Icarus 213 (2011) 12–23

Contents lists available at ScienceDirect

Icarus

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

Optimal dynamos in the cores of terrestrial exoplanets: Magnetic field generation and detectability

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ARTICLE INFO

Article history:
Received 29 April 2010
Revised 1 February 2011
Accepted 10 February 2011
Available online 17 February 2011

Keywords:
Extrasolar planets
Geophysics
Interiors
Magnetic fields
Radio observations

ABSTRACT

A potentially promising way to gain knowledge about the internal dynamics of extrasolar planets is by remote measurement of an intrinsic magnetic field. Strong planetary magnetic fields, maintained by internal dynamo action in an electrically conducting fluid layer, are helpful for shielding the upper atmosphere from stellar wind induced mass loss and retaining water over long (Gyr) time scales. Here we present a whole planet dynamo model that consists of three main components: an internal structure model with composition and layers similar to the Earth, an optimal mantle convection model that is designed to maximize the heat flow available to drive convective dynamo action in the core, and a scaling law to estimate the magnetic field intensity at the surface of a terrestrial exoplanet. We find that the magnetic field intensity at the core surface can be up to twice the present-day geomagnetic field intensity, while the magnetic moment varies by a factor of 20 over the models considered. Assuming electron cyclotron emission is produced from the interaction between the stellar wind and the exoplanet magnetic field we estimate the cyclotron frequencies around the ionospheric cutoff at 10 MHz with emission fluxes in the range 10^{-4}–10^{-7} Jy, below the current detection threshold of radio telescopes. However, we propose that anomalous boosts and modulations to the magnetic field intensity and cyclotron emission may allow for their detection in the future.

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1. Introduction

There is evidence of a large scale magnetic field maintained by internal dynamo action for every planet in the Solar System with the possible exception of Venus. In addition, magnetic fields have been measured on several satellites, but only Ganymede's field is likely to be generated by internal dynamo action. Detection of terrestrial exoplanet dynamos would provide important constraints on internal structures, dynamics, energetics, and the ubiquity of planetary magnetic fields in general. Dynamo action is maintained by efficient heat transfer from the deep interior so that planets with strong magnetic fields may also imply mobile-lid mantle convection. Furthermore, maintaining a habitable surface over long time scales may require a magnetic field to shield the atmosphere from mass loss, retain large amounts of water, and protect the surface from charged particles (Dehant et al., 2007; Lammer et al., 2007). Therefore, the search for terrestrial exoplanet magnetic fields is a critical component of the search for habitable planets.

Without a large scale planetary magnetic field, charged particles interact directly with the upper atmosphere and accelerate atmospheric mass loss. More frequent stellar flares and increased stellar wind flux associated with an active young star amplify these effects during the early stages of planetary evolution. Earth's strong magnetic field shields the atmosphere, promoting the retention of large amounts of water over geologic time scales. This may not have been the case for the other terrestrial planets. For example the measured D/H ratio in the venusian atmosphere indicates that it had more water in the past (Lammer et al., 2008), consistent with the absence of a strong magnetic field. Also, since maintenance of a strong magnetic field in a terrestrial planet likely requires large scale mantle convection, planets with strong magnetic fields may also maintain mobile-lid surface tectonics. Evidence for this coupling exists for both Venus and Mars, where some form of active surface tectonics may have ceased in conjunction with the extinction of dynamo action in their core and loss of water from their surfaces (Nimmo, 2002; Stevenson, 2001).

More than 400 extrasolar planets have been detected to date, with 20 planets less massive than $10M_E$ ($M_E = 1$ Earth-mass) (Schneider, 2010). Although the diversity among extrasolar planets has been surprising (Fischer, 2008; Howard et al., 2010), the key ingredients to sustaining a dynamo, an energy source (i.e. convection), rotation, and a large volume of electrically conducting fluid, are thought to be common planetary phenomenon. Numerical simulations indicate that planets in the 1–$10M_E$ regime with an

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0019-1035/$ - see front matter © 2011 Elsevier Inc. All rights reserved.
doi:10.1016/j.icarus.2011.02.010

Earth-like (terrestrial) composition that harbor large, mostly iron cores form readily within 3 AU of their host star (Laughlin et al., 2004; Ida and Lin, 2004), and are often referred to as "super-Earths". In this paper we explore the possibility of detecting the magnetic field of such a planet.

The magnetic planets in the Solar System emit intense electron cyclotron radiation at radio frequencies (1–100 MHz), which is generated by energetic solar wind electrons interacting with the planetary magnetic field. Cyclotron emission is modulated at the rotation period of the planet if the magnetic field contains non-axisymmetric components and has been used to estimate the rotation periods of the giant planets (e.g. Anderson and Schubert, 2007; Zarka et al., 2001). Massive extrasolar planets with strong magnetic fields are expected to produce detectable cyclotron emission at radio frequencies (Zarka, 2007), but little attention has been paid to the potential emission from terrestrial exoplanets. Several other techniques have been proposed to infer the presense of exoplanet magnetic fields, and evidence of the interaction between the magnetic field of the star and planet has been claimed in about 10 (e.g. Shkolnik et al., 2008).

