Rocks occurring in narrow shear zones (<4 m wide) in a gabbro–anorthosite near Nusfjord, Flakstadøy, Lofoten, Norway, include Cl-enriched mineral assemblages, Cl-free mineral assemblages and eclogite-facies assemblages. Mineral equilibria calculations suggest that the different mineral assemblages formed under similar pressure and temperature conditions, at P = 11–14 kbar and T = 650–700°C. One reason for the mineralogical variations of the shear zones is that the rocks evolved from three distinctly different protolith types. Interactions between the rocks and an externally derived Cl-bearing hydrous fluid during shear zone formation resulted in a strong fractionation of the hydrous fluid, and extreme compositional variations of the hydrous mineral phases that formed in equilibrium with the fluid. Parts of the shear zone rocks experienced multiple infiltrations of fluids of different compositions because of local fluctuations of the fluid phase during the fluid–rock interactions. During the deformation, the externally derived fluid was introduced through the transiently highly permeable central parts of the shear zones. The fraction of the fluid that did not escape was rapidly consumed during subsequent hydration reactions.

INTRODUCTION

The character of a fluid as a reactant, catalyst or transporting agent is strongly dependent on the Cl content, and considerable efforts have been made to establish the occurrence and composition of Cl-bearing fluids involved in crustal processes. Many processes involving Cl-enriched crustal fluids are, however, still incompletely understood, including the formation and evolution of such fluids. During interactions between Cl-bearing hydrous fluids and minerals, Cl is strongly partitioned into the fluid phase relative to the minerals. Because of this, Cl-bearing hydrous fluids are often strongly fractionated during fluid–rock interaction, i.e. the geochemical character of such fluids undergoes drastic changes as a function of reaction progress. The mineral assemblage in equilibrium with a fluid of an initial, non-fractionated, composition might be significantly different from the mineral assemblage in equilibrium with a fluid of an evolved and fractionated composition. It might be difficult, however, to decide whether the occurrence of different mineral assemblages in a rock is a result of variations in the chemistry of the fluid phase that equilibrated with the rock, or whether the various mineral assemblages represent differences in pressure and temperature during mineral growth.

Three narrow (<4 m wide) shear zones in a gabbro–anorthosite on the island of Flakstadøy, Lofoten, Norway (Fig. 1), provide an excellent opportunity to study the interaction between minerals and variably fractionated Cl-bearing fluids. The formation of the shear zones was associated with the influx of an externally derived Cl-bearing fluid, which resulted in complex textural and mineralogical variations in the shear zone rocks. In some
parts, metamorphic salt and Cl-enriched scapolite and hydrous silicates can be observed (Kullerud, 1995, 1996; Markl & Bucher, 1998; Markl et al., 1998; Kullerud & Erambert, 1999). In other parts, only Cl-poor hydrous silicates occur. Eclogite has also been described from the shear zones (Wade, 1985; Markl & Bucher, 1997).

It has previously been suggested that the various rocks of the shear zones represent three distinct stages of deformation and metamorphism of the gabbro–anorthositic protolith (e.g. Kullerud, 1992; Markl & Bucher, 1997; Markl et al., 1998): (1) an initial stage under amphibolite-facies conditions associated with the influx of an externally derived Cl-bearing fluid, resulting in the formation of Cl-enriched minerals; (2) reactivation of the shear zones under high-pressure conditions, resulting in the formation of eclogite; (3) reactivation of the shear zones associated with the influx of a Cl-poor fluid at amphibolite-facies conditions, resulting in the formation of Cl-free hydrous minerals.

Recent observations and data that are presented below, however, indicate that the diverse rock types of the shear zones represent three different protolith types. The diverse rock types and the contrasting mineral assemblages can best be explained by a single event of deformation and metamorphism. The mineralogical variations of the shear zone rocks were principally controlled by (1) distinct differences in protolith composition, and (2) extreme compositional variations of the fluid phase on a local scale.

**GEOLOGICAL SETTING**

The studied rocks occur within the Flakstadøya Basic Complex on the island of Flakstadøy, Lofoten, northern Norway (Fig. 1). The basic complex is composed of three lithological units (Romey, 1971): an eastern anorthosite zone, a central troctolite zone and a western norite zone. An Sm–Nd whole-rock isochron constrains the age of the complex to 1803 ± 112 Ma (Wade, 1985). West of Nusfjord (Fig. 1), Wade (1985) discovered two ductile shear zones of 3–4 m width within the norite zone and he gave a petrographic description of what he inferred to be an eclogite-facies assemblage (omphacite and garnet) within the shear zones. During recent fieldwork, one more shear zone, of <60 cm width, was discovered in the same area (shear zone III in Fig. 2).

The outer parts (1–1·5 m width) of the two shear zones of 3–4 m width (Fig. 3a) are characterized by deformation and breakdown of the igneous assemblage of the gabbro–anorthositic protolith. The rock is moderately deformed along the outermost margin of the shear zones (Fig. 3b), but the degree of deformation increases towards the central parts (Fig. 3c and d). In the zone 0·5–1·5 m from the margins the rock is generally highly deformed; however, domains rich in plagioclase and domains rich in mafic minerals, which represent deformed and altered igneous plagioclase and mafic minerals, respectively, can still be recognized. The central parts of the two shear zones are characterized by a dark, commonly garnet-bearing fine-grained rock (Fig. 3d). This rock type is commonly deformed and folded together with the coarser-grained rock of the outer parts of the shear zone. At one locality (see Fig. 2), a few layers of eclogite of 5–10 cm thickness have been observed in the central, highly deformed part of the shear zone (Fig. 3a).
It has previously been suggested that the dark and fine-grained rock that occurs in the central parts of the shear zones represents an extensively deformed and altered equivalent of the gabbro-anorthosite hosting the shear zones (Kullerud, 1995, 1996; Markl & Bucher, 1997). The field relationships of the newly discovered shear zone of 60 cm width (and the geochemical data discussed below), however, suggest that the rock within the central parts of the shear zones represents altered mafic dykes. In the northernmost part of the outcrop, shear zone III (Fig. 2) is composed of marginal zones (of 5–15 cm width) of the characteristically deformed...
gabbro–anorthosite and a central zone (of 30–35 cm width) of the dark and fine-grained rock type (Fig. 4a). The width of the shear zone decreases towards the SE, and in the southernmost part the rock type characteristic of the outer parts of the shear zones is absent, and the dark and fine-grained rock type shows a typical intrusive contact with the undeformed gabbro–anorthosite (Fig. 4b), indicating that the rock was emplaced as a dyke.

