

CONTRIBUTIONS TO THE MINERALOGY OF NORWAY

No. 5. Trace element variations in three generations of feldspars from the Landsverk I pegmatite, Evje, southern Norway.

BY

S. R. TAYLOR¹, K. S. HEIER² and T. L. SVERDRUP³

Abstract: Li, Na, K, Rb, Cs, Pb, Tl, Ca, Sr and Ba have been determined in potassium feldspars and coexisting albites of three different generations from one large pegmatite (Landsverk I, Evje, southern Norway). In addition Fe, Cr, Mn, Cu, Co, Ni, V, Sn and F were determined in a pink microcline and a green amazonite which were seen to grade into each other.

Some element concentrations and ratios reflect the different conditions of formation of the feldspars from the three generations. The data indicate that the amazonite formed through metasomatic action of a "pegmatitic rest liquid" (from which cleavelandite crystallised) on the pink microcline of the first generation. The trace element data also indicate that the youngest K- feldspar is not a late fractionation product of the pegmatite, but rather a late secondary mineralisation.

The cause of the colouring of the amazonite is discussed and it is concluded that it is not to be found in a significant difference in composition from the pink microcline for any of the elements determined.

Introduction.

Two of the authors (K. S. H. and S. R. T.) have made previously an extensive study of the distribution of trace elements in K-feldspars from a variety of pre-Cambrian basement rocks from southern Norway (TAYLOR and HEIER; 1958 a, b, 1960; HEIER and TAYLOR,

¹ Dept. of Geochemistry, University of Cape Town.

² Mineralogisk Museum, Oslo.

³ Norges geologiske undersøkelse, Oslo.

1959 a, b). Rock types studied included gneisses, augengneisses, anatectic granites, diapire granites, and small and large pegmatites. Extreme fractionation trends were discovered in the trace elements of the feldspars from large pegmatites. The present investigation of feldspars of different generations was undertaken to see whether these same trends would be useful in elucidating the history of a single large pegmatite, and if it would be possible to distinguish between feldspars of different generations by the trace element data.

Another aim of this study was to investigate the relations between the trace element contents of the different generations to establish whether they were genetically related in a single fractionation sequence, or whether the later generations were derived from replacement, either by solutions from within the pegmatite, related to earlier feldspars, or from a separate mineralisation. By this approach it was hoped to shed some light on the complexities of mineral paragenesis in pegmatites as discussed by many authors; for a selected bibliography and discussion, see JAHNS (1955).

Location and geological description.

The pegmatite Landsverk I, Evje, south Norway, is one of the numerous pegmatites occurring in the pre-Cambrian basement rocks of southern Norway (Fig. 1). It is located close to the road about 500 metres NNW of Landsverk. The pegmatite is described by ANDERSEN (1926, 1932) and BJØRLYKKE (1935). According to ANDERSEN the pegmatite has produced 20,000 tons of feldspar and more than 1000 tons of quartz. Mining was discontinued in 1928 but commenced again in 1958. The pegmatite occurs in a dark gabbroic (amphibolitic) rock. It strikes EW and the contacts toward the enclosing rock are irregular with apparent vertical dip. Part of the pegmatite is brecciated, the breccia consisting of fragments of pegmatite and the surrounding gabbro, cemented with quartz. The pegmatite extends for about 120 metres along the strike. It appears to pinch out towards the east and disappears under overburden towards the west. It has been mined for about 100 metres along the strike to a width of 10 to 20 metres, and in places to a depth of 10 metres. The feldspars examined by us are mainly taken from the dumps, but their position in the pegmatite can be deduced.

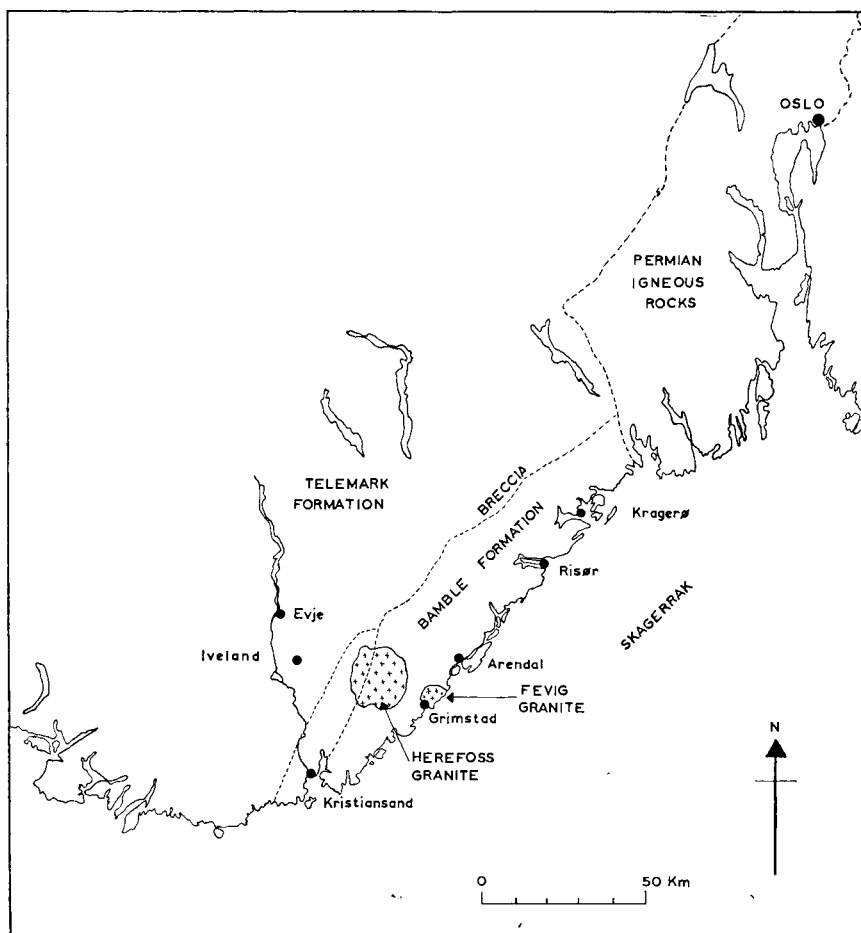


Fig. 1. Map showing the location of Evje.

The common minerals in the pegmatite are: quartz, potash feldspar, (often as large crystals several metres long), plagioclase (in small amounts compared to potash feldspars), biotite and muscovite. Less common minerals are: monazite, fergusonite and euxenite which occur with the mica. ANDERSEN also records the presence of uraninite, orthite, epidote, garnet, magnetite, calcite and the zeolite mineral leumontite.

The feldspar minerals.

Alkali feldspars of different colours are characteristic of the pegmatite, and it was the recognition of these which started the present study. Albite occurs with all the different coloured potash feldspars. On the basis of field evidence and thin section studies, the following sequence of feldspar generations has been established.

Generation 1. (G2) Pink microcline (10)¹ which is the common pegmatite feldspar. The other feldspars are rare by comparison. White plagioclase (16) occurs in association.

Generation 2. (G1) Amazonite (12). This feldspar has the characteristic green amazonite colour. A transition between the first generation microcline and the amazonite (Plate 2, Fig. 1) was sampled and analysed (Nos. 11.1–11.6). Cleavelandite occurs typically with the amazonite (17, 18).

