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Authors
Rignot, Eric
Bamber, Jonathan L
van den Broeke, Michiel R
et al.

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Recent Antarctic ice mass loss from radar interferometry and regional climate modelling

ERIC RIGNOT1,2,3*, JONATHAN L. BAMBER4, MICHEL R. VAN DEN BROEKE5, CURT DAVIS6, YONGHONG LI6, WILLEM JAN VAN DE BERG6 AND ERIK VAN MEIJGAARD7

1University of California Irvine, Earth System Science, Irvine, California 92697, USA
2Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California 91109, USA
3Centro de Estudios Científicos, Arturo Prat 514, Valdivia, Chile
4University of Bristol, Bristol BS8 1SS, UK
5Institute for Marine and Atmospheric Research (IMAU), Utrecht University, 3584 CC Utrecht, The Netherlands
6University of Missouri-Columbia, Columbia, Missouri 65211, USA
7Royal Netherlands Meteorological Institute (KNMI), 3732 GK De Bilt, The Netherlands
*e-mail: erignot@uci.edu

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Large uncertainties remain in the current and future contribution to sea level rise from Antarctica. Climate warming may increase snowfall in the continent’s interior12, but enhance glacier discharge at the coast where warmer air and ocean temperatures erode the buttressing ice shelves4–11. Here, we use satellite interferometric synthetic-aperture radar observations from 1992 to 2006 covering 85% of Antarctica’s coastline to estimate the total mass flux into the ocean. We compare the mass fluxes from large drainage basin units with interior snow accumulation calculated from a regional atmospheric climate model for 1980 to 2004. In East Antarctica, small glacier losses in Wilkes Land and glacier gains at the mouths of the Filchner and Ross ice shelves combine to a near-zero loss of 4 ± 61 Gt yr−1. In West Antarctica, widespread losses along the Bellingshausen and Amundsen seas increased the ice sheet loss by 59% in 10 years to reach 132 ± 60 Gt yr−1 in 2006. In the Peninsula, losses increased by 140% to reach 60 ± 46 Gt yr−1 in 2006. Losses are concentrated along narrow channels occupied by outlet glaciers and are caused by ongoing and past glacier acceleration. Changes in glacier flow therefore have a significant, if not dominant impact on ice sheet mass balance.

The mass balance of Antarctica is determined from the difference between two competing processes of ice discharge into the ocean by glaciers and ice streams and accumulation of snowfall in the vast interior, which are two large numbers affected by significant uncertainties212. Estimates of ice discharge have been sporadic in nature owing to the limited availability of ice velocity and thickness data at the grounding line of Antarctica, as well as precise knowledge of the grounding-line positions. Similarly, estimates of snowfall have been affected by uncertainties associated with the interpolation of sparse in situ data of varying quality and temporal coverage over the entire continent.

Here, we present a nearly complete map of surface velocities along the periphery of Antarctica (Fig. 1) obtained from interferometric synthetic-aperture radar (InSAR) data collected between 1992 and 2006 by the European Earth Remote Sensing (ERS-1 and 2), the Canadian Radarsat-1 and the Japanese Advanced Land Observing satellites. Our map covers all major outlet glaciers, ice streams and tributaries of importance for mass flux calculation, with ice velocity ranging from 100 to 3,500 m yr−1, at a precision of 5 to 50 m yr−1 (see the Methods section). Short-time variations in velocity, for example, due to ocean tides, are averaged out over the 24 to 46 day repeat period of our measurements. Velocities at the grounding line of fast-moving glaciers are assumed to be depth independent, which introduces errors of much less than 1% (ref. 3).