The formation of the shear zones was associated with the infiltration of an externally derived Cl-bearing fluid, which resulted in growth of Cl-rich biotite (Kullerud, 1995), Cl-rich amphibole (Kullerud, 1996), salt (halite–sylvite solid solution, Markl & Bucher, 1998; Markl et al., 1998) and Cl-rich scapolite (Kullerud & Erambert, 1999). The Cl-enriched minerals generally occur along the marginal parts of the shear zones. Cl isotope data indicate that the occurrence of the Cl-enriched minerals is a product of remobilized crustal Cl (Markl et al., 1997). The garnet–biotite geothermometer of Perchuk & Lavrent’eva (1983), in combination with a correction for the Cl content of biotite, suggests a temperature of ~580°C (calculated for P = 5 kbar) during the formation of the Cl minerals (Kullerud, 1995). However, it is shown below that more recent calibrations of the geothermometer indicate a significantly higher temperature during this metamorphic stage. Equilibrium between garnet, plagioclase, kyanite and quartz suggests a pressure in the range 8–11 kbar for the formation of the Cl-bearing minerals (Markl et al., 1998).

The observation of Cl-enriched amphibole included in the garnet of the eclogite-facies rock suggests that the eclogite-facies metamorphism post-dated the formation of the Cl-enriched minerals that occur along the marginal parts of the shear zones. Conditions of 680°C and 15 kbar for the eclogite-facies metamorphism (Markl & Bucher, 1997) were calculated for quartz-present conditions. However, new observations suggest low values of $a_{SiO_2}$ during the eclogite formation, indicating that the pressure was significantly below 15 kbar.

For this study, about 90 samples were collected, to cover the mineralogical, textural and geochemical variations of the shear zone rock. In addition, nine samples of the undeformed gabbro–anorthosite have been studied. The sample locations are given in Fig. 2.

TEXTURAL VARIATIONS OF THE STUDIED SAMPLES

The shear zone rocks show large textural and mineralogical variations. On the basis of mineral contents and textures, the samples from the outer parts of the shear zones, which represent the deformed equivalents of the gabbro–anorthosite, have been divided into four
types (types 1–4; see Fig. 3a). Type 1 represents the least deformed rock that commonly occurs along the outermost margins of the shear zones. Towards the central parts, the rock shows gradual transitions between the moderately deformed types 1, 2 and 3, and the highly deformed type 4. The transition from the deformed equivalents of the gabbro–anorthosite to the dyke rock and the eclogite, however, is abrupt (Figs 4 and 5j). Below, detailed descriptions of the igneous rock hosting the shear zones and the various types of shear zone rock are given.

Gabbro–anorthosite

The gabbro–anorthosite hosting the shear zones consists of ~60% plagioclase in addition to clinopyroxene, orthopyroxene, biotite, magnetite and ilmenite (Fig. 3a). Plagioclase commonly shows some extent of undulatory extinction. Occasionally, thin reaction zones (0.01–0.03 mm) can be observed along the grain boundaries between igneous plagioclase and igneous mafic minerals (Fig. 3b). Along the plagioclase–pyroxene contacts, the reaction zones consist of green amphibole, whereas garnet has formed along the plagioclase–biotite and plagioclase–oxide contacts. The extent of the reactions between the primary minerals of the gabbro–anorthosite was apparently highest in the zone of 1–2 m width adjacent to the shear zones, suggesting that the reactions were related to the formation of the shear zones.

Deformed gabbro–anorthosite, type 1

The rock is characterized by a slight rotation and incomplete break down of the primary igneous minerals. The original igneous plagioclase (An_{55}Or_{2}) has been altered to aggregates consisting of small grains (<0.1 mm) of secondary plagioclase (An_{20–55}), Cl-amphibole, Cl- and Ba-enriched biotite, epidote–clinozoisite, Fe–Ti oxides and occasionally Cl-enriched scapolite. Tiny needles (<0.03 mm long) of kyanite, and less commonly corundum, can also be observed in the altered igneous plagioclase. Along the contacts between igneous mafic minerals and the original igneous plagioclase, complex reaction zones composed of Cl-enriched amphibole, garnet, plagioclase and quartz have formed (Fig. 5c and d; see also fig. 4 of Kullerud [1996]). Commonly, the amphibole occurring in these reaction zones is chemically zoned. Green, low-Cl amphibole normally constitutes the inner parts of the reaction zones (adjacent to igneous mafic minerals), whereas blue–green intermediate-Cl amphibole constitutes the outer parts of the reaction zones (adjacent to the altered plagioclase). The blue, most Cl-enriched amphibole occurs within the aggregates of altered igneous plagioclase. The Cl-enriched amphibole, biotite and scapolite described by Kullerud (1995, 1996) and Kullerud & Ercembert (1999) occur in this rock type. Further, Markl & Bucher (1998) and Markl et al. (1998) described the occurrence of metamorphic salt (halite–sylite solid solution) in this rock.

Deformed gabbro–anorthosite, type 2

The rock is slightly more deformed than type 1. Igneous pyroxene has almost completely broken down to aggregates dominated by amphibole (Fig. 5e). The aggregates of mafic minerals commonly occur as slightly elongated lenses in a matrix of altered and recrystallized plagioclase. Abundant small grains (<0.1 mm) of amphibole, biotite, epidote–clinozoisite, kyanite and Fe–Ti oxides occur in the matrix of altered plagioclase. Small grains (<0.03 mm) of corundum can occasionally be observed together with kyanite in the altered plagioclase. The amphibole that replaces igneous mafic minerals is commonly green, whereas amphibole in the plagioclase matrix ranges in colour from green to blue–green.

Deformed gabbro–anorthosite, type 3

The rock is moderately deformed. Igneous pyroxene has been completely replaced by aggregates dominated by green amphibole, plagioclase, garnet and Fe–Ti oxides (Fig. 5f and k). Small grains of secondary clinopyroxene can occasionally be observed in the mafic aggregates (Fig. 5i). The original igneous plagioclase has been extensively overgrown; biotite, green amphibole, epidote–clinozoisite, Fe–Ti oxides, kyanite and minor amounts of corundum have occasionally replaced >50% of the mode of the mineral. Garnet is generally subhedral, and occurs both within the alteration product after igneous pyroxene, and within aggregates of altered plagioclase.

