Accelerating ice loss from the fastest Greenland and Antarctic glaciers

R. Thomas,[1] E. Frederick,[1] J. Li,[2] W. Krabill,[1] S. Manizade,[3] J. Paden,[2] J. Sonntag,[4] R. Swift,[1] and J. Yungel[3]

Received 28 February 2011; revised 7 April 2011; accepted 10 April 2011; published 24 May 2011.

[1] Ice discharge from the fastest glaciers draining the Greenland and Antarctic ice sheets – Jakobshavn Isbrae (JI) and Pine Island Glacier (PIG)– continues to increase, and is now more than double that needed to balance snowfall in their catchment basins. Velocity increase probably resulted from decreased buttressing from thinning (and, for JI, breakup) of their floating ice tongues, and from reduced basal drag as grounding lines on both glaciers retreat. JI flows directly into the ocean as it becomes afloat, and here creep rates are proportional to the cube of bed depth. Rapid thinning of the PIG ice shelf increases the likelihood of its breakup, and subsequent rapid increase in discharge velocity. Results from a simple model indicate that JI velocities should almost double to >20 km a^{-1} by 2015, with velocities on PIG increasing to >10 km a^{-1} after breakup of its ice shelf. These high velocities would probably be sustained over many decades as the glaciers retreat within their long, very deep troughs. Resulting sea-level rise would average about 1.5 mm a^{-1}. **Citation:** Thomas, R., E. Frederick, J. Li, W. Krabill, S. Manizade, J. Paden, J. Sonntag, R. Swift, and J. Yungel (2011), Accelerating ice loss from the fastest Greenland and Antarctic glaciers, *Geophys. Res. Lett.*, *38*, L10502, doi:10.1029/2011GL047304.

1. Introduction

[2] Low rates of sea-level rise during most of the 20th century imply that snowfall on the Greenland and Antarctic ice sheets was roughly balanced by melt-water runoff and seaward ice discharge. Outlet glaciers flow along valleys cutting through the fringes of coastal mountains in Greenland and West Antarctica and across the more mountainous East Antarctic bed. The central basin in Greenland is close to sea level, but that in West Antarctica is largely 1–2 km below sea level. Recent advances in remote-sensing techniques for the measurement of ice thickness and surface elevation, glacier velocities, and Earth's gravity field show progressively increasing mass loss from both ice sheets since the 1990s. Nearly all losses are from lower-elevation regions near the coasts, partly by increased Greenland melt as summer temperatures rise, but increasingly by ice discharge into the ocean from accelerating glaciers on both ice sheets. Most of these "dynamic" losses are from just a few major glaciers that are thinning as they accelerate along deep troughs either into floating ice tongues/ice shelves or directly into the ocean. The seaward extension of such glaciers behaves like a cork in a tilted bottle of wine: loosening allows the wine to leak out; removal causes a flood. Consequences depend on how far the bottle is tilted: how far inland a glacier "feels" the effects of ice-shelf breakup. Here, we present results from two of the world's fastest glaciers - Jakobshavn Isbrae (JI) in Greenland and Pine Island Glacier (PIG) in West Antarctica - showing rapid inland migration of these effects.

[3] Airborne surveys, with the NASA Airborne Topographic Mapper (ATM) [*Krabill et al.*, 2002] and U. Kansas ice-depth sounder [*Gogineni et al.*, 2008], have been made over JI almost annually since 1991, and over PIG in 2002, 04, 09 and 10. They show progressive increase in thinning rates and inland migration of the thinning zones, following breakup of the JI floating ice tongue and sustained thinning of the PIG ice shelf, indicating that, for these glaciers, the entire drainage basins may ultimately be affected.

