Elastic anisotropies of rocks in a subduction and exhumation setting

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Abstract. Subduction and exhumation are key processes in the formation of orogenic systems across the world, for example, in the European Alps. For geophysical investigations of these orogens, it is essential to understand the petrophysical properties of the rocks involved. These are the result of a complex interaction of mineral composition and rock fabric
10   including mineral textures (i.e. crystallographic preferred orientations). In this study we present texture-derived elastic anisotropy data for a representative set of different lithologies involved in the Alpine orogeny. Rock samples were collected in the Lago di Cignana area in Valtournenche, in the Italian Northwestern Alps. At this locality a wide range of units of continental and oceanic origin with varying paleogeographic affiliations and tectono-metamorphic histories are accessible. Their mineral textures were determined by time-of-flight neutron diffraction at the Frank Laboratory of Neutron Physics at
15   the JINR in Dubna, Russia. From these data the elastic properties of the samples were calculated. The data set includes representative lithologies from a subduction-exhumation-setting. In subducted lithologies originating from the oceanic crust, the elastic anisotropies range from 1.4 to 5.0% with average P-wave velocities of 7.01-8.24 km/s and VP/VS-ratios of 1.71-1.76. In the metasediments of the former accretionary prism the elastic anisotropies range from 4.7 to 8.2 %. This tectonic setting displays average P-wave velocities of 6.47-7.23 km/s and VP/VS-ratios of 1.60-1.76. Continental crust which is
20   incorporated in the collisional orogen shows elastic anisotropies ranging from 1.8 to2.8 % with average P-wave velocities of 6.42-6.51 km/s and VP/VS-ratios of 1.56-1.60. Our results suggest that mafic and felsic rocks in subduction zones at depth may be discriminated by a combination of seismic signatures: lower anisotropy and higher VP/VS ratio for mafic rocks, higher anisotropy and lower VP/VS ratio for felsic rocks and metasediments.

1 Introduction

25   During subduction, exhumation and continental collision rocks of different paleogeographic affiliations are brought together, strongly deformed and submitted to variable metamorphic conditions. Oceanic crust and its sedimentary cover, as well as continental fragments are frequently subducted before or during continental collision. This leads to a large variety of lithologies within the subduction zone from ultramafic rocks originating from mantle material, over mafic rocks originating from oceanic crust to felsic rocks which were either part of continental fragments or sedimentary cover of the oceanic rocks.
30   Deformation during subduction and exhumation causes the formation of crystallographic preferred orientations (CPO),

which in turn leads to elastic anisotropy of the rocks involved. The metamorphic overprint also influences their seismic velocities and Poisson's ratios. With increasing depth, the rocks undergo blueschist- to eclogite-facies metamorphic conditions. During exhumation the rocks are frequently exposed to further metamorphic overprint, depending on various factors such as exhumation speed and available water content (e.g., Peacock, 1993). These rocks which can now be found in
35 outcrops at the Earth's surface yield insights into the variable stages of subduction, collision, and exhumation, which can be compared to the geophysical investigations of such rocks at depth.

Such high-resolution geophysical imaging of 3D structures is currently taking place within the AlpArray initiative using a high-end seismological array within the Alpine orogen. For this project, knowledge of the elastic anisotropy of the imaged rocks is of great interest, because in conjunction with the VP/VS-ratio it is a key parameter for the interpretation of
40 geological structures in the subsurface. Receiver-function analysis allows separating crustal from mantle anisotropy in the Alps (Link and Rümpker, 2019). Because of the larger lithological heterogeneity of the crust, as compared to the mantle, information on the seismic properties of various crustal rocks is required for interpreting the crustal anisotropy pattern.

Elastic anisotropies of rocks can be calculated from mineral composition, single-crystal mineral properties, and the CPO of all involved mineral phases (Christoffel, 1874; Mainprice and Humbert, 1994). They are also influenced by other factors,
45 such as shape preferred orientation (SPO) of the minerals, their grain size, and microcracks in the rocks (Siegesmund and Vollbrecht, 1991; Siegesmund et al., 1993; Ullemeyer et al., 2011). With increasing depth, porosity and microcracks are closed, leading to CPO as the most important factor influencing elastic anisotropies in the deeper parts of subduction zones and collisional orogens.

In this study, we collected a set of rock samples from the Lago di Cignana area in Valtournenche, western Alps (figure 1a
50 and 1b). The samples are representative of various stages within a subduction- and exhumation-cycle, including rocks of oceanic and continental origin and variable degrees of metamorphism and deformation (figure 2). The CPOs of the mineral phases of the samples were acquired by time-of-flight (TOF) neutron diffraction. Measurements were conducted at the SKAT diffractometer of the Frank Laboratory of Neutron Physics at the JINR in Dubna, Russia (Ullemeyer et al., 1998, 2006, 2010; Ivankina et al., 2000; Ivankina et al., 2001; Nikitin et al. 2004; Kern et al., 2008; Keppler et al., 2014). This
55 method has the advantage of being able to measure bulk CPOs of all mineral phases in large, polyphase, coarse-grained samples simultaneously (Keppler et al., 2014). From the CPO and single-crystal elastic anisotropy data the seismic properties of all collected samples were calculated. These data provide a catalogue for seismic imaging and geodynamic models of the Alpine orogen as well as other subduction zones and continental collision zones around the world. Among the texture-derived properties presented are the P- and S-wave velocities, the VP/VS-ratio and the elastic anisotropies of each
60 sample.

2. Geological overview of the study area

The samples in this study were collected in the western Alps in northwestern Italy (figure 1a). In this area the Alpine orogen consists of tectonic nappes of continental and oceanic origin with various paleogeographic provenances and tectono-metamorphic histories. This nappe stack formed from continent-continent-collision of Europe and Adria occurring in the Palaeogene. The units in this nappe stack were accreted to the Adriatic margin during SE-directed subduction. The units from which the samples were collected are derived from the Piemont-Ligurian ocean and the Margna-Sesia continental fragment (Schmid et al, 2004), also termed Cervinia (Pleuger, et al., 2007).

At Lago Di Cignana (figure 1b), a cross-section through continental and oceanic units is very well accessible. Generally, the units exhibit a shallowly NW dipping foliation and SE-NW to E-W trending lineation (figure 1c). The oceanic Zermatt-Saas zone is found in the footwall, separated from the continental hanging wall unit of the Dent Blanche Nappe by the oceanic and metasedimentary Combin zone. These three units are separated by shear zones and faults and clear jumps in metamorphic grade can be found. For this study, metabasites (eclogite, blueschist, amphibolite and greenschist) were collected from the Zermatt-Saas zone. Furthermore, several metasedimentary rocks of varying composition were collected from the Combin zone as well as two gneiss samples from the Dent Blanche Nappe.

The rocks in this study followed different P-T paths. The Dent Blanche unit has experienced P-T conditions of 1.4-1.6 GPa and 450-520 °C (Ballèvre et al., 1986, Oberhänsli and Bucher, 1987; Angiboust et al., 2014; Manzotti et al., 2014). The blueschist-facies metamorphism was dated to 48-43 Ma by Rb-Sr geochronology (Angiboust et al., 2014) in the Arolla mylonites at the base of the Dent Blanche unit.

The Combin zone displays P-T conditions of 1.2 GPa and 450°C (Bousquet, 2008). The blueschist- to greenschist-facies metamorphism in the Combin zone has been dated between 48 and 36 Ma (Reddy et al., (1999, 2003)) by Rb-Sr and Ar/Ar.

In the Zermatt-Saas zone the P-T path involved prograde blueschist-facies conditions, HP to UHP peak conditions and retrograde amphibolite- and greenschist-facies overprints. Peak conditions reached 2.5-3 GPa and 550-600 °C (Bucher et al., 2005), locally UHP peak conditions of 3.2 GPa and over 600 °C (Groppo et al., 2009; Frezzotti et al., 2011). The UHP peak was dated to 40.6 ± 2.6 Ma by Sm-Nd garnet (Amato et al., 1999), to 44-43 Ma by the Ar/Ar step-heating method on phengite (Gouzu et al., 2006) and to 48.8 ± 2.1 Ma by Lu-Hf on garnet-omphacite-whole-rock (Lapen et al., 2003). The greenschist-facies retrograde overprint was dated to 38 ± 2 Ma by Rb-Sr whole-rock-phengite chronology (Amato et al., 1999).

3. Methods

CPOs were measured by the TOF neutron diffractometer SKAT at the Frank Laboratory of Neutron Physics at JINR in Dubna, Russia (Keppler et al., 2014; Ullemeyer et al., 1998). Neutron diffraction has several advantages over other texture analysis methods such as optical measurement and electron and x-ray diffraction analysis (Wenk et al. 1998; Ullemeyer et al. 2000; Nikitin and Ivankina, 2004). Among these advantages is the low absorption of thermal neutrons in matter. This allows the measurement of large sample volumes (up to 65 cm3) of complex material such as coarse-grained rocks containing

multiple low-symmetry mineral phases (Ullemeyer et al., 2006; Ullemeyer et al., 2010; Ivankina et al., 2005; Kern et al., 2008; Keppler et al., 2014).

95   The diffraction data are processed by Rietveld Texture Analysis (RTA) (Von Dreele, 1997; Matthies et al., 1997, Lutterotti et al., 1997). This method allows for the determination of the textures of all mineral phases in the sample by calculation of the orientation distribution function (ODF) using the E-WIMV algorithm (Lutterotti et al. 2004, Chateigner et al. 2019). The RTA is performed in the software Mineral Analysis Using Diffraction (MAUD: Lutterotti, 2010, Wenk et al., 2010). For further discussion on the functions and limitations of MAUD, see the publication of Wenk et al. (2012). For RTA,

100   knowledge of the mineral assemblage is of great importance. For this reason, thin sections of all samples were made and analyzed by polarization microscopy.

To calculate the P-wave velocities and other elastic properties of the samples the program BEARTEX (Wenk et al. 1998) was used. To calculate the bulk-rock elastic moduli, the orientation distribution function (ODF) of the mineral phases, knowledge of their single-crystal elastic constants and their volume fractions is necessary. The elastic moduli were then

105   calculated using the Christoffel equation (1) (Christoffel, 1874; Mainprice and Humbert, 1994):

$$Det\left|\Gamma_{ij} - V^2\delta_{ij}\right| = 0 \qquad (1)$$

In it the phase velocity of P-, S1- or S2-waves is represented by V, the Kronecker delta by δ_{ij} and the Christoffel Tensor by Γ_{ij}. The Equation of the Christoffel Tensor (2) is as follows:

$$\Gamma_{ij} = a_{ijkl}x_k x_1 \qquad (2)$$

110   In it a_{ijkl} represents the density-corrected elastic (stiffness) moduli and $x_k x_1$ represent the direction cosines of the wave propagation direction.

For the averaging, it has been shown by Hill (1952) that the widely used averaging schemes of Voigt (1928) and Reuss (1929) give upper and lower bounds of the elastic moduli, respectively. Furthermore, other averaging schemes have been proposed by Matthies and Humbert (1995) and Matthies (2012). However, since the values of the parameters a_{ijkl} are still

115   uncertain, the Voigt averaging scheme was used consistently throughout this study. It is known that the recalculated velocities are therefore maximum velocities only.

The ODF of all phases are exported from MAUD. The volume percentages of the phases are estimated from thin-sections and calculated in MAUD by RTA. The single-crystal elastic constants were taken from literature (quartz: Heyliger et al., 2003; albite: Brown et al., 2006; muscovite: Vaughan and Guggenheim, 1986; calcite: Dandekar, 1968; dolomite: Humbert

120   & Plique, 1972; hornblende: Aleksandrov and Ryzhova, 1961; epidote/zoisite/clinozoisite: Aleksandrov et al., 1974; garnet: Zhang et al., 2008; glaucophane: Bezacier et al., 2010; omphacite: Bhagat et al., 1992). In some cases, single-crystal elastic constants are not available for all phases (chlorite, actinolite, barroisite and clinozoisite), so similar minerals were selected as approximations. For samples containing chlorite we have selected the single-crystal elastic constant of muscovite (Vaughan and Guggenheim, 1986) to approximate chlorite. For clinozoisite the single-crystal elastic constants of epidote (Aleksandrov

125   et al., 1974) were selected, for the blueschist-facies amphibole barroisite those of glaucophane (Bezacier et al., 2010) and for

actinolite those of hornblende (Aleksandrov and Ryzhova, 1961). In most cases, the contents of the phases for which approximations were necessary, were below 10% of the total sample volume. At this low content, the CPOs of these minerals are of low relevance for the physical properties of the rock (Mainprice and Ildefonse, 2009). In RTA, amphiboles are generally very difficult to distinguish, since they are extremely similar crystallographically. For this reason, the

130 crystallographically best fitting amphibole structure was picked for refinement, which is not always the chemically fitting one in the sample. This has an insignificant influence on the calculated CPO due to the crystallographic similarity of the amphibole group (C2/m space group; Reynard et al. 1989).

