Deciphering viscous flow of frictional melts with the mini-AMS method

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ABSTRACT

The anisotropy of magnetic susceptibility (AMS) is widely used to analyze magmatic flow in intrusive igneous bodies including plutons, sills and dikes. This method, owing its success to the rapid nature of measurements, provides a proxy for the orientation of markers with shape anisotropy that flow and align in a viscous medium. AMS specimens typically are 25 mm diameter right cylinders or 20 mm on-a-side cubes, representing a volume deemed statistically representative. Here, we present new AMS results, based on significantly smaller cubic specimens, which are 3.5 mm on a side, hence~ 250 times volumetrically smaller than conventional specimens. We show that, in the case of frictional melts, which inherently have an extremely small grain size, this small volume is in most cases sufficient to characterize the pseudotachylite fabric, particularly when magnetite is present. Further, we demonstrate that the mini-AMS method provides new opportunities to investigate the details of frictional melt flow in these coseismic miniature melt bodies. This new method offers significant potential to investigate frictional melt flow in pseudotachylite veins including contributions to the lubrication of faults at shallow to moderate depths.

Keywords: pseudotachylite; anisotropy of magnetic susceptibility; fabric; frictional melt; fault, magma flow

1. Introduction

The anisotropy of magnetic susceptibility (AMS) provides invaluable insights into flow kinematics, deformation and emplacement of magmatic materials ranging from mafic to silicic melts as well as pyroclastic deposits (e.g., Ernst and Baragar, 1992; Archanjo et al., 1995; Cañón-Tapia et al., 1996; Ferré and Améglio, 2000). Like all petrofabric methods, the AMS method requires the analysis of a large enough number of grains (s.l., ferromagnetic particles as well as paramagnetic silicate minerals) to obtain statistically significant results (e.g., Borradaile and Henry, 1997). Because the average grain size in most igneous rocks ranges from a few millimeters to hundreds of microns in dimension, the volume of AMS specimens typically is about 10 cm³.

Fault pseudotachylytes generally form relatively thin generation veins, 0.1-10 mm in thickness (e.g., Sibson, 1975; Passchier, 1982; Di Toro and Pennachioni, 2004), along which melt forms and rapidly quenches during or shortly after an earthquake. AMS could be very useful for determining viscous flow in pseudotachylytes. However, because of their limited volumetric occurrence, the standard sample size of 10 cm³ is not suitable for AMS studies of pseudotachylytes. Previous attempts to use AMS of fault pseudotachylytes were performed on 10 cm³ (Nishioka and Ishikawa, 2005) or ~1 cm³ (Molina-Garza et al., 2009) specimens. These investigations proved somewhat inconclusive most likely because specimens were not small enough to completely exclude host-rock material adjacent to the vein. Recently, Ferré et al. (2015), successfully isolated the AMS fabric in the Val Gilba pseudotachylytes (Italian Alps) from that of the host-rock by measuring 1 cm³ specimens. These pseudotachylyte veins were relatively thick (>10 mm), which is unlike most of the reported pseudotachylytes in nature that are generally < 10 mm in thickness and consist of ultra-comminuted clasts embedded in an

extremely fine-grained quenched matrix. Therefore, to continue these types of investigations a method that measures AMS in very small specimens of pseudotachylyte is needed. In this paper, we measure AMS from 3.5 mm cubes ($\approx 0.04 \text{ cm}^3$ volume; 250 times smaller than standard size) and test the consistency of AMS fabric indicative of coseismic slip kinematics.

2. Geologic settings

The fault pseudotachylytes of this study come from four well-studied localities and are hosted by different rock types (Figure 1).

2.1. Hoping River, Taiwan

The Hoping River pseudotachylyte veins cut across the Tananao mylonitic gneiss complex, in eastern Taiwan, and formed less than 4.1 Ma ago (Chu et al., 2012). The host-rock consists of a quartzofeldspathic phengite, biotite and muscovite medium grained high-grade gneiss. The main slip plane forms a generation vein that varies in thickness from 0.1 to 10 mm, extends over 10 m length and records 0.22 m of seismic displacement (Korren et al., 2016). The overall orientation of this generation vein is $025^\circ, 35^\circ \text{ W}$. This vein also displays a clear slip direction, marked by offset piercing points and pseudotachylyte brushline striations at $292^\circ, 35^\circ$ (Ferré et al., Accepted). The oriented sample of this study (HP010P) comes from this vein where the local vein orientation is $146^\circ, 38^\circ \text{ SW}$. The normal sense of seismic slip is determined using offset pinning points and, as shown elsewhere (Ngo et al., 2012), by the obliquity of injection veins with respect to the slip plane. The pseudotachylyte – host rock contact is very sharp over a few microns. Widely dispersed feldspar microcrystallites (0.5 to 10 μm) generally display a radial growth pattern (Figure 2A). Microcrystallite sizes increase from the pseudotachylyte margin towards the vein center, which suggests increasing quench towards the margin. The

pseudotachylyte hosts 10 μm to 2 mm rounded clasts of quartz, alkali feldspar, epidote, and plagioclase embedded in a micaceous matrix.