Using double-difference interferometry, we mapped glacier grounding lines with a precision of 100 m all around Antarctica, except for eight glaciers south of 81° South where we used the Moderate Resolution Imaging Spectroradiometer (MODIS) mosaic13 with a precision of 1 km. Grounding-line thickness is derived from surface elevation assuming ice to be in hydrostatic equilibrium with sea water (see the Methods section). In selected parts of West Antarctica, we have direct measurements of ice thickness with a precision of 10 m instead (see Supplementary Information, Table S1). For surface elevation, we use a new digital elevation model (DEM) combining precise laser altimeter data from the Ice Cloud and land Elevation Satellite from 2003–2004, ERS-1/2 radar altimeter data from 1994 corrected for temporal changes in between13,14 and the new GGM02 geoid15. Comparison of the DEM with independent laser altimeter data at the grounding line of West Antarctica indicates a vertical precision in elevation of 0.15 ± 4 m. Surface elevation above mean sea level is then converted into solid-ice surface elevation after applying a firm depth correction16. We estimate the random error in inferred thickness to range from 80 to 120 m when accounting for uncertainties in grounding-line position, surface elevation, firm depth correction and geoid height. For verification, at the grounding line of Pine Island Bay glaciers, our thickness values are within 14 ± 60 m of direct thickness measurements16 ranging from 420 to 1,460 m.
Solid-ice fluxes are then calculated combining vector ice velocity and ice thickness, with a precision that is glacier dependent and ranges from 2 to 15% (see the Supplementary Information). The end points of the selected flux gates define the extent of the glacier drainage basins determined from the DEM. Individual drainage basins are grouped into large units labelled A–K.

Snowfall accumulation is from the RACMO2/ANT regional atmospheric climate model, at 55 km resolution, averaged for 1980–2004 (refs 17–19). Lateral forcings are taken from European Center for Medium-Range Weather Forecasting reanalyses (ERA-40) for the period 1980–2002, supplemented with European Center for Medium-Range Weather Forecasting operational analyses after August 2002. Comparisons with 1,900 independent field data show excellent agreement ($R = 0.82$) with the model 19. The model predicts higher coastal precipitation and wetter conditions in West Antarctica and the western Peninsula 20 than older maps obtained by interpolating limited field data using meteorological variables 21 or satellite passive microwave data 22. Few reliable in situ coastal accumulation data exist for comparison, but in the high-accumulation sector of the Getz Ice Shelf (basin FG), the model predicts precipitation levels consistent with a 2,030 mm yr$^{-1}$ record at Russkaya station (74°46′S, 136°52′W) for 1981–1989. Older maps yield accumulation levels 3 times lower, which imply a local mass balance 20 times more negative and high rates of glacier thinning that are not observed 2. The RACMO2/ANT accumulation values yield comparable losses for Pine Island and Thwaites glaciers, which is consistent with the similarity of their thinning rates; other maps yield twice more thinning for Thwaites. Finally, the model does not mix data from different time periods and fully incorporates temporal changes in snowfall between 1980 and 2004. A statistical analysis of absolute errors (see the Methods section) yields an uncertainty in accumulation varying from 10% in dry, large basins to 30% in wet, small coastal basins.

Ice flux and snowfall are compared for each glacier, for large basins A–K′, and for the Peninsula, East and West Antarctica. To include non-surveyed areas, we apply a scaling factor on the mass fluxes of each large basin A–K′ based on the percentage surveyed area versus total area to cover 100% of Antarctica (Table 1). In East Antarctica, we obtain a near-zero mass balance of $-4\pm 61$ Gt yr$^{-1}$. The J’K Filchner 22 and E’E Ross sectors are gaining mass, but this is compensated by the mass loss in Wilkes Land (basin CE) from the Philippi, Denman, Totten, Moscow University Ice Shelf, Cook Ice Shelf and David glaciers. Interestingly, all of these glaciers are marine based, that is, grounded well below sea level 2, and therefore

![Figure 1 Ice velocity of Antarctica colour coded on a logarithmic scale and overlaid on a MODIS mosaic](image_url)
more prone to instabilities. In West Antarctica, the well-known mass gain of the E’F’ Siple Coast basin is small compared with the combined mass loss from the F’G, HH and HH’ basins, which include the entire Amundsen and Bellingshausen sea coasts, and not just Pine Island Bay. The mass loss inferred from HH’ is much larger than in a previous survey that did not include many high-loss, small glaciers in the GF’ and HH’ basins and ongoing glacier acceleration in basin GH. Overall, the West Antarctic ice sheet lost 106 ± 60 Gt yr⁻¹ in the year 2000.

