Mush ado about the Ratagain Complex, NW Scotland: insights into Caledonian granitic magmatism and emplacement from magnetic fabric analyses

A. Lawrence*, M. Maffione and C.T.E. Stevenson

School of Geography, Earth and Environmental Sciences, University of Birmingham, Birmingham B15 2TT, UK
© 2022 The Author(s). This is an Open Access article distributed under the terms of the Creative Commons Attribution 4.0 License (http://creativecommons.org/licenses/by/4.0/). Published by The Geological Society of London for EGS and GSG. Publishing disclaimer: www.geolsoc.org.uk/pub_ethics

Abstract: The anisotropy of magnetic susceptibility (AMS) is used to reveal subtle mineral alignment fabrics in apparently isotropic crystalline lithologies, including granites. Such petrofabrics can be produced by emplacement-related magma flow or post-emplacement tectonic strain. However, discriminating between flow-related and tectonic fabrics using field observations alone may be challenging and is usually a broad and arbitrary interpretation. In this contribution, we employ a range of magnetic analyses to characterize the origin of the petrofabric in the c. 425 Ma Ratagain Complex, NW Scotland, a composite Late Caledonian granitic intrusion. Our detailed magnetic analyses reveal that whilst all intrusive units carry an ambient tectonic overprint, critically, this has not developed into an obvious tectonic fabric and contains a horizontal shortening component indicative of transpression. This appears at odds with the well-defined Silurian (Scandian phase) regional transpressional tectonic regime from c. 420–415 Ma onwards. Accordingly, we suggest that either the complex is younger than previously thought or that it existed as a crystal-mush close to the magmatic solidus for a protracted period after its initial emplacement. This study lays the foundations for much-needed further investigations into the detailed emplacement mechanisms, timescales and petrogenesis of individual granitic intrusions, to aid understanding of Late Caledonian tectonics.

Supplementary material: Supplementary data to this article are available at https://doi.org/10.6084/m9.figshare.c.5941375

Thematic collection: This article is part of the Early Career Research collection available at: https://www.lyellcollection.org/cc/SJG-early-career-research

Received 14 November 2021; revised 7 April 2022; accepted 11 April 2022

The anisotropy of magnetic susceptibility (AMS) is an important tool used for reconstructing petrofabrics (i.e. preferred orientations and distributions of minerals) in granitic rocks where visible flow markers are often weak or completely absent (Tarling and Hrouda 1993; Bouchez 1997). For over six decades it has been documented that the patterns of magnetic foliation and lineation in a granitic intrusion (i.e. the magnetic fabric) can be coaxial with the petrofabric of the intrusion (e.g. King 1966; Heller 1973; Tarling and Hrouda 1993; Borradaille and Henry 1997; Nédélec and Bouchez 2015; Temporim et al. 2020). So-called ‘normal’ magnetic fabrics have a maximum susceptibility axis ($K_1$) that is parallel to the petrofabric lineation, whilst the minimum susceptibility axis ($K_3$) is generally found perpendicular to the petrofabric foliation. In magnetite-series granites (Ishihara 1977) normal fabrics are characterized by multidomain (MD – greater than 0.1 µm diameter) ferromagnetic minerals including magnetite, maghemite and other related cubic iron oxides (Rochette et al. 1992).

However, the interpretation of AMS data can be complicated by the presence of single-domain (SD – smaller than 0.1 µm diameter) ferromagnetic grains, which can result in an inverse magnetic fabric where $K_1$ and $K_3$ are swapped, hence showing $K_3$ parallel to the lineation and $K_1$ perpendicular to the foliation (Rochette 1988). In granites containing mixed populations of MD and SD magnetic particles, the AMS ellipsoid is unlikely to accurately describe the petrofabric (Potter 2004). It has even been suggested that in granites containing similarly aligned MD and stable SD particles the competing anisotropy signals could cancel each other out, resulting in very low anisotropy despite the rock being strongly magnetic (Potter and Stephenson 1988). Furthermore, if a granite comprises magnetic mineralogical species formed at different stages in the emplacement history, then several magnetic fabrics may be recorded by the same intrusive body and can often overprint one another (Benn 2010). For example, secondary alteration products may yield a magnetic fabric that is geometrically unrelated to that acquired during emplacement of the granite (Martín-Hernández and Ferré 2007). Therefore, determination of the primary magnetic fabric, which can serve as a proxy for the petrofabric of the intrusion, cannot be achieved solely through AMS analyses (Bouchez 1997; Potter 2004; Stevenson 2004). Instead, an accurate and reliable fabric analysis requires the separation of the different contributions of the paramagnetic–diamagnetic and ferromagnetic minerals to the magnetic fabric, and the reconstruction of the nature, shape and concentration of the magnetic carriers. Employing such a multifaceted approach to distinguish between primary and secondary fabrics, can also help to verify whether an entire intrusive body was affected by regional strain during its emplacement (Silva and Raposo 2020).
When the magnetic fabric has been tested with various magnetic techniques, the interpretation of what the fabric records requires detailed petrological observations (e.g. Anderson et al. 2018). Even then, there may be uncertainty about the forces that created the fabric in terms of emplacement-related magma body forces (‘magma flow’ e.g. Stevenson et al. 2007), or post-emplacement ambient tectonic strain (Burton-Johnson et al. 2019) or a combination of the two (e.g. Petronis et al. 2012).

In this contribution we present the results of a detailed analyses of the Ratagain Granite Complex, which includes AMS, anisotropy of anhysteretic remanent magnetization (AARM), thermomagnetic experiments, isothermal remanent magnetization (IRM) acquisition and coercivity unmixing, partial anhysteretic remanent magnetization (pARM) acquisition, optical microscopy and scanning electron microscope (SEM) observations. The results allow us to reconstruct the nature, main orientation, evolution and precise timing of the magnetic fabric recorded in the Ratagain Complex and contributes to our understanding of both the different mineral sources that confer AMS and the interplays between magmatic and tectonic processes in granitic fabric development (e.g. Burton-Johnson et al. 2019). Our integrated rock magnetic characterization of the Ratagain Complex thus provides an important case study that can be applied in investigations of other composite, granitic intrusions worldwide.

**Geological background**

The lower Paleozoic Caledonian Orogen of Scotland and Ireland resulted from the oblique continental convergence of Laurentia, Avalonia and Baltica and associated closure of the Iapetus Ocean (e.g. Dewey 1969; Lambert and McKerrow 1976; Pickering et al. 1988; Soper et al. 1992; Dewey and Strachan 2003; Corfu et al. 2014). Whilst the orogenic cycle comprised various episodes, the three major events of relevance to the Scottish Caledonides are the Grampian (c. 490–460 Ma), the Scandian (c. 435–425 Ma) and the Acadian (c. 400–380 Ma).

The Grampian Event involved collision of an intra-oceanic island arc with Laurentia (e.g. Chew and Strachan 2014; Johnson et al. 2017) and the deformation and metamorphism of Neoproterozoic–Cambrian (Dalradian) sediments deposited on the Laurentian continental margin (Mark et al. 2020), culminating in the development of major NW-directed fold and nappe complexes. Syn-collapse magnetic activity took place in the Grampian Event, with the intrusion of two-mica S-type granites (‘Older Granites’ – Read 1961; Oliver et al. 2008) and associated gabbroic rocks.

The collision of Laurentia and Baltica in the Scandian Event caused considerable regional thrusting and folding of Neoproterozoic (Moine) sedimentary rocks and associated basement inliers north of the Great Glen Fault (GGF; see Fig. 1), but such deformational effects are difficult to trace southwards (Dallmeyer et al. 2001; Kinny et al. 2003). Following the end of Iapetus subduction, the latest stages of the Scandian saw the onset of intrusion and emplacement of the Late Caledonian suite, a set of relatively undeformed granitic (sensu lato) bodies and coeval lamprophyre and hornblende-rich dykes (i.e. ‘appinites’) that intrude much of the lower Paleozoic Caledonian Orogen of Scotland and Ireland (Fig. 1; Archibald and Murphy 2021). It is widely held that the emplacement of many Late Caledonian intrusions was modulated by the Scandian sinistral stress regime, with crustal-scale thrusts and shear zones within transensional basins serving as intrusive conduits for the upwelling magma (Hutton 1988; Hutton and Reavy 1992; Stevenson et al. 2008; Miles et al. 2016; Strachan et al. 2020; Archibald and Murphy 2021).

