The Glenelg-Attadale Inlier, NW Scotland, with emphasis on the Precambrian high-pressure metamorphic history and subsequent retrogression: an introduction and review

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Synopsis

The Glenelg-Attadale Inlier of the NW Highlands of Scotland represents the largest area of exposed basement within the Moine Supergroup and, thus, affords the best opportunity of studying the pre-Moine history and the relationship of the basement to the Moine. The Inlier can be divided into Western and Eastern Units, based on lithological associations, and the boundary is marked by a major ductile shear zone with a dominantly reverse (top-to-the-west) sense of shear. The presence of eclogites and of other eclogite facies rocks within the inlier is the only unequivocal examples of crustal eclogites exposed within the British Isles. Eclogites in the Eastern Unit (P-T *c.* 20 kbar and 750–780 °C) has been radiometrically dated (Sm–Nd garnet–omphacite–whole rock) to *c.* 1082 Ma and their retrogression to amphibolite facies (P-T conditions: minimum 14–16 kbar and *c.* 650 °C) have been dated (U–Pb zircon) to *c.* 995 Ma. This demonstrates that the Eastern Unit underwent tectonic thickening to *c.* 70 km during the Grenvillian orogeny and was decompressed into the mid-crust by the end of the Grenvillian orogenic cycle. Rare eclogites in the Western Unit appear to have been formed earlier in the Palaeoproterozoic (*c.* 1700–1750 Ma) and their juxtaposition against the Eastern Unit eclogites may be fortuitous or may reflect a fundamental tectonic manifestation of Wilson Cycle tectonics in orogenic belts. The basal Moine shows a deformed unconformable relationship against the Western Unit. However, the Eastern Unit is entirely enveloped by profound shear zones against the Western Unit and the Moine Supergroup, suggesting that the Eastern Unit is allochthonous. Direct dating of mylonites within the shear zone between the Western and Eastern Units, containing coaxially sheared Moine metasediments, suggests an age of *c.* 670 Ma and this may reflect the time at which the units were juxtaposed and the Moine infolded. Final deformation and metamorphism of the Glenelg-Attadale Inlier and the Moine occurred during the Caledonian orogeny, with major folding and extensional shearing operating sometime after *c.* 437 Ma, resulting in distinctive fold interference patterns. Terminal thrusting of the Moine and GAI westward over the Laurentian Caledonian foreland occurred at *c.* 435–430 Ma.

Introduction

The Glenelg-Attadale Inlier (GAI) of the NW Highlands of Scotland (Fig. 1) represents an enigmatic piece of basement exposed within the Neoproterozoic Moine Supergroup metasedimentary succession. The presence of coastal eclogites, the only unequivocal example preserved in the British Isles, has been known for some time but not truly understood in terms of their bearing on the tectono-metamorphic evolution of other basement units and the overlying metasediments in the NW Highlands of Scotland. Teall (1891) first recorded the presence of eclogite within the GAI and C. T. Clough (in Peach *et al.* 1910) first comprehensively mapped the area, recording numerous examples of eclogites. Alderman (1936) provided a detailed study of the petrography and some early geochemical analyses of the eclogites. Ramsay (1958); Sutton & Watson (1959) studied the structural evolution of the GAI and surrounding Moine. Ramsay (1958) in particular noted marked differences in the lithological associations between rocks lying to the east of a strip of Moine metasediments (the Moine strip; Fig. 2) within the GAI and those to the west. He referred to these as the eastern and western Lewisian, described here as the Eastern Unit (EU) and Western Unit (WU). Ramsay (1958) implied that the basement in the GAI is correlated with the Lewisian basement of the Caledonian foreland. The EU contains abundant eclogite (up to 25% of the outcrop) along with manganiferous, aluminous and calcareous pelites, marbles and intermediate to acid gneisses. In contrast, the WU contains predominantly acid gneisses (tonalite–trondhjemite–granodiorite), subordinate metabasites and no metasediments. Ramsay (1958) showed that the relationship between the basement and the overlying Moine was one of a basement-cover, with the Moine deposited unconformably upon the basement. This view was based on a small angular discordance (up to 15°) between the WU and Moine on the western margin of the area and the presence of deformed metaconglomerate at the base of the Moine.

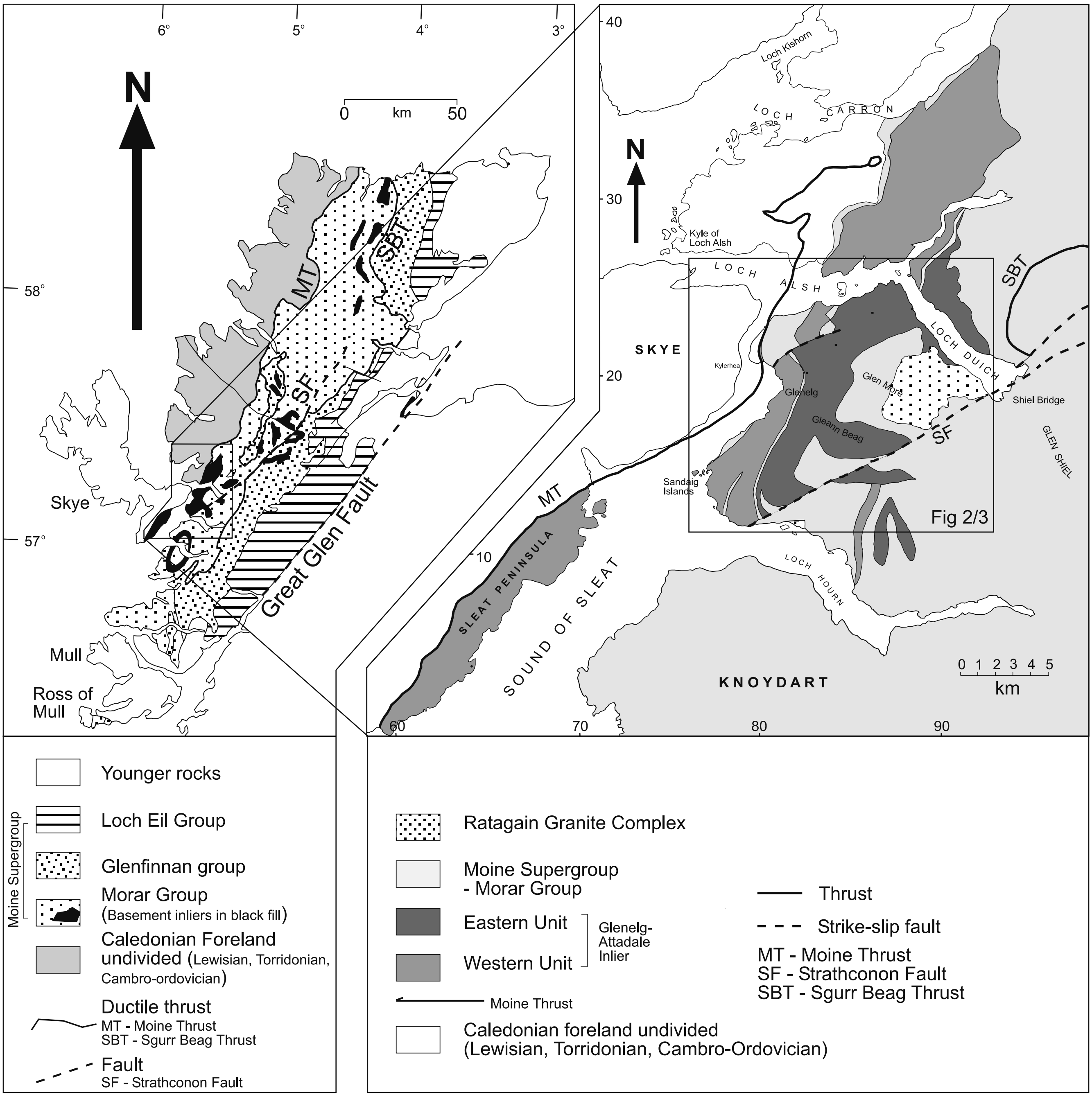

FIG. 1. Location maps. (**a**) Scotland with geology north of the Great Glen Fault. (**b**) Inset box geological map of Glenelg-Attadale Inlier modified after Peach *et al.* (1910); May *et al.* (1993). Note: grid lines are from the Great Britain Grid Reference and all are prefixed NG. See inset box for location of map in Fig. 2.

