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Modelling Earth's surface topography: Decomposition of the static and dynamic components

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ABSTRACT

Contrasting results on the magnitude of the dynamic component of topography motivate us to analyse the sources of uncertainties affecting long wavelength topography modelling. We obtain a range of mantle density structures from thermo-chemical interpretation of available seismic tomography models. We account for pressure, temperature and compositional effects as inferred by mineral physics to relate seismic velocity with density. Mantle density models are coupled to crustal density distributions obtained with a similar methodology. We compute isostatic topography and associated residual topography maps and perform instantaneous mantle flow modelling to calculate the dynamic topography. We explore the effects of proposed mantle 1-D viscosities and also test a 3D pressure- and temperature-dependent viscosity model. We find that the patterns of residual and dynamic topography are robust, with an average correlation coefficient (r) of respectively ~0.74 and ~0.71, upper-lower quartile ranges of 0.86–0.65 for residual topography and 0.83–0.62 for dynamic topography maps. The amplitudes are, on the contrary, poorly constrained. For the static component, the inferred density models of lithospheric mantle give an interquartile range of isostatic topography that is always higher than 100 m, reaching 1.7 km in some locations, and averaging ~720 m. Crustal density models satisfying the same compressional velocity structure lead to variations in isostatic topography averaging 350 m, with peaks of 1 km in thick crustal regions, and always higher than 100 m. The uncertainties on isostatic topography are strong enough to mask, if present, the contribution of mantle convection to surface topography. For the dynamic component, we obtain a peak-to-peak dynamic topography amplitude exceeding 3 km for all our mantle density and viscosity models. These extremely high values would be associated with a magnitude of geoid undulations that is not in agreement with observations. Considering chemical heterogeneities in correspondence with the lower mantle Large Low Shear wave Velocity Provinces (LLSVPs) helps to decrease the peak-to-peak amplitudes of dynamic topography and geoid, but significantly reduces the correlation between synthetic and observed geoid. The correlation coefficients between all our residual and dynamic topography maps (a total of 220 and 198, respectively) is <0.55 (average = ~0.19). The correlation slightly improves when considering only the very long-wavelength components of the maps (average = ~0.23). We therefore conclude that a robust determination of dynamic topography is not feasible since current uncertainties affecting crustal density, mantle density and mantle viscosity are still too large. A truly interdisciplinary approach, combining constraints from the geological record with a multi-methodological interpretation of geophysical observations, is required to tackle the challenging task of linking the surface topography to deep processes.

1. Introduction

The Earth's surface topography at long wavelengths is mostly governed by density heterogeneity in the crust and lithospheric mantle (Parsons and Daly, 1983). Dynamic processes in the deeper mantle also play a role, deforming the lithosphere and giving place to the so-called dynamic topography (e.g. Braun, 2010). The influence of mantle convection on the topography of a planetary body was first pointed out by Pekeris (1935) and subsequently confirmed with numerical analysis by McKenzie et al. (1974). During the 1980s, a series of studies, for instance the seminal paper by Hager et al. (1985), established the importance of the relation

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between large-scale density heterogeneity in the lower mantle and dynamic topography, that allows one to model the long-wavelength geoid to a good approximation. In recent years, efforts have been directed toward understanding the amount of observed topography effectively sustained by mantle dynamics. A precise estimation of dynamic topography would indeed place important constraints on the nature of mantle convection and its interaction with the surface. Investigations have been performed on both local and global scales (e.g. Becker et al., 2014; Davila and Lithgow-Bertelloni, 2013; Flament et al., 2013; Forte et al., 2010; Levandowski et al., 2014; Lithgow-Bertelloni and Silver, 1998; Panasyuk and Hager, 2000; Steinberger, 2007; Kaban et al., 2014), resulting in contrasting estimates of dynamic topography, especially concerning its amplitude. As an example, Becker et al. (2014) and Levandowski et al. (2014) significantly disagree on how the topography of the western U.S. is affected by mantle convection. The first group of authors assign an important role to dynamic effects in shaping the surface topography of the region. Levandowski et al. (2014) instead claim that most of the topography is isostatically supported except in correspondence of the Cascadia subduction zone. These inconsistencies are due to various sources: i) applying different methodologies to infer crust and mantle density, ii) different assumptions regarding the mechanical properties of crust and mantle, iii) different schools of thought in the definition of dynamic topography (e.g., Becker et al., 2014; Panasyuk and Hager, 2000; Pari and Peltier, 2000). The last point is not a minor source for misinterpretations, since it can lead to the comparison of residual topography maps obtained with different approaches, including or neglecting the effect of the lithospheric mantle on isostatic topography.

