First results on the LPO-derived seismic properties of iron ores from the Quadrilátero Ferrífero region, southeastern Brazil

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ABSTRACT

Determinations of the LPO-derived seismic properties of iron ores were carried out in five samples with contrasting mineralogy (hematite, magnetite and quartz) and deformed in different conditions. All the samples are seismically *quasi*-isotropic or weakly anisotropic, which reflect (i) an absent or weak crystallographic preferred orientation (CPO) in some samples; (ii) the high modal content of magnetite, (iii) the relatively weak anisotropy of elastic stiffness of hematite single crystal. Such variables induce a low anisotropic seismic behavior even in high strained iron ores with strong preferred orientation of hematite. A plane of seismic transversal isotropy parallel to the foliation of the aggregates is developed in hematite $\pm$ magnetite aggregates and in itabirites with strong CPO of hematites. In these high strained aggregates, some relationships between the crystallographic axes of hematite and the propagation velocities can be observed. The magnitudes of P and S-wave velocities derived of hematite CPO are lower than the values experimentally determined in iron ores. Such a difference probably reflects other microstructural variable which were not taken into account in the present contribution (e.g. shape preferred orientation).

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1. Introduction

Theoretical knowledge of the seismic properties of rocks in general is an important tool to understand seismic wave propagation in field-based studies. It allows the determination of body wave velocities and anisotropies, the tridimensional propagation directions and other geophysical parameters, such as Poisson and reflection coefficients (Anderson, 1989; Lowrie, 1998; Lillie, 1998) of these aggregates. Such characterizations are carried out in two principal forms. The experimental approach consists on the propagation of an acoustic wave of known frequency in at least three different directions through the material and determination of its elastic properties. Seismic wave velocities and anisotropies in different directions are calculated from these results (e.g. Burlini and Fountain, 1993; Mainprice, 2007a). The second possibility is to calculate the elastic properties of an aggregate, based on the orientation and on the single crystal elastic properties of constituent minerals, their densities and modal proportion of each phase in the aggregate (Ji and Mainprice, 1988; Mainprice and Nicolas, 1989; Babuska and Cara, 1991; Mainprice, 2007a). The seismic properties determinations based on the crystallographic preferred orientation (CPO) is important on the study of anisotropy in crustal and mantle rocks (e.g. Mainprice and Nicolas, 1989; Burlini and

Fountain, 1993; Ben Ismaïl and Mainprice, 1998; Lloyd and Kendall, 2004; Mainprice, 2007a).

The commonest applied geophysical methods in mineral exploration industry are the potential field and electro-magnetic techniques, which allow the determination of lateral extension and deep of relatively shallow ore deposits (e.g. Salisbury and Snyder, 2004). However, as pointed out by Salisbury et al. (2000), the discovery of new shallow large ore deposits is becoming rare. Therefore, the application of deep exploration methodologies is required in order to fulfill the needs of such materials to the next generations. Seismic refraction studies have been used since the beginning of the XX's on the exploration of petroleum fields. However, the application of such techniques in mineral industry is a relatively new tool, with increasing potential for the next years (e.g. Salisbury et al., 2000; Salisbury and Snyder, 2004; L'Heureux et al., 2005). Thus, in this paper, we present the first results of the indirect determination of some of the seismic properties of iron ores from the Quadrilátero Ferrífero region, one of the largest deposits of iron ore in the world. The results present here are based on the CPO of the constituent grains within each of the selected aggregates, and might be useful for correlation with field-based seismic data in iron ore deposits.

2. Geological setting

The specimens used in this study were collected in different sites of the Quadrilátero Ferrífero (QF) region (Fig. 1), within the Cauê formation of the Itabira group, an archean/paleoproterozoic metasedimentary sequence that lies in the southeastern boundary of the São

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Fig. 1. Geological map of the Quadrilátero Ferrífero region, showing the main petrological units and structures in the region, as well as the location of the samples. Thin dashed lines separate structural domains, as suggested by Rosière et al. (2001), whilst thick dashed lines delineated contrasting metamorphic domains (Pires, 1995). MS – Moeda Syncline; DBS – Dom Bosco Syncline; GS – Gandarela Syncline; IS – Itabira Syncline. Modified from Rosière et al. (2001).

Francisco Craton, SE Brazil (Babinski et al., 1995; Alkmim and Marshak, 1998). This sequence hosts large iron ore deposits in the form of itabirites and polycrystalline hematite with variable contents of magnetite. Itabirite is a local denomination to the metamorphic rocks in which continuous bands of quartz or calcite are regularly interlayered with bands of hematite + magnetite ($\pm$ goethite) and are equivalent to metamorphosed banded iron formations. Other common rock types include quartzites, phyllites, dolomitic marbles, and amphibolitic itabirites, which commonly host the iron ore deposits. The Itabira Group was deformed and metamorphosed in three principal deformation events (Marshak and Alkmim, 1989; Chemale et al., 1994; Alkmim and Marshak, 1998). During the first deformation episode, two sets of coaxial recumbent folds were generated during progressive deformation. Such folds were nucleated and amplified concomitantly to the shear and transposition of the original sedimentary layering, and are then superimposed by tight to open folds and discrete shear zones, resultant from the less intense deformation events D2 and D3. The intense percolation of fluids during the deformation were responsible to the intense breakdown of magnetite to hematite (Lagoeiro, 1998), feldspar to white mica + quartz (Hippertt, 1998) and quartz leaching, which allow the formation of

relatively large bodies of pure hematite + magnetite ($\pm$ goethite) with variable dimensions from decimeters to 400 m of diameter mainly in the eastern portion of the QF.

