The Palaeoproterozoic anatomy of the Lewisian Complex, NW Scotland: evidence for two ‘Laxfordian’ tectonothermal cycles

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Abstract: A new structural examination of Palaeoproterozoic high-P granulites on South Harris, NW Scotland, when integrated with previous geochronological, structural and metamorphic studies on key areas of the Lewisian Complex, suggests the existence of two distinct tectonothermal cycles within the Palaeoproterozoic ‘Laxfordian Event’, which on South Harris are separated by a >100 myr hiatus in deformation. The older cycle, from c. 1.91 to 1.85 Ga, records the development of an active continental margin on the Archaean gneisses that dominate the Complex, and the subsequent onset of continent–continent collision; this represents the continuation of the Nagssugtoqidian orogen of Greenland. Evidence for this is concentrated in allochthonous slivers of the former active continental margin displaced during the younger cycle. The younger cycle, around 1.75–1.65 Ga, began with thrust-related crustal thickening that initiated regionally extensive amphibolite-facies metamorphism and ductile deformation, which dominates the preserved ‘Laxfordian’ deformation history. This may be the peripheral expression of the accretion of the Malin block to the SW of the Lewisian, and represents the lateral continuation of the Labradorian–Ketilidian orogen of North America.

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The Lewisian Complex of NW Scotland comprises Archaean tonalite–trondhjemite–granodiorite (TTG) orthogneisses with subordinate Palaeoproterozoic additions. The traditional view of the Complex (Sutton & Watson 1950) is of a single entity that underwent an early phase of high-grade tectonothermal activity (the ‘Scourian’ event), which, following the emplacement of Palaeoproterozoic mafic and ultramafic dykes (Scourie dykes), was heterogeneously reworked during a younger lower grade (‘Laxfordian’) tectonothermal event. Within the mainland Lewisian, the Scourian was further subdivided into an older high-grade event (Badcallian) and a younger lower grade event (Inverian), the latter being similar in style to the Laxfordian (e.g. Evans & Lambert 1974).

More recently, based largely on apparent regional variation in metamorphic signatures and protolith ages, the Complex has been reinterpreted in terms of many discrete terranes, amalgamated during the Proterozoic along supposedly terrane-bounding shear zones (e.g. Friend & Kinny 2001; Kinny et al. 2005). However, because of the spatial variability inherent in metamorphism, igneous intrusion, etc., identifying regional differences does not, in itself, discriminate between telescoping of lateral and vertical heterogeneities within a single piece of crust and the accretion of separate crustal blocks. Making this distinction requires an understanding of the underlying controls on the observed variations. Given the likely original heterogeneity, protracted history and the presence of kilometre-scale poly-phase shear zones (e.g. Fettes et al. 1992; Wheeler 2007; Park 2010) it is almost inevitable that some of the apparent terrane structure is an artefact of dissection or telescoping of heterogeneous crust, rather than true terrane accretion. Nevertheless, the recognition of early Laxfordian volcanic arc and accreted oceanic material and high-P metamorphism does strongly suggest the presence of at least one Laxfordian suture (e.g. Baba 1998; Park et al. 2001; Mason et al. 2004b). The wealth of geochronological data now available, however, demonstrates that the ‘Laxfordian’ signature is complex and protracted, and probably encompasses more than one tectonothermal event (e.g. Park 2005; Love et al. 2010; Goodenough et al. 2013).

Comparison with the Palaeoproterozoic Nagssugtoqidian orogen of SE Greenland, with which the Lewisian is correlated (e.g. Buchan et al. 2000), reveals additional difficulties. Both cratonic blocks flanking the Nagssugtoqidian and their reworked equivalents comprise 2950–2630 Ma protoliths of varied age that in places record granulite-facies metamorphism, probably around 2750–2720 Ma (summarized by Kolb 2014). The Lewisian appears broadly similar and could be matched to either block, with comparable heterogeneous protolith ages and, in a number of locations, a c. 2760–2700 Ma signature in the U–Pb system, linked by some workers to high-grade metamorphism (Corfu et al. 1994, 1998; Zhu et al. 1997; Friend & Kinny 2001; Whitehouse & Bridgwater 2001; Mason et al. 2004b; Kinny et al. 2005; Kelly et al. 2008; Love et al. 2010; Crowley et al. 2014). Any Laxfordian suture is, therefore, unlikely to be indicated by an unambiguous change in the Archaean gneisses.

A potentially better way of examining the Palaeoproterozoic evolution of the Lewisian is from the perspective of Laxfordian orogenic processes and how these control such features as the observed architecture of the Complex and the distribution of tectonothermal signatures. This offers the possibility of correlating events in different areas at the process level, for example by matching collisional signatures or deformational regime, circumventing much of the lateral and vertical heterogeneity within the Complex.

The identification of collisional signatures (e.g. deformation associated with the termination of subduction-related magmatism and high-P metamorphism) also offers the possibility of discriminating between true terrane accretion or suturing and younger telescoping or dissection, even where the Archaean blocks involved are not distinct from each other. A more process-based view also allows some consideration of the feedbacks that inevitably must have existed between different processes or events and how these could potentially explain some of the local complexities of the geology.
Understanding the sequence of events that led to the current configuration of the Lewisian has wider implications for palaeo-geographical reconstructions and application of terrane theory elsewhere. For example, the ‘terrane’ configuration for the Lewisian (e.g. Kinny et al. 2005) is radically different from the relatively simple two-plate models for the Nagsugtoaqidian of SE Greenland (e.g. Kolb 2014), yet for the correlation with the Lewisian to stand, the assembly of one area has to be geometrically compatible with that of the other area.

Previous hypotheses for the assembly or structural modification of the Lewisian are numerous and often contradictory, ranging from predominantly transient juxtaposition of ‘terranes’ (e.g. Kinny et al. 2005), through reassembly of previously contiguous blocks (e.g. Mason & Brewer 2004; Kelly et al. 2008), to strike-slip shearing (Park 2010) or subduction accretion and continental collision (e.g. Baba 1999a; Park et al. 2001). Few studies are truly regionally integrated in the sense that they consider in parallel such factors as the geometric requirements of suturing or terrane accretion, the genesis and incorporation of juvenile igneous rocks, the driving mechanisms of metamorphism and deformation, and burial mechanisms of supracrustal rocks across the whole Complex. It is not, therefore, surprising that so many conflicting models, each representing an incomplete description of the same geology, should exist.

The primary objective of the present work is to move towards a more systematic, regionally integrated, process-oriented interpretation of the Laxfordian based on the working hypothesis (justified below) that the Lewisian was assembled and then reworked and dissected in two successive ‘Laxfordian’ tectonothermal cycles (broadly following Park 2005). To this end, a new structural study of a belt of high-P granite-facies gneisses on southwestern South Harris (Fig. 1) is presented and integrated with published work on key areas of the Outer Hebrides (e.g. Baba et al. 2012; Mason 2012; and references therein) and the mainland (e.g. Park 2010; Goodenough et al. 2013; and references therein). The rocks of South Harris have been chosen as the starting point because they include Laxfordian juvenile arc rocks metamorphosed at high pressure (e.g. Baba 1998), now separated from the dominant Archaean TTG gneisses by a major poly-phase shear zone (Mason 2012). South Harris is thus a likely archive for the processes that occurred prior to, during and subsequent to Laxfordian suturing.

Regional geology
Outer Hebrides

The Outer Hebrides largely comprise late Archaean TTG orthogneisses intruded by mafic dykes prior to extensive Laxfordian reworking in the Palaeoproterozoic (Fig. 1; Table 1). The dominant dykes (Outer Hebrides dykes) are chemically distinct from the mainland Scourie dykes, with a tentative 2.1 Ga age (Mason & Brewer 2004), although Davies & Heaman (2014) have also dated c. 2.4 Ga Scourie dyke equivalents on Lewis (Fig. 1). Laxfordian amphibolite-facies metamorphism culminated around 1675 Ma, locally with migmatization and granite intrusion, forming the Injection Complex of western Harris and Lewis (Fettes et al. 1992; Friend & Kinny 2001; Kelly et al. 2008). Within the TTG gneisses are belts of distinctive metagranitic–metasedimentary assemblages, including the Ness Assemblage on northern Lewis and the Harris Granulite Belt and Langavat Belt on South Harris (Fig. 1). The southeastern edge of the Outer Hebrides is affected by a brittle–ductile fault system (Outer Hebrides Fault Zone) that post-dates the main Laxfordian deformation and metamorphism (Fettes et al. 1992; Imber et al. 2002).

The Ness Assemblage and Harris Granulite Belt contain the relatively rare association of metadiorite and meta-anorthosite and record c. 1910–1870 Ma metamorphism (Cliff et al. 1983, 1998; Fettes et al. 1992; Timmerman et al. 2001; Whitehouse & Bridgewater 2001; Baba et al. 2012; Mason 2012). In the case of the Harris Granulite Belt, this metamorphism attained high-P granulite facies (Baba 1998); the 1910–1870 Ma metamorphism has not been recognized in the adjacent Archaean gneisses or Langavat Belt (Cliff et al. 1983, 1998; Friend & Kinny 2001; Whitehouse & Bridgewater 2001; Mason et al. 2004a; Kelly et al. 2008). Both the Ness Assemblage and Harris Granulite Belt are structurally above the Archaean gneisses (Mason 2012). The Langavat Belt is sandwiched along the interface between the Harris Granulite Belt and the Archaean gneisses to the NE (Mason et al. 2004a). The Langavat Belt contains metasedimentary rocks, which, based on their detrital zircon signature, are derived from the dominant Archaean TTG gneisses, but are now tectonically interleaved with these gneisses and metamorphosed under (Laxfordian) amphibolite-facies conditions (Mason et al. 2004a; Mason 2012). This is significant because it requires that at least some of the TTG gneisses were exhumed to the surface to source sediment, prior to the Laxfordian, and then reburied to mid-crustal depth along with the sedimentary rocks they sourced.

