Timing, relations and cause of plutonic and volcanic activity of the Siluro-Devonian post-collision magmatic episode in the Grampian Terrane, Scotland

J.C. NEILSON[1,3], B.P. KOKELAAR[1] & Q.G. CROWLEY[2,4]

[1]*Department of Earth and Ocean Sciences, University of Liverpool, Liverpool, L69 3BX, UK*
[2]*NERC Isotope Geosciences Laboratory, Keyworth, Nottingham, NG12 5GG, UK*
[3]*Present address: Dana Petroleum plc, 17 Carden Place, Aberdeen, AB10 1UR, UK*
[4]*Present address: School of Natural Sciences, Department of Geology, Trinity College, Dublin 2, Ireland*

(e-mail: p.kokelaar@liv.ac.uk)

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Abstract: Calc-alkaline magmatism in the Grampian Terrane started at ~430 Ma, after subduction of the edge of continental Avalonia beneath Laurentia, and it then persisted for ≥22 million years. ID-TIMS U-Pb zircon dating yields 425.0 ± 0.7 Ma for the Lorn Lava Pile, 422.5 ± 0.5 Ma for Rannoch Moor Pluton, 419.6 ± 5.4 Ma for a Fault-intrusion at Glencoe volcano, 417.9 ± 0.9 Ma for Clach Leathad Pluton in Glencoe, and in the Etive Pluton 414.9 ± 0.7 Ma for the Cruachan Intrusion and 408.0 ± 0.5 Ma for the Inner Starav Intrusion. The Etive Dyke Swarm was mostly emplaced during 418-414 Ma, forming part of the plumbing of a large volcano (≥2000 km^3) that became intruded by the Etive Pluton and was subsequently removed by erosion. During the magmatism large volumes (thousands of km^3) of high Ba-Sr andesite and dacite were erupted repeatedly, but were mostly removed by contemporaneous uplift and erosion. This volcanic counterpart to the 'Newer Granite' plutons has not previously been fully recognised. The intermediate magmas forming both plutons and volcanoes originated mainly by partial melting of heterogeneous mafic-to-intermediate lowermost crust that had high Ba-Sr derived from previous melting of LILE-enriched mantle, possibly at ~1.8 Ga. This crustal recycling was induced by heat and volatiles from underplated small-degree melts of LILE- and LREE-enriched lithospheric mantle (appinite-lamprophyre magmas). The post-collision magmatism and uplift resulted from breakoff of subducted oceanic lithosphere and consequent rise of asthenosphere.

Supplementary material: Geochronological methods, tabulated data and figures are available at http://www.geolsoc.org.uk/SUP00000.

Siluro-Devonian plutons populate a broad belt that extends ~300 km NE-SW across the Scottish Highlands, on either side of the Great Glen Fault. When intrusions of similar age and setting in Shetland and Donegal are included (Fig. 1), the belt extends to 750 km. The intrusions are the 'Newer Granites' of Read (1961) and are dominated by the compositionally distinct Argyll and Northern Highland Suite (high Ba and Sr, abundant appinites) and the Cairngorm Suite (relatively low Ba and Sr, few appinites) (Stephens and Halliday 1984; Halliday et al. 1985; Tarney and Jones 1994; Highton 1999; Fowler et al. 2001). 'Newer Granites' in the Southern Uplands and in England, comprising the Trans-Suture Suite of Brown et al. (2008), are younger than, and unrelated to, those further north. Owing to contemporary uplift and erosion, remnants of Siluro-Devonian volcanic rocks in the Highlands plutonic belt are relatively few and mostly preserved south of the Great Glen Fault, in the 'Grampian Terrane' of Gibbons and Gayer (1985). These include the centred or caldera volcanoes at Glencoe (Kokelaar and Moore 2006), Ben Nevis (Bailey 1960) and Etive (Anderson 1937), a thick and extensive sills-plus-lavas pile in Lorn (Kynaston and Hill 1908), and lavas, tuffs and fossiliferous hydrothermal deposits at Rhynie (Trewin and Thirlwall 2002 and references therein). Small outliers of lavas and tuffs in the north, southeast and south of the Grampian Terrane (Stephenson and Gould 1995, figure 26) indicate originally extensive volcanic cover(s) on a broadly low-relief basement.

Northwest of the Great Glen Fault, some plutons older than ~425 Ma were involved in late Scandian deformation: e.g. the 426 ± 2 Ma Strath Halladale Granite in the Northern Highland Moines (Fig. 1; Kocks et al. 2006). Southeast of the Great Glen Fault, in the Grampian Terrane, the intrusions and volcanic rocks are not pervasively deformed, but are faulted and tilted; they respectively cut and unconformably overlie Dalradian metamorphic basement. There seems consensus that sinistral strike-slip movement on the Great Glen Fault, conceivably a minimum of 700 km, occurred from ~428-410 Ma (e.g. Stewart et al. 1999, 2001; Dewey and Strachan 2005) and that this resulted in the juxtaposition of Scandian-deformed intrusions northwest of the fault with rocks showing no such deformation to the southeast. In the main Grampian belt there was no crustal thickening or tectonic overprint due to the $\geq$435-425 Ma Scandian orogeny, although there was uplift.

Fig. 1

The petrogenesis of the 'Newer Granites' of the Highlands, their closely associated appinites and lamprophyres, and the coeval hypabyssal and volcanic rocks, has generally, but not universally, been considered to have involved hydrous mantle melting above an *active* subduction zone that dipped beneath the Laurentian continental margin (Dewey 1971; Brown 1979; van Breemen and Bluck 1981; Soper 1986; Thirlwall 1988; Fowler et al. 2001; Oliver 2001; Woodcock and Strachan 2002; *cf*. Halliday and Stephens 1984). Some envisage an active continental destructive plate margin analogous to the Andean Pacific margin (e.g. Oliver et al. 2008). Pitcher (1982, 1987) and Watson (1984) believed that the 'Newer Granites' post-dated plate collision – in the Southern Uplands they effectively stitch the Iapetus suture – and they advocated a generalised origin involving decompression related to uplift and strike-slip pull-apart. Pitcher (1982) emphasised that the Caledonian granitoids were unlike those of the Andes, in both setting and composition.

The 'Newer Granites' of the Grampian Highlands are classified as I-type and high-K calc-alkaline (Halliday and Stephens 1984), with enriched-mantle signatures and with $^{87}Sr/^{86}Sr$ initial ratios <0.708 (average 0.706). There has long been consensus that the parental magmas were derived mainly from partial melting of metasomatised lithospheric mantle and lower crust (Thirlwall 1982; Halliday 1984; Stephens and Halliday 1984; Halliday et al. 1985; Tarney and Jones 1994; Fowler et al. 2001). This contrasts with earlier intrusions, ≥ 434 Ma, which are of S-type, with relatively high $^{87}Sr/^{86}Sr$ initial ratios >0.712 (average 0.716), and are considered (e.g. Oliver 2001) to result from uplift-decompression melting of crust (uplift ≥ 15-20 km) following the Grampian orogeny.

Refinements in spatial and temporal reconstructions of the Caledonides (e.g. Stone et al. 1987; Kneller 1991; Kneller et al. 1993; Dewey and Strachan 2003, 2005) cast doubt that the Grampian magmatism was directly related to active subduction. In this paper we distinguish a continuous ≥ 22 million-year magmatic episode that *started* just when subduction would have been impeded through plate collision(s), which contrasts with the discontinuous and protracted, syn-subduction assembly of the Andean batholith (e.g. Pitcher 1978). We re-emphasise the point made by Pitcher (1982) that the Caledonian plutons are unlike those of the Andean Cordillera; he referred to the Grampian and Donegal plutons as 'I- (Caledonian) type' in recognition of the dissimilarity. In a multi-element (R1-R2) tectono-magmatic discriminant diagram (Fig. 2a), the Grampian rocks do not plot with the Cordilleran I-type plutons in the 'pre-plate collision' field, but in the 'post-collision uplift' and 'late orogenic' fields. Also unlike the Andean magmatism (Pitcher et al. 1985), there was no precursor extensional basin, no prolific extrusion of basalt, and no early plutonic phase of gabbroic intrusions; a distinctive feature of the Grampian rocks is the paucity of basalts (Figs 2b and c) and gabbros (Pitcher 1982). The lamprophyres and appinites, however, mainly plot in the 'pre-plate collision' field.

Fig. 2.

Relationships between the Siluro-Devonian volcanoes, hypabyssal intrusions and plutons are uniquely well displayed in the southwest Grampian Highlands (Fig. 3), where cross-cutting relationships indicate protracted magmatism and progressive assembly of a batholith. We take the Strath Ossian, Rannoch, Clach Leathad and Etive plutons (Fig. 3) to constitute parts of a small batholith, named here Lochaber Batholith. Rannoch and Clach Leathad are both more extensive at depth than at outcrop (Kokelaar and Moore 2006; this paper) and the plutons combine at depth to form a continuous body with a southwesterly age progression. The plutons belong to the Argyll and Northern Highland Suite (Stephens and Halliday 1984), although they are not entirely of high Ba-Sr type (Fig. 2d). The extensive Etive Dyke Swarm, dominantly of intermediate to silicic compositions (Fig. 2c), but including lamprophyres, is related to a former centred volcano that was largely obliterated by emplacement of the Etive Pluton and subsequent erosion (see below). The rocks considered in this study are predominantly medium- or high-K calc-alkaline and include monzodiorite-granodiorite-monzogranite-granite plutons, small-volume appinite-lamprophyre intrusions, and volcanic suites of hypabyssal to extrusive trachybasalt, basaltic trachyandesite, trachyandesite, trachyte and rhyolite, with relatively sparse basalt, basaltic andesite and dacite (Figs 2b and c; after Le Maitre 1989); lavas and intrusions with shoshonitic affinity also occur. The prefix trachy- is mostly dropped, for brevity. The Dalradian metamorphic basement in this region is underlain by Proterozoic gneisses and mafic granulites, forming continental crust ~35 km thick. Basement rocks exposed to the west, on Islay and Colonsay (Rhinns Complex; Fig. 1), as well as magmatic rocks across the Grampian Terrane, indicate a mafic crustal component with subduction-arc affinities derived from the mantle at ~1.8 Ga (Halliday et al. 1985; Marcantonio et al. 1988; Dickin and Bowes 1991; Muir et al. 1992, 1994). Plutons widely across the region north of the Highland Boundary Fault have long been known to carry inherited zircons with ages of ~1.6 Ga (Pidgeon and Aftalion 1978). Nd, Sr and Pb isotope studies of the southwest Grampian rocks show that the underlying lithospheric mantle is heterogeneous: probably layered with tectonic interleaving and with variable metasomatism (Thirlwall 1982, 1986; Macdonald and Fettes 2007; Neilson 2008).

Fig. 3

This paper aims to establish the timing and duration of the Grampian volcanic and plutonic activity in the context of its tectonic setting, to constrain possible models for the petrogenesis. Recent studies have considered a discrete pulse of plutonism, based on poorly constrained age data (e.g. Atherton and Ghani 2002), or overlooked the evidence for a probable block to subduction just when the I-type magmatism was initiated (e.g. Oliver 2001; Oliver et al. 2008). We establish the importance of large volcanoes and hence previously unrecognised large volumes of intermediate magmas that bear on the interpretation of petrogenesis. Any petrogenetic model devised to account for the 'Newer Granites' throughout the magmatic belt (excluding the Southern Uplands and England), must also explain the generation of very large volumes of andesite.

Temporal constraints on the cause of the magmatism

The main problem with linking the magmatism to subduction in the conventional sense, i.e. with hydrous mantle melting due to dehydration of a down-going oceanic-lithosphere slab, is in their relative timing. Here we review the subduction history, before showing that magmatism in the southwest Grampian region was continuous for at least 22 million years *after* continental collision. Northwards subduction of Iapetus oceanic lithosphere began after the extended passive margin of Laurentia collided with a substantial intra-Iapetus island arc during the early-mid Ordovician (Dewey and Shackleton 1984; Dewey 2005). This collision was the Grampian event, when the Dalradian strata were tectonically stacked and metamorphosed, between 478 Ma and 460 Ma. It involved a subduction-polarity reversal, from southwards beneath the arc to northwards beneath it, towards Laurentia, and it resulted in the building of the Southern Uplands tectono-sedimentary complex (Leggett et al. 1979; Kelling et al. 1987; Stone et al. 1987; Soper et al. 1999; Oliver 2001; Kinny et al. 2003; Dewey and Strachan 2003, 2005; Dewey 2005).

There is no record in the Grampian Terrane of subduction-related magmatism *during* closure of the Iapetus Ocean. The terrane would have been several hundred kilometres north of the subduction-trench while the contemporary arcs and associated back-arc and fore-arc basins formed nearer the ocean, in the position of the Midland Valley Terrane and further oceanwards, now buried beneath thrusted and back-thrusted Southern Uplands strata (Stone et al. 1987; Heinz and Loeschke 1988; Bluck 1983, 2001).

Southwards subduction of Iapetus Ocean lithosphere beneath continental Avalonia ceased in Caradoc times (~450 Ma). By late Llandovery times, ~430 Ma, the edge of Avalonia had been subducted beneath the edge of Laurentia so that the Lake District Terrane became the site of a foreland basin (Stone et al. 1987; Freeman et al. 1988; Kneller 1991; Kneller et al. 1993). At this time, due to buoyancy constraints, continued subduction of continental Avalonia would have been resisted; by ~428 ± 1.9 Ma sinistral strike-slip movement occurred along the Great Glen Fault (recorded by the Clunes Tonalite; see Stewart et al. 2001). By mid-Wenlock times (~426 Ma) subduction had slowed (Kemp 1987; Kneller 1991) and, in the late Wenlock, sandstone turbidites of the *M. lundgreni* Biozone (Kemp 1987) were the last added to the Southern Uplands complex. At the same time, detritus reworked from the emergent complex extended far into the foreland basin, across the Iapetus Suture (Furness 1965; Kemp 1987; Soper and Woodcock 1990; Kneller et al. 1993). By the end of the Wenlock (~423 Ma), orthogonal shortening of the complex had switched to sinistral transpression and a southeast-vergent fold-and-thrust belt developed in the northern side of the foreland basin (Stone et al. 1987; Stone 1995). The thrust belt and foreland basin migrated south through the late Silurian, with syndepositional shortening during late Ludlow – Pridoli times, ~420-416 Ma. This migration has been related to evolution of a northwest dipping mid-crustal thrust with southward translation of an emergent (Acadian) mountain belt (Kneller and Bell 1993; Kneller et al. 1993). We conclude that by ~430 Ma the supply of hydrated oceanic lithosphere *into* the subduction zone had stopped and that further subduction of continental crust was resisted. This was ~5 million years after 'hard' continental collision between Baltica and Laurentia caused thrust and nappe deformation in Scandinavia, Greenland and northwest Scotland (the Scandian orogenic event; Soper et al. 1992). We show below that I-type magmatism in the Grampian Terrane began at ~430 Ma, just when subduction was resisted, and then continued for ≥22 million years. Magmatism directly related to subduction might have persisted for a few million years after 430 Ma, but other processes must be invoked to explain the intense post-collision magmatism that lasted at least 22 million years and produced a broad, linear belt of plutons and volcanoes where none had formed previously. We agree with Atherton and Ghani (2002) that an alternative to active subduction is required.

