

The Origin of Anorthosites and Related Rocks from the Lofoten Islands, Northern Norway: II. Calculation of Parental Liquid Compositions for Anorthosites

GREGOR MARKL^{1*} AND B. RONALD FROST²

¹INSTITUT FÜR MINERALOGIE, PETROLOGIE UND GEOCHEMIE, ALBERT-LUDWIGS-UNIVERSITÄT, ALBERTSTRASSE 23 B, D-79104 FREIBURG, GERMANY

²DEPARTMENT OF GEOLOGY AND GEOPHYSICS, UNIVERSITY OF WYOMING, LARAMIE, WY 82071, USA

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Clinopyroxene, orthopyroxene and plagioclase in the Proterozoic Lofoten anorthosites are inferred to be in equilibrium with ferrodioritic melts, on the basis of literature partition coefficients for these phases in equilibrium with basaltic melts. Ferrodiorites and high-Al gabbros occur as fine-grained dykes in spatial association with the ~1.8 Ga Lofoten anorthosites. They are linked to each other by an Fe-enrichment fractionation trend typical of tholeiites, and they are characterized by Al₂O₃ contents of 14–18.5 wt % and X_{Fe} [= FeO^{tot}/(FeO^{tot} + MgO)] between 0.5 and 0.8. The association of high-Al gabbros, ferrodiorites and anorthosites is common in many Proterozoic anorthosites. We infer that the high-Al gabbros to ferrodiorites represent the parental liquids from which these Proterozoic anorthosites formed. The Lofoten rocks show fractionation trends that are similar to those of high-Al gabbros and ferrodiorites from other Proterozoic anorthosite complexes. We interpret these trends in terms of incorporation of intermediate composition plagioclase into a typical tholeiitic magma with subsequent fractionation. The first process produces positive Eu anomalies and the high Al₂O₃ contents of the melts, and explains the lack of negative Eu anomalies in most of the ferrodiorites from Lofoten and Laramie.

KEY WORDS: *anorthosite; parental liquid; ferrodiorite; high-Al basalt; partition coefficients*

INTRODUCTION

Proterozoic massif-type anorthosites are huge intrusive bodies that contain hundreds to thousands of km² of nearly pure (>90%) plagioclase cumulates of intermediate composition (An_{50±10}; Ashwal, 1993, and references therein). Most Proterozoic anorthosites are not associated with significant amounts of mafic or ultramafic cumulates. They contain plagioclase that is relatively sodic (usually around An₅₀) and this makes them distinct from other anorthositic rocks (Archean anorthosites, lunar anorthosites, anorthosites in layered intrusions), which contain more calcic plagioclase. Proterozoic anorthosite complexes contain anorthosite *sensu stricto* (>90% plagioclase) as well as variable amounts of leuco-gabbroic, -troctolitic and -noritic rocks which respectively may contain up to 35% clinopyroxene, olivine or orthopyroxene.

The Proterozoic anorthosites of the Lofoten Islands in northern Norway (Heier, 1960; Green & Jorde, 1971; Romey, 1971; Griffin *et al.*, 1978; Wade, 1985) are associated with large amounts of temporally related intrusive mangerites, charnockites and gabbros (Markl *et al.*, 1998). There are two different anorthosite intrusions present. One is a relatively calcic complex, which occurs on the island of Flakstadøy (Flakstadøy Basic Complex, FBC), and the other is a relatively sodic complex, which is found at the Eidsfjord on the island of Langøy (Eidsfjord anorthosite). The FBC has the mineral assemblage cpx

*Corresponding author. Telephone: xx 49 761 203 6414. Fax: xx 49 761 203 6407.
e-mail: markl@ruf.uni-freiburg.de

\pm ol \pm opx + plag + ilm + mt, whereas the Eidsfjord anorthosite shows the assemblage opx \pm cpx + plag + ilm + mt without olivine. The FBC crystallized at pressures from 9.5 to 4 kbar at temperatures between 1185 and 1140°C, whereas the Eidsfjord anorthosite equilibrated at temperatures of 1100–1135°C with a maximum pressure of 7.3 kbar. Both crystallized at oxygen fugacities above fayalite–magnetite–quartz (FMQ).

The anorthosites of Lofoten are part of a NE–SW trending belt of Proterozoic anorthosites that extends from Scandinavia across North America. These anorthosites range in age from 920 to 1789 Ma (Ashwal, 1993, and references therein; Schärer *et al.*, 1996). The largest complexes occur within the Grenville and Nain–Churchill provinces of the Canadian Shield (Emslie, 1978, 1985; Wiebe, 1979, 1990, 1992; Morse, 1982; McLelland & Chiarenzelli, 1990; Owens *et al.*, 1994). Well-studied Proterozoic anorthosites from this belt include the Laramie Anorthosite Complex in Wyoming (e.g. Scoates & Frost, 1996, and references therein), the Nain Plutonic Suite in coastal Labrador (e.g. Morse, 1988, and references therein; Emslie *et al.*, 1994) and the Rogaland anorthosite complex, southwest Norway (e.g. Duchesne, 1984; Duchesne *et al.*, 1989, 1993; Schärer *et al.*, 1996).

There is still much debate as to how these immense bodies of nearly homogeneous mineralogy and plagioclase composition develop. The major problem is the lack of information on the compositions of the silicate melts from which the anorthosites crystallized. Two major theories have been proposed: the existence of hyperfeldspathic magmas (e.g. Wiebe, 1979, 1990; Morse, 1982) and two-stage polybaric crystallization of basaltic parent melts without involvement of hyperfeldspathic liquids (e.g. Emslie, 1978; Duchesne, 1984; Longhi & Ashwal, 1985; Wiebe, 1992; Ashwal, 1993; Emslie & Hegner, 1993; Longhi *et al.*, 1993; Emslie *et al.*, 1994; Scoates & Frost, 1996). The existence of a basaltic melt may be indicated by the presence of high-Al gabbros, ferrogabbros and ferrodiorites coeval with the anorthosites. These rocks have been suggested to represent parental or residual liquids of anorthosites (Olson & Morse, 1990; Mitchell *et al.*, 1995, 1996).

In this paper we use published partition coefficients and mineral compositions from the Lofoten anorthosites to calculate the composition of the parental melt from which the anorthosites crystallized. We then compare these calculated melt compositions with the compositions of fine-grained rocks from the Lofoten and other well-studied anorthosites to better constrain the origin of the parent melts to massif-type anorthosites.

WHOLE-ROCK CHEMISTRY

Analytical techniques

X-ray fluorescence (XRF) analyses of whole-rock samples and mineral separates of plagioclase, clinopyroxene and

orthopyroxene were carried out at the Institut für Mineralogie, Petrologie und Geochemie at the University of Freiburg on a Philips PW 1450/20 machine using natural standards; accuracy and detection limits are of the order of 0.1 wt % for major elements and 1–10 ppm (depending on the specific element) for minor elements. Sample preparation was performed at the same institute. For whole-rock measurements, about 1–3 kg of material, depending on the grain size, was crushed and milled in agate mills, and powder and Li borate fusion discs were prepared to measure trace and major elements. Mineral separates were prepared from 63–125 μ m fractions of the whole-rock samples; separations were made with heavy liquids, magnetic separators and a wave table. The resulting separates were again milled, and powder and fusion discs were prepared. Powders of six whole-rock samples were analysed for trace elements including La, Ce, Nd, Sm, Eu, Gd, Tb, Dy, Er, Yb and Lu by inductively coupled plasma mass spectrometry (ICP-MS) at the CRPG laboratory, Nancy, on a commercial basis.

Major and trace elements

Anorthositic rocks

Whole-rock analyses of anorthositic rocks typically reflect a cumulate mixture of various proportions of plagioclase, oxides and Fe–Mg silicates. Typically, it is difficult to derive genetic significance from whole-rock data, which in most cases only show that these rocks are cumulates (e.g. Haskin & Salpas, 1992). For example, a plot of wt % FeO vs wt % Al₂O₃ shows a linear relationship of decreasing Al₂O₃ with increasing FeO (Fig. 1). It is reasonable to assume that FeO in the rock is contributed by post-cumulate pyroxenes and oxides whereas the Al₂O₃ resides mostly in the plagioclase. Thus the linear pattern in Fig. 1 merely reflects different abundances of pyroxene and plagioclase. Similar trends are seen in other major oxides, indicating that the major element chemistry is controlled for the most part by the relative abundance of the minerals in the rock, rather than by the composition of the melt.