Sanders (1979, 1988, 1989) provided the first thermobarometric estimates for the EU eclogites and concluded that they reached peak metamorphic conditions of *c.* 15–18 kbar and 700–730 °C. Rawson *et al.* (2001) recorded the occurrence of websterite in the EU and provided P–T estimates of 20 ± 3 kbar and 730 ± 5 °C. Storey *et al.* (2005) investigated eclogite facies felsic gneisses and attained peak PT estimates of *c.* 20 kbar and 750–780 °C, in broad agreement with that recorded in the websterites. Storey *et al.* (2005) also attempted to define the P–T conditions of amphibolite facies retrogression of the EU eclogites and obtained estimates of *c.* 13 kbar and 650–700 °C.

The EU eclogites have been dated by the Sm–Nd garnet–omphacite–whole rock method, producing an isochron age of 1082 ± 24 Ma, interpreted as close to the peak age of eclogite facies metamorphism (Sanders *et al.* 1984). Thus, there is compelling evidence that the EU eclogites were formed during the Grenvillian orogeny. This is an important conclusion, not least because evidence of Grenvillian metamorphism is sparse in the British Isles, with the closest example being from the west coast of Ireland in the Annagh Gneiss Complex (Daly 1996).

The purpose of this article is to review the evidence for the tectono-metamorphic evolution of the GAI, with

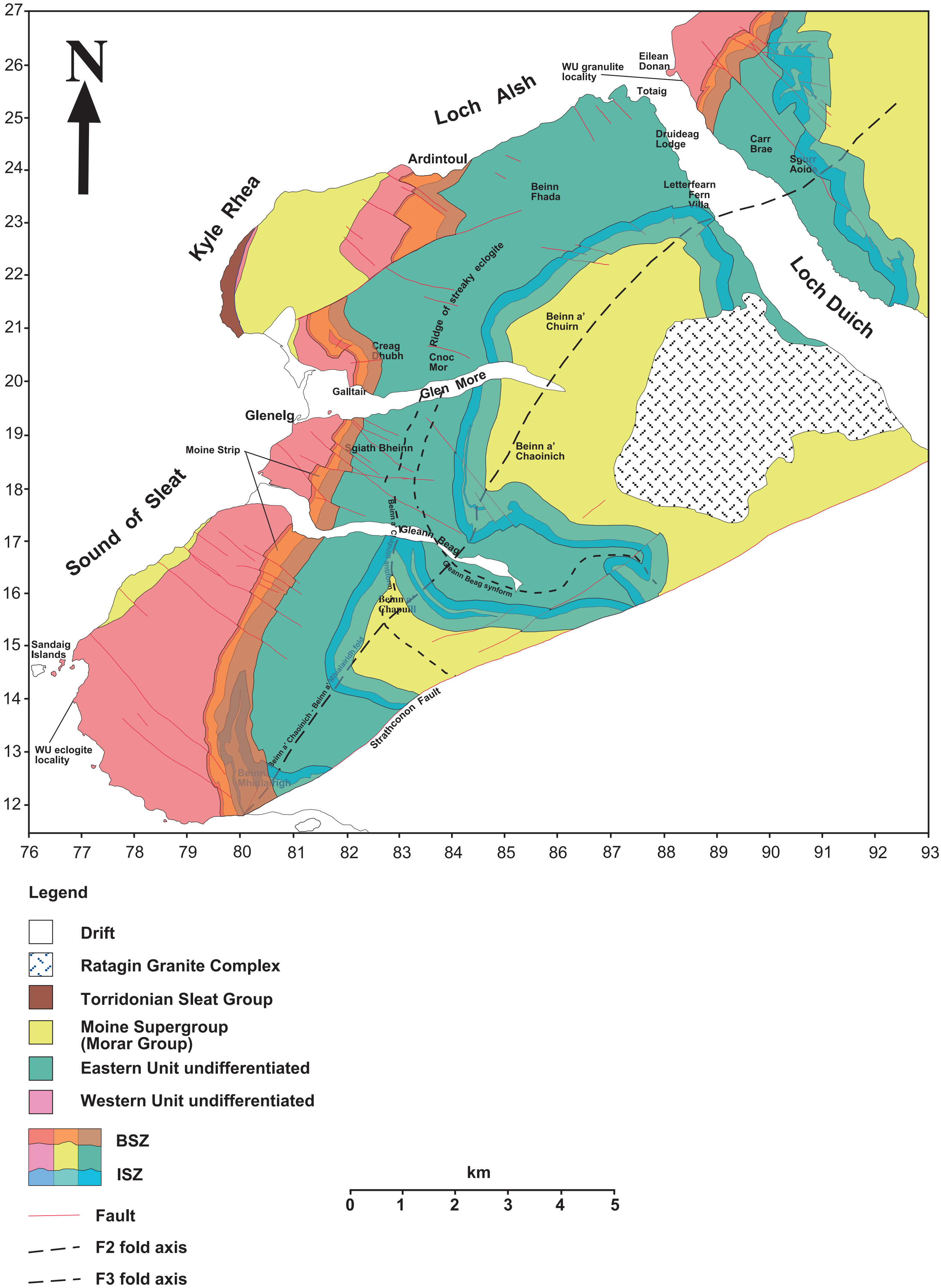

FIG. 2. Summary map of the GAI showing key areas and structures.

particular emphasis on Proterozoic high-pressure metamorphism, to allow researchers to follow the separate field guides that accompany this article and derive their own, more complete, interpretations.

Background to Proterozoic tectonics and sedimentation within the NW Highlands of Scotland

The topography of the NW Highlands of Scotland is dominated by the Caledonian Mountains, formed during the Lower Palaeozoic Caledonian orogeny. To the west, the Caledonian Foreland comprises generally low-lying topography with hummocky hills and intervening small lochs (Cnoc and Lochan topography). The foreland is dominated by deeply eroded Archaean to Palaeoproterozoic gneissic basement rocks of the Lewisian Gneiss Complex, overlain by unmetamorphosed Mesoproterozoic to Neoproterozoic continent-derived sediments of the Torridonian. To the east, the foreland is overridden by the Caledonides along the Moine Thrust Zone that carries the Moine Supergroup and minor slices of variably imbricated basement rocks, commonly termed Lewisianoid inliers, including the GAI.

The Moine Supergroup, sheared and infolded into the GAI, is a predominantly metasedimentary sequence comprising psammites, pelites and semipelites. It can be divided into a tripartite succession, comprising stratigraphically a lower Morar Group, the Glenfinnan Group and an upper Loch Eil Group (Johnstone 1975; Holdsworth *et al.* 1994 and references therein). The Morar Group forms part of the Moine Nappe whereas the Glenfinnan and Loch Eil Groups belong to the Sgurr Beag Nappe, separated from the Moine Nappe by the Sgurr Beag Thrust. Based on detrital zircon evidence, the Morar Group has a maximum age of *c.* 980 Ma (Peters 2001; Friend *et al.* 2003), the Glenfinnan Group a maximum age of *c.* 950 Ma (Friend *et al.* 2003) and the Loch Eil Group a maximum age of *c.* 900 Ma (Cawood *et al.* 2004; Emery 2005). The absolute age of sedimentation is constrained by the West Highland granite gneiss at *c.* 870 Ma (Friend *et al.* 1997) that intrudes the Sgurr Beag Nappe.

The Moine rocks are thought to have been deposited within two major rift basins, developed as half-grabens upon the Laurentian margin basement (Soper *et al.* 1998); that is, within an extensional continental setting. However, this model was criticized by Cawood *et al.* (2004) on the basis of sedimentological evidence. It has been suggested that the Neoproterozoic history of the NW Highlands of Scotland was one of continued extension without tectonic break, and with the conformable deposition of the Dalradian Supergroup upon the Moine Supergroup, culminating in the formation of the Iapetus Ocean (Soper *et al.* 1998; Dalziel & Soper 2001). However, a growing body of evidence suggests that a number of tectono-metamorphic events affected the Moine during the Neoproterozoic and at least some of these were compressional (see discussion below). Cawood *et al.* (2004) recently reviewed possible tectonic settings for the deposition of the Moine and concluded that it was most likely that it formed in an intracratonic basin within northeastern Laurentia. However, Kirkland *et al.* (2007) suggested that the Moine and putative correlative successions (the Sørøy Succession of the Kalak Nappe, Baltica; the Seve Nappe Complex) were deposited as successor basins on top of the developing Grenville orogen. This implies that the depositional loci of these basins were actually at the periphery of Rodinia in a free-ocean setting, allowing arc processes to operate.

