Journal of the Geological Society, *London*, Vol. **158**, 2001, pp. 7–14. Printed in Great Britain.

U–Pb geochronology of the Fort Augustus granite gneiss: constraints on the timing of Neoproterozoic and Palaeozoic tectonothermal events in the NW Highlands of Scotland

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Abstract: The West Highland granite gneiss suite in Inverness-shire, Scotland, represents a series of S-type, anatectic granites formed by partial melting of host Neoproterozoic metasediments of the Moine Supergroup. U–Pb (SHRIMP) dating of zircons from a member of the suite, the Fort Augustus granite gneiss, indicates that the granitic protolith to the gneiss was intruded at 870 ± 30 Ma. This is indistinguishable from the published age determined by the same method for the Ardgour granite gneiss at Glenfinnan, thus supporting the assumption that the various members of the West Highland granite gneiss are part of a single intrusive suite. The spread of ages from the zircon cores (1626–947 Ma) is interpreted to indicate a Proterozoic source terrain for the Moine sediments that were later melted to form the granitic protolith. A U–Pb age of 470 ± 2 Ma obtained for titanite in the Fort Augustus granite gneiss is interpreted to date amphibolite-facies metamorphism during the early to mid-Ordovician Grampian Orogeny. The emerging similarity in the timing of this event either side of the Great Glen Fault implies that this structure does not juxtapose crustal blocks with significantly different histories with respect to the Grampian Orogeny.

Keywords: U–Pb, zircon, titanite, Moine, granite gneiss.

The evolution of the Neoproterozoic Moine Supergroup within the Lower Palaeozoic Caledonian orogenic belt of NW Scotland has long been a matter of considerable controversy (e.g. Peach & Horne 1930). The major issues currently concern the presence, or otherwise, of Neoproterozoic orogenic events and the extent to which the Moine rocks are allochthonous with respect to the Laurentian foreland to the Caledonide belt (e.g. Soper 1994; Soper & England 1995; Bluck *et al*. 1997; Friend *et al*. 1997; Soper & Harris 1997; Rogers *et al*. 1998; Vance *et al*. 1998). Part of the controversy has lain in the problems inherent in dating polymetamorphic terrains whereby later events may have disturbed the geochronological record of earlier events. It is well established that the Moine Supergroup was regionally deformed and metamorphosed during the Ordovician–Silurian Caledonian orogeny (Powell & Phillips 1985 and references therein). Recent geochronological studies (e.g. Noble *et al*. 1996, 1997; Friend *et al*. 1997; Rogers *et al*. 1998; Vance *et al*. 1998; Highton *et al*. 1999; Millar 1999) have confirmed the broad conclusions of earlier work (e.g. Giletti *et al*. 1961; Brook *et al*. 1976, 1977; Brewer *et al*. 1979; Piasecki & van Breemen 1979, 1983; Powell *et al*. 1983) that the Moine Supergroup and related rocks south of the Great Glen Fault (the Central Highland Migmatites) were also affected by Neoproterozoic tectonothermal activity between *c*. 870 and *c*. 780 Ma. The results of integrated geochronological and metamorphic studies suggest that at least the latest stage of this tectonothermal activity (*c*. 820–790 Ma) was related to crustal thickening (Rogers *et al*. 1998; Vance *et al*. 1998; Phillips *et al*. 1999) rather than extension, as advocated by Soper (1994) and Soper & Harris (1997).

Despite increasing evidence that the Moine Supergroup records polyorogenic activity, there is considerable uncertainty over the actual age(s) of the various structures and related metamorphic assemblages in large parts of the outcrop. Largely because of the difficulty in mapping and correlating structures in strongly deformed, amphibolite-facies psammitic rocks, the age(s) and extent of Caledonian deformation fabrics are relatively poorly documented in the Moine Supergroup, as compared with the lithologically more diverse and more comprehensively investigated sectors elsewhere in the Appalachian–Caledonian orogen (e.g. Dunning *et al*. 1990; Cawood *et al*. 1994; Barr *et al*. 1995; Northrup 1997). More precise knowledge of the timing of Caledonian events is necessary in order to evaluate published structural correlations both within the Moine (e.g. Powell 1974) and, on a wider scale, across the entire Scottish Highlands. This will provide the database necessary to assess the significance of structures such as the Great Glen Fault which might represent a terrane boundary within the Scottish Caledonides (Harris 1995 and references therein).

