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Density structure of the mantle transition zone and the dynamic geoid

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ABSTRACT

Joint inversion of the observed geoid and seismic velocities has been commonly used to constrain mantle properties and convection flow. However, most of the employed tomography models have been obtained without considering the effect of the mantle transition zone (TZ). We use the new-generation tomography model of Gu et al. (2003), in which seismic velocity perturbations are estimated together with variations of the main TZ boundaries. In the inversion of this model with the observed geoid the velocity-to-density scaling factor and density jump at these discontinuities are determined simultaneously. By this, the mantle flow across TZ is defined self-consistently: the undulations of the TZ boundaries suppress or accelerate mantle currents depending on the determined density contrast. For the 410-km discontinuity we obtain the scaling factor and density jump, which are close to mineral physics predictions. Therefore, we conclude that these effects are decoupled in the tomography model. In contrast, the calculated density jump at the 660 discontinuity is approximately 4 times less than the PREM value. We suggest that this is the effect of multiple phase transitions within a depth range of 640–720 km. Under normal thermal conditions, the post-spinel phase transformation is relatively sharp (~5 km) and clearly visible in seismic models. On the other hand, the transition of majorite garnet to perovskite is much broader (up to ~50 km) and, therefore hardly detectable. Due to the different sign of the Clapeyron slope, the total gravity effect is drastically decreased. To fit the obtained results, the required value of the Clapeyron slope for the majorite garnet to perovskite transformation should be equal to about +1.7 MPa/K. In the cold zones the same effect might be produced by the transition from ilmenite to perovskite at 610–640 km depth, which is in agreement with multiple reflections revealed in regional seismic studies near the bottom of TZ (Deuss et al., 2006). The estimated amplitude of the mantle flow across TZ is about ±20 mm/year, which corresponds to the whole-mantle convection scheme. The calculated geoid better fits to the observed one than the obtained without considering the TZ effect.

1. Introduction

The observed geoid is one of the most important constraints on mantle parameters, first of all, density distribution and viscosity variations. However, determination of the Earth's mantle structure has an ambiguous solution when using only surface gravity data. A usual way to cope with such a problem is to combine gravity data with other geophysical data sets to produce the solution, which fits all the data sets and therefore contains fewer degrees of freedom. Seismic tomography models are used for these purposes most often. Since only several parameters (usually depth-dependent velocity-to-density scaling factors and radial viscosity variations) are determined from a large volume of the data, the solution is relatively stable. In global studies this technique is applied starting from the pioneer works by Hager and O'Connell (1981), Ricard et al. (1984), Richards and Hager (1984), Forte and Peltier (1987), King and Masters (1992), Corrieu et al. (1994) and many others.

Despite of the massive efforts, the obtained results are remarkably various and a generalized instantaneous model of the dynamic mantle does not exist at the moment. There might be several explanations for such an indefinite situation, but one of the principal factors is that seismic tomography models, which were analysed in the previous studies, do not include information about the mantle transition zone discontinuities. Up to now, most of the combined gravity-seismic models are based on the tomography data, in which the effects of velocity variations and phase boundaries are mixed (e.g. Forte and Perry, 2000; Steinberger and Calderwood, 2006 and many others). As it is shown in the following chapter, the effect of the phase boundary on seismic velocities depends on many factors, which are not well-defined. The phase boundary might even lead to strong artificial velocity in the vicinity of the phase transition. Inclusion of the phase boundaries in the model significantly influence the convection pattern and by this – the dynamic geoid. Defraigne and Wahr (1991) have developed a numerical technique

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to incorporate the effect of phase-boundaries into internal-loading dynamic models. It has been demonstrated that the effect of the phase boundaries on the dynamic geoid might be very significant (Defraigne et al., 1996). Moreover, these authors have found that using a density jump for the 660 discontinuity, which is prescribed in PREM (Dziewonski and Anderson, 1981) would lead to much larger impact to the observed geoid than it might be expected from the observed one. In this paper we use the new-generation tomography model of Gu et al. (2003), in which mantle velocities have been estimated in a joint inversion with the mantle transition zone (TZ) discontinuities (410 and 660 km). Employing joint inversion of all the parameters (seismic velocity anomalies, topography of the TZ discontinuities and observed geoid) we try to understand density structure of the TZ and to determine its impact to the dynamic geoid.

A role of the TZ in shaping mantle convection has been discussed for decades. However, up to now the problem of the convection style (a rate of the mass exchange across the TZ) is not resolved completely. In the global studies, dealing with a joint inversion of the geoid and seismic tomography, the style of mantle convection was usually predefined. Most previous works starting from (Hager and O'Connell, 1981) employing the whole-mantle model because this model generally provides better fit to the observed geoid (e.g. Corrieu et al., 1994; Thoraval and Richards, 1997). However, the layered model has been analysed in several studies (Wen and Anderson, 1997; Forte and Woodward, 1997; Cadek et al., 1997; Pari and Peltier, 1998). It has been demonstrated that this model can also provide a reasonable fit to the observed geoid and even better explanation for the dynamic topography. Čadek and Fleitout (1999) have developed a combined approach, according to which the whole-mantle and layered convection models are mixed in various proportions. Despite this model is artificial and simply depicts a fixed rate of the 660 permeability, which is the same over the whole Earth for all flow patterns, it may help improving the fit to the observed geoid. Steinberger (2007) used another approach to study the effect of the phase boundaries on the geoid; he estimated the topography implied due to temperature variations inferred from seismic velocity variations and Clapeyron slope. The main problem for most of the above models is that the calculated topography of the 660 discontinuity is remarkably different from observations (e.g. Čadek and Fleitout, 1999; Steinberger and Torsvik, 2007; Steinberger, 2007).

