Neoproterozoic to early Paleozoic extensional and compressional history of East Laurentian margin sequences: The Moine Supergroup, Scottish Caledonides

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ABSTRACT

Neoproterozoic siliciclastic-dominated sequences are widespread along the eastern margin of Laurentia and are related to rifting associated with the breakout of Laurentia from the supercontinent Rodinia. Detrital zircons from the Moine Supergroup, NW Scotland, yield Archean to early Neoproterozoic U-Pb ages, consistent with derivation from the Grenville-Sveconorwegian orogen and environs and accumulation post–1000 Ma. U-Pb zircon ages for felsic and associated mafic intrusions confirm a widespread pulse of extension-related magmatism at around 870 Ma. Pegmatites yielding U-Pb zircon ages between 830 Ma and 745 Ma constrain a series of deformation and metamorphic pulses related to Knoydartian orogenesis of the host Moine rocks. Additional U-Pb zircon and monazite data, and $^{40}Ar/^{39}Ar$ ages for pegmatites and host gneisses indicate high-grade metamorphic events at ca. 458–446 Ma and ca. 426 Ma during the Caledonian orogenic cycle.

The presence of early Neoproterozoic siliciclastic sedimentation and deformation in the Moine and equivalent successions around the North Atlantic and their absence along strike in eastern North America reflect contrasting Laurentian paleogeography during the breakup of Rodinia. The North Atlantic realm occupied an external location on the margin of Laurentia, and this region acted as a locus for accumulation of detritus (Moine Supergroup and equivalents) derived from the Grenville-Sveconorwegian orogenic welt, which developed as a consequence of collisional assembly of Rodinia. Neoproterozoic orogenic activity corresponds with the inferred development of convergent plate-margin activity along the periphery of the supercontinent. In contrast in eastern North America, which lay within the internal parts of Rodinia, sedimentation did not commence until the mid-Neoproterozoic (ca. 760 Ma) during initial stages of supercontinent fragmentation. In the North Atlantic region, this time frame corresponds to a second pulse of extension represented by units such as the Dalradian Supergroup, which unconformably overlies the predeformed Moine succession.

INTRODUCTION

The assembly and breakup of Neoproterozoic supercontinents Rodinia and Pannotia led to the development of a series of evolving sedimentary sources and sinks for sediment accumulation (Cawood et al., 2007a, 2007b). Neoproterozoic to early Paleozoic rift and passive-margin successions are well developed around the margins of Laurentia (Bond et al., 1984). They constitute a key piece of evidence that Laurentia lay at the core of the end Mesoproterozoic to early Neoproterozoic supercontinent of Rodinia (Dalziel, 1991; Hoffman, 1991). The rift and passive-margin successions are well preserved within, and their development heralds the initiation of, the Appalachian-Caledonian orogeny, which extends along the east coast of North America, through Ireland, Britain, East Greenland, and Scandinavia (Fig. 1A; Dalziel, 1997; Dewey, 1969; van Staal et al., 1998; Williams, 1984). The character of these successions, in terms of both original depositional lithology and age range, and subsequent orogenic deformational and metamorphic history, varies along and across the orogen and likely reflects variations in nature of basement lithologies, the nature and timing of the continental breakup record, and the timing and drivers of the orogenic record. In particular, and the focus of this paper, the early Neoproterozoic siliciclastic successions in Scotland, along with equivalent units in the East Greenland and Scandinavian Caledonides, record an early cycle of sedimentation and deformation that is absent from the North American Appalachian successions. Our aim is to present new data on provenance and timing of orogenic activity for the Moine Supergroup in Scotland, so as to constrain the development of the northeast Laurentian margin in the Neoproterozoic and to relate along-strike

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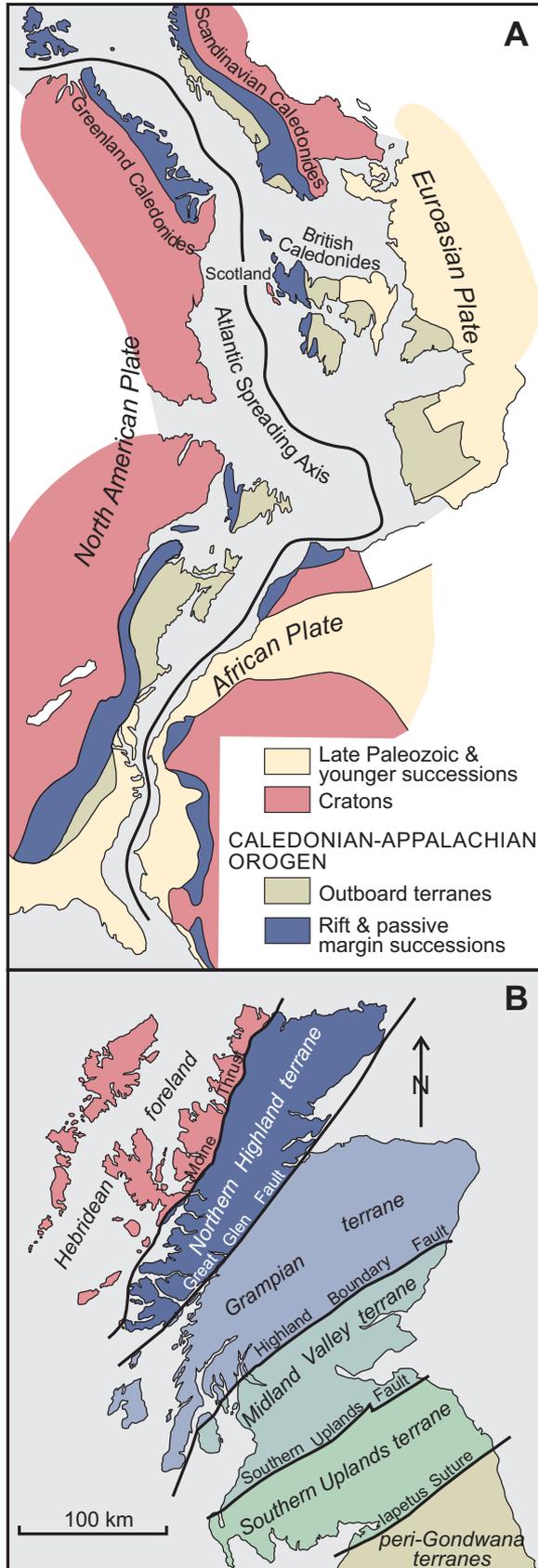
Figure 1. (A) Pangean reconstruction of North America, Europe, and Africa prior to opening of Atlantic (adapted from Williams, 1984). (B) Principal tectonostratigraphic units of northern England and Scotland.

changes in the margin's history to evolving tectonic settings both peripheral and internal to Rodinia.

REGIONAL SETTING

Laurentian rift and passive-margin successions within the Caledonian orogen in Scotland are preserved in three main belts, the Hebridean foreland, the Northern Highlands terrane, and the Grampian terrane (Fig. 1B). In the Hebridean foreland, a little-deformed Cambrian–Ordovician mixed siliciclastic and carbonate passive-margin succession is unconformable on pre-orogen Mesoproterozoic clastic rocks (Torridon, Sleat, and Stoer groups) and Archean to Paleoproterozoic basement of the Lewisian complex. It was this succession that first indicated the Laurentian origin of the foreland (Peach et al., 1907; Salter, 1859). The Northern Highlands terrane, between the Moine thrust and Great Glen fault, contains the Moine Supergroup, a clastic-dominated succession that accumulated during early Neoproterozoic lithospheric extension on the developing margin of Laurentia. The Grampian (or Central Highlands) terrane lies between the Great Glen and Highland Boundary faults and includes equivalents of the Moine Supergroup (Badenoch Group) and the unconformably overlying Dalradian Supergroup, a Neoproterozoic to early Paleozoic rift and passive-margin siliciclastic-dominated succession with some interstratified carbonates as well as bimodal mafic and minor felsic igneous rocks.

The Moine Supergroup (Fig. 2) consists of thick successions of strongly deformed and metamorphosed sedimentary rocks, now represented mainly by metasandstone and metapelite. The sedimentary precursors were deposited in a range of fluvial to marine environments (Bonsor et al., 2010, 2012; Glendinning, 1988; Strachan, 1986). Three lithostratigraphic subdivisions have been recognized, from inferred oldest to youngest, the Morar, Glenfinnan, and Loch Eil groups (Holdsworth et al., 1994; Johnstone et al., 1969; Soper et al., 1998; Strachan et al., 2002, and references therein). The Morar Group is located in the Moine Nappe, whereas the Glenfinnan and Loch Eil groups occur mainly within the structurally overlying Sgurr Beag Nappe. The East Sutherland migmatites, which lie in the Naver Nappe (Fig. 2), are thought to form part of the Moine Supergroup, although their exact lithostratigraphic affinities are uncertain. Tectonically emplaced inliers of Neoarchean ("Lewisianoid") orthogneisses (Fig. 2) are considered to represent the basement on which the Moine sediments were deposited (Friend et al., 2008; Holdsworth, 1989; Ramsay, 1958a; Strachan and Holdsworth, 1988).
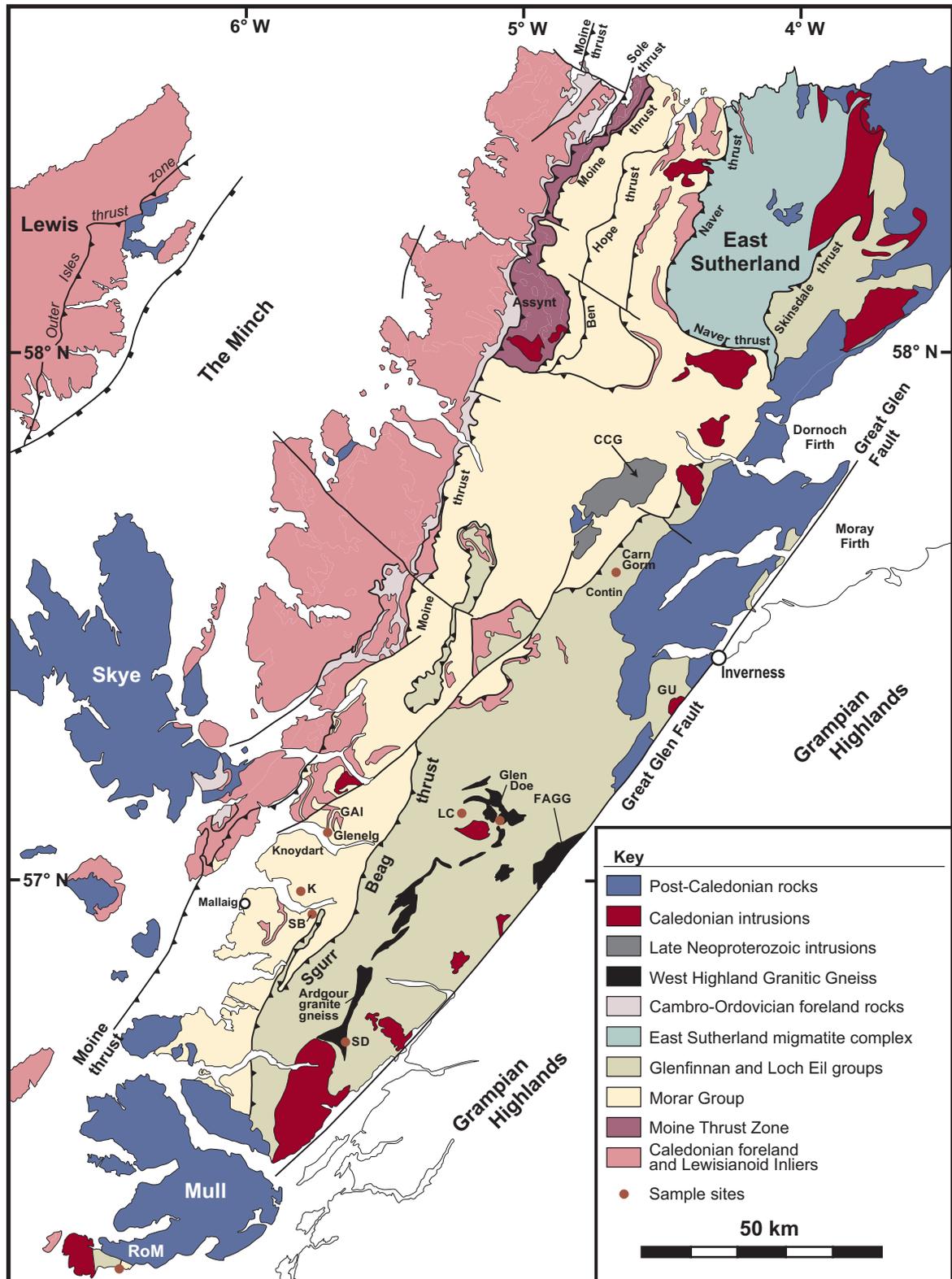
Figure 2. Geological map of northern Scotland (modified from Thigpen et al., 2010) showing the distribution of the Moine Supergroup, consisting of the Morar, Glenfinnan, and Loch Eil groups and East Sutherland migmatites, and the approximate location of analyzed samples (solid pink dot). The Moine Supergroup lies within the Northern Highlands terrane (Fig. 1B), which is separated from the Hebridean foreland to the northwest by the Moine thrust and the Grampian Highlands terrane to the southeast by the Great Glen fault. Abbreviations: CCG—Carn Chuinneag–Inchbae Granite; FAGG—Fort Augustus granite gneiss; GAI—Glenelg-Attadale Inlier; GU—Glen Urquhart; K—Knoydart; LC—Loch Cluanie (including Cruachan Coille a'Chait); RoM—Ross of Mull; SB—Sgurr Breac; SD—Sgurr Dhomhnuill.

Detrital zircons from Moine rocks have yielded a range of Neoarchean to earliest Neoproterozoic U-Pb ages consistent with derivation of sediment from the present-day NE Laurentia sector of Rodinia (Cawood et al., 2004, 2007b; Friend et al., 2003; Kirkland et al., 2008).

