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Caledonian and Knoydartian overprinting of a Grenvillian inlier and the enclosing Morar Group rocks: structural evolution of the Precambrian Proto-Moine Nappe, Glenelg, NW Scotland

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Abstract: The Grenville and Caledonian orogens, fundamental to building Laurentia and Baltica, intersect in northern Scotland. The Precambrian Glenelg Inlier, within the Scottish Caledonides, preserves a record of Grenvillian, Knoydartian and Caledonian orogenesis. Based on new mapping and re-interpretation of previous mapping, we present a structural model for the evolution of the Glenelg Inlier. The inlier can be divided into Western Glenelg gneiss comprising orthogneiss with no record of Grenville-age metamorphism, and Eastern Glenelg gneiss with ortho- and paragneiss, affected by Grenvillian eclogite-facies metamorphism. The basement gneisses and their original cover of psammitic, Neoproterozoic Morar Group (Moine) rocks were deformed by three generations of major ductile folds (F_1–F_3). In medium-strain areas F_2 and F_3 folds are broadly coaxial and both face to the west; in higher strain areas F_2 and F_3 folds are oblique to each other. By restoring post-F_1 folds and late faults, the Glenelg gneiss inliers are seen to form the core of a major recumbent SSE-facing F_1 isoclinal fold nappe – the Proto-Moine Nappe. The upper limb of this nappe is a thick, right-way-up sequence of moderately strained Morar Group rocks whereas the lower, inverted limb comprises intensely deformed, migmatitic Morar Group rocks. Within the constraints of published geochronology, the Proto-Moine Nappe is likely Pre-Caledonian and may have originated during the early Neoproterozoic Knoydartian Orogeny.

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The Grenville–Sveconorwegian and Caledonian–Appalachian orogens represent pivotal tectonic events in the evolution of the continental masses of Laurentia and Baltica (Fig. 1). The *c.* 1200 – 960 Ma Grenville–Sveconorwegian Orogen stretches from Mexico to Southern Sweden, and is one of the largest collisional belts on Earth (e.g. Rivers 1997, 2008; Tollo *et al.* 2004; Tohver *et al.* 2004). The Archaean–Palaeoproterozoic Glenelg Inlier in Scotland and the Annagh Gneiss Complex in Ireland provide small but crucial links between the North American and Scandinavian parts of the orogen (Sanders *et al.* 1984; Daly 1996; Brewer *et al.* 2003; Storey *et al.* 2010). The Glenelg Inlier of Scotland, however, occurs within the much later 470 – 400 Ma Caledonian Orogen (e.g. McKerrow *et al.* 2000; Woodcock & Strachan 2012; Chew & Strachan 2014; Dewey *et al.* 2015) and has thus been strongly reworked.

The Glenelg Inlier in the Northern Highlands of Scotland (Fig. 2) preserves Grenvillian-age eclogite-facies assemblages, but is surrounded by metasedimentary rocks of the early Neoproterozoic Morar Group (Fig. 2), the lowest tectono-stratigraphic unit of the Moine Supergroup (Holdsworth *et al.* 1994). Detrital zircon dating has shown that Morar Group deposition was later than, or possibly coeval with, the Grenville-age high-pressure metamorphism in the Glenelg Inlier (Brewer *et al.* 2003; Friend *et al.* 2003; Kirkland *et al.* 2008; Cawood *et al.* 2015). The Morar Group, together with the Sleat and Torridon groups in the Caledonian foreland, is now interpreted as a remnant of the foreland basin to the Grenville Orogen (Rainbird *et al.* 2001; Krabbendam *et al.* 2008, 2017; Bonsor *et al.* 2010, 2012; Krabbendam & Rainbird 2012).

The post-Grenvillian evolution of the Glenelg Inlier and surrounding Morar Group rocks is complicated and involved several episodes of tectonic reworking, including one or more phases of Knoydartian (Neoproterozoic, *c.* 825 – 720 Ma) activity, as well as Grampian (Ordovician) and Scandian (Silurian) deformation during Caledonian orogenesis (e.g. Van Breemen *et al.* 1974; Rogers *et al.* 1998; Vance *et al.* 1998; Kinny *et al.* 1999; Cutts *et al.* 2009, 2010; Bird *et al.* 2013; Cawood *et al.* 2015).

The Glenelg Inlier has been well studied in terms of its structure, petrology and geochronology (Alderman 1936; Ramsay 1957, 1963; Tanner 1965; Sanders *et al.* 1984; Sanders 1988, 1989; Rawson *et al.* 2001; Storey *et al.* 2004, 2005; Storey 2008*a*, *b*; Barber 2011). A number of structural studies have been carried out in the surrounding areas (Clifford 1957; Ramsay & Spring 1962; Tanner 1965; Powell 1966, 1974; Simony 1973; Poole & Spring 1974). This region has served as a natural laboratory for structural geology; in particular for the geometrical analysis of large- and small-scale fold interference patterns and folded lineations (Ramsay 1957, 1960, 1962, 1967). Nevertheless,

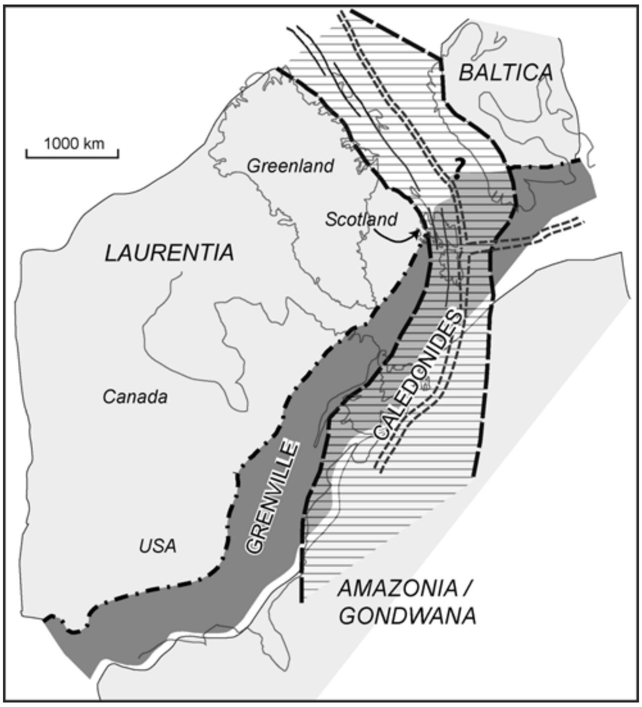

Fig. 1. Schematic reconstruction of the Grenville–Sveconorwegian Orogen (1200 – 960 Ma) with the trace of the Caledonian Orogen (470 – 400 Ma) superimposed (striped). Trace of the Iapetus Ocean indicated by double dashed line. Note that different parts of Amazonia/Gondwana collided with Laurentia during the Grenville and Caledonian orogens, respectively, and that the relative position of Baltica and Laurentia is that during the Paleozoic. Modified after Tohver *et al.* (2004), Cawood *et al.* (2007) and Bingen *et al.* (2008).

it remains uncertain which structures (folds, fabrics, thrusts) can be attributed to which orogenic event (e.g. Knoydartian, Grampian, Scandian) and where the limits or 'orogenic fronts' of these different orogenies were situated (e.g. Bird *et al.* 2013; Cawood *et al.* 2015; Dewey *et al.* 2015). Consequently, the structural setting and tectonic evolution of the Glenelg Inlier within the larger framework of the Scottish Caledonides has remained elusive (cf. Sanders *et al.* 1997; Temperley & Windley 1997; Storey 2008*a*; Storey *et al.* 2010; Barber 2011).

In this paper, we present an analysis of the geometry and structural evolution of the Glenelg Inlier and surrounding Morar Group rocks. In addition to referenced material, the datasets we use are: (1) detailed mapping by JGR and DD from the southern Glenelg Inlier and immediate surrounding Morar Group; (2) detailed mapping data from the Loch Hourn areas from PWGT (described in detail in Tanner 1965) and (3) reconnaissance and targeted detailed mapping in Knoydart and the northern Glenelg Inlier by MK, AGL and KMG.

Geological setting

The Moine Nappe and its setting

The Scottish Caledonides are dominated by two thrust systems: the lower Moine Thrust Zone in the west and the higher Sgurr Beag/Naver Thrust system further east (Fig. 2a; e.g. Tanner 1971; Strachan *et al.* 2010). The undeformed foreland of the Caledonides to the NW comprises Archean–Paleoproterozoic Lewisian gneiss, overlain by early Neoproterozoic Torridon Group sandstone, in turn unconformably overlain by a Cambro-Ordovician shelf sequence. The Moine Thrust Zone is a fold-and-thrust belt that imbricates this distinctive succession, and is capped by the brittle–ductile Moine Thrust (e.g. Peach *et al.* 1907; Elliott & Johnson 1980; Coward 1983; Butler 2010; Krabbendam & Leslie 2010; Thigpen *et al.* 2010).

On the Isle of Skye, west of Glenelg, the Moine Thrust Zone is dominated by the large-scale recumbent Lochalsh Syncline within the Kishorn Nappe (Bailey 1955; Coward & Potts 1985), which contains a thick sequence of Neoproterozoic siliciclastic sedimentary rocks of the Sleat and lower Torridon groups (Stewart 2002; Kinnaird *et al.* 2007; Krabbendam *et al.* 2017). The Moine Nappe, above the Moine Thrust, comprises a large (>50 × 200 km) outcrop of metasedimentary Morar Group rocks, with basement inliers including the Glenelg Inlier, mostly deformed under amphibolite-facies conditions (e.g. Barr *et al.* 1986; Strachan *et al.* 2010; Thigpen *et al.* 2013). In Sutherland, the northern Moine Nappe (*sensu lato*) comprises several thin (<1 – 2 km) ductile thrust slices, each containing strongly sheared Morar Group rocks and basement gneisses (e.g. Barr *et al.* 1986; Strachan & Holdsworth 1988; Holdsworth 1989; Holdsworth *et al.* 2007). In contrast, in Ross-shire (Fig. 2a), the central Moine Nappe is a single entity with large-scale west-vergent folds but no structural breaks (Leslie *et al.* 2010; Krabbendam *et al.* 2011). In this area, a clear Morar Group succession, many kilometres thick, has been documented, the lower part of which has been correlated with the Torridon Group west of the Moine Thrust (Krabbendam *et al.* 2008; Bonsor *et al.* 2010, 2012). A similarly thick sequence of Morar Group rocks occurs in Morar and Knoydart and, despite high-grade metamorphic overprint, the gross stratigraphies of Ross-shire, Knoydart and Morar can be correlated (Fig. 3) (Johnstone *et al.* 1969; Holdsworth *et al.* 1994; Krabbendam *et al.* 2014). The Glenelg Inlier occurs within the Moine Nappe, between these two areas of well-established Morar Group stratigraphy (Ramsay & Spring 1962).

The Moine Nappe Complex is structurally overlain by the ductile, amphibolite-facies Sgurr Beag Thrust (Fig. 2) (Tanner 1971, 1976; Powell *et al.* 1981; Roberts *et al.* 1987). The overlying Sgurr Beag Nappe comprises the pelitic Glenfinnan and the psammitic Loch Eil groups, and a number of slices of basement gneiss, typically below the Glenfinnan Group at or near the Sgurr Beag Thrust (e.g. Roberts *et al.* 1987; Strachan *et al.* 1988). From the Fannichs Inlier southward, the Sgurr Beag Thrust and Nappe are refolded by folds with axial planes that steepen towards the east (e.g. Roberts & Harris 1983; Roberts *et al.* 1987).

The Glenelg Inlier

The Glenelg Inlier (Figs 2 and 4), also termed the Glenelg–Attadale Inlier (e.g. Storey 2008*a*; Barber 2011), contains two lithologically very different assemblages, termed here 'Western Glenelg gneiss' and 'Eastern Glenelg gneiss'; previous names include 'Western Lewisian/Eastern Lewisian' or the 'Western Unit/Eastern Unit of the Glenelg–Attadale Inlier' (Ramsay 1957; Sutton & Watson 1958; Storey *et al.* 2004; Storey 2008*a*). As these two

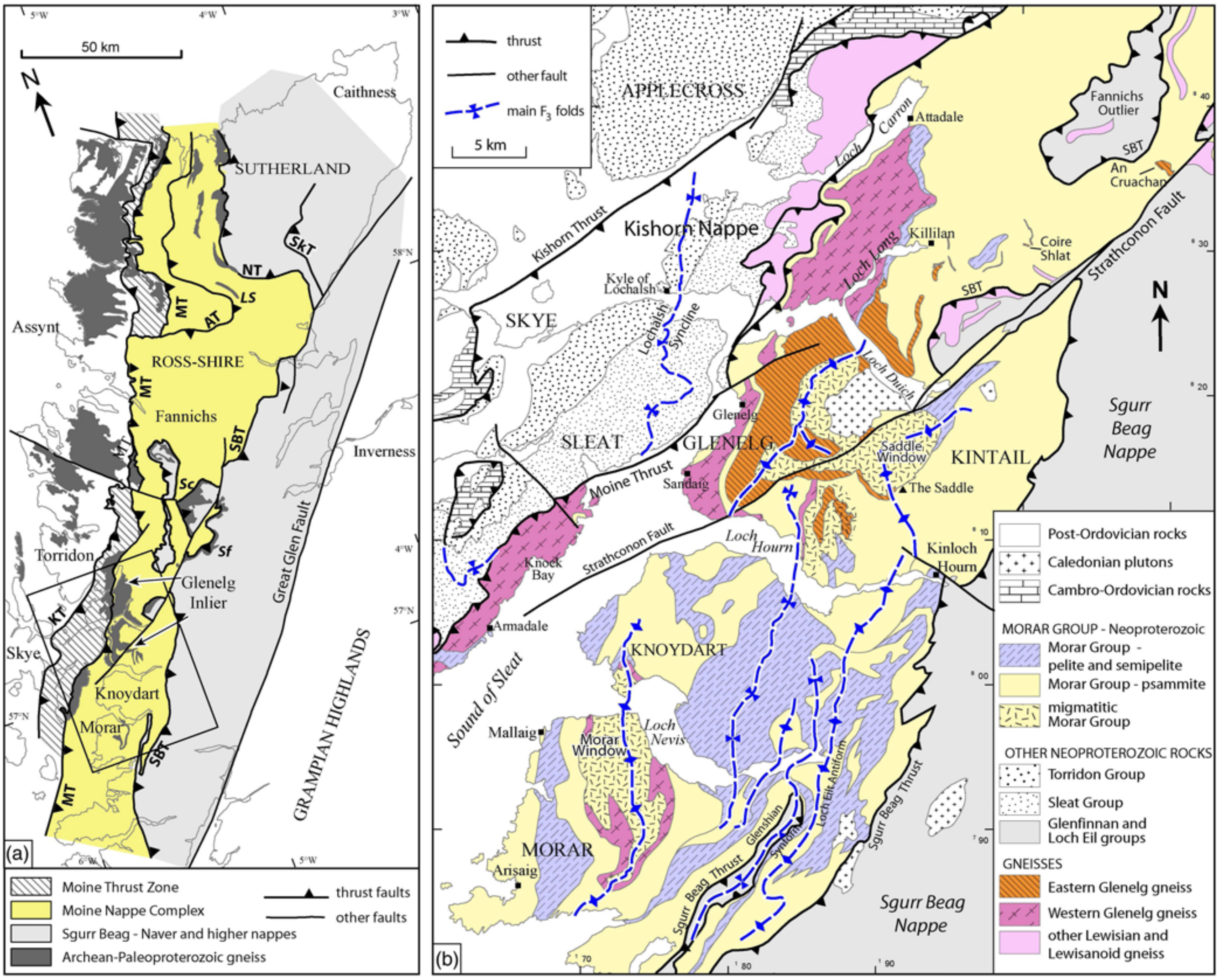

Fig. 2. (**a**) Main thrust sheets in Northern Highlands, with basement gneiss outcrops. Thrusts: AT, Achness Thrust; KT, Kishorn Thrust; MT, Moine Thrust; NT, Naver Thrust; SBT, Sgurr Beag Thrust; SkT, Skinsdale Thrust. Gneiss Inliers: LS, Loch Shin Inlier; Sc, Scardroy Inlier; Sf, Strathfarrar Inlier. After British Geological Survey (2007) and Strachan *et al.* (2010). Box is outline of (b). (**b**) Geological map of the Glenelg Inlier and surrounding area, showing distribution of Western and Eastern Glenelg inliers. After Geological Survey of Scotland (1909), British Geological Survey (1971, 1984), Powell (1974), Mendum (2009*b*) and mapping by the authors.

different gneiss assemblages occur in a number of separate inliers (*sensu stricto*), we first summarize their general lithological characteristics before assigning rocks within the different inliers to a particular assemblage. Comprehensive overviews of the lithology and petrology of the Glenelg Inlier were provided by Storey (2008*a*, *b*) and Mendum (2009*a*), and only a summary is given here.

