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Seismic anisotropy of upper mantle-lower continental crust rocks in Cabo Ortegal (NW Spain) from crystallographic preferred orientation (CPO) patterns

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Abstract: The development of crystallographic preferred orientation (CPO) in rocks as a consequence of deformation processes at high pressure and high temperature conditions is one of the sources of anisotropy in the propagation of seismic waves presently observed in deep seismic profiles of rocks in the lower crust and upper mantle. The CPO of 8 rock-forming minerals measured using electron back-scattered diffraction (EBSD) is compared with direct measurement of the propagation of seismic waves at high pressure in 5 rock specimens (felsic gneiss, eclogite, mafic granulite, metabasite and pyroxenite) representing the transition from the upper mantle to the lower crust in the Cabo Ortegal Complex.

Keywords: seismic anisotropy, EBSD, Moho, CPO, lower crust.

Direct access to rocks in the transition between the upper mantle and the lower crust is only possible in very limited exposures worldwide. The upper tectonic unit of the Cabo Ortegal Complex (NW Spain) allows the observation of the transition between the upper mantle and a relatively mafic lower crust of what has been interpreted as a thinned continental crust or an island arc (Galán and Marcos, 1997). This upper part of the complex is made up of an ordered sequence of ultramafic, mafic rocks and quartzo-feldspathic gneisses which records metamorphism in eclogite and high-pressure granulite facies conditions (770-900 ºC and >1.2-1.7 GPa) related to an Early Devonian subduction event (390-380 Ma; Ordóñez-Casado *et al.*, 2001) or two subduction events, in the Ordovician and Devonian (500-485 Ma and 390-385 Ma; Fernández-Suárez *et al.*, 2002). Field and microstructural observations suggest that this high-pressure metamorphic event occurred simultaneously with deformation and produced a pervasive tectonic fabric, which developed under a strain regime of bulk coaxial deformation (Llana-Fúnez *et al.*, 2004).

The objective of our work is to characterize the seismic properties of the rocks in the mantle-lower crust transition in Cabo Ortegal for the significance they may have when interpreting geophysical data such as reflection seismics and wide-angle data. We expect to gather information about the seismic properties of the Moho transition by using the analogy of Cabo Ortegal, following three different but complementary approaches: by inferring seismic velocities and anisotropies from the orientation of rock forming minerals, by measuring at high pressures (up to 600

