CHAPTER 4 - MINERAL CHEMISTRY AND METAMORPHIC CONDITIONS

Mineral chemistry and metamorphic conditions

The aims of this chapter are to present mineral chemical data, to describe the mineral chemistry in relation to parageneses in selected lithologies (with particular reference to high-pressure metamorphic types) and to evaluate the metamorphic conditions under which these parageneses formed in the light of published experimentally and theoretically derived phase equilibria. Finally, a synthesis of pressure-temperature-time (P-T-t) relationships will be presented.

Analytical methods and mineral recalculation schemes are described in Appendices 1 and 2. Results of geothermometry calculations are summarised in Table 4.20.

SECTION 4.01 The Eclogites

4.01.1 Summary of petrography

Detailed petrographic descriptions appear in Chapter 2, the salient features of which are summarised here.

The most common parageneses in the eclogites are:-

garnet + omphacite +/- amphibole +/- rutile
garnet + omphacite + phengite +/- quartz +/- rutile
garnet + omphacite + phengite + zoisite +/- quartz +/- rutile

garnet + omphacite + phengite + zoisite + kyanite +/- quartz +/- rutile

Segregations of zoisite + kyanite + quartz occur in some more aluminous types. Occasionally, Na-poor rocks have pale green or blue-green amphibole in textural equilibrium with omphacite and garnet (figure 2.18) which may become an essential or dominant phase, to give:-

garnet + amphibole + clinopyroxene +/- rutile.

The phases listed above will be referred to as 'matrix' phases.

Solid inclusions are common in the matrix phases (figures 2.19, 2.45). They are not associated with fractures in the garnets and commonly occur in samples where no alteration of garnet rims or matrix grains has occurred. Furthermore, they often show a systematic variation in mineralogy from core to rim, corresponding to a variation in garnet composition. Hence, the inclusions do not appear to be as a result of breakdown of the matrix phases, but rather seem to have been trapped and preserved. Garnet cores commonly contain dark-green amphibole, ilmenite, zoisite and quartz, whilst near garnet rims there are inclusions of omphacite, pale-green amphibole and rutile. Some garnets in the Basal Gneisses contain inclusions of glaucophane co-existing with omphacite (figure 2.56). These features are similar to those found elsewhere in Sunnfjord and by Nordfjord by Bryhni et al (1977) and Krogh (1980, 1982). In addition, zoisite and phengite are found as inclusions in omphacite and kyanite. Garnets frequently form atoll structures around phengite (figure 2.55).

Breakdown products of the matrix phases are variably developed and include diopside + plagioclase symplectites after omphacite, dark-green amphibole kelyphites (+/- magnetite) surrounding garnets and biotite + plagioclase symplectites after phengite (figure 2.55). Barroisitic amphiboles replace omphacite or pyroxene in symplectites and the resulting amphibole + plagioclase +/- quartz symplectites progressively coarsen to produce blocky intergrowths of large 'sieve-textured' amphiboles. Barroisites are often colour-zoned with darker rims by garnets. Kyanites are marginally altered to paragonitic white mica or fibrolitic sillimanite. Kyanites and co-existing zoisites have fine rims of calcic plagioclase.

Four stages of evolution are recognised in the eclogites on the basis of the observations above:-

Stage 1 - Dark-green amphibole +/- zoisite +/- ilmenite +/- quartz +/- garnet +/- white mica (inclusions)

Stage II - Omphacite + garnet +/- pale-green amphibole or glaucophane +/- rutile +/- zoisite +/- phengite (inclusions)

Stage III - Omphacite + garnet +/- zoisite +/- phengite +/- kyanite +/- quartz +/- rutile +/- amphibole (matrix)

Stage IV - Diopsidic clinopyroxene, biotite, plagioclase, biotite, barroisitic amphibole, dark-green amphibole, paragonite, fibrolite.

The chemistry of individual mineral species is now described.

4.01.2 Pyroxenes

Pyroxene analyses are presented in Table 4.1. Endmembers and ferric iron are calculated according to the method of Mysen and Griffin (1971) as discussed in Appendix 2. Compositions are plotted on the jadeite-acmite-augite diagram of Essene and Fyfe (1967) in figure 4.01. All the eclogite matrix (stage III) and inclusion (stage II) pyroxenes are omphacites with jadeite contents in the range 30-49 mole %. Silica contents are high (greater than 1.979 atoms per formula unit) and Na generally exceeds Al, indicating very low tetrahedral aluminium and some ferric iron. Acmite varies from zero to 13 mole %. Jadeite contents vary to some extent with bulk-rock composition with the highest values in rocks with high normative nepheline (eg phengite eclogite D116).

Clinopyroxenes in the garnet amphibolite D206, which are in textural equilibrium with amphiboles, are sodic augites (figure 4.01) with 20 to 26 mole % jadeite and 11-19 mole % acmite.

Matrix omphacites usually lack compositional zoning in jadeite or FeO/MgO with the exception of rare grains in the eclogitic band in the websterite at Gjörlanger (specimen 79/9a) which have diopsidic cores with up to 26 mole % calcium tschermak's end-member. These are probably relics of pre-eclogitic pyroxenes similar to those in the less altered websterite host-rocks.

Omphacite inclusions in garnets are compositionally similar to those in the matrix, but are sometimes slightly enriched in acmite and depleted in jadeite. In the phengite-eclogites from the Basal Gneisses the inclusions are slightly less jadeitic and more diopsidic than the matrix omphacites. This situation is also found in the eclogites from near Förde which are petrographically similar and occupy a similar tectonostratigraphic position at the top of the Jostedal complex (Krogh, 1980).

One analysis of a stage IV clinopyroxene from a symplectite is presented in Table 4.1. It has the composition $Jd_{0.14}$ $Di_{0.73}$ $Ac_{0.09}$ $Ts_{0.03}$ and co-exists with plagioclase of composition $Ab_{0.77}$ $An_{0.23}$.

4.01.3 Amphiboles

Amphibole compositions are set out in Table 4.2. Structural formula recalculations and estimates of ferric iron were carried out according to the method of Robinson et al.(1981) as outlined in Appendix 2. The amphiboles have been classified by the method of Leake (IMA Sub-committee on Amphiboles, 1978). Figure 4.02 shows that they range from actinolites to ferroan paragasites, barroisites and, in one case, taramite.

Two specimens were studied in some detail; a kyanite eclogite (D123) and a garnet amphibolite band from an eclogite (D206). Some analyses were also made from the eclogite host (D205) to the latter.

Sample D123 contains large garnets (up to 5 mm) with well developed amphibole inclusions in the cores (figures 2.45, 4.10a) along with minor zoisite and quartz (stage I). The amphiboles in the garnets vary in habit from small, anhedral subsequent grains to larger, prismatic grains. The zone next to the garnet rim (about 0.5 mm wide) is free of inclusions except for a single zoisite grain. In the eclogite matrix the omphacites are extensively replaced by a pale blue-green pleochroic amphibole.

In figure 4.02 the amphibole inclusions range from actinolites through actinolitic hornblende and magnesiohornblende with one analysis of tschermakitic hornblende. There is a general tendancy from the small, anhedral (resorbed?) grains to be more actinolitic and for the larger, prismatic grains to be more tschermakitic. The matrix amphiboles are barroisites. Mg/(Mg + Fe) shows little variation.

Some chemical features of the D123 amphiboles are plotted on figure 4.04. The inclusions show good positive correlations between tetrahedral aluminium (Al^{IV}) and high-charge octahedral cations ($(Al + Fe^{3+})^{OCT}$) (figure 4.04c) and between sodium in the M4 site (Na^{M4}) and $(Al + Fe^{3+})^{OCT}$ (figure 4.04b). The larger, prismatic and least resorbed grains (assumed to be the least altered and to represent earlier compositions) generally have higher $(Al + Fe^{3+})^{OCT}$ and Al^{IV}. A plot of A-site occupancy ($(Na + K)^{A}$) versus Al^{IV} (figure

4.04a) shows that these also have the highest $(Na + K)^A$. The more resorbed grains show a trend towards actinolitic compositions with decreasing $(Al + Fe^{3+})^{OCT}$, Al^{IV}, Na^{M4} and $(Na + K)^A$. An inflection in each trend implies that an early more rapid decrease in A-site occupancy and Na^{M4} occurs before more constant, low A-site values. This was accompanied by an early, more rapid decrease in Al^{IV} followed by a sharper rate of decrease in the $(Al + Fe^{3+})^{OCT}$ component.

The matrix amphiboles generally follow the early parts of the inclusion trends. They are all rim compositions of grains adjacent to garnets and show rimwards increase in strength of pleochroism. The trends probably represent a retrogressive decrease in edenite and tschermakite components.

The garnet amphibolite D206 has garnets containing two types of amphibole inclusions (figures 4.05 and 4.12a). Dark-green to olive-green and khaki pleochroic amphiboles are usually concentrated in garnet cores. These tend to be rather elongate and rounded. More euhedral, equant amphiboles with very pale-green to blue-green pleochroism occur near the garnet rims. The latter are associated with inclusions of lime-green clinopyroxene and are similar to the matrix (stage III) amphiboles.

In figure 4.06 the D206 amphiboles show very similar trends to those in D123 with the darker green variety being richer in Al^{IV}, $(Al + Fe^{3+})^{OCT}$, $(Na + K)^{A}$ and Na^{M4}, some approaching taramite in composition. The paler inclusions are more actinolitic, the palest resembling the compositions of the cores of matrix amphiboles. Rims of matrix amphiboles tend to be enriched in Al^{IV}, $(Al + Fe^{3+})^{OCT}$ and $(Na + K)^{A}$ so that they reverse the trend shown by the inclusions almost exactly.

A few analyses of amphiboles from bimineralic eclogite D205 are plotted on figure 4.06. Dark-green pleochrioc amphibole inclusions generally occur in garnet cores (figure 4.11a and 2.19). The example here is an aluminous barroisite closely approaching taramite in composition. Amphibole inclusions are generally absent in the outer zones of garnets, where omphacite and rutile are common. Matrix amphiboles are siliceous barroisites whose paragenetic relationships to omphacites are ambiguous but are occasionally clearer later. The D205 amphiboles follow the compositional trends of those from D206 but are rather more sodic.

A few amphiboles from kelyphites replacing garnets (stage IV) have been analysed. They range from tschermakitic hornblendes to rather aluminous barroisites (figure 4.02) and are enriched in Na and Al^{IV} relative to stage II and stage III amphiboles.

No reliable analyses of the glaucophane inclusions from the eclogite from the Basal Gneisses (D116) could be obtained, but their glaucophanitic composition was confirmed by their high Na and verylow Ca on microprobe X-ray scanning imagery.

4.01.4 Micas

Mica analyses are set out in Table 4.3. No estimates of ferric iron have been made. The two main types of mica are phengite and biotite. Minor late paragonite found in a kyanite eclogite was not analysed fully, but its sodic nature confirmed on X-ray imagery.

Phengites occur as matrix phases (stage III) and as large inclusions in atoll garnets (figure 2.55). The inclusions are often contiguous with matrix phengites.

Phengite analyses were carried out on a kyanite eclogite (D77), two quartz-phengite eclogites (79/38b and D116) and a siliceous vein at the margin of an eclogite bounding the Vardheia Unit at Asnes.

Compositions are plotted on figure 4.07. Celadonite substitutions are high with Si^{4+} between 6.56 and 6.96 atoms per formula unit. Iron contents are low with values of 100 Mg/(Mg + Fe) ranging from 68.7 to 80.2. The phengites are not obviously zoned although rims are sometimes more celadonitic than cores. There is, however, a general reciprocal relationship between tetrahedral Al and Mg in sample D77. Replacement of K by Na in phengites ranges from 5-17% with highest values in the phengite eclogite.

Biotites have only been observed in the phengite eclogites of the Basal Gneisses, where they form biotite-plagioclase symplectites which clearly replace phengite. They have 100 Mg/(Mg + Fe) values of 56 to 64 and are intermediate annite-siderophyllite solid solutions with Si^{4+} between 5.38 and 5.52 cations per formula unit. They co-exist with plagioclase of composition $An_{18.99}$ $Ab_{80.17}$ $Or_{0.84}$.

4.01.5 Zoisites

Zoisite analyses appear in Table 4.4. All iron is assumed to be ferric. Of the eclogites studied zoisite only occurs in D205, D77 and D123. Examples from the latter two have been analysed. They have rather constant compositions with only small amounts of pistacite substitution (0.11-0.14 cations per 12.5 $O,(OH)_2$). The zoisites are not significantly compositionally zoned.

4.01.6 Garnets

Garnet analyses appear in Table 4.5. Ferric iron has been estimated by charge balance and analyses recast into end-members by the method of Rickwood (1968). The garnets are zoned with darker red cores corresponding to areas of stage I amphibole inclusions, while paler red rims contain stage II omphacite or amphibole inclusions.

Microprobe analyses reveal that the eclogite garnets show strong chemical zoning, as shown in the diagram of Coleman et al (1965) in figure 4.09. Cores concentrate Mn and Fe relative to Mg and Fe/Mg and Mn decrease rimwards.

Compositional maps have been constructed from X-ray count traverses on the samples with well developed inclusion suites (D123, D205 and D206 - see figures 4.10, 4.11 and 4.12. The counts were crudely converted to mass % FeO, MgO, CaO or MnO by calibrating against full analyses of selected points.

The maps for D123 (figure 4.10) have a broad, bell-shaped profile for MnO with a gentle slope across the inclusion-bearing core, with more rapidly decreasing concentrations towards the rim, with a rather flatter zone in the inclusion-free rim. MnO ranges from 0.8 to 0.1 mass % rimwards.

The profile for FeO/MgO has a very flat, broad core corresponding to the inclusion zone, decreasing sharply towards the rim in the inclusion-free zone. FeO ranges from 26.5 to 21.5 mass % and MgO from 4.5 to 9.0 mass % rimwards.

CaO has a rather flat profile, showing a slight decrease from 9 mass % at the core to 8 mass % at the rim.

The maps for the bimineralic eclogite D205 (figure 4.11) show a bell-shaped profile for MnO with a maximum in the dark amphibole + ilmenite-bearing core (2.5%), flattening out near the omphacite-bearing rim at about 0.5%. There is a slight reverse in the zoning close to the rim. MgO increases rimwards as FeO decreases producing a bell-shaped profile for FeO/MgO which is steepest in the area between the amphibole and omphacite inclusions. FeO ranges from 29-24 mass % and MgO from 3-7 mass % rimwards. CaO has a rather flat profile in the core (8.5-9.0%) with a slight depression in the centre. In the omphacite zone there is a slight but sharp decrease before a slight increase at the rim, with a minimum value of 7.5 mass %.

The garnet amphibolite D206 (figure 4.12) has similar zoning features to D205 except that MnO has a very high peak (3.4%) in the core, decreasing very sharply to about 0.25% in the outer zone with the actinolitic amphiboles and clinopyroxenes. MnO then decreases gently to the rim (0.15%) but shows a slight increase within a few microns of the rim itself (up to 0.3%). FeO/MgO shows the same rapid decrease outside the core as is seen in D205, but the contour distribution is apparently controlled by the presence of large actinolitic amphibole and clinopyroxene inclusions. CaO shows a flat profile across the core (around 8.5%) with a ridge around the core up to 9.3%, then decreasing quite rapidly to about 8.0% at the rim.

In summary, the main features of the zoning patterns are as follows:-

Cores, with inclusions of stage I amphibole + zoisite +/- quartz +/- ilmenite, have high MnO, FeO and CaO, highest FeO/MgO and low MgO.

In the outer zone, containing stage II, inclusions of omphpacite, actinolitic amphibole and rutile, or being free of inclusions, there is a decrease in MnO, FeO, CaO and FeO/MgO with a concomittant increase in MgO. The profiles flatten out by the rim, but there may be an increase in MnO there.

4.01.7 Fe-Mg partitioning and geothermometry

The eclogites contain three mineral phases for which Fe-Mg exchange equilibria with garnet have been calibrated as geothermometers. These are:-

- Garnet + clinopyroxene (Banno, 1970; Rahiem and Green, 1974; Mori and Green, 1978; Ellis and Green, 1979 and Ganguly, 1979).

- Garnet + phengite (Krogh and Raheim, 1978; Green and Hellman, 1982).

- Garnet and hornblende (Graham and Powell, 1984).

The distribution co-efficient for Fe-Mg exchange between garnet and clinopyroxene ($K_D{}^{gt/cpx}$) is defined as:-

$(FeO/MgO)^{gt}/(FeO/MgO)^{cpx}$

(Raheim and Green, 1974 and Ellis and Green, 1979)

Values for $K_D{}^{gt/cpx}$ lie between 6.54 and 17.98 with most values between 12.0 and 14.0. Exceptions are the kyanite eclogite D77, where $K_D{}^{gt/cpx}$ lies consistently between 6.50 and 8.10 and quartz + phengite eclogite D181 with values of 8.83 and 11.83. All the values are derived from closely-spaced rim analyses as these are most likely to represent equilibrium values where garnets are zoned.

Ellis and Green (1979) have shown that $K_D{}^{gt/cpx}$ is affected by the grossular content of the garnet ($X_{gt}{}^{Ca}$) using their relationship.

$$K_D{}^{gt/cpx} = \frac{3104\ X_{Ca}{}^{gt}+3030+10.86P(kbar)}{T(^oK)} - 1.9034$$

Isopleths for K_D vs $X_{Ca}{}^{gt}$ at 16 kbar have been constructed on figure 4.13 and values for garnet-omphacite pairs are plotted. For any eclogite trends of K_D vs X_{Ca}^{gt} are roughly parallel to the isopleths. Inclusion-host pairs tend to show more scatter than rim-rim pairs, possibly as inclusions occupy a significant compositional range on a garnet zoning profile.

Most of the points on figure 4.13 lie between 540°C and 620°C, with the exception of D77 and D181 which cluster around 700°C and 670°C respectively.

The high values for D77 may be a result of the rather high Si concentrations in the omphacites (Table 4.1) which result in a rather high cation excess for Si after the endmember calculation (methods of Mysen and Griffin, 1971, also for Cawthorn and Collerson, 1974). Anomalously high Si values would seriously affect the Fe^{3+} calculation, producing low Fe^{3+}/Fe^{2+} ratios and low K_D values. Note that this will also result in a low acmite end-member proportion and higher jadeite. While these problems may arise from poor quality analyses, it is noteworthy that in one specimen of eclogite (D69) the cores of some omphacites contain minute needles of quartz. Smith and Cheeney (1980) have reported similar features from eclogites in East Greenland and consider the quartz needles to have exsolved from originally non-stoichiometric supersilicic omphacites. In D69 Si contents of the pyroxenes are lower where the quartz needles are present (Table 4.1). Calculations of ferric iron in such pyroxenes would result in artifically low Fe^{3+} values.

The eclogites from the Basal Gneisses have similar $K_D{}^{gt/cpx}$ values to those from the overlying units, but have lower $X_{Ca}{}^{gt}$ values. Their derived temperatures are thus some 50°C lower. Temperatures derived for inclusions are essentially similar to those for matrix pairs from the same specimen.

The geothermometric equation of Ellis and Green (1979) was derived from experiments performed at temperatures and pressures well above those inferred for the Dalsfjord eclogites and the calculations of temperatures involve some extrapolation from their calibration set. Krogh (1982) has re-evaluated the data of Ellis and Green using experimental results from temperatures down to 600°C. For a linear relationship between $K_D{}^{gt/cpx}$ and $X_{Ca}{}^{gt}$ has derived the relationship:-

$$\mathrm{Ln}K_D{}^{gt/xpx} = \frac{3104X_{Ca}{}^{gt}+2443+10.86P(\mathrm{kbar})}{T(^{o}C)} - 1.5049$$

Isopleths for this equation are subparallel to those in that of Ellis and Green (1979) but for any combination in K_D and $X_{Ca}{}^{gt}$ the derived temperatures are some 50°C lower.