2.2. Cima di Gratera, Corsica

The Cima di Gratera pseudotachylyte veins occur along the gabbro-peridotite contact of the Schistes Lustrés ophiolitic rocks, in Corsica (e.g., Austrheim and Andersen, 2004; Andersen et al., 2008; Magott et al., 2016). These pseudotachylytes post-date the Paleogene subduction of oceanic lithosphere (Jolivet et al., 2003), are cut by Miocene normal faults (Andersen et al., 2014), but have not been dated. The pseudotachylyte hosts clasts of olivine, orthopyroxene, diopside, and glass (Deseta et al., 2014a and b). The oriented sample for this study (COR006A) comes from a 10 m-long, continuous, up to 30 mm-thick generation vein collected about 20 m below the gabbro-peridotite contact in a Cpx-harzburgite host-rock. The host-rock / pseudotachylyte vein contact is sharp and shows small magnetite grains displaying a slight obliquity with the contact (Figure 2B). This vein was chosen from an area of fresh, moderately serpentinized host-rock. The orange color of the vein results from shallow (<1 mm thick) supergene alteration and the mini-cubes are dark grey in color. The orientation of this generation vein is 050° , 52° SE corresponding to the steep pseudotachylytes dip-slip reverse sense of shear of Magott et al. (2016).

2.3. Santa Rosa Mountains, California

The Santa Rosa Mountains pseudotachylyte veins occur in the Eastern Peninsular Ranges in California and formed ca. 59 ± 1 Ma (Wenk et al., 2000). One of the best exposed localities is located south of Highway 74 along Deep Canyon. At this locality, generation veins occur along the north-dipping contact between the Asbesto Mountain tonalite and the Palm Canyon Complex, with a general 25° north-dipping attitude (Ferré et al., 2014a). The oriented sample of

this study (SR108B) comes from a continuous 10 mm-thick generation vein that can be traced over 10 meters, cutting through a slightly hydrothermally altered biotite-hornblende-titanite tonalitic host-rock. The orientation of this generation vein is 092° , 38° N. The reverse sense of coseismic slip along this vein is determined by offset markers indicating 0.9 m of displacement. The sample of this study is a generation vein that exhibits compositional heterogeneity marked by varying modal concentration of magnetite grains, quartz clasts and microcrystallites (Figure 2C). The microcrystallites consist mainly of euhedral, calcic plagioclase ranging in size from 5 to 100 μm that locally form radiating microstructures. The large microcrystallites do not display a preferred orientation with respect to vein margin. The matrix consists of biotite and calcic amphibole. Magnetite grains are subrounded and range in size from 0.2 to 1.0 μm .

2.4. Outer Hebrides, Scotland

The Outer Hebrides pseudotachylyte veins crop out in northwest Scotland and are associated with the Outer Hebrides Fault Zone that cuts through Archean to Paleoproterozoic quartzofeldspathic gneisses (e.g., Sibson, 1975). This complex fault zone includes cataclasite bands and pseudotachylyte veins cutting through variably deformed host-rock. The general sense of shear along pseudotachylyte veins is top-to-northwest (thrusting). $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations on pseudotachylyte imply that seismic slip took place during the Caledonian orogeny at 430 ± 6 Ma (Kelley et al. 1994), although significantly older age determinations have also been reported (Sherlock et al., 2009). The depth of seismic rupture is estimated to 5-10 km (Sibson and Toy, 2006). The oriented sample for this study (SU2) is from 100 mm-thick generation vein, collected on the Rubha Àird a' Mhuile headland ~10 km west of the westernmost mapped trace of the Outer Hebrides Fault Zone. Overall, the generation vein plane is oriented 170° , 28° E and shows offset markers consistent with a top-to-West (thrusting) sense of shear. At the locality of

sampling, the vein orientation is slightly different (166°, 38°E). Under the microscope, this pseudotachylyte displays a slight banding parallel to the vein margin marked by zones of different clast percentage and elongated magnetite grains (Figure 2D). Skeletal plagioclase microcrystallites embedded in an amphibole-rich matrix are common in this vein.

