautala did the microanalyses. Pentti Hölttä was responsible for the field work in the Pihtipudas-Pielavesi area; he also studied the contact effects of hypersthene granite at Vaaraslahti. Pekka Wasenius participated in the field work in the Kuopio-Suonenjoki area and in compiling the metamorphic maps.

METAMORPHISM OF THE METAPELITE ZONE IN SOUTHEASTERN FINLAND

Conclusions concerning metamorphism in southeastern Finland are based on observations made during the studies on metamorphism in the Rantasalmi-Sulkava area and during bedrock mapping carried out by S. Lavikainen and O. Nykänen in the Tohmajärvi-Rääkkylä-Sim-

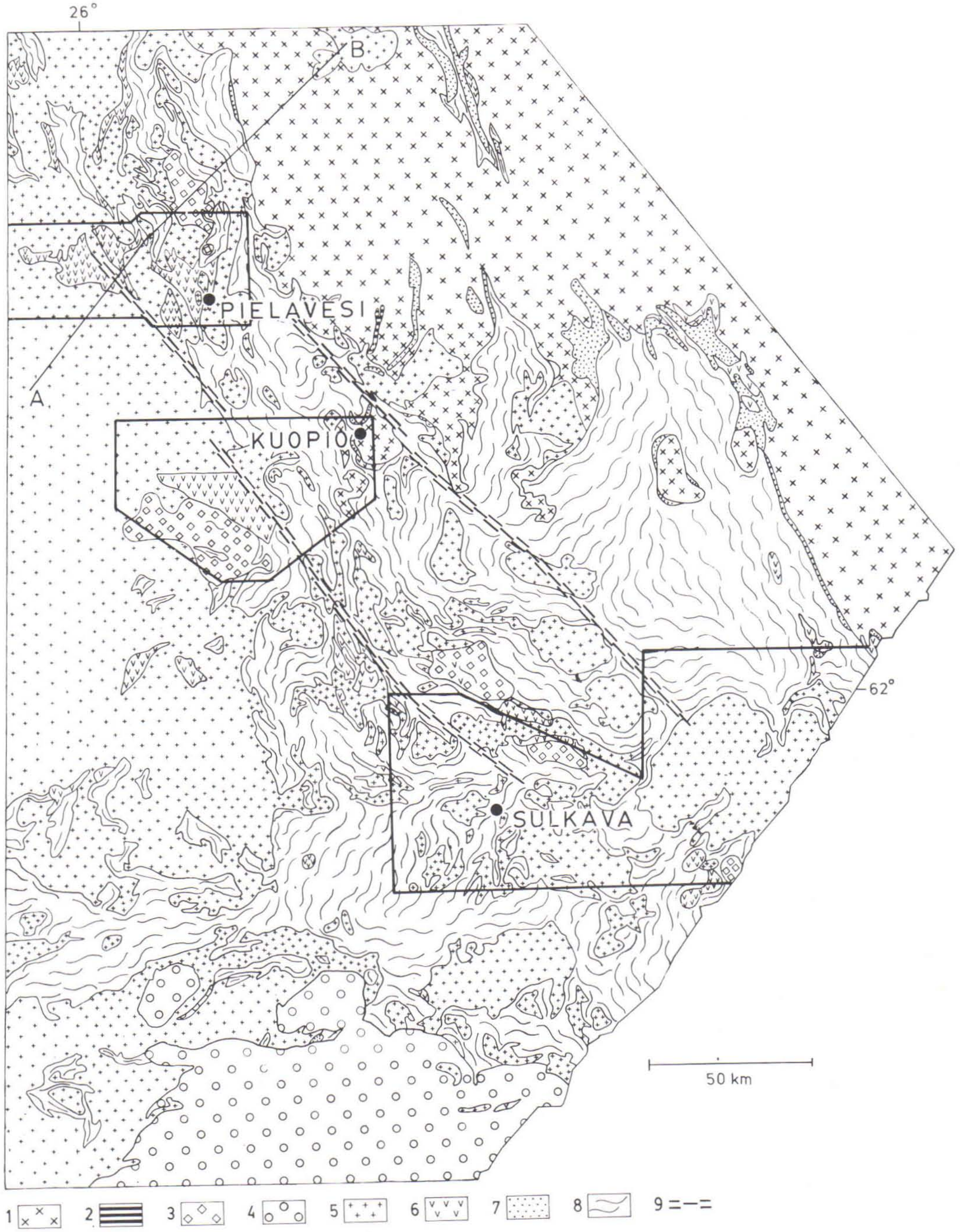

Fig. 2. Geological map of eastern Finland. 1. Archean granite gneiss, 2. carbonatite, 3. hypersthene granitoid, 4. rapakivi, 5. granitoid, 6. volcanite, 7. quartzite, 8. metapelite or veined gneiss, 9. fault or shear zone, A-B a part of the seismic profile SVEKA. Study areas delineated.

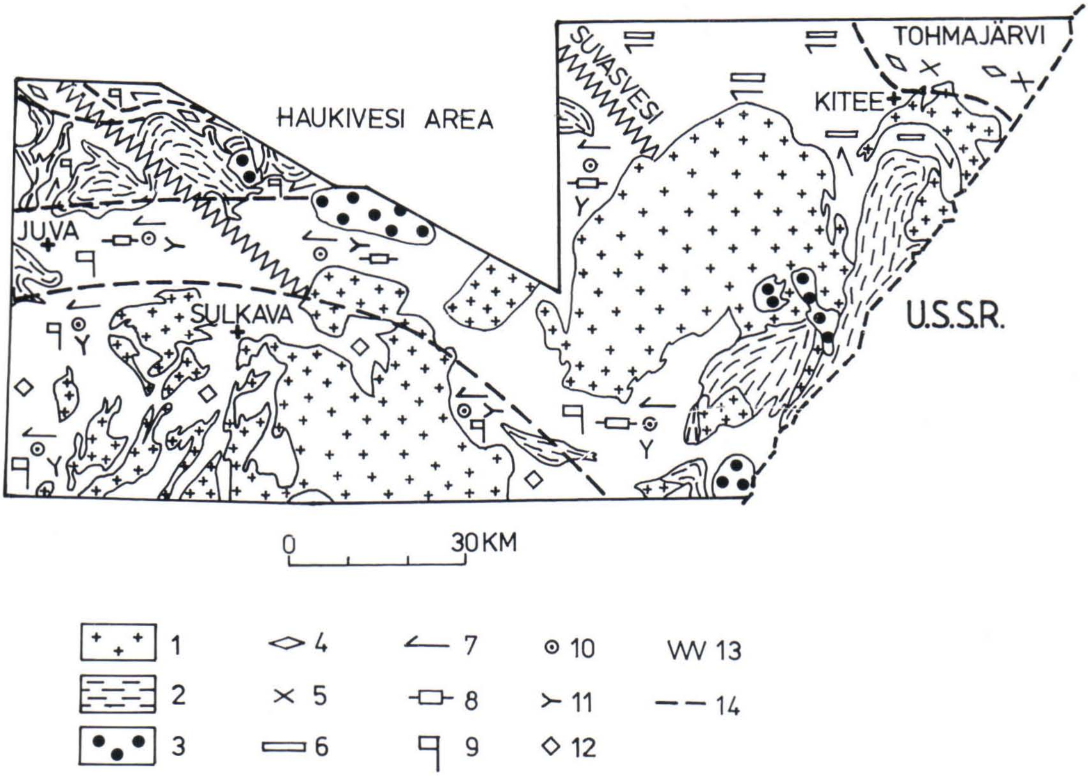

Fig. 3. Metamorphic map of southeastern Finland. 1. granitoid, 2. gneissic tonalite, 3. gabbro or diorite, 4. andalusite, 5. staurolite, 6. muscovite, 7. sillimanite, 8. biotite, 9. K-feldspar, 10. garnet, 11. cordierite, 12. hypersthene, 13. fault or shear zone, 14. isograd.

pele-Pihlajavesi area. The metamorphic maps of the study are shown in Figures 3 and 4. The metamorphic grade in southeastern Finland increases both westwards and southwards (Nykänen 1967, 1972, 1975, 1980, 1982, Korsman 1977).

The Tohmajärvi-Kitee and Rantasalmi area

The Tohmajärvi andalusite-staurolite mica schists in the east represent the lowest grade of metamorphism in the study area. At the boundary of the Tohmajärvi and Kitee zones staurolite and andalusite vanish from the rock almost simultaneously. West of Tohmajärvi, neither andalusite nor staurolite is encountered any more in the muscovite-sillimanite gneisses. Pseudomorphs filled with muscovite and sillimanite have been encountered in the area, however, and these can be interpreted as decomposed staurolite porphyroblasts (S. Lavikainen, pers.comm.).

The andalusite-sillimanite isograd has been mapped in the Rantasalmi area (Fig. 4). Muscovite is in equilibrium at the isograd, but stauro-

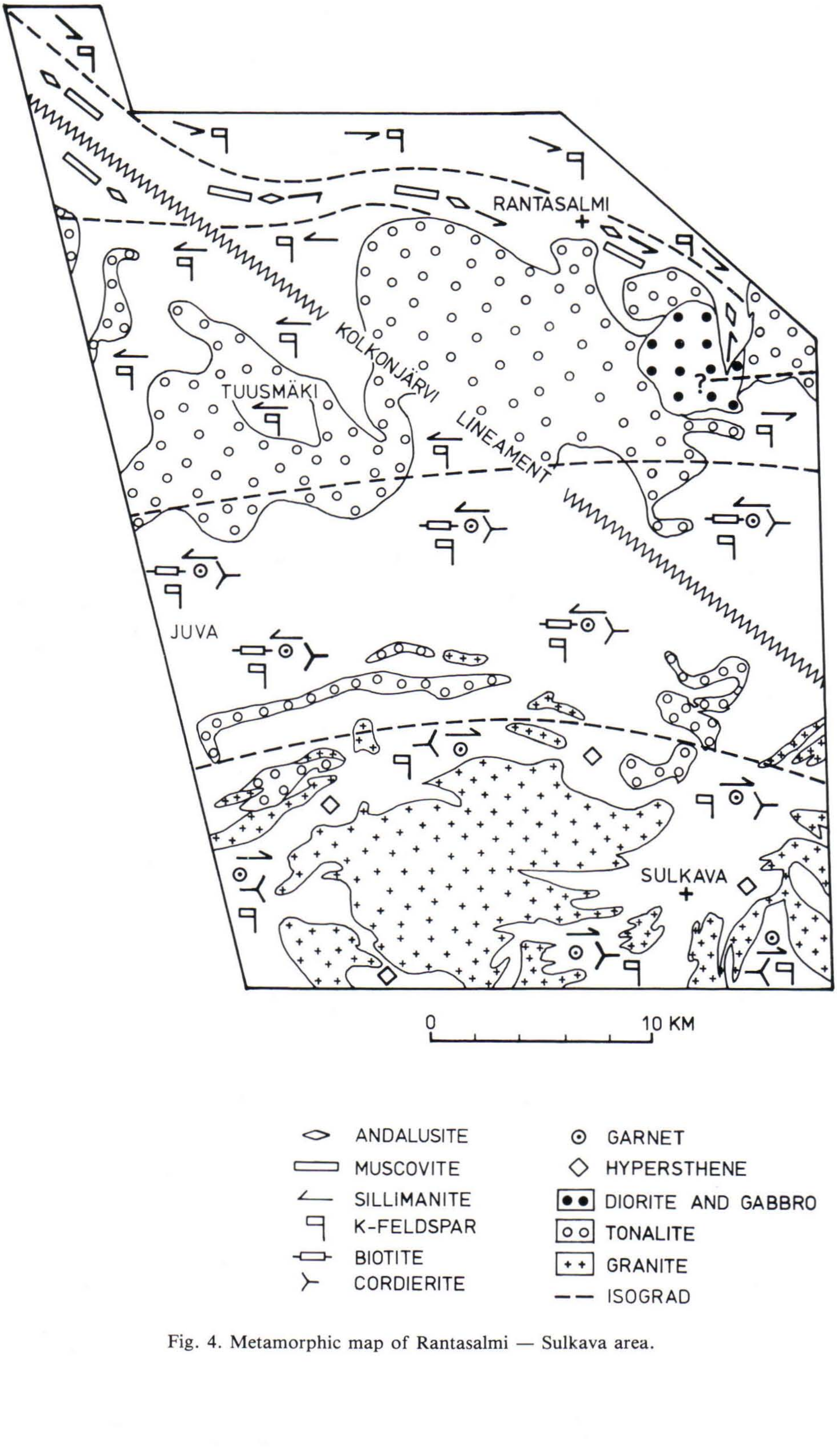

Fig. 4. Metamorphic map of Rantasalmi — Sulkava area.

lite has not been encountered in the andalusite-sillimanite mica schists. The K-feldspar-sillimanite isograd is located in the immediate vicinity of the andalusite-sillimanite isograd.

The following isograd reactions, presented in simplified form, have taken place in the least metamorphosed parts of the study area:

1) staurolite + muscovite + quartz = Al-silicate + biotite
2) muscovite + quartz = sillimanite + K-feldspar.

Andalusite alters into sillimanite in the southeastern part of the study area in the equilibrium field of muscovite, and the polymorphic change takes place under PT conditions very close to those under which staurolite vanishes from the muscovite-bearing metapelites. According to the phase equilibrium fields in Figure 5, staurolite vanishes at a pressure of 4kb and a temperature of 630°C. These values correspond to a geothermal gradient of 45°C/km. The geothermal gradient is the ratio of temperature to the depth at which the mineral assemblage equilibrated, assuming that the pressure is due principally to loading (1 kb = 3.33 km, e.g. Turner 1981).

In the Rantasalmi area the corresponding polymorphic change takes place very close to those conditions under which muscovite breaks down into K-feldspar and sillimanite. The breakdown reaction of muscovite took place in the Rantasalmi area at a temperature of 645°C and at a pressure of 3.4 kb (Fig. 5). The values correspond to a geothermal gradient of 55°C/km.

