

Estimates of present-day glacial rebound in the Lambert Glacier region, Antarctica

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Abstract. Changes in the ice load since the time of the Last Glacial Maximum (LGM) contribute to the present-day vertical motion of the Antarctic continent. The observation of these motions will reveal information on the ice load history. Predictions of uplift rates along a transect across the Lambert Glacier region, East Antarctica, from the coast to the southernmost rock outcrops of the Prince Charles Mountains have been computed for three deglaciation scenarios. The relative vertical velocities between sites on the transect are -7 to +7 mm/yr and are large enough to be detected from continuous GPS observations taken at permanent sites over several years. When available, such information will discriminate between the currently available models for deglaciation of East Antarctica.

Introduction

An important unresolved problem in understanding the glacial history of Antarctica is the retreat and reduction in volume of the East Antarctic Ice Sheet since the time of the last maximum glaciation about 20,000 years ago. Reconstructions of this ice history, largely unconstrained by quantitative information, vary significantly as seen by the three reconstructions discussed below. Because of the Earth's delayed response to deglaciation, precise measurements of present-day crustal rebound can be used to provide constraints on the ice models. This paper sets out to estimate rates of present uplift for different ice scenarios and to establish whether a GPS measurement campaign can distinguish between alternate ice models. The first steps have been taken in carrying out such measurements as part of an integrated geological-geodetic experiment.

Most studies which aimed to reconstruct the history of former ice sheets from their isostatic effects have used sea-level observations as the primary data source, because sea-level serves as a reference mark against which the vertical motion of the Earth's surface can be measured. This has been highly successful in Europe and North America, where the density of sea-level change data makes it possible to separate eustatic and isostatic contributions [Lambeck, 1993]. In contrast, sea-level records in Antarctica are few in number and typically span only the last 6,000 years [Zwartz *et al.*, 1998]. Other methods of measuring the iso-

static response to changes in ice thickness and distribution are therefore needed. Gravity anomaly measurements are one such method [Greischar and Bentley, 1980; Bentley and Wahr, 1998]; another is to measure the vertical motion of the crust directly.

The Lambert Glacier is one of the largest outlet glaciers in East Antarctica (Figure 1), draining approximately 10% of the East Antarctic ice sheet. Its history is thus significant to that of the East Antarctic ice sheet as a whole. It is flanked along its length by the Prince Charles Mountains, and discharges into Prydz Bay via the Amery Ice Shelf. Observations of the vertical extent of moraines deposited on mountains adjacent to the glacier directly constrain ice thickening in this area during the last glacial maximum [Mabin, 1992]. Ice thickness during moraine deposition stood higher by a maximum of ~600m near the present grounding zone of the Lambert Glacier, with more limited thickening both downstream and to the south and west towards the ice sheet interior. Maximum ice levels were maintained until 12 ka [Stone *et al.*, 1998]. The grounding zone is believed to have migrated about 500 km northwards during the LGM to a position beyond the present Amery Ice Shelf edge [O'Brien and Harris, 1996], indicating that substantial changes in the distribution of grounded ice have occurred.

Recent improvements in satellite geodesy allow field measurements of vertical motion from Global Positioning System (GPS) observations which are accurate enough to detect isostatic rebound [Scherneck *et al.*, 1998]. The Lambert Glacier region is ideally suited to measure differential vertical motion by geodetic methods because there are ice-free sites along a very long transect from the coast to 800 km inland in the Prince Charles Mountains. In this paper, we present estimates of rebound in the Lambert Glacier region, East Antarctica, and demonstrate that GPS observations at appropriate sites will discriminate between different scenarios of the ice sheet history in Antarctica. In particular, observations of vertical velocity from sites along the Lambert Glacier might constrain the timing of deglaciation and reveal whether the thickening indicated by moraines represents the major addition to the regional ice mass during the LGM, or whether the advance of the ice sheet adjacent to Prydz Bay was also significant. We also present preliminary results from GPS observations made in Lambert Glacier region.

