Mass changes of the Greenland and Antarctic ice sheets and shelves and contributions to sea-level rise: 1992–2002

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ABSTRACT. Changes in ice mass are estimated from elevation changes derived from 10.5 years (Greenland) and 9 years (Antarctica) of satellite radar altimetry data from the European Remote-sensing Satellites ERS-1 and -2. For the first time, the d*H*/d*t* values are adjusted for changes in surface elevation resulting from temperature-driven variations in the rate of firn compaction. The Greenland ice sheet is thinning at the margins (-42 ± 2 Gt a^{-1} below the equilibrium-line altitude (ELA)) and growing inland ($+53 \pm 2$ Gt a^{-1} above the ELA) with a small overall mass gain ($+11 \pm 3$ Gt a^{-1}; -0.03 mm a^{-1} SLE (sea-level equivalent)). The ice sheet in West Antarctica (WA) is losing mass (-47 ± 4 Gt a^{-1}) and the ice sheet in East Antarctica (EA) shows a small mass gain ($+16 \pm 11$ Gt a^{-1}) for a combined net change of -31 ± 12 Gt a^{-1} ($+0.08$ mm a^{-1} SLE). The contribution of the three ice sheets to sea level is $+0.05 \pm 0.03$ mm a^{-1}. The Antarctic ice shelves show corresponding mass changes of -95 ± 11 Gt a^{-1} in WA and $+142 \pm 10$ Gt a^{-1} in EA. Thinning at the margins of the Greenland ice sheet and growth at higher elevations is an expected response to increasing temperatures and precipitation in a warming climate. The marked thinnings in the Pine Island and Thwaites Glacier basins of WA and the Totten Glacier basin in EA are probably ice-dynamic responses to long-term climate change and perhaps past removal of their adjacent ice shelves. The ice growth in the southern Antarctic Peninsula and parts of EA may be due to increasing precipitation during the last century.

INTRODUCTION

The mass balances of the Greenland and Antarctic ice sheets are of interest because of their complex linkage to climate variability and their direct effects on sea-level change. In recent decades, the spatial distribution of mass input and output data has greatly improved as field observations have been complemented by advances in remote sensing and dynamic modeling. Approximately 399 Gt a^{-1} of ice is accumulated on the Greenland ice sheet above the equilibrium line, and approximately 1637 Gt a^{-1} on the Antarctic ice sheet (modified from Giovinetto and Zwally, 2000; Zwally and Giovinetto, 2001), which is equivalent to the removal of 5.6 mm a^{-1} from the oceans. The net mass balance is the difference between the mass input in the zone of net accumulation and the sum of the net ablation at the surface (including runoff), the direct discharge of ice into the ocean, and discharge of subglacial water across the grounding line. Uncertainties in previous mass-balance estimates (± 53.0 Gt a^{-1} for Greenland and ± 384 Gt a^{-1} for Antarctica (Huybrechts and others, 2001)) have been largely due to the difficulty of accurately measuring all the mass input and output fluxes (e.g. Rignot and Thomas, 2002).

Expected responses of the ice sheets to climate warming are both growth in thickness of the inland ice areas, due to increasing precipitation, and thinning near the margins, due to increasing surface melting (Huybrechts and others, 2001). In addition, dynamic ice thinning near the margins may be induced by processes such as removal or thinning of adjacent ice shelves or ice tongues (Thomas, 2003; Rignot and others, 2004; Scambos and others, 2004) as well as enhanced basal sliding due to surface meltwater reaching the ice–bedrock interface (Zwally and others, 2002c). Alley and others (2003) reviewed the state of knowledge of ice-sheet behavior from recent observational and modeling advances and suggested that the ice sheets may have a greater sensitivity to climate warming than previously considered.