Gabbro and ferrodiorite

In addition to the coarse-grained rocks, there are fine-grained mafic rocks, which we assume to represent chilled melts (Table 1). Of these samples, GM 405, 448, 406, 500, 508, 1120 and 1154 are dyke rocks, whereas GM 451, 443 and 420 are from bodies tens of metres in size. Eighteen additional analyses of evolved tholeiitic to ferrodioritic compositions from Lofoten dyke rocks have been given by Misra & Griffin (1972). The fine-grained samples listed here are interpreted to represent melt rather than cumulate compositions because of their fine grain size, the lack of Eu anomalies (where analysed, see

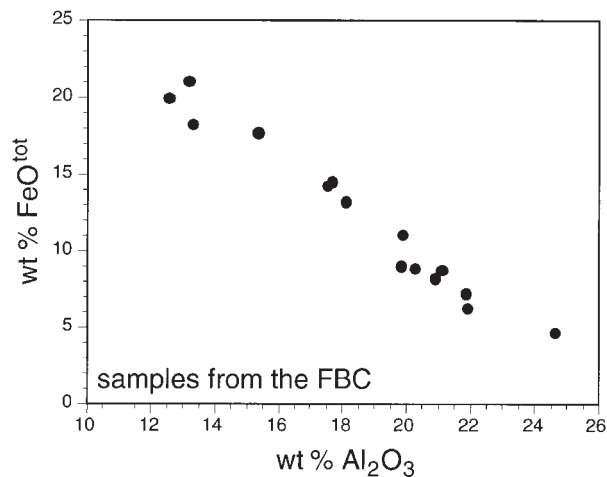


Fig. 1. Al₂O₃ vs FeO^{tot} plotted for anorthositic and related rocks from the FBC, Lofoten Islands. The linear dispersion of data clearly shows a cumulate trend and does not reflect a fractionation trend.

Fig. 2 and Table 2) and the lack of correlations between chemistry and the compositions of possible accumulating phases.

These fine-grained rocks are chemically distinct from the gabbroic anorthosites. The most important differences are that the anorthositic rocks have higher Al₂O₃ and Na₂O contents. Some of these fine-grained rocks (GM 448, 500, 420 and 443) have low X_{Fe} [= FeO^{tot} / (FeO^{tot} + MgO)] and have major element compositions that are similar to those of evolved tholeiites (e.g. Carmichael *et al.*, 1974), whereas others (GM 405, GM 406, GM 451, GM 1120) have higher X_{Fe} and have compositional affinities to ferrodiorites from other Proterozoic anorthosite complexes (e.g. high FeO^{tot}, TiO₂, P₂O₅; Emslie, 1980; Duchesne *et al.*, 1989; Mitchell *et al.*, 1995, 1996). In terms of their trace element chemistry, the latter group shows enrichment in Zr and Ba and depletion in Ni and Cr compared with the other fine-grained rocks and with the gabbroic anorthosites. GM 448, GM 1120 and GM 1154 were sampled from dykes within the Eidsfjord anorthosite. As this is of importance in the discussion below, the analysis of the gabbro sample BM-10 from the Laramie Anorthosite Complex in Wyoming (Mitchell *et al.*, 1995) is listed for comparison in Table 1.

CALCULATIONS WITH PARTITION COEFFICIENTS

Theory

The distribution of elements between a melt and co-existing crystals is given by the partition coefficient D_i , which is defined as

$$D_i = \frac{\text{concentration of element } i \text{ in crystal}}{\text{concentration of element } i \text{ in melt}}$$

The partition coefficient of an element may vary with temperature, pressure, composition of the melt and composition of the crystal, although changes in pressure and melt chemistry seem to exert a smaller influence than do changes in the other parameters. Based on experiments in the simplified system CaMgSi₂O₆-NaAlSi₃O₈-CaAl₂Si₂O₈, Blundy & Wood (1994) proposed a model that allows the calculation of the concentrations of all equally charged cations in the melt if their concentrations in the crystal, the partition coefficient for one element, the elastic moduli and the site sizes in the mineral are known. This model is based on the assumption that the structure of the crystal in equilibrium with a melt governs the partitioning between the melt and the crystal, and that the composition or structure of the melt is of much less importance. Because their experiments can account for the main network-modifying cations, this model system is also a good approximation of Fe-bearing basaltic compositions (see, e.g. Blundy, 1997). The D values and mineral analyses that were used in the calculations are given in Tables 3 and 4, and all other parameters are given in Table 5. Ionic radii are from Whittaker & Muntus (1970) and Blundy & Wood (1994); site radii were either adopted from Blundy & Wood (1994) or calculated as detailed below.

The use of the Blundy & Wood (1994) model is limited to 1+, 2+ and some 3+ elements because the partitioning behaviour of some 3+ and the 4+ elements is more complex, probably because of additional electron interactions. For this reason, and to better evaluate the validity of the results, we performed additional calculations using the partition coefficients from the works of Beattie *et al.* (1991) and Beattie (1993), who investigated partitioning of various major, minor and trace elements between orthopyroxene, clinopyroxene and basaltic melts. For comparison, we also calculated concentrations of K₂O, Rb, Sr and Ba using the partition coefficients of Simmons & Hanson (1978) and of Morse (1988), and of Al₂O₃, TiO₂ and FeO using the partition coefficients reported by Nolan & Morse (1986) that they derived for anorthosites or rocks related to anorthosites (Kiglapait troctolite). In the following, we will describe in detail which model and which values we used for the calculations with respect to the various minerals.

Calculation procedure

Plagioclase

The model of Blundy & Wood (1994) was used to calculate the concentration of Na₂O, K₂O, CaO and for

Table 1: XRF whole-rock data for selected gabbroic and ferrodioritic rocks from the Lofoten Islands; for comparison (see text), also an analysis of an LAC gabbro from Mitchell et al. (1995) is listed

Rock type:	Gabbro	Gabbro	Gabbro	Gabbro	Gabbro	Gabbro	Gabbro	Ferro-gabbro	Ferro-diorite	Ferro-diorite	Ferro-diorite	Ferro-diorite
Locality:	LAC, Wyoming	Børnhaugen	S Festvåg	Nesland	Hopen	Eidsfjord	Hopen	Hopen	Eidsfjord	Grønning	Fleignes	Skelfjord
Island:	Mitchell al. (1995)	et Austvågøy	Austvågøy	Flakstadøy	Austvågøy	Langøy	Austvågøy	Austvågøy	Langøy	Langøy	Langøy	Flakstadøy
Sample:	BM-10	GM 420	GM 443	GM 508	GM 406	GM 448	GM 500	GM 405	GM 451	GM 1120	GM 1154	GM 517
<i>wt %</i>												
SiO ₂	47.44	45.52	47.89	48.26	48.09	52.41	51.30	45.33	45.89	45.40	47.67	40.49
TiO ₂	1.65	0.90	1.13	1.55	2.38	1.09	0.88	1.36	2.92	2.57	1.98	2.61
Al ₂ O ₃	14.48	16.6	14.03	14.10	14.69	17.84	15.65	16.64	14.00	15.02	15.50	11.10
Fe ₂ O ₃ ^{tot}	13.89	14.31	14.88	14.88	16.04	11.34	10.74	15.44	20.06	16.62	16.69	22.82
MnO	0.17	0.15	0.21	0.23	0.22	0.16	0.16	0.14	0.18	0.21	0.19	0.25
MgO	11.19	9.08	8.02	6.48	5.32	6.01	7.09	5.39	3.54	5.37	6.01	6.05
CaO	8.77	10.62	12.15	9.39	8.31	8.01	9.90	10.78	7.44	7.68	8.44	9.43
Na ₂ O	2.55	1.81	2.54	3.41	3.72	3.53	3.61	3.50	3.37	3.80	3.20	2.75
K ₂ O	0.40	1.08	0.34	0.82	1.05	0.90	1.10	0.59	1.98	1.97	0.81	1.04
P ₂ O ₅	0.34	0.06	0.10	0.69	0.51	0.21	0.25	1.55	0.78	1.67	0.62	3.19
Total	101.18	100.28	99.84	99.97	100.42	101.65	100.84	100.92	100.43	100.31	101.11	99.87
X _{Fe}	0.53	0.59	0.63	0.67	0.73	0.63	0.58	0.72	0.84	0.74	0.71	0.77
<i>ppm</i>												
Rb	2	24	10	6	14	7	32	8	28	22	8	10
Sr	498	824	128	662	298	637	510	1390	595	841	650	518
Ba	450	388	33	773	477	812	534	421	2119	1527	796	587
Pb	n.d.	5	13	6	7	7	11	10	9	33	10	5
Y	19	9	22	28	30	14	18	14	32	38	21	45
Nb	n.d.	5	6	4	5	3	6	3	7	7	4	8
Zr	48	74	55	70	123	5	59	0	109	90	8	103
Hf	n.d.	5	n.d.	5	6	n.d.	n.d.	n.d.	n.d.	8	4	6
V	193	373	349	312	326	164	217	202	424	244	351	460
Ni	369	49	99	78	17	24	75	25	17	30	37	6
Cr	1235	159	216	419	86	144	333	81	16	37	95	113
Co	n.d.	61	42	50	48	43	43	54	46	40	49	64
Cu	38	61	n.d.	76	18	n.d.	n.d.	n.d.	20	30	24	41
Zn	116	89	n.d.	138	149	n.d.	n.d.	n.d.	144	167	159	198
Ga	n.d.	16	n.d.	18	21	n.d.	n.d.	n.d.	23	14	17	19
Th	n.d.	4	n.d.	3	3	n.d.	n.d.	n.d.	n.d.	8	7	0
U	n.d.	5	n.d.	7	8	n.d.	n.d.	n.d.	n.d.	0	0	5
Sc	25	24	30	23	25	15	16	20	26	17	18	26
La	11	9	4	32	31	6	11	12	41	24	12	72

n.d., not determined; $X_{Fe} = FeO^{tot}/(FeO^{tot} + MgO)$.

