Internal fabrics of the Idaho batholith, USA

A. Byerly, B. Tikoff, M. Kahn, B. Jicha, R. Gaschnig, and A.K. Fayon

Department of Geoscience, University of Wisconsin–Madison, Madison, Wisconsin 53706, USA
School of Earth and Atmospheric Science, Georgia Institute of Technology, Atlanta, Georgia 30332, USA
Department of Earth Sciences, University of Minnesota, Minneapolis, Minnesota 55455, USA

ABSTRACT

The Idaho batholith of the North American Cordillera is a large and long-lived silicic intrusive center. We studied fabrics at the regional scale within the different intrusive suites of the Idaho batholith, using microstructural characterization, anisotropy of magnetic susceptibility measurements, and shape preferred orientation analyses. Each studied outcrop was collocated with existing U-Pb zircon geochronology and ongoing (U-Th)/He zircon thermochronology results. Three new 40Ar/39Ar biotite ages, collocated with the existing U-Pb zircon ages, constrain the cooling rates within the batholith. The presence of dominantly magmatic microstructures allows us to interpret the results relative to the U-Pb zircon ages and observe spatial and temporal patterns of fabric development. The early (pre–80 Ma) intrusive suites exhibit solid-state microstructures that show a consistent orientation only in the western part of the batholith. Fabrics in the 83–67 Ma Atlanta peraluminous suite are well developed and consistently oriented (NW-striking; NE-dipping), recording localized contraction during magmatism. These fabrics in the 66–53 Ma Bitterroot peraluminous suite are weak and inconsistent in orientation, despite emplacement during regional contraction. We hypothesize that the lack of consistently oriented fabrics in the Atlanta lobe results from either: (1) topographic effects that caused local extensional/neutral strain environments in the upper parts of a crustal plateau; or (2) emplacement in thin, horizontal magmatic sheets. In contrast, fabrics within the 66–53 Ma Bitterroot peraluminous suite are well developed and consistently oriented (NW-striking; NE-dipping), recording localized contraction during magmatism.

INTRODUCTION

Large silicic intrusive complexes are fundamental building blocks of the continental crust. Understanding these large intrusive centers is essential to our understanding of the growth and recycling of continental crust. Many workers have evaluated the internal structure of individual plutons as constituent parts of larger batholiths to understand batholith construction and deformation through time (Tobisch et al., 1995; Borradaile and Henry, 1997; Benn et al., 1998, 2001; Launee and Cruden, 1998; McNulty et al., 2000; Neves et al., 2003; Mantani and Greiling, 2005; Sen and Mantani, 2006; Zak et al., 2007; Archanjo et al., 2008; DeCelles et al., 2009; Benn, 2010; Mantani et al., 2013; Cao et al., 2015; Paterson and Ducea, 2015). This approach of analyzing plutons as discrete, time-successive crustal additions provides insight into the broader magmatic and tectonic regime. Tobisch et al. (1995), for example, used the spatial and temporal distribution of magmatic versus solid-state fabric in the central Sierra Nevada batholith (Fig. 1) to reconstruct the evolution of this Cordilleran magmatic arc. They evaluated deformation within restricted areas and time frames defined by mapped plutonic boundaries. The fabrics, within the framework of known regional-scale plate trajectories and average strain rates, were then used to elucidate the orientation and magnitude of broad-scale tectonic strain. Similarly, Chardon et al. (1999) examined fabrics within compositionally and structurally distinct plutons in the Coast plutonic complex (Fig. 1). These syntectonic plutons record large-scale transpression and together were used to reconstruct the tectonic evolution of shear zones within the broader batholith. Results have shown that the growth and deformation record of batholiths provides magmatic and tectonic constraints that contribute to our understanding of the growth and recycling of continental crust.

Construction of a batholith can be understood by evaluating the growth history of individual plutons. Most granitic intrusions record some fabric, and quantitative evaluation of the macroscopically indisernible fabrics helps to elucidate the subtle strain recorded during magma crystallization (e.g., Bouchez, 1997). Magmatic fabrics, in particular, preserve orientations and magnitudes that result from the final increment of strain at the time of crystallization (e.g., Marre, 1986; Hutton, 1988; Paterson et al., 1989, 1998; Benn, 2010). The preservation of magmatic microstructures indicates that the magmatic fabrics were “locked in” during a limited time interval immediately prior to crystallization (e.g., Fowler and Paterson, 1997; Yoshinobu et al., 1998). By combining microstructural data with U-Pb zircon ages from individual plutons, the regional tectonic activity through time can be assessed.

The majority of the Idaho batholith lacks the typical internal plutonic contacts of other large silicic batholiths in the North American Cordillera (Fig. 1). Common structures in other magmatic arcs—mesoscale folds, kilometer-scale shear zones, mixing and mingling zones—are largely absent in the Atlanta lobe of the Idaho batholith. Rather, these granites are remarkably uniform in composition and structure, consisting primarily of biotite granodiorite and two-mica granite/granodiorite. The recent designation of magmatic suites within the batholith required a combination of geochronology and geochemistry (Gaschnig et al., 2010, 2011). Due to the apparent compositional homogeneity and cryptic structure, little attention has been paid to the tectonic record preserved within the batholith. Because of these factors, evaluating the first-order growth and deformation history of the Idaho batholith requires a different approach.
individual plutonic boundaries. Fabric orientations was chosen for the Idaho batholith because it is dependent on the priori ability to map individual plutonic boundaries. Fabric orientations and magnitudes were measured using shape preferred orientation (SPO) and anisotropy of magnetic susceptibility (AMS) techniques.

**GEOLOGICAL BACKGROUND**

The Idaho batholith of the North American Cordillera is a large silicic intrusive center, consisting of the Atlanta lobe in the south and the Bitterroot lobe in the north. The Atlanta lobe, the largest part of batholith, exhibits distinctively homogeneous composition and internal structure (Fig. 2). Gaschnig et al. (2010, 2011) constrained the emplacement and crystallization history and tracked the source and isotopic evolution of the crustal melt. The emplacement of the Idaho batholith and the intrusion of the subsequent Challis suite account for ~32,000 km² spatial extent at present, representing a 50 m.y. (98–43 Ma) episode of magmatism.

Prior to Idaho batholith magmatism, a spatially restricted N-S–trending series of plutons forming the suture zone suite of Gaschnig et al. (2010) was intruded in western Idaho, straddling the boundary between cratonic North America and accreted terranes (Manduca et al., 1992, 1993; Giorgis et al., 2008; Gaschnig et al., 2010). The suture zone suite was affected by the western Idaho shear zone, a major transpressional shear zone (e.g., Giorgis et al., 2008).

