Migmatite by partial fusion and short range hydrothermal transfer, British Columbia

Autor(en):   **Blattner, Peter**

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Migmatite by Partial Fusion and
Short Range Hydrothermal Transfer, British Columbia

By *Peter Blattner* (Lower Hutt, New Zealand)*)

With 11 figures and 5 tables in the text

Abstract

Variations of mineral chemistry with locality and with rock type have been determined for a typical migmatite. Distribution of Rb, Sr, and O^{18} between minerals suggests lowest recorded temperatures of 600° C. In view of well-known experimental data fusion may reasonably have started in the rocks; on account of limitations in the supply of volatiles it has probably remained incomplete. Partial pegmatoid melts, when crystallized, would be enriched in K-feldspar and would differ in feldspar chemistry from their surroundings. However, with the crystallization of melts volatiles are released. Hydrothermal transfer of cations at this stage may account for the mineralogical control of pegmatoid vein and host gneiss compositions in the present as well as in other occurrences. Elongate granite bodies, co-zonal with mesoscopic folds, are believed to have formed through local concentration of creep. A single batch of volatiles may have been transferred adiabatically from crystallizing pegmatoid melts to the newly forming granites. Local pre-migmatization chemistry has been partly preserved.

Zusammenfassung

In einem typischen Migmatit sind Unterschiede der Mineralchemie in Abhängigkeit von Lokalität und Gesteinstyp bestimmt worden. Die lokalen Verteilungen von Rb, Sr und O^{18} zwischen Mineralien, obwohl sicher teilweise retrograd, ergeben Temperaturen um 600° C. Auf Grund bekannter experimenteller Daten scheint es dann wahrscheinlich, dass der solidus der Gesteine erreicht worden ist. Mangel an H_2O müsste das Ausmass der Anatexis stark einschränken. Pegmatoide Teilschmelzen würden Kalifeldspat anreichern und sich in der Zusammensetzung der Feldspäte von ihrer Umgebung unterscheiden. Die Kristallisation würde jedoch H_2O freisetzen und die Einstellung hydrothermalen Gleichgewichts zwischen pegmatoiden Adern und Wirtgneiss ermöglichen. Die Monotonie der Plagioklaszusammensetzung in dem hier beschriebenen und in anderen Migmatitvorkom-

*) New Zealand D.S.I.R., Geological Survey, P.O. Box 30368, Lower Hutt, New Zealand.

men überrascht daher nicht und wird sich nur selten als Argument gegen Bildung durch Anatexis anwenden lassen. Den Faltenachsen B parallele, gangähnliche Granitkörper sind diskordant zu dem von Gneiss und pegmatoiden Adern gebildeten Flächengefüge und können sich durch lokal konzentrierte Kriechdeformation gebildet haben; eine gegebene Menge von H_2O kann adiabatisch von den kristallisierenden pegmatoiden Schmelzen an sich neu bildende granitische Schmelzen abgegeben worden sein. Reste der prä-migmatitischen Mineralchemie sind namentlich in Amphiboliteinschlüssen erhalten.

INTRODUCTION

Since SEDERHOLM (1907, 1967) drew attention to the deep-seated rocks he called migmatites, the subject has remained controversial in spite of the vast increase in available data and methods of study. The morphology of migmatites often suggests that material has been added to a pre-existing rock, but MEHNERT (1951), WHITE (1966), and in part LUNDGREN (1966) concluded that migmatization may have taken place in closed systems. HÄRME (1965) postulated addition of potassium during migmatization. However, data by YODER et al. (1957) and TUTTLE and BOWEN (1958), as well as the work of PLATEN (1965), have shown that rock portions enriched in K-feldspar can be expected from the partial fusion of An-bearing granitoid rocks. The expected enrichment could not explain the findings of WHITE (1966), who therefore rejected a partial fusion origin of the Palmer region migmatite. In many cases, evidence for plagioclase fractionation might strongly support partial fusion hypotheses. Studying veined gneisses, MEHNERT (1962), WHITE (1966), and MISCH (1968) have found no or only minor sodic shifts in favor of vein plagioclase. Partly on this basis MISCH ruled out a partial fusion origin, whereas MEHNERT assumed that uniform plagioclase compositions had been established after migmatization. Experiments by ORVILLE (1963) suggest possible mechanisms for subsolidus equilibration of feldspars over a distance.

GEOLOGICAL SETTING

The area studied, approximately 200 by 200 meters, lies in the deepest exposed level of the Thor-Odin gneiss dome, Monashee Mountains, British Columbia. The rocks are part of REESOR and MOORE's (in press) "veined granodiorite gneiss". For the most part they may consist of originally intrusive Mesozoic granodiorite. Formerly the whole "Shuswap Terrane" was considered to be Precambrian, possibly Archean basement, underlying the Windermere and early Paleozoic sedimentary sequence (JONES, 1959). The exposures lie at 50° 32′ 59″ N, 118° 04′ 53″ W, i. e. 3.3 km east of Mt Odin, at 7150 ft above sea level. They consist largely of fresh, glacially polished surfaces.

DESCRIPTION OF ROCKS

Fig. 1 is a sketch map with sample localities; structure elements are plotted in fig. 2. The diagram indicates N-S trending axes B_{S_1} of subhorizontal folds, with foliation planes S_1 dipping mostly about 35° WSW. Alternate, more steeply inclined limbs of the asymmetric folds are strongly attenuated, or their place is taken by flexural shear zones, defining a second preferred foliation plane S_2 (Fig. 3 a). S_1 and S_2 appear co-zonal, with B_{S_1} the zone axis. The
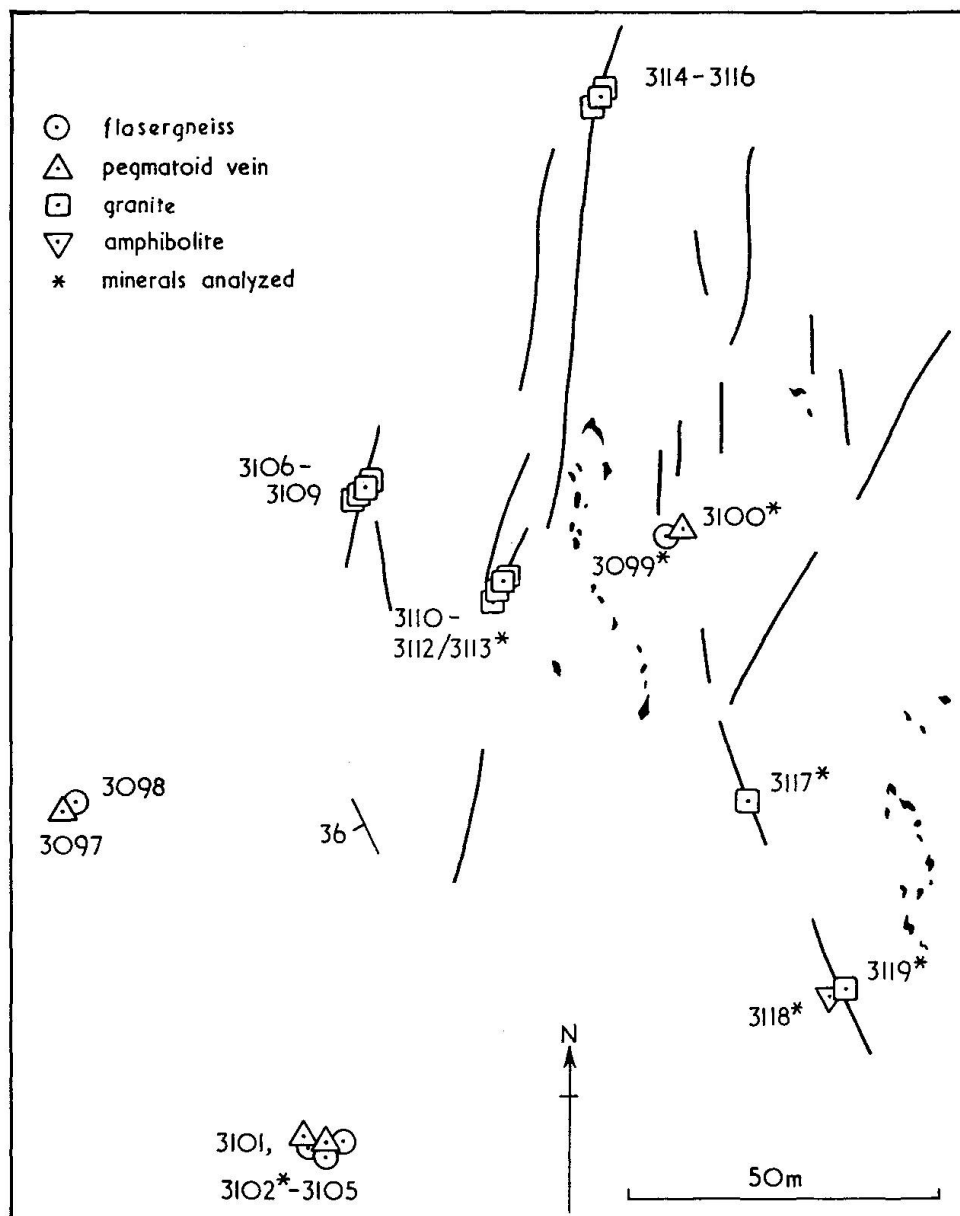