The Moine sequence has attracted numerous controversies over the years. The monotony of the psammitic lithologies has made metamorphic studies difficult and, likewise, the geochronology of metamorphic events. Until recently, it was considered that the Moine was first metamorphosed during the Grenvillian orogeny (Brook *et al.* 1976; Brewer *et al.* 1979) and was thus >1000 Ma in age. Re-dating of the West Highland Granite Gneiss and detrital zircon evidence (Friend *et al.* 1997) has disproved this theory but the age of the West Highland Granite Gneiss has helped to engender a new debate. What was originally considered to be a second period of Proterozoic metamorphism was defined by a suite of pegmatites, intruded widely throughout the Moine, and dated to 700–780 Ma (Van Breemen *et al.* 1974; 1978). These were considered to represent a major tectonothermal event that has come to be known as the Knoydartian event (Bowes 1968). The *c.* 870 Ma age of the West Highland Granite Gneiss was originally interpreted as a new age for the onset of Knoydartian orogenesis in a compressional setting (Friend *et al.* 1997). However, dating of associated basic rocks to *c.* 870 Ma (Miller 1999) suggests that this period was one of extension and high heat flow, and this has also been suggested, based on petrologic information (Zeh & Millar 2001). Prograde amphibolite facies ages of 820–790 Ma (peak P-T: 687–707 °C and 12.5–14.5 kbar; Vance *et al.* 1998), *c.* 737 Ma (greenschist to amphibolite facies; Tanner & Evans 2003) and *c.* 670 Ma (amphibolite facies and ductile compressional shearing; Storey *et al.* 2004) have questioned the validity of a single compressional orogenic event. For further reviews see Strachan *et al.* (2002) and Oliver (2002).

Within the Central Highland Migmatite Complex of the northern Highlands, the Dava Succession and Glen Banchor successions form the lowest structural part of the Grampian Group and may form the basement to the Grampian (Highton *et al.* 1999). The age of migmatization and gneissification of these rocks is *c.* 840 Ma (Highton *et al.* 1999), and shear zones that cut the migmatites and form along the contact between the migmatites and the overlying Grampian Group have been dated at *c.* 800 Ma (Noble *et al.* 1996). The best estimate for peak metamorphic conditions in this interval is 7–8 kbar and 500–600 °C (kyanite grade; Phillips *et al.* 1999). The Dava Succession also contains undated relicts of eclogites (Baker 1986) that may have formed in this interval. Whilst the origin of the Dava and Glen Banchor successions is equivocal, it is possible that they correlate with parts of the Moine Supergroup.

The cause of the Neoproterozoic tectono-metamorphic events within the Moine has recently been debated, with Cawood *et al.* (2004) suggesting, based on their preferred model with the Moine deposited in an intracontinental setting, that the compressional events were the result of far-field forces operating around the margin of Rodinia. In contrast, Kirkland *et al.* (2007), with their model of the Moine deposited as a successor basin outboard of Rodinia, suggested that arc processes and related accretion in a free-ocean setting could account for the pulses of compressional events. Growing evidence suggests that the Moine underwent a Neoproterozoic tectono-metamorphic evolution like that of basins of similar age in Scandinavia (the Sørøy Succession and the Seve Nappe Complex). In Scandinavia, tectono-metamorphic events are recorded at *c.* 840 Ma (Porsanger orogeny; Kirkland *et al.* 2006), *c.* 710 Ma (Snøfjord; Kirkland *et al.* 2006), *c.* 670 Ma (the Eidvågeid Paragneiss; Kirkland *et al.* 2007) and *c.* 637 Ma (the Skárjá granitic gneiss; Rehnstrom *et al.* 2002).

Ramsay (1958), based on the presence of conglomerates within the lowermost parts of the Morar close to the contact with the GAI (WU), along with up to 15° discordance between composite fabric in the GAI (WU) and the Morar, regarded the relationship between the GAI and the overlying Morar Group as unconformable. Holdsworth *et al.* (1994) discussed evidence elsewhere in the Scottish Highlands of conglomerates and cross-bedding within the Moine rocks close to the contact with basement inliers, which they argue indicates unconformable deposition of the Moine upon the contiguous basement. However, Temperley & Windley (1997) challenged this interpretation, at least with respect to the GAI, and presented evidence of a ductile shear zone (extensional detachment) between the GAI and the Morar Group. Storey *et al.* (2004) argued that a major shear zone is always present between the GAI (EU) and the Morar and that there is no strong evidence in this area for an unconformity. However, where the WU is in contact with the Moine the discordant contact and deformed metaconglomerate are likely to represent a modified unconformity. Hence, it is possible that only some of the exposed basement inliers within the Moine were exposed at the time of deposition of the Moine, with others representing allochthonous slices juxtaposed during later shearing events.

General lithological description

Eastern Unit

A detailed map of the EU is shown in Fig. 3 demonstrating the modal proportions of the different lithologies.

The most common lithology overall is a grey trondhjemitic gneiss, locally preserving migmatitic textures where the strain is low. The gneiss consists of banded quartz-feldspar and amphibole-biotite, the latter minerals sometimes occurring as clots or schlieren within the quartzo-feldspathic matrix. A conspicuous but less common trondhjemitic gneiss contains locally abundant garnet and occasional kyanite and omphacite, with evidence that they have undergone eclogite facies metamorphism. The second most common lithology is eclogite or its retrogressed equivalent amphibolite, forming up to 25% of the outcrop. A high proportion of the EU crust therefore had a basic protolith, contrasting with the adjacent WU. The EU also contains conspicuous metasediments, absent from the WU, in the form of aluminous metapelites that locally contain garnet and kyanite, manganiferous metapelites with garnet and phengitic mica, calc-silicate rocks with abundant epidote, and marbles with nodules of diopside, tremolite and forsterite. Associated with the sediments are rare eulysites, first described by Tilley (1936). These are rich in iron and manganese and occur either with an anhydrous assemblage containing fayalite, iron hypersthene, hedenbergite and pyroxmangite or in a hydrous group containing grunerite and other amphiboles such as cummingtonite and actinolite. Finally, minor exposures of ultrabasic lithologies locally include lenticular bodies of serpentinite of the order of 1–2 m in length. In one area, garnet-bearing olivine websterite formed during eclogite facies metamorphism (Rawson *et al.* 2001). In all, a range of lithologies can be demonstrated as being cofacial during peak eclogite facies metamorphism and indicate that apart from a few minor Neoproterozoic granitic melt veins, all of the lithologies within the EU were buried to around 70 km.

Western Unit

The WU is dominated by felsic trondhjemitic gneisses. These are commonly banded with variegated grey and pink felsic layers and variable proportions of amphibole, and with minor clots and larger lenticles (1–3 m diameter) of ultrabasic rock. Amphibolite sheets cut across the early banding that are superficially similar to Scourie dykes. These are speckled in appearance due to the presence of trondhjemitic leucosome patches formed during partial melting under granulite facies conditions. There are no metasedimentary rocks in the WU. Granulite facies metamorphism affected the unit at around 2.7 Ga (Storey 2002) and induced widespread partial melting of basic lithologies, leaving behind a residue of basic granulite. Eclogite facies metamorphism affected the WU but as eclogite is found in only one small area, evidence of this event affecting other lithologies is scant. The remaining basic rocks are predominantly amphibolites with evidence of being retrogressed granulites.

