



Pressure, temperature and lithological dependence of seismic and magnetic susceptibility anisotropy in amphibolites and gneisses from the central Scandinavian Caledonides

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ABSTRACT

As a petrofabric indicator, anisotropy of magnetic susceptibility (AMS) can potentially be used to infer seismic properties of rocks, and in particular seismic anisotropy. To evaluate the link between AMS and seismic anisotropy we present laboratory measurements of elastic wave velocities and anisotropy of magnetic susceptibility (AMS) for eight samples from the deep drilling investigation forming a part of the Collisional Orogeny in the Scandinavian Caledonides (COSC) project. The samples consist of a representative suite of mid crustal, deformed rock types, namely felsic and biotite-rich gneisses, and amphibolites (mafic gneisses). Compressional (P) and shear (S) waves were measured at confining pressures from room pressure to 600 MPa and temperature from room condition to 600 °C. Seismic anisotropy changes with increasing temperature and pressure, where the effect of pressure is more significant than temperature. Increasing pressure, considering the range of samples, results in an increase in mean wave speed values from 4.52 to 7.86 km/s for P waves and from 2.75 to 4.09 km/s for S waves. Biotite gneiss and amphibolite exhibit the highest anisotropy with P wave anisotropy (AVp) in the ranges of ~9% to ~20%, and maximum S-wave anisotropy exceeds 10%. In contrast, felsic gneisses are significantly less anisotropic, with AVp of <7% and AVs of <6%. Up to 20% anisotropy may be generated by microcracks at 600 MPa and 600 °C, which is likely originating from thermal expansion of anisotropic minerals. An agreement is found between AMS and seismic anisotropy, although this is only a case if mean magnetic susceptibility (k_{mean}) ranges between $\sim 1 \times 10^{-5}$ to $\sim 1 \times 10^{-3}$ [SI]. Such k_{mean} values are common in rocks dominated by paramagnetic matrix minerals. Based on our results we propose that such samples are the most likely to be useful for the prediction of seismic anisotropy based on their AMS.

1. Introduction

Laboratory measurements of seismic anisotropy provide petrophysical signatures that allow insight into the structure of the middle and lower crust. A considerable database of seismic properties of rocks exists today for samples with known composition and structure (e.g., Birch, 1960, 1961; Christensen, 1965, 1971, 1974, 1979; 1996; Fountain, 1976; Kern, 1982; Kern et al., 1991, 1997; Burke and Fountain, 1990; Burlini and Fountain, 1993; Ji et al., 2002). Over the last two to three decades the availability of geophysical observations of the seismic structure and anisotropy in different crustal settings (e.g., Hirn et al.,

1987; Mooney and Meissner, 1992; Christensen and Mooney, 1995; Shapiro et al., 2004; Schulte-Pelkum et al., 2005) has resulted in an increased interest in anisotropy sources in the middle and lower crust. Barruol and Kern (1996) experimentally inferred the effects of crystallographic preferred orientation on seismic anisotropy and shear wave birefringence for rocks that originated from the Ivrea-Verbano zone, which is a type section for the lower continental crust. Lloyd et al. (2009, 2011a, 2011b) and Tatham et al. (2008) argue that observed seismic anisotropy in the middle and lower crust can be explained mainly by CPO of mica (e.g., biotite and muscovite) and amphibole, based on microstructure-based calculations. These two groups of minerals are

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among the most anisotropic in the crustal mineral inventory and tend to develop strong CPO's in ductile deformed rocks (Fountain, 1976; Burlini and Fountain, 1993; Cholach and Schmitt, 2006; Ko and Jung, 2015). Almqvist and Mainprice (2017) provide a comprehensive overview of common minerals in the continental crust and how their microstructures affect seismic anisotropy.

A ductile deforming crust will result in the formation of a petrofabric, which is reflected in the expression of physical properties. It is well established that elements of petrofabric, such as crystallographic preferred orientation and shape preferred orientation, and seismic anisotropy are strongly correlated (e.g., Mainprice and Nicolas, 1989; Siegesmund et al. 1989; Mainprice, 1990; Burlini and Fountain, 1993; Barruol and Kern, 1996; Mainprice, 2000; Lloyd and Kendall, 2005; Mainprice, 2007; Tatham et al., 2008; Lloyd et al., 2009; Lloyd et al., 2011a; Lloyd et al., 2011b; Ji et al., 2013; 2014; 2015; Ko and Jung, 2015; Mainprice, 2015; Cyprych et al., 2017; Almqvist and Mainprice, 2017). Most of these are calculations based on quantitative constraints on crystallographic orientation data utilizing universal stage measurements, X-ray diffraction, neutron diffraction and electron backscatter diffraction (EBSD) analyses (e.g., Barruol and Mainprice, 1993; Barruol and Kern, 1996; Crosson and Lin, 1971; Valcke et al., 2006; Wenk et al., 2012). In contrast to the aforementioned methods, anisotropy of magnetic susceptibility (AMS) provides a rapid and fairly inexpensive technique (Borradaile and Henry, 1997; Borradaile and Jackson, 2010). AMS is related to the sum of the magnetic contributions from all minerals in a sample, although it is more strongly influenced by ferromagnetic minerals, such as magnetite and pyrrhotite (e.g., Hrouda et al., 1971; Henry, 1983; Rochette et al., 1992; Lagroix and Borradaile, 2000). For magnetic fabric studies, the motion in the flow plane is expected to be parallel to the maximum principal axis (k_{max}) because the motion aligns minerals within the magnetic foliation plane (k_{max} - k_{int}) and parallel to the flow direction. This suggests that k_{max} is parallel to the mineral lineation and the minimum principal axis (k_{min}) aligns with the pole to foliation. This information provides access to specifics of magmatic fabrics in igneous rocks (e.g., Khan, 1962; Knight and Walker, 1988; Ellwood and Whitney, 1980; Bazargan et al., 2019; Vachon et al., 2021) and their tectonic settings and deformation conditions (e.g., Graham, 1966; Borradaile and Henry, 1997; Borradaile and Jackson, 2010; Mamtani et al., 2011). Furthermore, the pressure, temperature and deformation conditions may lead to new mineral growth and therefore modify the magnetic fabric (Borradaile and Jackson, 2004). The fast measurement procedure of AMS allows the evaluation of many samples, enabling a statistical approach to study the petrofabric.

Our study uses samples retrieved through the Collisional Orogeny in the Scandinavian Caledonides (COSC) continental drilling project (Lorenz et al., 2015, 2021). Reflection and applied seismological studies were essential to defining the subsurface of the first of the investigated drilling sites (COSC-1) (Juhlin et al., 2016; Hedin et al., 2012; Hedin et al., 2014; Hedin et al., 2016; Elger et al., 2021). Simon et al. (2017, 2019) presented seismic tomography and anisotropy results from surface and in the borehole to illustrate and show the anisotropy effect with case scenarios of different Thomsen parameters (Thomsen, 1986) to better explain and present the seismic wave data collected at the surface and measurements in the borehole. Additionally, several laboratory petrophysical studies has investigated the seismic and magnetic properties of the COSC-1 drill core (Wenning et al., 2016; Merz et al., 2019; Kästner et al., 2020; 2021; Almqvist et al., 2021). Merz et al. (2019) studied petrofabrics using AMS in the lower part of the borehole (1910 to 2450 m) and showed the transition of magnetic mineral and periodic appearances of mylonites in mica schists and amphibole at different depth. Notably, they demonstrated that metamorphic and deformation processes related to emplaced on the Lower Seve Nappe generated new ferromagnetic minerals, including magnetite, pyrrhotite, ilmenite and hematite.

Here, we provide new data on laboratory seismic properties and AMS measurements for the samples. AMS data is compared to seismic

anisotropy to evaluate under which conditions it can be used as a proxy for seismic anisotropy.

2. Geological setting and sample material

The COSC-1 drill site is located in central-western Sweden (Jämtland County), situated in the central Scandinavian Caledonides (Fig. 1; Lorenz et al., 2015). Tectonostratigraphically, the drilled rock unit belongs to the Lower Seve Nappe, which is part of the Middle Allochthon (Fig. 1; Gee et al., 2010; Gee et al., 2020). The COSC-1 borehole is dominated by interlaying of gneisses and amphibolites, occurring on cm- to 10's m-scale, from the surface to ~2200 m depth. The protoliths for the gneisses are sedimentary rocks formed from passive margin sediments of the Iapetus Ocean whereas the amphibolites are interpreted to originate from dyke swarms that intruded the sediments in the passive margin (Bergman et al., 2012; Gee et al., 2010; Gee, 1978). Such interlayering of metasedimentary gneisses and amphibolites may result in excellent seismic reflections, as shown for example in the Grenville Province, Canada Ji et al. (1993; 1997). During orogeny, the Lower Seve Nappe rocks were incorporated in the subduction channel during closure of the Iapetus Ocean and experienced eclogite peak metamorphic conditions (Giuntoli et al., 2018, 2020; Holmberg, 2017). During subsequent emplacement into the crust, the unit experienced metamorphic re-equilibration at amphibolite to greenschist facies conditions (Giuntoli et al., 2020).

Accordingly, the petrophysical character of these rocks is relevant to the seismic properties and anisotropy of tectonically deformed middle to lower continental crust. Even though there are outcrops of equivalent rocks available, such rocks are expected to have been modified through weathering and erosion resulting in changes in mineralogy, open and mineral-filled fractures and micro-cracks. These latter features may significantly influence physical properties and thus the pristine drill core samples hold an advantage over outcrop derived samples.

2.1. Sample selection and preparation

Eight representative samples were selected from the COSC-1 drill core, originating from depths ranging from 538 m to 1243 m (Fig. 1). All samples are metamorphic and exhibit a distinct gneissic foliation characterized by alternating layers with different mineral compositions as well as mineral aggregate lineations (Piazolo and Passchier, 2002; Paaschier and Trouw, 2005). Samples were chosen to represent the compositional and mineralogical variation (Table 1).

From the drill core, cubic samples (43x43x43 mm) were prepared for ultrasonic measurements. Sample cubes were cut according to the fabric elements of the rocks (i.e., lineation, foliation), where the sample reference frame is defined as follows: X – parallel to the mineral lineation in the foliation plane; Y – perpendicular to the mineral lineation in the foliation plane; Z – normal to the foliation plane. In this paper, we refer to all samples based on their core related number and depth. Thin sections were prepared from the sample cube subsequent to the ultrasonic experiment and cut perpendicular to the foliation and parallel to lineation (i.e., XZ-plane in the sample reference system).

2.2. Sample description

Samples are divided into rock types based on their general mineral composition, into felsic gneisses (samples 248, 309, 362), intermediate, bt-rich gneiss, i.e., biotite gneiss (sample 316) and mafic gneisses (193, 274, 340). We note that the general terms felsic, intermediate to mafic are used in a purely descriptive manner indicating dominant composition in terms of mineral abundance (e.g., felsic: dominated by quartz and feldspar). No genetic connotation is intended, i.e., magmatic or metamorphic origin. Table 1 provides a mineral modal composition for all samples.

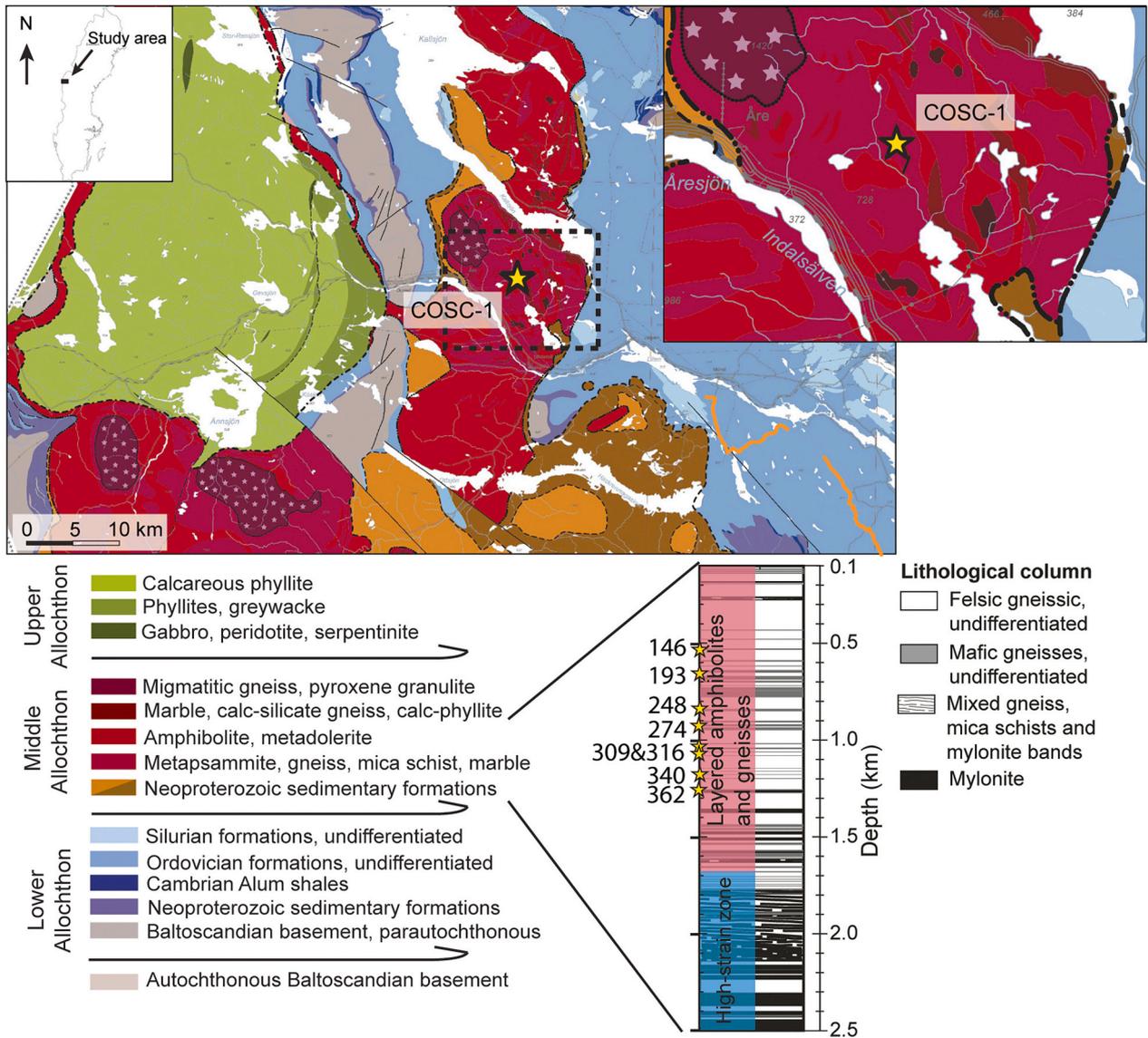


Fig. 1. COSC-1 borehole location (star) shown on a regional tectonostratigraphic map of the central Scandinavian Caledonides (based on [Gee and Sturt, 1985](#)). A simplified lithological column is provided in the lower half of the figure, which shows the locations of the samples investigated in this study and the position of the COSC-1 borehole in the generalized tectonostratigraphy of the central Scandinavian Caledonides (modified from [Hedin et al., 2016](#)).