We utilized the divisions of Gaschnig et al. (2010), who distinguished intrusive suites of the Idaho batholith based on a combination of lithology, geochronology, and geochemistry (Fig. 2). The main phases of the batholith are: (1) the early metaluminous and border zone suites (98–85 Ma); (2) the Atlanta peraluminous suite (83–67 Ma); (3) the late metaluminous suite (75–69 Ma); and (4) the Bitterroot peraluminous suite (66–53 Ma; Gaschnig et al., 2010, 2011). Magmatism continued in Idaho with the Challis intrusive suite (51–43 Ma; Gaschnig et al., 2010, 2011), which is not considered to be part of the Idaho batholith.

The border zone suites consist of a series of N-S–trending tabular bodies located east of the western Idaho shear zone and suture zone plutons. Early metaluminous plutons were emplaced simultaneously with the border zones suite, but they are exposed in the Atlanta lobe of the Idaho batholith as roof pendants and septa within the younger phases and along the eastern boundary of the batholith. The Atlanta peraluminous suite is the largest phase of the batholith, making up the majority of the Atlanta lobe, and it is relatively homogeneous in structure and composition (biotite granodiorite and two-mica granite). Evolved Hf and Nd isotopes and evolved major-element compositions with limited range suggest that magma was derived almost exclusively from preexisting continental crust and received little material input from the mantle (Gaschnig et al., 2011). The Bitterroot lobe consists of a late metaluminous suite (75–69 Ma) and the Bitterroot peraluminous suite (66–53 Ma), and it was initiated during the waning stages of emplacement of the Atlanta lobe.
peraluminous suite. Finally, the Challis suite (51–43 Ma) is a regionally extensive intrusive and extrusive group that intrudes earlier Idaho batholith granites.

ASSESSING FABRIC DEVELOPMENT IN GRANITIC ROCKS: METHODS AND RESULTS

Background

Fabrics in plutons provide important information about the emplacement and postemplacement deformational history of an igneous body; these fabrics are grouped into magmatic fabrics and solid-state fabrics (e.g., Paterson et al., 1989; Benn et al., 2001; de Saint-Blanquat et al., 2001). Magmatic fabrics form via the mechanical alignment of anisotropic phases in the presence of melt (e.g., Marre, 1986; Paterson et al., 1989, 1998). Fabric development may result from internal (e.g., flow, mixing, thermal convection) or external (regional strain) processes. Previous workers have interpreted magmatic fabrics in plutons as a record of either magmatic processes at the emplacement level (Cruden and Launeau, 1994; Tobisch and Cruden, 1995; Archanjo et al., 1999; Ferré et al., 1999; de Saint-Blanquat et al., 2001, 2006; Parada et al., 2005; Stevenson et al., 2007; Naibert et al., 2010; Archanjo and Campanha, 2012; Gutiérrez et al., 2013; Morgan et al., 2013) or regional strain during syntectonic emplacement (Brun and Pons, 1981; Brun et al., 1990; Archanjo et al., 1995; Leblanc et al., 1996; Benn et al., 1997, 1999, 2001; Sen et al., 2005; Mamtani et al., 2013). Solid-state (or subsolidus) fabrics in granitic intrusions are formed by crystal-plastic deformation of phases in the absence of melt. Solid-state fabrics in granites are generally used to track distributed regional-scale strain and to document kinematics of localized shear zones. In some cases, solid-state fabrics are used to infer synemplacement deformation, such as the deformation of a marginal facies of an intrusion by an internal facies (e.g., de Saint Blanquat et al., 2001). Solid-state fabrics superimposed on magmatic fabrics represent an overprinting of emplacement-level processes (e.g., Gleizes et al., 1998; Pignotta and Benn, 1999; Zak et al., 2007) or time-transgressive tectonic fabrics in a cooling pluton (e.g., Bouchez et al., 1990; Karlstrom et al., 1993; Aranquen et al., 1996; Mamtani and Greiling, 2005; Baxter et al., 2005; Henry et al., 2009; Majumder and Mamtani, 2009; Sant’Ovaia et al., 2010).

In the case of magmatic deformation, granites provide superb evidence for the timing of fabric development. Igneous rocks that contain >5% melt during cooling are significantly weaker than fully crystallized rocks (e.g., Vigneresse and Tikoff, 1999; Mecklenburgh and Rutter, 2003; Rosenberg and Handy, 2005). Thus, deformation preferentially partitions into semimolten igneous rocks. In contrast, once a granitic body has crystallized, it is typically competent relative to other sedimentary and metamorphic rock types (e.g., Paterson et al., 1989). Thus, magmatic fabrics can be used to infer deformation related to either emplacement processes or regional deformation during a limited time interval following emplacement and prior to complete crystallization. If both the crystallization age of the granite and a cooling history are known, the history of deformation can be constrained.

Sampling and Field Fabrics

Samples were collected from field sites that had previously been dated using U-Pb on zircon (Unruh et al., 2008; Giorgis et al., 2008; Gaschnig et al., 2010, 2013; Braudy et al., 2016). Field-based fabric within the Idaho batholith is often very subtle; foliations were observed at some locations, and lineations were not observed except in the border zone suite. Field-based magmatic foliation throughout the Idaho batholith is defined by the alignment of biotite and, locally, potassium feldspar. In contrast, solid-state foliations are defined by elongate quartz in addition to aligned biotite.
Microstructures

Microstructural Characterization and Classification

The three categories of microstructures in this study are: (1) magmatic; (2) slight solid-state overprint; or (3) solid state. This classification follows from other studies in which microstructures have been used qualitatively to document the conditions of deformation (e.g., Paterson et al., 1989; Bouchez et al., 1990; Bouchez and Gleizes, 1995; Mamantin and Greiling, 2005; Benn, 2010; Sant’Ovaia et al., 2010). A detailed list of microstructural observations for all samples in this study is given in Byerly (2014).

Criteria for magmatic microstructures include: (1) quartz grains that are consistently coarse grained, anhedral with relatively straight grain boundaries, and lack subgrains; (2) alkali and plagioclase feldspar grains that are coarse grained and pervasively exhibit growth twins; and (3) biotite grains that are undeformed and distributed throughout the sample.

The slight solid-state overprint microstructure contains magmatic microstructures, but it also includes some of the following subsolidus microstructures: undulatory extinction of quartz, bulging or interlobate grain boundaries, quartz-filled fractures in feldspar, and deformation twinning in feldspar (Blumenfeld and Bouchez, 1988; Paterson et al., 1989). Quartz grains are large and equant, but grain boundaries are bulging rather than straight. Carlsbad growth twinning is preserved in plagioclase feldspar, and talc twinning is preserved in alkali feldspar. The lack of a well-developed SPO in these samples suggests that solid-state overprinting was minimal.