In this paper we report results of the U–Pb dating of zircon and titanite from the Fort Augustus granite gneiss (Fig. 1). The zircon data indicate that the granitic protolith of the Fort Augustus granite gneiss is similar in age to that of the Ardgour granite gneiss (Friend *et al*. 1997), thus supporting the assumption that the two bodies are part of the same intrusive suite. The titanite data record an age for Lower Palaeozoic

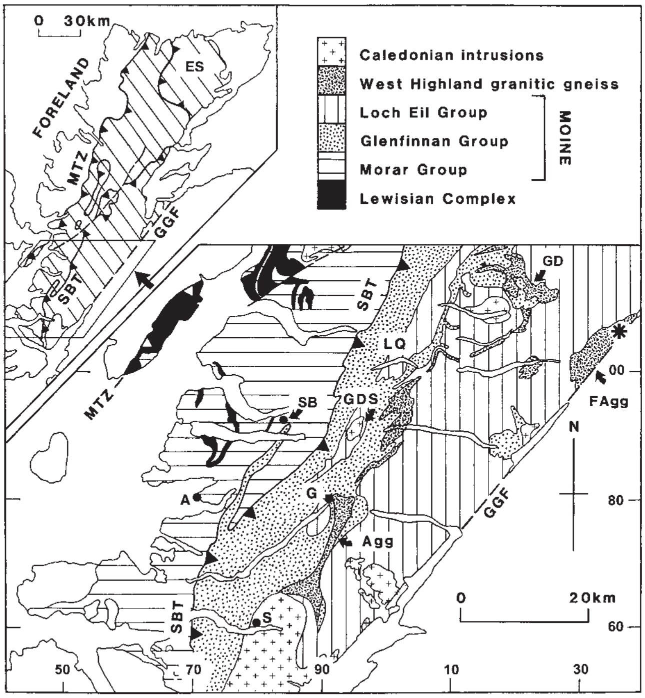

Fig. 1. Sketch map of the southern outcrop of the Moine Supergroup (see inset for location) showing the position of the Fort Augustus granite gneiss (Fagg) in relation to the Ardgour granite gneiss (Agg) and the rest of the West Highland granitic gneiss suite. Other abbreviations: A, Ardnish; G, Glenfinnan; GD, Glen Doe; GDS, Glen Dessary syenite; GGF, Great Glen Fault; LQ, Loch Quoich; MTZ, Moine Thrust Zone; S, Strontian; SB, Sgurr Breac; SBT, Sgurr Beag Thrust. Location of sample RC2184 indicated by asterisk.

metamorphism that is very similar to those ages obtained from recent isotopic studies elsewhere in the Scottish Highlands. This has implications for the age of deformation fabrics and structural correlations both within the Moine outcrop and across the Great Glen Fault.

Geological setting

The Neoproterozoic Moine Supergroup of NW Scotland is tectonically bounded to the west by the Moine Thrust Zone and to the southeast by the Great Glen Fault (Fig. 1). The Moine mainly comprises thick sequences of psammites and pelites that are subdivided into the Morar, Glenfinnan and Loch Eil groups (Holdsworth *et al.* 1994 and references therein). Inliers of basement gneisses occur as thrust slices and in the cores of nappe-scale, isoclinal folds. These gneisses may represent fragments of the basement that floored the Moine sedimentary basin, and have been correlated with the Lewisian Complex of the Caledonian foreland (Holdsworth *et al.* 1994; Soper *et al.* 1998). The Moine rocks record polyphase deformation and metamorphism up to amphibolite facies (e.g. Ramsay 1958; Brown *et al.* 1970; Tobisch *et al.* 1970). The results of integrated field mapping and geochronology provide evidence for three major tectonothermal events.

Formation of the West Highland granite gneiss c. 870 Ma. The junction between the Glenfinnan and Loch Eil groups is generally closely coincident with the outcrop of a series of highly deformed and metamorphosed granites known collectively as the West Highland granite gneiss (Johnstone 1975).

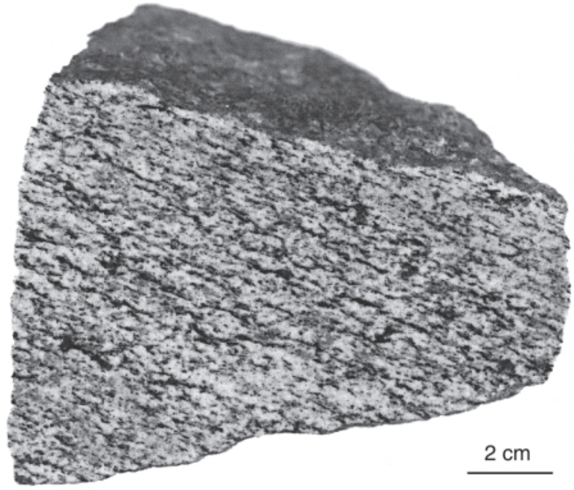

Fig. 2. Hand specimen of the sample dated, showing the planar foliation defined by sub-parallel alignment of biotite and quartzofeldspathic aggregates.

