Geological significance of $^{40}\text{Ar}/^{39}\text{Ar}$ mica dates across a mid-crustal continental plate margin, Connemara (Grampian orogeny, Irish Caledonides), and implications for the evolution of lithospheric collisions

Anke M. Friedrich and Kip V. Hodges

Abstract: The Connemara region is a world-class example of a regional-scale, high-temperature metamorphic terrain. Its rock record documents formation of a bi-vergent orogenic wedge and associated calcalkaline magmatism in a arc-continent collisional setting (Grampian orogeny), for which a protracted evolution was inferred based on a >75 Ma spread in U–Pb, Rb–Sr, and K–Ar mineral ages. In contrast, geological field observations imply a simple relationship between syntectonic magmatism, bi-vergent deformation, and Barrovian-type metamorphism. We explore the significance of the spread in apparent cooling ages using $^{40}\text{Ar}/^{39}\text{Ar}$ mica thermochronometers of varying grain sizes and composition, collected across metamorphic grades ranging from staurolite to upper sillimanite. We integrated geological and previously published geochronological evidence to identify a 32 Ma range (ca. 475–443 Ma) of permissible cooling ages and distinguished them from those dates not related to cooling after high-temperature metamorphism. Variations in $^{40}\text{Ar}/^{39}\text{Ar}$ dates at a single locality are ≤10 Ma, implying rapid cooling (≥6–26 °C/ Ma) following metamorphism and deformation. A distinct cooling age variation (≥15 Ma) occurs on the regional scale, consistent with spatial differences in the metamorphic, magmatic, and deformational evolution across Connemara. This cooling record relates to a lateral thermal gradient (30 °C/km) in an evolving arc–continent collision, rather than to differential unroofing of the orogen. Our results imply that the large (≥250 Ma) spread in thermochronometers commonly observed in orogens does not automatically translate into a protracted cooling history, but that only a small number of thermochronometers supply permissible cooling ages.

Résumé : La région de Connemara des Calédonides irlandaises est un exemple de classe mondiale de terrain métamorphique de haute température d’envergure régionale. Sa formation est reliée au magmatisme calcoalcalin dans le contexte d’une collision arc-continent, pour laquelle une évolution prolongée a été inférée à la lumière de la distribution sur plus de 75 Ma d’âges U–Pb, Rb–Sr, et K–Ar sur différents minéraux. Une telle évolution ne correspond pas aux observations géologiques sur le terrain, qui indiqueraient plutôt une relation simple entre un magmatisme, une déformation et un métamorphisme de type barroviens syntectoniques. Nous avons examiné la signification de la distribution des âges apparents de refroidissement à l’aide de thermochronomètres $^{40}\text{Ar}/^{39}\text{Ar}$ sur micas de granulométries et compositions variées, prélevés à différents degrés de métamorphisme allant de la zone à staurolite à la zone à sillimanite supérieure. Nous avons intégré les données géologiques et de l’information géochronologique déjà publiée pour délimiter une fenêtre de 32 Ma (de 475 à 443 Ma environ) pour les âges de refroidissement possibles et avons distingué ces derniers des âges non reliés au refroidissement ayant suivi le métamorphisme de haute température. La variation des âges $^{40}\text{Ar}/^{39}\text{Ar}$ en un même endroit est ≤10 Ma, ce qui indiquait un refroidissement rapide (≥6 à 26 °C/ Ma) suivant le métamorphisme et la déformation. Des variations claires des âges de refroidissement (≥15 Ma) existent à l’échelle régionale, correspondant aux variations spatiales de l’évolution métamorphique, magmatique et de la déformation à l’échelle de la région de Connemara. Cet historique de refroidissement est relié à un gradient latéral de température et de vitesse de déformation dans une collision arc-continent dynamique, plutôt qu’à la dénudation tectonique différentielle de l’orogène. Nos résultats indiquent que la large distribution (≥50 Ma) des thermochronomètres couramment observée dans les ceintures orogéniques ne reflète pas nécessairement une histoire de refroidissement prolongé, mais que seul un petit nombre de thermochronomètres produisent des âges de refroidissement admissibles pour un contexte donné. [Traduit par la Rédaction]

Introduction

Studies of thermal histories of ancient metamorphic terrains using thermochronologic techniques frequently yield large ranges of apparent ages, sometimes spanning hundreds of millions of years (Onstott and Peacock 1987; Mezger et al. 1991; Hodges and Bowring 1995). Such ranges are often interpreted in terms of a protracted cooling history, assuming reasonable closure temperatures for mineral-isotopic systems (e.g., Cliff 1985; Heaman and Parrish 1991; Hodges 1991, 2014; Baxter 2010). However, studies of the mechanisms by which radiogenic isotopes are lost from mineral systems (e.g., Lee 1995) suggest that assigning an exact closure...
temperature to any particular mineral sample is not entirely straightforward (Watson and Baxter 2007). For example, ranges of dates provided by a single mineral-isotopic system may reflect local differences in deformational history or mineral chemistry rather than protracted cooling (e.g., Mulch et al. 2005; Bröcker et al. 2013). One approach to assessing the importance of this problem is to compare thermochronologic data for minerals of known composition with independent constraints on the thermal and deformational history of the metamorphic terrain from which they were collected.

In this paper, we present thermochronologic data for the Neoproterozoic metamorphic complex of Connemara, one of the world’s best exposed sections of middle crust—a type-example of an island-arc continent collision (Grampian orogeny, e.g., Dewey 2005; Dewey and Ryan 2016). At Connemara, a protracted cooling history spanning at least 75 million years has been inferred from earlier Rb–Sr and K–Ar results (e.g., Elias et al. 1988, Cliff et al. 1996). The significance of this “Grampian tail” (Dewey and Manga 1999) continues to play an important role for tectonic models of island arc–continent collisions, involving subduction polarity-reversal and establishment of Andean-type margins, such as the one documented in western Ireland (e.g., Dewey 2005). Connemara provides a special opportunity to compare the thermochronologic record as documented in metamorphic rocks (e.g., Yardley et al. 1987) with very tight constraints on the thermal and deformational history of the area provided by the U–Pb geochronology of synmetamorphic intrusive igneous rocks of the Connemara complex (Friedrich et al. 1999a, 1999b). We employed the 40Ar/39Ar method on muscovite, biotite, and phlogopite from metapelitic and calc-silicate rocks sampled at different metamorphic grades, ranging from staurolite to upper sillimanite. The results were then evaluated against independent constraints. We found a substantial spread in 40Ar/39Ar dates, largely consistent with the results from previous studies, but the distribution of dates seems less a simple straightforward (Watson and Baxter 2007). For example, ranges of dates provided by a single mineral-isotopic system may reflect local differences in deformational history or mineral chemistry rather than protracted cooling (e.g., Mulch et al. 2005; Bröcker et al. 2013). One approach to assessing the importance of this problem is to compare thermochronologic data for minerals of known composition with independent constraints on the thermal and deformational history of the metamorphic terrain from which they were collected.

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Intrusive history

Intrusions of the Connemara complex were either broadly synchronous with or postdate crustal deformation. The oldest plutons, consisting of ultramafic and gabbroic rocks, occur in both northern (Dawros–Currywongaun–Doughruagh complex, Fig. 1; Kanaris-Sotiriou and Angus 1976) and southern Connemara (e.g., Cashel–Lough Wheelaun intrusion, Leake 1989). A younger intrusive suite, which consists mainly of quartz diorite and granite, intruded older mafic plutons only in southern Connemara. The gabbros of northern Connemara were emplaced syntectonically into the same tectonostratigraphic level as the southern Connemara gabbros (Leake and Tanner 1994). Wellings (1998) suggested that the Dawros–Currywongaun–Doughruagh complex intruded during D2 deformation. If he is correct, a recent U–Pb zircon age for a gabbro from the complex constrains the age of D2 in northern Connemara as 474 Ma (Friedrich et al. 1999a). The Cashel–Lough Wheelaun intrusion of southern Connemara intruded just prior to or early during D3 deformation (Tanner 1990) at 470 Ma (U–Pb zircon age, Friedrich et al. 1999a). A minimum age for both D3 and D4 deformation is provided by the undeformed Oughterrard granite (e.g., Tanner et al. 1997), which has a U–Pb xenotime crystallization age of 463 Ma (Friedrich et al. 1999a).

Metamorphic history

From north to south across Connemara, Grampian metamorphic conditions increase progressively from upper greenschist to upper amphibolite facies (Fig. 1). Four major zones, trending east–west, have been identified based on the distribution of prograde pelitic mineral assemblages (Barber and Yardley 1985; Yardley et al. 1987; Boyle and Dawes 1991): (1) the garnet–staurolite zone, (2) the staurolite–sillimanite transition zone, (3) the sillimanite–muscovite zone, and (4) the sillimanite–K-feldspar zone. In southern Connemara, near the principal outcrop region of quartz-diorites of the Connemara igneous complex, muscovite is no longer stable and the diagnostic metapelitic assemblage becomes sillimanite + biotite + cordierite + K-feldspar + quartz + plagioclase. Many metasedimentary outcrops are migmatitic in the southernmost parts of this zone, suggesting that temperatures were high enough to promote widespread anatexis.
Despite the simplicity of the metamorphic pattern at Connemara, textural relationships suggest that the observed assemblages grew during two prograde events. Minerals diagnostic of the garnet-staurolite assemblage help define the S2 schistosity in northern Connemara, whereas sillimanite-muscovite and sillimanite-K-feldspar assemblages help define the S3 schistosity in central and southern Connemara (Barber and Yardley 1985; Boyle and Dawes 1991; Yardley et al. 1987, and references therein). This has led previous workers to define an M2 event on the basis of the northern Connemara assemblages and an M3 event on the basis of central and southern Connemara assemblages. Estimated peak M2 conditions in northern Connemara were 550 ± 50 °C and 6–8 kb (Boyle and Dawes 1991). Peak M3 temperatures were at least 650 °C in the middle of the sillimanite-K-feldspar zone, increasing to ~750 °C in southernmost Dalradian outcrops (Barber and Yardley 1985). Near the main outcrop area of the quartz diorites, M3 pressures during anatexis reached ~5.5 kb, but may have declined to <3 kb during crystallization of the leucosome.