Uranium-Lead (U-Pb) zircon dating

U-Pb zircon dating was undertaken to facilitate understanding of the duration, timing and rates of activity in the Siluro-Devonian magmatic episode. Previous studies have mostly been limited and variable with regard to both the magmatic bodies studied and the precision of the dating techniques. Thirlwall (1988) has provided the best available regional synthesis of the timing of what he referred to as 'Late Caledonian' magmatic activity, including Shetland, the Northern and Grampian Highlands, the Midland Valley and the Southern Uplands; he considered all except that in the latter terrane to be related to active subduction beneath Laurentia between 425 Ma and 408 Ma. Here we focus closely on a single element of that large area, with the benefit of new plate-tectonic reconstructions and more precise dating techniques. The chosen field area is where *relative* age relationships between various magmatic centres are uniquely well constrained. Samples of one lava and five intrusions were selected: Lorn Lava Pile, Rannoch Moor Pluton, Glencoe Fault-intrusion, Clach Leathad Pluton, and the Cruachan and Inner Starav intrusions of the Etive Pluton (Fig. 3). The Clach Leathad Pluton intrudes the Glencoe caldera volcano and has only recently been recognised as separate from, and older than, the Cruachan Intrusion of the Etive Pluton to the south (Kokelaar and Moore 2006; see below). The relationships indicate that the Rannoch pluton is the oldest intrusion and Inner Starav the youngest; the age relations of the Lorn Lava Pile to Glencoe and Rannoch were unknown. Appinites and the Ballachulish Pluton have already been dated using U-Pb techniques, at ~429-427 Ma (Rogers and Dunning 1991; Fraser et al. 2004).

Analytical Techniques

Zircon fractions were analysed by Isotope Dilution Thermal Ionization Mass Spectrometry (ID-TIMS) at the NERC Isotope Geosciences Laboratory (NIGL), Nottingham, UK and at the Jack Scatterly Geochronology Laboratory, University of Toronto. A full account of analytical protocols is given in the online supplementary publication: http://www.geolsoc.org.uk/SUPXXXXX.

New U-Pb zircon ages

Thirty-four U-Pb zircon TIMS analyses were made on the six samples and are presented in supplementary Table 1, with concordia diagrams in Figure 4 and final results in Figure 5. Sample sites are located with 8-figure references to the Ordnance Survey National Grid 100-km squares NM and NN. The geological context and immediate implications of the new dates are discussed with each result; the broader implications are considered afterwards. The new dates are entirely consistent with the relative age relationships known from fieldwork.

Fig. 4

Fig. 5

Top of the Lorn Lava Pile 425.0 ± 0.7 Ma. Sample JNL86 is from the uppermost dacite (Carn Gaibhre, NM 9740 2595; Fig. 5), near the top of the 600 m-thick lava pile. It is purple and fine grained, with sparse phenocrysts of plagioclase and hornblende, altered to sericite and chlorite respectively, and with minor apatite, opaque oxides, zircon and titanite. The zircons picked for geochronology were acicular to prismatic (200x70x70 µm to 120x50x50 µm), colourless and translucent, commonly with magmatic c-axis inclusions (Supplementary Fig. 1a). A range of zircon morphologies and sizes exists in this sample, including crystals with visible cores (Supplementary Fig. 1b). Due to the small size of acicular zircons that were picked, to avoid inherited cores, multi-grain fractions were generally used. Z-2 and Z-6 yield a concordia age of 425.0 ± 0.7 (concordance and equivalence MSWD 1.9, probability 0.13) and mean $^{206}Pb/^{238}U$ age of 425.0 ± 0.6 Ma (MSWD 2.5, probability 0.11), while Z-1 and Z-5 yield an older concordia age of 430.6 ± 1.3 Ma. Regressing all four gives a lower intercept age of 425.5 ± 4.7 Ma. The 425.0 ± 0.7 Ma age is considered to represent the true eruption age, with the apparently older zircons involving an inherited component. It is noteworthy that although an apparent age range of ~5 million years occurs in these Lorn dacite zircons, the analyses are concordant suggesting that individual fractions either represent single-age populations, or at least populations so close in age that the analyses are not discordant. The former is confirmed by Cathode Luminescence (CL) imaging (Supplementary Fig. 1a); the analysed acicular zircons with c-axis melt inclusions do not have any inherited cores, nor do they show any resorption textures. Accumulation of the Lorn lava pile would have taken ≤1 million years, consequently the onset of the eruptions is placed at no more than ~426 Ma; there are no known significant unconformities within the pile. Previous (Rb-Sr) ages for Lorn include a lava at 423.5 ± 6.1 Ma and, from the Ferry Plug intrusion southwest of Oban (Fig. 5), a range from 413.6 ± 6 to 426.9 ± 5 Ma (Thirlwall 1988).

The Lorn Lava Pile is underlain by conglomerates that rest unconformably on the Dalradian. They contain abundant clasts of quartzite, boulders of andesite (≤2 m) and biotite granite (≤1 m), and cobbles or pebbles of quartz-porphyry, granodiorite and granite. The clasts suggest that either there was earlier andesite volcanism somewhere in the palaeo-Great Glen drainage system to the northeast (palaeocurrents were towards southwest; Morton 1979; Neilson 2008), or the andesitic boulders were derived from the developing Lorn pile, having been reworked into a canyon that cut down to the Dalradian basement. The large size of the boulders suggests either that they were locally derived or that the fluvial drainage was very steep. The palaeocanyon or valley floor where the boulder conglomerates are preserved shows no evidence for steepness; associated sandstones and siltstones record tranquil overbank and shallow lacustrine sedimentation. The provenance of the silicic igneous clasts is uncertain, but plutons were emplaced in the general vicinity, up-palaeovalley, and these may have been unroofed and eroded before the influx of sediments: e.g. Clunes Tonalite 427.8 ± 1.9 Ma (Stewart et al. 2001) and Ballachulish Igneous Complex 427 ± 1 Ma (Fraser et al. 2004). At the time of emplacement of the Ballachulish Pluton there was 9-10 km of cover above the level of the present outcrop (Pattison and Harte 1997), but a shallow upper part may have been exhumed within as little as ~2 million years (e.g. Harayama 1992), as later with the Rannoch Moor Pluton (see below). The appinites at Garabal Hill (Nockolds 1941) and at Rubha Mhor near Ballachulish (Bowes and Wright 1967) had also been emplaced (~429-427 Ma; Rogers and Dunning 1991), but no such rock has been recognised in the Lorn conglomerates. The boulders show that pluton unroofing, and possibly also andesitic eruptions, predated 426-425 Ma. The interval required for (rapid) unroofing places the age of plutonic intrusion at no younger than 428-427 Ma, consistent with Ballachulish as a potential source.

Two silicic ignimbrites within the Lorn Lava Pile, east of Benderloch (NM 91 37), were tentatively linked by Roberts (1963) to eruptions from the Glencoe caldera volcano. However, explosive silicic eruptions there started some 5 million years after accumulation of the Lorn pile (~420 Ma; see below) and there is no obvious source for the ignimbrites. As strike-slip on the Great Glen Fault continued long after emplacement of the Lorn pile, it is probable that some substantial counterpart of the pile northwest of the fault has been removed. The minor pipe-like intrusions on southern Kerrera and to the south of Oban (Fig. 5) are not major feeders of the Lorn pile and here the lowermost sheets are sills that were emplaced into unlithified wet sediments (Neilson 2008). The Kilmelford intrusive suite, south of the Lorn outcrop (Fig. 3), is dominated by biotite-hornblende diorites and granodiorite, and has shoshonitic affinity (Zhou 1987) similar to Lorn (see below). Conceivably Kilmelford could represent a plutonic counterpart to the Lorn extrusive pile, as suggested by Tarney and Jones (1994).

The sedimentary strata beneath the Lorn lavas and sills around Oban and on Kerrera have been renowned for their fossils of fish, eurypterids, millipedes, ostracods, plant remains and sporomorphs, which have been assigned to an uppermost Silurian or lowermost Devonian age (Lee and Bailey 1925; Morton 1979; Marshall 1991; Wellman 1994; Wellman and Richardson 1996). The new radiometric age for the Lorn lavas suggests that their sedimentary substrate is of an older Silurian age than originally assigned, implying earlier occurrences for the flora and fauna; 426-425 Ma is in the Homerian Stage of the Wenlock (Gradstein et al. 2004).

Rannoch Moor Pluton 422.5 ± 0.5 Ma. Sample JNGC88 is a foliated granodiorite from the main body of the pluton (G2 of Jacques and Reavy 1994), 1.5 km from the outer contact (NN 2610 5570; Fig. 5). It contains hornblende and biotite, with minor apatite, titanite, opaque oxides and zircon. The analysed zircons were single-grain fractions, acicular or squat (250x90x90 μm to 300x20x80 μm), colourless and translucent. Four analyses [Z-5, Z-6, Z-8 and Z-9] define a concordia age of 422.5 ± 0.5 Ma (concordance and equivalence MSWD 0.88, probability 0.52) and a mean $^{206}Pb/^{238}U$ age of 422.7 ± 0.4 Ma (MSWD 0.67, probability of fit 0.57); the consistent data are taken as the true crystallization age.

The Rannoch crystallisation age of ~422.5 Ma post-dates the Lorn eruptions by some 2.5 million years and pre-dates the onset of centred volcanism at Glencoe. Boulders of foliated Rannoch Moor granodiorite in conglomerates beneath the basal ignimbrite at Glencoe show that the Rannoch pluton was unroofed before the onset of Glencoe caldera volcanism (Kokelaar and Moore 2006). Older conglomerates at Glencoe, beneath the Basal Andesite Sill-complex, contain granite, granodiorite and 'basic plutonic rock resembling kentallenite' (Bailey and Maufe 1916), and, on the reasonable assumption that fluvial drainage was from the east or southeast along the Glencoe Lineament (Kokelaar and Moore 2006), the Rannoch Moor Pluton is the best contender as their source. The boulders would have been derived from levels of the pluton shallower than those presently exposed. Assuming a minimum depth of some 1.5-2 km to the top of the pluton and 2 million years as the minimum time for unroofing (Harayama 1992), the conglomerates could only have been deposited at or after 420.5 Ma. This is the probable maximum age of the andesite sills and certainly the maximum age of the (somewhat later) onset of the caldera volcanism. Flora and sporomorphs (Kidston and Lang 1924; Wellman 1994) from siltstones overlying the conglomerates were considered to be early Devonian (late-early or early-late Lochkovian) in age, which, from the timescales of Gradstein et al. (2004) and Kaufmann (2006), would be ~415 Ma. However, 415 Ma is the age obtained for the Cruachan Intrusion of the Etive Pluton (Fig. 5), which post-dates both the Clach Leathad Pluton (~418 Ma) and the Glencoe volcano (~419 Ma). Considering the radiometric ages from Rannoch and Glencoe, with the time required to unroof Rannoch and derive the boulders, a ~420 Ma age for the fossiliferous siltstones and onset of centred volcanism at Glencoe is indicated. 420 Ma is in the Ludfordian Stage (Ludlow) of the Silurian (Gradstein et al. 2004).

Glencoe Fault-intrusion 419.6 ± 5.4 Ma. Sample JNGC76 is from the Gleann Charnan Fault-intrusion in the southwest of the Glencoe Caldera-volcano Complex (NN 1354 5103; Fig. 5). It is a hornblende-biotite quartz diorite, with minor apatite, titanite, opaque oxides and zircon. The zircons are typically acicular (120x90x50 μm to 100x80x60 μm), colourless and translucent. With the exception of Z-8, U-Pb data for this sample plot on a Pb-loss trajectory from ~419 Ma; regression through the data yields an upper intercept age of 419.6 ± 5.4 Ma (MSWD 1.01). Glencoe Fault-intrusions widely show evidence of hydrothermal alteration (e.g. Garnham 1988; Kokelaar 2007), which may have caused extensive Pb loss.

This intrusion was selected for analysis because it is one of the few with an inferred extrusive counterpart within the Glencoe Volcanic Formation. It probably represents a feeder of the Bidean nam Bian Andesite Member (Kokelaar and Moore 2006; Neilson 2008). This member lies close to the top of the preserved volcanic pile and records a *single* intra-caldera eruption of compositionally distinct batches of andesitic and dacitic magmas totalling ≥12 km^3. Thirlwall (1988) obtained a $^{40}Ar/^{39}Ar$ plateau age of 421 ± 4 Ma from hornblende separated from this member. On the basis of the age of the onset of caldera volcanism, at ~420 Ma (see above), the upper intercept age of 419.6 ± 5.4 Ma is regarded as the emplacement age for the Gleann Charnan Fault-intrusion. Z-8 gives a $^{206}Pb/^{238}U$ age of 421.4 ± 0.7 Ma, which is interpreted as recording inheritance, because it is highly unlikely that unroofing of the Rannoch Moor Pluton and subsequent development of much of the Glencoe caldera-volcano succession could have occurred in ~1 million years. Furthermore, the intercept age of ~419 Ma is more readily reconciled with the ~418 Ma of the Clach Leathad Pluton (Fig. 5). This pluton intrudes the youngest preserved part of the Glencoe Volcanic Formation and is not chilled in contacts with a (different) fault-intrusion, in the southeast. The evolution of the Basal Andesite Sill-complex and centred caldera volcano at Glencoe, up to the time of its intrusion by the Clach Leathad Pluton, occurred within ~2.5 million years. The unconformity between the Basal Andesite Sill-complex and the overlying ignimbrite (Lower Etive Rhyolite) represents approximately 0.5 million years (Kokelaar and Moore 2006), so that caldera volcanism up to emplacement of the Clach Leathad Pluton lasted a maximum of 2 million years. During this time there were five large effusions of andesite-dacite and at least seven caldera-forming eruptions of rhyodacite-rhyolite (Moore and Kokelaar 1998).

Clach Leathad Pluton 417.9 ± 0.9 Ma. Sample JNE5 was collected from the summit of Clach Leathad (1099 m; NN 2470 4910; Fig. 5) in the main body of the pluton. It is a pink and grey, hornblende-biotite monzogranite, containing minor apatite, titanite, opaque oxides and zircon. The zircons display various morphologies, but those analysed were either acicular or squat (220x80x60 μm to 280x120x100 μm), colourless and translucent. Five analyses [Z-27, Z-28, Z-29, Z-30 and Z-31] yield a concordia age of 417.9 ± 0.9 Ma (concordance and equivalence MSWD 2.7, probability 0.01) and a mean $^{206}Pb/^{238}U$ age of 417.8 ± 0.9 Ma (MSWD 3.5, probability of fit 0.01), interpreted to represent the time of crystallization.

This date of ~418 Ma is the accepted age of the Silurian-Devonian boundary (Kaufmann 2006) and is when magmatic activity centred at Glencoe ceased. It marks the upper age limit of the subsequent protracted emplacement of the Etive Dyke Swarm. Mineral alignment fabrics in the Clach Leathad monzogranite adjacent to the earliest Etive dykes show that the latter were emplaced *during* advanced solidification of the pluton, before its full crystallisation (Jacques 1995, p. 356).

The Clach Leathad Pluton was described (e.g. Bailey 1960) as a northern lobe of the Cruachan Intrusion, and hence as part of the Etive Pluton to the south. However, field relationships between the plutonic intrusions and with dykes of the Etive swarm led Kokelaar and Moore (2006) to propose that the Clach Leathad Pluton is a separate intrusion, centred at Glencoe and substantially older than the Cruachan Intrusion. This reinterpretation is justified by the Clach Leathad Pluton proving to be ~3 million years older than the Cruachan Intrusion (Fig. 5). The intervening time was when the centred Etive Volcano developed (see below).

Cruachan Intrusion (Etive Pluton) 414.9 ± 0.7 Ma. Sample JNE8 is from the main body of the intrusion, 800 m from its outer contact, in the region of Glen Ure (NN 0710 4740; Fig. 5). It is a pale grey hornblende-biotite granodiorite with minor apatite, titanite, opaque oxides and zircon. The analysed zircons were either acicular or squat (250x60x50 μm to 310x110x80 μm), colourless and translucent. Four analyses [Z-33, Z-36, Z-37 and Z-38] yield a concordia age of 414.9 ± 0.7 Ma (concordance and equivalence MSWD 2.6, probability 0.012) and a mean $^{206}Pb/^{238}U$ age of 414.7 ± 0.7 Ma (MSWD 3.9, probability of fit 0.01), with the data taken to be the best estimate of the true crystallization age. Two further [Z-34 and Z-35] analyses are discordant due to Pb loss.

