



**HAL**  
open science

# Texturing in the Earth's inner core due to preferential growth in its equatorial belt

Renaud Deguen, Philippe Cardin, Sebastien Merkel, R.A. Lebensohn

► **To cite this version:**

Renaud Deguen, Philippe Cardin, Sebastien Merkel, R.A. Lebensohn. Texturing in the Earth's inner core due to preferential growth in its equatorial belt. 2011. hal-00560109v1

**HAL Id: hal-00560109**

**<https://hal.science/hal-00560109v1>**

Preprint submitted on 28 Jan 2011 (v1), last revised 11 Jun 2011 (v2)

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

# Texturing in the Earth's inner core due to preferential growth in its equatorial belt

Renaud Deguen<sup>a,b</sup>, Philippe Cardin<sup>a,\*</sup>, Sébastien Merkel<sup>c</sup>, Ricardo A. Lebensohn<sup>d</sup>

<sup>a</sup>*Institut des Sciences de la Terre (ISTerre), CNRS, Université Joseph-Fourier Grenoble 1, Grenoble, France.*

<sup>b</sup>*Earth and Planetary Sciences, Johns Hopkins University, Baltimore, Maryland, USA.*

<sup>c</sup>*Unité Matériaux et Transformations (UMET), CNRS, Université Lille 1, Villeneuve d'Ascq, France.*

<sup>d</sup>*Materials Science and Technology Division, Los Alamos National Laboratory, Los Alamos, New Mexico, USA.*

---

## Abstract

We propose an extension of the model by Yoshida et al. (1996), where deformation in the inner core is forced by preferential growth in the equatorial belt, by taking into account the presence of a stable compositional stratification. Stratification inhibits vertical motion, imposes a flow parallel to isodensity surfaces, and concentrates most deformation in a shallow shear layer of thickness  $\sim B^{-1/5}$ , where  $B$  is the dimensionless buoyancy number. The localization of the flow results in large strain rates and enable the development of a strong texture of iron crystals in the upper inner core. We couple our dynamical model with a numerical model of texture development and compute the time evolution of the lattice preferred orientation of different samples in the inner core. With sufficient stratification, texturing is significant in the uppermost inner core. In contrast, the deeper inner core is

---

\*Corresponding author

*Email address:* philippe.cardin@ujf-grenoble.fr (Philippe Cardin )

unaffected by any flow and may preserve a fossil texture. We then investigate the effect of an initial texture resulting from solidification texturing at the ICB. In the present inner core, the deformation rate in the shallow shear layer is large and can significantly alter the solidification texturing, but the solidification texture acquired early in the inner core history can be preserved in the deeper part. Using elastic constants from *ab initio* calculations, we predict different maps of anisotropy in the modern inner core. A model with both solidification texturing and subsequent deformation in a stratified inner core produce a global anisotropy in agreement with seismological observations, both in magnitude and geographical distribution, with a weak anisotropy in the uppermost layer and stronger anisotropy in the deeper parts.

*Keywords:* inner core, anisotropy, HCP iron, texturation, crystallization

---

## 1. Introduction

2       The inner core of the Earth exhibits a noticeable anisotropy in P-wave ve-  
3       locity and attenuation (Poupinet et al., 1983; Morelli et al., 1986; Woodhouse  
4       et al., 1986; Souriau, 2007), which may reflect structural and/or dynamical  
5       complexity. The main observation is a  $\sim 3\%$  anisotropy with cylindrical  
6       symmetry, with the fast axis aligned with the axis of rotation of the Earth or  
7       possibly slightly tilted (Su et al., 1996). The degree of inner core anisotropy  
8       is increasing with depth (Souriau, 2003), with no strong anisotropy in the up-  
9       per 150-200 km (Song and Helmberger, 1995), and perhaps still different [but  
10       poorly constrained (Calvet et al., 2006)] seismic properties in an innermost  
11       inner core (Ishii and Dziewoński, 2002; Niu and Chen, 2008). There is also a  
12       growing consensus on the presence of an hemispherical asymmetry in the seis-

13 mic properties of the upper part of the inner core (Tanaka and Hamaguchi,  
14 1997): the Western hemisphere (west of the Greenwich meridian) appears  
15 more anisotropic than the Eastern hemisphere. This observation can be due  
16 to lateral variations of the local seismic properties (Niu and Wen, 2001),  
17 because of a different degree of crystal alignment, for instance, or as orig-  
18 inating from a longitudinal variation of the thickness of the quasi-isotropic  
19 upper layer, which would have to vary from  $\sim 100$  km in the western hemi-  
20 sphere to  $\sim 400$  km in the eastern hemisphere (Garcia and Souriau, 2000).  
21 An anisotropy in inner core seismic attenuation is also observed (Souriau and  
22 Romanowicz, 1996, 1997; Yu and Wen, 2006a) and shows hemispherical vari-  
23 ations (Garcia, 2002; Yu and Wen, 2006b) in reminiscence with the pattern  
24 of the anisotropy.

25 The seismic anisotropy of the inner core is most often interpreted as result-  
26 ing from the lattice preferred orientation (LPO) of elastically anisotropic iron  
27 crystals. Processes for creating iron LPO in the inner core include solidifica-  
28 tion at the inner core boundary (ICB) (Karato, 1993; Bergman, 1997; Brito  
29 et al., 2002), plastic deformation (Jeanloz and Wenk, 1988; Karato, 1999;  
30 Wenk et al., 2000; Buffett and Wenk, 2001), or stress-induced recrystalliza-  
31 tion (Yoshida et al., 1996). It is not, however, unlikely that a combination  
32 of several mechanisms are - or have been - active. To what extent solidifica-  
33 tion texturing can be reworked by subsequent deformation might well be a  
34 critical point to understand the observed radial variations in anisotropy and  
35 attenuation.

36 A number of deformation models have been proposed, including thermal  
37 convection (Jeanloz and Wenk, 1988; Wenk et al., 1988, 2000; Weber and

38 Machetel, 1992; Buffett, 2009), continuous deformation forced by aspherical  
39 growth (Yoshida et al., 1996), and flow induced by the outer core magnetic  
40 field (Karato, 1999; Buffett and Wenk, 2001). Most of these mechanisms have  
41 significant drawbacks. In particular, their viability depends on the thermal  
42 and chemical state of the inner core. Apart from the mechanism of Buffett  
43 and Wenk (2001), all involve a large radial flow (Jeanloz and Wenk, 1988;  
44 Yoshida et al., 1996; Karato, 1999; Wenk et al., 2000) and will be inhibited by  
45 a stable stratification in the inner core (Buffett and Bloxham, 2000; Deguen  
46 and Cardin, 2009).

47 In this study, we concentrate our efforts on connecting geodynamics and  
48 simulations of LPO development in mineral aggregates using the viscoplas-  
49 tic self-consistent (VPSC) method of Lebensohn and Tome (1993). We will  
50 restrict ourself to a single phase of iron, hexagonal-closed-packed (*hcp*-Fe),  
51 using a single set of plastic deformation mechanisms, described therein, and  
52 a single set of elastic moduli (Vočadlo et al., 2009). We extend the dy-  
53 namical model proposed by Yoshida et al. (1996) to include stable density  
54 stratification in the inner core (Deguen and Cardin, 2009), and simulate the  
55 development of LPO associated with the deformation.

56 In part 2, we briefly discuss the thermal and chemical states of the inner  
57 core, and introduce our dynamical model. This inner core model couples  
58 preferential crystallization in the equatorial region (Yoshida et al., 1996) and  
59 density stratification. The density stratification arises from inner core growth  
60 and evolution of the concentration of light elements in the fluid outer core  
61 over time (Deguen and Cardin, 2009). In part 3, from our numerical models  
62 of core formation, we track the deformation history of markers inside the

63 inner core and calculate the evolution of LPO in Fe using the VPSC method.  
 64 In part 5, using published value of elastic moduli of *hcp*-Fe, we produce maps  
 65 of P wave velocity anisotropy in the modern inner core.

## 66 2. Dynamical model

67 In this study, we focus on the effect of thermal or chemical stratification  
 68 on the dynamics of the inner core. The thermal state of the inner core  
 69 is discussed in section 2.1, the chemical stratification is derived in 2.2, the  
 70 dynamic equations are presented in 2.3, and the resulting computations are  
 71 discussed in 2.4.

### 72 2.1. Thermal state of the inner core

73 The thermal state of the inner core results from a competition between  
 74 cooling at the ICB and extraction of the inner core internal heat. The inner  
 75 core would be stably stratified (subadiabatic) if its growth is slow enough  
 76 to allow thermal conduction to evacuate most of the internal heat. Deguen  
 77 and Cardin (2009) introduced a simplified criterion for thermal stability in  
 78 the inner core, which is based on the assumption that the cooling rate of the  
 79 core is constant (A similar approach has been given by Buffett (2009)). In  
 80 this model, the geotherm in the inner core is subadiabatic (stably stratified)  
 81 if

$$\tau_{\text{ic}} > \tau_{\kappa} \left( \frac{dT_s}{dT_{\text{ad}}} - 1 \right), \quad (1)$$

82 where  $\tau_{\text{ic}}$  is the age of the inner core,  $\tau_{\kappa} = r_{\text{icb}}^2 / (6\kappa)$  is the present thermal  
 83 diffusion time,  $r_{\text{icb}}$  is the radius of the inner core,  $\kappa$  the thermal diffusivity,  
 84 and  $dT_s/dT_{\text{ad}}$  the ratio of the Clapeyron slope to the adiabat. This criterion

85 provides a first-order estimate of the age of the inner core required for thermal  
 86 stratification and explicitly describes the sensitivity of the thermal stratifi-  
 87 cation limit to the relevant parameters. Moreover, it is shown in Deguen and  
 88 Cardin (2011) that this criterion remains valid when a more realistic model  
 89 of core thermal history is used (Labrosse et al., 2001).

90 Uncertainties on  $\tau_\kappa$  are mainly due to uncertainties on the thermal con-  
 91 ductivity of iron at inner core conditions. The relatively large value ( $k = 63$   
 92  $\text{W.m}^{-1}.\text{K}^{-1}$  in the outer core at the ICB,  $k = 79 \text{ W.m}^{-1}.\text{K}^{-1}$  in the in-  
 93 ner core) favored by Stacey and Anderson (2001) has been recently revised  
 94 downward by Stacey and Loper (2007). Stacey and Davis (2008) give  $k = 36$   
 95  $\text{W.m}^{-1}.\text{K}^{-1}$  for the inner core.

96 The ratio  $dT_s/dT_{\text{ad}}$  can be estimated by using Lindeman’s law,

$$\frac{dT_s}{dP} = \frac{2(\gamma - 1/3)T}{K_T}, \quad (2)$$

97 and writing the adiabatic temperature gradient as

$$\frac{dT_{\text{ad}}}{dP} = \frac{\gamma T}{K_S}, \quad (3)$$

98 where  $\gamma$  is the Gruneisen parameter,  $P$  the pressure,  $T$  the temperature,  $K_T$   
 99 and  $K_S$  the isothermal and adiabatic bulk modulus, respectively. Using the  
 100 thermodynamic identity  $K_S = K_T(1 + \gamma\alpha T)$ , we find

$$\frac{dT_s}{dT_{\text{ad}}} = \frac{2(\gamma - 1/3)(1 + \gamma\alpha T)}{\gamma}. \quad (4)$$

101 With the values and uncertainties of thermo-physical properties given in  
 102 table 1, we find  $\tau_\kappa = 1.4 \pm 0.7 \text{ Gy}$ , and  $dT_s/dT_{\text{ad}} = 1.65 \pm 0.11$ . This gives  
 103 a critical age for thermal stratification of  $0.9 \pm 0.6 \text{ Gy}$ . The important point  
 104 here is that the range of critical ages we obtain fully overlaps with what

105 models of core thermal evolution predict for the age of the inner core, e.g.  
 106  $1 \pm 0.5$  Gy in Labrosse et al. (2001) and  $1.15 \pm 0.75$  Gy in Nimmo (2007).  
 107 This means that the uncertainties are such that whether the inner core is  
 108 subadiabatic or superadiabatic is not known. In this paper, we assume that  
 109 it is subadiabatic and that the inner core has a stable thermal stratification.

## 110 2.2. Chemical stratification in the inner core

111 The composition of the core remains uncertain. However, a comparison  
 112 between outer core density (Dziewonski and Anderson, 1981) and that of  
 113 liquid iron at core conditions (Laio et al., 2000) indicates that the outer core  
 114 material is 6-10% lighter than pure iron. The presence of other elements is  
 115 also inferred from seismological estimates of the density jump at the inner  
 116 core boundary,  $\Delta\rho_{icb} = 800 \pm 200$  kg m<sup>-3</sup> (Masters and Gubbins, 2003;  
 117 Souriau, 2007). The density jump due to solidification alone ( $\approx 200$  kg m<sup>-3</sup>,  
 118 Laio et al. (2000)) is not sufficient to explain the observed value, implying  
 119 that, as most alloys, the core mixture partitions upon freezing.

120 If the outer core is assumed to be well mixed, the evolution of the con-  
 121 centration of the light element  $c^\ell(t)$  in the liquid outer core is given by the  
 122 Rayleigh distillation law (e.g. Albarède, 1996),

$$c^\ell(t) = c_0^\ell \left[ 1 - \frac{M_{ic}(t)}{M_0} \right]^{D-1}, \quad (5)$$

123 where  $c_0^\ell$  is the concentration in the core prior to inner core nucleation,  $M_{ic}$   
 124 the mass of the inner core, and  $M_0$  the mass of the core. For simplicity, we  
 125 approximate  $M_{ic}(t)/M_0 \approx (r_{icb}(t)/r_c)^3$ . The light element concentration in



Table 1: Parameters used in this study.

Parameter	Symbol	Value
Core radius <sup>a</sup>	$r_c$	3480 km
Inner core radius <sup>a</sup>	$r_{icb}$	1221 km
Solidification temperature <sup>b</sup>	$T_{icb}$	$5600 \pm 500$ K
Gruneisen parameter <sup>c</sup>	$\gamma$	$1.4 \pm 0.1$
Thermal expansivity <sup>c</sup>	$\alpha_T$	$1.1 \pm 0.1 \times 10^{-5}$ K <sup>-1</sup>
Heat capacity <sup>d</sup>	$c_p$	$800 \pm 80$ J.kg <sup>-1</sup> .K <sup>-1</sup>
Density in the inner core <sup>a</sup>	$\rho$	13 000 kg.m <sup>-3</sup>
Thermal conductivity <sup>e</sup>	$k$	36 – 79 W.m <sup>-1</sup> .K <sup>-1</sup>
Initial concentration <sup>b</sup>	$c_0$	5.6 wt%
Partition coefficient <sup>b</sup>	$D$	0.8
Chemical expansivity <sup>f</sup>	$\alpha_c$	-1

<sup>a</sup> From PREM (Dziewonski and Anderson, 1981).

<sup>b</sup> From Alfè et al. (2002).

<sup>c</sup> From Vočadlo (2007).

<sup>d</sup> From Stacey and Davis (2008).

<sup>e</sup> From Stacey and Anderson (2001) and Stacey and Davis (2008).

<sup>f</sup> See text (section 2.2).