Western Glenelg gneiss

The Western Glenelg gneiss assemblage consists solely of meta-igneous rocks. These include hornblende-bearing quartzo-feldspathic grey gneiss with abundant inclusions of meta-mafic rocks, mainly of amphibolite, but also including hornblendite, pyroxene-bearing mafic high-pressure granulite and very rare eclogite (Barber & May 1976; Sanders 1988; Storey *et al.* 2005; Storey 2008*a*, *b*). Strain is highly variable: agmatitic textures are locally preserved, with intimate intermingling of mafic and felsic meta-igneous material, suggesting very low strain (Fig. 5a). Elsewhere the gneisses show an intensely developed ‘tramline’ banding, showing extremely high finite strains (Fig. 6c). Locally, amphibolite sheets cross-cut an earlier gneissic layering and likely represent metamorphosed dykes, similar to the ‘Scourie Dykes’ of the Caledonian Foreland (Ramsay 1957). Western Glenelg gneiss is similar to gneisses exposed below the Moine Thrust near Lochcarron (Barber & May 1976) and a correlation with the Lewisian Gneiss Complex seems valid. Both the protolith of the felsic gneiss as well as an early granulite-facies metamorphism are of Archean age (Friend *et al.* 2008; Storey *et al.* 2010). High-pressure granulite-to-eclogite facies metamorphism occurred at *c.* 1.75 Ga (Storey *et al.* 2010), but there is no evidence for Grenvillian metamorphism.

Eastern Glenelg gneiss

The Eastern Glenelg gneiss assemblage contains abundant paragneiss in addition to orthogneiss. Rocks of sedimentary origin include kyanite–garnet-bearing pelitic gneiss, graphitic pelite, forsterite marble with large conspicuous nodules of diopside, tremolite and forsterite (Fig. 5b), and a variety of

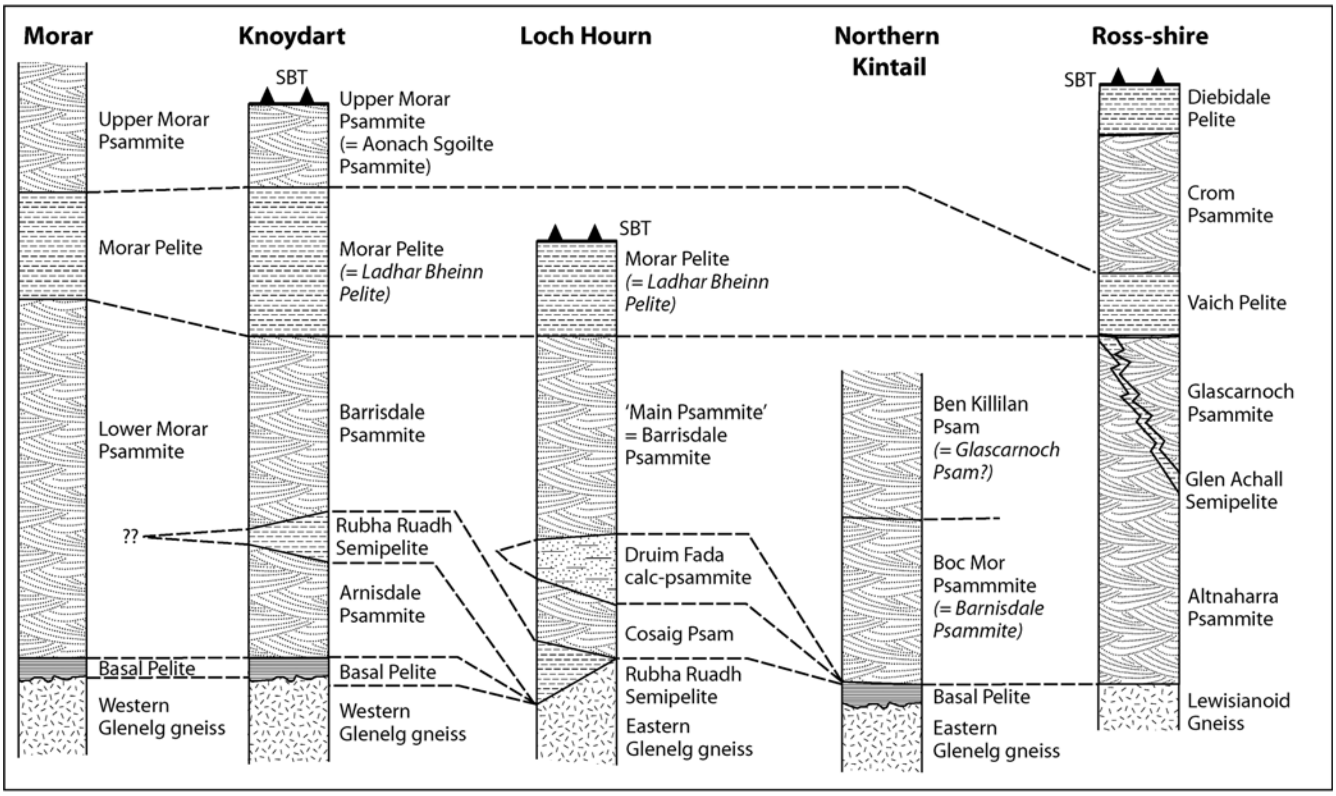

Fig. 3. Stratigraphy of the Morar Group in different sub-areas, from the SW (Morar) to the NE (Ross-shire). Morar: Johnstone *et al.* (1969); Knoydart: Ramsay & Spring (1962); Loch Hourn: Tanner (1965); Northern Kintail: May *et al.* (1993); Ross-shire: Bonsor *et al.* (2012). SBT, Sgurr Beag Thrust.

gneissose semipelitic to psammitic rocks (Peach *et al.* 1910; Alderman 1936; Sanders 1988; Rawson *et al.* 2001; Storey *et al.* 2005; Storey 2008*a*, and our observations). Eclogite boudins or layers (Fig. 5c), commonly retrogressed to symplectitic eclogite or amphibolite, are abundant in both paragneisses and orthogneisses. Many eclogite bodies probably originated as mafic intrusions, but some thin, laterally extensive eclogite layers in paragneiss have gradational boundaries, and may have a volcanic or volcaniclastic origin. Felsic eclogite, comprising kyanite, quartz and K-feldspar occurs locally (Sanders 1988; Storey *et al.* 2005; Storey 2008*b*). Rare ultramafic rocks include garnet-bearing websterite (Rawson *et al.* 2001).

Peak-metamorphic conditions have been estimated at 17 – 23 kbar at 730 – 780°C (Sanders 1989; Rawson *et al.* 2001; Storey *et al.* 2005), although Sajeev *et al.* (2010) suggested higher estimates. The eclogite-facies metamorphism has been dated at 1082 ± 82 Ma (Sm–Nd on garnet; Sanders *et al.* 1984), and subsequent amphibolite-facies retrogression, involving near-isothermal decompression (Storey *et al.* 2005), has been dated at 995 ± 8 Ma (U–Pb on zircon and titanite; Brewer *et al.* 2003). These dates provide unequivocal evidence for Grenvillian high-pressure metamorphism, followed by late-Grenvillian exhumation, in the Eastern Glenelg gneiss.

The Morar Group and contacts with the basement

The Morar Group is dominated by a broad tripartite grouping (Fig. 3) of Lower Morar Psammite Formation–Morar Pelite Formation–Upper Morar Psammite Formation or their lateral equivalents (e.g. Richey & Kennedy 1939; Johnstone *et al.* 1969; Holdsworth *et al.* 1994; Bonsor *et al.* 2012). In Knoydart, Ramsay & Spring (1962) further subdivided the Lower Morar Psammite Formation into four units, from base to top: the Basal Pelite, Arnisdale Psammite, Rubha Ruadh Semipelite and Barrisdale Psammite (see also Krabbendam *et al.* 2014). Deposition of the Morar Group occurred after *c.* 1000 Ma, as indicated by the youngest recorded detrital zircons (Friend *et al.* 2003; Kirkland *et al.* 2008; Cawood *et al.* 2015). Although much of the Morar Group is strongly deformed, sedimentary structures are locally well preserved in low strain zones (Fig. 5d), and sedimentological analysis points to deposition in both braided-river fluviatile and shallow-marine environments (Glendinning 1988; Bonsor & Prave 2008; Krabbendam *et al.* 2008; Bonsor *et al.* 2010, 2012). The Morar Group has an original unconformable relationship against Western Glenelg gneiss, with good evidence for a metaconglomerate at the contact at Attadale [NG 913 377] and SW of Glenelg village [NG 793 174] (Figs 2 and 4 for locations), as well as angular relationships between gneissosity and bedding SSW of Glenelg village (Peach *et al.* 1910; Ramsay 1957; Barber 2009) and near Slisneach in NW Knoydart (Fig. 5e; Krabbendam *et al.* 2014). In contrast, no unequivocal evidence for an unconformable contact between the Morar Group and the Eastern Glenelg gneiss has been found.

Migmatitic Morar Group rocks are psammitic, semipelitic and pelitic rocks that are strongly recrystallized, showing coarse-grained gneissose and commonly migmatitic textures (Fig. 5f; see also Richey & Kennedy 1939; Kennedy 1955; Lambert 1958; Lambert & Poole 1964; Simony 1973; May

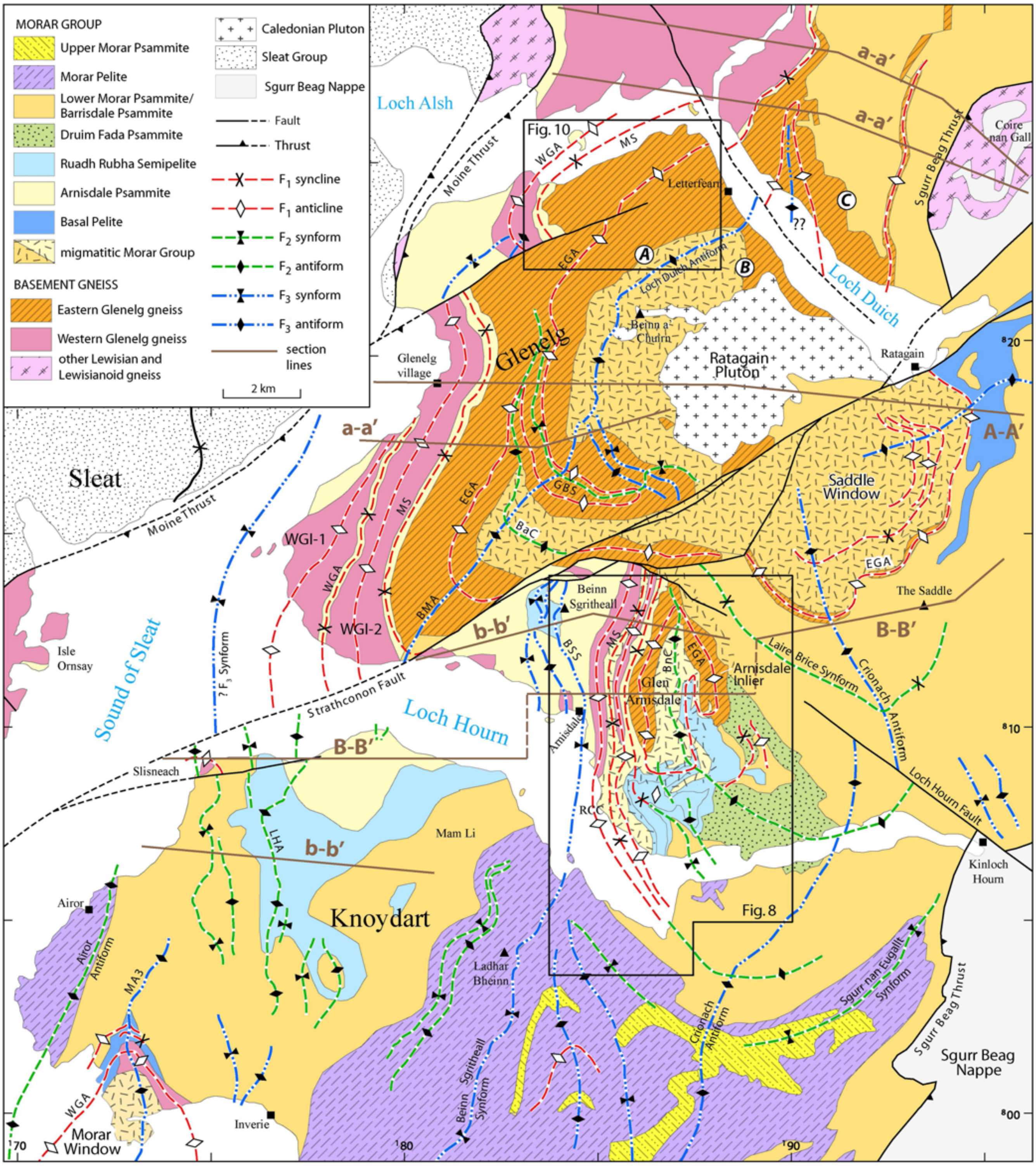

Fig. 4. Geological map of the Glenelg Inlier and surrounding area, with Morar Group units subdivided and major fold traces shown. After British Geological Survey (1971, 1984), Ramsay & Spring (1962), Tanner (1965), Simony (1973) and mapping by the authors. Abbreviations for fold axial traces: BaC, Beinn a' Chapuill Antiform (F2); BnC, Beinn nan Caorach Antiform (F2); BSS, Beinn Sgritheal Synform (F3); BMA, Beinn Mhialairigh Antiform (F3); GBS, Gleann Beag Synform (F2); MA3, Morar Antiform (F3); MS, Median Syncline (F1); LHA, Loch Hourn Antiform (F2); WGA, Western Glenelg Anticline (F1); EGA, Eastern Glenelg Anticline (F1). Inliers: WGI-1; WGI-2: Western Glenelg Inlier 1 and 2; RCC, Rubha Camas na Cailinn. Boxes show outlines of Figures 8 and 10. Main section lines (A–A′ and B–B′) and supporting section lines (a–a′ and b–b′) of Figure 9 are indicated. Points ***A***, ***B*** and ***C*** described in text.

et al. 1993). These rocks were termed 'Sub-Moine' by Richey & Kennedy (1939), but were subsequently recognized as belonging to the Morar Group (e.g. Kennedy 1955; Tanner 1965). Quartz and quartz–feldspar segregations with stromatic or augen textures occur locally; pegmatite is also common. Sedimentary structures and younging evidence have generally been obliterated by pervasive recrystallization and deformation and localized partial melting. In the remainder of this paper, these rocks are referred to as 'migmatitic Morar Group'.