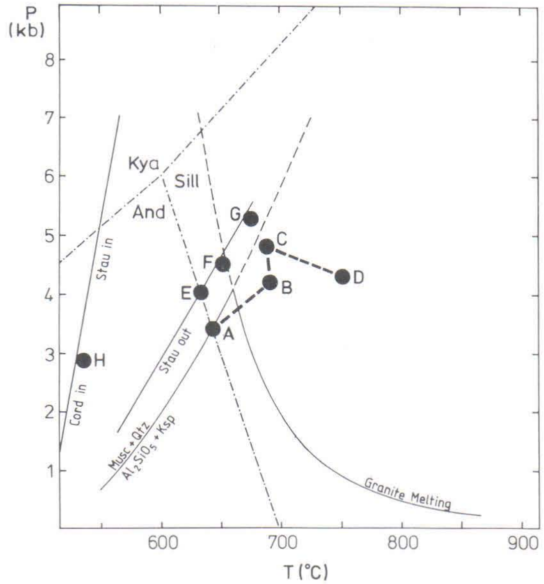

Fig. 5. Crystallization conditions determined from the study areas. A. Rantasalmi, B. Juva (north), C. Juva (south), D. Sulkava, E. Tohmajärvi—Kitee, F. Kuopio and Lampaanjärvi, G. Rautalampi and Pielavesi, H. Pihtipudas. The equilibrium fields of Al-silicates, staurolite and cordierite are from Winkler (1979). The equilibrium curve for muscovite is from Kerrick (1972) and the melting curve of granite from Novgorod and Shkodzinski (1974).

The Juva area

The grade of metamorphism in the Rantasalmi K-feldspar-sillimanite gneiss zone continues to increase southwards. The abundance of muscovite diminishes, and the K-feldspar-cordierite assemblage is equilibrated as the zone finally passes into the biotite-garnet-cordierite zone, called in the present study the Juva area. The boundary of the zones is marked first by anatexis and almost simultaneously by a breakdown reaction of biotite in which garnet and cordierite are in equilibrium with each other

3) biotite + sillimanite + quartz = cordierite + garnet + sillimanite + K-feldspar.

The crystallisation temperature of the northern part of the biotite-garnet-cordierite zone was determined before the present investigation as 660°C (Korsman 1977) and its pressure as

4 kb. The values correspond to a geothermal gradient of 50°C/km. The persistence of the metamorphic zoning and the almost equal values of the geothermal gradients suggest that the migmatic garnet-cordierite gneisses of the northern part of the Juva area belong to the same metamorphic facies series as the K-feldspar-sillimanite gneisses. The andalusite-sillimanite polymorphic change in the Tohmajärvi-Kitee area, on the other hand, took place at somewhat higher pressure than in the Rantasalmi area.

The TiO_2 content of the biotite and the X_{Mg} of the coexisting garnet and cordierite increase southwards in the Juva zone. The first signs of hypersthene in the southern part are encountered in rocks depleted in aluminium in relation to alkalies and calcium. Although the metamorphic grade in the zone appears to increase southwards the Mg-Fe partition coefficients, do not show differences in the crystallisation temperatures between the northern and southern parts of the zone (Table 1 and 2). The crystallisation conditions of the Juva zone were determined from the compositions of garnet and cordierite according to Holdoway and Lee (1977).

Most of the experimental and theoretical investigations indicate that the abundance of magnesium increases in garnet and cordierite as pressure increases whereas the crystallisation temperature affects the partition coefficient. The partial pressure of water does not have any influence on the partition coefficient; the lowered partial pressure of water, however, affects the compositions of the phases in that the X_{Mg} of the phases increases. It can be seen from Table 2 and Figure 5 that the whole of the Juva zone underwent metamorphism at a uniform temperature but that the crystallisation pressure was higher in the southern part than in the northern part of the zone. It has been assumed that the partial pressure of water in the northern part of the zone was 0.4 P_{tot}, because this value does not contradict the findings obtained with other methods (Korsman 1977). The crystallisation of hypersthene and the increase in the TiO_2 content of biotite in the southern part of the Juva zone indicate a drop in the partial pressure of water; hence the partial pressure of water in the southern part is presumed to be 0.2. Although the values of the partial pressure of water were not derived through direct determinations, they do not affect the interpretation given in the present study, provided Holdoway and Lee's concept of the garnet-cordierite barometer and thermometer is correct.

The foregoing implies that temperature is buffered in the Juva zone. On the other hand, the assemblage qtz-ksp-sill-bio-cord-gar, univariant in relation to pressure and temperature, has been encountered throughout the zone even though the biotite in the southern part appears to avoid sillimanite. The partial pressure of water was near total pressure still in the K-feldspar-sillimanite zone (Korsman 1977). Nevertheless, theoretical and experimental investigations indicate that the above six-phase assemblage is not in equilibrium unless the partial pressure of water is considerably lower than the total pressure. The breakdown reaction of biotite in the study area is indeed associated with anatexis. The water released in the breakdown reaction of muscovite promoted anatexis, because the system was not open in relation to water (Korsman op.cit.). By drying the system, anatexis furthered the breakdown reaction of biotite. Anatexis and the breakdown reaction of biotite did not advance very far in the Juva zone, however. The slowdown in the breakdown reaction of biotite is exhibited in the high abundance of biotite in the southern part of the zone as well. The poor progress in metamorphism is demonstrated by the K-feldspar porphyroblasts encountered throughout the Juva zone. The pretectonic nature of these porphyroblasts suggests that some of them were already formed during the breakdown reaction of muscovite (Fig. 7). One reason why anatexis could not advance was the lack of water; chiefly, however, the combination of anatexis and the breakdown reaction of biotite buffered the rise in temperature. If anatexis and the breakdown reaction of biotite had advanced far in the Juva zone, the reaction would have required a very great amount of heat, c. 300-400 cal/°cm^3; the amount of heat required for the total disappearance of muscovite, for example, was less than 100 cal/°cm^3 in the K-feldspar-sillimanite zone. The figures are only relative and were calculated according to Kerrick (1972), England and Richardson (1977) and Holdoway and Lee (1977). The buffering of temperature is exhibited in the almost uniform crystallisation temperature between the southern and northern parts of the zone, even though the compositions of garnet and cordierite indicate that the southern part represents a deeper level of the crust than does the northern part.

The above model shows that only a substantial change in the factors influencing metamorphism, e.g. heat flow, could have caused the breakdown reaction of biotite to be completed and at the same time to have led to pronounced formation of anatectic melt. If metmorphism occurs in a normal heat flow as a result of tectonically thickened crust, then the advanced metamorphism of pelites requires much time as well (cf. Walther and Orwille 1982).

The Sulkava area

The biotite-garnet-cordierite zone of Juva is bound in the south by the garnet-cordierite-sillimanite Sulkava zone, which underwent metamorphism under conditions of granulite facies and at a considerably higher temperature than the Juva zone. Hypersthene generally consti-

Table 1. Microprobe analyses of cordierite and garnet in samples from Vaaraslahti, Pielavesi and Rautalampi (total iron as FeO).

	Vaaraslahti						Pielavesi			
Sample	216PSH78		154-2PSH82		108PSH78		60-3PSH82		120-2PSH83	
	cord	gar	cord	gar	cord	gar	cord	gar	cord	gar
SiO_2	48.4	37.3	47.0	36.9	47.0	36.9	48.9	37.6	48.7	38.2
Al_2O_3	33.1	21.2	32.8	20.7	32.9	21.1	33.7	21.2	33.3	21.3
FeO	9.4	33.7	9.4	34.7	9.9	34.9	7.3	32.3	7.3	30.7
MnO	0.1	0.5	0.1	0.0	0.0	1.4	0.1	0.7	0.0	0.7
MgO	8.4	5.6	8.1	4.6	8.3	4.0	9.5	6.1	9.8	6.6
CaO	0.0	1.6	0.0	1.7	0.0	1.5	0.0	1.4	0.0	1.6
	99.4	99.9	97.4	98.6	98.1	99.8	99.5	99.3	99.1	99.1

	Rautalampi									
Sample	86BPSH81		224PSH81		50PTW80		202PTW79		150APTW80	
	cord	gar	cord	gar	cord	gar	cord	gar	cord	gar
SiO_2	49.3	38.6	48.8	39.1	48.7	37.7	48.8	39.1	49.1	38.3
Al_2O_3	33.2	21.8	33.8	22.1	33.2	21.2	33.6	22.3	33.1	22.4
FeO	7.1	32.1	6.5	31.6	6.1	29.9	6.0	30.2	6.2	31.0
MnO	0.1	1.2	0.1	1.6	0.1	1.9	0.1	1.7	0.1	1.2
MgO	9.5	6.2	9.9	6.6	10.0	7.4	10.2	7.3	10.0	7.4
CaO	0.0	1.3	0.0	0.9	0.0	1.0	0.0	1.3	0.0	1.1
	99.2	101.2	99.1	101.9	98.1	99.1	98.7	101.9	98.5	101.4

tutes an equilibrated phase in aluminium-deficient rocks in the Sulkava zone. The breakdown reaction of biotite and anatexis are far advanced. Biotite is no longer an equilibrated phase with sillimanite, the abundance of biotite being very low in sillimanite-bearing gneisses, and garnet, cordierite and sillimanite are principally restites in migmatites. Pretectonic K-feldspar porphyroblasts are no longer encountered in the Sulkava zone. The values of the Mg-Fe partition coefficients of the cordierite-garnet pair also indicate a very high crystallisation temperature (Table 2). In contrast, the compositions of garnet and cordierite indicate a lower crystallisation pressure than in the southern part of the Juva zone. The partial pressure of water in the Sulkava zone was estimated at 0.2 when assessing the crystallisation conditions from the compositions of garnet and cordierite.

From the foregoing we can say that the Sulkava zone forms a thermal dome in relation to the Juva zone. According to the theoretical study by England and Richardson (1977), the metamorphic zoning of a tectonically thickened crust is related to the rate of erosion. The deeper the metamorphism begins, the more slowly the temperature rises. The part of the crust metamorphosed at the highest temperature might then crystallise at a markedly lower pressure than that prevailing when recrystallisation began. The development of the geothermal gradient suggests that metamorphism in the study area

Table 2. The X_{Mg} of cordierite and garnet, the K_{D-Mg} between the cordierite and garnet and the PT conditions calculated with the method of Holdoway and Lee (1977) from Juva, Sulkava, Vaaraslahti and Pielavesi — Rautalampi. The values of the Juva and Sulkava samples were calculated from the analyses published by Korsman (1977).

Sample	cord X_{Mg}	gar X_{Mg}	K_{D-Mg}	T C°	P Kb
Juva (north)					
33KC	46.4	11.2	6.9	689	4.2
34KC	46.7	11.4	6.8	693	4.2
Juva (south)					
39KC	62.6	19.7	7.2	680	4.7
48KC	63.8	20.1	7.0	686	4.8
43KC	64.1	20.3	7.0	686	4.8
Sulkava					
52GC	58.2	20.5	5.4	765	4.2
53GC	59.4	19.8	5.9	739	4.4
54GC	58.8	20.8	5.4	765	4.3
55GC	57.9	18.9	5.9	739	4.3
57GC	57.5	19.6	5.6	755	4.2
Vaaraslahti					
216PSH78	61.2	21.6	5.7	747	4.4
154-2PSH82	60.3	18.2	6.8	691	4.5
108PSH78	59.9	15.7	8.0	648	4.7
Pielavesi — Rautalampi					
60-3PSH82	69.6	23.8	7.3	678	5.1
120-2PSH83	70.5	26.0	6.8	697	5.1
86BPSH81	70.2	24.0	7.5	671	5.2
224PSH81	72.8	25.5	7.8	660	5.4
50PTW80	74.2	28.5	7.2	682	5.4
202PTW79	74.9	27.9	7.7	664	5.5
150APTW82	73.9	28.2	7.2	682	5.4

took.place largely as described above. Especially the combination of anatexis and the dehydration reaction of biotite appears to have buffered the rise in temperature and slowed down the development of metamorphism in the study area. Nevertheless, the most substantially thickened part of the crust underwent considerable heating. The heated and granitised crust was uplifted by gravitation as the progressive recrystallisation continued. The area of granulite facies thus underwent metmorphism at lower pressure than the immediate environment, as is shown by the compositions of garnet and cordierite.

Detailed metamorphism investigations have not been conducted in the environment of the Sulkava area, except in the northern part. Observations made during geological mapping, however, indicate that the metamorphic grade increases towards the thermal dome from the east, the south or the west (Nykänen 1982, Lavikainen pers.comm., Vorma 1964, Meriläinen 1966, Simonen and Niemelä 1980). Koistinen (1979) has noted that the metamorphic grade of some gneisses at Taipalsaari, 70 km south of Sulkava, corresponds to that in the northern part of the Juva area.

The relationship of metamorphism to granitoids

Observations on the development of the geothermal gradient in the metapelite zone of southeastern Finland suggest that metamorphic zoning is associated with the tectonically thickened part of the crust. It is therefore appropriate to examine the extent to which the granitoids affected the increase in the grade of metamorphism.

The granitoids in the Rantasalmi-Sulkava area can be broadly classified as follows:

- gneissose tonalites
- migmatising microcline granite
- tourmaline-bearing albite pegmatite dykes
- Hirvensalo granodiorite

The gneissose tonalite is encountered mainly in the K-feldspar sillimanite zone as the abundance of tonalite diminishes southwards (Fig. 4). The tonalite exhibits the same foliation planes as the surrounding schists. The metamorphic grade of tonalite likewise increases towards the south. The rock is thus premetamorphic in relation to progressive metamorphism, and it does not show any thermal influence on its environment. Tonalite is almost entirely lacking in the Sulkava zone, which has undergone more intense metamorphism.

The abundance of microcline granite increases from the northern margin of the Juva zone southwards. There is a conspicuous change in the rate of increase at the boundary of the Juva and Sulkava zones. As was pointed out earlier, microcline granite is, at least partly, a product of anatexis and metamorphism. The crystallisation of microcline granite was associated with intense retrograde metamorphism; no observations have been made of its direct progressive influence.