Deformed gabbro–anorthosite, type 4

The rock is highly deformed and foliated. The foliation is defined by continuous bands of 0.5–2 mm width and elongated lenses of alternating plagioclase-dominated and amphibole-dominated compositions (Fig. 5g and h). Occasionally, bands of corundum and plagioclase can be observed (Fig. 6). Bands and lenses dominated by biotite, Fe–Ti oxides and clinozoisite are also observed. In some high-strain zones, rounded grains of garnet and amphibole occur in a mylonitic matrix of amphibole and plagioclase (Fig. 5h). K-feldspar is a common constituent.
Fig. 5. Microtextures of the studied rocks. Scale bar represents 1 mm unless otherwise indicated. Abbreviations of mineral names from Bucher & Frey (1994). (a) Photomicrograph (polarized light) of the undeformed gabbro-anorthosite (sample G5), showing a reaction zone of garnet along the contacts between primary igneous biotite and plagioclase. Amphibole has formed along the contact between primary igneous magnetite and plagioclase. Orthopyroxene and plagioclase. Amphibole has formed along the contact between primary igneous orthopyroxene and plagioclase. Garnet has formed along the contact between primary igneous orthopyroxene and plagioclase. (b) Backscatter electron image of the undeformed gabbro-anorthosite (sample G6), showing a reaction zone of garnet along the contacts between primary igneous biotite and plagioclase, and between primary igneous magnetite and plagioclase. Amphibole has formed along the contact between primary igneous orthopyroxene and plagioclase. Garnet has formed along the contact between primary igneous orthopyroxene and plagioclase. (c) Photomicrograph (polarized light) showing the alteration of primary igneous plagioclase (left-hand side) and primary igneous clinopyroxene (right-hand side) from sample A1. The primary igneous plagioclase has recrystallized and reacted to form a matrix dominated by small grains of plagioclase and Cl-enriched amphibole. Along the original grain boundary between the igneous minerals, a reaction zone dominated by amphibole has formed. The amphibole of the reaction zone is strongly zoned, with low Cl contents adjacent to the clinopyroxene and increasing Cl content towards plagioclase. The Cl-poor amphibole adjacent to clinopyroxene shows a symplectic intergrowth with quartz. (d) Photomicrograph (polarized light) showing the alteration of primary igneous plagioclase (light area) and primary igneous clinopyroxene (dark area) along the marginal parts of the shear zones (sample A1). A complex reaction zone consisting of amphibole, quartz and garnet has formed between the primary igneous minerals. Primary igneous plagioclase has been altered to a fine-grained matrix of plagioclase and Cl-enriched amphibole. Small amounts of biotite, Fe–Ti oxides, epidote and kyanite can also be observed in the matrix of altered plagioclase. (e) Photomicrograph (polarized light) showing the textural type 2 of the deformed gabbro-anorthosite (sample B11). The primary igneous mineral assemblage has almost completely broken down. Relics of igneous pyroxene are, however, observed in the central parts of aggregates of mafic minerals (dominantly amphibole and garnet). Igneous plagioclase (light area) has broken down to a fine-grained matrix of plagioclase, amphibole, biotite, kyanite, magnetite and epidote. (f) Photomicrograph (polarized light) showing the textural type 3 of the deformed gabbro-anorthosite (sample C3). The primary igneous assemblage has been completely replaced. The igneous mafic minerals have been replaced by aggregates consisting of amphibole, garnet and Fe–Ti oxides. Igneous plagioclase has been replaced by...
in the most K-enriched samples (see below) of the shear zone rock types 3 and 4 (Fig. 5i).

**Dyke**

The rock is dominated by elongated grains or aggregates of green amphibole of up to 3 mm length and subhedral grains of garnet (0·5 mm in diameter) in a fine-grained matrix of plagioclase and amphibole (Fig. 5j and l). The amphibole aggregates generally define a weak foliation. The garnet cores contain numerous tiny inclusions that are too small to identify, whereas the garnet rim generally is free of inclusions.

Occasionally, grains of colourless clinopyroxene can be observed in the rock (Fig. 5m). The clinopyroxene is generally overgrown by green amphibole along the rim and internal fractures. It is inferred that this clinopyroxene represents a relic of the igneous mineral assemblage of the dyke rock.

similar minerals to those in (e). (g) Photomicrograph [polarized light] showing the highly deformed type 4 of the gabbro–anorthosite (sample B18). Close-up photomicrograph of inset frame is given in (h). (b) Photomicrograph [polarized light] showing zones of high deformation (sample B18). Rounded grains of garnet occur in dark bands consisting of very fine-grained amphibole and plagioclase. It should be noted that garnet adjacent to the high-deformation zone is euhedral. (i) Backscatter electron image from sample KF9 showing the occurrence of K-feldspar and secondary clinopyroxene. (j) Photomicrograph [polarized light] showing the contact between the deformed equivalent of the gabbro–anorthosite (left-hand side) and the dyke (right-hand side) from sample L4. (k) Photomicrograph [polarized light] showing the details of the deformed equivalent of the gabbro–anorthosite (textural type 3) from sample L4. The primary igneous pyroxene has been completely replaced by aggregates dominated by amphibole. (l) Photomicrograph [polarized light] showing details of the altered dyke from sample L4. Garnet and coarse grains or aggregates of amphibole occur in a fine-grained matrix of amphibole and plagioclase. (m) Photomicrograph [polarized light] of relic igneous clinopyroxene (outlined mineral grains) from the dyke rock (sample B7). The matrix is dominated by small grains of green amphibole and plagioclase. (a) Photomicrograph [polarized light] showing the eclogite (sample Nu1). Omphacite has been partly replaced by symplectite.
Eclogite
The rock is dominated by garnet and pale green omphacite [Fig. 5n, see also fig. 4a of Markl & Bucher (1997)]. Chlorine-bearing apatite is abundant in the garnet-omphacite rock (Fig. 7a). Aggregates of rutile, ilmenite and plagioclase occur in narrow bands (<0.5 mm thick; Fig. 7b). We see no evidence that the plagioclase occurring in these bands has formed during breakdown of omphacite. On the contrary, plagioclase, rutile and ilmenite occur in textural equilibrium with garnet and omphacite, suggesting that the minerals belong to the same metamorphic assemblage. Small grains of blue Cl-enriched amphibole and omphacite commonly occur as inclusions in garnet (Fig. 7b).

Thin reaction zones consisting of plagioclase, magnetite and amphibole are common in fractures in omphacite and along grain boundaries between omphacite and garnet (Fig. 7a). In parts of the rock, omphacite has been replaced by a symplectite consisting of low-Na clinopyroxene and plagioclase [Fig. 5n; see also Markl & Bucher (1997)]. In other parts, omphacite has been undeformed gabbro–anorthosite and garnet (Fig. 7a). In parts of the rock, omphacite has been replaced by a symplectite consisting of low-Na clinopyroxene and plagioclase [Fig. 5n; see also Markl & Bucher (1997)]. In other parts, omphacite has been replaced by green Cl-free amphibole (Fig. 7c–g).

Markl & Bucher (1997) noted that omphacite had formed in fractures in garnet, and inferred that brittle deformation of garnet was related to the eclogite event. They suggested that omphacite growth outlasted garnet growth and the post-garnet brittle deformation. The omphacite-filled fracture shown in Fig. 7c–g, however, suggests an even more complex growth history of the omphacite. The omphacite-filled fracture crosscuts, i.e. it is younger than, the matrix that consists of garnet, plagioclase and green amphibole, which has partly replaced earlier formed omphacite. It should also be noted that the backscatter image of the fracture-filling clinopyroxene (Fig. 7b) shows variations in grey shades, suggesting that the mineral is compositionally inhomogeneous. Similar variations can be observed for the matrix omphacite. Generally, the grey shade is slightly lighter along the rim and along internal fractures than in the unfractured core of the mineral. This may suggest that the original composition of the mineral has later been altered during exchange reactions with an infiltrating fluid.