126 the inner core,  $c(r)$  becomes

$$c(r) = Dc_0^\ell \left( 1 - \left( \frac{r}{r_c} \right)^3 \right)^{D-1}, \quad (6)$$

and the corresponding profile of differential compositional density is then

$$\Delta\rho_c = \alpha_c\rho[c(r) - c(0)], \quad (7)$$

$$= \alpha_c\rho Dc_0^\ell \left[ \left( 1 - \left( \frac{r}{r_c} \right)^3 \right)^{D-1} - 1 \right], \quad (8)$$

$$\approx \alpha_c\rho D(1 - D)c_0^\ell \left( \frac{r}{r_c} \right)^3, \quad (9)$$

127 where  $D$  is the partition coefficient (the ratio of the concentration in the  
 128 solid to that in the liquid),  $\rho$  the mean density in the inner core, and  $\alpha_c =$   
 129  $(1/\rho)(\partial\rho/\partial c)$  is the compositional expansion coefficient. Equation (9) shows  
 130 that the chemical stratification is approximately proportional to  $D(1 - D)$ :  
 131 stratification would be maximum if  $D = 0.5$ , and is small if the distribution  
 132 coefficient is either small or close to one.

133 The stratification depends on the nature and abundance of the light el-  
 134 ements present in the core. The composition of the core has been a long  
 135 standing issue and is still controversial, but recent models favour O, Si and  
 136 S as the most plausible alloying elements. *Ab initio* calculations of the par-  
 137 titioning behavior of O, Si and S by Alfè et al. (2002) suggest that Si and S  
 138 both partition weakly (with similar partition coefficients,  $D^{\text{Si,S}} = 0.8$ ) while,  
 139 in contrast, O partitions strongly,  $D^{\text{O}} = 0.02$  (values of  $D$  are converted  
 140 from molar ratios to mass ratios). These values favor the presence of sulfur  
 141 or silicium in the inner core. Alfè et al. (2002) estimate that the outer core  
 142 contains 5.6 wt. % of Si and/or S and 2.5 wt. % of O, in good agreement

143 with the geochemical model of Allègre et al. (1995), and that the inner core  
144 contains 4.4 wt. % of Si/S and negligible amount of O. Badro et al. (2007) used  
145 additional constraints from experimentally measured compressional wave ve-  
146 locity of Fe-Si, Fe-S and Fe-O alloys. In their model, S is unlikely to be a  
147 major component of the inner core, the outer core contains 2.8 wt. % Si and  
148 5.3 wt. % O, and the inner core is constituted of 2.3 wt. % Si and 0.1 wt.  
149 % O.

150 The chemical expansion coefficient can be estimated from the partial  
151 atomic volumes of Fe, Si, S and O calculated by Alfè et al. (2002). We  
152 find  $\alpha_c^{\text{Si}} = -0.91$ ,  $\alpha_c^{\text{S}} = -0.67$  and  $\alpha_c^{\text{O}} = -1.3$  in the limit  $c \rightarrow 0$  (Deguen  
153 and Cardin, 2011).

154 With the distribution coefficients calculated by Alfè et al. (2002), the  
155 chemical models discussed above, and the estimates of  $\alpha_c$ , we find that the  
156 density stratification associated with Si/S is much larger than that associated  
157 with O. Therefore, in our model, the inner core is considered as a Fe-(Si,S)  
158 binary mixture, with a distribution coefficient equal to 0.8. With the value  
159 listed in table 1, the difference in density between the center of the Earth and  
160 in the ICB is  $\Delta\rho_c \sim -5 \text{ kg m}^{-3}$ .

161 This model includes a number of approximations:

- 162 (i) The pressure and temperature dependence of the partition coefficient  
163 and the compositional expansion are neglected.
- 164 (ii) We also neglect possible variations of the effective partition coefficient  
165 resulting from changes in the efficiency of melt extraction by com-  
166 paction (Sumita et al., 1996) and interdendritic convection (Loper,  
167 1983; Worster, 1997) in a mushy layer at the ICB. As noted by Al-

168 boussière et al. (2010), this may significantly decrease the magnitude  
169 of the chemical stratification.

170 (iii) We suppose a perfect mixing in the outer core, thus ignoring the possi-  
171 ble accumulation of a light element rich liquid below the CMB (Fearn  
172 and Loper, 1981) or the presence of a dense layer in the lowermost liquid  
173 outer core (Souriau and Poupinet, 1991; Alboussière et al., 2010).

174 (iv) Chemical interactions between melt in the outer core and silicate mate-  
175 rial at the CMB, which might buffer the composition of the outer core  
176 (Buffett et al., 2000), are assumed negligible.

177 The compositional density variation of our model reflects a relative en-  
178 richment of light elements at the top of the inner core as it grows. It is  
179 the internal imprint of the enrichment of the outer core in those same light  
180 elements induced by the partitioning during the growth of the inner core.  
181 The variation of concentration of light element in the inner core generates  
182 an internal solute flux by diffusion. The solute diffusivity is not known but  
183 is unlikely to be much larger than the self-diffusion of iron, estimated below  
184  $10^{-12}\text{m}^2.\text{s}^{-1}$  in the inner core (Van Orman, 2004). Chemical diffusion is neg-  
185 ligible as the diffusion time is larger than  $10^6$  Gyr for the inner core, and the  
186 compositional profile (9) is valid on the time scale of the inner core formation  
187 ( $\approx 1$  Gyr).

### 188 *2.3. Dynamic equations for a stratified inner core*

189 We build a model for a thermally and chemically stratified inner core  
190 according to the two previous subsections. In these conditions, the inner core  
191 stays at rest except if external forcings, such as pressure forcings (Yoshida

192 et al., 1996) or magnetic forcings (Karato, 1999; Buffett and Bloxham, 2000;  
 193 Buffett and Wenk, 2001), are imposed. In this paper, we follow the ideas of  
 194 Yoshida et al. (1996) and focus on how differential growth of the inner core  
 195 generates motion in a stratified inner core. For simplicity, we consider only  
 196 the chemical contribution to the density stratification.

### 197 *2.3.1. Equatorial growth of the inner core*

198 The growth of the inner core is primarily due to its cooling by action of  
 199 convective motions in the liquid outer core, also responsible of the geodynamo  
 200 generation. The predominance of the axial geomagnetic dipole throughout  
 201 the Earth’s history demonstrates the permanent key role of the Coriolis force  
 202 in the fluid motion in the outer core. Thermal convection in the outer core  
 203 is made of columnar vortices aligned with the axis of rotation. This quasi  
 204 geostrophic flow, first predicted by Busse (1970) and later confirmed by nu-  
 205 merical or experimental means (Zhang, 1992; Dormy et al., 2004; Cardin and  
 206 Olson, 1994) has a strong influence on convective heat transfer (Aubert et al.,  
 207 2008; Aurnou et al., 2008).

208 Accordingly, it is expected that more heat is extracted from the equato-  
 209 rial band than from the polar regions of the inner core by the geodynamo.  
 210 Following Yoshida et al. (1996), we model this by introducing a spherical  
 211 harmonic degree 2 dependence in the rate of crystallization of the inner core  
 212  $u$ ,

$$u(\theta, t) = u_{icb}(t) \left( 1 - S_2 \frac{3 \cos^2 \theta - 1}{2} \right), \quad (10)$$

213 where  $\theta$  is the colatitude and  $S_2$  is a dimensionless parameter that measures  
 214 the differential growth. Calculations by Aubert et al. (2008) show a variation

215 of the ICB heat flux of +30% in the equatorial band and -30% in polar re-  
 216 gions. In other words, the crystallization rate at the equator is approximately  
 217 twice larger than that at the poles, which is equivalent to  $S_2 = 2/5$ .

218 Preferential growth of the inner core in the equatorial belt produces an  
 219 out-of-equilibrium topography, which, if not sustained by heterogeneous so-  
 220 lidification, would relax toward hydrostatic equilibrium. The problem is very  
 221 similar to that of post-glacial rebound, and the timescale of viscous relax-  
 222 ation  $\tau_\eta$  can be calculated by extending to spherical geometry the classical  
 223 calculation of topography relaxation in a semi-infinite domain (*e.g.* Turcotte  
 224 and Schubert, 2002). We find

$$\tau_\eta = \frac{19}{5} \frac{\eta}{\Delta\rho_{\text{icb}} g r_{\text{icb}}} \simeq \left( \frac{\eta}{10^{18} \text{ Pa}\cdot\text{s}} \right) \left( \frac{1221 \text{ km}}{r_{\text{icb}}} \right)^2 \times 40 \text{ year}, \quad (11)$$

225 where  $\eta$  is the dynamic viscosity of iron in the inner core, and  $g$  the acceler-  
 226 ation of gravity. Since the relaxation timescale is very small in comparison  
 227 to the timescale of boundary conditions evolution and inner core growth,  
 228 the inner core topography must be in a quasi-steady state, with the growth  
 229 rate anomaly balanced by the continuous relaxation of the ICB topography  
 230 (Yoshida et al., 1996). The radial velocity at the ICB is therefore prescribed  
 231 to be equal to the opposite of the anomalous solidification rate.

232 This gives boundary conditions for the radial velocity; boundary condi-  
 233 tions for the horizontal velocity are given by the assumption that the ICB is  
 234 a shear stress free boundary.

235 The newly crystallized material has a solute concentration  $c(r_{\text{icb}}(t))$  given  
 236 by (6).

237 *2.3.2. Equations of motion in a growing inner core*

238 We consider an incompressible fluid of constant viscosity  $\eta$  in a spherical  
 239 domain ( $r < r_{icb}$ ). The Boussinesq approximation is valid and the momentum  
 240 equation is written as :

$$\mathbf{0} = -\nabla p + \alpha_c \bar{\rho} c \mathbf{g} + \eta \nabla^2 \mathbf{v}, \quad (12)$$

241 where  $p$  is the dynamic pressure,  $\alpha_c c = (\rho - \bar{\rho})/\bar{\rho}$  is the density perturba-  
 242 tion from the static density field. Inertial terms have been neglected as the  
 243 Reynolds number of the flow is vanishingly small. Taking the curl of the  
 244 Stokes equation (12) gives

$$\mathbf{0} = -\frac{\alpha_c \bar{\rho} g'}{\eta} \frac{\partial c}{\partial \theta} \mathbf{e}_\phi + \nabla^2 \nabla \times \mathbf{v}, \quad (13)$$

245 where  $g$  has been assumed to be linear in radius, and  $g' = dg/dr$ . Taking  
 246 advantage of the incompressibility of the velocity field, we can introduce  
 247 the poloidal ( $P$ ) and the toroidal ( $T$ ) scalars to describe the velocity field  
 248  $\mathbf{v} = \nabla \times (T\mathbf{r}) + \nabla \times \nabla \times (P\mathbf{r})$  which are projected onto the basis of the  
 249 axisymmetric spherical harmonics  $(P, T) = \sum (P_l(r), T_l(r)) Y_l^0(\theta)$ . Taking  
 250  $\mathbf{e}_r \cdot \nabla \times (13)$  and  $\mathbf{e}_r \cdot \nabla \times \nabla \times (13)$ , we find

$$D_l^2 P_l = \frac{\alpha_c \bar{\rho} g' r}{\eta} c_l \quad \text{and} \quad D_l^2 T_l = 0, \quad (14)$$

251 where  $c_l$  is the degree  $l$  spherical harmonic component of the solute concen-  
 252 tration field and  $D_l$  is the Laplacian operator defined by  $\frac{\partial^2}{\partial r^2} + \frac{2}{r} \frac{\partial}{\partial r} - \frac{l(l+1)}{r^2}$ .

253 At the ICB, we deduce from (14) and (10) that the toroidal flow is zero and  
 254 that, consequently, the velocity field is fully described by its poloidal part. To  
 255 write (14) in a dimensionless form, we scale the length by  $r_{icb}(t)$ , the velocity

256 by  $u_{icb}(t)$  and the solute concentration by the difference of concentration  
 257  $\Delta c(t) = c(r_{icb}) - c(0)$  between the concentration at the top of the growing  
 258 inner core and the initial concentration at the center. The poloidal part of  
 259 (14) becomes

$$D_t^2 \tilde{P}_l = B \tilde{c}_l, \quad (15)$$

260 where  $\tilde{\phantom{x}}$  stands for the dimensionless variables and  $B$ , the dimensionless buoy-  
 261 ancy number, is defined as

$$B = \frac{\alpha_c \bar{\rho} g' \Delta c(t) r_{icb}^3(t)}{\eta u_{icb}(t)}. \quad (16)$$

262  $B$  increases significantly during the growth of the inner core as a result of  
 263 the combined effect of increasing stratification ( $\Delta c \propto r_{icb}^3$  and  $g \propto r_{icb}$ ), de-  
 264 creasing solidification rate, and increasing length scale  $r_{icb}$ , which makes the  
 265 viscous transfert of momentum less efficient. If we assume that the inner  
 266 core radius grows as the square root of time (simplified context of 2.1),  $B$   
 267 evolves as  $r_{icb}^7$ . Since  $\alpha_c$  is negative,  $B$  is negative, which corresponds to a  
 268 stratified stable system. The actual value of  $B$  in the inner core suffers from  
 269 very large uncertainties on the solid-state viscosity of iron in those condi-  
 270 tions (Yoshida et al., 1996; Buffett, 1997; Van Orman, 2004), with published  
 271 estimates ranging from  $10^{11}$  Pa.s to  $10^{21}$  Pa.s. With  $\Delta \rho_c = -5 \text{ kg m}^{-3}$  (see  
 272 2.2) and a typical  $u_{icb} = 3 \times 10^{-11} \text{ m.s}^{-1}$  (1 mm/yr), the present value of  $B$   
 273 (noted  $B^* = B(\tilde{t} = 1)$ ) would be

$$B^* = -10^6 \times \left( \frac{10^{18} \text{ Pa.s}}{\eta} \right). \quad (17)$$

274 We explore a large range of  $B$  value, going from  $B = 0$  (no stratification,  
 275 Yoshida et al. (1996)) to  $B^* = -10^9$  (very stratified and moderately viscous  
 276 inner core).



277 *2.3.3. Light element transport equation*

278 The solute transport equation is written as

$$\frac{\partial c}{\partial t} + \mathbf{v} \cdot \nabla c = 0. \quad (18)$$

279 In order to simplify the numerical implementation, we solve the problem in  
 280 a spherical domain with no moving boundary. This is done naturally with  
 281 our choice of length scale which is the time dependent radius of the inner  
 282 core  $r_{icb}(t)$ . The radial domain is defined by  $\tilde{r} = r/r_{icb}(t) \in [0, 1]$  with a fixed  
 283 boundary at  $\tilde{r} = 1$ . The time derivative in the new coordinates system  $\tilde{\mathcal{R}}$  is  
 284 written (Crank, 1984) as

$$\frac{\partial}{\partial t} \Big|_{\tilde{\mathcal{R}}} = \frac{\partial}{\partial t} \Big|_{\mathcal{R}} - \frac{\partial \tilde{r}}{\partial t} \Big|_{\mathcal{R}} \frac{\partial}{\partial \tilde{r}} \Big|_t = \frac{\partial}{\partial t} \Big|_{\mathcal{R}} + \tilde{r} \frac{u_{icb}}{r_{icb}} \frac{\partial}{\partial \tilde{r}} \Big|_t. \quad (19)$$

285 This allows the equation of solute transport to be written in the new refer-  
 286 ential as

$$\frac{\partial \tilde{c}}{\partial \tilde{t}} = \frac{u_{icb} \tau_{ic}}{r_{icb}} (\tilde{r} \tilde{\mathbf{e}}_{\mathbf{r}} - \tilde{\mathbf{v}}) \cdot \tilde{\nabla} \tilde{c} - \frac{\dot{\Delta} c \tau_{ic}}{\Delta c} \tilde{c}, \quad (20)$$

287 where time has been scaled by the age of the inner core,  $\tau_{ic}$ , and other  
 288 variables as in section 2.3.2.

289 The first term in the RHS of equation (20) comes from the moving bound-  
 290 ary transformation and corresponds to an apparent inward advection in the  
 291 new referential. The last term in the RHS comes from the time dependence  
 292 of the prescribed concentration scale  $\Delta c(t)$  ( $\dot{\Delta} c$  stands for the time deriva-  
 293 tive of  $\Delta c$ ). If we assume that the inner core radius grows as the square root  
 294 of time, and that the compositional evolution takes the simplified form (9),  
 295 then  $(u_{icb} \tau_{ic})/r_{icb} = 1/2$  and  $(\dot{\Delta} c \tau_{ic})/\Delta c$  simplifies to  $3/2$ .