Eclogite facies and high-pressure granulite facies petrology

Eastern Unit

The EU comprises a variety of lithologies that have demonstrably been metamorphosed to eclogite facies. Eclogite *sensu stricto* generally comprises a granoblastic polygonal growth of garnet and omphacite, with minor rutile/ilmenite intergrowths and quartz. However,

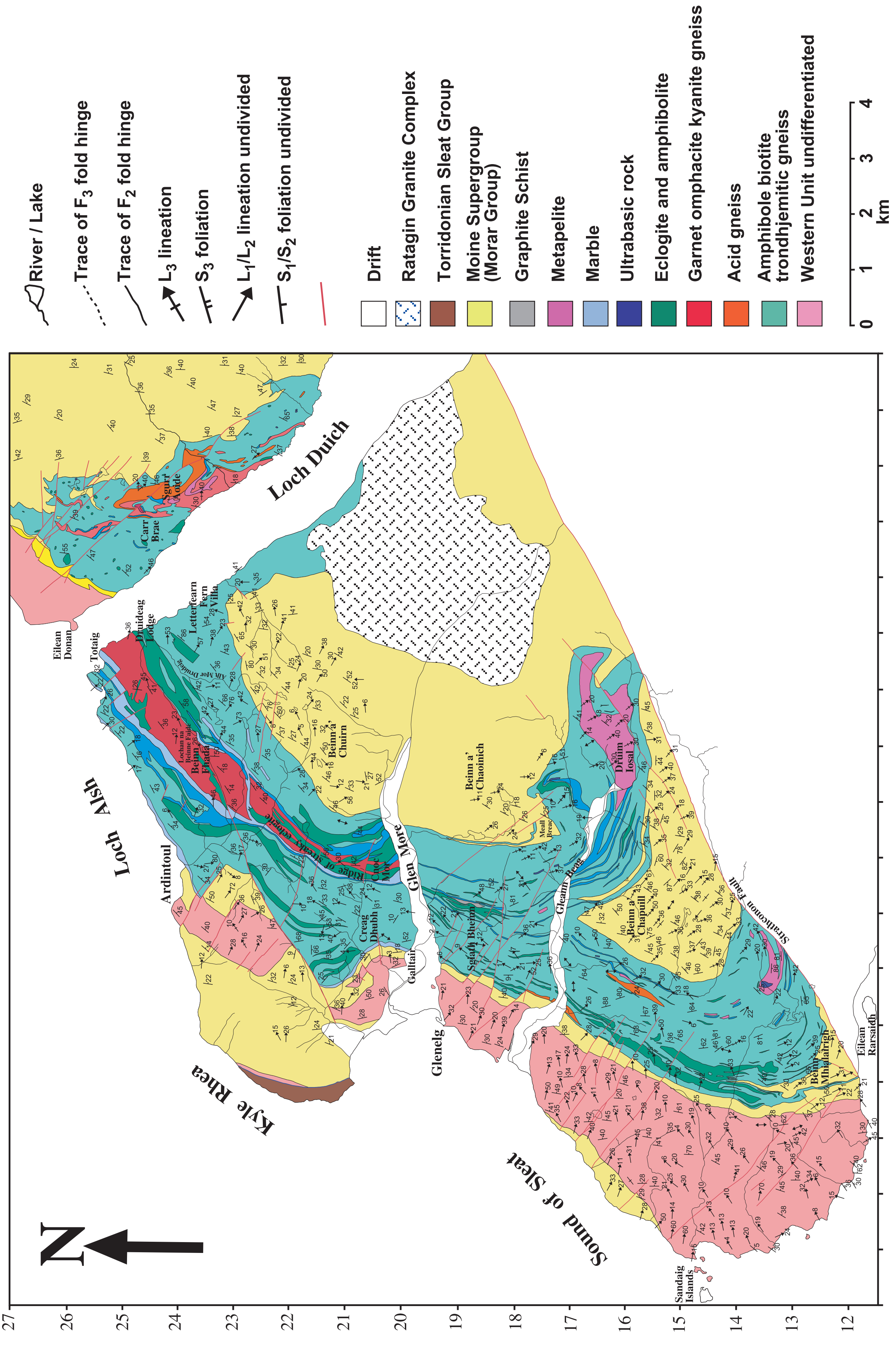

FIG. 3. Detailed map of GAI with lithological division of EU. Structural data also included. Maps modified after Ramsay (1958; south of Glen More), Sutton & Watson (1959; north of Glen More, south of Loch Duich) and May *et al.* (1993; north of Loch Duich).

almost ubiquitously within the EU it shows some degree of retrogression, with omphacite being replaced by symplectites of diopside, plagioclase and quartz during early anhydrous decompression. Omphacite and garnet typically display some replacement by amphibole (pargasite) and plagioclase, and the rutile/ilmenite overgrowths are rimmed and partially replaced by framboidal titanite. Later retrogression is recognized as epidote replacing garnet, presumably during cooling of the EU.

Felsic eclogite facies rocks, first recorded by Sanders (1988), were referred to as streaky eclogite. They occur predominantly along a 2–3 km ridge, striking NNE–SSW, to the north of Glen More (marked on map as ridge of streaky eclogite; Fig. 2). They are characterized by ubiquitous white threads composed of quartzo-feldspathic material containing kyanite, quartz, oligoclase, K-feldspar, minor scapolite and biotite. The streaks were considered by Sanders (1988) to represent sites of former plagioclase that have reacted with the surrounding eclogite to produce aligned kyanite, that locally forms augens surrounding knots of eclogite. There are two important implications from these observations: the prograde reaction occurred during deformation, and the rock was already at eclogite facies prior to shearing. More recent observations by the author have shown that the deformation was non-coaxial as kyanite grains not only grow preferentially, with their *c*-axes aligned parallel to the principal stretching direction, but are also asymmetric with fish-type morphology, attesting to non-coaxial deformation. It is not clear what the original feldspathic streaks represent, but one interpretation, based on textural relationships, is that they represent partial melt veins, formed at or prior to eclogite facies metamorphism. The retrograde reaction history of the streaks resulted in oligoclase growing between kyanite and quartz, and kyanite and omphacite. Sanders (1988) considered the reaction to be: Jd-rich omph+qtz+ky=olig+Jd-poor omph+grt. Fe–Mg exchange thermometry yields similar results during the prograde breakdown of plagioclase and the retrograde regeneration at around 740 °C (Sanders 1988). The peak pressure conditions of the prograde reaction are >16 kbar, whereas the retrograde reaction is estimated as 14–15 kbar (Sanders 1988). An estimate of the P-T conditions prior to the prograde reaction and shearing is provided by measurements on omphacite and garnet cores, which yield a temperature of around 700 °C, with an assumed pressure of around 13–14 kbar, based on XJd of *c*. 0.2 and the inferred presence of plagioclase (Sanders 1988). In this case, both the prograde and retrograde reactions occurred pseudo-isothermally within the eclogite facies stability field and define what Sanders (1988) referred to as a minor (pressure) excursion. Felsic gneisses (trondhjemites) that have a Late Archaean protolith age (Storey 2002) also show evidence of having been metamorphosed to eclogite facies conditions. Sanders (1979) first noted the presence of felsic eclogite facies gneisses interlayered with eclogite within the EU. The gneisses have a granoblastic polygonal texture and contain omphacite, garnet, oligoclase and quartz, with local kyanite, K-feldspar, rutile and biotite. These are mapped as garnet omphacite kyanite gneiss on Fig. 3. Storey *et al.* (2005) estimated the peak P-T conditions of these rocks as 20 ± 3 kbar and 750–780 °C, based on the garnet anorthite diopside silica (GADS) barometer (pyr+2grs+3qtz=3ano+3diop) of Powell & Holland (1988) and the Powell (1985) calibration of the Fe–Mg grt–cpx exchange thermometer.

