Geol. Mag. **142** (5), 2005, pp. 605–615. © 2005 Cambridge University Press
doi:10.1017/S001675680500110X Printed in the United Kingdom

P–T conditions of Grenville-age eclogite facies metamorphism and amphibolite facies retrogression of the Glenelg–Attadale Inlier, NW Scotland

C. D. STOREY*†, T. S. BREWER* & S. TEMPERLEY*

*Department of Geology, University of Leicester, University Road, Leicester LE1 7RH, UK

(*Received* 1 July 2004; *accepted* 28 April 2005)

Abstract – Peak and retrograde *P–T* conditions of Grenville-age eclogites from the Glenelg–Attadale Inlier of the northwest Highlands of Scotland are presented. Peak conditions are estimated as *c.* 20 kbar and 750–780 °C, in broad agreement with previous work. The eclogites subsequently followed a steep decompression path to *c.* 13 kbar and 650–700 °C during amphibolite facies retrogression. Peak eclogite facies metamorphism occurred > 1080 Ma and retrogression at *c.* 995 Ma, suggesting fairly sluggish uplift rates of < 0.3 km/Ma and cooling rates of < 1.25 °C/Ma, when compared with other parts of the Grenville orogeny and/or modern orogens. However, current poor constraints on the timing of peak metamorphism mean that these rates cannot be used to interpret the geodynamic evolution of this part of the orogen. The *P–T–t* data, together with petrology and the field relationships between the basement rocks of the Glenelg–Attadale Inlier and the overlying Moine Supergroup, mean that it is difficult to support the currently held view that an unconformable relationship exists between the two. It is suggested that more data are required in order to re-interpret the Neoproterozic tectonic evolution of the northwest Highlands of Scotland.

Keywords: *P–T* conditions, eclogite, retrograde, Moinian, Scotland.

1. Introduction

The Glenelg–Attadale Inlier (GAI) is the largest tract of Sub-Moine basement within the Moine Nappe of northern and western Scotland (Fig. 1). The inlier is bound to the west by the Caledonian Moine Thrust and an intervening thin sliver of Moine rocks, and to the east by the Moine Supergroup metasedimentary rocks (Fig. 1). This area is a critical locality in understanding the nature of the basement in the hinterland of the Caledonian orogen, and how it might relate to the basement in the foreland, that is, the Lewisian Complex (Fig. 1).

The Glenelg–Attadale Inlier contains eclogites and their retrogressed equivalents (Teall, 1891; Peach *et al.* 1910; Alderman, 1936; Mercy & O'Hara, 1968). Ramsay (1958) and Sutton & Watson (1959) divided the Glenelg–Attadale Inlier into two separate lithological units, with a zone of high-strain and discontinuous slivers of Moine psammites between the two. The Western Unit (WU) is composed largely of felsic gneisses, intruded by mafic sheets and enclosing mafic–ultramafic pods and larger bodies, some of which preserve granulite facies assemblages. The Eastern Unit (EU) comprises banded mafic, intermediate and felsic gneisses, metasediments including marbles, and eclogite. The eclogite is best preserved in the interiors of thick sheets and/or pods. The eclogite sheets and pods are generally sheared and retrogressed to amphibolites at the margins of such bodies. A single rare occurrence of eclogite in the Western Unit has been reported by Sanders (1979).

The eclogites in the Glenelg–Attadale Inlier are the only unequivocal occurrence of eclogite facies rocks found within the British Isles. Several issues remain unresolved, such as their regional extent, their relationship to surrounding rocks in the inlier and to the surrounding Moine Supergroup rocks, and their *P–T–t* evolution. The latter point is the subject of this study. Furthermore, it is unclear as to what the affinities are between the eclogite-bearing units and the Lewisian basement, whether the eclogites were reworked during the Grenville orogeny or were formed as part of this orogeny. This latter point is beyond the scope of the present study. Previous *P–T* studies on these eclogites (Sanders, 1988, 1989) concluded that eclogites in the Eastern Unit attained peak metamorphic conditions of *c.* 15–18 kbar and 700–730 °C, while Rawson, Carswell & Smallwood (2001) estimated *P–T* conditions of 'orogenic' type peridotite within the Eastern Unit as 20 ± 3 kbar and 730 ± 50 °C, suggesting a slightly greater depth of burial. In terms of tectonic interpretations, the determination of the post-peak retrograde *P–T–t* path is critical. In this paper we present a new study of the

---

† Author for correspondence. Current address: Department of Earth Sciences, The Open University, Walton Hall, Milton Keynes MK7 6AA, UK; e-mail: c.storey@open.ac.uk

