

Petrogenesis of the Oslo Region Larvikites and Associated Rocks

by ELSE-RAGNHILD NEUMANN

Mineralogisk-Geologisk Museum, Sarsgt. 1, Oslo 5, Norway

(Received 16 April 1979; in revised form 8 June 1979) -

ABSTRACT

The Larvik ring-complex is a possible key area for comprehension of the petrogenesis of the felsic rocks of the Oslo Region.

The ring-complex formed by a series of intrusions of weakly over- to weakly under-saturated intermediate magmas differing slightly in degree of iron-enrichment. Subsequent removal of ternary feldspar + olivine \pm clinopyroxene + Fe-Ti-oxides + apatite produced markedly over- and under-saturated derivatives. The lardalites seem to represent accumulates.

The formation of the approximately saturated intermediate melts which invaded the upper crust may best be explained by fractional crystallization of basaltic melts at pressures between about 7 and 10 kb, that is in the lower part of the crust. Within this pressure range both alkali and transitional basalts may form saturated felsic derivatives by fractionation. Small variations in P_{total} , $P_{\text{H}_2\text{O}/\text{CO}_2}$ or f_{O_2} during fractionation in the lower crust could give rise to the minor differences in silica activity and degree of iron-enrichment observed among the least evolved Larvik rocks.

The Oslo Region magmatic rocks show a positive correlation between calculated magma densities and the extensional stress at the time of emplacement, as inferred from other geological parameters. The extensive fractionation which gave rise to the large volumes of light intermediate magmas was probably caused by the density-filtering effect of the light continental crust during periods of low extensional stress.

INTRODUCTION

DURING the Permian period, southeastern Norway suffered an intracrustal rifting episode accompanied by extensive magmatic activity, which led to the formation of the Oslo Region alkaline province (Fig. 1). The main period of magmatism lasted from about 295 to 245 m.y. ago (Sundvoll, 1978*a*). It started with eruptions mainly of basalts and rhomb porphyry lavas, first from fissures, later from central volcanoes, and terminated with the emplacement of large intermediate to silicic batholiths (Ramberg & Larsen, 1978*a*) to such shallow depth that they locally intruded the lower part of the Permian lava pile (Brøgger, 1980).

Felsic rocks of monzonitic composition (larvikites, kjelsåsites, and rhomb porphyries) make up about 40 per cent of the exposed surface and subsurface rocks (Barth, 1945; Ramberg & Larsen, 1978*a*). Such high proportions of relatively highly evolved magmatic rocks are typical of continental rifts (Bailey, 1974). Gravity studies of the Oslo Region have, however, revealed the existence of large masses of dense rocks at depth along the entire Oslo Region (Ramberg, 1976). When the estimated volume of these mafic to ultramafic rocks is included in the frequency calculations, the proportion of felsic rocks is reduced to about 10 per cent, which is compatible with derivation by fractionation or partial melting of a basaltic parent (Ramberg, 1976).

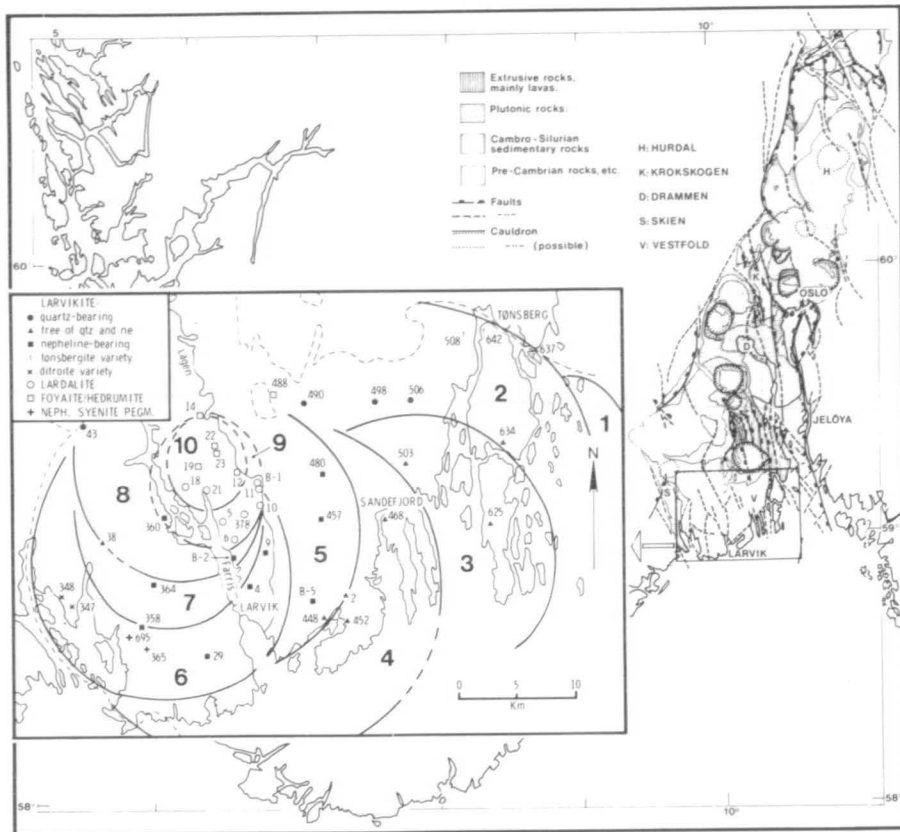


FIG. 1. Generalized map of the Larvik ring-complex showing sample localities, after maps by Holtedahl & Dons, 1960; Petersen, 1977, and Ramberg & Larsen 1978a.

It has been suggested elsewhere that the high proportion of mafic or ultramafic rocks at deeper levels, and of intermediate to silicic rocks at and near the surface of the crust, results from density filtering of the magmas by the continental crust (Neumann, 1978).

The extrusive mafic rocks in the Oslo Region range from ankaramites to quartz tholeiites, whereas the intermediate larvikites and rhomb porphyry lavas are approximately silica-saturated. Most of the highly evolved rocks are moderately to strongly oversaturated; strongly undersaturated ones are only found in the Larvik ring-complex. All these rocks are enriched in incompatible elements, including light rare earth elements (REE). Larvikites and lardalites (plagifoyaites) have low, very uniform Th/U and $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios (0.7038–0.7041). The same seems to be true for the basalts and rhomb porphyry lavas, although fewer data are as yet available on the extrusive rocks. These data on the intrusive and extrusive rocks have been interpreted to show derivation from an 'enriched' mantle source (or sources), and lack of detectable contamination from the surrounding rocks (Finstad, 1972; Raade, 1973, 1978; Weigand, 1975; Sundvoll, 1978a, b). However, Jacobsen & Wasserburg (1978) have found small variations in initial $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ values among 12 basalts, intermediate and felsic volcanic rocks and

plutonic rocks from the Oslo Region. They find these variations compatible with derivation of the basalts from three different 'undepleted' to slightly 'depleted' type mantle sources, and suggest that small amounts of crystal contamination may be important in the genesis of the intermediate rocks. The nordmarkites (syenites) and ekerites (alkali granites) have slightly higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than the intermediate rocks, and more variable incompatible element ratios. The geochemistry of many of these rocks is compatible with fractionation of a 'larvikitic' magma; contamination, however, cannot be totally disregarded (Raade, 1973, 1978; Neumann *et al.*, 1977; Sundvoll, 1978a). Significant contamination is only indicated in a fine-grained to porphyritic monzonitic rock-type called akerite, which generally occurs as a border facies between intrusive bodies and Cambro-Silurian sedimentary rocks, and in intrusive bodies with appreciable amounts of sedimentary xenoliths (*e.g.* the Grefsen nordmarkite; Brøgger, 1890; Barth, 1945; Oftedahl, 1946, 1953; Sæther, 1962; Raade, 1973, 1978; Sundvoll, 1978a).

The petrogenetic relationship between the different plutonic rocks in the Oslo Region has been discussed in a number of papers. Brøgger (1898) and Barth (1945) proposed, on the basis of whole rock major element analyses, that the rocks fall on two main evolutionary trends, one towards silica-oversaturation (larvikite → nordmarkite → ekerite), the other towards undersaturation (larvikite → lardalite → nepheline syenite pegmatite; Fig. 2). Bailey & Schairer (1966) believed the geochemical evolution to be governed by the composition of the fractionating feldspar. On the basis of Oftedahl's analyses of separated feldspars (1948), indicating Fe_2O_3 concentrations up to 2.99 wt. per cent in larvikites but only about

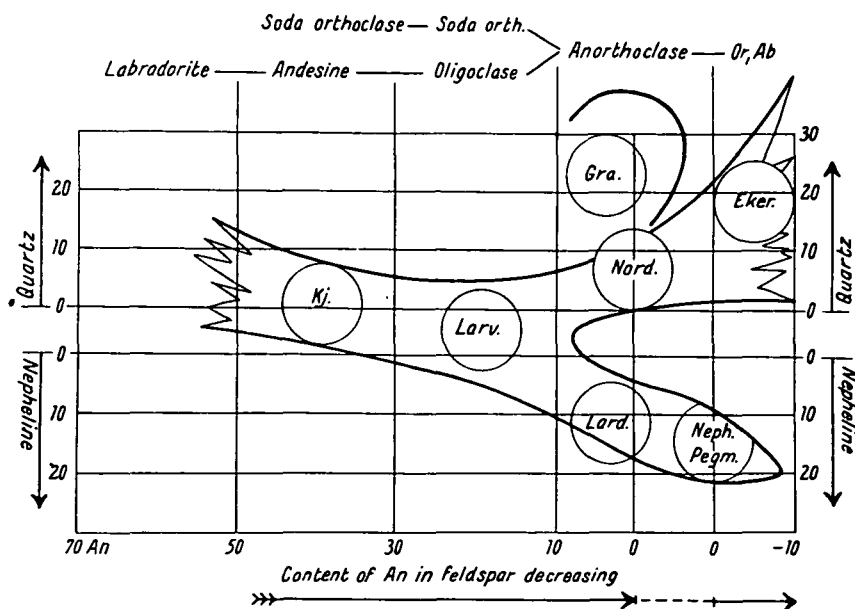


FIG. 2. Barth's (1945) 'family tree' indicating petrogenetic relationships among the main Oslo plutonic rocks. Kj = kjelsåsité, Larv = larvikite, Lard = lardalite, Neph Pegm = nepheline pegmatite, Nord = nordmarkite, Eker = ekerite, Gra = granite.

1 wt. per cent in the nordmarkites, they suggested that removal of Fe_2O_3 by feldspar ($\text{NaFe}^{3+}\text{Si}_3\text{O}_8$) instead of by clinopyroxene ($\text{NaFe}^{3+}\text{Si}_2\text{O}_6$) from a silica-saturated or even oversaturated magma (larvikitic), might result in undersaturated derivatives (lardalites), whereas the separation of iron-poor feldspars would lead to oversaturation. Differences in the concentrations of iron in feldspar might, according to Bailey & Schairer (1966), result from small differences in P_{O_2} , $P_{\text{H}_2\text{O}}$ or P_{total} . However, T - f_{O_2} determinations (Neumann, 1976) do not support this hypothesis.

Recent trace element analyses have revealed the existence of a series of evolutionary trends (Raade, 1973, 1978; Neumann, 1976; Neumann *et al.*, 1977). Alternative modes of formation for particular rock types have also been suggested. Raade (1973, 1978) believed the quartz-bearing larvikite-variety tønbergite to be metasomatically enriched in SiO_2 , K_2O , U and Th. Petersen (1978) has proposed that the lardalites formed by progressive partial melting of undersaturated larvikites and not by fractionation as previously proposed.

None of the above papers discusses what appears to the author to be a central problem: 'Why are the intermediate rocks approximately silica-saturated?'

The present petrological and geochemical studies were therefore initiated in an attempt to establish the evolutionary history of these intermediate rocks. The Larvik ring-complex (SW Oslo Region) is considered to be a key area because it contains both over- and undersaturated intermediate rocks and was therefore chosen for detailed study. In the course of the study it has proved possible to explain both the highly uniform degree of silica saturation in the intermediate rocks and the strongly divergent trends towards both over- and undersaturation in the more strongly evolved rocks of the pluton complex.

