Introduction

Large-scale folds (corrugations) parallel to tectonic transport on extensional detachments are documented for both oceanic and continental core complexes (e.g., Cann et al., 1997; Osmundsen and Andersen, 2001; Jolivet et al., 2004; Tani et al., 2011), but their formation and importance in crustal-scale processes are still debated. There is a link to orogenic processes at deep-crustal levels and especially eclogitised (or serpentinised) dense crustal roots, which are seen as fundamental for the potential energy of the crust and its buoyancy (e.g., Ahrens & Schubert, 1975; Austrheim, 1991; Dewey et al., 1993). Elogites are rare at the Earth’s surface, hence the possibility to study exhumed lowermost crust is limited to a few areas, with western Norway as a classical province (Fig. 1). There, eclogites are present both in Caledonian nappes (e.g., Bergen Arcs) and in the sub-Caledonian parautochthonous basement (Western Gneiss Region; WGR). According to models proposed by Andersen et al. (1994) and Dewey et al. (1993), eclogites formed at extreme depths (>100 km) beneath overriding thrust sheets near the centre of the Caledonian orogen. Subsequent unroofing of the eclogites occurred during extension of the thrust welt, through successive stages (Andersen & Jamtveit, 1990; Engvik & Andersen, 2000; Labrousse et al., 2004; Hacker et al., 2010) of horizontal shortening, vertical flattening and subhorizontal stretching, and large-magnitude extensional shearing and exhumation of the footwall of the Nordfjord–Sogn Detachment (NSD). Alternatively, or additionally, as proposed by Terry et al. (2000), thrust stacking at deep-crustal levels concurrent with transcurrent movements played a major role in addition to the extension in the northwesternmost part of the WGR. Other generally proposed mechanisms for exhumation of subducted crust include thrusting accompanied by erosion (Avigad, 1992) or upper-plate extension (Jolivet et al., 1994), and buoyancy forces driving combined reverse and normal faulting (Chemenda et al., 1997).

In this contribution we examine the eclogitisation and unroofing models for western Norway (e.g., Johnston et al., 2007; Hacker et al., 2010). Our extensive dataset is from the large Engebøfjellet Eclogite body, and surrounding rocks, the former mapped in great detail (more than 15 km of core have been drilled) by DuPont and the Geological Survey of Norway (NGU) because of...
Beneath the detachment, the WGR consists of various deep-crustal orthogneisses and migmatitic gneisses intercalated with supracrustal rocks. It contains numerous enclaves of meta-anorthosite, metagabbro/amphibolite, ultramafic rocks and, in the west and northwest, subordinate bodies of eclogite (Milnes et al., 1997). The eclogite protoliths are generally gabroic, but some eclogite-facies rocks are dioritic or granodioritic (e.g., Korneliussen et al., 1998; Engvik et al., 2000). The ages of the WGR protoliths vary, but are mainly Mesoproterozoic (e.g., Kullerud et al., 1986; Skår, 1998; Austrheim et al., 2003; Krogh et al., 2011).

Figure 1. (A) Geological outline of West and Central Norway, modified from Braathen et al. (2000). (B) Simplified bedrock map of western South Norway. The box locates the study area. UHP and HP zones of eclogite-facies metamorphism. BA – Bergen Arc; D – Dalsfjord, M – Møre region, MTFC – Møre–Trøndelag Fault Complex; N – Nordfjord region, NSD – Nordfjord–Sogn detachment, S – Sunnfjord Region.
Typically, mineral assemblages in country-rock gneisses in the western part of the WGR are amphibolite facies or, locally, granulite facies (e.g., Krogh & Carswell, 1995; Korneliussen et al., 1998; Hacker et al., 2010). A temperature gradient from <550°C in the southwest (Sunnfjord) to c. 800°C in the northwest (Nordfjord and Møre) is recorded for eclogite-facies relics (Labrousse et al., 2004; Hacker et al., 2010). A concurrent increase in pressure is found from Sunnfjord with peak pressure at c. 18 kbar (e.g., Cuthbert et al., 2000) to the Nordfjord area with P ≥ 28 kbar, as indicated by the presence of coesite-bearing UHP eclogites (Smith, 1984; Wain, 1997) and of microdiamonds (Dobrzhinetskaya et al., 1995).

The Caledonian rocks in the hanging wall of the NSD zone can be divided into nappes of pre-Caledonian basement with an overlying cover sequence, exotic nappes, and unconformable, Devonian, suprapeternal basins of low-grade, coarse-clastic sediments (e.g., Andersen & Jamtveit, 1990; Furnes et al., 1990; Osmundsen & Andersen 2001). The initial Scandinian thrusting was towards the SE, whereas late-Caledonian deformation included a reversal in transport towards the western hinterland, due to extension of the orogenic welt (Andersen, 1998; Fossen, 2000, 2010). The Devonian basins are interpreted to have formed during the extension prior to, or during, folding of the Devonian sediments and other rocks into a series of E–W-trending and south-verging synclines and anticlines (Roberts, 1983; Chauvet & Seranne, 1994; Krabbendam & Dewey, 1998; Braathen, 1999; Osmundsen & Andersen 2001; Sturt & Braathen, 2001; Osmundsen et al., 2006; Johnston et al., 2007).

The present position of high-pressure rocks in western Norway shows that the WGR was exhumed from significant crustal depths during the late stages of the Caledonian orogeny (e.g., Andersen & Jamtveit, 1990; Andersen et al., 1994; Hacker et al., 2010). This decompression may have been initiated during upper and middle crust extension at 410–400 Ma (Andersen, 1998; Fossen & Dallmeyer, 1998; Fossen & Dunlap, 1998), contemporaneous with, or closely following, the UHP / HP eclogitisation of the lower crust at around 415 to 400 Ma (Mørk & Mearns, 1986; Terry et al., 2000; Carswell et al., 2003; Root et al., 2004, Young et al., 2007; Glodny et al., 2008; Kylander-Clark et al., 2009; Krogh et al., 2011).

The Engebøfjellet Eclogite is an elongated lens with an E–W long axis. Its length exceeds 3 km, whereas its width is at most approximately 1 km. In map-view the body is asymmetric with a sigmoidal shape (Figs. 2, 3A). Surrounding rocks have been divided into the Hegreneset and Helle complexes by Korneliussen et al. (1998). The Hegreneset complex comprises tonalitic to dioritic and granodioritic mica-bearing gneisses, and maflc to ultramafic rocks, some eclogitised like the Engebøfjellet eclogite. The Helle complex consists mostly of migmatitic granitic to granodioritic gneisses that apparently intrude the Hegreneset complex. All rocks are deformed, and a pronounced foliation is generally parallel to transposed lithological contacts. The overall structure in the region is that of a broad E–W trending and double-plunging antiform (Fig. 2), the Førdefjord antiform, which is tighter in the west near Engebøfjellet. This regional structure probably relates to or was enhanced during formation of the synclines hosting the Devonian basins. The map pattern also shows that the antiform was decapitated by the later, openly folded, NSD zone (Fig. 2, cross-section X–X”).

**Lithology and structure at Engebøfjellet**

The eclogite lens consists of several rock types (Fig. 3A), of which two are massive. A ferrogabbroic eclogite (or 'ferro-eclogite') forms a wide core and several smaller bodies along the southern margin of the main lens. It can be divided into a darker garnet- and rutile-rich (truly ferrogabbroic) and a Fe- and Ti-poorer 'intermediate' type. A leucogabbroic eclogite (or 'leuco-eclogite') dominates in the west. These massive rock types have more foliated counterparts. The well-foliated and layered margin of the ferro-eclogite, some metres wide, defines a transition to more strained parts of the leuco-eclogite. This foliated rock is present in a broad area around the ferro-eclogite, as illustrated in the cross-sections. Other foliated rocks include (i) quartz-phengite gneiss, (ii) marginal rocks surrounding the major eclogite body, with eclogite lenses and layers in a sheared quartz-mica matrix, and (iii) a unit with alternating dm- to m-thick eclogite, amphibolite and quartz-rich layers, all commonly retrograded. Unit (iii) is well developed north of the main body. Rocks found around the main eclogite are amphibolites, dioritic gneiss and granitic augen gneiss that include numerous lenses of retrograded eclogite.

The overall preservation of the eclogite parageneses is fairly good in both the massive rock types and their foliated counterparts (except iii). In the huge volume represented by the central part of this body, retrogression is limited mainly to the amphibolite-facies shear zones (Fig. 3A), where near complete recrystallisation occurred, and adjacent areas with local static retrogression (see Table 1).

Macro- and mesoscopic overprinting structural relationships are found throughout the study area. A structural evolution linked to metamorphic development designated D₃ to D₅ (Fig. 4) can be established based on refolding of folds and cross-cutting foliations/fractures. The Engebøfjellet Eclogite is bounded by amphibolitic D₅-sheared and foliated country-rock gneisses (Fig. 3A) characterised by a well-developed fabric with a mineral stretching lineation (L₂, for mineralogy, see below). Throughout the area, D₅ structures have a consistent orientation, conforming to regional observations (e.g., Engvik and Andersen, 2000; Osmundsen and Andersen, 2001). The S₂-foliation strikes E–W and dips steeply, generally to the north or subordinately to the south. The
Cross-cutting relationships are exemplified by two internal foliations ($D_1$ & $D_2$) of the eclogite lens, which formed prior to the surrounding foliation ($D_3$): they are

$L_3$ stretching lineation has a moderate to subhorizontal plunge either to the east or to the west, as is also the case for the $F_3$-folds.