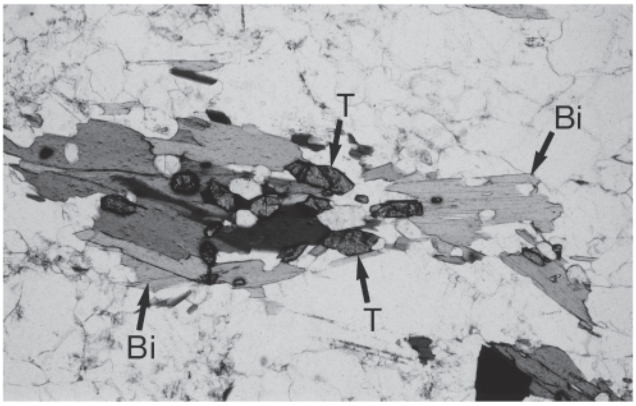

Fig. 3. Titanite (T) associated with biotite (Bi) defining the planar foliation in the Fort Augustus granite gneiss (horizontal field of view *c.* 4 mm).

We accept the conclusions of Barr *et al.* (1985) that they represent S-type, anatectic granites derived by partial melting of Moine metasediments during regional high-grade metamorphism. Zircons from the granite orthogneiss and associated migmatitic segregations at Glenfinnan have yielded near identical U–Pb ages which have been interpreted as indicating that anatexis occurred at 873 ± 7 Ma (Friend *et al.* 1997). The migmatitic segregations are locally discordant to the gneissic banding within the host granite orthogneiss (Barr *et al.* 1985) which suggests that early deformation in this part of the Moine occurred at *c.* 870 Ma (Friend *et al.* 1997). Millar (1999) obtained a U–Pb zircon age of 873 ± 6 Ma for a schistose metagabbro within the Glen Doe granitic gneiss, part of the West Highland granite gneiss. He argued that this gabbro was originally intruded into the granitic protolith of the gneiss, and that both were then affected by the regional D_1 tectonothermal event. The implicit assumption that the granitic gneiss formed during an orogenic event has been questioned by Soper (1994) and Soper & Harris (1997), who have suggested (1) that the granitic protolith was emplaced during regional extension and development of the Moine sedimentary basins (see also Millar 1999) and (2) that all deformation and metamorphism is of Caledonian age.