Acadian transpression is ascribed to the collision of Iberia/Armorica with Avalonia and closure of the Rheic Ocean at c. 400–390 Ma (Woodcock et al. 2007). In Scotland, the Acadian is bracketed at 404–394 Ma (Miles et al. 2016; Woodcock et al. 2019) and was characterized by northward-directed, orthogonal contractual and sinistral strike-slip movements along major tectonic structures, including the Southern Upland, Highland Boundary and Great Glen faults (e.g. Phillips et al. 1995; Jones et al. 1997; Mendum and Noble 2010; see also Fig. 1). The intensity and effects of deformation were focused along the aforementioned fault zones and decreased with distance away from them (Mendum 2012).

The Ratagain Complex is a granitic Late Caledonian intrusion located along the shores of Loch Duich in the western Highlands of Scotland (Fig. 2; Nicholls 1951; Hutton et al. 1993). The complex was emplaced into the metamorphic rocks of the Neoproterozoic Morar Group and the Archean–Paleoproterozoic Lewisian Gneiss basement of the Eastern Glenelg inlier (Krabbendam et al. 2018) in the hanging wall of the Moine Thrust. The complex is adjacent to the Strathconon Fault, one of many Silurian–Devonian NE–SW sinistral strike-slip structures associated with the GGF (Watson 1984; Law and Johnson 2010; Krabbendam et al. 2018).

With a composition transitional between the alkaline syenites of NW Scotland and the more commonplace metaluminous cale-alkaline granites found in central and eastern areas (Fowler et al. 2008; Bruand et al. 2014), and geochemical characteristics typical of high Ba–Sr granitic magmatism (Fowler and Henney 1996; Fowler and Rollinson 2012; Searle 2022), the origin of the Ratagain Complex has been attributed to slab break-off processes based on regional geological syntheses (Atherton and Ghani 2002; Fowler et al. 2008; Neilson et al. 2009; Fowler and Rollinson 2012; Miles et al. 2016; Archibald and Murphy 2021).

**Petrography**

A range of intrusives, including granite (Fig. 3a, e), quartz monzonite (Fig. 3b, c, d and h), quartz monzodiorite (Fig. 3f, g), monzonite, monzodiorite, granodiorite and syenite occur within the postulated boundaries of the complex. All the above lithologies are characterized by quartz, plagioclase and alkali feldspars as the main felsic minerals and biotite, amphibole and relic clinopyroxene as the main mafic minerals.

The complex also varies in texture from fine-grained and equigranular, to coarse-grained and porphyritic. Coeval mafic microgranular enclaves (MMEs) occur in all intrusive units (Fig. 3d, f) ranging from a few centimetres up to 5 m in diameter; such features are suggestive of magma mingling
and possible mixing processes involving juvenile mafic to ultramafic material (Fowler et al. 2008).

Microstructures

In all the intrusive units of the Ratagain Complex, magmatic microtextures (Vernon 2004) can be observed in thin section. The magmatic microtextures are typified by compositionally zoned euhedral to subhedral plagioclase phenocrysts (Fig. 4a), euhedral to subhedral biotite (Fig. 4e), subhedral to anhedral K-feldspar and anhedral quartz, which exhibits no or rare undulatory extinction (Fig. 4i). With the exception of one locality (57° 13′ 22.86″ N, 5° 29′ 12.91″ W, Fig. 4e; see also Fig. 3c), the biotite and quartz grains are randomly oriented and do not define a magmatic flow foliation.

In samples from the northeastern edge of the complex (Fig. 2, see also Fig. 3g, h), coexisting magmatic microtextures and low-temperature ($T < 400^\circ$C) solid-state microstructures are evident. The low-temperature microstructures comprise highly altered plagioclase, with phenocrysts displaying partial to complete sericitization (Fig. 4d), broken or folded grains of biotite (Fig. 4h), and quartz showing bulging recrystallization or sutured grain boundaries (Fig. 4k).

By contrast, shear-related medium- ($400^\circ$C $< T < 500^\circ$C) and high-temperature ($T > 650^\circ$C) solid-state microstructures are weakly developed and only discernible in a few thin sections. The medium-temperature microstructures feature occasional myrmekite (Fig. 4c), kinked plagioclase twins, cleavage traces in biotite (Fig. 4g) and quartz subgrains (Fig. 4i). The high-temperature deformation microstructures display deformation twinning in plagioclase (Fig. 4b), chessboard extinction in quartz (Fig. 4j) and bent biotite.

Internal structures

Hutton and McErlean (1991) mapped visible, steeply dipping foliations consisting of aligned plagioclase feldspar throughout the Ratagain Complex. The authors interpreted...
these as pre-full crystallization (PFC) fabrics and attributed them to mid-Silurian deformation (Hutton and McErlean 1991; Hutton et al. 1993; McErlean 1993) and they interpreted the mapped fabric pattern to show two NW–SE-trending sinistral shear zones. The authors attributed these PFC fabrics to mid-Silurian deformation. However, the mesoscopic fabrics described by these workers are typically weak (Fig. 3b) and difficult to identify in the field.

Small-scale cataclasite-breccia zones (Fig. 3g) within the intrusion have been taken as evidence of Early Devonian post-emplacement, low-temperature, solid-state deformation (Hutton and McErlean 1991). These minor shear zones are restricted to the NE of the intrusion, in close proximity to Loch Duich, and are cut by numerous quartz–fluorite–calcite veins (Fig. 3h).

Contact relationships

The shape of the Ratagain Complex has not been discussed in detail, but McErlean (1993) depicted a steep-sided body with a gently dipping roof. McErlean (1993) recorded gently SE-dipping contacts with host rocks (albeit irregular) in the NW. These appear to be approximately parallel to the presumed roof of the intrusion close to the Strathconan Fault and suggest that the intrusion may exhibit an overall gently dipping, roughly tabular geometry, with a floor cropping out in the NW and a roof in the SE.

Furthermore, with the exception of sharp contacts in the SE (Fig. 3e) the boundaries observed between the Lewisian country rocks and the Ratagain Complex have a highly irregular and diffuse or gradational nature (Fig. 3a), with no obvious deflection or shouldering of the country rocks and no thermal aureole.

Geochronology

A range of ages are reported for the Ratagain Complex. A Devonian U–Pb zircon age of c. 365 Ma was yielded by Pidgeon and Aftalion (1978) in a monzonite, although the authors considered this to be anomalously young for a Late Caledonian intrusion, invoking Pb loss from the zircons as an explanation. Turnell (1985) yielded biotite–whole-rock and biotite–feldspar–whole-rock ages ranging between 408 and 416.8 Ma, yielding a mean age of 415 ± 5 Ma for the complex, which Thirlwall (1988) readjusted to 419 ± 3 Ma. U–Pb baddeleyite dating undertaken by Rogers and Dunning (1991) suggests that crystallization of the Ratagain Complex occurred in the Silurian at 425 ± 3 Ma.

All in all, the Ratagain Complex provides the opportunity to investigate magnetic mineral fabrics across a visually isotropic intrusion of small areal extent (17 km²) but remarkably varied composition (Stephenson et al. 1999) where rock magnetic techniques have not been applied previously.

Methods

Overview

In total, 100 oriented block samples were obtained from exposures across the Ratagain Complex, including one
syenite sample, six samples of granite, three granodiorite samples, seven samples of monzodiorite, 12 monzonite samples, 38 samples of the quartz monzodiorite, 32 quartz monzonite samples, and one quartz-rich granitoid sample. Oriented block samples were cored (21.5 mm diameter) in the laboratory. The number of cores per block sample varied from 3 to 10, being conditional on the state of weathering; in one sample (AL025) the internal jointing and disintegration due to weathering was such that no cores could be obtained. The cores were then cut into 22 mm-long sub-specimens; in total between 5 and 18 sub-specimens were yielded per block sample.

Rock powders were obtained by cutting a rectangular fragment c. 100 × 50 mm in dimensions from each block.
sample and removing any visibly weathered surfaces. Coarse crushing of the fragments was achieved using a jaw crushe, whilst further crushing to yield fine powders (200 mesh) was undertaken on a TEMA T750k laboratory disc mill.

