Evolution of a metamorphic fluid during progressive metamorphism in the Joroinen–Sulkava area, southeastern Finland, as indicated by fluid inclusions

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Abstract

Fluid inclusions in the progressively metamorphosed rocks of the Joroinen–Sulkava area, located in the south-eastern end of the Raahe–Ladoga zone near the Archaean–Proterozoic boundary, south-eastern Finland, fall into four main categories: (1) H_2O-rich, (2) CO_2-rich, (3) mixed H_2O–CO_2 and (4) CH_4–N_2 inclusions. The samples were collected from quartz veins associated with different deformation phases (D_2–D_4) and from metapelites. The progressive stage of metamorphism took place mainly during the D_2 deformation. The age of metamorphism and D_2 deformation becomes younger with increase in metamorphic grade from amphibolite to granulite facies.

Regional distribution of the different fluid types indicates a change in fluid regime from H_2O to CO_2-dominant during the progressive stage of the metamorphism. H_2O entered preferentially into the anatectic melt. The possibility of CO_2 infiltration from deeper crust can not be excluded, because granulite facies rocks occur most probably below the lower grade zones. A zone enriched in CH_4–N_2 fluids is located near the lineament zones caused by the D_3 deformation. This fluid type dominates the Au-bearing D_2–D_3 quartz lenses in the K-feldspar–sillimanite zone. Density data of early CO_2 inclusions in combination with estimates of metamorphic temperatures (645–750 °C) in the different metamorphic zones indicate a pressure range of 3.0–4.5 kbar, which is consistent with data derived from mineral geobarometry. The diversity of fluid types encountered in the D_2–D_4 quartz veins are a result of the passage of different fluids through veins at different times without re-equilibrating with the wall rocks. However, it is supposed that the CH_4–N_2 fluid is derived from a CO_2-rich fluid with $X_{CH_4} \leq 0.4$ by re-equilibration during its passage through the rocks.

KEYWORDS: metamorphic fluid, fluid inclusions, progressive metamorphism, Finland.

Introduction

METAMORPHIC fluids consist mainly of molecular compounds of the system C–O–H–N. The distribution of these fluids in the continental crust and upper mantle was modelled by Touret and Dietvorst (1983). There is some dispute about the role of CO_2 in high-grade metamorphism (i.e. syn- or postmetamorphic fluids, their origin: see Lamb *et al.*, 1987; Hollister, 1988) and about the extent to which a free volatile-rich fluid phase is present during metamorphism (i.e. wet versus dry metamorphism: Fyfe *et al.*, 1978; Thompson, 1983). However, on one point there is a general agreement; veins in metamorphic terrains represent fractures along which fluids escape. Yardley (1986) has pointed out that quartz veins in pelitic schists develop when a period of fluid release coincides with a deformation event.

The Joroinen–Sulkava area in south-eastern Finland was selected for fluid inclusion study because of its unique geology and well-developed metamorphic zoning. A number of studies have been made by several authors on different aspects of local and regional geology. Progressive metamorphism in the area has been studied in detail by Korsman (1977), Korsman *et al.* (1984, 1988). The evolution of both metamorphism and deformation has been related to time by Korsman and Kilpeläinen (1986) and Kilpeläinen (1988). Several U–Pb zircon and monazite ages from the area have been reported by Korsman *et al.* (1984) and Vaasjoki and Sakko (1988). In addition,

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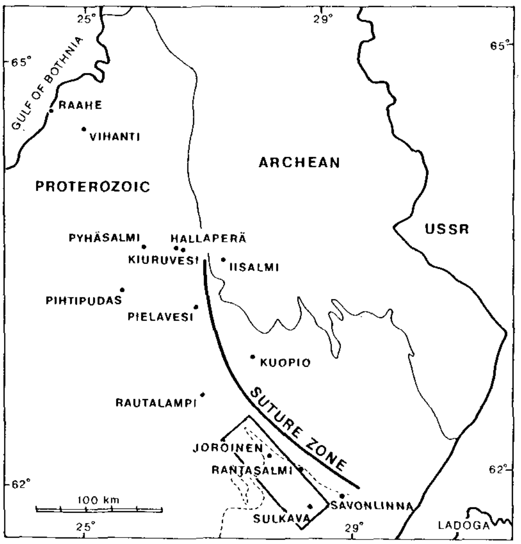

FIG. 1. Geographical location of the study area (rectangle). The progressively metamorphosed Joroinen–Sulkava area is outlined with a dashed line (modified from Korsman *et al.*, 1988).

Haudenschild (1988) has carried out K–Ar age determinations on biotite and muscovite and Korsman *et al.* (1984) on biotite and hornblende.

The aim of this paper is to characterize the metamorphic fluids associated with polyphase deformation of rocks, from amphibolite to granulite facies, in the Joroinen–Sulkava area. Data on compositional variations and regional distribution of different fluid types are given. Estimates of their relative abundances in different rock types are also presented. Finally, evolution of the metamorphic fluids, including interpretation of pressure–temperature conditions, is outlined.

Tectonometamorphic evolution

The Joroinen–Sulkava area is located at the south-eastern end of the Raahe–Ladoga zone near the Archaean–Proterozoic boundary (Fig. 1). The metamorphic grade increases from north to south as established by the observation of the metapelites (Korsman, 1977, Fig. 2). The area between Joroinen and Sulkava is tilted to the north by 1–3° (Korsman, 1977) and this is responsible for the well-developed zonality associated with progressive metamorphism. Sulphide and gold-bearing quartz veins and lenses occur in the northern part of the study area (Makkonen and Ekdahl, 1988).