Tourmaline-bearing pegmatite veins occur in the K-feldspar sillimanite zone and to the north of it in the muscovite-mica schist zone. Tourmaline is a metamorphic index mineral since it is no longer encountered in the migmatitic metapelites. Since boron causes a marked drop in the solidus temperature of the granitic system (Chorlton and Martin 1978) it is possible that tourmaline-bearing albite pegmatites, too, are products of metamorphism. Moreover, the veins occur in less intensely metamorphosed zones; therefore, the pegmatites might be either segregations of early anatexis or the final crystallisation products of microcline granite that intruded into the metamorphic structure at a relatively high level.

The Hirvensalo granodiorite in the margin of the Sulkava thermal dome is postmetamorphic (Korsman and Lehijärvi 1973). Apart from the magmatic orientation it is practically unorientated. The postmetamorphic nature of granodiorite is further manifested in its tendency to cut the microcline granite. Moreover, andalusite has crystallised at the contact with granodiorite (P. Pitkänen, pers. comm.).

There is no indication on the map that metamorphic zoning is associated with granitoids in the eastern part of the study area either. As in the Rantasalmi-Sulkava area, the abundance of migmatising potassium granite appears to increase towards the Sulkava thermal dome.

The foregoing suggests that the granitoids are not connected with metamorphic zoning in any way other than that the abundance of migmatising potassium granite increases as the metamorphic crystallisation temperature rises. The observation corroborates the view that zoned metamorphism was caused by tectonically thickened crust.

Metamorphic structure and development

At the present level of erosion the muscovite-bearing mica schists of the Rantasalmi-Sulkava area grade into garnet and cordierite-bearing migmatites over a distance of 10 to 12 km. Since the change in the metamorphic grade is continuous and the geothermal gradient is 50°C/km, the isograd surfaces have tilted 1°—3° towards north. A corresponding change in the metamorphic grade has been observed in the eastern part of the study area over a distance of 40 km. The isograd surfaces are therefore likely to be almost horizontal.

The extensions of the isograd surfaces above and below the erosion level are naturally only inferred; nevertheless, they are justified by the continuity of the change in the metamorphic grade and the regularity in the metamorphic geothermal gradient. In any case, the change in the intensity of the metamorphic grade reflects the angle between the isograd surfaces and the level of erosion.

The regular change in the metamorphic grade of the least metamorphosed zones in the study area and the rectilinear continuity of the isograds are suggestive of the post-tectonic character of the progressive metamorphism in relation to the formation of the macrostructures. The isograd surfaces might thus have preserved the almost horizontal position attributed to the tectonically thickened crust. The relatively young age of the metamorphism is indicated by the premetamorphic character of the Tuusmäki tonalite in relation to progressive metamorphism. The tonalite at Tuusmäki is exposed on the level of erosion at the axis culmination, that is, the tonalite forms a domelike structure. Geological mapping (Vorma 1971), Korsman 1973) indicates that the Tuusmäki tonalite is located in a complicated structure that, from observations made at Tuusmäki, constitutes an anticlinorium. The isograds of progressive metamorphism in the K-feldspar-sillimanite gneiss zone cut the Tuusmäki structural elements. Sillimanite and especially K-feldspar are post-tectonic in relation to the oldest structural elements (Fig. 6), which can be observed in both the Tuusmäki tonalite and the schists surrounding the tonalite. It should be noted, however, that the development of schistosity in the progressively metamorphosed area was not episodic but progressive. The external and internal schistosities of the porphyroblasts in the study area belong to the same stage of deformation, and the internal schistosity represents only an earlier episode of deformation than does the external schistosity (cf. Fig. 7). The recrystallisation of

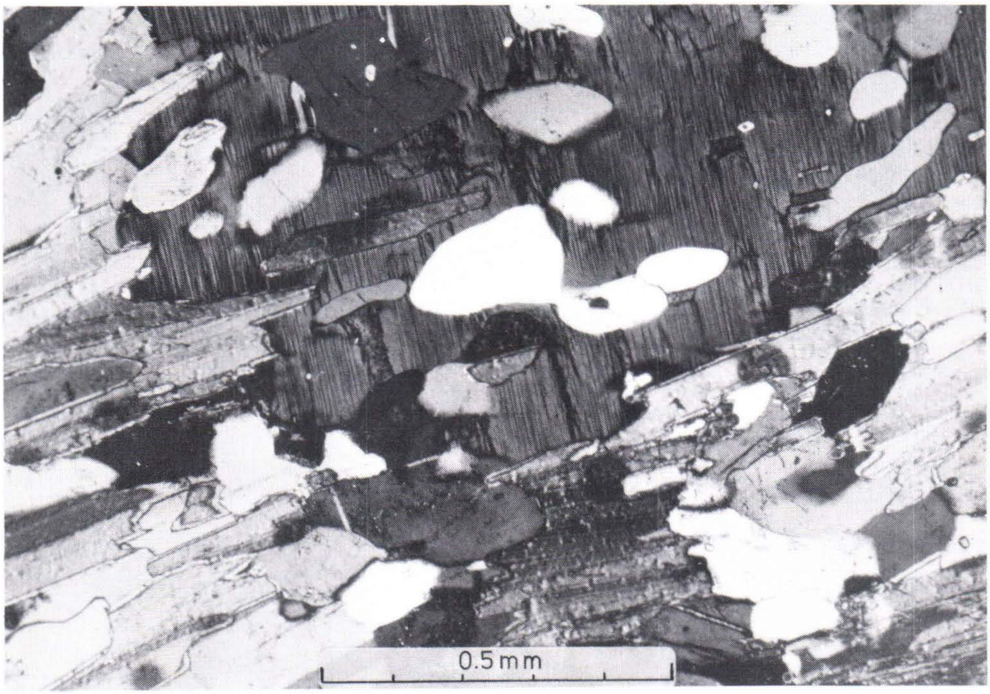

Fig. 6. K-feldspar overprinting foliation.

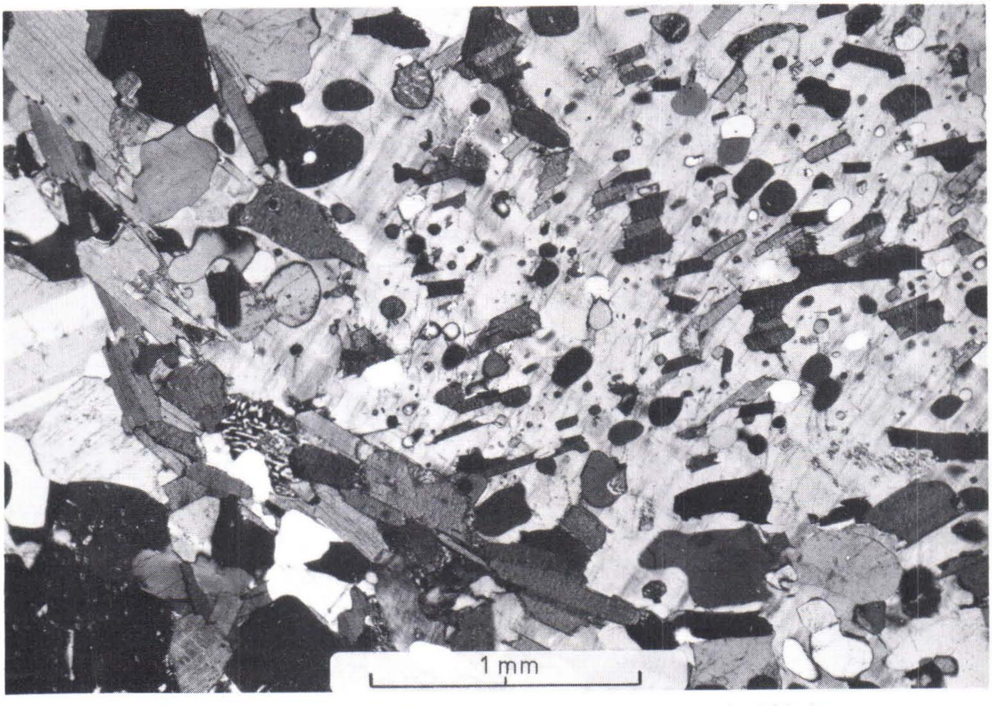

Fig. 7. Partly pretectonic K-feldspar in relation to coarse-grained biotite.

muscovite points to the post-tectonic character of the progressive metamorphism. In the least metamorphosed areas the muscovite is lameller and exhibits a foliation plane parallel to that of biotite. Muscovite recrystallises as porphyblasts a short way before the K-feldspar-sillimanite isograd, and is now post-tectonic in relation to the plane of schistosity to which it was syntectonic a little farther north (cf. Guidotti 1970). This observation implies that the muscovite-bearing metapelites in the study area did not take part in the progressive metamorphism. And, indeed, the metapelites exhibit deformation that is older than that in the gneissic tonalites. Hence one stage of metamorphism already preceded the intrusion of the tonalites.

Structurally, the Juva zone constitutes a synclinorium, with an axial plane trending 260°—270°, between two dome areas. Garnet and cordierite were crystallised syntectonically in relation to the axial plane foliation. K-feldspar, on the other hand, cystallised before the formation of the axial plane foliation trending 260°—270°, as demonstrated by the pretectonic K-feldspar porphyroblasts (Fig. 7). The Juva zone evidently underwent the same metamorphic stages as did the K-feldspar-sillimanite zone. Progressive metamorphism and the associated deformation have left their mark on the Juva zone much more decisively than on its less metamorphosed environment. The above indicates that the axial plane foliation trending 260°—270° is younger than the elements of the Tuusmäki dome structure. The temperature is buffered in the Juva zone, but the composition of cordierite in equilibrium with garnet shows that the southern part of the area crystallised at higher pressure than did the northern part, in other words, the southern part represents a deeper level of the crust than the northern part.

Crystallisation in the area of the Sulkava thermal dome probably started deeper and at a higher pressure than suggested by the compositions of garnet and cordierite. It is important for the metamorphic structure and the evolution of the whole area that the Sulkava area constitutes the core of the thermal dome in relation to its environment; structurally, too, the Sulkava area constitutes the core of the dome. The subhorizontal foliation planes envelop microcline granites in the high-temperature core. Outside the dome the foliation planes grow steeper yet encircle the core of the dome (Simonen 1980). These features corroborate the concept of gravitative uplifting of the core. Intense anatexis and migmatisation are closely connected with the evolution of the core. This explains the pronounced banding visible at Sulkava, where the paleosome consists primarily of quartz, sillimanite, garnet and cordierite. The Sulkava area has naturally undergone many phases of metamorphism and deformation, but, as demonstrated by their compositions, garnet and cordierite were equilibrated during a synkinematic phase of the genesis of the dome. The microcline granite has crosscutting contacts and it was crystallised during intense retrograde metamorphism (Korsman 1977). Because the above enables us to date the development of metamorphism to at least some extent in relation to granitoids, some radiometric age data are given for the Rantasalmi-Sulkava area.

The gneissose tonalite of Tuusmäki belongs to the 1880 Ma age group (Table 3 and Fig. 8), whereas the postmetamorphic granodiorite of Hirvensalo is one of the 1800 Ma old granitoids (Table 4 and Fig. 9). The analytical data on zircons and monazites separated from the paleosome and neosome of the Säviönsaari migmatite are given in Table 5 and Figure 10. The migmatite, which is located in the area of the Sulkava thermal dome, is composed of garnet, cordierite and sillimanite-bearing paleosome cut by microcline granitic neosome.

Long euhedral (fraction C) and blunt (fraction D) crystals of zircon have been encountered in the neosome. Both types of crystal give the same diffusion model age (1859 Ma and 1857 Ma correspondingly). The mixed fraction of the crystal types is represented by analysis point E (untreated) and point F (preleached uncrushed in cold 5 % HF for ten minutes). The euhedral, coarse crystals from fractions. 4.0—4.2 (G) and 3.6-3.8 g/cm^3 (I) fit the same isochron even in case the fraction I, having almost 3000 ug/g uranium, was pulverized and preleached in HF. The data for six zircon fractions show distinct but different degrees of discordance on the concordia plot and follow the content of uranium from nearly 600 to 3000 ug/g.

It has been shown that differential leaching of uranium and lead may occur in the leach experiment. However, much common lead was removed from fractions F and I and the residue of fraction I is certainly more concordant. ^{207}Pb-^{206}Pb ages increase regularly along the concordancy from fraction I (1780 Ma) to fraction D (1823 Ma). The HF-washed fraction F is offset showing ^{207}Pb-^{206}Pb age of 1807 Ma. Nevertheless, the approximate collinearity of the six fractions points to a chord having an upper intersection age of 1833 ±16 Ma (2 sigma) and 1823 ± 10 Ma if monazite (fraction A) is included. The analysis points of monazite are slightly above the concordia, the fraction B leached in HF still more.

Several factors indicate that the zircon of the paleosome represents metamorphic crystallisation: it has many crystal faces; it is transparent and it is of uniform quality. Its uranium content is low and its age is highly concordant.

Monazite has been separated from two samples of the paleosome. One of them gives a concordant age of 1835 Ma; the other is slightly discordant, being about 1843 Ma. Two monazites have been analysed from the neosome; both give an age of about 1817 Ma.

The other observations reported above indicate that the metamorphism in the Sulkava thermal dome area culminated after the emplacement of the 1880 Ma quartz diorites, but that progressive metamorphism terminated before the emplacement of the 1800 Ma Hirvensalo granodiorite. Even the retrograde phase associated with the microcline granite is older than the

Table 3. U-Pb analytical data on zircons from Tuusmäki tonalite (A596).