Table 1
Modal mineral composition (%) of samples in the study.

Lithological group	Rock type	Sample no.	Qz	Pl	Kfs	Cal	Amph (hbl)	Cpx (Di)	Bi	Ms	Other (Ep, Ap, opq)
Felsic gneiss	qtz-cal gneiss	248	50	≤5	≤5	35				≤5	
	qtz-feldspar gneiss	309	50	35	15						
	qtz-feldspar gneiss	362	25	50–60		10–15		10			
Intermediate gneiss	Bt gneiss	316	30	25			15	25		≤5	
Mafic gneiss	Bt bearing amphibolite	340		25			60		15		
	Amphibolite	274		25	5		60		3–5		5
	Amphibolite	193		35			65				
Layered felsic-mafic gneiss	qtz feldspatic gneiss layer	146	60	25	10	3–5		≤2			

Mineral abbreviations from [Whitney and Evans \(2010\)](#): Qz - quartz; Pl - plagioclase; Kfs - potassium feldspar; hbl - hornblende; Di - diopside; Bi - biotite; Mu - muscovite; Ep - epidote; Ap - apatite; opq - opaque.

2.3. Felsic gneiss (FG) samples

Sample 248 (839.2 m) is dominated by medium to coarse-grained quartz and calcite, with small amounts of feldspar and muscovite, and accessory opaque phases (Table 1). Calcite is seen as large porphyroclasts (up to 1 mm) surrounded by a matrix of medium-grained (down

to 100 μm) quartz and calcite, suggesting some degree of dynamic recrystallization. The elongation of some calcite grains defines a weak foliation barely noticeable in hand specimens but apparent in thin sections (Fig. 2a). Quartz and opaque minerals do not show significant shape preferred orientation (SPO).

Sample 309 (1021.9 m) is dominated by quartz and plagioclase

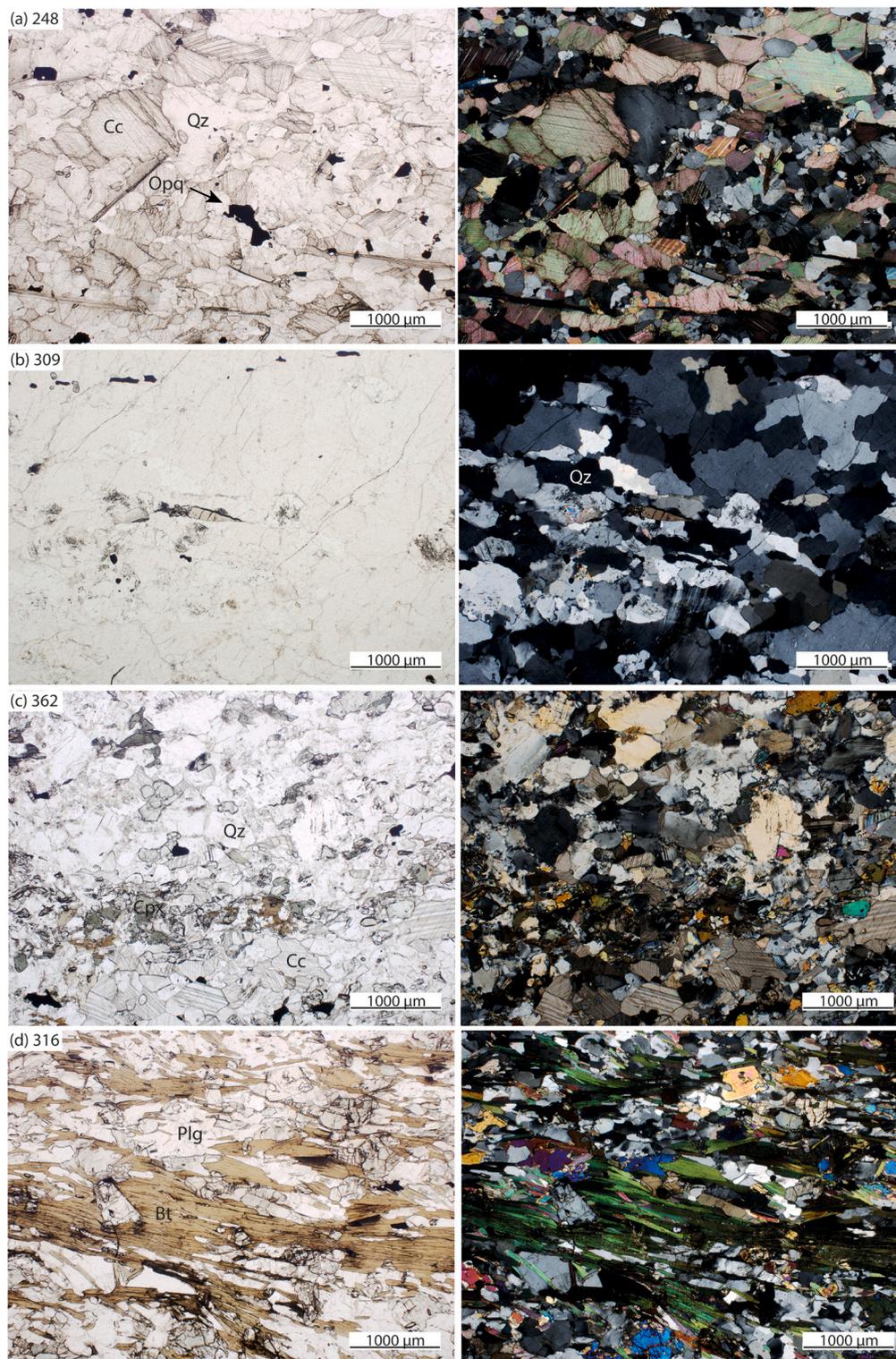


Fig. 2. Photomicrographs of the (a–c) felsic gneisses (samples 248, 309, 362) and the (d) intermediate biotite-gneiss (sample 316), taken in plane-polarized (left side) and cross-polarized (right side) light. Mineral abbreviations follow [Whitney and Evans \(2010\)](#); Amp: amphibole; Plg: plagioclase feldspar; Cpx: clinopyroxene; Qz: quartz; Opq: opaque (generally iron-oxide or iron-sulphide).

feldspar; smaller amounts of K-feldspar, muscovite and opaque phases are also observed ([Table 1](#)). Quartz grains are medium to coarse-grained (a few hundred μm to $>1\text{ mm}$), while plagioclase is dominantly medium-grained (a few hundred μm). Grain boundaries of quartz are highly irregular and lobate and some grains exhibit undulose extinction, which suggests significant grain boundary migration, dislocation movement

and subgrain formation as main deformation mechanisms. Even though individual grains show elongation, the foliation is defined by alternating bands of fine- to medium-grained plagioclase-quartz-muscovite-opaque phases and bands of $>90\%$ coarse-grained quartz identified in hand specimen and thin section ([Fig. 2b](#)).

Sample 362 (1243.1 m) is mainly composed of plagioclase, K-

feldspar and quartz, with smaller amounts of calcite, clinopyroxene and accessory opaque phases (Table 1). A clear foliation is observed in both hand specimens and thin-section. The foliation is defined by the alternation of thick bands of fine to coarse-grained feldspar with only minor clinopyroxene and bands of fine-grained mixtures of feldspar and clinopyroxene (Fig. 2c). In the thick bands, feldspar shows a clear bimodal grain size distribution, with up to 0.8 mm porphyroclasts surrounded by

100 to 300 μm size feldspar grains. Grains show some degree of undulose extinction and elongation (SPO) of feldspar porphyroclasts is noticeable. In fine-grained bands, the clinopyroxene shows a moderate SPO with elongation subparallel to band boundaries (Fig. 3c).

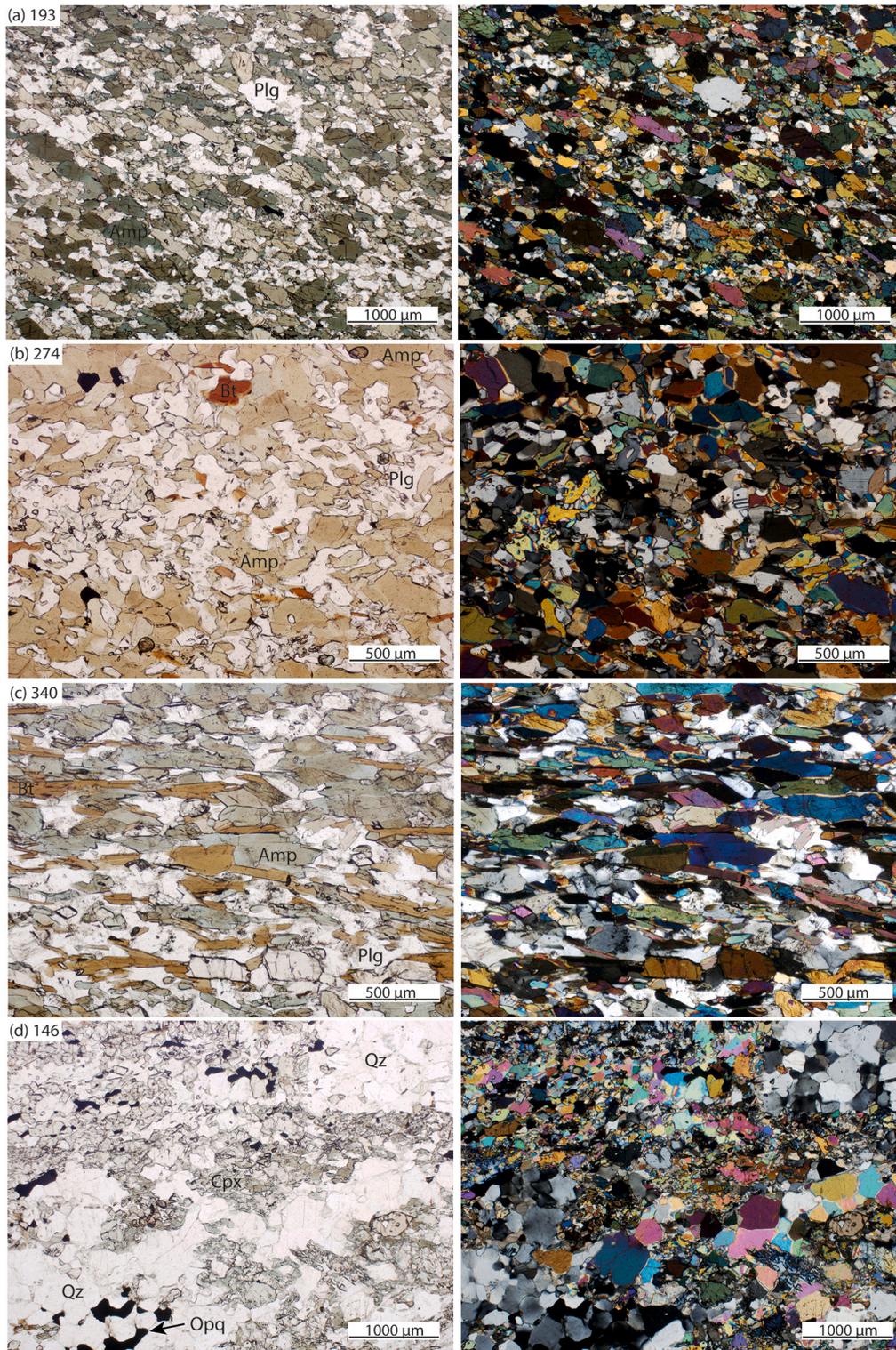


Fig. 3. Photomicrographs of (a–c) mafic gneisses (samples 193, 274, 340) and (d) layered felsic-mafic gneiss (sample 146), taken in plane-polarized (left side) and cross-polarized (right side) light. Mineral identification follow Whitney and Evans (2010); Amp: amphibole; Bt: biotite; Plg: plagioclase feldspar; Qz: quartz.

2.4. Intermediate biotite-gneiss (IBG) sample

Sample 316 (1080.0 m) is dominated by plagioclase feldspar, quartz and biotite in nearly equal parts (Table 1). A clear foliation is apparent, defined both by the preferred orientation of elongate biotite grains and planar bands dominated by biotite and bands with equal amounts of medium-grained biotite and feldspar. In these bands, biotite and feldspar show a clear SPO. However, these are less pronounced than the strong SPO seen in the biotite dominated layers (Fig. 2d); grain sizes are medium to large in biotite-dominated layers (~100 μm to >1 mm). Basal planes of biotite can be identified and are shown to be parallel to the foliation plane (arrows; Fig. 2d).

2.5. Mafic gneiss (MG) samples

Sample 193 (678.2 m) is dominated by equigranular, medium-grained (50 μm–400 μm) amphibole and plagioclase with minor clinopyroxene (Table 1). Amphibole exhibit a clear SPO of tabular shaped grains whose orientation defines the foliation (3a); plagioclase also shows an SPO similar to that of amphibole. Foliation planes are evenly spaced and identifiable in the hand specimen.

Sample 274 (916.9 m) is composed of ~50% amphibole, ~40% plagioclase, as well as minor biotite, K-feldspar and accessory opaque phases (Table 1). Grain size ranges for amphibole from ~300 μm to >1 mm, while the average grain size for feldspars is smaller than for

amphibole. Grains are strongly interlocking and there is only a weak SPO of amphibole discernible in areas dominated by amphibole (Fig. 3b).

Sample 340 (1176.3 m) consists mainly of amphibole and feldspar and up to 15% biotite. Grain size is largely uniform and ranges from ~500 μm to >1 mm. Undulose extinction is seen in all three phases and all phases show a strong SPO, which defines the foliation (Fig. 3c). The SPO is most pronounced for biotite, while feldspar and amphibole show similar SPO strength. Basal planes of biotite can be identified and are shown to be parallel to the foliation plane. Foliation planes are clearly defined and tightly spaced.