Indirect constraints on the age of M2 and M3 are provided by the U-Pb ages of synmetamorphic intrusive rocks. Based on the work of Wellings (1998), S2 and the M2 assemblages that define it are at least partly coeval with intrusion of the Currywongaun gabbro at 474 Ma (Friedrich et al. 1999a). The M3 event overlapped in time with emplacement of the bulk of the Connemara complex (Yardley et al. 1987), which occurred during the 470–463 Ma interval (Friedrich et al. 1999a, 1999b). While textural evidence suggests that amphibolite-facies metamorphism should be attributed to two separate events at Connemara, the two were separated in time by no more than a few million years and the prograde M2–M3 transition is probably best regarded simply as a temperature increase during progressive orogeny.

**Post-Grampian extensional faulting, erosion, sedimentation, igneous activity, and late-stage fluid flow**

The high-grade Dalradian metasedimentary rocks and the Connemara intrusions are overlain unconformably by unmetamorphosed shallow marine sedimentary rocks of lower Silurian age.
Fig. 3. Representative field photographs and photo micrographs of the deformation sequence in Connemara. (A) S1, schistosity is only preserved as inclusion trails within garnet. (B) S2, schistosity is the dominating penetrative fabrics at the low-metamorphic zones of northern Connemara, but a weak F3 overprint is usually observed. Exposure from the shore at Tully Mountain. (C) Sample AF17, a muscovite-garnet metapelite in which muscovite + chlorite define a retrograde mineral assemblage, probably equivalent to the M3 assemblages farther south. (D) Impure marble of the Lakes Marble Formation. This picture is from Cur Hill, at the staurolite-sillimanite transition zone, and shows the typical F3 folds that occur at a variety of scales throughout Connemara. F3 folds refold minor F2 fold axes. (E) The main S3 foliation plane in a staurolite-sillimanite–garnet schist from the staurolite-sillimanite transition zone at Cur Hill (sample AF79). (F) Outcrop “migmatitic” metasedimentary rocks, ca. 50 m northeast of the parking lot at the Alcock and Brown landing site memorial. The field relationships show the S2 schistosity folded around a F3 fold-axis in a metasedimentary rock. This block has been disrupted subsequent to F3 folding, as indicated by its random orientation relative to the strong anatectic foliation defined within the anatectic metapelite rock. The foliation (S3) is defined by alignment of paleosome and leucosome. (G) Outcrop F2–F4 fold generations. F2 and F3 fold axial traces are subparallel, whereas F4 folds are almost orthogonal to the older folds. F4 folds are accompanied by granitic pegmatite. Outcrop at the head of Errislanaan peninsula, ca. 150 m southwest of the lighthouse.

(Fig. 1; e.g., Leake and Tanner 1994; Ryan and Dewey 2011; Clift et al. 2004). This unconformity marks the end of the Grampian orogeny in the geologic record at Connemara. The unconformity is estimated to have formed at ca. 443 Ma, but no later than 435 Ma (Clift et al. 2004; Tucker and McKerrow 1995), such that the associated hiatus may be up to 29 Ma long. The real hiatus duration is shorter, because Dalradian rocks are overlain locally by a Late Ordovician alluvial cobble-conglomerate (Lough Mask Formation; Derryveeney Formation; Graham et al. 1989; Clift et al. 2004; Ryan and Dewey 2011). Direct evidence for erosional unroofing of the Dalradian sequence is documented in northeastern Connemara and South Mayo by the first arrival of metamorphic clasts in the sedimentary record of the Rosroe and Mweelrea Formations between ca. 468 Ma and at least 462 Ma (cf. fig. 13.5 in Ryan and...
Dewey 2011). By Upper Silurian to Devonian time, the Connemara metamorphic complex was buried again under ~3-4 km of marine sedimentary rocks (Graham et al. 1989). At this time, a major post-Grampian granitic intrusion—the Galway batholith—intruded older plutons of the Connemara igneous complex (e.g., Burke 1957; Leake and Tanner 1994). The thermal pulse associated with batholith intrusion led to the development of a regionally extensive hydrothermal system (Jenkins et al. 1992), which may be responsible for local retrogression of M2–M3 metamorphic assemblages throughout Connemara.

Previous thermochronology

K–Ar and Rb–Sr mineral dates ranging from 490 to 415 Ma were interpreted as representing a prolonged period of unroofing (e.g., Giletti et al. 1961, Dewey et al. 1970; Elias et al. 1988; Dewey 2005) or local resetting related to hydrothermal alteration (e.g., Leggo et al. 1966; Miller et al. 1991). A discordant U–Pb zircon date of ca. 490 Ma for the Cashel–Lough Wheelaun gabbro (Jagger et al. 1988), intruded just before or during the early stages of D3, has been used to suggest that M3 metamorphism began at about that time (Cliff et al. 1996). This interpretation gained support from slightly younger K–Ar hornblende dates from some outcrops (Elias et al. 1988). Mica K–Ar and Rb–Sr cooling dates clustered between 460 and 450 Ma in southern Connemara, leading previous workers to suggest two important episodes of cooling in the region—one between 490 and 480 Ma, and the other between 460 and 450 Ma, separated by an interval of relatively slow cooling (e.g., Elias et al. 1988). However, not all early thermochronologic data fit into this simple model. One K–Ar hornblende date of 420 Ma was reported from northern Connemara, and several southern Connemara samples yielded mica dates in the 420–410 Ma age range (e.g., Elias et al. 1988). Such anomalies led to the development of complex models for regional cooling involving differential uplift along high-angle faults and localized reheating events related to hydrothermal activity (Elias et al. 1988). In particular, Elias et al. (1988) found a general southward increase of hornblende K–Ar dates (of up to 480 ± 10 Ma) with increasing metamorphic grade and concluded that they cooled before the northern lower-grade rocks.

An alternative model was proposed by Miller et al. (1991). These authors determined K–Ar dates for 21 hornblende and several micas from across Connemara, obtaining a large range of dates similar to that reported by Elias et al. (1988). In one 30 m wide quarry in southern Connemara, hornblende dates for amphibolites containing abundant quartz + epidote veins ranged over 70 million years. Miller et al. (1991) found a weak positive correlation between K–Ar hornblende dates and 87Sr/86Sr values for the same amphibolites, suggesting that the range of hornblende dates represented variable amounts of Ar loss related to hydrothermal activity. In this scenario, K–Ar hornblende dates provide little constraint on the thermal evolution of the Connemara region.

U–Pb geochronologic analyses of abraded single-zircon crystals from intrusive rocks across Connemara (Friedrich et al. 1999a, 1999b) helped to place constraints on the range of viable interpretations of K–Ar and Rb–Sr thermochronologic data. For example, a more precise redetermination of the age of the Cashel–Lough Wheelaun gabbro and a new determination of the age of the Currywogaun gabbro of southern Connemara demonstrate that peak M2–M3 metamorphism is no older than 475–470 Ma (Friedrich et al. 1999a), such that older published K–Ar hornblende dates must represent excess 40Ar contamination. Moreover, given the very brief duration of Grampian orogenesis at Connemara, we would expect that all amphibole and mica cooling dates would cluster within a few million years, between perhaps 460 and 450 Ma (Friedrich et al. 1999a). The Ordovician–Silurian boundary provides a lower bound on the post-Grampian cooling history. In this paper, we report new 40Ar/39Ar geochronologic data obtained in an attempt to better understand the extent and cause of the large range of K–Ar and Rb–Sr ages obtained by previous workers.

Methods

We collected muscovite, biotite, and phlogopite from metapelitic and calcilutite rocks from small subareas within each metamorphic zone (Fig. 4). We assumed that the effective closure of a mineral-isotopic system is a volume diffusion process, governed by effective diffusion dimension and chemical composition (McDougall and Harrison 1988; Watson and Baxter 2007; Baxter 2010; Hodges 2014). Although studies suggest that fast diffusion pathways can complicate the geometry and rate of diffusion in minerals (Lee 1995), the effective diffusion dimension of mica crystals empirically appears to be similar to the physical grain size (Hames and Bowring 1994). In the hope of recovering a larger portion of the cooling history at each locality, we concentrated on analyzing micas with a range of composition and grain size. We separated muscovite, biotite, and phlogopite from crushed or uncrushed rock and purified following standard procedures (Appendix A; Hodges et al. 1994).

We performed both high-precision incremental heating analyses of large samples (several milligrams) with a resistance furnace, and microanalysis of between 1 and 10 individual crystals with an Ar-ion laser, which yields most of its energy in the blue-green visible spectrum (Appendix A; supplementary data, Table ST14; Fig. SF2f). To control the effects of sample composition and grain size, we determined the compositions of between three and five representative crystals of each analyzed sample (Tables ST1, ST2, and ST3), and measured the grain size of each crystal analyzed (Tables 1–3). The grain sizes of biotite, muscovite, and phlogopite from northern Connemara are relatively uniform and small (typically 300–500 μm in diameter). More significant grain size variations are found in southern Connemara, with the largest grains (>6000 μm diameter) occurring in some muscovite-bearing pegmatites.

All age calculations were based on assuming an initial 40Ar/36Ar ratio of modern atmosphere (295.5). Although previous work suggests that excess 40Ar contamination does exist in Connemara in minerals—particularly in amphibole as noted above—attempts to isolate such components and correct for them using inverse isotope correlation diagrams were hampered by uniformly high radiogenic Ar yields for all samples. Fortunately, this characteristic means that calculated dates are relatively insensitive to our choice of initial 40Ar/39Ar. We interpret the dates reported here as cooling ages, but the presence of small amounts of excess 40Ar may make some of the dates close overestimates of the time of cooling below the effective closure temperature interval. 39Ar weighted mean ages and their 2σ errors were calculated for all total fusion and incremental release analyses. These ages represent the 39Ar volume-averaged total gas age of a sample. For incremental heating analyses, a plateau date is reported if at least three consecutive steps overlap within 2σ error and contain over 50% of the 39Ar released. For total fusion analyses, results from multiple analyses of an individual sample were combined statistically to give an apparent age distribution for the sample. These are shown as normalized probability distribution plots (e.g., Fig. SF2). Interpretation of the total fusion analyses was guided by the shape of the probability distribution. If the distribution showed a single mode, without secondary peaks or significant skewness, we assumed that the 39Ar weighted mean date of all analyses represents the cooling age of the sample. If the apparent age distribution is multi-modal or skewed, the 39Ar weighted
mean date may be geologically meaningless, and we interpret distinct peaks as the best indicator of age components represented by the fusion analyses. Skewness of peaks toward older dates can often suggest contamination by excess $^{40}$Ar, assuming that the excess component is distributed inhomogeneously such that multiple total fusion analyses of a sample will yield ages with variable excess. In this case, we regard the sample mode as a close minimum estimate of a samples’ cooling age (e.g., Fig. SF2).

