Flow and electrical anisotropy in the upper mantle:
finite-element models constraints on the effects of olivine
crystal preferred orientation and microstructure

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Abstract

Long-period magnetotelluric (MT) data shows that electrical conductivity in the upper mantle is highly anisotropic. Agreement with electrical conductivity directions and seismic anisotropy fast directions suggests that anisotropic diffusion of hydrogen in olivine crystals controls the anisotropy of electrical conductivity in the upper mantle, since both seismic waves and diffusion are faster along the [1 0 0] axis of olivine. Thus, MT electrical anisotropy data, like seismic anisotropy, may be flow patterns in the upper mantle. However, observed electrical anisotropies are significantly higher than seismic ones, the influence of strain-induced crystal preferred orientations of olivine on upper mantle bulk electrical conductivities, the macroscopic electrical conductivity anisotropy of a series of naturally and experimentally deformed peridotites anisotropic finite-element model. These models, which fully take into account the microstructure: crystal and shape orientations, based on orientation maps obtained by indexation of electron back-scattered diffraction (EBSD) patterns, macroscopic electrical anisotropy factors ranging from 3 to 16. The intensity of electrical anisotropy depends on spatial distribution of the various crystal orientations may significantly enhance influence the anisotropy in strongly mantle rocks. The strongest anisotropy factors (>10) occur in mantle rocks in which deformation by dislocation creep not only crystal but also strong shape preferred orientations, even if the latter is masked by recrystallization. Higher factors (>100) observed in a few MT experiments imply however an additional, as yet unknown, mechanism controlling conduction at asthenospheric depths in these regions.

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1. Introduction

Long-period magnetotelluric (MT) data shows that electrical conductivity in the upper mantle, in particular the asthenosphere, is highly anisotropic. Anisotropy factors (electrical anisotropy is defined as the ratio between highest and lowest electrical conductivity, $A = \sigma_{max}/\sigma_{min}$) obtained for sublithospheric depths (100–200 km) range from >2 beneath central Australia (Simpson, 2001) to >100 in Central Germany (Leibecker et al., 2002; Gatzemeier and Moorkamp, 2005). Different processes may produce this anisotropy: (1) aligned melt lenses resulting from partial melting at the lithosphere–asthenosphere boundary, (2) aligned

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fecting films (e.g., graphite or sulphides) along fractures, or (3) anisotropic diffusion of hydro-
long oriented olivine crystals.

As by far the most abundant (~50–70 vol.%) single mineral. As the principal interconnected
conductivity is assumed to dominate the bulk
vity of the upper mantle down to the transition
km depth (Duba and Constable, 1993; Xu et
Moreover, agreement between electrical and
isotropy fast directions in regions as varied as
belt in Canada (Ji et al., 1996; Sénéchal et
the Appalachians (Wannamaker et al., 1996),
province (Eaton et al., 2004), central Australia
, and central Germany (Gatzemeier and
suggests that seismic and electrical
have a common cause. Seismic anisotropy
mantle is linked to coherent orientation of
tals over length scales of tens to hundreds of
m in the upper mantle (Nicolas and Christensen,
price et al., 2000). Like the propagation of
of shear waves), H+ dif-
er along the [1 0 0] axis of the olivine crystal
and Mackwell, 1990, Mackwell and Kohls.
Thus, preferred orientation of olivine [1 0 0]
roduce electrical anisotropy in the astheno-
ents at high temperature and moderate pres-
well as extensive data on naturally deformed
ks show that olivine at upper mantle con-
forms essentially by diolosion creep with
ip on the [1 0 0] (0 1 0) system, developing
preferred orientations (CPO) characterized of the [1 0 0] axis with the flow direction
(0 1 0) axis normal to the flow plane (see
mosi et al., 2000). Thus, if H+ diffusion
the dominant electrical conductivity mech-
electrical conductivity directions map flow
the asthenospheric mantle, similarly to fast
or P- or Rayleigh waves
directions. The high electrical conductiv-
at asthenospheric depths (≥0.1 S/m) (e.g.,
, 1995; Simpson, 2001; Gatzemeier and
which cannot be explained based on
cuctivity of dry olivine, provide evi-
dominant role of H+ diffusion in olivine in the
duction at these depths (Hirth et al., 2000).
versely, in contrast to seismic anisotropy data,
ulk electrical anisotropies (Simpson, 2003;
and Moorkamp, 2005) are significantly
A fundamental parameter to this question is how the
electrical anisotropy produced by anisotropic H+ dif-
sion in individual olivine crystals is transferred to the
poly-crystal (rock) scale.