Fig. 5. Field photographs of lithologies of Glenelg gneiss and Morar Group. (**a**) Agmatitic texture in Western Glenelg gneiss, showing relic magma mixing of original felsic and mafic rocks. Glenelg village. Hammer 40 cm long. BGS Photo P707935. (**b**) Marble with calc-silicate nodules; Eastern Glenelg gneiss. North of Gleann Beag. BGS Photo P779470. (**c**) Layered mafic and felsic eclogite, folded; Eastern Glenelg gneiss. BGS Photo P779489. (**d**) Coarse psammite with trough cross-bedding, truncated by planar-bedded psammite, right-way-up. Barrisdale Psammite, Morar Group, central Knoydart (Mam Li). BGS Photo P779591. (**e**) Unconformity of thinly bedded psammite and semipelite (Morar Group) above intermediate gneiss (Western Glenelg gneiss), preserving angular relationship between bedding and gneissose fabric. Slisneach, NW Knoydart. BGS Photo P779573. (**f**) Coarse psammite with quartzo-feldspathic segregations and sigma-clast of deformed pegmatite. Migmatitic Morar Group, just east of Eastern Glenelg gneiss; NW of Beinn a' Chuirn. BGS Photo P779499.

Faulting and intrusion

The eastern part of the Glenelg Inlier was intruded by the late-Caledonian monzonite–monzodiorite Ratagain Pluton, dated at 425 ± 3 Ma (Rogers & Dunning 1991). Intrusion of the pluton is thought to be associated with movement along the Strathconon Fault (Hutton & McErlean 1991), one of a family of Silurian–Devonian NE–SW sinistral strike-slip faults found throughout the Scottish Highlands (Watson 1984). The fault cuts and displaces the Ratagain Pluton (Fig. 4) and associated dykes (Peach *et al.* 1910; Hutton & McErlean 1991) with an apparent sinistral displacement.

Deformation phases

The analysis, definition and correlation of different deformation phases in polydeformed terranes are fraught with difficulty and can lead to confusion (e.g. Forster & Lister 2008). Throughout the Moine, different authors

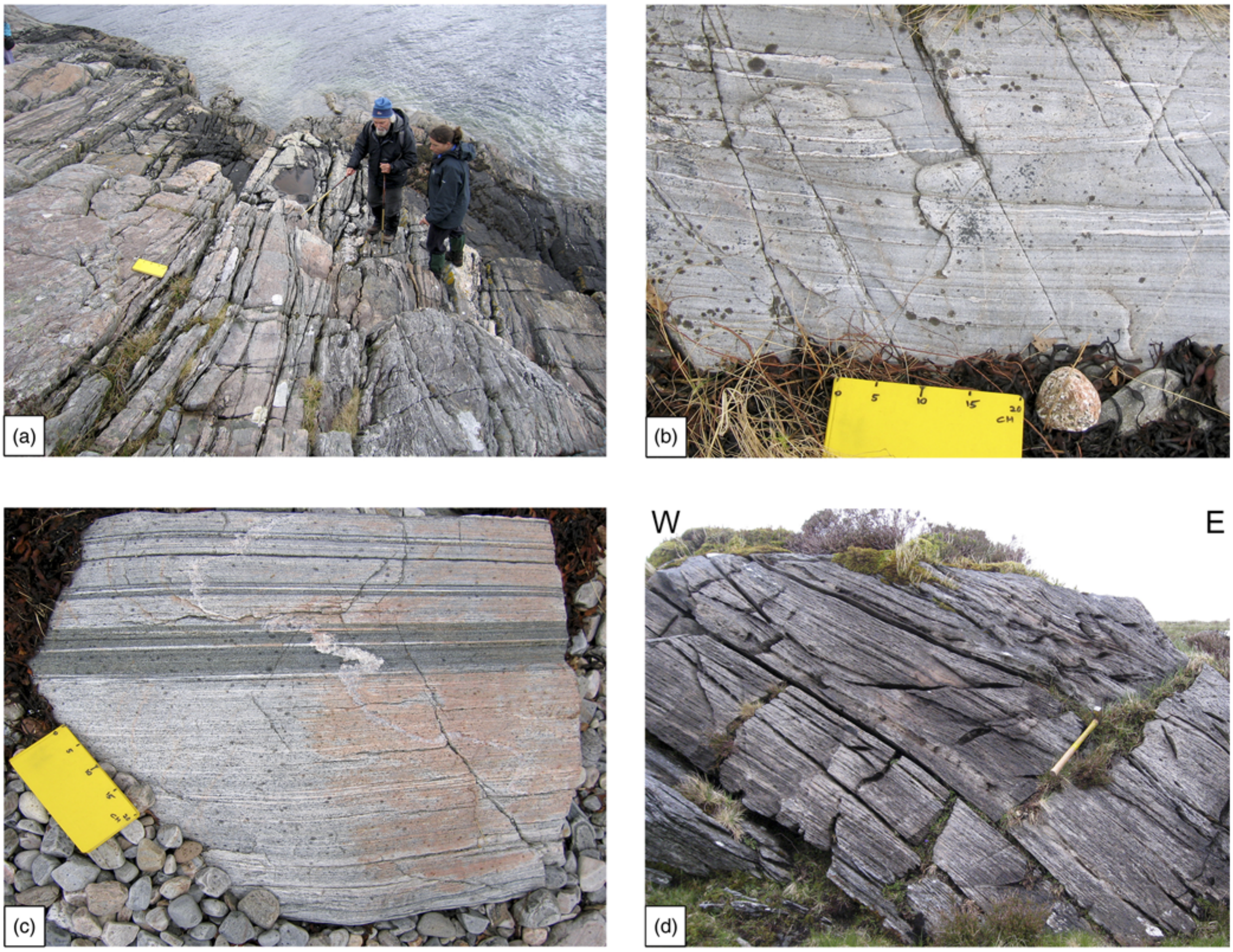

Fig. 6. Field photographs of early structures: F_1 folds and shear zones. (**a**) Isoclinal F_1 folding of Morar Group psammite and Western Glenelg gneiss. Viewed downward to south. Rubha Camas na Cailinn. BGS Photo P707471. (**b**) Isoclinal F_1 folds within Morar Group psammite, looking west; between Coran village and Rubha Camas na Cailinn. BGS Photo P707458. (**c**) Very finely banded, ultramylonitic gneiss, cut by thin pegmatite, western margin of WGI-1 inlier, Western Glenelg gneiss, south of Arnisdale village. BGS Photo P707451. (**d**) Strongly sheared, talc-amphibolite mylonite along eastern boundary of Eastern Glenelg gneiss, Inverinate Shear Zone, west of Beinn a' Chuirn. Hammer 40 cm long; view to NNE. BGS Photo P779508.

have used different deformation-phase schemes (overviews in Powell (1974) and Storey (2008*a*)). The deformation phases referred to in this paper are characterized as follows.

(1) **D_1** is represented by isoclinal folds (Fig. 6a and b), that are refolded by D_2 folds and/or overprinted by D_2 fabrics. These folds are commonly only recognizable by careful mapping and recognition of repetition within the sequence, and interfolding of basement gneiss and metasediment. Because of strong subsequent deformation and recrystallization, linear features related to F_1 folds are rare. Where abundant (e.g. in Glen Arnisdale, Fig. 4), they are characterized by quartz rodding that is lacking in feldspar (Ramsay 1963). Quartz vein development was particularly strong prior to F_1, so that many F_1 folds deform quartz veins but not quartzo-feldspathic segregations, which only developed after F_1 folding (Tanner 1965, p. 95 ff.). This suggests that the metamorphic grade of D_1 did not reach that of the peak-metamorphic conditions during D_2.

(2) **D_2** structures are characterized by amphibolite-facies fabrics (S_2 and L_2) that are penetrative in most lithologies. In psammitic rocks, S_2 is characterized by a quartz-flattening fabric and alignment of dispersed flakes of biotite, whereas in pelitic rocks the S_2 fabric is normally a well-developed mica-dominated schistosity. In calc-silicate rocks most metamorphic minerals (garnet, zoisite, biotite, amphiboles) grew or recrystallized during F_2, although some hornblende growth outlasted F_2 deformation (Tanner 1976). Within Eastern Glenelg gneiss, earlier-formed kyanite, staurolite and garnet remained stable, whereas eclogite minerals were further retrograded to garnet amphibolite (Alderman 1936; Storey 2008*a*). Within Western Glenelg gneiss, garnet remained stable. In the Basal Pelite and at some locations in the Western Glenelg Inlier (Sandaig, Attadale and Knock Bay on Skye, see Fig. 2 for locations), hornblende occurs either as prismatic crystals defining the local D_2 mineral lineation or as *garbenschiefer* in which randomly orientated porphyroblasts lie in the S_2 fabric. Low in

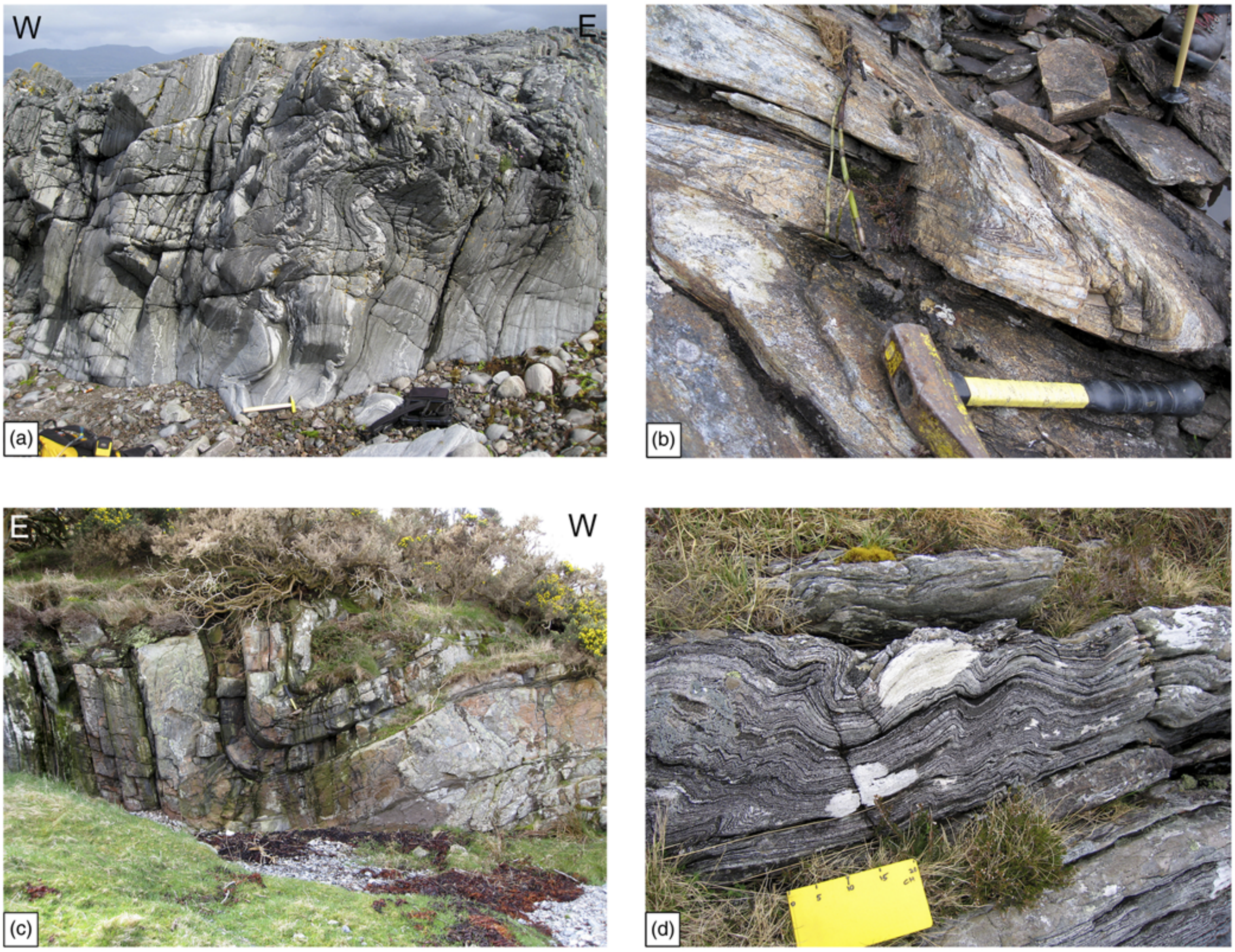

Fig. 7. Field photographs of D_2 and D_3 structures. (**a**) F_2 folds, modest strain. Open–close, recumbent folds; in west-facing hinge zone of Airor Antiform; basal part of Morar Pelite Formation. View to north; near Airor, west Knoydart. BGS Photo P779625. Height of outcrop *c.* 3 m. (**b**) F_2 folds, high strain. Tight to isoclinal F_2 folds in Western Glenelg gneiss, 2.5 km SSW of Glenelg village. BGS Photo P707999. (**c**) Hinge zone of F_3 Beinn Sgritheall Synform, south of Arnisdale village, looking south. BGS Photo P707850. Height of outcrop *c.* 5 m. (**d**) Minor F_3 folds in thinly banded Western Glenelg gneiss, Rubha Camas na Cailinn. BGS Photo P707465.

the structural pile, D_2-related metamorphism was locally accompanied by migmatization and the formation of quartz–feldspar–mica segregations (Tanner 1965, pl. 23). However, some quartz–feldspar–mica segregations and pegmatite intrusions cross-cut F_2 folds, but are deformed by F_3 folds (Tanner 1965, pl. 24). In summary, peak metamorphism was broadly coeval with D_2, but some mineral growth and partial melting outlasted D_2 deformation (Ramsay 1963; Tanner 1965, 1976).

F_2 fold styles are varied and depend strongly on lithology and finite strain. In thick-bedded psammite and low–medium strain zones, as in western Knoydart, F_2 folds are open to close, near-concentric folds (Fig. 7a) with poor fabric development and a dominance of intersection over stretching lineations (Krabbendam *et al.* 2014). In higher strain areas (e.g. in Glen Arnisdale), F_2 folds are tight to isoclinal, similar in shape (Fig. 7b) and associated with a strong linear mineral orientation and fold-axis-parallel quartz–feldspar–mica rods.

(3) **D_3** folds are recognized by folding or crenulation of all penetrative fabrics (Fig. 7c and d). Only minor recrystallization occurred during D_3; new fabric formation, normally a crenulation cleavage, is restricted to hinge zones. D_3 structures thus developed at lower metamorphic temperatures than the D_2 structures. F_3 folds are typically open to close; no tight or isoclinal F_3 folds have been observed.