The following evolutionary history of the rock can be inferred from Fig. 7: (1) formation of Cl-bearing amphibole; (2) growth of garnet + omphacite (+ rutile + ilmenite + plagioclase + apatite); (3) hydration, omphacite is replaced by green Cl-free amphibole; (4) growth of omphacite in fractures; (5) chemical alteration of omphacite during exchange reactions between fluid and mineral. The symplectitization of omphacite and the formation of the reaction zones of plagioclase, magnetite and amphibole along the grain boundaries between omphacite and garnet occurred during stage (3) or later.

COMPOSITIONAL VariATIONS OF CLINOPYROXENE
Electron microprobe analyses of clinopyroxene from the eclogite and of the primary igneous clinopyroxene of the undeformed gabbro–anorthosite and the deformed equivalents of the gabbro-anorthosite have previously been reported by Kullerud (1992) and Markl & Bucher (1997). For the present study, additional analyses of omphacite have been obtained. We have also analysed the chemically altered domains of this mineral (Fig. 7b). Further, the small grains of secondary clinopyroxene from the extensively altered equivalents of the gabbro-anorthosite (Fig. 5i) and the inferred primary igneous clinopyroxene of the dyke rock (Fig. 5m) have been analysed. The analyses reported here were carried out on a Jeol JSM 830 scanning electron microscope equipped with an Edax energy dispersive (EDS) detector, at the University of Tromsø.

Figure 8 shows a Jd–Acm–Quad representation of the clinopyroxene compositions. The clinopyroxene of the undeformed gabbro-anorthosite shows compositions close to quadrilateral pyroxene (Quad). The relics of igneous clinopyroxene in the dyke rock and the deformed equivalents of the gabbro-anorthosite show X_{Jad} in the range 3–15%, which is close to the composition of the symplectic pyroxene of Markl & Bucher (1997). The small grains of secondary clinopyroxene from the extensively altered equivalents of the gabbro-anorthosite show similar compositions.

The clinopyroxene occurring in the matrix of the eclogite and the fracture-filling clinopyroxene have been given different symbols in Fig. 8. The domains of the mineral that show dark grey shades on backscatter images overlap with the omphacite analyses from Kullerud (1992), which show a scatter with X_{Jad} in the range 10–30% and X_{Acm} in the range 0–15%. The three omphacite analyses of Markl & Bucher (1997) have slightly higher jadeite contents (with X_{Jad} ~ 30%) and very low acmite contents. The domains of the clinopyroxene that show light grey shades on the backscatter images, however, show significantly lower values of X_{Jad}. The EDS line scan profile across the fracture-filling clinopyroxene given in Fig. 7b indicates that the transition from high-Na to low-Na compositions is gradual. Similar gradual compositional variations can be observed within the matrix clinopyroxene.

WHOLE-ROCK COMPOSITIONAL DATA
Seventy samples of the shear zone rocks, including samples from three detailed profiles across the shear zone (samples A0–A10, C1–C12 and KF1–KF17), and
nine samples of the undeformed gabbro–anorthosite were analysed for major, minor and trace elements by X-ray fluorescence (XRF) on a Philips PW 1400 instrument at the Department of Geology, University of Tromsø. An additional 22 samples, including samples of eclogite and samples collected along the profile L1–L21, were analysed on a Philips PW 1480 instrument at the Geological Survey of Norway. The concentrations of the trace elements in the 70 shear zone samples were reanalysed at the Geological Survey to avoid inter-laboratory inconsistencies. The trace element data from the laboratory at the Geological Survey have been used in the following figures and calculations, except for data for Rb, Nb, Ce, Cl and S. The concentrations of these elements were generally below the detection limit at the laboratory at the Geological Survey, and analyses from the instrument in Tromsø have been used.

The size of the samples was commonly in the range 300–400 g for the gabbro–anorthosite and 100–300 g for the shear zone rock. For major and minor element analyses, fused pellets containing a mixture of rock powder and lithium tetraborate flux were used (mixed in a ratio of 1:7 for the analyses carried out in Tromsø, and in a ratio of 1:8 for the analyses carried out at the Geological Survey). Trace element analyses were carried out on pressed powder pellets. Special care was taken to avoid Cl contamination of the samples during preparation. A selection of whole-rock analyses is given in Table 1.

The undeformed gabbro–anorthosite
To study the extent of element mobilization during the formation of the shear zones, it is necessary to constrain...
Table 1: Selected whole-rock XRF analyses of the undeformed gabbro–anorthosite and the shear zone rock

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<th>Sample:</th>
<th>G3</th>
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<th>KF1</th>
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SiO₂  48.47  49.01  48.73  47.76  47.52  48.03  49.06  49.19  47.90  47.34  47.96  47.13
TiO₂  1.54  1.23  1.54  1.47  1.51  1.46  1.29  1.38  1.44  1.24  1.20  0.86
Al₂O₃  17.18  19.84  19.05  18.45  18.82  18.76  19.29  17.94  19.19  18.65  19.69  17.13
Fe₂O₃*  14.63  12.21  13.04  13.24  13.00  12.79  11.35  12.06 12.02  12.02  10.59  12.28
MnO  0.19  0.12  0.15  0.16  0.16  0.16  0.14  0.15  0.15  0.15  0.13  0.18
MgO  6.28  4.43  4.81  5.12  4.97  5.11  4.40  4.96  4.76  5.50  4.57  7.11
Na₂O  3.12  3.35  3.25  3.22  3.18  2.95  3.13  3.00  3.22  2.93  2.97  3.02
K₂O  0.50  0.51  0.45  1.34  1.33  1.65  1.74  1.67  1.06  1.42  2.08  0.95
P₂O₅  0.03  0.02  0.03  0.04  0.04  0.03  0.03  0.06  0.04  0.04  0.09  0.27
Total  100.01  100.02  100.12  99.67  99.50  99.51  99.22  99.14  99.11  99.17  99.00  100.16
Rb  0.3  0.0  1.4  17.4  15.8  20.4  19.0  17.0  10.8  13.8  18.0  7.8
Sr  870  1010  1009  958  840  782  1305  1016  1045  1064  1686  673
Y  6  6  10  5  7  6  6  9  11  9  7  16
Zr  19  11  14  13  15  17  16  23  59  27  23  43
Nb  0  0  0  0  0.3  0.4  0.3  1.4  1.2  0.6  0.3  2.0
Ba  654  662  434  962  873  1082  1224  1100  714  895  1114  543
Ce  5  4  15  6  16  13  4  11  13  19  2  25
V  287  274  322  322  305  296  263  289  365  277  258  237
Cr  216  116  204  181  175  183  153  190  197  209  189  296
Co  47  34  51  44  46  45  36  45  38  35  44  34
Ni  35  49  76  76  73  68  59  61  64  61  57  66
Zn  125  72  114  112  105  109  89  96  99  92  79  97
Cu  45  86  163  100  104  117  79  138  97  54  107  48
Sc  27  21  25  24  23  20  19  16  23  29  27  22
Ga  28  28  29  26  27  24  22  22  25  26  27  21
Cl  4  28  111  97  98  78  99  111  72  126  n.a.  n.a.
S  490  1002  86  117  169  172  191  268  151  146  n.a.  n.a.