GEOLOGICAL SETTING OF THE LARVIK RING-COMPLEX

The Larvik ring-complex is the largest body of monzonitic rocks in the Oslo Region (Fig. 1). The pluton consists of a number of semi-circular sections, with chilled cross-cutting contacts indicating a progressive decrease in age towards the west (Petersen, 1977). The oldest, easternmost ring-sections (RS-1, 2) consist of quartz-bearing larvikite (including the variety tønbergite). Westward the larvikites change gradually to become silica-undersaturated. Ring-sections 9 and 10 are made up of lardalite and foyaite/hedrumite (nepheline syenite). The ring-complex is cut by numerous pegmatite dikes, nepheline syenite pegmatites in the western part (Brøgger, 1890), and quartz syenite pegmatites in the eastern part (G. Raade, pers. comm., 1978). The larvikites and lardalites in the ring-complex have Rb-Sr ages of 277 ± 3 and 269 ± 5 m.y. respectively (Sundvoll, pers. comm., 1978).

The different rock types in the Larvik complex have been described by Brøgger (1890, 1898), Raade (1973), Neumann (1976), and Petersen (1978). Mafic mineral compositions are discussed by Neumann (1976); the feldspars are briefly discussed below.

Sample localities are shown in Fig. 1. In tables and diagrams the ring-section number is commonly included in the sample number; e.g. 5-480 denotes sample 480 in ring-section 5.

ANALYTICAL PROCEDURES

The feldspar analyses (Table 2) reported in this paper were done on an ARL-EMX microprobe at the Central Institute of Industrial Research, Oslo, using a series of natural and synthetic standards. The raw data were corrected for dead-time and background. Corrections for mass absorption and fluorescence were made by a procedure described by Bence & Albee (1968), using correction factors from Bence & Albee (1968) and Albee & Ray (1970). The feldspars in the analyzed rocks are exsolved to perthites. Each reported analysis is the mean of 3 or more scanning analyses across rim or core in zoned feldspars, or across the whole grain where the primary feldspar appears to have been homogeneous. Several grains were analyzed in each sample.

The whole rock analyses (Table 3) were done by flame photometry (Na), atomic absorption (Mg), and X-ray fluorescence analyses of fused pellets (other major elements). Fe^{2+} was determined by titration. Trace elements were analyzed by instrumental activation analysis methods described by Gordon *et al.* (1968) and Brunfelt & Steinnes (1969). Precision is estimated by 'pooled variance' ($\epsilon\%/\sqrt{2}$) of duplicate analyses, where

$$\begin{aligned} \epsilon\% &= \sqrt{\text{Sp}^2} \cdot 100/\bar{x}, & \text{Sp}^2 &= (S_1^2 + S_2^2 + \dots + S_N^2)/N, \\ S_i^2 &= (X_i^A - \bar{x}_i)^2 + (X_i^B - \bar{x}_i)^2, & \text{and} & \quad \bar{x} = (\bar{x}_1 + \bar{x}_2 + \dots + \bar{x}_N)/N. \end{aligned}$$

DIFFERENTIATION MODELLING

The interpretation of the analytical data is partly based on a comparison of observed with computed trace element behavior assuming fractional crystallization.

Trace element enrichment factors (E_c) are calculated on the basis of the equation

$$E_c = c_L/c_0 = F^{(D-1)} \quad (\text{Greenland, 1970}),$$

where c_0 and c_L are the concentrations of a given trace element in the original and residual melt respectively, F is the fraction of melt remaining, and D is the bulk distribution coefficient for a given element.

$$D = \sum_1^N w \cdot K^{\alpha/L},$$

where W is the weight fraction of phase α and $K^{\alpha/L}$ is the mineral/liquid partition coefficient. F and W_1, W_2, \dots, W_N are calculated from major element compositions on the basis of a modified version of the petrographic mixing computer program written by Bryan *et al.* (1969).

Mineral/liquid partition coefficients (P.C.) are listed in Table 1. Anorthoclase rim/rhomb porphyry matrix P.C.s have been determined by A. O. Brunfelt and B. T. Larsen (pers. comm., 1978). With the exception of Eu and Ba, these data agree well both with those determined for natural anorthoclase in phonolite lavas on Ross Island (Sun & Hanson, 1976), and with experimental data on plagioclase at 1150 °C (Drake & Weill, 1975). The larvikites and lardalites crystallized under

TABLE 1
Partition coefficients, mineral-melt

	<i>Olivine</i>	<i>Augite</i>	<i>Anorth.</i>	<i>Alk</i>	<i>Ore</i>	<i>Apatite</i>
La	(0.008)	(0.12)	0.23	(0.060)	0.05	(15.0)
Ce	0.009	0.15	0.21	0.044	0.05	18.0
Sm	0.011	0.50	0.15	0.018	0.05	29.3
Eu	0.010	0.51	1.4	1.13	0.05	20.5
Tb	0.013	0.68	0.13	0.011	0.05	25.6
Yb	0.023	0.62	0.15	0.012	0.05	13.1
Lu	0.026	0.56	0.12	0.006	0.05	11.2
Rb		0.032	0.33	0.95		
Sr		0.516	4.19	2.7		5.5
Ba		0.131	1.49	10		0.073
Ref.	1, 3	1	2	1, 3	6	4, 5

1. Arth, 1976.
2. Phenocryst rim/rhomb porphyry lava, Oslo Region, A. O. Brunfelt & B. T. Larsen, pers. comm.
3. Irving, 1978.
4. Nagasawa, 1970.
5. Nash, 1972; Nash & Wilkinson, 1971.
6. Zielinski, 1975.

$T-f_{O_2}$ conditions close to those defined by the QFM-buffer (Neumann, 1976). As the rhomb porphyry lavas appear to have cooled under somewhat more oxidizing conditions (Ramberg & Larsen, 1978a), a $Eu/Eu^* = 10$ ($Eu^* =$ interpolated Eu^{3+} value), suggested by the experimental data of Drake & Weill (1975) to be pertinent for QFM-conditions, was used in the calculations. Also the Ross Island anorthoclases gave $Eu/Eu^* \sim 10$. Partition coefficients for the same elements entering other mineral phases were chosen from the literature on the basis of the following requirements: (a) they should be representative of systems of intermediate compositions; (b) they should lie in the overlapping field between value-sets determined experimentally and from natural porphyritic lavas.

Hf, Ta, Th, and U generally behave as strongly incompatible elements (Ferrara & Treuil, 1974; Leeman *et al.*, 1976; Baker, 1978; Baker *et al.*, 1978). These elements are below referred to as SIE. Bulk distribution coefficients (D) for SIE have been set to zero to obtain maximum enrichment factors for different F -values.

RESULTS

Feldspars

The compositions of feldspars in various Oslo Region plutonic rocks are listed in Table 2. The microprobe data confirm the earlier observations (Brøgger, 1890, 1898; Barth, 1945; Oftedahl, 1948; Muir & Smith, 1956; Smith & Muir, 1958) that the l arvikites carry a Ca-bearing alkali feldspar. These may have cores of plagioclase, often partly resorbed. The alkali feldspar is unmixed to crypto-, micro- or macropertthite. Also the least silicic nordmarkites frequently carry plagioclase coated by alkali feldspar, whereas the quartz-rich nordmarkites and ekerites, like

TABLE 2
Average feldspar compositions

	Cation per cent			Weight per cent		Rock
	An	Ab	Or	$Fe_2O_3^{TOTAL}$	MgO	Na + K
						Al
LARVIK PLUTON						
<i>Larvikite</i>						
4-2 plag. cores	31.5	65.5	3.0	0.53	0.14	0.82
alk. coat.	11	64	25			
5-448	12	68	20	0.39	0.14	0.79
6-4	14	65	21	0.04	0.02	0.80
6-9	10	65	25	0.03	0.03	0.81
6-29	7	66	27	0.06	0.02	0.86
8-B-2	15	66	19			0.73
<i>Lardalite</i>						
9-6	2	61	37	0.18	0.04	0.93
9-B-1	4	68	28	0.09	0.01	0.93
10-21	3	69	28	0.18		0.91
<i>Foyaité/hedr.</i>						
10-PØ-77	0.3	53	46.7	0.25	0.04	1.05
488	0.3	56	43.7	0.20		1.08
OTHER PLUTONS (see Neumann, 1976)						
<i>Larvikite</i>						
285 plag.	27.5	66.0	6.5	0.27	0.03	0.65
alk.	5.5	33.0	61.5			
290 plag.	26.5	66.5	7.0	0.37	0.08	0.71
alk.	3.5	32.0	64.5			
905 plag.	28	67	5	0.41	0.08	0.58
alk.	5	31	64			
916 plag.	29	66	5	0.29	0.04	0.68
alk.	5	35	60			
1101 plag. cores	32	64	4	0.29	0.10	
alk. coat.	25	65	10			
<i>Nordmarkite</i>						
113	1.5	65.0	33.5	0.58	0.01	1.17
126 plag. cores	33.5	65.0	1.5	0.19	0.05	0.93
alk. coat.	2.8	58.80	40.12			
162	0.2	~59	~41	0.63	0.01	1.13
176 plag. cores	25	70	5	0.28	0.02	0.86
alk. coat.	4.13	56.72	40.15			
303	0.2	~59	~41	0.59	0.01	1.06
<i>Ekerite</i>						
144	0	~52	~48	1.23	0.09	1.08
271	0	~64	~36	1.00	0.04	1.03
743	0.5	~54	~46	0.94	0.01	1.23
<i>Oftedal, 1948</i>						
Larv. L5. Larvik pl.	13.6	70.2	16.2	2.99	0.67	
Nordmark. N1	3.0	61.6	35.4	1.03		
N2	2.4	66.1	31.5			
<i>Muir & Smith, 1956</i>						
Larvikite No. 1	12.2	61.2	26.7	0.38	0.07	
No. 8	16.8	58.8	24.4			

the nepheline-rich lardalites, foyaite/hedrumites and nepheline syenite pegmatites carry only alkali feldspar.

The highest iron contents (0.58–1.23 wt. per cent) were found in feldspars of peralkaline silica-oversaturated nordmarkites and ekerites. This finding is consistent with previously reported feldspar–host rock compositional relationships. The numerous feldspar analyses compiled by Smith (1974) include a few electron microprobe determinations of Fe_2O_3 -concentrations ≥ 1.0 wt. per cent. The host

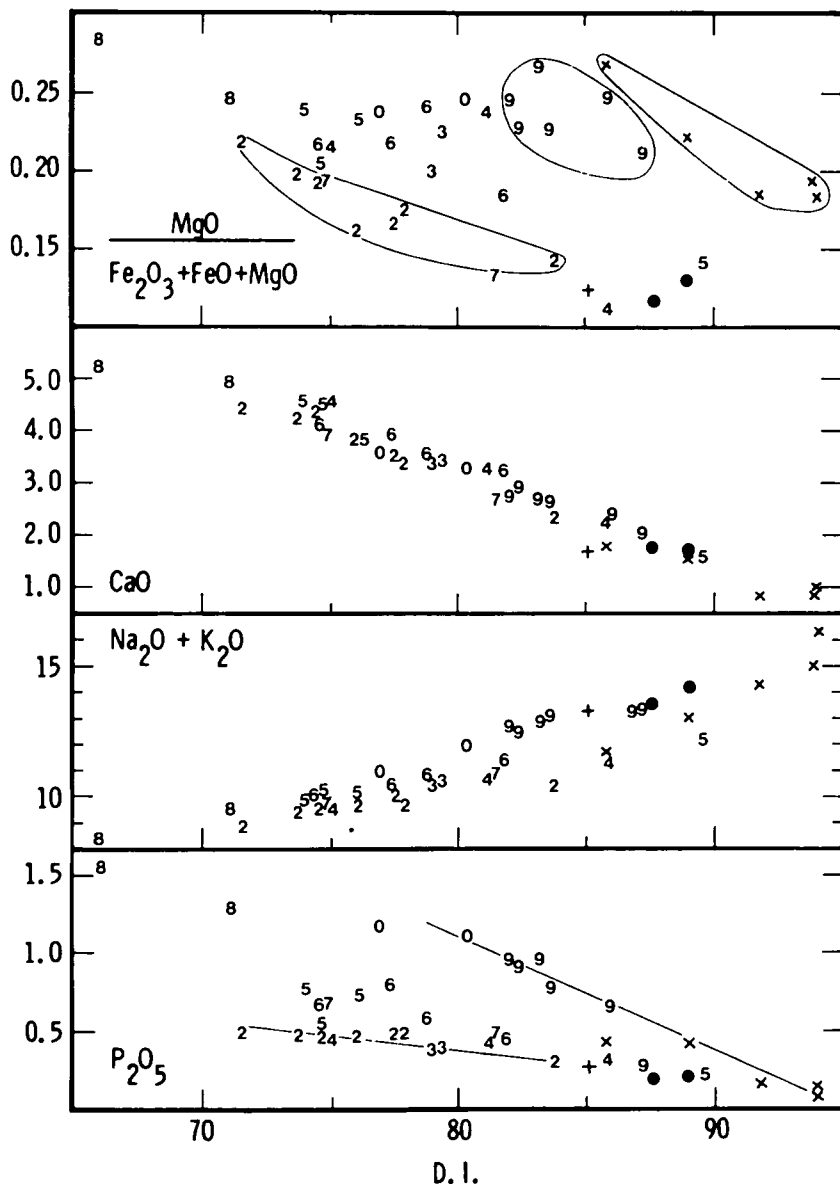


FIG. 3. Selected major element parameters plotted against differentiation index (D.I. = normative $q + \text{or} + \text{ab} + \text{ne}$). The numbers 2–8 = larvikites from ring-sections 2–8. 9 and 0 = lardalites from ring-sections 9 and 10 respectively. x and + = foyaite/hedrumites inside and outside ring-section 10 respectively. Filled circles = ditroites.