The Quadrilátero Ferrífero region is characterized by the increasing of metamorphic and deformational conditions from west to east (Pires, 1995; Rosière et al., 2001 – Fig. 1). The low strain western domain was deformed in lower greenschist facies (Pires, 1995), where large sets of megasynclines and anticlines are also accompanied by the presence of subtle shear zones, and the banded iron formations still preserve sedimentary structures (Rosière et al., 2001). The increase of metamorphic conditions towards east, reaching lower amphibolite facies, is accompanied by the development of large sets of thrusts and transcurrent faults of different relative ages, and probably reflects the superimposition of tectonic events during the Paleoproterozoic and Neoproterozoic (Chemale et al., 1994; Almeida et al., 2005).

3. Specimens details

The samples collected in the western portion of the QF (CF-04 and MP-09) are predominantly formed by hematite ($\approx$ 60–70%) and magnetite ($\approx$ 30–40%), with goethite present as an alteration mineral.

The specimens are apparently isotropic, without flattened foliation or stretching lineation. Hematite and magnetite grain sizes are usually fine and vary between 20 to 30 μm (Fig. 2a), whilst larger crystals are scarce and made of magnetite, with grains $\leq$ than 100 μm . Grain boundaries are usually very irregular (Fig. 2b) and lobate contacts are observed between hematite crystals. The transformation occurred along octahedral planes {111} as can easily be seen by the triangular interlock of hematite stripes within the magnetite crystals (Fig. 2c). In isolated places, aggregates of small granular crystals of hematite with straight and sharp grain boundaries can occur, but crystals do not show any evidence of intracrystalline deformation. Although in hand samples iron oxide aggregates look massive, in fact fine hematite platy grains are strongly oriented in bands alternately distributed in the iron oxide rocks (Fig. 2d).

The increasing of metamorphic and deformational conditions towards the eastern portion of the QF is responsible to extensive breakdown of magnetite to hematite in a fluid-rich environment (Lagoeiro, 1998). As observed in the samples CE-39 and CE-41,

magnetite crystals occur only as relicts within granular and platy hematites. In relatively low strain domains, millimetric bands of granular hematite crystals are irregularly interlayered with bands of platy hematite (Fig. 2e). In high strain zones or in some fold hinges affecting hematite-rich bands, platy hematite is dominant, and granular forms only occur within low-strain, lens-shaped domains (Fig. 2f) (Rosière et al., 2001; Morales et al., in press). Grain size is predominantly very fine, between 15 to 30 μm , with rare isolated grains of 90–100 μm . Grain boundaries are usually sharp and straight in platy hematites, whereas in granular crystals the boundaries are usually more irregular and even lobate.

The sample of itabirite used in the present study present strong differences between the microstructures of hematite and quartz. Whilst hematite occurs predominantly as relatively large plates with an average axial ratio of 10:1 in XZ surface (visual determination), quartz grains are just slightly flat, with an axial ratio of 3:1 in the section parallel to the lineation and normal to the foliation. Hematite grain boundaries are usually straight and sharp and materialize the

Fig. 2. Photomicrographs of the iron ore samples CF-04 (a and b), MP-09 (c), CE-39b (d), CE-41 (e) and MR-08 (f), showing the main microstructures and mineralogy of the samples used to determine the seismic properties based on the orientations of hematite, magnetite and quartz. With the exception of the sample CF-04, all the others were sectioned parallel to the lineation and normal to the foliation. Cross polarized light, details in the text.

foliation plane (Fig. 2g). Quartz grain boundaries are also sharp and well developed and just in a few places it is possible to observe the occurrence of curved grain boundaries. Undulose extinction and subgrain boundaries are common features, mainly observed in large crystals. In the interface between quartz and hematite bands, quartz grain boundaries seem to mimetize the hematite fabric, becoming more flat than in relation to the internal parts of the band (Fig. 2h). Geochemical analyses of the studied samples presented by Lagoeiro et al. (submitted for publication) suggest that whereas the itabirites are chemically similar to the Lake Superior iron formations, the hematite bearing rocks are different due the smaller contents of quartz. The banded iron formations of QF region are characterized by a

small content (<1%) of CaO, Al_2O_3 , K_2O , MnO, MgO, Na_2O , TiO_2 and P_2O_5 , as well as trace elements.

