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Timing of Recent Accelerations of Pine Island Glacier, Antarctica

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[1] We have used Interferometric Synthetic Aperture Radar (InSAR) data and sequential Landsat imagery to identify and temporally constrain two acceleration events on Pine Island Glacier (PIG). These two events are separated by a period of at least seven years (1987–1994). The change in discharge between two flux gates indicates that the majority of the increase in discharge associated with the second acceleration originates well inland (>80 km) from the grounding line. An analysis indicates that changes in driving stress consistent with observed thinning rates are sufficient in magnitude to explain much of the acceleration. *INDEX TERMS:* 0933 Exploration Geophysics: Remote sensing; 1827 Hydrology: Glaciology (1863); 4556 Oceanography: Physical: Sea level variations; 9310 Information Related to Geographic Region: Antarctica. **Citation:** Joughin, I., E. Rignot, C. E. Rosanova, B. K. Lucchitta, and J. Bohlander, Timing of Recent Accelerations of Pine Island Glacier, Antarctica, *Geophys. Res. Lett.*, *30*(13), 1706, doi:10.1029/2003GL017609, 2003.

1. Introduction

[2] The West Antarctic ice sheet has been the focus of numerous glaciological studies because of its potential to raise sea level by as much as 5–6 m. Early attention was focused on potential instabilities in the Ross Ice Streams [*Alley and Bindschadler*, 2001]. Although this region has receded significantly over the Holocene [*Conway et al.*, 1999], recent results show it is now thickening [*Joughin and Tulaczyk*, 2002]. Advances in spaceborne remote sensing now make it possible to study the more inaccessible drainages of the Amundsen Sea sector. Radar altimetry and InSAR results have revealed thinning over much of the Amundsen Sea sector, suggesting that these glaciers yield the largest Antarctic contribution to sea level rise [*Rignot and Thomas*, 2002; *Shepherd et al.*, 2002].

[3] Some of the most dramatic thinning has been observed on PIG. Radar altimetry has revealed thinning rates greater than 3 m/yr over the period from 1992 to 1999 [*Shepherd et al.*, 2001]. One section of the grounding line, the point where the ice loses contact with the land and begins to float, receded by 5 km from 1992 to 1994 [*Rignot*, 1998]. This thinning was found to coincide with an 18% increase in the ice flow velocity from 1992 to 2000 [*Rignot et al.*, 2002]. Results by *Rosanova and Lucchitta* [2001] indicate that the acceleration extends back to the mid 1970s. Here we examine all available data to better constrain the timing of the acceleration and to examine a mechanism for how the acceleration might have propagated rapidly inland.

2. Data and Methodology

[4] We used InSAR data from the European Remote Sensing (ERS-1 and ERS-2) satellites, which cover the period from 1992–2000. The data from 1992 and 1994 were acquired during the ERS-1 Ice Phases with temporal baselines of 3 and 6 days. The acquisitions from 1995 to 2000 were 1-day Tandem ERS-1/2 pairs. When we averaged data for the 1995–1996 and 1999–2000 periods, we obtained estimates that are representative of dates in early 1996 and 2000, respectively. Thus, in the remainder of this paper we refer to these as the 1996 and 2000 data. In addition, we examined velocities from the tracking of features in Landsat images [*Rosanova and Lucchitta*, 2001; *Scambos et al.*, 2003].

[5] In much of the 1992 and 1994 data, the fringes were too dense for phase unwrapping. In these cases, we used speckle tracking rather than interferometry [*Michel and Rignot*, 1999]. Speckle-tracked offsets were computed on a dense grid (94-by-485 m spacing) and smoothed to ~2.5 km resolution. For the Tandem data, we used phase for the across-track component and offsets for the along-track component. The velocity estimation algorithms are described by *Joughin* [2002].