The Glenfinnan and Loch Eil groups were intruded at ca. 870 Ma by the igneous protoliths of the West Highland Granitic Gneiss (Fig. 2) and associated metabasic rocks (Dalziel, 1966; Dalziel and Johnson, 1963; Dalziel and Soper, 2001; Friend et al., 1997; Millar, 1999; Rogers et al., 2001). Geochronological evidence indicates that the Moine Supergroup, and likely correlatives east of the Great Glen fault (Badenoch Group), together with the intrusive protoliths of the West Highland Granitic Gneiss and associated mafic rocks, was affected by a series of Neoproterozoic tectonothermal events between ca. 840 Ma and ca. 725 Ma (Long and Lambert, 1963; Piasecki, 1984; Piasecki and van Breemen, 1983; Noble et al., 1996; Highton et al., 1999; Rogers et al., 1998; Vance et al., 1998; Tanner and Evans, 2003; Cutts et al., 2009, 2010). These events are referred to collectively as "Knoydartian" (Bowes, 1968), although the documented duration of >100 m.y. presumably incorporates multiple discrete events rather than a single protracted event (Cawood et al., 2010).

The late Neoproterozoic to early Paleozoic breakup of Pannotia and development of the Iapetus Ocean was associated with the intrusion of granites dated at 600–580 Ma (U-Pb zircon) into the Moine Supergroup (Kinny et al., 2003b; Oliver et al., 2008). East of the Great Glen fault, the depositional history of the mid-Neoproterozoic to Cambrian Dalradian Supergroup reflects progressive rifting and basin deepening on the Laurentian margin of Iapetus (Anderton, 1985). The initial closure of the Iapetus Ocean during the early- to mid-Ordovician (480–470 Ma) was marked by the Grampian orogenic event (Lambert and McKerrow, 1976), equivalent to the Taconic event in the Appalachian orogen (Dewey and Shackleton, 1984; Rogers, 1970). This resulted from collision of the Laurentian margin with a volcanic arc, probably represented by the Midland Valley terrane (Fig. 1B), obduction of ophiolites, and widespread deformation and Barrovian metamorphism of the Moine and Dalradian successions (Chew and Strachan, 2013, and references therein; Cutts et al., 2010; Dewey and Ryan, 1990; Kinny et al., 1999; Oliver, 2001; Rogers et al., 2001; Soper et al., 1999). A further tectonothermal event at ca. 450 Ma has been identified on the basis of Lu-Hf and Sm-Nd dating of metamorphic garnet within the western Moine rocks (Bird et al., 2013). The final stages of closure of the Iapetus Ocean during the Silurian were associated with the collision of Baltica with East Greenland and the segment of Laurentia that contained the Moine Supergroup (the Scottish Promontory of Dalziel and Soper, 2001), resulting in the Scandian orogenic event (Coward, 1990; Dallmeyer et al., 2001; Dewey and Strachan, 2003). This resulted in regional-scale ductile thrusting, folding, and Barrovian metamorphism of the Moine Supergroup, culminating in westward displacement on the Moine thrust zone (Fig. 2) at ca. 430–425 Ma (Freeman et al., 1998; Goodenough et al., 2011; Kinny et al., 2003a; Strachan and Evans, 2008).

The Moine Supergroup was established as an important area for the study of multiple deformation events before geochronological investigations revealed unsuspected complexity in its orogenic history (e.g., Clifford, 1959; Fleuty, 1961; Ramsay, 1958a, 1958b, 1960). Many parts of the Moine Supergroup have been mapped in detail, with deformational sequences interpreted in the context of D_1–D_x events, each of which might be associated with a particular set of tectonic structures (e.g., Brown et al., 1970; Powell, 1974; Tobisch et al., 1970). The isotopic dating of deformed igneous intrusions has proved useful in assigning tectonic structures to certain orogenic events (e.g., Kinny et al., 2003a; Kocks et al., 2006; Strachan and Evans, 2008). However, in other cases, it has been difficult to relate isotopic data obtained from metamorphic porphyroblasts and/or accessory phases to tectonic fabrics, meaning that the age(s) of the latter have remained ambiguous (e.g., Cutts et al., 2010). Furthermore, the relative paucity of modern isotopic data means that the ages of tectonic structures and associated metamorphic assemblages are poorly constrained over large tracts of the Moine Supergroup.

The present study reports new U-Pb zircon data from: (1) Morar Group and possible Glenfinnan Group metasedimentary lithologies, aimed at further evaluating the provenance of these units, and (2) a range of meta-igneous intrusions and felsic melts, aimed at either establishing or refining the ages of protoliths and/or deformation and high-grade metamorphism. In addition, U-Pb monazite and $^{40}Ar/^{39}Ar$ muscovite data were obtained from three samples to place further constraints on the timing of major tectonothermal events.

SAMPLE DESCRIPTIONS AND FIELD RELATIONSHIPS

Metasedimentary Units

Samples of metasandstone and metapelite were obtained from the Moine rocks on the Ross of Mull (Figs. 2 and 3), which have been correlated with the Morar and Glenfinnan groups (Holdsworth et al., 1987). Whereas on the mainland, the two groups are everywhere thought to be separated by the Sgurr Beag thrust (Tanner, 1970), on the Ross of Mull they are interpreted to form a continuous sequence (Holdsworth et al., 1987). To evaluate further the nature of the contact between the two groups at this locality, four samples were collected, in stratigraphic order (Fig. 3): Lower Shiaba Psammite (MG01, sampled at NM 44554 18971 [the location of each sample is linked to a UK National Grid Reference]), Upper Shiaba Psammite (MG04, sampled at NM 42829 18367), Scoor Pelitic Gneiss (RS01–10, sampled at NM 4201 1882), and the Ardalanish Striped and Banded unit (MG03, sampled at NM 39627 18699). Correlation of these units with the Morar and Glenfinnan groups on the mainland is shown in Figure 3. Descriptions of the selected samples are given in the GSA Data Repository.[1] In addition, a sample of the Morar Group (KD07–02) was obtained from the Knoydart Peninsula at NM 79698 96119 (Fig. 2). The unit sampled is the regionally extensive Ladhar Bheinn Pelite (Ramsay and Spring, 1962), thought to be the northern continuation of the Morar Pelite (Holdsworth et al., 1994), and to lie stratigraphically between Lower and Upper Shiaba Psammite samples MG01 and MG04 (Fig. 3; see Data Repository text [footnote 1]).

Meta-Igneous Intrusions: West Highland Granitic Gneiss

Ardgour granitic gneiss (Sgurr Dhomhnuill facies; sample "SD Gneiss"). The West Highland Granitic Gneiss comprises a series of highly deformed and metamorphosed granitic intrusions that mainly occur close to the boundary between the Glenfinnan and Loch Eil groups, and also further east adjacent to the Great Glen fault (Fig. 2; Johnstone, 1975). Barr et al. (1985) concluded that the granitic gneisses represent S-type, anatectic granites derived by partial melting of Moine metasedimentary rocks during regional high-grade metamorphism. In contrast, Dalziel and Soper (2001) and Ryan and Soper (2001) argued that the igneous protoliths were pretectonic and intruded during extensional rifting. Previously reported ages for the gneiss include a thermal ionization mass spectrometry (TIMS) zircon age of 873 ± 7 Ma for a segregation pegmatite from the Ardgour granitic gneiss (Friend et al., 1997) and a U-Pb zircon (secondary ion mass spectrometry [SIMS]) age of 870 ± 30 Ma

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[1]GSA Data Repository item 2014324, analytical methods, cathodoluminescence imaging, is available at http://www.geosociety.org/pubs/ft2014.htm or by request to editing@geosociety.org.

Figure 3. Schematic strati-graphic columns for the Moine succession showing the inferred lithostratigraphic relationships between the thrust sheets depicted in Figure 2 and rock units exposed on the Ross of Mull (after Holdsworth et al., 1987). Also shown is the Moine-equivalent Badenoch Group, which is inferred to be uncon-formably overlain by the Dal-radian succession to the south of the Great Glen fault. Solid black triangles show approxi-mate stratigraphic position of samples analyzed for detrital zircons. WHGG—West High-land granite gneiss.
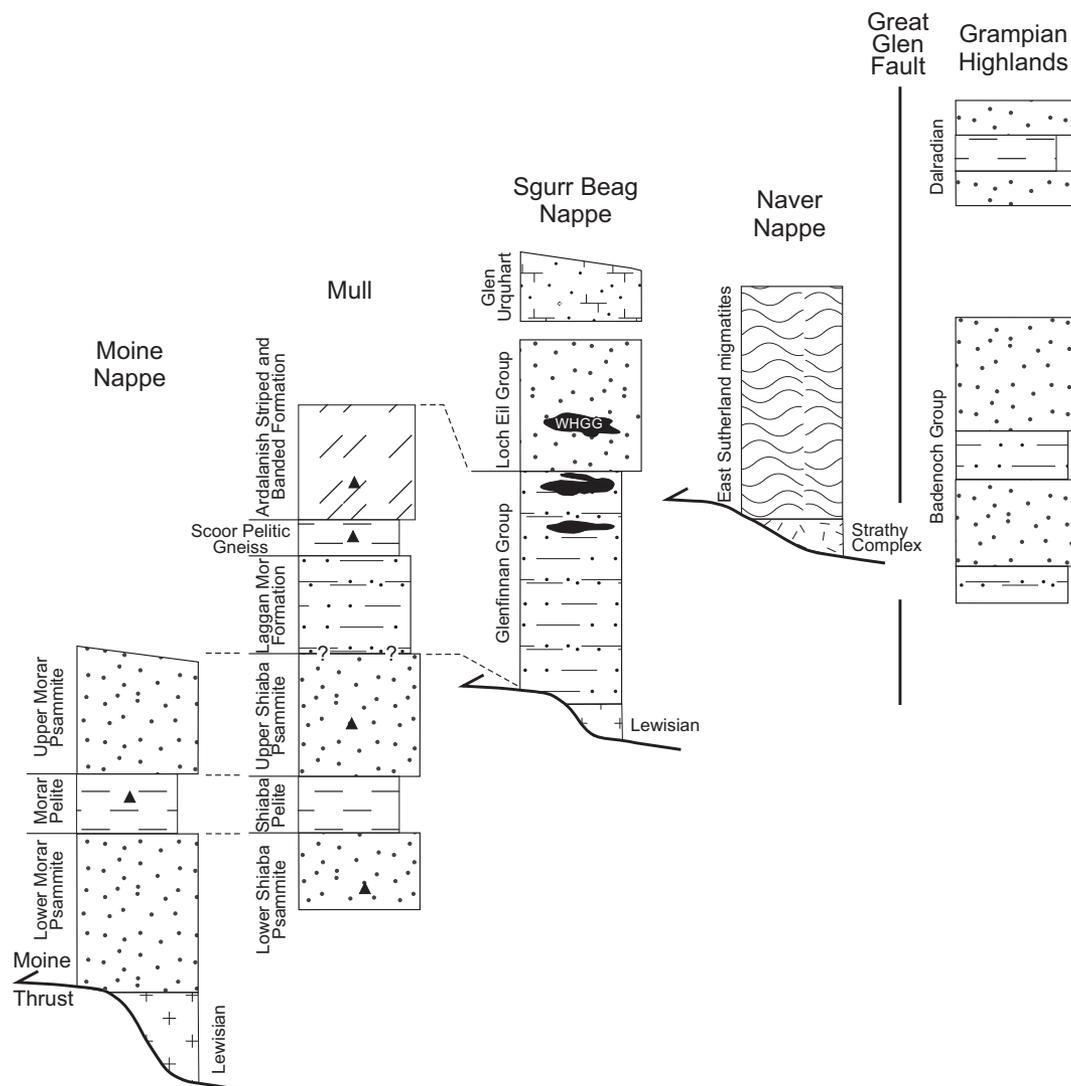
for the Fort Augustus granitic gneiss (Fig. 2; Rogers et al., 2001).

To more accurately ascertain the age of the igneous protolith of the Ardgour granitic gneiss, a sample of nonmigmatitic granitic gneiss was obtained from NM 84677 66189 (Fig. 2). In this southern part of the body, the granitic gneiss commonly contains numerous 2–3 cm K-feldspar augen, the "Sgurr Dhomhnuill" facies of Harry (1953; see Data Repository text [footnote 1]).

Glen Doe granitic gneiss and metagabbro. The Glen Doe granitic gneiss is the northern-most body of the West Highland Granitic Gneiss (Fig. 2) and occurs in association with meta-gabbros and metadolerites (Barr et al., 1985; Millar, 1990; Millar, 1999; Peacock, 1977). Field relationships indicate that the igneous protoliths of the granitic gneisses and the metagabbros were intruded more or less contemporaneously and are pretectonic relative to regional defor-

mation and metamorphism of the host Moine rocks (Dalziel and Soper, 2001; Millar, 1990; Millar, 1999). A U-Pb zircon age of 873 ± 6 Ma obtained from a metagabbro and the mid-ocean-ridge basalt (MORB) affinities of spatially asso-ciated metadolerites are key lines of evidence indicating that the ca. 870 Ma event was domi-nated by extensional rifting and bimodal mag-matism (Dalziel and Soper, 2001; Millar, 1999).

To test the hypothesis that igneous protoliths of the granitic gneisses and the metagabbros were of approximately the same age, four samples were obtained from the River Doe sec-tion (Fig. 2). Two of these were collected from the main body of the granitic gneiss: samples SH-02–18B (NH 21643 12642) and D24 (NH 21162 12661). A third granitic gneiss sample (D48) was a xenolith within metagabbro (NH 21864 12583). The fourth sample (D93) was a foliated metagabbro collected at NH 21848 12598 (see Data Repository text [footnote 1]).

Felsic Melts

Cruachan Coille a'Chait pegmatite (MS07–01). The Cruachan Coille a'Chait pegmatite (Fig. 2) was sampled at NH 11532 11103. It is the largest of a suite of crosscutting, foliated pegmatites intruding interbanded cross-bedded metasandstones and metapelites assigned to Loch Eil Group (Millar, 1990). The pegma-tite is a subvertical sheet, up to 2 m in thick-ness, that cuts obliquely across the subvertical composite S_0/S_1 gneissic banding within host Moine lithologies (Fig. 4A). The S_0/S_1 fabric contains intrafolial isoclinal folds, but these do not deform the contact between the pegmatite and its host rocks. A penetrative S_2 schistos-ity is defined within the pegmatite by aligned muscovite grains that wrap quartz-feldspar aggregates; this fabric is oblique to S_0/S_1 and subparallel to the contact between the pegma-tite and host rocks. The field evidence there-fore suggests that the pegmatite was intruded

Figure 4. Field sketches of pegmatites intruding Moine sedimentary rocks in the vicinity of Loch Cluanie (after Millar, 1990). (A) Cruachan Coille a'Chait pegmatite (NH 1532 11103) exposed on small hill above the shores of the loch. The pegmatite crosscuts an early D1 fold in psammites of the Loch Eil Group and carries a variable but generally well-developed S2 fabric. (B) Pegmatite exposed on shores of loch (NH 10151 11456) intrusive into pelitic gneisses of the Glenfinnan Group.
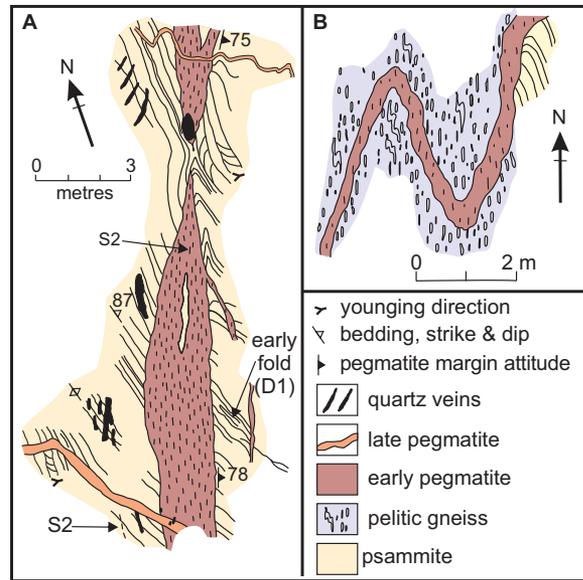
between the D_1 and D_2 deformation events identified in this area.