Figure 2. (A) Bedrock map of the Førdefjord region. For details, see Korneliussen et al. (1998). Lines of cross-sections are indicated. (B) N–S cross-sections of the Førdefjord region. The two lower sections ($Y$–$Y''$ and $Z$–$Z''$) are modified from Korneliussen et al. (1998).
Figure 3. (A) Bedrock map of the Engebøfjellet Eclogite and country rocks, simplified from Korneliussen et al. (1998). Sampling locations (arrows), from E to W: 14 (quarry - samples E7A, E7B, E10, E11, ME38/96A), 12 (E2), 15 (ME39/96), 16 (ME1/97, ME2/97). Other samples (labelled Ex/x) are from drillcores, see Korneliussen & Erambert (1997). Inset: outline of the first-order eclogite lens, with deduced sense of shear. (B) Cross-sections of the Engebøfjellet Eclogite and country rocks. The profiles are located in Fig. 3A. The rock distribution in the subsurface was constructed using both drillhole data and down-plunge projection of observed structures. Since most folds and cut-off lines have steep plunges that change along strike, the method is regarded as uncertain: deeper portions of the cross-sections are interpretative, although consistent with drillhole and gravimetric data. The margin of the first-order eclogite lens shows lateral changes in rock composition: this boundary is outlined between the various cross-sections.
Table 1. Summary of metamorphic parageneses at Engebøfjellet, Sunnfjord, WGR.

<table>
<thead>
<tr>
<th>Metamorphic stage</th>
<th>Bulk composition</th>
<th>Texture</th>
<th>Paragenesis *</th>
<th>Selected samples</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eclogite-facies D1 + D2</td>
<td>ferrogabbroic+ intermediate</td>
<td>foliated, fine grained</td>
<td>Grt + Omp + Amp(Bar) + R+Qz ± Py + Qz + Ph + Czo/Ep + Dol + Ap</td>
<td>E3/159.8, E4/973.3, E4/130.3, E2, ME1/97</td>
<td>Vein in ME38/96: Qz, Amp, Dol, Pty, Bt, Al + Omp</td>
</tr>
<tr>
<td></td>
<td>ferrogabbroic</td>
<td>coarse-grained</td>
<td>Grt + Omp + Amp(Bar) + Dol + Qz + Rt + Py + Czo/Ep + Ph</td>
<td>E103/103.3 A and B</td>
<td>Grt pseudoblasts; Dol-rich</td>
</tr>
<tr>
<td></td>
<td>leucogabbroic</td>
<td>coronitic/pseudomorphic</td>
<td>Grt + Amp + Amp(Bar) + Ph + Pg + Zr + Czo + Ab + R + Py</td>
<td>ME1/97, E3/101.9</td>
<td>Ab in felsic aggregates</td>
</tr>
<tr>
<td></td>
<td>leucogabbroic</td>
<td>foliated</td>
<td>Grt + Amp(Bar) + Omp + P + P + Czo + Qz + R + Qt + Dol, Ab</td>
<td>E3/33.9, E8/35.9.5, E5, E10/74.5A, (E7B)</td>
<td>post-D1 Omp porphyrobl. (E3/33.9; E10/74.5A)</td>
</tr>
<tr>
<td>Qz-rich leuco-eclogite</td>
<td>foliated, coarse-grained</td>
<td>Grt + Ph + P + Dol + Qt + R + (Ab)</td>
<td>E8/22.9</td>
<td>Ab included in helicitic Grt; minor Chl II</td>
<td></td>
</tr>
<tr>
<td>Grt-amphibolite facies D3</td>
<td>layered (felsic/mafic)</td>
<td>foliated</td>
<td>Grt + Amp(Fe-Pg) + Ep + Pl + Ms + Bt + Qz + Ilm + Ap</td>
<td>E7A, E10</td>
<td>Bt + Ms in E7A; Bt only in E10</td>
</tr>
<tr>
<td></td>
<td>leucogabbroic</td>
<td>foliated</td>
<td>Amp + Mg-Hbl + Czo/Ep + Grt + Pl + Qt + Rt + Py ± Ph ± Pg</td>
<td>E10/74.5B and C, E7B</td>
<td>+ symplectites (after Omp?, reaction 1)</td>
</tr>
<tr>
<td></td>
<td>leucogabbroic</td>
<td>foliated</td>
<td>Amp + Mg-Hbl + Grt + Czo/Ep + Pl + Qt + Rt + Py</td>
<td>E11, E8/163.2 (cores)</td>
<td>Grt included in Amp cores; transition D3 to D4</td>
</tr>
<tr>
<td>Amphibolite facies D4</td>
<td>leucogabbroic</td>
<td>Amp + Mg-Hbl + Czo/Ep + Grt + Qt + Bt + Py</td>
<td>E11, E8/163.2 (rims + shear bands)</td>
<td>transition D3 to D4</td>
<td></td>
</tr>
<tr>
<td></td>
<td>leucogabbroic</td>
<td>thin shear bands</td>
<td>Bt + Ep + Amp + Pl + Qt + Py + Qz</td>
<td>E10/74.5B and C, E7B</td>
<td></td>
</tr>
<tr>
<td>Greenschist facies D5</td>
<td>fracture/vein filling</td>
<td>Amp(Act) + Ep + Qt + Ab + Chl + Cal + Ilm + Ttn + Mag</td>
<td>E7/074.0</td>
<td>ferrogabbroic sample</td>
<td></td>
</tr>
</tbody>
</table>

Ecdelgites with advanced coronitic retrogression and unfoliated amphibolites:

<table>
<thead>
<tr>
<th>Retrogression stage **</th>
<th>Bulk composition</th>
<th>Texture/assemblages</th>
<th>Main reactions or product assemblage ***</th>
<th>Selected samples</th>
<th>Comments/Other reactions ***</th>
</tr>
</thead>
<tbody>
<tr>
<td>D3</td>
<td>ferrogabbroic/felsic layer</td>
<td>Grt + Omp + Amp + Pl + Ms + Bt + Qz + Ilm + Ap</td>
<td>(1), (2), (3), (5), (6)</td>
<td>E4/65.9</td>
<td>reaction (1) complete</td>
</tr>
<tr>
<td>D4</td>
<td>ferrogabbroic/felsic layer</td>
<td>Grt + Omp + Amp + Pl + Ms + Bt + Qz + Ilm + Ap</td>
<td>(1), (3), (4a), (4b), (5), (6)</td>
<td>E4/72.9</td>
<td>reaction (1) complete; (4a)/(4b) incipient</td>
</tr>
<tr>
<td>ferrogabbroic</td>
<td>symplectites D4</td>
<td>Amp + Ep + Pl + Ilm + Qt + Py + Cal + Ap</td>
<td>E4/81.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>layered Qz-rich</td>
<td>ferrogabbroic</td>
<td>Grt + Amp + Pl + Qt + Qt + Ab + Cal + Ilm + Ttn + Mag</td>
<td>E1/164.2</td>
<td>reaction (1) complete; Dol + Qt = Tlc + Cal</td>
<td></td>
</tr>
</tbody>
</table>

Notes: * Symbols according to Whitney & Evans (2010), in addition Bar = barroisite.
** Retrogression stage attributed to D3 or D4 according to stability of Grt.
*** Pseudomorphic replacements: (1) Omp = Amp + Pl ± Cpx II (2) Grt + Omp = Grt II + (m) + Amp + Pl (3) Ph = Bt + Pl + Ep (4a) Grt = Amp + Pl ± Ep in mafic layer (4b) Grt = Ep + Bt + Pl ± Amp in felsic layer (5) Amp + Bt = Amp II + Pl (6) Rt = Ilm ± Ttn
## Tectonic evolution of Engebøfjellet Eclogite

<table>
<thead>
<tr>
<th>Stage</th>
<th>Observed structures</th>
<th>Characteristics</th>
</tr>
</thead>
</table>
| D1    | F1-fold            | **Eclogite facies**  
|       |                    | horizontal shortening(?)  
|       |                    | isoclinal folding  
|       |                    | sub-penetrative fabric + banding |
| post-D1 (D2?) | S1 omn needles | **Eclogite facies**  
|       |                    | static growth of prismatic Omp |
| D2    | cleavage F2         | **Eclogite facies**  
|       |                    | coaxial vertical shortening (restored position)  
|       | S1=S0              | tight to isoclinal folds  
|       |                    | penetrative foliation in high-strain zones  
|       | S2-foliation       | cleavage in folds |
| D3    | F3-fold C cleavage  | **Garnet amphibolite facies**  
|       | S3-shear zones     | non-coaxial shortening  
|       |                    | tight + sheath folds w/ cleavage  
|       |                    | sheared fold-limbs  
|       |                    | major rotational strain during sinistral shear. |
| D4    | S4-shear band       | **Amphibolite facies**  
|       |                    | continued non-coaxial deformation (?)  
|       |                    | m-wide shear zones |
| D5    | tension joints      | **Greenschist facies**  
|       |                    | N-S shortening and/or E-W extension  
|       |                    | vertical extension fractures |
| D6    | N map view          | **Lower greenschist facies?**  
|       | of fault system    | (Ep+Qtz+Ank)  
|       |                    | N-S shortening on conjugate strike-slip faults  
|       |                    | and bisecting normal faults |

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Figure 4. Summary diagram of the structural and metamorphic evolution of the Engebofjellet Eclogite.
6B) has variable mode, mineralogy and preferred mineral orientation. Near the margins of the massive to S₁-foliated bodies there is a clear angular relationship between the two eclogitic fabrics. Superimposed NW–SE-oriented D₁ structures occur as open to tight F₂-folds of the D₁-fabric, with development of axial-plane cleavage. With increasing strain, F₂-folds become isoclinal and are commonly sheared out to lenses or boudins, whereas the S₁-foliation in the fold limbs is entirely transposed into the penetrative S₂-foliation. In the latter case, a mineral stretching lineation (L₂) is developed.

