Mid-Devonian sinistral transpressional movements on the Great Glen Fault: the rise of the Rosemarkie Inlier and the Acadian Event in Scotland.

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Abstract

The Rosemarkie Inlier is a small fault-bounded lens of interleaved Moine psammites and possible Lewisianoid orthogneisses with distinctive leucogranite veins and pods that lies adjacent to the Great Glen Fault (GGF). The basement rocks and most of the leucogranites are strongly deformed and tightly folded with foliations generally steeply dipping and a locally well-developed NE-plunging rodding lineation. Mid-Devonian sandstone and conglomerate unconformably overlie the inlier on its western side. Monazite from a deformed leucogranite vein gave a mean ID-TIMS $^{207}Pb/^{235}U$ age of 397.6 ± 2.2 Ma and acicular zircons gave a compatible concordant ID-TIMS U-Pb age of 400.8 ± 2.6 Ma, dating emplacement as mid-Devonian. Xenocrystic zircons from the leucogranites and complex zoned zircons from two adjacent tonalitic gneisses gave LA-MC-ICP-MS concordant ages between 2720 and 2930 Ma confirming their Archaean Lewisianoid origin. Leucogranite emplacement is interpreted to mark the onset of Acadian transpression and sinistral strike-slip movement on the GGF that resulted in multi-phase deformation and oblique exhumation of the Rosemarkie Inlier. The sequence and structure of the Early Devonian Meall Fuar-mhonaidh Outlier, 32 km farther SW along the GGF, are also linked to this tectonic event, which was apparently localised along the main terrane-bounding faults in Scotland.

End of Abstract

The Great Glen Fault (GGF) is a major geological and topographical feature that transects the Highlands of Scotland, separating the Grampian Highlands to the southeast from the Northern Highlands to the northwest. The fault passes offshore into the Moray Firth (Fig. 1), where it forms a major sub-vertical structure (Andrews *et al*. 1990). The fault can be traced for a further 23 kilometres northeast into the West Moray Firth Basin into deformed Mesozoic strata (Bird *et al*. 1987; Underhill 1991). The full role of the GGF in the geological history and tectonic development of the Scottish Highlands remains unclear, but it appears to have acted as a Neoproterozoic basin-bounding fault (Banks & Winchester 2004) and has undoubtedly been the focus of significant sinistral movements during the Palaeozoic (Johnstone & Mykura, 1989; Stewart *et al*. 1999). Minor sinistral, dextral and vertical movements followed in Mesozoic and Cenozoic times (Rogers *et al*. 1989; Andrews *et al*. 1990; Underhill & Brodie, 1993; Roberts & Holdsworth, 1999). Exposure along the Great Glen is generally poor, but farther SW, mylonites and blastomylonites attest to ductile shearing at mid-crustal levels (9-16 km) with later cataclasite, phyllonite and breccia development reflecting shallower level brittle movements (Stewart *et al*. 1999).

The Rosemarkie and Cromarty inliers crop out adjacent to the GGF surrounded by Devonian rocks (Fig. 1). The Rosemarkie Inlier, some 2 km wide and 9 km long, lies on the Black Isle adjacent to the GGF whose trace runs up to 500 m offshore (Fig. 2). It exposes deformed amphibolite-facies psammites, subsidiary semipelites, amphibolitic mafic bodies and laminated felsic and mafic gneisses, all cut by abundant, typically salmon pink, leucogranite veins and sheets (Rathbone & Harris 1980; Fletcher *et al*. 1996). Exposure is effectively limited to coastal outcrops (Fig. 3a), with a few weathered inland outcrops. To the SW and NE the inlier is fault-bounded, but on its NW side mid-Devonian sandstones and conglomerates unconformably onlap the inlier. Palynological data from this Orcadian sequence show that its basal beds were deposited in the late Eifelian at c. 393 Ma (Marshall *et al*. 2007). The main psammitic and gneissose lithologies are similar to those of the Neoproterozoic Moine succession and Archaean Lewisianoid inliers of the Northern Highlands respectively, but they differ considerably from the nearest Moine rocks, Loch Eil Group psammites that crop out some 20 km to the NW.

The rocks of the inlier are widely altered, fractured and crushed with breccia and gouge developed. The exception to this is a section below Learnie Farm between [NH 760 712] and [NH 767 620] where brittle deformation was limited and earlier ductile deformation structures are seen clearly. Here, Rathbone & Harris (1980) recognised four discrete phases of ductile deformation with the leucogranites deformed by the latter three phases. The majority of the leucogranite veins show clear evidence of intrusion at an early stage of the main deformation (D_2) (Rathbone & Harris 1980). The deformation phases in the inlier have been correlated with those of the Moine rocks and hence the leucogranites were thought to be of Ordovician age or older (Rathbone 1980).

The inlier was exhumed after emplacement of the leucogranite veins but prior to deposition of the adjacent late Eifelian to Givetian sequence. To provide a lower age constraint on deformation and exhumation, samples of leucogranites were collected for U-Pb isotopic dating. Samples of the adjacent gneisses were also taken to ascertain their relationship with the Lewisianoid gneisses that underlie the Moine Supergroup of the Northern Highlands. Stratigraphical units and related ages quoted in this paper are based on the current International Commission on Stratigraphy chart that largely follows Gradstein *et al*. (2004).

Tectonic Setting

Studies of the later stages of the Caledonian Orogeny in the Northern Highlands have focused on the formation of the Moine Thrust Zone at its western margin, and on the uplift and intrusion history of its interior and its southeast side. The Moine Thrust Zone was active mainly during the Scandian event between c. 437 Ma and c. 430 Ma (Johnson *et al*. 1985; Dallmeyer *et al*. 2001; Goodenough *et al*. 2006) with later extensional movements continuing spasmodically to 408 Ma (Freeman *et al*. 1998). Dewey & Strachan (2003) argued that Scandian deformation is absent from the Grampian Highlands to the SE of the GGF, and hence postulated that at least 700 km of sinistral movement took place along the GGF between 425 Ma and 395 Ma. The waning phases of the Caledonian Orogeny in the Northern Highlands were marked by intrusion of granitoid plutons coeval with regional uplift and significant lateral movements on the main NE-trending faults (Watson 1984). Documented examples include the Clunes Tonalite (428 ± 2 Ma; Stewart *et al*. 2001) and the Strontian Pluton (425 ± 3 Ma; Rogers & Dunning, 1991), both linked to movements on the GGF (Hutton 1988), and the Ratagain Pluton (425 ± 3 Ma; Rogers & Dunning 1991), linked to movements on the Strathconon Fault (Hutton & McErlean 1991).

The main development of the Acadian Orogeny lies in eastern North America where Avalonia collided with Laurentia; the resultant deformation, metamorphism and related igneous intrusion events lasted from 420 – 395 Ma (Van Staal *et al*. 1998; Van Staal & Whalen 2006; Zagorevski *et al*. 2007). In the British Isles the Acadian event resulted from early collision of an Armorican microcontinent with the Avalonian part of Laurussia at the northwest margin of the Rheic Ocean. The collision caused the Midland platform (microcraton) to indent the Welsh Basin and Late Palaeozoic basins of central and northern England (Woodcock *et al*. 2007). Acadian deformation and related metamorphism occurred in the Lake District, Wales, and southern Britain between 400 and 390 Ma (Soper & Woodcock 2003; Sherlock *et al*. 2003). Acadian volcanic rocks seem to be absent from the British Isles and 'Acadian' granites are restricted to the Southern Uplands and the Lake District. The nature of the Acadian event in Scotland is equivocal, although structures in the Midland Valley, e.g. the Strathmore Syncline, have been attributed to mid-Devonian (c. 400 Ma) sinistral transpression (Soper *et al*. 1992; Jones *et al*. 1997).

In Baltica the Acadian event is absent and there is only evidence of an extended history of Devonian uplift and exhumation of the Western Gneiss region of Norway (Krabbendam & Dewey, 1998; Johnston *et al*. 2007; Walsh *et al*. 2007) and formation of large sinistral transtensional basins (Osmundsen & Andersen, 2001; Eide *et al*. 2005). Uplift and extension here lasted from at least 410 Ma through to 370 Ma.

The lithology and structure of the Rosemarkie Inlier

Hugh Miller (in 1885) and John Horne (in 1890) mapped the Rosemarkie Inlier during the primary geological survey and brief descriptions of the lithologies and petrography are given in the Geological Survey of Scotland Memoir for the area (Horne, 1923). Horne noted the distinctive character of the rocks and the abundance of alkali-felspar-rich granitic material. He commented on their similarities to the Moine psammites, but also speculated that the hornblendic felsic gneisses may equate to the 'Lewisian floor' to the Moine succession (i.e. the Lewisianoid gneisses). He even suggested (p. 58) that the rocks may be a 'distinct group of Moine rocks brought up by the Great Glen Fault'. Subsequently, P. A. Rathbone carried out detailed work on the inlier as part of his Ph.D. (Rathbone 1980; Rathbone & Harris 1980) and A. J. Highton remapped the southern part of the Rosemarkie Inlier and questioned the nature of the protolith of the felsic and mafic gneisses (Fletcher *et al*. 1996). Much of the structural data used here has been abstracted from Rathbone (1980).

