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U–Pb geochronology of deformed metagranites in central Sutherland, Scotland: evidence for widespread late Silurian metamorphism and ductile deformation of the Moine Supergroup during the Caledonian orogeny

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Abstract: Within the Caledonides of central Sutherland, Scotland, the Neoproterozoic metasedimentary rocks of the Moine Supergroup record NW-directed D_2 ductile thrusting and nappe assembly, accompanied by widespread tight-to-isoclinal folding and amphibolite-facies metamorphism. A series of metagranite sheets which were emplaced and penetratively deformed during D_2 have been dated using SHRIMP U–Pb geochronology. Zircon ages of 424 ± 8 Ma (Vagastie Bridge granite), 420 ± 6 Ma (Klibreck granite) and 429 ± 11 Ma (Strathnaver granite) are interpreted to date emplacement, and hence regional D_2 deformation, during mid- to late Silurian time. Titanite ages of 413 ± 3 Ma (Vagastie Bridge granite) and 416 ± 3 Ma (Klibreck granite) are thought to date post-metamorphic cooling through a blocking temperature of *c*. 550–500 °C. A mid- to late Silurian age for D_2 deformation supports published models that have viewed the internal ductile thrusts of this part of the orogen as part of the same kinematically linked system of foreland-propagating thrusts as the marginal Moine Thrust Zone. The new data contrast with previous interpretations that have viewed the dominant structures and metamorphic assemblages within the Moine Supergroup as having formed during the early to mid-Ordovician Grampian arc–continent orogeny. The mid- to late Silurian D_2 nappe stacking event in Sutherland is probably a result of the collision of Baltica with the Scottish segment of Laurentia.

Keywords: Scotland, Moine Supergroup, Caledonian orogeny, zircons, titanites, metagranites.

Caledonian orogenesis (*sensu lato*) in the North Atlantic region was associated with the closure of the Iapetus Ocean during the Ordovician–Silurian 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 in early to mid-Ordovician time, and preceded by some 40–50 Ma final amalgamation of crustal blocks in late Silurian time. Tectonic models for the Caledonian orogeny therefore depend critically upon the dating of regional deformation events, which allows distinction between these two phases of orogenic activity.

Geological resolution of various parts of the Appalachian–Caledonian orogen has benefited in recent years from the application of high-precision geochronological techniques, which has influenced new tectonic models for areas such as Newfoundland (e.g. Cawood *et al*. 1995), New Brunswick (e.g. van Staal & de Roo 1995), Nova Scotia (e.g. Barr *et al*. 1995) and western Ireland (e.g. Friedrich *et al*. 1999a, 1999b). In contrast, the geochronological database available for Caledonian metamorphic events in northern Scotland still largely comprises a relatively small number of Rb–Sr whole-rock isochron determinations together with scattered Rb–Sr, U–Pb and K–Ar mineral analyses (e.g. Giletti *et al*. 1961; Brown *et al*. 1965*a*, *b*; Miller & Brown 1965; van Breemen *et al*. 1974, 1978, 1979*a*, *b*; Brewer *et al*. 1979; Aftalion & van Breemen 1980; Kelley 1988; Freeman *et al*. 1998; Kinny *et al*. 1999; Rogers *et al*. 2001). In northern Scotland, amphibolite-facies metasedimentary rocks of the Neoproterozoic Moine Supergroup are disposed in a series of east-dipping, Caledonian ductile thrust nappes (Fig. 1a; Barr *et al*. 1986). Published syntheses based on rather sparse isotopic data assign development of the internal ductile thrusts and associated Barrovian metamorphic assemblages to the early to mid-Ordovician Grampian arc–continent orogeny (e.g. Kelley & Powell 1985; Powell & Phillips 1985; Harris 1995). In contrast, the age of low-grade marginal thrusting and emplacement of the Moine rocks onto the Laurentian foreland along the Moine Thrust Zone (Fig. 1a) is more firmly assigned to the late Silurian (van Breemen *et al*. 1979*a*; Kelley 1988; Freeman *et al*. 1998) and probably resulted from the collision of Baltica with the Scottish segment of Laurentia (Coward 1990).

In this paper we present the results of an integrated structural and geochronological study of deformed metagranites within the Moine Supergroup in the Loch Naver area, central Sutherland (Fig. 1b). Recent structural and metamorphic investigations and U–Pb zircon dating in north and central Sutherland provide a detailed framework for tectonic analysis of this part of the Caledonian belt (Holdsworth 1987, 1989*a*, *b*, 1990; Holdsworth & Strachan 1988; Moorhouse & Moorhouse 1988; Moorhouse *et al*. 1988; Strachan & Holdsworth 1988; Holdsworth & Grant 1990; Burns 1994; British Geological Survey 1997; Holdsworth

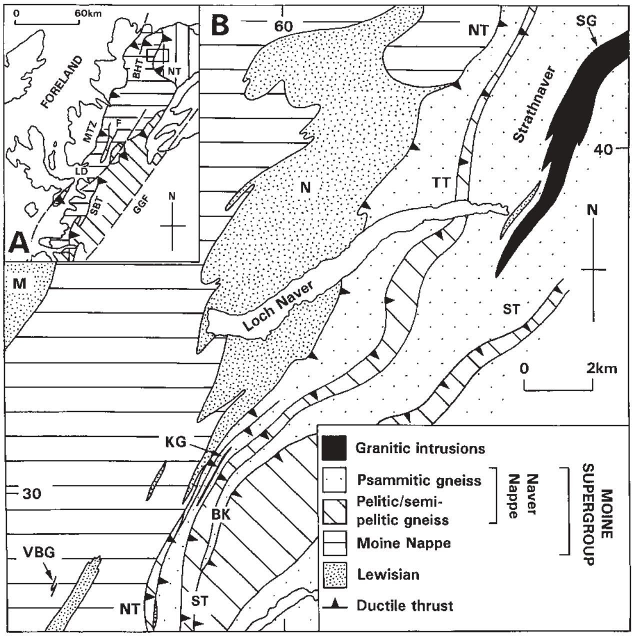

Fig. 1. (**a**) General disposition of the nappe structures NW of the Great Glen Fault (GGF) and above the Moine Thrust (MTZ), showing the Sgurr Beag Thrust (SBT), the Naver Thrust (NT) and the Ben Hope Thrust (BHT). Horizontal lines, Moine nappe; diagonal lines, Sgurr Beag nappe; vertical lines, Naver nappe. LD, Loch Duich; F, Fannich. Small box represents Loch Naver area. (**b**) General geology of area around Loch Naver, central Sutherland. The regional foliation dips gently to moderately to the east or SE (see text for discussion and Fig. 2). Arrows point to the locations of the Strathnaver granite (SG), the Vagastie Bridge granite (VBG) and the Klibreck granite (KG). N, Naver basement inlier; M, Meadie basement inlier; NT, Naver Thrust; TT, Torrisdale Thrust; ST, Swordly Thrust; BK, Ben Klibreck.

et al. 1999, 2002; Kinny *et al.* 1999; Friend *et al.* 2000). The data reported here place constraints on the age of ductile thrusting and related folding and fabric development within this part of the Caledonides. This has implications for the timing of related deformation events elsewhere in the Moine Supergroup, and hence for regional tectonic models for the Caledonian orogeny in this part of the North Atlantic region.