3. Methods

Mini-cubes (≈ 3.5 mm on a side) were cut from oriented generation veins of pseudotachylyte, as far as possible from large injection veins that could have affected melt flow geometry. All the mini-cubes were prepared from the same side of the vein, along the vein margin, with the exception of SU2 (Scotland) where samples were respectively at 0 mm (4 mini-cubes) and 4 mm (3 mini-cubes) from the vein-host rock contact. The mini-cubes were cut so that one of their faces was parallel to the host-rock / pseudotachylyte contact and with a horizontal cube edge to facilitate sample orientation. The samples were placed in acrylic cubic custom-made samples holders (Figure 3).

Low-field, room-temperature magnetic susceptibility measurements of these mini-cubes were performed using a KLY-4S Kappabridge magnetic susceptometer at a field strength of 450 A/m and a frequency of 875 Hz. The field of 450 A/m was chosen to increase the measured magnetic moment. Digital data was processed with the Agico software and K_1 , K_2 and K_3 parameters were plotted on equal-area stereonetts using the Anisoft 4.2 program. Jelinek statistics on the directional data (Jelinek, 1981) are provided as confidence angles α (Table 1) and plotted as ellipses of confidence for $\alpha \leq 25^\circ$ in stereonetts. Borradaile (2003) showed that these statistics provide meaningful results only for $\alpha \leq 25^\circ$.

Magnetic hysteresis measurements were performed on the mini-cubes of this study on a Princeton Measurements vibrating sample magnetometer (VSM) 3900-04 at Southern Illinois University. An applied field of up to 1 Tesla was used to magnetically saturate all samples. Above magnetic saturation, the slope of a hysteresis loop represents the sum of paramagnetic and diamagnetic contributions (in the absence of antiferromagnetic minerals). We use the high-field slope above 70% of the maximum applied to correct for these contributions and calculate the magnetic hysteresis parameters and the paramagnetic contribution ($K_{\text{para}}/K_{\text{m}}$). Figure 4 shows the hysteresis loop before and after slope correction. The magnetic hysteresis parameters, saturation magnetization (M_{s}), saturation remanent magnetization (M_{r}), magnetic coercivity (H_{c}) and coercivity of remanence (H_{cr}) are given for one representative specimen per locality (Fig. 4).

4. Experimental results

4.1. Rock magnetism

The four studied pseudotachylytes display a broad range of magnetic susceptibilities (K_{m}) and magnetic hysteresis behaviors (Table 1, Figures 4 and 5). For each locality, however, hysteresis properties are consistent among specimens of different veins and among specimens of the same vein. The Hoping River pseudotachylytes of Taiwan display relatively low values of mean $K_{\text{m}} \approx 540 \pm 97 \cdot 10^{-6}$ [SI]. These pseudotachylytes exhibit a dominantly paramagnetic behavior with $K_{\text{para}}/K_{\text{m}} = 99.1\%$ (Figure 4A), and a minute ferromagnetic contribution (Figure 4A inset). This ferromagnetic component, however, is too small to yield reliable M_{r} , H_{c} and H_{cr} data. The Cima di Gratera pseudotachylytes of Corsica show slightly higher values of mean $K_{\text{m}} \approx 762 \pm 95 \cdot 10^{-6}$ [SI] compared with the Hoping River locality. These pseudotachylytes exhibit a mixed paramagnetic and ferromagnetic hysteretic behavior, with $K_{\text{para}}/K_{\text{m}} = 16.5\%$ (Figure 4B).

The ferromagnetic component shows magnetic hysteresis properties indicative of pseudo-single domain (PSD) magnetite. The Santa Rosa pseudotachylytes of California display the greatest values of magnetic susceptibility of all four groups with mean $K_m \approx 74079 \pm 8312 \cdot 10^{-6}$ [SI]. These pseudotachylytes exhibit a primarily ferromagnetic behavior with $K_{para}/K_m = 3.4$ % (Figure 4C). As shown by Ferré et al. (2005, 2012, 2014a) and our new results (Figure 4C), this group is dominated by PSD magnetite. The Outer Hebrides pseudotachylytes of Scotland also show very large values of magnetic susceptibility with mean $K_m \approx 52317 \pm 11384 \cdot 10^{-6}$ [SI]. In addition, these pseudotachylytes have the most ferromagnetic behavior with $K_{para}/K_m = 1.8$ % only (Figure 4D). The ferromagnetic component shows magnetic hysteresis properties indicative of PSD magnetite.

