New SIMS U–Pb zircon ages from the Langavat Belt, South Harris, NW Scotland: implications for the Lewisian terrane model

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Abstract: Secondary ionization mass spectrometry (SIMS) U/Pb dating of zircon has been applied to rocks of the Lewisian Gneiss Complex of South Harris (Outer Hebrides, NW Scotland), yielding insights into the complex geological evolution of this region and enhancing our understanding of terranes in the Lewisian Gneiss Complex. Results show that tonalite–trondhjemite–granodiorite magmatism occurred in the Tarbert Terrane between c. 2850 and 2830 Ma, equivalent to similar rocks in other parts of the Outer Hebrides. Magmatism was accompanied or closely followed by high-grade metamorphism at c. 2830 Ma, the first reported occurrence of such an event at this time. These rocks were affected by a subsequent high-grade metamorphic event at c. 2730 Ma, which has correlates both in the Outer Hebrides and in the mainland Lewisian Gruinard Terrane. These data suggest that the ‘terrane’ forming the Outer Hebrides and parts of the mainland Lewisian were once contiguous and were fragmented by post-2730 Ma events. The c. 1700–1660 Ma ages from felsic sheets and pegmatites within the Langavat Shear Zone may record the reassembly of these crustal fragments.

The Archaean high-grade Lewisian Gneiss Complex represents a classic geological terrain that has played an important role in developing our understanding of the generation of granulites, as well as the behaviour of the middle–lower continental crust. Along with the Nuuk region of southern West Greenland, the Lewisian has also been a key location driving the extrapolation of the ‘terrane’ concept to explain the existence of plate tectonics in the Archaean (Friend et al. 1987, 1988; Kinny & Friend 1997; Friend & Kinny 2001). Central to the application of this ‘terrane’ concept to the Lewisian Gneiss Complex is the precise and accurate definition of the ages of protoliths and the metamorphic and deformation events that subsequently affected them (Corfu et al. 1994, 1998; Friend & Kinny 1995, 2001). However, interpretation of the timing of magmatic and metamorphic events and therefore correlations within the complex has been highly controversial, with recent debate driven by the interpretation of complex U–Pb zircon data (Corfu 2007; Friend et al. 2007). In particular, interpretation of event histories has commonly centred on an apparent ‘absence’ of data from single rocks and areas as a criterion for distinguishing between ‘terranes’ (e.g. Love et al. 2004). Therefore, detailed characterization of sample materials and careful isotopic measurement are an important part of unravelling these complex overprinting events.

This paper presents the results of new U–Pb microbeam dating of zircon grains in high-grade Lewisian Gneiss Complex rocks from South Harris, NW Scotland, which reveal a complex history of metamorphism and deformation. The U–Pb zircon data produced have shed new light on the Archaean tectonothermal evolution of the tonalite–trondhjemite–granodiorite (TTG) gneisses on South Harris. In particular, they call into question interpretations that suggest the Outer Hebrides evolved as a series of crustal fragments unrelated to each other and to other ‘terranes’ on mainland NW Scotland before being amalgamated in the Palaeoproterozoic (e.g. Friend & Kinny 2001; Kinny et al. 2005).

Regional geology

Lewisian Gneiss Complex

The Lewisian Gneiss Complex (Fig. 1) is predominantly composed of Archaean high-grade TTG orthogneisses with protoliths as old as c. 2960–3030 Ma (Friend & Kinny 2001) and lesser occurrences of metasediments and metavolcanic rocks, variably reworked during the late Archaean and Proterozoic (Kinny et al. 2005, and references therein). Early interpretations (e.g. Sutton & Watson 1951) viewed the Lewisian as a single contiguous block of Archaean crust that experienced three major orogenic cycles (Sutton & Watson 1962, 1969; Evans & Tarney 1964; Evans 1965; Mooribath & Park 1971; Davies & Watson 1977; Park & Tarney 1987; Heaman & Tarney 1989; Park et al. 1994). The development of high spatial resolution, single-grain U–Pb geochronology (e.g. ion microprobe) integrated with zircon grain imaging techniques (cathodoluminescence (CL), back-scattered electron (BSE) imaging) allowed the discrimination of multiple episodes of zircon growth or isotopic disturbance in Lewisian rocks that experienced complex tectonothermal histories (e.g. Friend & Kinny 1995). Such work suggested that the Lewisian did not comprise a single crustal block, but possibly distinct Archaean crustal units, or terranes, which were assembled at various times during the Proterozoic (Kinny & Friend 1997; Whitehouse et al. 1997; Friend & Kinny 2001; Whitehouse & Bridgwater 2001; Love et al. 2004).

The number of discrete terranes within the Lewisian, the nature of each terrane’s tectonothermal history and the timing of amalgamation or subsequent tectonic reworking are strongly
debated (Love et al. 2004; Kinny et al. 2005; Corfu 2007; Friend et al. 2007). Lewisian evolutionary models vary from simpler scenarios with few terranes that were dissected by subsequent events (e.g. that recently proposed by Park (2005), which questioned the validity of a multiple terrane model) to more complex models involving six mainland Archaean terranes and potentially four or more Outer Hebridean terranes that were assembled during the Proterozoic (Kinny et al. 2005). Crucially, the division of crustal blocks into discrete, apparently unrelated terranes is dependent on the interpretation of complex U–Pb zircon datasets produced using different but equally valid techniques (Corfu et al. 1994, 1998; Love et al. 2004), and has often relied on the absence of specific age data as evidence for these divisions (Kinny & Friend 1997; Friend & Kinny 2001; Kinny et al. 2005). Although the validity of these divisions within the Lewisian is still a matter of debate (Corfu 2007; Friend et al. 2007), the terranes proposed by Friend & Kinny (2001) and Kinny et al. (2005) are useful when discussing the regional character of the complex, and are summarized in Figure 1.

According to this framework, the Outer Hebrides may be divided into four or more discrete terranes (Nis, Tarbert, Roineabhal Terranes, Uist Block), in addition to other blocks (Barra Block, Corodale Gneiss) that may reflect distinct crustal units or fragments of mainland terranes (Fig. 1; Kinny et al. 2005). The following description will focus on two of these terranes: the Tarbert and Roineabhal Terranes.