Lithology

The Rosemarkie Inlier consists of grey, flaggy, typically thinly banded siliceous to micaceous psammites with subsidiary semipelites and pelites. Thin quartzofeldspathic lenticles impart a weakly gneissose appearance to the rocks. The psammites are interleaved with laminated to thinly banded felsic and mafic gneisses on scales varying from a few centimetres to ten of metres. The felsic gneisses consist essentially of quartz-plagioclase-biotite with variable hornblende content and are interlaminated with abundant amphibolitic mafic gneisses. Thicker amphibolitic mafic units and hornblendic ultramafic lenses, locally with agmatitic net-veins, also occur within these gneisses, features characteristic of the basement Lewisianoid inliers within the Moine succession. No obvious shear zones or dislocations can be identified at psammite-gneiss contacts and in places the distinction is quite cryptic. Discrete mafic amphibolite sheets and lenses are also common in the psammites, semipelites and the gneisses. These lithologies all contain a strong layer parallel fabric that is folded by F_2, F_3 and F_4 folds. The rocks show evidence of pervasive recrystallisation with quartz, feldspar, hornblende, biotite and muscovite defining a composite S_1-S_2 fabric. The metamorphic assemblages are characteristic of lower amphibolite facies metamorphism although index minerals are largely absent and retrogression effects are widespread. Elongate garnet porphyroblasts are developed in the pelitic units and deformed by the D_2 crenulation fabric, whereas the abundant small muscovite and shimmer aggregate porphyroblasts overprint the main S_2 foliation (Fletcher *et al.*1996).

Pink to red, foliated and lineated leucogranite sheets, lenses and veins are diagnostic of the Rosemarkie Inlier. The intrusions are typically 0.3 to 1 m wide but range from a millimetre up to 5 m in thickness. They are generally parallel sided and show sharp planar contacts with the country rocks (Fig. 3a), but some very thin veins do show diffuse margins. Although strongly deformed, the leucogranite veins are clearly discordant to the host banding in numerous instances (Rathbone & Harris, 1980). Typically, the angle of discordance is <5° but locally high angles are seen. The granite is variable from fine-grained to coarse-grained and partly pegmatitic; in parts it contains pink potash feldspars, typically augened. Its mineralogy is essentially quartz, potash feldspar and plagioclase, with minor muscovite and biotite and accessory zircon, magnetite or ilmenite, and rare apatite, monazite and titanite. Secondary chlorite (after biotite), zoisite, carbonate and rarely sodic amphibole are developed (Fletcher *et al.* 1996). In some areas white muscovite-bearing leucogranite sheets and veins intrude the Moine psammites and semipelites; they show similar features to the pink veins. Although many of the leucogranite veins show evidence of strong deformation, others show deformation features, lower strains and mineralogies indicative of lower temperature and brittle shearing, suggesting they were emplaced at higher crustal levels.

Structure

Rathbone & Harris (1980) recognised four deformation phases in the Rosemarkie Inlier. The planar fabrics parallel to the compositional banding and fine-scale interleaving of Moine psammites and Lewisianoid gneisses were attributed to the D_1 event. Tight to isoclinal minor folds are moderately abundant and are attributed to D_2. A related planar schistosity (S_2) and associated lineation (L_2) are pervasively developed. F_3 open to tight folds demonstrably refold the F_2 folds and S_2 fabrics and are abundant on a small- and medium-scale. Their axes normally plunge moderately to the NE and verge towards the SE. However, Rathbone & Harris (1980) noted that F_3 fold hinges are commonly curvilinear through up to 120°. Their axial planes are typically upright with an S_3 schistosity widely developed. D_4 folds control much of the variation in strike and dip. They also plunge towards the northeast but their vergence is towards the northwest. Their axial planes are generally upright, and only rarely is an associated cleavage developed. F_4-F_3 and F_3-F_2 fold interference patterns are seen in the psammite-semipelite lithologies and in the felsic and mafic gneisses. At [NH 765 615] interleaved Moine and Lewisianoid rocks are folded by a very tight metre-scale fold (F_2) that is in turn refolded by a D_3 synform. The leucogranite veins contain a strong L-S fabric defined by strongly attenuated quartz and feldspar with minor thin stringers of biotite and sparse muscovite development. The lineation and foliation is contiguous with L_2 and S_2 in the adjacent Moine and Lewisianoid rocks where it is defined by quartz, feldspar and in the mafic rocks, hornblende alignment. L_2 in the leucogranite is a millimetre-scale rodding of quartz and pink feldspar; it locally dominates to give an L-tectonite. In places there are spectacular F_2-F_3 interference folds involving the leucogranite veins (Fig. 3b). Rathbone & Harris (1980, Figure 4) documented examples of fold interference patterns involving the leucogranite veins and also showed that the prominent quartz lineation (L_2) was locally modified by later D_3 structures.

Fenitization, carbonate veining, brecciation and minor faulting dominate the southern exposures in the Rosemarkie Inlier, but on the coastal section below Learnie Farm later brittle deformation effects are minimal. Here, the banding/foliation strikes mainly between northeast and north, dips range from vertical to moderately eastwards, and L_2 plunges northeast at moderate angles (37° to 050°) (Fig. 4). L_2 is co-linear with the majority of the F_2 and F_3 axes, as indicated by the distribution of poles to foliation/bedding.

Leucogranite Textures

In thin section the foliated and lineated leucogranites from Learnie Shore are dominated by quartz and feldspar ‘ribbons’, the latter being lenticular, typically measuring 8 to 12 mm long and 0.3 mm to 1 mm wide in the S-L plane. The potash and plagioclase feldspars are disaggregated and fragmented with some sericitisation and new quartz growth. In the quartz ribbons grain-size reduction has occurred giving rise to recrystallized aggregates 0.03 mm to 0.6 mm across, that exhibit tessellate grain boundaries, strain shadows, fine inclusion trails and extensive sub-grain development. Small ragged' biotites, in part altered to chlorite, in parts form discontinuous trails marginal to the feldspar laminae. Ilmenite or magnetite form irregular altered aggregates and may have released iron to give the prominent pink to red feldspar colouring. The 'ribbons' have formed in response to the strong deformation with the elongate feldspars now effectively being porphyroclasts. Thin sections normal to the lineation show potash feldspar and plagioclase porphyroclasts commonly 3 to 4 mm across. They show strain twinning, embayed margins and locally marginal myrmekite development. The typical porphyroclast size and dimensions of the ribbons suggest that the leucogranite was originally a medium- or even coarse-grained granite. The petrographic features in the leucogranites are compatible with strong deformation under lower amphibolite-facies conditions. As the lineation clearly defines the stretching direction (Ls), Rathbone (1980) measured the shapes of the quartz aggregates in the deformed leucogranite using them to define the principal planes of the strain ellipsoid. Taking the quartz as initially equant, a common circumstance in granites, he obtained X:Y:Z values of 18:2.5:1. This strain is prolate with a *k* value of 2.88 (Flinn 1962).

In parts the leucogranite shows an augen texture with potash feldspars up to 1 cm across. Some appear to reflect relic feldspar crystals but other have grown during deformation and recrystallization. Although some augen show δ and σ tails that imply a shear sense, many have a neutral geometry. M Stewart (*pers. com*. 2004) noted that the shear sense was generally consistent within individual intrusions and in the pink leucogranite sheets was generally sinistral. In contrast the white muscovite-bearing sheets generally had internal and marginal fabrics implying a dextral shear sense. Stewart also recorded that there was evidence of later more brittle dextral shearing, generally focused at the margins of the leucogranite veins and in the more pelitic units. This deformation was accompanied by extensive chlorite growth.

Cromarty Inlier

The Cromarty Inlier, measuring c. 9 km x 2.7 km, lies immediately northeast of the Rosemarkie Inlier, nestling against the GGF (Fig. 1), and is similarly onlapped by the late Eifelian sandstones and conglomerates. The inlier exposes variably siliceous to micaceous psammites with subsidiary semipelites, cut by garnet amphibolite bodies: the metasedimentary lithologies have been assigned to the Moine Supergroup (Rathbone & Harris 1980). The pelitic units are garnetiferous and contain quartzofeldspathic segregations and white-mica aggregates with sparse relict fibrolite and more rarely ragged, strained kyanite blades (Rathbone & Harris 1980). Thin quartzofeldspathic veins and porphyroblasts and segregations of pink feldspar are common and there is evidence of *in situ* mobilisation of the psammites and semipelites. Several dyke-like masses of red pegmatitic granite of unknown age, some over 6 metre thick, cross cut the mafic and metasedimentary rocks.

The structures in the Cromarty Inlier correlate in part with those in the Rosemarkie Inlier. Rathbone & Harris (1980) reported that a single early planar fabric (S_1-S_2?) is folded by the dominant F_3 folds. In the southern part of the inlier F_3 axial planes are sub-vertical and first trend NE-SW but swing to E-W and then SE-NW at South Sutor stacks (Fig.2). F_3 axes similarly plunge gently NE and also swing to plunge moderately NW. In the northern part of the inlier F_3 axial traces trend SE-NW and axial planes dip NE at c. 60°. F_3 axes plunge gently to moderately both NW and SE. The folds have no consistent vergence and as in the Rosemarkie Inlier, the F_3 axes are curvilinear. F_4 folds are developed on several scales; their axes plunge consistently NE and they have steeply dipping axial planes.

Devonian Rocks

At the southwest end of the Rosemarkie Inlier is a thin, wedge-shaped and fault-bounded sliver of early Devonian Lower Old Red Sandstone (ORS) rocks that is overlapped to the west by mid-Devonian conglomerate (Fig. 2). The succession, termed the Den Siltstone Formation by Fletcher *et al*. (1996), consists of indurated breccio-conglomerates and chocolate brown to green-grey siltstones and silty sandstones. Lithologically, they are similar to Struie Group that crops out farther northwest (Trewin & Thirlwall 2002). The rocks dip moderately to very steeply westwards, but are affected by small-scale faulting with gouge commonly developed and sub-horizontal slickensides locally present. The bounding faults to this Lower ORS sequence extend into the overlying mid-Devonian succession.