Following Dodson (1973), we calculated closure temperatures for micas from existing constraints for Ar diffusivity and grain size by assuming a cooling rate (Table 4). Estimated closure temperatures used in this study are shown in Table 5.

Results of $^{40}$Ar/$^{39}$Ar and microprobe analyses

All sample locations are shown in Fig. 4, their coordinates are given in Tables 1–3. Results of geochemical microprobe and $^{40}$Ar/
Table 1. Summary of 40Ar/39Ar analyses for muscovite, biotite, and phlogopite from northern Connemara.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rocktype</th>
<th>Locality (Nat’l. Grid Ref.)</th>
<th>Mineral</th>
<th>Grain radius (range) (µm)</th>
<th>No. of grains per analysis or weight</th>
<th>40Ar weighted mean date (Ma)</th>
<th>2σ error (Ma)</th>
<th>Mode (range) (Ma)</th>
<th>Plateau or flat segment (Ma)</th>
</tr>
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<tbody>
<tr>
<td>AF7</td>
<td>Metapelite 693.636</td>
<td>Muscovite</td>
<td>250–355</td>
<td>1–4</td>
<td>476.4</td>
<td>3.7</td>
<td>471–476</td>
<td>—</td>
<td></td>
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<tr>
<td>AF7</td>
<td>Metapelite 670.637</td>
<td>Biotite</td>
<td>250–355</td>
<td>1–5</td>
<td>468.4</td>
<td>3.1</td>
<td>469–476</td>
<td>—</td>
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<tr>
<td>AF12</td>
<td>Metapelite 663.607</td>
<td>Biotite</td>
<td>125–250</td>
<td>2–5</td>
<td>481.5</td>
<td>4.3</td>
<td>471–483</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF15</td>
<td>Mylonite —</td>
<td>Muscovite</td>
<td>—</td>
<td>—</td>
<td>452.0</td>
<td>1.1</td>
<td>452.5±1.4</td>
<td>—</td>
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</tr>
<tr>
<td>AF49</td>
<td>Metapelite 804.574</td>
<td>Muscovite (crappy)</td>
<td>125–150</td>
<td>3.0 mg</td>
<td>450.7</td>
<td>1.2</td>
<td>455, 495</td>
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Garnet-staurolite zone

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rocktype</th>
<th>Locality (Nat’l. Grid Ref.)</th>
<th>Mineral</th>
<th>Grain radius (range) (µm)</th>
<th>No. of grains per analysis or weight</th>
<th>40Ar weighted mean date (Ma)</th>
<th>2σ error (Ma)</th>
<th>Mode (range) (Ma)</th>
<th>Plateau or flat segment (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AF67</td>
<td>Semi-pelite 760.554</td>
<td>Phlogopite (w/altered)</td>
<td>125–150</td>
<td>2.1 mg</td>
<td>399.0</td>
<td>3.0</td>
<td>448.7±1</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF37</td>
<td>Calcsilicate 842.540</td>
<td>Phlogopite</td>
<td>100–150</td>
<td>4.7 mg</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF35</td>
<td>Metapelite 806.506</td>
<td>Biotite</td>
<td>125–150</td>
<td>0.4 mg</td>
<td>454.7</td>
<td>3.3</td>
<td>459.8±2.3</td>
<td>—</td>
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</tr>
</tbody>
</table>

Staurolite-sillimanite zone

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rocktype</th>
<th>Locality (Nat’l. Grid Ref.)</th>
<th>Mineral</th>
<th>Grain radius (range) (µm)</th>
<th>No. of grains per analysis or weight</th>
<th>40Ar weighted mean date (Ma)</th>
<th>2σ error (Ma)</th>
<th>Mode (range) (Ma)</th>
<th>Plateau or flat segment (Ma)</th>
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</thead>
<tbody>
<tr>
<td>AF35</td>
<td>Metapelite 849.538</td>
<td>Biotite</td>
<td>125–250</td>
<td>1–7</td>
<td>472.2</td>
<td>4.3</td>
<td>466–474</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF37</td>
<td>Calcsilicate 842.540</td>
<td>Phlogopite</td>
<td>100–150</td>
<td>4.7 mg</td>
<td>476.3</td>
<td>4.5</td>
<td>483</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF66</td>
<td>Calcsilicate 760.554</td>
<td>Phlogopite (w/altered)</td>
<td>125–150</td>
<td>2.1 mg</td>
<td>399.0</td>
<td>3.0</td>
<td>485±1.6 (432)</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF67</td>
<td>Semi-pelite —</td>
<td>Muscovite</td>
<td>—</td>
<td>—</td>
<td>452.1</td>
<td>2.4</td>
<td>448.7±1.8</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF70</td>
<td>Metapelite 914.508</td>
<td>Biotite</td>
<td>125–250</td>
<td>2.8 mg</td>
<td>459.0</td>
<td>0.7</td>
<td>466.8±1.0</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF73</td>
<td>Metapelite 915.508</td>
<td>Biotite</td>
<td>200–350</td>
<td>1–2</td>
<td>441.2</td>
<td>1.0</td>
<td>443, 473</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF79</td>
<td>Metapelite —</td>
<td>Biotite</td>
<td>150–350</td>
<td>1–2</td>
<td>472.8</td>
<td>1.0</td>
<td>473±1</td>
<td>—</td>
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</table>

Note: Preferred values are noted by bold type.

Table 2. Summary of 40Ar/39Ar analyses for biotite and phlogopite from central Connemara.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rocktype</th>
<th>Locality (Nat’l. Grid Ref.)</th>
<th>Mineral</th>
<th>Grain radius (range) (µm)</th>
<th>No. of grains per analysis or weight</th>
<th>40Ar weighted mean date (Ma)</th>
<th>2σ error (Ma)</th>
<th>Mode (range) (Ma)</th>
<th>Plateau or flat segment (Ma)</th>
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</thead>
<tbody>
<tr>
<td>AF28</td>
<td>Calcsilicate 835.490</td>
<td>Phlogopite</td>
<td>250–350</td>
<td>3–6</td>
<td>468.3</td>
<td>2.5</td>
<td>443, 479</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF38</td>
<td>Calcsilicate 836.492</td>
<td>Biotite</td>
<td>225–450</td>
<td>4</td>
<td>453.9</td>
<td>1.7</td>
<td>442, 474, 472</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF35</td>
<td>Calcsilicate 811.500</td>
<td>Phlogopite</td>
<td>300</td>
<td>1–4</td>
<td>473</td>
<td>2.0</td>
<td>460–465</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF34</td>
<td>Metapelite 805.501</td>
<td>Biotite</td>
<td>300–400</td>
<td>Furnace</td>
<td>444.9</td>
<td>2.9</td>
<td>448.5±2.0</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF35</td>
<td>Metapelite 806.506</td>
<td>Biotite</td>
<td>150–350</td>
<td>1–5</td>
<td>459.3</td>
<td>6.9</td>
<td>460–465</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF38</td>
<td>Calcsilicate 718.531</td>
<td>Phlogopite</td>
<td>400–1250</td>
<td>1</td>
<td>466.9</td>
<td>2.2</td>
<td>472</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>AF52</td>
<td>Calcsilicate 856.505</td>
<td>Phlogopite</td>
<td>125–250</td>
<td>9.4 mg</td>
<td>454.8</td>
<td>1.0</td>
<td>456.7±2.6</td>
<td>—</td>
<td></td>
</tr>
</tbody>
</table>

Note: Preferred values are noted by bold type.

39Ar analyses are presented in Tables ST1–ST3, and ST4, respectively, and the Argon mineral ages are summarized in Tables 1–3. Figures SF1 and SF2 show the plateau and total fusion results. The spatial distribution of the 40Ar/39Ar dates are shown in Fig. 5. In general, 40Ar/39Ar dates of biotite and muscovite are the oldest in the sillimanite–K-feldspar zone in southern Connemara (Table 1), younger in central Connemara (Table 2), and the youngest in the migmatitic rocks of southern Connemara (Table 3) where they also show the largest range in dates.

Interpretation of 40Ar/39Ar mica dates across Connemara

Because the interpretation of these results depends on their regional geological context, the data collected in this study are summarized and interpreted individually and in geographic order, from the garnet–staurolite zone in northern Connemara to the staurolite–K-feldspar zone in southern Connemara (Figs. 5 and 6). Below we also provide a justification for our sampling strategy in the context of earlier published work.

Northernmost garnet–staurolite zone

To test the interpretation of Boyle and Dawes (1991), based on the Elias et al. (1988) K–Ar data, that cooling from ~500–300 °C in northern Connemara occurred over the 470–430 Ma interval, we analyzed coexisting muscovite and biotite from two garnetiferous rocks from northernmost Connemara.

Furnace incremental heating experiments for muscovite (XKms = 0.73) and biotite (XAnm = 0.47) from sample AF16 yield essentially flat spectra but no statistically defined plateaus (Table 1; Fig. SF1). The weighted mean 40Ar/39Ar dates for the flat segments of these spectra are 471.0 ± 1.0 and 462.5 ± 1.1 Ma for muscovite and biotite, respectively. Total fusion results for larger muscovite (XKms = 0.67) and biotite crystals from metapelite AF7 show symmetric apparent age distributions with weighted mean dates of 476.4 ± 3.7 and 468.4 ± 3.1 Ma for these minerals, respectively (Table 1; Fig. 5; Fig. SF1).