2.6. Layered felsic-mafic gneiss (LG)

Sample 146 (538.6 m) is strongly banded, where banding is defined by granoblastic layers of feldspar, quartz and minor opaques (felsic layer) and layers of plagioclase-clinopyroxene gneiss (mafic layer). These layers are planar and well defined in hand specimen, ranging from 0.5 to 3 cm in thickness. The quartz-feldspar-rich layers are generally coarse-grained, although in some cases fine-grained grains surround coarser grains. The mafic layer is generally finer-grained than the quartz-feldspar layer, with a clear bimodal grain size distribution of the plagioclase and clinopyroxene. Besides the layering, there is little evidence for SPO among most of the phases (Fig. 4d) except for elongate grains of interstitial opaque phases present in both types of layers.

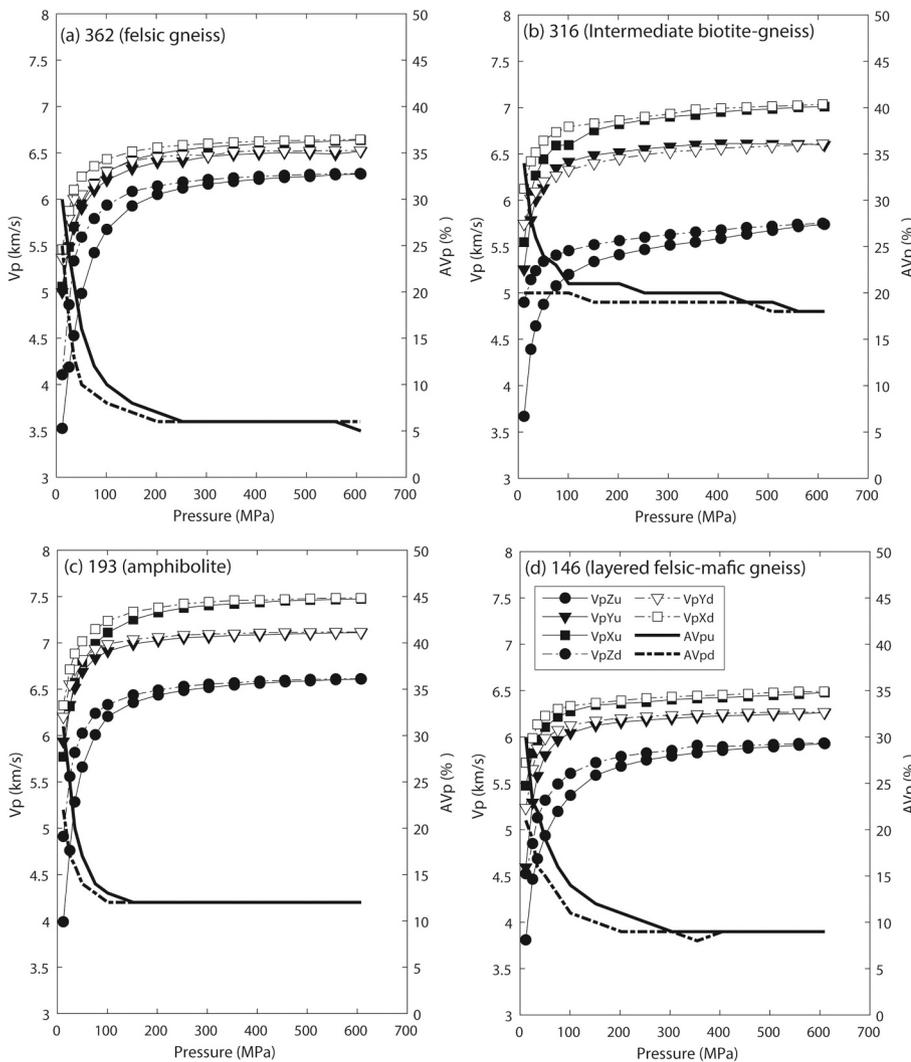


Fig. 4. P-wave velocity (Vp; left y-axis) and anisotropy of Vp (AVp, right y-axis) as a function of confining pressure for four selected samples investigated in this study, (a) felsic gneiss (362), (b) intermediate biotite-gneiss (316), (c) amphibolite (mafic gneiss; 193) and (d) layered felsic-mafic gneiss (146). The filled circles show Vp during pressurization and the open circles represent Vp during depressurization. Letters u and d in the legend indicate measurements performed during pressurization and depressurization, respectively.

Nevertheless, in the mafic band, the cm-scale banding is mirrored by the elongation of clusters of feldspar and/or clusters of clinopyroxene.

3. Methodology

3.1. Ultrasonic velocity measurements

Experiments were performed with a multi-anvil apparatus at the Institute for Geosciences at Kiel University (Germany), allowing for measurements of compressional (V_p) and orthogonally polarized shear wave velocities (V_{s1} , V_{s2}). Ultrasonic velocities (P- and S-wave) were experimentally measured using the ultrasonic pulse transmission technique (Birch, 1960; Kern and Wenk, 1990). The shear-wave transducers were oriented such that the predominant S_1 and S_2 particle motions were either parallel or perpendicular to foliation and lineation (Kern et al., 1997; Motra and Stutz, 2018). Transducers were operated at 2 MHz and 1 MHz for P-wave and S-waves, respectively. The cumulative errors in both V_p and V_s were estimated to be <1% of the measured wave speed. Length and volume changes of the sample cube, resulting from changes in principal stresses, are obtained by the resulting piston displacement, which was used to correct the time-of-flight of the ultrasonic wave. Measurements were first performed at room temperature and increasing pressure up to 600 MPa, in step-wise increasing pressure increments. After the measurements during loading, the confining pressure was reduced and measurements were made during unloading (de-pressurization). Temperature-dependent measurements were made at maximum confining pressure (600 MPa), by increasing temperature from room conditions at 100 °C steps up to 600 °C. After each pressure increment, a five-minute stabilization period was used before reading the result and 30 min for each temperature increment - until stable conditions are reached. The rationale behind the choice for pressure and temperature stems from the conditions expected in the crust from a setting of exhumed crust (low P and low T), buried orogenic crust (e.g., the Himalayas and the Tibetan Plateau) so that the data can be compared to acquired seismic data from both ancient, exhumed and modern orogens. In addition, initially pressurizing the sample to 600 MPa followed by increasing the temperature to 600 °C, reduces the amount of thermally induced cracking of the sample during elastic wave speed measurements. Each set of results for a sample is comprised of nine velocities, measured along the three perpendicular axes (X, Y, Z): three P-wave velocities and six S-wave velocities. Each S-wave has two polarizations, s_1 and s_2 , where $s_1 > s_2$. The ultrasonic measurements were complemented by density measurements, which were measured by dividing the measured sample volume by its mass.

Based on the measurements, we calculate the difference in shear wave speed (dV_s), the anisotropy of the P- (% AV_p) and S- waves (% AV_s), and the mean V_p and V_s ,

$$dV_s = V_{s1} - V_{s2} \quad (1)$$

$$\%AV_p = \frac{V_{pmax} - V_{pmin}}{V_{pmean}} * 100 \quad (2a)$$

$$\%AV_s = \frac{V_{smax} - V_{smin}}{V_{smean}} * 100 \quad (2b)$$

$$V_{pmean} = \frac{V_{pmax} + V_{pint} + V_{pmin}}{3} \quad (3a)$$

$$V_{smean} = \frac{V_{smax} + V_{sint} + V_{smin}}{3} \quad (3b)$$

The fastest V_p and V_s directions represent V_{pmax} and V_{smax} , whereas the slowest axes are V_{pmin} and V_{smin} . Additionally, the Thomsen parameters ϵ and γ Thomsen (1986), were calculated

$$\epsilon = \frac{C_{11} - C_{33}}{2C_{33}} \quad (4)$$

$$\gamma = \frac{C_{66} - C_{44}}{2C_{44}} \quad (5)$$

The C_{11} , C_{33} , C_{44} and C_{66} are elastic constants that depend on the orientation of the sample. We assume that each sample has a transverse isotropic or orthorhombic symmetry. Elastic constants were then calculated using ultrasonic wave speed along the X, Y and Z axes, and density. It was not possible to calculate the Thomsen δ parameter, which requires measurements at 45° to the symmetry axis in the sample to obtain the C_{13} elastic constant.

3.2. Anisotropy of magnetic susceptibility (AMS)

AMS measurements were performed on rock cubes that were cut from the blocks initially used for ultrasonic wave speed measurements. Each block was cut into eight smaller cubes with ~20 mm sides. Measurements were made using the MFK1-FA susceptibility bridge (AGICO, Czech Republic) at the Department of Earth Sciences at Uppsala University. The measurements give the orientation and magnitude of three principal axes of the AMS ellipsoid ($k_{max} \geq k_{int} \geq k_{min}$), which are calculated from the second-rank symmetric magnetic susceptibility tensor. All measurements were made in the semi-automatic spinning mode, with an applied alternating field of 200 A/m and a frequency of 976 Hz. The values of k_{max} , k_{int} and k_{min} , based on up to eight separate cubes, were used to calculate the following AMS parameters.

$$\mu_1 = \ln(K_{max}), \mu_2 = \ln(K_{int}), \mu_3 = \ln(K_{min}), \mu_m = \ln(K_{mean}) \quad (6a)$$

$$K_{mean} = \frac{K_{max} + K_{int} + K_{min}}{3} \quad (6b)$$

$$P_j = \exp \sqrt{(2(\mu_1 - \mu_m)^2 + (\mu_2 - \mu_m)^2 + (\mu_3 - \mu_m)^2)} \quad (7)$$

$$T_j = \frac{2\mu_2 - \mu_1 - \mu_3}{\mu_1 - \mu_3} \quad (8)$$

$$F = \frac{K_{int}}{K_{min}} \quad (9)$$

$$L = \frac{K_{max}}{K_{int}} \quad (10)$$

The degree of anisotropy, P_j (Eq. (7)), is indicative of the magnitude of anisotropy, which reflects generally the shape preferred orientation (SPO), and crystallographic preferred orientation (CPO) (Borradaile and Jackson, 2004). Hence, a high P_j implies the presence of stronger fabric (defined by both SPO and CPO) intensity. Even though P_j does not directly measure the strain magnitude, it may indirectly be a useful indicator for principal axes of strain and the intensity of deformation (e.g., Borradaile and Henry, 1997). The shape of the anisotropy (Eq. (8)) is indicated by T_j and ranges from -1 to 1, which indicates prolate or oblate shape, respectively. A value of $T_j = 0$ represents a neutral triaxial ellipsoid with the difference of k_{max} and k_{int} being proportional to the difference between k_{int} and k_{min} . F and L represent the magnetic foliation and lineation in the sample.

For a better understanding of the source minerals that contribute to AMS, magnetic susceptibility was measured as a function of temperature. This test is used to identify the magnetic minerals present in the rock. The tests constitute continuous susceptibility measurements with increasing temperature, from room conditions up to 700 °C, followed by a decrease in temperature back to room conditions. For this measurement, sample material was crushed to fine-grained pieces with a milling machine and further crushed to a powder with an agate pestle and mortar. Measurements were done with the MFK1-FA, equipped with the CS-4 furnace add-on, in an inert argon gas atmosphere. The results for these measurements are presented in the online supplementary material

4. Results

4.1. Ultrasonic wave velocities (V_p and V_s)

Representative P- and S- waves velocities as a function of pressure and temperature are shown in Figs. 5 to 8. Equivalent seismic wave speed figures for all samples are provided in the online supplementary material S2 (Figs. S2–S5). Tables 2 and 3 summarize the seismic properties of all samples.

4.1.1. Compressional wave (V_p)

4.1.1.1. Felsic gneiss samples. All felsic gneiss samples show similar ultrasonic wave speeds and AVp, where the latter ranges from 4 to 6% at the highest applied pressure (600 MPa) and temperature (600 °C) (Tables 2 and 3). Sample 248 show an increase in $V_{p\text{mean}}$ as a function of pressure, from 4.71 to 6.41 km/s, from room conditions up to 600 MPa. The minimum and maximum P-wave velocities are 6.22 and 6.59 km/s along the Z- and X-axis, respectively. From these differences in velocity, the P-wave anisotropy at 600 MPa is 5.4%. Increasing temperature from room conditions to 600 °C (at a fixed pressure of 600 MPa), shows a drop in $V_{p\text{mean}}$ from 6.38 to 6.09 km/s. The minimum and maximum V_p are 5.91 and 6.26 km/s, resulting in an AVp of 5.8%. For sample 309, mean V_p increases from 4.93 to 6.18 km/s as a function of pressurization. At

600 MPa confining pressure, the minimum and maximum V_p are along the Z and X-axes, 6.00 and 6.32 km/s, respectively, with AVp of 5.3%. $V_{p\text{mean}}$ decreases slightly when increasing temperature to 600 °C, from 6.18 to 5.97 km/s. Minimum and maximum velocities are 5.80 and 6.08 km/s, along the Z – and X-axis, corresponding to an AVp of 4.7%. In Sample 362, $V_{p\text{mean}}$ increases from 4.53 to 6.48 km/s from room conditions to 600 MPa (Fig. 4a). The minimum and maximum V_p at 600 MPa pressure are 6.27 and 6.64 km/s along the Z- and X- axes, respectively. From these differences, the AVp is 5.6%. A slight decrease in V_p along with X, Y and Z-axes is observed as a function of temperature, and AVp is ~6% at 600 °C.

4.1.1.2. Intermediate, biotite-gneiss. The biotite-rich sample (316) shows generally comparably V_p to the mafic gneisses (below); AVp is the highest observed for any sample. $V_{p\text{mean}}$ increases from 4.82 to 6.45 km/s, from room pressure to 600 MPa. P-wave velocities at 600 MPa along the Z- and X- axes are 5.74 and 7.01 km/s resulting in an AVp of 19.7% (Fig. 4b). $V_{p\text{mean}}$ drops from 6.44 to 6.32 km/s when increasing temperature to 600 °C (Fig. 5b), and AVp increases to 23.8% (from the minimum and maximum V_p of 5.50 and 7.01 km/s).