Knoydart pegmatite (KD07–04). The Knoydart pegmatite was sampled at NM 79707 96103, where it occurs within the Ladhar Bheinn Pelite of the Morar Group (Fig. 2). The first indication that the Moine Supergroup had been affected by Precambrian metamorphism was provided by Rb-Sr muscovite ages of ca. 740 Ma obtained from the pegmatite by Giletti et al. (1961). The Knoydart "pegmatite" consists of a series of concordant sheets and veins of foliated pegmatite, ~2–3 m in length and ~0.5 m thick, within migmatitic pelitic gneiss (Hyslop, 2009b; see Data Repository text [footnote 1]). The pegmatite contains a coarsely developed foliation defined by quartz-feldspar aggregates that is parallel to their margins and a composite $S_0/S_1/S_2$ foliation in host pelitic gneisses. Although the field evidence is not clear-cut, the view adopted here is that the pegmatites formed more or less in situ by recrystallization and segregation during the development of the S_1 migmatitic banding within the host pelitic gneisses (see also Hyslop, 2009b).

Carn Gorm pegmatite (SH-03–04A). The Carn Gorm pegmatite (Fig. 2) was sampled at NH 4388 6289, where it occurs within pelitic gneisses assigned to the Glenfinnan Group (Wilson, 1975). Long and Lambert (1963) reported muscovite Rb-Sr ages from the pegmatite of ca. 747, 721, and 662 Ma. The Precambrian age of the pegmatite was confirmed by van Breemen et al. (1974), who obtained Rb-Sr muscovite ages between 755 Ma and 727 Ma. The pegmatite contains a coarsely developed foliation that is parallel to the margins of the pegmatite and to the S_0/S_1 gneissic fabric in the host rocks (Wilson, 1975). Adjacent to the pegmatite, the host pelitic gneisses contain abundant muscovite, quartz veins, and lenticular quartzofeldspathic segregations (see also Kennedy et al., 1943). Field evidence suggests that the pegmatite formed as a result of in situ recrystallization and segregation during high-grade metamorphism of the host Moine rocks (Hyslop, 1992; Long and Lambert, 1963; van Breemen et al., 1974).

Loch Cluanie pegmatite (C51). The Loch Cluanie pegmatite was sampled at NH10151 11456, where it occurs within pelitic gneisses of the Glenfinnan Group. The pegmatite is a sheet less than 1 m in width and tightly folded by N-S–trending upright F_3 folds (Fig. 4B). At this locality, pelitic gneisses of the topmost Glenfinnan Group were strongly reworked during the D_3 event. However, on the flat limbs of F_3 folds, earlier isoclinal fold structures are preserved in boudinaged psammite horizons, and rare cross-bedding indicates younging upwards and to the east. The pegmatite cuts a composite S_0/S_1 fabric, but its relationship to the pre-F_3 isoclinal fold structures is not seen. The pegmatite appears to carry a much weaker axial-planar S_3 schistosity than the surrounding pelitic gneiss. The field evidence suggests that the pegmatite was intruded before the D_3 event, but its relationships to earlier structures are somewhat ambiguous.

Glenelg pegmatite (SH-03–01C). The Glenelg pegmatite was sampled at Rudha Camas na Caillin (NG 8504 0795) south of Arnisdale (Fig. 2), where Lewisianoid basement has been interfolded isoclinally with Moine psammites of the Morar Group (Ramsay, 2010). This episode of deformation is the earliest to affect the Moine rocks and has therefore been designated "D_1"

(Ramsay, 1958a, 2010). Subsequent "D_2" deformation affected both Lewisianoid basement and its Moine cover and resulted in widespread tight folding on all scales, development of a pervasive "S_2" schistosity and "L_2" lineation, and formation of granitic segregations (Ramsay, 1958a). Granitic segregations are deformed by D_2 folds, but more commonly they appear to have been intruded during the final stages of D_2. These syn- to late-D_2 segregations are typically no more than 10–15 cm thick at maximum and are typically slightly boudinaged, but they are generally undeformed internally and thus do not carry either S_2 or L_2. The sample was obtained from an ~15-cm-thick pegmatitic segregation that occurs as a concordant sheet within Lewisianoid orthogneisses.

ANALYTICAL METHODS

Detrital zircons from the metasedimentary units were analyzed in situ using the high-resolution ion microprobe at Curtin University. To minimize bias, separated detrital zircons were placed randomly on the mount and analyzed sequentially. Zircons from the Ardgour and Glen Doe granitic gneisses, the Glen Doe metagabbros, and the Knoydart and Cruachan Coille a'Chait pegmatites were analyzed using the ion microprobes at Curtin University (sensitive high-resolution ion microprobe [SHRIMP]) and NORDSIM (Cameca 1270). Isotopic dilution–thermal ionization mass spectrometry (ID-TIMS), undertaken at the University of Texas at Austin, was used to investigate zircons from the Ardgour and Glen Doe granitic gneisses and the Glenelg and Carn Gorm pegmatites (the TIMS data set for the latter also including monazite analyses). TIMS analyses and laser-ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) analyses of the Loch Cluanie pegmatite were carried out at the Natural Environment Research Council Isotope Geosciences Laboratory. To further constrain the timing of metamorphic events, muscovite from the Knoydart pelite and the Cruachan Coille a'Chait pegmatite was investigated using the $^{40}Ar/^{39}Ar$ dating technique at the Western Australian Argon Isotope Facility at Curtin University. A detailed outline of analytical methods, cathodoluminescence (CL) imaging, and U-Pb data tables are given in the GSA Data Repository (DR text; Tables DR1–DR6; Fig. DR1 [see footnote 1]). A summary of age results is presented in Table 1.

Results

For the SIMS data, individual concordia ages were calculated using Isoplot (Ludwig, 2003) for the detrital zircon data set. This approach

TABLE 1. SAMPLE NUMBERS, LITHOLOGIC UNIT, GRID REFERENCE, AND AGE OF ANALYZED SAMPLES

Sample no.	Stratigraphic unit	Grid reference	Age (Ma)
MG01	Lower Shiaba Psammite	NM 44554 18971	1011 ± 57
MG04	Upper Shiaba Psammite	NM 42829 18367	1259 ± 35
RS01-10	Scoor Pelitic Gneiss	NM 4201 1882	954 ± 48
MG03	Ardalanish Striped and Banded unit	NM 39627 18699	899 ± 51
KD07-02	Ladhar Bheinn Pelite	NM 79698 96119	997 ± 19
KD07-02	Ladhar Bheinn Pelite	NM 79698 96119	458 ± 4 (muscovite)
SD gneiss	"Sgurr Dhomhnuill" facies of Ardgour granitic gneiss	NM 84677 66189	862 ± 11
SH-02-18B	Glen Doe granitic gneiss	NH 21643 12642	840 ± 23
D24	Glen Doe granitic gneiss	NH 21162 12661	869 ± 8
D48	Glen Doe granitic gneiss (xenolith in metagabbro)	NH 21864 12583	880 ± 7
D93	Metagabbro of Glen Doe granitic gneiss	NH 21848 12598	742 ± 19 (rim)
KD07-04	Knoydart pegmatite	NM 79707 96103	780 ± 6
SH-03-04A	Carn Gorm pegmatite	NH 4388 6289	740 ± 8 (zircon + monazite)
SH-03-04C	Metasedimentary gneiss host to Carn Gorm pegmatite	NH 4388 6289	454 ± 1 (monazite)
MS07-01	Cruachan Coille a'Chait pegmatite	NH 11532 11103	793 ± 24
MS07-01	Cruachan Coille a'Chait pegmatite	NH 11532 11103	426 ± 2 (muscovite)
C51	Loch Cluanie pegmatite	NH10151 11456	830 ± 3
SH-03-01C	Glenelg pegmatite	NG 8504 0795	446 ± 4

Note: Age of sedimentary units is for youngest detrital grain. All ages are concordia U-Pb zircon ages, except Carn Gorm pegmatite and host metasedimentary gneiss ages, which are U-Pb monazite, and Ladhar Bheinn Pelite and Cruachan Coille a'Chait pegmatite, which include both U-Pb zircon and Ar/Ar muscovite plateau ages.

was used because it precludes the practice of selecting the best $^{238}U/^{206}Pb$ or $^{207}Pb/^{206}Pb$ for individual analyses when the data set displays a large range of ages (Ludwig, 1998). Only individual concordia ages with robust statistical parameters are given, with uncertainties at the 2σ level. Pooled concordia ages, intercepts, and weighted averages were calculated using Isoplot (Ludwig, 2003) and are given at the 95% confidence level.

The probability density distribution diagrams of the ages of detrital zircons were constructed from the individual concordia ages and statistical parameters (mean square of weighted deviates [MSWD] and probability of fit [P]) following the approach of Nemchin and Cawood (2005). This involves double weighting of the data based on the probability of concordance and errors highlighting the most concordant data (probability of concordance > 0.05).

U-Pb Zircon Data from Metasedimentary Units

The zircon grains have an average size of 50–200 μm. The grains display rounded morphologies with a length:width ratio in the range 3:1–2:1. Internal structures of the grains, revealed by CL images, are dominated by complex oscillatory zoning (Fig. DR1 [see footnote 1]). Rounded homogeneous central parts of grains that can be readily interpreted as inherited cores are uncommon. Faint zoning and multistage dissolution and growth zoning features are widely observed. In all five samples, the zircons are interpreted as detrital grains that were incorporated into the sedimentary protolith during deposition. Only in one sample (KD07–02) were local thin homogeneous overgrowths observed that could be interpreted as having formed in situ, either by metamorphic overgrowth or fluid-related alteration under amphibolite-facies con-

ditions (Hanchar and Miller, 1993). Three grains in this sample were large enough to allow separate core and rim analyses.

Lower Shiaba Psammite (MG01). In total, 79 analyses from 79 grains yielded ages ranging from ca. 2640 Ma to 1000 Ma (Table DR1 [see footnote 1]). The youngest reliable concordia age was 1011 ± 57 Ma (MSWD = 0.20, P = 0.67). Twenty-one analyses show discordance higher than 10%, including three that are significantly displaced from concordia (Fig. 5). The other analyses plot on or close to the concordia curve but show a spread from ca. 1800 Ma to ca. 1000 Ma. The U and Th concentrations show large variations (U = 36–8630 ppm, Th = 18–3126 ppm, Th/U = 0.13–1.11; Table DR1 [see footnote 1]). None of these variations can be related to age. On the frequency distribution diagram (Fig. 5), a main peak is observed at ca. 1750 Ma, with subordinate peaks at ca. 1600 Ma, 1500 Ma, 1200 Ma, and 1050 Ma.

Upper Shiaba Psammite (MG04). Ages from 64 analyses obtained from 64 grains range from ca. 2870 Ma to 865 Ma (Fig. 5). The youngest reliable concordia age is 1259 ± 35 Ma (MSWD = 2.05, P = 0.15). Twelve analyses have a discordance higher than 10%, with eight significantly displaced from the concordia curve. The other analyses display a cluster between 1800 Ma and 1600 Ma (Fig. 5). U and Th concentrations are rather high and display large variations (18–4536 ppm U and 4–893 ppm Th). Two analyses show low Th/U (0.03–0.04) and U contents higher than 2000 ppm, suggesting postcrystallization resetting or alteration. The ages of these grains do not represent the age of their formation and these data were not included into the density probability plot. As for the previous sample, the chemical characteristics are not correlated with age. On the frequency plot, a dominant population is present at ca. 1750 Ma,

and a minor population is seen at ca. 1500 Ma (Fig. 5).

Scoor Pelitic Gneiss (RS01–10). Ages for 82 analyses from 82 grains range from 2718 Ma to 954 Ma. The youngest reliable individual concordia age yielded 954 ± 48 Ma (MSWD = 3.10, P = 0. 10). Ten analyses show discordance higher than 10%, but the data plot close to the concordia diagram, forming two distinct groups. A small cluster of data is observed at 2600 Ma, with the remaining data showing a large spread from 1800 Ma to 1000 Ma (Fig. 5). U and Th concentrations are moderate (U = 12–976 ppm and Th = 8–543 ppm) compared to the other samples. Only one analysis shows high U and Th concentrations (3025 ppm and 1623 ppm, respectively) and yields a reliable concordia age of 1433 ± 13 Ma (MSWD = 0.14, P = 0.71). On the frequency plot (Fig. 5), a predominant population is present at 1800 Ma, and minor peaks are at 2600 Ma, 1600 Ma, 1400 Ma, and at ca. 1100 Ma.

