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Distinct differences now exist between the mainland and Outer Hebridean outcrops of the Lewisian Gneiss Complex, with the boundary probably coinciding with the Outer Hebrides Fault Zone. This separation combined with more reliable geochronology from neighbouring Palaeoproterozoic regions suggests that the assumption of continuity between the Nagssugtoqidian and Lapland-Kola Belts through the Lewisian Gneiss Complex may be incorrect.

Evidence for Palaeoproterozoic terrane assembly in the Lewisian Gneiss Complex on the Scottish mainland south of Gruinard Bay

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Abstract: The so-called 'southern region' of the mainland Lewisian Gneiss Complex is shown to be a package of isotopically distinct tectonostratigraphic units, from south to north the Rona, Ialltaig and Gairloch terranes. These were deformed during and following their accretion to the previously assembled Gruinard, Assynt and Rhiconich terranes to the north. SHRIMP U/Pb zircon geochronology has identified the 2730 Ma granulite facies metamorphic signature of the Gruinard Terrane in a tonalitic gneiss from the Tollie antiform, which is in sheared contact with the Palaeoproterozoic Gairloch Terrane (Loch Maree Group and Ard Gneiss) along the Creag Bhan Belt. Current structural interpretations of the region suggest the tectonic contact may be folded over the Carnmore antiform, coinciding with the high-strain zone along the Gruinard Belt. South of the Gairloch Terrane the Archaean gneisses of the Rona Terrane show a quite different pattern to those of the Gruinard Terrane, yielding older zircon protolith ages (ca. 3135-2880 Ma), and no evidence for the 2730 Ma Gruinard granulite facies metamorphism. Instead they show evidence for metamorphism and melting at ca. 2953 Ma and ca. 2700 Ma in two different low-strain zones. Contained within the deformed southern margin of the Gairloch Terrane the Ialltaig terrane occurs, bounded by the Kerry shear zone. This is shown to be an exotic slice of granulite facies Palaeoproterozoic crust, rather than Archaean basement as previously thought. The ca. 1877 Ma granulite facies metamorphism of the Ialltaig block is the youngest event unique to a single terrane in this part of the Complex, making it an upper estimate for the timing of amalgamation with surrounding tectonic units. U-Pb titanite ages of 1670 ± 12 Ma and ca. 1660 Ma for low-strain zones at Diabaig are interpreted to be cooling through the titanite closure temperature after the amphibolite facies reworking of these southern terranes and the southern margin of the Gruinard Terrane.

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1. Appreciation

Northwest Scotland was, arguably, the birthplace of the understanding of gneiss complexes (Peach et al. 1907), and it is fitting that Bill Compston's development of the SHRIMP instrument has helped immensely in the quest to determine at least a more understandable history for the Lewisian Gneiss Complex. Approximately 50 years ago the Lewisian of northwest Scotland was, on the basis of field evidence alone, originally considered a contemporaneous section of variably reworked basement gneisses (e.g. Sutton and Watson, 1951). The development of whole rock isotopic dating techniques led to the production of many 'isochrons' most of which contained lithologies from all parts of the complex that in reality had no geological meaning (e.g. Hamilton et al., 1979) and so complicated the story. Bulk zircons had been tried but the problems of dealing with complex grains, including inheritance, also meant that these ages were suspect (e.g. Pidgeon and Bowes, 1972). These different spurious ages from the different methods became accepted whereas ages that were better constrained, being obtained from coherent sample sets but more difficult to interpret (e.g. Humphries and Cliff, 1982), were forgotten. The major difficulty with whole rock dating of gneiss complexes was that the samples had to be collected on the assumption that the samples were cogenetic and so this possibility tended to become a false reality. The breakthrough in unravelling this conundrum came with Bill Compston putting together the team that developed the revolutionary SHRIMP I instrument to date single zircons (e.g. Compston et al., 1984) and the age of ion microprobe zircon geochronology had began. For the first time this instrument allowed an individual sample to be independently dated using its constituent zircon crystals and then those results compared to another sample. Whether a series of samples were cogenetic no longer needed to be assumed, it could be established and so the technique has the advantage that once cogenetic suites are identified whole rock isotopic studies could be carried out in the safe knowledge that like was being compared with like. SHRIMP provided a key with which to take the next steps in understanding the protoliths of the TTG rocks in the Lewisian Gneiss Complex. Additionally, zircons may also preserve part of their subsequent metamorphic history through the growth of rims and recrystallisation zones. Again this information can be accessed using ion probe techniques and so more parts of the story can be unravelled. This technique is now routine in understanding the ages of the components of all gneiss complexes throughout the world. The contribution SHRIMP has made cannot be underestimated as it has led to a much better constrained data set for modelling crustal genesis and so giving an insight into the way that Earth has evolved.

2. Introduction

Using SHRIMP ion probe geochronology the Lewisian Gneiss Complex has been re-interpreted to comprise several tectonically-bounded terranes (Kinny & Friend, 1997; Friend & Kinny, 2001; Love et al. 2004; Kinny et al., 2005a) that preserve different protolith and metamorphic histories (Fig. 1). These studies have rightly proven controversial (e.g. Park, 2005a,b; Booroah and Bowes, 2009) as the terrane proposal challenges long-held views. None-the-less, whilst some of the details may be debated (Mason et al., 2004) and, inevitably that in the light of new mapping some modifications have been made (Goodenough et al., in press), the principle has stood up.

Following on from the earlier work (Fig. 1) and the nomenclature proposals of Kinny et al. (2005a), this contribution presents the ion-probe geochronological data from the region south of Gruinard Bay (Fig. 2) with the geochronology of important components of the southern portion of the Lewisian from south of Gruinard Bay to Loch Torridon. U-Pb zircon age constraints are presented for gneisses from some of the lower strain zones, and compared to samples from the Loch Maree Group and from gneisses further north. U-Pb titanite analyses have also been undertaken in an attempt to place constraints on the timing of Palaeoproterozoic amphibolite facies metamorphism. Results are discussed with reference to the terrane accretion model for the mainland Lewisian Complex (Kinny et al., 2005a) and how this impacts existing Palaeoproterozoic tectonic reconstructions in the North Atlantic (e.g. Park, 1994, 1995, 2005a).

3. Previous work and geological description

The Lewisian Gneiss Complex was historically the first Precambrian gneiss complex to be broadly understood through detailed field mapping as representing a section through the lower crust (e.g. Peach et al., 1907). This early work was refined by Sutton and Watson (1951) who, following the pioneering work in southern Greenland (Noe Nygaard, 1948), used the Scourie dyke suite to separate the mainland Lewisian gneisses into a region where dykes are undeformed (the Scourian), and regions where dykes have been strongly deformed and metamorphosed and the host gneisses reworked (the Laxfordian). Considered in terms of two orogenic cycles separated by a period of dyke emplacement, this evolutionary model remained in place for ca. 50 years, and formed not only the basis of subsequent Lewisian studies, but studies of other Precambrian gneiss terranes in general. It was generally understood that increasing grades of metamorphism were related to increasing pressure and temperature at depth, the positioning of pyroxene granulites and lower-grade amphibolite facies rocks along major shear zones in the mainland Lewisian was considered to represent juxtaposed layers in a vertical section of crust. However, doubts about the validity that the Lewisian Gneiss Complex represented a single unit of variably reworked crust were already in existence from field and geochemical studies for example, by Bowes (1962) and Sheraton et al. (1973).

The origins of the gneisses themselves generated much discussion. Peach et al. (1907) initially considered a plutonic origin, but later favoured the view that the gneisses were melted sediments a view also held by Sutton and Watson (1951) in their model for the complex. It was later suggested (e.g. Pidgeon & Bowes, 1972) that the gneisses were of felsic volcanic origin, a view supported for some units by the geochemical data of (Sheraton et al., 1973). However, it subsequently became accepted that the majority of the grey quartzo-feldspathic gneisses were deformed plutonic rocks of tonalite-trondhjemite-granodiorite (TTG) affinity (e.g. Bridgwater et al., 1973; Moorbath, 1977; Bowes, 1978; Rollinson and Windley, 1980). The acceptance that the main gneisses comprised largely plutonic TTG rocks (see summary in Park et al., 1994) indicated that there was a need for a reappraisal of the complex as a whole.

3. 1. The northern part of the Lewisian Gneiss Complex

The original subdivision of the mainland Lewisian Gneiss Complex was into ‘northern’, ‘central’ and ‘southern’ regions based primarily on structural and metamorphic differences (Sutton & Watson, 1951). With the application of modern precise SIMS U/Pb zircon techniques it was subsequently found that the type Scourian gneisses of the ‘central region’ had protolith ages of c.3030-2960 Ma (Friend & Kinny, 1995), while north of Loch Laxford the gneisses of the ‘northern region’ yielded protolith ages of 2840-2680 Ma (Kinny & Friend, 1997). Both Humphries & Cliff (1982) and Corfu et al. (1994) had both recognised an important isotopic event in the ‘central region’ at ca. 2490 Ma. Humphries & Cliff (1982) interpreted their age as cooling from granulite facies whilst Corfu et al. (1994) interpreted the U-loss to be the timing of the Inverian hydrous retrogression. However, this ca. 2490 Ma age signature was subsequently found in totally unretrogressed granulite facies gneisses at the two ‘type’ localities of Pidgeon and Bowes (1972) at Badcall Bay and Kylesku (Friend and Kinny, 1995). As this was inconsistent with the age representing retrogression it was consequently attributed to dating the granulite facies event itself. This was then became consistent with the Sm-Nd mineral ages obtained earlier by Humphries & Cliff (1982) from granulite facies assemblages. Subsequently, this ca. 2490 Ma age was also found throughout granulite and retrogressed granulite samples in the Asynt terrane (Kinny and Friend, 1997). As with the geochemistry of Sheraton et al. (1973), no evidence from the zircons supports the gneisses north of Loch Laxford ever being metamorphosed to granulite facies. Instead, the gneisses here have different protoliths and are intruded by a suite of granites at ca.1855 Ma, which are only present on the north side of the Laxford Front (Fig.1) but not to the south (Friend & Kinny, 2001; Goodenough et al., 2009). To account for the differences between the Assynt and Rhiconich terranes it was suggested that the two were unrelated pieces of crust, which had evolved separately prior to their juxtaposition along the Laxford shear zone. Together with titanite data from Corfu et al. (1994), a shared event at ca. 1740 Ma was identified in both terranes (Kinny & Friend, 1997), interpreted to provide a minimum age for the timing of amalgamation. A further lower-grade metamorphic disturbance is also recognised at 1670 Ma from resetting of the U/Pb system in rutile (Corfu et al., 1994) and from titanite dating (Love, 2004). This common metamorphism was called the Somerledian (Kinny et al., 2005a).

More recently it has been shown that the 'central' region (*sensu* Sutton & Watson, 1951) comprises two different granulite facies terranes, with distinct zircon age signatures for each (Love et al., 2004). In the Assynt terrane, the 2490 Ma Badcallian (*sensu* Kinny et al., 2005) granulite facies U-Pb isotopic system disturbance is recognised in zircon as far south as Lochinver (Fig. 1). South of Lochinver, variably retrogressed granulite facies gneisses contain zircons that yield a ca. 2730 Ma age signature, interpreted to be the timing of granulite facies metamorphism, defining the Gruinard terrane. Around Gruinard Bay, protolith ages for the gneisses are between 2860-2825 Ma and the same 2730 Ma metamorphic disturbance can be recognised in granulites as far north as the Strathan Line (Fig. 2). It has been proposed that the Gruinard terrane granulite facies metamorphism is called the Inchinaian (Kinny et al., 2005a). Additionally, there have been no ca. 2490 Ma overgrowths found on the zircons of the Gruinard terrane. The 2730 Ma Inchinaian event is thus considerably older than the 2490 Ma Badcallian (*sensu* Kinny et al., 2005a) event, with which it was once correlated – they cannot be the same metamorphic event. Therefore, it is interpreted that the Assynt and Gruinard terranes were unrelated until after ca. 2490 Ma. Presently, the best interpretation of the data is that the boundary between the two coincides with the amphibolite facies, Inverian-age Strathan Line described by Evans (1965) and Evans & Lambert (1974) as an important boundary between flat-lying rocks structurally above, and steeply inclined rocks structurally below. Consequently, a plausible explanation of the pre-Scourie dyke suite Inverian retrogression is that it represents the relatively localised effects of terrane amalgamation on the granulites of the southern margin of the Assynt terrane and the northern margin of the Gruinard terrane. These two terranes were already juxtaposed by the emplacement of the Scourie dyke suite (Love, 2004; Kinny et al., 2005a).

3. 2. The Lewisian Gneiss Complex south of Gruinard Bay

The Lewisian gneisses south of Gruinard Bay (Fig. 1) have been collectively referred to as the 'southern region' since the earliest subdivisions of the complex (e.g. Sutton and Watson, 1951; Park et al. 1994). The geology of the Lewisian Gneiss Complex between Gruinard Bay and Loch Torridon (Fig. 2) is well documented through numerous structural and geochemical studies (e.g. Peach et al.,1907; Sutton and Watson, 1951; Park, 1964, 1966, 1970, 2002; Moorbath and Park, 1971; Cresswell, 1972; Crane, 1978; Park et al., 1987; Holland and Lambert, 1995). The position of the Gruinard Front is rather ill-defined, having been mapped in different places. More recently, Park

(2002) describes the contact as an Inverian shear zone separating reworked gneisses to the south from relict Archaean structures to the north. The zone represents a transition in the Complex south of which the effects of retrogression and reworking become more pronounced until they are essentially total (e.g. Crane, 1978). Along the zone of most intense shearing lies a highly deformed band of supracrustal rocks known as the Gruinard Belt, which has been related to the Loch Maree Group through the Carnmore antiform (e.g. Park et al., 2001; Park, 2005b), although these rocks have also been proposed to be of Archaean in age (Park, 2002). Gneisses north of the Gruinard Belt are typically retrogressed to amphibolite facies and contain dykes with discordant margins and interpreted as equivalent to the Inverian (e.g. Davies, 1977). As post-dyke strain decreases, these rocks grade into pyroxene granulites at the eastern end of Gruinard Bay (e.g. Crane, 1978; Field, 1978; Corfu et al., 1998). Along the Gruinard Belt and further south towards Fionn Loch (Fig. 2), post-dyke deformation becomes intense with dykes rotated into parallelism with the northwest-trending subvertical foliation that typifies the region (e.g. Park, 2002). The rocks are characterised by strong NW-trending sub-vertical fabrics and large upright folds formed during amphibolite facies reworking of the Archaean gneisses (e.g. Park, 1964, 2002). Establishing the chronology of is segment of the complex has always been recognised as being difficult (e.g. Holland and Lambert, 1995). Previous isotopic studies of the 'southern region' gneisses using K-Ar (e.g. Evans and Park, 1965; Moorbath and Park, 1971) and Rb-Sr (e.g. Bikerman et al., 1975) have yielded typically imprecise ages spanning the Late Archaean and Palaeoproterozoic. Further, due to the susceptibility of these systems to disturbance (e.g. fluid access during faulting, ancient and recent weathering) and the use of non-cogenetic suites these ages are now of questionable value. Bulk zircon U-Pb studies on the gneisses provided Archaean ages for components in some of the low-strain augen (e.g. Lyon et al., 1973; Chamberlain et al., 1986). However, these were carried out before the complexities of zircon behaviour were understood and, neglected the potentially complex internal relationships between cores and metamorphic rims. Therefore, they reflect mixtures and bring the obtained ages into question.