2. Jakobshavn Isbrae (JI)

[4] Jakobshavn Isbrae has a balance discharge (equivalent to total snowfall within its catchment basin) of about 30 km^3 ice a^{-1} [*Echelmeyer et al.*, 1991], and converges into a rapidly moving trunk ~4km wide, that flows into a deep fjord on the west coast of Greenland. Until recently, a 15-km floating glacier tongue was wedged between the fjord walls. VHF-band radar surveys (J. Plummer et al., A high-resolution bed elevation map for Jakobshavn Isbræ, West Greenland, submitted to *Journal of Glaciology*, 2011) show the fastest part of the glacier flowing in a deep trough, more than 1000m below sea level (Figure 1). Between 1850 and 1962, the calving front retreated ~25 km up the fjord, and then stabilized to within 3 km until the mid-1990s. During the 1980s and early 1990s, the glacier had a small positive mass balance [*Echelmeyer et al.*, 1991]. Then, probably in 1997, the glacier began to thin [*Thomas et al.*, 2003] at rates that increased to 15 m a^{-1} near the calving front, where its speed almost doubled to >12 km a^{-1} by 2003 as the floating tongue finally broke up, with continued increases since [*Joughin et al.*, 2008], (Figure 1). Progressive retreat of the grounding line resulting from the rapid thinning reduced the basal and lateral drag acting on the glacier [*Thomas*, 2004], and by 2005 the glacier was thinning by >2 m a^{-1} at a distance of 50 km from the calving front, increasing to >5 m a^{-1} between 2005 and 2007.

[1]SIGMA Space Inc., NASA Wallops Flight Facility, Wallops Island, Virginia, USA.
[2]Center for Remote Sensing of Ice Sheets, University of Kansas, Lawrence, Kansas, USA.
[3]URS Corporation, NASA Wallops Flight Facility, Wallops Island, Virginia, USA.
[4]URS Corporation, NASA Goddard Space Flight Center, Greenbelt, Maryland, USA.

0094-8276/11/2011GL047304

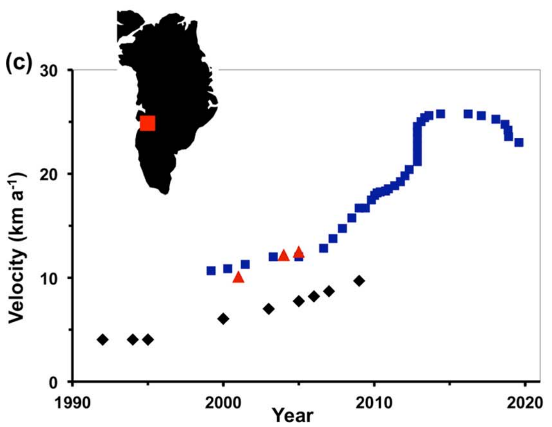

Figure 1. (a) Surface elevations along JI: 1993&97 (red), 2003 (black); 2010 (blue); smoothed bed elevations (brown), and glacier velocities in 2001 (green). The 2005 calving front is at ~15 km distance. (b) Rates of surface-elevation change (dS/dt): 1993–97 (black); 1997–2003 (red); 2003–05 (purple); 2005–07 (blue); 2007–10 (green). (c) Measured (red) and modeled (blue) JI velocity at the calving ice front, and measured (black) at a location about 5 km upstream from the 2005 ice front [*Joughin et al.*, 2008]. Location of JI is shown on inset map of Greenland.

[5] Surface and bed elevations along JI (Figure 1) indicate that observed thinning by 15 m a^{-1} near the calving front caused a grounding-line retreat by 1.3 km between 2005 and 2009. Because the grounding line is now retreating over a bed that deepens further inland, continued thinning at this rate will cause an additional retreat of 4 km by 2013, and a sustained retreat by an average of ~0.5 km a^{-1} over subsequent decades.

3. Pine Island Glacier (PIG)

[6] With a balance discharge of about 70 km^3 a^{-1} [*Rignot*, 2008], and speed in the mid 1970s of 2.2 km a^{-1}, PIG is one of the most active glaciers in the world. Its width is about 30 km where it flows along a deep trough (Figure 2) into a 50-km long ice shelf bounded on each side by grounded ice. Glacier speed near its grounding line increased by 20% or more by the mid 1990s, and continued to increase to almost 4 km a^{-1} by 2007 [*Joughin et al.*, 2010], when its annual discharge had increased to about 120 km^3 a^{-1} of ice [*Rignot*, 2008]. Analyses of satellite and aircraft data show that the ice shelf, glacier, and much of its catchment basin have thinned over the last two decades [*Shepherd et al.*, 2001; *Thomas et al.*, 2004a], and the grounding line receded by 5 km between 1992 and 1996 [*Rignot*, 1998], and a further 20 km by 2009 [*Joughin et al.*, 2010].