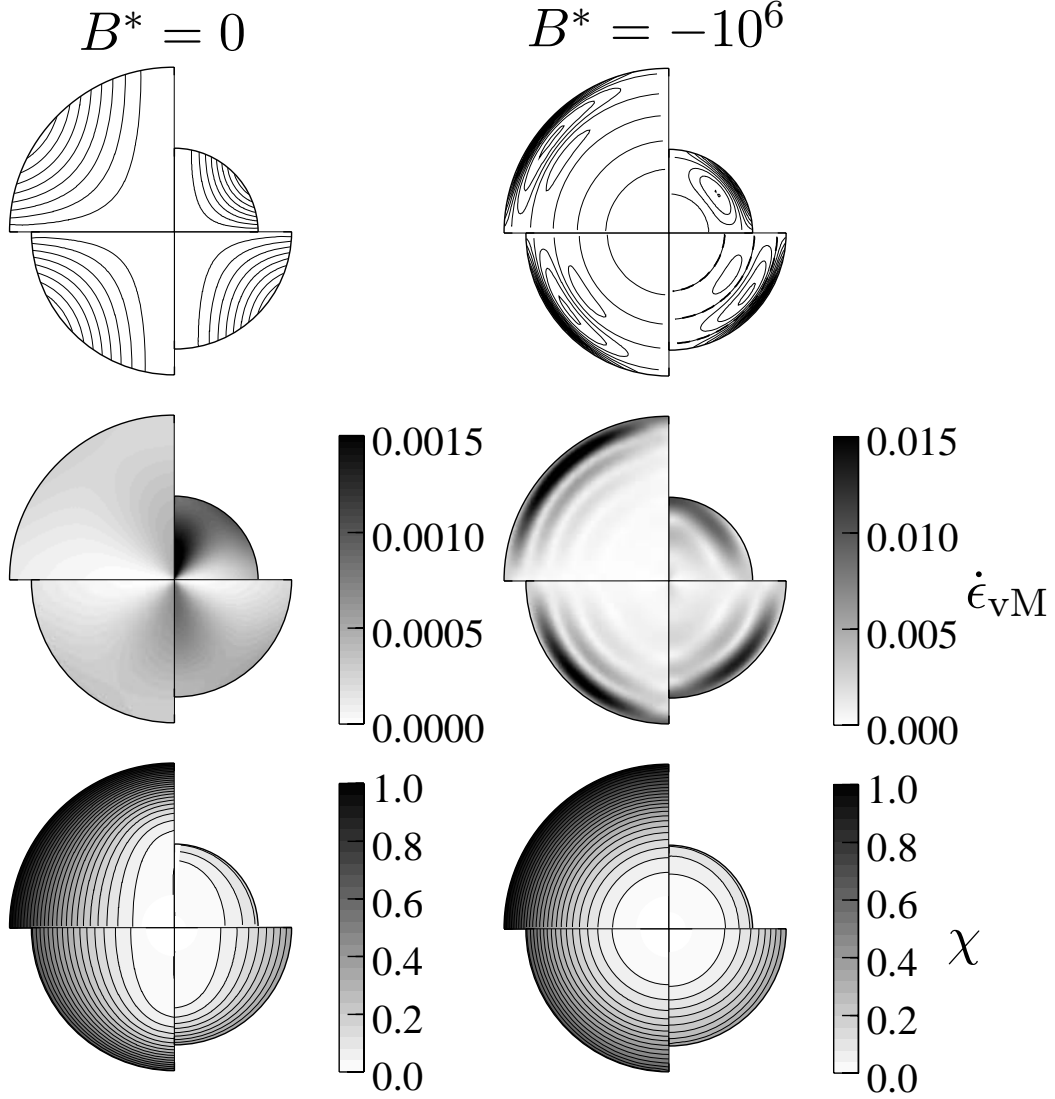


Figure 1: Time evolution of the flow induced by preferential growth of the inner core at the equator for two values of the dimensionless buoyancy number,  $B^* = 0$  (no stratification) and  $B^* = -10^6$  (stratified flow). All subfigures are divided into four quadrants, which correspond to non-dimensional times  $\tilde{t} = 0.25, 0.5, 0.75$  and  $1$ , starting clockwise from the upper right quadrant; the radius of each quadrant reflects the value of  $r_{icb}(t)$ . a, Contours of the stream function  $\psi$ . b, von Mises equivalent strain rate  $\dot{\epsilon}_{vM}$ , in  $\text{Myr}^{-1}$ . c, Non-dimensional light-element concentration  $\chi = c/\Delta c(\tilde{t} = 1)$ . The computations were made with  $S_2 = 2/5$ , which corresponds to a solidification rate two times higher at the equator than at the poles.

296 *2.3.4. Numerical implementation and simulations*

297 We implement the two equations (15,20) into a meridional plane, with  
 298 a spectral description for the horizontal dependence and a finite difference  
 299 scheme for the radial description. The radial mesh can be contracted in the  
 300 outermost part of the domain for large  $B$ . Boundary conditions at  $\tilde{r} = 1$   
 301 are :  $\forall l \neq 2, \tilde{P}_l = 0, \tilde{P}_2 = S_2$ , and  $\forall l, \frac{\partial^2 \tilde{P}_l}{\partial \tilde{r}^2} = [2 - l(l + 1)]\tilde{P}_l$  (shear stress  
 302 free condition),  $\tilde{c}_0 = 1$  and  $\forall l \neq 0, \tilde{c}_l = 0$ . The non linear term  $\tilde{\mathbf{v}} \cdot \tilde{\nabla} \tilde{c}$   
 303 is evaluated in the physical space at each timestep. A semi-implicit Crank-  
 304 Nickolson scheme is implemented for the time evolution of the linear terms  
 305 and an Adams-Bashforth procedure is used for the non-linear term. We  
 306 typically use time steps of order  $10^{-4}$ . This problem presents no major  
 307 numerical difficulty.

308 We start the simulations with a small inner core ( $\tilde{r} = 0.1$ ) and let evolve  
 309 the system to the final time  $\tilde{t} = 1$ . We track some fluid particles to com-  
 310 pute their trajectories and their stress tensor  $\dot{\epsilon}$  by bilinear interpolation in  
 311 the physical space. These quantities are saved and used as inputs for the  
 312 mineralogical model.

313 To visualize the dynamic effect in the inner core, we plot the isocontours of  
 314 the stream function  $\psi$  computed from the poloidal component of the axisym-  
 315 metric velocity to exhibit the flow pattern,  $\psi = r \sin \theta \frac{\partial P}{\partial \theta}$ . The deformation  
 316 is shown by the plot of the von Mises equivalent strain rate  $\dot{\epsilon}_{\text{vM}} = f(P)$  (a  
 317 measure of the deviatoric strain rate (Tome et al., 1984; Wenk et al., 2000)).  
 318 The normalised concentration of the solute  $c/\Delta c(\tilde{t} = 1)$  is also plotted. All  
 319 results are shown in Fig. 1.

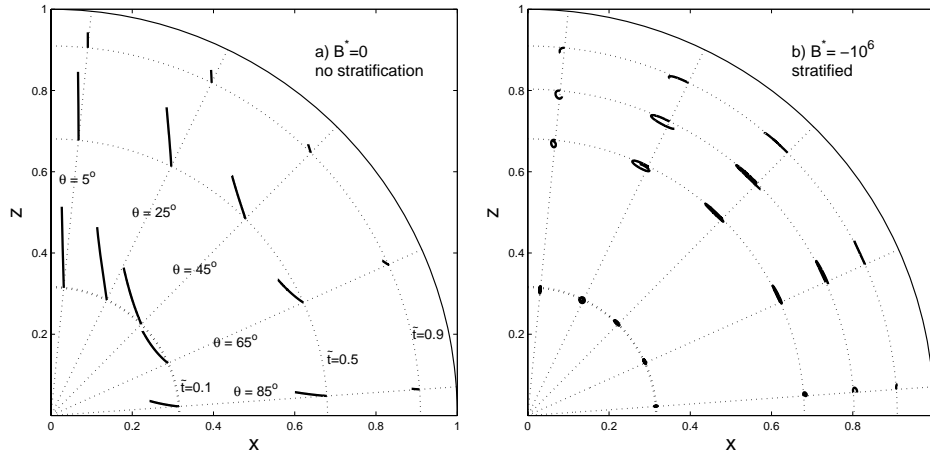


Figure 2: Trajectories within the inner core. Twenty samples are introduced at the inner core boundary at different colatitude,  $\theta = 5, 25, 45, 65$  and  $85$  degrees, and different times,  $\tilde{t} = 0.1, 0.5, 0.7$ , and  $0.9$ . (a) model with no stratification ( $B^* = 0$ ), the time  $\tilde{t} = 0.7$  is omitted for clarity. (b) model with stratification ( $B^* = -10^6$ ).

## 320 2.4. Numerical results

### 321 2.4.1. Neutral flows, $B = 0$

322 The case  $B = 0$  (Fig. 1) corresponds to no back reaction of the archimedean  
 323 force on the flow. Yoshida et al. (1996) have derived analytically the solution  
 324 and we can compare our numerical result shown in figure 1 to their analyt-  
 325 ical model. The absence of stratification allows the flow to develop into the  
 326 whole inner core. Continuous relaxation of the dynamic topography results  
 327 in a flow from the equator to the poles with a quadripolar pattern. The  
 328 average strain rate is rather small  $\dot{\epsilon} \approx 3 \times 10^{-18} \text{ s}^{-1} \approx 0.1 \text{ Gyr}^{-1}$ . In other  
 329 words, it takes 1 Gyr to accumulate a deformation of 10% in the material  
 330 sample. The deformation induced texture is expected to be weak, as will be  
 331 seen in section 3, except at the center of the inner core where strain rate can

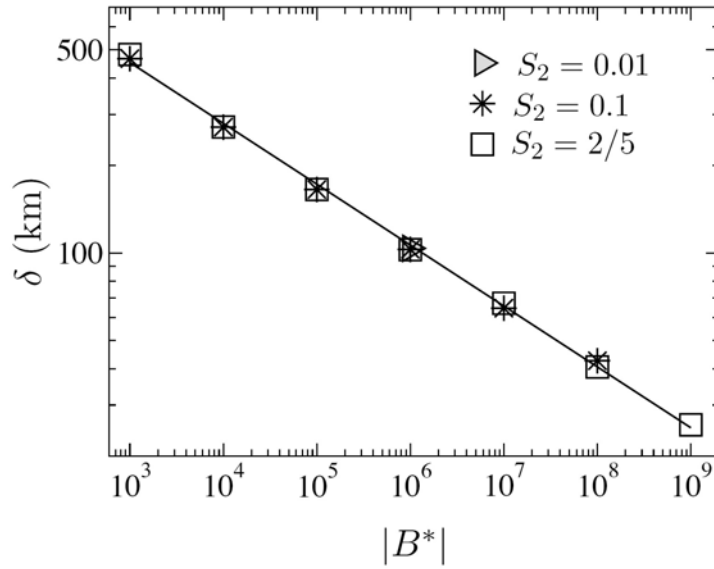


Figure 3: Evolution of the thickness  $\delta$  of the uppermost stratified layer with the dimensionless buoyancy number  $B^*$  for different values of differential growth parameter  $S_2$  (Eq. 10).

332 reach values of  $1.5 \text{ Gyr}^{-1}$  during the first stage of the inner core growth.

333 Note the strong deformation of the isoconcentration surfaces in the final  
 334 stage. It shows how the spherical boundary of the inner core in the past  
 335 is modified to take an axial oblate shape in the final inner core. This mo-  
 336 tion is also apprehended by the trajectories of samples set at the ICB and  
 337 transported by the flow during its later evolution as shown in figure 2a.

### 338 2.4.2. Stratified flows

339 Calculations for a stratified flow ( $B^* = -10^6$ ) are also presented in Fig. 1.  
 340 As explained in section 2.3, the magnitude of  $B$  increases from 0 to  $B^*$  during  
 341 the growth of the inner core. The effect of the stratification is negligible early  
 342 in the inner core history, but becomes rapidly important as the inner core

343 grows and  $B$  increases. The stratification tends to prevent radial motion  
344 because deformation of isodensity surfaces induces restoring archimedean  
345 forces. As the stratification strengthens, the flow is forced to follow isodensity  
346 surfaces and becomes quasi-horizontal. The radial penetration is strongly  
347 reduced and the flow is progressively confined in a thin shear layer below the  
348 ICB, of thickness  $\delta$ . An important consequence of the localization of the flow  
349 is that the magnitude of the strain rate which is associated becomes much  
350 larger, as can be seen in figure 1.

351 Momentum is transmitted deeper in the inner core by viscous entrain-  
352 ment, and the flow takes the form of a vertical series of elongated cells with  
353 vorticity of alternating sign. The specific geometry of the flow results from  
354 the competing effects of the stratification, which inhibits vertical motion,  
355 and of the symmetry of the forcing, which implies that the horizontal veloc-  
356 ity must vanish in the equatorial plane and on the N-S axis. In the vicinity  
357 of the equatorial plane and of the N-S axis, the flow is deflected and forced  
358 to be locally vertical. Stratification limits this vertical flow which is forced  
359 to rotate further until it becomes horizontal again (the deformation of the  
360 isodensity surfaces acts as a localized source of vorticity). The thickness of  
361 the resulting cells depends on the magnitude of the local stratification. The  
362 associated strain rate vanishes quickly in depth and is mainly concentrated  
363 in between the two first shallower cells.

364 Figure 2b shows the trajectories of different particles introduced at the  
365 surface of the inner core at different time during its growth. The trajectories  
366 are quasi horizontal below the ICB, with a maximum of amplitude at mid  
367 latitude. Deeper within the inner core, the particles follow the return flow

368 associated with the viscous counter cell. At polar and equatorial latitudes,  
 369 as well as in the deepest parts, the radial and horizontal displacement are  
 370 equivalent and the particles follow almost circular trajectories.

371 *2.4.3. Scaling laws for the thickness of the layers*

372 Simulations with different values of the differential growth  $S_2$  have been  
 373 performed (shown in figure 3). The amplitude of the flow is found to be  
 374 proportional to  $S_2$  but, perhaps surprisingly, the dimensionless depth  $\tilde{\delta}$  of  
 375 the alternating layers does not depend on  $S_2$  when  $S_2 < 1$ . Nevertheless,  
 376 as shown in figure 3, the thickness of the layer varies with  $B$  and we find  
 377  $\tilde{\delta} \approx 1.6|B|^{0.20 \pm 0.01}$  from a fit for  $3 < \log |B| < 9$ .