[6] Our error estimates ranged from about 5–20 m/yr on the grounded ice. This excludes uncertainty caused by slope errors in the correction for vertical displacements under a surface-parallel-flow assumption [*Joughin et al.*, 1998]. Most the data were collected along the same satellite track, so these common errors largely cancel when evaluating velocity changes. In a few cases, there were large "streak errors" [*Gray et al.*, 2000]. Most of the data are visibly free of such errors. There are a few isolated areas, however, where streaks contribute errors of up to ~40 m/yr beyond that indicated by the error estimates. We corrected tidal displacements on the floating ice using a tide model [*Padman et al.*, 2002]. We assumed a 10-cm uncertainty in the differential tide estimates, which yields the major contribution to the error on floating ice (Figure 2).

[7] We averaged estimates by year(s) for 1992 (2 pairs), 1994 (1 pair), 1995–1996 (5 pairs), and 1999–2000 (2 pairs). Figure 1 shows the 1992 velocity estimate.

3. Discussion

[8] Figure 2 shows no significant acceleration from 1992 to 1994. There was acceleration from 1994 to 1996 averaging 51 m/yr over the grounded portion of the longitudinal profile.

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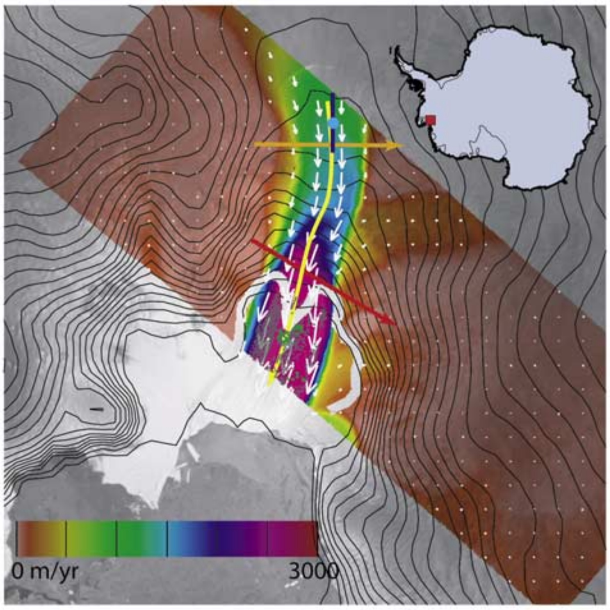

Figure 1. Velocity (color, white vectors) derived from two 1992 pairs. The underlying image is from the RADARSAT Antarctic Mapping mission [*Jezek*, 1999]. Surface topography is shown with 50-m black contours. Color lines are the profiles plotted in Figure 2, with arrows indicating the direction of increasing distance. Gray and green dots show the location of the 1987 and 1974 Landsat velocity points, respectively. Velocity was not estimated in the grounding zone because of the strong gradients in tidal displacements, so this region appears as a nearly-white data-free band.

The increase near the upstream end was about 40 m/yr, with peaks of up to 100 m/yr near the grounding line. The roughly 5-km-scale spatial variability in the velocity differences (Figure 2a) likely reflects residual errors from the InSAR slope corrections. The average increase from 1992 to 2000 was 230 m/yr along the longitudinal profile, with the difference in speed increasing in a roughly linear fashion from about 140 m/yr at the upstream end of the profile to a peak of about 350 m/yr near 60 km. Although the estimates are less reliable on the floating ice, the changes are comparable to those on the grounded ice.

[9] The increase from 1996 to 2000 is in good agreement with that measured by *Rignot et al.* [2002], using essentially the same data. The velocity change from 1992 to 1996 (Figure 2), however, is about half that measured by *Rignot et al.* [2002] using 1992 data collected along a track direction nearly orthogonal to flow. As a result, the sensitivity to displacement was small, potentially causing the large difference.

[10] *Rignot et al.* [2002] established that a significant acceleration took place during the 1990s. Because both intervals they examined showed acceleration, however, they were unable to establish the timing of the acceleration onset. The results in Figure 2 indicate that no significant acceleration occurred from 1992 to 1994 and that the change began sometime between 1994 and 1996. The data do show a slight deceleration on the floating ice from 1992 to 1994, but this could plausibly be accounted for by the unmodeled errors.