At the macro-scale, the incremented D₂ and D₃ strain is displayed as a predominant subvertical orientation of sheared lithological contacts (Fig. 3B). In more detail, starting in the east (cross-section A–A', Fig. 3A, B), the eclogite foliation (S₁) was folded during D₃ into a F₃ synform-antiform pair, both folds with elliptical shape. They verge slightly to the south in that they have open to tight inter-limb angles and the axial surfaces are steeply inclined to the north. In detail, folding of the S₁ foliation resulted in steeply dipping S₁ orientations with strike varying from E–W to N–S (Fig. 6C). A vaguely defined F₃-fold axis plunges steeply to the north, whereas mesoscopic parasitic F₁ folds have a moderate to subvertical plunge to the east (Fig. 6G). Near the fold-hinges, a spaced cleavage, subparallel to the axial surface, is developed. D₂ shear zones in the fold-limbs modify the fold geometry and are accompanied by retrogression of the eclogite into garnet amphibolite. A metre-wide shear zone, striking WNW–ESE (Fig. 6J), deviates from this pattern by cutting the fold-pair and the other amphibolite-facies shear zones, hence it probably post-dates this deformation. This D₄ shear zone, with a weak, subhorizontal, mineral stretching lineation, caused retrogression of the eclogites and garnet amphibolites.

Farther west (cross-section B–B”, Fig. 3), steeply north-dipping shear zones (D₂ ± D₃) divide the area into several bodies. The northern sheared eclogite contains mesoscopic F₁ fold-hinge zones surrounded by well-foliated, sheared-out limbs. This is well expressed by a distinct zone of quartz-phengite gneiss, 5–10 m wide, which forms a marker band that is isoclinally folded within the foliation. The central mass of ferro-eclogite contains a well-preserved D₁-fabric (cross-section C–C”). F₁-folds of the locally preserved (but metamorphosed) magmatic layering (=S₀) are consistent with significant strain and transposition, seen as a foliation that parallels the limbs and axial surfaces of F₁ folds (Fig. 5A, B). Towards the boundary of the ferro-eclogite, the D₁-fabric is progressively deformed into tight to isoclinal, elliptical F₁ folds. A well-developed foliation (S₁) is found in fold limbs; fold hinges develop a crenulation or spaced S₂ cleavage. Mesoscopic F₂-folds in the central part of the first-order eclogite lens plunge steeply NW and SE. They plot along a vague great circle that strikes ESE–WNW and dips steeply north (Fig. 6F). Near the boundary to the leuco-eclogite, a penetrative S₂ fabric predominates.
The $S_1$ fabric dips NNE and SSW, whereas the $L_2$ stretching lineation has a moderate plunge to the WNW, or subordinately to the ESE to SE (Fig. 6D, E).

In the west (cross-section D–D', Fig. 3), a distinct $F_1$ fold-pair of the $S_1$–$S_2$ foliations in the ferro-eclogite and the contact towards the leuco-eclogite is present. This is well displayed by the $S_1$ fabric, which plot along a great circle.

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**Figure 6.** Stereoplots of structural data presented in equal-area, lower-hemisphere stereonets. (A) $S_1$ foliation (poles) in the eastern part of the Engebøfjellet Eclogite. Star indicates the general fold axis (as in plot b and c), based on the distribution of the foliation along a great circle. (B) $S_1$ foliation (poles) of the central part of the Engebøfjellet Eclogite. Dots are recorded $F_1$ fold axes. (C) $S_1$ foliation (poles) from the western part of the Engebøfjellet Eclogite. (D) $L_2$ lineation (triangles) and $S_2$ foliation (squares; poles) from the central and eastern parts of the Engebøfjellet Eclogite. (E) $L_2$ lineation (triangles) and $S_2$ foliation (squares; poles) from the western part of the Engebøfjellet Eclogite. (F) $F_2$ fold axes from the entire Engebøfjellet Eclogite. (G) $L_3$ lineation (triangles), $S_3$ foliation (squares; poles) and $F_3$ fold axes (circles) from the entire Engebøfjellet Eclogite and surrounding wall rocks. (H) Summary plot of average structural orientations for the $D_1$, $D_2$ and $D_3$ structures. (I) Plot of restored (by rotation; axis 265/0, plane 265/70) $D_1$, $D_2$ and $D_3$ structures in H. (J) $D_4$ foliation and $L_4$ lineation from a shear zone near the eastern end of the Engebøfjellet Eclogite. (K) Poles to $D_5$ fractures from the entire Engebøfjellet Eclogite (filled squares). (L) Slip-linear plot of $D_6$ brittle faults.
This circle defines a fold axis that plunges moderately ENE, in accordance with orientations of parasitic folds (Fig. 6C). The macroscopic F3 fold-structure is tight to isoclinal and cylindrical, with an elliptical fold shape and an axial surface steeply inclined to the NNE. A crenulation cleavage is well developed in the knee of the fold. The fold-pair is terminated by an S3 shear-zone to the north.

Two types of structure have no map-scale expression. One type is the very common tensile joints/fractures (SS) that cross-cut all fabrics (S1 to S4) of the first-order lens (Fig. 5A, B). The fractures are normally 0.1 to 1 cm wide and <3 m long. They have a consistent N–S, subvertical orientation throughout the Engebofjellet Eclogite (Fig. 6K); however, similar fractures in eclogite lenses of the southern limb of the Førdefjord anticline strike NW–SE. These fractures thus constitute a marker that may be folded on a regional scale (see discussion). Mesoscopic brittle faults (D1) truncate the D2 fractures. The faults show cm-scale separation of marker beds or bands, and net-slip is suggested from slip-lines on the fault surfaces. Two types of fractures are present; conjugate strike-slip faults and bisecting normal faults. The strike-slip fault, all with steep dips, strike NNE–SSW (dextral) and NNW–SSE (sinistral). The normal faults strike N–S and dip steeply east or west (Fig. 6L).

Petrography

Eclogite-facies parageneses and fabrics

All rocks in the Engebofjellet eclogite body underwent eclogite-facies metamorphism and recrystallisation; only ghost magmatic textures may be found. The degree of textural equilibration of the eclogites depends on their D1 and D2 deformational history. In unstrained leucogabbroic rocks (massive leuco-eclogite), microcrystalline aggregates (Ph + Pg + Ab + Zo + Qz; abbreviations for minerals according to Whitney & Evans, 2010) are presumably pseudomorphs after plagioclase. Small granular amphibole and minor mica replaced mafic phases. Garnet coronas separating felsic and mafic domains highlight former grain boundaries of the parent gabbro (Fig. 7A). With increasing strain, the felsic aggregates recrystallise then disappear (foliated leuco-eclogites). Ferro-eclogites are generally completely recrystallised.

Eclogite-facies parageneses comprise garnet, omphacite, amphibole, phengite, paragonite, clinozoisite/epidote, quartz, carbonate, rutile, apatite and pyrite (Table 1). Accessory minerals are allanite and zircon. Ferro-eclogites are usually rich in garnet (from c. 25 to 55% modal), rutile (true ferrogabbro rocks: >3 to c. 10%), clinopyroxene and amphibole, with minor epidote, phengite and quartz. Leuco-eclogites are rich in phengite and paragonite, clinozoisite and often in amphibole and quartz. In most eclogites, volatile-bearing phases are abundant: amphibole (up to c. 40%), mica, clinozoisite/epidote; carbonate is in some cases a major constituent. These eclogites, especially the ferro-eclogites, are generally fine-grained (with garnet size ≤0.5 mm). Notable exceptions (e.g., coarse-grained ferro- or leuco-eclogites with centimetric poikilitic or helicitic garnet, Table 1 and Fig. 7D) are typically found at the contacts between ferro- and leuco-eclogites. Eclogite-facies veins (quartz-rich, with variable amounts of garnet, carbonate, amphibole, omphacite, pyrite, apatite and allanite) are abundant in the eastern part of the body. Their internal foliation is parallel to the eclogite facies foliation (S3 or S4) of the wall-rock. Segregations in pressure shadows of eclogite boudins and along axial planes of F1 or F2 folds contain quartz, omphacite and rutile.