378 This scaling can be explained as follow. The thickness of the first cell  
 379 should correspond to the depth at which the buoyancy forces resulting from  
 380 the deformation of isodensity surfaces balance the viscous forces, by the mean  
 381 of pressure, and prevent any further vertical motion. The uppermost cell has  
 382 a dimensionless horizontal elongation equal to 1 and a dimensionless thickness  
 383  $\tilde{\delta}$ . From the dimensionless incompressibility equation, we can deduce that  
 384 the horizontal component of the velocity in the layer  $\tilde{v}_\theta$  is  $O(S_2 \tilde{\delta}^{-1})$  since  
 385 the vertical component  $\tilde{v}_r$ , imposed by the boundary condition is of order  $S_2$   
 386 ( $S_2 u_{icb}$  in dimensional unit). The vorticity  $\nabla \times \mathbf{v}$  is then  $O(S_2 \tilde{\delta}^{-2})$ . The  
 387 dimensionless vorticity equation (13) writes

$$\nabla^2 \nabla \times \mathbf{v} = B \frac{\partial \tilde{c}}{\partial \theta} \mathbf{e}_\phi, \quad (21)$$

388 from which a scaling for  $\tilde{\delta}$  can be derived. The scaling of  $\partial \tilde{c} / \partial \theta$  can be  
 389 found by considering the deformation by the flow in the upper cell of a newly  
 390 cristallized, initially horizontal, isocompositional surface. This isocompo-

391 sitional surface is progressively tilted by the flow and, at a time  $\delta\tilde{t}$  after  
 392 solidification, the resulting horizontal compositional gradient is of order

$$\frac{\partial\tilde{c}}{\partial\theta} \sim \left| \frac{\partial\tilde{c}}{\partial\tilde{r}} \right| \tilde{v}_r \delta\tilde{t} \sim S_2 \delta\tilde{t} \quad (22)$$

393 since  $|\partial\tilde{c}/\partial\tilde{r}| = O(1)$  and  $\tilde{v}_r = O(S_2)$ . It reaches a maximum in the transition  
 394 zone between the two upper cells, at a depth  $\tilde{\delta}$ , where vorticity changes sign.  
 395 If the deformation velocity is small compared to the mean growth rate of the  
 396 inner core, material crystallized at the ICB is buried to a depth  $\delta\tilde{t}$  in a time  
 397  $\delta\tilde{t} = \tilde{\delta}/u_{\text{icb}} = \tilde{\delta}$ . The maximum horizontal compositional gradient is then  
 398  $\partial\tilde{c}/\partial\theta = O(S_2 \tilde{\delta})$ . Using this scaling in equation (21), we deduce that the  
 399 thickness of the shear layer scales as

$$\tilde{\delta} \sim |B|^{-1/5}, \quad (23)$$

400 which has no dependence in  $S_2$ . The scaling exponent estimated above from  
 401 the numerical simulations is in very good agreement with the value  $1/5$  pre-  
 402 dicted by the scaling analysis.

403 The scaling law (23) implies that the strain rate in the uppermost shear  
 404 layer should scale as

$$\dot{\epsilon} \sim \frac{\tilde{u}_\theta}{\tilde{\delta}} \sim S_2 |B|^{2/5}. \quad (24)$$

405 This shows that stratification is necessary to reach a large deformation in  
 406 a time shorter than the age of the inner core. The  $2/5$  power is small but  
 407 the buoyancy number can easily reach values large enough to concentrate the  
 408 strain in the uppermost layer. With  $B^* = -10^6$ , the shear layer has a current  
 409 thickness  $\delta \sim 100$  km, and strain rates of  $0.015 \text{ Myr}^{-1}$  are obtained in the  
 410 uppermost layer. This corresponds to a horizontal cumulative deformation of



411 more than 100% in a material sample during its burying below the dynamical  
412 superficial layers. Note that since  $B$  is an increasing function of time,  $\delta$   
413 decreases and  $\dot{\epsilon}$  increases with time.

### 414 3. Mineralogical model

415 It is generally accepted that the stable phase of pure iron at inner core  
416 conditions is hexagonal close packed (*hcp*) (e.g. Mao et al., 1990; Ma et al.,  
417 2004; Dewaele et al., 2006). The presence of light elements in the core is said  
418 to stabilize cubic phases high temperature (Vočadlo et al., 2003; Dubrovinsky  
419 et al., 2007; Côté et al., 2008). Kuwayama et al. (2008) also argued that a  
420 transition from an *hcp* to an *fcc* phase could explain the enigmatic presence  
421 of the innermost inner core (Ishii and Dziewoński, 2002; Niu and Chen, 2008),  
422 although the range of composition at which this can occur is tight. For the  
423 sake of simplicity, in this paper, we will restrict ourself and assume that the  
424 inner core is composed of pure *hcp*-Fe.

#### 425 3.1. Plastic properties

426 To this day, the determination of active deformation mechanisms in iron  
427 at core conditions remains an active field of research (Poirier and Price,  
428 1999; Wenk et al., 2000; Merkel et al., 2004; Miyagi et al., 2008; Liermann  
429 et al., 2009) and more experiments will be needed to properly constrain those  
430 parameters.

431 At room temperature, plastic deformation in *hcp*-Fe is controlled by a  
432 dominant  $(0001)\langle\bar{1}2\bar{1}0\rangle$  basal slip, with a contribution of  $\{10\bar{1}0\}\langle\bar{1}2\bar{1}0\rangle$  pris-  
433 matic slip, and other minor contributions (Wenk et al., 2000; Merkel et al.,  
434 2004). At higher temperature, pyramidal  $\langle c+a\rangle$ ,  $\{2\bar{1}\bar{1}2\}\langle 2\bar{1}\bar{1}3\rangle$  becomes more

Slip type	Plane	Direction	CRSS
Basal	(0001)	$\langle\bar{1}2\bar{1}0\rangle$	0.5
Prismatic	$\{10\bar{1}0\}$	$\langle\bar{1}2\bar{1}0\rangle$	1.0
Pyramidal $\langle a \rangle$	$\{10\bar{1}1\}$	$\langle\bar{1}2\bar{1}0\rangle$	3.0
Pyramidal $\langle c + a \rangle$	$\{2\bar{1}\bar{1}2\}$	$\langle 2\bar{1}\bar{1}3 \rangle$	2.0

Table 2: Slip systems and their critical resolved shear stress (CRSS) of *hcp*-Fe used in our simulations.

435 dominant (Miyagi et al., 2008) and allows an easier rotation of the Fe grains.  
436 In agreement with those experimental results, we assume easy basal slip and  
437 allow significant contribution of prismatic and pyramidal  $\langle c+a \rangle$  slip (Table 2).

438 LPO in Fe polycrystals were simulated using the Los Alamos viscoplastic  
439 self-consistent (VPSC) code of Lebensohn and Tome (1993). The VPSC  
440 model treats each grain as an inclusion in a homogeneous but anisotropic  
441 medium that has the average properties of the polycrystal. It is intermediate  
442 between the Taylor model that enforces strain compatibility and the Sachs  
443 model that is based on stress equilibrium. One to one grain interaction  
444 or intergranular heterogeneities are not accounted for directly, but using a  
445 mean field approach. As deformation proceeds, crystals deform and rotate to  
446 generate preferred orientation. By applying different critical resolved shear  
447 stresses (CRSS) to slip systems and twin modes, the model will favor one  
448 deformation mode over another. It was already used for inner core modeling  
449 by Wenk et al. (2000) and Buffett and Wenk (2001) and has been reliable  
450 for simulating textures of many low symmetry materials (Wenk, 1999).

451 For each deformation step, we extracted the velocity gradients for markers

452 placed inside our inner core formation models. Those gradients are then used  
453 as inputs for simulating texture evolution of a 3000 grains Fe aggregate using  
454 VPSC. Orientations in the aggregates are characterized by an orientation  
455 distribution functions (ODF). The ODF is a probability function for finding  
456 an orientation and it is normalized over the whole orientation space to unity.  
457 An aggregate with a random distribution function has a probability of one  
458 for all orientations, or one multiple of a random distribution (m.r.d.). If  
459 preferred orientation is present, some orientations have probabilities higher  
460 than one and others lower than one but orientation probabilities can never  
461 be less than zero. Three dimensional ODF are not easy to visualize. Here,  
462 we will display pole figures of the  $\langle a \rangle$  and  $\langle c \rangle$  directions. In order to match  
463 the meridional representation of the flow, we place the  $z$ -direction (axis of  
464 rotation of the Earth) at the top of the figure, the  $x$ -axis (cylindrical radial  
465 direction) at the right of the figure, and the  $y$ -axis (azimuthal direction) at  
466 the center.

### 467 3.2. Initial texture

468 Simulations were run with both a random initial texture and a pretexture  
469 due to solidification. It seems quite likely that a texture can be frozen-in at  
470 the ICB during solidification (Bergman, 1997; Brito et al., 2002), in partic-  
471 ular if the inner core grows dendritically, as predicted by theoretical analysis  
472 (Fearn et al., 1981; Shimizu et al., 2005; Deguen et al., 2007). Dendritic  
473 crystallization of *hcp* materials typically leads to cristallographic preferred  
474 orientation with the basal planes of the crystals parallel to the gradient of  
475 temperature (*e.g.* Bergman et al., 2000, 2003). Therefore, pretexture due to  
476 solidification at the ICB was represented by an ODF with the majority of

477 the  $\langle c \rangle$  axis in the horizontal plane and the  $\langle a \rangle$  axis perpendicular to the ICB  
478 (Fig. 4).

### 479 3.3. Time evolution of texturing

480 Figure 5 shows an example of outputs of the dynamical model assuming  
481 stratified flow ( $B^* = -10^6$ ), for a polycrystalline aggregate introduced at  
482 time  $\tilde{t} = 0.5$ , colatitude  $\theta = 45^\circ$ , and radius 0.96. Our calculations assume  
483 an incompressible fluid and thus  $\sum_i \dot{\epsilon}_{ii} = 0$ . For this particular tracer, de-  
484 formation is dominated by the  $xx$  and  $zz$  components of the deformation  
485 tensor with  $\dot{\epsilon}_{xx} \approx -\dot{\epsilon}_{zz}$ . This corresponds to simple shear applied at  $45^\circ$  with  
486 respect to the Earth reference frame, in the direction of flow (Fig. 5).

487 The rigid body rotation of the aggregate is given by  $\omega_y$ . This component  
488 is important in magnitude and induces a rotation of aggregate around the  
489 azimuthal axis that affects the texture in the material.

490 Corresponding textures simulated for an *hcp*-Fe aggregates are shown in  
491 Fig. 6 assuming crystallization textures of Fig. 4. Texture develops very  
492 quickly as the aggregate is in the top layer and migrates toward the pole.  
493 At time  $\tilde{t} = 0.6$ , texture is fully developed, with the  $a$  and  $c$ -axes of the ag-  
494 gregate mainly distributed along the North-South and East-West directions,  
495 respectively.

## 496 4. Evolution of lattice preferred orientation during inner core growth

### 497 4.1. Neutral flow, $B=0$

498 In this case, we do not account for stratification, and flow can develop  
499 into the whole inner core (section 2.4.1, Fig. 2a). At the very center of the

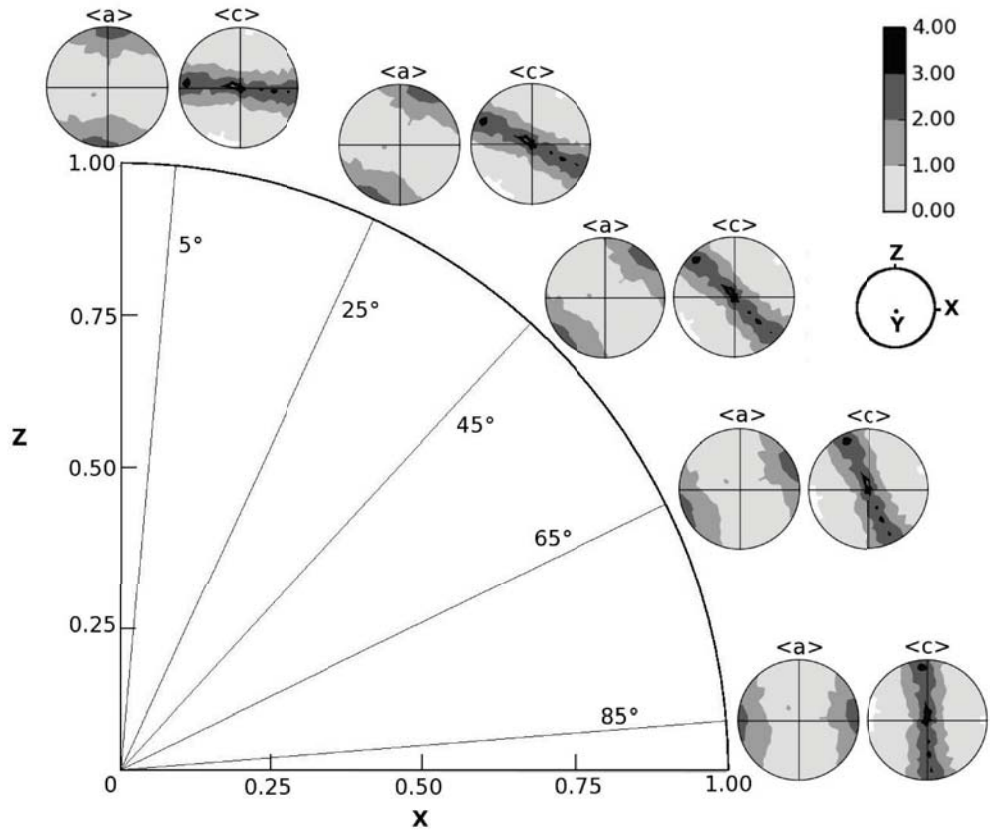


Figure 4: Pretexture due to solidification at the ICB, expressed using pole figures of the  $\langle a \rangle$  and  $\langle c \rangle$  directions, for colatitudes of 5, 25, 45, 65, and 85°. Linear scale, equal-area projection, contours in multiples of a random distribution (m.r.d.). Maximum intensity of the  $\langle a \rangle$  and  $\langle c \rangle$  pole figures are 3.1 and 4.3 m.r.d., respectively.

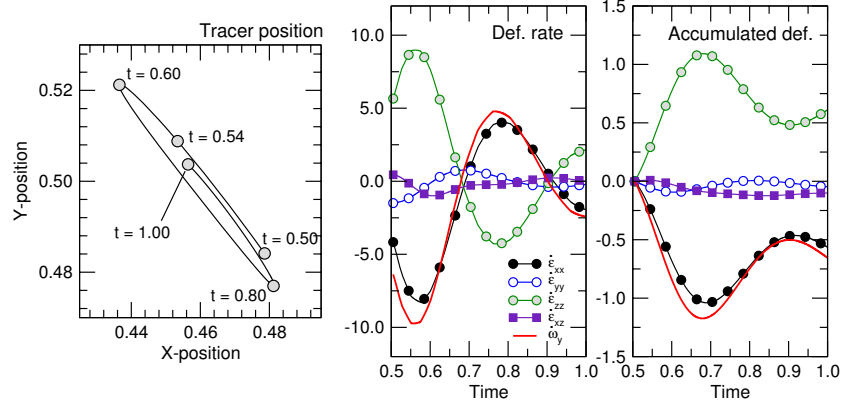


Figure 5: Trajectory, components of the strain rate tensor ( $\dot{\epsilon}_{xx}$ ,  $\dot{\epsilon}_{yy}$ ,  $\dot{\epsilon}_{zz}$ ,  $\dot{\epsilon}_{xy}$ ), rigid body rotation ( $\dot{\omega}_y$ ), and accumulated deformation ( $\epsilon_{xx}$ ,  $\epsilon_{yy}$ ,  $\epsilon_{zz}$ ,  $\epsilon_{xy}$ ) and rotation ( $\omega_y$ ) vs. time for a particle introduced at time  $\tilde{t} = 0.5$ , colatitude  $45^\circ$ , and radius 0.96.

500 inner core, plastic deformation is large enough to modify the orientation of  
 501 the aggregates at the beginning of the inner core formation. Later on, and  
 502 consequently further out in the inner core, deformation becomes weaker and  
 503 is not sufficient to induce any texture development.