[7] Figures 1 and 2 show the JI and PIG beds to deepen inland from the glacier grounding lines and, for PIG, bed depth approximately doubles within a distance of 20 km. Within this region, the glacier in 2002 was only lightly grounded, leading to the prediction that continued thinning would lead to its flotation and an associated velocity increase to about 4 km a^{-1} [*Thomas et al.*, 2004b], as confirmed by subsequent observations (Figure 2). Figure 2c shows what appear to be partially-grounded "ice rumples" near the location identified as the 2002 grounding line (at zero distance), with the actual grounding line in 2002 and 2004 about 20 km upstream, at the inland end of the low-slope "ice plain" [*Corr et al.*, 2001]. Drag between the ice rumples and the seabed probably dammed up the glacier immediately upstream to cause the ice plain. Progressive un-grounding of the ice rumples then favored thinning of upstream ice resulting in a 5-km grounding-line retreat between 2004 and 2009, to a region where the bed is more than 1 km below sea level. In addition, the entire ice shelf has been thinning: comparison of ATM surveys in 2002, 04, and 09 shows a sustained drop in surface elevations averaging almost 1 m a^{-1}. This suggests that the ice shelf thinned by perhaps as much as 9 m a^{-1} during this period, decreasing its buttressing effects, and favoring more glacier acceleration and thinning.

[8] Such rapid ice-shelf thinning is most probably caused by warming of deeper ocean waters that penetrate beneath the ice shelf [*Shepherd et al.*, 2004] which, if thinning continues, may become vulnerable to breakup and rapid collapse similar to that of ice shelves around the Antarctic Peninsula [*Vaughan and Doake*, 1996]. If this happens, the grounding line will continue to retreat, and glacier velocity will increase substantially, and below we try to estimate upper limits for future velocities for JI, and for PIG following ice-shelf breakup. Various efforts are underway to improve 3-dimensional ice-flow models towards more reliable predictions of ice-sheet contributions to SLR, but results will be delayed by uncertainties in ice stiffness, basal sliding "laws", glacier boundary conditions, and incomplete numerical description of the system of equations being solved. So here, we apply a simple volume-continuity model near the calving fronts of the glaciers, where the ice dynamics are comparatively straightforward, in order to make early estimates of maximum future velocities for JI, and for

PIG after it loses its ice shelf, for a range of ice-stiffness values.

4. Glacier Acceleration After Ice-Shelf Breakup

[9] On these glaciers, thinning rates ($-dH_i/dt$) are determined primarily by the balance between thickening (by snow accumulation and advection of thicker ice from upstream) and creep thinning:

$$dH_i/dt \sim A + V\ dH_i/dx + H_i(e^{\varepsilon z} - 1) \quad (1)$$

with local vertical creep rate: $\varepsilon_z = -(1 + \alpha)\varepsilon_x$, where ε_x is the creep rate in the direction of motion, and

$$\varepsilon_x \sim \{(S_{hi} - S_{hw} - S_d)/2kB\}^3 \quad (2)$$

assuming local ice dynamics are similar to those for ice shelves. A = snow accumulation; V = glacier speed; dH_i/dx = thickness slope in the upstream direction; S_{hi} = hydrostatic ice stress (proportional to ice thickness, H_i); S_{hw} = back pressure caused by water pressure (proportional to water depth, H_w) acting on the ice front or floating ice tongue; S_d = resistive stress caused by drag between the downstream glacier and its bed and margins; k (determined by velocity convergence/divergence α) = 2 for parallel flow ($\alpha = 0$); and B = temperature/ice fabric-dependent hardness parameter averaged over ice thickness.

