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On the residual isostatic topography effect in the gravimetric Moho determination

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ABSTRACT

In classical isostatic models, a uniform crustal density is typically assumed, while disregarding the crustal density heterogeneities. This assumption, however, yields large errors in the Moho geometry determined from gravity data, because the actual topography is not fully isostatically compensated. Moreover, the sub-crustal density structures and additional geodynamic processes contribute to the overall isostatic balance. In this study we investigate the effects of unmodelled density structures and geodynamic processes on the gravity anomaly and the Moho geometry. For this purpose, we define the residual isostatic topography as the difference between actual topography and isostatic topography, which is computed based on utilizing the Vening Meinesz–Moritz isostatic theory. We show that the isostatic gravity bias due to disagreement between the actual and isostatically compensated topography varies between −382 and 596 mGal. This gravity bias corresponds to the Moho correction term of −16 to 25 km. Numerical results reveal that the application of this Moho correction to the gravimetrically determined Moho depths significantly improves the RMS fit of our result with some published global seismic and gravimetric Moho models. We also demonstrate that the isostatic equilibrium at long-to-medium wavelengths (up to degree of about 40) is mainly controlled by a variable Moho depth, while the topographic mass balance at a higher-frequency spectrum is mainly attained by a variable crustal density.

1. Introduction

Gravimetric methods for a Moho determination become more commonly used especially after the advent of dedicated satellite gravimetry missions, such as the Challenging Mini-satellite Payload (CHAMP; Reigber et al., 2005), the GRavity field and Climate Experiment (GRACE; Tapley et al., 2005) and the recent European Space Agency (ESA) mission, the Gravity field and steady-state Ocean Circulation Explorer (GOCE; see ESA, 2008). The main reason is a lack or total absence of seismic data in some parts of the world, while the satellite-derived gravity field has a global coverage with well-known stochastic properties. Various methods have been developed and applied for this purpose; for overview we refer readers, for instance, to Bagherbandi and Sjöberg (2013). Airy (1855) assumed a variable depth of compensation, while a different isostatic principle was proposed by Pratt (1855). He assumed that the topographic masses are compensated by lateral density changes. Both these isostatic theories assume a local compensation scheme. Vening Meinesz (1931) modified the Airy's theory by introducing a regional isostatic mechanism based on a thin plate lithospheric flexure model. He presented a concept of the Moho determination based on assuming that the Bouguer gravity anomalies are fully isostatically compensated by a variable crustal thickness (while considering a uniform crustal density). This principle was also adopted in the Parker-Oldenburg's method for the gravimetric Moho determination (Oldenburg, 1974; see also Gómez Oritz and Agarwal, 2005). Later, Moritz (1990) generalized the Vening Meinesz's theory for a global compensation and applied a spherical approximation to the problem. The principal disadvantage of these isostatic concepts is that the actual crustal density structure is often disregarded. Computations of the Bouguer gravity anomalies and the compensation attraction are realized only for constant values of the crustal density and the Moho density contrast. The condition of the complete isostatic equilibrium then does not hold exactly. Furthermore, the isostatic gravity anomalies comprise also the gravitational signal of the sub-crustal structures. A method of filtering the gravitational signature of the mantle heterogeneities was presented by Bagherbandi and Sjöberg (2012,

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2013). They used the a priori seismic Moho model to treat this residual gravitational signal in the combined gravimetric–seismic approach. Bagherbandi et al. (2013) applied this approach for a global recovery of the Moho geometry. They demonstrated that the application of the gravity corrections due to major known crustal density structures (topography, bathymetry, ice, sediments, and density variations within the crystalline crust) as well as the non-isostatic correction improved significantly the agreement between gravimetric and seismic models (by means of the RMS of their differences).

Whereas the signature of crustal density structures is mainly presented at medium-to-higher frequencies of the gravity field spectrum, the signal of deep mantle heterogeneities has mostly a long-wavelength character. These two signals can be separated to some extent by filtering (cf. Bagherbandi and Sjöberg, 2012). There are, however, additional aspects which should be taken into consideration in gravimetric methods for a Moho modeling. The isostatic mass balance depends on loading and effective elastic thickness, rigidity, rheology of the lithosphere and viscosity of the asthenosphere (cf. Watts, 2001). Moreover, geodynamic processes such as post-glacial rebound, present-day glacial melting, plate motion, mantle convection and crustal flexure contribute to the time-dependent isostatic balance (see e.g., Kaban et al., 2003, 2004; Watts, 2001). A separation of these sources by filtering is, however, not feasible, because their spectral characteristics are not well determined. The only possible way to eliminate these signals from isostatic gravity data is to apply some simplified theoretical models and assumptions.