The prograde stage of metamorphism was preceded by a very weak metamorphism associated with the D_1 deformation (Korsman *et al.*, 1988). D_1 deforms the primary structures (S_0) as isoclinal folds in the andalusite–muscovite zone (Kilpeläinen, 1988). Metamorphism took place below the equilibrium field of andalusite (Korsman *et al.*, 1988).

The D_2 deformation is displayed as tight F_2 folding with associated S_2 schistosity, whose intensity increases with increasing metamorphic grade. In the Sulkava thermal dome area, the schistosity is marked by an alignment of garnet, cordierite and quartz (Kilpeläinen, 1988). As he demonstrated, both metamorphism and D_2 deformation become younger with increasing metamorphic grade. Prograde metamorphism in the area took place mainly during the D_2 deformation.

The D_3 deformation is seen as asymmetric folds of varying size. Mineral growth does not accompany the deformation until south of the K-feldspar–sillimanite zone, where biotite has grown in the S_3 crenulation seams (Kilpeläinen, 1988). The metamorphic zones are overprinted by the D_3 deformation (Kilpeläinen, 1988; Korsman *et al.*, 1988) which is why Kilpeläinen (1988) concluded that the deformation took place 1.83–1.80 Ga ago. This is also the time span reported by Korsman *et al.* (1984) for the culmination of prograde metamorphism. Outside the Sulkava thermal dome, Nironen (1989) inferred that the D_3 deformation occurred after prograde metamorphism, because the D_3 appears to vanish towards the dome (see Fig. 2). The D_3 structures in the dome area may have been destroyed by uplift of the dome caused by melt-enhanced deformation during anatexis (Nironen, 1989).

The youngest D_4 deformation was not accompanied by mineral growth. The deformation is represented by semi-open folding and by crenulation of older structures. D_4 structures cut the 1.80 Ga-old granitoids (Kilpeläinen, 1988).

Crystallization conditions for the area are known from earlier studies (Korsman, 1977; Korsman *et al.*, 1984); near the K-feldspar–sillimanite isograd the metamorphic temperature estimate is 645 °C at $P = 3.4$ kbar. These values are based on the phase equilibrium fields of Al_2SiO_5 (Winkler, 1979) and on the equilibrium curve for muscovite (Kerrick, 1972). In the northern and southern part of the garnet–cordierite–sillimanite–biotite zone, temperature estimates are 690 °C at $P = 4.2$ kbar and 685 °C at $P = 4.8$ kbar respectively, and in the Sulkava thermal dome area, 750 °C at $P = 4.3$ kbar. These data are derived from garnet–cordierite thermobarometry (see Holdoway and Lee, 1977).

Progressive metamorphism in the area

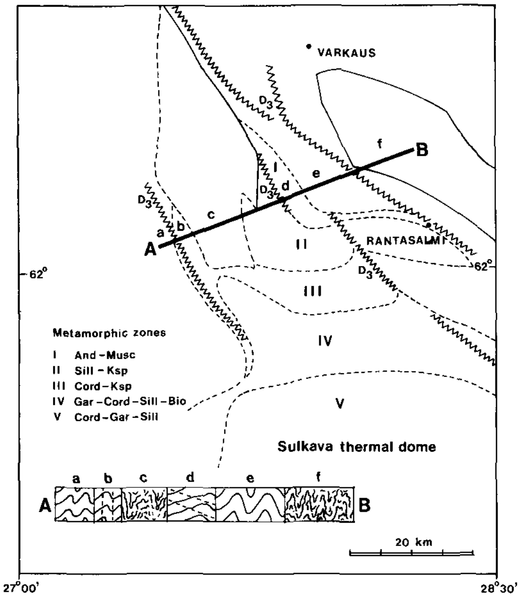

FIG. 2. Tectono-metamorphic map of the Joroinen–Sulkava area. The progressive stage of the metamorphism (D_2) in the area was associated with well-developed metamorphic zoning (I–V). The lineament zones caused by the D_3 deformation are marked by the serrated lines (modified from Korsman *et al.*, 1988).

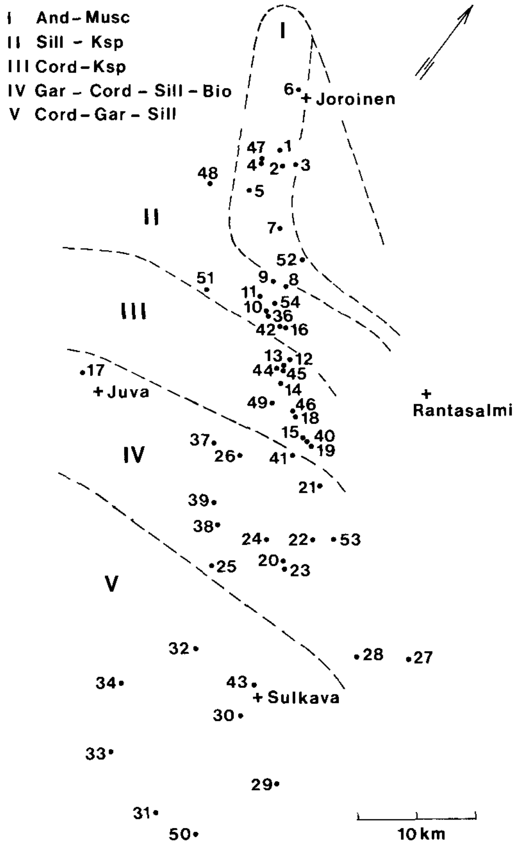

FIG. 3. Fluid-inclusion sample localities in the Joroinen–Sulkava area.