Zircon fraction	Concentration		Atom ratios						Radiometric ages		
d = density Ø = size in μm HF = preleached in HF	ug/g		blank corrected			corrected for blank and common lead			Ma		
	^{238}U	^{206}Pb radio-genic	$\frac{^{206}Pb}{^{204}Pb}$	$\frac{^{207}Pb}{^{206}Pb}$	$\frac{^{208}Pb}{^{206}Pb}$	$\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{207}Pb}{^{206}Pb}$	$\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{207}Pb}{^{206}Pb}$
A596, tonalite; Tuusmäki, Rantasalmi											
A d > 4.6	311.7	82.96	3431	.11774	.09296	.3076 ± 16	4.826 ± 27	.11379 ± 17	1729 ± 9	1789 ± 5	1861 ± 5
B 4.0 < d < 4.2 Ø < 70	774.8	175.96	2810	.11722	.10275	.2625 ± 13	4.067 ± 22	.11239 ± 12	1502 ± 7	1647 ± 4	1838 ± 4
C 4.2 < d < 4.6	494.6	123.66	3543	.11749	.09113	.2890 ± 15	4.528 ± 24	.11366 ± 11	1636 ± 8	1736 ± 5	1858 ± 3
D 4.2 < d < 4.6	469.4	119.80	3750	.11731	.09024	.2950 ± 15	4.624 ± 25	.11370 ± 8	1666 ± 8	1753 ± 4	1859 ± 3
E 4.2 < d < 4.6; HF	277.9	77.74	13061	.11631	.09192	.3233 ± 17	5.138 ± 28	.11528 ± 12	1805 ± 8	1842 ± 5	1884 ± 4

Table 4. U-Pb analytical data on zircons and titanites from Hirvensalo granodiorite (A267).

Zircon fraction	Concentration		Atom ratios						Radiometric ages		
d = density Ø = size in μm HF = preleached in HF	ug/g		blank corrected			corrected for blank and common lead			Ma		
	^{238}U	^{206}Pb radiogenic	$\frac{^{206}Pb}{^{204}Pb}$	$\frac{^{207}Pb}{^{206}Pb}$	$\frac{^{208}Pb}{^{206}Pb}$	$\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{207}Pb}{^{206}Pb}$	$\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{207}Pb}{^{206}Pb}$
A267, granodiorite; Hirvensalo, Anttola											
A Ø > 70	298.3	76.33	2924	.11347	.11808	.29579	4.4372	.10881	1670 ± 11	1719 ± 8	1779 ± 12
borax fusion						± 210	± .0428	± 59			
B Ø < 70	298.0	76.21	1039	.12200	.14351	.29560	4.4375	.10888	1669 ± 12	1719 ± 8	1780 ± 9
borax fusion						± 237	± .0437	± 48			
C titanite	84.5	22.46	283	.15637	.83322	.30737	4.5819	.10812	1727 ± 12	1745 ± 14	1768 ± 25
borax fusion						± 239	± .0781	± 137			
C titanite/HF	85.3	23.14	380	.14376	.80669	.31356	4.6637	.10788	1758 ± 8	1760 ± 10	1764 ± 19
						± 166	± .0542	± 99			
D d > 4.6	206.6	51.38	1820	.11583	.13196	.28747	4.2937	.10833	1628 ± 7	1692 ± 5	1771 ± 5
						± 148	± .0245	± 21			
E 4.2 < d ± 4.6	311.2	76.59	1853	.11565	.12325	.28449	4.2475	.10829	1613 ± 7	1683 ± 5	1771 ± 5
						± 148	± .0239	± 18			
E 4.2 < d < 4.6; HF	285.1	71.29	5445	.11182	.11174	.28900	4.3558	.10932	1636 ± 7	1703 ± 4	1788 ± 3
						± 150	± .0233	± 7			
G 4.2 < d < 4.6; HF	285.2	74.38	3247	.11376	.11930	.30136	4.5522	.10956	1698 ± 8	1740 ± 5	1792 ± 5
						± 156	± .0252	± 16			
H d > 4.6; HF	302.8	79.99	3587	.11296	.11511	.30533	4.5955	.10916	1717 ± 8	1748 ± 5	1785 ± 3
						± 158	± .0246	± 9			

Fractions A-C recalculated from Korsman and Lehijärvi, 1973

Table 5. U-Pb analytical data on zircons from Säviönsaari migmatite (A14, neosome and A22, paleosome).

Zircon fraction d = density Ø = size in μm HF = preleached in HF	Concentration ug/g ^{238}U	Concentration ug/g ^{206}Pb radiogenic	Atom ratios, blank corrected $\frac{^{206}Pb}{^{204}Pb}$	Atom ratios, blank corrected $\frac{^{207}Pb}{^{206}Pb}$	Atom ratios, blank corrected $\frac{^{208}Pb}{^{206}Pb}$	Atom ratios, corrected for blank and common lead $\frac{^{206}Pb}{^{238}U}$	Atom ratios, corrected for blank and common lead $\frac{^{207}Pb}{^{235}U}$	Atom ratios, corrected for blank and common lead $\frac{^{207}Pb}{^{206}Pb}$	Radiometric ages Ma $\frac{^{206}Pb}{^{238}U}$	Radiometric ages Ma $\frac{^{207}Pb}{^{235}U}$	Radiometric ages Ma $\frac{^{207}Pb}{^{206}Pb}$
A14, neosome in migmatite; Säviönsaari, Sulkava											
A monazite; d > 4.2	2701	764.5	10848	.11227	6.925	.3271 ± 17	5.007 ± 50	.11101 ± 8	1824 ± 8	1820 ± 4	1816 ± 3
B monazite; d > 4.2 HF	2818	804.5	14448	.11206	6.556	.3300 ± 17	5.055 ± 27	.11112 ± 11	1838 ± 9	1828 ± 4	1818 ± 4
C d > 4.2 long crystals	828.4	171.16	835.3	.12603	.08342	.2388 ± 13	3.613 ± 21	.10975 ± 21	1380 ± 7	1552 ± 5	1795 ± 5
D d > 4.2 subhedral	642.6	151.75	706.0	.13067	.11996	.2729 ± 15	4.194 ± 28	.11144 ± 33	1555 ± 8	1672 ± 5	1823 ± 8
E d > 4.2 unsettled	695.1	156.59	1005	.12420	.10015	.2604 ± 14	3.973 ± 21	.11069 ± 8	1491 ± 7	1628 ± 4	1810 ± 3
A 22, paleosome in migmatite; Säviönsaari, Sulkava											
A monazite; d > 4.2 Ø > 160	1769	503.3	30743	.11265	6.519	.3289 ± 17	5.087 ± 27	.11220 ± 6	1832 ± 8	1834 ± 5	1835 ± 3
B monazite; d > 4.2 Ø > 160	2332	671.2	28580	.11317	5.281	.3326 ± 17	5.168 ± 27	.11269 ± 7	1851 ± 9	1847 ± 5	1843 ± 3
C monazite; d > 4.2 Ø > 160; HF	2195	642.6	45973	.11317	5.549	.3383 ± 18	5.265 ± 28	.11288 ± 6	1878 ± 9	1863 ± 5	1846 ± 3
D d > 4.6	389.6	107.97	14185	.11115	.07285	.3203 ± 17	4.867 ± 26	.11019 ± 8	1791 ± 8	1796 ± 4	1802 ± 3
E d > 4.6; HF	382.0	105.75	44956	.11062	.05924	.3200 ± 17	4.866 ± 26	.11032 ± 5	1789 ± 8	1796 ± 5	1804 ± 2
F d > 4.2; HF not crushed	594.1	145.26	2554	.11578	.09144	.2826 ± 15	4.304 ± 38	.11045 ± 70	1604 ± 8	1694 ± 8	1807 ± 14
G 4.0 < d < 4.2 Ø > 160	1138.0	223.675	582.5	.13335	.09536	.2272 ± 12	3.445 ± 50	.11000 ± 130	1319 ± 6	1514 ± 11	1799 ± 24
I 3.6 < d < 3.8 Ø 160; HF pulverized	2957	536.78	2358	.11464	.03356	.2098 ± 11	3.149 ± 20	.10885 ± 32	1227 ± 5	1444 ± 5	1780 ± 7

$\lambda_{^{238}U} = 1.55125 \times 10^{-10}/a$
$\lambda_{^{235}U} = 9.8485 \times 10^{-10}/a$
Atomic ratio $^{238}U/^{235}U = 137.88$

$(^{206}Pb/^{204}Pb)_{measured}$:

A14A	10092 ± 262	C	32780 ± 3828
B	13508 ± 671	D	12768 ± 842
A22A	25910 ± 1970	E	32179 ± 1763
B	24570 ± 932	F	14304 ± 754

(fraction A22A from a separate rock sample)

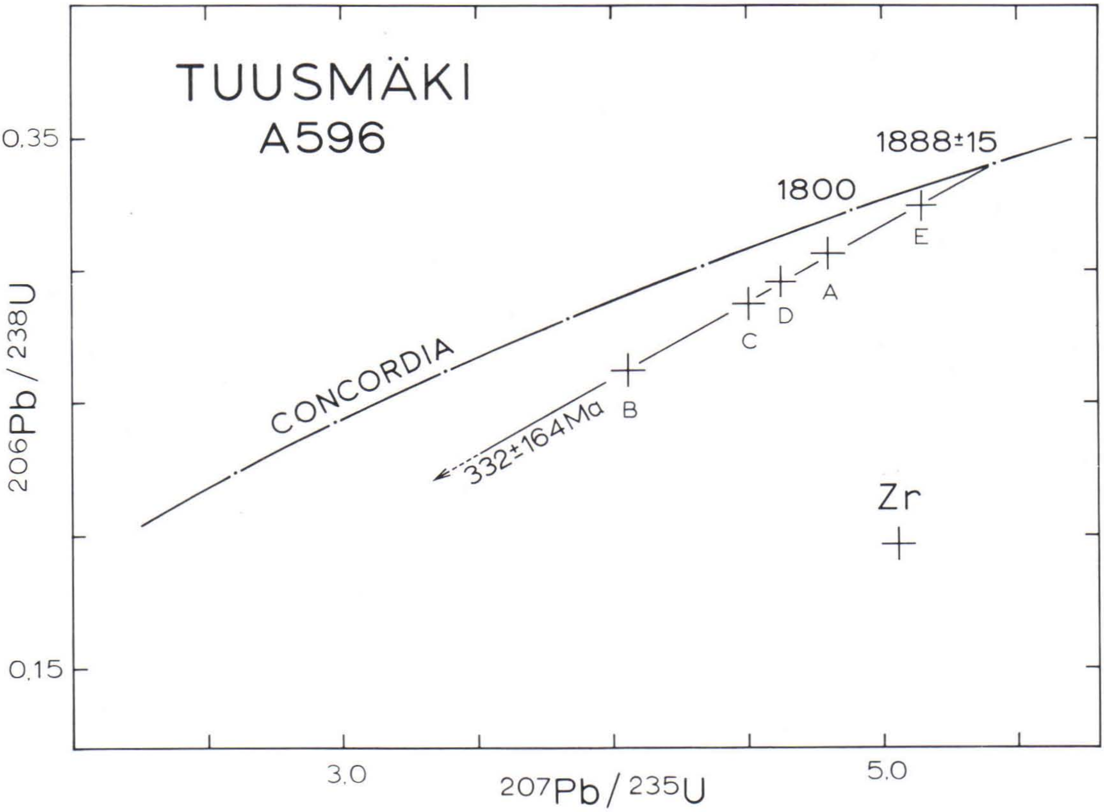

Fig. 8. Concordia diagram for zircons from the Tuusmäki tonalite. The diskordia array through data points gives an upper concordia intercept at 1888 ± 15 Ma (2 sigma error).

Hirvensalo granodiorite. There is no conflict between the dates obtained for the Säviönsaari migmatite and other observations. The discrepancy in the ages of the Säviönsaari paleosome and neosome implies that they have a different crystallisation history. This was not altogether unexpected, taking into consideration the intrusive nature of the neosome in relation to the paleosome. The age of zircon in the paleosome indicates that relatively intense recrystallisation took place 1810 Ma ago. The age of the monazite in the neosome (1817 Ma) is comparable, within the limits of analytical error, with the age of the zircon in the paleosome. These ages come very close to that of the retrograde phase, provided that the age of zircon in the Hirvensalo granodiorite (1800 Ma) reflects the emplacement of the granodiorite. The age of the zircon in the neosome 1833 ± 16 Ma and of the monazite in the paleosome (1840 Ma) might be related to the culmination stage of metamorphism. As was stated earlier, the thermal dome of Sulkava has a complex metamorphic history. Intense metamorphic crystallisation in associated with the uplifting of the dome, as shown by the compositions of garnet and cordierite. Nevertheless, gravitative uplifting could not have taken place until a fairly late stage of metamorphism. The age of zircon in the neosome was determined from six discordant data. Some zircons have a brown core and a light mantle. Our intention is to continue the work by getting more analytical

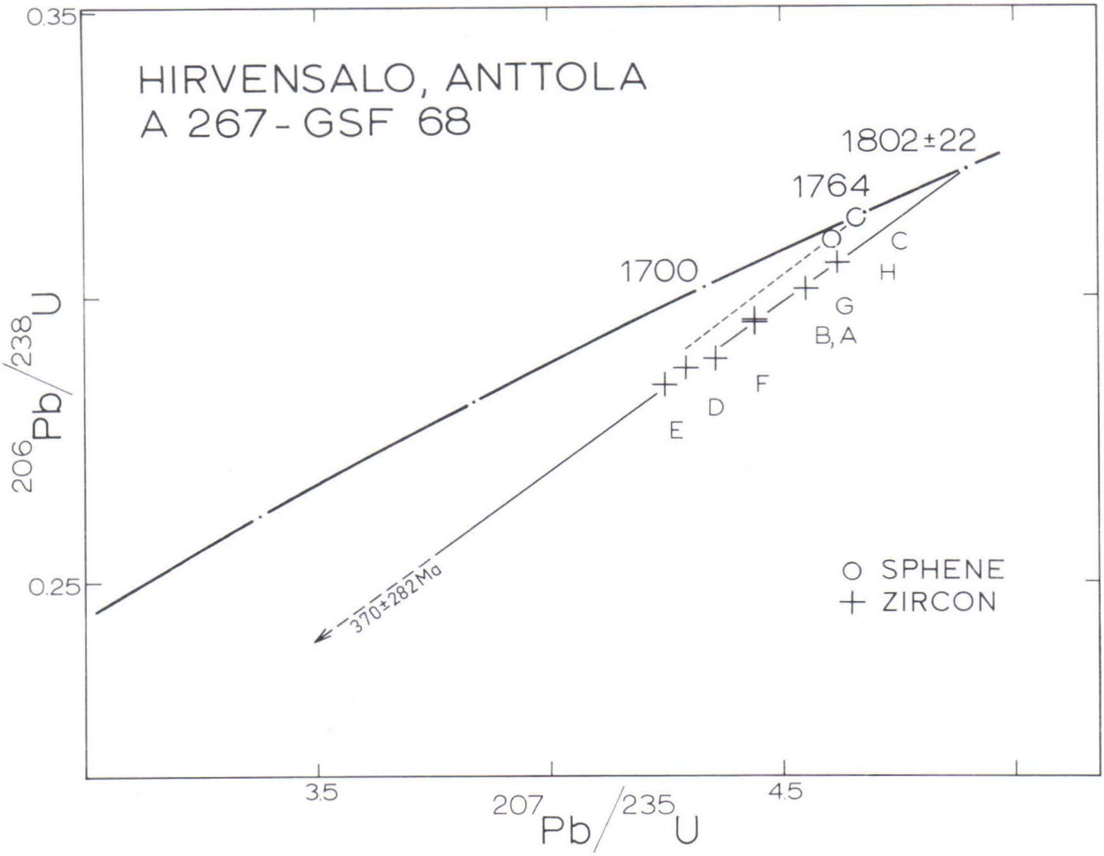

Fig. 9. Concordia plot for zircons and titanites from the Hirvensalo granodiorite.

data on the neosome and by investigating zircons from the other metamorphic zones in the Rantasalmi-Sulkava area.

