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'On the eclogites of Norway'—65 years later

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Abstract

The Western Gneiss Region (WGR) of Norway consists largely of Proterozoic orthogneisses, but also contains paragneisses, peridotites, anorthosites, gabbros and coarse-grained intermediate-acid 'rapakivi granites'. All of these lithologies enclose eclogites. Structural and isotopic data suggest that many of these rocks were juxtaposed by early Caledonian thrusting prior to eclogite formation at *ca.* 425 Ma.

Low-P protoliths can be demonstrated for many eclogites. Prograde metamorphism to eclogite facies is demonstrated by inclusion suites within garnet grains and zoning of eclogite minerals. The regional distribution of K_D (gnt/cpx) and X_{jd}^{cpx} shows a decrease in T_{max}, and in the corresponding P, away from the present coastline. The lowest values (500 °C, 10 kbar) are found in the Sunnfjord area and the highest ($\sim$ 800 °C) along the coast of Sunnmøre and Nordmøre.

Maximum pressures were reached at temperatures 100–200° < T_{max}. This P–T–t path is consistent with the preservation of jadeite-rich cpx (and possible coesite) in the coastal regions. Earlier overestimates of P_{max}, based on partitioning of Al between opx and gnt, resulted from combination of early low-T (low-Al) opx and T values derived from cpx/gnt equilibrated at T_{max}. Despite pervasive later amphibolitization, high-P assemblages (phengite + kyanite, omphacite + quartz) are locally preserved within gneisses near the coast. The high-P metamorphism can be explained by westward subduction of the Baltic continental plate beneath the Greenland plate, during the Caledonian orogeny.

At least some of the Mg-Cr garnet peridotites of the WGR were derived from low-P protoliths (spinel $\pm$ chlorite peridotites, enclosing high-Al pyroxenites). While Sm–Nd mineral ages of most eclogites cluster around 425 Ma, garnet peridotites and their enclosed garnet pyroxenites give Proterozoic Sm–Nd mineral ages (1700–1000 Ma). The tectonic position of the Mg-Cr garnet peridotites, relative to the Caledonian high-pressure metamorphism, remains to be resolved.

KEYWORDS: eclogite, gneisses, metamorphism, peridotites, Norway.

Introduction

SIXTY-FIVE years ago, Eskola (1921) published a classic treatise which described several types of eclogites from the Western Gneiss Region (WGR) of Norway, and clearly recognized their high-pressure origin. Since then, eclogite studies have grown into a minor industry, especially after the advent of plate tectonics underlined the importance of eclogites for understanding tectonic processes at destructive margins. However, relatively little was done on Norwegian eclogites for nearly 40 years after Eskola. O'Hara and Mercy (1963) studied garnet peridotites in the WGR, and concluded that they were tectonically introduced fragments of the upper mantle. By inference, this conclusion was extended to all of the eclogites of the WGR (O'Hara *et al.*, 1971; Lappin, 1966; Lappin and Smith, 1978), thus leading to an interesting controversy.

During the last 10–12 years, extensive work has been carried out on the Norwegian eclogites, especially by staff, students and many visiting researchers at the Mineralogisk-Geologisk Museum, Oslo (where Eskola did his work). This effort has led to an overall synthesis with three major conclusions: (1) most (all?) of the eclogites were derived from low-pressure protoliths; (2) the high-pressure metamorphism occurred essentially *in situ*, in a

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continental collision zone similar to the present Himalayas; (3) this metamorphism was part of the Caledonian (500–350 Ma) orogeny.

A summary of the data and arguments up to 1981 is given by Griffin *et al.* (1985). That summary is now partly outdated, due to the long publication delay and the rapid pace of research. This short review is intended to bring the 1981 version up to date, and to serve as a guide to some recent literature. Other references are given by these papers and by Griffin *et al.* (1985). No attempt will be made to cover some specialized topics, such as the detailed mineralogy of the eclogites, which do not bear directly on the synthesis presented here.

Occurrence of eclogites

Eclogites are now known from many areas of the Scandinavian Caledonides (Fig. 1), and studies from all of these areas have contributed to this synthesis. The nappes of the Scandinavian Caledonides have been grouped into the Lower, Middle, Upper and Uppermost Allochthons (Roberts and Gee, 1985). The eclogites of northern Norway and Sweden occur in the Uppermost, and