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**Fig. 1.** Sketch map of the Lewisian Gneiss Complex of NW Scotland (after Kinny et al. 2005) showing the division of units into ‘terranes’ according to the model of Friend & Kinny (2001). The areas underlain by grey shading are inferred in this study to have a similar crustal evolution and therefore should not be strictly considered separate terranes. OHFZ, Outer Hebrides Fracture Zone; SZ, shear zone.
Tarbert Terrane

The Tarbert Terrane (Friend & Kinny 2001), also referred to as the Northern Gneisses (Mason et al. 2004a), forms the majority of the islands of Lewis and Harris (Figs 1 and 2; Friend & Kinny 2001). The terrane is predominantly composed of weakly banded amphibolite-facies TTG gneisses with middle to late Archaean ages (c. 3125–2818 Ma; Pidgeon & Aftalion 1972; Mooribath et al. 1975; Whitehouse 1990; Friend & Kinny 2001; Whitehouse & Bridgwater 2001; Mason et al. 2004a). These gneisses were deformed and metamorphosed during the Archaean, and although the timing is unclear, this tectono thermal event is inferred to have post-dated the intrusion of c. 2818 Ma gneisses (Mason et al. 2004a). On North and South Harris the TTG gneisses are locally affected by in situ anatexis and extensively injected by pegmatitic granite sheets (Jehu & Craig 1927; Dearnley 1962; Bowes 1971; Myers 1971; Mason et al. 2004b), which have U–Pb zircon ages between c. 1683 and 1657 Ma (Friend & Kinny 2001; Mason et al. 2004b). Deformation at amphibolite- and greenschist-facies conditions is inferred to have occurred at ≤1675 Ma (Friend & Kinny 2001), although this deformation may have initiated prior to or contemporaneously with granite sheet intrusion (Mason et al. 2004b).

Roineabhal Terrane

The Roineabhal Terrane (Fig. 2; Friend & Kinny 2001), which includes the South Harris Igneous 'Complex' and the Leverburgh Belt metasediments, has been interpreted as remnants of a Palaeoproterozoic continental volcanic arc (Baba 1997; Whitehouse & Bridgwater 2001; Mason et al. 2004a). This terrane is variably interpreted to include (Friend & Kinny 2001) or not...
include (Mason et al. 2004b) the Langavat Belt, a boundary zone between the Tarbert Terrane and the South Harris Igneous ‘Complex’. The c. 1890–1875 Ma South Harris Igneous ‘Complex’ is a calc-alkaline plutonic suite (Jehu & Craig 1927; Davidson 1943; Dearney 1962, 1963; Fettes et al. 1992; Whitehouse & Bridgewater 2001; Mason et al. 2004a), that has a geochemical signature consistent with formation in an arc setting (Fettes et al. 1992; Bridgewater et al. 1997). Leverburgh Belt metamasides and metavolcanic rocks (Baba 1997) were deposited between c. 2000 Ma and c. 1880 Ma (Cliff et al. 1998; Friend & Kinny 2001; Whitehouse & Bridgewater 2001), and the complex was metamorphosed at high-P granulite-facies conditions ($T > 900 ^\circ$C; Baba 1998, 1999; Schenk & Timmerman 1997) between c. 1890 and 1870 Ma (Friend & Kinny 2001).

Langavat Belt and terrane assembly

Mapped and described in detail by Mason et al. (2004b), the Langavat Belt is a linear, mixed zone of quartzofeldspathic, metabasic and ultramafic rocks, with lesser metasediments. Sta Series metasediments (Fig. 2) have detrital zircon grains with ages that suggest they are younger than the c. 2840 Ma Tarbert Terrane orthogneisses immediately to their north, and were deposited in the late Archaean or early Proterozoic (Mason et al. 2004b). Another unit of deformed quartzofeldspathic rocks within the Langavat Belt, the Borve Series (Fig. 2), is interpreted to reflect intensely tectonized tonalites (Mason et al. 2004b). The Langavat Belt preserves amphibolite-facies assemblages, and did not reach the higher-T metamorphic conditions seen in the Leverburgh Belt.

The ‘Langavat Shear Zone’ (Fig. 2) has resulted in intense and penetrative deformation of the Langavat Belt and southern margin of the Tarbert Terrane (Mason et al. 2004b). At least two distinct phases of deformation are preserved in the Langavat Shear Zone, the first characterized by ductile, hornblende-bearing L–S fabrics, and the second by discrete brittle–ductile structures that rework the earlier ductile features (Mason et al. 2004b). These deformation phases are separated in time by the intrusion of c. 1675–1657 Ma granitic pegmatite sheets (Friend & Kinny 2001; Mason et al. 2004b), suggesting that the initiation of the Langavat Shear Zone must predate these ages (Mason et al. 2004b), and that by inference terrace assembly occurred at an age older than 1675 Ma (Friend & Kinny 2001).

On the basis of a late Archaean–early Proterozoic depositional age for the Sta Series metasediments and inferred tonalitic origin for the Borve quartzofeldspathic gneisses, Mason et al. (2004b) suggested that the Langavat Belt represents a deformed succession of metasediments and metavolcanic rocks that are tectonically interleaved with late Archaean orthogneiss, as originally proposed by Coward & Park (1987).

Sample relationships to Langavat Shear Zone deformation

Sampling for this study focused on samples that would place better age constraints on the early magmatic and metamorphic evolution and subsequent deformation within the Langavat Shear Zone. In particular, this sampling aimed to examine the relationships between and within the older grey gneisses and the variably deformed remnants of cross-cutting granitic sheets. To the NE of the Langavat Shear Zone (Fig. 2b), deformed layered felsic, intermediate and mafic orthogneiss are intruded by weakly deformed granitic dykes or sheets. With increasing proximity to the inlet at Borve Lodge (Fig. 2b) the intensity of strain increases until highly deformed rocks, similar to those described by Mason et al. (2004b) as constituting the Langavat Shear Zone, are encountered. The approximate boundary of the Langavat Shear Zone is marked on Figure 2b. Granitic dykes throughout this zone cut tightly to isoclinally folded layering and layer-parallel foliations in the host layered gneisses. In addition, these dykes are commonly boudinaged and may be folded. In the Finsbay area (sample HA001; Fig. 2c), and the Borve area (samples HA004–HA006; Fig. 2b), the intensity of strain continues to increase towards the SW. Granitic sheets become increasingly attenuated, until they are difficult to distinguish from deformed felsic orthogneiss layers, and rocks locally preserve mylonitic fabrics with feldspar porphyroclasts. Regrression of amphibolite-facies assemblages to chlorite and/or epidote may be associated with these mylonite zones, and may correlate with the Type 2 tectonites of Mason et al. (2004b).

The six samples analysed within this study were taken from within or immediately outside the Langavat Shear Zone (Fig. 2). Five samples (HA001, HA004, HA005, HA006, HA010) were taken from localities within the Langavat Shear Zone, with a further sample (HA003) taken from a less deformed area to the NE of the Langavat Shear Zone and within the northern gneisses of the Tarbert Terrane (Fig. 2).

Analytical methods

Zircon grains were separated using traditional techniques and mounted in epoxy along with fragments of the 91500 and Temora 2 zircon standards. Polished grain mounts were imaged in detail (transmitted and reflected light microscopes, Phillips XL30 Scanning Electron Microscope; School of GeoSciences, University of Edinburgh). Prior to ion microprobe analysis, mounts were repolished using felt laps then cleaned ultrasonically in beakers of petroleum spirit followed by methanol. U/Pb dating of zircons was carried out at the Edinburgh Ion Microprobe Facility (EIMF), using the Cameca ims-1270 ion microprobe (EMMAC, School of GeoSciences, University of Edinburgh). Analytical procedures are similar to those described by Schuhmacher et al. (1994) and Whitehouse et al. (1997), and full details are available online at http://www.geol soc.org.uk/ SUP18316.