It is known that at low confining pressures, close to the Earth's surface, the elastic properties of rocks are greatly influenced by microcracks and crack fabrics, as well as a shape preferred orientation (SPO). This has been shown in multiple studies

135 (Kern et al., 2002, 2008, Pros et al., 2003, Ivankina et al., 2005, Ullemeyer et al., 2006). Further studies have even shown that the influence of microcracks is still visible up to a confining pressure of 1 GPa (Christensen, 1974, Ullemeyer et al., 2011). In this study the elastic properties are calculated solely on the basis of mineral texture and single-crystal data without taking microcracks into account. Since the microcracks are closed at great depth, the calculated elastic properties can be seen as representative of these conditions.

4. Composition and microfabrics of the samples studied

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In this study we have related the subduction and exhumation settings to several structural positions, each containing characteristic lithologies (figure 2). These positions are (1) the oceanic crust of the subducting plate, (2) the sedimentary cover of the subducting plate as well as the accretionary prism and finally (3) the continental crust of a microcontinent. The eclogite, blueschist, amphibolite and greenschist of the Zermatt-Saas zone belong to the first of these. The metasediments of

145 the Combin zone represent the second position and finally the gneisses of the Dent-Blanche represent the position of the deformed continental crust. However, the samples do not always represent the peak conditions of metamorphism but contain minerals grown during various steps of the P-T path. In most samples presented here a clear chlorite-forming greenschist overprint is visible, even if the rock originated at higher metamorphic conditions. This is especially true for the metabasic rocks in this study, which often contain hornblende, actinolite and chlorite, indicating the amphibolite- and greenschist-facies

150 overprint, respectively.

4.1 Eclogite

The eclogite (MJS17) is composed of omphacite (30 vol.-%), garnet (30 vol.-%), clinozoisite (15 vol.-%), glaucophane (10 vol.-%), albite (10 vol.-%), and rare white mica (< 5 vol.-%), and retrograde chlorite (< 5 vol.-%) (table 1, figure 3a). Mesoscopically the fabric is strongly dominated by a clear foliation defined by omphacite and glaucophane, surrounding

155 larger grains of garnet. Elongated omphacite and glaucophane grains form the stretching lineation and are visible in the foliation plane. On the microscopic level the eclogite is strongly foliated with idiomorphic garnet grains surrounded by

layers of omphacite, glaucophane, clinozoisite and feldspar. The grainsize distribution is bimodal with a mostly fine-grained matrix of omphacite, glaucophane and some feldspar surrounding larger garnet grains. The presence of albite shows a greenschist-facies overprint.

4.2 Blueschist

The blueschist (MJS16) comprises barroisite (40 vol.-%), albite (30 vol.-%), clinoziosite (15 vol.-%), garnet (10 vol.-%) and epidote (5 vol.-%). The foliation in the blueschist is clearly pronounced on the mesoscopic scale (table 1). The stretching lineation is formed by elongated barroisite and feldspar grains in the foliation plane. On the microscopic scale a foliation formed by barroisite can clearly be seen. The grain-size distribution is bimodal, with a finer grained matrix of amphibole and feldspar, and larger, strongly retrogressed, inclusion-rich and fractured garnets therein.

4.3 Amphibolite

The composition of the amphibolite (MJS26; table 1) is dominated by hornblende (40 vol.-%). Feldspar (25 vol.-%), clinozoisite (15 vol.-%), and chlorite (10 vol.-%) are also present in the matrix. Some large, fractured garnet (5 vol.-%) grains are found as accessory minerals, which display strong greenschist-facies retrogression, with abundant chlorite surrounding them. Hornblende defines a weak foliation in MJS26. Both hornblende and feldspar grains form a stretching lineation on the foliation plane.

4.4 Greenschist

In the greenschist MJS36, the main constituent minerals are chlorite (30 vol.-%), clinozoisite (30 vol.-%), and actinolite (20 vol.-%). Feldspar is visible throughout the matrix (table 1, figure 3b). Actinolite defines the foliation of the sample, while chlorite grains show a weaker preferred orientation. In the foliation, amphibole grains can be seen forming a stretching lineation. The greenschist is fine-grained and shows a very homogeneous mineral distribution.

4.5 Metasediments

The composition of the metasediments (MJS18, MJS20, MJS22) is variable (table 1), but they all display a clear foliation and stretching lineation (figure 3c). In sample MJS18 the matrix consists of dolomite (60 vol.-%), quartz (20 vol.-%), calcite (10 vol.-%) and white mica (10 vol.-%). The foliation is pronounced and defined by dolomite and calcite. Layers of finer and coarser grained calcite, dolomite and quartz alternate. Mica fish are found interlayered in the matrix. In the layers of coarser grained calcite, dolomite and quartz, a weak oblique foliation, in addition to the main foliation can be seen. The stretching lineation is visible on the foliation plane, in elongated calcite and dolomite grains. In the calc-micaschist (MJS20) a clear foliation is defined by white mica. Alternating layers of coarser and finer grained calcite (15 vol.-%), and quartz (40 vol.-%) are separated by bands of white mica (30 vol.-%) and chlorite (10 vol.-%). Larger grains of strongly fractured garnet (5 vol.-%) are present. These garnet grains give the fabric a knotty appearance, with the other minerals flowing around them. The

calcschist MJS22 is made up of quartz (55 vol.-%), calcite (30 vol.-%) and chlorite (15 vol.-%). Compositional layering is clearly visible, with thick bands of very coarse-grained quartz, alternating with bands of fine-grained quartz and calcite with chlorite fishes. The coarser grains of both quartz and calcite form a weak oblique foliation.

4.6 Gneiss

The composition of the two gneisses (MJS34, MJS35) is quite similar with the main mineral components in both being quartz (40-57 vol.-%), feldspar (27-35 vol.-%), white mica (7-15 vol.-%), and chlorite (9-10 vol.-%) (table 1, figure 3d). The fabric of both samples is also very similar. They are strongly foliated with a lineation formed by quartz and feldspar. Sample MJS35 contains a 5 mm thick quartz band, which is a feature quite common in this unit in general. The other sample has a more homogeneous distribution of minerals. Microscopically the foliation is formed by mica, quartz, and feldspar. Mica shows a pronounced SPO of mineral fishes aligned in the foliation plane. Where chlorite is present it is interwoven with white mica between the quartz and feldspar layers.

5. Results

5.1 Crystallographic preferred orientations

5.1.1 Eclogite

In the eclogite (MJS17) omphacite shows a distinct CPO and has the highest texture strength of all phases in the sample with an F2 index (Bunge, 1982; see complete list in appendix A) of 3.49. The c-axes display the highest intensity clusters with a maximum parallel to the stretching lineation. The a-axes display maxima in a girdle perpendicular to the stretching lineation. The omphacite b-axes form a weak girdle perpendicular to the stretching lineation with a clear maximum normal to the foliation plane (figure 4). The glaucophane in eclogite has a weak CPO with an F2 index of 1.19. The c-axes display a maximum parallel to the stretching lineation. The a-axes develop a girdle perpendicular to the stretching lineation with a maximum paralleling the kinematic y-axis. The glaucophane b-axes display two maxima, the stronger of which being normal to the foliation plane and a secondary maximum parallel to the stretching lineation (figure 4). Clinozoisite has weak textures with an F2 index of 1.01. The c- axes display a maximum paralleling the kinematic y-axis. The main b-axes maximum is parallel to the stretching lineation (figure 4). The texture of white mica in the eclogite is quite weak (F2 index of 1.02). However, the basal planes show an alignment to the foliation plane (figure 4). Garnet and albite show random textures.

5.1.2 Blueschist

In the blueschist (MJS16), barroisite displays the strongest textures with an F2-index of 2.36. The strongest textures are seen in the c-axes, which exhibit a maximum parallel to the stretching lineation. The a-axes in barroisite display maxima in a girdle distribution perpendicular to the lineation direction with a maximum parallel to the kinematic y-axis. The b-axes

display maxima in a similar girdle perpendicular to the stretching lineation with highest densities normal to the foliation plane (figure 4). Clinozoisite and epidote CPOs are very similar and relatively weak. Both show an F2 index of 1.02 and b-axes with a maximum almost parallel to the stretching lineation. A-axes and c-axes show girdles perpendicular to the lineation (figure 4). Garnet and albite exhibit random textures.

5.1.3 Amphibolite

In the amphibolite MJS26, the strongest textures are found in the hornblende, with an F2 index of 1.62. The hornblende c-axes show the strongest CPO with a distinct maximum parallel to the stretching lineation. The b-axes are distributed in a girdle normal to the stretching lineation, with a maximum normal to the foliation plane. The a-axes also display a girdle normal to the stretching lineation (figure 5). For chlorite, the F2 index is 1.10. Chlorite c-axes show a girdle perpendicular to the lineation with a maximum parallel to the foliation normal (figure 5). In albite the texture strength is low, with an F2 index of 1.07. However, a-axes display a girdle perpendicular to the stretching lineation and a weak maximum normal to the foliation plane while b-axes display a broad, weak maximum parallel to the stretching lineation (figure 5). The textures in garnet and clinozoisite are nearly random.

5.1.4 Greenschist

In the greenschist MJS36, the strongest textures are found in chlorite with an F2 index of 1.34. Chlorite basal planes are oriented parallel to the foliation plane (figure 5). The textures of actinolite are of intermediate strength with an F2 index of 1.21. The c-axes display a strong maximum parallel to the stretching lineation and a weak girdle distribution in the foliation plane. The a-axes show highest intensities with a maximum normal to the foliation plane. The b-axes display a slightly weaker texture with a maximum parallel to the foliation normal, connected by a weak, irregular girdle normal to the stretching lineation. Furthermore, secondary maxima are slightly oblique to the stretching lineation (figure 5). Albite has an F2 index of 1.18 and a-axes display the strongest CPO with a maximum parallel to the foliation normal. The b-axes are concentrated in a girdle in the foliation plane with weak maxima therein. The c-axes display a weak CPO with diffuse maxima (figure 5). Clinozoisite shows a random CPO.

5.1.5 Metasediments

In the metasediment samples (MJS18, MJS20, MJS22) the CPOs are quite distinct and often stronger than in the samples described above.

In the dolomitic micaschist MJS18, the textures of muscovite are strongest with an F2 index of 2.18. Its basal planes display a strong alignment with the foliation plane (figure 6). The textures of dolomite are of intermediate strength with an F2 index of 1.14. The c-axes display a maximum normal to the foliation plane. The a-axes show a girdle distribution in the foliation plane with a maximum parallel to the stretching lineation (figure 6). In calcite the textures are slightly stronger than in dolomite and have an F2 index of 1.28. The c-axes display a strong maximum perpendicular to the foliation plane and a-axes

display a girdle distribution in the foliation plane (figure 6). The CPO of quartz is weak with an F2 index of 1.04. The c-axes display small girdles around the foliation normal. A-axes show a weak diffuse maximum oblique to the stretching lineation. Prism and rhomb planes display almost random distributions (figure 6).

250   In the calcareous micaschist (MJS20), muscovite and chlorite show strong textures with F2 indices of 2.83 and 2.17, respectively. Both minerals yield a strong alignment of their basal planes in the foliation (figure 6). Calcite textures are weak, with an F2 index of 1.08. The c-axes display a maximum normal to the foliation and the a-axes are concentrated in a girdle in the foliation plane (figure 6). The texture of quartz in is of intermediate strength with an F2 index of 1.14. The c-axes display small circle distributions around the foliation normal, while the a-axes display two maxima at the periphery,

255   each at low angle to the stretching lineation. [01-11] shows one maximum normal to the foliation and a second set of maxima in the foliation plane. [10-10] displays a diffuse maximum nearly parallel to the stretching lineation, while [10-11] displays maxima clustered around the foliation normal and in the foliation plane parallel to the y-direction. Garnet shows a random CPO.

