



Slow postseismic recovery of geoid depression formed by the 2004 Sumatra-Andaman Earthquake by mantle water diffusion

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[1] Earthquakes are accompanied with mass redistributions and cause changes in gravity field and shape of geoid, an equipotential surface coincident with the mean sea surface. Such coseismic changes were detected by satellite gravimetry after the 2004 Sumatra-Andaman Earthquake, but little has been known on what happens on geoid after the earthquake. Here we report slow postseismic recovery of coseismic geoid depression from satellite measurements. This cannot be explained with simple afterslip or viscous relaxation of Maxwellian upper mantle. It suggests the relaxation of coseismic dilatation and compression by the diffusion of supercritical H₂O abundant in the upper mantle. Such a self-healing system of coseismic geoid undulations, a brand-new role of water in mantle, would significantly reduce the amount of permanent shifts of the Earth's rotation axis by earthquakes. **Citation:** Ogawa, R., and K. Heki (2007), Slow postseismic recovery of geoid depression formed by the 2004 Sumatra-Andaman Earthquake by mantle water diffusion, *Geophys. Res. Lett.*, *34*, L06313, doi:10.1029/2007GL029340.

1. Introduction

[2] The Earth's monthly gravity fields recovered by GRACE (Gravity Recovery and Climate Experiment) satellite, launched in 2002, have been improving our knowledge on time-variable mass distributions on the Earth including lithospheric rebound by on-going ice melting [Tamisiea *et al.*, 2005], seasonal hydrological cycles in river basins [Tapley *et al.*, 2004], water loss by drought in tropical regions [Crowley *et al.*, 2006], and seasonally changing mass in semi-closed ocean basins [Fenoglio-Marc *et al.*, 2006]. The great Sumatra-Andaman Earthquake, 2004 December 26, ruptured the eastern boundary of the Indian Plate extending >1000 km from Sumatra to the Andaman Islands causing devastating tsunami in the Indian Ocean [Ammon *et al.*, 2005]. This earthquake has been studied with various approaches such as tsunami observations with tide gauges [Tanioka *et al.*, 2006] and satellite altimetry [Hirata *et al.*, 2006], crustal movement with Global Positioning System (GPS) [Vigny *et al.*, 2005], and ionospheric disturbances [Heki *et al.*, 2006]. This was also the first earthquake whose coseismic gravity changes have

been detected by the GRACE satellite measurements [Han *et al.*, 2006].

2. Co- and Postseismic Changes in Geoid

2.1. Calculation of Coseismic Changes in Geoid Height

[3] Earthquakes modify geoid height by two kinds of mass distribution perturbations, (1) deformation of layer boundaries with density contrasts, and (2) the density changes of rocks. Uplift and subsidence of the ocean floor and Moho cause (1) and coseismic dilatation and compression of crustal and mantle rocks cause (2). Here we calculate such coseismic geoid height changes a little differently from Han *et al.* [2006]. First we assume fault parameters by Banerjee *et al.* [2005] and calculate surface and Moho uplift/subsidence and volume strain in an elastic homogeneous half space [Okada, 1992]. Difference from a more realistic approach assuming spherical and layered Earth would be <10% because mass perturbations contributing to geoid deformation mostly occur within 100 km from the fault [Cummins *et al.*, 1998]. Secondly, we calculated geoid height changes by integrating scalar potentials at the geoid surface from these perturbing masses ($=Gdm/r$, where dm is the small perturbing mass at distance r from the point on geoid, and G is the universal gravity constant) and by dividing them with gravitational acceleration. Following the work of Han *et al.* [2006], ocean floor was assumed 3 km below geoid and Moho 20 km further below, and the distance r was calculated assuming spherical Earth. We used the same mantle, crust, and seawater densities as in the work of Han *et al.* [2006].

[4] Figure 1 shows the calculated coseismic geoid height changes and their east-west profile. The uplift/subsidence signature is confined near the fault, while dilatation/compression signature extends farther away. The total geoid change is downward throughout the region although a small ridge of geoid is formed along the arc (peak at ~93E). Spatial averaging, often performed to suppress noisy high degree/order components in GRACE data [Wahr *et al.*, 1998], blurs such short wavelength components (see dashed curves in Figure 1b). Therefore the coseismic geoid change observed with GRACE, using a realistic averaging radius, would look like a plain oval depression. Dominance of downward geoid changes is the manifestation of the essence of plate subduction, a downward movement of material. Such a change in mass distribution would cause coseismic shift of the Earth's principal axis of inertia (the North Pole moves toward the epicenter if earthquakes occur in the northern hemisphere) [Gross and Chao, 2006], and contributes to the excitation of the Chandler wobble.