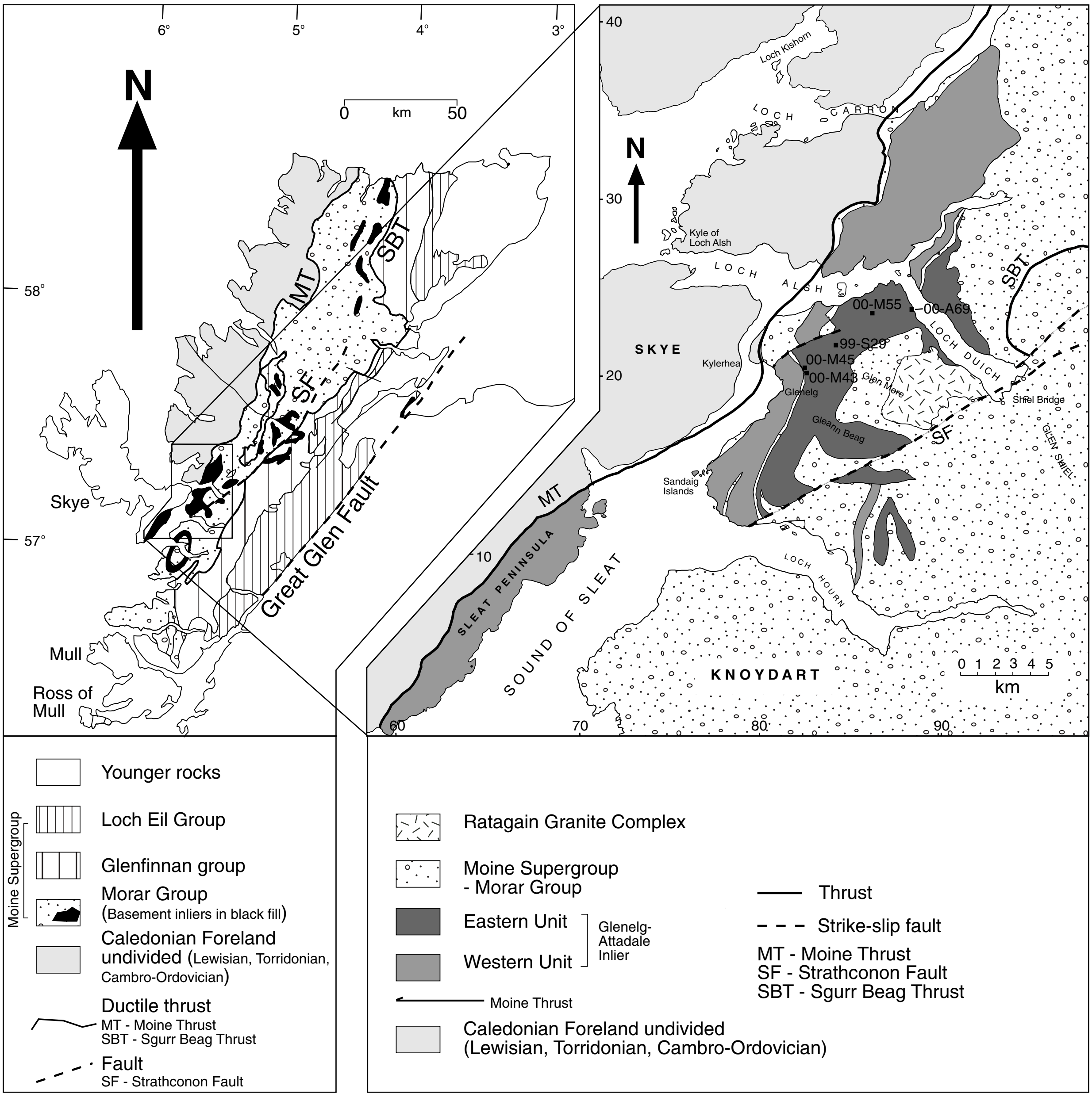


Figure 1. Location maps. Left side is Scotland with geology north of the Great Glen Fault. Inset box is geological map of Glenelg–Attadale Inlier modified after Peach *et al.* (1910) and May *et al.* (1993). Note grid lines for the inset are from the UK National Grid and all are prefixed NG.

decompression (retrogession) history of the Glenelg eclogites, by petrographic and microprobe analysis of samples and estimation of *P*–*T* conditions.

In terms of timing of eclogite facies metamorphism and retrogression, two studies are pertinent. Sanders, van Calsteren & Hawkesworth (1984) reported mineral-whole rock Sm–Nd analyses on two samples of Eastern Unit eclogite that produced isochron ages of 1082 ± 24 Ma and 1010 ± 13 Ma. They interpreted the older age to be close to the date of the peak of eclogite facies metamorphism. Brewer *et al.* (2003) concluded that upper amphibolite facies retrogression of the Eastern Unit occurred at 995 ± 8 Ma by studying the U–Pb systematics of zircon and titanite. They therefore showed that retrograde metamorphism (*c.* 1000 Ma) occurred within the Grenville orogenic cycle.

2. Location and petrography of samples

All samples studied were collected from the Eastern Unit and their locations are shown on Figure 1.

2.a. Felsic gneisses

Two samples of felsic gneiss that contain omphacite were collected. Sample 00-M45 [NG 8255 2050] was from a felsic layer about 30 cm thick within an interlayered succession of garnet–clinopyroxene–plagioclase–kyanite gneiss and eclogite in the vicinity of Creag Dubh. Sample 99-S29 [NG 841 218] was collected from the west side of Coire an Daimh from an outcrop containing layered felsic to mafic gneisses. The felsic layers, from which sample 99-S29 was collected, consisted of abundant plagioclase, garnet, omphacite and kyanite, and the mafic layers contained abundant omphacite with garnet and plagioclase.

2.b. Eclogite

Two samples of partially retrogressed eclogite were collected (Fig. 1). Sample 00-M43 [NG 827 201] was collected from the same area as acid gneiss 00-M45 and represents a more basic layer from the interlayered succession. Sample 00-M55 [NG 861 235] is from close to the southwest end of Loachan Beinne Faide and is an outcrop of eclogite close to outcrops of ultrabasic rock described by Rawson, Carswell & Smallwood (2001). Also, from the southwest shore of Loch Duich [NG 8845 2380], a sample of retrogressed streaky eclogite was collected from rock types similar to those described by Sanders (1988). This sample, 00-A69, has an L–S tectonic fabric defined by a grain-shape alignment of relict omphacite grains, which have been largely replaced by symplectitic intergrowths of amphibole–plagioclase–quartz. Also contributing to the fabric are garnet ribbons, comprising aggregates of equant to elongate crystals, and quartzo-feldspathic streaks, originally containing kyanite.

3. Petrography and mineral chemistry of samples

Polished thin-sections were studied and mineral compositions were determined at the University of Leicester using a JEOL 8600 electron microprobe fitted with a wavelength dispersive detector. The accelerating potential was 15kV, with a beam current of 3×10^{-8} A and a spot diameter of 10 μm for feldspar and 5 μm for all other minerals. Representative mineral analyses are presented in Table 1.

3.a. Felsic gneisses

3.a.1. Petrography

The sampled felsic gneisses contain garnet, omphacite, quartz, plagioclase, K-feldspar, diopside, kyanite, biotite, amphibole, rutile, zircon, allanite and epidote. Omphacite and plagioclase occur, along with rutile, as inclusions in garnet, and in the matrix, where they share straight grain boundaries with garnet. This suggests that the rock is cofacial with the eclogites, although strictly this rock is a high-pressure granulite since plagioclase is present in equilibrium with omphacite and garnet (Carswell, 1990). Kyanite occurs sporadically as thin 20–30 μm long blades within aggregates of plagioclase. Retrogression can be recognized as: (a) thin symplectitic rims between matrix garnet, omphacite and plagioclase composed of diopside, plagioclase and quartz, (b) randomly orientated needles of diopside (10–20 μm) along with exsolved perthite rods (5–10 μm length) within matrix plagioclase grains and (c) minor amphibole locally replacing omphacite, diopside and garnet at grain boundaries.

3.a.2. Mineral chemistry

In sample 99-S29, omphacite in contact with plagioclase is generally resorbed and is altered to thin (*c.* 50 μm thickness) symplectitic intergrowths of clinopyroxene and plagioclase. The highest Na and Al values are preserved in the cores of larger matrix grains whilst there is a decrease in Na and Al at the expense of Ca, Mg and Fe towards the rims. X_{Jd} values (determined using the IMA classification: Morimoto, 1988) are *c.* 0.4 in the core regions (Table 1) and 0.33 at rims in contact with symplectite. The garnet cores retain the highest X_{Fe2+} whilst there is an increase in X_{Ca} from core to rim and a negligible but slight increase in X_{Mg} at the rim. Plagioclase is most sodic in the cores of matrix grains with X_{Ab} *c.* 0.8 (oligoclase) with a slight increase in the albite component at the rims.

In sample 00-M45, the cores of matrix omphacite and plagioclase grains retain the highest Na values and garnet cores the highest Mg no. (Mg/(Mg + Fe)), indicating that the cores represent the highest *P–T* conditions. Plagioclase is found to be oligoclase (X_{Ab} 0.835; Table 1), whilst omphacite contains approximately X_{Jd} 0.35–0.45. The mineral chemistry is similar to that from 99-S29.