The two eclogite-facies ‘events’, D1 and D2, recognised in the field are not easily distinguished from one another in thin-section. The S1 and S2 foliations, and the F1-related cleavage, are defined by the shape orientation of omphacite, amphibole, clinozoisite, white micas, oxide strings, quartz ribs and apatite (Fig. 7C). Elongated omphacite and amphibole form wedge-shaped lenses, surrounded by a finer grained matrix, especially along S2. This suggests dynamic recrystallisation during eclogite-facies shearing. In some leuco-eclogites, late amphibole porphyroblasts (up to 3–4 cm in length) grow over S2 (Fig. 7B). They are randomly oriented and commonly found near F1 fold-hinges. Such overgrowth is not documented for S1 (Fig. 4), thus this omphacite seems to post-date S1 and pre-date S2. The lack of orientation suggests growth under static conditions. Similarly, in intermediate and leuco-eclogites, post-D1 (post-D2?) amphibole porphyroblasts are common.

Subhedral garnet has a poikilitic core and generally an inclusion-poor rim. In coarse-grained ferro-eclogites (Table 1, Fig. 7D), poikiloblast cores include amphibole, carbonate, clinozoisite, rutile and pyrite; the irregular overgrowth contains omphacite in addition to the previous phases. Helicitic garnet in coarse, quartz-rich, leuco-eclogites contains quartz, carbonate, rutile, amphibole and albite (albite is found only here and within felsic aggregates in leuco-eclogites).

Amphibolite-facies parageneses and fabrics

In D3 shear zones and shear bands, as well as along the axial planes of F3 folds, major minerals are green amphibole, garnet, clinozoisite-epidote, plagioclase, ilmenite and quartz, and subordinate white mica, biotite, carbonate and chlorite. All minerals are anhedral to subhedral to subangular. Garnet coronas separating felsic and mafic domains highlight former grain boundaries of the parent gabbro (Fig. 7A). With increasing strain, the felsic aggregates recrystallise then disappear (foliated leuco-eclogites). Ferro-eclogites are generally completely recrystallised.
too, there are indications of mineral growth during deformation (Fig. 7F).

**Greenschist-facies fracturing**

The $D_5$ tensile fractures contain fibrous amphibole (actinolite), albite, epidote, quartz, calcite, chlorite, magnetite and titanite. The adjacent wall-rock shows retrogression of omphacite, blue-grey amphibole and garnet to symplectic green amphibole, epidote, plagioclase and magnetite, as well as transformation of rutile to ilmenite or more rarely titanite.

The final stage of deformation (brittle faults of the $D_6$...
stage) did not result in any significant recrystallisation of the eclogites and nearby rocks. Mineral growth is limited to fracture walls, commonly coated with fibrous epidote or quartz and locally with brown ankerite.

Mineral chemistry

Some 30 samples of eclogites, equilibrated D₄ and D₅ amphibolites and D₅ fractures (Table 1) were selected for electron microprobe study, covering the range in textural, mineralogical and chemical variations observed within the Engebofjellet Eclogite (see Korneliussen & Erambert, 1997). Samples with static retrogression were excluded, since they are not well suited for geothermobarometry. Analytical procedures and recalculation methods are described in the Appendix. Selected mineral analyses are presented in Table 3.

Garnet in eclogites is almandine-rich, with Xₘ₄ (=Mg/(Mg + Fe)) increasing from layered quartz-rich eclogites (Alm₄₃.₃₋₄₄.₆ Spₛ₉.₉₋₁₁.₄ Prₚ₄.₇₋₅.₉ Ga₁₄₋₁₅ A₄₋₅.₉) via ferro-eclogites (fine- and coarse-grained: Alm₅₆₋₆₉.₂ Sps₂₋₇.₃ Prp₄.₇₋₁₀.₈ Grs₁₇₋₁₉.₇ Adr₁.₇₋₅.₃) to leuco-eclogites (Alm₄₆₋₆₉.₈ Sps₆₋₈.₃ Prₚ₄.₆₋₆.₃ Grs₉₋₁₄.₄ A₄₋₅.₉). Rims have generally the highest Mg, Fe and lowest Mn content. Poikiloblasts in ferro-eclogites (Fig. 7D) show a bell-shaped Mn profile, with Mg increasing and Ca and Fe decreasing towards the rims (Fig. 8). This ‘prograde’ zoning is not paralleled by a clear evolution in inclusion mineralogy: all are similar to matrix eclogitic phases, although omphacite is found only near or within the overgrowth. A similar zoning is found in helicitic garnet. Garnet in amphibole-poor D₄ amphibolite layers overlaps in composition with eclogite-facies garnet in ferрогаббро samples (core analyses: Alm₄₁₋₄₃.₆ Sps₆₋₈.₃ Prₚ₅₋₅.₉ Grs₁₄₋₁₆.₃ A₄₋₅.₉) via ferro-eclogites (fine- and coarse-grained: Alm₅₆₋₆₉.₂ Sps₂₋₇.₃ Prp₄.₇₋₁₀.₈ Grs₁₇₋₁₉.₇ A₄₋₅.₃) to leuco-eclogites (Alm₄₆₋₆₉.₈ Sps₆₋₈.₃ Prₚ₄.₆₋₆.₃ Grs₉₋₁₄.₄ A₄₋₅.₉). Rims tend to be more homogeneous than cores, reflecting a tendency towards chemical equilibrium.

Clinopyroxene is an omphacite (Morimoto et al., 1988) tending towards aegirine-augite in ferro-eclogites. Its compositional range is J₃₋₆₋₈₋₁₀₉₋₁₃ Aₑ₈₋₁₀₋₁₄₋₁₇ Cₐ₆₋₇₋₈₋₁₀ in ferrogabbroic and intermediate eclogites to J₉₋₁₂₋₁₅ Aₑ₆₋₈₋₁₀₋₁₂ Cₐ₆₋₇₋₈₋₁₀ in leuco-eclogites. Higher Jₐ values are recorded in amphibole-rich layers and intermediates eclogites than in fine-grained ferro-eclogites, and increase in Xₘ₄ from ferro- and intermediates eclogites to leuco-eclogites (Fig. 9). Rims are generally higher in Al (Al⁴₋₅ and Al⁴₋₆) and (Na+K) and lower in Si, Naₘ₄, and often in Xₘ₄. However, in fine-grained ferro-eclogites, Naₘ₄ increases towards the rim. Since Naₘ₄ (glaucophane substitution) increases with pressure (Graham et al., 1989), this could reflect a pressure increase during early ferro-eclogite crystallisation, before a decrease during leuco-eclogite (with post-D₅ – or post-D₇? - amphibole porphyroblasts) crystallisation (e.g., head of the P-T loop). Inclusions in garnet poikiloblasts from coarse ferro-eclogites range from barroisite to magnesio-katophorite and taramite, a range largely accounted for by re-equilibration of inclusion rims with host garnet. Profiles consisting of the largest inclusion cores (to avoid disturbance due to rim re-equilibration) were analysed across garnet to test whether their compositional evolution matched the ‘prograde’ zoning of the garnet. Barroisites within Mn-rich garnet cores, however, are similar to those near or within the overgrowth (e.g., coexisting with omphacite), both plotting generally mid-way within the compositional range for all inclusions (see Fig. 9). Moreover, this reversal in amphibole composition may follow opposite trends along different profiles within a single garnet. Amphibole in recrystallised D₅ amphibolites is a ferro-pargasite or ferro-tschermakite. Amphibole in D₄ amphibolites is a magnesio-hornblende. In both D₅ and D₄ samples, zoning towards the rim corresponds to an increase in Al (mostly Al⁴₋₅), Ca and usually Naₘ₄, and a decrease in Si, Naₘ₄, and Xₘ₄, a continuation of the trend already observed in eclogites. Some porphyroclast cores in D₅ samples that have relatively high Naₘ₄ (Fig. 9) and include garnet are likely relics of the D₅ stage (Table 1). Fibrous amphibole in D₄ veins is a ferro-actinolite.

Foliation phengite in all eclogites and small crystals in felsic segregations in leuco-eclogites has a Si core value of 6.62–6.89 apfu, decreasing to 6.36 apfu towards the rim. Phengitic muscovite in S₃ (sample E7A) has lower Si (6.45–6.60 apfu in cores) and Mg+Fe (0.79–1.03 apfu) than phengite in eclogites, indicating a decrease in the celadonite component. S₃ biotites (samples E10 and E7A) have Al = 2.84–3.29 apfu and Xₘ₄ = 0.43–0.56, and no systematic zoning. Later biotites in these samples (in D₄ shear planes with biotite ± chlorite) show no systematic difference in composition from S₃ biotites, although they tend to plot at the low Al and high Mg end of the compositional range.

