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Tomography-based mantle flow beneath Mongolia-Baikal area

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ABSTRACT

Recent progress in seismic tomography of Asia allows us to explore and understand more clearly the mantle flow below the Mongolia-Baikal area. We present a tomography-based model of mantle convection that provides a good match to the residual topography. The model provides predictions on the present-day mantle flow and flow-induced asthenospheric deformation which give us new insights on the mantle dynamics in the Mongolia-Baikal area. The predicted mantle flow takes on a very similar pattern at the depths shallower or deeper than 400 km and almost opposite flow directions between the upper (shallower than 400 km) and lower (deeper than 400 km) parts. The flow pattern could be divided into the 'simple' eastern region and the 'complex' western region in the Mongolia. The upwelling originating from about 350 km depth beneath Baikal rift zone is an important possible drive force to the rifting. The seismic anisotropy cannot be simply related with asthenospheric flow and flow-induced deformation in the entire Mongolia-Baikal area, but they could be considered as an important contributor to the seismic anisotropy in the eastern region of Mongolia and around and in Sayan-Baikal orogenic belt.

1. Introduction

The mantle flow beneath the Mongolia-Baikal area has been mainly inferred from the upper mantle heterogeneity, SKS splitting, azimuthal anisotropy and radial anisotropy (e.g., Barruol et al., 2008; Lebedev et al., 2006; Dricker et al., 2002; Gao et al., 1997, 1994; Vinnik et al., 1992; Silver and Chan, 1991). These inferred mantle flow maps are very rough and probably different from each other. For example, Lebedev et al. (2006) inferred that the asthenospheric flow ascends from the 200 km depth beneath the Siberian craton and flows horizontally from NW to SE towards the shallower depths beneath the rift in order to explain the origin of volcanism in the Baikal rift zone. While Petit et al. (1998) surmised that a narrow and hot mantle plume beneath the Siberian craton originates from the 670 km phase boundary and reaches the bottom of the resistant Siberian plate and follows its borders in the Baikal area, which results in a NW mantle flow in the Siberian craton. Therefore, in order to reveal the mantle dynamic process under the Mongolia-Baikal area more clearly, mantle flow models are necessary. So far, the amount of this work has been remained relatively limited. Liu (1978) and Huang and Fu (1983), using a set of satellite geo-potential harmonics and the mantle flow model proposed by Runcorn (1967), obtained the sublithospheric mantle flow patterns within China and its adjacent area. Taking regional isostatic gravity anomalies as constraints on a regional-scale upper mantle convection model, Xiong et al. (2010) obtained the sublithospheric mantle flow velocity and flow-induced stress fields in the Mongolia-Baikal area. However, there were obvious differences between the flow directions under the Mongolia presented by Liu (1978) and Huang and Fu (1983). In addition, mantle viscosity was assumed constant in the flow model of Liu (1978), Huang and Fu (1983) and Xiong et al. (2010), which is not realistic because mantle viscosity has not only strong depth-dependence but also lateral variations by 2–4 orders in the upper and lower mantle and 1–2 orders in the middle mantle (e.g., Zhong et al., 2000; Ranalli, 2001; Mitrovica and Forte, 2004; Liu and Stegman, 2011). Mantle viscosity has very important effects on mantle flow patterns and thermal state (e.g., Christensen, 1984; Christensen and Harder, 1991; Tackley, 2000; Zhong et al., 2000), and then on geoid, gravity anomaly and topography (e.g., Mckenzie, 1977; Richards and Hager, 1989; Koch and Ribe, 1989; Zhang and Christensen, 1993; Ye and Wang, 2003). For instance, gravity and topographic anomalies can be positive or negative over mantle upwellings depending on viscosity structure (Mckenzie et al., 1977). If mantle viscosity does not change with depth, the effects of lateral viscosity variations (LVVs) on geoid are very small and could even be neglected. If mantle viscosity is depth-dependent, there are significant effects on the higher order (>3)

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modes of geoid (Zhang and Christensen, 1993). Both the uncertain flow patterns and dynamic properties of the deep Earth motivate us to explore more reasonable mantle flow beneath the Mongolia-Baikal area.

To date, some depth-dependent viscosity profiles (e.g., Ricard et al., 1993; Forte and Mitrovica, 1996; King and Masters, 1992; Lambeck et al., 1996; Ricard and Bai, 1991; Mitrovica and Forte, 2004) have been presented and some seismic tomography models (e.g., Simmons et al., 2009; Koulakov, 2011) have approached the horizontal resolution needed to address the regional mantle dynamics in the Mongolia-Baikal area, which allows us to set up an updated model of mantle convection. In this study, our first challenge is to present and elucidate the tomography-based mantle flow and its geodynamic implications in the Mongolia-Baikal area and the subsequent challenge is to elucidate the relationship between asthenospheric flow and seismic anisotropy.

2. Model setup

In order to determine the mantle flow field below the Mongolian-Baikal area, we use the 3D spherical finite element code for mantle convection, CitcomS (Zhong et al., 2000). The code solves for an incompressible Newtonian fluid within a spherical mantle shell. Table 1 lists the basic model parameters used in this study.