Regional setting

The Moine rocks exposed in central Sutherland comprise two distinct structural units, the Moine and Naver nappes separated by the Naver Thrust (Fig. 1b; Moorhouse 1977; Holdsworth & Strachan 1988; Moorhouse & Moorhouse 1988; Moorhouse *et al.* 1988; Strachan & Holdsworth 1988). Both the Moine and Naver nappes are imbricated internally by subordinate ductile thrusts (e.g. Torrisdale Thrust, Swordly Thrust, Fig. 1b). Moine rocks within the structurally lower Moine nappe are mainly unmigmatized psammites that locally preserve sedimentary structures. The Naver nappe is dominated by strongly deformed and migmatized Moine psammitic and pelitic paragneisses. Concordant amphibolite sheets within the Moine rocks are interpreted as deformed and metamorphosed basic intrusions. Tectonic interleaving of Moine metasediments with basement gneisses assigned to the late Archaean–Palaeoproterozoic Lewisian Complex of the Laurentian foreland occurred during Caledonian ductile thrusting and folding (Strachan & Holdsworth 1988). The larger basement units, such as the Naver and Meadie inliers (Fig. 1b), occupy the cores of major isoclinal antiforms. Other basement units, such as the inlier exposed 3 km SW of Ben Klibreck (Fig. 1b), are structurally located in the hanging walls of ductile thrusts.

Throughout the Moine nappe and the lower part of the Naver nappe in north and central Sutherland, the dominant tight-to-isoclinal, reclined folds, ductile thrusts and associated fabrics are generally related to a D_2 phase of deformation (e.g. Soper & Brown 1971; Soper & Wilkinson 1975; Barr *et al.* 1986; Moorhouse & Moorhouse 1988; Moorhouse *et al.* 1988; Strachan & Holdsworth 1988; Holdsworth 1989*a*, 1990; Holdsworth & Grant 1990; Alsop & Holdsworth 1993; Burns 1994; Alsop *et al.* 1996). In the Naver nappe, D_2 folds deform migmatitic layering. Regional D_2 ductile thrusting resulted in development of a composite S_0–S_1–S_2 foliation, which intensifies into broad belts of mylonites associated with discrete ductile thrust surfaces. A prominent mineral lineation (L_2) defines the Caledonian thrust transport direction. This displays a change in orientation from ESE-plunging in the west adjacent to the Moine Thrust to SSE-plunging in the east adjacent to the Naver Thrust (Barr *et al.* 1986; Alsop *et al.* 1996). Shear sense indicators suggest overthrusting towards the WNW to NNW (Holdsworth & Grant 1990; Holdsworth *et al.* 2002). A number of lines of evidence indicate that D_2 was accompanied by regional amphibolite-facies metamorphism at temperatures higher than *c.* 550–600 °C. These include: the presence of syn-D_2 garnet and staurolite within Moine pelites, the widespread parallelism of hornblende with L_2 in mafic rocks, the identification of post-D_2 kyanite within Lewisian metasedimentary rocks of the Meadie inlier, and syn- to post-D_2 sillimanite in Moine pelites of the Naver nappe (Strachan & Holdsworth 1988; Holdsworth 1989a; Burns 1994; Holdsworth *et al.* 2002; Strachan, unpublished data; see also Friend *et al.* 2000).

The Naver nappe and structurally high levels of the Moine nappe are intruded by a suite of concordant granitic sheets (Fig. 1b; Read 1931; Brown 1967, 1971; Barr 1985; Holdsworth & Strachan 1988). Some of the sheets analysed by Brown (1967) are peraluminous and may be of crustal derivation. However, other members have chemical characteristics comparable with the Northern Highland high Ba–Sr granitoids, which are thought to have been derived in part from the mantle (e.g. Fowler *et al.* 2001; Kocks & Fowler, unpublished data). They mostly range from 10 cm to 30 m in thickness, although the Strathnaver granite is significantly larger (Fig. 1b). The intrusions occasionally incorporate metasedimentary xenoliths, and contacts with host rocks are sharp. Some members of the suite cut D_2 folds but all carry the composite S_2 schistosity and the L_2 linear fabric, and some sheets are strongly boudinaged. Although no quantitative metamorphic study has yet been undertaken, there is widespread petrographic evidence for thorough recrystallization within the amphibolite facies. These observations are consistent with a syn-D_2 age of emplacement for the granites, broadly synchronous with displacements along the Naver Thrust (Barr 1985; Holdsworth & Strachan 1988).

Recent metamorphic studies carried out along the well-exposed coastal section of north Sutherland indicate that the Caledonian orogeny comprises two tectonothermal events each defined by a separate P–T loop (Friend *et al.* 2000). The earliest resulted in high-pressure granulite-facies metamorphism, relics of which are preserved within the cores of amphibolite sheets that intrude Moine gneisses of the Naver nappe. High-pressure metamorphism was accompanied by melting and migmatization. U–Pb (SHRIMP) dating of zircons from Moine migmatites within the Naver and Kirtomy nappes of north Sutherland indicates that melting occurred at 467 ± 10 Ma and 461 ± 13 Ma, respectively (Kinny *et al.* 1999). This event corresponds to the early to mid-Ordovician Grampian orogenic event recognized elsewhere in Scotland and Ireland (Lambert & McKerrow 1976; Friedrich *et al.* 1999*a*, *b*; Soper *et al.* 1999; Oliver *et al.* 2000), which has been correlated broadly with the Taconic event in the Appalachians. The later event resulted in renewed crustal thickening and was associated with development of the D_2 structures and related metamorphic assemblages outlined above. However, it is at present uncertain whether these two events are both part of the Grampian event, or whether they are entirely unrelated. To resolve this problem, we focus on the field relations, structure and geochronology of syn-D_2 metagranite sheets that intrude the Moine metasediments above and below the Naver Thrust in central Sutherland. Three have been examined in detail, the Vagastie Bridge, Klibreck and Strathnaver granites (Fig. 1b).

Geology of the syn-D_2 metagranites

Vagastie Bridge granite

The Vagastie Bridge granite is a composite intrusion, formed of four subconcordant, anastomosing sheets (Holdsworth & Strachan 1988). It is at least 500 m long and attains a maximum thickness of *c.* 50 m (Fig. 1b). Most of the intrusion comprises a coarse-grained, pink gneissic granite containing numerous augen of recrystallized perthitic orthoclase up to 1.25 cm across. The matrix typically comprises plagioclase (An_{15-20}), K-feldspar (orthoclase), quartz, hornblende, biotite, secondary chlorite and accessory titanite, zircon and magnetite. Trails of euhedral to anhedral titanite grains are generally associated with the biotite layers that lie parallel to the gneissic foliation. Subordinate granodiorite and diorite are also present, and are characterized by a reduction in feldspar and quartz, and a corresponding increase in the proportion of hornblende. The diorite component also contains isolated grains of clinopyroxene (diopside), which are rimmed and partially replaced by green hornblende.

The field relations of the Vagastie Bridge granite have been documented in detail by Holdsworth & Strachan (1988). Although the intrusion is broadly concordant with lithological banding (tectonically modified bedding) in the host psammites of the Moine nappe, F_2 tight-to-isoclinal minor folds are consistently cut by granite sheets and veins. Emplacement therefore clearly postdated F_2 folding. All the intrusive sheets carry a strong L>S fabric defined by alignment of hornblende and/or biotite and augen of recrystallized feldspar. The planar component of this fabric is parallel to S_2 in the host rocks, and the linear component is parallel to F_2 fold axes and the L_2 lineation (Fig. 2a). Granite emplacement is therefore considered to have mainly post-dated F_2 folding, but to have predated a late component of flattening co-planar with S_2 and extension collinear with L_2 (Holdsworth & Strachan 1988).