The initial composition of the rock. Therefore, nine samples of the undeformed gabbro–anorthosite were analysed. The rock shows some variations in the contents of the major elements (Fig. 9). The contents of Al, Ca and Na are negatively correlated with Mg, whereas Fe and Ti correlate positively with Mg. SiO₂ ranges between 46 and 49 wt%, whereas K and Mn show constantly low concentrations.

The variations of the chemical components shown in Fig. 9 can be related to variations in the modal contents of the primary igneous minerals of the gabbro–anorthosite. The contents of Al, Ca and Na principally reflect variations in the modal content of plagioclase, whereas the contents of Fe, Mg, Ti and Mn reflect modal variations of the mafic minerals. The K content of the rock is relatively constant, suggesting a uniform distribution of biotite, which is the only K-bearing phase present in addition to plagioclase (An₂₅Or₃₅, Kullerud & Erambert, 1999). Titanium is principally stored in Fe–Ti oxides and biotite. As the biotite content of the rock is relatively constant, the variations in the Ti content can be related to modal variations of the Fe–Ti oxides.

The shear zone rocks

The compositional variations of the rocks occurring in the shear zones are shown in the element vs TiO₂ diagrams in Fig. 10, which also include the data from the gabbro–anorthosite. The shear zone samples representing the deformed equivalents of the gabbro–anorthosite show large compositional overlaps with the gabbro–
Table 1: continued

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G-A, gabbro–anorthosite; Def., deformed; Ecl., eclogite, n.a., not analysed. Major elements in wt %; trace elements in ppm.

*Total iron as Fe₂O₃.

anorthosite. However, several of these samples show large variations in the contents of K, Ba, Rb, Sr, Nb, Cr and Ni compared with the gabbro–anorthosite. It can also be observed that almost all of the shear zone samples show higher contents of Cl, but lower contents of S than the undeformed gabbro–anorthosite. The samples of the dyke rock and the eclogite show compositions that are distinctly different from the gabbro–anorthosite. The dyke rock can be classified as olivine tholeiitic basalt following the scheme of Irvine & Baragar (1971). Compared with the gabbro–anorthosite and the dyke rock, the eclogite is characterized by a significantly lower content of Si and higher contents of Ti and Fe.

The variations of a selection of chemical components in the shear zone rock across four profiles (for locations see Fig. 2) are shown in Fig. 11. The average compositions of the undeformed gabbro–anorthosite (+1σ) have been indicated for the respective components. The samples from the dykes are indicated with vertical bars of dark grey shading. Eclogite was not observed along any of the sampled profiles shown in Fig. 11.

The following observations in Fig. 11 should be noted:

1) The contacts between the deformed equivalents of the gabbro–anorthosite and the dyke rock are abrupt boundaries between rocks of different chemical composition.

2) The samples of the deformed equivalents of the gabbro–anorthosite from profiles I and KF (except for the samples from the outermost margin along profile KF) show significantly enriched concentrations of K, Rb
and Ba relative to the undeformed gabbro–anorthosite.

3. The Cl content is generally highest (500–2500 ppm) in the least deformed marginal parts of the shear zones. Most of the shear zone samples (including the samples of the dykes), however, show concentrations in the range 100–200 ppm, which is considerably above the Cl content of the undeformed gabbro–anorthosite (<40 ppm).

4. Along profile C, the concentrations of Ni and Cr increase from the NE margin through the deformed equivalents of the gabbro–anorthosite towards the dyke. A similar increase in the concentrations of Ni and Cr towards the SW can be observed along profile A.

5. The Sr content of the deformed equivalent of the gabbro–anorthosite displays large variations compared with the undeformed gabbro–anorthosite along profile C. It should be noted, however, that these variations are correlated with the variations in Si and Al.

POSSIBLE MECHANISMS FOR BULK-ROCK CHEMICAL CHANGES DURING SHEAR ZONE FORMATION

Markl et al. (1998) showed on the basis of whole-rock compositional data that Cl and Ba were introduced to the shear zones together with H2O during metamorphism and deformation. Markl et al. pointed out that other elements possibly were added (e.g. Na, K, Fe, Mg and Si), but in amounts too small to be detected by whole-rock XRF analyses given the variations in mode of the protolith, and concluded that the shear zone formation occurred under nearly isochemical conditions. The data presented in Figs 10 and 11 are, to a large extent, in agreement with the conclusion of Markl et al. (1998). However, the large variations in the contents of K, Rb, Sr, Cr and Ni, in addition to Cl and Ba, must be more closely examined. Another problem that needs to be addressed is whether the eclogite represents an extensively altered equivalent of the gabbro-anorthosite or the dyke rock, or whether it formed from a different protolith.

Infiltration of K, Ba and Rb

The chemical variations across the shear zones (Fig. 11) suggest that K, Rb and Ba were introduced with the externally derived fluid phase along profiles KF and L during shear zone formation. The introduction of K resulted in the formation of K-feldspar (Fig. 5), whereas Ba was principally bonded in biotite (see Kullerud, 1995; Markl et al., 1998). Rubidium was probably incorporated in both K-feldspar and biotite. Along profiles A and C, and along the margins of profile KF, the concentrations of K, Rb and Ba were little affected by the infiltrating fluid. The chemical differences between the various parts of the shear zones may indicate large variations in the K, Rb and Ba contents of the fluid that infiltrated the rock. Alternatively, the chemical differences reflect lateral variations in the total amount of a relatively homogenous K-, Rb- and Ba-bearing fluid that reacted with the rock. The textural variations along profile KF indeed suggest that the integrated fluid flux was considerably higher in the central parts of the shear zones than along the margins. If this interpretation is correct, the K, Rb and Ba contents of the rocks might be used to monitor the total amount of fluid that reacted with the different parts of the shear zones. It should be noted that the samples enriched in K, Rb and Ba are devoid of Cl-enriched minerals (e.g. scapolite, Cl-amphibole and Cl-biotite).