Graham and Powell (1984) have recently empirically calibrated a geothermometer based on Fe-Mg exchange between garnet and hornblende. The thermometer was calibrated against the garnet-clinopyroxene thermometer of Ellis and Green (1979), assuming Fe^{2+} = total Fe in hornblende. Isopleths for $LnK_D{}^{gt/hbl}$ have been constructed for a range of temperatures on figure 4.14, using Graham and Powell's (1984) expression:-

$$LnK_D{}^{gt/hbl} = \frac{2880+3280\ X_{Ca}}{T(^oK)} - 2.426$$

where $K_D{}^{gt/hbl} = (X_{Fe}/X_{Mg})/(X_{Fe}/X_{Mg})$

$K_D{}^{gt/hbl}$ varies from 2.65 to 13.09 with most values between 8.0 and 11.0. Very low K_D's correspond to dark hornblendic inclusions (stage I). The more actinolitic amphiboles (inclusions and matrix)define a trend with K_D's varying from 7.4-10.2 with some scatter up to 13.2 for inclusions. Stage III amphiboles with lower K_D values tend to be those with high Al^{IV} and $(Na + K)^A$ and the most intense pleochroism and may not be in equilibrium with the garnet due to retrogressive decomposition.

The actinolitic amphibole inclusions in D123 (stage I) define temperatures of 511-570°C, while the matrix barroisites with the lowest Al^{IV} and A-site occupancy give temperatures of 533-615°C.

Stage II actinolitic amphiboles in amphibolite D206 give temperatures of 469-552°C, while matrix (stage III) amphiboles with low Al^{IV} and $(Na + K)^A$ give 495-535°C.

Estimates for inclusion-host pairs and matrix pairs are not significantly different. The amphibole-garnet temperatures are, on average, some 70°C lower than those for both stage II and stage III garnet-omphacite pairs from the same specimen. This may be due to differences in amphibole composition between the calibration set of Graham and Powell (1984) and those under study. Making a correction of +70°C to the estimates for the stage I amphiboles (which are very similar in composition to those in D206), it transpires that the transition from stage I to stages II and III involved very little change in temperature.

Compositions of garnet + phengite pairs have been plotted on figure 4.15. The isopleths are constructed from the garnet-phengite geothermometer of Green and Hellman (1982) using their expression for a basaltic bulk composition:-

$$\mathrm{Ln}K_D{}^{gt/ph} = \frac{5170 + 0.036P\ (\mathrm{bars})}{T(^\circ K)} - 4.17$$

where $K_D{}^{gt/ph} = (FeO/MgO)^{gt}/(FeO/MgO)^{ph}$

K_D values range from 4.6 to 9.5, corresponding to temperatures between 622-734°C, averaging 675°C. These values assume that Fe^{2+} = total Fe. Attempts to calculate ferric iron (eg by the method of Laird and Albee, 1981) produced highly variable or even negative values. Whole rock ferric iron in these rocks is fairly low and a large proportion of it is likely to have been taken up as acmite in pyroxene, so that ferric iron in the phengites may be fairly low. However, the phengites usually contain less than 2.5 mass % total iron oxide and if only half of this is ferric, the error in temperature estimates may be up to 100°C.

Estimates of temperature using Green and Hellman's (1982) equation are 70-130°C higher than for co-existing garnet + omphacite pairs at 16 kbars, which is probably the result of uncertainties in ferric iron in the phengites outlined above.

4.01.8 Mineralogical evolution and metamorphic conditions

Figure 4.17 is a petrogenetic grid with reaction curves constraining the equilibrium P-T conditions of the eclogite parageneses. The evolution of the eclogites is summarised below in relation to the four petrographic stages.

Stage I

Evidence for the early evolution of the eclogites is preserved within the garnets in the form of inclusion suites and compositional zoning.

Raheim and Green (1975) have interpreted zoning in some eclogite garnets as resulting from growth zoning during prograde evolution of garnet and omphacite, the changes in composition reflecting changing $K_D^{gt/cpx}$ for Fe/Mg as pressure and temperature changed. However, if the mineral assemblage which is in equilibrium with garnet changes during its growth, this might be expected to have a significant effect on its composition (see the review of Tracy, 1982). At any temperature the K_D for Fe/Mg exchange between garnet and amphibole is much lower than that for garnet and clinopyroxene (compare figures 4.13 and 4.14), so that the Fe/Mg ratio of a garnet co-existing with amphibole will be much lower than that for a garnet + clinopyroxene pair.

As the reservoir for iron and magnesium in the Dalsfjord eclogites is likely to be large in relation to the volume of garnet the observed zoning is unlikely to be a result of simple reaction partitioning (fractionation). Hence, the zoning can be ascribed to the change from a garnet-bearing amphibolite to an eclogite. This is consistent with the steep increase in FeO/MgO ratio between the amphibole and omphacite-bearing zones in eclogite D205 (figures 4.11a and b). The bell-shaped profile for FeO/MgO in garnet amphibolite D206 may either be due to increasing temperatures (as garnet was in equilibrium with amphibole through much of its growth) or to the effect of stage II clinopyroxene increasing the K_D between garnet and the bulk matrix. The latter interpretation is favoured as analogy with the stage I amphiboles of kyanite eclogite D123 suggests that the transition to stages II and III occurred with little or no change in temperature.

The steep decrease in MnO concentrations in all the garnets analysed is attributable to fractionation. Bulk rock Mn is low in the eclogites and as MnO is strongly portioned into garnet the matrix reservoir for Mn would have been rapidly depleted. The local increase in MnO at garnet rims may be a result of decomposition of the cores of outgoing amphiboles which would have been slightly enriched in Mn if volume diffusion in amphiboles was slow (Tracy, 1982).

Temperatures for garnet amphibole pairs in stage I are 540 +/- 30°C, or if adjusted for consistence with garnet-omphacite pairs (see sub-section 4.01.7) may be approximately 70°C higher. Minimum pressures are given by the minimum stability of zoisite + quartz at 3-4 kbars between 550-600°C (Newton, 1965).

The transition from stage I to stage II can be qualitatively modelled with reference to the high-pressure stability limits of amphibole. At low pressures the decomposition of amphiboles has a positive slope in the P-T plane (Ernst, 1966 and Gilbert, 1966). However, at high water vapour pressures (P_{H2O}) amphibole decomposition curves bend sharply backwards to a shallow negative slope (Hill and Boettcher, 1968 and Lambert and Wyllie, 1970) due to the high compressibility of water vapour and the replacement of low density reaction products (pyroxenes and plagioclase) to high density phases (garnet and pyroxenes) (Essene et al, 1970).

Essene et al (op cit) indicated that pure tremolite may be stable up to pressures in excess to 25 kbar P_{H2O} where it breaks down to diopside + enstatite + quartz + vapour or talc + diopside. Natural hornblendes in basaltic systems may decompose over a range of pressures between 10 kbars to 25 kbars in which they may co-exist with garnet and pyroxene (figure 4.16). The shallow slope of the breakdown curve allows the amphibole to eclogite transition to occur up-pressure which is consistent with the Fe-Mg exchange equilibria for the eclogites described here. The compositional trends exhibited by the amphibole inclusions in figures 4.04 and 4.06 indicate that the breakdown of amphibole involved progressive removal of Na and Al to produce a progressively more actinolitic amphibole (in equilibrium with garnet + omphacite + H_2O) by a reaction resembling:-

2 pargasite + n tremolite = n-2 tremolite + diopside + 2 jadeite + 2 garnet + H_2O

The occurence of interbanded garnet amphibolites and eclogites, as near Solvik (samples D205 and D206), which are apparently in textural equilibrium suggests that these lithologies are co-facial. The amphibolites are consierably less sodic and silicic than the eclogites, producing a lower normative albite content. It seems possible that if the albit component of the rock (either in solution in amphiobole or as a separate phase) has a relatively low concentration, tremolitic amphibole will remain stable, whereas at higher normative albite contents omphacite will be stable relative to amphibole. This is supported by the results of Essene et al (1970), whose experiments on an albite-rich quartz tholeiite cxontained significant amounts of clinopyroxene and garnet well below maximum pressure for stability of tremolitic amphibole (figure 4.16). As discussed below, P_{H2O} was probably equal to lithostatic pressure in this area.

Stages II and III

Garnet-clinopyroxene geothermometry (Ellis and Green, 1979) gives temperatures of 597 +/- 30°C for eclogites from the Gjorlanger and Flekke Units and 566 +/- 24°C for those from the Basal Gneisses (calculated at 16 kbars) for both stages II and III.

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The maximum pressure for stage II is constrained by the minimum stability of almandine-rich garnet and rutile, which is 10.6 kbars at 600°C (Bohlen et al, 1983).

Minimum pressures for stage III are given by co-existing omphacite ($DiJd_{37-40}$) and quartz in quartz-bearing eclogites (12.5 kbars at 600°C, from Holland, 1980, 1983). Co-existence of kyanite and omphacite ($DiJd_{40}$ for most reliable analyses) give a minimum pressure of 16.2 kbars at 600°C (Holland, 1979).

The presence of omphacite-filled veins in the coronites at Solvik (see sub-sections 2.03.2 and 4.02.1) indicates that fluid pressure was equal to lithostatic pressure during eclogite formation. Maximum temperatures for eclogite stability are, therefore, given by the solidus for a fluid saturated eclogite. The solidus curve of Hill and Boettcher (1968) for a water saturated eclogite has an intersection with the curve for $DiJd_{40}$ + kyanite + paragonite + H_2O at approximately 635°C and 16 kbar (figure 4.17). The solidus is not significantly affected by high CO_2 concentrations above 15 kbars. Minimum temperatures for the eclogites are by the occurence of zoisite + kyanite + quartz segregations. The reaction curve for Lawsonite = zoisite + kyanite + quartz + H_2O (Newton and Kennedy, 1963 and Perkins et al, 1980) passes through 550°C at 16 kbars. These maximum and minimum temperatures are consistent with the garnet-clinopyroxene geothermometry.

The presence of glaucophane as a stage II mineral only in the Basal Gneiss eclogites is interesting in the light of their rather lower garnet-clinopyroxene temperatures. Omphacite inclusions in a garnet co-existing with glaucophane give temperatures of 566°C and 547°C (Ellis and Green, 1979). However, the work of Maresch (1977) indicates that the maximum stability for glaucophane is about 525°C above 10 kbars. If the recalibration of Ellis and Green's (1979) geothermometer by Krogh (1982) is employed, temperatures of 504°C and 486°C are derived, which are more consistent with the results of Maresch. However, this calibration often gives temperatures in the stability field of lawsonite at 16 kbars for the Gjörlanger Unit eclogite and may not be applicable here. The breakdown of glaucophane is probably divariant in P-T space and the amphibole may be stabilised at greater temperatures by tschermak's exchange with phengite-like micas (Koons, 1982). In the absence of analyses of these ampiboles, this is difficult to confirm. It is noteworthy that in the Förde area, some 40 km to the north-east, the Naustdal eclogite contains barroisitic amphiboles and gives temperatures of 630 +/- 35°C at 15.0 +/- 2.5 kbars, while glaucophane-bearing eclogites at Kvineset give lower temperatures similar to those for the Basal Gneiss eclogites at 540 +/- 35°C, 12.5 +/- 2.5 kbars (Krogh, 1980). Glaucophanitic amphibole thus seems to be unstable above about 550°C in these rocks.

Minimum pressures for the Basal Gneiss eclogites are given by the minimum stability of omphacite ($DiJd_{42}$) which is 11.3 kbar at 550°C. Extrapolation of the minimum stability of phengite (Si^{4+} = 6.8) + quartz after Velde (1967) gives approximately 9 kbars at 550°C.

Stage IV

The presence of dipside + plagioclase symplectites after omphacite indicates a drop in pressure. Petrographic observations indicate that symplectite formation is most effective in the presence of quartz, as quartz-free eclogites are often less symplectised. Symplectite formation in quartz-free eclogites may require intergranular transport of silica in a fluid phase, or oxidation of the omphacite (Mysen and Griffin, 1973). The presence of small patches of quartz along grain boundaries in some otherwise quartz-free eclogites supports the former mechanism.

A quartz-bearing eclogite from the Basal Gneisses contained symplectites with sodic augite ($DiJd_{14.3}$ - Table 4.1., D116). The calculated Fe^{3+}/Fe^{2+} ratio of the symplectite pyroxene is not significantly higher than that for stage III omphacites. The curve for the minimum stability of omphacite ($DiJd_{15}$) is plotted on figure 4.17. At a maximum temperature of 630°C constrained by the 'wet' eclogite solidus (Hill and Boettcher, 1968) a maximum pressure for symplectite formation of 10 kbar is derived.

In the kyanite eclogites the marginal breakdown of kyanite to paragonite in the presence of omphacite implies a drop in pressure below the stability curve for $DiJd_{40}$ + kyanite + paragonite + H_2O at about 16 kbar. The rims of anorthitic plagioclase around zoisite against quartz in zoisite + kyanite + quartz segregations suggests the reaction:-

zoisite + kyanite + quartz = anorthite + H_2O

The presence of anorthite rather than margarite + quartz implies temperatures in excess of 625^oC, but the upper temperature limit is constrained by the 'wet' eclogite solidus at 660^oC. The reaction curve of Jenkins (1984) gives maximum pressures of 8.2 to 9.0 kilobars at these temperatures.

Phengite grains in some quartz eclogites have fine rims of K-feldspar and extremely fine needles of what may be sillimanite (figure 2.16). If this identification is correct it indicates the reaction:-

muscovite + quartz = K-feldspar + sillimanite + H_2O

indicating temperatures were in excess of 650^oC at 7 kbars (Evans, 1965). However, the presence of a significant celadonite content in the mica may shift the reaction to lower temperatures (Velde, 1965). The breakdown of kyanite to fibrolitic sillimanite (figure 2.47) gives temperatures greater than 517 +/- 25^oC at pressures below 4 to 7 kbars (Robie and Hemingway, 1984).

In summary, a P-T path for the eclogites can be constructed (figure 4.17). An early amphibolite-facies assemblage (stage I) at approximately 550-600°C and probably below 10 kbars was compressed, with only a minor temperature increase, to pressures above 16 kbars at 597 +/- 30°C (stages II and III). This was followed by rapid decompression with little temperature change (possibly a small increase) to pressures below 7 kbars, prior to ultimate cooling to surface conditions.

The P-T path for the Basal Gneiss eclogites is less well resolved, but a peak temperature of 566 +/- 24°C at pressure in excess of 12.5 kbars was followed by steep decompression to less than 10 kbars. No pre-eclogitic (stage I) evolution can be observed.

SECTION 4.02 Corona-textured metabasic rocks

4.02.1 Summary of petrography

The lithologies described here are the meta-anorthositic troctolites or gabbros of the Flekke unit (Sections 2.03.2, 3.04.4 and Enclosure 1). They vary from virtually unaltered igneous assemblages to true eclogites. Texturally they are unfoliated or have only very weak or localised planar fabrics. As a result evidence for original igneous textures has been preserved even where the most complete mineralogical transformations have occurred.

Little altered cumulate igneous anorthositic troctolites were described in Section 2.03.2. The igneous assemblage is plagioclase + olivine + hypersthene + augite +/- spinel +/- opaque oxide (Figure 2.09). Sub-solidus recystalisation has been restricted to the development of 2-pyroxene + spinel symplectites between olivine and plagioclase (Figure 2.10), clouding of plagioclase by zoisite and growth of very narrow garnet coronas at the contacts between plagioclase and mafic minerals. The garnet postdates the zoisite.

Relics of igneous clinopyroxenes are found in the coronites at Balsarvika. The pyroxenes (Figure 2.26) lie within robust shells of red garnet up to 5 mm thick and tend to be single crystals up to 3 cm across. They are heavily clouded with extremely fine specks of Fe-Ti oxide in their cores. Towards the garnet shell the oxide specks coarsen and become less common so that the pyroxene is clearer. Some of the oxides are altered to rutile and sub-grain boundaries form in the pyroxene. Immediately adjacent to the garnet shell the pyroxene clears completely and consists of a granoblastic mosaic of pale-green omphacite intergrown with clots of rutile.

The white material which surrounds the garnet shells (Figure 2.22) consists of intimate, very fine-grained intergrowths of jadeite + zoisite + paragonite (Figure 2.25) which has probably replaced plagioclase. The garnet shells are crowded with small inclusions of jadeite or colourless omphacite or, more rarely, paragonite. Where augites are still preserved the garnet extends right up to the pyroxene rim.

Where augites have been entirely replaced by omphacite + rutile the inner half of the garnet shell is free of jadeite inclusions, although coarser green omphacite embays, or is included by, the garnet here. This indicates that the garnet has overgrown both the jadeite + zoisite + paragonite masses and the clinpyroxene cores to the coronas.

Patches of essentially bimineralic eclogite occur within the jadeite-bearing coronites (Figure 2.24). These are, in fact, coronites, but instead of the jadeite + zoisite + paragonite intergrowths there are masses of pale-green or colourless omphacite, individual grains of which have a rather "feathery" internal structure. They are assumed to have replaced plagioclase. The garnet corona shells consist of masses of interlocking euhedral garnets whose cores are crowded with small colourless omphacite inclusions, but whose rims are clear. Rare inclusions of amphibole or plagioclase occur also. The inclusion-rich garnets extend right up to the inner garnet shell border with omphacites forming the cores to the coronas. These omphacites are darker green than the feathery variety, are often colour zoned with paler cores and coexist with rutile.

The coronites of Aurevagen, near Hellevik, are also true eclogites which are petrographically similar to the bimineralic type at Balsarvika. Grass-green granoblastic omphacites and rutile occupy corona cores. Garnet shells are granular, with concentrations of colourless omphacite inclusions in garnet of the outer part of the shell. Outside the garnet shell is a granular mosaic of very fine-grained colourless omphacite intergrown with kyanite (Figure 2.14). Discreet euhedra of garnet are common in these areas. They are also full of omphacite inclusions. Rare blades of zoisite are occasionally intergrown with omphacite and kyanite.

No traces of igneous olivine or orthopyroxene have been found in the coronites except for the little altered igneous varieties themselves. However, their previous existence is indicated by the assemblage omphacite + talc +/- Ca-amphibole in the cores of some coronas.

Petrographic evidence shows that the garnet coronas have formed by reactions between titaniferous augite (or olivine or orthopyroxene) and plagioclase. During corona growth augite was transformed to omphacite and plagioclase either broke down to jadeite + zoisite + paragonite or omphacite + kyanite +/- zoisite. Garnet has generally encroached across the areas previously containing plagioclase (as indicated by the abundant jadeite or colourless omphacite inclusions) but also locally encroached across the igneous augite.

The chemistry of the coronite minerals is described below.

4.02.2 Pyroxenes

Compositions of pyroxenes are set out in table 4.6. Clinopyroxene compositions have been plotted on Figure 4.18.

Pyroxenes in the anorthositic troctolites are augites and hypersthenes. Their compositions are plotted on a pyroxene quadrilateral diagram (Figure 4.19) using the projection scheme of Lindsley (1983) to remove the effects of non-quadrilateral components. Igneous hypersthenes are zoned with cores relatively enriched in Ca and Ti. Al and Fe/Fe + Mg show no significant variation. Igneous augites are also zoned with cores less calcic than rims, but fairly constant Al, Ti and Fe/Fe + Mg. Hypersthenes in 2-pyroxene + spinel + symplectites are low in Ti, Al and Ca relative to the igneous type. Symplectic augites are similarly depleted in Ti and Al relative to igneous augites but enriched in Ca. Fe/Fe + Mg in the symplectic pyroxenes is essentially the same as that in the igneous varieties. Jadeite concentrations in the igneous clinopyroxenes are very low (0-1.5 mol %) and jadeite/tschermak's ratios all fall outside the eclogite field of White (1964) (Figure 4.18a).