Klibreck granite

The Klibreck granite is located 100 m structurally above the Naver Thrust on the steep, western slope of Ben Klibreck (Fig. 1b). It forms a well-defined lenticular sheet that is traceable laterally for *c.* 2 km, and is *c.* 30 m thick at maximum. Contacts with host Moine gneisses are sharp. Most of the intrusion is a pink, equigranular, medium-grained metagranite or metagranodiorite, comprising variable proportions of plagioclase (An_{20-32}), K-feldspar, quartz and biotite, with accessory titanite, magnetite and zircon. Titanite occurs as elongate aggregates of euhedral to anhedral grains, either included within or associated with the biotite layers that lie parallel to the foliation. Biotite is locally replaced by secondary chlorite. Minor, decimetre-scale patches of hornblendic mafic material have been interpreted as a minor dioritic facies of the intrusion, and are petrographically identical to those of the Vagastie Bridge granite (Holdsworth & Strachan 1988). The margins of the intrusion locally truncate isoclinal F_2 folds within host Moine gneisses. A penetrative L>S fabric is defined within the intrusion by the subparallel alignment of biotite grains, which wrap elongate aggregates of recrystallized plagioclase and K-feldspar. The planar component of this fabric is parallel to S_2 in the host gneisses, and the linear component is parallel to L_2 (Fig. 2b). The field relations are therefore interpreted as indicating that the Klibreck granite was emplaced during the later stages of D_2.

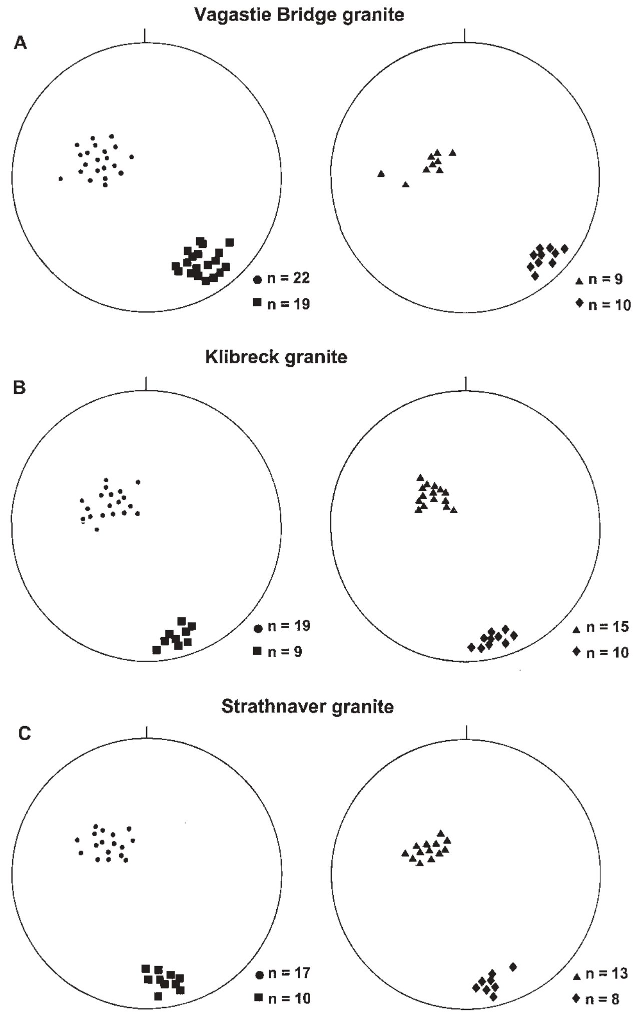

Fig. 2. Equal area, lower hemisphere stereographic projections of structural data from the (**a**) Vagastie Bridge, (**b**) Klibreck and (**c**) Strathnaver granites and their Moine host rocks. In all cases, the stereonets on the left represent the structural data obtained from the Moine host rocks (●, poles to composite $S_0-S_1-S_2$ foliation; ■, L_2 lineation), and the stereonets on the right the data obtained from the granites (▲, poles to S_2 foliation; ◆, L_2 lineation).

Strathnaver granite

The Strathnaver granite is located within the Naver nappe, and is sparsely exposed in the river and on the eastern slopes of Strathnaver (Fig. 1b). It has an overall lenticular geometry and is *c*. 500 m thick at maximum. Contacts with host Moine gneisses of the Naver nappe are commonly sheeted, but invariably sharp, concordant and gently dipping where best exposed at the south end of the intrusion. Local highly micaceous patches within the intrusion are interpreted as remnants of xenoliths of Moine semi-pelitic gneiss. The metagranite is typically medium-grained, pink and equigranular, with a heterogeneously developed D_2 L–S fabric defined by the subparallel alignment of trails of biotite flakes interspersed within a granoblastic groundmass of recrystallized quartz, plagioclase (An_{20-30}) and subordinate K-feldspar. Subordinate coarser-grained variants are present locally and display a characteristic augen texture defined by lenticular, recrystallized feldspar megacrysts. Biotite is commonly replaced by secondary chlorite; accessory minerals include titanite, zircon and magnetite. Titanite is typically closely associated with biotite grains. The planar component of this deformation fabric is parallel to S_2 in the host gneisses, and the linear component is parallel to L_2 (Fig. 2c). Although the temporal relationship between emplacement and F_2 folding is uncertain because of poor exposure, the intrusion is intepreted as essentially part of the same syn-D_2 suite as the Vagastie Bridge and Klibreck intrusions.

U–Pb geochronology

Analytical techniques

Zircons separated from the rock samples were mounted in epoxy resin, sectioned, imaged by cathodoluminescence (CL), and dated using the SHRIMP-II ion microprobe at the Isotope Science Research Centre, Curtin University, Perth. Operating conditions for SHRIMP were routine, namely 25 μm analytical spot size, primary beam current 2–5 nA, mass resolution 5000R (1% valley), and sensitivity for Pb isotopes *c*. 15 c.p.s. ppm^{-1} nA^{-1}. Correction of isotope ratios for common Pb was based on the measured ^{204}Pb, representing in all cases <5% correction to the ^{206}Pb counts (see %c206, Table 1), with the common Pb composition modelled upon that of Broken Hill ore Pb. Pb/U isotopic ratios were corrected for instrumental inter-element discrimination using the observed covariation between $^{206}Pb/^{238}U$ and UO/U (Compston *et al*. 1984, 1992) determined from interspersed analyses of the Perth standard zircon CZ3 (564 Ma; $^{206}Pb/^{238}U = 0.0914$). The uncertainty in Pb/U ratios associated with this correction procedure was in the range 1.8–2.1%, whereas uncertainties in Pb/Pb ratios were generally lower, being governed principally by counting statistics. Isotope ratios and corresponding ages (calculated using standard decay constants) are listed in Table 1, together with 1σ uncertainties. Unless otherwise stated, all ages discussed in the text are given with 95% confidence limits.

U–Pb analysis of titanite was performed by thermal ionization mass spectrometry (TIMS) at the Scottish Universities Environmental Research Centre at East Kilbride. Techniques followed those 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 two separates of the Klibreck granite and one of the Vagastie granite are presented in Table 2.