3. 2. The Loch Maree Group

The 'southern region' also contains the largest exposures of supracrustal material in the mainland Lewisian, the Loch Maree Group. This is exposed in two belts, separated by the Loch Maree Fault (LMF) and the TTG gneisses in the Tollie antiform (e.g. Park, 1964, 1970b, 2002; Fig. 1). The Loch Maree Group comprises a sequence of interbanded, low amphibolite facies. Palaeoproterozoic amphibolites and metasedimentary rocks interpreted as volcanic, volcaniclastic and sedimentary rocks (Park, 1964, 2002, Park et al., 2001). In the eastern belt, early fabrics in the Loch Maree Group are folded about the Letterewe synform (LS Fig. 1) and brought into parallelism with a northwest-trending steep foliation. On the southwestern limb of the synform the belt is truncated by the LMF. The western belt is bounded on both sides by gneisses with highly tectonised contacts. At

Gairloch, the dip of the Loch Maree Group suggests that it may form part of a corresponding synform to the Tollie antiform that has been sheared-out along the Creag Bhan belt (e.g. Park, 1970b, 2002). The north-eastern side is a zone of mylonitised Loch Maree Group units and TTG gneiss which grades into intensely deformed TTG gneiss (Buainichean gneiss, Park, 1964) coincident with the sheared southern limb of the Tollie antiform. Internally this belt also includes a completely shear-bounded tectonic slice of gneisses interpreted to represent Scourian basement, known as the Ialltaig gneiss (Park 1964). This block comprises largely retrogressed granulite facies rocks with rare relict orthopyroxene and conventionally is considered a tectonically isolated slice of Archaean basement correlated with the Scourian granulites (e.g. Park 1964, 2002). The Gairloch shear zone is defined by Park (2002) as comprising the entirety of the rocks south of the southeast limb of the Tollie anitiform to the reworked Archaean basement gneisses to the southwest of Gairloch. A significant shear zone is present within the southern margin of the Loch Maree Group, which divides to contain the Ialltaig gneiss block (e.g. Park, 1964, 2002). This shear zone, named here the Kerry Shear Zone (Fig. 2), occurs just to the north of the Shieldaig shear zone that separates the Loch Maree Group from the Archaean gneisses of the Rona terrane.

Early Sm-Nd modelling provided indications of an Archaean component (O'Nions et al., 1983), and an ion microprobe detrital zircon study on a metasedimentary unit in the Loch Maree Group identifying Proterozoic (2.2-2.0 Ga) and Archaean (3.06-2.48 Ga) components (Whitehouse et al., 1997). An important constraint on the age of the Loch Maree Group was the dating of the Ard Gneiss, a discordant body of augen gneiss (Park, 1964) that has yielded a precise TIMS zircon protolith age of 1903 ± 3 Ma (Park et al. 2001). This age is interpreted to date the timing of emplacement of the granodioritic protolith into the Loch Maree Group during peak metamorphic conditions (Park et al. 2001).

The work of Park et al. (2001) and Park (2002) suggested the Loch Maree Group protoliths are intercontinental in origin, comprising both oceanic and terrestrial rocks and interpreted in terms of a subduction-related collisional model whereby basaltic oceanic crust has been accreted to the base of an overriding plate and tectonically imbricated with oceanic and terrestrial sedimentary material.

Adjacent to the south side of the Loch Maree Group is an enigmatic block of gneisses that show retrogressed granulite facies textures and mineralogies called the Ialltaig gneiss (Park, 1964, 2002). Even though the Ialltaig gneiss occurs as a totally shear-bounded block juxtaposed with amphibolite facies rocks, because of indications of granulite facies assemblages it was described as being Scourian (Park, 1964, Park, 2002).

3. 3. The Lewisian Gneiss Complex south of the Loch Maree Group

Apart from the early work of Sutton and Watson (1951), the gneisses of the Torridon have been the subject of a number of field studies (e.g. Cresswell, 1972; Park et al., 1987; Wheeler et al., 1987; Naimitullah and Park, 1990). These latter have been mainly concerned with elucidating the Proterozoic structural history of the area. Intense Palaeoproterozoic aged deformation of the 'southern region' has folded the Loch Maree Group and surrounding TTG gneisses together, largely destroying earlier structures and fabrics except in small augen of lower strain (e.g. Park 2002; Fig. 2). The models proposed by Park (2005a,b) suggest that there are two different Archaean crustal plates with Palaeproterozoic material trapped between them. Therefore, evaluating the origins of the gneisses bounding the Loch Maree Group is an important step in determining how the southern mainland Lewisian evolved during the Archaean and Palaeoproterozoic. Similar ages and metamorphic histories for the gneisses either side of the Loch Maree Group would support a model in which there was a rifting phase in the history of the 'southern region'. Alternatively, if the gneisses bordering the Loch Maree Group were of different origins, the 'southern region' might be an amalgamation of several different pieces of crust.

South of Gairloch (Figs. 1, 2) the highly deformed gneisses along the Shieldaig shear zone become progressively less deformed the further south of the Loch Maree Group one goes. Nearest the Loch Maree Group interlayered tonalitic gneisses and dykes are both foliated, whereas several hundred meters further south only the gneisses are highly strained, which also reveals that the pre-dyke state of the gneisses is very variable. Decreasing post-dyke strain grades into the Ruadh Mheallan block, where predominantly pre-dyke structures are preserved (e.g. Cresswell, 1972; Park, 2005a). South of the Ruadh Mheallan block, post-dyke strain isolates numerous smaller low-strain augen, (Chamberlain et al., 1986; Wheeler et al., 1987), which are typically bordered by dykes truncating the internal fabric but lying parallel to the main external foliation.

With the exception of relict orthopyroxene in the Ialltaig gneiss (Park 1964), no granulite facies assemblages were reported from the low strain zones south of the Loch Maree Group. Retrogression to amphibolite facies assemblages appears complete, presumably occurring in conjunction with the development of the external high-strain fabrics. However, evidence that these low-strain augen have been metamorphosed to granulite facies, as has been the traditional assumption, is reliant upon the recognition of opalescent blue quartz and the rather nebulous texture of some of the gneisses. Using some of the textural criteria for retrogressed granulites (McGregor and Friend, 1997), the current study has recognised dehydration partial melt veins in some of the thin bands of mafic gneiss/amphibolite, which are identical to those observed in the granulite facies gneisses east of Gruinard Bay (Love et al. 2004). Gneisses in other low-strain augen are clearly metatexitic, and contain both intrusive and *in situ* melt relationships.

4. Analytical procedures

U-Pb analyses were performed using the sensitive high-resolution ion microprobe mass spectrometer SHRIMP II located at the John de Laeter Centre for Mass Spectrometry at Curtin

University of Technology. Routine procedures for SHRIMP U-Pb analysis were observed (Nelson 1997). A 2-4 nA primary beam was typically used for all zircon analyses, with the spot size ranging between 20 and 30 μm. Mass resolution in excess of 5000 (1% valley), and sensitivity for Pb isotopes in the range of 12 to 15 cps ppm^{-1} nA^{-1} are typical. Masses 196 (Zr_2O), 204 (^{204}Pb), 204.1 (background), 206 (^{206}Pb), 207 (^{207}Pb), 208 (^{208}Pb), 238 (U), 248, (ThO) and 254 (UO) were measured sequentially by peak stepping, over seven cycles of the mass spectrum. Common Pb corrections on isotopic ratios were based on the measured ^{204}Pb, which typically result in less than a 3 % correction to the ^{206}Pb counts in low to moderate U zircon (%c206), with the composition of the common Pb modelled on Broken Hill ore Pb. Pb/U isotopic ratios were corrected for instrumental inter-element discrimination using the observed covariation between $^{206}Pb/^{238}U$ and UO/U (Compston et al. 1984, 1992) obtained from regular analyses of the Perth standard zircon CZ3 (564 Ma; $^{206}Pb/^{238}U$=0.0914).

For titanite analysis similar operating conditions were observed as for zircon. Masses 200 ($CaTi_2O_4$), 204 (^{204}Pb), 204.1(background), 206 (^{206}Pb), 207 (^{207}Pb), 208 (^{208}Pb), 248 (ThO), 254 (UO) and 270 (UO_2) were measured (Kinny, 1997) by peak stepping over seven cycles for each analysis. The co-variation between ^{206}Pb/UO and UO_2/UO was used to correct for inter-element fractionation determined from interspersed analyses of the Khan titanite standard (518 ± 2 Ma; $^{206}Pb/^{238}U$ = 0.083671, (Kinny et al., 1994). Isotope ratios and their corresponding ages calculated from the standard decay constants of (Steiger and Jager 1977) are presented in Table 1.

Where analyses have been grouped a chi-square test was applied in order to assess the relative effects of analytical sources of error, such as counting statistics, and possible geological sources of error. A chi-square of 1 or less indicates that scatter about the weighted mean determined from the grouped analysis can be accounted for by analytical error alone. Chi-square values in excess of 1 indicate that analyses are not normally distributed about the weighted mean and that other sources of geological error are present within the grouped population. In these cases, the 95 % confidence error is based on the observed, rather than the expected scatter about the weighted mean $^{207}Pb/^{206}Pb$ ratio of pooled analyses. The complex nature of the majority of zircons means that ages quoted in the text consistently have chi-square values greater than unity. The variable resetting of recrystallised zircons has resulted in a more restricted use of statistical population analysis for the typical spreads of age data. Chi-square values of less than 2.5 have been taken as the limit for age quotation, with errors in data tables and concordia diagrams expressed at the 1σ level.

5. Sample details and results

5.1. Gruinard Bay and north of the Loch Maree Group

Sample GL01/12 is a banded tonalitic gneiss from the road cutting at the south-western part of Gruinard Bay (NG 940902), in the region where granulite facies gneisses have suffered pre-dyke amphibolite facies retrogression but are relatively unaffected by post-dyke deformation and metamorphism (Fig. 2). While studies of the Gruinard Bay area have identified the gneisses to be predominantly trondhjemitic in composition, tonalitic gneisses are also present, often recognised by more pronounced banding due to their higher mafic mineral content (e.g. Fowler, 1986; Whitehouse et al., 1997b). The tonalitic gneisses have been described as forming agmatite structures within the trondhjemitic gneisses (e.g. Crane, 1978), which implies the banded tonalitic gneisses were formed (and presumably metamorphosed) prior to the intrusion of the trondhjemitic gneisses (Rollinson and Fowler, 1987). In the main road cutting large amphibolite blocks (agmatite) are clearly contained within trondhjemitic gneisses, however the relationship between banded tonalitic gneisses and trondhjemitic gneisses often less clear. There are areas where fairly uniform gneisses (considered to be trondhjemitic) grade into areas containing diffuse mafic banding which could be considered tonalitic due to its higher mafic mineral content. Differences between banded tonalitic gneiss and homogenous trondhjemitic gneiss are complicated by areas where in situ partial melting in the gneisses appears to have generated additional localised trondhjemitic melts. GL01/12 is considered to be representative of the “banded tonalitic gneisses” that are apparently intruded by the protoliths of the more abundant trondhjemitic gneisses. A sample of homogeneous trondhjemitic gneiss from the same cutting has been analysed previously (Love et al., 2004, sample GL00/07) by the same technique and provides an excellent comparison to GL01/12, which represents the most distinct sample of banded tonalite that could be obtained.

Zircons from GL01/12 are pale pink ovoid grains with glassy rims distinguishable from slightly turbid cores. Cathodoluminescence (CL) imaging reveals cores of variably preserved, oscillatory-zoned, primary igneous zircon surrounded by brighter luminescent rims (Fig. 3). The rims mostly exhibit a brighter CL response and contain homogenised oscillatory igneous banding (transgressive zones) identical to that recognised in zircon from granulites and retrogressed granulites elsewhere in the Lewisian (Friend and Kinny, 1995; Love et al., 2004). Most of the zircons in sample had bright zircon rims in comparison to GL01/12, an important distinction as the samples are spatially close together. Results from tonalitic gneiss GL01/12 (Table 1) show a significant spread of concordant and discordant age data (Fig. 4). Three of the measurements from oscillatory-zoned cores form a cluster that is interpreted to record the timing of igneous protolith emplacement, estimated at ca. 2860 Ma from $^{207}Pb/^{206}Pb$ ratios. The rims show a full spread of concordant and near-concordant analyses with only a few preserving the 2730 Ma signature for the Gruinard terrane granulite facies metamorphism (Love et al., 2004). A few show even younger ages interpreted to reflect resetting. In comparison, the trondhjemitic gneiss analysed previously under identical conditions (Love et al 2004), contained almost ubiquitous resetting of the U/Pb system in zircon rims, despite being sampled only several metres away from GL01/12 (Fig. 4 upper inset).

North of Gairloch, tonalitic gneiss sample GL00/06 was taken from a small quarry on the A832 within the Tollie antiform (NG 832782). This sample has two zircon suites. First, there are pale-pink, oval-shaped zircons with oscillatory-zoned cores, and recrystallisation rims consistent with those above and elsewhere in the Lewisian (Fig. 3). The second suite are pale pink >200 μm irregular fragments of much larger zircons probably broken during sample processing. Using CL there are regions of oscillatory zones, and broad regions exhibiting transgressive zoning. Both suites have been combined for SHRIMP analyses.

U-Pb zircon determinations for the sample GL00/06 (Table 2) form two distinct age populations corresponding to measurements from well-preserved oscillatory-zoned zircon, and recrystallised zircon (Fig 4). Analyses show a typical age spread associated with Lewisian granulites and ex-granulite facies gneisses, with a clustering of points at either end of a discordia line sub-parallel to concordia. Measurements forming the older cluster are typically several hundred ppm U, with Th/U of less than 1, and form a strong grouping at ca. 2825 Ma (max $^{207}Pb/^{206}Pb$ ages 2840 ± 6 Ma). The younger analyses are lower in U (<100 ppm), with measurements from the most intensively recrystallised rims having Th/U of >1.7 and ages consistent with a major metamorphic disturbance at ca. 2730 Ma. For comparison a previously analysed tonalitic granulite from the Gruinard Terrane (Love et al., 2004, GL00/09) is presented (Fig. 4 lower inset) showing the strong similarities in protolith and metamorphic age signatures.

5.2. South of the Loch Maree Group- the Rona terrane

To characterise the gneisses south of the Loch Maree Group, low-strain augen were investigated near Gairloch (Park, 1964), Diabaig (Cresswell, 1972) and Kenmore (Niamitullah and Park, 1990). Since these low strain augen preserve the pre-dyke history of the gneisses better, they provide the best targets for geochronological studies. North of Diabaig, from the low strain area examined by Creswell (1972), sample GL00/03 of homogeneous tonalite with apparent intrusive characteristics was obtained at NG 829598, and a sample of an older phase of banded tonalitic gneiss, GL00/04, at NG 830604 (Fig.2).

Zircons from GL00/03 are glassy, light brown and prismatic, and often accompanied by iron oxide development along fractures in the grains. Under CL imaging the zircons luminesce poorly (Fig. 3), showing oscillatory zoning and only one grain with an observable core and rim relationship. High common Pb in the analyses (Table 3) has hampered determination of a precise age from sample

GL00/03, the points defining a discordia with an imprecise concordia upper intercept of ca. 2700 Ma (Fig. 5). U content ranges from several hundred to >1000 ppm, and Th/U values are consistently around 1 or less. Despite being unable to define a precise age, the ca. 2700 Ma estimate corresponds with the 2697 ± 10 Ma age for rims from gneiss sample GL00/04, with which it appears to have an intrusive relationship, and so is interpreted as reflecting a metamorphic event.

Titanite was also found in GL00/03 (Table 4). These are brownish grains and on analysis showed three Archaean-aged grains of ca. 2550 Ma in age, while the dominant population of 20 grains formed a strong cluster around 1660 Ma (Fig. 7). Th/U ratios are typically <1 (including the Archaean grains), and all ages are thought to reflect metamorphic cooling. The older grains have probably been shielded from younger resetting by more refractory phases in the gneiss.

In sample GL00/04, zircons are prismatic to ovoid, and pale pink to brown in colour. CL shows oscillatory-zoned protolith zircon, with well-developed rims that generally give a darker CL response than cores (Fig. 3). The embayed nature of the cores and the presence of thin bright reaction fronts suggests that the rims are melt grown rather than formed by solid-state processes. Analyses from sample GL00/04 (Table 5) shows two age groupings, corresponding to zircon cores and rims (Fig. 5). Core analyses are typically a few hundred ppm U and have Th/U > 0.4, with the four oldest ^{207}Pb/^{206}Pb measurements yielding a weighted mean ^{207}Pb/^{206}Pb age of 2881 ± 10 Ma. Zircons rims are typically enriched in U (several hundred ppm to >1000 ppm), and have Th/U ratios of <0.25. Eight measurements form a slightly discordant array with a mean weighted ^{207}Pb/^{206}Pb age of 2697 ± 10 Ma, interpreted to date the addition of melt-grown zircon.

On the coast 300m south of the Diabaig pier, the rocks comprise metatexites and inhomogenous diatexites developed from tonalitic and granitic gneiss with both in situ melting and intrusive relationships preserved. Samples of the least melted tonalitic gneiss (GL01/01) and granitic melt (GL01/02) both come from NG 799596. GL01/01 has large pale-pink glassy zircons that are typically subhedral, with well-developed crystal faces and blunted terminations. They show fine oscillatory zoning in CL images (Fig. 3) with occasional fine thin rims of new zircon growth, thought to be associated with the melt intrusion. Zircons from GL01/02 are brown and typically elongate, distinctly different in appearance to those from the host tonalitic gneiss. Under CL they appear to have corroded cores and broad dark rims (Fig. 3) that are interpreted to reflect new melt-grown zircon.