In the jadeite-bearing coronites at Balsarvika (specimen D194) the relict igneous clinopyroxenes (Figure 2.28) show a trend of increasing jadeite component. The least altered pyroxenes are sodic augites with high Ti and Jd/Tsch less than 0.5. As the pyroxenes become clearer of oxide clouding and greener in colour Ti, Fe, Mg and Ca decrease and Si, Al^{VI} and Na increase, giving an increasing Jd/Tsch ratio passing into the eclogite field of White (1964) in Figure 4.18a. The Na/Ca ratio shows its strongest increase where tetrahedral Al is at its lowest, adjacent to the garnet shell. Jadeite concentrations reach nearly 30 mol % in the most altered pyroxene. Acmite shows a systematic decrease as the pyroxene becomes more omphacitic (Figure 4.18b).

Clinopyroxenes in corona cores without relict augite (specimen 6-5.5) are all omphacites with very low A^{IV} (0.032-0.011 cations per formula unit) and jadeite concentrations of 34-46 mol %. Jadeite content does not vary in a systematic pattern within such corona cores.

Clinopyroxenes in the white material around the garnet coronas are nearly pure jaeite (specimen D187, table 4.6) with a composition $Di_{0.04}$ $Jd_{0.96}$ $Acm_{0.00}$. No quartz has been observed coexisting with the jadeite.

Inclusions in the garnet coronas vary from nearly pure jadeite (92% Jd) to omphacite. These omphacites are significantly more jadeitic than those in the corona cores (specimen 6-5.5-1; Di_{37-43} Jd_{50-52} Acm_{2-7}) and have slightly higher Fe/Fe + Mg ratios.

In the bimineralic eclogites at Balsarvika (specimen D192) the green clinpyroxenes in corona cores are omphacites. They are often colour zoned with pale green cores and darker green rims, the rims being relatively enriched in acmite and depleted in jadeite. They fall in the compositional range Di_{49-57} Jd_{35-47} $Acm_{1.5-7.0}$
The Si contents are very high (1.992-2.004 cations per formula unit) allowing only very low tschermaks contents.

The feathery, pale green to colourless clinopyroxenes in the bimineralic eclogites are also omphacites with jadeite concentrations falling in the higher end of the range for those in corona cores. Compositions lie in the range

$Di_{47.51}$ $Jd_{45.49}$ $Acm_{2.0-5.8}$
Tschermak's molecule is very low with Si in the range 1.987-2.000 cations per formula unit.

Clinopyroxene inclusions in the bimineralic eclogites are omphacites with slightly higher jadeite concentrations than in matrix grains ($Di_{43-49}Jd_{45-50}$ $Acm_{1.5-32}$). One apparently sound analysis was less jadeitic with the composition Di_{53} Jd_{36} Acm_{8}. As in the jadeite-bearing coronites, Fe/Fe + Mg is higher in inclusions than in matrix omphacites.

Omphacites filling veins in the bimineralic eclogites (specimen D208) have the composition $Di_{50}Jd_{41}Acm_{7}$ and rather lower Fe/Fe + Mg ratios than in the host rock. Symplectite clinpyroxenes in the veins have the compositional range Di_{69-78} Jd_{14-22} Acm_{2-5}.

Clinopyroxenes in the kyanite-bearing coronites from Aurevagen (specimen D132) are all omphacites which show similar variation to those in the bimineralic eclogites at Balsarvika. Green, coarser grained omphacites in the cores of coronas have the compositional range Di_{51-55} $Jd_{39-46}Acm_{1.9-5.0}$. Some grains are slightly zoned with more omphacitic cores. Finer-grained paler coloured ophacites surrounding the garnet coronas have similar compositions ($Di_{50}Jd_{43-47}$ $Acm_{0.5-4.3}$) to those with the coronas, but inclusions in garnets are significantly more jadeitic ($Di_{40-42}Jd_{51-53}Acm_{1.1-1.6}$)

4.02.3 Garnet

Garnet analyses are presented in table 4.7 and compositions are plotted on Figure 4.20.

In the jadeite-bearing coronites the garnet coronas are massive. The example studied here (6-5.5) rims granoblastic omphacite. It has an outer zone with abundant omphacite (+/- albite) inclusions and an inner zone which is free of inclusions. In Figure 4.20a the inclusion-free zone is obviously enriched in grossular with respect to the inclusion-rich zone. A microprobe traverse across the garnet shell (Figure 4.21) shows that the inclusion-free zone is richer in Ca than inclusion-rich outer zone, while Fe shows a complimentary outward increase in Fe. Fe/Mg varies sympathetically with Ca but increases at the rim, which is amphibolitised.

A traverse across a garnet corona in kyanite-bearing coronite D132 (Figure 4.22) shows a trend of increasing grossular outwards, with fairly large fluctuations. This zoning pattern is the opposite to that described above. Both Fe and Mg decrease outwards, but Fe/Mg increases outwards and has fluctuations in phase with those of Ca. Analyses of discreet garnets by the outer garnet shell rim and in the pale omphacite matrix (Figure 4.20b) show a strong enrichment of grossular in their cores. The cores of this discreet garnets are crowded with omphacite inclusions. Fe/Mg decreases rimwards.

Garnet coronas in bimineralic coronite D192 consist of masses of discreet garnets. Individual garnets show a trend decrease of grossular near the rim, the sharp peak in the core corresponding to a zone of omphacite inclusions (Figures 4.20c and 4.23). However, two garnets containing hornblende inclusions have cores enriched in almandine relative to rims (Figure 4.20b), a trend seen in the eclogites (Section 4.01).

4.02.4 Zoisite

Zoisite analyses are presented in table 4.8. Zoisites in jadeite-bearing coronites are almost pure zoisite and end-member compositions with up to 0.050 cations per formula unit Fe^{3+}. A zoisite from a kyanite-bearing coronite, which coexists with omphacite, is similar in composition to those in the eclogites with 0.123 cat/fu Fe^{3+}.

4.02.5 Paragonite

Analyses of paragonites appear in table 4.9. They contain only small amounts of muscovite and margarite in solid solution

(K 0.060-0.095 cat/fu, Ca 0.049-0.059 cat/fu).

4.02.6 Other phases

Analyses of plagioclases and spinels are given in table 4.10. Olivine and plagioclase analyses from an anorthositic troctolite have been presented in table 3.16. The olivines have the composition Fo_{70} and plagioclases An_{60}. Spinels from the two-pyroxene + spinel symplectites are pleonastes with $Fe^{2}{}_{+}/Fe^{2+}$ + Mg close to 0.50. Spinels of apparently igneous origin are also pleonastes with slightly higher Fe^{2+}/Fe^{2+} + Mg and Cr than the symplectic variety.

Rare plagioclase inclusions in garnet coronas are nearly pure albites (Ab_{96-99}). Plagioclase coexisting with clinopyroxene in symplectised vein omphacites at Balsarvika is of albite-oligoclase composition (Ab_{89-92}).

4.02.7 Relationship of parageneses to bulk rock chemistry

The mineral assemblages of the coronites show some systematic variations with bulk-rock chemistry, as shown on Figure 4.24. The meta-troctolites which retain igneous assemblages are characterised by relatively high MgO/MgO + FeO^T (FeO^T = total iron oxides) and a low normative colour index (NCI). Bimineralic eclogites have low MgO/MgO + FeO^T and a relatively high NCI. Jadeite-bearing coronites have high normative nepheline values, while kyanite-rich varieties have very low or zero normative nepheline, with lower NCI and higher normative Ab/Ab + An than the bimineralic eclogite type. Amphibole-rich varieties have high MgO/MgO + FeO^T and moderate to high NCI. In addition, kyanite-bearing coronites have higher normative plagioclase contents than the bimineralic type.

Bulk-rock and mineral relationships can also be seen on an ACF diagram (Figure 4.25). Although this is only a crude paragenetic diagram, it can be seen that bimineralic eclogites have compositions lying along the join garnet-clinopyroxene, while kyanite-bearing coronites lie within the three-phase field defined by garnet, clinopyroxene and kyanite.

4.02.8 Evolution of the coronites

Petrographic observations and bulk-rock chemistry show that the coronites developed from coarse-grained anorthositic troctolites and gabbros, probably of cumulate type. The most "primitive" variety, the anorthositic troctolites, have retained little altered igneous mineralogy. Although their primary precipitate minerals were olivine and plagioclase, orthocumulus hypersthene, augite and spinel seem to have been the phases present at solidus conditions, along with plagioclase. The common coexistence of unaltered olivine and plagioclase indicates that they may also have been a stable pair on the solidus. Hence final solidification of the troctolites lay close to the intersection of the solidus and the reaction curve for

orthopyroxene + clinopyroxene + spinel = olivine + plagioclase

The results of Green and Ringwood (1967) place this point at approximately 1200°C, 9 kbars for an olivine tholeiite.

Variations in pyroxene chemistry give some clues to the sub-solidus evolution of these troctolites. Referring again to Figure 4.19, pyroxene compositions are superimposed upon isotherms for the two-pyroxene solvus (at 10 kbars) of Lindsley, 1983. The clinopyroxene limb has a trend of decreasing temperature (decreasing X_{En}) from 900°C for rims of igneous pyroxenes to 710°C for symplectic pyroxenes (errors are probably at least +/- 50°C - Lindsley, 1983). On the orthopyroxene limb igneous cores lie on the 1050°C isotherm and the symplectite type gives similar temperatures to coexisting clinopyroxenes at 680°C. Rims to igneous orthopyroxenes give significantly lower temperatures than coexisting clinopyroxenes (650°C). This is a well-known feature of granulite-facies orthopyroxenes, but the reason for it is not well understood (Lindsley, 1983).

The observed variations in aluminium in the pyroxenes give some indications as to their early P-T history. Gasparik

(1984) has recently calibrated isopleths of tschermak's content of coexisting pyroxenes in the system $CaO-MgO-Al_2O_3-SiO_2$ (CMAS). While the CMAS system is not directly applicable to the troctolites, some useful inferences can be made. In the anorthite + forsterite field tschermak's (X_{TS}) isopleths are very steep in the P-T plane (Figure 4.26) for both ortho- and clinopyroxenes. In the spinel + anorthite field tschermaks isopleths are slightly less steep. For each type of pyroxene rims have only slightly less X_{TS} than cores, but symplectites have significantly less X_{TS} than igneous grains. Assuming that the igneous grains lay just within the anorthite + forsterite field and the symplectites in the spinel + anorthite field the variations is consistent with virtually isobaric cooling as shown schematically on Figure 4.26. Significant pressure increase should have produced an increase in X_{TS}. Decompression would have made crossing the An + Fo = Sp + An boundary less likely, but cannot be entirely ruled out.

The cores of the igneous pyroxenes must have suffered some re-equilibration during cooling from igneous conditions as the maximum recorded temperatures are about 150°C lower than the solidus of Green and Ringwood (1967). A reconstructed P-T path is given on Figure 4.26. The occurrence of tiny garnets between the pyroxenes and plagioclase implies a late pressure increase, but the effect of garnet growth on pyroxene composition is uncertain as fine rims of amphibole lie between the garnets and the pyroxenes.

The next observed stage in the formation of the coronites is the breakdown of augite and plagioclase seen in the jadeite bearing rocks at Balsarvika.

The nepheline normative nature of these rocks and the lack of quartz coexisting with jadeite suggests the reaction
nepheline + albite = 2 jadeite.
This can be combined with two other reactions;
4 anorthite + H_2O = 2 zoisite + kyanite + quartz and
jadeite + kyanite + H_2O = paragonite.
which model the breakdown of plagioclase to the observed assemblage jadeite + zoisite + paragonite. The normative nepheline content is too small to entirely consume all the silica generated in this reaction so that some free SiO_2 must have been released. The presence of paragonite rather than kyanite implies that the plagioclase decomposition took place at pressures below the equilibrium curve for
paragonite + jadeite + kyanite + H_2O
at 24-25.5 kbars, 600^oC (Holland, 1979).

The augites show a trend of decreasing tschermak's molecule and increasing jadeite, which occurs on the periphery of the original augites next to the garnet corona (Figure 2.26). Garnet can form from tschermak's molecule bythe reactions;

$CaMgSi_2O_6$+$MgAl_2SiO_6$ = $CaMg_2Al_2Si_3O_{12}$, and
diops Mg-tsch garnet
$3CaAl_2SiO_6$+$2SiO_2$ = $Ca_2Al_2Si_3O_{12}$+$2Al_2SiO_5$
Ca-tsch garnet Ky

In the latter reaction silica may have been provided by plagioclase breakdown, while the kyanite generated would react with jadeite to produce paragonite, which occurs as inclusions in the garnet shells. Breakdown of Ca-tschermak's may explain the slightly higher Ca concentrations in the inner part of the garnet shells.

The increasing jadeite content of the augites can be explained by movement of a jadeite component across the corona, giving the reaction
jadeite + diopside = omphacite
To avoid significant volume increase of the developing omphacite there must have been an opposing transfer of diopside component out of the augite, to produce more garnet;
diopside + anorthite = garnet + SiO_2
(petrographic evidence suggests that the pyroxenes did not increase their size during their transformation).

The rutile associated with the omphacite was probably produced by reaction of the exsolved ilmenite with the augite;
1 tschermak's + 2 ilmenite + 2 SiO_2 = 1 garnet + 2 rutile
The growth of this type of corona is illustrated

schematically in Figure 4.27. Transfer of mass across the corona requires diffusion of Na^+, Ca^{2+}, Mg^{2+}, Fe^{2+}, Al^{3+} and Si^{4+} on a scale of a few millimeters. The lack of free SiO_2 in the rocks requires either that the balance of silica producing and consuming reactions used up all the available silica or that there was a net loss of silica from the system. The latter might be expected to have produced quartz-veining, which is absent within the coronites.

The kyanite-bearing coronites from Aurevagen have low values of normative nepheline. In areas where the igneous phase is assumed to have been plagioclase, omphacite occurs as inclusions in garnet and kyanite. True jadeite is absent and zoisite is rare.

These features imply that the earliest reaction during the breakdown of plagioclase was the formation of omphacite rather than jadeite. This required the input of a diopside component, which may have been partly provided by breakdown of augite, but in view of the troctolitic normative composition of these rocks it was probably largely contributed by the breakdown of olivine. Early breakdown of olivine by the reaction

Olivine + plagioclase = aluminous pyroxenes + spinel

has not been directly observed in this type of coronite, but its occurrence seems likely in view of the presence of 2-pyroxene + spinel symplectites in the troctolitic anorthosites

discussed above. Further reactions such as
6 orthopyroxene + anorthite + spinel = diopside + garnet
4 orthopyroxene + anorthite = diopside + garnet + quartz
would generate the necessary diopside to produce omphacite after plagioclase and simultaneously begin to form the garnet corona shell. Transfer of jadeite component across the corona would have produced omphacite within the corona core, while excess silica generated by the reactions mentioned above was possibly consumed by breakdown of remaining enstatite and olvine;
6 enstatite + 2 SiO_2 + 2 H_2O = talc
3 olivine + 5 SiO_2 + 2 H_2O = talc
tremolite = 3 enstatite + 2 diopside + SiO_2 + H_2O

These reactions are nearly all mutually interdependent and therefore probably occurred simultaneously by transfer across the original plagioclase/mafic grain interface. The lack of free quartz in the plagioclase sites and the silica-consuming reactions inferred for the corona core require a net tansfer of silica across the corona. Hence entry of silica into solution in a fluid (aqueous?) phase was probably necessary to trigger the reactions.

Formation of highly calcic garnets and kyanite (with omphacite inclusions) indicates the breakdown of the remaining anorthite content of the plagioclase, according to the reaction
anorthite = Ca-garnet + kyanite + SiO_2

The sharp decrease in Ca near the rims of these garnets (table 4.7, Figures 4.20b and 4.22) may be due to a reduction in Ca supply when all the plagioclase was used up. The omphacite inclusions are more enstatite-rich than matrix grains and the low-Ca rims to the garnets may be partly a result of consumption of this enstatite.

The omphacite inclusions are also significantly more jadeitic than matrix grains (Figure 4.18), exceeding $DiJd_{50}$ in some cases. Thus a decrease in jadeite content with time seems to have occurred, probably as a result of equilibration of pyroxene compositions across the corona. Due to the very high equilibration pressures required for the coexistence of kyanite with $DiJd_{50}$ omphacites in the presence of water vapour (nearly 20 kbars at 600°C - Holland, 1979a) it seems unlikely that kyanite had developed until the Ca-rich garnets grew. The appearance of kyanite required that either some paragonite should have formed or the jadeite content of the omphacites be reduced (it will be argued in Section 4.03 that pressures did not exceed those necessary for the breakdown of albite to pure jadeite + quartz).

The bimineralic eclogites show very similar features to the kyanite-bearing variety. Their lack of kyanite can be attributed to their higher normative colour index, as less anorthite component would have been available for breakdown to Ca-garnet + kyanite + SiO_2. The other variable jadeite contents of the omphacites in sample D192 (table 4.6) can be attributed to poor equilibration. The rather high jadeite contents of some grains (up to $DiJd_{49}$) do not necessarily imply extreme pressures as no coexisting quartz has been observed.

The omphacite-filled veins within the patches of bimineralic coronites (Figure 2.24) are strong evidence for high fluid pressures during the metamorphism which produced the coronites and eclogites. The abundance of hydrous phases in these rocks indicates that this was dominantly water vapour. For the veins to be held open long enough for omphacite to grow P_{H2O} must have equalled lithostatic pressure (P_{total}) and permeability must have been great enough for fluids and associated dissolved ions to reach the fractures and maintain them open. The implied enhancement of ionic mobility by such a fluid is consistent with the rather coarse grain size of the coronas and the presence of garnets up to a meter in diameter in the same area (see Kolderup and Rosenqvist, 1950). Similar veins are known from eclogites in the Förde area (Krogh, 1980) and the Austrian Alps (Holland, 1979a).

There may be one of two explanations as to why the veins stop at the edges of the bimineralic coronites where they meeting the jadeite-bearing variety:-

1) Transformations to eclogite facies assemblages involved

greater volume loss in the bimineralic type than in the jadeite-bearing type, so that fracturing was caused by differential contraction.

2) Fracturing was caused by very high pore fluid-pressures in a differential stress-field, and only the parts of the fractures traversing bimineralic coronites suffered omphacite infilling. In the jadeite-bearing coronites the plagioclase breakdown products were in equilibrium with their externally imposed conditions (P_{total}, P_{H2O}, T). However, due to their different composition, the plagioclase sites in the bimineralic coronites required considerable input of externally supplied cations in order to produce an equilibrium mineral assemblage. If the dissolved cation content of the fluid was controlled by chemical potentials driven by such factors, only the parts of the fractures transecting developing bimineralic coronites would have been filled with fluids capable of precipitating the vein-fill. Outside these domains the fractures would have simply closed up as the fluid escaped.

The observed breakdown of the vein omphacite to more diopsidic pyroxene and plagioclase in the absence of free quartz would have required a source of silica, as the more diopsidic pyroxene is, if anything, more reduced than the omphacites (table 4.6, D208). Minute interstitial blebs of quartz are found by the vein margin, perhaps indicating high mobility of silica.

4.02.9 Conditions of metamorphism

Metamorphic conditions are summarised on a petrogenetic grid in Figure 4.29.

The early P-T history of the coronas has been discussed above. In summary, following cooling from solidus conditions (approximately 1200°C, 9 kbars) the development of 2 pyroxene + spinel symplectites indicate attainment of intermediate-pressure granulite facies conditions at about 700°C by virtually isobaric cooling.

Maximum temperature for the formation of the coronites is given by the solidus for "wet" eclogite. As maximum pressures are constrained by the absence of any jadeite + quartz assemblages in the gneiss (see Section 4.03) maximum conditions for coronite formation are the intersection of the curve for albite + jadeite + quartz (Holland, 1980) and the solidus (Hill and Boettcher, 1968) at 640°C, 17.3 kbars. Minimum pressures are given by the stability of jadeite in the absence of quartz. The equilibrium curve for nepheline + albite + jadeite passes through 600°C at 9.3 kbars (Gasparik, 1985).