In the calcschist MJS22 all mineral phases display medium texture strength (appendix A). In calcite the F2 index is 1.34. The

260   c-axes display a strong maximum parallel to the foliation normal. A-axes are concentrated in the foliation plane with a maximum parallel to the stretching lineation (figure 6). In quartz the textures are stronger than in calcite, with an F2 index of 1.54. The quartz c-axes display a weak sinistrally rotated girdle with a strong maximum offset from the kinematic y-axis by ca. 30°. A weaker density cluster is found perpendicular to this girdle. The a-axis maximum is located close to the periphery of the pole figure slightly sinistrally rotated away from the x-direction. The poles of the prism and rhomb planes display

265   weaker maxima in girdles paralleling the c-axes girdle (figure 6). Chlorite has an F2 index of 1.52 and displays a strong alignment of its basal plane in the foliation (figure 6).

5.1.6 Gneiss

In the gneisses (MJS34, MJS35) CPOs are similar to those of the metasediments (figures 6 and 7).

In MJS34 the muscovite and chlorite display basal planes parallel to the foliation plane. The textures are strongest for

270   chlorite with an F2 index of 2.15. Muscovite has an F2 index of 1.27. Quartz shows an F2 index of 1.18 and its c-axes are distributed in small circles around the foliation normal. The a-axes display a weak and diffuse girdle distribution in the foliation plane. In albite a very weak (F2 index of 1.03) but distinct CPO can be seen. The a-axes display a maximum parallel to the stretching lineation (figure 7).

In MJS35 both muscovite and chlorite display clear alignment of the basal planes in the foliation plane. Their F2 indices are

275   1.13 and 1.53, respectively. The CPOs in quartz are weaker (F2 index of 1.06) than in MJS34. C-axes display an asymmetric crossed girdle distribution. The a-axes display diffuse intensity clusters at the periphery around the stretching lineation. In albite the CPO is nearly random (F2 index = 1.01), with a faint maximum of the a-axes in the lineation direction (figure 7).

5.2 Elastic anisotropy

P-wave anisotropies for the various lithologies were calculated from the ODFs of constituent mineral phases, single-crystal
280   data of the respective phase and their volume fraction within the sample using the software BEARTEX (Wenk et al., 1998).
Graphical representations of S-wave velocity distributions are not shown in this study, since the differences between
maximum and minimum velocities in a sample are on the order of 0.01 km/s and deemed insignificant.

5.2.1 Eclogite

In the eclogite (MJS17) the average P-wave velocity (VP) is 8.2 km/s and the average S-wave velocity (VS) is 4.8 km/s
285   resulting in a VP/VS-ratio of 1.72. The elastic anisotropy is 1.4%, which is defined by omphacite (30 vol.-%, table 1) and
glaucophane (10 vol.-%, table 1). The fast velocities correlate with the c-axis maximum of omphacite as well as of
glaucophane parallel to the stretching lineation. The lowest P-wave velocities are found parallel to the foliation normal and
intermediate velocities can be found parallel to the kinematic y-axis (figures 4 and 8).

5.2.2 Blueschist

290   In the blueschist the average VP is 7.6 km/s, with a maximum of 7.8 km/s and minimum of 7.5 km/s. The average VS is 4.5
km/s, resulting in a VP/VS-ratio of 1.71. The P-wave anisotropy is of medium strength with 3.7 % and defined by barroisite
(40 vol.-%, table 1) and clinozoisite (15 vol.-%, table 1). The distribution of the fast velocities matches the orientation of the
barroisite c-axes and the clinozoisite b-axes. Lowest P-wave velocities can be found in a girdle distribution perpendicular to
the stretching lineation (figures 4 and 8).

295   ### 5.2.3 Amphibolite

The amphibolite has an elastic anisotropy of 3.1 %. The VP range from 7.0 km/s to 7.3 km/s, with an average of 7.1 km/s.
The average VS is 4.0 km/s, resulting in a VP/VS-ratio of 1.76. Maximum VP are distributed parallel to the stretching
lineation, with a girdle of low velocities in the kinematic y-z-plane, perpendicular to the stretching lineation (figure 8). The
maximum velocity correlates well with the c-axes of hornblende (40 vol.-%, table 1) in the amphibolite (figures 5 and 8).

300   ### 5.2.4 Greenschist

In the greenschist MJS36, the average VP is 7.0 km/s, with a maximum of 7.2 km/s and a minimum of 6.8 km/s. The average
VS is 4.1 km/s, resulting in a VP/VS-ratio of 1.73. The P-wave anisotropy is high with a value of 5 %. The highest VP are
found in the foliation plane, with a maximum parallel to the stretching lineation. Lowest VP are found parallel to the
foliation normal (figure 8). The highest velocities are parallel to the actinolite (20 vol.-%, table 1) c-axes and distributed
305   similarly to the chlorite (30 vol.-%, table 1) basal planes (figures 5 and 8).

5.2.5 Metasediments

The metasedimentary schist samples (MJS18, MJS20 and MJS22) show a VP of 6.5-7.2 km/s and a VS of 4.0-4.1 km/s. The P-wave anisotropies range from 4.7-8.2 % and VP/VS-ratios average at 1.66 (figure 8).

In the dolomitic micaschist (MJS18) the average VP of 7.2 km/s is high for a metasedimentary sample. The VP anisotropy is
310 6.7%, the highest determined in the metasediments. The average VS is 4.1 km/s, resulting in a VP/VS-ratio of 1.76. Maximum VP is parallel to the stretching lineation lowest VP is found parallel to the foliation normal (figure 8).

In the calcareous micaschist (MJS20) the average velocities are lower than in the previously described sample. The average VP is 6.7 km/s and the average VS is 4.1 km/s, resulting in a VP/VS-ratio of 1.63. The P-wave anisotropy in the calcareous micaschist is very strong at 8.2%. Highest VP is concentrated in the foliation plane, lowest VP is found parallel to the
315 foliation normal (figure 8).

In the calcschist (MJS22) the average VP is 6.5 km/s with an anisotropy of 4.7%. The average VS is 4.0 km/s, resulting in a VP/VS-ratio of 1.60. Highest P-wave velocities are concentrated in a broad girdle parallel to the foliation plane with lowest P-wave velocities perpendicular to it (figures 6 and 8).

5.2.6 Gneiss

320 In the gneiss samples (MJS34, MJS35) the P-wave anisotropies and velocity-distributions are quite similar (figure 8).

In MJS34 the average VP is 6.5 km/s and the average VS is 4.1 km/s resulting in a VP/VS-ratio of 1.6. The P-wave anisotropy of 2.8% is low, and the velocity-distribution displays fast velocities in the foliation plane and a minimum velocity perpendicular to it (figure 8).

In MJS35 the average VP is slightly lower than in MJS34 at 6.4 km/s. The average VS is 4.1 km/s, resulting in a VP/VS-
325 ratio of 1.56. The P-wave anisotropy is 1.8%. Similar to MJS34, the highest velocities are found in the foliation plane. However, a weak maximum is visible almost parallel to the stretching lineation (figure 8).

6. Discussion

The elastic anisotropy of rocks is influenced by several factors, among which are its mineral composition, its grain fabric, possible microcracks, and the CPO of its constituent minerals. The CPO constrains a large part of the anisotropy and mostly
330 results from rock deformation. According to Mainprice and Ildefonse (2009) only mineral phases which make up more than 10% of the sample, have an important influence on the elastic properties of the rock in question. In our calculations we have, however, included all mineral phases down to a volume fraction of 5 %.

6.1 Metabasic rocks

6.1.1 CPO development

335 In the metabasic rocks of eclogite, blueschist, amphibolite and greenschist facies metamorphic conditions the compositions are dominated by pyroxenes (omphacite), amphiboles (glaucophane, barroisite, hornblende and actinolite) and chlorite (table 1).

The CPOs of omphacite, glaucophane, barroisite and hornblende have a strong effect on the elastic anisotropy of the eclogite, the blueschist and the amphibolite. In previous studies, the CPO geometry of omphacite has been interpreted as

340 being the result of corresponding strain geometry (Helmstaedt et al., 1972; Kurz et al., 2004; Neufeld et al., 2008; Müller et al., 2011; Keppler et al., 2016). L-type (lineation dominated) fabrics are interpreted as the result of constrictional strain, S-type (foliation dominated) fabrics as the result of flattening strain and the transitional fabrics as indicators of plain strain deformation (e.g. Helmstaedt et al., 1972). SL- and LS-type fabrics have been described by Godard and Van Roermund (1995). Omphacite in MJS17 displays an L-type fabric, with c-axes paralleling the stretching lineation and a- and b-axes

345 paralleling the kinematic y- and z-axes, respectively (figure 4). Keppler (2018) suggests that this correlation of strain geometry to CPO geometry found in omphacite is applicable for amphiboles as well. The amphibole CPO in three of the four metabasites (eclogite, blueschist, and amphibolite; figure 4 and 5) can therefore also be interpreted as L-type or LS-type fabrics, since the c-axes are aligned in lineation direction, while a-axes and b-axes are closer to a girdle distribution perpendicular to the lineation. In the greenschist the actinolite c-axes display a maximum paralleling the stretching lineation

350 with a girdle extending into the foliation plane, while the a-axes display a maximum perpendicular to the foliation plane (figure 5), which points to a transitional SL-type fabric.

Since the omphacite, glaucophane and barroisite c-axes are parallel to both the stretching lineation and one another, the assumption can be made that they formed in the same prograde deformation event. A further possibility would be retrograde mimetic overgrowth of omphacite by glaucophane (McNamara et al., 2012). There is considerable debate on timing of CPO-

355 formation in eclogites, with proponents of formation during subduction, during exhumation, or both, including concurrent switches in strain geometry (Zulauf 1997; Kurz et al., 2004; Kurz, 2005; Neufeld et al., 2008; Müller et al., 2011; Keppler et al., 2016; Keppler, 2018). The CPOs of clinozoisite and epidote frequently have similarities to those of omphacite and the amphiboles. The clinozoisite b-axes display a maximum paralleling the stretching lineation, matching the omphacite and amphibole c-axes. Keppler et al. (2017) suggest concordant CPOs due to opposing crystallographic axis definitions and

360 further conclude that clinozoisite CPOs can likewise be classified in S-type, L-type and intermediated fabrics (Keppler et al. 2018; Puelles et al. 2017). In this case the CPOs of clinozoisite would suggest L-type fabrics hinting at constrictional strain and matching those already described for omphacite and the amphiboles. Albite preferentially displays weak CPOs in the metabasites. However, in the amphibolite and greenschist the weak albite texture of the b- and c-axes corresponds to the stretching lineation, while the a-axis maximum is aligned nearly normal to the foliation plane (figure 5). The slip-systems in

365 plagioclase are highly complex and a correlation with the strain geometry is difficult (Hacker and Christie, 1990; Prior and

Wheeler, 1999). When chlorite or white mica are present, they mostly display a strong alignment of the basal planes in the foliation plane. The similarities in the CPOs of chlorite, amphibole and albite in the greenschist suggest a formation during the same deformation, which likely took place during the retrograde exhumation.

6.1.2 Elastic anisotropies

370 The elastic anisotropies and the VP-distribution in the metabasites are primarily controlled by the influences of omphacite, amphibole and chlorite. These minerals result in two variations of elastic anisotropies and VP-distributions. On the one hand there is a group of results with lower anisotropies (eclogite: 1.4 %; blueschist 3.7%; amphibolite: 3.1 %) and highest VP paralleling the lineation direction and on the other hand there is the greenschist, with a higher anisotropy (5%) and VP-distributed in a girdle in the foliation plane.