Ardalanish Striped and Banded Formation (MG03). Detrital ages for 68 analyses from 68 grains yielded ages ranging from ca. 1880 Ma to 490 Ma. Twenty analyses are >10% discordant. The individual analyses form two groups on the concordia diagram (Fig. 5). The older group spans 1800–1400 Ma, and the younger spans ca. 1200 Ma to 900 Ma. The U concentrations of the grains range from 45 to 2087 ppm, and the Th content ranges from 13 to 806 ppm. Two analyses are distinct from these two groups, plotting at ca. 500 Ma and 700 Ma. The younger analysis is from the rim of a grain, which has a low Th content (1 ppm) and a Th/U ratio of 0.01, unusual for zircon (Rubatto, 2002). As with the Upper Shiaba Psammite data, this rim might have been altered by postcrystallization processes and has not been included in the density plot. The ca. 700 Ma analysis has a discordance

Figure 5. (Left column) Concordia diagrams based on [204]Pb-corrected zircon U-Pb data for samples of psammitic and pelitic rocks from the Ross of Mull and Knoydart. Error ellipses are shown at the 2σ level. Analytical data are available from the GSA Data Repository (see text footnote 1); *n* = number of zircon analyses in each sample. (Right column) Probability density distribution diagrams for detrital zircon data from psammitic and pelitic samples from the Ross of Mull and Knoydart. Dashed vertical lines separate boundaries of the Paleoproterozoic, Mesoproterozoic, and Neoproterozoic and are taken at 1600 Ma, 1000 Ma, and 545 Ma. Solid gray line highlights 1750 Ma age. The small arrows indicate the age of the youngest detrital zircon; *n* = number of zircon analyses in each sample. Analytical data are available from the GSA Data Repository (see text footnote 1).
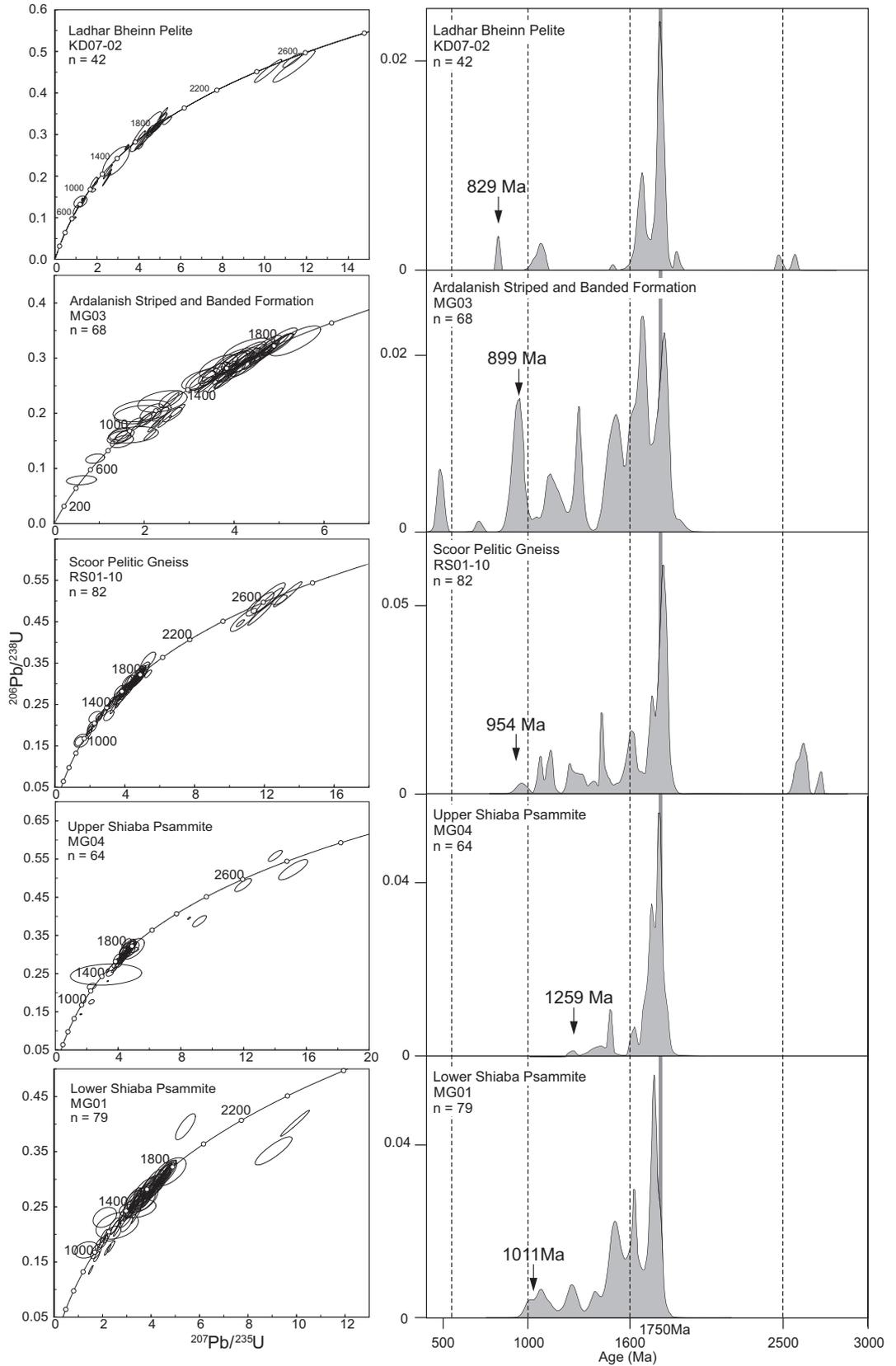
>10% and poor statistical parameters (Table DR1 [see footnote 1]). The youngest reliable individual concordia age obtained was 899 ± 51 Ma (MSWD = 0.90, *P* = 0.34). No correlation between the chemical composition and age was observed. The frequency diagram (Fig. 5) shows two main peaks at 1800 Ma and ca. 1650 Ma and four minor peaks at ca. 1500 Ma, 1300 Ma, 950 Ma, and ca. 500 Ma.

Ladhar Bheinn Pelite (KD07–02). Forty-two analyses were obtained from 39 zircon grains. Nine grains have a U concentration higher than 1000 ppm (Table DR1 [see footnote 1]). The ages range from 2687 Ma to 616 Ma, with the latter age obtained from the inner part of a rim and yielding poor statistical parameters. Plotted on a concordia diagram and a frequency density plot (Fig. 5), the data set displays age clusters at 2600–2400 Ma, 1800–1600 Ma, 1200–1000 Ma, and ca. 850 Ma. The youngest age population at ca. 850 Ma is defined by only three analyses, with only one reliable individual concordia age yielding 829 ± 8 Ma (MSWD = 2.71, *P* = 0.10). These data were obtained from rims or external parts of cores and are consistent with crystallization of zircon during mid-Neoproterozoic Knoydartian metamorphism (Cutts et al., 2010; Vance et al., 1998).

U-Pb Zircon Data from Meta-Igneous Intrusions

Ardgour granitic gneiss (Sgurr Dhomhnuill). Using TIMS data, nine zircon fractions and two monazite fractions were analyzed (Table DR2 [see footnote 1]). The Z4 fraction was a single acicular grain, the typical morphology of magmatic origin (Dalziel, 1963), and Z3 was a euhedral prism with concentric (magmatic) zonation in CL. Three analyses of single or small multigrain zircon fractions and two analyses of multigrain monazite fractions constitute a well-defined discordia yielding upper- and lower-intercepts ages of 862 ± 11 Ma and 433 ± 3 Ma (2σ; Fig. 6A), respectively, with robust statistics (MSWD = 0.44, *P* = 0.72). Fraction Z2, which lies slightly below the discordia line, apparently included a small inherited core or experienced minor Pb loss. Since the upper intercept is defined by the fractions Z3 and Z4, which show magmatic features, this age at 862 ± 11 Ma is interpreted as the crystallization age of the magmatic protoliths. The lower intercept is well constrained by two overlapping analyses of monazite that yield a concordia age of 433 ± 1 (MSWD = 0.00056, *P* = 0.98). Because monazite is a typical metamorphic mineral, the age at 433 Ma ± 3 Ma is interpreted to be related to a metamorphic event. Four fractions show ages older than 862 Ma (fractions Z5, Z6, Z7, and Z8). CL images of these grains (before dissolu-

tion) reveal rounded cores with euhedral overgrowths. Three fractions, Z5, Z7, and Z8, define a poorly correlated discordia line, yielding an upper-intercept age at 1748 ± 93 Ma (MSWD = 9.8, *P* = 0, 2σ; Fig. 6A) when anchored at 862 ± 11 Ma. The Paleoproterozoic intercept and the presence of rounded zircon cores support the interpretation of inheritance from the Moine metasedimentary host rocks (see also Friend et al., 1997; Rogers et al., 2001; Table DR1 [see footnote 1]).

Using SIMS data, 28 grains were analyzed at Curtin University (Table DR1 [see footnote 1]). They yielded ages ranging from ca. 1710 Ma to 620 Ma. The data scatter along and below the concordia curve, and a set of points defines a discordia line with an upper intercept at 1716 ± 120 Ma and a lower intercept at 831 ± 36 Ma but with poor statistics (Fig. 7; MSWD = 2.80, *P* = 0). Only three cores were investigated and yielded reliable concordia ages of 864 ± 55 Ma (MSWD = 0.001, *P* = 0.98), 1132 ± 64 Ma (MSWD = 0.25, *P* = 0.62), and 1709 ± 47 Ma (MSWD = 0.077, *P* = 0.78). The 25 analyses from grain rims are scattered on the concordia curve (Fig. 7) but fail to define a discordia with reliable intercepts. However, the data cluster around 800 Ma and yield a concordia age of 853 ± 11 Ma (95% confidence, MSWD = 2.50, *P* = 0.12). This is within error of the TIMS age reported previously, which is taken as the age of crystallization of the magmatic protolith.

Glen Doe granitic gneiss and metagabbro. TIMS (Texas) and SIMS (Curtin SHRIMP) data obtained from sample SH-02–18B are of uncertain significance (Tables DR1 and 2 [see footnote 1]). TIMS analyses of five zircon fractions or single grains (Z1, Z5, Z6, Z7, Z8) yielded an upper intercept of 798 ± 15 Ma and a lower intercept of 432 ± 65 Ma (MSWD = 1.90, *P* = 0.13; Fig. 6B). These could correspond, respectively, to the protolith crystallization age and major disturbance of isotopic systems during the Caledonian orogeny. Similarly, TIMS fractions Z2, Z3, and Z4 lie above discordia line defined by fractions Z1 and Z5–Z8 (Fig. 6B), consistent with a mid-Neoproterozoic event causing zircon growth or partial resetting. TIMS fractions Z9, Z10, and Z11 suggest incorporation of inherited cores.

SHRIMP analyses were carried out on zircon cores and rims from the same sample (Table DR2 [see footnote 1]). The rims are scattered along and above the concordia curve defining a discordia line with reliable lower and upper intercepts at 585 ± 160 Ma and 736 ± 64 Ma (MSWD = 0.61, *P* = 0.79), respectively (Fig. 8). The cores yielded a reliable concordia age of 840 ± 23 Ma (95% confidence, MSWD = 2.20, *P* = 0.14; Fig. 8).

Additional analytical data were obtained from SIMS (NORDSIM) analyses of granitic gneiss samples D24 and D48 (Table DR3 [see footnote 1]; Figs. 9A and 9B). Zircons from both samples form euhedral prisms with typical igneous CL zonation. Fifteen grains were analyzed from granitic gneiss sample D24. Of these, five analyses were discordant or showed evidence for Pb loss. The remaining 10 analyses yielded a concordia age of 869 ± 8 Ma (MSWD = 0.084, *P* = 0.77). Fifteen grains were also analyzed from sample D48, a granitic xenolith within metagabbro at Glen Doe. All 15 analyses yield a concordia age of 880 ± 7 Ma (MSWD = 0.53, *P* = 0.82). The SIMS zircon ages for D24 and D48 therefore overlap within analytical uncertainty. They are interpreted as dating crystallization of the magmatic protolith, consistent with the protolith ages for the Ardgour granitic gneiss of 862 ± 11 Ma (this paper) and 873 ± 7 Ma (Friend et al., 1997).

Zircons from deformed metagabbro sample D93 show igneous zonation under CL but frequently have thin, CL-bright rims. Fourteen spots were analyzed from this sample (Table DR3 [see footnote 1]) and show a spread of ages (^{206}Pb/^{238}U ages between 878 Ma and 731 Ma). Older analyses are obtained from grain cores; however, these show a range in ^{206}Pb/^{238}U ages suggesting either variable Pb loss, or mixing of core and rim during ablation. Six analyses from overgrowths yield a concordia age of 742 ± 19 Ma (MSWD = 0.037, *P* = 0.85; Fig. 9C). This is interpreted as dating Knoydartian deformation and metamorphism of the metagabbro.

U-Pb Zircon and Monazite Data from Felsic Melts

Cruachan Coille a'Chait pegmatite (MS07–01). Forty-two grains were extracted from this sample, and 51 analyses were made by SHRIMP at Curtin University (Table DR1 [see footnote 1]). Zircons display an elongated and prismatic shape and are dark brown in color, typical of strongly metamict grains, and they are black in CL. The average size of the grains is approximately ~150 μm, but the largest reach 800 μm in length. The average length:width ratio of the grains is ~5:1.

Plotted on the concordia diagram, the data display a large spread. The obtained ages range from 988 Ma to 371 Ma, but only 18 analyses are concordant (discordance lower than 10%). Among the discordant data, 20 of them are reverse discordant (discordance lower than –10%; Table DR1 [see footnote 1]). Note that the data form a cluster at ca. 860 Ma (Fig. 10A).