In this article, we use finite-element modelling
(Garboczi, 1998) to evaluate macroscopic electrical
isotropy associated with intracrystalline diffusion of
H+ in olivine in three naturally deformed peridotites
and a synthetic dunite deformed in simple shear at
high temperature and pressure. These models fully
ake into account the microstructure of the rocks: both
ystal and shape preferred orientations are explicit-
described based on orientation maps obtained by
exion of electron back-scattered diffraction (EBSD)
erns. Predicted anisotropies are then compared (i)
to anisotropies predicted by simple averaging models
(Voigt–Reuss–Hill), which are traditionally used to
culate seismic and thermal properties of upper mantle
ks (e.g., Ben Ismail and Mainprice, 1998; Tommasi
al., 2001), and by random resistor networks (Simpson
and Tommasi, 2005) and (ii) to electrical anisotropies
ferred from long period MT data.

2. Modelling electrical properties of composite
media

The prediction of effective electrical properties of
composite media from the properties and microstruc-
ture of its components has applications in many areas,
from material sciences to geophysics (e.g., Archie, 1942;
Schmeling, 1986; Coverdale et al., 1995). A large num-er of approaches may be used to predict the electrical
conductivity of a composite. In the absence of informa-
tion on texture, effective medium theory using simple
duction models can be applied to obtain bulk
electrical properties of the composite. These models
describe heterogeneous, two-phase materials where a
phase of often high electrical conductivity is embed-
ed into a low-conductive, interconnected matrix. The
fective conductivity of such medium is dependent
only not of the electrical conductivity of the constituents,
but also on their volume fraction, geometry and inter-
nnectivity. The simplest mixing models are the series
and parallel models (e.g., Schulgasser, 1976). In reality,
ither of them is able to describe precisely the
ixture of materials. More complex and realistic expressions
dcribing the conductivities of mixtures with isolated
high conductivity phases have been developed, like the
axwell (1881) model for isolated conducting spheres
and
ions on a two-phase mixture are given by the Hashin and Shtrikman (1962) model. Other results, such as Andrade’s (1952) effective medium theory, show that the effective conductivity (Del Rio et al., 1992) of two-phase 2D materials, or the simple geometric average (Shankland and Duba, 1990), lie between upper and lower bounds. An effective-medium theory of a mixture with phases of anisotropic conductivities was derived by Bernasconi (1974). Extending Hashin’s (1993) effective-medium approximation to a mixture with different conductivity distributions among the orthogonal directions, this approach allows calculation of effective conductances in three orthogonal directions. A further extension allows describing more complex structures (Toledo et al., 1992).

Phase percolates in a mixture, i.e., if it is electrically connected along the entire length of the model, it has the ability to greatly influence the overall properties (Patrick, 1973). In this case, effective medium theory can be used to determine a homogeneous distribution becomes relevant (Gueguen and Palciauskas, 1994). Percolation and renormalization methods describe the activity of so-called clusters, i.e., areas with specific properties, on networks with extreme conditions of the physical properties of their constituents (Aharony, 1992). These methods allow the conductance of networks with statistical distribution of open or connected bonds. Random networks (e.g., Maddon, 1976; Bahr, 1997) or lattices (Bigalke, 1999) have been successful in calculating electrical conductance of heterogeneous materials. For modelling anisotropic conductivity, the 3D combined effective medium and percolation approaches (Gatzemeier and Tommasi, 2005) by introducing a topological parameter: the percolation.

These methods have been used to model electrical conductivities of crustal and mantle rocks under various conditions (e.g., Shankland and Waff, 1977; Jodicke, 1999; Roberts and Tyburczy, 1999; 2000b). For the upper mantle, most studies consider the anisotropic conductivities described by (i) a single phase system dominated by the electrical conductivity of dry olivine (e.g., Constable et al., 1992; Dubble, 1993; Xu et al., 2000b) or (ii) a two-phase system with dry olivine as a matrix with an embedded inhomogeneous conductivity phase, like partial melt (e.g., Roberts and Tyburczy, 1999). Indeed, olivine, which is by far the most abundant upper mantle mineral, shows little variation in electrical conductivity under dry conditions: the largest and smallest electrical conductivities may be described as isotropic, even in presence of strong olivine lattice preferred orientations.