U/Pb ratios were calibrated against measurements of the Geostandards 91500 zircon (Wiedenbeck et al. 1995; c. 1062.5 Ma). Measurements over single ‘sessions’ gave a standard deviation on the $^{206}$Pb/$^{238}$U ratio of single repeats of 91500 of about 1% (1σ). Analyses of a secondary, external reference standard (Temora 2) during the analytical sessions yielded a mean $^{206}$Pb/$^{238}$U age of 418.4 ± 1.6 Ma (MSWD = 1.8; 95% confidence limit; $n = 68$). Correction for in situ common Pb has been made using measured $^{204}$Pb counts above that of detector background (0.03–0.08 c.p.s.; 0.2–1.5 ppb) and using modern-day composition of common Pb. Uncertainty on this correction (typically <15 ppb of $^{204}$Pb) is included in the calculation of errors on the U/Pb and Pb/Pb ratios. Data were processed offline using in-house data reduction spreadsheets (R. W. Hinton).

Uncertainties on ages quoted in the text and in tables (see Supplementary Publication) for single analyses (ratios and ages) are at the 1σ level. All uncertainties in calculated group ages are reported at 95% confidence limits. Plots and age calculations have been made using the computer program ISOPLOT/EX v3 (Ludwig 2003).
Zircon morphology and U–Th–Pb data

HA004: deformed tonalite

Sample HA004 is a deformed tonalitic orthogneiss from within the transition zone that marks the approximate boundary to the Langavat Shear Zone (Fig. 2). A well-developed foliation is defined by trails of biotite, locally retrogressed to chlorite, interleaved with elongate quartz and feldspar ribbons. The assemblage includes accessory monazite, apatite and titanite. Zircon grains occur as elongate, stubby or equant, subhedral to euhedral forms, each preserving similar internal zoning characteristics (Fig. 3a and b). Rare grains contain rounded, moderate to strongly luminescent inner cores that exhibit internal oscillatory and planar zoning patterns truncated by oscillatory zoned mantles (Fig. 3a). These inner cores are interpreted to be inherited. Moderate to weakly luminescent oscillatory zoned grains define the dominant population, and occur as both cores and mantles to inherited grains. Oscillatory zoning is typically parallel to external crystal faces of euhedral grains, and may be locally truncated and embayed along curved fronts by younger rims. These rims are typically too narrow for ion microprobe analysis; they are variably weakly, moderately and highly luminescent, and either are homogeneous or preserve patchy, sector or planar banding (Fig. 3a–d).

A total of 18 analyses have been made in 16 zircon grains. Most have low common Pb (typically <0.20% of 206Pb) although three analyses are somewhat higher (11-1, 10.1%; 15-1, 0.59%; 16-1, 0.50%). Ages form a broad cluster near c. 2830 Ma and show minor scatter to younger ages. One analysis older than this cluster (4-1: c. 2931 Ma) is from a grain inferred to reflect an inherited xenocrystic zircon. The 15 analyses made in oscillatory zoned zircon grains have Th/U ratios that range between 0.01 and 0.77 (data are given in the Supplementary Publication, see p. 000). Five of the 15 analyses, which hit zones of alteration or cracking, have been rejected from age calculations (5-1, 11-1, 12-2, 14-1, 15-1). The remaining 10 analyses define a relatively tight cluster of concordant or nearly concordant data that intersects the Wetherill concordia curve at c. 2830 Ma and a weighted average of 207Pb/206Pb ages is 2832 ± 7 Ma (95% confidence limit; MSWD = 5.9; Fig. 4). This average 207Pb/206Pb age is interpreted to reflect magmatic crystallization of the tonalite precursor to the orthogneiss. Of the remaining two analyses, one (3-2: 2821 ± 8 Ma, Th/U = 0.37) was made in a moderate to highly luminescent, sector zoned band that transgresses zoning within an oscillatory zoned grain, possibly reflecting metamorphic modification of zircon after crystallization. The other (16-1: 1691 ± 4 Ma, Th/U = 0.35) is from a rare, weakly luminescent grain with patchy concentric zoning that may reflect partially recrystallized magmatic zoning (Schaltegger et al. 1999; Hoskin & Black 2000).

HA006: intensely deformed intermediate gneiss

Sample HA006 is an intensely deformed, layered intermediate gneiss within the Langavat Shear Zone. Layering is defined by coarse-grained and equigranular to moderately elongate, amphibole-rich domains interlayered with more leuocratic amphibole–biotite–plagioclase–quartz and biotite–plagioclase–quartz layers and minor quartz ribbons. On the basis of internal zoning patterns, three populations have been distinguished in the zircon grains. Population 1 zircon occurs as small, rounded and resorbed inner cores that may be weakly or moderately luminescent, and are interpreted to be inherited xenocrysts (Fig. 3h–j). Typically stubby to moderately elongate in shape (e.g. Fig. 3e and f), Population 2 zircon shows a variation from weakly to highly luminescent and displays clear oscillatory zoning. These grains have near ubiquitous overgrowths (Population 3; Fig. 3h–j) that may embay and truncate zoning, but also form conformable to zoning (Fig. 3i). Population 3 zircon (Fig. 3h), which may form as entire grains or mantles, can be homogeneous, weakly zoned, planar banded or, less commonly, sector zoned with ‘tire-tree’ forms. Planar banded and sector zoned zircon in this sample is indicative of one or more episodes of zircon growth at high metamorphic grade (after Vavra et al. 1996, 1999).

In total, 47 analyses have been made in 34 zircon grains. Common Pb is typically low (<0.3% of 206Pb), although it is as high as 4% in rare analyses. Th/U ratios typically range between 0.09 and 0.93 (one analysis, 2-1c, gave 1.33), with no trend apparent based on zircon type or inferred magmatic or metamorphic origin. Two analyses have been rejected as they comprise overlaps between zones (3-2, 30-1). Seven analyses from three xenocrystic zircon grains (Population 1) show inheritance ages ranging between c. 3645 and 2923 Ma (Fig. 5a), suggesting a significant, if minor, mid- to early Archaean component in the crustal source of the igneous protolith. The remaining data plot along concordia between c. 2900 and 2800 Ma. Although data from Populations 2 and 3 overlap, Population 2 tends to occupy the older end of this spectrum in comparison with Population 3 (Fig. 5a). Twenty-five analyses made in oscillatory zoned zircon (Population 2) form a dense cluster that intersects the concordia curve at c. 2850 Ma (Fig. 5b). A weighted average of all 207Pb/206Pb ages is 2845 ± 8 Ma (95% confidence limit; MSWD = 13). Eliminating three analyses on the basis of excess scatter (2-1, 4-2, B55-1) gives an almost identical result with a minimal reduction in MSWD (2846 ± 8 Ma; 95% confidence limit; MSWD = 9.8). The high MSWD suggests scatter that is well outside analytical uncertainty, probably associated with isotopic disturbance during thermal events subsequent to crystallization. This average age is therefore considered to be a minimum crystallization age.