4.1.1.3. Mafic gneiss samples. For sample 193, mean V_p increases from 5.23 to 7.07 km/s, from room pressure to 600 MPa (Fig. 4c). At 600 MPa pressure, the minimum and maximum V_p are 6.61 and 7.49 km/s along

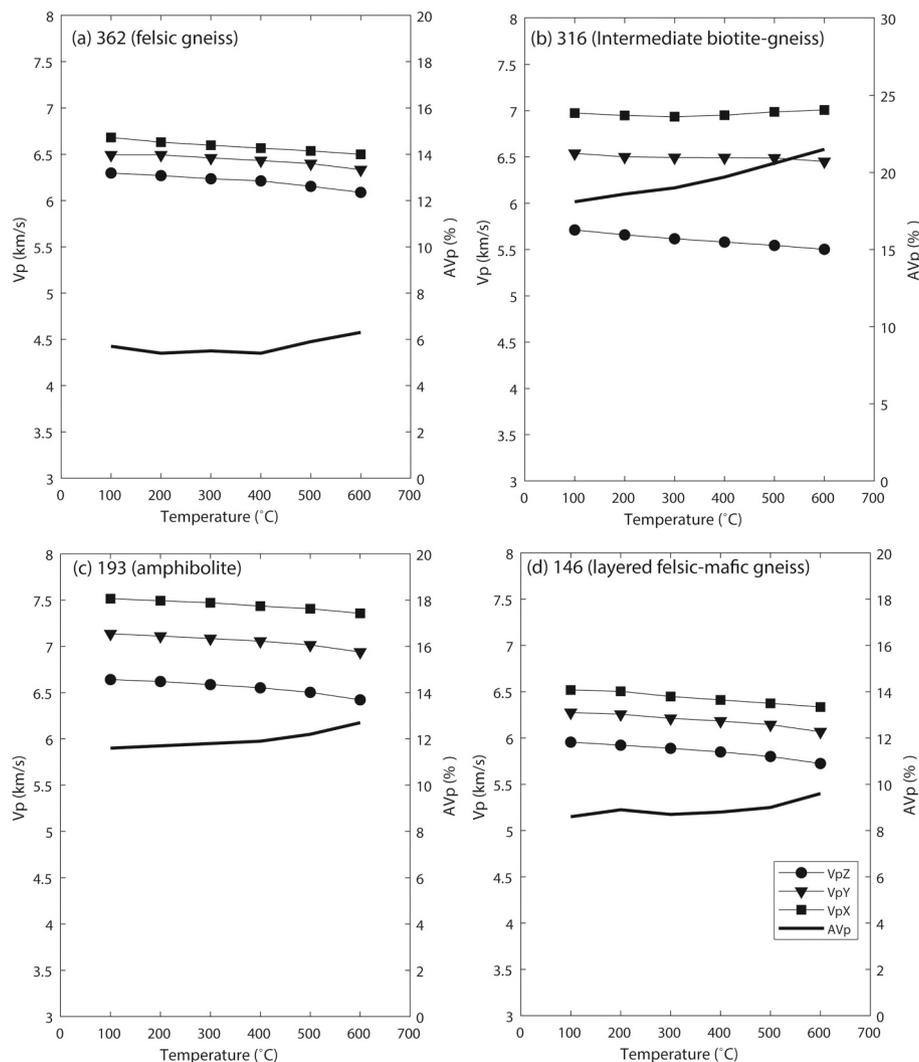


Fig. 5. P wave velocity (V_p ; left y-axis) and anisotropy of V_p (AVp, right y-axis) as a function of temperature for samples (a) 362 – felsic gneiss; (b) 316 – Intermediate biotite-gneiss; (c) 193 – amphibolite; (d) 146 – layered felsic-mafic gneiss. All measurements were made at a fixed pressure of 600 MPa.

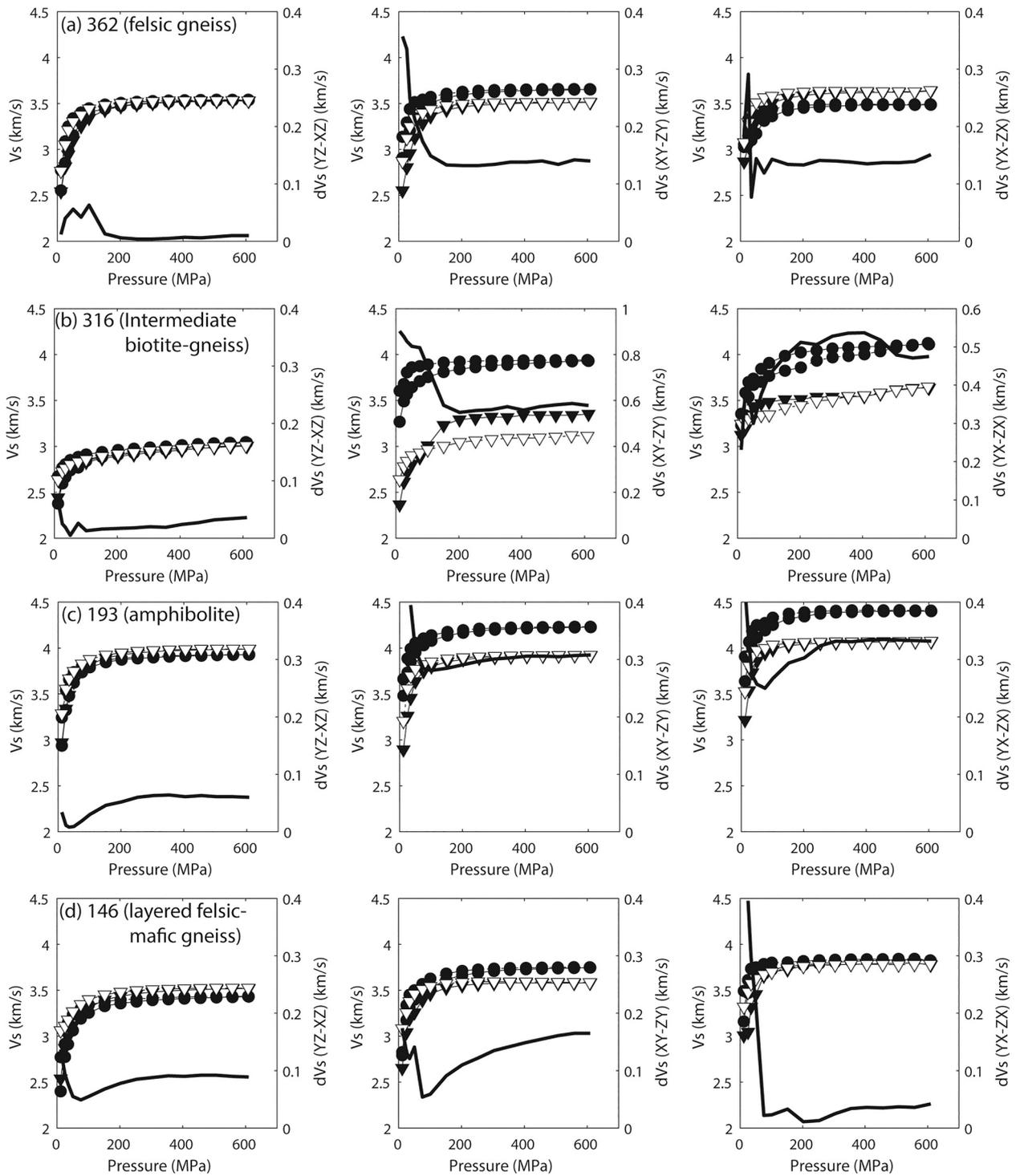


Fig. 6. Shear wave velocities and shear wave splitting (dV_s) for selected, representative samples (a) 362 – felsic gneiss; (b) 316 – Intermediate biotite-gneiss; (c) 193 – amphibolite; (d) 146 – layered felsic-mafic gneiss. Measurements were carried out at room temperature as a function of pressure. The filled circles show V_s during pressurization and the open circles represent V_s during depressurization. The black lines indicate the shear wave splitting value (dV_s) for polarized shear waves, as a function of pressure and is read on the right y-axis.

the Z- and X- axes, respectively; AV_p at 600 MPa is 12.2%. P-wave velocity shows a slight drop due to increasing temperature to 600 °C for $V_{p,mean}$, from 7.11 to 6.91 km/s. At 600 °C, the minimum and maximum velocities are 6.42 and 7.36 km/s; correspondingly AV_p increased to 13.6% (Fig. 5c). For sample 274 $V_{p,mean}$ increases from 5.52 to 6.89 km/s, during pressurization from room conditions to 600 MPa. The minimum and maximum P-wave velocities at 600 MPa pressure are 6.65 and 7.33 km/s along the Y- and X-axis, resulting in 9.7% AV_p (note that V_p

along with Y- and Z-axes are similar for this sample). $V_{p,mean}$ shows a drop from 6.9 km/s to 6.89 km/s when increasing temperature to 600 °C. The minimum and maximum velocities at 600 °C are 6.52 and 7.21 km/s, along the same axes as during pressurization, resulting in 10.0% AV_p . In sample 340 $V_{p,mean}$ increases from 4.70 to 7.05 km/s. At 600 MPa pressure the minimum and maximum V_p are 6.33 and 7.60 km/s along the Z- and X- axes, respectively, resulting in an AV_p of 17.9%. The increasing temperature to 600 °C leads to a drop in $V_{p,mean}$

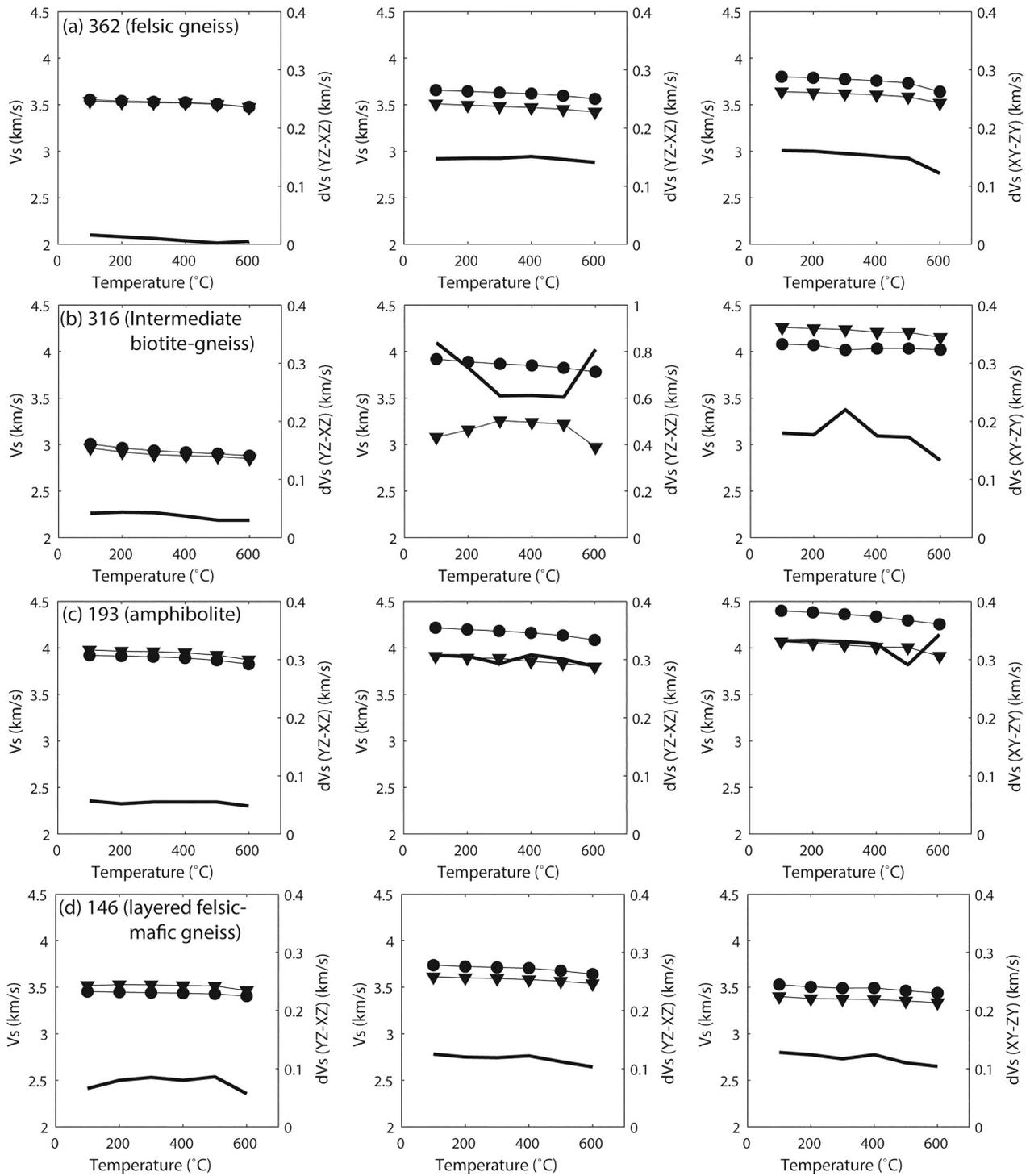


Fig. 7. Shear wave velocities and shear wave splitting (dVs) as a function of temperature for samples (a) 362 – felsic gneiss; (b) 316 – Intermediate biotite-gneiss; (c) 193 – amphibolite; (d) 146 – layered felsic-mafic gneiss. Measurements were performed at 600 MPa confining pressure. The black lines indicate the shear wave splitting value (dVs) for polarized shear waves, as a function of temperature and is read on the right y-axis.

from 7.08 to 6.91 km/s. Minimum and maximum velocities are 6.12 and 7.54 km/s, resulting in an AVp of 18.0%.

4.1.1.4. Layered felsic-mafic gneiss. Sample 146 results show that $V_{p\text{mean}}$ changes from 5.23 to 6.98 km/s, as a function of pressure (Fig. 4d). At 600 MPa pressure, the minimum and maximum P-wave velocities are 5.94 and 6.49 km/s. P wave anisotropy at 600 MPa is 8.9%. P-wave velocity shows a slight drop due to the temperature increase, from 6.26 to 6.04 km/s. At 600 °C the minimum and maximum

velocities were 5.73 and 6.33 km/s, along the X-axis and Z-axis, respectively (Fig. 6d); AVp at 600 MPa and 600 °C increased to 10.1%.