Following Mason (2012), the architecture of the Outer Hebrides is considered in terms of a major (but now cryptic owing to later ductile deformation; see below) thrust sheet. The Harris Granulite Belt and Ness Assemblage are the remnants of this, with the Archaean gneisses beneath. The Langavat Belt represents a former duplex or imbricate zone beneath the thrust sheet (Mason et al. 2004a; Mason 2012; Fig. 1). The thrust sheet carried with it the distinctive anorthosite–diorite assemblage and the c. 1910–1870 Ma metamorphic signature that characterizes the Ness Assemblage and Harris Granulite Belt; it simultaneously buried the Archaean gneisses and Langavat Belt, establishing the prerequisite mid-crustal conditions for the widespread (~200 km across strike; Fettes et al. 1992) amphibolite-facies metamorphism and ductile deformation that dominates the observed geology. The thrust sheet has now been dissected and contorted by younger ductile deformation and (not unsurprising, given the present mid-crustal exposure) largely removed by erosion, exposing the underlying Archaean gneisses that form most of the Outer Hebrides (Mason 2012; Table 1). This thrust sheet is part of a major structure, here termed the Gairloch–Langavat Thrust System, which is also considered to affect the mainland (see below); this should not be confused with the Langavat Shear Zone and Gairloch Shear Zone (see below and Table 1), which are distinct younger structures.

Following the inferred thrusting, perhaps progressively, the most important of the younger deformation phases was associated with top-to-the-NW or -WNW shearing and gravitational flattening, which led to near-ubiquitous and often intense deformation of the Archaean gneisses and the Outer Hebrides dykes cutting them (equivalent to F 2 of the traditional regional deformation history; e.g. Coward et al. 1970; Graham & Coward 1973; Fettes et al. 1992; Table 1), as well as the formation of large-scale shear zones. The most significant of these is probably the Langavat Shear Zone, which is NW–SE striking, steeply SW dipping and approximately coincident with the Langavat Belt, although it extends somewhat into the adjacent rocks (Mason et al. 2004a). The Langavat Shear Zone downthrows the thrust sheet to the SW leaving a klippe, the Harris Granulite Belt, preserved in the hanging wall, consistent with the geophysical evidence that the Harris Granulite Belt has no lower crustal presence and extends only to c. 7 km depth (Westbrook 1974). The same extensional movement probably also helped control the location of the Injection Complex through foot-wall uplift (Mason 2012). The Ness Assemblage is preserved as a second klippe by younger down-warping (Mason 2012).

Based on the similarity of the TTG gneisses and Outer Hebrides dykes on either side of the Harris Granulite Belt, Mason & Brewer (2004) and Kelly et al. (2008) interpreted the Outer Hebrides in
terms of break-up and reassembly of a single piece of TTG gneiss, with the Harris Granulite Belt forming a magmatic arc trapped during reassembly. This relies on the argument that the Harris Granulite Belt separates the TTG gneisses of Harris–Lewis and the Uists (Fig. 1). However, kinematic evidence suggests that the Langavat Shear Zone was initiated in oblique extension (part of the top-to-the-WNW shearing mentioned above; Mason 2012), indicating insertion of the Harris Granulite Belt into the TTG gneisses from above, rather than it being trapped between two TTG blocks. This is consistent with the geophysical evidence that the Harris Granulite Belt has no lower crustal presence, and also negates any argument that the TTG gneisses of Harris–Lewis and the Uists have ever been separated since the Outer Hebrides dykes were emplaced (Fig. 1; Table 1). Tectonic insertion of the Harris Granulite Belt from above would require prior transport of the Harris Granulite Belt over the TTG gneisses and Langavat Belt on a structure that predates the Langavat Shear Zone (i.e. transport by the Gairloch–Langavat Thrust System). Prograde metamorphism during the initial extensional movement of the Langavat Shear Zone corroborates prior burial beneath the Harris Granulite Belt, as the Langavat Belt metasediments cannot easily have been buried to mid-crustal depth to drive metamorphism, and interleaved with the TTG gneisses, during extension (Mason 2012). This elimination of the Langavat Shear Zone as the primary transport mechanism for the Harris Granulite Belt leads almost inexorably to the correlation with the Ness Assemblage, as the Harris Granulite Belt must have been transported from somewhere and the Ness Assemblage is the only candidate for a lateral continuation of the Harris Granulite Belt on the Outer Hebrides. The Ness Assemblage itself presents a similar geometric constraint in that it is essentially flat-lying (although with some later modification; Mason 2012) and above the TTG gneisses to the SW (Fig. 1). Transform or transcurrent terrane accretion (Friend & Kinny 2001; Kinny et al. 2005) on upright structures analogous to the Langavat Shear Zone provides an unsatisfactory explanation for such flat-lying structures.

**Mainland**

The mainland also comprises mainly TTG gneisses, cut by Scourie dykes. The central region of the mainland comprises relatively
unmodified Archaean granulite-facies gneisses in which Laxfordian reworking is mostly restricted to discrete shear zones (e.g. Coward 1990). The northern and southern regions of the mainland show generally more pervasive Laxfordian reworking (Wheeler 2007; Goodenough et al. 2010). Proterozoic supracrustal rocks and associated intrusions occur in the Loch Maree Group (Fig. 1), which approximately marks the boundary between the central and southern region (Park et al. 2001).

The central part of the mainland is possibly at a higher structural level than is presented on the Outer Hebrides and southern mainland, and, consequently, escaped pervasive Laxfordian reworking (Park 2005; Mason 2012). Correspondingly, the Loch Maree Group sits structurally beneath the central region gneisses but above the generally more pervasively reworked southern region, and was overthrust by the central region (Park et al. 2001). The Loch Maree Group marks the lateral continuation of the thrust system (Gairloch–Langavat Thrust System) carrying the Harris Granulite Belt and Ness Assemblage (Mason 2012; Fig. 1). The Loch Maree Group is associated with a sliver of Proterozoic granulite-facies gneisses (Iallaig Gneiss; Park 2002; Love et al. 2010) and falls within a gently dipping (unnamed) top-to-the-NW ductile shear zone, which has been heavily modified by younger upright folding and dextral shearing associated with the Gairloch Shear Zone (Park et al. 2001; Park 2010). The history is comparable with that of the Langavat Belt–Harris Granulite Belt on South Harris (Table 1).

Based on the SW dip of the Laxford Shear Zone (Fig. 1), the central mainland gneisses have also been considered to be above the more heavily reworked northern region, NE of the Laxford Shear Zone (e.g. Park 2005; Goodenough et al. 2010). However, geophysical surveys across the Laxford Shear Zone (Bott et al. 1972) suggest that this is a local, not regional relationship, with downthrown equivalents of the central mainland granulites continuing to the NE of the Laxford Shear Zone, concealed at a few kilometres depth (Fig. 1). This is consistent with inferred Inverian thrust-sense movement on the Laxford Shear Zone (e.g. Coward & Park 1987), which would downthrow the former granulite-facies isograd to the NE. If correct, the reworked northern mainland gneisses are at a different structural level from the reworked gneisses of the southern mainland and Outer Hebrides (Fig. 1).

*This is the regional deformation history established by Coward et al. (1970) and subsequently adopted or modified by many other workers such as Fettes et al. (1992).
| Tectonothermal History | Facies Tectonites that in many cases are derived from the 1675–1657 Ma granite pegmatites. |
|------------------------|--------------------------------------------------------------------------------------------|
| Dilution of E-W-striking HGB D1 shear zones, pegmatic emplacement | NW-plunging facies (Type 2 tectonites) |
| HGB D1 (Langavat Granulite Zone) | NW assigned | Amphase facies (Type 1 tectonites) Origin of high-P metamorphism (M1) |
| Either no equivalent overprint, or obliterated by D3 overprint | Lang D1 | Prograde amphibolite facies; significant syn-tectonic garnet growth; metamorphic peak post-Lang D2 |
| Emplacement of HGB or Langavat Belt - little or no internal deformation in HGB | Lang D1 | Sub-garnet grade; mostly obliterated by Langavat Belt Zone and prograde formation |

Table 1 (continued)

| Deformation phase | Deformation Metamorphic grade in main supracrustal rocks (leucosome group) | Fabric development phase | Source |
|-------------------|---------------------------------------------------------------------------|--------------------------|---------|
| Lang D3 | Replacive green-schist facies | NW-plunging facies, SW-dipping foliation (Type 2 tectonites) | Reactivation of the Langavat Shear Zone; top-to-the-NW oblique dextral-normal movement associated with restricted brittle–ductile deformation; reworking of the Lang D2 mylonites, development of minor 2-geometry drag folds and supercritical foliation parallel reactivated shear zones; possibility of Greenian age | Mason & Brewer (2005) |
| Dilution of E-W-striking HGB D1 shear zones, pegmatic emplacement | Lang D1 | Retractive green-schist facies | Reactivation of the Langavat Shear Zone; top-to-the-SW oblique sinistral thrust movement associated with a reactivated region of brittle–ductile deformation; localized mylonites development from granite pegmatites, with the formation of well-developed but spatially restricted sinistral shear sensa criteria and minor x-geometry drag folds; probably partly coeval with the granite pegmatites | Mason & Brewer (2005) |
| HGB D1 (Langavat Granulite Zone) | NW assigned | Amphibole facies (Type 1 tectonites) | Widespread occurrence of amphibolite-facies tectonites that delimit the Langavat Shear Zone; the Type 2 tectonites are un-annealed greenschist-facies tectonites that in many cases are derived from the 1675–1657 Ma granite pegmatites. |
| Either no equivalent overprint, or obliterated by D3 overprint | Lang D1 | Prograde amphibolite facies; significant syn-tectonic garnet growth; metamorphic peak post-Lang D2 | Initiation of the Langavat Shear Zone in oblique top-to-the-NW dextral extension; intense ductile deformation; first well-preserved fabrics in Lower Laxfordites dykes | Mason (2012) |
| Emplacement of HGB or Langavat Belt - little or no internal deformation in HGB | Lang D1 | Sub-garnet grade | Imbrication of the supracrustal rocks with Archaean tonalite: Mason (2004a), Kiglies ner; burial of the main supracrustal rocks to mid-crustal level; onset of prograde metamorphism; assembly of the Langavat Belt; initiation of the Gairloch–Langavat Thrust System | Mason (2012) |

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Levisian ‘inliers’ occur within the Caledonian orogen to the SE of the Moine Thrust, and although they are displaced and difficult to correlate in detail, they broadly match the Lewisian of the Caledonian foreland (e.g. Friend et al. 2008; Fig. 1). The most significant of these is the Glenelg–Attadale inlier, which is notable for the occurrence of c. 1750 Ma ‘Laxfordian’ eclogite-facies metamorphism within Archaean TTG gneisses (Storey et al. 2010).