Fig. 1. Migmatite 3.3 km east of Mt. Odin; sketch map and sampling plan. Lines are shear zones S_2, small black areas are amphibolite inclusions. Foliation with dip 36 is typical S_1.
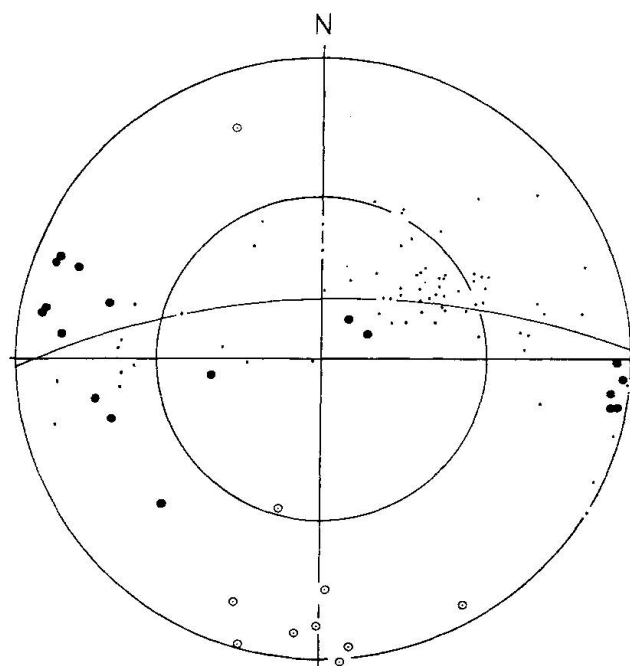
Fig. 2. Projection of fabric elements: Poles to foliation planes S_1 (small dots), axes of mesoscopic folds B_{S_1} (open circles), and shear planes S_2 (solid circles). Equal area, lower hemisphere.

Table 1. *Analyzed rock samples*

No.	Rock type	Position
3097	pegmatoid vein	} contiguous
3098	flasergneiss	
3099*	flasergneiss	} contiguous
3100*	pegmatoid vein	
3101	pegmatoid vein	
3102*	flasergneiss	} contiguous
3103	contact 3101–3102, including mafic selvedge	
3104	flasergneiss	} 20 cm apart
3105	pegmatoid vein	
3106		0 cm S of 3106
3107		10 cm S of 3106
3108		20 cm S of 3106
3109		200 cm S of 3106
3110	granite (drill cores)	0 cm S of 3110
3111		10 cm S of 3110
3112*		200 cm S of 3110
3113*		210 cm S of 3110
3114		0 cm S of 3114
3115		100 cm S of 3114
3116		110 cm S of 3114
3117*	granite	
3118*	amphibolite, inclusion (0.3 m diameter) in granite	} contiguous
3119*	granite	

* = mineral analyses available for this rock
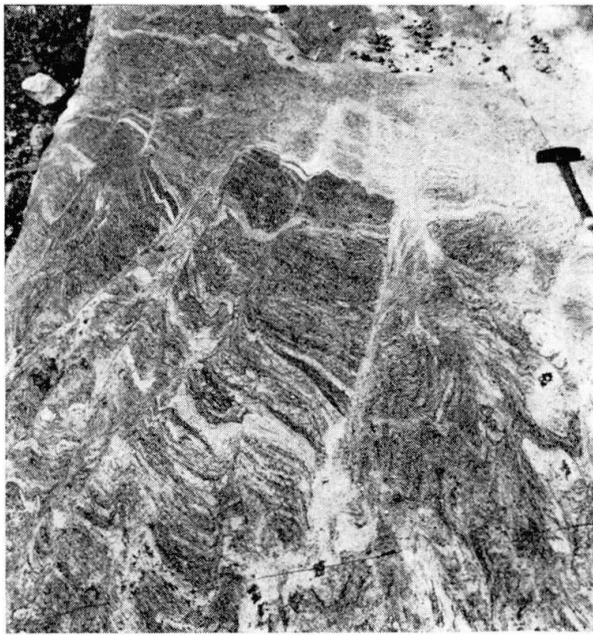
a)

b)

c)

d)

Fig. 3. Field relations. a: flasergneiss with pegmatoid veins ($//S_1$) and flexural shears ($//S_2$). b: intensely transposed S_1. c: small granite body; scale as in next frame. d: agmatite in shear zone; granite "dike" is one meter wide; its contact is seen to the left of and parallel to the hammer handle. Disregard glacial striae.
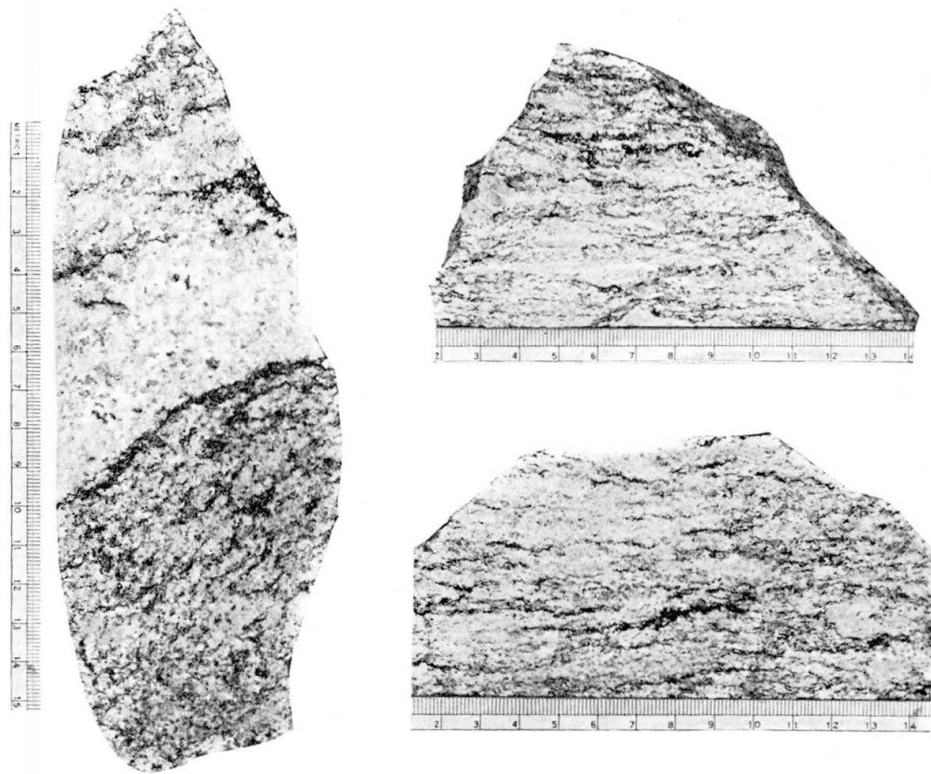
Fig. 4. Sections of flasergneiss. On left with pegmatoid vein and selvedge; section at high angle
to B_{S_1}. Scales in centimeters.

few subhorizontal shears present confirm this view. A direction of shearing in
S_2 is not obvious. Most shears end within short walking distance and the
displacement is less than about one meter.

BLATTNER (1965) has drawn attention to the strong anisotropy of typical
migmatite fabrics ("B-migmatites"). As in the present case, such fabrics may
be typified only if the structural coordinates are taken into account.

The most abundant rock type is a hornblende-bearing granitoid biotite
gneiss (Figs. 3 and 4) with a coarse, mainly planar flaser structure (NIGGLI,
1954). "Granitoid" denoting an approximate composition rather than struc-
ture, the term *granitoid flasergneiss* is used. Within the flasergneiss strings of
isolated *amphibolite inclusions* occur (Fig. 1), which may once have formed
continuous layers or sheets (former dikes?). Subconcordant *pegmatoid layers*
or *veins* of less than about 0.3 m thickness occur throughout the flasergneiss
(Fig. 3a, 3b). In some places selvedges of mafic minerals (i. e. mainly biotites)
are developed alongside the veins (Fig. 4).

The rocks considered so far appear to have been affected as a unit by the
folding B_{S_1} and the shearing S_2, and outside of the shear zones are largely
concordant. The shear zones themselves are not uniformly developed. Over
tens of square meters intense transposition of S_1 planes has given rise to com-
plex looking rock mixtures (Fig. 3b). In other areas the shearing has been

confined to more regularly spaced, parallel planes (Fig. 3a). Irregular aggregates of granite with a massive fabric are present along shear zones, often as if penetrating or replacing adjacent flasergneiss and pegmatoid veins (Fig. 3c). These *granites* constitute elongate bodies up to 1 meter wide, and some look like intrusive dike rocks, giving rise in one place to agmatite structures (Fig. 3d). For clarity the expression "vein" will be used for the pegmatoid veins only.