In spite of the considerable effort involved in investigating the origin of the Earth's surface topography, uncertainties in crustal and mantle density are often neglected in its modelling. The aim of the present work is to explore the effects of such uncertainties in order to ascertain whether it is possible to disentangle, on a global scale and at long wavelengths, the isostatic and dynamic components of topography. A series of crustal density models has been obtained in a previous work (Guerri et al., 2015). Here we infer mantle density structures from seismic tomography models applying an approach based on thermodynamic modelling of mantle chemical composition. We then use the lithosphere density to compute the isostatic component of topography and to estimate the dynamic support of topography as the difference between observed topography and computed isostatic component (residual topography maps). Furthermore, we perform instantaneous mantle flow modelling in order to compute the surface deflection due to mantle dynamics. We test different viscosity profiles and a 3D pressure- and temperature-dependent viscosity distribution.

2. Data and methods

2.1. Inferring mantle density structure

The main constraint on mantle density structure comes from seismic tomography models. Mantle density is often obtained from shear wave speed by applying constant conversion coefficients (e.g. Becker et al., 2014; Lithgow-Bertelloni and Silver, 1998; Panasyuk and Hager, 2000). This approach, although straightforward, makes impossible to account for the effects of pressure, temperature and composition on the seismic to density conversion function. We rely instead on a methodology involving thermodynamic modelling and constraints from mineral physics. First, we compute pressure-temperature-composition (P-T-C) tables for seismic velocities and density. Second, we interpret relative *Vs* and *Vp* variations from tomography models for temperature, and consequently density, assuming a compositional structure.

2.1.1. Thermodynamic modelling and anelasticity

We determine the identity, amount and composition of stable phases, as well as the elastic properties of the mineralogical assemblage, as a function of P-T, for four chemical compositions (Table 1): i) a pyrolite (*pyr*) from Workman and Hart (2005); ii) a modified pyrolite (*pyr_mg85*), where the *Mg# (%Mg/(%Fe + %Mg))* is lowered from the original ~89–85, thus increasing the percentage of iron and decreasing magnesium; iii) a mixed pyrolite and basaltic composition (50% each, both from Workman and Hart (2005)) (*pyr-bas_50*); iv) a depleted lithospheric composition (*lith*) by Griffin et al. (2009). The computation is performed with the thermodynamic modelling code Perple_X (Connolly, 2009), adopting the thermodynamic formalism and thermoelastic properties dataset of Stixrude and Lithgow-Bertelloni, (2005, 2011). The stable mineralogy as a function of P-T-C is obtained by minimizing the Gibbs free energy of the system. The elastic properties of the assemblage are computed starting from a dataset of thermoelastic parameters of pure end-members. Pressure effects on the isothermal bulk and shear moduli are modelled with a third-order Eulerian finite strain expansion and temperature effects with a Mie-Grüneisen correction (Stixrude and Lithgow-Bertelloni, 2005). The moduli of each mineral phase of the assemblage are computed from the moduli of their end-members (corrected for the effects of P-T) weighted for the molar proportion in which they appear in the respective phase. The Voigt-Reuss-Hill averaging scheme (Hill, 1952) is adopted to get bulk and shear moduli for the whole rock. Finally, the elastic wave speeds of the aggregate are computed as:

$$V_P = \left(\frac{k + 4/3\mu}{\rho}\right)^{1/2} \tag{1}$$

$$V_S = \left(\frac{\mu}{\rho}\right)^{1/2} \tag{2}$$

Results are given on 2D P-T grids (one for each composition) spanning the entire P-T range of the mantle. Fine steps are considered for both pressure and temperature, 10 MPa and 10 K respectively, in order to properly account for phase transitions and variations in the composition of stable phases.

The modelling does not account for anisotropy and provides purely elastic (anharmonic) moduli. At seismic frequencies, energy dissipation due to anelastic processes involving viscous deformation does occur (e.g. Jackson, 2000). Anelastic effects are not negligible when interpreting seismic tomography models (Karato, 1993) and become important especially when the temperature approaches the rock solidus, increasing the sensitivity of *Vs* to variations in temperature (Cammarano et al., 2003). We modify the purely elastic velocities obtained from the thermodynamic computation to take anelasticity into account. Anelastic effects as a function of P, T and seismic wave frequency are modelled as in Cammarano et al. (2003):

$$Q_S = B\omega^a exp\left(\frac{agT_S(P)}{T}\right) \tag{3}$$

$$g = \frac{H(P)}{RT_S(P)} \tag{4}$$

where Q_S is the shear quality factor (the inverse of seismic attenuation), ω is seismic frequency, α is an exponent regulating frequency dependence, H is activation enthalpy, T_S is the solidus temperature and B a normalization factor. The bulk quality factor (Q_K) is assumed to be very large (1000 in the upper and 10,000 in the lower mantle) and constant. We apply the correction directly