Tectonothermal History

A widely distributed c. 2.73 Ga U–Pb age signature (Fig. 1) of probable metamorphic origin (Corfu et al. 1994, 1998; Zhu et al. 1997; Kinny et al. 2005; Kelly et al. 2008; Love et al. 2010; Crowley et al. 2014), ‘Inverian’ deformation that maintains a NW–SE structural grain throughout much of the mainland (Evans & Lambert 1974; Park 2002; Wheeler 2007; Goodenough et al. 2010) and the wide spread occurrence of 2.4 Ga Scourie-type dykes and (the possibly younger) Outer Hebrides dykes (Mason & Brewer 2004; Davies & Heaman 2014; Fig. 1) suggest that many of the Archaean components of the Lewisian share at least parts of their pre-Laxfordian history. Whether the Laxfordian represents the reworking of a single crustal block or the (re-)assembly of blocks with a common or comparable earlier history must be considered.

The Gairloch–Langavat Thrust System explains a significant part of the Laxfordian geology, but not the origin of the c. 1910–1870 Ma metamorphic signature in the Harris Granulite Belt and Ness Assemblage. The Harris Granulite Belt underwent high-P metamorphism (Baba 1998), but, critically, was substantially exhumed before thrusting placed it over the Langavat Belt; the main supracrustal rocks of the Langavat Belt only record relatively low-P prograde amphibolite-facies metamorphism (Mason 2012). The fact that the Harris Granulite Belt was both buried and exhumed before being placed over the Langavat Belt (and the markedly younger high-P metamorphism at Glenelg), suggests the existence of two distinct tectonohermal cycles: an early one in which the Harris Granulite Belt was buried, metamorphosed and then exhumed; a younger one in which the Harris Granulite Belt was displaced to its present...
position and the underlying rocks were buried, extensively deformed and metamorphosed. Therefore, in this work, Laxfordian chronology is divided into an Early and Late Laxfordian cycle (Table 1).

The Harris Granulite Belt

The Harris Granulite Belt predominantly constitutes c. 1.9 Ga protoliths metamorphosed between 1.91 and 1.87 Ga, shortly after formation (Cliff et al. 1983, 1998; Friend & Kinny 2001; Timmerman et al. 2001; Whitehouse & Bridgwater 2001; Mason et al. 2004b; Baba et al. 2012). The central part of the Harris Granulite Belt comprises the Leverburgh Belt, a range of psammitic to pelitic metasedimentary gneisses with minor calc-silicates and marble (Dearnley 1963). These are flanked to the NE by a c. 10 km wide body of metadiorite, and to the SW by a body of ‘metanorite’ of comparable size (Fettes et al. 1992). A c. 3 km long wedge-shaped body of meta-anorthosite occurs within the SE end of the Leverburgh Belt (Fig. 2). Numerous smaller bodies of metabasite (‘garnet metagabbro’) are also common.

Baba (1998, 1999b) recognized several stages of metamorphism. Inclusions in garnet cores in pelitic gneisses record early prograde amphibolite-facies metamorphism (M1) with stable sillimanite, which progressed to ultrahigh-temperature granulite-facies conditions of 9–11 kbar at 900–980°C. This was followed by (M2) high-P granulite-facies metamorphism with stable kyanite and conditions of 13–14 kbar at 850–900°C. This was succeeded by retrograde decompression (M3) that produced orthopyroxene–cordierite (in pelites) and orthopyroxene–plagioclase (in metabasites) coronae around M1 garnets. This was followed by localized amphibolite-facies retrogression (M4). Baba (1998, 1999b) considered that peak T preceded peak P; Hollis et al. (2006) presented additional P–T estimates supporting this.

The NE contact of the Harris Granulite Belt dips steeply SW. The Langavat Belt separates the Harris Granulite Belt from the much larger area of Archaean TTG gneisses to the NE (Figs 1 and 2). The Langavat Belt comprises metabasic, ultramafic and supracrustal rocks interleaved with Archaean tonalitic gneisses, deformed by the Langavat Shear Zone. Although superficially similar to the Harris Granulite Belt, the supracrustal rocks of the Langavat Belt have a different zircon provenance (Mason et al. 2004a) and record only peak amphibolite-facies metamorphism between 590°C at 3.7 kbar and 641°C at 6.1 kbar (Mason 2012). The Harris Granulite Belt is largely retrogressed to amphibolite facies for c. 2 km adjacent to the Langavat Belt. A major structural discontinuity occurs either between the Harris Granulite Belt and the Langavat Belt, or within the upper part of the Langavat Belt (Mason 2012).

Little is known about the SW contact of the Harris Granulite Belt, which is mostly unexposed (Dearnley 1963; Fettes et al. 1992). Palmer (1971) and Graham (1980) regarded the contact as a NE-dipping shear zone, based on the limited available outcrops. The structure of the Harris Granulite Belt is complicated by the lithological contrast between the main meta-igneous bodies and the Leverburgh Belt. In general, the Leverburgh Belt has a relatively uniform, near-vertical NW–SE-striking composite foliation. This is often mirrored in the margins of the main meta-igneous bodies, but internally they have more complex trends and, where
least deformed, a steeply NW-dipping foliation is seen, orthogonal to that in the Leverburgh Belt (Fig. 2; Graham 1980; Fettes et al. 1992). Dearnley (1963) considered that the structure of both the Harris Granulite Belt and Langavat Belt was controlled by largescale folding. Subsequent more detailed structural work (Graham 1980; Fettes et al. 1992; Mason 2012) has not substantiated this interpretation.

Two areas of the Harris Granulite Belt have been examined in detail here: the area around Roineabhal including most of the anorthosite and the adjacent rocks and the Toe Head peninsula (Fig. 2). The former area has been mapped in detail (Fig. 3), and Graham (1980) presented a detailed map of the latter area. The NE margin of the metadiorite was mapped along with the Langavat Belt in a previous component of this work (Mason 2012). These areas were focused on because they contain locally well-preserved granulite-facies deformation fabrics in a number of different lithologies and distinct younger amphibolite-facies ductile deformation. Given the likely plate margin origin of the Harris Granulite Belt (e.g. Baba 1998), linking the granulite-facies deformation to the intrusive and metamorphic history is of critical importance, as the plate margin the Harris Granulite Belt occupied must inevitably have been involved in subsequent suturing. Studying the subsequent amphibolite-facies ductile deformation provides a potential link to the history of the Langavat Belt and the underlying Archaean TTG gneisses.

**Anorthosite**

The meta-anorthosite forms a wedge-shaped outcrop superficially resembling an isoclinal fold. Contacts are not generally well exposed, but a lack of topographic deflection indicates that they are mainly near vertical. Where contacts are exposed they are usually concordant and are often formed by minor shear zones; quartzose intercalations are common where the contact is against felsic rocks. On Ha Cleit (Fig. 3), the anorthosite forms veins in adjacent metabasic rocks and contains metabasic ‘xenoliths’. This relationship is undeformed even though the anorthosite is foliated, suggesting that the apparent intrusive contact is due to remobilization of the anorthosite after the foliation formed. Minor pyroxenitic net-vein intrusions occur within the margins of the anorthosite and are associated with extensive local brecciation of the host anorthosite (attributed to explosive intrusion of the minor pyroxenites; Witty 1975). They are not seen within the central part of the anorthosite, or the
adjacent metasedimentary rocks. The pyroxenite net-veins should not be confused with later minor pseudotachylite veins or breccia that also occur sporadically (e.g. Macauidière & Brown 1982); although superficially similar, the former are easily distinguishable in the field by their coarse, high-grade mineral assemblages.

The anorthosite shows considerable variation in lithology and includes the following main types.

(1) Pure anorthosite: apparently structureless and largely devoid of mafic minerals.

(2) Spotted or schlieren anorthosite: anorthosite with approximately equidimensional mafic (garnet + clinopyroxene) where fresh, hornblende where retrogressed) patches typically a few centimetres in diameter, which have been flattened to differing extents into schlieren, defining a foliation.

(3) Massive garnet anorthosite: relatively mafic, massive ‘anorthosite’ with abundant garnet porphyroblasts. Small aggregates of clinopyroxene (or hornblende where retrogressed) are also common.

(4) Laminated metagabbro: finely banded or laminated feldspathic metabasite with garnet porphyroblasts.

(5) Banded anorthosite: ‘anorthosite’ showing centimetre to <1 m scale mineralogical or textural banding, or alternation of lithologies. The banding is usually somewhat irregular and discontinuous; cyclical layering is rare.

Attempts were made to delimit the various lithologies as structural markers on the basis that igneous layering had been reported previously (Witty 1975; see Fettes et al. (1992) for published summary), but this was hampered by a lack of lateral continuity. The ‘banded anorthosite’ and ‘laminated metagabbro’ show some semblance of igneous layering, which is generally now foliation-parallel, except in the hinges of occasional minor folds to which the foliation is axial planar (Fig. 4a). Three vague zones (approximating to the divisions of Fettes et al. (1992)) can, however, be seen. In the NW, the peripheral part of the anorthosite is characterized by laminated metagabbro bands up to c. 20 m thick and typically 50–100 m long, interspersed with other facies. The second domain comprises the majority of the mapped area and includes all the facies described except the laminated metagabbro bands. The third area forms the central part of the body at the SE end of the mapped area; it comprises almost entirely schlieren or spotted anorthosite. The boundaries of these areas are diffuse. On Ha Cleit, where the laminated metagabbro bands are most abundant, the bands are markedly oblique (c. 50°) to the contact of the anorthosite and to the inner boundary of the domain they collectively define. The NE boundary of the zone of schlieren anorthosite is flanked by a semi-continuous band of pure anorthosite; this is not repeated on the opposite side (Fig. 3).