All of these deformation phases affected Morar Group rocks as well as Glenelg gneisses and produced minor fold structures as well as regional-scale folds. This deformation-phase scheme is broadly consistent with that reported in Ramsay (1960, 1963, 2010*a*), Ramsay & Spring (1962), Tanner (1965, 1971) and Powell (1966, 1974). Differences with other published schemes occur because some studies did not account for the early isoclinal D_1 folds (e.g. Ramsay 1957), did not separate the F_1 isoclinal infolding of gneiss and psammite from the later F_2 folds (e.g. Storey *et al.* 2004; Storey 2008*a*) or may not have linked fold generations to specific mineral growth phases (e.g. Barber 2011). We stress that the deformation-phase scheme above applies only to the southern Moine Nappe as discussed here. Deformation phases in the Moine outcrop as a whole cannot *a priori* be correlated across major thrusts and may only be valid within

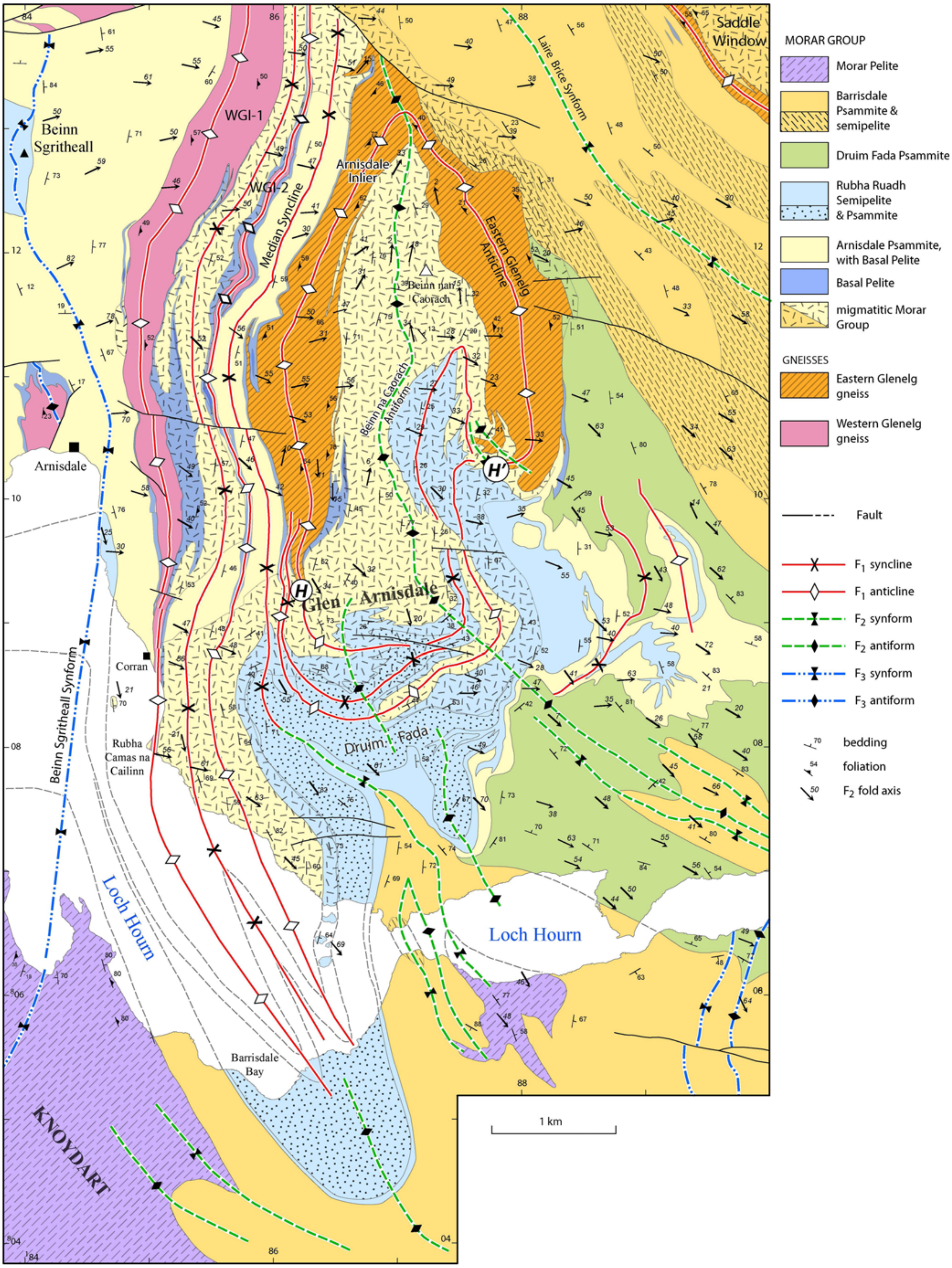

Fig. 8 . Detailed geological map of Glen Arnisdale, mapping by PWGT and JGR. Points ***H*** and ***H′*** are discussed in the text.

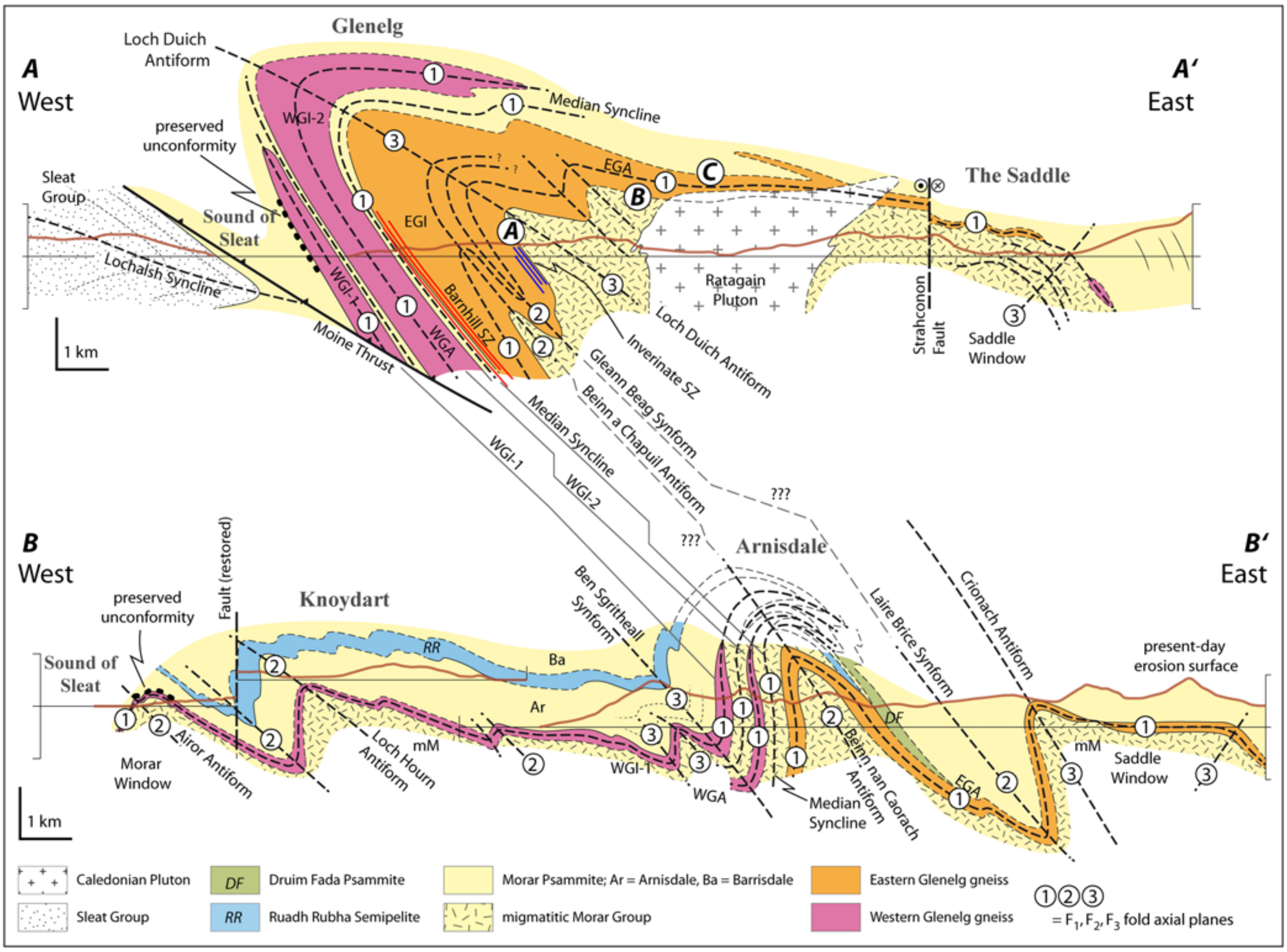

Fig. 9 . Schematic cross-section of the Glenelg Inlier north of the Strathconon Fault (A–A′) and south of the Strathconon Fault (B–B′). Cross-sections were constructed along two main section lines (A–A′ and B–B′), with supporting cross-section (a–a′ and b–b′) marked on Figure 4, stacked on to each other to portray greater structural depth. Linkages between fold structures are discussed in the text. Note that the axial planes of the Beinn a' Chapuill folds are highly oblique to plane of section. Points ***A***, ***B*** and ***C*** are discussed in the text. EGA, Eastern Glenelg Anticline; EGI, Eastern Glenelg Inlier; mM, migmatitic Morar group; WGA, Western Glenelg Anticline; WGI-1, WGI-2, Western Glenelg Inlier 1 and 2, respectively.

a particular thrust nappe (Holdsworth *et al.* 2006; Leslie *et al.* 2010; Krabbendam *et al.* 2011; cf. Tanner 1971; Powell 1974; Mendum 2009*a*; see also discussion in Strachan *et al.* 2010). Grenvillian and possibly earlier deformation within the Glenelg gneisses prior to Morar Group deposition was significant, but is not considered here; see Storey (2008*a*) for an analysis of this earlier deformation. The absolute ages of the different deformation phases are discussed later.

Disposition of gneiss units and migmatitic Morar Group

Western Glenelg gneiss makes up the main Western Glenelg Inlier which comprises two separate strips (WGI-1 and WGI-2), separated by a thin unit of highly deformed Morar Group psammite (Fig. 4). North of Glenelg village this thin psammite strip thins out and a single band of Western Glenelg gneiss continues north to Lochcarron (Fig. 2b), where its northern termination unfortunately occurs below sea level (Barber 2011). Basement gneiss on southern Sleat directly above the Moine Thrust is also Western Glenelg gneiss (Fig. 2b). SE of the Strathconon Fault, two narrow strips of Western Glenelg gneiss occur east of Arnisdale village (Figs 4 and 8); these are the equivalent to the two WGI-1 and WGI-2 strips NW of the fault (Ramsay 2010*a*). Smaller inliers of Western Glenelg gneiss occur along the north coast of Loch Hourn west of Arnisdale village; these occur at a similar structural level as a small inlier of Western Glenelg gneiss at Slisneach, in NW Knoydart (Fig. 4).

In SW Knoydart a hook-shaped inlier of Western Glenelg gneiss (Ramsay & Spring 1962; Krabbendam *et al.* 2014) forms the northern part of the Morar Window, which continues further south of Loch Nevis. This latter gneiss inlier is exposed in a series of narrow outcrops that almost surround the Morar Window (Fig. 2b): these gneisses comprise felsic and mafic orthogneiss with mafic sheets only (Lambert & Poole 1964), suggesting correlation with Western Glenelg gneiss.

Eastern Glenelg gneiss, with its distinct metasedimentary lithologies, makes up the main Eastern Glenelg Inlier. This inlier occurs between the Strathconon Fault and Loch Duich, and continues north to terminate just south of Loch Long (Fig. 2b; May *et al.* 1993; Storey 2008*a*; Barber 2011). A thin, partially dismembered sliver of gneiss directly south of Kilillan appears to be continuous with the main Eastern Glenelg Inlier (British Geological Survey 1984). Inliers further NE are small or poorly surveyed, but crystalline marble is reported in Coire Shlat (Fig. 2b; May *et al.* 1993), marble and calc-silicate are reported in Coire nan Gall

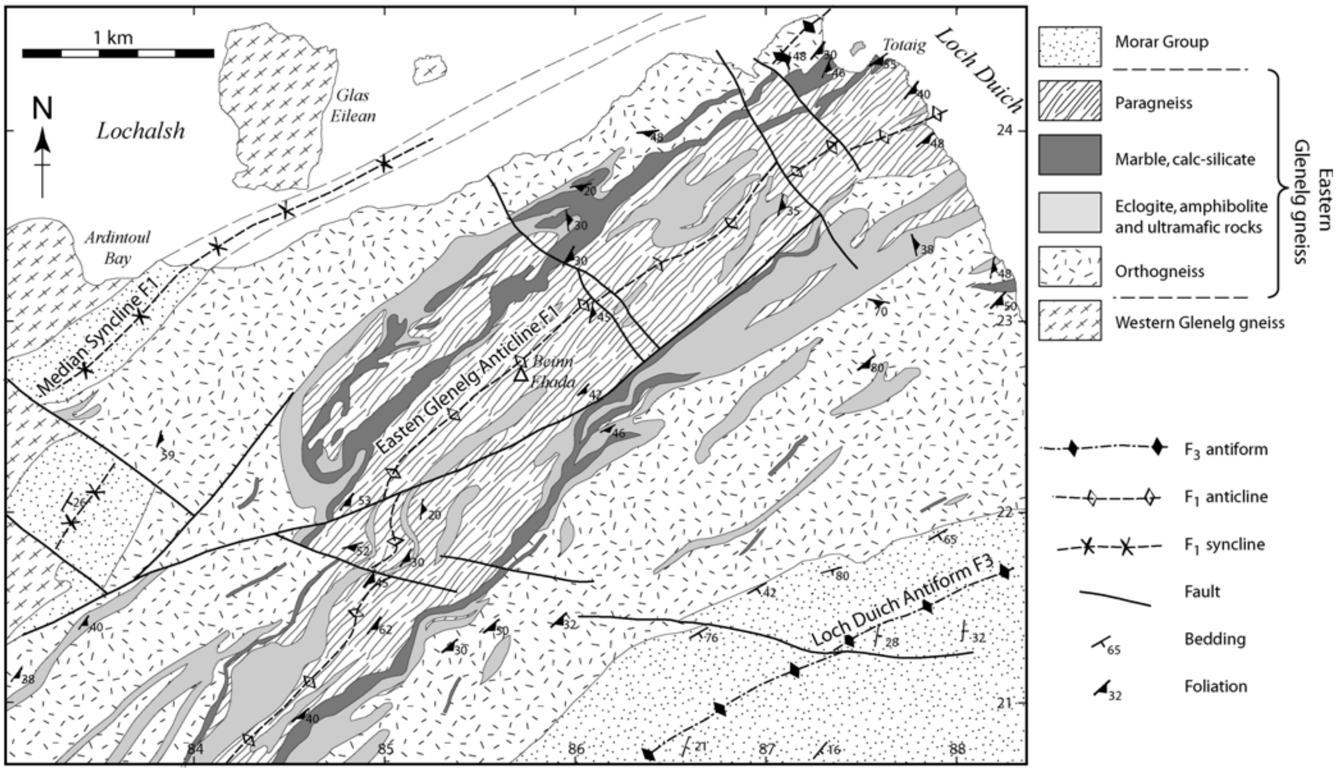

Fig. 10. Detail of northern part of the Eastern Glenelg Inlier, showing subdivision of lithologies within the Eastern Glenelg Inlier. The isoclinal Eastern Glenelg Anticline is interpreted on the basis of symmetry of lithologies; the close Loch Duich Antiform is constrained on the basis of folded fabrics – see text. After Sutton & Watson (1958), Storey (2008*a*) and mapping by the authors.

