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Changes in ice dynamics and mass balance of the Antarctic ice sheet

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The concept that the Antarctic ice sheet changes with eternal slowness has been challenged by recent observations from satellites. Pronounced regional warming in the Antarctic Peninsula triggered ice shelf collapse, which led to a 10-fold increase in glacier flow and rapid ice sheet retreat. This chain of events illustrated the vulnerability of ice shelves to climate warming and their buffering role on the mass balance of Antarctica. In West Antarctica, the Pine Island Bay sector is draining far more ice into the ocean than is stored upstream from snow accumulation. This sector could raise sea level by 1 m and trigger widespread retreat of ice in West Antarctica. Pine Island Glacier accelerated 38% since 1975, and most of the speed up took place over the last decade. Its neighbour Thwaites Glacier is widening up and may double its width when its weakened eastern ice shelf breaks up. Widespread acceleration in this sector may be caused by glacier ungrounding from ice shelf melting by an ocean that has recently warmed by 0.3 °C. In contrast, glaciers buffered from oceanic change by large ice shelves have only small contributions to sea level. In East Antarctica, many glaciers are close to a state of mass balance, but sectors grounded well below sea level, such as Cook Ice Shelf, Ninnis/Mertz, Frost and Totten glaciers, are thinning and losing mass. Hence, East Antarctica is not immune to changes.

Keywords: Antarctica; interferometry; Southern Ocean; polar climate

1. Introduction

The evolution of the Antarctic ice sheet is a problem of considerable societal importance because of its influence on global climate and sea level (Church *et al.* 2001). Ice sheets have taken tens of thousands of years to form and we have grown accustomed to the conservative view that they may change with eternal slowness. The observational record of Antarctic glaciers is short and limited because humans have seldom visited many Antarctic glaciers (Swithinbank *et al.* 1988). In the last decade, the advent of satellite missions has revolutionized our knowledge of ice dynamics in Antarctica. These tools have provided observations of ice motion in most places for the first time, and the measurements have

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One contribution of 14 to a Discussion Meeting Issue 'Evolution of the Antarctic Ice Sheet: new understanding and challenges'.

already been repeated over the last decade. The results reveal glaciological changes of surprising rapidity and magnitude (Rignot & Thomas 2002).

Changes in the Antarctic Peninsula have been particularly significant because of pronounced regional warming which triggered ice shelf collapse (King 1994). The results have important implications for the future of Antarctica in a warmer climate. Further south, where air temperatures are below freezing and glaciers drain large sectors, important glaciological changes have been detected as well, in areas suspected to be vulnerable to rapid change (Rignot 1998; Shepherd *et al.* 2001). The Pine Island Bay sector of West Antarctica is one of these sectors, with the potential to raise global sea level by 1 m and trigger widespread retreat of West Antarctic ice.

East Antarctica rests for the most part on bedrock grounded above sea level, and hence is more stable. Observations of its evolution have been less complete and less conclusive than in West Antarctica (Wingham *et al.* 1998; Rignot & Thomas 2002). Recent observations of changes in surface elevation, however, suggest that it is not immune to changes (Davis *et al.* 2005).

Here, I review observations of changes in ice dynamics gathered in the last decade in West and East Antarctica, mostly from radar satellite remote sensing. The review includes an update of prior results based on data acquired in 2005. I discuss how the observed glacier changes may relate to climate, what we have learned from them, what are the remaining uncertainties and what the implications are for the contribution to sea-level rise from the Antarctic ice sheet.

2. Methodology and datasets

(*a*) *Satellite techniques*

Until the 1980s, ice velocity was measured at discrete locations by arduous field expeditions far from the grounding zone (Swithinbank *et al.* 1988). The advent of optical satellites, such as Landsat, allowed researchers to measure ice velocity from space over large areas using image processing techniques (Bindschadler & Scambos 1991). Few data were available, however, for a comprehensive survey of Antarctic glaciers. In 1991, the European Space Agency launched European remote sensing (ERS-1) satellite, a synthetic aperture radar satellite that imaged Antarctica down to 81° south until year 2000. It was followed by ERS-2 in 1995, which was flown in tandem with ERS-1 in 1995–1996. In 1995, the Canadian Space Agency launched Radarsat-1, which provided the first complete coverage of Antarctica (Jezek 1999), and a demonstration of radar interferometry (interferometric synthetic-aperture radar, InSAR) with a 24-day repeat cycle in 1997. In 2000, Radarsat-1 collected a comprehensive set of interferometric data down to 80° south, with sporadic data acquisitions thereafter (Jezek *et al.* 2003).

ERS-1/2 InSAR measures the displacement of the ice surface in the radar-looking direction (Goldstein *et al.* 1993). Three look directions are required to measure ice velocity in vector form. In practice, two crossing tracks are combined with the assumption that ice flows parallel to the ice surface, which is a reasonable approximation in coastal Antarctica (Joughin *et al.* 1998). The precision in velocity mapping varies with data quality, but typically reaches 10–20 m yr^{-1}.

In complement, speckle tracking measures ice velocity in vector form as fractions of image pixels in the along and across track directions (Michel & Rignot 1999). The precision of this technique exceeds 1/128th of a pixel, which corresponds to a velocity of 10–20 m yr^{-1} with Radarsat-1 24-day repeat data.

One chief advantage of ERS-1/2 InSAR is its ability to locate glacier grounding lines (Rignot 1996). The grounding line is where a glacier reaches the ocean and becomes afloat. It is the natural boundary where ice fluxes to the ocean ought to be calculated for estimating the state of mass balance of an ice sheet. Downstream of the grounding line, ice melting by the ocean is vigorous enough to remove a large fraction of the volume of ice (Rignot & Jacobs 2002). Near the grounding line, the assumption that ice is in hydrostatic equilibrium with the ocean waters yields estimates of ice thickness from satellite measurements of surface elevation.

ERS-1/2 and Radarsat-1 InSAR have revealed grounding line positions with a precision two orders of magnitude better than in the past, i.e. within 100 m instead of tens of kilometres. On slow-moving areas, surface features provided a reasonable description of those positions, but on outlet glaciers or stream flow biases as large as 150 km on Lambert Glacier have been found (Rignot 2002*b*). This means that grounding line ice thickness is often greater than presumed, ice discharge is larger and glacier mass balance is more negative (Rignot & Thomas 2002).

