

Micro-scale element migration during eclogitisation in the Bergen arcs (Norway): A case study on the role of fluids and deformation

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Abstract

Understanding chemical element mobility during high-pressure metamorphism is paramount to the knowledge of the transformations occurring during the course of eclogite-facies metamorphism. In particular, the role of deformation and fluid circulation appear essential. In order to better decipher this role, we investigated two eclogite-facies samples from the Lindås Nappe in the Bergen arcs (Norwegian Caledonides), which are related to a fluid/deformation metamorphic event. This study, based on detailed microtextural investigations, coupled with major and trace element analyses, either in situ EMPA and LA-ICP-MS or by conventional ICP-MS nebulisation on mineral fractions, illustrates the complexity of chemical reactions accompanying the eclogite- and amphibolite-facies metamorphism. A single eclogite-facies mineral species two eclogite-facies (i.e. phengite and epidote) depicts various chemical features depending on its textural location at thin-section scale. The correlation with the mineralogical reaction occurring in each textural site demonstrates that the compositional variations are inherited from the precursor minerals. Thus, the composition of eclogite-facies minerals is locally controlled by the composition of few hundred-micron domains. On the contrary, alteration phases of the amphibolite facies (symplectite and calcic amphibole) display clear enrichment in LILE, Pb, Sr and LREE compared to their precursor minerals (omphacite and garnet). This supports that the transport of elements at hand-sample scale was enhanced during the retrogression and elements were efficiently delocalised from one textural site to another.

The contrasting behaviour of element mobility between peak eclogite-facies metamorphism and retrogression can be explained in terms of rate processes and fluid/deformation activity. During eclogitisation, deformation allows the formation of textural and mineralogical microdomains. Contemporaneously the fluid circulation favours the development of heterogeneities by transport of elements in excess towards areas protected from deformation where phengite-bearing quartz lenses crystallise. This, together with the very fast recrystallisation of eclogite-facies minerals, results in a heterogeneous redistribution of elements at sample scale and in a chemical disequilibrium between different microdomains of hundred-micron size. During retrogression in the amphibolite facies, coeval with long-term ductile deformation, the fluids act as an efficient vector to redistribute the elements from a given textural site to another and, thus, allow to reach, at hand-sample scale, a new chemical equilibrium between the different crystallising phases.

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1. Introduction

Advances in analytical techniques now allow rapid, accurate and precise in situ analyses of trace elements, which are powerful to link element behaviour to petrogenetical processes (e.g. Tribuzio et al., 1996; Bea and Montero, 1999; Rubatto and Hermann, 2003). However some geochemical analyses such as partition coefficient determination or geochronological investigations, still need a larger sampling scale (mineral fractions) because minerals of interest are not rich enough in target elements, and also because isotope radiometric analyses are not easily performed in situ at the present time (i.e. Rb/Sr and Sm/Nd datings). When performing routine mineral-fraction dating or mineral partition coefficient determinations, the validity of the results depends, in particular, on the achievement, at a sample scale, of an isotopic and chemical equilibrium between the different mineralogical phases. Several previous studies have outlined the difficulty to obtain meaningful isotopic data from mineral separates, which is often related to a hand-sample scale chemical disequilibrium (Mørk and Mearns, 1986; Thöni and Jagoutz, 1992, 1993; Cliff et al., 1998; Zheng et al., 2002). The lack of penetrative deformation and intensive fluid circulation are the main reasons commonly put forward to explain such an incomplete re-equilibration process. The importance of synmetamorphic deformation to achieve a complete redistribution of chemical elements at sample scale has been previously discussed by several authors (i.e. for in situ analyses, Messiga et al., 1995; Sassi et al., 2000; Zack et al., 2002). It has been shown that deformation acts as a catalyst for recrystallisation processes thus favouring element redistribution and that fluid circulation enhances element mobility, allowing a better homogenisation of the system (Walter and Wood, 1984; Brodie and Rutter, 1985; Austrheim, 1986; Rubie, 1986; Wayte et al., 1989; Austrheim, 1990; Rubie, 1990; Zhang and Liou, 1997; Straume and Austrheim, 1999; Wain et al., 2001; John and Schenk, 2003).

The present work is focused on the redistribution of chemical elements that took place in eclogites from the Bergen Arcs system (Norway). Previous studies have demonstrated that deformation and fluid circulation are the main factors responsible for the transformation of dry granulites in eclogites during the Caledonian orogeny (e.g. Austrheim, 1986). The objective is two-folds: (1) investigate elements behaviour at thin-section scale to estimate the influence of parameters such as deformation and fluid circulation on element redistribution in the course of HP metamorphism; but also, (2) to test the scale of homogeneity of a single mineral phase in a hand-

specimen sample in view to constrain the chemical characteristics of the mineral fractions used for other investigations such as radiometric dating. Thus, detailed in situ electron microprobe (EMP) major elements analyses and in situ trace elements analyses, by laser-ablation inductively coupled-plasma mass-spectrometry (LA-ICP-MS), were coupled to conventional-ICPMS trace element analyses of mineral fractions. Several mineralogical fractions (up to nine) of a single mineral phase have been separated as a function of grain-size, density and magnetic properties. These fractions are constituted by phengite, further used for Rb/Sr dating and by garnet, used for Sm/Nd geochronology. The different micro textural sites and related mineralogical reactions operating during the eclogite-facies metamorphic event and the subsequent retrogression in the amphibolite-facies have been determined. Precise and extensive analyses of chemical composition (major and trace elements) have been performed on inherited granulite-facies minerals, neoformed eclogite-facies minerals and secondary amphibolite-facies minerals. The scale of chemical element redistribution has been estimated for different textural sites appearing during the eclogite- and amphibolite-facies metamorphic transformations and the role of infiltrating fluid (as a vector for transport of elements at sample scale) has been evaluated. At last, the effects of a deformation synchronous to a metamorphic event have been investigated in relation to element chemical redistribution.

2. Geological setting

The Bergen Arcs System (BAS), in the south-western part of the Norwegian Caledonides, is characterised by a series of Caledonian nappes, thrust onto a Precambrian parautochthonous gneissic complex (Fig. 1). Samples studied in the present work come from one of these thrust sheets, the Lindås Nappe, and were collected on the Holsnøy Island.

The Lindås Nappe extends over an area of more than 1000 km² and mainly consists of a Proterozoic anorthositic complex affected by two distinct metamorphic/deformation events. A first penetrative Grenvillian (Sveconorwegian) metamorphic event took place in the granulite facies (0.8–1.1 GPa, 780–900 °C; Austrheim and Griffin, 1985; Kühn, 2002) at ca. 930 Ma (Bingen et al., 1998, 2001) and a fluid/deformation Caledonian metamorphic event occurred in the eclogite facies (1.8 to 2.1 GPa, ~ 700 °C; Jamtveit et al., 1990) at ca. 425 Ma (Bingen et al., 1998, 2001; Glodny et al., 2002; Kühn, 2002) followed by retrogression in the amphibolite facies (1 to 1.2 GPa, 650–690 °C; Boundy

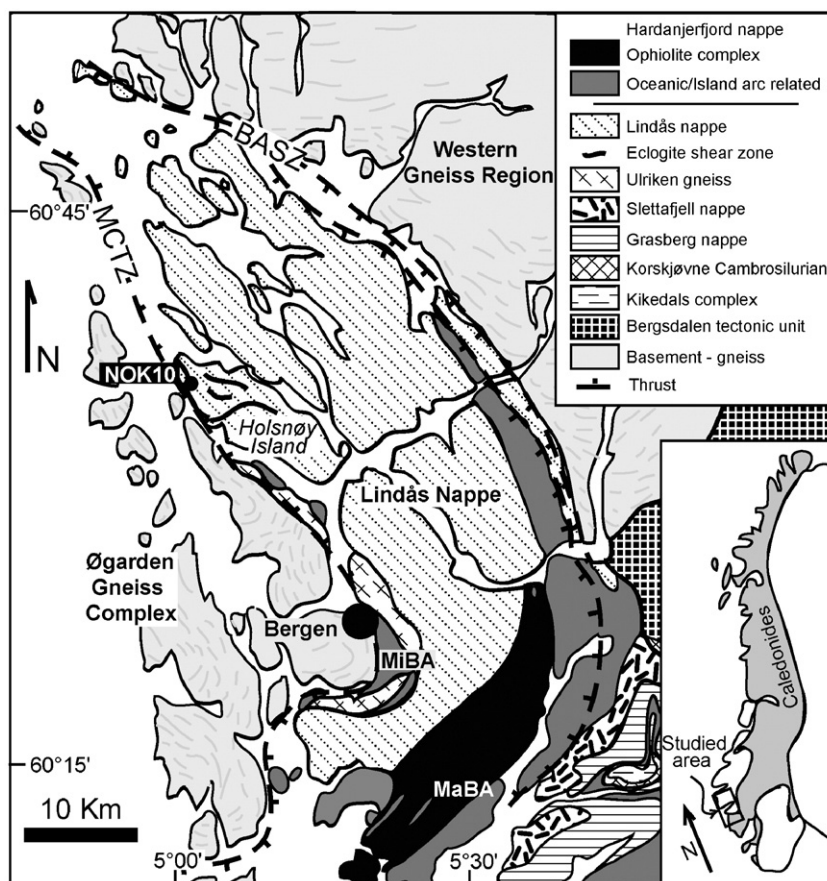


Fig. 1. Geological map of the Bergen Arcs area with location of the sampling site (NOK10). Modified after Boundy et al. (1996) and Kühn et al. (2000).

et al., 1996; Kühn, 2002) at ca. 411 Ma (Austrheim and Råheim, 1981; Austrheim, 1990; Bingen et al., 2001; Glodny et al., 2002). The Lindås Nappe exposes an alternation of anorthosites s.s., gabbroic anorthosites and anorthositic gabbros locally intruded by jotunite (hypersthene monzodiorite or monzodiorite) and mangerite (hypersthene monzonite; Austrheim and Griffin, 1985; Austrheim, 1990).

The most significant features recorded by the rocks from the Lindås Nappe during the eclogite-facies metamorphism (Austrheim and Griffin, 1985; Austrheim, 1986; Austrheim, 1990; Jamtveit et al., 1990; Andersen et al., 1991a,b; Klaper, 1991; Boundy et al., 1992; Erambert and Austrheim, 1993; Austrheim and Boundy, 1994; Kühn, 2002) can be summarized as follows: (1) The entire Lindås Nappe underwent a metamorphism in the eclogite facies with preservation of metastable granulites. (2) The eclogitisation is located along zones of tectonic weakness that developed at every scale, i.e. from mineral- to regional-scale. (3) The eclogitisation affected all granulites types, independent-

ly on the nature of their protolith (anorthosite, gabbro or jotunite). (4) A large part of the identified eclogite-facies reactions needs an excess of water. All together, these various observations led the authors to conclude that the partial eclogitisation of the granulites from the Lindås Nappe is mainly controlled by deformation and hydrous fluid percolation.

The favoured origin for the eclogitising fluids is the dehydration of sedimentary rocks below granulites during the subduction and the related exhumation stages (Jamtveit et al., 1990; Mattey et al., 1994; Van Wyck et al., 1996). The fluids have percolated along fractures and shear zones and the resulting volume of impregnated rocks exceeds several cubic kilometres (Jamtveit et al., 1990). However a local consumption of the hydrous fluid during eclogite-facies reactions, resulting in the chemical evolution of the fluid in a closed system (Boundy et al., 1997; Svensen et al., 1999; 2001; Kühn, 2002), is favoured to a wide fluid flow passing through the Lindås Nappe via zones of weakness. Moreover, contrasting

fluid compositions at the different stages of the Caledonian metamorphism have been recognised (Andersen et al., 1990, 1991a,b; Matthey et al., 1994; Van Wyck et al., 1996). This suggests several events of fluid infiltration.

3. Analytical techniques

Whole rock major element analyses were performed at the Centre of Mineral Analyses from Lausanne University in Switzerland on a Philips PW 2400XRF spectrometer. Precision on oxide percent measured is about 1% and accuracy as well as the reproducibility was achieved by analyses of rock standards from various lithologies (BHVO-1, basic rock; STM, syenite; AGV, andesite; VH1, granodiorite). Trace and Rare Earth Element determinations on whole-rocks and mineral fractions were measured by conventional nebulisation ICP-MS using a VG Plasmaquad 2 (precision of 3–5%) at ISTEEM, Montpellier II University. Digestion procedures and standards used are similar to those described in Ionov et al. (1992). The detection limit is of ca. 1 ppt, and blank values range from 1 ppb to 100 ppt for LILE, 10 ppt to 1 ppt for REE and, ~10 ppt for HFSE. International standards BEN and UBN ($n=9$) have been repeatedly analysed during this study and the measured values are in close agreement with standard reference values. The accuracy of trace element analyses for REE-poor minerals such as phengite was controlled by measurements of duplicates at a one-year timespan (duplicates are labelled “re”, Tables III, IVD in Electronic Supplementary Material). The reproducibility of LREE is more precise than that obtained for HREE due to the very low contents of HREE in phengites (around 1 ppb).

In situ trace elements analyses were achieved by laser ablation (LA-) ICP-MS (ISTEEM, Montpellier II University) using an Excimer ArF laser ($\lambda=193$ nm) coupled to a quadrupole VG Plasmaquad 2 ICP mass spectrometer. Analyses were performed on c.150 μm thick sections and the laser was operated at a repetition rate of 5 Hz using an energy density of 15 J/cm². Laser spot sizes were 51 or 77 μm depending of the element concentrations in the mineral analysed. Calibration was done using NIST-612 and NIST-610 standard glasses using the values reported by Pearce et al. (1997). During analyses, oxide level, measured using the ThO/Th ratio, was below 0.5%. Concentration values were adjusted using Si and Ca (EPM analyses) as internal standards. Precision was in the range 5–10% (1 σ RSD%) for most elements.