resembles the metamorphism of tectonically thickened crust (Korsman *et al.*, 1984). However, the intense rise in temperature and the formation of the dome cannot be explained solely by a tectonically thickened crust. According to Korsman and Kilpeläinen (1986), the heat pulse of the Sulkava thermal dome could be related to rifting of the crust during the tensional stage, as suggested by some 1.80 Ga-old granitoids. Recently, many authors have argued that a well-focussed passage of hot fluids of deep-seated origin can cause local heating in the crust (see Walther and Wood, 1986). This model is favoured also by Schreurs (1984) for the West Uusimaa Complex in SW Finland.

Location and selection of the samples

The samples were collected from quartz veins associated with different deformation phases (D_2–D_4), from metapelites, metamorphosed tonalites and D_3–D_4 granitoids. Location of the sampling sites are shown in Fig. 3.

Quartz veins varying in width, abundance and age are common in the metapelites. They are usually rare in other rock types. Where folds are present, relative ages of quartz veins are readily apparent. Elsewhere they exhibit the effects of specific deformations and may therefore be dated. Quartz is usually the only vein mineral. Coarse andalusite prisms are found in some D_2–D_3 veins in the metaturbidites. In the less intensively metamorphosed zones, tourmaline- and beryl-bearing pegmatites (D_3–D_4) are common.

Quartz veins associated with D_1 deformation, which preceded progressive metamorphism, have not been found. The oldest quartz veins occur in the andalusite–muscovite and K-feldspar–sillimanite zones, where they are easily recognized. The veins are 0.3–2.0 cm in thickness with a maximum length of about 20 cm. According to Kilpeläinen (pers. comm., 1988) they were formed in *en échelon* tension fractures and deformed during the D_2 event.

The D_3 quartz veins are restricted to the K-feldspar–cordierite zone. Typically, these veins have a strong lineation. In addition, there are younger quartz veins in the garnet–cordierite–

sillimanite–biotite zone, which are interpreted as D_3–D_4 veins. They are rather coarse-grained, varying in thickness from 2.0 to 6.0 cm. The veins lack lineation and are folded by F_3 and F_4.

The metamorphic zones also contain large masses of quartz in the form of veins and lenses up to 1.0 m in width and several metres in length. In part they are cross-cutting, but sometimes they exhibit boudinage structures. These veins and lenses are concordant with the S_2 foliation in the lower-grade zones, where they are deformed during the D_3 stage. Thus, they are interpreted as D_2–D_3 in the north, but as D_3–D_4 in the south.

Analytical method

Microthermometric determinations were performed on a Fluid Inc. heating/freezing stage ('the Reynolds stage', see Shepherd *et al.*, 1985). Double-polished wafers, 250 to 300 μm in thickness were prepared for fluid-inclusion analysis. The stage was calibrated against an ice bath (0 °C), liquid nitrogen (−196.8 °C) and a set of synthetic fluid-inclusion standards (Syn Flinc, U.S.A.). Corrected temperatures are accurate to ±0.1 °C in the range +0.0 to −56.6 °C and ±1.0 to 1.5 °C in the range −90 to −160 °C. The measurements are reproducible to ≤0.2 °C at low temperatures (+30 to −60 °C), but at high temperatures (≤450 °C) the error is probably ≤3.0 °C.

Fluid-inclusion types

The main fluids encountered in the quartz veins (D_2–D_4) and in the matrix quartz of the metapelites are H_2O, CO_2 and CH_4–N_2. Fluid inclusions have been classified according to the different fluid types and their relative ages based on modes of occurrence (Table 1); from the oldest (I) to the youngest (VII) (see Touret, 1981; Touret and Dietvorst, 1983; Roedder, 1984). According to the classification suggested by Van den Kerkhof (1988), the most commonly observed fluid inclusion types in the Joroinen–Sulkava area are types H1 (CH_4–N_2) and H3 (CO_2). However, no systematic observations of the phase transitions in CO_2–CH_4–N_2 inclusions at very low temperatures were made.

Regional distribution

It is commonly reported that the transition from amphibolite to granulite facies is accompanied by an increase in the abundance of CO_2-rich inclusions relative to H_2O-rich inclusions (Touret, 1971, 1972; Konnerup-Madsen, 1977; Schreurs, 1984; Vry and Brown, 1986). However, as pointed out by Touret (1981), the extent to which the different fluid inclusion types overlap metamorphic isograds is complicated by several factors, especially by anatexis. In the Joroinen–Sulkava area, the regional distribution of the different fluid-inclusion types and their relative abundance indicate that prograde metamorphism is marked by a progressive change in fluid composition from H_2O-dominant inclusions in the north to CO_2-dominant in the south (Fig. 4). However, CO_2-rich inclusions are also well developed in the lower-grade zones.

The relative proportions of different fluid-inclusion types were obtained by visually estimating their relative abundance in 10 to 20 quartz grains in different parts of each sample, as described by Konnerup-Madsen (1977) and Touret (1977, 1981).