2.1. Mesh parameters

Our study area covers the range of 14° (41°–55°N) and 33° (87°–120°E) in latitude × longitude (Fig. 1). To avoid artificial return flow from side walls, we have chosen a wide enough box, with the nearest vertical boundary being >1000 km away from any part of our study area except for its northern boundary due to the limitation of the domain of the seismic model used here. Our regional model adopts a mesh with 129 × 193 × 65 nodes in latitude × longitude × depth, covering a physical domain of 30° (25°–55°N) × 60° (70°–130°E) × 670 km, respectively. The mesh is largely centered on our study area with mean grid spacing in all three dimensions.

2.2. Rheology

We use both depth and temperature dependent viscosity (Eq. (1)).

$$\eta(r,T) = \eta(r)\exp\left(\frac{E}{T+T_0} - \frac{E}{1+T_0}\right) \tag{1}$$

Where, $\eta(r)$ is the depth-dependent effective viscosity inferred from joint inversions of global convection-related observables and glacial isostatic adjustment data associated with the response of the Earth to melting of the Laurentide and Fennoscandian ice loads (Mitrovica and Forte, 2004). In order to understand the viscosity-dependence of our model, we also test other five depth-dependent viscosity profiles (Supplementary Fig. S2) proposed by Richard et al. (1993), Forte and Mitrovica (1996), King and Master (1992), Lambeck et al. (1996) and Richard and Bai (1991). The results (Supplementary Fig. S3–S6) show that the mantle flow patterns at the depths shallower than 400 km beneath Mongolia-Baikal area have less viscosity-dependent (see the Supplementary Material in detail). E is activation energy and assumed 120 kJ/mol (Watts and Zhong, 2000). T is temperature split up into laterally averaged part T_r which only depends on r and the lateral temperature variation (LTV) δT. T_0 is surface temperature of 273 K. The presence of LTV δT (derived from seismic structure in the present paper) leads to large amplitude lateral viscosity variations (LVV) which will be superimposed on the depth-dependent effective viscosity.

Table 1
Summary of model parameters.

Parameter	Value
Gravitational acceleration	9.81 m s^{-2}
Reference mantle density	3300 kg m^{-3}
Reference viscosity	10^{21} Pa s
Thermal diffusivity	10^{-6} m^2 s^{-1}
Temperature change across the upper mantle	1600 K

2.3. Density perturbations and lateral temperature variation

Regardless of many available seismic models (e.g., Ritzwoller et al., 2002; Huang and Zhao, 2006; Yakovlev et al., 2007; Simmons et al., 2009; Li and van der Hilst, 2010; Koulakov and Bushenkova, 2010; Koulakov, 2011; Li et al., 2013), only two ones are our candidates here. One is a global model, called TX2008, which was constructed via a joint inversion of global seismic and geodynamic data sets in which mineral physical constraints on the thermal dependence of seismic wave velocities and density were explicitly incorporated (Simmons et al., 2009), the other is a regional seismic model of P and S anomalies in the upper mantle beneath Asia, here called IVAN2011, which was constructed based on the tomographic inversion of travel time data from the revised ISC catalog for the years 1964–2004 (Koulakov, 2011). Comparing the mantle structure below the Mongolia-Baikal area (Fig. 2), we found that the spatial resolution of model IVAN2011 is much higher than that of model TX2008, so model IVAN2011 was adopted to derive density (and corresponding temperature) perturbations within the mantle in this study. Model IVAN2011 could produce the seismic P and S velocity anomalies (Fig. 2). The results of checkerboard tests presented by Koulakov (2011) show that the spatial resolution for the P tomographic model is at least 1° higher than that for the S model, so we ultimately decided to employ the seismic P wave velocity anomalies to drive mantle density anomalies.

The density-velocity conversion factor used here is determined by analyzing the standard deviation of the data set of the residual topography subtracting our predicted dynamic topography and the correlation between our predicted dynamic topography and the residual topography. If the maximum correlation coefficient and the smallest standard deviation are achieved in the meantime, the corresponding velocity-to-density conversion factor is adopted in the present paper. Here the residual topography is computed by subtracting the topography isostatically compensated in the crust from actual topography (Fig. 3a). The CRUST1.0 crustal model (available on line at http://igppweb.ucsd.edu/~gabi/crust1.html) used. Other details of the computation were described by Steinberger et al. (2001).

The maximum correlation coefficient of 0.44 and the smallest standard deviation of 0.41 km are achieved at the velocity-to-density conversion factor of 0.3 (Fig. 4), and the corresponding model predicted dynamic topography is shown as Fig. 3b, so 0.3 is adopted to derive the thermally-induced density perturbations $\delta\rho$ in the mantle. Thus we can approximate the lateral temperature variations δT required to generate the derived thermal density anomalies using the standard relationship:

$$\delta T = -\frac{\delta\rho}{\alpha(r)} \tag{2}$$

where $\alpha(r)$ is the thermal expansivity which was proved to be a decrease with depth (Chopelas and Boehler, 1989; Calderwood, 1999). In the present paper, we take the thermal expansivity profile of Calderwood (1999).