The AMS of the subsamples was measured at the University of Birmingham Palaeomagnetic Unit and Magnetic Anisotropy (PUMA) laboratory using a KLY4S Kappabridge operating at 875 Hz. Measurements were made at room temperature at 300 Am$^{-1}$. Statistical analysis of the data was carried out using ANISOFT 4.2 and 5 software (Chadima and Jelinek 2009), whilst the programme ‘Stereonet 11.3.0 64-bit version’ (Allmendinger et al. 2012; Cardozo and Allmendinger 2013) facilitated statistical interpretation. Interpolations of the three main parameters (namely $K_m$, $P_j$ and $T$) were made using QGIS, an open-source geographic information system (GIS) application that supports the visualization and analysis of geospatial data (QGIS.org 2021).

To test the efficacy of our AMS analyses of the Ratagain Complex, gain further insights into the magnetic mineralogy and verify whether the entire intrusive body was affected by regional strain during its emplacement, we employed magnetic approaches that are widely used in sedimentary palaeomagnetic studies (e.g. Martín-Hernández et al. 2004) but less commonplace in investigations of granitic intrusions (e.g. Raposo and Gastal 2009; Silva and Raposo 2020). Complementary thermomagnetic, rock magnetic (pARM and IRM), and magnetic fabric analyses (AARM) were undertaken on 16 representative sites. In addition, based on a comparison between AMS and AARM results, we selected three samples, representative of domains with distinct AMS fabrics, for further study of the magnetic carriers through SEM, optical microscopy, IRM experiments and pARM acquisition.

Fig. 4. Microstructural features of Pl, Bt and Qz in the Ratagain Complex: (a) twinned magmatic Pl phenocryst; (b) deformation twinning in Pl; (c) lobes of myrmekite advancing into Pl; (d) Pl phenocryst partially replaced by sericite with remnant growth twinning; (e) magmatic flow fabric with aligned Bt grains; (f) high-temperature solid-state fabric with Bt parallel to thin bands of recrystallized (mylonite) Qz; (g) low-temperature fabric with kinked cleavage in Bt; (h) kinked and folded cleavage in Bt; (i) magmatic Qz with slight undulatory extinction; (j) chessboard extinction and dynamic recrystallization of Qz aggregates; (k) Qz subgrains bordering larger Qz grains; (l) sutured grain boundaries in Qz. All photomicrographs taken in cross-polarized light; all scale bars measure 200 μm. Mineral abbreviations after Whitney and Evans (2010).
Anisotropy of magnetic susceptibility (AMS)

All materials, rocks included, produce a magnetization at temperatures above absolute zero (0 K) when in an applied magnetic field (Tarling and Hrouda 1993). How magnetizable a material is, or its magnetic susceptibility, relates to the strength of the induced magnetization and the strength of the applied field, as follows:

\[ M = KH \]

where \( M \) is the induced magnetization or, in more precise terms, the magnetic dipole moment per unit volume (in A m\(^{-1}\)); \( H \) is the magnetic field (in A m\(^{-1}\)); and \( K \) is the (volumetric) susceptibility (dimensionless).

The AMS records the bulk magnetic fabric of the rock that results from the preferred orientation and magnetic behaviour of its constituent minerals (Rochette et al. 1992; Hrouda 2007). The different forms of magnetic behaviour exhibited by rock-forming minerals can be broadly classified as diamagnetic, paramagnetic, ferromagnetic, ferrimagnetic and antiferrimagnetic (Tarling and Hrouda 1993). In most cases, the rock magnetic fabric is dominated by the contribution from the distribution and shape-preferred orientation of Fe-bearing minerals (chiefly magnetite and biotite) and can be used as a proxy for the rock fabric (e.g. Tarling and Hrouda 1993; Bouchez 1997; Dunlop and Özdemir 1997).

AMS is measured by applying a weak magnetic field to symmetrical drill-core samples, which have been drilled in the laboratory from oriented blocks or directly oriented in the field (Tarling and Hrouda 1993). If the magnetization induced in a core sample is constant in strength, regardless of the orientation of the sample within the applied field, then it is said to be magnetically isotropic. However, the majority of rock samples are magnetically anisotropic; the strength of the induced magnetization varies depending on the direction in which the field is applied (Tarling and Hrouda 1993; Bouchez 1997).

AMS is graphically represented by an ellipsoid of magnetic susceptibility (Rochette et al. 1992), which has the three principal axes represented by the maximum \((K_1)\), intermediate \((K_2)\), and minimum \((K_3)\) susceptibility axis, where \(K_1 > K_2 > K_3\).

A standardized way of presenting AMS data utilizes the following parameters of the ellipsoid:

\[ K_m = (K_1 + K_2 + K_3)/3 \]
\[ L = K_1/K_2 \]
\[ F = K_2/K_3 \]

The AMS ellipsoid is determined by the corrected degree of anisotropy \((P_j)\), and the shape factor \((T)\), which are given in the following equations (Jelinek 1981):

\[ P_j = \exp \left[ \frac{1}{2} \left( \frac{1}{\eta_1 - \eta_m} + \frac{1}{\eta_2 - \eta_m} + \frac{1}{\eta_3 - \eta_m} \right) \right] \]

\[ T = \left[ \frac{1}{2} \frac{\eta_2 - \eta_1 - \eta_3}{(\eta_1 - \eta_3)} - 1 \right] \]

where \(\eta_1\) is \(\ln K_1\), \(\eta_2\) is \(\ln K_2\), \(\eta_3\) is \(\ln K_3\) and \(\eta_m\) is \((\eta_1 + \eta_2 + \eta_3)/3\). \(T\) varies between \(-1\) and 1, where a value greater than 0 tends toward oblateness and a value below 0 is in the prolateness field. \(T\) of 1 is oblate and \(T\) of \(-1\) is prolate, \(T\) of 0 is triaxial.

AMS results

The complete results from the AMS study are presented in Supplementary Table S1 and are stereographically projected in Figure 5. Summary stereograms showing poles to the AMS foliation throughout the complex is also shown in Figure 6. Susceptibility values have an average of 19.22 × 10\(^{-3}\) SI. The average \(P_j\) value is 1.17, reflecting an anisotropy of around 17% (Fig. 7). The shape of the AMS ellipsoid is extremely skewed towards oblateness, with a small minority of samples showing negative \(T\) values (Fig. 7). There is no systematic spatial variation of these parameters across the Ratagain Complex (Fig. 8).

AARM

The interpretation of AMS as a direct proxy of the rock fabric is not straightforward in granites with composite fabrics wherein several magnetic minerals are present (e.g. Bouchez 1997; Stevenson 2004). Analysis of such mixed magnetic fabrics necessitates separating the respective subfabrics. A range of methods, which essentially involving measuring samples at varying temperatures or field strengths, have been devised to isolate the different magnetic components and quantify their contribution to the bulk susceptibility (see Owens and Bamford 1976; Rochette et al. 1992; Richter and van der Pluijm 1994).

AARM analysis involves measuring the anisotropy of the ARM imparted over a specific pattern of directions in a rock sample. Unlike AMS, which represents the susceptibility contributions of all minerals in the rock, including those exhibiting ferromagnetic, diamagnetic and paramagnetic behaviour (e.g. Borradaile and Gauthier 2001), AARM is provided solely by remanence-bearing minerals (e.g. Potter 2004; Silva and Raposo 2020) and therefore distinguishes the contributions of ferromagnetic minerals.

ARM is acquired by simultaneously exposing the sample to a decaying alternating field (AF) and a weak direct current (DC field) (e.g. Borradaile and Jackson 2010). As with AMS, the AARM provides an ellipsoid with the principal axes \(AARM_{max} > AARM_{int} > AARM_{min}\) in which \(AARM_{max}\) is parallel to the magnetic lineation and \(AARM_{min}\) is the pole to the magnetic foliation (e.g. Jackson 1991).

AARM measurements were carried out with an AGICO JR5A spinner magnetometer at the University of Birmingham. ARM was imparted over six directions parallel to \(X, Y\) and \(Z\) (and opposite) axes of the specimen using an ASC D2000 magnetizer, and adopting an AF of 100 mT and a DC field of 0.05 mT. Between each ARM acquisition the sample was fully demagnetized with a tumbling system at 100 mT AF using an AGICO LDA5 demagnetizer. Results from the AARM study are presented in Table 1, together with representative AMS results for comparison.