It should be remembered that observations from Rantasalmi, particularly those from the northern part of the area, show that the evolution of metamorphism started before the emplacement of the Tuusmäki quartz diorite over 1880 Ma ago.

The K-Ar dating of biotite and hornblende (Table 6) ascribes ages of 1749—1773 ± 52 Ma (biotite) and 1770—1796 ± 54 Ma (hornblende) to the samples from the muscovite zone and the K-feldspar-sillimanite zone. The ages of the biotite in the samples from the Juva and Sulkava migmatite areas are 1702—1721 ± 51 Ma. From what was noted concerning the relationship of the age of metamorphism to the ages of the granitoids it is clear that the K-Ar datings do not represent the age of metamorphism but are associated with the cooling of the bedrock. Although there is a considerable error in the age data, there would seem to be an age difference between the migmatised zones and the less metamorphosed zones. Hence the K-Ar datings imply that the migmatised areas evolved later than their environment. If the age difference is a true one, it corroborates the concept that the temperature was buffered and that the reactions slowed down at the onset of melting.

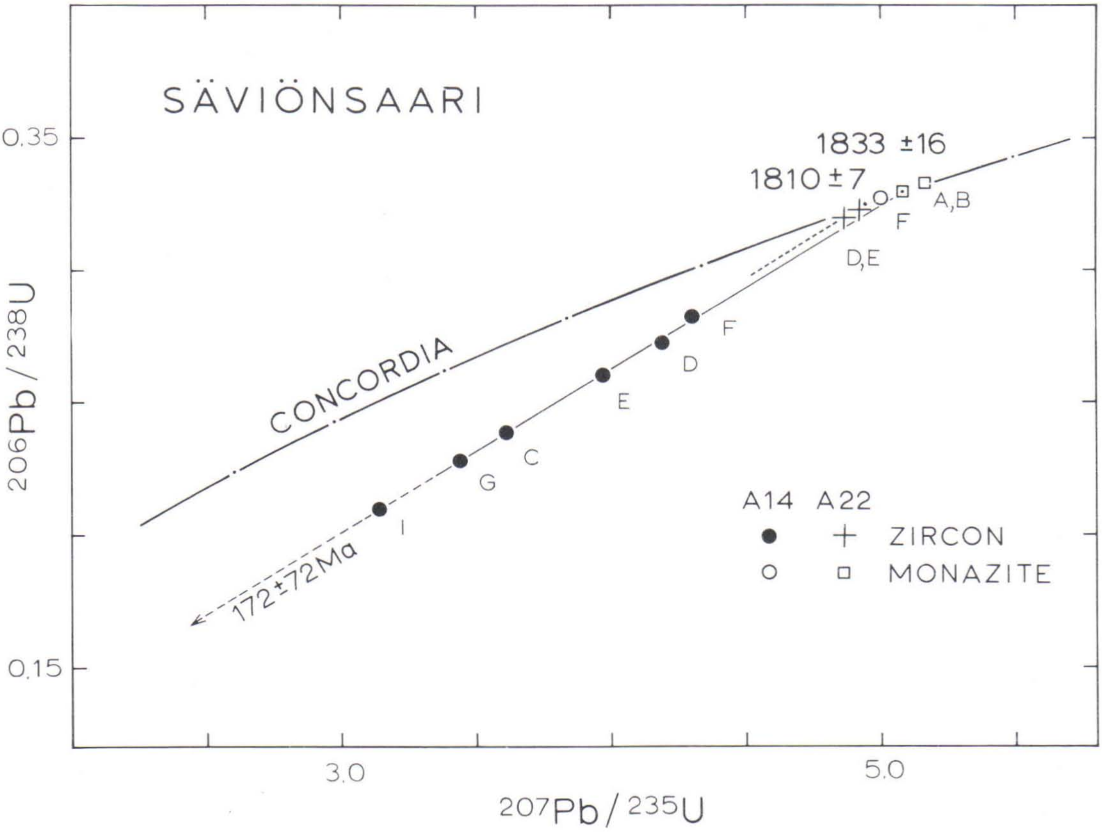

Fig. 10. Concordia plot for zircons and monazites from paleosome (A22) and neosome (A14) in migmatite on the island Sāviönsaari.

The regularity and continuity of the change in the metamorphic grade further indicate that faulting had little impact on the metamorphic structure: either the movements associated with faulting were of such magnitude that they cannot be verified through the variation in the metamorphic grade or then the faults are premetamorphic.

The Kolkonjärvi lineament, which trends 320° in the Rantasalmi-Sulkava area, is characterised in many places by mylonitisation and metasomatic alterations. The K-feldspar-sillimanite isograd appears to cut the lineament (Fig. 4), but the andalusite-sillimanite isograd turns as it approaches the lineament so that it is parallel to it. Furthermore, to the east of the lineament, the metamorphic grade increases both southwards and northwards from the andalusite-sillimanite isograd but to the west of the lineament the metamorphic grade increases only southwards. The rise in the metamorphic grade towrds the north is, however, associated with the Haukivesi magmatic area (cf. Neuvonen *et al.* 1980).

Nykänen (1975 b) has described a sudden change in metamorphic grade from the Pistalanjärvi-Raikuu line, which is part of the Suvasvesi lineament trending 320° (Fig. 3), and which Halden (1982) has interpreted as a wrench fault. To the east of the fault there are muscovite-

Table 6. K-Ar datings of biotite and hornblende from the Rantasalmi — Sulkava area. ($\lambda\beta$: 4.962 × 10^{-10}/y, $\lambda\varrho$: 0.581 × 10^{-10}/y ^{40}K = 0.01167 atomic per cent.

Sample	% K	vol. ^{40}Ar	% ^{40}Ar rad.	Age (Ma)
Muscovite zone (Rantasalmi)				
OOBPTW83 biotite	7.63 ± 0.05	89.33	99.4	1773 ± 53
		89.563	99.6	
13B biotite	7.79 ± 0.04	89.389	99.4	1750 ± 52
		89.529	99.4	
K-feldspar-sillimanite zone Tuusmäki				
8TTK83 biotite	7.57 ± 0.04	87.112	99.3	1751 ± 52
		86.898	99.3	
168AP69 hornblende	0.225 ± 0.03	2.6893	90.1	1796 ± 54
		2.6943	95.0	
A596 hornblende	0.580 ± 0.003	6.7859	96.4	1770 ± 53
		6.7833	90.5	
biotite	7.41 ± 0.04	85.208	98.8	1749 ± 52
		84.823	98.7	
Biotite-cordierite-garnet zone Juva				
18KAK73 biotite	7.54 ± 0.08	84.166	99.4	1716 ± 51
		83.909	99.2	
128KAK73 biotite	7.86 ± 0.05	88.123	99.0	1721 ± 52
		87.823	99.3	
129BKAK73 biotite	7.85 ± 0.05	86.485	99.3	1702 ± 50
		86.201	99.1	
Garnet-cordierite-sillimanite zone				
66/KK169 biotite	7.36 ± 0.06	81.500	99.0	1712 ± 51
		81.899	99.3	

bearing mica schists and to the west of it garnet and cordierite-bearing gneisses.

The metapelite zone of southeastern Finland is characteristed by well developed metamorphic zoning. Progressive is not directly related to magmatic activity. The area of the intensely anatectic thermal dome was metamorphosed at a very high temperature but at low pressure. Taken as whole, metamorphism in the study area was a fairly long-lasting process. The features described above might well refer to metamorphism in a tectonically thickened crust.

There are indications in the study area of subhorizontal isograd surfaces, as is also suggested by the buffering of temperature in the Juva zone. Nevertheless, the formation of the Sulkava thermal dome and the heavy rise in temperature cannot be attributed to a tectonially thickened crust alone.

METAMORPHISM OF THE SAVO SCHIST BELT

Pelites and volcanites metamorphosed under conditions of granulite facies have been encountered in a belt trending N-S in the Kiuruvesi-Rautalampi zone. The volcanites, which are characterised by the low abundance of calcium and alkalies, are composed of quartz, garnet, cordierite and hypersthene. In chemical composition the rocks resemble hydrothermally altered volcanites (Huhtala 1979). Studies were conducted along two profiles across the belt of the granulite facies, one in the Pihtipudas-Pielavesi area and the other in the Kuopio-Rautalampi area (Fig. 2). The study areas belong partly to the main sulphide ore zone, whose principal fea-

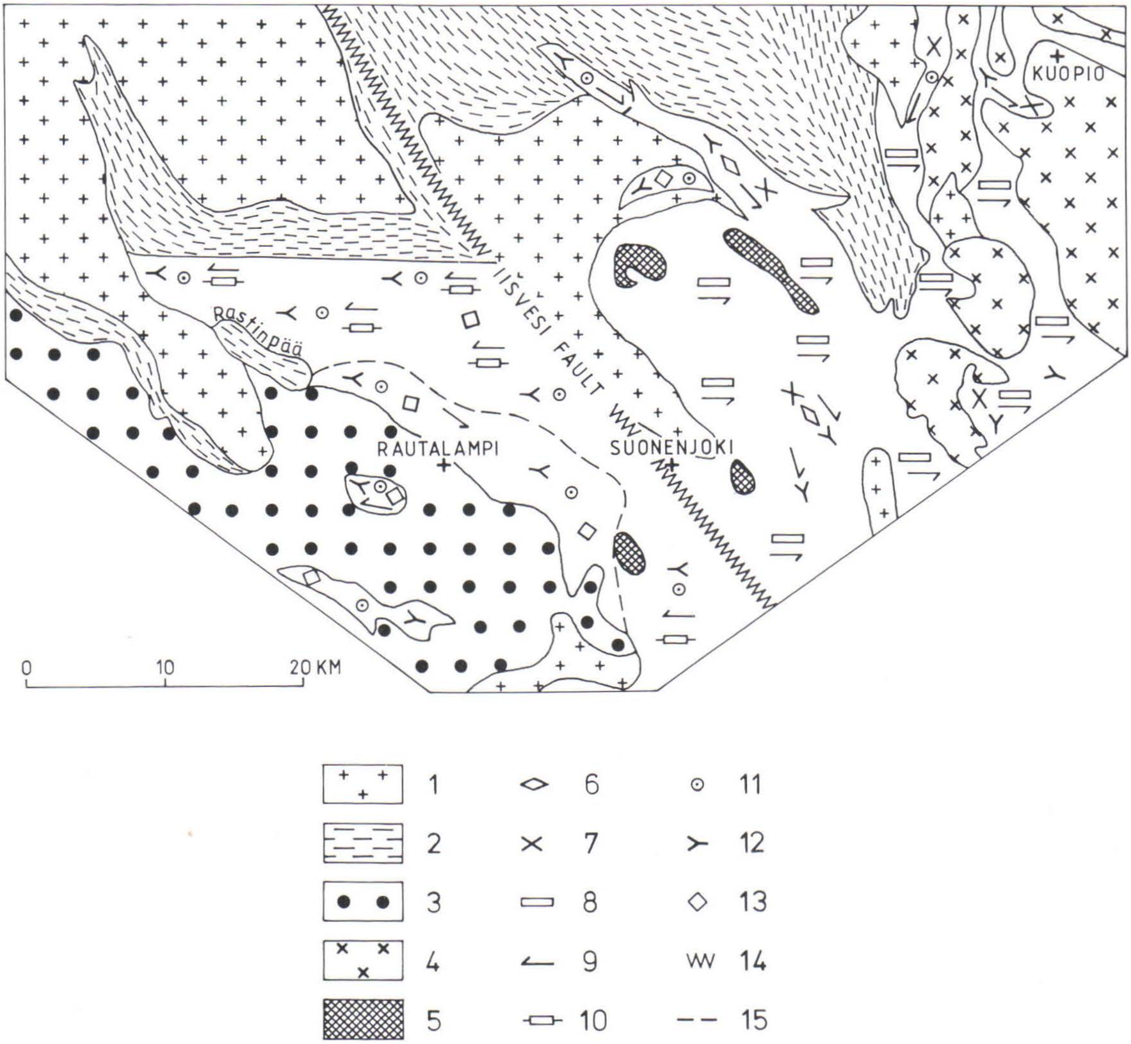

Fig. 11. Metamorphic map of the Kuopio — Rautalampi area. 1. granitoid, 2. gneissic tonalite, 3. hypersthene granitoid, 4. Archean granite gneiss, 5. gabbro and diorite, 6. andalusite, 7. staurolite, 8. muscovite, 9. sillimanite, 10. biotite, 11. garnet, 12. cordierite, 13. hypersthene, 14. fault, 15. isograd.

tures have been described in numbers of papers (e.g. Neuvonen *et al.* 1980); worth mentioning here are the gravimetric trench, the change in the isotope composition of sulphide lead and the ore potential of the area. The study areas are characterised by poorly developed, extensive metamorphic zoning, and have therefore been subdivided into metamorphic blocks.