The Meall Fuar-mhonaidh Outlier

The Meall Fuar-mhonaidh Outlier exposes a c. 2 kilometre-thick sequence of Lower ORS sandstones and conglomerates and minor siltstones and mudstones (Fig. 5) (Mykura & Owens 1983). The fault-bounded outlier measures 15 km long and c. 3 km wide and lies adjacent to the GGF and Loch Ness, some 32 kilometres SW of the Rosemarkie Inlier (Fig.1). Mykura & Owens (1983) collected a siltstone sample from the Drumbuie Burn by Drumnadrochit for palynological studies. This yielded fragmentary plant material and miospores indicating a late Emsian or early Eifelian age. The dominant lithology is a red-brown to purple, micaceous sandstone with thin mudstone partings and minor siltstone and conglomerate interbeds. Thick units of pink to red-brown, arkosic, gritty, coarse-gained sandstone occur in the southern part of the outlier and around Urquhart Castle (Fig. 5). Prominent lenticular units of poorly sorted and unbedded conglomerate and breccio-conglomerate, 50m to 400m thick, interdigitate with the sandstones. Moine psammites form most of the conglomerate clasts with subordinate semipelite, vein quartz, granite-gneiss, microgranite and some locally derived sandstone. Conglomerate-sandstone contacts are generally sharp and planar, but at [NN 4620 2215] a large ‘flame’ structure is developed; a 3-4m wide septum of steeply dipping silty sandstone penetrates the overlying conglomerate for some 50 metres.

At the northeast end of the outlier near the top of the succession lies the Creag Nay Conglomerate, a highly lensoid, clast-supported, breccio-conglomerate unit, bounded on its eastern side by a steep fault. It consists of large angular clasts of psammite and pink to orange leucogranite in a coarse-grained sandstone matrix. Mykura & Owens (1983) interpreted the unit as the product of a proximal debris flow derived from the east.

The Meall Fuar-mhonaidh sequence is folded into a broad syncline with minor folds and steeper dips developed in the southeast part of the outlier closer to the GGF (Fig. 4).

Minor fold axes typically plunge gently to the E and NE or to the W and SW. Cross-section restorations show that overall shortening across the outlier totals some 25%. At the southern end of the outlier Mykura & Owens (1983) recorded that the Devonian rocks have been thrust and faulted against brecciated and cataclastic Moine psammites. At NH 449 188 shattered pebbly sandstones are separated from the underlying Moine psammites by a mylonite zone that dips c. 25° SE. Mykura & Owens (1983) also postulate a thrust at the Devonian-Moine boundary immediately north of Loch a'Bhealaich to explain its sinuous nature and the steep dips in the overlying conglomerates and sandstones. The northwestern boundary of the outlier is unexposed but its overall orientation suggests that it is a steeply dipping NE-trending fault. Faults in the outlier generally trend north and NE and postdate the folding and thrusting.

The succession preserved in the Meall Fuar-mhonaidh Outlier represents a fluviatile and lacustrine sequence, rapidly deposited in late Emsian times in a restricted rift basin with marginal alluvial fans (Mykura & Owens 1983). Contractional deformation linked to movements along the GGF may have overlapped the later stages of sedimentation, and subsequently generated folding, localised thrusting, and faulting in the outlier. The structural pattern is compatible with a positive flower structure linked to sinistral transpression along the GGF.

Geochronology

Previous studies

Rathbone (1980) reported strongly discordant zircon U-Pb data defining a chord with a lower intercept at 384 Ma from a leucogranite vein below Learnie [NH 766 618]. The bulk zircon fractions were analysed by R.A. Cliff at Leeds University and consisted of old grains, neocrystalline grains, and composite grain cores and rims. The data were obtained when zircons were not processed to reduce the effects of lead loss, and hence this age likely underestimated the true age of new zircon growth. The leucogranites were interpreted as deformed by a D_2 event that was correlated directly with the deformation sequence seen in the Moine succession. Hence, as the rocks were altered and lay adjacent to the GGF the lower intercept age was discounted as due to subsequent leaching and consequent lead loss.

Sampling and analytical techniques

For this study four samples of leucogranite and two of adjacent thinly banded hornblendic felsic gneisses were collected from the shore section below Learnie Farm (Figure 3a). Zircon and monazite were recovered from the c. 3 kg samples using standard crushing, heavy liquid and isodynamic magnetic separation techniques. Zircons from two leucogranite samples were not analysed as the grains proved insufficiently robust to survive either air or chemical abrasion due to abundant cracks and probable high U contents. Sample location grid references are given in Tables 1 and 2.

Mineral grains selected for TIMS analysis and zircons for LA-MC-ICP-MS analysis were hand-picked under ethanol and only the highest quality crack-free grains were chosen. Cathodoluminescence images of the main types of zircon grains in the leucogranite and gneisses are shown in Fig. 6. Zircons selected for TIMS analysis were abraded following Krogh (1982) to reduce Pb loss. All minerals selected for TIMS analysis were washed in distilled 2N HNO_3 at c. 60° C and ultra-pure water, spiked with a $^{205}Pb/^{235}U$ tracer and dissolved in ultra-pure acids, and processed through chemistry following Krogh (1973) with modifications as described by Corfu & Noble (1992). Data were mainly obtained on a VG 354 mass spectrometer fitted with an ion-counting Daly detector, with some data obtained on a Triton mass spectrometer using an ion-counting secondary electron multiplier. Procedural blanks were <10 pg and <0.1 pg for Pb and U, respectively. Raw data were reduced using PbDat (Ludwig 1993). The common Pb isotope composition used in data reduction was estimated using the two-stage model of Stacey & Kramers (1975).

Zircons from the samples GX 1732 and 1737 were selected for analysis by LA-MC-ICP-MS and subjected to chemical abrasion (CA) following Mattinson (2005), as this has been shown to improve concordance of the ablated minerals (M. Horstwood. pers. comm. 2006). The grains were annealed at 900 °C for 60 hrs and then leached at 180 ° C in 29M

HF for 10 hrs to remove domains that could contribute to Pb loss. Zircons from sample GX1732 were not significantly affected; whereas some GX1737 grains were reduced in volume by up to c. 30% (see Fig. 6 e). The zircons were then mounted in 25 mm diameter epoxy resin discs and polished to remove c. 40% of the grain thicknesses to yield cross sections. Grains were imaged in backscatter electron (BSE) and cathodoluminescence (CL) modes using a scanning electron microscope (SEM) to examine the zircon internal zonation, thus allowing selection of appropriate areas for laser ablation analysis. Data were obtained on a Nu HR MC-ICP-MS using analytical protocols based on Horstwood *et al*. (2003). Raw data were reduced using an in-house Excel spreadsheet. TIMS and MC-ICP-MS reduced data were plotted using Isoplot (Ludwig 2003). Sample information and U-Pb data are summarized in Tables 1 and 2, and plotted on Figure 7. TIMS errors in the data tables and plotted on concordia diagrams are quoted at the 2σ level.

Results

Two distinctive zircon morphologies were recognised in the leucogranites. Stubby, internally complex-zoned, faceted to rounded, partly resorbed zircons were the most abundant type, but well-faceted, acicular zircons were also present (see Fig. 6). Titanite was the principal secondary U-bearing accessory phase in the leucogranites, except in GX 1734 where monazite was present.

The acicular zircons typically have large aspect ratios, up to 10:1, and show strong oscillatory compositional zoning, visible both under the binocular microscope and by CL (Fig. 6a). Cores occur in all but the smallest acicular grains. U contents in the acicular zircons are relatively high (c. 700 ppm, Table 1) with the highest U-zones (grey low CL areas - Fig. 6a) occurring at the grain tips. Multi-grain fractions of acicular zircons from sample GX 1731 were selected for analysis by TIMS to constrain the emplacement age of the leucogranites. Each fraction comprised abraded small grains, <50 μm long, exhibiting only simple oscillatory zoning. Given their small size and delicate elongate form, the grains were only lightly abraded, making total elimination of Pb-loss difficult. Additional analyses using CA-TIMS were not pursued as the high-U domains that best represent new zircon growth during leucogranite emplacement and crystallization would have been removed during the preliminary leaching steps.

Monazite was found only in leucogranite sample, GX 1734, where it forms sharply faceted greenish-yellow euhedral crystals and crystal fragments up to 100 μm long. Fe-oxide inclusions are common in most of the monazite grains but only inclusion-free crystals were selected for analysis. This high quality monazite is considered to be the most reliable and preferred mineral for dating the leucogranite crystallization at Rosemarkie. Although zircon is abundant in the leucogranites, its near ubiquitous inheritance is a serious impediment to achieving concordant TIMS ages. A similar problem has previously been noted in Himalayan leucogranites (e.g. Noble & Searle 1995).

One acicular zircon fraction from sample GX 1731 is reversely discordant but overlaps the concordia curve, yielding a concordia age (Ludwig 2003) of 400.8 ± 2.6 Ma (Fig. 7a). The second zircon fraction is normally discordant and slightly younger than 400 Ma. Its position on the concordia plot is consistent with Pb-loss coupled with a small amount of inherited older zircon present.

The monazite data are slightly reversely discordant, which results from excess ^{206}Pb, as is normally found in pristine monazite that does not show Pb-loss (Schärer 1984). The data spread along concordia but all four analyses overlap within error. The crystallization age is best constrained by the average monazite $^{207}Pb/^{235}U$ age of 397.6 ± 2.2 Ma based on all of the data (MSWD = 1.4). This age is consistent with the concordia age obtained from the GX 1731 acicular zircons.