4.2. Anisotropy of magnetic susceptibility

The AMS scalar parameters, P' , T and K_m , are controlled by the nature of the magnetic phases responsible for anisotropy as well as their spatial arrangement (i.e., fabric) of these grains (e.g., Borradaile and Jackson, 2004). The corrected degree of magnetic anisotropy P' (Jelinek, 1981) describes the strength of the magnetic fabric. Overall, this parameter is relatively moderate for magmatic materials except in the Taiwanese specimens for which P' shows higher values (up to 1.344) (Table 1; Figure 5). Our discussion will show that these anomalously high P' values arise from low signal-to-noise ratios, and hence are meaningless. Specimens from the other three localities have consistent P' values for specimens from the same set: for Corsica, $P' = 1.060 \pm 0.028$; for California, $P' = 1.041 \pm 0.06$; and for Scotland, $P' = 1.100 \pm 0.018$. The shape factor T (Jelinek, 1981) provides information on the symmetry of the AMS fabric. For $T < 0$, the AMS ellipsoid is prolate, whereas for $T > 0$, the AMS ellipsoid is oblate. In turns, the AMS fabric symmetry reflects the fabric of magnetic grains, PSD magnetite in this case. The T values for the

Taiwanese specimens (Figure 5B and C) are deemed statistically unreliable ($\alpha > 25^\circ$). For the other three localities, T is generally consistent between specimens of the same set. Figure 5 shows that specimens from Corsica and California tend to have oblate AMS fabrics, whereas specimens from Scotland tend to have prolate AMS fabrics (Figure 5B and C). The sets of specimens, however, define slightly different symmetries: the Taiwanese vein shows a scattered fabric (Figure 6A); both Corsican and Scottish veins have plano-linear (triaxial) symmetries (Figure 6B and D); and the Californian set defines a dominantly planar fabric (Figure 6C). The three AMS scalar parameters, P' , T and K_m show no significant correlation amongst themselves (Figure 5).

The AMS directional parameters are provided as trend and plunge of K_1 , K_2 and K_3 axes (Table 1) and depicted on stereographic projections (Figure 6). The mean directional values for each locality are provided with their confidence angles in Table 1 and shown in the stereonets when $\alpha \leq 25^\circ$. The number of specimens needed per locality to define a statistically meaningful fabric depends primarily on consistency of the AMS fabric. For the Taiwan locality (Figure 6A), the AMS fabric is highly scattered because the AMS directions are not reproducible because of a low signal-to-noise ratio. This AMS fabric also appears incompatible with the orientation of the pseudotachylyte vein because its magnetic foliation (plane normal to K_3) is at high angle to the pseudotachylyte generation vein (locally oriented $146^\circ 38^\circ$ SW). For Corsica (Figure 6B), the AMS fabric is characterized by a well-defined magnetic foliation (plane normal to K_3) and a moderately well defined magnetic lineation (K_1). The K_2 axes are not well defined and therefore the Jelinek statistics (mean direction and ellipse of confidence) are not plotted for this axis. The magnetic foliation ($033^\circ, 67^\circ$ SE) plots at $\approx 21^\circ$ from the pseudotachylyte generation plane ($050^\circ, 52^\circ$ SE). The magnetic lineation ($038^\circ, 13^\circ$) plots 17° from the pseudotachylyte generation plane

(050°, 52° SE). For California (Figure 6C), the AMS fabric is characterized by very well-defined magnetic foliation and magnetic lineation. The magnetic foliation (083°, 49° N) and magnetic lineation (266°, 03°) plot at $\approx 13^\circ$ and 3° respectively from the pseudotachylyte generation plane (092°, 38° N). For Scotland (Figure 6D), the AMS fabric is also characterized by a very well-defined magnetic foliation and a well-defined lineation. The magnetic foliation (162°, 66° E) and magnetic lineation (119°, 57°) plot at $\approx 28^\circ$ and 24° respectively from the pseudotachylyte generation plane (166°, 38° E). Since the mini-cubes were cut at the same distance from the vein margin (except for Scotland), the distribution of AMS axes is deemed independent from sampling geometry. The scatter of AMS axes seems to decrease when the magnetic susceptibility increases.

5. Discussion

5.1. Mineralogic origin of the AMS

The interpretation of AMS rests on determining the mineralogic origin of magnetic susceptibility and its anisotropy (Rochette et al., 1992; Borradaile and Henry, 1997; Borradaile and Jackson, 2004). Quartz and feldspars have small negative magnetic susceptibilities, $\approx -10^{-5}$ [SI], whereas ferromagnesian silicates, such as biotite, pyroxene, or olivine, are paramagnetic and have moderate magnetic susceptibilities, $\approx 10^{-3}$ [SI], while ferromagnetic (s.l.) minerals, such as magnetite, have magnetic susceptibilities between about 1 and 3 [SI] (Dunlop and Özdemir, 2009).