The zircons are metamict with high U and low Th contents and display reverse discordance and Pb loss leading to discordance (Fig. 10A). As a
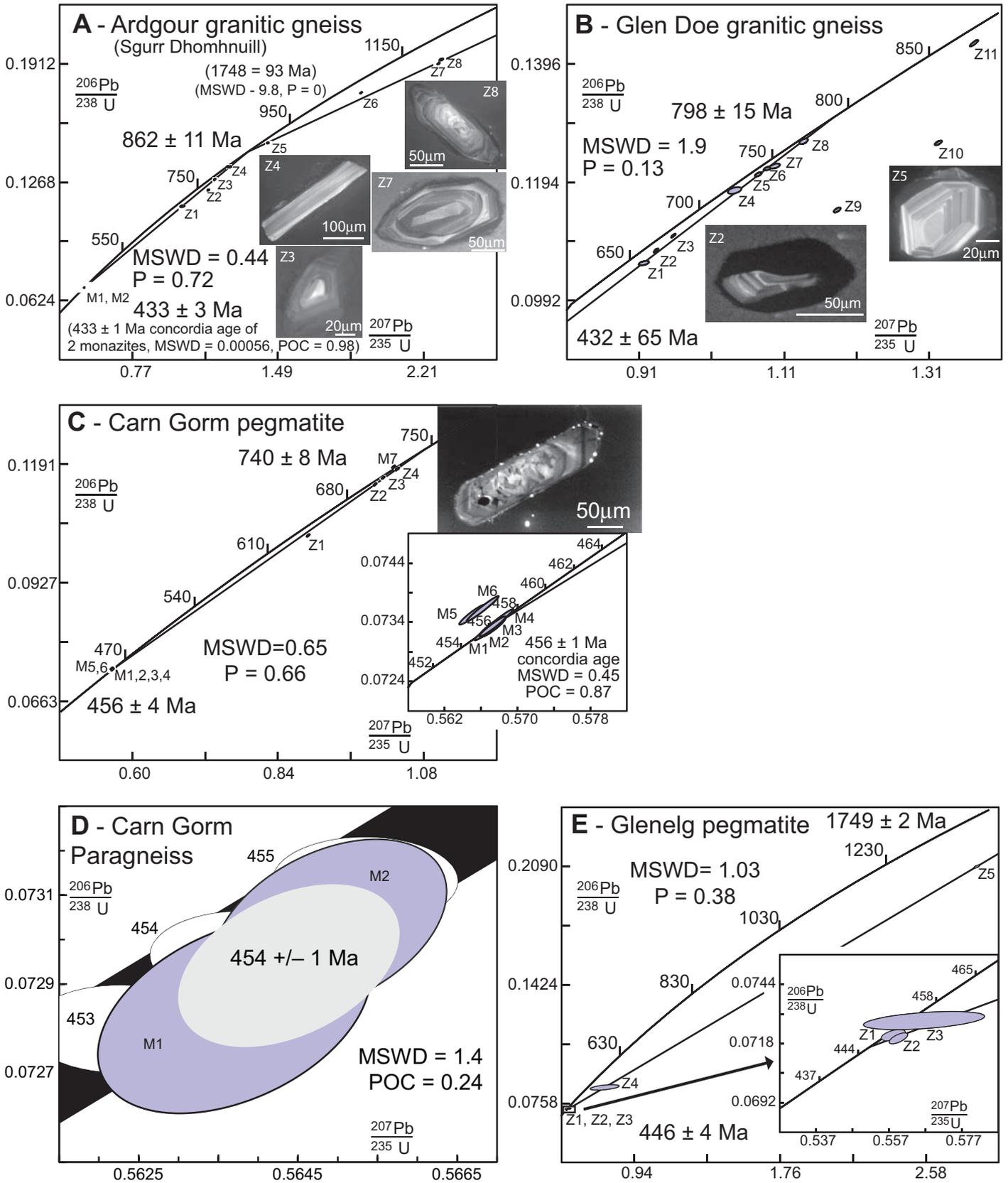
Figure 6. Concordia plots for thermal ionization mass spectrometry (TIMS) data from (A) Ardgour granitic gneiss (Sgurr Dhomhnuill), (B) Glen Doe granitic gneiss (sample SH-02–18B), (C) Carn Gorm pegmatite, (D) Carn Gorm paragneiss, (E) Glenelg pegmatite. *P*—probability of fit; POC—probability of concordance and equivalence; MSWD—mean square of weighted deviates.

Figure 7. (A) Inverse concordia plot of secondary ion mass spectrometry (SIMS) data for Ardgour granitic gneiss. (B) Enlarged section of the concordia curve highlighting analyses of grain rims. *P*—probability of fit; MSWD—mean square of weighted deviates.

them are reverse discordant (discordance lower than –10%; Table DR1 [see footnote 1]). As with MS07–01, zircons are metamict and have a high U concentration, which is likely responsible for the reverse discordance and the scattering of the data (Fig. 10C). The $^{238}U/^{206}Pb$ ratios are likely biased; hence, we applied the same approach as for sample MS07–01. The younger population yielded a weighted average age of 462 ± 21 Ma (95% confidence, *N* = 4, MSWD = 0.92, *P* = 0.43), and the older population yielded an age of 780 ± 6 Ma (95% confidence, *N* = 9, MSWD = 0.69, *P* = 0.70). This latter age is interpreted as the age of crystallization of the pegmatite.

Carn Gorm pegmatite (SH-03–04A). Zircons are euhedral and acicular, with cloudy, significantly altered cores and pristine bipyramidal overgrowths. CL imaging of polished grains revealed oscillatory concentric zonation in the overgrowths. The shape and zoning of the grains suggest that the overgrowths formed during crystallization of the pegmatite. Euhedral tips were broken off several acicular grains and analyzed by TIMS (Table DR2 [see footnote 1]). Several monazite fractions were also analyzed. Monazite fractions (M1, M2, M3, M4) define a concordia age of 456 ± 1 Ma (MSWD = 0.45, *P* = 0.87 concordance and equivalence; Fig. 6C). The combined zircon and monazite data (except reversely discordant monazite fractions M5, M6, M7) define a well-correlated line with an upper intercept of 740 ± 8 Ma and a lower intercept of 456 ± 4 Ma (MSWD = 0.65, *P* = 0.66). As the zircon fractions fall near the upper intercept, ca. 740 Ma is interpreted to be the crystallization age. This age is consistent with the published Rb-Sr age of 730 ± 20 Ma obtained from large muscovite books within the pegmatite (van Breemen et al., 1974). Several monazite fractions lie near the upper and lower intercepts and are reversely discordant, likely due to the unsupported ^{206}Pb produced from high initial ^{230}Th concentrations. Four monazite fractions overlap at the lower intercept, which is interpreted to indicate the timing of metamorphism associated with the deformation of the pegmatite during the Caledonian orogenesis. Two monazite fractions from a sample of the host metasedimentary gneiss (SH-03–04C) collected adjacent to the pegmatite (NH 4379 6297) also yielded a U-Pb concordia age of 454 ± 1 Ma (MSWD = 1.4, *P* = 0.24 concordance and equivalence; Fig. 6D), similarly interpreted to correspond to Caledonian metamorphism.

Loch Cluanie pegmatite (C39). ID-TIMS and LA-ICP-MS data were obtained at the NERC Isotope Geosciences Laboratory (Tables DR4 and DR5 [see footnote 1]). Zircons are elongate prisms between 100 and 200 μm in length, with aspect ratios between 5:1 and 10:1.

consequence of the differing matrix behavior of the sample and standard, the $^{238}U/^{206}Pb$ ratio is likely biased, and so must be any discordia line or individual concordia age derived from it. A much more robust approach to determining the age of the sample is to use $^{207}Pb/^{206}Pb$ data, which do not suffer from these effects. Therefore, we selected the data with a $^{207}Pb/^{206}Pb$ ratio of 0.068, which form the cluster at 860 Ma, and these analyses yielded an age of 865 ± 4 Ma (95% confidence, *N* = 14, MSWD = 0.76, *P* = 0.70; Fig. 10B). This age is, however, inconsistent with structural relations outlined previously (see also Fig. 4) that indicate the pegmatite is associated with the Knoydartian orogenic event, which to date has yielded ages in the range 840 Ma to 725 Ma. The discordia line through the data

set yields intercepts at 793 ± 24 Ma and 383 ± 46 Ma (MSWD = 19, *P* = 0). The age of the upper intercept is reasonable on the basis of regional relations but is not statistically reliable.

Knoydart pegmatite (KD07–04). SIMS data were collected at Curtin University (Table DR1 [see footnote 1]). Zircons display prismatic or subprismatic shapes. They commonly show metamict features, are black in CL, and have high U concentrations (1574–7874 ppm). Sixteen analyses were made in 16 grains, yielding ages ranging from ca. 1017 Ma to 409 Ma (Fig. 10C). The data display two clusters at ca. 800 Ma and 500 Ma, the former including the majority of the analyses. Four analyses including the oldest age are very discordant (percentage of discordance higher than 10%). Two of

Figure 8. Inverse concordia plot of secondary ion mass spectrometry (SIMS) data for the Glen Doe granitic gneiss (SH-02–18B). (B) Enlarged section of the concordia curve showing analyses from grain cores.

an upper-intercept age of 1749 ± 12 Ma, which might reflect the contribution of inherited cores. Storey et al. (2004) mentioned a 1750 Ma U-Pb zircon age from the nearby western portion of the Glenelg-Attadale Inlier (Fig. 2).

$^{40}Ar/^{39}Ar$ Muscovite Data

Muscovites from two samples were analyzed using the $^{40}Ar/^{39}Ar$ technique to place constraints on the late cooling history of the region (Table 2; Table DR6 [see footnote 1]). Muscovites from sample KD07–02 (Ladhar Bheinn Pelite) yielded a statistically robust plateau age of 457.9 ± 3.5 Ma (MSWD = 0.6, *P* = 0.8; Fig. 11A). This is interpreted to reflect cooling following Ordovician (Grampian) metamorphism of the Ladhar Bheinn Pelite. In contrast, muscovites extracted from sample MS07–01 (Cruachan Coille a'Chait pegmatite) yielded a plateau age of 425.9 ± 1.9 Ma (MSWD = 0.5, *P* = 0.94; Fig. 11B). This age is interpreted to reflect cooling following Silurian (Scandian) metamorphism of the pegmatite and its host Moine rocks.

DISCUSSION

Provenance and Age of the Moine Supergroup

Detrital zircons from the Moine successions on Mull and at Knoydart (Figs. 2 and 3) range in age from Archean to early Neoproterozoic, with most grains yielding late Paleoproterozoic and a range of early to late Mesoproterozoic ages (Fig. 5). The overall age range and distribution of specific age peaks are similar to those obtained from previous analyses of Moine metasedimentary rocks (Cawood et al., 2004; Friend et al., 2003; Kirkland et al., 2008), as well as broadly coeval successions around the North Atlantic (Cawood et al., 2007b, 2010). The overall detrital zircon age signatures of the Moine samples, in combination with the south-to-north paleoflow of the southern Morar and Loch Eil groups (Glendinning, 1988; Strachan, 1986), are consistent with derivation from the Grenville-Sveconorwegian orogen and environs. In the late Mesoproterozoic to early Neoproterozoic, this source region formed a major mountain belt and supplied detritus along the axis of the Moine basin (Cawood et al., 2010).

The distribution and frequencies of ages within the samples analyzed from the Moine succession display a number of temporal trends. Archean grains are only present in minor quantities (<10%) in the Morar and lower Glenfinnan groups and have not yet been recorded in samples from the upper Glenfinnan, Loch Eil, and Glen Urquhart successions, and they are also absent from Moine-equivalent strata (Badenoch Group)

They are colorless, and no cores were observed. Three grains were analyzed by chemical abrasion (CA) ID-TIMS (Fig. 9D, inset). Despite having undergone chemical abrasion, the resulting analyses show evidence of minor Pb loss, lying on a normal discordia with an upper intercept of 830 ± 3 Ma (MSWD = 0.046). Twenty-three analyses of zircon grains using laser-ablation single-collector ICP-MS plot on or close to concordia (all are <5% discordant). However, the data again fall on a trend indicating minor Pb loss. A concordia age defined by the seven grains with least apparent Pb loss gives an age of 827 ± 5 Ma (MSWD = 3.1), within error of the ID-TIMS age. The best estimate of the age

of the Loch Cluanie pegmatite is given by the ID-TIMS age of 830 ± 3 Ma.

Glenelg pegmatite (SH-03–01C). Five zircon (single grains and multigrains) fractions were analyzed by TIMS (Table DR2 [see footnote 1]). Fractions Z1 and Z3 are euhedral tips (igneous overgrowths) that were broken off of larger grains and analyzed separately. Fractions Z4 and Z5 are analyses of whole grains. Z4 is a single grain, and Z5 contains three small grains. Zircon fractions Z1, Z2, and Z3 plot close to or on the concordia curve (Fig. 6E). Together with the two other fractions, they define a discordia line with a lower-intercept age of 446 ± 4 Ma, which can be interpreted as the crystallization age, and

Figure 9. Data for Glen Doe granitic gneisses and metagabbro, and Loch Cluanie pegmatite: (A) concordia plot of granite gneiss sample D24, (B) concordia plot of granite gneiss sample D48, (C) concordia plot of foliated metagabbro sample D93, and (D) concordia plot of Loch Cluanie pegmatite sample C39. Analyses of Glen Doe zircons were undertaken by secondary ion mass spectrometry (SIMS), whereas Loch Cluanie zircons were analyzed by inductively coupled plasma–mass spectrometry (ICP-MS; main diagram) and thermal ionization mass spectrometry (TIMS; inset). Error ellipses are shown at 2σ level. *P*—probability of fit; MSWD—mean square of weighted deviates.

east of the Great Glen fault (Fig. 12). Late Paleoproterozoic detritus is particularly prevalent in the Morar and Glenfinnan groups, with prominent peaks at around 1750 Ma and 1650 Ma. Mesoproterozoic ages include peaks at around 1550–1500 Ma and a range of ages between 1300 and 1000 Ma (Figs. 6 and 12). The proportion of Mesoproterozoic detritus increases, relative to Paleoproterozoic detritus, in the younger units. The East Sutherland migmatites within the Naver Nappe also contain a relatively high proportion of Mesoproterozoic detritus (Fig. 12), indicating derivation from a similar source and/or temporal equivalence to the upper Glenfinnan and younger units of the Moine succession.