Somewhat higher minimum pressures are indicated by the coexistence of omphacite ($DiJd_{43-47}$) and kyanite. The curve for paragonite + kyanite + $DiJd_{45}$ + H_2O lies at 18 kbars, 600°C with a shallow negative slope. This pressure seems rather high in view of the argument outlined above. Possible explanations are that errors in Fe^{3+} calculation are causing overestimation of jadeite in the omphacites, or that the omphacites were not in equilibrium with the kyanites, the latter seeming to have overgrown the omphacite rather than to have grown alongside it.

Compositions of garnet/clinopyroxene pairs for two coronites from Balsarvika are plotted on Figure 4.28, against the isotherms of Ellis and Green (1979). Despite the obvious gross disequilibrium textures of the coronites and the chemical variability of the minerals the apparent temperatures are remarkably uniform, lying in the range 553 +/- 25°C (16 kbars), with most points lying between 500-550°C. These temperatures are significantly lower than those derived from the eclogites in the Gjörlanger and Flekke units (597 +/- 30°C). The coronites and eclogites have a similar range of $K_D^{Gt/cpx}$ but the coronites have rather lower X_{Ca}^{Gt} values. This feature has also been noted by Messiga et al (1983) for similar rocks in the western Ligurian ophiolite (Voltri Group), where Fe-Ti rich metagabbroic eclogites have similar K_D's but lower grossular contents in garnets than Mg-rich eclogites. The bimineralic coronites described here are more ferroan than the eclogites selected for geothermometry

($MgO/MgO + FeO_{total}$ = 0.25 and 0.32 respectively. As noted by Messiga et al (1979) the calibration of Ellis and Green was based on experiments run at higher temperatures than ophiolitic (or "crustal") eclogites. It is possible that at lower temperatures the K_D is less sensitive to grossular in garnet, or that some bulk-rock effect is operative. If the temperature difference is real the Fe/Mg exchange system must have closed some 50°C lower in the coronites than the eclogites.

In summary the conditions of eclogite-facies metamorphism deduced for the coronites are broadly consistent with those for the nearby eclogites although they are less well constrained and prone to problems of determining what are equilibrium assemblages. They are more useful in that plagioclase breakdown, garnet growth and increasing jadeite content of pyroxenes clearly demonstrates an up-pressure evolution of originally plagioclase-bearing crustal rocks.

The link between the granulite-facies stage and the beginning of compression is unclear from these rocks alone. The relics of amphibolite-facies assemblages in the eclogites clearly differentiates them from the coronites. However, small inclusions of amphibole are found in garnets in one of the coronites. The enclosing garnet has zoning similar to that found in the eclogites proper with increasing pyrope rimwards, which contrasts with the grossular dominated zoning elsewhere in the coronites (Figure 4.20). This shows the importance of the pre-existing mineralogy of the metabasites in controlling the zoning and provides a link between the two lithologies. It would seem that the eclogites have previously been basaltic rocks which were prevasively amphibolitised prior to ultimate high-pressure metamorphism while the coronites escaped pervasive amphibolitisation.

Temperatures of 550-600°C have been inferred for eclogite Stage I. Hence further cooling from granulite-facies (about 700°C) to amphibolite-facies conditions must have occurred prior to the crystallisation of the eclogites. Whether this cooling was isobaric or involved pressure variation is not clear from mineralogical considerations alone.

4.02.10 Comparison with other coronites in eclogite-bearing terrains

Coronitic metadolerites are well known to the north of Nordfjord in Western Norway (Gjelsvik, 1950; Griffin and Raheim, 1973; Törudbakken and Raheim, 1981 and Mörk, 1982) and coronitic meta-anorthosites occur in the Bergen Arcs (Griffin, 1972; Austrheim and Griffin, 1982). In both areas these rocks are characterised by multishell coronas; for instance olivine is rimmed successively by orthopyroxene, clinopyroxene, garnet and pyroxene + spinel against plagioclase. Griffin (1972) and Griffin and Raheim (1973) attributed these to a continuous process of cooling and compression, followed by strong decompression.

The Dalsfjord coronites show some features in common with the metadolerites and anorthosites. Clinopyroxenes show progressive increase in jadeite and loss of tschermak's molecule, followed by exsolution of plagioclase and decrease in jadeite (Figure 4.18). Garnets are zoned with outward increase in Ca (Figure 4.22). Early reaction of olivine and plagioclase to pyroxenes + spinel is observed in all these coronites, but is only found in the least altered types in the Dalsfjord area.

The main differences are that in the more advanced stages of corona formation no olivine or orthopyroxene is found (although their chemistry suggests that it was originally abundant), but omphacite is commonly associated with talc or Ca-amphibole in corona cores. Furthermore where coronas are well developed plagioclase is entirely absent except as a retrogressive mineral. In the other Norwegian coronites plagioclase is a persistent mineral until late in the corona development sequence, whereas in the Dalsfjord rocks petrographic evidence suggests that plagioclase breaks down very early on. Plagioclase is usually replaced by Na-clinopyroxene +/- zoisite +/- kyanite +/- paragonite with a rim of garnet, unlike the metadolerites described by Griffin and Raheim (1973) and Mörk (1982) where it is gradually encroached upon by garnet. Where garnet encroaches upon the plagioclase sites in the Dalsfjord coronites it overgrows omphacite or jadeite.

The Dalsfjord coronites share some features in common with those described from the Alps. Miller (1970) has described kyanite-bearing eclogites from the Oetztal Alps, Austria with coronitic texture. Both plagioclase and mafic sites now contain omphacite and omphacite commonly occurs as inclusions in garnets. In metagabbros feldspars are replaced by a "saussurite" of kyanite + zoisite +/- quartz +/- clinopyroxene. Fe-Ti metagabbros from Western Liguria have chloromelanitic omphacite + rutile replacing igneous augite with thick coronas of garnet and fine grained omphacite + zoisite + paragonite replaces plagioclase. Olivine and orthopyroxene are replaced by omphacite + talc +/- Ca-amphibole (Ernst, 1976; Messiga et al, 1983). Many of the Alpine eclogites are thought to have formed at relatively low temperatures (for instance the Voltri eclogites of Liguria give temperatures in the range 483-544°C - Messiga et al, 1983). Their similarity with the Dalsfjord coronites confirms the rather low equilibration temperature of the latter in comparison with those to the north of Nordfjord which give temperatures around 700°C (Griffin and Raheim, 1973 - note that more recent estimates of P-T conditions on the coast north of Nordfjord by Griffin et al, 1985 are as high as 800°C, (18-20 kbars). Such low temperatures would have resulted in the stability of the observed talc + amphibole rather than orthopyroxene or olivine in the corona cores. The different crystallisation sequence, with early breakdown of plagioclase to Na-pyroxene at Dalsfjord may be due to crossover of garnet-forming equilibria with omphacite-forming equilibria such that in an up-pressure evolution earlier production of omphacite relative to garnet is favoured at low temperatures. This is shown schematically in Figure 4.30.

SECTION 4.03 Gneisses and other lithologies

Mineral compositions for these rocks are in tables 4.11 to 4.18 inclusive.

4.03.1 Early, high-grade assemblages

The assemblages described here are essentially anhydrous and occur as relics in domains of low strain. Field and petrographic observations show that these assemblages pre-date the S_1 foliation which is associated with the crystalisation of eclogites, garnet + phengite-bearing gneisses and chlorite-harzburgites. They all lack garnet as a "primary" phase, although garnet is frequently found forming corona-structures around pyroxenes and oxides.

Metabasic dykes are found within charnockitic gneisses near Flekke and Gjörlanger (figures 2.01, 2.41, 2.42). The metabasites are olivine-normative and contain the assemblage:-

plagioclase + orthopyroxene + clinopyroxene + K-feldspar +/- Fe-Ti oxide +/- spinel +/- biotite

The host gneisses, where least deformed and hydrated contain:-

plagioclase + quartz + K-feldspar + orthopyroxene + clinopyroxene +/- Fe-Ti oxide +/- biotite

Orthopyroxenes are hypersthenes and clinopyroxenes are augites and salites (table 4.11). The latter are generally pigmented with a khaki hue, but locally become clearer and pale green in colour. Orthopyroxenes frequently contain parallel exsolution lamellae of colourless clinopyroxene. They also contain rounded inclusions of pigmented clinopyroxene, possibly formed by granule exsolution. Plagioclases and K-feldspars are heavily clouded with zoisite and mica and reliable analysis was not possible. Oxides are intergrowths of ilmenite and magnetite. Biotites are Ti-rich (3.65 mass % TiO_2 - table 4.16). No spinel was found in the slides used for microprobe analysis, but where they were observed, their deep green, semi-opaque appearance indicates that they are pleonastes.

The association of pyroxenes and spinel in the olivine-normative dyke-rocks is characteristic of intermediate pressure granulites (Green and Ringwood, 1967) at minimum pressures of 5-9kbar between 700-1000°C. If the olivine - out curve of Green and Ringwood (1967) is extrapolated parallel to that for the CMAS system (Gasparik, 1984) rather than linearly (see figure 4.29) somewhat lower pressures at 700°C (about 5 kbar are indicated). The absence of primary garnet in equilibrium with the pyroxenes in the dykes and the more basic gneisses indicates pressures less than 12.5 +/- 1.0 kbar at 1000°C and 7.5 +/- 1.0 kbar at 700°C when referred to the results of Green and Ringwood (1967) for a quartz tholeiite. The appearance of garnet in experiments on a dry adamellite (Green and Lambert, 1965) occurs over a similar range of P-T conditions; these results are applicable to the more K-feldspar rich charnockitic gneisses.

Equilibration temperatures are not easy to estimate for the charnockitic rocks due to alteration and disequilibrium of their minerals. Results of thermobarometry studies on charnockites give conditions in the range 600-1000°C and 2-12 kbars (see compilation in table 2 of Wendlandt, 1981). Most estimates seem to lie in the range 700-800°C. Geothermometry based on compositions of coexisting pyroxenes can be used to define equilibration temperatures of charnockites. In a recent assessment of two-pyroxene thermometry in granulites Stephenson (1985) concluded that the methods of Kretz (1982) and Lindsley (1983) gave results most consistent with independent estimates of temperature, although the latter method, which is solved graphically, has rather poor precision. Both these methods have been applied to a K-feldspar-poor charnockitic gneiss (sample D99) and a metabasic dyke-rock (D134). The compositions are plotted on figure 4.31, superimposed upon the isotherms for the graphical thermometer of Lindsley (1983).

In the gneiss, exsolved lamellar and granular augites show little variation in composition and both are similar to the rim composition of a sample showing no exsolution features. Hypersthenes show more variation, with the host to the lamellar augites having higher diopside contents than hosts to granular augite or unexsolved grains. Temperatures on the augite limb are somewhat lower for granule-bearing and unexsolved samples (600°C and 660°C respectively) but the host to the lamellar augite varies from 690-830°C. Coexisting pyroxenes give temperatures of 777-846°C for the Ca-transfer equation of Kretz (1982) and 772-851°C for his Fe-Mg exchange equation, in good agreement with the augite limb of Lindsley's (1983) model. The characteristically lower temperatures of the orthopyroxene limb are commonly found in granulite-facies rocks (Lindsley, 1983; Stephenson, 1985).

Analyses from the basic dyke pyroxenes show that in augites the lowest diopside concentrations are in grain cores, while rims and unpigmented parts of grains are most diopsidic, approaching salite. Cores of smaller grains are less diopsidic than in larger ones. The most diopsidic hypersthene is from a core analysis, although some core values, along with a rim analysis, are less calcic. An augite/hypersthene core pair gives 900/850°C (Lindsley, 1983), 856°C (Kretz, 1982, Ca-transfer) but only 607°C for the Kretz (1982) Fe-Mg exchange thermometer. A rim/rim pair (same grains) gives 590°C and 690°C (Lindsley cpx and opx respectively) and only 399°C (Kretz Ca-transfer) and 503°C (Kretz Fe-Mg exchange). Two other single hypersthene core analyses give temperatures of 600°C and 660°C while single pyroxene temperature estimates for small augite cores and a rim give 650-680°C and unpigmented areas in augites lie close to the 500°C isotherm in figure 4.31

(Lindsley, 1983). The pigmented core of a large augite falls on the 1000°C isotherm.

In short, at high temperatures the Kretz (1982) and Lindsley (1983) 2-pyroxene thermometers give reasonably consistent results, but at lower temperatures agreement is less good. Hypersthenes give rather lower temperatures than augites in many cases. Lindsley (1983) considered errors for the graphical thermometer to be +/- 50°C. The values quoted here are at 10 kbar. Using the 5 kbar graphical thermometer gives results within the error range. Kretz considered that for small numbers of samples (less than 4) precision is probably +/- 120°C. Hence the exsolved pyroxenes of the charnokitic gneiss give temperatures of about 800 +/- 120°C while cores of pyroxenes in the dyke give 900-1000°C, with diffusive re-equilibration (related to distance from grain boundaries) to as low as 550 +/- 120°C. Rim temperatures are in broad agreement with garnet-corona/augite rim estimates by the method of Ellis and Green (1979) of 592-621°C (16 kbars) for the dyke rock, suggesting that the low temperatures were associated with garnet growth. Garnet analyses appear in table 4.14. The higher temperatures are consistent, within error, with those derived from the pyroxenes of the anorthositic troctolite discussed in the previous section.

Consideration of the minimum stability of augite + hypersthene + plagioclase with respect to hornblende might be useful in defining the P-T field for the charnockites. Experiments by Allen et al (1975) have located the upper stability of hornblende in an andesite and an olivivne tholeiite to be 970°C and 1000°C at 10kbars respectively where P_{H2O} = P_{total} on the magnetite-wustite buffer. These temperatures are rather high when compared with the pyroxene thermometry outlined above and are well above the solidi for these lithologies under water-saturated conditions (Stern et al, 1975). However, there is much recent evidence to suggest that many granulite-facies rocks formed under conditions of reduced water activity and probably in the presence of a fluid with high X_{CO2} (see, for instance, Hansen et al, 1984 and references therein).

Low water activities would have considerably reduced the stability of amphibole. The results of provisional calculations by Spear (1981) indicate that complete dehydration of amphibole can occur at 700°C at 2 kbars if X_{H2O} = 0.1. Rare modal biotite in the rocks discussed here indicates that small amounts of water were present during their crystalisation. However, the paragenetic association of biotite is rather ambiguous in the light of common hydration of the granulite phases. Without better delineation of fluid composition (or saturation) errors in minimum temperature estimates are at least as great as for pyroxene thermometry.

Similar considerations apply to determination of the melting temperatures of the charnockites. Experiments in the system $KAlSiO_4$ - Mg_2SiO_2 - H_2O - CO_2 by Wendlandt (1981) show that vapour-free and CO_2-saturated melting of sanidine + quartz + enstatite occurs at close to 1000°C between 1 bar and 20 kbar, whilst decomposition of phlogopite + sanidine + quartz + vapour ($X_{CO2} = 0.5$) occurs at 750°C between 5-10 kbars. The very small amounts of modal biotite and the high pyroxene core temperature estimates indicates that water activity was low and the higher end of this range of temperatures is aplicable to the more potassic charnockitic gneisses.

Intrusive websterites in the Gjörlanger area (figures 2.01, 2.54) have an early assemblage:-
clinopyroxene + orthopyroxene + spinel +/- olivine.

In samples 79/8 and D45, the clinopyroxenes are diopsides with up to 3.3 mass % Al_2O_3 and the orthopyroxenes are bronzitic with up to 2.8 mass % Al_2O_3 (table 4.11). Spinels are dominantly chromitic magnetites with up to 12% of the chromite endmember, although rare green Cr-poor pleonastes occur as relics in masses of chlorite (table 4.17 - see appendix II for estimation of ferric iron). A single olivine analysis (table 4.12, 79/8 OAA) has the composition Fo_{61}.

The assemblage pyroxene + spinel + olivine is an intermediate pressure granulite-facies assemblage in ultramafic rocks. In the CMAS system the stability of this assemblage relative to anorthite and forsterite lies above 4 kbars, 700°C and 7 kbars, 1000°C (Gasparik, 1984). Addition of FeO to the CMAS system in amounts found in the Websterites produces little change in the curve and the low Na_2O contents of the rocks are unlikely to have had any significant effect (see discussion in Herzberg, 1978). The upper stability of the websterite assemblage is defined by the lower stability of garnet. In CMAS garnet appears above 15-16 kbars, 700-1000°C (Gasparik, 1984). Addition of FeO to the system stabilises more almandine rich garnet at rather lower pressures; for instance in the experiments of Green and Ringwood (1967) at 1100°C garnet appeared at 15.2 +/- 0.6 kbars in a basalt with Mg/Mg + Fe = 0.82, whilst a basalt with Mg/Mg + Fe = 0.70 (close to values for the websterites - see table 3.7) evolved garnet at 12.9 +/- 0.6 kbars.

Pyroxene thermometry gives rather variable results, although they are not incompatible with those from the charnokitic rocks. Pyroxene compositions are plotted on the isotherms of Lindsley (1983) for 10 kbars. Highest temperatures are derived from diopside cores at 910 and 990°C. Most other values, from cores, rims and a colourless recrystalised diopside lie closee to 500°C. Bronzites from sample 79/8 occupy the same range of isotherms as those in the charnockitic rocks (although errors are likely to be greater as the isotherms are closer together) with a range of 600-800°c, the rim value being the lowest. Bronzites from sample D45 give values consistently around 650yoC, but lying near the end of the plotted isotherms. Core temperatures are, characteristically, lower in bronzites than diopsides. The Ca-transfer thermometer of Kretz (1982) gives rather lower temperatures, between 146°C and 690°C. It is noteworthy that the best agreement between the Kretz (1982) and Lindsley (1983) thermometers occurs when Fe^{3+} in the clinopyroxene is lowest. Recalculation of Kretz (1982) temperatures with Fe = total Fe markedly improves the agreement (using D45 PBK a change from 690°C to 778°C occurs, against a Lindsley value of 990°C). Three reasons can be put forward to explain the discrepancy:-

1) Kretz (1982) made erroneous estimates of the ferric iron contents of his pyroxene calibration data set (Skaergaard and Quairading granulite Ca-pyroxenes).

2) The Fe^{3+} content of the pyroxenes discussed here has been overestimated. In this respect it is noteworthy that plotting compositions recalculated to Fe = Fe total on Lindsley's (1983) projection produces only a small temperature change (from 990°C to 940°C for sample D45 PBK).

3) Kretz (1982) does not allow for non-quadrilateral components in his geothermometer. Adjusting his calculation to allow for Fe^{3+}, Al, Cr and Na as for Lindsley (1983) increases the apparent temperature to 782°C.

It is possible that a combination of (2) and (3) above may be responsible for the discrepancy. Recalculating sample D45 PBK on this basis produces a temperature of 842°C and 940°C for Kretz (1982) and Lindsley (1983) respectively. Note finally that the Kretz (1982) Fe-Mg exchange thermometer, when recalculted for Fe = Fe total in clinopyroxene gives D45 PBK a temperature of 656°C, close to the Lindsley (1983) orthopyroxene temperature of 650°C. Fe-Mg exchange between pyroxenes may continue to lower temperatures than Ca-transfer (Saxena, 1983) and diffusive re-equilibration of orthopyroxene may continue to lower temperatures than clinopyroxene (Lindsley, 1983). Hence there may be some internal consistency in these results. However, a much larger data set is required to evaluate their precision before firmer conclusions can be drawn. Pyroxene thermometry is nonetheless taken to indicate early high temperatures of 950 +/- 120°C with later cooling to 500-700°C.