Rawson *et al.* (2001) first reported ultrabasic rocks within the EU that contained evidence of high-pressure metamorphism. They discovered limited, isolated exposures of garnet-bearing olivine websterite and websterite. The field relations, and particularly the lack of contacts with surrounding eclogites and gneisses, do not allow derivation of their tectonic evolution (i.e. were they tectonically emplaced 'alpine-type' or cumulates of the basaltic magma that was protolith to the EU eclogites?). On the basis of mineralogy, texture and composition, Rawson *et al.* (2001) interpreted the websterites as more likely derived from banded ultrabasic-basic granulite facies gneisses equivalent to the Scourian gneisses of the Lewisian Complex and therefore as metamorphosed igneous intrusions rather than fragments of mantle. There are therefore, and unusually for a high-pressure terrane, no unequivocal examples of lithospheric mantle present within the GAI. The garnet-bearing olivine websterite contains two pyroxenes, olivine, garnet, amphibole and minor magnetite and spinel. Websterites have similar mineralogy but lack garnet and, generally, the olivine has been replaced by serpentine and the pyroxenes by amphibole as a retrograde reaction. Rawson *et al.* (2001) derived a P-T estimate of 20 ± 3 kbar and 730 ± 50 °C from the garnet-bearing olivine websterite, in agreement with the estimate from the felsic eclogite facies gneisses by Storey *et al.* (2005).

Kyanite–garnet–biotite gneisses are common in the EU within pelitic units mapped on Figs. 2 and 3 and commonly contain distinctive mauve-coloured garnet up to 10 mm in diameter within a foliation comprising biotite, kyanite, phengitic white mica, plagioclase, orthoclase and quartz. Rawson (2003) studied the composition of the major minerals by electron microprobe and discovered that the white mica is not phengite *sensu stricto*, but is closer to the phengite end-member than to muscovite (>3.2–3.6 Si atoms per formula unit). The implication of this is that the phengite is a high-pressure phase associated with eclogite facies metamorphism. However, in applying the phengite barometer of Massone & Schreyer (1987) the pressure estimates varied wildly from *c*. 7 to 17 kbar. Clearly the compositions must have been altered during retrogression if the higher pressure estimates really are a relict from high-pressure metamorphism. The kyanite and mica present indicate that these rocks were originally aluminous pelites that in all probability were cofacial with the other eclogite facies rocks in the EU.

Amphibolite facies metamorphism within retrogressed eclogites has a P-T estimate of *c*. 13 kbar and 650–700 °C (Storey *et al.* 2005) based on the GAPQ (garnet–amphibole–plagioclase–quartz) barometer of Kohn &

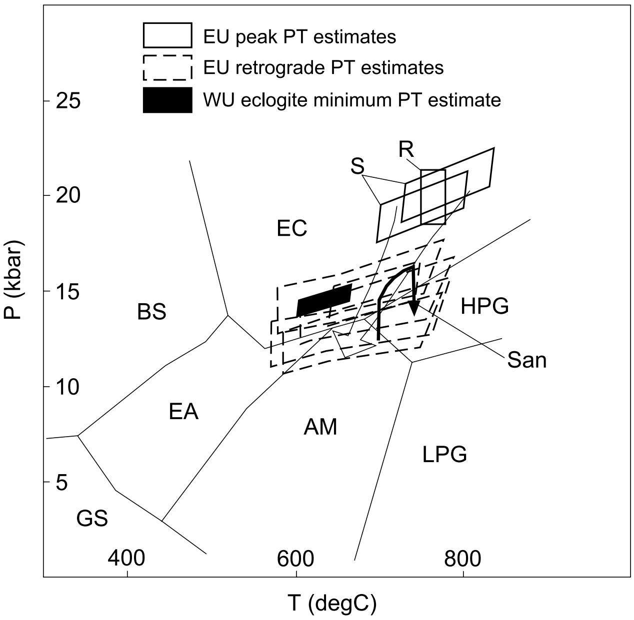

FIG. 4. P-T estimates of eclogite facies rocks from the EU; peak and amphibolite facies retrograde estimates shown. Temperature with minimum P estimate shown also for WU eclogite. AM, amphibolite facies; BS, blueschist facies; EA, epidote–amphibolite facies; EC, eclogite facies; GS, greenschist facies; HPG, high-pressure granulite facies; LPG, low-pressure granulite facies. R, P-T estimate from EU garnet websterite of Rawson *et al.* (2001); S, peak P-T conditions of EU eclogite facies felsic gneisses of Storey *et al.* (2005); San, syn-eclogite facies P-T loop of EU streaky eclogites of Sanders (1988). EU Amphibolite facies retrograde P-T estimates are from Storey *et al.* (2004). WU eclogite facies PT estimates are from Storey (2002).

Spear (1990), the amphibole–plagioclase thermometer of Holland & Blundy (1994) and the amphibole–garnet Fe–Mg exchange thermometer of Graham & Powell (1984).

P-T conditions, along with possible P-T paths for the EU during Grenvillian orogenesis, are presented in Fig. 4. This demonstrates that maximum conditions of around 20 kbar (around 70 km burial depth) and 750–780 °C were attained at close to 1080 Ma ago, followed by decompression to around 13 kbar and 650–700 °C by *c.* 1000 Ma.

Western Unit

Amphibolites and felsic gneisses within the WU have not been studied for P-T conditions, generally due to the lack of useful phases with which to calculate phase equilibria. Felsic gneisses generally comprise feldspar, quartz, amphibole, biotite and zircon. Amphibolites contain amphibole, feldspar, minor quartz, local retrograde biotite and relict garnet.

The occurrence of eclogite in a limited area close to Sandaig in the WU (Fig. 2) was first reported by Sanders (1972) at NG 767 140. The eclogite has a granoblastic polygonal texture and contains garnet, omphacite, rutile/ilmenite and quartz. Retrogression is common, forming transgressive veins in which all omphacite has been converted to amphibole. At the margins of the eclogite pods, where they contact with surrounding felsic gneisses, amphibolitization is also complete. Similar veins also occur within fairly pristine eclogite, but in these omphacite commonly has a thin rim of amphibole and retrogression is only partial. An earlier phase of retrogression is recognized where omphacite is partially replaced by symplectites of diopside, plagioclase and quartz, imparting a ‘fingerprint texture’ to the relict omphacite grains that are also commonly rimmed by a thin overgrowth of amphibole. It is difficult to obtain P-T estimates from these rocks due to the lack of useful phases, but minimum pressures obtained from the Jd component within the omphacite returns pressures of 13–14 kbar at around 650 °C (Storey 2002), which must be considered a minimum estimate. An attempt to estimate conditions of the initial symplectite-forming retrogression by use of the GADS barometer on the symplectite phases assumed to be in equilibrium with garnet rim compositions returned an estimate of *c.* 14–16 kbar at *c.* 650 °C (Storey 2002). By inference, the eclogite peak pressure must be higher, so a minimum of 14–16 kbar is suggested as the best available estimate. Temperature estimates, based on co-existing omphacite and garnet, using the Fe–Mg exchange thermometer, return 600–700 °C from the WU eclogite. The P-T estimate is shown in Fig. 4 for comparison.

Granulite facies assemblages in mafic protoliths occur within restricted areas in the NE of the WU outcrop, although nearly all metabasites within the area demonstrate evidence of having been previously metamorphosed to granulite facies. A pod of mafic granulite near Eilean Donan Castle (Fig. 2) contains an assemblage including clinopyroxene (diopside), garnet, plagioclase and quartz with a granoblastic polygonal texture, along with minor rutile/ilmenite, and is thus a high-pressure granulite (Ringwood 1975). Minor orthopyroxene and primary brown amphibole are also locally present within the WU mafic granulites in equilibrium with diopside, garnet and plagioclase. The mafic granulites are typically associated with partial melt (leucosome) patches and veins, and it is considered that much of the trondhjemitic melt forming veins (agmatite) ubiquitously within WU amphibolites formed by partial melting during the granulite facies event. Retrogression is well developed in the granulites and the minor patches of preserved pristine granulite are difficult to find. Retrogression occurs as thin symplectitic rims of blue-green amphibole (pargasite) and plagioclase along diopside–garnet–plagioclase contacts. In more progressively retrogressed areas the diopside has been partially to completely replaced by pargasite and the rutile/ilmenite intergrowths are rimmed by framboidal titanite. In addition, the plagioclase grains are ubiquitously clouded by fine-grained, randomly orientated laths of zoisite. Rawson (2003) considered that this may have been the result of partial reaction during later eclogite facies metamorphism. Whilst this idea is appealing, it is by no means proven and there remains no unequivocal petrologic evidence that the WU granulites have subsequently been metamorphosed to eclogite facies.