The Kuopio and Rautalampi blocks

The metamorphic blocks of Kuopio in the east and Rautalampi in the west can be distinguished in the Kuopio-Rautalampi area. The metamorphic map of the area is depicted in Figure 11. Quartzites, volcanites and metapelites are encountered on the Archean granite gneiss basement in the Kuopio dome area in the east. The mineral assemblage typical of pelites is qtz-olig-sill-bio-stau-cord. Staurolite has only been met with as relics considerably smaller in grain size than cordierite. Pseudomorphs after staurolite filled with sillimanite have been encountered in the Kuopio area (Aumo 1983) and the abundance of staurolite declines westwards. Hence the metamorphic grade appears to increase slightly towards the west. Muscovite-bearing greywackes occur right at the western margin of the Kuopio block. The greywackes do not contain staurolite, not even as relics, but they do exhibit muscovite porphyroblasts that might be pseudomorphs after staurolite. The phase pair staurolite-muscovite have not therefore been encountered in equilibrium in the Kuopio block. The decomposition of staurolite started in the muscovite-bearing samples with the reaction

1) staurolite + muscovite + quartz = sillimanite + biotite.

The reaction did not continue to the end, however, because there was insufficient muscovite. And so staurolite continued to decompose with the reaction

4) staurolite + quartz = cordierite + sillimanite.

There are no observations from the Kuopio block of the K-feldspar-sillimanite isograds. Crystallisation therefore appears to have taken place in the Kuopio block at a slightly higher temperature than that at which the decomposition reaction of staurolite begins (reaction 1). Staurolite appears to have faded out in the equilibrium field of sillimanite, that is, above the equilibrium boundary of andalusite-sillimanite. Accordingly, the geothermal gradient of the Kuopio block might have been lower than in corresponding areas in the metapelite zone of southeastern Finland (Fig. 5).

Andalusite occurs commonly in the Kuopio block but as a very fine-grained variety. It crystallised during the retrograde metamorphic phase when cordierite was intensely biotitised. Myrmekitic intergrowths of quartz and andalusite can be seen at the margins of the biotitised cordierite grains (Fig. 12). Small staurolite grains are also visible at the margins of altered cordierite grains.

The porphyric microcline granite of Suonenjoki is located in the western part of the Kuopio block. The granite is a chlorite-bearing and postmetamorphic rock that has metamorphic windows with the mineral assemblage qtz-orthocl-hyp-gar-cord as inclusions (Fig. 11).

From the Kuopio block towards the Rautalampi block there is a jump in the metamorphic grade from the stability field of muscovite to that of hypersthene accompanied by equilibration of the pair garnet-cordierite. The Kuopio and Rautalampi blocks are in fact separated from each other by the mylonitised Iisvesi fault, which is several kilometres wide and trends 320°. Nevertheless the hypersthene zone proper of the Rautalampi block is closely associated with the hypersthene-bearing intrusions. Hypersthene-bearing gneisses are encountered only

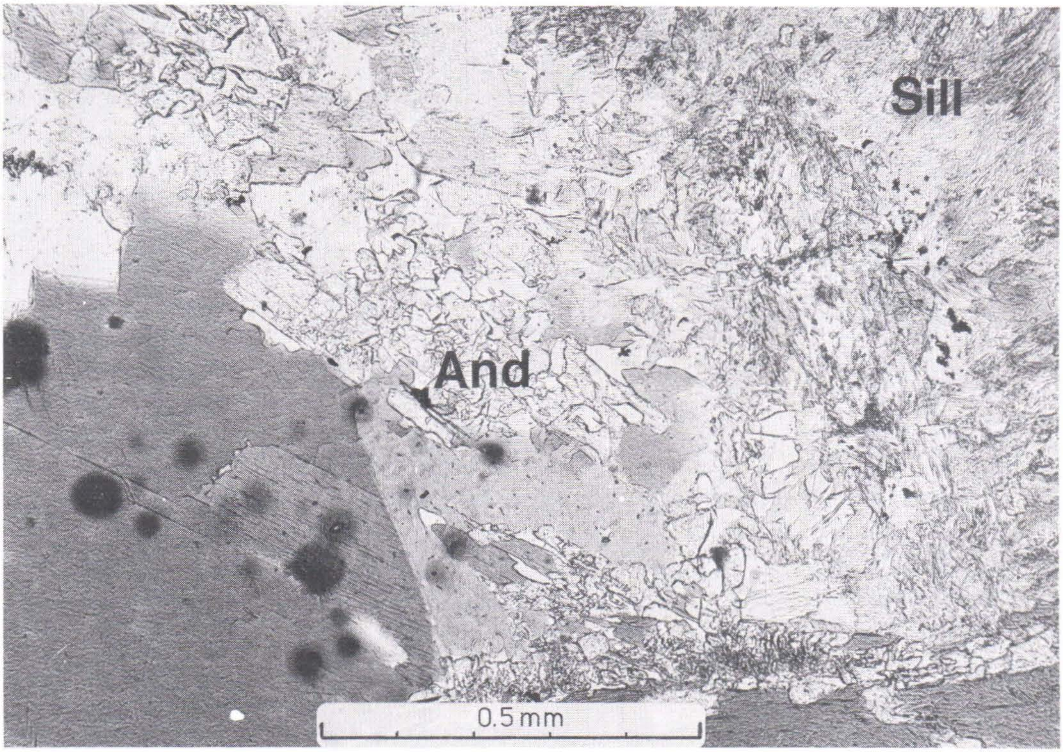

Fig. 12. Cordierite altered into andalusite and biotite.

sporadically in the eastern part of the block.

Mineral assemblages typical of the gneisses in the hypersthene zone are qtz-hyp-gar-cord and qtz-ant-cord-gar. In the northern part of the block in particular retrograde metamorphism has completely obliterated high-grade mineral assemblages. The compositions of coexisting garnet and cordierite (Tables 1 and 2) indicate that the pressure of crystallisation was much higher in the Rautalampi block than in the area of the Sulkava granulite facies but that the temperature of crystallisation was surprisingly low in the Rautalampi block. According to the experimental studies of Holdoway and Lee (1977), a low temperature of crystallisation indicates low partial water pressure ($0.2P_{tot}$). The breakdown reaction (reaction 3) of biotite certainly began in the Rautalampi block, but it did not get far past the beginning, because biotite is present in great abundance in the pelitic rock and it is in contact with sillimanite. The crystallisation conditions of the Kuopio and Rautalampi blocks are presented in Figure 5.

Metamorphism under dry conditions is also suggested by the occurrence of orthoclase in the Rautalampi hypersthene zone.

The metamorphic blocks of the Pihtipudas Pielavesi area

The following metamorphic blocks can be distinguished in the Pihtipudas-Pielavesi area from east to west: Lampaanjärvi, Vaaraslahti, Pielavesi, Korppinen and Pihtipudas (Fig. 13).

The Lampaanjärvi block is located, as is the Kuopio block, in the marginal zone of the Archean granite gneiss area. The two blocks are also metamorphically alike. Porphyroblasts after staurolite filled with muscovite and sillimanite are encountered at Lampaanjärvi, and the staurolite has continued to fade out with the crystallisation of cordierite. The cordierite is biotitised. A myrmekitic intergrowth of andalusite and quartz is associated with the biotitisation. Typical of the Lampaanjärvi block are the numerous shears trending north-south and the related epidotisation.

Lithologically, the Vaaraslahti block is similar to the Lampaanjärvi block. A fault has not been observed between the blocks but the difference in metamorphic grade between them is sharp and pronounced. The contact aureole of the hypersthene granite in the Vaaraslahti block exhibits the equilibrated assemblage qtz-ksp-sill-bio-gar-cord. Hypersthene is also in equilibrium in the Vaaraslahti block. Schistosity disappears from the rock adjacent to the contact. Hence the contact metamorphism of the Vaaraslahti granite overprints the deformation and metamorphism of the Lampaanjärvi block. The Vaaraslahti block is characterised by biotitisation of cordierite and the myrmekitic intergrowth of andalusite and quartz. The breakdown reaction of biotite has not advanced far in the Vaaraslahti block. The presence of orthoclase in the block refers to dry metamorphism. The compositions of coexisting garnet and cordierite indicate that metamorphism took place at a pressure of 4.5 kb ($P_{H_2O} = 0.2P_{tot}$). The crystallisation temperature on the other hand drops with increasing distance from the contact with granite; sample 216PSH78 is from nearest the contact and sample 108PHS78 from farthest away (Table 2).

Since the Vaaraslahti and Lampaanjärvi blocks are metamorphically alike except for the contact effect of granite, staurolite should disappear from the Lampaanjärvi block at a pressure of 4.5 kb (Fig. 5). This is an agreement with the observation made in the Kuopio block that the geothermal gradient in the marginal zone of the Archean granite gneiss area was gentler than in the Rantasalmi area.

There is a sharp change in lithology between the Vaaraslahti and Pielavesi blocks. Rocks typical of the Pielavesi block are norites, hypersthene granitoids and hypersthene or sillimanite-bearing garnet-cordierite gneisses. Metapelites proper have not been encountered in the block. The compositions of the coexisting garnet and cordierite are of the same order of magnitude as in the Rautalampi block (Tables 1 and 2), in other words, the blocks crystallised under similar conditions. The compositions of garnet and cordierite differ substantially from those of the Vaaraslahti garnet and cordierite. The crystallisation conditions of the Pielavesi block indicate that it represents a deeper erosional level than the Vaaraslahti block (Fig. 5). Contact aureoles of hypersthene granitoids as distinct as those at Vaaraslahti have not been encountered in the Pielavesi block: another indication that the hypersthene granitoids were emplaced at different depths in the blocks. Some outcrops at the boundary of the blocks suggest that there is a fracture zone between the blocks.

The boundary between the Pielavesi and Korppinen blocks is not straight but a tortuous and partly mylonitised seam. The change in metamorphic grade between the blocks is sharp because the equilibrium field of hypersthene passes straight into that of muscovite. Nevertheless the Korppinen block is characterised by intense retrograde metamorphism associated with sericitisation, cloritisation and epidotisation. The relics in the eastern part of the block indicate that the crystallisation conditions preceding the retrograde phase were close to those of the K-feldspar-sillimanite isograd. Garnet, cordierite, sillimanite and alkali feldspar are in equilibrium right at the eastern margin of the block. The assemblage is encountered only sporadically and intensely muscovitised. The above indicates that the Karppinen block formerly represented, at least partly, a higher metamorphic grade than

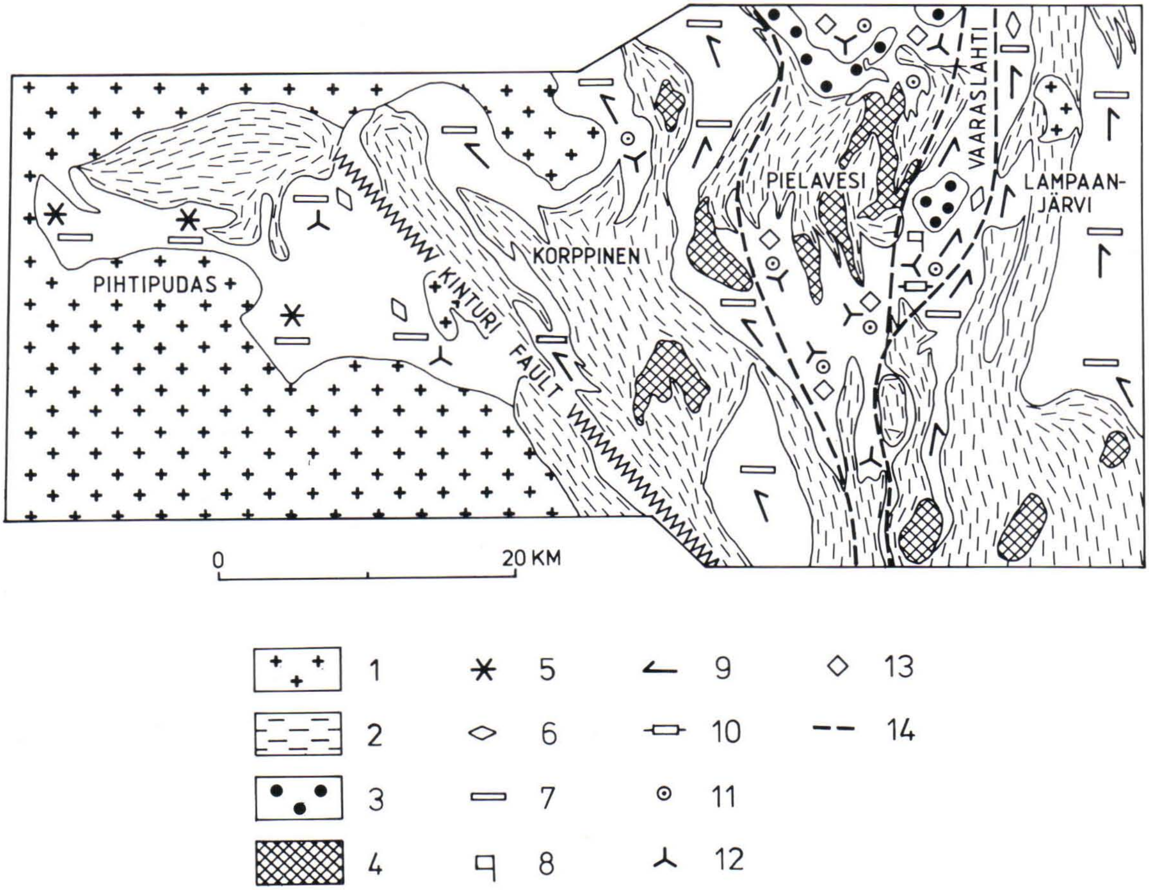

Fig. 13. Metamorphic map of Pihtipudas — Pielavesi area. 1. granitoid, 2. gneissic tonalite, 3. hypersthene granitoid, 4. gabbro, 5. epidote, 6. andalusite, 7. muscovite, 8. K-feldspar, 9. sillimanite, 10. biotite, 11. garnet, 12. cordierite, 13. hypersthene, 14. isograd.

can be inferred nowadays. Lithologically the Korppinen block differs substantially from the Pielavesi block. Hypersthene-bearing intrusions have not been encountered in the Korppinen block, and there are only random occurrences of anthophyllite-cordierite rock. A considerable proportion of the supracrustal rock consists of biotite-plagioclase gneisses, often with tourmaline. Hornblende and epidote-bearing gneissic quartz diorites predominate in the area.