Clinozoisite/epidote has a Fe content (Ps= 100 Fe⁴⁺/(Fe³⁺ + Al)) increasing from Ps₂₋₄₋₅₋₆₋₇ in foliated leuco-eclogites to Ps₂₋₄₋₅₋₆₋₇ in intermediate eclogites and Ps₂₋₄₋₅₋₆₋₇ in ferro-eclogites. In felsic pseudomorphs in leuco-eclogite, it is a zoisite/clinozoisite (Ps₄₋₅). Inclusions do not differ in composition from matrix grains, even in garnet porphyroblasts from ferro-eclogites. Epidote in recrystallised D₄ amphibolites has Ps₂₋₄₋₅₋₆₋₇ in
Figure 8. Garnet compositions in atomic proportions of Fet+Mn–Mg–Ca in (A) eclogites (cores and rims) and (B) amphibolites D3 (+ relict Grt in amphibolites D4) compared with compositional fields for eclogite-facies garnets. (C) Zoning profile in garnet porphyroblast from coarse-grained ferro-eclogite (location of profile, see Fig. 7D).
Temperature estimates

Considering Fe\(^{3+}\) and Mg partitioning between garnet and clinopyroxene, the complex growth zoning of garnet and the erratic zoning in clinopyroxene make it unlikely that core compositions could be in equilibrium. Thus, only analyses at stable grain boundaries (matrix grains or inclusion-host pairs) have been used. A major problem is the accuracy of Fe\(^{3+}\) estimates for Na-rich pyroxene: assuming stoichiometry, the calculated Fe\(^{3+}\)/Fe\(^{2+}\) of 0.33 to 0.95 results in a spread in K\text{d} and temperatures of 290–623°C (Ellis & Green, 1979; Table 2). The highest Fe\(^{3+}\) values are unrealistic, as in the nearby Naustdal eclogite, whose mineralogy is similar to the present ferro-/intermediate eclogites, from which a wet chemical analysis of omphacite gave a Fe\(^{3+}\)/Fe\(^{2+}\) ratio of c. 0.6 (Binns, 1967). Assuming Fe\(^{3+}\)/Fe\(^{2+}\) = 0.6 in clinopyroxene from Engebofjellet and \(P = 15\) kbar, temperature estimates range from 535 to 645°C (Ellis & Green, 1979) with a peak around 590°C for ferro-/intermediate eclogites. Temperature estimates using Powell (1985) and Krogh-Ravna (2000a) are systematically lower by c. 25°C and 80°C, respectively (Table 2). All these temperatures may be too low, if the assumed Fe\(^{3+}\)/Fe\(^{2+}\) ratio is still too high for these ferro-eclogites.

D\(_1\) amphibolites (of leucogabbroic composition). Yellow epidote in D\(_4\) veins is Py\(_{31-38}\).

Carbonate in eclogites is a dolomite/ankerite (\(X_{\text{Mg}} = 0.58\) to 0.87). Carbonate in D\(_1\) and D\(_2\) amphibolites, as well as in D\(_3\) veins, is calcite. Plagioclase in felsic aggregates from leuco-eclogites is albite (An\(_{02-05}\)), as are inclusions in garnet from coarse quartz-rich eclogite (An\(_{01}\)). Plagioclase in D\(_1\) amphibolites is An\(_{16-33}\) and An\(_{12-20}\) in D\(_4\) amphibolites. In D\(_3\) veins, it is albite An\(_{16}\). Chlorite in D\(_1\) veins is a clinochlore with Si = 5.94–6.04 apfu, Al = 4.04–4.06 apfu and \(X_{\text{Mg}} = 0.53\).

P–T evolution

Eclogite-facies development: D1 and D2

No significant variation in mineral compositions could be correlated to variations in fabrics associated to the two, successive, eclogite-facies deformation phases. Hence, the P–T estimates below are taken as representative of both the D\(_1\) and the D\(_2\) phases.

\[
D_1 \text{ amphibolites (of leucogabbroic composition). Yellow epidote in } D_4 \text{ veins is } Py_{31-38}.
\]

\[
\text{Carbonate in eclogites is a dolomite/ankerite (} X_{\text{Mg}} = 0.58 \text{ to } 0.87\text{). Carbonate in } D_1 \text{ and } D_2 \text{ amphibolites, as well as in } D_3 \text{ veins, is calcite. Plagioclase in felsic aggregates from leuco-eclogites is albite (} An_{02-05}\text{), as are inclusions in garnet from coarse quartz-rich eclogite (} An_{01}\text{). Plagioclase in } D_1 \text{ amphibolites is } An_{16-33} \text{ and } An_{12-20} \text{ in } D_4 \text{ amphibolites. In } D_3 \text{ veins, it is albite } An_{16} \text{. Chlorite in } D_1 \text{ veins is a clinochlore with } Si = 5.94-6.04 \text{ apfu, } Al = 4.04-4.06 \text{ apfu and } X_{\text{Mg}} = 0.53.\]

\[
P–T evolution
\]

\[
\text{Eclogite-facies development: } D1 \text{ and } D2
\]

No significant variation in mineral compositions could be correlated to variations in fabrics associated to the two, successive, eclogite-facies deformation phases. Hence, the P–T estimates below are taken as representative of both the D\(_1\) and the D\(_2\) phases.
### Table 2. Pressure and temperature estimates on eclogites and amphibolites from Engebøfjellet

#### ECLOGITE FACIES (D1 + D2)

<table>
<thead>
<tr>
<th>Fe-Mg partitioning between Grt and Cpx (a)</th>
<th>XMg Cpx</th>
<th>XMg Grt</th>
<th>XMn Grt</th>
<th>XCa Grt</th>
<th>Kd</th>
<th>T(°C) (°)</th>
<th>P(15 kbar)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ferro-eclogites + intermediate eclogites</td>
<td>0.732-0.925</td>
<td>0.116-0.300</td>
<td>0.002-0.023</td>
<td>0.196-0.281</td>
<td>11.5-32.3</td>
<td>378-601</td>
<td>315-541</td>
</tr>
<tr>
<td>Leuco-eclogites</td>
<td>0.843-0.985</td>
<td>0.236-0.314</td>
<td>0.006-0.016</td>
<td>0.196-0.267</td>
<td>11.8-155</td>
<td>293-622</td>
<td></td>
</tr>
</tbody>
</table>

Fe-Mg partitioning between Grt and Amp (Graham & Powell, 1984) (b)

<table>
<thead>
<tr>
<th>XMg Cpx</th>
<th>XMg Grt</th>
<th>XMn Grt</th>
<th>XCa Grt</th>
<th>Kd</th>
<th>T(°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.62-0.76</td>
<td>0.157-0.281</td>
<td>4.9-7.3</td>
<td>15</td>
<td>399-674</td>
<td>648-718</td>
</tr>
</tbody>
</table>

Fe-Mg partitioning between Grt and Ph (c)

<table>
<thead>
<tr>
<th>XMg Ph</th>
<th>XMg Grt</th>
<th>Kd</th>
<th>P(kbar)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.940-1.190</td>
<td>1.410-1.464</td>
<td>1.5</td>
<td>559-674</td>
</tr>
</tbody>
</table>

#### AMPHIBOLITE FACIES D4

<table>
<thead>
<tr>
<th>XMg Cpx Ph barometer (Waters &amp; Martin, 1993; 1996) (d)</th>
<th>XMg amp</th>
<th>XMg Grt</th>
<th>Kd</th>
<th>P(kbar)</th>
</tr>
</thead>
<tbody>
<tr>
<td>D3 rims</td>
<td>0.30-0.38</td>
<td>0.35-0.46</td>
<td>0.14-0.18</td>
<td>0.02-0.03</td>
</tr>
</tbody>
</table>

#### GREENSCHIST FACIES D5

<table>
<thead>
<tr>
<th>XMg Grt</th>
<th>XFe3+ Amp</th>
<th>Ep</th>
<th>XAlm</th>
<th>XPrp</th>
<th>XSp</th>
<th>XCa</th>
<th>XGr</th>
<th>P(kbar)</th>
</tr>
</thead>
<tbody>
<tr>
<td>E7A</td>
<td>An21-26</td>
<td>0.558-0.609</td>
<td>0.106-0.127</td>
<td>0.017-0.20</td>
<td>0.286-0.297</td>
<td>P1</td>
<td>9.4-10.2</td>
<td>9.7-10.5</td>
</tr>
<tr>
<td>E10</td>
<td>An26</td>
<td>0.579-0.615</td>
<td>0.101-0.117</td>
<td>0.034-0.35</td>
<td>0.259-0.290</td>
<td>P2</td>
<td>10.2-11.3</td>
<td>10.1-10.3</td>
</tr>
</tbody>
</table>

Fe-Mg partitioning between Grt and Cpx (e)

<table>
<thead>
<tr>
<th>XMg Grt</th>
<th>XMn Grt</th>
<th>XCa Grt</th>
<th>Kd</th>
<th>T(°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.33-0.36</td>
<td>0.436-0.469</td>
<td>0.497-0.533</td>
<td>0.238-0.309</td>
<td>0.277-0.446</td>
</tr>
</tbody>
</table>

Fe-Mg partitioning between Grt and Amph (Triboulet, 1992)

<table>
<thead>
<tr>
<th>XMg Amp</th>
<th>XMg Grt</th>
<th>XFe3+</th>
<th>Ep</th>
<th>XAlm</th>
<th>XPrp</th>
<th>XSp</th>
<th>XCa</th>
<th>XGr</th>
<th>P(kbar)</th>
</tr>
</thead>
<tbody>
<tr>
<td>E7A</td>
<td>An18-26</td>
<td>0.273-0.301</td>
<td>0.525-0.540</td>
<td>1.2-8 kbar</td>
<td>E7A</td>
<td>E7A</td>
<td>E7A</td>
<td></td>
<td></td>
</tr>
<tr>
<td>E10</td>
<td>An19-29</td>
<td>2.49-2.89</td>
<td>1.7-1.9</td>
<td>1.1-1.5</td>
<td>300-525</td>
<td>7.8</td>
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</tbody>
</table>