some samples Rb, Ba, Sr, MgO and FeO, from the analyses of plagioclase separates and of electron microprobe (EMP) point analyses. In accordance with the data

of Blundy & Wood (1994), we set D_{Ca} to unity for bivalent elements. The combination of the model of Blundy & Wood (1991) for D_{Sr} with the model of Blundy & Wood

Table 2: REE concentrations (in ppm) in selected gabbroic and ferrodioritic rocks from the Lofoten Islands

ppm	GM 405	GM 443	GM 448	GM 451	GM 500	GM 508	GM 517
La	18.94	4.00	15.59	40.56	18.52	25.27	58.37
Ce	40.82	10.90	31.47	83.77	39.70	57.38	128.94
Pr	5.21	1.67	3.95	10.76	5.14	7.92	16.92
Nd	25.14	8.66	17.45	46.98	21.05	34.98	79.05
Sm	5.10	2.56	3.32	9.25	4.17	7.01	14.64
Eu	1.49	0.90	1.60	3.73	1.50	2.13	3.45
Gd	4.13	3.21	3.05	7.81	3.87	6.20	11.76
Tb	0.49	0.54	0.41	1.07	0.56	0.85	1.62
Dy	2.73	3.45	2.22	5.57	2.87	4.52	8.73
Ho	0.48	0.77	0.44	1.08	0.61	0.89	1.51
Er	1.22	2.09	1.14	2.90	1.62	2.29	4.07
Tm	0.18	0.30	0.16	0.45	0.23	0.34	0.59
Yb	1.00	2.09	1.16	2.50	1.46	2.07	3.38
Lu	0.13	0.34	0.18	0.41	0.24	0.31	0.53

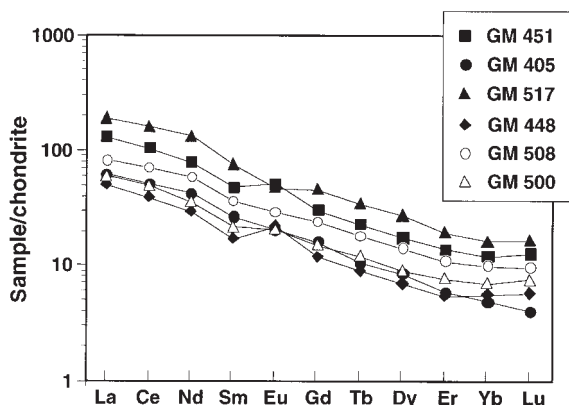


Fig. 2. REE patterns of ferrodioritic rocks from Lofoten (whole-rock chemistry is reported in Table 1). (Note the absence of significant Eu anomalies, which is discussed in the text.)

(1994) gave us another constraint on D values of 2+ elements and thus allowed us to calculate the size of the Ca site ($r_{\text{Ca}}^{\text{Ca}}$) in plagioclase according to the expression

$$r_{\text{Ca}}^{\text{Ca}} = \left[\frac{\{-\ln[\exp\{26800 - 26700[278 \cdot 2X_{\text{An}}^{\text{Plag}} / (262 \cdot 23 + 15 \cdot 97X_{\text{An}}^{\text{Plag}})]\} / RT] / D_{\text{Ca}}\} RT}{4\pi EN_{\text{A}}} - \frac{[0 \cdot 333(r_{\text{Sr}}^3 - r_{\text{Ca}}^3)]}{[0 \cdot 5(r_{\text{Ca}}^2 - r_{\text{Sr}}^2)]} \right] \quad (1)$$

where R is the gas constant, T is the temperature in K, E is Young's modulus and N_{A} is Avogadro's constant. D_{Sr} was calculated from the expression

$$RT \ln D_{\text{Sr}} = 26800 - 26700X_{\text{An}} \quad (2)$$

from Blundy & Wood (1991). The calculated sizes $r_{\text{Ca}}^{\text{Ca}}$ were always within the range given by Blundy & Wood

(1994) and hence support our choice of D_{Ca} . To determine D for univalent elements, the value of D_{Na} between plagioclase and melt was fitted to the Na content of the melt as estimated from clinopyroxenes. From this we calculated D_{K} from the expressions of Blundy & Wood (1994). For comparison, we also calculated Rb, K, Ba and Sr from the partition data of Morse (1988) and of Simmons & Hanson (1978). The results of these calculations are given in Table 3. The values for K_2O and Rb from the Blundy & Wood model are significantly higher than those derived from the other data (by 0.5–2.5 wt %), whereas the Sr and Ba values are considerably lower (by 100–400 ppm).

Clinopyroxene and orthopyroxene

Experimentally determined partition coefficients of Beattie *et al.* (1991) and Beattie (1993) were used to estimate the contents of CaO , MgO , FeO , Na_2O , Sr, Ni, Co, TiO_2 , Al_2O_3 and SiO_2 in the melt from bulk and microprobe analyses of pyroxenes. We used the data of Beattie *et al.* (1991) and Beattie (1993) because the crystallization temperatures of the Lofoten rocks are similar to those used in his experimental work. For orthopyroxene, D_{Ca} was adjusted so that the results agreed with those from the plagioclase calculations. The D values of Mn, Mg, Ni and Co were then calculated from the expressions of Beattie *et al.* (1991). For the FeO calculations from orthopyroxene, we estimated the ferric iron from mineral formula calculations (normalization of the analyses to two cations and six oxygens) and used the Fe^{2+} content for the calculations. For clinopyroxene, we used data from Beattie's (1993) experiments at 10

Table 3: Calculated concentration of some major, minor and trace elements in the melt coexisting with plagioclase from the Lofoten anorthosites and the K_D values used (see text for explanation of the calculation procedure)

Sample:	GM 447	GM 449	GM 452	GM 452	GM 531	GM 532	GM 535	GM 255	GM 263	GM 265
	Plag	Plag	Plag	Plag	Plag	Plag	Plag	Plag	Plag	Plag
Island:	Langøy	Langøy	Langøy	Langøy	Flakstadøy	Flakstadøy	Flakstadøy	Flakstadøy	Flakstadøy	Flakstadøy
Method:	XRF	XRF	XRF	EMP	XRF	XRF	XRF	EMP	EMP	EMP
SiO ₂	56.25	56.07	55.90	56.76	54.52	54.31	54.23	56.26	56.00	55.88
TiO ₂	0.03	0.05	0.04	0.00	0.03	0.04	0.04	0.05	0.00	0.00
Al ₂ O ₃	27.18	27.62	27.14	27.12	28.43	28.58	28.69	27.88	27.83	28.29
FeO	0.25	0.34	0.24	0.00	0.16	0.15	0.18	0.15	0.04	0.08
MnO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00
MgO	0.00	0.06	0.03	0.00	0.06	0.03	0.16	0.00	0.00	0.06
CaO	9.41	9.90	9.34	9.24	10.7	10.85	10.91	10.13	10.20	10.45
Na ₂ O	5.59	5.45	5.71	6.21	4.76	4.65	4.90	5.30	5.35	5.18
K ₂ O	0.65	0.34	0.47	0.34	0.87	0.81	0.54	0.26	0.35	0.21
Total	99.58	100.06	99.06	99.67	99.74	99.62	99.86	100.04	99.78	100.11
Rb	2.4	2.9	1.2	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.
Sr	1150	1245	1147	n.m.	1265	1205	1274	n.m.	n.m.	n.m.
Ba	686	408	469	n.m.	494	460	502	n.m.	n.m.	n.m.
<i>Partition coefficients used for the model of Blundy & Wood (1994)</i>										
D_{Na}	1.35	1.35	1.40	1.51	1.40	1.35	1.43	1.55	1.55	1.50
D_{Ca}	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00
<i>Calculated concentrations in liquid [partition coefficients of Beattie et al. (1991), Blundy & Wood (1991, 1994) and Beattie (1993)]</i>										
FeO	—	—	—	—	—	—	—	—	13.91	16.31
MgO	—	7.07	6.39	—	6.46	—	—	—	—	6.16
CaO	9.40	9.90	9.34	9.68	10.70	10.90	10.90	10.10	10.70	11.10
Na ₂ O	4.14	4.04	4.08	4.11	3.40	3.44	3.43	3.42	3.45	3.45
K ₂ O	2.62	1.47	2.10	1.40	3.76	3.65	2.34	1.11	1.17	0.87
Rb	64	85	43	—	—	—	—	—	—	—
Sr	362	416	318	—	456	444	468	—	—	—
Ba	812	572	557	—	861	854	920	—	—	—
<i>Calculated in liquid [partition coefficients of Morse (1988)]</i>										
K ₂ O	1.04	0.54	0.75	0.54	1.40	1.30	0.87	0.42	0.57	0.34
Rb	5	6	2	—	—	—	—	—	—	—
Sr	767	830	765	—	843	803	849	—	—	—
<i>Calculated concentrations in liquid [partition coefficients of Simmons & Hanson (1978)]</i>										
K ₂ O	2.18	1.13	1.58	1.13	2.92	2.71	1.81	0.87	1.18	0.70
Rb	17	21	9	—	—	—	—	—	—	—
Sr	628	680	627	—	691	658	696	—	—	—
Ba	1163	692	795	—	837	780	851	—	—	—

kbar as an approximation for our calculations. As a comparison, we calculated concentrations of Al₂O₃, FeO and TiO₂ in the coexisting liquid from the partitioning data reported by Nolan & Morse (1986). For the FeO calculations, again the ferric/ferrous ratio was calculated as above; as Nolan & Morse (1986) gave a value for only

$D_{FeO/MgO}$, we chose the MgO content of the liquid as calculated from the Beattie *et al.* (1991) expressions and then calculated FeO from the combination of these two values. These values are significantly higher (by 0.5 to several wt %) than those calculated from the Blundy & Wood model.