2.2. Observations with GRACE

[5] By combining the Stokes' coefficients in the GRACE level 2 data [Bettadpur, 2003], monthly geoid height at a

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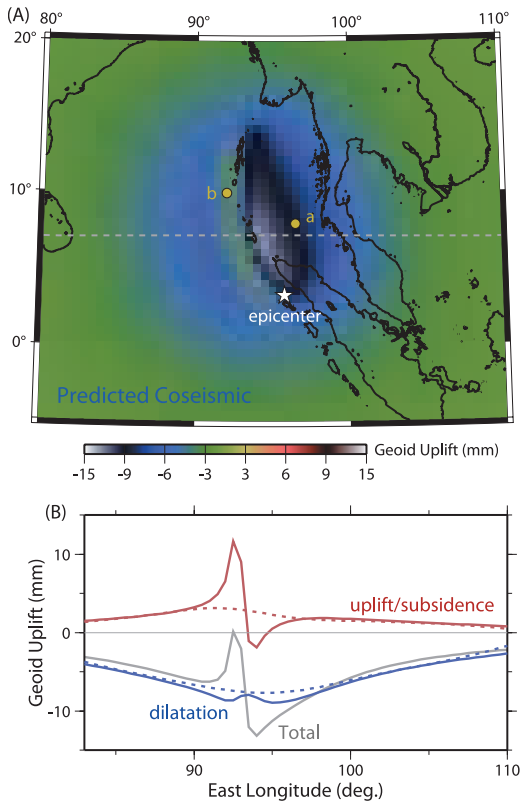


Figure 1. (a) Predicted coseismic change in geoid height and (b) its profile along the dashed line (N7.0) in Figure 1a. Contributions from uplift/subsidence of the ocean floor plus Moho, and those from dilatation plus compression of crust/mantle, and the total change, are shown by red, blue, and gray curves, respectively in Figure 1b (dashed ones are those through the 350 km isotropic Gaussian filter [Wahr *et al.*, 1998]). Open star and yellow circles in Figure 1a indicate, respectively, the epicenter and points where geoid height time series are shown in Figure 2.

desired point is calculated by Wahr *et al.* [1998, equation (1)]. We used such data (Release 1 from Center for Space Research, University of Texas) from 2002 April to 2006 October. Figure 2 shows geoid height time series at points *a* and *b* (see Figures 1a, 3a, and 3b for positions). There we used the coefficients with degree/order complete to 80 and isotropic 350 km Gaussian filter to reduce short-wavelength noises [Wahr *et al.*, 1998]. In addition to moderate amount of secular trend, the time series show strong seasonal (annual and biannual) signals that reflect yearly repeating hydrological and oceanographic cycles. To model these time series, we added two earthquake-related parameters, coseismic offset of the geoid height (unit: mm), and postseismic slow geoid height change proportional to a function, $1 - e^{-t/\tau}$, where τ is a time constant and t is the time after the earthquake.

[6] Figure 2a shows a large coseismic downward jump of geoid consistent with Figure 1. Figure 2b shows a smaller coseismic jump, but its more notable feature is the significant coseismic change in trend and its decay. The trend change is positive (postseismic geoid upheaval), and the time constant τ as long as 0.6 year resulted in the best fit

(Figures 2c and 2d). Figures 3a and 3b shows, respectively, values of the coseismic geoid height changes and cumulative postseismic changes over the 1-year period, for $1^\circ \times 1^\circ$ grid points. In order to stabilize the solution, the time constant of the decay of the postseismic changes τ is fixed to 0.6 year everywhere. Typical post-fit residuals of the geoid height time series were 1–2 mm. Geographic peaks of the coseismic and postseismic changes show some shift; the peak in coseismic geoid depression is located in the Andaman Sea while the postseismic geoid upheaval is the largest along the arc. In Figure 3c, we compare profiles of the observed and the 350 km-filtered predicted geoid changes (filtering is done by disassembling predicted geoid height changes shown in Figure 1a into Stokes' coefficients and by reassembling them after multiplying with the Gaussian filter in wave number domain [Wahr *et al.*, 1998]). They are fairly consistent throughout the region, suggesting overall appropriateness of the above described procedures.