Major and minor element mineral analyses were performed at Montpellier II University on a microprobe CAMECA SX100. Analytical conditions were a 20 kV accelerating voltage and 10 nA beam current. Both synthetic and natural SiO₂, Al₂O₃, TiO₂, Fe₂O₃, Cr₂O₃, albite, orthose, forsterite, wollastonite and rhodonite were used as calibration standards. Fe³⁺ in garnet, omphacite, amphibole, and epidote have been calculated after normalisation to 12 O and 8 cations, 6 O and 4 cations (Cawthorn and Collerson, 1974), 23 O and 13 cations (Leake et al., 1997) and 12.5 O and 8 cations (Fe_{total}=Fe³⁺), respectively. Mineral and end-member abbreviations are after Kretz (1983), except calcic amphibole (Ca-Am), phengite (Ph) and symplectite (Sy).

Table 1
Whole rock geochemistry of eclogites NOK10a and NOK10d

	NOK 10a: ky-rich eclogite	NOK 10d: ph-rich eclogite
SiO ₂	48.71	50.97
Al ₂ O ₃	22.42	16.76
Fe ₂ O ₃	7.36	11.67
MnO	0.07	0.18
MgO	4.63	4.76
CaO	11.77	9.74
Na ₂ O	3.18	3.05
K ₂ O	0.61	1.94
TiO ₂	0.24	0.60
P ₂ O ₅	0.54	0.03
Sr	0.10	0.10
Total	99.63	99.80
Rb	0.98	4.6
Sr	979	1106
Y	0.91	7.9
Zr	1.5	11
Nb	0.065	0.178
Cs	0.008	0.027
Ba	320	1169
La	1.5	4.5
Ce	2.8	8.9
Pr	0.35	1.29
Nd	1.62	6.87
Sm	0.31	1.75
Eu	0.51	1.13
Gd	0.27	1.78
Tb	0.034	0.25
Dy	0.19	1.52
Ho	0.034	0.31
Er	0.87	0.85
Tm	0.012	0.12
Yb	0.73	0.80
Lu	0.012	0.14
Hf	0.058	0.53
Pb	1.12	4.4
Th	0.016	0.005

Major elements expressed as weight percent oxides and trace elements in ppm.

4. Petrological results

4.1. Sample presentation

This study focused on two samples collected on the Island of Holsnøy (Fig. 1). In this area, eclogites occur as: (1) “static eclogites” which consist in a metasomatic front of few centimetres-thick propagating on both side of brittle fractures in the granulites; (2) “eclogite breccias” where granulite blocks are embedded by eclogite shear zones with granulites partially eclogitised at the edge; (3) “eclogite shear zones” of various scale (from 10 cm to 100 m thick and several kilometres long); (4) “pseudotachylytes” (Austreim and Boundy, 1994; Austreim, 1986, 1990; Klaper, 1991; Boundy et al., 1992). In order to address the effect of deformation on element migration, samples from occurrence (2) and (3) were chosen. Sample NOK10a comes from an anorthositic protolith and was collected at the edge of a granulitic boudin showing strong eclogitic shearing (occurrence (2)). Sample NOK10d was taken from a gabbroic protolith and was collected at the core of an eclogitic shear-zone (occurrence (3)), few meters away from sample NOK10a. Both samples are about 25 dm³ in size and are mostly equilibrated during the eclogite-facies metamorphism. They are only partially retrogressed in the amphibolite facies (Caledonian metamorphism), but relics of the granulite-facies Grenvillian metamorphism are still present. Sample NOK10a displays an alternation of thin white and green crenulated layers (1–2 mm wide) defining the

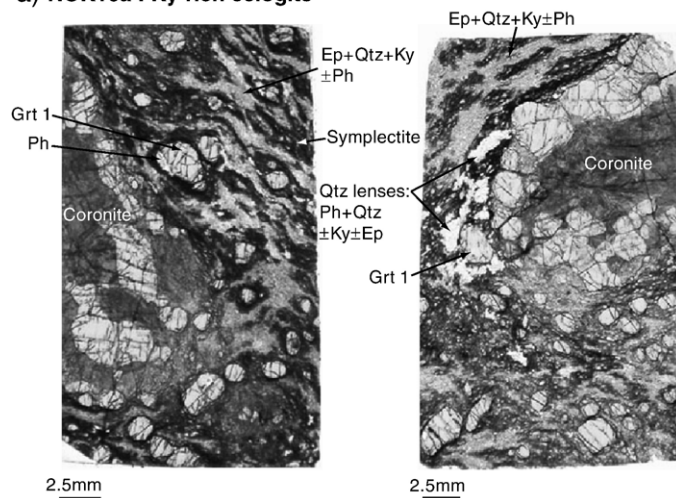
eclogitic foliation. The green layers are made of omphacite and also contain garnet surrounded by phengite; the white layers mostly contain epidote, kyanite and rare phengite. Coronites (pyroxenes surrounded by garnet corona) are also observed and are elongated in the eclogitic foliation (~1 cm wide, ~2–3 cm long). Sample NOK10d displays green augens (~3 mm wide, ~5 mm long) made of omphacite, surrounded by an anastomosed network of white layers (~1 mm wide) defining the eclogitic foliation. These layers contain epidote, phengite and quartz. Garnet (up to 2 mm wide) is located mostly at the edge of the omphacite augens. NOK10a is kyanite-rich and phengite-poor, whereas NOK10d is kyanite-free and phengite-rich. NOK10a is significantly richer in Al₂O₃ and CaO (Table 1) compared to NOK10d, the latter containing higher Fe₂O₃ and K₂O amounts (Table 1).

4.2. Texture and mineral assemblages

4.2.1. Inherited structures from the original granulite

Both samples contain textural and mineralogical relics from the previous Sveconorwegian granulitic stage. In the Ky-rich sample (NOK10a), these relics consist of coronites constituted by a core of diopside with minor hypersthene and by a rim of garnet (Figs. 2a and 3a). The coronites have been stretched during the eclogite shearing and offer only few clues on the eclogite-facies overprint that will be discussed latter. Garnet inherited from the granulite stage occurs

a) NOK10a : Ky-rich eclogite



b) NOK10d : Ph-rich eclogite

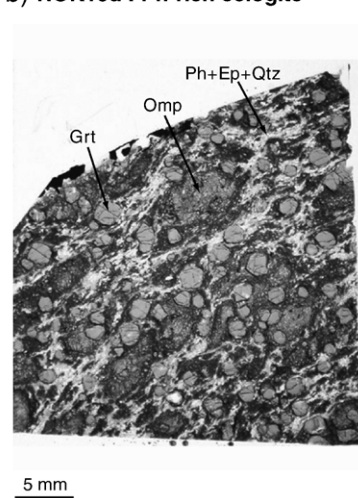


Fig. 2. 150 µm-thick section photomicrographs of the two eclogite samples NOK10a (a) and NOK10d (b) in natural light. Abbreviations after Kretz (1983), except for phengite (Ph).

in both samples. In the Ky-rich sample, it is partially obliterated while it is fully preserved in the Ph-rich sample (Figs. 3e and 4e). Finally, the original granulite layering expressed by mm-scale variations in the modal proportion of garnet-diopside and plagioclase, though partially or totally replaced by eclogite-facies minerals and stretched during coeval deformation, is still visible (Fig. 2a, b).

4.2.2. Progression of the eclogitisation

The plagioclase-rich layers from the granulite have been totally replaced, in both samples, by medium grained epidote-rich layers during eclogitisation. The eclogite-facies assemblages in these layers are $\text{Ep} + \text{Ky} + \text{Qtz} \pm \text{Ph}$, in Ky-rich sample and $\text{Ep} + \text{Ph} + \text{Qtz}$ in Ph-rich sample. Eclogite-facies minerals are equigranular ($\sim 100 \mu\text{m}$) and aligned in the eclogitic foliation. Epidote

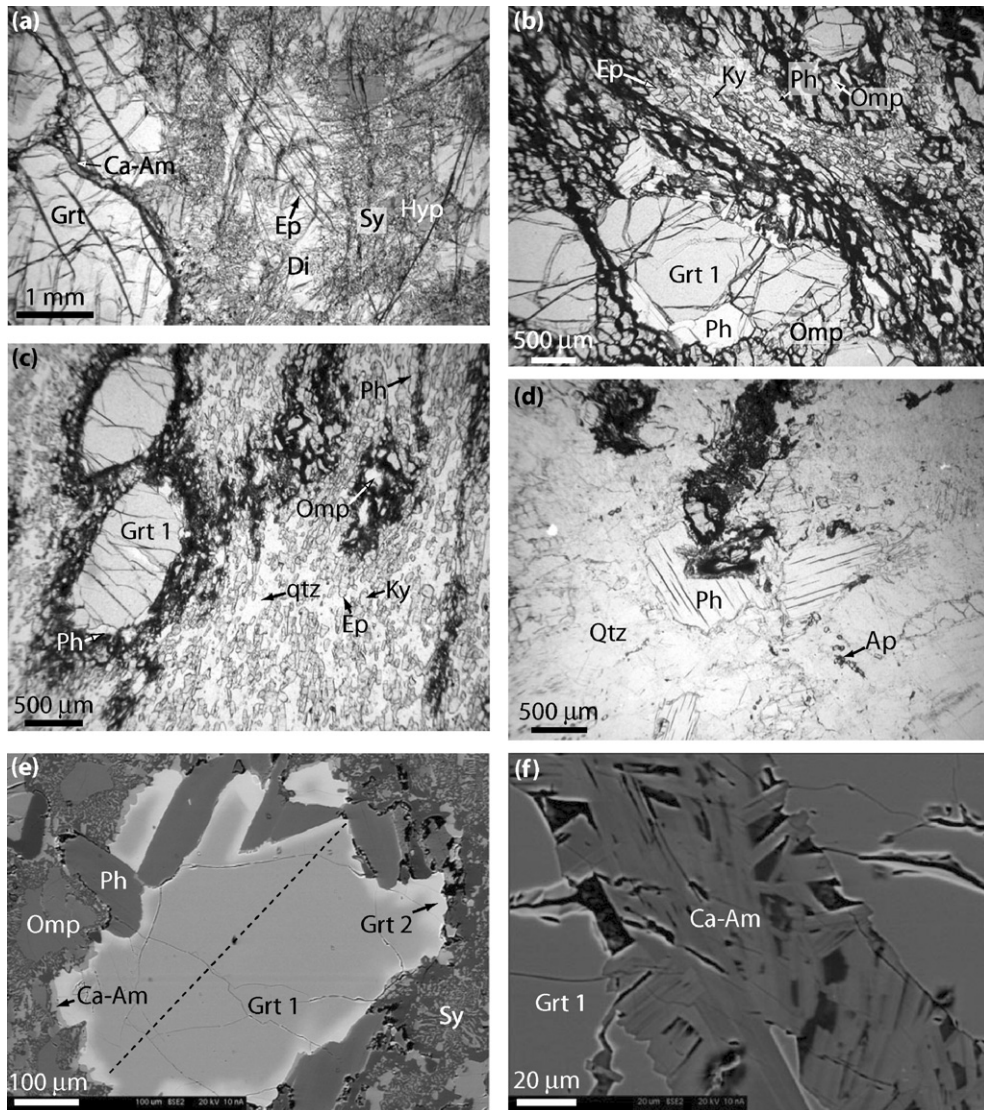


Fig. 3. 30 μm -thin section photomicrograph of sample NOK10a in natural light (NL) and back-scattered electron (BSE). (a, NL) Detail of a coronite constituted by diopside and hypersthene at the core and rimmed by garnet. The pyroxene inherited from the granulite stage is partially eclogitised. Epidote needles appear in diopside partially destabilised in omphacite (Sy for symplectite). (b, LN) Garnet/omphacite area where omphacite is partially converted in symplectite (Amp+Ab) at the edge. Phengite replaces garnet at the edge. (c, LN) The eclogitic foliation is outlined by the assemblage $\text{Ep} + \text{Qtz} + \text{Ky} + \text{Ph}$. Garnet augen (Grt1) inherited from the granulite stage is rimmed by retrograde symplectite. (d, LN) Detail of quartz lenses with large phengite grains. (e, BSE) Two distinct garnet generations: the granulite-facies garnet (Grt1) forms the dark core and the eclogite-facies garnet (Grt2) forms the clear rim. Dashed lines give the orientation of profile shown in Fig. 5a. Note the occurrence of phengite at the garnet rim. (f, BSE) Detail of retrograde calcic-amphibole filling garnet fracture.