In the Antarctic Peninsula, the H’I and I’I basins of Palmer Land are near balance, despite a reported increase in snowfall, but basin II’ of Graham Land is out of balance. On the east coast, the Larsen A and B glaciers experienced an abrupt acceleration (300% on average) in 2002, which increased their mass loss from 3 ± 1 in 1996 and 2000, to 31 ± 9 Gt yr⁻¹ in 2006 (ref. 11). Farther south, airborne laser altimetry data suggest that the Larsen C glaciers are close to balance. But on the west coast, the glaciers have experienced widespread ice-front retreat, enhanced melt and continuous speed up. We have no thickness data for these glaciers, and there is no floating section. A 12% speed up in 10 years, enhanced melt and a net accumulation of 42 ± 14 Gt yr⁻¹ suggest a loss of 7 ± 4 Gt yr⁻¹ in 1996, 10 ± 5 Gt yr⁻¹ in 2000 and 13 ± 7 Gt yr⁻¹ in 2006. The combined loss for the Peninsula then becomes 25 ± 45 Gt yr⁻¹ in 1996, increasing by 140% in 2006 to 60 ± 46 Gt yr⁻¹ (Table 2).

Changes in surface elevation in basin FH’ for 1995–2005 (Fig. 2) reveal broad-scale, centimetre-level variations in snowfall in the interior (wetter conditions in H’H, and drier in F’E (ref. 17)), but pronounced, metre-scale thinning concentrated in narrow channels occupied by outlet glaciers and extending in the flow direction across the entire coastal range. The strong, widespread correlation between ice thinning and ice velocity (>50 m yr⁻¹), for example, on the Berg, Ferrigno, Venable, Pine Island, Thwaites, Smith and Getz glaciers, indicates that thinning is caused by the velocity of glaciers being well above that required to maintain mass balance, that is, ice stretches longitudinally, which causes it to thin vertically. In basin GH, we find that Pine Island Glacier accelerated 34% in 1996–2006, Smith 75%, Pope 20%, Hayes 27% and Thwaites is widening. The mass flux from basin GH thereby increased 21% since 1996 and the mass loss doubled from 41 ± 27 Gt yr⁻¹ in 1996 to 64 ± 27 Gt yr⁻¹ in 2000 and 90 ± 27 Gt yr⁻¹ in 2006 (Table 2). This is the largest loss in Antarctica. In contrast, we detect no glacier acceleration in basins

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**Table 1 Mass balance of Antarctica in gigatonnes (10¹² kg) per year for the year 2000. Area: area surveyed. Input: snow accumulation ±σ of surveyed area. Outflow: grounding-line ice flux ±σ of surveyed area. Net: mass balance calculated as Input minus Outflow, ±σ. Net+: mass balance scaled on the basis of total area (Area+) versus area surveyed (Area), except for basin II’ where we use refs 8,9. Input+: total snowfall in Area+. Mass losses for 1996 and 2006 differ from those in the year 2000 in basins GH and II’ (see Table 2).**