Sample and zircon characteristics

The sample of Vagastie Bridge granite from which zircons were separated (99-1) was collected in the river section 10 m east of Vagastie Bridge (Fig. 1b; National Grid Reference NC 53502825). Abundant zircons were recovered from the sample. They are euhedral to subhedral prismatic grains up to several hundred microns in size, ranging from colourless to pale pink or yellow in colour. CL imaging revealed the presence of a discrete structural core in most grains, surrounded by a broad mantle of finely oscillatory zircon (Fig. 3, top). The sample of Vagastie Bridge from which titanites were separated (RC2124) was collected from the same locality.

The sample of Klibreck granite (RC2122) from which both zircons and titanites were separated was collected *c*. 1300 m NNW of the summit of Ben Klibreck (Fig. 1b; NC 58153110). Zircons from the sample are euhedral to subhedral prismatic grains, including some particularly narrow, delicate grains of high elongation ratio. Colours ranged from colourless to pale yellowy or orangey brown. CL imaging again showed the presence of a discrete, brightly luminescent structural core in most grains (Fig. 3, middle), although a number of the particularly elongate grains lacked any visible core.

The sample of Strathnaver granite (99-2) was collected from a road cutting on the east side of Strathnaver (Fig. 1b; NC 69104270). Zircons are subhedral prismatic grains, mostly under 200 μm in length. Colours ranged from yellowish to reddish brown, with some grains having darker interiors. As in both of the previous samples, structural cores were visible in CL, often highly resorbed and anhedral in outline. The surrounding mantles are clearly zoned, but poorly luminescent, owing to their very high U contents (Fig. 3, bottom).

Results and interpretation

Vagastie Bridge granite

The results of SHRIMP spot analyses of zircons from Vagastie Bridge sample 99-1 are listed in Table 1 and summarized on U–Pb concordia plots in Figures 4 and 5. Analyses of structural cores fall into two age groups: an older population preserving late Palaeoproterozoic to Mesoproterozoic ages (Fig. 4), and five younger cores, four of which form an approximately concordant group with a mean $^{206}Pb/^{238}U$ age of 430 ± 19 Ma (mean square weighted deviation (MSWD) 1.14), the fifth slightly older (shaded zircon analyses, Fig. 5). Analyses of the oscillatory zoned mantles surrounding these cores, which we interpret as magmatically crystallized zircon, form a second coherent population characterized by higher U content and somewhat lower Th/U than both groups of cores (Table 1). They form a concordant group (unshaded analyses, Fig. 5) with a slightly (but not significantly) younger mean $^{206}Pb/^{238}U$ age of 424 ± 8 Ma (MSWD 1.16, two outliers omitted). The results of TIMS analysis of a bulk fraction of titanites from a different sample of Vagastie Bridge granite (RC2124) are listed in Table 2 and plotted together with the zircon data in Figure 5 (black ellipse). The titanite fraction has a mean $^{206}Pb/^{238}U$ age of 413 ± 3 Ma. There is thus a small, but significant difference between the ages indicated by the magmatic zircon and titanite populations. We interpret the mean $^{206}Pb/^{238}U$ age of the zircon rims in sample 99-1 as giving the time of magmatic crystallization. The slightly younger titanite date from sample RC2124 reflects either a real difference in the crystallization ages of the two samples or, more probably, the effects of post-crystallization cooling through the somewhat lower blocking temperature of titanite. The zircon cores are considered to represent restitic zircons inherited from rocks that contributed to the granitic melt. They therefore record the ages of events affecting the source rocks before anatexis.

Table 1. *SHRIMP U–Pb age data for zircons from the Vagastie Bridge, Klibreck and Strathnaver granites*