On the other hand, under high temperature and hydrated conditions, electric conduction in the upper mantle is dominated by H+ diffusion in olivine (Karato, 1990). According to the Nernst–Einstein equation, which relates ionic conductivity to charge carrier concentration and mobility, conductivities are proportional to diffusivities of the charge carrying species:

\[
\sigma_i^{H^+} = \frac{f c_i^{H^+} D_i q_i^2}{k_B T},
\]

where \(\sigma_i^{H^+}\) is the electrical conductivity, \(f\) a numerical (correlation) factor approximately equal to unity, \(c_i^{H^+}\) the concentration (m⁻³), \(D_i\) the diffusivity along direction \(i\) and \(q_i\) is the electrical charge of the charged species (Karato, 1990). Hydrogen diffusion in single crystal olivine is fast and highly anisotropic (\(D_{[100]} \approx 20 D_{[010]} \approx 40 D_{[001]}\), Kohlstedt and Mackwell, 1998). This conduction mechanism leads therefore to high electrical conductivity and anisotropy at the crystal scale and, if these properties are preserved at the polycrystal and larger scales, high conductivity and anisotropy in the upper mantle (e.g., Lizzaralde et al., 1995; Simpson, 2003; Gatzemeier and Moorkamp, 2005).

To evaluate the bulk electrical conductivity anisotropy that may result from anisotropic H⁺ diffusion in the upper mantle, Simpson and Tommasi (2005) developed a model based on random resistor networks that accounts for realistic upper mantle mineralogical compositions, crystal preferred orientations, and rates of H⁺ self-diffusion in olivine and pyroxenes. This model predicts a mean electrical anisotropy factor less than 3 for a peridotite mantle. However, this model, as all methods described above, does not take into account the microstructure of the samples.

### 3. FEM modelling of effective electrical conductivities of polycrystalline aggregates

In a 3D finite-element model (FEM), the effective electrical conductivity tensor of a polycrystalline aggregate depends on the full microstructure, i.e., on the orientation and shape of the crystals that compose the aggregate, on their spatial distribution, and on their electrical conductivity tensors. The microstructural information on the individual grain orientations and the number of grains per unit volume are factored into the model to calculate the bulk electrical conductivity of the aggregate.
The orientation of the crystallographic lattice is described by a macroscopic reference frame. Each grain is represented by a set of neighbouring pixels with similar composition and orientation; a grain may be defined by a few hundred pixels, depending on the resolution of the image. The crystal orientation data is used to orient the conductivity tensor of each pixel. Bulk electrical conductivity of the aggregate is then obtained using an infinite-element modelling (FEM) by calculating electrical conductivity densities that are produced by a constant current (Garboczi, 1998).

The procedure employed in this study to obtain the anisotropy for a given polycrystalline aggregate is divided into the following steps:

1. Starting from thin sections of naturally or experimentally deformed mantle rocks, the texture and CPO are determined by crystallographic orientation mapping using the electron-backscattered diffraction (EBSD) approach, in which the orientation of the crystal lattice relative to the macroscopic reference frame is obtained by backscattered electron imaging of electron-backscattered diffraction patterns from a thin sample (EBSD).

2. Each pixel in a EBSD map contains therefore information about both the mineral phase and the orientation of the crystal lattice of the grain it belongs to, given as a set of Euler angles (Bunge, 1982).

3. In the present study we use EBSD data obtained from 2D thin sections but this method can be applied for 3D blocks as well by 3D-EBSD.

4. Orientation mapping and indexing are performed automatically, the EBSD map contains information on the misinterpreted diffraction patterns or artifacts due to fractures in the sample. Noise-reduction techniques are used for cleaning the map of isolated noise (spikes) that differ in composition or orientation from their neighbours. The orientation of each grain relative to these points is compared to the orientation of its eight nearest neighbours and replaced with the orientation of the neighbour that displays the lowest indexation error in the EBSD map. A similar procedure is used to fill up the gaps, i.e., areas without a complete determination of the crystal orientation. Correction of the diffraction pattern was not possible in some cases. In order to fill the procedure a minimum of five nearest neighbours is imposed. Small orientation deviations were allowed.

5. In step 3, each mineral phase of the EBSD map is associated with a diagonal electrical conductivity tensor, the elements of which contain the conductivities along the main crystallographic axes. Depending on the mineral phase and Bunge–Euler angle of each pixel, the single crystal conductivity tensor is rotated into the macroscopic reference frame (XYZ of the model) and assigned to the element. In this way a two- or, in cases using 3D-EBSD data, a three-dimensional model is obtained that has in each point a 3D electrical conductivity tensor, which is a function of the local composition and crystallographic orientation.