3.b. Eclogite

3.b.1. Petrography

The most well-preserved eclogites in the Glenelg–Attadale Inlier comprise a coarse granoblastic polygonal texture of omphacite and garnet with minor quartz and rutile. Omphacite occurs as inclusions in garnet and interstitially to garnet, and small garnet inclusions are also occasionally included in omphacite. All samples show some evidence of retrogression, but the degree of retrogression varies considerably from sample to sample. An early stage of retrogression is recognized as fine lamellar symplectites of diopside, oligoclase and quartz replacing omphacite. The presence of garnet as a stable phase during this phase of retrogression implies a short-lived high-pressure granulite facies overprint (O'Brien, 1997). Later intergrowths of pargasite and oligoclase form rims around the garnet, omphacite and symplectite. They also replace omphacite along the prismatic cleavage and garnet along fractures, although locally garnet retains

Table 1. Electron microprobe analyses

Sample	99-S29			00-M45			00-M55									00-M43			00-A69					
Mineral	O	G	P	O	G	P	A	A	A	P	P	P	G	G	G	O	P	G	P	G	A	P	A	G
Info	c	c	c	c	c	c	r	s	r	r	s	r	r	s	r	c	c	c	s	s	s	s	s	s
PT calc	a	a	a	b	b	b	c	d	e	c	d	e	c	d	e	f	f	f	g	g	g	h	h	h
O	6	12	32	6	12	32	23	23	23	32	32	32	12	12	12	6	32	12	32	12	23	32	23	12
Si	1.90	3.00	11.32	1.93	2.98	11.39	6.37	6.44	6.47	11.57	11.20	11.26	3.00	3.00	3.00	1.98	11.02	3.00	11.14	3.01	6.40	11.39	6.38	3.01
Ti	0.01	0.00	0.00	0.01	0.01	0.00	0.16	0.04	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.10	0.00	0.02	0.01
Al	0.53	1.96	4.65	0.42	1.99	4.60	2.30	2.23	2.27	4.39	4.55	4.65	1.98	1.98	1.98	0.41	4.47	2.00	4.75	1.97	2.36	4.57	2.45	1.99
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.01	0.00	0.00	0.00
Fe2	0.15	1.55	0.00	0.23	1.53	0.01	1.58	1.97	1.81	0.04	0.42	0.08	1.63	1.66	1.68	0.22	0.31	1.66	0.05	1.57	1.81	0.06	2.14	1.63
Mn	0.00	0.04	0.00	0.00	0.03	0.00	0.01	0.01	0.02	0.00	0.00	0.00	0.04	0.05	0.05	0.00	0.01	0.06	0.00	0.04	0.01	0.00	0.01	0.06
Mg	0.46	0.70	0.00	0.46	0.74	0.00	2.65	2.51	2.52	0.01	0.00	0.00	0.60	0.64	0.60	0.65	0.44	0.88	0.03	0.51	2.35	0.00	2.13	0.33
Ca	0.60	0.75	0.69	0.62	0.74	0.62	1.78	1.79	1.82	0.42	0.63	0.70	0.74	0.68	0.70	0.41	0.86	0.40	0.88	0.90	1.88	0.57	1.85	0.96
Na	0.35	0.01	3.26	0.36	0.00	3.35	0.80	0.81	0.84	3.59	3.47	3.39	0.00	0.01	0.00	0.29	3.22	0.00	3.20	0.00	0.58	3.44	0.58	0.00
K	0.00	0.00	0.09	0.00	0.02	0.04	0.15	0.02	0.02	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.04	0.00	0.20	0.03	0.24	0.00
Total	4.00	8.01	20.03	4.03	8.04	20.01	15.80	15.82	15.80	20.03	20.27	20.10	8.01	8.01	8.01	3.96	20.35	8.00	20.10	8.00	15.71	20.06	15.79	7.99
Fe^{2+}	0.14	1.51		0.14	1.43		1.33	1.27	1.40				1.61	1.63	1.64	0.22		1.65		1.56	1.67		1.70	1.63
Fe^{3+}	0.00	0.04		0.09	0.10		0.24	0.70	0.41				0.03	0.03	0.04	0.00		0.00		0.0	0.14		0.43	0.00
G-X_{Fe2+}		0.51			0.49								0.54	0.54	0.55			0.55		1.52				0.54
G-X_{Mg}		0.23			0.25								0.20	0.21	0.20			0.30		0.17				0.11
G-X_{Ca}		0.25			0.25								0.25	0.23	0.23			0.13		0.30				0.32
G-X_{Mn}		0.01			0.01								0.01	0.02	0.02			0.02		0.01				0.02
G-Mg no.		0.31			0.34								0.27	0.28	0.27			0.35		0.25				0.17
O-X_{Jd}	0.39			0.36												0.35								
P-X_{Ab}			0.81			0.84				0.89	0.85	0.83					0.79		0.78			0.85		

Analyses recalculated to cation totals for the appropriate number of oxygens in the mineral formulae. Ferric/Ferrous Iron calculated by charge/balance considerations using the method of Droop (1987). Minerals: G – garnet, O – omphacite, P – plagioclase, A – amphibole, Analysis information: c – core, r – rim, s – symplectite, *PT* calculation: letters refer to analyses used together to calculate *P–T* estimates reported in Figure 2.

straight boundaries with pargasite. This growth of pargasite and oligoclase indicates further retrogression to upper amphibolite facies conditions following the high-pressure granulite facies overprint. One sample (00-M43) contains coarse veins comprised of dark green amphibole 1–2 mm thick, within which relict omphacite, garnet and rutile crystals occur. The veins are sharply defined, and away from the veins a coarse, granoblastic polygonal texture of omphacite, garnet, rutile and quartz occurs.

Streaky eclogites were shown by Sanders (1988) to comprise K-feldspar, oligoclase, quartz, fine-grained kyanite and local minor scapolite and biotite within the streaks. The remainder of the rock comprises granoblastic omphacite and garnet with minor quartz and rutile. Quartz and perthitic K-feldspar form interstitially to kyanite, but a thin retrograde rim of oligoclase separates the first two from the kyanite. Omphacite is generally less common than garnet at the borders of the streaks, but where it does occur, it partially overgrows kyanite and there is a thin intervening rim of oligoclase separating these two phases. The boundaries between the quartzo-feldspathic streaks and the garnet are distinctly lobate, with the quartz and plagioclase embayed into the garnet. At the contact zones between the streaks and non-streaky areas, the garnet displays a corona-like border of subhedral neoblasts and new rims, which locally partially overgrow kyanite within the streaks. The new rims also contain inclusions of titanite, omphacite and quartz, indicating that they are coeval with the neoblasts forming the reaction corona and that this reaction occurred in the eclogite facies stability field. Plagioclase in the matrix locally shares straight boundaries with garnet and omphacite neoblasts, indicating that it grew in equilibrium with these phases. Rutile occurs as inclusions in old garnet porphyroblasts, but in the matrix it only occurs as relict cores inside titanite neoblasts, indicating that the eclogite facies paragenesis in the felsic streak did not contain rutile. Omphacite inclusions have not been found in the old garnet porphyroblasts, but quartz is present. In the mafic part of sample 00-A69, garnet and omphacite display a coarse granoblastic polygonal texture with minor interstitial quartz and rutile as inclusions in omphacite and garnet.