That the shearing S_2 has taken place as a separate tectonic event, after completion of the folding, seems improbable in view of the orientations of S_2. It is noted that the present structure has a counterpart on a samller scale in schists with an early foliation cut by strain-slip cleavages (TURNER and WEISS, 1963): both S_2 and B_{S_1} are likely to have formed in one sequence of deformation. In the present case, however, formation of the "strain-slip cleavage" appears linked to the formation of a new rock type, i. e. granite.

Mineralogy and Textures

For convenience, typical modal compositions have been estimated from mineral separations and chemical analyses (Table 2). According to STRECK-

Table 2. *Modes of typical rocks*

Sample no.	3099 f	3100 p	3117 g	3118 a	3119 g
Quartz	33	31	34	4	33
K-feldspar	19	43	28	nil	17
Plagioclase	38	24	30	22	36
Biotite	8	1	6	11	9
Hornblende	1	0.0	1	62	4
Sphene	trace	nil	trace	0.3 ⎫	
Allanite and zircon	0.3	nil	0.3	nil ⎬	0.5
Apatite	0.1	0.0	0.2	0.1	0.3

Letters f, p, g, a for flasergneiss, pegmatoid vein, granite, amphibolite

Table 3. *Plagioclase compositions*

Sample no.	3099 f	3100 p	3117 g	3118 a
Wt. % An (U-stage)	22–27*	22–29*	17–29*	23–32
Estimated weighted average	26	26	27	28
Wt. % An (Chemical analysis)	25.0	25.2	27.3	26.6
Electron microprobe: Range of seven spot counts CaK$_\alpha$ (counts per second; $\sigma = 0.3$)	11.8–12.9			11.4–12.8

* = also minor albitic rims. Letters f, p, g, a as in table 2

EISEN (1967) the compositions of samples nos. 3100 and 3117 correspond to "granites", those of nos. 3099 and 3119 to "granodiorites".

Ranges of plagioclase compositions have been determined and average compositions estimated by U-stage and electron microprobe methods (Table 3). An erratic occurrence of elements, as found in some cases by GONI (1966), might cause difficulties in interpreting the compositions of bulk rocks and mineral concentrates. Electron microprobe spot analyses (sample no. 3099) have broadly confirmed the value K_D^{Ba} (Bio./K-feld.) = 0.27 found chemically (0.34 by microprobe) and 26 random spot analyses of K-feldspar show a simple, approximately normal distribution of Ba counting rates between 0.52 and 0.83 cps ($2\sigma = 0.1$ cps). Both observations support the assumption that essentially all of this element contained in the present rocks occurs throughout the crystal structures.

The complex granoblastic fabric appears uniform for the granitoid rocks. Myrmekite is always present and locally plagioclase crystals have narrow albitic rims. Relationships between relatively sodic and calcic domains of plagioclase are irregular and antiperthitic inclusions are frequent. K-feldspars are "cross-hatched" in places (small crystals and parts of crystals) and sometimes have fine perthite lamellae. Concentrates from all granitoid rock types gave triclinicities $\Delta = 12.5 \, [d_{(131)} - d_{(1\bar{3}1)}]$ of $\lesssim 0.2$. In the amphibolite, myrmekite and albitic rims of plagioclase are lacking, along with K-feldspar, and the otherwise dark brown biotites and dark green hornblendes are somewhat lighter colored. Iron oxides are rare and together with chlorite, omnipresent in small amounts, may be of secondary origin.

SAMPLING AND ANALYSIS

Samples (Fig. 1, table 1) consisted of two or more kilograms of rock, obvious discontinuities of structure or composition having been removed. The four samples of pegmatoid veins weighed between 0.5 and 1.5 kilograms. Sampling was facilitated by a light, portable drill (GAUCHER and MEILLEUR, 1965). Drill cores weighed between 100 and 150 grams. Metal from the drill and saw was removed. 60 mesh size fractions were used for mineral separation. Concentrates were purified by centrifuging and with a "Frantz" magnetic separator. Impurities were estimated with the microscope; the estimates for hornblendes were $< 1\%$ biotite, $< 0.5\%$ allanite and zircon, $< 0.2\%$ quartz and feldspar. The only more serious impurity found were $< 2\%$ K-feldspar in quartz-plagioclase concentrates.

The following *methods of analysis* were used, at the Geological Survey of Canada unless stated otherwise (for analysts see table 4): Gravimetry (Ca, Mg), titrimetry (Fe), flame photometry (Na, K), spectrophotometry (Ti, Mn, P, F after distillation), atomic absorption (Mg, Li), and a modified Penfield method for H_2O^+ (MAXWELL, 1968; ABBEY, 1967) – Emission spectrography: Internal standard technique with Stallwood air jet, by W. H. Champ (Ca, Ti, Mn, Sr, Ba, Sc, Y, Yb, V, Cr, Co, Ni, Cu). – X-ray fluorescence:

Fusion technique using an ARL multichannel spectrograph (major elements, Ti, Mn; MAXWELL, 1968).

Rb and Sr in bulk rocks were determined on a single channel XRF spectrograph. The electron microprobe used was an MAC model 400, supervised by G. R. Lachance. Many results were obtained by more than one procedure.

Rb and Cs in biotites and K-feldspars were determined by atomic absorption at Chemistry Division, and relative O^{18}/O^{16} ratios according to CLAYTON and MAYEDA (1963) at the Institute of Nuclear Sciences, New Zealand D.S.I.R. For $\delta\,O^{18}_{SMOW}$ see CRAIG (1961).

Precision and Accuracy

Many of the analyses (Tables 4, 5) have been related to analyses of international rock standards as discussed by BLATTNER (1970). In this way the *accuracy* of data is open to evaluation, while at the same time the problem of absolute values is eliminated from geochemical comparison.

A minimum analytical *precision* believed to have been attained is expressed in the quoted figures, if the last large type digit is taken to vary within less than 10 units ($2\,\sigma$). This may suggest somewhat too good precision for Mg and Ti, and on the other hand will fail to do justice to some of the data. Attention is drawn to the good precision of the emission spectrographic Sr and Ba determinations. It is due to the near perfect line matches Sr 4607/Eu 4627 and Ba 4554/Eu 4523, Eu being the internal standard. Relative precision for single determinations is about 5%, but almost all analyses were repeated at least once. Similarly good precision has been reached for Rb in minerals, and for Li; relative precision for Na_2O, K_2O, and FeO is about 2%. Ba determinations in biotite were difficult because of Ti interference with Eu.

PHYSICAL AND CHEMICAL CONDITIONS

Element fractionations between minerals of a rock may at best be indicators of physical conditions at a specific time (mainly of temperatures; KRETZ, 1961; ALBEE, 1965). Alternatively, compositions of different minerals may correspond to different sets of conditions (e. g. O'NIONS et al., 1969), and give false distribution coefficients. Analyses that may be interpreted thermometrically are summarized in fig. 5. A retrograde effect (GREENLAND et al., 1968; JAMES and CLAYTON, 1962) can be expected alone or in addition to disequilibrium effects. Temperatures indicated in fig. 5 are therefore expected to be on average lower than peak temperatures attained in the rock system. The three independent estimates lie between 550 and 750° C. A variation of temperature conditions between rock types might be expected to occur in rocks containing intrusive veins or dikes, but is not shown consistently by the present data. Estimated temperature resolution for Rb, Sr, and O^{18} fractionations used is 50° C, i. e. for the present rocks precise temperatures are expected to agree within 50° C for each of these species. If distribution coefficients were false in the above
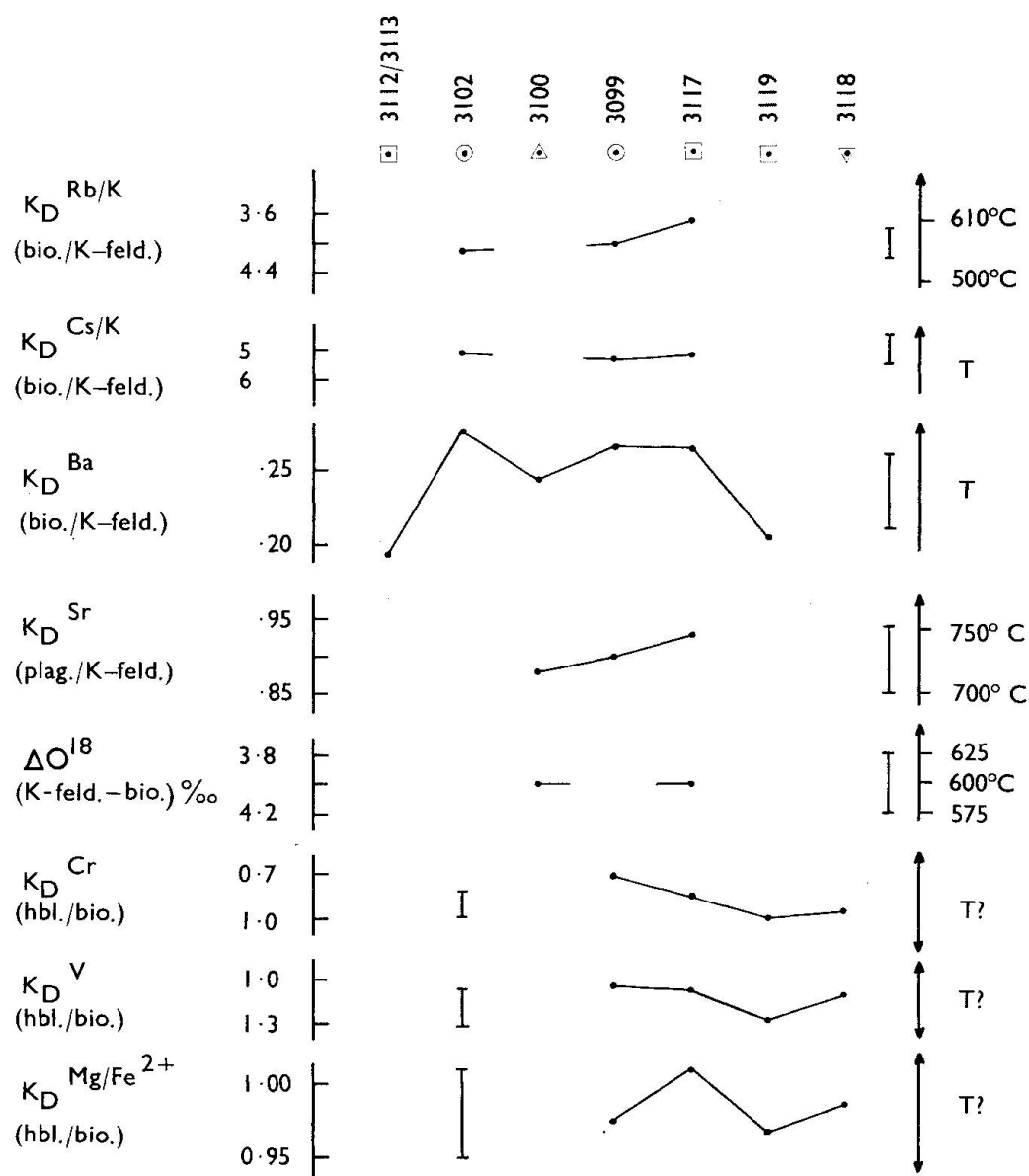
Fig. 5. Lateral variation of distribution coefficients. Limits of error suggest possible temperature resolution of various "thermometers". Left hand scales are linear; symbols for rock types as in fig. 1. Sources of temperature calibration in text.

