Physics of the Earth and Planetary Interiors 243 (2015) 22–29

Contents lists available at ScienceDirect

Physics of the Earth and Planetary Interiors

journal homepage: www.elsevier.com/locate/pepi

The fluid dynamics of inner-core growth

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ARTICLE INFO

Article history:
Received 2 October 2014
Received in revised form 11 March 2015
Accepted 6 April 2015
Available online 10 April 2015

Keywords:
Geodynamo
Core processes
Thermal evolution
Inner-core anisotropy

ABSTRACT

Aspherical growth of the inner core has been suggested as a mechanism to produce seismic anisotropy through alignment of crystal lattices. This mechanism is viable if the response to aspherical growth occurs by slow viscous deformation. The inner core can also respond by melting and solidification at the boundary if flow in the liquid core can redistribute latent heat over the surface. We use a numerical geodynamo model to quantitatively assess the process of melting and solidification, and find that the response to aspherical growth occurs primarily through phase transitions when the viscosity of the inner core is 10^{21} Pa s or higher. A lower inner-core viscosity favors viscous adjustment, but the associated stresses may be too low to produce substantial crystal alignment. Independent of the primary relaxation mechanism, we expect a persistent and large-scale flow of the liquid core over the surface of the inner core. The predicted flow should be large enough to affect the crystal orientation of hcp-iron alloys during solidification, yet the absence of detectable seismic anisotropy in the top 60–80 km is suggestive. Either the mechanism of flow-induced alignment does not apply in the core or the intrinsic anisotropy of hcp iron at inner-core conditions is weak. Future seismological modeling using the predicted distribution of lattice preferred orientation might establish whether this texture is detectable with current observations.

1. Introduction

Cooling and solidification of the liquid iron core causes growth of the inner-core at a rate of roughly 10^{-3} m yr^{-1} (e.g. Nimmo, 2007). Spatial variations in the growth rate are expected when latent heat and chemical impurities are not removed uniformly from the interface. One factor that contributes to aspherical growth is due to the effects of rotation on convection in the liquid core. Heat transport is promoted in the equatorial region (Zhang, 1992; Sumita and Olson, 2000), allowing preferential solidification at equator. Departures from hydrostatic equilibrium drive slow viscous flow through the interior of the inner core, which can produce alignment of the crystallographic axes of solid iron (Yoshida et al., 1996). The combination of persistent aspherical growth and viscous adjustment offers an attractive explanation for the presence of seismic anisotropy in the inner core (see Deuss, 2014, for a recent review).

Deviations in the shape of the inner core also perturb temperature in the liquid core. Even small variations in temperature are capable of driving flow because of the viscosity of liquid iron is very low (de Wijs et al., 1998; Pozzo et al., 2013). When the flow is vigorous enough to redistribute latent heat over the surface of the inner core, the boundary can adjust by melting and solidification, rather than by slow viscous flow through the interior. One way to assess the relative importance of these two mechanisms is to estimate the timescales for relaxation toward a hydrostatic equilibrium. In one end-member case relaxation occurs solely by viscous flow in the interior of the inner core. The relevant timescale is set by the inner-core viscosity and the density jump across the inner-core boundary. In the other end-member case relaxation occurs through phase changes at the inner-core boundary. Here the relaxation time depends on the size of temperature anomalies associated with a non-hydrostatic shape and the strength of fluid motion that results from these anomalies.

In this study we use a numerical geodynamo model to investigate the fluid motion induced by thermal anomalies at the inner-core boundary. We quantify the transport of heat and assess the latent heat that must be removed or added to restore the boundary to its equilibrium position. The timescale for phase change is compared with the timescale for viscous relaxation (e.g. Cathles, 1975) to assess the primary adjustment mechanism. A low inner-core viscosity favors viscous adjustment, but the resulting stresses may be too small to cause alignment of iron crystals. On the other hand, a high viscosity, comparable to that proposed by Yoshida

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http://dx.doi.org/10.1016/j.pepi.2015.04.001
0031-9201/

Table 1
List of physical properties.

Property	Symbol	Value	Units
Bulk modulus (adiabatic)	K_S	1.3×10^{12}	Pa
Bulk modulus (isothermal)	K_T	1.2×10^{12}	Pa
Density Contrast at r_i	$\Delta\rho$	600	kg m^{-3}
Density of Fluid (average)	ρ	10^4	kg m^{-3}
Gravity at r_i	$g(r_i)$	4.4	m s^{-2}
Gravity at r_o	$g(r_o)$	10	m s^{-2}
Gruneisen parameter	γ	1.4	–
Latent heat	H	10^6	J kg^{-1}
Melting temperature	T_L	5500	K
Radius of inner core	r_i	1.22×10^6	m
Radius of outer core	r_o	3.48×10^6	m
Specific heat	C_p	800	J K^{-1} kg^{-1}
Thermal expansion	α	10^{-5}	K^{-1}

et al. (1996), favors adjustment by melting and solidification. Under these circumstance we expect a substantial reduction in preferential growth of the inner core and the associated development of elastic anisotropy would likely be suppressed.