The texture of quartz and plagioclase embayments into garnet could be interpreted as the product of melting. In this case it is possible that the quartzo-feldspathic streaks may have crystallized from melts, which were generated during the peak of metamorphism. However, the source of the partial melt is unclear. The field evidence and descriptions supplied by Sanders (1988), where he notes that the streaks form a ‘pervasive network of ramifying, sub-parallel white threads which may comprise more than half the rock’ but are heterogeneously developed throughout basic layers, are consistent with the melt hypothesis. Sanders (1988) interpreted the streaks as sites of former plagioclase, which reacted with neighbouring garnet and omphacite during a period of ductile shearing.

The oligoclase rims on kyanite and resorption of omphacite are interpreted as a result of near-isothermal decompression (Sanders, 1988), and are compatible with an initial phase of high-pressure granulite facies overprinting of the eclogite. Further retrograde reactions occur as symplectites of amphibole, plagioclase and quartz that clearly pseudomorph earlier grains (probably omphacite) that share straight to embayed boundaries with garnet. Relict garnet porphyroblasts display an atoll structure, with small euhedral garnet neoblasts surrounding each garnet core. The neoblastic garnets are intergrown with amphibole–plagioclase–quartz symplectites. This textural evidence indicates that garnet was in equilibrium with amphibole, plagioclase and quartz during this stage of retrogression. The garnet neoblasts are also arranged in strings that define the foliation and lineation along with titanite. The titanite occasionally contains relict cores of rutile. The fabric is probably a relict of the eclogite facies fabric described above; the retrograde upper amphibolite facies textures suggest that the replacement of the eclogite facies paragenesis occurred statically.

3.b.2. Mineral chemistry

In eclogite sample 00-M55, garnets are chemically zoned, particularly at the contacts with omphacite, where there is usually a thin intervening rim (50–100 μm) of diopside/amphibole and plagioclase. The garnets display an increase in Fe and Ca and a decrease in Mg from core to rims. This is interpreted as a retrograde re-equilibration with the decompression paragenesis diopside/amphibole and plagioclase (Florence & Spear, 1991). Omphacite is unzoned (X_{Jd} *c.* 0.3), but at the rims there are signs of retrograde change, with decrease in Na and increase in Ca, Mg and Fe. The retrograde amphibole is calcic and transitional between edenite and pargasite (Leake *et al.* 1997) and the plagioclase is sodic (oligoclase) with X_{Ab} 0.8–0.9 (oligoclase; Table 1).

In sample 00-M43, garnet and omphacite in areas without amphibole veins (X_{Jd} 0.25–0.3) have relatively homogeneous mineral compositions, except where there is a thin symplectitic rim of plagioclase and clinopyroxene and/or amphibole between the grains. In this case, the garnet has a lower Mg no. at the contact with the symplectite. In areas traversed by amphibole veins, the garnets and omphacites are homogeneous and X_{Jd} is consistently low at *c.* 0.25, suggesting re-equilibration with amphibole. The intervening amphibole veins are rather more complex. Where the vein lies between omphacite and garnet the Mg no. in vein amphibole increases away from garnet towards omphacite. In contrast, Na and Ca are homogeneous in vein amphibole, except at contacts with either garnet or omphacite where Na tends to show a small increase. This suggests a degree of solid-state re-equilibration with surrounding phases following the veining. Generally, the eclogite facies minerals are most likely re-equilibrated in this sample and particularly in areas where amphibole veining is intense, whilst the amphibole compositions show signs of re-diffusion following formation. The amphibole is transitional between pargasite and edenite, whilst the preservation of omphacite (with fairly low X_{Jd} compared with other eclogites) may suggest that the amphibole formed syn-eclogite facies, if the omphacite and garnet re-equilibration is indeed related to the amphibole veins. Alternatively, the omphacite and garnet may have re-equilibrated prior to amphibole veining, but in less intensely veined areas, higher X_{Jd} values are preserved in omphacite. Overall, in sample 00-M43 the evidence presented is compelling for extensive disequilibrium, formed both by 'dry' symplectitic forming reactions and 'wet' vein forming reactions.