We interpret the 40Ar/39Ar dates of all four analyses as cooling ages. The age differences among the four minerals are consistent with differences in closure temperatures related to differences in grain size. Collectively, the data are consistent with cooling at a moderate rate over the ca. 476–463 Ma interval, and we found no evidence to support protracted cooling in northernmost Connemara.

Garnet-staurolite and staurolite–sillimanite transition zones

Previously published K–Ar ages from these zones range between 450 and 400 Ma for biotite and between 470 and 430 Ma for muscovite (Elias et al. 1988). To evaluate the significance of this spread in dates, we analyzed muscovites, biotites, and phlogopites from a variety of metasedimentary rocks (Tables 2 and 3; Fig. 5; Figs. SF1 and SF2).

Most 40Ar/39Ar analyses yield 39Ar weighted mean dates older than 470 Ma, but a few muscovites, one biotite, and one of the phlogopites yield 40Ar/39Ar dates that are significantly younger. We first report the >470 Ma dates, and then discuss the significance of the younger results.
Table 3. Summary of $^{39}$Ar/$^{40}$Ar analyses for muscovite and biotite from southern Connemara.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Mineral</th>
<th>Grain radius (mm)</th>
<th>No. of grains per analysis or weight</th>
<th>Mode of grain releases</th>
<th>No. of analyses</th>
<th>No. of $^{39}$Ar weighted mean date (Ma)</th>
<th>$^{39}$Ar weighted mean date (Ma)</th>
<th>Error (Ma)</th>
<th>Mode error (Ma)</th>
<th>Plateau of flat segment (Ma)</th>
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<tbody>
<tr>
<td>AF16</td>
<td>Metapelite</td>
<td>&gt;1000</td>
<td>10 Mapping</td>
<td>10</td>
<td>10</td>
<td>236.8 ± 3.8</td>
<td>236.8 ± 3.8</td>
<td>5.1</td>
<td>3.3</td>
<td>453 ± 11</td>
</tr>
<tr>
<td>AF18</td>
<td>Metapelite</td>
<td>833 ± 45</td>
<td>10 Laser step</td>
<td>10</td>
<td>10</td>
<td>457.4 ± 2.7</td>
<td>457.4 ± 2.7</td>
<td>5.1</td>
<td>3.3</td>
<td>445 ± 12</td>
</tr>
<tr>
<td>AF22</td>
<td>Metapelite</td>
<td>833 ± 45</td>
<td>10 Laser step</td>
<td>10</td>
<td>10</td>
<td>457.4 ± 2.7</td>
<td>457.4 ± 2.7</td>
<td>5.1</td>
<td>3.3</td>
<td>445 ± 12</td>
</tr>
<tr>
<td>AF35</td>
<td>Phlogopite</td>
<td>&gt;1000</td>
<td>10 Laser step</td>
<td>10</td>
<td>10</td>
<td>457.4 ± 2.7</td>
<td>457.4 ± 2.7</td>
<td>5.1</td>
<td>3.3</td>
<td>445 ± 12</td>
</tr>
<tr>
<td>AF37</td>
<td>Phlogopite</td>
<td>833 ± 45</td>
<td>10 Laser step</td>
<td>10</td>
<td>10</td>
<td>457.4 ± 2.7</td>
<td>457.4 ± 2.7</td>
<td>5.1</td>
<td>3.3</td>
<td>445 ± 12</td>
</tr>
<tr>
<td>AF38</td>
<td>Phlogopite</td>
<td>&gt;1000</td>
<td>10 Laser step</td>
<td>10</td>
<td>10</td>
<td>457.4 ± 2.7</td>
<td>457.4 ± 2.7</td>
<td>5.1</td>
<td>3.3</td>
<td>445 ± 12</td>
</tr>
<tr>
<td>AF40</td>
<td>Biotite</td>
<td>833 ± 45</td>
<td>10 Laser step</td>
<td>10</td>
<td>10</td>
<td>457.4 ± 2.7</td>
<td>457.4 ± 2.7</td>
<td>5.1</td>
<td>3.3</td>
<td>445 ± 12</td>
</tr>
</tbody>
</table>

Note: Preferred values are noted by bold type.

Table 4. $^{39}$Ar diffusion data for musc.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>$D_a$ (cm²/s)</th>
<th>$E_a$ (kcal/mol K)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Biotite</td>
<td>0.54</td>
<td>0.77</td>
<td>47.1±1.5</td>
</tr>
<tr>
<td>Muscovite</td>
<td>0.71</td>
<td>0.4</td>
<td>50.5±2.2</td>
</tr>
<tr>
<td>Phlogopite</td>
<td>0.00039</td>
<td>43</td>
<td>57.9±2.6</td>
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</tbody>
</table>

Note: Total fusion analyses of biotite ($X_{Ann} = 0.41$) from one staurolite–garnet schist (AF12), biotites from three staurolite–garnet–sillimanite schists (AF35, $X_{Ann} = 0.37$; AF70, $X_{Ann} = 0.43$; AF79, $X_{Ann} = 0.45$), and phlogopites from two calcisilicate rocks (AF37, $X_{Ann} = 0.06$; AF40, $X_{Ann} = 0.01$) yielded similar dates (Table 1; Fig. SF2). Biotite AF12 has a $^{39}$Ar weighted mean date of 478.0 ± 5.1 Ma, which broadly overlaps with the sample mode of 471–483 Ma. Total fusion analyses of the other muscials yield $^{39}$Ar weighted mean dates of 472.2 ± 4.3 (AF35), 474.6 ± 5.4 (AF70), 476.3 ± 4.5 (AF37), and 472.9 ± 2.4 Ma (AF40). More precise incremental release analyses of biotite AF70 and AF79 yielded $^{39}$Ar weighted mean dates of 466.8 ± 1.0 and 473.1 ± 1.1 Ma for ~70% of the released gas. In all cases except the AF40 phlogopite, these dates broadly overlap with the sample modes of the respective probability distributions, so we interpret them to yield approximate cooling ages (cf. Table 1). In contrast, the probability distribution of phlogopite AF40 shows distinctive modes at 455 and 495 Ma, inconsistent with the $^{39}$Ar weighted mean age, indicating that the sample is probably contaminated by heterogeneously distributed excess $^{40}$Ar.

Younger $^{40}$Ar/$^{39}$Ar dates were obtained for musc in samples collected in or near fault zones. Muscovite AF15 from a mylonitic metasedimentary rock collected within the Renville–Bofin Slide (Fig. 5). Conventional furnace analysis of this sample yielded a plateau date of 452.5 ± 1.4 Ma. Given the probability that the mylonitic fabric developed at low temperature (for example, there is no evidence for plastic deformation of feldspars in thin section) and the fact that the muscovite helps define the mylonitic fabric, we interpret 452 Ma as a close estimate of the time of mylonitization. Two additional samples were collected near smaller faults that offset the Renville–Bofin Slide. The incremental release spectra of muscovite ($X_{KMs} = 0.6$) and biotite ($X_{Ann} = 0.39$) from a strongly foliated garnet schist rock (AF49) displayed flat segments with mean dates of 452.3 ± 1.0 and 458.1 ± 2.3 Ma, respectively (Table 1; Fig. SF2). Muscovite from another garnet schist (AF67, $X_{KMs} = 0.64$) has a plateau date of 448.7 ± 1.8 Ma. The similar
muscovite dates suggest an important phase of post-M3 deformation and dynamic recrystallization at ca. 450 Ma. The older AF49 biotite date probably reflects excess ⁴⁰Ar.

Another anomalously young date was provided by phlogopite from a calc-silicate rock (AF66). Its spectrum displayed a flat segment with a ³⁹Ar weighted date of 418.5 ± 1.6 Ma. This date is similar to many K–Ar dates obtained by Elias et al. (1988) from the same region. Most likely, it reflects local resetting by hydrothermal fluids related to intrusion of the post-orogenic Galway batholith.

Sillimanite–K-feldspar zone

In this zone, we concentrated on biotites and phlogopites from metasedimentary samples collected near Derryclare Lough area in central Connemara (Fig. 5; Figs. SF1 and SF2; Table 2). Biotite AF34 (Xₐₙₜ = 0.29) from a sillimanite–K-feldspar–biotite gneiss has a plateau date of 448.5 ± 2.0 Ma. Phlogopite AF38 (Xₐₙₜ = 0.04) from a metasomatic diopside rock yields a plateau date of 455.8 ± 1.3 Ma, consistent with a U–Pb date of ca. 462 Ma for titanite from the same sample (Friedrich 1998; Friedrich et al. 1999b). Phlogopite AF52 (Xₐₙₜ = 0.02) from another metacarbonate rock has a more complicated release spectrum with a flat segment corresponding to a date of about 457 Ma. We interpret the plateau and flat segment dates of these three samples as cooling ages.

These cooling ages are significantly younger than those of trioctahedral micas with similar compositions and grain sizes from the garnet–staurolite and staurolite–sillimanite transition zones of northern Connemara (Tables 1 and 2; Fig. 5). The boundary between the cooling age provinces broadly coincides with the muscovite-breakdown isograd and the crest of the Connemara antiform.

Several additional samples from the sillimanite–K-feldspar zone yielded ⁴⁰Ar/³⁹Ar dates that were complicated and difficult to interpret unambiguously. The frequency distribution of laser fusion dates for phlogopite AF27 (Xₐₙₜ = 0.05) shows distinctive modes at 479 and 443 Ma. Petrographically, this sample contains phlogo-
pites of two different grain sizes, one somewhat larger than 500 μm and one significantly smaller. It may be that the two modes in the release spectrum correspond to different cooling ages for the two phlogopites. However, the ca. 479 Ma date seems inconsistent with other data from the area, particularly U–Pb titanite dates that may record cooling through a much higher closure temperature (≥650 °C; Friedrich et al. 1998) roughly 15 million years later. This older mode may reflect some degree of excess 40Ar contamination, but Friedrich (1998) provided textural evidence that the titanites crystallized hy-

**Fig. 6.** Time–space diagram of 40Ar/39Ar dates for the Connemara region. The data are grouped into four categories, based on their relationship to the geological context. Constraints on the timing of major regional heating significantly above the Argon–mica closure temperatures were provided by U–Pb zircon dating of the Cashel–Lough Wheelaun and the Currywongaun mafic-ultramafic plutons (Friedrich et al. 1999a). Older 40Ar/39Ar dates are interpreted to yield excess Argon (Category 4). The shaded areas (Category 1) contain all permissible cooling ages as constrained by the age of the magmatic complex, peak metamorphism (Friedrich et al. 1999b), fluid infiltration (Category 3), and the Silurian hiatus. Category 2 dates may represent extensional unroofing of the Connemara orogenic wedge. The north–south axis is not drawn to scale.
drothermally at low ambient temperatures, perhaps related to far-reaching fluid circulation related to emplacement of the Oughterard granite.