Table 4: Calculated concentration of some major, minor and trace elements in the melt coexisting with pyroxenes from the Lofoten anorthosites and the K_D values used (see text for explanation of the calculation procedure)

Sample:	GM 452 Cpx	GM 450 Opx	GM 263 Cpx	GM 265 Cpx	GM 524 Cpx	GM 265 Opx
Island:	Langøy	Langøy	Flakstadøy	Flakstadøy	Flakstadøy	Flakstadøy
Method:	EMP, reintegr.	XRF	EMP, reintegr.	EMP, reintegr.	XRF	EMP, reintegr.
SiO ₂	51.14	51.45	51.45	51.71	49.90	52.12
TiO ₂	0.47	0.38	0.46	0.41	0.73	0.25
Al ₂ O ₃	3.23	2.84	4.40	4.16	4.03	3.23
FeO	10.68	21.62	9.16	9.19	13.96	16.39
MnO	0.25	0.38	0.16	0.26	0.35	0.41
MgO	13.33	22.26	15.93	16.24	14.78	24.58
CaO	19.76	1.93	17.98	17.54	15.88	2.56
Na ₂ O	0.61	0.11	0.44	0.47	0.62	0.03
K ₂ O	0.00	0.02	0.00	0.00	0.16	0.00
Total	99.47	100.99	100.00	100.00	100.47	99.57
Fe ³⁺ /(Fe ²⁺ + Fe ³⁺)	0.12	0.13	0.10	0.09	0.29	0.09
Sr	n.m.	36	n.m.	n.m.	64	n.m.
Ba	n.m.	28	n.m.	n.m.	69	n.m.
Ni	n.m.	126	n.m.	n.m.	48	n.m.
Cr	n.m.	364	n.m.	n.m.	147	n.m.
Sc	n.m.	34	n.m.	n.m.	64	n.m.
Co	n.m.	107	n.m.	n.m.	54	n.m.
Zn	n.m.	202	n.m.	n.m.	74	n.m.
<i>Partition coefficients used (see text for detailed discussion of sources)</i>						
D_{Si}	1.05	—	1.05	1.05	1.05	—
D_{Ti}	0.40	—	0.40	0.40	0.40	—
D_{Al}	0.23	—	0.23	0.23	0.23	—
D_{Fe}	—	1.21	—	—	—	1.41
D_{Mn}	—	1.40	—	—	—	1.64
D_{Mg}	2.50	4.04	2.50	2.50	2.50	4.73
D_{Ca}	2.10	0.20	2.10	2.10	2.10	0.25
D_{Na}	0.15	—	0.15	0.15	0.15	—
D_{Ni}	—	5.44	—	—	4.61	5.44
D_{Co}	—	2.35	—	—	2.03	2.35
<i>Calculated concentrations in liquid</i>						
SiO ₂	48.70	n.c.	49.00	49.20	47.50	—
TiO ₂	1.18	n.c.	1.15	1.03	1.83	—
Al ₂ O ₃	14.04	n.c.	19.13	18.09	17.52	—
FeO, corrected for						
Fe ³⁺	—	15.54	—	—	—	10.58
MnO	—	0.27	—	—	—	0.25
MgO	5.33	5.51	6.37	6.50	5.91	5.19
CaO	9.41	9.65	8.56	8.35	7.56	10.24
Na ₂ O	4.07	n.c.	2.93	3.13	4.13	—
Sr	—	—	—	—	467	—
Ni	—	23	—	—	10	—
Co	—	46	—	—	27	—
<i>Calculated concentrations in liquid [partition coefficients of Nolan & Morse (1986)]</i>						
Al ₂ O ₃	16.14	14.20	21.98	20.82	20.15	16.16
TiO ₂	1.94	1.58	1.92	1.71	3.04	1.05
FeO at fixed X_{Fe}	21.35	26.76	18.32	18.40	27.91	17.30
X_{Fe} , corrected for						
Fe ³⁺	18.80	23.21	16.58	16.66	19.82	15.73

For comparison, also values calculated with the partition coefficients of Nolan & Morse (1986) are reported; here, FeO could only be calculated when value of MgO (for which our calculated value was taken) was fixed. In the lowermost row, the calculated FeO was corrected for the Fe³⁺/Fe²⁺ ratio of the pyroxene used. n.m., not measured.

Table 5: Parameters used for modelling the anorthositic parental melts

Crystal-related parameters		Ion-related parameters	
<i>Young's moduli (kbar)</i>		<i>Ionic radii (Å)</i>	
E^{1+} (plag)	595	r_{K^+}	1.5
E^{2+} (plag)	1116	r_{Na^+}	1.18
E^{1+} (cpx)	454	r_{Rb^+}	1.605
E^{2+} (cpx)	1608	$r_{Ca^{2+}}$	1.12
		$r_{Mg^{2+}}$	0.88
		$r_{Fe^{2+}}$	0.86
<i>Sizes of lattice sites (Å)</i>		$r_{Mn^{2+}}$	0.96
		$r_{Ba^{2+}}$	1.455
r_o^{1+} (plag)	1.235	$r_{Si^{2+}}$	1.26
r_o^{2+} (plag)	see text	$r_{Ni^{2+}}$	0.77
r_o^{1+} (cpx)	1.02	$r_{Zn^{2+}}$	0.89
r_o^{2+} (cpx)	1.02	$r_{Co^{2+}}$	0.83

Young's modulus for the 1+ site of plagioclase was chosen to be intermediate between the values for pure albite and pure anorthite as provided by Blundy & Wood (1994). For Young's modulus of the plagioclase 2+ site we chose the value of Blundy & Wood (1994) run 90-6 plagioclase, which is nearly intermediate between the values for pure anorthite and pure albite. For the clinopyroxene 1+ site, we chose again an intermediate value between the two values for diopside provided by Blundy & Wood (1994), and for its 2+ site, we adopted their value for subcalcic augite.

Problematic elements

Plagioclase mineral separates consistently have higher Fe than microprobe analyses. This is probably caused by the presence of small (micrometre-sized) inhomogeneously distributed grains of Fe oxides in the plagioclase. For FeO content of plagioclase, therefore, the microprobe analyses were used in the calculations because they are believed to give more accurate results. The estimate of Fe in the melt from clinopyroxene using the data of Beattie (1993) gives unreasonably high results. We discount these results because there is a strong dependence of D_{Fe} on the X_{Fe} of the clinopyroxene (Beattie, 1993), and our clinopyroxenes are more Fe rich than those used in Beattie's (1993) experiments.

There are also discrepancies between estimations of K_2O in the melt made from plagioclase separates and those made from EMP analyses, especially in the case of the FBC. Based upon the composition of plagioclase separates, the calculated K_2O content of the FBC melt would be of the order of 2–4 wt %, whereas that of the Eidsfjord melt would be 1–2.5%. In contrast, calculations based upon microprobe analyses give more reasonable values of 0.9–1.4 wt% K_2O in both intrusions. We believe that these high K_2O values and related high

Table 6: Reconstructed anorthositic parental melt compositions from the Lofoten anorthosites

Reconstructed melt	Flakstadøy FBC	Langøy Eidsfjord	Derived from melt
SiO ₂	48.57	48.70	Cpx
TiO ₂	1.34	1.18	Cpx
Al ₂ O ₃	18.25	14.04	Cpx
FeO ^{tot}	10.58	15.54	Opx
MnO	0.25	0.27	Opx
MgO	5.99	5.42	Opx/Cpx
CaO	9.71	9.58	Plag/Opx/Cpx
Na ₂ O	3.42	4.09	Plag
K ₂ O	1.05	1.43	Plag
Sum	99.16	100.25	
FeO ^{tot} / (FeO ^{tot} +MgO)	0.64	0.74	
Rb	—	43	Plag
Sr	456	365	Plag/Cpx
Ba	—	557	Plag
Co	27	46	Opx
Ni	10	23	Opx
Ni	2	—	OI

concentrations of Ba and Rb are caused by the fact that the mineral separates included small amounts of late K-feldspar, which had formed from an interstitial liquid that was considerably enriched in K, Ba and Rb compared with the liquid from which the rest of the plagioclase and the mafic minerals crystallized. We therefore used the concentrations of K_2O in plagioclase as determined by the EMP analyses for the estimated liquid composition (Table 6). For the same reason, we do not report values for Ba and Rb in the calculated melt of the FBC.