AARM results

Almost all AMS and AARM tensors are coaxial (Fig. 9), as the \(K_1, K_2\), and \(K_3\) susceptibility axes are sub-parallel to the \(AARM_{max}, AARM_{int},\) and \(AARM_{min}\) axes, respectively. A discrepancy between the AMS and AARM tensors is,
Fig. 5. Map of the Ratagain Complex showing the sampling locations and the AMS stereograms at each site. Stereograms are lower-hemisphere equal-area projections showing the magnetic lineation (red squares), foliation (great circles) and the principal axis normal to foliation (blue circles) for each location. Grid locations provided in Supplementary data.
however, observed at sites AL002 and AL009, where the AARM$_{\text{max}}$ axis is parallel to the $K_1$ and $K_2$ axes, respectively. At site AL002, a susceptibility carried by SD to pseudo-single-domain (PSD) magnetite grains could account for the observed inverse fabric (Rochette 1988). The non-coaxiality between the AMS and AARM tensor observed at site AL009 (Fig. 9) may be explained by the low degree of anisotropy (see Table 1) characterizing these sites, which may have affected the calculation of the AMS and AARM tensors. Another possible explanation for the abnormal fabric at site AL009 is that there are coexisting normal and inverse anisotropies, due to the presence of MD and SD grains, hence producing a possible ‘intermediate fabric’ (Rochette et al. 1992).

We can therefore conclude that the other 14 sites showing coaxiality between the AMS and AARM tensors yield a normal magnetic fabric. Furthermore, the coaxiality of the AMS and AARM tensors suggests that the ferromagnetic and paramagnetic matrix-forming grains may have the same preferred orientation. It is therefore likely that the ferromagnetic mineral assemblages and matrix-forming minerals were aligned under the same magmatic or tectonic episode, implying that the ferromagnetic minerals are primary and magmatic in origin as opposed to secondary and hydrothermal.

**Thermomagnetic curves**

The variation of magnetic susceptibility with temperature can provide insights into the magnetic mineralogy of rocks (Orlický 1990). High-temperature thermomagnetic analyses were performed in an Argon-controlled atmosphere using an AGICO KLY-4S Kappabridge coupled with a CS-3 apparatus. Powdered samples from the 16 representative sites were subjected to a heating–cooling cycle from room temperature to 700°C. Curie temperatures were calculated using the graphical method of Petrovský and Kapička (2006).

**Thermomagnetic results**

All sites record a significant fall in magnetic susceptibility at temperatures between 580 and 650°C (Fig. 10), which corresponds to the Curie temperature ($T_c$) of both magnetite and maghemite (Dunlop and Özdemir 1997). These $T_c$ values preclude the presence of titanomagnetite, iron sulfides and hematite as potential contributors to the AMS and AARM fabrics (Orlický 1990; Dunlop and Özdemir 1997). Most of the samples show similar non-reversible heating curves that exhibit a broad peak at temperatures above 300°C, which can be taken as further evidence of maghemite. Sites AL067 and AL081 show a $T_c$ of 580 and 590°C,
Table 1. Anisotropy of magnetic susceptibility (AMS) and anisotropy of anhysteretic remanent magnetization (AARM) data from representative sites from the Ratagain Complex

| Site | N  | \(K_m\) (10\(^{-3}\) SI) | \(D\) | \(I\) | \(L\) | \(F\) | \(P_j\) | \(T\) | \(R_m\) (10\(^{-3}\) A m\(^{-1}\)) | \(D\) | \(I\) | \(L\) | \(F\) | \(P_j\) | \(T\) |
|------|----|--------------------------|------|------|------|------|------|------|--------------------------|------|------|------|------|------|------|
| AL002| 6  | 21.2                     | 292.8| 16.6 | 1.078| 1.147| 1.240| 0.293| 3.574                    | 203.7| 4.2  | 1.323| 1.505| 1.999| 0.189|
| AL009| 6  | 6.1                      | 241.9| 66.2 | 1.010| 1.041| 1.055| 0.608| 2.869                    | 142.3| 33.7 | 1.078| 1.064| 1.150| -0.151|
| AL011| 11 | 22.1                     | 179.9| 12.2 | 1.011| 1.170| 1.206| 0.875| 5.748                    | 174.5| 13.7 | 1.137| 1.283| 1.469| 0.327|
| AL017| 10 | 55.1                     | 073.9| 37.6 | 1.003| 1.009| 1.013| 0.502| 13.46                    | 109.6| 39.7 | 1.050| 1.106| 1.169| 0.320|
| AL031| 10 | 26.5                     | 001.8| 26.1 | 1.026| 1.114| 1.153| 0.612| 3.284                    | 355.0| 16.8 | 1.105| 1.161| 1.294| 0.162|
| AL038| 9  | 14.2                     | 177.6| 5.1  | 1.027| 1.126| 1.167| 0.631| 2.919                    | 174.9| 7.5  | 1.129| 1.214| 1.378| 0.230|
| AL045| 8  | 0.4                      | 208.3| 37.3 | 1.017| 1.030| 1.048| 0.272| 377.7                    | 197.2| 52.6 | 1.044| 1.081| 1.132| 0.222|
| AL048| 5  | 20.6                     | 174.5| 11.1 | 1.039| 1.222| 1.293| 0.680| 4.605                    | 161.1| 7.6  | 1.218| 1.366| 1.671| 0.227|
| AL054| 7  | 2.4                      | 225.3| 51.2 | 1.017| 1.056| 1.078| 0.534| 1.129                    | 222.4| 31.6 | 1.010| 1.04  | 1.113| -0.112|
| AL062| 9  | 17.3                     | 037.9| 43.4 | 1.061| 1.171| 1.252| 0.452| 3.332                    | 026.6| 44.7 | 1.170| 1.520| 1.821| 0.467|
| AL067| 9  | 2.8                      | 035.0| 60.7 | 1.055| 1.238| 1.343| 0.599| 1.409                    | 109.8| 23.9 | 1.120| 1.088| 1.226| -0.098|
| AL079| 6  | 19.4                     | 022.9| 42.3 | 1.053| 1.202| 1.282| 0.563| 5.267                    | 026.8| 27.9 | 1.144| 1.404| 1.631| 0.438|
| AL081| 5  | 12.8                     | 334.9| 45.5 | 1.045| 1.283| 1.373| 0.698| 2.502                    | 334.5| 41.2 | 1.250| 1.564| 1.982| 0.350|
| AL094| 7  | 56.8                     | 104.4| 27.6 | 1.068| 1.222| 1.321| 0.505| 9.256                    | 116.2| 23.3 | 1.160| 1.656| 1.989| 0.584|
| AL096| 6  | 34.8                     | 068.0| 81.4 | 1.052| 1.278| 1.373| 0.657| 6.098                    | 070.1| 68.0 | 1.216| 1.467| 1.814| 0.352|
| AL100| 6  | 7.8                      | 058.3| 66.9 | 1.062| 1.080| 1.147| 0.122| 1.799                    | 056.1| 64.4 | 1.148| 1.181| 1.368| 0.087|

\(N\) is the number of samples taken per site; \(K_m\) is the mean magnetic susceptibility (in SI units); \(R_m\) refers to the mean magnitude of AARM (measured in A m\(^{-1}\)); AARM\(_{max}\) is the mean maximum remanent magnetization axis; \(D\) is declination; \(I\) is inclination; \(L\) is the lineation parameter; \(F\) is the foliation parameter; \(P_j\) is the degree of anisotropy; and \(T\) is the shape parameter (Jelinek 1981).
respectively, indicating the occurrence of pure, or slightly oxidized magnetite (Dunlop and Özdemir 1997). The remaining 14 sites show $T_c$ ranging between 605 and 650°C, which indicates the occurrence of partially to heavily oxidized (i.e. maghemized) magnetite.

The lack of significant Hopkinson peaks (Hopkinson 1889) in most thermomagnetic curves suggests that MD magnetic particles dominate over SD particles, which are thought to be responsible for this peak (Dunlop 1974). The predominantly gentle or negligible rise in susceptibility with temperature is consistent with relatively large MD magnetite grains being the dominant carrier (e.g. Liss et al. 2004; Burton-Johnson et al. 2019). The form of the Hopkinson peak is controlled by the magnetic blocking temperature and the Curie temperature where the increased susceptibility with temperature is enhanced between these two temperatures. The presence of a relatively narrow Hopkinson peak in the heating curves of AL038 and AL054 (Fig. 10) indicates a greater contribution from PSD or SD magnetite grains. In addition, the cooling curves of these samples display a marked increase in susceptibility in the vicinity of 600°C and much higher susceptibilities than the heating curve, suggesting the likely creation of secondary magnetite at high temperatures (Hrouda 1994).