3. Model of Postseismic Geoid Changes

3.1. Afterslip and Viscous Relaxation

[7] The 1-year postseismic geoid uplift is statistically significant, and is comparable in amount to the coseismic subsidence (Figure 3). This is the first detection of postseismic gravity/geoid changes with satellite gravimetry and its mechanism merits investigation. Interplate thrust earthquakes are usually followed by afterslips. They often release seismic moment comparable to main ruptures [Heki *et al.*, 1997]. Continuous GPS observations suggested that this earthquake is followed by an afterslip, with the moment not smaller than a few tens of percents of the main rupture [Vigny *et al.*, 2005; Hashimoto *et al.*, 2006]. Afterslips cause surface velocities similar in direction to coseismic jumps although actual directions are different in small scales owing to complementary distribution of co- and postseismic fault slips. They occur with time constants from days to years. The polarity reversal between co- and postseismic geoid changes found here suggests that simple afterslip model cannot explain the geoid behaviors in question.

[8] Viscous relaxation of the mantle also causes slow surface displacements. For example, current oceanward crustal movements in Chile and western Argentina are explained by viscous flow of mantle rocks with Maxwell viscoelasticity (with the viscosity η of 2.5×10^{19} Pa s) induced by the 1960 Chilean Earthquake [Hu *et al.*, 2004]. Assuming similar viscosity structure in Sumatra, together with elastic modulus ϵ of 40 GPa, the Maxwell time η/ϵ becomes ~ 20 years, i.e. this mechanism would hardly explain the 0.6 year timescale changes discussed here. A Kelvin element viscosity, representing delayed elasticity in a biviscous rheology, of $\sim 5 \times 10^{17}$ Pa s is suggested to explain postseismic crustal deformation as beautifully as afterslip [Pollitz *et al.*, 2006]. It is unclear, at the moment, if this successfully explains a part of the postseismic geoid recovery. In this study, we make a case for the third mechanism, water diffusion around the down-dip end of the fault.

3.2. Diffusion of Water

[9] Pore fluid diffusion causes postseismic changes opposite to coseismic ones with relatively short timescales [Jónsson *et al.*, 2003]. Dilatation and compression associ-

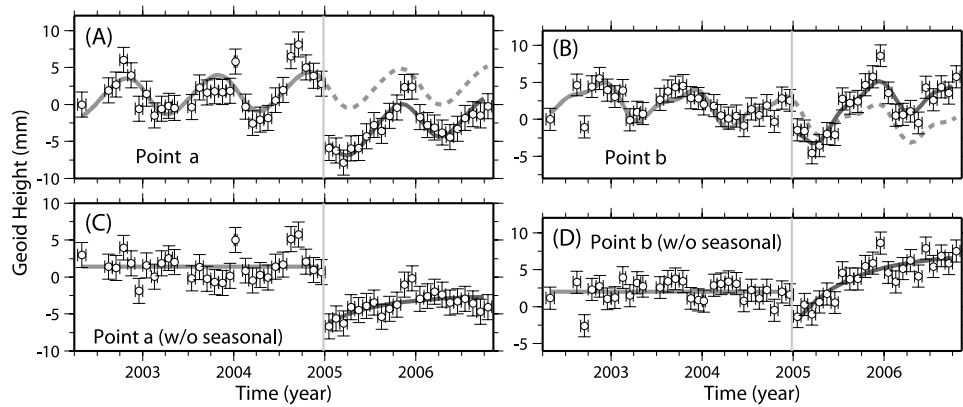


Figure 2. Time series of geoid height at the two points (a) *a* and (b) *b* (Figure 1a) calculated using GRACE level-2 data and 350 km isotropic Gaussian filter. Vertical gray lines show the occurrence time of the 2004 Sumatra-Andaman Earthquake. Light and dark gray curves indicate modeled time series before and after the earthquake (the former are extrapolated to the postseismic period as dashed curves to visualize their differences from the latter). Seasonal components are assumed same throughout the period. The error bars indicate time periods represented by individual data (horizontal bar) and a-posteriori 1σ measurement error to bring χ^2 unity after regression (vertical bar). (c and d) Same time series are redrawn after removing secular and seasonal components.

ated with faulting disturb pore fluid pressure, and the fluid diffuses from the region with increased pressure (compression part) to the decreased region (dilatation part) [Nur and Booker, 1972]. This has been supposed to occur in the uppermost crust, but we here suggest that it may occur deep in the mantle by supercritical water that abundantly exists in subduction zones.