[11] Results by *Rosanova and Lucchitta* [2001] suggested that PIG acceleration extends back at least to 1974. There are no PIG InSAR data prior to 1992, so we compared the InSAR data with velocities from Landsat image pairs spanning the intervals from 1973 to 1975 (1974 estimate) [*Rosanova and Lucchitta*, 2001] and 1986 to 1988 (1987 estimate) [*Scambos et al.*, 2003]. For the 1987 estimate, we compared velocities at 579 points on the floating ice with the 1992 InSAR data. The results yielded a mean increase of 34 m/yr with a standard deviation of 35 m/yr (see also Figure 2a). Considering the Landsat errors and the InSAR errors on the floating ice, we concluded that there was statistically negligible acceleration between 1987 and 1992.

[12] There were 25 points from 1974 that overlap with the InSAR data to yield an average increase from 1974 to 1992 of 286 m/yr with a standard deviation of 44 m/yr. The mean

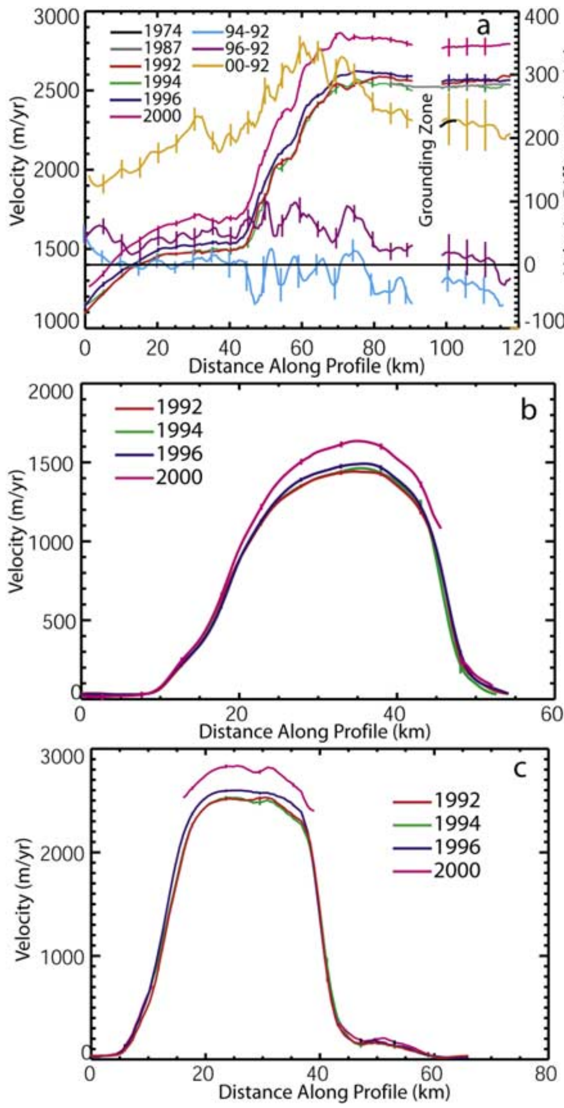

Figure 2. Ice flow speed along (a) the yellow longitudinal profile, and the (b) orange and (c) red transverse profiles in Figure 1. Arrows in Figure 1 show the direction of increasing distance. The 1974-data appear for only a short interval at about 100 km along the longitudinal profile. Error bars are plotted every several kilometers. In some cases they may be too small to be distinguishable.

difference is more than a factor of six larger than the standard deviation, indicating that the acceleration was well above the random error. The Landsat data could have a systematic error, but this would require an unrealistically large co-registration error of greater than 7 pixels. Finally, there could be absolute Landsat location errors of up to a few kilometers. To assess the impact of such errors, we shifted the Landsat data by 5 km in several directions. In all cases, the 1974 to 1992 differences were above 200 m/yr. Thus, the data indicate that a significant acceleration ($\sim$286 m/yr) took place from 1974 to 1992, with most of the change occurring between 1974 and 1987.