Spot	U	Th	Th/U	%c^{206}Pb	^{207}Pb/^{206}Pb	±1σ	^{206}Pb/^{238}U	±1σ	^{207}Pb/^{235}U	±1σ	Age	±1σ
Vagastie Bridge granite zircons, old cores											^{207}Pb/^{206}Pb age	
7a	135	57	0.42	0.48	0.1011	0.0022	0.2853	0.0053	3.975	0.120	1644	40
4a	49	96	1.97	2.19	0.0975	0.0036	0.2669	0.0053	3.588	0.159	1577	70
2a	462	327	0.71	0.18	0.0949	0.0007	0.2500	0.0044	3.272	0.065	1526	14
3a	128	43	0.33	0.86	0.0779	0.0019	0.2048	0.0041	2.199	0.072	1144	47
Vagastie Bridge granite zircons, young cores											^{206}Pb/^{238}U age	
5a	133	87	0.66	1.37	0.0580	0.0040	0.0732	0.0015	0.586	0.043	455	9
1a	122	110	0.91	4.46	0.0557	0.0079	0.0705	0.0016	0.541	0.079	439	10
6a	116	97	0.84	2.69	0.0552	0.0058	0.0705	0.0014	0.536	0.059	439	9
9b	82	50	0.60	3.47	0.0464	0.0077	0.0675	0.0015	0.431	0.073	421	9
10a	81	48	0.59	3.08	0.0441	0.0071	0.0673	0.0015	0.409	0.067	420	9
Vagastie Bridge granite zircons, magmatic rims											^{206}Pb/^{238}U age	
9a	676	162	0.24	0.86	0.0544	0.0012	0.0707	0.0013	0.530	0.016	440	8
10b	580	119	0.20	0.60	0.0539	0.0012	0.0696	0.0012	0.518	0.015	434	7
1b	746	241	0.32	1.13	0.0538	0.0013	0.0695	0.0012	0.515	0.016	433	7
7b	1559	458	0.29	0.19	0.0556	0.0007	0.0684	0.0012	0.525	0.012	427	7
11	726	138	0.19	0.40	0.0552	0.0013	0.0684	0.0012	0.521	0.016	427	7
3b	750	268	0.36	0.85	0.0549	0.0012	0.0683	0.0012	0.517	0.015	426	7
12	772	238	0.31	0.31	0.0563	0.0009	0.0680	0.0012	0.528	0.013	424	7
2b	796	220	0.28	0.33	0.0554	0.0009	0.0676	0.0011	0.517	0.013	422	7
5b	840	162	0.19	0.13	0.0554	0.0007	0.0669	0.0011	0.511	0.012	417	7
8	542	132	0.24	0.38	0.0550	0.0011	0.0666	0.0011	0.505	0.014	416	7
6b	1135	322	0.28	0.79	0.0569	0.0011	0.0660	0.0012	0.518	0.015	412	7
4b	689	149	0.22	1.35	0.0566	0.0018	0.0650	0.0012	0.507	0.019	406	7
Klibreck granite zircons, cores											^{207}Pb/^{206}Pb age	
36	157	41	0.26	0.00	0.10893	0.00063	0.3043	0.0053	4.571	0.084	1782	11
53	242	110	0.46	0.02	0.10857	0.00052	0.3179	0.0051	4.760	0.080	1776	9
45a	126	75	0.60	0.04	0.10779	0.00084	0.3073	0.0055	4.567	0.089	1762	14
50a	861	399	0.46	0.52	0.10064	0.00040	0.2810	0.0042	3.900	0.060	1636	7
51	149	84	0.56	0.05	0.09979	0.00076	0.2876	0.0050	3.957	0.075	1620	14
35a	379	140	0.37	0.13	0.09766	0.00053	0.2305	0.0036	3.104	0.051	1580	10
40	309	132	0.43	0.25	0.09407	0.00065	0.2246	0.0036	2.913	0.050	1510	13
64	164	80	0.48	0.04	0.09354	0.00074	0.2611	0.0045	3.367	0.064	1499	15
65a	107	85	0.79	3.56	0.09261	0.00267	0.2358	0.0047	3.011	0.105	1480	55
24	287	122	0.42	0.07	0.09251	0.00053	0.2514	0.0040	3.207	0.054	1478	11
6a	128	144	1.13	0.38	0.09050	0.00122	0.2192	0.0040	2.736	0.062	1436	26
57	153	64	0.42	0.04	0.08200	0.00073	0.1935	0.0033	2.188	0.042	1246	17
71	131	69	0.52	0.18	0.08099	0.00124	0.1890	0.0035	2.111	0.051	1221	30
77	162	47	0.29	0.00	0.08035	0.00075	0.1453	0.0026	1.609	0.032	1206	18
70	143	86	0.60	0.07	0.07980	0.00093	0.1962	0.0035	2.159	0.046	1192	23
Klibreck granite zircons, magmatic rims											^{206}Pb/^{238}U age	
6b	5875	1320	0.22	0.03	0.05558	0.00015	0.0694	0.0010	0.532	0.008	432	6
61	2764	377	0.14	0.06	0.05568	0.00027	0.0685	0.0010	0.525	0.008	427	6
32	2417	468	0.19	0.32	0.05529	0.00035	0.0681	0.0010	0.519	0.008	425	6
63	687	524	0.76	0.12	0.05557	0.00060	0.0680	0.0011	0.521	0.010	424	6
39a*	597	518	0.87	0.03	0.05565	0.00063	0.0677	0.0011	0.519	0.010	422	6
66	3123	238	0.08	0.01	0.05553	0.00022	0.0676	0.0010	0.518	0.008	422	6
50b	2301	635	0.28	0.01	0.05547	0.00027	0.0667	0.0010	0.510	0.008	416	6
39b*	586	485	0.83	0.40	0.05454	0.00087	0.0667	0.0011	0.502	0.011	416	6
45b	1662	244	0.15	0.04	0.05573	0.00034	0.0665	0.0010	0.511	0.008	415	6
65b	3058	434	0.14	0.19	0.05599	0.00029	0.0664	0.0010	0.512	0.008	414	6
35b	2015	402	0.20	0.37	0.05569	0.00040	0.0664	0.0010	0.510	0.008	414	6
Strathnaver granite zircons, old cores											^{207}Pb/^{206}Pb age	
2b	476	212	0.45	0.12	0.1165	0.0007	0.3093	0.0055	4.969	0.096	1903	11
Strathnaver granite zircons, young cores											^{206}Pb/^{238}U age	
1a	175	81	0.46	1.83	0.0524	0.0032	0.0756	0.0015	0.546	0.036	470	9
6b	84	18	0.21	4.72	0.0509	0.0079	0.0749	0.0017	0.526	0.084	465	10
5b	131	77	0.59	1.18	0.0627	0.0040	0.0735	0.0015	0.635	0.044	457	9
Strathnaver granite zircons, magmatic rims											^{206}Pb/^{238}U age	
2a	2455	17	0.01	0.22	0.0558	0.0004	0.0712	0.0012	0.548	0.010	443	7
3	2364	95	0.04	0.86	0.0558	0.0007	0.0706	0.0012	0.544	0.012	440	7
9	4007	275	0.07	0.63	0.0566	0.0005	0.0699	0.0012	0.545	0.011	436	7
6a	2458	95	0.04	1.18	0.0555	0.0010	0.0692	0.0013	0.530	0.014	431	8
4	1050	202	0.19	1.99	0.0569	0.0018	0.0676	0.0014	0.531	0.021	422	8
5a	1629	10	0.01	0.37	0.0565	0.0008	0.0676	0.0013	0.527	0.013	422	8
8	2447	17	0.01	0.21	0.0550	0.0004	0.0673	0.0011	0.511	0.010	420	7
10	1000	9	0.01	0.39	0.0559	0.0011	0.0667	0.0012	0.514	0.015	416	7
1b	3021	24	0.01	0.20	0.0564	0.0007	0.0646	0.0011	0.506	0.011	404	7
7	1693	61	0.04	0.30	0.0560	0.0006	0.0640	0.0011	0.493	0.011	400	7

%c^{206}Pb is the percentage of common (non-radiogenic) ^{206}Pb estimated from the measured ^{204}Pb/^{206}Pb ratio. U and Th concentrations are ppm. Errors are 1σ. Ages were calculated using standard decay constants.

*Grain 39 wholly magmatic; no core.

Table 2. *TIMS U–Pb data for titanites from the Klibreck and Vagastie Bridge granites*

Fraction number and description	Wt (µg)*	U (ppm)	Pb rad. (ppm)	Total common Pb (pg)	Measured		Atomic ratios			Ages (Ma)			$\rho^{\S}$
					$^{206}Pb/^{204}Pb^{\dagger}$	$^{208}Pb/^{206}Pb^{\ddagger}$	$^{206}Pb/^{238}U^{\ddagger}$	$^{207}Pb/^{235}U^{\ddagger}$	$^{207}Pb/^{206}Pb^{\ddagger}$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	
Klibreck granite (RC 2122)													
1 Colourless	474	286	18.2	3100	199.8	0.0523	0.06675 ± 40	0.5085 ± 34	0.05525 ± 15	417	417	422	0.911
2 Colourless	383	374	23.6	1700	373.4	0.0453	0.06660 ± 38	0.5119 ± 34	0.05575 ± 18	416	420	442	0.864
Vagastie Bridge granite (RC 2124)													
3 Brown	403	66	6.38	920	137.4	0.6354	0.06617 ± 53	0.4932 ± 86	0.05406 ± 81	413	407	374	0.531

*Uncertainty in weight ±6 µg (2σ).
†Corrected for fractionation and spike, 50 pg Pb blank (isotopic composition ^{208}Pb:^{207}Pb:^{206}Pb:^{204}Pb is 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 Stacy & Kramers (1975).
§$^{207}Pb/^{235}U - ^{206}Pb/^{238}U$ error correlation coefficient (Ludwig 1993).

Klibreck granite

The results of SHRIMP zircon analyses from the Klibreck granite sample RC2122 are plotted in U–Pb concordia diagrams in Figures 4 and 6. All of the structural cores identified in the zircons yielded Proterozoic ages, ranging from *c*. 1200 to 1800 Ma, and overlapping the recorded age range of old cores in the Vagastie Bridge zircons (Fig. 4). These cores are considered to represent restitic zircon inherited from the rocks that contributed to the granitic melt. The surrounding rims, together with whole grains lacking a structural core, form an approximately concordant group of analyses with a calculated mean $^{206}Pb/^{238}U$ age of 420 ± 6 Ma (MSWD 0.6, analysis 6b with extremely high U content excluded) (Fig. 6). These are interpreted as the principal magmatic population. The results of TIMS analysis of two titanite fractions from this sample are shown in Figure 6 (black ellipses). They have a mean $^{206}Pb/^{238}U$ age of 416 ± 3 Ma, which is statistically indistinguishable from the age given by the magmatic zircon rims.

