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Effects on the long-wavelength geoid anomaly of lateral viscosity variations caused by stiff subducting slabs, weak plate margins and lower mantle rheology

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ABSTRACT

Instantaneous flow numerical calculations in a three-dimensional spherical shell are employed to investigate the effects of lateral viscosity variations (LVVs) in the lithosphere and mantle on the long-wavelength geoid anomaly. The density anomaly model employed is a combination of seismic tomography and subducting slab models based on seismicity. The global strain-rate model is used to represent weak (low-viscosity) plate margins in the lithosphere. LVVs in the mantle are represented on the basis of the relation between seismic velocity and temperature (i.e., temperature-dependent rheology). When highly viscous slabs in the upper mantle are considered, the observed positive geoid anomaly over subduction zones can be accounted for only when the viscosity contrast between the reference upper mantle and the lower mantle is approximately 10^3 or lower, and weak plate margins are imposed on the lithosphere. LVVs in the lower mantle exert a large influence on the geoid pattern. The calculated geoid anomalies over subduction zones exhibit generally positive patterns with quite high amplitudes compared with observations, even when the low activation enthalpy of perovskite in the lower mantle is employed. Inferred weak slabs in the lower mantle may be explained in terms of recent mineral physics results, highlighting the possibility of grain-size reduction due to the postspinel phase transition.

1. Introduction

The geoid anomaly observed on the Earth's surface (Fig. 1a) reflects density anomalies and rheological structure in the present-day mantle. The longest-wavelength geoid with spherical harmonic degrees of 2 and 3 reveals that positive geoid amplitude peaks exist on the Africa-Atlantic regions, beneath which there are no known subducting plates, and the westernmost part of the Pacific plate, where the Australian and Pacific plates are subducting (Fig. 1b). Consequently, it is likely that the locations of the peak positive anomaly are not related to either (1) contemporary plate-tectonic mechanisms and associated mantle downwellings (i.e., subduction zones) or (2) mantle upwellings inferred from hotspot distributions at the surface (Fig. 1b) and low seismic velocity regions in the lower mantle (Fig. 1d). In contrast, when the longest-wavelength components are subtracted from the observed geoid anomaly, broad positive geoid highs appear over entire subduction zones, especially the circum-Pacific trench belt (Fig. 1c). This implies that the shorter-wavelength geoid anomaly may be strongly affected by plate tectonic processes and the locations of subducting plates.

Using an *a priori* numerical model of density anomalies and viscosity structure in the Earth's mantle as input to fluid dynamical models of mantle flow (i.e., the instantaneous flow model), we can calculate geoid anomalies and compare them with observations (Hager, 1984). However, analytical methods using propagator matrices are restricted to radially symmetric viscosity structures, because of mathematical complexities arising from mode coupling associated with laterally variable viscosity (e.g., Richards and Hager, 1989; Hager and Clayton, 1989).

On the other hand, plate tectonic processes induce distinct lateral viscosity variations (LVVs) in the mantle and lithosphere. Seismic tomography models illustrate that almost all subducting slabs reach the 660-km phase boundary, and that some of them penetrate into the lower mantle (Dziewonski, 1984; Tanimoto and Anderson, 1990; Fukao et al., 1992; van der Hilst et al., 1997). This indicates that the existence of LVVs may be due to "stiff" (high-viscosity) subducting plates. At the same time, plate margins, including "diffuse plate boundaries" (Gordon, 2000), induce LVVs in the lithosphere. The effective viscosity of diffuse oceanic/continental boundaries is at least one order of magnitude smaller than that of the stable plate interior (Gordon, 2000). Such a "weak" (low-viscosity) plate margin may have the potential to

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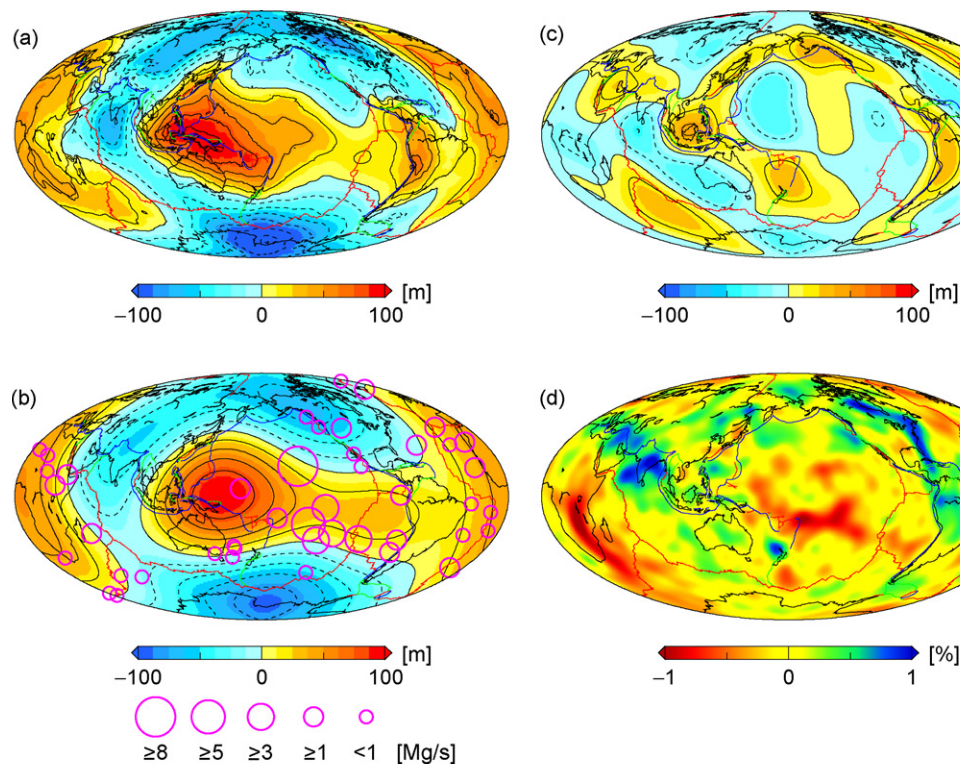