4. Determination of CPO

The indirect determination of the seismic properties through the elastic constants of the minerals present in a given rock requires the knowledge of orientation of such minerals within the aggregates (Mainprice et al., 1990; Babuska and Cara, 1991; Mainprice and Humbert, 1994). In this paper, we explore the automated-mode functions of the electron back-scattered diffraction technique in a scanning electron microscope (EBSD/SEM – Prior et al., 1999) to

Fig. 3. (color image) – Examples of EBSD maps of the samples CF-04 (a), MP-09 (b), CE-39b (c), CE-41 (d) and MR-08 (e), showing all Euler angles. In these maps, variations in colors are directly related to variations in crystallographic orientation. Maps generated through the program Tango (Channel 5) Note the relatively increasing of grain size from the orientation map (a) to (c), and axial ratio from (a) to (d) in the case of aggregates of hematite + magnetite. (black and white image) – Examples of EBSD maps of the samples CF-04 (a), MP-09 (b), CE-39b (c), CE-41 (d) and MR-08 (e), showing all Euler angles. In these maps, variations in shades of gray are directly related to variations in crystallographic orientation. Maps generated through the program Tango (Channel 5) Note the relatively increasing of grain size from the orientation map (a) to (c), and axial ratio from (a) to (d) in the case of aggregates of hematite + magnetite.

Fig. 4. Single crystal seismic properties and individual pole figures for the main crystallographic forms of hematite, magnetite and quartz, as well as the elastic constants used to carry out such determinations. The calculations and plotting were done via Single_ANIS, Single_VpG and PF2k (Mainprice, 2007b).

characterize the orientation of the minerals in each of the oriented specimens briefly described above. All the samples were reduced to small blocks of $1.5 \times 1.0 \times 0.7$ cm, cut parallel to the stretching lineation and normal to the foliation planes in the cases of more deformed specimens, or just normal to the foliation in the case of absent lineation. These samples were mounted in polyester moulds and mechanically polished with diamond pastes of different grains sizes. As the analysis on the EBSD/SEM requires very flat samples, a chemical–mechanical polishing process using an alkaline solution of colloidal silica for all the specimens was also carried out, which took between 8 to 24 h, depending on the mineralogical composition of each specimen (Fynn and Powell, 1979; Morales et al., 2007). The orientation data were acquired in a SEM JEOL JSM 5510 with a Nordlys HKL-Oxford EBSD detector accomplished to the HKL-Channel 5™ suite of programs, operating at the Laboratory of Microscopy and Micro-analysis (MICROLAB), Department of Geology, Federal University of Ouro Preto. Measurements of preferred orientation of hematite crystals were carried out using the automatic beam scanning mode on a predefined areas of $\sim 3500 \times 2750 \mu\text{m}$ (three areas for each sample). The main setup parameters of the SEM/EBSD system were exactly the same for all the samples and were: accelerating voltage of 20 keV; spotsize of 70; tilt angle of 70° ; stepsize of $10 \mu\text{m}$. The minimum number of bands detected in the electron diffraction patterns was 4, whilst the maximum was 5, and the chosen mean angular deviation (MAD) = 2° . However, all the pole figures and seismic properties were calculated using an $\text{MAD} \leq 1^\circ$, increasing the precision of data. The average percentage of correctly indexed points for all the is 80%. Representative orientation maps of each specimen are presented in Fig. 3.

All the mathematics evolved in the calculation of single-crystal and CPO-derived seismic properties is extensively reviewed by Mainprice (1990, 2007a) and Mainprice et al. (2000) and will not be addressed here. The increase of temperature and pressure modify the elastic stiffness of single-crystals, usually with unknown derivatives. Thus, we adopted the convention of Mainprice (2007a) and Tatham et al. (2008) and use the elastic constants of hematite, magnetite (Landolt-Bornstein, 2007) and quartz (McSkimin et al., 1965) single-crystals determined in environmental conditions.

5. Seismic properties

5.1. Single-crystals

Hematite single crystal has three maxima parallel to the pole of $\{10\bar{1}4\}$ rhombs (Vs1, Vs2 and AVs), and one maxima (Vp) parallel to the prism $\{10\bar{1}0\}$ pole. It also has two minima (Vs1 and AVs) parallel to $[c]$ -axis and two minima (Vp and Vs2, respectively) parallel to poles of rhombs $\{10\bar{1}4\}$ and prism $\{11\bar{2}0\}$. The trigonal symmetry is reflected in stereoplots as a triple repetition of maxima or minima around $[0001]$, above the position of rhomb poles or subparallel to prismatic poles $\{10\bar{1}0\}$. Magnetite single crystal, on the other hand, has two maxima parallel to $[100]$ (Vs1 and Vs2), one maxima parallel to $[111]$ and one maxima parallel to $[110]$ (Vp and AVs, respectively). There are also two minima parallel to $[100]$ direction (Vp and AVs), and two minima parallel to $[110]$ and $[111]$ (Vs1 and Vs2, respectively). Quartz single crystal is highly anisotropic in relation to the seismic properties. Trigonal symmetry is reflected in the pole figures as the triple repetition of Vp and Vs2 maximum and Vs1 minimum parallel to the