Geochronology

Eastern Unit

Trondhjemitic gneisses of the EU are of Late Archaean age (Storey 2002), whereas zircons from the eclogites mostly have Hf_{TDM} ages of around 2.0 Ga (Brewer *et al.* 2003), suggesting that the basic protoliths may have been younger than the gneisses and possibly synchronous with deposition of the sediments prior to eclogite facies metamorphism. However, there is some evidence that the gneisses underwent a Mid-Proterozoic metamorphic event at around 1.75 Ga (Storey 2002; R.A. Strachan, pers. comm.).

Determination of the age of eclogite facies metamorphism was first attempted by Miller *et al.* (1963), who produced K–Ar ages from omphacite of 1515 ± 104 Ma. Precise Sm–Nd garnet–omphacite–whole rock isochron ages of 1082 ± 24 and 1010 ± 13 Ma were produced by Sanders *et al.* (1984), who interpreted the older of the two as being close to the peak of eclogite facies metamorphism and the latter as the date of retrogression. This implies that the original K–Ar date suffered from problems with excess Ar. The age of retrogression to amphibolite facies was confirmed by Brewer *et al.* (2003), who dated retrograde zircon, growing within the amphibolite facies paragenesis, by the U–Pb method and derived an age of 995 ± 8 Ma. The geochronology of the eclogites demonstrates that they were formed and retrogressed during the Grenville orogenic cycle and this implies that part of the British Isles was inboard of the Grenville Front during the assembly of Rodinia.

Western Unit

Felsic gneisses of the WU are of Late Archaean age (Storey 2002) and superficially could be correlated with the EU based on this evidence alone. The granulite facies metamorphism has proved problematic in terms of timing, with the assumption that it was coeval with the eclogite (e.g. Barber & May 1976). However, recent U–Pb zircon geochronology has indicated a Late Archaean (*c.* 2700 Ma) age for granulite facies metamorphism (Storey 2002).

It has commonly been assumed, with no corroborating geochronological evidence, that the WU eclogite is coeval with eclogites in the EU. However, new evidence has suggested a more complex high-pressure evolution for the GAI. Zircon U–Pb data from the eclogite suggests an age of *c.* 1700–1750 Ma (Storey 2002), whilst garnet–omphacite Lu–Hf data indicate an age of *c.* 1680 Ma (Storey *et al.* unpublished data). These suggest that the eclogites of the WU and EU cannot be correlated and imply a separate high-pressure metamorphism for the WU only during the Palaeoproterozoic.

More recent Sm–Nd and Lu–Hf garnet–clinopyroxene dating of the mafic granulite indicates that the system was strongly disturbed during *c.* 1700–1750 Ma reworking at eclogite facies. The isotopes perhaps provide the strongest evidence that the granulites were later metamorphosed to eclogite facies.

Dating of structures

The *c.* 995 Ma age of retrograde zircon growth in the EU eclogite is considered to be linked, on petrographic grounds, to the uplift of the eclogites into the mid-crust during D_2 (Brewer *et al.* 2003; Storey *et al.* 2004). However, the D_2 Barnhill Shear Zone (BSZ) contains synkinematic Morar Group psammites that have a maximum depositional age of *c.* 980 Ma (Peters 2001). The BSZ contains synkinematic titanite, which has been dated by U–Pb methods to 669 ± 31 Ma (Storey *et al.* 2004). The data could be interpreted as a cooling age due to the small grain size of the titanite and so is considered a minimum age for movement along this shear zone. Zircon U–Pb dating of a granite sheet that cuts across S_2 has a lower intercept age of *c.* 670 Ma (Storey *et al.* 2004) and so confirms that D_2 must be at least this old and not Lower Palaeozoic in age as previously suggested (Barber & May 1976).

Storey *et al.* (2004) attempted to date D_3 using minor granitic veins that cross-cut S_2 but were deformed by F_3. In one case, titanite within a pegmatite folded by F_3 has a U–Pb titanite age of 437 ± 6 Ma and indicates that D_3 is younger than this.

Structural evolution of the GAI

A detailed map, with structural measurements, is presented in Fig. 3.

Four major phases of deformation can be recognized within the GAI: D_2 and D_3 are associated with major folding episodes and shearing, whilst D_1 and D_4 are associated with the development of shear zones. A major current controversy is the relative contribution of shearing versus folding during D_2 and D_3.

Ramsay & Spring (1962) suggested that folding during compression could account for all of the structural relationships in the area relating to D_2 and D_3 (their D_1 to D_3). By contrast Sutton & Watson (1959); Barber & May (1976) contended that the earliest recognized phase of deformation (D_2 in this study, D_1 *sensu* Ramsay 1958), that involved interleaving of the Moine and GAI basement, was achieved by compressional shearing (i.e. thrusting). The structural models of Ramsay (1958) and Sutton & Watson (1959) did not account for the exhumation of the eclogites in the EU and, at the time, their age and P-T conditions were not understood. Barber & May (1976) produced a structural model involving overthrusting of the EU onto the WU (top-to-the-west) that they interpreted as having occurred during the Lower Palaeozoic Caledonian orogeny. However, this failed to take into account the presence of eclogite in the WU which implies that there may be no regional deep-over-shallow relationship. Ramsay (1958) considered that, following interleaving, the later events were purely related to folding episodes in a compressional environment.

Temperley & Windley (1997) first mooted the idea that the earliest phase of deformation common to the GAI and Moine was extensional, with the Moine juxtaposed against the GAI in the hanging wall of a top-to-the-east extensional detachment as the eclogite facies basement was decompressed from great depth in the footwall. This was accompanied by folding, verging and facing down towards the east. Temperley & Windley (1997) present kinematic data supporting this hypothesis, demonstrating the non-coaxial nature of the early deformation, and argue that their model is the only one that can account for the decompression of eclogites towards the surface, noting that earlier models had failed to take into account the significant metamorphic break between the GAI and the Moine. However, there are problems with the model as it fails to take into account strips of Moine psammites that occur dominantly between the EU and the WU, but also within the WU. It also fails to acknowledge compressional (top-to-the-west) kinematic features within the EU and the Moine strip (Storey *et al.* 2004) that are apparently at the same grade as the early extensional deformation. Indeed, the recent knowledge that the Morar Group was deposited after *c.* 980 Ma (Peters 2001), younger than the *c.* 995 Ma amphibolite facies retrogression (Brewer *et al.* 2003), further rules out this hypothesis.

Storey *et al.* (2004) presented evidence that the boundary between the EU and WU (the BSZ; Fig. 2) was a major ductile shear zone with a reverse sense of motion (top-to-the-west) that can be linked, on petrographic grounds, to the decompression of eclogites in the EU. The *c.* 669 Ma minimum age of shearing in the BSZ confirms that the event was of Neoproterozoic age and not Lower Palaeozoic, as previously assumed (Barber & May 1976).

A summary of previous work and structural interpretations is presented in Table 1.

D_1 structures

The earliest recognized deformation fabric within the WU comprises a gneissic banding cut by speckled amphibolite sheets that have undergone later deformation and shearing. The banding is irregular and cannot be used to infer any larger-scale structure in the WU during the Late Archaean. The earliest structures recognized within the EU are those formed at peak eclogite facies conditions. This fabric does not occur in the adjacent WU. Hence, the earliest fabric in the WU is termed D_{1W} and that in the EU D_{1E}. Occasional L-dominated fabrics are present in eclogites, defined by a shape alignment of omphacite and elongate garnet. A dominant area of D_1 deformation was first described by Sanders (1988). A ridge of streaky eclogite, some 3–4 km long, occurs north of Glen More, striking NNE along a prominent topographic feature (Fig. 2). The eclogite comprises a strong L_1 rodding and elongation lineation formed by quartzo-feldspathic streaks, containing needles of kyanite and aligned omphacite grains. The plunge of the lineation lies between 10° and 40° clockwise of the dominant SE-dipping compositional layering along the length of this ridge as a result of F_3 refolding.