The pyroxene + olivine + spinel assemblage has been transformed to the hydrous minerals Ca-amphibole + chlorite, while diopside rims have lost their characteristic dusty appearance and become clear and colourless, sometimes recrystalising to a much finer, granular diopside. Amphibole and chlorite is replaced by garnet, which also forms corona-like rims around pyroxenes. Analyses of garnets are presented for reference in table 4.14, amphiboles in table 4.15 and chlorites in table 4.16. The relationships of these replacements are petrographically complex, but some inferences can be made. Chlorite probably evolved by the reaction:-

H_2O = chlorite + olivine + spinel

Amphibole could have been produced by:-

clinopyroxene + orthopyroxene + spinel + H_2O = amphibole + olivine

Jenkins (1981) has determined the intersection of these reactions at an invariant point in the system CaO - MgO - Al_2O_3 - SiO_2 - H_2O (CMASH) at 825 +/- 10°C, 9.3 +/- 0.5 kbar. Hence in CMASH hydration of a pyroxene - forsterite - spinel assemblage requires cooling to temperatures of less than 825°C. Alternatively, if the anhydrous assemblage was stabilised at low P_{H_2O} the chlorite + amphibole may have developed by hydration at constant T and P below 825°C. In the complex iron-bearing system chlorite stability is displaced to lower temperatures (McOnie et al, 1975) but as the chlorites in the websterites have Fe/Fe + Mg less than 0.2 this effect is likely to be small. Fe may also shift the stability of amphibole to lower temperatures, but this may be counteracted by the presence of significant Al and Na (Obata and Thompson, 1981).

The appearance of garnet in CMASH is controlled by reactions emanating from an invariant point defined by the intersection of the chlorite decomposition curve with the reaction:-

clinopyroxene + orthopyroxene + spinel = pyrope + forsterite

Jenkins (1981) located this invariant point at 865 +/- 5°C, 15.2 +/- 0.3 kbars. Obata and Thompson (1981) infer from Schreinemaker's analysis that garnet-forming reactions involving hydrous phases at temperatures lower than this invariant point have rather shallow slopes in the P-T plane. The exact paragenetic relationships of the garnet are not easily determined petrographically, but the above observations suggest that garnet could have formed by dominantly up-pressure evolution to over 15 kbars from reactions involving amphibole, chlorite, pyroxenes and possibly, olivine and spinel, as shown schematically in figure 4.32. Applying the geothermometer of Ellis and Green to garnet-pyroxene pairs in the websterite gives temperatures of 579-649°C, while a garnet amphibole pair gives 583°C by the method of Graham and Powell (1984). These results are consistent with results from eclogitic bands in the pyroxenites (section 4.01). In summary, an evolution of the websterites from temperatures in excess of 900°C and pressures between 7-13 kbars was followed by cooling to 500-650°C and hydration, followed by compression to over 15 kbars at 580-650°C during the eclogite-facies metamorphism.

The mineralogy of the chlorite-harzburgites is not discussed in detail here, but some points are of interst. Analyses of olivines, anthophyllites, chlorites and talc, and oxides are given in tables 4.12, 4.15, 4.16 and 4.17 respectively. Petrographic descriptions appear in sub-section 2.03.3. Relics of pleonaste spinel are occasionally observed among the chlorites, suggesting that chlorites formed from an early anhydrous assemblage:-

olivine + enstatite + spinel +/- Fe-Ti oxide

Hence the chlorite harzburgites seen to have evolved by hydration of spinel harzburgites. The strong chlorite-enstatite schistosity is correlated with the S_1 schistosity found in the gneisses and eclogites. Similar rocks, with slightly more magnesian chlorites have been described from the Lepontine Alps (Trommsdorf and Evans, 1969). Consideration of adjacent pelites and marbles indicated maximum temperatures for the harzburgites of 600-650°C, indicating that P_{H2O} was less that P_{total} as enstatite and olivine were stable with respect to talc + olivine at these temperatures. Similar conditions are, therefore, inferred for the Dalsfjord harzburgites, with the temperatures of about 600°C being consistent with those for the eclogites. The common presence of carbonate in textural equilibrium with chlorite and enstatite may indicate that water activity was lowered by the presence of CO_2. Late post-tectonic replacement of enstatite by anthophylite + olivine then talc + olivine implies a late increase in P_{H2O} and/or drop in temperature (see Jenkins, 1981 figure 7 and review of Evans, 1977 with references therein).

4.03.2 Omphacite and phengite-bearing gneisses

Omphacite and garnet are commonly found within the "massive green gneisses" and the "K-poor gneisses" of the Gjörlanger unit. As discussed in Chapters 2 and 3 these gneisses have gradational relationships with granodioritic augen gneisses (more or less altered charnockitic rocks) and are broadly dioritic in composition. Omphacite usually occurs as small relics in masses of extremely fine grained amphibole-plagioclase symplectites, which is a characteristic of quartz-bearing eclogitic rocks in Dalsfjord area.

The petrography of the low-K gneisses was described in sub-section 2.04.3. Pre-symplectite minerals associated with the S_1 foliation are:-
omphacite + garnet + quartz + zoisite + kyanite + phengite + paragonite + rutile

This assemblage does not appear to represent a paragenesis, however, as the minerals are grouped into mafic and non-mafic domains, apparently as a result of shearing of a more isotropic rock.
Coexisting phases are:-
Garnet + omphacite + rutile + quartz
Kyanite + zoisite + phengite + quartz
Zoisite + paragonite + quartz

Note that neither (omphacite + kyanite) nor (omphacite + paragonite + kyanite) coexist.

Garnet and omphacite compositions (tables 4.14 and 4.11, sample D24) are plotted on the diagrams of Coleman et al (1965) and Essene and Fyfe (1967) in figures 4.33 A and B respectively. Garnet rim compositions fall within the range of eclogite garnets described in section 4.01 but are less magnesian than the rim compositions of the latter. Omphacites contain 45-52% jadeite, lying at the most jadeitic extreme of the eclogite omphacite compositions.

An analysis of a phengite (table 4.16) is plotted on figure 4.07. It is slightly more iron-rich (Fe^{2+} = Fe total) than the eclogite phengites, but Si^{4+} (6.635 cat/fu) lies in the range of the eclogite analyses. An analysis of a clinozoisite (table 4.18) shows it to be fairly pistacitic with 0.412 cat/fu Fe^{3+}. However, this is a rim analysis. Unfortunately, no successful core analyses were made, but cores have much lower birefringence than rims and are assumed to be closer to pure clinozoisite. A paragonite analysis (table 4.16) contains significant muscovite and margarite substitution, with 7.5% replacement of Na by Ca and 9.1% replacement of Na by K.

The estimated P-T conditions for the low-K, omphacite-bearing gneisses are summarised on figure 4.34. Unaltered garnet/omphacite contacts are only rarely preserved in these rocks, but one unaltered contact gives a temperature of 605°C (16 kbars) which is consistent with the results for eclogites. Assuming a 2% maximum error for silica analysis, this estimate has error brackets of +/- 103°C.

Maximum temperatures are constrained by the absence of melting features in these gneisses. Stern et al (1975) have determined the water-saturated solidus for a tonalite, which is similar to that for a gabbro and lies at 625^{o}C at 15 kbars. Minimum temperatures are defined by the lower stability of the assemblages clinozoisite + paragonite + quartz and clinozoisite + kyanite + quartz with respect to lawsonite, jadeite and albite, as determined by Holland (1979b) and Perkins et al (1980). These relationships are complicated by the presence of small amounts of muscovite and margarite in solid solution. The steep slope of the paragonitic limb of the muscovite-paragonite solvus in the T-X plane (Eugster et al, 1971) dictates that small amounts of muscovite in solution can stabilise the mica to significantly higher temperatures. Conversely, margarite in solid solution will tend to stablise the paragonite to lower temperatures (Franz et al, 1977) hence the effects of these impurities probably effectively cancel each other out. It is, therefore, concluded that the observed assemblage paragonite + zoisite + quartz indicates a minimum temperature of 515^{o}C at 16 kbars (Holland, 1979b) whilst melting in this assemblage with P_{H2O} = P_{total} occurs at about 68o^{o}C above 10.8 kbars (Franz and Althaus, 1977). The assemblage clinozoisite + kyanite + quartz indicates slightly higher minimum temperatures (550^{o}C at 16 kbars - Perkins et al, 1980) and a minimum pressure of 8.2 kbars at 625^{o}C (Jenkins, 1984).

Minimum pressure can also be estimated from the omphacite composition. The minimum stability of $DiJd_{50}$ + quartz occurs at 13.35 kbars at 600°C (using the data of Holland, 1980 and assuming jadeite activity = X_{Jd} as in Holland, 1983). Maximum pressure is taken from the maximum stability of albite relative to jadeite + quartz, giving 16.25 +/- 0.5 kbars at 600°C. It is noteworthy here that lobate intergrowths of two apparently omphacitic pyroxenes have been observed in some of the massive green and K-poor gneisses (figure 2.48), indicating unmixing. Recent work (Holland, 1983) indicates that on the join diopside-jadeite the critical temperature for the solvus occurs below 600°C. Based partly on the calorimetric results of Holland (1983), Gasparik (1985) has determined the critical temperature at 565°C at composition $Di_{42.4}Jd_{57.6}$. The critical temperature seems to be increased by increasing acmite component, but complete miscibility occurs above about 700°C for omphacites with 3-12% acmite (Carpenter, 1980; Carpenter and Smith, 1981). Unfortunately, no samples of such pyroxenes could be obtained suitable for probe analysis. However, their occurrence is consistent with the estimated P-T conditions. Further work on these omphacites (and perhaps some of the rather "feathery" omphacites in other lithologies) with XRD structure refinement and transmission electron microscopy may prove useful in the finer resolution of the P-T-t history of this area by relating microstructure, phase compositions and space-group symmetry (see, for instance, Rossi et al, 1983).

veined by the gneiss at the contact (see section 2.02) and hence the eclogite is autocththonous with respect to the gneiss. The S_1 foliation postdates the veining.

The gneiss contains the assemblage:-
quartz + albite + clinozoisite + phengite + garnet +/- sphene.
The albite has the composition Ab_{98} (table 4.13). The clinozoisites (table 4.18) are usually zoned with low birefringent cores and higher birefringent, more ferric rims. Rims tend to be more ferric (pistacitic) where the phengite is altered, indicating that the unaltered phengites originally coexisted with the more aluminous clinozoisite core compositions. The least pistacitic clinozoisites in this gneiss are significantly more ferric than those in the eclogites, for instance a clinozoisite core in D178 has $Fe^{3+}/Fe^{3+} + Al = 0.174$, whereas eclogite D77 zoisites are close to 0.043.

Phengite compositions have been plotted on figure 4.07. They are significantly more iron-rich than those from eclogites of the underlying Flekke and Gjörlanger units, but have similar Si^{4+} contents (no attempt has been made to estimate ferric iron contents, but whole-rock ferric iron is low (0.25 mass % and much of it may be locked up in clinozoisite). There does not seem to be any strong systematic variation in composition, although core compositions tend to be the most siliceous and magnesian compared with rims of altered grains or secondary grains. Alteration of the phengites involves static breakdown to a fine intergrowth of khaki-green biotite (table 4.16), K-feldspar and quartz with minor new white mica. Si^{4+} contents of phengite cores lie in the range 6.58-6.62. Na contents are less than 0.01 cat/fu and Ca contents less than 0.002 cat/fu.

Garnet compositions are plotted on figure 4.33A. They have very low pyrope contents where compared with eclogite garnets, reflecting the very low whole-rock MgO content. However, these garnets are significantly enriched in grossular relative to the eclogite garnets and also have rather higher Mn contents (table 4.14). These compositions are similar to those for garnets in granitic gneisses in the Forde area (Krogh, 1980). Rims tend to be depleted in Ca relative to cores, which is also found in the Förde gneiss garnets.

P-T conditions for the psammitic gneiss are summarised on figure 4.35. Equilibration temperatures for two garnet-phengite rim pairs have been estimated according to the method of Green and Hellman (1982) assuming that Fe^{2+} = Fe total. At 10 kbars the values are 607 and 625°C, or 639 and 658°C at 16 kbars. These are maximum temperatures due to the assumption about ferric iron. If 2/3 rds of the iron is assumed to be ferric, the temperature for the higher of the two values quoted above reduces to 580°C. As the whole-rock Fe^{3+}/Fe_{total} ratio is only 0.22 and much of the ferric iron may be taken up in epidote then a temperature somewhere in the range 580-658°C (quoted at 16 kbars) seems to be reasonable.

Velde (1965, 1967) has shown that phengite + biotite + K-feldspar + quartz is a high-pressure, low temperature assemblage, with the Si content of the phengite increasing with increasing pressure or decreasing temperature (see also Massone and Schreyer, 1979). using the silica curves of Velde (1967) the phengites in the gneiss give a pressure of approximately 9.0 kbars at 600°C (Si = 3.3). This is a minimum pressure in the absence of biotite + K-feldspar.

Preliminary experimental results of Rao and Johannes (1979) for the reaction:-
Staurolite + annite + quartz = almandine + muscovite + H_2O
gives a minimum pressure of 6.5 kbars for the stability of almandine-rich garnet and phengite. However, this curve was unreversed and rather poorly constrained and does not allow for celadonite and grossular substitutions.

The garnets in the psammitic gneiss are rich in grossular (25-31%). Green and Mysen (1972), in a dicussion of eclogites and gneisses to the north of Nordfjord, concluded that the occurrence of grossular-rich almandine garnet in the presence of quartz is an indicator of high pressures, probably in excess of 8 kilobars at temperatures over 600°C.

The absence of melting features affecting the S_1 fabric phases in the Vardheia unit gneisses indicates that the granite or pelite solidus conditions were not exceeded. The granite solidus under water-saturated conditions lies at 600-630°C between 5-15 kbars (Stern et al, 1975) which is very similar to the water saturated curve for the beginning of melting in a pelite (Thompson, 1982).

Maximum pressures can be inferred from the presence of nearly pure albite in the gneisses and the absence of jadeite + quartz. Albite appears to form granular ribbons lying in the foliation in the same manner as the quartz and, apart from some internal strain, is in textural equilibrium with coexisting quartz, mica and garnet. It shows no evidence of replacing a pyroxene. The absence of jadeite + quartz indicates pressures below 16.4 kbars at 600°C. As the foliation in the gneiss can be correlated with that in the underlying eclogite and as their contact indicates that the eclogite is autochthonous, then it can be concluded that the gneiss assemblage:-

quartz + albite + clinozoisite + phengite + garnet +/- sphene

is cofacial with that in the eclogite. Estimated equilibration temperatures are compatible for the two lithologies and although minimum pressure estimates for the gneiss are somewhat lower than those for the eclogites. Phengite + garnet + clinopyroxene + quartz are known to coexist with jadeitic or omphacitic pyroxenes in other high-pressure terrains (eg Sesia-Lanzo Zone, Western Alps; Compagnoni, 1977) indicating that they can exist at high pressures. Hence the presence of albite in the gneisses indicates that pressures did not exceed about 16 kbars at the peak of eclogite-facies metamorphism.

Section 4.04 Conclusions - a P-T path for the Hellevik-Flekke area

In this chapter evidence has been presented for four metamorphic episodes which have affected the rocks of the Hellevik-Flekke area.

An eclogite-facies metamorphism, which transformed basic intrusive rocks into various types of eclogite, had peak metamorphic conditions at 597 +/- 30°C and 16.2-16.8 kbars (figure 4.17) except in the Basal Gneisses, where temperatures of 566 +/- 24°C at a minimum pressure of 11.3 kbars was attained. As field evidence shows that metabasites, gneisses, websterites and metaperidotes were all mutually autochthonous the conditions for the eclogites must also apply to the other lithologies. Estimates of P-T conditions for some of these other lithologies are broadly consistent with this hypothesis.

The eclogite-facies assemblages evolved along two mineralogical pathways:-

1. transformation of pre-existing amphibolites, and

2. transformation of pre-existing igneous rocks.

(Granulite-facies assemblages were very probably also transformed but there is less direct evidence to show how this occurred). There is rare evidence that the igneous rocks suffered some amphibolitisation prior to eclogitisation, which provides a link between the two types and confirms that both types of transformation took place under the same set of changing conditions.

Conditions of amphibolite-facies metamorphism prior to the formation of eclogites were 550-600°C with pressures constrained between 4-10 kbars. This implies a steep compressional path to the high-pressure peak with only a small increase in temperature. This does not apply to the Basal Gneiss eclogites, however, whose pre-eclogite facies evolution is not represented in its mineralogy.

After the eclogite-facies peak conditions were reached a strong decompression occurred with a pressure drop of some 10 kbars with little change in temperature. This was presumably followed by cooling to surface conditions as the terrain reached its present position.

The resulting P-T path, starting at the amphibolite-facies conditions, has a tight "hairpin" shape which is rather similar to that deduced for other Norwegian eclogites.

The earliest recorded metamorphic episode in the area was at granulite-facies conditions. Pressures are constrained at 7-13 kbars. Temperature estimates from two-pyroxene thermometry are rather variable, ranging from above 950°C down to less than 600°C. Peak granulite-facies conditions were probably between 750-1000°C, with the higher temperatures possibly being relict igneous values. Cooling to roughly 600°C is indicated by some rim compositions, which is broadly consistent with temperatures for the later eclogite-facies metamorphism.

The best preserved anorthositic troctolites show evidence for virtually isobaric cooling from solidus temperatures and pressures (1200°C at below 9 kilobars) to granulite-facies conditions at 900-710°C. It might then be suggested that the basic-ultrabasic complex of the Flekke Unit was intruded into country-rocks under intermediate-pressure granulite-facies conditions.

The link between the granulite-facies metamorphism and the subsequent evolution to the eclogite-facies peak is difficult to determine. The simplest model would be a continuous evolution by simple cooling, with localised formation of amphibolites prior to compression during the eclogite-forming event.

However, the picture is complicated by the presence of the zoisite rocks found at Flekke, which might be interpreted as metarodingites (see sections 2.03.4, 3.04.3 and figure 2.29).

If this interpretation is correct then these rocks must have been close to the surface of the crust at some time after the intrusion of the basic-ultrabasic complex and before the high-P metamorphism. If this was intruded under granulite-facies conditions some considerable erosion is implied, requiring the presence of a significant unconformity in this area. As the base of the Vardheia unit has an intrusive contact with the Flekke Unit it cannot be a candidate for such an unconformity. No other such unconformity is known, although any existing one may easily be obscured by strong deformation.

Some evidence of low-grade metamorphism, possibly at greenschist or even glaucophane - lawsonite-schist facies might be expected if these rocks had closely approached the surface prior to eclogite-facies metamorphism. However, no signs of this are found, the lowest-grade assemblages being of the amphibolite facies. This is not unusual for terrains having undergone strong 'prograde' evolution to eclogite-facies. For instance, Messiga et al (1983) could only infer the presence of early low-grade assemblages in ophiolitic metagabbros which presumably underwent ocean-floor metamorphism prior to subduction. In the Sesia-Lanzo Zone, Western Alps, which is believed to have occupied a high crustal level prior to high-pressure Eo-Alpine metamorphism (Lardeaux et al, 1982), only relics of an early granulite-facies assemblage are found.

Another possibility which must be considered is that the zoisite precursors formed after intrusion of the basic-ultrabasic complex, but before the granulite-facies metamorphism. This would allow for a continuous evolution from granulites via amphibolites to eclogites, but means that the igneous assemblages and textures must have survived two tectonometamorphic cycles virtually unaltered, which does not seem very likely.

Hence the link between the granulite- and eclogite-facies events remains uncertain. The determination of a continous or interupted evolution is important for tectonic considerations (see Carswell and Cuthbert, in preparation). What can be said is that the eclogite-facies protoliths in the Hellevik-Flekke area were at least as deep as mid-crustal levels (4-10 kbars or 13-33 kilometers) prior to their up-pressure evolution.

The P-T conditions deduced for the various lithologies are summarised on figure 4.36 with alternative P-T paths for continuous evolution or post-granulite exhumation as outlined above.