Previous studies on artificial and natural pseudotachylytes show that the breakdown of iron-rich minerals during frictional melting results in coseismic nucleation of PSD magnetite (e.g., Nakamura et al., 2002; Ferré et al., 2005; 2012; 2014a; Michima et al., 2009; Chou et al., 2012).

In some cases, maghemite may also form through the breakdown of lepidocrocite (Fukuchi et al., 2005). Because magnetite has a very high magnetic susceptibility, ≈ 3 [SI], even when present in small concentration, this phase dominates the magnetic susceptibility. Pseudotachylyte veins may also host comminuted magnetite clasts derived from the host-rock. In this study, our strategy is deliberately to minimize the contribution from clasts by sampling clast-poor to clast-free (macroscopically) generation veins (Figure 1).

The pseudotachylytes from the Hoping River locality in Taiwan have a magnetic susceptibility and magnetic hysteresis properties dominated by paramagnetic minerals such as biotite, muscovite and phengite (Figure 4A). The ferromagnetic contribution to magnetic susceptibility (0.9%) is attributed to fine-grained hematite, ≈ 200 nm in length, as this mineral has been previously documented by Chu et al. (2012) (their Figure 9) in these pseudotachylytes. Hematite of this size falls in the superparamagnetic to single domain range (Dunlop and Özdemir, 2009) and is not sufficiently abundant to contribute appreciably to magnetic susceptibility. The poor reproducibility of AMS directions and scalar parameters, even during repeated measurement of the same specimen, indicates a low signal-to-noise ratio and result in erratic directional data. This instrumental artifact also explains the P' values that are anomalously high for paramagnetic rocks.

In contrast, the pseudotachylytes from Corsica, California, and Scotland display magnetic hysteresis properties indicating that PSD magnetite is the dominant magnetic phase (Table 1 and Figure 4). This is confirmed by our back-scattered electron microscopy data (Figures 2B, C and D) and previous studies on Californian pseudotachylytes (Ferré et al., 2005, 2012, and 2014a). The ferromagnetic contributions to magnetic susceptibility ($K_{\text{ferro}}/K_{\text{m}}$) in these pseudotachylytes from Corsica, California, and Scotland are respectively 73.5%, 96.6% and 98.2% (Figure 4B, C,

and D). Despite the fact that magnetite grain size tends to be larger than 1 μm , the contribution of single-domain (SD) magnetite cannot be entirely ruled out, particularly in the case of Scotland ($M_r/M_s \sim 0.34$, $H_c \sim 43$ mT, $H_{cr}/H_c \sim 2.3$). Yet, the AMS fabrics are not inverted as shown by the high angle of K_1 with respect to the pseudotachylyte plane. The even distribution, small size and relative roundness of magnetite grains is strikingly similar to that of artificially produced pseudotachylytes where magnetite is neoformed (e.g., Nakamura et al. 2002). Rock fabrics carried by PSD magnetite arise from the 3-D magnetostatic anisotropy, i.e., shape-preferred orientation (SPO) of magnetite grains (e.g., Arbaret et al., 2013).

5.2. AMS acquisition mechanism

Deformation within a pseudotachylyte occurs across a wide range of conditions beginning with high differential stress (σ_D) / low temperature at the onset of seismic rupture, followed by low σ_D / very high temperature during frictional melting, and finishing with low σ_D / moderate temperature at the end of slip. Frictional melt temperature has been estimated to 1300-1400°C for Corsica (Andersen and Austrheim, 2006), 1600-1700°C for California (Ferré et al., 2012), and 1100-1200°C for Scotland (Sibson, 1975). Nakamura et al. (2002) showed that magnetite typically forms in pseudotachylytes through the breakdown of mafic silicates. The elongate shape of magnetite grains in our samples (Figure 2) also supports this interpretation. These shapes most likely result from plastic deformation at temperatures close to magnetite melting ($\approx 1538^\circ\text{C}$), possibly by dislocation creep as shown by Till et al. (2012) and Till and Moskowitz (2013; 2014). As discussed by Mamtani et al. (2011), however, dislocation creep is unlikely to occur at temperatures similar to that of granitic magmas ($\approx 600^\circ\text{C}$). Once elongate magnetite grains have formed, their SPO may also be strengthened by grain rotation in a viscous medium until seismic slip has ceased, in a manner similar to that described by Mamtani et al. (2011) for

SPO acquisition in granitic magma. Frictional melting requires a significant normal stress during seismic slip to occur (e.g., Di Toro et al., 2006), hence the coseismic strain regime is most likely transpressional. Therefore, although the vorticity of coseismic deformation in pseudotachylyte generation veins is generally poorly understood, coseismic shear should be expected to have a significant pure shear component.