(*b*) *Velocity mapping*

Figure 1 shows a compilation of Antarctic ice velocity derived from ERS-1/2 1996 and Radarsat-1 2000 data processed by the author and I. Joughin. The mapping is sparse in the interior, but coastal sectors are well sampled and more relevant for mass balance assessments. This is because ice is not flowing uniformly toward the coast but is channelized through narrow gateways occupied by ice streams and glaciers. The gateways are several tens of kilometres wide, 1–3 km thick, with ice velocity ranging from a few hundred metres to 3 km yr^{-1}. The glaciers occupy 10% of the coastline but drain 90% of the ice sheet. Most West Antarctic glacier velocities have been measured. In East Antarctica, mapping is complicated by a reduced coverage from ERS-1/2 and the lower signal coherence of Radarsat-1 data.

Pine Island Bay (figure 2), Siple Coast, and the Lambert basin have been mapped in most detail using InSAR. In Pine Island Bay, the arborescent structure of tributaries illustrates the complex interconnectivity of ice flow from the coast to the deep interior (Bamber *et al.* 2000). Perturbations initiating at the coast may affect the entire drainage basin more rapidly than if ice flow were uniform. Many glacier systems exhibit similar complexities, but Pine Island Bay and Siple Coast are among the most complex.

Figure 3 shows the first velocity map of the Antarctic Peninsula derived from Radarsat-1 data collected in year 2000. Prior to this map, velocity measurements were limited to few glaciers. Drainage basins are smaller than in West Antarctica and with a fixed limit imposed by the steep sides of the central spine of the Peninsula in its northern sector, Graham Land. The fishbone structure of Graham Land illustrates that most ice present at the last glaciation is now gone. Flow velocities are larger on the west where few ice shelves remain than on

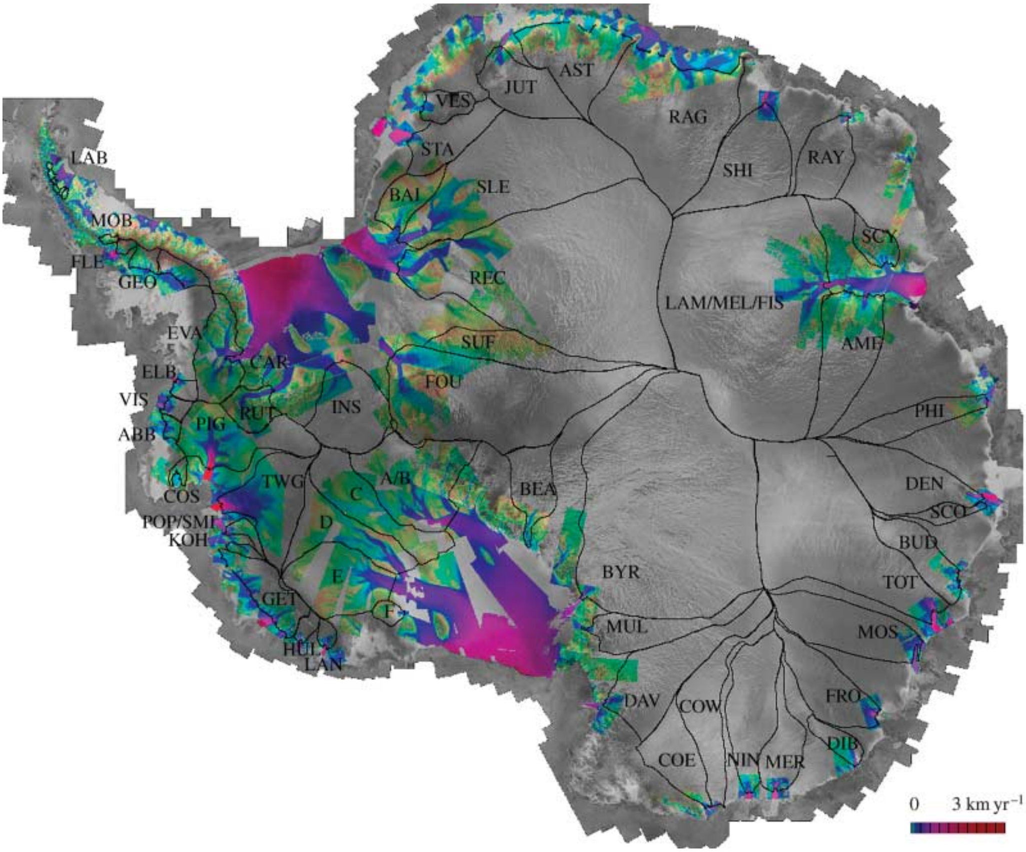


Figure 1. Antarctic ice velocity derived from ERS-1/2 and Radarsat-1 data overlaid on Antarctic Mapping Mission (AMM)-1 Radarsat-1 mosaic at 5 km spacing. Velocities for Ross, Filchner, Ronne and Amery ice shelves, and Siple Coast ice streams are from I. Joughin; others are from the author. Catchment basin boundaries are black. Abbreviations for glaciers/ice streams are as follows: Larsen B (LAB), Mobil Oil Inlet (MOB), Fleming (FLE), George VI (GEO), Eltanin Bay (ELB), Venable Ice Shelf (VIS), Abbot Ice Shelf (ABB), Cosgrove Ice Shelf (COS), Pine Island (PIG), Thwaites (TWG), Pope/Smith (POP/SMI), Kohler (KOH), Getz (GET), Hull (HUL), Land (LAN), Siple Coast ice streams (A–F), Beardmore (BEA), Byrd (BYR), Mulock (MUL), David (DAV), east and west Cook Ice Shelf (COE, COW), Ninnis (NIN), Mertz (MER), Dibble (DIB), Frost (FRO), Moscow University ice shelf (MOS), Totten (TOT), Budd coast (BUD), Scott (SCO), Denman (DEN), Philippi (PHI), American Highland (AME), Lambert/Mellor/Fisher (LAM), Scylla (SCY), Rayner (RAY), Shirase (SHI), Princess Ragnhild (RAG), Princess Astrid (AST), Jutulstraumen (JUT), Veststraumen (VES), Stancomb-Wills (STA), Bailey (BAI), Slessor (SLE), Recovery (REC), Support-Force (SUF), Foundation (FOU), Institute (INS), Rutford (RUT), Carlson (CAR) and Evans (EVA).

the east coast where ice shelves are present, and of similar magnitude when ice shelves are absent. This similarity in glacier velocity is surprising, given the strong east–west gradient in accumulation (Turner *et al.* 2002), and is probably a result of present-day mass balance conditions. Glacier velocities along George VI Ice Shelf and toward Larsen D and E are comparatively low, only a few hundred metres per year.