Generation 3. (G3) "Brick red" microcline (14) occurs in the brecciated part of the pegmatite as:

- a. coating on the surface of the other feldspars.
- b. fillings along tiny cracks and fissures in the other feldspars.
- c. replacing the albite lamellae of the other feldspars (13, 15).

A sugar grained albite (19, 20) often occurs intermixed with the G3 microcline and is assumed to be of the same age.

Potash feldspars of all three generations are microclines. Triclinicity values as indicated by X-ray powder methods (GOLDSMITH and LAVES, 1954) are approximately 0.80 for G1 and G2 feldspars. For the G3 feldspar the triclinicity value is as high as 1.10, indicating a much more ordered structure for the latter.

Samples 11.1 to 11.6 are different transitional stages between G1 and G2 feldspars. The samples were drilled out at 1 cm intervals from the same handspecimen (Plate 2, Fig. 1). The size of the albite lamellae is fairly large compared to the diameter of the drill holes so that the minor chemical differences between the six samples cannot be claimed to give the correct picture of the amazonite-microcline transition. However, From Fig. 2 it can be seen that these feldspars constitute a Na rich group of alkali feldspars. The pure amazonite (12) is high in Na. The microcline (10) plots in the albite poor end. Thus,

¹ Sample numbers in Tables 1 to 4.

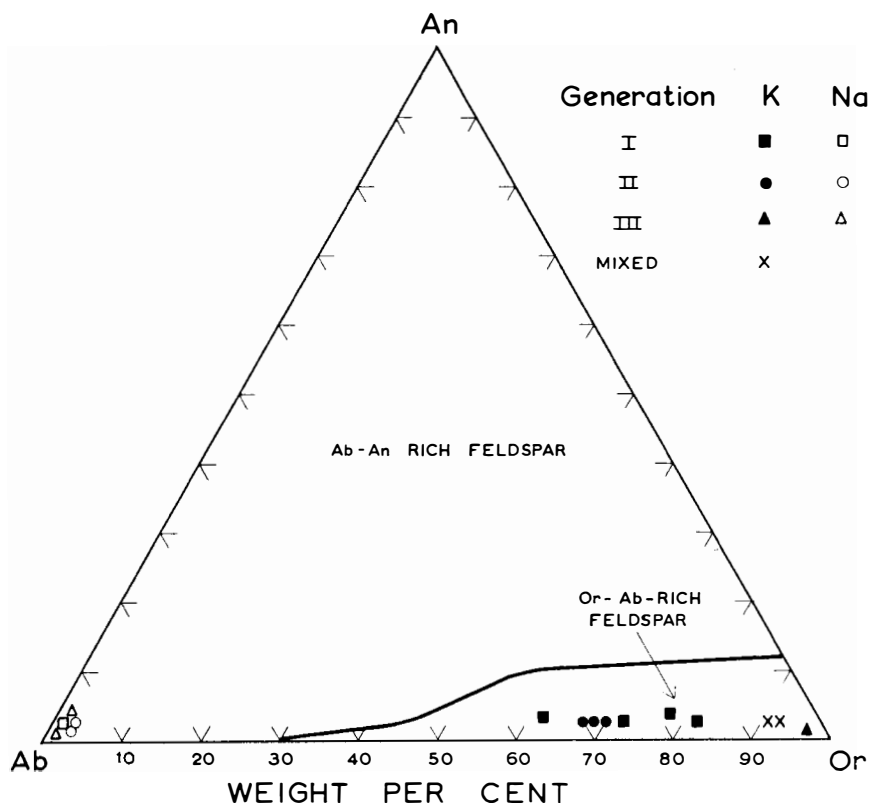


Fig. 2. Feldspar compositions from Table 3 plotted on the projection of the quaternary system Or-Ab-An-H₂O at 50 00 bars H₂O pressure (Yoder et al, 1957).

the amazonites are not lower in Na than the G1 microcline which would be expected if they crystallized of a pegmatite melt or vapour at lower temperatures than the G1 microcline.

OFTEDAL (1957) on the basis of heating experiments with amazonite, concluded that the "colour centra" in amazonite formed below 300° C, probably at about 250° C in already crystallized feldspar, and that this is in accordance with field observations which indicate that the green colour originated in connection with the formation of later cleavelandite veins. This association between amazonite and cleavelandite is always present in Norwegian pegmatites of which the authors have firsthand knowledge.

The very low Na content of the G3 microcline (Table 1, Fig. 2) shows this feldspar to have formed at very low temperatures (it occurs together with albite). This feldspar, besides occurring on cracks and as a coating on the other feldspars, is typically found to replace the albite lamellae in the other K- feldspars. Thus the albite associated with it may be recrystallised material of the albite lamellae in the older perthites.

In summary the following sequence of feldspar formation is used as a basis for the discussion of trace element distribution.

1. Crystallisation of microcline and albite of generation 1 (G1).
2. Crystallisation of the "pegmatitic rest liquid" to form cleavelandite, and metasomatic alteration of G1 microcline into amazonite (G2). An albitisation of the G1 plagioclase may have taken place simultaneously, but is difficult to prove. The G1 plagioclase (albite) contains no more Ca than the cleavelandite (Table 1).
3. Brecciation of a portion of the pegmatite.
4. Crystallisation of the generation 3 (G3) potash feldspar in cracks and replacing albite lamellae in G1 and G2 feldspars. Some albite is assumed to belong to this stage as well.

Many authors, for example SCHALLER (1927) and LANDES (1933) have discussed the role of replacement in pegmatites. JAHNS (1955) gives a summary of the various views. It will be shown later that the third generation feldspars are so unrelated in trace element composition to those of the preceding generations as to make a separate origin necessary. They cannot be considered as a late fractionation product related to the preceding generations, nor may they be derived from reaction of a rest liquid with the earlier formed crystals.

Sample preparation.

Methods of mineral separation and sample preparation were similar to those described previously (TAYLOR and HEIER, 1958 a; HEIER and TAYLOR, 1959 a).

To obtain samples from the specimen showing the transition from pink microcline to green amazonite, a series of drillings (11.1–11.6) 1 cm apart were made (see Plate 2, Fig. 1). These samples were found to be contaminated with Fe, Cu, and Ni, even after magnetic separation. Because of this contamination, three samples (11.7–11.9) for

the determination of these elements were taken in the areas shown on Plate 2, Fig. 1. Maximum precautions were taken to avoid contact with metal. Because of this it was not possible to take a series of small samples similar to the drilled ones, and instead, one sample of the amazonite and two of the microcline were broken off from a sawed section about $\frac{1}{4}$ " thick. A new chisel was used to break off the initial pieces along the cleavages. Edges which came in contact with the chisel were discarded and small cleavage fragments were crushed in a large agate mortar which had been cleared with A. R. HCl and HNO₃. A hand magnet was used at this stage to extract any metal; none was observed. Samples were then ground in a mechanical agate mortar to pass 120 mesh Terylene bolting cloth held in Perspex holders. This powder was run through a Frantz magnetic separator set at 20° forward and 15° back tilt and 1.5 amp. current, but no magnetic fraction was observed to separate. It is believed that adequate precautions were taken to avoid contamination in these samples. That these are necessary is shown by the extreme difficulty of removing metallic contamination from the drilled material (samples 11.1—11.6). Repeated running of these samples through the magnetic separator and treatment with a hand magnet was unable to reduce the Fe contamination below about several hundred ppm. With drilled samples, of course, the metallic contamination is of very small particle size, and the mineral powder tends to adhere to the metallic nuclei.

