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High resolution local Moho determination using gravity inversion: A case study in Sri Lanka

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ABSTRACT

The seismic data incorporated in global Moho models are sparse and therefore the interpolation of global Moho depths on a local area may give unrealistic results, especially in regions without adequate seismic information. Gravity inversion is a useful tool that can be used to determine Moho depths in the mentioned regions. This paper describes an interactive way of local Moho depth determination using the gravity inversion method constrained with available seismic data. Before applying inversion algorithms, the Bouguer gravity data is filtered in various stages that reduce the potential bias usually expected in Moho depth determination using gravity methods with constant density contrast assumption. A test area with reliable seismic data is used to validate the results of Moho computation, and subsequently the same computation procedure is applied to the Sri Lankan region. The results of the test area are in better agreement with seismically determined Moho depths than those obtained by global Moho models. In the Sri Lankan region, Moho determination reveals a fairly uniform thin crust of average thickness around 20 km. The overall result suggests that our gravity inversion method is robust and may be suitable for local Moho determination in virgin regions, especially those without sufficient seismic data.

1. Introduction

The boundary between the Earth's crust and mantle is termed Mohorovičić discontinuity, commonly called Moho. Generally, seismic and isostatic–gravimetric methods are used to determine Moho. Seismic methods usually provide a realistic way of imaging crust–mantle interface (Moho). However, the limited coverage of seismic data, due to the cost and effort involved in collecting the data, makes the application difficult for Moho estimation in some regions.

An alternative way of determining Moho surface could be the use of gravity data in isostatic–gravimetric methods. The main principle of isostatic methods, which define topographic compensation masses, is based on equal-pressure and equal-mass hypotheses (Heiskanen and Moritz, 1967). Two different methods have been developed based on these hypotheses: Pratt–Hayford system (1854) and Airy–Heiskanen system (1855). The former assumes varying crustal density with uniform depth, termed as level of compensation. The latter supposes varying crustal thickness with uniform density, which forms roots and anti-roots. Both systems are highly idealized and assume the compensation to be local (Bagherbandi and Sjöberg, 2012b). By introducing regional instead of local compensation, Vening Meinesz modified the Airy system in 1931 assuming that the crust is a homogeneous elastic plate floating on a viscous mantle (Bagherbandi and Sjöberg, 2012b) and later the system was further generalized and discussed by Moritz (1990) considering spherical effects.

The homogeneity of plates, however, is far from reality since many geophysical evidences have proved that the density variations inside the plates and various non-isostatic effects (Bagherbandi, 2011b; Bagherbandi and Sjöberg, 2012a) introduce isostatic anomalies which violate the fundamental assumption of the inverse problem of isostasy (Sjöberg, 2009). The isostatic–gravimetric methods of Moho estimation deal with the isostatic anomalies together with a suitable isostatic model and recently the Vening Meinesz–Moritz isostatic model (Sjöberg, 2009) has been used as it is the most realistic model among the traditional isostatic models (Bagherbandi, 2011a). Isostatic anomalies can also be used as a classical-spectral approach of Moho estimation (Braitenberg et al., 1997; Zadro and Braitenberg, 1997; Braitenberg et al., 2000).

The pure gravimetric method of Moho estimation is based on an inversion of the gravity data. Usually, Bouguer anomalies are used for the inversion. The advantage of gravity inversion is the accessibility of gravity data since the satellite gravity data together with the terrestrial data provide a seamless coverage for the whole globe (Braitenberg et al., 2004; Shin et al., 2007; Block et al., 2009). The global geopotential models are also an important source

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of gravity information (Bagherbandi, 2011c; Bagherbandi, 2012; Reguzzoni and Sampietro, 2012). The most important role of gravity inversion method is to extract the appropriate anomalies related to Moho deflection. Many techniques have been employed to isolate anomalies associated with these Moho discontinuities (Chakraborty and Agarwal, 1992; Lefort and Agarwal, 2000). Spectral methods are used to filter short and long wavelength effects of gravity stem from intracrustal/superficial inhomogeneities and deep seated sources (Sjöberg, 2009). To decide which wavelengths should be filtered out or which range of wavelengths should be used is very difficult and complex. It usually depends on the target depth, spectrum analysis of the gravity anomaly, and other geophysical or geological information (Jin et al., 1994; Shin et al., 2006; Gómez-Ortiz et al., 2011).

In this paper, we discuss a methodology of high resolution Moho depth determination using gravity inversion in smaller regions with limited seismic data coverage (Ex. Sri Lanka). The global Moho models (for example, CRUST2.0 (Bassin et al., 2000)) are sparse and therefore the interpolation of global Moho depths on a local area may give some unrealistic Moho depth estimations, especially in the abovementioned regions. Gravity inversion is a powerful tool that can be used to determine Moho depths with high resolution. It becomes more realistic with the abundance of gravity data covering the whole globe. We also describe the importance of gravity data filtering in order to estimate more realistic Moho depth that has better consistency with seismic data. Our filtering process comprises three steps: short wavelength correction, long wavelength correction, and the extraction of Bouguer anomaly spectrum that mainly represents the average Moho deflection. The final step is our main contribution of gravity data filtering that reduces the potential under/lower-estimation of Moho depths usually anticipated in gravity methods.