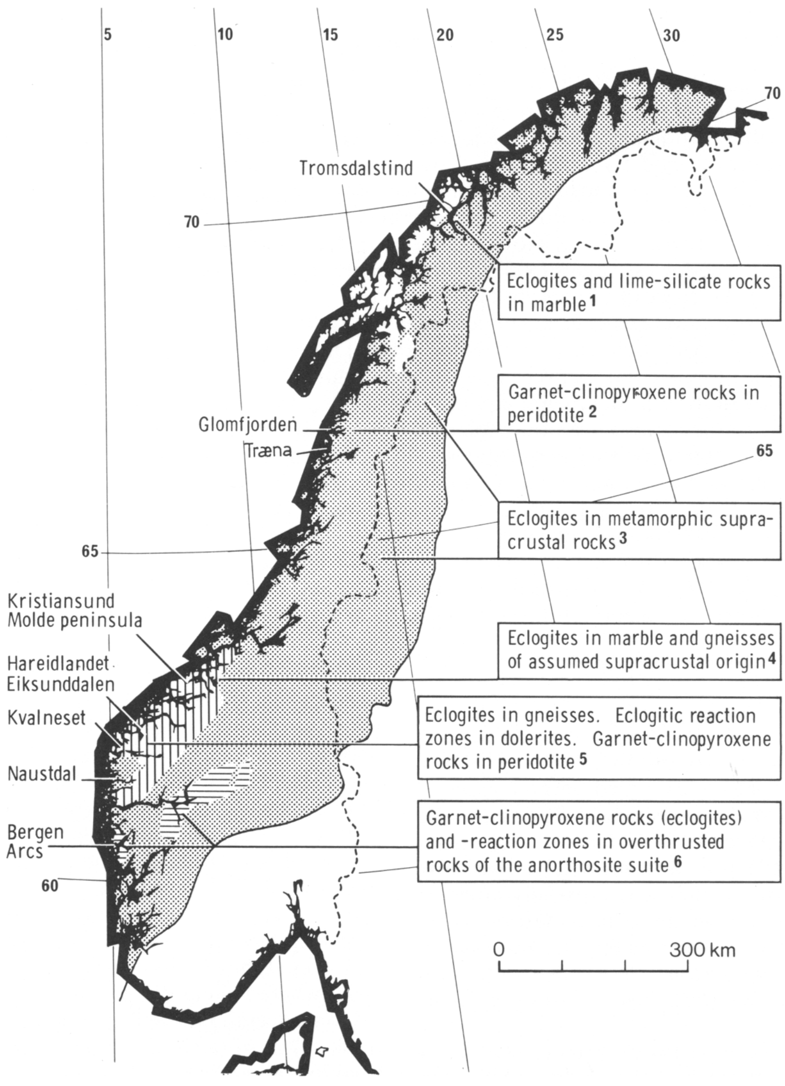

FIG. 1. Occurrences of eclogites and related rocks in the Scandinavian Caledonides.

Middle Allochthons (Roberts and Gee, 1985; Krogh *et al.*, 1987), while those in the Bergen Arcs (south-western Norway) are in levels corresponding to the Middle Allochthon. The classic WGR area represents an autochthonous or parautochthonous basement culmination in the middle of the orogen. An important observation is that eclogites are found only *within* this orogen. The eclogites in the northern Scandinavian nappes appear to have formed under similar conditions and by similar processes, to those in the WGR (van Roermund, 1985; Andreasson *et al.*, 1985; Krogh *et al.*, 1987) and this review will concentrate on the WGR.

The WGR is a heterogeneous terrain, dominated by intermediate migmatitic orthogneisses. Within these gneisses occur lesser amounts of paragneisses (including marbles and quartzites), anorthosites, augen gneisses (some derived from rapakivi granites), gabbros and peridotites. Eclogites occur in all of these rock types. The terminology used for the eclogites reflects the attention originally paid to the garnet peridotites: eclogites within the peridotites are generally called 'internal', while those enclosed by other rock types are 'external' or 'country-rock' eclogites. It is essential to distinguish these two groups; much confusion has been caused by extending conclusions drawn from study of 'internal' eclogites, to infer the origin of 'external' eclogites. As will be shown below, the two types probably are of different age, as well as origin.

Carswell *et al.* (1983) distinguished two types of garnet peridotite in the WGR. An 'Fe–Ti' type (olivine Fo_{67-82}) forms layers in layered eclogite–garnet websterite bodies of the Eiksunddal type (Schmidt, 1963; Jamtveit, 1986), as well as small isolated pods. The 'Mg–Cr' peridotites are more widespread, forming bodies from metres to kilometres in diameter. Garnet peridotite typically occurs as small relict areas, surrounded by and grading into a retrograde chlorite–amphibole peridotite (Medaris, 1980, 1984; Carswell, 1986). The classic 'internal eclogites' occur in two ways: garnet clinopyroxenite/garnet websterite layers grading into garnet peridotite, and as sharply bounded, dike-like bodies of eclogite or garnet websterite ($\pm$olivine). Carswell (1981) gives details of these occurrences in the Almklovdalen peridotite. A third type of 'internal eclogite' was recognized by Griffin and Qvale (1985); these are boudinaged dykes of superferrian (FeTi) basalt composition. Dike-like bodies of similar composition have been reported in the peridotites of Bjørkedalen (Brastad, 1983, 1985) and Gurskebotn (Jamtveit, 1984).

The typical or classic 'external' eclogites form pods and lenses ranging in length from decimetres to hundreds of metres. They typically have amphibolitized margins, foliated parallel to the contacts. These rocks, and their retrograded equivalents, are ubiquitous over large areas of the WGR (Eskola, 1921). Movement along their contacts, related to folding and boudinage, has eliminated most original contact relations between the eclogites and their host rocks. However, Griffin and Carswell (1985) describe one location where backveining of the eclogite by anatectic melts from the gneisses is preserved, and the veins contain high-*P* phases (omphacite, garnet, phengite, kyanite). Krogh (1980*a*) has also described intrusive contacts between eclogite and gneiss. This sort of relationship suggests that many of the eclogites may have originated as dolerite dykes. In many areas, granite pegmatites have intruded both eclogites and their country rocks prior to boudinage of the mafic rocks. The pegmatites are preserved within the eclogites, but disrupted where they cross the contacts. Such observations imply that many of the 'tectonic contacts' between eclogites and their wall rocks are due to movement at a very late stage in the retrogression history of the complex.

In addition to the typical lenses and pods, several very large bodies of 'external' eclogite are known. The Hareidland eclogite (Mysen and Heier, 1972; Grønlie *et al.*, 1972) measures roughly $6 \times 1 \times 0.3$ km, and the Eiksunddal body (Jamtveit, 1986, 1987*a*, *b*) $1 \times 0.5 \times 0.5$ km. Within some supracrustal sequences, eclogites occur as dm-thick layers within quartzites and calc-silicate rocks (Griffin and Råheim, 1973; Krogh, 1980*b*). On the Molde peninsula, eclogite layers 10–100 m thick occur within and overlying regionally extensive, massive marble bodies (Hernes, 1954; Harvey, 1985; Reksten, 1986).