Analytical methods.

(a) The alkali elements.

Na and K were determined by flame photometer. Li, Rb and Cs were determined by emission spectroscopy using DC arc anode excitation. The method was similar to previously published methods (TAYLOR and HEIER, 1958 a, HEIER and TAYLOR, 1959 a) except that Kodak IL plates were used in place of Kodak IIN. For the plagioclase feldspars, anodes with an internal diameter of $1/8$ " (3.2 mm) and a depth of 5 mm were used to increase sensitivity.

(b) Lead, thallium, and copper.

The method was similar to that described by HEIER and TAYLOR (1959 a) except that Kodak 103-0 plates were used in place of Ilford

N. 30. Slightly larger electrodes 1/8" (3.2 mm) internal diameter and 5 mm deep were used; the optical system comprised a Hilger E.958 lens focussed directly on the slit; K 3447,7 was used as internal standard line for K-feldspars and K 4044 for the plagioclase feldspars.

Samples 11.1 to 11.6 which were drilled from the sample shown in Plate 2, Fig. 1 were found to be contaminated by Fe, Ni and Cu. The nickel contamination was particularly serious for the determination of Tl since interference on Tl 3775.7 was observed by Ni 3775.6 (500 intensity listing in the MIT tables). The control line was Ni 3783.5 (500). This interference was observed in all the drilled samples (particularly in 11.4 and 11.5), but not otherwise. Tl 3519.2 was used to determine Tl in these samples; this line was fortunately resolved from Ni 3519.8 (500) which was also present in samples 11.1 to 11.6. No interference of Pb 4057 by Mn 4057 (80) was observed except for the standard granite G-1 (Ahrens and Fleischer, 1960). The control line was Mn 4058 (80). Critical timing of the exposure is of course necessary to avoid CN interference on Tl 3775 and Pb 4057. Copper was also determined in samples 11.7 to 11.9. Cu 3247 and 3274 were the analytical lines.

(c) *Ca, Sr and Ba.*

These elements were determined in a similar fashion to the method described by HEIER and TAYLOR (1959 b) except that Kodak 103-0 plates were used, and National Carbon Co. grade SP-2 graphite powder was used in place of Johnson-Matthey JM-9.

(d) *Cr, Fe, Mn, V, Co, Ni.*

Since sample 11 shows a transition from green amazonite to pink microcline, the opportunity was taken to see if any significant differences existed in the concentrations of various elements which have been suggested as being responsible for the green colour (see p. 152). Anodes were cut from 3/16" (4.8 mm) National Carbon Co. Special graphite (Cat. No. L.3806) and were 1/8" (3.2 mm) internal diameter by 5 mm deep. Cathodes were 1/8" (3.2 mm) National Carbon Co. special carbon (Cat. No. L.3863). Blank electrodes were arced on the same plates as the samples. The samples were mixed with two parts of the carbon-palladium mix used for the determination of Ca, Sr and Ba. (One part by wt. of Johnson-Matthey "Specpure" $(\text{NH}_3)_4$

$\text{Pd}(\text{NO}_3)_2$ and nine parts of National Carbon Co. SP-2 graphite). The same arcing conditions were used as in the determination of Ca, Sr and Ba except that the optical system was changed to give maximum intensity (Hilger E.958 lens focussed on slit). Pd 4473 was the internal standard line. Fe 4326, Mn 4030 and Cr 4254 were read. V 4379 was not observed in any samples.

A detection limit of less than 1 ppm was obtained. An exposure was also made to cover the range from 2750 to 4650 Å using quartz optics, but Ni 3414 and Co 3453 were not detected in samples 11.7 to 11.9. It is believed that the conditions above were of sufficient sensitivity to have detected any cation present in amounts sufficient to cause the green colouration.

(e) *Fluorine.*

The samples were mixed with one part of Johnson-Matthey "Specpure" CaCO_3 and one part of National Carbon Co. SP-2 graphite powder and arced at 3 amp DC using anode excitation. A Hilger large quartz and glass spectrograph was used with glass optics. Ilford HP3 plates were employed. Lower electrodes were cut from National Carbon Co. Special Graphite Spectroscopic (Cat. No. L3806) 3/16" rod. The crater was 1/8" (3.2 mm) diameter and 2 mm deep. Upper electrodes (cathodes) were 1/8" (3.2 mm) National Carbon Special Carbon rod (Cat. No. L3863). The bandhead CaF 5291 was used. The standard granite G-1 (700 ppm F) and the standard diabase (200 ppm F) (AHRENS and FLEISCHER, 1960) served as standards.

Analytical data.

The concentrations of the elements determined are given in Table 1, and the calculated theoretical feldspar compositions in Table 3. Table 3 shows that Or, Ab, and An, are the only feldspar components present in amounts greater than 1 per cent. The Ba- feldspar is in one case (14) as high as 0.97 per cent, but is typically much less. The Rb- feldspar component is typically next in importance after Or, Ab, and An, ranging between 0.22 and 0.67 per cent. Or, Ab, and An, may be regarded as major elements in the feldspar structure, and their concentrations govern the distribution of the other elements (see HEIER and TAYLOR, 1959 b). The feldspars are plotted on the

Table 1. Analytical data for feldspars from Landsverk I pegmatite. (n d = not detected).

Sample No. (K-feldspars)	Generation	Li ppm	Na %	K %	Rb ppm	Cs ppm	Pb ppm	Tl ppm	Ca %	Sr ppm	Ba ppm
10 (131d)	1	1.1	1.25	11.26	980	n d	300	11	0.35	33	235
11.1 (429.6)	1	1.0	3.15	8.67	960	32	110	24	0.47	140	235
11.2 (429.5)	1	0.8	1.54	10.96	1130	39	135	27	0.50	92	155
11.3 (429.4)	1	0.8	2.17	10.15	1130	37	125	29	0.39	70	105
11.4 (429.3)	2	0.7	2.55	9.55	1110	53	245	73	0.33	34	27
11.5 (429.2)	2	0.6	2.55	9.55	980	50	115	38	0.27	41	28
11.6 (429.1)	2	1.2	2.41	9.78	1370	54	76	33	0.26	36	35
12 (131a)	2	0.6	2.61	9.48	1170	70	115	99	0.38	54	55
13 (137b)	2-3	1.8	0.37	12.97	1210	74	99	65	0.40	105	380
14 (137e)	3	1.9	0.25	13.45	580	n d	1.4	8.9	0.22	43	3550
15 (137a)	1-3	2.4	0.61	12.77	890	n d	115	10	0.28	27	640
(Na-feldspars)											
16 (134b)	1	0.4	8.42	0.23	19	n d	13	1.0	0.51	53	19
17 (131b)	2	0.6	8.52	0.31	92	7.0	43	1.2	0.87	62	49
18 (501)	2	0.5	8.31	0.31	67	6.9	23	1.4	0.21	15	25
19 (133a)	3	0.3	8.31	0.15	7.9	n d	5.9	1.1	0.43	36	16
20 (500)	3	0.4	8.61	0.14	9.5	n d	5.0	0.7	0.16	13	12

Table 2. Sample numbers and descriptions of feldspars from the Landsverk I pegmatite.