D_2 structures

These structures are the first that are common to the EU, WU and Moine, and occurred at upper amphibolite facies conditions, as the EU was decompressed into the mid-crust following deep burial. D_2 comprises a pervasive, generally eastward-dipping, composite S_2 foliation that has an associated generally eastward (down-dip) plunging L_2 mineral elongation lineation (where not refolded). A plethora of minor F_2 folds are associated with this episode, as well as large-scale F_2 folds, the Beinn a'Chapuill–Beinn nan Caorach antiform and the Glen Beag synform (see Ramsay 1958). The axes of the F_2 folds are commonly sub-parallel to the L_2 stretching lineation but also markedly curvilinear. Hence, D_2 comprised penetrative non-coaxial ductile shearing with accompanying folding. Within the EU the L_2 mineral stretching lineation is almost ubiquitous and most of the basement is highly strained. Low strain windows do occur, particularly within rheologically stronger eclogite layers and boudins. In particular, it is possible to trace the transformation from pristine eclogite within boudins towards the edges, where the assemblage passes towards amphibolite and a strong L-S tectonic fabric occurs. The curvilinear minor F_2 fold hinges, and their rotation into parallelism, or near-parallelism, with the elongation direction, also argue for non-coaxial shear-related folding. Strain intensifies towards the edges of the EU outcrop where it is in contact with the Moine to the east and the Moine strip / WU to the west, and within 1 km of these contacts the fabric is mylonitic to ultramylonitic and, hence, the contacts are intense shear zones (Fig. 2). Storey *et al.* (2004) referred to the shear zone between the EU and the WU as the Barnhill Shear Zone, and that to the east, between the EU and Moine, as the Inverinate Shear Zone (ISZ). Storey *et al.* (2004) noted that the BSZ appears to be dominated by compressional top-to-the-west kinematic indicators, whilst the ISZ is dominated by apparently extensional top-to-the-east features. These have the overall effect of describing the EU as a wedge bounded at its lower surface by a thrust and at its upper surface by an extensional shear, and may account for the emplacement of the EU into the overlying Moine as it was decompressed from the deep crust. Direct dating of synkinematic titanite in EU mylonite within the BSZ controversially suggests a minimum age of shearing of 669 ± 31 Ma (Storey *et al.* 2004). The discovery within the Morar Group of detrital zircons of *c.* 980 Ma age (Peters 2001) implies that D_2 actually comprises two distinct events, since the earlier deformation at *c.* 995 Ma cannot, therefore, have affected the Moine; the later shearing event, producing synkinematic titanite at a minimum of *c.* 669 Ma, must have affected both the Moine and EU and may have been responsible for their first juxtaposition. There remains the possibility that the GAI was exhumed to the surface and the Moine deposited unconformably upon it at some time prior to *c.* 669 Ma. However, it should be emphasized that everywhere where the EU shares contact with the Moine there is always an intense shear zone present. Moreover, there is no petrological evidence of a

TABLE 1
Comparison of interpretation of structural evolution of GAI for various authors.

Ramsay (1958)	Sutton & Watson (1959)	Ramsay & Spring (1962)	Barber & May (1976)	May *et al.* (1993)	Temperley & Windley (1997)	Storey *et al.* (2004) and herein
Unconformable deposition of Moine on exposed basement	Unconformable deposition of Moine on exposed basement	Unconformable deposition of Moine on exposed basement	Unconformable deposition of Moine on exposed basement	Unconformable deposition of Moine on exposed basement	Deposition of Moine in rift basin allochthonous to GAI. >0.84 Ga	D_{1W} Early gneissic banding in the WU. Archaean. D_{1E} Eclogite facies ductile shearing top-to-the-west 1.1–1.0 Ga
Interleaving of Moine and basement via ductile shearing	Interleaving of Moine and basement via ductile shearing	Interleaving of Moine and basement via isoclinal folding	D_1 interleaving of Moine and basement. Late Precambrian	D_1M/D_1L interbanding of Moine and basement. ?Late Precambrian. D_2L interbanding of Moine and basement. ?Late Precambrian	Top-to-the-east high grade extensional shearing and related folding within the GAI. 1.08–0.84 Ga	D_2 decompression of eclogites into mid-crust. 1.0 Ga. Deposition of Moine (on exposed GAI?). 0.98–0.87 Ga. D_2 interleaving of Moine and GAI by isoclinal folding and compressional shearing. Concomitant extensional shearing accommodating extrusion of EU as a wedge. Progressive development of F_2 folds facing down to the east. 1.0–0.67 Ga
First phase of folding facing down to the east	First phase of folding facing down to the east accompanied by ductile shearing	Second phase folding facing down to the east	D_2 NE-plunging folds and rodding. Late Precambrian			
Second-phase folding, verging and facing down to the SE	Second-phase folding, verging and facing down to the SE accompanied by ductile shearing	Third-phase folding verging and facing down to the SE	D_3 SE-plunging folds and rodding at amphibolite facies. Late Precambrian	D_3L SE-plunging folds and rodding at amphibolite facies. ?Late Precambrian	Top-to-the-SE ductile extensional shearing and related large-scale folding juxtaposing Moine and GAI. Neoproterozoic	D_3 SE-plunging folds and lineation, large-scale folding verging and facing down to the SE; ductile top-to-the-SE shearing at amphibolite facies. 437–425 Ma
	Development of Moine Thrust		D_4 mylonitization and ESE plunging lineation. Post Llanvirn	D_4L/D_2M mylonitization, ESE to SE plunging lineation and rodding. Ductile thrusting. Caledonian	Low temperature shearing in Moine Thrust nappe translates GAI and Moine onto Caledonian foreland. Caledonian	D_4 greenschist facies brittle–ductile top-to-the-NW shearing translates GAI and Moine onto Caledonian foreland along Moine Thrust. 437–425 Ma

second, prograde amphibolite facies metamorphic event within the EU and this would necessarily be present within this model. It seems, on balance, highly likely that the Moine is allochthonous with respect to the EU, but autochthonous with respect to the WU.

D_3 structures

These structures are clearly lower grade and less ductile than D_2: in minor S_3 shears amphibole is replaced by biotite and the grade is that of lower amphibolite facies. This phase of deformation was responsible for large-scale refolding of F_2 folds and resulted in NE–SW trending major fold axes throughout the area, the Loch Hourn–Loch Duich antiform and the Ben Sgriol synform, described by Ramsay (1958). Temperley & Windley (1997) also provided strong evidence for SE-directed extensional shearing during D_3 in the form of a pronounced L_3 mineral stretching lineation, asymmetric boudins and shear bands cutting through D_2 structures, and F_3 folds with curvilinear hinges in places approximating parallelism with L_3. These structures are commonly observed along the exposed road section (A87) on the north side of Loch Duich. L_3 plunges moderately towards the SE throughout the area and shear bands, in some places very pronounced, tend to have fairly steep attitudes with clear displacements down-dip towards the SE. The pegmatite dated to *c.* 437 Ma that cross-cuts S_2 but is folded by F_3 indicates that D_3 is younger than this.

D_4 structures

The latest major phase of deformation was brittle–ductile and formed at greenschist facies conditions. Local minor shear zones occur throughout the area and these are accompanied by a new L_4 mineral stretching lineation, generally formed by chlorite, plunging gently towards the SE and cutting across all earlier structures. These zones can be easily related to movements associated with the Moine Thrust Zone. Other minor deformation is in the form of kink bands and minor box-folding. Movements related to the Moine Thrust are believed to have an age of *c.* 435–430 Ma (Halliday *et al.* 1987; Johnson *et al.* 1985; Kelley 1988; Freeman *et al.* 1998) and thus bracket the age of D_3 fairly tightly.

Discussion and conclusions

To aid the discussion, a list of dates of events is given in Table 2.