375 The compositions, anisotropies and VP-distributions of the first group are primarily controlled by omphacite and amphibole and secondarily by clinozoisite, epidote and garnet. In the amphiboles, highest P-wave velocities in the single-crystals correlate almost perfectly with the c-axes, only deviating by a few degrees. Since the c-axes of the amphiboles in these metabasites parallel the stretching lineation, this results in a velocity maximum in lineation direction. The influence of omphacite is slightly more complicated, since in the omphacite single-crystals the highest velocity is not parallel to the c-

380 axis and instead can be found between the c- and a-axes. This discrepancy has little influence in this case since it also results in a velocity maximum parallel to the stretching lineation, but it should be kept in mind when considering anisotropy strength and distribution. Epidote and clinozoisite CPOs in the blueschist likely contribute to the same pattern of high velocities in lineation direction. Even though garnet has a near random CPO and does not contribute directly to the anisotropy, the high garnet content in the eclogite sample (30 vol.-%, table 1) and the very high isotropic P-wave velocities

385 in garnet single-crystals increase the average seismic velocities in MJS17. In the blueschist and the amphibolite in which the garnet is only present to 5-10% this effect is negligible. Albite can also influence the bulk anisotropy of the metabasites due to its strong single-crystal velocity anisotropy (Brown et al., 2006; Brown et al., 2016). The highest velocities in albite are parallel to the b- and c-axes (Brown et al., 2006), both of which display weak maxima parallel to the stretching lineation in the metabasites, further contributing to the high velocities parallel to the stretching lineation (figure 4 and 5). In the

390 amphibolite, the chlorite basal planes display a girdle distribution perpendicular to the stretching lineation with a maximum perpendicular to the foliation (figure 5). This CPO reflects a rotation of the chlorite basal planes around an axis paralleling the stretching lineation. Due to the lowest velocities in sheet silicates normal to the basal plane, the CPO of chlorite in this sample contributes to the low velocity girdle perpendicular to the stretching lineation. The eclogite contains white mica, however, due to its very low volume fraction of 5 % it is considered insignificant for the bulk anisotropy (Mainprice and

395 Ildefonse, 2009).

In the greenschist the influence of chlorite dominates the composition and the anisotropies as well as the VP-distribution. In sheet silicates the highest velocities in the single crystals are distributed parallel to the basal planes. The strong alignment of these basal planes in the foliation plane results in high elastic anisotropy and highest velocity in the foliation plane very

similar to that observed in the metasediments (figures 5, 6 and 8). Albite in the greenschist also contributes to this pattern
400   due to the maxima of b- axes and c-axes within the foliation plane. Actinolite CPOs in the greenschist display a transitional
SL-type fabric and similarly contribute to the high VP-distribution in the foliation plane.

When comparing the elastic anisotropy results to anisotropy data from literature we find a very wide range of anisotropy
values for these lithologies. For the eclogite our data (1.4 %) fits well inside the bounds of 0.6% to 5.0% by Llana-Fúnez and
Brown (2012), 0.4-3.7 % by Keppler et al. (2015), 2.9% by Bezacier et al. (2010) and 0.6-1.7 % by Wang et al. (2009). In a
405   study by Cao et al. (2013) P-wave anisotropies of 1.4-10.2 % were calculated from CPOs into which our value of 1.4%
would also fit. Similarly, our data from the amphibolite (A: 3.1%) is good in agreement with previous studies on
amphibolites. Siegesmund et al. (1989) report an elastic anisotropy of 6.4 %, Ivankina et al. (2005) of 0.59-5.51 %,
Ullemeyer et al. (2006) of 4.4 % and Keppler et al. (2015) of 2.6 %. In the case of the blueschist, however, the elastic
anisotropy (A: 3.7%) is significantly lower than that presented in literature. Bezacier et al. (2010) report a P-wave anisotropy
410   of 16.1 % for a blueschist, while Cao et al. (2013) report even higher elastic anisotropies of 13.4-24.6 %. There are several
possible explanations for this. The first of these is that in many studies, blueschist samples are collected from highly
deformed bands resulting in stronger anisotropies. In our study, however, a sample with a fabric that is representative for the
lithology within the Alps was selected, resulting in weaker overall mineral CPOs and a lower anisotropy. Furthermore, in the
two studies mentioned the amphibole in the blueschist was glaucophane, while in the case of the sample in this study it is
415   mostly barroisite. This was addressed in part when calculating the elastic properties of the blueschist since no single-crystal
data for barroisite were available so the single-crystal data of glaucophane (Bezacier et al. 2010) were selected as a suitable
substitute. However, glaucophane may still react differently to deformation than barroisite, possibly resulting in stronger
CPOs leading to higher elastic anisotropies. Finally, few studies are available on the elastic anisotropies of greenschists. Due
to the high chlorite-content (30 vol.-%; table 1) which contains a lot of water, measurements by neutron diffraction are very
420   difficult. Measurements by synchrotron diffraction are possible, but limited to very fine grain sizes. When comparing the
elastic anisotropy of 5 % in the greenschist of this study, it can be compared to other lithologies containing such high sheet
silicate fractions like the metasediments and gneisses mentioned below (figure 8).

The VP/VS-ratios in the metabasites range from 1.71 in the blueschist, over 1.72 in the eclogite, 1.73 in the greenschist and
up to 1.76 in the amphibolite. Theses VP/VS-ratios are the direct result of the VP/VS-ratios of their constituent minerals and
425   their respective volume fractions (table 1). The lower ratios of 1.69 in the glaucophane are balanced out by the slightly
higher ratios of pyroxene, garnet, clinozoisite, epidote, chlorite and albite whose ratios range between 1.73-1.74. The ratio in
the amphibolite is particularly high due to the high hornblende content of 40 vol.-% (table 1) with its very high VP/VS-ratio
of 1.84.

6.2 Metasediments

430 **6.2.1 CPO development**

The small circles of quartz c-axes around the foliation normals of samples MJS18 and MJS20 might indicate flattening strain (Lister and Hobbs, 1980; Schmid and Casey, 1986) (figure 6). This is also supported by the distribution of quartz a-axes, but the textures are quite weak. The c-axis pole figures of MJS18 and MJS20 are indicative of basal <a> slip and rhomb <a> slip in quartz (cf. Schmid and Casey, 1986) (figure 6). In MJS22 the quartz c-axes display an inclined single girdle pointing to

435 simple shear. This distribution is also reflected by the quartz a-axes, prism and rhomb planes and indicative of multi-slip activity, possibly a combination of rhomb <a> slip, prism <a> slip and perhaps some basal <a> slip (e.g., Schmid and Casey, 1986; Stipp et al., 2002) (figure 6). In all metasediments the basal planes of white mica and chlorite display a strong alignment in the foliation plane. In all three metasedimentary samples the CPOs of calcite c- and a-axes exhibit axial symmetry with respect to the foliation normal as a symmetry axis (figure 6). This symmetry in calcite has first been

440 interpreted as indicative of pure shear deformation (Wenk et al., 1987; Kern and Wenk, 1983). However, more recently, there was considerable debate on this topic (De Bresser, 1989; Ratschbacher et al., 1991; Erskine et al., 1993; Burlini et al., 1998; Leiss et al., 1999; Bestmann et al., 2000). The maximum of the calcite a-axes can either parallel the stretching lineation or the kinematic y-axis or even display multiple maxima between the two (Ratschbacher et al., 1991; Punturo et al., 2005; Bestmann et al., 2000; Bestmann et al., 2006; Trullenque et al., 2006). Very few studies have focused on dolomite

445 CPOs. Delle Piane et al. (2009) show that dolomite CPOs very closely reflect those of calcite, but deformation itself strongly influence the texture development when both phases deform together. Dolomite in MJS18 displays very similar CPOs to calcite, with the same axial symmetry in the c- and a-axes (figure 6).

6.2.2 Elastic anisotropies

The elastic anisotropies of the metasediments (MJS18, MJS20 and MJS22) range from 4.7-8.2 % and are among the highest

450 presented in this study (figure 8). These anisotropies directly correlate to the pronounced foliation and high mica and chlorite contents found in the metasediments, as well as the strong influence of calcite and dolomite CPOs on the samples. This correspond to elastic anisotropy data from similar metasediments reported in other studies such as Weiss et al. (1999) with 4.56-6.53 %, Erdman et al. (2013) with 2.3-11.4 %, and Keppler et al. (2015) with 5.1-7.4 %.

In our study, the orientation of white mica and chlorite (10-40 vol.-%) basal planes and the very high single-crystal

455 anisotropy of white mica, which is used for both sheet silicates, mainly contribute to the elastic anisotropy in the metasediments. The V_P distribution in muscovite single crystals is dominated by its highest velocity parallel and lowest velocity normal to the basal plane (figures 6 and 8). As micas preferentially align with increasing strain in the foliation, the result is a strong tendency for high P-wave velocities in the foliation plane in deformed mica-rich rocks.

The composition of the metasediments varies substantially. Calcite and dolomite combined range from 15 vol.-% up to 70

460 vol.-% (table 1). The CPOs of both of these minerals strongly influence the P-wave anisotropy of the metasediments. They

result in high velocities in the foliation plane with a maximum parallel to the stretching lineation. Further they contribute to the velocity minima normal to the foliation plane. This is the result of the calcite single crystal data in which the c- and a-axes are the slow and fast directions, respectively (Puntero et al., 2005). The single-crystal anisotropy of dolomite is quite similar to that of calcite. Hence, both similarly contribute to the bulk anisotropy of sample MJS18. The contribution of

465 quartz to the P-wave anisotropy is weak due to the near random CPO observed for the poles of the rhombs, which represent the fastest direction in the quartz crystal. Furthermore, the three-fold symmetry of the quartz single crystal makes any influence on the velocity distribution complicated to assess. The garnet present in MJS20 displays a random texture and therewith has no significant influence on the elastic anisotropy.

The VP/VS-ratios in the metasediments range from 1.60 to 1.76. The highest ratio is found in sample MJS18 at 1.76, while

470 the ratios of the other two samples range from 1.60-1.63. The high ratio of MJS18 is the result of the very high dolomite content of 60 vol.-% and the 10 vol.-% of calcite in this sample. Both dolomite and calcite have very high single crystal VP/VS-ratios of 1.851 and 1.918, respectively. These are significantly higher than those of quartz (20 vol.-%) and white mica (vol.-10%) at 1.476 and 1.726, respectively. For the other two samples the velocity ratio is dominated by quartz (40-55 vol.-%), resulting in their low ratios.

475 **6.3 Gneiss**

6.3.1 CPO development

The small circles around the foliation normal, visible in the quartz c-axes and the distribution of a-axes of MJS34 indicate flattening strain with combined activity of rhomb <a> slip and basal <a> slip (Lister and Hobbs, 1980; Schmid and Casey, 1986) (figure 7). In MJS35, the type I crossed girdle in the quartz c-axes is indicative of plane strain with combined basal

480 <a>, rhomb <a> slip and prism <a> slip (e.g., Schmid and Casey, 1986). White mica and chlorite CPO in both MJS34 and MJS35 bear a strong relation to the foliation plane (figure 7). Albite in MJS34 and MJS35 shows very weak CPOs, but albite a-axes in both display a weak but visible maximum paralleling the stretching lineation (figure 7). Due to the random distribution of the b-axes and c-axes a clear assignment cannot be made, however this CPO pattern is closest to an axial A type, indicating that the activated plagioclase slip system is [100](010) and/or [100](001) (Satsukawa et al., 2013).

485 **6.3.2 Elastic anisotropies**

The elastic anisotropy in the gneisses MJS34 and MJS35 is primarily constrained by the CPOs of mica and chlorite (25 vol.-% and 15 vol.-% respectively, table 1) and secondarily by quartz and albite. However, the influence of quartz is far weaker than that of the sheet silicates as the pole figure maxima of quartz do not match well to the bulk elastic anisotropy diagram. The strong single-crystal anisotropy of muscovite and chlorite in combination with their strong CPO, results in the high

490 velocities found in the foliation plane and the corresponding perpendicular velocity minima (figures 7 and 8). The VP/VS-ratio of the gneiss samples ranges from 1.56 to 1.60. As in the metasediments this low ratio is the result of the very low ratio

of 1.48 in quartz (40 vol.-% and 60 vol.-% respectively) and the intermediate ratios in albite (1.73) and the sheet silicates (1.73; 25 vol.-%) in the samples (appendix 2; table 1). The velocity distribution is very similar to that of the metasediments, with high velocities in the foliation plane and low velocities perpendicular to this plane (figure 8), but the elastic anisotropy
495   is lower. When comparing the results of our study to those previously published, we see a very wide range of P-wave anisotropy data reported in gneisses. Values range from low to intermediate anisotropies such as 1.76-8.73 % by Weiss et al. (1999), 1.72-2.99 % by Ivankina et al. (2005), 0.9-3.2 % by Ullemeyer et al. (2006), and 7.9 % by Kern et al. (2008), up to high to very high anisotropies such as 4.1-20.1 % by Erdman et al. (2013) and 8.0-11.3 % by Llana-Fúnez and Brown (2012). As gneisses predominantly display weak CPOs similar to our results, our low elastic anisotropy values are common
500   for gneisses.