[10] Glacier acceleration and thinning is caused by an increase in creep thinning following a reduction in S_d resulting from thinning or breakup of a floating ice tongue, or by progressive un-grounding and breakup of near-frontal parts of the glacier [*Thomas*, 2004]. These changes, in turn, result from changes in the ocean near and beneath the floating ice [*Holland et al.*, 2008], which can increase rates of melting from beneath the floating ice to as much as two orders of magnitude more than accumulation or melt rates on the upper surface. Such changes cause rapid thinning of floating ice tongues, and their ultimate breakup. On both JI and PIG, the continued thinning and glacier acceleration shown in Figures 1 and 2 suggest that reduction in S_d as the calving front retreats, outweighs the effects of decreasing S_{hi} as the glaciers thin. The grounding line on each glacier is retreating, reducing the drag between the glacier and its bed, and the PIG ice shelf is thinning by several meters each year, reducing drag between the ice shelf and its sides. Moreover, the partially-grounded ice rumples shown in

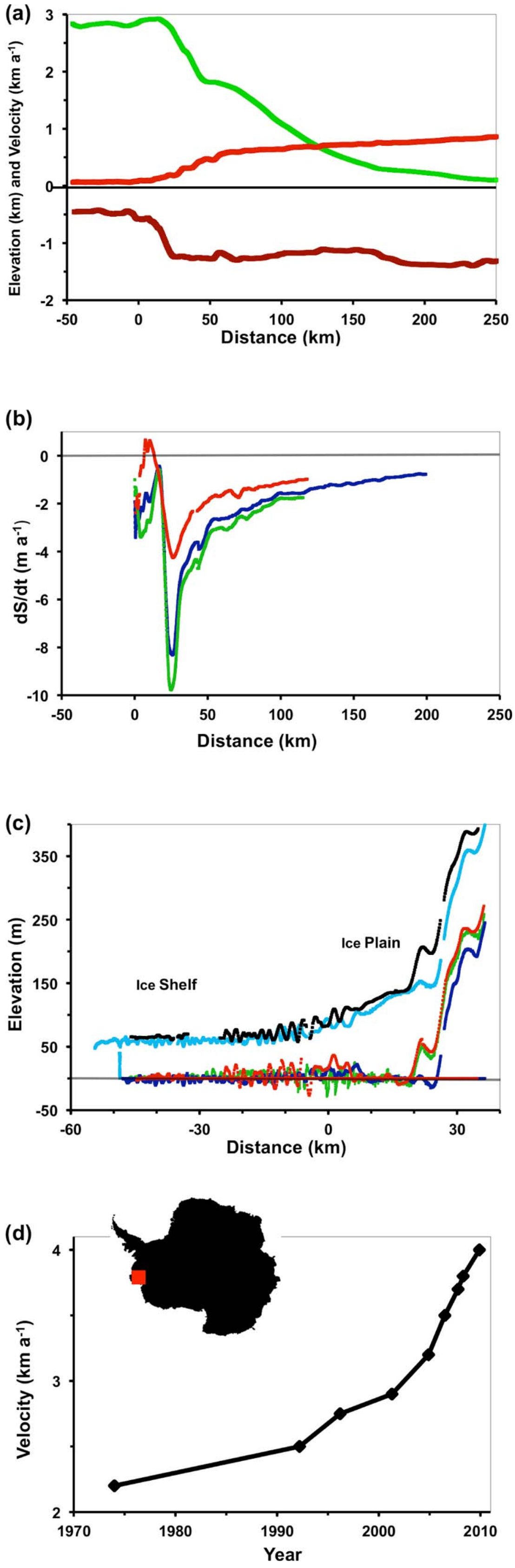

Figure 2. (a) PIG surface (red) (2002 ATM survey) and bed elevations (brown) [*Blake et al.*, 2010], and 1996 ice velocities (green) [*Rignot*, 1998]. Distance is from the 2002 grounding line, where the glacier became afloat to form an ice shelf. (b) Values of dS/dt for 2002–04 (red), 2004–09 (green), and 2002–09 (blue). (c) Surface elevations in 2002 (black) and 2009 (light blue), and freeboard above the surface elevation at which ice would float free from bedrock in 2002 (red); 2004 (green); 2009 (blue). (d) Velocity near the grounding line since the 1970s [*Rignot*, 2008; *Joughin et al.*, 2010]. Location of PIG is shown on inset map of Antarctica.