During low-temperature microthermometric observations relative abundances were checked to confirm the results. Excluded from the estimates are monophase aqueous inclusions (Type VII, Table 1), which are clearly very late post-metamorphic in origin. In the pie diagrams, CO_2-rich inclusion types (I–III) have been grouped together. Early type I and late type III CO_2 inclusions represent only a very small proportion of the total carbonic inclusions.

On a regional scale, the main fluid types (H_2O, CO_2, CH_4–N_2) exhibit considerable spatial variation. In the D_2 quartz veins, the abundance of CO_2 inclusions increases markedly when passing from the andalusite–muscovite to K-feldspar–sillimanite zone. The corresponding increase in CO_2/H_2O inclusion ratio is also shown by inclusions in the matrix quartz of the metapelites. However, the relative proportion of CO_2 inclusions is much higher in the D_2 and D_3 quartz veins than in their host rocks. In the granulite facies domain (Sulkava), CO_2 is the dominant fluid type in the host rock. The change in fluid composition coincides with the initial migmatization of the metapelites. Partial melting of the metapelites begins in the K-feldspar–cordierite zone just before equilibration of garnet and cordierite (Hölttä *et al.*, 1988).

In the northern part of the study area, D_2–D_3 quartz lenses contain almost exclusively CO_2-rich inclusions. However, in the Pirilä gold deposit (K-feldspar–sillimanite zone) the dominant fluid in the gold-bearing D_2–D_3 smoky quartz lenses is CH_4–N_2. In the northern part of the garnet–cordierite–sillimanite–biotite zone, the proportion of CH_4–N_2 inclusions in the narrow and strongly folded D_3–D_4 quartz veins is very high compared with the corresponding veins in the southeastern part of the zone, where the dominant fluid phase

Table 1. Fluid inclusion types in D_2-D_4 quartz veins and in matrix quartz of metapelites in the Joroinen - Sulkava area.

Inclusion type	Composition	Size (µm)	Mode of occurrence	Remarks
I	$CO_2(\pm CH_4-N_2)$	< 5	Isolated, single inclusions. Rarely in small groups.	Fluid density from intermediate to high. In D_2 and D_2-D_3 quartz veins $X_{CH_4}<0.1$. Scarce. No daughter minerals.
II	$CO_2(\pm H_2O)$	< 13	Intragranular inclusion trails.	Fluid density intermediate. In D_3 quartz vein (nr 12) inclusions contain anisotropic daughter minerals. $X_{H_2O}<0.4$.
III	CO_2	< 15	Intergranular inclusion trails.	Fluid density very low. In general relatively rare fluid type. No daughter minerals.
IV	CH_4-N_2	< 20	Intergranular inclusion trails.	Fluid density very low. Locally most common fluid type in the northern part of the study area. X_{N_2}= 0.2-0.5 (<0.7).
V	H_2O-NaCl	< 10	Mostly solitary inclusions.	A number of inclusion generations: (1)~25- 50 vol.% gas ~70-100 vol.% gas ≤ 5- 20 vol.% gas Tubular or negative crystal shaped inclusions. Rare.
		< 15	Intra- and intergranular inclusion trails.	(2)~10- 20 vol.% gas More or less regular shaped, rounded and relatively flat inclusions.
		< 20	Intra- and intergranular inclusion trails.	(3)≤ 5- 10 vol.% gas More or less irregular shaped, angular and very flat inclusions. Daughter minerals (NaCl) are very rare in all types of H_2O-NaCl inclusions.
VI	H_2O-CO_2	< 20	Usually in groups among younger type V H_2O-NaCl inclusions.	(1) 30- 90 vol.% CO_2 Ragged and very flat inclusions. CO_2 phase has a liquid homogenization.
		< 10	Isolated, single inclusions.	(2) 40- 70 vol.% CO_2 Tubular or negative crystal shaped inclusions. CO_2 phase has a gaseous homogenization. Very rare.
VII	H_2O	<100	Very weakly healed microfractures.	Only one phase (liquid) at room temperature. Found in varying amounts in all samples.

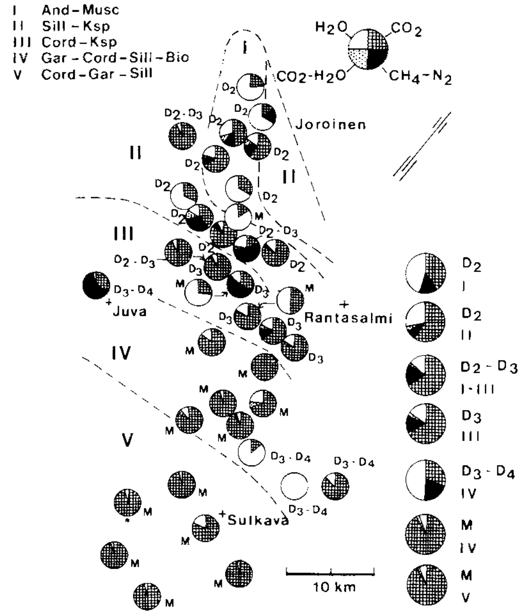

FIG. 4. Regional distribution of fluid inclusion types and their relative abundance within the different metamorphic zones (I–V). D_2–D_4 refers to quartz veins associated with the different deformation phases and M to matrix quartz of the host rock (i.e. metapelite).