Finally, the metamorphic episodes are correlated with the various structural elements in table 4.19. Metamorphic events are designated as follows, from earlier to later:-

M_1 - granulite facies
M_2 - amphibolite facies
M_3 - eclogite facies
M_4 - retrogression to amphibolite facies

CHAPTER 5 - SUMMARY OF CONCLUSIONS AND DISCUSSION OF REGIONAL SIGNIFICANCE AND TECTONIC IMPLICATIONS

Summary of conclusions and discussion of reginal significance and tectonic implications

In this final chapter the main conclusions are reviewed in the light of the aims set out in Chapter 1. The resulting geological model is compared with results for other parts of the Basal Gneiss Region. The consequences of deductions about the eclogite-facies metamorphism for tectonothermal models of the Basal Gneiss Complex are discussed. Finally, some suggestions for future work are put forward.

Section 5.01 Summary of conclusions: a geological synthesis for the Hellevik-Flekke area

It was stated in Chapter 1 that the main aim of this Thesis is to evaluate the relationships between eclogites and their host gneisses and their comparative metamorphic histories. This has involved detailed fieldwork, bulk-rock geochemical analysis and petrographic and mineral chemical studies. The results, it was suggested, would help to elucidate the mechanisms for formation and emplacement of the eclogites and hence aid in the tectonothermal modelling of the Basal Gneiss Complex.

Much of the work on eclogites and related high-pressure metamorphic rocks in the BGR has been carried out to the north of Nordfjord and the Sunnfjord eclogites are relatively poorly known (see Binns, 1967 and Krogh, 1980 for work in the Dalsfjord area). This work provides new data for eclogites and related rocks which have not been previously studied.

The geological history of the Hellevik-Flekke area, with emphasis on the Gjörlanger, Flekke and Vardheia units, is outlined below:-

The earliest discernable events in the area were the deposition of the sedimentary precursors to the paragneisses (probably as variably argillaceous sands) and the intrusion of a suite of calc-alkaline igneous rocks ranging from basalts to granodiorites. The basalts often cut the more acid rocks in dykes and sills. The comparative geochemistry of the calc-alkaline rocks suggests an origin in a continent-margin magmatic arc environment.

The calc-alkaline rocks were invaded by intrusions of websterite. Preliminary U-Pb zircon work on anatectic backveins in the websterites give an age of approximately 1500 Ma, which is a minimum age for the calc-alkaline rocks. These two groups of rocks now form the Gjörlanger Unit.

The Gjörlanger Unit and the sediments of the Vardheia Unit underwent high-grade metamorphism at intermediate-pressure granulite-facies (M_1) and there is evidence that at some stage the Vardheia Unit suffered migmatisation. During, or shortly before, this metamorphism the contact between the Gjörlanger and Vardheia Units was the locus of a layered basic-ultrabasic intrusion of broadly leucotroctolitic composition similar to those associated with mid-Proterozoic massif anorthosites.

The granulite-facies metamorphism was probably responsible for the depletion of Rb from the Gjörlanger unit gneisses. Metamorphic conditions (750-1000°C, 7-13 kbars) indicate enhanced heat-flow at depths corresponding to normal thicknesses of stable continental crust.

If the calc-silicate pods in the chlorite-peridotites at Flekke are interpreted as metarodingites, a period of uplift and cooling into the stability field of serpentine followed the intrusion of the basic-ultrabasic complext (Flekke Unit). The lack of any corroborative evidence makes this conclusion rather tentative, however. What is more certain is that a period of amphibolite-facies metamorphism (M_2) followed the granulite-facies (M_1) event, at temperatures of 550-600°C and 4-10 kbars, as indicated by inclusions in eclogite garnets.

The M_2 amphibolite-facies metamorphism was followed by (and may have been continuous with) a period of intense tectonism and increase in lithostatic pressure, leading to the formation of eclogites, garnet amphibolites, omphacite-bearing gneisses, phengite-garnet gneisses, garnetiferous websterites and probably chlorite-peridotites associated witha pervasive planar fabric (S_1) and tight, often rootless isoclinal recumbant folds (F_1, F_2). Peak high pressure metamorphic conditions were 597 +/- 30^{o}C at 16.2-16.8 kbars, corresponding to loading by approximately 55 kilometers of overlying continental crust. This eclogite facies metamorphism (M_3) was followed by a rapid near-isothermal decompression (M_4) during which only localised shearing took place. Breakdown of high-pressure phases was generally static, with symplectites replacing omphacite and phengite and fibrolite replacing kyanite. Minimum recorded pressures are approximately 6 kilobars, indicating a pressure drop of some 10 kilobars which corresponds to the removal of approximately 33 kilometers of overburden prior to the commencement of significant cooling.

The age of the high-pressure metamorphism is not well constrained by the results embodied in this Thesis. However, the mapping of Kildal (1970) shows that the "Cambro-Silurian schists" were involved in fold episodes shared by the underlying eclogite-bearing gneisses. Should this prove to include the F_1 and F_2 episodes a lower Palaeozoic ("Caledonian") age for M_3 and M_4 are indicated. Geochronological studies on U-Pb in zircon and Nd-Sm and Rb-Sr on various eclogite minerals (Krogh, 1973; Griffin and Brueckner, 1980, 1982; Gebauer et al, 1982) indicate that some eclogites in the BGR either crystalised, or cooled through the Nd-Sm blocking temperature, at about 400 Ma BP. However, Griffin and Brueckner (1982) noted that eclogites which give equilibration temperatures below 700°C (such as those in Sunnfjord - Krogh, 1980 and this work) usually show poor equilibration of REE between omphacite and garnet, making age determinations difficult.

Griffin and Brueckner (1980) noted that Rb-Sr mineral ages from the BGR (usually from amphibolite-facies minerals, eg Brueckner, 1972) are usually similar to the Sm-Nd ages, but the latter are often slightly older (maximum difference 70 Ma). From these results it seems likely that the strong decompression may have occupied a period of a few tens of millions of years.

The appearance of the rocks underlying the Hellevik-Flekke area at the surface had occurred by middle Devonian times, which is the age of the basin-fill sandstones overlying the gneisses and over Palaeozoic schists at Solund and Kuamshesten, close to the study area (Kildal, 1970).

It remains to mention the Basal Gneisses. Little can be said about their early history except that the composition of some orthogneisses suggests an origin either as part of the anorthosite-mangerite-rapakivi granite suite of anorogenic intrusive rocks, or possibly highly evolved magmatic arc intrusives. Textures of little-deformed gneisses are comparable with some of the charnockitic rocks of the Gjörlanger Unit, indicating an early granulite-facies metamorphism.

Compositions of phengite-bearing eclogites in the Basal Gneisses are highly altered but indicate alkaline affinities. They show no signs of pre-eclogitic igneous or amphibolite-facies assemblages, inclusions in garnets being omphacite, glaucophane and phengite. Peak eclogite facies metamorphic conditions were 566 +/- 24^{o}C at a minimum pressure of 11.3 kbars. Temperatures were slightly lower than those derived from eclogites and gneisses in the overlying units.

The Basal Gneisses are separated from the Gjörlanger and Flekke Units by a zone of strong attenuation, which has been regionally mapped as a "thrust" (or perhaps more correctly a "slide" zone as no distinct fault can be discerned) by Skjerlie and Pringle (1978). The age of movement on this thrust is not certain, but the planar fabric lies parallel to S_1 elsewhere and is folded by F_3 and F_4. The most strongly attenuated monzonitic gneisses contain relics of phengite parallel to S_1 as well as some garnet. It seems possible, therefore, that the slide zone was operative during the eclogite-facies metamorphism.

The most important findings of the work in the Hellevik area, with respect to elucidating the origin of eclogites in the BGC, are:-

1. Eclogites can be shown to have untectonised contacts with quartzo-feldspathic gneisses. The contacts take the form of sharply defined dyke walls or they may be gradational or interveined with the gneiss.
2. Corona-textures indicate that eclogites have formed from basaltic igneous rocks which originally equilibrated at low pressures (less than 9 kbars). The igneous rocks can be shown to form parts of a layered basic/ultrabasic intrusion which was intruded into its present position and hence was autochthonous prior to the formation of eclogites.
3. High-pressure assemblages have formed in non-basaltic rocks including the low-K dioritic gneisses, psammitic gneisses and intrusive websterites. These assemblages

can be shown to have formed by transformation of earlier granulite-facies assemblages either by corona growth or tectonic recrystalisation. Estimates of P-T conditions for these rocks are compatible with those for the metabasaltic eclogites.

4. Bulk-rock geochemical studies indicate that eclogitised dyke-rocks, grey amphibolite-facies gneisses, charnockitic relics and omphacite-gneisses are genetically related, confirming their gradational or mutually intrusive relationships in the field. The chemical alteration of the eclogites indicates that chemical "fingerprinting" of magma-types for such rocks is highly dubious, particularly where tectonism has removed all traces of their relationships to their host rocks.
5. A P-T path for the evolution of the eclogites can be defined which shows a very high dP/dT for both prograde and retrograde paths, producing a "hairpin" shaped trajectory. From its starting point in the amphibolite-facies to its endpoint as recorded in the mineralogy of the rocks a maximum temperature range of 80°C is indicated, although this is more poorly constrained at lower pressures.
6. Bulk-rock chemistry has a strong effect on the observed eclogite-facies parageneses. A striking example is the interbanding of apparently co-facial garnet-omphacite and garnet-amphibole rocks, the later of which, when taken out of context, would seem to be a fairly unremarkable garnet-amphibolite.

Section 5.02 Regional context

The general geological succession in the Hellevik-Flekke area is similar to that which has been mapped in several areas to the south of Nordfjord. A fairly monotonous "basement" of orthogneisses and migmatites, commonly cut by pegmatites, is usually found to underlie a more heterogeneous sequence of augen-orthogneisses, paragneisses, meta-anorthosites and metabasites with conspicuously few pegmatites. Bryhni (1966) recognised this subdivision, calling the basement the "Jostedal Complex" and the overlying heterogeneous sequence the "Fjordane Complex". Similar relationships have been found in the Tafjord-Grotli area (Strand, 1969; Brueckner, 1977), the Breimsvatn area to the north of Fördefjord (Bryhni and Grimstad, 1970) and Sunnfjord (Skjerlie, 1969; Kildal, 1970; Skjerlie and Pringle, 1978). A correlation with the mapping of Skjerlie (1969) was suggested in Section 2.01.

The age relationships of the Jostedal and Fjordane Complexes have been a matter of some discussion for many years. For instance, Strand and Kulling (1972) seemed to prefer an origin for the heterogeneous gneisses as arkoses, greywackes, volcanics and dolerites equivalent to those in the "Sparagmite Region" of Central Southern Norway, which are of "Eocambrian" or latest Precambrian age. However, a perceptive review by Carswell (1973) indicated that many of these rocks are probably of Proterozoic age (or even older), based on the presence of anorthosites and granulites and some isotopic data. The presence of autochthonous and parautochthonous late Precambrian sandstones lying on basement in the Oppdal area (Krill, 1980 - see Chapter 1 of these Thesis) indicates that both these groups of rocks may be present.

A Proterozoic whole-rock age has been derived from Fjordane complex lithologies: Brueckner (1972) derived a whole-rock Rb-Sr age of 1253 +/- 100 Ma for gneisses by Holsavatn, Sunnfjord (10 km west of Förde) which have been assigned to the Holsen Group of the Fjordane Complex by Skjerlie (1969) and Skjerlie and Pringle (1978).

The preliminary Rb-Sr and U-Pb results obtained in this study indicate a Proterozoic age for intrusion of some of the lithologies corresponding to the Veuring Complex of Skjerlie and Pringle (1978). The presence of relict granulite-facies assemblages ("charnockites") is also indicative of a Proterozoic age by analogy with similar rocks in southern Norway. This author is not aware of any lower Palaeozoic charnockites elsewhere in the Caledonian orogenic belt. Kildal (1970) has allocated rocks corresponding to the Fjordane Complex to "late Precambrian to Cambro-Silurian age" but this appears now to be erroneous and from the conclusions drawn above they appear to be Proterozoic.

The age of the Flekke unit is less easy to determine. On the Geological Map of Norway (Holtedahl and Dons, 1960) and the map of Kildal (1970) it is given a "Caledonian" age. However, its similarity to mid-Proterozoic leucotroctolites elsewhere (see Chapter 3) and its traces of a granulite-facies overprint indicate a Proterozoic age for this also. These metagabbros and ilmenite-ores have for many years been referred to as "igneous rocks of the Bergen-Jotun Division" (eg Kolderup, 1928) implying a similarity to rocks of the mid-Proterozoic anorthosite-mangerite-rapakivi granite suite.

The structural features of the Hellevik-Flekke area are similar to those found elsewhere in the BGR, with early recumbant isoclinal folding followed by later, more open folding (eg Brueckner, 1977 for the Tafjord-Grotli area, about 150 km NE of the Dalsfjord area).

In short, the lithostratigraphy of the Hellevik-Flekke area is comparable with that of a large part of the BGR, particularly to the south of Nordfjord. Its structural history is also similar to other parts of the BGR. Most of the rocks seem to have had a Proterozoic history, which is in line with geochronological and geological evidence for neighbouring areas.

Two further lithological features of the Hellevik-Flekke area are comparable with those in other parts of the BGR. The intrusive websterites were shown to have some bulk-chemical similarities to the well known orthopyroxene eclogites of the coastal region to the north of Nordfjord (see Lappin and Smith, 1979 and Carswell et al, 1985). Green and Mysen (1972) reported the occurrence of websterites on the island of Kvamsoy, near Stadlandet, with small amounts of olivine, early turbid pyroxenes, interstitial garnet and local "saussuritised" plagioclase. Small bodies of enstatite eclogite occur in the same area. The descriptions of these websterites are reminiscent of the websterites near Gjörlanger and Flekkefjord.

Smith (1982) described veins of phengite + eastonite +/- albite with rare ziron and monazite within an orthopyroxene eclogite at Aarsheimneset, near Selje. These may be similar to the phengite-rich veins in the websterite by Gjörlangerfjord. Some orthopyroxene eclogites have pyroxenes with very high $^{87}Sr/^{86}Sr$ ratios, possibly as a result of extensive interaction with the surrounding gneisses (Griffin and Brueckner, 1982). The evidence from Gjörlanger is consistant with such interaction if the mica veins are interpreted as anatectic backveins of the host gneisses.

The discovery of a leucotroctolitic basic-ultrabasic complex in the Hellevik-Flekke area encourages comparisons with other large basic bodies in the BGR. Schmitt (1964) described a large (0.4-1.0 km^3) layered basic body at Eiksundal, south of Alesund. The body now consists dominantly of hypersthene eclogite interlayered with diopside-bearing and diopside-free garnet peridotites as being enriched in Fe and Ti, with abundant oxide-phases. Normative compositions of the eclogites indicate that low-pressure igneous assemblages would have been dominated by plagioclase with subordinate olivine and hypersthene and small amounts of augite and oxides. Despite its rather higher metamorphic grade (750°C, 20 kbars - Carswell et al, 1983) it seems rather similar to the Flekke unit.

It is instructive to compare the derived peak metamorphic conditions for the eclogites with those elsewhere in the BGR. Krogh (1980) has studied eclogites in the nearby Naustdal area, near Förde. As was noted in Section 4.01 the barroisite-bearing Naustdal eclogite gives temperatures of 630 +/- 35°C at minimum pressures of 15.0 +/- 2.5 kbars while glaucophane-bearing eclogites (eg that at Kvineset) give rather lower temperatures of 540 +/- 35°C at 12.5 +/- 2.5 kbars (minimum P). It is also noteworthy that the Naustdal eclogite lies in the lower part of the Fjordane Complex (from the map of Kildal, 1970) while the lower temperature eclogites at Kvineset, Sande and Erdal lie in the Basal Gneisses. It, therefore, appears that equilibration temperatures of eclogites in the Basal Gneisses (Jostedal Complex of Bryhni, 1966) are significantly and consistantly lower than thoe in the overlying Fjordane Complex. Skjerlie and Pringle (1978) have called the Fjordane Complex rocks in this area the "Sunnfjord Nappe" on the basis of the thrust zone which forms its lower junction. Hence the temperature difference may have some tectonic cause. The temperatures are plotted on a simplified geological map of the Sunnfjord area on Figure 5.01.

A regional variation in $K_D^{Fe/Mg}$ for garnet and omphacite was discovered by Bryhni et al (1977) and Krogh (1977a) which indicated an increase in maximum equilibration temperatures towards the northwest with maximum values along the coast between Stadlandet and Kristiansand and isotherms parallel to the coast. Griffin et al (1985) have recalculated the K_D values for 59 localities as temperatures by the method of Ellis and Green (1979). Their map is reproduced in figure 5.02. They confirmed the northwestward increase in T and found that P increased sympathetically with T (a trend also noted by Krogh, 1977a). Maximum conditions were estimated at 800°C, 20 kbars at the coast, decreasing to 500°C, 12 kbars in Sunnfjord.

However, as noted by Griffin et al (in press) the distribution to the south of Nordfjord in the Sunnfjord area is somewhat more complex than a simple coastwards increase in P and T. First of all a number of major east-west faults transect the region and secondly, there seems to be a discontinuity in the trend across the lower boundary of the Sunnfjord Nappe. Hence the continuity of the isotherms across Nordfjord is in some doubt, and the simple straight 600°C contour of Griffin et al (in press) for the Sunnfjord area must be considered to be an oversimplification.

Finally, it is instructive to compare the garnet zoning profiles for the eclogites studied here with those elsewhere in the BGR. Garnets with "prograde" zoning (decreasing Mn and Fe, increasing Mg rimwards) like those described here are known from the Förde area (Krogh, 1980, 1982) and Nordfjord (Bryhni and Griffin, 1971; Krogh, 1980), where maximum equilibration temperatures lie in the range 550-700°C. Inclusions of Si-poor amphibole commonly lie in the cores of the garnets, along with clinozoisite, quartz, paragonite, plagioclase, biotite and K-feldspar (Krogh, 1980, 1982).

Eclogites in the westernmost coastal areas between Nordfjord and Kristiansund often have rather flat zoning profiles in garnets (Krogh, 1980). A typical example comes from Langavatnet, Moldefjord (data of M A Harvey in Cuthbert et al, 1983, appended in this Thesis). Mg/Mg + Fe has a flat profile, with a marked decrease bringing about 50 microns from the rim. These garnets tend to contain inclusions of omphacite, quartz and rutile throughout (Mysen and Heier, 1972; Krogh, 1982). Omphacite in these garnets tends to have higher jadeite concentrations than matrix grains (Mysen and Heier, 1972) whereas in the Sunnfjord eclogites omphacite inclusions are less jadeitic than, or similar to, matrix pyroxenes (Krogh, 1980, 1982 and this work). These features are taken to indicate that the matrix phases equilibrated at lower pressures than the inclusion assemblage in the coastal eclogites of Möre and Romsdal. The narrow "retrogressive" rims are typical of late diffusive re-equilibration during cooling (as opposed to growth zoning) (Tracy, 1982).

These diffusive readjustments make it unclear whether the inclusion assemblage equilibrated at higher or lower temperatures than the matrix phases. One can only conclude that their present "frozen-in equilibrium" indicates final lower temperatures before the system closed.

An instructive exception to the flat profiles of the coastal eclogite garnets comes from Haröysund, near Molde, as described by Krogh (1982). The garnets have weak "prograde" zoning with inclusions of Ti-rich, Si-poor hornblende, plagioclase, biotite and magnetite in the cores. A noteable feature of this eclogite is that the garnets are very coarse-grained (up to2.5 cm across). It can be concluded from this that at least some of the eclogites in this area have developed by prograde metamorphism of amphibolites (as suggested for eclogites to the south and southeast) but that the high temperatures have generally wiped out the zoning by encouraging more rapid diffusion and reaction rates so that only very coarse-grained rocks have retained the early assemblages. Estimates of equilibration temperatures for the coastal eclogites of Möre and Romsdal are generally in the range 700-800°C (Griffin et al, in press).

