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Distinguishing large-scale secular surface mass loading from glacio-isostatic adjustment and plate tectonics in global coordinate and gravity field time series

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1. Introduction

Continental water storage (CWS), including ground water, snow and ice, is a key climate variable which impacts many parts of Earth system science and engineering. Changes in CWS will affect Earth's geocentre, gravity field and shape through surface mass loading. If other effects on these geodetic observables, including tidal and other sources of surface mass loading and perhaps technique-specific errors, can be eliminated, then it should be possible to obtain consistent multi-technique estimates of regional CWS. It is possible to discriminate surface mass loading in the frequency domain because it is the only phenomenon that changes the geodetic observables at decadal to seasonal periods (we are here limited by the time span of our data). CWS is the largest source of surface mass loading (Figure 1); atmospheric pressure loading can be modelled and removed comparatively easily, and loading due to changes in ocean dynamics is negligible. However, longer-term (secular) changes in CWS cannot so readily be discriminated from other geophysical phenomena which deform the Earth, notably plate tectonics and glacio-isostatic adjustment (GIA), which we address here.

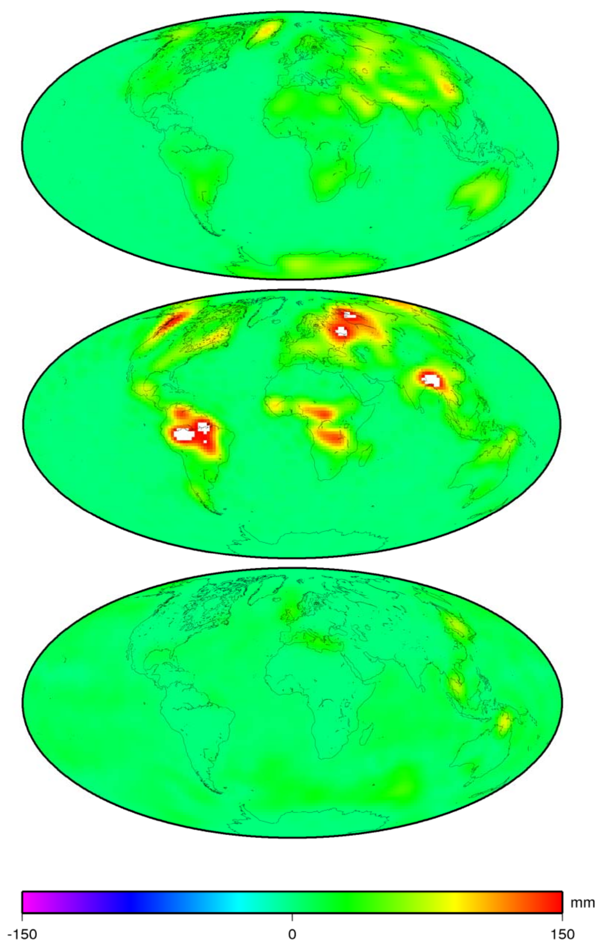

Figure 1. Annual amplitude of surface mass load variations due to atmospheric pressure (top), continental water storage (middle), and ocean bottom pressure (bottom), all expressed as equivalent thickness of sea water using a common colour scale.

2. Surface mass loading theory

A surface mass load T which depends on position Ω may be described as an equivalent layer of sea water with thickness expressed as a spherical harmonic expansion of coefficients T_{nm}:

$$T(\Omega)=\sum_{n=0}^{\infty}\sum_{m=0}^{n}\sum_{\Phi}^{\{C,S\}}T_{nm}^{\Phi}Y_{nm}^{\Phi}(\Omega)$$

where $Y_{nm}^{\Phi}(\Omega)$ are spherical harmonic functions. For a spherical elastic Earth this will lead to a change in gravitational potential $V(\Omega)$, with coefficients (Farrell, 1972):

$$V_{nm}^{\Phi}=k'_n\frac{3g\rho_s}{(2n+1)\rho_E}T_{nm}^{\Phi}$$

where g is the gravitational acceleration and ρ_s and ρ_E are the mean densities of the Earth and of sea water. Accompanying this there will be vertical displacements $H(\Omega)$ and lateral displacements $\nabla\Psi(\Omega)$, with

$$H_{nm}^{\Phi}=h'_n\frac{3\rho_s}{(2n+1)\rho_E}T_{nm}^{\Phi}\quad;\quad\Psi_{nm}^{\Phi}=l'_n\frac{3\rho_s}{(2n+1)\rho_E}T_{nm}^{\Phi}$$

Love-Shida numbers h'_n, k'_n and l'_n (Figure 2a) may be taken from a standard Earth model such as PREM (Dziewonski & Anderson, 1981). In practice, the observed variability of surface mass loads (Figure 1) makes it preferable to combine spherical harmonics into a set of modified basis functions that allow continental load variation whilst maintaining a mass-conserving, gravitationally-equilibrated ocean response (Figure 2b; Clarke *et al.*, 2007), an approach analogous to "fingerprinting" (e.g. Tamisiea *et al.*, 2001).