The more equant, multi-faceted zircons from sample GX 1731 and zircons with similar morphology from the gneisses GX 1732 and 1737 share the same general internal compositional zoning characteristics, as revealed by CL imaging. Normal igneous oscillatory zoning is absent and most grains show broad bands of contrasting luminescence (e.g. Fig. 6b, c). Some internal zones have boundaries that suggest original crystal faces and resorption features (Fig. 6b, c) or vague primary compositional zoning (Fig. 6c). However, for the most part the textures indicate complete internal recrystallization with weak sector (e.g. Fig. 6d) or chaotic zoning (e.g. Fig. 6e) being the most pronounced features. Sample GX 1737 also has a number of intergrown and completely recrystallized zircons (Fig. 6f). These textures are characteristic of rocks known to have been metamorphosed under granulite-facies conditions (Corfu *et al.* 2003), attesting to a high-grade metamorphic history for the Rosemarkie gneisses. Similar textures have been noted in zircons from the high-metamorphic grade gneisses from the mainland and offshore Lewisian Gneiss Complex (Corfu *et al.* 1998; Whitehouse & Bridgwater 2001; Love *et al.* 2004).

LA-MC-ICP-MS data from these complex zoned zircons are listed in Table 2 and summarized in Figure 7b. Data were obtained from the cores of the zircons, and in general only a single compositional zone was sampled. In a few instances the pit (c. 25 μm diameter) did sample across several compositional zones but there was no significant difference in isotope ratio or calculated age. Three main observations can be drawn from the data: sample GX 1732 is an Archaean gneiss; sample GX 1737 contains Archaean zircons that are distinctly younger than those in GX 1732; sample GX 1737 experienced new zircon growth or complete metamorphic resetting during the Palaeoproterozoic.

Zircon analyses from sample GX 1732 plot mainly as concordant to reversely discordant, giving ages between 2932 ± 8 Ma and 2808 ± 9 Ma. The older concordant analyses are from grains with central regions preserving vestiges of oscillatory zoning within euhedral grain outlines, surrounded by the broad banding. The younger concordant analyses are from zircons with broad sector and fir-tree zonation or from grains with roughly homogeneous and low CL. This age pattern suggests that the gneiss formed either from an igneous protolith emplaced at c. 2900 Ma with subsequent metamorphism, or at c. 2800 Ma under high-grade metamorphic conditions with c. 2900 Ma inheritance. The age of granulite-facies metamorphism certainly extended to c. 2800 Ma, by which time many of the zircons had undergone significant recrystallization.

The reversely discordant GX 1732 grains have similar $^{207}Pb/^{206}Pb$ ages to the concordant grains, and correlate with ablations from low CL regions dominated by broad sector or chaotic zoning. In contrast, the normally discordant data do not correlate with a particular CL texture. These zircon data form an array consistent with the main period of Pb-loss occurring between 0 Ma and 400 Ma, and do not show evidence of Proterozoic Pb-loss (see Fig. 7b).

Sample GX 1737 has Archaean zircons that show the affects of significant Pb-loss or new zircon growth in the Palaeoproterozoic. Unlike GX 1732, none of the Archaean zircons in this rock show reverse discordance. Concordia ages range from 2781 ± 13 Ma to 2719 ± 10 Ma, indicating a younger protolith age than GX 1732. The discordant, largely Archaean grains fall in an array towards ~1750 Ma (Fig. 7b). A second Pb-loss array is outlined by concordant to moderately discordant analyses. A regression through these data (see Table 2) yields an upper intercept age of 1746 ± 31 Ma with a lower intercept anchored at the GX 1734 monazite age of 398 Ma (MSWD = 2.6). An unconstrained regression yields intercepts of 1740 ± 16 Ma and 233 +140/-150 Ma (MSWD = 2.0). The data and CL textures are consistent with the gneiss being generated from an Archaean protolith and metamorphosed under granulite- or upper amphibolite-facies conditions between c. 2780 and 2720 Ma, followed by further upper amphibolite-facies metamorphism during pervasive Laxfordian reworking at c. 1745 Ma.

Finally, the cores of a few rounded zircons from sample GX 1731 were analysed by LA-MC-ICP-MS merely to determine the nature of inheritance in the leucogranite. Both Archaean age and ca. 1700 Ma grains were observed, consistent with their derivation from the adjacent Proterozoic-Archaean gneisses of the inlier.

Implications of Dating

The zircon U-Pb data from the two felsic and mafic gneisses sampled in the Rosemarkie Inlier clearly show their Lewisianoid affinity. Protolith ages for GX 1732 range from c. 2930 Ma to 2810 Ma, with evidence of a granulite- or upper amphibolite-facies metamorphic overprint at c. 2810 Ma. GX 1737 shows younger protolith ages between 2780 Ma and 2720 Ma with evidence of Laxfordian recrystallisation at c. 1745 Ma. Friend *et al*. (2008) presented similar Archaean ages for the Borgie, Farr and Ribigill Lewisianoid inliers of north Sutherland. The variability of ages shown by the Rosemarkie samples is surprising given their proximity. It suggests the inlier contains structurally interleaved slivers that represent different parts of the Lewisianoid basement to the Moine succession. Interleaving of Moine and Lewisianoid rocks occurred prior to emplacement of the leucogranites and the subsequent 'D_2' event. The planar nature of the fabric and basement-cover contacts and lack of small- or medium-scale F_1 folding suggests that this interleaving represents part of a 'D_1' ductile shear zone (see Harris, 1978).

The U-Pb TIMS zircon and monazite ages of 401 Ma and 398 Ma respectively obtained from the two leucogranite samples date their emplacement as late Emsian. Taking an average value of 399 Ma for leucogranite vein intrusion, and given that the inlier was unconformably overlain by conglomerate and sandstones by late Eifelian times at c. 393 Ma (Marshall *et al*. 2007), exhumation of the inlier and related deformation are restricted to some 6 million years. Leucogranite deformation textures, metamorphic assemblages in the Moine pelitic rocks, and the fabrics and fold geometries now exposed are all indicative of mid-crustal levels. Hence, c. 12-15 km uplift apparently occurred in a maximum time frame of 6 Ma. However, given that this period included planation of the topography and generation of the overlying erosional surface, 4 to 5 million years would be a more realistic estimate. Implied exhumation rates for the Rosemarkie Inlier thus range from 2 mm/year to about 4 mm /year. The early Devonian Lower ORS succession is present on the NW side of the GGF but absent from its immediate SE side except offshore. However, outliers of Lower ORS rocks are present farther east in the Grampian Highlands and there is little apparent difference in topographical level of the basal unconformity across the GGF. Hence the Rosemarkie Inlier has apparently behaved as an extruded, constricted, elongate 'pip' linked to sinistral transcurrent movements on the GGF.

We now review the available structural and strain data and the regional structure to try and explain the evolution of the Rosemarkie Inlier, particularly with respect to the history of the adjacent GGF.

Structural Model

The local structure of the Learnie shore section is described above and the foliation and lineation orientations shown in Fig. 4. The foliations strike NE, orientated some 8° clockwise of the trend of the GGF, and dip moderately to steeply NW. The only measured strain values from the inlier indicate a strongly constrictional strain (Rathbone 1980) with a k value of 2.88. This obliquity of foliation, NE-plunging stretching lineation, and related D_2-D_4 prolate strain, suggest a strong component of transpression or transtension (e.g. Sanderson & Marchini, 1984; Dewey *et al*. 1998; Fossen & Tikoff, 1998). Given the proximity of the GGF and localised nature of the uplift this seems a likely circumstance. Oblate to plane strains are more generally characteristic of transpression, whereas prolate to plane strains are more typically developed in transtension (Tikoff & Teyssier 1994; Jones *et al*. 2004). Although both types of deformation can give rise to steeply dipping fabrics, they are more commonly developed in transpression. Steeply dipping lineations generally form during transpression, whereas shallow dipping, commonly horizontal lineations are typical of most transtensional situations (see Fossen & Tikoff, 1998; Krabbendam & Dewey, 1998). In the Rosemarkie Inlier the steeply dipping foliation, moderately dipping lineation (37° to 050°), and documented rapid uplift are compatible with transpression but not with transtension. The boundary conditions in effect require that the inlier is extruded, an unreasonable circumstance for a regional strain field, but plausible in small domains. It is proposed that the inlier formed at a restraining bend of the GGF due to a fault step-over to the NW that developed in mid-Devonian times. The fault geometry and the internal structure of the inlier indicate that transpressional uplift accompanied significant sinistral lateral movements on the GGF.

Theoretical transpressional deformation models and field examples on several scales have been amply documented (e.g. Robin & Cruden 1994; Lin *et al*. 1998; Jones *et al*.

2004). Although specific cases can be modelled, most authors have found it difficult to describe the transpressional deformation fully, even in specific well-documented geological examples. The variations in boundary conditions, convergence angles, vorticity, strain and fabric development, strain rates, and the common occurrence of strain partitioning, all impose limits on the accuracy of the model (Robin & Cruden 1994; Jones & Tanner 1995; Lin *et al*. 1998). In the Rosemarkie Inlier the bounding faults are either unexposed or have been reactivated subsequent to mid-Devonian deformation and uplift; brittle deformation effects and alteration affect much of the exposed section. However, the structural geometry, strong prolate strain, and poorly developed and apparently contradictory kinematic indicators seen in the Learnie shore section do constrain the possible transpressional models. Tikoff & Fossen (1999) provided 12 reference 3-D deformation models applicable to thrust and transpressional/transtensional deformation. The Rosemarkie deformation features and prolate strain fit well in the widening shear or widening/shortening shear categories, dominated by vertical or possibly oblique extrusion. Robin & Cruden (1994) described transpressional shear zones from Canada and Sweden and derived dynamic theoretical models of the stress and strain distribution in a vertical transpression zones. Again, the Rosemarkie Inlier would fit well as a sinistral transpressive zone with a strong 'Press' (i.e. vertical or steep extrusion) component. One is left to speculate as to whether the inlier was extruded with fixed bounding faults or became thinner as extrusion occurred. Oblique transpression models introduce further complications, particularly with regard to strain and vorticity variations, both across the zone and at different vertical levels (see Robin & Cruden 1994, Figure 12).