sense, their constancy would still indicate pervasive thermal equilibrium over the period of recording.

Temperature calibrations in fig. 5 are from the following sources:

1. For Rb/K (Biotite/K-feldspar) from BESWICK and EUGSTER (1968). The calibration applies strictly for phlogopite and sanidine only. In the present temperature range a small positive correction has since been made (20° C; BESWICK, pers. comm.).

2. For Sr (Plagioclase/K-feldspar) from Slade (in: VIRGO, 1968).

3. Based on data from TAYLOR and EPSTEIN (1962), and on calibrations reviewed by TAYLOR (1967), a temperature of 600° C has been assigned to the value of $4^0/_{00}$ Δ O^{18} (K-feldspar-biotite). This estimate is tentative, because of the variable OH^- content of biotites. However, for constant OH^- content the temperature sensitivity of $50°$ $C/0.4^0/_{00}$ Δ will be nearly correct.[1])

The migmatite is overlain by pelitic rocks carrying sillimanite. According to RICHARDSON et al. (1969) and ALTHAUS (1967) this indicates a minimum temperature, at 6 kb, of ~600° C. According to REESOR and MOORE (in press) the load pressure (P_t) before metamorphism should have been above 5 kb for the present area.

An estimate of the oxygen fugacity, obtained from biotite and hornblende compositions (Fig. 9) allows construction of a breakdown curve for idealized biotite with present Fe/Mg ratios ($Fe/Fe + Mg \leqq 0.7$) in the presence of quartz (Fig. 10; WONES and EUGSTER, 1965; EUGSTER and WONES, 1962).

Chemical Environments

If the recorded distribution coefficients date from a time or period of pervasive thermal equilibrium they also show that chemical equilibrium was far from pervasive. Minerals of the amphibolite represent a different chemical environment from that of the granitoid rocks with respect to most elements (Fig. 6a, 6c; also, notably, Sc, Y, Yb, Co, Ni, Cu). Among the granitoid rocks Ba in analysed biotite and K-feldspar varies by a factor of up to two with locality (Fig. 6d). The Ba contents of K-feldspar may also be deduced from bulk rock analyses. Results (Fig. 7) confirm lateral variations that affect both flasergneiss and pegmatoid veins. Correction factors for Fig. 7 are based on the mineral analyses; any systematic difference between K-feldspar of gneiss and veins is probably less than 5% of the Ba present. Inspection of tables 4 and 5 shows that the Sr abundance pattern is similar to that of Ba, whereas Rb is more evenly distributed.

Feldspar compositions in Ab, Or, and An cannot be shown to differ between any of the rock types, including the amphibolite (Table 3). Weight norms have been calculated for bulk analyses of the granitoid rocks as follows:

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[1]) Sine this paper has gone to press, the oxygen isotope fractionation between quartz and K-feldspar has been determined for two of the rocks. The results are, for no. 3100 Δ (qz.–K-feld.) = $1.45^0/_{00}$, and for no. 3117 Δ (qz.–K-feld.) = $1.65^0/_{00}$. According to the calibrations quoted by TAYLOR (1967), these values correspond to 630 and 540° C, respectively. They agree surprisingly well with the "Rb-temperatures" obtained for rocks nos. 3099 (which is contiguous with 3100), and 3117.

$$Ab = \% \ Na_2O \times 8.4_8$$

$$Or = [\% \ K_2O - (\% \ FeO \times 0.4)] \times 5.9_2$$

$$An = [\% \ CaO - (\% \ P_2O_5 \times 1.3)] \times 4.9_7$$

$$Bi = \% \ FeO \times 5.0$$
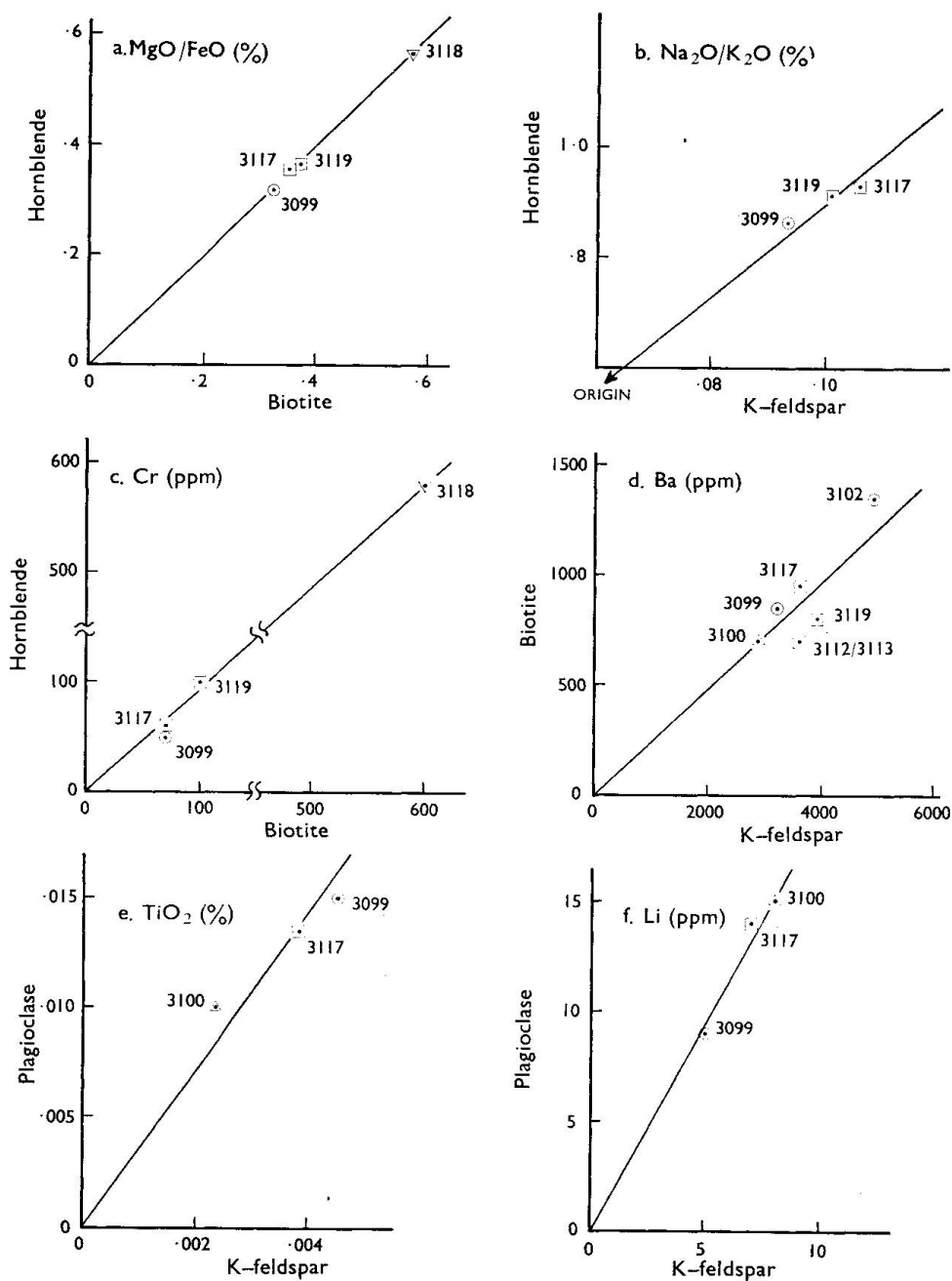
$$Q = 100 - (Ab + Or + An + Bi)$$