(Fig. 4; May *et al.* 1993), whereas the An Cruachan gneiss inlier 10 km further NW [NH 09 35] is reported to contain graphitic schist and eclogitic rock (Peach *et al.* 1913), consistent with the Eastern Glenelg gneiss assemblage. A number of 'Lewisianoid' gneiss inliers further NE (e.g. Scardroy, Strathfarrar and Loch Shin inliers; Fig. 2a) also contain layers of quartzite, calc-silicate gneiss, crystalline marble and/or graphitic schist (Read *et al.* 1926; Sutton & Watson 1953; Strachan *et al.* 2002; Ramsay 2010*b*), suggesting possible affinity with Eastern Glenelg gneiss.

South of the Strathconon Fault, an arcuate inlier of Eastern Glenelg gneiss with eclogite and forsterite marble occurs near Glen Arnisdale, termed here the Arnisdale Inlier (Figs 4 and 8). The thin gneiss sliver that surrounds the Saddle Window (Fig. 4) contains forsterite-bearing marble, calc-silicate, pelitic gneiss and rare eclogite (Simony 1973; May *et al.* 1993) and thus also comprises Eastern Glenelg gneiss. No rocks resembling the Eastern Glenelg gneiss have been found SE of Glen Arnisdale or The Saddle.

Migmatitic Morar Group rocks occur in (1) the Morar Window in the core of the Morar Antiform, (2) the Saddle–Ratagain Window, (3) in Glen Arnisdale. In Glen Arnisdale the migmatitic front obliquely cross-cuts well-mapped stratigraphic boundaries of the Morar Group, confirming the migmatitic rocks are indeed Morar Group (Kennedy 1955). All occurrences are bound by, or are closely associated with, basement gneiss (Figs 2b, 4 and 8).

Major structures

The main fold traces are shown on Figure 4 and in cross-sections on Figure 9.

Major F_1 folds

Exceptional exposures of outcrop-scale F_1 folds occur at Rubha Camas na Cailinn, on the north shore of Loch Hourn (Fig. 8), where Morar Group psammite and Western Glenelg gneiss (WGI-1) are isoclinally interfolded (Fig. 6a; Ramsay 1963, 2010*a*; see also description in Mendum 2009*a*); these folds are demonstrably overprinted by F_2 folds, fabrics and lineations. The F_1 folds here plunge SSE. SE of Rubha Camas na Caillin, the Morar Group strip between the Western and Eastern Glenelg gneiss opens out southwards into a clear F_1 synclinal structure, the *Median Syncline* (Fig. 8) and involves stratigraphically higher units of the Morar Group in its hinge zone. These tight to isoclinal F_1 synclines clearly equate with those responsible for the thin strips of Morar Group alternating with gneiss that occur NW of the Strathconon Fault. Thus, both Western and Eastern Glenelg gneisses occur in the cores of isoclinal F_1 anticlines, termed the *Western Glenelg Anticline* and *Eastern Glenelg Anticline*, respectively (Figs 8 and 9).

NW of the Strathconon Fault, evidence for large-scale F_1 folding is provided by the symmetrical disposition of lithologies in the main Eastern Glenelg Inlier (Fig. 10). SW of Loch Duich this inlier has a core of paragneiss that is flanked on both sides firstly by a near-continuous strip of forsterite marble and calc-silicate rock, then by grey orthogneiss with abundant eclogite layers, and finally by Morar Group psammite. In the far north of the main Western Glenelg Inlier near Attadale (Fig. 2b), Barber (2011) reported isoclinal F_1 folds on a *c.* 100 m scale, folded by recumbent F_2 folds with strong axial planar and linear fabric.

Returning to the SE side of the Strathconon Fault, the gneiss inliers that contain Western Glenelg gneiss in western Knoydart and Morar have Morar Group rocks both structurally above and below them. The Slisneach gneiss inlier is demonstrably folded by F_2 folds, and the Morar inlier by the F_2/F_3 Morar Antiform (Krabbendam *et al.* 2014). This is consistent with these gneiss inliers forming cores of F_1 anticlines, although it is possible that some shearing/thrusting may have occurred (e.g. Ramsay & Spring 1962; Lambert & Poole 1964; Powell 1974).

In summary, all Glenelg gneiss inliers occur in the cores of early F_1 anticlines, whereas Morar Group psammite occurs consistently in the cores of F_1 synclines (Figs 8 and 9).

Early shear zones

The western margin of the main Eastern Glenelg Inlier is marked by a zone of ultramylonite, tens of metres wide (Ramsay 1957), originally mapped by Clough (Geological Survey of Scotland 1909) as ‘extremely attenuated bands’ and termed the *Barnhill Shear Zone* by Storey *et al.* (2004). Felsic gneisses are reduced to fine-grained and finely laminated quartz–feldspar rocks, while hornblendic and garnetiferous rocks are now seen as olive-grey banded ultramylonite. Strong deformation also affected the ‘Median Strip’ of Morar Group psammite between the main Western and Eastern Glenelg inliers (Storey *et al.* 2004; Storey 2008*b*) and adjacent parts of the Western Glenelg gneiss, and also affects the Western Glenelg gneiss near Arnisdale; an example is shown in Figure 6c. A similar high-strain zone occurs on the eastern boundary of the Eastern Glenelg Inlier, near Beinn a’Chuirn; the *Inverinate Shear Zone* of Storey *et al.* (2004). This shear zone locally comprises mylonitic amphibolite (Fig. 6d), demonstrating that shearing occurred after eclogite decompression. Morar Group psammitic rocks east of the contact here are, however, coarse grained, strongly recrystallized and migmatitic (Fig. 5f). This suggests that migmatization post-dated shear zone development.

F_2 folds; transport direction during D_2

Numerous outcrop- and regional-scale F_2 folds occur throughout the area. Large F_2 folds can be mapped using fold vergence changes of minor F_2 folds, and cleavage vergence between bedding and the S_2 fabric (e.g. Bell 1981). However, the intensity of the D_2 strain, F_2 fold geometry, character of the axial planar foliation and the orientation of F_2 fold axes vary considerably from place to place. To assess the orientation of F_2 fold axes with respect to the Caledonian transport direction, we plot first L_2 stretching lineations in areas not affected by F_3 folding, e.g. from Balmacara and Armadale areas (Fig. 11a and b). The stretching lineations in these areas are orientated WNW–ESE (mean: azimuth 108°/plunge 28°), consistent with the regional WNW Caledonian transport direction (*c.* 290°; e.g. McClay & Coward 1981). We then rotate F_2 fold axes on the limbs of F_3 folds to the horizontal around average F_3 fold axes (Fig. 11b–d). It then appears that F_2 fold axes in Knoydart, on the flat limb of the Beinn Sgritheall Synform, are at high angles to the regional transport direction (Fig. 11e). Conversely, F_2 fold axes in Glen Arnisdale and around Glenelg are oriented sub-parallel to the regional transport direction (Fig. 11f and g).

The Knoydart peninsula is dominated by a series of kilometre-scale west-facing and west-vergent F_2 folds (Figs 4 and 9) with north–south-trending fold hinges and consistently east-dipping axial planes and fabrics. Named F_2 folds include the *Loch Hourn* and *Airor antiforms* (Krabbendam *et al.* 2014). Large areas of Knoydart preserve sedimentary structures, including delicate soft-sediment deformation structures (Fig. 5d). L_2 lineations are mainly intersection lineations and trend parallel to the main fold axes; L_2 stretching lineations are rare (Krabbendam *et al.* 2014). F_2 folds are open to close buckle/flexure folds (Fig. 7a), developed at high angles to the transport direction (Fig. 11e). All these features attest to low-to-medium D_2 strain in western Knoydart.

In contrast, in Glen Arnisdale and around Glenelg, D_2 strain was intense, with generally tight F_2 folds (Fig. 7b) accompanied by well-developed L_2 stretching lineations, subparallel to F_2 fold axes. F_2 fold axes and L_2 lineations plunge ESE to SE, subparallel to the inferred transport direction (Fig. 11f and g). The F_2 folds are interpreted to have rotated towards the transport direction as a result of high D_2 shear strain compared to Knoydart. Major, mappable F_2 folds include the *Beinn nan Caorach Antiform* and *Laire Brice Synform* south of the Strathconon Fault and the *Beinn a’ Chapuill* and *Gleann Beag* folds north of the Strathconon Fault (Ramsay 1960; Tanner 1965). The Beinn a’ Chapuill and Gleann Beag folds face broadly to the south, where mapped. Such folds are likely to be curvilinear, but individual major folds are not well enough exposed to test this possibility.

Major F_3 folds

The archetypal F_3 fold is the *Beinn Sgritheall Synform* (Ben Sgriol Synform in older publications – current Gaelic spelling is followed here) south of the Strathconon Fault, analysed in detail by Ramsay (1960). Its hinge zone is well exposed south of Arnisdale village (Fig. 7c), where all penetrative fabrics are folded around the hinge zone, which itself has only a crenulation cleavage as an axial planar fabric (Ramsay 1960, 2010*a*). The axial plane dips *c.* 60° to the east. The fold has a flat to gently east-dipping western limb whereas the eastern limb is subvertical to overturned, steeply east-dipping (Figs 8 and 9).

Some 8 km further east, the *Crionach Antiform* (Tanner 1965, p. 68) forms a large-scale, north–south-trending synform–antiform pair with the Beinn Sgritheall Synform. The fold pair has an overall westerly vergence. Both folds plunge gently to the south, so that higher structural levels are exposed further south (e.g. lower Morar Group, followed by Morar Pelite). Both folds are large-scale structures that can be traced southwards from the Strathconon Fault for 20 – 30 km or more (Fig. 2b). The steep common limb is some 5 – 7 km wide and contains the majority of exposed gneiss south of the Strathconon Fault.

North of the Strathconon Fault, the major *Loch Duich Antiform* occurs west of the Ratagain Pluton (Figs 4 and 9). Around its hinge zone, exposed near Letterfearn, the fold axis plunges 10 – 20° to the NE. All penetrative fabrics in the Eastern Glenelg gneiss, as well as the main S_2 and L_2 fabrics in migmatitic Morar psammite, are folded over, and parasitic D_3 folds are concentric in style. The Loch Duich Antiform is thus unequivocally an F_3 fold (cf. Barber 2011). The

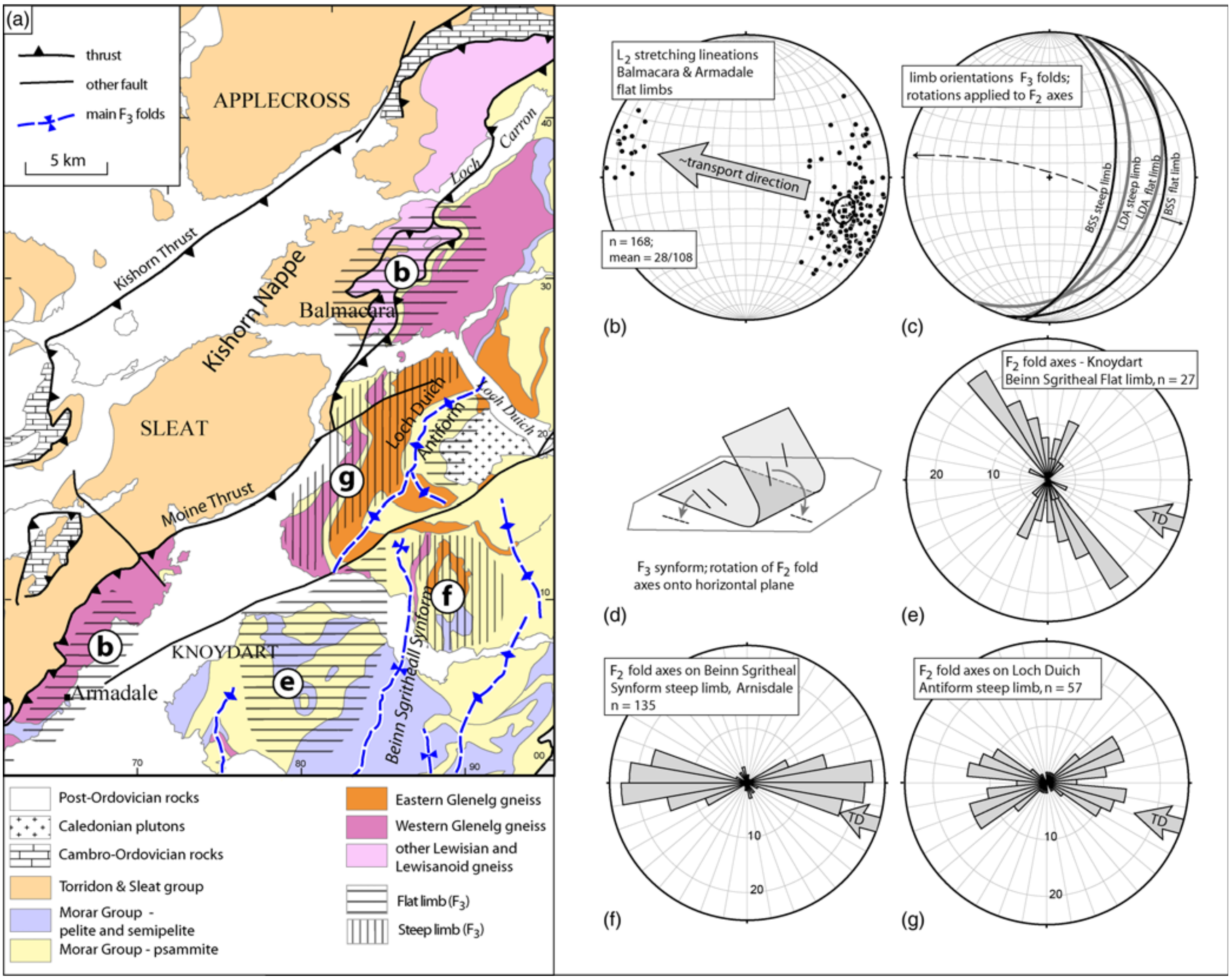

Fig. 11. Orientation analysis of F_2 lineations and fold axes. (**a**) Map showing steep and flat limbs of F_3 folds, used to delineate domains for orientation analysis. Circled letters refer to the respective stereogramme or rose diagrams. (**b**) L_2 stretching lineation from Balmacara and Armadale (Skye), taken as representative for F_2 transport direction. Lower hemisphere projection, equal area. (**c**) Mean limb orientation of steep and flat limbs of the Beinn Sgritheall Synform (BSS) and the Loch Duich Antiform (LDA), based on bedding/foliation orientation in the limbs. These have been used to re-orientate F_2 fold axes on to horizontal plane. (**d**) Schematic diagram, showing rotation of F_2 fold axes from flat and steep limbs of an F_3 synform to the horizontal. (**e**) F_2 fold axes rotated on to horizontal, from Knoydart on flat limb of Beinn Sgritheall Synform. (**f**) F_2 fold axes rotated on to horizontal, from Glen Arnsidale within steep limb of Beinn Sgritheall Synform. (**g**) F_2 fold axes rotated on to horizontal, from Glenelg within steep limb of Loch Duich Antiform.

antiform has an east-dipping axial plane, with a steep to overturned western limb and a gently dipping eastern limb. In the steep, overturned western limb (***A*** on Figs 4 and 9), migmatitic Morar Group overlies the Eastern Glenelg Inlier (with the contact sheared along the Inverinate Shear Zone), whereas on the gently dipping eastern limb (***B*** on Figs 4 and 9) the Eastern Glenelg Inlier lies structurally *above* migmatitic Morar Group (e.g. between Letterfearn and the Ratagain pluton). The F_3 Loch Duich Antiform is thus *locally* a synclinal antiform, facing down to the east (Figs 4 and 9). This local downward facing confirms that the F_3 fold has refolded an earlier, larger-scale fold, most likely the F_1 Eastern Glenelg Anticline.

