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Towards quantifying the contribution of the Antarctic ice sheet to global sea level change

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Abstract. At present, the mass balance of the Antarctic Ice Sheet (AIS) and its contribution to global sea level change are poorly known. Current methods to determine AIS mass balance as well as the inherent uncertainties are discussed. Special emphasis is placed on the increasingly important role of regional atmospheric climate models, which can reduce the uncertainties in surface accumulation, the correction for the firn layer depth and density in ice thickness calculations and moreover help in interpreting surface elevation changes in terms of accumulation and firn density variability. Some recent advances in these fields of research are presented.

1. INTRODUCTION

The Antarctic ice sheet (AIS) is the largest freshwater source on Earth, larger by an order of magnitude compared to the Greenland ice sheet and even by two orders of magnitude compared to all other glaciers and ice caps combined. The Antarctic climate is too cold to allow for significant melt and runoff, and model studies suggest that the first order effect of a warmer climate on AIS mass balance would be to increase snowfall on the ice sheet and hence lead to sea level lowering [1]. But recent results from both Antarctica and Greenland strongly suggest that fringing ice shelves play a central role in modulating the dynamic behaviour of the inland ice [2-6]. In turn, these ice shelves have proven to be very sensitive to changes in atmospheric/oceanic temperature [7, 8]. Ice shelf disintegration through enhanced meltwater ponding and basal melt may, in the future, be responsible for unexpectedly rapid changes in mass balance of the entire AIS.

Presently, three methods are used to estimate AIS mass balance: i) repeated weighing of the ice sheet using remotely sensed gravity anomalies, ii) equating mass input (accumulation) and output (solid ice flux) for individual ice drainage basins and iii) remotely sensed elevation changes in time (radar/laser altimetry). All three methods have their advantages and deficiencies, while methods ii) and iii) are partly complementary. Several recent studies used these techniques to estimate the mass balance of the East Antarctic ice sheet (EAIS), the West Antarctic ice sheet (WAIS) and their sum (Table 1) [9-12]. Despite three of the studies employing the same data set and the rapid, unambiguous thinning of some coastal glaciers in West Antarctica [6, 13], there is clearly no consensus over the contribution of the AIS to global sea level change. Although the three elevation change studies suggest that the mass of the EAIS increases (Table 1), other work based on ice cores and regional climate modelling indicates no significant increase in Antarctic precipitation over the last 25-50 years [14-16]. Results of studies 1-3 are also in conflict with recent (and particularly since 1999) sea level rise and ocean freshening trends [17].

This lack of consensus is due to the paucity of ice dynamical and climate data from Antarctica. For example, the average area density of accumulation observations on the Antarctic ice sheet is $1/6,000\,\mathrm{km}^2$, sixty times less than what is deemed necessary to capture precipitation variability at middle latitudes. This gap cannot be closed simply by increasing the number of *in-situ* observations, for these are expensive and time-consuming: a smart combination of new observations, modelling and remote sensing techniques must be used. In this paper we highlight the role of regional atmospheric climate modelling to narrow down the uncertainty margins in AIS mass balance estimates.

Table 1. Estimates of Antarctic ice sheet mass balance in 10^{12} kg per year. EAIS = East Antarctic Ice Sheet, WAIS = Wast Antarctic Ice Sheet, AIS = Antarctic Ice Sheet.

Study	*Ref.*	*Method*	*Period*	*EAIS*	*WAIS*	*AIS*
Davis and others (2005)	[9]	Radar altimetry	1992-02	$+45 \pm 7$	-	-
Zwally and others (2005)	[10]	Radar altimetry	1992-02	$+16 \pm 11$	-47 ± 4	-31 ± 12
Wingham and others (2006)	[11]	Radar altimetry	1992-02	$> +27$	< 0	$+27 \pm 29$
Velicogna and Wahr (2006)	[12]	Gravity data	2002-05	0 ± 56	-148 ± 21	-152 ± 80