The spotted anorthosite qualitatively provides natural strain ellipsoids and attests to the variation in deformation. Small areas near the centre of the body appear little deformed, but elsewhere ‘spots’ with aspect ratios of 10:1 are not uncommon, indicating considerable flattening associated with the foliation. Likewise, rare isoclinal minor folds within the banded anorthosite, to which the foliation is axial planar (Fig. 4a), indicate considerable local strain. Based on the deformation of presumed igneous features (mafic spots, banding), the foliation is assumed to be tectonic, but with a component of transposed igneous layering in the banded facies of the anorthosite. The foliation trends within the anorthosite are complex (see below and Fig. 3) and bear little relation to the lithological variation or contact. Despite the fold-like outcrop pattern, neither foliation nor lithological variation gives any clear indication that the anorthosite is isoclinal folded at the large scale, as claimed by Dearnley (1963) and Witty (1975).

Metanorite

The metanorite on the Toe Head peninsula (Fig. 2), apart from patchy retrogression to amphibolite facies, is a relatively uniform mafic–intermediate two-pyroxene granulite. A prominent, generally steep east–west-striking foliation is defined by the elongation of the pyroxenes (or hornblende where retrogressed) and turns into parallelism with the contact as the Leverburgh Belt is approached. The contact with the Leverburgh Belt is a steeply NE-dipping shear zone, although in places the deformation is restricted to the paragneisses. On the Toe Head peninsula, the metanorite has a discontinuous anorthosite marginal zone up to about 20 m thick (Fig. 4b). The typical metanorite passes NE into garnetiferous metanorite free of orthopyroxene, which then locally grades into impure anorthosite or leucocratic metabasite, often with thin pyroxene-rich ultramafic layers. Similar marginal rinds have been noted from intrusions elsewhere (e.g. Platten 1991) suggesting that, although sheared, the NE contact of the metanorite is the original intrusive boundary. Also on the Toe Head peninsula (e.g. [NF 9696 9145]), the metanorite contains occasional angular xenoliths of banded felsic rock resembling Leverburgh paragneiss. The xenoliths show only slight crenulation of their internal banding associated with foliation in the metanorite, suggesting that strain is fairly low. The metanorite is cross-cut by minor pyroxenite net veins or dykes.

Metadiorite

The metadiorite is a composite body and at many outcrops two components can be seen: an older slightly finer grained more mafic component that is cross-cut and veined by a slightly coarser and marginally more leucocratic component. The main foliation is
NE–SW striking and steeply NW dipping within much of the body and produces only slight distortion of the internal intrusive relationships. The NE contact against the Langavat Belt is intensely sheared and may be entirely tectonic. The NE–SW-striking foliation is progressively dextrally transposed as the Langavat Belt is approached (Graham 1980). The SW contact is locally also sheared, but intrusive relationships with the adjacent metabasite are still locally preserved (e.g. [NF 966 947]).
Supracrustal rocks

The main mass of the Leverburgh Belt forms the central part of the Harris Granulite Belt. Mapping by Dearlely (1963) and Baba (1997) indicates substantial lateral variation in the lithologies present, with single lithological horizons not traceable for more than c. 2 km along-strike. This lateral variation may betray the presence of early, now largely effaced folds, but the lack of coherent marker horizons makes this difficult to establish. Dearlely considered there to be tight NW–SE-trending folds present and attributed some reversals of dip in the NW–SE-striking composite foliation to such folds. However, these folds of the foliation have not been substantiated (e.g. Graham 1980), nor has any evidence of them been observed here.

Deformation fabrics and structures

Two generalized groups of structures and fabrics can be recognized within the Harris Granulite Belt: an older group of fabrics preserved within granulite-facies mineral assemblages and a younger group of retrograde amphibolite-facies structures and fabrics. Within the metanorite and the margins of the meta-anorthosite, for example on Ha Cleit (Figs 2 and 3). The only observed occurrences of the NE–SW–striking S1 foliation from the Leverburgh Belt are within garnet–kyanite–rich restites in some pelitic migmatites mapped in Figure 3. The younger foliation (S2) is NW–SE striking and steeply inclined. It is more common within the Leverburgh Belt and the margins of the main meta-igneous bodies, and wraps around or cuts across the S1 domains. S1 relics within the pelites are locally crenulated by S2.

In much of the metanorite of the Toe Head peninsula, the foliation is east–west striking and turns clockwise to NW–SE as the contact is approached. Based on the orientation, the foliation is assumed to be S1, that has been partially transposed, with the S2 overprint increasing towards the contact (see also Graham 1980, fig. 11, which suggests that the foliation was originally NE–SW striking in the metanorite). Likewise, in parts of the anorthosite, S1 relics can be seen to turn towards a NW–SE orientation, suggesting that much of the foliation in the anorthosite is partially transposed S1; the generally chaotic trends are due to an incomplete S1 overprint on S2.

Both S1 and S2 are preserved within near-pristine anhydrous garnet–kyanite–microcline and garnet–clinopyroxene–plagioclase granulite-facies assemblages, respectively in pelitic migmatites and the meta-anorthosite. In the metanorite on the Toe Head peninsula, partially transposed S1 (east–west striking foliation) is seen within orthopyroxene–clinopyroxene–plagioclase granulite-facies assemblages, and in thin ultramafic layers forming part of the marginal facies, S1 is preserved in garnet–clinopyroxene-dominated assemblages. Within the metanorite on Baloval Scarista (Fig. 2) S1 is seen in orthopyroxene–clinopyroxene–plagioclase granulite-facies assemblages. In the metanorite and the southwestern margin of the anorthosite respectively, partially transposed S1 and S2 are cross-cut by virtually undeformed mafic–ultramafic net-veins or small dykes (Fig. 4c). Within the banded facies of the anorthosite, isoclinal minor folds to which S3–S4 is axial planar are sporadically developed (Fig. 4a).

For c. 300 m SW of the meta-anorthosite, S3 is associated with a steeply inclined to moderately SE-plunging lineation (L3), defined by elongate garnet aggregates (stretched migmatite restite) and elongate garnet, kyanite and orthopyroxene crystals (Fig. 4e). L3 appears to be restricted to the SW flank of Roineabhial.

Post-granulite-facies deformation

Three main sets of post-granulite-facies fabrics and structures can be recognized. The oldest (S1, L1) comprises a steeply dipping NW–SE-striking foliation associated with a near-vertical but often poorly developed mineral lineation or quartz-feldspathic roding. Development of these fabrics is associated with extensive (usually total) retrogression to amphibolite facies. S1 and L1 become pervasive in the NE part of the metadiorite and can be directly traced into the fabrics of the Langavat Shear Zone. Further SW, they are only locally developed and appear to define diffuse and relatively superficial shear zones within the NE margin of the anorthosite and between Rodel and Leverburgh (Fig. 2). Minor folds and other mesoscopic structures such as shear bands and S–C fabrics are infrequent.

The second group of fabrics (S2, L2) comprise a NW–SE-striking foliation associated with a subhorizontal to moderately NW-plunging lineation, typically defined by biotite aggregates or fine quartz-feldspathic roding, and is mostly restricted to the Leverburgh Belt. Direct overprinting relationships between L1 and L2 are only rarely seen, but along Glen Rodel (between Leverburgh and Rodel; Fig. 2) and on the coastal section of the Bag Steinigie metasediments (a raft of Leverburgh-type metasediments within the metadiorite; Fig. 2) a weak subhorizontal lineation is superimposed on locally penetrative steep L1. These fabrics are often associated with near-vertical NW–SE-striking shear zones, the most significant of which is c. 30–40 m wide and flanks the metanorite on its NE side. Although clearly retrogressive (to amphibolite facies), deformation appears to have been poorly hydrous, as even in highly sheared rocks remnants of the granulite-facies assemblages often survive as porphyroclasts. S–C fabrics, shear bands, disharmonic drag folds, σ-geometry pods and porphyroclasts are common.

The final set of structures comprises conjugate sets of east–west-striking sinistral and north–south-striking dextral (both near-vertical) shear zones (Fig. 4b). These cross-cut the above structures and have displacements up to about 300 m, but usually much less. The shears may be very narrow (some superficially resemble faults), but they are always marked by the presence of amphibolite-facies tectonite. Many of the sinistral zones in the eastern part of the Leverburgh Belt host late granitic pegmatites (Fig. 3).

Kinematic criteria and deformation phases

The presence of foliation domains highly oblique to the regional NW–SE strike in all of the major components of the Harris Granulite Belt suggests that S1 (D1 deformation with respect to the meta-igneous bodies) was originally present throughout the Harris Granulite Belt. The present complex foliation pattern is attributed to the partial destruction of S2, which occurred preferentially in the weaker metasedimentary rocks and the margins of the meta-igneous bodies. The kinematic regime of D2 is not known, but there are no obviously related major structures, only the NE–SW-striking S2 foliation.

The meta-anorthosite shows a complex foliation trend that displays several superimposed deflections of S1. In places these deflections define an irregular, discontinuous, box-like antiform (Fig. 3) that bears little relationship to the geometry of the lithological contact. In general, S1 is deflected anticlockwise (apparent sinistral) towards the SW contact, and becomes subparallel to the composite S2–S4 foliation in the Leverburgh Belt. This deflection
is primarily related to \( S_4 \) (\( D_2 \) deformation) and predates the ultra-

crystalline net veins, which are largely undeformed along the SW mar-
gin of the meta-anorthosite. A single \( \sigma \)-geometry fresh 

orthopyroxene, apparently associated with \( L_4 \), in magnesian pelites 
(described below), SW of the anorthosite suggests the same sense of 

movement. However, \( L_4 \) may be the result of \( S_4 \) overprinting on 

\( S_1 \) (an intersection lineation), rather than a transport direction, so 

the apparent sinistral shear sense should be treated with caution.

Deflection of \( S_4 \) towards the NE contact of the anorthosite is 

commonly clockwise (apparent dextral) and is associated with 

map-scale \( \sigma \)-geometry deflections and shear band development 

(Fig. 3). Within the NE part of the meta-anorthosite, retrogression 
is extensive and the net veins are deformed and amphibolitized, 
demonstrating that much of the deflection of \( S_4 \) is younger than in 

the SW (Fig. 4d). Based on occasionally developed steep lineations, 
dip-slip movement is inferred. This is consistent with the apparent 
dextral deflection in a SW-up regime, as the deflected \( S_4 \) foliation has a 
NW dip. Based on the style of deformation and steep linea-
tion, this movement is associated with \( S_4-L_3 \) (\( D_3 \) deformation).