6. In step 4, the electrical anisotropy of the conductivity model is calculated using the finite-element code elecfem3d (Garboczi, 1998). As input files the program expects a 3D digital image consisting of cubic pixels in which each pixel can have a different phase characterized by an arbitrary symmetry conductivity tensor. The size of the digital image can vary in each direction and is limited only by the available computational resources. Periodic boundary conditions are applied to avoid influence on the calculation from the sample limits (end effect, Nye, 1985). These boundary conditions imply that the sample is representative for a larger region and hence that the model may be extended by repeating periodically the input digital image. In this paper we use digital images obtained from thin sections, and hence a 2D microstructure, but tensor (3D) conductivity properties. Extending the model along z direction via periodic boundary conditions results in a 3D conductivity model with a microstructure invariant along z-direction. For an applied (external) electric field \( \vec{E}^{ext} = (E_x, E_y, E_z) \) the program calculates the resulting average current density \( \vec{j}^{ext} = (j_x, j_y, j_z) \). These current densities are related to the electrical field by the effective electrical conductivity tensor \( \vec{j}^{ext} = \sigma^{eff} \vec{E}^{ext} \).

\[
\sigma^{eff} = \frac{j_i}{E_i} = \frac{1}{E_i} \int \sigma dV,
\]

(2)

For an effective conductivity tensor with orthorhombic symmetry (diagonal form if \( x, y, z \) are main crystallographic axes) \( \sigma^{eff} = \frac{j_i}{E_i} \) returns the main effective electrical conductivities. The electrical anisotropy is the ratio between the conductivities along two (orthogonal) directions follows as \( A = \frac{\sigma^{eff}}{\sigma^{eff}} = \frac{j_i E_j}{j_j E_i} \).

Comment: The last step, transferring the currents to the geometry of the model, is done by filling the digital image with values of \( \sigma^{eff} \) that are obtained as \( \sigma^{eff} = \frac{j_i}{E_i} \) from the effective conductivity tensors. The angle between the eigenvectors of the effective conductivity tensors is determined to be the angle between the orientations of the crystals that contribute to the respective element.

In summary, the workflow starts with EBSD to get the orientation data. This data is then used to determine the orientation of each grain relative to its nearest neighbours. The orientation is corrected by filling up gaps and correcting for noise. The orientation is then used to calculate the electrical conductivity tensor for each pixel. Finally, the anisotropy of the conductivity model is calculated and the currents are transferred to the geometry of the model.
element program. In the general case, the bulk conductivity tensor of an aggregate, which is described by crystals with orthorhombic tensor properties, not orthorhombic but triclinic, resulting in six three independent coefficients. However, for O the bulk conductivity tensor has a simpler form than can be approximated by an orthorhombic symmetrical case, alignment of the macroscopic coordinate system with the CPO results in small off-diagonal elements in the electrical currents calculated along \( j_x, j_y \), which are directly used to obtain approximated conductivities. The models presented in this study fulfill this condition, since the EBSS maps were performed on samples parallel to the flow direction and normal plane, which contain the maximum concentration of both the [1 0 0] and [0 1 0] crystallographic planes. In the general case, a set of numerical solutions for different sample orientations (\( \geq 6 \)) is required to calculate the six independent coefficients of the triclinic conductivity tensor (Nye, 1985). A full 3D model is then applied to each sample.

**Evaluating electrical conductivity anisotropy in mantle rocks**

By the method described in the previous section we can calculate the influence of hydrogen diffusion on conductivity and anisotropy in four mantle peridotites with varying microstructures and olivine CPO. Generally speaking, the electrical conductivity is the conduction of all charge carriers: \( \sigma = \sum \sigma_i \), and mechanisms active in mantle rocks, we consider only conduction due to hydrogen diffusion in the presence of water, is assumed to dominate conductivity (Gatzemeier and Tommasi, 2005).

For olivine we use hydrogen diffusivities along the crystallographic axes based on results of Mackwell et al. (1990) and Kohlstedt and Mackwell (1998):

\[
10 \times D_{{\text{ol}}[010]}^{\text{MK90}} \approx 10 \times D_{{\text{ol}}[010]}^{\text{MK98}} \quad (\text{MK90})
\]

\[
20 \times D_{{\text{ol}}[010]}^{\text{MK90}} \approx 40 \times D_{{\text{ol}}[010]}^{\text{MK98}} \quad (\text{MK98})
\]

(3)

where diffusion is also anisotropic in Mg-rich enstatites (Stalder and Skogby, 2003). We assume that diffusion is anisotropic between those along olivine [0 1 0] axes:

\[
A = \frac{\sigma_i}{\sigma_j} = \frac{D_i}{D_j}. \quad (5)
\]

According to Eq. (1) electrical conductivity depends not only on hydrogen diffusivities but also on hydrogen concentration. Water concentration in the upper mantle is still controversial and absolute value for the electrical conductivities due to H\(^+\) diffusion cannot be obtained. However, anisotropy \( A \) is given by the ratio between the electrical conductivity in two orthogonal directions. According to Eq. (1), if hydrogen concentration is assumed to be the same in all crystals, the anisotropy does not depend on the hydrogen concentration, but only on the ratio between the diffusivities in the two directions.