5.3. AMS-based kinematic analysis of coseismic slip

The Hoping River pseudotachylytes of Taiwan do not yield an accurate AMS fabric for two reasons: i) these rocks do not host magnetite formed during seismic slip and ii) the mini-cubes are clearly too small to reliably integrate the fabric of a large enough number of paramagnetic grains. The mini-cubes of this study are about 20 times smaller volumetrically than the small cubes used by Ferré et al. (2015).

For the pseudotachylytes from Corsica, California and Scotland, we have demonstrated that the AMS arises from magnetite SPO, and thus interpret the AMS principal directions to represent magnetite finite strain axes. Plastic shear experiments show that the relationship between magnetite strain and bulk strain depends on deformation partitioning between magnetite grains and the surrounding matrix (Till et al., 2012; Till and Moskowitz, 2013, 2014). Because frictional melts have low viscosities, on the order of 10^1 to 10^4 Pa.s (e.g., Spray, 1993), mechanical coupling between magnetite grains and their matrix should be small, at least before complete melt solidification. Under these conditions, the AMS of pseudotachylytes should record viscous flow, like in any other magma body (e.g., Knight and Walker, 1988; Ernst and Baragar, 1992; Grégoire et al., 1995). Complications may arise, however, due to coseismic strain vorticity, i.e., the partition between simple shear and pure shear. In the cases of Corsica, California and Scotland, the angle between the magnetic foliation (plane normal to K_3) defined

by the AMS tensor and the pseudotachylyte plane is less than 28° . This geometric relationship suggests that the magnetic fabric records viscous drag along the vein margin as an imbrication fabric similar to that described elsewhere by Knight and Walker (1988) in basaltic dikes and by Ferré et al. (2015) in other pseudotachylytes. It should be noted, however that in the dike case, imbrication shows opposite sense of shear on either side of the dike, whereas in a pseudotachylyte, since the intrusion walls are in shear motion, the imbrication would be expected to reflect the same shear sense. In a perfect simple shear regime, the magnetic lineation (K_1) should be perpendicular to the intersection of the magnetic foliation and the pseudotachylyte margin, at a decreasing angle as shear strain increases. Following Geoffroy et al. (2002), who observed a similar geometry in mafic dikes from Greenland, we propose using the direction perpendicular to the intersection between magnetic foliation and the vein margin, along the pseudotachylyte plane, as the most reliable coseismic slip indicator. We attribute the skewness of K_1 to the pure shear component of coseismic viscous flow. The imbrication between the AMS fabric and the pseudotachylyte plane can also be used to determine seismic slip sense (Ferré et al., 2015). The coseismic slip directions and slip senses obtained using this method are reported with corresponding focal mechanisms in Figure 7. The sense of slip deduced from the mini-AMS data is identical that observed macroscopically for Corsica (Magott et al., 2016), California (Rowe et al., 2012) and Scotland (Campbell et al., 2014).

6. Conclusions and recommendation for further studies

The mini-AMS method requires pseudotachylyte specimens to be precision-cut with a stainless steel diamond wire saw, a process that can be relatively time intensive. The measurement in a classic Kappabridge instrument also requires the use of low magnetic

susceptibility, custom-made sample holders. The signal for pseudotachylytes dominated by paramagnetic mineral phases is insufficient to be reliably measured. This situation is, however, rather exceptional because frictional melting generally results in magnetite formation. It would be nonetheless advantageous to test the magnetic assemblage of pseudotachylytes using magnetic hysteresis experiments before performing a mini-AMS study. The mini-cubes need to be free of post-deformational alteration in order for the AMS to represent coseismic deformation. In addition, the mini-cubes should also host as little as possible host-rock clasts because the latter may complicate the pattern of viscous flow. The area of the pseudotachylyte vein investigated has to be relatively far away from injection veins since proximity to the latter may disturb the flow pattern of the frictional melt. At least 10 mini-cube specimens from a single vein are needed to obtain acceptable and interpretable directional statistics.

The consistency of AMS fabrics is remarkable considering the small size of the specimens, yet this consistency makes sense when one considers the extremely fine-grained nature of the material and the large shear strain involved in seismic slip on a plane. Despite all the limitations listed above, the mini-AMS method successfully resolves the viscous flow fabric of fault pseudotachylytes at an unprecedentedly high resolution ($\sim 43 \times 10^{-3} \text{ cm}^3$). This new method allows the determination of coseismic slip direction and slip sense. These results can then be used to quantify the focal mechanisms of ancient earthquakes. Further studies will most likely attempt to decipher the details of frictional melt flow patterns taking advantage of the high resolution of this new method.