Inner Starav Intrusion (Etive Pluton) 408.0 ± 0.5 Ma. Sample JNGC87 is from near the centre of the intrusion, in the vicinity of Ben Starav, >1 km from the contact with the Porphyritic Outer Starav Intrusion (NN 1375 4310; Fig. 5). It is a biotite monzogranite, with minor apatite, titanite, opaque oxides and zircon. The analysed zircons were either acicular or squat (240x90x90 μm to 150x90x80 μm), colourless and translucent. Three analyses [Z-10, Z-41 and Z-42] yield a concordia age of 408.0 ± 0.5 Ma (concordance and equivalence MSWD 0.12, probability 0.99) and a mean $^{206}Pb/^{238}U$ age of 408.0 ± 0.3 Ma (MSWD 0.11, probability of fit 0.89), taken as the crystallization age. One fraction [Z-4] is discordant due to minor-Pb loss while another [Z-5] indicates some inheritance.

The duration of assembly of the Etive Pluton is defined by the age of the Inner Starav Intrusion at 408.0 ± 0.5 Ma and some age between those of the Clach Leathad Pluton, 417.9 ± 0.9 Ma, and the Cruachan Intrusion, 414.9 ± 0.7 Ma. The earliest component of the Etive Pluton to cut the Clach Leathad Pluton is the Meall Odhar Intrusion (Figs 5 and 6; Kokelaar and Moore 2006) against which the Cruachan Intrusion shows no chilled margin. Consequently the age for the Meall Odhar Intrusion is no older than ~416 Ma. Thus the Etive Pluton was assembled in ~8 million years. During this time the Lintrathen ignimbrite was emplaced in the southeast of the Grampian Terrane, at ~415 Ma, and volcanic activity with hydrothermal mineralisation formed the renowned plant-bearing Rhynie chert, at ~412 Ma (Trewin and Thirlwall 2002; S. Parry pers. comm. 2008).

Fig. 6

The Inner Starav Intrusion was the last major body of magma of the southwest Grampian suite emplaced in this region. 408 Ma is in the late Pragian (Gradstein et al. 2004) or early Emsian stage (Kaufmann 2006) of the Lower Devonian, after which there was folding in the Midland Valley and development of a widespread unconformity in Scotland. The magmatism of the Trans-Suture Suite of 'Newer Granites' in the Southern Uplands was initiated at ~400 Ma (Brown et al. 2008), at least 5 million years after the main activity in the southwest Highlands had ended.

New ages recently published by Oliver et al. (2008) and concerning rocks that we have analysed are in conflict with the geological field relationships. Their data, derived from sensitive high-resolution ion microprobe (SHRIMP) U-Pb analyses of zircons, place Rannoch Moor Pluton at 409 ± 8 Ma (in contrast to our 422.5 ± 0.5), which would make it one of the youngest elements of the southwest Grampian suite, despite the fact that Rannoch is cut by Etive dykes that predate the Outer Starav Intrusion of the Etive Pluton, which they have dated at 415 ± 6 Ma. Also, ages published by other authors and utilised by Oliver et al. (2008), for example the 406 ± 6 and 398 ± 2 Ma respectively for rhyolite and andesite at Glencoe (Fraser et al. 2004), are similarly impossible to reconcile with the geological field evidence and with our new results (Glencoe ~420.5-418 Ma; see above).

Nature and timing of appinite-lamprophyre magmatism

Appinites and high-K calc-alkaline lamprophyres, respectively coarse-grained and porphyritic varieties of incompatible-element-rich rocks with abundant hydrous mafic phases and commonly some K-feldspar, ultimately derive from small-degree partial melts of enriched (K-amphibole and/or phlogopite-bearing) sub-continental lithospheric mantle (e.g. Rock 1991; Fowler and Henney 1996). That kentallenite (phlogopite and K-feldspar-bearing picrite), appinite and lamprophyre intrusions are genetically related was first recognised in the southwest Grampian region, by Hill and Kynaston (1900). The high contents of incompatible trace elements (especially Ba and Sr) in these rocks, and in some mafic enclaves and dioritic components in plutons belonging to the Argyll and Northern Highland Suite, have long been taken to indicate a petrogenetic link (e.g. Stephens and Halliday 1984; Macdonald et al. 1986; Rock and Hunter 1987; Atherton and Ghani 2002). Several authors have suggested that high Ba-Sr 'Newer Granites' may derive from the mantle-derived melts primarily by fractional crystallisation with crustal assimilation (e.g. Rock and Hunter 1987; Fowler et al. 2001). That syenite and granite can derive from fractional crystallisation of lamprophyric magma has been demonstrated in a dyke (Macdonald et al. 1986) and interpreted in hybrid appinite-syenite-granite complexes (Fowler 1988; Fowler and Henney 1996).

The southwest Grampian 'Appinite Suite' was originally considered to have been emplaced during a relatively short time, generally before intrusion of the 'late Caledonian' granites (Read 1961; Wright and Bowes 1979; Rogers and Dunning 1991). More widely in the magmatic belt, the appinites are thought to have been intruded before or during emplacement of the granites with which they are (physically) associated (e.g. Pitcher and Berger 1972; Fowler et al. 2001; Atherton and Ghani 2002; Macdonald and Fettes 2007). In the southwest Grampian region we find that small-degree mantle melts that would form appinite or lamprophyre were available through much of the ≥22 million-year magmatic episode, although seemingly their arrival in the upper crust was most common during early stages.

The oldest U-Pb zircon age for magmatic crystallisation in the 'post-collision' rocks of the southwest Grampian region is 429 ± 2 Ma (Fig. 5; Rogers and Dunning 1991), from Garabal Hill appinitic diorite in the Glen Fyne igneous complex (Nockolds 1941). This age was derived from three discordant zircon analyses and the ± 2 Ma error seems low. Re-evaluation of the data gives a 428 ± 9.8 Ma upper intercept age derived from a discordia through the three zircon fractions. Air abrasion (Krogh 1982) of these fractions probably failed to remove all of the effects of Pb-loss. It is likely that volcanism started before plutonic crystallisation, which would push the initiation of magmatism perhaps to ~430 Ma.

The Rubha Mor appinitic intrusion (427 ± 3 Ma; Rogers and Dunning 1991) was considered, as was the type kentallenite nearby, to have been emplaced before the adjacent Ballachulish Pluton (Fig. 5). However, the subsequent determination of 427 ± 1 Ma for that pluton (Fraser et al. 2004), which has gradational contacts with other appinites (Weiss and Troll 1989), suggests that the Rubha Mor and Ballachulish intrusions were practically contemporaneous. The original 427 ± 3 Ma age was derived from a single titanite fraction and re-evaluation of the data gives the Rubha Mor titanite a $^{206}Pb/^{238}U$ age of 426.5 ± 4.2 (2σ, includes decay constant errors).

Shoshonitic lavas ($SiO_2 \leq 57.0$ wt.%; $K_2O \geq Na_2O - 2.0$) that are petrographically and compositionally indistinguishable from lamprophyre, and minor lamprophyre intrusions, occur in the Lorn Lava Pile (see below), dated here at 426-425 Ma.

Magma-mixing and mingling relationships between the Clashgour appinitic intrusion and its host monzogranite indicate the co-existence of the contrasting magmas here (Supplementary Fig. 2). The Clashgour outcrop (Figs 3 and 5) is most probably a part of the Rannoch Moor Pluton, which is dated here at 422.5 ± 0.5 Ma, but it could be part of the younger Clach Leathad or Etive plutons. Thus magmas that formed appinite were available for at least 6 million years in the early part of the episode.

Lamprophyre dykes are present widely within the Etive Dyke Swarm (e.g. Kynaston and Hill 1908), which formed from 418 Ma onwards for more than 3 million years (see below). One lamprophyre dyke cuts the contact between the 415 Ma Cruachan Intrusion and the younger Outer Starav Intrusion, and is chilled against both (Bailey and Maufe 1916, p. 124), and another, thin and amygdaloidal, lamprophyre dyke occurs close to the middle of the 408 Ma Inner Starav Intrusion (Fig. 5; 1.4 km south-southwest of the summit of Ben Starav). The latter contains phenocrysts of olivine (pseudomorphs), augite, hornblende and phlogopite, with K-feldspar, plagioclase, apatite, magnetite and calcite (see also Kynaston and Hill 1908, p. 117) and it might record the last phase of Siluro-Devonian magmatism in the region. Thus, mafic small-degree partial melts of lithospheric mantle that would form appinite or lamprophyre were present through at least 15 million years, ~430-<415 Ma, and possibly throughout the ≥22 million-year post-collision magmatic episode.

Etive Dyke Swarm and associated Etive Volcano

The Etive Dyke Swarm extends over 100 by 20 km, with the Etive Pluton near its centre (Fig. 3). Although the swarm is well known, it has not been well understood; there has been a tendency (e.g. Morris et al. 2005) to stress a relationship with the developing Etive Pluton while overlooking evidence that the swarm was well populated with dykes before the plutonic intrusion at this level. Here it is proposed that the swarm originated from the plumbing of a developing centred volcano that eventually became intruded by the pluton and was subsequently mostly removed by uplift and erosion.

Analyses of Etive dykes are plotted in Figure 2c and d. Both compositionally and petrographically the (fine-grained) dykes are indistinguishable from southwest Grampian volcanics, being closely matched by the diverse rocks of the earlier-emplaced Lorn Lava Pile. Numerous authors, e.g. Anderson (1937), following Bailey and Maufe (1916) and Kynaston and Hill (1908), have distinguished ‘felsite and porphyry’: dacites and rhyolites with or without quartz and feldspar phenocrysts, ‘prophyrites’: andesites and dacites with phenocrysts of hornblende or biotite and plagioclase, ‘microdiorites’ and ‘lamprophyres’. No systematic compositional sequence through time has ever been properly established and, by analogy with Lorn (see below), no systematic change need be expected.

As Etive dykes intrude the Cruachan Intrusion (Fig. 6), it is evident that the main development of the Etive Dyke Swarm took more than 3 million years, between the cooling of the Clach Leathad Pluton at ~418 Ma and after crystallization of the Cruachan Intrusion at ~415 Ma. Few dykes intrude the more central components of the Etive Pluton, but some do cut the Inner Starav Intrusion (408 Ma), including lamprophyre (see above). Uncertainty as to when, after 408 Ma, the last dykes were intruded causes some of the uncertainty in the overall duration of the Grampian post-collision magmatic episode. This is expressed as 'equal to or more than' (≥) 22 million years, which takes initiation as ~430 and the ending at, or some time after, ~408 Ma.

Morris et al. (2005) derived $^{40}Ar – ^{39}Ar$ ages of 415 ± 1.8 and 414 ± 2 Ma directly from Etive dykes and concluded that they were emplaced at ~415 Ma, somehow marking the end of magmatism and the Caledonian orogeny here. However, the full age range of the dykes was not determined systematically from the field relationships (e.g. Fig. 6) and the significantly younger age of the Inner Starav Intrusion, dated here at 408 Ma (see above), was not recognised. Further, assignation of the magmatism to the end of the Caledonian orogeny is distinctly problematic when neither the tectono-magmatic setting nor the intended meaning of 'orogeny' is clarified.

An instructive feature of the Etive Dyke Swarm is that its northeastern limb extends without deflection across the position of the previously influential northwest-trending crustal discontinuity known as the Glencoe Lineament. This crust-penetrating fault and shear zone (1) facilitated diverse magmas to rise rapidly to the surface at the Glencoe caldera volcano, (2) repeatedly influenced caldera-collapse, and (3) persistently formed a graben that directed a major river through the caldera (Moore and Kokelaar 1997, 1998; Kokelaar and Moore 2006). To explain the lack of deflection of the dykes, it is proposed that they propagated laterally northeastwards from the Etive centre at shallow crustal levels. They were not influenced by the Glencoe Lineament because they were emplaced within the solidified Clach Leathad Pluton, which had obliterated the northwest-trending discontinuity in the upper crust. The latter intrusion is more extensive at depth than at outcrop, extending outwards at least as far as beneath the Glencoe Fault-intrusions (Figs 3 and 5; British Geological Survey 2005).

It is proposed that the Etive Dyke Swarm was directly involved in the growth of an extensive and thick intermediate-to-silicic volcanic pile at and around the site subsequently intruded by the Etive Pluton; we unimaginatively name this volcanic pile – Etive Volcano. That there were vents and volcanic strata at the surface above the dyke swarm seems intuitively obvious, but the considerable thickness of the volcanic pile forming Etive Volcano has not previously been appreciated. In attempting to reconcile the original 3-6 km of cover above the level of outcrop of the Etive Pluton (Droop and Treloar 1981; Moazzen and Droop 2005; Droop and Moazzen 2007) with the apparently shallow, 1-2 km-depth, dome-shaped and miarolitic roof-zone outcrop of the neighbouring and earlier Clach Leathad Pluton, Kokelaar and Moore (2006) suggested that the Etive Pluton may have been displaced upwards, by at least ~1.5-2 km, relative to the Clach Leathad Pluton. However, there is no evidence of a continuous fault between the two intrusions, which are, in effect, stitched by unbroken Etive dykes and by contact alteration of (earlier) truncated dykes (Fig. 6).

The evident difference between the original thicknesses of cover of the two plutons is interpreted here as due to considerable thickening of the cover across the region after emplacement of the Clach Leathad Pluton and before emplacement of the Cruachan Intrusion. The greatest density of Etive dykes lies outside the Etive Pluton, implying that many, if not the majority, predate it (Fig. 6; see also British Geological Survey (2005) and the Geological Survey of Scotland (Oban) Sheet 45 (1907)). It is inferred that the earlier Etive dykes, and probably a central vent complex (since obliterated), built a substantial volcano south of the defunct Glencoe caldera volcano and Clach Leathad Pluton. The thickness of the Etive Volcano pile would have had to have been at least ~1.5-2 km. Such a thickness above the pre-existing Lorn Lava Pile, itself ≥600 m thick and densely populated with Etive dykes, would account for the greenschist-grade hydrothermal metamorphism of the Lorn lavas. The Etive pile, which was entirely removed by erosion (the screen in the south of the Etive Pluton may be a vestige of the Lorn pile), constituted ≥2000 km^3 of intermediate-to-silicic lavas and tuffs. Like Glencoe with its Clach Leathad Pluton, and Ben Nevis, Etive is consistent with the relationship – volcano followed by pluton – supporting our hypothesis of similar patterns for many if not all ‘Newer Granite’ plutons, from Donegal to Shetland. Etive demonstrates the point that, because of uplift and erosion, intermediate-composition magmas have been grossly underestimated, with significant implications for the overall petrogenesis.

Geochemistry

Southwest Grampian magmas, which formed for ≥22 million years after continental collision, were dominated by intermediate to silicic compositions with great major-element diversity (Fig. 2 a-c) and with considerable variations in the abundance of incompatible trace elements (ITE: LILE, LREE, Nb, Zr and Y; e.g. Figs 2d and 7). Most southwest Grampian magmas had the high Ba and Sr that characterises the Argyll and Northern Highland plutonic suite, although low Ba-Sr types more characteristic of the Cairngorm plutonic suite were also emplaced, both as plutons and extrusions (Figs 2d and 7b). The plutons and volcanic successions show overlapped compositions (e.g. Fig. 2a and d); the plutons are more restricted compositionally than the extrusions and they present a greater proportion of silicic compositions at outcrop: predominantly intermediate to silicic (granodiorite, monzogranite and granite) *vs*. predominantly intermediate (andesite, quartz latite and dacite). Some of the relative compositional restriction may reflect sampling fewer rocks from the plutons, although it can be interpreted as due to a greater prevalence of hybridisation of coexisting diverse magmas and remobilized cumulates during pluton growth. Key features, which are mainly recognised from studies of the volcanoes, are (1) compositionally diverse magma batches were emplaced together or separately in apparently haphazard succession (Fig. 7b and c), and (2) there appears little correlation between pairs of ITE (e.g. Figs 2d and 7b and c); binary ‘scattergrams’ indicative of highly variable ITE ratios in different magma batches *are characteristic*. Strong tectonic control of magma ascent and eruption, via crust-penetrating faults or shear zones, has been established (e.g. Watson 1984; Moore and Kokelaar 1997, 1998) and, to explain the key features, we suggest (1) both simultaneous and successive sampling of compositionally diverse magmas that co-existed at depth, and (2) that many of the co-existing magmas were not directly related genetically. Occurrence of numerous distinct and not obviously related batches of magma was reported by Thirlwall (1979, 1981), especially for Lorn.

