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Regional perturbations in a global background model of glacial isostasy

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ABSTRACT

In Schotman and Vermeersen (2005, Earth Planet. Sci. Lett. 236), the effect of crustal and asthenospheric low-viscosity zones (LVZs) on geoid heights was shown, as predicted by models of glacial-isostatic adjustment (GIA). The governing equations were solved analytically in the spectral domain, which makes the method used accurate and fast. However, it does not allow for (large) lateral variations in earth stratification. As the properties of shallow LVZs can be expected to vary laterally, we have developed a finite-element model based on Abaqus. Global (spherical-3D) finite-element models are currently not capable of providing high-resolution predictions, which we expect due to the shallowness of the LVZs. We therefore use a regional (flat-3D) model and compute geoid heights from the predicted displacements at density boundaries by solving Laplace's equation in the Fourier-transformed domain. The finite-element model is not self-gravitating, but we compare the results with a self-gravitating spectral model under the assumption that the lack of self-gravitation is partly compensated by the lack of sphericity, and that long-wavelength differences largely cancel out when using perturbations, which are the difference between a model with and without an LVZ. We show that geoid height perturbations due to an LVZ can be computed accurately, though the accuracy deteriorates somewhat with the depth of the LVZ. Moreover, we show that horizontal rates of displacement, though not accurate for total displacements, are accurate for perturbations in the near field. We show the effect of lateral variations in the properties of the LVZ and in lithospheric thickness, and compute geoid height perturbations for Northern Europe based on a simple laterally heterogeneous model. The model is forced with a realistic ice-load history and a eustatic ocean-load history. The errors introduced by using a eustatic ocean-load history instead of realistic oceans are generally smaller than 10%, but might be critical for perturbations due to crustal LVZs.

1. Introduction

The process of glacial-isostatic adjustment (GIA), in which the solid earth is deformed by load changes due to variations in the volume of continental ice sheets, is commonly associated with long-wavelength phenomena (spherical harmonic degree < 20) as land-uplift in Scandinavia ("postglacial rebound", e.g. Milne et al., 2001), part of the geoid low in Hudson Bay-area (e.g. Tamisiea et al., 2007), and variations in the position and speed of the earth's axis of rotation (polar wander and changes in length-of-day, e.g. Mitrovica et al., 2005). These phenomena are mainly influenced by the thickness of the lithosphere and the viscosity of the mantle, and can to a large extent be explained using spherical, radially stratified (*laterally homogeneous* or 1D) viscoelastic earth models, which often use a spherical harmonic expansion to solve the governing equations in the spectral (SP) domain. In the case of lateral variations, the transformed differential equations in SP methods can no longer be solved separately for each degree due to mode coupling (Wu, 2002), which complicates the use of the SP method, especially for large viscosity gradients. Recently, *laterally heterogeneous* (3D) spherical earth models based on the finite-element (FE) method have been developed, showing the effect of lateral variations in lithospheric thickness and mantle viscosity (Wu and van der Wal, 2003; Zhong et al., 2003; Wu et al., 2005; Latychev et al., 2005).

The upcoming GOCE satellite gravity mission, planned for launch by ESA summer 2008, is predicted to measure the static gravity field with centimeter accuracy at a resolution of 100 km or less (Visser et al., 2002). This makes the detection of *high-resolution* signals as expected from shallow earth structures possible. In GIA, high-harmonic signatures (spherical harmonic degree > 20) can be induced by shallow layers (depth < 200 km) with low viscos-

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ity (~ 10^{18}–10^{20} Pas, e.g. Schotman and Vermeersen, 2005). Crustal low-viscosity zones (CLVZs) are layers in the lower crust which have viscosities that are significantly smaller than the high-viscosity, for GIA studies effectively elastic, upper crust and lithospheric mantle. As the presence and properties of CLVZs depend strongly on thermal regime (Meissner and Kusznir, 1987; Ranalli and Murphy, 1987; Watts and Burov, 2003), these can in principle not be modeled using 1D-stratified earth models. In Northern Europe for example, CLVZs are not expected in the cold and thick crust of the Baltic Shield, but might be present in surrounding younger continental shelves (Schotman and Vermeersen, 2005). Asthenospheric LVZs (ALVZs) can probably be found below the oceanic lithosphere and perhaps also below the continental shelves, but not below the Baltic Shield (Schotman and Vermeersen, 2005; Steffen and Kaufmann, 2005).

As these shallow low-viscosity layers induce high-resolution signals, the use of a 3D-spherical FE model is considered to be not feasible yet. We therefore test if predictions from a 3D-flat FE model, based on the commercial package Abaqus (Wu, 2004), are accurate enough for computing geoid height *perturbations*. Perturbations are differences between predictions from a *perturbed* background model and predictions from the background model itself. Here we assume that we know the background model, consisting of a laterally homogeneous *earth stratification* and *ice-load history*, from previous GIA studies. As we concentrate on Northern Europe, this background model will be mainly based on Fennoscandian studies (Lambeck et al., 1998; Milne et al., 2001). The presence and properties of LVZs are, via thermal regime, closely related to the thickness of the lithosphere. We are therefore also interested in constraining both the properties of LVZs and the thickness of the lithosphere simultaneously. As an additional constraint, apart from high-resolution satellite-gravity data, we will consider the use of 3D-velocities as for example measured by the BIFROST network (Milne et al., 2001), which has an accuracy of 0.8–1.3 mm/yr in the vertical direction and 0.2–0.4 mm/yr in the horizontal direction (Johansson et al., 2002).