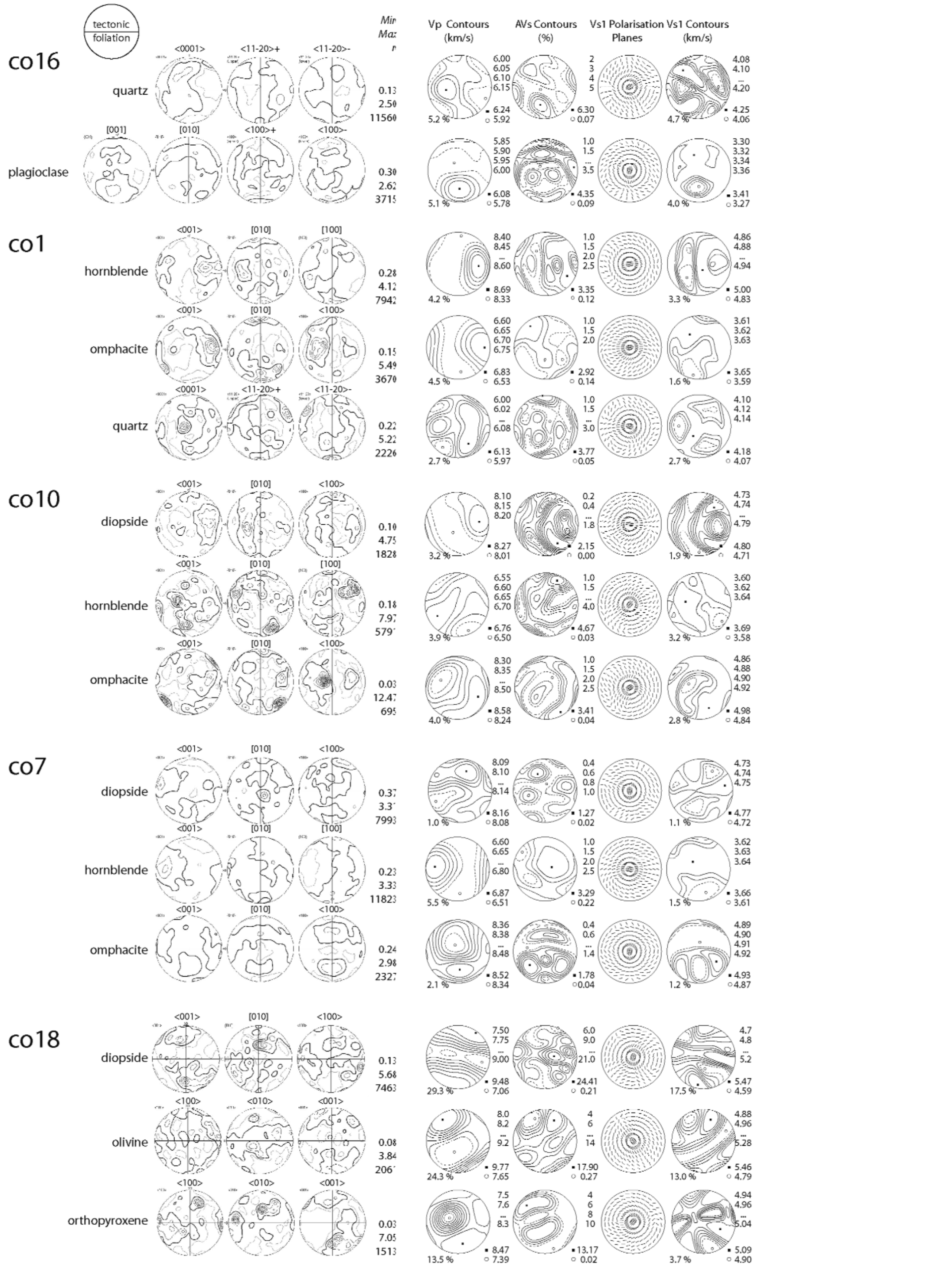

Figure 1. CPO and properties of seismic P- and S- waves for rock-forming minerals in 5 selected samples: co16 is a quartz-feldspathic gneiss from the Tarroiba cliff, co1 is an eclogite from the lighthouse of Cabo Ortegal, co10 and co7 are a mafic granulite and a metabasite from two localities near San Andrés de Teixido and co18 is a pyroxenite layer within the ultramafics from Vigía Herbeira. Crystallographic directions are indicated in the figure for each mineral phase. The contouring in CPO patterns is every multiple of uniform distribution with the exception of the first contour of 0.5, in grey. Contour 1.0 is a thick black line, which commonly outlines the patterns. The contouring for Vp, AVs (anisotropy shear wave) and Vs1 (fast) is done independently for each stereogram. The first contour is always a discontinuous line. Maximum and minimum are indicated by a black square and an open circle. The small tick lines in the Vs1 polarization planes indicate the orientation of the fast wave, Vs1.

MPa) the propagation velocities of seismic waves in rock cores, and by running a detailed geophysical profile across this transition in the field. We present here preliminary results on the seismic properties of 5 selected samples, firstly inferred from the crystallographic orientation of the main mineral phases, measured by electron back-scattered diffraction (EBSD), and secondly directly measured on the same five samples at high pressure at the Laboratory for High-Pressure Research at NRC, Canada.