Overprinting relationships

F_1–F_2 fold interference: Glen Arnisdale

Some of the most structurally complex ground in the area occurs in the steep eastern limb of the Beinn Sgritheall Synform in the Glen Arnisdale area (Fig. 8; described in more detail in Tanner 1965). The effects of the rugged terrain on the outcrop pattern add to the complexity. Two thin slivers of Western Glenelg Inlier (WGI-1 and WGI-2) and one slice of Eastern Glenelg gneiss occupy the cores of major isoclinal F_1 anticlines, separated by F_1 isoclinal synclines cored by Morar psammite. These structures are overprinted by the large-scale F_2 Beinn nan Caorach Antiform (BCA), illustrated on Figures 8 and 9. As a consequence, a great variety of F_1/F_2 fold interference patterns occurs at both outcrop- and map-scale that provides a ready means of distinguishing the two generations of minor folds. Type 3 (co-axial) interference patterns (see Ramsay 1963, 1967; Ramsay & Huber 1987, p. 492, for definitions of fold interference types) between F_1 and F_2 are relatively common across the area (Fig. 12a; Tanner 1965, pls 10, 11); Type 1 (dome-and-basin) occur locally (Tanner 1965, pls 3, 5, 13, 21); and composite Type 1/3 patterns (Fig. 12b) also occur. Type 2 ('mushroom') forms occur rarely (Tanner 1965, pls

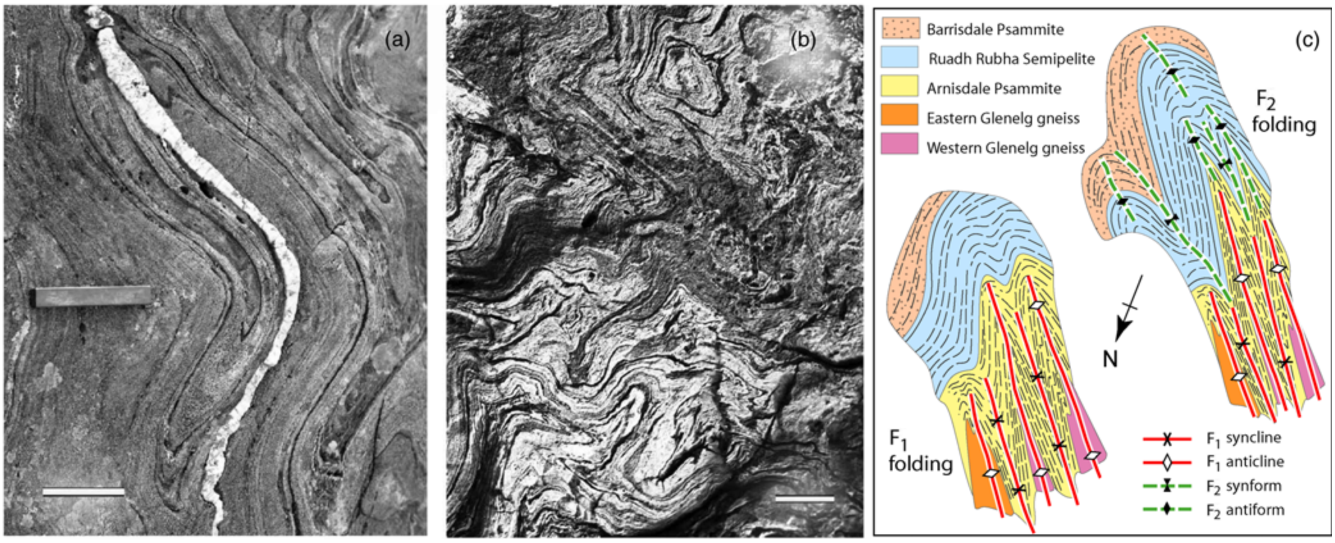

Fig. 12. F_1/F_2 fold interference. The white scale bar in both photos is 10 cm long. (**a**) F_1 isoclinal folds, together with an axial-planar quartz vein, folded by an open F_2 fold; F_1/F_2 axial planes at high angles. Horizontal surface in Coire Mhicrail. (**b**) Tight F_1 folds folded during D_2 and forming a Type 1/Type 3 interference pattern. (**c**) Schematic structural map, oriented down-plunge to the SSE, based upon the geology of Arnisdale – Loch Hourne area, showing tightening of F_1 folds by F_2 folds, resulting in propagation of F_2 fold axial planes to higher stratigraphic levels (after Tanner 1965, p. 124).

14, 22; Ramsay 1967) but clearly reflect the geometry of the Arnisdale Inlier as a whole. A quartz-ribbon L_1 lineation commonly occurs parallel to the hinges of minor F_1 folds, and both elements are deformed into sinusoidal curves on S_2, arranged symmetrically about the steeply SE-plunging D_2 stretching lineation (Ramsay 1960). Most fold patterns are very intricate (Fig. 12a; Tanner 1965, pls 5, 21) and thus likely to be true fold interference patterns rather than sheath folds, but locally isolated eyed structures occur (e.g. Tanner 1965, pl. 9) which likely formed as single-phase F_2 sheath folds, compatible with the strongly developed L_2 stretching lineations throughout the area.

The large arcuate outcrop of the Arnisdale Inlier resulted from Type 2 refolding of the F_1 isoclinal Eastern Glenelg Anticline–with its core of Eastern Glenelg gneiss–by the F_2 BCA (Fig. 13a). The axial plane of the F_2 BCA dips moderately east. The fold is periclinal; in the north its fold axis plunges 20 – 30° north, whereas in the south it plunges 20 – 40° south (Fig. 8). Well-developed L_2 stretching lineations and sub-parallel F_2 fold hinges plunge steeply (40 – 70°) to the east or SE across the area. An exception to this pattern occurs in the hinge zone of the F_2 BCA (Fig. 8), where fold hinges and stretching lineations plunge north or NE. This deviation may be due to the fact that an early axis-parallel lineation formed in the hinge zone of the BCA and was not subsequently rotated towards the stretching direction, as were the lineations on the limbs of the fold and in the enclosing Morar Group rocks.

Southward from Glen Arnisdale, the F_2 BCA changes from a single 'simple' tight structure into a number of individual folds (traceable for a few kilometres at most) in the more distinctly layered Morar Group. The complex outcrop pattern of the Rubha Ruadh Semipelite south of the Arnisdale Inlier is likely the result of F_1/F_2 fold interference patterns superimposed upon an equally complex stratigraphy of interbedded and interfingering sandstone and siltstone units. More complex kilometre-scale fold interference patterns occur in the area, as described in detail in Tanner (1965).

South of Loch Hourn, F_1 axial traces can be followed southward to higher structural and stratigraphic levels, but the same folds change into F_2 folds and are seen to refold minor F_1 folds (Figs 8 and 12c; Tanner 1965, p. 125). This suggests (Tanner 1965) that (1) F_1 folds die out to higher stratigraphic levels, and (2) that subsequent F_2 deformation resulted in a tightening of the F_1 folds and was accompanied by propagation of folding to higher structural levels, resulting in local co-axial overprinting.

East of the Arnisdale Inlier, the trace of a probable F_2 synform, the Laire Brice Synform (Tanner 1965, p. 132), occurs in poorly exposed ground between the Arnisdale Inlier and The Saddle (Figs 4 and 8). This F_2 synform is complementary to the F_2 BCA. The axial plane of the F_1 Eastern Glenelg Anticline, as exposed in Glen Arnisdale, likely continues eastward (folded by the Laire Brice Synform) to reappear as the isoclinal anticline that forms the thin arcuate outcrop of Eastern Glenelg gneiss of The Saddle (Fig. 9, and below).

F_2/F_3 overprinting: Beinn a' Chapuill

Across much of the Moine Nappe, e.g. in Knoydart and Ross-shire, major F_2 and F_3 folds trend broadly north–south, F_2 axial planes are gently east-dipping and F_3 axial planes have a somewhat steeper dip (Powell 1966; Krabbendam *et al.* 2011, 2014). Overprinting relationships in these cases are either of 'Type 0' (coaxial and coplanar) or 'Type 3' (coaxial but non-coplanar).

In contrast, around Beinn a' Chapuill, north of the Strathconon Fault, strongly non-coaxial and non-coplanar F_2 and F_3 overprinting occurs. Here the NE–SW-trending F_3 Beinn Mhialairigh Antiform refolds the F_2 Beinn a' Chapuill and Gleann Beag folds in a 'Type 2' fold interference pattern, described in detail in Ramsay (1957) and shown schematically in Figure 13b. All major fabrics, fold axes and lineations related to D_2 are folded over by the F_3 Beinn Mhialairigh Antiform. F_2 hinges near the hinge zone at Beinn

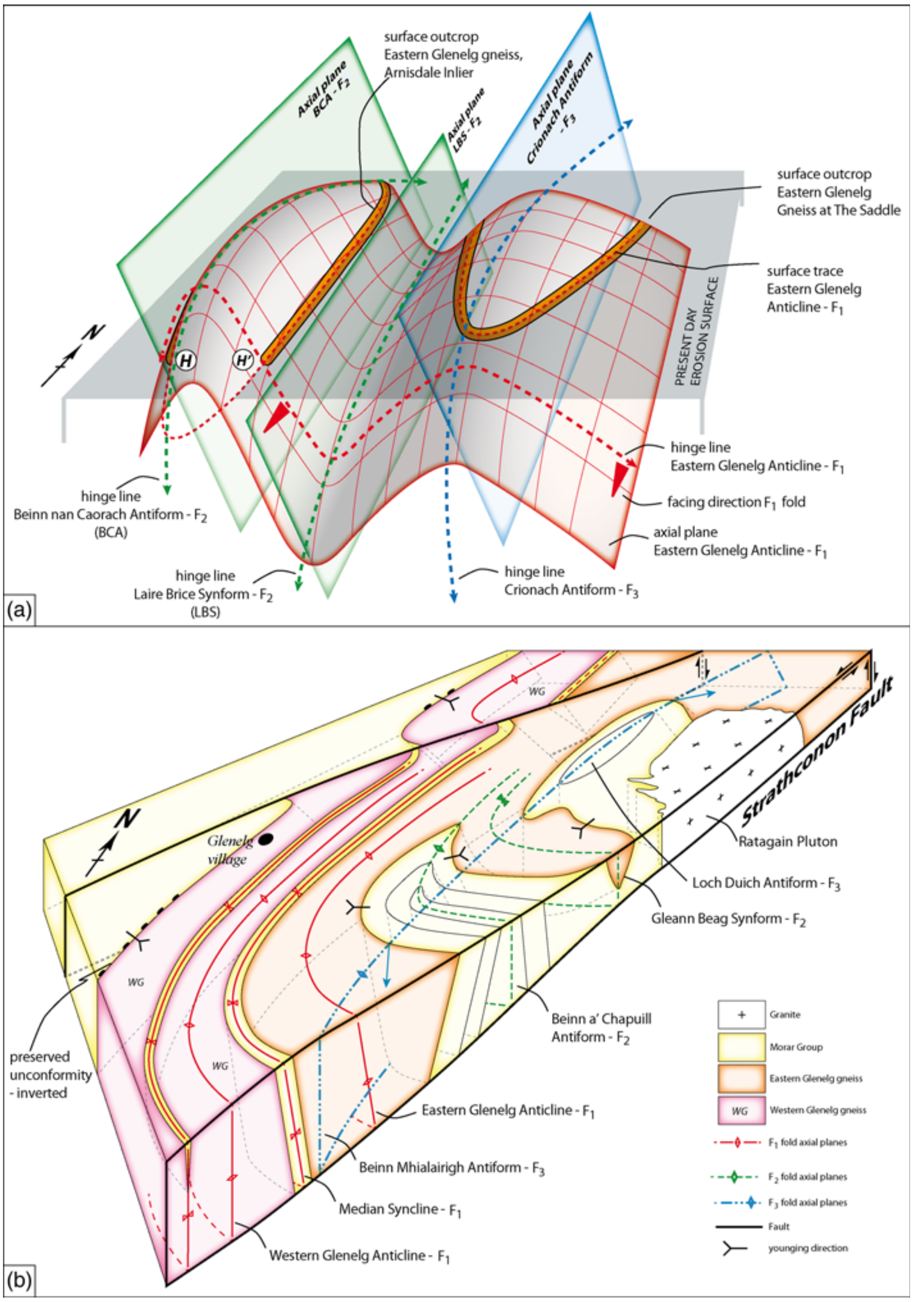

Fig. 13. Schematic block diagrams illustrating the geometry of: (**a**) F_1/F_2 overprinting around the Arnisdale Inlier, south of the Strathconon Fault; (**b**) F_2–F_3 overprinting relationships around the Beinn a' Chapuill fold, north of the Strathconon Fault. Points ***H*** and ***H′*** are discussed in the text.

a' Chapuill plunge steeply to the east as the large-scale F_2 folds face sideways towards the south. These folds are generally associated with axial plane crenulations and small-scale 'Type 2' interference patterns (Ramsay 1962). Although the major F_3 Beinn Mhialairigh fold has a reclined form over much of its trace, it becomes antiformal towards its southeastern limit before it is cut by the Strathconon Fault.

F_1/F_3 overprinting: the Saddle and Morar windows

The area between The Saddle and the Strathconon Fault has not been investigated for this study and our interpretation relies on mapping reported in Simony (1973) and May *et al.* (1993). The Saddle Window is bounded by a thin sheet of Eastern Glenelg gneiss (Simony 1973; May *et al.* 1993). The gneiss has an arcuate outcrop (Fig. 4) and structurally overlies migmatitic Morar Group rocks forming the Saddle Window. These latter rocks include mixed migmatitic and gneissose psammitic and pelitic rocks with additional, smaller gneiss slivers in cores of isoclinal folds (Simony 1973; May *et al.* 1993). The arcuate outcrop shape of the Saddle Window is primarily the result of two major F_3 antiforms, trending at right angles to each other in a box-fold-type arrangement (Simony 1973). The major north–south-trending antiform is the regional-scale F_3 Crionach Antiform (Figs 4 and 8) (Tanner 1965).

Morar Group psammite occurring structurally above The Saddle Window shows relatively low strain and locally preserves sedimentary structures (Tanner 1965). The situation is patently similar to that in the Morar Window, and the gneiss sliver in The Saddle area most likely marks the core of a large-scale F_1 isoclinal anticline, possibly with a sheared-out lower limb (see also Simony 1973). Thus, The Saddle Window results from an isoclinal F_1 fold, refolded by more upright F_3 box-folds.

South of The Saddle Window, the hinge and eastern limb of the Crionach Antiform contain a thick but broadly eastward-younging sequence of Morar Group psammite, truncated by the Sgurr Beag Thrust further east (Figs 2b and 4; Tanner 1965). South of Loch Hourn, the Crionach Antiform is responsible for the large-scale arcuate change in strike of the Morar Pelite. The well-defined base of the Morar Pelite can be reliably correlated over large distances (Tanner 1965) and the psammite of The Saddle area likely equates with the Barrisdale Psammite of Knoydart (Figs 3 and 4). The Crionach Antiform folds three major F_2 folds, in ascending order: the Laire Brice Synform, the Beinn nan Caorach Antiform (both described above), and the Sgurr nan Eugallt Synform (Tanner 1965). This latter synform (labelled erroneously as an F_3 fold by Powell 1974) is associated with a penetrative fabric, numerous congruous F_2 minor folds, and refolds a single population of isoclinal folds. The synform plunges to the SE and faces sideways to the SW; and, as a consequence, Morar Group psammites are inverted in the direct footwall of the Sgurr Beag Thrust south of Kinloch Hourn.