Fig. 5. Calculated velocities and 3D distribution of compressional waves (Vp), fast and slow shear waves (Vs1 and Vs2) and shear wave birefringence (AVs) for the sample CF-04, as well as the pole figures of poles to $[0001]$, $\{10\bar{1}0\}$, $\{11\bar{2}0\}$ and $\{10\bar{1}4\}$ planes of hematite and $[100]$, $[110]$ and $[111]$ directions of magnetite. Seismic properties were calculated using the Voigt–Reuss–Hill averaging scheme and were plotted on the lower hemisphere. Pole figures of magnetite and hematite were plotted on the lower hemisphere, equal-area net, non-polar data, multiples of uniform distribution.

pole of rhomb planes, and AVs maximum parallel to the poles of second-order prisms. Vs2 maxima and AVs minima are both parallel to the quartz [0001]-axis (Fig. 4c).

5.2. Aggregates

The calculated seismic properties for the aggregates are summarized in Table 2 and were calculated through the softwares ANISch5 and VpG (Mainprice, 2007b), via Voigt–Reuss–Hill averaging scheme (Hill, 1952).

The samples collected in the western portion of the Quadrilátero Ferrífero region (Figs. 5 and 6), present a weak or absent CPO, which is reflected in their *quasi*-isotropic seismic properties. A comparison between the crystallographic fabrics pole figures and the seismic properties of the sample CF-04 demonstrates, however, that the $V_{p\max}$, $V_{s1\min}$, and $V_{s2\min}$ are parallel to the maximum concentration of poles to {10 $\bar{1}$ 0} of hematite and parallel to the maximum of [110] and [111] of magnetite. $V_{s2\max}$ and $V_{p\min}$ are parallel to the maximum concentration of [100] direction of magnetite. The seismic isotropic behavior is confirmed by similar values between maximums and minimums of Vs1 and Vs2, as well as by their non-orthogonal propagation directions, as expected in anisotropic rocks.

The distribution of the seismic waves in the sample MP-09 is different than the ones observed in the sample CF-04. In this sample, hematite fabric becomes slightly stronger, and modal proportion of magnetite decreases. This result in a slight increase of the anisotropy of the P and S waves, but the specimen can still be described as seismically isotropic. The poles to basal planes of hematite are smoothly concentrated parallel Z in the reference frame, whilst poles to prisms {10 $\bar{1}$ 0} and {11 $\bar{2}$ 0} are distributed along a wide girdle

subparallel to the foliation plane, and poles to {10 $\bar{1}$ 4} occurs along an irregular conical girdle around Z. Magnetites do not show preferred orientation. In relation to the crystallographic fabrics, $V_{p\max}$ is parallel to the maximum concentration of poles to {1010} of hematite, whilst $V_{s1\max}$ is subparallel to the maximum concentration of hematite [0001] axes and $V_{s2\max}$ is parallel to the maximum concentration of poles to {1014} rhombs (Fig. 6). The development of an incomplete and wide plane of transversal isotropy for P-waves reflects a weak alignment of hematite plates subparallel to the foliation in this sample.

The hexagonal symmetry of the seismic wave propagation becomes stronger in the highly deformed iron ores (samples CE-39b and CE-40). In these samples, [0001] axes are strongly concentrated parallel to the pole of the foliation, whereas the poles to {10 $\bar{1}$ 0} and {1120} are distributed along a continuous girdle, slightly oblique to the foliation plane, and rhomb poles are plotted along a conical girdle around Z (Figs. 7 and 8). Magnetites occur in modal proportions of $\approx$ 15% in these samples. Directions [100] (CE-39b) or [111] (CE-41) are subparallel to X. Distribution of [100] occurs along two discontinuous great circle girdles normal to the XY plane, whereas [110] axes concentrated in three single maxima and one small circle girdle around the lineation in the sample CE-41 (Fig. 8). In these samples, a plane of transverse isotropy subparallel to the foliation plane of reference system, for $V_{p\max}$, $V_{s1\min}$, and $AV_{s\min}$, is developed. A secondary plane of isotropy occurs parallel to the conical girdles of poles to {1014} rhomb, for the distribution $V_{s1\max}$, $V_{s2\max}$, distribution is controlled by the poles of a secondary rhomb (Fig. 8).

For the determination of seismic properties of itabirites (MR-08), we have considered the elastic constants of quartz and hematite, assuming a modal composition of 50% of each phase. These two single-crystals are seismically anisotropic and possess strong differences in

Fig. 6. Sample MP-09. See caption for Fig. 5.