4.1.2. Shear wave velocities (Vs_1 , Vs_2) and shear wave splitting (dVs)

4.1.2.1. Felsic gneiss samples. As with P-wave velocities, shear wave speeds increase as a function of pressure and decrease slightly as a function of increasing temperature. Shear wave splitting is generally highest along X- and Y-axes, whereas minimum splitting occurs along

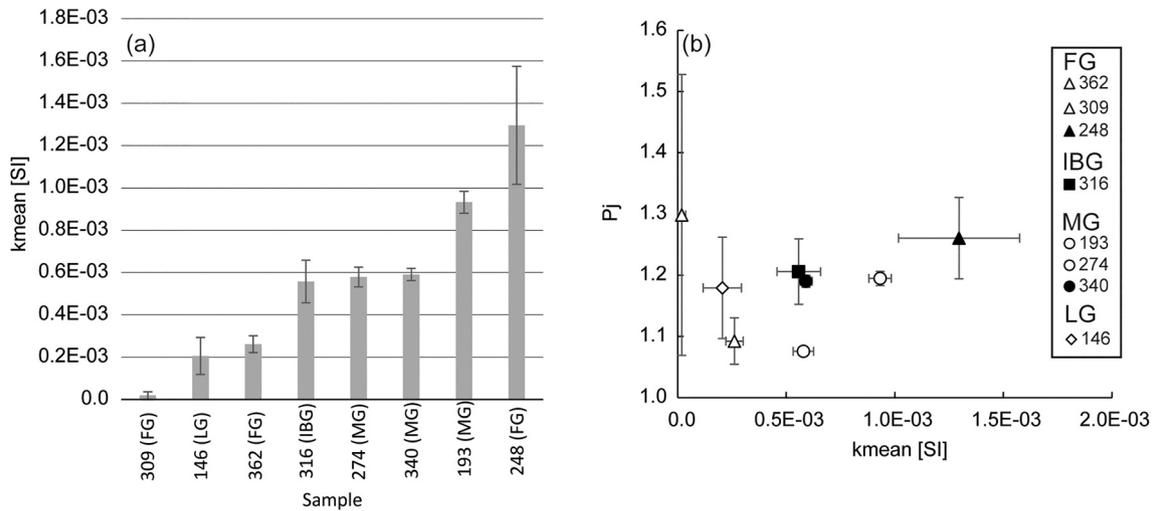


Fig. 8. Mean susceptibility of 63 specimens cut out of the eight cubic samples. a) mean susceptibility against the degree of anisotropy, P_j . The error bars represent two standard deviations and were obtained from the measurements of the AMS sub-samples. b) mean susceptibility for each sample with two standard deviations (obtained from measurement of the AMS sub-samples).

the Z-axis. Results are shown in Figs. 6 and 7 and in Tables 2 and 3.

4.1.2.2. Felsic gneiss samples. At 600 MPa, sample 248 exhibits dVs along X, Y and Z axes of 0.008, 0.099 and 0.119 km/s, and at 600 MPa and 600 °C, the dVs has increased slightly, to 0.01, 0.11 and 0.14 km/s. Sample 309 exhibits fairly low dVs and at room temperature and maximum confining pressure, the dVs along the X, Y and Z axes are 0.153, 0.003 and 0.151 km/s, respectively, and at 600 MPa and 600 °C, the respective axes show dVs of 0.033, 0.003 and 0.109 km/s. Shear wave splitting for sample 362 (Felsic gneiss) is minimum along the X-axis; splitting along the Y-axis is highest and along the Z-axis it is intermediate (Fig. 7a). This is similar for room temperature conditions and 600 MPa confining pressure, and applying temperature to 600 °C (Fig. 8a). The dVs under pressure condition are 0.01, 0.14 and 0.151 km/s and in fixed 600 MPa pressure and 600 °C are 0.005, 0.141 and 0.122 km/s along X-, Y- and Z-axes, respectively.

4.1.2.3. Intermediate, biotite-gneiss. Sample 316 exhibits maximum splitting along the X- and Y-axes. This trend remains similar from room temperature and increasing pressure with dVs of 0.475, 0.578 (Fig. 7b) and 0.036 km/s to fixed 600 MPa pressure and 600 °C (Fig. 8b) with dVs of 0.03, 0.809 and 0.133 km/s along X, Y and Z-axes.

4.1.2.4. Mafic gneiss samples. For sample 274, dVs remains similar at maximum confining laboratory conditions pressure (600 MPa), compared to maximum pressure and temperature (600 °C). The dVs under these conditions along X, Y and Z axes are 0.01, 0.22, 0.19 km/s and 0.02, 0.10, 0.24 km/s, respectively. Sample 193 shows maximum shear wave splitting along the X-axis and minimum splitting along the Z-axis. The differences between V_{s1} and V_{s2} at room temperature and 600 MPa are, along X, Y and Z axes, 0.06, 0.31 and 0.33 km/s, respectively (Fig. 7c). At 600 MPa and 600 °C, are 0.048, 0.288 and 0.343 km/s (Fig. 8c). Sample 340, shows maximum splitting along Y and X-axes and minimum splitting along the Z-axis. Its dVs 600 MPa and room temperature, along the X-, Y- and Z-axes, are 0.008, 0.633 and 0.737 km/s and at 600 MPa pressure and 600 °C temperature in X, Y and Z axis are 0.004, 0.637 and 0.722 km/s.

4.1.2.5. Layered felsic-mafic gneiss sample. Sample 146 shows relatively small shear wave splitting overall, with maximum splitting along the Y-axis, whereas the smallest splitting occurs along the X- and Z-axes. This trend remains similar at 600 MPa (Fig. 7d) and 600 °C (Fig. 8d),

although it is somewhat reduced.

4.2. Magnetic mineral composition and anisotropy of magnetic susceptibility (AMS)

Based on thermomagnetic analyses (online supplementary material, S1) and mean susceptibility values, all samples show the presence of ferromagnetic minerals (e.g., magnetite, pyrrhotite, hematite), although in varying amounts. The analysis is done in order to identify if potential ferromagnetic minerals, in addition to Fe-bearing paramagnetic (matrix) phases (such as amphibole and biotite), can be contributors to the AMS. Sample 248 show a significant presence of ferromagnetic minerals, based on bulk susceptibility (Fig. 8a, b) and thermomagnetic analysis. Six out of eight samples show a decrease in susceptibility as a function of temperature around 300–320 °C suggesting the presence of pyrrhotite or possibly conversion of maghemite, which is unstable at higher temperatures (Fig. S1). Seven out of the eight samples shows a susceptibility drop at ~580 °C, which indicates that magnetite is present.

In regards to the AMS, for all samples the k_{\min} axis plots sub-parallel to the pole to foliation (Fig. 9; Z sample axis). The k_{\min} axis is less well grouped for samples 248 and 309 (felsic gneisses) and deviates somewhat from the pole to foliation. k_{\max} and k_{int} axes plot in the X-Y plane, and are typically clustered. Although the mineral lineation is marked by the sample X-axis, there is not an apparent consistent relationship between the k_{\max} -axis and X-axis. A few samples display k_{\max} being sub-parallel to the X-axis, including 274 (mafic gneiss), 309 (felsic gneiss) and 316 (intermediate biotite-gneiss), whereas other samples show k_{\max} sub-perpendicular to the lineation, such as 146 (layered felsic-mafic gneiss), 193 (mafic gneiss), and 248 (felsic gneiss). Samples 274 (amphibolite) and 362 (felsic gneiss) show K_{\max} between the X- and Y-axes.

Further information regarding the petrofabric, using AMS, can be obtained from plots of the degree of anisotropy, shape parameter and the magnetic lineation and foliation (Fig. 10). Between them, there are six samples showing a foliation dominated AMS, and two samples that show a neutral triaxial magnetic fabric. Samples 274 (amphibolite) and 362 (felsic gneiss) are close to the neutral line (where L and F are equal). Samples 309 and 248 (felsic gneisses) plot differently compared to other samples (Fig. 10a), with a $P_j > 1.25$ and $T_j < 0.2$. (See Fig. 11.)

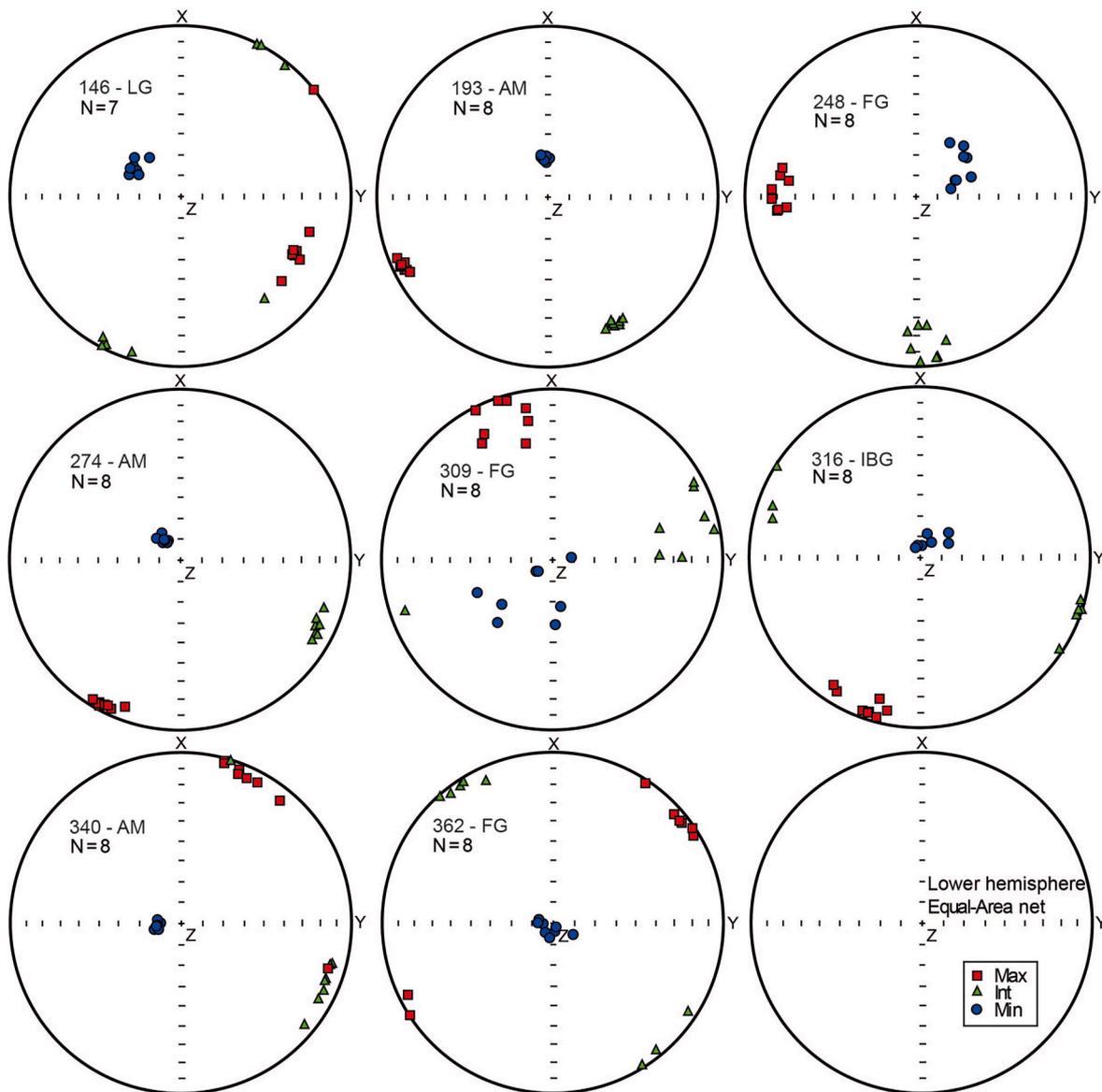


Fig. 9. Anisotropy of magnetic susceptibility plotted in lower hemisphere equal area nets for subsamples (N) of the original cubes used for ultrasonic velocity experiments. X, Y and Z axes correspond to the sample reference frame used for the ultrasonic measurements. Abbreviations in the figure are provided in the sample description are FG: gneiss; IBG: intermediate biotite gneiss; LG: layered gneiss; AM – amphibolite.

5. Discussion

5.1. Influence of pressure, temperature and microcracks on seismic anisotropy

There are several possible sources for the generation of seismic anisotropy, most notably the existence of microcracks, crystallographic preferred orientation (CPO) and shape preferred orientation (SPO), as well as layering and compositional banding of minerals and rocks (Babuska and Cara, 1991). The exponential increase in V_p as a function of pressure in Figs. 5 and 7 represent the zone of crack closure pressure, which ranges from up to 150 MPa to 350 MPa, in line with observations reported by Ji et al. (2007) and Sun et al. (2012) for drill core samples. Generally, microcracks are prone to close at confining pressures exceeding 100–200 MPa (Kern et al., 1997; Kern and Wenk, 1990). The closure pressure varies somewhat for each sample, but typically the relationship between V_p and V_s as a function of pressure tends to become linear above 200 MPa (Figs. 5, 7). Additionally, the velocity as a function of crack closure pressure varies depending on the direction of

measurement. Compare, for example, V_p as a function of pressure along X and Z axes in Fig. 5 (a, b; felsic gneiss 362 and intermediate biotite gneiss 316), which show markedly different pressurization and depressurization curves along the two axes. During depressurization, the cracks gradually re-open, although not to the same extent as before pressurization, resulting in hysteresis (Ji et al., 2007). The experimental data show that this hysteresis effect is largest along the Z-axis (Fig. 5), which is the axis that is normal to the orientation of the foliation plane, suggesting that this plane exhibits the largest amount of interaction with microcracks; i.e., microcracks tend to orient parallel to the foliation plane. Similar effects were observed during the temperature increase (Figs. 6 and 8), indicating that at higher temperature microcracks appear to be generated and thus reduces V_p and V_s .