A change in the volume of continental ice sheets leads to a global change in relative sea level (RSL), which is defined as the difference between a change in the ocean bottom topography and in the equipotential surface that coincides with mean sea level, the geoid. This makes GIA essentially a global process. In this global process, mass is conserved and redistributed gravitationally self-consistently, i.e. the mass redistribution on the surface and in the interior of the earth affect the gravitational potential, which in turn affects the mass redistribution. This is called *self-gravitation* and leads for example to a higher than *eustatic* (=uniformly distributed ice-mass equivalent) RSL near an ice sheet, because the mass of the ice sheet attracts the nearby ocean. Another effect is the long-wavelength effect in the solid earth: If we place an ice mass on the North Pole, not only the top of the earth will depress, but also the opposite side. This can be understood by considering the northward movement of the center of mass due to accumulation of mass at the North Pole, the subsequent flattening of the North Pole due to outward movement of the solid earth and the resulting change in the gravitational potential. Both effects are illustrated in Fig. 1 for an elliptical ice load, with a height at the center of 2500 m and a radius of 8°, which is on the North Pole for 10 kyrs.

In Section 6 we test if we can use a eustatic ocean-load history to force the flat, non-self-gravitating FE model. The reason we do not want to include self-gravitation in the flat model is that we then have to iterate several times over a glacial cycle (see Section 2, and Wu, 2004, for a spherical model), which is very demanding in terms of CPU time. We assume that we can neglect self-gravitation in the solid earth, because in a flat model the lack of self-gravitation is partly compensated by the lack of sphericity (Amelung and Wolf, 1994) and because we assume that some long-wavelength phenomena largely cancel out when using perturbations. Using an axisymmetric Heaviside load, we test in Section 4 the accuracy of predictions from the FE model by comparing differences with a spherical, self-gravitating SP model with the expected accuracies of the GOCE mission and the realized accuracy of the BIFROST network. Using the RSES ice-load history (Lambeck et al., 1998) and a simple model of Northern Europe, with CLVZs in relatively young continental crust and ALVZs below young continental and oceanic areas, we try to deduce in Section 5 if it is important to consider lateral heterogeneities in LVZs. We start with a short description of the theory, especially how we compute geoid heights from the output of the flat FE model (Section 2), followed by a description of the FE and SP model used in this study (Section 3).

2. Theory

The linearized, *elastic* equation of motion for geophysical problems in which inertia can be neglected is (e.g. Wu, 2004; Sabadini and Vermeersen, 2004, p. 5):

$$\vec{\nabla}\cdot\vec{\vec{\sigma}}_\delta - \vec{\nabla}(\vec{u}\cdot\rho_0 g_0 \vec{e}_r) - \rho_\delta g_0 \vec{e}_r - \rho_0 \vec{\nabla}\phi_\delta = 0 \quad (1)$$

with $\vec{\vec{\sigma}}$ the stress tensor, $\vec{u}$ the displacement vector, ρ the density, g the gravity acceleration, ϕ the gravity potential, and where the subscripts ('0', 'δ') denote the *initial* and *incremental* state (Wolf, 1998), respectively. For an *incompressible* material (Poisson's ratio $\nu = 0.5$), $\rho_\delta = 0$ and the third term (*internal buoyancy* due to material compressibility) vanishes. In this case, the gravity potential ϕ_δ

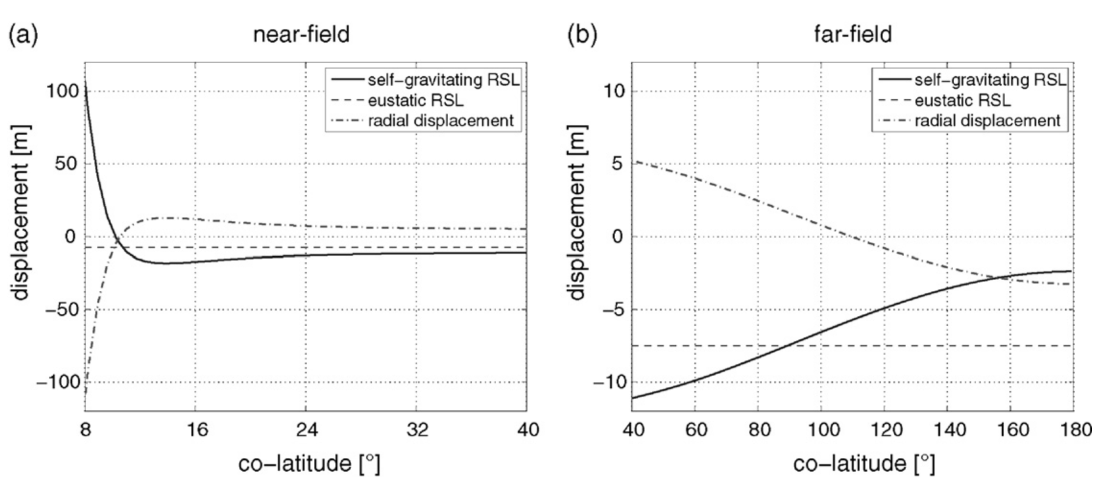

Fig. 1. Near- (a) and far- (b) field effect of an elliptical iceload (8° radius) on the North Pole after 10 kyrs of loading.