Criteria for solid-state microstructures are primarily based on the size, shape, and structure of quartz. Quartz microstructures include elongate grains that define a consistent SPO, pervasive undulatory or checkerboard extinction, a bimodal grain-size distribution, and bulging and interlobate grain boundaries. Larger quartz grains are mantled by small, recrystallized grains; quartz also displays subgrains. Solid-state deformation is also evidenced by the presence of kinked biotite grains and deformation twins in plagioclase feldspar grains. Though these microstructures all reflect solid-state strain, they formed at a range of temperatures. For instance, medium- to high-temperature (400–700 °C) solid-state deformation is inferred from checkerboard extinction and anisotropically deflected quartz grains due to recrystallization by subgrain rotation and grain boundary migration (Hirth and Tullis, 1992; Lloyd and Freeman, 1994; Stipp et al., 2002), whereas bulging quartz grain boundaries and patchy undulatory extinction can form at significantly lower temperatures (~300–400 °C, e.g., Stipp et al., 2002).

Observed Microstructures

The U-Pb zircon crystallization ages provided by Uruhu et al. (2008), Giorgis et al. (2008), Gaschnig et al. (2010, 2013), and Braudy et al. (2016) allow us to evaluate these microstructural observations within a temporal framework, which is added to the spatial framework given by the distribution of the samples. We discuss each magmatic suite, from oldest to youngest (border zone, early metaluminous Atlanta suite, late metaluminous, Bitterroot suite), to elucidate the tectonic history of magmatism in the Idaho batholith.

Samples from the border zone and early metaluminous suites exhibit both a magmatic and slight solid-state overprint or solid-state fabrics (Fig. 3). Feldspar grains commonly exhibit deformation twinning and anhedral morphology. Biotite locally defines a SPO and is found in aggregates. Foliations are NS-to-NNE striking, parallel to the regional trend of the western Idaho shear zone. Samples from the early metaluminous suite locally show strong fabric development, but the directions are not regionally consistent.

The Atlanta peraluminous suite preserves dominantly magmatic microstructures (Fig. 3). Grain size is commonly coarse, with blocky grains, suggesting that the original magmatic microstructure is preserved. Though some quartz grains have serrated grain boundaries and undulatory extinction, they are generally equant and sometimes preserve 120° triple junctions. Alkali feldspar is commonly subhedral and blocky with the twin twinning, and oscillatory zoning is found in many plagioclase feldspar grains. Mica is distributed throughout samples, and the mica cleavage is rarely kinked or fractured. The Atlanta peraluminous suite locally records slight solid-state overprint, but there is no trend in the spatial distribution. We interpret fabric preserved in the rocks of the Atlanta peraluminous suite to represent fabric acquired in the magmatic state, immediately before full crystallization.

Solid-state microstructures were identified in the Bitterroot peraluminous and late metaluminous suites, indicating that magmatic microstructures were modified or obliterated by subsolidus deformation. Quartz in samples from this suite commonly displays undulatory or checkerboard extinction, is occasionally dynamically recrystallized, and in extreme cases is elongate and defines an SPO (Fig. 3). Alkali feldspar is commonly subhedral to anhedral and occasionally fractured. Plagioclase feldspar often displays deformation twins (Fig. 3). In samples with a weak solid-state overprint (e.g., 06RMG03), magmatic microstructures still dominate. In samples with strong solid-state fabric (e.g., MC25-91), subsolidus microstructures dominate, and any fabric formed prior to crystallization has been drastically altered or obliterated.

SHAPE PREFERRED ORIENTATION

Methodology

These weak and dominantly magmatic fabrics, as characterized in the previous paragraphs, were quantified by SPO. Fabric in intrusive rocks can be defined by the alignment of minerals, alignment of mineral aggregates, deformed phases such as quartz ribbons, or deformed markers such as enclaves. Fabric in rocks that preserve original igneous texture is defined by the early crystallizing phases, which are able to rotate into alignment in the presence of melt. In the Idaho batholith granites, fabrics are generally defined by aligned biotite.

The method of SPO analysis provides a quantitative fabric orientation, shape, and strength for each sample. Orientation is given by a fabric ellipsoid with axes K1 ≥ K2 ≥ K3 that defines the lineation (long axis = K1) and the pole to the foliation (short axis = K3) in a geographic reference frame. The shape of the fabric is defined by the shape parameter, \( T = [(\log[K2/K3] - \log[K1/K3])/(\log[K2/K3] + \log[K1/K3])] \), which ranges from 1 (oblate) to 0 (prolate), representing constrictional or flattening deformation, respectively (Jelinek, 1981). The strength of the fabric is defined by the degree of anisotropy, \( P' = (K1/K3) \). Because all intrusive suites in the Idaho batholith contain biotite, this phase was used as the basis for SPO comparison.

Three mutually perpendicular thin sections were prepared for 46 samples. Each sample was oriented in the geographic coordinate system: one horizontal face, one north-striking vertical face, and one east-striking vertical face. Each thin section was photographed using a digital camera attached to a microscope. Biotite was isolated in binary images using Imaged (National Institute of Health’s public domain image processing software). These thin sections were also compared to unoriented thin sections used during earlier geochronological-based studies (Gaschnig et al., 2010; Braudy et al., 2016).

Binary images of biotite grains were analyzed using Patrick Launeau’s SP02003 and Ellipsoid2003 software, utilizing the intercept method (Launeau et al., 1990; Launeau and Robin, 1996). The intercept method is used for determining the alignment of phases within a series of mutually perpendicular two-dimensional planes, which are then combined into a three-dimensional ellipsoid. The number of intercepted grain boundaries is counted along a set of parallel scan lines that are rotated around the image by a specified angular increment. The
direction in which these intercept counts is the
lowest is the direction of preferred orientation.
The aggregate preferred orientation of grains is averaged to determine the average SPO within a
given thin section. SPO2003 analyzes the inertia
and intercept tensor of each thin section indi-
vidually, producing a two-dimensional ellipse
that quantifies the fabric strength and orientation
for the phase of interest. The three ellipses from
the different faces of each sample are combined
into a three-dimensional ellipsoid using Ellip-
soid2003. The result is an average fabric ellip-
soid that can be characterized for each sample
by orientation, shape, and degree of anisotropy.

Results

The results of SPO analysis of biotite suggest
weak fabrics with a wide range of fabric shapes. The degree of anisotropy (P') of SPO fabrics
determined using the intercept method ranges
from 1.06 to 1.23. SPO foliations of the border zone suite also strike north and dip steeply to the east. The early metaluminous plutons on the eastern and
northern borders of the Atlanta lobe show no con-
sistent trend in SPO foliation or lineation orienta-
tion. The degree of anisotropy for fabrics deter-
mined using the intercept method ranges from
-0.85 to 0.78.