[13] The results suggest that from 1974 to 2000 PIG has undergone at least two accelerations of roughly comparable magnitude. These accelerations were separated by at least a 7-year period (1987–1994) of apparently steady flow. We find no evidence of major deceleration, although we cannot rule out paired acceleration/deceleration events missed by the coarse temporal sampling. The combined accelerations yield roughly a 22% change on the floating ice between 1974 and 2000, with perhaps slightly larger changes just above the grounding line.

[14] Figure 2 indicates that the more recent acceleration extends well inland. Although the increase is smaller at the upper end of the profile, the thickness there is larger. This suggests that the acceleration is causing inland thinning. To determine if this is the case, we estimated fluxes at the inland transverse (orange) profile (Figure 1). The flux through this gate changed from 64.0 to 71.5 km^3/yr over the period from 1992 to 2000. The uncertainties on these estimates are large because of the limited amount of thickness data in this region [*Lythe and Vaughan*, 2001]. The errors are relatively small, however, when evaluating changes in flux because thickness errors that are common to each estimate cancel. Assuming a 20% thickness error, the discharge increase through the upper gate from 1992 to 2000 was 7.5 ± 1.5 km^3/yr.

[15] Next, we evaluated the flux just above the grounding line (red profile, Figure 1) through a transect of measured ice thickness [*Crabtree and Doake*, 1982], which has an uncertainty of about 10%. The flux changed from 64.8 km^3/yr [1992] to 74.8 km^3/yr [2000], yielding a flux increase of 10.0 ± 1 km^3/yr. Comparison with the flux increase through the inland gate indicates that about three quarters of the mass loss caused by the increased discharge originates above the inland gate. This is not inconsistent with the larger thinning rates near the grounding line [*Shepherd et al.*, 2002], because the increased inland discharge is distributed over a much larger area.

[16] A notable aspect of the recent acceleration is that it propagated over the full 120-km length of the profile in only a few years. Glacier speed can be influenced in two ways [*Van der Veen*, 1999]. The first way is through a change in the driving stress, τ_d. The second way is through a change in the resistive stresses. There is little evidence of a change in width on the grounded ice. Locally the longitudinal stress gradients can be large (e.g., opposing pushes and pulls), but average close to zero over the grounded part of the glacier. This leaves changes in basal shear stress, τ_b, or changes in the sliding law that relates τ_b to flow speed as the prime means by which resistive stresses may have an effect.

[17] Subtle changes in τ_d may explain deceleration observed on Whillans Ice Stream [*Joughin et al.*, 2002]. The conditions for much of the fast moving part of PIG are different because τ_d is large (>100 kPa) compared to the Ross Ice Streams (1 to 20 kPa). At and above the upstream profile in Figure 1, however, τ_d is significantly smaller, as is evident from the elevation contours. This departure from a typical ice sheet profile in the upstream region suggests a weak bed.

[18] To examine the conditions in the area with a potentially weak bed, we used a simple model for the centerline ice-stream velocity, U, which is given by [*Raymond*, 1996]

$$U = \frac{2AH}{(n+1)}(\tau_d - \tau_b)^n \left(\frac{W}{2H}\right)^{n+1} \tag{1}$$

where W is the ice stream half-width, H is the thickness, and A and n are the flow law parameters [*Van der Veen*, 1999]. We applied this model at the location indicated with a light blue circle in Figure 1, with $n = 3$, $A = 2.9 \times 10^{-16}\ s^{-1}\ kPa^1$, $H = 1740$ m, $W = 19$ km, and $\tau_d = 30$ kPa averaged over the 25-km long dark blue line in Figure 1 and across the ice stream width. Using Equation 1, we found τ_b = 7.5 kPa matched U = 1280 m/yr. While only a simple model with considerable uncertainty, this result suggests a weak bed comparable to that of the Ross Ice Streams [*Kamb*, 2001].