In the Ba vs K diagram in Fig. 12a, the analyses of the undeformed gabbro-anorthosite and ~50% of the analyses of the deformed equivalents of the gabbro-anorthosite form a cluster with concentrations of Ba ~500 ppm and K ~4000 ppm (~0.5 wt % K2O). The other half of the analyses of the deformed gabbro-anorthosite scatter around a trend with a slope corresponding to a Ba/K ratio of 3.3 × 10⁻² (r = 0.79) towards higher concentrations of Ba and K. The analyses of the dyke rock define an almost parallel trend with a Ba/K ratio of 6.3 × 10⁻². In the Rb vs K diagram (Fig. 12b), the analyses of the gabbro-anorthosite and its
Fig. 10. Element vs TiO$_2$ diagrams showing the compositional variations of the undeformed gabbro-anorthosite and the various types of shear zone rock. Grey shaded squares, undeformed gabbro-anorthosite; ●, incompletely hydrated equivalents of the gabbro-anorthosite (types 1 and 2); ○, completely hydrated equivalents of the gabbro-anorthosite (types 3 and 4); ×, dyke rock; +, eclogite.
Fig. 11. Variations in the concentrations of selected chemical components along four profiles across the shear zones. The horizontal dashed lines and light grey fields indicate the average concentrations of the undeformed gabbro–anorthosite ($\pm 1\sigma$). Vertical grey fields indicate the dyke rock. The textural types are indicated at the bottom of the diagram. In the Cl diagrams: ■, samples containing Cl-enriched amphibole, biotite and scapolite.
calculated mixing model reflects the compositional variations of the shear zone rocks to a large extent. We therefore infer that the K, Ba and Rb contents of the shear zone rocks monitor the total amount of the externally derived K-, Ba- and Rb-bearing fluid that interacted with the various parts of the shear zones.

The above calculations indicate that up to 1.4 g of Ba + K was added per 100 g rock by the fluid phase during shear zone formation. To calculate the total amount of fluid that infiltrated the rock, the concentrations of K, Ba and Rb of the fluid must be constrained. Previous workers on the shear zones (Kullerud, 1995, 1996; Markl & Bucher, 1998; Markl et al., 1998) have argued that the Cl content of the infiltrating fluid initially was low, but increased as a result of the fluid-rock interactions (see below). Markl et al. (1998) calculated a composition of \[ X_{\text{H}_2\text{O}} = \frac{\text{H}_2\text{O}}{\text{H}_2\text{O} + \text{Cl}} \] of 0.99 for the least Cl-enriched fluid that reacted with the shear zone rocks. This composition can be assumed to represent the initial composition of the externally derived fluid. For the following calculations of the total amount of fluid that infiltrated the rock, we have assumed that the Cl content of the initial fluid was charge balanced by K, Ba and Rb. Under the assumption that the variations in the contents of K, Ba and Rb of the shear zone rocks represent mixtures between the K, Ba and Rb of their protoliths and the K, Ba and Rb of the infiltrating fluid, the relative proportions of the components in the fluid phase can be calculated from the data in Fig. 12a and b accordingly: K/Ba/Rb is > 500 ppm/30 ppm/1 ppm. This corresponds to molar proportions of K/Ba/Rb of ~1100 mol/20 mol/1 mol. For a fluid with \[ X_{\text{H}_2\text{O}} = 0.99 \], the concentrations of the dissolved components can be calculated as outlined above: Cl = 2.04 \times 10^{-2} \text{ mol/mol H}_2\text{O}, K^+ = 1.97 \times 10^{-2} \text{ mol/mol H}_2\text{O}, \text{ Ba}^{2+} = 3.58 \times 10^{-4} \text{ mol/mol H}_2\text{O}, \text{ and Rb}^{2+} = 1.79 \times 10^{-3} \text{ mol/mol H}_2\text{O}. \] The amount of a fluid of this composition that was necessary for the inferred changes in the K, Ba and Rb contents of the rocks along profile KF is shown in Fig. 14. The model suggests that the most K-, Ba- and Rb-enriched parts of the shear zones were infiltrated by ~70 g H$_2$O per 100 g rock. For comparison, ~1 g H$_2$O is sufficient to react 100 g of the nearly anhydrous gabbro-anorthosite completely to amphibolite. It is therefore inferred that the permeability of the K-, Ba- and Rb-enriched parts of the shear zones was high during the fluid infiltration. It should be noted, however, that the fluid might have contained additional components such as Na, although their effect on the bulk-rock composition was small. Therefore, the calculations presented in Fig. 14 represent minimum values of the amount of fluid that was necessary to

![Fig. 12. Variations in the concentrations of Ba, Rb, Cl and K of the shear zone rocks. Symbols as in Fig. 10. (See text for discussion)](image-url)
The variations of Cl across the shear zones

Chlorine shows, in contrast to K, Rb and Ba, most enriched concentrations in the least altered and least deformed marginal parts of the shear zones, i.e. in the rock type that contains Cl-enriched amphibole, biotite, scapolite and salt (Kullerud, 1995, 1996; Markl & Bucher, 1998; Markl et al., 1998; Kullerud & Erambert, 1999). Samples containing Cl-enriched amphibole (Cl >1 wt %) have been indicated in Fig. 11. Amphiboles occurring in the dyke rock and in the more extensively hydrated and deformed equivalents of the gabbro-anorthosite generally contain <0.2 wt % Cl (Kullerud, work in progress). The uncorrelated variations of Cl and K (Fig. 12c) suggest that the Cl content of the shear zone rocks was controlled by a mechanism significantly different from the mechanism controlling the K, Rb and Ba contents.

The formation of Cl-enriched biotite, amphibole, scapolite and salt in the least altered rock, which occurs along the marginal parts of the shear, has been discussed by Kullerud (1995, 1996), Markl & Bucher (1998), Markl et al. (1998) and Kullerud & Erambert (1999). The suggested mechanism, which has been termed the ‘desiccation mechanism’ (Markl & Bucher, 1998), can be summarized as follows:

1. During the deformation the rock was infiltrated by a Cl-bearing hydrous fluid.
2. The influx of the externally derived fluid stopped shortly after the initial fluid saturation of the grain boundaries, i.e. the fluid–rock system was closed.
3. Reactions between the primary igneous minerals and the grain-boundary fluid resulted in a successive consumption of the fluid phase and formation of amphibole and biotite.
4. Cl has a pronounced preference for hydrous fluids relative to silicates. Therefore, the first hydrous minerals that formed in equilibrium with the fluid were Cl poor.
5. Continued hydration reactions (with preferential extraction of H₂O from the fluid) in the closed fluid–rock system resulted in a successive increase in the Cl content of the fluid phase and a gradual development of fluid-deformed marginal parts of the shear zones, i.e. in the rock type that contains Cl-enriched amphibole, biotite, scapolite and salt.
6. Cl-enriched amphibole, biotite, scapolite and salt formed in equilibrium with the last highly saline fluid that was present along the grain boundaries.