To further assess the different loading and unloading effects we plot V_p variations along different sample axes, e.g., $V_p(X)-V_p(Z)$, during pressurization to 600 MPa and depressurization to room pressure (Fig. 12). This projection of data is useful to show the development of sample anisotropy, comparing different sample axes, as a function of pressure (hence reflecting micro-crack closure as a function of

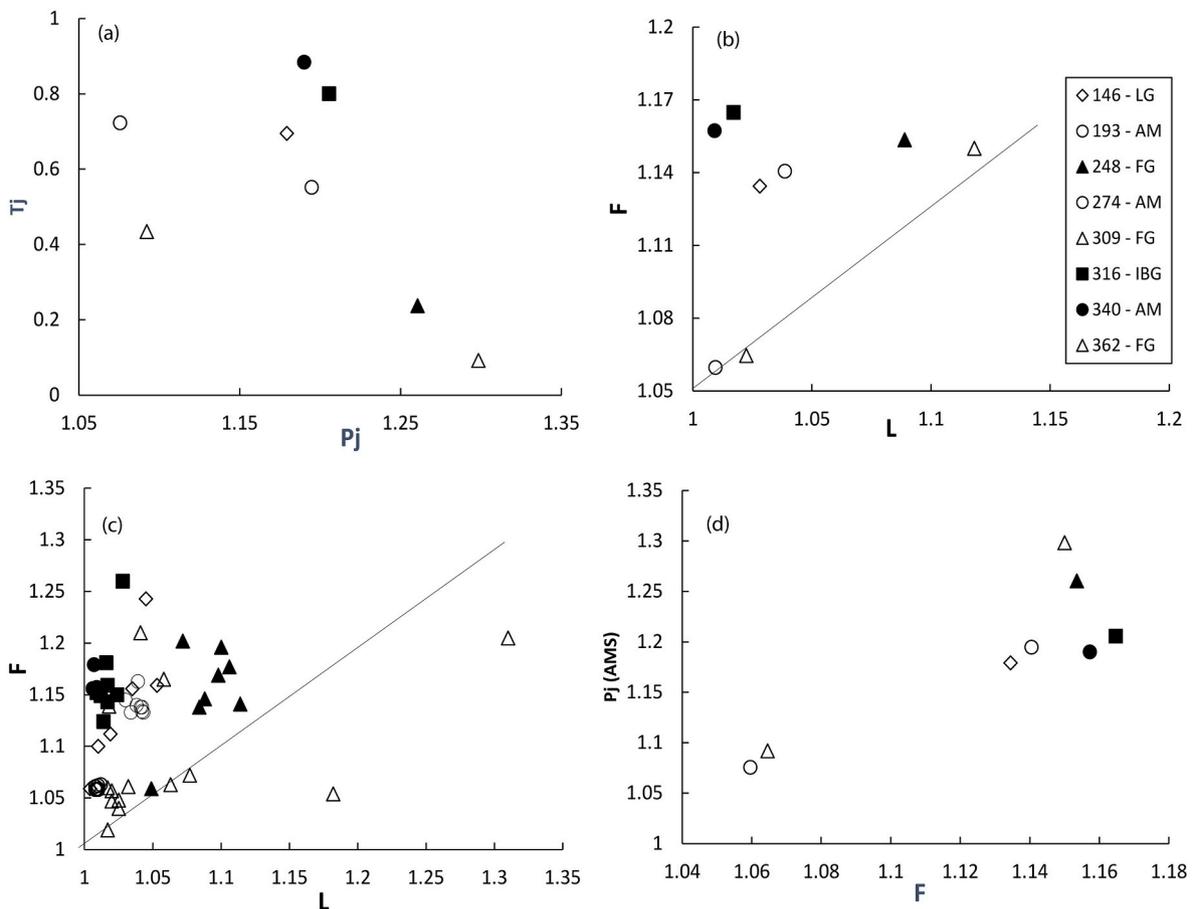


Fig. 10. Anisotropy of magnetic susceptibility parameters, P_j , T , L and F , shown against each other; a) Shape of the anisotropy (T_j) vs. the degree of anisotropy (P_j); b) mean value of foliation and lineation of all samples and c) magnetic foliation (F) and lineation (L); d) magnetic foliation to the degree of the anisotropy. The data plotted in (a), (b) and (d) represent the average value based on the set of sub-samples from each cubic sample used for ultrasonic measurements.

pressurization and depressurization). Notably, the intermediate biotite gneiss (Fig. 12b) stands out among the tested samples as it does not show a reversible path for V_p during pressurization and depressurization. For this sample, the difference between V_p along X, Y and Z axes remain similar upon depressurization, which most likely indicates that micro-cracks do not re-open during depressurization.

All samples were re-measured at room pressure and room temperature the day following the pressurization and depressurization experiment (i.e., the sample was allowed to remain in the pressure vessel at ambient pressure conditions), to investigate possible effects of residual stresses. Although not shown in a figure, samples showed that they had accumulated residual stresses having only partially released the stress the day after the loading measurement. The difference between V_p measurements before and after the pressure cycling ranged from 5% to 20%, depending on the sample, which we believe is the result of closure (and slow re-opening) of pre-existing intercrystalline cracks since we did not observe any localized damage or deformation in the thin sections (e.g., Figs. 3 and 4; note that thin sections were prepared from the sample cube subsequent to the ultrasonic measurements). However, for all samples, the largest differences were observed normal to the foliation.

With an increase in pressure from room conditions to 600 MPa, there is a successive diminishing in P- and S- wave anisotropy (Fig. 5). However, when pressure remains fixed at 600 MPa and temperature increases from room conditions to 600 °C, the P-wave anisotropy increases by up to 4% from its value at 600 MPa and room temperature (e.g., Fig. 6b). This observation is similar to the trend observed during depressurization, suggesting that changes in anisotropy are due to the reopening of intercrystalline cracks. The increase in AVp by applying

temperature ranges from 0.4% in sample 248 (felsic gneiss) up to 4% in sample 316 (intermediate biotite gneiss; Fig. 6). Anisotropy of P- and S-wave velocities increases to a greater extent in the biotite-gneiss and amphibolites compared to the layered sample (146) and felsic gneisses, as a function of temperature. The layered sample 146 (layered felsic-mafic gneiss) follows a similar trend of AVp as a function of temperature as other samples (in particular amphibolites). At 600 MPa, this sample has a relatively high AVp = 8.9%, and AVp increased to 10.1% as a function of pressure and temperature (at 600 MPa and 600 °C). This may indicate that thermal expansion, or the temperature dependence of V_p and V_s , of different minerals, plays a significant role in increasing AVp and AVs. One possibility is that the directional dependence of thermal expansion in amphibole (Tribaudino et al., 2008) can affect the anisotropy of elastic wave measurements. Also, the anisotropic thermal expansion of biotite seems to be one of the most logical reasons for an increase in seismic anisotropy as a function of temperature for samples 316 (intermediate biotite gneiss) and 340 (biotite-bearing amphibolite). Siegesmund et al. (2008) has noted the considerable difference of thermal expansion along the c-axis ($\alpha = 17.3 \times 10^{-6} \text{ K}^{-1}$) and within the basal plane ($\alpha = 9.65 \times 10^{-6} \text{ K}^{-1}$) of biotite. This difference, of nearly two times higher expansion along the c-axis, may be of importance also in influencing seismic anisotropy (at least on a laboratory scale). Notably, samples 309 (felsic gneiss), and 193 (amphibolite) showed a reduction of AVs during temperature-dependent measurements, compared to measurements made solely at confining pressure, and thus deviate from the trend exhibited by other samples.

Samples 316 and 340 contain a significant amount of biotite (>10%), with an AVp of 19.7% and 17.9%. Although sample 340 is dominated by

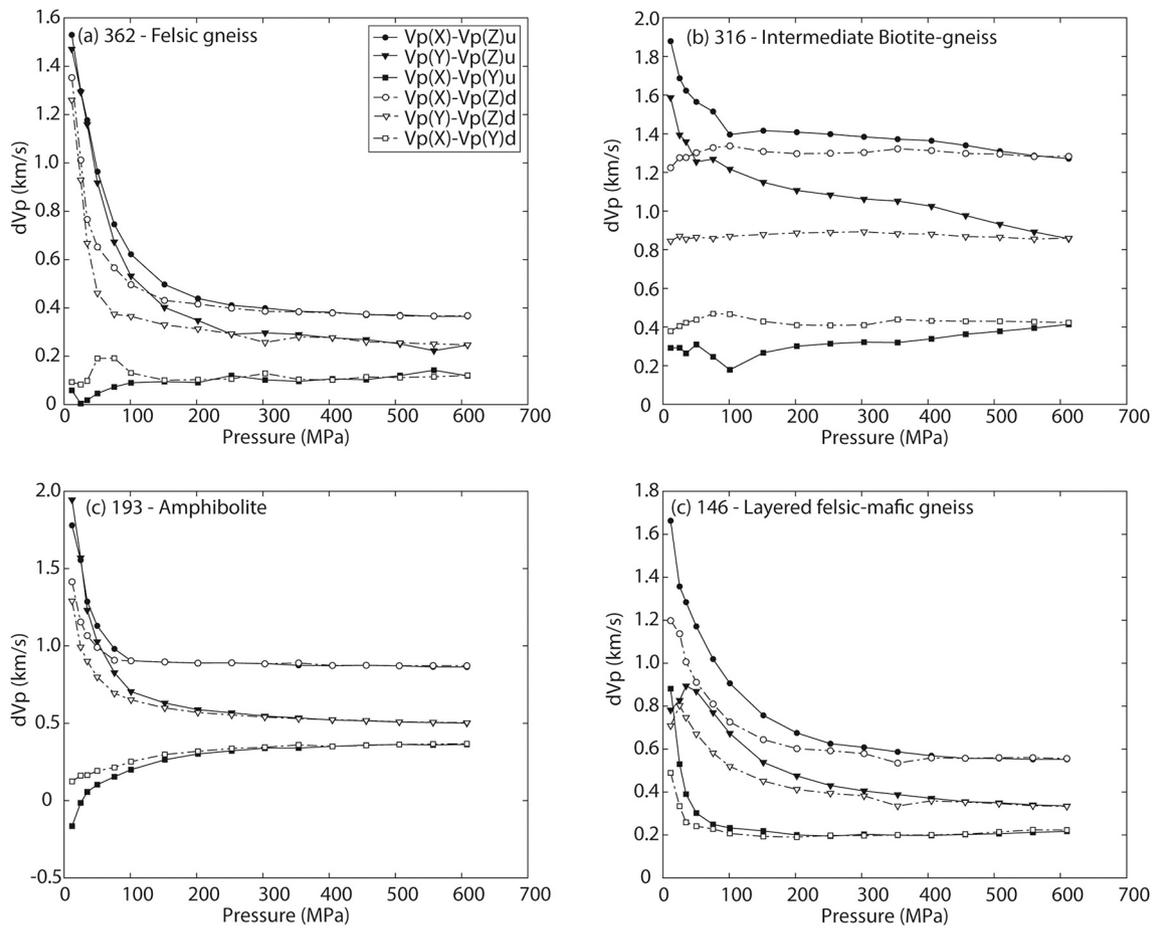


Fig. 11. Differences in Vp along different sample directions, expressed by $V_p(X)-V_p(Z)$, $V_p(Y)-V_p(Z)$ and $V_p(X)-V_p(Y)$, for samples (a) 362 (Felsic gneiss), (b) 316 (intermediate biotite-gneiss), (c) 193 (amphibolite) and (d) 146 (layered felsic-mafic gneiss). The letter ‘u’ and ‘d’ in the legend indicates pressurization and depressurization, respectively.

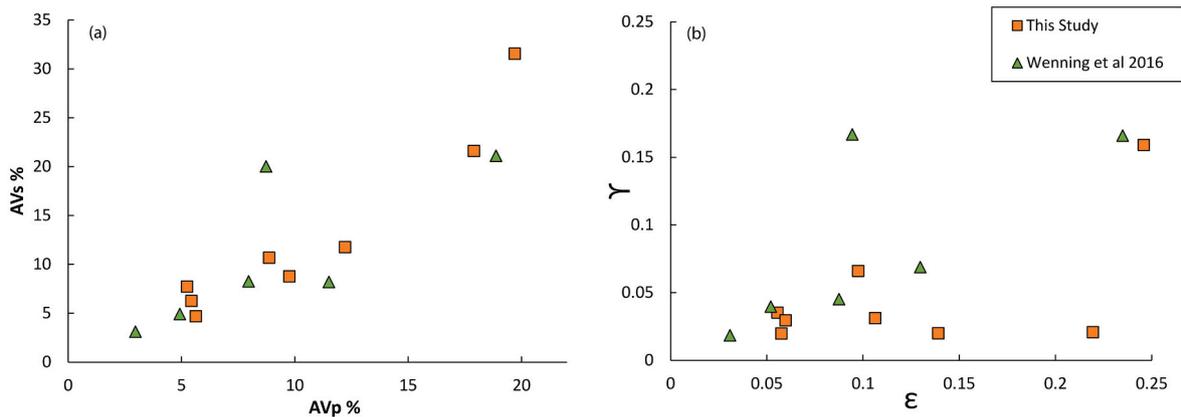


Fig. 12. Elastic wave parameters calculated from ultrasonic wave propagation data. These values are calculated directly from elastic wave speed and density results. a) Percentages of AVp (%) and AVs (%) plotted against each other (data from this study and Wenning et al., 2016). b) Relationship between Thomsen parameters (ϵ and γ), (data from this study and Wenning et al., 2016). Wenning et al. (2016) carried out measurements up to 260 MPa confining pressure; in this study, the comparable data were measured up to 600 MPa.

amphibole, the observed AVp is high considering the high single-crystal AVp (up to 27%) of amphibole (hornblende; Aleksandrov and Ryzhova, 1961; Aleksandrov et al., 1974; Brown et al., 2016), and the additional influence of biotite is very likely. Biotite has very high single-crystal anisotropy, with AVp ~64% (Aleksandrov and Ryzhova, 1961). Qualitatively, SPO of biotite is apparent for both samples 316 and 340 (e.g., Fig. 3d), which is likely to coincide with the CPO and hence produce a

high seismic anisotropy.

5.2. Application of COSC-1 sample results to seismic anisotropy of the middle to lower continental crust

5.2.1. Elastic wave speeds from COSC-1 laboratory studies

Two other laboratory studies have presented data on elastic wave

velocity and anisotropy from the COSC-1 project (Wenning et al., 2016; Kästner et al., 2020). These previous studies used three one-inch diameter bored plugs from each drill core sample; each plug was taken along a different axis, representing the structural X, Y and Z axes, from which elastic wave anisotropy could be calculated. Wenning et al. (2016) provided AVp, AVs and calculated Thomsen parameters, at confining pressures up to 260 MPa. Kästner et al. (2020) provided p-wave velocities of sixteen core samples with measurements made up to 250 MPa confining pressure and compared the results with on-drillcore Vp measurements at ambient conditions and borehole sonic logging data. The present work extends these conditions by measurements up to 600 MPa and additionally adding temperature-dependent measurements, up to 600 °C. A comparison of data obtained from the three studies is given in Figs. 12 and 13. AVp (%) vs AVs (%) data and a comparison of calculated Thomsen parameters are given in Fig. 12a, b (only based on Wenning et al., 2016 and this study). There is a fair agreement between AVp and AVs (Fig. 12a), even though the peak confining pressures are rather different (comparing 260 MPa and 600 MPa). Note that the anisotropy that was obtained at 600 MPa in this study, is compared with the crack-free back-extrapolated anisotropy (labelled AVp₀; AVs₀), which was calculated by Wenning et al. (2016) and Kästner et al. (2020) (Figs. 12 and 13). For the Thomsen parameters, there is relatively little change in γ , and values typically range up to 0.06 (sample 316, intermediate biotite gneiss, shows a deviation with $\gamma \sim 0.15$), while there is a larger range in ϵ . This is due to the smaller differences between the fastest and slowest S-wave velocities, which influences γ . In comparison, there are relatively larger differences in Vp among samples, leading to a wider range in ϵ .