Zircons from sample GL01/01 yielded a relatively simple data set (76-252 ppm U, Th/U < 1; Table 6) with only 4 obviously discordant analyses, and produced a mean ^{207}Pb/^{206}Pb age of 3135 ± 5 Ma from 12 analyses (Fig. 5). The discordant grains may have suffered minor Pb-loss during ancient times, or undergone disturbance during the melting episode. Sample GL01/02 forms the granitic component of the metatexite containing GL01/01. Analyses of zircon cores in GL01/02 yield ages in excess of 3000 Ma (Table 7) interpreted from similar U content and Th/U ratios, as being direct inheritance from the tonalitic gneiss. Oscillatory-zoned melt-grown zircon contains several hundred ppm U, variable amounts of common Pb, and most analyses have Th/U <10.5. A concordant cluster of 8 analyses yielded a mean $^{207}Pb/^{206}Pb$ age of 2953 ± 8 Ma (Fig. 5).

Subsequent, lower-grade metamorphic events were assessed using U-Pb analyses of titanite discovered amongst the accessory phases. Eighteen U-Pb analyses from sample GL01/01 (Table 8) defined a mean $^{207}Pb/^{206}Pb$ age of 1670 ± 12 Ma with no outliers (Fig. 7). A low U content (11-203 ppm) leads to large individual error boxes, although the fairly uniform Th/U (0.12-0.28) and lack of outliers provides strong evidence for either crystallisation in a metamorphic event or, pre-existing titanite that was uniformly reset.

5.3. Age determinations within the Loch Maree Group

Within the Loch Maree Group the Ard gneiss is an intrusive porphyritic granitoid that constrains the early deformation (e.g. Park, 1964) and has yielded as age of 1903± 3Ma (Park et al., 2001). It is thus and important marker regarding constraints on the regional metamorphism and structure. The Ialltaig block, comprising retrogressed granulite facies gneissic rocks, which is regarded as a tectonic slice of Scourian rocks and so Archaean in age (e.g. Park, 1964, 2002) is clearly of importance in constraining the extent of possible Archaean gneisses in proximity to this accepted piece of Proterozoic crust .

5.3.1. Ialltaig gneiss

Regrettably, all of the gneisses in the Ialltaig block are at least partially retrogressed (e.g. Park, 1964, 2002) and so the zircons are open to disturbance. Sample GL00/05 (NG 806734) is as little retrogressed from granulite facies gneiss as could be obtained from the Ialltaig block. The sample is nebulitic with quartz-seived hornblende interpreted as pseudomorphs after orthopyroxene (e.g. McGregor and Friend, 1997) and contains trace garnet as reported by Park (2002). Sample GL00/05 has pale pink to brown, glassy zircons that are ovoid, and under CL have observable cores and rims (Fig. 3). Oscillatory zones are common in the cores, while homogeneous rims of variable CL response occur, often accompanied by transgressive zones.

Th/U ratios in the protolith zircon in sample GL00/05 are less than 1, and show a systematic decrease with decreasing age (Table 9). The zircons yielded a range of $^{207}Pb/^{206}Pb$ zircon ages from ca. 2052 ± 80 Ma to ca. 1855 ± 7 Ma (Fig. 6). The ages obtained from the cores are too scattered to make a precise determination from, but indicate a minimum protolith age of ca. 2050 Ma. Homogenous rims of varying U content (generally >100 ppm) are well-developed on many zircon grains, preserving evidence for recrystallisation, with 7 measurements producing a mean $^{207}Pb/^{206}Pb$ age of 1877 ± 13 Ma and this is interpreted to date the granulite facies metamorphism. The granulite facies event is thus presumed to have taken place some 200 Ma after the emplacement of the protoliths leaving considerable room for debate over the early deformational and metamorphic history of these gneisses and their entrained mafic enclaves.

5.3.2. Ard gneiss

A sample of the Ard Gneiss from near Charlestown pier, sample GL01/15, NG 806750, is an augen granitoid developed from a coarse-grained, porphyritic protolith (see Park et al., 2001). GL01/15 yielded zircons that are pale pink to brown and prismatic, often with sharp pyramidal terminations. They appear uniformly dark in CL images and have well-developed oscillatory zoning (Fig. 3). Titanites were also obtained from the Ard gneiss (GL01/15). All three titanite suites are pale-brown in colour and xenoblastic in character, suggesting a metamorphic origin. Titanites from the Ard gneiss (GL01/15) are considerably less inclusion rich than the other two samples and are slightly darker in colour. Backscatter electron (BSE) imaging of titanites from all samples shows that their internal morphology is without growth structures, giving a uniform response.

Results from zircons in sample GL01/15 of the Ard gneiss showed a slight amount of reverse discordance (Table 10), with the 6 most concordant analyses of oscillatory zoned zircon yielding a mean ^{207}Pb/^{206}Pb age of 1900 ± 11 Ma (Fig. 6). The U content in the zircon is fairly uniform (334-506 ppm), as are Th/U ratios (0.19-0.29). This age is indistinguishable from that obtained earlier for the Ard Gneiss (Park et al. 2001).

Titanites from the same sample were a homogeneous pale brown colour and showed even luminescence in CL. They yielded an age of 1890 ± 13 Ma for 16 analyses (Fig. 6 inset) and, from fairly uniform U contents (40-84 ppm) and Th/U ratios (4.8-7.3), are thought to reflect cooling of the original igneous titanite (Table 11).

6. Discussion

Two major problems considerably hampered the earlier debates over the origins of the gneisses comprising the mainland Lewisian Gneiss Complex. The original work (Peach et al., 1907) and the subsequent work of Sutton and Watson (1951) relied solely upon lithologies, structure and the metamormorphic state of the rocks for correlation, which of course was all that was then available. Consequently, the correlations of the different rock groups made were always suspect. This suspect base was then compounded by a series of whole rock and bulk mineral isotopic data, which in reality had no validity, being obtained to corroborate the field model without any major reassessment. For example, in the bulk zircon U/Pb dating study of Pidgeon and Bowes (1972), the gneisses were interpreted to have had felsic volcanic protoliths that contained no primary or inherited zircon. The age of the granulite facies was deduced on the basis that all of the zircons had grown during the metamorphism. Once it was later accepted that the gneisses were deformed TTG plutonic rocks this possibility of no primary zircon became extremely unlikely because most ordinary TTG protoliths contain copious primary igneous zircon. Nonetheless, the granulite facies ‘age’ derived from the Pidgeon and Bowes (1972) work was carried through the subsequent literature (e.g. Chapman and

Moorbath, 1977; Lyon and Bowes, 1977; Whitehouse, 1988). It took a radical reappraisal to demonstrate that the 'age' obtained by Pidgeon and Bowes (1972) had no geological significance (Friend and Kinny, 1995). This reappraisal of the components Lewisian Gneiss Complex has been continued here with further demonstrations that some of the old assumptions had no validity.

6. 1. Gneisses of the Gruinard Bay area

The gneisses in the southern part of Gruinard Bay had previously been dated using very carefully collected suites of co-genetic rocks (Whitehouse et al., 1996) and confirmed using ion probe techniques (Whitehouse et al., 1997b). These studies yielded protolith ages of ca. 2800 Ma for the tonalitic gneisses. This set of ages was broadly corroborated by Love et al. (2004), although in this latter study unretrogressed granulite facies rocks were dated which gave a somewhat simpler data set, yielding an age of ca. 2730 Ma for the event and the preservation of a slightly older ca. 2860 Ma magmatic protolith age. This age, found in some zircon cores from GL01/12, is indistinguishable from the 2858 ± 11 Ma protolith estimate (Love et al., 2004) suggesting that these tonalitic protoliths were probably emplaced during the same period of crust generation from ca. 2860-2800 Ma. GL01/12 does not contain any of the 2905 ± 15 Ma inherited zircon that forms such a large component in the zircon suite from sample GL00/07, which is an important difference.

All of the gneisses in the Gruinard Bay area contain the 2730 Ma Inchinaian metamorphic age signature (Love et al., 2004; Kinny et al., 2005a,b). However, some tonalitic and trondhjemitic gneiss samples in this study have yielded younger protolith ages of ca. 2825 Ma (see also Love et al., 2004), which are closer in age to those recorded by Whitehouse et al. (1997b). It thus appears that the protoliths of these samples appear to have been generated later than those of apparently similar granulite facies gneisses north of Gruinard Bay, and further highlight the ambiguity between tonalitic and trondhjemitic gneisses.

Present work does not suggest any major boundary running through Gruinard Bay but the shearing separating the low strain from higher strain areas is difficult to evaluate. Definite separation of the gneisses into trondhjemitic or tonalitic compositions is also difficult, since metamorphism, partial melting and retrogression have all had an effect on the original protolith relationships.

6.2. The southern margin of the Gruinard terrane

Presently, there is a major change in the geology along the line of the Shieldaig shear zone on the southern side of the Loch Maree Group (Fig. 2) and so the Gruinard terrane rocks are certainly north of that. The problems in determining the exact position of the boundary are exacerbated because of the dissection of the Loch Maree Group into two outcrops by the Loch Maree Fault and the intervening Tollie antiform. Additionally, there is essentially total retrogression of the gneisses. The

Tollie tonalitic gneiss sample shows an age pattern consistent with that of the granulite facies tonalitic gneisses from the Gruinard Terrane (Love et al., 2004; this paper), with protolith zircon yielding maximum $^{207}Pb/^{206}Pb$ ages of ca. 2825 Ma. The younger ages correspond to recrystallised zircon domains, and show a signature consistent with ca. 2730 Ma metamorphic ages from the Gruinard Terrane (Kinny et al., 2005a). Metamorphic and protolith age data from GL00/06 are interpreted to show an isotopic link between the Tollie gneiss and Gruinard Terrane, despite any evidence for granulite facies metamorphism at Tollie being destroyed by amphibolite facies reworking. However, preserved granulite facies, orthopyroxene-bearing gneisses have been reported by Park (1964, 2002) in a region of lower post-dyke strain on the northeast side of the Tollie antiform, southwest of Loch Maree. This was interpreted as being a relict of an Archaean granulite facies event (Park 2002). The evidence presented here would tend to support this being the ca. 2730 Ma granulite facies event observed in Gruinard Bay (Love et al., 2004).

Linking the Tollie gneiss with the Gruinard Terrane does, however, complicate current structural interpretation of the 'southern region' gneisses. The problem involves the interpretation of the Gruinard Belt and its relationship with the Loch Maree Group. Two views have been expressed which are either that the supracrustal rocks of the Gruinard Belt are possibly Archaean (see Park, 2002, Fig. 1.3), or that they are equivalent to the Loch Maree Group, i.e. Palaeoproterozoic (Park et al., 2001; Park, 2005b). In the cross-section of Park et al. (2001) the Gruinard Belt is considered to be an attenuated extension of the Loch Maree Group, forming part of the NE limb of the Carnmore Antiform (Fig. 8). If this case is considered, the boundaries between the Loch Maree Group and the surrounding gneisses would also be folded over the Carnmore antiform, and the Gruinard Belt could represent an important tectonic boundary. The implications of this interpretation, following Park (2005b) are that the Gruinard terrane lies structurally above the Loch Maree Group. Therefore, the TTG gneisses in the core of the Letterewe synform would most likely be from the Gruinard Terrane, and more importantly, that the gneisses in the Carnmore antiform, structurally underneath the Loch Maree Group, may be different from those at Tollie and Gruinard. These possibilities are all capable of being tested using geochronology.

At Tollie, the TTG gneisses appear to be to same as the Gruinard terrane recording an event equated to the ca. 2730 Ma granulite facies metamorphism. The boundary of the TTG gneisses with the Loch Maree Group is heavily tectonised, and the associated synform to the Tollie antiform has been sheared out along the Creag Bhan belt of Park (1964, 2002). The traditional interpretation of the Loch Maree Group at the location of the Tollie antiform, was that it lay at a structurally higher level, above the present land surface, and was now unseen (e.g. Park et al., 2001; Park, 2002, 2005b). If the interpretation above, that the rocks are equivalent to the Gruinard terrane and structurally above the Loch Maree Group is correct, the implication is that within the Tollie antiform the Loch Maree Group rocks are below the surface not above (Fig. 8). Alternatively, the possibility that the Gruinard Belt comprises highly strained Loch Maree Group that was simply invaginated into the Gruinard terrane cannot be ignored. In this case the 2730 Ma granulite facies even could be found both above and below the Loch Maree Group in the Tollie anitiform. This implies that there are no constraints upon the structural position of the Loch Maree Group and it could be either at a deeper level or be structurally above the current level of exposure.

It is certainly possible that the supracrustal rocks in the Gruinard Belt could be completely unrelated to the Loch Maree Group, as suggested by Park (2002, Fig. 1.3). On this basis they would simply represent material caught up in the Archaean TTG gneisses probably during emplacement rather than during later deformation, as in other high-grade gneiss complexes (e.g. southern West Greenland or Labrador). The small amount of exposed supracrustal material in the Gruinard Belt contains deformed amphibolite of more than one generation, intercalated with mica schists and quartzite (e.g. Corfu et al., 1998), and despite being clearly supracrustal material, there is no strong lithological resemblance to many of the other lithologies in the Loch Maree Group (e.g. calc-silicates, quartzo-feldspathic intrusive rocks). While the Gruinard Belt is deformed by post-dyke events, it also lacks some of the higher-deformation features of the Loch Maree Group such as broad mylonite belts, and semi-brittle “crush belts”.

The regional deformation is constrained by the emplacement of mafic dykes correlated with the Scourie dyke suite (e.g. Park, 2002), and also more reliable whole rock Rb-Sr age of ca. 1660 Ma for the Tollie pegmatites (Holland and Lambert, 1995) corresponding with a zircon age of 1694 ± 5 Ma (Park et al., 2001).

6.3. The Loch Maree Group

The geology of the Loch Maree Group has become clearer over recent times with the understanding that the rocks have affinities to the ca. 1900 Ma collisional orogenic complexes of Fennoscandia (Park, 1994, 2002; Park et al., 2001). The geochronology has been fairly well established with detrital zircons indicating that Archaean crust 3060-2480 Ma was contributing some sediment and that Palaeoproterozoic grains provide the minimum age of deposition at ca. 2000 Ma (Whitehouse et al., 1997a). The early tectonic evolution of the Loch Maree Group is constrained by the 1903 Ma age for the Ard gneiss, which only shows the latest phase of deformation (Park et al., 2001). This age is corroborated here with both single zircon and titanite ages. With these reasonably tight constraints there would appear to be little scope for any major argument over the main sequence of events.

6. 4. Ialltaig gneiss

There is, however, considerable room for debate over the Ialtaig gneisses (Fig. 2). This low strain block within the southern margin of the Loch Maree Group southeast of Gairloch is surrounded by mylonitic shear zones that separate it from the enclosing amphibolite Loch Maree Group rocks and is described as displaying well-preserved Scourian structures (e.g. Park, 1964, 1973, 2002). Thus the traditional interpretation was that it represented a piece of retrogressed granulite facies Archaean basement. The data presented here demonstrate that major problems can arise due to incorrect correlation of individual blocks on the basis of metamorphic assemblage and structures alone, showing again that it is imperative that geochronology is also utilised. This allows the Ialltaig block to be recognised as an exotic slice of high-grade Palaeoproterozoic crust tectonically isolated within the southern margin of the amphibolite facies Loch Maree Group supracrustal rocks. The oscillatory-zoned cores of the analysed zircons yielded maximum ages of ca. 2000 Ma, which is interpreted to be an estimate of the protolith age, while recrystallised rims of lower U zircon yielded an age of 1877 ± 13 Ma, interpreted to represent the timing of granulite facies metamorphism. The Ialltaig gneiss is thus shown to be an independent piece of crust that is unrelated in age to any of the dated Loch Maree Group components (Love, 2004). The question is when was the Ialltaig terrane emplaced along the structure referred to here as the Kerry shear zone? There is no debate that the metamorphic grade of the Loch Maree Group has never exceeded low amphibolite facies conditions (e.g. Droop et al., 1999). The Ard gneiss constrains the peak metamorphism to have occurred before ca. 1900 Ma (Park et al., 2001). Since the 1877 ± 13 Ma granulite facies metamorphism of the Ialltaig gneiss is younger than this event the tectonic intercalations that took place in this area can only be post-1877 Ma. The evolutionary model of tectonic thickening suggested by Droop et al. (1999) is perfectly acceptable for the 'early Laxfordian' structures described because of the extremely well-constrained metamorphic data. However, this event cannot be that responsible for the metamorphism of the Ialltaig gneisses as it is too early. Whilst it is possible that it could be a piece of juvenile TTG continental crust from the magmatic arc that formed at ca. 2000 Ma, the metamorphism appears to imply a granulite facies event that is much younger.