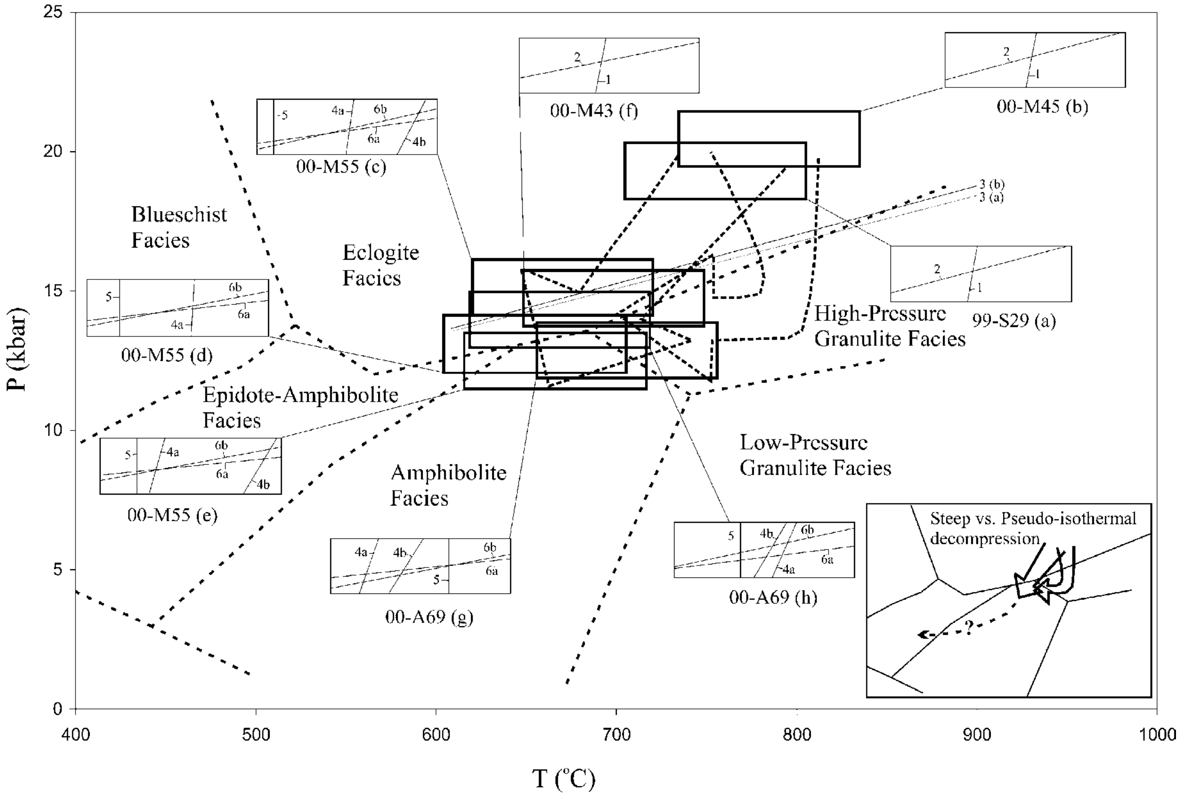
In sample 00-A69 (retrogressed streaky eclogite), in the streaks, the margins of the streaks comprise symplectitic intergrowths of amphibole (pargasite–edenite), plagioclase (oligoclase X_{Ab} *c.* 0.8; Table 1) and garnet. The garnet chemistry is variable in these areas, with the cores of larger (100–200 μm) grains having a higher Mg no. In areas away from plagioclase segregations, relict garnet porphyroblasts *c.* 1–2 mm in diameter are zoned from core to rim with X_{Mg} and X_{Fe} decreasing and X_{Ca} increasing. The relict porphyroblasts are surrounded by small garnet neoblasts that extend into and define the foliation, which are in contact with pargasite, oligoclase, titanite and occasionally small relict diopside grains. The titanite is a low-Al type and therefore formed at lower pressure than eclogite facies (C. D. Storey, unpub. Ph.D. thesis, Univ. Leicester, 2002; Brewer *et al.* 2003; Oberti *et al.* 1991). Garnet neoblasts most likely grew during eclogite facies metamorphism in equilibrium with omphacite and plagioclase as the textures and assemblages are identical to those described by Sanders (1988), whereas kyanite in the leucocratic areas was originally part of the eclogite facies paragenesis and reacted to form plagioclase and garnet during decompression within the eclogite facies. The main difference with this interpretation is that the titanite described, occasionally containing cores of rutile, is interpreted as retrograde. Further evidence for this exists, as rutile occurs as inclusions in garnet along with omphacite and is demonstrably part of the eclogite facies assemblage, whereas titanite only occurs in the matrix (with its occasional rutile cores) along with plagioclase and amphibole. Sanders (1988) also reported titanite as a product of a high-pressure excursion within the eclogite facies in samples that he studied. Evidence for re-equilibration of garnet with plagioclase and amphibole exists as symplectitic intergrowths of these three minerals. In addition, garnet has low Mg no. where either garnet neoblasts are in contact with the symplectites or where thin intergrowths of garnet occur in the symplectites.

4. Thermobarometry

P–T estimates of the peak of eclogite facies metamorphism is possible in rocks of felsic composition where plagioclase co-exists with omphacite and garnet (samples 99-S29 and 00-M45) using the GADS barometer (pyrope + 2 grossular + 3 quartz = 3 anorthite + 3 diopside: Powell & Holland, 1988) and the Grt–Cpx Fe–Mg exchange thermometer (Powell, 1985). In addition, the X_{Jd} content of omphacite provides minimum *P* estimates in all of the eclogites (Holland, 1983) when used in conjunction with the Grt–Cpx thermometer. The preferred calibration of the GADS barometer is that of Powell & Holland (1988), since they employ a robust error propagation scheme and estimate an error of ± 1 kbar for random uncertainties and ± 3 kbar for overall uncertainty. The Grt–Cpx thermometer of choice is that of Powell (1985) for similar reasons to those for the barometer and also because previous authors have used this calibration in defining *P–T* estimates for the Eastern Unit (Sanders, 1988; Rawson, Carswell & Smallwood, 2001), therefore, a degree of internal consistency can be preserved between different *P–T* estimates. X_{Jd} minimum pressures are based on the X_{Jd} isopleths of Holland (1983).

For the retrograde history, where diopside–plagioclase–quartz symplectites have formed at the expense of omphacite and the garnet rims have re-equilibrated with these symplectites, the GADS barometer and Grt–Cpx thermometer may also provide *P–T* estimates. Where amphibole–plagioclase–quartz has formed in a symplectite or as replacement rims around omphacite, both Mg and Fe calibrations of the GAPQ barometer by Kohn & Spear (1990) are found to be satisfactory (Mg calibration: 6 anorthite + 3 tremolite = 2 grossular + pyrope + 3 tschermackite + 6 quartz; Fe calibration: 6 anorthite + 3 Fe-actinolite = 2 grossular + almandine + 3 Fe-tschermackite + 18 quartz). Temperature is defined by: (a) the amphibole–plagioclase thermometer, with two calibrations based on the reactions: edenite + 4 quartz = tremolite + albite (tremolite calibration) and edenite + albite = richterite + anorthite (richterite calibration) (Holland & Blundy, 1994), and (b) the amphibole–garnet Fe–Mg exchange thermometer (Graham & Powell, 1984). There are five independent calibrations, which should intersect in *P–T* space (± uncertainties) and therefore provide a reliable estimate of both the degree of equilibration and *P–T* conditions.

The best *P–T* estimates with approximate uncertainties (± 1 kbar for barometers and ± 50 °C for thermometers) are shown in Figure 2. Peak *P–T* from two different rocks (99-S29 and 00-M45) independently give similar results of 750 °C/20 kbar and



Geothermobarometers/Reactions:

1) Grt-Cpx Fe-Mg exchange thermometer, Powell (1985)
2) Grt-Cpx-Plag (GADS) barometer, Powell & Holland (1988)
3) Jd minP (Xjd isopleths) barometer, Holland (1983)
4) Amph-Plag thermometer, Holland & Blundy (1994)
   a) Edenite-Tremolite reaction
   b) Edenite-Riebeckite reaction
5) Grt-Hbl Fe-Mg exchange thermometer, Graham & Powell (1984)
6) Grt-Amph-Plag (GAPQ) barometer, Kohn & Spear (1990)
   a) Mg end-member
   b) Fe end-member

Samples and analyses used:

Example 99-S29 (a) displayed below each flyout box - 99-S29 is the sample number; in parenthesis is the corresponding set of microprobe analyses used to calculate the thermobarometric curve, which can be found in Table 1

Figure 2. *P–T* grid displaying *P–T* estimates with fixed uncertainties of ± 50 °C and ± 1 kbar superimposed on petrogenetic grid for basaltic compositions of Oh & Liou (1998), with divisions between metamorphic facies shown as hatched lines. Large arrows display possible trajectories for decompression path from eclogite facies to amphibolite facies.