2. METHODS AND UNCERTAINTIES

2.1 Repeated weighing of the ice sheet

This method uses the Gravity Recovery And Climate Experiment (GRACE) satellite, and is completely independent of the other two methods. AIS mass balance estimates obtained from the GRACE satellite cover both the EAIS and WAIS, but only for a short period of time (2002-2005, Table 1). The results imply a dramatic mass loss for the WAIS ($-148 \pm 21\,\mathrm{km}^3$, equivalent to 0.4 mm of sea level rise per year) but approximate balance for the EAIS [12]. Apart from the uncertainties deriving from the short time period over which the trends have been calculated (3 years), several corrections must be applied to GRACE data, each introducing additional uncertainties. An example is the correction for upward motion of the Earth's crust after the last glacial maximum (postglacial rebound), which is poorly constrained over Antarctica. Another example is the atmospheric mass correction, for which the air pressure at the surface of the ice sheet must be known at better than 1 hPa precision. Given the sparse network of meteorological stations in Antarctica one must resort to modelled meteorological fields, which are known to be unreliable. This method will not be further discussed here.

2.2 Equating mass input and output

This method relies on estimating the difference between snowfall over a catchment area (mass input) and the outgoing ice flux along the catchment gate at the ice sheet grounding line (mass output). It requires techniques from various disciplines: i) meteorological modelling and in-situ observations to obtain the surface accumulation distribution, ii) satellite radar interferometry to obtain the flow speed of the narrow glaciers through the flux gates at the ice sheet grounding line and iii) satellite altimetry to accurately delineate the ice drainage basins as well as to obtain the elevation/thickness of the floating glacier. Because the resulting mass imbalance represents the difference between two large terms, this method is sensitive to uncertainties in the individual components. That is why until now, this method has mainly been applied to individual AIS drainage basins with sufficient data coverage [18, 19]. However, using remotely sensed ice velocities/thickness and elevation data and an ever-improving surface accumulation distribution, an AIS-wide assessment of the mass balance will soon be feasible.

Another problem is created by using a multitude of data sources covering different time intervals: local variations in mass balance may be driven by short or long term changes in ice velocity and/or accumulation that may or may not be related to recent climatic forcing. Combining this method with remotely sensed elevation change measurements (method iii, see below) can help elucidate whether an imbalance has a dynamic or snowfall driven origin and whether it is a long-term, secular trend or a short-term perturbation.

2.3 Remotely sensed elevation changes in time

This method assumes a direct relation between changes in ice sheet elevation (as observed by satellite radar/laser altimeters) and changes in ice volume. Datacoverage is not perfect: no data exist for the area south of 81.5°, while in the steeply sloping coastal margins of the ice sheet the signal to noise ratio is too small to obtain accurate height changes. This is especially problematic for the narrow, fast flowing outlet glaciers, which are not adequately resolved by present-day sensors but which are expected to react most rapidly to environmental changes. With the loss of CryoSat this problem will not be resolved for at least another 5-7 years.

Another major uncertainty is introduced by the fact that the altimeter time series are too short to infer the underlying cause of elevation changes (ice dynamics vs. snowfall). There are indications that decadal changes in accumulation and firn densification are a major source for the observed elevation changes [6]. Firn densification rate is a strong non-linear function of firn temperature; Zwally and others [10] corrected the observed elevation trends for densification changes using a firn densification model driven by AVHRR-derived surface temperatures [19], but owing to a lack of data they had to assume a constant accumulation rate. The three recent studies using radar altimetry listed in Table 1 suggest that over the 11-year period 1992-2002, the EAIS is gaining mass, while the WAIS looses a similar amount. These results are somewhat in conflict with recently observed salinity changes, which would require a freshwater source in the southern hemisphere [17].

3. REDUCING THE UNCERTAINTIES USING A REGIONAL ATMOSPHERE MODEL

Regional atmospheric climate models run at higher resolution than global models, and are therefore well suited to narrow down the uncertainties in AIS mass balance calculations. Below we describe three applications using output of the regional Antarctic climate model RACMO2/ANT at 55 km horizontal resolution [21], namely i) the surface accumulation distribution, ii) the firn depth correction and iii) elevation changes owing to firn densification and accumulation variability.