The remaining 13 analyses (Population 3 zircon) form a broad cluster at c. 2830 Ma with scatter along the concordia curve to ages younger than c. 2700 Ma (Fig. 5a and c). These data define a discordia trend that has an upper intercept with the concordia curve at 2846 ± 43 Ma (lower intercept 1738 ± 380 Ma; MSWD = 2.1). It is important to note that the older ages in this population are taken from grains with internal zoning features that are characteristic of growth at high temperature (e.g. sector zoned mantles that truncate zoning in cores; Fig. 3g). As such, this upper intercept age is considered to reflect a real thermal event in which new zircon growth has occurred, and is not interpreted to reflect partial recrystallization of c. 2850 Ma magmatic zircon by an event younger than c. 2800 Ma. In addition, close inspection of Population 3 data with respect to internal zoning indicates that two of the younger ages (27-1: c. 2749 Ma, 21-1: c. 2734 Ma) come from banded overgrowths on the tips of grains. It is therefore possible that the outer rims and tips reflect a younger metamorphic event than that indicated by the upper intercept age (c. 2834 Ma) and that scatter along concordia in data from Population 1 and 2 zircon grains may result from this overprinting event.

HA010: intensely deformed intermediate gneiss

Sample HA010 is an intensely deformed intermediate gneiss that preserves plagioclase porphyroclasts enveloped by coarse-grained green–brown biotite and quartz–plagioclase ribbons. The amphibi
bolite-facies assemblage in the rock is locally retrogressed to an epidote, white mica and scapolite-bearing assemblage. On the basis of SEM imaging, three generations of zircon have been identified in HA010. Population 1 comprises weakly to moderately luminescent, oscillatory zoned (magmatic) mantle on an inherited xenocrystic core. (b–d) Zircon grains with oscillatory zoning patterns. (e–j) HA006, intermediate orthogneiss. (e, f) Stubby, oscillatory zoned (magmatic) grains. (g) Zircon grain with rounded, brightly luminescent core rimmed by a weakly to moderately luminescent, sector zoned rim. (h) Moderately luminescent, sector zoned rim formed on a resorbed core with partially preserved oscillatory zoning. (i) Zircon grain fragment in which three zones are analysed: a small, resorbed inherited xenocrystic core; brightly luminescent, oscillatory zoned mantle; and a banded, weakly luminescent rim. (j) Multi-generation zircon with an inherited xenocrystic core, partially preserved oscillatory zoned inner mantle (moderately luminescent), brightly and weakly luminescent outer mantles with weakly developed sector zoning, and a semi-continuous banded outer rim. Analysis 3-1 may have partially overlapped the xenocrystic core.

Fig. 3. SEM images of representative and illustrative zircon grains from orthogneiss samples HA004 and HA006. CL (upper or left image in each zircon pair) and BSE (lower or right) images are presented for each example. Scale bars all represent 50 μm. (a–d) HA004, tonalite. (a) Weakly luminescent oscillatory zoned (magmatic) mantle on an inherited xenocrystic core. (b–d) Zircon grains with oscillatory zoning patterns. (e–j) HA006, intermediate orthogneiss. (e, f) Stubby, oscillatory zoned (magmatic) grains. (g) Zircon grain with rounded, brightly luminescent core rimmed by a weakly to moderately luminescent, sector zoned rim. (h) Moderately luminescent, sector zoned rim formed on a resorbed core with partially preserved oscillatory zoning. (i) Zircon grain fragment in which three zones are analysed: a small, resorbed inherited xenocrystic core; brightly luminescent, oscillatory zoned mantle; and a banded, weakly luminescent rim. (j) Multi-generation zircon with an inherited xenocrystic core, partially preserved oscillatory zoned inner mantle (moderately luminescent), brightly and weakly luminescent outer mantles with weakly developed sector zoning, and a semi-continuous banded outer rim. Analysis 3-1 may have partially overlapped the xenocrystic core.
(Fig. 6d) and planar banded (Fig. 6c and e) mantles and rims that may truncate zoning in oscillatory zoned grains. Rims are continuous and most extensively developed at grain terminations. Grains commonly preserve an inner, moderately to highly luminescent inner rim, as well as a weakly to moderately luminescent outer rim that is not always present. Most grains also preserve an outer, highly luminescent rim (Population 3) that is typically a few microns in thickness but may be more extensively developed at grain terminations.

In this sample, 29 analyses have been made in 24 grains. These typically have low common Pb (<0.15% of $^{206}\text{Pb}$), although in one analysis this is exceptionally high (18-1: 0.61%...
of $^{206}$Pb). Th/U ratios are spread between 0.15 and 2.25 (U 31–434 ppm) but do not define any systematic trend with respect to zircon morphology or inferred origin. However, Population 2 zircon (inferred metamorphic origin) has the largest spread and highest Th/U values (e.g. 2.25 for 7-1). Twelve analyses made in oscillatory zoned (Population 1) zircon yield ages that scatter along or near concordia between c. 2830 and 2740 Ma (Fig. 5d and e). These data define an array that intersects the concordia curve at 2759 ± 36 Ma (lower intercept 2218 ± 650 Ma; MSWD = 1.3). However, the scatter of data along concordia
indicates a degree of isotopic mobility within grains subsequent to magmatic crystallization. Therefore, this upper intercept is interpreted to be an estimate of the minimum age for the crystallization of the protolith to the orthogneiss.

Twelve analyses made in sector zoned and planar banded zircon (Population 2), have ages between c. 2750 and 2640 Ma (Fig. 5f), and an upper intercept age of 2729 ± 14 Ma (lower intercept 1559 ± 300 Ma; MSWD = 0.38). By rejecting two analyses because of excess scatter, a ‘concordia’ age (Ludwig 2003) can be calculated: 2735 ± 7 Ma (20; MSWD = 0.006), as well as a weighted average 207Pb/206Pb age of 2735 ± 7 Ma (95% confidence limit; MSWD = 2.3). A further five analyses have been made in weakly banded outer rims (Population 3). After rejection of one analysis because of minor overlap with an inner rim (8-2: c. 2767 Ma), the remaining four analyses define a ‘concordia’ age of 2713 ± 10 Ma (MSWD = 0.56).

**HA003: felsic sheet**

Sample HA003 is a felsic sheet that lies within layering in moderately deformed layered grey orthogneiss in the southern extent of the Tarbert Terrane (Fig. 2). The rock preserves a foliation defined by trails of biotite, plagioclase and quartz, with local retrogression to form an epidote–chlorite–quartz-bearing assemblage. Zircon is dominated by a single population of euhedral, elongate and weakly luminescent grains that have cores characterized by broad homogeneous domains that are overgrown by mantles and rims with well-defined oscillatory zoning (Fig. 6h and m). Some grains preserve rare, small and rounded highly luminescent inherited cores that contain zoning truncated by the weakly luminescent phase.