Methods

Rock specimens were sampled sufficiently large to allow drilling of three mutually perpendicular 1 inch diameter by 3 inch long cores. Although most specimens were not oriented in the field, the thin section blocks were oriented with respect to structural framework (foliation and, when present, lineation). Ultra-polishing with colloidal silica (syton) in the selected thin section blocks was required for EBSD analysis. In this first set of samples a CamScan FEG scanning electron microscope was used with a horizontal stage and an electron beam tilted 70º. All five EBSD maps are stage maps, with a step size of the order of ~50 µm, slightly smaller than the mineral grain size (>200 µm), so that a sufficient number of grains could be analyzed. Data acquisition and mineral phase indexing was done using HKL software (now Oxford Instruments).

CPO data of quartz, plagioclase (anorthite), omphacite, garnet (almandine), hornblende, clinopyroxene (diopside), olivine and orthopyroxene (enstatite) was then processed using Mainprice free software (Mainprice and Humbert, 1994) to calculate first the seismic properties of individual mineral aggregates (Fig. 1) and then the bulk seismic anisotropy of samples according to EBSD-measured mineral modal abundances (Fig. 2). Mineral densities of end members in solid solutions were used and elasticity constants were taken from the literature: Heyliger *et al.* (2003) for quartz (d=2.6466 g cm^{-3}), Aleksandrov *et al.* (1974) for plagioclase (d=2.61 g cm^{-3}), Bhagat and Bass (1992) for omphacite (d=3.327 g cm^{-3}), Babuska *et al.* (1978) for garnet (d=4.16 g cm^{-3}), Aleksandrov and Ryzhova (1961) for hornblende (d=3.298 g cm^{-3}), Collins and Brown (1998) for diopside (d=3.327 g cm^{-3}), Chai *et al.* (1997) for enstatite (d=3.306 g cm^{-3}) and Abramson *et al.* (1997) for olivine (d=3.355 g cm^{-3}).