Fig. 7

Magmas with primitive compositions were volumetrically scarce in the upper crust of the southwest Grampian region. A few appinite-lamprophyre intrusions have Mg-numbers ≥68 (100 Mg/(Mg+Fe^{2+})), but most rocks have ≤62. There are very few magmatic bodies with SiO_2 ≤50 wt.% (Fig. 2; Thirlwall 1979, 1981) and only a few, amongst the Lorn lavas, with between 50 and 52 wt.% SiO_2. Apart from the appinites, no sample we have analysed has Ni >220 ppm or Cr >500 ppm, and thus it is inferred that most of even the more mafic magmas that reached the upper crust were significantly evolved relative to compositions of primary mantle melts. Thirlwall (1981) considered that the moderately high Ni and Cr contents in the few basalts, and in some basaltic andesites, indicated modified mantle melts with some cumulates.

The sampled appinite-lamprophyre intrusions include mafic and intermediate compositions that indicate plagioclase-absent fractional crystallisation, consistent with storage at depths ≥30 km in the presence of water (Neilson 2008). They generally contain high Ni, Cr, V and MgO, coupled with elevated LILE and LREE abundances, high $Na_2O + K_2O$, and K_2O/Na_2O generally ≥1. Figure 2d shows that most of the sampled appinitic rocks maintain a fairly constant Sr/Ba ratio (~1.2), consistent with their direct genetic relationship.

Roughly one third of the Lorn samples (33 out of 95) show *compositional* similarities to the appinites-lamprophyres, in having several of the features: Ni >100 ppm, Cr >200 ppm, V >100 ppm, Sr >1000 ppm, Ba >100 ppm, Rb >60 ppm, La >49 ppm, Ce >79 ppm, Nd >35 ppm, $Na_2O + K_2O$ >8 wt.%, and K_2O/Na_2O >0.95 (Fig. 7b; Neilson 2008); about one fifth are shoshonitic (17 samples, including 11 intrusions, have SiO_2≤57.0 wt.% and K_2O≥Na_2O-2.0; after Le Maitre 1989). Thus, superficially, it appears possible that some of the Lorn magmas could be related to small-degree mantle melts directly via fractional crystallisation. Similarly at Glencoe a few lavas show appinite-lamprophyre affinities (Neilson 2008). However, in contrast to the appinites and lamprophyres, most of the compositionally similar volcanic rocks contain few phenocrysts, in many cases lacking hydrous phases (amphibole, biotite/phlogopite) and including plagioclase. Importantly, in binary plots involving incompatible trace elements, for example Figures 2d and 7b, the volcanic rocks produce ‘scattergrams’ in which the majority of compositions cannot be related to the appinitic series by crystal-liquid fractionation. The volcanic rocks contain the signature(s) of an ultimate origin in enriched lithospheric mantle, but most cannot be related directly to the contemporary appinite-lamprophyre magmas. The geochemical similarities are interpreted as reflecting a common ultimate source – the lithospheric mantle – but with the extrusive and intrusive intermediate to silicic rocks mostly derived by partial melting of the crust and *not* directly related to mantle melts by fractional crystallisation.

Petrogenesis in the new time-frame

The timing of events in the Siluro-Devonian magmatic episode in the southwest Grampian region, and the inferred concurrent tectonic processes, are summarised in Figure 8. Episodes of voluminous eruption of diverse intermediate magmas were common; the deeply eroded Grampian metamorphic basement was repeatedly covered by extensive volcanic fields that mostly were eroded away quickly because of uplift, for example, as recorded by contemporary conglomerates in the Midland Valley (Bluck 1984). The Lorn Lava Pile (~426-425 Ma) represents a minimum of 300 km^3 (probably >>1000 km^3) of mainly andesite and dacite emplaced some 5 million years before the Basal Andesite Sill-complex and its associated andesitic lavas at Glencoe (≤420.5 Ma). Glencoe caldera volcano and its associated Fault-intrusions record several further major eruptions of andesite and dacite, during ~420-418 Ma, and then, during ~418-≤415 Ma, >2000 km^3 of intermediate-to-silicic magmas were emplaced to form the Etive Dyke Swarm with associated Etive Volcano. The several plutons, Rannoch, Clach Leathad and Etive, also represent large volumes of intermediate to silicic magmas.

Fig. 8

Mafic magmas that would form appinite-lamprophyre were certainly available for at least 15 million years. Small volumes were emplaced in the upper crust together with the intermediate to silicic magmas around and within the plutons, in the Lorn Lava Pile and in the Etive Dyke Swarm (and hence also in Etive Volcano). Although both appinite-lamprophyre and intermediate-silicic magmas utilised the same pathways to the upper crust, as noted by Watson (1984), the compositions indicate that they were not genetically related in the same way as in the hybrid appinite-syenite-granite intrusions that record fractional crystallisation (e.g. Ach'uaine hybrid; Fowler 1988). A relative volume problem arises (e.g. Tarney and Jones 1994; Fowler et al. 2001) in attempting to attribute the large volumes of intermediate-silicic magmas to derivation by fractionation from appinite-lamprophyre parent magmas, albeit involving some crustal assimilation. Fractional crystallisation alone to form huge volumes of andesite and dacite from mantle-derived magma would require very substantial volumes of mafic and ultramafic cumulates to form at or within the base of the crust, as in the mafic sills of an active 'deep crustal hot zone' (see Annen et al. 2006). Despite frequent tapping of diverse andesite-dacite magma batches along crust-penetrating faults, little mafic magma was erupted or intruded and mafic cumulate xenoliths are unknown. The substantial volume of intermediate to silicic magmas emplaced over ≥22 million years is most easily reconciled with extensive partial melting of crust.

The considerable variability (characteristic scatter) of the ITE (e.g. Figs 2d and 7) is interpreted as primarily reflecting partial melting of an inhomogeneous but mainly incompatible-element-rich, mafic-to-intermediate lower crustal underplate *and* fractional crystallisation of the mantle-derived appinitic-lamprophyric melts. Some variability attributable to crustal assimilation is inevitable. The lower crustal melting, at ≥950 °C (e.g. see Petford and Gallagher 2001; Annen et al. 2006) and in the absence of plagioclase, would have needed a substantial source of heat and volatiles; it also entailed uplift (see below). The persistent mantle-derived (appinite-lamprophyre) magmas are interpreted as providing the required heat and volatiles. This new petrogenetic model is illustrated in Figure 9.

Fig. 9

There are additional reasons for invoking mainly crustal melting together with some fractionation of hydrous mantle melts. (1) Partial melting of mafic-to-intermediate lower crust can simply account for the considerable variability of incompatible trace-element contents, including the generation of low Ba-Sr magmas, via heterogeneity of the crust and differing degrees of its melting. An origin primarily involving fractional crystallisation would be difficult to reconcile with the compositional diversity and the lack of any systematic compositional trend through time (Figs 2 and 7); it would have to entail periodic tapping of numerous bodies of magma at widely differing stages of fractional crystallisation, without accidental incorporation of cumulate. Presumably, mafic restite is less easily incorporated in segregating partial melts, relative to the cumulate fragments that are commonly included where fractional crystallisation has occurred. (2) Appinitic fractionation series typically have strongly potassic intermediate daughter compositions (syenite-trachyte), rather than (trachy-) andesite. In nature, extrusive counterparts of high-K calc-alkaline appinites and lamprophyres are rare, although some shoshonitic lavas may fit this role (e.g. Rock 1991). In the Grampian case, there are only a few lavas, in the Lorn pile, that are petrographically and chemically closely similar to lamprophyres and are potassic (shoshonitic: $SiO_2 \leq 57.0$ wt.%; $K_2O \geq Na_2O-2.0$). (3) The high Ba-Sr link (e.g. Rock and Hunter 1987; Fowler and Henney 1996) and 'volume problem' are simply resolved if the mafic-to-intermediate lower crust was formed from enriched sub-continental lithospheric mantle during a *previous* episode of underplating. Mafic crust with subduction-arc affinities was produced from the mantle at ~1.8 Ga (see above; Marcantonio et al. 1988; Dickin and Bowes 1991; Muir et al. 1992, 1994) and this seems a possible source in the later crustal melting.

The apparently minor occurrence of little-evolved lithospheric mantle melts (mafic lamprophyres and appinite assemblages) in the upper crust does not preclude their substantial presence at depth, where they would have provided a viable advective heat and volatile source to partially melt the mafic-to-intermediate crustal underplate. In early stages such magmas commonly reached the upper crust, but their ascent may have become increasingly restricted by their greater density than the zones with evolved magmas that formed above them. The model proposed here involves repetitive intrusion of the lower crust by mantle melts throughout the magmatic episode, conceivably reflecting the duration of availability of the asthenospheric heat source. Repetitive intrusion of sheets of volatile-bearing mantle melt would be an efficient mechanism in crustal melting (e.g. Petford and Gallagher 2001) and would inevitably have led to some hybridisation of the respective magmas, contributing further to the plethora of compositions in existence. Recent studies of oxygen isotopes in zircons collected from dioritic rocks associated with 'Newer Granites' of the Grampian Terrane (Loch Nagar pluton) also have indicated a petrogenesis primarily involving crustal melting and the mixing of different batches of melts (Appleby et al. 2008).

The model proposed here involves both crustal growth, by addition from the lithospheric mantle, and a considerable amount of crustal recycling (Fig. 9). The magmatism followed the collision of Laurentia with Baltica and was initiated at the time of the collision of Laurentia with Avalonia. It is inferred that breakoff and sinking of the subducted Iapetus lithospheric slab allowed buoyant ascent of 'dry', relatively hot asthenosphere through the developing break to impact the hydrous, enriched lithospheric mantle above (see also Atherton and Ghani 2002). The ascent of asthenosphere must have been of sufficient scale and longevity to account for the ≥22 million-year duration of the magmatic episode, although it yielded no magma directly (see Halliday et al. 1985), because it was dry. In cases of slab breakoff, the duration of magmatism is related to the scale and longevity of the asthenospheric flow that is caused by the sinking slab and which leads to heating and partial melting of the sub-continental lithospheric mantle. In the European Alps, for instance, the duration of the Tertiary slab-breakoff-related magmatism was ~17 million years. Shallower breakoff leads to a larger anomaly and more melting (Davies and von Blankenburg 1995).

Discussion

As well as being impossible to reconcile with continuing active subduction, it is also clear that the Grampian volcanoes were unlike typical subduction-related varieties (e.g. Ewart 1976; Gill 1981). The Grampian andesites and dacites are relatively phenocryst-poor, were erupted in unusually large volumes and with unusually low viscosity (e.g. at Glencoe, Bidean nam Bian single cooling unit of ≥12 km^3), and predominantly formed laterally extensive lava sheets and sills; the ≥600 m-thick Lorn pile has commonly been described as comprising plateau lavas (e.g. Pitcher 1982). Big stratovolcanoes with broad volcaniclastic ring-plains are not recorded. Large volumes of magma were tectonically facilitated to ascend rapidly to the surface, with little opportunity for storage, fractionation, degassing or cooling at shallow levels (Kokelaar and Moore 2006).

Our geochemical and field data together establish that the Grampian Siluro-Devonian plutons and volcanic rocks are essential counterparts, the similar compositions of which reflect essentially the same petrogenesis in this episode of post-collision magmatism. Evidence that centred volcanoes were substantially obliterated by genetically related plutons, and that substantial thicknesses of lavas and tuffs were removed by erosion, much of it due to uplift during the magmatism (Kokelaar and Moore 2006; Neilson 2008), leads us to advocate that similar volcanic activity originally occurred widely across the entire Grampian belt of 'Newer Granites', including Donegal (e.g. Pitcher and Berger 1972) and Shetland (e.g. Flinn et al. 1968). This proposal is consistent with the compositional overlap of the Grampian rocks and Donegal plutons (Fig. 2a).

Most previous authors concerned with understanding the origin of the Caledonian 'Newer Granites' have invoked either subduction or some relation to the uplift and major strike-slip faulting that is known to have accompanied the magmatism. Those advocating the latter (e.g. Pitcher 1982; Watson 1984) were influenced by those 'Newer Granites' that occur in the Southern Uplands and England, clearly too close to the Iapetus suture for them to be subduction-related. These more southerly plutons are now recognised (Brown et al. 2008) as younger than those further north, and as unrelated. This paper establishes that the Grampian magmatism was not related to on-going subduction and that its onset at the time of continental collision strongly suggests this as causative, as recognised by Atherton and Ghani (2002). Our model explains the well known and distinctive association of appinite-lamprophyre magmas with the high Ba-Sr granitoids and resolves the supposed 'volume problem' while highlighting previously unrecognised huge volumes of intermediate magmas. Magma chemistry, volumes and timing in relation to tectonism preclude explanations for the magmatism in terms of lithosphere delamination or post-orogenic uplift-related decompression melting. Slab breakoff is the best petrogenetic option available, although it is impossible to prove unequivocally.

Atherton and Ghani (2002) were the first to advocate slab breakoff, to account for 'late granite syn-collisional magmatism' in the Scottish Highlands and in Donegal. They invoked progressive slab detachment that was initiated by collision between Baltica and Laurentia and then produced a linear uplift belt with a singular, peaked pulse of plutonic magmatism related to migration of the heat source (towards Donegal). In the southwest Grampian study area no such peak is distinguished, but instead a sustained episode involving repeated developments of volcanoes and plutons over ≥22 million years within an area that has a length scale similar to the crustal thickness. There is no sign of any protracted hiatus in the magmatic episode, or of any protracted waxing or waning stage, although assimilation of silicic crust appears to have increased towards the end (Neilson 2008), which is not surprising. Our model also differs from that of Atherton and Ghani (2002) in recognising and accounting for the very large volumes of intermediate magmas that cannot be directly related to the appinite-lamprophyre magmas. In their model, Atherton and Ghani (2002, figure 7) first derive the appinitic magmas from the mantle to form a crustal underplate and then partially melt this underplate to form the granites, whereas, for reasons detailed above, we invoke persistent underplating and persistent melting of previously formed mafic crust (Fig. 9). Atherton and Ghani's model involves mainly crustal growth, whereas the model presented here involves considerable crustal recycling. It is not clear whether slab breakoff beneath the Grampian Terrane was a mechanical result of the subduction-resistance of continental Avalonia, or due to lateral propagation of slab detachment from the earlier collision involving Baltica. Slab segmentation may have interrupted any lateral propagation of detachment and our data are inadequate to resolve this uncertainty.

Uplift and erosion were rapid and considerable (evident from Ballachulish, Rannoch Moor and Etive plutons; see above), with differential movements on major northwest- and southwest-trending faults controlling the local preservation of thick volcanic successions (Lorn and Glencoe) (Fig. 3). It is inferred that the uplift was stimulated by early rebound consequent upon breakoff of the subducted dense lithospheric slab, tumescence over the buoyantly rising asthenosphere, magmatic underplating, and crustal heating.

It seems likely that a linear domain of asthenosphere rising through a slab break would, as well as causing uplift, form a linear zone of melting of mantle and lower crust that in turn would produce a linear belt of magmatism, thermal weakening, and focussing of tectonic strain in the mid- and upper-crust (Fig. 9). The crust-penetrating Great Glen Fault and similar parallel faults in the Grampian region (Fig. 3) would then be *consequent upon* existence of the hot zone, even defining the dimensions of its main locus of activity, with the crustal weakness being exploited on the switch to plate-scale strike-slip. Both strike-slip and substantial normal displacements were common on faults such as the Etive-Laggan, Ericht-Laidon, Tyndrum and Garabal faults (Treagus 1991). The latter fault is inferred to have plumbed the early (~428-429 Ma) magmas of the Garabal Hill-Loch Fyne Complex (Rogers and Dunning 1991) and the Etive-Laggan Fault, and its splays, clearly developed ≥500 m of downthrow to the west shortly before facilitating major magma ascent during caldera volcanism at

Glencoe (Kokelaar and Moore 2006). Logically one might expect the asthenospheric plume to have broadened its influence through time, but our data are inadequate to test for this. Our view, that uplift was consequent upon melt-forming processes and that fault development and activity were focused by the magmatism, is the virtual opposite of the former proposals (e.g. Pitcher 1982; Watson 1984) that the magmatism passively resulted from uplift or was triggered from above by vertical and horizontal fault-block movements.