Our main goal is to explore the detectability of low mass, terrestrial exoplanet magnetic fields. This requires a number of important assumptions because the likely internal structure and thermal evolution of terrestrial exoplanets is unknown. The first models of radial variation of density in planets with end member compositions were constructed in the pioneering work of Zapolsky and Salpeter (1969). More recently preliminary internal structure models of super-Earths have been constructed to obtain simple scaling laws for the planetary and core radius as a function of planetary mass (e.g. Valencia et al., 2006; Seager et al., 2007; Sotin et al., 2007). These models either ignore the thermal state of the mantle and core and phase transitions therein or assume a core–mantle boundary (CMB) heat flux proportional to planet mass. Exoplanets may have important differences from the terrestrial planets in the Solar System, therefore we focus on optimizing a potentially observable feature of terrestrial exoplanets: their magnetic field. The most promising targets are those with extremely intense magnetic fields, likely maintained by efficient heat transfer out of the core. We compute internal structure profiles and develop a thermal convection model that maximizes core heat flow, which we refer to as the optimal state for dynamo action.

Before introducing the specifics of the model it is helpful to list the main assumptions and idealizations of the optimal model. We assume an Earth-like composition, structure, surface temperature and pressure, and rotation rate. We assume the mantle and core are convecting rigorously, and that there is no large scale melting in the mantle. We propose that as the planet cools it may pass through an optimal state for dynamo action before evolving into a sub-optimal state, perhaps similar to the present-day Earth.

We describe the internal structure model in Section 2, the optimal thermal model in Section 3, and magnetic field generation in Section 4. The main results are presented in Section 5, and considerations of magnetic field detectability are in Section 6. Finally, we summarize our main conclusions and discuss the prospects for detecting an exoplanet magnetic field in the near future in Section 7.

2. Internal structure model

The internal structure modeling technique employed here is very similar to those of Valencia et al. (2006), Sotin et al. (2007) and Seager et al. (2007). The following set of equations are solved in a spherical shell of thickness dr and are then integrated over the full radius of the planet R subject to boundary conditions. The continuity equation describing the change in mass $m(r)$ within radius r, Poisson's equation for gravity g, the hydrostatic equation for pressure P, and the Adams–Williamson equation for density ρ are,

$$dm(r)/dr = 4\pi r^2 \rho(r) \tag{1}$$

$$dg(r)/dr = 4\pi G\rho(r) - 2Gm(r)/r^3 \tag{2}$$

$$dP(r)/dr = -\rho(r)g(r) \tag{3}$$

$$d\rho(r)/dr = -\rho^2(r)g(r)/K_S(r) \tag{4}$$

where $K_S(r) = \rho(\partial P/\partial\rho)_S$ is the isentropic bulk modulus and G is the gravitational constant. We write K_S in terms of the isothermal bulk modulus K_T as,

$$K_S(r) = K_T(r)[1 + \alpha(r)\gamma(r)T(r)] \tag{5}$$

where α is thermal expansivity, γ is the Gruneisen parameter, and T temperature. The equation of state (EOS) we use to relate K_T to ρ is the third order Vinet EOS (Oganov, 2007; Vinet et al., 1989),

$$K_T = K_0 x^{-2/3}[1 + (1 + \theta x^{1/3})(1 - x^{1/3})]\exp[\theta(1 - x^{1/3})] \tag{6}$$

where the zero subscript refers to the zero pressure value of a quantity, $x(r) = \rho(r)/\rho_0$, and $\theta = 3/2(K_0' - 1)$, where K_0' is the zero pressure derivative of K_T. The adiabatic temperature gradient,

$$dT_{ad}(r)/dr = -\rho(r)g(r)\gamma(r)T(r)/K_S(r) \tag{7}$$

describes the increase in temperature with depth in a well-mixed layer. The depth-dependence of the remaining thermodynamic parameters γ and α are parameterized by

$$\gamma(r) = \gamma_0(x(r))^{-\gamma_1}, \quad \alpha(r) = \alpha_0(x(r))^{-3} \tag{8}$$

where γ_0, γ_1, and α_0 are constant within each compositional layer (Table 1). The variation of α with density in (8) is based on high pressure experiments (Chopelas and Boehler, 1992; Merkel et al., 2000).

We impose surface conditions on each model of $P_0 = 1$ atm, $\rho_0 = 3226$ kg m^{-3}, and $T_0 = 300$ K. Conditions at the center ($r = 0$) of each model require that the mass and gravity go to zero and the other variables (e.g. ρ and T) remain smooth and finite. The internal structure Eqs. (1)–(8) are integrated from the surface inwards and the surface radius R is modified until the conditions are satisfied at the center, with a typical error in R that corresponds to about one part in 10^4.

2.1. Layers

We include up to five layers in the model: a peridotite upper mantle, a perovskite mid-mantle, a post-perovskite lower mantle, and a solid or liquid metallic core. We do not include a spinel structure as in the transition zone of the Earth's mantle, because this layer is less than 300 km thick. There are four possible transitions or discontinuities: peridotite to perovskite, perovskite to post-perovskite in the mantle, a core–mantle boundary (CMB) where the material changes from silicates to iron, and an iron solidus boundary denoted the inner core boundary (ICB). The pressure at which the olivine transitions to perovskite (in the spinel structure) is a function of temperature described by Ito and Takahashi (1989)

$$P(T) = P_{pd0} - \gamma_{pd}T \tag{9}$$

where the reference pressure is $P_{pd0} = 28.3$ GPa and the Clapeyron slope is $\gamma_{pd} = 2.8$ MPa K^{-1}. The pressure at which perovskite transforms to the higher density post-perovskite phase is described by

$$P(T) = P_{ppv0} + \gamma_{ppv}(T - T_{ppv0}) \tag{10}$$

where the reference pressure and temperature are $P_{ppv0} = 124$ GPa and $T_{ppv0} = 2500$ K, and the Clapeyron slope is $\gamma_{ppv} = 8$ MPa K^{-1} (Hernlund and Labrosse, 2007).

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