780 °C/20.5 kbar respectively (Fig. 2), well within the inferred uncertainities. Rim compositions were found to offset the *P–T* estimates relative to cores to higher values of both pressure and temperature, but various calibrations of the Grt–Cpx thermometer (Råheim & Green, 1974; Krogh, 1988; Ai, 1994; Krogh-Ravna, 2000) yielded a large spread in temperature when rim compositions were used, indicating that the rims are not in equilibrium and are therefore unsuitable for *P–T* calculation. The *P–T* results agree favourably with those calculated by Rawson, Carswell & Smallwood (2001) on ultramafic rocks (20 ± 3 kbar and 730 ± 50 °C), are slightly higher than those reported by Sanders (1988; *P* = 15 to 17 kbar and 740 °C), and are taken as a reliable estimate of peak eclogite facies conditions.

Retrograde conditions are calculated from minerals in several rocks of different composition (00-M43, 00-M55 and 00-A69). Sample 00-M43 contains considerable evidence for disequilibrium, particularly where amphibole veining is intense. However, in unveined areas, symplectitic intergrowths of diopside, plagioclase and quartz are in contact with re-equilibrated garnet rims and these provide a reasonable estimate of 15 kbar and 700 °C. A spread of temperatures calculated from Grt–Cpx calibrations (as with peak calculations from rim compositions, described above) of *c.* 100 °C may suggest that the temperature estimate is not the most reliable. *P–T* estimates of 650–700 °C and 15 kbar are provided from 00-M55 where symplectic intergrowths of amphibole, plagioclase and quartz are in contact with re-equilibrated garnet rims. Five independent calibrations intersect at one point (Fig. 2) and agree favourably with the *P–T* estimate from 00-M43 above. Sample 00-A69 provides lower *P* estimates of 13 kbar and 650–700 °C. The textural evidence for symplectic growth of garnet, amphibole, plagioclase and quartz as well as the good agreement of the five different calibrations suggests that this *P–T* estimate is fairly robust. The *P–T* paths shown in Figure 2 have trajectories that pass through the slightly higher retrograde estimates, and one possible interpretation is that the rocks have locally re-equilibrated at different stages during exhumation and that the estimates of 15 kbar and 700 °C during exhumation may well also be correct for those particular samples. Figure 2 displays a steep decompression path, which is consistent with rapid exhumation of eclogite, which is also implied by the presence of the symplectites. It is also possible that the initial decompression path followed a steeper path into the high-pressure granulite facies, and that a longer period of pseudo-isobaric cooling produced re-equilibration between the phases, resulting in the preservation of temperatures lower than high-pressure granulite facies (Fig. 2).

5. Discussion and implications

The data presented here are consistent with previous *P–T* estimates of peak eclogite facies metamorphism, which occurred at or prior to *c.* 1080 Ma (Sanders, van Calsteren & Hawkesworth, 1984). Our new data imply, however, that the *c.* 20 kbar pressure calculation from the garnet websterites by Rawson, Carswell & Smallwood (2001) more closely represents peak eclogite facies metamorphism than the *c.* 17 kbar suggested by Sanders (1988, 1989). Temperature estimates from sample 99-S29 and 00-M45 (c.750 °C and *c.* 780 °C respectively) are in agreement within error with previous estimates (Sanders, 1988, 1989; Rawson, Carswell & Smallwood, 2001). The *P–T* estimates from the post-peak high-pressure granulite and upper amphibolite facies phases are the first attempt to establish the retrograde exhumation history of the eclogites and co-facial lithologies. Insights into the amphibolite facies retrograde conditions are particularly important given that Brewer *et al.* (2003) now date this at *c.* 995 Ma. It is most likely that the *P–T* conditions following thermal re-equilibration of the Eastern Unit at amphibolite facies were *c.* 13 kbar and 650–700 °C. In terms of the decompression history, either partial re-equilibration of the Eastern Unit rocks occurred during decompression, or the rocks cooled under isobaric conditions following initial isothermal decompression into the high-pressure granulite facies stability field.

The steep decompression path implies rapid uplift of the Eastern Unit into the mid-crust following eclogite facies metamorphism, which is consistent with the high-pressure granulite facies overprint recorded by symplectite formation (O'Brien, 1997). However, presently available geochronology suggests a gap of 87 ± 25 Ma between peak and retrograde metamorphism. The maximum thermobarometric differential during this period is therefore 7 kbar and 100 °C, equivalent to *c.* 25 km denudation (tectonic and/or erosional) or an average uplift rate of < 0.3 km/Ma with a cooling rate of < 1.25 °C/Ma. This would be considered as particularly sluggish in terms of modern orogens. For example, *c.* 15 km/Ma has been suggested to account for isothermal decompression in the Sikkim Himalayas, India (Ganguly *et al.* 2000). However, given the uncertainties currently assigned to both the geochronological and geothermobarometric data, it is not prudent to discuss the geodynamic implications much further than this.

It is perhaps worthwhile considering the geodynamics of the Grenville orogen elsewhere to see if the data from the Glenelg–Attadale Inlier could correspond within that context. In eastern North America, the main phase of Grenville metamorphism, known as the Ottawan pulse (Rivers, 1997), occurred between 1100 Ma and 1000 Ma. However, there is a significant amount of new evidence to suggest that at least two distinct tectono-metamorphic events were imposed during this period. The youngest of these events occurred at *c.* 1000 Ma and is known as the Rigolet orogen (Rivers, 1997). In addition, Cox, Indares & Dunning (2002) concluded that metamorphism within the orogen is diachronous, but explain this phenomenon as a result of proximity to the Grenville Front, whereby cooling ages young to the southwest away from this boundary. In this case, exhumation and cooling rates are most rapid close to the Grenville Front as a result of NW-directed thrusting, whereas further southwest, where exhumation was achieved by extension above the thickened orogenic pile to the southeast, uplift and cooling were rather more sluggish, exacerbated by punctuated high-*T* events presumably related to magmatism. In the Manicouagan Imbricate Zone of the eastern Grenville Province, Cox, Indares & Dunning (2002) allude to an early pseudo-isothermal decompression of eclogites, followed by a prolonged slower period of cooling, contemporaneous with rapid cooling within units at higher tectonic levels. They suggest that the slower cooling rates of the eclogite, following initial rapid uplift, were a result of tectonic exhumation of the upper tectonic levels by thrusting over the deeper rocks, effectively keeping the deeper tectonic levels deeper and hotter for a longer period of time, resulting in slower cooling rates.