Low-*P* protoliths of external eclogites

One major result of recent work has been the recognition that most, if not all, external eclogites have been derived by high-*P* prograde metamorphism of low-pressure protoliths. This conclusion is contrary to interpretations of these rocks as fragments of subcontinental or suboceanic mantle (O'Hara *et al.*, 1971; Lappin and Smith, 1981), and is based on three lines of evidence.

(1) *Lithologic associations/geochemistry.* Eclogites occurring as integral parts of supracrustal associations, as mentioned above, are demonstrably of shallow origin, though in many cases it is difficult to distinguish between sedimentary or volcanic protoliths. Eclogites (often Ky-bearing) occur as layers in anorthosites; these eclogites show gradational contacts and geochemical affinities with the anorthosites, including positive Eu anomalies (Brastad, 1985; Erambert, 1986). They appear to have formed by metamorphism of olivine+

pyroxene + plagioclase rocks, which cannot have formed at $P > 8$ kbar. Positive Eu anomalies have been observed in several other kyanite eclogites, suggesting that their protoliths were plagioclase-rich cumulates (Gebauer *et al.*, 1985; Griffin, unpubl. data).

The Hareidland eclogite has marked layering, in which garnet-rich layers have higher Mg/(Mg + Fe) than pyroxene-rich layers (Mysen and Heier, 1972). The layering thus does not reflect accumulation of eclogite phases, but it can be described in terms of oliv + plag *vs.* cpx + plag ± Mg^+ cumulates. The Eiksunddal body consists of interlayered eclogites, garnet websterites ('orthopyroxene eclogites') and garnet peridotites. Detailed geochemical modelling (Jamtveit, 1986, 1987*a*) shows that the layering can be modelled by accumulation of olivine + plagioclase + minor magnetite, with varying proportions of trapped liquid. The parental liquid is Fe- and Al-rich, and corresponds to the compositions of many eclogitized gabbros found in the WGR. Disruption of this type of complex has probably produced many of the smaller eclogites and 'orthopyroxene eclogites' in the WGR (Carswell *et al.*, 1983; cf. Lappin and Smith, 1978). Sr-isotope studies show that the Eiksunddal garnet websterites, like the smaller bodies, have been heavily contaminated by crustal Sr prior to or during eclogite-facies metamorphism (Brueckner, 1977; Griffin and Brueckner, 1985).

(2) *Arrested reactions.* The partial conversion of gabbroic rocks (oliv + plag, cpx + plag, opx + plag) to eclogite, via a coronite stage, is observed throughout the WGR (Gjelsvik, 1952; Griffin and Råheim, 1973; Griffin and Heier, 1973). Mørk (1985*a*, *b*; 1986) has provided detailed descriptions of the microchemical and microstructural aspects of this metamorphic process, and has discussed the kinetic controls. Griffin and Råheim (1973) interpreted these reactions as the result of isobaric cooling. However, Sm-Nd dating (Mørk and Mearns, 1986) has shown that the metamorphism (*ca.* 400 Ma) occurred long after the intrusion of the gabbros (*ca.* 1250 Ma), and must be regarded as prograde.

In the Bergen Arcs, Proterozoic anorthositic granulites are transformed into kyanite eclogites in shear zones and along microshears (Austrheim and Griffin, 1985). This work, like those of Mørk, demonstrates the importance of deformation and fluid access in promoting eclogite-forming reactions, and emphasizes the possible metastable persistence of dry mineral assemblages through high-P metamorphism.

(3) *Inclusion suites and mineral zoning.* Amphibolite-facies inclusion suites (Ca-amphibole, epidote, biotite, paragonite, plagioclase, magnetite, ilmenite, quartz) have been recognized in the cores of garnets from many external eclogites (Krogh, 1982). The transition from amphibolite facies to eclogite facies has occurred during the growth of the garnets; the garnet rims contain inclusions of the eclogite assemblage. These inclusion suites are clear evidence that many eclogites formed by prograde metamorphism of amphibolites. Krogh (1980*a*) has described glaucophane-schist facies inclusion suites in the cores of eclogites from Sunnfjord.

Many eclogite garnets, including most of those with amphibolite-facies inclusion suites, also show strong chemical zoning. The typical patterns show cores with high Mn and/or Fe, and rims with higher Mg/Fe (Bryhni and Griffin, 1971; Krogh, 1980*a*, 1982). This pattern is interpreted as the result of growth during prograde metamorphism; detailed analysis of Fe/Mg distribution between inclusions and matrix, variations in amphibole chemistry, and zoning of individual cpx grains in garnet support this interpretation (Krogh, 1982). Reversals of this zoning pattern in the outermost rims of some garnets are interpreted as a retrograde effect. Ca tends to show either little zoning, or a decrease from core to rim of the garnets. Bryhni and Griffin (1971) noted the similarity of the high-Ca cores to the garnets of some amphibolites. Mørk (1985) and Jamtveit (1986, 1987*b*) suggest that strong zoning of Ca is related to early growth of garnet in equilibrium with plagioclase, while the low-Ca cores of garnets unzoned in Ca were in equilibrium with amphibole-albite assemblages. In either case, these zoning profiles imply the initial growth of garnet in a low-P assemblage, followed by high-P metamorphism.

In many eclogites from the coastal areas, the garnets are homogeneous, with only a thin rim of lower Mg/Fe. These all yield high T (> 700 °C, see below).

P–T estimates

The careful application of geothermobarometry to a large number of eclogites in the WGR has made a significant contribution to our understanding of the high-P metamorphism. Krogh (1977*a*, *b*) showed that there is a zonal distribution of K_D [(Fe/Mg in gnt)/(Fe/Mg in cpx)] across the WGR. This has later been improved by the accumulation of more samples, and converted to a temperature distribution (Fig. 2).