Fig. 1. (a–c) Observed geoid anomaly at spherical harmonic degrees of (a) 2–360, (b) 2 and 3, (c) 4–12, based on the EGM96 potential model (Lemoine et al., 1998) after correction for the hydrostatic shape (Nakiboglu, 1982). The contour intervals are 20 m. In (b), the distribution of 44 hotspots is shown by purple open circles, whose sizes represent the magnitude of the buoyancy flux of each hotspot. The buoyancy flux data are taken from several papers (Davies, 1988; Sleep, 1990; Schilling, 1991; Davies, 1992; Ribe and Christensen, 1999; Steinberger, 2000). Small hotspots of unknown buoyancy flux are not shown. (d) S-wave seismic velocity anomaly (δv_s) in the lower mantle (1507 km depth) from the SMEAN model (Becker and Boschi, 2002). In (a)–(d), plate boundaries are shown for reference.

affect the degree of mechanical coupling between the lithosphere and subducting slabs sinking into the mantle. These two factors of LVVs need to be considered in numerical models.

Using a numerical modeling technique, we can address models incorporating LVVs and plate configuration in three-dimensional (3D) spherical shell geometry. Plate rheology variations, arising due to stiff plate interiors and weak plate boundaries, significantly affect the long-wavelength geoid anomalies (Zhong and Davies, 1999; Yoshida et al., 2001). Zhong and Davies (1999) have shown that coupling between stiff subducting plates and weak slabs can explain the observed geoid anomaly better than stiff slabs alone. In these calculations, a subduction history model (Ricard et al., 1993; Lithgow-Bertelloni and Richards, 1998) is used to construct the density anomaly model. However, such subduction history models may lead to discrepancies with the actual slab distributions and morphologies observed in seismic tomography models. In particular, subducting slab geometries in the upper mantle inferred from subduction history modeling are somewhat broader horizontally than the geophysically observed horizontal scales of slabs.

Moresi and Gurnis (1996) has undertaken regional instantaneous flow modeling of geoid anomalies in a 3D Cartesian geometry, and suggested that the geoid is very sensitive to lateral strength variations of subducted slabs. They concluded that, a low slab viscosity in the lower mantle comparable to that of the surround mantle is required to account for the observed geoid high over the subduction zone. Our previous work (Yoshida, 2004) has shown, on the basis of a 2D Cartesian mantle convection model with self-consistent subducting plates, that the long-wavelength geoid anomaly is significantly affected by LVVs in the mantle: that is, by stiff subducting slabs and weak plate margins. However, the effects of such LVVs in 3D spherical shell geometries are not yet clear. Therefore it is important to examine which mechanism is more important in determining long-wavelength geoid anomaly patterns.

In this paper, we have examined the possible effects of LVVs on the long-wavelength (spherical harmonic degree $\ell \leq 12$) geoid stemming from stiff subducting slabs, weak plate margins and lower mantle rheology, using the instantaneous flow model in a 3D spherical shell domain. The density anomaly model used in this study has been obtained from two advanced geodynamic models; a high-resolution tomographic model and a subducting slab model based on seismicity. The global strain-rate model is used to constrain the LVV in the lithosphere [while the LVV in the lower mantle is inferred using a plausible relation between seismic velocity and temperature (i.e., temperature-dependent viscosity).

2. Model description

2.1. Numerical methods

Instantaneous mantle flow in a 3D spherical shell of 2871 km thickness is computed numerically under the Boussinesq approximation. The non-dimensionalized equations governing the instantaneous mantle flow with spatially variable viscosity are the conservation equations of mass and momentum:

$$\nabla \cdot \boldsymbol{v} = 0, \tag{1}$$

$$-\nabla p + \nabla \cdot \{\eta(\nabla \boldsymbol{v} + \nabla \boldsymbol{v}^{tr})\} + Ra_i \zeta^3 \delta\rho \boldsymbol{e}_r = 0, \tag{2}$$

where ∇ is the differential operator in spherical polar coordinates (r, θ, ϕ), $\boldsymbol{v}$ the velocity vector, p the dynamic pressure, η the viscosity, $\delta\rho$ the density anomaly, $\boldsymbol{e}_r$ the unit vector in the r-direction,