504 With a random crystallization texture, strong LPO concentrate at the  
 505 center of the inner core (Fig. 7, aggregates introduced at  $\tilde{t} = 0.1$ ). The inner  
 506 core itself could be seen as a sample in a pure shear deformation with a com-  
 507 pression in the equatorial plane and an extension along the polar axis. This  
 508 explains why the samples crystallized at  $\tilde{t} = 0.1$  all show the same LPO.  
 509 The distribution of  $c$ -axis shows a pronounced maximum  $10^\circ$  to  $15^\circ$  away  
 510 from the equatorial plane while  $a$  axis concentrate toward the poles and a  
 511 secondary maximum at colatitudes of  $60^\circ$ . The  $10^\circ$  to  $15^\circ$  deviation of  $c$ -axis  
 512 from the equatorial plane may be surprising at first, but can be related to  
 513 deformation experiments on hcp-Fe as well as numerical modeling of com-

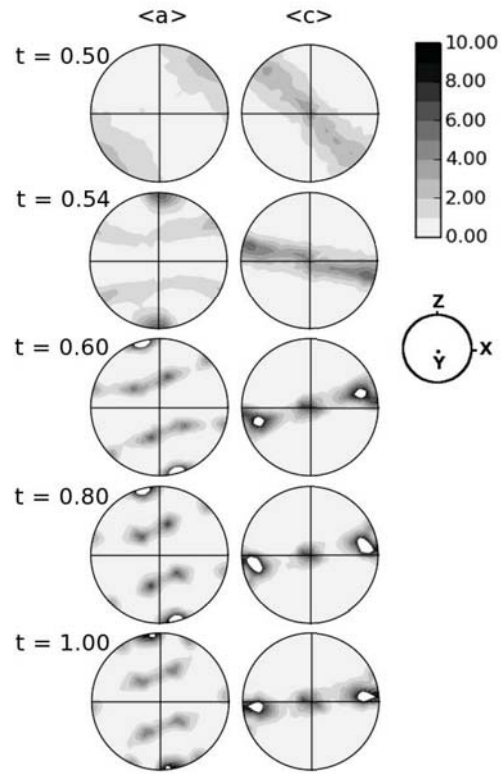


Figure 6: Pole figures of the  $\langle a \rangle$  and  $\langle c \rangle$  directions illustrating the LPO simulated for an *hcp*-Fe aggregate introduced in the inner core time  $\tilde{t} = 0.5$ , colatitude  $45^\circ$ , and radius 0.96 (Fig. 5), assuming a crystallization texture as in Fig. 4. Grey scale in m.r.d.

Model with no stratification, random starting texture

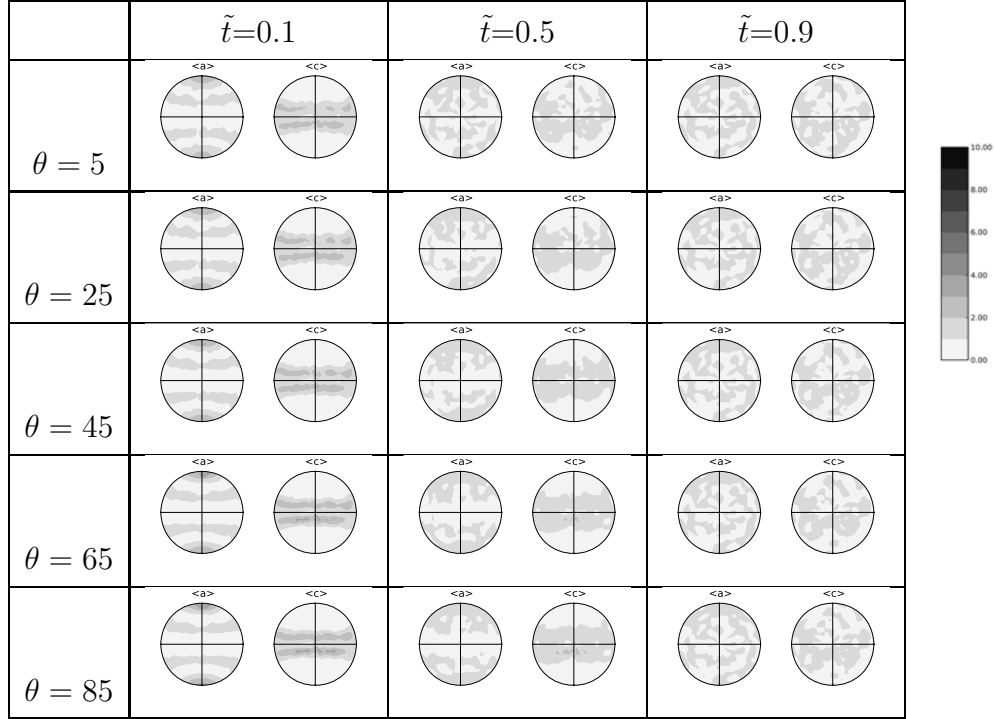


Figure 7: Pole figures of the  $\langle a \rangle$  and  $\langle c \rangle$  directions representing present day LPO in hcp-Fe aggregates introduced at colatitudes  $\theta = 5, 25, 45, 65$  and  $85^\circ$  and times,  $\tilde{t} = 0.1, 0.5$ , and  $0.9$ . Inner core growth model with no stratification and assuming random texture after crystallization. Corresponding trajectories can be seen in Fig. 2a. Linear scale, equal-area projection, contours in multiples of a random distribution (m.r.d.), axis as in figure 6.



Model with no stratification and initial texture

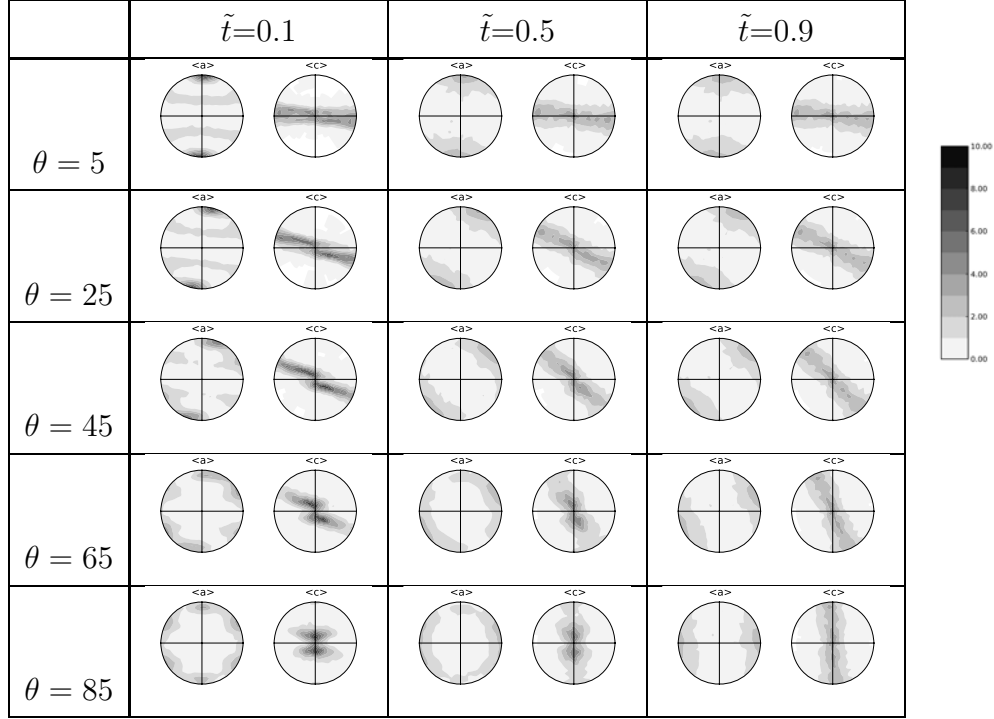


Figure 8: Pole figures of the  $\langle a \rangle$  and  $\langle c \rangle$  directions representing present day LPO in hcp-Fe aggregates introduced at colatitudes  $\theta = 5, 25, 45, 65$  and  $85^\circ$  and times,  $\tilde{t} = 0.1, 0.5,$  and  $0.9$ . Inner core growth model with no stratification and assuming crystallization textures as in Fig. 4. Corresponding trajectories can be seen in Fig. 2a. Linear scale, equal-area projection, contours in multiples of a random distribution (m.r.d.), axis as in figure 6.

514 pression experiments (Merkel et al., 2004) in which compressions textures do  
515 not align exactly with 0001 but are shifted  $10^\circ$  to  $15^\circ$  away from it.

516 If one assumes an initial crystallization texture as in Fig. 4, a long time  
517 is necessary for the textures to be affected by the flow, as can be seen in  
518 Fig. 8. Only aggregates in the central portion of the inner core (introduced  
519 at  $\tilde{t} = 0.1$  and  $\tilde{t} = 0.5$  in Fig. 8) are significantly affected by the deformation.  
520 Aggregates in the outer section of the core (introduced at  $\tilde{t} = 0.9$  in Fig. 8)  
521 preserve their crystallization texture. Close to the axis of rotation, the plastic  
522 deformation is compatible with the pretexturing and reinforce the alignment  
523 of the  $a$ -axis with the geographic axis (e.g.  $\tilde{t} = 0.1$ ,  $\theta = 5^\circ$ ). In the equatorial  
524 region (e.g.  $\tilde{t}=0.1$ ,  $\theta = 85^\circ$ ), the  $c$ -axis originally aligned nearly perpendicular  
525 to the  $z$  axis tend to move perpendicular to the meridional plane, inducing  
526 an hexagonal distribution of the  $a$ -axis in the meridional plane. This is a  
527 signature of the hexagonal structure of the crystal and its sliding planes.

#### 528 4.2. Stratified flow, $B = -10^6$

529 For a stratified flow, and assuming random crystallization textures, aggre-  
530 gates tend to develop strong textures in the outer layers of the core (Fig. 9)  
531 where maximum plastic deformation is attained (Fig. 1). Near the center  
532 ( $r < 0.5$ , aggregates introduced at  $\tilde{t} = 0.1$ ), textures are weak and the sig-  
533 nature of plastic deformation can not be detected. This can be attributed  
534 to the very moderate buoyancy number and the small corresponding strain  
535 rate (24) at the beginning of inner core formation. Later in the history of the  
536 inner core, the buoyancy number increases and the material acquires LPO  
537 by simple shear in the very thin superficial layer in a relatively short time  
538 scale.

Model with stratification and no starting texture

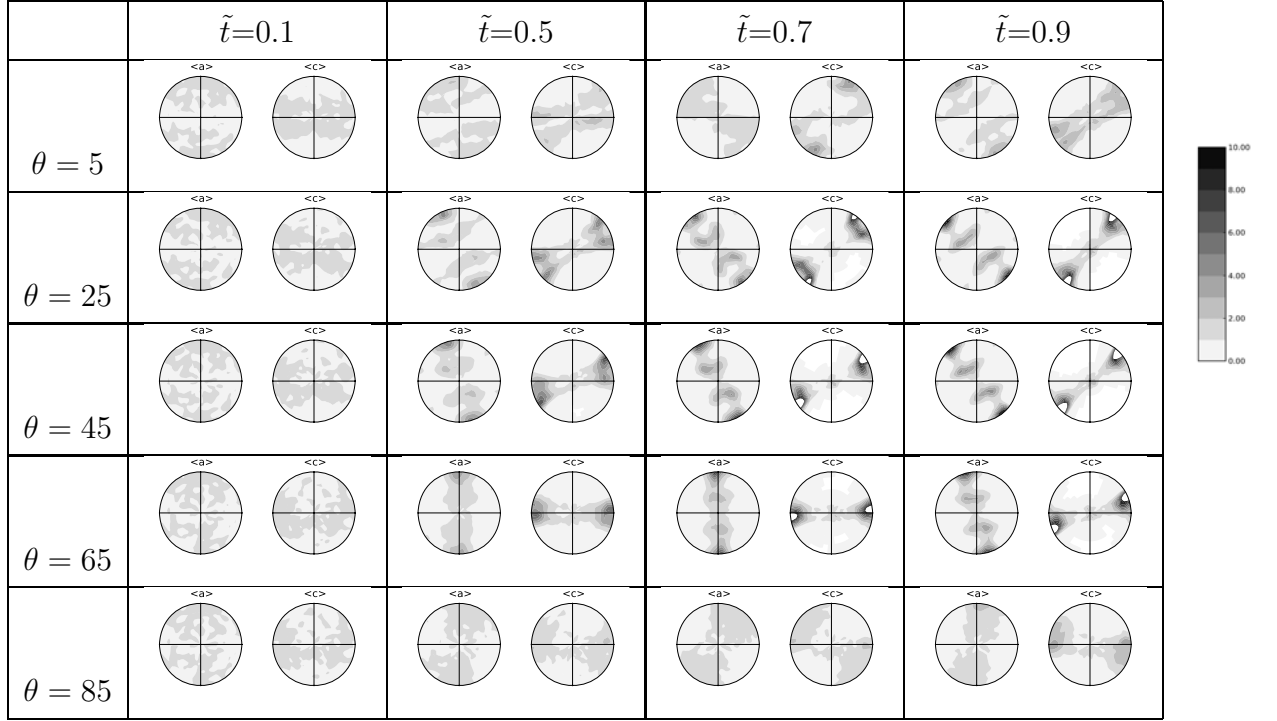


Figure 9: Pole figures of the  $\langle a \rangle$  and  $\langle c \rangle$  directions representing present day LPO in hcp-Fe aggregates introduced at colatitudes  $\theta = 5, 25, 45, 65$  and  $85^\circ$  and times,  $\tilde{t} = 0.1, 0.5, 0.7,$  and  $0.9$ . Inner core growth model with with stratification  $B = -10^6$  and assuming random starting textures. Corresponding trajectories can be seen in Fig. 2b. Linear scale, equal-area projection, contours in multiples of a random distribution (m.r.d.), axis as in figure 6.

Model with stratification and an initial texture

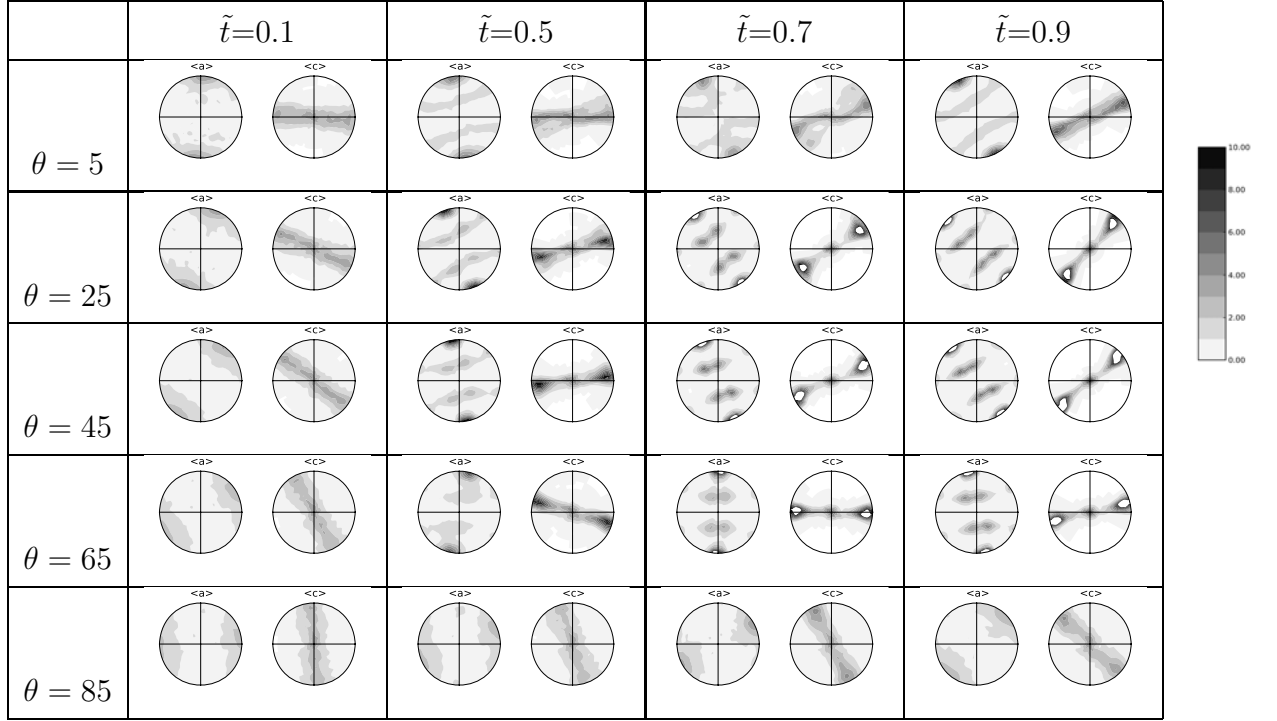


Figure 10: Pole figures of the  $\langle a \rangle$  and  $\langle c \rangle$  directions representing present day LPO in hcp-Fe aggregates introduced at colatitudes  $\theta = 5, 25, 45, 65$  and  $85^\circ$  and times,  $\tilde{t} = 0.1, 0.5, 0.7$ , and  $0.9$ . Inner core growth model with stratification  $B = -10^6$  and assuming crystallization textures as in Fig. 4. Corresponding trajectories can be seen in Fig. 2b. Linear scale, equal-area projection, contours in multiples of a random distribution (m.r.d.), axis as in figure 6.