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Figure Captions

Figure 1. Location with GPS coordinates of the four pseudotachylyte localities and samples investigated in Taiwan (A), Corsica (B), California (C), and Scotland (D). The white dash lines on the sample show where mini-cube specimens were taken from. The mini-cubes were cut from 3 successive and parallel rows in the direction perpendicular to the plane of the photograph, hence all mini-cubes are at the same distance from the vein margin, except for Scotland where the 7 mini-cubes were stacked as shown.

Figure 2. Back-scattered electron microscopy of the four pseudotachylytes investigated. A. Taiwan – This pseudotachylyte vein did not show any magnetite grains. The plagioclase snowflake-shaped microcrystallites attest to frictional-melt rapid quenching. B. Corsica – Small, scattered elongate grains of magnetite occur throughout the pseudotachylyte generation vein, here near the host-rock contact. The dash line marks the sharp boundary between host-rock and pseudotachylyte. C. California - Small, oblong grains of magnetite occur throughout the pseudotachylyte generation vein. D. Scotland - Small, elongated grains

of magnetite occur throughout the pseudotachylyte generation vein, here near the host-rock contact. We note that the largest magnetite grain (the largest white object with an arrow pointing at it) is most likely a clast inherited from the host-rock. The thicker dash line marks the sharp boundary between host-rock and pseudotachylyte; the thinner dash line marks the slight compositional banding.

Figure 3. **A.** Frame and dimension sketch of the mini-AMS sample holders. The numbers refer to millimeters [with inches in between brackets]. Ø4.00 refers to the toll diameter in millimeters. The cubic sample sits exactly at the center of this structure. Both upper and lower parts of the sample holder are symmetrical. **B.** Acrylic mini-AMS sample holder with Agico orientation scheme reported as arrows. **C.** Assembled mini-AMS sample holder. The mini-cube is stabilized in the sample holder using vacuum grease Dow Corning ©112. The arrows correspond to the Agico orienting scheme for the Kappabridge instrument.

Figure 4. Magnetic hysteresis properties of representative mini-cube specimens. The specimens from Taiwan display a fundamentally paramagnetic behavior (A). The percentage of paramagnetic susceptibility in the specimens from Corsica (B), California (C) and Scotland (D) is low to very low.

Figure 5. Anisotropy of magnetic susceptibility (AMS) scalar parameters. **A.** Corrected degree of magnetic anisotropy (P') vs magnetic susceptibility (K_m); **B.** Shape parameter (T) vs

magnetic susceptibility (K_m); and C. Shape parameter (T) vs corrected degree of magnetic anisotropy (P'). In all diagrams, specimens from the four localities plot in distinct domains.

Figure 6. Anisotropy of magnetic susceptibility (AMS) directional data. A. Taiwan – The magnetic fabric at this locality is very dispersed and unreliable. B. Corsica – The magnetic fabric is relatively clustered with the exception of axis K_2 . C. California – The magnetic fabric at this locality is very consistent. D. Scotland - The magnetic fabric at this locality is very consistent. All stereonetts plotted in geographic coordinate system, equal area, lower hemisphere, using Stereonet 6.3. (Allmendinger, 2005). Jelinek's statistics calculated using Anisoft 4.2 software and plotted for for $\alpha > 25^\circ$. K_1 (filled squares), K_2 (grey triangles), and K_3 (circles) data and means (larger symbols) are also reported in Table 1. The α confidence angles in Table 1 provide a statistical measure of dispersion for all AMS axes.

Figure 7. Earthquake focal mechanism diagrams showing the interpretation of the magnetic fabric as a coseismic slip direction and slip sense criteria. The seismic slip direction is deduced from the pseudotachylyte plane and the magnetic foliation (plane to K_3). The oblique magnetic lineation (K_1) results from strain vorticity.

Table Captions

Table 1. AMS scalar and directional data of four pseudotachylyte localities.