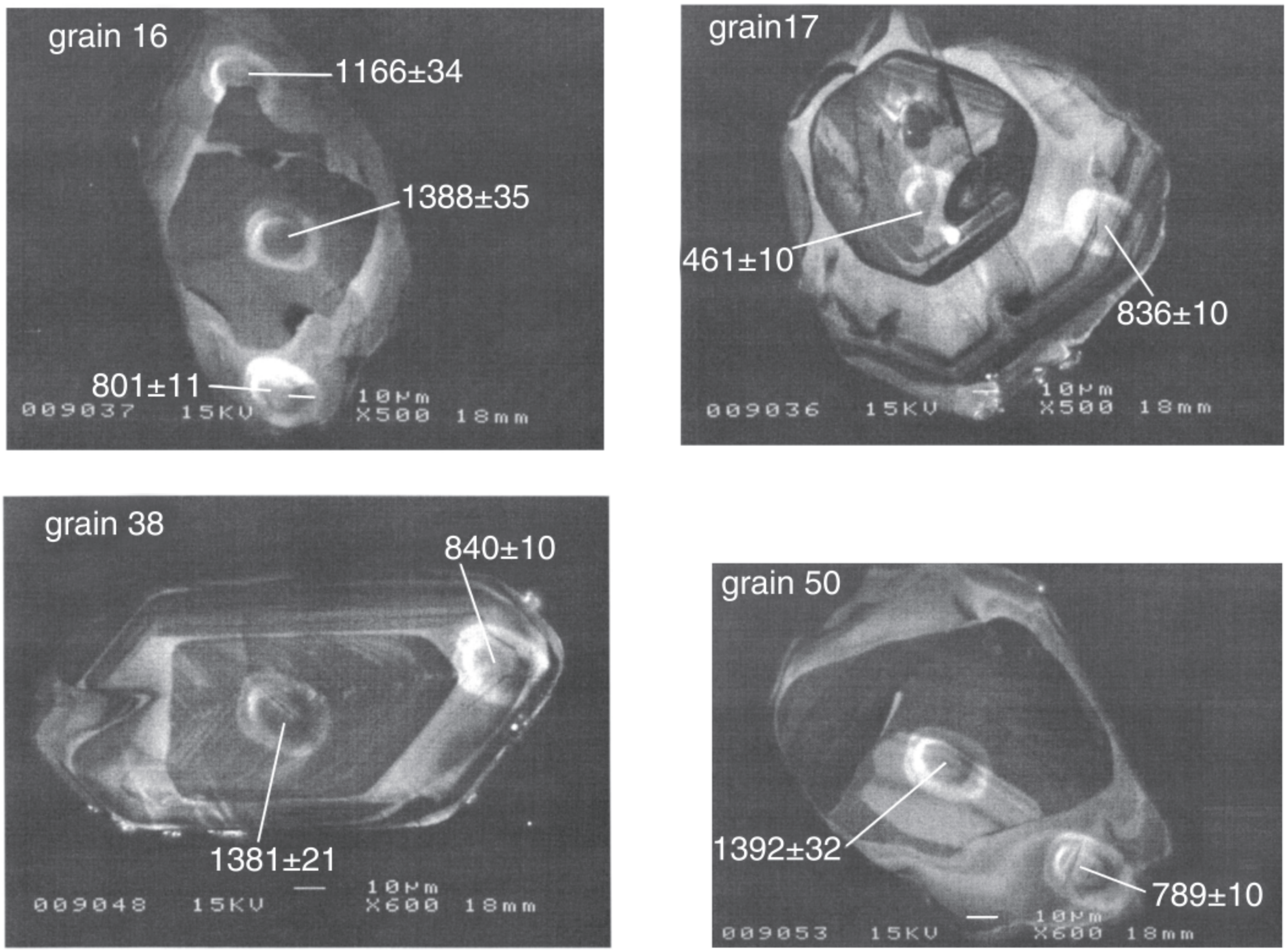

Fig. 4. Examples of CL images of zircons showing ion probe pits from the Fort Augustus granite gneiss. (**a**) grain 16; (**b**) grain 17; (**c**) grain 38; (**d**) grain 50.

Knoydartian deformation and metamorphism c. 830–780 Ma. Early garnet-grade metamorphism of the Morar Group in Inverness-shire has been dated at 820–790 Ma (Sm–Nd garnet; Vance *et al.* 1998). Growth of garnet post-dated early nappe-scale interleaving of Moine rocks and gneissic basement (Powell 1974). The peak pressures of 12.5–14.5 kbar and temperatures of *c.* 650–700°C obtained from the dated garnets are consistent with crustal thickening. This is broadly coeval with the syntectonic generation and intrusion of pegmatites at Ardnish (*c.* 827 Ma) and Sgurr Breac (*c.* 784 Ma) (Fig. 1; U–Pb monazite; Rogers *et al.* 1998). A large number of Rb–Sr muscovite ages between *c.* 776 and 674 Ma for such pegmatites (recalculated from Giletti *et al.* 1961; Long & Lambert 1963; van Breemen *et al.* 1974, 1978; Piasecki & van Breemen 1983; Powell *et al.* 1983; Piasecki 1984) also broadly document this event (Rogers *et al.* 1998).

Caledonian deformation and metamorphism c. 470–420 Ma. Caledonian orogenesis (*sensu lato*) in the North Atlantic region was associated with the closure of the Iapetus Ocean during Ordovician and Silurian times and the convergence of three crustal blocks, Laurentia, Baltica and Eastern Avalonia (e.g. Soper & Hutton 1984; Pickering *et al.* 1988; Soper *et al.* 1992). Early orogenic activity along the Iapetan margins of both Laurentia and Baltica resulted from a series of arc–continent collisions that occurred during initial oceanic closure. It is generally considered that the Moine Supergroup underwent pervasive deformation and metamorphism up to amphibolite facies during the early to mid–Ordovician Grampian arc–continent orogeny (e.g. Kelley & Powell 1985; Powell & Phillips 1985 and references therein). This was followed in the middle to late Silurian by collision of the Scottish segment of Laurentia with Baltica (Coward 1990) to form the Moine Thrust Zone at *c.* 430–420 Ma (van Breemen *et al.* 1979; Kelley 1988; Freeman *et al.* 1998).

Geology of the Fort Augustus granite gneiss

The Fort Augustus granite gneiss constitutes the easternmost body of the West Highland granite gneiss (Johnstone 1975). Between Strontian and Glen Doe (Fig. 1), the granite gneisses follow closely the boundary between the Glenfinnan and Loch Eil groups whereas between Loch Quoich and the Great Glen Fault (Fig. 1), the granite gneisses transgress into the Loch Eil Group. The Fort Augustus granite gneiss forms an elongate mass, truncated to the southeast by the Great Glen Fault (Fig. 1). Only the upper contacts of the mass are seen; these are sharp and concordant with the regional foliation in the Loch Eil Group (May & Highton 1997).