Fig. 7. Sample CE-39b. See caption for Fig. 5.

density ($d_{\text{hematite}} = 5.255 \text{ g/cm}^3$ and $d_{\text{quartz}} = 2.65 \text{ g/cm}^3$). In this sample, hematite CPO is similar to the presented by sample CE-41 (Fig. 9). [0001]-axes of quartz are distributed along a continuous type-I crossed girdle centered in Y, whereas the poles to {1010} are distributed along an oblique girdle in relation to the foliation plane of the reference frame. The maximum concentration of poles of {2110} is parallel to the maximum of {1010} poles, but the girdles have different geometries, whereas poles to rhombs {1011} and {0111} are concentrated in single maximums in intermediate positions within the stereonet. The distribution of seismic waves in this aggregate (Fig. 9) is similar to the presented by the sample CE-41 (Fig. 8), with small variations in the magnitudes of velocities and anisotropies (Table 2).

6. Discussion

CPO-derived seismic properties of iron ores demonstrates that these rocks are seismically *quasi*-isotropic or weakly anisotropic, even in the case of high deformed aggregates. The lack of seismic anisotropy in these rocks seems to be the result of three different, but also concurrent aspects of such aggregates.

The first aspect is the absent or weak CPO, which results in very similar propagation velocities for both P- and S-waves in contrasting directions. This is observed in the distribution of seismic wave velocities of the samples CF-04 and MP-09, where the maximum and minimum propagation velocities are very close to each other. Also, the propagation direction of P- and S-waves is directionally independent, and no tridimensional variations on the velocities are observed, as demonstrated for the lack of contour velocities on the stereonets (Figs. 5 and 6).

The elastic properties of an isotropic medium are usually described by two independent coefficients, known as the Lamé constants λ (Lamé's coefficient) and μ (shear modulus — e.g. Nye, 1957; Montagner and Guillot 2002; Mainprice, 2007a) plus the density ρ . The first two parameters can be ascribed in the format of two-index Voigt system (Eq. (1)), as suggested by Mainprice (2007a):

$$\begin{aligned} C_{11} &= C_{22} = C_{33} = \lambda + 2\mu \\ C_{12} &= C_{23} = C_{13} = \lambda \\ C_{44} &= C_{55} = C_{66} = \mu \end{aligned} \quad (1)$$

Analyzing the elastic constant matrix for the sample CF-04, we can extract those values (Eq. (2)).

$$\begin{bmatrix} 2.3854 & .5301 & .5375 & -.0002 & -.0011 & -.0032 \\ .5301 & 2.4014 & .5461 & .0006 & -.0012 & -.0019 \\ .5375 & .5461 & 2.3937 & -.0021 & .0001 & -.0032 \\ -.0002 & .0006 & -.0021 & .9224 & .0014 & .0000 \\ -.0011 & -.0012 & .0001 & .0014 & .9274 & .0007 \\ -.0032 & -.0019 & -.0032 & .0000 & .0007 & .9199 \end{bmatrix} \quad (2)$$

In the above matrix, values of C_{44} , C_{55} and C_{66} are very close to each other ($\mu \approx 0.92$), which means that the shear wave velocity will be approximately the same in any direction. A similar but not direct consideration can be made in relation to the elements C_{12} , C_{23} and C_{13} ($\lambda \approx 0.535$), which have very similar values and give an approximately indication of the incompressibility of the bulk aggregate. The direct comparison between such values and the magnitude of propagation velocities of P-waves is not simple as in the case of shear waves, but it is expected that the velocity of compressional waves should be almost

Fig. 8. Sample CE-41. See caption for Fig. 5.

the same in any direction of the aggregate, as observed for the samples CF-04 and MP-09 (Figs. 5 and 6).

Another aspect that might influence the isotropy of elastic properties of these aggregates is the presence of a cubic mineral (magnetite) in the studied samples. Usually, isometric single crystals are optically isotropic, but can be anisotropic to many physical properties (e.g. Ahrens, 1995; Bascou et al., 2001; Xu et al., 2006; Mainprice et al., 2007a). The anisotropy of these properties, however, is usually lower than that of non-cubic minerals (Fig. 4). The relationships between the seismic properties of single-crystals in similar crystallographic directions are also an important characteristic. For example, the $Vs1_{max}$ is parallel to hematite [0001] axis, whereas for magnetite, this position is exactly the point of $Vs1_{min}$ (parallel to [100] direction). Thus, if a crystal of hematite has its [0001] axis parallel to the [100] direction within an aggregate, it is expected a general decreasing of $Vs1_{max}$, due a destructive interference of different single crystals properties. In contrast, constructive interference can also be observed, as in the case of $Vs2_{max}$, which is parallel to the poles of {1014} of hematite and [111] of magnetite, both lying in the same position of single-crystal pole figures (Fig. 4).