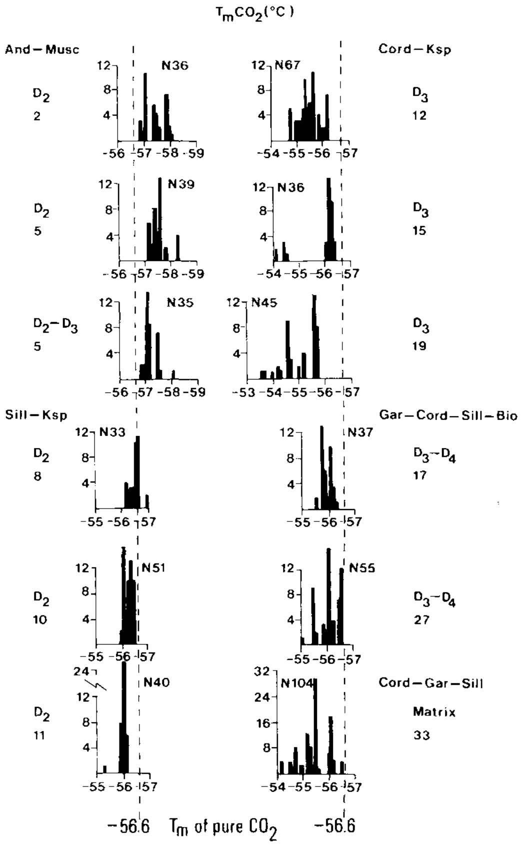

FIG. 5. Melting temperatures of CO_2-rich inclusions (Types I–II) in the D_2–D_4 quartz veins and in the quartz of the host rock. Numbers refer to the samples listed in Fig. 3. The T_m-temperatures are corrected to a pure CO_2 standard (Syn Flinc).

consists solely of H_2O (Type V, Table 1). Excluding the Pirilä gold deposit, coarse-grained and boudinaged D_2–D_4 quartz lenses do not contain CH_4–N_2 inclusions. This inclusion type is also absent from the matrix quartz of the metapelites. The relatively narrow zone enriched in CH_4–N_2 fluids is located near the lineament zones caused by the D_3 deformation (see Fig. 2).

Microthermometry

Melting temperatures of CO_2-rich inclusions (Types I–II) in the quartz veins and in the quartz grains of their host rocks (i.e. metapelites) increase with increasing metamorphic grade (Fig. 5). This is clearly shown by inclusions occurring in the D_2 quartz veins in the less intensively metamorphosed zones. In the andalusite–muscovite zone, the depression of melting points below the triple point of pure CO_2 indicates the presence of additional components such as CH_4 and/or N_2. The amount of these components in the CO_2 inclusions decreases with increasing metamorphic grade. However, the melting temperatures are close to pure CO_2, showing a maximum melting point depression of 1.8 °C. This is equivalent to an admixture of maximum $X_{CH_4} = 0.1$ (Burruss, 1981; Shepherd *et al.*, 1985). As indicated by Fig. 5, the purest CO_2 inclusions occur in the K-feldspar–sillimanite zone. This is also the case for Al_2SiO_5-bearing pelitic rocks from Mica Creek, British Columbia (Stout *et al.*, 1986). For the relatively dense (0.9 g/cc) type II CO_2 inclusions with a maximum 20 vol.% undetected H_2O, the molar proportion of H_2O is about 0.4 (see Schwartz, 1989). Estimated molar proportion of N_2 in CH_4–N_2 inclusions inferred from the phase equilibria (Van den Kerkhof, 1988) varies most commonly between 0.2 and 0.5, with a maximum value of about 0.7 (see also Table 1).

Homogenization temperatures of CO_2 inclusions (I–III) vary from −28 to +30 °C (Fig. 6). Usually, the distribution of T_h-temperatures is

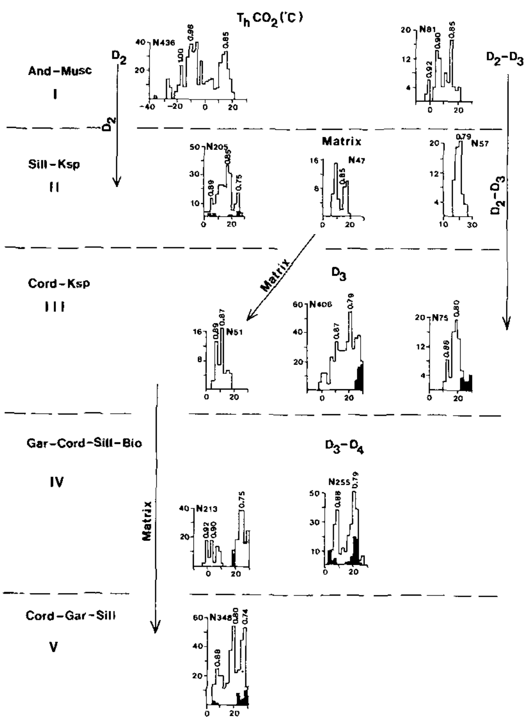

FIG. 6. Histograms of homogenization temperatures of CO_2-rich inclusions (I–III) in the D_2–D_4 quartz veins and in the host rock (black; type III inclusions with a homogenization into vapour state) in the different metamorphic zones (I–V).