It is perhaps not surprising that rocks from the Glenelg–Attadale Inlier, a relatively minor sliver of exposed basement, preserve only two points in the thermotectonic history of Grenville orogenesis. It is plausible, given the data from North America, that the Glenelg–Attadale Inlier may have been located away from the Grenville Front during collisional orogenesis, and so underwent the much slower cooling and uplift rates that typify rocks to the southwest away from the Grenville Front, in the higher pressure orogenic interior (Cox, Indares & Dunning, 2002). Indeed, the initial phase of isothermal decompression followed by slower cooling of high-pressure rocks in the Grenville could well accord with the situation in the Glenelg–Attadale Inlier. However, it is acknowledged that better age constraints on the peak eclogite facies metmorphism are required in order to arrive at an enhanced geodynamic understanding of the orogen in the northwest Highlands of Scotland before any reliable wider correlations can be made.

This study also bears on the relationship of the Glenelg–Attadale Inlier to the surrounding Moine Supergroup metasediments. The latter were deposited between *c.* 1000 and 870 Ma (Friend *et al.* 2003), although recent work has suggested that the Loch Eil Group, the structurally and stratigraphically highest level of the Moine Supergroup, has a maximum deposition age of 883 ± 35 Ma (Cawood *et al.* 2004). In this study we are concerned with the lowermost structural and stratigraphic part of the Moine Supergroup, known as the Morar Group, which does not contain zircons unequivocally younger than *c.* 1000 Ma (Friend *et al.* 2003). Traditionally, rocks of the Morar Group are interpreted as being deposited unconformably upon basement represented by the Glenelg–Attadale Inlier (Ramsay, 1958). Brewer *et al.* (2003) note that the exhumation of the Eastern Unit eclogites was achieved partly by ductile shearing, with the shear zone along the contact of the Eastern Unit and Western Unit representing a structure during this process. However, this shear zone contains discontinuous slivers of Morar Group metasediments that appear to have undergone kinematically identical shearing to that seen in the eclogite at amphibolite facies (Storey, Brewer & Parrish, 2004) and, hence, most likely during the same event as the retrogression of the Eastern Unit. Given the *c.* 995 Ma age of retrogression and coeval shearing of the Eastern Unit (Brewer *et al.* 2003), there is very little time available to exhume and erode the Glenelg–Attadale Inlier, deposit the entire Morar Group and then metamorphose the entire sequence to amphibolite facies by *c.* 995 Ma. Moreover, the petrological evidence indicates that the *c.* 995 Ma age is dating retrogression from eclogite facies rather than prograde metamorphism related to crustal thickening. So, in order to explain both the petrological evidence and the geochronological evidence, this would require another phase of eclogite facies metamorphism of both basement and cover following the earlier > 1080 Ma phase affecting only the basement (Sanders, van Calsteren & Hawkesworth, 1984), for which there is no evidence from the Glenelg–Attadale Inlier and certainly no evidence from the Moine. One possible alternative is that at least the Morar Group of the Moine Supergroup is allochthonous with respect to basement of Eastern Unit type. Although undated, the eclogite of the Western Unit is assumed to be coeval with those from the Eastern Unit, in which case the assumption of an unconformity between the entire Glenelg–Attadale Inlier and the Morar Group must be treated with caution.

6. Conclusions

Peak *P–T* conditions within the Eastern Unit of the Glenelg–Attadale Inlier are confirmed as being in the region of *c.* 20 kbar and 750–780 °C, in agreement with the previous estimate of Rawson, Carswell & Smallwood (2001). Estimates of retrograde amphibolite facies *P–T* conditions suggest that local re-equilibration occurred within different samples, providing a partial record of the retrograde *P–T* path, or that an initially isothermal decompression path has been overprinted by thermal re-equilibration to lower temperatures in an isobaric environment. With a lowest *P–T* estimate of *c.* 13 kbar and 650–700 °C, the calculated retrograde path in either case is steep, which suggests decompression during relatively rapid uplift. However, the trajectory of the path after this point in the retrograde evolution is unclear. Peak eclogite facies metamorphism occurred at > 1080 Ma (Sanders, van Calsteren & Hawkesworth, 1984), whereas amphibolite facies retrogression occurred at *c.* 995 Ma (Brewer *et al.* 2003). These values translate to an average uplift rate of < 0.3 km/Ma with a cooling rate of < 1.25 °C/Ma. This is a rather low cooling rate compared to other parts of the Grenville orogeny and a low uplift rate compared with modern orogens. However, the initial phase of uplift could have been rapid, followed by a prolonged period of slower, pseudo-isobaric cooling, which would be analogous to the Manicouagan Imbricate Zone of the eastern Grenville province (Cox, Indares & Dunning, 2002). The possibility remains that there may be discrete events within the *c.* 1100–1000 Ma Grenville orogenic cycle that are not recorded within the small exposure of rocks within the Glenelg–Attadale Inlier.

When the data of Sanders, van Calsteren & Hawkesworth (1984) are taken into account, the evidence presented in this study brings into question the hypothesis that the Moine Supergroup was deposited unconformably upon basement of Glenelg–Attadale Inlier type. Further work is now required to understand the relationship of the Moine Supergroup with the basement and the *P–T–t* evolution of the Moine, in order to understand when and how the basement and Moine 'cover' were juxtaposed.

Acknowledgements. Many thanks to Rob Wilson at Leicester for assistance with EPMA. CS acknowledges a studentship from the University of Leicester.

References

AI, Y. 1994. A revision of the garnet–clinopyroxene Fe^{2+}–Mg exchange geothermometer. *Contributions to Mineralogy and Petrology* **115**, 467–73.

ALDERMAN, A. R. 1936. Eclogites in the neighbourhood of Glenelg, Inverness-Shire. *Quarterly Journal of the Geological Society of London* **92**, 488–528.

BREWER, T. S., STOREY, C. D., PARRISH, R. R., TEMPERLEY, S. & WINDLEY, B. F. 2003. Grenvillian age exhumation of eclogites in the Glenelg–Attadale Inlier, NW Scotland. *Journal of the Geological Society, London* **160**, 565–74.

CAWOOD, P. A., NEMCHIN, A. A., STRACHAN, R. A., KINNY, P. D. & LOEWY, S. 2004. Laurentian provenance and an intracratonic tectonic setting for the Moine Supergroup, Scotland, constrained by detrital zircons from the Loch Eil and Glen Urquhart successions. *Journal of the Geological Society, London* **161**, 861–74.

CARSWELL, D. A. 1990. Eclogites and the eclogite facies: definitions and classifications. In *Eclogite Facies Rocks* (ed. D. A. Carswell), pp. 1–14. Glasgow and London: Blackie.

COX, R. A., INDARES, A. & DUNNING, G. R. 2002. Temperature-time paths in the high-P Manicougan Imbricate zone, eastern Grenville Province: Evidence for two metamorphic events. *Precambrian Research* **117**, 225–50.

DROOP, G. T. R. 1987. A general equation for estimating Fe^{3+} concentrations in ferromagnesian silicates and oxides from microprobe analyses, using stoichiometric criteria. *Mineralogical Magazine* **51**, 431–5.

FLORENCE, F. P. & SPEAR, F. S. 1991. Effects of diffusional modification of garnet growth zoning on *P–T* path calculations. *Contributions to Mineralogy and Petrology* **107**, 487–500.