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Figure Captions

Figure 1. The ‘Newer Granites’ of Scotland and Ireland (black). The intrusions in the Southern Uplands belong to the ‘Trans-Suture Suite’ attributed by Brown et al. (2008) to a transtensional origin; their ages range from 400-390 Ma, they have significant S-type characteristics and they are not petrogenetically related to the ‘Newer Granites’ further north that are the subject of this paper. SH is the Strath Halladale Granite; RC is the Rhinns Complex.

Figure 2. (a) R1-R2 diagram for comparison of volcanic and plutonic rock compositions (De la Roche et al. 1980), showing tectono-magmatic discriminant fields applicable to granitoids (after Batchelor and Bowden 1985). Southwest Grampian pluton compositions (22 analyses) lie entirely within the Donegal ‘granites’ field (from Ghani 1997; Atherton and Ghani 2002), but barely overlap with the Coastal Batholith of Peru (Pitcher et al. 1985). Southwest Grampian volcanic compositions show considerable diversity (77 analyses from Glencoe, 95 from Lorn), with compositions on both sides of the ‘critical line of silica saturation’ (R1/R2=1). The Southwest Grampian pluton compositions lie almost entirely within the compositional field of the volcanics. Unlike the majority of Southwest Grampian volcanic and plutonic compositions, the Appinite Suite plots for the most part in the pre-plate collision field (13 analyses). (b) Southwest Grampian volcanic compositional field showing rock classification and contours of SiO_2 wt.% (from De la Roche et al. 1980). (c) Detail of the Southwest Grampian volcanic compositions (triangles) and Etive Dyke Swarm compositions (diamonds: including data from Maufe and Bailey 1916; Garnham 1988; MF Thirlwall, unpublished). (d) Sr *vs*. Ba plot for Southwest Grampian volcanics, plutons, appinites and Etive Dyke Swarm. Such binary ‘scattergram’ plots of incompatible trace-element data are a distinctive feature of the magmatism, most simply reconciled with the occurrence of numerous batches of magma that had no direct genetic relationship. LowBaSr granitoid field from Tarney and Jones (1994).

Figure 3. Southwest Grampian Siluro-Devonian post-collision magmatic rocks: plutons, volcanoes and dykes of the study region are shown with surrounding plutons and volcanoes. The Ben Nevis Dyke Swarm is not shown.

Figure 4. U-Pb concordia diagrams. (a) Lorne Lava Pile; (b) Rannoch Moor Pluton; (c) Glencoe Fault-intrusion; (d) Clach Leathad Pluton; (e) Cruachan Intrusion; (f) Inner Starav Intrusion.

Figure 5. Regional map showing sample sites with the previous (unshaded boxes) and new U-Pb zircon ages (shaded boxes).

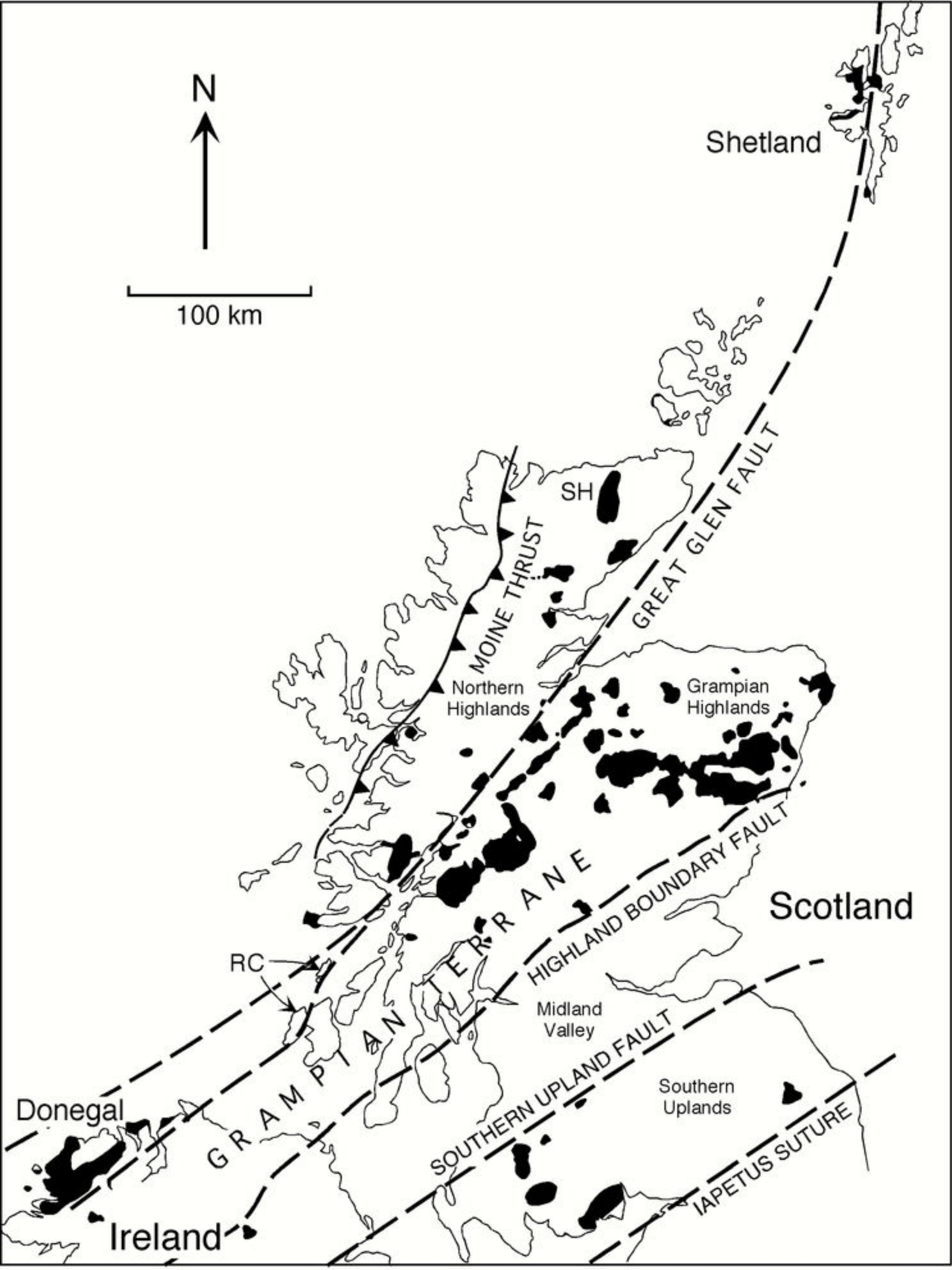
Figure 6. Clach Leathad Pluton (pale green), dated at 418 Ma, is cut by the Meall Odhar and Cruachan Intrusions (reds) of the Etive Pluton, dated at 416-415 Ma (see text). Some Etive dykes (purple) are cut and metamorphosed by the Meall Odhar Intrusion, one dyke emanates from that intrusion and several dykes cut it and the Cruachan Intrusion. The outcrop of the older pluton is close to its original domed roof, formed at 1-2 km depth, whereas the Etive Pluton outcrop and wide metamorphic aureole (not shown) indicate significantly thicker original cover, perhaps 3-6 km (Droop and Treloar 1981; Moazzen and Droop 2005; Droop and Moazzen 2007). Vertical relative displacement between the plutons is precluded by the dykes that (1) are truncated and altered by the Meall Odhar Intrusion, and (2) emanate from or cut straight across the contacts. The figure is based on original Geological Survey mapping (reported by Bailey 1960) compiled in a new map (British Geological Survey 2005). FI is Fault-intrusion (of Glencoe); DAQ and DAS are Dalradian quartzites and schists, respectively. Map coordinates relate to Ordnance Survey 100 km square NN.

Figure 7. (a) Harker plots of representative incompatible trace-element compositions in Southwest Grampian volcanic, plutonic, appinite-lamprophyre and dyke rocks (Neilson 2008). (b) Compositional variation in the Lorn Lava Pile and associated intrusions, demonstrating that despite a wide range of states of compositional ‘evolution’, here in MgO *vs*. SiO_2, there is no evolutionary trend through time, expressed as MgO *vs*. stratigraphic height in the ≥600 m succession (intrusions are plotted at height equals zero). Widely different magma batches were emplaced successively and haphazardly during ≤1 million years, with widely varying ratios of incompatible trace elements (here Sr and Ba). Such ‘scattergram’ plots involving incompatible trace elements are a general characteristic of the magmatism (e.g. Fig. 2d). (c) Contrasting compositional batches of magma emplaced together in a single caldera-wide lake of ≥12 km^3 of magma erupted via crust-penetrating faults. The samples derive from four vertical transects spaced across Glencoe caldera volcano; the

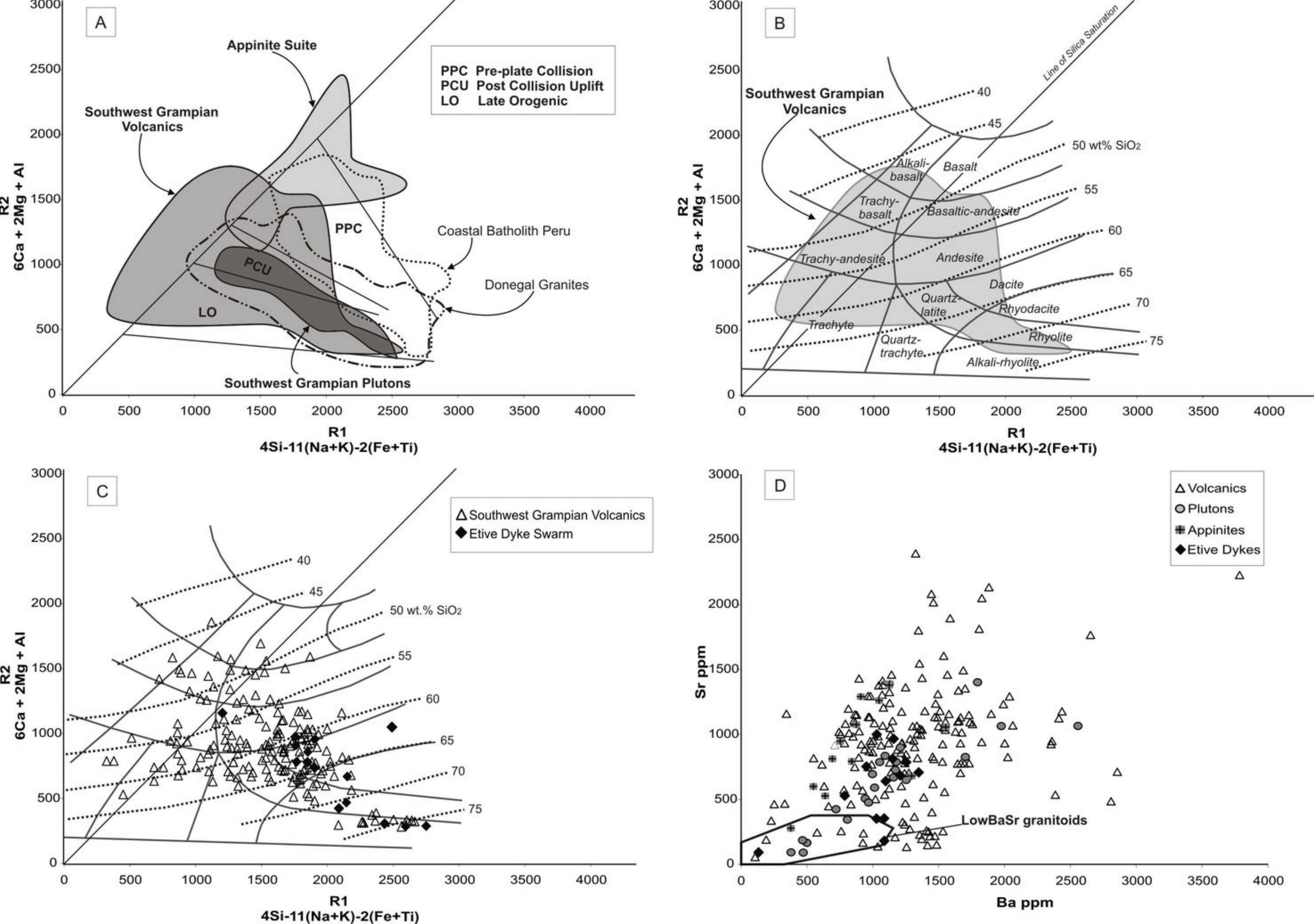
rocks constitute a single cooling unit, locally ≥350 m thick, and no internal contacts are discernible (Kokelaar and Moore 2006; Neilson 2008).

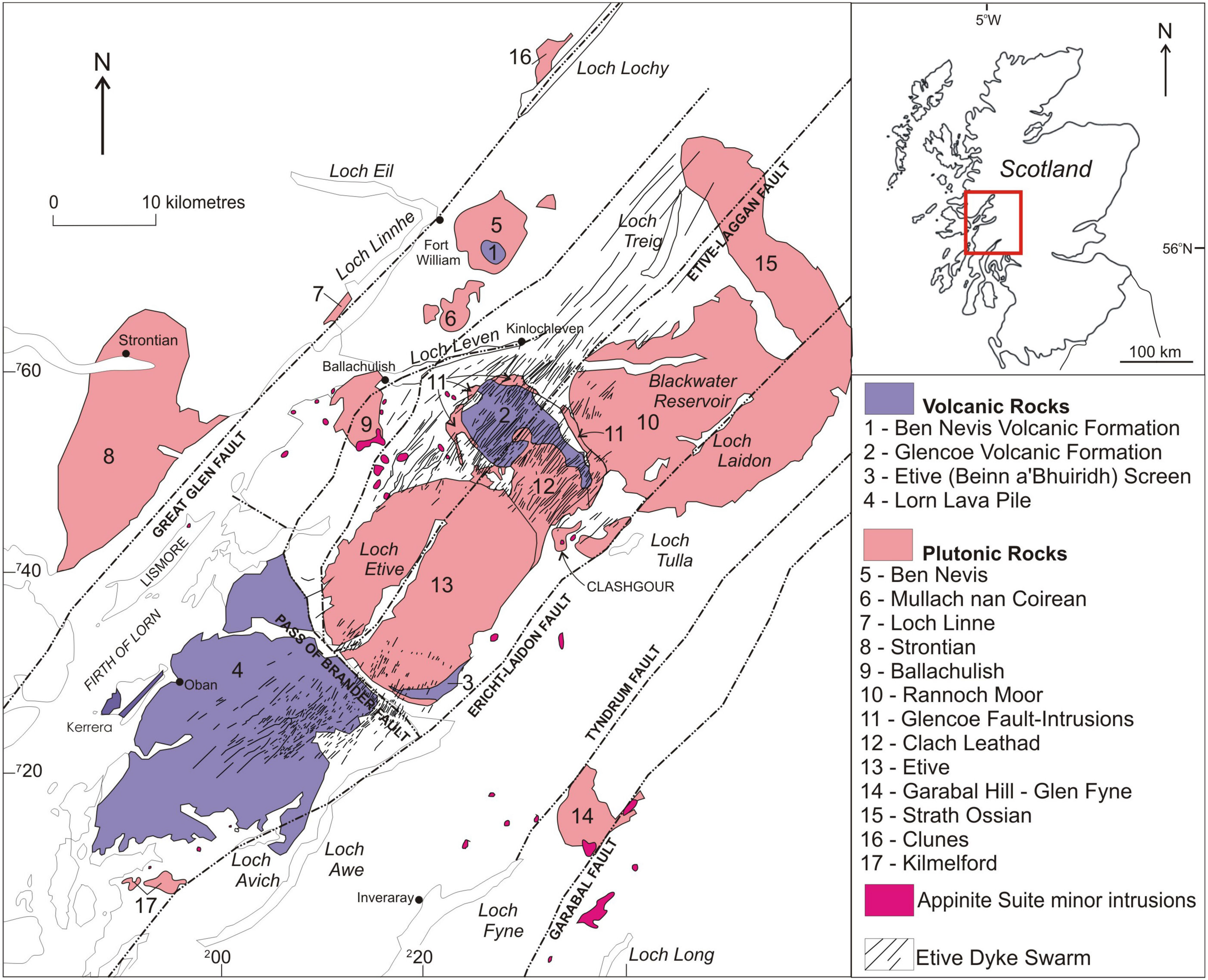
Figure 8. Time-lines for Siluro-Devonian tectonic and magmatic activity relevant to the southwest Grampian region. The Silurian-Devonian boundary is at 418 Ma (Kaufmann 2006). Age ranges of Ballachulish and Rannoch Moor plutons are less certain than others. Etive Volcano and Etive Dyke Swarm were mostly formed during the first ~4 million years; the extensive ranges of these account for the few dykes, including lamprophyre, that cut the Inner Starav Intrusion dated at 408 Ma (see text). [1]Onset age derived from recalculation of results of Rogers and Dunning (1991). [2]After Fraser et al. (2004). [3]After Dewey and Strachan (2005).

