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The effect of earth rheology and ice-sheet size on fault slip and magnitude of postglacial earthquakes

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ABSTRACT

Moderate past and present-day seismicity is observed in formerly deglaciated regions of North America and Europe. An understanding of the occurrence of these earthquakes is important for estimating the seismic risk within these areas as well as areas affected by recent ice-sheet melting, in particular Greenland and Antarctica.

We have developed a new finite-element approach that allows us to estimate the fault throw for areas once covered by a continental ice sheet. The simulation is initialized by loading a two-dimensional earth model with an ice sheet. The model incorporates a stress field consisting of rebound, horizontal background and vertical stresses, as well as a fault that can accommodate slip. The sensitivity of fault throw and activation time is tested with respect to lithospheric and crustal thickness, viscosity structure of upper and lower mantle, and ice-sheet thickness and width, as well as fault location and angle.

Single-event seismic displacements of up to 18.5 m are obtained, approximately equivalent to an earthquake with a moment magnitude of 8.5. The thickness of the crust and lithosphere are major parameters affecting the total magnitude of fault slip, whereas the size of the ice sheet primarily affects the activation time. Most faults start to move close to the end of deglaciation, and movement typically stops for our simulations after one thrusting/reverse earthquake. However, in our simulations faults with a dip of 60° also show several fault movements before and after the end of deglaciation.

1. Introduction

Stress is associated with the load of an ice sheet that causes surface subsidence underneath the load during glaciation and uplift during and after ice-sheet melting. The process of subsidence and uplift together with all related phenomena (e.g. mass redistribution, changes in gravity, moment of inertia, bending of the lithosphere) is called glacial isostatic adjustment (GIA). The change in the vertical displacement observed by GPS data and the change in the sea level are the most common observations of GIA (e.g. Steffen and Wu, 2011).

Areas affected by GIA are typically interspersed with faults, which were activated during or after the last deglaciation (e.g. Kujansuu, 1964; Lagerbäck, 1978; Olesen, 1988; Dyke et al., 1991; Muir-Wood, 2000; Fenton, 1994; Stewart et al., 2000). The postglacial onset of fault movements is clear, as faults offset scratch marks left by the movement of ice and offsets Pleistocene sediments (e.g. Shilts et al., 1992; Lagerbäck and Sundh, 2008). From the timing of the fault and the time scale of stress-inducing processes (see below), it is highly likely that slip on these postglacial faults is induced by the GIA process; we therefore refer to these features as glacially induced faults (GIF).

GIFs are found in cratonic areas of North America and Europe (e.g. Hoffman, 1989; Lagerbäck and Sundh, 2008; Olesen et al., 2013) that were affected by continent-wide glacial cycles during the Last Ice Age, but are almost ice-free today. Although background tectonic stresses in these stable cratonic areas are assumed to vary only on geological time scales (more than 1 million years; Luttrell and Sandwell, 2010), a glacial cycle lasts only about 100 ka (Shackleton et al., 1990; Marshall, 1998). Therefore, during a glacial cycle the tectonic background stress regime is assumed to be constant, so any abrupt observed changes in seismic activity cannot be explained by changes in tectonic stress. Furthermore, as predicted by theory, areas that still remain presently glaciated, such as Greenland and Antarctica, show almost no seismic activity below the ice sheet (e.g. Johnston, 1987; Wu and Hasegawa, 1996; Munier and Fenton, 2004), whereas in areas affected by strong ice melting higher seismicity is observed (Sauber and Molnia, 2004; Larsen et al., 2006; Voss et al., 2007; Dahl-Jensen et al., 2010).

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The dip angle of GIFs in North America and Europe varies between 45° and 80° (e.g. Fenton, 1994; Juhlin et al., 2009; Brandes et al., 2012). This high angle of dip suggests that the faults initially had a normal sense of displacement. However, observed fault offsets show a thrust movement. Tectonic stresses in North America and Europe are mainly in the thrust regime, and only a few locations are dominated by strike-slip or normal regimes (Stein et al., 1979, 1989; Adams, 1989; Slunga, 1991; Zoback, 1992; Muir-Wood, 2000; Lund and Zoback, 1999; Mazotti and Townend, 2010; Steffen and Wu, 2011; Steffen et al., 2012). Furthermore, GIA stresses alone are not large enough to create new fractures (Quinlan, 1984), which indicates that GIFs are reactivated fault zones. GIFs are also not found everywhere in deglaciated regions, but are localized in certain areas, e.g. Lapland Province in Scandinavia, and coastal areas in North America. Measured fault slip varies between a few metres in southern Canada and northeastern United States, to about 100 m in the Canadian Arctic (e.g. Kujansuu, 1964; Lagerbäck, 1978; Olesen, 1988; Dyke et al., 1991; Muir-Wood, 2000; Fenton, 1994; Stewart et al., 2000; Munier and Fenton, 2004).

Currently, northern Europe and eastern North America are also characterized by a moderate seismicity, which is not expected for stable continental regions. However, the reason for this recent seismicity is still under investigation. Bungum et al. (2010) and Pascal et al. (2010) suggest changes in tectonic and potential stresses as a trigger for earthquakes, but GIA stress might be still a factor in current seismicity.