For numerical analysis, we use two different regions: first a test area which will be used to validate the gravity inversion method, and secondly the Sri Lankan region. Hence, the main purpose of this paper is to determine the Moho depths for the Sri Lankan region. We use the Parker–Oldenburg gravity inversion method (Parker, 1973; Oldenburg, 1974; Gómez-Ortiz and Agarwal, 2005) because of its simplicity in computation. The algorithms of this inversion scheme are discussed in the next section. The filtering process of Bouguer gravity data is explained in Section 3. The numerical results of the test and target area (Sri Lanka) are given in Section 4 and finally some conclusions of this study are drawn.

2. Gravity inversion

One of the fundamental challenges of geophysical study is to determine the geometry of subsurface structure (density interfaces) from gravity anomaly. One such important application is to map crust–mantle boundary (Moho) from surface gravity anomaly. The solution of this problem can be categorized into two main modeling techniques: forward and inverse (Ebbing et al., 2001). The latter is only considered in this article.

In inverse method, the parameters of the perturbing body are computed directly from the observed anomaly. This method becomes more popular after inclusion of Fast Fourier Transform (FFT) algorithm. Braitenberg et al. (1997) proposed an iterative method from isostatic gravity anomalies with an initial isostatic Moho. One of other important FFT applications in gravity forward modeling was presented by Parker (1973). He expressed the total gravitational anomaly due to uneven, non-uniform layer of material in terms of its Fourier transform as a result of the sum of Fourier transforms of powers of the surface causing the anomaly. The corresponding inverse scheme was presented by Oldenburg (1974). He rearranged Parker's forward algorithm to determine the density interface from the observed gravity anomaly. Although the original form of Parker–Oldenburg's algorithms is two-dimensional (2D), its three-dimensional (3D) application with large data sets in geophysical field can still be frequently found now (Gómez-Ortiz and Agarwal, 2005; Shin et al., 2006, 2007; Block et al., 2009; Bagherbandi, 2011a). The Parker–Oldenburg algorithms, which we used in this study, are briefly described below.

Parker (1973) presented the relation between the vertical gravity effect, Δg and its causative mass topography, $h(\vec{r})$ with two-dimensional form in wavenumber domain as

$$F(\Delta g) = -2\pi G\rho e^{(-|k|z_0)}\sum_{n=1}^{\infty}\frac{|k|^{n-1}}{n!}F[h^n(\vec{r})], \tag{1}$$

where $F(\Delta g)$ is the Fourier transform of the gravity anomaly, G is the gravitational constant, ρ is the density contrast across the interface, k is the wave number, $h(\vec{r})$ is the depth to the interface (positive downwards) and z_0 is the mean depth of the horizontal interface. $\vec{r}$ denotes the projection of the position $r = (x, y, z)$ onto $x - y$ plane. Parker (1973) showed that Eq. (1) is convergent when max $(h(\vec{r}) < z_0$ and $z_0 > 0$.

Oldenburg's rearrangement (Oldenburg, 1974) of Parker's scheme for the 3D case can be written as (Shin et al., 2007)

$$F[h(\vec{r})] = -\frac{F[\Delta g(\vec{r})]e^{(|k|z_0)}}{2\pi G\rho} - \sum_{n=2}^{\infty}\frac{|k|^{n-1}}{n!}F[h^n(\vec{r})]. \tag{2}$$

The Eq. (2) can be used to compute the depth to the undulating interface (Moho) from gravity anomaly by means of an iterative process, starting with an initial value of $h(\vec{r})$, for example, $h(\vec{r}) = 0$. The final Moho surface depends on the pre-set parameters, the mean Moho depth z_0 and the density contrast of the interface ρ. In order to monitor the convergence of Eq. (2), Oldenburg (1974) used the following equation

$$S_n = \underset{\text{overall k}}{\max}\left|\frac{|k|^{n-1}}{n!}F[h^n(\vec{r})]\right|. \tag{3}$$

The summation of Eq. (2) is performed until the criterion, $S_n/S_1 < E$ is satisfied, where E is a sufficiently small number. However, in practical computation, the above convergence criteria is not enough, because the term $e^{(|k|z_0)}$, which is included in Eq. (2), highly affects short wavelength. Therefore, the short wavelength features, usually coming from intra-crustal density inhomogeneity or noise, should be filtered out before the inversion process since not only would they cause problems on the stability of inversion process, but their signals may also unrelated to our target depth. As we mentioned earlier, to decide the appropriate wavelength range of the features correspond to Moho deflection is complex. This is even more difficult if the area combining both continental and marine regions since the target depth of marine regions is significantly lower than that of the continental areas.

Both Oldenburg (1974) and Nagendra et al. (1996) used a cosine filter like Eq. (4) for gravity data filtering in wave number domain. This high-cut filter, $HCF(k)$ is defined by

$$HCF(k) = \begin{cases} 1, \text{for}(k/2\pi) < WH \\ \frac{1}{2}\left[1 + \cos\left(\frac{k-2\pi WH}{2(SH-WH)}\right)\right], \text{for } WH \leqslant (k/2\pi) \leqslant SH \\ 0, \text{for}(k/2\pi) > SH \end{cases}, \tag{4}$$

where k is the wave number (frequency $\times\ 2\pi$), WH and SH are smaller and greater frequency parameters. These frequency parameters may be determined from spectral analysis of the gravity data. The filter (Eq. (4)) cuts off the frequencies higher than SH and fully passed the frequencies greater than WH. The intermediate frequencies are partially passed. This filter is regularly used in Parker–Oldenburg gravity inversion scheme. Butterworth filter could be an