A similar interpretation can be proposed for data from another phlogopite separate (AF32, $X_{\text{ann}} = 0.03$; Fig. SF2c); although its laser fusion dates have a $^{39}$Ar weighted mean of 473.7 ± 2.3 Ma, the frequency distribution is positively skewed away from a single mode between 460 and 465 Ma. Biotite AF28 ($X_{\text{ann}} = 0.42$) from a biotite–sillimanite–garnet schist has a $^{39}$Ar weighted mean date of 453.9 ± 21.7 Ma, but its population distribution displays three modes. The youngest, at 442–447 Ma, is similar to the young mode for the AF27 phlogopite, and this range may reflect localized resetting of Ar systems by hydrothermal fluid interactions or dynamic recrystallization during deformation.

**Migmatitic portions of the sillimanite–k-feldspar zone**

Our $^{40}$Ar/$^{39}$Ar research on magmatic rocks of the sillimanite–K-feldspar zone was designed to augment data presented elsewhere by Friedrich et al. (1999a, 1999b) pertinent to the detailed thermal evolution of southern Connemara. Our earlier work suggested relatively rapid cooling over the 460–377 °C temperature interval between 460 and 454 Ma near the northern distribution limit of anatectic melt products in the sillimanite–K-feldspar zone. As described in the following paragraphs, the $^{40}$Ar/$^{39}$Ar dates for samples collected closer to the main outcrop region of the Connemara igneous complex are notably younger than the mica ages reported by Friedrich et al. (1999b) and much younger than the crystallization ages of igneous phases in the Connemara complex.

Previous workers found abundant evidence that the K–Ar hornblende chronometer cannot be applied with confidence to amphibolites from southern Connemara (Miller et al. 1991). The most commonly proposed explanation for the tremendous range in hornblende apparent ages was the disturbance of isotopic systematics by hydrothermal fluids related to intrusion of the ca. 400 Ma Galway batholith and related “post-orogenic” plutons (Leggo et al. 1996). To seek additional support for this hypothesis, we analyzed micas from metapelitic schists and older granitic dikes collected at various distances from late plutons. In general, the muscovites yielded relatively simple release spectra with plateaus or flat segments, whereas the biotite results were much more complex.

Biotite AF22 is from a schist collected 10 km away from the post-orogenic Roundstone granite (Fig. 5). Petrographic analysis provided no evidence of hydrothermal alteration (Fig. SF3). Laser fusion analyses yielded symmetric frequency distributions with $^{39}$Ar weighted mean dates of 453 and 440 Ma for the ~400 and ~150 μm size fractions, respectively (Table 2), consistent with a lower closure temperature for the smaller size fraction. Substantially older than the 420 Ma crystallization age of the granite, these dates probably represent cooling after M3 metamorphism.

Collected 5 km south of the AF22 outcrop, AF18 is a garnet–K-feldspar schist containing biotite ($X_{\text{ann}} = 0.53$), secondary muscovite ($X_{\text{ann}} = 0.74$) that replaces K-feldspar, and chlorite that fills fractures in garnet (Fig. 4). Incremental heating analysis of coarse (900 μm) muscovite yielded a plateau date of 456.5 ± 2.2 Ma, whereas a finer-grained fraction (450 μm) had a plateau date of 453.6 ± 0.9 Ma (Fig. SF2c). Laser fusion analyses of fragments of a 450 μm single crystal of this muscovite had a $^{39}$Ar weighted mean date of 457.4 ± 3 Ma, which lies within the uncertainty of the two other analyses. The similarity of these ages with those obtained for M3 micas further north in the “migmatite” zone of Connemara suggests that retrogression in AF18 occurred at a relatively high temperature. Laser total fusion analyses for two aliquots of biotite ($X_{\text{ann}} = 0.53$) from this sample have $^{39}$Ar weighted mean ages of 489.1 ± 3.3 and 275.6 ± 5.1 Ma (Table 3; Fig. SF1d). To better understand this complex behavior, we also performed a furnace incremental heating experiment on a 10 mg biotite aliquot with the resistance furnace, and a laser incremental heating experiment on a single 750 μm crystal (Table 3; Fig. SF1d). Both experiments yielded similar $^{39}$Ar weighted mean dates of 446.8 ± 2.3 and 445.3 ± 2.1 Ma, respectively. In both cases, the spectra displayed similar stair-step shapes with age components of ca. 508 and ca. 365 Ma (Fig. SF1f).

Given U–Pb constraints that M3 occurred over the 460–474 Ma interval, the 508 Ma component must reflect contamination with excess $^{40}$Ar. The ca. 365 Ma component is much younger than other biotite cooling ages from this area and does not correspond to a previously identified thermal event. This $^{40}$Ar/$^{39}$Ar date, however, may correspond to one of several low-temperature fluid alteration events that affected the Connemara region after emplacement of the Galway granites (cf. Jenkin et al. 1992).

We also analyzed micas from two samples collected less than 5 km east of the Roundstone granite (Fig. 4). Sample AF45 is from the paleosome of an anatectic pelitic gneiss that surrounds pegmatitic leucosome AF44. Furnace incremental heating analysis of the biotite ($X_{\text{ann}} = 0.49$) from the leucosome yields a plateau date of 418.4 ± 6.5 Ma. Given the crystallization age of this pegmatite (467 ± 2 Ma, U–Pb zircon; Friedrich et al. 1999b), this $^{40}$Ar/$^{39}$Ar date probably reflects complete resetting during intrusion of the Roundstone granite, and possibly limited resetting at ca. 350 Ma or younger.

The $^{40}$Ar/$^{39}$Ar systematics of the biotite ($X_{\text{ann}} = 0.49$) in AF45 are more complex (Figs. SF1 and SF2d). Laser fusion analyses of aliquots of up to seven grains have a $^{39}$Ar weighted mean date of 434.5 ± 3.1 Ma. However, frequency distribution plots show three distinct modes at ca. 420, 437, and 463 Ma (Fig. SF1f). Laser incremental heating analysis of another aliquot yielded a stair-step profile with a flat segment at 420.1 ± 6.5 Ma and a minimum age of 327 ± 20 Ma (Fig. SF1f). The simplest interpretation of this spectrum is that the sample experienced a major Ar re-equilibration event at ca. 420 Ma followed by a second, less profound event at ≤330 Ma. Muscovite ($X_{\text{ann}} = 0.63$) in AF45 occurs as fine-grained pimite and as large (up to 3000 μm) randomly oriented, secondary muscovite books (Fig. SF3d). Laser incremental heating analyses of different size fractions yielded simple plateaus or flat segments corresponding to dates between 439 and 429 Ma (Fig. SF1f; Table 3). Large (700 μm) crystals gave the oldest dates, whereas the smallest crystals (200 μm) gave the youngest dates. Despite the positive correlation between grain size and apparent age, these cannot be closure ages related to simple cooling after M3 metamorphism because AF45 is from a structural level below the older Silurian unconformity (Fig. 4). Conceivably, the young ages imply partial resetting associated with intrusion of the Roundstone granite.

**Significance of the $^{40}$Ar/$^{39}$Ar mica age pattern**

The $^{40}$Ar/$^{39}$Ar mica dates for Connemara fall into four interpretable groups (Fig. 6), based on the U–Pb geochronological constraints of the main intrusive phases (Friedrich et al. 1999a), the timing of peak metamorphism (Friedrich et al. 1999b), and other previously published constraints from the geological record (e.g., Cliff et al. 1993, 1996; Ryan and Dewey 2011; Chew and Strachan 2013, Wellings 1998; Boyle and Dawes 1991; Yardley et al. 1987). The last Category can be defined because the ca. 400–420 Ma post-orogenic granites are the youngest, well-dated, potential causes for resetting, and because dates much older than ca. 475 Ma are inconsistent with U–Pb zircon constraints on the age of the early Grampian Cashel–Lough Wheelaun gabbro (Friedrich et al. 1999a).

**Category 1 ages**

Micas that record the thermal relaxation of the Grampian orogen show a spatial pattern opposite that proposed by Elias et al. (1988) (Figs. 5 and 6), but our study does not include any argon results for hornblende. The oldest reliable Category 1 ages—ranging from 478 to 467 Ma—were obtained for garnet–staurolite zone and staurolite–sillimanite transition zone samples collected in northern Connemara. To the south, toward the Connemara...
igneous complex, Category 1 biotite ages decrease to as young as 453–440 Ma in the highest temperature portions of the sillimanite–K-feldspar zone. The only complication in this pattern occurs along the extreme northern margins of the Dalradian outcrop region, where biotite ages from staurolite-zone rocks are slightly younger (468–463 Ma) than those a few kilometers to the southeast in the staurolite–sillimanite transition zone (478–467 Ma). One possible explanation for the relatively young ages in northernmost Connemara is limited argon degassing related to late extensional deformation (Boyle and Dawes 1991).

Category 1 ages are defined as related to cooling subsequent to the M3 event in northern Connemara and to cooling following anatexis in southern Connemara (Figs. 6 and 7). No significant discontinuity occurs in the pattern across the sillimanite isograd (Figs. 5 and 6), which has been suggested by some authors as corresponding to the effective northern distribution limit of prograde M3-metamorphic assemblages (Yardley 1976; Tanner and Shackleton 1979). Even if the northern limit of prograde M3-metamorphism extends only as far north as sillimanite-in isograd, as has been proposed elsewhere (Yardley et al. 1987), the thermal effects of M3 extended throughout Connemara (cf. Boyle and Dawes 1991).