Following this model, a Cl-enriched brine may evolve from a low-Cl fluid during fluid–rock interaction, provided that the amount of the fluid infiltrating the rock is too low for a complete hydration of the rock, i.e. the fluid–rock ratio must be low. In this case the grain-boundary fluid will be consumed (unless it escapes) before...
the rock is completely hydrated. The final free fluid that is present along the grain boundaries will inevitably be highly saline and amphiboles forming in equilibrium with this fluid will be Cl rich.

Within the extensively hydrated and deformed Cl-poor parts of the shear zones, the fluid–rock ratio was considerably higher than along the Cl-enriched margins (e.g. Fig. 14). In this situation, the rock underwent complete hydration long before the fluid phase evolved to Cl-enriched compositions. It can therefore be assumed that the low-Cl amphiboles that are typical for the central shear zone rock (Cl <0.2 wt %) formed in equilibrium with a fluid with a composition close to the initial composition of the externally derived fluid.

Variations of Sr, Cr and Ni
The variations in the concentrations of Cr and Ni along profile C (Fig. 11) may suggest that these components were transported by diffusion from the dyke rock (~400 ppm Cr, ~90 ppm Ni) and into the surrounding gabbro–anorthositic (~150 ppm Cr, ~40 ppm Ni) during the shear zone formation. It cannot be ruled out, however, that the variations in the Cr and Ni contents of the shear zone rocks reflect modal variations of their protoliths. The irregular variations of Sr along profile C (Fig. 11) cannot be explained as a result of diffusion processes or fluid–rock interaction. Most likely, the variations in the concentrations of Sr along this profile reflect variations in the modal content of plagioclase of the protolith of the rock. This interpretation is supported by the positive correlations between Sr and the concentrations of Al and Si along the profile.

The eclogite
If the shear zone eclogite was formed from a protolith similar to the gabbro–anorthosite or the dyke rock, an extensive exchange of chemical components between fluid and rock must have occurred during the reactions (see Fig. 10). A quantification of the amount of chemical components added or removed from the protolith during such fluid–rock interactions can be carried out following the ‘isocon’ method of Grant (1986). For the present problem, however, this method gives values that are highly unrealistic (removal of 28 g SiO₂ and 12.5 g Al₂O₃ to react 100 g of a gabbro–anorthositic protolith to eclogite; addition of 12 g FeO and 2 g TiO₂ to react 100 g of the dyke rock to eclogite), considering the near-isochemical alteration that was inferred for the rocks elsewhere in the shear zones (see above). We therefore suggest that the eclogite did not form during element exchange between the infiltrating fluid and a protolith similar to the gabbro–anorthosite or the dyke rock.

The formation of eclogite-facies pseudotachylites in shear zones (Austrheim & Boundy, 1994; Austrheim et al., 1996) may provide a mechanism responsible for significant changes in the bulk-rock composition during eclogite formation. Until such a mechanism has been proven and quantified, however, we prefer the explanation that the protolith of the eclogite was chemically distinct from both the gabbro–anorthosite and the dyke rock. Most likely, the protolith of the eclogite corresponds to xenoliths of the Fe- and Ti-enriched varieties of the rocks that occur in the Flakstadøy Basic Complex, as suggested by Markl & Bucher (1997).

PRESSURE AND TEMPERATURE CONDITIONS DURING SHEAR ZONE FORMATION
It has previously been suggested that the shear zones developed during at least three stages of deformation and metamorphism (e.g. Kullerud, 1992; Markl & Bucher, 1997; Markl et al., 1998). During the first stage, the deformation was associated with the infiltration of a Cl-bearing hydrous fluid, which resulted in the formation of Cl-enriched minerals. The formation of eclogite has been related to a second stage of deformation under high P–T conditions. Finally, the replacement of eclogite-facies clinopyroxene by symplectite and the large abundance of low-Cl amphibole in the shear zone rock have been related to a second reactivation, which was associated with the infiltration of a low-Cl hydrous fluid. Important for this interpretation of the shear zone evolution was the assumption that all of the rock types that occur in the shear zones evolved from one common type of protolith, namely the gabbro–anorthosite.

The present data, however, suggest that the various rock types occurring in the shear zones evolved from three different protolith types. This may suggest that the mineralogical and textural differences between the various rock types (e.g. the presence or absence of eclogite-facies mineralogy) were related to differences in the composition of the protoliths and variations in the chemistry of the infiltrating fluid phase, and that all of the rock types formed under the same P and T conditions. To determine this, the P and T conditions during growth of the various mineral assemblages have been re-examined, taking account of the new observations and data presented above.

Formation of the Cl-bearing mineral assemblages
Temperature conditions of ~580°C at 5 kbar (box I in Fig. 15) have previously been calculated for the formation
Formation of the eclogite

Pressures of >14 kbar at ~680°C during the formation of the eclogite assemblage in the Nusfjord shear zones have previously been calculated assuming $a_{SiO_2} = 1$ (Markl & Bucher, 1997; box III in Fig. 15). The temperature calculations were based on the Fe–Mg exchange between garnet and clinopyroxene, whereas the calculated pressures were based on the reaction between albite, clinopyroxene and quartz.

The precise value of $a_{SiO_2}$ during eclogite formation is unknown, because neither quartz, nor kyanite, nor corundum has been observed in this rock type. However,
we find it not likely that $a_{\text{SiO}_2}$ in the narrow bands of the relatively SiO$_2$-poor eclogite was considerably higher than in the surrounding rock. Figure 15 shows isopleths for the reaction plagioclase = clinopyroxene + quartz for $a_{\text{SiO}_2} = 0\cdot7$ and $a_{\text{SiO}_2} = 0\cdot5$ (curves 11a and 11b). The isopleths were calculated following Holland (1980), using the jadeite activity model of Holland (1990), the albite activity model of Newton (1983) and the mineral compositions $X_{\text{Na}}^{\text{clino}} = 0\cdot25$ and $X_{\text{Na}}^{\text{quartz}} = 0\cdot75$.