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Glacial rebound modelling

Vertical motion predictions are presented here for three well-known reconstructions of Antarctic ice change. The ANT3 model [Nakada and Lambeck, 1988] contributes 37 m to eustatic sea-level rise since the LGM, and is based on the

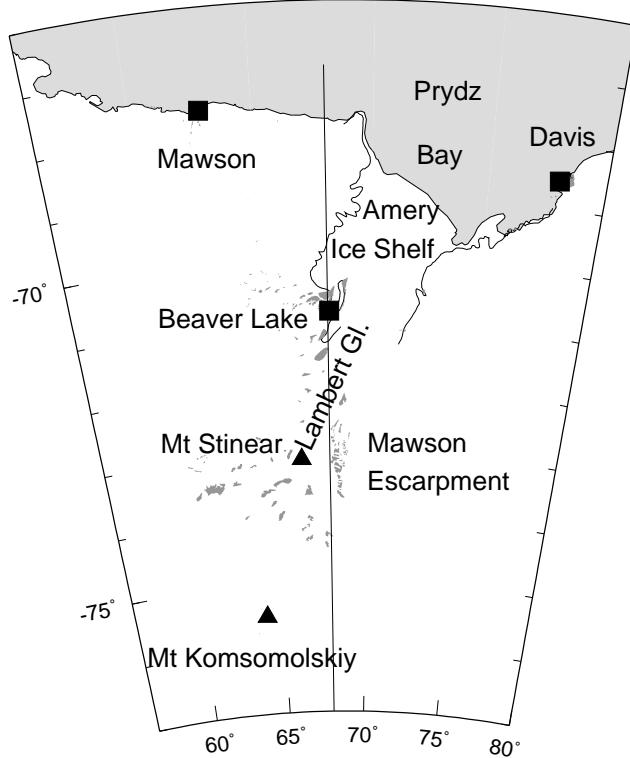


Figure 1. Map of the Lambert Glacier region, showing rock outcrops and the transect along which predictions of vertical velocity have been calculated. The locations of existing GPS sites (squares) and preferred locations of additional GPS sites (triangles) are plotted.

difference between the maximum ice sheet reconstruction of *Denton and Hughes* [1981] and the present-day ice thickness of *Drewry* [1982]. The ICE-3G model [*Tushingham and Peltier*, 1991] contributes 25 m to eustatic sea-level rise. A third model, HUY, contributes 13 m of eustatic sea-level rise and is derived from a thermo-mechanical numerical model of the ice sheet, forced by changing temperature and sea-level over the last glacial cycle [*Huybrechts*, 1990]. Of the three models, the change in ice thickness for HUY (Figure 2) agrees best with the moraine evidence.

In models ANT3 and ICE-3G, the timing of deglaciation is imposed according to constraints from records of eustatic sea-level at far-field sites. The ANT3 model melts mostly from 14 – 6 ka BP; ICE-3G from 9 – 4 ka. The variation between the models is due to the respective authors' differing corrections for the isostatic component of sea-level change at far-field sites. The HUY model has a much later maximum glaciation, at 9 ka BP, and melts mostly after 6 ka.

We calculated the present day vertical motion of the Earth's surface in response to each scenario of ice sheet history using the method of *Mitrovica et al.* [1994], performed to spherical harmonic degree 256 with two iterations of the water load. Both viscous and elastic components of the isostatic rebound are included. Variation in three earth model parameters was considered: the elastic thickness of the lithosphere h_{lith} , and the upper and lower mantle viscosities η_{um} and η_{lm} . h_{lith} was allowed to vary from 50 – 100 km, η_{um} from $(3 - 5) \times 10^{20}$ Pa s, and η_{lm} from $(5 - 10) \times 10^{21}$ Pa s.

This range encompasses the optimum models derived from studies in Europe and Australia [*Lambeck and Nakada*, 1990; *Lambeck et al.*, 1998]. Depth dependence of density and elastic parameters are based on the PREM [*Dziewonski and Anderson*, 1981].