However, the lack of preferred orientation cannot be used as an argument to the weak seismic anisotropy presented by the samples CE-39, CE-40 and MR-8, all of them showing strong CPO of hematite. CPO is a well-known cause of anisotropy of seismic waves, either in mantle or crustal rocks (e.g. Mainprice and Nicolas, 1989; Burlini and Fountain, 1993; Ben Ismail and Mainprice, 1998; Mainprice et al., 2000; Montagner and Guillot, 2002; Lloyd and Kendall, 2005). The weak anisotropy resultant from the hematite CPO results of the relatively weak anisotropic behavior of hematite single-crystal elastic properties. The elastic stiffness of a mineral is controlled by the density of the elements within the structure of the mineral, its degree of structural

packing and by chemical bonding in which the elements are connected to each other (Love, 1944; Nye, 1957). Hematite ($\alpha\text{-Fe}_2\text{O}_3$) crystals have a relatively dense hexagonal closest packing structure (Blake and Hessevick, 1973), similar to the cubic closest packing found in magnetite crystals. This implies that the elastic properties of hematite single crystal will not be as much anisotropic as the elastic properties of a crystal of quartz or muscovite (Nye, 1957). For example, the anisotropy of P-waves is almost 28% in quartz single crystal and 11% in the case of hematite, whereas AVs is $\approx 43\%$ to quartz and 18% to hematite single crystal (Table 1 and Fig. 4). Thus, if the anisotropy of the elastic properties of a single-crystal is not strong, then the seismic properties of an aggregate of this mineral resultant from its CPO will be equally poor anisotropic (e.g. Mainprice et al., 2000; Lloyd and Kendall, 2005). Even in the case of a strong preferred orientation, such as the samples CE-41 and MR-8, the crystals are not perfectly aligned, and that small differences between the crystallographic orientation of minerals induce a reduction of anisotropy values.

Despite the low anisotropic seismic behavior of the high strained hematite ores, a plane of transverse isotropy for seismic wave propagation parallel to the foliation of hematite aggregates is developed in high deformed iron ores. This effect reflects a hexagonal-type seismic wave distribution whose symmetry axis is the pole of the foliation. The development of such planes reflects the strong alignment of hematite basal planes parallel/subparallel to the foliation in these samples. Such kind of distribution of seismic waves is very common in mica-bearing rocks with relatively high preferred orientations (Nishizawa and Yoshino, 2001; Takanashi et al., 2001). In the case of platy hematites, the alignment of basal planes is materialized by the strong alignment of hematite basal grain-boundaries parallel to the foliation, which generates a strong shape fabric. The presence of a shape preferred orientation can also induce

Fig. 9. Calculated velocities and 3D distribution of Vp, Vs1, Vs2 and AVs for an itabirite, assuming a modal proportion of 50% of hematite and 50% of quartz (sample MR-08). Pole figures showing the distribution of poles to 001, {10-10}, {11-20} and {10-14} of hematite, and [0001], {2-110}, {10-10}, {10-11} and {01-11} of quartz. Seismic properties were calculated using the Voigt–Reuss–Hill averaging scheme and were plotted on the lower hemisphere. Pole figures of hematite and quartz were plotted on the lower hemisphere, equal-area net, non-polar data, multiples of uniform distribution.

an effect of anisotropy of seismic waves (e.g. Mainprice and Nicolas, 1989; Burlini and Kunze, 2000; Wendt et al., 2003; Lloyd, 2004) and, in some cases, can make a considerable contribution to the whole effect of anisotropy of a given aggregate (e.g. Burlini and Kunze, 2000). Thus, it is possible that experimental measurements of the elastic constants of hematite aggregates (Ji et al., 2002; Liebermann and Schreiber, 1968) might indicate an effect of anisotropy which cannot be quantified in the determinations based on orientations of the hematite/magnetite and quartz.

Experimentally determined P-wave velocities at different pressures in different directions in Lake Superior iron ores (compiled in Ji et al., 2002) demonstrate that there is a tendency of increase of compressional wave velocities following the increase in pressure. A comparison between the experimentally determined P-wave velocities and CPO-derived ones demonstrates that the former velocities are lower than the CPO-derived ones for the hematite+magnetite aggregates. However, as the Lake Superior iron ores are structurally

and compositionally similar to the itabirites (regular interlayering between bands of quartz and iron oxides), the experimentally determined velocities of compressional waves in X and Z are similar to the Vp derived from preferred orientation of quartz and hematite in these directions. Seismic wave velocities experimentally determined by Liebermann and Schreiber (1968) for hematite aggregates are similar or even higher than the values tabled in Table 2. Such a variation between CPO-derived and experimental seismic velocities seems to be related to the presence or absence of quartz in these rocks.

7. Geological implications

It is commonly understood that metamorphic, igneous and even sedimentary rocks can be seismically anisotropic (e.g. Mainprice and Nicolas, 1989; Lloyd and Kendall, 2005; Valcke et al., 2006). From our observations, however, this seems not to be the case for iron ores, at least for the lattice preferred orientation-derived seismic anisotropy,

Table 1

Summary of the velocities and anisotropies of compressional (Vp), fast and slow shear waves (Vs1 and Vs2, respectively) and the main propagation direction in hematite, magnetite and quartz single crystals