Fig. 6. Elements in mineral pairs from various rock types and localities; weight fractions. Symbols as in fig. 1. Straight lines as visual aid.
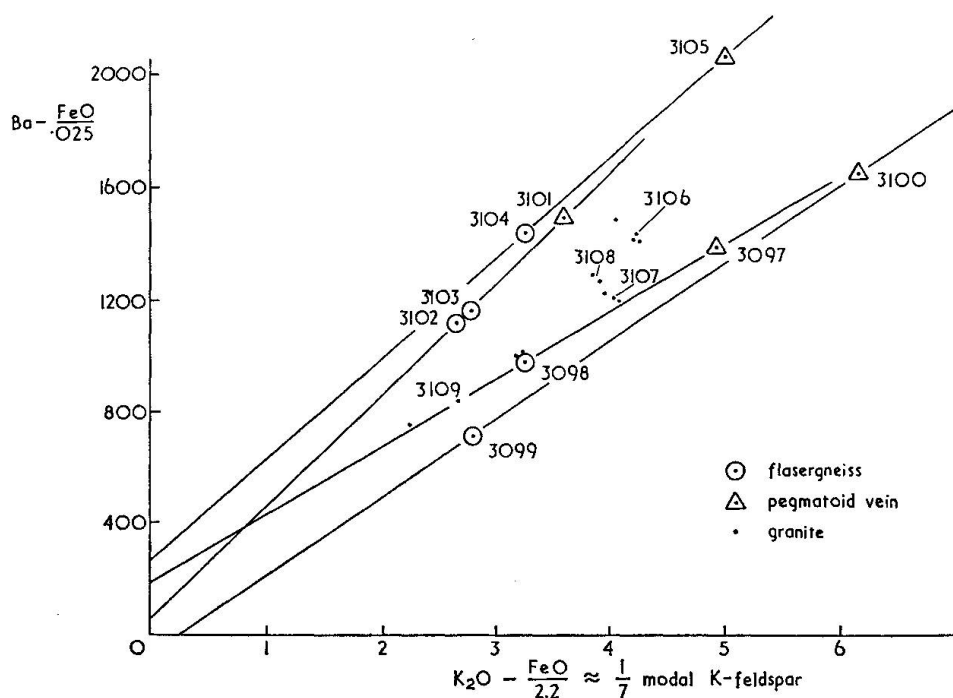
Fig. 7. Bulk rock Ba and K contents, corrected for Ba (Ordinate) and K (Abscissa) bound in biotite. Straight lines connect contiguous vein and gneiss samples and indicate approximately equal Ba/K ratios. For nos. 3101–3105 Ba/K is about 1.5 times that of the other localities. Numbered granites are drill cores from within two meters distance (Table 1). K_2O and FeO in percent, Ba in ppm, both by weight.
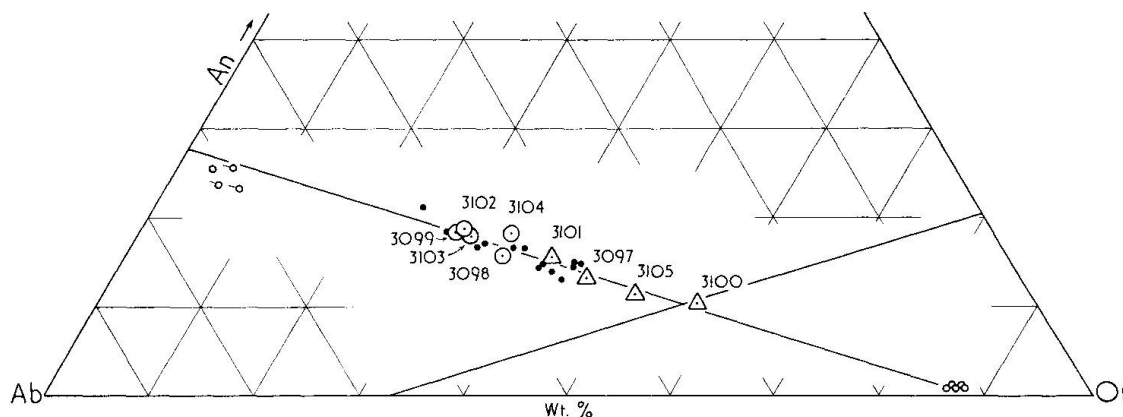
Fig. 8. Bulk rock compositions projected from Q of the Ab-Or-An-Q tetrahedron; weight norm according to text. Symbols as in fig. 7; small circles for analyzed feldspars; for plagioclase errors due to K-feldspar impurities are indicated. Cotectic line from fig. 11.

and are plotted in fig. 8. Considering that about 1% hornblende contained in flasergneiss and granites has been neglected in the calculations, the straight line in fig. 8 does not differ significantly from the tie-line between analyzed feldspars. The straight line configuration limits a possible systematic difference of feldspar compositions between rock types to $\lesssim 3\%$ Ab. Smaller lateral variations are suggested by the mineral analyses (Fig. 6 b).

Fig. 9. Fe^{2+}, Fe^{3+}, and Mg in hornblende-biotite pairs. Dashed curves are estimates for buffered pure Fe-Mg biotites from WONES and EUGSTER (1965), after recalculation to weight percent.

It is apparent from fig. 7 that Ba in K-feldspar of granite does not vary much, whereas the modal K-feldspar of granites does within small areas. All chemical data support the view that the granites are identical in composition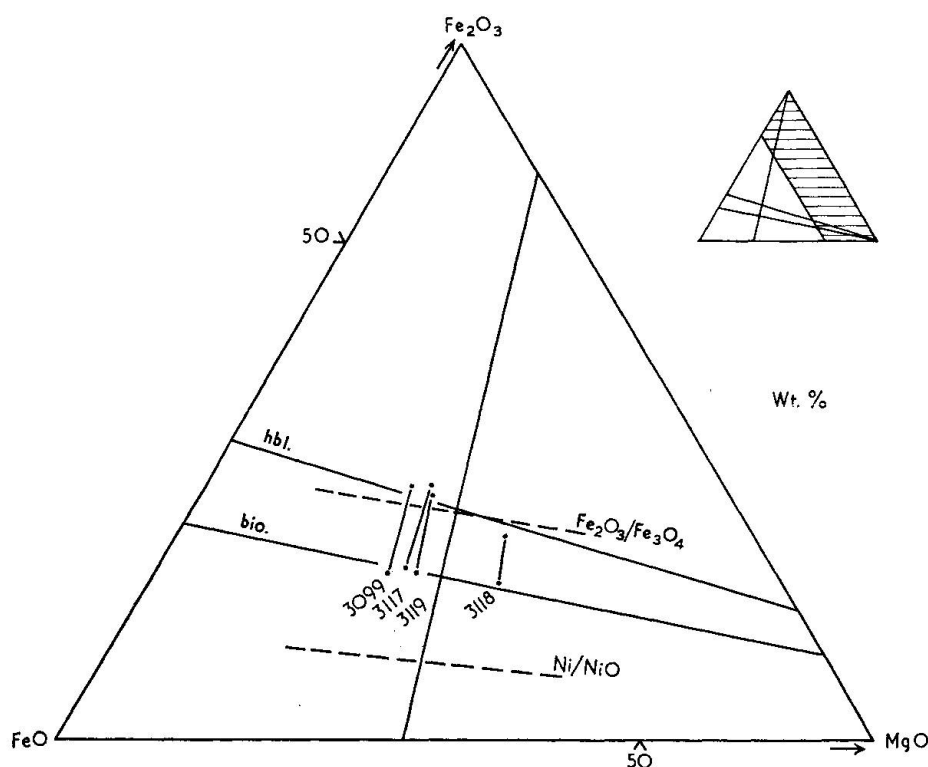 with a mixture of the three other rock types, usually of about two parts of flasergneiss and one part of pegmatoid veins.

Volatiles. The data of fig. 9 suggest that the oxygen fugacity has been the same for granitoid rocks and the amphibolite. This probably means that if the amphibolite has originated as an intrusive dike rock, it has later exchanged oxygen with its surroundings. Oxygen isotopic compositions of hornblendes (Table 4) are compatible with this view. Overlying the migmatite by about 100 m a metaconglomerate horizon occurs. In order to place a limit on the effective range of oxygen exchange during metamorphism, acid plagioclase from the metaconglomerate matrix was analysed for $\delta\,O^{18}$ and compared to the K-feldspar from the migmatite. For isotopic equilibrium the plagioclase oxygen should have been somewhat lighter than that of the K-feldspar (O'NEIL and TAYLOR, 1967). However, possibly in accordance with its metasedimentary origin, the plagioclase was found to be heavier by $2.2^{0}/_{00}$ $\delta\,O^{18}$ than the K-feldspar.