In samples AL002 and AL011, the cooling curve exhibits a narrow Hopkinson peak, suggesting that some secondary SD magnetite has been generated during heating. The absence of a clear Hopkinson peak at site AL002 is surprising, as its inverse AMS tensor would indicate the occurrence of SD magnetite.

**Partial anhysteretic remanent magnetization (pARM) acquisition**

As natural rock samples can be dominated by a single magnetic phase, such as magnetite, with a wide range of grain sizes (Liu et al. 2005), estimating the relative distributions of SD, PSD and MD grains will in turn help to characterize the contributions of these different fractions to the bulk magnetic fabric. Partial ARM (pARM) measurements can be used to examine the distribution of grain size fractions by virtue of their coercivity – the resistance of individual magnetic grains to an opposing magnetic field (Butler 1992, p. 35) – as coercivity is strongly related to grain size (Jackson et al. 1988).

A stepwise pARM was imparted on three specimens from representative sites (AL002, AL009 and AL048) using a D2000 magnetizer, and the remanence was then measured with a JR5A magnetometer. Similar to the methodologies of Trindade et al. (2001) and Bolle et al. (2003), following AF tumbling demagnetization of the specimens at 100 mT, pARM was imparted at 15 AF steps with windows 5 mT in width from 5 to 50 mT, and 10 mT in width from 50 to 100 mT. A full demagnetization at 100 mT was performed before each pARM step.
**pARM results**

The pARM spectra for all three samples show a well-defined peak at 5 mT (Fig. 11). According to Jackson et al. (1988), a pARM peak at 5 mT corresponds to >25 μm grains, hence MD grains. The pARM curve from site AL009 displays an additional peak between 50 and 80 mT, corresponding to PSD-MD grains in the range of 0.1–0.7 μm (Jackson et al. 1988). For site AL002 this further evidence for MD grains is, once again, at odds with its inverse AMS fabric, which would require the occurrence of SD grains. Conversely, the normal magnetic fabric displayed by site AL048 is consistent with the occurrence of MD magnetite revealed by pARM analysis.

The occurrence of both PSD–SD and MD grains at site AL009 is consistent with our hypothesis of a dual (i.e. normal and inverse) magnetic fabric being responsible for the ‘intermediate’ fabric.

**Optical and scanning electron microscopy**

Optical microscopy and SEM can characterize opaque magnetic minerals and, when the latter is combined with an energy-dispersive X-ray (EDX) detector, further insights on the distribution and concentration of elements within these minerals may be provided. Optical microscopy studies
were carried out at the University of Birmingham using a stereomicroscope Zeiss Axioplan 2 Imaging microscope with a Canon EOS 550D Digital SLR Camera. The EDX analyses and SEM images were obtained using a Hitachi TM3000 instrument coupled with Quantax 70 software at the University of Liverpool, UK.

Photomicrographs show that the magnetite of the Ratagain Complex occurs as highly dispersed accessory crystals (Fig. 12). Although volumetrically variable and differing in grain size, the magnetite is ubiquitous across all rock types constituting the intrusion from quartz monzodiorites (Fig. 12a) to syenites (Fig. 12f).

Backscattered electron (BSE) images (Fig. 13) show three different size fractions of magnetite were present in all samples, which can be broadly categorized as MD, PSD and SD (Day et al. 1977). MD magnetite occurs as large, individual, 20–100 μm subhedral grains (Fig. 13a) and accounts for 75% of the magnetite observed in thin section. Approximately 15% of the magnetite comprises PSD grains of 5–20 μm size (Fig. 13b), which also appear as individual microphenocrysts. The SD fraction, accounting for only 5% or so of all the observed magnetite, consists of elongated grains less than 1 μm wide and equant grains >0.05 μm wide (Fig. 13c), which typically occur as inclusions inside the feldspars.

EDX analyses of selected magnetite grains from each of the grain size fractions (Fig. 13) indicate that the magnetite is nearly pure, with a marked absence of
Fig. 10. Thermal variation of the low-field magnetic susceptibility ($\chi$) from representative samples. Heating curves are red, cooling curves are blue.
Fig. 10. Continued.

Mush ado about the Ratagain Complex
titanium or other elements. This could be accounted for by the exceptional amount of titanite in these rocks (Fig. 12c, e), into which titanium may have partitioned preferentially prior to the crystallization of magnetite. A high oxygen fugacity ($f_{O_2}$) would also favour the crystallization of pure magnetite (Ishihara 2004; dos Santos Dias et al. 2019), but a detailed discussion of the magmatic redox conditions and evolution is beyond the scope of this paper.

Isothermal remanent magnetization (IRM) acquisition

The exposure of a rock containing ferromagnetic minerals to a strong magnetic impulse at room temperature produces an IRM (Dunlop and Özdemir 1997). Stepwise acquisition of an IRM in a rock sample up to its saturation isothermal remanent magnetization (SIRM) and the coercivity unmixing methods that use these curves (Kruiver et al. 2001; Heslop et al. 2002; Maxbauer et al. 2016) have frequently been used to determine the nature and relative concentration of the magnetic carriers. In particular, the unmixing methods help separate the contributions of magnetic minerals with known coercivities (Dunlop 1972; Sant’Ovaia et al. 2015; Maxbauer et al. 2016).

IRM experiments were performed on the same three representative samples from sites AL002, AL009 and AL048 used for the pARM experiments and SEM observations. IRM was imparted with an ASC Scientific IM 10-30 impulse magnetizer by applying 39 increasing peak fields from 10 to 1000 mT. The remanence after each IRM step was measured with a JR5A spinner magnetometer.

IRM results

IRM curves show that all representative samples yield steep acquisition between 10 and 150–200 mT (Fig. 14a). For
samples AL002 and AL048, the acquisition curves flatten thereafter, indicating complete saturation and the dominance of low-coercivity (or soft) magnetic fractions. Sample AL009, however, does not reach saturation at the maximum applied field of 1 T, which indicates the occurrence of high-coercivity (or hard) fractions. To characterize the nature and relative contribution of the various magnetic fractions in these samples we unmixed the coercivities using the Max UnMix online software (Maxbauer et al. 2016).

Our modelling (Fig. 14b–d; Table 2) indicates that a soft coercivity fraction (component 1) is present in all samples,
with variable coercivity (Bh) between 2.05 and 39.66 mT (0.31–1.60 mT in log10 units) and dispersion parameters (DP) between 0.51 and 1.19. An intermediate coercivity fraction is present in samples AL002 and AL009 in variable proportions lower than 20%, and is characterized by higher Bh values of 107.89 and 2922.10 mT (2.03 and 3.47 mT in log10 units) and DP of 0.30 and 0.36. A hard fraction is present in site AL009 only, and is characterized by a Bh of 3885.42 mT (3.59 mT in log10 units) and a DP of 0.75.

Considering these coercivity spectra together with the SEM observations and the thermomagnetic analyses, we infer that the soft fractions correspond to pure magnetite of variable grain size, with site AL048 showing the coarsest magnetite (indicated by its exceptionally low coercivity). The intermediate fraction is likely represented by partially maghematized magnetite and the hard fractions may be represented by magnetite with a higher degree of maghemitization.

**Discussion**

**Interpretation of the magnetic fabric**

Overall, our rock magnetic analyses and microscopy indicate that the intrusive units of the Ratagain Complex contain ferromagnetic minerals of variable grain size from SD to MD, but mainly MD. Mean magnetic susceptibilities of around $20 \times 10^{-3}$ SI confirm that magnetite is the dominant ferromagnetic mineral (Ishihara 1977). Thermomagnetic analyses (Fig. 9) further reveal that the magnetite is Ti-poor and that, locally, the magnetite grains have been oxidized, and partially maghemitized, most probably due to weathering. However, our optical microscope and SEM observations of well-defined, subhedral to euhedral magnetite grains indicate that the magnetite of the Ratagain Complex is mostly primary and magmatic in origin; any hydrothermal alteration following intrusion and emplacement of the granite is not pervasive enough to have compromised the original magnetic mineralogy.