The Atlanta peraluminous suite exhibits
highly variable fabric strength, shape, and ori-
entation (Fig. 5). SPO foliation and lineation
within the Atlanta peraluminous suite have no

Figure 3. Representative microstructures in the Idaho batholith. Left: Magmatic microstructures include coarse-grained, roughly
equant quartz, and both growth twins and concentric zonation in feldspar. Center: Samples with weak solid-state overprint show
similar characteristics to magmatic microstructures, but with some interlobate grain boundaries and undulatory extinction. Right:
Solid-state microstructures include elongate quartz grains with undulatory extinction, subgrains, and recrystallized grains.

Figure 4. Results of shape preferred orientation (SPO) analyses. Plot of shape parameter (T) vs.
degree of anisotropy (P'), indicating no correlation
between the fabric strength and shape parameter.
consistent trend in orientation. The degree of anisotropy for fabrics determined using SPO ranges from 1.03 to 1.16. Fabrics are dominantly weakly oblate, with a range in T of -0.5 to 0.75. Both the strength and shape of SPO fabrics are variable through time, showing no clear trend through the 16 m.y. of Atlanta peraluminous magmatism.

Foliation in the Bitterroot lobe strike NNW and dip moderately to the NE, parallel to those measured in outcrop. Lineations are consistently subhorizontal. The degree of anisotropy for fabrics determined using SPO ranges from 1.08 to 1.12. Fabrics are dominantly weakly oblate, with shape parameter values from -0.28 to 0.28.

ANISOTROPY OF MAGNETIC SUSCEPTIBILITY (AMS)

Methodology

Magnetic susceptibility is a dimensionless proportionality defined as the ratio of the magnetization induced in a sample to the strength of the applied field (e.g., Hrouda, 1982). Susceptibility is commonly directional, and AMS is a second-rank tensor that can be represented by an ellipsoid. The AMS ellipsoid has three mutually perpendicular axes, where K1 ≥ K2 ≥ K3. The long axis, K1, represents the magnetic lineation, and the short axis, K3, is the pole to the magnetic foliation. The orientation, shape, and strength of the magnetic susceptibility ellipsoid (AMS ellipsoid) serves as a proxy for petrofabric measurements (Jelinek, 1981; Hrouda, 1982; Tarling and Hrouda, 1993; Bouchez, 1997; Borradaille and Jackson, 2004) because magnetic axes of common paramagnetic and ferromagnetic phases are parallel to shape axes. Bulk susceptibility (K1 + K2 + K3)/3 varies by orders of magnitude depending on the class of magnetic material (e.g., diamagnetic, paramagnetic, and ferromagnetic, in order of increasing bulk susceptibility). Diamagnetic phases, such as quartz, have low-magnitude, slightly negative intrinsic susceptibilities that contribute little to the composite AMS signal if paramagnetic or ferromagnetic grains are present. Paramagnetic phases such as biotite have a moderately large positive susceptibility that will control the AMS if only diamagnetic and paramagnetic phases are present. The much higher intrinsic susceptibility of ferromagnetic grains (three orders of magnitude greater than paramagnetic phases) will dominate the signal when present (e.g., Hargraves et al., 1991; Grégoire et al., 1995).

The AMS signal of granite is commonly dominated by either paramagnetic (biotite and/or hornblende as the dominant magnetic phases) or ferromagnetic (magnetite and/or other iron

Figure 5. Foliation (above) and lineation (below) results from shape preferred orientation (SPO, right) and anisotropy of magnetic susceptibility (AMS, left) analyses. Foliation strikes consistently N-S in the border zone suite on the western boundary of the batholith. Foliation strikes consistently NW in the Bitterroot lobe. In the Atlanta lobe, fabric shows no trend in orientation. Lineations are highly variable, except in the Bitterroot peraluminous suite, where they are shallow to subhorizontal.
oxides as the dominant magnetic phase) minerals. The AMS signal in granitic rocks is due to either grain shape or crystallographic anisotropy. In paramagnetic materials, the crystallographic axes are parallel to the magnetic axes. In biotite, the crystallographic axes are also parallel to the shape axes. AMS fabrics formed by biotite grains, therefore, are a proxy for the grain SPO. AMS fabrics resulting from ferromagnetic grains may be due to either grain shape or grain distribution (e.g., Hargraves et al., 1991; Stephenson, 1994; Grégoire et al., 1995). The presence of single-domain ferromagnetic grains may result in magnetic interactions with each other during AMS analyses. In this case, the AMS ellipsoid will reflect an aggregate anisotropy rather than grain shape anisotropy (e.g., Bouchez, 1997). For all multidomain ferromagnetic materials, the AMS signal is controlled by grain shape. The aggregate AMS represents the average orientation and fabric shape of all of the magnetic grains (both ferromagnetic and paramagnetic) within the measured volume. Quantification of the degree and orientation of this anisotropy allows for rapid, nondestructive measurement of the average SPOs of grains within a sample.

Low-field AMS was measured for samples from 47 sites (42 of which correspond to SPO samples sites). AMS measurements were made at the University of Wisconsin–Madison on an AGICO KLY-3 Kappabridge magnetic susceptibility bridge. Between 5 and 10 specimens (1-in.-diameter core [2.54 cm]) prepared from at least three separate cores were analyzed from each site. Using AGICO Anisoft 4.2 software and the tensor statistical techniques of Jelínek and Kropáček (1978), susceptibility was determined for each of the three measurement axes, and data were combined into AMS ellipsoids.

**Magnetic Characterization Methods**

Magnetic characterization of the samples was carried out at the Institute for Rock Magnetism at the University of Minnesota, Twin Cities. Bulk susceptibility results from AMS analyses were used to select representative samples from each magmatic suite. Susceptibility versus temperature measurements were carried out on 17 samples (Fig. 6). For samples with high magnetic susceptibility, hysteresis measurements were carried out on 17 samples (Fig. 6). For samples with high magnetic susceptibility, hysteresis measurements were performed (Fig. 7). Details of the analyses are given in Byerly (2014).

**Magnetic Characterization Results**

Results from high-temperature susceptibility measurements confirm the presence of a ferromagnetic phase in samples with high bulk magnetic susceptibility (Fig. 6). The early metaluminous and Bitterroot suite samples show a consistent magnetic unblocking temperature of ~575 °C, characteristic of magnetite (Fig. 6). Thus, the AMS signal from high-magnetic-susceptibility samples is controlled by ferromagnetic minerals. Some samples (Atlanta, late metaluminous suites) record a gradual change in susceptibility across the temperature range, indicative of paramagnetic mineralogy, but with a larger change at ~575 °C (see Byerly, 2014). For these samples, we interpret the AMS signal to be controlled by a paramagnetic component with a secondary ferromagnetic component.