The Korppinen and Pihtipudas blocks are separated from each other by the mylonitized Kinturi fault trending 320°. The assemblage qtz-ab-olig-ep-hbl is encountered in the metavolcanites in the western part of the Pihtipudas block and that of qtz-olig-musc-bio-tourm in the metapelites. Albite fades out eastwards and cordierite appears in the metapelites in the equilibrium field of andalusite. The metamorphic grade in the Pihtipudas block increases from west to east from the epidote amphibolite facies to the lower amphibolite facies. Although the progressive metamorphism of the Korppinen block is overprinted by a strong retrograde phase, once the Kintturi fault has been crossed there is a marked change in the metamorphic grade. After all, it means transition from the andalusite equilibrium field and the first isograd of cordierite direct to the equilibrium field of sillimanite and to above the equilibrium field of muscovite. In Figure 5 the geothermal gradient of Pihtipudas is inferred to be 55°C/km, a value corroborated by observations on the occurrence of andalusite and staurolite outside the study area west of Pihtipudas.

Metamorphic evolution

The Savo schist belt is characterised by the metamorphic blocks. Internally the blocks exhibit uniform metamorphic features, but between the blocks the transition from one metamorphic grade to another is stepwise rather than continuous. Except for parts of the Vaaraslahti and Lampaanjärvi blocks, even adjacent blocks display completely different metamorphic features.

The Kuopio and Lampaanjärvi blocks, which are located at the margin of the Archean granite gneiss area, are metamorphically similar. Metamorphism took place in the equilibrium field of muscovite but above the stability field of staurolite in the muscovite-bearing samples. The Vaaraslahti and Lampaanjärvi blocks represent more or less the same erosional level. Observations in Vaaraslahti indicate that the crystallisation pressure was 4.5 kb. In the marginal zone of the Archean granite gneiss area staurolite fades out above the andalusite-sillimanite equilibrium boundary. The above indicates that the geothermal gradient of the marginal zone was 44°C/km. This value is lower than that for the Rantasalmi area, but it is about the same as that for the Tohmajärvi-Kitee area. At Vaaraslahti the geothermal gradient has risen as a result of magmatic activity. The biotitisation of cordierite and the associated crystallisation of andalusite are characteristic features of the blocks at the margin of the Archean granite gneiss areas.

The conditions of granulite facies were attained in the Rautalampi, Pielavesi and Vaaraslahti blocks. Metamorphism of the granulite facies is related in time and space to the proximity of hypersthene-bearing granitoid intrusions. The Pb-U age of zircon from the hypersthene granite of Vaaraslahti and Rautalampi is 1880 Ma (Salli 1983, Fig. 14, Table 7). The metamorphism in the blocks of the Savo schist belt that were metamorphosed under the conditions of granulite facies thus preceded the progressive metamorphism in the metapelite zone of southeastern Finland and the granulite facies metamorphism in Sulkava. On the other hand, the contact metamorphism in the Vaaraslahti hypersthene granite overprints the older metamorphism. Hence the metamorphism associated with hypersthene granites was preceded by an older metamorphic episode (cf. p. 16). The zircon and monazite in the Vaaraslahti granite are of roughly the same age (Salli op.cit.). This indicates that the areas were equilibrated relatively early owing to intense metamorphism, because monazite often manifests specifically the age of metamorphism (Kouvo 1982).

The crystallisation temperature determined from the pair garnet-cordierite is relatively low (660°—700°C) for metamorphism of the granulite facies. Indeed, the principal effect of the metamorphism associated with hypersthene granitoids was to dry the system. The common occurrence of orthoclase in areas of the Savo schist belt that were metamorphosed under granulite facies conditions points to metamorphism under somewhat similar dry conditions as described by Vorma (1972) from contact aureoles of Wiborg rapakivi. The magnesium contents in the Pielavesi and Rautalampi garnets are substantially higher than those in the Vaaraslahti samples. The differences in the compositions of the phases appear to be mainly due to variations in the emplacement depth of the hypersthene granites. There are no contact aureoles like those at Vaaraslahti around the Rautalampi and Pielavesi hypersthene granites. Crystallisation of andalusite related to biotitisation of cordierite has not been detected in the samples from Pielavesi and Rautalampi, indicating that at the crystallisation of andalusite the blocks were not yet in their present position in relation to each other.

The Pielavesi block is bounded in the west by the lithologically different Korppinen block. The progressive episode of metamorphism in the Korppinen block has extended to above the equilibrium field of muscovite but the earlier metamorphism is overprinted by an intense retrograde phase. Hypersthene granitoids have not been encountered in the block; its characteristic granitoids are gneissic hornblende-bearing tonalites with gradual contacts with amphibolites. The evolution of the Korppinen block is problematic. One question that has to be settled is whether the progressive metamorphism in the block preceded the metamorphism due to hypersthene granitoids. The granulite facies blocks of Rautalampi and Pielavesi are bounded by the very gneissic tonalite of Rastinpää and Kirkkosaari, and at Rautalampi in particular the Rastinpää tonalite (Fig. 11) envelops the hypersthene granite (age 1880 Ma). Observations in the field indicate that the gneissic tonalite is older than the hypersthene granitoids, a concept that is corroborated by U-Pb datings of zircon in tonalite. Age determinations at Rastinpää give an age of 1920 Ma (Fig. 15, Table 8). The Rastinpää gneiss can be compared with the Kettuperä gneiss north of Pielavesi with a Pb-U age of 1930 Ma (Helovuori 1979). The Rastinpää gneiss is intensely migmatised, indicating a high metamorphic grade; however, it underwent retrograde metamorphism during the emplacement of hypersthene granitoids. The complex evolution of the Rastinpää gneiss makes it difficult to

Table 7. U-Pb analytical data on zircons from Haukilampi hypersthene granite (A218).

Zircon fraction	Concentration		Atom ratios						Radiometric ages		
d = density Ø = size in µm HF = preleached in HF	ug/g		blank corrected			corrected for blank and common lead			Ma		
	238_{U}	206_{Pb} radiogenic	$\frac{206_{Pb}}{204_{Pb}}$	$\frac{207_{Pb}}{206_{Pb}}$	$\frac{208_{Pb}}{206_{Pb}}$	$\frac{206_{Pb}}{238_{U}}$	$\frac{207_{Pb}}{235_{U}}$	$\frac{207_{Pb}}{206_{Pb}}$	$\frac{206_{Pb}}{238_{U}}$	$\frac{207_{Pb}}{235_{U}}$	$\frac{207_{Pb}}{206_{Pb}}$
A218, porphyritic hypersthene granite; Haukilampi, Suonenjoki											
A d > 4.6; Ø > 70	302.4	86.05	27282	.11554	.09253	.3290 ± 17	5.218 ± 28	.11505 ± 7	1833 ± 9	1855 ± 5	1880 ± 3
B d > 4.6; Ø > 70 HF	284.7	83.03	53234	.11550	.09567	.3370 ± 19	5.356 ± 31	.11525 ± 8	1872 ± 9	1877 ± 5	1883 ± 3
C 4.2 < d < 4.6 Ø > 70; HF	514.0	148.20	67954	.11573	.08047	.3332 ± 22	5.308 ± 36	.11553 ± 5	1853 ± 10	1870 ± 6	1888 ± 3

Table 8. U-Pb analytical data on zircons from Rastinpää gneissic tonalite (A217).

Zircon fraction	Concentration		Atom ratios						Radiometric ages		
d = density Ø = size in μm HF = preleached in HF	ug/g		blank corrected			corrected for blank and common lead			Ma		
	^{238}U	^{206}Pb radiogenic	^{206}Pb/^{204}Pb	^{207}Pb/^{206}Pb	^{208}Pb/^{206}Pb	^{206}Pb/^{238}U	^{207}Pb/^{235}U	^{207}Pb/^{206}Pb	^{206}Pb/^{238}U	^{207}Pb/^{235}U	^{207}Pb/^{206}Pb
A217, gneissic tonalite; Rastinpää, Rautalampi											
A d > 4.6; Ø > 70	357.9	94.04	32791	.11687	.10749	.3037	4.876	.11646	1709 ± 8	1798 ± 5	1902 ± 2
						± 17	± 27	± 5			
B d > 4.6; Ø < 70	342.3	90.73	23481	.11663	.11621	.3063	4.902	.11605	1722 ± 8	1802 ± 4	1896 ± 3
						± 16	± 26	± 6			
C 4.2 < d < 4.6;	526.6	146.26	34153	.11696	.11110	.3210	5.158	.11656	1794 ± 8	1845 ± 4	1904 ± 3
Ø > 160						± 16	± 27	± 9			
D 4.2 < d < 4.6;	547.6	134.51	15457	.11775	.10095	.2839	4.575	.11687	1610 ± 7	1744 ± 4	1909 ± 5
Ø < 160						± 15	± 25	± 18			
E 4.0 < d < 4.2;	776.2	210.25	29531	.11670	.10004	.3130	5.017	.11625	1755 ± 8	1822 ± 5	1899 ± 3
Ø < 160						± 16	± 26	± 5			
F 3.8 < d < 4.0	1204	329.53	29927	.11609	.09996	.3164	5.044	.11563	1771 ± 8	1826 ± 4	1889 ± 3
Ø < 160						± 16	± 27	± 8			
G 4.2 < d < 4.6	382.9	107.60	29695	.11800	.14392	.3248	5.264	.11755	1813 ± 9	1863 ± 5	1919 ± 3
Ø > 160; HF						± 17	± 29	± 8			
H 4.2 < d < 4.6	468.5	134.92	31275	.11737	.11720	.3329	5.366	.11694	1852 ± 9	1879 ± 5	1910 ± 3
Ø < 160; HF						± 17	± 29	± 5			
I 4.0 < d < 4.2	558.0	153.57	27430	.11778	.14032	.3181	5.144	.11729	1780 ± 8	1843 ± 5	1915 ± 3
Ø > 160; HF						± 16	± 27	± 5			
J 3.8 < d < 4.0	793.2	216.82	43709	.11719	.12799	.3159	5.091	.11688	1769 ± 8	1834 ± 4	1909 ± 3
Ø < 160; HF						± 16	± 27	± 6			

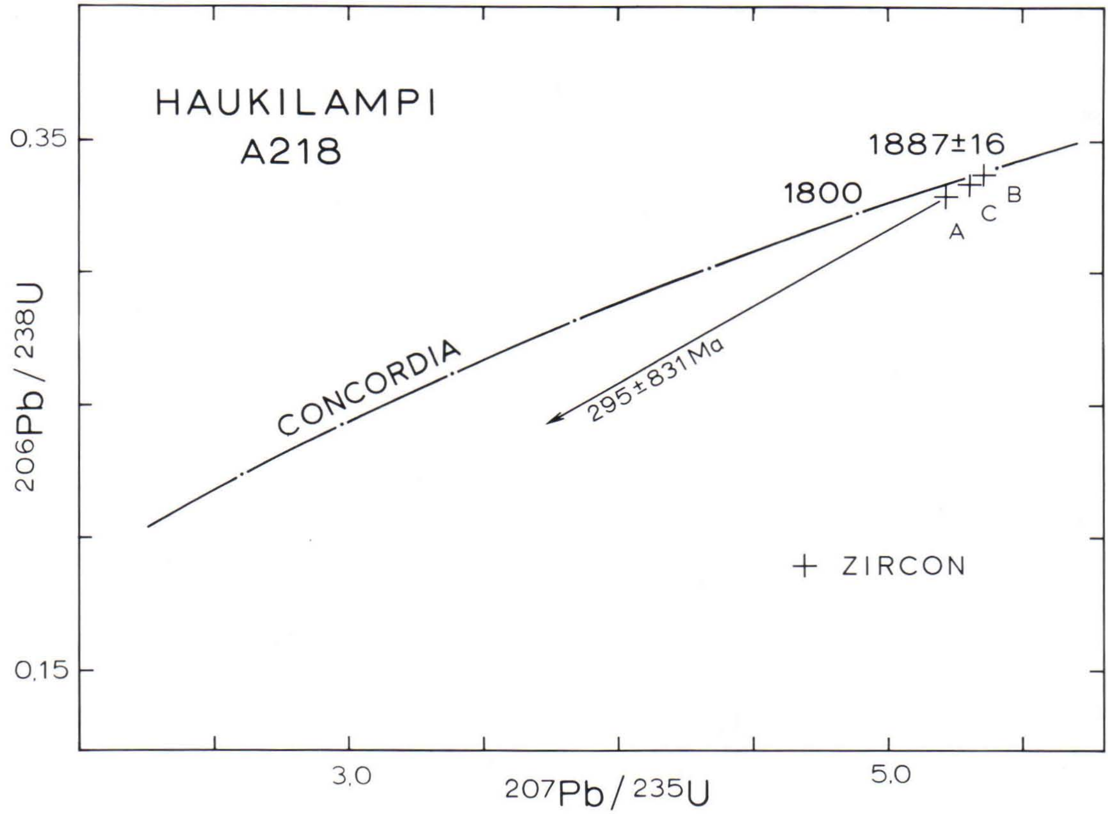

Fig. 14. Concordia plot for zircons from the Haukilampi hypersthene granite.

establish what episode the age refers to, but it implies early metamorphism antedating the granulite facies metamorphism at 1900 Ma.