This ca. 1877 Ma high-grade event might appear close to ca. 1870 Ma suggested for the granulite facies metamorphism in the juvenile Palaeoproterozoic rocks in the Leverburgh Belt of the Roineabhal terrane on South Harris (e.g. Cliff et al., 1983; Friend and Kinny, 2001; Whitehouse and Bridgwater, 2001). However, this correlation causes problems with matching the major gneiss units across the Minch Fault and the Outer Hebrides Fault Zone (Friend and Kinny, 2001) and needs further geochronological data to clarify the position.

6.5. Gneisses south of the Loch Maree Group – the Rona Terrane

The gneiss complex on the southern side of the Loch Maree Group at Gairloch shows a quite different history to the TTG gneisses from Tollie, and the Gruinard Terrane as a whole, and here are collectively termed the Rona Terrane. The main fabrics observed are all related to post-dyke amphibolite facies shearing (e.g. Cresswell, 1972; Park et al., 1987; Wheeler et al., 1987) that has caused considerable modification of the pre-dyke rocks. Only in rare parts of some low strain regions have any hint of earlier metamorphic assemblages been preserved.

In a low-strain zone along the coast at Diabaig (Fig. 2), tonalitic gneiss GL01/01 yielded a protolith age of 3135 ± 5 Ma and forms the host rock that melted and was back-veined by granitic melt produced at 2955 ± 8 Ma. The zircons from the tonalitic gneiss showed limited development of metamorphic rims thought to have developed during anatexis and granitic melt intrusion. The palaeosome components in GL01/01 and GL01/02 thus represent some of the oldest components of the Lewisian Gneiss Complex. It is plausible that the protolith ages recorded are similar to some of the oldest grains from the Assynt Terrane. However, whilst there are several generations of melt veins intruding the Assynt Terrane gneisses they are largely trondhjemitic and not granitic (e.g. Rollinson and Windley, 1980) and were not generated at ca. 2955 Ma, which appears to be a new metamorphic event.

In the low strain zone north of Diabaig that links to the Ruadh-Mheallan block the zircons from GL00/03 show marked recent Pb-loss with an imprecise ca. 2700 Ma concordia upper intercept the only age constraint that could be obtained. Banded tonalitic gneiss GL00/04 yielded an age for zircon cores of 2881 ± 10 Ma, with an age of 2697 ± 10 Ma obtained from recrystallised rims. The apparent intrusive nature of GL00/03 and the development of rims in GL00/04 suggest that high-grade metamorphic conditions may have prevailed at ca. 2700 Ma, at which time GL00/03 was emplaced.

The low strain zones that were investigated on the north side of Loch Torridon do not show a strong correlation in age data, but have yielded protolith ages that are considerably older than the gneisses to the north of the Loch Maree Group. The strong deformation and lack of consistent isotopic signatures means that no further sub-division of the Rona Terrane is possible at this time. Since the Gairloch shear zone of Park (2002) is a large complex structure affecting both the Gairloch and Rona Terranes, the boundary between these two terranes lies within it. Park (1964, 2002) mapped a zone of high-strain TTG gneisses at the very edge of the Loch Maree Group, which is named the Shieldaig shear zone (Fig. 2), and is taken as the line of division. Torridonian rocks obscure much of this area and it is thus uncertain whether there is any imbrication along the high strain zone.

6.6. Terrane amalgamation and metamorphism

The proposal for a series of separate tectonic terranes within the Lewisian was challenged by Park (2005a). The terrane model was established in West Greenland and has stood the test of time (see Friend et al., 1988; Friend and Nutman 2005 and references therein). A criticism of Park (2005a) was that the Lewisian terranes are not very big. In parts of the West Greenland Archaean craton the terranes there are not very big as a result of complex folding and attenuation. For example, in one stretch of approximately 4km, three different terranes can be traversed that contain two different granulite facies metamorphic events. Therefore, if this part of West Greenland was further dissected by younger fault structures it could provide a complicated setting as seen in Northwest Scotland. It is also the case that many of the terrane boundaries are not the sites of lost oceanic areas and the explanation is that assembly was through strike-slip shuffling of blocks of continental crust. That clear differences can be identified between the different blocks in the Lewisian Gneiss Complex suggest that they are terranes. Park (2005a) proposes an upper and lower 'two plate' model with the Loch Maree Group sandwiched between representing a suture. This is an acceptable model for the formation of the arc that created the Loch Maree Group. This creation of the volcanic arc must have occurred soon after ca. 2000 Ma, the youngest detrital age from zircons (Whitehouse et al., 1997a) and the ca. 1900 Ma age for the Ard Gneiss (Park et al., 2001). This model omits the ca. 1877 Ma granulite facies metamorphism in the Ialltaig gneisses which indicates that amalgamation occurred after this date.

In terms of the development of the Loch Maree Group, the U-Pb zircon ages presented here are in general agreement with the accretionary model for it as presented by Park et al. (2001), where the amalgamation and metamorphism of the 'southern region' occurred during a protracted post-dyke event. Four phases of deformation (D_1-D_4, Park 1964, Park et al. 1987, 2001) are recognised, with the D_1/D_2 phase attributed to the development of an accretionary prism and the initial effects of continent-continent collision. This produced a composite fabric recognised throughout the highly deformed gneisses around Torridon, the supracrustals of the Loch Maree Group, and gneisses north of the Loch Maree Group. D_1/D_2 fabrics have been refolded during the D_3 retrogression phase to produce the Tollie, Letterewe and Carnmore fold closures, with D_4 deformation concentrated along narrow zones of shearing and cataclasis.

The timing of D_1/D_2 amalgamation and metamorphism in the 'southern region' has been determined from the Ard gneiss protolith age to be ca. 1900 (Park et al., 2001) and confirmed by the data presented here, based on the interpretation that emplacement was syntectonic. The intrusion of the Ard gneiss protolith into the Loch Maree Group is considered to have occurred during peak metamorphism, due to its lack of strong D_1/D_2 fabrics and the inclusions of minor amphibolite and marble bands (Park et al. 2001). This was modelled from the preserved assemblages by Droop et al. (1999) as 530 ± 20°C at 6.5 ± 1.5 kbar. Alternatively, a limited development of D_1/D_2 fabrics in the Ard gneiss may be explained by the relative competency contrast between a competent granodiorite sheet and incompetent, already foliated supracrustal rocks, with the latter accommodating more D_1/D_2 strain. Entrained supracrustal material in the Ard gneiss protolith would be expected, and during metamorphism and deformation would inevitably become metamorphosed and banded in the granodioritic gneiss. The Ard gneiss clearly has an intrusive origin. Given that fabrics thought to be D_1/D_2 are recognised in both the Loch Maree Group and the surrounding TTG gneisses, all of these would have had to have developed after the ca. 1877 Ma peak of metamorphism in the Ialltaig gneiss. It is thus suggested that some of the fabrics that have been correlated are in fact different events.

Metamorphism and folding during D_3 is thought to have occurred during retrogression from D_1/D_2 peak P-T (Droop et al., 1999). The ca. 1670 Ma titanite ages here are thought to coincide with D_3 cooling through the U-Pb closure temperature, and there is a 1694 ± 5 Ma U-Pb age from a post-

D_2, pre-D_3 pegmatite from Tollie (Park et al., 2001), and a cluster of K-Ar cooling ages at ca. 1700 Ma (Moorbath & Park, 1971) providing an upper limit on the timing of D_3 folding.

The D_1-D_4 history defined for the 'southern region' has traditionally been considered contemporaneous with the "Laxfordian" event defined from Loch Laxford, however recent geochronology (Corfu et al., 1994, Kinny and Friend, 1997) has restricted the "Laxfordian" event to the juxtaposition of the Assynt and Rhiconich terranes along the Laxford Front at or prior to ca. 1740 Ma (Kinny & Friend, 1997). The widespread ca. 1670 Ma metamorphic ages that have been reported throughout the Lewisian (e.g. Corfu et al., 1994, Kinny & Friend, 1997, Friend & Kinny, 2001) now appear to reflect the final unifying event for the terranes now forming the Lewisian Gneiss Complex. On the Scottish mainland the Rona, Ialltaig, and Gairloch Terranes were tectonically amalgamated with and deformed against the combined Gruinard, Assynt and Rhiconich terranes at c.1670 Ma, which resulted in a low-grade metamorphic resetting of the U-Pb system in rutile (Kinny & Friend, 1997) in the Assynt Terrane. A revision of the use of the term Laxfordian for describing the post-dyke history of the 'southern region' is therefore required, since metamorphism occurred later than in the type Laxfordian gneisses near Loch Laxford.

6.7. Implications for Proterozoic reconstructions

The pre-Grenvillian supercontinent (Fig. 9) is highlighted by a continuity of mid-Proterozoic features (Bridgwater & Windley, 1973), with the Lewisian Gneiss Complex believed to form part of a key orogenic zone linking the Nagssugtoqidian of Greenland with the Lapland-Kola Belt of Baltica (e.g. Park, 1994; Buchan et al., 2000; Park et al., 2001). A review of palaeomagnetic data by Buchan et al. (2000) shows a pole age match at 1.235 Ga between Laurentia and Baltica, which places Baltica adjacent to Eastern Greenland (Fig. 9). In this fit, the Labradorian belt of Laurentia and the Gothian belt of Baltica are aligned, and are interpreted to have been part of a continuous active margin upon which successive magmatic arcs were accreted between 1.83 and 1.55 Ga. The direction of accretion is approximately normal to the Proterozoic fabrics developed in the Lewisian gneisses, and provides a regional context for the compressional regime that affected the Lewisian during the Palaeoproterozoic. While the interpreted direction and timing of terrane accretion in the Lewisian may be consistent with this, the disparate nature of the Lewisian terranes prior to 1.83 Ga questions pre-1.83 Ga reconstructions that assume the Lewisian to be in existence in its assembled form (and part of a larger Proterozoic belt) at that time.

Current understanding of Proterozoic North Atlantic tectonics suggest that prior to the development of the Labradorian and Gothian belts the region was in a collisional regime with the main direction of convergence about 40° counterclockwise (Fig. 9) with respect to that of the Labradorian and Gothian belts (Park, 1994, 1995). Subduction occurred along the western and northern margins of the North Atlantic Craton, and also between the Kola and Karelia Cratons. This widespread subduction is believed to coincide with the development of the Nagssugtoqidian and Lapland-Kola belts respectively, although any relationship between the two at this stage, if any, is uncertain. In the reconstruction of Park (1994, 1995) the development of the Nagssugtoqidian and Lapland Kola belts between 1.9 and 1.85 Ga are depicted separately, though tentatively linked through a zone of predominantly strike-slip movement. Continuity between the Nagssugtoqidian and Lapland-Kola belts has been inferred by their broadly equivalent geology and relationship to the Lewisian Gneiss Complex, but may need revising in light of recent geochronology from all three belts. In particular the high spatial density of SHRIMP U/Pb zircon constraints for the Lewisian has shown that the assembled terranes are narrow; typically <50km in thickness and often considerably less. If the Archaean and Proterozoic components were disparate prior to 1.83 Ga then there is no longer the necessity that a 200-300 km wide and 2000-3000 km long orogenic belt existed before c.1.85 Ga.

One traditional link between all North Atlantic Palaeoproterozoic belts has been the presence of juvenile crust within the reworked Archaean gneisses, with age signatures (of varying precision) for supracrustal rocks throughout the North Atlantic Proterozoic belts in general agreement (e.g. Daly et al., 2001). With ion microprobe zircon geochronology, the Palaeoproterozoic juvenile terranes may be further resolved, indicating that they may not be coeval as traditionally thought. This is highlighted well by recent precise geochronology of the supracrustal belts of the Lewisian Gneiss Complex (Whitehouse et al., 1997a, Park et al., 2001, Friend & Kinny, 2001), and further supported by geochronological differences between the supracrustal rocks of East Greenland and the Lapland Kola Orogen.

7. CONCLUSIONS

It is shown that the concept of the Lewisian Gneiss Complex comprising different blocks of crust can be extended southwards to the Loch Torridon area. The age constraints on the assembly of the complex are best in and around the Gairloch Terrane and it is shown to be Palaeoproterozoic.

The distinct isotopic signature of the Gruinard Terrane (Love et al. 2004) can be recognised in the mainland Lewisian gneisses as far south as the Tollie antiform, which is in sheared contact with the Gairloch Terrane (combined Loch Maree Group supracrustal rocks and the Ard gneiss). Uncertainty with respect to the structural interpretation means the exact nature of the contact is unclear. It is currently assumed that the Gruinard Belt is an extension of the Gairloch Terrane over the Carnmore antiform, which implies that the contact between the Gairloch and Gruinard Terranes would also be folded. However, structurally this requires the TTG gneisses forming the Carnmore antiform to be part of another unrecognised component, rather than Gruinard gneisses.

Rather than comprising Archaean basement, as previously considered, the Ialltaig gneisses are shown to have Palaeoproterozoic protoliths. Further, they were metamorphosed to granulite facies at ca.1877 Ma, and must have been tectonically emplaced into the margin of the Loch Maree Group in the Gairloch Terrane after this date. This metamorphism is the youngest single event unique to a terrane in the southern part of the Complex and is used as the upper estimate for the final amalgamation of the southern terranes. There are thus some major structural events that have not yet been fully understood in the evolution of this part of the Lewisian Gneiss Complex.

The Rona Terrane extends from the southern margin of the Gairloch Terrane, marked by the Shieldaig shear zone, to the southern limits of mainland Lewisian exposure south of Loch Torridon. Augen of low Palaeoproterozoic strain have yielded protolith ages between ca. 3135 Ma and ca.2880 Ma, and there is evidence that the older components underwent anatexis at ca. 2955 Ma. No evidence of the ca. 2730 Ma Gruinard Terrane metamorphic zircon signature has been observed. The gneisses do appear once to have been to granulite faices but the event is presently undated.

Metamorphic titanite from the Rona Terrane have yielded ages of c.1670 Ma, and are considered to date the timing of the main deformation which has affected the southern part of the Complex. This matches the ca.1670 Ma metamorphic ages that have been reported elsewhere in the mainland Lewisian Gneiss Complex and is interpreted here as reflecting the final assembly or stabilisation of the complex – called the Somerledian (Kinny et al., 2005a).

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Figure captions

Figure 1. Geological sketch map of the mainland Lewisian Gneiss Complex of northwest Scotland showing existing terrane subdivisions (Friend & Kinny 2001, Love et al. 2004). Abbreviations: CSZ, Canisp Shear Zone; FL, Fionn Loch; GB, Gruinard Belt; LMF, Loch Maree Fault; LSZ, Laxford Shear Zone; SL, Strathan Line; SSZ, Shieldaig Shear Zone.

Figure 2. Geological sketch map of the mainland Lewisian Gneiss Complex south of Gruinard Bay with sample locations and ages indicated. The boxes at Gruinard Bay with sample numbers commencing MW are data from Whitehouse et al. (1997b), with unlabelled boxes indicating data from Love et al. (2004). Abbreviations: CA, Carnmore Antiform; KSZ, Kerry shear zone; LMF, Loch Maree Fault; LS, Letterewe synform; SSZ, Shieldaig shear zone; TA, Tollie antiform.

Figure 3. Representative cathodoluminescence images of sectioned zircons from the Southern mainland Lewisian gneisses. Abbreviations: p protolith zircon, r recrystallised (metamorphic) zircon, n new zircon growth (metamorphic), i inherited zircon.

Figure 4. Conventional Wetherill concordia diagrams showing U/Pb zircon results from gneisses north of the Loch Maree Group. Inset: results from Love et al. 2004 for comparable Gruinard Terrane samples taken under identical analytical conditions.

Figure 5. Conventional Wetherill concordia diagram showing U/Pb zircon results from gneisses in zones of low Proterozoic strain south of the Loch Maree Group.