We assume identical hydrogen concentrations in all crystals. Therefore, olivine and enstatite conductivity tensors can be normalized by \( \sigma_{[100]} \), the electrical conductivity due to H\(^+\) diffusion in olivine along the olivine [1 0 0] axis. The resulting electrical conductivity tensors contain electrical conductivities relative to \( \sigma_{[100]} \), without further requirement about absolute water content in the minerals. Using these normalized conductivities allows to calculate bulk electrical conductivities relative to \( \sigma_{[100]} \) and anisotropies of the sample.

**4.1. Sample descriptions**

The studied samples are three naturally deformed peridotites and a synthetic dunite deformed in simple shear at high temperature and pressure. The selected samples have olivine contents (85–100%) that are significantly higher than typical asthenospheric mantle compositions (50–70% olivine), because the aim of the study is to evaluate the effect of the intrinsic anisotropy of H\(^+\) diffusion in olivine and of the olivine preferred orientations on the macroscopic electrical anisotropy. The studied peridotites display a wide range of microstructures (variations in grain size and shape, Fig. 1) that allows testing the effect of the microstructure on the electrical conductivity.

EPTA3 is a dunite xenolith (>95% olivine) from the Torre Alfina volcanic field in the Appenines (Italy) that shows a coarse-grained microstructure, characterized by mm-scale anhedral olivine grains (1–10 mm); the larger grains show widely spaced subgrain boundaries normal to the grain elongation (aspect ratio 2:1). This microstructure, characterized by a very weak shape

Fig. 1. Crystallographic orientation maps obtained by indexation of electron backscattered diffraction (EBSD) patterns for the four studied peridotites. Gray scale indicates the angular misorientation of the olivine [100] axis relative to the macroscopic X direction, which corresponds to the lineation (maximum grain elongation) in the naturally deformed peridotites FRB1359 and EPT A3, to the extension direction in harzburgite 900A87, and to the shear direction in dunite PO342. All maps were performed in the XZ structural plane, i.e., parallel to the flow direction and normal to the flow plane.

FRB1359

3 mm = 200 steps

900A87

50 µm = 50 steps

EPT A3

2 mm = 20 steps

PO342

25 µm = 50 steps

FRB1359 is a harzburgitic xenolith (82% olivine, 15% enstatite, 3% garnet) from the Premier Mine in South Africa. This sample is a high-temperature mylonite, which shows an almost fully recrystallized and highly re-equilibrated microstructure, partially recrystallized olivine porphyroclasts are locally preserved. Enstatite occurs as cm-scale porphyroclasts elongated parallel to the foliation with very fine-grained recrystallization tails (<20 µm). The elongation of the olivine and pyroxene porphyroclasts and the alignment of the small tabular olivine crystals give rise to a well-developed foliation and lineation. These mylonites are interpreted as formed by high strain rates under very...

High temperature conditions ($T > 1400^\circ$C, Boullier and Ohimoto, 1975). The EBSD map used as the input for the analysis was performed in a domain composed essentially of recrystallized polygonal olivine grains.

Fig. 7 is a fine-grained mylonitic harzburgite (75% enstatite # = 91; 20% opx; 3% cpx; 2% sp) collected in a low-temperature mylonitic zone in the Wadi Khubta ophiolite that was further experimentally deformed in extension at 1200°C and 300 MPa at a rate of $10^{-5}$ s$^{-1}$ (Ben Ismail, personal commun.). It shows a well-developed lineation parallel to the strain direction marked by the shape-preferred orientation of olivine porphyroclasts (aspect ratios up to 10 and the orientation of fined-grained olivine-rich, respectively. Olivine porphyroclasts are up to 1 cm long and show ondulose extinction, closely spaced (100) subgrain walls, and irregularly shaped subgrain boundaries. Enstatite porphyroclasts are up to 1 cm long and the average grain size is 2–4 mm. The recrystallized matrix, which is mainly composed of olivine grains up to 100 mm in diameter, represents more than 70% of the sample. EBSD mapping was performed in a domain devoid of enstatite with the X-axis parallel to the imposed extension direction.

PO342 is a synthetic dunite (100% olivine) deformed in simple shear at 1200°C and 300 MPa at a strain rate of $6 \times 10^{-5}$ s$^{-1}$ up to a shear strain of 5 (Bystricky et al., 2000). It is composed by a very homogeneous recrystallized matrix (95% volume) composed by 3 μm-wide equant or slightly elongated grains that surround two types of porphyroclasts. The first is formed by highly elongated ribbon grains (e.g., the black crystal that cuts the upper right domain of the EBSD map, Fig. 1) that are in an orientation for easy slip on (0 1 0) [1 0 0]. Their aspect ratios, which may attain 1:25, are consistent with the finite strain imposed to the sample. The second type of porphyroclasts (white in the map, Fig. 1) show lower aspect ratios (1:10), more subgrains and deformation features, and have an oblique lattice orientation. Both types of porphyroclasts have average grain areas equal to those of the starting grains (equivalent diameter of 20 μm), suggesting that their shapes are due entirely to strain by dislocation glide. EBSD mapping was performed in a thin section secant to the cylinder within...