Comparison of the results with Lofoten rocks and with MELTS calculations

For several reasons, we are confident that the calculated melts (Table 6) come close to the composition of the melt from which both the FBC and the Langøy anorthosites formed. First, the calculated liquids from the Lofoten anorthosites strongly resemble ferrodioritic rocks from Lofoten. Dykes of ferrodioritic and high-Al basaltic composition are found throughout the Lofoten Islands (Table 1) and additional data have been reported by Misra & Griffin (1972). The calculated liquids of the FBC and the Eidsfjord anorthosite are both similar to GM 508 (although the FBC melt is more aluminous), a fine-grained basaltic dyke from the FBC. The Eidsfjord

melt closely resembles the ferrodioritic rock GM 1154 from the margin of the Eidsfjord anorthosite (Table 1) in that they have both similar contents of many major and minor elements and similar Na/Ca and Al/Si ratios. The calculated melts are also similar to some of the dykes reported by Misra & Griffin (1972) that are net-veined by anatectic material from the country rocks. This material has an apparent age (1795 ± 20 Ma, Rb-Sr whole rock, Misra & Griffin, 1972) similar to that of the mangerites and anorthosites. Pending more precise U-Pb geochronological work and because of the similarity between our calculated melts and some of these dykes, we propose that these dykes may represent basic melts coeval with the anorthosites and mangerites and that observed differentiation trends in these rocks may reflect the differentiation of the basic magma. A plot of major element oxides against X_{Fe} ($= \text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}} + \text{MgO})$) in the fine-grained rocks from Lofoten is shown in Fig. 3. X_{Fe} is chosen as a monitor of differentiation in this figure because SiO_2 does not change much during crystallization of the ferrodiorites. There is a continuous trend of increasing Fe_2O_3 , P_2O_5 , K_2O and TiO_2 , and of decreasing Al_2O_3 , CaO and MgO with increasing X_{Fe} . Also shown in Fig. 3 are the calculated melts for the FBC and Langoy anorthosites. They clearly lie within the trend defined by the fine-grained rocks and closely resemble some individual dyke rock samples.

Another reason why we have confidence in the veracity of our results is that we used the MELTS program of Ghiorso *et al.* (1994) to calculate the composition of the phases that would be in equilibrium with our inferred melts at the temperatures and pressures that we estimated for the formation of these anorthosites (Markl *et al.*, 1998). The calculated phase compositions of plagioclase, olivine and clinopyroxene are very similar to those observed in the FBC and the Eidsfjord anorthosite (Table 7). Small discrepancies in the oxygen fugacities result from the fact that we have poor control of the ferric iron content of the pyroxenes and hence poor control of the ferric iron content of the calculated melts. This probably explains the predicted existence of olivine in the MELTS calculations at 4 kbar for the Eidsfjord melt, even though olivine seems to be lacking in these rocks.

DISCUSSION

Validity of the calculations

Choice of partition coefficients

Morse (1992) cast doubts on the validity of the Blundy & Wood (1991) formulation for augite-poor, anorthositic liquids. He argued that in the Kiglapait intrusion the formulation for Sr of Blundy & Wood (1991) led to strong discrepancies between the calculated and his reconstructed liquid evolution trend. Blundy & Wood (1992)

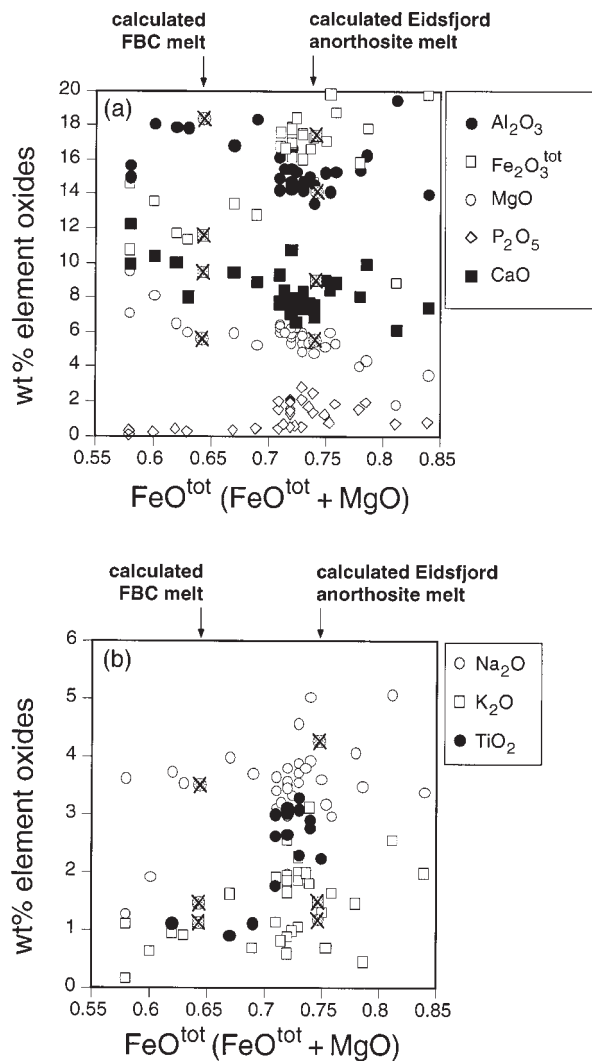


Fig. 3. (a) and (b) Comparison of basic dykes, fine-grained gabbroic rocks and ferrodiorite GM 451 from the Lofoten Islands with the reconstructed melts for the Flakstadøy Basic Complex (FBC) and the Eidsfjord anorthosite. Data used in this plot are taken from Misra & Griffin (1972) or are given in Tables 1 and 2. Crossed symbols indicate calculated melts.

argued in a reply that Morse's dataset was dependent on strictly closed system behaviour and pointed out that his results were in contrast to the results of all other studies they used for the derivation of their model. Furthermore, recent experimental work by Blundy (1997) on the Kiglapait intrusion suggests that the calculated liquid composition of Nolan & Morse (1986) and hence the partition coefficients calculated therefrom may not be correct. Hence, the discrepancy between the results calculated with the partition coefficients of Morse (1988) and Nolan & Morse (1986) and those calculated from Blundy & Wood (1994) (Tables 3 and 4) are likely to be caused by

Table 7: Comparison between mineral compositions that are observed in the Lofoten anorthosites and those that are calculated from the reconstructed parental melts using the MELTS program of Ghiorso et al. (1994)

	X_{An}^{plag}	X_{Or}^{plag}	X_{En}^{cpx}	X_{Wo}^{cpx}	X_{Fo}^{ol}
<i>Flakstadøy Basic Complex</i>					
Calculated compositions					
7 kbar, 1178°C, QFM	53	4	50	27	
4 kbar, 1123°C, QFM	47	6	50	30	65
4 kbar, 1127°C, QFM – 1	51	5	55	30	64
Observed compositions	45–57	1–8	52–53	35–36	66–72
<i>Eidsfjord anorthosite</i>					
Calculated compositions					
4 kbar, 1109°C, QFM – 1	44	7	41	34	54
Observed compositions	44–48	2–9	43	<38	QUIIF: 58

As the Eidsfjord anorthosite lacks olivine the theoretical olivine in equilibrium was calculated using QUIIF (see text for explanation) for comparison with the MELTS results.

errors in Morse's partition coefficients. Blundy (1997) furthermore showed that the partition coefficients calculated after Blundy & Wood (1994) are suitable for rocks and liquids of the anorthosite kindred, and he suggested that inversion using appropriate D values is a better means of calculating parental melt compositions rather than summation over exposed cumulates assuming closed system fractionation behaviour.

Validity and accuracy of the results

Because whole-rock analyses of cumulate rocks are sensitive more to variations in mineral abundance than to the composition of the melt from which they were derived, we have chosen to calculate the composition of the coexisting melt from the composition of the constituent minerals. This technique, though, is prone to some difficulties, which arise from the fact that anorthosites as cumulate rocks have had a long crystallization history and the melt from which the minerals crystallized certainly changed composition as crystallization proceeded. Calculations using cumulus and post-cumulus minerals will therefore give an average of the liquid composition that was in equilibrium with the anorthosites during its evolution. However, there are several arguments that support the validity of our results:

(1) the calculated melts are similar to gabbroic to ferrodioritic rocks observed in the Lofoten area;

(2) using the MELTS program and our calculated melt composition we can closely match the composition of the phases in the rocks;

(3) the calculated totals come close to 100%, even though there is no reason why they should (without normalization!);

(4) where element concentrations in the liquid are calculated from various minerals and from various sets of partition coefficients (e.g. Ca from plagioclase and from clinopyroxene, or Sr from plagioclase and clinopyroxene), the results show only low variance (Tables 3 and 4).