Table 1. *SHRIMP U–Pb data for zircons from the Fort Augustus granite gneiss (RC2184)*

Spot	U	Th	Th/U	Pb	%c206	$^{207}Pb/^{206}Pb \pm 1\ \sigma$		$^{206}Pb/^{238}U \pm 1\ \sigma$		$^{207}Pb/^{235}U \pm 1\ \sigma$		Age (Ma) ± 2 σ	
												$^{207}Pb/^{206}Pb$	
Zircon cores													
6.1	512	64	0.13	136	0.11	0.1001	0.0005	0.2696	0.0035	3.721	0.053	1626	9
7.1	158	57	0.36	44	0.24	0.0974	0.0012	0.2632	0.0035	3.534	0.066	1574	22
10.1	159	95	0.60	50	0.48	0.0944	0.0010	0.2814	0.0037	3.663	0.066	1517	20
17.1	412	57	0.14	97	0.06	0.0917	0.0005	0.2435	0.0031	3.080	0.045	1461	10
42.1	233	102	0.44	59	0.15	0.0898	0.0007	0.2385	0.0031	2.953	0.047	1421	14
50.1	100	18	0.18	22	0.13	0.0884	0.0015	0.2263	0.0030	2.760	0.063	1392	32
16.1	102	46	0.46	25	0.25	0.0883	0.0016	0.2300	0.0031	2.799	0.067	1388	35
38.1	201	84	0.42	50	0.37	0.0880	0.0010	0.2358	0.0031	2.859	0.052	1381	21
41.1	516	183	0.36	123	0.05	0.0872	0.0005	0.2310	0.0030	2.776	0.040	1365	10
7.2*	194	21	0.11	31	0.56	0.0814	0.0013	0.1643	0.0021	1.844	0.040	1232	31
28.3	36	18	0.50	8	2.00	0.0807	0.0049	0.2034	0.0030	2.262	0.146	1213	121
16.2*	146	61	0.42	25	0.44	0.0788	0.0014	0.1624	0.0021	1.764	0.040	1166	34
46.1	244	80	0.33	48	0.39	0.0786	0.0010	0.1942	0.0025	2.106	0.040	1163	24
12.1	274	376	1.37	64	0.09	0.0781	0.0008	0.1790	0.0023	1.927	0.033	1149	19
11.1	103	126	1.22	23	0.95	0.0765	0.0023	0.1697	0.0023	1.790	0.063	1108	61
32.1	284	135	0.48	53	0.16	0.0752	0.0008	0.1769	0.0023	1.835	0.032	1074	21
28.1	58	47	0.81	13	1.56	0.0751	0.0037	0.1943	0.0027	2.012	0.107	1072	100
8.1	60	64	1.07	14	1.44	0.0740	0.0036	0.1857	0.0026	1.895	0.099	1042	98
15.1	169	79	0.46	30	0.40	0.0731	0.0013	0.1700	0.0022	1.714	0.039	1017	35
39.1	69	27	0.40	13	1.59	0.0725	0.0031	0.1741	0.0024	1.739	0.080	999	86
15.3	121	43	0.36	20	0.80	0.0707	0.0020	0.1570	0.0021	1.529	0.051	947	59
												$^{206}Pb/^{238}U$	
Zircon rims													
42.2	94	39	0.41	15	1.26	0.0638	0.0027	0.1466	0.0020	1.290	0.060	882	11
6.2	153	43	0.28	23	1.43	0.0652	0.0020	0.1443	0.0019	1.297	0.045	869	11
15.2	129	60	0.47	20	0.71	0.0652	0.0020	0.1427	0.0019	1.283	0.045	860	11
46.2	236	92	0.39	34	0.15	0.0697	0.0010	0.1405	0.0018	1.350	0.027	848	10
32.2	184	57	0.31	27	0.59	0.0682	0.0015	0.1404	0.0018	1.321	0.035	847	10
38.2	124	40	0.32	18	0.61	0.0668	0.0016	0.1392	0.0018	1.281	0.037	840	10
17.2	167	49	0.29	23	0.45	0.0664	0.0012	0.1384	0.0018	1.266	0.030	836	10
41.2	123	32	0.26	17	0.95	0.0627	0.0022	0.1341	0.0018	1.159	0.045	811	10
39.2	156	33	0.21	21	0.73	0.0669	0.0020	0.1329	0.0017	1.225	0.041	804	10
16.3	57	22	0.38	8	2.65	0.0557	0.0047	0.1324	0.0019	1.016	0.089	801	11
50.2	216	57	0.27	28	0.65	0.0626	0.0016	0.1302	0.0017	1.124	0.033	789	10
2.1	424	2	0.00	31	0.50	0.0576	0.0012	0.0803	0.0010	0.638	0.016	498	6
28.2	488	9	0.02	33	0.27	0.0581	0.0011	0.0726	0.0009	0.581	0.014	452	6
10.2	408	34	0.08	19	1.29	0.0573	0.0023	0.0474	0.0006	0.375	0.017	299	4

*Denotes partial overlap onto rim.