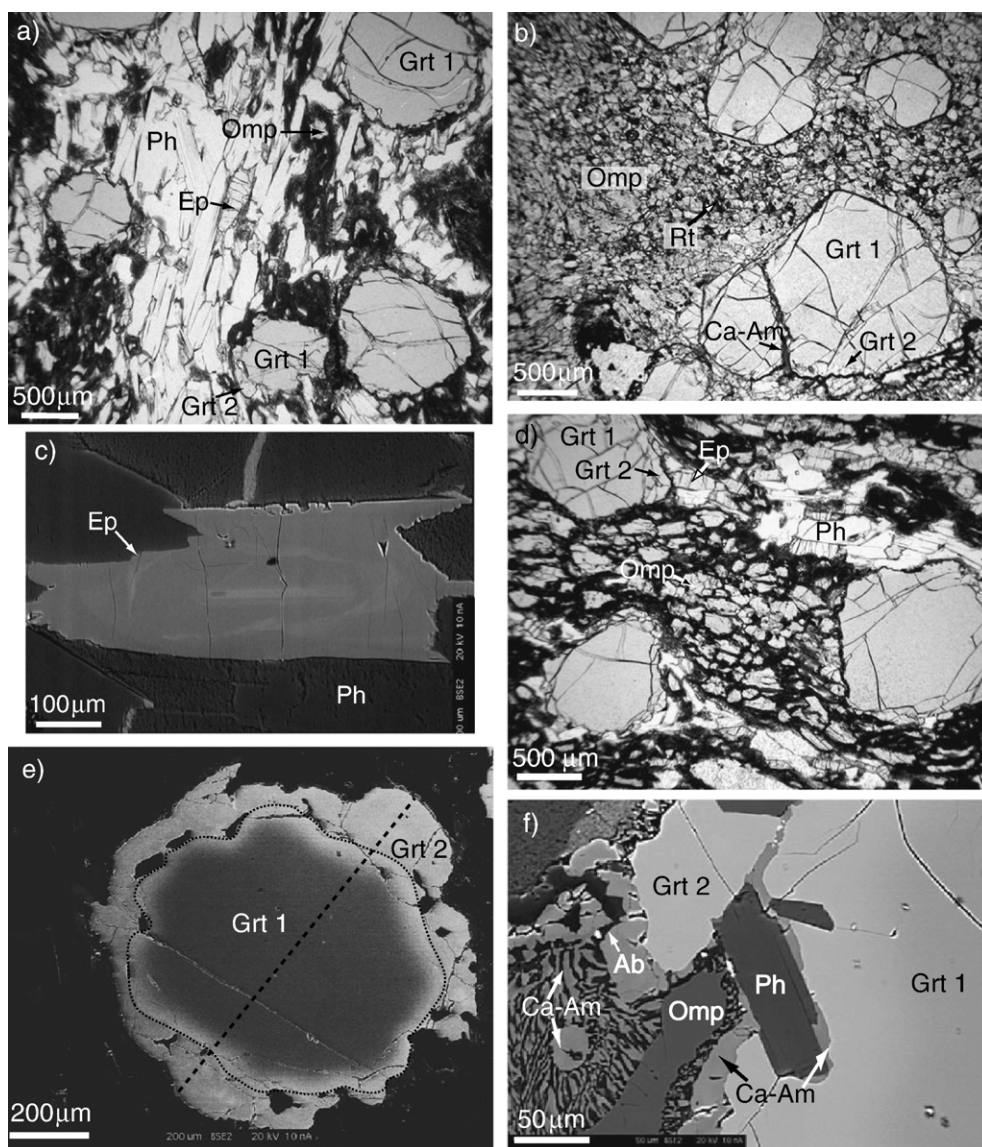
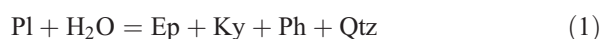


Fig. 4. 30 μm -thin photomicrograph from sample NOK10d in natural light (NL) and back-scattered electron (BSE). (a, NL) Detail of the eclogitic foliation characterised by the assemblage $\text{Ph} + \text{Ep} + \text{Qtz}$. (b, LN) Detail of garnet omphacite augen. Numerous rutile crystals occur between the omphacites. (c, LN) Epidote surrounded by phengite. (d, LN) Garnets from the foliation flanked by epidote- and phengite-rich areas (e, BSE) Granulite-facies garnet (Grt1) showing an eclogite-facies garnet (Grt2) overgrowth. The white black dotted-line delineates the granulite-facies garnet. The dashed line gives the orientation of the profile shown in Fig. 5b. (f, BSE) Details of inclusions in secondary eclogite-facies garnet. The breakdown of eclogite-facies minerals during exhumation in the amphibolitic facies can be seen.

displays patchy optical domains, with 10 μm -width acicular core and two overgrowth areas. The eclogitisation reaction (1), for the Ky-rich sample, can be written as follows (non stoichiometric):



In the Ph-rich eclogite, plagioclase has been replaced according to reaction (2).



The presence of mica and epidote implies participation of H_2O in this reaction. The K-rich component allowing phengite crystallisation arises from exsolution microtextures of K-feldspar antiperthitic grains present in the crystalline network of the granulite-facies plagioclase (Austrheim, 1990). The excess of Na from these reactions enters omphacite. Finally, the Fe–Mg component of

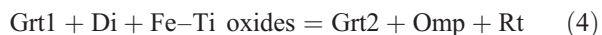
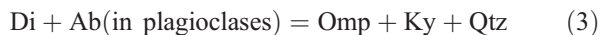
Table 2

Selected representatives EMP analyses of minerals from the Ky-rich eclogite (sample NOK10a)

NOK 10a: Ky-rich eclogite	Garnet		Epidote			Phengite							
	Grt 1	Grt 2	Coronite	Foliation core	Foliation rim	Core	Rim	Core	Rim	Core	Rim	Rim	Core
	Pyrope	Almandin	Czo	Czo	Czo	Ph from epidote area	Ph from garnet area	Ph from quartz area	Ph from quartz area	Ba-rich phengite			
SiO ₂	40.78	39.32	37.90	39.42	39.06	46.75	45.95	47.20	46.65	47.02	46.70	46.45	47.24
TiO ₂	0.23	0.24	0.26	0.05	0.04	0.52	0.82	0.45	0.43	0.33	0.34	0.38	0.62
Al ₂ O ₃	23.11	22.28	29.80	32.86	32.37	33.52	33.00	33.52	33.44	34.80	34.53	32.29	32.17
Fe ₂ O ₃	0.00	0.00	5.27	1.58	2.16	—	—	—	—	—	—	—	—
MgO	13.33	8.08	0.20	0.03	0.00	1.70	1.68	1.76	1.61	1.48	1.31	2.07	2.18
FeO	14.62	22.03	—	—	—	1.62	1.63	1.64	1.99	1.20	1.50	1.83	1.63
MnO*	0.24	0.78	0.01	0.02	0.02	0.01	0.01	0.01	0.00	0.00	0.00	0.00	0.00
CaO	7.61	7.30	24.53	21.18	21.10	0.00	0.03	0.01	0.03	0.02	0.00	0.01	0.02
BaO	—	—	0.00	—	—	0.80	0.85	—	—	0.70	0.62	1.16	1.06
Na ₂ O	0.02	0.03	0.00	0.02	0.02	1.38	0.51	1.51	0.81	2.14	1.26	0.75	1.07
K ₂ O	0.00	0.02	0.00	0.00	0.00	8.96	10.42	8.17	9.26	7.83	8.97	9.72	9.32
Total	99.94	100.08	97.97	95.16	94.77	95.27	94.89	94.27	94.22	95.52	95.22	94.66	95.31
<i>Cations for 12 oxygens</i>			<i>12.5 oxygens</i>			<i>11 oxygens</i>							
Si	3.00	3.00	2.93	3.05	3.04	3.12	3.10	3.15	3.13	3.10	3.11	3.14	3.16
Ti	0.01	0.01	0.02	0.00	0.00	0.03	0.04	0.02	0.02	0.02	0.02	0.02	0.03
Al	2.00	2.00	2.72	3.00	2.97	—	—	—	—	—	—	0.86	0.84
AlIV	—	—	—	—	—	0.88	0.90	0.85	0.87	0.90	0.89	1.72	1.70
AlVI	—	—	—	—	—	1.76	1.73	1.78	1.78	1.81	1.81	—	—
Fe ³⁺	0.00	0.00	0.31	0.09	0.13	—	—	—	—	—	—	—	—
Mg	1.46	0.92	0.02	0.00	0.00	0.17	0.17	0.17	0.16	0.15	0.13	0.21	0.22
Fe ²⁺	0.90	1.41	—	—	—	0.09	0.09	0.09	0.11	0.07	0.08	0.10	0.09
Mn*	0.02	0.05	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca	0.60	0.60	2.04	1.76	1.76	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ba	—	—	0.00	—	—	0.02	0.02	—	—	0.02	0.02	0.03	0.03
Na	0.00	0.00	0.00	0.00	0.00	0.18	0.07	0.20	0.11	0.27	0.16	0.10	0.14
K	0.00	0.00	0.00	0.00	0.00	0.76	0.90	0.69	0.79	0.66	0.76	0.84	0.80
Total	7.99	7.99	8.04	7.90	7.91	7.01	7.02	6.96	6.97	6.99	6.98	7.02	7.00
Xalm	0.30	0.47											
Xpyr	0.49	0.31											
Xgro	0.20	0.20											
Xspe	0.00	0.02											
Xczo			0.70	0.91	0.88								
Xep			0.30	0.08	0.12								
Na+K						0.94	0.97	0.89	0.90	0.93	0.92	0.94	0.93
Mg#						0.65	0.65	0.66	0.59	0.69	0.61	0.67	0.70
Na#						0.19	0.07	0.22	0.12	0.29	0.18	0.11	0.15

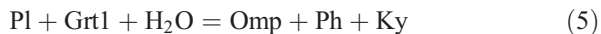
phengite comes from the local breakdown of granulite-facies diopside in omphacite.

In the granulite-facies garnet–diopside-rich layers, diopside has been entirely replaced by omphacite according to reaction (3) for Ky-rich sample and according to reaction (4) for Ph-rich sample.



Reaction (4) points to the crystallisation of secondary garnet (Grt2) surrounding primary granulite-facies garnet

(Grt1) (Fig. 4e–f). The boundary between Grt1 and Grt2 is outlined by an inclusion-rich corona of eclogite-facies minerals (i.e. phengite and omphacite). By contrast to the Ph-rich sample, garnet from the Ky-rich sample displays an irregular rim suggesting its breakdown according to reaction (5) and its partial preservation (Fig. 3e).



This converges with previous observations on the Holsnøy eclogites (Erambert and Austrheim, 1993).

In coronites from the Ky-rich sample, eclogitisation can be traced by the exsolution of epidote lamella in

Table 2 (continued)

NOK 10a: Ky-rich eclogite	Clinopyroxene		Amphibole		Plagioclase
	Coronite	Foliation	Garnet veins		Symplectite
	Diopside	Omphacite	Pargasite	Mg-taramite	Albite
SiO ₂	50.04	54.89	37.83	39.48	65.98
TiO ₂	0.57	0.12	0.13	0.14	0.00
Al ₂ O ₃	8.77	15.03	20.59	19.52	21.68
Fe ₂ O ₃	4.33	0.00	—	—	—
MgO	13.19	6.09	11.21	9.88	0.00
FeO	0.00	4.22	12.96	13.73	0.01
MnO*	0.02	0.04	0.20	0.28	0.00
CaO	20.08	11.02	7.68	8.82	1.74
BaO	0.03	0.00	—	—	—
Na ₂ O	2.44	7.51	3.98	3.61	9.93
K ₂ O	0.04	0.00	0.21	0.28	0.13
Total	99.51	98.93	94.79	95.74	99.47
<i>6 oxygens</i>			<i>23 oxygens</i>		<i>8 oxygens</i>
Si	1.82	1.96	5.62	5.82	2.90
Ti	0.02	0.00	0.02	0.02	0.00
Al	0.38	0.63	3.60	3.39	1.12
AlIV	—	—	—	—	—
AlVI	—	—	—	—	—
Fe ³⁺	0.12	0.00	0.80	0.85	—
Mg	0.72	0.32	2.48	2.17	0.00
Fe ²⁺	0.00	0.13	0.80	0.85	0.00
Mn*	0.00	0.00	0.03	0.04	0.00
Ca	0.78	0.42	1.22	1.39	0.08
Ba	0.00	0.00	—	—	—
Na	0.17	0.52	1.15	1.03	0.85
K	0.00	0.00	0.04	0.05	0.01
Total	4.00	3.98	15.76	15.59	4.96
%Jd		53.3			
NaA			0.68	0.48	
NaB			0.47	0.56	
Ca B			1.22	1.39	
(Ca+Na)B			1.69	1.95	
(K+Na)A			0.72	0.53	
Mg#			0.76	0.72	
Xan					0.09
Xab					0.90
Xor					0.01

*as Mn₂O₃ and Mn³⁺ for epidote. Mg# = Mg/(Mg + Fe²⁺); Na# = Na/(Na + K); X_{czo} = (Al - 2)/(Fe³⁺ + Al + Cr³⁺ - 2) and X_{ep} = Fe³⁺/(Fe³⁺ + Al + Cr³⁺ - 2) (Franz and Liebscher, 2004).

diopside, about 10 µm-width, and by partial transformation of diopside into omphacite (Fig. 3a).

Quartz lenses are usually observed in the pressure shadows of coronites in the Ky-rich sample (Figs. 2a and 3d). These lenses mainly contain quartz, large phengite flakes (over 500 µm) and rare small kyanite grains (50 µm). The eclogitisation process produced a large quantity of SiO₂ (reactions 1 and 5), which partially precipitated in situ (Jamtveit et al., 1990). The silica excess is carried out by fluids and precipitated in lower fluid pressure zones such as the coronites pressure shadows. In the

Holsnøy area, quartz veins are also observed at an outcrop-scale in fractures or at the edge of shear-zones (Austrheim and Griffin, 1985; Andersen et al., 1990; Jamtveit et al., 1990; Boundy et al., 1992; Kühn, 2002).

4.2.3. Progression of the amphibolitisation

In both eclogites, decompression resulted in the formation of symplectite and reaction rims around primary eclogite-facies phases (mainly garnet and omphacite; Fig. 3b,c,f; Fig. 4a,d,f). Omphacite broke down to symplectites made of amphibole and plagioclase. The breakdown

Table 3

Selected representatives EMP analyses of minerals from the Ph-rich eclogite (sample NOK10d)

NOK10d: Ph-rich eclogite	Garnet		Epidote		Phengite		
	Grt 2	Grt 1	Core	Rim	Foliation core	Foliation rim	Included in Grt 2
	Almandin	Almandin	Czo	Czo	Phengite	Phengite	Phengite
SiO ₂	38.67	39.35	38.63	38.63	47.30	46.27	46.31
TiO ₂	0.05	0.13	0.17	0.27	0.81	0.81	0.84
Al ₂ O ₃	21.96	21.79	28.62	25.87	31.60	31.20	30.64
Fe ₂ O ₃	0.00	1.18	7.20	10.37	–	–	–
MgO	5.26	9.73	0.15	0.06	1.98	1.83	1.85
FeO	24.76	22.42	–	–	2.34	3.06	3.69
MnO*	0.81	0.73	0.06	0.11	0.01	0.00	0.03
CaO	8.72	5.20	20.39	20.97	0.00	0.01	0.04
Na ₂ O	0.00	0.00	0.04	0.00	1.19	0.47	0.70
K ₂ O	0.00	0.00	0.00	0.07	8.61	9.74	8.26
H ₂ O	–	–	1.90	1.89	4.45	4.38	4.35
Total	100.22	100.53	97.16	98.24	98.29	97.77	96.68
<i>Cations for 12 oxygens</i>			<i>12.5 oxygens</i>		<i>11 oxygens</i>		
Si	3.00	2.98	3.05	3.06	3.19	3.16	3.19
Ti	0.00	0.01	0.01	0.02	0.04	0.04	0.04
Al	2.01	1.95	2.66	2.42	2.51	2.51	2.49
Fe ³⁺	0.00	0.07	0.43	0.62	–	–	–
Mg	0.61	1.10	0.02	0.01	0.20	0.19	0.19
Fe ²⁺	1.61	1.42	–	–	0.13	0.18	0.21
Mn*	0.05	0.05	0.00	0.01	0.00	0.00	0.00
Ca	0.72	0.42	1.72	1.78	0.00	0.00	0.00
Na	0.00	0.00	0.01	0.00	0.16	0.06	0.09
K	0.00	0.00	0.00	0.01	0.74	0.85	0.73
Total	8.00	8.00	7.90	7.91	6.97	6.99	6.94
Xalm	0.54	0.48					
Xprp	0.20	0.37					
Xgrs	0.24	0.14					
Xsps	0.02	0.02					
Xczo			0.61	0.40			
Xep			0.39	0.60			
Na+K					0.90	0.91	0.82
Mg#					0.60	0.52	0.47
Na#					0.17	0.07	0.11

progresses from grain joints towards the cores. In natural light, the alteration zones form asymmetrical dark eyes around garnet in the Ky-rich sample and omphacite augens in the Ph-rich sample. In the coronites, Ca-rich amphibole preferentially appears in fractures and at garnet grain boundaries. Where the amphibolitisation is more penetrative, the fractures in garnet are filled with secondary calcic-amphibole needles. The amphibolite-facies alteration of white mica resulted in its partial consumption.