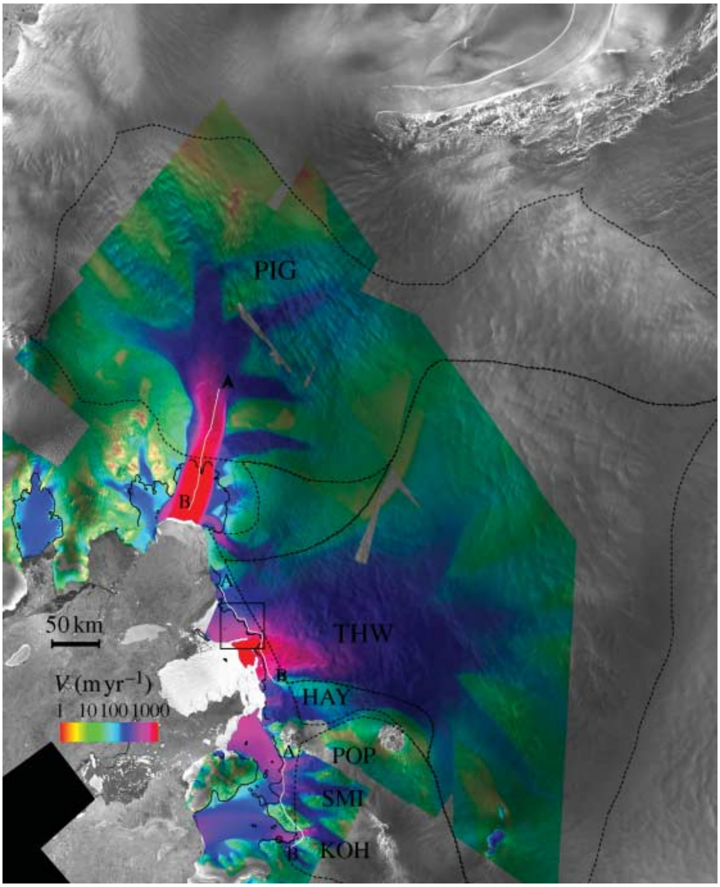


Figure 2. Ice velocity of glaciers draining into Pine Island Bay, West Antarctica from ERS-1/2 1996 interferometry, overlaid on a Radarsat-1 map of radar brightness. Grounding line is thin black; drainage boundaries extending from the NASA/CECS thickness data collected in 2002 are dotted black. Ice velocity is displayed on a logarithmic scale, as in figure 1. Profiles A and B shown in figures 5 and 6 are white lines. A thin black box delineates the location of figure 7.

3. Results

(a) *Antarctic mass flux and budget*

Ice velocity mapping in Antarctica, combined with a digital model of elevation derived from ERS-1 radar altimetry (Bamber & Bindschadler 1997) and a map of snow accumulation (Giovinetto & Zwally 2000) permitted a major revision of earlier ice sheet mass budgets from Budd & Smith (1985) and Bentley & Giovinetto (1991). In that revision, Pine Island and Lambert glaciers are not thickening but thinning (Rignot & Thomas 2002). The Pine Island Bay sector of West Antarctica exhibits the largest negative mass balance of all Antarctica. Satellite and airborne altimetry confirm that it is thinning 10 cm yr^{-1} in the interior (Wingham *et al.* 1998) to 3–4 m yr^{-1} at the coast (Shepherd *et al.* 2001, 2002). Recent airborne surveys show that most ice loss (60%) is concentrated in a narrow band (50–100 km wide) along the coast, channelized in valleys and depressions occupied by ice streams and glaciers (Thomas *et al.* 2004*a*). This pattern of change is similar to that observed in Greenland (Krabill *et al.* 2004).

The glaciers draining West Antarctica into Ronne Ice Shelf are close to a state of mass balance. Uncertainties remain in the drainage basin and ice discharge of Foundation ice stream (Lambrecht *et al.* 1999; Rignot & Thomas 2002), but there is no indication of major change in ice dynamics in this sector (Doake *et al.* 2001).

The glaciers draining into Ross Ice Shelf exhibit a positive mass budget (Joughin & Tulaczyk 2002) caused by the slow down of ice streams discovered by Rose (1979). The slow down is explained by an aclimatic process of meltwater freezing/melting at the base of the ice streams. The positive ice balance of Siple Coast is more than three times smaller than the negative ice balance of Pine Island Bay, so that overall West Antarctica is losing mass (Rignot & Thomas 2002).

In East Antarctica, most glaciers are closer to a state of mass balance than assumed in the past, but there are exceptions. Totten Glacier, the largest discharger of ice in East Antarctica, is almost certainly in a state of negative mass balance (Rignot & Thomas 2002). It flows into a 120 km confined ice shelf with an extensive grounding zone similar to Pine Island's ice plain (Corr *et al.* 2001; Rignot 2002*a*), which makes it prone to rapid change. In Davis *et al.* (2005), Totten Glacier and Moscow University Ice Shelf are thinning rapidly, along with most glaciers in Wilkes Land, such as Mertz, Ninnis and Frost.

Several glaciers (e.g. Lambert, David, Shirase) await more precise grounding line thicknesses and mean accumulation values to improve confidence in the mass budget results. Byrd Glacier's positive mass budget remains uncertain due to poor topographic control and thickness data. Stearns & Hamilton (2005) report that the glacier slowed down in recent decades, which would be consistent with its estimated positive mass budget. Overall, with the available data and uncertainties in mean accumulation and thickness, it is difficult to determine even the sign of mass balance of East Antarctica.