Based on the comparison between different samples and different sets of partition coefficients, we estimate that the calculated major element concentrations (SiO_2 , CaO, FeO, Al_2O_3) are precise to about 1–2 wt %. The Fe content may be subject to larger uncertainties because of the effects of Fe^{3+}/Fe^{2+} uncertainties. MgO, K_2O and Na_2O are probably precise to about 0.5–1 wt %. Sr gives a precision on the order of 30 ppm or 8–10 relative % when results from plagioclase and from clinopyroxene calculations are compared, and this may be representative for the trace elements. The calculated CIPW norm of these liquids is slightly Ne-normative when low ferric/ferrous ratios are assumed. At higher ferric/ferrous ratios and if a P_2O_5 content of 0.5–1 wt % in the liquid is assumed, the calculated melts are only weakly Ne-normative or not at all, and hence resemble the observed ferrodiorite compositions from Lofoten which are Hy- or slightly Ne-normative.

The reconstructed melts from the FBC and the Eidsfjord anorthosite (Table 6) have similar SiO_2 , TiO_2 and K_2O contents, and both are enriched in FeO^{tot} and Al_2O_3 compared with tholeiitic basalts. The calculated Eidsfjord melt is much more evolved than the FBC melt

as suggested by all common compositional indicators (higher X_{Fe} , Na_2O , K_2O , lower CaO , lower An in plagioclase; see also Markl *et al.*, 1998). Hence, we infer that the Eidsfjord anorthosite equilibrated with a more evolved melt than did the FBC.

The magmatic evolution from basalts to ferrodiorites on the Lofoten Islands as recorded by olivine and feldspar

As noted above, the fine-grained rocks from the Lofoten Islands are continuous in composition from high-Al gabbro to ferrodiorite. There is evidence for melts with even higher MgO contents as indicated by the composition of magnesian olivines in the Moskenesøy cumulate (Fig. 4). The presence of an early Mg-rich basaltic magma is hence recorded by the mafic cumulates (Markl *et al.*, 1998), whereas the more differentiated melts are recorded by the gabbros, anorthosites and ferrodiorites. The olivine composition in the Moskenesøy cumulate is Fo_{82-83} . Progressive Fe enrichment is recorded by the Ramberg gabbro on Flakstadøy (olivine with Fo_{77}), the anorthosites GM 263 and GM 265 with olivine of Fo_{72} and Fo_{68} , respectively, and the anorthosite GM 139 with olivine of Fo_{66} . In the Eidsfjord anorthosite, olivine is absent, but the calculated equilibrium composition from QUIIF (Andersen *et al.*, 1993) would have been Fo_{58} (at lower a_{SiO_2} than actually observed). The apparent general fractionation trend reflected by the whole-rock (Fig. 3) and olivine (Fig. 4) compositions is also followed by the plagioclase compositions. Plagioclase evolved from An_{69} in the Ramberg gabbro through normative An_{63} and An_{56} in the gabbros near Bjørnhaugen on Austvågøy (GM 420) and on Gimsøy (GM 440) to the FBC plagioclase (An_{57-47}) and the Eidsfjord anorthosite plagioclase (An_{48-44}). The ferrodiorite GM 451 has a normative ternary feldspar of $\text{An}_{17}\text{Or}_{23}$ (Markl *et al.*, 1998).

Comparison of the calculated melt compositions with rocks from other Proterozoic anorthosite complexes

Figure 5 plots the compositions of fine-grained rocks from Lofoten vs X_{Fe} together with a field for 50 analyses of gabbroic and ferrodioritic rocks from Harp Lake, Labrador (Emslie, 1980) and with 63 analyses from the Laramie Anorthosite Complex in Wyoming (Mitchell *et al.*, 1995, 1996). The fine-grained rocks from Lofoten have compositions very similar to those of the gabbros and ferrodiorites from both complexes (Fig. 5). The rocks from all three complexes clearly define similar evolution trends from high-Al gabbros to low-Al gabbros and to ferrodiorites in the majority of major and trace elements. The Lofoten rocks show the bend to high P_2O_5 , K_2O

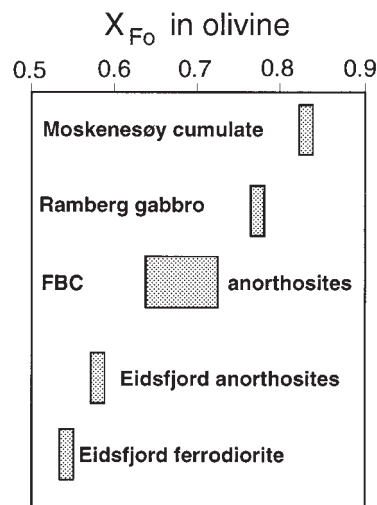
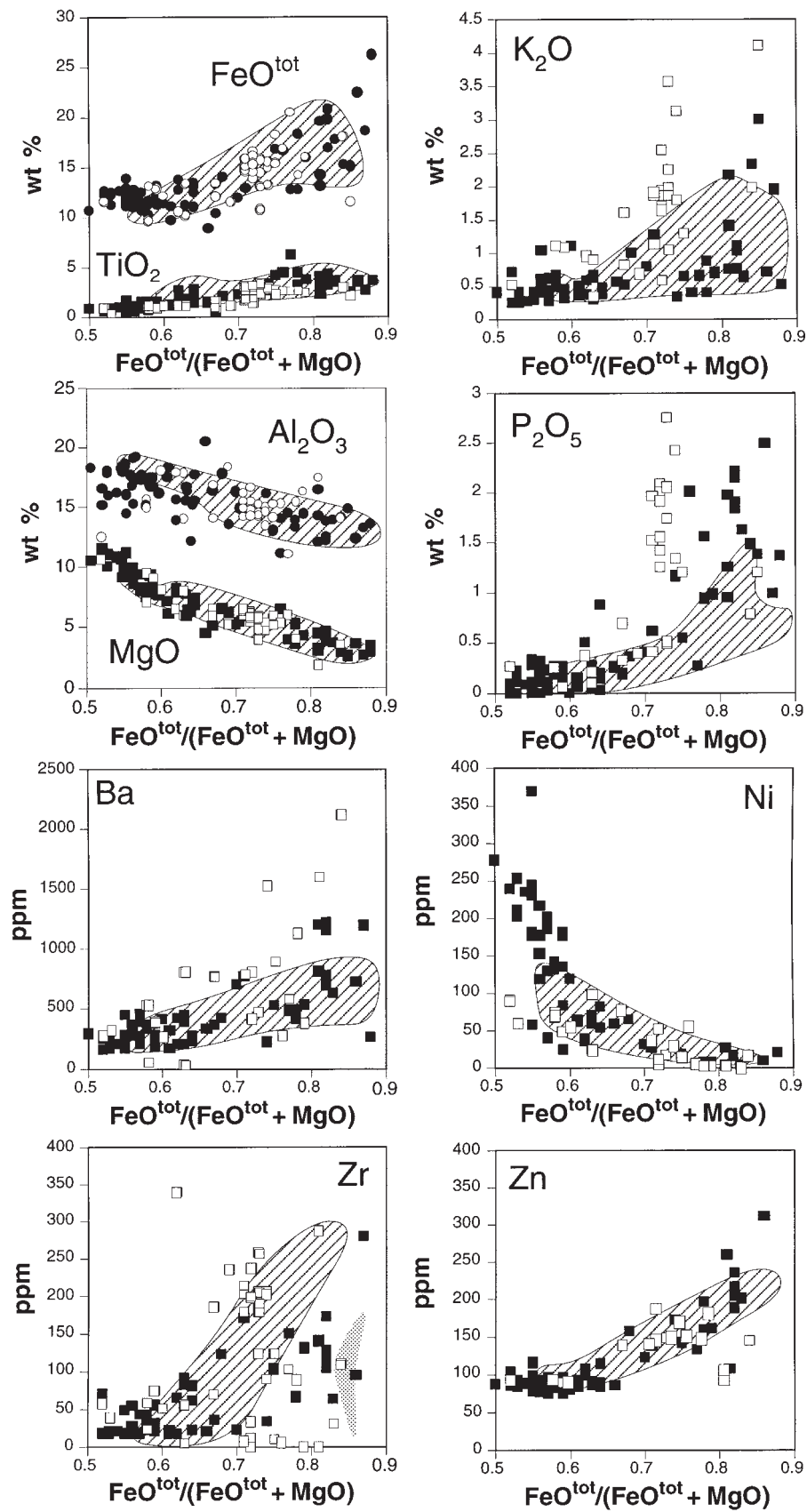


Fig. 4. The composition of olivine in various rock types from the Lofoten Islands. For rocks lacking olivine (Eidsfjord anorthosite, Eidsfjord ferrodiorite) the theoretical olivine in equilibrium was calculated using QUIIF (see text for explanation).

and Zr at lower X_{Fe} than the Laramie and Harp Lake rocks (Fig. 5). These discrepancies, though, do not negate the similar evolutionary trend from basaltic to ferrodioritic compositions in both areas, as they are most probably due to contamination of the evolving ferrodioritic melt with crustally derived melts at various stages of their evolution. Similarly (but not shown in Fig. 5), the Lofoten rocks closely resemble high-Al-Fe gabbros from the Adirondacks in New York (Olson & Morse, 1990), ferrodioritic rocks from the Nain Complex in Labrador (Wiebe, 1990) and ferrodiorites from Rogaland, Norway (Duchesne *et al.*, 1989). The Lofoten fine-grained rocks and the calculated melts evidently have equivalents in many large Proterozoic anorthosite complexes. Emslie (1980), Olson & Morse (1990), Wiebe (1990) and Mitchell *et al.* (1996) have interpreted these rocks as residual liquids of anorthosites, whereas Duchesne *et al.* (1989) have explicitly ruled out any comagmatic relationship.