5.2.2. Implications for crustal seismic anisotropy

One of the main aims of the COSC-1 project is to understand how the Middle Allochthon Lower Seve Nappe rocks were emplaced into the middle crust during the Caledonian orogeny (Gee et al., 2010; Gee et al., 2013; Lorenz et al., 2015). The rocks experienced large scale ductile flow during emplacement at middle crustal conditions at amphibolite to greenschist grade metamorphic conditions. The laboratory conditions we apply in this study (600 MPa and 600 °C) appear appropriate to recreate the mid-crustal conditions that these rocks experienced during emplacement (and subsequent to their emplacement), and may therefore be of use when comparing lab results to present-day settings where middle crustal flow currently occur (e.g., Himalaya-Tibet). The results of AVp and AVs for amphibolites and biotite bearing gneiss and amphibolite are generally around 10% or higher. Felsic gneisses samples are comparatively less anisotropic, with AVp of <7% and AVs of <6%. These results share notable similarities with other experimental studies on cubic samples employing multi anvil apparatus to pressures of 600

MPa (e.g., Kern, 1990; Babuska and Cara, 1991; Barruol and Mainprice, 1993; Barruol and Kern, 1996; Kern et al., 1999; Kern et al., 2002; Kern et al., 2008).

The impact of the results in this study concerns both the intrinsic (i.e., CPO and SPO) and extrinsic sources (i.e., microcracks) of anisotropy. There are several studies on samples at elevated pressure, where the influence of open micro-cracks and pores are gradually reduced (e.g., Wyllie et al., 1958; Todd and Simmons, 1972; Eberhart-Phillips et al., 1989; Ji et al., 2007; Ullemeyer et al., 2011; Madonna et al., 2012). It is generally assumed that most microcracks are closed above 200 MPa, which would in nature correspond to a depth of $\sim 7\text{--}8$ km (Emmermann and Lauterjung, 1997). However, from the KTB well in Germany, it is known that fluid-filled fractures and pores can remain open until 9 km depth (Kern et al., 1991). Sun et al. (2012) studied samples collected from the Chinese Continental Scientific Drilling (CCSD) cores and reported that most microcracks were newly formed by the release of in-situ stresses in the borehole. These microcracks are relatively clean without secondary mineral fillings and have smaller aspect ratios, and are thus more easily closed under the applied hydrostatic pressure conditions in the laboratory. After the cracks are fully closed, the core samples that contain little weathering products, are much less sensitive to pressure and thus have lower pressure derivatives and smaller hysteresis than the samples collected from the surface outcrops. To understand the effects of cracks better we need to consider their closure pressure carefully. Equations derived by Walsh (1965), based on continuum elasticity theory, indicates that crack-closure is closely related to the shape of the microcrack, and in particular its aspect ratio. Higher aspect ratio cracks, or pores, may remain open to pressures >1 GPa pressure as experimentally shown by Christensen (1974). In fact, it is impossible to close high aspect ratio pores (i.e., spherical pores) or cracks fully during pressurization (Walsh, 1965). A lower aspect ratio of cracks leads to an increase in rock compressibility, which is reduced during pressurization. Additionally, Walsh (1965) showed that low aspect ratios generally increase elastic anisotropy in rock prior to closure during pressurization. The influence of temperature on the opening of microcracks and elastic anisotropy is nonlinear, considering measurements from room conditions up to 600 °C. There is an apparent reduction in Vp and Vs, but a general increase in anisotropy. We can only speculate as to the origin of the increase in seismic anisotropy, but thermal expansion due to alignment (CPO) of the minerals appear as a plausible explanation, as is noted in Section 5.1.

Cracks and pore spaces are important sources of anisotropy in the upper crust, but tend to diminish as we move deeper into the crust (Schijns et al., 2012; Audet, 2015). Therefore, it is important to investigate the influence of microcracks versus intrinsic anisotropy. Intrinsic anisotropy does not depend on microcracks but on constituting minerals

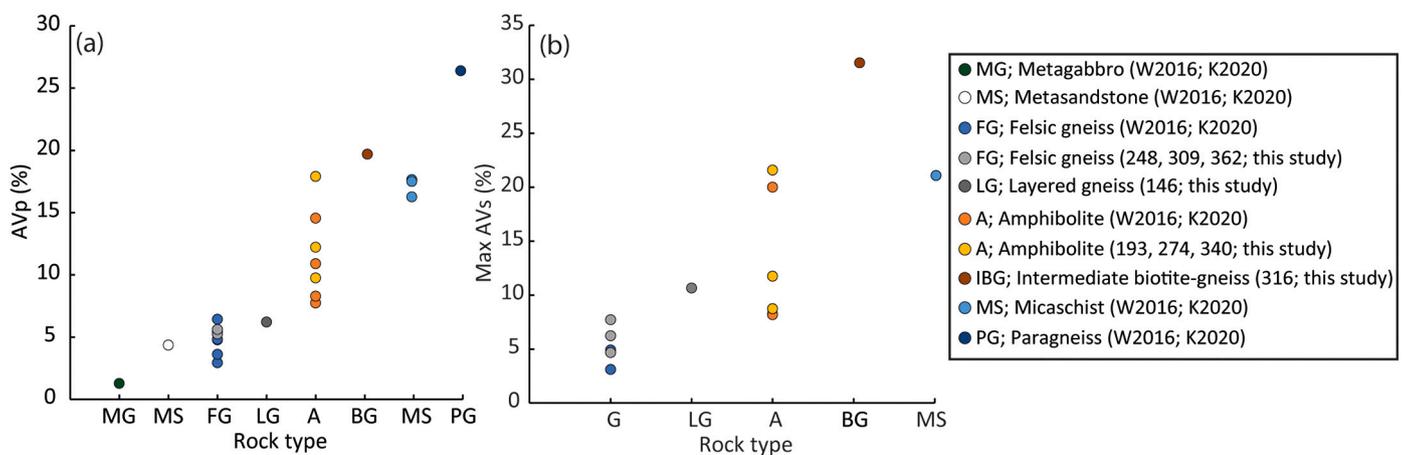


Fig. 13. A comparison of (a) AVp and (b) Max AVs from this study and the studies of Wenning et al. (2016) and Kästner et al. (2020). Note the different sizes of the circles, where large circles are from this study and smaller circles are from Wenning et al. (2016) and Kästner et al. (2020).

and the fabric of the rock while extrinsic anisotropy is crack-related. We observed that intrinsic anisotropy is containing 20% to 25% of the initial anisotropy values of biotite gneiss and amphibolite, felsic gneiss and mixed layered samples due to the assumption of microcracks and pores effects diminishes and are negligible by 600 MPa confining pressure. The pressure range of the crack closure in metamorphic rocks has been the subject of investigations in different studies to better understand the middle and lower crust, as well as the upper mantle (Anderson et al., 1974; Ji et al., 2003; Ji et al., 2007). A general increase in seismic anisotropy is furthermore observed due to the increase in temperature at 600 MPa confining pressure. In our samples, we observed up to 3 to 4 times reduction in anisotropy when comparing ambient and maximum confining laboratory pressure conditions (600 MPa). Additionally, seismic anisotropy generally increased slightly when increasing temperature to 600 °C. The most drastic change in anisotropy occurs for sample 316 (intermediate biotite-gneiss), where AVp increases from 19.7% (at 600 MPa and room temperature) to 23.8% (at 600 MPa and 600°C). The origin of the increase in anisotropy results mainly from the linear decrease in Vp along with the sample Z-axis (normal to foliation) as a function of increasing temperature, whereas Vp remains more or less unchanged along the sample X-axis (parallel to mineral lineation). Therefore, we suggest that at conditions representative for the middle and lower crust, ~80- to 90% of anisotropy exhibited by the biotite bearing specimens, felsic gneiss and amphibolites are due to intrinsic sources of anisotropy, including CPO and SPO, whereas micro-cracking can still contribute up to 20% of anisotropy (even at mid-crustal conditions). Clearly, crack and pore space at depth will not be empty or dry, as in experiments performed in this study. The presence of any type of pore fluid is likely to increase the effect of crack or porosity influence on seismic anisotropy since the effective pressure can be reasoned to be lower than the confining pressure. Even if the influence is smaller than at upper-crust levels, the presence of micro-cracks should therefore be considered in observed seismic anisotropy at middle and lower crustal conditions.

5.3. Relationship between AMS and elastic wave anisotropy

As derived seismic parameters are in principle analogous to parameters used to describe AMS, it is reasonable to hypothesize that AMS, as a petrofabric indicator, can be used as a proxy for elastic wave anisotropy. However, the latter type of measurement is simplified because it is only represented by measured values along three axes and not the complete elastic tensor. A positive correlation between seismic anisotropy and magnetic susceptibility anisotropy was shown experimentally by Punturo et al. (2017) on metamorphic rocks, although their samples were weakly anisotropic. In contrast, Biedermann et al. (2020) noted that the relationship between seismic and magnetic anisotropy is not straightforward, given that the former is governed mainly by the bulk mineral composition, whereas the latter is controlled to a great extent by small amounts of ferromagnetic minerals. A complicating factor, observed particularly in shear zones, is the obliquity between the CPO and SPO (Ji et al., 1996). Such obliquity can potentially yield a significant difference between AMS and seismic anisotropy, in particular when bulk magnetic susceptibility exceeds 1×10^{-3} SI, and the magnetic fabric is likely to be controlled by the SPO of ferrimagnetic iron oxides (Rochette et al., 1992).

Here, we discuss the results of the measurements of AMS in comparison with the ultrasonic wave data. The rationale for this approach is that we can use AMS data as a proxy for the petrofabric (CPO and SPO) and link this to elastic wave anisotropy, based on the inference that AMS provides a suitable proxy for SPO and CPO of the major rock-forming minerals (Owens and Rutter, 1978; Punturo et al., 2017). The aim is to provide the conditions for when AMS is a suitable proxy for seismic anisotropy. It has to be acknowledged that magnetic susceptibility and elastic wave velocity values of the minerals are generated partly from the same source, mainly because of the CPO of paramagnetic (matrix)

minerals, but also in part from different sources. The magnetic susceptibility and its anisotropy are related to mineral composition and mineral chemistry, whereas elastic wave anisotropy results mainly from CPO (Almqvist et al., 2013; Biedermann et al., 2020). In addition, magnetic susceptibility is strongly influenced by the minor presence of ferromagnetic minerals, whereas the elastic wave velocities arise mainly from the bulk mineral composition. Lagroix and Borradaile (2000) have noted that mafic silicate rocks mostly have magnetic susceptibility higher than predicted based on their chemical composition. This is due to magnetite inclusions in the silicate grains, which may also increase the magnetic anisotropy. To illustrate the relationship between AMS and elastic wave anisotropy, Fig. 14 shows a comparison of parameters obtained from measurements of AMS and elastic wave velocities.

Mica and amphibole-dominated rocks represent important sources of anisotropy in the deformed middle and lower crust. Between our measured samples, the biotite bearing and amphibole-rich samples clearly have the highest seismic anisotropy at ambient temperature and 600 MPa confining pressure conditions as well as 600 MPa and 600 °C. Although it was not measured directly, this most likely reflects the effect of strong crystallographic preferred orientation, which can be qualitatively inferred from Fig. 4 and also Fig. 14, for the discussion on the general correlation between AMS and elastic wave anisotropy. Mica and amphibole single crystals also show considerable AMS and importantly their mean susceptibility is in the paramagnetic-dominated susceptibility range and hence their principal axes of susceptibility coincide with the crystallographic axes and CPO of a sample (e.g., Martín-Hernández and Hirt, 2003; Biedermann et al., 2015; Biedermann, 2018). In the case of mica, including biotite and muscovite, there is a relatively straightforward relationship between the principal axes of susceptibility and the crystallographic axes. The k_1 and k_2 axes are oriented within the basal plane of the mica, which hosts the $a[100]$ - and $b[010]$ - crystallographic axes, whereas the minimum susceptibility (k_3) is oriented normal to the basal plane, along the c -axis $[001]$. This petrofabric and AMS relationship makes it simple to compare with seismic anisotropy since there is an apparent direct agreement between physical properties and crystallography. Amphibole, in contrast, is more challenging to directly relate AMS and seismic anisotropy, because the highest susceptibility (k_1) and highest elastic stiffness (e.g., c_{11}) may not coincide. AMS in single-crystal amphiboles mainly depends on the chemical composition and location of the iron the amphibole (Biedermann, 2018). In addition, the problem with inclusions of iron-oxides within dia- and paramagnetic silicate grains will affect AMS (Lagroix and Borradaile, 2000; Feinberg et al., 2006), although with negligible influence on seismic anisotropy. The approach of using AMS as a proxy for seismic anisotropy must therefore be made cautiously, even in the qualitative case.

Fig. 14(a,c) shows the relationship between the degree of magnetic anisotropy and seismic anisotropy. Among the eight samples, six shows a moderate linear relationship between the degree of magnetic anisotropy [P_j (AMS)] and anisotropy of ultrasonic waves (AVp and AVs). The P_j (AMS) correlate better with $P_j(V_p)$ than $P_j(V_s)$. Between our samples, the best fits are recognized for the biotite bearing and amphibole-rich samples, 316 and 340 (Fig. 14a, c). Samples 248 and 309 do not follow the linear relationship and show lower seismic anisotropy, making it challenging to directly compare their AMS and seismic anisotropy (Fig. 14a, c). Based on the thermomagnetic curves (online supplementary material) it is observed that samples 248 contain magnetite (and other potential ferromagnetic minerals). Magnetite is very magnetic and even a small modal proportion ($\ll 1\%$) can dominate the magnetic susceptibility of rock; a fraction of magnetite can change the AMS by changing the shape (T_j) and degree of anisotropy (P_j) (Rochette et al., 1992). Such a small amount of magnetite will not influence the ultrasonic wave velocity, although it is very influential with regard to magnetic susceptibility. In contrast, sample 309 (felsic gneiss) shows a mean susceptibility that is close to zero (Fig. 9a, b). The high degree of anisotropy is a result of the low mean susceptibility (close to zero), and not the rocks' inherent magnetic anisotropy (Hrouda, 1986,