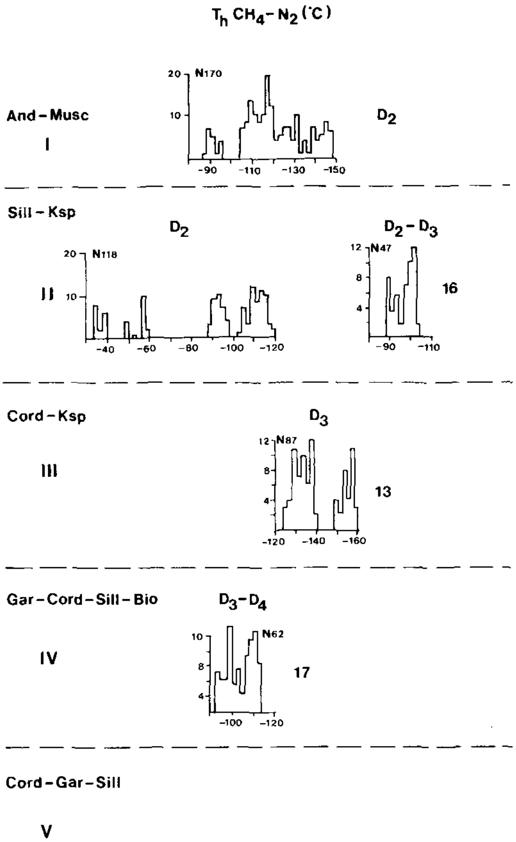

FIG. 7. Histograms of homogenization temperatures of CH_4–N_2 inclusions (Type IV) encountered in the D_2–D_4 quartz veins near the lineament zones caused by the D_3 deformation.

bimodal in each sample. Lower and higher T_h-temperatures correspond to older (I) and younger (II–III) CO_2 inclusions respectively. Homogenization of CO_2 inclusions occur into the liquid state, except in type III inclusions, which exhibit homogenization into the vapour state. CO_2 densities corresponding to T_h-temperatures are presented in Table 2. Very dense type I CO_2 inclusions encountered in some D_2 quartz veins do not represent pure CO_2 fluid. Thus, densities derived from their T_h-temperatures must be considered as maximum values. In general, the homogenization temperatures increase with increasing metamorphic grade. T_h-temperatures for CO_2 inclusions in the matrix quartz of the metapelites show no notable change with metamorphic grade. In the K-feldspar–sillimanite and K-feldspar–cordierite zones, the density of early type I CO_2 inclusions in the quartz grains of their host rocks as well as in the D_2 and D_3 quartz veins are similar. T_h-temperatures for the younger type II CO_2 inclusions and their corresponding peaks in the T_h-histograms are very uniform, both in the quartz veins and in the matrix quartz of the metapelites throughout the study area.

CH_4–N_2 inclusions develop a rim of liquid around the inclusion walls on cooling below −90 °C. Generally, this liquid disappears on heating to −160 and −87 °C (Fig. 7). However, some hydrocarbon inclusions have T_h-temperatures of −34 to −60 °C, well above the critical temperature of CH_4 (−82.6 °C). Because there is no solid CO_2 phase below −56.6 °C they are considered CO_2-free. As pointed out by Van den Kerkhof (1988), T_h of CH_4–N_2 inclusions is gradually lowered with higher amounts of N_2. In the Joroinen–Sulkava area, T_h-temperatures of the inclusions show no apparent uniform or gradual change with increasing metamorphic grade.

The salinity of aqueous inclusions (Type V) tends to decrease with increasing metamorphic grade and with decreasing age of the quartz veins,

Table 2. Density of fluid inclusion types I-VII in the Joroinen-Sulkava area.

Inclusion type	Composition	D_2-D_4/ Matrix	Metamorphic zone	Fluid density g/cc
I	$CO_2(\pm CH_4-N_2)$	D_2	I	1.05 - 0.98
I	- " -	D_2	II	0.89 - 0.86
I	- " -	D_2-D_3	I	0.92 - 0.90
I	- " -	D_2-D_3	III	0.87 - 0.86
I	- " -	D_3	III	0.91 - 0.87
I	- " -	D_3-D_4	IV	0.90 - 0.88
I	- " -	M	II	0.88 - 0.86
I	- " -	M	III	0.90 - 0.88
I	- " -	M	IV	0.92 - 0.88
I	- " -	M	V	0.90 - 0.87
II	$CO_2(\pm H_2O)$	D_2	I	0.93 - 0.81
II	- " -	D_2	II	0.81 - 0.78
II	- " -	D_2-D_3	I	0.88 - 0.79
II	- " -	D_2-D_3	II	0.81 - 0.79
II	- " -	D_2-D_3	III	0.85 - 0.75
II	- " -	D_3	III	0.85 - 0.75
II	- " -	D_3-D_4	IV	0.85 - 0.79
II	- " -	M	II	0.84 - 0.81
II	- " -	M	III	0.88 - 0.86
II	- " -	M	IV	0.79 - 0.74
II	- " -	M	V	0.85 - 0.75
III	CO_2	D_2-D_4/M	I-V	<0.35
IV	CH_4-N_2	D_2-D_4	I-IV	<0.30
V	H_2O-NaCl	D_2-D_4/M	I-V	0.97 - 0.52
VI	H_2O-CO_2	D_2-D_4/M	I-V	0.90 - 0.50
VII	H_2O	D_2-D_4/M	I-V	1.00 - 0.98

from 6.0 eq.wt.% NaCl (average) in the andalusite–muscovite zone (D_2 veins) to 1.9 eq.wt.% NaCl (average) in the granulite facies (matrix, Fig. 8). Salinities were calculated using the equation of Potter *et al.* (1978). Homogenization temperatures of H_2O inclusions vary from <120 to about 420 °C. However, there is an almost complete overlap of T_h-temperatures for the different groups (Fig. 8). Anomalously high T_h-temperatures of aqueous inclusions in some samples may be due to necking-down phenomenon or re-equilibration to *PT* conditions after their entrapment (see Pecher, 1981). Thus the variation in T_h of H_2O inclusions may reflect incomplete equilibration and not necessarily a range of trapping conditions.