6.4 Elastic anisotropy data in the context of a subduction and exhumation setting

In this study we examine the elastic properties of a wide array of lithologies in a subduction and exhumation setting. As previously mentioned and is visible in figure 2 we have divided this setting into several structural positions. These are the oceanic crust of the subducting plate undergoing different metamorphic facies, the sedimentary cover of the subducting
505   plate, which is at least partly scraped off and transferred into the accretionary prism and finally the deformed continental crust. This continental crust can be part of a microcontinent (as in the case of the Dent Blanche unit) or can be the continental crust on the opposite side of the ocean, which enters the subduction zone subsequently to the oceanic crust leading to continental collision. This scenario is presented in the schematic cross-section in figure 2. The structural positions vary in a range of criteria such as degree of deformation and mineral composition. The different degrees of deformation and
510   metamorphism lead to different texture strengths and variable mineral content. When comparing the different structural positions, samples can be related in order of SiO2-content, with quartz- and feldspar-rich felsic rocks on one end of the spectrum, and the olivine-rich ultramafic rocks on the opposite end of the spectrum. This has been discussed extensively in a very insightful study by Almqvist and Mainprice (2017) for the earth's crust. Generally, a trend of lower velocities in the felsic and higher velocities in the mafic to ultramafic rocks can be observed. In this study we further discuss another criterion
515   of great importance on the elastic properties which is the influence of CPO of the constituent minerals on the elastic anisotropy. This might differ from the compositional effect in the spectrum from felsic, over mafic to ultramafic. Concerning elastic anisotropies, the sheet-silicate content and its alignment caused by deformation show a pronounced influence.

During the prograde metamorphism in subduction zones the basaltic oceanic crust undergoes blueschist-facies metamorphism leading to the formation of glaucophane (in some cases barroisite), epidote, omphacite and some accessory
520   minerals. Glaucophane and feldspar usually dominate the composition, of which glaucophane greatly influences the anisotropy and the velocity distribution. Blueschists display strongly varying elastic anisotropies of 3.7-24.6 % (Cao et al., 2013; Bezacier et al., 2010). The deformation during subduction leads to CPO formation in amphibole. In most cases plane strain is dominant. The orientation of the c-axes of amphibole parallel to the stretching lineation results in the highest velocity parallel to the lineation. Dominant constriction in the blueschist of the present study produces a similar pattern (see

525 figures 4 and 8). At greater depth the oceanic crust reaches eclogite facies metamorphic conditions. There, omphacite and garnet are the rock-forming minerals (table 1) and omphacite has the strongest influence on the elastic anisotropies. The elastic anisotropies in eclogites are lower than those in the blueschists and display values of 0.4-10.2 %, due to the higher elastic anisotropy of amphiboles compared to pyroxenes (Llana-Fúnez and Brown, 2012; Keppler et al., 2015; Cao et al., 2013; Bezacier et al., 2010; Wang et al., 2009). The highest P-wave velocities remain parallel to the stretching lineation, due

530 to the omphacite CPO yielding an alignment of [001] in lineation direction during plane strain as well as constriction (see figures 4 and 8).

During exhumation the high pressure rocks frequently experience retrogression as they pass through the amphibolite facies, where they are overprinted or completely retrogressed. In the amphibolite-facies rocks, hornblende together with feldspar and epidote dominates the composition. These minerals result in anisotropies of 0.59-6.4 % (Siegesmund et al., 1989;

535 Ivankina et al., 2005; Ullemeyer et al., 2006; Keppler et al., 2015), of which the amphiboles have the strongest influence. The velocity distribution in the amphibolites is very similar to those observed in the blueschists and eclogites, with highest velocities found parallel to the stretching lineation (figure 8). Further along the exhumation path, the rocks enter greenschist facies conditions, where further retrogression is possible. In the greenschist facies, chlorite is formed and together with actinolite, epidote and feldspar, makes up the greenschists. The high chlorite content results in intermediate to high

540 anisotropies of 5 %, presented in this study and a velocity distribution with high velocities in the foliation plane (figure 8). This is directly caused by the sheet silicate chlorite and provides greenschist VP patterns very similar to those of the metasediments and the deformed continental crust. Very few studies have been made on the elastic properties of greenschists and further research is necessary to better constrain this lithology. The present study, however, shows that greenschist facies metabasites constitute a special case concerning their elastic properties. While VP is still higher than that of metasediments

545 and comparable to the other metabasites, the velocity distribution resembles that of the metasediments. Due to their frequent occurrence in subduction settings and collisional orogens this needs to be considered in seismic investigations.

In the overlying metasediments VP anisotropies display values of 2.3-11.4 % (Erdman et al., 2013; Keppler et al., 2015). The composition of these metasediments in the present study is highly variable. However, quartz, feldspar, mica, chlorite, calcite and dolomite are the most abundant mineral phases and also those controlling the elastic properties. The velocity distribution

550 shows highest velocities in the foliation plane with a velocity-minimum normal to it, as seen in the greenschists, due to the high contents in sheet silicates as well as of calcite and dolomite.

In the deformed continental crust the composition is dominated by feldspar, quartz, mica and chlorite, represented by the gneisses in this study. In the gneisses the elastic anisotropies are similar to those observed in the metasediments, with values of 1.72-20.1 % (Ivankina et al., 2005; Ullemeyer et al., 2006; Erdman et al., 2013; Kern et al., 2008; Weiss et al., 1999) even

555 though our samples are at the lower end of this range. The velocity distribution of the gneisses displays highest velocities in a girdle in the foliation plane, similar to the greenschists and metasediments. This results from the high sheet silicate-contents of this lithology (figure 8). In both the metasediments and the gneisses the metamorphic facies does not change the elastic properties beyond the greenschist facies up to HP conditions, since the single-crystal anisotropies in most of the

constituent minerals are constant at higher metamorphic conditions (e.g. quartz, feldspar, calcite and dolomite). Mica shows

560 changes in chemistry with rising pressures and quartz transitions to coesite at UHP conditions, but due to limited single-crystal anisotropy data of these minerals their influence cannot be determined.

With these structural positions and their characteristic elastic properties in mind a better interpretation of seismic data is feasible. Lower V_P/V_S-ratios with high elastic anisotropies hint at felsic compositions and high sheet silicate content, while the higher V_P/V_S-ratios and overall lower to intermediate elastic anisotropies indicate more mafic lithologies, with

565 exception of the zones of both high velocity-ratios and high anisotropies characteristic of deformed greenschists (Almqvist and Mainprice, 2017). In the majority of the mafic rocks a dominance of high velocities parallel to the stretching lineation can be observed. In comparison, the metasediments and gneisses display a distinctly different velocity distribution with the highest velocities in the foliation plane. But again, greenschists are an exception. This mafic lithology displays high velocities in the foliation plane with an intermediate to high P-wave anisotropy of 5% reminiscent of the metasediments. The

570 knowledge of the characteristic properties of the different structural positions can help to better identify thickness and boundaries of the predominantly mafic subducting plate and the felsic continental margins. It can further assist to better constrain the anatomy of subduction channels.

6.5 Common issues for calculated elastic properties

Precise single-crystal data are not available for all mineral phases, but our fitting represents a good approximation. Firstly,

575 the selected substitute minerals are very similar to those in the samples, especially with respect to their crystallography. Secondly, the goal of our models is not to simulate specific rock samples, but the rock types that are representative distinct sections of the subduction and exhumation cycle. Hence, for the missing mineral phases substitute single crystal data were selected, which we considered to be acceptable matches due to the metamorphic and petrological context they can be found in. As mentioned above, this was done for clinozoisite, for which we selected the single-crystal data of epidote (Aleksandrov

580 et al., 1974), as well as for chlorite, for which we selected the data of muscovite (Vaughan and Guggenheim, 1986). Furthermore, the almandine garnet single-crystal data of Zhang et al. (2008) was used for all garnet-bearing samples, even if different garnet compositions were present in the samples. A similar simplification was made for the amphiboles. In samples with glaucophane or barroisite, we used glaucophane single-crystal data by Bezacier et al. (2010), while for all other amphiboles we utilized the single-crystal elastic constants for hornblende from Aleksandrov and Rhyzova (1961).

585 Throughout this study the Voigt averaging scheme was used consistently. This model is based on the "equal strain" assumption and represents the upper bound for the stiffness of a polycrystalline material. As a result, the velocities presented are therefore maximum velocities only. This does not correspond to the "equal stress" assumption of the Reuss model that represents the lower stiffness boundary. Neither method takes grain shape, layering, microcracks, fractures or pores into account (Vasin, et al. 2017). Other methods such as the self-consistent method GeoMIXself (GMS) by Matthies (2010;

590 2012) can take CPOs, morphologies and SPOs of the mineral grains into account. However, only based on CPO data the Voigt averaging scheme is deemed appropriate to estimate representative samples. Furthermore, the Voigt averaging scheme

has widely been used on the subject and ensures therefore comparability with most other published texture derived elastic anisotropy data.

A further issue, which requires discussion is the influence of microcracks on the elastic properties of the rocks (Siegesmund 595 and Vollbrecht, 1991; Siegesmund et al., 1993; Ullemeyer et al. 2011, Vasin et al., 2014; 2017). The elastic anisotropies presented in this study are texture-derived and do not take the presence of microcracks into account. Instead, they depend on the mineral composition, single-crystal mineral properties and the CPOs of the respective phases. This means that the calculated data reflect the elastic anisotropies of rocks at great depth, in which microcracks are closed due to high confining pressure. The influence of microcracks can well be observed in studies in which CPO-derived calculated results have been 600 compared to elastic anisotropy measurements in laboratory experiments at high confining pressure. Some of these studies have demonstrated that the influence of microcracks is still detectable up to confining pressures of 1 GPa (Christensen, 1974; Ullemeyer et al., 2011). However, with increasing pressure and closing of the microcracks, the results far better resemble the CPO-derived results and the influence of microcracks decreases significantly (Ivankina et al., 2005; Lokajicek et al., 2017; Vasin et al., 2017).

7. Conclusion

Using time-of-flight neutron diffraction, texture data from a wide range of compositionally variable and complex rocks were collected. These are representative of different positions for subduction and exhumation in a collisional orogen. From these data texture-derived elastic anisotropies and other petrophysical properties were modelled. The following conclusions can be made:

610 1. Composition and CPO have a strong effect on the rock texture and the related elastic properties in subduction and exhumation settings. Felsic rocks with high sheet silicate and calcite or dolomite contents preferentially display the fastest P-wave velocities in the foliation plane independently of the orientation of the stretching lineation. The mafic rocks with high amphibole and pyroxene contents in this study mostly display highest P-wave velocities parallel to the stretching lineation, with exception of greenschist, in which the high chlorite content results in the fast velocities in the foliation plane.

615 2. Among mafic rocks the elastic anisotropies vary strongly by composition. The elastic anisotropies increase from a composition dominated by pyroxenes (1.4 %), over amphiboles (3.1-3.7 %) and are highest in the chlorite dominated greenschist (5 %).

3. The described structural positions have varying elastic properties (elastic anisotropy, VP/VS-ratio, P-wave-velocity distribution), often overlapping each other. However, in many cases the combination of properties can be utilized to 620 distinguish these positions from one another. In this study the mafic rocks of oceanic crustal origin such as eclogite, blueschist and amphibolite display lower anisotropies (1.4-3.7 %) with higher VP/VS-ratios (1.71-1.76) and highest velocities parallel to the stretching lineation. An exception to this is the greenschist (5 %; 1.73), which displays properties more characteristic to the felsic rocks of the continental crust and the metasedimentary cover. These gneisses, micaschists

and calcareous schists display higher elastic anisotropies (1.8-8.2 %), lower VP/VS-ratios (1.56-1.76) and fast P-wave-
625 velocities distributed in the foliation plane.

Acknowledgments

This study was funded by the German research foundation (DFG-grant No. KE 2268/2-1) as part of the DFG priority programme "Mountain Building Processes in 4 Dimensions". Fruitful discussions within the priority programme are gratefully acknowledged. We wish to thank the Frank Laboratory of Neutron Physics (FLNP) at the Joint Institute of
630 Neutron Research (JINR) in Dubna, Russia for the texture measurements made possible at SKAT.