Three explanations could satisfy the general north-to-south younging of Category 1 ages. The first, which is consistent with some U–Pb titanite and monazite dates of less than 470 Ma in central and southern Connemara (Cliff et al. 1996; Friedrich et al. 1999b; Friedrich 1998) is that the older 40Ar/39Ar dates of northern Connemara represent contamination by excess 40Ar. We cannot disprove this interpretation with the data at hand; our samples were so radiogenic that the use of isotope correlation diagrams to evaluate this problem was impractical. However, the consistency of most northern Connemara 40Ar/39Ar ages from a single subarea is not a pattern typical of pervasive excess 40Ar contamination. Most well-documented examples of sample suites contaminated with excess 40Ar are characterized by irreproducible results for multiple aliquots of the same sample, wide variations in dates from single outcrops, and (or) complex release spectra (Harrison and McDougall 1981; von Blanckenburg and Villa 1988; Arnaud and Kelley 1995). Because many of the young titanites found in Connemara are thought to have grown during late- or post-M3 metasomatic events (Friedrich 1998; Friedrich et al. 1999b) and because we cannot preclude a similar origin for the few young monazites dated in the region, we see no reason to reject the 40Ar/39Ar evidence for the M3 metamorphism in northern Connemara at ca. 474–470 Ma.

A second possible explanation for the north–south age trend is that the region was unroofed diachronously—first in the north—about the east–west-trending Connemara antiform (e.g., Elias et al. 1988). This interpretation is consistent with the coincidence of the abrupt break in cooling ages with the axial trace of this antiform, but inconsistent with evidence from (i) paleomagnetic studies that northern Connemara had already cooled to below ~510–580 °C before folding (Morris and Tanner 1977; Robertson 1988), and (ii) structural studies which show that the antiform...
formed as a late (D4) feature (Leake et al. 1983). Furthermore, structural, petrologic, and thermobarometric constraints on palaeodepth trends during M3 suggest synchrony throughout Connemara region.

We favor a third explanation consistent with all available geochronologic and structural data: that intrusion of intermediate to felsic phases of the Connemara igneous complex over the ca. 468–463 Ma interval maintained M3 temperatures high in southern Connemara while regional cooling occurred in northern Connemara. Textural relations in dated metamorphic rocks further suggest that structures and fabrics commonly attributed to D3 are older in the north than in the south; for example, micas that define the S3 schistosity in northern Connemara had cooled well below ~400 °C while synkinematic micas were still growing at upper amphibolite facies in southern Connemara. While this model requires high lateral temperature gradients of up to 20 °C/km, we suggest that such a thermal regime would be consistent with an arc-collision setting like that of Connemara.

Results of thermal modeling studies further support our explanation. Reverdatto and Polansky (2004) found that a lateral thermal gradient of ~14 °C/km may be sustained over 5–6 Ma if a basaltic-type heat source is assumed to have occurred structurally above the Connemara complex. Dewey and Ryan (2016) concluded that Barrovian-type PT conditions, as observed in Connemara, may be achieved in under 5 Ma by an actively thrusting hot hanging wall over colder footwall. In their model, significant heating of the footwall is restricted to within 7 km of the thrust contact.

**Category 2 ages**

Category 2 ages may provide close minimum estimates on the age of the Renvyle–Bofin Slide and related late structures in central and northern Connemara (Figs. 5–7). If this interpretation is correct, extensive deformation in this part of the Grampian orogen was taking place no more than about 15 million years after peak metamorphism associated with large-scale crustal shortening and Barrovian-type metamorphism in the island arc – continent collisional environment.

**Category 3 ages**

Most Category 3 ages record resetting of the Ar systematics of micas throughout much of Connemara at ca. 420 Ma (Figs. 6 and 7). This event cannot be regarded as part of the Grampian orogeny, because, both, north- and south-vergent structures are much older and Dalradian rocks had been covered unconformably by marine Silurian sedimentary deposits by 435 Ma (Leake and Tanner 1994). Instead, the spread of Category 3 ages may be related to intrusion of granites that resulted in context of northward subduction beneath the Laurentian margin from Late Ordovician to Devonian time (Mayoian phase of Dewey and Ryan 2016). If this interpretation is correct, the thermal effects of the post-orogenic magmatic event extended far beyond southern Connemara, perhaps through a regional metasomatic mechanism similar to that invoked for older ca. 462 Ma metasomatic deposits in Connemara (Yardley et al. 1991; Cliff et al. 1993). Such a large regional extent of hydrothermal activity related to emplacement of the Galway batholith has been suggested by Jenkin et al. (1992) for retrograde alteration in southern Connemara.

Two other anomalous dates may be related to similar alteration events. The frequency distribution plots for AF27 phlogopite and AF28 biotite display modes between 450 and 440 Ma. These dates are significantly younger than those obtained for compositionally similar minerals at the same structural level in the sillimanite–K-feldspar zone. Neither AF27 or AF28 show evidence of post-metamorphic dynamic recrystallization, but irregularly spaced ~1–3 mm quartz-filled fractures occur orthogonal to the S3 foliation defined by biotite and phlogopite in these samples. Therefore, we tentatively attribute the 450–440 Ma dates to retrograde resetting. An extremely young near-plateau segment on the incremental heating spectrum for AF45 biotite (ca. 330 Ma) suggests that some alteration may have occurred long after intrusion of the Galway batholith and related plutons. However, the AF45 biotite incremental heating analysis was poor, with very large uncertainties, and laser analyses of the same sample yielded discrepant results.

**Thermal and structural evolution of Connemara**

We integrate the new 40Ar/39Ar data along with previously published geochronologic, structural, metatexitic, and paleomagnetic data and suggest a refined thermal and tectonic model of the Grampian orogeny at Connemara. We refer to three simplified developmental north–south cross sections (Fig. 8), which were constructed from youngest to oldest (see Appendix B), while discussion of the cross section follows geological convention, from oldest to youngest. The oldest section (Fig. 8A), drawn to represent the structural pattern at ca. 474–470 Ma, the time of syntectonic (D2) magmatism (see fig. 6 in Friedrich et al. 1999b and references therein), is dominated by a regional-scale recumbent fold of D2 age (Yardley et al. 1987). The D2 event, responsible for the fold nappe and the S2 schistosity throughout Connemara, was at least in part synchronous with M2 metamorphism. Although direct constraints on the initiation age and duration of D2 and M2 are limited, Wellings (1998) suggested that the ca. 474 Ma Currywon gaun gabbro intruded during the regional D2 and M2.

Numerous 40Ar/39Ar cooling ages for M3 sillimanite-grade micas in northern Connemara are within uncertainty of this age or only slightly younger at 475 Ma (Figs. 5, 6, and 7), so we conclude that the transition from D2/M2 to D3/M3 in northern Connemara occurred within a few million years. On a regional scale, this transition may simply represent a deformational continuum at progressively higher temperatures.

Both D3 and M3, as defined by F3 folds and S3 axial planar schistosity in Dalradian rocks, affected all of Connemara (Fig. 9; e.g., Barber and Yardley 1985; Treloar 1985; Yardley et al. 1987). Both D3 and M3 appear to have been very short-lived phenomena in northern Connemara (ca. 470 Ma), but they may have lasted 7–8 million years in southern Connemara as a result of more prolonged and more voluminous magmatic activity (470–463 Ma; Leake 1989; Friedrich et al. 1999b, 1999c).

M3 and D3 reached their peak intensities at 468–466 Ma in southern Connemara, coincident with widespread quartz diorite magmatism and anatexis of metasedimentary rocks (Fig. 8B; Tanner et al. 1997; Friedrich et al. 1999b). At that time, a very high thermal gradient persisted across Connemara, with temperatures of ≥750 °C in the southernmost sillimanite–K-feldspar zone and <300 °C in northern Connemara (Figs. 7, 8B, and 9). D3 structures of this age include an intense foliation in anatectic metapelitic rocks (Fig. 3F) and wide ductile zones, especially in southwestern Connemara above the Mannin thrust (Leake et al. 1983), both of which contrast significantly with the older F3 folds and tectonite fabrics that developed at lower temperatures throughout Connemara.

Figure 8C represents the regional structure after development of the D4 Connemara antiform, but before late extensional deformation. In combination with paleomagnetic data, the 40Ar/39Ar cooling ages provide tight constraints on the age of the antiform. Paleomagnetic analyses of mafic rocks indicate that the northern Connemara gabbros cooled below ~510–580 °C, the nominal range for the Curie temperature of magnetite, before folding of the Connemara antiform, whereas the southern Connemara gabbros cooled below this temperature range after development of the antiform (Morris and Tanner 1977; Robertson 1988). The north–south variation in Category I 40Ar/39Ar mica ages (Fig. 7) thus brackets formation of the Connemara antiform to between ca. 468 and ca. 463–462 Ma. Additional confirmation of the lower age bracket comes from the U–Pb crystallization age of 462.5 Ma for the post-F4 Oughterard granite (Tanner et al. 1997; Friedrich...
Fig. 8. Schematic evolutionary cross-sections summarizing the major thermal and structural evolution of the Connemara region in three time slices. Refer to Appendix B for details regarding the restorations (Friedrich 1998). The three panels emphasize the tight constraints on timing of deformation, metamorphism, and cooling. (A) Restoration starts at the time of syn-kinematic emplacement of mafic and ultramafic plutons (Wellings 1998). The main penetrative metamorphic fabrics (D2 and D3) formed at this time interval. The orientation of the regional-scale isograde implies more voluminous mafic magmatism in southern Connemara. (B) The interval between 468 and 467 Ma marks the time of peak Barrovian metamorphism with anatexis, a switch from northerly to southerly vergence of orogenic fabrics and structures, and a shift in magmatism from mafic to felsic (e.g., Leake and Tanner 1994). Radiometric dates are from Friedrich et al. (1999a, 1999b). [Colour online.]
The persistence of “late” M3 high-temperatures in southern Connemara during development of the F4 Connemara antiform may help explain the asymmetry in deformational style across the crest of the antiform, with more intense deformation and shearing in the southern limb (locally referred to as “the steep belt”, e.g., Leake and Tanner 1994). Formation of the antiform was linked kinematically with ductile deformation of the quartz diorites and gabbros of southern Connemara along an incipient Mannin thrust-shear zone that eventually resulted in formation of the brittle Mannin thrust.