TABLE 3
Major and trace element analyses

Larvikites														
R.S	2							3		4			5	
Wt. per cent	642 ¹	498	506	490	634	637 ¹	508 ¹	503	625	452	2 ¹	468	2867	457
SiO ₂	57.77	57.60	58.31	58.32	59.30	59.50	60.34	59.82	58.42	58.21	59.71	59.65	56.15	56.55
TiO ₂	1.18	1.27	1.17	1.09	1.15	1.33	1.22	1.08	1.14	1.09	1.00	1.60	1.75	1.32
Al ₂ O ₃	18.65	18.34	18.56	18.47	18.71	17.10	15.84	19.85	18.68	18.34	18.59	16.64	17.97	19.36
Fe ₂ O ₃	4.42	2.49	2.37	2.69	2.01	5.13	5.66	1.92	2.09	2.62	1.90	2.86	3.04	2.02
FeO	1.30	3.06	3.02	3.00	3.10	1.30	0.29	2.20	2.62	2.21	2.33	2.80	3.25	3.09
MnO	0.11	0.13	0.13	0.15	0.13	0.15	0.12	0.13	0.15	0.10	0.14	0.23	0.20	0.15
MgO	1.60	1.35	1.29	1.07	0.98	1.35	0.99	0.95	1.37	1.05	1.31	0.67	1.99	1.32
CrO	4.47	4.24	4.13	3.86	3.56	3.34	2.34	3.41	3.46	3.52	3.25	2.30	4.57	4.43
Na ₂ O	5.16	5.33	5.65	5.77	5.86	5.22	4.54	6.11	6.22	5.58	6.54	5.60	6.10	5.67
K ₂ O	3.66	4.10	3.91	3.92	4.18	4.41	5.85	4.37	4.34	4.97	4.11	5.73	3.67	4.41
P ₂ O ₅	0.49	0.47	0.46	0.47	0.47	0.48	0.30	0.37	0.39	0.44	0.42	0.31	0.76	0.55
H ₂ O	0.75	0.58	0.66	0.59	0.87	0.83	0.83	0.61	—	0.42	0.58	0.89	0.53	0.67
TOTAL	99.56	98.96	99.66	99.40	100.32	100.28	98.32	100.82	98.88	99.40	99.88	99.28	99.98	99.54
D.I.	71.6	73.7	74.6	76.0	77.5	77.9	83.8	79.0	79.4	80.0	81.2	85.9	74.0	74.6
MgO/(Fe ₂ O ₃ + FeO + MgO)	0.219	0.196	0.193	0.161	0.166	0.175	0.143	0.199	0.225	0.215	0.236	0.111	0.240	0.205
Norm Q	3.3	1.1	0.9	1.0	0.8	4.7	6.4	—	—	—	—	1.2	—	—
Norm Ne	—	—	—	—	—	—	—	—	2.2	—	0.9	—	2.7	3.2
ppm														
Sc	8	9	8	8	7	10	8	7	7	6	6	10	11	8
Cr	59		32	n.d.	37	25	10		10				17	
Co	11	11	9	8	8	8	6	6	7	5	3		10	9
Rb	99	125	115	84	103	141	200	99	115	91	148		59	166
Sr	1260	590	960	980	620	710	210	690	710	620	54		970	860
Cs	1.0	1.3	1.6	1.1	1.4	1.3	1.5	1.4	1.9	0.9	1.0		0.6	1.8
Ba	1070	840	850	1150	1100	1350	630	930	890	1090	153		890	920
La	127	168	156	141	151	141	185	123	121	122	138		116	127
Ce	250	290	310	260	270	290	290	160	230	110	230		240	260
Sm	16.9	21.0	18.8	17.4	17.6	18.9	23.7	12.7	12.9	8.2	12.5		14.4	12.8
Eu	5.7	4.5	6.0	6.7	5.4	6.9	4.2	4.0	4.2	6.4	4.7		4.5	3.7
Tb	1.7	2.1	2.7	2.0	1.9	2.3	2.8	1.8	1.7	1.0	1.4		1.6	1.5
Yb	6	14	3	3	7	7	10	4	3	3	5		3	3
Lu	0.69	1.36	1.25	0.82	0.76	1.12	1.23	0.52	0.64	0.37	0.52		0.55	0.55
Hf	20	34	26	20	11	21	34	12	15	3.6	8.5		12	12
Ta	8.2	11.8	13.6	9.9	10.1	10.1	14.4	8.4	9.7	2.7	7.1		6.6	8.4
Th	16.4	24.1	21.2	14.4	16.8	21.0	32.3	15.4	17.9	3.7	17.8		9.6	16.6
U	4.9	6.6	6.3	4.7	5.3	6.4	8.0	3.2	4.2	1.7	4.5		2.4	4.6
Eu/Eu*	1.14	0.57	0.96	1.23	1.00	1.16	0.61	0.97	0.92	2.51	1.21		0.90	0.96

¹ Tensbergite.

² P.V. = pooled variance ($\epsilon\%/\sqrt{2}$, see text). P.V.s for the trace elements are estimated from sample replicates analyzed at different times.

³ The high number is caused by interference by La in REE-rich samples.

⁴ Major element data from Neumann, 1976.

⁵ REE data from Neumann *et al.*, 1977, Th-U from Raade, 1973.

TABLE 3 (continued)

	Larvikites								Ditroite		Neph. Sy. Pegm.			
	5		6				7		8		6		6	
	448 ¹	480	358 ²	9 ¹	4 ¹	29 ¹	364	38	360	B-2 ¹	347	348	365	695
SiO ₂	58.23	60.01	57.18	56.89	57.33	58.41	50.71	58.60	55.72	55.54	55.91	57.12		
TiO ₂	1.52	0.77	1.21	1.37	1.18	1.29	2.11	1.76	1.21	1.63	0.84	0.92		
Al ₂ O ₃	17.90	18.66	19.77	18.22	19.01	18.89	17.75	16.86	19.45	18.64	20.07	20.48		
Fe ₂ O ₃	2.55	1.97	1.27	2.06	1.66	1.93	4.51	2.75	2.21	2.54	1.69	1.30		
FeO	3.87	1.68	3.70	3.77	3.05	3.06	4.04	3.70	2.89	4.36	3.70	3.50		
MnO	0.20	0.18	0.16	0.18	0.15	0.17	0.22	0.27	0.17	0.21	0.22	0.17		
MgO	1.94	0.60	1.34	1.61	1.49	1.13	3.35	0.90	1.22	2.25	0.69	0.65		
CaO	3.85	1.59	4.37	3.91	3.56	3.22	5.21	2.77	3.98	4.94	1.76	1.68		
Na ₂ O	5.98	6.97	6.24	6.30	6.51	6.92	5.70	5.56	5.48	5.94	7.75	7.92		
K ₂ O	4.05	5.24	3.73	4.07	4.16	4.50	2.71	5.28	4.15	3.54	5.84	6.28		
P ₂ O ₅	0.72	0.21	0.66	0.79	0.58	0.46	1.54	0.48	0.66	1.29	0.19	0.20		
H ₂ O		0.89	0.54		0.55		1.24	1.40	1.46		1.24	0.56		
TOTAL	100.80	98.77	100.17	99.17	99.23	99.98	99.09	101.52	98.60	100.88	99.90	100.78		
D.I.	76.1	89.6	74.5	77.4	78.8	81.8	74.8	81.5	66.0	71.1	87.6	89.0		
MgO/(Fe ₂ O ₃ + FeO + MgO)	0.232	0.141	0.216	0.216	0.240	0.185	0.193	0.133	0.282	0.246	0.116	0.130		
Norm Q	—	—	—	—	—	—	—	0.3	—	—	—	—		
Norm Ne	—	5.1	3.5	4.0	5.2	7.3	0.1	—	3.1	2.4	21.5	22.6	+	+
ppm														
Sc	9	7		10	6	5	9	17	6	10	6	6	4	4
Cr	19			<2	42	23				21	4	11	n.d.	n.d.
Co	10	2		7	7	9	8	4	18	11	3	3	6	5
Rb	102	178		90	92	131	95	96	52	71	291	330	451	613
Sr	680	520		970	880	690	1030	290	1230	1220	69	39	35	300
Cs	1.2	2.4		0.6	1.0	1.6	0.8	0.5	0.3	0.3	4.1	1.1	1.7	20
Ba	1120	1060		1150	1160	1090	1170	630	1390	1160	75	108	290	126
La	151	137	111	121	123	117	131	109	145	128	201	130	251	433
Ce	270	250	220	170	220	240	190	170	210	270	420	250	620	780
Sm	17.4	9.1	12.6	13.5	11.2	9.6	13.8	12.7	15.0	14.8	20.2	14.8	26.8	34.3
Eu	4.8	2.1	4.6	4.2	4.7	3.9	4.1	4.0	4.6	5.4	2.1	1.40	2.2	3.1
Tb	2.1	1.4	1.3	1.4	1.1	1.0	1.5	1.7	1.8	1.6	2.2	1.6	2.0	3.7
Yb	6	5	4	4	5	4	4	3	2	2	17	9	20	5
Lu	1.04	0.86	0.56	0.48	0.61	0.63	0.57	0.40	0.66	0.48	2.5	1.2	3.5	6.4
Hf	16	30		13	16	18	10	7.5	5.1	5.6	59	45	30	36
Ta	13.2	15.0		7.9	10.6	12.2	8.0	6.7	5.8	4.6	43	17	55	39
Th	19.3	30.5	14.2	12.0	20.2	18.7	11.2	6.2	10.4	6.0	50	15.1	90	186
U	5.5	11.6	3.8	3.2	5.5	4.6	3.1	1.9	2.4	1.6	19	9.4	37	23
Eu/Eu*	0.88	0.70	1.20	1.05	1.43	1.36	0.97	0.98	0.97	0.92	0.34	0.31	0.29	0.29

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TABLE 3 (continued)