Figure 6. Conventional Wetherill concordia diagram showing U/Pb zircon results from the Ialltaig Gneiss and Ard Gneiss. Inset: U/Pb titanite results from the Ard Gneiss.

Figure 7. Conventional Wetherill concordia diagram showing U/Pb titanite results from GL00/03 and GL01/01.

Figure 8. Schematic cross-section through the southern part of the mainland Lewisian Gneiss Complex showing the possible relationship between the Loch Maree Group and Gruinard Belt (modified after Park, 2002; Park et al., 2001). Gaps in the present coverage of geochronology data make the nature of the gneisses around the Carnmore antiform uncertain. Abbreviations: CBB, Creag Bhan crush belt; GB, Gruinard Belt; IT, Ialtaig terrane; KSZ, Kerry shear zone; SSZ, Shieldaig Shear Zone.

Figure 9. Proterozoic tectonic reconstructions of the North Atlantic region at 1.265 Ga (after Buchan et al., 2000) and 1.9-1.85 Ga (after Park, 1995). In the 1.265 Ga reconstruction the mainland Lewisian Complex is depicted to be a part of the LKO (Baltica) while the Outer Hebrides are linked to East Greenland (Laurentia). The region approximately occupied by the Lewisian terranes is modified to show the probable discontinuity of Palaeoproterozoic features. Abbreviations: E Nag/ W Nag East and West Nagssugtoqidian respectively; LKO, Lapland Kola Orogen; EGC, East Greenland Craton; NAC, North Atlantic Craton.

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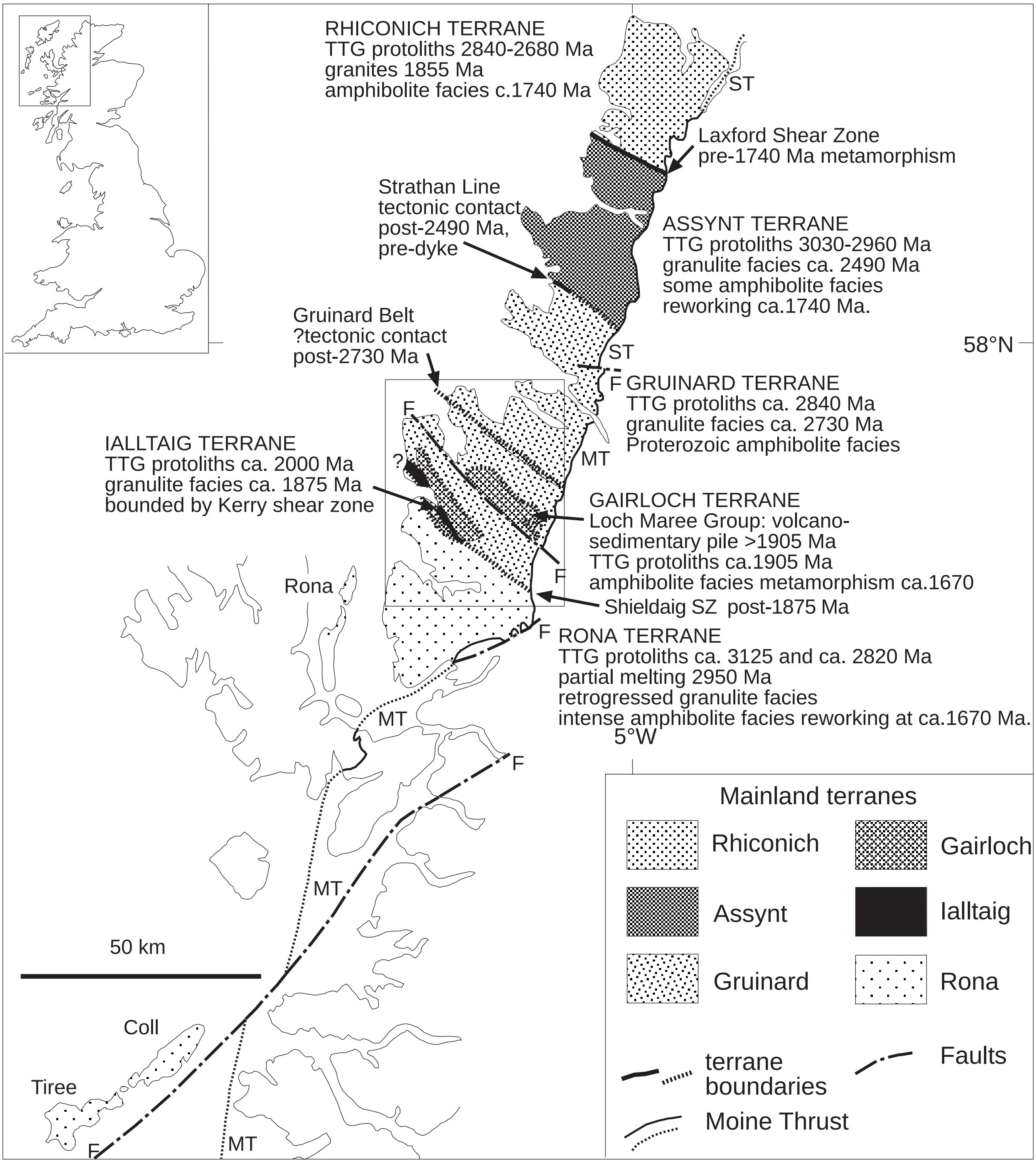
Figure