Sample no.	Generation	Description of samples
10 (131d)	1	Reddish brown (pink) microcline, parts of which are covered on the surface by G3 microcline. Care was taken to avoid contamination. Coexists with plagioclase (16).
11.1 (429.6)	1	G1 microcline.
11.2 (429.5)	1	
11.3 (429.4)	1	
11.4 (429.3)	2	Fig. 2. Transition between G1 and G2.
11.5 (429.2)	2	
11.6 (429.1)	2	G2 microcline (amazonite).
12 (131a)	2	Typical green amazonite occurring in contact with and in assumed equilibrium with cleavelandite (17).
13 (137b)	2-3	Green amazonite which is partly replaced by G3 microcline. The latter appears to replace preferentially the albite lamellae in the amazonite.

Table 2. (Cont.)

Sample no.	Generation	Description of samples
14 (137e)	3	"Brick red" microcline which occurs as a coating on G1 microcline. This type also occurs penetrating and replacing G1 and G2 microclines (13 and 15). It also occurs intergrown with white "sugar-grained" albite (19).
15 (137a)	1-3	Typical G1 microcline which is partly replaced by G3 microcline.
16 (134b)	1	White plagioclase crystals occurring with G1 microcline (10).
17 (131b)	2	Cleavelandite of very weak bluish colour, in contact with amazonite (12).
18 (501)	2	Light blue "sugar grained" plagioclase, not in contact with microcline, but chemically and physically similar to cleavelandite (17).
19 (133a)	3	White "sugar grained" albite intergrown with G3 microcline (14).
20 (500)	3	Light blue "sugar grained" albite not in contact with microcline but chemically similar to G3 plagioclase (19).

projection of the quaternary system Or., Ab., An., H₂O at 5000 bars water vapour pressure (YODER et al, 1957) in Fig. 2. The potash feldspars show a wide spread in composition from Or. 61 to Or. 96, whereas the coexisting plagioclases have a rather uniform composition of about An. 1 to 5.

Lithium, Tin, Beryllium and Fluorine.

Lithium is present in nearly all the feldspars, but always in very small amounts. It averages 1.2 ppm in the potash feldspars in good agreement with the average of 1 ppm found by HEIER and TAYLOR (1959 a). The plagioclases contain less Li, averaging 0.43 ppm. The G3 microcline and the G1—G3 and G2—G3 mix are higher in Li than the other feldspars.

OFTEDAL (1957) suggested that the new elements introduced into the pegmatite during the formation of the cleavelandite were chiefly F, Li, Be, Sn. No introduction of Li is indicated in our samples. Tin and beryllium were not detected in any samples. The detection limit

Table 3. Theoretical feldspar "molecules" calculated from analytical data.

Sample no. (K-feldspars)	Ab	Or	RbAl Si ₃ O ₈	CsAl Si ₃ O ₈	PbAl ₂ Si ₂ O ₈	TlAl Si ₃ O ₈	An	SrAl ₂ Si ₂ O ₈	BaAl ₂ Si ₂ O ₈	Sum
10 (131d)	14.25	80.14	0.29	—	0.06	0.002	2.43	0.012	0.064	97.25
11.1 (429.6)	35.9	61.70	0.36	0.009	0.024	0.005	3.26	0.052	0.064	101.38
11.2 (429.5)	17.56	78.00	0.43	0.011	0.029	0.006	3.47	0.034	0.042	99.58
11.3 (429.4)	24.74	72.24	0.43	0.010	0.027	0.006	2.71	0.026	0.029	100.22
11.4 (429.3)	29.07	67.97	0.42	0.015	0.053	0.016	2.29	0.013	0.007	99.85
11.5 (429.2)	29.07	67.97	0.37	0.014	0.025	0.008	1.87	0.015	0.008	99.35
11.6 (429.1)	27.47	69.60	0.52	0.015	0.016	0.007	1.80	0.013	0.010	99.45
12 (133a)	29.75	67.47	0.67	0.020	0.025	0.021	2.64	0.021	0.015	100.63
13 (137b)	4.22	92.31	0.46	0.021	0.021	0.014	2.78	0.039	0.10	99.97
14 (137e)	2.85	95.72	0.22	—	—	0.002	1.53	0.016	0.97	101.31
15 (137a)	6.95	90.88	0.59	—	0.025	0.002	1.94	0.010	0.18	100.58
(Na-feldspars) .										
16 (134b)	95.99	1.64	0.01	—	0.003	—	3.54	0.020	0.005	101.21
17 (131b)	97.13	2.21	0.035	0.002	0.009	—	6.04	0.023	0.013	105.46
18 (501)	94.73	2.21	0.025	0.002	0.005	—	1.46	0.006	0.007	98.45
19 (133a)	94.73	1.07	0.003	—	0.001	—	2.98	0.013	0.004	98.80
20 (500)	98.15	0.99	0.004	—	0.001	—	1.11	0.015	0.003	100.26

is difficult to set precisely, but is not greater than 50 ppm. Fluorine was not detected in the amazonite (11.7), nor in accompanying microcline (11.8 and 11.9) (Plate 2, Fig. 3). The detection limit is about 100 ppm.

Rubidium.

The behaviour of Rb in potash feldspars from the pre-Cambrian basement rocks, including pegmatites, from southern Norway is discussed by HEIER and TAYLOR (1959 a). Fig. 3 shows plots of ppm Rb vs. per cent K for the Evje pegmatite feldspars. The first-generation microclines and albites have K/Rb ratios which fall just within the limits of "normal" ratios. (For discussions of the significance of "normal" and "anomalous" or "abnormal" K/Rb ratios, see TAYLOR et al, 1956, TAYLOR and HEIER, 1958 a and b; HEIER and TAYLOR 1959 a). The third generation microcline and albite likewise have "normal" K/Rb ratios. It will be seen readily from Fig. 3 that in each case the K- feldspar and the coexisting albite have similar ratios (within the limits of precision of the data), and that the G3 feldspars contain distinctly less Rb (higher K/Rb ratios) than those of the first generation. Sample 15, a mixture of G1 and G3 feldspars, has a ratio intermediate between them. The amazonite (G2) is somewhat enriched in Rb, as are the transitional stages between the microcline and the amazonite (11.1–6).

The second generation cleavelandite (17, 18) is likewise enriched in rubidium. This is the first well authenticated case of a plagioclase feldspar displaying enrichment in rubidium relative to potassium. The K and Rb spectra are shown in Plate 1. The spectra of the other plagioclases and the standard diabase W-1 (AHRENS and FLEISCHER, 1960) are shown for comparison.

Three possibilities exist to account for the Rb enrichment in a plagioclase: (a) The cleavelandite crystallised from a medium enriched in Rb (and Cs, see next section) and these elements were able, with K, to enter the lattice in proportions comparable to their concentrations in the liquid. (It is difficult to devise a reasonable process to enrich Rb and Cs selectively, relative to K, in the albite lattice).

(b) The Rb (and Cs) concentrations come from intergrowths of amazonite. The low bulk content of potassium does not support this explanation. No such intergrowths were visible in thin sections.