---

**Fig. 14.** (a) Isothermal remanent magnetization (IRM) acquisition curves for representative samples. The red, blue and green lines represent AL002, AL009 and AL048, respectively. (b–d) Coercivity unmixing analysis of IRM acquisition curves for the representative samples. Grey circles are the IRM data and the grey shaded area represents error envelopes of 95% confidence intervals. The contribution from all magnetic components is represented by the gold line and shaded area. The blue, purple and green lines and corresponding shaded areas represent the first, second and third component, respectively.
Out of the 16 representative sites selected to test the magnetic fabric using high field measurements, only one (AL002) shows an inverse fabric, in which the $K_1$ and $K_3$ axes are swapped. This is typically caused by the occurrence of SD magnetic grains (Rochette 1988); however, both pARM and IRM experiments on this sample exclude the presence of a significant proportion of SD grains, suggesting that the inverse fabric in this case may be yielded by PSD grains. The sole intermediate AMS fabric (site AL009) can be interpreted either in terms of a low degree of anisotropy or an inverse fabric carried by SD magnetite grains partly camouflaging the normal fabric yielded by the MD magnetite (Borradaile and Gauthier 2001). Such samples are rare in the entire dataset. Therefore, the bulk magnetic fabric of the Ratagain Complex revealed by our AMS measurements can thus be taken as representative of the shape preferred alignment of magnetite grains and likely parallel to the the main silicate petrofabric i.e. the foliation and lineation defined by mineral shape alignment (e.g. Bouchez 1997; O’Driscoll et al. 2007; Archanjo et al. 2012).

### A possible emplacement mechanism for the Ratagain Complex

The AMS ellipsoids determined in this study are predominantly oblate and the fabric orientations are consistent with a steeply dipping foliation (where $K_3$ is gently plunging) (Fig. 5), indicating that there was a flattening ambient strain field as this magma crystallized. Plotting all $K_1$ axes confirms the preponderance of a steeply subvertical orientation of the magnetic lineation and thus vertical stretching (Fig. 6a). Plotting all $K_3$ shows a general NE–SW strike (Fig. 15b), although with a bimodal north–south and NE–SW spread owing to the sigmoidal trace of the fabric. Following the reasoning of Burton-Johnson et al. (2019) and Di Chiara et al. (2020), the ambient tectonic strain field operating during the emplacement of the Ratagain Complex is NW–SE shortening. Modelling by Tikoff and Greene (1997) showed that transpression may produce vertical stretching lineations which may be relevant in terms of Caledonian tectonic events involving transpression.

Because the $T$ values are predominantly positive, we place more confidence in the magnetic foliation. From Figure 16 we can see that the magnetic fabric is consistent with the weak visible fabric described by McErlean (1993); the magnetic lineation, although much weaker, is mainly subvertical (Fig. 16).

Combining the magnetic and visible fabrics reveals a markedly sigmoidal pattern (Fig. 15a). The sigmoidal combined fabric trace may represent two opposing magmatic lobes (Fig. 15b) (e.g. Stevenson et al. 2008). However, the fabric cross-cuts internal boundaries inferred by previous workers (Nicholls 1951; Hutton et al. 1993). This would suggest that the fabric was imposed at the latter stages or at the end of emplacement and it follows that the mainly oblate shape of the AMS records a flattening ambient strain field. We find it more likely that the sigmoidal pattern records a sinistral magmatic shear zone (Fig. 15c) much like the interpretation of McErlean (1993), who proposed that sinistral slip on the Strathconan Fault and a parallel shear zone to the NW linked by faults acted to permit space for...
magma intrusion during sinistral transpression and thus controlled the emplacement of the Ratagain Complex.

**Post-emplacement fabric development**

With the exception of one field locality (57° 13′ 23.41″ N, 5° 24′ 49.63″ W; see also Fig. 3g), the rocks of the Ratagain Complex do not evidence solid-state deformation in thin-section, hand-specimen or field exposures and also lack visible fabrics. The magmatic texture preserved in the majority of studied thin sections (see Fig. 4) and the undeformed appearance of the MMEs (Fig. 3f) provide strong evidence against deformation in the magmatic-flow stage as well as extensive solid-state grain-shape modification after full crystallization of the intrusion. This is further supported by the occurrence of flow structures in some outcrops, which are decoupled from tectonic structures in the metamorphic country rocks (Fig. 3c).

The field observations outlined above suggest that the timing of fabric acquisition for the Ratagain Complex was restricted to a brief window post or at the end of its emplacement, when the granite was near the magmatic solidus but still contained substantial interstitial melt to accommodate most of the tectonic stress and shield crystallizing grains from substantial deformation (e.g. Liu et al. 2018). This is further corroborated by our microstructural observations, which clearly evidence that whilst the petrofabric of the Ratagain Complex records a deformational overprint, critically, this overprint has not developed into a solid-state tectonic fabric.

---

**Fig. 15.** (a) Interpolated fabric trace for the Ratagain Complex. The base map is simplified from Figure 2 and fabric traces are based on magnetic fabrics from this study, combined with weak visible fabrics described by McErlean (1993). (b) Interpretation of the fabric pattern in (a) considering the emplacement of opposing lobes of magma similar to Stevenson et al. (2008). (c) Interpretation of the fabric pattern in (a) considering the strain recorded by a sinistral shear zone under a generally horizontal shortening or flattening strain field.

**Fig. 16.** (a) Map showing the direction and plunge of the magnetic lineation ($K_1$). Light green arrows are a $K_1$ lineation plunging 0–45°; dark green arrows are a $K_1$ lineation plunging 46–89°; and green crosses are a vertical $K_1$ lineation. (b) Map showing the strike of the visible foliation, magnetic foliation and our overall interpretation of the fabric pattern. Light green lines are visible foliations mapped by McErlean (1993); light blue lines are normal magnetic foliations; red lines are intermediate or inverse magnetic lineations; dark blue continuous lines are interpolations between the foliation data points.
Timing of emplacement

The petrofabric of the Ratagain Complex carries an ambient tectonic overprint, which is most likely to have developed post or at the end of emplacement. Our data provide insights on the regional kinematics and timing of deformation relative to granite emplacement as the significant horizontal shortening component of the overprint can be ascribed to NW–SE shortening and is consistent with transpression.

The Late Caledonian suite of Britain and Ireland was emplaced in a complicated and varied tectonic setting that included alternating transtensional and transpressional strike-slip deformations (Jacques and Reavy 1994; Miles et al. 2016). The major deformation phases that could potentially have produced the observed overprint are Iapetan (Scandian) transpression between c. 425 and 420 Ma, and Acadian transpression from 404 to 394 Ma (Miles et al. 2016). Whilst the favoured age of crystallization (425 ± 3 Ma, Rogers and Dunning 1991) and interpretation of the Ratagain Complex as a late Silurian intrusion (Hutton and McErlean 2016). The major deformation phases that could potentially have produced the observed overprint are Iapetan (Scandian) transpression between c. 425 and 420 Ma, and Acadian transpression from 404 to 394 Ma (Miles et al. 2016). Whilst the favoured age of crystallization (425 ± 3 Ma, Rogers and Dunning 1991) and interpretation of the Ratagain Complex as a late Silurian intrusion (Hutton and McErlean 1993; Hutton et al. 1993; McErlean 1993) may seem consistent with emplacement during the Iapetus transpression, this hypothesis presents several significant problems.

1. Both previous (Fowler et al. 2008; Fowler and Rollinson 2012) and contemporary geochemical studies, the later of which will be presented in a companion paper, evidence that the Ratagain Complex bears the characteristic elemental and isotopic signatures of post-collisional, slab break-off magmatism (see Moyen et al. 2017; Hildebrand et al. 2018), precluding intrusion and emplacement during active Iapetan subduction.

2. As Iapetus subduction was followed by orogen-wide transtension and thereafter a return to transpression in the Acadian, it follows that if the Ratagain Complex was emplaced during the late Silurian, the intrusion would reflect these respective kinematic events in the form of multiple tectonic fabrics (see Benn 2010). Even if Iapetus transpression is backdated to c. 430 Ma, as postulated by some authors (Archibald et al. 2022), the Ratagain Complex would nonetheless record multiple tectonic fabrics, as opposed to a single, moderate tectonic overprint, if it was emplaced and crystallized at c. 425 Ma.