Graham (1980) documented similar apparent dextral deflection in 
the metadiorite against the Lagavat Shear Zone. The \( S_4-L_3 \), fabrics 
in the NE margin of the metadiorite are traceable into dextral-thrust 
fabrics in the Lagavat Shear Zone (Lang- \( D_3 \); Table 1). This shear 
sense is consistent with the deflection observed by Graham in the 
metadiorite and that observed here in the meta-anorthositic; \( D_3 \) is thus 
peripheral deformation associated with the Lagavat Shear Zone.

Kinematic criteria associated with \( S_4-L_3 \) (\( D_3 \) deformation) are 
abundant and include drag folds, shear bands and \( \sigma \)-type porphyro-
clasts, but indicate inconsistent shear sense. The main zone of \( S_4-
L_3 \) deformation flanking the metanorite contains both sinistral and 
dextral kinematic criteria, but without an overall obvious preferred 
 shear sense, or an indication of a systematic overprinting relation-
ship that would suggest more than one distinct phase of movement. 
A possible kinematic interpretation is that \( D_3 \) was a NE–SW coaxial 
flattening event in which NW–SE-striking shear zones developed to 
accommodate the heterogeneous development of \( S_4 \). Domains 
relatively unaffected by \( S_4 \), such as the metanorite, thus detached 
from domains where \( S_4 \) developed, such as the Leverburgh Belt, to 
accommodate the greater shape change in the latter areas; hence, in 
the case of the metanorite, strain was concentrated at the original 
intrusive contact. The oblique conjugate shears have east–west-
striking sinistral and north–south-striking dextral members, again 
indicating NE–SW flattening and NW–SE extension. They may 
well represent the latter stage of the same (\( D_3 \)) deformation. The 
subhorizontal \( L_3 \) stretching lineation, NW–SE-striking \( S_4 \) foliation 
and no overall preferred shear sense are consistent with the same 
coaxial strain geometry that the conjugate shears imply.

The extreme NE margin of the anorthosite locally displays appar-
ent sinistral deflection (Fig. 3). This is smaller scale and more 
peripheral than the \( D_3 \), apparent dextral deflection, and is considered 
to be a younger overprint. On Ha Cleit, \( D_3 \) apparent dextral shear 
zones are also folded by apparent sinistral drag-folds, and mylonitic 
retrogressed rocks NE of the anorthosite contain small-scale appar-
ent sinistral criteria. The style of deformation is similar to that in the 
 shear zone bordering the metanorite; hence, it is correlated to \( D_4 \).
The consistent sinistral \( D_3 \) shear sense in this area is attributed to the 
strike of the NE boundary of the anorthosite being slightly closer to 
east–west than the NE boundary of the metanorite; that is, during 
NE–SW coaxial shortening it had a consistent sinistral shear couple 
acting across it, whereas the boundary of the metanorite did not.

Relationship between metamorphism, deformation and intrusion

The three main metaigneous bodies all contain \( S_4 \), indicating they 
predate the end of \( D_3 \). Variation in \( S_4 \) fabric intensity between the 
intrusions suggests that they may collectively be syntectonic. \( S_4 \) in 
the metadiorite on Bolavlar Scarista (Fig. 2) and partially trans-
posed \( S_4 \) in the metanorite are defined by orthopyroxene–clinopy-
roxene–plagioclase. This is a probable \( M_4 \), ultrahigh-temperature 
semblage (Baba 1998), suggesting that \( M_4 \) is coeval with, or 
younger than, the \( S_4 \) foliation. The remobilization of the anorthosite 
contact seen on Ha Cleit largely post-dates \( S_4 \) foliation and can 
tentatively be attributed to peak \( M_4 \) metamorphism. \( S_4 \) in the 
 marginal facies of the metanorite and the anorthosite is locally defined 
in \( M_4 \), high-\( P \) garnet-clinopyroxene-bearing assemblages, thus \( S_4 \) 
predates the end of \( M_4 \).

Fe–Al-rich pelitic (terminology follows Baba 1997) migmatite 
lenses on the SW flank of Reoinchealh (NG0378 8541, Fig. 4e and f) 
comprise quartz-poor, garnet–kyanite-rich (\( M_4 \), high-\( P \) 
pelitic assemblage; Baba 1998) restite pods in more felsic leuco-

some. The restites comprise melanocratic garnet–kyanite-domi-
nated domains and more leucocratic areas richer in K-feldspar. \( L_2 \) 
is prominently defined by stretched migmatitic textures and by the 
alignment of kyanite associated with \( M_4 \), high-\( P \) metamorphism. 
NW–SE-striking \( S_2 \) foliation is defined by crude banding and the 
alignment of kyanite. Some restites contain NE–SW-striking \( S_4 \) 
 remnants defined by kyanite, which are somewhat crenulated by 
\( S_4 \). Retrograde fabrics are relatively scarce in these restites.

In sample SH01-45 (Fig. 4e and f), a pristine \( M_4 \), granulite, 
kyanite associated with \( M_4 \), high-\( P \) metamorphism defines an \( L_2 \) 
mineral lineation. The melanocratic garnet–kyanite-dominated 
portion of the rock is fractured and dilated perpendicular to this 
lineation, with K-feldspar locally leucocratic material filling the 
gaps. The alignment of the kyanite parallel to \( L_2 \) and its fracturing 
perpendicular to \( L_2 \) suggest that the kyanite is coeval with the 
lineation. This indicates that \( D_3 \) deformation overlaps with \( M_4 \) 
kyanite growth. The \( D_3 \) stretching of migmatitic textures and the 
preservation of \( S_2 \) within the restites, but not the leucosome, sug-
gest that migmatization largely predates \( D_3 \) deformation (although 
some syn-\( L_2 \) leucosome mobility is also indicated), with the com-
petent restites armouring \( S_2 \) relics against \( D_3 \). The timing of mig-
matization suggests that peak \( M_4 \), ultrahigh-temperature 
metamorphism predates \( D_3 \) deformation. Because \( S_3 \) cuts \( S_4 \) and 
is coeval with \( M_4 \), high-\( P \) metamorphism, and \( S_3 \) occurs within the 
restites of presumed \( M_4 \) migmatites, the metasedimentary rocks 
also suggest that \( S_3 \) is probably coeval with \( M_4 \), high-\( T \) metamor-
phism. This corroborates the relationships seen in the meta-
igneous rocks and helps to confirm the equivalence of \( S_3 \) and \( S_2 \) 
between the Leverburgh Belt and the meta-igneous bodies.

Quartzose magnesian orthopyroxene–kyanite-bearing pelitic 
gneiss around [NG 0392 8553] contain a retrograde biotite-defined 
foliation (in part \( S_4 \)), but small lenses preserve granulite-facies 
assemblages in a near-pristine state, including the rare assemblage 
orthopyroxene–kyanite–quartz (Fig. 4g). \( L_3 \) is preserved as a kyanite 
and orthopyroxene mineral lineation. Hollis et al. (2006) obtained a 
\( P-T \) estimate transitional between \( M_3 \), ultrahigh-temperature meta-

morphism and \( M_2 \), high-\( P \) metamorphism for a similar assemblage. 
The preservation of \( L_3 \) in a probable transitional assemblage, the 
Fe–Al-rich pelite described above, and the 90° change in foliation 
orientation between \( D_4 \) and \( D_3 \) suggest that \( M_3 \) is linked to the onset 
of \( D_3 \). The largely undeformed net-veins that cut \( S_4 \) in the anorthosite 
(Fig. 4e) and metanorite contain garnet–clino pyroxene-dominated 
\( M_3 \), high-\( P \) assemblages (Baba 1998), indicating that the latter stages 
of \( M_3 \) were static; \( M_2 \), outlasted \( D_3 \) deformation.

The magnesian pelites locally contain \( M_3 \), orthopyroxene–cordonierite 
decompression coronae partially replacing garnet (Fig. 4h), but the 
cordierite and secondary orthopyroxene do not contribute to any of the deformation fabrics present in the rock. Baba 
(1998) similarly noted the lack of contribution of \( M_2 \), retrograde 
minerals to deformation fabrics. The static conditions characteris-
tic of the latter stages of \( M_3 \), high-\( P \) metamorphism thus continued
during the M₄ part of the retrograde path. Deformation appears to have resumed only during D₃, once the Harris Granulite Belt had been exhumed to amphibolite facies, indicating a hiatus in deformation between D₂ and D₃.

**Discussion**

*A tectonic model for the Harris Granulite Belt*

Previous work (Cliff et al. 1983, 1998; Baba 1997, 1998; Whitehouse & Bridgwater 2001; Mason et al. 2004b; Baba et al. 2012) suggested that the Harris Granulite Belt is a continental magmatic arc derived mainly from a c. 1.9Ga juvenile source, but with some input of Archaean material. A c. 2.49Ga U–Pb zircon age, previously interpreted as the intrusion age of the anorthosite (Mason et al. 2004b), also provides indirect evidence of a ≈1.9Ga base ment to the Harris Granulite Belt. In light of the similarity in deformation history between the anorthosite and the rest of the Harris Granulite Belt, the vague concentric zonation suggesting that the anorthosite contact is a reworked intrusive boundary, the presence of an anorthositic facies in the 1890Ma metanorite (Mason et al. 2004b) and a c. 1.9Ga depositional age for the Leverburgh Belt (Whitehouse & Bridgwater 2001), the interpretation of the c. 2.49Ga age as a meta-igneous inheritance, rather than intrusion age (Mason et al. 2004b), is now preferred. The similarity of this age to intense disturbance of some of the central mainland TTG gneisses (e.g. Corté et al. 1994) is noteworthy. Archaean zircons reset at 2.49Ga would be a candidate for such meta-igneous xenocrysts. The Ness Assemblage too contains Archaean components (Whitehouse & Bridgwater 2001); indeed, reworked Archaean TTG appears to be the main component of the Assemblage, particularly in the northern part (Mason 2012). This is significant, because it suggests that the Harris Granulite Belt and Ness Assemblage represent a section through an active margin established on the Archaean gneisses that dominate the Lewisian, rather than a separate accreted arc terrane.