3.1 Surface accumulation

Errors in surface accumulation compilations based on interpolation of *in-situ* observations may differ by as much as 20% on the basin scale [22-24]. To make matters worse, the quality of many 'older' (pre-1980) *in-situ* observations has recently been questioned (C. Genthon, personal communication 2006), which affects *all* existing observation-based accumulation compilations.

A more observation-independent approach is to nudge output of a (regional) meteorological model towards fewer, but strict quality-controlled observations. This ensures that data sparse regions still get a physically correct accumulation magnitude [15], while potentially remaining erroneous measurements do not negatively impact large areas of the domain. Van den Berg and others [21] showed that even without any nudging, modelled accumulation from RACMO2/ANT compares very well to 1900 *in situ* quality-controlled observations over the AIS ($R = 0.82$). This is remarkable if one keeps in mind that the accumulation periods are different and that the model is not nudged towards observations. Fig. 1 shows the accumulation distribution after nudging in 500 m elevation intervals. The surface accumulation field shows detail on scales much smaller than the irregularly distributed observations, confirming the added value of this approach. Van den Broeke and others [25] showed that processes that are not yet explicitly modelled in RACMO2/ANT, such as the erosion by and sublimation of drifting snow, could account for the remaining differences between modelled and observed accumulation. Modelling these processes explicitly and further increasing the model resolutions is a line of future research.

Fig. 1 compares the modelled and observed accumulation. The model captures the steep accumulation gradient between the wet coastal zone and the dry interior. It also reliably reproduces the accumulation gradients over the flat ice shelves. New accumulation data should confirm the band with very high accumulation rates (>2000 kg m^{-2} yr^{-1}) along coastal West Antarctica and the west coast of the Antarctic Peninsula.

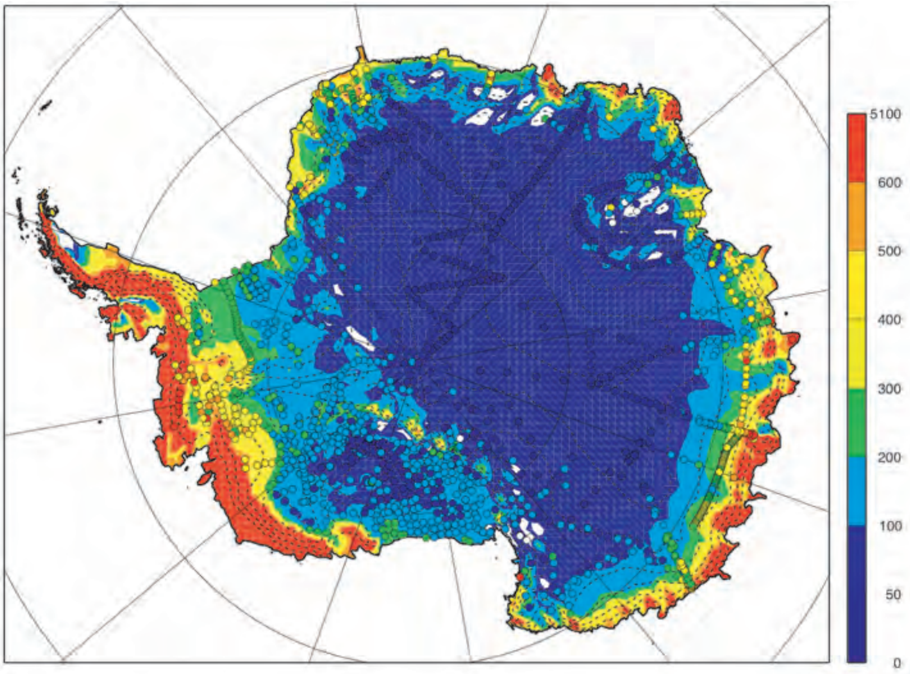

Figure 1. Observed (circles) and modelled accumulation in kg m $^{-2}$ year^{-1}.