Searle *et al*. (1998) documented transpressional tectonics along the dextral Karakoram Fault Zone in Ladakh where a fault splay gives rise to a restraining bend. The resulting inlier (the Pangong Range), which consists of Cretaceous-age migmatitic and high-grade metamorphic ortho- and paragneisses (Searle *et al*. 2010), was exhumed between 18 and 11 Ma. Leucogranite bodies (Tangtse Granite) were intruded at the start of transpression and show S-C fabrics and a prominent lineation that plunges 20° to the NW. Searle *et al*. (1998) used the offset of geological features and the lineation orientation to conclude that lateral slip totalled 56 km and vertical exhumation some 20 km during this c. 7 Ma Miocene event, giving average lateral slip rates of 8.3 mm/year and vertical uplift rates of 3.0 mm/year.

If we take the strain values obtained by Rathbone (1980) as representative of the deformation during transpressional uplift of the Rosemarkie Inlier, then we can obtain a crude estimate of the amounts of uplift and lateral movement. We must assume that:

i) the strongly deformed leucogranites were emplaced at the start of deformation, as suggested by the field relationships

ii) the strain is representative of the inlier as a whole,

iii) the inlier is 2 km wide.

Approximating the inlier as a simple prolate ellipsoid and restoring it to an unstrained state (Fig 8) implies uplift of c. 15 km and sinistral displacement of c. 29 km. Despite the admitted simplicity of the assumptions (see above), these values are of a sensible order and compatible with the structural and metamorphic state at the current level of outcrop. They fit with the structural and geochronological evidence that the Rosemarkie Inlier was extruded obliquely as a rising but deforming body (elongate ‘pip’) coeval with a short-lived (4 to 6 Ma) mid-Devonian transpressional event. Lateral slip rates of 4.8 – 7.25 mm/year and vertical uplift rates of 2.3 – 3.65 mm/year are implied. These rates are of the same order as those reported from the Yammouneh and related faults in the Lebanon (Gomez *et al*. 2007; Butler *et al*. 1998), but less that those reported from major plate-bounding strike-slip fault zones such as the Alpine Fault in New Zealand (Walcott, 1998) or the Karakoram Fault (see above).

The Cromarty Inlier possibly represents a further ‘pip’ that was exhumed from somewhat shallower crustal levels, implying that the restraining bend developed on the GGF but was partitioned into blocks by linking faults, perhaps indicating that step-overs developed sequentially to the northeast as sinistral movements occurred. Similarly the fault-bounded sliver of Lower ORS rocks that lies adjacent to the Precambrian rocks of the Rosemarkie Inlier on its southwestern side (Fig. 2.) appears to represent only limited uplift during transpression.

Discussion

Tectonic Implications

Transpressional uplifts are a well-documented phenomenon linked to restraining bends in strike-slip fault zones. The Rosemarkie Inlier is a good small-scale example of a sharp restraining bend within a cratonic strike-slip fault system based on older crustal faults and removed from active plate boundaries. Mann (2007) provided an extensive overview and classification of restraining- and releasing bends related to active and ancient strike-slip fault systems. He noted that the bends act as ‘concentrators’ of intraplate stresses and the related uplifts affect relatively small rhomboidal step-over areas. Such bends are typically short-lived as they are bypassed by subsequent faulting or become extinct with continuing lateral movement. Examples of small-scale focused uplift have been reported from the San Gabriel Mountains adjacent to the San Andreas Fault (Buscher & Spotila 2007), and from the Ocotillo Badlands (8 km x 2 km step-over) along the active Coyote Fault in Southern California (Segall & Pollard, 1980). In contrast, Paul *et al*. (1999) described a c. 500 Ma example from the northern Flinders Ranges (South Australia) where localised sinistral transpressional uplift occurred during the Delamerian Orogeny. Uplift was accompanied by very high heat flows with the exhumed metamorphosed basement rock assemblages implying that temperatures of 500-550°C were attained at depths of c. 10 km.

Late Silurian – Mid-Devonian evolution of the Great Glen Fault (GGF)

As noted above the GGF has had a lengthy history of movement dominated in Silurian and Devonian times by sinistral lateral displacements. Most authors favour at least 200 kilometres of late Silurian sinistral movement based on the offset of Caledonian regional tectonic features (e.g. Dewey & Strachan, 2003) and prominent geophysical reflectors in the upper mantle lithosphere (Hall *et al*. 1984; Snyder & Flack 1990). The GGF appears to have acted as a near planar sub-vertical structure from Neoproterozoic to early Devonian times (see Stewart *et al*. 1999, 2001), but its subsequent geometry in the Moray Firth area is more complex. The Rosemarkie Inlier lacks evidence of linear or planar fabrics formed during the late Silurian sinistral movements along the GGF at c. 425 Ma. The planar fabrics that predate the leucogranite sheets and veins lie near parallel to bedding and relate to the earlier interleaving of Moine and Lewisianoid rocks. Hence, the formation of the restraining bend, related step-overs and main structures in the inlier are mid-Devonian in age. So why did the geometry of the GGF change at this time?

Watson (1984) estimated that regional uplift of the Inverness area totalled some 10-15 km during the late Silurian-early Devonian period. The bulk of uplift was completed prior to deposition of the Lower ORS lacustrine and fluvial sandstone, siltstone and conglomerate succession that commenced in the Emsian at c.407 Ma. Thick bituminous mudstone, siltstone and impure limestone units are developed both in the Beauly-Strathpeffer area (Mykura & Owens 1983), NW of the GGF, and also beneath the Mesozoic units of the Inner Moray Firth, SE of the GGF (Marshall & Hewett 2003). These lacustrine rocks form part of the Struie Formation and attain over 1000m in thickness. The Lower ORS facies distribution and sedimentology define a pattern of restricted fault-controlled basins with proximal infill marking a period of extension (Trewin & Thirlwall 2002; Marshall & Hewett 2003). Possibly the GGF was reactivated in transtension at this time (Dewey & Strachan, 2003). During end Caledonian uplift and extension new faults were formed and the structural template was changed. Hence, when sinistral transpression occurred in the mid-Devonian as a result of a northward directed compressive 'pulse', lateral movement stepped northwest. This short-lived deformation event (399-393 Ma) signalled a marked change in the applied stress direction, apparently focused on the GGF and in the adjacent early Devonian basins where localised thrusting, folding and faulting occurred.

Late Palaeozoic – Mesozoic evolution of the Great Glen Fault (GGF)

Underhill & Brodie (1993) evaluated the structural geology of Easter Ross and the Moray Firth basin and described a sequence of NNE-trending folds and faults lying NW of the GGF trace in Middle and Upper ORS rocks around Tain. They concluded that the structures, developed during Permo-Carboniferous inversion, reflected major faults in the underlying early Devonian and older rocks. Limited dextral movements occurred on the adjacent GGF. However, the main offshore extension of the GGF was inactive during widespread extension in the Permo-Triassic and Jurassic and the fault zone was only reactivated as part of a transtensional flower structure in the late Cretaceous or early

Cainozoic. Fault movements from Jurassic times onward were focused at the margins of the Moray Firth Basin (Andrews *et al*. 1990). Bird *et al*. (1987) showed that during the Mesozoic lateral movements transferred to the northwest onto the Helmsdale Fault giving rise on the Sutherland Terrace to localised extensional deformation during sinistral movements, and inversion and localised compression during dextral movements. This migration of movement resulted in formation of a series of step-overs linked in the Mesozoic sequences by moderate to gently dipping thrusts or extensional faults. Hence, the onshore and offshore record shows that strike-slip movements migrated northwestwards onto parallel faults with their timing dependent on the regional plate tectonic geometry and the prevailing stress orientations (Underhill & Brodie, 1993). The northeast extension of the GGF into the Moray Firth Basin was locked for much of the Late Palaeozoic and Mesozoic. Underhill and Brodie (1993) concluded that because the GGF was a vertical structure and lay orthogonal to the northwest-southeast extensional strain field, strike-slip reactivation only occurred when extensional slip vectors changed to become near-parallel to the fault., i.e. in the Permo-Carboniferous and Cainozoic. Thus, the mid-Devonian transpressional event marked a major change in the pattern of movement on the GGF, with the locus of fault movement migrating progressively northwest; this pattern continued into Mesozoic times with lateral and vertical movements becoming focused on the Helmsdale Fault (Roberts & Holdsworth 1999).

The generation of the Rosemarkie Inlier

It is proposed that the Rosemarkie Inlier was generated by oblique extrusion at a sharp restraining bend developed on the GGF between about 399 Ma and 393 Ma. The formation of the step-over reflected the regional stress conditions that prevailed during this sinistral transpressional event, the increased frictional resistance to renewed movement along the pre-existing GGF trace, and the newly formed structural template. This short-lived mid-Devonian transpressional event marked the end of late Caledonian uplift, extension and sinistral transtension in the late Silurian and early Devonian (see Dewey and Strachan 2003). The event was coeval with the Acadian compressional event that is widely developed in England and Wales (Woodcock & Soper 2006; Woodcock *et al*. 2007) and even recorded in western Ireland (Meere & Mulchrone 2006). In contrast, in West Norway and possibly also in Shetland there is evidence of widespread extension, transtension, and strike-slip fault movements during this period, which was dominated by the uplift and erosion of the emerging Caledonide chain (Krabbendam & Dewey 1998; Walsh *et al*. 2007; Fossen, 2010). Although the Acadian event resulted from the onset of collisional activity in the developing Rheic Ocean south of Avalonia we argue below that it also affected Scotland giving rise to localised sinistral transpression focused along the main terrane-bounding faults and extending as far north as the Moray Firth.