Representative results of hysteresis measurements from high-magnetic-susceptibility samples are shown in Figure 7. Out of the 17 analyses, 16 contain multidomain magnetite, and one contains pseudo-single-domain magnetite (Fig. 8). These results suggest that multidomain magnetite will dominate the AMS signal, consistent with most magnetite-bearing granitic rocks (e.g., Uyeda et al., 1963; Bouchez, 1997). As such, we anticipate a direct correlation between magnetic and shape fabrics. For samples that were not characterized using these techniques, the bulk magnetic susceptibility determined during AMS measurements was used to classify samples in the University of Wisconsin–Madison on an AGICO KLY-3 Kappabridge magnetic susceptibility bridge. Between 5 and 10 specimens (1-in.-diameter core [2.54 cm]) prepared from at least three separate cores were analyzed from each site. Using AGICO Anisoft 4.2 software and the tensor statistical techniques of Jelínek and Kropáček (1978), susceptibility was determined for each of the three measurement axes, and data were combined into AMS ellipsoids.

**Magnetic Characterization Methods**

Magnetic characterization of the samples was carried out at the Institute for Rock Magnetism at the University of Minnesota, Twin Cities. Bulk susceptibility results from AMS analyses were used to select representative samples from each magmatic suite. Susceptibility versus temperature measurements were carried out on 17 samples (Fig. 6). For samples with high magnetic susceptibility, hysteresis measurements were performed (Fig. 7). Details of the analyses are given in Byerly (2014).

**Magnetic Characterization Results**

Results from high-temperature susceptibility measurements confirm the presence of a ferromagnetic phase in samples with high bulk magnetic susceptibility (Fig. 6). The early metaluminous and Bitterroot suite samples show a consistent magnetic unblocking temperature of ~575 °C, characteristic of magnetite (Fig. 6). Thus, the AMS signal from high-magnetic-susceptibility samples is controlled by ferromagnetic minerals. Some samples (Atlanta, late metaluminous suites) record a gradual change in susceptibility across the temperature range, indicative of paramagnetic mineralogy, but with a larger change at ~575 °C (see Byerly, 2014). For these samples, we interpret the AMS signal to be controlled by a paramagnetic component with a secondary ferromagnetic component.

Representative results of hysteresis measurements from high-magnetic-susceptibility samples are shown in Figure 7. Out of the 17 analyses, 16 contain multidomain magnetite, and one contains pseudo-single-domain magnetite (Fig. 8). These results suggest that multidomain magnetite will dominate the AMS signal, consistent with most magnetite-bearing granitic rocks (e.g., Uyeda et al., 1963; Bouchez, 1997). As such, we anticipate a direct correlation between magnetic and shape fabrics. For samples that were not characterized using these techniques, the bulk magnetic susceptibility determined during AMS measurements was used to classify samples in

![Figure 6. Representative hysteresis loops from a sample from each suite. The nonlinear hysteresis loops indicate a ferromagnetic component of the rock for the early metaluminous (and border zone) and Bitterroot peraluminous suites. Dominantly paramagnetic behavior occurs in Atlanta peraluminous and late metaluminous suites.](https://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/1001569/283.pdf)
Figure 7. Representative magnetic susceptibility vs. temperature plots for a sample from each suite. Gray line is the heating curve; black line is the cooling curve. A dramatic decrease in magnetic susceptibility at ~575 °C indicates the presence of magnetite (a ferromagnetic mineral). The Atlanta peraluminous and late metaluminous suites contain some magnetite, but significantly less than the early metaluminous (and border zone) and Bitterroot peraluminous suites.
the paramagnetic-dominant and ferromagnetic-dominant fields.

AMS Results

Bulk susceptibility of measured samples from the Idaho batholith ranges from 13 to $17,400 \times 10^{-6}$ SI, with an average of $3724 \times 10^{-6}$ SI (Fig. 9). The high bulk susceptibility indicates that the AMS signals of many granites are dominated by the ferromagnetic component. Only the early metaluminous (98–85 Ma) and Atlanta peraluminous (83–67 Ma) suites have sites that are dominated by the paramagnetic component. The degree of anisotropy ($P'$) ranges from 1.014 to 1.881. For all samples in this study, the shape parameter, $T$, ranges from $-0.805$ to 0.941, encompassing a wide range of fabric shapes. The average fabric shape is slightly oblate (Fig. 9).

The AMS fabrics of the suture zone and border zone suites exhibit N–S–oriented foliation with a steep E dip. These fabrics are similar to field measurements and fabrics determined using SPO analyses. The bulk magnetic susceptibility in these early phases of the batholith ranges from paramagnetic to ferromagnetic. Early metaluminous granites show no consistency in orientation. The corrected degree of anisotropy from AMS results ranges from 1.05 to 1.28. Fabric shapes are dominantly oblate, with a range of $-0.36$ to 0.72.

The fabrics measured using AMS in the Atlanta peraluminous suite are consistently very weak and highly variable in orientation (Figs. 5 and 9). Bulk susceptibility is variable across the Atlanta lobe, suggesting both ferromagnetic (eight samples) and paramagnetic (10 samples) minerals dominate the AMS signal. The corrected degree of anisotropy from AMS results ranges from 1.02 to 1.32. Fabric shapes are weakly prolate in the center of the suite, and oblate toward the margins. The shape parameter ranges from $-0.35$ to 0.52. The AMS foliation differs, in some cases, from the SPO foliation. In general, the mismatch is due to the presence of very weak fabrics, such that the two different
methods are recording small deviations from random fabric. In a few cases, the mismatch is due to large degrees of variation within a sample site, leading to a poorly constrained average AMS ellipsoid (samples 07RMG45 and 10RMG20). Magnetic fabrics of the Bitterroot lobe depend on the intrusive suite, with the Bitterroot peraluminous suite having stronger and more consistently oriented AMS fabrics than the late metaluminous suite (Figs. 5 and 9). The two late metaluminous suite sites analyzed in this study have low-anisotropy (average 1.07), oblate fabrics. Foliation is consistent in the Bitterroot peraluminous suite, and AMS fabrics are in general agreement with field measurements and results from SPO measurements. AMS foliations consistently strike NW and dip to the NE, and measurements are consistent in the field, in AMS, and in SPO measurements. The degree of anisotropy from AMS results ranges from 1.22 to 1.53. Fabric shapes are dominantly oblate, with a shape parameter range of −0.08 to 0.52.