| Sector | Area (10⁶ km²) | Input (Gt yr⁻¹) | Outflow (Gt yr⁻¹) | Net (Gt yr⁻¹) | Net+ (Gt yr⁻¹) | Area+ (10³ km²) | Input+ (Gt yr⁻¹) |
|--------|----------------|------------------|-------------------|--------------|---------------|----------------|-----------------|
| J’K Filchner | 1,698 | 93 ± 8 | 75 ± 4 | 18 ± 9 | 19 ± 10 | 1,780 | 100 ± 9 |
| KK Riser Larsen | 218 | 42 ± 8 | 45 ± 4 | −3 ± 9 | −3 ± 11 | 246 | 50 ± 10 |
| K’A Jutulstraumen | 159 | 26 ± 7 | 28 ± 2 | −1 ± 8 | −1 ± 9 | 178 | 32 ± 9 |
| AA Queen Maud Land | 615 | 60 ± 9 | 60 ± 7 | 0 ± 11 | 0 ± 12 | 622 | 62 ± 9 |
| A’B Enderby Land | 354 | 39 ± 5 | 40 ± 2 | −1 ± 5 | −1 ± 9 | 645 | 115 ± 14 |
| BC Lambert | 1,197 | 73 ± 10 | 77 ± 4 | −4 ± 11 | −4 ± 12 | 1,332 | 87 ± 12 |
| CC Philippi, Denman | 434 | 81 ± 13 | 87 ± 7 | −7 ± 15 | −11 ± 24 | 702 | 137 ± 22 |
| C’D Totten, Frost | 1,053 | 198 ± 37 | 207 ± 13 | −8 ± 39 | −9 ± 43 | 1,162 | 261 ± 49 |
| DD Cook, Mertz, Ninnis | 563 | 92 ± 14 | 94 ± 6 | −2 ± 16 | −2 ± 19 | 691 | 136 ± 21 |
| D’E Victoria Land | 267 | 20 ± 1 | 22 ± 3 | −2 ± 4 | −3 ± 6 | 450 | 62 ± 4 |
| EE Transantarctic | 1,441 | 61 ± 10 | 49 ± 4 | 11 ± 11 | 13 ± 13 | 1,639 | 89 ± 15 |
| East Antarctica 2000 | 7,998 | 786 ± 48 | 785 ± 20 | 1 ± 52 | −4 ± 61 | 9,447 | 1,131 ± 69 |
| E’F’ Siple Coast | 751 | 110 ± 7 | 80 ± 2 | 31 ± 7 | 34 ± 8 | 845 | 130 ± 8 |
| F’G Getz, Hall, Land | 119 | 108 ± 28 | 128 ± 18 | −19 ± 33 | −23 ± 39 | 140 | 128 ± 33 |
| GH Pine Is., Thwaites | 393 | 177 ± 25 | 237 ± 4 | −61 ± 26 | −64 ± 27 | 417 | 196 ± 28 |
| HH’ Ferrigno, Abbot | 55 | 51 ± 16 | 86 ± 10 | −35 ± 19 | −49 ± 27 | 78 | 71 ± 22 |
| JJ’ Ronne | 933 | 142 ± 11 | 145 ± 7 | −4 ± 13 | −4 ± 14 | 1,028 | 165 ± 13 |
| West Antarctica 2000 | 2,251 | 588 ± 49 | 676 ± 22 | −88 ± 54 | −106 ± 60 | 2,508 | 690 ± 57 |
| H’I English Coast | 92 | 71 ± 21 | 78 ± 7 | −7 ± 23 | −7 ± 24 | 98 | 77 ± 23 |
| II’ Graham Land | 13 | 15 ± 5 | 20 ± 3 | −5 ± 6 | −15 ± 8 | 78 | 125 ± 46 |
| I’I East Palmer Land | 11 | 8 ± 4 | 9 ± 2 | −1 ± 4 | −6 ± 18 | 52 | 32 ± 14 |
| Antarctic Peninsula 2000 | 116 | 94 ± 21 | 107 ± 8 | −13 ± 23 | −28 ± 45 | 228 | 234 ± 53 |
| Antarctica 2000 | 10,365 | 1,469 ± 87 | 1,568 ± 31 | −100 ± 78 | −138 ± 92 | 12,183 | 2,055 ± 122 |