(*b*) *Antarctic Peninsula ice dynamics*

The Antarctic Peninsula has experienced regional warming six times the global average over the last century (Vaughan *et al.* 2001). This warming collapsed its floating ice shelves with a combination of high surface melting (Scambos *et al.* 2000), enhanced bottom melting (Shepherd *et al.* 2003) and a domino effect during collapse (MacAyeal *et al.* 2003). Collapsing ice shelves do not displace mean sea level because they are already afloat in the ocean, but they may change glacier flow upstream.

After the collapse of Larsen A in 1995, De Angelis & Skvarca (2003) found that the glaciers accelerated markedly and left bits of glacier ice hanging on cliff walls 40 m above the new glacier surface. Drygalski Glacier was flowing three times faster in 2000 than prior to the collapse (Rott *et al.* 2002). In 2005, I find that Drygalski Glacier flows another 25% higher compared to 2000 (figure 4*a*). The glacier acceleration is therefore sustained more than 10 years after the removal of the ice shelf.

In 2002, Larsen B ice shelf collapsed in three weeks following more than 10 000 years of stability (Domack *et al.* 2005). Following the collapse, Hektoria/Green/Evans, Crane and Jorum accelerated eight and two times, respectively (Rignot *et al.* 2004*b*; Scambos *et al.* 2004). In 2005, I find that Crane accelerated by a

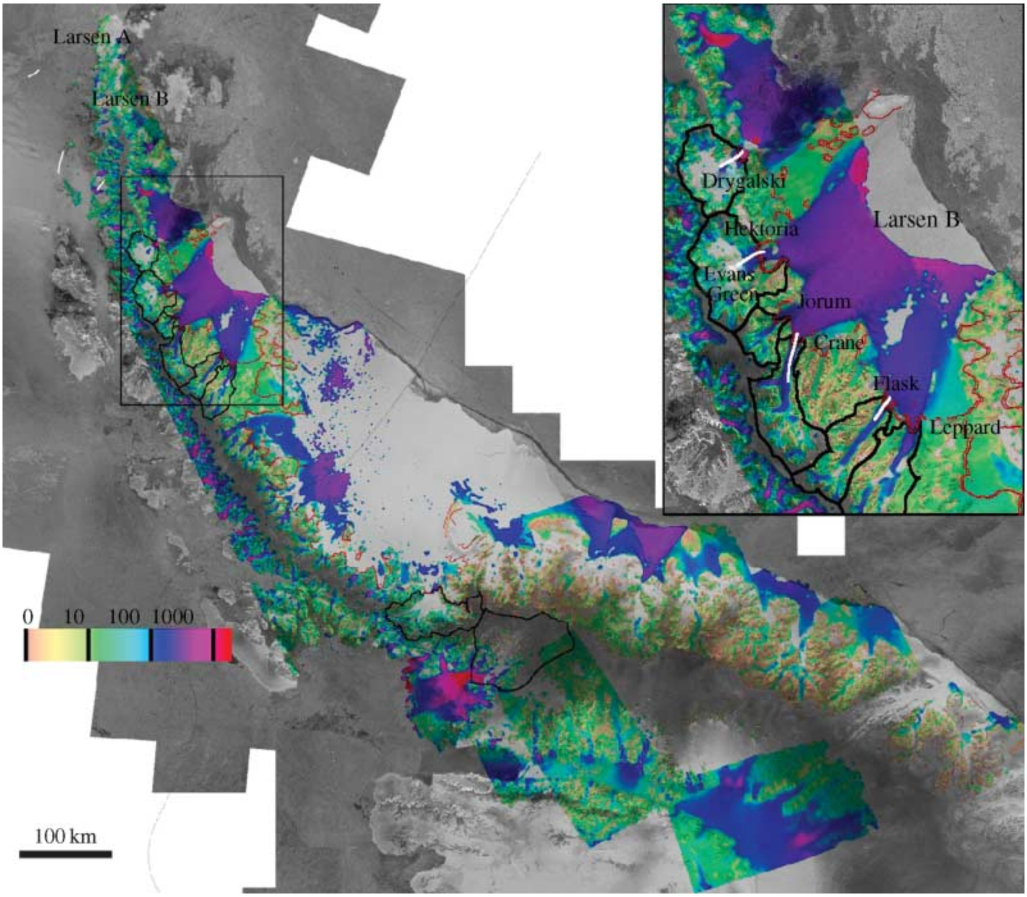


Figure 3. Antarctic Peninsula ice velocity from Radarsat-1 data in 2000. Drainage boundaries for selected basins are black. Grounding lines are red. Inset shows glaciers flowing into former Larsen A and B ice shelves. White lines denote profiles shown in figure 4.

factor of 2 compared to 2000 (figure 4*b*–*d*), and is now calving inland of its 1996 grounding line. Hektoria/Green/Evans slowed down 500 m yr^{-1}, but are still 700% out of balance. Further south, Flask and Leppard glaciers have remained at the same speed since 1996, as they are still buttressed by remnant parts of Larsen B not affected by the 2002 break up.

On the west coast of Graham Land, Fleming and other glaciers have been flowing steadily since 1992. The main episode of ice shelf removal took place in the 1980s (Doake & Vaughan 1991). The subsequent lack of change in glacier flow direction was used as an indicator that ice shelves do not matter by Vaughan (1993), but recent data showed that the region is 80% out of balance and Fleming Glacier flows 50% faster than in 1974, 50 km inland of the grounding line (Rignot *et al.* 2004*b*). Glacier acceleration at the grounding line has probably been greater than 50%. The surprising feature of Fleming Glacier is the stability of its flow despite its negative mass balance. Hence, glaciers may reach stable flow regimes even when pushed out of equilibrium. It also shows that ice shelf removal has irreversible, long-term consequences on ice flow.

The glaciers draining the northern part of the Antarctic Peninsula are certainly contributing to sea-level rise at present. The third international report on climate change neglected their contribution due to lack of data. The glaciers draining the east coast from Drygalski to Leppard lost $27 \pm 9\ \mathrm{km}^3\ \mathrm{yr}^{-1}$ ice in 2002 (Rignot *et al.* 2004*b*) and $34 \pm 10\ \mathrm{km}^3\ \mathrm{yr}^{-1}$ ice in 2005. On the west coast, all ice fronts have been retreating in the last 40 years (Cook *et al.* 2005) and air temperatures are warmer than on Fleming Glacier. If these glaciers loose a comparable percentage of their total accumulation to the ocean, then the contribution of Graham Land to sea-level rise is in the range of $0.1\ \mathrm{mm\ yr}^{-1}$ (Rignot *et al.* 2004*b*).