	Lardalite						Foyaité/Hedrumite						P.V. (%) ^a		
	9			10			10			11				488 ¹	
	10	5	B-1	11	6 ¹	378	18	21 ¹	23	19	14	22			12
SiO ₂	52.91	53.34	54.51	53.33	55.55	56.35	52.37	52.38	58.69	57.76	56.20	56.19	55.90	56.98	0.3
TiO ₂	1.53	1.45	1.49	1.28	1.12	0.73	1.67	1.41	1.11	1.20	0.85	0.96	0.61	1.22	0.8
Al ₂ O ₃	19.57	19.29	19.69	21.30	20.38	22.13	18.96	18.81	18.04	19.09	22.27	22.01	23.18	17.31	0.5
Fe ₂ O ₃	2.45	2.61	2.16	2.04	2.06	1.08	2.82	2.62	1.81	1.52	1.57	0.90	1.30	3.10	0.7
FeO	3.70	3.33	3.42	2.79	2.42	1.69	3.86	3.67	2.37	2.39	0.98	1.22	0.67	1.95	
MnO	0.21	0.20	0.18	0.17	0.16	0.10	0.21	0.21	0.23	0.19	0.14	0.12	0.09	0.25	1
MgO	2.00	1.75	2.02	1.42	1.46	0.76	2.07	2.04	1.53	1.11	0.58	0.51	0.44	0.71	4
CaO	2.77	2.92	2.68	2.65	2.32	2.01	3.59	3.30	1.81	1.55	0.97	0.83	1.03	1.70	0.4
Na ₂ O	7.90	7.63	7.93	8.42	8.15	8.43	6.75	7.55	6.07	7.26	8.68	9.06	10.57	7.61	<1
K ₂ O	4.68	4.83	4.93	4.63	5.04	4.85	4.19	4.32	5.64	5.73	5.66	5.94	5.70	5.71	0.5
P ₂ O ₅	0.95	0.89	0.94	0.77	0.64	0.28	1.20	1.11	0.43	0.42	0.16	0.14	0.08	0.26	2
H ₂ O	0.76	0.88		0.99	0.75	0.71	1.22		1.29	1.07	1.04	1.83	0.58	1.04	<1
TOTAL	99.43	99.12	99.95	99.79	100.05	99.12	98.91	97.42	99.02	99.29	99.10	99.71	100.15	97.84	
D.I.	82.0	82.4	83.2	83.6	85.9	87.2	76.9	80.3	85.8	89.0	91.8	93.9	94.0	85.1	
MgO/(Fe ₂ O ₃ + FeO + MgO)	0.245	0.228	0.266	0.227	0.246	0.215	0.237	0.245	0.268	0.221	0.185	0.194	0.183	0.123	
Norm Q	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Norm Ne	21.7	19.9	20.5	24.5	20.6	21.0	12.8	18.9	3.1	13.0	24.4	28.1	37.5	14.7	
ppm															
Sc	5	6	9	5	4	3	12	6	11	8	8	1	12	8	2
Cr	36	36	34	nd	37		44	21	22	39			nd.	<2	
Co	9	9	1	8	5	3	6	10	2	3	1	1	1	2	
Rb	120	130	161	115	131	164	78	138	127	189	166	159	129	134	5
Sr	750	820	630	690	760	760	420	950	170	190	190	58	nd.	320	7
Cs	1.1	1.5	1.1	0.9	1.0	1.5	0.6	2.8	1.4	2.2	2.1	2.1	0.22	1.1	12
Ba	880	940	1400	1070	1150	1060	1710	1220	790	570	86	116	53	460	4
La	122	118	119	110	95	63	89	112	130	110	112	136	101	119	3
Ce	220	240	230	200	130	170	150	240	220	190	190	210	260	160	11
Sm	11.1	11.6	15.5	9.5	7.4	4.3	9.6	9.7	10.2	9.4	11.0	17.1	13.7	12.6	3
Eu	3.4	3.0	6.8	3.5	2.7	2.5	5.2	4.0	3.0	3.3	2.1	3.0	2.2	2.9	8
Tb	1.4	1.3	1.0	1.1	1.0	0.5	1.1	1.5	1.5	1.2	1.6	1.9	1.8	1.4	8
Yb	2	3	5	5	3	4	5	6	6	6	2	5	2	4	~40
Lu	0.57	0.65	0.70	0.47	0.58	0.29	0.39	0.82	0.68	0.79	0.37	0.19	0.23	0.58	12
Hf	13	19	12	9	21	21	3.9	24	10	18	13	14	5.7	15	14**
Ta	12.2	15.1	13.2	10.4	14.0	13.1	5.7	21	8.5	16.8	20	29	21	11.3	7
Th	15.9	22.9	16.9	13.6	22.9	22.9	6.3	42.8	17.0	30.1	28.0	26.9	2.4	14.5	4
U	3.5	5.9	3.9	3.1	5.6	5.7	2.0	10.2	4.7	8.8	5.7	7.2	2.0	3.8	8
Eu/Eu*	0.94	0.89	1.65	1.16	1.14	2.00	1.74	1.25	0.92	1.12	0.58	0.56	0.51	0.74	

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rocks of these iron-rich feldspars (Carmichael, 1960, 1967; Nash *et al.*, 1969; Sigurdsson, 1971) all have peralkaline affinities $(\text{Na} + \text{K})/\text{Al} \geq 0.95$). In such rocks Fe^{3+} will tend to substitute for Al because of the low activity of the latter (Rosenqvist, 1951). However, no such concentration of iron was found in the feldspars of the strongly silica-undersaturated, peralkaline foyaite/hedrumites 488 and PØ-77.

These new microprobe determinations of Fe_2O_3 in larvikite thus agree with those reported by Muir & Smith (1956), but are considerably lower than the value obtained by Oftedahl (1948) on the basis of wet chemical analysis of separated feldspars (Table 2). Similar discrepancies between electron microprobe analyses and analyses of mineral separates have been reported by Ribbe & Smith (1966), and Smith & Ribbe (1966). This is due to the fact that the feldspars contain inclusions of opaques (2–4 per cent) (occasionally also of olivine and clinopyroxene). The opaques occur mainly as scattered single grains of euhedral to anhedral Fe–Ti oxides (ilmenite–haematite_{ss} and magnetite–ulvöspinel_{ss}) of varying size. Less frequently the opaques form bands of droplike inclusions with a diameter of a few microns ($\ll 1$ per cent), which seem to consist only of iron oxide. Whereas the single grains are undoubtedly inclusions of primary Fe–Ti-oxides, the droplike opaques probably represent iron exsolved from the feldspar structure. Although the initial Fe_2O_3 -contents of the feldspars were therefore probably slightly higher than the values reported in Table 2, the microprobe analyses clearly disprove the hypothesis of Bailey & Schairer (1966) that the different evolutionary trends are related to differences in the Fe_2O_3 -concentration in the feldspars.

Rock analyses

Major and trace element data on the analyzed rocks are listed in Table 3. For each ring-section, the samples are listed in the order of increasing differentiation index (D.I. = normative $q + \text{or} + \text{ab} + \text{ne}$).

The analyzed samples have low concentrations of the compatible elements Sc, Ni, Cr, and Co, are rich in the excluded elements Rb, Cs, SIE (Hf, Ta, Th, U) and

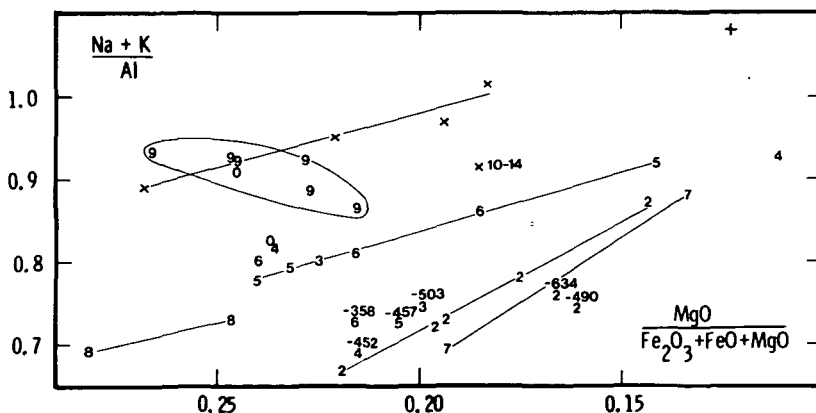


FIG. 4. Agpaite index $((\text{Na} + \text{K})/\text{Al})$ plotted against $\text{MgO}/(\text{Fe}_2\text{O}_3 + \text{FeO} + \text{MgO})$. Samples falling markedly below their ring-segment trend are marked by complete sample numbers. Symbols as in Fig. 3.

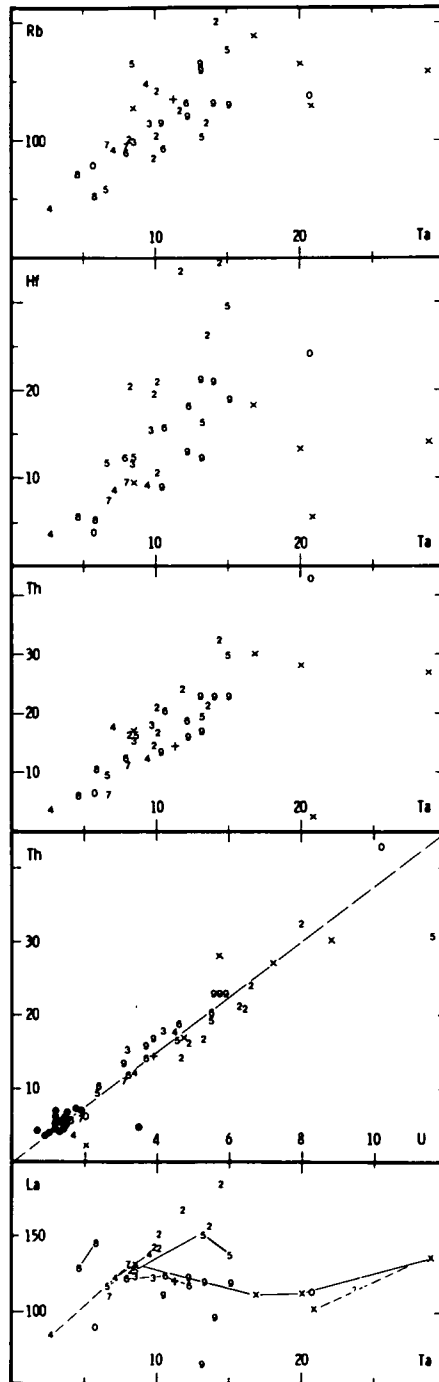


FIG. 5. Incompatible element (ppm) variations among the Larvik rocks. Symbols as in Fig. 3, except for the filled circles in the Th-U diagram which represent basaltic rocks from the Skien-Vestfold area (analyzed by Weigand, 1975). The line gives the mean Th/U ratio of 137 larvikites in the Larvik pluton (analysed by Raade, 1973).

REE, and in light relative to heavy REE, and thus have trace element concentrations typical of highly evolved alkaline rocks.

The rocks in the Larvik ring-complex vary considerably in $MgO/(Fe_2O_3 + FeO + MgO)$ ratio and P_2O_5 -concentrations relative to D.I. (Fig. 3), and in agpaitic index $((Na + K)/Al)$ relative to $MgO/(Fe_2O_3 + FeO + MgO)$ (Fig. 4), and can thus hardly belong to one single cogenetic rock series. As shown below, the analytical data rather suggest the existence of several separate evolutionary trends.

Larvikites

The larvikites in different ring-sections have overlapping ranges of D.I. and incompatible element concentrations (Table 3, Figs. 3, 5, 6). Within each ring-section the 'least evolved' rocks (that is, those with the lowest D.I. and highest MgO - and CaO -concentrations and $MgO/(Fe_2O_3 + FeO + MgO)$ ratios) also have least normative quartz or nepheline, the lowest Rb - and Sr -concentrations, and have semi-linear REE_{sample}/REE_{chondrite} patterns with no, or weak positive Eu -anomalies (Table 3, Fig. 6). With increasing D.I. the MgO -, CaO - and Sr -concentrations, and the $MgO/(Fe_2O_3 + FeO + MgO)$ ratios decrease; incompatible element concentrations increase; REE-patterns become weakly concave with negative

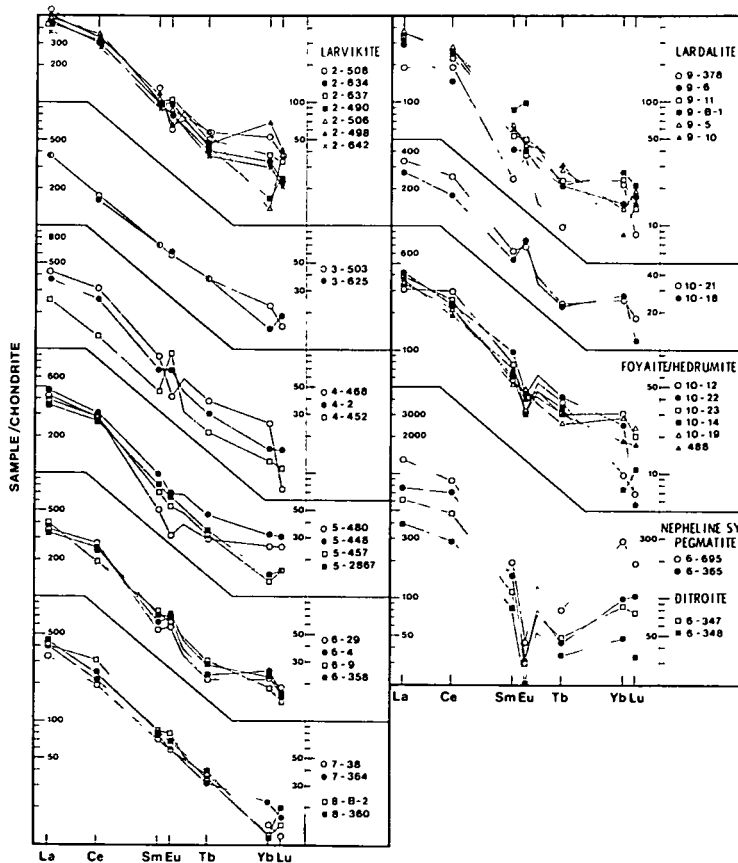


FIG. 6. Chondrite normalized REE patterns for rocks from the different ring-sections.

