ARTICLE IN PRESS
EPSL:14511
Earth and Planetary Science Letters ••• (••••) •••–•••
Contents lists available at ScienceDirect
Earth and Planetary Science Letters
www.elsevier.com/locate/epsl

Formation, stratification, and mixing of the cores of Earth and Venus

Seth A. Jacobson [a,b,*], David C. Rubie [a], John Hernlund [c], Alessandro Morbidelli [b], Miki Nakajima [d]

[a] Universität Bayreuth, Bayerisches Geoinstitut, D-95440 Bayreuth, Germany
[b] Observatoire de la Côte d'Azur, Laboratoire Lagrange, Bd. de l'Observatoire, CS 34229, F-06304 Nice Cedex 4, France
[c] Tokyo Institute of Technology, Earth-Life Science Institute, 2-12-1-IE-1 Ookoyama, Meguro-ku, Tokyo, 152-8550, Japan
[d] Carnegie Institution of Washington, Department of Terrestrial Magnetism, 5241 Broad Branch Rd., NW, Washington, DC, 20015-1305, USA

ARTICLE INFO

Article history:
Received 6 January 2017
Received in revised form 5 May 2017
Accepted 9 June 2017
Available online xxxx
Editor: B. Buffett

Keywords:
planet formation
planetary differentiation
core formation
early Earth
early Venus
geodynamo

ABSTRACT

Earth possesses a persistent, internally-generated magnetic field, whereas no trace of a dynamo has been detected on Venus, at present or in the past, although a high surface temperature and recent resurfacing events may have removed paleomagnetic evidence. Whether or not a terrestrial body can sustain an internally generated magnetic field by convection inside its metallic fluid core is determined in part by its initial thermodynamic state and its compositional structure, both of which are in turn set by the processes of accretion and differentiation. Here we show that the cores of Earth- and Venus-like planets should grow with stable compositional stratification unless disturbed by late energetic impacts. They do so because higher abundances of light elements are incorporated into the liquid metal that sinks to form the core as the temperatures and pressures of metal-silicate equilibration increase during accretion. We model this process and determine that this establishes a stable stratification that resists convection and inhibits the onset of a geodynamo. However, if a late energetic impact occurs, it could mechanically stir the core creating a single homogenous region within which a long-lasting geodynamo would operate. While Earth's accretion has been punctuated by a late giant impact with likely enough energy to mix the core (e.g. the impact that formed the Moon), we hypothesize that the accretion of Venus is characterized by the absence of such energetic giant impacts and the preservation of its primordial stratifications.

1. Introduction

Earth's magnetic field is generated inside its convecting fluid outer core, and paleomagnetic evidence indicates that it has persisted since at least 4.2 Ga (Tarduno et al., 2015). Seismological probing of the core suggests that it consists mostly of iron and nickel with approximately 10 wt% light elements (i.e., an uncertain mixture of Si, O and S and potentially others such as H and C) (see Poirier, 1994, for review). Besides possible stratified layers at the very top (Helffrich and Kaneshima, 2010; Buffett, 2014) and bottom (Gubbins et al., 2008) of the outer core, the average structure is consistent with isentropic compression of a homogenous liquid (Hirose et al., 2013). Dynamical constraints suggest that the bulk of Earth's outer core is exceptionally well-mixed, exhibiting density fluctuations of order one part in a billion or less relative to an hydrostatic equilibrium profile (Mandea et al., 2012). However, it is not known how Earth's core achieved this high degree of homogeneity and whether such a high degree of homogeneity is expected for all terrestrial planets.

Terrestrial planets like Earth grow from a series of accretion events characterized by collisions with planetesimals and planetary embryos, most of which had cores of their own. In other words, Earth's core is not created in a single stage but from a series of core forming events (multistage core formation is reviewed in Rubie and Jacobson, 2016). A core formed over multiple stages is not in chemical equilibrium with the mantle since each core addition equilibrates with only part of the mantle (Deguen et al., 2011; Rubie et al., 2015). Moreover, the core is not necessarily chemically homogenous or isentropic at the end of planet formation. Only further processing within the core removes the signatures of multistage core formation and creates the practically homogenous core observed today.

In order to determine the chemical state of the core during and after planet formation, we linked a terrestrial planet formation model, a planetary differentiation model, and a core growth model together (Section 2). From these linked models, we obtained thermal and compositional profiles of the cores of Earth and Venus. We find that the memory of multistage core formation remains

* Corresponding author.
E-mail address: sethajacobson@gmail.com (S.A. Jacobson).

http://dx.doi.org/10.1016/j.epsl.2017.06.023

as a distinct compositional stratigraphy within the core. While convection may occur within certain layers, some boundaries between layers resist convection, require conductive heat transport, and create multiple convective cells within the core. However, we also determined that the density profile of the core has a strong dependence on the efficiency of impact driven core mixing (Section 3). If the impact energy from planetary accretion events is efficiently converted into turbulent mixing of the core, then the core is mechanically mixed and homogenized. Otherwise, the density structure is preserved within the core. As a consequence, a planet with this preserved stable stratification may not be able to produce an Earth-like geodynamo (Section 4). We hypothesize that such an internal structure is still present in Venus, whereas the core of Earth was sufficiently mixed by the Moon-forming impact (Section 5).