In Palmer Land, snow accumulation has increased by 10–20% (Raymond *et al.* 1996) in recent decades compared to the last century. The interior is thickening rapidly according to satellite radar altimetry (Davis *et al.* 2005). The interior gain in mass may therefore compensate the loss of mass at the coast. George VI Ice Shelf is experiencing vigorous melting from the ocean and its ice front is retreating (Lucchitta & Rosanova 1998). Wilkins Ice Shelf is slowly disappearing. On the east coast, the mass balance of Larsen D and E glaciers is unknown. We know very little about ice flow changes in this part of the Antarctic Peninsula.

(*c*) *Pine Island Bay ice dynamics*

Pine Island Bay was identified in 1979 as prone to trigger widespread retreat of ice in West Antarctica (Hughes 1981). The glacier eluded observations for decades and Hughes' hypothesis fell into oblivion when a calculation was published that the glacier was thickening (Bentley & Giovinetto 1991). In 1997, it was found that the glacier grounding line had been mislocated, with the consequence that the glacier mass budget was not positive but negative. More important, the glacier grounding line was observed to be retreating rapidly at $1\ \mathrm{km\ yr}^{-1}$. Satellite data also revealed that the ice shelf in front of Pine Island Glacier experiences some of the highest rates of bottom melting in Antarctica: $58 \pm 8\ \mathrm{m\ yr}^{-1}$ near its grounding zone and $24 \pm 4\ \mathrm{m\ yr}^{-1}$ on average for the whole ice shelf. This is two orders of magnitude larger than melting underneath the Ross and Ronne ice shelves. This suggests an enormous influence of thermal forcing from the ocean on the ice shelf mass budget (Jenkins *et al.* 1997), and highlights ocean warming as the most likely cause for ice shelf thinning and grounding line retreat (Rignot 1998). Shepherd *et al.* (2001) showed that glacier thinning is felt far inland of the grounding line and is concentrated in areas of fast flow. At the same time, we found that the glacier was accelerating and that the acceleration was sufficient to explain the observed thinning (Rignot *et al.* 2002).

The two studies differed in the amount of mass loss by a factor of 2. In 2002, NASA and Centro de Estudios Cientificos (CECS) conducted an airborne survey of Pine Island Bay glaciers with radio echo sounding, laser altimetry and global positioning system precision navigation. The results confirmed the larger negative balance from the mass flux method (Rignot *et al.* 2005) and revealed thinning rates several times larger than those reported from satellite radar altimetry (Thomas *et al.* 2004*a*). One reason for the difference is that the glaciers have been accelerating and hence thinning has been increasing with time. A second reason is that the amount of spatial averaging used by the two methods differs by orders of magnitude (tens of metres versus tens of kilometres). Finally, satellite radar altimetry may underestimate thinning in the trunk of narrow glaciers because of its coarse spatial resolution (10 km) and its sensitivity to surface slope.

The flow evolution shown in figure 5 spans from 1974 to present. The glacier velocity increased 38% in 31 years. The acceleration rate increased from 0.8% yr^{-1} in 1974–1987 (10.4% in 13 years) to 4% yr^{-1} in 2002–2005 (13.2% in 3 years). The acceleration was small (0.6%) in 1987–1992 (Joughin *et al.* 2003) and 2001–2002 (0.3%), but observable on a monthly basis in 2000 (Rignot *et al.* 2002) and from 2001 to 2005. The acceleration is detected more than 100 km upstream of the grounding line.

The mass balance of Pine Island Glacier was $-9\pm8\ km^3\ yr^{-1}$ ice in 1996, assuming a 10% uncertainty in mean accumulation (Rignot *et al.* 2004*a*), -18 ± 8 in 2000 and $-27\pm8\ km^3\ yr^{-1}$ ice in 2005. A set of wide, concave-shaped crevasses, crossing the whole glacier 120 km from the grounding line (Lucchitta *et al.* 1995), is observed to have migrated 10 km further upstream in 2004 compared to 2001. The spreading of cracks to that location indicates that flow acceleration is propagating upstream.

Changes on Thwaites Glacier are quite different. In 1996–2000, the centre peak velocity of Thwaites Glacier was decreasing with time, but the glacier was widening (Rignot *et al.* 2002). This is confirmed in our recent analysis which includes 1992 and 2005 data (figure 6*a*). The peak velocity is decreasing, but the eastern ice shelf velocity is increasing, effectively widening the glacier. New cracks appeared on that ice shelf in 2003, rapidly evolving into 300 m wide, 10 km long rifts in 2005 (figure 7). The rifts originated downstream of the grounding line. The mottled appearance of the ice shelf prior to rifting clearly revealed the presence of bottom crevasses almost reaching the surface. A model simulation—not reproduced here—shows that the surface expression of bottom crevasses is only visible when they reach three-fourths of the column thickness. The subsequent opening of crevasses into rifts is consistent with glacier speed up. The longitudinal strain rate increased 70% between 1992 and 2005 at that location, causing ice to fracture through the entire ice shelf column. As Thwaites Glacier continues to widen up, its mass loss could double. The glacier mass balance changed from $-36\pm7\ km^3\ yr^{-1}$ ice in 1996 to $-40\pm7\ km^3\ yr^{-1}$ in 2000 and $-42\pm7\ km^3\ yr^{-1}$ ice in 2005.