As noted above, many of the garnets are zoned, while pyroxenes typically are homogeneous. The temperatures given in Fig. 2 are maximum values, derived using the highest Mg/Fe in each garnet

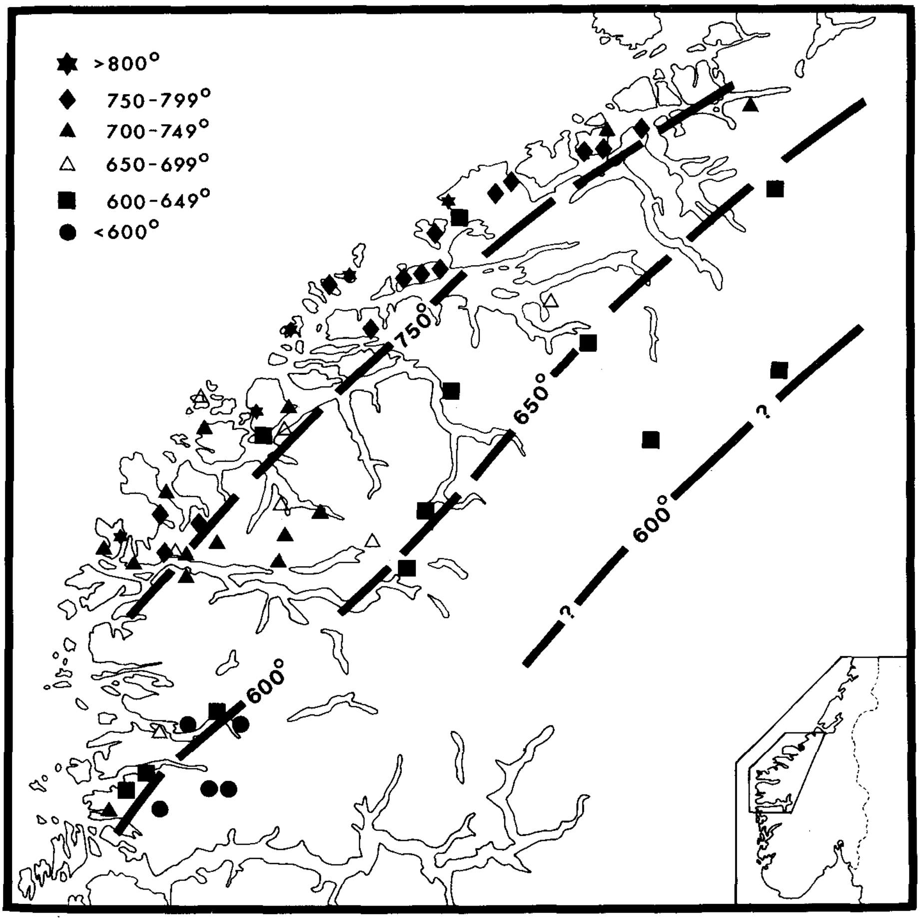

FIG. 2. Regional distribution of T_{max} in the Western Gneiss Region, based on the geothermometer of Ellis and Green (1979) at an assumed P = 15 kbar. Data given by Griffin *et al.* (1985).

(near-rims in prograde-zoned ones, cores in retrograde-zoned ones). A major uncertainty in the temperatures arises from the need to calculate Fe^{3+} in the pyroxenes; overestimation will give too low a temperature. However, in the present case, many low-T eclogites have no calculated Fe^{3+} in the pyroxenes, while many high-T ones contain significant Fe^{3+}. Large variations in the compositions of the eclogite minerals also can lead to uncertainties in T estimates (Koons, 1984), but most of the samples used here fall in a narrow range of compositions. The gradient shown in Fig. 2 is therefore regarded as real. The absolute values of the isotherms are a function of the geothermometer used (Ellis and Green, 1979). However, the T values given here agree reasonably well with those obtained from other methods (Krogh and Råheim, 1978; Harley, 1984).

The isotherms show the easternmost occurrence of the indicated T. The occasional lower-T values to the west may reflect either problems with determination of Fe^{3+} in pyroxene, relict mineral compositions that never equilibrated to higher T, or undetected retrograde effects. There is a marked difference in the isotherm pattern in the main part of the WGR and in the Sunnfjord area to the south ('Naustdal' on Fig. 1). These areas are separated by a major E-W transcurrent fault (Bryhni, pers. comm.). The lower temperatures in Sunnfjord are consistent with the preservation of glaucophane in eclogites there (Krogh, 1980*a*). There is some evidence in this area that the high-T eclogites are in a nappe cover, and low-T ones in the basement, but the geologic relations are not clear.

The regional increase of T_{max} toward the coastline is paralleled by an increase in P (Fig. 3). The