Baba et al. (2012) dated the onset of metamorphism in the Leverburgh metasediments to 1909±3 Ma (Timmerman et al. 2001) but also dated a c. 1910Ma monzane age) and regarded the main meta-igneous bodies as the heat source for triggering M1; Cliff et al. (1983, 1998) dated the end of granite-facies metamorphism to c. 1870Ma. The observed relationship between migmatization and S₁, and the main intrusions and S₁, suggests that the late D₁ emplacement of the metadiorite and metanorite is a credible heat source for peak M₁. However, the onset of M₁ predates the metanorite and metadiorite, respectively dated at 1890±1Ma and 1888±2Ma (Mason et al. 2004b), by some 20myr. The c. 10kbar pressure estimate for M₁ (Baba 1998) also requires burial of the Leverburgh Belt to c. 35km, with early M₁ amphibolite-facies inclusions in garnet cores (Baba 1998) indicating a former geothermal gradient requiring significant burial-related heating at 35km depth. The Ness Assemblage underwent coeval metamorphism (Mason 2012, and references therein) but metadiorites and anorthosites are only minor components, which are unlikely to have provided significant metamorphic heating. A more robust explanation is that metamorphism was initiated by tectonic thickening c. 1910Ma, perhaps associated with accretion of material to the active margin; superimposed intrusion-related heating in the Harris Granulite Belt at 1888–1890Ma drove peak M₁, and the antilockwise P–T path (Baba 1998).

Subduction-related magmatism is well dated at 1888–1890Ma in South Harris, at 1903Ma (And Gneiss) in the Loch Maree Group (Park et al. 2001) and at 1886Ma in the equivalent rocks (Ammassilik Intrusive Complex) in SE Greenland (Kalsbeek et al. 1993). The 1910–1870Ma metamorphism begins well before the youngest known subduction-related magmatism, but ends some 20myr after subduction-related magmatism had apparently been terminated. In this context the D₃–M₄ to D₃–M₄ transition in the Harris Granulite Belt makes sense if the 1910–1870Ma metamorphism is a composite event, beginning with accretion- or arc-related metamorphism (M₄), rapidly followed by a subduction-terminating continent–continent collision; D₄–M₄ represents crustal thickening at the onset of this suturing collision. Exhumation of the Harris Granulite Belt and Ness Assemblage around 1860–1870Ma (Table 1) appears to mark the end of this (Early Laxfordian) tectonohermal cycle.

**Relationship to the Langavat Belt**

The deformation history of the Langavat Belt is summarized in Table 1; Langavat deformation phases are prefixed ‘Lang’. Two main phases of intense ductile shearing associated with the Langavat Shear Zone and younger than the Outer Hebrides dykes are recognized (Lang-D₃ and Lang-D₄). Lang-D₃ marks the initiation of the Langavat Shear Zone in top-to-the-WNW dextral extension and Lang-D₄ records top-to-the-north dextral-reverse reactivation. The tectonohermal history of the Langavat Belt up to and including Lang-D₄, recorded within its main supracrustal rocks, is one of prograde metamorphism resulting in relatively low-pressure mid- to upper amphibolite-facies conditions (Mason 2012); the associated deformation is recorded in highly annealed (Type 1) tectonites (Table 1). Primary burial of the supracrustal rocks and the initiation of prograde metamorphism are attributed to now cryptic Lang-D₃ thrusting that predates the Langavat Shear Zone (Table 1; Mason et al. 2004a; Mason 2012). Two phases of brittle–ductile reactivation of the Langavat Shear Zone (Lang-D₄ and Lang-D₅) are also recognized, which affect 1675–1657Ma granite pegmatites cutting the Type 1 tectonites (Mason & Brewer 2005; Table 1). Lang-D₄ records top-to-the-SE sinistral-reverse movement and Lang-D₅ records top-to-the-NW dextral extension. Both are recorded in unannealed greenschist-facies (Type 2) tectonites (Table 1). Lang-D₅ is probably of Grenvillian age (Mason & Brewer 2005, and references therein) and is not considered further here. The termination of amphibolite-facies metamorphism and the transition to brittle–ductile deformation after Lang-D₅ is attributed to the onset of exhumation.

Lang-D₄ is well represented in the Harris Granulite Belt by D₄, but Lang-D₅ is poorly represented (possibly absent) in the Harris Granulite Belt. The pre-D₄ deformation in the Harris Granulite Belt is associated with high-P and ultrahigh-temperature metamorphism, whereas the D₄ deformation is associated with retrogression of the high-grade assemblages. Comparable high-grade metamorphism and equivalents of the concomitant D₃ and D₂ deformation are absent from the Langavat Belt; the tectonohermal histories of the two belts converged only after D₄, once granulite-facies metamorphism in the Harris Granulite Belt had ceased.

Age constraints on the cessation of granulite-facies metamorphism (e.g. Cliff et al. 1983) indicate that D₃ (in the Harris Granulite Belt) is older than around 1870Ma, whereas age constraints from the Langavat Belt place D₃–Lang-D₃ at ≤1746Ma (Mason 2012; Table 1). This confirms the field and metamorphic evidence for a hiatus in deformation between D₃ and D₄ in the Harris Granulite Belt. It also indicates that at least the latter part of the amphibolite-facies metamorphism recorded in the Langavat Belt (i.e. that coeval with D₃–Lang-D₃) is substantially younger than the granulite-facies metamorphism in the Harris Granulite Belt. Fluids released by prograde metamorphism of the Langavat Belt probably facilitated the retrogression and deformation of the Harris Granulite Belt. Thus, relict granulite-facies assemblages armoured the Harris Granulite Belt against the Langavat Shear Zone during Lang-D₄, inhibiting this deformation, but, by Lang-D₅, the Harris Granulite Belt had been sufficiently weakened by retrogression to succumb to Lang-D₅ hydrous shearing. By D₅, prograde metamorphism in the Langavat Belt had ceased (Table 1),
limiting fluid supply; D₃ is, thus, poorly hydrous compared with D₂. Considering the amphibolite-facies metamorphism is prograde with respect to the Langavat supracrustal rocks but retrograde with respect to the Harris Granulite Belt, and the hiatus in deformation, two distinct tectonothermal cycles are supported. Some thought does, however, have to be given to why so much variation exists in the post-granulite-facies deformation history, if these events are to be considered as part of a single Late Laxfordian cycle.

The combined post-granulite-facies deformation history of the Harris Granulite Belt and Langavat Belt suggests progressive or incremental rotation of the stress field. Lang-D₂, the oldest well-preserved event, records top-to-the-WNW transport, which had rotated to top-to-the-north transport (north–south shortening) by D₃-Lang-D₄ (Table 1). Further rotation led to NE–SW flattening in D₄ producing a NW–SE-striking foliation, NW–SE-striking shear zones with inconsistent shear sense, and north–south-striking dextral and east–west-striking sinistral conjugate shears. Further rotation led to east–west flattening, which diluted many of the east–west-striking D₃ shear zones allowing pegmatite emplacement, and the Langavat Shear Zone was reactivated with a sinistral-thrust sense (Lang-D₄). Superimposed on this, there almost inevitably were backslips associated with Lang-D₃ thrusting and subsequent exhumation, as crustal thickening and exhumation inevitably must have altered gravitational loading. Extension and prograde amphibolite-facies metamorphism during Lang-D₄, and widespread coeval gravitational flattening of the Archaean gneisses are probable consequences of loading by major Lang-D₃ thrusting (Mason 2012; Table 1). In turn, modest Lang-D₃ and Lang-D₄ compression can be considered the response to Lang-D₃–erosion–exhumation reducing the overburden opposing the horizontal tectonic stresses. In the case of Lang-D₄, this would account for a compressional event on the retrograde path, as at some stage exhumation will inevitably also terminate prograde metamorphism. The post-granulite-facies tectonothermal history is consistent with a single progressive cycle of crustal thickening and exhumation initiated by Lang-D₃ and regulated by a rotating stress field.

The timing of Lang-D₃ thrusting of the Harris Granulite Belt over the Langavat Belt is not directly constrained. However, Lang-D₃, is the apparent trigger for prograde syn-Lang-D₂ metamorphism, which itself represents the prograde path of the amphibolite-facies metamorphism that prevailed during Lang-D₂, indicating that Lang-D₃ and Lang-D₂ are allied, and Lang-D₂ is dated to ≤1746 Ma (Mason 2012; Table 1). Similarly, the combination of Lang-D₂ extension superimposed on Lang-D₃ thrusting is a probable control on where the Injection Complex formed, and this too is probably dated to 1704–1657 Ma (Friend & Kinny 2001; Kelly et al. 2008; Table 1). This circumstantially places the entire (Late Laxfordian) tectonothermal cycle of the Langavat Belt some 100 myr after the 1910–1870 Ma high-grade (Early Laxfordian) metamorphism in the Harris Granulite Belt; ample time for static exhumation of the Harris Granulite Belt from high-P conditions, prior to Lang-D₃ emplacement over the Langavat Belt (Table 1).

Integration with the rest of the Lewisian Complex

The northern part of the mainland Lewisian (Fig. 1) records granite magmatism along the Laxford Shear Zone (Goodenough et al. 2010), an Inverian reverse-sense shear zone reactivated during the Laxfordian (Coward 1990; Goodenough et al. 2010). Friend & Kinny (2001) and Goodenough et al. (2013) dated these granites to c. 1855 and c. 1880 Ma, and, based on their mineralogy, Goodenough et al. (2013) suggested that they possibly contain a juvenile magmatic component. Coward (1990) identified two phases of Laxfordian deformation following the earlier Inverian movement: an earlier oblique dextral-thrust phase considered to predate the granites and a younger sinistral-extensional phase possibly coeval with the granites. Younger, undeformed (presumably post-tectonic) granites dated at c. 1770 and c. 1790 Ma (Goodenough et al. 2013) place a lower age limit on significant deformation within and to the NE of the Laxford Shear Zone. If correct, this history is consistent with a mid-crustal expression of the Early Laxfordian history seen in the Harris Granulite Belt: the earlier Laxfordian deformation recognized by Coward (1990) may equate to D₃–M₂ thickening in the Harris Granulite Belt, whereas the younger extensional event and granite magmatism correspond to the crustal collapse–exhumation that ended high-P metamorphism in the Harris Granulite Belt. The possible juvenile component in the c. 1880 Ma granites may represent broadly back-arc magmatism with respect to the Harris Granulite Belt. However, given the younger 1855 Ma age, some of this input may represent the final products of a moribund subduction zone, which had essentially already been terminated by the M₃–D₄ collision seen in the Harris Granulite Belt, around 1880 Ma.