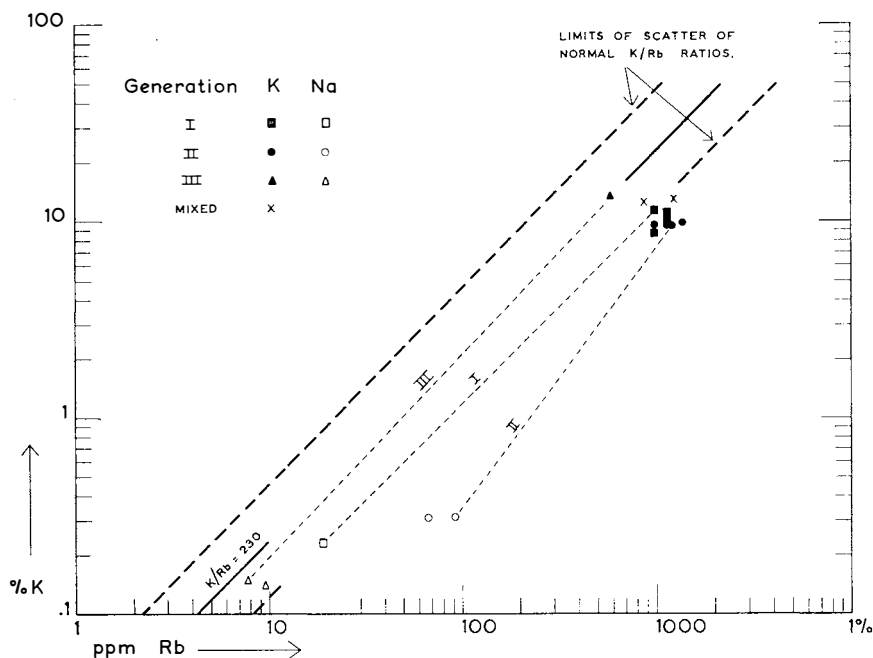


Fig. 3. K/Rb relationship. The similarity of K/Rb ratios for plagioclase and K- feldspars of the same generation should be noted. Generation 2 plagioclase (Cleavelandite) has slightly lower K/Rb ratios than the coexisting amazonite (see text).

(c) Adequate precautions to avoid contamination with the amazonite were taken, and in any event the K/Rb ratio of the cleavelandite is lower than that of the amazonite. This may also be seen from Fig. 3. The fact that the cleavelandite is relatively richer in Rb than the coexisting amazonite is in accord with the suggestion that the amazonite results from the action of the cleavelandite liquid on the first generation microcline. It is interesting to observe that the albite lattice is able to accommodate Rb and Cs (see next section) in amounts greater than normal. The concentration levels are of course, low; this may enable the crystal to reflect the composition of the liquid from which it is crystallising without apparent rejection of the larger ions: in any events there seems to be no selective effect based on the lower coordination of Na in albite compared with K in K- feldspar (TAYLOR and HEIER, 1958 b).

The relatively low amount of Rb in the third generation feldspars indicates that they are not derived from the "residual pegmatitic liquid", in which the Rb content would be expected to increase over that in the second generation. From the evidence of the K/Rb ratios, the G3 feldspars are not connected genetically with those of the preceding generations.

Cesium.

Cesium has a distribution pattern very similar to that of Rb in the Evje pegmatite, and its distribution can be interpreted in the same way. In the K- feldspars the Cs concentrations are high in G2

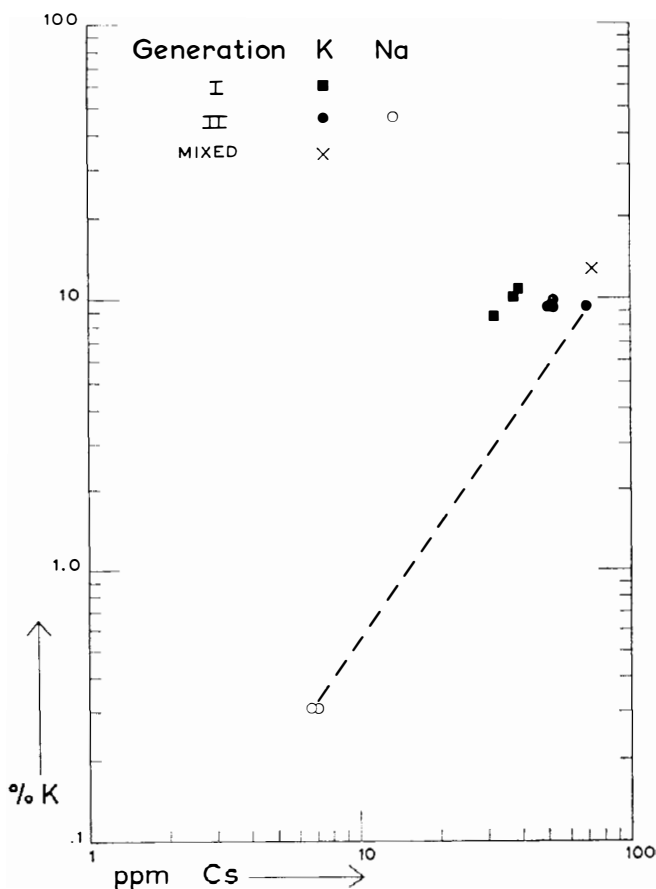


Fig. 4. K/Cs relationship. The cleavelandites have the same K/Cs ratio as the amazonites. The other plagioclases do not contain detectable Cs (< 2 ppm) and hence have much greater K/Cs ratios than the coexisting K- feldspars.

(70 ppm), not detected in G3, and not detected or low in G1. The G1—G2 transitions are intermediate in Cs content between the pure members, ranging from 32 to 54 ppm.

The presence of cesium in the cleavelandite is noteworthy. It was not detected in the other albites. The K/Cs ratio in the cleavelandites is 446 and is less than the K/Cs ratio of 1354 in the coexisting amazonite. The relative enrichment of Cs in the cleavelandite compared to the amazonite is very similar to that for rubidium. This can be easily seen from Figs. 3 and 4 where the slope of the cleavelandite-amazonite join for Cs is 56° (Fig. 4) and for Rb 53° (Fig. 3). Thus both the K/Rb and K/Cs ratios indicate that the cleavelandite is more highly fractionated than the amazonite. There is, of course, a considerable degree of enrichment in Cs compared with the abundant crustal rocks, where K/Cs ratio is commonly about 7000 (AHRENS, EDGE and TAYLOR, 1960).

Lead.

The first generation K- feldspar has the high lead content typical of pegmatite feldspars. WEDEPOHL (1956) found an average of 100 ppm in K- feldspars from pegmatites compared with an average of 25 ppm in K- feldspars from normal igneous rocks. The amazonite is slightly lower in lead than the G1 microcline, but the third generation feldspar is distinctly low in Pb.

The first generation albite contains 13 ppm lead. WEDEPOHL (1956) found an average of 10 ppm for albite, with some cleavelandites containing up to 32 ppm. Our cleavelandites, coexistent with the amazonite, contain 23 and 43 ppm Pb. HEIER (1960) found the Pb content of the plagioclases to be similar to or slightly higher than that of coexisting K- feldspars in amphibolite and granulite facies rocks. The distribution of lead in the Evje pegmatite feldspars is rather similar to rubidium and cesium. The geochemical behaviour of lead in K- feldspars has been interpreted by TAYLOR and HEIER (1960) mainly in terms of relative bond strengths. Lead tends to be concentrated in the later fractions. Divalent strontium of nearly the same size, is enriched in the earlier fractions, and its behaviour can, to a first approximation, be explained by the classical principle of capture (HEIER and TAYLOR, 1959 b). The contrasted behaviour of lead is attributed to the more covalent character of the Pb—O bond.

Thallium.