Therefore, implementation of seismically determined variations of the main TZ boundaries into geodynamic models is likely the most promising way to overcome the existing problems. Unfortunately existing seismic models of these boundaries are remarkably different (Steinberger, 2007). In contrast to previous models, the TZ topography model of Gu et al. (2003) has been obtained in a joint inversion with the tomography model. Therefore it should be more consistent. These results open innovative possibilities to investigate the role of the TZ boundaries in mantle dynamics. In this case we do not have to identify the convection style explicitly. We apply a joint inversion of the seismic velocities and 410, 660 topography with the observed geoid to determine simultaneously the velocity-to-density scaling factor and density jump at the TZ discontinuities. After construction of the mantle density model, the TZ boundaries should control the convection style self-consistently suppressing or intensifying mantle flow through the boundary.

2. Velocity-to-density scaling in the vicinity of a phase boundary

The mantle transition zone is extensively studied by seismic methods. Usually two techniques are employed for these purposes. Study of precursors to the SS and PP waves are widely used to determine global variations of the TZ boundaries (e.g. Shearer, 2000; Deuss, 2009). The 660 discontinuity is the most important for a global dynamic modelling. However, existing studies based on available date often provided controversial results, which could not be explained using a simple mineralogical model of 660 discontinuity as described in PREM (Deuss et al., 2006; Deuss, 2009). Normally, the 660 discontinuity is globally detected by the long-period S660S precursors, but the long-term P660P precursors are available only for some places (e.g. Estabrook and Kind, 1996; Shearer and Flanagan, 1999). On the other hand, the short-period observations show that the 660-km discontinuity is sharp and prominent in some isolated places, where it is not visible in the long-period PP precursors (Benz and Vidale, 1993). Furthermore, studies of the converted waves often provide the results, which are different from the precursors studies (e.g. Vinnik et al., 1997).

All together, these results evidence that the 660 discontinuity is characterized by a complicated structure and is often represented by several boundaries in a depth range from 640 km to 720 km. It has been suggested that this structure might be related by multiple phase transitions in this depth range (e.g. Deuss et al., 2006). In addition to the commonly accepted transition from ringwoodite to perovskite and magnesiowüstite there should be additional boundaries.

These features would inevitably influence the mantle flow and dynamic geoid. However, up to now most studies of the global dynamic model of the Earth were based on standard velocity models, which are constructed using body and surface waves tomography methods. It is usually assumed that these models also reflect principal features of the mantle transition zone providing artificial seismic velocity anomalies, which might be properly scaled to density anomalies used in the dynamic modelling. One of the principal parameters, determined in the inversion, is a depth dependant velocity-to-density scaling factor (SF), which defines 3D density structure of the mantle based on seismic models. Let us determine first the "effective" scaling factor (s_{effect}) for the traditional seismic models, in which the effects of in situ velocity variations and TZ topography are mixed.

We assume that temperature variations (ΔT) near any single phase boundary control variations of the material properties: seismic velocity (ΔV), density ($\Delta\rho$) and topography of the TZ (Δh):

$$\frac{\Delta\rho}{\rho} = \alpha\Delta T, \quad \frac{\Delta V}{V} = \frac{\alpha}{s}\Delta T, \quad \Delta h = \gamma_h \Delta T, \tag{1}$$

where α is the thermal expansion coefficient, $s = \partial\ \ln(\rho)/\partial\ \ln(V)$ is the in situ velocity-to-density scaling factor (SF) and $\gamma_h = dh/dT = -\rho^{-1}g^{-1}dp/dT$ the Clapeyron slope for depth. The resolution of seismic tomography models is always limited; therefore it provides velocities, which characterize some depth range (Z) depending on the parameterization, damping and data coverage. Normally it is about 100–150 km at the depths 400–700 km (e.g. Gu et al., 2003). As a result, the variations of seismic velocity in the vicinity of a phase boundary represent a cumulative effect of in situ velocity changes and variations of the boundary dividing layers with basically different velocities. This effect might be described with sufficient accuracy by the simple equation:

$$\frac{\Delta V_{\text{effect}}}{V} = \frac{\alpha}{s}\Delta T + \frac{\Delta h}{VZ}\Delta V_{tz} = \frac{\alpha}{s}\Delta T + \frac{\gamma_h \Delta T}{VZ}\Delta V_{tz}, \tag{2}$$

where $\Delta V_{tz} = V_2 - V_1$ (Fig. 1). Therefore, the average velocity variations near TZ depend on many factors, which could be variable and not defined precisely. In the case of the negative Clapeyron slope, the effective velocity variations might be of different sign for the same temperature anomaly or close to zero. The situation is even more complicated when seismic velocities and TZ boundary variations are not only controlled by temperature variations but also by other factors (composition, water content, etc.).