2. Numerical geodynamo model

We use the geodynamo model *Calypso* (Matsui et al., 2014) to estimate the flow driven by a non-hydrostatic inner core. This flow is superimposed on a background convective flow, which maintains an internal magnetic field. We consider an incompressible and electrically conducting fluid in a spherical shell that rotates at constant angular velocity Ω. The inner, r_i, and outer, r_o, radii of the shell are chosen to have an Earth-like geometry (see Table 1). Convection is driven by an unstable temperature difference, ΔT, between the top and bottom boundaries. Allowing for small thermal anomalies on the bottom boundary modifies the convective flow to account for the influence of a non-hydrostatic inner core (see Section 3).

The equations for temperature, T, velocity, $\mathbf{V}$, and magnetic field $\mathbf{B}$ are written in non-dimensional form using $L = r_o - r_i$ as a characteristic length scale and L^2/ν as the characteristic timescale, where ν is the kinematic viscosity. Temperature and velocity are scaled by ΔT and ν/L, respectively, while the magnetic field is scaled by $\sqrt{\rho\mu\Omega\eta}$, where ρ is the fluid density, μ is the magnetic permeability and η is the magnetic diffusivity. The resulting governing equations are specified by four dimensionless parameters

$$E = \frac{\nu}{\Omega L^2}, \quad Pr = \frac{\nu}{\kappa}, \quad Pm = \frac{\nu}{\eta}, \quad Ra = \frac{\alpha g(r_o)\Delta T L}{\nu\Omega} \tag{1}$$

which include the Ekman number, E, the Prandtl number, Pr, the magnetic Prandtl number, Pm and a modified Rayleigh number, Ra. Here α is the coefficient of thermal expansion, $g(r)$ is the radially dependent gravity and κ is the thermal diffusivity. The mantle is assumed to be electrically insulating, whereas the inner core $(r < r_i)$ can be either insulating or electrically conducting with the same conductivity as the fluid. We find that the electrical properties of the inner core have only a small influence on the flow.

2.1. Thermal boundary conditions

Temperature anomalies δT on the bottom boundary are associated with radial displacements δr of the inner-core surface. This correspondence is defined in terms of the local geotherm, T_A, and the melting temperature, T_L (see Fig. 1). The boundary temperature is constrained to remain on the melting curve when the interface is shifted in radius. A positive δr gives a boundary temperature below the average value of T_A at the same depth. Conversely, a negative δr corresponds to a positive temperature anomaly. A quantitative relationship between δT and δr can be written as

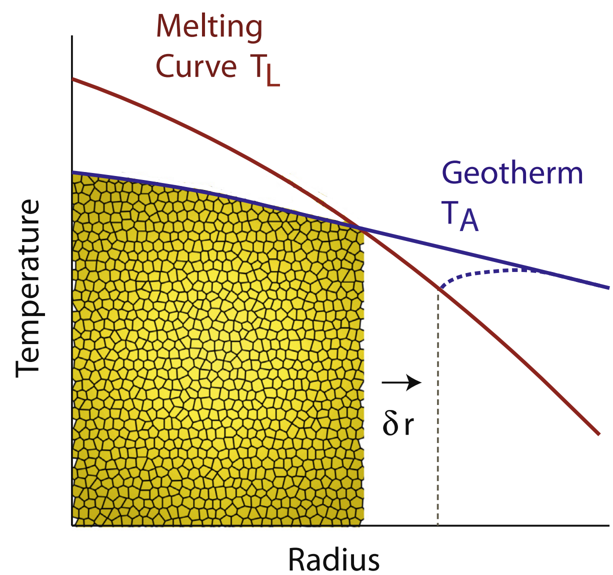

Fig. 1. Schematic illustration of thermal structure near the inner-core boundary. The intersection of the geotherm, T_A, and the melting temperature, T_L define the location of the inner-core boundary. When a radial displacement, δr, shifts the location of the boundary, the resulting melting temperature lies below the average geotherm. Consequently, a positive δr produces a negative temperature anomaly on the boundary. Conversely, a negative δr causes a positive temperature anomaly on the boundary.

$$\delta T = -\rho g(r_i)\left(\frac{dT_L}{dP} - \frac{dT_A}{dP}\right)\delta r \tag{2}$$

where the pressure derivative of T_L is based on Lindemann's law (e.g. Stacey and Davis, 2008)

$$\frac{dT_L}{dP} = \frac{2(\gamma - 1/3)T_L}{K_T} \tag{3}$$

and the pressure derivative of T_A is given by

$$\frac{dT_A}{dP} = \frac{\gamma T_L}{K_S} \tag{4}$$

under the assumption that T_A is an isentrope (e.g. Braginsky and Roberts, 1995). Other quantities in (3) and (4) include the Gruneisen parameter, γ, the isothermal and adiabatic bulk moduli, K_T and K_S, and the fluid density, ρ (see Table 1).

Temperature in an incompressible (Boussinesq) fluid can be viewed as a perturbation from an isentrope. Thus ΔT represents the temperature difference across the liquid core in excess of that predicted for the isentrope. As a result, the boundary conditions on T require

$$T = 0 \tag{5}$$

at the core-mantle boundary $r = r_o$ and

$$T = \Delta T + \delta T \tag{6}$$

at the inner-core boundary, $r = r_i$. We adopt a constant value for ΔT, specified by the choice of Ra, and consider several values for δT. We also assume that the temperature anomaly has a spatial dependence of the form

$$\delta T = \delta T_2 P_2(\cos\theta) \tag{7}$$

where $P_2(\cos\theta)$ is the Legendre polynomial of degree 2 and θ is colatitude. A similar dependence is assumed for the radial displacement