3. The late Silurian to early Devonian in the Scottish Caledonides is understood to have been a period of brittle, upper crustal thrusting and strike-slip faulting along the GGF and related structures (Strachan et al. 2020; Searle 2022). Given the proximity of the Ratagain Complex to the Strathconan fault, an offset of the GGF, one would expect the primary magmatic fabric to be reworked or obliterated by a well-developed, solid-state fabric imparted by the region-wide strike-slip faulting. Our thin-section petrographical analyses and field observations, do not show this and instead demonstrate that in the majority of the Ratagain Complex the primary magmatic-state fabric has been preserved.

For these reasons, we find it difficult to unequivocally place the final emplacement and crystallization of the majority of the Ratagain Complex at c. 425 Ma during Iapetus transpression. We instead suggest two hypotheses to satisfy the field, petrological and rock magnetic evidence.

a. The Ratagain Complex is younger than previously reported and could be a syn-Acadian intrusion. Although zircon age spectra for the Late Caledonian suite compiled by Miles et al. (2016) suggest that volumes of granitic emplacement were suppressed during the Acadian, emplacement within transpressional settings can be facilitated by transpressional sites formed at releasing bends (e.g. Glazner 1991), as has recently been suggested for the Late Caledonian Shan, Skiddaw and Weardale granites, Lake District (Miles and Woodcock 2018; Woodcock et al. 2019). In our specific case, we postulate that a fault splay from the Strathconan Fault, subjacent to the Ratagain Complex emplacement site, may have interacted with the Strathconan Fault to create a local transtensional jog or opening, in an otherwise transpressional setting, to facilitate magma ascent from a deeper-seated mid- or lower-crustal source region. We envisage that the magma was sufficiently crystallized to accommodate ambient strain during the Acadian but, owing to both the presence of interstitial melt and the relatively short duration of the Acadian Event, the development of a significant solid-state deformational fabric was unable to take place before the cessation of the transpression at c. 394 Ma.

b. The Ratagain Complex was emplaced incrementally with crystallization of different magma batches over a protracted period. Rather than dating the final crystallization of the complex, Rogers and Dunning’s (1991), 25 ± 3 Ma U–Pb baddeleyite age for the complex could instead record pre-emplacement magmatic activity in a deeper part of the transcrustal crystal-mush system (see Cashman et al. 2017; Jackson et al. 2018; Sparks et al. 2019). The intrusion of new magma pulses into partially crystallized magma mush could have accommodated most of the ambient tectonic strain (Liu et al. 2018) during the period of Silurian strike-slip faulting and offers a plausible explanation as to why the Ratagain Complex bears no obvious solid-state fabric, has no thermal aureole and is decoupled from tectonic structures in the host Precambrian rock (Fig. 2a, e). Whilst direct evidence for multiple injections of magma, especially internal contact boundaries, may be lacking in the field, this is resolved by a model in which the remelting and/or recrystallization of successive magma pulses resulted in the contacts becoming diffuse (Miller et al. 2011; Liu et al. 2018). On the other hand, indicative features of new injections of magma, such as the MMEs and complex zoning patterns in feldspar phenocrysts identified in this study (Figs 2d, f and 3a), were preserved.

A tentative model of emplacement

Summing up, our field, microstructural and magnetic data would support emplacement of the Ratagain Complex in a
compressive (likely transpressive) system. As the steeply dipping magnetic fabric pattern of the complex cross-cuts any internal boundaries, and there are also no apparent pulses in the fabric (cf. Stevenson et al. 2007), we interpret this fabric pattern as recording an ambient strain field operating post or at the very end of emplacement. The mostly irregular contact boundaries – such as that of the western granite, which has not been discussed in great detail by previous workers – could also be taken as representative of sheeting linked to wedging and sill-like intrusions with broken bridges (Schofield et al. 2012; Eide et al. 2017), similar to those recorded at the eastern contact of the Ross of Mull Granite (Petronis et al. 2012). We therefore tentatively suggest that the juxtaposition of the considerable compositional heterogeneity with diffuse or indistinguishable internal boundaries observed in the Ratagain Complex reflects incremental emplacement of a sheeted intrusive complex (Fig. 17).

Conclusions

(i) Strong agreement with visible fabrics (Hutton and McErlean 1991) and our detailed rock magnetic analyses confirming that MD ferromagnetic minerals are the main contributors to the magnetic fabric, allows us to interpret our AMS data as normal fabrics where the $K_3$ axis represents the pole to the magnetic foliation and the $K_1$ axis the magnetic lineation. The magnetic fabric of the Ratagain Complex as revealed by AMS can thus be considered a valid proxy for the petrofabric.

(ii) Further evidence for the predominant occurrence of a ‘normal’ magnetic fabric in the Ratagain Complex is provided by the coaxiality of AARM and AMS tensors; only two sites out of 16 showed an inverse (AL002) and intermediate (AL009) fabric, where the ARM max axis coincides with the $K_3$ and $K_2$ susceptibility axis, respectively.

(iii) Thermomagnetic, IRM and pARM acquisition experiments, together with optical and scanning electron microscopy indicate that MD magnetite and minor maghemite (or partially maghemitized magnetite) are the main contributors to the magnetic susceptibility, and hence to the magnetic fabric.

(iv) Thin-section petrological investigations evidence predominantly magmatic textures with a tectonic overprint, but critically there is no obvious solid-state fabric. Together with the observation that the magnetite of the Ratagain Complex is magmatic (as indicated by the coaxiality between the AMS and AARM data and our microscopy studies) we propose that the magnetic fabric of the Ratagain Complex was formed post-emplacement, when the granite was near solidus temperature and in a crystal-rich but
theologically magmatic state. Sinistral transpressive strain imparted a tectonic overprint but did not persist long enough to form a solid-state fabric.

(v) Previous emplacement models for the Ratagain Complex do not fully correlate with all the field, petrographic and magnetic observations presented in this study, leading us to propose two hypotheses for its emplacement and relationship to regional kinematics. We suggest that either the Ratagain Complex is younger than previously suggested and records only Acadian transpression, or that it was intruded and emplaced incrementally over a protracted time period, whereby incremental remobilization by successive magma pulses accommodated most of the ambient tectonic strain, enabling the preservation of the primary magmatic fabric whilst simultaneously preventing the full development of a solid-state tectonic fabric.

(vi) Our work demonstrates that using a suite of rock magnetic analyses can provide insights on the detailed emplacement history of a granitic complex, particularly its relationship to regional tectonic strain, and allows us to infer temporal relations between different kinematic events that have taken place.

Acknowledgements

We thank Mike Fowler and Craig Storey for their initial suggestion to investigate the Ratagain Complex and all their advice since. Thanks go to Paul Hands for producing excellent thin sections for petrological advice on specialized stereographic software. AL would also like to thank Jon Butler, R.F. 1992. Slab breakoff: a model for Caledonian, Late Granite syn-collisional magmatism in the orthotectonic (metamorphic) zone of Scotland and Donegal, Ireland. Lithos, 62, 65–85, https://doi.org/10.1016/S0024-4937(02)00111-4

Benn, K. 2010. Anisotropy of magnetic susceptibility fabrics in syntectonic plutons as tectonic strain markers: The example of the Caucho pluton, Megane Terrane, Nova Scotia. Transactions of the Royal Society of Edinburgh: Earth and Environmental Science, 100, 147–158, https://doi.org/10.1017/S1755691000016028

Bolle, O., Diet, H. and Trindade, R.F. 2003. Magnetic fabrics in the Holm granite (Vest-Agder, southernmost Norway): implications for the late evolution of the Sveconorwegian (Grenvillian) orogens of SW Scandinavia. Precambrian Research, 121, 221–249, https://doi.org/10.1016/S0301-9268(03)00113-5

Borradaile, G.J. and Gauthier, D. 2001. AMS-detection of inverse fabrics without AARM, in ophiolite dikes. Solid Earth, 28, 3517–3520, https://doi.org/10.1029/2001GL012976

Borradaile, G.J. and Henry, B. 1997. Tectonic applications of magnetic susceptibility and its anisotropy. Earth-Science Reviews, 42, 49–93, https://doi.org/10.1016/S0012-8252(96)00044-5

Borradaile, G.J. and Jackson, M. 2010. Structural geology, petrofabrics and magnetic fabrics (AMS, AARM, AIRM). Journal of Structural Geology, 32, 1519–1551, https://doi.org/10.1016/j.jsg.2009.09.006