The GAI represents the largest tract of basement exposed within the Neoproterozoic Moine Supergroup. The WU and EU are superficially similar Late Archaean gneiss complexes, but there are important lithological differences between the two. The WU comprises felsic gneisses of tonalite–trondhjemite–granodiorite association along with subordinate metabasites (amphibolites, minor granulites and eclogites) and minor ultrabasites. The EU, in contrast, contains up to 25% eclogite, aluminous, calcareous and manganiferous metapelites, marbles, ultrabasites and eulysites hosted within Late Archaean felsic, trondhjemitic gneisses. The EU and WU are separated by the BSZ, a major ductile shear zone with a dominant top-to-the-west sense of movement. The granulites in the WU were formed in the Late Archaean (Storey 2002), whereas eclogite facies reworking most likely occurred at around 1700–1750 Ma; the only discernible affect of this event, apart from formation of minor amounts of eclogite, is major loss of Pb from zircons in WU gneisses and granulites (Storey 2002). There is very little evidence of the WU having been metamorphosed to eclogite facies during the Grenvillian orogeny and, indeed, very little evidence of it having been metamorphosed to any strong degree at all at that time. In stark contrast, the EU eclogites are clearly Grenvillian in age and many of the other EU lithologies are cofacial and, by extension, coeval. Thus, the BSZ may represent a major tectonic discontinuity that represents the Grenville Front in Scotland, with Grenvillian high-pressure crust (the EU) being squeezed westward towards the foreland as it was decompressed into the mid-crust at around 995 Ma, before final juxtaposition against the WU and Moine, potentially at or before around 670 Ma via ductile shearing and concomitant folding.

The Morar Group has a maximum age of 980 ± 2 Ma (Peters 2001). At higher stratigraphic and structural levels within the Moine Supergroup, the Glenfinnan and Loch Eil Groups contain evidence for having been deposited at a younger maximum age of *c.* 950 Ma and 900 Ma, respectively (Cawood *et al.* 2004; Emery 2005), and have been intruded by the West Highland Granite Gneiss at *c.* 870 Ma (Friend *et al.* 1997; Rogers *et al.* 2001), thus constraining the deposition to the period *c.* 900–870 Ma. However, the Glenfinnan Group is separated from the Morar Group by a ductile thrust, the Sgurr Beag Thrust, and so stratigraphic continuity may not be guaranteed and a maximum age of *c.* 980 Ma for the Morar Group is considered more robust. This datum suggests that the juxtaposition of the Moine and the GAI cannot have occurred during the initial decompression of the GAI into the amphibolite facies mid-crust during D_2 at *c.* 995 Ma (Brewer *et al.* 2003). Therefore, it further suggests that D_2 deformation along the BSZ (minimum age *c.* 669 Ma) must have been formed by two separate events: one prior to Moine deposition and one after. The latter was responsible for the juxtaposition of the Moine and GAI. This also permits the possibility that the Moine and GAI have an unconformable relationship, although the field relations suggest that any unconformity was more likely only between the WU and the Moine. This, in turn, further adds to the evidence for Neoproterozoic compressional deformation and prograde metamorphism affecting the Moine (Morar Group), with events recorded at *c.* 820–790 Ma (peak P-T: 687–707 °C and 12.5–14.5 kbar; Vance *et al.* 1998) and *c.* 737 Ma (greenschist to amphibolite facies; Tanner & Evans 2003).

TABLE 2
Chronology of events affecting the GAI and Moine as discussed in the text.

Age (Ma)	Error (Ma)	WU	BSZ	EU	Moine	Ref.
>2700		TTG protoliths and mafic dykes		TTG protoliths		
c. 2700		Granulite facies metamorphism and partial melting				
1750–1700		Eclogite facies metamorphism and major reworking		?Metamorphic event		
≥1082	24			Eclogite facies peak metamorphism		Sanders *et al*. (1984)
1010	13			Amphibolite facies retrogression		Sanders *et al*. (1984)
995	8			Amphibolite facies retrogression new zircon growth		Brewer *et al*. (2003)
980	2				Maximum age of Morar Group	Peters (2001)
c. 950					Maximum age Glenfinnan Group	Friend *et al*. (2003)
c. 900					Maximum age Loch Eil Group	Cawood *et al*. (2004); Emery (2005)
c. 870					West Highland Granite Gneiss	Friend *et al*. (1997)
840	11				Anatexis in Central Highland Migmatite Complex	Highton *et al*. (1999)
804	13				Shearing in the Grampian Shear Zone	Noble *et al*. (1996)
820–790					Prograde amphibolite facies metamorphism in the Morar Group at 12.5–14.5kbar and 687–707°C	Vance *et al*. (1998)
737	5				Prograde amphibolite facies metamorphism in the Moine	Tanner & Evans (2003)
669	31		Amphibolite facies top-to-the-west compressional shearing			Storey *et al*. (2004)
<437	6			D_3 folding	D_3 folding	Storey *et al*. (2004)

A major episode of refolding, resulting in the characteristic interference fold patterns that mark the western boundary of the GAI (Figs. 1 and 2), first described by Ramsay (1958, 1963), was shown to be Lower Palaeozoic in age (*c.* 437 Ma; Storey *et al.* 2004) and, therefore, attests to the major Caledonian orogenic event that affected the Scottish Highlands and culminated in the formation of the Moine Thrust at *c.* 435–430 Ma.

The juxtaposition of disparate diachronous high-pressure terranes may seem coincidental, particularly as no other unequivocal crustal eclogites occur within the British Isles. However, it may reflect a more fundamental property of orogenic belts in general. Eclogites within the eastern Alps, for example, are diachronous: earlier Variscan MORB eclogites, formed from *c.* 347 to 336 Ma (Thöni 2006), are considered to represent north Apulian (Austroalpine) crust, whereas in the structurally lower Penninic unit, exposed within the Tauern Window, eclogites were formed during the Alpine orogeny (<45–31 Ma; Thöni 2006). Potentially, the Wilson Cycle can account for cyclical formation of thickened orogenic belts that produce eclogites during collision of continents that then rift before colliding again, producing younger eclogites. The eclogitic units may be juxtaposed by shearing during tectonism related to orogenic re-organization, as in the case of the eastern Alps and, potentially, within the GAI. Shearing and folding, products of contractional deformation, at or before *c.* 670 Ma suggests that plate reorganization during the break-up of Rodinia was active and diachronous within the Scottish portion of Laurentia.

Previous ages of prograde amphibolite facies metamorphism within the Moine of 820–790 Ma (peak P-T: 687–707 °C and 12.5–14.5 kbar; Vance *et al.* 1998) and greenschist to amphibolite facies metamorphism of *c.* 737 Ma (Tanner & Evans 2003), suggest that compressional events occurred in short pulses within the Scottish Highlands. Events of similar age occur in the Caledonides of Scandinavia, the *c.* 840 Ma Porsanger orogeny and tectono-thermal events at *c.* 710 Ma, 670 Ma (Kirland et al. 2006, 2007) and *c.* 637 Ma (Rehnstrom *et al.* 2002). It is suggested that the Moine may be correlated with the Sørøy Succession of the Kalak Nappe Complex of Arctic Norway (Kirkland *et al.* 2007) and that this rifted from Laurentia–Baltica following break-up of Rodinia to form an exotic mobile belt, which was subsequently sutured to the margin of Baltica (Kirkland *et al.* 2007). This can account for the tectono-metamorphic events affecting the Moine during the Neoproterozoic, as there would have been sufficient free ocean to allow arc mechanisms and accretion to operate and, hence, short-lived contractional deformation and metamorphism (Kirkland *et al.* 2007). However, an alternative hypothesis suggests an intracontinental setting, with contractional deformation driven by far-field effects around the margins of Rodinia (Cawood *et al.* 2007). By extension, the 820–790 Ma event recorded in the Morar Group, was clearly significant, resulting in medium- to high-pressure metamorphism (Vance *et al.* 1998) that may be more compatible with accretion along a continental margin than with intracontinental deformation. Relicts of eclogites within the sub-Grampian Dava Succession of the Scottish Central Highland Migmatite Complex (Baker 1986) that may correlate with parts of the Moine Supergroup, are intriguing and, as yet, undated. The age of migmatization of *c.* 840 Ma (Highton *et al.* 1999) and subsequent shearing at *c.* 800 Ma (Noble *et al.* 1996), at kyanite-grade metamorphic conditions (Phillips *et al.* 1999), places a large question mark over whether these eclogite relicts too are part of the Neoproterozoic tectono-metamorphic evolution of the Highlands. If that were true, it would be difficult to reconcile their formation within an intracontinental setting.

Major folding and associated extensional shearing of the GAI occurred during the Caledonian orogeny, sometime after *c.* 437 Ma, resulting in large-scale interference folds. The GAI and overlying Moine were transported westward in the hanging wall of the Moine Thrust Zone onto the Caledonian Foreland at around 435–430 Ma during the terminal Scandian orogeny.

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