The predicted vertical motion due to the northern hemisphere ice sheet model ARC3 [*Nakada and Lambeck*, 1988] was calculated using the same globally-consistent method and added to each of the Antarctic models (ANT3 and HUY). ICE-3G already includes the northern hemisphere ice sheets. Adding the ARC3 model to ANT3 and HUY in this way results in a different total eustatic sea-level for each model: 126 m for ANT3+ARC3; 102 m for HUY+ARC3; and 115 m for ICE-3G. However, because the northern hemisphere ice sheets are so far from the region of this study, the vertical motion estimate is mainly due to the water load of their 89 m contribution to eustatic sea-level rise. This effect contributes less than 1 mm/yr to the predicted vertical velocity, so scaling it in order to give the different models equal eustatic sea-levels does not alter the predicted pattern of uplift.

We have not included the response of the Earth's crust to the modern ice sheet mass balance. Recent calculations [*James and Ivins*, 1998] suggest that this may be contributing up to several mm/yr in some parts of Antarctica. In particular, their J92 model predicts significant rebound in

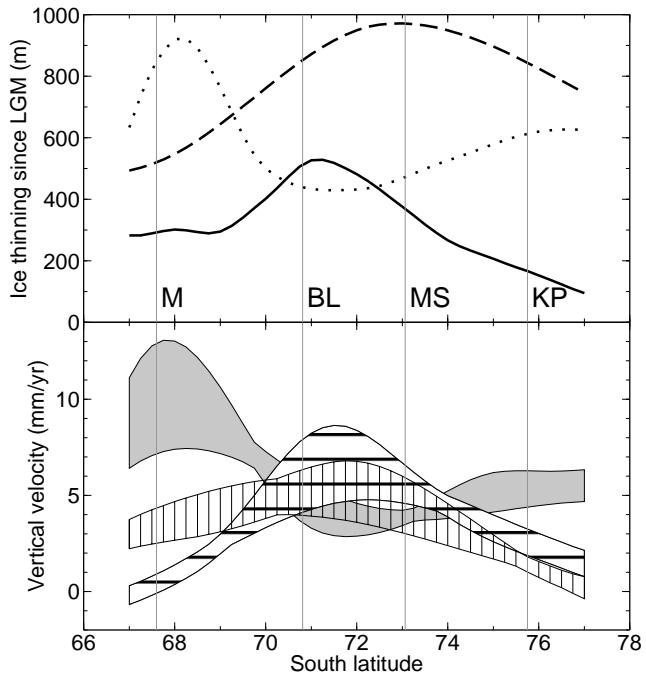


Figure 2. Reduction in ice thickness (upper panel) and predicted present vertical motion (lower panel) along the Lambert Glacier transect shown in Figure 1, using the range of Earth rheology models described in the text. Dashed line and vertical hatching = deglaciation model ANT3, solid line and horizontal hatching = HUY, dotted line and grey shading = ICE-3G. The locations of current and proposed GPS observation sites are indicated: M = Mawson Station, BL = Beaver Lake, MS = Mount Stinear, KP = Komsomolskiy Peak.

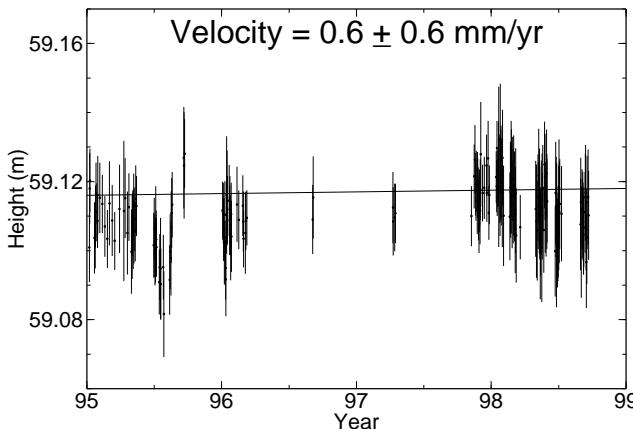


Figure 3. Vertical velocity at GPS site MAW1 (Mawson Station), derived from observations from 1995 to the present. The gaps in the time series represent data which have not yet been processed.

the Lambert Glacier region because large negative mass balance values are assigned to areas immediately inland of all ice shelves, including the Amery. In the J92 model, the effect of modern mass balance at Davis Station is larger than the effect of rebound from deglaciation; however, over most of the Lambert Basin, particularly to the west, the long-term rebound is the dominant signal. Moreover, the assignment of a negative mass balance to the Lambert region in the J92 model is not supported by glaciological measurements in the region [Higham et al., 1997]. Nevertheless, the omission of this effect introduces an uncertainty of around 2 mm/yr to the estimates in this paper.