Several studies have been conducted to better understand the spatial and spectral characteristics of the isostatic gravity field. Tenzer et al. (2009, 2011a,b, 2012, 2014), for instance, evaluated the gravitational contributions of major known crustal density structures and investigated their spectral and spatial characteristics. Bagherbandi (2011) demonstrated that the isostatic equilibrium can be attained by a variable Moho geometry only at a long-wavelength spectrum approximately up to a spherical harmonic degree of 60 (which corresponds to a half-wavelength of 3 arc-deg, or about 330 km on equator). Different estimates were given in earlier studies. Zhong (1997) reported the long wavelengths larger than 800 km, Sjöberg (1998b) mentioned 500 km, and Haagmans (2000) estimated that the isostatic equilibrium is attained at long wavelengths exceeding 200 km.

The spectral characteristics of a particular isostatic model used for a gravimetric determination of the Moho geometry can be investigated by comparing power spectra of the seismic and gravimetric Moho models. Alternatively, spectral analyses can be done for the actual topography and the respective topography predicted using an isostatic model. Following this principle, we utilize here the Vening Meinesz–Moritz (VMM) isostatic theory in the definition of the isostatic topography. We then evaluate deviations of the isostatic topography from the actual topography, which represent the residual isostatic topography (RIT). In numerical studies we investigate some spectral characteristics of global isostatic mechanisms based on the analysis of the RIT power spectrum. Moreover, the RIT values are used to study the isostatic equilibrium of the continental and oceanic crustal structures. Finally, we demonstrate how the RIT correction can improve the Moho results. Theoretical definitions given in Section 2 are applied in numerical studies of Section 3. Results and findings are concluded in Section 4.

2. Methodology

In this section we derive the expressions for computing the RIT and its effects on the gravity anomaly and the Moho geometry. Since all expressions are defined in a frequency domain, these quantities can be computed using harmonic coefficients of global gravity and crustal structure models.

2.1. Residual isostatic topography

As stated by Flament et al. (2013), the Earth's topography is mainly in the isostatic equilibrium by lateral differences in the mass density structure within the crust and the lithosphere. Following this isostatic principle, we define the RIT as a difference between the actual and isostatic (solid) topography. Hence, we write

$$\mathrm{RIT} = H - H_{IT}, \tag{1}$$

where H is the topographic height, and H_{IT} is the respective height computed based on the applied isostatic model. Furthermore, we extend the definition of the RIT in Eq. (1) also to the bathymetric depths. The VMM isostatic model (Sjöberg, 2009) was utilized to evaluate the isostatic topography and to determine the RIT effects on the gravity anomaly and the Moho geometry. It is expected that the isostatic topography deviates from the actual topography mainly due to a simple assumption in the gravimetric–isostatic model which facilitates a definition of the Bouguer gravity anomaly (cf. Bagherbandi et al., 2013). The RIT is then attributed to unmodelled crustal and sub-crustal density structures as well as additional geodynamic processes which cannot readily be described by classical isostatic models.

According to the VMM isostatic model (cf. Vening Meinesz, 1931; Moritz, 1990), the Moho depth is determined based on the assumption that the Bouguer gravity anomaly Δg_B is fully compensated by the attraction A_C. The isostatic gravity anomaly Δg_I is then defined by (Sjöberg, 2009)

$$\Delta g_I = \Delta g_B + A_C \approx 0. \tag{2}$$

The isostatic equilibrium in Eq. (2) is, however, not fulfilled exactly due to reasons mentioned in the previous paragraph.

We further decomposed the compensation attraction A_C in Eq. (2) into the mean and residual compensation terms A_{C_0} and dA_C respectively. We then write

$$\begin{aligned} A_C &= A_{C_0} + dA_C \\ &= k\left[\iint_\sigma \int_R^{R-T_0} \frac{r^2(r-r_P t)}{l_P^3} dr d\sigma + \iint_\sigma \int_{R-T}^{R} \frac{r^2(r-r_P t)}{l_P^3} dr d\sigma\right], \end{aligned} \tag{3}$$

where $k = G\Delta\rho$, G is the Newton's gravitational constant, $\Delta\rho$ is an apparent density anomaly which occurs within a depth interval between the mean Moho depth T_0 and the (actual) Moho depth T (cf. Moritz, 1990), R is the Earth's mean radius, and σ is the unit sphere. The Euclidean spatial distance between two points r and r_P is computed from $l_P = \sqrt{r_P^2 + r^2 - 2rr_P t}$, and the argument $t = \cos\psi$ is defined for the respective spatial distance ψ. The Bouguer gravity anomaly in Eq. (2) is obtained from the gravity anomaly Δg by applying the Bouguer gravity reduction. Hence

$$\Delta g_B = \Delta g - 2\pi G\rho H, \tag{4}$$

where ρ is the (average) topographic mass density.

Substituting from Eqs. (3) and (4) to Eq. (2), we arrive at

$$\Delta g_B + A_C = \Delta g - 2\pi G\rho H + A_{C_0} + dA_C = 0, \tag{5}$$

The compensation attraction terms A_{C_0} and dA_C are derived in a spectral domain by expanding the integral terms on the right-hand side of Eq. (3) into a harmonic series. After solving the radial integral and applying a binomial series, the compensation term A_{C_0} becomes (cf. Sjöberg, 2009)

$$A_{C_0} = \frac{4\pi kR}{3}[(1-\tau_0)^3 - 1], \tag{6}$$