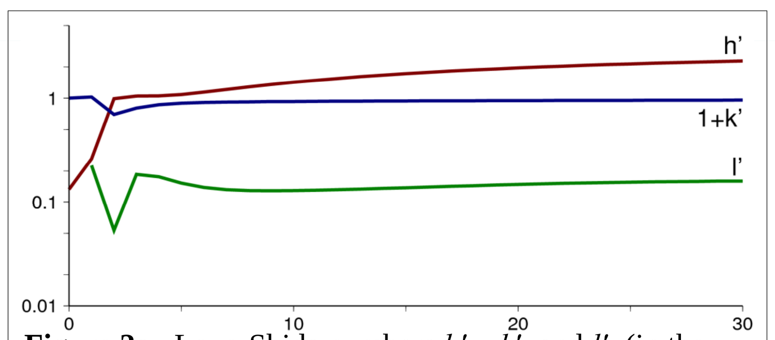

Figure 2a. Love-Shida numbers h'_n, k'_n and l'_n (in the CF frame) as a function of spherical harmonic degree n.

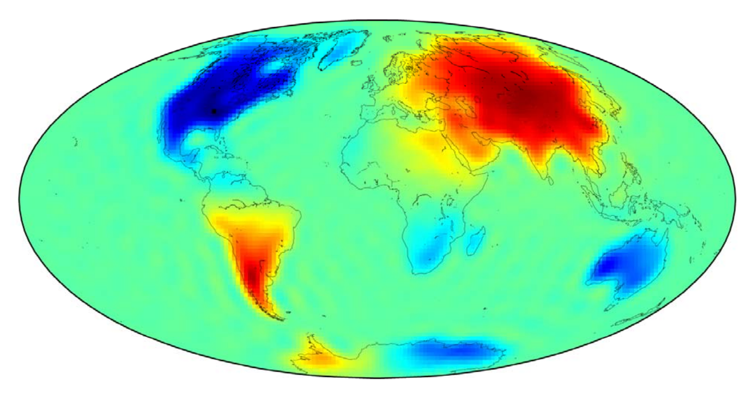

Figure 2b. Example basis function (masked degree 2, order 1, sine) showing attenuated but non-uniform (mass conserving, gravitationally consistent) oceanic load.

3. Tectonic plates and boundary zones

Earth's surface is divided into around a dozen major plates separated by boundary zones which vary from near-zero to a few hundred kilometres in width (Figure 3). In plate tectonic theory, the secular motion of sites within the stable plate interiors is purely lateral and may be described by rigid-body rotation about the plate's Euler pole. Provided the plates are large enough and are sufficiently densely monumented, this allows ready separation of plate tectonic motion from that due to large-scale loading, because the latter incorporates lateral deformation and vertical motion. However, sites in the plate boundary zones and any others known to be experiencing local motions cannot contribute to the estimation of the plate and surface mass load models. Furthermore, sites in the plate interiors may suffer from glacio-isostatic adjustment (Box 4).

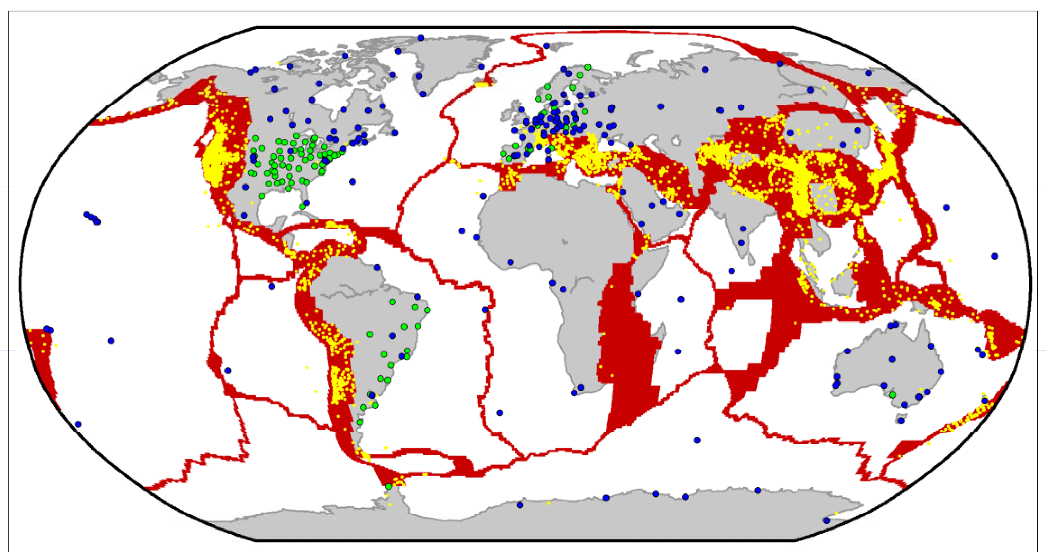

Figure 3. Long-running continuous GPS sites (blue/green) on rigid plate interiors. Other campaign and continuous GPS sites (yellow) in deforming plate boundary zones (red) are also shown. After Kreemer *et al.* (2007).