Geobarometry

The density of early type I CO_2 inclusions, together with the estimates for metamorphic temperatures in the different metamorphic zones, indicate a pressure range of 3.0–4.5 kbar (Fig. 9). This is consistent with the data derived from mineral geobarometry and metamorphic assemblages (Korsman *et al.*, 1984). Thus, the high-density data of early type I CO_2 inclusions when considered in conjunction with pressure–temper-

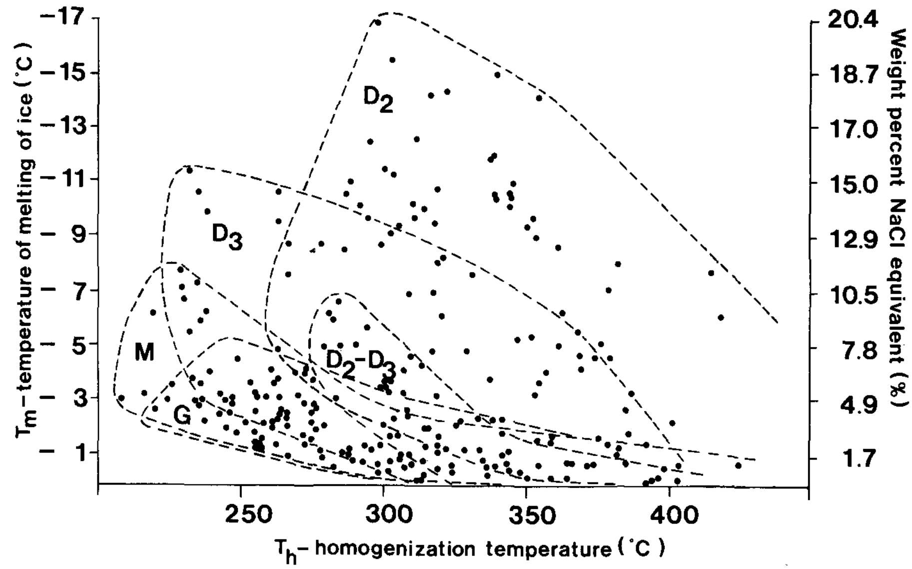

FIG. 8. Temperatures of homogenization versus salinity for the aqueous inclusions (Type V) in the D_2–D_3 quartz veins and in the host rock (M and G: Gar–cord–sill–bio and cord–gar–sill zone respectively).

ature data from mineral chemistry strongly confirm their synmetamorphic origin during the different stages of progressive metamorphism. Isochores for CO_2 were taken from Touret and Bottinga (1979).

In the andalusite–muscovite zone, the early type I CO_2 inclusions contain up to 10 mole% $CH_4(\pm N_2)$. This may result in an overestimation of the entrapment pressure by about 1.0 kbar at 600 °C (see Newton, 1986) as shown by the solid arrows in the Fig. 9. However, the D_2 veins also contain nearly pure younger type II CO_2 inclusions whose density indicates a trapping pressure of 3.7 kbar at 600 °C. This is approximately concordant with the pressure estimate of 3.4 kbar from mineral equilibria calculations (Korsman *et al.*, 1984).

Density of early CO_2 inclusions in the matrix quartz of the metapelites in the different metamorphic zones indicate a steady pressure increase with increasing metamorphic grade. In the different metamorphic zones, the densities of CO_2 inclusions in the quartz veins and in the host rocks are almost similar. Thus, the quartz veins of different relative ages seem to be closely connected with the progressive metamorphism of the area. These observations support the idea that the relative age of quartz veins become younger with increasing metamorphic grade and with decreasing age of metamorphism. The D_2–D_3 quartz lenses in the K-feldspar–sillimanite zone contain only younger type II CO_2 inclusions. The inferred pressure estimate is somewhat lower than that derived from solid phase estimates.

Mineral geobarometry indicates a pressure increase inside the Juva zone (gar–cord–sill–bio, Fig. 9) from north to south. However, the crystallization temperature remained unchanged. In the host rocks the density of early type I CO_2 inclusions increases towards the south, but their density does not correspond with pressure estimates derived from minerale quilibria. This may be due to an intense temperature increase associated with the formation of the Sulkava thermal dome, which probably caused decrepitation of early inclusions and re-equilibration to give higher molar volume values (i.e. lower densities). In contrast, early type I CO_2 inclusions in host rocks of the granulite facies domain probably represent synmetamorphic and associated peak-metamorphic fluids.