The nature of the Acadian Event in Scotland

The Devonian succession in Scotland contains evidence of periods of uplift and possible tectonic activity that separate it into three distinct sequences. These were termed the Lower, Middle and Upper Old Red Sandstone (ORS) by Murchison (1859) and the terms are still in use today, albeit with considerably modifications (see Trewin & Thirlwall 2002).

Midland Valley

The Midland Valley terrane is separated from the Highlands on its northwest side by the Highland Boundary Fault Zone (HBFZ) and from the Southern Uplands on its southeast side by the Southern Uplands Fault (SUF) (Fig. 8). Within this terrane, fluvial and lacustrine Lower ORS rocks of late Silurian to early Devonian age are widely developed (Bluck 2000). In Strathmore the sequence consists mainly of fluvial sandstones and conglomerates with volcanic rocks in its lower parts. These include the distinctive dacitic Lintrathen Tuff (Porphyry), dated at 415 ± 6 Ma (Thirlwall 1988). In its uppermost parts thick conglomerate units are developed locally adjacent to the HBFZ. A prominent example is the c.1500 m thick Strathfinella Hill Conglomerate near Fettercairn that represents a proximal alluvial fan deposit (Haughton & Bluck 1988). Unlike most conglomerate units in the Lower ORS sequence that consist largely of reworked quartzite cobbles derived from the northeast, this unit contains first cycle metamorphic and volcanic clasts, derived from the northwest. It is dominated by Dalradian psammite and semipelite clasts that can be matched readily in the nearby

Grampian Highlands and clearly records syn-depositional uplift of the adjacent Highlands. The conglomerate passes rapidly to the southeast into siltstones and mudstones which have been dated as Emsian from miospores (Richardson *et al*. 1984). The Lower ORS sequence is folded into the Strathmore Syncline and Sidlaw Anticline (Fig. 8) and overlapped unconformably by the late Devonian Upper ORS succession. Hence fault movement, deformation, uplift, and erosion are bracketed as mid-Devonian in age. The Lintrathen Tuff crops out on both sides of the HBFZ, but shows an apparent sinistral offset of some 34 kilometres. The structural features and timing of deformation are consistent with an Acadian sinistral transpressional event focused along the HBFZ during the mid-Devonian (see Jones *et al*. 1997 for kinematic analysis; Tanner 2008). Deposition of the youngest Emsian parts of the sequence appears to have overlapped with fault movements along the HBFZ.

Lower ORS sandstones, conglomerates, and andesitic and basaltic volcanic rocks also crop out near the southeast margin of the Midland Valley (Smith 1995) (Fig. 8). Again the volcanic rocks yield a Lochkovian age, here c. 412 Ma (Thirlwall, 1988). The rocks were deformed during a mid-Devonian tectonic event whose effects become more intense towards the SUF. Deformation resulted in the formation of kilometre-scale, asymmetrical anticlines and synclines whose axes trend NE to ENE, oblique to the SUF. They form *en échelon* arrays and Smith (1995) interpreted the fold pattern as indicative of sinistral transpression focused along the SUF. Floyd (1994) presented evidence for some 12 km of sinistral offset of structures just north of the Loch Doon granite pluton.

Meall Fuar-mhonaidh Outlier

Lower ORS sandstones and conglomerates are preserved in the Meall Fuar-mhonaidh outlier adjacent to the GGF some 32 km southwest of Rosemarkie (see above for details of lithology and structure) (Fig. 5). The c. 2km thick sequence was deposited rapidly in a fault-bounded basin (Mykura & Owens 1983). At its northeast end the Craig Nay Conglomerate contains large angular clasts of psammite and pink-orange leucogranite veins that match those exposed in the Rosemarkie Inlier. Given its highly proximal nature and high stratigraphical position, it is proposed that this conglomerate unit formed by erosion of the inlier, which at the time was situated immediately to the northeast. This would date the start of exhumation of the inlier and sinistral movement on the GGF as late Emsian in accord with the age of leucogranite emplacement. Note that the eroded material would be derived from a much higher crustal level than that presently exposed. The fold pattern, limited WNW-directed thrusting and decreasing strain away from the GGF in the outlier are all compatible with the development of a positive flower structure linked to sinistral transpression. Although this deformation cannot be dated with certainty here its low grade and structural pattern are best explained as due to the Acadian event.

Rosemarkie and Easter Ross

Deformation also affects the Lower ORS sequence northwest of the GGF on the Black Isle and in Easter Ross. Rogers *et al* (1989) placed the sequence in the late Emsian and noted that its northwestern bounding faults, the Torr Achilty and Glaick-Polinturk faults, show evidence of limited compressional and strike-slip reactivation. Localised thrusting occurs at the base of the succession, e.g. at Contin, and farther north on Struie Hill, where the gently ESE-dipping Struie Thrust forms a prominent feature marked by low grade mylonites (Armstrong, 1964). The thrust lies some 25 kilometres northwest of the GGF trace, and 20 km from the Cromarty Inlier (Fig.8). Underhill and Brodie (1993) deduced that the thrust formed as a consequence of footwall collapse linked to inversion along the Polinturk Fault. They interpreted the resultant flower structure as Permo-Carboniferous, linked to Variscan deformation in the Moray Firth, but its geometry is also compatible with Acadian transpression. In Easter Ross the Middle ORS sandstone and conglomerate sequence (with fish beds) overlies the Lower ORS with slight to moderate angular unconformity and is in turn overlain by Upper ORS sandstones. The whole sequence is folded by the large open Black Isle Syncline, a probable Permo-Carboniferous age structure.

The angular unconformable Middle ORS – Lower ORS boundary can be traced northeastwards into the Golspie and Badbea basins, but at Sarclet (by Wick) in Caithness the two successions are conformable (Trewin & Thirlwall, 2002), possibly documenting the northward waning of Acadian tectonic effects.

Conclusions

The Rosemarkie Inlier consists of Moine psammites and semipelites and Lewisianoid felsic and mafic gneisses, all intruded by abundant pink leucogranite veins. Zircon U-Pb LA-MC-ICP-MS data from two gneiss samples give Archaean protolith ages between 2930 and 2720 Ma; zircon morphologies are consistent with their formation in high grade gneisses at deep crustal levels. One sample contains evidence of significant zircon growth at c. 1745 Ma, indicative of Laxfordian reworking. Hence, the inlier exposes structurally interleaved Moine and Lewisianoid rocks, effectively providing a 'snapshot' of the deeper levels of the Caledonian orogen in this area. The interleaving and related planar fabrics predate leucogranite emplacement and may be late Silurian (Scandian), early Ordovician (Grampian) or Neoproterozoic (Knoydartian) in age. Similar relationships are found at the Sgurr Beag Thrust some 28 km to the WNW (Grant & Harris 2000).

Monazite and zircon U-Pb TIMS data from the leucogranite veins show that they were emplaced into the Moine and Lewisianoid rocks at 399 Ma. The veins are discordant to the early fabrics (S_1) but are strongly deformed and folded in three structural phases (D_2-D_4). Metamorphic assemblages and the structural style of the main D_2 deformation are compatible with their formation at depths of 12-15 km. The inlier is overlain unconformably by mid-Devonian (Eifelian – Givetian) sandstones and conglomerates whose deposition commenced at c. 393 Ma; thus deformation and exhumation are restricted to a maximum time frame of 6 Ma, implying local uplift rates of 2-4 mm/year.

The structure of the Rosemarkie Inlier is dominated by a generally steep NE-trending foliation (S_2), moderately NE-plunging lineation (L_2/L_3) and strongly constrictional strains (Rathbone 1980). These features are compatible with its extrusion as an elongate 'pip' at a sharp restraining bend of the Great Glen Fault (GGF) during sinistral transpression. The Rosemarkie and adjacent Cromarty inliers represent fault-bounded step-overs, formed as the locus of sinistral lateral movement on the GGF migrated onto sub-parallel faults farther to the northwest.

It is suggested that in early Devonian (Emsian) the Rosemarkie Inlier lay adjacent to the Meall Fuar-mhonaidh Outlier, now situated some 32 km to its SW. The Lower ORS sequence in the outlier contains a highly proximal conglomerate unit at its northeast end whose clasts match the lithologies of the Rosemarkie Inlier. It is proposed that the final stages of Lower ORS sedimentation in the outlier overlapped with the initial exhumation of the inlier and thus the onset of significant lateral fault movement at c. 399 Ma. The 32 km offset is broadly compatible with the strain values obtained from the deformed leucogranites and the structural geometry in the inlier.

The mid-Devonian sinistral transpressional event identified at Rosemarkie is interpreted as a manifestation of the Acadian Event, a short-lived northward-directed compressional pulse generated between 400 and 390 Ma by the collision of an Armorican microcontinent with Avalonia (Woodcock *et al*. 2007). In Scotland this pulse was focused on the main terrane-bounding fault zones, namely the Southern Upland, Highland Boundary and Great Glen fault zones. It was generally partitioned into sinistral strike-slip movements on the faults and related orthogonal compressional deformation. Intensity of deformation is greatest adjacent to the fault zones and decreases with distance away from them. Deformation was preferentially taken up by the Lower ORS sequences that had accumulated in nearby fault-bounded extensional basins. Adjacent to the HBFZ there seems to have again been an overlap of fault movement and the later phases of ORS sedimentation. Positive flower structures were formed on the northwest side of the GGF in the Meall Fuar-mhonaidh Outlier and in the Lower ORS succession in Easter Ross, e.g. the Struie Thrust. The Lower ORS – Middle ORS unconformity can be traced northwards as far as Caithness, possibly reflecting the waning effects of the Acadian Event.