Neighbouring glaciers are changing even more rapidly than Pine Island and Thwaites glaciers, as first shown by Shepherd *et al.* (2002) and Zwally *et al.* (2002). These glaciers are largely out of balance despite being buttressed by an ice shelf (Rignot & Thomas 2002). Kohler Glacier accelerated 12% since 1992 and flows four times faster than that required to maintain mass balance (figure 6*b*). Smith Glacier accelerated 85% since 1992 and 35% since 1996. Pope Glacier shows no change in speed. As a result of the acceleration, the ice discharge from Haynes/Pope/Smith/Kohler glaciers increased from $46\pm3\ km^3\ yr^{-1}$ ice in 1996 to $57\pm3\ km^3\ yr^{-1}$ ice in 2005. The mass loss increased from 33 ± 2 to $44\pm3\ km^3\ yr^{-1}$ ice. Changes in these glacier basins per unit area are therefore greater than for Pine Island and Thwaites glaciers, despite the presence of sizeable ice shelves in front of them. In total, the mass loss from the Pine Island sector increased from $81\pm17\ km^3\ yr^{-1}$ ice in 1996 to $114\pm18\ km^3\ yr^{-1}$ ice in 2005. This total does not include the mass loss along the Getz and Abbot ice shelves, which will be published elsewhere.

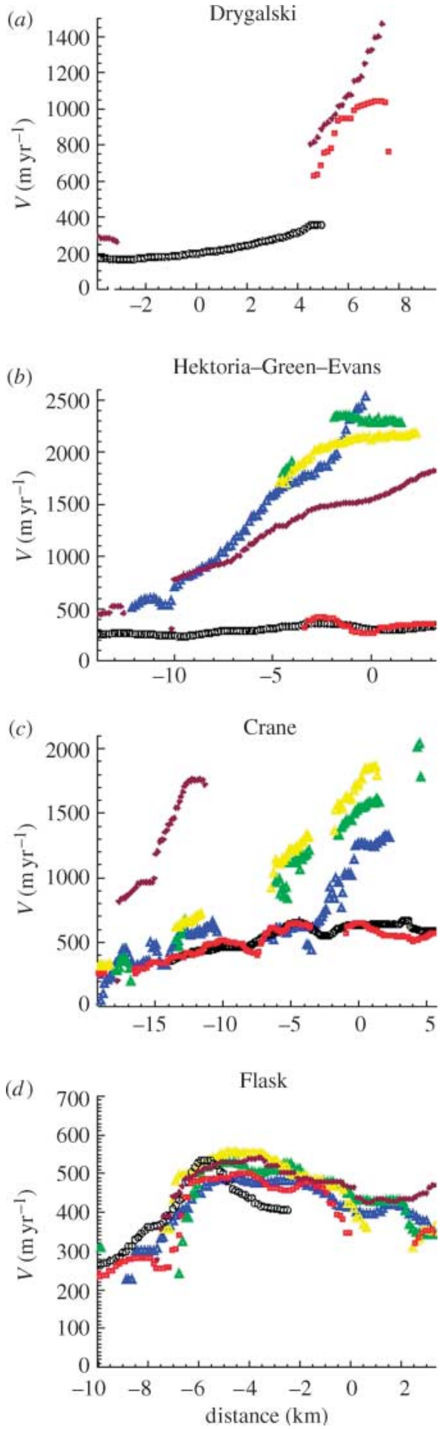


Figure 4. Glacier acceleration after the collapse of Larsen A and B for (*a*) Drygalski, (*b*) Hektoria/Green/Evans, (*c*) Crane and (*d*) Flask glaciers. Black square is 1996, red square is 2000, blue triangle is 2003, yellow triangle is 2003, green triangle is 2003 and purple diamond is 2005.

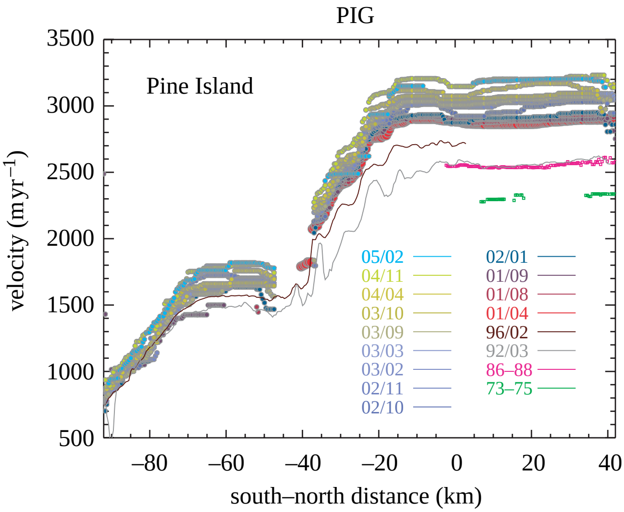


Figure 5. Glacier acceleration of Pine Island Glacier from 1974, 1987, 1992, 1996 (ERS-1/2) and 2001–2005 (Radarsat-1) versus distance along profiles A and B in figure 2. Grounding line in 1996 is at 0 km. The legend indicates the colour for each year/month.

(*d*) *Interpretation of flow changes in Pine Island Bay*

The evolution of Pine Island Glacier was compared to numerical models by Schmeltz *et al.* (2002), Thomas *et al.* (2004*b*) and Payne *et al.* (2004). Schmeltz *et al.* (2002) quantitatively explained the speed up through forward numerical modelling as resulting from ice shelf thinning and rifting. The model predicted a doubling in glacier speed if the ice shelf were abruptly removed completely. Thomas *et al.* (2004*b*) showed that the ungrounding of the ice plain upstream of the grounding line was the most significant control on glacier speed, and that the corresponding perturbation in force balance could propagate far upstream quickly. Pine Island Glacier's ice plain (Crabtree & Doake 1982; Corr *et al.* 2001) is an area of slightly grounded ice, only a few tens of metres above floatation. The ungrounding of the ice plain has a pronounced effect on the glacier force balance, longitudinal stretching and flow speed. Thomas *et al.* (2004*b*)'s model qualitatively explains why Pine Island and Smith glaciers—which include large ice plains (Rignot *et al.* 2004*a*)—retreat faster than Thwaites and Kohler glaciers—which have fewer areas near floatation. The timing and magnitude of the glacier changes were analysed in more detail by Payne *et al.* (2004).

One implication of these observations is that ice shelves do not need to experience surface melting, rifting or collapse to cause significant glacier change. A thinning ice shelf in front of a weakly grounded glacier may trigger a rapid retreat with important consequences. Other large glaciers in Antarctica are in this configuration, e.g. Totten Glacier. Conversely, a thickening ice shelf could slow down glacier flow and reverse the process.