At ca. 462 Ma, the waning stages of Connemara igneous complex activity maintained temperatures in the southern part of the region relatively high, perhaps more than 150 °C higher than temperatures in northern Connemara. However, extensive hydrothermal systems extended northward from the igneous complex, leading to local resetting of mica chronometers and the crystallization of metasomatic assemblages such as the titanite-bearing diopside rocks described by Yardley et al. (1991) and dated by Friedrich (1998). Similar episodes of metasomatic or hydrothermal alteration events appear to have occurred more recently in the history of the region, consistent with the occurrence of Category 3 dates (Fig. 6).

The youngest structures for which we have age constraints are the extensional faults of northern Connemara. Last movement along the Renyle–Bofin Slide occurred at ca. 452 Ma based on the 40Ar/39Ar muscovite crystallization age of mylonite AF15. This age is consistent with the recent reinterpretation of the structural age of this fault to be post-D3 (Wellings 1998). It seems likely that extensional faulting entirely postdates Grampian deformation at Connemara, but it was concurrent with rapid exhumation and cooling.

Discussion

We assimilated thermochronological (Fig. 7), structural (Fig. 8), and thermobarometric data in a refined PT-path (Fig. 9) and established relationships between the regional features and their geodynamic context (Fig. 10). We then discuss the arc-continent collision model proposed by Dewey (2005) and Dewey and Ryan (2016).

Constraints on the PT evolution of deformation

At Connemara, Dalradian metasedimentary strata of Laurentian affinity document progressive deformation (D1–D4) and metamorphism, which captured the transition from north- to south-vergent orogenic deformation under clockwise PT conditions (Fig. 9). First, multi-phase north-vergent deformation occurred under prograde- and peak-pressure conditions and coeval with emplacement of calcalkaline mafic to ultramafic plutons. The extra heat source lead to Barrovian metamorphism, crustal anatexis, and formation of migmatite under peak-temperature, but decreasing pressure conditions. Next, progressively localized south-vergent deformation accommodated crustal-scale shortening under decreasing PT conditions. Last, exhumation, extensional deformation, granitic intrusion, and hydrothermal circulation document the waning stages of the orogeny.

The thermal peak outlasted north-vergent fold-nappe formation such that cooling ages (Category I, Fig. 6) provide a lower limit on the timing of this deformation (D3). But because southern Connemara also experienced Barrovian metamorphism at higher temperatures, only the Category 1 cooling ages (475–470 Ma) of northern Connemara are useful to define the timing of D3 (Figs. 7 and 9). Exhumation with decompression of about 4 kbar must have occurred relatively rapid to limit cooling between ca. 474 and 472 Ma to <150 °C (Figs. 7 and 9). This yields an exhumation rate of ~7 km/Ma, which is comparable to exhumation of modern orogens such as Taiwan (Liu et al. 2000; Willett et al. 2003; Byrne et al. 2011).

In southern Connemara, U–Pb geochronological constraints place the anatectic Barrovian event at 468–467 Ma, coeval with intermediate quartz-diorite intrusions (Friedrich et al. 1998). Therefore, significant cooling from peak temperatures (~750 °C, Yardley et al. 1987) to the highest closure temperature inferred for...
Fig. 10. Schematic profiles of the island–arc continent collision (Grampian orogeny) in Connemara at five time intervals. (A) Plate boundary configuration across an intra-oceanic subduction zone prior to collision of the inferred Lough Nafooey arc with the Laurentian continent (Dewey 2005). (B–D) Reconstructed within the frame of the island arc – continent collision model (Dewey 2005; Dewey and Ryan 2016) and refined by the constraints provided by our data (e.g., Figs. 6, 8, and 9). We assume that the subducting Laurentian lithosphere steepened as it entered the sub-asthenospheric mantle, which acted as a geometric quasi-pin point to further reconstruction. The Laurentian slab and its extended margin (Dalradian) thereby moves under the oceanic island–arc lithosphere, equivalent to ophiolite obduction in the geological record as suggested by Dewey and Ryan (2016). (B) The sharp bend in the downdropping Laurentian plate facilitated detachment of crustal nappes (D2 and D3) and their northward emplacement over Laurentia, marking the onset of collision. (C) Continuing ocean-ward motion of Laurentia, driven by asthenospheric flow, resulted in south-vergent lithospheric necking and formation of upward-propagating lithospheric mantle shear zones, in both plates, followed by (D) thrusting of Laurentian mantle–lithosphere over oceanic mantle–lithosphere, demarking deep-seated collision and formation of the Grampian orogenic wedge. Continuing plate convergence results in further southward thrusting of Laurentia, slab detachment, upward propagation of mantle shear-zones, and final emplacement of the orogenic wedge over the former volcanic island arc. As the bi-vergent wedge shortens at depth, its rocks move upwards and are subject to erosion, horizontal extension, and collapse. (E) An Andean-type margin may have established in western Ireland (e.g., Dewey 2005) by ca. 450 Ma. By ca. 440 Ma, marine sedimentary rocks were deposited onto extended and exhumed mid-crustal rocks terminating the possibility of any protracted cooling at Connemara following the Grampian orogeny. Note: arrows indicate relative plate motion, schematically, not to scale.

mica in this study (−430 °C) occurred between ca. 468 and 460 Ma at a rate of −40 °C/Myr. Category I 40Ar/39Ar mica thermochronometers record cooling from −430 °C to <320 °C between ca. 460 and 452 Ma, at −10 °C/Myr. A lateral thermal gradient of up to 30 °C/km persisted between ca. 472 and 460 Ma (Figs. 7, 8B, and 9).

Formation of the major south-verging thrust and range-scaled immit postdate crustal anatectics, but it was finished by the time of granite intrusion in eastern Connemara at ca. 463 Ma (Friedrich et al. 1999a, 1999b). Therefore, D4 structures formed between 467 and 463 Ma. The interval between ca. 468 and 467 Ma best defines the reversal in polarity, i.e., the time at which activity on north-vergent structures diminished, while it increased along south-vergent structures.

**Geometric and kinematic boundary conditions**

To reconstruct the geometric context in which D2 and D3 fold nappes formed, we combined structural and P-Tt information (Figs. 8 and 9). The overall geometry of the Grampian structures at Connemara define the mid-crustal core of a bi-vergent orogenic wedge. A basal north-verging decollement (not exposed) may be inferred based on the existence of north-vergent D2 and D3 fold nappes (Fig. 8A). The basal south-verging structure is represented by the Mannin thrust and related features, such as the Clifden steep belt and the Connemara antiform (Fig. 8C). Decompression and exhumation started along basal north-vergent structures, as D2 and D3 nappes began to detach at ca. 474 Ma to a depth equivalent of ~5 kbar at ca. 470 Ma, and continued facilitated by south-vergent motion between ca. 467 and 463 Ma (Figs. 9 and 10). In southern Connemara, significant decompression occurred at elevated temperatures (above 400 °C), while in northern Connemara at least some decompression occurred at temperatures below 300 °C (cf. Boyle and Dawes 1991), resulting in an environment that favors brittle fracturing and hydrothermal fluid circulation.

To examine the relationship between the regional-scale structures and their geodynamic context, we rescaled the Connemara orogenic wedge to plate-boundary dimensions (cf. Fig. 8A and 10A). In Fig. 10, we suggest a permissible sectional configuration that would result from collision between the Laurentian margin and an intraoceanic island arc, as proposed by Dewey (2005). An important modification to this model was recently proposed by Ryan and Dewey (2011) and tested by Dewey and Ryan (2016). These authors no longer consider Connemara as an “outlier terrane”, which was emplaced to its present position following the Grampian orogeny, but that the Connemara crust should be considered in-situ relative to the Laurentian margin of western Ireland; it represents the thinned Laurentian margin which was overriden by the postulated Lough Nafooey island arc system, sedimentary remnants of which are preserved in the South Mayo basin to the north (Fig. 1; e.g., Ryan and Dewey 2011). This suggestion solves a long-standing problem in Connemara geology, thus allowing us to relate field observations and our analytical results to refine the proposed model (Fig. 10).

**Island arc – continent collision at Connemara (Grampian orogeny)**

The key elements of the arc-continent collision model proposed by Dewey (2005) are as follows: (i) northward-directed obduction and collision of an intraoceanic island-arc system (Lough Nafooey arc and South Mayo forearc region, Dewey 2005) as the passive continental Laurentian plate margin with its Dalradian sedimentary cover rocks entered an oceanward-facing subduction zone, followed by (ii) subduction polarity reversal, (iii) extensional collapse, roll back (cf. Clift et al. 2004), and (iv) formation of an active Laurentian margin. Dewey and Ryan (2016)’s suggestion that hot island-arc mantle of the overriding plate acted as a heat source to calcalkaline plutonism and Barrovian high-temperature metamorphism in the lower plate, is a key to connect local- and regional-scale observables. Below we critically address the evolution of Connemara in this context, referring to Fig. 10.

By 490 Ma, an intraoceanic island arc had formed (Lough Nafooey arc; Dewey and Mange 1999) and no later than 478 Ma the Laurentian continental margin and its Dalradian strata entered the subduction zone. By 474 Ma, a major low-angle shear zone formed at the plate interface (Fig. 10B), successively incorporating Dalradian cover strata as they detached from their basement. The sequence deformed internally as it moved north- and upwards, defining an incipient orogenic wedge. Exhumation may have occurred by some combination of thrusting and erosion of the oceanic arc in the hanging wall, which is consistent with thick deposits of volcanic arc debris in the South Mayo basin (e.g., Sheekey Formation, Ryan and Dewey 2011). Based on geometric arguments illustrated in Fig. 10B, we postulate that peak pressures should be higher in exposed southern Connemara rocks, although only limited evidence for this has been found to date (cf. Leake and Tanner 1994).