The mid-Devonian (Acadian) sinistral transpression marked a significant change in the kinematics of the Great Glen Fault. Prior to this event in the late Silurian and early Devonian the fault appears to have been a planar structure and a focus for sinistral lateral movements, firstly in transpression (Stewart *et al*. 1999), but mainly in transtension

(Dewey & Strachan 2003). The end Caledonian uplift and formation of small scale basins in the early Devonian altered the structural geometry, particularly in the Moray Firth area. Hence, when the far-field Acadian effects reached northern Scotland, the GGF formed a restraining bend to facilitate the migration of lateral movement northwestwards. This pattern of fault migration was repeated in late Palaeozoic and Mesozoic times during transtensional and transpressional events, both sinistral and dextral.

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Figure Captions

Fig. 1. Generalised geology of the area around Inverness showing the location of the Rosemarkie and Cromarty inliers.

Fig. 2. Geology of the Rosemarkie Inlier

Fig. 3 **(a)** Learnie foreshore showing the steeply dipping Moine psammites and Lewisianoid gneisses intruded by a prominent, near concordant, pink leucocratic microgranite vein. The hammer (38 cm long) marks the vein. Locality for GX 1731, 1732. [NH 7620 6124]. **(b)** Interference fold ($F_2 + F_3$) of thin leucogranite vein in Moine psammites and semipelites. [NH 767 617].

Fig. 4. Structural data from the Rosemarkie Inlier below Learnie Farm (lower hemisphere projection).

Fig. 5. Geological map of the Meall Fuar-mhonaidh Outlier showing its stratigraphy and structure (modified after Mykura & Owens 1983).

Fig. 6. Cathodoluminescence images. **(a)** acicular zircon (leucogranite - GX 1731) showing oscillatory zoning and high U (low CL) tip, **(b)** xenocrystic zircon (leucogranite - GX 1731) showing remnants of a now recrystallized primary zircon surrounded by a metamorphic zircon rim, **(c)** and **(d)** complex zoned zircons (gneiss - GX 1732) showing a possible primary zircon outline and embayed surface (c) and completely recrystallized internal structure (d), **(e)** and **(f)** chemically abraded complex zoned zircons; the conjoined grains both give ages of c. 1740 Ma (gneiss - GX 1737).

Fig. 7. Concordia diagrams showing: **(a)** ID-TIMS data for samples GX 1731 and 1734; light grey ellipses are GX 1731 zircons, medium grey ellipses are GX 1734 monazites. **(b)** LA-MC-ICP-MS data for samples GX 1731, 1732 and 1737. Black ellipses are GX 1731, medium grey ellipses are GX 1732, and white ellipses are GX 1737. Reference lines are 2781 – 1740 Ma and 1740 – 398 Ma.

Fig. 8. Strain modelling, see text for details.

Fig. 9. Acadian features in the British Isles. GGF – Great Glen Fault, HBF – Highland Boundary Fault, SUF – Southern Upland Fault. Ro – Rosemarkie Inlier, MF – Meall fuar-mhonaidh Outlier, SHC – Strathfinella Hill Conglomerate.

Table 1. *Zircon and monazite U-Pb ID-TIMS data for samples GX 1731, 1732*

	Concentrations †				Atomic ratios								Ages (Ma)			
Fractions *	Weight (µg)	U (ppm)	Pb (ppm)	Com -mon Pb(pg)	^{206}Pb/ ^{204}Pb §	^{208}Pb/ ^{206}Pb #	^{206}Pb/ ^{238}U #	err	^{207}Pb/ ^{235}U #	err	^{207}Pb/ ^{206}Pb #	err	^{206}Pb- ^{238}U	^{207}Pb- ^{235}U	^{207}Pb- ^{206}Pb	ρ**
GX 1731 Leucogranite [NH 7620 6124]																
1. zr, cl-pbr, 5:1, abr, 50-70 µm (13)	4.4	647.0	46.67	11.2	1042	0.1844	0.0642403	0.67	0.4829	0.80	0.05452	0.35	401.4	400.1	392.5	0.90
2. zr, pbr, 4:1, abr, 80-100 µm (2)	7.7	724.0	49.48	16.4	1351	0.1636	0.06315	0.34	0.4827	0.53	0.05544	0.39	394.8	399.9	429.9	0.67
GX 1734 Leucogranite [NH 7641 6144]																
3. mo, 2:1 eu, 80 µm (1)	1.0	1325	1286	5.3	1026	16.24	0.06417	0.52	0.4778	0.56	0.05401	0.21	400.9	396.6	371.4	0.93
4. mo, 2:1, eu, 90 µm (1)	1.3	935.0	977.0	7.4	683.3	17.5	0.06434	0.56	0.4783	0.66	0.05392	0.33	401.9	396.9	367.7	0.87
5. mo, 2:1, sub, 80 µm (1)	1.0	870.0	690.0	5.9	623.6	12.86	0.06492	0.77	0.4835	0.82	0.05401	0.29	405.5	400.5	371.5	0.93
6. mo, 1:1 eu, 70 µm (1)	1.2	850.0	1157	4.7	903.9	23.23	0.06410	0.35	0.4795	0.48	0.05425	0.33	400.5	397.7	381.3	0.74

* mo = monazite, zr = zircon; l:w aspect ratio; abr = air abraded; eu = euhedral, sub = subhedral; cl = colourless, pbr = pale brown; length (µm); (x) = number of grains analyzed.

† Maximum errors are ± 20%. Weights were measured on a Cahn C32 microbalance or calculated from grain dimensions measured on binocular microscope photos.

§ Measured ratio corrected for mass fractionation and common Pb in the ^{205}Pb/^{235}U spike.

Corrected for mass fractionation, spike, laboratory blank Pb and U, and initial common Pb (Stacey and Kramers 1975; calculated at 400 Ma). The laboratory blank Pb composition is ^{206}Pb/^{204}Pb:^{207}Pb/^{204}Pb:^{208}Pb/^{204}Pb = 17.46: 15.55: 37.32. Quoted errors are 2 σ (% for atomic ratios, absolute for ages).