The sensitivity of ice shelf bottom melting to thermal forcing from the ocean is 10 m yr^{-1} per degree (Rignot & Jacobs 2002). This is similar to that measured *in situ* for melting icebergs dragged in seawater (Russell-Head 1980).

Hence, the 3–4 m yr^{-1} thinning of Pine Island ice shelf could be due to ocean waters 1/3 °C warmer than what is required to maintain the ice shelf in a state of mass balance.

Jacobs *et al.* (1996) found that warm circumpolar deep water intrudes onto the continental ice shelf in Pine Island Bay. Jacobs *et al.* (1992) subsequently showed that the signature of ice shelf melting in the Amundsen Sea is observed in a 40-year time-series of oceanographic data collected in the Ross Sea. A warmer ocean is the only plausible explanation for the simultaneous melting of all ice shelves, the simultaneous glacier speed up from an entire region and the absence of ice shelf thickening in response to faster flow (Shepherd *et al.* 2004; Payne *et al.* 2004). The connection between a warmer ocean and global climate, however, remains unclear, as there is little data on ocean temperature change in this region. Most likely, oceanic changes address a much larger sector than Pine Island Bay (Jacobs & Comiso 1997).

(*e*) *East Antarctica ice dynamics*

East Antarctica rests on a huge, high plateau. Rather than responding to climate, East Antarctica may be viewed as creating its own climate. Wingham *et al.* (1998) detected no significant change in ice sheet elevation in 1992–1996. Davis *et al.* (2005)'s longer time-series suggest that the interior is thickening by 48 $km^3\ yr^{-1}$. The result broadly confirms that precipitation in the interior increases in a warmer climate (e.g. Church *et al.* 2001), but few details are available along the coast. Elevation changes are assumed to represent changes in snowfall. Changes observed over fast moving portions of the ice sheet, however, are more likely caused by ice, so that the conversion from height change to mass change should employ a density of 900 kg m^{-3} instead of 350 kg m^{-3}. This means that coastal mass losses are significantly underestimated.

Many glaciers in Davis *et al.* (2005)'s map are thinning, in particular where basins are grounded well below sea level. One example is Cook Ice Shelf (figure 8*a*), another is Totten Glacier (figure 1). The eastern side (COE) flows into an ice shelf at 500 m yr^{-1}. The western side (CWO) along Cape Freshfield calves at its grounding line and flows up to 1800 m yr^{-1}. The western ice shelf calved off in the 1980s (Frezzotti *et al.* 1998) and flows too fast to be in balance with accumulation. Figure 8*b* shows no change in velocity between 1996 and 2000 on Cook Ice Shelf, so thinning caused by ice dynamics must have initiated prior to 1996. The nearby Ninnis ice tongue broke off in 2000 and has been retreating since 1980 (Frezzotti *et al.* 1998), which suggests the possibility of a common oceanic origin for the evolution of glaciers in this region.

Totten Glacier velocity was only measured in 1996. Historic measurements were conducted several hundred kilometres inland (Young *et al.* 1989). We have no information on flow changes since 1996. Australian researchers report that the glacier thinned 10 m in the last 16 years (I. Allison 2004, personal communication). Further west, we detect no change in velocity on Lambert/Fisher/Mellor glaciers between 1996 and 2000. The interior region is slightly thickening (Davis *et al.* 2005), but the grounding line mass budget is negative (Rignot & Thomas 2002).

The information we now have about changes in ice dynamics in East Antarctica is limited, but indications from satellite radar altimetry are that there are dynamic changes taking place along the coast that require further study.

4. Discussion and conclusions

The last 10 years of observation of Antarctica with satellites have solved a significant part of the vexing problem of knowing how fast ice flows toward the ocean and have revealed spatial details in flow velocity which exceeded expectations. The results revealed that the ice sheet is changing, at a pace more rapid than anticipated. Changes in the Antarctic Peninsula caused by regional warming are large and illustrate the important role of ice shelves in controlling the mass balance of an ice sheet. Assuming that ice shelves have no influence on ice discharge is no longer justified by recent observations. Flow changes consecutive to ice shelf removal appear to be long lived, on the time-scale of decades, and have a long-term, irreversible impact on the glaciers.

Predictive models of the evolution of Antarctica in a warmer climate dismissed the importance of ice shelves because direct evidence of their mechanical importance on ice stream flow had been missing. Moreover, changes in bottom melt rates under different climate scenarios have been difficult to derive and employ in these predictions. Observations of the evolution of glaciers in Antarctica suggest that this is a significant shortcoming of the models. The mass balance of Antarctica in the next centuries may in fact strongly depend on the evolution of its ice shelves.

Ocean temperatures exert a dominant control on ice shelf evolution. Bottom melt rates are of orders of magnitude larger than surface sublimation/accumulation rates. Enhanced bottom melting is a most efficient way to remove an ice sheet, even with moderate levels of ocean warming (Warner & Budd 1998; Huybrechts & de Wolde 1999). The temperature of the world's ocean in the upper few hundred metres increased 0.3 °C over the last century (Levitus *et al.* 2000), which is large enough to impact ice shelves and glaciers. Few ice shelves have advanced in the last decades, most have retreated or remained stable (Kim *et al.* 2001). This suggests a greater likelihood for ice shelf thinning and enhanced glacier flow than ice sheet growth.

The sector of greatest impact on the mass balance of Antarctica is Pine Island Bay and its surrounding regions in the Amundsen and Bellingshausen Sea. Changes have been detected by independent techniques and verified *in situ* with airborne surveys. The oceanic origin of the glacier evolution is likely but the timing of the change is unknown. It may take decades for oceanic perturbations to migrate far upstream as shown by Payne *et al.* (2004). The basin-wide thinning of Thwaites Glacier and other glaciers probably originated well before the advent of satellites, and has been pervasive over a large area for many years.