539 In the uppermost shear layer, at mid latitude, the  $c$ -axis tends to orientate  
540 along the radial direction (the local vertical), perpendicular to the shear  
541 plane. The  $a$ -axis are mainly in the local horizontal plane with a primary  
542 orientation in the direction of the shear and a secondary direction at  $60^\circ$  off  
543 the meridional plane. In deeper portions of the core, rigid body rotation and  
544 simple shear change these orientations but by less than  $20^\circ$ .

545 In the equatorial disk and the polar cone, Fe aggregates exhibit a weak  
546 texture: the vertical migration of material in these region is associated with  
547 pure shear but is much smaller. The pure shear deformation would tend  
548 to align the  $c$ -axis with the major compression direction, and  $a$ -axis in the  
549 extension direction. From the scaling laws deduced in section 2.4.3, we can  
550 show that the deformation associated with the pure shear is a factor  $\tilde{\delta}$  smaller  
551 than the simple shear at mid latitude. In those two regions, rigid body rota-  
552 tion of the aggregates plays an important role and rotates the deformation  
553 texture around the  $y$ -axis (clockwise in the equatorial disk and the opposite  
554 and the polar region).

555 In summary, for a stratified inner core, large texturing is produced mainly  
556 in the thin outer layer at mid latitude and is preserved for  $r > 0.5$ .

557 Assuming initial crystallization texture as in Fig. 4, the starting texture  
558 is preserved in the central portion of the core (Fig. 10, aggregates introduced  
559 at  $\tilde{t} = 0.1$ ). Along the axis of rotation (Fig. 10,  $\theta = 5^\circ$ ) and the equatorial  
560 plane (Fig. 10,  $\theta = 85^\circ$ ), the solidification texture is rotated by the rigid  
561 body rotation of the sample, as shown in Fig. 2b. In the superficial shear  
562 layer, the texture is heavily transformed to textures resembling that of the  
563 Fig. 9, when a random crystallization texture was assumed.

## 564 **5. Predicted anisotropy of the inner core**

### 565 *5.1. Elastic properties of hcp-Fe at inner core conditions*

566 Calculation of seismic wave velocities inside crystal aggregates requires  
567 the knowledge of single crystal elastic moduli. The elastic properties of iron  
568 are difficult to evaluate experimentally at inner core conditions (e.g. Mao  
569 et al., 1998; Fiquet et al., 2001; Merkel et al., 2005; Antonangeli et al., 2006;  
570 Badro et al., 2007; Mao et al., 2008) while they remain a challenge for the  
571 most advanced *first principles* calculations (e.g. Stixrude and Cohen, 1995;  
572 Laio et al., 2000; Steinle-Neumann et al., 2001; Gannarelli et al., 2003, 2005;  
573 Vočadlo et al., 2009; Sha and Cohen, 2010). From those experiments and cal-  
574 culations, the anisotropy of *hcp* iron remains unclear. Low temperature, high  
575 pressure calculations (Stixrude and Cohen, 1995; Laio et al., 2000; Steinle-  
576 Neumann et al., 2001; Gannarelli et al., 2003, 2005; Vočadlo et al., 2009;  
577 Sha and Cohen, 2010) suggest that P-wave propagation is faster along the  
578 *c*-axis than along the *a*-axis, in agreement with low pressure/low temperature  
579 *hcp* analogs. Experimental determination of the elastic constants suggest a  
580 fast direction lying at an intermediate angle between the *a* and *c*-axis (Mao  
581 et al., 1998; Merkel et al., 2005), but the technique that was used includes se-  
582 rious artifacts related to stress heterogeneities induced by plastic deformation  
583 (Antonangeli et al., 2006; Merkel et al., 2009). At inner core pressures and  
584 temperatures, Steinle-Neumann et al. (2001) found a reversal of anisotropy,  
585 with the *a*-axis faster than the *c*-axis but the validity of their calculation has  
586 been questioned (Gannarelli et al., 2003, 2005). Yet recent *ab initio* calcula-  
587 tions at inner core pressure and temperature by Vočadlo et al. (2009) and Sha  
588 and Cohen (2010) again suggest that anisotropy in iron changes at high tem-

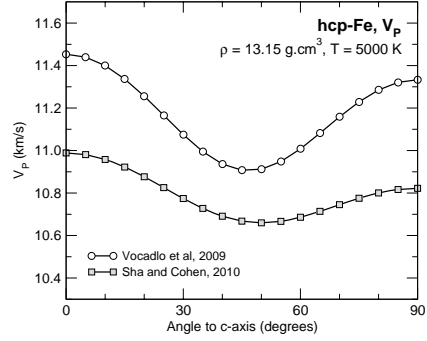


Figure 11: P-wave velocity in hcp-iron at 5000 K and core density according to the *ab initio* calculations of Vočadlo et al. (2009) and Sha and Cohen (2010) as a function of propagation direction with respect to the *c*-axis. Results of Sha and Cohen (2010) have been interpolated to a density of  $\rho = 13.15 \text{ g.cm}^{-3}$  and temperature  $T = 5000 \text{ K}$ .

589 perature. Results of the two calculations differ, both for the average *P*-wave  
 590 velocity and the amplitude of anisotropy (Table 3 and Fig. 11). However,  
 591 both calculations find that *P*-waves travel faster along the *c* than the *a* axis,  
 592 with a pronounced minimum  $45^\circ$  away from *c*. Here, we will use the set of  
 593 elastic moduli of Vočadlo et al. (2009). Calculations using the elastic moduli  
 594 of Sha and Cohen (2010) would not change the style of anisotropy, but only  
 595 the absolute *P*-waves velocities and the amplitude of anisotropy.

### 596 5.2. Model with no stratification and a random crystallization texture

597 Figures 12a shows the present day P wave velocities inside the inner core  
 598 obtained for a growth model with no stratification and assuming a random  
 599 crystallization texture. In the central portion, P wave velocity is faster along  
 600 the NS axis than any equatorial or mid latitude direction. In the outer  
 601 portions of the core, anisotropy is weak.

602 As in Yoshida et al. (1996), the orientation of anisotropy has a cylindrical

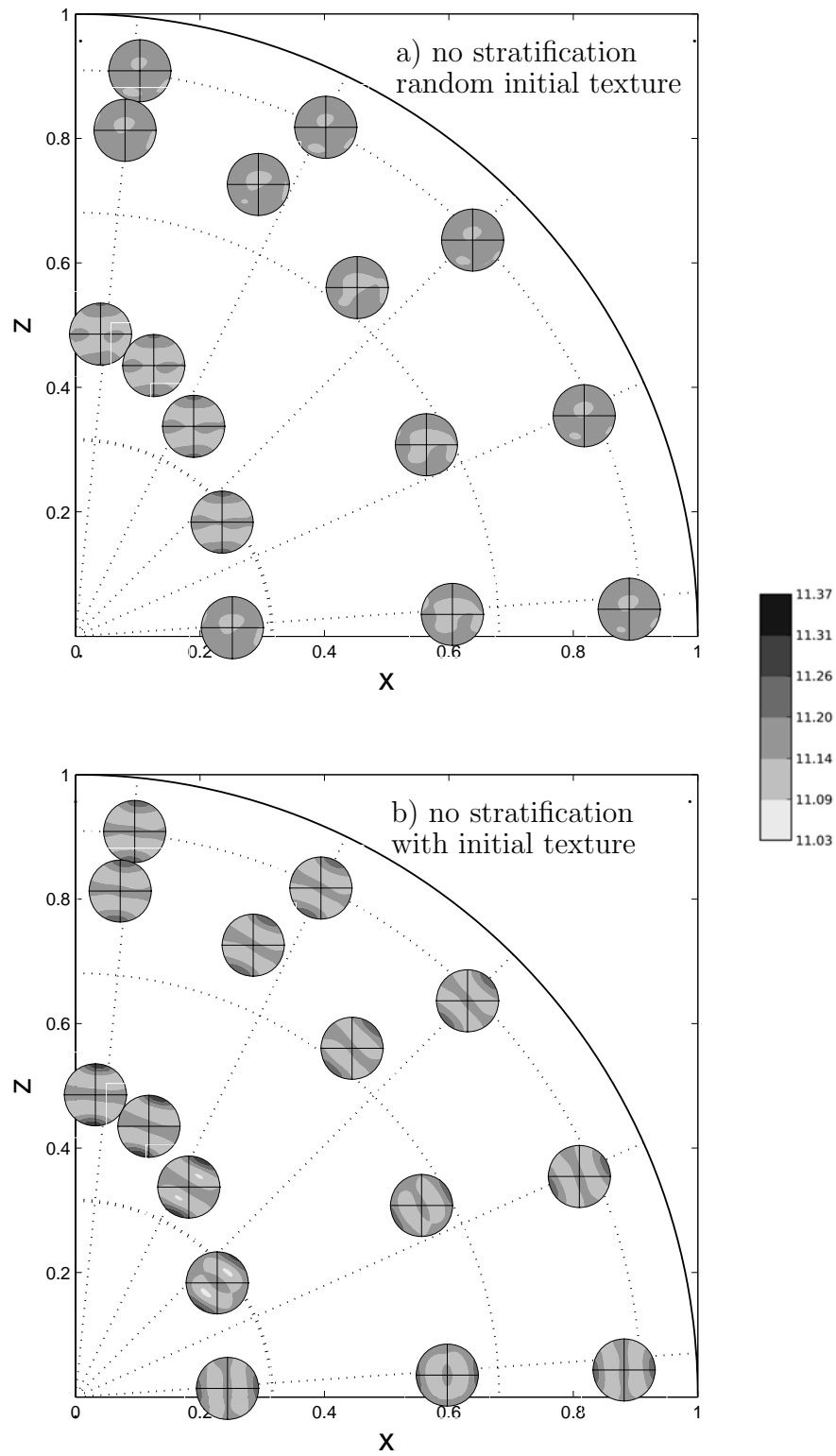


Figure 12: Present day P wave velocity anisotropy inside the inner core for a model with no stratification ( $B = 0$ ). (a) Model with a random solidification texture, corresponding to the hcp-Fe textures of Fig. 7. (b) Model with initial solidification texture, corresponding to the hcp-Fe textures of Fig. 8. Velocity scale are in km/s. Each contour corresponds to a 0.5% change in velocity.



	Vočadlo <i>et al</i> (2009)	Sha and Cohen (2010)
$c_{11}$	1689	1540
$c_{33}$	1725	1588
$c_{12}$	1186	1019
$c_{23}$	990	920
$c_{44}$	216	254
$\rho$	13.154 g.m <sup>-3</sup>	13.15 g.cm <sup>-3</sup>

Table 3: Interpolated isothermal elastic constants (in GPa) for hcp-Fe at 5000 K.  $c_{66} = (c_{11} - c_{12})/2$

603 symmetry with the fastest direction along the N-S axis, in agreement with  
604 seismological observations, but the predicted amplitude of anisotropy is sig-  
605 nificantly weaker than observed. At most, the P-wave velocity is 1% faster  
606 along the NS axis in the very central part ( $r < 0.5$ ). As such, global travel  
607 time anisotropy would be significantly less than 1%. This is far from the 3%  
608 anomaly observed by seismologists (Souriau, 2003). In addition, the slowest  
609 direction in this model is at about 45° from the N-S axis.

610 Under our assumptions (equatorial growth model, plastic model, and hcp-  
611 Fe elastic properties), this simple model has three major weaknesses:

- 612 (i) The direction of anisotropy is correctly predicted, but the cumulated  
613 deformation is not sufficient to generate the observed inner core anisotropy.
- 614 (ii) Seismological observations indicate that the uppermost inner core has  
615 distinct seismological properties. This model does not generate any  
616 specific superficial layer at the top of the inner core.
- 617 (iii) This model requires that the axis of rotation of the inner core did not

618 significantly wobble during the time life of the inner core.

619 *5.3. Model with no stratification and a non-random crystallization texture*

620 Solidification texturing can produce strong anomalies at the top of the  
621 inner core. In this model, directional variations of P-wave velocities can be  
622 seen throughout the core, with anisotropy amplitudes reaching about 1.5%  
623 (Fig. 12b). In most of the core, the calculated fast direction of propagation is  
624 mostly radial, with a secondary maximum in the plane perpendicular to the  
625 radial direction. This configuration poorly explains the seismological data as  
626 ray paths integrations would not show a strong variation of travel time with  
627 the ray angle.

628 *5.4. Model with stratification and a random crystallization texture*

629 Stratification inhibits radial flows, concentrates the deformation in a su-  
630 perfacial horizontal layer and produces important simple shear that can gen-  
631 erate strong LPO and large seismic anisotropy. Figure 13a shows the present  
632 day P-wave velocity distribution inside the inner core for a model with mod-  
633 erate stratification and assuming a random crystallization texture.