Ten analyses have been made in 10 grains, focusing on homogeneous cores or oscillatory zoned mantles in the weakly luminescent zircon population. Analyses are typically low in common Pb (<0.03% of 206Pb) with one analysis relatively high (9-1, 0.86%). Zircon grains have Th/U ratios that range between 0.40 and 0.79, and moderate to high concentrations of U (590–2277 ppm) and Th (228–1643 ppm) that are positively correlated. The majority of the analytical population, although close to concordant, is reversely discordant (up to ~6%; Fig. 7a). Reversely discordant zircon secondary ionization mass spectrometry (SIMS) data may potentially result from a number of processes, including local loss of U from the site of a zircon analysis. However, it is more commonly interpreted to be caused by an apparently higher abundance of Pb that can be the result of localized diffusive enrichment in the zircon lattice (Harley & Black 1997; Compston 1999) or enhanced ion yield of Pb during analysis of metamict and/or high-U zircon relative to the pristine matrix of a standard zircon (Black et al. 1986; McLaren et al. 1994; Wiedenbeck 1995). In either case, the loss of U or gain in Pb should not involve fractionation between the different U or Pb isotopes. Provided no ancient Pb loss has occurred, such as during a thermal event subsequent to crystallization, and any discordant data array projects toward a zero age, a weighted average of 207Pb/206Pb ages may still be a robust and accurate age estimate of the zircon population. A weighted average of all 10 analyses produces a 207Pb/206Pb age of 1704 ± 3 Ma (95% confidence limit; MSWD = 2.2), interpreted to be the crystallization age for the felsic sheet.

**HA005: granitic pegmatite boudin (Borve)**

Sample HA005 is a boudinaged granitic pegmatite dyke, preserved within deformed layered gneisses in the Langavat Shear Zone transition zone close to sample HA004 (Fig. 2). The rock is locally retrogressed, with biotite replaced by medium-grained chlorite. On the basis of morphology, zircon grains can be divided into two populations. Population 1 is characterized by...
moderately to highly luminescent grains with variable and at times complex internal zoning patterns. These include oscillatory zoned grains up to hundreds of microns in diameter (e.g. Fig. 6m) that typically show alteration along partially healed fractures, and homogeneous, patchy zoned and sector zoned or planar banded mantles and rims that embay the oscillatory zoned luminescent cores. Population 1 grains may be rimmed by weakly luminescent rims. On the basis of age patterns and structural context, this population is inferred to be xenocrystic. Population 2 is characterized by weakly luminescent zircon that may form whole grains, or rims on Population 1 zircon (Fig. 6j–l). Both grains and rims are typically euhedral, but show localized rounding of terminations. Grains typically preserve intensely altered, weakly luminescent porous and inclusion-rich inner cores that are overgrown by clean, oscillatory zoned mantles (Fig. 6i).

A total of 24 analyses have been made in 21 grains. Of these, nine analyses of Population 1 zircon have been made in both oscillatory zoned and planar banded mantles or rims. 207Pb/206Pb ages range between c. 2956 and 2713 Ma, reflecting magmatism and metamorphism in the source region of the pegmatite at or before this time (e.g. Fig. 6m). The remaining 15 analyses have been made in Population 2 outer mantles and rims that preserve oscillatory zoning. These zircon grains have tightly clustered Th/U ratios ranging between 0.03 and 0.05, and high to very high U concentrations (1377–7645 ppm) that positively correlate with Th. These grains are also elevated in Hf (HfO2 = 2.94–4.46 wt%) that positively correlate with Th/U ratios ranging between 0.03 and 0.05, and high to very high U concentrations (1377–7645 ppm) and elevated Hf concentrations (HfO2 = 2.23–5.90 wt%), identical to morphologically equivalent grains in orthogneiss or felsic sheet samples (HfO2 typically between 1 and 1.5 wt%), which may reflect the late-stage refractory nature of the pegmatite. In comparison with orthogneiss samples, analyses are relatively high in common Pb (6–110 ppm; 204Pb; 0.02–0.34% of 206Pb), and typically show slight to moderate reverse discordance (=8%). Three analyses are significantly more discordant, one showing extreme reverse discordance (B3-1, −73%) and two that are normally discordant (9-1 and 15-1, 23% and 19%, respectively; Fig. 7b). Therefore it is probable that the Population 2 zircon is metamict and that at least some or all of the reverse discordance is an analytical artefact.

Although lying on an extension of a trend through the entire Population 2 dataset, the extremely reversely discordant analysis B3-1 was rejected from age analysis. The remaining 14 analyses define a discordia curve that has an upper intercept with the concordia curve at 1664 ± 6 Ma and a lower intercept at 525 ± 50 Ma (MSWD = 2.81; Fig. 7b). A weighted average of 12 207Pb/206Pb ages (rej ecting points 9-1 and 15-1) gives a poorly constrained age of 1683 ± 9 Ma (95% confidence limit; MSWD = 30). Rejecting a further five points because of excess scatter gives a more tightly constrained age of 1681 ± 2 Ma (95% confidence limit; MSWD = 0.97; n = 7). However, the robustness of this average 207Pb/206Pb age is somewhat questionable. For reversely discordant data produced by zero age fractionation to be entirely reliable, the zircons must not have been affected by significant ancient Pb loss. However, the regression through U–Pb data suggests that some ancient Pb loss has occurred, possibly at c. 540 Ma. As such, the timing of crystallization of the pegmatite is tentatively interpreted to have occurred at a minimum age of c. 1681 Ma.

HA001: granitic pegmatite boudin (Fionnsbhagh)

Sample HA001 is a deformed and boudinaged granitic pegmatite dyke that cuts folded layering and an axial planar foliation in intermediate–mafic orthogneiss near Finsbay, and is lithologically and structurally equivalent to HA005. The pegmatite is composed of a coarse-grained quartz–K-feldspar–plagioclase assemblage with minor chlorite. On the basis of SEM imaging, three zircon populations can be defined. Population 1 zircon is characterized by moderately to brightly luminescent, oscillatory zoned grains that may be well preserved or partly altered. These grains typically contain homogeneous, patchy zoned or planar banded and sector zoned rims with thicknesses ranging from a few to tens of microns. In addition, some grains preserve weakly luminescent, discontinuous outer rims (Population 3). Population 2 zircon comprises cores to subhedral, weakly luminescent grains that are typically easily distinguished from rims and mantles. They have patchy or convolute zoning, although rare examples preserve oscillatory zoning, and are commonly intensely cracked and contain empty ‘pores’ and BSE-bright inclusions of thorite and uraninite (Fig. 6f and g). Weakly luminescent mantles and rims on these cores (Population 3) range in thickness from a few microns to >50 μm, and are more extensively developed on grain terminations. These rims typically preserve well-defined oscillatory zoning and may either embay cores or in some cases include multiple small relics of core material (Fig. 6f and g).