Figure 10 shows the variation in the AMS fabric through time. The AMS results are plotted using the degree of anisotropy (P’). The timing is constrained through the U-Pb zircon ages: The presence of magmatic microstructures and the relatively upper-crustal emplacement of the Idaho batholith allow us to assume that the crystallization age is approximately the time of fabric development. As such, we note that within the Idaho batholith, strong fabrics formed at two different periods. The first period occurred during emplacement of the border zone and early metaluminous suite, during or immediately after deformation in the western Idaho shear zone. The second period occurred during emplacement of the Bitterroot peraluminous belt.

### 40Ar/39Ar THERMOCHRONOLOGY

#### Methodology

Three samples, previously dated for U-Pb zircon crystallization ages, were analyzed for 40Ar/39Ar closure ages to constrain the timing of the solid-state fabric development. Each sample was crushed, sieved, and separated for individual biotite phenocrysts. Biotite phenocrysts along with the 28.201 ± 0.046 Ma Fish Canyon tuff sanidine standard (Kuiper et al., 2008) were irradiated for 40 h at the Oregon State University TRIGA reactor in the Cadmium-Lined In-Core Irradiation Tube. Single biotite phenocrysts were incrementally heated using a 25 W CO2 laser in the WiscAr Geochronology Laboratory. Gas cleanup and isotopic analysis using the MAP 215–50 mass spectrometer followed the methods of Jicha et al. (2006). The ages reported herein are shown with 2σ analytical uncertainties (including J uncertainty) and were calculated using the decay constants of Min et al. (2000).

#### Results

### Suture Zone Suite

Sample 10RMG011 (Fig. 2) is a biotite-hornblende tonalite from within the western Idaho shear zone (Sage Hen orthogneiss of Braudy et al., 2016), with a zircon U-Pb age of 93.0 ± 3.4 Ma (Gaschnig et al., 2010). The biotite plateau 40Ar/39Ar of 69.01 ± 0.22 Ma (Fig. 11). Sample 07RMG56 (Fig. 2) is a biotite granodiorite of the Atlanta peraluminous suite with a zircon U-Pb age of 76.3 ± 2.6 Ma. This sample yielded a plateau biotite 40Ar/39Ar of 46.89 ± 0.15 Ma (Fig. 11).

Samples in the Atlanta peraluminous suite with similar crystallization ages yielded two very different average cooling rates: a rapid rate for 10RL896, between 86.0 °C/m.y. and nearly

![Figure 10. Plot of anisotropy of magnetic susceptibility (AMS) fabric strength through time. Fabrics in the Atlanta peraluminous suite (83–67 Ma) and the Challis intrusive suite (53–41 Ma) are consistently weak. The strongest fabrics are found in the border zone, early metaluminous, and Bitterroot peraluminous suites.](https://www.gsapubs.org/doi/10.3139/106.1002569.283.pdf)
DISCUSSION

Constraints on Regional Tectonics during Construction of the Idaho Batholith

The integration of fabric data with existing zircon (U-Pb) and new biotite ⁴⁰Ar/³⁹Ar ages constrains timing of fabric development (Fig. 11). The zircon (U-Pb) and biotite ⁴⁰Ar/³⁹Ar ages give maximum and minimum ages of fabric development, respectively. The AMS and SPO thus record the deformational regime during the time period between zircon crystallization and cooling through the argon closure temperature in biotite.

Suture zone and border zone suites. The suture zone suite samples were affected by the western Idaho shear zone and are characterized by strong solid-state foliation. Fabrics measured using SPO and AMS in samples from these early plutons are strong, N-S–striking and E-dipping foliation and down dip lineation, consistent with field measurements (Fig. 12). The border zone suite shows similar fabrics. The ca. 91 Ma Payette River tonalite, a part of the border zone suite and located immediately east of the

Figure 12. Tectonic evolution of the Idaho batholith, shown for different times (98–85 Ma, 83–67 Ma, and 75–53 Ma). Pink indicates location of active magmatism. Dark black line indicates the location of the cross section that characterizes the interaction between magmatism and tectonism for that time interval. Lower-hemisphere equal-area projections of the poles to foliation for shape preferred orientation (SPO) and anisotropy of magnetic susceptibility (AMS) results are given below. Early and late phases of the batholith record fabrics that formed due to regional strain. Fabrics in the Atlanta peraluminous suite are relatively weak. The sample locations are labeled in Figure 2. LCL—Lewis and Clark line.
western Idaho shear zone, records a N-S–striking fabric that grades eastward into a parallel magmatic fabric. Fabrics in this unit, therefore, resulted from waning western Idaho shear zone deformation that ceased at 90 Ma (Giorgis et al., 2008). In contrast, the younger metaluminous granites of the border zone suite exhibit N-S–striking foliation, parallel to the western Idaho shear zone fabrics, but they lack a consistent downdip lineation. We interpret these fabrics to result from continued regional contraction until 85 Ma (Giorgis et al., 2008).

**Early metaluminous suite.** The early metaluminous plutons (98–85 Ma) are found on the northern and eastern boundaries of the Atlanta lobe (Fig. 2) and as roof pendants and septa in younger batholith units. Fabrics measured in the early metaluminous plutons are weaker and are not parallel to the strong fabrics of the border zone suite (Fig. 12). Given that the ages of the early metaluminous suite plutons and the border zone suite plutons overlap, regional deformation at this time was likely partitioned. Transpression deformation associated with the western Idaho shear zone may have been restricted to the western edge of the Idaho batholith.

**Atlanta peraluminous suite.** The 83–67 Ma Atlanta peraluminous suite is lithologically homogeneous and structurally isotropic at the outcrop scale. Magmatic microstructures are preserved in many samples, and solid-state overprinting, where present, is weak. Foliation in the Atlanta peraluminous suite show no consistent trend in orientation and no clear progression in fabric orientation through time, suggesting a lack of distributed regional deformation (Fig. 12). AMS results suggest generally a low degree of anisotropy and poorly constrained AMS ellipsoids. Further, fabric measurements for the same sample site differ in strength and orientation when comparing the AMS and SPO results. In summary, the granodiorites of the Atlanta peraluminous suite appear isotropic.

The approximately N-S–trending Deadwood structure (Johnson Peak–Profile Gap shear zone of Lund, 2004) does occur within Atlanta lobe. Because the timing of deformation is presently unconstrained, we do not incorporate this zone into a model of Atlanta peraluminous suite development.