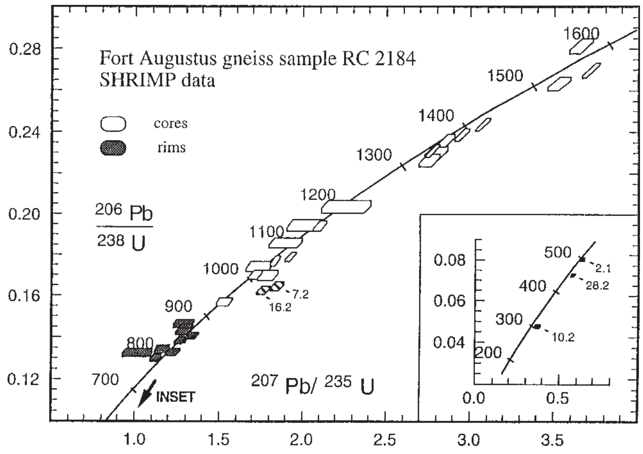

Fig. 5. U–Pb concordia diagram for zircons from the Fort Augustus granite gneiss RC2184. Analyses 7.2 and 16.2 are highly discordant and are interpreted as mixtures.

The granite gneiss sample dated (RC2184, collected at National Grid Reference [NN371071], Fig. 1) is mainly of monzonitic composition, containing quartz, plagioclase, K-feldspar and biotite, with accessory muscovite, titanite, hornblende, garnet, apatite, zircon and orthite. The main foliation in the granite gneiss is a sub-horizontal to gently dipping fabric defined by oriented micas, biotite-rich laminae, elongate mafic aggregates and quartzofeldspathic segregations (Fig. 2; May & Highton 1997). However, locally the biotite fabric and the quartzofeldspathic segregations are folded by mesoscopic, recumbent 'D_2' isoclinal folds which are accompanied by development of a new flat-lying, axial planar mica fabric. The foliation in the granite gneiss is therefore at least in part composite in origin. D_2 fold hinges are parallel to a north–south-trending mineral and extension lineation defined by aligned quartz–feldspar aggregates and biotite grains. The variation in dip and orientation of the foliation within the body defines a broad D_3 antiform (Barr *et al.* 1985; May & Highton 1997). The dated sample is dominated by a planar biotite foliation; titanite is closely associated with the biotite, often elongate along the foliation, occurring either included within or associated with biotite grains (Fig. 3).

Zircon morphologies

High-integrity zircons were hand-picked from the least magnetic zircon fractions separated from the Fort Augustus granite gneiss. The selected grains comprise colourless to pale brown, clear prisms; although most had euhedral pyramidal terminations, some also had more rounded, corroded ends. Cathodoluminescence (CL) imaging revealed that all grains contain a discrete core whose internal zonation was truncated by a surrounding brightly luminescent rim (Fig. 4).

Analytical techniques

U–Pb analyses of zircon were performed on the SHRIMP™ II instrument at Curtin University using techniques described by Friend *et al.* (1997) except that in this study the uncertainty in the normalization of Pb/U ratios to standard zircon CZ3 was 1.27%, and a primary beam current of 3 nA was used, resulting in a sensitivity for Pb isotopes of *c.* 50 cps/ppm. The analytical results are presented in Table 1 with 1σ errors quoted, and displayed in Fig. 5 where the error boxes are shown at the 1σ level. All errors on ages quoted in the text are 2σ.

U–Pb analysis of titanite was performed by thermal ionization mass spectrometry (TIMS) at SURRC following techniques described by Rogers *et al.* (1998) with the exception that the chemical separation of U and Pb was achieved using HBr as an elutant (Rogers & Dunning 1991 and references therein). Analytical data for the titanite separate are presented in Table 2.

