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Rapid bedrock uplift in the Antarctic Peninsula explained by viscoelastic response to recent ice unloading

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ABSTRACT

Since 1995 several ice shelves in the Northern Antarctic Peninsula have collapsed and triggered ice-mass unloading, invoking a solid Earth response that has been recorded at continuous GPS (cGPS) stations. A previous attempt to model the observation of rapid uplift following the 2002 breakup of Larsen B Ice Shelf was limited by incomplete knowledge of the pattern of ice unloading and possibly the assumption of an elastic-only mechanism. We make use of a new high resolution dataset of ice elevation change that captures ice-mass loss north of 66°S to first show that non-linear uplift of the Palmer cGPS station since 2002 cannot be explained by elastic deformation alone. We apply a viscoelastic model with linear Maxwell rheology to predict uplift since 1995 and test the fit to the Palmer cGPS time series, finding a well constrained upper mantle viscosity but less sensitivity to lithospheric thickness. We further constrain the best fitting Earth model by including six cGPS stations deployed after 2009 (the LARISSA network), with vertical velocities in the range 1.7 to 14.9 mm/yr. This results in a best fitting Earth model with lithospheric thickness of 100–140 km and upper mantle viscosity of 6×10^{17}–2×10^{18} Pa s – much lower than previously suggested for this region. Combining the LARISSA time series with the Palmer cGPS time series offers a rare opportunity to study the time-evolution of the low-viscosity solid Earth response to a well-captured ice unloading event.

1. Introduction

Rapid changes in climate in the Antarctic Peninsula (AP) over the past 50 years have led to the retreat and eventual collapse of several major ice shelves (Fig. 1), such as Prince Gustav by 1993–1995 (Rott et al., 1996), Larsen A in 1995 (Rott et al., 1996), and Larsen B in 2002 (Rack and Rott, 2004) (see Cook and Vaughan (2010) for a complete summary). In response to ice shelf collapse, tributary glaciers have exhibited acceleration and thinning (e.g., De Angelis and Skvarca, 2003; Rignot et al., 2004; Scambos et al., 2004) and this dynamic ice loss induces a solid Earth response which may be observed in GPS records.

The study of Thomas et al. (2011) identified markedly-increased uplift in GPS coordinate time series from the Northern Antarctic Peninsula (NAP) that they associated with ice unloading related to the breakup of Larsen B Ice Shelf in 2002. This uplift was best captured in the near-continuous cGPS record at Palmer station which exhibited an increase in uplift rate from 0.1 mm/yr prior to 2002.2, to 8.8 mm/yr thereafter. Thomas et al. (2011) suggested that the effect was due to the elastic response of the solid Earth but they were not able to satisfactorily reproduce the increased uplift rates with an elastic model, which they suggested was at least partly due to the weakly defined magnitude and spatial pattern of ice-mass loss in their model.

The NAP lies in a complex tectonic setting which passes from active subduction along the South Shetland Trench, located north

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Table 1
Location of cGPS stations, observing period, and observed uplift velocities.

Site	Latitude (°)	Longitude (°)	Observing period	cGPS observed uplift (mm/yr)
Palmer (PALM)	−64.78	−64.05	1998.5–2013.0	6.6±2.1 (2009.0–2013.0 only)
Cape Framnes (CAPF)	−66.01	−60.56	2010.1–2013.0	4.5±2.9
Duthier's Point (DUPT)	−64.81	−62.82	2009.3–2013.0	12.8±2.1
Foyn Point (FONP)	−65.25	−61.65	2010.1–2013.0	14.9±2.7
Hugo Island (HUGO)	−64.96	−65.67	2009.3–2013.0	1.7±3.3
Robertson Island (ROBI)	−65.25	−59.44	2010.1–2013.0	7.8±2.9
Vernadsky (VNAD)	−65.25	−64.25	2010.1–2013.0	5.8±2.4