Comparison of liquid evolution trends

Melt evolution as recorded by the fine-grained dyke rocks in the Lofoten Islands and the calculated melts is remarkably similar to evolution trends of tholeiitic and high-Al troctolitic liquids. Figure 6a compares the evolution of the liquids from Lofoten and the Laramie Anorthosite Complex with the trends reported from the Kiglapait intrusion in Labrador (Morse, 1981) and from the Skaergaard intrusion (Wager & Brown, 1967) and the Basistoppen sill in Greenland (Naslund, 1989). Figure 6b compares the Lofoten trend with the Skaergaard trend as calculated by Hunter & Sparks (1987) and with the



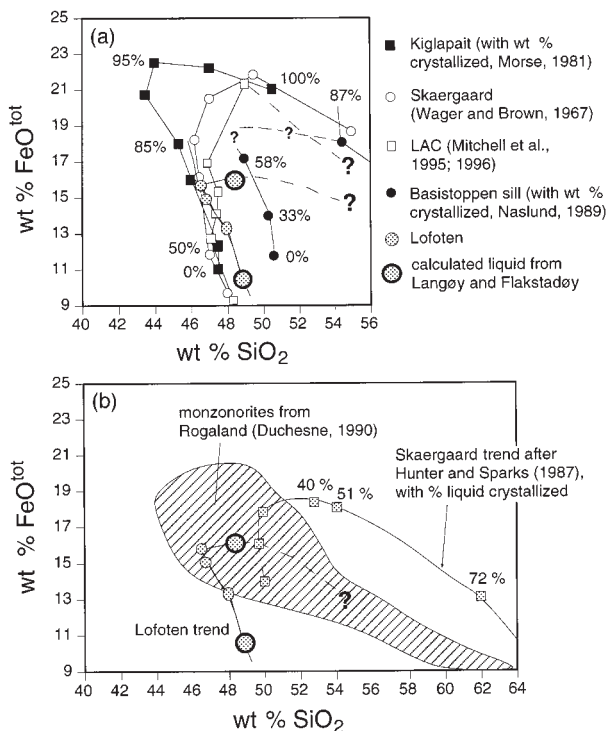


Fig. 6. (a) Fractionation trends in terms of FeO^{tot} and SiO₂ of the Lofoten and Laramie Anorthosite Complex high-Al gabbros (this study; Misra & Griffin, 1972; Mitchell *et al.*, 1995, 1996) and, for comparison, of the Skaergaard intrusion (Wager & Brown, 1967), the Kiglapait intrusion (Morse, 1981) and the Basistoppen sill (Naslund, 1989). (b) Fractionation trend of the Lofoten rocks as in (a), compared with the trends in the Rogaland monzonorites (Duchesne, 1990) and with the Skaergaard evolution as calculated by Hunter & Sparks (1987).

broad trend spanned by the Rogaland monzonorites of Duchesne (1990). With the exception of the Rogaland monzonorites, which were interpreted by Duchesne (1990) to be lower-crustal anatectic melts, all trends in Fig. 6a and b show initial, slightly variable Fe enrichment at nearly constant SiO₂ followed by SiO₂ enrichment and Fe depletion. The extensions of the Lofoten Islands and the Laramie Complex trends to higher SiO₂ values cannot be reliably retrieved and are therefore shown dashed in Fig. 6a and b. Hunter & Sparks (1987) and Toplis & Carroll (1996) pointed out that the amount of Fe enrichment depends on the composition and on the *f*_{O₂} of the parental magma. This argument further supports that the Lofoten rocks and calculated melts record an evolution trend typical of tholeiites.

Comparison with the calculations of Naslund (1989) and Morse (1981) suggests that the fraction of melt crystallized until the point where the Lofoten high-Al basalts or ferrodiorites attain their highest FeO^{tot} contents lies between 50 and 75 wt %. As the ferrodiorites are the liquids from which the anorthosites formed, there is a huge amount of these liquids to be crystallized after the formation of the anorthosites. During further fractionation and/or assimilation processes, these ferrodiorites may, for example, be used to produce the large amounts of mangerites and charnockites commonly observed to accompany the anorthosites and ferrodiorites in Proterozoic anorthosite complexes (e.g. Fuhrman *et al.*, 1988; Emslie & Hunt, 1990; Emslie, 1991; Emslie *et al.*, 1994).

GENERAL CONCLUSIONS

Two types of high-Al basalts

The parental melts of Proterozoic anorthosite complexes are high-Al basalts that fractionated to ferrodiorites (e.g. Emslie, 1980; Olson & Morse, 1990; Mitchell *et al.*, 1995, and references therein). It is important to note, however, that high-Al basalts associated with anorthosite complexes are chemically distinct from those that are characteristic of subduction environments (e.g. Myers *et al.*, 1986). The subduction-related high-Al basalts are higher in Al₂O₃, have relatively high *X*_{Fe} (e.g. Yoder & Tilley, 1962; Singer *et al.*, 1992) and much lower TiO₂ and Ni compared with the least evolved high-Al gabbros from Proterozoic anorthosite complexes (Emslie, 1980; Mitchell *et al.*, 1995). In addition, they do not exhibit the strong Fe enrichment that is characteristic of the anorthosite-related high-Al gabbros. Many of the high-Al gabbros from Proterozoic anorthosites show strong positive Eu anomalies, which are not characteristic of high-Al basalts from subduction environments (e.g. Myers *et al.*, 1986; Myers, 1988). Because of these geochemical differences, it is evident that the high-Al gabbros related to Proterozoic anorthosites have an origin distinct from that of the subduction-related high-Al basalts.

The origin of the high-Al basalts associated with anorthosite complexes can be evaluated using the analyses of high-Al gabbros from the Laramie Complex, which have a wide range of compositions (Mitchell *et al.*, 1995). One way to show this compositional range is on a plot of MgO (which monitors the crystallization of ferromagnesian

Fig. 5. Element oxide vs FeO^{tot}/(FeO^{tot} + MgO) plots of high-Al gabbros and ferrodiorites from the Laramie Anorthosite Complex (Mitchell *et al.*, 1995, 1996, filled symbols), of high-Al gabbros, low-Al gabbros and ferrodiorites from the Harp Lake Complex in Labrador (Emslie, 1980; hatched field), and of high-Al basalts and ferrodiorites from the Lofoten Islands [this study and 18 samples from Misra & Griffin (1972), open symbols]. (See text for discussion.)

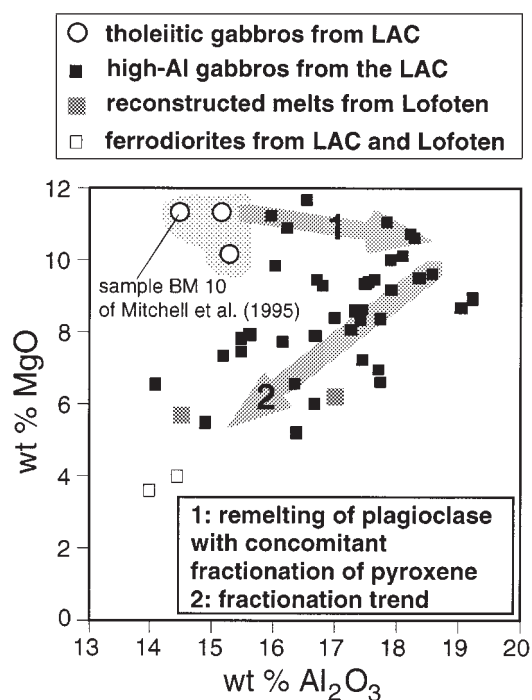


Fig. 7. MgO vs Al_2O_3 in fine-grained gabbros of the Laramie Anorthosite Complex (Mitchell *et al.*, 1995). This diagram shows the evolution trend from tholeiitic basalts to high-Al basalts by incorporation of earlier formed plagioclase and contemporaneous fractionation of clinopyroxene and/or orthopyroxene (1), and the subsequent fractionation trend of the high-Al basalt to ferrogabbros and ferrodiorites (2). (See text for discussion.)

silicates) vs Al_2O_3 (which monitors the crystallization of plagioclase) (Fig. 7). There are two apparent trends on this diagram. The first shows variable Al_2O_3 at slightly decreasing, but high values of MgO, whereas the second starts at high Al_2O_3 and MgO values and shows decreasing MgO with decreasing Al_2O_3 . We interpret these trends to indicate that two major processes were involved in the magmatic evolution of these rocks. Trend (1) shows a transition from a magma that is of normal tholeiitic composition to high-Al basalt, and Trend (2) is the differentiation trend from high-Al basalt to ferrodiorite (Figs 3 and 5).