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**Table 2 Mass balance in gigatonnes (10¹² kg) per year for 1996 and 2006 of basins II’ and G H, West Antarctica, the Peninsula and the entire Antarctic ice sheet.**

| Sector | Outflow (Gt yr⁻¹) | Net (Gt yr⁻¹) | Net+ (Gt yr⁻¹) |
|--------|------------------|--------------|---------------|
| GH Pine Is. Thwaites 1996 | 215 ± 3 | −39 ± 25 | −41 ± 27 |
| GH Pine Is. Thwaites 2006 | 261 ± 4 | −85 ± 26 | −90 ± 27 |
| West Antarctica 1996 | 654 ± 22 | −66 ± 53 | −83 ± 59 |
| West Antarctica 2006 | 700 ± 23 | −112 ± 54 | −132 ± 60 |
| II’ Graham Land 1996 | 20 ± 3 | −5 ± 6 | −12 ± 7 |
| II’ Graham Land 2006 | 49 ± 3 | −34 ± 6 | −47 ± 9 |
| Peninsula 1996 | 107 ± 8 | −13 ± 23 | −25 ± 45 |
| Peninsula 2006 | 136 ± 10 | −42 ± 24 | −60 ± 46 |
| Antarctica 1996 | 1,546 ± 30 | −78 ± 78 | −112 ± 91 |
| Antarctica 2006 | 1,621 ± 32 | −153 ± 78 | −196 ± 92 |
CDW onto the continental shelf and trigger glacier acceleration, but this hypothesis cannot be confirmed at present.

Our results provide a nearly complete assessment of the spatial pattern in mass flux and mass change along the coast of Antarctica, glacier by glacier, with lower error bounds than in previous incomplete surveys, and a delineation of areas of changes versus areas of near stability. Over the time period of our survey, the ice sheet as a whole was certainly losing mass, and the mass loss increased by 75% in 10 years. Most of the mass loss is from Pine Island Bay sector of West Antarctica and the northern tip of the Peninsula where it is driven by ongoing, pronounced glacier acceleration. In East Antarctica, the loss is near zero, but the thinning of its potentially unstable marine sectors calls for attention. In contrast to major increases in ice discharge, snowfall integrated over Antarctica did not change in 1980–2004 (ref. 27) and even slightly increased in areas of large loss17. We conclude that the Antarctic ice sheet mass budget is more complex than indicated by the temporal evolution of its surface mass balance. Changes in glacier dynamics are significant and may in fact dominate the ice sheet mass budget.

**METHODS**

**FRN DEPTH CORRECTION**

Ice thickness, \( H \), is deduced from surface elevation above mean sea level with reference to the GGM02 geoid45, \( h = H + (h \cdot \Delta H) \cdot \rho_{\text{sea}} / (\rho_{\text{sea}} - \rho_{\text{ice}}) \), where the density of sea water, \( \rho_{\text{sea}} = 1,028 \text{ kg m}^{-3} \), and at 34 p.s.u. salinity, 1 km depth, the density of solid ice, \( \rho_{\text{ice}} = 917 \text{ kg m}^{-3} \), and \( \Delta H \) is the first depth correction. For \( H = 1 \text{ km} \) a 4 m uncertainty in \( \Delta H \) introduces a 4% uncertainty in thickness and flux. Earlier work assumed a constant firn depth correction. We calculate \( \Delta H \) from a firm densification model19 driven by surface density using 25-year-average air temperature, snow accumulation and wind speed from RACMO2/ANT. \( \Delta H \) varies from 0 to 20 m. Its precision is 2–3 m on the basis of a comparison with firm core data at the critical densities of 350 and 880 kg m\(^{-3}\).