200–300 /H9262m to its outer edge, the X-axis is parallel to the direction. Samples show olivine crystal preferred orientation (CPO, Fig. 2) characterized by a strong concentration of [0 1 0] parallel to the flow direction (inferred from the lineation in the EPTA3 and FRB1359, respectively). The [0 1 0] direction is aligned to the flow plane, with some dispersion in a normal to the flow direction in all samples except sample 3. These CPO are typical of deformation creep under high temperature conditions, characterized by activation of \([1 0 0] \{0 k l\}\) syste-
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105.4% for the FB1359 \((J=5)\) to 139.1% for EPTA3 \((J=10)\). PO342 that has the strongest CPO \((J=13)\) shows a slightly lower electrical conductivity anisotropy (132%).

4.2. FEM results

The finite-element modelling evaluates the current densities along the sample coordinate axes \(x, y, z\) for an applied electric field \(E = (1, 1, 1)\). Current densities and electrical field are related via the electrical conductivity tensor (Eq. (2)). Since a detailed description of the microstructure is only available for the XZ plane (2D) and the extension of the model in the third dimension is ensured essentially by the periodic boundary conditions, we only consider the current densities in the \(x\) and \(y\) directions. Comparing anisotropies of 2D and 3D random resistor networks, Labendz (1999) only observed dependence on models’ dimension and intrinsic electri-
Magnetotelluric (MT) data provide information about depth-dependence of horizontal electrical conductivities. Thus, MT data only provides constraints on the electrical anisotropy within the horizontal plane. Orientation maps were obtained in thin sections cut perpendicular to the foliation (flow plane which normal is the Z structural direction) and parallel to the lineation (flow or X structural direction). The models evaluate the anisotropy in this plane, for which the microstructure is fully described. They may be used therefore to estimate the horizontal electrical anisotropy in a vertical shear zone, in which both the flow direction and the normal to the flow plane are horizontal. To estimate the horizontal electrical anisotropy that would be produced in a horizontal shear zone, like the one that would form in the asthenosphere by viscous drag in response to the plate motion (Tommasi, 1998), we need to calculate the anisotropy within the flow plane. To evaluate this anisotropy, we run a second series of models in which we assume that (i) the microstructure (grain shapes and orientation spatial distribution) in the flow plane is similar to the one in the plane normal to the foliation that was analyzed by EBSD mapping and (ii) the CPO is rotated.

By evaluating orientations intensities (J factor), Voigt–Reuss–Hill (VRH), and finite-element (FEM) electrical anisotropies of the studied samples, all samples show similar and strong LPO. Thus, as far as grain boundaries are not the dominant charge carriers, the discrepancy between anisotropy in the $x$–$y$ plane in the present 2.5D models and in a full 3D model is expected to be small. Electrical conductivity anisotropy is thus calculated as the ratio between the current densities along $x$ and $y$ directions as given in Step 4 for models using either the MK90 or the KM98 olivine H$^+$ diffusivity tensor (Eq. (3)).

The electrical anisotropy is calculated using the MK90 and KM98 olivine H$^+$ diffusivity tensors. The anisotropy in the X–Z plane is calculated using the VRH method, while the one in the Y–Z plane is calculated using the FEM method.
models, is better observed in models that evaluate the anisotropy in the XZ plane (normal to the foliation) using KM98 H⁺ diffusivity data. In this dataset, anisotropy factors vary between 5 and 7 for samples FRB1359 (J = 5) and EPTA3 (J = 11), respectively.

Anisotropy factors calculated for the XY plane, i.e., within the flow plane, are usually lower than those obtained in the XZ plane, but the difference is usually small and the trend may be even reversed for those samples displaying a strong dispersion of [0 1 0] and [0 0 1] normal to the flow direction, like PO342 or FRB1359. Calculated anisotropy factors do not differ significantly for models using MK90 or KM98 H⁺ diffusivity data in spite of the significant lower intrinsic anisotropy of the olivine crystal of the latter data, which has a single crystal anisotropy factor of 40 instead of 100 (Fig. 4). However, models using MK90 data tend to display a stronger difference between the anisotropies observed within and normal to the flow plane.