Al content in Amph coexisting with Ep + Pl + Gz (Plyusnina, 1982)

<table>
<thead>
<tr>
<th>XMg Grt</th>
<th>XFe3+</th>
<th>Ep</th>
<th>XAlm</th>
<th>XPrp</th>
<th>XSp</th>
<th>XCa</th>
<th>XGr</th>
<th>P(kbar)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.14-0.15</td>
<td>0.425-0.494</td>
<td>0.108</td>
<td>0.014-0.25</td>
<td>0.384-0.401</td>
<td>0.409-0.503</td>
<td>P5</td>
<td>10.4-11.2</td>
<td>10.1-10.7</td>
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Fe-Mg partitioning between Grt and Cpx:

<table>
<thead>
<tr>
<th>XMg Cpx</th>
<th>XMg Grt</th>
<th>XMn Grt</th>
<th>XCa Grt</th>
<th>Kd</th>
<th>T(°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.62-0.76</td>
<td>0.157-0.281</td>
<td>4.9-7.3</td>
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</table>

Fe-Mg partitioning between Grt and Cpx:

<table>
<thead>
<tr>
<th>XMg Cpx</th>
<th>XMg Grt</th>
<th>XMn Grt</th>
<th>XCa Grt</th>
<th>Kd</th>
<th>T(°C)</th>
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<tbody>
<tr>
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<td>15</td>
<td>399-674</td>
<td>648-718</td>
</tr>
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</table>
Table 3. Microprobe analyses of minerals in eclogites and amphibolites.

<table>
<thead>
<tr>
<th>SAMPLE</th>
<th>E4/97.3</th>
<th>ME1/97</th>
<th>E3/33.9</th>
<th>E103/103.3</th>
<th>ME1/97</th>
<th>E3/33.9</th>
<th>E8/35.9</th>
<th>E103/103.3</th>
<th>ME1/97</th>
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<tbody>
<tr>
<td>Gt</td>
<td>Gt</td>
<td>Cpx</td>
<td>Amp</td>
<td>Amp inclusions (core) in Gt</td>
<td>Ph</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>rim w. Cpx</td>
<td>rim w. Cpx</td>
<td>core</td>
<td>interm.</td>
<td>overgrowth</td>
<td>rim w. Gt</td>
<td>rim w. Cpx</td>
<td>rim w. Gt</td>
<td>core</td>
<td>interm.</td>
</tr>
<tr>
<td>SiO₂</td>
<td>37.99</td>
<td>38.53</td>
<td>38.98</td>
<td>37.63</td>
<td>37.67</td>
<td>37.67</td>
<td>54.99</td>
<td>55.79</td>
<td>56.78</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.07</td>
<td>0.00</td>
<td>0.07</td>
<td>0.15</td>
<td>0.08</td>
<td>0.03</td>
<td>0.09</td>
<td>0.03</td>
<td>0.07</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.03</td>
<td>0.00</td>
<td>0.04</td>
<td>0.00</td>
<td>0.01</td>
<td>0.00</td>
<td>0.03</td>
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<td>99.99</td>
<td>100.31</td>
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The table continues with similar data for other minerals.
Partitioning of Mg–Fe\textsuperscript{2+} between phengite and garnet rims indicates temperatures of 599–674°C (Krog & Råheim, 1978) and 648–718°C (Green & Hellman, 1982) for 15 kbar. These estimates might be too high, since Fe\textsuperscript{3+} in phengite was neglected. Partitioning of Mg–Fe\textsuperscript{2+} between matrix amphibole and garnet rims gives 588 to 677°C (all eclogite types) assuming Fe\textsuperscript{3+} = Fe\textsuperscript{t} in amphibole (Graham & Powell, 1984). Since garnet with similar ‘prograde’ zoning and inclusion pattern has been used to determine the prograde path of eclogites in the Sunnfjord and Nordfjord areas of the WGR (e.g., Cuthbert & Carswell, 1990), we attempted to trace a possible temperature change during growth of garnet poikiloblasts (e.g., Figs 7D, 8) using amphibole inclusion cores (rims have re-equilibrated with adjacent garnet) and corresponding garnet zones (based on garnet zoning). Resulting temperature estimates vary from 566–657°C (high-Mn garnet core) to 534–621°C (intermediate zone) and 575–577°C (near/within overgrowth, with amphibole inclusions). No systematic change in temperature is then found, and all estimates are in the range obtained on matrix eclogite minerals.

The spread in calculated temperature for the garnet-clinopyroxene thermometer and uncertainty on the Fe\textsuperscript{3+}/Fe\textsuperscript{t} ratio do not allow precise estimates. However, considering all estimates above, a peak temperature of around 600°C seems likely during eclogitisation. This conclusion is supported by the actual eclogite-facies mineralogy. The presence of barroisite, not of glaucophane or edenite-pargasite, indicates temperatures above 550°C (Maresch, 1977) and most likely below 700°C (e.g., medium-T eclogites, Mottana et al., 1990). A lower limit of 500–550°C (at P = 15–17 kbar) is also indicated by the absence of lawsonite within felsic aggregates in leuco-eclogites (e.g., above the reaction curve Lws + Jd = Zo + Pg + Qz Holland, 1979).

**Pressure estimates**

The highest Jd content is recorded in omphacites from leuco-eclogites (Jd 45) where this mineral coexists with quartz. In the absence of coexisting plagioclase (albite is found only within felsic aggregates in leuco-eclogites, e.g., not in equilibrium with omphacite), this indicates a minimum pressure of c. 15 kbar at 600°C (Holland, 1980). The presence of albite (An 05) in fine-grained pseudomorphs from leuco-eclogites suggests that pressure never exceeded 16–17 kbar (for T ≤600°C) although the possibility of disequilibrium breakdown of plagioclase (i.e., persistence at higher pressure) cannot be ruled out. The garnet-clinopyroxene-phengite barometer (Waters & Martin, 1993 + updated calibration, 1996) gives 15.5–18.0 kbar at 600°C assuming Fe\textsuperscript{2+} = Fe\textsuperscript{t} in clinopyroxene (or 16–18.5 kbar at 550°C). However, the estimated pressure decreases drastically (ΔP = 2–3 kbar) when Fe\textsuperscript{2+} is used instead of Fe\textsuperscript{t}.

Support for these pressure estimates can be drawn from comparison with the experimental study of Poli (1993) on the amphibolite-eclogite transition (at 550–650°C and 8–26 kbar) for a basaltic composition close to the bulk compositions of the leuco-eclogites at Engebøfjellet.

**Table 3.** Continued.

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<tr>
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<td>*: Structural formula on 23 oxygens and Sum cations (Na+K) = 15</td>
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(Korneliussen et al., 1998). The foliated eclogite (E8/35.9, Table 1), poor in Mn and K and with an equilibrated paragenesis of Grt + Omp + Amp (barroisite) + Pl + Hbl + Qz ± Bt ± Ms, although calibrated several geobarometers on coexisting Grt + Amp give at 600°C (at 15 kbar) deduced from the stability ofPg + Ze + Qz + Ab in pseudomorphs from leuco-eclogite (Holland, 1979).

**Exhumation path: \( D_3 \), \( D_4 \) and \( D_5 \)**

**Stage \( D_3 \) (Garnet amphibolite facies)**

Close attainment of equilibrium is indicated by the relative homogeneity of rim compositions in \( D_3 \) shear-zone amphibolites. Several geothermobarometers have been calibrated that can be applied to such parageneses (Grt + Amp + Ep + Pl + Ilm + Qz ± Bt ± Ms), although the absence of reliable solid-solution models for amphibole and mica renders their use hazardous. For example, the garnet-amphibole geothermometer (Graham & Powell, 1984) gives unrealistically high temperatures and greater spread (617–1034°C, using Fe\(^{3+}\) in amphibole, Table 2), probably because the \( D_3 \) amphiboles with high Fe\(^{3+}\) and Ca M4 values are not in the recommended compositional range for this calibration. Even higher temperatures and greater spread (617–1034°C, using Fe\(^{2+}\) in amphibole) is obtainable with the calibration of Plyusnina (1982), giving compositions of adjacent amphibole and plagioclase coexisting with amphibole near Jd50, as well as a maximum temperature of 750°C (at 15 kbar) deduced from the stability ofPg + Ze + Qz + Ab in pseudomorphs from leuco-eclogite (Holland, 1979).

**Stage \( D_4 \) (Epidote amphibolite facies)**

Application of Plyusnina (1982) to rim compositions of adjacent amphibole and plagioclase coexisting with epidote and quartz in samples E11 and E8/163.2 results in 500–525°C and 7–8 kbar. The pressure estimates for \( D_4 \) and \( D_5 \) consistently bracket the experimentally determined garnet-out boundary for leucogabbroic compositions (\( P = 9 \) kbar, Poli, 1993).

**Stage \( D_5 \) (Greenschist-facies fracturing)**

The composition of amphibole coexisting with Ab + Chl + Ep + Qz gives 300°C and 3–4 kbar (Sample E7/074.0, Table 2) using the empirical calibration of Triboulet (1992). These estimates are broadly consistent with the expected stability field of such a calcite-bearing assemblage in a Fe-bearing system (Liou et al., 1985; Cho & Liou, 1987).