The c. 1770–1790 Ma undeformed granites correlate to the hiatus in deformation seen in the Harris Granulite Belt. Titanite ages of c. 1740 and c. 1670 Ma (Corfu et al. 1994; Friend & Kinney 1995; Goodenough et al. 2013) from the northern-central part of the mainland are comparable with deformation ages for the Langavat Shear Zone (Table 1) and the emplacement of the Injection Complex, indicating a shared Late Laxfordian history. However, the lack of strong deformation after the 1770–1790 Ma granites were emplaced indicates that, unlike on the Outer Hebrides, the Late Laxfordian was weakly expressed in the northern mainland (see Clayman of Dash & Bowes 2014) and what Laxfordian deformation did occur was Early Laxfordian in the present terminology.

Goodenough et al. (2013) attributed the 1880 Ma granites to an early cycle of arc magmatism broadly correlated with South Harris (Early Laxfordian in the present terminology), but considered the 1770–1790 Ma granites to indicate crustal melting associated with the onset of a new tectonothermal cycle. The former interpretation is consistent with the present work. However, given that the 1770–1790 Ma granites are reported to be largely undeformed (Goodenough et al. 2013) and correspond to a period of quiescence on South Harris, they are perhaps better interpreted as anastistic melts related to post-Early Laxfordian exhumation, rather than the initiation of the Late Laxfordian.

The Loch Maree Group and Ialltaig gneisses show more obvious similarities to South Harris. The Ialltaig gneisses record c. 1.875 Ga granite-facies metamorphism (Love et al. 2010) and contain a NE–SW-striking foliation (Park 2002). Park regarded this foliation to be of Scourian (late Archaean age) but a subsequent c. 2 Ga protolith age estimate (Love et al. 2010) refutes this idea; this fabric is a potential correlative of S₁ in the Harris Granulite Belt. The 1790 Ma granite pegmatites (Mason 2012) may repre-
gneisses to the NE and SW by thrusting during and following accretion. Both Park et al. (2001) and Droop et al. (1999, and references therein) recognized the same early ductile deformation associated with prograde amphibolite-facies metamorphism and consistent with top-to-the-NW or -WNW transport, but reached different conclusions as to its significance. Park et al. (2001) considered that the early deformation and metamorphism was approximately coeval with arc magmatism (Ard Gneiss c. 1903 Ma) and with the accretion of the Loch Maree Group and, hence, records very early orogenic events; Early Laxfordian in the present terminology. This was based on the assumption that the Ard Gneiss was syntectonic, because it contains weaker deformation fabrics than the surrounding supracrustal rocks (Park et al. 2001). This is a non-unique interpretation that equally could be explained by competence variation. Conversely, Droop et al. (1999) suggested that metamorphism is of Barrovian type related to prior crustal thickening and that top-to-the-NW shearing occurred in its final stages (their D4 phase) under retrograde conditions, consistent with exhumation or collapse of previously thickened crust and relatively late orogenic events. However, they were unable to identify the crustal thickening mechanism.

The Droop et al. (1999) hypothesis is entirely consistent with the Late Laxfordian tectonothermal history of the Langavat Belt, in which the top-to-the-NW deformation and coeval metamorphism are the responses to prior burial by thrusting. In light of the revised ages for the Ialltaig gneiss (Love et al. 2010), the top-to-the-NW deformation is now somewhat problematic for the Park et al. (2001) interpretation. Where dated Early Laxfordian granulite-facies assemblages and fabrics are preserved in the Ialltaig Gneiss and Harris Granulite Belt (the latter including high-P arc rocks), they are in low-strain zones (see above and Park 2002), anti-correlated to the top-to-the-WNW deformation. If the top-to-the-NW deformation really was Early Laxfordian and associated with the accretion of the Loch Maree Group around 1900 Ma as claimed by Park et al. (2001), there is no obvious reason for such an anti-correlation to exist. There is also no known high-P metamorphism associated with the top-to-the-NW deformation to link this deformation to crustal duplication of the scale of that invoked by Park et al. (2001) in their post-accretion continent—continent collision. Nevertheless, the main ideas of Park et al. (2001) can be retained with the modification that the thrust-related burial of the Loch Maree Group is reassigned to the Late Laxfordian Gairloch–Langavat Thrust System. In this scenario, the top-to-the-NW deformation and coeval metamorphism is part of the Late Laxfordian signature seen on South Harris, unrelated to and overwritten on the subduction-accretion history of the Loch Maree Group protoliths. The Ialltaig Gneisses and Harris Granulite Belt survive as low-strain zones relative to the top-to-the-NW deformation as a result of armouring by Early Laxfordian high-grade assemblages.

Additional evidence supporting this modification of the Park et al. (2001) model comes from the observation of Wheeler (2007, fig. 2) that the degree of Laxfordian reworking in the TTG gneisses SE of the Loch Maree Group (excepting the relatively narrow zone of high-strain gneisses in the SW margin of the Gairloch Shear Zone) generally decreases towards to the NE, towards the Loch Maree Group. This is not what would be expected on approach to an intact Laxfordian suture (as in the Park model). However, this would be consistent with the present geometry being established by Late Laxfordian thrust stacking, long after original accretion; the general SW increase in reworking marks the overall transition into the deeper, hotter, more easily deformed part of the Late Laxfordian orogen, more comparable with the section visible on the Outer Hebrides (Fig. 1).

The subsequent deformation in the Loch Maree Group (considered to be part of a second Laxfordian collision event by Park 2005; see below) is dominated by a combination of dextral shearing associated with the Gairloch Shear Zone and the formation of large NW–SE-trending upright folds under greenschist-facies conditions (Park 2010). It is not clear if these events have exact correlates on South Harris, but the transition from north–south to NE–SW flattening (Lang-D2–Harris Granulite Belt D2, and Harris Granulite Belt D3) regime could account for most structures seen in the Loch Maree Group, although deformation occurred under lower grade conditions than on South Harris. Titanite ages of 1660–1670 Ma (Love et al. 2010) from just south of the Loch Maree Group versus ages of 1642 and c. 1560 Ma from the Outer Hebrides (Mason et al. 2004b) do, however, seem to corroborate that the Gairloch Shear Zone exited amphibolite facies earlier than South Harris.

Tectonic evolution of the Lewisian

Accreted oceanic material within the Loch Maree Group and arc components within the Harris Granulite Belt provide strong evidence for the existence of an Early Laxfordian suture. Indeed, the combination of abrupt change of structural trends, high-P metamorphism and the apparent termination of arc magmatism around 1880 Ma provides a strong candidate for the signature of the collision that amalgamated the Lewisian Complex. A number of supposed terrane boundaries (notably the Langavat Shear Zone and straights of the Gairloch Shear Zone) younger than 1880 Ma have been invoked (Kinny et al. 2005; Love et al. 2010) and, to overcome the problem of a lack of juvenile arc rocks after c. 1880 Ma associated with these boundaries, predominantly strike-slip Late Laxfordian amalgamation is considered (Kinny et al. 2005). However, features such as major thrusting, widespread ductile deformation, prograde metamorphism and burial of metasedimentary rocks indicate that the Late Laxfordian is primarily a continental crustal thickening event, with subordinate strike-slip movement, not a strike-slip event per se. The idea of strike-slip terrane accretion can thus be rejected on geometric grounds, as a dominantly strike-slip regime does not offer the prerequisite driving mechanisms for the observed Late Laxfordian signature. A better interpretation is that the Lewisian was amalgamated at c. 1880 Ma in the Early Laxfordian and then underwent intense structural modification in the Late Laxfordian, as a result of a second collision external to the Complex (see below).

Interpretation of the Laxfordian configuration ultimately hinges on the significance that is assigned to the pre-Laxfordian geology; specifically, whether features such as the wide distributions of the c. 2.73 Ga U–Pb age signature, apparent ‘Inverian’ deformation and mafic dykes (Fig. 1) are taken to indicate that the TTG gneiss formed a single coherent plate at the onset of the Laxfordian or several (re) assembled plates with a comparable pre-Laxfordian history. The lack of large cratonic domains free of significant Proterozoic rifting within the Lewisian, such as those seen in Greenland, makes this difficult to directly assess. However, the presence of active margin remnants (Harris Granulite Belt, Ness Assemblage, Loch Maree Group) along the Gairloch–Langavat Thrust System indicates that this structure probably marks the vestige of the Early Laxfordian suture. Nevertheless, the lack of widespread c. 1880 Ma metamorphism in the Archaean TTG gneisses, coeval with the M2 high-P metamorphism in the Harris Granulite Belt, indicates that substantial telescoping has dissociated the Harris Granulite Belt from whatever it originally collided with during D2–M2. Indeed, besides the allochthonous slivers along the Gairloch–Langavat Thrust System itself, the only part of the Early Laxfordian belt not excised (based on the presence of Early Laxfordian granites and deformation) seems to be the northern mainland. This geometry can be explained by the suturing of two Archaean plates, one with an active margin, during the Early Laxfordian (Fig. 5a), followed by Late Laxfordian thrust stacking, which more or less juxtaposed the Early Laxfordian forelands, excising much of the Early Laxfordian
A. Post-Early Laxfordian Suturing

The similarity of the zircon age from the South Harris anorthosite to the c. 2.49Ga disturbance of the central mainland (e.g. Corfu et al. 1994) suggests that it was the NE Archaean block (Fig. 5a) that the active margin developed on, implying NE-directed subduction beneath the central and northern mainland; a juvenile component in the 1880Ma granites along the Laxford Shear Zone, perhaps in part representing back-arc magmatism, is consistent with this (see Goodenough et al. 2013). The greater concentration of derrite and anorthosite in the Harris Granulite Belt compared with the Ness Assemblage, which appears to be largely reworked Archaean TTG, formed the metadiorite and anorthosite bodies of the Ness Assemblage and the Harris Granulite Belt. In the latter case, these were emplaced into the already hot Leverburgh metasediments c. 1890Ma, driving local granulite-facies metamorphism.