5. Geochemical results

5.1. Major element mineral chemistry

Representative EMP analyses of main minerals from samples NOK10a and NOK10d are reported in

Tables 2 and 3, but an extensive database is available as Electronic Supplementary Material (Table I and II).

5.1.1. Garnet

Garnet cores are pyrope-rich, inherited from granulite (Grt1), and display contrasting compositions in the two samples. The garnet core composition (Alm 30–40 Prp 45–52 Grs 18–20; Table 2, Figs. 3e and 5a) of the Ky-rich sample is typical of granulite-facies garnets derived from an anorthositic protolith. In the Ph-rich sample, the garnet core composition (Alm 48–49 Prp 34–37 Grs 14–12; Table 3, Figs. 4e and 5b) is similar to granulite-facies garnets derived from a gabbroic protolith (Austrheim and Griffin, 1985; Cohen et al., 1988; Boundy et al., 1992; Erambert and Austrheim, 1993; Matthey et al., 1994). In the Ky-

Table 3 (continued)

NOK10d: Ph-rich eclogite	Clinopyroxène		Plagioclase	Amphibole	
	Included in Grt 2	Foliation	Symplectite	Alteration of Grt 2	Alteration of omphacite
	Omphacite	Omphacite	Albite	Taramite	Fe-Actinolite
SiO ₂	54.22	54.66	64.95	39.43	52.98
TiO ₂	0.20	0.19	0.02	0.44	0.01
Al ₂ O ₃	13.37	12.78	21.14	17.80	1.84
Fe ₂ O ₃	–	–	–	–	–
MgO	5.60	5.84	0.40	7.27	8.70
FeO	6.83	6.62	0.61	16.67	13.34
MnO*	0.06	0.04	0.03	0.11	0.18
CaO	8.72	8.92	2.86	7.00	13.60
Na ₂ O	6.74	6.74	9.26	3.55	3.61
K ₂ O	0.01	0.00	0.06	0.95	0.00
H ₂ O	–	–	–	1.92	1.97
Total	95.74	95.77	99.31	95.14	96.23
6 oxygens			8 oxygens		23 oxygens
Si	2.00	2.02	2.88	6.15	8.04
Ti	0.01	0.01	0.00	0.05	0.00
Al	0.58	0.56	1.10	3.27	0.33
Fe ³⁺	–	–	–	–	–
Mg	0.31	0.32	0.03	1.69	1.97
Fe ²⁺	0.21	0.20	0.02	2.18	1.69
Mn*	0.00	0.00	0.00	0.01	0.02
Ca	0.35	0.35	0.14	1.17	2.21
Na	0.48	0.48	0.80	1.07	1.06
K	0.00	0.00	0.00	0.19	0.00
Total	3.94	3.94	4.97	15.79	15.33
%Jd	54.49	53.67			
Xan			0.15		
Xab			0.85		
Xor			0.00		
Na A				0.60	1.06
Na B				0.47	0.00
Ca B				1.17	1.98
(Ca+Na)B				1.64	1.98
(Na+K) A				0.79	1.06
Mg#				0.44	0.54

*as Mn₂O₃ and Mn³⁺ for epidote. Mg# = Mg/(Mg + Fe²⁺); Na# = Na/(Na + K).

rich sample, granulite-facies garnet was partially resorbed during the eclogite-facies overprint. The chemical variations over 20 μ m within the rims (garnet 2 in Fig. 5a) attest to partial reequilibration by penetrative diffusion in the eclogitic facies (Alm 45–49 Prp 28–32 Grs 21–22; Table 2, Figs. 3e and 5a). In the Ph-rich sample, garnet shows an overgrowth of eclogite-facies garnet (Grt2) richer in Ca and Fe (Alm 53–55 Prp 19–21 Grs 22–26; Table 3, Figs. 4e and 5b). Between both domains, Grt1 rims display a continuous zonation characterised by a progressive increase of the Fe, Ca and Mn contents and a decrease of the Mg# toward the rim (Alm 55–58 Prp 23–26 Grs

16–20; Table 3). The transect through Grt1 depicts a diffusion profile possibly recorded during the cooling of granulite after garnet reached the peak temperature in the lower crust (Erambert and Austrheim, 1993), or related to reequilibration of garnet 1 rim with garnet 2 (Figs. 4e and 5b).

5.1.2. Epidote

Two groups of monoclinic epidote (referred as “epidote” elsewhere in the text) have been distinguished in the Ky-rich sample: epidote needles appearing in coronites as exsolutions from diopside, and epidote neoblasts that replace plagioclase in the main foliation.

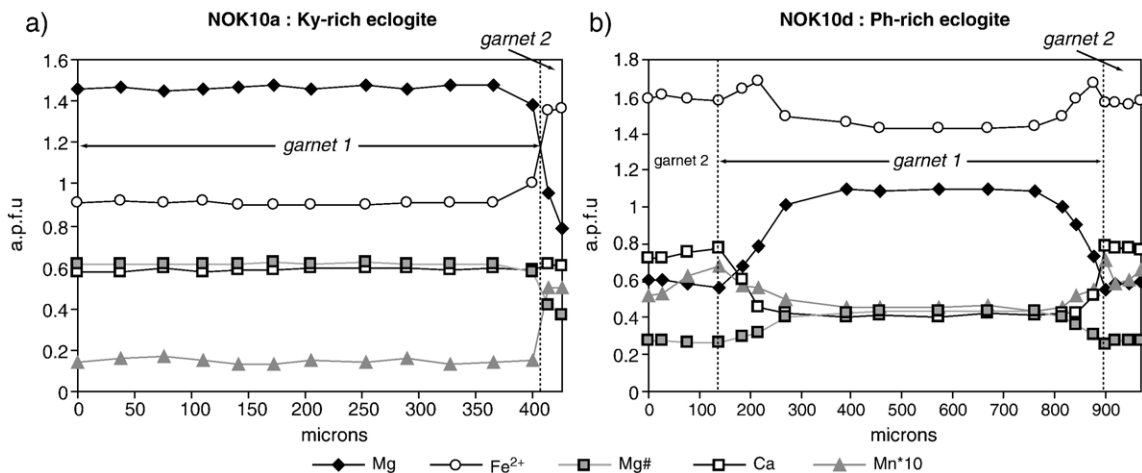


Fig. 5. Electronic microprobe profiles for representative garnet grains from samples NOK10a (a) and NOK10d (b). The orientation of the profiles is shown on Figs. 3e and 4e.

Epidote from coronites is richer in Fe³⁺ (X_{ep} 0.18–0.30; Table 2, Fig. 6a) than epidotes from the main foliation (X_{ep} 0.08–0.10; Table 2, Fig. 6a). In the Ph-rich sample, only one type of epidote has been observed and replaces plagioclase in the foliation. It is Fe-rich and Al- and Ca-poorer (X_{ep} 0.39–0.43; Table 3) than the epidote replacing plagioclase in the Ky-rich sample. In both samples, the optical zoning is not mirrored by significant compositional zoning except at the rim where the amphibolite-facies overprint is outlined by a slight increase of X_{ep} (up to 0.61 for the Ph-rich sample and 0.13 for the Ky-rich sample; Tables 2, 3). The acicular cores are easily recognizable on BSE mode (Fig. 3c) and correspond to epidote needles which appear in plagioclase during the first stage of the granulite conversion into eclogites (Austrheim and Griffin, 1985; Kühn, 2002).

5.1.3. Phengite

In the Ky-rich sample, three textural groups of phengite were identified: (1) phengite replacing (with epidote) plagioclase in the main foliation (Figs. 2a and 3c) (2) phengite replacing garnet (Figs. 2a and 3b,e) (3) phengite located in quartz lenses (Figs. 2a and 3d). A fourth group has been defined on the basis of chemical composition and corresponds to barium-rich phengite located in quartz lenses ($0.76 < \text{BaO}\% < 1.48$; Table 2). In the Ph-rich sample, only two types of phengite have been discriminated: (1) phengite replacing (with epidote) plagioclase in the main foliation (Figs. 2b and 4a,d) and (2) phengite included in garnet (Fig. 4f).

Phengites of both samples are characterised by a low tschermak substitution (Tables 2, 3). In the Ky-rich

sample, Si=3.09–3.15 a.p.u.f. and Fe+Mg=0.21–0.30 a.p.u.f. In the Ph-rich sample, Si=3.16–3.21 a.p.u.f. and Fe+Mg=0.3–0.4 a.p.f.u. The phengites are not zoned and only minor K increase and Na decrease are recorded at the extreme rims, on the last 10 μm (Tables 2, 3). In the Ky-rich sample, Na decreases from 0.30 (core) to 0.10 (rim) a.p.u.f. and in Ph-rich sample, Na decreases from 0.18 (core) to 0.07 a.p.f.u. in a single grain. This feature is attributed to low-temperature alteration in the amphibolite facies (Guidotti and Sassi, 2002). In the Ph-rich sample, phengite included in garnet is Fe richer (~ 0.20 vs. ~ 0.15 a.p.f.u.; Table 3; Fig. 6b).

Phengitic substitution influences the Na# since substitution of Fe²⁺ and Mg²⁺ by Al³⁺ in the octahedral site favours the incorporation of the largest cations such as K⁺ to Na⁺ (e.g. Bailey, 1984; Guidotti, 1984; Guidotti et al., 1992; Domanik and Holloway, 1996; Guidotti and Sassi, 1998a,b; Brigatti and Guggenheim, 2002). Following this process, all phengites analysed in the Ky-rich sample describe a well-defined trend in a Na# vs. (Si+Mg+Fe²⁺–3) diagram (Fig. 6c). Only the rim analyses, related to alteration in the amphibolite facies, do not follow this trend (not represented in Fig. 6c). After crystallisation in the eclogitic facies, phengite and paragonite substitutions are decoupled during cooling in the amphibolite facies. All core analyses define a narrow range of values, which testifies that all phengites crystallised under the same eclogite-facies conditions. Slight compositional differences can be seen between the different groups of phengite when considered as a function of their textural sites as defined above (Fig. 6c). The four defined fields overlap each other, but phengites tend to be less paragonitic and more phengitic in the

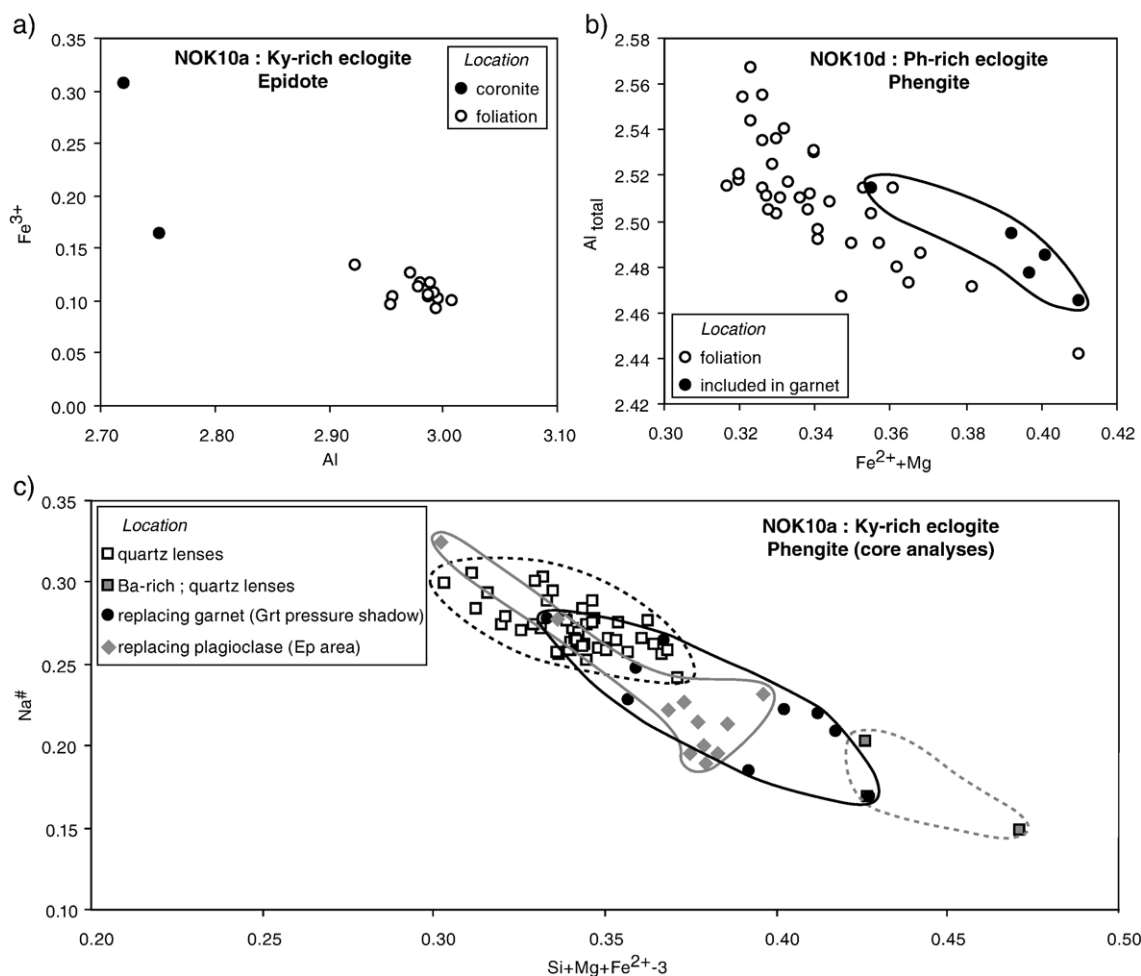


Fig. 6. (a) Fe^{3+} vs. Al diagram for epidotes from sample NOK10a showing the different chemical composition of eclogite-facies epidotes: epidotes replacing diopside (in coronite) and epidotes replacing plagioclase (in foliation). (b) Al_{total} vs. $Fe^{2+}+Mg$ diagram for phengites from sample NOK10d. Note, for the eclogite-facies phengites, the difference in composition between those located in the main foliation and those included in garnet. (c) Na# vs. $(Si+Mg+Fe^{2+}-3)$ diagram for core analyses of eclogite-facies phengites located in different textural sites (sample NOK10a). $Na\# = Na/(Na+K)$.

following order: phengite from quartz lenses, phengite replacing plagioclase, phengite replacing garnet, and Ba enriched phengite.