F and H_2O determinations (Table 4) suggest measurable lateral variations of the ratio F/H_2O.

Table 4. *Mineral analyses*

	3099 h[1]	3099 b[1]	3099 p	3099 k	3100 b	3100 p	3100 k	3102 b	3102 k	3112/3113 b	3112/3113 k	3117 h[1]	3117 b[1]	3117 p	3117 k	3118 h[1]	3118 b[1]	3118 p	3119 h	3119 b	3119 k	G-1[2]	W-1[3]	CAAS-[4] Syenite / G-2[5]
(percent)																								
Fe_2O_3	7.8_5	5.5_2			6.7_9			4.3_7		5.6_6		7.7_4	5.7_5			5.9_0	5.4_8		7.4_4	5.4_2				
FeO	17.4_6	20.9_0			19.9_5			20.4_0		19.4_2		16.6_5	20.5_5			14.8_5	17.1_5		16.6_7	19.7_6				
MgO	5.53	6.79			9.75			7.80		8.00		5.97	7.32			8.35	9.75		6.05	7.41				
CaO	11.0_5	0.15	4.8_6	0.14	0.20	4.7_9	0.14	9.0_2	0.17	8.9_6	0.11	11.1_5	0.21	5.1_6	0.14	11.5_0	0.92	5.2_5	11.1_0	0.20	0.14			2.60
K_2O	1.51	9.0_4	0.81[2]	14.5_6	8.3_1	1.19[2]	14.3_8	8.4_0	14.2_5		14.2_5	1.66	8.7_1	0.86[2]	13.9_6	1.42	8.1_6	0.56	1.68	8.8_6	14.2_1			3.50
Na_2O	1.30	0.08	8.40	1.36	0.10	8.2_6	1.44	0.08	1.44	0.06	1.36	1.54	0.07	8.1_0	1.48	1.38	0.16	8.4_6	1.54	0.07	1.44			
TiO_2	1.44	3.89	0.0150	0.0045	3.94	0.0100	0.0023	3.82	0.0047	3.30	0.0027	1.47	3.74	0.0135	0.0038	1.36	2.91	0.0285	1.47	3.65	0.0050	0.25	1.07	0.48
MnO (0.005)	0.65	0.39			0.39			0.39				0.58	0.39			0.37	0.32		0.65	0.39		0.027	0.176	0.40
F	0.20	0.36										0.19	0.32			0.25								
H_2O^+	1.2_1	2.1_4										1.4	3.1			1.5_1	3.7_4							
(ppm)																								
Li	10	230	9.0	5.0	255	15	8.0	200	8.0	250	5.0	10.0		14.0	7.0	8.0	180		9.0	195	3.0			170
Rb		79_5		320				69_5	270				78_5		340									
Cs		27						21	6_5						6_3									
Sr (20)			39_5	44_8		37_5	42_5		61_4					41_5	44_0			36_0			50_5	250	200	
Ba (15)	42	80_6	200	320_5	70_5	310	290_6	135_5	490_9	70_5	360_0	45	95_0	220	360_8	40	70_5	98	45	80_5	390_5	1250	200	35 / 15
Sc (15)	17_5	15										22_5	35			17_9	25					3		70
Y (30)	36_0											31_5				36_0						3		70
Yb (5)	35											30				35								
V	23_0	22_0			19_5			29_0		23_0		24_0	22_0			31_0	28_5		31_0	24_5			24_5	80
Cr (40)	50	70			50			80		50		60	70			58_0	60_5		10_5	50			13_5	45
Co (40)		50			50			50		50						100	115		50				80	
Ni (50)																20_5	33_5						80	
Cu (10)		14			14			14		14			20			15	100			14			110	24
(permil deviation)																								
$\delta\,O^{18}_{SMOW}$		3.8 ± 0.3				2.5 ± 0.1	6.55 ± 0.1					4.3 ± 0.3	2.7 ± 0.1		6.75 ± 0.1	4.1 ± 0.3								7.7 ± 0.3

Letters h, b, p, k, for hornblende, biotite, plagioclase, K-feldspar.
Dash (—) means not found at limit given in brackets.
No symbol means no determination. Analysts: Mg, Li, S. Abbey; Na, K, H_2O^+, J. L. Bouvier; Fe, F, Ti, Mn, $\delta\,O^{18}$, and trace elements except Li, Rb, Cs, P. Blattner; Rb, Cs (table 4), R. L. Goguel and P. Blattner. Also "rapid" determinations by staff, Analytical Chemistry Section, Geological Survey of Canada.
Footnotes: [1] Mg and Ca in biotites and Ca in hornblendes determined in main portion by Bouvier and Blattner. *Full analysis and formula available from author.* [2] Plagioclase-quartz concentrates contain up to two percent K-feldspar. [3] FLEISCHER (1969). [4] SINE et al. (1969). [5] FLANAGAN (1969).

Table 5. *Bulk rock analyses*

	3098 f	3099 f	3102 f	3104 f	3103 f/p	3097 p	3100 p	3101 p	3105 p	3106 g	3107 g	3108 g	3109 g	3110 g	3111 g	3112 g	3113 g	3114 g	3115 g	3116 g	3117 g	3118 a	3119 g	G-1	W-1	CAAS-Syenite
(percent)																										
Fe_2O_3	0.6	0.58					0.08							0.5	0.5	0.5					0.9	4.5_2	1.1			
FeO	1.9	1.78	2.1	1.9	1.9	0.30	0.29	0.20	0.20	1.4	1.2	1.5	1.6	0.8	1.0			0.9	1.2	1.4	1.1	10.8_5	2.3			
MgO			0.61	0.81			0.09	0.11								0.44	0.50			0.48		6.20	0.90			
CaO	1.8	2.2	2.1	2.1	2.0	1.7	1.3	1.6	1.3	1.8	1.9	1.8	2.2	2.1	2.1	2.2	2.2	1.8	1.8	1.8	1.6	8.8	2.5			2.60
K_2O	4.1	3.60	3.6	4.1	3.6	5.1	6.30	3.7	5.1	4.9	4.6	4.6	3.4	3.6	3.7	4.5	4.5	4.7	4.8	4.7	4.60	1.80	3.3			3.50
Na_2O	3.1	3.50	3.2	3.0	3.1	3.2	2.40	2.7	2.6	3.2	3.2	3.2	3.5	3.5	3.4	3.4	3.5	2.9	2.8	2.8	3.00	2.70	3.5			
TiO_2	0.40	0.37	0.37	0.34	0.38	0.06	0.05	0.03	0.02	0.21	0.15	0.16	0.26	0.18	0.22	0.22	0.24	0.15	0.20	0.22	0.25	1.17	0.41			
MnO	0.042	0.045	0.049	0.047	0.046	0.007	0.009	0.006	0.004	0.030	0.030	0.027	0.043	0.025	0.026	0.031	0.029	0.025	0.028	0.028	0.040	0.260	0.076			
P_2O_5	0.070	0.060	0.080	0.070	0.070	0.020	<0.02	0.020	<0.02	0.080	0.060	0.070	0.080	0.050	0.050	0.090	0.090	0.060	0.070	0.070	0.070	0.040	0.110			
(ppm)																										
Li		30	33			16	13								24		26				24	23	32	22_5		
Rb	22_5	19_5	16_0	15_0	16_0	90				16_0				12_5				16_0			16_5	35	15_0	22_5	218	
Sr	320	280	340	42_5	360	385	375	370	43_0	325	290	290	300	325	310	360	340	330	310	330	275	70	290	124_5	285	
Ba	106_5	78_5	121_5	152_5	124_5	140_5	167_5	151_5	208_5	150_5	126_5	133_5	90_5	103_5	102_5	127_5	135_5	145_5	147_5	155_5	124_5	100	84_5	124_5	185	

For analysts and comments see table 4. Letters f, p, g, and a stand for flasergneiss, pegmatoid vein, granite, and amphibolite.

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GENESIS OF MIGMATITE

On structural grounds alone the period corresponding to the temperature estimates may have been close to that of migmatization. High temperature mineral assemblages could not have survived a later deformation intense enough to form the present fabric; on the other hand, entirely post-deformational heating to above 550° C is unlikely. That melts have played a part is then probable on the basis of experiments summarized in figs. 10 and 11, from YODER et al. (1957), TUTTLE and BOWEN (1958), BURNHAM and JAHNS (1962), LUTH et al. (1964), and KLEEMAN (1965).

Conditions for fusion. In view of the present analyses for fluorine, H_2O will be considered the only volatile component that has affected fusion temperatures (MUNOZ and EUGSTER, 1969; WYLLIE and TUTTLE, 1961; PLATEN, 1965). Curve 2 in fig. 10 is the estimated solidus (KLEEMAN, 1965) for the present rocks, excepting only the amphibolite. Agreement is good with the experimental curve by PIWINSKII (1968), for granodiorites with most sodic plagioclase compositions (rims) of An 38 and An 28.