Most significant changes are taking place along the coast. No ice sheet mass budget should be assumed complete until the entire coastal sector has been surveyed. Glacier flow above equilibrium conditions near the coastline can easily offset an increase in snowfall over the vast interior. Similar to what is observed in Greenland, Antarctica is thickening slightly in the interior and thinning along

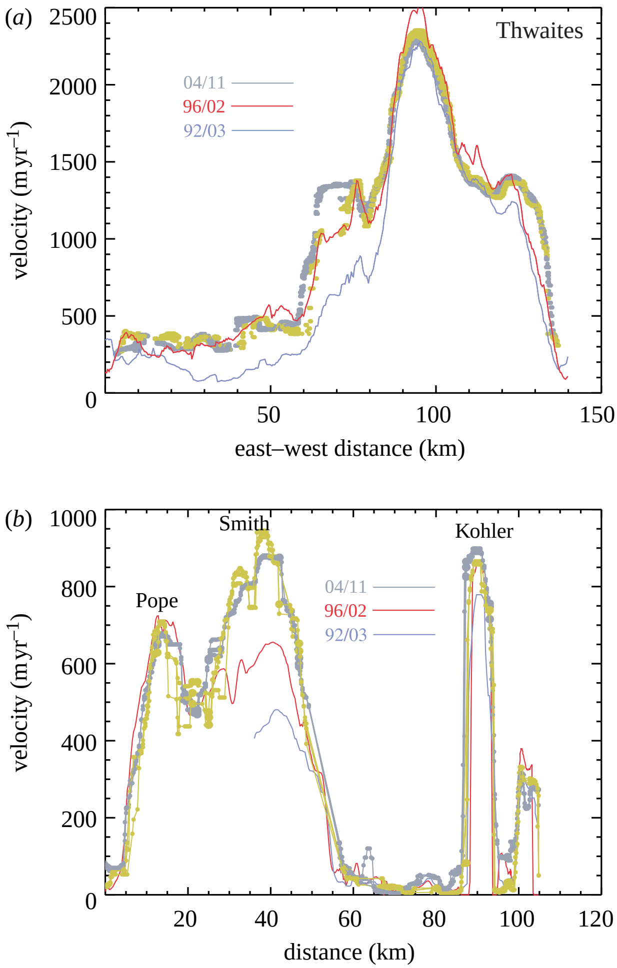


Figure 6. Glacier velocity of (*a*) Thwaites Glacier and (*b*) Pope, Smith and Kohler glaciers in 1992 (blue), 1996 (red), 2004 (grey) and 2005 (yellow) versus distance along profiles A and B shown in figure 2.

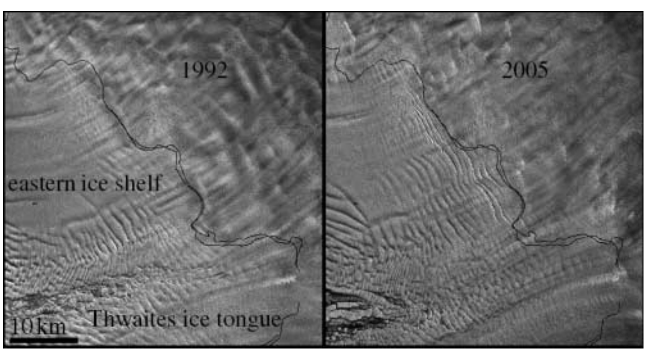
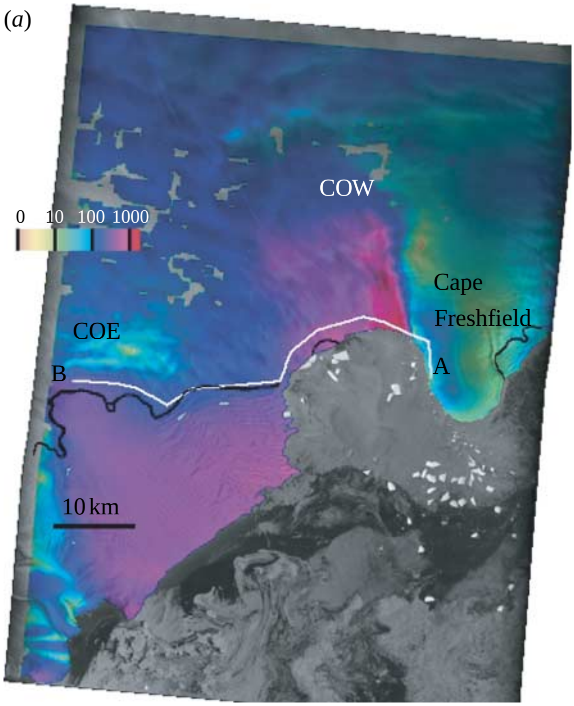


Figure 7. Eastern ice shelf of Thwaites Glacier in 1992 versus 2005 showing new rifts and cracks in 2005 near the 1996 grounding line (black line).



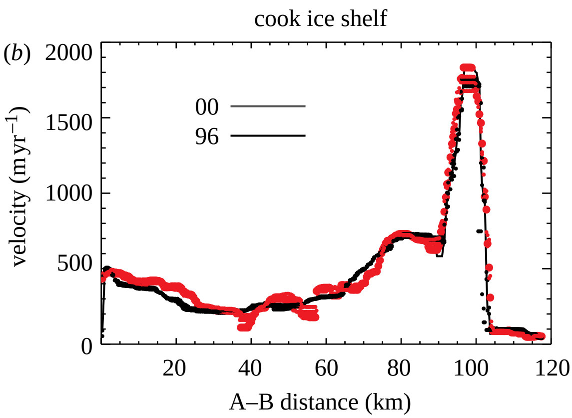


Figure 8. (*a*) Ice velocity of Cook Ice Shelf derived from ERS-1/2 data in 1996, with grounding line position in black. (*b*) Velocity variations in 1996–2000 along profiles A and B shown in white line in (*a*).

the coast, with thinning concentrated along glaciers. As in Greenland, it is likely that the evolution of Antarctica in a warmer climate will depend more on the evolution of its ice streams and glaciers than on the evolution of snowfall in the interior.

Satellite techniques have been crucial to gain insights into the mass balance and ice dynamics of what remains a vast and largely unknown part of the world. These observations must continue, in combination with airborne surveys and field surveys, to extend the data record in time, improve its quality, and gain confidence in the nature of the observed changes and their long-term significance.

New data on flow velocity, grounding line ice thickness, basal conditions and mean accumulation are needed to allow modellers to assimilate them into modern numerical schemes capable of interpreting present-day changes and forecasting the future evolution of Antarctica more realistically.

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