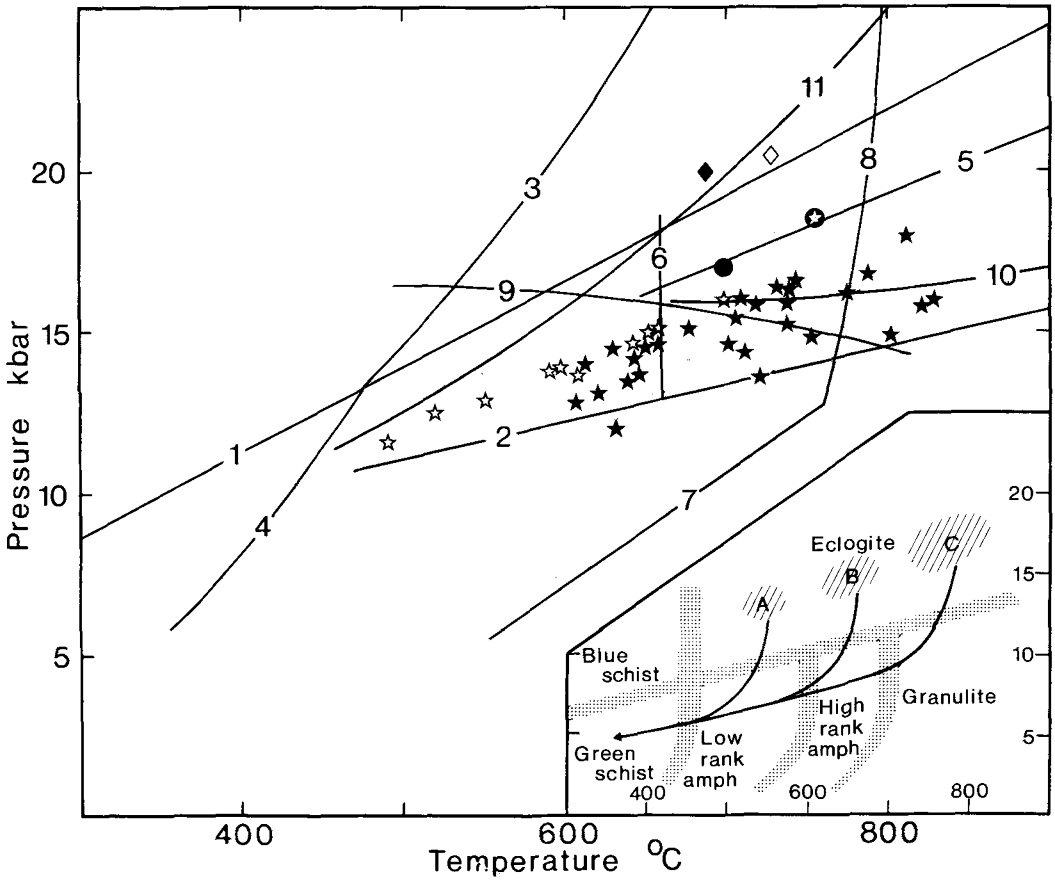

FIG. 3. P–T_{max} relations in eclogites from the Western Gneiss Region, after Griffin *et al.* (1985). Open stars: eclogites from Sunnfjord. Filled stars: eclogites from northern part of WGR. Diamonds: average gnt-peridotite estimates from Medaris (1980) and Carswell and Gibb (1980). Filled circle: average external opx-eclogite with Fe^{2+} in cpx analysed (Carswell *et al.*, 1985). Circled star: Same, with Fe^{3+} calc. Curves: 1 and 2: Lower stabilities of $Jd_{100}+Q$ and $Jd_{20}+Q$, respectively (Holland, 1980). 3 and 4: Lw+Jd+Zo+Pg+Q+Vp and Lw+Ab+Zo+Pg+Q+Vp, respectively (Holland, 1979). 5: Plag out in high-Al basalt/andesite, based on data for relevant compositions at 1100 °C, extrapolated at 20 bars/°C (Green and Ringwood, 1972). 6: Pg+Zo+Q+L (Franz and Althaus, 1977). 7 and 8: Zo+Ky+Q+An+Vp and Zo+Ky+Q+Vp+L, respectively (Boettcher, 1970). 9: Pg+$DiJd_{40}$+Ky+Vp (Holland, 1979). 10: Spinel lherzolite/garnet lherzolite transition (O'Hara *et al.*, 1971). 11: P, T path, Zagros, top of slab (Bird *et al.*, 1975).

pressures shown here are minimum values for P at T_{max}, since they assume the coexistence of albite with omphacite. In the case of the paragonite-bearing samples, these may not be far from the true P at T_{max}. Griffin *et al.* (1985) and Krogh (1982) assumed that P_{max} coincided with T_{max}, and that this 'piezothermal array' reflected the maximum pressures reached across the WGR. However, theoretical models (England and Richardson, 1977; England and Thompson, 1984; Thompson and England, 1984) require that T_{max} is reached during uplift of high-P metamorphic terrains, and this implies that P_{max} may have been higher, at some lower T, for each point shown in Fig. 3. Jamtveit (1986, 1987*b*) made a detailed study of inclusion suites and mineral zoning in the Eiksunddal eclogites and garnet websterites. He derived a very steep prograde metamorphic path, with P_{max} = *ca.* 25 kbar at $T \leqslant 600$ °C, while at T_{max} = *ca.* 750 °C, P = *ca.* 18 kbar. This P–T path corresponds well with the theoretical models mentioned above. It is also consistent with the preservation of jadeite-rich pyroxenes (Smith *et al.*, 1980; Jamtveit, 1986, 1987*b*) and possible coesite (Smith, 1984) in eclogites from the highest-T part of the area, though the pressures required are not as extreme as those suggested by

Smith (1984). This is especially so if coesite has been stabilized by moderate degrees of strain (*ca.* 30%; Green, 1972).

Even higher pressures have been proposed for external garnet websterites (30–40 kbar; Lappin and Smith, 1978, 1981). The orthopyroxenes in these rocks typically contain low-Al cores and higher-Al rims. This has been interpreted as the effect of a decrease in P, and the high pressures derived by combining analyses of opx cores with temperatures (T_{max}) derived from cpx + gnt cores. This procedure is probably invalid. There are two ways to produce low-Al orthopyroxenes at low pressure, and both have been observed in the WGR: (1) coronitic replacement of olivine by opx, followed by recrystallization (Mørk, 1985*a* and *b*, 1986), (2) low-T growth of opx, in equilibrium with either chlorite or garnet (Jamtveit, 1984; Carswell *et al.*, 1985). On the P–T path derived by Jamtveit (1986, 1987*b*), opx cores in equilibrium with garnet at P_{max} (25 kbar) will have < 0.5 wt.% Al_2O_3, while opx rims equilibrated at T_{max} will have $\sim 1\%$ Al_2O_3, as observed in many 'orthopyroxene eclogites'.