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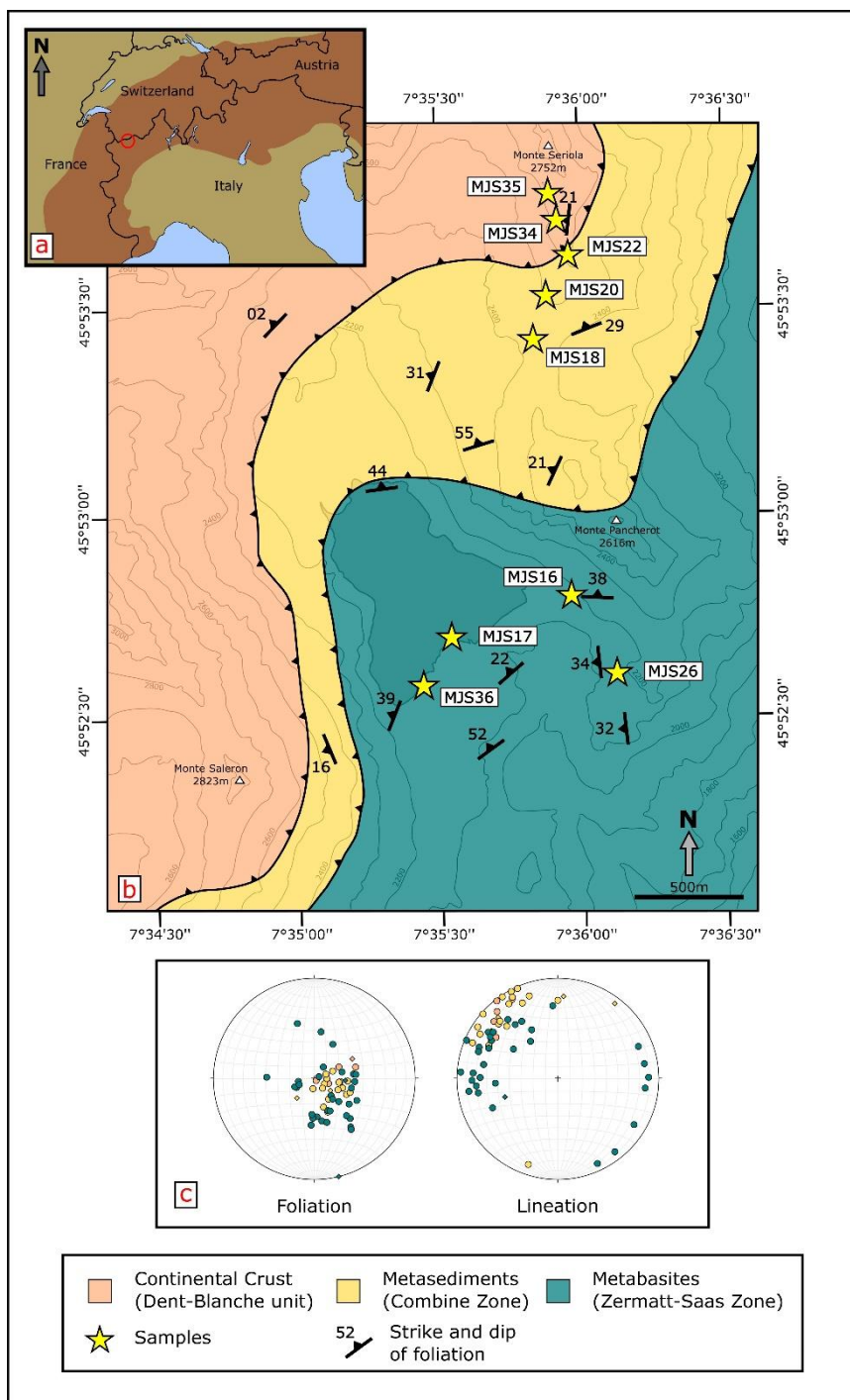
Figure 1: (a) Miniature map displaying approximate location of the study area in the European alpine orogen. (b) Geological map of the study area at Lago di Ciganan, in the Western Italian Alps modified after the map of Kirst and Leiss, 2016. Sample locations indicated by stars. (c) Stereoplots of foliation normal and stretching lineation as equal area projections in the lower hemisphere; data compiled from own measurements (diamonds) and from Kirst and Leiss (2016) (circles); Colors match those used in the map.
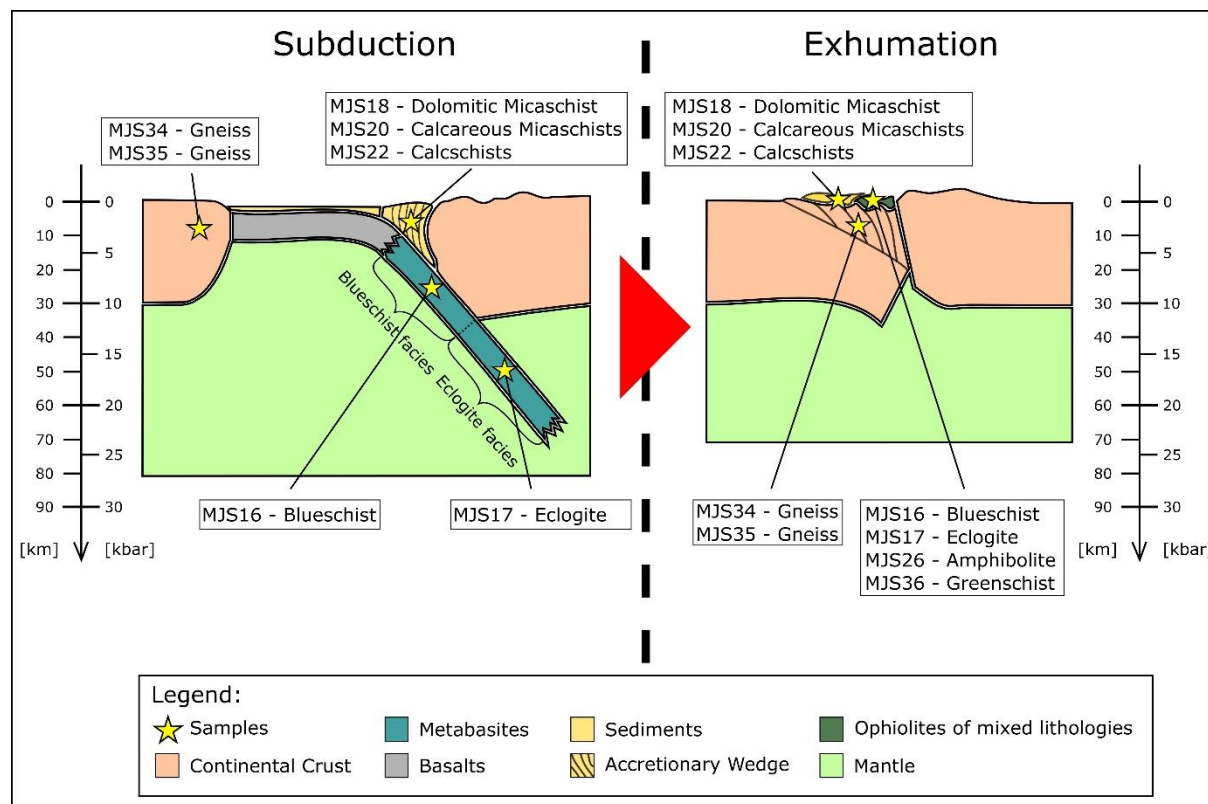
Figure 2: Schematic Cross section of subduction and exhumation in a collisional orogeny. The stars show locations which the
855 **samples represent within this setting. Colors correlate to map in figure 1.**
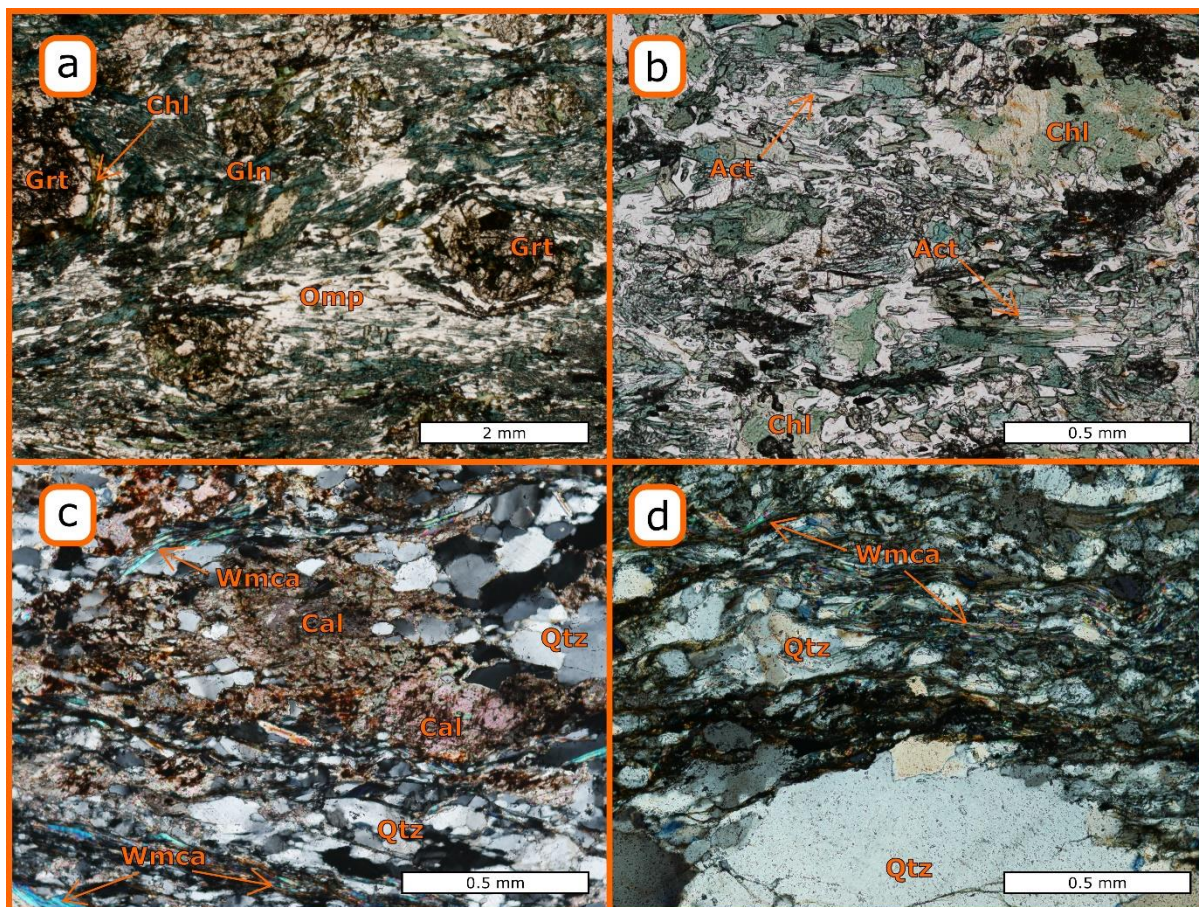
Figure 3: Thin-section photographs of typical microstructures in four representative samples. (a) eclogite (MJS17) with foliation of glaucohpane (Gln) and omphacite (Omp) flowing around larger grains of garnet (Grt) with some retrograde chlorite (Chl) visible at its rim. (b) greenschist (MJS36) with actinolite (Act) and chlorite (Chl) visible in the foliation and larger chlorite grains found unoriented grown over the foliation. (c) calcareous micaschist (MJS20) with quartz (Qz), calcite (Calc) and white mica (Wmca) oriented in the foliation. (d) gneiss (MJS35) displaying strongly foliated fine grained quartz (Qz) and white mica (Wmca) larger coarse grained quartz. Sections are oriented perpendicular to the foliation and parallel to the stretching lineation. (a) and (b) are shown under plane polarized lighte, while (c) and (d) are shown under x-polarized light. For sample descriptions see table 1.