5.1.4. Other minerals

Eclogite-facies clinopyroxenes is omphacite according to the Morimoto et al. (1988) classification (Jd₅₃ NOK10a, Jd_{53–57} NOK10d; Tables 2, 3). For the Ky-rich sample, omphacite is only occurs in the foliation whereas it appears both as inclusions in garnet and in the matrix of the Ph-rich sample. However, there is no chemical distinction between these different types of omphacite (Table 3). Plagioclase from the amphibolite stage (breakdown of omphacite in symplectite) is albitic in both samples (Tables 2, 3). Amphibole resulting from the breakdown of garnet and omphacite is calcic (Tables 2, 3).

5.2. Trace element analyses

5.2.1. Whole rock

REE patterns of both samples display a progressive decrease from La to Lu and a positive Eu anomaly (Table 1; Fig. 7a). The LREE enrichment, as well as the Eu anomaly is more pronounced in the Ky-rich sample than in the Ph-rich sample. This discrepancy is related to the original content of plagioclase in the two distinct protoliths. Both samples have similar trace element patterns (Fig. 7b) with high Sr and Ba contents, and low Rb concentration (Table 1).

5.2.2. Garnet

Four garnet fractions from the Ky-rich sample and three garnet fractions from the Ph-rich sample were

analysed. In-situ analyses of pure-garnet areas (mostly Grt1) along with garnet alteration zones (breakdown in calcic-amphibole) were also carried out. Only one analysis of eclogite-facies garnet (Grt2) in sample NOK10d has been achieved (Fig. 8a–b; all trace elements data are reported in an Electronic Supplementary Material, Table III for the Ky-rich sample and Table IV for the Ph-rich sample).

REE patterns of the different garnet fractions from the two samples are very similar. A good agreement between Grt1 in situ analyses and mineral fractions is observed. Therefore the thin eclogite-facies rim does not significantly affect the global REE budget of the garnet. REE patterns depict a strong depletion in LREE and a positive Eu anomaly for garnets from both samples. Compared to garnet from basic rocks, garnet from the Ky-rich sample exhibits a slight decrease from the MREE to the HREE (e.g. Griffin and Brueckner, 1985; Messiga et al., 1995; Amato et al., 1999; Sassi et al., 2000; Luais et al., 2001; Xiao et al., 2001; Spandler et al., 2003). This feature results from the relatively high depletion in HREE of the protolith ($Lu_N=0.5$ in NOK10a). Grt2 from the Ph-rich sample differs from Grt1 by a stronger depletion in LREE. Moreover, in the Ph-rich sample, differences between pure garnet and alteration zones in garnet are clearly observed (Fig. 8a–b). Alteration zones display a strong enrichment in LREE compared to pure garnet, a positive Ce anomaly and a similar HREE pattern. Moreover, in both samples, alteration zones bear excesses in Rb, Ba, Pb and Sr compared to unaltered garnet analysed in situ (Fig. 9a–b). Alteration products located in fractures significantly contribute to the global budget of LILEs in the garnet fraction analyses. Indeed, all the LILE values measured for garnet fractions are intermediate between those of

unaltered garnets and of alteration zones analysed in situ (Fig. 9a–b).

5.2.3. Epidote

Epidotes from the eclogite stage have been found in two distinct textural sites in the Ky-rich sample: in the main foliation and in the coronites. An epidote fraction from the foliation has been separated from the Ky-rich sample for conventional ICP-MS analysis, and epidotes from the foliation and from the coronites were analysed in situ. However, because of the laser spot size used (51 μm), in situ analyses of epidotes from coronites ($\sim 10 \mu\text{m}$ needles) were partially contaminated by surrounding diopside. Nevertheless, diopside is very poor in REE (below detection limits when analysed in situ), and thus the diopside contamination lowers the absolute value concentration in REE, but does not affect the general shape of the REE pattern. Epidote occurs only in the main foliation in the Ph-rich sample and was analysed in situ.

In the Ky-rich sample, mineral fraction and in situ analyses display a good correlation (Fig. 8c). Epidote replacing plagioclase in both samples (epidote from the foliation) displays REE pattern with a progressive decrease from LREE to HREE and a positive Eu anomaly in agreement with characteristics commonly observed for eclogite-facies epidotes (Sassi et al., 2000; Spandler et al., 2003; Frei et al., 2004). Epidote from coronites is slightly different and shows a more pronounced depletion in HREE, a subhorizontal La–Nd portion and a less pronounced Eu anomaly. In the Ph-rich sample, two groups of epidote have been distinguished on the basis of their La/Lu ratio (Group 1: $La/Lu > 100$, Group 2: $La/Lu < 100$; see Fig. 8d). Epidotes replacing plagioclases are the Sr richest minerals among those of the

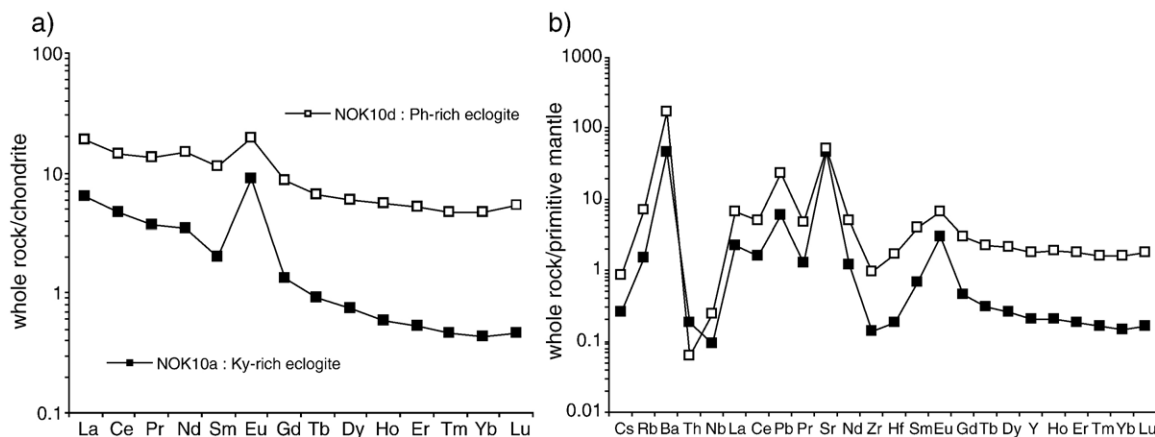


Fig. 7. (a) Rare earth element patterns normalised to chondrite for samples NOK10a and NOK10d. (b) Primitive-Mantle normalised trace element patterns of NOK10a and NOK10d. Normalisations after Sun and McDonough (1989).

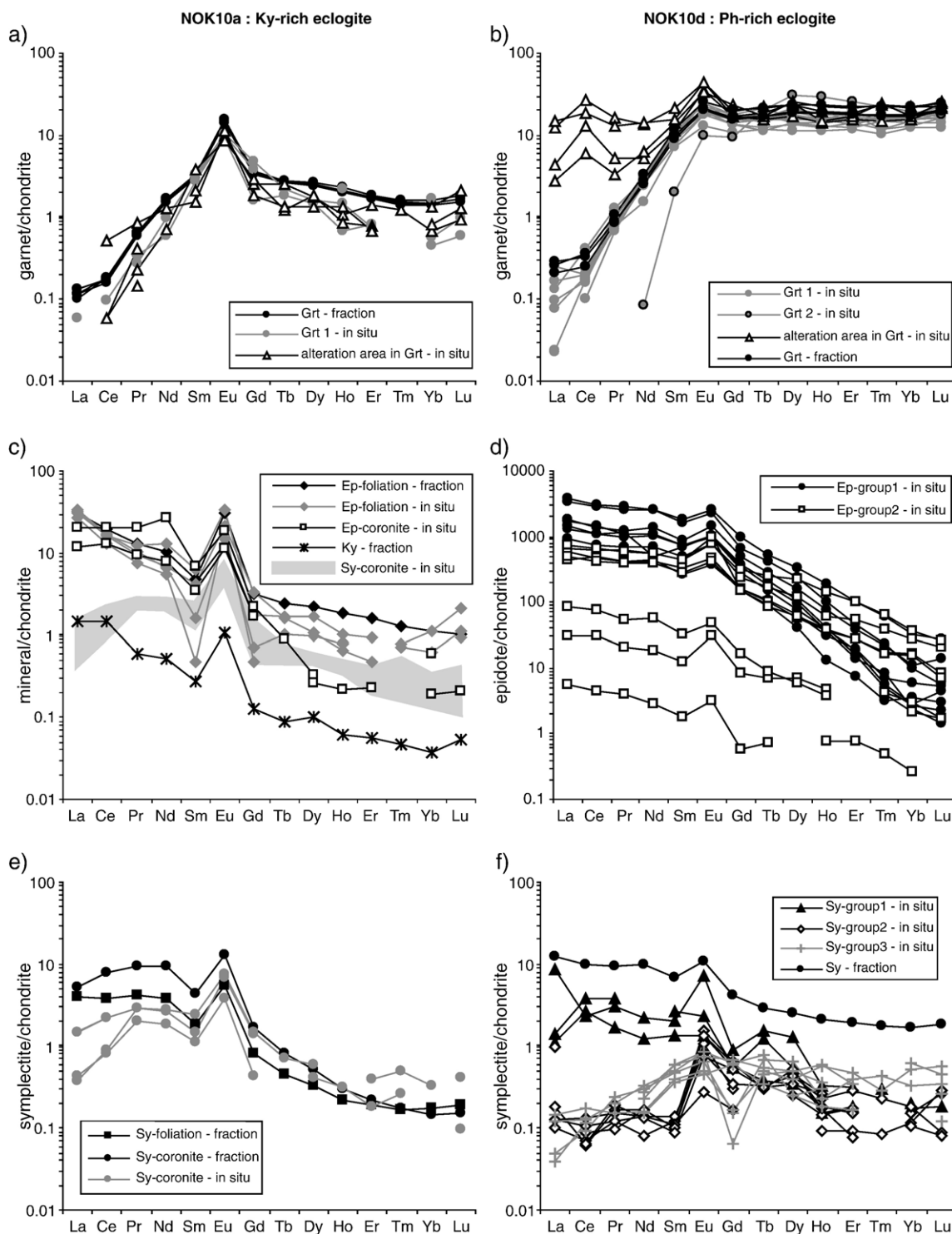


Fig. 8. REE patterns normalised to chondrite of various minerals from samples NOK10a and NOK10d analysed by conventional ICP-MS method (labelled as “fraction” in the caption) or by LA-ICP-MS (labelled as “in situ” in the caption). Normalisations after [Sun and McDonough \(1989\)](#). (a) garnet (Grt) patterns from NOK10a; (b) garnet (Grt) patterns from NOK10d; (c) epidote (Ep), symplectite (Sy) and kyanite (Ky) patterns from NOK10a; (d) epidote (Ep) pattern from NOK10d; (e) symplectite (Sy) patterns from NOK10a; (f) symplectite (Sy) patterns from NOK10d.