Since the first results using altimeter surveys of ice-sheet elevation changes to estimate changes in ice volume and mass balance (Zwally, 1989), there has been an increasing use of altimetric measurements of elevation change (d*H*/d*t*) to improve upon estimates from mass-flux studies. Recent results include the detection of significant thinning of the margins of the Greenland ice sheet, which was attributed to increases in both melting and dynamic thinning (Abdalati and others, 2001; Krabill and others, 2004), thickening of the EA ice sheet, attributed to increases in precipitation (Davis and others, 2005), and growth of the interior of Greenland (Johannessen and others, 2005). Nevertheless, a comprehensive assessment of the current mass balance of the ice sheets has not been made, due in part to the limited performances of satellite radar altimeters over the steeper ice-sheet margins and in part to limitations in the spatial and temporal coverage of airborne laser-altimeter surveys. Elevation changes are also caused by temporal variations in the rate of firn compaction (Zwally, 1989; Braithwaite, 1994; Arthern and Wingham, 1998; Zwally and Li, 2002), but corrections for this effect have not previously been made.

In this study, we extend the analysis of radar altimeter data from the two European Remote-sensing Satellites (ERS-1 and -2) to 90.0% of the Greenland ice sheet, 77.1% of the Antarctic ice sheet and 81.8% of the Antarctic ice shelves. In Greenland, we use results of Airborne Topographic Mapper