** ^{207}Pb/^{235}U - ^{206}Pb/^{238}U error correlation coefficient calculated following Ludwig (1993).

Table 2. *Zircon U-Pb LA-MC-ICP-MS data for samples GX1731, 1732, 1737*

	Concentrations [†]		Atomic ratios [#]							Ages[‡] (Ma)							
Analysis	U (ppm)	Pb (ppm)	$^{206}Pb/^{238}U$	err	$^{207}Pb/^{235}U$	err	$^{207}Pb/^{206}Pb$	err	ρ**	$^{206}Pb/^{238}U$	err	$^{207}Pb/^{235}U$	err	$^{207}Pb/^{206}Pb$	err	Conc. age	Err (MSWD)
GX 1731 Leucogranite [NH 7620 6124]																	
1731-1	664	179	0.3030	0.8	4.425	0.8	0.1059	0.1	0.99	1717	24	1706	13	1730	3		
1731-2	365	108	0.3234	0.7	4.747	0.7	0.1064	0.1	0.98	1776	21	1807	12	1739	5		
1731-3	295	127	0.4725	0.8	11.46	0.9	0.1759	0.4	0.91	2562	33	2495	16	2615	12		
1731-4	267	108	0.4552	0.9	10.91	1.0	0.1738	0.4	0.92	2515	35	2418	18	2595	12		
1731-5	150	63.4	0.4490	0.9	11.02	0.9	0.1780	0.2	0.99	2525	35	2391	17	2635	5		
1731-6	80.8	33.4	0.4486	1.4	10.99	1.4	0.1777	0.3	0.98	2523	55	2389	26	2632	9		
GX 1732 Hornblende-biotite felsic gneiss [NH 7620 6124]																	
1732-1	17.8	9.6	0.5603	0.9	15.68	1.1	0.2030	0.7	0.81	2858	43	2868	21	2851	22	2854	20 (0.5)
1732-2	31.7	16.7	0.5610	0.9	15.90	1.0	0.2056	0.4	0.90	2871	40	2871	18	2871	14	2871	14 (<0.1)
1732-3	18.6	7.8	0.4390	0.8	12.05	1.1	0.1991	0.8	0.73	2608	31	2346	20	2819	25		
1732-4	61.5	31.8	0.5498	0.9	15.63	1.0	0.2062	0.2	0.97	2855	42	2824	18	2876	7		
1732-5	51.6	26.7	0.5536	0.9	15.64	1.0	0.2049	0.3	0.96	2855	42	2840	18	2866	9	2864	11 (0.5)
1732-6	13.2	6.3	0.5161	1.0	14.51	1.3	0.2039	0.9	0.74	2784	43	2683	25	2858	29		
1732-7	9.3	4.5	0.5225	1.0	13.92	1.6	0.1932	1.2	0.63	2744	45	2710	30	2770	41		
1732-8	37.7	18.7	0.5413	0.8	15.02	0.9	0.2013	0.4	0.91	2817	36	2789	17	2836	12		
1732-9	28.9	15.1	0.5533	0.8	15.46	1.0	0.2026	0.5	0.87	2844	38	2839	18	2847	15	2837	16 (1.2)
1732-10	26.8	14.3	0.5701	0.9	16.70	1.0	0.2125	0.4	0.89	2918	40	2908	18	2925	14	2923	15 (0.6)
1732-11	22.4	11.8	0.5487	0.8	15.28	0.9	0.2019	0.5	0.83	2833	35	2820	18	2842	17	2831	13 (5.9)
1732-12	61.1	32.5	0.5615	0.7	16.02	0.7	0.2070	0.2	0.95	2878	33	2873	14	2882	8	2881	10 (0.3)
1732-13	16.5	7.9	0.5087	0.8	14.09	1.1	0.2008	0.7	0.75	2756	36	2651	21	2833	24		
1732-14	76.4	40.5	0.5492	0.9	15.92	0.9	0.2102	0.2	0.98	2872	40	2822	17	2907	6		
1732-15	29.0	14.6	0.5329	0.8	14.78	0.9	0.2012	0.4	0.88	2801	37	2753	18	2836	14		
1732-16	45.6	24.1	0.5666	0.8	15.78	0.9	0.2020	0.3	0.94	2864	37	2894	16	2842	9		
1732-17	13.5	7.2	0.5754	0.9	15.78	1.3	0.1989	0.8	0.76	2864	44	2930	24	2817	27		
1732-18	125	64.1	0.5596	0.8	15.80	0.8	0.2047	0.2	0.98	2865	38	2865	16	2864	5	2864	8.3 (<0.1)
1732-19	75.6	35.3	0.5166	1.5	14.06	1.5	0.1974	0.2	0.99	2754	64	2685	28	2805	7		
1732-20	43.7	20.9	0.5251	1.1	14.13	1.2	0.1951	0.4	0.95	2758	50	2721	22	2786	12		
1732-21	72.5	33.5	0.4974	0.9	13.46	0.9	0.1963	0.2	0.96	2713	37	2603	17	2796	8		
1732-22	103	54.6	0.5789	1.0	17.04	1.0	0.2135	0.2	0.99	2937	48	2944	20	2932	5	2933	7.1 (0.9)
1732-23	109	52.4	0.5415	1.0	14.78	1.0	0.1980	0.2	0.99	2801	47	2790	20	2809	6	2809	8.7 (0.7)
1732-24	55.2	29.3	0.5590	1.1	15.85	1.2	0.2056	0.2	0.98	2868	53	2863	22	2871	8	2871	12 (0.1)
1732-25	40.3	20.3	0.5507	0.9	15.27	1.0	0.2012	0.3	0.94	2832	42	2828	19	2836	11		
1732-26	30.1	16.4	0.5863	0.8	16.75	0.9	0.2073	0.4	0.89	2921	38	2974	17	2884	13		
1732-27	86.3	43.8	0.5634	1.1	15.34	1.1	0.1975	0.3	0.96	2837	49	2881	21	2806	10		
1732-28	31.1	16.8	0.5923	0.8	16.87	0.9	0.2066	0.4	0.90	2927	39	2999	17	2879	13		
1732-29	20.5	10.7	0.5701	0.9	15.85	1.1	0.2016	0.6	0.85	2868	42	2908	20	2839	18		
1732-30	42.2	21.3	0.5521	0.8	15.50	0.8	0.2036	0.3	0.92	2846	34	2834	15	2855	10	2852	12 (1.3)
1732-31	95.3	40.4	0.4585	1.0	12.45	1.0	0.1969	0.2	0.98	2639	39	2433	18	2801	7		
1732-32	22.1	9.0	0.4440	1.3	11.89	1.5	0.1942	0.7	0.88	2596	52	2369	27	2778	23		
1732-33	103	50.9	0.5400	1.1	14.84	1.2	0.1993	0.2	0.99	2805	51	2783	22	2820	6	2819	8.8 (0.16)
1732-34	130	63.2	0.5093	1.1	13.86	1.1	0.1974	0.2	0.99	2740	48	2654	21	2805	5		
GX 1737 Hornblende-biotite felsic gneiss [NH 7649 6155]																	
1737-1	83.5	33.9	0.4345	1.2	10.71	1.3	0.1788	0.5	0.92	2499	47	2326	24	2642	17		

1737-2	85.2	31.3	0.3998	1.2	8.007	1.4	0.1453	0.7	0.89	2232	46	2168	25	2291	22		
1737-3	272	104	0.4183	1.1	9.533	1.1	0.1653	0.2	0.98	2391	42	2253	20	2511	8		
1737-4	349	159	0.4947	1.2	12.81	1.2	0.1878	0.1	0.99	2666	50	2591	22	2723	5		
1737-5	168	43.9	0.2882	1.2	4.153	1.3	0.1045	0.6	0.87	1665	33	1632	22	1706	24		
1737-6	353	102	0.3183	1.1	5.275	1.1	0.1202	0.4	0.94	1865	34	1782	19	1959	13		
1737-7	478	214	0.4893	1.1	12.00	1.1	0.1779	0.1	0.99	2605	47	2568	21	2634	4		
1737-8	362	163	0.5029	1.1	13.09	1.1	0.1887	0.2	0.99	2686	47	2626	21	2731	6		
1737-9	196	73.0	0.4088	1.1	9.138	1.1	0.1621	0.3	0.95	2352	40	2209	20	2478	12		
1737-10	264	93.2	0.3907	1.2	9.154	1.3	0.1699	0.4	0.95	2354	43	2126	23	2557	13		
1737-11	84.7	33.9	0.4339	1.2	10.67	1.3	0.1784	0.6	0.89	2495	46	2323	24	2638	20		
1737-12	147	50.7	0.3809	1.8	8.014	1.9	0.1526	0.5	0.97	2233	64	2080	33	2375	16		
1737-15	315	140	0.4943	1.2	12.46	1.3	0.1828	0.3	0.97	2640	52	2589	24	2678	11		
1737-16	119	48.3	0.4479	1.2	10.40	1.3	0.1684	0.4	0.95	2471	48	2386	23	2542	13		
1737-17	127	51.1	0.4328	1.2	10.01	1.3	0.1677	0.5	0.93	2435	47	2318	24	2534	16		
1737-18	223	104	0.5042	1.1	13.45	1.1	0.1935	0.2	0.98	2712	47	2632	21	2772	7		
1737-19	216	82.2	0.4161	1.1	9.086	1.2	0.1584	0.4	0.95	2347	42	2243	21	2438	12		
1737-20	118	35.8	0.3344	1.2	6.711	1.3	0.1455	0.6	0.89	2074	38	1860	23	2294	21		
1737-21	234	91.1	0.4437	1.3	10.92	1.3	0.1785	0.3	0.97	2516	50	2367	24	2639	10		
1737-22	125	47.1	0.4299	1.3	10.24	1.5	0.1728	0.7	0.87	2457	50	2305	27	2585	24		
1737-23	302	84.1	0.3072	1.3	4.604	1.4	0.1087	0.5	0.93	1750	38	1727	22	1778	19		
1737-24	63.7	31.0	0.5304	1.2	13.92	1.3	0.1903	0.5	0.92	2744	54	2743	25	2745	17	2745	17 (<0.1)
1737-25	108	47.1	0.4741	1.2	11.32	1.3	0.1732	0.4	0.94	2550	49	2502	23	2588	14		
1737-26	99.8	49.0	0.5397	1.1	14.48	1.2	0.1945	0.4	0.95	2781	51	2782	22	2781	12	2781	13 (<0.1)
1737-27	152	62.3	0.4467	1.3	11.18	1.3	0.1815	0.3	0.97	2538	50	2380	24	2667	11		
1737-28	199	95.9	0.5256	1.3	13.58	1.3	0.1874	0.2	0.98	2721	57	2723	25	2719	8	2719	10 (<0.1)
1737-29	449	218	0.5163	1.3	13.81	1.3	0.1939	0.1	0.99	2736	56	2684	24	2776	4		
1737-30	255	126	0.5309	1.1	13.84	1.1	0.1892	0.2	0.99	2739	50	2745	21	2735	6	2735	9 (0.16)
1737-31	97.6	41.6	0.4535	1.2	11.29	1.3	0.1806	0.4	0.94	2547	50	2411	24	2658	15		
1737-32	109	46.7	0.4621	1.1	11.50	1.2	0.1804	0.4	0.94	2564	45	2449	22	2657	14		
1737-33	281	101	0.3832	1.2	8.997	1.3	0.1703	0.3	0.98	2338	44	2091	23	2561	9		
1737-34	168	72.8	0.4636	1.2	11.43	1.3	0.1788	0.4	0.95	2559	51	2456	24	2641	14		
1737-35	44.5	13.8	0.3296	1.2	5.123	2.0	0.1127	1.6	0.58	1840	37	1837	33	1844	59		
1737-36	107	49.0	0.4852	1.2	12.37	1.3	0.1849	0.4	0.95	2633	51	2550	24	2697	14		
1737-37	216	38.2	0.1901	1.2	2.710	1.9	0.1034	1.5	0.62	1331	25	1122	28	1686	56		
1737-38	110	33.3	0.3244	1.3	4.839	1.5	0.1082	0.8	0.86	1792	41	1811	25	1769	28		
1737-39	101	30.2	0.3185	1.1	4.731	1.4	0.1077	0.8	0.80	1773	35	1782	23	1761	31		
1737-40	68.0	32.3	0.5075	1.2	12.91	1.3	0.1846	0.5	0.91	2673	50	2646	24	2694	17		
1737-41	201	93.9	0.4990	1.2	12.49	1.2	0.1815	0.2	0.98	2642	51	2610	22	2666	8		
1737-42	118	33.0	0.3060	1.2	4.467	1.5	0.1059	0.8	0.84	1725	37	1721	24	1730	29		

† Errors are c. ± 10%. # Measured ratios not corrected for common Pb. ** $^{207}Pb/^{235}U$ - $^{206}Pb/^{238}U$ error correlation coefficient calculated following Ludwig (1993).

‡ Age errors quoted at the 1σ level. Concordia ages and goodness of fit expressed as MSWD were calculated using Ludwig 2003.