634 The amplitude of anisotropy is large (3%) but is mostly localized in  
635 the outer portion of the inner core, away from the equator and the pole.  
636 The directions of fast propagation are mainly oriented along the direction  
637 of shear, horizontal relative to ICB, with a slight rotation as the material  
638 plunges deeper into the core. This rotations induces a shift of fast direction  
639 of propagation from the local horizontal towards a North South direction  
640 (e.g.  $r \approx 0.7$ ). Note the quadrangular shape of the anisotropy figure which

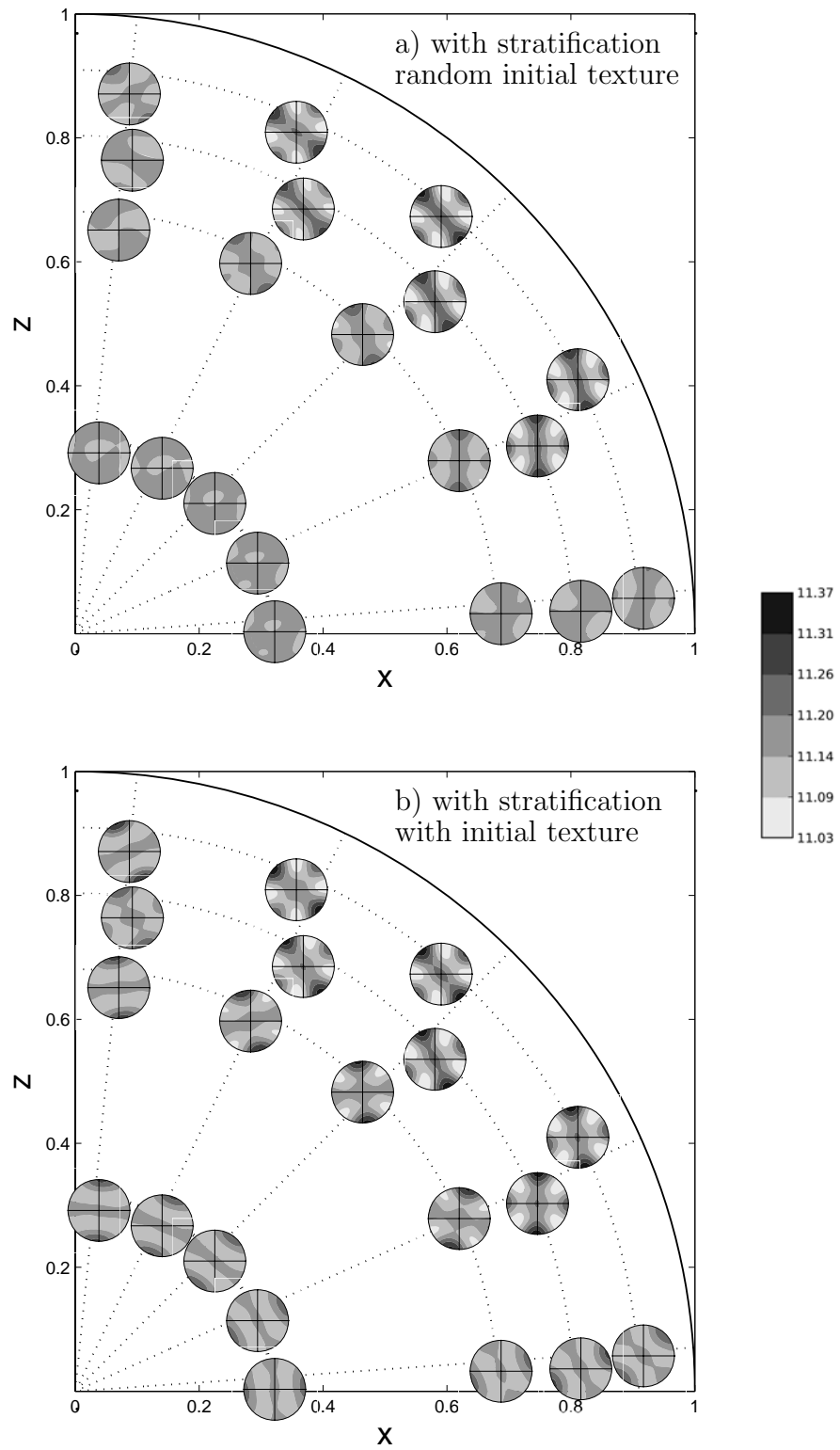


Figure 13: Present day P wave velocity anisotropy inside the inner core for a model with stratification ( $B = -10^6$ ). (a) Model with a random solidification texture, corresponding to the hcp-Fe textures of Fig. 7. (b) Model with initial solidification texture, corresponding to the hcp-Fe textures of Fig. 8. Velocity scale are in km/s. Each contour corresponds to a 0.5% change in velocity.

641 is the reminiscence of the P wave velocity minimum at  $45^\circ$  away from  $a$  and  
642  $c$  in the single crystal (Fig. 11).

643 In the center ( $r < 0.5$ ), anisotropy is weak. Lack of stratification early in  
644 the inner core history and stratification in the later stages of core formation  
645 inhibit deformation, therefore LPO, in this region. This lack of anisotropy  
646 in the deep inner core is in strong contradiction with seismic observations.

#### 647 *5.5. Model with stratification and a non-random crystallization texture*

648 As in the previous model, there is large anisotropy in the outside portion  
649 of the inner core, away from the equator and the pole (Fig. 13b). In these  
650 region, the crystallization texture is erased and the observed anisotropy is the  
651 result of the large shear flow induced by stratification. Below  $r < 0.5$ , LPO  
652 associated with the solidification process are preserved. Along the equator  
653 and the N-S axis, the anisotropy induced by solidification is slightly rotated.  
654 Indeed, in those regions, aggregate are submitted to significant rigid-body  
655 rotation, but little plastic deformation (e.g. Fig.2).

656 The strong radial anisotropy in the uppermost layer would be difficult to  
657 detect in seismology. N-S and E-W velocities are equivalent. Moreover, it  
658 would be difficult to compare horizontal and radial traveltimes in a superficial  
659 layer  $\sim 100$  km thick. Indeed, most studies of the uppermost inner core are  
660 based on comparison of almost horizontal rays with different orientations  
661 and would not capture our predicted anisotropy. As such our model is not  
662 inconsistent with observations of isotropy in the uppermost inner core.

663 Textures in the outermost inner core are a result of intense plastic defor-  
664 mation, and this process is fast (less than 10 Myr). In contrast, anisotropies  
665 in the deep inner core, along the equator or the N-S axis, are mostly fossil

666 and would therefore be very sensitive to wobbles of the inner core.

## 667 **6. Conclusion**

668 A fundamental question regarding the inner core dynamics remains whether  
669 it is stably stratified or not. Because of the large uncertainties on the age of  
670 the inner core and on most of thermo-physical parameters of the core, and on  
671 most of thermo-physical parameters of the core, in particular on the thermal  
672 conductivity, it is currently unknown whether the inner core has a stable or  
673 unstable density stratification. If the inner core is unstably stratified, con-  
674 vection can take the form of a 'convective translation' (Alboussière et al.,  
675 2010; Monnereau et al., 2010) if the viscosity is larger than  $\sim 10^{18}$  Pa.s (Al-  
676 boussière et al., 2010). If the viscosity is smaller, it would be a more classical  
677 plume regime (Weber and Machetel, 1992; Deguen and Cardin, 2011) .

678 Here, we assumed that the inner core is stably stratified. Stratification  
679 prevents any strong vertical motions within the inner core and forces the flow  
680 to be as horizontal as possible. The flow induced by aspherical growth of the  
681 inner core is localized in a thin shear layer below the ICB, where the strain  
682 is potentially large enough to produce a significant texture.

683 We successfully coupled our model of inner core dynamics with a visco-  
684 plastic deformation model to compute LPO of hcp-Fe in the inner core and  
685 predict the resulting seismological anisotropy. We emphasize on the fact,  
686 that although the procedure we developed is quite general, the exact result  
687 is function of a number of assumptions regarding the thermo-chemical state  
688 of the inner core, the stable phase of iron at inner core condition, its rhe-  
689 ology, and its elastic properties. The choices we have made are consistent

690 with current beliefs and the most recent experimental and numerical stud-  
691 ies, but there is still no consensus on these questions. It will be important  
692 to test the sensitivity of our results to the mineralogical assumptions of the  
693 model. Other possible phases of Fe, different slip systems or different elas-  
694 tic properties may lead to different results and deserve more numerical and  
695 experimental studies.

696 We summarize the principal results of texture calculations.

697 (i) Without stratification ( $B^* \rightarrow 0$ , large viscosity and small stratifica-  
698 tion):

- 699 • The degree of deformation-induced LPO increases with depth. Without  
700 solidification texturing, the outer part of the inner core is isotropic and  
701 the central region exhibits a cylindrical anisotropy.
- 702 • Deformation texturing seems too weak however to explain a  $\sim 3\%$   
703 anisotropy.
- 704 • In addition, since the texturing timescale is comparable to the age of  
705 the inner core, the development of the deep anisotropy requires that  
706 there have been no significant inner core polar wander.
- 707 • Solidification textures, if present, are only weakly reworked by subse-  
708 quent deformation. The final structure of the inner core would therefore  
709 reflect the initial geometry of the solidification texture. With the solid-  
710 ification texture assumed here, the final texture would be spherically  
711 symmetric, in obvious disagreement with the seismological observa-  
712 tions. Solidification texturing might explain the cylindrical anisotropy  
713 of the inner core only if, as proposed by Bergman (1997), the initial  
714 texture exhibits a N-S cylindrical symmetry due to anisotropic heat

715 flow in the outer core.

716 (ii) With a stable stratification :

- 717 • Plastic deformation in the shear layer can erase a solidification texture,  
718 and produce a strong deformation texture and seismic anisotropy at  
719 mid latitudes.
- 720 • The resulting texture in the outer part of the inner core has a spherical  
721 symmetry, and is therefore not able to explain the global cylindrical  
722 anisotropy of the inner core.
- 723 • This layer would appear isotropic to body wave seismology, but might  
724 be detectable in normal mode data.
- 725 • In this model, it is possible to produce a layered inner core, with a deep  
726 fossil texture inherited from solidification texturing, and a deformation  
727 texture in the outer part of the inner core. The relative thickness of the  
728 two layers depend on the value of the buoyancy number  $B^*$ . A larger  
729 value of  $B^*$  would result in a smaller region of fossilized solidification  
730 texture. As in the case with no stratification, the deep inner core would  
731 have a cylindrical anisotropy only if the initial solidification texture has  
732 a N-S cylindrical symmetry (Bergman, 1997).

733 This last model seems to be the most compatible with the seismological  
734 data.

735 Hemispherical seismic variations (Tanaka and Hamaguchi, 1997; Garcia  
736 and Souriau, 2000; Deuss et al., 2010) should also be studied with non ax-  
737 isymmetric models of inner core growth as suggested by the experiments of  
738 Sumita and Olson (1999, 2002) and the numerical simulations of Aubert et al.

739 (2008).

740 *Acknowledgments.* All computations presented here were performed at the  
741 Service Commun de Calcul Intensif de l'Observatoire de Grenoble. We  
742 thank the undergraduate students Jonathan Serafini and Charlotte Reg-  
743 nier who participated to the study. R. Deguen was partly supported by  
744 NSF grant EAR#0909622. S.Merkel was supported by ANR program DiUP,  
745 N. ANR-07-JCJC-0136. The study was granted by the program SEDIT of  
746 INSU/CNRS.

## 747 **References**

748 Albarède, F., 1996. Introduction to geochemical modeling. Cambridge Uni-  
749 versity Press, Cambridge UK.

750 Alboussière, T., Deguen, R., Melzani, M., 2010. Melting induced stratifica-  
751 tion above the Earth's inner core due to convective translation. *Nature*  
752 466, 744–747.

753 Alfè, D., Gillan, M. J., Price, G. D., Apr. 2002. Ab initio chemical poten-  
754 tials of solid and liquid solutions and the chemistry of the Earth's core.  
755 *J. Chem. Phys.*116, 7127–7136.

756 Allègre, C. J., Poirier, J.-P., Humler, E., Hofmann, A. W., 1995. The chemical  
757 composition of the Earth. *Earth and Planetary Science Letters* 134, 515–  
758 526.



- 759 Antonangeli, D., Merkel, S., Farber, D. L., Dec. 2006. Elastic anisotropy in  
760 hcp metals at high pressure and the sound wave anisotropy of the Earth's  
761 inner core. *Geophys. Res. Lett.* 33, 24303–+.
- 762 Aubert, J., Amit, H., Hulot, G., Olson, P., Aug. 2008. Thermochemical flows  
763 couple the Earth's inner core growth to mantle heterogeneity. *Nature* 454,  
764 758–761.
- 765 Aurnou, J., Heimpel, M., Allen, L., King, E., Wicht, J., Jun. 2008. Convec-  
766 tive heat transfer and the pattern of thermal emission on the gas giants.  
767 *Geophysical Journal International* 173, 793–801.
- 768 Badro, J., Fiquet, G., Guyot, F., Gregoryanz, E., Ocelli, F., Antonangeli,  
769 D., D'Astuto, M., Feb. 2007. Effect of light elements on the sound velocities  
770 in solid iron: Implications for the composition of Earth's core. *Earth and*  
771 *Planetary Science Letters* 254, 233–238.
- 772 Bergman, M. I., 1997. Measurements of elastic anisotropy due to solidification  
773 texturing and the implications for the Earth's inner core. *Nature* 389, 60–  
774 63.
- 775 Bergman, M. I., Agrawal, S., Carter, M., M., M.-S., 2003. Transverse solidi-  
776 fication textures in hexagonal close-packed alloys. *J. Crystal Growth* 255,  
777 204–211.
- 778 Bergman, M. I., Giersch, L., Hinczewski, M., Izzo, V., Jan. 2000. Elastic and  
779 attenuation anisotropy in directionally solidified (hcp) zinc, and the seismic  
780 anisotropy in the Earth's inner core. *Physics of the Earth and Planetary*  
781 *Interiors* 117, 139–151.

- 782 Brito, D., Elbert, D., Olson, P., Feb. 2002. Experimental crystallization of  
783 gallium: ultrasonic measurements of elastic anisotropy and implications for  
784 the inner core. *Physics of the Earth and Planetary Interiors* 129, 325–346.
- 785 Buffett, B. A., Aug. 1997. Geodynamic estimates of the viscosity of the  
786 Earth’s inner core. *Nature* 388, 571–573.
- 787 Buffett, B. A., Nov. 2009. Onset and orientation of convection in the inner  
788 core. *Geophysical Journal International* 179, 711–719.
- 789 Buffett, B. A., Bloxham, J., Dec. 2000. Deformation of Earth’s inner core by  
790 electromagnetic forces. *Geophys. Res. Lett.* 27, 4001–4004.
- 791 Buffett, B. A., Garnero, E. J., Jeanloz, R., Nov. 2000. Sediments at the Top  
792 of Earth’s Core. *Science* 290, 1338–1342.
- 793 Buffett, B. A., Wenk, H.-R., Sep. 2001. Texturing of the Earth’s inner core  
794 by Maxwell stresses. *Nature* 413, 60–63.
- 795 Busse, F. H., 1970. Thermal instabilities in rapidly rotating systems. *Journal*  
796 *of Fluid Mechanics* 44, 441–460.
- 797 Calvet, M., Chevrot, S., Souriau, A., Jun. 2006. P-wave propagation in trans-  
798 versely isotropic media: II. Application to inner core anisotropy: Effects  
799 of data averaging, parametrization and a priori information. *Physics of the*  
800 *Earth and Planetary Interiors* 156, 21–40.
- 801 Cardin, P., Olson, P., Mar. 1994. Chaotic thermal convection in a rapidly  
802 rotating spherical shell: consequences for flow in the outer core. *Physics of*  
803 *the Earth and Planetary Interiors* 82, 235–259.

- 804 Côté, A. S., Vočadlo, L., Brodholt, J. P., 2008. Light elements in the core:  
805 Effects of impurities on the phase diagram of iron. *Geophys. Res. Lett.* 35,  
806 5306–+.
- 807 Crank, J., 1984. *Free and moving boundary problems*. Clarendon Press, Ox-  
808 ford UK.
- 809 Deguen, R., Alboussière, T., Brito, D., 2007. On the presence and structure  
810 of a mush at the inner core boundary of the Earth. *Phys. Earth Planet.*  
811 *Inter.* 274, 1887–1891.
- 812 Deguen, R., Cardin, P., Jun. 2009. Tectonic history of the Earth’s inner core  
813 preserved in its seismic structure. *Nature Geoscience* 2, 419–422.
- 814 Deguen, R., Cardin, P., 2011. Thermo-chemical convection in the Earth’s  
815 inner core. *Geophysical Journal International*, submitted.
- 816 Deuss, A., Irving, J. C. E., Woodhouse, J. H., May 2010. Regional Variation  
817 of Inner Core Anisotropy from Seismic Normal Mode Observations. *Science*  
818 328, 1018–.
- 819 Dewaele, A., Loubeyre, P., Occelli, F., Mezouar, M., Dorogokupets, P. I.,  
820 Torrent, M., 2006. Quasihydrostatic equation of state of iron above 2 mbar.  
821 *Phys. Rev. Lett.* 97, 215504.
- 822 Dormy, E., Soward, A. M., Jones, C. A., Jault, D., Cardin, P., Feb. 2004.  
823 The onset of thermal convection in rotating spherical shells. *Journal of*  
824 *Fluid Mechanics* 501, 43–70.