Strathnaver granite

Zircons from the Strathnaver granite yielded similar results to those from the Vagastie Bridge granite insofar as two distinct ages of cores were found, together with a younger magmatic population. Analyses of three cores form an approximately concordant cluster of points with a mean $^{206}Pb/^{238}U$ age of 464 ± 26 Ma (shaded analyses, Fig. 7), whereas analysis 2b of an old core yielded a Proterozoic age (minimum *c*. 1900 Ma, Fig. 4). The zircon rim population was again relatively U rich, with all points having very low Th/U ratios (Table 1). They form an approximately concordant group (unshaded analyses, Fig. 7) with a significant spread in $^{206}Pb/^{238}U$ ages, which may reflect some recent Pb loss. Exclusion of the three analyses with lowest $^{206}Pb/^{238}U$, and analysis 2a, results in a mean age estimate for the remaining six rim analyses of 429 ± 11 Ma (MSWD 1.18). This date is interpreted as the approximate age of crystallization of the granite, the distinctly older cores interpreted as representing inherited restitic zircon.

Discussion and conclusions

Age of D_2 deformation and associated metamorphism

The zircon age estimates obtained for the metagranites of 424 ± 8 Ma (Vagastie Bridge granite), 420 ± 6 Ma (Klibreck granite) and 429 ± 11 Ma (Strathnaver granite) are interpreted to date emplacement of the intrusions during mid- to late Silurian time. These ages overlap within error, consistent with the three intrusions belonging to the same suite. Given that the metagranites are thought to have been emplaced during regional D_2 deformation, amphibolite-facies metamorphism and NW-directed displacement along the Naver Thrust, the zircon ages are also interpreted as demonstrating a mid- to late Silurian age for this event in central and west Sutherland. There is a clear temporal distinction between this late Silurian orogenic event and the mid-Ordovician tectonothermal event identified from the SHRIMP dating of zircons in migmatites in east Sutherland (Kinny *et al*. 1999; Friend *et al*. 2000).

The titanite age of 416 ± 3 Ma for the Klibreck granite is indistinguishable from the 420 ± 6 Ma mean age for the Klibreck magmatic zircons, whereas the 413 ± 3 Ma age for the Vagastie Bridge titanites is marginally younger than the corresponding magmatic zircon population. The titanite ages also

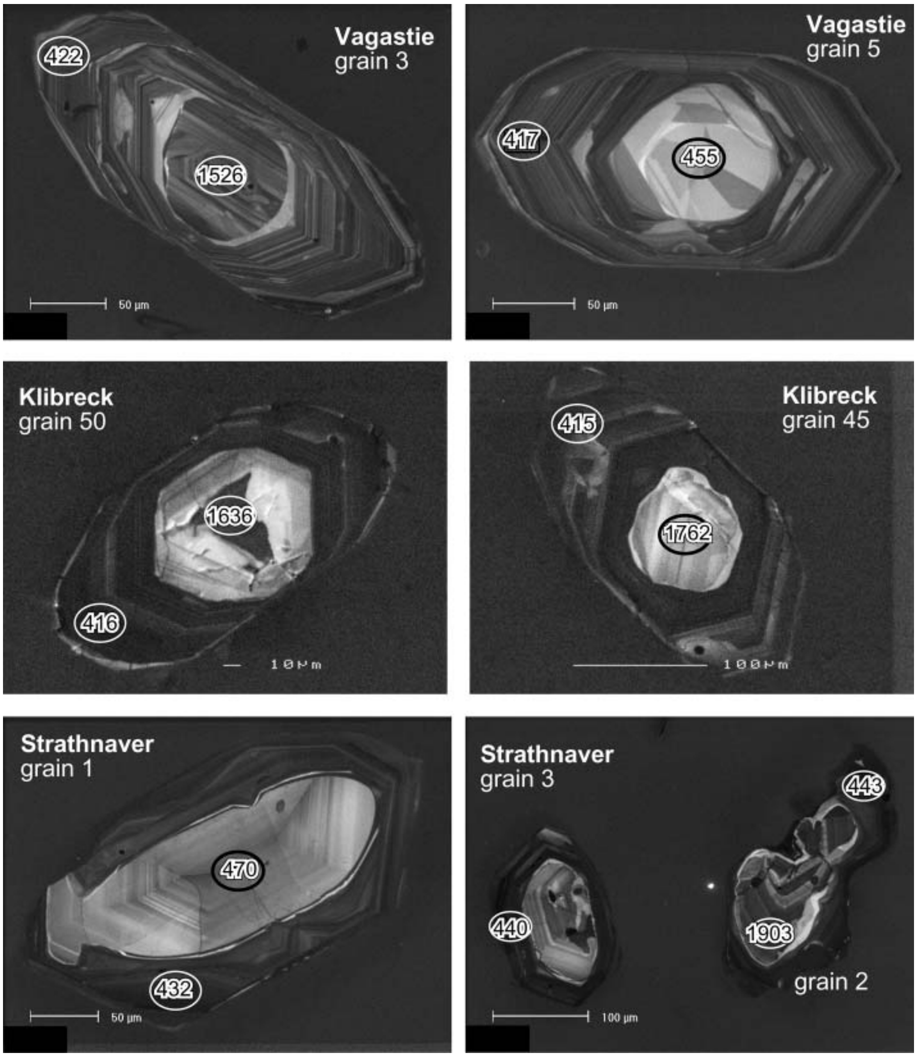

Fig. 3. Cathodoluminescence images of selected representative zircons from the Vagastie Bridge (top), Klibreck (middle) and Strathnaver (bottom) granites, showing distinct core and rim structures. SHRIMP analytical sites are shown as ellipses together with the indicated U–Pb age in Ma (data from Table 1).

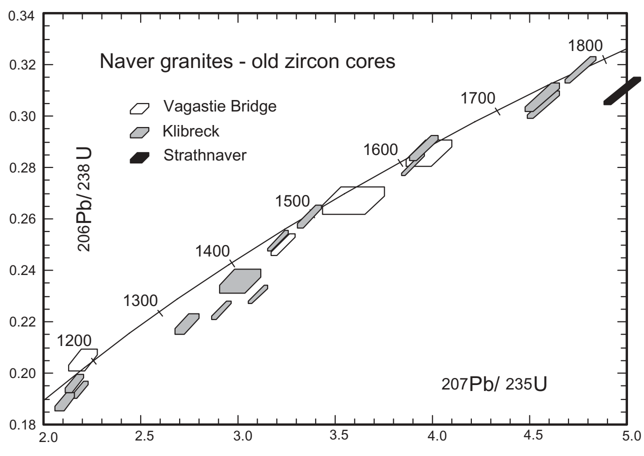

Fig. 4. Combined U–Pb concordia diagram showing SHRIMP analyses of older inherited zircon cores from the Vagastie Bridge, Klibreck and Strathnaver granites (1σ error boxes, as given in Table 1).