*Figure 1

SG-D-16-00064R_Fig1

*Figure 2

SG-D-16-00064R_Fig2

***Figure 3**

SG-D-16-00064R_Fig3

*Figure 4

SG-D-16-00064R_Fig4

*Figure 5

*Figure 6

*Figure 7

SG-D-16-00064R_Fig7

Table 1

Table 1. AMS scalar and directional data of four fault pseudotachylite localities

Localities	Spec Name	K_m	P'	T	d_{K1}	i_{K1}	d_{K2}	i_{K2}	d_{K3}	i_{K3}
$10^{-6} [SI]$										
Taiwan	HP010P-223A	546	1.344	0.213	150	87	314	3	44	1
	HP010P-352C	462	1.283	-0.378	341	60	180	28	86	8
	HP010P-463C	368	1.199	0.690	1	62	145	23	241	15
	HP010P-473B	372	1.260	-0.505	87	39	234	46	343	17
	HP010P-622A	508	1.261	-0.037	105	44	229	30	339	31
	HP010P-623A	640	1.272	-0.344	90	41	290	47	189	10
	HP010P-712B	550	1.203	-0.640	114	24	235	50	9	30
	HP010P-733A	627	1.259	0.064	328	75	229	2	139	15
	HP010P-742C	540	1.246	-0.018	73	59	230	30	326	10
	HP010P-752F	610	1.286	-0.251	336	55	144	35	238	6
	HP010P-753B	611	1.167	-0.262	104	54	259	34	357	12
	HP010P-763E	644	1.250	-0.307	321	60	172	27	75	13
	$\bar{\alpha}$ conf. angle ($^{\circ}$)				<u>43</u>	<u>12</u>	<u>51</u>	<u>30</u>	<u>46</u>	<u>12</u>
Corsica	tensorial mean		1.134	-0.279	76	75	298	12	206	10
	scalar mean	540	1.253	-0.148						
	stdev	97	0.046	0.359						
	COR006A01	701	1.042	0.475	196	68	37	21	304	7
	COR006A02	603	1.046	-0.349	42	37	217	53	310	2
	COR006A03	656	1.049	-0.062	69	51	191	23	295	30
	COR006A04	652	1.071	0.247	24	8	137	70	291	18
	COR006A05	851	1.041	0.069	228	21	40	69	137	3
	COR006A06	789	1.064	-0.089	35	14	134	31	283	55
	COR006A07	757	1.048	0.105	215	12	75	75	307	10
	COR006A08	919	1.124	0.663	68	30	176	28	301	47
	COR006A09	846	1.040	0.678	46	7	173	79	315	9
	COR006A10	787	1.034	0.319	30	21	217	69	121	2
	COR006A11	748	1.051	-0.334	26	7	230	83	116	3
	COR006A12	831	1.106	0.490	41	10	149	61	306	27
	$\bar{\alpha}$ conf. angle ($^{\circ}$)				21	8	26	18	24	8
	tensorial mean		1.049	0.261	38	13	157	63	303	23
	scalar mean	762	1.060	0.184						
	stdev	95	0.028	0.354						
California	SR108B1	70300	1.041	0.623	81	8	342	52	177	38
	SR108B2	72730	1.051	0.484	76	8	337	47	174	42
	SR108B3	82360	1.048	0.205	77	3	344	46	170	44
	SR108B4	64420	1.043	0.414	261	4	356	51	168	39
	SR108B5	63040	1.044	0.394	271	12	12	41	168	46
	SR108B6	62880	1.035	0.323	80	6	344	45	176	44
	SR108B6	73410	1.033	0.266	72	10	333	43	172	46
	SR108B8	83290	1.035	0.312	273	16	26	54	173	32
	SR108B9	86850	1.042	0.313	83	3	351	39	177	51
	SR108B10	69480	1.035	0.248	274	16	21	48	171	38
	SR108B11	73130	1.034	0.348	94	14	350	44	197	43
	SR108B12	75610	1.049	0.249	277	23	34	46	170	35
	SR108B13	85530	1.046	0.282	272	9	13	49	175	40
	$\bar{\alpha}$ conf. angle ($^{\circ}$)				14	4	14	4	5	4
	tensorial mean		1.039	0.387	266	3	0	49	173	41
	scalar mean	74079	1.041	0.343						
	stdev	8312	0.006	0.114						
Scotland	SU2A	43030	1.111	-0.571	117	58	356	18	257	26
	SU2B	56640	1.110	-0.282	118	56	357	19	257	27
	SU2C	72970	1.129	-0.554	123	59	20	8	285	30
	SU2D	38240	1.099	-0.311	128	52	339	34	239	15
	SU2E	52190	1.090	-0.558	107	59	342	19	243	23
	SU2F	56410	1.086	-0.663	112	57	337	25	237	21
	SU2G	46740	1.076	-0.497	126	53	348	29	246	21
	$\bar{\alpha}$ conf. angle ($^{\circ}$)				4	2	16	4	16	2
	tensorial mean		1.098	-0.534	119	57	352	21	252	24
	scalar mean	52317	1.100	-0.491						
	stdev	11384	0.018	0.142						