Figure 9. Schematic diagram illustrating the proposed petrogenetic model, involving slab breakoff and consequent rise of relatively hot and dry asthenosphere, partial melting of LILE- and LREE-enriched lithospheric mantle, new crustal addition of magmas that form appinite-lamprophyre, and crustal recycling by partial melting of mafic-to-intermediate lower crust to form plutons and volcanoes. Ascent of diverse magmas to the upper crust was facilitated by major faults that formed in this region as a consequence of the thermal weakening of the crust here and the onset of plate-scale sinistral strike-slip. Considerable uplift resulted from the slab detachment, from the buoyant rise of asthenosphere, and from the heating of the lithosphere. The model shows a putative early to middle stage of the magmatic episode and it is inferred that the asthenospheric plume broadens through time.



N
100 km
Shetland
SH
MOINE THRUST
GREAT GLEN FAULT
Northern Highlands
Grampian Highlands
HIGHLAND BOUNDARY FAULT
GRAMPIAN TERRANE
Scotland
RC
Midland Valley
SOUTHERN UPLAND FAULT
Southern Uplands
IAPETUS SUTURE
Donegal
Ireland


A
Appinite Suite
PPC Pre-plate Collision
PCU Post Collision Uplift
LO Late Orogenic
Southwest Grampian Volcanics
PPC
PCU
LO
Coastal Batholith Peru
Donegal Granites
Southwest Grampian Plutons
R2
6Ca + 2Mg + Al
R1
4Si-11(Na+K)-2(Fe+Ti)
B
Line of Silica Saturation
Southwest Grampian Volcanics
40
45
50 wt% SiO2
55
60
65
70
75
Alkali-basalt
Basalt
Trachy-basalt
Basaltic-andesite
Trachy-andesite
Andesite
Dacite
Quartz-latite
Rhyodacite
Trachyte
Quartz-trachyte
Rhyolite
Alkali-rhyolite
C
Southwest Grampian Volcanics
Etive Dyke Swarm
50 wt.% SiO2
D
Volcanics
Plutons
Appinites
Etive Dykes
LowBaSr granitoids
Sr ppm
Ba ppm


N
0 10 kilometres
Loch Lochy
16
Loch Eil
Loch Linnhe
Fort William
5
1
7
6
Loch Treig
ETIVE-LAGGAN FAULT
15
Strontian
Kinlochleven
Loch Leven
Ballachulish
11
2
11
10
Blackwater Reservoir
Loch Laidon
9
8
12
GREAT GLEN FAULT
LISMORE
Loch Etive
13
Loch Tulla
CLASHGOUR
FIRTH OF LORN
PASS OF BRANDER FAULT
ERICHT-LAIDON FAULT
4
Oban
Kerrera
3
TYNDRUM FAULT
14
GARABAL FAULT
Loch Awe
Loch Avich
17
Inveraray
Loch Fyne
Loch Long
760
740
720
200
220
5°W
N
Scotland
56°N
100 km
Volcanic Rocks
1 - Ben Nevis Volcanic Formation
2 - Glencoe Volcanic Formation
3 - Etive (Beinn a'Bhuiridh) Screen
4 - Lorn Lava Pile
Plutonic Rocks
5 - Ben Nevis
6 - Mullach nan Coirean
7 - Loch Linne
8 - Strontian
9 - Ballachulish
10 - Rannoch Moor
11 - Glencoe Fault-Intrusions
12 - Clach Leathad
13 - Etive
14 - Garabal Hill - Glen Fyne
15 - Strath Ossian
16 - Clunes
17 - Kilmelford
Appinite Suite minor intrusions
Etive Dyke Swarm

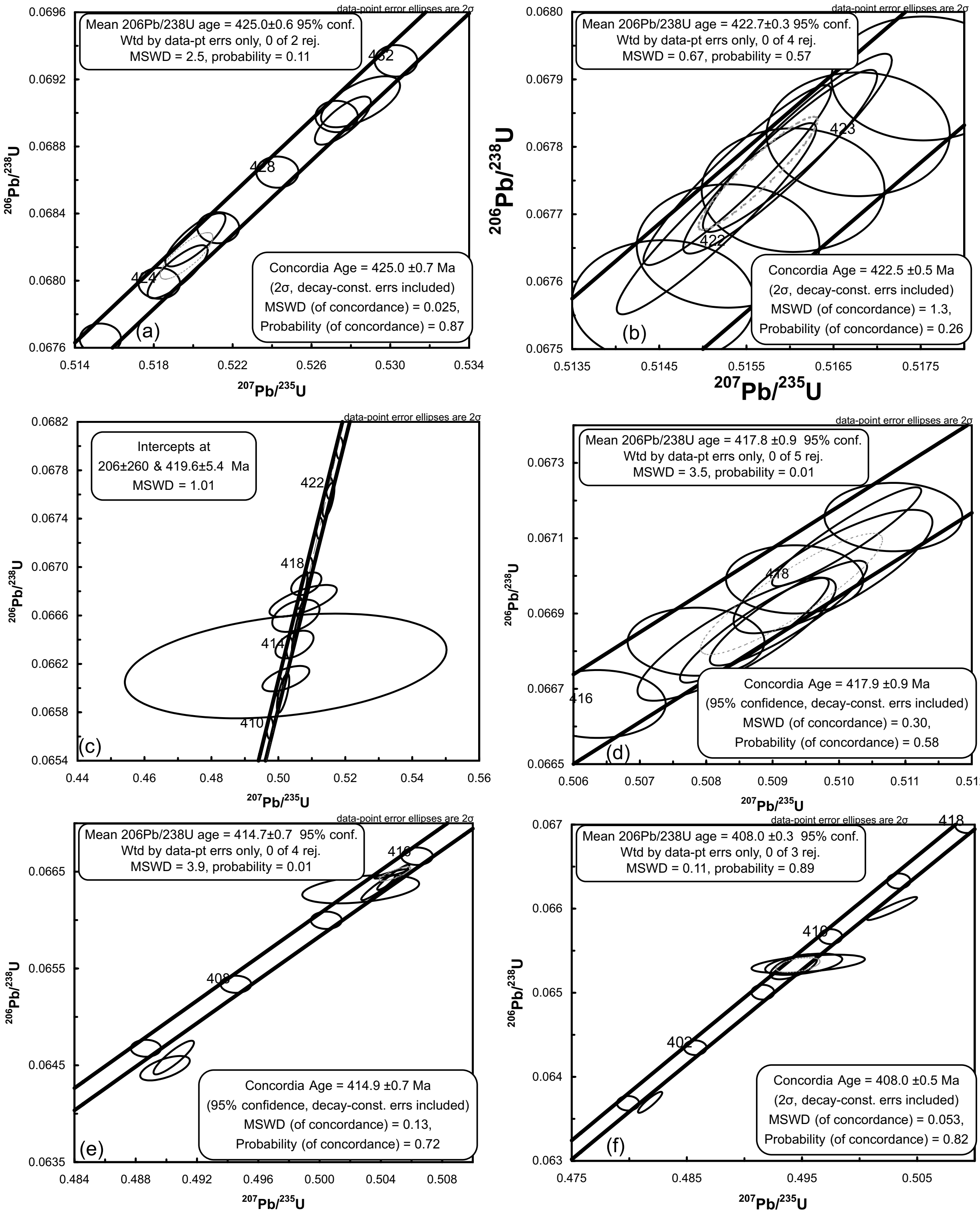
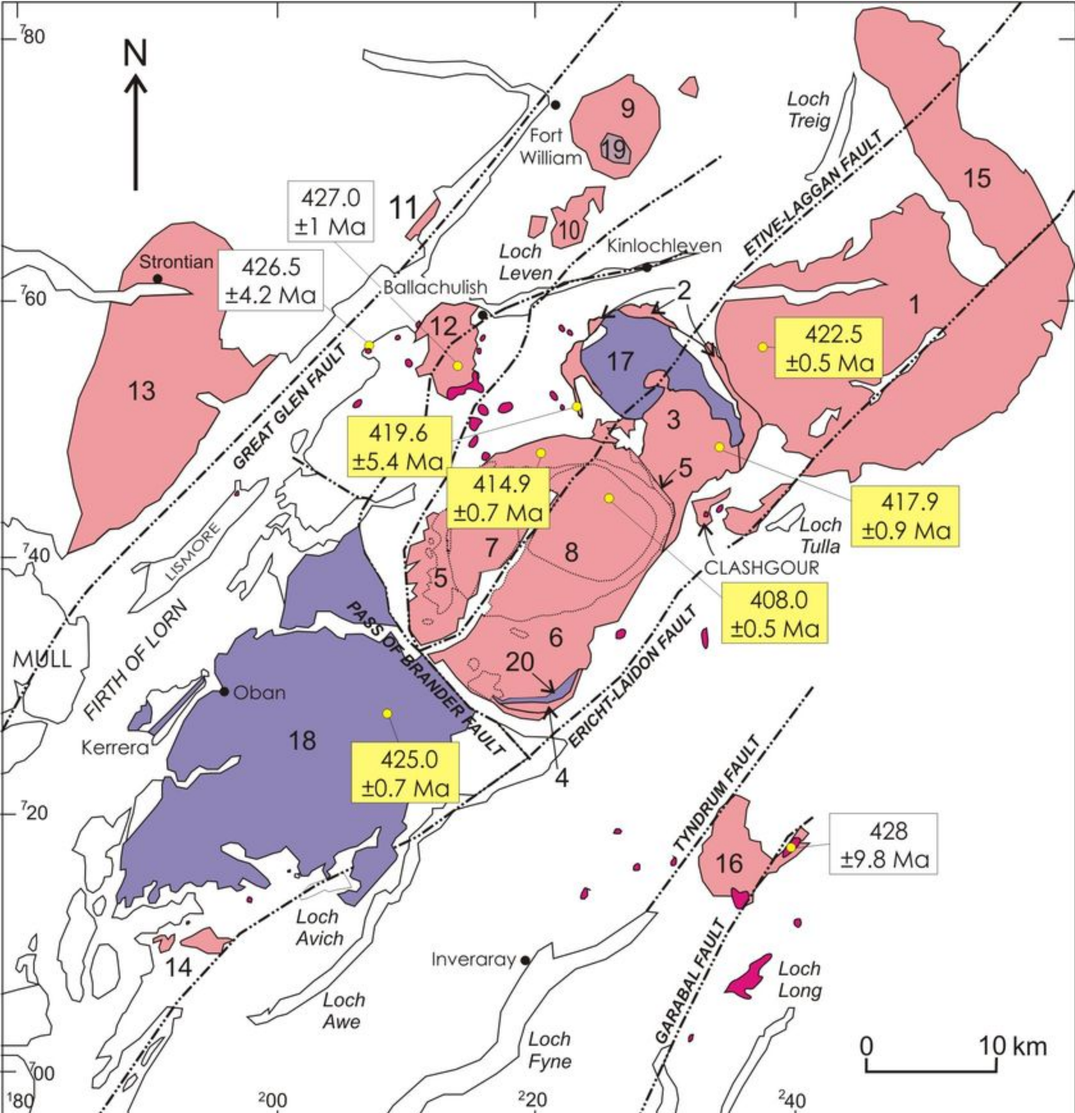


Figure 4. U-Pb concordia diagrams. A: Lorne Lava Pile; B: Rannoch Moor Pluton; C: Glencoe Fault Intrusion; D: Clach Leathad Pluton; E: Cruachan Intrusion; F: Inner Starav Intrusion.



Plutonic Rocks

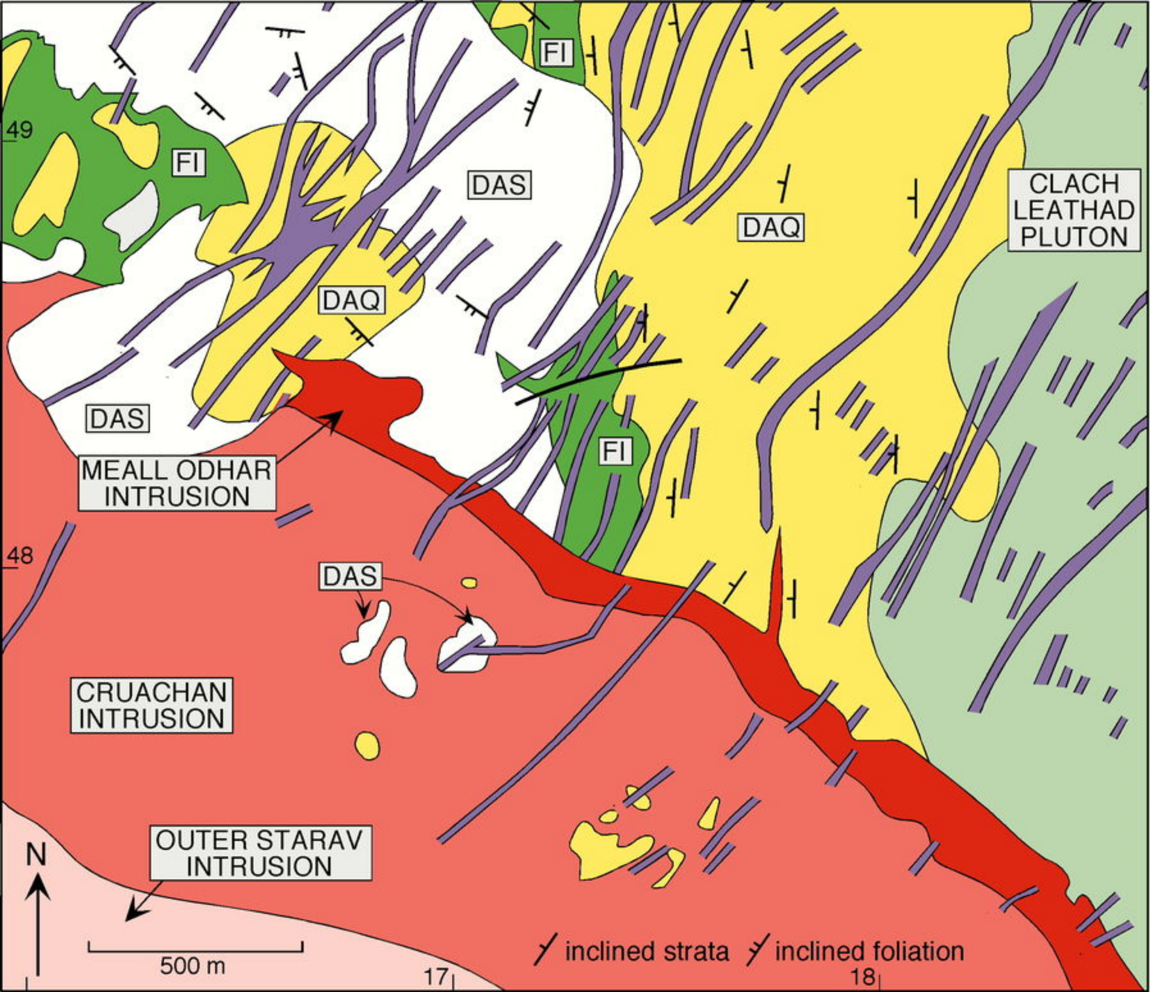
1 Rannoch Moor
2 Glencoe Fault-intrusions
3 Clach Leathad
4 Quarry Intrusion
5 Meall Odhar Intrusion
6 Cruachan Intrusion
7 Porphyritic Outer Starav
8 Inner Starav
9 Ben Nevis
10 Mullach nan Coirean
11 Loch Linne
12 Ballachulish
13 Strontian
14 Kilmelford
15 Strath Ossian
16 Garabal Hill - Glen Fyne