- 825 Dubrovinsky, L., Dubrovinskaia, N., Narygina, O., Kantor, I., Kuznetsov, A.,  
826 Prakapenka, V. B., Vitos, L., Johansson, B., Mikhaylushkin, A. S., Simak,  
827 S. I., Abrikosov, I. A., 2007. Body-Centered Cubic Iron-Nickel Alloy in  
828 Earth's Core. *Science* 316, 1880–.
- 829 Dziewonski, A. M., Anderson, D. L., Jun. 1981. Preliminary reference Earth  
830 model. *Physics of the Earth and Planetary Interiors* 25, 297–356.
- 831 Fearn, D., Loper, D., Roberts, P., 1981. Structure of the Earth's inner core.  
832 *Nature* 292, 232–233.
- 833 Fearn, D. R., Loper, D. E., Jan. 1981. Compositional convection and strati-  
834 fication of earth's core. *Nature* 289, 393–+.
- 835 Fiquet, G., Badro, J., Guyot, F., Requardt, H., Krisch, M., 2001. Sound  
836 velocities in iron to 100 gigapascals. *Science* 291, 468–471.
- 837 Gannarelli, C. M. S., Alfè, D., Gillan, M. J., 2003. The particle-in-cell model  
838 for ab initio thermodynamics: implications for the elastic anisotropy of  
839 the Earth's inner core. *Physics of the Earth and Planetary Interiors* 139,  
840 243–253.
- 841 Gannarelli, C. M. S., Alfè, D., Gillan, M. J., Sep. 2005. The axial ratio of  
842 hcp iron at the conditions of the Earth's inner core. *Physics of the Earth  
843 and Planetary Interiors* 152, 67–77.
- 844 Garcia, R., Sep. 2002. Constraints on upper inner-core structure from wave-  
845 form inversion of core phases. *Geophysical Journal International* 150, 651–  
846 664.

- 847 Garcia, R., Souriau, A., Oct. 2000. Inner core anisotropy and heterogeneity  
848 level. *Geophys. Res. Lett.* 27, 3121–3124.
- 849 Ishii, M., Dziewoński, A. M., 2002. The innermost inner core of the earth:  
850 Evidence for a change in anisotropic behavior at the radius of about 300  
851 km. *Proceedings of the National Academy of Science* 99, 14026–14030.
- 852 Jeanloz, R., Wenk, H., Jan. 1988. Convection and anisotropy of the inner  
853 core. *Geophys. Res. Lett.* 15, 72–75.
- 854 Karato, S., Dec. 1993. Inner Core Anisotropy Due to the Magnetic Field-  
855 Induced Preferred Orientation of Iron. *Science* 262, 1708–1711.
- 856 Karato, S., Dec. 1999. Seismic anisotropy of the Earth’s inner core resulting  
857 from flow induced by Maxwell stresses. *Nature* 402, 871–873.
- 858 Kuwayama, Y., Hirose, K., Sata, N., Ohishi, Y., 2008. Phase relations of iron  
859 and iron–nickel alloys up to 300 gpa: Implications for composition and  
860 structure of the earth’s inner core. *Earth and Planetary Science Letters*  
861 273 (3-4), 379–385.
- 862 Labrosse, S., Poirier, J.-P., Le Mouél, J.-L., 2001. The age of the inner core.  
863 *Earth planet. Sci. Lett.* 190, 111–123.
- 864 Laio, A., Bernard, S., Chiarotti, G. L., Scandolo, S., Tosatti, E., Feb. 2000.  
865 *Physics of Iron at Earth’s Core Conditions. Science* 287, 1027–1030.
- 866 Lebensohn, R. A., Tome, C. N., 1993. Self-consistent anisotropic approach  
867 for the simulation of plastic deformation and texture development of poly-

868 crystals: application to zirconium alloys. *Acta Metallurgica et Materiala*  
869 41, 2611–2624.

870 Liermann, H.-P., Merkel, S., Miyagi, L., Wenk, H.-R., Shen, G., Cynn, H.,  
871 Evans, W. J., 2009. New experimental method for in situ determination  
872 of material textures at simultaneous high-pressure and temperature by  
873 means of radial diffraction in the diamond anvil cell. *Rev. Sci. Instrum.*  
874 80, 104501.

875 Loper, D. E., 1983. Structure of the inner core boundary. *Geophys. and*  
876 *Astrophys. Fluid Dyn.* 25, 139–155.

877 Ma, Y., Somayazulu, M., Shen, G., Mao, H., Shu, J., Hemley, R., 2004. In  
878 situ X-ray diffraction studies of iron to Earth-core conditions. *Phys. Earth*  
879 *Planet. Inter.* 143-144, 455–467.

880 Mao, H.-K., Shu, J., Shen, G., Hemley, R. J., Li, B., Singh, A. K., Dec.  
881 1998. Elasticity and rheology of iron above 220GPa and the nature of the  
882 Earth's inner core. *Nature* 396, 741–743.

883 Mao, H. K., Wu, Y., Chen, L. C., Shu, J. F., Jephcoat, A. P., 1990. Static  
884 compression of iron to 300 GPa and Fe<sub>0.8</sub>Ni<sub>0.2</sub> alloy to 260 GPa: Implica-  
885 tions for composition of the core. *J. Geophys. Res.* 95 (B13), 21737–21742.

886 Mao, W. L., Struzhkin, V. V., Baron, A. Q. R., Tsutsui, S., Tommaseo,  
887 C. E., Wenk, H.-R., Hu, M. Y., Chow, P., Sturhahn, W., Shu, J., Hemley,  
888 R. J., Heinz, D. L., Mao, H.-K., 2008. Experimental determination of the  
889 elasticity of iron at high pressure. *J. Geophys. Res.* 113, B09213.

- 890 Masters, G., Gubbins, D., Nov. 2003. On the resolution of density within the  
891 Earth. *Physics of the Earth and Planetary Interiors* 140, 159–167.
- 892 Merkel, S., Shu, J., Gillet, P., Mao, H.-K., Hemley, R. J., May 2005. X-ray  
893 diffraction study of the single-crystal elastic moduli of  $\epsilon$ -Fe up to 30 GPa.  
894 *Journal of Geophysical Research (Solid Earth)* 110 (9), 5201–+.
- 895 Merkel, S., Tomé, C., Wenk, H.-R., 2009. A modeling analysis of the influence  
896 of plasticity on high pressure deformation of hcp-co. *Phys. Rev. B* 79,  
897 064110.
- 898 Merkel, S., Wenk, H., Gillet, P., Mao, H., Hemley, R. J., Jul. 2004. Deforma-  
899 tion of polycrystalline iron up to 30GPa and 1000K. *Physics of the Earth*  
900 *and Planetary Interiors* 145, 239–251.
- 901 Miyagi, L., Kunz, M., Knight, J., Nasiatka, J., Voltolini, M., Wenk, H., Nov.  
902 2008. In situ phase transformation and deformation of iron at high pressure  
903 and temperature. *Journal of Applied Physics* 104 (10), 103510–+.
- 904 Monnereau, M., Calvet, M., Margerin, L., Souriau, A., 2010. Lopsided growth  
905 of earth’s inner core. *Science* 328, 1014–1017.
- 906 Morelli, A., Dziewonski, A. M., Woodhouse, J. H., 1986. Anisotropy of the  
907 inner core inferred from PKIKP travel times. *Geophys. Res. Lett.* 13, 1545–  
908 1548.
- 909 Nimmo, F., 2007. *Treatise on Geophysics*. Vol. 8. Elsevier, Ch. Energetics of  
910 the Core, pp. 33–65.

- 911 Niu, F., Chen, Q., Oct. 2008. Seismic evidence for distinct anisotropy in the  
912 innermost inner core. *Nature Geoscience* 1, 692–696.
- 913 Niu, F., Wen, L., Apr. 2001. Hemispherical variations in seismic velocity at  
914 the top of the Earth’s inner core. *Nature* 410, 1081–1084.
- 915 Poirier, J. P., Price, G. D., 1999. Primary slip system of  $\epsilon$ -iron and anisotropy  
916 of the Earth’s inner core. *Phys. Earth Planet. Inter.* 110, 147–156.
- 917 Poupinet, G., Pillet, R., Souriau, A., 1983. possible heterogeneity of the  
918 earth’s core deduced from PKIKP travel-times. *Nature* 305 (5931), 204–  
919 206.
- 920 Sha, X., Cohen, R. E., 2010. Elastic isotropy of  $\epsilon$ -Fe under earth’s core con-  
921 ditions. *Geophys. Res. Lett.* 37, L10302.
- 922 Shimizu, H., Poirier, J.-P., Le Mouél, J.-L., 2005. On crystallization at the  
923 inner core boundary. *Phys. Earth Planet. Inter.* 151, 37–51.
- 924 Song, X., Helmberger, D. V., Jun. 1995. Depth dependence of anisotropy of  
925 Earth’s inner core. *J. Geophys. Res.* 100, 9805–9816.
- 926 Souriau, A., Jan. 2003. The seismological picture of the inner core: structure  
927 and rotation. *Comptes Rendus Geoscience* 335, 51–63.
- 928 Souriau, A., 2007. *Treatise on Geophysics*. Vol. 1. Elsevier, Ch. The Earth’s  
929 Core, pp. 655–693.
- 930 Souriau, A., Poupinet, G., Nov. 1991. The velocity profile at the base of the  
931 liquid core from PKP(BC+Cdiff) data: An argument in favor of radial  
932 inhomogeneity. *Geophys. Res. Lett.* 18, 2023–2026.



- 933 Souriau, A., Romanowicz, B., 1996. Anisotropy in inner core attenuation: a  
934 new type of data to constrain the nature of the solid core. *Geophys. Res.*  
935 *Lett.* 23, 1–4.
- 936 Souriau, A., Romanowicz, B., Apr. 1997. Anisotropy in the inner core: rela-  
937 tion between P-velocity and attenuation. *Physics of the Earth and Plane-*  
938 *tary Interiors* 101, 33–47.
- 939 Stacey, F., Davis, P., 2008. *Physics of the Earth*. Cambridge University Press,  
940 Cambridge UK.
- 941 Stacey, F., Loper, D., 2007. A revised estimate of the conductivity of iron  
942 alloy at high pressure and implications for the core energy balance. *Phys.*  
943 *Earth Planet. Inter.* 161, 13–18.
- 944 Stacey, F. D., Anderson, O. L., 2001. Electrical and thermal conductivities  
945 of Fe-Ni-Si alloy under core conditions. *Phys. Earth Planet. Inter.* 124,  
946 153–162.
- 947 Steinle-Neumann, G., Stixrude, L., Cohen, R. E., Gülseren, O., Sep. 2001.  
948 Elasticity of iron at the temperature of the Earth’s inner core. *Nature* 413,  
949 57–60.
- 950 Stixrude, L., Cohen, R. E., Mar. 1995. High-Pressure Elasticity of Iron and  
951 Anisotropy of Earth’s Inner Core. *Science* 267, 1972–1975.
- 952 Su, W., Dziewonski, A. M., Jeanloz, R., Dec. 1996. Planet Within a Planet:  
953 Rotation of the Inner Core of Earth. *Science* 274, 1883–1887.

- 954 Sumita, I., Olson, P., 1999. A Laboratory Model for Convection in Earth's  
955 Core Driven by a Thermally Heterogeneous Mantle. *Science* 286, 1547–  
956 1549.
- 957 Sumita, I., Olson, P., Aug. 2002. Rotating thermal convection experiments in  
958 a hemispherical shell with heterogeneous boundary heat flux: Implications  
959 for the Earth's core. *Journal of Geophysical Research (Solid Earth)* 107,  
960 2169–+.
- 961 Sumita, I., Yoshida, S., Kumazawa, M., Hamano, Y., 1996. A model for  
962 sedimentary compaction of a viscous media and its application to inner-  
963 core growth. *Geophys. J. Int.* 124, 302–324.
- 964 Tanaka, S., Hamaguchi, H., Feb. 1997. Degree one heterogeneity and  
965 hemispherical variation of anisotropy in the inner core from PKP(BC)-  
966 PKP(DF) times. *J. Geophys. Res.*102, 2925–2938.
- 967 Tome, C., Canova, G., Kocks, U., Christodoulou, N., Jonas, J., 1984. The  
968 relation between macroscopic and microscopic strain hardening in f.c.c.  
969 polycrystals. *Acta Metallurgica* 32 (10), 1637 – 1653.
- 970 Turcotte, D. L., Schubert, G., 2002. *Geodynamics*. pp. 472. ISBN  
971 0521661862. Cambridge, UK: Cambridge University Press.
- 972 Van Orman, J. A., 2004. On the viscosity and creep mechanism of Earth's  
973 inner core. *Geophysical Research Letters* 31, 20606–+.
- 974 Vočadlo, L., 2007. *Treatise on Geophysics*. Vol. 2. Elsevier, Ch. The Earth's  
975 Core: Iron and Iron Alloys, pp. 91–120.

- 976 Vočadlo, L., Alfè, D., Gillan, M. J., Wood, I., Brodholt, J., Price, G. D.,  
977 2003. Possible thermal and chemical stabilization of body-centred-cubic  
978 iron in the Earth's core. *Nature* 424, 536–539.
- 979 Vočadlo, L., Dobson, D. P., Wood, I. G., Nov. 2009. Ab initio calculations of  
980 the elasticity of hcp-Fe as a function of temperature at inner-core pressure.  
981 *Earth and Planetary Science Letters* 288, 534–538.
- 982 Weber, P., Machetel, P., Nov. 1992. Convection within the inner-core and  
983 thermal implications. *Geophys. Res. Lett.*19, 2107–2110.
- 984 Wenk, H., Baumgardner, J. R., Lebensohn, R. A., Tomé, C. N., Mar.  
985 2000. A convection model to explain anisotropy of the inner core. *J. Geo-*  
986 *phys. Res.*105, 5663–5678.
- 987 Wenk, H. R., 1999. A voyage through the deformed earth with the self-  
988 consistent model. *Model. Simul. Mater. Sci. Engin.* 7, 699–722.
- 989 Wenk, H. R., Matthies, S., Hemley, R. J., Mao, H. K., Shu, J., 2000. The  
990 plastic deformation of iron at pressures of the Earth's inner core. *Nature*  
991 405, 1044–1047.
- 992 Wenk, H. R., Takeshita, T., Jeanloz, R., Johnson, G. C., Jan. 1988. Develop-  
993 ment of texture and elastic anisotropy during deformation of hcp metals.  
994 *Geophys. Res. Lett.*15, 76–79.
- 995 Woodhouse, J. H., Giardini, D., Li, X.-D., 1986. Evidence for inner core  
996 anisotropy from free oscillations. *Geophys. Res. Lett.* 13, 1549–1552.

- 997 Worster, M. G., 1997. Convection in Mushy Layers. *Annual Review of Fluid*  
998 *Mechanics* 29, 91–122.
- 999 Yoshida, S., Sumita, I., Kumazawa, M., Dec. 1996. Growth model of the  
1000 inner core coupled with the outer core dynamics and the resulting elastic  
1001 anisotropy. *J. Geophys. Res.*101, 28085–28104.
- 1002 Yu, W., Wen, L., May 2006a. Inner core attenuation anisotropy. *Earth and*  
1003 *Planetary Science Letters* 245, 581–594.
- 1004 Yu, W., Wen, L., Jul. 2006b. Seismic velocity and attenuation structures in  
1005 the top 400 km of the Earth’s inner core along equatorial paths. *Journal*  
1006 *of Geophysical Research (Solid Earth)* 111, 7308–+.
- 1007 Zhang, K., Mar. 1992. Spiralling columnar convection in rapidly rotating  
1008 spherical fluid shells. *Journal of Fluid Mechanics* 236, 535–556.