Eu-anomalies; and the rocks in RS-2 become more strongly silica *oversaturated*, whereas those in RS-5 and RS-6 become more strongly *undersaturated* (Table 3, Fig. 6). In the apaitic index versus $\text{MgO}/(\text{Fe}_2\text{O}_3 + \text{FeO} + \text{MgO})$, the AFM, and the Fo–Q–Ne diagrams (Figs. 4, 7 and 8), the rocks tend to define separate trends.

The significance of the major element variations within each ring-section was tested by petrographic mixing modelling. The results of these mixes are used to calculate theoretical trace element enrichment factors ($\text{element}_{\text{daughter}}/\text{element}_{\text{parent}}$) as outlined above. The results are listed in Table 4 together with observed trace element enrichment factors. The compositional variations are roughly consistent with a closed system model involving fractionation/accumulation of feldspar plus minor amounts of olivine \pm clinopyroxene + ore + apatite. The variable, and in some cases, excessively high observed, relative to calculated, enrichment factors for

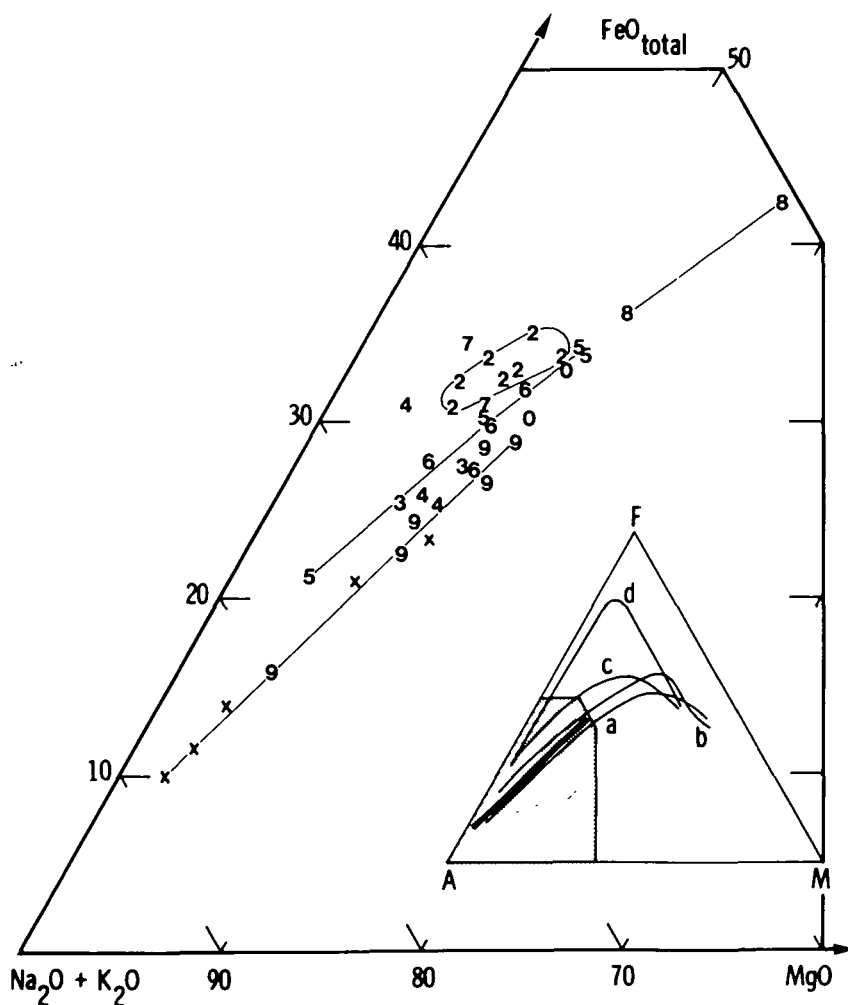


FIG. 7. Alkali– $\text{FeO}_{\text{total}}$ –MgO variations among the Larvik rocks. Symbols as in Fig. 3. In the inserted figure the composite Larvik plutonic trend is marked by a thick line, whereas the thin lines represent the chemical variations among (a) highly alkaline lavas and (b) plutonic rocks of Tahiti (McBirney & Aoki, 1968); (c) undersaturated alkaline volcanic rocks from East Otago, New Zealand (Coombs & Wilkinson, 1969); and (d) tholeiitic rocks in the Skaergaard intrusion (Carmichael *et al.*, 1974).

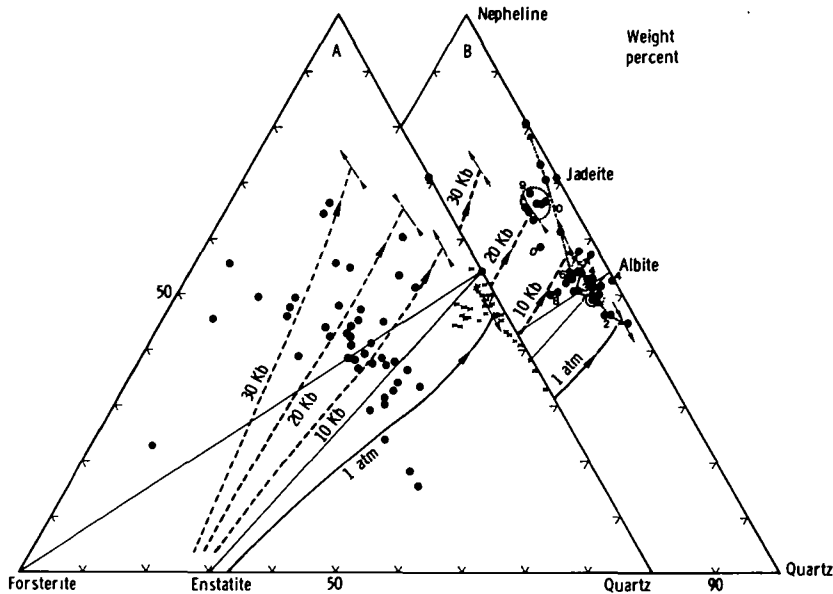


FIG. 8. Whole rock analyses projected into the nepheline–forsterite–quartz compositional plane, together with the forsterite–enstatite liquidus boundary at 1 atm (Schairer & Yoder, 1961), and 10, 20 and 30 kb (Kushiro, 1968). (A) Basaltic rocks (dots) analyzed by Weigand (1975), and a larvikite–nordmarkite–ekerite series (crosses) from the Kikut pluton, analyzed by R. Wilson and M. P. Annis (unpublished data). (B) Rocks from the Larvik pluton. Samples from the same ring-section are encircled (dashed lines) or joined (dash-dot lines), and marked with the ring-section number. Arrow-heads indicate directions of increasing D.I.

SIE indicate, however, that the petrographic mixing is an oversimplification of the crystallization history of the rocks, which involved re-equilibration of the feldspars during cooling, some crystallization of minor phases like zircon, and probably also magma mixing (discussed below) and other processes.

It is not immediately obvious whether the ‘least evolved’ larvikites may be regarded as approximately representative of melt compositions, or if they have suffered some accumulation of feldspar. The larvikites obviously represent a stage of transition from weakly positive to negative Eu-anomalies. During crystallization of an intrusive body of intermediate composition, a certain degree of feldspar settling may be expected, causing relative depletion of Eu in those parts from which feldspar has been separated, and relative enrichment where feldspar has accumulated. Accumulation of feldspar would also lead to a decrease in the concentrations of incompatible elements (by dilution) and an increase in Sr (plagioclase), Sr and Ba (anorthoclase), or Ba (K-feldspar) (Arth, 1976; Sun & Hanson, 1976; Irving, 1978).

Weak positive Eu-anomalies may, however, also be induced in a magma. REE data on natural rock series indicate that during the early and intermediate stages of fractional crystallization the effect of preferred removal of Eu by plagioclase is counteracted by, and sometimes even exceeded by the effect of rejection of Eu by simultaneously crystallizing clinopyroxene, apatite and/or kaersutite with negative Eu-anomalies (Zielinski & Frey, 1970; Price & Taylor, 1973; Zielinski, 1975; Sun & Hanson, 1976; Kyle & Rankin, 1976; Baker *et al.*, 1977).

TABLE 4

Representative mixing calculations tested on trace elements

Parent	Daughter	Larvikites		Foyatie/Hedrumites		Lardalites	
		Obs.	Calc.	Obs.	Calc.	Obs.	Calc.
2-642	508	4-2	468	9-10	5	9-10	B-1
4-2	452	5-2867	480	10-23	14	10-23	378
6-4	29	6-4	29	10-23	14	10-23	378
11	± 0.16	± 0.16	± 0.16	± 0.05	± 0.09	± 0.09	± 0.05
-11	0.56	0.56	0.56	0.90	0.88	0.88	0.98
+1	(1.34)	(1.34)	(1.34)	± 0.03	± 0.03	± 0.03	± 0.26
+95	± 0.04	± 0.04	± 0.04	± 0.04	± 0.04	± 0.04	(1.13)
-85	0.84	0.84	0.84	0.84	0.84	0.84	(0.98)
-86	± 0.03	± 0.03	± 0.03	± 0.03	± 0.03	± 0.03	(0.98)
-4	0.64	0.64	0.64	0.64	0.64	0.64	(0.98)
-2	± 0.03	± 0.03	± 0.03	± 0.03	± 0.03	± 0.03	(0.98)
-4	0.64	0.64	0.64	0.64	0.64	0.64	(0.98)
-2	± 0.03	± 0.03	± 0.03	± 0.03	± 0.03	± 0.03	(0.98)
0.84	± 0.04	± 0.04	± 0.04	± 0.04	± 0.04	± 0.04	(0.98)
October	October	October	October	October	October	October	October
30, 2014	30, 2014	30, 2014	30, 2014	30, 2014	30, 2014	30, 2014	30, 2014
Trace element enrichment factors (element daughter/element parent).	Trace element enrichment factors (element daughter/element parent).	Trace element enrichment factors (element daughter/element parent).	Trace element enrichment factors (element daughter/element parent).	Trace element enrichment factors (element daughter/element parent).	Trace element enrichment factors (element daughter/element parent).	Trace element enrichment factors (element daughter/element parent).	Trace element enrichment factors (element daughter/element parent).
Rb	1.4	1.4	1.4	1.4	1.4	1.4	1.4
Sr	0.3	0.3	0.3	0.3	0.3	0.3	0.3
Ba	0.47	0.47	0.47	0.47	0.47	0.47	0.47
La	1.3	1.3	1.3	1.3	1.3	1.3	1.3
Ce	1.2	1.2	1.2	1.2	1.2	1.2	1.2
Sm	1.4	1.4	1.4	1.4	1.4	1.4	1.4
Eu	0.7	0.7	0.7	0.7	0.7	0.7	0.7
Tb	1.7	1.7	1.7	1.7	1.7	1.7	1.7
Yb	1.7	1.7	1.7	1.7	1.7	1.7	1.7
Lu	1.8	1.8	1.8	1.8	1.8	1.8	1.8
Hf	1.7	1.7	1.7	1.7	1.7	1.7	1.7
Th	1.4	1.4	1.4	1.4	1.4	1.4	1.4
U	1.6	1.6	1.6	1.6	1.6	1.6	1.6

¹ Per cent relative to initial system.