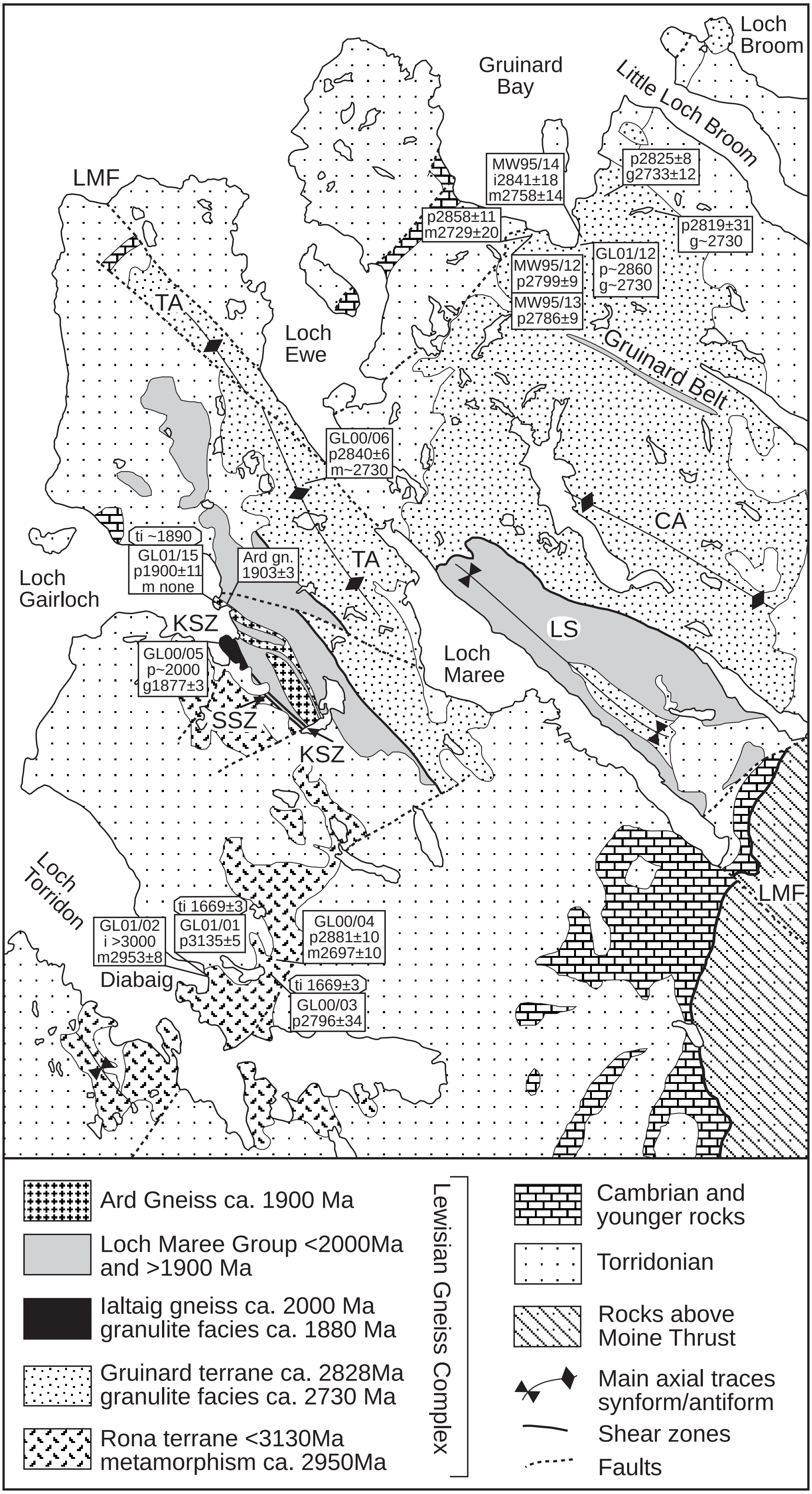
18/09/2009 Love, Friend & Kinny

Fig. 1



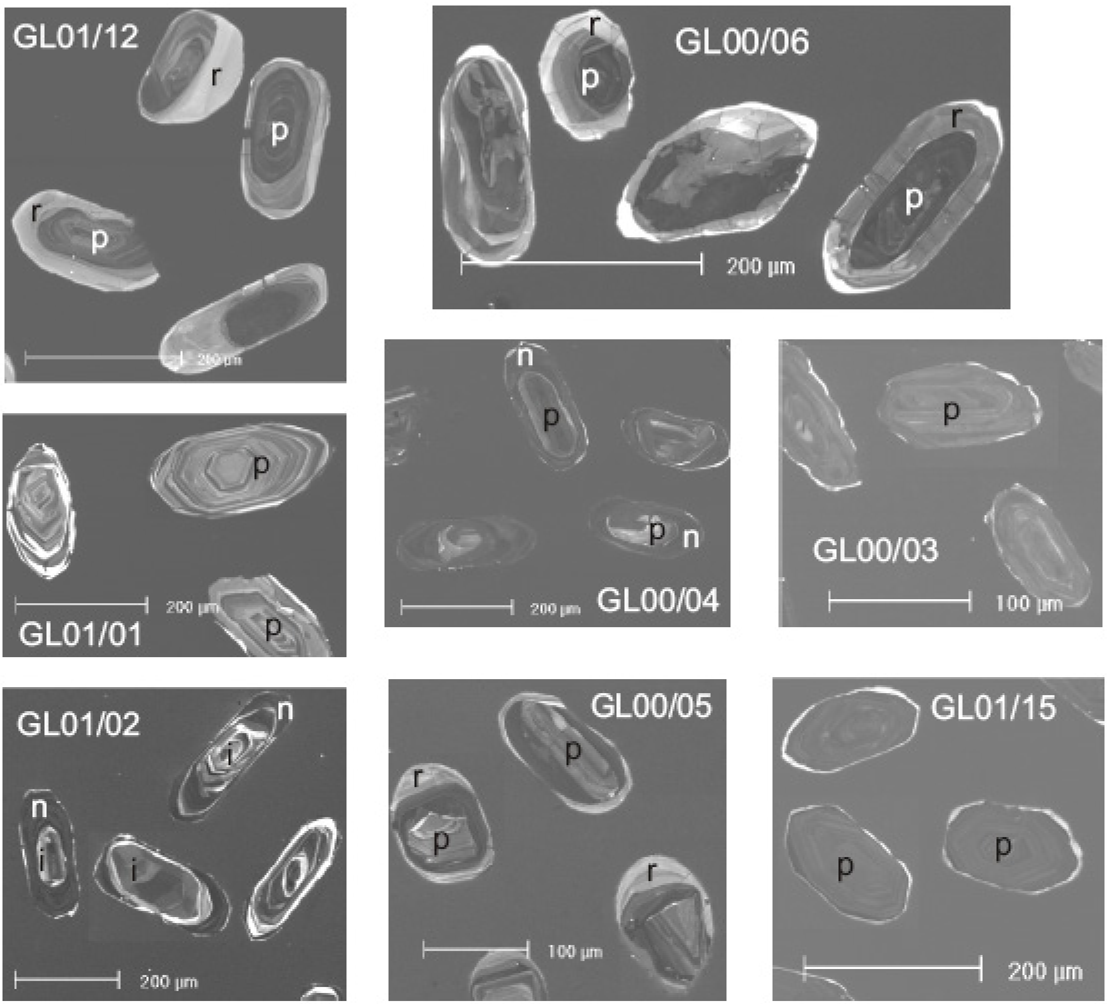
Figure

Love, Friend & Kinny Fig. 2



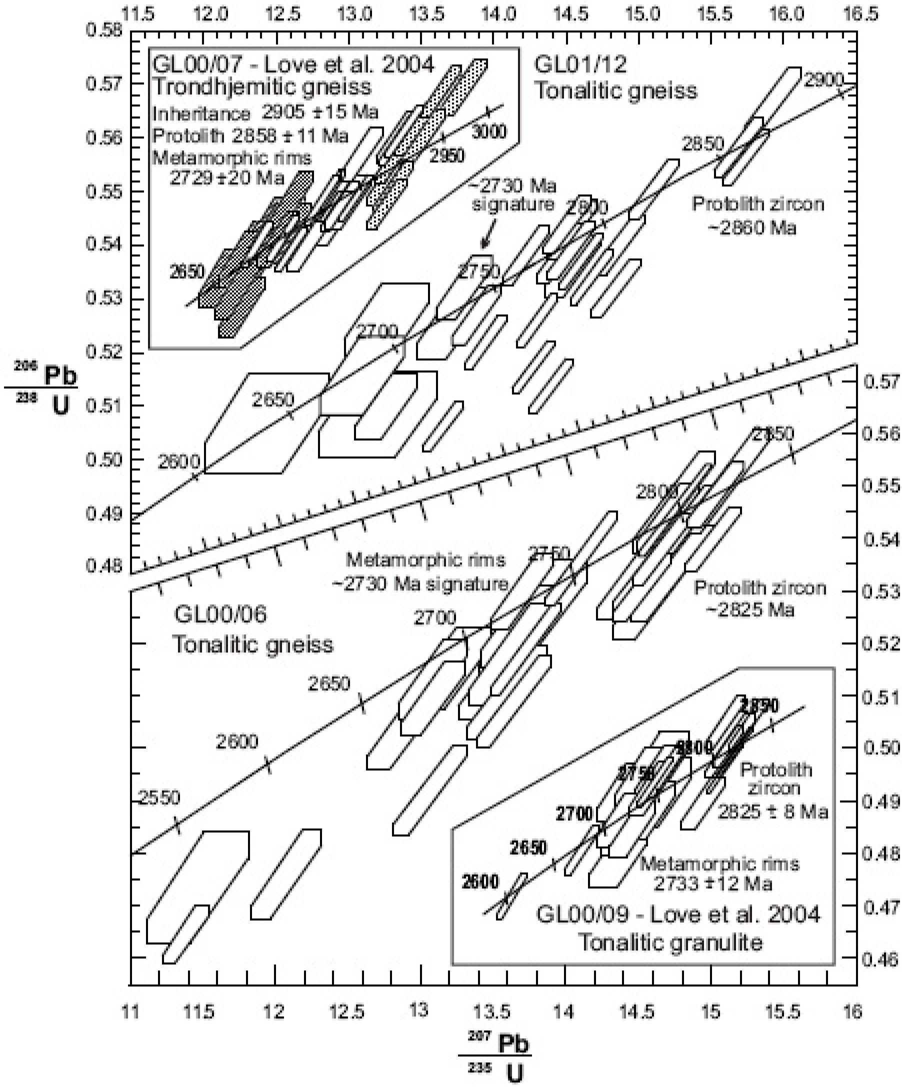
Figure

Love, Friend & Kinny **Figure 3**



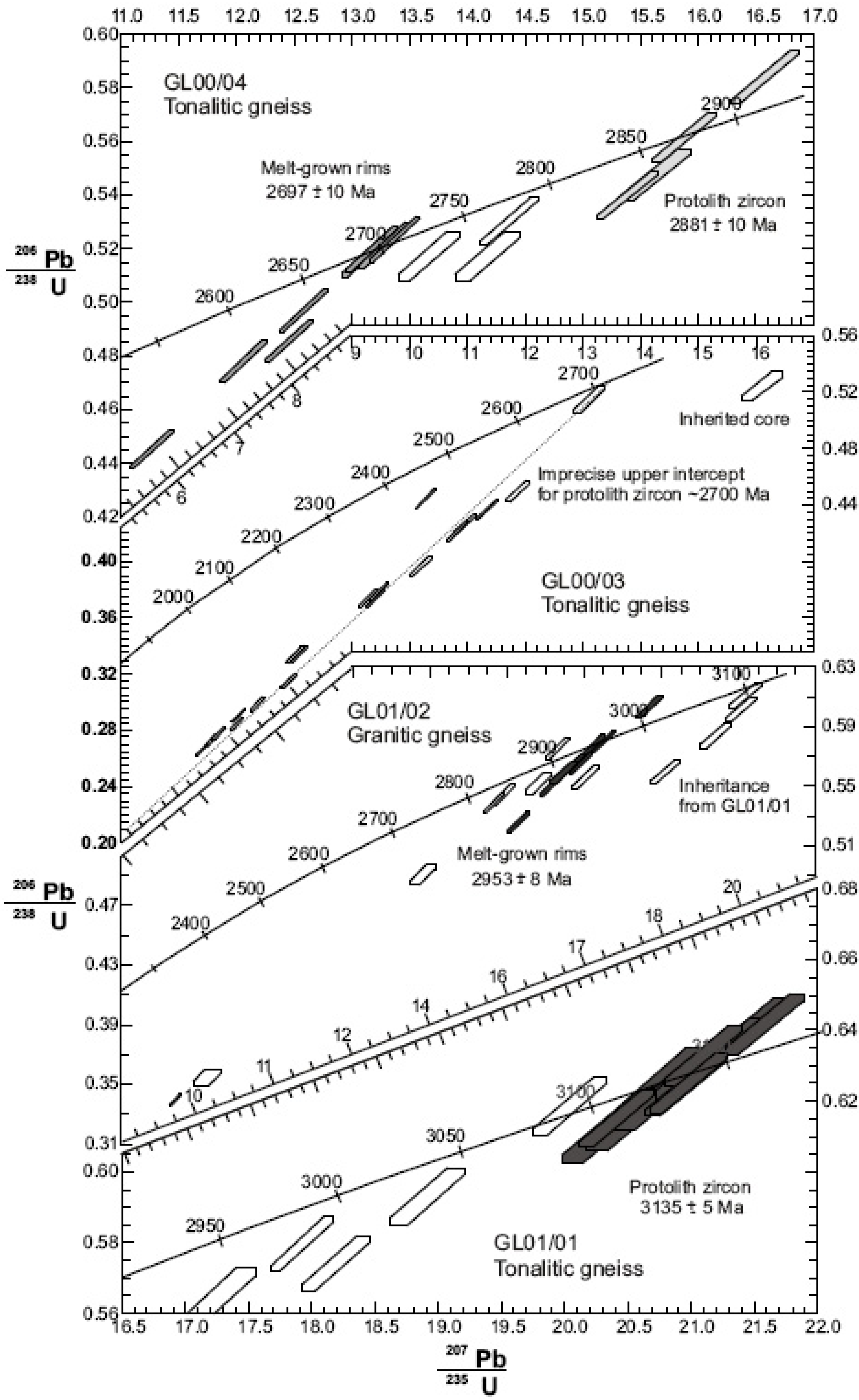
Figure

Love, Friend & Kinny Figure 4



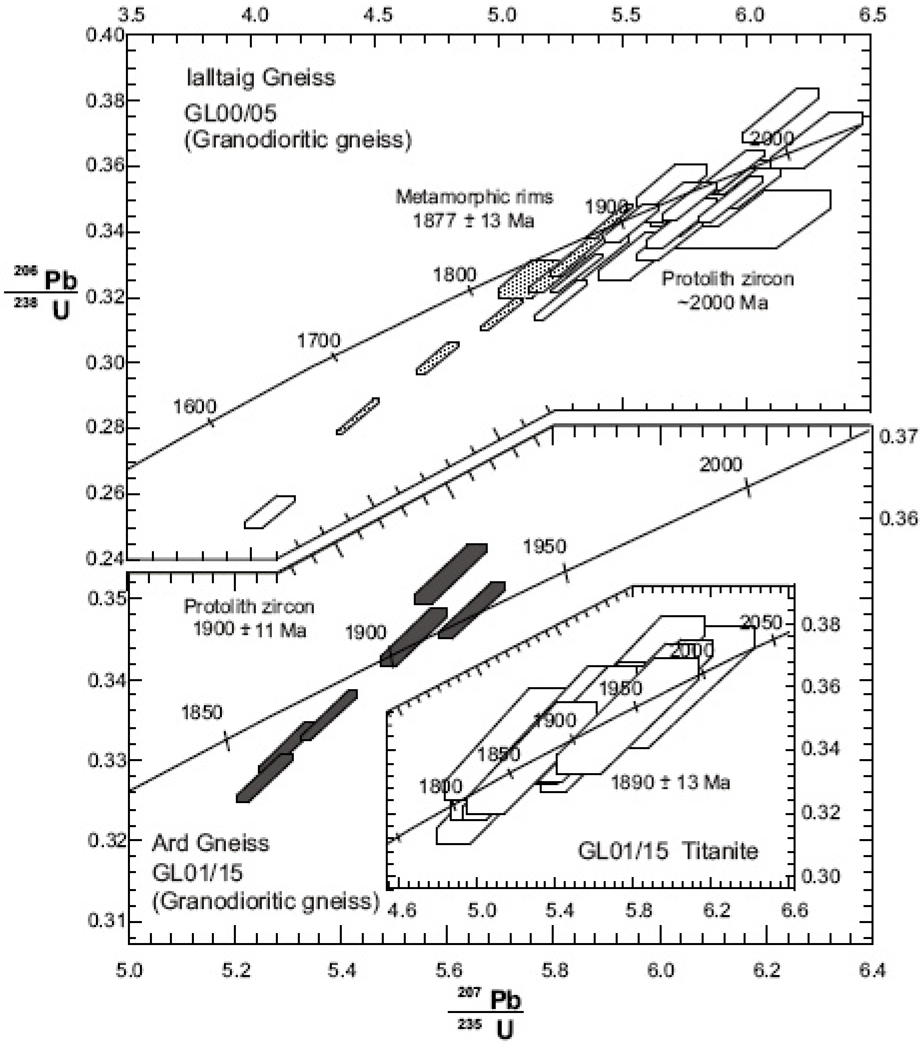
Figure

Love, Friend & Kinny   Figure 5



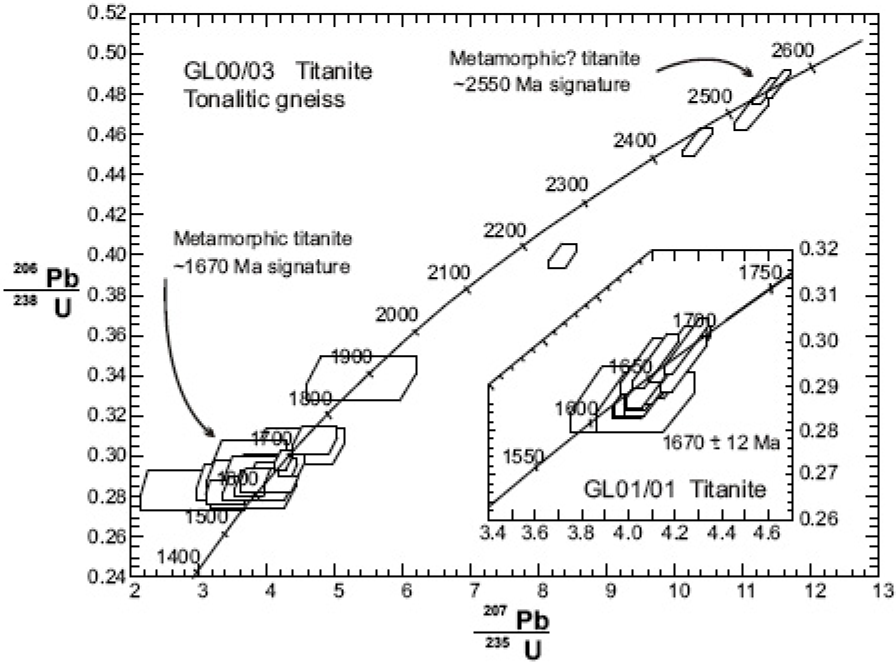
Figure

Love, Friend & Kinny **Figure 6**



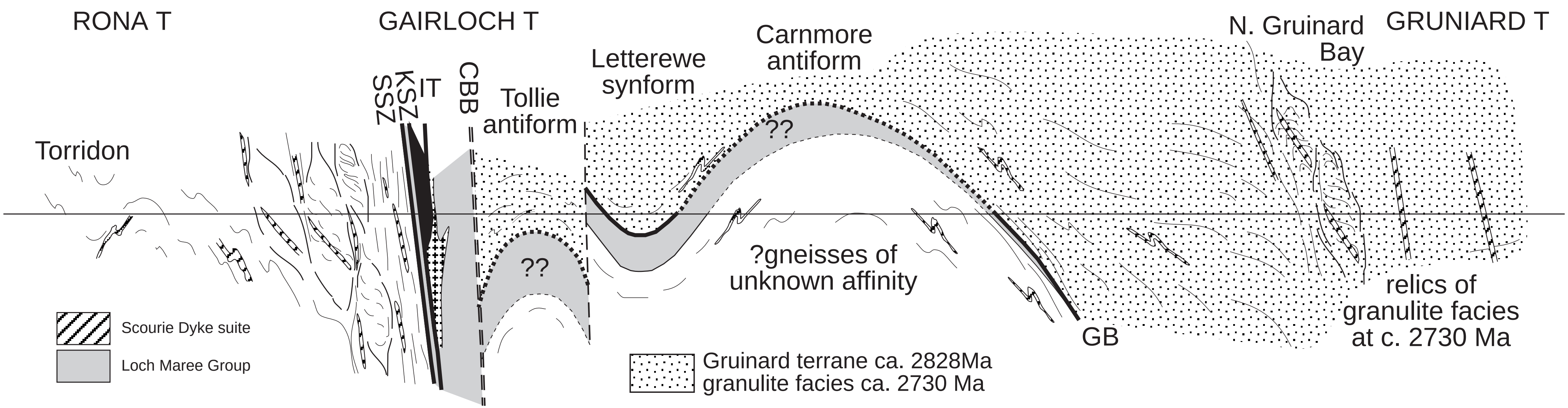
Figure

Love, Friend & Kinny **Figure 7**



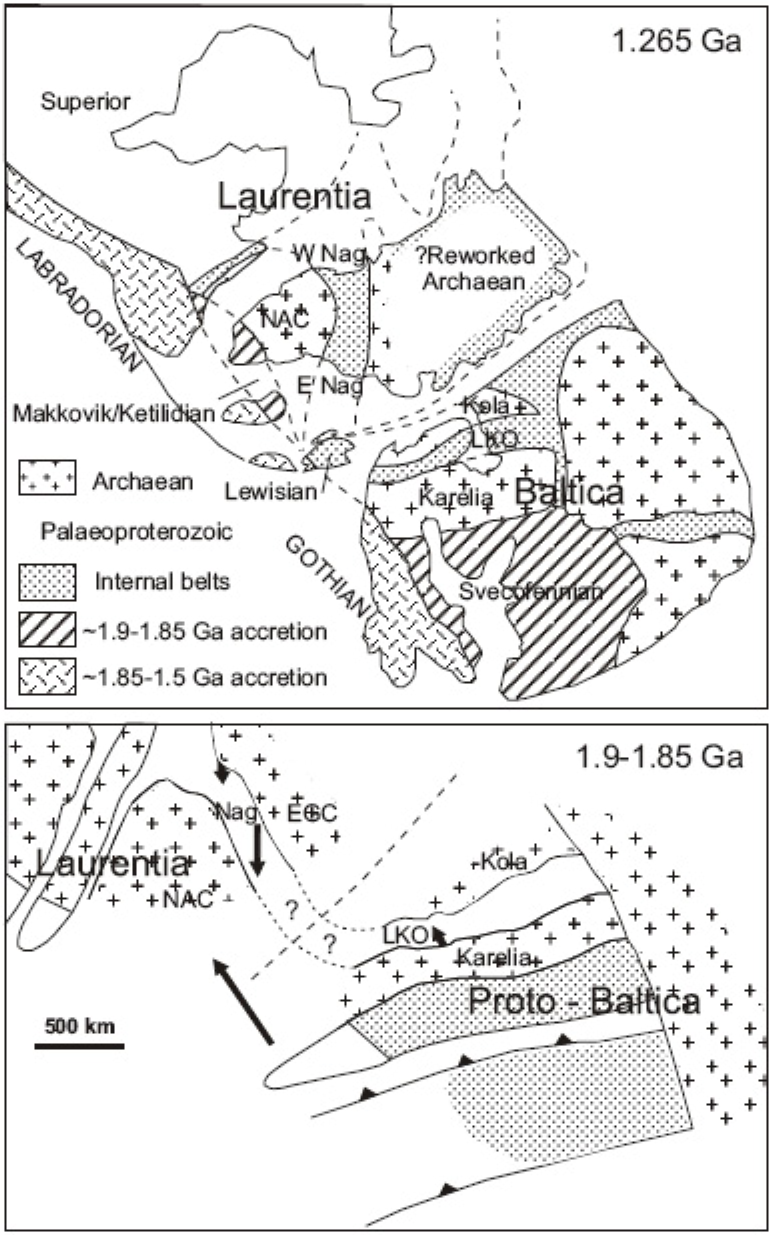
Figure

Love, Friend & Kinny Fig. 8



Figure

Love, Friend & Kinny **Figure 9**



Table

Table 1. SHRIMP U-Pb isotopic data for GL0l/12. *%c206* is the percentage of non-radiogenic ^{206}Pb estimated from the measured ^{204}Pb/^{206}Pb ratio.
Abbreviations for spot types: *oz oscillatory zones, p protolith zircon, tz transgressive zones, x recrystallised.*
GL01/12 amphibolite facies tonalitic gneiss, western Gruinard Bay NG 940902.

Spot	Type	U	Th	Th/U	%c206	^{207}Pb/^{206}Pb	±1σ	^{206}Pb/^{238}U	±1σ	^{207}Pb/^{235}U	±1σ	^{207}Pb/^{206}Pbage	±1σ
GL01-12-1	*f, tz*	42	93	2.23	0.42	0.1881	0.0013	0.5321	0.0060	13.800	0.191	2725	11
GL01-12-2	*tz*	102	113	1.11	0.22	0.1935	0.0007	0.5436	0.0054	14.502	0.161	2772	6
GL01-12-3	*tz*	108	34	0.31	0.10	0.1990	0.0007	0.5504	0.0056	15.102	0.170	2818	6
GL01-12-4	*p*, oz	110	37	0.34	0.09	0.2038	0.0006	0.5583	0.0055	15.691	0.169	2857	5
GL01-12-5	*f, tz*	16	29	1.83	0.87	0.1842	0.0028	0.5156	0.0074	13.095	0.291	2691	25
GL01-12-6	*tz*, oz	221	56	0.25	0.09	0.2024	0.0007	0.5320	0.0054	14.843	0.166	2845	6
GL01-12-7	*tz*, oz	71	84	1.19	0.15	0.1943	0.0008	0.5429	0.0056	14.546	0.169	2779	7
GL01-12-8	*tz*	76	69	0.91	0.15	0.1996	0.0008	0.5401	0.0055	14.865	0.171	2823	7
GL01-12-9	*tz*	173	41	0.24	0.33	0.1956	0.0006	0.5061	0.0048	13.650	0.140	2790	5
GL01-12-10	*tz*, oz	145	39	0.27	0.07	0.1938	0.0006	0.5219	0.0050	13.945	0.145	2774	5
GL01-12-11	*f, tz*	19	39	2.02	1.06	0.1829	0.0028	0.5259	0.0072	13.262	0.288	2679	25
GL01-12-12	*tz*, oz	93	95	1.02	0.14	0.1956	0.0007	0.5390	0.0054	14.537	0.160	2790	6
GL01-12-13	*f, tz*	8	26	3.14	1.75	0.1779	0.0048	0.5069	0.0093	12.436	0.431	2634	45
GL01-12-14	*f, tz*	28	41	1.47	0.18	0.1892	0.0019	0.5253	0.0066	13.704	0.236	2735	17
GL01-12-15	*tz*, oz	51	53	1.04	0.45	0.1967	0.0013	0.5362	0.0058	14.543	0.194	2799	11

Table

Table 2. SHRIMP U-Pb isotopic data for GL00/06. %c206 is the percentage of non-radiogenic ^{206}Pb estimated from the measured ^{204}Pb/^{206}Pb ratio.
Abbreviations for spot types: oz oscillatory zoning, tz transgressive zoning, x recrystallised, c core, r rim
GL01/06 amphibolite facies tonalitic gneiss, Tollie NG 832782.