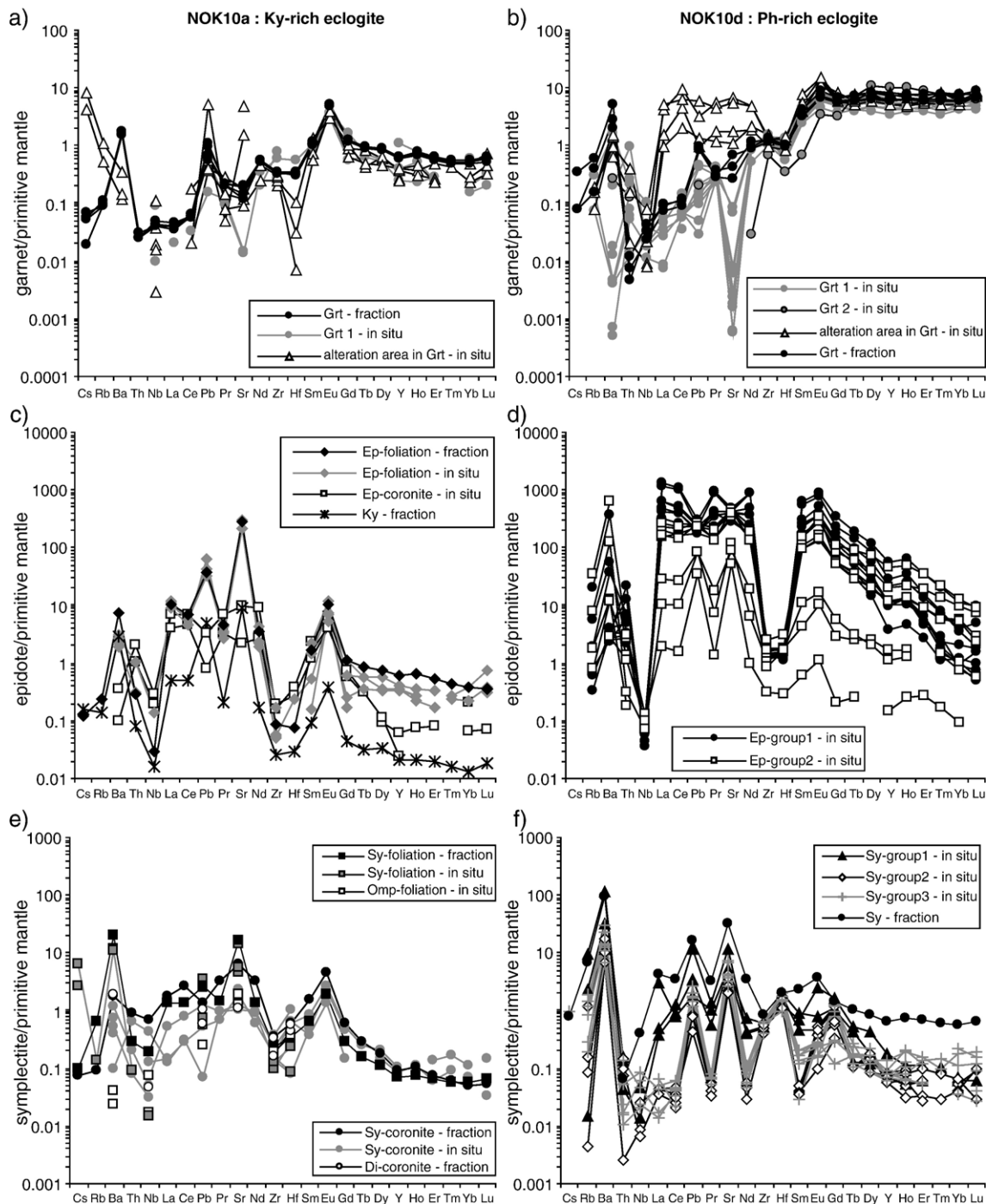


Fig. 9. Spider diagrams normalised to primitive mantle of various minerals from samples NOK10a and NOK10d analysed by conventional ICP-MS method (labelled as “fraction” in the caption) or by LA-ICP-MS (labelled as “in situ” in the caption). Normalisations after Sun and McDonough (1989). (a) garnet (Grt) patterns from NOK10a; (b) garnet (Grt) patterns from NOK10d; (c) epidote (Ep) and kyanite (Ky) patterns from NOK10a; (d) epidote (Ep) pattern from NOK10d; (e) symplectite (Sy) patterns from NOK10a; (f) symplectite (Sy) patterns from NOK10d.

eclogite-facies assemblages with concentrations of over 4500 ppm and up to 10,000 ppm in the Ph-rich sample (Fig. 9c–d).

5.2.4. Pyroxene and symplectite

The analysed pyroxene fractions are not pure since it was not possible to extract enough unaltered

omphacite and diopside. Therefore the fractions are referred as symplectite from the foliation and symplectite from the coronites (NOK10a). Moreover, pure omphacite and diopside were analysed in situ (but REE were below detection limit), as well as the symplectite areas located in the main foliation (breakdown of omphacite in the amphibolite facies) and in coronites (breakdown of diopside in the eclogitic facies).

All symplectites are characterised by LREE enrichment and a positive Eu anomaly (Fig. 8e). Symplectite from the coronites has a bell-shaped La–Sm pattern that contrasts with the sub-horizontal pattern of symplectite from the foliation. Bulk analysis of symplectite from the coronites displays higher LREE contents than those determined in situ. This can be explained by a higher concentration in epidote needles in the fraction than in the areas analysed in situ. Symplectite from the foliation is richer in LILE than pure omphacite analysed in situ, whereas symplectite from coronites is richer in Sr than pure diopside (Fig. 9e). This likely reflects the crystallisation of epidote that concentrates Sr compared to the initial diopside (Frei et al., 2004). In the Ph-rich sample, most of the symplectite areas analysed in situ give “zig zag” patterns (Figs. 8f and 9f), reflecting the compositional variation of the area sampled during ablation. However, based of their shape, three kinds of pattern can be distinguished. Symplectites forming the

Group 1 yield REE patterns decreasing from LREE to HREE (close to that of the symplectite fraction). Group 2 consists of analyses with irregular patterns displaying a plateau for the LREE and HREE and an MREE-enrichment. Lastly, group 3 consists of analyses yielding depletion in LREE and a mostly flat pattern from MREE to HREE. The diversity of the REE patterns, as well as their irregularity, is attributed to sampling by the laser beam of several phases (namely Pl, Omp and Amp) in varying proportions due to different degree of alteration. Bell shaped patterns (group 3) are considered as closely reflecting the pattern of the omphacite (e.g. Messiga et al., 1995; Xiao et al., 2001; Hermann, 2002), while the others (groups 1 and 2) are intermediate between pure omphacite and the analysed symplectite fraction.

5.2.5. Kyanite

Bulk analyses of kyanite (Figs. 8c and 9c) have similar REE patterns than epidote from the foliation, but lower REE and LILE contents.

5.2.6. Phengite

Nine phengite fractions were separated as a function of grain size, density and magnetic properties in the Ky-rich sample (Table 4). Disregarding grain size, these parameters reflect intrinsic features of the phengites, especially variations in chemical composition. It is thus expected that in each fraction, phengites

Table 4

Detail of grain size, density and magnetism used to separate the phengite fractions in the studied samples

Phengite fraction	Grain size (µm)	Density	Frantz isodynamic parameters		
			Counter slope	Intensity (A)	Attraction
<i>Sample NOK10a: Ky-rich eclogite</i>					
1-a	100-125	<2.9	10°	Max	attracted
1-b	100-125	<2.9	5°	Max	attracted
1+b	100-125	2.9<-<3.3	15°	1.3	unattracted
2-a	125-250	<2.9	5°	Max	attracted
2-b	125-250	<2.9	5°	Max	unattracted
2+a	125-250	2.9<-<3.3	10°	Max	unattracted
2+b	125-250	2.9<-<3.3	10°	Max	attracted
4-a	250-400	<2.9	12.5°	Max	attracted
4-b	250-400	<2.9	2.5°	Max	unattracted
<i>Sample NOK10d : Ph-rich eclogite</i>					
1-	100-125	<2.9	-	-	-
2+a/2+are	125-250	2.9<-<3.3	15°	1.5	attracted
2+b/2+bre	125-250	2.9<-<3.3	15°	1.5	unattracted
2+c	125-250	2.9<-<3.3	-	-	-
4+a	250-400	2.9<-<3.3	-	-	-
4+b	250-400	2.9<-<3.3	15°	1.25	unattracted
4+c/4+cre	250-400	2.9<-<3.3	15°	1.15	attracted
4+d	250-400	2.9<-<3.3	15°	1.15	unattracted

from a specific textural sites (phengite replacing plagioclase in the main foliation or phengite replacing garnet or phengite precipitated in quartz lenses) is preferentially concentrated since the EMPA data reveals variation in composition of each phengite group (Fig. 6c). In the Ph-rich sample, eight fractions of phengite were analysed and three of them were duplicated (Table 4). In-situ analyses were performed in both samples, but REE were below the detection limit. Because of the very high Ba content of the phengites analysed (>5000 ppm) and their low REE concentrations, Ba-oxide-hydroxide interferences appear on MREE (Nd, Sm, Eu, Gd). These interferences cannot be corrected for and have repercussions on the HREE (particularly for Yb). Therefore these elements are not considered further in the discussion.

In the Ky-rich sample, three phengite groups are defined according to their REE patterns (Fig. 10a). Phengite from group 1 (fractions 1–b and 1+b) differs from the others by its higher HREE contents. Phengite from group 2 (fractions 2+a, 2+b and 4–b) is characterised by a slight positive Ce anomaly ($Ce_N/Ce^* = 1.2$ to 1.8). The four remaining fractions (1–a, 2–a, 2–b and 4–a) are then referred as group 3. The phengites from the Ph-rich sample have almost identical REE patterns with a decrease from La to Ho followed by a slight increase up to Lu (Fig. 10b).

Regarding the LILE, there is no clear distinction between the three groups of phengite described in the Ky-rich sample (Fig. 10c). There is good agreement between in situ and bulk analyses for Rb and Sr (Fig. 10c–d) and phengite appears to be the most

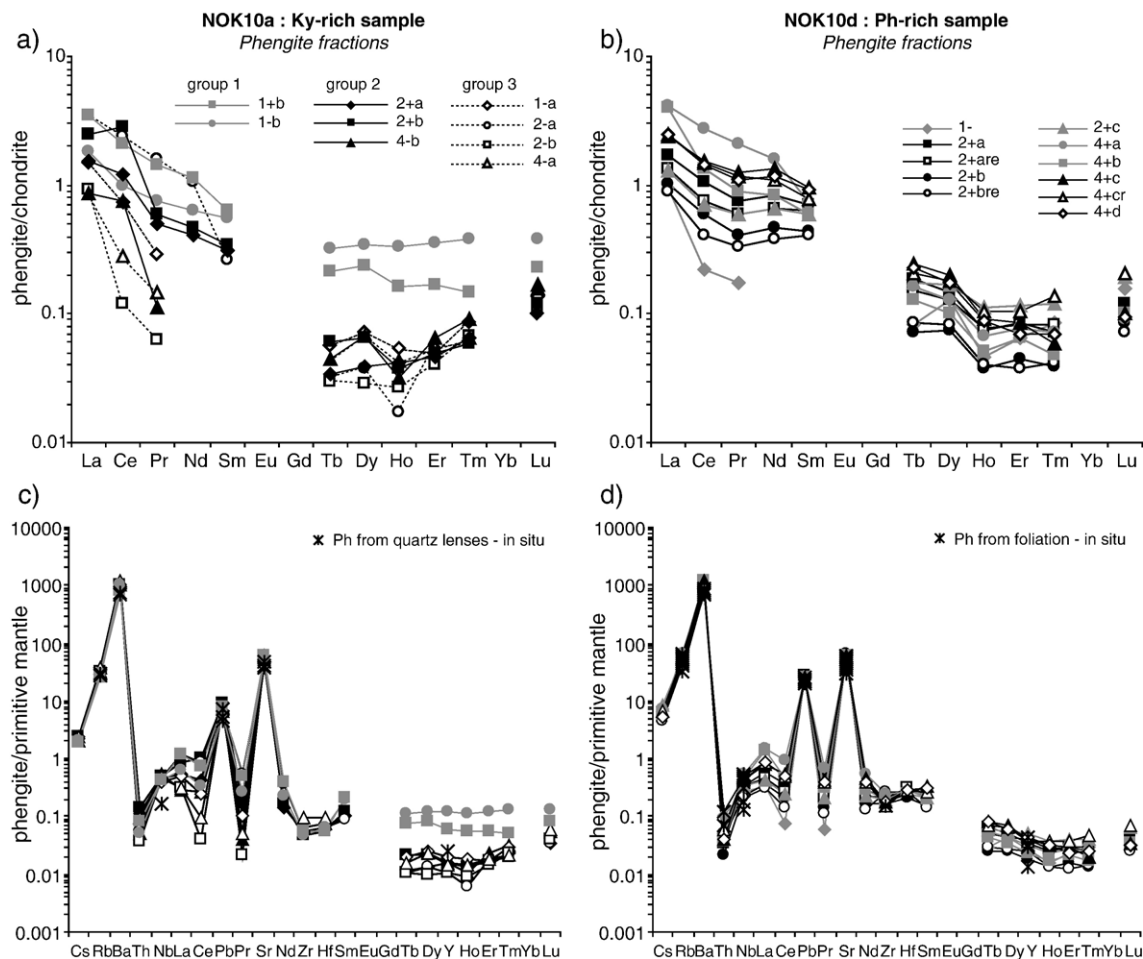


Fig. 10. REE patterns normalised to chondrite (a–b) and spider diagrams (c–d) normalised to primitive mantle for phengite fractions from sample NOK10a (a, c) and NOK10d (b, d) analysed by conventional ICP-MS method and by laser ablation (LA-ICP-MS). Normalisations after Sun and McDonough (1989).

fractionating phase for Rb in the eclogite-facies assemblage (over 17 ppm).