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In the agpaite index ((Na + K)/Al) versus MgO/(Fe₂O₃ + FeO + MgO) diagram (Fig. 4), some larvikites (2-490, 2-634, 3-625, 4-452, 5-457, and 6-358) plot considerably to the low (Na + K)/Al side of their ring-segment-trend. Most of these samples also have low SIE and high Sr and Ba concentrations, and Eu/Eu* \geq 1.0, or higher than those in other rocks with similar MgO/(Fe₂O₃ + FeO + MgO) ratios (Table 3). It is thus strongly suggested that these samples contain cumulate feldspar. Other rocks with Eu/Eu* > 1.0, like 6-4 and 6-29, may well be normal members of fractionation series (Table 4).

In ring-sections 5, 6 (and 10) marked increases in SIE-concentrations and D.I. are accompanied by moderate or no increases in REE-concentrations. This fact (illustrated for La/Ta in Fig. 5) indicates that REE in many larvikites (and in the foyaite/hedrumites) have bulk distribution coefficients approaching unity, and can thus not be regarded automatically as incompatible elements. The petrographic mixing calculations imply that fractionation of apatite is responsible for removal of REE from the melts, and for the development of weakly concave REE-patterns observed in some of the highly evolved larvikites, and in the analyzed ditroites and nepheline syenite pegmatites (Table 4, Fig. 6). The importance of apatite in controlling REE abundances and patterns in differentiates has earlier been discussed by Zielinski & Frey (1970), Price & Taylor (1973), Zielinski (1975) and Sun & Hanson (1976).

The analyzed samples of the quartz-bearing larvikite-variety tønbergite seem to represent normal members of a fractionated series of rocks (Table 3 and 4; Figs. 3–8). No indications of metasomatic enrichment of SiO₂, K₂O, Th or U, as suggested by Raade (1973, 1978), were found.

Ditroites (gneissic nepheline syenites) and nepheline syenite pegmatites

The ditroites and nepheline syenite pegmatites in ring-segment 6 are strongly enriched in incompatible elements, including Rb and REE, and depleted in Sr and Ba relative to the larvikites (Table 3). Their chondrite normalized REE-patterns are strongly concave with pronounced negative Eu-anomalies (Fig. 6), compatible with a similar, but higher degree of evolution than the most evolved larvikites.

Lardalites

The strongly silica-undersaturated lardalites have significantly higher MgO/(Fe₂O₃ + FeO + MgO) ratios and P₂O₅ concentrations than the larvikites, but have similar concentrations of incompatible elements, Sr and Ba (Table 3, Figs. 3 and 5). A *positive* correlation exists between agpaite indexes and MgO/(Fe₂O₃ + FeO + MgO) (Fig. 4). Chondrite normalized REE-patterns are concave with weakly negative to strongly positive Eu-anomalies (Fig. 6). High positive Eu-anomalies tend to be accompanied by relatively high Sr and/or Ba, low SIE-concentrations, and low agpaite indexes (Table 3).

The RS-9 lardalites show a positive correlation between P₂O₅ and REE-concentrations (Table 3) demonstrating strong control of apatite over REE. In most diagrams, however, the lardalites tend to scatter, rather than to define trends. SIE/SIE ratios are fairly constant, whereas the REE/SIE ratios vary considerably

(Fig. 5). As the lardalites have higher $\text{MgO}/(\text{Fe}_2\text{O}_3 + \text{FeO} + \text{MgO})$ ratios than the larvikites, overlapping ranges of incompatible element concentrations, and positive Eu-anomalies (Table 3, Figs. 3, 5 and 6), they can hardly be formed from a larvikitic magma by fractionation, or by partial melting of crystallized larvikite. The petrographic mixing calculations (Table 4) suggest instead that the lardalites may be related to one another through a complex scheme of addition and/or subtraction of feldspar, apatite and feric minerals in different proportions (using sample 9-10 with the lowest D.I. as a reference), thus strongly suggesting that accumulation has been an important factor in their formation. The accumulation hypothesis cannot be tested quantitatively as the composition of the melt into which minerals have accumulated is unknown. The geochemical data presented above are, however, roughly consistent with accumulation in a strongly undersaturated melt highly enriched in incompatible elements, with a concave REE-pattern, and probably with a negative Eu-anomaly. Negative Eu-anomalies are found in the other strongly undersaturated rocks in the area, and would explain why two of the lardalites have weak *negative* Eu-anomalies although they supposedly have suffered some accumulation. It is unlikely that the lardalites are directly related to the foyaite/hedrumites which appear to be younger (Petersen, 1978), and to have crystallized under more oxidizing conditions (Neumann, 1976). In contrast to the lardalites, the foyaite/hedrumites have highly variable ratios between Hf, Ta, Th, and U. It is, however, possible that some of the nepheline syenite pegmatites in the area were squeezed out from the crystal mush which gave rise to the lardalites.

Foyaite/hedrumites

The foyaite/hedrumites have 3 to 38 per cent normative nepheline (Table 3). Increasing degree of undersaturation is accompanied by decreasing $\text{MgO}/(\text{Fe}_2\text{O}_3 + \text{FeO} + \text{MgO})$ ratios and MgO-, CaO-, TiO_2 -, P_2O_5 -, Sr- and Ba-concentrations, and increasing Rb/Sr and $(\text{Na} + \text{K})/\text{Al}$ ratios (Figs. 3 and 4). Increasing degree of undersaturation is also accompanied by a change in chondrite-normalized REE patterns from concave with $\text{Eu}/\text{Eu}^* \sim 1$, to convex with $\text{Eu}/\text{Eu}^* \ll 1$ (Fig. 6). The SIE, however, show a rather erratic behaviour (Table 3, Fig. 5). A positive correlation between these elements and D.I. is observed among the least evolved foyaite/hedrumites, while in the most highly evolved ones SIE are relatively depleted.

Although the differentiation modelling supports the qualitative geochemical evidence that the foyaite/hedrumites probably are related to each other through fractionation of feldspar plus Fe–Mg-silicates. Fe–Ti-oxides and apatite, their compositional variations have not been successfully reproduced (Table 4). This may signify that these rocks do not represent a strictly cogenetic series, or that marked variations in mineral compositions (particularly aegirine-augites and sodic amphiboles) have not been carefully enough simulated. The relative depletion of heavy REE and SIE in the most highly evolved rocks is probably caused by separation of sphene and/or zircon which occur in the foyaite/hedrumites as a major and a minor accessory mineral respectively. Both minerals have strong affinities for elements with high charges and moderately large ionic radii (Kostov,

1968; Nagasawa, 1970). Separation of sphene and zircon may thus effectively deplete a differentiate in such elements. A wide variety of minerals rich in Zr, Nb, SIE and REE (including cappelinite, chevkinite, eucolite, karyocerite, l avenite, perrierite, polymignite, mosandrite, and thorite) has been described from nepheline syenite pegmatites in the western part of the Larvik pluton (Br ogger, 1890; Segalstad & Larsen, 1978*a, b*), testifying to high activities of these elements in the strongly undersaturated residual melts.

Incompatible element ratios

Ratios between strongly incompatible elements (SIE) do not tend to change significantly during partial melting or fractionation (*e.g.* Gast, 1968; Shaw, 1970; Ferrara & Treuil, 1974). The degree of spread in SIE ratios between different fractionated rock series is therefore taken to reflect the degree of heterogeneity in their mantle source region (*e.g.* Ferrara & Treuil, 1974; Hanson, 1977).

With the exception of the most highly evolved foyaite/hedrumites, the analyzed rocks have fairly uniform ratios between Ta, Th and U whereas Rb/Ta and Hf/Ta ratios vary moderately, and REE/SIE markedly (Fig. 5). This finding is in agreement with Raade's (1973) comprehensive study of Th–U–K variations among the Oslo Region plutonic rocks. He found low uniform Th/U ratios among the least evolved plutonic rocks (including larvikites and lardalites), and higher and more variable ratios among the more highly evolved rocks (nordmarkites, ekerites and biotite granites). One hundred and thirty-seven larvikites (including t onsbergites) from the Larvik pluton have a mean Th/U = 3.80 ± 0.40 , 12 lardalites: 3.72 ± 0.10 , 10 foyaite/hedrumites: 4.23 ± 1.05 (analyzed by Raade, 1973, 1978). Basaltic lavas from Skien and Vestfold have similar, but rather variable Th/U ratios (Weigand, 1975). The mean of 16 samples is 4.21 ± 1.21 .

A slight tendency for the quartz-bearing larvikites from RS-2 to fall off the main trends in the SIE diagrams (Fig. 5) has been tested on the basis of Raade's Th/U data. These data show no significant differences in Th/U ratios between rocks from different ring-segments. The difference in Hf/Ta ratios between RS-2 rocks and other rocks are somewhat more pronounced than those exhibited by Th/U ratios (Fig. 5), but the number of analyses is too limited to make the data conclusive. It is also clear from the previous discussion that extensive fractionation of REE relative to SIE was produced by late stage separation of apatite. At most there can have been weak variations in REE/SIE ratios prior to intrusion.

Magma densities

Magma densities for the Oslo Region magmatic rocks have been calculated on the basis of chemical analyses after the method of Bottinga & Weill (1970), assuming that rock \approx magma, a temperature of 1100 °C, dry conditions for the basaltic rocks and 1050 °C, dry conditions for the intermediate rocks. The results for the basaltic rocks are presented in Fig. 9. Bottinga & Weill (1970) estimated magma density to decrease about 0.01 g/cm³ per 100 °C temperature increase, and to decrease 0.02 g/cm³ on addition of 1.3 mole per cent H₂O at 1 kb to a 'dry' basaltic magma. Small variations in magma temperature and water content will

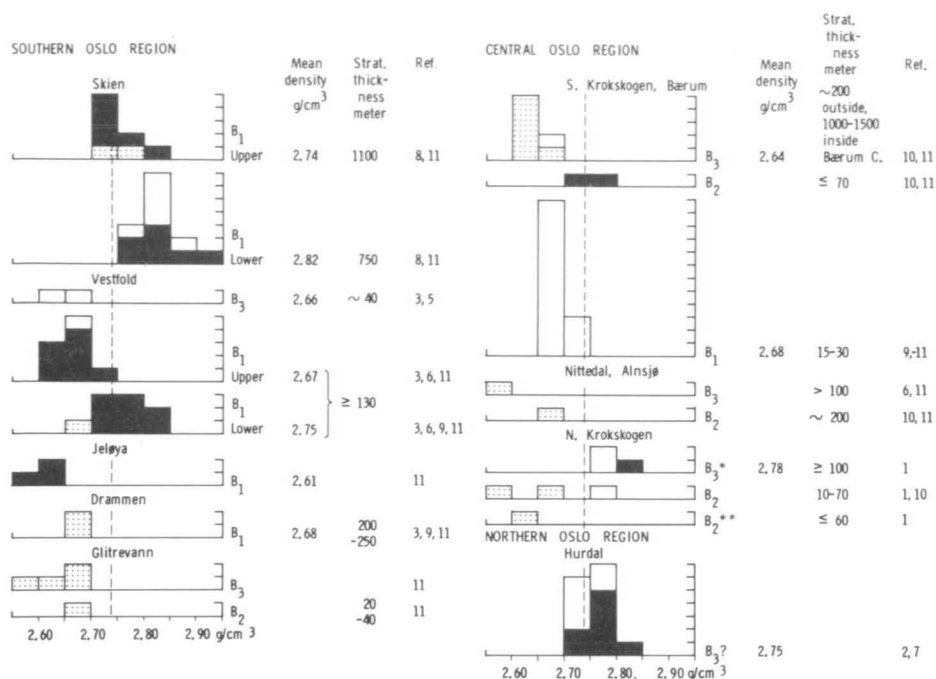


FIG. 9. Variations in magma densities among the Oslo Region basaltic rocks, calculated after a method by Bottinga & Weill (1970) assuming 1100 °C, dry conditions. The colours indicate the dominating phenocryst phase: white = aphyric basalt, light gray = plagioclase b., dark gray = plagioclase-pyroxene b., black = pyroxene b. Chemical analyses and lava stratigraphy by ¹Larsen, 1978; ²Nystuen, 1975; ³, ⁴Oftedahl, 1952, 1953; ⁵Oftedahl & Petersen, 1978; ⁶, ⁷Ramberg & Larsen, 1978a, b; ⁸Segalstad, 1978; ⁹Sundvoll, pers. comm., 1978; ¹⁰Sæther, 1962; ¹¹Weigand, 1975. ~B₃*: at B₃ stratigraphic position. <B₂**>: below B₂ stratigraphic position.

thus not affect calculated magma densities seriously. The highest calculated densities were obtained for aphyric basalts, and basalts mainly with clinopyroxene phenocrysts. As no corrections have been made for phenocryst contents, calculated clinopyroxene basalt densities are minimum values for the erupted material.