Even though the *very* high pressure estimates may be regarded as spurious, it appears now that $P \approx 25$ kbar may have been reached in some of the coastal areas of the WGR. P_{max} was probably correspondingly lower in the inland areas, so that heating during uplift was correspondingly less (cf. England and Thompson, 1984; Thompson and England, 1984). The regional distribution of T_{max} implies that the eclogite-facies metamorphism was regional in extent, and that the terrain behaved as a coherent unit during the metamorphism. This of course implies that the metamorphism affected the country rocks as well as the eclogites.

High-*P* metamorphism in the gneisses

One of the major arguments for the tectonic introduction of the eclogites into the gneisses, and against *in situ* metamorphism, has been the apparent lack of evidence for high-P metamorphism in the gneisses. Although granulite-facies relics (of unknown age) survive throughout the coastal strip, the gneisses are generally in the amphibolite facies. The common occurrence of garnet amphibolites has also been taken as evidence against high-P metamorphism of the enclosing gneisses. However, Eskola (1921) recognized the garnet amphibolites as strongly retrograded eclogites, and as part of a widespread and pervasive retrograde metamorphic overprint that affected gneisses, eclogites and peridotites (Medaris, 1984; Carswell, 1986).

While petrographic studies on the gneisses have been sadly neglected until recently, considerable evidence does exist that the gneisses have experienced high-P metamorphism. Omphacite relics occur in the felsic gneisses in the form of clinopyroxene–plagioclase and amphibole–plagioclase symplectites (Mysen and Heier, 1972; Fig. 4*a*) and irregular aggregates of plagioclase dotted with quartz and pyroxene and/or amphibole (Fig. 4*b*), and as inclusions in garnet (Jamtveit, 1986, 1987*b*). Calcic ($Gros_{20-35}$) garnets apparently coexist with sodic plagioclase + kyanite + qtz in gneisses and quartzites (Griffin and Mørk, 1981), implying pressures $\geqslant 19$ kbar. Finally, pseudomorphs of biotite + kspar + kyanite after phengite (Fig. 4*c*) appear to be widespread in the gneisses. The significance of this microstructure was first recognized by Heinrich (1982). The phengite-producing reaction in the gneisses is a prograde, water-conserving reaction; breakdown of the phengite during decompression releases water, which will hydrate pyroxenes in the gneisses, as well as the rims of enclosed eclogite pods. Thus petrographic evidence does exist to support the *in situ* model, and to explain the widespread retrogression of the gneisses, without even requiring a large outside supply of water. Koons and Thompson (1985) have demonstrated that mineral assemblages in non-mafic rocks subjected to HP metamorphism will 'revert' to amphibolite-facies assemblages more rapidly than the enclosed mafic eclogites during uplift.

Timing of high-*P* metamorphism

Opinions on the age of the WGR, and of the eclogite-facies metamorphism, have swung from Caledonian to Precambrian and back again several times since Eskola's work. Extensive geochronological work over the past 10–15 years has established that the gneisses, and most of the supracrustal rocks, of the WGR are Proterozoic in age, and have suffered metamorphic events at *ca.* 1800–1600 Ma (Southwest Scandinavian orogeny; Gorbatschev, 1980), *ca.* 1200–1000 Ma (Sveconorwegian) and 450–350 Ma (Caledonian). Gabbroic protoliths to eclogites have been dated to *ca.* 1550 Ma (Rb-Sr: Tørudbakken, 1981), *ca.* 1250 Ma (Sm-Nd; Mørk and Mearns, 1986) and *ca.* 550 Ma (Sm–Nd; Griffin and Brueckner, 1985).

The first reliable dates on the eclogite mineral paragenesis itself were Sm–Nd cpx–gnt ages on external eclogites (Griffin and Brueckner, 1980, 1985). Five pairs average 425 ± 20 Ma; one sample with a date of *ca.* 880 Ma contains two generations of garnet. Similar Sm–Nd ages have been obtained on several other external eclogites (Mearns, 1984,

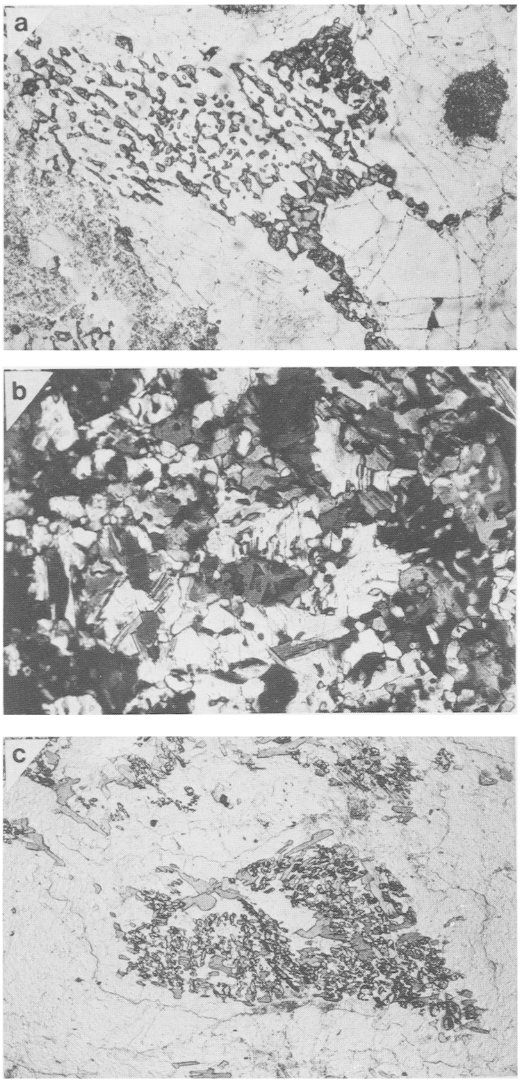

FIG. 4. Petrographic evidence of high-P metamorphism in felsic rocks from the Western Gneiss Region. (*a*) Diopside-oligoclase symplectite after omphacite, in Kspar-plag-qtz-gnt-bio gneiss, Bergsøya. Width of view 2.5 mm. (*b*) Irregular aggregate of oligoclase with dispersed droplets of quartz and flakes of biotite; breakdown of jadeitic pyroxene? Plagioclase-quartz-biotite gneiss, Fiskåbygd. Width of view 1 mm. (*c*) Biotite-Kspar-kyanite intergrowth, pseudomorph after phengite. Kspar-quartz-plagioclase gneiss, Bergsøya. Width of view 2.5 mm.