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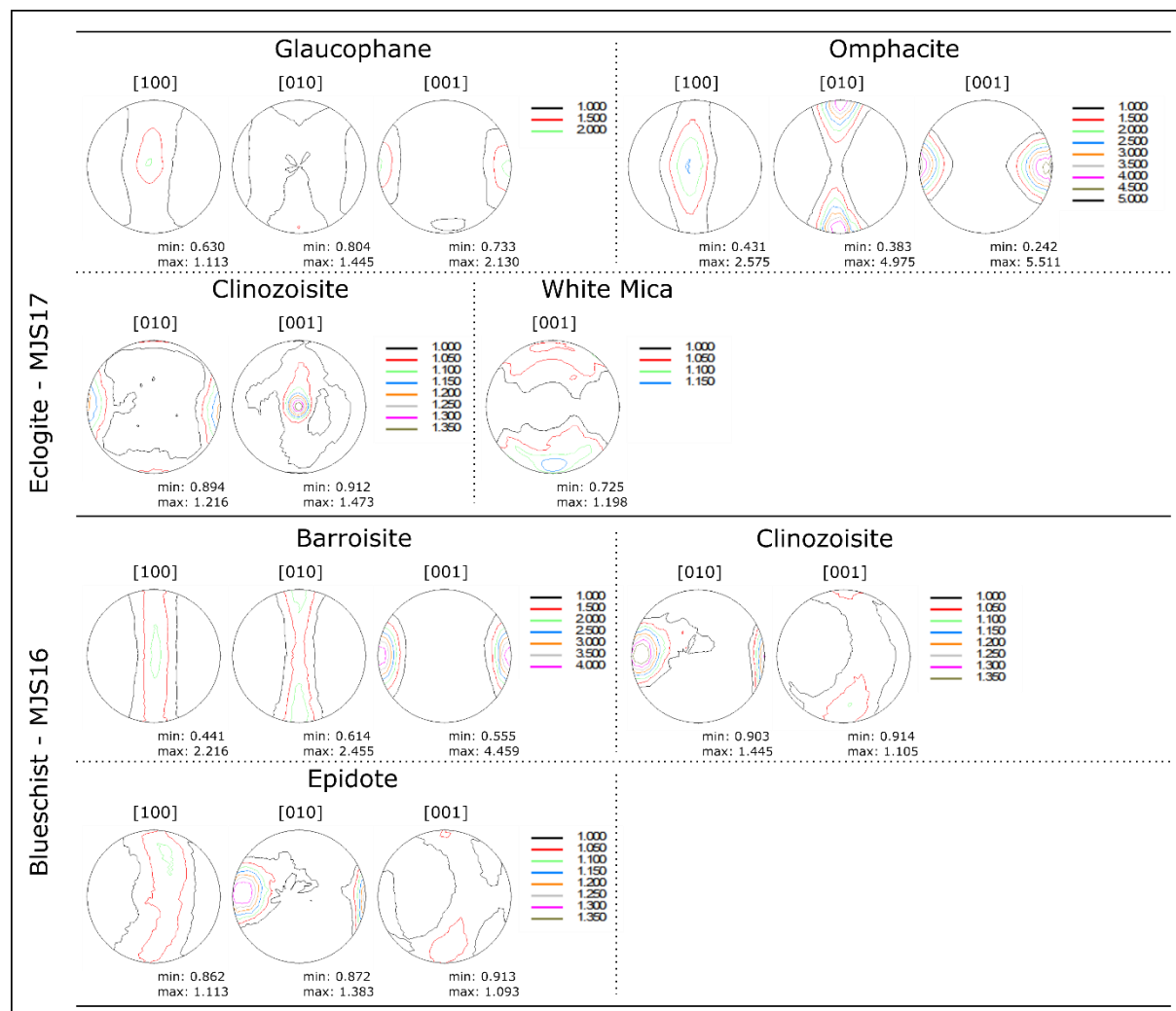
Figure 4: Calculated pole figures of important mineral phases in the eclogite (MJS17) and the blueschist (MJS16). The stretching lineation (kinematic x-direction) is oriented EW in the pole figures, the foliation normal (kinematic z-direction) is oriented NS and the kinematic y-direction (perpendicular to x- and z-directions) is normal to the pole figure plane. The pole figures are lower hemisphere equal area projections. The contour levels display multiples of random distribution and minimum and maximum values are listed below the pole figure.
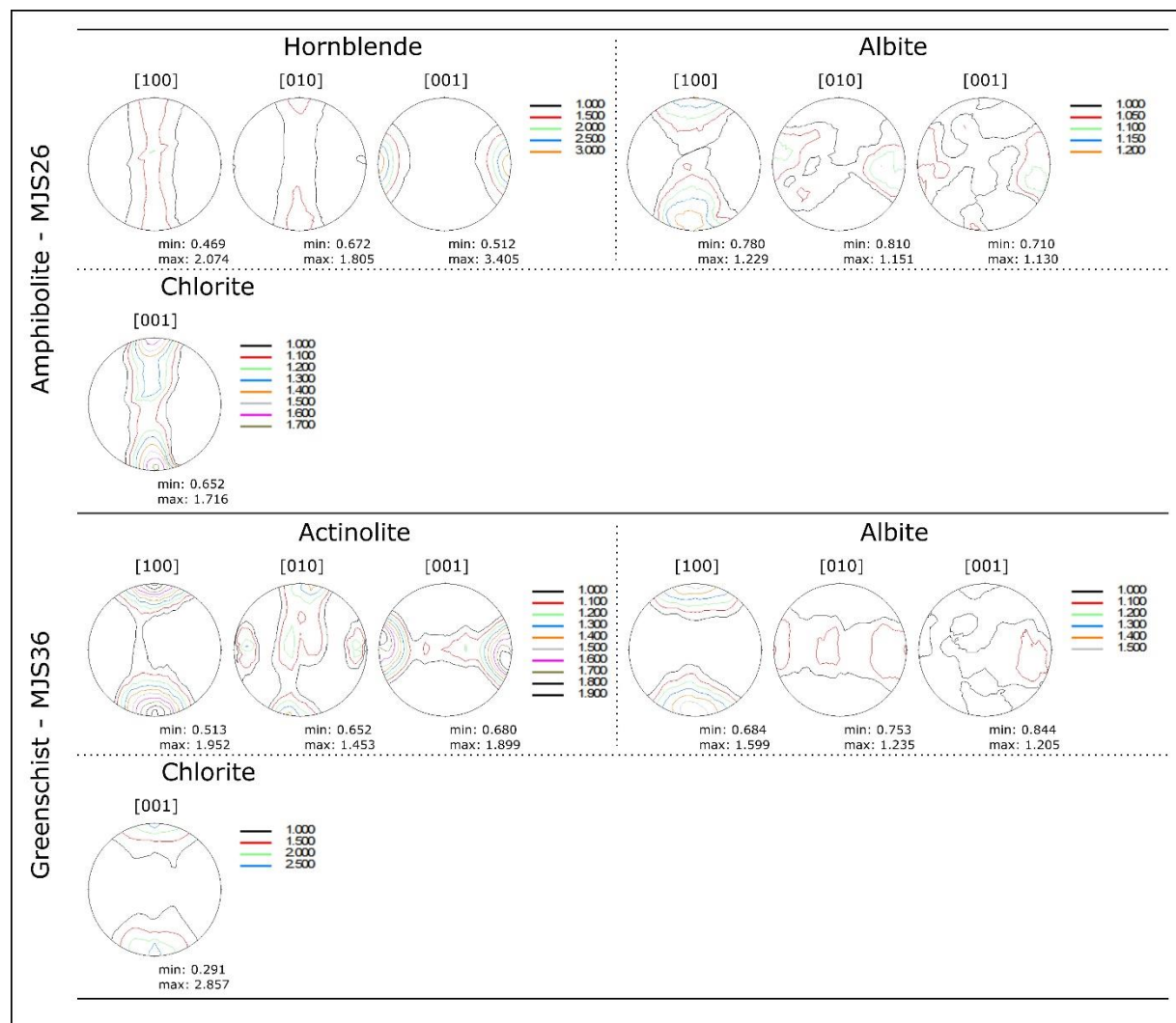
Figure 5: Calculated pole figures of important mineral phases in the amphibolite (MJS26) and the greenschist (MJS36). The stretching lineation (kinematic x-direction) is oriented EW in the pole figures, the foliation normal (kinematic z-direction) is oriented NS and the kinematic y-direction (perpendicular to x- and z-directions) is normal to the pole figure plane. The pole figures are lower hemisphere equal area projections. The contour levels display multiples of random distribution and minimum and maximum values are listed below the pole figure.
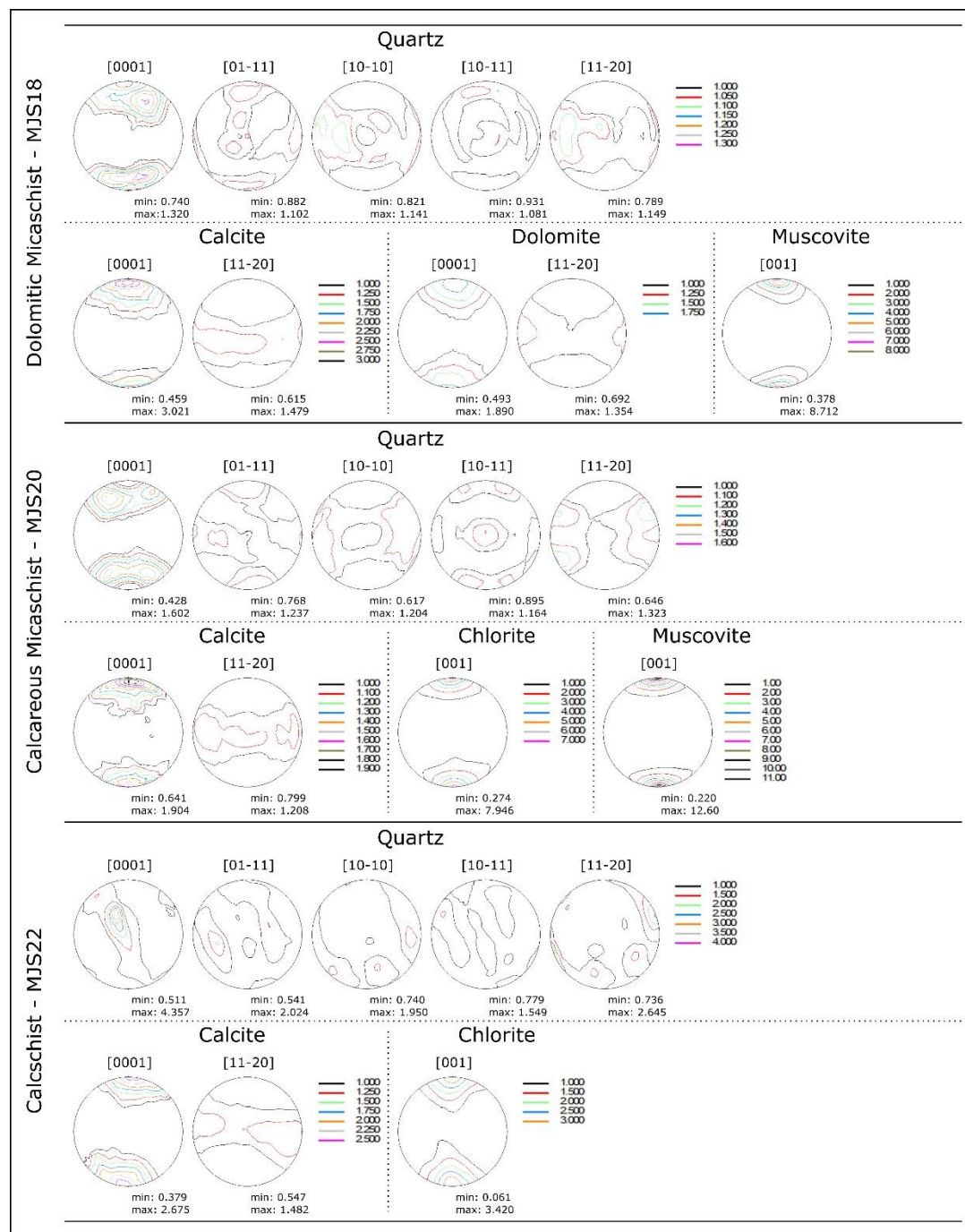
Figure 6: Calculated pole figures of important mineral phases in the dolomitic micaschist (MJS18), the calcareous micaschist (MJS20) and the calcschist (MJS22). The stretching lineation (kinematic x-direction) is oriented EW in the pole figures, the foliation normal (kinematic z-direction) is oriented NS and the kinematic y-direction (perpendicular to x- and z-directions) is

normal to the pole figure plane. The pole figures are lower hemisphere equal area projections. The contour levels display multiples of random distribution and minimum and maximum values are listed below the pole figure.