6. Discussion

6.1. Influence of fluid circulation on bulk chemistry

The bulk composition of the NOK10a and NOK10d samples has been compared to other eclogites and granulites from the Lindås Nappe (Rockow et al., 1997) in order to tackle the consequences of fluid circulation on the whole rock bulk chemistry during the Caledonian event. Based on statistical comparison of granulite/eclogite pairs derived from various protolith, Rockow et al. (1997) demonstrated that, except for an H₂O supply, no significant loss or gain of elements is associated to fluid infiltration. Granulite and eclogite pairs from Rockow et al. (1997), classified as a function of their protolith nature, are reported together with the two studied samples (Fig. 11). In the Sm vs. Nd diagram (Fig. 11a), all analyses plot on the same correlation line defining the original magmatic series. This substantiates that no fractionation between these elements occurred as a consequence of syn-eclogite-facies fluid infiltration. In a Ba vs. Sr diagram (Fig. 11b) two correlation lines are observed, one defined by the jotunites ($r^2=0.88$) and the second by the anorthosites and gabbros ($r^2=0.89$). Our Ky-rich sample plots close to the anorthosite+gabbro line, whereas our Ph-rich sample is located on the jotunite trend and, with some gabbros, lies on a vertical line pointing to a Ba enrichment for an almost constant Sr content. In a Rb vs. Cs (Fig. 11c) diagram, gabbro and anorthosite data from Rockow et al. (1997) together with our data plot in a narrow field of low Rb and Cs values, well distinct from the jotunite field characterised by higher Rb contents. Only one eclogite-facies gabbro studied by Rockow et al. (1997) displays Cs in excess. The Ba enrichment observed for Ph-rich sample NOK10d may derive from fluid circulation and leaching of the surrounding granulites. One likely candidate could be, for instance, the jotunites, which are richer in Ba than the anorthosites and gabbros (Fig. 11b). However, as mentioned above, the jotunites are also characterised by high Rb concentration and thus, as observed for Ba, a Rb enrichment would have also been expected, which is not the case. Another potential

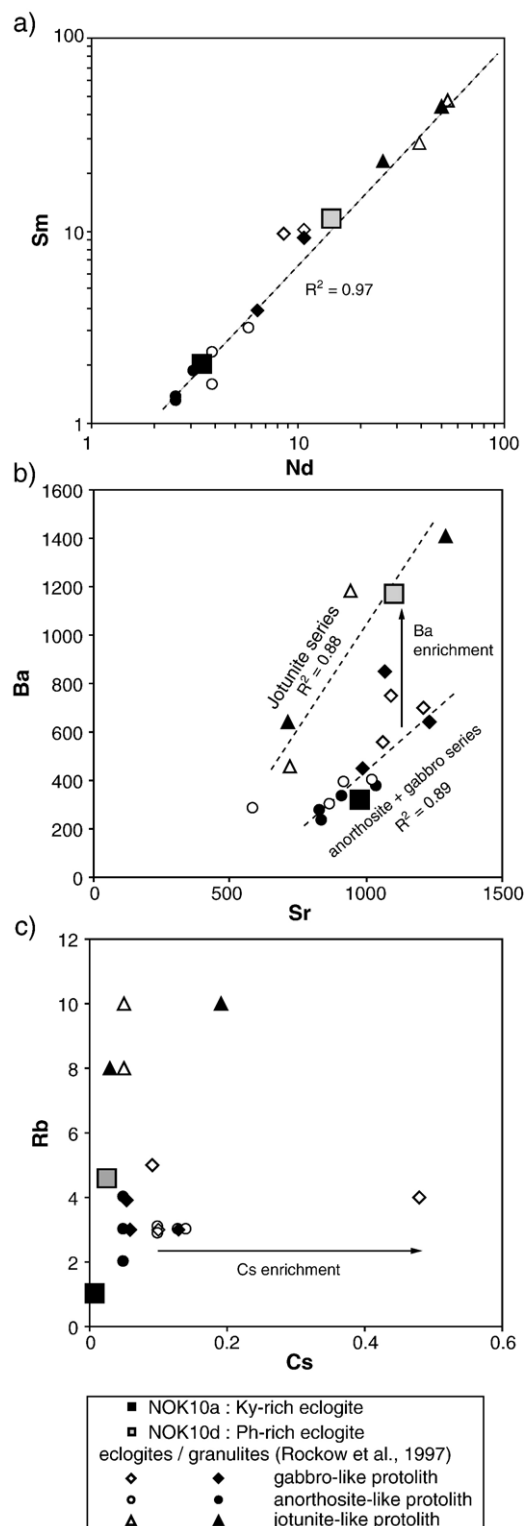


Fig. 11. (a) Nd–Sm, (b) Ba–Sr, (c) Rb–Cs correlation diagrams for samples NOK10a and NOK10d. Granulites and eclogites from the Lindås Nappe (data from Rockow et al., 1997) are also reported for comparison. The concentrations are in ppm.

Ba source, already proposed by Boundy et al. (2002), could be the breakdown, during the first stage of the eclogitisation, of scapolite, which locally occurs in granulites. However, scapolite is not only a Ba-rich mineral but is also very rich in Sr (Teertstra and Sherriff, 1997). The metasomatic marbles from the Lindås Nappe (which are a product of the breakdown of scapolite) are indeed considerably enriched in Sr (concentrations up to 11,000 ppm, Boundy et al., 2002), whereas no significant Sr enrichment has been observed in our sample. Therefore the apparent enrichment in Ba observed for sample NOK10d is most likely attributed to the heterogeneous composition of the protolith gabbros (Rockow et al., 1997). In conclusion, the studied samples did not suffer from any significant gain or loss of elements as a result of fluid circulation during Caledonian times.

6.2. Relationships between chemical composition of a mineral and crystallisation process

In both studied samples, the preservation of textures and minerals inherited from the granulite protoliths (coronites, main foliation and garnet) can be recognized. Moreover the chemical analyses of a same mineral phase but located in different textural sites yield significant differences in major and trace element composition. This feature can be related to the nature of the precursor minerals and is discussed in details for epidote and phengite in the Ky-rich sample.

6.2.1. Composition of epidote

Chemical analyses reveal that epidote from the main foliation is richer in Al than epidote located in coronites (Fig. 6a) and is characterised by Pb and Sr positive anomalies ($Pb_N/Pb^* = 6.4\text{--}17.5$; Fig. 9c). On the contrary epidote from the coronites is characterised by a Pb negative anomalies ($Pb_N/Pb^* = 0.2\text{--}0.5$; Fig. 9c) and a stronger depletion in HREE compared to LREE (Fig. 8c). Epidote from the main foliation crystallised from the breakdown of plagioclase after reaction 1 and displays similar REE patterns than plagioclase from basic granulites (Pride and Muecke, 1981). This is in agreement with epidote/plagioclase partition coefficients being almost equal for every REE (Frei et al., 2004). Therefore, epidote crystallised from plagioclase exhibits similar REE patterns, with, nevertheless, higher REE content than plagioclase. Epidote from the coronites formed at the expense of granulite-facies diopside. Epidote/clinopyroxene partition coefficients for REE predict that epidote fractionates much more LREE than clinopyroxene but have similar HREE

partitioning coefficients (Sassi et al., 2000; Spandler et al., 2003; Frei et al., 2004). Therefore, epidote and omphacite crystallised from diopside yield similar HREE patterns, but epidote is significantly LREE-richer than omphacite. This is substantiated by the comparison between the REE patterns of epidote from the coronites on one hand and the REE patterns of symplectites from the coronites (area of diopside breakdown in omphacite) analysed in situ on the other hand (Fig. 8c). Eventually, the chemical features of epidote from the foliation are typical of plagioclase (Al rich, enrichment in LREE compare to HREE, positive Eu, Pb and Sr anomalies), whereas those of epidote from coronites are typical from diopside previously crystallised in equilibrium with garnet (Fe rich, HREE poor, negative Pb anomaly). Thus, it appears that epidote from the eclogite stage inherited the chemical characteristics of its precursor mineral. This means for instance, that epidote from the coronites did not benefit from the contribution of elements coming from the breakdown of the plagioclase in the main foliation despite the presence of a fluid phase during the eclogitisation. Epidote crystallised from elements available on site.

6.2.2. Composition of phengite

In the Ky-rich sample, the EMP analyses of phengite emphasize slight compositional variations (variation of the phengitic substitution and of the Na#, Fig. 6c) between the different textural groups. Each type of eclogite-facies phengite crystallised following three different processes. The first involves the breakdown of plagioclase (reaction (1)). This reaction is water consuming and came off synchronous crystallisation of epidote, kyanite, phengite and quartz. The second is also water consuming (reaction (5)), which involves the breakdown of granulite-facies garnet and plagioclase to crystallise omphacite, kyanite and phengite. In reactions (1) and (5), the K content necessary for phengite crystallisation is supplied by antiperthitic K-feldspar located in the crystalline network of the plagioclase. The third process involves the leaching, by the eclogitising fluid, of elements in excess released during metamorphic reactions affecting various textural sites. These elements (mostly Si and alkalis) precipitated in the pressure shadows of coronites where the fluid pressure was the lowest to form phengite-bearing quartz lenses. The breakdown of isolated grains of diopside in the plagioclase-rich layers has to be considered in complement of reaction (1) to explain the Fe–Mg component of phengite. Thus, depending on their proximity to diopside grains during its crystallisation, phengite should be more or less Fe–Mg rich. Moreover, in

reaction (1), the Na liberated from plagioclase breakdown will be partially incorporated in phengite but the excess will be either incorporated in omphacite (produced from the breakdown of the isolated grains of diopside) or relocalised by the fluid (in quartz lenses for instance). Therefore, the occurrence or absence of diopside/omphacite grains in the crystallisation site of phengite from the epidote area will influence its Na, Fe and Mg composition. This process explains the large spread of compositions observed for phengite from the epidote area (Fig. 6c). Phengite crystallised after garnet breakdown (reaction (5)), is characterised by the highest phengitic substitution (Fig. 6c). In the course of this reaction, a greater quantity of Fe–Mg is available (coming from garnet breakdown) compared to reaction (1). Most of these components, and also Na released from plagioclase breakdown, are incorporated in omphacite, but a part is incorporated in phengite. Thus it is consistent to find lower Na# and higher phengitic substitution for phengite which crystallised after this reaction. The crystallisation of phengite in the quartz lenses implies the leaching and transport of elements by the fluids. The elements available are partly elements in excess released from eclogite-facies metamorphic reactions occurring in various textural sites, but are mainly elements released from reaction (1) (Si and Na, Jamtveit et al., 1990). This can explain the lower phengitic substitution of these phengites and their higher Na content.

Following these observations, it can be inferred that phengite derived from garnet (reaction (5)) will not have the same REE signature than phengite which crystallised from plagioclase (reaction (1)). Unfortunately, no in situ data for REE in phengite have been obtained because of its extremely low REE contents (La < 1 ppm). We thus rely on phengite fractions. It is not possible to straightforwardly link the different phengite fractions analysed to the textural groups recognised in-situ because the only data (concentrations in Rb, Sr, Ba and Pb) obtained, both for phengite analysed in-situ and phengite fractions, unable us to discriminate the different groups. Therefore we can only attempt to interpret the chemical features of the different phengite fractions. The nine phengite fractions from sample NOK10a display very similar Rb, Ba, Pb and Sr concentration. The main source for these elements in reactions (1) and (5) is the plagioclase (the participation of garnet is negligible as revealed by in situ analyses). The leaching of plagioclase by the fluids during its breakdown might also be the main source of these elements for phengite in the quartz lenses. Therefore, it is expected that these elements will

not be able to discriminate the different phengite groups. Considering the REE, plagioclase is the main source for these elements, but garnet in reaction (5) could have influenced the HREE composition of the secondary eclogite-facies minerals. Indeed, among the three groups of phengite fractions, which have been subdivided according to their REE characteristics, fractions defining group 1 (fractions 1+b and 1–b) are clearly enriched in HREE. Thus, these fractions most likely correspond to fractions dominated by phengite crystals formed at the expense of garnet according to reaction (5). A second group corresponds to phengite fractions (fractions 2+ a, 2+b and 4–b) characterised by a positive Ce anomaly. This anomaly indicates that Ce mainly occurs as Ce⁴⁺ rather than Ce³⁺ (e.g. Bau, 1991). The oxidation of Ce⁴⁺ in Ce³⁺ requires high *f*O₂ conditions, and Boundy et al. (2002) have demonstrated that the fluid phase associated to the partial eclogitisation of granulites in the Lindås Nappe was very oxidizing. Therefore, it seems likely that fractions with positive Ce anomalies correspond to fractions dominated by phengite grains from the quartz lenses, which precipitated from the fluid associated to the eclogitisation. The remaining fractions (fractions 1–a, 2–a, 2–b, 4–a), which have neither Ce anomaly nor high HREE content, could potentially be representatives of phengite which crystallised with epidote after plagioclase breakdown.

In the Ph-rich sample, most phengite crystals crystallised in the foliation from a single mineralogical reaction, which involved the breakdown of plagioclase (reaction (2)). Therefore it is not surprising that all fractions have similar REE patterns. Only very fine-grained phengite in garnet displays slightly different compositions. This phengite, richer in Fe (Fig. 6b), was formed partially at the expense of garnet. However, its contribution in the budget of the phengite fractions should be negligible.

6.2.3. The significance of Ba-rich phengite

Barium-rich phengite with contrasted composition compared to the other phengite groups (lower Na# and higher phengitic substitution, Fig. 6c) has been recognised in the Ky-rich sample. It is worth pointing out that this phengite occurs as very fine grains (few microns) plated to large, Ba-poor, phengite flakes. Both types of phengite are enclosed in quartz aggregates and this precludes that compositional variations are linked to retrogression in the amphibolite facies. A possible explanation is that their chemical characteristics are controlled by the fluid, which composition evolved with time. Guidotti and Sassi (2002) discussed the potential

relationship between the activity of H₂O in the fluid phase and the Na# of dioctahedral micas. The latter can decrease as water activity decreases and thus, phengite precipitated from the originally H₂O-rich fluid tends to be Na# higher than phengite crystallised later from an evolved H₂O-poor and Ba-rich fluid. Kullerud (1995) described a similar phenomenon for biotites that crystallised from a fluid evolving in composition as the crystallisation process advances (fluids tend to become more concentrated in Cl and Ba). A way to test this hypothesis in our model would be to analyse Cl in phengites since it has been demonstrated that fluids became concentrated in Cl in the course of eclogitisation reactions (Svensen et al., 2001; Kühn, 2002). An overview on the Ba content of phengite may be found in Harlow (1995) and Guidotti and Sassi (1998b).

6.3. Implications for redistribution of elements during eclogitisation

6.3.1. Scale of equilibrium achievement

The major and trace element analyses of eclogites NOK10a and NOK10d reveal that a single mineral species from the eclogite-facies has different chemical features in different sites at thin-section scale (in situ investigation) and at hand-sample scale (mineral fractions study). A single mineral phase, which crystallised after various processes in the course of the high pressure metamorphic event, displays different chemical features (i.e. phengites from NOK10a), while a population which crystallised after a single reaction is much more homogeneous (i.e. phengites from NOK10d). The chemical composition of the secondary eclogite-facies minerals depends on the nature of the precursor minerals involved in their crystallisation. Although they inherit some of the features of the precursor minerals (for instance: Al content in epidote, Na and Fe–Mg in phengite, and possibly HREE content in phengite), their composition also depends on the nature of coexisting eclogite-facies minerals in their crystallisation site (Na buffering by omphacite in phengite). This suggests a process of in situ recrystallisation where the composition of secondary eclogite-facies is controlled locally by the composition of microdomains, a few hundred microns in size (Messiga et al., 1995; Guidotti and Sassi, 1998a; Sassi et al., 2000; Zack et al., 2002). Chemical equilibrium is achieved at mineral scale (in microdomains), but not at sample scale. All the observations reported in the present study (inheritance of granulite-facies structures, variation of eclogite-facies mineral composition as a function of mineralogical reactions, occurrence of

equilibrium microdomains) show that fluids and deformation associated to the eclogite-facies metamorphism did not allow element rehomogenisation at sample scale. In other words, the rocks behaved as closed systems at the sample scale, and the fluid composition was internally controlled by the rocks.