**Discussion**

**Basis for the exhumation model**

Deformation and metamorphism of the Engebøfjellet Eclogite and the surrounding rocks are separated into stages \( D_1 \) to \( D_5 \) (Fig. 4). Several aspects are crucial for kinematic analyses of the stages: On the map-scale, the truncating nature of shear zones and/or marginal, large drag-related folds may indicate the separation across the zone, or overall vergence. On the meso- and micro-scale, shear-sense indicators may be used to confirm the local shear, such as: asymmetrical shear folds, composite ductile shear fabrics, outcrop-scale stacked units and foliation/bedding duplexes, winged porphyroclasts and shear-related veins and fractures (e.g., Hamner & Passchier, 1991). In addition, sense-of-slip along brittle faults can be deduced by slip-related lineations (see Petit, 1987).

The \( D_1 \) stage is expressed as isoclinal folds of the primary layering, and a dominant eclogitic S-tectonite type transposition foliation. A subhorizontal E–W orientation of the \( F_1 \) axes (Fig. 6A) may indicate N–S shortening, but segmentation of the eclogite lens into individually rotated and deformed bodies during superimposed deformation undermines a reliable interpretation. The following eclogitic \( D_2 \) stage (Fig. 4) reveals steeply plunging tight to isoclinal folds of \( S_1 \), with an associated spaced cleavage. More highly strained parts show a penetrative, LS- to SL-foliation and a distinct WNW–ESE-oriented stretching lineation. This foliation represents the dominant fabric in the Engebøfjellet Eclogite. Kinematic indicators in the \( D_2 \) high-strain zones reveal both dextral and sinistral shear-senses for separate zones, respectively. The general lack of asymmetry combined with the structural orientations indicates subhorizontal (in present position, see below) coaxial N–S shortening during the \( D_2 \) phase.

The amphibolite-facies \( S_1 \) foliation represents the main fabric in the country rocks around the eclogite body. In the eclogite, this stage is characterised by folds of \( S_1/S_2 \) with a spaced cleavage. Distinct, metre-wide, LS-type shear zones (\( S_2 \)) were generated in the fold limbs, or reactivated the \( S_2 \) foliation. Abundant \( D_1 \) shear-sense indicators are consistent with non-coaxial sinistral shear along the subhorizontal E–W stretching lineation (Figs 5D, 6G). The \( D_3 \) stage is revealed by localised metre-wide shear zones. Shear-sense indicators are similar to those of...
the D₂ stage, i.e., subhorizontal sinistral shear; hence, the D₃ phase may represent a progressive continuation of the D₂ stage.


The P–T domains for the various stages (D₁–D₅) are plotted in Fig. 10. Peak conditions for eclogite facies at Engebøfjellet are c. 600°C and 15–18 kbar (undifferentiated D₁ and D₂). These estimates are similar to those previously proposed for eclogites in this inner Sunnfjord area and in accordance with the general metamorphic gradient for eclogite-facies metamorphism in the WGR (Krogh & Carswell, 1995). When the same geothermobarometric methods are used, our P–T estimates for D₁+D₂ at Engebøfjellet are compatible to those obtained on various Sunnfjord eclogites by other authors. However, newer P–T estimates using THERMOCALC (or similar thermodynamic database methods) indicate significantly higher pressures (although similar or somewhat higher temperatures) for the eclogite-facies event. At Kvineset in inner Sunnfjord, the P–T conditions for the eclogite facies of 517–561°C and 16–17 kbar estimated by Cuthbert et al. (2000) using Grt-Cpx and Grt-Cpx-Ph geothermobarometry have been recalculated by Labrousse et al. (2004) to 561–613°C and 20–22 kbar using THERMOCALC. In the Dalsfjord area, some 25 km south of Engebøfjellet, Foreman et al. (2005) estimated a peak temperature of 537–633°C and pressure of 17±2 kbar on the Drøsdal eclogite body using, respectively, Grt-Amp thermometry and Grt-Cpx-Ph barometry, but higher T=720–830°C and P=19–21 kbar using THERMOCALC. In the same area, Engvik & Andersen (2000) estimated that eclogitisation at Vårdalneset occurred at 677°C and 16 ± 2 kbar using Grt-Cpx thermometry and Grt-Cpx-Ph barometry, while Engvik et al. (2007) recalculated these conditions to 635°C and 23 kbar using Grt-Omp-Ky-Ph-Qz thermobarometry and THERMOCALC. In the Solund–Hyllestad–Lavik area, preferred P–T estimates by Hacker et al. (2003) for the Lavik mafic eclogites are 700°C and 23 kbar, using THERMOCALC. It is to be noted that THERMOCALC recalculation of the peak metamorphic conditions for the high-pressure Hyllestad mica schists (consisting of Grt + Ph + St + Ky + Cld +Qz+ Rt) to 575°C and 16 kbar by Hacker et al. (2003) differ very little from the original estimates of 570–600°C and 15 kbar obtained by Chauvet et al. (1992).

Considering the similarity of P–T estimates obtained by ‘classical’ geothermobarometry for all these localities and Engebøfjellet, it is likely that recalculation of our analyses with THERMOCALC would similarly lead to higher pressure estimates of around 22–23 kbar for D₁ + D₂. However, such high pressure estimates will be in conflict with observed phase relations at Engebøfjellet and absolute constraints provided by experimental reaction boundaries: a maximum limit for the pressure of 17–18.5 kbar at T= 600–700°C indicated by the persistence of albite in leuco-eclogites (reaction boundary Jd-Ab-Qz; Holland, 1980). Again, because of the similarity of the older P–T estimates, it is not likely that eclogitisation at Engebøfjellet would have happened at a much lower pressure than for the other Sunnfjord eclogite bodies. In mafic eclogites, with mostly divariant mineral assemblages, both classical thermobarometry and P–T grid calculations using thermodynamic databases (e.g., THERMOCALC) are fraught with the same problems (estimation of Fe³⁺ content and poorly constrained solid-solution models for Fe³⁺ and Na-bearing pyroxene, amphiboles and micas). Considering this, we have tried to better constrain the P–T estimates by direct comparison with experimental work (as outlined in 6.1 above). Improved solid-solution models for these minerals might in the future help to resolve the discrepancy between ourestimates and the newer P–T estimates for the Sunnfjord eclogites (cf., Diener & Powell, 2012, and references therein).

Preferred estimates on retrograde shear-zone amphibolites are c. 525–550°C and 9–12 kbar (D₃ stage) and 500–525°C and 7–8 kbar (D₄). Greenschist-facies fracturing (D₅) occurred at c. 300°C and 3–4 kbar. The resulting retrograde P–T path is similar to the one proposed by Krogh & Carswell (1995) for the nearby Naustdal eclogite, although at slightly lower temperature. In particular, no increase in temperature is recorded during initial decompression, a notable difference from the path established for Naustdal. The only other P–T path defined with some detail in the Sunnfjord area has been obtained on the high-pressure Hyllestad mica schists (Hacker et al., 2003; Labrousse et al., 2004). It is consistent with our P–T data for D₁ and D₃; the whole P–T path for Engebøfjellet (to D₅) showing possibly a slight shift to lower temperature. Our P–T estimates for D₁+D₂ would fit nicely as a higher P extension to the P–T path drawn by Hacker et al (2003) for these mica schists but differ drastically from their P–T estimates for the Lavik eclogites from the same area, as discussed above.

No indisputable record of the prograde path was found. Even ‘prograde’ zoned garnet poikiloblasts fail to demonstrate an amphibolite-facies precursor. The lack of correlation between garnet zoning and compositional trends for amphibole inclusions (see also the lack of trend in temperature estimates, although this argument is rather frail) may indicate that these strongly zoned garnets grew entirely under eclogite-facies conditions
(with a possible correlation: core = D₁, rim/matrix = D₂); the garnet zoning being controlled by chemical fractionation rather than by drastic P–T changes. If this is true for those rocks found near lithological contacts and most susceptible to react under metamorphic events prior to eclogite metamorphism, this suggests that most of the present ferrogabbroic rocks may have undergone direct transformation from metastable gabbro to eclogite, as this is undoubtedly the case for the leuco-eclogites with pseudomorphic gabbroic texture.

**Exhumation model**

In order to evaluate the kinematic data of the Engebofjellet Eclogite in a regional perspective, the initial orientation of the various fabrics has to be established. The Engebofjellet Eclogite is located in the northern limb of a regional E–W-trending antiform, which is slightly double-plunging and tighter to the west. This structure is located in the lower plate, beneath distinct mylonites defining the NSD zone (e.g., Braathen et al., 2004), and in rocks that have not been involved in the deformation of this detachment. Since the tight antiform is truncated by the openly folded overlying detachment, this regional folding occurred prior to formation of the NSD zone. This touches on a general discussion as to how the NSD zone should be defined as, for instance, Marques et al. (2007) argued for an increased vertical shortening with increasing metamorphic conditions (increased depth towards the east) in the overall zone separating the WGR from the upper plate (see below).

