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Snow- and ice-height change in Antarctica from satellite gravimetry and altimetry data

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ABSTRACT

We combine the surface-elevation and surface-mass change derived from Envisat data and GRACE solutions, respectively, to estimate regional changes in air and ice content of the surface of the Antarctic Ice Sheet (AIS) between January 2003 and October 2010. This leads, upon certain assumptions, to the separation of the rates of recent snow-accumulation change and that of ice-mass change. We obtain that the height of ice in Thwaites and Pine Island glaciers sectors decreases (≤ -15.7 cm/yr) while that in the Kamb glacier sector increases (≥ 5.3 cm/yr). The central part of the East AIS is mostly stable while the whole Dronning Maud Land coast is dominated by an increase in snow accumulation. The Kemp land regions show an ice-mass gain that accounts for 67–74% of the observed rates of elevation change in these regions. A good agreement is obtained over 68% of the investigated area, mostly in the East AIS, between our estimated rates of snow accumulation change and the predicted rates of the monthly surface mass balance derived from a regional atmospheric climate model.

1. Introduction

The Antarctic Ice Sheet (AIS) is the largest reservoir of ice on Earth. Accurately defining the present-day mass balance of the AIS yields vital information on its contribution to sea level rise (e.g. Cazenave et al., 2009). Understanding all the processes involving mass redistribution such as changes of ice thickness and snow accumulation or the isostatic adjustment induced by surface-mass changes is required to correctly estimate the mass balance of the ice sheet. Indeed, quantifying the ice-sheet mass balance still remains challenging as no single process dominates at the whole of ice-sheet scale, rather several processes compete to differing degrees at the basin scale with regional variations, leading to multiple mass redistribution patterns (Shepherd et al., 2012).

The radar altimeter onboard Envisat measured the variations of the surface elevation of the AIS between 2002 and 2010. Altimetry data are thus very useful to understand the changes in the geometry of the ice sheet (e.g. Rémy and Parouty, 2009). However, to obtain a mass balance from the rate of volume change derived from these data, one needs to use appropriate densities (e.g. Gunter et al., 2009; Slobbe et al., 2009; Li and Zwally, 2011; Lee et al., 2012). These apparent densities are difficult to assess in Antarctica mostly because in-situ measurements are sparse and varying surface masses are not purely ice or snow (Li and Zwally, 2011).

The Gravity Recovery And Climate Experiment (GRACE) satellite mission provides the space and time variation of the Earth gravity field since 2002. These variations are in part caused by the redistribution of mass on the surface of the Earth and consequently can be used to study the redistribution of mass in Antarctica. More specifically surface-mass changes in Antarctica are estimated by correcting GRACE data for the secular effect induced by the response of the Earth to the late Pleistocene deglaciation (Glacial Isostatic Adjustment, GIA) (e.g. King et al., 2012; Shepherd et al., 2012; Whitehouse et al., 2012b; Velicogna and Wahr, 2013). Using GRACE data, recent studies have shown that the AIS is subject to a change of mass at a rate ranging between −137 and 41 Gt/yr depending on the time interval considered (King et al., 2012; Shepherd et al., 2012; Barletta et al., 2013; Sasgen et al., 2013; Velicogna and Wahr, 2013). These studies have also shown that the mass in the East AIS (EAIS) is increasing while that of the West AIS (WAIS) is decreasing at a much higher rate. Interannual variations are also observed in the AIS (e.g. Ramillien et al., 2006; Horwath et al., 2012).

Horwath et al. (2012) compare linear rates of change in the AIS surface height with those of surface mass (expressed as elevation change in ice equivalent) estimated using Envisat measurements

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and GRACE solutions, respectively. They find consistent patterns of change in the ice sheet, the differences being attributed to the changes in air content of the firn. They obtain a correlation coefficient of 0.8 between the elevation and surface mass time series showing that the interannual signals observed by both methods are also consistent. They resolve from both data sets an accumulation event in West Antarctica occurring in 2005 September/October. The results obtained by Horwath et al. (2012) demonstrate that Envisat and GRACE satellites are observing the same geophysical processes over the ice sheets. Therefore, combining satellite gravity and altimetry data should lead to a better understanding of the processes affecting the distribution of surface masses in Antarctica (e.g. Wahr et al., 2000; Riva et al., 2009; Gunter et al., 2013). Wahr et al. (2000) and Velicogna and Wahr (2002) combine 5 years of simulated satellite laser altimetry and gravity measurements and estimate the uncertainties associated with the GIA signal and snow compaction on estimating the mean Antarctic mass balance. Their results show that the effect of the time variability of snow accumulation is the largest source of uncertainty. Riva et al. (2009) combine data from the laser altimeter of the Ice, Cloud, and land Elevation Satellite (ICESat) and the GRACE solutions between March 2003 and March 2008 in Antarctica. They estimate the changes in surface heights due to GIA and deduce changes in the firn depth and ice thickness. They use two spatially varying density maps, one modelling the lateral variation of the effective rock density, the second dealing with a priori surface densities. The largest surface density is that of the ice whereas the lowest ranges between 300 and 450 kg/m^3, a value of 600 kg/m^3 is used for Kamb glacier. These surface densities employed by Riva et al. (2009) are a simplification of the actual apparent densities required to separate the variations resulting from snow accumulation or ablation and ice-mass change. For example, they find that the rate of mass change in the Kamb glacier region is due to variations of the firn depths while glaciological evidence favours an ice increase induced by the shutdown of the Kamb ice stream about 130 years ago (e.g. Joughin and Tulaczyk, 2002; Smith et al., 2005; Rignot et al., 2008; Pritchard et al., 2009). Similarly, mass change in the Amundsen sea catchment in West Antarctica is attributed to ice-mass change while results of Lee et al. (2012) suggest that both ice and snow mass vary. Gunter et al. (2013) have recently updated the study by Riva et al. (2009) to estimate GIA and ice-mass change in Antarctica.