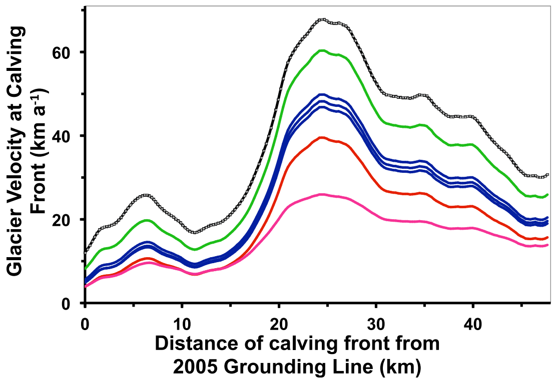

Figure 3. Estimated velocities of Jakobshavn Isbrae at the calving front as it retreats along its deep fjord, for ice stiffness (B) corresponding to ice temperature ~−4°C (black); −6°C (green); −8°C (blue and purple); −10°C (red), all for parallel flow, local thinning rates of 15 m a^{-1}, and surface slope = 0.03 radians except purple (0.05). Middle blue is for thinning rate of 50 m a^{-1}, and lower blue curve is for convergent glacier flow, with $\alpha = -0.5$.

Figure 2c appear to be thinning sufficiently to float free from the seabed.

[11] For JI, with no floating ice tongue, and on the PIG ice shelf, $S_d \sim 0$ in equation (2) near their calving fronts. Equation (2) then becomes:

$$\varepsilon_z \sim [(1 - \rho_w/\rho_i) g H_w / 2kB]^3 \tag{3}$$

where ρ_i and ρ_w are ice and water density, and g is gravity acceleration. Observations of current glacier velocities and surface and basal topography near the calving fronts give estimates of ε_z, H_w, and k from which we infer B. For JI, we estimated a value of $B \sim 250$ to 300 kPa $a^{1/3}$ equivalent to an ice temperature of −4° to −6°C, which is warmer than expected from local surface temperatures, but consistent with measurements of deep ice temperatures [*Thomsen and Thorning*, 1992; *Iken et al.*, 1993], and interpretations of blue ice in icebergs calving from JI as indicating a temperate-ice thickness of about 700 m [*Lüthi et al.*, 2009]. Moreover, B is the hardness averaged over the full ice thickness, parts of which are heavily fractured by deep surface crevasses and very probably deeper water-filled cracks at the bed. $B \sim 0$ for this fraction of the depth, reducing its depth-averaged value. For PIG, we estimated $B \sim 450$ to 500 kPa $a^{1/3}$, equivalent to an ice temperature of −14 to −17°C, which is broadly consistent with local air temperatures and basal temperatures of about −2°C.

5. The Future

[12] Both JI and PIG flow in deep troughs that slope upwards towards the grounding line, which is retreating into progressively deeper water. This will increase H_w at the calving front, which increases the dynamic imbalance of the glacier, leading to even higher discharge velocities and further inland migration of the rapidly thinning zone. This process could be delayed if a floating ice tongue were to become re-established in the fjord, but this is unlikely based on JI, where the very rapid, highly fractured glacier breaks into small fragments as it becomes afloat. Using different values of B, we solved equation (3) to give estimates of ε_z for appropriate values of H_w at the calving front of each glacier as it retreats along the deep fjord. Solution of equation (1) then gives the glacier velocity (V_f) at the calving front for selected values of dH_i/dt and α. Results are shown in Figures 3 and 4.