The variation in the orientation of D₅ subvertical fractures may be used as a guide to relatively young folding. They strike N–S in the Engebofjellet eclogite, i.e., in the northern limb of the Førdefjord antiform, and NW–SE in the southern limb (Fig. 6J). This minor difference indicates that the region was significantly folded prior to D₅ fracturing and subsequent minor folding. In conclusion, the D₅ fractures were probably

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**Figure 10. Model of structural evolution and P–T path for Engebofjellet.** For deformation stages D₁ to D₆, see Fig. 4 and text. Arrows indicate the orientation of the shortening or stretching axis. UC – upper crust; LC – lower crust; DT – detachment; F – fold.
generated during N–S shortening and/or E–W extension, and could be explained as cross-fold fractures that were slightly folded during progressive strain.

Since no unique temporal strain marker of certain Devonian age can be found in the lower plate, reconstructions are debatable. However, if the north limb of the antiform is rotated to horizontal (rotation axis 265/0, rotation plane 265/70; i.e., disregarding the slight westerly plunge), the D2 to D4 fabrics restore to near-horizontal attitudes (Fig. 6H, I). In this case, the D2 eclogitic fabric records coaxial, near-vertical shortening and approximately horizontal E–W stretching, followed by D5 and D6, amphibolitic, top-W, sub-horizontal shear. It is noteworthy that this restoration reveals a good fit with the Dalsfjord eclogites (Engvik & Andersen, 2000). If valid, one implication is that the lower plate was subjected to amphibolite-facies top-W shear prior to regional tight folding, the folds later being decapitated by the NSD zone.

As discussed by Braathen (1999), the following D2 brittle faults correlate with a later stage of N–S contraction and E–W extension/extrusion. This fracture system can relate to regional folding as conjugate shear and cross-fold fractures. Since the fault system has an equal character in both the upper and the lower plates, it suggests that regional uniformity was established at this time; a latest Devonian – Early Carboniferous age has been indicated (discussed in Braathen, 1999).

Based on the kinematics and structural restoration of the Engebøfjellet region, the following sequence of events can be envisaged (Fig. 10): (i) D1 phase of eclogitisation of uncertain origin, (ii) D2 phase of eclogitisation, consistent with vertical flattening and horizontal E–W stretching, (iii) amphibolite-facies, highly ductile, top-W shear (D2–D3) in the WGR, (iv) regional E–W folding (D3), (v) truncation of fold(s) by renewed top-W shear, this time on the more discrete NSD zone, (vi) regional E–W folding of the NSD plus upper- and lower-plate rocks (D4), and (vii) truncation of the folded NSD zone by narrow, low-angle, (semi-)brittle faults at the base of the upper-plate rocks (e.g., Braathen, 1999; Braathen et al., 2004). Some authors argue that these late faults are gently folded (e.g., Torsvik et al., 1997; Krabbendam & Dewey, 1998). This evolution in a general sense mimics that proposed by Johnston et al. (2007). However, the overall link between regional folding and detachment development remains enigmatic: did the extensional detachment(s) repeatedly lock during transpression and unlock (reactivate) during transtension, or did it/they experience progressive development? Our observations support a step-wise evolution, as illustrated in Fig. 10.

In linking the regional P–T path with the tectonic model outlined above, two main P–T segments appear. First, significant decompression from c. 18 to 7 kbar occurred under near-isothermal conditions (c. 600°C to 525°C).

Such a slow thermal response most likely relates to extremely rapid uplift, linked to the period between our stages D1/D2 and D3. Geochronology of the region can constrain this event, and includes: (i) a 405±5 Ma age for eclogitisation (D2?) (Terry et al., 2000; Root et al., 2004, Glodny et al., 2008) (ii) a titanite cooling (c. 500–550°C) age of c. 395 Ma (Tucker et al., 1987; Terry et al., 2000), (iii) 40Ar/39Ar white-mica ages of around 400 Ma (<400 Ma under, and >400 Ma above the NSD) in the Sunnfjord region (e.g., Andersen, 1998), and (iv) a 40Ar/39Ar biotite age of 395 Ma obtained from a pegmatite in a top-W shear zone in the Dalsfjord area (Eide et al., 1999). When these ages are applied to the Engebøfjellet area, they indicate that roughly 30 km of unroofing occurred in <10 Myr, between 405 and 395 Ma. Farther north, in the Nordfjord and More area, even more substantial unroofing took place (Terry et al., 2000). The second P–T segment (D4 to D5) was characterised by cooling from c. 525°C to 300°C and lower (brittle conditions), associated with a fairly small drop in pressure (3–4 kbar, from 7 to 3 kbar or lower). According to Eide et al. (1999), using 40Ar/39Ar data (feldspar, phengite, biotite), a phase of rapid cooling took place in Sunnfjord in Early Carboniferous time (c. 360 Ma), which may well apply to our D4 stage. This brackets the time span of this P–T segment to a period of c. 30 Myr.

**Comparison with previous models for the WGR**

The prolonged tectonic scenario for Engebøfjellet shows similarities with the proposed setting of nearby areas. Along Dalsfjord to the south, Engvik & Andersen (2000) envisage that two eclogite-facies fabrics were formed by bulk horizontal shortening and vertical stretching of the deep crust, related to plate convergence, followed by rapid decompression during orogenic extensional collapse. There is a clear similarity in estimated pressure for the Dalsfjord (Engvik & Andersen, 2000; Foreman et al., 2005) and Engebøfjellet eclogites (if estimated by the same method; see discussion above); both reveal two eclogite fabrics related to the same pressure regime (c. 18 kbar). The estimated P–T domain for the retrograde amphibolite fabric (c. 550°C and 10 kbar) is also strikingly similar, thus we suggest that the crustal position of these units was more or less identical through the D2–D3 stages. Northwards, in the Molde region, i.e., marginal to the Møre–Trøndelag Fault Complex (MTFC), cooling ages on titanite from the WGR (c. 402–395 Ma; see Labrousse et al., 2004; Hacker et al., 2010) and monazite U–Th–Pb ages of eclogites (407–395 Ma; see Labrousse et al., 2004; Hacker et al., 2010) suggest that regional uplift and homogeneous cooling occurred in a very short period of time. Interestingly, titanite from isolastically down-folded Caledonian nappes is not reset, revealing an Ordovician age, implying that a major extensional detachment existed between the WGR and the nappes prior to the intense folding (Tucker et al., 1997). Furthermore, Robinson (1995) concluded that the folding was contemporaneous with or followed by extension in a sinistral shear field.
According to Terry et al. (2000), early UHP eclogites (407 Ma) were thrust upward and retrograded to HP eclogite, before becoming incorporated in amphibolite-facies top-W and sinistral shear at approximately 395 Ma.

Conclusions

To conclude, our results show similarities to the tectonic framework established both to the south and to the north. In particular, the D2 eclogite-facies and D1-D2 amphibolite-facies setting for Engebøfjellet is consistent with that in Dalsfjord, where eclogitic coaxial as well as top-W shear structures are retrograded by amphibolitic top-W shear zones (Engvik & Andersen, 2000). Northwards, folding plays a major role in the amphibolitic top-W shear zones (Engvik & Andersen, 2000). Terry et al., 2000; Labrousse et al., 2004; Hacker et al., 2010), as we have also found for the Engebøfjellet Eclogite.

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References


Green, T.H. & Hellman, P.L. 1982: Fe-Mg partitioning between coexisting garnet and phengite at high pressure, and comments on a garnet-phengite geothermometer. Lithos 15, 253–266.


Holland, T.B.J. 1979: Experimental determination of the reaction paragonite = jadeite + kyanite + water, and internally consistent thermodynamic data for part of the system Na2O-Al2O3-SiO2-H2O, with applications to eclogites and blueschists. Contributions to Mineralogy and Petrology 68, 293–301.


Appendix

Analytical methods

Minerals were analysed on a CAMECA Camebax Microbeam electron microprobe at the Mineralogical Museum, Oslo. Analytical conditions were accelerating voltage 15 kV, counting time 10 s on peak. Beam current of 10 or 20 nA and focused beam were used for analysis of garnet, pyroxene, amphibole, epidote and sphe re, 10 nA and raster size of 10 or 20 μm for feldspar, micas and carbonate. Natural and synthetic silicates and oxides were used as standards, and analyses corrected according to the PAP procedure. Structural formulae and Fe$^{3+}$ were calculated assuming stoichiometry on 24 oxygen and 16 cations for garnet, 64 cations for pyroxene, amphibole, epidote and sphe re, 10 nA and raster size of 10 or 20 μm for feldspar, micas and carbonate. Among the possible calculation schemes for amphibole (e.g., Leake et al., 1997), only those on 23 oxygen (Fe$_{33}$ = Fe$^{3+}$) and on total cations excluding Ca,Na,K= 13 (and for some calcic amphiboles, sum cations excluding Na,K= 15) satisfied crystallochemical constraints for most amphiboles. The calculation scheme on 13 cations was chosen when plotting amphibole compositions, corresponding here to the maximum allowed Fe$^{3+}$ and Na$_{4}$ apfu; relative compositional trends are, however, little affected by the recalculation scheme adopted. Abbreviations for mineral names are taken from Whitney & Evans (2010) with the addition of Bar (barroisite), Aeg (aegirine) and CaTs (Ca-tschermak component)