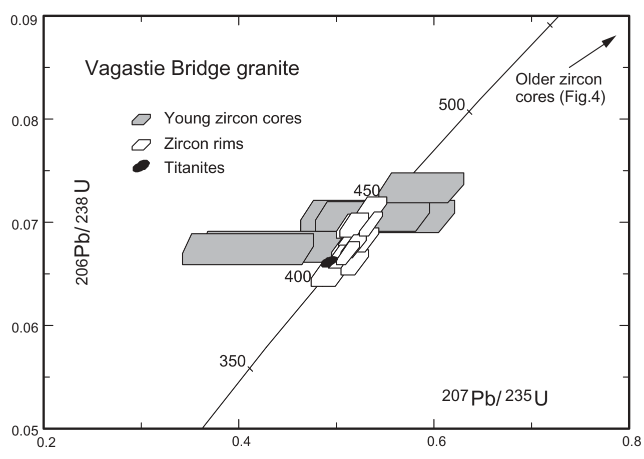

Fig. 5. U–Pb concordia diagram showing the results of SHRIMP analyses of young zircon cores and rims from the Vagastie Bridge granite (1σ error boxes) and TIMS analyses of titanites (2σ error ellipse). Analyses of older inherited zircon cores are shown in Figure 4.

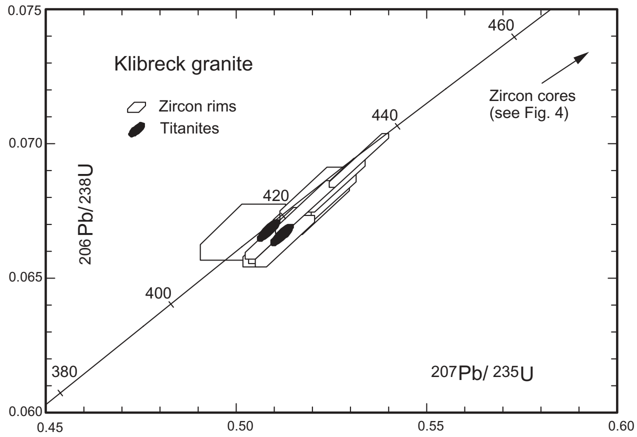

Fig. 6. U–Pb concordia diagram showing the results of SHRIMP analyses of magmatic zircon rims from the Klibreck granite (1σ error boxes) and TIMS analyses of titanites (2σ error boxes). Analyses of older inherited zircon cores are given in Figure 4.

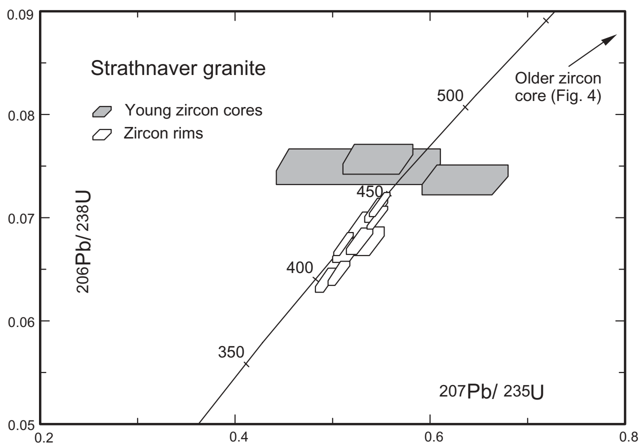

Fig. 7. U–Pb concordia diagram showing the results of SHRIMP analyses of young zircon cores and rims from the Strathnaver granite (1σ error boxes). A plot of analysis 2b, an older inherited core, is given in Figure 4.

closely compare with $^{40}Ar/^{39}Ar$ hornblende and muscovite mineral ages in the age range *c*. 420–410 Ma obtained from the vicinity of the Naver Thrust along the north coast of Sutherland (Dallmeyer *et al*. 2001) and interpreted to record post-metamorphic cooling during exhumation of the nappe pile. The closure temperature of titanite is generally considered to be *c*. 550–500 °C, comparable with that of argon in hornblende (Cliff 1985). Following their syntectonic emplacement and crystallization, the granites are presumed to have cooled rapidly (in probably <1 Ma) to the ambient amphibolite-facies temperatures of the host rocks, and then to have cooled more slowly during regional exhumation (e.g. Tribe & D'Lemos 1996).

Significance of the ages of the zircon cores

The petrogenesis of the granites is a subject of continuing research. The relative proportions of crustal v. lower-crustal or mantle contributions to the melts are at present uncertain. However, the Palaeo- to Mesoproterozic ages obtained from the cores of the zircons compare closely with those obtained from Moine migmatites in Sutherland (Kinny *et al*. 1999), and the West Highland Granite Gneiss in Inverness-shire, which is also thought to have resulted from the melting of Moine sediments (Friend *et al*. 1997; Rogers *et al*. 2001). In both these cases, the zircon cores were interpreted as detrital grains that had been incorporated within the Moine sediments at the time of their deposition. The ages of these zircons were thus interpreted as dating the source rocks for the Moine sediments. The similar age range of zircon cores obtained from the granites of central Sutherland suggests that either the latter were derived at least in part from the melting of Moine metasediments at deeper crustal levels, or that Moine rocks were incorporated into the granitic magma during its ascent through the crust.

The Lower Palaeozoic ages obtained from zircon cores in the Strathnaver and Vagastie Bridge granites are interpreted as follows. The age of 464 ± 26 Ma obtained for zircon cores in the Strathnaver granite is distinctly different from the estimated intrusion age of 429 ± 11 Ma. The core age (although the error is large) is interpreted to date high-grade metamorphism of host Moine gneisses of the Naver nappe during the early to mid-Ordovician Grampian orogeny. It has already been shown elsewhere in Sutherland that this event was associated with migmatization of Moine rocks structurally above the Naver Thrust at *c*. 470–460 Ma (Kinny *et al*. 1999). In the case of the Vagastie Bridge granite, there is overlap between the age of 430 ± 19 Ma obtained from zircon cores and the indicated intrusion age of 424 ± 8 Ma from zircon rims. Further analyses are required to assess whether or not these ages record chronologically separate events.

Regional correlations and implications for the timing of Caledonian events elsewhere in northern Scotland

The mid- to late Silurian age of *c*. 435–420 Ma that has been demonstrated for regional D_2 deformation and metamorphism within the Moine and Naver nappes of central and west Sutherland is essentially the same as that proposed previously for displacements along the Moine Thrust Zone as a result of isotopic age determinations from syntectonic intrusions (Van Breemen *et al*. 1979b) and mylonites (Freeman *et al*. 1998). This lends support to field-based structural models that have viewed the Naver and Moine thrusts and associated fold and thrust structures (e.g. Ben Hope Thrust; Fig. 1a) as part of the same kinematically linked system of foreland-propagating deformation (e.g. Soper & Brown 1971; Soper & Wilkinson 1975; Barr *et al*. 1986; Butler 1986; Holdsworth 1989a; Alsop *et al*. 1996; Holdsworth *et al*. 2002).