3.2 Ice thickness and firn correction depth

For most Antarctic glaciers, ice thickness at the grounding line has not been directly measured and must be calculated from accurate satellite elevation observations and a floatation criterion. The floatation criterion equates the mass of the displaced seawater with the added weight of the combined ice/firn column:

$$(h_i + h_f - h_{asl})\,\rho_{sw} = h_f\rho_f + h_i\rho_i \equiv \rho_i H_i \tag{1}$$

where h_i, h_f and ρ_i, ρ_f are the thicknesses and average densities of the ice and firn layers, respectively, h_{asl} is the elevation of the floating glacier surface above sea level and ρ_{sw} is the density of seawater which for this application can be assumed constant. The equivalent ice thickness (H_i) is obtained by compressing the firn layer until it has the density of glacier ice (917 kg m^{-3}). An expression for H_i in terms of h_{asl} is obtained by eliminating h_i:

$$H_i = \frac{(h_{asl} - \Delta h)\rho_{sw}}{\rho_{sw} - \rho_i}; \quad \Delta h = h_f(1 - \rho_f/\rho_i) \tag{2}$$

where Δh is the *firn depth correction*, defined as the difference between the combined ice/firn column and the equivalent ice thickness. It is clear from (2) that to convert h_{asl} to H_i, Δh must be known at the grounding line, i.e. the firn depth and average firn density must be known.

Assessing the depth and average density of the firn layer requires medium-deep firn cores that reach the firn-ice transition at 450 − 150 m depth. These cores are rare in Antarctica and seldom drilled at the grounding line (Fig. 2). Therefore, the spatial distribution of Δh along the grounding line is poorly known at present, and Δh is often assumed constant in ice flux calculations. For a typical ice thickness of 500 m, ($h_{asl} \approx 54$ m), an uncertainty in Δh of 4 m introduces a 7.4% uncertainty in the ice flux estimate, which is a large number compared to uncertainties in the other components of the solid ice flux calculation, e.g. surface elevation from satellite laser altimetry (uncertainty <0.5 m or <1% for this

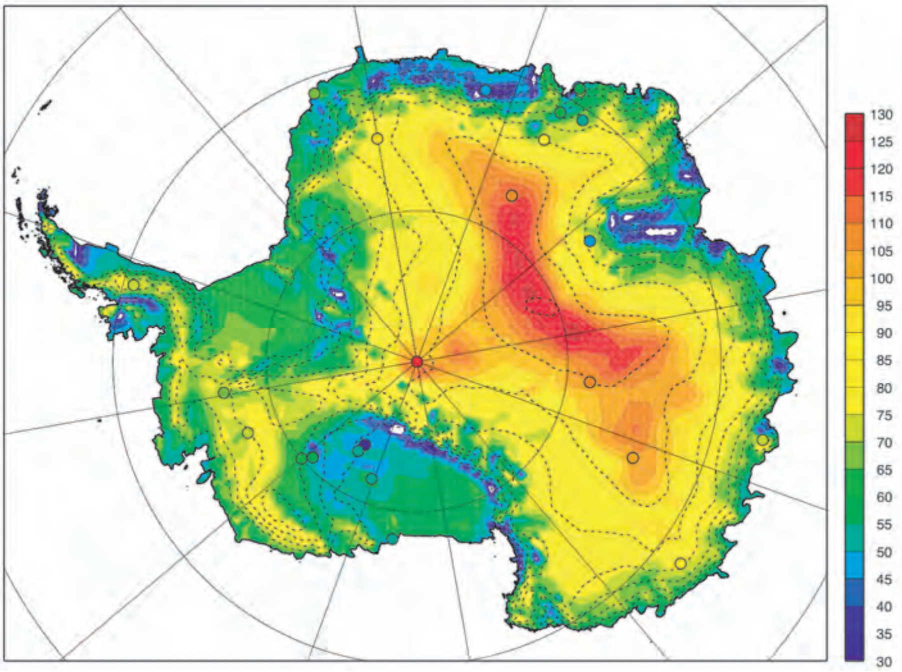

Figure 2. Observed (circles) and modelled depth of the firn layer h_f in m, taken to be the 830 kg m^{-3} density level.

example, if GLAS sensor onboard ICESAT can be used, but <5 – 10 m or 10 – 19% for this example if ERS1/2 data must used, [26]) and glacier velocity from radar interferometry (<5% uncertainty in column ice speed, [13]).