As Dalradian rocks were thrust below the oceanic arc – mantle, a transient inverted thermal gradient established. The hot oceanic arc – mantle acted as a heat source, which resulted in formation of calcalkaline mafic and ultramafic intrusions and—with a few million years—to Barrovian-type metamorphism (Dewey and Ryan 2016). These conditions were established by 468 Ma (Fig. 8; Friedrich et al. 1999a). Thermal modeling of this scenario by Reverdatto and Polansky (2004) and Dewey and Ryan (2016) implies that a hot mantle heat source must have been adjacent to, or obliquely on top of the Connemara metamorphic and magmatic complex from a southerly direction (Figs. 10B and 10C). At 468–467 Ma, a lateral thermal gradient of ~20–30 °C/km developed across Connemara (Figs. 7, 8B, and 10C). This gradient is a key supportive element of the model that a hot upper plate, which was continuously moving northwards, provided the heat for high-temperature metamorphism and anatectis in southern Connemara, while northern Connemara—being farther from the major
Heat source—cooled rapidly to below 300 °C. Figure 10C provides an internally consistent geometric solution for this scenario. The elevated temperatures also facilitate localization of deformation and disruption of earlier D3 fabrics by formation of incipient south-vergent deformation within the steep belt of southern Connemara (Fig. 8B).

To meet the inferred PT conditions requires actively obducting upper plate no thicker than ~20–25 km (Dewey and Ryan 2016). Therefore, the upper portion of the downgoing plate must be in low-angle contact with the overriding arc lithosphere, which limits any asthenospheric mantle that may be trapped at the interface (Fig. 10B). As collision proceeds, the oceanic arc – lithosphere detaches from its sublithospheric mantle along a postulated intramantle shear zone. This zone occurs near the contact with the mantle and is facilitated by a fairly sharp bend in the downgoing Laurentian plate, which could form by slab-steeping above a slow sub-asthenospheric mantle. This lithospheric bend is the geometric basis for the upcoming collision and formation of the second vergence direction, which enabled the reversal of subduction polarity. A similar geometric configuration has been observed under Taiwan (Carena et al. 2013).

The switch from north- to south-vergent structures in Connemara is one of the most fundamental features upon which the arc-continent collision model is built (Dewey 2005). Integration of structural, metamorphic, and geochronological data bracket the time of this polarity reversal to a short interval between 468 and 467 Ma (Figs. 9 and 10B; Friedrich et al. 1999b). In Fig. 10C, we illustrated a geometrically permissible solution, in which lithospheric necking in the downgoing plate occurs as a response to continuing plate convergence and incipient continent-ward subduction of the oceanic mantle – lithosphere.

Reversal of subduction polarity may be completed only once the former subducting plate tears away (Fig. 10D). This change in vergence of compressional structures is driven by the deep-seated collision between subducted Laurentian mantle – lithosphere with oceanic mantle – lithosphere in a mobile asthenospheric environment.

Timing of extension and slab roll-back

Dewey (2005) and Ryan and Dewey (2011) argued that extensional collapse of the orogen occurred at 466 Ma, based on the sudden appearance of staurolite-grade detritus in the sedimentary record of South Mayo. Because these authors cannot reconcile this observation in any other way, they argue that extensional faulting is an effective way to unroof high-grade metamorphic rocks, suddenly, without having first eroded lower-grade rocks. If correct, the first appearance of staurolite would mark the onset of extensional collapse, and in turn, slab roll-back as the underlying cause of extension (upper portion of the Rosroe Formation: 466 Ma; figs. 3 and 4C in Dewey 2005; fig. 13.2 in Ryan and Dewey 2011; fig. 3 in Dewey et al. 2015; fig. 3 in Dewey and Ryan 2016). This scenario implies that extensional collapse and slab roll-back would have occurred prior to formation of the south-vergent D4 structures, such as the Mannin thrust, which we find difficult to reconcile with the kinematic requirements that the deeper portions of the orogenic wedge were under compression at that time (Fig. 10C).

The presently exposed mid-crustal section in Connemara appears to have undergone the transition from shortening to extension at the scale of rock samples no earlier than 463 Ma. In the field, this transition may be marked by the post-D4 intrusion of granites and related evidence of fracturing and hydrothermal fluid circulation. Thus, extensional collapse of the orogenic wedge, and thus its cause, postdate formation of D4 south-vergent structures. A lower bound to significant late Grampian extension would be provided by Category 1 cooling ages for southern Connemara, which may reflect extensional unroofing of the orogen until ca. 452 Ma (Fig. 9).

We therefore suggest that the sharp contact between ophiolitic and staurolite-grade metamorphic detritus in the stratigraphic section of South Mayo at 466 Ma marks the time at which erosional unroofing of the orogen exhumed the major thrust that juxtaposed the hot ophiolitic material and arc mantle against high-grade metamorphic rocks of the lower plate. The hypothesis of Dewey and Ryan (2016) that hot oceanic mantle was thrust over the Dalradian rocks and heated them from above is consistent with this conclusion. After 463 Ma, slab roll-back may have enhanced extension and morphologic decay of the orogenetic lid, requiring only one major phase of extension at ca. 450 Ma. The Silurian nonconformity may be a late surface expression of this process.

Timing of post-Grampian processes

Our geochronological results confirm that Grampian events at Connemara took place progressively and within a few million years (Friedrich et al. 1999a, 1999b), while our thermochronological results imply that cooling following the Grampian orogeny was locally rapid and finished long before deposition of marine Silurian rocks, as originally proposed by Fitch et al. (1964), but opposed by Dewey (2005). We postulate that the Grampian “tail”—our Category 3 (Fig. 6)—is not related to slow cooling related to denudation and erosion as proposed by Dewey (2005), and conclude that Silurian thermochronological dates bear no relevance to models of Grampian orogeny; they may relate to post-Grampian magmatism and hydrothermal circulation related to active continental margin processes, which were established by that time along the Laurentian margin (Fig. 10E).

Conclusions

The spatial pattern of the 40Ar/39Ar data presented here supports the previously reported large spread in mineral dates from the Connemara region. However, consideration of our data in the context of other petrographic, geochronologic, and stratigraphic constraints, suggests that some dates represent regional cooling subsequent to metamorphism, others record episodes of dynamic recrystallization, still others reflect alteration episodes, and a few are so contaminated by excess 40Ar as to be geologically meaningless. The regional cooling ages display patterns that can be linked directly to temporal and spatial variations in deformation, metamorphism, and magmatism. Northern Connemara cooled below ~300 °C by ca. 470 Ma, whereas southern Connemara was not at such low temperatures until ca. 450 Ma. This phenomenon probably was related to large lateral thermal gradients in an evolving arc-continent collisional setting. One important implication of the Connemara study is that regional trends in cooling ages may reflect regional variations in the orogenic temperature structure (such as distance to the heat source) rather than regional variations in the unroofing history, as is commonly assumed.

The significance of the geologic record to proper interpretation of mineral cooling ages cannot be overemphasized. If we knew less about the stratigraphic, intrusive, and structural history of Connemara, it would be a simple matter to interpret the broad spread in mineral ages from this region in terms of protracted cooling over nearly one hundred million years. This realization is a reminder that geochronology is of limited value in regions that have not been characterized adequately through careful field mapping and structural analysis.

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Appendix A. Analytical methods

We obtained muscovite, biotite, and phlogopite separates by standard crushing, sieving, Wilfley, magnetic, and heavy liquid separation. Two representative crystals of each sample were used to determine the major element compositions using a JEOL 733 electron microprobe operating with a beam current of 10 nA and an accelerating voltage of 15 kV. The remaining separates were cleaned ultrasonically with distilled water and wrapped in aluminium foil before irradiation at the McMaster University Reactor in Hamilton, Ontario. Fast neutron flux was monitored using MMhb hornblende (520.4 Ma, Samson and Alexander 1987). Corrections for interfering reactions were accomplished by including K$_2$SO$_4$ and CaF$_2$ and KCl salts in the irradiation packages. All samples were analyzed by a MAF 215-50 gas source mass spectrometer at MIT. Gas was extracted from the sample in three ways: furnace incremental heating, laser incremental heating, and laser fusion. The first method involved experiments on 5–50 mg samples in a double-vacuum resistance furnace for a period of 10 min per increment. Laser incremental heating was performed with a defocused, Coherent Ar-ion laser on 1–10 crystal aliquots. Progressive heating was accomplished by firing the laser for 2 min intervals at successively higher power levels until each sample melted. Laser total fusion of 1–30 crystals was achieved by firing a defocused laser beam for 30 s at high power (typically >20 W). Representative systems blanks for laser microanalysis for the M\(\alpha\)e 40, 39, 38, 37, 36 (moles) were 9 x 10$^{-18}$, 2 x 10$^{-17}$, 8 x 10$^{-18}$, 1 x 10$^{-17}$, and 8 x 10$^{-18}$.

Appendix B. Construction of schematic serial cross sections

The Connemara antiform plunges about 10° eastward over a distance of 80 km, exposing a roughly 10 km thick crustal section (Leake and Tanner 1994). This down-plunge view allows great insight into the three-dimensional structure of the Connemara antiform and the gabbro and gneiss complex, which facilitated a simple cross section construction. We started with the present-day configuration and added all metamorphic, palaeomagnetic, and geochronological constraints. Next, we used published knowledge about the timing and sense of motion along the main structures. Restoration of both older time slices, prior to 460 Ma, involved several assumptions. The Silurian unconformity is parallel to the structurally highest unit, rather than cutting across previously folded layers. This implies that the unconformity was filled together with the Dalradian block sometime after the Gramian orogeny. We assumed that the vertical offset on the Renyle–Bofin Slide of northern Connemara was minor, because there is no significant break in metamorphic grade across the fault. The geometry of unfolding of the F4 Connemara antiform was based on paleomagnetic data that required a 20° southwestern tilt of the F3 fold axes prior to F4 folding (Robinson 1988; Morris and Tanner 1977). Paleomagnetic constraints on the pre-F4 orientation of regional fold axes and geometric considerations require restoration of the main thrust late during the development of the Connemara antiform, a relationship shown schematically by a steep incipient Mannin thrust (Fig. 8). The present-day surface trace is shown on the cross sections as a reference line.