The existence of the first trend implies that there is a mechanism by which tholeiitic basalts may evolve to the high-Al basalts of anorthosite complexes. As suggested by Wiebe (1990), we propose that resorption of plagioclase and contemporaneous fractionation of ortho- and/or clinopyroxene (which may be tentatively regarded as the megacrysts observed in many anorthosite complexes) during rise of the magma or during influx of new, hot magma is responsible for the formation of this type of high-Al basalts. We observe two justifications for this process: first, resorption of plagioclase combined with fractionation of, for example, Cr- or Ti-spinel will increase the melt in Al_2O_3

whereas MgO can stay relatively unchanged. Second, resorption of plagioclase also increases Eu in the melt, producing the positive Eu anomalies observed in many of the Proterozoic high-Al gabbros. The strongly positive Eu anomalies of the high-Al gabbros may also explain why ferrodioritic rocks often have low Eu anomalies or even lack them entirely (Mitchell *et al.*, 1996; Scoates & Frost, 1996). It has long been argued that these rocks cannot be residual after extensive crystallization of plagioclase because they lack a negative Eu anomaly. Apart from the process of plagioclase resorption, the derivation of melts with strongly positive Eu anomalies from the mantle is difficult to reconcile with the known mantle mineralogy and geochemistry.

The process of plagioclase remelting may produce a variety of high-Al basaltic magmas. Depending on the amount and composition of the plagioclase and the fractionating pyroxenes and on the amount of fractionation in the tholeiitic basalt before resorption of plagioclase, the high-Al basalts can show various X_{Fe} , CaO or SiO_2 values. Thus, then the samples with the lowest Al_2O_3 and highest MgO should represent the least evolved magma. Such rocks are found in the Laramie Anorthosite Complex (e.g. sample BM-10), where they have compositions typical of tholeiites. Sample BM-10 has no Eu anomaly, high Ni and Cr contents, ~11 wt % MgO and ~14 wt % Al_2O_3 (Table 1). Extrusive rocks with nearly identical composition are found in Hawaii (Macdonald & Katsura, 1964) and on the Hebrides (Carmichael *et al.*, 1974), the former representing an oceanic, the latter a continental tholeiitic province.

Plagioclase compositions and density contrasts

Proterozoic anorthosites are characterized by plagioclase of the composition $\text{An}_{45}\text{--}\text{An}_{55}$ and rarely contain plagioclase above An_{60} . This observation may be explained by the experimental constraints of Fram & Longhi (1992) and Longhi *et al.* (1993), according to which more calcic plagioclase is not a liquidus phase of anorthositic parental melts at high pressures. However, we propose an additional process that may control plagioclase composition in Proterozoic anorthosites. At low pressures the density contrast between more calcic plagioclase and basaltic melt is too small to permit efficient accumulation at the top of the deep-seated magma chamber. Figure 8 shows results of calculations with the MELTS program of Ghiorso *et al.* (1994), where the densities of the least fractionated high-Al gabbros, of tholeiites, of the calculated melts, of a ferrodiorite and of various plagioclase compositions were calculated at 10 and 4 kbar. Figure 8 shows that

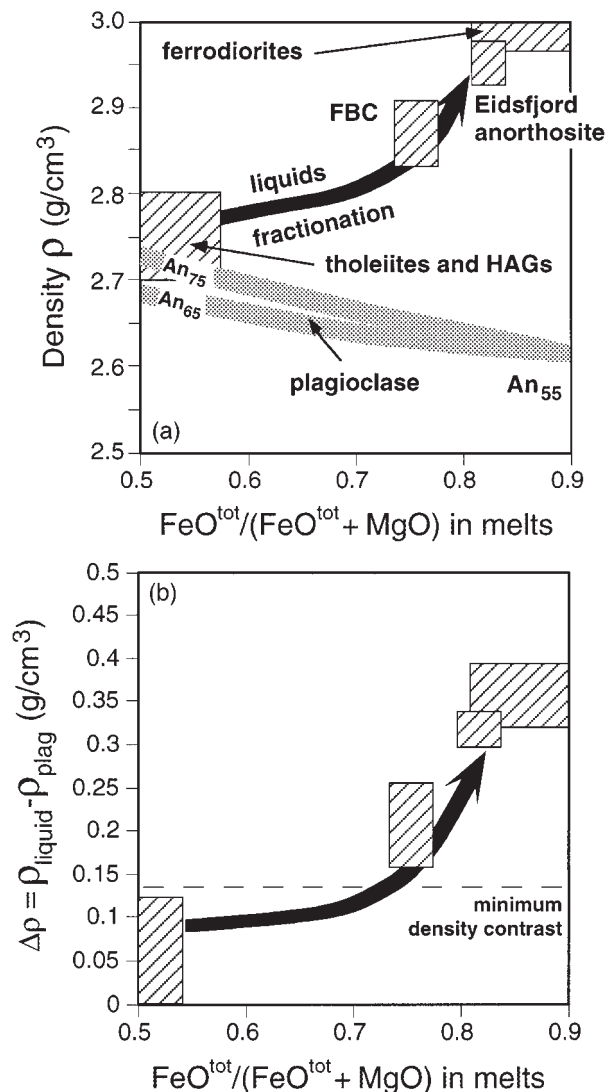


Fig. 8. (a) The density of tholeiitic liquids like BM-10 and primitive high-Al gabbros, of the reconstructed anorthositic melts and of ferrodioritic liquids (calculated using MELTS; Ghiorso *et al.*, 1994) plotted against the $\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}} + \text{MgO})$ ratio of the liquid as a monitor of differentiation. Also shown is the density of plagioclase that can be in equilibrium with these melts at pressures between 4 kbar (An_{75}) and 10 kbar (An_{65}). (b) The density contrast between plagioclase and the coexisting liquid. It should be noted that the plagioclase compositions found in anorthosites record a density contrast of 0.15 or higher, indicating that at lower contrasts or high $X_{\text{An}}^{\text{plag}}$ the plagioclase is not able to accumulate at the top of a deep-seated magma chamber.

the density contrast during early stages of fractional crystallization is around 0.1–0.15 g/cm³, until the plagioclase reaches a composition of about An_{60} and, especially, until the melt reaches high values of Fe and of X_{Fe} . Thus, plagioclase of $\text{An}_{>60-65}$ probably is not able to float in primitive tholeiitic magmas.

The original magma to start with: a tholeiitic basalt

The arguments above imply that the parental melts to anorthosite complexes are not of extraordinary composition; they are tholeiitic. The controlling parameter in the formation of Proterozoic anorthosites is probably not the melt composition, but the specific crystallization history (resorption of plagioclase in melts), and this in turn may be strongly influenced by the tectonic setting. The most obvious way for resorption to happen is as a result of movement of a high-temperature tholeiitic magma through a deep-seated magma chamber containing intermediate plagioclase. Because intermediate plagioclase will crystallize from relatively unfractionated tholeiitic melt only at high pressures (Fram & Longhi, 1992; Longhi *et al.*, 1993), this leads directly to the two-stage model of anorthosite formation (e.g. Emslie, 1978; Duchesne, 1984; Longhi & Ashwal, 1985; Wiebe, 1992; Ashwal, 1993; Emslie & Hegner, 1993; Longhi *et al.*, 1993; Emslie *et al.*, 1994; Scoates & Frost, 1996). Plagioclase stable at high pressures could be remelted either because of a pressure decrease (Longhi *et al.*, 1993) related to convective lithospheric thinning (Corrigan & Hanmer, 1997) or because of injection of new hot magma into the magma chamber in a plume-related environment (e.g. Hoffman, 1989). Under high-pressure conditions at the crust-mantle boundary, the stable crystallizing plagioclase would be of intermediate composition and would float in the crystallizing magma; continuous fractionation of these high-Al melts can eventually lead to anorthositic cumulates buoyant enough to rise to high crustal levels as crystal-rich mushes (Fram & Longhi, 1992; Wiebe, 1992; Ashwal, 1993; Longhi *et al.*, 1993; Emslie *et al.*, 1994; Lafrance *et al.*, 1996) whereas the melt entrained in this crystal mush continues to fractionate. During ascent and emplacement at mid-crustal levels, the successively more evolved liquids may be removed at various stages of fractionation because of filter-pressing. Our results strongly support that these liquids are the ferrodiorites observed in most anorthosite complexes.

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