In the Morar Group, the youngest grains yield ages of ca. 1011 and ca. 1259 Ma (this paper),

ca. 1070 Ma and 1022 Ma (Kirkland et al., 2008), ca. 1032 Ma (Friend et al., 2003), and ca. 980 Ma (Peters, 2001). In the Glenfinnan Group, the youngest grains are ca. 954 Ma and ca. 899 Ma (this paper), ca. 1009 Ma (Kirkland et al., 2008), ca. 947 Ma (based on inherited detrital grains in West Highland Granitic Gneiss; Friend et al., 2003), and ca. 917 Ma (Cutts et al., 2010). In the Loch Eil and Glen Urquhart successions, the youngest grains are ca. 962 Ma and ca. 883 Ma, respectively (Cawood et al., 2004), and in the Badenoch Group, the youngest grain is ca. 900 Ma (Cawood et al., 2003). The youngest detrital grain in the East Sutherland migmatitic succession of the Moine is ca. 926 Ma (Kinny et al., 1999). These results suggest that the Morar Group accumulated after 1000–980 Ma, with a

significant time break (>50 m.y.) prior to deposition of the Glenfinnan and Loch Eil groups after 900–880 Ma. However, our samples collected across the apparently continuous stratigraphic contact between the Morar and Glenfinnan groups on the Ross of Mull (Holdsworth et al., 1987) do not indicate any marked difference in provenance. The increase in Mesoproterozoic detritus and early Neoproterozoic grains in the Glenfinnan Group may simply represent evolution of the Grenville-Sveconorwegian source region to include younger phases of the orogenic welt. Nonetheless, the possibility still remains that a fundamental temporal break between these two successions has been obscured at this locality by the effects of deformation and metamorphic recrystallization.
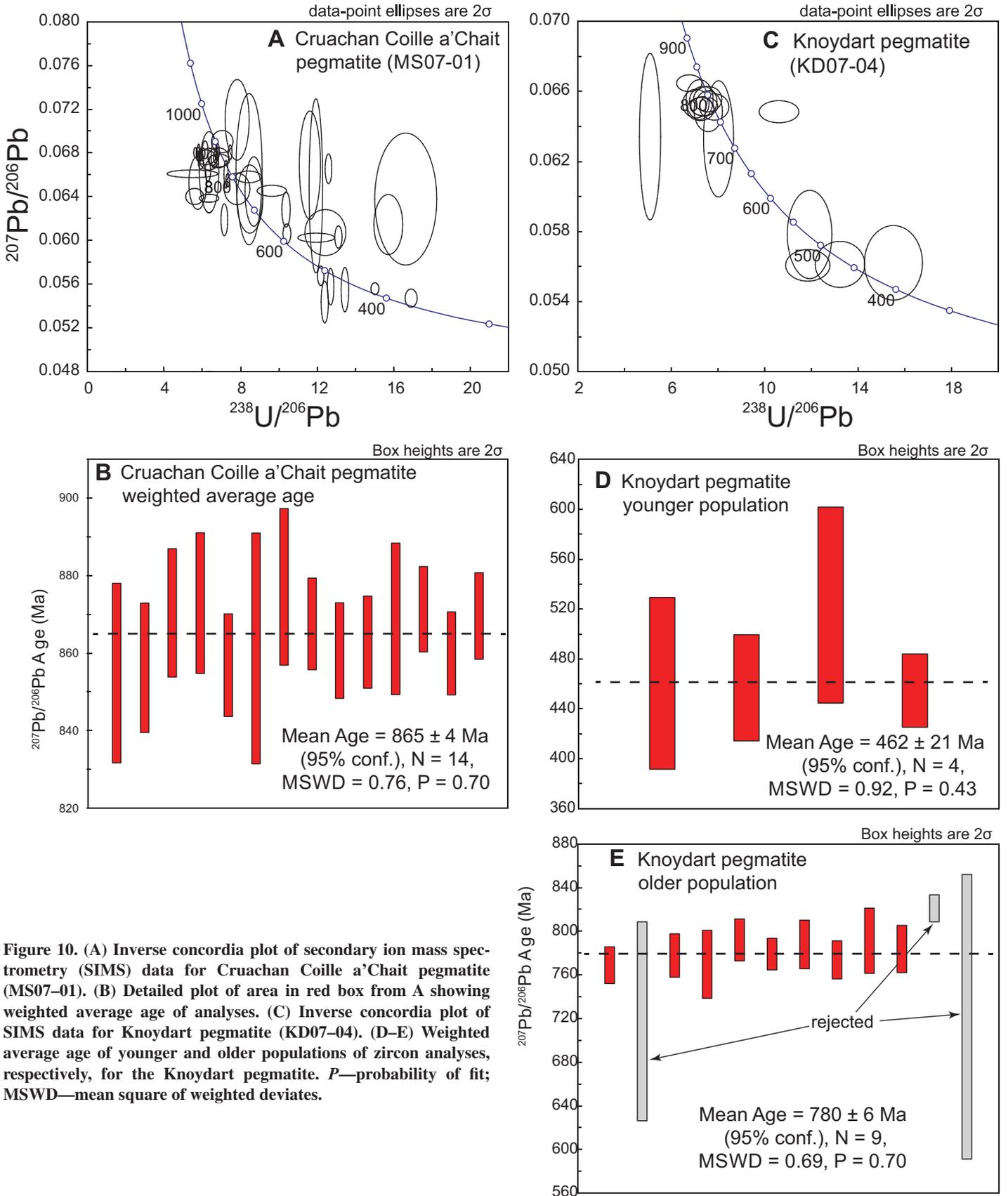
Figure 10. (A) Inverse concordia plot of secondary ion mass spectrometry (SIMS) data for Cruachan Coille a'Chait pegmatite (MS07–01). (B) Detailed plot of area in red box from A showing weighted average age of analyses. (C) Inverse concordia plot of SIMS data for Knoydart pegmatite (KD07–04). (D–E) Weighted average age of younger and older populations of zircon analyses, respectively, for the Knoydart pegmatite. *P*—probability of fit; MSWD—mean square of weighted deviates.

TABLE 2. ^{40}Ar/^{39}Ar RESULTS

	General characteristics			Plateau characteristics		
Sample no.	Mineral	Integrated age (Ma, ±2σ)	Plateau age (Ma, ± 2σ)	Total ^{39}Ar released (%)	MSWD	P
MS07-01	Muscovite	425.8 ± 2.0	**425.9 ± 1.9**	100.0	0.49	0.94
KD07-02	Muscovite	457.3 ± 4.1	**457.9 ± 3.6**	100.0	0.64	0.80

Note: Summary table indicating integrated, plateau/miniplateau and isochron ages for the Moines Formation samples. Mean square of weighted deviates (MSWD) for plateau, percentage of ^{39}Ar degassed used in the plateau calculation, number of analysis included in the isochron, and ^{40}Ar/^{36}Ar intercept are indicated. Plateau age calculated using trapped ^{40}Ar/^{36}Ar is indicated. Analytical uncertainties on the ages are quoted at 2 sigma (2σ) confidence levels and at 1σ for the ^{40}Ar/^{36}Ar intercept. Bold data indicate the accepted age for a given sample.

Source of 950–900 Ma Detritus

Although the source of the Moine sediment has generally been ascribed to the Grenville orogenic province (Cawood et al., 2004, 2007b; Kirkland et al., 2008), this cannot account for the youngest 950–900 Ma detritus. The termination of Grenville orogenic activity at ca. 1.0 Ga was followed by cooling and uplift, as recorded by ^{40}Ar/^{39}Ar mineral ages (Gower and Krogh, 2002; Rivers, 1997, 2012). In contrast, the Sveconorwegian province shows a similar overall age signature of late Mesoproterozoic tectonism and plutonism, but with orogenic activity extending as late as 920 Ma (Bingen et al., 2008, and references therein). Thus, the Sveconorwegian province provides a potential southern source for the 950–900 Ma detritus within the Moine succession.

A related issue concerns the tectonic setting of early Neoproterozoic deformation and metamorphism along the margin of eastern Laurentia as recorded in Svalbard, East Greenland, and the Shetland Islands. One viewpoint is that collision of Laurentia and Baltica formed a putative third arm of the Grenville-Sveconorwegian belt (Lorenz et al., 2012; Park, 1992). Isotopic data from a shear-bounded basement inlier within the Northern Highlands terrane at Glenelg (Fig. 2) yield evidence for eclogite-facies metamorphism at around 1080 Ma on the basis of Sm-Nd mineral and whole-rock dating (Sanders et al., 1984), with retrograde zircon growth in the eclogites at 1010 ± 13 Ma (Brewer et al., 2003) and at least partial exhumation on the boundary shear zone at 669 ± 31 Ma on the basis of a titanite age (Storey et al., 2004). Mesoproterozoic metamorphism of the inlier predates deposition of Moine succession, based on constraints from the youngest detrital zircons. Given the major and minor faults (e.g., Fig. 1B) that separate the inlier from the autochthonous foreland, and significant strike-slip displacement of the Northern Highland terrane during Caledonian orogenesis (Dewey and Strachan, 2003), it is difficult to evaluate the original paleogeographic position of the inlier and the tectonic significance of this isolated Mesoproterozoic age.

The alternative viewpoint to an extension of the Grenville orogen northward through Scotland, Greenland, and Norway is that early Neoproterozoic (980–920 Ma) tectonothermal activity along the present-day eastern Laurentian margin was related to development of an external accretionary plate margin, the Valhalla orogen of Cawood et al. (2010; see also Gasser and Andresen, 2013; Kirkland et al., 2011). Cawood et al. (2010) argued on the basis of paleomagnetic data (see also Cawood and Pisarevsky, 2006; Elming et al., 2014) from end Mesoproterozoic to early Neoproterozoic units that Baltica lay south of East Greenland, and this conclusion, combined with lithotectonic analysis, implies that this North Atlantic region did not occupy an internal location between Laurentia and Baltica but rather developed on the margin of the supercontinent (Fig. 13). Furthermore, Cawood et al. (2010) considered the 980–920 Ma tectonothermal activity discrete both tectonically and spatially from the Grenville orogeny and referred to it as the Renlandian orogeny of the Valhalla orogen. Calc-alkaline igneous activity and high-grade metamorphism along this plate margin, such as that documented in Svalbard, East Greenland, Shetlands, and East Sutherland at 960–915 Ma (Cutts et al., 2009; Gasser and Andresen, 2013; Leslie and Nutman, 2003; Kinny and Strachan, 2010, personal commun.), would have provided further potential sources for the detritus of this age within the Moine.

Mafic and Felsic Magmatism at 870 Ma

New U-Pb zircon ages for the Glen Doe and Sgurr Dhomhnuill intrusions of the West Highland Granitic Gneiss suite, and near coeval metagabbro intrusions at Glen Doe (Figs. 6–9), confirm a widespread pulse of magmatism at around 870 Ma within the Sgurr Beag Nappe (see also Friend et al., 1997; Millar, 1999; Rogers et al., 2001). These meta-igneous bodies intrude the Glenfinnan and Loch Eil groups (Figs. 2 and 3) and provide a broad younger age limit on their accumulation. Furthermore, the intrusions display all structural events recorded

Figure 11. Argon data for Ladhar Bheinn Pelite (KD07–02) and Cruachan Coille a'Chait pegmatite (MS07–01).
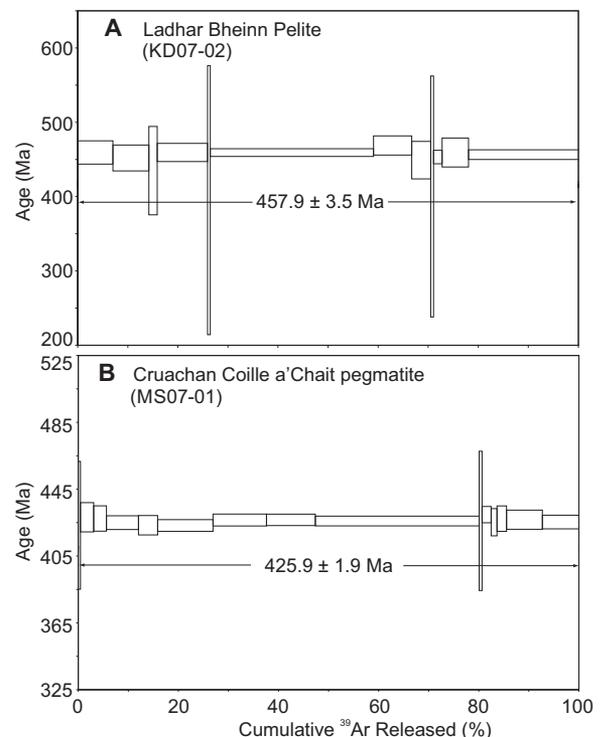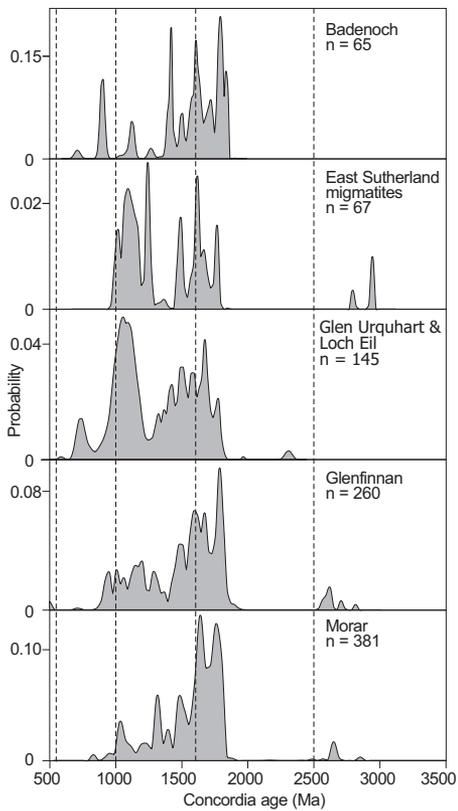
Figure 12. Probability plots comparing detrital zircon ages from lithostratigraphic units of the Moine Supergroup and correlatives within the sub-Grampian basement south of the Great Glen fault (Badenoch Group). Sources of data: Morar Group (this paper; Friend et al., 2003; Kirkland et al., 2008); Glenfinnan Group (this paper; Friend et al., 2003; Kirkland et al., 2008); Loch Eil Group and Glen Urquhart succession (Cawood et al., 2004); East Sutherland migmatites (Friend et al., 2003; as modified by Cawood et al., 2004); and Badenoch Group (Cawood et al., 2003).

in the Moine metasedimentary rocks, indicating that the timing of igneous activity provides an older age limit on Knoydartian deformation (Dalziel and Soper, 2001). The detrital age signature of the Glenfinnan and Loch Eil groups is similar to the inheritance pattern of zircons within the granitic gneiss bodies, indicating that the latter were indeed derived from melting of the former (Friend et al., 1997, 2003; Rogers et al., 2001) as suggested earlier on the basis of geologic and structural mapping and comparison of zircon morphologies (Dalziel, 1963, 1966). The bimodal nature of the magmatism at this time is consistent with mafic magmatic underplating of the crust, leading to widespread crustal melting (Fowler et al., 2013; Ryan and

Soper, 2001). The MORB geochemical affinities of the mafic phases associated with the West Highland Granitic Gneiss suggest crustal extension and thinning (Millar, 1999) and may indicate propagation of the inferred spreading center associated with the Asgard Sea (Cawood et al., 2010) into the Laurentian margin (Fig. 13).