1986; Mørk and Mearns, 1986), and are supported by Caledonian U/Pb ages on zircons (concordant zircons, Krogh *et al.*, 1974; lower intercepts; Gebauer *et al.*, 1985).

It might be argued that these Sm-Nd ages represent resetting of the Nd isotopic systems in assemblages that are really much older. This seems very unlikely on crystal-chemical grounds alone (Griffin and Brueckner, 1985). Furthermore, work on the shear-zone eclogites of the Bergen Arcs (Austrheim and Griffin, 1985) shows that the garnet granulites retain Proterozoic Sm-Nd mineral ages within metres of the eclogitized shear zones and even in augen within the shear zones (Cohen *et al.*, in prep.). The shear zone eclogites give lower Palaeozoic mineral ages. Mørk and Mearns (1986) show that Sm-Nd re-equilibration among the minerals in coronite dolerites is only achieved after complete recrystallization to an equilibrated metamorphic fabric. It therefore seems reasonable to interpret the Sm-Nd mineral ages as the time of crystallization and homogenization of the garnet and pyroxene, probably near T_{max}.

Krill (1980, 1985) has provided important evidence on the timing of the HP metamorphism relative to the development of the Caledonian orogen. In the Oppdal area on the eastern edge of the WGR, he defined a sequence of nappes including, from bottom to top: (1) arkoses (parautochthonous), (2) augen gneisses, rapakivi granites and anorthosites, (3) arkoses with dolerite dykes, (4) Palaeozoic mica schists and amphibolites, (5) Palaeozoic phyllites and greenschists. The lithologic associations in these nappes are similar to those recognized as disrupted lenses and 'stripes' in the western part of the WGR, and locally preserved as a recognizable lithostratigraphy (Bryhni, 1966; Bryhni and Grimstad, 1970). Rb-Sr whole-rock dating of the Blåhø group (unit 4) to *ca.* 580 Ma demonstrates that the nappe transport is Caledonian.

Following thrusting, this nappe pile was recumbently folded with its basement. As the Blåhø unit is traced westward, its metamorphic grade increases (as shown by biotite-garnet thermometry), until the amphibolite units are converted to eclogites. This occurs at $T \sim 600$ °C; these eclogites are the points farthest east on Fig. 2. Eclogites are found in the basement a few km away; these also yield temperatures of 600-650 °C (Griffin *et al.*, 1985). Krill (1985; pers. comm.) has later traced several other infolded supracrustal sequences westward across the WGR, and this work is being followed up by several thesis students at Mineralogisk-Geologisk Museum.

These observations suggest that early Caledonian overthrusting created a pile of nappes above the present WGR, similar to those seen farther north along the Caledonides (Roberts and Gee, 1985). This nappe pile was then progressively disrupted by flattening and infolding with the basement, during the HP metamorphism. However, it is important to note that the WGR is not a 'melange' (as suggested by Smith, 1980) in the sense that this term is used in blueschist terrains such as the Franciscan. Some distinctive units (anorthosites, quartzites) may be traced for tens of kilometres both along and across the trends of the isotherms in Fig. 2 (Bryhni, 1966; Krill, 1985; Sigmond *et al.*, 1984). Much more mapping will be required to reconstruct the Caledonian history of this terrain.

General tectonic model

The regional pattern of pressure and temperature in the eclogites, the relics of HP metamorphism in the basement, and Krill's structural observations in the eastern part of the WGR all suggest that the WGR has been deeply buried during the late stages of the Caledonian orogeny. The most obvious mechanism for this is the depression of the Baltic Shield beneath the overriding Greenland plate. This model has been summarized in an elegant cartoon by Cuthbert *et al.* (1983). The outer (coastal) parts of the WGR have been depressed to depths of at least 70–80 km, the inner parts to 30–40 km.

The Sm–Nd mineral ages on external eclogites are interpreted as placing the T_{max} at *ca.* 425 ± 20 Ma; this probably postdates the maximum burial by a few Ma at most. Uplift from T_{max} was relatively rapid, aided by transport of the Upper Allochthon to the east, and by underthrusting of more continental crust from the east. The top of the underthrust material may be shown by the mid-crustal low-velocity zone reported by Mykkeltveit *et al.* (1980).

Rb–Sr and K–Ar mineral ages suggest that the uplift was mainly over by *ca.* 390 Ma (Griffin and Brueckner, 1985; Lux, 1985; Mørk and Mearns, 1986; Griffin and Krill, unpubl.). In the northern part of the WGR, uplift was accompanied by a severe flattening that developed a strong NE–SW foliation, and produced blastomylonites on a regional scale. These blastomylonites are cut by an undeformed granite pegmatite which gives a Rb–Sr mineral isochron of 345 ± 2 Ma (Griffin, unpubl.); this date probably marks a definitive end to major deformation in the WGR.