GPS observations

To obtain sufficient precision to detect isostatic rebound from GPS observations, it is necessary to have a permanent installation and to observe for a long period of time. Permanent instruments installed by the Australian Survey and Land Information Group (AUSLIG) have operated at Australia's Davis and Mawson stations to the east and west of the Lambert Glacier (Figure 1) since 1993. Analysis of GPS data observed at Mawson since 1995 is still in progress. A least squares linear regression of the preliminary height estimates indicates a vertical velocity of 0.6 ± 0.6 mm/yr (Figure 3); however, estimating accurate vertical velocities using GPS requires careful consideration of many possible sources of error (e.g. definition of the terrestrial reference frame, atmospheric pressure loading effects and geocentre motion), and we are investigating the sensitivity of the vertical velocities to these influences. We therefore consider our preliminary velocity to be nominal at this stage and emphasise that the quoted uncertainty is the formal error.

In January 1998, a new geodetic site was installed near Beaver Lake on the southwest of the Amery Ice Shelf ($70^{\circ}49'S$ $68^{\circ}04'E$). The scatter in height computed from the first 25 days of data is ± 7.5 mm for each daily solution. Tregoning et al [1999] estimated the expected precision of uplift rate at this site by simulating height estimates for a 5 year period, applying an uncertainty of ± 10 mm to each

daily estimate and then computing the uncertainty of the slope of a linear regression through all the estimates. They concluded that isostatic rebound of < 1 mm/yr should be able to be detected after a few years of observations at the site.

Results and Discussion

Predictions of vertical motion were calculated in the Lambert region along a north-south transect at 68°E longitude, from 67° to 77°S (Figure 1). This line passes close to the region of maximum ice thinning in the ANT3 and HUY scenarios. In the ICE-3G scenario, the maximum change in ice thickness occurs along the Mawson Coast and in Enderby Land, to the west of the transect line. Estimates along a transect through Mawson Station are not significantly different. The predictions (Figure 2) based on the three models are distinctly different at the north end of the transect, with uplift rates ranging from 0 to 13 mm/yr, whilst at Beaver Lake all three models predict similar intermediate values. The predicted uplift at Beaver Lake relative to Mawson Station is thus $\sim +7$ mm/yr for the HUY model, ~ -7 mm/yr for ICE-3G, and ~ 0 mm/yr for ANT3. This relative uplift rate is large enough to be detected from rates derived from continuous GPS height measurements at the currently installed sites of Beaver Lake and Mawson, potentially allowing us to differentiate between these models. Furthermore, additional sites located in the southern Prince Charles Mountains and on Mt Komsomolskiy ($\sim 76^{\circ}\text{S}$) would provide better constraints on all ice models and, when combined with other geomorphological constraints, would lead to an improved understanding of the ice history of the region.

The estimated rate of uplift at Mawson from 3 years of GPS observations is 0.6 ± 0.6 mm/yr (Figure 3), which supports the HUY model, indicating only minor change in the ice thickness outside the Lambert Basin. However, this GPS estimate is preliminary and very sensitive to small changes in the definition of the terrestrial reference frame of the GPS global network. The effect of these changes and other sources of error can be reduced by deriving the relative uplift rates between Mawson and Beaver Lake. There is currently insufficient GPS data available from the Beaver Lake site to estimate an accurate rate of uplift; hence we cannot yet draw any conclusions about the validity of the ice sheet models.

The present vertical velocity due to postglacial rebound is not simply a function of the former ice thickness - the timing of deglaciation and possible glacial readvances since the LGM are also significant. Geomorphological studies conducted in the Prince Charles Mountains over several recent field seasons will provide constraints on these factors when the programme of exposure-age dating is completed [Stone et al., 1998]. Conversely, observations of vertical motion, interpreted by comparison with rebound calculations, can constrain the ice history in parts of the region where there is no geomorphic evidence. In this way, the best possible ice sheet reconstruction can be obtained.

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