Sixty-three analyses of rhomb porphyries and larvikites analyzed by Brøgger, 1933; Neumann, 1976; this paper) give a mean magma density of 2.48 ± 0.04 g/cm³ at 1050 °C, dry. Assuming 40 per cent feldspar phenocrysts of the composition An₄₀Ab₅₃Or₇ (simulating some of the rhomb porphyries), the density increases to 2.55 g/cm³ (Neumann, 1978).

DISCUSSION

Pressures of fractionation

The geochemical data described above imply that each ring-segment in the Larvik pluton represents the intrusion of (at least) one batch of magma of intermediate composition, which underwent some fractionation *in situ*. This conclusion is supported by compositional variation trends among the clinopyroxenes (Neumann, 1976).

The near saturated character of the least evolved rocks in the Larvik pluton is typical of all the Oslo rift rocks of 'intermediate' composition. Larvikites contain small amounts of nepheline or quartz; both minerals may be absent (Brøgger, 1890, 1933; Barth, 1945; Sæther, 1962; Raade, 1973; Neumann, 1976, 1978). Also the rhomb porphyries are approximately silica-saturated; small amounts of normative quartz are somewhat more common than small amounts of normative nepheline (Brøgger, 1933; Oftedahl, 1946, 1978; Ramberg & Larsen, 1978a). The nordmarkites (including Grefsen syenite and pulaskite) range from quartz-free to quartz-rich types (Brøgger, 1890, 1933; Barth, 1945; Sæther, 1945; Raade, 1973; Neumann, 1976). Transitions from near-saturated larvikites (or nordmarkites), to mildly or strongly oversaturated more highly evolved rocks (as found in the Larvik pluton), have been observed in plutons in all parts of the Oslo Region (Brøgger, 1890; Barth, 1945; Oftedahl, 1953, 1978; Sæther, 1962; Raade, 1973; Nystuen, 1975; Sørensen, 1975; Neumann, 1976; M. P. Annis and R. Wilson pers. comm., 1977; Neff & Khalil, 1978).

The basaltic rocks, on the other hand, which are found partly below (B_1), and partly alternating (B_2 - B_3) with the rhomb porphyries, range in composition from strongly undersaturated alkali basalts to quartz tholeiites (Fig. 5). In the Skien district (W of the Larvik pluton) only B_1 basalts occur, consisting mainly of a thick pile of strongly to mildly alkaline basalts and basanites (Weigand, 1975; Segalstad, 1978). In Vestfold both B_1 and younger basalts are mildly alkaline (Oftedahl, 1978; Oftedahl & Petersen, 1978). In the northern part of the Oslo Region, B_1 is represented by a single flow of quartz tholeiite, whereas the younger flows (B_2 - B_3) consist mainly of mildly alkaline basalts, but include flows of olivine tholeiite (Weigand, 1975; Ramberg & Larsen, 1978a).

With the exception of quartz tholeiite (B_1) from Krokskogen, with its continental plateau type trace element geochemistry, the basaltic rocks, the larvikites and the rhomb porphyry lavas have similar chondrite-normalized REE-patterns, Th/U and low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Finstad, 1972; Sundvoll, 1978a, b). This is compatible with derivation of the intermediate rocks by fractional crystallization of mantle-derived basaltic magmas, or by remelting of gabbros.

Experimental partial melting of materials of assumed mantle compositions (quartz peridotite, picrite and mantle eclogite) (e.g. Green & Ringwood, 1967; Kushiro, 1968; Green, 1973; Mysen & Boettcher, 1975; Mysen & Kushiro, 1976; Stern & Wyllie, 1978) indicates that direct derivation of the intermediate rocks by partial melting in the mantle is an unlikely mechanism.

Although no single basalt flow may be assumed to be the parent magma of any of the evolved rocks, it is most likely that the mildly alkaline to transitional nature of the majority of extruded *mafic lavas* is representative of the compositional range of *the mafic magmas that invaded the upper mantle and crust* during the period of differentiation of the evolved rocks. The roughly silica-saturated nature of the rhomb porphyries and larvikites therefore strongly suggests that the intermediate magmas that crystallized to form these rocks evolved under such conditions that melts with a range of alkali olivine (and olivine tholeiitic) basaltic compositions would differentiate to yield silica-saturated felsic derivatives of fairly uniform

compositions. This is only possible under a limited range of moderately low pressures. Experimental investigation of the diopside–forsterite–anorthite system (Presnall *et al.*, 1978) has demonstrated that between about 4 and 12 kb, alkali basalt can differentiate to tholeiitic basalt by fractional crystallization of phase assemblages including aluminous clinopyroxenes. At lower pressures a thermal divide exists between alkaline and tholeiitic basalts (represented in a simplified manner by the diopside–forsterite–anorthite join). At pressures above about 12 kb it may be possible to produce silica-undersaturated derivatives from tholeiitic basalts.

Moderately low pressures are also required for fractionation of olivine tholeiitic melts to yield *silica-saturated* felsic residues. Experimental studies in relevant synthetic systems ($\text{Mg}_2\text{SiO}_4\text{--SiO}_2\text{--X}$, where X is CaMgSiO_4 , CaAl_2O_4 or NaAlSiO_4) indicate that under dry conditions at $P = 1$ atm the Fo–En_{ss} liquidus boundary lies in the field of silica-oversaturation (Andersen, 1914; Bowen, 1914; Schairer & Yoder, 1961, 1962). At higher pressures the liquidus boundary shifts away from silica, and enters the field of silica-saturation at 5–7 kb in the $\text{Mg}_2\text{SiO}_4\text{--SiO}_2\text{--NaAlSiO}_4$ system (Fig. 8), and at about 10 kb in the $\text{Mg}_2\text{SiO}_4\text{--SiO}_2\text{--CaMgSiO}_4$ and $\text{Mg}_2\text{SiO}_4\text{--SiO}_2\text{--CaAl}_2\text{O}_4$ systems (Kushiro, 1968). Higher pressures yield silica-undersaturated felsic derivatives in these systems. The position of phase boundaries in *natural* systems may be roughly estimated on the basis of experimental investigations of the systems $\text{MgO--SiO}_2\text{--X}$, $\text{CaO--SiO}_2\text{--X}$, $\text{FeO--SiO}_2\text{--X}$, Fo--Q--Di--Y , and Fo--Q--Ne--Y at 1 bar and 20 kb, where X represents major and minor oxides, and Y H_2O , CO_2 or P_2O_5 (Kushiro, 1972–1974; Eggler, 1974). Kushiro and Eggler find that the presence of oxides of monovalent cations (H_2O , K_2O , Na_2O) causes the liquidus field of forsterite to expand relative to that of enstatite_{ss}, and those of enstatite_{ss}, pseudowollastonite and fayalite to expand relative to silica minerals, whereas the presence of oxides of polyvalent cations (CO_2 , TiO_2 , P_2O_5) has the opposite effect. The oxides of divalent cations and Al_2O_3 only have a minor effect. These conclusions are strongly supported by Wendlandt's (1977) investigation of the $\text{Mg}_2\text{SiO}_4\text{--SiO}_2\text{--KAlSiO}_4$ system, and by the melting experiments of Mysen & Boettcher (1975) on mantle peridotite at various H_2O and CO_2 partial pressures. There is thus in principle a wide range of magma composition—volatile partial pressure—total pressure conditions that will allow the salic part of the olivine–enstatite liquidus boundary to fall within the field of silica-saturation. Natural basalts (and nephelinites) from different localities, however, show a rough positive correlation between K_2O (+ Na_2O), P_2O_5 and TiO_2 (e.g. Green & Ringwood, 1967; Green, 1971; Hekinian, 1974; Sun & Hanson, 1975; Pearce, 1976), and contain a complex volatile phase including both H_2O and CO_2 . It may therefore be assumed that the effects of mono- and polyvalent components will tend to cancel each other, leaving the behavior of the $\text{Mg}_2\text{SiO}_4\text{--SiO}_2\text{--CaMgSiO}_4\text{--CaAl}_2\text{O}_4\text{--NaAlSiO}_4$ system under dry conditions as acceptable first approximations to those of most natural basaltic systems.

During fractional crystallization Na and K are enriched in the residual magma until the last stages of differentiation, whereas the concentrations of P and Ti generally begin to decrease at an earlier stage, with the onset of crystallization of

apatite and Fe–Ti-oxides (*e.g.* Ferrara & Treuil, 1974, Leeman *et al.*, 1976, Baker *et al.*, 1977). The solubility of CO_2 is, furthermore, much lower than that of H_2O in silicic melts, although it is higher in H_2O -rich than in H_2O -poor melts (Eggler *et al.*, 1974). It is therefore likely that the Fo–En_{ss} liquidus boundary will tend to shift towards higher silica (as compared with the experimentally determined ones for simple systems) during advanced stages of fractionation. Such a tendency seems to be followed by the Oslo Region intermediate rocks. The larvikites and rhomb porphyry lavas have on the average 47 ± 6 per cent normative albite, and nepheline is typically found in the undersaturated rocks. These rocks may therefore best be discussed on the basis of the Fo–Q–Ne synthetic system (Fig. 8). The larvikites (and probably also the rhomb porphyries) appear to have crystallized under fairly dry conditions (Neumann, 1976), but they are highly enriched in K_2O relative to TiO_2 and P_2O_5 (Table 3). It is therefore likely that the olivine–pyroxene liquidus boundary in these natural intermediate systems at any given pressure will lie on the silica-rich side of that determined for the simple Fo–Q–Ne system at the same pressure (Fig. 8).

We are interested in the pressure range within which the low temperature part of the Fo–En_{ss} liquidus boundary and invariant point lies in, or very close to, the field of silica-saturation (bounded by the Mg_2SiO_4 – $\text{NaAlSi}_3\text{O}_8$ and MgSiO_3 – $\text{NaAlSi}_3\text{O}_8$ joins in Fig. 8). By simple interpolation between the liquidus boundaries determined for 1 atm and 10 kb (Fig. 8), it is concluded that this requirement in the pure Fo–Q–Ne system is met by a pressure range of roughly 5–8 kb. According to the above discussion, slightly higher pressures, *ca.* 7–10 kb, are probably more realistic for the formation of the near silica-saturated intermediate residual melts in the Oslo Region.

It is an interesting fact that no undersaturated rocks are found among the youngest intrusions in the Oslo Region. This fact may reflect the effect of a further relative increase of monovalent cations towards the end of the period of magmatism, but may equally well signify that the ‘magma accumulation’ region in the lower crust expanded to shallower depths with time. Not enough data are presently available to test these possibilities.

According to Ramberg’s (1976) density distribution model for the crust below the Oslo Region, 7 and 10 kb correspond to depths of 24 and 35 km.

The regional positive gravity anomaly associated with the Oslo rift has been described and discussed in detail by Ramberg (1976). He concludes that the gravity anomaly is caused by a body of dense material in the crust below the Oslo Region. Ramberg’s preferred structural model places the dense body in the lower crust with its top and bottom at about 20 and 34 km (Moho) depths respectively, and a mean density contrast to the surrounding lower crust of 0.1 g/cm^3 . Ramberg (1976) has also proposed that the dense material may represent cumulates. The estimated depth range for the fractionation leading to evolved approximately saturated melts supports Ramberg’s suggestion that the dense body is situated at the base of the crust, and that it partly or mainly consists of cumulates.

The small differences in silica activities and $\text{MgO}/(\text{Fe}_2\text{O}_3 + \text{FeO} + \text{MgO})$ ratios that appear to have existed in the evolved magmas prior to intrusion into the upper

crust, may be inherited from the mafic parent magmas or induced by small differences in total pressure, $P_{\text{H}_2\text{O}/\text{CO}_2}$ and f_{O_2} during fractionation, thus implying derivation from a series of magma chambers in the lower crust, rather than one major homogeneous body of magma. The effect of varying magma compositions, P_{total} and $P_{\text{H}_2\text{O}/\text{CO}_2}$ on the position of the liquidus boundary has been discussed above. The degree of iron-enrichment in a magma is strongly dependent upon f_{O_2} (Osborn & Watson, 1977). The possible influence of f_{O_2} on the degree of iron-enrichment in the Larvik rock series is indicated by the fact that the foyaites/hedrumites which fall on the trend of lowest iron-enrichment (Fig. 7) also seem to have crystallized under the most oxidizing conditions (Neumann, 1976). It does not, however, explain the similarly low degree of Fe-enrichment in the lardalites which appear to have crystallized under reducing conditions (QFM-buffer) (Neumann, 1976).