[13] Figure 3 shows calving-front speed is little affected by thinning rates and lateral convergence, but is strongly affected by local surface slope and ice stiffness. The overall undulating velocity pattern results from bed topography, with highest speeds over bed that slopes upwards inland. Recent JI behavior is consistent with a depth-averaged stiffness appropriate to warm ice, and if this continues then calving-front velocities should initially be as high as shown by the black curve in Figure 3. But it is likely that ice stiffness will increase as colder ice is carried by the fast-moving glacier to its calving front, shifting velocities progressively to the green, blue, and possibly red curves. Velocities would be further slowed if the surface slope near the calving front increases from its present value (about 0.03 radians) because of the decrease in thinning rates farther inland (Figure 1). This should result in a slope increase to about 0.05 radians in 100 years, which would reduce calving-front velocities towards those shown by the purple curve. However, if the glacier continues to thin as shown in Figure 1, the calving front will retreat by about 50 km over the same time period, during which average ice velocities are likely to be 30 km a^{-1} or more, with total ice discharge exceeding that required to balance upstream snowfall by more than 100 km^3 a^{-1}.

[14] Figure 4 shows a substantial increase in PIG velocities following total breakup of the ice shelf, with a strong influence from surface slope and ice temperature on glacier velocities during retreat. The influence of bed slope is

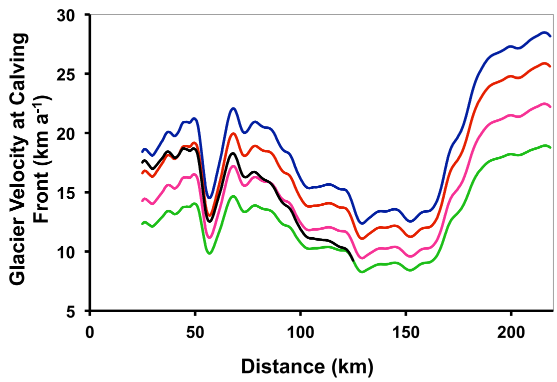

Figure 4. Estimated velocities at the PIG calving front, after ice-shelf breakup, as the glacier retreats along its deep fjord: for ice stiffness corresponding to temperature ~−17°C, parallel flow, local thinning rate of 10 m a^{-1}, and surface slope = 0.02 radians (blue); 0.03 radians (green); and purple for −20°C ice stiffness. Increasing thinning rates to 50 m a^{-1} reduces velocities by about 10%, as does convergent glacier flow, with $\alpha = -0.5$ (red). The black curve takes account of a progressive increase in slope as the calving front retreats. Distances are from the 2002 grounding line.

highlighted by the sharp dip in velocities at 60-km distance, coinciding with the marked bump on the bed shown in Figure 2. Depth-averaged ice temperatures are probably close to the assumed −17°C along much of the glacier, but results for −20°C ice hardness are also shown. Surface slopes are lower here than on JI, rising from about 0.005 radians on the partially floating "ice plain" to 0.02 radians inland from the 2009 grounding line. If the grounding line were to retreat by about 1 km a^{-1}, the thinning rate gradient inland from the grounding line (Figure 2) would increase the surface slope to about 0.03 radians in 100 years. With these assumptions, loss of the ice shelf in the near future would increase PIG velocity to follow the black curve in Figure 4, migrating from the blue/red curves towards the green curve after 100 years. During this period, velocities would average about 15 km a^{-1} with total ice discharge exceeding balance values by more than 600 km^3 a^{-1}, or enough to raise sea level by 1.5 mm a^{-1}. But all depends on if and when the ice shelf breaks up. Since 2002, the drop in surface elevation discussed earlier suggests that the ice shelf may be thinning by as much as 9 m a^{-1}, making it progressively more vulnerable to breakup. Meanwhile, reduction in marginal drag on the thinning ice shelf, and loss of basal drag at the retreating grounding line should result in continued acceleration, slowed somewhat by a decrease in hydrostatic stresses as the glacier thins.