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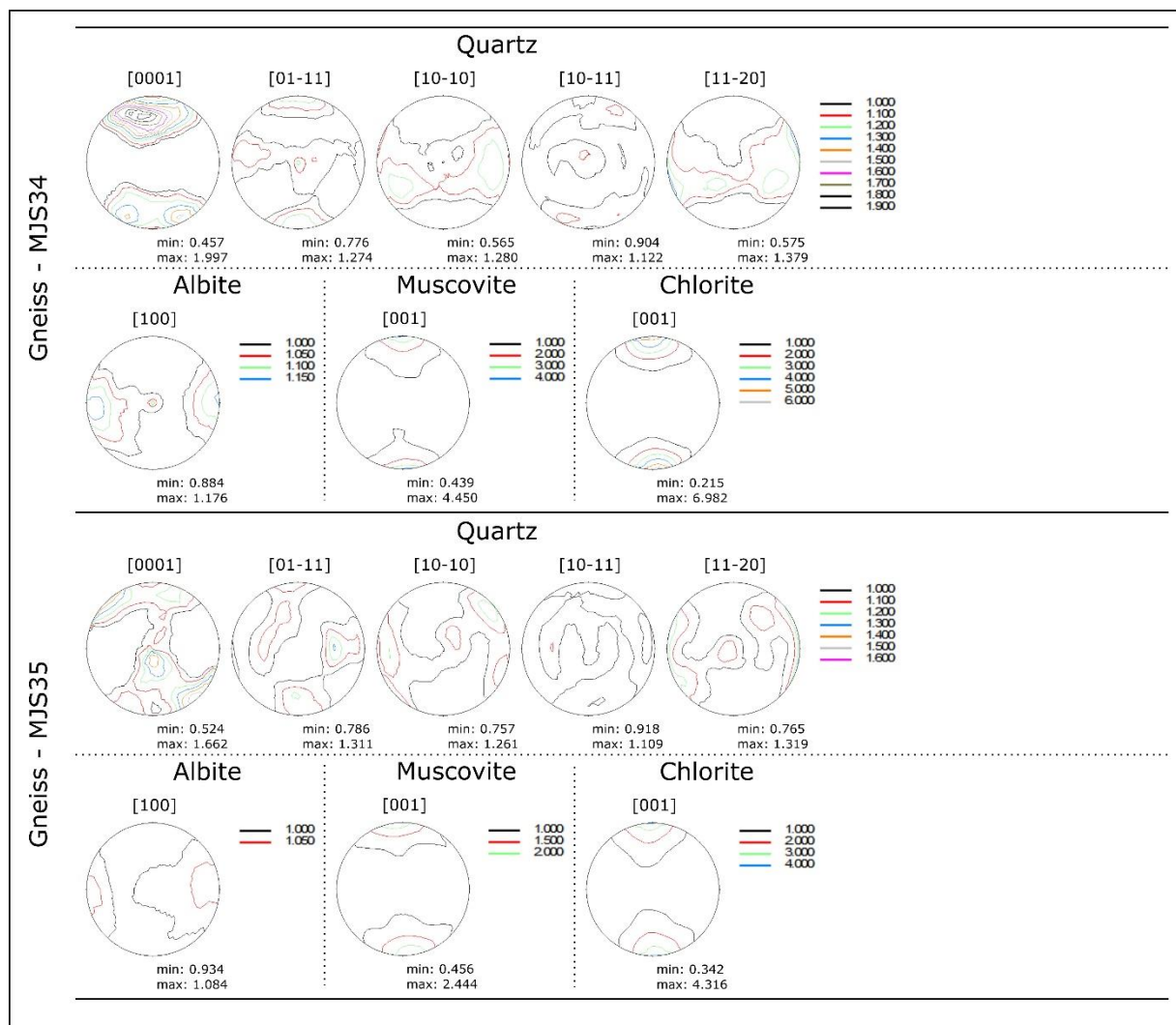
Figure 7: Calculated pole figures of important mineral phases in the gneisses (MJS34) and (MJS35). The stretching lineation (kinematic x-direction) is oriented EW in the pole figures, the foliation normal (kinematic z-direction) is oriented NS and the kinematic y-direction (perpendicular to x- and z-directions) is normal to the pole figure plane. The pole figures are lower hemisphere equal area projections. The contour levels display multiples of random distribution and minimum and maximum values are listed below the pole figure.

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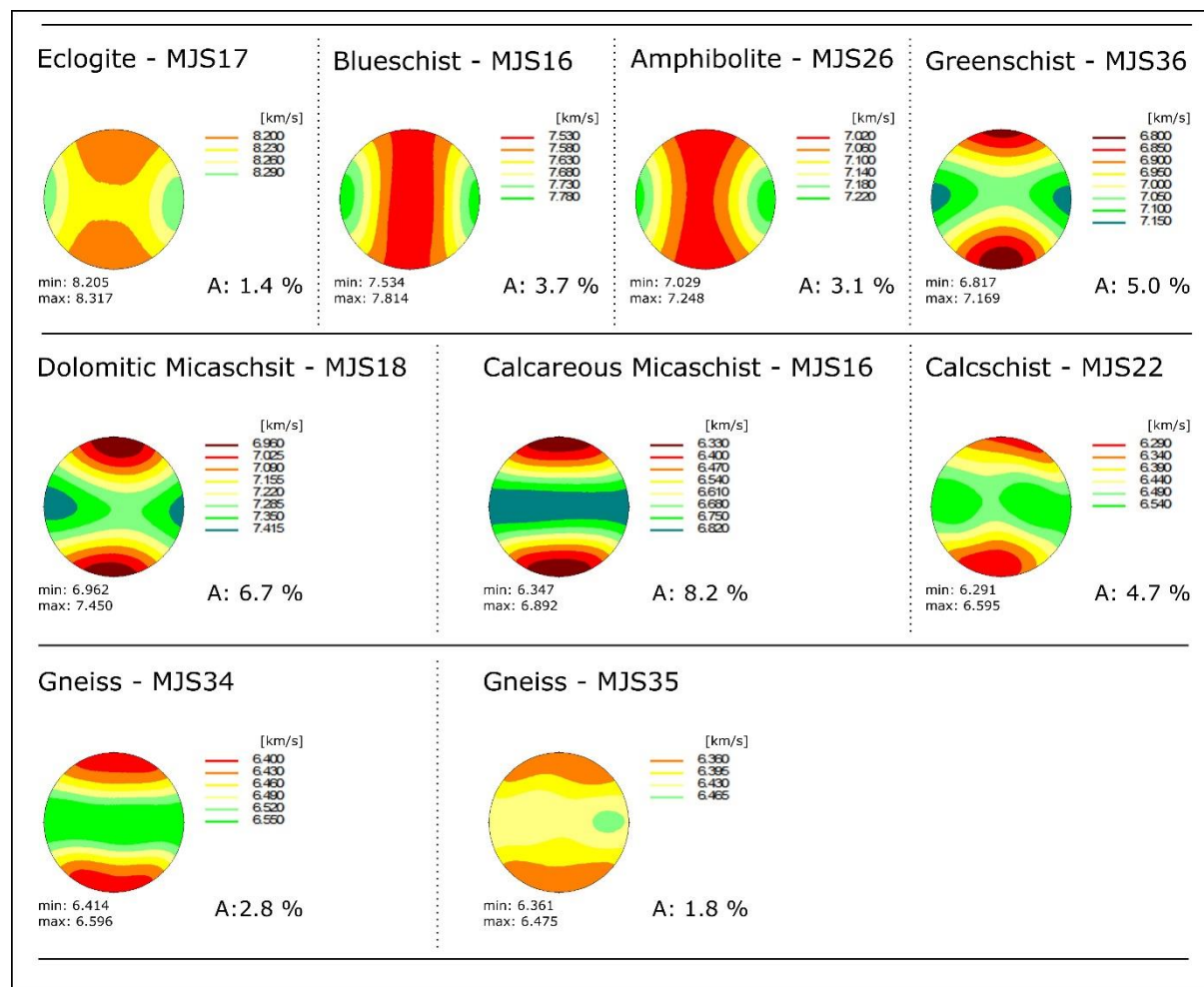
Figure 8: Texture derived calculated P-wave velocity distributions of all samples in the study. All velocities in [km/s] with maximum and minimum values as well as P-wave anisotropy (in %) given below the pole figure.

Sample	Lithology and discription	Composition
MJS16	Strongly foliated blueschist	40 % barroisite, 30% albite, 10% almandine, 15% clinozoisite, 5% epidote
MJS17	(U)HP eclogite with a clear foliation and lineation and large garnets	30% omphacite, 30% garnet, 15% clinozoisite, 10% glaucophane, 10% albite, 5% white mica
MJS18	Dolomitic micaschist with calcite and a strong mica-related foliation	60% dolomite, 20% quartz, 10% calcite, 10% muscovite
MJS20	Garnet-bearing calcareous micaschist	40% quartz, 30% muscovite, 15% calcite, 10% chlorite, 5% garnet
MJS22	Well foliated calcschist	55% quartz, 30% calcite, 15% chlorite
MJS26	Weakly retrogressed amphibolite	40% hornblende, 25% albite, 20% clinozoisite, 10% chlorite, 5% garnet
MJS34	Weakly retrogressed gneiss	40% quartz, 35% albite, 15% muscovite, 10% chlorite
MJS35	Quartz-rich weakly retrogressed gneiss	57% quartz, 27% albite, 9% chlorite, 7% muscovite
MJS36	Actinolite-bearing greenschist	30% chlorite, 30% clinozoisite, 20% albite, 20% actinolite

Table 1: Description and modal composition of samples in the study.

Sample	Lithology	avarage VP (km/s)	min VP (km/s)	max VP (km/s)	VP-anisotropy (%)	avarage VS (km/s)	min VS1 (km/s)	max VS1 (km/s)	min VS2 (km/s)	max VS2 (km/s)	VP/VS
MJS16	Blueschist	7.63	7.53	7.81	3.7	4.47	4.49	4.52	4.42	4.49	1.71
MJS17	Eclogite	8.24	8.21	8.32	1.4	4.78	4.78	4.81	4.77	4.78	1.72
MJS18	Dolomitic micaschist	7.23	6.96	7.45	6.7	4.12	4.06	4.24	4.02	4.10	1.76
MJS20	Calcareous micaschist	6.65	6.35	6.89	8.2	4.09	3.97	4.25	3.94	4.09	1.63
MJS22	Calcschist	6.47	6.29	6.60	4.7	4.04	4.00	4.14	3.97	4.04	1.60
MJS26	Amphibolite	7.10	7.03	7.25	3.1	4.04	4.05	4.08	4.00	4.06	1.76
MJS34	Gneiss	6.51	6.41	6.60	2.8	4.07	4.02	4.13	4.00	4.08	1.60
MJS35	Gneiss	6.42	6.36	6.48	1.8	4.12	4.10	4.16	4.07	4.13	1.56
MJS36	Greenschist	7.01	6.82	7.17	5.0	4.05	4.01	4.14	3.98	4.04	1.73

920 **Table 2: P-wave (VP) and S-wave (VS) velocities, elastic anisotropies (A%) and VP/VS-ratios of all samples, modelled from CPOs.**

Sample	Phase	F2 index
MJS16	almandine	1.01
	albite	1.01
	barroisite	2.36
	clinozoisite	1.02
	epidote	1.02
MJS17	omphacite	3.49
	almandine	1.00
	glaucophane	1.19
	albite	1.02
	clinozoisite	1.01
	white mica	1.02
MJS18	quartz	1.04
	dolomite	1.14
	calcite	1.28
	muscovite	2.18
MJS20	quartz	1.14
	muscovite	2.83
	calcite	1.08
	chlorite	2.17
	almandine	1.02
MJS22	calcite	1.34
	quartz	1.54
	chlorite	1.52
MJS26	almandine	1.01
	hornblende	1.63
	albite	1.07
	clinozoisite	1.01
	chlorite	1.09
MJS34	quartz	1.18
	albite	1.03
	chlorite	2.15

	muscovite	1.27
MJS35	quartz	1.06
	muscovite	1.13
	albite	1.01
	chlorite	1.53
MJS36	albite	1.18
	actinolite	1.21
	clinozoisite	1.03
	chlorite	1.34

Appendix A: F2-indicies after Bunge (1982) for all mineral phases in the study.