[19] Equation 1 can be used to determine the sensitivity of acceleration with respect to temporal changes in τ_d [*Joughin et al.*, 2002]. Assuming changes in slope are more significant than changes in thickness, an approximation for acceleration caused by changes in driving stress is given by

$$\dot{U}_{\tau_d} \approx \frac{3AH(\tau_d - \tau_b)^2}{2}\left(\frac{W}{2H}\right)^4 \dot{\tau}_d. \tag{2}$$

This approximation with the parameters given above indicates that $\dot{\tau}_d$ = 0.185 kPa/yr yields $\dot{U}_{\tau_d} = 33$ m/yr^2, which compares well with the observed change ($\sim$140 m/yr change over 6 years). This amounts to an increase in slope of 1.2×10^{-4} per year, which is equivalent to a thinning induced relative elevation change of 0.3 m/yr over the 25-km dark blue line in Figure 1. For comparison, the thinning rate varies from about 3.5 to 0.7 m/yr over the first 100 km above the grounding line [*Shepherd et al.*, 2001]. Thus for the potentially weak-bedded region, acceleration may be driven by thinning-induced changes in τ_d.

[20] The analysis above applies to a plastic bed with a distinct yield stress [*Raymond*, 1996]. A hard-bed sliding model (e.g., $U \propto \tau_b^2$) is more appropriate for the region of high τ_d [*Raymond*, 1996]. Equation 1 is a closed-form solution for a plastic bed obtained from a differential equation [*Raymond*, 1996]. For a hard bed, we can numerically integrate the differential equation with a sliding law of the form$U \propto \tau_b^2$. Using W = 17 km, τ_d = 122 kPa, H = 1550 m, and U = 2230 m/yr yields $\dot{U}_{\tau_d} = 39$ m yr^{-2} for a change of 1 kPa/yr. Such a change in τ_d requires a relative change in elevation of 2.4 m/yr over the 30 km of the longitudinal profile with the greatest acceleration (40 to 70 km). If most of the change in the thinning rate is concentrated along this section, then changes in τ_d could account for a significant fraction, though perhaps not all, of the acceleration. Waves of thinning propagating upstream at rates of 7 to 40 km/yr by this mechanism (e.g., thinning induced changes in τ_d) have

been modeled for the Ross Ice Streams [*Alley et al.*, 1987; *Bindschadler*, 1997].

[21] Changes at the bed may also influence velocity, potentially providing a positive feedback. Large τ_b and velocity for the strong-bed region imply large basal melt ($\sim$0.5 m/yr). In turn, the increase in speed implies a significant increase in basal melt. The characteristics of the fast moving part of the glacier (e.g., $\tau_d > 0.25\tau_b$) indicate that PIG operates in what *Raymond* [2000] terms the drainage-limited regime, whereby velocity depends on the ability of the drainage system to accommodate changes in melt. Thus, failure of the drainage system to keep up with a melt increase could lead to increased water pressure and lubrication, which could propagate relatively quickly up-glacier. For example, a mini-surge occurred along nearly the entire length of Ryder Glacier, Greenland over a period of days to weeks [*Joughin et al.*, 1996]. Although the source of increased water (drainage of supra-glacial lakes) to the bed was different, the Ryder mini-surge demonstrates that changes in basal water pressure can potentially propagate rapidly along the length of a large outlet glacier.

[22] *Rignot* [2002] suggested that the PIG thinning was tied to acceleration on the floating ice, possibly in response to an intrusion of Circumpolar Deep Water onto the continental shelf. Analysis of changes visible in a 24-year time series of Landsat imagery indicates that significant widening and thinning have occurred on the PIG floating section since 1973 [*Bindschadler*, 2003]. That study also identified periods of enhanced rifting that may be related to the acceleration events we observe. If such changes cause strong thinning at the grounding line, then the thinning might retreat inland, by some combination of the mechanisms just described.

[23] While we have observed significant increases in discharge over a 26-year period, this is a relatively short period on glaciological timescales. At present, we have observations over an insufficient period to tell whether PIG discharge oscillates about a point of stable mass balance or we are witnessing the recent onset of a long-term draw-down of this sector of the ice sheet. Because the increased discharge from PIG appears to account for the observed thinning, it is interesting to speculate whether thinning in other Amundsen Sea drainages could be the result of similar changes in flow speed. The dramatic changes that have been observed and the potential impact on sea level argue for a greater scientific focus on PIG and the other glaciers of the Amundsen Sea Sector.

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