6. Conclusions and Discussion

[15] Results from a simple volume-balance model predict a discharge-velocity increase to >20 km a^{-1} on JI by 2015, followed by a decrease by ~20% before peaking at perhaps 50 km a^{-1} by 2050–60. Future PIG behavior will be determined primarily by the ice shelf, which appears to have thinned by 10s of meters since 2002. Following its breakup, PIG discharge velocities are also predicted to rise substantially. Although these results represent probable upper limits to future ice velocities on these two glaciers, they include consideration of potential effects that would reduce the velocities. Recent behavior of JI confirms the rapid and sustained inland migration of glacier thinning and acceleration, following loss of its floating ice tongue, required to feed the accelerating discharge velocities. Moreover, JI predictions (Figure 3) showing a rapid velocity increase by 2015 (Figure 1c), will soon be tested, providing the opportunity to observe the consequences of ice-shelf breakup. However, most Greenland glaciers flow along far shorter deep fjords into the ocean, and are unlikely to undergo sustained acceleration similar to JI. But much of the ice-sheet bed in West Antarctica is more than 1 km below sea level. Here, most large glaciers flow into the Ross and Filchner-Ronne ice shelves, each about as big as Texas and unlikely to break up any time soon. But there are several large glaciers near PIG that flow into the Amundsen Sea via quite small ice shelves, and these are also vulnerable to acceleration similar to that suggested here for PIG. Together, they could raise sea-level by several mm a^{-1}, highlighting the need to understand why many ice shelves around Antarctica are thinning, and whether this thinning is likely to continue. Modeling the future behavior of their tributary glaciers will also require more extensive surveys of bed topography at high spatial resolution, and of rates of surface-elevation change over deep-based glaciers in both West and East Antarctica.

[16] **Acknowledgments.** Financial support for ATM surveys was provided by NASA's IceBridge Project and Cryospheric Sciences Program, and we thank the dedicated NASA P3 and DC8 flight crews, whose efforts allowed the safe and efficient collection of this data over some of the most isolated and extreme regions on this planet. Students, staff, and faculty at the Center for Remote Sensing of Ice Sheets (CReSIS) assisted with development of radars and associated signal processing techniques, with partial support from NASA grants NAG5-12659, NAG5-12980, NNX09AR77G and NNG10HP19C, and NSF grants OPP-01222520 and ANT-0424589.

[17] The Editor thanks two anonymous reviewers for their assistance in evaluating this paper.

References

Blake, W., L. Shi, J. Meisel, C. Allen, and P. Gogineni (2010), Airborne 3D basal DEM and ice thickness map of Pine Island Glacier, in *2010 IEEE International Geoscience and Remote Sensing Symposium*, pp. 2503–2506, Inst. of Electr. and Electron. Eng., New York.

Corr, H., C. Doake, A. Jenkins, and D. Vaughan (2001), Investigations of an 'ice plain' in the mouth of Pine Island Glacier, Antarctica, *J. Glaciol.*, *47*(156), 51–57, doi:10.3189/172756501781832395.

Echelmeyer, K., T. Clarke, and W. Harrison (1991), Surficial glaciology of Jakobshavns Isbrae, west Greenland: Part I. Surface morphology, *J. Glaciol.*, *37*(127), 368–382.

Gogineni, S., C. Leuschen, J. Li, L. Smith, J. Plummer, and A. Hoch (2008), Radar sounding of fast-flowing glaciers, *Eos Trans. AGU*, *89*(53), Fall Meet. Suppl., Abstract C11E-02.

Holland, D., R. Thomas, B. de Young, M. Ribergaard, and B. Lyberth (2008), Acceleration of Jakobshavn Isbræ triggered by warm subsurface ocean waters, *Nat. Geosci.*, *1*(10), 659–664, doi:10.1038/ngeo316.

Iken, A., K. Echelmeyer, W. Harrison, and M. Funk (1993), Mechanisms of fast flow in Jakobshavns Isbræ, West Greenland: Part I. Measurements of temperature and water level in deep boreholes, *J. Glaciol.*, *39*(131), 15–25.

Joughin, I., I. Howat, M. Fahnestock, B. Smith, W. Krabill, R. Alley, H. Stern, and M. Truffer (2008), Continued evolution of Jakobshavn Isbrae following its rapid speedup, *J. Geophys. Res.*, *113*, F04006, doi:10.1029/2008JF001023.

Joughin, I., B. Smith, and D. Holland (2010), Sensitivity of 21st century sea level to ocean-induced thinning of Pine Island Glacier, Antarctica, *Geophys. Res. Lett.*, *37*, L20502, doi:10.1029/2010GL044819.