Appinite Suite minor intrusions

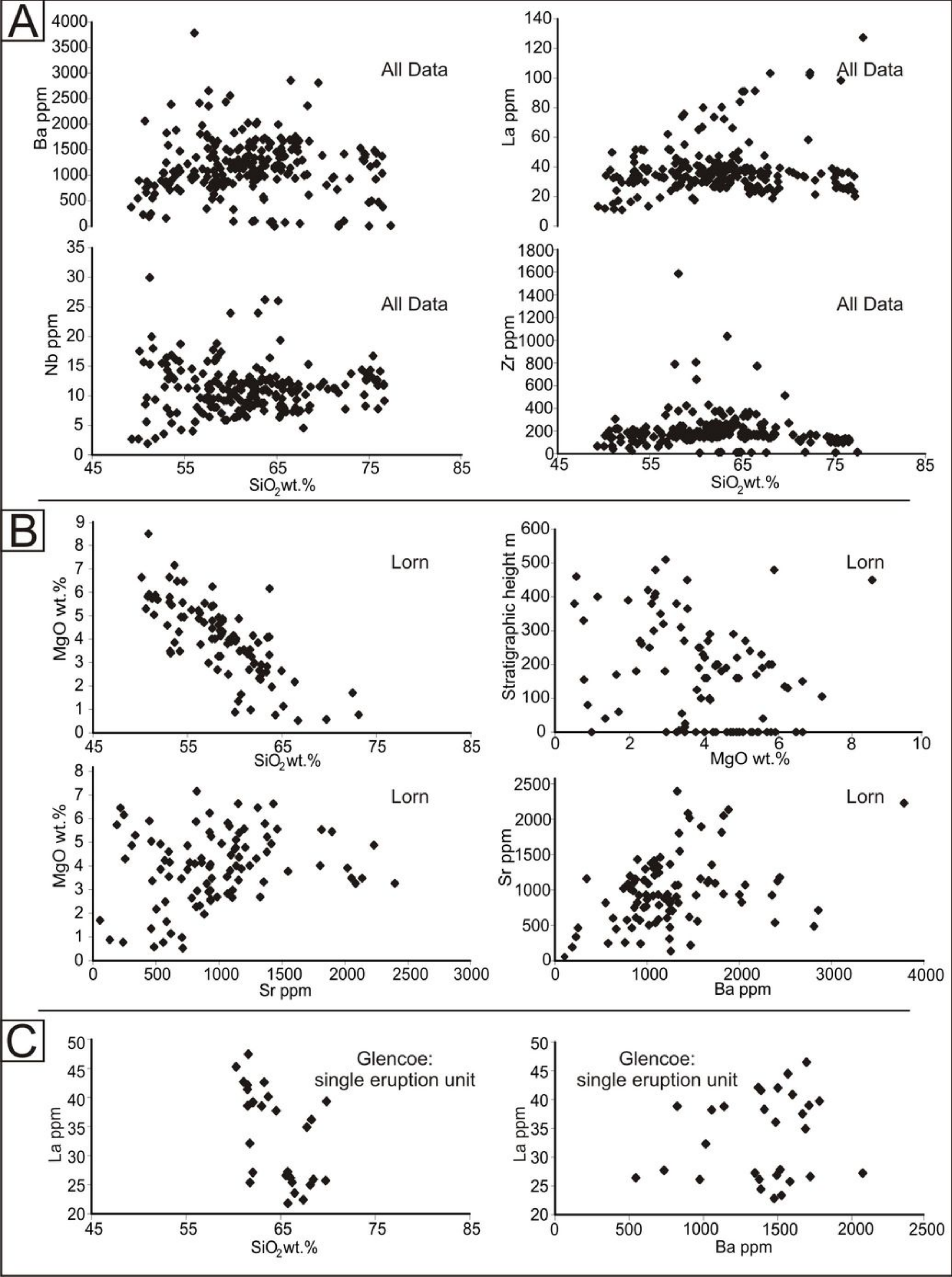
Sample site

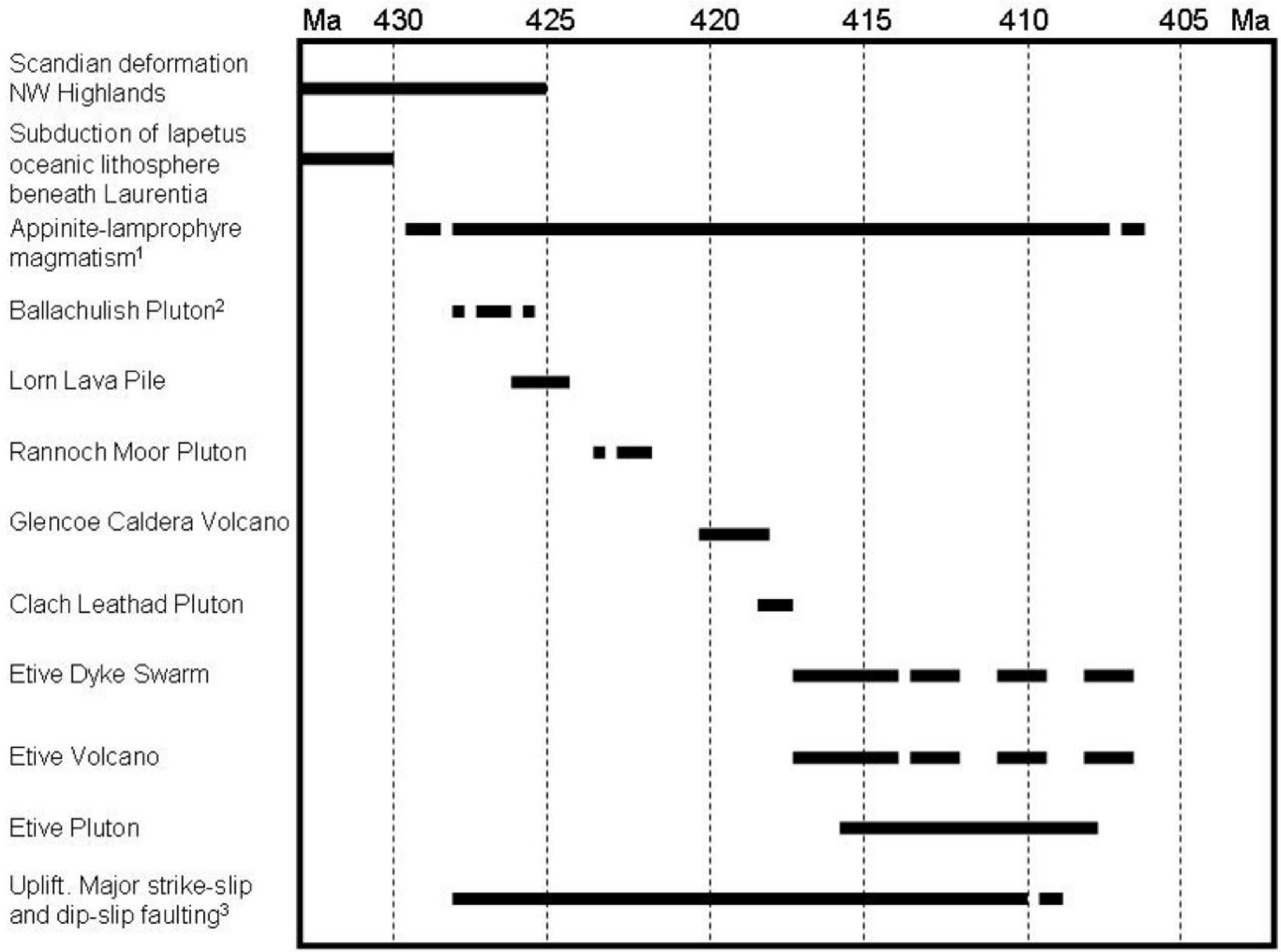
Volcanic Rocks

17 Glencoe Volcanic Formation
18 Lorn Lava Pile
19 Ben Nevis
20 Beinn a'Buiridh Screen

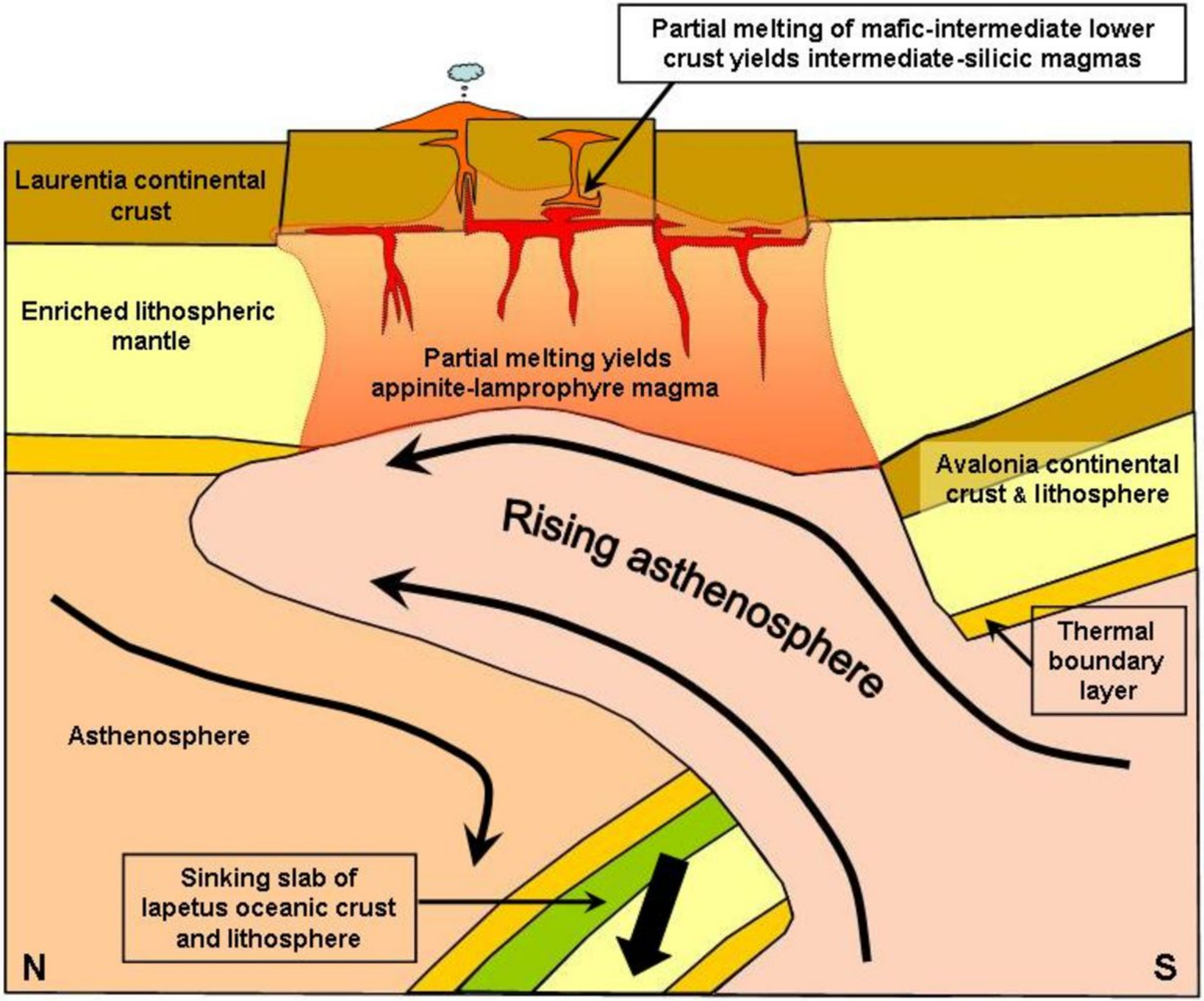


FI
FI
FI
DAS
DAS
DAS
DAQ
DAQ
CLACH
LEATHAD
PLUTON
MEALL ODHAR
INTRUSION
CRUACHAN
INTRUSION
OUTER STARAV
INTRUSION
N
500 m
49
48
17
18
inclined strata
inclined foliation


A
All Data
Ba ppm
La ppm
Nb ppm
Zr ppm
SiO2 wt.%
B
Lorn
MgO wt.%
Stratigraphic height m
Sr ppm
Ba ppm
C
Glencoe:
single eruption unit
La ppm
SiO2 wt.%
Ba ppm



Ma
430
425
420
415
410
405
Ma
Scandian deformation NW Highlands
Subduction of Iapetus oceanic lithosphere beneath Laurentia
Appinite-lamprophyre magmatism[1]
Ballachulish Pluton[2]
Lorn Lava Pile
Rannoch Moor Pluton
Glencoe Caldera Volcano
Clach Leathad Pluton
Etive Dyke Swarm
Etive Volcano
Etive Pluton
Uplift. Major strike-slip and dip-slip faulting[3]


Partial melting of mafic-intermediate lower crust yields intermediate-silicic magmas
Laurentia continental crust
Enriched lithospheric mantle
Partial melting yields appinite-lamprophyre magma
Avalonia continental crust & lithosphere
Rising asthenosphere
Thermal boundary layer
Asthenosphere
Sinking slab of Iapetus oceanic crust and lithosphere
N
S

Analytical Techniques

Zircon fractions were analysed by Isotope Dilution Thermal Ionization Mass Spectrometry (ID-TIMS) at the NERC Isotope Geosciences Laboratory (NIGL), Nottingham, UK and at the Jack Scatterly Geochronology Laboratory, University of Toronto. Analytical procedures were largely as described in Noble et al. (1993). Zircon crystals were separated from <355 μm bulk-rock powder using standard vibrating-table, specific gravity and magnetic techniques. Crystal fractions for analysis were picked under a binocular microscope. Initially a batch of zircons was air abraded (Krogh 1982), but subsequently a modified chemical abrasion technique was used. Bulk zircon fractions were annealed at between 800 and 900 ^{o}C in quartz glass beakers for between 48 and 60 hours. The zircon crystals were ultrasonically washed in 4N HNO_3, rinsed in ultra-pure water, then further washed in warm 4N HNO_3 prior to rinsing with distilled water to remove surface contamination. The annealed, cleaned bulk zircon fractions were then chemically abraded in 200 μl 29N HF and 20 μl 8N HNO_3 between 120 and 180 ^{o}C for 12 hours to minimise or eliminate Pb-loss (technique modified after Mattinson 2005). Chemically abraded zircons were washed several times in ultra-pure water, cleaned in warm 3N HCl for several hours on a hot-plate, rinsed again in ultra-pure water and 8N HNO_3 and split into single grain fractions ready for dissolution. Mixed ^{205}Pb – ^{235}U or ^{205}Pb – ^{233}U – ^{235}U tracers were used to spike all fractions. The first batch of zircons, which was air abraded, underwent chemical separation procedures prior to mass spectrometry. Subsequent chemically abraded batches were not subjected to ion-exchange procedures, but were converted to chloride and loaded onto degassed rhenium filaments in silica gel, following a procedure modified after Mundil et al. (2004). Isotope data at the Jack Scatterly Geochronology Laboratory were collected using a VG-354 equipped with a Daly ion-counter, whereas at NIGL a Thermo Electron Triton equipped with a new generation of MassCom Secondary Electron Multiplier was utilised (Noble et al. 2006). A minimum of 100 ratios were collected for Pb and 60 for U. Pb ratios were scrutinised for any evidence of organic interferences, which were determined to be negligible. Errors were calculated using numerical error propagation (Ludwig 1980). Isotope ratios were plotted using Isoplot version 3 (Ludwig 1993, 2003); error ellipses reflect 2σ uncertainty. Total procedural blanks for three separate batches of chemistry between October 2004 and April 2006 were 2.0 to 0.2 pg for Pb and 0.3 to 0.1 pg for U. Samples were blank corrected using the measured blank ^{204}Pb:^{206}Pb:^{207}Pb:^{208}Pb ratio of 1:18.70:15:15:36.82. Correction for residual common lead above analytical blank was carried out using the Stacey-Kramers common lead evolutionary model (Stacey and Kramers 1975).