In the absence of significant melting, the steady-state firn densification rate depends mainly on temperature, accumulation rate and wind speed. Using a steady-state firn densification model [27] in combination with annual mean accumulation, surface temperature and near surface wind speed (1980-2004) from RACMO2/ANT we calculated the vertical density profile in Antarctica. As an example, Fig. 2 shows the resulting depth of the firn layer. Owing to the wide range of (near) surface climate conditions over the AIS, firn depth shows large spatial variability [10, 28, 29], which are well reproduced by the model. In the calm, dry and cold interior, densification is slow and the firn-layer thickness exceeds 100 m. In the windier, wetter and milder coastal zone, densification is more rapid and the firn layer shallower, typically 40 – 60 m. In regions with active katabatic winds and low precipitation rate, the firn layer may have been completely removed by snowdrift erosion and/or sublimation, exposing the glacier ice at the surface [25, 30].

From these results, a first-order estimate of the firn depth correction along the grounding line can be obtained, for use in solid ice flux calculations.

3.3 Elevation changes owing to temperature and accumulation variability

We used the same firn densification model, but in a time-dependent fashion, forced by 6-hourly temperature and accumulation rate from the regional atmospheric climate model RACMO2/ANT over the period 1980-2005. To minimize spin-up effects, the model was run for this 25-year period twice, but only the last 25 years were considered for analysis. As an example, Fig. 3 shows the results for an arbitrary location in coastal East Antarctica. This site has no melting in summer, moderately low temperatures (annual mean surface temperature 249.7 K), relatively high accumulation (522 kgm^{-2}