FRIEND, C. R. L., STRACHAN, R. A., KINNY, P. D. & WATT, G. R. 2003. Provenance of the Moine Supergroup of NW Scotland: evidence from geochronology of detrital and inherited zircons from (meta)sedimentary rocks, granites and migmatites. *Journal of the Geological Society, London* **160**, 247–57.

GANGULY, J., DASGUPTA, S., CHENG, W. & NEOGI, S. 2000. Exhumation history of a section of the Sikkim Himalayas, India: records in the metamorphic mineral equilibria and compositional zoning of garnet. *Earth and Planetary Science Letters* **183**, 471–86.

GRAHAM, C. M. & POWELL, R.1984. A garnet-hornblende geothermometer: calibration, testing, and application to the Pelona Schist, Southern California. *Journal of Metamorphic Geology* **2**, 13–31.

HOLLAND, T. J. B. 1983. The experimental determination of activities in disordered and short-range ordered jadeitic pyroxenes. *Contributions to Mineralogy and Petrology* **82**, 214–20.

HOLLAND, T. & BLUNDY, J. 1994. Non-ideal interactions in calcic amphiboles and their bearing on amphibole–plagioclase thermometry. *Contributions to Mineralogy and Petrology* **116**, 433–47.

KOHN, M. J. & SPEAR, F. S. 1990. Two new geobarometers for garnet amphibolites, with applications to southeastern Vermont. *American Mineralogist* **75**, 89–96.

KROGH, E. J. 1988. The garnet–clinopyroxene Fe–Mg geothermometer – a reinterpretation of existing experimental data. *Contributions to Mineralogy and Petrology* **99**, 44–8.

KROGH-RAVNA, E. 2000. The garnet–clinopyroxene Fe^{2+}–Mg geothermometer: an updated calibration. *Journal of Metamorphic Geology* **18**, 211–19.

LEAKE, B. E., WOOLLEY, A. R., ARPS, C. E. S., BIRCH, W. D., GILBERT, M. C., GRICE, J. D., HAWTHORNE, F. C., KATO, A., KISCH, H. J., KRIVOVICHEV, V. G., LINTHOUT, K., LAIRD, J., MANDARINO, J., MARESCH, W. V., NICKEL, E. H., ROCK, N. M. S., SCHUMACHER, J. C., SMITH, D. C., STEPHENSON, N. C. N., UNGARETTI, L., WHITTAKER, E. J. W. & YOUZHI, G. 1997. Nomenclature of amphiboles: report of the subcommittee on amphiboles of the International Mineralogical Association Commission on New Minerals and Mineral Names. *Mineralogical Magazine* **61**, 295–321.

MAY, F., PEACOCK, J. D., SMITH, D. I. & BARBER, A. J. 1993. *Geology of the Kintail district.* Memoir of the British Geological Survey, Scotland, Sheet 72W and part of 71E.

MERCY, E. L. P. & O'HARA, M. J. 1968. Nepheline normative eclogite from Loch Duich, Ross-Shire. *Scottish Journal of Geology* **4**, 1–9.

MORIMOTO, N. 1988. Nomenclature of pyroxenes. *Mineralogical Magazine* **52**, 535–50.

O'BRIEN, P. J. 1997. Garnet zoning and reaction textures in overprinted eclogites, Bohemian Massif, European Variscides: a record of their thermal history during exhumation. *Lithos* **41**, 119–33.

OBERTI, R., SMITH, D. C., ROSSI, G. & CAUCIA, F. 1991. The crystal-chemistry of high-aluminium titanites. *European Journal of Mineralogy* **3**, 777–92.

OH, C. W. & LIOU, J. G. 1998. A petrogenetic grid for eclogite and related facies under high-pressure metamorphism. *Island Arc* **7**, 36–51.

PEACH, B. N., WOODWARD, H. B., CLOUGH, C. T., HARKER, A. & WEDD, C. B. 1910. *The geology of Glenelg, Lochalsh and South East part of Skye.* Memoir of the Geological Survey of Scotland, Sheet 71.

POWELL, R. 1985. Regression diagnostics and robust regression in geothermometer/geobarometer calibration: the garnet–clinopyroxene geothermometer revisited. *Journal of Metamorphic Geology* **3**, 231–43.

POWELL, R. & HOLLAND, T. J. B. 1988. An internally consistent dataset with uncertainties and correlations: 3. Applications to geobarometry, worked examples and a computer program. *Journal of Metamorphic Geology* **6**, 173–204.

RÅHEIM, A. & GREEN, D. H. 1974. Experimental determination of the temperature and pressure dependence of the Fe–Mg partition coefficient for coexisting garnet and clinopyroxene. *Contributions to Mineralogy and Petrology* **48**, 179–203.

RAMSAY, J. G. 1958. Moine-Lewisian relations at Glenelg, Inverness-shire. *Quarterly Journal of the Geological Society of London* **113**, 487–520.

Rawson, J. R., Carswell, D. A. & Smallwood, D. 2001. Garnet-bearing olivine–websterite within the Eastern Glenelg Lewisian of the Glenelg Inlier, NW Highlands. *Scottish Journal of Geology* **37**, 27–34.

Rivers, T. 1997. Lithotectonic elements of the Grenville Province: review and tectonic implications. *Precambrian Research* **86**, 117–54.

Sanders, I. S. 1979. Observations on eclogite- and granulite-facies rocks in the basement of the Caledonides. In *The Caledonides of the British Isles – reviewed* (eds A. L. Harris, C. H. Holland and B. E. Leake), pp. 97–100. Geological Society of London, Special Publication no. 8.

Sanders, I. S. 1988. Plagioclase breakdown and regeneration reactions in Grenville kyanite eclogite at Glenelg, NW Scotland. *Contributions to Mineralogy and Petrology* **98**, 33–9.

Sanders, I. S. 1989. Phase relations and P–T conditions for eclogite-facies rocks at Glenelg, north-west Scotland. In *Evolution of metamorphic belts* (eds J. S. Daly, R. A. Cliff and B. W. D. Yardley), pp. 513–17. Geological Society of London, Special Publication no. 43.

Sanders, I. S., van Calsteren, P. W. C. & Hawkesworth, C. J. 1984. A Grenville Sm–Nd age for the Glenelg eclogite in northwest Scotland. *Nature* **312**, 439–40.

Storey, C. D., Brewer, T. S. & Parrish, R. R. 2004. Late-Proterozoic tectonics in northwest Scotland: one contractional orogeny or several? *Precambrian Research* **134**, 227–47.

Sutton, J. & Watson, J. 1959. Structures in the Caledonides between Loch Duich and Glenelg, north-west Highlands. *Quarterly Journal of the Geological Society of London* **114**, 231–55.

Teall, J. J. H. 1891. On an eclogite from Loch Duich. *Mineralogical Magazine* **9**, 217–18.