The garnet–clinopyroxene geothermometer expressions of Ai (1994), Berman et al. (1995) and Ravna (2000) have been used on the average compositions of garnet and clinopyroxene from the Nusfjord shear zone eclogite [data from Markl & Bucher (1997); curves 12–14 in Fig. 15]. For these calculations the Fe$^{3+}$ content of clinopyroxene was calculated according to Fe$^{3+} = \text{Na} - \text{Al}^{VI}$ (= acmite content of clinopyroxene). Combination of the garnet–clinopyroxene equilibrium curves with the equilibrium curves for the reaction plagioclase = clinopyroxene + quartz (curves 11a and 11b) gives pressures and temperatures for the formation of the eclogite in a range that overlaps with the conditions calculated for the formation of the Cl-bearing mineral assemblages (box VI in Fig. 15). We therefore infer that the formation of the Cl-bearing mineral assemblages in the deformed equivalents of the gabbro–anorhasite, and the garnet–omphacite-bearing assemblage of the eclogite, might have formed at similar pressure and temperature conditions. The alteration along the grain margins and internal fractures of both the matrix omphacite and the fracture-filling omphacite (Fig. 7h) may be a result of a change in the fluid activity at a late stage of the eclogite-forming event. If this is correct, the reactions that can be observed in the eclogitic rock can be related to multiple infiltrations of fluids of different compositions, at isothermal and isobaric conditions. Similar multiple events of fluid infiltration were documented by Markl et al. (1998) in the Cl-enriched parts of the deformed gabbro–anorthosite. On the basis of the modal and compositional variations of Cl-bearing amphibole they showed that parts of the rock were infiltrated by three or more fluids of different salinity during the formation of the Cl-bearing amphiboles.

Probably the last reaction that occurred in the eclogitic rocks was the breakdown of omphacite to a symplectite consisting of Na-poor clinopyroxene and plagioclase. Markl & Bucher (1997) suggested that the reaction occurred during isothermal decompression to $\sim 11$ kbar.

### Retrogression

The formation of Cl-poor amphibole in the completely hydrated equivalents of the gabbro–anorhasite and the replacement of the omphacite of the eclogite by Cl-poor amphibole have previously been related to a late reactivation of the shear zones, associated with the infiltration of a Cl-free hydrous fluid (e.g. Kullerud, 1993, 1996). In a previous section, however, it was shown that the Cl-poor mineral assemblages occurring in the completely hydrated equivalents of the gabbro–anorhasite and the Cl-enriched assemblages that occur along the shear zone margins probably were related to the same event of fluid infiltration, i.e. the mineral assemblages formed at similar pressure and temperature conditions. The two generations of omphacite shown in Fig. 7c–g (matrix omphacite and fracture-filling omphacite) are compositionally indistinguishable, indicating that the minerals formed at similar metamorphic conditions. Most likely, the replacement of the matrix omphacite by Cl-free amphibole represents a local and temporary change in the composition and abundance of the fluid phase. Thus, it can be inferred that the two generations of omphacite and the Cl-free amphibole formed under similar pressure and temperature conditions. Possibly, also, the formation of the reaction zones between garnet and omphacite consisting of plagioclase, magnetite and amphibole was an effect of changes in the chemical character of the grain boundary fluid during isothermal and isobaric conditions.

**A MODEL FOR THE SHEAR ZONE FORMATION**

The whole-rock geochemical data presented above suggest that the shear zone rocks formed during interaction between an externally derived fluid and three different protolith types. The compositional variations of the minerals of the shear zones, in particular the amphiboles, indicate large lateral variations in the composition of the shear zone fluid. The formation of the various rock types of the shear zones (rocks containing the Cl-enriched mineral assemblages, eclogitic rocks and rocks containing Cl-free amphibole) can be modelled as a result of only one event of deformation and metamorphism, at pressure and temperature conditions characteristic for the lower parts of a thick continental crust ($P = 11–14$ kbar, $T = 650$–$700^\circ$C). In previous studies shear zone rock formation has been ascribed to three distinctly different metamorphic stages. We also infer that the shear zone formation was associated with only one event of influx of the externally derived fluid. Interaction between the fluid and the shear zone rock, however, resulted in local composition gradients in the fluid phase. During the reactions, the free fluid phase was probably moving along the grain boundaries. Because of this, parts of the shear zone rock experienced several pulses of influx of fluids of different composition.
The source of the fluid that infiltrated the shear zones is unknown. However, because of the high reactivity of hydrous fluids in the nearly anhydrous lower crust (e.g. Frost & Bucher, 1994; Yardley & Valley, 1997), it is not likely that the fluid was transported over a long distance. Probably, the highly reactive fluid originated from a source close to the shear zones, and was introduced through the transiently highly permeable central parts of the shear zones. The free fluid phase existed only for a geologically short period, because it was rapidly consumed during hydration reactions in the shear zone rock.

A model for the formation of the shear zones is given in Fig. 16:

1. Before deformation, the gabbro-anorthosite was intruded by basaltic dykes (Fig. 16a). The protolith of the eclogite is uncertain; probably, it corresponds to xenolithic material, compositionally more mafic than the gabbro-anorthosite.

2. Deformation was focused along the pre-existing dykes (Fig. 16b).

3. Externally derived Cl-bearing fluid was channelled along the central, most extensively deformed parts of the shear zones (zone I in Fig. 16c). In these parts of the shear zones, the original igneous mineral assemblages were completely hydrated, indicating that the fluid-rock ratio was high. The low Cl content of the hydrous minerals reflects the low initial Cl content of the externally derived fluid.

4. Along the marginal parts of the shear zones the permeability was low. The fluid phase was diffusing along grain boundaries from the highly permeable central zone towards the margins (zone II in Fig. 16c). The fluid-rock ratio was very low (0.01–0.02; see above), which resulted in a complete consumption of the free fluid phase during the subsequent reactions between the fluid and the primary igneous minerals. During fluid-rock interaction, the composition of the fluid (e.g. the Cl content) was successively changing according to the desiccation mechanism (see above). This resulted in the formation of salt and amphibole, biotite and scapolite with highly variable Cl content along the shear zone margins.

5. The distributions of the amphibolite and eclogite-facies assemblages within the shear zones were controlled by the bulk-rock chemistry of the protoliths. At the pressure and temperature conditions prevailing during deformation, eclogite-facies assemblages formed within the mafic xenoliths whereas the gabbro-anorthosite and the basaltic dykes transformed to amphibolite-facies rocks. The mineral textures of the eclogitic rock suggest that the rock experienced several pulses of influx of fluids of different compositions. Inclusions of Cl-rich amphibole in garnet suggest that an evolved Cl-rich fluid was present.
at an early stage during the alteration of this rock. The formation of the garnet–omphacite-bearing assemblage reflects a change in the composition of the fluid phase. Possibly, the free fluid phase was totally consumed during the formation of the Cl-rich amphibole, i.e. the garnet–omphacite-bearing assemblage formed under fluid-absent conditions. The replacement of omphacite by Cl-free amphibole indicates that a Cl-free fluid was introduced to the rock. Subsequent changes in the fluid composition or the development of anhydrous conditions resulted in further growth of omphacite. This was followed by fluid infiltration along grain boundaries and internal fractures of omphacite that resulted in element exchange between fluid and mineral. The replacement of omphacite by sympletite during isothermal decompression was probably the last reaction that occurred in the rock.

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