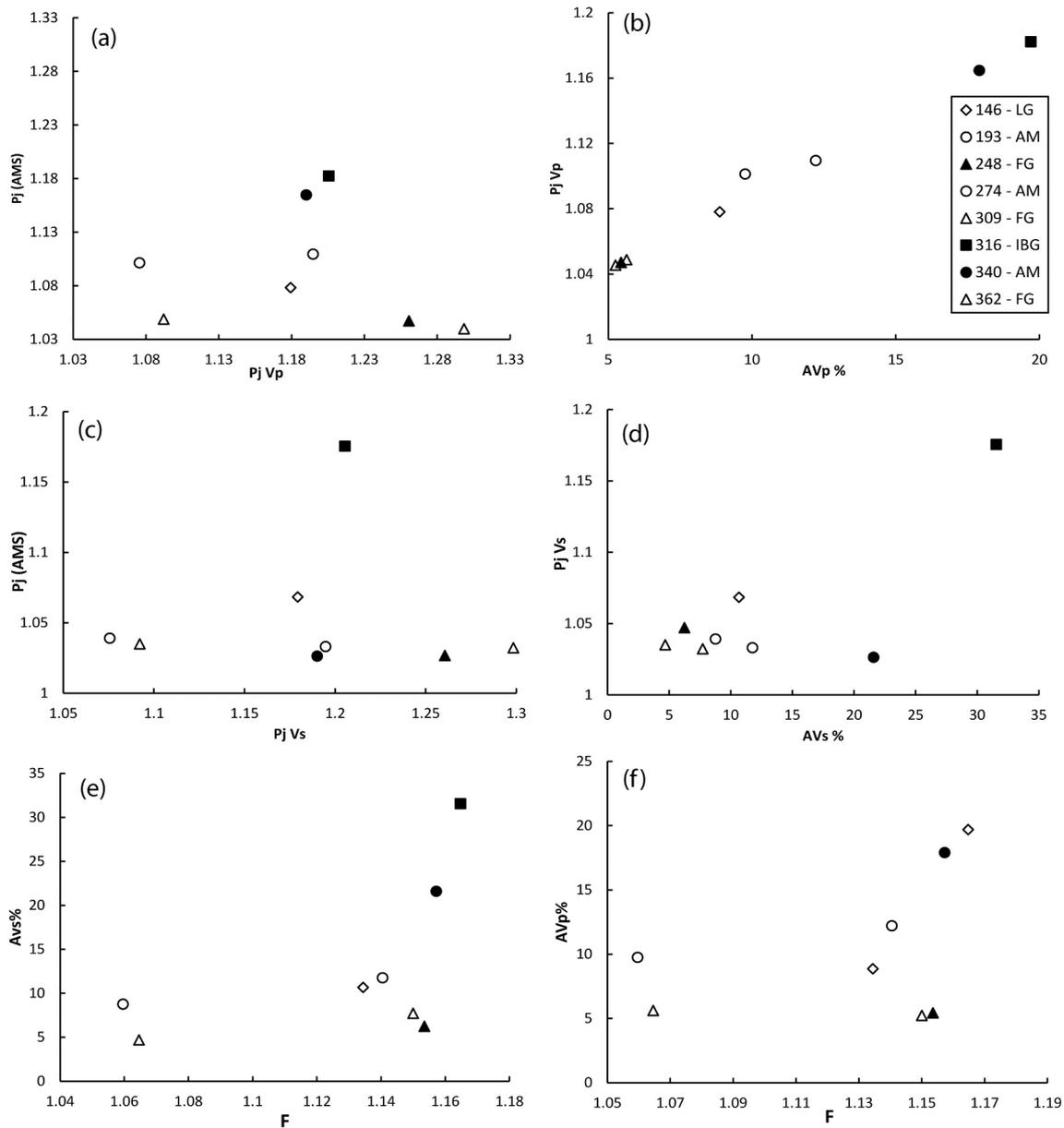


Fig. 14. A comparison of seismic anisotropy with magnetic susceptibility. a) degree of anisotropy of magnetic susceptibility compared to the degree of anisotropy of compressional wave anisotropy; b) degree of anisotropy of Vp compared to AVp; c) degree of anisotropy of magnetic susceptibility compared to the degree of anisotropy of shear wave anisotropy; d) degree of anisotropy of Vs compared to AVs; e) AVs as a function of magnetic foliation; f) AVp as a function of magnetic foliation.

2004). The layered felsic-mafic gneiss (sample 146) lies between the mafic and felsic gneiss samples.

To assess if the degree of anisotropy of the fabric directly influences the nature of the anisotropy of magnetic susceptibility, we do the following: 1) calculate the “foliation” of magnetic susceptibility (Eq. (10)); 2) plot the magnetic foliation relative to elastic wave anisotropy (AVp%, AVs%; Fig. 14e, f). P wave velocity data indicate a more linear relation, and straight-forward relationship, with AMS among samples in this study compared shear wave velocity measurements. The additional solutions available for the splitting of shear waves along the three sample axes is likely adding to the complexity of relating AVs to AMS.

With this study, we show that there is a relation between AMS and seismic anisotropy. However, this relation can be affected significantly by highly (ferro)magnetic minerals or a mean magnetic susceptibility close to zero. The existence of these highly magnetic minerals significantly increases the mean susceptibility and the absence of

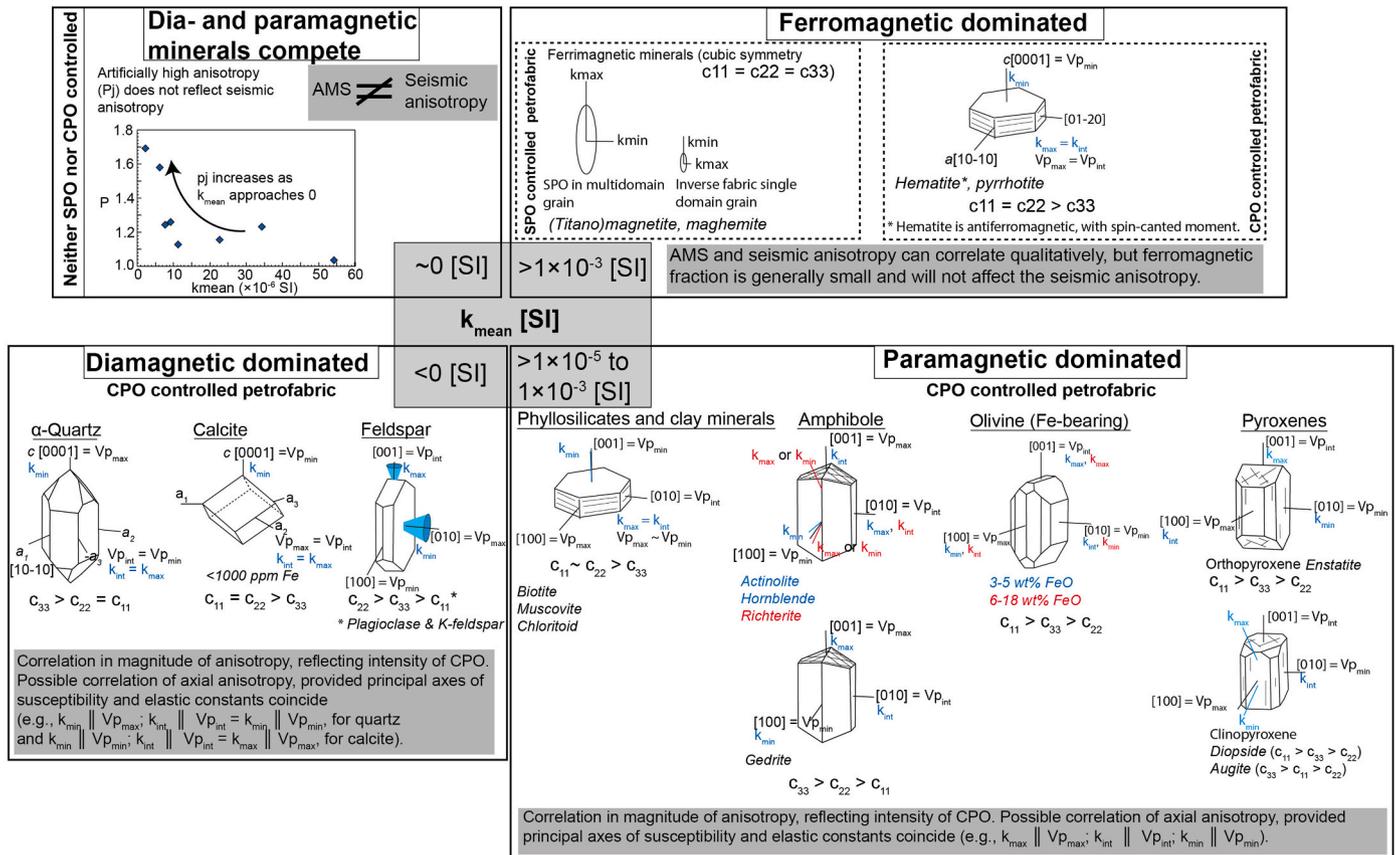
paramagnetic minerals. These effects cannot be seen in the ultrasonic data during laboratory measurements, simply because these highly magnetic minerals are not present in large enough volume to affect the elastic wave velocities. Additionally, a combination of paramagnetic and diamagnetic minerals can lead to a susceptibility close to 0 [SI]. A rock that consists of such a combination of minerals, with mean magnetic susceptibility close to zero, is not suitable as a proxy for seismic anisotropy. Samples that do not contain ferromagnetic minerals in appreciable amounts, with mean susceptibility between 1×10^{-5} to 1×10^{-3} SI appear to be the most suitable samples to study the seismic anisotropy and AMS relationship, giving rise to a relatively linear relation between AVp and the degree of magnetic anisotropy (Pj). However, caution is needed when relating the orientation of principal axes of susceptibility and orientation of maximum and minimum Vp. It should be noted that there is a systematic challenge in comparing AMS with elastic wave speed data as AMS has a large range of values (varying over

several orders of magnitude) compared with the range of elastic wave velocities which ranges well within one order of magnitude; consequently, any observed trends are thus tentative, although potential exists with careful analysis to correlate AMS and seismic anisotropy through the petrofabric.

Fig. 15 shows a flow chart that aims to indicate how seismic anisotropy (and elastic constants) relate to magnetic susceptibility and the principal axes of susceptibility. Note that this diagram presents the case for selected minerals in the crust, and not an exhaustive attempt to relate seismic anisotropy and AMS for all minerals. It serves with the aim to illustrate what needs to be taken into consideration when making a comparison between seismic anisotropy and AMS. In some cases, the relationship between the two physical properties is straightforward, such as for phyllosilicate dominated rocks, where maximum axes of Vp and k coincide. Provided that such rocks are governed by the matrix minerals (paramagnetic and diamagnetic cases), there should also be a reasonable relationship between the degree of anisotropy, as both properties are governed by CPO. The relationship breaks down for cases when susceptibility is close to 0 [SI], as the calculation of Pj results in artificially high values. In addition, for strongly magnetic rocks, where $k_{mean} > 1 \times 10^{-3}$ [SI], it is likely that the ferromagnetic mineral fraction will overtake the contribution to AMS, making a direct comparison

between AMS and seismic anisotropy difficult. A qualitative relationship may still exist, however, if there is a coincidence between the SPO (magnetite)/CPO (hematite, pyrrhotite) of the ferromagnetic phases compared to the matrix phases that control seismic anisotropy.

When the magnetic susceptibility is dominated by matrix minerals, such as in the diamagnetic and paramagnetic cases in Fig. 15, is necessary to consider 1) the magnitude or degree of anisotropy and 2) the directional relationship between principal axes of k and Vp. In regards to the magnitude of anisotropy, there should be a good relationship between seismic and magnetic anisotropy, regardless of the orientation of the principal axes, since it is only the strength of the petrofabric (CPO in this case) that is of relevance. However, in regards to the orientation of the principal axes of seismic and magnetic anisotropy, the relationship is considerably more complicated (Fig. 15). For instance, as noted above, the principal axes agree well in regards to direction in rocks dominated by phyllosilicates. In contrast, the principal axes in amphibole dominated rocks show a complicated relationship, with many possibilities for orientation of the principal axes of magnetic susceptibility (even though the elastic constants show regular directional properties). This is likely the reason why we observe a good relationship between the degree of susceptibility anisotropy (Pj) and AVp (%) (Fig. 14a, with exception of felsic gneiss samples 248 and 309), but a more unclear relationship



between principal axes of Vp and k (Fig. 10).

6. Conclusion

We conducted laboratory measurements to investigate the effects of pressure and temperature on ultrasonic Vp and Vs for amphibolite and gneisses, which are representative of a middle crustal orogenic setting. AMS measurements were made on the same samples to evaluate the relation between AMS and seismic anisotropy. The results and points in the discussion are summarized in the following remarks:

- Results in this study illustrate how micro-cracks affect the (elastic) deformation of samples as well as enhancing anisotropy at pressures from ambient condition to 600 MPa and temperature from ambient conditions to 600 °C. Therefore, laboratory wave velocities measured at confining pressures below 200 MPa may be significantly influenced by microcracks. However, even above 200 MPa, there is likely a contribution of microcracks to elastic anisotropy. In particular, when temperature matches confining pressure (i.e., ~ 1 MPa/1 °C), there could be a considerable influence of microcracks on seismic anisotropy, up to 20% at mid-crustal conditions (600 MPa and 600 °C).
- Seismic measurements confirm strong relations of velocity anisotropy, shear wave splitting and shear wave polarization to the structural grain of the rocks (foliation, lineation).
- Ultrasonic laboratory measurements of COSC-1 samples, from this and previous studies, illustrate that seismic anisotropy is an important parameter that affects our interpretation of Caledonian orogenic processes and crustal dynamics. These results may be applied to other orogen settings with ongoing collisions, such as the Himalaya-Tibet orogen, where significant anisotropy is observed in the mid and lower crust.
- Amphibolites and biotite bearing specimens are significantly anisotropic with AVp ranging from 8.9% to 19.7%, and maximum AVs exceeding 10%. Felsic gneisses are considerably less anisotropic, with AVp of <7% and AVs of <6%.
- Among the eight studied samples, only one sample (309, felsic gneiss) showed a reduction in P-wave and S-wave anisotropy as a function of increasing temperature.
- There is a general correlation between seismic anisotropy and AMS in samples where paramagnetic minerals control the AMS. Added complication is introduced by the presence of ferromagnetic minerals in the felsic gneisses (sample 248) or artificially high degree of anisotropy generated in samples with k_{mean} close to zero (felsic gneiss, sample 309).
- To better understand the relationship between seismic anisotropy and AMS, we constructed a flow chart that describes how these two properties relate in terms of mean (bulk) susceptibility and mineral composition. The aim of this chart is to raise awareness and improve the interpretability of AMS data, used as a petrofabric indicator and indirectly as an indicator for seismic anisotropy.

Credit author statement

Mohsen Bazargan (MB), Bjarne Almqvist (BA), Sandra Piazzolo (SP), Hem Motra (HM) and Christoph Hieronymus (CH) designed and developed the concept for the study together. Samples were collected by MB and BA. Laboratory ultrasonic and magnetic measurements were made by MB, HM and BA at Christian Albert University of Kiel and Uppsala University. MB, BA and SP prepared the first draft of the manuscript. All authors contributed to the editing of the manuscript.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence

the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.tecto.2021.229113>.

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