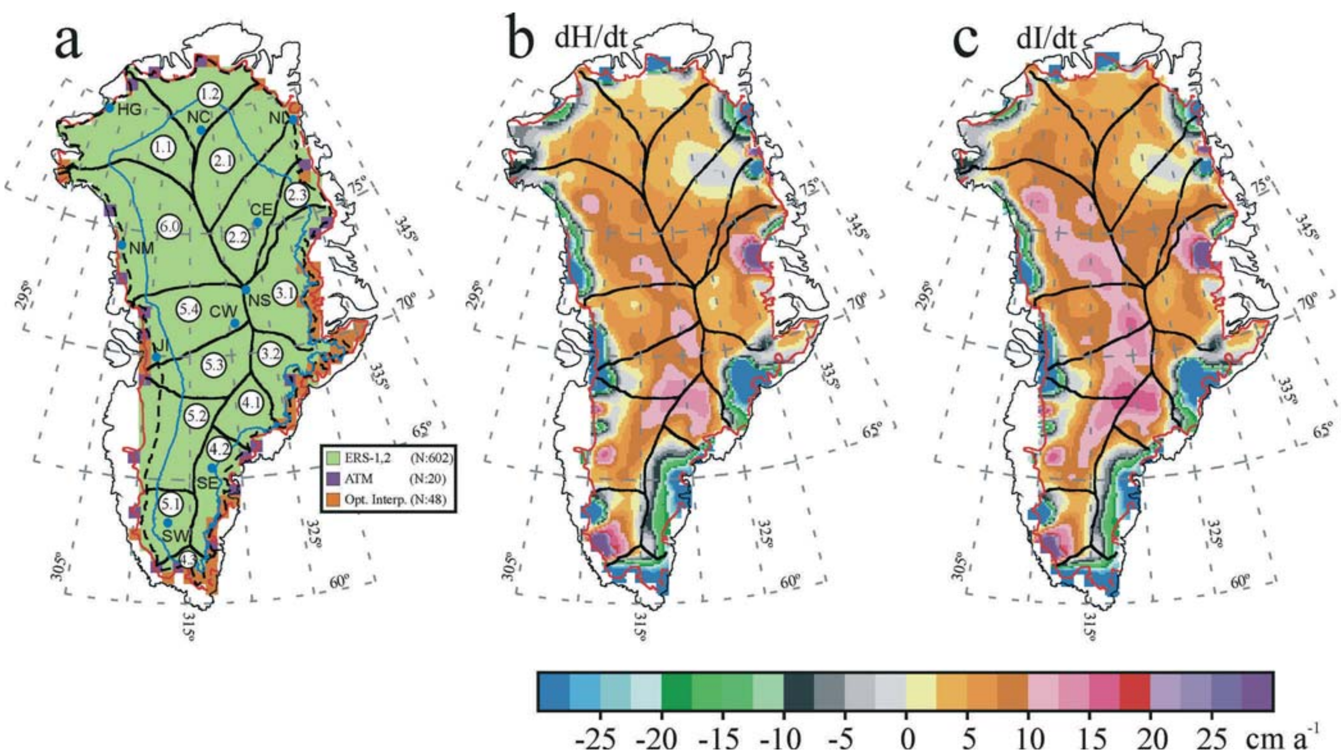

Fig. 1. Greenland. (a) Distribution of surface elevation change data by source, derived from ERS-1 and -2 radar altimetry, ATM (closest-neighbor interpolation from airborne surveys), and obtained by optimal interpolation: ice terminus of coterminous ice sheet (red), equilibrium line (black dashes), 2000 m elevation contour (blue), drainage divides (black), drainage system designation (number in circles), and location of $H(t)$ series depicted in Figure 3a (labeled blue full circles). (b) Distribution of elevation change (dH/dt). (c) Distribution of ice-thickness change (dI/dt).

(ATM) laser-altimeter surveys to increase coverage of the margins. We also use optimal-interpolation procedures to provide nearly complete spatial coverage of the ice sheets and shelves (Figs 1a and 2a). For the first time, we use a firn compaction model with a 20 year record of satellite-based surface temperatures to calculate corrections for elevation changes due to changes in the rate of firn compaction caused by temporal variations in firn temperature and near-surface melting.

OBSERVED AND INTERPOLATED ELEVATION CHANGES (dH/dt)

Measurements of ice surface elevations (H) from ERS-1 and -2 radar altimeters are compiled as elevation time series, $H(t)$, from which elevation change (dH/dt) values are derived (Fig. 3). The ERS altimeters operated in either ocean mode with a resolution of 45 cm per range gate or ice mode with a resolution of 182 cm per range gate. We use only ice mode data, because of a spatially variant bias between the modes and the greater spatial and temporal coverage of the ice mode data. We applied our V4 range-retracking algorithm, atmospheric range corrections, instrument corrections, slope corrections and an adjustment for solid tides (Zwally and Brenner, 2001). Instrument corrections include subtraction of a 40.9 cm bias from ERS-1 elevations to account for a different instrument parameter used for ERS-2 (Femenias, 1996) and corrections for drifts in the ultra-stable oscillator and bias changes in the scanning point target response that are obtained from the European Space Agency. We use the DUT DGM-E04 orbits, which have a radial orbit precision of 5–6 cm (Scharroo and Visser, 1998).

The time periods are from mid-April 1992 to mid-October 2002 for Greenland and to mid-April 2001 for Antarctica. The series are constructed for gridpoints nominally 50 km apart from sets of elevation differences measured at orbital crossovers using time periods of 91 days. Our methods enable us to obtain useful $H(t)$ series over more of the ice-sheet area than some other analyses have. For most gridpoints, crossovers within a 100 km circle centered on the gridpoint are used, and within 200 km for a few points. Crossover selection is also limited to elevations within ± 250 m of the elevation at the gridpoint center, which for slopes >1/200 restricts the selected areas to bands along elevation contours where the elevation changes tend to be spatially coherent. The first sequence of elevation differences at crossings between period T_1 and all successive T_i is combined with the second sequence of those for crossings between T_2 and all successive T_i and so forth for all additional sequences, which are then combined in one $H(t)$ (Zwally and Brenner, 2001). The resulting $H(t)$ series use all independent crossovers, including inter-satellite crossovers, which greatly increases the number of crossovers and the accuracy of the results (more crossovers and longer time intervals), as compared to using only the first sequence or only intra-satellite crossovers. For Greenland, we have 16×10^6 crossovers from ERS-1/ERS-1, 52×10^6 from ERS-2/ERS-2, and 59×10^6 from ERS-1/ERS-2. For Antarctica, we have 157×10^6 crossovers from ERS-1/ERS-1, 276×10^6 from ERS-2/ERS-2, and 419×10^6 from ERS-1/ERS-2, whereas only crossovers from ERS-1/ERS-1 and ERS-2/ERS-2 are used by Davis and others (2005).

We obtain dH/dt values for 602 (90%) of the 670 gridpoints on Greenland ice and 4085 (79%) of the 5175