The metamorphic grade in the Pihtipudas block increases steadily eastwards as far as Kinturi fault. The nature of the metamorphism is similar to that of the zoned metamorphism in southeastern Finland. The 1880 Ma radiometric age of zircon in the tonalite of Pihtipudas (Aho 1979) is considerably older than the age of titanite, 1800 Ma, in the same rock (Aho op.cit.). This, too, reflects the difference in the nature of metamorphism between the Pihtipudas and Pielavesi blocks.

As stated above, the granulite facies metamorphism of the Savo schist belt is related to hypersthene granitoids; no other granitoids had any impact on the increase in the grade of metamorphism, although some metamorphic windows have been encountered in the porphyric granite of Suonenjoki. The porphyric microcline granite contains muscovite and chlorite whereas the windows have the assemblage qtz-orthocl-gar-cord-hyp. The porphyric granite probably transported the windows from depth to their present location.

The metamorphic blocks of the Savo schist belt are separated from each other by faults. The Kuopio and Rautalampi blocks are separated by the Iisvesi fault. The boundary of the blocks is marked by a gravimetric trench and the contact of the Suonenjoki porphyric granite. The boundary between the Pihtipudas and Korppinen blocks is similar to the former; these

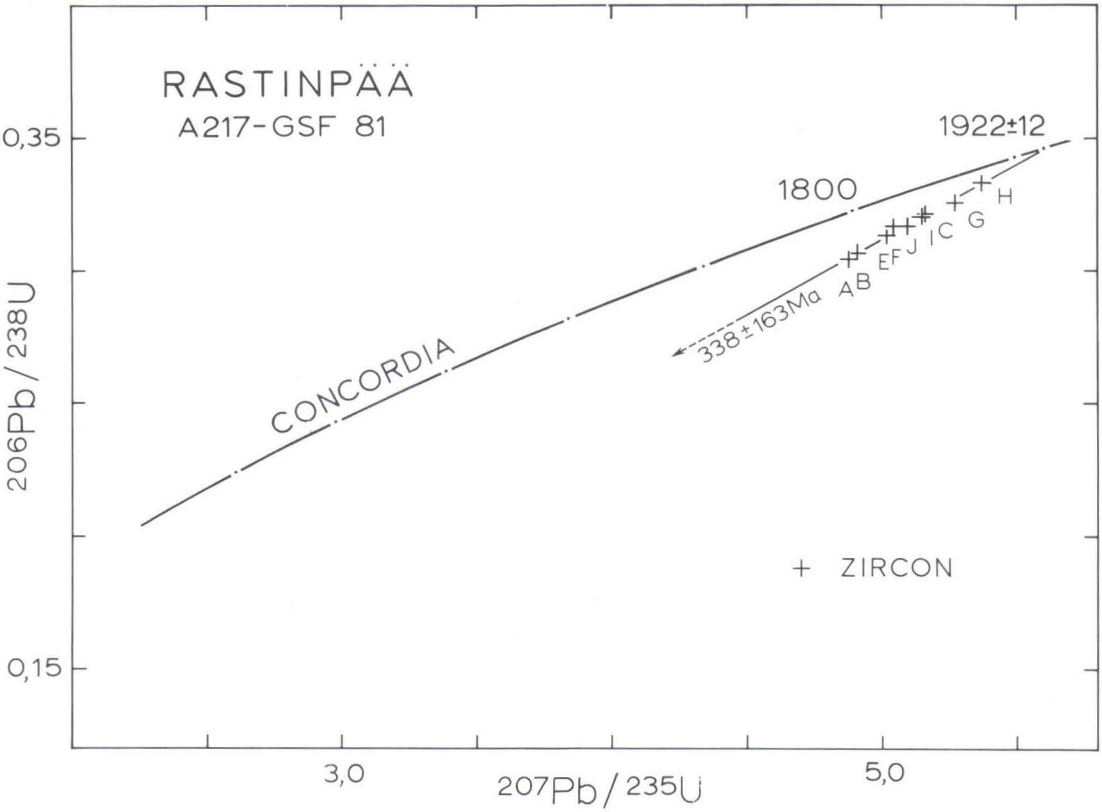

Fig. 15. Concordia plot for zircons from the Rastinpää gneissic tonalite.

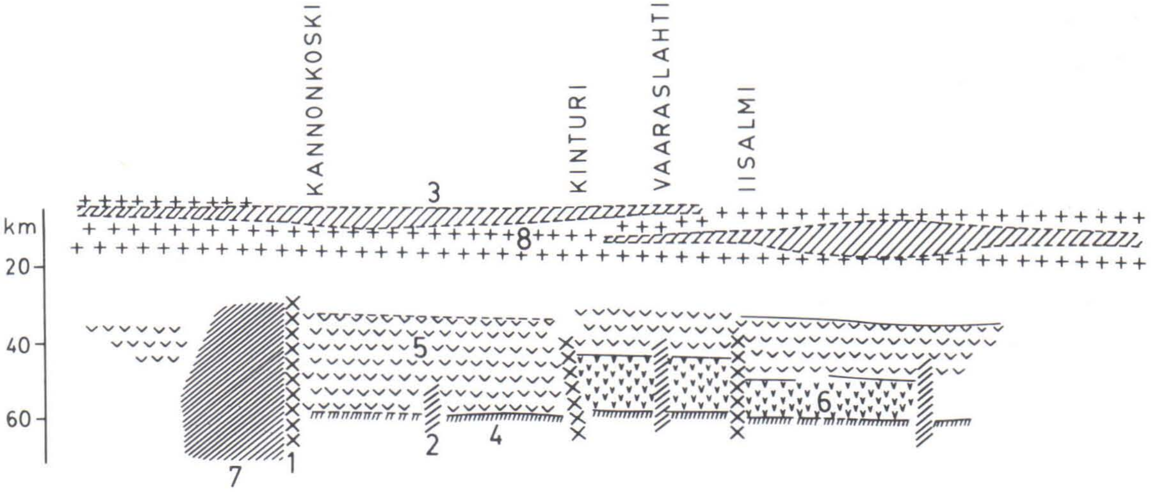

Fig. 16. Interpretation model of seismic SVEKA profile according to Luosto et al. (1982) (cf. Fig. 13). 1. deep-seated fracture, 2. tectonic anomaly in lower crust, 3. low-velocity zone, 4. Moho discordancy surface, 5. lower part of crust, 6. high-velocity zone, 7. shade area, 8. upper part of crust.

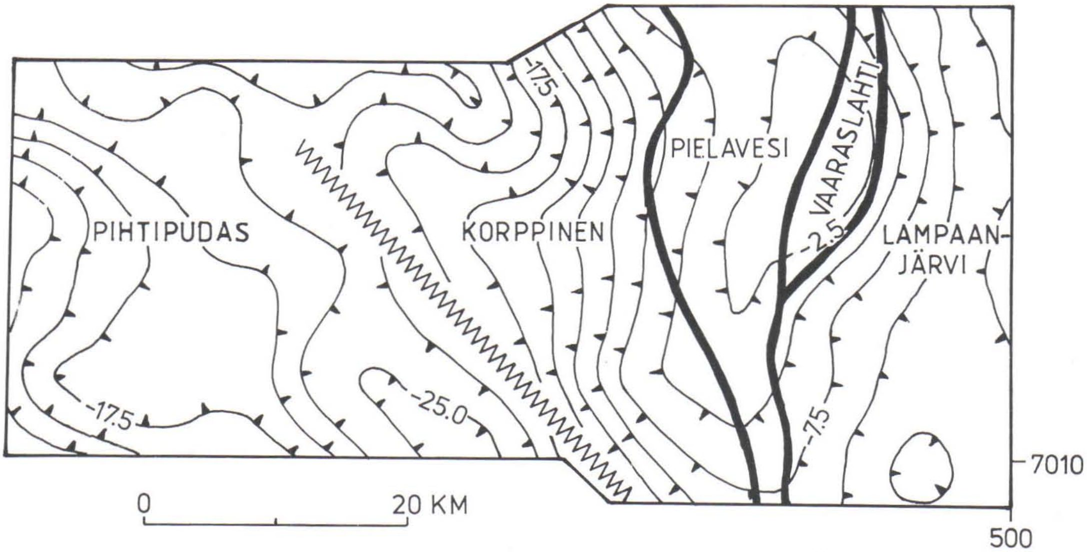

Fig. 17. The metamorphic blocks of the Pihtipudas — Pielavesi area and a Bouguer anomaly map (IGSN17) compiled at the Geophysics Department of the Geological Survey of Finland according to data measured by the Finnish Goedetic Institute.

two are separated by the Kinturi fault located near by a gravimetric trench (Fig. 17). The Kinturi fault also cuts the Pielavesi block in the south. The Pielavesi and Vaaraslahti blocks are separated by a fault trending mainly N-S. The boundary between the Korppinen and Pielavesi blocks is neither tectonically nor metamorphically well developed, but is occupied by a meandering, mylonitised seam.

The block-like character of the Savo schist belt is reflected not only in metamorphic features but in other geological features as well. The isotope composition of the Pielavesi and Rautalampi sulphide lead differs from that of the Pihtipudas sulphide lead in that the former contains more mantle lead than the latter (M. Vaasjoki 1981). The boundary between the Pihtipudas and Korppinen blocks (Fig. 16) is clearly revealed by deep seismic sounding (Luosto *et al.* 1982). Seismic survey also indicates the presence of a tectonic anomaly between the Pielavesi and Vaaralahti blocks. The block of the Pielavesi granulite facies overlaps the gravity maximum almost completely (Fig. 17). According to Elo (1983), the positive Bouguer anomaly cannot be explained by the mean density of the surficial bedrock alone.

The metamorphic blocks and deep-seated fractures of the Savo schist belt might well indicate metamorphism of a rift zone formed at the margin of the Archean granite gneiss area. Magmatism, which obviously originated in the mantle, is responsible in time and place for the granulite facies metamorphism of the rift zone.

DISCUSSION

The metamorphism of the Savo schist belt differs unmistakeably from the zoned metamorphism of southeastern Finland. The differences in the nature of the metamorphism also indicate a difference in the structure and evolution of the crust. Metamorphism has imprinted the crust with specific structural features; therefore, not only the establishing of the geothermal gradient but also the study of the isograd surfaces and the continuity of metamorphic zoning in general are crucial when delineating the evolution of the crust with the aid of metamorphic investigations. The Savo schist belt is characterised by a progressive phase of metamorphism that evolved early and is one reason why the impact of faults on the position of isograd surfaces is so strong in the belt. One phase of metamorphism preceded the emplacement of the intrusives at 1880 Ma; the metamorphism of the granulite facies, however, is related to the hypersthene granites dated at 1880 Ma. The marginal zone of the Archean granite gneiss area is characterised by uniform metamorphism that took place mainly in the equilibrium field of muscovite. Biotitisation of cordierite with associated myrmekitic crystallisation of andalusite are typical features of the metamorphism in the marginal zone. The Savo schist belt with its deep-seated fractures and metamorphic blocks might well indicate a rift zone in which deep-seated magmatism has led in both time and place to metamorphism of the granulite facies. The marginal zone of the Archean granite gneiss area appears to represent a somewhat shallower level of the crust than do the granulite facies blocks of Pielavesi and Rautalampi. Thus, the pelitic rocks might overlie the volcanites of Rautalampi and Pielavesi in stratigraphy. The metamorphism of the Pihtipudas area differs from that of the rift zone and resembles the metamorphism of the metapelite zone of southeastern Finland. Hence, the rift zone, which also includes the marginal blocks of the Archean gneiss area, is located at the boundary between the Pihtipudas and Archean gneiss blocks.

The picture given by metamorphic studies of the evolution of the Savo schist belt resembles in broad features the model that Marttila (1976) formulated for the geological evolution of the Kiuruvesi area.

The zoned metamorphism of the metapelites of southeastern Finland evolved after the emplacement episode at 1880 Ma, and is therefore younger than that of the Savo rift zone. Some of the ages obtained indicate that a very high temperature was reached in the area of the Sulkava thermal dome between 1850 and 1810 Ma. The zoned metamorphism terminated before the emplacement of the garnitoids of the the 1800 Ma age group. The zoned metamorphism is characteristed by the high geothermal gradient, which is 55°C/km in the least metamorphosed zones. The corresponding value for the Kuopio and Lampaanjärvi blocks and the Tohmajärvi-Kitee area of southeastern Finland is lower, 45°/km. The difference in metamorphism between the rift zone and the metapelite zone is manifested clearly in the areas of granulite facies. A temperature of 750°C was reached at Sulkava at a pressure of 4 kb, whereas at Pielavesi and Rautalampi a temperature of 660°—700°C corresponds to a pressure of 5.1—5.5 kb. In the Savo schist belt, in particular, faults trending 320° have affected the geometry of the metamorphic zoning.

The zoned metamorphism of southeastern Finland resembles the model worked out by England and Richardson (1977) for the metamorphism of tectonically thickened crust. Although the Sulkava area underwent granulite facies metamorphism very late, it was not possible within the scope of the present study to determine the age relationship between the metamorphism of the K-feldspar-sillimanite zone and the Juva zone. In terms of macrocstructure, the metamorphism of the K-feldspar-sillimanite zone is mainly post-tectonic and that of the Juva zone syntectonic; but then the more strongly developed structures of the Juva zone are younger than the corresponding structures in the K-feldspar-sillimanite zone. In other words, the area of the Sulkava thermal dome might have acted as some kind of thermal centre in relation to its environment.

Although the reduction in the rate of erosion played a part in furthering the evolution of the Sulkava thermal dome, the basic cause of the zoned metamorphism still evades us. As England, discussing the evolution of the thermal Tauern window in the Alps, points out (England 1978, p. 37): »It is possible that there is a feature common to the mechanical development of many continent-continent collision with causes depression of the overthrust region for some time after movement, ceases, and the removal of this feature would then initiate the erosion. Such a feature could be the negative buoyancy of oceanic lithosphere attached to be subducted basement which at some stage would become detached from it and sink.»

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The manuscript was read critically by A. Vorma. The manuscript was translated into English by Mrs Gillian Häkli.

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