Seismic waves magnitudes	HEMATITE	MAGNETITE	QUARTZ
Vp _{max.} (km/s)	6.86	7.47	7.03
Vp _{min.} (km/s)	6.14	7.16	5.32
AVp (%)	11.1	4.2	27.7
Vs1 _{max.} (km/s)	4.59	4.33	5.11
Vs1 _{min.} (km/s)	4.03	4.08	3.74
AVs1 (%)	13.1	5.9	31.1
Vs2 _{max.} (km/s)	4.42	4.33	4.69
Vs2 _{min.} (km/s)	3.81	3.93	3.3
AVs2 (%)	14.8	9.8	34.8
AVs _{max.} (%)	18.66	9.78	43.19
AVs _{min.} (%)	0.00	0.00	0.00
Density (g/cm ³)	5.255	5.26	2.65
Propagation directions	HEMATITE	MAGNETITE	QUARTZ
Vp _{max.}	// to [0001]	// to [111]	// to a rhomb pole
Vp _{min.}	// to {10-14} pole	// to [100]	sub// to {10-10} pole
Vs1 _{max.}	// to {10-14} pole	// to [100]	// to {2-1-10} pole
Vs1 _{min.}	// to [0001]	// to [111]	// to a rhomb pole
Vs2 _{max.}	// to {10-14} pole	// to [100]	// to the [0001]
Vs2 _{min.}	// to {11-20} pole	// to [110]	// to {2-1-10} pole
AVs _{max.}	// to {10-14} pole	// to [110]	// to {2-1-10} pole
AVs _{min.}	// to [0001]	// to [100]	// to the [0001]

because even in the high-strained aggregate that we have determined the seismic properties, their anisotropy are relatively weak. Despite of that, such a behavior can still be useful on the identification of iron ores through seismic studies. If we compare the results of iron ores with the results of sedimentary rocks (generally weakly anisotropic), the effect of anisotropy seems to be enough to be detected from field seismic data (e.g. Alkhalifah and Rampton, 2001; Valcke et al., 2006). This means that the velocities of seismic waves will dependent on the local propagation direction, and the polarization will depend of the type of seismic wave and the symmetric relationships of the elastic properties of the minerals that form these rocks. The main variables that control seismic anisotropy includes orientated cracks, faults and porosity, and the type and nature of their infilling material, shape and lattice preferred orientation of mineral phases, and their internal distribution in rocks layering, etc. (e.g. Montagner and Guillot, 2002; Mainprice et al., 2000; Valcke et al., 2006). However, the relative importance to each of these factors on the anisotropic seismic properties of iron ores is poorly understood. As far as we know, the characterization of the elastic properties of polycrystalline hematites had been carried out only by Liebermann and Schreiber (1968). Besides, measurements of P-wave velocities in contrasting pressures of Canadian iron ores were compiled in Ji et al. (2002). In both cases, the elastic stiffness of the aggregates and the seismic velocities were determined via ultrasonic procedures, but none of these works have a complete description of the mineralogy, structures, microstructures and crystallographic preferred orientation present in the rocks. Such information is essential if ones wish to determine what causes anisotropy in any type of rock. Besides, ultrasonic methods take into account all the bulk micro and mesostructural variables, which in almost cases are difficult to separate, even in the case of controlled experimental conditions, as the determinations of seismic properties in different temperatures and pressures (see the compilation carried out by Ji et al., 2002). Another limitation of this approach is the usual assumption that all the samples have hexagonal symmetry whose axis lies in a vertical position. In many cases, the specimens do not have such symmetry pattern, which invariably will lead to incorrect determinations of the seismic velocities, anisotropies and propagation directions. Therefore, the study of the effect of the lattice preferred orientation of hematite, magnetite and quartz on the seismic properties of iron ores seems to be a good first approach to examine one of these variables separately.

It is important to consider how the seismic waves propagate through an iron-ore deposit and through the host rocks. The LPO-derived seismic properties of iron ores are usually more isotropic than the anisotropy of seismic waves of quartzites (e.g. Lloyd and Kendall, 2005) schists (e.g. Takanashi et al., 2001) and marbles (e.g. Burlini and Kunze, 2000), which are common lithotypes usually associated to the banded iron formations and polycrystalline aggregates of hematite + magnetite. The seismic velocities of compressional and shear waves are also higher than the values present by these rocks, which can be useful on the identification of iron ore deposits based on seismic surveys. However, the identification of these deposits through the propagation velocities of seismic waves in different types of rocks and their anisotropies will only be allowed depending on the size and deep of the deposit, and the type of seismic survey used to do the deposit identification. Teleseismic and earthquake waves allow the identification of large scale structures with relatively low frequencies and consequently low resolution, mainly due the wavelength of these types of seismic waves, of an order of 1000 km. On the other hand, wide-angle controlled seismic-wave sources can give a better resolution deep and lateral resolution of structures and even ore deposits, but with a minor special extend. The studied iron ores are *quasi*-isotropic or weakly anisotropic, and are hosted by usually more anisotropic rocks. However, to be detected as an isotropic body 'sandwiched' between anisotropic rocks, the isotropy have to be uniform thought the ore body and persist in a relatively large lateral extension to be detectable with respect to the seismic wavelength (as in the case of shear zones, see Lloyd and Kendall, 2005). Taking into account only the seismicity generated during an earthquake episode, such a ore body would have to have some kilometers in extension. This is not common in the case of 'high-grade' iron ore bodies (hematite + magnetite aggregates), which usually have just up to few hundreds of meters of lateral extension. Banded iron formation deposits with hematite, magnetite and quartz in contrast, can be kilometric in extent and could be detectable in such a case. However, seismic reflection modes at interfaces between the ore deposits and their hosts provide a much better resolution (of decametric to hectometric order) of either lateral and deep extend of ore deposits and could be even used on the identification of high-grade iron ore deposits. Also, the interface between a host anisotropic rock and the *quasi*-isotropic or weakly anisotropic ore could be detectable in seismic survey due the differences on seismic reflections. As pointed out by Lloyd and Kendall (2005), there is a direct relationship between the reflection coefficients and the degree of preferred orientation (and hence seismic properties) of the aggregates. They also pointed out that there is an azimuthal variation of the reflections in the direction of the shear zone they have studied. Therefore, if the iron-ore bodies have an isotropic seismic behavior, and