U–Pb results

The zircon cores generally yield concordant or near concordant SHRIMP data with $^{207}Pb/^{206}Pb$ ages ranging from 947 ± 59 to 1626 ± 9 Ma (Table 1). These fall into three main groupings with $^{207}Pb/^{206}Pb$ ages: (i) greater than 1461 Ma; (ii) between 1365 and 1421 Ma; (iii) between 947 and 1231 Ma. Most rim analyses plot around concordia with the $^{207}Pb/^{238}U$ ages varying between 882 ± 22 and 789 ± 19 Ma. Three rim analyses (2.1, 10.2 and 28.3, see inset of Fig. 4) have distinctly younger $^{206}Pb/^{238}U$ ages between 300 and 500 Ma, and also have very low Th/U ratios (<0.08) compared to the rest of the analyses. In addition, two analyses (16.2 and 7.2) gave discordant data with $^{207}Pb/^{206}Pb$ ages of 1167 and 1231 Ma; although CL imaging shows that these analyses were from rim areas, the discordant position of the data suggests that the ion beam may have sampled some core material. Alternatively, these grains may contain rim material which is older than the dominant rim material in the other zircons, and that this has lost some Pb subsequent to its crystallization.

The U–Pb TIMS analysis of titanite 'pancakes' from the Fort Augustus granite gneiss yields a concordant datum with a $^{206}Pb/^{238}U$ age of 470 ± 2 Ma. The use of the $^{206}Pb/^{238}U$ age here, and also in the case of the younger zircon ages above, is because this ratio is less sensitive to any uncertainties in the correction of either the laboratory common Pb or the initial common Pb in the minerals.

Discussion

Age of the Fort Augustus granite gneiss

In common with many other U–Pb zircon studies, the zircon cores are interpreted as representing inherited material (e.g. Pidgeon & Aftalion 1978; Rogers *et al.* 1989; Friend *et al.* 1997). In the case of the Fort Augustus granite gneiss, these cores are interpreted to be derived from the metasedimentary host rocks, whereas the rims are considered to be zircon which grew during its crystallization. The spread of the rim data along the concordia curve could suggest that the analyses may either include some differently aged material or do not reflect an undisturbed population. As CL images were used to guide choice of location of analysis site it is concluded that the latter explanation is correct, hence making an accurate determination of the age of the Fort Augustus granite gneiss less certain. The titanite data clearly reflect the effects of Caledonian metamorphism, whose influence has been widely documented throughout the Scottish Highlands (e.g. van Breemen *et al.* 1974; Aftalion & van Breemen 1980; Powell & Phillips 1985).

The zircon rim data (excluding the three very young ages) fall into two distinct groups, with $^{206}Pb/^{238}U$ ages ranging 836–882 Ma and 789–814 Ma. Given the evidence that this body has experienced a Caledonian tectonothermal event, it is possible that the spread of the data for the zircon rims reflects partial Pb loss, and hence that the distribution of the $^{206}Pb/^{238}U$ ages is skewed. In this case the age of the oldest rim

Table 2. *U–Pb titanite data for Fort Augustus granitic gneiss RC2184*

Fraction number and description	Wt (µg)*	U (ppm)	Pb rad. (ppm)	Total common Pb (pg)	Measured		Atomic ratios			Ages (Ma)			ρ§
					$^{206}Pb/^{204}Pb$†	$^{208}Pb/^{206}Pb$‡	$^{206}Pb/^{238}U$‡	$^{207}Pb/^{235}U$‡	$^{207}Pb/^{206}Pb$‡	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	
1 Colourless pancakes	397	144	11.9	22	12108	0.2189	0.07559 ± 37	0.5869 ± 30	0.05631 ± 8	470	469	465	0.958

*Uncertainty in weight ± 6 µg (2 σ).
†Corrected for fractionation and spike, 50 pg Pb blank (isotopic composition ^{208}Pb: ^{207}Pb: ^{206}Pb: ^{204}Pb=36.88: 15.49: 17.35: 1), and 10 pg U blank.
‡Corrected for fractionation, spike, blank and initial common Pb calculated from the model of Stacey & Kramers (1975).
§$^{207}Pb/^{235}U$–$^{206}Pb/^{238}U$ error correlation coefficient (Ludwig 1993).

would most closely approximate the maximum estimate for the age of the magmatic event in which the granitic protolith crystallized; such reasoning is identical to that commonly employed when estimating the crystallization age of polymetamorphic Archaean rocks from SHRIMP data (e.g. Friend & Kinny 1995). Taking the seven analyses in the older of the two groups gives a mean $^{206}Pb/^{238}U$ age of 854 ± 17 Ma whereas the three oldest $^{206}Pb/^{238}U$ ages give a mean age of 870 ± 30 Ma. The three rim analyses with 300–500 Ma $^{206}Pb/^{238}U$ ages also have very low Th/U ratios compared with the other analyses. This suggests that the young ages may be largely caused by growth of younger metamorphic zircon, rather than simply just Pb loss from the older (*c.* 870 Ma) grains. The lowest ages, however, imply that some post-Caledonian Pb loss must also have occurred. Such young Pb loss could also have contributed to the spread in the *c.* 870 Ma data, although the fact that these data lie along and close to the concordia curve suggests that the effects of such recent Pb loss have been minimal.