Spot	Type	U	Th	Th/U	%c206	^{207}Pb/^{206}Pb	±1σ	^{206}Pb/^{238}U	±1σ	^{207}Pb/^{235}U	±1σ	^{207}Pb/^{206}Pbage	±1σ
GL00/06-1	oz,c	84	73	0.88	0.27	0.2001	0.0015	0.5303	0.0097	14.629	0.300	2827	12
GL00/06-2	x, tz	110	107	0.97	0.85	0.1841	0.0015	0.4759	0.0084	12.077	0.245	2690	13
GL00/06-3	oz	139	73	0.52	0.65	0.1981	0.0009	0.5405	0.0094	14.766	0.274	2811	8
GL00/06-4	x, tz	93	87	0.93	0.44	0.1931	0.0014	0.5105	0.0092	13.594	0.275	2769	12
GL00/06-5	oz, c	117	100	0.86	0.23	0.2003	0.0012	0.5333	0.0094	14.728	0.285	2829	10
GL00/06-6	x, tz, r	27	46	1.72	0.99	0.1757	0.0032	0.4733	0.0105	11.464	0.348	2612	30
GL00/06-7	oz, c	101	93	0.92	0.15	0.1991	0.0011	0.5507	0.0098	15.115	0.291	2818	9
GL00/06-8	x, tz	31	61	1.98	0.21	0.1903	0.0026	0.5246	0.0114	13.768	0.370	2745	22
GL00/06-9	oz	539	766	1.42	0.26	0.1963	0.0004	0.5453	0.0089	14.757	0.248	2795	4
GL00/06-10	x, tz	518	209	0.40	0.96	0.1919	0.0007	0.5189	0.0087	13.727	0.241	2758	6
GL00/06-11	oz	114	151	1.33	0.23	0.1956	0.0011	0.5467	0.0097	14.742	0.283	2790	9
GL00/06-12	x, tz	134	6 98	0.07	0.48	0.1392	0.0005	0.2279	0.0037	14.374	0.075	2217	6
GL00/06-13	x, tz	128	52	0.41	0.22	0.1945	0.0010	0.5087	0.0089	13.642	0.258	2780	9
GL00/06-14	oz	113	109	0.96	0.20	0.1926	0.0011	0.4919	0.0086	13.067	0.250	2765	9
GL00/06-15	x, tz, r	62	80	1.29	0.94	0.1907	0.0015	0.5178	0.0096	13.614	0.285	2748	13
GL00/06-16	x, tz	266	90	0.34	0.59	0.1909	0.0006	0.5359	0.0090	14.105	0.247	2750	6
GL00/06-17	oz	74	56	0.76	0.07	0.1968	0.0011	0.5311	0.0064	14.410	0.200	2800	9
GL00/06-18	x, tz	95	82	0.86	0.75	0.1911	0.0016	0.5113	0.0060	13.472	0.208	2752	14
GL00/06-19	x, tz	30	56	1.88	0.09	0.1850	0.0016	0.5130	0.0074	13.084	0.231	2698	14
GL00/06-20	x, tz	43	68	1.57	0.09	0.1862	0.0014	0.5097	0.0070	13.086	0.215	2709	13
GL00/06-21	oz, c	98	91	0.93	0.16	0.1978	0.0010	0.5325	0.0062	14.525	0.193	2808	8
GL00/06-22	x, tz	28	54	1.92	0.19	0.1893	0.0017	0.5233	0.0078	13.659	0.248	2736	14
GL00/06-23	oz	110	92	0.83	0.04	0.2017	0.0008	0.5398	0.0063	15.012	0.190	2840	6
GL00/06-24	x, tz	25	47	1.86	0.28	0.1859	0.0023	0.5152	0.0079	13.208	0.276	2707	21
GL00/06-25	oz, c	72	82	1.13	0.11	0.1960	0.0012	0.5441	0.0066	14.704	0.210	2793	10
GL00/06-26	oz	99	84	0.85	0.00	0.1988	0.0009	0.5483	0.0063	15.031	0.192	2817	7
GL00/06-27	x, tz, r	25	46	1.83	0.00	0.1876	0.0012	0.5297	0.0074	13.700	0.222	2721	11
GL00/06-28	oz	79	62	0.78	0.00	0.1977	0.0008	0.5379	0.0069	14.659	0.203	2807	7
GL00/06-29	x, tz	31	56	1.82	0.02	0.1854	0.0017	0.5025	0.0066	12.849	0.216	2702	15
GL00/06-30	oz	67	50	0.74	0.00	0.1985	0.0009	0.5305	0.0063	14.524	0.192	2814	7
GL00/06-31	x, tz	69	162	0.90	0.24	0.1777	0.0010	0.4644	0.0055	11.379	0.157	2632	10

Table

Table 3. SHRIMP U-Pb zircon isotopic data for GL00/03. %c206 is the percentage of non-radiogenic ^{206}Pb estimated from the measured $^{204}Pb/^{206}Pb$ ratio.
Abbreviations for spot types: oz oscillatory zones.
GL00/03 weakly-foliated amphibolite facies tonalitic gneiss, zone of low post-dyke strain near Diabaig NG 829598.

Spot	Type	U	Th	Th/U	%c206	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$ age	±1σ
GL00/03-1	oz	518	565	1.09	0.20	0.1897	0.0007	0.4228	0.0071	11.060	0.194	2740	6
GL00/03-2	oz	699	95	0.14	0.37	0.1832	0.0006	0.3770	0.0062	9.525	0.163	2682	5
GL00/03-3	oz	807	339	0.42	1.14	0.1759	0.0007	0.2775	0.0046	6.729	0.117	2614	7
GL00/03-4	oz	624	358	0.57	0.08	0.1920	0.0005	0.4325	0.0071	11.450	0.195	2759	4
GL00/03-5	oz	649	179	0.28	0.25	0.1834	0.0009	0.3140	0.0055	7.941	0.148	2684	8
GL00/03-6	oz	1144	615	0.54	0.29	0.1843	0.0006	0.3722	0.0061	9.458	0.163	2692	6
GL00/03-7	oz	708	384	0.54	1.68	0.1757	0.0009	0.2716	0.0045	6.580	0.117	2613	8
GL00/03-8	oz	695	336	0.48	1.85	0.1794	0.0009	0.2841	0.0047	7.028	0.125	2647	8
GL00/03-9	oz	553	161	0.29	0.28	0.1952	0.0009	0.4456	0.0075	11.992	0.216	2787	8
GL00/03-10	oz	694	442	0.64	0.28	0.1757	0.0006	0.3321	0.0057	8.047	0.146	2613	6
GL00/03-11	oz	1209	495	0.41	0.19	0.1754	0.0004	0.2659	0.0044	6.432	0.108	2610	4
GL00/03-12	oz	783	810	1.03	0.35	0.1893	0.0009	0.3920	0.0071	10.230	0.198	2736	7
GL00/03-13	oz	970	822	0.85	1.30	0.1815	0.0006	0.3728	0.0062	9.330	0.161	2667	6
GL00/03-14	oz	1003	6	0.01	0.07	0.1696	0.0003	0.4420	0.0072	10.335	0.173	2554	3
GL00/03-15	oz	619	135	0.22	0.97	0.1767	0.0007	0.3332	0.0056	8.117	0.144	2622	7
GL00/03-16	oz	692	336	0.49	0.93	0.1801	0.0009	0.2977	0.0049	7.395	0.131	2654	8
GL00/03-17	oz	468	344	0.74	0.41	0.2258	0.0021	0.5214	0.0100	16.230	0.360	3022	15
GL00/03-18	oz	1862	985	0.53	0.66	0.1767	0.0004	0.2896	0.0047	7.056	0.118	2622	4
GL00/03-19	oz	334	82	0.24	1.53	0.1870	0.0014	0.5120	0.0095	13.200	0.275	2716	12
GL00/03-20	oz	548	148	0.27	0.13	0.1900	0.0006	0.4176	0.0070	10.939	0.190	2742	5

Table

Table 4. SHRIMP U-Pb titanite isotopic data for GL00/03. *%c206* is the percentage of non-radiogenic ^{206}Pb estimated from the measured $^{204}Pb/^{206}Pb$ ratio. GL00/03 amphibolite facies tonalitic gneiss, zone of low post-dyke strain near Diabaig NG 829598.

Spot	U	Th	Th/U	%c206	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$ age	±1σ
GL00/03-1	130	57	0.44	2.27	0.1028	0.0026	0.2989	0.0048	4.235	0.132	1674	46
GL00/03-2	58	36	0.62	6.35	0.0957	0.0061	0.2869	0.0061	3.787	0.264	1543	120
GL00/03-3	19	5	0.26	14.90	0.1147	0.0161	0.3396	0.0115	5.371	0.798	1875	256
GL00/03-4	127	58	0.46	1.32	0.1622	0.0017	0.4587	0.0071	10.262	0.204	2479	18
GL00/03-5	154	59	0.39	4.11	0.1501	0.0027	0.4016	0.0061	8.313	0.209	2347	31
GL00/03-6	33	17	0.51	7.53	0.0934	0.0092	0.2932	0.0075	3.777	0.394	1496	186
GL00/03-7	69	44	0.64	2.93	0.1071	0.0037	0.2961	0.0056	4.374	0.180	1751	63
GL00/03-8	47	10	0.22	4.38	0.1105	0.0070	0.3087	0.0069	4.702	0.326	1807	115
GL00/03-9	27	15	0.54	17.22	0.1061	0.0147	0.3060	0.0090	4.479	0.648	1734	256
GL00/03-10	41	35	0.87	5.76	0.1005	0.0069	0.2910	0.0066	4.032	0.302	1633	128
GL00/03-11	22	6	0.26	23.98	0.0763	0.0208	0.2838	0.0100	2.987	0.835	1104	569
GL00/03-12	20	11	0.55	11.98	0.0970	0.0153	0.2837	0.0095	3.796	0.627	1568	299
GL00/03-13	30	9	0.31	11.81	0.0906	0.0122	0.3002	0.0085	3.748	0.530	1437	260
GL00/03-14	102	62	0.60	1.47	0.1695	0.0022	0.4726	0.0079	11.043	0.248	2552	22
GL00/03-15	70	37	0.53	11.47	0.0877	0.0079	0.2828	0.0058	3.419	0.325	1376	174
GL00/03-16	127	55	0.43	1.91	0.1057	0.0028	0.2953	0.0047	4.304	0.140	1727	49
GL00/03-17	36	20	0.57	7.01	0.0974	0.0093	0.2947	0.0075	3.957	0.404	1575	180
GL00/03-18	39	24	0.61	8.88	0.1003	0.0104	0.2859	0.0077	3.954	0.437	1630	194
GL00/03-19	32	33	1.04	8.28	0.0889	0.0106	0.2910	0.0074	3.566	0.446	1402	230
GL00/03-20	38	13	0.34	11.62	0.1004	0.0099	0.2905	0.0070	4.021	0.421	1631	185
GL00/03-21	196	153	0.78	1.09	0.1685	0.0012	0.4846	0.0067	11.256	0.182	2543	12
GL00/03-22	37	26	0.71	7.21	0.0933	0.0073	0.2847	0.0065	3.662	0.310	1494	149
GL00/03-23	208	161	0.77	0.89	0.1700	0.0011	0.4882	0.0066	11.444	0.179	2558	10
GL00/03-24	24	9	0.40	11.04	0.0890	0.0134	0.2872	0.0090	3.522	0.554	1403	291
GL00/03-25	57	29	0.50	5.08	0.0966	0.0058	0.2888	0.0061	3.846	0.256	1559	114
GL00/03-26	58	43	0.74	5.51	0.1056	0.0057	0.2890	0.0059	4.206	0.254	1724	100

Table

Table 5. SHRIMP U-Pb zircon isotopic data for GL00/04. %c206 is the percentage of non-radiogenic ^{206}Pb estimated from the measured $^{204}Pb/^{206}Pb$ ratio.
Abbreviations for spot types: oz oscillatory zoning, tz transgressive zoning, c core, r rim.
GL00/04 amphibolite facies tonalitic gneiss, zone of low post-dyke strain near Diabaig NG 830604.

Spot	Type	U	Th	Th/U	%c206	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$age	±1σ
GL00/04-1	oz,c	559	80	0.14	8.03	0.1856	0.0004	0.5237	0.0086	13.400	0.225	2703	4
GL00/04-2	tz, r	1157	56	0.05	25.76	0.1768	0.0004	0.3518	0.0057	8.577	0.143	2623	3
GL00/04-3	oz, c	93	52	0.55	34.69	0.1993	0.0014	0.5171	0.0093	14.211	0.285	2821	11
GL00/04-4	tz, r	778	852	1.10	97.62	0.1712	0.0008	0.2810	0.0046	6.631	0.117	2569	7
GL00/04-5	tz, r	678	89	0.13	7.66	0.1839	0.0004	0.4969	0.0082	12.598	0.213	2688	4
GL00/04-6	oz, c	108	93	0.86	16.01	0.1921	0.0011	0.5173	0.0091	13.698	0.264	2760	10
GL00/04-7	tz, r	1105	65	0.06	88.29	0.1427	0.0006	0.2252	0.0037	4.433	0.078	2261	8
GL00/04-8	tz, r	308	58	0.19	8.27	0.1843	0.0006	0.5184	0.0086	13.174	0.229	2692	5
GL00/04-9	tz, r	433	60	0.14	7.63	0.1852	0.0005	0.5213	0.0086	13.312	0.226	2700	4
GL00/04-10	oz, c	265	111	0.42	4.06	0.2067	0.0007	0.5840	0.0099	16.645	0.295	2880	6
GL00/04-11	oz, c	232	131	0.56	11.91	0.1970	0.0007	0.5303	0.0089	14.403	0.254	2801	6
GL00/04-12	tz, r	616	84	0.14	4.61	0.1864	0.0004	0.4854	0.0079	12.472	0.209	2710	4
GL00/04-13	tz, r	127	53	0.42	3.50	0.2078	0.0011	0.5475	0.0096	15.687	0.298	2888	9
GL00/04-14	tz, r	490	42	0.09	10.11	0.1832	0.0005	0.4780	0.0079	12.074	0.205	2682	4
GL00/04-15	tz, r	579	338	0.58	89.16	0.1873	0.0008	0.3550	0.0058	9.169	0.161	2719	7
GL00/04-16	tz, r	625	79	0.13	5.23	0.1842	0.0004	0.5199	0.0085	13.205	0.223	2691	4
GL00/04-17	tz, r	1075	42	0.04	11.07	0.1836	0.0004	0.4452	0.0072	11.268	0.188	2685	4
GL00/04-18	oz, c	456	479	1.05	47.73	0.2074	0.0006	0.5400	0.0089	15.442	0.263	2885	5
GL00/04-19	oz, c	226	172	0.76	12.37	0.2059	0.0007	0.5616	0.0095	15.946	0.283	2874	6

Table

Table 6. SHRIMP U-Pb zircon isotopic data for GL01/01. %c206 is the percentage of non-radiogenic ^{206}Pb estimated from the measured $^{204}Pb/^{206}Pb$ ratio.
Abbreviations for spot types: oz oscillatory zoning.
GL01/01 amphibolite facies tonalitic gneiss, zone of low post-dyke strain near Diabaig NG 799596.