Krabill, W., E. Frederick, S. Manizade, C. Martin, J. Sonntag, R. Swift, R. Thomas, and J. Yungel (2002), Aircraft laser altimetry measurements of changes of the Greenland ice sheet: Technique and accuracy assessment, *J. Geodyn.*, *34*, 357–376, doi:10.1016/S0264-3707(02)00040-6.

Lüthi, M., M. Fahnestock, and M. Truffer (2009), Calving icebergs indicate a thick layer of temperate ice at the base of Jakobshavn Isbræ, Greenland, *J. Glaciol.*, *55*(191), 563–566, doi:10.3189/002214309788816650.

Rignot, E. (1998), Fast recession of a West Antarctic Glacier, *Science*, *281*(5376), 549–551, doi:10.1126/science.281.5376.549.

Rignot, E. (2008), Changes in West Antarctic ice stream dynamics observed with ALOS PALSAR data, *Geophys. Res. Lett.*, *35*, L12505, doi:10.1029/2008GL033365.

Shepherd, A., D. Wingham, J. Mansley, and H. Corr (2001), Inland thinning of Pine Island Glacier, West Antarctica, *Science*, *291*(5505), 862–864, doi:10.1126/science.291.5505.862.

Shepherd, A., D. Wingham, and E. Rignot (2004), Warm ocean is eroding West Antarctic ice sheet, *Geophys. Res. Lett.*, *31*, L23402, doi:10.1029/2004GL021106.

Skvarca, P., W. Rack, H. Rott, and T. Donangelo (1999), Climatic trend and the retreat and disintegration of ice shelves on the Antarctic Peninsula: An overview, *Polar Res.*, *18*(2), 151–157, doi:10.1111/j.1751-8369.1999.tb00287.x.

Thomas, R. (2004), Force-perturbation analysis of recent thinning and acceleration of Jakobshavn Isbræ, Greenland, *J. Glaciol.*, *50*(168), 57–66, doi:10.3189/172756504781830321.

Thomas, R., W. Abdalati, E. Frederick, W. Krabill, S. Manizade, and K. Steffen (2003), Investigation of surface melting and dynamic thinning on Jakobshavn Isbrae, Greenland, *J. Glaciol.*, *49*, 231–239, doi:10.3189/172756503781830764.

Thomas, R., et al. (2004a), Accelerated sea level rise from West Antarctica, *Science*, *306*, 255–258, doi:10.1126/science.1099650.

Thomas, R., E. Rignot, P. Kanagaratnam, W. Krabill, and G. Casassa (2004b), Force-perturbation analysis of Pine Island Glacier, Antarctica, suggests cause for recent acceleration, *Ann. Glaciol.*, *39*, 133–138, doi:10.3189/172756404781814429.

Thomsen, H., and L. Thorning (1992), Ice temperature profiles for western Greenland, technical report, Groenl. Geol. Unders, Copenhagen.

Vaughan, D. G., and C. S. M. Doake (1996), Recent atmospheric warming and retreat of ice shelves on the Antarctic Peninsula, *Nature*, *379*, 328–331, doi:10.1038/379328a0.

E. Frederick, W. Krabill, R. Swift, and R. Thomas, SIGMA Space Inc., NASA Wallops Flight Facility, Bldg. N-159, Wallops Island, VA 23337, USA. (earl.b.frederick@nasa.gov; william.b.krabill@nasa.gov; robert.n.swift@nasa.gov; robert_thomas@hotmail.com)

J. Li and J. Paden, Center for Remote Sensing of Ice Sheets, University of Kansas, Nichols Hall, Room 325, 2335 Irving Hill Rd., Lawrence, KS 66044, USA. (jiluli@ku.edu; paden@cresis.ku.edu)

S. Manizade and J. Yungel, URS Corporation, NASA Wallops Flight Facility, Bldg. N-159, Wallops Island, VA 23337, USA. (serdar.s.manizade@nasa.gov; james.k.yungel@nasa.gov)

J. Sonntag, URS Corporation, NASA Goddard Space Flight Center, Code 614.1, Greenbelt, MD 20771, USA. (john.g.sonntag@nasa.gov)