In sample HA001, 21 analyses have been made in 20 zircon grains, yielding two broad age clusters that directly relate to zircon morphology. Common Pb in most analyses is typically low (<0.33% of 206Pb), with higher common Pb analysis obtained from sites that intersected cracks (e.g. 8–1, 8.98%). Four analyses made in moderately to highly luminescent (Population 1) grains have Th/U ratios ranging between 0.06 and 0.20. 207Pb/206Pb ages from these analyses lie between c. 2750 and 2675 Ma. Based on textural and isotopic data these grains are interpreted to be inherited xenocrysts. The remaining 17 data come from weakly luminescent grains, with locations chosen to analyse oscillatory zoned mantles and mantles. These data have low Th/U ratios (0.02–0.05, with one analysis at 0.13), high U concentrations (1764–3318 ppm) and elevated Hf concentrations (HfO2 = 2.23–5.90 wt%), identical to morphologically equivalent grains in HA005. Three analyses (B8-1, B11-1, B13-1) that intersected cracks or straddle partially altered cores have been rejected from age considerations. The majority of the remaining data are reversely discordant (<−5%) but lie along a well-defined chord that has an upper intercept with the concordia curve at c. 1670 Ma and projects towards the origin (0 Ma; Fig. 7c). This data geometry suggests that reverse discordance is most probably a function of zero age fractionation as a result of analysis of metamict zircon, and for reasons discussed for sample HA003, an average 207Pb/206Pb age should represent the age of pegmatite crystallization. The weighted average 207Pb/206Pb age for the population (n = 14) is poorly constrained at 1677 ± 6 Ma (95% confidence limit; MSWD = 25). Rejecting a further two analyses because of excess scatter (B4-1, B10-1) gives a statistically more robust result of 1669 ± 3 Ma (95% confidence limit; MSWD = 2.6, n = 12; Fig. 7c).

Discussion

Archaean magmatism

Oscillatory zoned zircon grains from the three orthogneiss samples have ages that suggest crystallization of protoliths near the end of the Mesoproterozoic, ages that correlate well with previous studies on TTG rocks from the Tarbert Terrane (see Table 1). For example, the age of sample HA006 (2845 ± 7 Ma) is equivalent to a U–Pb TIMS age of 2855 ± 16/–14 Ma for a grey gneiss sample from North Harris (Mason et al. 2004a). Although only a minimum crystallization age can be inferred
from oscillatory zoned zircon from sample HA010 (2765 Ma; Fig. 5e), the correlation between the oldest concordant analysis within this population and the calculated age for magmatic zircon from sample HA004 (2832 ± 7 Ma) would suggest that the protoliths to these orthogneisses crystallized during the same c. 2830 Ma magmatic event. This age is also equivalent to a number of reported ages of TTG rocks within the Tarbert Terrane (Pidgeon & Aftalion 1972; Friend & Kinny 2001; Mason et al. 2004a). The abundance of TTG crystallization ages between c. 2860 and c. 2820 Ma would suggest that these rocks form the dominant component of at least the southern parts of the Tarbert Terrane.

Similar crystallization ages are also reported for many TTG rocks across the Lewisian. These include rocks within the Uist Block (Fig. 1; c. 2834 Ma: Whitehouse & Bridgewater 2001; 2760 Ma: Mason et al. 2004a), Gruinard Terrane (c. 2800 Ma: trondhjemitic, c. 2825 Ma tonalite: Love et al. 2004), and Rhiconich Terrane (c. 2840 Ma: Friend & Kinny 2001).

**Archaean tectonothermal events in the Outer Hebrides**

TTG orthogneisses dated here record evidence of at least two episodes of metamorphic zircon formation (i.e. growth or recrystallization) at c. 2845 Ma and c. 2730 Ma. These provide the first direct record of Archaean tectonothermal events in rocks from the Tarbert Terrane.

Mantles and rims formed on oscillatory zoned zircon in HA006 preserve internal zoning features that are interpreted to have formed during high-grade metamorphism and have ages that show evidence for a degree of post-formation disturbance. However, these data define a coherent Pb-loss trend, the upper intercept age of which is interpreted to reflect the timing of growth of new zircon, or recrystallization of pre-existing zircon. This occurred at, or soon after, the crystallization of the igneous protolith (Fig. 5c). In addition to this sample, a single analysis of a sector zoned (metamorphic) zircon domain in a HA004 zircon has an age of c. 2821 Ma. Metamorphic layering and fabrics in the grey (northern) gneisses led Mason et al. (2004a) to suggest that a high-grade event must have affected the terrane at some time following the intrusion of c. 2818 Ma tonalite. The ages presented here may indicate that the intrusion of the protolith to parts of the grey gneisses was concurrent with metamorphism that occurred at or prior to c. 2818 Ma. The closely spaced nature of TTG intrusion events in the Tarbert Terrane between c. 2860 Ma and c. 2820 Ma and the possibility for incomplete resetting of igneous zircon ages during metamorphic recrystallization make it difficult to reliably assign these ages to particular events. However, it is clear that metamorphism was associated with, or occurred soon after, magmatic accretion in the Tarbert Terrane by c. 2830–2820 Ma.

Although the nature of the c. 2830–2820 Ma metamorphic zircon ages is somewhat ambiguous, evidence for a second Archaean event in the Tarbert Terrane is less so. The internal zoning patterns of mantles and rims on c. 2830 Ma magmatic zircon in HA010 are interpreted to reflect growth at high metamorphic grades, the age of which is well constrained by a ‘cordillera’ age of 2735 ± 7 Ma. In addition, analyses of weakly banded outer rims on these grains give an average age of 2713 ± 10 Ma, the event significance of which is unclear. These ages have not previously been recorded or inferred for the Tarbert Terrane, although a minimum protolith age of >2760 Ma for TTG gneiss on Uist (Fig. 1; Mason et al. 2004a) may reflect minor resetting of c. 2830 Ma zircon by an event at c. 2730 Ma. In addition, c. 2730 Ma granulite-facies metamorphism has been reported for the ‘Barra Block’ (Kinny et al. 2005).

**Archaean tectonothermal events in the Lewisian**

Although not previously reported for the Tarbert Terrane, c. 2740–2730 Ma ages have been published for other locations in the Lewisian Complex, some of which are also somewhat controversial. A subject of some debate is the potential for a high-grade metamorphic event of this age in the Assynt Terrane (Fig. 1). U–Pb thermal ionization mass spectrometry (TIMS) zircon ages from a deformed trondhjemitic sheet at Badcall Bay were interpreted by Corfu et al. (1994) to reflect the magmatic crystallization at c. 2720 Ma, and coupled with the age of metamorphic zircon from mafic gneiss led to the conclusion that the (Assynt Terrane) region experienced high-grade metamorphism at ≈2710 Ma. This interpretation was dismissed by Love et al. (2004) on the basis of U–Pb sensitive high-resolution ion microprobe (SHRIMP) ages of zircon from the same trondhjemitic. These workers instead interpreted a dense, near-cordillera smear of ages between c. 2895 Ma and c. 2480 Ma to reflect crystallization of the sheet following a high-grade event at ≈2900 Ma and intense resetting during c. 2480 Ma Badcallian metamorphism. The poor precision of ion microprobe ages (relative to TIMS) makes the interpretation of such data arrays problematic, and the resolution or absence of c. 2720 Ma ages in these data difficult to establish (see Corfu 2007). However, interpretation of the event history of the Assynt Terrane must also consider c. 2745–2710 Ma monazite core ages reported from Scourie metasediment (Zhu et al. 1997), which correlate