Tl⁺ is the same size as Rb⁺. AHRENS (1948) showed a close coherence between Rb and Tl to exist in rocks. HEIER and TAYLOR (1959 a) in an investigation of K- feldspars from the pre-Cambrian basement rocks of southern Norway found that Tl became enriched with respect to Rb as fractionation proceeded. This was shown by the steeper slope of the Tl-Rb plot in our case compared with that of AHRENS (1948). In particular, the large pegmatite feldspars were enriched in Tl with respect to Rb. This behaviour, which is similar to that of cesium, has been ascribed by TAYLOR and HEIER (1960) to the more covalent character of the Tl-O bond compared to the Rb-O bond. The larger size of cesium is the cause of its enrichment in the later fractions.

Figure 5 shows a plot of Tl against Rb. It should be compared with Fig. 8 of HEIER and TAYLOR (1959 a). Tl is relatively enriched, compared to Rb, in the amazonite, and in the G1-G2 and G2-G3 transitions. The first and third generation feldspars have normal amounts of Tl for the amount of Rb present.

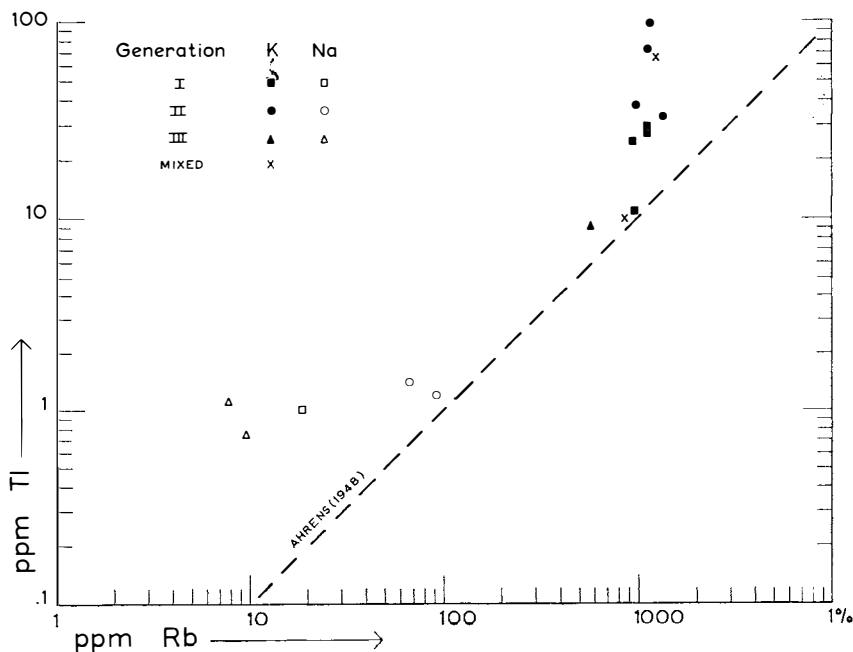


Fig. 5. Rb/Tl relationship. The average line of the Rb/Tl diagram from Ahrens (1948) is shown for comparison.

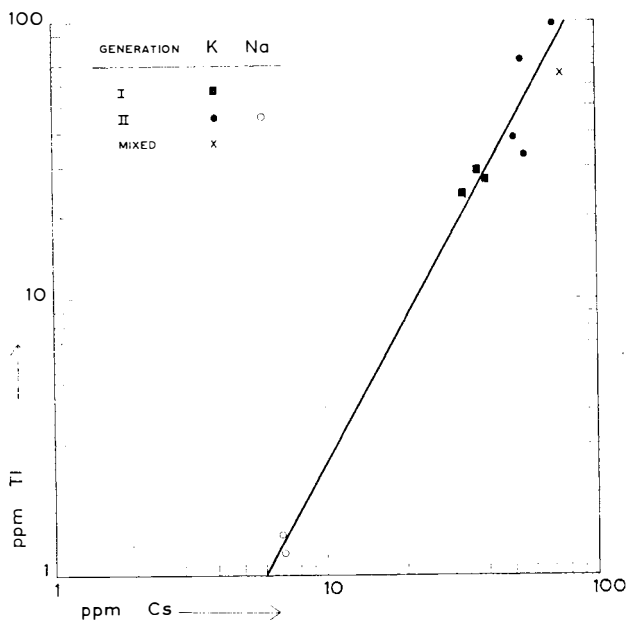


Fig. 6. Tl/Cs relationship. A close association is seen.

The Na- feldspars have the low amounts of Tl typical of the group, and the levels are comparable to published data (AHRENS, 1948; SHAW, 1952, 1957). In view of the fractionated character of the cleavelandites their low Tl content is surprising. SHAW (1952) gives a value of 6.8 ppm for a cleavelandite and this sample was available to us as a standard through the courtesy of Dr. D. M. Shaw. We can find no adequate reason to account for the low Tl content in our samples. If the cleavelandite lattice can accomodate Cs, there seems no reason why it should not accomodate Tl at a level of a few ppm.

Fig. 6 shows the Tl—Cs relation. In those samples where both elements were detected, a moderately close relationship exists. Compared to cesium, the amazonites show slight enrichment in Tl. This effect is also shown by the G1—G2 transition samples.

Calcium, Strontium and Barium.

The distribution of Ca, Sr and Ba in K- feldspars has been discussed by HEIER and TAYLOR (1959 b). They found in a fractionation series the sequence of entry of these elements in K- feldspars to be

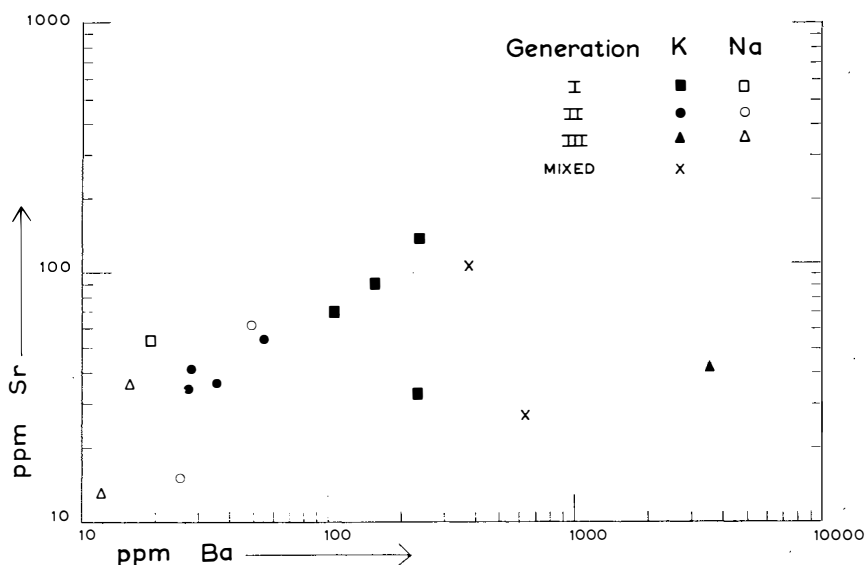


Fig. 7. Ba/Sr relationship.

in the order Ba, Sr, Ca, so that Ba decreases more rapidly than Sr, and Sr more rapidly than Ca, so that the Ba/Sr and Sr/Ca ratios both decrease with increasing fractionation. This was explained in terms of the relative strength of the Ba-O, Sr-O and Ca-O bonds.