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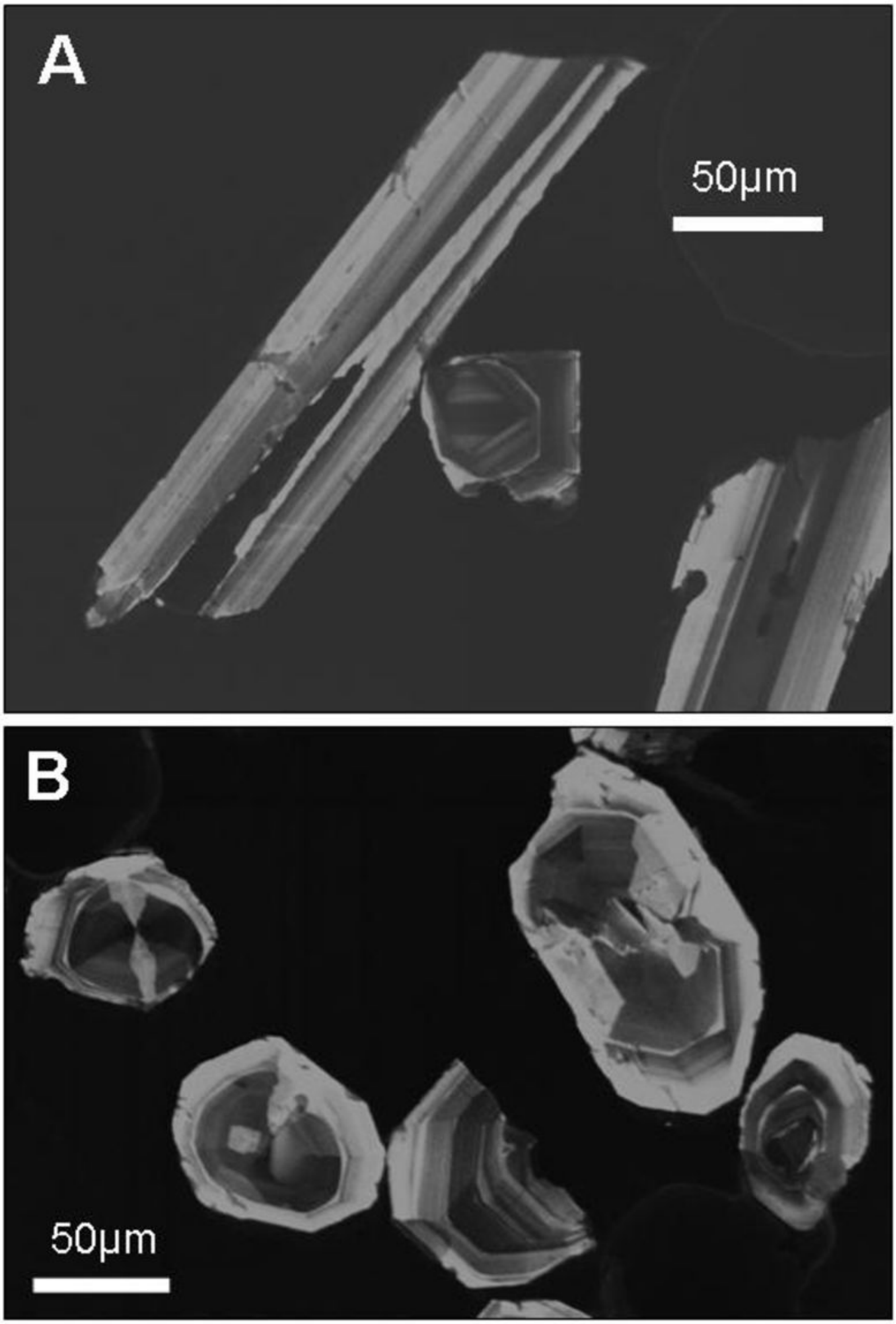
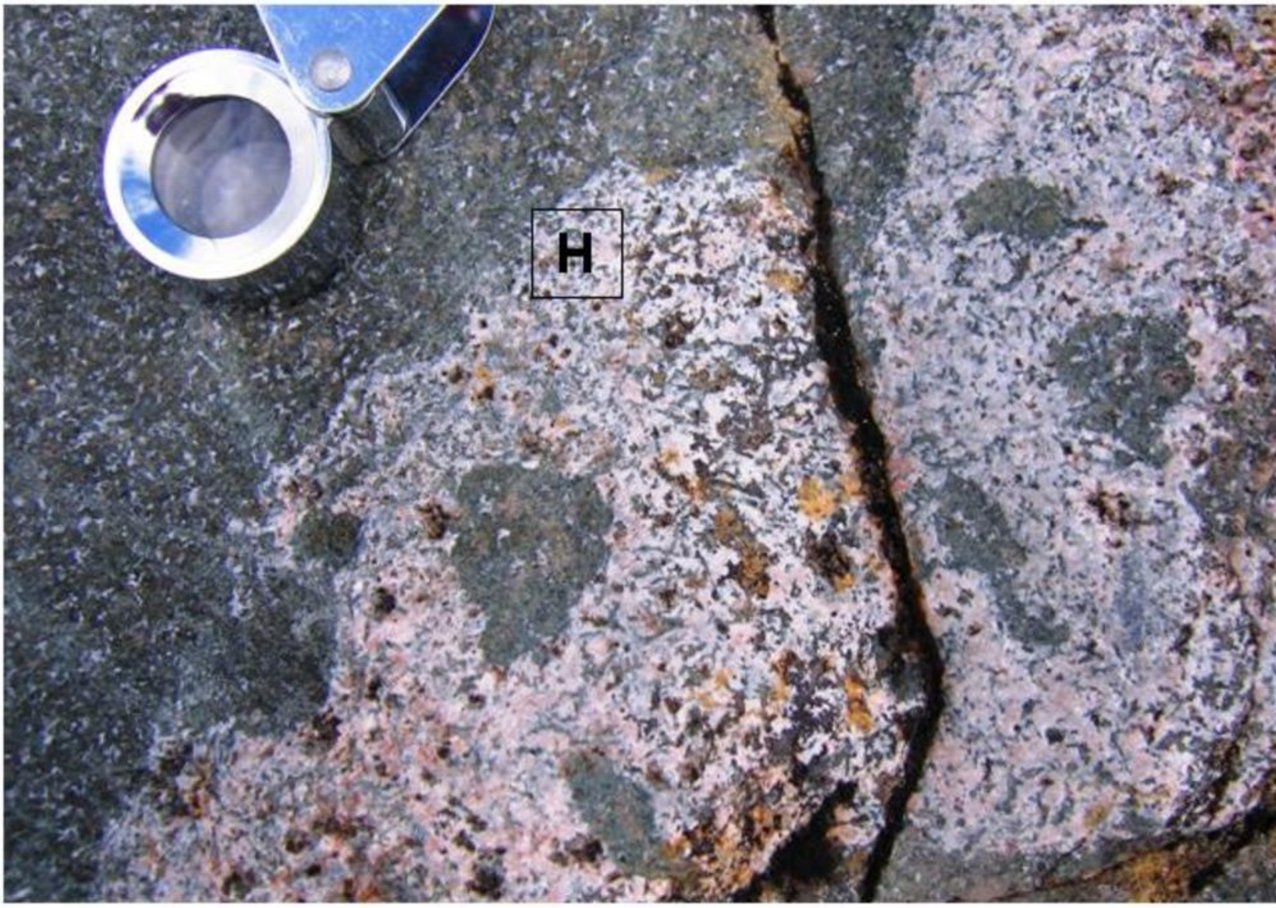
Supp Captions

Supplementary Table 1. *U-Pb Isotope Dilution Thermal Ionisation Mass Spectrometry (ID-TIMS) data*

Supplementary Figure 1. Cathode luminescence images of zircons from Lorn dacite sample JNL86. (a) Representative of the analysed acicular zircons with simple magmatic zoning, c-axis magmatic inclusions, and no inherited cores. (b) Representative of the distinctly more equant zircons with magmatic overgrowth on inherited cores, which were excluded from analysis.

Supplementary Figure 2. Textures due to mingling of appinitic and monzogranitic magmas in the Clashgour intrusion (NN 228 434). Lobate to crenulated contacts and streaky to irregular-shaped inclusions distinguish coexistence as fluids, probably crystal-bearing, while rather 'blurred' contacts, labelled H, indicate local hybridisation.


A
50μm
B
50μm


H


H

Table 1.

Sample	Fraction	Weight (µg)	U (ppm)	Cm-Pb (ppm) ‡	Ratios				
					$^{206}Pb/^{204}Pb$ †	$^{207}Pb/^{206}Pb$*	2σ%	$^{206}Pb/^{238}U$*	2σ%
JNL86 - Lorn	Z1	1.0	328	4.5	210.87	0.05550	0.24	0.06907	0.18
	Z2	4.1	243	1.6	895.18	0.05552	0.14	0.06824	0.19
	Z5	7.6	90	3.3	482.95	0.05503	0.14	0.06895	0.17
	Z6	4.5	163	4.9	408.31	0.05541	0.14	0.06805	0.19
JNL88 - Rannoch Moor	Z5	1.1	1095	0.8	1463.03	0.05519	0.06	0.06769	0.17
	Z6	3.2	635	0.7	2468.14	0.05520	0.06	0.06781	0.16
	Z8	0.6	1820	1.0	946.35	0.05517	0.06	0.06776	0.16
	Z9	0.3	885	3.1	647.62	0.05520	0.06	0.06778	0.16
JNGC76 - Glencoe Fault Intrusion	Z1	1.7	125	2.5	91.41	0.05503	7.81	0.06618	0.53
	Z2	1.6	149	8.7	1639.34	0.05517	0.86	0.06636	0.14
	Z3	2.1	962	12.5	900.90	0.05502	1.08	0.06660	0.16
	Z4	9.4	193	6.9	1732.79	0.05515	1.06	0.06608	0.13
	Z5	1.1	290	10.5	1661.72	0.05514	0.68	0.06686	0.12
	Z6	3.9	307	3.3	6587.27	0.05516	1.52	0.06672	0.16
	Z7	1.1	437	0.3	6582.16	0.05514	0.09	0.06648	0.21
	Z8	1.1	373	8.1	232.64	0.05529	0.22	0.06755	0.17
	Z9	1.3	509	7.2	401.67	0.05506	0.15	0.06681	0.19
	Z10	1.1	415	7.8	259.28	0.05513	0.21	0.06596	0.20
JNE5 - Clach Leathad	Z27	2.8	524	0.9	6731.17	0.05519	0.14	0.06701	0.20
	Z28	0.6	1025	1.3	2061.83	0.05521	0.09	0.06690	0.17
	Z29	1.7	309	0.7	3097.45	0.05518	0.13	0.06683	0.20
	Z30	3.3	313	1.1	4102.95	0.05513	0.07	0.06710	0.16
	Z31	3.0	245	1.9	1630.51	0.05519	0.08	0.06686	0.17
JNE8 - Cruchan	Z33	8.9	761	66.4	361.65	0.05497	0.57	0.06632	0.18
	Z34	4.0	603	1.7	6725.00	0.05512	0.22	0.06447	0.16
	Z35	8.0	550	4.4	4045.31	0.05508	0.12	0.06458	0.22
	Z36	7.0	291	2.8	3018.55	0.05506	0.07	0.06648	0.06
	Z37	6.3	468	1.8	7028.43	0.05510	0.06	0.06644	0.16
	Z38	10.6	3	21.5	4097.55	0.05508	0.07	0.06634	0.16
	Z4	1.4	254	0.4	3439.20	0.05507	0.12	0.06370	0.16
	Z5	2.5	489	2.2	2378.29	0.05540	0.16	0.06600	0.17
JNGC87 - Starav	Z10	7.1	350	3.3	14055.72	0.05500	0.75	0.06534	0.14
	Z41	8.4	452	3.8	12397.28	0.05496	0.50	0.06534	0.16

	Z42	4.5	175	0.9	3819.94	0.05509	0.10	0.06531	0.19

All errors are 2σ (per cent for ratios, absolute for ages)
‡ Total common Pb in analysis, corrected for spike and fractionation (0.09%/amu)
† Measured ratio, corrected for spike and Pb fractionation
* Corrected for blank Pb, U and common Pb (Stacey and Kramers 1975)

			Ages (Ma)					
$^{207}Pb/^{235}U$*	2σ%	Rho	$^{207}Pb/^{206}Pb$	2σ abs	$^{206}Pb/^{238}U$	2σ abs	$^{207}Pb/^{235}U$	2σ abs
0.52849	0.31	0.66	432.3	0.4	430.5	0.8	430.8	1.3
0.52011	0.24	0.81	433.0	3.1	425.4	0.8	426.8	1.0
0.52759	0.22	0.78	413.6	3.1	429.6	0.7	427.1	0.9
0.51921	0.24	0.81	428.8	3.2	424.5	0.8	425.1	1.0
0.51516	0.18	0.94	420.0	1.3	422.2	0.7	421.9	0.8
0.51609	0.17	0.93	420.0	1.4	422.9	0.7	422.5	0.7
0.51540	0.17	0.93	419.1	1.4	422.6	0.7	422.1	0.7
0.51589	0.18	0.91	420.3	1.2	422.8	0.7	422.4	0.7
0.50219	7.82	0.31	413.6	1.8	413.1	2.1	413.2	26.9
0.50481	0.92	0.48	419.2	19.2	414.2	0.6	415.0	3.1
0.50523	1.14	0.42	413.1	24.0	415.6	0.7	415.2	3.9
0.50224	1.14	0.60	418.2	23.7	412.5	0.5	413.4	3.9
0.50833	0.73	0.45	417.9	15.2	417.2	0.5	417.3	2.5
0.50741	1.62	0.69	418.6	33.5	416.4	0.7	416.7	5.5
0.50547	0.23	0.93	418.0	1.9	414.9	0.9	415.4	1.0
0.51495	0.28	0.64	423.8	4.8	421.4	0.7	421.8	1.2
0.50721	0.24	0.80	414.7	3.3	416.9	0.8	416.6	1.0
0.50140	0.30	0.70	417.7	4.7	411.8	0.8	412.7	1.2
0.50991	0.24	0.80	419.8	3.2	418.1	0.8	418.4	1.0
0.50924	0.19	0.89	420.6	1.9	417.5	0.7	417.9	0.8
0.50846	0.24	0.84	419.5	2.9	417.1	0.8	417.4	1.0
0.51011	0.18	0.93	417.7	1.5	418.7	0.7	418.5	0.8
0.50873	0.18	0.91	419.8	1.7	417.2	0.7	417.6	0.8
0.50268	0.62	0.39	411.2	12.7	413.9	0.7	413.5	2.6
0.48991	0.27	0.61	417.0	4.8	402.7	0.6	404.9	1.1
0.49048	0.22	0.88	415.7	2.7	403.4	0.9	405.2	0.9
0.50469	0.18	0.92	414.7	1.5	414.9	0.2	414.9	0.7
0.50481	0.17	0.94	416.4	1.3	414.7	0.7	414.9	0.7
0.50383	0.18	0.93	415.7	1.5	414.0	0.7	414.3	0.7
0.48181	0.18	0.90	414.8	2.8	398.5	0.6	399.4	0.7
0.50282	0.35	0.94	428.4	3.6	412.4	0.7	413.3	1.5
0.49554	0.81	0.50	412.3	16.8	408.0	0.5	408.7	2.7
0.49515	0.55	0.48	410.7	11.0	408.0	0.6	408.4	1.9

0.49427	0.34	0.60	415.7	2.3	407.8	0.8	407.8	1.4