A comparison of VRH and FEM results show that FE electrical anisotropy factors systematically exceed the VRH ones, except for sample 90OA87, for which FEM and VRH models using MK90 diffusivity data give similar results in the XZ plane (Fig. 5). VRH models should thus provide good lower bound estimations of upper mantle electrical anisotropies. In general the FEM results confirm the tendency observed for the VRH averages of increasing electrical anisotropy with increasing CPO intensity. However, the FEM results show that the spatial distribution and, hence, the interconnectivity of the various CPO components can significantly enhance the electrical anisotropy.
A closer analysis of the FEM results for sample 900A87 highlights the effect of these two factors on the electrical anisotropy:

Anisotropies calculated for sample PO342 are similar to those predicted by the VRH model. Indeed, PO342 and PO342 show a weak variation in J-h, that leads to similar values for P-, S-wave, and VRH electrical anisotropies (Fig. 2), but significantly in their FE electrical anisotropies (Fig. 3). High j_x and low j_y current densities for PO342 in comparison with EPT3 (Table 1) indicate that concurrent diminished electrical conductivity along the lower current density direction may explain the anisotropy of PO342 (Fig. 1), which is characterized by elongated lenses of similar CPO parallel to the glide direction X (porphyroclasts or recrystallized domains that had originated from a single porphyroclast). These highly elongated porphyroclasts correspond to grains in easy glide orientations, characterized by [1 0 0] semi-parallel shear direction (X). These lenses form thus a complex pathway along X and Y, and conductivity path along Y, resulting in a higher electrical anisotropy than the one that would be produced by a single CPO with a random distribution of the various CPO components.

The electrical anisotropy calculated using MK90 data for 900A87 in the plane normal to the Z plane is more than twice as high as in the Y plane, XY (Fig. 4). Analysis of the current distributions reveals that the high anisotropy of 7.6 is due to a high current density along X (j_x = 0.528) and low current density along Z (j_y = 0.069). This may be explained by the shape preferred orientation of large olivine porphyroclasts, like the one in the centre of the map (Fig. 1), which forms a continuous olivine pathway with [1 0 0] and [0 1 0] parallel to X and Z, respectively. For MK90 data, this interconnected pathway enhances the current density along X because [1 0 0] is the axis of the highest conductivity. At the same time it blocks current along Z (y direction in the FEM model) [0 1 0] is the direction of lowest conductivity of the olivine crystal. In contrast, the anisotropy calculated in the XY plane (R-900A87 with original microstructure as 900A87) is 3. The interconnected porphyroclast pathway gives an almost identical current density in both X and Y directions of the FEM model. Since in the MK90 data, [0 0 1] has intermediate diffusivity values, this leads to an enhancement of the conductivity along the y direction of the FEM model resulting in a higher current (j_y = 0.173) and lower anisotropy in the flow plane. This effect is not observed for the models calculated using MK98 data, because of the lower anisotropy and smaller difference in olivine H^+ diffusivity between [0 1 0] and [0 0 1] directions.

Comparison between electrical conductivity anisotropy factor predicted by FEM models and random resistor networks also suggests that both the CPO and microstructure produced by plastic deformation in the mantle contribute to electrical anisotropy. Anisotropy factors calculated of 100 random resistor networks range from <1 to 1.5, i.e., maximum anisotropies similar to those predicted by FEM, but display an average value of 3 (Simpson and Tommasi, 2005). The present models suggest that orientation distributions producing high anisotropies are more common in deformed mantle rocks than those producing low anisotropies. Thus, averaging a large number of random resistor networks tends to underestimate the actual electrical anisotropy factors that result from a given olivine CPO. The highest electrical anisotropy will be observed in samples in which dislocation glide produced elongated grains even if these grains completely recrystallize, since these “fossil porphyroclasts” will still form elongated lenses with similar orientations, as in the experimentally deformed dunite PO342 (Bystricky et al., 2000). On the other hand, fast grain boundary migration during deformation will tend to hinder the development of high anisotropies, leading to a more random distribution of the most conductive orientations, like in sample EPT3. Grain boundary migration is favored by high temperatures and low strain rates. The strongest anisotropy factors (>10) should thus occur in mantle rocks deformed under “lithospheric” conditions, i.e., at lower temperatures and/or high strain rates, in which grain boundary migration is limited.