Plots of ppm Ba vs. ppm Sr in the Evje pegmatite feldspars are shown on Fig. 7. There is no correlation between the Ba/Sr ratios of the K-feldspars from the different generations. The unique position of the G3 microcline is indicated by its abnormal ratio. It is extremely enriched in Ba. Both the G1—G3 and G2—G3 mixtures show the presence of the G3 microcline by being high in Ba. It is considered that the G3 microcline is not genetically related to the major pegmatite crystallisation. High Ba contents are typically associated with the high temperature, least fractionated, K-feldspars in normal rocks. However, also adularia of low temperature veins are typically high in Ba, and the Ba-feldspar, Celsian, is a product of low temperatures of formation. The amazonite (G2) and the G1—G2 transition feldspars have rather uniform Ba/Sr ratios. It was suggested that the amazonite formed through reactions between the G1 microcline and the "pegmatite rest liquid" from which cleavelandite crystallised. Fig.

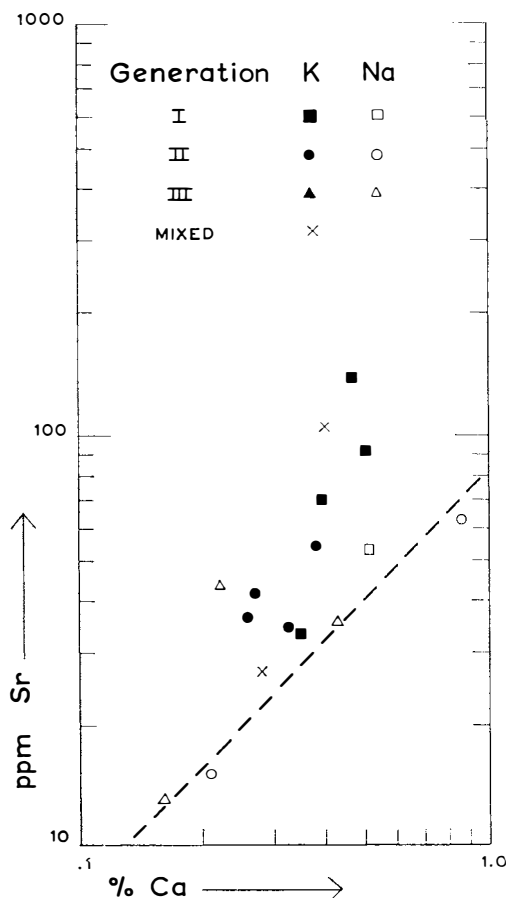


Fig. 8. Ca/Sr relationship. A close association is apparent in the plagioclase feldspars, but not in the K-feldspars.

7 gives some support for this assumption in that the amazonites and the amazonite-G1 transition series gradually changes towards typical plagioclase ratios by increasing "amazonitization".

As can be seen from Fig. 7 barium is a more sensitive indicator of the different generations than is strontium. The first generation K-feldspars contain 100–250 ppm Ba, the second generation less than 60 ppm and the third over 3000 ppm. These variations are not reflected by the plagioclase, probably for structural reasons. The cleavelandites are higher in Ba than the albites, and contain about the same concentrations as the co-existing amazonite.

The Sr/Ca ratios are not useful in distinguishing the various generations. The interesting feature here is the Ca–Sr coherence shown by the plagioclase feldspars (Fig. 8) and the relative enrichment in Sr relative to Ca displayed by the K-feldspars.

The green colour of amazonite.

GOLDSCHMIDT et al (1933), TOLMACHEV and FILIPOV (1935) and AHRENS (1945, p. 229) noted the large amounts of Rb in amazonites. KAPUSTIN (1939) correlated the green colouration with the amount of Rb; this was also the opinion of ZAVARITSKY (1943). BOROVIK-ROMANOVA and KALITA (1958) state that Rb is not responsible for the green colour. GOLDSCHMIDT (1954) noted that since both Rb-poor amazonites and Rb-rich red microclines are not uncommon, Rb could

not be responsible, but thought that perhaps Tl, normally a close associate of Rb might be the cause. He found no trace of Cr, Ni or Cu in amazonites. He did not think that the Pb present would affect the colour. ZIRHOV (personal communication with Professor L. H. AHRENS) considers that the substitution of ($\text{Pb}^{2+}\text{Al}^{3+}$) for (K^+Si^{4+}) takes place in microclines, but that in amazonites Pb^{2+} substitutes for 2K^+ , with a defect lattice resulting, which is responsible for the change in colour. ELISEEV (1949) and BASSETT (1956) attribute the green colour to ferrous iron.

A good opportunity exists in sample 11 (see Plate 2, Fig. 1). to examine the ideas which have been proposed to account for the green colour of amazonite, since in this sample a transition between green amazonite and pink microcline takes place.

Table 4. Comparison of the amounts of Fe, Cr, Mn, and Cu in pink microcline and green amazonite Plate 2, (Fig. 1).

Sample No.	Fe ppm	Cr ppm	Mn ppm	Cu ppm
11.7 green	290	0.11	10	3.8
11.8 pink	480	0.26	14	1.0
11.9 pink	470	0.23	15	1.2

The following elements were not detected in any of the samples (approximate detection limits are given in brackets).

F (100), Sn (50), Be (50), Co (5), Ni (1), V (1).

On a series of drilled samples 1 cm apart (11.1 to 11.6) we have determined Li, Na, K, Rb, Cs, Pb, Tl, Ca, Sr and Ba. Samples 11.7 to 11.9 were taken with extreme precautions to avoid metallic contamination and were used for the determination of Cu, Fe, Mn and Cr. Figures for these elements are given in Table 4. V, Co and Ni were not detected in these samples. An examination of the spectra of these three samples from 2470 to 9600 Å revealed no differences other than those tabulated. The only significant differences are that the amazonite contains rather more Ca and much less Sr and Ba than the microcline. These elements are rather critical indicators of fractionation (TAYLOR and HEIER, 1960; HEIER and TAYLOR, 1959 b). However, this difference is not more than that observed rather com-

monly in pegmatite feldspars of the same colour (see for example, data given by TAYLOR and HEIER, 1960, and HEIER and TAYLOR, 1959 b).

In Plate 2, Fig. 2 the alkali metal spectra are shown for 11.7, (green), 11.8 and 11.9 (red). These spectra show the uniform amounts of Rb and Cs. In Plate 2, Fig. 3 the spectra of the standard granite G-1 with 700 ppm F, the standard diabase W-1 with 200 ppm F (AHRENS and FLEISCHER, 1960), and samples 11.7 (green) and 11.8 (red) are shown. The CaF bandhead at 5291 Å may be seen clearly in G-1, less distinctly in W-1, and is absent in both the amazonite and the microcline.

Amazonites are commonly higher in Rb, Cs, Pb, Tl and lower in Ca, Sr and Ba than normal pegmatite feldspars, and are clearly products of an extreme fractionation process. We are lead to the conclusion that the cause of the green colouration is not to be found in a significant difference in composition from the microcline from the elements: Li, Na, K, Rb, Cs, Pb, Tl, Ca, Sr, Ba, Ga, Al, Si, Cu, Fe, Mn, Cr, V, Co, Ni, Sn, Be, and F.

It seems that the change in colour must be caused by physical differences, such as lattice defect or strain, rather than by chemical differences.

Acknowledgements.