Later generations of younger type II and III CO_2 inclusions with decreasing fluid density indicate a possible post-metamorphic *PT* path for the Joroinen–Sulkava area. To prevent decrepitation of the higher density CO_2 inclusions in the quartz veins and their host rocks during unloading, it must be assumed that the *PT* path never dropped to pressures more than 1–2 kbar below the pressures indicated by the isochores for high-density inclusions at any given temperature (see Naumov

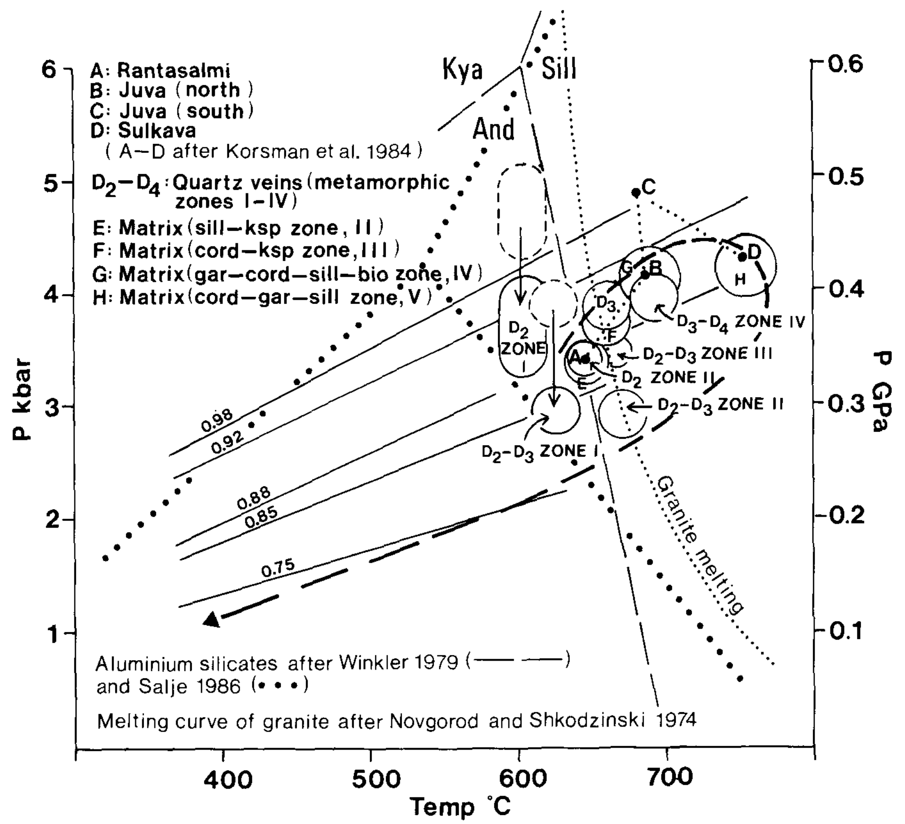

FIG. 9. A model of the evolution of the metamorphic fluids in the Joroinen–Sulkava area. Density of early type I CO_2 inclusions together with estimates of metamorphic temperatures (A–D) in the different metamorphic zones indicate a pressure range of about 3.0–4.5 kbar during the prograde stage of metamorphism (E–H). The relative ages for the quartz veins (D_2–D_4) are closely connected with the progressive metamorphism of the area. Fluid densities (dashed circles) derived from the T_h-temperatures of the impure, early type I CO_2 inclusions (X_{CH_4} ≤ 0.1) encountered in some D_2 and D_2–D_3 quartz veins and lenses (and–musc zone) are not representative, as indicated by the solid arrows. The possible post-metamorphic *PT* path for the area is indicated by the broken arrow.

et al., 1966; Leroy, 1979; Hollister *et al.*, 1979; Selverstone *et al.*, 1984; Rudnick *et al.*, 1984).

Conclusions

In the Joroinen–Sulkava area the regional distribution of the different fluid inclusion types may be explained by anatexis and very gently dipping isograd surfaces. A model of progressive enrichment of CO_2 into the fluid phase and the loss of H_2O into the anatectic melts (see Touret and Dietvorst, 1983) is supported by the abundant H_2O-rich inclusions in the granitic leucosome. However, the possibility of CO_2-infiltration from deeper crust can not be excluded. It is most probable that the granulite facies rocks occur below the lower-grade zones.

The quartz veins mark the preferred pathways for escaping fluids generated during prograde metamorphism. Vein formation is considered to have taken place under a limited range of *PT* conditions during different stages of metamorphism. Further, the quartz veins are characterized by a diversity of fluid types (H_2O, CO_2, CH_4–N_2), which would have been miscible at the prevailing metamorphic temperatures (i.e. >645 °C). However, the observed fluid inclusions contain only one fluid type (usually CO_2 or H_2O) with little or no mixing (excluding some late composite H_2O–CO_2 inclusions). Thus, it is likely that the fluid inclusion types in these veins are a result of different fluids having passed through the veins at the different times without re-equilibrating with the wall rocks.

In the Sulkava thermal dome area, the influx of synmetamorphic (D_2) CO_2-rich fluid was followed and heavily overprinted by a second influx of CO_2-rich fluids of deep-seated origin associated with the intrusion of microcline granite (updoming) during the D_4 deformation. This is supported by a preliminary study of fluid inclusions in the microcline granite. The granite contains several successive generations of relatively dense (>0.85 g/cc) CO_2-rich inclusions.

The mode of occurrence and very low-density of CH_4–N_2 inclusions suggest their entrapment late in the metamorphic evolution of the rocks. This fluid type may be considered as a residual fluid (see Van den Kerkhof, 1988), which had concentrated near the lineament zones caused by the D_3 deformation. The composition of the original fluid may be related by the early type I CO_2-rich inclusions ($X_{CH_4} \leq 0.4$) encountered in the microcline granite and associated quartz–feldspar veins in the granulite facies domain.

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