In summary, we interpret the three oldest $^{206}Pb/^{238}U$ ages as indicating that the granitic protolith to the gneiss was intruded at 870 ± 30 Ma. This is indistinguishable from the at 873 ± 7 Ma age of the Ardgour granite gneiss (Friend *et al.* 1997) and from the 873 ± 6 Ma age for the metagabbros associated with the Glen Doe granite gneiss (Millar 1999), thus supporting the assumption that the various members of the West Highland granite gneiss are part of a single intrusive suite. The data also reinforce the existence of a *c.* 870 Ma tectonomagmatic event within the Moine which appears to be distinct from the later widespread *c.* 830–780 Ma Knoydartian event (Rogers *et al.* 1998; Vance *et al.* 1998). Within the Central Highlands, although there is geochronological evidence for the latter event (Noble *et al.* 1996; Highton *et al.* 1999), no evidence for the earlier, 870 Ma, event has yet been reported. The present data do not, however, help to resolve the ongoing debate as to whether the melting event was associated with orogeny (Friend *et al.* 1997) or crustal extension and development of the Moine sedimentary basins (Soper & Harris 1997; Millar 1999).

Source of the Fort Augustus granite gneiss

The spread of ages from the zircon cores (1626–947 Ma) is similar to that obtained from the Ardgour granite gneiss (Friend *et al.* 1997), though slightly more restricted. Given the evidence presented above for small amounts of Pb loss from the rims it is possible that the cores have also been influenced by such processes. Nonetheless, the data set clearly highlights the absence of Palaeoproterozoic and particularly Archaean detrital grains in the Moine sediments at the level that was melted to form the Fort Augustus granite gneiss. This supports the conclusion of Friend *et al.* (1997) that this part of the Moine Supergroup was not sourced from Archaean basement or from younger rocks which contained a substantial amount of Archaean zircons. The Fort Augustus granite gneiss contains a higher proportion of zircon cores with Grenvillian ages than the Ardgour granite gneiss, though this may simply reflect a sampling or analytical bias.

Age of Lower Palaeozoic deformation and metamorphism of the Moine Supergroup

The *c.* 470 Ma titanite age is interpreted as broadly corresponding to the age of amphibolite-facies metamorphism in the Fort Augustus area during the early to mid-Ordovician Grampian Orogeny (*c.* 480–460 Ma; Soper *et al.* 1999; Oliver *et al.* 2000). The titanite age places a minimum age on the development of the planar mica fabric in the dated sample. The fabric could have formed during the Grampian Orogeny, but it is also possible that static, mimetic growth of titanite overprinted an older, Neoproterozoic Knoydartian fabric of similar age to that demonstrated in the Ardgour granite gneiss (Friend *et al.* 1997). Further data are thus needed to resolve the age of the fabrics within the Fort Augustus granite gneiss.

In the context of the Scottish Caledonides, the *c.* 470 Ma titanite age compares closely with U–Pb zircon ages for Moine migmatites in Sutherland (Kinny *et al.* 1999), and for syn-metamorphic gabbros and crustally derived granites in the Dalradian rocks of the NE Grampian Highlands (Aftalion & Kneller 1987; Rogers *et al.* 1994; Oliver *et al.* 2000). Similar ages for the timing of metamorphism and syn-tectonic granite magmatism have been documented in the Dalradian of Connemara (Cliff *et al.* 1996; Friedrich *et al.* 1999). This event is generally considered to reflect the collision of a volcanic arc with the Laurentian margin during the middle Ordovician and has been correlated with the Taconic event of the Appalachians (e.g. Dewey & Ryan 1990). It was formerly thought that there were significant differences in the timing of early Caledonian metamorphism and deformation either side of the Great Glen Fault (Harris 1995). However, the emerging similarity in the timing of early to mid-Ordovician metamorphism either side of the fault implies that this structure does not juxtapose crustal blocks with significantly different histories with respect to the Grampian Orogeny.

C.R.L.F. and R.A.S. acknowledge research funding from Oxford Brookes University enabling fieldwork and the SHRIMP™ analytical work to be carried out. A. Kelly and V. Gallagher are thanked for expert technical assistance at SUERC. R. E. Holdsworth is thanked for assistance in sample collection. This work was supported by NERC Grant GR3/7222 to GR. The paper has benefitted from reviews by S. Daly and E. Phillips.

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Received 9 November 1999; revised typescript accepted 13 June 2000.
Scientific editing by Ray Burgess.