4. Glacio-isostatic adjustment

GIA is a slow, steady deformation of the Earth, caused by visco-elastic relaxation of the mantle in response to the progressive redistribution of surface mass during the last deglaciation (Figure 4). GIA can be modelled using an ice load history coupled with a layered Earth model. Present-day surface motion is both lateral and vertical; maximum vertical velocities are 2-3 times maximum lateral velocities, although the ratio at a location varies. Unlike plate tectonics, there are no undeforming areas which can be used to constrain present-day surface mass loading alone.

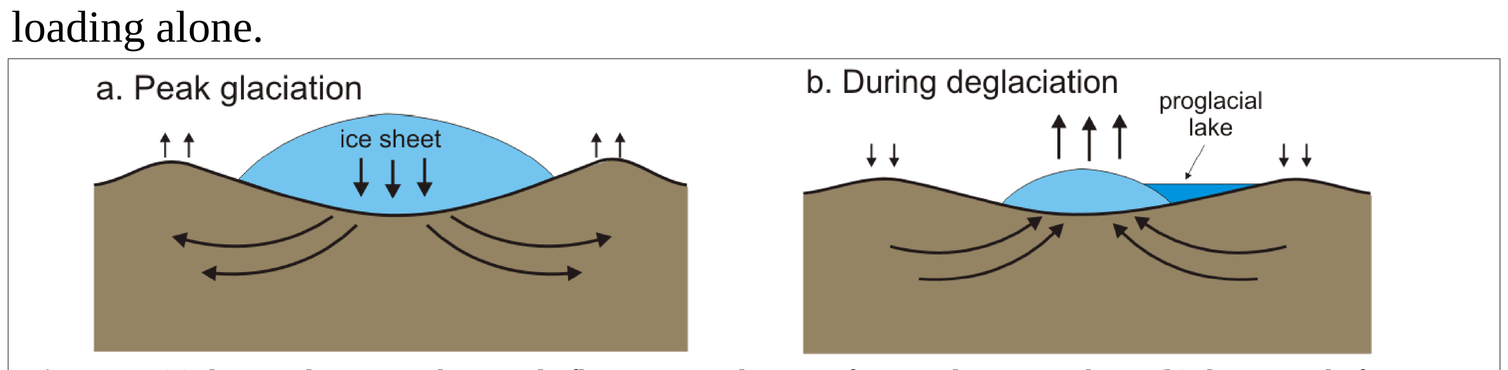

Figure 4. (a) during glaciation, the mantle flows viscously away from under an ice sheet; (b) during and after deglaciation, reverse mantle flow takes place until isostatic equilibrium is restored.

5. Estimation strategies

We will use IGS "repro1" and ILRS weekly coordinate and geocentre data, and monthly gravity field data from GRACE and SLR. Two strategies are possible:

(1) Step-by-step. Remove the GIA signal from the coordinate time series using an *a priori* model (or ensemble of models), then estimate a rigid plate model and use the residuals from this to determine the surface mass load (*e.g.* van der Wal *et al.*, 2009). The GIA-detrended gravity data can be used either with the coordinate residuals or independently.

(2) Simultaneous. Wahr *et al.* (2000) show a relationship between surface uplift and geoid height change $N(\Omega)$ due to GIA, which in our case can be used to eliminate or isolate GIA because we have both observables. Simultaneous, model-independent estimation of plate tectonics, GIA and CWS is therefore possible at low spherical harmonic degrees.

Wahr *et al.* (2000, eqs. 8-9) show that

$$H_{nm}^{\Phi,PGR}=\frac{2n+1}{2}N_{nm}^{\Phi,PGR}$$

However, we observe the total uplift and geoid height change:

$$H_{nm}^{\Phi,TOT}=H_{nm}^{\Phi,PGR}+H_{nm}^{\Phi,CWS}$$

$$N_{nm}^{\Phi,TOT}=N_{nm}^{\Phi,PGR}+N_{nm}^{\Phi,CWS}$$

But for surface mass loading,

$$H_{nm}^{\Phi,CWS}=\frac{h'_n}{1+k'_n}N_{nm}^{\Phi,CWS}$$

which leads us to

$$N_{nm}^{\Phi,CWS}=2(1+k'_n)\frac{H_{nm}^{\Phi,TOT}-\frac{2n+1}{2}N_{nm}^{\Phi,TOT}}{2h'_n-(2n+1)(1+k'_n)}$$

and hence

$$T_{nm}^{\Phi}=\frac{(2n+1)\rho_E}{3\rho_S}\cdot\frac{H_{nm}^{\Phi,TOT}-(2n+1)N_{nm}^{\Phi,TOT}}{2h'_n-(2n+1)(1+k'_n)}$$

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