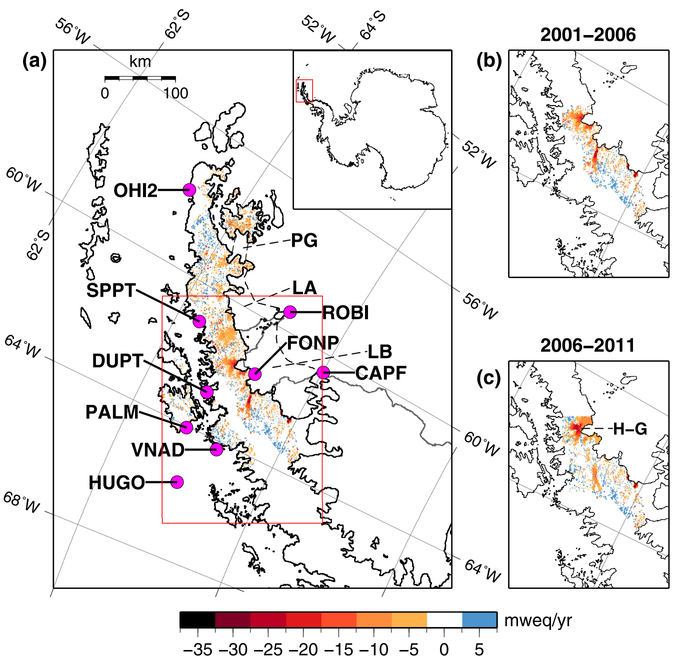

Fig. 1. Observed ice-mass change rate given in metres water equivalent per year. a) The full study area with cGPS locations shown as pink circles and former ice shelf locations as dashed black lines (Prince Gustav (PG), Larsen A (LA) and Larsen B (LB)). Values in the Larsen B area (see Fig. 1b) represent the mean rate of change for the period 2001–2006, values elsewhere represent the mean rate of change for the period 2003–2009. Inset shows location of the study area. b) Ice-mass change for Larsen B only using 2001–2006 data. c) Ice-mass change for Larsen B only using 2006–2011 data. H-G is the Hektoria-Green drainage basin. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

of the South Shetland Islands, to passive margin west of 65°W at the intersection of the Hero Fracture Zone with the Shetland Platform (see Fig. S1 in the supplementary material). The Bransfield Basin separates the South Shetland Islands from the NAP and has active volcanism along a mid-axial region, suggesting a back arc tectonic setting (Barker and Austin, 1998; Barker et al., 1991). The conversion from active subduction to passive margin, and hence mantle conditions more likely reflective of low viscosity, took place relatively recently at ~ 4 Ma before present along the AP margin just south of Hero Fracture Zone (Barker et al., 1991). One of the GPS stations used in this study (ROBI, see Section 2.1) lies along a presumed incipient rift axis as expressed by the active volcanic chain of the Seal Nunataks (González-Ferrán, 1983). However, there is little known in relation to the mantle or crustal configuration beneath the Seal Nunataks region. The reader is referred to Barker (1982) and Larter and Barker (1991) for a full tectonic history of the region. Due to the active tectonic setting of the region, the mantle is likely to have a relatively low viscosity compared with other locations undergoing deformation in response to changes in ice-mass e.g. East Antarctica or Fennoscandia. Using a combination of inferred ice history, GPS and GRACE data, Ivins et al. (2011) suggested this region has a relatively thin lithosphere (20–45 km) and a low viscosity mantle ($3–10 \times 10^{19}$ Pa s). Due to the low viscosity nature of the upper mantle, the Earth's viscous response to ice-mass change in the AP is much more rapid than in other regions of Antarctica, and post-1995 unloading events may hence be contributing to the observed uplift in the NAP through viscoelastic rebound. Likewise, assuming a Maxwell rheology, there may be very little, to no, residual response of the NAP to unloading events associated with recession from the Last Glacial Maximum.

In this study we use cGPS data from the NAP to constrain a local model of solid Earth response to a high resolution present-day mass loss field (Scambos et al., in review). By comparing the modelled elastic uplift time series with the observed cGPS uplift time series from Palmer Station we show that elastic uplift alone cannot reproduce the cGPS observations. A viscoelastic model is then used to predict uplift based on a range of Earth models, and results are compared with the Palmer record to obtain the range of best fitting models. Finally, the Earth model is further refined using six shorter cGPS time series from the NAP (see Table 1).

2. Data

2.1. GPS

Fig. 1 shows the locations of available cGPS sites in the NAP. Of these, we used the seven sites closest to the region of ice-mass change (see Table 1). Palmer is a long term station with an excess of 15 years of data, and the remaining six sites were installed in 2009–2010 as part of the LARISSA project (LARsen Ice Shelf System, Antarctica, 2013) (http://www.hamilton.edu/expeditions/larissa). We did not use the record from O'Higgins (a compilation of three records from two adjacent stations, OHIG, OHI2, OHI3; labelled OHI2 in Fig. 1) as a constraint as it lies far from the region of largest mass loss and as such may be affected by potential lateral heterogeneity in Earth structure. Spring Point (Bevis et al., 2009) (SPPT in Fig. 1) was also excluded due to the lack of data at this site; however we compare our results with both of these records in the supplementary material.

The cGPS data from 1998 through to the end of 2012 were processed using a Precise Point Positioning strategy (Zumberge et al., 1997) within GIPSY-OASIS v6.1.2. Homogeneously reprocessed (as of 2012) satellite clocks and fiducial-free orbits were fixed to values provided by the Jet Propulsion Laboratory. The receiver clocks, tropospheric zenith wet delay, tropospheric gradients and station coordinates were estimated in standard ways (e.g. Thomas et al., 2011). For the troposphere, we adopted the Vienna Mapping Function v1 (Boehm et al., 2006) and we assumed hydrostatic zenith delays based on the European Centre for Medium-Range Weather Forecasts. Higher order ionospheric effects (Petrie et al., 2011) were not corrected. Solid Earth tides were corrected according to the IERS 2010 conventions (Petit and Luzum, 2010) and ocean tide loading displacements were corrected based on the TPXO7.2 ocean tide model (Egbert et al., 2009), which has been shown to perform very well in this region (King et al., 2011), using the SPOTL