[10] For a thrust earthquake, a pair of increased/decreased pore pressure changes is formed at the down-dip end of the fault (Figure 4a), where flow from the red to blue parts would occur following the pressure perturbation. Gravity would also encourage such upward flow in order to recover hydrostatic equilibrium disturbed by the density decrease of the shallow layer. Such mantle water flow will continue

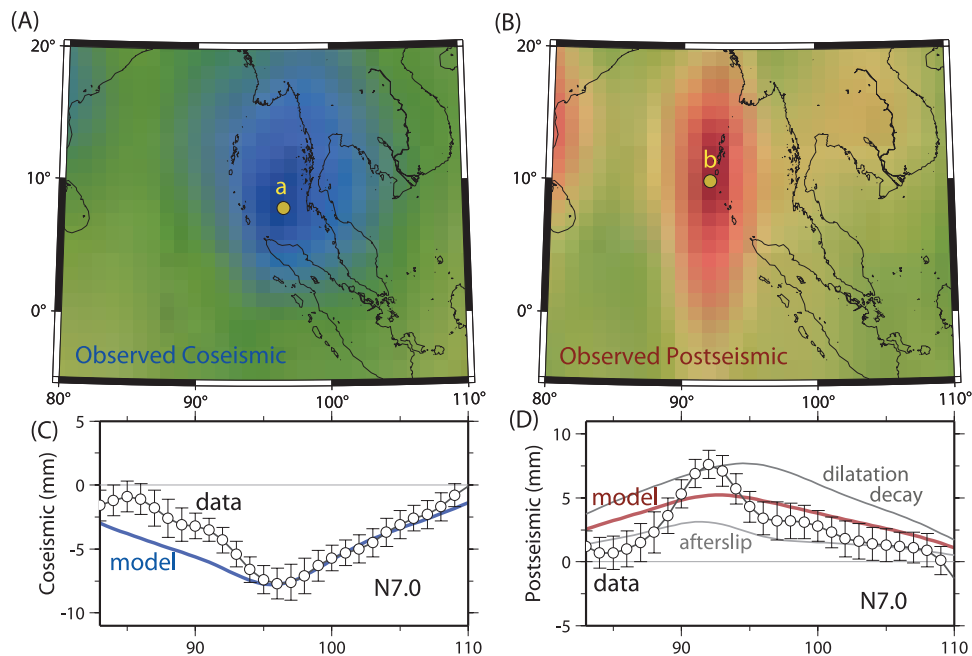


Figure 3. Geographic distribution of (a) coseismic jump of geoid height, and (b) 1-year cumulative postseismic geoid height change, are shown with the same color scheme as Figure 1a. (c and d) Their profiles along the dashed line in Figure 1a are shown. 1σ error bars there are determined a-posteriori based on the post-fit residuals of the geoid height time series (Figure 2). The model in Figure 3c is obtained by performing 350 km isotropic Gaussian filter to the predicted total geoid changes in Figure 1a. The two gray curves in Figure 3c indicate possible contributions (also filtered) from the uplift/subsidence by the afterslip (with the 1-year cumulative moment as large as the main rupture), and from the total decay of coseismic dilatation. The model in Figure 3d is obtained by combining the halves of the contributions from these two.

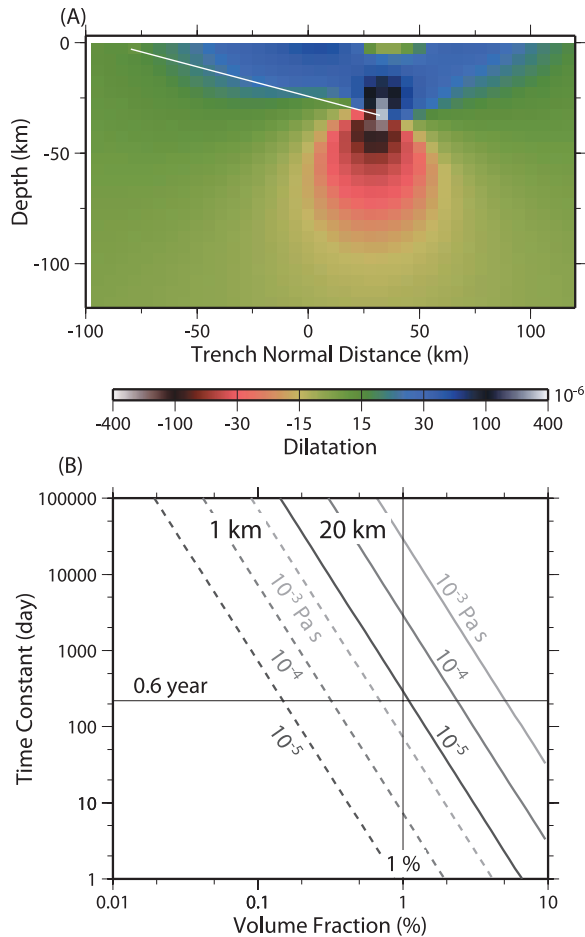


Figure 4. (a) Cross section of coseismic dilatation (blue) and compression (red) perpendicular to the fault strike at $\sim N7.0$. The white line indicates the fault [Banerjee *et al.*, 2005]. (b) Time constant of the pore pressure diffusion for distances 1 km and 20 km as functions of the volume fraction of pore fluid. Three kinds of viscosity of the pore fluid, 10^{-3} , 10^{-4} , and 10^{-5} Pa s, are assumed.