Due to the lack of pervasive and consistent deformation, we envision the fabrics of the Atlanta peraluminous suite to represent magma flow within a series of intrusions. Because there has been no significant subsolidus deformation, the magmatic structure of these individual intrusions has likely been preserved. The margins of the intrusions may also have been obliterated due to local remelting of intrusive contacts during intrusion of subsequent pulses of magma. Our interpretation of discrete, but obscured, individual plutons is consistent with the protracted growth history of the Atlanta peraluminous suite as evidenced by the ubiquitous and complexly zoned zircon grains analyzed in recent studies (Gaschnig et al., 2010, 2011).

**Late metaluminous and Bitterroot peraluminous suites.** Construction of the 75–53 Ma Bitterroot lobe exhibits a renewed cycle of low-volume metaluminous magmatism followed by higher-volume peraluminous magmatism. Only two samples were collected from the late metaluminous suite, and they have weak fabrics with orientations that are not consistent with nearby measurements (Fig. 12).

Fabrics in the Bitterroot peraluminous suite are some of the strongest in the batholith. Foliation strikes NW–SE, and lineations are shallowly plunging. Due to the consistency of foliation orientation, the higher degree of anisotropy, and the presence of some solid-state fabrics, we interpret the fabric of the Bitterroot peraluminous suite to represent distributed regional strain. This regional deformation—and the Bitterroot lobe itself—is limited to northern Idaho (e.g., Foster et al., 2001). Regional deformation did occur in northern Idaho at this time, as evidenced by movement on multiple, large-scale structures in the region, such as (1) the Lewis and Clark line and (2) the Coolwater culmination (Wallace et al., 1990; Lund et al., 2008). The Lewis and Clark line is a WNW–ESE–trending zone of faults and shear zones that cuts through northern Idaho and western Montana (Wallace et al., 1990; Sears and Hendrix, 2004). Foliation orientations in the Bitterroot peraluminous suite are parallel to the Lewis and Clark line, which experienced sinistral transpression during the intrusion of the Bitterroot lobe (e.g., McClelland and Oldow, 2007). Contractional, NE-oriented deformation is also documented on the Coolwater culmination, to the southwest of the Bitterroot lobe, at this time (86–61 Ma) (Lund et al., 2008). The Coolwater culmination is interpreted to have formed by wedging of oceanic crust into the Syringa embayment (Fig. 2) on the margin of Laurentia, and this interpretation is consistent with NE-SW–directed shortening (Lund et al., 2008). Given the lack of evidence of deformation anywhere in the Atlanta lobe, Late Cretaceous–Paleogene strain appears to have been localized in the Syringa embayment (Schmidt et al., 2016).

**85–70 Ma Orogenic Plateau in Idaho**

It is difficult to account for the weak fabrics of the Atlanta peraluminous suite (Atlanta lobe) of the Idaho batholith. The Atlanta peraluminous suite was emplaced within a regional contractional regime, as evidenced by ongoing deformation in the Sevier fold-and-thrust belt to the east (e.g., Wiltshchko and Dorr, 1983; DeCelles and Mitra, 1995). With the degree of shortening attributed to the Sevier orogeny—an estimated 220 km of shortening occurred on the Sevier fold-and-thrust belt (DeCelles and Coogan, 2006)—we would expect to see strong tectonically imparted fabrics, both magmatic and solid state, in the voluminous Atlanta peraluminous intrusive suite.

The Atlanta lobe of the Idaho batholith likely intruded into a crustal plateau. A fault restoration model by Coney and Harms (1984) shows that the Cordilleran crust was up to 60 km thick in central Idaho. This finding is consistent with recent seismic data (K. Davenport, 2016, personal commun.), which constrain the current depth of the Moho to 37–40 km below current exposure levels in the Atlanta lobe of the Idaho batholith. Using the 3–4 kbar emplacement pressures for the exposed early metaluminous suite (98–85 Ma; Gaschnig et al., 2010), based on the AI-in-hornblende geobarometer (Jordan, 1994), the maximum emplacement depth for the Atlanta peraluminous suite is ~12 km. The geophysical and geobarometry data together constrain the thickness of the Idaho batholith to be ~50 km during emplacement of the Atlanta suite. This crustal plateau is also consistent with central Idaho being a major sediment source for Late Cretaceous paleoivers draining to basins in California, Wyoming, and Washington (e.g., Dumitrzu et al., 2016).

The Atlanta lobe is part of a continent-scale two-mica granite belt that is immediately inboard of the other North American Cordilleran batholiths (Fig. 1; Miller and Bradflish, 1980). The belt stretches ~1300 km from Arizona to Canada and is evidenced by 41 muscovite-bearing, dominantly quartz monzonite or granitic plutonic units. These rocks have generally high silica (>70%) and high ^87Sr/^86Sr (0.7086–0.734), indicating a more peraluminous composition than the Cordilleran arc batholiths (Miller and Bradflish, 1980). The Idaho batholith is the greatest contiguous volume in this discontinuous belt. The geochemical data from the Atlanta peraluminous suite are consistent with the interpretation of emplacement in overthickened crust (Gaschnig et al., 2011). The zircons from the Atlanta peraluminous suite display extensive inheritance and multiple generations of Cretaceous growth (Gaschnig et al., 2010, 2016), in addition to Precambrian inheritance (Gaschnig et al., 2013). The evolved Nd and Hf isotopes imply only limited, if any, mantle input (Gaschnig et al., 2011). Together, these data suggest large-scale crustal melting, which we interpret to have been caused by crustal thickening. In contrast, the other Cordilleran arc batholiths...
(e.g., Sierra Nevada, Coast plutonic complex, Peninsular Range) have more voluminous mafic components, and the granitoids are dominantly metaluminous, suggesting contribution of juvenile mantle material (e.g., Bateman, 1992; Todd et al., 1998; McNulty et al., 2000; Ortega-Rivera, 2003; Gehrels et al., 2009).

The weak fabrics of the Atlanta peraluminous suite can potentially be explained by emplacement in a crustal plateau. The crustal thickening required to generate a plateau may lead to contrasting styles of deformation within an orogen. In overthickened crust, a vertical gradient can develop in lithospheric deformation, transitioning from contraction at the plateau roots to extension in the upper crust, with an intervening plane of neutrality (Molnar and Lyon-Caen, 1988; Rey et al., 2001). Crustal thickening in the Andean and Himalayan plateaus has been shown to lead to gravitational instability, resulting in extensional accommodation in the upper crust (e.g., Hodges et al., 1992; Fleisch and Kreeger, 2010). The lack of strong tectonic fabric in the Atlanta peraluminous suite suggests that the current exposure level may reflect this transition between contraction and extension (Fig. 12). This interpretation accounts for ongoing shortening in the foreland as a product of lateral variations in gravitational potential energy (e.g., Molnar and Lyon-Caen, 1988).