A process involving crystallization of early-formed basaltic magmas near the mantle/crust boundary and subsequent partial remelting to form felsic melts cannot be discarded on a geochemical basis. This model would, however, require an abnormally high geothermal gradient and large supplies of heat during the period of partial melting. Such conditions would be created during the period when basaltic magmas intruded the upper mantle and lower crust (and fractionated there), as outlined above. Partial remelting of gabbro may therefore have given rise to some intermediate melts, but is unlikely as the main mechanism.

Fractionation apparently continued *after* intrusion of the felsic magmas into the upper crust. This low pressure fractionation of weakly over- and weakly under-saturated melts gave rise to more strongly over- and under-saturated derivatives respectively.

The lardalites formed by accumulation of feldspar, nepheline, Fe–Mg-silicates and apatite in a highly evolved, strongly under-saturated melt.

The evolutionary history of the intermediate rocks outlined above is also supported by the feldspar data. Ternary feldspars as found in the larvikites and nordmarkites are only stable at low fluid pressures (Tuttle & Bowen, 1958), and must have formed *after* emplacement of the magmas into the upper crust. With increasing P_{fluid} , the miscibility between An–Ab and Or decreases, and at 4.25 kb the solvus and solidus intersect on the Or–Ab(–H₂O) join (Yoder *et al.*, 1957; Morse, 1970). The intermediate magmas would thus have plagioclase on the liquidus during fractionation in the lower crust. During ascent to the upper crust these magmas might become superheated, resulting in partial or total resorption of the plagioclase before the onset of crystallization of ternary feldspar at shallow depths.

Crustal density filtering

The above discussion suggests that in the Oslo Region, fractionation leading to magmas of approximately silica-saturated intermediate compositions took place in the lower crust. It does, however, not explain *why* fractionation was located there, or *which factors* determined the degree of evolution of each batch of magma before ascent to the upper crust. The following data bear on these problems:

1. The Oslo Region, like many other continental rifts, has a high percentage of surface and subsurface rocks of intermediate and acid compositions.
2. The tectonic style in the Oslo Region changed from linear high-extension types (fissure eruptions, rift valley formation) during the early stages, to circular low-tension types (central-volcanoes, ring-complexes, randomly oriented plutons) during the last stages of Permian rifting (Ramberg & Larsen, 1978*a*, Ramberg & Spjeldnæs, 1978). This shift is accompanied by a change from extrusion of dense mafic to intermediate lavas, to intrusion of intermediate and finally silicic magmas.
3. The data presented in Fig. 9 suggests a positive correlation between (calculated) magma density and stratigraphic thickness among the basaltic rocks.
4. Rb–Sr ranges are 286–275 m.y. for the intermediate to silicic intrusives (including cauldron ring-dykes and central dome intrusives) in the southern part of the rift, and 280–247 m.y. for those in the northern part of the Region, suggesting that the rift has a trend of development with time from southwest to northeast (Sundvoll, 1978*a*). A similar trend is found among the basaltic rocks. In the Skien and Vestfold areas, the thickest unit of basalts is found at the base of the lava pile (B₁). These basalts also give higher calculated magma densities than the younger ones (Fig. 9). In the Krokstogen and Hurdal areas the thickest basalt units (with the highest calculated magma densities) are found among younger basalts (at the B₂ and B₃ stratigraphic positions) (Fig. 9).

These data strongly suggest that accumulation and fractionation of mafic melts near the mantle/crust boundary primarily resulted from the tendency of the continental crust to act as a density filter. The range of differentiation indexes, and average D.I., of magmatic rocks has been found to increase from mid-ocean ridges, through aseismic ridges and ocean islands, to continental rifts (*e.g.* Hekinian, 1974; Bailey, 1974), and thus appears to be dependent upon the density (and thickness) of the crust which has to be penetrated. (Magma density decreases as D.I. increases.) It has been proposed (Ramberg, 1970; Roberts, 1970; Gill, 1973; Baker *et al.*, 1978) that dense mafic magmas may only be admitted to the upper part of the continental crust in periods of high crustal extension. In tectonically quiet periods, dense magmas will be unable to rise through the 'light' crust, and will be restrained at the mantle/crust boundary or in the lower crust. Here the magmas will fractionate until their residues have low enough densities to rise buoyantly through the crust, giving rise to ring dykes, ring intrusions and intrusive bodies of random orientation. According to this model the intermediate melts differentiated to the stage at which their density (and viscosity) allowed them to rise buoyantly through the crust. This model furthermore suggests that the extrusion of large volumes of B₁ basalts in the southern, and B₂ and B₃ basalts in the northern part of the Oslo Region, took place during periods of relatively high extensional stresses, thus supporting the suggestion of a northeastward trend of evolution.

The correlation found to exist between calculated magma densities and variations in tensional stresses along the Oslo rift, and with time, as deduced from

other geological parameters, furthermore suggests that although magma density is not the only parameter influencing a magma's ability to ascend, it may be a useful semi-quantitative measure of variations in extensional stresses in areas of intra-plate volcanism.

Homogenization during fractionation in the lower crust

Uniform ratios between pairs of incompatible elements, as found in the Larvik plutonic rocks, are generally taken as an indication of derivation from a homogeneous mantle source region (e.g. Ferrara & Treuil, 1974; Hanson, 1977). Data on evolved rocks from a continental rift should, however, be regarded with some caution. It seems probable that the 'magma accumulation region' typically found near the mantle/crust boundary in continental rifts consists of numerous magma chambers, where fractionation takes place (Neumann & Ramberg, 1978). Such a structure is indicated by the data on the Larvik ring-complex, by local and short term alternations between magmatic products belonging to different main rock series observed in e.g. Gulinskaya, East Siberia (Logatchev, 1978), the Kenya rift (Baker *et al.*, 1978), and the Rio Grande rift (Baldrige, 1978), and by seismic surveys of the Rio Grande rift (Bridwell, 1978) and the Rhine Graben (Mueller, 1978). It is, however, also clear that the 'magma accumulation region' acts as an open system (like O'Hara's (1977) periodically refilled magma chamber), where mafic magmas more or less continuously flow in from the melting region deeper in the mantle, and somewhat evolved melts rise to extrude, or to intrude shallower parts of the crust. Some magma mixing must necessarily also take place when the melts rise slowly through the 'magma accumulation region' in the lower crust. Partial remelting of previously crystallized gabbros, and fractionation in a periodically refilled magma chamber will (whether each batch of injected mafic magma has the same or different compositions) tend to cause an excessive enrichment of residual elements, and significant variations in ratios between strongly and weakly incompatible elements in the residual melts (O'Hara, 1977). Mixing of melts would, in contrast, have a homogenizing effect on their compositions.

It is thus clear that in continental rifts, moderately to strongly evolved rocks provide rather unreliable information about trace element ratio variations in the source region.

CONCLUSIONS

In tectonically quiet periods between periods of high tectonic activity, and during the final stages of the Permian rifting episode in the Oslo Region, basaltic melts rising from the mantle source region were unable to penetrate the light crust, and accumulated near the mantle/crust boundary. Here fractional crystallization proceeded to the stage at which the residual melts had enough buoyancy to force their way to the surface or the upper crust. Within the pressure range of the lower part of the thinned crust in the Oslo rift (Moho depth: ~35 km or ~10 kb), no

thermal barrier existed between alkali and tholeiitic basalts, and the strongly pressure-dependent olivine–pyroxene liquidus boundary lay within, or close to, the field of silica-saturation. A combination of crystal fractionation of alkali to transitional basalts at roughly 7–10 kb, and crustal density filtering thus gave rise to the high proportion of approximately silica-saturated intermediate rocks in the Oslo Region, leaving large masses of dense cumulates together with gabbros in the lower crust. Small differences in silica activity and degree of iron-enrichment in the intermediate magmas which ascended from the lower crust signify derivation from chemically dissimilar parent magmas and/or evolution under slightly different conditions of crystallization (such as P_{total} , $P_{\text{H}_2\text{O}/\text{CO}_2}$, f_{O_2}), in a series of magma chambers in the lower crust. The compositional similarities between the intermediate rocks (rhomb porphyry lavas, larvikites, least evolved foyaite/hedrumites and nordmarkites) thus resulted from a fairly uniform degree of fractionation under a limited range of physicochemical conditions rather than evolution from similar parent magmas.

The accumulation of mafic melts near the mantle/crust boundary must have increased the geothermal gradient considerably in the upper lithosphere in the rift area, and is likely to have caused some partial melting in the crust. Contamination by such melts may be as important a factor in the formation of hybrid rocks, *e.g.* akerites and Grefsen nordmarkite, as assimilation of xenoliths (of country-rock) by monzonitic and syenitic melts.

During the ascent to the upper crust, a thermal divide was established along the diopside–forsterite–anorthite join and the forsterite–orthopyroxene liquidus boundary shifted into the field of silica-oversaturation. Subsequent low pressure fractionation forced the evolutionary trends away from the field of silica saturation, and gave rise to strongly over-, and undersaturated derivatives, depending on the silica activity in the melt at the time of ascent.

It is likely that the evolutionary course outlined for the Oslo rift has been followed in many cases of continental rifting, with the pressures in the lower part of a thinned crust playing the dominant role in determining the composition of the evolved rocks. Seismic and gravimetric studies have shown that under active rifts (*e.g.* the Baikal, the East African, and the Rio Grande rifts) a zone of low-velocity, low-density material is typically found below the thinned crust, corresponding to the masses of dense material found at the mantle/crust boundary or forming part of the lower crust under paleorifts (the Gardar and Oslo rifts) (Fairhead, 1976; Ramberg, 1976; Bridwell, 1978; Logatchev *et al.*, 1978; Mueller, 1978; Neumann & Ramberg, 1978; Upton & Blundell, 1978). Evolutionary trends from approximately saturated intermediate, to both over- and undersaturated highly evolved rocks, as described above for the Oslo rift, are also reported from many other continental rifts (MacDonald, 1974), *e.g.* the Kenya rift (Baker *et al.*, 1978), and the Gardar rift (Upton, 1974; Upton & Blundell, 1978). Baker *et al.* (1977) have, furthermore, concluded that a basalt–benmoreite–trachyte suite from the Gregory rift, Kenya, formed by fractionation within the stability field of plagioclase, but at a high enough pressure for the stability field of clinopyroxene to be markedly expanded relative to its low pressure extent.

The larvikites, the lardalites and the least evolved foyaite/hedrumites in the Larvik complex have fairly uniform ratios among the strongly incompatible elements Rb, Hf, Ta, Th and U. These similarities may reflect derivation from parent magmas melted out of a homogeneous mantle source region. However, complex processes (such as fractionation in periodically refilled magma chambers, partial remelting of previously crystallized alkaline gabbros, and mixing between magmas) probably occurred in the complex system of magma chambers and feeding channels which appears to have existed in the lower crust during the rifting episode. Such processes are likely to obscure initial ratio differences and cause a higher degree of enrichment of incompatible elements than does simple fractionation. Thus the 'enriched' Oslo Region magmatic rocks may well be derived from an 'undepleted' type mantle source, as suggested by Jacobsen & Wasserburg (1978) on the basis of Sm–Nd and Rb–Sr isotope data.

Furthermore, there seems to be a positive correlation between calculated magma densities for the Oslo Region rocks, and degree of extensional stress at the time of extrusion (and intrusion) as implied by various parameters, suggesting that magma densities may be used as semi-quantitative indicators of relative stress differences.

ACKNOWLEDGEMENTS

I am indebted to B. T. Larsen and B. Sundvoll for keeping me informed about their latest analytical results, and for many long discussions, each of which has increased our understanding of the various aspects of rifting and magmatism in the Oslo Region. I am also indebted to M. P. Annis and R. Wilson who have kindly let me use their unpublished data on the Kikut plutonic rocks. I thank A/S Norsk Varekrigsforsikrings Fond for financing my microprobe and neutron activation analyses. Finally, but not least, I gratefully acknowledge W. L. Griffin and B. B. Jensen for constructive criticism of the manuscript and helpful suggestions for improvements.

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