Discussion

Main folding south of the Strathconon Fault

The overall structure of the region is discussed first, illustrated by two cross-sections (Fig. 9) and two block diagrams (Fig. 13), to better illustrate the geometric relations in areas of non-coaxial overprinting. The structure of Knoydart is dominated by an internally coherent, right-way-up sequence of Morar Group rocks, several kilometres thick (Figs 4 and 9; see also Ramsay & Spring 1962; Krabbendam *et al.* 2014). This sequence is deformed by large-scale west-facing and west-vergent, broadly co-axial F_2 and F_3 folds (e.g. Morar and Loch Hourn antiforms) that trend north–south (Krabbendam *et al.* 2014). This whole package overlies a thin sheet of Western Glenelg gneiss, interpreted as marking the hinge of a regional-scale isoclinal F_1 fold. The lower limb of this F_1 fold comprises migmatitic Morar Group, exposed in the Morar Window, and most likely underlies most of Knoydart.

Further east, the Morar sequence is deformed by a regional-scale, west-vergent F_3 fold pair (Beinn Sgritheall

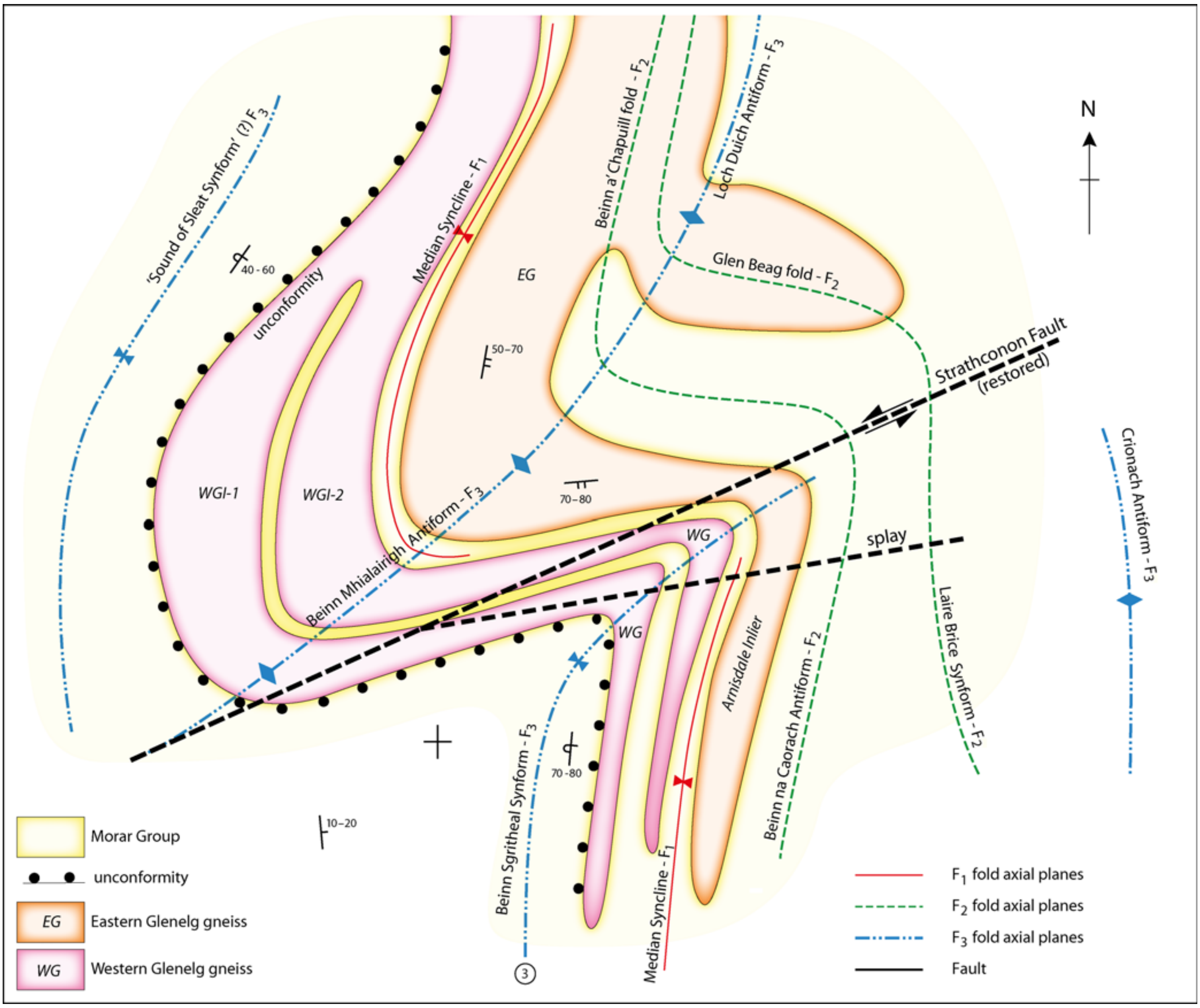

Fig. 14. Schematic reconstruction prior to movement along the Strathconon Fault; see discussion in text.

Synform and Crionach Antiform, Fig. 9). The steep common limb of these F_3 folds contains two isoclinal F_1 fold cores of Western Glenelg gneiss (WGI-1 and WGI-2) and the Arnisdale Inlier (Eastern Glenelg gneiss). The Western Glenelg gneiss in the F_1 fold cores are thought to connect to the Western Glenelg gneiss in the Morar Window, via the small outcrops of Western Glenelg gneiss along Loch Hourn and at Slisneach (Figs 4 and 9). Prior to F_3 folding, the gneiss-cored F_1 isoclines in Glen Arnisdale were positioned structurally *below* the right-way-up sequence of Knoydart, and the Glen Arnisdale Inlier would be structurally *below* F_1 folds cored by Western Glenelg gneiss (Fig. 9).

Still further east, the Eastern Glenelg gneiss of the Saddle Window is most likely connected to the Arnisdale Inlier, being folded by the F_2 Laire Brice Synform and the F_3 Crionach Antiform (Fig. 13a). The Saddle Window is thus analogous to the Morar Window, except in that it contains Eastern Glenelg gneiss in its F_1 fold core rather than Western Glenelg gneiss. In both cases, migmatitic Morar Group rocks occur in the lower limb of isoclinal F_1 folds and were brought to higher structural levels by F_3 antiforms.

Facing direction of F_1 folds

The F_1 Eastern Glenelg Anticline in the core of the arcuate Arnisdale Inlier was likely originally recumbent. It was subsequently folded by the periclinal F_2 Beinn nan Caorach Antiform, with the F_1 fold axis originally at a high angle to the F_2 fold axes (Fig. 13a). This resulted in Type 2 (mushroom- or crescent-shape, depending on the erosion level; Ramsay & Huber 1987, p. 496) fold interference patterns. In the case of the Arnisdale Inlier, a tight crescent is exposed. Despite refolding, the F_1 hinge line of a particular horizon (in this case the Eastern Glenelg gneiss/Morar contact) becomes exposed as two hinge points on the surface (***H*** and ***H′*** on Figs 8 and 13a). The connecting line between the two hinge points ***H*** and ***H′*** gives an average regional trend for the original F_1 fold axis (Ramsay & Huber 1987, p. 496). It follows that the regional trend of the F_1 Eastern Glenelg Anticline was WSW–ENE. Since older rocks (Eastern Glenelg gneiss) occur north of hinge points ***H*** and ***H′*** and younger rocks to the south, it follows that the facing of the main F_1 fold in this area is towards the SSE, very different from the regional westward facing of subsequent F_2 and F_3 folds. On a regional scale, this broad SSE F_1 facing is consistent with the presence of other thin Eastern Glenelg gneiss occurrences to the NNW of this line (e.g. around the Saddle Window), but their absence further SSE.

Links across the Strathconon Fault

The Strathconon Fault, which cuts the entire inlier, complicates the interpretation of the earlier ductile folding, as linking all structures across the fault is complex and uncertain. Broadly speaking, the following units and structures can be correlated across the fault (Figs 9 and 14):

(1) the two main inliers of Western Glenelg gneiss WGI-1 and WGI-2,
(2) the main inliers containing Eastern Glenelg gneiss (Eastern Glenelg Inlier and Arnisdale Inlier);
(3) the F_1 Median Syncline between the main Eastern and Western Glenelg inliers;

Furthermore, the main F_2 folds south of the fault, e.g. Beinn nan Caorach Antiform and Laire Brice Synform are likely equivalent to, and continuous with, the Beinn a' Chapuill and Gleann Beag folds north of the fault.

These links suggest sinistral movement along the Strathconon Fault: Figure 14 shows a schematic reconstruction, after removing the effects of the Strathconon Fault and its splays. F_3 fold links across the fault, however, are more problematic: the Beinn Sgritheall Synform has no obvious counterpart north of the Strathconon Fault, whereas the Beinn Mhialairigh Antiform has no obvious counterpart to the south. It is possible that the F_3 Beinn Mhialairigh Antiform is a continuation, or was nucleated upon, the F_2 Loch Hourn Antiform on Knoydart (Ramsay & Spring 1962). A synform complementary to the Beinn Mhialairigh Antiform likely occurs in the Sound of Sleat (a putative Sound of Sleat Synform, Fig. 4), which also explains the outcrop of Western Gneiss rocks on Sleat, north of Armadale (Fig. 2). However, the reconstruction of Figure 14 suggests that this synform is not continuous with the Beinn Sgritheall Synform. It appears that the F_3 folds either side of the Strathconon Fault were arranged in an *en echelon* fashion, with the steep southern limb of the Beinn Mhialairigh Antiform functioning as an east–west-trending linking structure. Unfortunately the terminations of the Beinn Mhialairigh Antiform to the SW and the Beinn Sgritheall Synform to the NE are not exposed or have been eroded, respectively, so that the exact *en echelon* arrangement remains uncertain. We note that the *en echelon* offset arrangement broadly coincides with the change from thick gneiss inliers north of the Strathconon Fault to thin gneiss inliers south of the fault.

Regardless of the manner in which the F_3 folds are linked across the Strathconon Fault, the critical constraint is that the widest outcrops of Glenelg gneiss occur in steep overturned limbs of a set of regional-scale west-vergent F_3 folds, which certainly include the Loch Duich Antiform and the Beinn Sgritheall Synform (Fig. 8). From here it follows that many field relations involving Glenelg Gneiss (e.g. unconformities, D_1 folds and shear zones) were inverted by F_3 folding, which is critical to the reconstruction of the earlier evolution of the region.

F_3 folding; the original disposition of the F_1 folds and position of migmatite

South of the Strathconon Fault, migmatitic Morar invariably occurs structurally beneath basement gneiss in F_1 fold cores. The question is whether this is also the case north of the fault. The F_3 Loch Duich Antiform is a key structure north of the Strathconon Fault (Figs 4 and 9). Most of the outcrop of Glenelg gneiss north of the Strathconon Fault occurs in the steep western limb of this antiform, so that the original F_1/F_2 geometry has been overturned. Thus, prior to F_3 folding, the unconformable Morar Group–Western Glenelg Inlier contact along the Sound of Sleat was right-way-up. Conversely, before F_3 folding, the strongly mylonitic eastern contact (the Inverinate Shear Zone) of the Eastern Glenelg gneiss with the

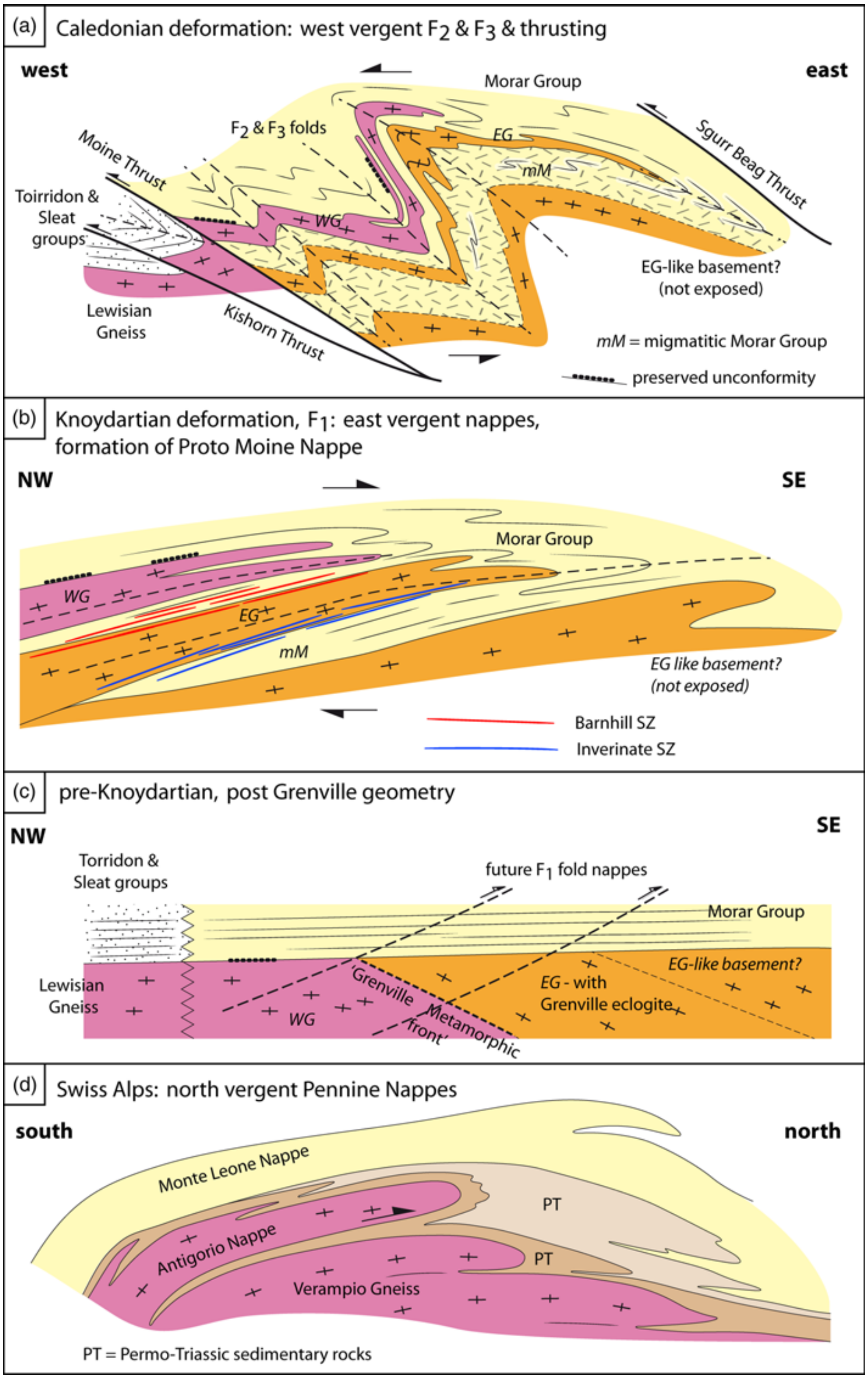

Fig. 15. Schematic structural evolution of the Glenelg Inlier. (**a**) Schematic cross-section showing west-vergent Caledonian F_2 and F_3 folding. (**b**) Cross-section reconstructed prior to F_2 and F_3 folding: Knoydartian F_1 Proto-Moine Nappe development. (**c**) Suggested pre-Knoydartian/post-Grenville geometry. (**d**) Schematic cross-section across Pennine Nappes, Swiss Alps, after Steck (2008). See text for further discussion. EG, Eastern Glenelg gneiss; EG, Eastern Glenelg gneiss; SZ, shear zone; WG, Western Glenelg gneiss.

migmatitic Morar Group was overturned and positioned structurally *below* the Glenelg Inlier. This latter point may not have been appreciated in previous studies (e.g. Temperley & Windley 1997; Storey *et al.* 2004), but is important for the interpretation of this shear zone.