until the driving forces (pressure gradient and gravity) vanish under the new hydrostatic equilibrium. By then, the dilatation signal in geoid will be relaxed while dilatation/compression of the solid part (crustal and mantle rocks) remains almost intact. Two questions arise here, (1) if mantle water can diffuse that fast, and (2) if mantle water is ample enough to allow such an adjustment. As for question (2), the coseismic volume strain is mostly within 10^{-5} level (Figure 4a), and the weight percent of H_2O down there is much larger than this as discussed later. So movement of just a small portion of the total water could compensate density perturbation caused by such small dilatation/compression.

[11] The question (1) needs some discussions. The pressure difference lets the pore pressure obey diffusion equation with the time constant τ of $L^2/4c$, where L is typical travel distance of fluid and c is the pore pressure diffusivity [Nur and Booker, 1972]. The diffusivity is given by $c = k/\eta\beta$, where k is permeability, η is the viscosity of pore fluid, and β is bulk compressibility [Nur and Booker, 1972]. We

inferred the permeability k after McKenzie [1984], in which k scales with ϕ^3 (ϕ is the volume percent of the pore fluid). Figure 4b shows τ as a function of ϕ , and for different values of L and η . The quantity β was fixed to 2.5×10^{-11} Pa $^{-1}$. Due to ambient high pressure and temperature, water (or H_2O) behaves as supercritical fluid whose viscosity is closer to gas (10^{-5} Pa s level) than to liquid (10^{-3} Pa s level). The time constant τ of 0.6 year is realized with the volume percent of H_2O of $\sim 1\%$ and with the viscosity $\sim 10^{-5}$ Pa s. Water in oceanic crust of subducting slabs is released at the depth of a few tens of kilometers (depending on thermal structure), which then diffuses into the wedge mantle and serpentinize mantle rocks [Iwamori, 1998]. Recent petrologic studies suggest water content as high as 0.5 weight percent or more in the upper mantle of back arc [Kelley *et al.*, 2006]. Mantle water diffusion can therefore take place on a <1 year timescale required to fit postseismic geoid changes with reasonable hydraulic parameters.

4. Discussion

[12] Afterslip causes uplift/subsidence and dilatation/compression similar to the main rupture, and may leave geoid depression signature similar to Figure 1a. However, such geoid depression might recover to some extent by simultaneous diffusion of mantle water as the afterslip slowly goes on. Therefore geoid signal of afterslip may look different, i.e. composed more of uplift/subsidence signature and less of dilatation/compression signature. Figure 3d shows the “model” postseismic geoid uplift profile, calculated as the mixture of the decay of the coseismic dilatation signature (inverse of the blue dashed curve in Figure 1b, shown here as a dark gray curve), and of afterslip without dilatation (same as the red dashed curve in Figure 1b, shown here in light gray). The model assumes that an afterslip with the one-year cumulative moment release equal to one half of the main rupture has occurred, and that one half of the coseismic dilatation signature has disappeared. This model is not unique, e.g. we could consider less afterslip and more decay of the coseismic dilatation, and vice versa. The reality would lie somewhere, but current resolution of GRACE does not allow further discussion. We did not consider viscous mantle relaxation here, but its contribution to geoid changes should grow in time and merits long-term attention.

[13] The mantle water flow causes, by poroelastic rebound, crustal deformation that partly cancels afterslip GPS signatures, and it may cause underestimate of seismic moment released by afterslip. It would also cause postseismic uplift of the Andaman and Nicobar Islands, which is opposite to what the viscoelastic model predicts [Pollitz *et al.*, 2006], but is consistent with postseismic GPS observations there [Freymueller *et al.*, 2006]. Proper treatment of the poroelastic rebound would thus lead to better assessments of afterslips and viscous relaxations. From global point of view, it would call for downward revision of coseismic shifts of the Earth’s figure axis, whose estimate is based on the conventional view of mass redistribution in earthquakes [Gross and Chao, 2006]. Water plays numbers of important roles in subduction zones. It causes non-volcanic tremors in the down-dip end of seismogenic zones [Obara, 2003], and moves in the wedge mantle encouraging

magma genesis by lowering the solidus temperature [Iwamori, 1998]. Satellite gravimetry may have revealed its brand-new role, postseismic healing of coseismic geoid undulations by its diffusive adjustment.

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