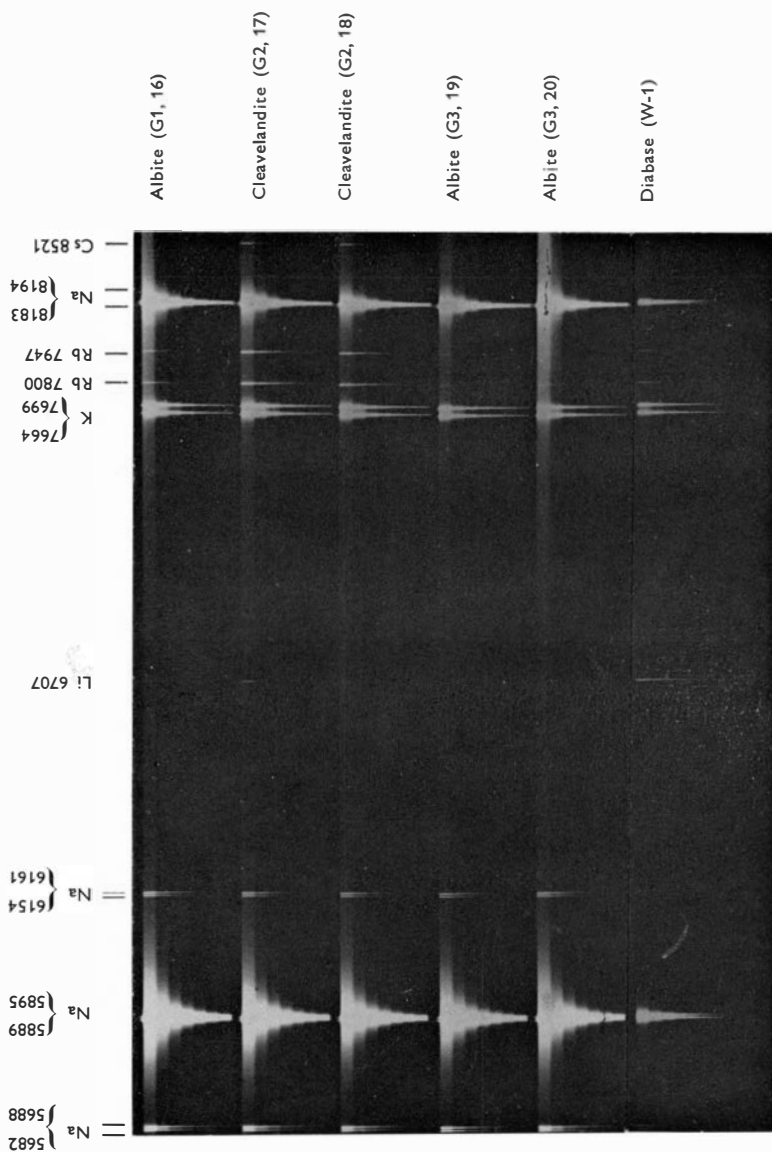
We are grateful to Miss Maureen Sachs for assistance with sample preparation and the spectrographic determinations; to Mr. B. Bruun and Miss K. Haugen for the Na and K determinations; to Professor I. Oftedal for some preliminary spectrographic determinations; to Professors L. H. Ahrens and T. F. W. Barth, and Dr. H. Neumann for reading the manuscript and for constructive suggestions.

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Alkali element spectra of plagioclase feldspars. Samples 16, 19 and 20 have normal K/Rb ratios and no Cs. The spectrum of the standard diabase, W-1 (Ahrens and Fleischer, 1960) is shown for comparison. It can be seen to have a similar K/Rb ratio to samples 16, 19 and 20. The cleavelandite samples (17, 18) are enriched in Rb relative to K, and have detectable Cs.

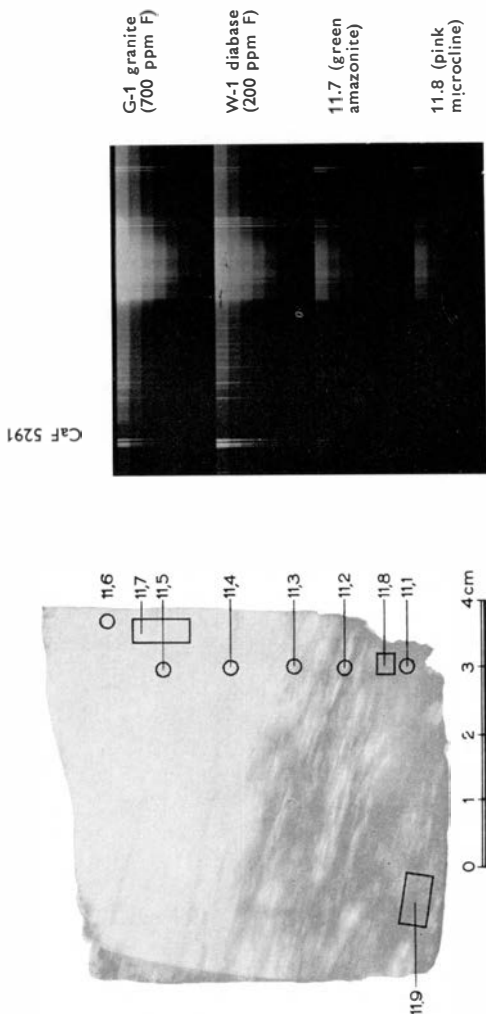


Fig. 1

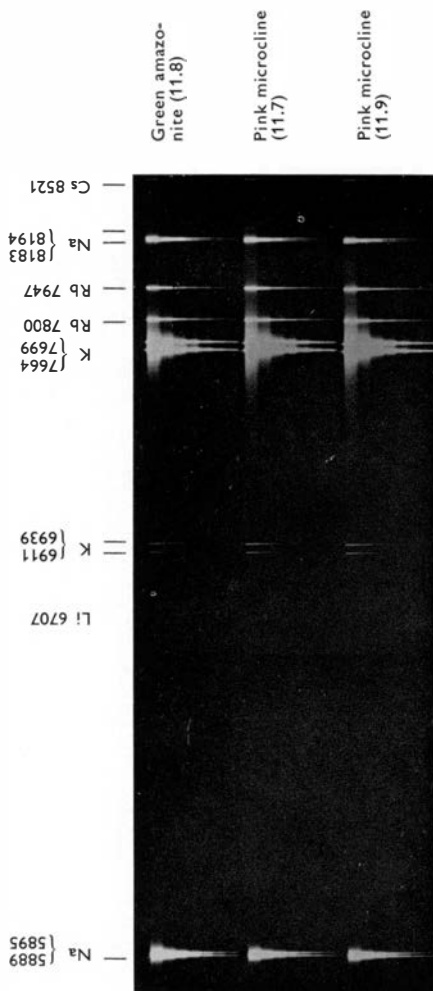


Fig. 2

Fig. 1. Transition between green amazonite (light) and pink microcline (dark). The position of the samples are indicated.

Fig. 2. Alkali metal spectra of green amazonite (sample 11.7) and adjacent pink microcline (11.8 and 11.9). The spectra are very similar.

Fig. 3. The CaF bandhead at 5291 Å can be seen easily in the standard granite G-1 (700 ppm F) and less clearly in the standard diabase W-1 (200 ppm F) (Ahrens and Fleischer, 1960). It is not present in either the green amazonite (11.7) or in the pink microcline (11.8). The other band spectra are due to CaO (CaCO₃ was added to all samples).