It follows that the flat limbs of the major F_3 folds are most representative of the original pre-D_3 orientation of D_1 structures. The migmatitic Morar Group between the Eastern Glenelg Inlier and the Ratagain Pluton originally occurred in the lower limb of a recumbent F_1 fold, just like in Knoydart and in the Saddle Window.

The ground NE of Loch Duich was not examined during this study and it is therefore uncertain how the Loch Duich Antiform may be traced across the loch. Nevertheless, SW of Loch Duich (at ***B*** on Figs 4 and 9), in the eastern flat limb of the Loch Duich Antiform, the Eastern Glenelg Inlier–Morar contact is inverted, whereas the same contact NE of Loch Duich (***C*** on Figs 4 and 9) is right-way-up (Clifford 1957; May *et al.* 1993; Barber 2011). This suggests (1) that these two contacts are at very different structural levels and do not represent a single shear zone (cf. Storey 2008*a*, *b*) and (2) the likely presence of an early, probable F_1 isoclinal fold along Loch Duich and in the volume now occupied by the Ratagain

Pluton. This latter putative F_1 fold may connect with the F_1 fold core of the Saddle Window, although this would be difficult to prove.

Overall, a geometry emerges of a composite, recumbent F_1 fold nappe complex with at least three distinct F_1 folds: the Western Glenelg Anticline, the Median Syncline and the Eastern Glenelg Anticline (Figs 4, 8 and 9). The upper limb of this nappe complex comprises Morar Group with a coherent stratigraphy and preserved sedimentary structures, whereas the lower limb, at the structurally deepest levels, is partially migmatitic.

Role of shear zones

As described above, shear zones occur between the main Western and Eastern Glenelg inliers along the Median Syncline (Barnhill Shear Zone) and to the east of the Eastern Glenelg Inlier (Inverinate Shear Zone) (Figs 6d and 9). Temperley & Windley (1997) reported top-to-east shear sense indicators in the Inverinate Shear Zone, followed by top-to-the west shearing; Storey *et al.* (2004) report top-to-the west shear sense indicators in the Barnhill Shear Zone. However, the interpretation of these shear zones is controversial (Sanders *et al.* 1997; Temperley & Windley 1997; Storey *et al.* 2004). Temperley & Windley (1997) interpreted the top-to-the east shearing as extensional, associated with the collapse of the Grenville Orogen and responsible for the exhumation of the eclogite-facies Eastern Glenelg gneisses. However, dating of eclogite exhumation of the Eastern Glenelg Gneiss at *c.* 995 ± 8 Ma (Brewer *et al.* 2003) and detrital zircon dating of the Morar Group metasedimentary rocks, with the youngest zircons between 1070 and 1010 Ma (Friend *et al.* 2003; Kirkland *et al.* 2008; Cawood *et al.* 2015), invalidate this interpretation. Morar Group deposition evidently occurred *after* exhumation (e.g. Krabbendam *et al.* 2017) and cannot have formed the direct overburden to the eclogite: the eclogite-bearing Eastern Glenelg gneiss must have been exhumed from beneath an unidentified overburden. Nevertheless, at some time prior to D_1–D_3 deformation, a major shear zone may well have separated eclogite-bearing Eastern Glenelg gneiss from basement gneisses with lower pressure assemblages. Within the Grenville Province in eastern Canada, shear zones with both thrust and extensional movement invariably separate high-pressure gneiss units from low–medium-pressure gneiss (e.g. Rivers 2008). It is possible that the Barnhill Shear Zone represents the (intensely reworked) remnants of such a late-Grenvillian shear zone and, in effect, marks the position of a Grenville (Metamorphic) Front, albeit reworked by post-Grenvillian deformation.

The early history as outlined above, however, is less likely for the Inverinate Shear Zone, situated further east and well away from the Western Gneiss Inlier. What is clear is that both shear zones are folded around, and thus predate, the major F_2 and F_3 folds (Fig. 9). The structural analysis presented here shows that during D_1 folding the Inverinate Shear Zone had gneiss in its hanging wall and migmatitic Morar Group in its footwall (Fig. 15b; cf. Temperley & Windley 1997; Storey *et al.* 2004). The parallelism of the shear zones with the traces of F_1 isoclinal folds suggest that the shear zones operated as high strain zones along sheared-out limbs during the development of the very attenuated isoclinal F_1 folds (Fig. 15b); in other words, they are (in part) syn-D_1 shear zones. The Barnhill Shear Zone may have reactivated an earlier Grenvillian shear zone, whereas the Inverinate Shear Zone in the east operated along the Morar Group/Eastern Glenelg contact, in the overturned long limb of an isoclinal F_1 fold. The reconstruction in Figure 15b shows a large-scale recumbent nappe complex, cored by gneiss, with more intense shearing at its base than at its top.

A kinematic model for the Proto-Moine Nappe

The overall model of the evolution of the Glenelg Inlier is proposed as follows (Fig. 15). At some point after Grenvillian orogenesis, a basement partially comprising Grenvillian high-pressure gneisses became covered by siliciclastic sediments of the Sleat, Torridon and Morar Group (Fig. 15c) as part of the Grenville foreland basins (see Krabbendam *et al.* 2017). We surmise that the Eastern Glenelg gneiss with Grenvillian eclogite was structurally positioned over Western Glenelg gneiss and Lewisian gneiss further NW, and that more Eastern Glenelg-like gneiss occurred further SE, but this geometry is very poorly constrained. During F_1, presumably during the Knoydartian (see below), a recumbent fold nappe complex developed – termed here the Proto-Moine Nappe. The uppermost limb of the fold nappe comprises a thick slab of relatively low-strain, right-way-up Morar Group with a well-preserved stratigraphy, whereas the lower limb consists of intensely deformed rocks including shear zones. The facing of the fold nappe, as far as the data allow, is towards the SSE, but it is uncertain whether or not this represented the transport direction. However, it appears that the Western Glenelg gneiss was thrust over Eastern Glenelg gneiss (Fig. 15b), supporting an overall southern or eastern transport direction. Narrow, tight infolds of metasedimentary rocks and gneiss are part of the overall structure, similar to that of the Pennine nappes of the Swiss Alps (Fig. 15d), where thin slivers of Permo-Triassic sedimentary rocks are infolded in between the Antigoria and Verampio basement nappes (e.g. Steck 2008). By analogy with these basement-carrying nappes in the Alps, it is likely that more basement occurs at structurally lower levels (Fig. 15b); but this is not exposed in the Scottish Highlands.

The F_1 Proto-Moine Nappe is of significant size: the outcrop and reliable subcrop of the gneiss core and migmatitic lower limb extend over an area of *c.* 20 × 55 km, stretching from southern Morar to Attadale (Fig. 2b). If the upper limb, comprising right-way-up Morar Group, is coherent across Kintail towards the Fannichs and further north (which is by no means certain), then the upper limb would include the entire outcrop of Morar Group rocks stretching from Ardnamurchan in the south to the north coast in Sutherland.

The F_1 fold nappe was subsequently refolded by F_2 and F_3 folds (Fig. 15a), presumably during the Caledonian (see below). Peak-metamorphism occurred broadly during F_2, with migmatization occurring consistently at the deepest structural levels. In areas of medium-strain, F_2 and F_3 folds are coaxial and face and verge towards the west; whereas in areas of high D_2 strain, F_2 fold axes rotated towards parallelism with the WNW transport direction. The F_3 anticlinal structures have brought the lower limb of the F_1 fold nappe to higher levels.

The Moine Thrust cuts obliquely across F_2 and F_3 folds (e.g. the Morar Antiform and possibly the Beinn Sgritheall Synform), showing that it formed considerably later than the F_1 folds. Thus the development of the Proto-Moine Nappe significantly predates the Moine Thrust, and the Moine Nappe was already a nappe before the Caledonian Moine Thrust developed.

Timing: linking the structural evolution to the geochronological record

The geochronology of the complex series of tectonic events in the Northern Highlands remains controversial, with a number of unresolved issues (e.g. Oliver 2002; Bird *et al.* 2013; Cawood *et al.* 2015; Dewey *et al.* 2015). The age of Morar Group deposition is constrained by their youngest detrital zircons from the Morar Group (1070 – 1010 Ma; Friend *et al.* 2003; Kirkland *et al.* 2008; Cawood *et al.* 2015) and the *c.* 870 Ma bimodal magmatism affecting the structurally higher and younger Loch Eil and Glenfinnan groups (Friend *et al.* 1997; Millar 1999; Cawood *et al.* 2015). Subsequently, the Moine Supergroup was subjected to several orogenic events: Knoydartian, Grampian and Scandian (e.g. Strachan *et al.* 2010; Cawood *et al.* 2015).

In the Glenelg/Knoydart area, geochronological ages associated with Knoydartian tectonothermal activity suggest two phases. The first ('Knoydart I') phase occurred around 830 – 820 Ma, recorded by U–Pb monazite ages on pegmatite and Sm/Nd ages from garnet (Rogers *et al.* 1998; Vance *et al.* 1998). The second ('Knoydart II') phase occurred around 790 – 780 Ma, based on U–Pb monazite and U–Pb zircon ages on pegmatite (Rogers *et al.* 1998; Cawood *et al.* 2015) and garnet Sm/Nd ages from the Ladhar Bheinn pelite (Vance *et al.* 1998). A third, even later phase of Knoydartian activity around 740 – 720 Ma has thus far only been recorded in the Sgurr Beag Nappe or in its direct footwall (e.g. Van Breemen *et al.* 1974; Tanner & Evans 2003; Cutts *et al.* 2010).

As to Caledonian activity, the separation between Grampian (*c.* 470 – 460 Ma; early Ordovician) and Scandian (440 – 420 Ma; Silurian) affecting the northern Scottish Caledonides has recently been challenged by emerging evidence for late-Ordovician *c.* 460 – 440 Ma activity, suggesting either a distinct 'Grampian-II' phase or a more protracted, continuous Caledonian Orogeny without distinct phases (Bird *et al.* 2013; Cawood *et al.* 2015). Garnets from Morar Group pelitic rocks close to the Western Glenelg gneiss have yielded *c.* 460 – 450 Ma ages (Lu–Hf and Sm–Nd; Bird *et al.* 2013); and syn-late D_2 pegmatite from Rubha Camas na Cailinn has yielded a U–Pb zircon age of 446 ± 5 Ma (Cawood *et al.* 2015). Storey *et al.* (2004) obtained a U–Pb (titanite) 437 ± 6 Ma age from a post-D_2/pre-D_3 pegmatite intruding the Eastern Glenelg gneiss. All these ages are from structural levels close to or within the Glenelg basement gneisses. Overall we tentatively interpret these dates to indicate that peak-metamorphic D_2 in the southern Moine Nappe activity occurred during the period 460 – 440 Ma ('Grampian II'), i.e. later than the 470 – 460 Ma Grampian activity within the overlying Sgurr Beag and Naver nappes (e.g. Kinny *et al.* 2003; Bird *et al.* 2013), but earlier than the 440 – 420 Ma Scandian deformation that appears to dominate the northern Moine Nappe (Kinny *et al.* 2003; Alsop *et al.* 2010; but see Bird *et al.* 2013). D_3 folding, occurring at cooler conditions, is likely to be Scandian and was closely followed by the development of the Moine Thrust Zone in early-mid-Silurian (Freeman *et al.* 1998; Goodenough *et al.* 2011).

Dates higher up in the structural sequence are rare, but Cawood *et al.* (2015) report a muscovite *c.* 460 Ma $^{40}Ar/^{39}Ar$ age from the Ladhar Bheinn/Morar pelite from the synclinal core of the Beinn Sgritheal Synform, interpreted to indicate cooling from a Grampian event, but also suggesting an absence of resetting high up in the structural pile during the 'Grampian II' phase.

With the currently available geochronological data, we propose the west-verging D_2 and D_3 folds developed during the Caledonian Orogeny, with D_2 commencing in the Late Ordovician ('Grampian II') and D_3 continuing into the Silurian (Scandian). This suggests that the SSE-facing Proto-Moine Nappe developed during the Knoydartian Orogeny.

Conclusions

(1) Basement inliers in the southern Moine Nappe can be divided into two distinct types: *Western Glenelg gneiss*, dominated by orthogneiss; and *Eastern Glenelg gneiss*, containing significant paragneiss and having experienced Grenvillian eclogite-facies metamorphism. The Eastern Glenelg gneiss also occurs in The Saddle Inlier and possibly the An Cruachan Inlier, so extending its known outcrop further east.

(2) The Western and Eastern Glenelg gneisses occur in several inliers arranged in the cores of a number of stacked anticlinal F_1 isoclines that together form a recumbent D_1 fold nappe complex, termed here the *Proto-Moine Nappe*, a precursor to the Caledonian Moine Nappe Complex.

(3) The upper limb of the Proto-Moine Nappe comprises a slab >5 km thick of medium-strain Morar Group metasedimentary rocks with a coherent stratigraphy.

(4) By restoring the major F_2 and F_3 folds it can be shown that strongly deformed, migmatitic Morar Group rocks occur in the lower limb of the F_1 fold nappe. Shear zones are interpreted to have resulted from intense shearing along the long limbs of isoclinal F_1 folds. These shear zones occur between F_1 isoclinal folds and, after F_2/F_3 restoration, structurally below the F_1 fold-nappe complex. No unequivocal structures related to the Late-Grenvillian exhumation of the Eastern Glenelg gneiss are preserved, although it is possible that some shear zones reactivated pre-existing, late-Grenvillian shear zones.

(5) Restoration of F_1/F_2 fold interference patterns show that F_1 folds were SSE facing, consistent with the spatial distribution of gneiss inliers of Eastern Glenelg gneiss affinity. It is uncertain whether or not this represented the transport direction of the F_1 fold nappe.

(6) Prior to the onset of F_2 and F_3 folding, the Eastern Glenelg gneiss inliers within the F_1 Proto-Moine

Nappe complex occurred at structurally lower levels than the inliers of Western Glenelg gneiss. Western Glenelg gneiss bodies were likely thrust over Eastern Glenelg gneiss bodies.

(7) The core and lower limb of the F_1 Proto-Moine Nappe is proven over an area (outcrop and reliable subcrop) of *c.* 20 × 50 km. If the upper limb is coherent north of Kintail, and stretches further north, the nappe would cover an area of some 70 × 220 km.

(8) The F_1 Proto-Moine Nappe is refolded by F_2 and F_3 folds; these are typically west-facing and west-vergent over much of the area. In areas of high F_2/F_3 strain, however, F_2 fold axes have been rotated towards a dominant WNW transport direction; in these areas F_2/F_3 fold interference is non-coaxial.

(9) The best fit with the available geochronological data suggests that the SSE-facing F_1 Proto-Moine Nappe developed during Knoydartian (mid-Neoproterozoic) orogenesis, whilst the west-facing and west-vergent F_2 and F_3 folds affecting the Glenelg Inlier are Caledonian in age, with F_2 probably developed during the Late-Ordovician 'Grampian II' phase and D_3 during Silurian Scandian orogenesis. The F_1 Proto-Moine Nappe thus developed *well before* the Scandian Moine Thrust.

(10) The Moine Nappe is a poly-orogenic feature: it originally formed as the Proto-Moine Nappe during Knoydartian orogenesis and was refolded and transported again during Caledonian orogenesis in a different direction.

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