The difficulty with this interpretation is that the level of neutral extension only extends a limited distance down from the surface of a crustal plateau. The depth of the transition is given by \( z = \text{horizontal stress/(density} \times \text{gravitational acceleration)} \) (Molnar and Lyon-Caen, 1988). For density = 2700 kg/m\(^3\) and gravitational acceleration = 9.8 m/s\(^2\), this equation becomes horizontal stress/26,460 Pa. Estimates for horizontal stress from crustal plateaus are 25 MPa (Peruvian Andes; Richardson and Cobelintz, 1994), 7–10 MPa (Himalaya; Panthi, 2012), 4–25 MPa (eastern margin of the Tibetan Plateau; Meng et al., 2015), and ~10 to 20 MPa (eastern margin of the Tibetan Plateau; Styron and Hetland, 2015). Assuming a horizontal stress of 20 MPa, the depth where vertical stress is equal to horizontal stress (“neutral” stress) is ~755 m. However, Copley et al. (2009) suggested that if mountains were supported by the ductile part of the lithosphere (e.g., lithospheric mantle), then stresses of 80–240 MPa are supported. If so, the depth of “neutral” stress is ~3–9 km. Thus, it is permissible that upper-crustal magmatism in a crustal plateau could occur in a neutral environment.

An alternative model is that the magmas of the Atlanta lobe were intruded as thin, horizontal sheets. This mode of emplacement has been documented in mafic-silicic layered intrusions studied by R. Wiebe and colleagues based on exposures in Maine, United States (e.g., Wiebe, 1993; Wiebe and Collins, 1998). Horizontal layering has since been recognized in dominantly silicic systems (Miller and Miller, 2002; Harper et al., 2004; Waight et al., 2007). Further, individual intrusive horizontal sheets have been recognized in the shallow (<5 km depth) silicic intrusions of the Henry Mountains, Utah (e.g., Horsman et al., 2005; Morgan et al., 2008). The lack of clear plutonic boundaries in the Atlanta lobe of the Idaho batholith is permissive of this geometry. Each of these magma sheets would crystallize and cool in a geologically short (~10^6 yr) period of time at upper-crustal (<10 km) depths (e.g., Nyman et al., 1995). As such, these magma sheets, which are relatively viscous, may not record regional contraction in a manner similar to a larger pluton. A difficulty with this model is that ductile shear zones are commonly reported from the margins of thin, fast-cooling meter-scale tabular intrusions (Fémines et al., 2004; Horsman et al., 2005; Creixell et al., 2006).

We cannot discriminate between these two options, or the possibility of their simultaneous operation, as the cause of the ubiquitous weak fabrics recorded in the Atlanta lobe. In order to fully evaluate these scenarios, sampling and geochronologic, microstructural, and fabric analysis would need to be done at a much finer spatial scale.

**CONCLUSIONS**

The Idaho batholith consists of series of intrusive suites (Gaschnig et al., 2010) that exhibit distinctive variations in fabric development. We were able to constrain the orientation and magnitude of tectonic strain through time within the batholith, because the granites: (1) record dominantly magmatic fabrics; (2) cooled relatively quickly (~3 m.y.) through ~350 °C as constrained by ⁴⁰Ar/³⁹Ar closure ages on biotite; and (3) have crystallization ages known from U–Pb zircon analyses (Gaschnig et al., 2010, 2013; Braudy et al., 2016). Fabrics in the suture and border suites, on the western margin of the Idaho batholith, were affected by Late Cretaceous deformation associated with the western Idaho shear zone. These fabrics are typically N-S oriented and show solid-state to slightly solid-state microstructures. The Atlanta peraluminous suite, despite being emplaced during orogenesis, records very weak and randomly oriented fabrics, and it contains only magmatic microstructures. The Bitterroot peraluminous suite exhibits strong and regionally consistent NW-oriented fabrics, with some solid-state microstructures. In general, the SPO fabrics determined from biotite and the AMS fabrics are consistent throughout the Idaho batholith, with the exception of the weakly developed fabrics in the Atlanta peraluminous suite. The lack of strong fabrics in the voluminous Atlanta peraluminous suite, in combination with the crustal thickness estimates and geochemical analyses, is consistent with emplacement in an orogenic plateau and/or sill-like emplacement.

**ACKNOWLEDGMENTS**

We thank Michael Kedenburg and Joey Lane for their help as field assistants for A. Byerly, C.E. Bate and N. Garbaldo for their patience and support, and F. Heron and Michael O’Keefe for their constructive reviews. We are grateful to three anonymous reviews whose comments significantly improved the quality of the manuscript. This work was supported by National Science Foundation grants EAR-0644260 and EAR-1251877 to B. Tikoff (University of Wisconsin–Madison).
Launeau, P., Bouchez, J.-L., and Benn, K., 1990, Shape pre-
Launeau, P., and Robin, P., 1996, Fabric analysis using the
Karlstrom, K., Miller, C. F., Kingsbury, J. A., and Wooden, J.
Lund, K., Snee, L., and Evans, K., 1986, Age and genesis of
Kuiper, K., Deino, A., Hilgen, F., Krijgsman, W., Renne, P., and
Richards, R.M., and Coblenz, D.D., 1994, Stress model-
ROD017.1.
Rosenberg, C.L., and Handy, M.R., 2005, Experimental de-
Sant’Ovaa, H., Olivier, P, and Ferreira, N., 2010, Magmatic structures and kinematic implications of the Variscan granites from central Portugal (Serra da Estrela and Cas-
Schmidt, K.L., Lewis, R.S., Vervoort, J., Stetsen-Lee, T.A., Mi-
chels, D.Z., and Tokilo, B., 2016, Tectonic evolution of the Syrtaki detachment in the central North American Co-
dillarier accretionary boundary: Lithosphere, v. 110.133 (1/5), 1 (in press).
Sen, K., Majumder, S., and Mantami, M.A., 2005, Degree of magmatic anisotropy as a strain intensity gauge in fer-
Shields, J.M., Girty, G.H., Kimbrough, D.L., and Martín-Barajas,
Smith, J.M., 2004, Lewis and Clark line and the Ap-
Snodgrass, R.L., 1973, The determination of magma flow dire-
Spangenberg, W.G., ed., Metamorphism and
tion in the Tarawega Bay Granite, southwestern New Zealand; using anisotropy of magnetic susceptibility: Geo-
Storrey, R.H., and Hetland, E.A., 2015, The weight of the moun-
tains: Constraints on tectonic stress, friction, and fluid pressure in the 2008 Wenchuan earthquake from esti-
Tarling, D., and Hiruda, F., eds., 1993, Magnetic Anisotropy of
lithologies of the East Idaho sheet in the central