5. Comparison of calculated electrical anisotropies with electromagnetic field studies data

Agreement between electrical and seismic anisotropy fast directions has been observed in many regions worldwide, like the Greenville belt in Canada (Ji et al., 1996; Sénéchal et al., 1996), the Appalachians (Wannamaker et al., 1998), the Caledonian and the Adirondacks (van der Meer et al., 2000), the Great Basin of North America (Haxby et al., 2000), the Canadian Shield, the Andean orogen (e.g., Magalhães and Sampaio, 1998), the easternmost part of Brazil (Bastos et al., 2000), the Hawaiian Islands (Lee et al., 1998), and the Great Bear-Lowe belt in Canada (Barr et al., 2003). This agreement is commonly interpreted as an indicator of the predominant orientation of the porphyroclasts (e.g., Sénéchal et al., 1996; van der Meer et al., 2000).
Duba, 2000), and central Germany (Leibecker et al., 2005). How- 
tever, the observed anisotropies at the base of the asthenosphere vary between 3 below central Australia, northern Bavaria, >35 for Fennoscandia and central Europe. The Fennoscandian data show abrupt changes of anisotropy in some areas without a clear orientation of strike angle, Lahti et al., 2005) due to conductivity heterogeneities, which may have enhanced the electrical anisotropy for those regions. Another conduction mechanism is required.

Besides partial melt, small amounts of highly conductive phases like graphite (Duba and Shankland, 1982; Deines, 2002) and sulphides (e.g., Ducea and Park, 2000) may enhance upper mantle conductivities if they form an interconnected network along grain boundaries. However, to produce a high electrical anisotropy, if the highly conductive phase is equally distributed along grain boundaries, a very strong shape preferred orientation is required in order to engender a significant enhancement of interconnectivity along the lineation (X direction in sample coordinate system) and concurrent reduction of interconnectivity perpendicular to it (Y direction). Yet, as observed in sample EPTA3, under high temperature, low-stress conditions that are expected to prevail under at asthenospheric depths, fast synkinematic grain boundary migration leads to development of weak shape preferred orientations.

Anisotropy could nevertheless be produced if the highly conductive phases were not evenly distributed along all grain boundaries. Analyses of the spatial arrangement of melt pockets in experimentally sheared peridotite + basalt assemblages (Holtzman et al., 2003) and of the crystallization of secondary clinopyroxenes in naturally deformed mantle rocks (Tommasi et al., 2004) suggests nevertheless that fluids tend to align in lens shaped pockets parallel or at low angle to the shear plane. In a subhorizontal shear zone, the highly conductive phases should therefore be concentrated in subhorizontal lenses and hence produce no anisotropy in long period MT data. In contrast, in a subvertical shear zone, e.g., a lithospheric strike slip fault, highly conductive aligned in the shear plane may add to the intrinsic anisotropy of olivine, resulting in strong electrical anisotropy.

A general outcome of the FEM results is that the weak anisotropy is not caused by low conductivities along the flow direction (X) but rather by high electrical conductivities perpendicular to it. From the calculated current densities (Table 1), we may estimate the factor $r_T$, which is the electrical conductivity along a certain direction with respect to the electrical conductivity along the olivine $[100]$-axis. The samples with high anisotropy show for the high-conductivity direction $[\alpha]$, a ratio $\lambda_T$ of about 0.45–0.55. The bulk electrical conductivity along that direction is therefore expected to be about half the intrinsic electrical conductivity along olivine $[100]$-axis, $\sigma_{[100]}^{\text{ol}}$. Assuming saturated conditions (Lizarralde et al., 1995), $H^+$ diffusion can result in...
The finite-element modelling to quantify the anisotropy due to anisotropic intracrystalline diffusion in upper mantle rocks as a function of CPO and microstructure, which are described in section maps obtained by indexation of electron backscattered diffraction (EBSD) patterns. The studies have varied microstructures and olivine textures of upper mantle deformation, with intensifying the entire range of CPO intensities usually in the upper mantle. Highest conductivity is parallel to X direction, which is the flow direction inferred from both the crystals elongation and alignment of olivine [1 0 0] crystallographic axes. Motion of the experimentally sheared dunite that is strongly anisotropic (>10), macroscopic electrical anisotropy factors range between 3 and 8. The anisotropy of electrical anisotropy depends on a first order of the olivine crystal preferred orientation, the relation saturates for strong crystal presentations. Comparing the FEM results with others from random resistivity network modelling or averaging (e.g., VRH averages) shows that the anisotropy and, hence, the interconnectivity of CPO components in mantle samples significance their electrical anisotropy. These models are lower bound estimates of upper mantle anisotropies. The strongest anisotropy factors could occur in mantle rocks in which deformed or recrystallization has produced not only crystal shape preferred orientations, even if the masked by recrystallization.

Vertical conductivity anisotropy factors <10, similarly inferred from MT data in Australia or Germany, may thus result from anisotropic H+ diffusion in olivine in peridotites displaying olivine presentations typical of upper mantle deformation. Anisotropy factors observed in other regions, like Germany, imply that intracrystalline H+ diffusion is the dominating conduction mechanism. However, the high anisotropies at asthenospheric depths cannot be explained by the presence of aligned simulations in combination with laboratory measurements of electrical conductivities on naturally deformed peridotites at high temperature and high water fugacity conditions are thus required to understand this data.

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