The distribution of zoning types can be related to the temperature distribution from the foregoing. Disequilibrium of REE between garnet and omphacite also seems to be related to lower equilibration temperatures. Humphries and Cliff

(1982) have attempted to calculate closure temperatures for the Sm-Nd system for grossular and pyrope-rich garnets. Maximum closure temperatures (for very rapid cooling) indicated closure temperatures between 600-775°C, with higher temperatures for more grossular-rich garnets. Humphries and Cliff (1982) also estimated the closure temperature for Fe-Mg diffusion in garnet from the data of Freer (1981) as 725 +/- 30°C for a relatively slow cooling rate of 5°C Ma^{-1}. These results are consistant with the distribution of temperature and zoning types in the BGR, indicating that zoning will tend to have been eradicated at temperatures in excess of about 700°C at suitable cooling rates and length scales for diffusion. A consequence of this is that to the northwest of the 700°C isotherm calculated equilibration temperatures will be minimum values related to closure of the system. Further southeast, and in Sunnfjord, temperatures will have been frozen in during garnet growth and are closer to actual temperatures. This is corroborative evidence that temperatures did not greatly increase during decompression of the eclogites described here, as this would have wiped out any zoning.

Section 5.03 Implications for the tectonothermal evolution of the BGC

5.03.1 "Foreign" versus "*in-situ*" metamorphism.

Before going on to discuss how the results set out in this Thesis can be used to unravel the tectonothermal evolution of the BGC, an appraisal is made in this sub-section of the features distinguishing eclogites which have been tectonically introduced into their host gneisses as solid intrusions ("foreign" eclogites) from those which have been formed by metamorphism within their present host rocks ("in-situ" eclogites). Recognition of either foreign or in-situ origin has important consequences for reconstruction of the tectonic history of an eclogite-bearing terrain.

The hypothesis that some eclogites have been tectonically emplaced within their present host rocks stems from the observation that they represent a much higher grade of metamorphism than their host gneisses, which usually consist of amphibolite-facies assemblages in western Norway. In addition, the contact between the eclogite and the gneiss is frequently sheared and a retrogressive shell of amphibolite surrounds the fresh eclogite core. Such features are characteristic of the eclogite "pods" which are commonly found in the BGR. Coleman et al (1965) used these features to define their "Group B" eclogites and noted that these rocks were "difficult to assess properly because they seem to be geologically out of place in their associated rocks".

The origin and possible mode of emplacement of "foreign" eclogites has been discussed by Lappin (1966, 1977), Lappin and Smith (1978), O'Hara (1976) and Smith (1980, 1981).

These authors favoured upthrusting ("deep level obduction") of basic and ultrabasic material with pre-existing eclogite-facies assemblages from the deep crust (Lappin, 1966; Smith, 1980, 1981), the upper mantle (Lappin and Smith, 1978; Smith 1980, 1981) or subducted oceanic crust (O'Hara, 1976). All these phenomena are thought to have occurred in the Western Alps (Rivalenti et al, 1981; Zingg, 1983; Ernst and Dal Piaz, 1978; Ernst, 1981) but such mechanisms are usually related to the emplacement of large, regionally mappable tectonostratigraphic units and not small bodies like the majority of eclogites exposed in the BGR.

In the long-standing controversy regarding the foreign versus in-situ origin for Norwegian eclogites the term "crustal eclogites" has often been used synonymously with "in-situ" (eg Krogh, 1977a). Evidence presented for "crustal" eclogite formation includes demonstration of prograde metamorphism of low-pressure protoliths (eg dolerites, amphibolites) to form eclogites (eg Bryhni et al, 1977; Krogh, 1977a). However, although this precludes an upper mantle origin for the eclogites, it does not exclude tectonic emplacement of oceanic crust. Krogh (1977b) suggested that the discovery of "blueschist" mineralogy (glaucophane eclogite) in Sunnfjord supported a "crustal" origin despite the fact that blueschists are well known for occurring as exotic blocks in lower-grade schists.

The ultimate aim of resolving the "in-situ" or "foreign" origin of eclogites is to determine whether or not the large volumes of host gneisses have been subjected to unusually high lithostatic pressures, along with the eclogites. The only evidence that can uniquely demonstrate this is:-

1) to show that the eclogite was autochthonous with respect to its host rocks prior to eclogite-facies metamorphism and/or
2) to demonstrate that the host rocks have assemblages of the same mineral-facies as the eclogite which they enclose.

Both these kinds of evidence have been demonstrated in this Thesis, along with corroborative evidence of prograde metamorphism of low-P protoliths. This adds to a growing body of evidence that a large volume of the BGC has been subjected to unusually high pressures, corresponding to depths of burial well below the normal moho depth of stable continental crust. The evidence for the rest of the BGC has been reviewed in Cuthbert et al (1983) which is appended as enclosure 2 here.

5.03.2 Tectonothermal evolution

A number of publications in recent years have attempted to model the thermal evolution of thickened continental crust (eg England and Richardson, 1977; Richardson and England, 1979; Rubie, 1984; England and Thompson, 1984). Using a simple model involving a doubling of crustal thickness along a single thrust fault, the following general predictions can be made:-

1) If thrusting takes place at plate-tectonic rates little increase in temperature can be expected while pressure increases in the underthrust plate (Oxburgh and Turcotte, 1974).
2) The thrusting will produce a perturbation of the steady-state continental geotherm by introducing relatively cold rocks to great depths (figure 5.03a).
3) Crustal thickening will tend to produce an elevated land surface due to the relative buoyancy of the crustal rocks. This will encourage erosion (and perhaps gravity tectonics) which will act to return the crust to its normal thickness (Anhert, 1970; England and Richardson, 1977).
4) A combination of heat flux from the mantle and radiogenic self-heating will act to restore an equilibrium steady-state geotherm to the thickened crust. This is referred to as "thermal relaxation".
5) The thermal history of any point in the thickened crust (its "P-T path") is a result of competition between the rate of thermal relaxation and the rate of uplift. The most important variables controlling these are the amount of crustal thickening, the rate and length scale of erosion, the supply of heat from the mantle and internal radiogenic energy and the thermal conductivity of the

crustal rocks (England and Thompson, 1984).

6) Modelling P-T paths for reasonable values of these parameters using the simple overthrust model indicates that shortly after reaching its maximum depth, a point in the lower plate will begin to decompress and at the same time, suffer an increase in temperature. The maximum temperature reached is found to be at some pressure significantly lower than the maximum pressure attained. As erosion brings the point closer to the surface cooling begins, which continues until exhumation is complete (England and Richardson, 1977). This simple model is illustrated in figure 5.03b, c.

7. Points deeper in the crust (which reach higher pressures) will take longer to reach the surface and tend to suffer more heating on the uplift path. The locus of temperature maxima of the P-T paths for rocks from various depths is known as the "metamorphic field gradient" (Spear et al, 1984).

8) The mineral assemblage which is preserved at the surface after exhumation is likely to be that which formed on the negatively sloping or isothermal segment of the P-T path, hence the maximum recorded pressures will be those which occurred near the temperature maxima. The recorded pressure may, therefore, be significantly lower than the true maximum pressure (England and Richardson, 1977; Thompson and England, 1984). However, conditions pre-dating T_{max} may be derived from mineral

inclusions in porphyroblasts or, if the closure temperature for diffusion was not exceeded, from zoning profiles. The relationship between P-T paths and types of mineralogical information is shown schematically in figure 5.03d. It is apparent that mineral assemblages giving "Peak" P-T conditions will define the metamorphic field gradient, and such a gradient can be defined within an area of significant structural relief.

9. Finally, the metamorphic field gradient may not simply be a result of the hysteretic P-T path described aove, but rather result from the convergence of thermal conditions in adjacent plates across the thrust surface. Hence hot material at the base of the upper plate will tend to cool perhaps almost isobarically prior to decompression, whilst cool material at the top of the lower plate will tend to be heated prior to or during decompression (Spear et al, 1984; Carswell and Cuthbert, in press). This is illustrated in figure 5.03e.

The P-T path derived from the eclogites of the Hellevik-Flekke area is significantly different from those derived from the simple numerical models outlined above in that the decompression path is effectively isothermal (within the bounds of resolution of mineral equilibria). Before attempting to ascertain what factors may be responsible for this difference, it is important to show that it is, in fact, a true P-T path.

Assuming that the petrographic analysis ana the mineral equilibria used to define the various points in the P-T plane are not grossly in error, it can be reasonably accepted that those points actually represent P-T conditions through which the rocks have passed. The possibility remains that some point on the P-T path is not represented in the mineral assemblages, for intance the maximum-pressure conditions as outlined in point 8 above.

As was discussed in Chapter 4, the inclusion suites in garnets and zoning of garnets constrains their pre-matrix history (pre-stage III). Omphacite inclusions (stage II) have similar jadeite contents to matrix omphacites (stage III) and pre-eclogitic amphiboles (stage I) define temperatures only slightly lower than stage II omphacites. The preservation of growth zoning implies that temperatures never rose high enough to obliterate relics of pre-eclogitic assemblages and mineral compositions. It can, therefore, be inferred that the history of these rocks is reasonably completely recorded and that T_{max} approximately coincided with P_{max}.

"Hairpin"-shaped profiles are, in fact, well known from eclogite-bearing terrains (Miller, 1977; Maresch and Abraham, 1980; Ernst, 1981; Lardeaux et al, 1982). Avoidance of the decompressional heating phase required by the simple overthrust models (England and Richardson, 1977) requires either a very low heat input into the system, low conductivity or very rapid rates of erosion and uplift.

Rubie (1984) has developed a model for the Sesia-Lanzo zone of the Western Alps in which temperatures are retarded by subduction of oceanic crust beneath the underthrust eclogite-facies terrain. Although the temperatures derived for the eclogites described here are rather higher than those of Sesia-Lanzo, the fact that considerable volumes of obducted oceanic crust overlie the BGC in the Dalsfjord area (Furnes et al, 1976) make this a viable mechanism for retarding thermal relaxation there. Rubie (1984) states that this mechanism removes the necessity for the very rapid uplift rates otherwise required to preserve eclogite-facies assemblages. However, the ubiquitous occurrence of symplectic breakdown products of omphacite and phengite, together with the geochronological constraints outlined in Section 5.01 suggest that uplift rates were quite fast. Cuthbert et al (1983) (enclosure 2) estimated a rate of approximately 1 mm yr^{-1}, which is similar to estimates for the present-day Himalayan belt (Zeitler et al, 1982).

The presence of eclogites in the Basal Gneisses of the Sunnfjord area which have equilibration temperatures rather lower than those in the Fjordane Complex (figure 5.01) may indicate that the Jostedal Complex consisted of a body of rather cool rocks which were overthrust by the .warmer Sunnfjord Nappe. This could also have retarded thermal relaxation of the BGC. However, such underthrusting would not have been as long-lived as oceanic subduction, so that the retardation would have been less effective (Rubie, 1984).

Even so, the blanketing effect of a wedge of cool underthrust continental crust in delaying the influx of mantle-derived heating may have been sufficient if uplift was rapid enough.

It was noted in Chapter 3 that the gneisses of the Gjörlanger Unit had the characteristic low Rb/Sr ratios typical of ancient granulite-facies terrains. Low levels of radiogenic heat-producing elements may have been instrumental in retarding thermal relaxation as radioactive self-heating would have been at low levels.

Hossack (1984) has found that the Devonian basins of the west coast of Norway probably formed by extensional movements along low-angle detachments. The detatchment surfaces appear to have been rejuvenated thrust-planes lying below the mangerite-syenite nappes such as the Dalsfjord Nappe. These nappes may form the tail end of the outcrop of the huge Jotun Nappe which lies to the south-east of the study area for this Thesis, on the other side of the Gaularfjell anticlinorium (Skjerlie, 1969; Hossack, 1983). Late extensional tectonics may have been a very effective way of removing the overburden to the BGC and hence encourage rapid uplift.

Summing up the conclusions for the Hellevik-Flekke area eclogites, it can be stated that their clearly demonstrable "in-situ" metamorphic origin implies that all the lithologies were buried to depths of about 55 kms. By analogy with evidence from other parts of the BGR this occurred during the

Caledonian Orogeny. The similarity of lithostratigraphy to that over wide areas of the BGR, particularly to the south of Nordfjord, indicates that this conclusion can be applied to very large volumes of rock. An "in-situ" origin for eclogites in the Basal Gneisses has not been demonstrated here, but Krogh (1980) reported intrusive or gradational contacts between eclogites and gneisses in a similar lithostratigraphic position in the Förde area.

The form of the P-T path for eclogites in the Gjörlanger and Flekke Units requires rapid exhumation and/or retardation of thermal relaxation. A number of features which can be observed in the Dalsfjord area are consistent with either of these mechanisms, but it is not certain which was the most effective. The presence of a possible structural and metamorphic break between the Basal Gneisses (Jostedal Complex) and the Fjordane Complex (Sunnfjord Nappe) may suggest that the BGC consisted of at least two major structural elements during the Caledonian orogeny, the latter having been thrust over the former.

The observed regional P-T variation across the BGR (Griffin et al, in press) corresponds to a metamorphic field gradient. For the sake of simplicity, it is assumed that the Sunnfjord eclogites conform to this gradient, but that their position has been affected by later movements. Krogh (1977a) suggested that the gradient was a result of continental collision in the Precambrian during which the BGC was overthrust by the margin of another continent. Evidence that the eclogite-facies metamorphism was Caledonian (Griffin and Brueckner, 1980) shows that the collision event was also Caledonian and the collision was a result of closure of the Iapetus ocean between Greenland and Baltica. Griffin et al (in press) have suggested that the P-T gradient resembles that for the top of the underthrusted slab in the Zagros collision zone and hence that the present erosion surface of the BGR, closely corresponds to an ancient plate surface. However, the exposed part of the BGC is a culmination which is likely to expose considerable structural relief. If this is so it is more likely that the gradient is a result of differential uplift of increasingly more deeply underthrust rocks, the most deeply buried being the most northwesterly. This implies that the peak metamorphic assemblages are diachronous, becoming older towards the coast. As the Jostedal complex is the lowest structural element in the BGC it is likely that increasing pressure and temperature corresponds to progressively higher structural levels. This apparent contradiction is solved if underthrusting was at a relatively steep angle and uplift hinged somewhere in the region of the Jotun Nappe.

A tectonic model for the metamorphic evolution of the BGC has been published by Cuthbert et al (1983 - enclosure 2 here), based on the petrological features of the BGC, regional tectonastratigraphy, analogies with present-day continent-collision zones and geophysical evidence for crustal structure. The reader is referred to the paper for the detailed discussion, but the important features are that the BGC appears to be allochthonous and its present position is a result of the thrusting of a crustal wedge beneath it. This would have provided the final impetus for uplift of the eclogites and allowed them to appear at the surface above a normal thickness of continental crust. The underthrust wedge (which may or may not correspond to the Jostedal Complex) would have blanketed mantle heat flux, allowing the effectively adiabatic uplift observed in the mineralogical P-T paths.

The preservation of the relatively low-temperature eclogites of Sunnfjord is, according to this model, a result of rapid uplift above an insulating crustal wedge allied to its relatively high position in the climactic collisional thrust-sheet assembly. The coastal eclogites of Möre og Romsdal were more deeply buried and hence took longer to uplift. As a result they suffered more thermal relaxation, generally erasing prograde garnet zoning as a result of the high temperatures attained.

The present-day preserved metamorphic field gradient may be a result of differential uplift tilting the underthrust plate during its isostatic recovery. This tilting would have been a direct result of the underthrusting of buoyant crustal material below the emerging BGC in a similar manner to that obserbed in the Himalaya (Windley, 1983; Carswell and Cuthbert, in press).

Section 5.04 Recommendations for further research

The Hellevik-Flekke area has proved to be a useful area for studies of the BGC as it has a wide range of lithologies and disequilibrium on all scales provides a window through which to view a long and complex crustal history.

The most obvious requirement for further research is more fieldwork. Indeed this is a major requirement for most of the BGR as fieldwork to date is at best patchy and at worst based on outdated lithostratigraphic and structural principles. The major tectonostratigraphic units of the BGR have yet to be correlated regionally and the overall structural history of the area, except in the east near Oppdal, is very poorly described in the literature, generally being only at the reconnaisance stage. Tectonic models of the BGC based on eclogite petrology can only be regarded as working hypotheses until a coherent regional geological picture has been developed.

The recent work of Krill (1980) and Gee (1980) in working out the relationships of the BGC and the overlying allochthon in the Oppdal-Surnadal area has resulted in significant advances in knowledge of the evolution of the Scandinavian Caledonides. However, a similar tectonostratigraphy outcrops near the west coast in the Dalsfjord area and this has received very little attention, despite the fact that the lower Palaeozoic schists are intimately involved with eclogite-bearing gneisses. A detailed study of the relationships of the schists to the gneisses should provide useful data on their comparative metamorphic histories and may help to resolve the nature of the overburden to the BGC during eclogite formation.

The reconnaisance structural studies carried out for this Thesis indicate that microstructural studies of fabrics in relation to mineral growth have promise. This is because of the common occurrence of high-pressure parageneses in the gneisses. The phengite-bearing paragneisses of the Vardheia Unit are particularly useful in this respect as they should be relatively easy to date isotopically with the Rb-Sr method, hence avoiding disequilibrium problems with Sm-Nd encountered with eclogites.

A number of mineralogical studies can be suggested. First of all, wet chemical determination of Fe^{3+} in minerals used for Fe-Mg exchange thermometry would give more confidence to temperature estimates. Secondly, modelling of amphibole breakdown at high pressures should better define the amphibolite-eclogite transition. Estimation of amphibole-garnet-pyroxene equilibria may aid quantitative modelling of the garnet zoning to provide estimates of cooling rates. Microstructural studies of omphacites, particularly those showing exsolution features should also aid in determination of P-T-t paths. More analyses of pyroxene-pairs in granulite-facies assemblages should improve the precision of pyroxene thermometry and Fe-Ti oxide thermometry on samples from Sördal would provide further points on the P-T trajectory.

The possibility that the Basal Gneiss eclogites formed at lower temperatures than those from overlying units deserves further scrutiny. It should be possible to map T estimates in the area in order to better resolve the discontinuity.

Finally, the geochemical studies in this Thesis have tended to be of a comparative nature. It should be possible to use the bulk-rock chemical data (particularly trace-elements) to model the petrogenesis of a number of the lithologies described.

While it is obviously desirable to quantify parameters involved in modelling orogenic processes with analytical laboratory work, the major conclusions of this Thesis are based on detailed fieldwork and petrographic studies. It is the firm belief of this author that the Dalsfjord area still has much to offer the field geologist. Increasingly sophisticated work on the eclogites of the BGR has become the focus of activity on the area in recent years at the expense of basic geological groundwork. A period of consolidation is now required which concentrates on the less spectacular, but volumetrically vastly more abundant gneisses and other lithologies. Only in this way can arguments on the history of the area become less polarised (see Harte, 1983) and models for its origin become more objective.

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Appendix 1 Analytical methods

In this appendix brief descriptions of the laboratory methods used to obtain bulk-rock and mineral analyses are given.

A1.1 "Wet chemical analysis

A1.1.1 Sample preparation

All bulk-rock analysis was performed on powdered rock. Sample sizes varied from approximately 0.2-1.0 litres of the solid rock, with larger sample sizes for coarser-grained rocks wherever possible.

Samples had weathered surfaces removed with a hammer or diamond-saw. They were then crushed into cm-sized chips with a jaw-crusher, thoroughly mixed and divided and a large handful-sized split was crushed to powder in a "TEMA" tungsten-grinder for about 20 seconds. Long crushing times were avoided to prevent sample oxidation. Samples were stored in screw-top glass jars.

Powders used for analysis were oven-dried at 110°C for 24 hours and stored in a dessicator jar.

A1.1.2 Determination of combined water

"Penfields method" was used. About 0.5g of sample was placed in a pre-weighed glass Penfield tube, weighed to determine the sample weight, then strongly heated over a Meker burner for 10-15 minutes. The water driven off was condensed on the cooler end of the tube, aided by a collar of wetted tissue-paper wrapped around the tube. The tube was drawn into two parts and the closed end containing the sample discarded. The remainder of the tube was stoppered to prevent evaporation, allowed to cool, then weighed. The tube was oven-dried at 110°C for 1-2 hours, cooled and re-weighed to determine the amount of water. Combined water (H_2O^+) is given by:-

$$\%H_2O^+ = \frac{\text{weight of water (g)} \times 100}{\text{sample weight (g)}}$$

Precision of this method was poor, partly due to loss of water by evaporation. Some low values in pyritic samples may have been due to retention of water on sulphate sublimates on the inner walls of the tube. Errors are estimated at +/- 20%.