**SNOW ACCUMULATION ERROR**

Snow accumulation is the arithmetic average of the values given in refs 18,19. We use 1,900 in situ independent observations, \( \text{SMB}_{\text{obs}} \), to calculate absolute errors. The error for the observations is modelled as \( \text{SMB}_{\text{obs}} = 5 \pm 0.15 \text{ SMB}_{\text{obs}} \) in kg m\(^{-2}\) yr\(^{-1}\) where the second term accounts for the uncertainty associated with spatial variability. The error for the modelled values, \( \text{SMB}_{\text{mod}} \), is modelled as \( \text{SMB}_{\text{mod}} = 9 \pm 0.10 \text{ SMB}_{\text{mod}} + 0.0003 \text{ SMB}_{\text{mod}} \in 	ext{kg m}^{-2} \text{yr}^{-1} \). This representation reflects that the model is well calibrated with many good observations for low and medium values, but that the relative and absolute errors increase for high values where few reliable observations exist. The relative error is maximized at 30% for \( \text{SMB}_{\text{obs}} > 557 \text{ kg m}^{-2} \text{yr}^{-1} \). Coefficients for the modelling of \( \text{E}_{\text{sn}} \) were optimally chosen by examining the distribution of differences, \( \text{SMB}_{\text{obs}} - \text{SMB}_{\text{mod}} \), normalized by the total error margin, that is, the squared sum of the error for the observations and the modelled values, \( \text{SMB}_{\text{obs}} - \text{SMB}_{\text{mod}} \), divided by the error for the modelled values. With our selection of coefficients, we obtain a normal distribution with \( \sigma = 1 \), which provides strong statistical support for the error analysis. To calculate accumulation uncertainty at the basin scale, we also account for the spatial autocorrelation of errors. The correlation length of (SMB\(_{\text{obs}}\)–SMB\(_{\text{mod}}\)) varies from 161 km below 2,000 m elevation to 300 km above 2,000 m. Combining these correlation lengths with the error modelling, we obtain total errors in AK basins (Table 1) ranging from 10% in large, dry basins to 30% in wet and smaller coastal basins. These errors represent our most likely estimate of absolute errors, not the 95% confidence interval. Previous attempts at defining accumulation errors only addressed interpolation errors.

**VELOCITY AND MASS FLUX ERRORS**

Ice velocity is measured with speckle tracking on Radarsat-1 24 day, Japanese Advanced Land Observing PALSAR 46 day (basin GF) and ERS-1 9 day (basin H'I') repeats, and interferometrically using ascending/descending ERS-1/2 tandem pairs (basin HG) with an ERS-1/2 precision of 2–5 m yr\(^{-1}\); and with a combination of interferometric phase and speckle tracking (basin D'C), with a precision of 20–50 m yr\(^{-1}\). Systematic errors are negligible compared with random errors because we use stagnant areas for calibration and combine multiple tracks with different look directions. The unknown positive bias between surface and vertically integrated velocity is much less than 1%.

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**Figure 2** Changes in surface elevation (centimetres per year) for basin F'H'. Values were derived from ERS-2 and Envisat radar altimeter data for 1995–2005 and colour coded from red (thinning) to blue (thickening), overlaid on 50 m yr\(^{-1}\) velocity contours in green. Areas of fast flow in F'H' correspond to areas of concentrated thinning, in contrast with other glaciers.
Systematic errors in thickness are less than 1%, whereas random errors range from 10 to 120 m (see Supplementary Information, Table S1). The percentage error in mass flux is calculated as the sum of the percentage error in velocity and the percentage error in thickness. This is appropriate for plug flow or U-shaped velocity profiles, which is the case for most large Antarctic glaciers. For glaciers that approach a V-shaped velocity profile, our errors may be underestimated by a factor of 2. Errors in Table 1 and Supplementary Information, Table S1 are only random errors. Systematic errors are not known but small, so that actual errors may be slightly higher. Decadal changes in velocity were only available for glaciers mentioned in the text, ref. 11 or Table 2, and were assumed to be zero elsewhere.

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Correspondence and requests for materials should be addressed to E.R. Supplementary Information accompanies this paper on www.nature.com/naturegeoscience.

Author contributions

All authors discussed the results and commented on the manuscript. J.R.G. led the remote sensing analysis, development of the paper and integration of the results, J.L.B. provided a digital elevation model and analysed its accuracy, M.R.B., W.J.B. and E.M. contributed calculations of snow accumulation, firn depth correction and associated errors and C.D. and Y.L. analysed elevation changes from satellite radar altimeter data.

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