By c. 1880Ma, subduction had ended that the onset of continent–continent collision; this collision deeply buried parts of the continental margin, such as the Harris Granulite Belt, producing high-P metamorphism, whereas many areas (Loroch Mearc Group, Langavat Belt, most Archaean TTG gneisses) remained at high level, temporarily escaping intense metamorphism. Other areas such as the Laxford Shear Zone and northern mainland were affected, but much more peripherally than the Harris Granulite Belt. By 1860–1870Ma, subduction had ended and the over-thickened crust collapsed, exhuming the Harris Granulate Belt and Ness Assemblage and ending the c. 1901870Ma metamorphism. Along with the granite magmatism and extensional movement associated with the Laxford Shear Zone, this marks the tail end of the Early Laxfordian cycle and the start of a period of relative quiescence. Eclogite-facies metamorphism around 1750Ma in the Glenelg–Attadale inlier (Storey et al. 2010; Fig. 1), suggests that the Lewisian Complex was involved in a second major collision. It seems likely that this collision drove the major thrusting that initiated the Late Laxfordian tectono thermal cycle (Table 1).

Comparison with the Nagssugtoqidian orogen

Pre-Atlantic reconstructions correlate the Lewisian Complex with the Nagssugtoqidian orogen of SE Greenland (e.g. Buchan et al.
2000), a collisional orogen between the North Atlantic Craton in the SW and the Central Greenland Craton in the NE (Park 2005; Nutman et al. 2008; Kolb 2014). The Lewisian has close similarities to SE Greenland. Both areas have large domains of reworked Archaean gneisses above which are displaced Proterozoic arc remnants and relatively well-preserved Archaean granulites (Nutman et al. 2008). Arc rocks in SE Greenland (Ammassalik Intrusive Complex and associated supracrustal rocks) are almost identical in terms of protoloth age to the Harris Granulite Belt; both contain a 1.9 Ga supracrustal rocks substantially derived from a Proterozoic source and c. 1.89 Ga diorite–norite intrusions and syn-intrusion metamorphism (Kalsbeek et al. 1993). High-P metamorphism within the Nagssugtoqidian orogen at 1867 ± 28 Ma (Nutman et al. 2008) is also comparable with the inferred timing of M2 in the Harris Granulite Belt. These similarities support correlation with the Early Laxfordian. The interpretation of the Early Laxfordian presented here is consistent with two-plate type interpretations of SE Greenland (Nutman et al. 2008; Kolb 2014) and suggests that the Lewisian straddles the continuation of the Nagssugtoqidian suture. The timing of the inferred continent–continent collision c. 1870–1880 Ma is also consistent (see above and Kolb 2014).

**More than Nagssugtoqidian deformation?**

Although correlation of the Early Laxfordian with the Nagssugtoqidian seems clear, it does lead to the question: what caused the extensive Late Laxfordian tectonothermal activity? A likely candidate has been discussed by Park (2005); namely, the c. 1740 Ma accretion of the Malin block. This forms much of the Scottish basement to the SW of the Lewisian Complex (Morton et al. 2014) and effectively forms the continuation of the Labradorian–Ketilidian–Gothian orogen (Zhao et al. 2004; Park 2005). In SE Greenland the Ketilidian orogen is separated from the Nagssugtoqidian orogen by the North Atlantic Craton. However, this largely thins out by the point at which the Lewisian is reached (Buchan et al. 2000), such that there would be little or no foreland separating these orogens. Park (2005) attributed only limited folding and shearing to the collision of the Malin block. However, the present work suggests a second major collision, in which the Lewisian Complex, representing the dissected vestiges of the Nagssugtoqidian and its former foreland, was telescoped and reworked within a younger orogen, probably the Labradorian–Ketilidian–Gothian orogen as in the model of Park (2005). Much of the Proterozoic ‘terrane’ structure of the Lewisian can be considered to be superficial; a result of this second collision dissecting the earlier configuration, rather than true accretion. Thus, the Lewisian is far less complex than models such as that of Kinny et al. (2005) suggest, and requires no special consideration in palaeogeographical reconstructions to accommodate an abnormally large number of small ‘terranes’.

**Conclusions**

Although the ideas discussed here are speculative, a substantial proportion of the Palaeoproterozoic evolution of the Lewisian Complex can be explained in terms of a relatively small number of overprinting processes or events associated with two Laxfordian tectonothermal cycles. The Early Laxfordian cycle is equivalent to the Nagssugtoqidian orogen of Greenland, and was driven by the development of an active continental margin on a Archaean block and subsequent continent–continent collision between about 1.91 and 1.85 Ga. Most of the Archaean gneisses themselves appear to have been relatively unaffected and lay either in the foreland or at least in the peripheral part of the belt.

The Late Laxfordian cycle probably equates to the Labradorian orogen and marks a second collision between the Lewisian and the Malin block to the SW. Thrusting telescoped the Early Laxfordian belt, buring the former SW foreland in the process, triggering regionally extensive amphibolite-facies metamorphism and deformation; as a result of subsequent erosion of overthrust Early Laxfordian material, the Early Laxfordian belt was largely excised. Much of the Palaeoproterozoic ‘terrane’ structure of the Lewisian can be attributed to the dissection and restructuring of the Early Laxfordian configuration, rather than true terrane accretion. This leaves only two probable major Palaeoproterozoic boundaries: the continuation of the Nagssugtoqidian (i.e. Early Laxfordian) suture and the suture associated with the Malin block and the c. 1750 Ma Glenelg eclogites.

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**References**

Baba, S. 1997. Geology and geochemical characteristics of the Leverburgh Belt in South Harris, Outer Hebrides, northwest Scotland. Journal of Geosciences, Osaka City University, 40, 119–143.

Baba, S. 1998. Proterozoic anorthosite crust: a P–T path of the Lewisian Complex of South Harris, Outer Hebrides, NW Scotland. Journal of Metamorphic Geology, 16, 819–841.

Baba, S. 1999a. Evolution of the Lewisian complex in south Harris Northwest Scotland and its relation to the north Atlantic craton in the Palaeoproterozoic (2.0 Ga). Journal of Geosciences Osaka City University, 42, 115–125.

Baba, S. 1999b. Sapphirine-bearing orthopyroxene–kyanite/sillimanite granulites from South Harris, NW Scotland: Evidence for Proterozoic UHT metamorphism in the Lewisian. Contributions to Mineralogy and Petrology, 136, 33–47, http://dx.doi.org/10.1007/s004100050522.

Baba, S., Dunkley, D.J., Hokada, T., Horie, K., Suzuki, K. & Shiiraiashi, K. 2012. New SHRIMP U–Pb zircon ages and CHIME monazite ages from South Harris granulites, Lewisian Complex, NW Scotland: Implications for two stages of zircon formation during Palaeoproterozoic UHT metamorphism. Precambrian Research, 200–203, 104–128.

Bott, M.H.P., Holland, J.G., Storri, P.G.G. & Watts, A.B. 1972. Geophysical evidence concerning the structure of the Lewisian of Sutherland, N.W. Scotland. Journal of the Geological Society, London, 128, 599–610, http://dx.doi.org/10.1144/gsja.128.6.0599.

Buchan, K.L., Mertanen, S., Park, R.G., Posenen, L.J., Elmng, S.A., Abrahamsen, N. & Bylund, G. 2000. Comparing the drift of Laurentia and Baltica in the Proterozoic: the importance of key palaeomagnetic poles. Tectonophysics, 319, 167–198, http://dx.doi.org/10.1016/S0040-1951(00)00329-9.

Cliff, R.A., Gray, C.M. & Hulma, H. 1983. A Sm–Nd isotopic study of the South Harris Igneous Complex, the Outer Hebrides. Contributions to Mineralogy and Petrology, 82, 91–98, http://dx.doi.org/10.1007/BF00371178.

Cliff, R.A., Rex, D.C. & Gaise, P.G. 1998. Geochronological studies of Proterozoic crustal evolution in the northern Outer Hebrides. Precambrian Research, 91, 401–418, http://dx.doi.org/10.1016/S0301-9268(98)00060-6.

Corfu, F., Heaman, L.M. & Rogers, G. 1994. Polymetamorphic evolution of the Lewisian complex, NW Scotland, as recorded by U–Pb isotopic compositions of zircon, titanite and rutile. Contributions to Mineralogy and Petrology, 117, 215–228.

Corfu, F., Crane, A., Moser, D. & Rogers, G. 1998. U–Pb zircon systematics at Grunard Bay, northwest Scotland: Implications for the early orogenic evolution of the Lewisian complex. Contributions to Mineralogy and Petrology, 133, 329–345, http://dx.doi.org/10.1007/s0041000500456.

Coward, M.P. 1990. Shear zones at the Laxford front, NW Scotland and their significance in the interpretation of lower crustal structure. Journal of the Geological Society of London, 147, 279–286, http://dx.doi.org/10.1144/gsjgs.147.2.0279.

Coward, M.P. & Park, R.G. 1987. The role of mid-crustal shear zones in the Early Proterozoic evolution of the Lewisian. In: Park, R.G. & Taner, J. (eds) Evolution of the Lewisian and Comparable Precambrian High Grade Terranes. Geological Society, London, Special Publications, 27, 127–138, http://dx.doi.org/10.1144/gsl.sp.1987.027.01.11.

Coward, M.P., Francis, P.W., Graham, R.H. & Watson, J. 1970. Large-scale tectonics of the Outer Hebrides in relation to those of the Scottish mainland. Tectonophysics, 10, 425–435, http://dx.doi.org/10.1016/0040-1951(70)90120-4.

Crowley, Q.G., Key, R. & Noble, S.R. 2014. High-precision U–Pb dating of complex zircon from the Lewisian Gneiss Complex of Scotland using an incremental CA-ID-TIMS approach. Gondwana Research, 27, 1381–1391.

Dash, B. & Blowes, D.R. 2014. Lewisian Complex of Strath Dionard–Rhiconich and its significance in the early history of the NW Highlands of Scotland. Scottish Journal of Geology, 50, 27–47, http://dx.doi.org/10.1144/sgj2013-006.