The aim of this paper is to estimate the spatial variation in the change of ice thickness and snow accumulation in Antarctica and to assess the theoretical difficulties associated with the estimation of both quantities. Based on the results of Horwath et al. (2012), we developed a new methodology combining Envisat and GRACE data (Sections 2–3) to determine the rates of snow and ice-height change (Section 4). Then, we discuss our results (Section 5) and conclude (Section 6).

2. Elevation and surface-mass changes

We use Envisat altimetric observations from August 2002 to October 2010 (i.e. up to 85 repeat passes) to estimate monthly variations of the AIS surface elevation. The reference elevation is taken as March 2004 as this corresponds to the first cycle with the most complete coverage. The along-track processing, that allows a dense data set, is explained in Flament and Rémy (2012) (see Supplementary material). Grids of 5 km × 5 km resolution are binned in 0.25° × 0.25° resolution grids to be compared with GRACE data. Rates of surface elevation change (on the 5 km grid) are shown in Fig. 1-a.

Surface-mass changes in Antarctica are determined using the release 2 of the GRACE solutions provided by the Groupe de Recherche de Géodésie Spatiale (GRGS) (Bruinsma et al., 2010).

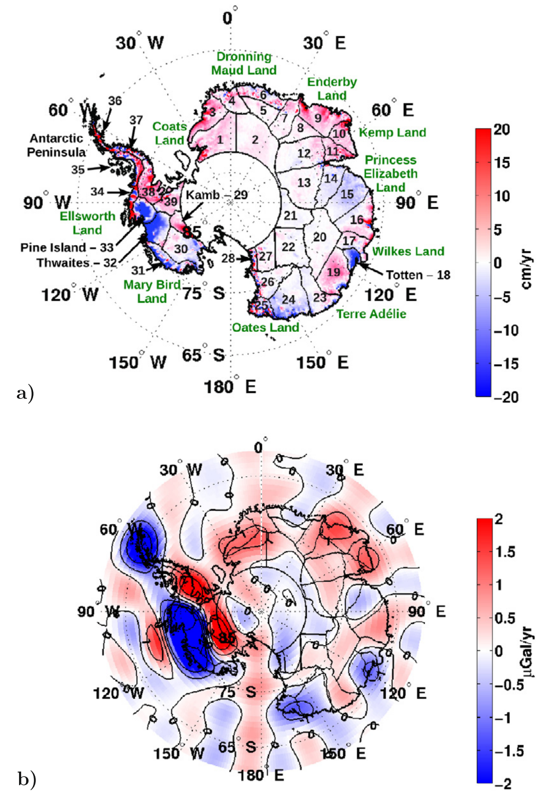

Fig. 1. Rates of elevation (a) and gravity (b) variations estimated from Envisat data and GRACE solutions, respectively. Positive and negative values are for mass gain and loss, respectively. Region boundaries used in this study are in black. Names of the glacier and land mentioned in the text are written in black and green, respectively. Note: 1 µGal = 10 nm/s^2. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

These solutions are available (http://bgi.grgs.fr) as fully normalized spherical harmonic coefficients, or Stokes coefficients, of the 10-day gravity potential field up to the harmonic degree 50. This is the maximum degree for which the GRGS provides harmonic coefficients and it corresponds to a spatial resolution of ∼400 km. The GRGS solutions are stabilized during their generation process by gradually constraining the coefficients of degree 2 through degree 50 to the coefficients of the static field (Bruinsma et al., 2010). This results in significant reduction of the noise characterized by South–North stripes. We add the degree one terms using those estimated by Swenson et al. (2008), available from January 2003. To ensure a temporal consistency with the AIS elevation changes, we estimate monthly variations of the gravity potential field with respect to that of March 2004. The rate of change derived from the GRACE data is shown in Fig. 1-b.

We remove the GIA contribution to monthly maps of variation in elevation and gravity potential using the W12a model of Whitehouse et al. (2012b). This model is driven by a deglaciation history developed using a numerical ice-sheet model constrained by geological evidence of past ice extent (Whitehouse et al., 2012a). It currently provides a good fit to present uplift rates in Antarctica observed using the Global Positioning System (Thomas et al., 2011). The W12a model is released with upper and lower bounds which are later used here to compute the contribution of GIA to the uncertainty in our estimated rates.

We use the map of drainage basins of Zwally and Giovinetto (2011) and subdivide and modify several of their basins to define smaller regions (Fig. 1-a) according to the pattern of the elevation and gravity rates obtained from Envisat and GRACE (Fig. 1), respectively, or according to topographic features. We compute regional elevation changes, h_{env}, time series by averaging monthly maps