2. Establishing the structure of the core from accretion

In order to understand the growth of Earth's core, we used previously published simulations of the growth of Earth from the accumulation of planetesimals and planetary embryos out of the terrestrial protoplanetary disk (Jacobson and Morbidelli, 2014). These simulations are described in detail in the supplementary information. For clarity, we focus on the results of a well-studied simulation, 4:1-0.5-8, which is the same as that examined in Rubie et al. (2015, 2016). We passed the accretion histories of each planet to a planetary differentiation model, in which we calculated the chemical evolution of each planet's mantle and core as described in Rubie et al. (2011, 2015, 2016). This model uses data from high pressure laboratory experiments as well as a mass balance and element partitioning approach to calculate the composition of core forming liquids after each accretion event. Any equilibrated metal liquid continues sinking to the core due to the high density contrast between metal and silicate, while equilibrated silicate material is mixed with the rest of the mantle.

We calculated reference core density, mass, gravity, and pressure profiles using an iterative process. After every core addition, we constructed a two-layer planet model using a pair of Murnaghan equations of state for a silicate mantle and a metallic core. This reference density profile as a function of pressure P was fitted to the mantle and the liquid outer core of the preliminary reference Earth model (PREM; Dziewonski and Anderson, 1981):

$$\rho_{\mathrm{ref}}(r) = \begin{cases} 1669\,(18.89 + 5.517P(r))^{1/5.517} & \text{if } r > R_{\mathrm{CMB}} \\ 1438\,(195.7 + 3.358P(r))^{1/3.358} & \text{if } r \le R_{\mathrm{CMB}} \end{cases} \tag{1}$$

where the reference density ρ_{ref} is measured in kg m^{-3} and the pressure P is measured in GPa. Both the mass of the planet M and the mass of the core M_{C} are known from the planetary accretion model, so from the following equations, we determined the radius of the core R_{CMB} and the radius of the planet R.

$$M = 4\pi \int_{R_{\mathrm{CMB}}}^{R} \rho_{\mathrm{ref}}(r') r'^2 dr' + M_{\mathrm{C}} \tag{2}$$

$$M_{\mathrm{C}} = 4\pi \int_{0}^{R_{\mathrm{CMB}}} \rho_{\mathrm{ref}}(r') r'^2 dr' \tag{3}$$

Then we used the following equations to determine the gravitational acceleration and pressure profiles.

$$g(r) = \frac{4\pi G}{r^2} \int_{0}^{r} \rho_{\mathrm{ref}}(r') r'^2\, dr' \tag{4}$$

$$P(r) = \int_{r}^{R} \rho_{\mathrm{ref}}(r') g(r')\, dr' \tag{5}$$

This iterative procedure needed an initial guess, so we used an uncompressed ($P = 0$ GPa) density profile to initially calculate the core and surface radii given the core and planet mass. We iterated through the equations above until the relative difference between successive density, gravity and pressure profiles added in quadrature is less than 10^{-6}, which typically took about 10 iterations. The core growth model calculates perturbations to this reference model due to the varying thermal and compositional properties of each core addition.

2.1. Establishing the thermal structure of the core

As new core forming liquids sink through the mantle, they are heated by adiabatic compression and released gravitational potential energy. Immediately after equilibration, the metallic liquids have a temperature T_{eq}, which is approximately halfway between the peridotite solidus and liquidus at the metal-silicate equilibration pressure P_{eq}. As this material sinks to the core–mantle boundary, it is adiabatically compressed and so heats up to a temperature at the core–mantle boundary T_{CMB} of:

$$T_{\mathrm{CMB}} = T_{\mathrm{eq}} + \left.\frac{dT}{dP}\right|_{\mathrm{S}} (P_{\mathrm{CMB}} - P_{\mathrm{eq}}) \tag{6}$$

where P_{CMB} is the pressure at the core–mantle boundary and $dT/dP|_{\mathrm{S}} = 7.7$ K GPa^{-1} is the adiabatic temperature gradient for core fluids. Furthermore, gravitational potential energy is released as the denser core fluids sink through the less dense silicate mantle. If this heat is fully retained, then the temperature of the core addition when it reaches the core–mantle boundary is:

$$T_{\mathrm{CMB}} = T_{\mathrm{eq}} + \left.\frac{dT}{dP}\right|_{\mathrm{S}} (P_{\mathrm{CMB}} - P_{\mathrm{eq}}) + \frac{g_{\mathrm{eq}} r_{\mathrm{eq}} - g_{\mathrm{CMB}} R_{\mathrm{CMB}}}{4\pi c_P} \tag{7}$$

where $c_p = 825$ J kg^{-1} K^{-1} is the estimated specific heat capacity at constant pressure for core fluids, g_{eq} and g_{CMB} are the gravitational accelerations at the radius of equilibration r_{eq} and the core–mantle boundary R_{CMB}, respectively. As the core continues to grow, layers already within the core continue to adiabatically compress and increase in temperature:

$$T = T_{\mathrm{CMB}} + \left.\frac{dT}{dP}\right|_{\mathrm{S}} (P - P_{\mathrm{CMB}}) \tag{8}$$

where T is the temperature of the layer in the core at pressure P.

It is unclear how much of the released gravitational potential energy is retained within the sinking core addition as heat, so we examine this process in light of two end-member scenarios. In the high temperature end-member model corresponding to Eq. (7), all generated heat from adiabatic compression and sinking in the gravitational potential is retained within the newly formed layer of liquid metal. Alternatively, in the low temperature end-member model corresponding to Eq. (6), the new core addition is heated only by adiabatic compression; all of the released gravitational potential energy is assumed to be transported away in the silicate mantle. Reality likely lies between the low and high temperature end-member models, however both establish a nearly isothermal core structure (see Fig. 1(a) and (b)).

These two end-members would leave the mantle, particularly at the core–mantle boundary, in different thermal states. In the low temperature end-member model, the mantle would be very hot and thermal energy is unlikely to be vigorously transported across the core–mantle boundary, whereas for the high temperature end-member model, the mantle would be cooler and so thermal energy