Bouchou, J.L. 1997. Granite is not anisotropic: an introduction to AMS studies of granitic rocks. In: Bouchou, J.L., Hutton, D.H.W. and Stephens, W.E. (eds) Granite: from Segregation of Melt to Emplacement Fabrics. Kluwer Academic, Dordrecht, 95–112

Brown, P.E., Ryan, P.D., Soper, N.J. and Woodcock, N.H. 2008. The Newer Granite problem revisited: a transboundary origin for the Early Devonian Trans-Suture Suite. Geological Magazine, 145, 235–256, https://doi.org/10.1017/S0016756807004219

Brunaud, E., Storey, C. and Fowler, M. 2014. Accessory mineral chemistry of high Ba–Sr granites from Northern Scotland: constraints on petrogenesis and records of whole-rock Signatures. Journal of Petrology, 55, 1619–1651, https://doi.org/10.1093/petrology/egt057

Buttle, A.J., Long, G. and Murchison, C.G., Muraszko, J., Harrison, R.J. and Jordan, T.A. 2019. Tectonic strain recorded by magnetic fabrics (AMS) in plutons, including Mt Kinabalu, Borneo: a tool to explore past tectonic regimes and syn-magmatic deformation. Journal of Structural Geology, 119, 50–60, https://doi.org/10.1016/j.jsg.2018.10.003

Butler, R.F. 1992. Paleomagnetism: Magnetic Domains to Geologic Terranes. Blackwell Scientific, Oxford, UK

Cardozo, N. and Allmendinger, R.W. 2013. Lithospheric coalitions and the Late Caledonian–Iapetan Orogeny: implications for the ‘Newer Granite’ suite of the Scottish and Irish Caledonides. Geological Society of America Special Paper, 554, 1–XXX, https://doi.org/10.1130/2013.2554(15)

Atherton, M.P. and Ghiati, A.A. 2002. Slab breakoff: a model for Caledonian, Late Granite syn-collisional magmatism in the orthotectonic (metamorphic) zone of Scotland and Donegal, Ireland. Lithos, 62, 65–85, https://doi.org/10.1016/S0024-4937(02)00111-4
Jackson, M.D., Blundy, J. and Sparks, R.S.J. 2018. Chemical differentiation, cold storage and remobilization of magma in the Earth’s crust. Nature, 564, 405–409, https://doi.org/10.1038/s41586-018-0746-2
Jackson, M. and Reay, D.J. 2019. Large-scale plutonism and major lineaments in the SW Scottish Highlands. Journal of the Geological Society, London, 151, 955–969, https://doi.org/10.1144/jgs2017-06955
Jelinek, V. 1981. Characterization of the magnetic fabrics of rocks. Tectonophysics, 79, 763–767, https://doi.org/10.1016/0040-1951(81)90170-4
Johnson, T.E., Kirkland, C.L., Viefh, D.R., Fischer, S., Reddy, S.M., Evans, N.J. and Johnson, M.D. 2017. Zircon geochronology reveals polyphase magmatism and crustal anatexis in the Bucan Block, NE Scotland: implications for the Grampian Orogeny—Geoscience Frontiers, 8, 1469–1478, https://doi.org/10.1016/j.gsf.2017.02.002
Jones, R.R., Holdsworth, R.E. and Bailey, W. 1997. Lateral extrusion in transpression zones: the importance of boundary conditions. Journal of Structural Geology, 19, 1201–1217. https://doi.org/10.1016/S0191-8141(97)00034-5
King, R.F. 1966. The magnetic fabric of some Irish granites. Geological Journal, 5, 43–46, https://doi.org/10.1002/gj.3500505106
Kim, P., Strachan, R., Friend, C., Kocks, H., Rogers, G. and Paterson, B. 2003. U-Pb geochronology of deformed megagranites in central Sutherland, Scotland: evidence for widespread late Silurian metamorphism and ductile deformation of the Moine Supergroup during the Caledonian orogeny. Journal of the Geological Society, London, 160, 259–269, https://doi.org/10.1144/0016-7490-1487
Krabbendam, M., Ramsay, J.G., Leslie, A.G., Tanner, P.W.G., Dietrich, D. and Goodenough, K.M. 2018. Caledonian and Knydorartian overprinting of a Grenvillian inlier and the evolution of the greenstone belt of the Precambrian Proto-Moine Nappe, Glenieg, NW Scotland. Scottish Journal of Geology, 54, 13–35, https://doi.org/10.1144/sjg2017-006
Kruiver, P.P., Dekkers, M.J. and Heslop, D. 2001. Quantification of magnetic coercivity components by the analysis of acquisition curves of isothermal remanent magnetisation. Earth and Planetary Science Letters, 189, 269–276, https://doi.org/10.1016/S0012-821X(01)00367-3
Lambert, R.S.J. and McKerrow, W.S. 1976. The Grampian orogeny. Scottish Journal of Geology, 12, 271–292, https://doi.org/10.1016/s0305-0484(01)80207-1
Law, R.D. and Johnson, M.R.W. 2010. Microstructures and crystal fabrics of the Moine Thrust zone and Moine Nappe: history of research and changing tectonic interpretations. Geological Society, London, Special Publications, 355, 443–503, https://doi.org/10.1144/SP135.21
Liss, D., Owens, W.H. and Hutton, H.W. 2004. Palaeomagnetic results from the Whin Scill complex: evidence for a multiple intrusion event and revised virtual geomagnetic poles for the late Carboniferous for the British Isles. Journal of the Geological Society, London, 161, 927–938, https://doi.org/10.1144/0016-7464-156
Liu, H., Martelet, G. et al. 2018. Incremental emplacement of the late Jurassic Midcrustal, Lopolith-Like Qitianlan Pluton, South China, revealed by AMS and Bouguer Gravity Data. Journal of Geophysical Research: Solid Earth, 123, 9249–9268, https://doi.org/10.1029/2018JB015761
Liu, Q., Yu, Y., Pan, Y., Zhuo, R. and Zhao, X. 2005. Partial anhysteretic remanent magnetization (pARM) of synthetic single- and multidomain magnetites and its paleoenvironmental significance. Chinese Science Bulletin, 50, 2381–2384, https://doi.org/10.1007/s11434-003-1813-5
Mark, D., Rice, C., Hole, M. and Condron, D. 2020. Multi-chronometer dating of the Souther Head complex: rapid exhumation terminates the Grampian Event of the Caledonian Orogeny. Journal of Geophysical Research: Solid Earth, 115, 108–110, https://doi.org/10.1029/2010JG001759
Martin-Hernández, F. and Ferri, E.C. 2007. Separation of paramagnetic and ferromagnetic anisotropies: a review. Journal of Geophysical Research: Solid Earth, 112, B03105, https://doi.org/10.1029/2006JB004340
Martin-Hernández, F., Lüneburg, C.M., Aubourg, C. and Jackson, M. 2004. Magnetic fabric: methods and applications—an introduction. Geological Society, London, Special Publications, 238, 1–7, https://doi.org/10.1144/1476-6403
Mile, S.P.L. 2004. 238.01.01
Maxbauer, D.P., Feinberg, J.M. and Fox, D.L. 2016. MAX UnMix: a web application for unmixing magnetic coercivity distributions. Computers & Geosciences, 95, 140–145, https://doi.org/10.1016/j.cageo.2016.07.009
McErlain, M.A. 1993. Granitoid Emplacement and Deformation: A Case Study of the Thorr Pluton, Ireland, With Contrasting Examples from Scotland. PhD thesis, Durham University.
M pendum, J.R. 2012. Late Caledonian (Scandian) and Proto-Variscan (Acadian) orogenic events in Scotland. OUGS Journal, 33, 37–51.
Mendum, J.R. and Noble, S.R. 2010. Mid-Devonian sinistral transpressional movements on the Great Glen Fault: the rise of the Rosemarkie Inlier and the Acadian Event in Scotland. Tectonics, London, Special Publications, 335, 161–187, https://doi.org/10.1144/SP135.5
Miles, A.J. and Woodcock, N.H. 2018. A combined geochronological approach to investigating long lived granite magmatism, The Shap granite, UK. Lithos, 304–307, 245–257, https://doi.org/10.1016/j.lithos.2018.02.012
Miles, A.J., Woodcock, N. and Hawkesworth, C. 2016. Tectonic controls on postsubduction granite genesis and emplacement: the late Caledonian suite of Britain and Ireland. Gondwana Research, 39, 250–260, https://doi.org/10.1016/j.gr.2016.02.006