The regime of volatiles will critically affect any scheme of migmatization that involves fusion. P_{H_2O}-T-P_{total} relations have been summarized in fig. 10. For the present purpose planes of equal H_2O content of melts ("isohydrons") have been drawn through the $P_{H_2O} < P_t$ volume (Fig. 10, inset)[2], regardless of the fact that melts to be considered vary somewhat in their anhydrous compositions. Fusion surfaces for the respective compositions intersect the $P_{H_2O} = P_t$ plane in the solid curves 1.–4. The isohydron traces II of fig. 10 may be visualized as intersections of the subhorizontal isohydron planes with a given solidus surface. They have been projected parallel to the P_{H_2O} axis onto the $P_{H_2O} = P_t$ plane of Fig. 10 and are valid only for the specific anhydrous composition on whose fusion curve they originate.

Idealized regimes for fusion may be considered, differing in the way in which the partial pressure of volatiles is controlled (i. e., externally or internally), and in the relationship of P_t and $P_{vol.}$ (presence or absence of a gas phase). For example, in a closed system with a gas phase initially present, a course of fusion could be as follows: For $P = 5$ kb and present rock compositions small initial melts correspond to a point A in fig. 10 and are H_2O saturated. For equilibrium and a bulk H_2O concentration of 5%, feldspars could be completely molten in the presence of quartz at 790° C and a hypothetical $P_{H_2O} = P_t = 1.4$ kb (solid arrows and point Z, fig. 10). The melt would then be only just saturated with H_2O at zero pore volume in the crystalline residue. In reality (broken arrows and Z'), excess solid pressure would develop,

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[2] These are in fact surfaces, concave upward and sloping toward NW in inset. For recent data on the $P_{H_2O} < P_t$ volume of a related system, see MILLHOLLEN (1971).
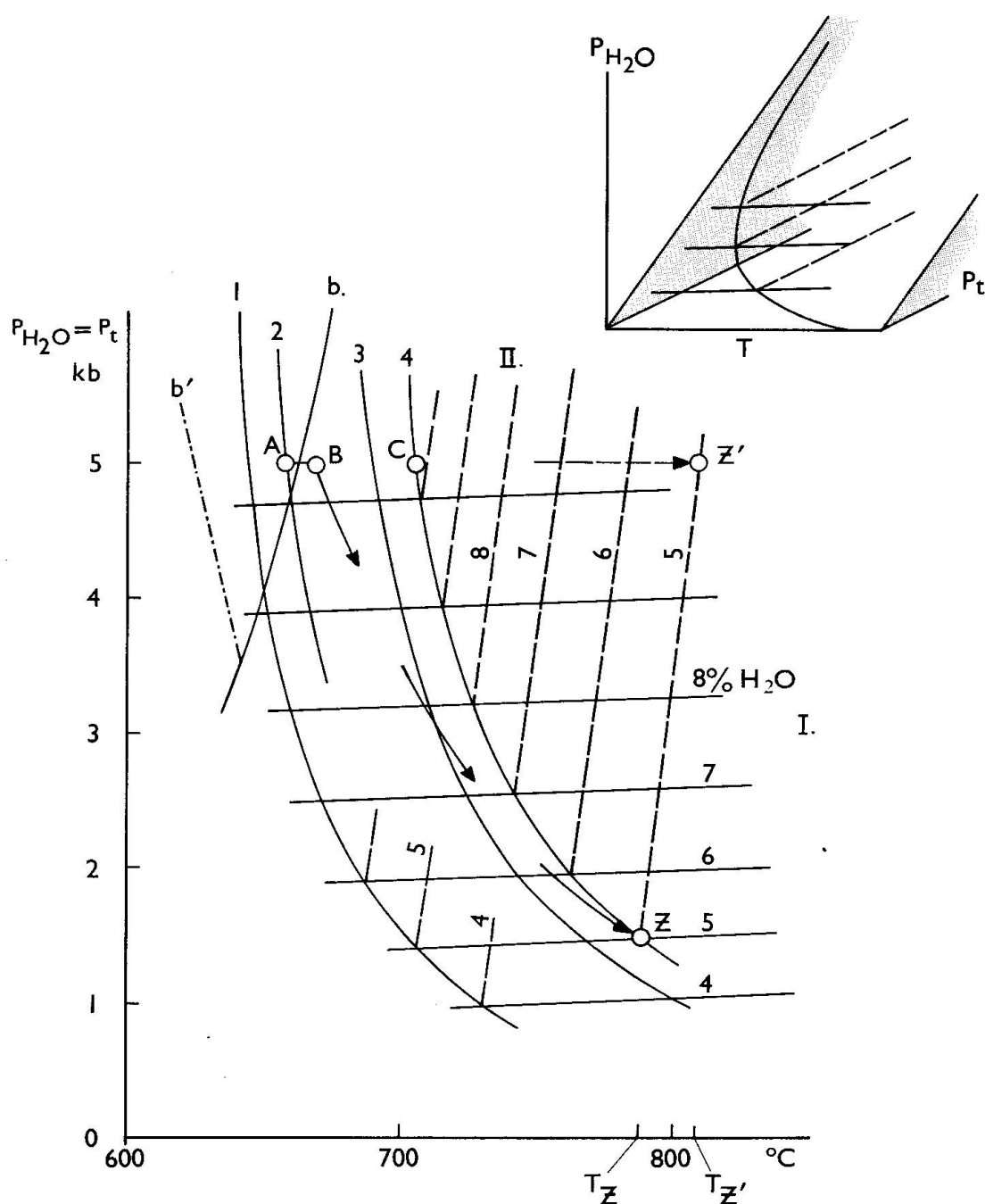
Fig. 10. Melting relations for $P_{H_2O} = P_{vol.} \leqq P_t$. Solid curves for $P_{H_2O} = P_t$. Curve 1: Ab-Or-Q minimum, 2: Solidus for present average feldspar compositions with quartz, 3: Ab-Q eutectic, 4: Estimated liquidus for present feldspars and saturation with quartz. Sources in text, p. 168. Isohydron planes are marked in inset by solid and dashed traces. Dashed traces have been projected // P_{H_2O} onto the $P_{H_2O} = P_t$ plane (isohydron traces II). b and b′ mark breakdown curves for biotite (Text p. 165 and 172). See text for A, B, C, etc.
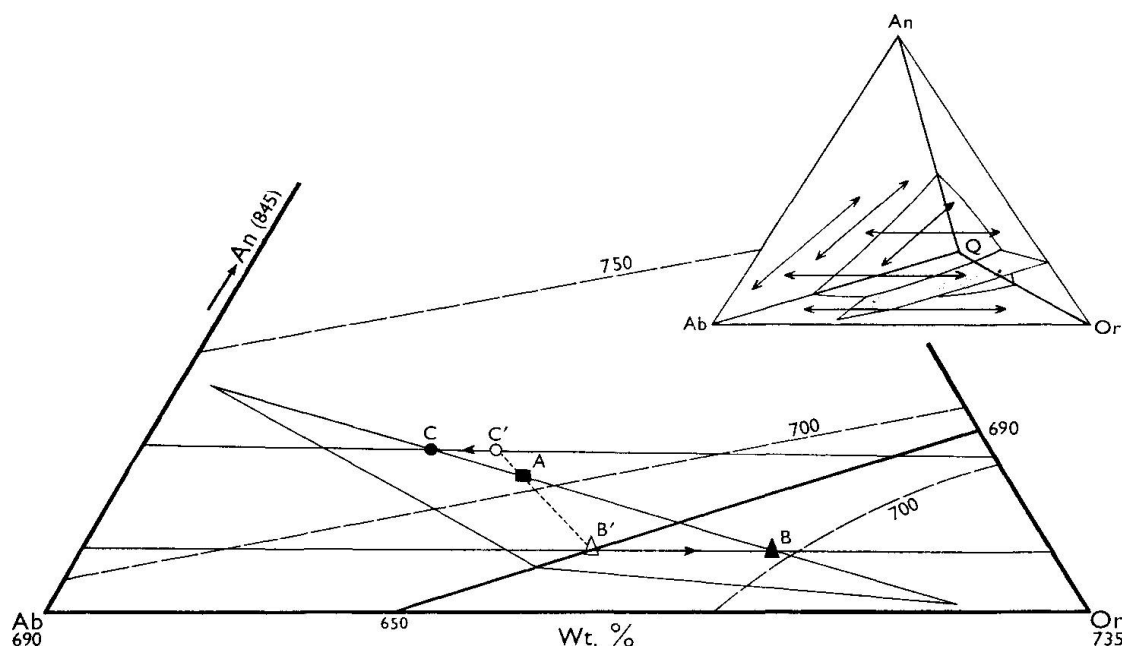
Fig. 11. Cotectic line and liquidus temperatures (°C) for quartz saturated surface at 5 kb P_{H_2O} = P_t projected from Q of Ab-Or-An-Q tetrahedron (KLEEMAN, 1965). Three phase triangle for present feldspar compositions. Near horizontals (horizontals for mol percent) are exchange paths for Na^+ ⇌ K^+ after ORVILLE (1963). See text for A, B, C etc.
Inset shows cotectic surfaces (from WINKLER, 1967) with exchange paths superimposed: for Na^+ ⇌ K^+ // Ab-Or-An face, for Ca^{++} ⇌ $2 Na^+$ // Ab-Q-An face.

diverting the liquidus to isohydron traces II and raising the liquidus temperature to $T_{Z'}$, for 5 kb P_t and 5% H_2O. The gas would be used up and the melt first undersaturated at a point such as B. The melting interval in this case is ∼150° C and it would be even larger if less H_2O was present. In the presence of excess H_2O, of course, the liquidus would be reached at C of fig. 10, the melting interval being as short as 60° C.