Ages and Significance of Felsic Pegmatites—Knoydartian Orogeny

The U-Pb zircon age of 830 ± 3 Ma for the Loch Cluanie pegmatite falls within the older range of Knoydartian metamorphic and pegmatite ages (Fig. 14). Although the Cruachan Coille a'Chait pegmatite is not reliably dated, it is important because the pegmatite crosscuts a preexisting gneissic fabric (Fig. 4A), thus demonstrating a Neoproterozoic age for the earliest deformation and metamorphism of its host Moine rocks. A possible age of ca. 793 Ma falls within the early phase of the Knoydartian orogeny. The U-Pb zircon age for the Knoydart pegmatite (786 ± 4 Ma) similarly falls within the older range of Knoydartian metamorphic and pegmatite ages, its intrusion age comparing closely to that of the U-Pb monazite age of the Sgurr Breac pegmatite (784 ± 1 Ma; Rogers et al., 1998) at a similar structural level within the Morar Group ~7 km to the southeast (Fig. 2). Both pegmatites are concordant with gneissic fabrics in their host rocks and are inferred to have formed more or less in situ as a result of segregation during a high-grade metamorphic event (Hyslop, 2009b; Rogers et al., 1998). Independent evidence for high-grade metamorphism at ca. 820–790 Ma is provided by Sm-Nd garnet ages from Morar Group pelites (Vance et al., 1998) and U-Pb ages from monazite inclusions within zoned garnets in the Glenfinnan Group near Glen Urquhart (Cutts et al., 2010). In contrast, the U-Pb zircon age of 740 ± 8 Ma for the Carn Gorm pegmatite falls within the younger range of Knoydartian metamorphic ages (Fig. 14). Its field relations are comparable with the Knoydart and Sgurr Breac pegmatites, consistent with a further episode of high-grade metamorphism leading to in situ melting. The new age for the Carn Gorm pegmatite is within error of the U-Pb zircon age of 742 ± 19 Ma obtained from metamorphic rims within deformed Glen Doe metagabbro. Independent evidence for a younger high-grade metamorphic event at ca. 740 Ma is provided by titanite ages of 737 ± 5 Ma from the SW Morar Group (Tanner and Evans, 2003) and U-Pb ages from monazites present as inclusions within zoned garnets in the Glenfinnan Group near Glen Urquhart (Cutts et al., 2010).

Available Knoydartian age data span from 840 to 725 Ma (Table DR7 [see footnote 1]) and
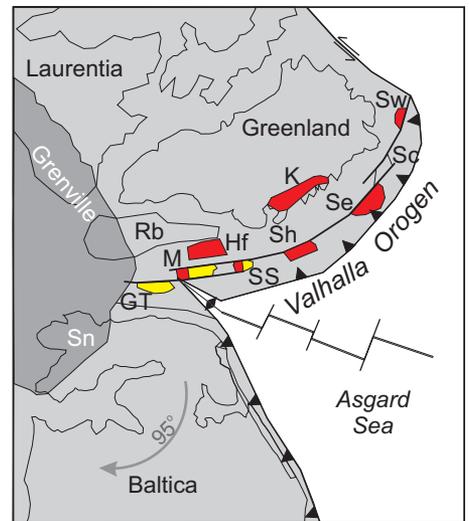
Figure 13. Neoproterozoic reconstruction of eastern Laurentia and Baltica prior to opening of Iapetus Ocean (adapted from Cawood et al., 2010). Baltica is shown in post–1000 Ma position with respect to Laurentia after it has undergone some 95° of clockwise rotation creating the Asgard Sea. Highlighted are the successions associated with the Valhalla orogen. Red blocks correspond to sedimentary and deformational sequences associated with Renlandian orogenic cycle of Valhalla orogen, and yellow blocks correspond to Knoydartian cycle. Abbreviations: GT—Grampian terrane (Badenoch Group); M—Moine succession; Hf—Hebridean foreland; K—Krummedal succession; Rb—Rockall Bank; SS—Svaerholt and Sørøy successions; Sh—Shetland Islands; Sn—Sveconorwegian orogen; Se—Eastern terrane of Svalbard; Sc—Central terrane of Svalbard; Sw—Western terrane of Svalbard.

fall into an older (840–780 Ma) and younger (740–725 Ma) set separated by a 40 m.y. gap, suggesting that the Knoydartian incorporates at least two distinct events. Only the Moine and Sgurr Beag thrust sheets show evidence for both older and younger phases of the Knoydartian event, whereas the Grampian terrane (Badenoch Group; Fig. 3) only shows evidence for the older event. This is consistent with age data from the Dalradian Supergroup, the lower parts of which are thought to have been deposited unconformably at ca. 750 Ma on the Badenoch Group (Robertson and Smith, 1999), overlapping with the youngest Knoydartian deformation and metamorphism in the Northern Highlands terrane (Fig. 14). Thus, the Badenoch Group was unlikely to have been in direct stratigraphic continuity with the main Moine succession, with

Figure 14. Time-space plot of age constraints on principal Neoproterozoic metasedimentary units and of tectonothermal events within the thrust sheets of the Moine Supergroup, Scotland. Numbers on data points refer to the following sources: 1—Friend et al. (2003), U–Pb zircon age of 1032 ± 12 Ma for youngest detrital grain in Morar Group, Moine Nappe; 2—Kirkland et al. (2008), U-Pb zircon age of 1022 ± 24 Ma for youngest detrital grain in Morar Group, Moine Nappe; 3—this paper, U-Pb age of 1011 ± 57 Ma for youngest detrital grain in lower psammite, Morar Group, Moine Nappe; 4—Peters (2001), U-Pb zircon age of 980 ± 4 Ma for youngest detrital grain in Morar Group, Moine Nappe; 5—this paper, U-Pb zircon age of 899 ± 51 Ma for youngest detrital grain in Glenfinnan Group, Moine Nappe; 6—Kirkland et al. (2008), U-Pb magmatic zircon overgrowth age of 842 ± 20 Ma on detrital Morar Group zircon, Moine Nappe; 7—Rogers et al. (1998), U-Pb monazite ages of 827 ± 2 Ma and 781 ± 1 Ma for the synmetamorphic Ardnish and Sgurr Breac pegmatites intrusive into Morar Group; 8—Vance et al. (1998), Sm-Nd garnet whole-rock ages of 823 ± 5 Ma and 788 ± 4 Ma for growth zones in garnet from the Morar Group; 9—this paper, U-Pb zircon age of 786 ± 7 Ma for Knoydart pegmatite, Moine Nappe; 10—Tanner and Evans (2003), U-Pb titanite age of 737 ± 5 Ma from calc-silicate pod in Morar Group, Moine Nappe; 11—Storey et al. (2004), U-Pb titanite age of 669 ± 31 Ma occurring within, and inferred to date, a shear zone between eastern and western units of the Glenelg-Attadale Inlier, which also contains Morar Group rocks; 12—Burns et al. (2004), Nd depleted mantle model age of around 1000 Ma for the Strathy Complex, possible basement assemblage to the East Sutherland Moine succession in the Naver Nappe; 13—Kinny and Strachan (2010, personal commun.), U-Pb zircon age of ca. 965 Ma for protolith to orthogneiss intrusive into Moine rocks of Naver Nappe; 14—Friend et al. (2003), U-Pb zircon age of 926 ± 68 Ma for youngest detrital grain from within the Kirtomy migmatites in East Sutherland Moine succession, Naver Nappe; 15—Kirkland et al. (2008), U-Pb zircon age of 1009 ± 22 Ma for youngest detrital grain in sample of Glenfinnan Group, Sgurr

Beag Nappe; 16—Cawood et al. (2004), U-Pb zircon age of 962 ± 32 Ma for the average of four analyses from youngest detrital grain in sample of Loch Eil Group, and U-Pb zircon age of 883 ± 35 Ma for the average of two analyses from youngest detrital grain in sample of Glen Urquhart psammite, Sgurr Beag Nappe; 17—Friend et al. (2003), U-Pb zircon age of 947 ± 59 Ma for youngest detrital grain in sample of granite gneiss incorporating metasediments of the Glenfinnan Group, Sgurr Beag Nappe; 18—Cutts et al. (2010), U-Pb zircon age of 917 ± 13 Ma for youngest detrital grain from sample of Glenfinnan Group, Sgurr Beag Nappe and U-Pb zircon age of 725 ± 4 Ma and inductively coupled plasma–mass spectrometry (ICP-MS) monazite ages of 825 ± 18 Ma, 782 ± 11 Ma, and 724 ± 6 Ma for monazite from migmatites within the Glenfinnan Group, Sgurr Beag Nappe; 19—this paper, U-Pb zircon ages of 869 ± 8 Ma and 860 ± 18 Ma for Glen Doe granitic gneiss, 880 ± 7 Ma for granitic xenolith in Glen Doe metagabbro, and 742 ± 19 for metamorphic rims on zircon grains within the metagabbro; 20—U-Pb zircon ages of felsic and mafic igneous bodies emplaced into the Glenfinnan and Loch Eil groups of the Sgurr Beag Nappe give ages of 873 ± 7 Ma for the Ardgour granite gneiss (Friend et al., 1997), 873 ± 6 Ma for mafic sheets (Hyslop, 2009a), and 870 ± 20 Ma for the Fort Augustus granite gneiss (Rogers et al., 2001); 21—this paper, thermal ionization mass spectrometry (TIMS) and secondary ion mass spectrometry (SIMS) U-Pb zircons ages for samples of Sgurr Dhomhnuill phase of Ardgour granite gneiss of 863 ± 4 Ma and 852 ± 10 Ma, respectively; 22—this paper, U-Pb zircon ages for Cruachan Coille a'Chait pegmatite, Loch Cluanie of ca. 793 ± 24 Ma and 830 ± 3 Ma for pegmatite on shore of Loch Cluanie; 23—this paper, U-Pb monazite age for Carn Gorm pegmatite of 756 ± 4 Ma; 24—van Breemen et al. (1974), Rb-Sr muscovite age for Carn Gorm pegmatite of 730 ± 20 Ma; 25—Highton et al. (1999), U-Pb zircon ages of 926 ± 6 Ma for youngest detrital grain and 840 ± 11 Ma for melt crystallization and new zircon growth in Badenoch Group, and the succession is part of the sub-Grampian basement to the Dalradian Supergroup and inferred to be an equivalent to the Moine succession (Piasecki, 1980); 26—Cawood et al. (2003), U-Pb zircon age of 900 ± 17 Ma for youngest detrital grain from sample of Badenoch Group; 27—Noble et al. (1996), U-Pb monazite ages of 806 ± 3 Ma, 808 +11/–9 Ma, and 804 +13/–12 Ma for pegmatites and mylonitic rocks from the Grampian shear zone separating sub-Grampian basement and Dalradian Supergroup; 28—approximate age of 750 Ma based on Rb/Sr ages of muscovites in pegmatites from Badenoch Group (Piasecki and van Breemen, 1983, and references therein); 29—Halliday et al. (1989) and Dempster et al. (2002) have determined U-Pb zircon ages for the Tayvallich Volcanics and inferred comagmatic intrusive rocks of around 600–595 Ma. Abbreviations: DS—Dalradian Supergroup; ES—East Sutherland Moine succession; BG—Badenoch Group, sub-Grampian basement; GL—Glenfinnan and Loch Eil groups, including Glen Urquhart psammite; GF—Glenfinnan Group; MG—Morar Group; SC—Strathy complex.
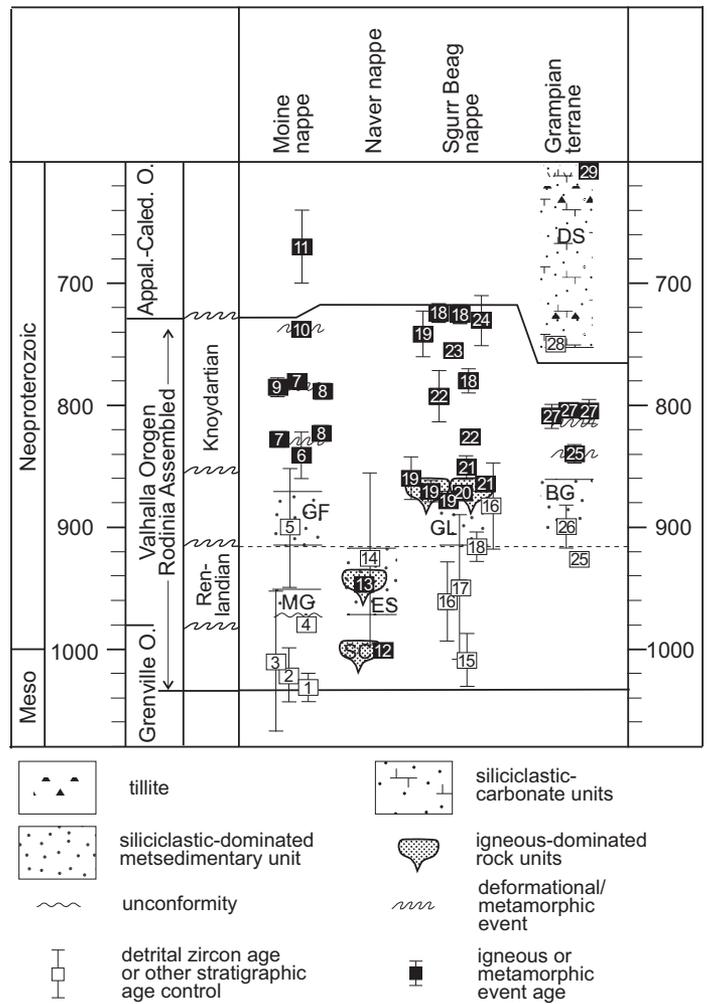
juxtaposition on the intervening Great Glen fault occurring during the Paleozoic Caledonian orogeny (cf. Dewey and Strachan, 2003).

Implications for Caledonian Tectonic Models

New U-Pb zircon and monazite data for the Carn Gorm pegmatite and its host gneisses indicate a high-grade metamorphic event at ca. 456 Ma. This is younger than published ages for the Ordovician Grampian orogenic event elsewhere in the eastern Moine Supergroup. These include a U-Pb TIMS titanite age of 470 ± 2 Ma from the Fort Augustus granitic gneiss (Rogers et al., 2001) and a U-Pb SIMS zircon age of 463 ± 4 Ma from a synkinematic pegmatite at Glen Urquhart (Cutts et al., 2010). The ca. 456 Ma event in the Carn Gorm pegmatite and host gneisses is perhaps more likely to be associated with the ca. 450 Ma tectonothermal event recently identified as having affected large tracts of the Morar Group (Bird et al., 2013). Of particular relevance to the present study is a Sm-Nd garnet age of 450 ± 2 Ma obtained from the Morar Group basal pelite at Glenelg (Bird et al., 2013). The U-Pb zircon age of 446 ± 2 Ma for the Glenelg pegmatite therefore confirms Late Ordovician high-grade metamorphism and pegmatite formation as well as pervasive "D2" deformation in this part of the Morar Group. The "D2" folds and associated mineral lineations in the Glenelg area had been correlated previously with similarly oriented "D2" structures within the Morar Group further north in Sutherland known to be of Silurian (Scandian) age (Kinny et al., 2003a). However, the new data presented here show that this is incorrect, emphasizing the difficulties in using fold style and fabric orientation as the basis for correlation, even over relatively small areas. As is likely to be the case for Knoydartian events, there are considerable variations in the relative intensities of the various Caledonian metamorphic events across the Moine Supergroup. This is reinforced by the $^{40}Ar/^{39}Ar$ data for the Ladhar Bheinn Pelite, which indicate no substantial reheating since ca. 458 Ma, whereas muscovites within the Cruachan Coille a'Chait pegmatite further east were presumably reset at ca. 426 Ma, during the Scandian orogenic event. The U-Pb zircon lower-intercept age of 433 ± 3 Ma obtained from the Ardgour granitic gneiss reflects substantial reheating within the central outcrop of the Moine succession during the Scandian orogenic event.