The region of the WGR affected by the HP metamorphism is at least 50×300 km (Fig. 2), but the continental collision zone may have been much longer. van Roermund (1985) has described a corresponding zonation of P–T estimates in the eclogites of the Seve nappe in northern Sweden. His reconstruction of the original positions of these nappes places them outside of north Norway, along the trend of the isotherms in Fig. 2. However, preliminary Sm–Nd data (Mørk, pers. comm.) and $^{39}Ar/^{40}Ar$ data (Dallmeyer and Gee, 1986) on these eclogites suggest they may be somewhat older (500 Ma?) than the WGR eclogites. Krogh *et al.* (1987) suggest that eclogite-facies conditions may have been reached around 600 Ma ago in the Tromsø eclogites. Multiple subduction/transport episodes, or migration of the locus of subduction with time, may be implied.

The garnet peridotite problem

A major unresolved question is—where do the Mg–Cr garnet peridotites fit into this model? Garnet peridotites are known from the coastal strip of the WGR, and from the Seve nappes in northern Sweden (van Roermund, 1985; pers. comm.). Carswell and Gibb (1980) and Medaris (1984) interpreted the zoning of garnet and orthopyroxene in the WGR peridotites as reflecting decreasing P and T from very high values. However, Carswell *et al.* (1983), Jamtveit (1984) and Carswell (1986) have presented evidence that at least some of these peridotites originated as low-pressure assemblages (spinel $\pm$ chlorite peridotite) and subsequently equilibrated at upper mantle/deep crust conditions (*ca.* 750 °C, 20 kbar). Medaris (1984) and Carswell (1986) have discussed the petrographic evidence which defines the very complex history of these rocks, involving at least two stages of equilibration in the garnet–peridotite field, followed by retrograde breakdown to spinel peridotite, then to chlorite + amphibole + opx assemblages.

All of the known occurrences of garnet peridotite, except that in Tafjord, occur in the highest-T part of the WGR terrain ($T \geqslant 750$ °C). Their estimated temperatures are similar to those in nearby external eclogites, and the estimated pressures for this part of the terrain are well within the stability field of the garnet peridotite assemblage. Griffin *et al.* (1985) therefore suggested that the garnet peridotites also had low-P protoliths, and were metamorphosed to high-pressure assemblages together with the surrounding terrain, during the Caledonian orogeny. Griffin and Qvale (1985) and Jamtveit (1984) supported this interpretation by showing that the superferrian internal eclogites in Almklovdalen and Gurskebotn had formed by prograde metamorphism from (rodingitized?) amphibolites. However, this model does not explain the Tafjord occurrences (Carswell, 1968) which give temperatures $\geqslant 750$ °C (Medaris, 1984) even though they lie near the 650° isotherm of Fig. 2.

Furthermore, this model has received a severe blow from recent Sm–Nd dating of garnet peridotites and garnet pyroxenites from Almklovdalen. A peridotite yields a gnt–cpx–w.r. age of *ca.* 1700 Ma (Mearns, 1986) while two garnet pyroxenites give *ca.* 1475 Ma (Jacobsen and Wasserburg, 1980) and *ca.* 1050 Ma (Mearns, 1986). The existence of more than one generation of garnet and pyroxene in these rocks (Medaris, 1984; Carswell, 1986) makes the detailed interpretation of these data difficult. Mearns (1986) cautiously suggests that the 1700 Ma date may be close to the real age of the garnet peridotite assemblage, while the pyroxenite dates reflect varying degrees of Sveconorwegian and Caledonian disturbance.

At the moment, two general interpretations of these data are possible.

(1) The garnet peridotites evolved from spinel peridotite protoliths in the subcontinental mantle *ca.* 1700 Ma ago, and were stored there until incorporation into the deep crust during the Caledonian 'subduction'. This implies a relatively cool sub-continental geotherm.

(2) The garnet peridotites were formed during a *ca.* 1700 Ma HP metamorphism and introduced into the upper crust then, or during the Sveconorwegian episode. They have partially re-equilibrated (Carswell's (1986) Stage III), but essentially survived, during the Caledonian HP metamorphism. This interpretation may be strengthened by reports of Sveconorwegian Sm-Nd ages for the Loch Duich eclogites in the Scottish Caledonides (Sanders *et al.*, 1984).

Obviously, further isotopic work on these rocks is essential; the results will have major tectonic implications.

Conclusions

(1) The 'external' eclogites of the WGR are the products of prograde metamorphism on a regional scale. The area affected measures at least 50 × 300 km. If, as seems likely, the eclogites in the nappes of northern Norway and Sweden were formed during the same orogenic period, the area affected by HP metamorphism is over 1000 km long.

(2) The HP metamorphism of the WGR was the result of continental collision during Silurian time. The Baltic shield was overridden from the west by the Greenland plate; the overridden terrain consisted of a Proterozoic basement complex and a pile of nappes, emplaced earlier in the Caledonian orogeny.

(3) Maximum depths of *ca.* 70-90 km are recorded by external eclogites near the present coast, shallower depths (30-40 km) by the eastern and southern parts of the WGR.

(4) The tectonic position of the garnet peridotites and their 'internal' eclogites (garnet pyroxenites) is still ambiguous: isotopic evidence indicates that the earliest garnet peridotite assemblages were formed in Proterozoic time.

Further understanding of the evolution and tectonic setting of these HP rocks will require detailed mapping of large areas of the WGR, to delineate the disrupted basement/cover relations, and the relation of the HP rocks to these units. Detailed, high-precision isotopic studies will be necessary to define the ages and origins of the protoliths, and the fine structure of the HP metamorphism. In particular, understanding of the time relations of the garnet peridotites to the other HP assemblages will be essential. Isotopic studies of the transported HP rocks in the nappes will tell a great deal about the evolution of the orogen as a whole.

It seems reasonable to predict that the WGR, and the HP rocks in the nappes, will serve for many years as a laboratory in which to study processes in the deeper levels of Himalayan-type plate collision zones.

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