Recently, numerical models have been developed to study the initiation of GIFs and to model the timing and the amount of throw (e.g. Hampel and Hetzel, 2006; Steffen et al., 2014). In the model by Hampel and Hetzel (2006), the faults are not only induced by GIA, but are also induced by converging plate motion, whereas a new two-dimensional (2D) model of Steffen et al. (2014) allows the determination of the throw of a pre-existing fault in a glaciated and deglaciated area solely due to GIA. In this paper we use the new 2D model, which is based on GIA results from North America, to address the following questions:

- Is the amount of fault slip related to the width and thickness of the ice sheet?
- Can observed fault throw data constrain estimates for maximum ice thicknesses for global ice models?
- How sensitive is fault slip to the thickness of the lithosphere and crust, and the viscosity structure of the mantle? In accordance with common practice for GIA studies, the term lithosphere refers to the outer shell of the Earth, which has an elastic rheology (e.g. Ranalli, 1995).
- How do calculated fault slips compare to those observed in tectonically active regions?

The purpose of the paper is thus to investigate the general behavior of faults within glaciated regions rather than to focus on specific variations of different regions on the Earth. Although, several parameters of earth and ice models are tested to identify key parameters affecting fault slip and activation time, this is not specific to a particular region.

Before the results are presented and discussed with respect to the questions raised above, background information on fault stability and model setup are explained.

2. Fault stability and state of stress

Fault stability analysis is a necessary component to understand the reactivation of existing faults in areas affected by ice sheets. Here, we use Coulomb Failure Stress (CFS) to evaluate the stability of a fault. Assuming negligible cohesion and neglecting the pore-fluid factor, the CFS is defined as follows (Harris, 1998; Steffen et al., 2014):

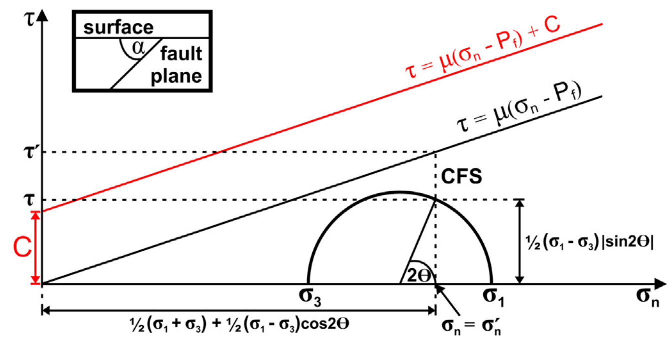

Fig. 1. Coulomb fault stability (CFS) defined as the shear-stress difference between the line of failure and the Mohr circle. The red line refers to the line of failure when the rock mass has a cohesion, and the black line without cohesion, which is used in this study. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

$$\mathrm{CFS} = \frac{1}{2}\big[(\sigma_1 - \sigma_3)|\sin 2\Theta|\big] - \frac{\mu}{2}\big[(\sigma_1 + \sigma_3) + (\sigma_1 - \sigma_3)\cos 2\Theta\big], \tag{1}$$

which depends on the magnitudes of maximum and minimum principal stresses σ_1 and σ_3, and the angle Θ between the normal to the fault plane and the direction of maximum principal stress (Fig. 1). The dip angle of the fault α to the surface depends on Θ (see for more details Steffen et al., 2014). Negative CFS refers to a stable condition, whereas a positive CFS corresponds to instability.

Neglecting pore-fluid pressure, the overburden pressure S_V using density ρ_{layer}, gravity g_{layer}, and depth z is as follows (e.g. Twiss and Moores, 2007):

$$S_V = \int \rho_{layer} g_{layer} \, dz. \tag{2}$$

The horizontal stress S_H depends on the overburden pressure S_V and a tectonic background stress component, which is determined by assuming a critically stressed fault before glaciation and with negligible cohesion along the fault. Zoback and Townend (2001) showed that the assumption of a critically stressed crust provides a good approximation for continental regions, and results from Steffen (2013) indicate that fault equilibrium before glaciation is a necessary condition to obtain fault movement. Tests with a negative CFS before glaciation show that GIFs are confined to low-angle faults below the ice-sheet centre (Steffen, 2013). A positive CFS, which refers to instability before glaciation, predicts seismic activity during glacial maximum that is not observed for currently deglaciating regions (e.g. Greenland) and already deglaciated regions (e.g. eastern North America).

The calculated tectonic background stress varies in magnitude for different dip angles, if the fault is assumed to be in equilibrium before the glacial cycle started and non-optimally orientated faults are allowed to be reactivated. Furthermore, it is assumed for simplicity that only a single fault exists and the rock mass has a higher cohesion than the fault. These assumptions lead to an equation for the horizontal background stress S_H depending on the angle of the fault (Steffen et al., 2014):

$$S_H = \frac{([\mu_{back} - \mu_{back}\cos 2\Theta + |\sin 2\Theta|]S_V)}{-[\mu_{back} + \mu_{back}\cos 2\Theta - |\sin 2\Theta|]}. \tag{3}$$

The coefficient of friction can be used as a value for the internal frictional behavior of the whole crust (μ_{back}), and as the friction between two surfaces acting against each other, e.g. at a fault, which would be called μ_{fault} (e.g. Nüchter and Ellis, 2010). μ_{back} is