6.3.2. Parameters controlling microdomain segregation

Field observations (Austrheim, 1990; Boundy et al., 1992; Austrheim et al., 1997) and modelling (Bjørnerud et al., 2002) suggest that conversion of anhydrous granulite into eclogite was mainly a catastrophic event in terms of rate process. Indeed, granulite brought into eclogite-facies conditions during the early stages of collision, remained metastable until seismic faulting allowed fluid percolation which triggered metamorphic reactions. The large overstepping of the equilibrium conditions before the introduction of the fluid into the system favoured very fast nucleation of eclogite-facies minerals where the granulites were impregnated (Rubie, 1983, 1998). This, together with the common observation that eclogite-facies minerals are unzoned, pleads for recrystallisation process by local dissolution/precipitation with low element migration. This can explain the compositional heterogeneities of phengite replacing plagioclase together with epidote. This composition is mainly representative of the local environment of phengite crystallisation (microdomain). Moreover, in Ky-rich sample NOK10a, the texture observed for the replacement of granulite-facies garnet by a phengite + omphacite + kyanite assemblage (Fig. 3e) strongly suggests a pseudomorphic-like replacement. This indicates that the replacement phases precipitated on the site of dissolution of the reactant (garnet; Carmichael, 1969). This process involves element migration only at very small scale (scale of the reacting phase), and thus easily explains the inheritance of chemical features of the precursor minerals by secondary eclogite-facies minerals.

The efficiency of fluids involved in metamorphic processes for promoting element mobility is mostly dependent on the fluid/rock ratio (Walter and Wood, 1984). In the case discussed here, the fluid itself did not modify the composition of the system, except for an H₂O supply as revealed by whole rock analyses (no loss or gain of elements between initial granulite and secondary eclogite). However, there are elements in excess which were released from eclogite-facies metamorphic reactions (mostly occurring after plagioclase breakdown), but which did not escape the system (hand-sample scale) via the fluids. On the contrary, they were locally reprecipitated at sample scale in

protected areas where fluid pressure dropped (quartz lenses in shadow pressure of coronites). This low-scale migration of the elements carried by the fluid phase pleads for a small amount of fluid, which appears to be quickly oversaturated (Ferry, 2000). It seems then that after its introduction, the fluid was locked in the system and that its composition evolved in the course of hydration reactions as suggested by the occurrence of Ba-rich phengite at the edge of Ba-poor phengites crystals in quartz lenses. Therefore it appears that in the case of a low fluid/rock ratio, the fluid instead of promoting homogenisation of the rock composition will localise heterogeneities (quartz lenses). In the same way, ductile deformation triggered by the appearance hydrous minerals like phengite (Austrheim, 1990; Boundy et al., 1992; Austrheim et al., 1997; Bjørnerud et al., 2002; Jolivet et al., 2005) will also localise heterogeneities by segregation of textural sites (segregation of epidote+phengite shearing layers and garnet+omphacite augens). At this point, the texture of the protolith appears to be an important factor to take into account. Indeed, Zack et al. (2002) emphasised that chemical and isotopic disequilibrium observed in the case of high-pressure metamorphic rocks from the Alps are usually reported for eclogites derived from gabbroic protoliths. In contrast, they showed that high-pressure rocks derived from basalts characterised by homogeneous microlitic texture do not reveal these types of disequilibrium. Therefore the protolith texture coupled to ductile deformation and fluid circulation appears to control the amplitude of chemical rehomogenisation of a sample. Thus, for granuloblastic and granular textures, deformation will tend to bound textural microdomains that also correspond to domains of chemical equilibrium, as it has been demonstrated in our study.

6.4. Mobility of elements during retrogression (amphibolitisation)

The chemical composition of secondary amphibolite-facies phases has been compared to composition of the primary eclogite-facies phases in order to describe the element mobility during the retrogression processes. Different mineral phases recorded a more or less penetrative retrograde stage. Indeed, the alteration in the amphibolite facies is outlined by areas, a few microns in size, developed at the rim of phengite and epidote. The omphacite is partially to totally converted into symplectite (Ca-Am+Ab), and the garnet is corroded along rims and fractures and replaced by calcic amphibole. As reported above, alteration phases

of garnet and omphacite show a clear enrichment in LILE in both samples (Figs. 9a–b,e–f and 12a–d). Moreover, whereas in the Ky-rich sample alteration phases of garnet inherited the REE composition of precursor garnet, in the Ph-rich sample, alteration phases of omphacite and garnet are strongly enriched in LREE (Fig. 8a–b,f) and are mainly constituted by calcic amphibole. The strong enrichment in LILE and LREE observed in alteration zones from the Ph-rich sample implies that a LILE and LREE-rich phase broke down during the amphibolite-facies alteration. A likely candidate may be epidote, which is the main reservoir for LREE, Pb and Sr. Rb and Ba probably come from the breakdown of phengite, but this contribution is less important. Therefore the decrease of Sr, Pb and REE concentrations observed in epidote II from the Ph-rich sample (Figs. 8d and 9d) could correspond to more or less advanced stages of alteration in the amphibolite facies. It seems then that elements have been redistributed from one textural site to another during the retrogression, and thus that element mobility was enhanced during retrogression. The amphiboles are hydroxyl minerals that crystallise after garnet and omphacite in the presence of H₂O. Breakdown of epidote can partially be the source for the H₂O component, but also, Andersen et al. (1991a) demonstrated the occurrence of an H₂O-rich free fluid phase in the amphibolite-facies. Moreover, Kühn (2002) reports that the LOI content in the amphibolites is higher than in the eclogites, which suggests a fluid/rock ratio higher during retrogression than during the eclogite-facies stage. A higher fluid/rock ratio during the retrogression, resulting from a second influx of H₂O-rich fluid, is a reasonable explanation for the increase of element mobility at sample scale during retrogression. Moreover, it has been demonstrated that ductile deformation, triggered by the crystallisation of hydrous minerals, remains active during all the exhumation and the retrogression in the amphibolite-facies and was localised in eclogites that are less competent than granulites (Jolivet et al., 2005). Deformation enhances fluid circulation by creating new pathways but also favours recrystallisation kinetics (e.g. Walter and Wood, 1984; Rubie, 1990, 1998; Straume and Austrheim, 1999; Engvik et al., 2001). Active deformation will then promote element redistribution between destabilised sites and sites of new mineral growth (epidote sites vs. garnet/omphacite sites) by coupled processes of dissolution, transport (via the fluid) and recrystallisation. Therefore, fast recrystallisation (less than 1 Ma, Bjørnerud et al., 2002), opposite to long term deformation (at least 15 Ma delay between eclogitisation

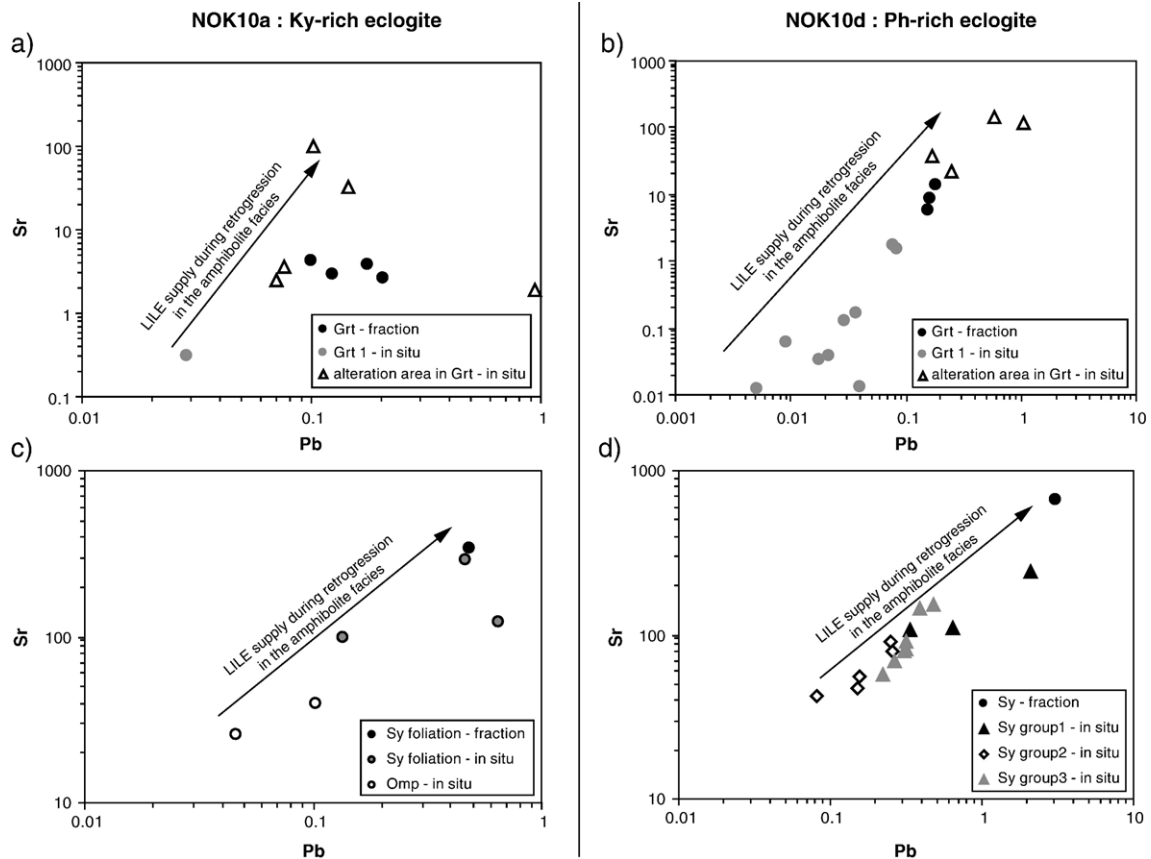


Fig. 12. Sr–Pb correlation diagrams showing the concentration of LILE in garnet (a–b) and symplectite+omphacite (c–d) from samples NOK10a and NOK10d. Analyses were done either on mineral fractions by conventional ICP-MS method (labelled as “fraction” in the caption) or in situ by laser ablation (labelled as “in situ” in the caption).

and amphibolitisation, Austrheim and Råheim, 1981; Austrheim, 1990; Bingen et al., 1998, 2001; Glodny et al., 2002; Kühn, 2002), can also explain the contrasting behaviour in elements migration between eclogite- and amphibolite-facies stages.

7. Conclusion

In the course of this study, focused on the precise recognition of the different textural sites and on the analyses of element repartition in mineral phases from the various metamorphic stages (Grenvillian granulite-facies event, Caledonian eclogite-facies metamorphism and subsequent retrogression in the amphibolite-facies), the impact of the deformation and of fluid circulation on element redistribution at sample scale, has been estimated. A contrasting behaviour for element redistribution between the eclogite-facies stage and the amphibolite-facies retrogression has been demonstrated.

During the eclogite-facies metamorphism, the major and trace element signature of the different minerals reflects mineralogical reactions, which occurred in situ (at the scale of the granulite-facies precursor mineral). This is attributed to processes of recrystallisation by dissolution/precipitation, which operated very quickly (<1 Ma, Bjørnerud et al., 2002) after the impregnation of metastable dry granulite by hydrous fluids. As a result, the composition of eclogite-facies minerals is locally controlled by the composition of few hundred-micron domains (or microsites) in which they crystallised, without fluid circulation or deformation, which can efficiently rehomogenise the composition of a mineral phase at hand-sample scale. On the contrary, ductile deformation promoted the segregation of textural and mineralogical microdomain and fluid circulation favoured hand-sample scale heterogeneities by transport of elements in excess (released during eclogite-facies reactions) toward protected areas where phengite-bearing quartz lenses crystallised.

The study of alteration phases formed during the amphibolite-facies retrogression (symplectite, and alteration zones in garnet) shows that they are generally richer in LILE, Pb and Sr and sometimes in LREE compared to the former eclogite-facies minerals. The origin of these elements is attributed to the breakdown of epidote. Therefore during the amphibolite-facies stage, element mobility is much more important than during eclogitisation because elements concentrated in some specific textural sites (epidote layers) are relocated to other textural sites (garnet+omphacite layers). The enhancement of element mobility during the retrogression is attributed to recrystallisation processes associated to long lasting (tenths of millions years) ductile deformation and increase in the fluid/rock ratio. It appears then that the recrystallisation time scale is a very important factor for element migration, and that, if too short, even deformation will not be able to erase heterogeneities and fluid circulation may be ineffective.

As a consequence of the microdomain segregation process, several phengite fractions separated from a single sample, and related to the same metamorphic stage are heterogeneous. However some of their chemical features allow correlating these fractions to the different populations recognised in situ. Thus, for further investigation like Rb/Sr dating, it will be possible to relate isotopic characteristics to petrogenetic processes. The analyses of several garnet fractions reveal that they are very homogeneous. However, if the comparison between in situ and mineral fractions analyses shows that the REE budget of the garnet fraction is well representative of the garnet itself (no interference related to sub-microscopic inclusions), it also shows that the LILE budget of the garnet fraction is the result of a mixing between pure garnet and alteration phases which are located in grain fractures. This will have repercussion on Rb/Sr isotopic data which can be obtained during geochronological analyses of these various fractions.

The methodology applied here, which consists of in situ analyses by EMP and LA-ICP-MS of the various mineral populations recognised at thin-section scale, coupled to analyses of several fractions of a same mineral species by conventional ICP-MS, is very powerful to discuss the heterogeneities of a mineral population. This coupling enables a precise interpretation of the chemical features depicted by the mineral fractions in comparison to the recrystallisation processes described for the mineral species of interest and can be really useful in order to interpret isotopic data obtained on these fractions.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at [doi:10.1016/j.lithos.2006.10.001](https://doi.org/10.1016/j.lithos.2006.10.001).

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