Spot	Type	U	Th	Th/U	%c206	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$ age	±1σ
GL01/01-1	oz	89	53	0.60	0.19	0.2421	0.0014	0.6199	0.0087	20.693	0.328	3134	9
GL01/01-2	oz	77	69	0.90	0.07	0.2409	0.0015	0.6108	0.0090	20.293	0.339	3126	10
GL01/01-3	oz	90	28	0.31	0.34	0.2217	0.0014	0.5647	0.0080	17.264	0.279	2993	10
GL01/01-4	oz	170	105	0.62	0.22	0.2297	0.0010	0.5739	0.0077	18.178	0.265	3050	7
GL01/01-5	oz	156	109	0.70	0.18	0.2439	0.0009	0.6407	0.0085	21.541	0.308	3145	6
GL01/01-6	oz	176	65	0.37	0.11	0.2350	0.0010	0.6179	0.0082	20.018	0.288	3086	7
GL01/01-7	oz	141	53	0.38	0.11	0.2434	0.0009	0.6243	0.0084	20.947	0.300	3142	6
GL01/01-8	oz	91	29	0.32	0.15	0.2409	0.0015	0.6252	0.0087	20.763	0.330	3126	10
GL01/01-9	oz	160	20	0.13	0.07	0.2416	0.0009	0.6321	0.0084	21.060	0.301	3131	6
GL01/01-10	oz	76	35	0.46	0.18	0.2424	0.0013	0.6291	0.0089	21.024	0.332	3136	9
GL01/01-11	oz	208	107	0.51	0.39	0.2406	0.0009	0.6145	0.0081	20.384	0.290	3124	6
GL01/01-12	oz	252	214	0.85	0.12	0.2243	0.0007	0.5793	0.0076	17.914	0.248	3012	5
GL01/01-13	oz	176	126	0.72	0.49	0.2434	0.0010	0.6234	0.0085	20.922	0.308	3142	7
GL01/01-14	oz	132	83	0.63	0.05	0.2427	0.0009	0.6339	0.0087	21.216	0.313	3138	6
GL01/01-15	oz	124	91	0.73	0.25	0.2428	0.0010	0.6395	0.0087	21.411	0.316	3139	7
GL01/01-16	oz	116	331	2.84	0.26	0.2313	0.0012	0.5926	0.0082	18.894	0.289	3061	8
GL01/01-17	oz	84	48	0.58	0.33	0.2419	0.0014	0.6141	0.0088	20.479	0.332	3132	9

Table

Table 7. SHRIMP U-Pb zircon isotopic data for GL01/02. %c206 is the percentage of non-radiogenic ^{206}Pb estimated from the measured $^{204}Pb/^{206}Pb$ ratio.
Abbreviations for spot types: oz oscillatory zoning, ic inherited core, r rim
GL01/02 amphibolite facies granitic melt, zone of low post-dyke strain near Diabaig NG 799596.

Spot	Type	U	Th	Th/U	%c206	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$age	±1σ
GL01/02-1	oz, ic	129	80	0.62	0.60	0.2424	0.0013	0.5860	0.0081	19.583	0.300	3136	8
GL01/02-2	oz,r	433	198	0.46	0.37	0.2164	0.0006	0.5276	0.0067	15.741	0.212	2954	5
GL01/02-3	oz, ic	107	55	0.52	0.18	0.2412	0.0012	0.6041	0.0084	20.095	0.306	3128	8
GL01/02-4	oz,r	607	45	0.07	0.74	0.1933	0.0007	0.3398	0.0043	9.055	0.123	2770	6
GL01/02-5	oz	369	55	0.15	0.44	0.2153	0.0006	0.5791	0.0074	17.194	0.232	2946	5
GL01/02-6	oz	398	178	0.45	0.03	0.2168	0.0006	0.5672	0.0072	16.957	0.226	2957	4
GL01/02-7	oz,ic	143	46	0.33	0.66	0.2402	0.0011	0.5619	0.0076	18.605	0.275	3121	7
GL01/02-8	oz,r	39	22	0.55	2.02	0.1978	0.0040	0.3547	0.0058	9.676	0.265	2808	33
GL01/02-9	oz,r	112	10	0.09	0.16	0.2217	0.0012	0.5580	0.0077	17.053	0.263	2993	9
GL01/02-10	oz	316	47	0.15	0.24	0.2151	0.0006	0.5519	0.0071	16.366	0.221	2944	5
G101/02-11	oz	70	20	0.29	1.14	0.2045	0.0018	0.4921	0.0071	13.876	0.247	2862	15
G101/02-12	oz	277	82	0.30	0.12	0.2151	0.0006	0.5670	0.0073	16.813	0.227	2944	4
G101/02-13	oz, ic	112	75	0.67	0.12	0.2385	0.0012	0.6134	0.0085	20.173	0.309	3110	8
G101/02-14	oz	125	24	0.20	1.59	0.2113	0.0014	0.5539	0.0074	16.135	0.251	2915	11
G101/02-15	oz,r	202	32	0.16	0.44	0.2073	0.0008	0.5779	0.0076	16.520	0.233	2885	6
GL01/02-16	oz,r	115	32	0.28	0.26	0.2148	0.0012	0.5612	0.0077	16.623	0.257	2942	9
GL01/02-17	oz	168	54	0.32	0.32	0.2052	0.0009	0.5462	0.0072	15.455	0.224	2868	7
GL01/02-18	oz	452	216	0.48	0.05	0.2169	0.0005	0.5820	0.0074	17.409	0.232	2958	4
GL01/02-19	oz	266	55	0.21	0.09	0.2190	0.0008	0.6056	0.0080	18.287	0.256	2973	6
GL01/02-20	oz	591	99	0.17	0.04	0.2046	0.0005	0.5408	0.0069	15.257	0.204	2863	4

Table

Table 8. SHRIMP U-Pb titanite isotopic data for GL00/01. *%c206* is the percentage of non-radiogenic ^{206}Pb estimated from the measured $^{204}Pb/^{206}Pb$ ratio.
GL01/01 amphibolite facies tonalitic gneiss, zone of low post-dyke strain near Diabaig NG 799596.

Spot	U	Th	Th/U	%c206	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$ age	±1σ
GL01/01-10	92	17	0.18	1.77	0.1033	0.0015	0.2931	0.0062	4.173	0.113	1684	27
GL01/01-11	82	14	0.17	1.89	0.1017	0.0016	0.2932	0.0063	4.112	0.115	1655	29
GL01/01-12	114	15	0.14	1.30	0.1036	0.0012	0.2970	0.0062	4.241	0.106	1689	21
GL01/01-13	81	15	0.19	3.03	0.1019	0.0020	0.2888	0.0062	4.057	0.125	1658	36
GL01/01-14	91	16	0.18	1.85	0.1004	0.0015	0.2944	0.0063	4.077	0.113	1632	28
GL01/01-15	90	19	0.21	1.56	0.1028	0.0014	0.2941	0.0063	4.169	0.112	1676	26
GL01/01-16	11	2	0.16	11.12	0.1063	0.0106	0.2830	0.0086	4.147	0.448	1737	184
GL01/01-17	168	20	0.12	1.07	0.1009	0.0009	0.2956	0.0061	4.114	0.097	1641	16
GL01/01-18	34	5	0.13	5.50	0.1031	0.0045	0.2862	0.0068	4.067	0.212	1680	80
GL01/01-19	110	14	0.12	1.53	0.1021	0.0015	0.2909	0.0062	4.095	0.113	1663	28
GL01/01-20	203	25	0.12	0.85	0.1030	0.0008	0.2988	0.0061	4.242	0.096	1678	14
GL01/01-3	156	18	0.12	0.93	0.1041	0.0009	0.2949	0.0061	4.234	0.099	1699	16
GL01/01-4	160	27	0.17	1.11	0.1022	0.0009	0.2971	0.0061	4.186	0.099	1664	17
GL01/01-5	24	7	0.28	5.81	0.1010	0.0053	0.2869	0.0073	3.997	0.245	1643	98
GL01/01-6	113	17	0.15	1.54	0.1019	0.0012	0.2911	0.0061	4.090	0.104	1659	23
GL01/01-7	93	17	0.18	1.57	0.1025	0.0016	0.2895	0.0062	4.090	0.114	1669	28
GL01/01-8	72	10	0.14	2.17	0.1017	0.0018	0.2896	0.0063	4.058	0.121	1655	33
GL01/01-9	107	17	0.16	1.46	0.1026	0.0013	0.2922	0.0061	4.135	0.106	1672	23

Table

Table 9. SHRIMP U-Pb zircon isotopic data for GL01/05. %c206 is the percentage of non-radiogenic ^{206}Pb estimated from the measured $^{204}Pb/^{206}Pb$ ratio.
Abbreviations for spot types: *oz oscillatory zoning, tz transgressive zoning, x recrystallised,* c *core, r rim*
GL01/05 retrogressed granulite facies gneiss, Ard Ialltaig, Gairloch NG 806734.

Spot	Type	U	Th	Th/U	%c206	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$ age	±1σ
GL01/05-01	oz,c	166	80	0.4804	4.87	0.1186	0.0013	0.3755	0.0080	6.142	0.155	1936	20
GL01/05-02	oz,c	174	61	0.3495	0.79	0.1223	0.0013	0.3375	0.0065	5.691	0.133	1990	19
GL01/05-03	x, tz, r	823	85	0.1033	2.97	0.1134	0.0004	0.2834	0.0053	4.432	0.086	1855	7
GL01/05-04	oz	134	35	0.2605	0	0.1193	0.0007	0.3191	0.0062	5.250	0.110	1946	10
GL01/05-05	x, tz, r	430	159	0.3687	1.53	0.1162	0.0014	0.2546	0.0050	4.077	0.100	1898	22
GL01/05-06	oz	49	40	0.8149	39.17	0.1267	0.0058	0.3440	0.0088	6.007	0.333	2052	81
GL01/05-07	oz, c	178	75	0.4219	10.37	0.1214	0.0020	0.3547	0.0072	5.936	0.165	1976	30
GL01/05-09	oz, c	179	69	0.3839	2.27	0.1214	0.0019	0.3324	0.0071	5.563	0.157	1977	29
GL01/05-10	oz	129	59	0.4582	0	0.1215	0.0011	0.3426	0.0076	5.739	0.143	1978	16
GL01/05-11	oz, c	437	21	0.0485	5.04	0.1159	0.0009	0.3276	0.0063	5.236	0.113	1894	14
GL01/05-12	oz, c	186	78	0.4182	0	0.1230	0.0008	0.3499	0.0070	5.935	0.130	2001	12
GL01/05-13	x, tz, r	612	27	0.0449	7.76	0.1158	0.0006	0.3331	0.0063	5.319	0.108	1893	9
GL01/05-14	oz	186	91	0.4877	26.85	0.1239	0.0019	0.3680	0.0085	6.286	0.183	2013	27
GL01/05-15	oz	560	125	0.2241	0	0.1178	0.0005	0.3273	0.0062	5.316	0.105	1923	7
GL01/05-16	oz, c	302	30	0.1	2.5	0.1180	0.0010	0.3320	0.0067	5.399	0.124	1925	15
GL01/05-17	oz, c	109	35	0.3232	27.01	0.1168	0.0018	0.3540	0.0065	5.698	0.145	1907	28
GL01/05-18	oz	202	89	0.4408	24.76	0.1186	0.0012	0.3490	0.0060	5.708	0.119	1935	18
GL01/05-19	oz	150	59	0.3893	28.62	0.1237	0.0014	0.3479	0.0063	5.931	0.133	2010	20
GL01/05-20	oz	137	58	0.4208	32.48	0.1197	0.0015	0.3477	0.0061	5.739	0.130	1952	22
GL01/05-21	oz, c	333	33	0.1004	15.13	0.1171	0.0010	0.3428	0.0057	5.536	0.109	1913	15
GL01/05-22	oz	162	72	0.4462	9.87	0.1215	0.0013	0.3452	0.0063	5.781	0.129	1978	19
GL01/05-23	oz, c	247	143	0.5777	21.34	0.1203	0.0010	0.3591	0.0060	5.953	0.117	1960	15
GL01/05-24	oz	175	86	0.4899	13.05	0.1226	0.0013	0.3534	0.0061	5.974	0.127	1994	19
GL01/05-25	x, tz, r	825	24	0.0286	15.14	0.1154	0.0005	0.3150	0.0050	5.013	0.085	1887	8
GL01/05-26	x, tz, r	781	30	0.0383	7.47	0.1152	0.0005	0.3432	0.0054	5.449	0.092	1882	8
GL01/05-27	oz, c	244	107	0.437	10.58	0.1197	0.0009	0.3494	0.0059	5.769	0.111	1952	14
GL01/05-28	x, tz, r	646	52	0.08	24.5	0.1143	0.0007	0.3014	0.0048	4.749	0.084	1868	11
GL01/05-29	oz, c	146	51	0.3462	9	0.1233	0.0013	0.3532	0.0062	6.006	0.129	2005	19
GL01/05-30	x, tz, r	122	34	0.2786	60.93	0.1144	0.0020	0.3257	0.0059	5.137	0.137	1871	32

Table

Table 10. SHRIMP U-Pb zircon isotopic data for GL01/15. %c206 is the percentage of non-radiogenic ^{206}Pb estimated from the measured $^{204}Pb/^{206}Pb$ ratio.
Abbreviations for spot types: *oz oscillatory zoning*
GL01/15 Ard Gneiss, granodiorite, Gairloch NG 806750.

Spot	Type	U	Th	Th/U	%c206	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$age	±1σ
GL01/15-1	oz	506	133	0.26	0.06	0.1175	0.0004	0.3485	0.0035	5.648	0.063	1919	6
GL01/15-2	oz	374	110	0.29	0.14	0.1164	0.0005	0.3453	0.0035	5.539	0.064	1901	7
GL01/15-3	oz	445	97	0.22	0.05	0.1152	0.0007	0.3874	0.0051	6.150	0.093	1882	11
GL01/15-4	oz	428	110	0.26	0.07	0.1159	0.0004	0.3316	0.0031	5.300	0.053	1894	5
GL01/15-5	oz	479	133	0.28	0.10	0.1163	0.0004	0.3278	0.0030	5.257	0.053	1901	6
GL01/15-6	oz	416	81	0.19	0.04	0.1115	0.0008	0.3749	0.0049	5.764	0.090	1824	13
GL01/15-7	oz	495	121	0.24	0.10	0.1163	0.0003	0.3354	0.0031	5.379	0.053	1900	5
GL01/15-8	oz	334	81	0.24	0.06	0.1152	0.0005	0.3531	0.0037	5.609	0.067	1883	8

Table

Table 11. SHRIMP U-Pb titanite isotopic data for GL00/15. *%c206* is the percentage of non-radiogenic ^{206}Pb estimated from the measured $^{204}Pb/^{206}Pb$ ratio.
GL01/15 Ard Gneiss, granodiorite, Gairloch NG 806750.

Spot	U	Th	Th/U	%c206	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$age	±1σ
GL01/15.1	84	485	5.8	2.73	0.1132	0.0029	0.3273	0.0172	5.109	0.313	1852	46
GL01/15-2	68	372	5.4	3.21	0.1122	0.0030	0.3356	0.0177	5.190	0.322	1835	49
GL01/15-3	83	608	7.3	1.99	0.1130	0.0021	0.3356	0.0175	5.231	0.301	1849	33
GL01/15-4	65	352	5.4	1.79	0.1180	0.0022	0.3553	0.0187	5.782	0.335	1927	34
GL01/15-5	47	274	5.8	3.15	0.1095	0.0033	0.3417	0.0181	5.159	0.330	1791	55
GL01/15-6	42	216	5.1	5.96	0.1148	0.0046	0.3483	0.0185	5.515	0.390	1877	73
GL01/15-7	58	314	5.4	1.86	0.1146	0.0023	0.3481	0.0183	5.499	0.321	1873	36
GL01/15-8	50	244	4.9	2.79	0.1136	0.0031	0.3372	0.0178	5.280	0.329	1857	49
GL01/15-9	45	243	5.3	3.32	0.1195	0.0034	0.3508	0.0185	5.780	0.365	1949	51
GL01/15-10	42	203	4.8	3.09	0.1215	0.0036	0.3602	0.0191	6.036	0.386	1979	53
GL01/15-11	40	212	5.3	2.25	0.1160	0.0031	0.3630	0.0193	5.805	0.363	1895	48
GL01/15-12	67	429	6.4	1.85	0.1150	0.0021	0.3497	0.0183	5.545	0.319	1880	32
GL01/15-13	57	326	5.7	1.84	0.1190	0.0023	0.3446	0.0181	5.654	0.328	1941	34
GL01/15-15	47	233	5.0	2.13	0.1166	0.0027	0.3474	0.0183	5.584	0.337	1905	42
GL01/15-16	68	326	4.8	2.18	0.1141	0.0023	0.3450	0.0180	5.427	0.316	1866	36
GL01/15-17	53	348	6.6	1.90	0.1194	0.0025	0.3564	0.0187	5.866	0.345	1947	37