Overall, the geochronological data suggest that the Moine Supergroup was affected by significantly more orogenic events than would be indicated by D-numbers alone. This can be explained in various ways: The early isoclinal folds may themselves be polyphase (Strachan et al., 2010), deformation as well as metamorphism varied spatially in intensity (see earlier herein), and it is also probable that some major folds and ductile thrusts have a composite history (Bird et al., 2013).

Along-Strike Laurentian Margin Comparisons

Neoproterozoic siliciclastic-dominated sequences are widespread along the eastern continental margin of Laurentia, stretching from the southern Appalachians to northern Greenland (Rankin et al., 1993; Strachan et al., 2012; Watt and Thrane, 2001; Williams et al., 1995). These rocks are generally related to rifting associated with the initiation of the breakout of Laurentia from the supercontinent Rodinia (Bond et al., 1984). The Moine Supergroup and correlative successions in Shetlands, East Greenland, Svalbard, and Norway are amongst the earliest record of Neoproterozoic lithospheric extension and subsidence. These successions differ from those in eastern North America by their older age of sediment accumulation, between ca. 1000 and 870 Ma, and evidence for at least two pulses of Neoproterozoic deformation and metamorphism, the Renlandian and Knoydartian orogenies at 980–920 Ma and 845–725 Ma,

respectively (Figs. 14 and 15; Cawood et al., 2010). In eastern North America, the oldest successions related to Rodinia breakup occur in the Blue Ridge and are dated as mid-Neoproterozoic, accumulating between ca. 760 and 700 Ma (Aleinikoff et al., 1995; Evans, 2000; Tollo et al., 2004). Correlatives of the Blue Ridge succession include the Dalradian Supergroup in the Grampian Highlands terrane in Scotland, the Eleonore Bay Supergroup in East Greenland, and the Murchinsonfjorden and Sofiebogen successions in Svalbard (Cawood et al., 2007b, 2010). These North Atlantic successions are inferred to have been deposited unconformably on the predeformed correlatives of the Moine Supergroup (Robertson and Smith, 1999; Sonderholm et al., 2008). Understanding the origin of these early Neoproterozoic successions around the North Atlantic, and their absence from eastern North America, is important in constraining Laurentian paleogeography during the breakup of Rodinia. Cawood et al. (2010) have argued that regions incorporated into the Appalachian orogen in eastern North America occupied an internal location within Rodinia during the early Neoproterozoic, whereas rocks dispersed around the North Atlantic realm and subsequently constituting part of the Caledonian orogen occupied an external location (Figs. 13 and 15). This external location was brought
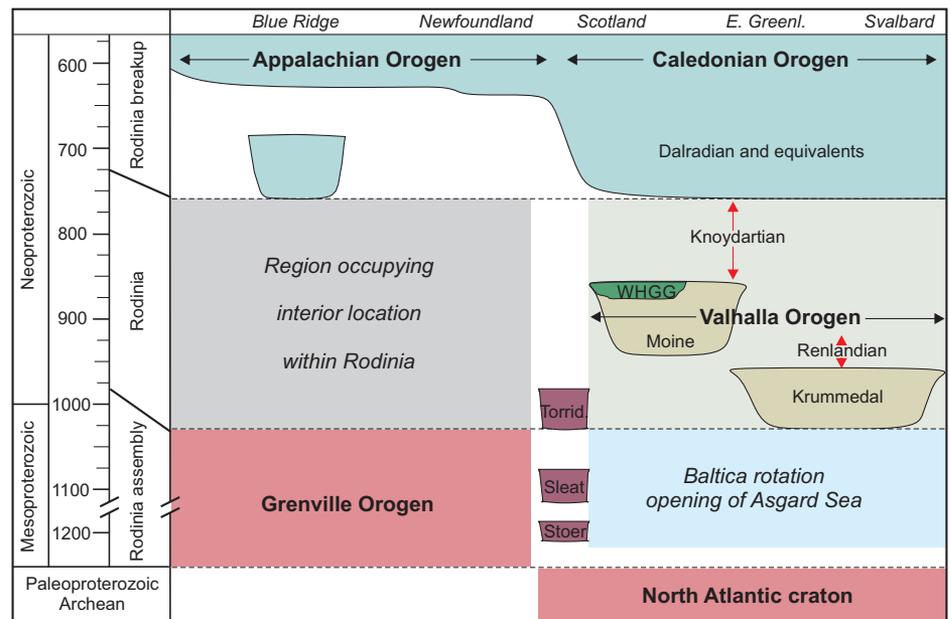
Figure 15. Schematic time-space plot of orogenic successions and events extending from the southeast United States, through the Atlantic provinces of Canada, across Ireland and the UK to East Greenland and Svalbard. The Grenville orogen formed during Rodinia assembly, the Valhalla orogen developed along the margin of an assembled Rodinia, and the Appalachian-Caledonian orogen was initiated during Rodinia breakup. WHGG—West Highland granite gneiss; Torrid—Torridonian.

about by the end-Mesoproterozoic ~95° clockwise rotation of Baltica with respect to Laurentia. Throughout much of the Mesoproterozoic, the present-day northern margin of Baltica was adjacent to East Greenland, resulting in a linear Grenville-Sveconorwegian orogen (e.g., Gower et al., 1990; Karlstrom et al., 2001). End-Mesoproterozoic rotation of Baltica resulted in its Scandinavian margin facing Scotland, the Rockall Bank, and southeast Greenland, a position it maintained until the opening of the Iapetus Ocean at the end of the Neoproterozoic (Cawood et al., 2001; Cawood and Pisarevsky, 2006). Rotation resulted in oroclinal bending of the Grenville-Sveconorwegian orogen in the south and opening of the Asgard Sea in the north (Figs. 13 and 15) such that the early Neoproterozoic successions occupied a peripheral location (in sense of Murphy and Nance, 1991) with respect to Rodinia (see also Gasser and Andresen, 2013; alternative interpretation in Lorenz et al., 2012). This peripheral location on the margin of Laurentia provided both a site for extension and sediment accumulation, represented by the Moine Supergroup, and for deformation, metamorphism, and magmatism associated with crustal thickening during Renlandian and Knoydartian orogenesis, which were in part driven by subduction following closure of the interior ocean, all as part of the Valhalla orogenic cycle (Fig. 14). Cutts et al. (2010, and references therein) determined maximum pressure and temperature conditions for Moine rocks during Knoydartian orogenesis of 700 °C and 0.9 GPa. In contrast, eastern North America at this time (1000–760 Ma) occupied an intracratonic position within Rodinia (Fig. 15), and only toward the end of this period did it start to undergo lithospheric extension and sedimentation, along with associated igneous activity as preserved in the Appalachian Blue Ridge (e.g., Tollo et al., 2004). In Scotland, this second phase of extension, represented by the Dalradian succession, marks the opening of the Iapetus Ocean and initiation of the Caledonian orogen and is inferred to have commenced at a similar time (ca. 760 Ma; Prave et al., 2009) to that in the Appalachians (Figs. 14 and 15).

During the early Paleozoic, the eastern Laurentian Neoproterozoic successions were affected by a series of approximately coeval orogenic events associated with closure of the Iapetus Ocean, although the tectonic drivers for, and hence the intensities of, these events vary along strike. Late Cambrian to Late Ordovician orogenic events were generally short-lived (Dewey, 2005) and resulted from accretion of crustal ribbons to the Laurentian margin. In the Canadian Appalachians, these are referred to collectively as "Taconic" (Rogers, 1970; Williams, 1995),

comprising three geodynamically distinct orogenic events (van Staal and Barr, 2012, and references therein; van Staal et al., 2007, 2009). The Late Cambrian (ca. 495 Ma) "Taconic 1" event represents the west-directed obduction of an oceanic arc onto peri-Laurentian crust in Newfoundland (van Staal et al., 2007; Waldron and van Staal, 2001) but has no counterpart in the Irish–Scottish–East Greenland Caledonides. In contrast, the Early Ordovician (ca. 480–470 Ma) "Taconic 2" event, associated with ophiolite obduction onto the Laurentian margin and arc-continent collision (Cawood and Suhr, 1992; Cawood et al., 1995; Chew et al., 2010; Roberts, 2003; van Staal et al., 1998), is the main Ordovician orogenic phase in the Appalachians and Caledonides. It correlates with the Grampian orogenic event that regionally deformed and metamorphosed the Laurentian Neoproterozoic successions of Scotland and Ireland (Dewey and Ryan, 1990; Oliver et al., 2000; Soper et al., 1999). Moine rocks in East Sutherland were migmatized (Kinny et al., 1999) during metamorphism up to 650–700 °C and 11–12 kbar (Friend et al., 2000), and elsewhere in the eastern Moine rocks, there was growth of new garnet (Bird et al., 2013), monazite (Cutts et al., 2010; this study), and titanite (Rogers et al., 2001). In Newfoundland, the Late Ordovician (ca. 460–450 Ma) "Taconic 3" orogenic event resulted from the accretion to the Laurentian margin of the Popelogan-Victoria arc (van Staal et al., 2009, and references therein). This occurred broadly coeval with the ca. 450 Ma metamorphic event identified in the western Moine rocks. In Newfoundland, this was associated with garnet-grade metamorphism (Vance and O'Nions, 1990), and in the western Moine rocks, it was associated with synkinematic growth of garnet (Bird et al., 2013), ductile deformation, and segregation of granitic pegmatites and veins (this study). In Scotland, this orogenic event has similarly been attributed to the accretion to the Laurentian margin of an arc or microcontinental fragment (Bird et al., 2013). In the Scandinavian Caledonides, the Laurentian-derived Uppermost Allochthon in Scandinavia (Baltica) contains evidence for arc-microcontinent accretion events that are approximately coeval with Taconic 2 and 3 and correlative events in Ireland-Scotland (Corfu et al., 2003; Roberts, 2003; Roberts et al., 2007).

Silurian orogenic events and final closure of the Iapetus Ocean resulted from the collision of an amalgamated Gander-Avalonia-Baltica crustal block with eastern Laurentia. The mid-Silurian (430–422 Ma) Salinic event in Newfoundland occurred approximately coeval with the Erian event in western Ireland (Dewey et al., 1997), although the two vary in their intensity. Whereas in Newfoundland this collision

resulted in strong deformation and metamorphism up to migmatite grade (Cawood et al., 1994; D'Lemos et al., 1997; Dunning et al., 1990; van Staal et al., 1994), in Ireland-Scotland, it was relatively "soft," with minor deformation and low-grade metamorphism associated with sinistrally oblique docking of Gander/Avalonia with Laurentian terranes along the Iapetus suture (Soper and Woodcock, ~500–700 km further north of their present position relative to the Grampian terrane, and they record the effects of the Silurian (435–425 Ma) Scandian orogenic event that resulted from the collision of Baltica with this segment of the Laurentian margin (Dewey and Strachan, 2003). The Moine rocks were affected by regional-scale ductile deformation and metamorphism up to 650 °C and 5–6 kbar (Friend et al., 2000). Late Silurian to Early Devonian sinistral displacement along the Great Glen fault juxtaposed the Northern Highland terrane against the Grampian terrane, which was largely unaffected by any Silurian deformation or metamorphism (Dewey and Strachan, 2003).

CONCLUSIONS

Principal conclusions of our revised geochronology of the Moine Supergroup are:

(1) Detrital zircons from the Moine successions on Mull and at Knoydart in NW Scotland range in age from Archean to early Neoproterozoic. The data reported here, in combination with regional paleoflow data, add weight to the general consensus that the Moine sediments were derived post–1000 Ma from the erosion of the Grenville-Sveconorwegian orogen and environs.

(2) New U-Pb zircon ages for the Glen Doe and Sgurr Dhomhnuill intrusions of the West Highland Granitic Gneiss suite and coeval metagabbro intrusions confirm a widespread pulse of extension-related bimodal magmatism at ca. 870 Ma within the Sgurr Beag Nappe. These bodies provide an older age limit on Neoproterozoic "Knoydartian" orogenic activity, as they record all structural events preserved in their host rocks.

(3) Pegmatites yielding U-Pb zircon ages of 830 ± 3 Ma, ca. 793 Ma, 786 ± 7 Ma, and 740 ± 8 Ma constrain a series of deformation and metamorphic pulses related to Knoydartian orogenesis of the host Moine rocks.

(4) U-Pb zircon, monazite, and ^{40}Ar/^{39}Ar ages for pegmatites and host gneisses record reworking at ca. 458–446 Ma and ca. 426 Ma during the Caledonian orogenic cycle.

(5) The geochronological data demonstrate that some previously published structural correlations are incorrect, emphasizing the difficulties in basing these on fold style and fabric orientation, even over relatively small areas. The

Moine Supergroup was affected by significantly more orogenic events than might be suspected by the analysis of tectonic structures in any single area. Deformation and metamorphism varied spatially in intensity, and it is probable that some major folds and ductile thrusts have a composite history.

(6) The presence of early Neoproterozoic siliciclastic sedimentation and deformation in the Moine and equivalent successions around the North Atlantic and their absence along strike in eastern North America reflect contrasting Laurentian paleogeography during the breakup of Rodinia. The North Atlantic realm occupied an external location on the margin of Laurentia and accumulated detritus (Moine Supergroup and equivalents) derived from the Grenville-Sveconorwegian orogen. Neoproterozoic orogenic activity corresponds with the inferred development of convergent plate-margin activity along the periphery of the supercontinent. In contrast, in eastern North America, which lay within the internal parts of Rodinia, sedimentation did not commence until the mid-Neoproterozoic (ca. 760 Ma) during initial stages of supercontinent fragmentation, and there is no evidence of orogenic activity of this age.

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