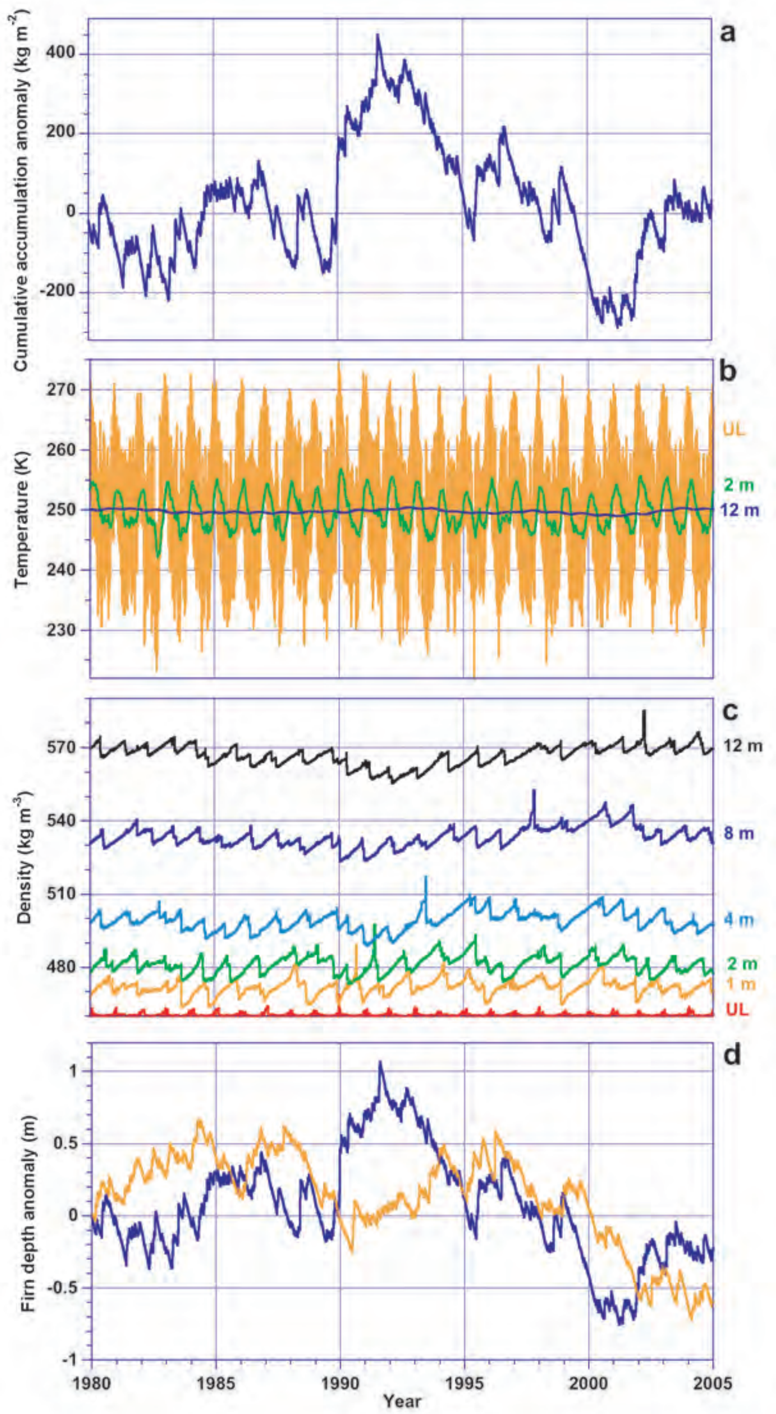

Figure 3. Modelled cumulative accumulation anomaly (a), Upper Layer (UL), 2 m and 12 m snow temperature (b), UL, 1 m, 2 m, 4 m, 8 m and 12 m snow density (c) and firn depth anomaly (d, blue line) for an arbitrary coastal points in East Antarctica. Yellow line in d) represents another arbitrary location in East Antarctica.

year^{-1}) and strong near-surface katabatic winds (annual mean 10 m wind speed 13.1 m s^{-1}). Fig. 3a shows the cumulative accumulation deviation from the mean. Clearly, decadal accumulation variability is present, causing cumulative deviations as large as 80% of the annual accumulation. Fig. 3b shows that seasonal temperature variations are only important in the upper 15 m of the firn. In response to this annual temperature wave the density also shows a clear annual cycle, with more rapid densification in summer (Fig. 3c). Interannual temperature variations also have an effect, especially warm summers. For instance, the anomalously high summer surface temperatures in 1989/90 (Fig. 3b) caused a sharp density peak that is seen to slowly travel downward in the firn, reaching a depth of 12 m in 2002.

Fig. 3d shows surface elevation change resulting from non-steady densification and surface accumulation variability. *If we assume a constant vertical motion at the firn-ice and ice-bedrock boundaries, this signal represents the observed vertical displacements at the ice sheet surface.* From comparing Fig. 3a and Fig. 3d it is clear that at this site, temporal accumulation variability is the dominant mechanism explaining variations in surface height, the influence of temperature effects being only of second order. Elevation trends in Fig. 3d are as large as 10 cm per year, at the high end of the range of values reported from remote sensing studies [10-12]. Curiously, the period that is characterized by a pronounced surface lowering coincides more or less exactly with the time frame in which the ERS-1 satellite was active and on which the estimates of studies 1-3 in Table 1 are based (1992-2002). Note that this trend is not representative for the full period (1980-2005), nor for a larger area: the yellow line in Fig. 3d represents another arbitrary site in East Antarctica, showing entirely different characteristics.

These preliminary calculations show that estimating AIS mass balance from satellite altimetry alone is undesirable; it should be combined with method ii). Once the above calculations have been applied over the whole ice sheet (>4000 gridpoints for the present configuration), the effects of firn densification and accumulation variability can be isolated and subtracted from the observed vertical velocity, yielding the real mass imbalance. This can then be compared to results of methods i) and ii).

4. SUMMARY AND CONCLUSIONS

The balance state of the Antarctic Ice Sheet (AIS) and its contribution to global sea level change are poorly known at present. Three methods are currently used to determine the AIS mass balance, two of which involve surface accumulation and the (time-space) structure of the firn layer. Major uncertainties in these methods are caused by i) the poorly known present-day surface accumulation distribution, ii) the poor constraint on the firn layer depth and density at the grounding line to convert elevation to ice thickness and iii) the poor understanding of accumulation variability and time-dependent firn densification in the interpretation of remotely-sensed surface elevation changes. Recently, some progress has been made in minimizing these uncertainties, using a mix of atmospheric and firn densification modelling. Several of these efforts have been presented in this paper.

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