Table 2

Summary of the maximum and minimum propagation velocities of Vp, Vs1 and Vs2 waves, individual anisotropies and degree of birefringence of shear waves (AVs) determined from the crystallographic orientation of hematite, magnetite and quartz within the studied iron ore specimens

Properties	CF-04	MP-09	CE-39b	CE-41	MR-8
Vp _{max.} (km/s)	6.76	6.67	6.67	6.72	6.16
Vp _{min.} (km/s)	6.72	6.60	6.55	6.49	5.99
AVp (%)	0.50	1.10	1.80	3.40	2.70
Vs1 _{max.} (km/s)	4.21	4.19	4.25	4.36	4.13
Vs1 _{min.} (km/s)	4.19	4.15	4.20	4.14	3.98
AVs1 (%)	0.50	1.10	1.40	5.10	3.70
Vs2 _{max.} (km/s)	4.20	4.18	4.20	4.20	4.02
Vs2 _{min.} (km/s)	4.18	4.15	4.16	4.10	3.95
AVs2 (%)	0.40	0.90	0.90	2.60	1.80
AVs _{max.} (%)	0.52	1.08	1.64	4.70	4.03
AVs _{min.} (%)	0.02	0.02	0.02	0.05	0.05
Density (kg/m ³)	5257	5256	5256	5256	3952
Hematite (%)	67.15	72.90	85.26	86.57	50.00
Magnetite (%)	37.85	27.10	14.74	13.43	0.00
Quartz (%)	0.00	0.00	0.00	0.00	50.00

the host rocks have a relatively high degree of preferred orientation, then multi-azimuth wide angle reflection seismic data could generate significant data to the identification of such ore bodies interlayered with anisotropic aggregates.

8. Conclusions

The determination of the LPO-derived seismic properties of hematite, magnetite and quartz in five samples of iron ore of contrasting mineralogy and degrees of deformation allow the following conclusions:

(a) Iron ores can be considered as seismically *quasi-isotropic* or weakly anisotropic aggregates even when they present a high degree of preferred orientation of hematite, the principal phase in all the analyzed phases. Such behavior is resultant of three concurrent main factors that, depending on the analyzed sample, and can occur simultaneously. These factors include (i) the random distribution of hematite and magnetite within the aggregates, mainly in the rocks collected in the western domain of the Quadrilátero Ferrífero; (ii) the high content of magnetite, a isometric mineral with weak anisotropy of the seismic properties and (iii) the relatively low degree of anisotropy of the elastic stiffness of the hematite single crystal, and consequently low degree of seismic anisotropy, leading to a weak effect of anisotropy of seismic properties in the iron ores with strong preferred orientation. Despite of that, some relationships between the maximum and minimum concentrations of hematite crystallographic axes and seismic velocities propagation directions can be made in such rocks.

(b) A plane of transversal seismic isotropy whose axis is parallel to the pole of the foliation is developed parallel to the foliation of the high-strained aggregates, reflecting a hexagonal-type of seismic distribution, similar to the observed in mica-bearing rocks. The strong shape fabric materialized by the grain boundary alignment in these rocks might have some influence on the seismic properties and experimental measurements are required to complement the data calculated here.

(c) The magnitudes of CPO-derived P-wave velocities in hematite + magnetite aggregates are similar to values experimentally determined by Liebermann and Schreiber (1968) in hematite-rich aggregates. However, they are lower than the values compiled by Ji et al. (2002) for banded iron formations, but are similar to the velocities of P-waves of itabirites. Such a variation reflects the quartz content in banded iron formations and itabirites, differently of the almost pure iron oxide aggregates;

(d) The differences of propagation velocities iron ores, in comparison with quartzites, schists and marbles (common iron ore hosts), as well as the differences on the anisotropies of these rocks may be useful on the identification of iron ore bodies using multi-azimuth wide angle reflection data. However, a successful application of seismic data on the identification of iron-ore bodies remains to be seen, and will depend on the dimensions of the bodies, the resolution of data and the spatial coverage of the seismic surveys.

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