A1.1.3 Determination of CO_2

Samples were first tested for carbonates by mixing with dilute HCl in a test tube to see if any gas was evolved. If any effervescence occurred the sample was analysed. Samples with the prefix "79/" were analysed by absorption by soda asbestos. Samples with the prefix "D" were analysed by non-aqueous titration.

For the absorption method about half a gram of sample was heated in a bulb-flask containing 25 ml $H_3P_2O_4$. The gas evolved passed through a condenser into the absorption train: a flask of sulphuric + chromic acid mixture absorbed any evolved water; a U-tube, one half filled with magnesium perchlorate, the other half with manganese dioxide, removed sulphates, and finally two U-tubes, each filled with one third calcium chloride and two thirds soda-asbestos trapped the CO_2 (the $CaCl_2$ removed any remaining traces of water). To encourage complete removal and absorption of Co_2 a small electric air-pump drew the gases through the train. The two soda-asbestos tubes were weighed before and after each run and any increase in weight calculated as evolved CO_2:-

$$\%CO = \frac{\text{weight increase of tubes (mg)} \times 100}{\text{sample weight (mg)}}$$

The $CaCl_2$ and soda asbestos were changed if the second tube showed any significant weight increase.

The non-aqueous titration method (Carbon Method 4 in BS Handbook No 19 (1970)) evolved CO_2 in the same way as above. The gases were passed into a flask containing 150 ml dimethylforamide, 5 ml monoethylamine and 2 ml 0.1%

thymolphthalien indicator. Passage of the gas was aided by an electric air pump and a U-tube filled with soda-asbestos placed before the reaction vessel prevented influx of atmospheric CO_2. Between the reaction-vessel and the titration flask an absorption train of silica gel, calcium chloride, magnesium perchlorate and magnesium dioxide removed any H_2O, SO_2 or SO_3. The air output was cleaned of the toxic organic fumes with a charcoal tower.

Before starting the reaction the system was flushed through with air to remove any CO_2. The sample was reacted and the CO_2 bubbled through the titration mixture. Prior to reaction, the mixture had been titrated to produce a blue colour. Addition of CO_2 clears the colour. After reaction (about 10 minutes) the mixture was titrated bck to the blue end-point. Air was drawn through the system for a further 5 minutes to ensure complete evaluation of CO_2 with further titration if necessary. The:-

$$\%CO_2 = \frac{\text{volume of titrant (ml)} \times 4.4}{\text{weight of sample (g)} \times 10}$$

The non-aqueous titration method was quicker than the absorption method, but great care was needed with the hazardous chemicals used. The non-aqueous titration was more precise, but within error gave results comparable with the absorption method for the sample.

A.1.1.4 Determination of ferrous iron

The cold digestion method of Wilson (1955) was used. Ferric iron was calculated by subtraction of FeO from the total iron measurement obtained by XRF analysis (see below).

About half a gram of rock sample was placed in a polythene screw-top bottle with 10 ml 40% HF and 4 ml ammonium metavanadate solution. This was then put aside for about three weeks (experiments showed that for garnet or spinel-bearing samples up to 19 days digestion was required to fully react the sample). Reaction of metavanadate with Fe^{2+} reduces V^{5+} to V^{3+}. The original V^{5+} in the reagent was titrated against a standard $Fe(NH_4)SO_4.6H_2O$ solution. After digestion the samples were neutralised in a solution of HBO_3 and titrated to determine the remaining amount of V^{5+}. FeO was calculated by:-

$$\%FeO = \frac{V_B - V_R}{\text{Sample wt (mg)}} X\ C$$

where V_B = volume of titre for blank NH_4VO_3 reagent, V_R = volume of titre for digested sample and C is determined by titration of the ferrous ammonium sulphate against potassium dichromate, such that:-

$$C = \frac{\text{vol potassium dichromate titre (ml)}}{25}$$

Some Fe-Ti oxide-rich samples were very iron-rich. For these samples twice the amount of NH_4VO_3 was used. Some of these samples contained significant V (up to 4500 ppm). No correction was made for this as these amounts were considered negligable compared to the total amount of ferrous iron in the samples.

Precision for ferrous iron determination is considered to be better than +/- 10% on the basis of repeated analyses.

A1.1.5 Determination of sodium

Sodium was determined by atomic absorption spectrometry as the Phillips 1220 X-ray fluorescence spectrometry gave poor precision for Na on the Cr-tube and $NaNO_3$ was used as an oxidant in XRF sample preparation.

Solutions for atomic abosorption analysis were prepared by fusion of samples in lithium metaborate. About 0.05 grams of sample were mixed with 0.3 grams lithium metaborate powder ina pre-fired graphite crucible. The mixture was fused in a muffle furnace at 1050°C for 15 minutes. The molten sample was carefully emptied into a 300 ml plastic bottle containing 15 ml distilled water, 5 ml concentrated HCl and a magnetic stirrer. The mixture was placed on the stirrer for about ½ hour or until the fusion pellet has dissolved. 5 ml of 10% KCl was added and the mixture made up to 250 ml in a graduated flask. A blank and standard was run with each set of samples.

The Perkin-Elmer atomic absorption spectrometer was auto-zeroed with a blank, then calibrated with appropriate standards (BCS 309, W1, G1, G2 and/or DRN). Results were read off as weight % Na_2O. An average of three readings for each sample was taken.

A1.2 Instrumental methods

A1.2.1 X-ray fluorescence spectrometry

The major elements Si, Al, Ti, Fe, Mg, Ca, K, Mn, P and S were determined on fused glass discs using a technique modified after Norrish and Hutton (1969). The flux used (Spectroflux 105) contains the heavy X-ray absorber lanthanum oxide which helps to reduce the variation in mass absorption coefficients in different rocks. Analysis was carried out on a Phillips 1220 spectrometer with a Cr-tube under standard operating conditions. Background intensities were measure on a blank of spec-pure SiO_2 prepared in the same way as the samples.

Specimens were prepared as follows:-
0.373 g of sample was mixed with 0.027 g
$NaNO_3$ and 2.0g of flux and fused in a Pt-Rh-Au crucible at 1000°C for 10 minutes. The mixture was cooled, re-weighed and the ignition loss calculated, allowing for the decomposition of $NaNC_2$. The mixture was then re-fused and poured onto a graphite disc on a hot plate at 230°C and

quenched and molded into a 31 mm diameter disc of glass with an aluminium plunger. Discs were polished with a cloth prior to analysis to remove any surface hydration bloom.

Samples prefixed "79/" were analysed for the trace elements V, Cr, Mn, Ni, Cu, Zn, Rb, Sr, Y, Zr, Ba and Pb using self-supporting pressed powder pellets on the Phillips 1220 spectrometer with a tunsten X-ray tube. Undiluted pellets avoided the problem of the low intensities for trace elements and interference by the La absorber on the heavy elements.

Samples prefixed "D" were analysed using glass discs prepared with lithium borate flux (Spectroflux 100A). Sample discs were prepared in a similar way to that used for major elements except that no oxidant is used, 0.5 g of sample was used and 2.0 g of flux. Siliceous samples tended to be rather brittle and were strengthened by backing with an adhesive paper disc to avoid loss of any samples in the spectrometer.

The samples were analysed on a Phillips PW 1400 automated X-ray fluorescence spectrometer using a rhodium tube. Results were in good agreement with those done on pressed powder pellets.

A1.2.3 Electron microprobe analysis

Mineral analysis was carried out on a Cambridge Instruments Microscan IX automated electron-probe microanalyser. Samples were carbon-coated polished thin sections. Samples were analysed with a beam diameter of approximately 2 microns with an accelerating voltage of 15 kV. Standards were natural and synthetic minerals and pure metals. The instrument was calibrated before running each set of samples and calibration was accepted if precision was better than +/- 1% for all elements. Counts were corrected for X-ray absorption, characteristic fluorescence and atomic number effects (ZAF correction) according to the method of Sweatman and Long (1969). Elements analysed were Si, Ti, Al, Cr, Fe, Mg, Mn, Ca, Na, K, Ba and V. As total Fe was determined as FeO some method of estimating ferric iron is required. This is discussed in Appendix 2.

A1.2.4 Rb-Sr Isotopic Analysis

This work was performed at the Mineralogisk-Geologisk Museum, Oslo.

Samples were prepared by weighing 500 g of powder into a 100 ml Teflon beaker with a little distilled water. A number of acids were successively added and evaporated off:-

20 ml HCl; 5 ml HNO_3 plus 5 ml HF; 5 ml 6M HCl. 3 to 5 ml 2.5 HCl were then passed through a series of ion-exchange

columns to separate the Sr. The sample was loaded onto a tungsten filament and placed in the spectrometer.

Analyses were performed on a VG Micromass 30 mass spectrometer. Peaks for ^{88}Sr, ^{87}Sr and ^{86}Sr were measured. Variable mass discrimination in ^{87}Sr/^{89}Sr was corrected by normalising ^{86}Sr/^{88}Sr = 0.1194 (Faure and Hurley, 1963). Regression lines were calculated using the technique of York (1969) embodied in the 150CH V02-02 computer program available at the Museum.

Appendix 2 Mineral Recalculation Schemes

This appendix outlines the methods by which mineral analyses were re-cast into structural formulae. As microprobe analysis does not discriminate between ferrous and ferric iron some method of estimating amounts of these must be applied to iron-bearing minerals, particularly those used in thermobarometry. These methods are discussed in the relevant sections and comments are made on the likely uncertainties which they introduce into thermobarometry.

A2.1 Garnet

Coleman et al (1965) noted that eclogite garnets have generally low contents of ferric iron. Scrutiny of "wet" chemical determinations for eclogite garnets from western Norway (eg Eskola, 1921) indicates that ferric iron contents are generally less than 1.5 mass %. Ferric iron was estimated here by assuming stoichiometry and using a simple charge-balance calculation. The generally low Fe^{3+} values are consistant with the comments made above.

Microprobe analyses of garnets tended to give rather high analysis totals, although despite this the structural formulae were generally close to stoichiometry. This may be due to the use of a more pyrope-rich standard, which is not closely similar in composition to the rather almandine-rich garnets studied here.

Garnet end-members, which were used in the ternary garnet diagrams, were calculated by the method of Rickwood (1968). This method attempts to maximise the cations allocated to end-member molecules and tends to leave atoms which may form mineral inclusions unallocated (Fe^{2+}, Fe^{3+}, Ti, H). The calculation order for the elements analysed in this work is:-
Uvarovite, andradite, pyrope, spessartine, grossular, almandine, schorlomanite.

The Fe^{3+} and end-member calculations were performed by a FORTRAN IV computer program written by D A Carswell.

Errors in Fe^{3+} (and, therefore, Fe^{2+}) are largely due to erros in Si analysis. As the garnets are fairly iron-rich (usually more than 20% FeO) these errors will generally be relatively small, producing virtually negligable uncertainty in K_D values for garnet and coexisting phases.

Analysis quality was judged by the number of unallocated cations left after the end-member calculation, as suggested by Rickwood (1968). Total allocated cations was generally better than 99%.

A2.2 Clinopyroxenes

Omphacites analysed for this study are rather iron-poor (generally less than 7% total iron as FeO). Also the number of Al^{VI} cations (Al^{VI} = al - (2-Si)) is usually less than the number of Na cations, indicating the presence of some acmite molecule. As omphacites from low-temperature eclogites are often chloromelanites one might expect that a significant proportion of the total iron might be ferric, which has important consequences for geothermometry.

A number of methods have been proposed for estimating Fe^{3+} in pyroxenes, based on charge balance (Ryburn et al, 1976) or end-member molecule calculations (eg Cawthorn and Collerson, 1974). Charge-balance methods have the disadvantage of being very sensitive to uncertainties in Si analysis, often producing large uncertainties in Fe^{2+} values in low Fe pyroxenes. Furthermore, charge-balance calculations are often done on the basis of 4 cations per formula unit, but, as Cawthorn and Collerson (1974) pointed out, omphacites often have cation totals deviating from 4, causing possible erroneous estimates of the charge deficit.

Two methods were investigated here. The method of Cawthorn and Collerson (1974) (hereafter called "CC") calculates the end-members in the order:-
jadeite, acmite, Ca-Ti-tschermak, Ca-tschermak, wollastonite, enstatite, ferrosilite.

Fe^{3+} is equal to acmite, or the amount of Na left after jadeite has been calculated. This has the advantage that it does not depend on the silica analysis for calculation of Al^{IV}, with its knock-on effect for Fe^{3+}. The method requires that acmite and tschermak's are mutually exclusive. This may have some crystal-chemical basis, as Rossi et al (1983) have found that bond-distance studies on omphacites mitigate against the presence of any tetrahedral Al. Si contents of most of the omphacites in eclogites studied here are in excess of 1.98, suggesting that tetrahedral Al is very low. However, Si contents from probe analysis are consistently slightly less than 2.000, rather than having some statistical variation around that number, perhaps indicating the presence of small amounts of Al^{IV}. Furthermore, exclusion of Al^{IV} at the expense of acmite may mask variations in tschermak's molecule in transitional augite/omphacites such as those found in the coronites. Finally, the CC method trends to discard iron if it does not fit the stoichiometry, which may significantly effect the Fe/Mg ratio.

The other method investigated was that of Mysen and Griffin (1973), modified after Essene and Fyfe (1967). In this method, Al^{IV} is added to Si to bring Z = 2.000. Then end-members are calculated in the order:-

Ca-Ti-tschermak's, Ca-tschermak's, jadeite, ureyite, acmite.

Acmite is calculated as Na-Al^{VI} + Al^{IV} after Ca-Ti-tschermak's has been calculated. The calculation is performed on the basis of 6 anions, so that allocation of some iron as Fe^{3+} renders this basis invalid. Hence the calculation is reiterated until there is no further change in Fe_2O_3 at the 0.1% level.

Mysen and Griffin (1973) considered that calculated $Fe^{3+}/Fe^{3+} + Fe^{2+}$ was within 20% of analytical values for a composition of 50 eclogite pyroxenes.

As this method allowed calculation of Ts and Ac together, unlike the CC method, and as it makes no assumptions about the number of cations, it was preferred and used for all clinpyroxenes in this study. Uncertainty due to Si analyses was considered a necessary evil in the search for systematic variations in tschermak's molecule. In general, for good quality analyses, the Mysen and Griffin and CC methods gave comparable Fe^{2+} values, with CC Fe^{2+} being slightly higher and jadeite also being higher. This would tend to give higher pressures and higher garnet/cpx temperatures. Comparison of the two methods was used as a basis for assessing analytical quality. Calculations were performed on a FORTRAN IV programe "PYROX" written by D A Carswell.

For the worst case of error in Si being 2%, error in temperature estimates by the Ellis and Green (1979) geothermometer was up to +/- 100°C. Precision of Si calibration was usually closer to 1% (on a jadeite standard), so that realistic uncertainty was probably closer to +/- 50°C.

A2.3 Orthopyroxenes

Orthopyroxenes were recalculated into structural formulae on the basis of 6 oxygens. All iron was assumed to be ferrous. Trials with the Mysen and Griffin (1973) Fe^{3+} calculation and charge-balance considerations tended to confirm this conclusion.

A2.4 Amphiboles

Simple charge-balance calculations for Fe^{3+} cannot be used for amphiboles due to uncertainties in the amount of (OH) and the possible vacancy in the A-site. In fact an estimate of Fe^{3+} was not required for geothermometry (method of Graham and Powell, 1984) providing that the amphiboles were sufficiently similar to those in the calibration data set. Howeer, some estimate of Fe^{3+} was necessary to help characterise the amphiboles, as ferric iron can significantly affect the site occupany, in particular the A and B site populations.

The approach used here is that of Robinson et al (1982), in which the amphibole analysis (calculated to cation proportions on the basis of 23(O,OH)) is normalised to one of a number of cation totals. Each normalisation is valid for certain types of amphibole (see below). The number of positive charges for the chosen normalisation is totalled, treating all Fe as Fe^{2+}. Charges are then added to bring the total to 46, and for each added charge one Fe^{2+} atom is converted to Fe^{3+}. The mineral formula is then calculated in the order:-

T-site, C-sites, B-site, A-site.

The formula is then evaluated according to the following crystal-chemical limits:-

K : A-site only
Na : A-site and M4(B) site only
Ca : M4 site only
Mn : C-sites (M1, M2, M3) or M4
Fe^{2+} : C-sites or M4 site
Mg : C-sites or M4 site
Fe^{3+} : C-sites only
Ti : M2-site only
Al : M2 or T-sites
Si : T-sites only

If any of these limits were violated the normalisation was discarded and another tried and so on until one (or more) satisfactory result appeared. If more than one normalisation was good the results were averaged if they were in close agreement, or some objective criterion was used to choose one or the other recalculation.

The normalisations used were as follows:-

1) All iron as FeO: this almost never worked.
2) Total cations to 16: (all sites filled). This might be

suitable for very taramitic amphiboles.

3) Total cations = 15 exclusive of Na, K: This excludes all Na + K from the B-site and excludes Ca from the A-site, also maximising Mn, Fe and Mg in M4. This works well for FeMg amphiboles.
4) Total cations = 15 exclusive of K: This excludes Na from A, forcing it into B. This works well for glaucophane and was often successful for low Na actinolitic amphiboles, where Ca is not forced into C by a lot of Na.
5) Total cations = 13 excluding K, Na, Ca: this was successful for many calcic and calcic-sodic amphiboles, dividing Na between B and A. It excludes Mn, Fe^{2+} and Mg from B, eliminating the "cummingtonite" component.

As with the pyroxene recalculations, this method is highly sensitive to variability in silica analysis.

The formulae were calculated with a BASIC computer program written by this author.

A2.5 Micas, chlorites, talc

Micas were recalculated to $22(O,(OH)_2)$. Attempts to estimate ferric iron (eg method of Laird and Albee, 1981) gave very erratic results, often with negative ferric iron. It may be that a similar approach to that used for amphiboles could be more successful. However, the problems of silica analytical precision are exacerbated by the possibility of beam damage volatilising Na, K and bound water.

Talc analyses were recast on the basis of $22(O(OH)_2)$ with all iron as FeO. Cation totals were usually greater than 13.97, close to the ideal value of 14.

Chlorite analyses were recalculated to $28(O(OH)_2)$ with all iron ferrous. Cation totals were between 19.81 and 20.40, usually close to 19.99, in good agreement with the ideal value of 20.00.

A2.6 Epidote group

Zoisites and epidotes were recalculated to structural formulae on the basis of $12.5(O(OH)_2)$ with all iron as Fe^{3+}. Si analyses tended to be slightly high and cation totals slightly low (average 7.96), although some analyses were more satisfactory, for no obvious reason.

A2.7 Oxides

Oxide analyses were recalculated on the basis of both the ilmenite-haematite (R_2O_3) and the magnetite-ülvöspinel (R_3O_4) formulae, with ferric iron calculated by charge balance for each formula. The method giving the most satisfactory formula, with the analysis total closest to 100%, was selected and the other discarded. The calculations were performed on a FORTRAN IV computer program written by F G F Gibb.

A2.8 Feldspars

Feldspars were recalculated to 8 oxygens with all iron ferric. Totals ranged from 4.96 to 5.06 in reasonable agreement with the ideal value of 5.00.

A2.9 Olivine

Olivine analyses were recast into structural formula on the basis of 4 oxygens. Cation totals varied from 2.99 to 3.01 in close agreement with the stoichiometric value of 3.00. Analytical totals were often slightly high, probably due to their rather iron-rich nature in comparison with the standard olivine, which was very forsteritic.