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**Table 1. Summary of ages and events recorded by zircon grains in the South Harris rocks**

| Age (Ma)       | Event type                                      | Sample         | Correlation                        |
|---------------|------------------------------------------------|----------------|-----------------------------------|
| 3645–2920     | Inherited magmatic zircon                      | HA006          | Assyt Terrane                     |
| 2845 ± 8, 2832 ± 7, >2760 | Tonalitic magmasim                             | HA006, HA004, | Uist Block, Gruinard Terrane, Rhiconich Terrane |
| 2846 ± 43     | High-T metamorphism                            | HA010          |                                   |
| 2735 ± 7      | High-T metamorphism                            | HA010          | Uist Block, Barra Block, Gruinard Terrane, Assyt Terrane? |
| 2713 ± 10     | Metamorphism?                                   | HA010          |                                   |
| c. 1700–1650  | Metamorphism, anatexis and partial resetting of zircon | HA004, HA006, |                                   |
| 1704 ± 3      | Felsic sheet                                    | HA003          |                                   |
| 1669 ± 2.6, 1681 ± 2.2 | Felsic pegmatite                              | HA001, HA005, |                                   |
|               |                                                | Uist Block, Rona Terrane, Rhiconich Terrane |                                   |

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NEW U–Pb AGES FROM THE LEWISIAN COMPLEX 977
with the synedfomational trondhjemite age interpretations proposed by Corfu et al. (1994).

More conclusive evidence for tectonothermal activity at c. 2730 Ma comes from the Gruinard Terrane (Fig. 1). Syntectonic trondhjemitic magmas were interpreted to have crystallized at c. 2758 Ma (Whitehouse et al. 1997) and c. 2730 Ma (Corfu et al. 1998). As these sheets cut deformed c. 2800 Ma tonalite, c. 2850 Ma zircon cores in these trondhjemites were interpreted by Whitehouse et al. (1997) to be inherited xenocrysts. In contrast, Love et al. (2004) interpreted these older ages to reflect the timing of tonalite (c. 2830 Ma) and trondhjemite (c. 2858 Ma) emplacement, and considered rim ages of c. 2730 Ma in both rocks to reflect growth of metamorphic zircon. Despite this disagreement, there is consensus that a high-grade metamorphic event affected the Gruinard Terrane at c. 2730 Ma, directly correlating with the same event age documented in this study from the Tarbert Terrane.

Palaeproterozoic events: the Langavat Shear Zone

The ages for a felsic sheet (c. 1704 Ma; HA003) and two granitic pegmatite dykes (c. 1669 Ma; HA005, HA001) are consistent with ages already reported for lithologically similar rocks that have the same structural contexts in South and North Harris (Friend & Kinny 2001; Mason et al. 2004b; Mason & Brewer 2005). Friend & Kinny (2001), who dated granitic and pegmatite sheets in North Harris (c. 1683 Ma) and within the Langavat Shear Zone (c. 1674 Ma), have argued that these sheets do not occur in the Rhoinnebhal Terrane to the south of the Langavat Shear Zone and therefore that they provide an apparent upper age limit to the juxtaposition of the Tarbert and Rhoinnebhal Terranes, which must have occurred after the injection of these felsic sheets and pegmatites (<1675 Ma). However, detailed field mapping and U/Pb zircon dating by Mason & Brewer (2005) contradicts these findings, suggesting not only that granitic pegmatites intrude the Rhoinnebhal Terrane (see also Dearnley 1963) but also that they are syntectonic with respect to the Langavat Shear Zone itself (Mason et al. 2004b).

The c. 1704 Ma felsic sheet dated in this study, sampled to the north of the Langavat Shear Zone, is interpreted to form part of the ‘Harris injection complex’ (Fettes et al. 1992), an extended area of migmatized Archaean gneisses and felsic sheets within the Tarbert Terrane. Friend & Kinny (2001) dated c. 1672 Ma zircon crystallized within metatexite developed in a c. 3.1 Ga North Harris tonalitic gneiss. They suggested that this age reflected the development of the injection complex and that older (c. 1715 Ma) ages derived from discordant zircon analyses in TTG gneisses (Pidgeon & Aftalion 1972) and granites (van Breeman et al. 1971) represent apparent ages. However, the data presented here suggest that portions of the injection complex are at least as old as c. 1704 Ma, and that the older ages produced by Pidgeon & Aftalion (1972) and van Breeman et al. (1971) are not unreasonable. Furthermore, the structural context of granitic pegmatite dykes in the Langavat Shear Zone (Mason et al. 2004b) indicates that deformation associated with the Langavat Shear Zone initiated prior to c. 1675 Ma, and as such is most probably causally linked to the generation of the injection complex on Harris. We interpret that the development of the injection complex, which reflects amphibolite-facies metamorphism and melting during contractional deformation between the Tarbert and Rhoinnebhal Terranes (Friend & Kinny 2001), commenced prior to c. 1704 Ma. Deformation in the Langavat Shear Zone then evolved through time to develop ductile to more brittle structures (after Mason et al. 2004b; Mason & Brewer 2005), punctuated by the injection of late-crystallizing melts (c. 1670–1685 Ma pegmatite dykes; Friend & Kinny 2001; Mason et al. 2004b; Mason & Brewer 2005; this study).

The c. 1704 Ma and c. 1675 Ma Palaeoproterozoic ages from Harris have correlates elsewhere in the Lewisian Complex, indicating a degree of regional impact to these events on Harris. A titanite ‘cooling age’ of c. 1642 Ma from tonalitic gneisses of South Uist (Mason et al. 2004a), finds similar correlatives in the mainland ‘Rona Terrane’, where titanite ages are interpreted to infer amphibolite-facies reworking at c. 1670 Ma (Love et al. 2004). Further mainland imprints of this event are found in the Rhiconich Terrane near Laxford Bridge, where an amphibolite-facies overprint at c. 1670 Ma (‘Somerledian’; Kinny et al. 2005) has been inferred on the basis of titanite and rutile ages (Corfu et al. 1994). In addition, rutile ages of c. 1685–1680 Ma (Heaman & Tarney 1989; Corfu et al. 1994) have been reported for the Scourie area in the Assynt Terrane. Such widespread occurrence of c. 1685–1640 Ma ages across the Lewisian suggests that a degree of tectonothermal reworking of the other mainland and outer Hebridean terranes was associated with the higher-grade deformation and metamorphism experienced on Harris at this time.

Implications for the Lewisian Complex: a revision of the terrane concept?