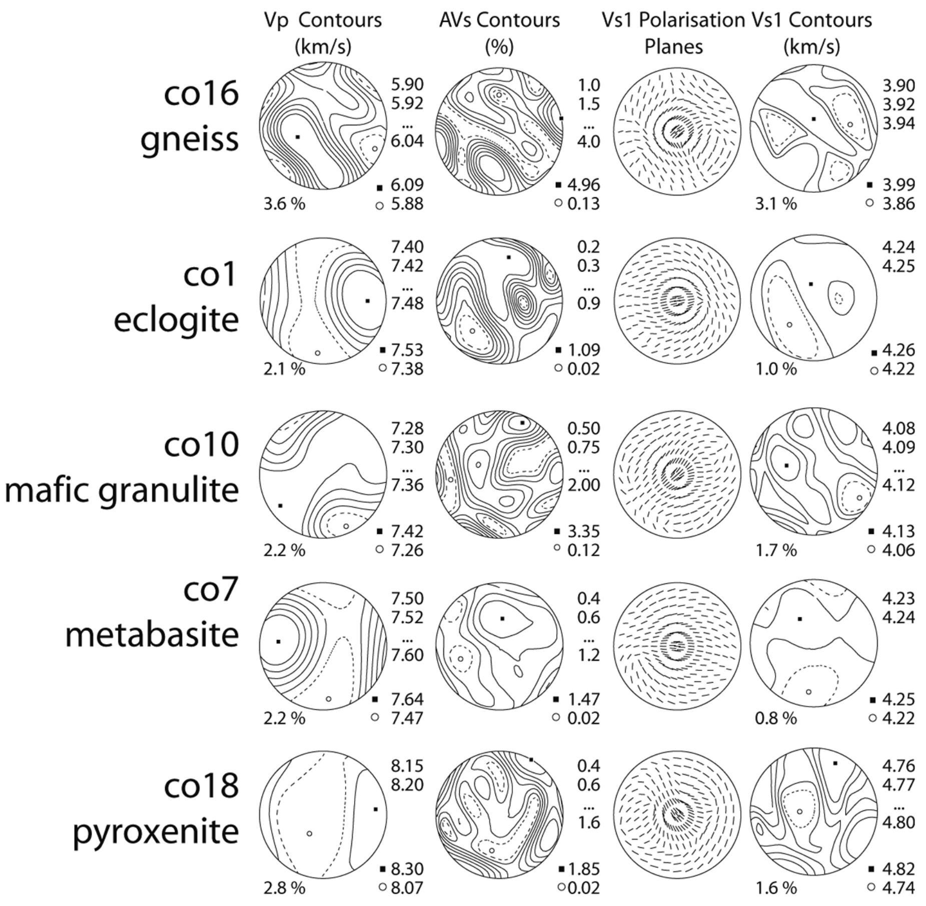

Figure 2. Bulk seismic properties of the samples calculated using the modal fractions as obtained by EBSD analysis and including garnet.

CPO and Vp-Vs data

CPO patterns for most rock forming minerals of the 5 selected samples are shown in figure 1. CPO in garnet has also been measured but it is not included in the figure due to the lack of crystallographic fabric and its very poor anisotropy for the propagation of seismic waves.

CPO patterns are relatively weak for quartz and plagioclase in the gneisses, despite the relative homogeneous microstructure in the sample. In all the other samples, the microstructure is not so homogeneous in that there are isolated large crystals (e.g. garnet, orthopyroxene, clinopyroxene). Both the CPO patterns and the calculated seismic velocities include all points. Thus, some of the CPO patterns may show clustering of points around large crystals, as may be the case in some hornblende and diopside (and quartz in the eclogite) grains. This clustering disappears in the stereoplots of seismic velocities. The propagation of P-waves seems to bear some relationship with the structural framework, mostly the foliation (since a lineation is rarely developed).

Discussion and conclusions

The development of anisotropy in the propagation of seismic waves (Vp and Vs) in rocks at depth is partly controlled by properties of rock-forming mineral phases when deformation processes produce a crystallographic fabric. Our EBSD work on selected and representative rocks across the mantle-lower crust transition in Cabo Ortegal show that quartz dominates velocity patterns in gneisses, omphacite in eclogites and hornblende in mafic granulites. The bulk anisotropy of Vp of all rock specimens measured with the EBSD is between 2-4% (Fig. 2), in keeping with laboratory measurements on the same samples

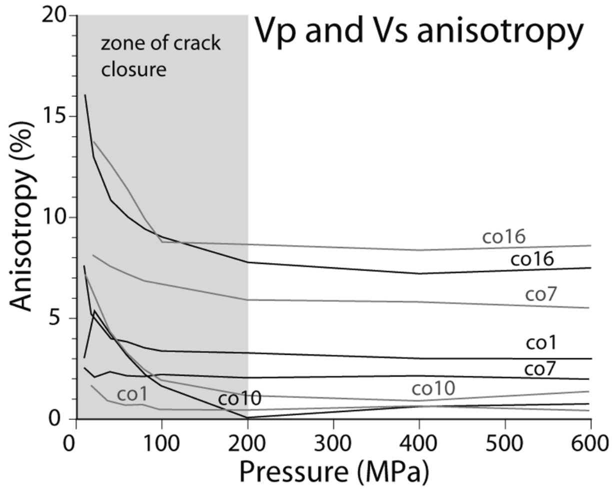

Figure 3. Anisotropy in the propagation of seismic waves with pressure, measured directly on drilled rock cores from the same samples. The reduction of anisotropy illustrates the effect of crack and small planar pore closure with increasing confining pressure.

(Fig. 3), with the only exception of the felsic gneiss where one of the main mineral constituents and most anisotropic, biotite, has not been measured.

The generally relatively weak CPO in the main mineral phases may be the product of the strain geometries developed during deformation, the dispersive nature of late high temperature recrystallization processes, and the abundance of garnet. In ultramafic rocks, relatively strong anisotropy in all phases interacts destructively when added together and renders the pyroxenites the least anisotropic of the sequence.

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