Extensive tracts of the western Moine further south (e.g. Fannich, Loch Duich, Fig. 1a) are dominated by north- to NW-trending mineral lineations and associated tight-to-isoclinal, reclined folds that developed during displacement along the Sgurr Beag Thrust (Kelley & Powell 1985; May *et al*. 1993), and are identical to, and structurally continuous with, the regional D_2 structures described in central and north Sutherland (Fig. 8). The Naver Thrust has been generally regarded as the northern extension of the Sgurr Beag Thrust (e.g. Strachan & Holdsworth 1988). In view of these similarities it might therefore be reasonable to infer that much of the western Moine further south also records mid- to late Silurian deformation and metamorphism. However, the prevailing view has been that the Sgurr Beag Thrust and associated structures and amphibolite-facies metamorphic assemblages formed during the early to mid-Ordovician Grampian orogenic event (e.g. Kelley & Powell 1985; Powell &

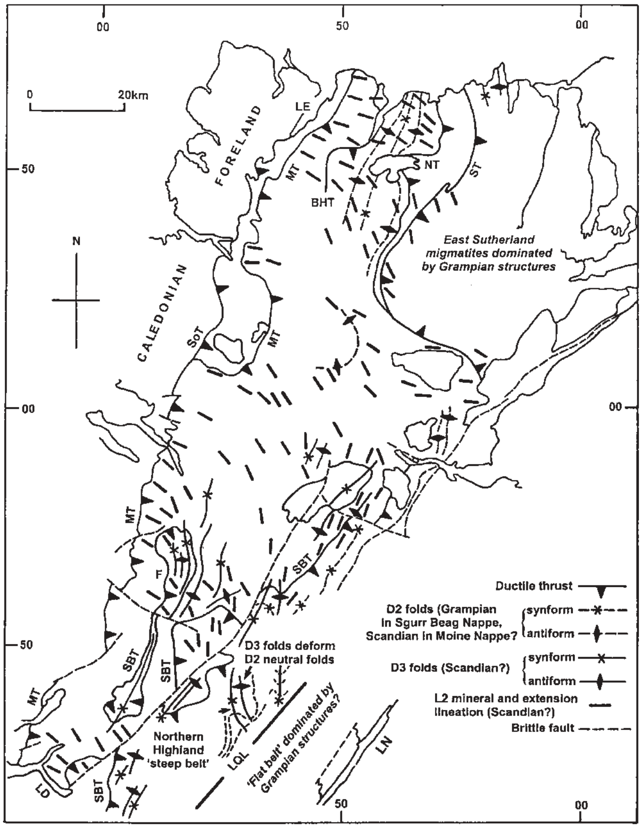

Fig. 8. Caledonian structures within the Moine rocks of Ross-shire and Sutherland, showing structural correlations and probable relative intensities of the Grampian and Scandian orogenic events; the dominance of the Scandian event in the western Moine should be noted (from Strachan *et al*. 2002). F, Fannich; LD, Loch Duich; LE, Loch Eriboll; MT, Moine Thrust; SBT, Sgurr Beag Thrust; NT, Naver Thrust; BHT, Ben Hope Thrust; ST, Swordly Thrust, LQL, Loch Quoich Line, LN, Loch Ness; SoT, Sole Thrust.

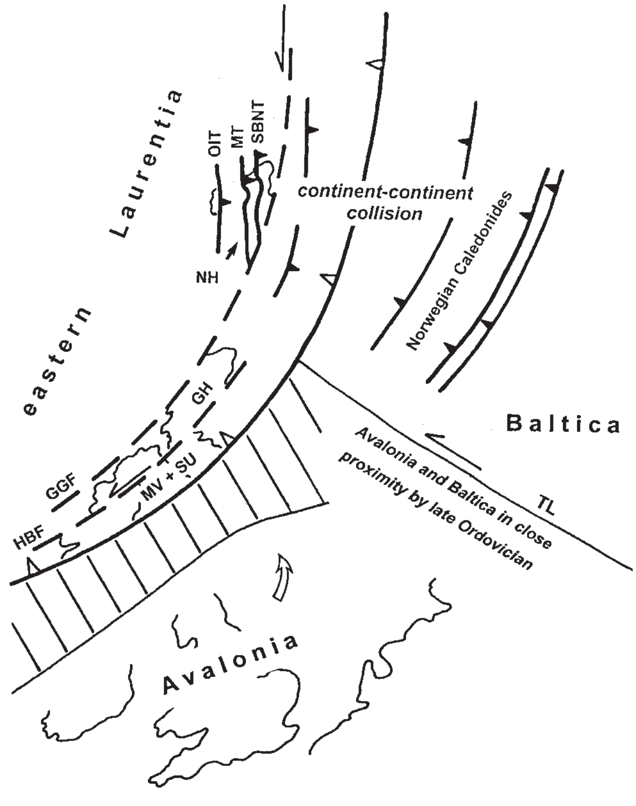

Fig. 9. Schematic plate tectonic framework for the Silurian at *c*. 430 Ma showing how regional thrusting in the Northern Highlands of Scotland (NH) resulted from Laurentia–Baltica collision (from Strachan *et al*. 2002). It should be noted that this Scandian orogenic event does not appear to have affected the Dalradian rocks of the Grampian Highlands (GH). The dashed lines represent major strike-slip faults that developed between the Late Silurian and Early Devonian. Displacement along the Great Glen Fault (GGF) resulted in the juxtaposition of the Grampian and Northern Highlands. OIT, Outer Isles Thrust; MT, Moine Thrust; SBNT, Sgurr Beag–Naver Thrust; HBF, Highland Boundary Fault; TL, Tornquist Line; MV, Midland Valley; SU, Southern Uplands.

Phillips 1985; Harris 1995). According to Powell & Phillips (1985), ductile thrusting was followed by tight upright folding to form the West Highland steep belt, and a prolonged period of post-tectonic cooling from *c*. 453 Ma to *c*. 405 Ma. This interpretation was based mainly on Rb–Sr muscovite ages of *c*. 450 Ma obtained from syn- to post-tectonic pegmatites (Van Breemen *et al*. 1974), a Rb–Sr whole-rock age of 467 ± 20 Ma for a pelite within the Moine nappe and thought to date amphibolite-facies regional metamorphism (Brewer *et al*. 1979), and the view of Powell *et al*. (1981) that displacement along the Sgurr Beag Thrust occurred during the same metamorphic event. Reappraisal of this interpretation is appropriate in the light of not only the new data from Sutherland but also recent isotopic dating south of the Great Glen Fault that has shown the Grampian event to have been short-lived (*c*. 480–465 Ma) and followed by very rapid exhumation (Oliver *et al*. 2000).

Implications for Caledonian tectonic models

It is generally accepted that the Moine Thrust formed during the Scandian orogenic event, which resulted from the collision during the Silurian (*c*. 435–420 Ma) of Baltica and Laurentia (Coward 1990; Fig. 9). The same collision resulted in widespread Barrovian metamorphism, ductile deformation and nappe stacking in East Greenland and Scandinavia (Anderson *et al*. 1992; Dallmeyer *et al*. 1994; Andersen *et al*. 1998; Fig. 9). To date, northern Scotland has seemed anomalous in that the apparent sole effect of this major collision was the formation of the Moine Thrust Zone with no documented isotopic evidence for internal deformation and metamorphism of the Moine rocks at this time. The isotopic evidence presented here from Sutherland for widespread late Silurian deformation and ductile thrusting indicates greater similarities with the Caledonides of East Greenland and Scandinavia than suspected previously. In contrast, recently published isotopic data from the Dalradian Supergroup south of the Great Glen Fault show that the dominant Lower Palaeozoic deformation and metamorphic events occurred during the early to mid-Ordovician Grampian orogeny at *c*. 480–465 Ma (Oliver *et al*. 2000), with no evidence for the late Silurian tectonometamorphic event identified in Sutherland. Evidently, this segment of the orogen must have been located away from, and to the south of, the main collision (Fig. 9). The location within the Scottish Caledonides north of the Great Glen Fault of the eastern limit of late Silurian reworking of Grampian structures and metamorphic fabrics is uncertain at present. It may lie within the poorly known Moine rocks of east and SE Sutherland, or alternatively be defined by a late-orogenic structure such as the Great Glen Fault.

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