Pegmatoid veins. Compositions of flasergneiss and pegmatoid veins suggest that a solidus-liquidus relationship may once have existed between the two rock types, although their feldspars are now identical in composition. In the process of crystallization hydrous quartzo-feldspathic melts will release dissolved volatiles. If at the time of crystallization of pegmatoid melts enough K-feldspar was left in the crystalline residues, bulk compositions of veins and residues may have been shifted along horizontals in the molar Ab-Or-An projection, until, at subsolidus equilibrium, they both lay on one two-feldspar tie-line (Fig. 11). Similar Na^+ ⇌ K^+ transfer equilibration has been obtained experimentally by ORVILLE (1963) and there is no reason to doubt the presence of Cl in addition to F in metamorphic and plutonic rocks generally (e. g. WHITE, 1966; DODGE et al., 1968). If melt compositions had left the cotectic line, subsolidus equilibration between veins and residues consisting of plagioclase and quartz only, apart from mafic minerals, could be obtained by transfer

of Ca^{2+} and Na^+, and local precipitation and dissolution of quartz as dictated by feldspar stoichiometry (Arrows in fig. 11, inset).

If transfer equilibration is allowed for, possible compositions of former melts and residues may be reconstructed for any fixed weighted average composition of the two. Clearly the actual rock samples may represent impure melt or impure crystalline residue, and extreme or extrapolated compositions should be used in such a reconstruction. For example, if point A in fig. 11 represents the average composition, B that of veins, and C that of residues, we obtain, on the basis of $Na^+ \gtrless K^+$ transfer, B' for melts and C' for original residues.

Granites. The granites are the most transgressive rock type structurally. However, in most instances, intrusion from further than several meters seems impossible, and in this case (as against that of pegmatoid veins) the lack of channels cannot be ascribed to later structural transposition. On structural evidence, too, granites may therefore be expected to represent a mixture of pre-existing wall rocks, formed in the late shear zones S_2. A possible mechanism may be outlined following OROWAN (1960). At high temperatures and low deformation rates, creep will concentrate in sections of a body, rather than affect it uniformly. A type of faulting takes place which does not involve fracturing, but generates frictional heat continuously. Orowan points out that owing to pore formation the strained zones may take up volatiles from the surroundings. Also, plastic strain will increase the free energy of solids, but not of liquids, and will thereby depress fusion temperatures. Thus, there could be several reasons why rocks of the same composition should melt in shear zones, but not outside.

Amphibolites have been at least partly preserved as inclusions in the granites (Fig. 3d). Their solidus temperature at constant $P_{vol.}$ lies more than $80°\,C$ above that of the granitoid rocks (Fig. 11). In that interval amphibolite inclusions could have been protected by the surrounding liquid from the effects of shearing.

DISCUSSION

On structural grounds the pegmatoid veins could have been massive intrusions only if emplaced early in the deformation sequence S_1–S_2. Subsequent to such intrusion, transfer equilibration similar to that proposed would have been required to establish mineralogical control of vein and gneiss compositions. The granites are of local origin in the same sense, and in addition they cannot have been massive intrusions in any case. Discussing the process of migmatization we will therefore assume a pre-existing rock mass of about the present composition, containing basic rocks to be transformed into amphibolite.

Whereas the one-time presence of melts in the migmatite seems certain, the occurrence of wholesale melting at any given time may be excluded-for the very reason for which the term "migmatite" was applied by SEDERHOLM: rocks differing in composition or fabric, or both, are associated on a mesoscopic scale. In the presence of excess volatiles failure of wholesale melting to occur could have been due to a temperature peak inside the then narrow melting interval for the granitoid rocks. GREENWOOD (1961) has discussed the possibility of externally controlled P_{H_2O} at $P_{vol.} < P_t$. At constant $P_{vol.} < P_t$ the melting interval, although shifted to higher temperatures, would be of about the same width (Fig. 10). On the other hand, appreciable fusion at constant $P_{vol.}$ would mean a large turnover of gas, and shortage of volatiles, with the melting interval extended in consequence, seems a more obvious reason for fusion to have remained partial. For reactions involving volatiles, it is even tempting to consider the migmatite as an effectively closed system, with all volatiles originally bound in hydrous minerals.

Had $P_{vol.}$ originally been smaller than P_t, the breakdown of hydrous minerals would have begun on a curve such as b' in fig. 10 (YODER, 1955). For the same combination of P_t and $P_{vol.}$ melting could not have begun until the temperature reached the corresponding isohydron trace II. However, in a closed system, the evolution of volatiles from hydrous minerals would have first served to raise $P_{vol.}$, and for total pressures above the intersection of curves b and 2 of fig. 10, volatiles could have become available to melts before $P_{vol.}$ had reached P_t. LUNDGREN (1966) has argued a similar case for muscovite-bearing rocks.

For structural reasons, the granites must have formed after the pegmatoid veins, and the difference between these two features may be ascribed to changes with time in physical conditions and deformation pattern. For example, pegmatoid melts may have formed and accumulated during a temperature increase when deformation was slight, whereas when the shears S_2 started to form, temperatures may already have been inside or near the melting interval. There are two ways in which the same batch of volatiles may have been directly passed from the pegmatoid to the granite melts. Crystallization of the pegmatoid melts may have been due to a regional decrease in temperature, and the associated change in rheological properties of the rock may have induced shearing and formation of the granites. Alternatively, loss of volatiles to the newly-forming shear zones may have caused adiabatic crystallization of existing pegmatoid melts. Either sequence could lead to models of zone melting induced by volatiles.

The isotopic analyses of feldspars from the migmatite and from nearby metasediments (p. 167/168) support the view that volatiles have been in short supply during metamorphism. The rocks contain up to 10% modal biotite and minor hornblende. However, even at 800° C (above the intersection of curve 2, fig. 10, with the 4% isohydron), less than 10% of the migmatite at a

time could have existed as melts, with $4\%\ H_2O$, whereas the visually estimated volume of pegmatoid veins is at least 25%. Although the problem is avoided easily enough by assuming an additional supply of volatiles, it may alternatively point to formation of the pegmatoid veins themselves in more than one stage.

Transfer equilibration, at whatever stages it may have taken place, should have affected amphibolites as well as the granitoid rocks, within the limits of active mechanisms. Unless we take it as incidental, the presence of the same plagioclase in the amphibolite and in the granitoid rocks indicates that $Ca^{2+} \rightleftarrows 2\ Na^+$ has been active as well as $Na^+ \rightleftarrows K^+$. Exchanges of Sr and Ba might then also be reasonably expected. Nevertheless, few other elements seem to have been fully exchanged between minerals of the amphibolite and those of the surrounding rocks. Differences in Ti and Li between flasergneiss and pegmatoid veins (Fig. 6e, 6f) may indicate a fractionation of these elements in partial fusion, and lack of a sufficiently active exchange mechanism afterwards.

Whereas transfer equilibration is considered to have established uniform mineral compositions with regard to Na, K, Ca, etc. on a scale of about 0.5 m, it seems difficult to propose an explanation (such as a steady temperature gradient) for the concentration of Ba and Sr in the area of samples nos. 3101–3105 at the same time. This variation and some other minor ones could be remnants from the time before migmatization.

CONCLUSION

Only a framework for the genesis of the present migmatite has been outlined. A more detailed picture will require more specific data, and a comparison of several morphologically similar migmatite occurrences seems desirable. Mineral textures, if they date from a period after the consolidation of melts, may tell only little about the process of migmatization.

The veined gneiss described by WHITE (1966) could well fit a partial fusion/ $Na^+ \rightleftarrows K^+$ transfer model, and it may be unnecessary in that case to invoke an unspecified process of metamorphic differentiation. For some compositions in the plagioclase volume of the Ab-Or-An-Q tetrahedron, both $Na^+ \rightleftarrows K^+$ and $Ca^{2+} \rightleftarrows 2\ Na^+$ transfer may fail to completely obliterate solidus-liquidus relationships. Such partly preserved relationships in nature could therefore considerably strengthen hypotheses involving reciprocal transfer.

The critical feature of migmatites is the interposition of different rock types on a mesoscopic scale. Whether metamorphism has altered the bulk composition of the present rock complex, seems a question only indirectly related to the origin of its mixed-rock character.

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