The realization that the Lewisian Complex of NW Scotland was potentially composed of not one, but multiple crustal fragments assembled at different times through the complex’s history (Friend & Kinny 1995, 2001; Kinny & Friend 1997), revolutionized our understanding of and approach to the area and provided a valuable framework within which to address problems relating to its tectonothermal evolution. The application of terrane theory, as applied both to the Lewisian and previously to the Nuuk region of southern West Greenland (e.g. Friend et al. 1987, 1988), requires that for the existence of terranes to be established, the area must be divided into discrete crustal blocks bounded by faults (or higher grade equivalents) with each block preserving clearly distinct protolith ages and tectonothermal histories prior to amalgamation, and a common history thereafter. However, potential flaws in this theory as applied to high-grade terrains such as the Lewisian include terrane distinctions made on the basis of metamorphic grade, and most critically, the ‘absence’ of evidence. For example, a criterion used to separate the Rhiconich Terrane from the Assynt and Gruinard Bay Terranes is the lack of granulite-facies assemblages in the former (Friend & Kinny 2001; Love et al. 2004), an interpretation that ignores the potential for multiple crustal levels of a single terrain to be juxtaposed. More critical are interpretations made on the basis of a lack of evidence, in particular geochronological evidence, such as the distinction between the Assynt and Gruinard Terranes. In this case, the Assynt Terrane is interpreted by some workers (Friend & Kinny 1997; Love et al. 2004) to have not ‘seen’ the c. 2730 Ma granulite-facies event prevalent in the Gruinard Terrane. The strength of such an interpretation heavily depends on the clarity of U/Pb zircon data, which in the Assynt terrane is hotly contested (e.g. Love et al. 2004; Corfu 2007; Friend et al. 2007). In these cases the at times highly unreactive nature of zircon in some metamorphic environments coupled with its ability to be reset in others masks our ability to adequately interpret event histories.

The poly-metamorphic nature of the Lewisian has commonly resulted in U/Pb zircon data from many areas being ambiguous
and notoriously difficult to interpret (Friend & Kinny 1995; Kinny & Friend 1997; Love et al. 2004). As such, the interpretation of crystallization and thermal event ages has caused controversy over the applicability of the ‘terranes’ theory to the Lewisian. The new data presented in this study shed light on the interpretation that the Tarbert Terrane is a distinct crustal unit from other Hebridean and mainland terranes.

TTG gneisses of the Tarbert terrane have protolith ages that predominantly fall between c. 2850 and 2820 Ma (Friend & Kinny 2001; Mason et al. 2004a; this study), with rare older c. 3125 Ma components (Friend & Kinny 2001). The Tarbert Terrane therefore shares protolith ages with the Uist and Barra Blocks, as well as the mainland Gruinard Bay and Rhiconich Terranes (Whitehouse et al. 1997; Friend & Kinny 2001; Whitehouse & Bridgwater 2001; Mason et al. 2004a). Links between the Tarbert Terrane rocks and the Uist Block are further supported by the geochemical similarity of mafic dykes in these two areas, which are interpreted to have formed through rifting of the Archaean gneisses, possibly at c. 2.0 Ga, prior to reassembly at >1675 Ma (Mason & Brewer 2004).

Although any link to the Rhiconich terrane on event grounds is difficult, further evidence implies that links may exist with other areas in the Outer Hebrides and with the Gruinard Terrane. Within the Uist Block, the protolith ages of tonalitic gneisses are partially reset by an event younger than c. 2760 Ma (Mason et al. 2004a), and granulite-facies reworking is reported for the Barra Block at c. 2730 Ma (Love et al. 2004; Kinny et al. 2005). The latter was tentatively linked by Kinny et al. (2005) to the Gruinard Terrane, which also was reworked by granulite-facies metamorphism at this time (Whitehouse et al. 1997; Corfu et al. 1998; Love et al. 2004). With solid evidence for similarity in protolith ages, the comparable timing of high-grade metamorphism at c. 2730 Ma in the Tarbert and Gruinard Terranes, and supporting evidence for events at this time in the Uist and Barra Blocks, we argue that there is little basis for a division of these areas into separate ‘terranes’. Instead, the present disposition of these ‘blocks’ reflects the fragmentation and reshuffling of an originally coherent crustal section by events subsequent to c. 2730 Ma.

The correlation of ages between the Tarbert Terrane and Uist–Barra Blocks for c. 2730 Ma metamorphism and older magmatic protoliths also calls into question the context of the Roineabhail Terrane (Fig. 1). The similarities in magmatic and metamorphic history that exist between the Tarbert and Uist rocks, in particular the occurrence of chemically equivalent mafic dykes in both areas that have continental rift affinities (Mason & Brewer 2004), are suggestive that these ‘terranes’ were contiguous. It is important to note here that the 1890–1870 Ma magmatic and metamorphic ages recorded in the South Harris Igneous ‘Complex’ and Leverburgh Belt metamadies of the Roineabhail Terrane (Friend & Kinny 2001; Mason et al. 2004a) are not, as yet, reported for the Tarbert Terrane, Uist and Barra Blocks or Langavat Belt. Therefore, on an ‘absence of evidence’ basis there is no reason to assume that the Roineabhail Terrane was in place (in fact, its linkage with adjoining areas is problematic) prior to c. 1700 Ma. The lack of any evidence in the Tarbert Terrane for events related to the development of the South Harris Igneous ‘Complex’ arc rocks indicates that the latter’s construction and evolution most probably occurred some distance from the rocks against which it is now juxtaposed. A simplified scenario for the current disposition of these areas involves the translation of the Roineabhail Terrane during transpressional reassembly of the Uist and Tarbert crustal fragments. So, although this supports the interpretation that the Roineabhail rocks constitute a terrane, it does not require the separate treatment of the Uist and Tarbert Blocks.

Conclusions
Detailed U–Pb isotopic dating of zircons from South Harris in the Outer Hebrides (NW Scotland) has yielded new insights into the geological history of this part of the Lewisian Gneiss Complex, and into the evolution of the complex as a whole. Important conclusions are as follows.

(1) There is a possible Palaeo- to Eoarchaean source component to some Outer Hebrides TTG orthogneisses.

(2) A protracted episode of TTG magmatism occurred in the Tarbert Terrane between c. 2850 and 2820 Ma, similar to that in other areas of the Lewisian. This TTG magmatism was accompanied by high-grade metamorphism, the first such event reported for the Tarbert Terrane.

(3) A subsequent episode of zircon growth occurred during high-grade metamorphism at c. 2730 Ma, which correlates with tectonothermal events in other Outer Hebridean and Mainland ‘terranes’, and indicates that the number of discrete, unrelated crustal blocks within the Lewisian complex should be revised. A comparison of U–Pb zircon data suggests that the Archaean-aged TTG-dominated crustal blocks forming the Outer Hebrides and the mainland ‘Gruinard Terrane’ were once contiguous and were subsequently fragmented by post-2730 Ma events.

(4) The ages of syn-deformation pegmatites are consistent with previous estimates (≥1670 Ma). However, a c. 1704 Ma age for a felsic sheet from the southern edge of the Harris Inlection Complex places a minimum age on the initiation of contractional deformation between the Roineabhail and Tarbert Terranes.

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