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Recent changes in the flow of the Ross Ice Shelf, West Antarctica

Christina L. Hulbe [a,*], Ted A. Scambos [b], Choon-Ki Lee [c], Jennifer Bohlander [b], Terry Haran [b]

[a] *School of Surveying, University of Otago, PO Box 56, Dunedin, 9054, New Zealand*
[b] *National Snow and Ice Data Center, University of Colorado, Boulder, CO, USA*
[c] *Division of Polar Earth System Sciences, Korean Polar Research Institute, Incheon, Republic of Korea*

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ABSTRACT

Comparison of surface velocities measured during the Ross Ice Shelf Geophysical and Glaciological Survey (RIGGS, 1973 to 1978) and velocities measured via feature tracking between two Moderate-resolution Imaging Spectroradiometer (MODIS) mosaics (compiled from 2003/4 and 2008/9 images) reveals widespread slowing and minor areas of acceleration in the Ross Ice Shelf (RIS) over the approximately 30 year interval. The largest changes ($-13\ \mathrm{ma}^{-2}$) occur near the Whillans and Mercer Ice Streams grounding line in the southernmost part of the ice shelf. Speed has increased over the interval (up to $5\ \mathrm{ma}^{-2}$) between the MacAyeal Ice Stream grounding line and the shelf front, and along the eastern shelf front. Changes in ice thickness computed using ICESat laser altimetry are used together with a well-tested model of the ice shelf to investigate underlying causes of change in the flow of the ice shelf over time. The observed transients represent a combination of recent forcings and ongoing response to ice stream discharge variations over the past millennium. While evidence of older events may be present, the modern signal is dominated by shorter time scale events, including the stagnation of Kamb Ice Stream about 160 years ago, recent changes in basal drag on the Whillans Ice Stream ice plain and, perhaps, iceberg calving. Details in embayment geometry, for example the shallow sea floor below Crary Ice Rise, modulate the spatial pattern of ice shelf response to boundary condition perturbations.

1. Introduction

The West Antarctic Ice Sheet (WAIS) has the potential for rapid and significant change due to its marine character and fast-flowing ice streams (Alley and Whillans, 1991). Such change is of interest in part because it would affect sea level immediately. The region of the WAIS draining into the Ross Sea was grounded nearly to the edge of the continental shelf at the Last Glacial Maximum, about 20 000 YBP and retreat from that farthest extent appears to have been episodic, not simply directed by an external climate forcing (Conway et al., 1999; Shipp et al., 1999). The prolonged retreat appears to be modulated by the shape of the sea floor and variation in the rate of ice discharge from the interior (Anderson et al., 2002; Hulbe and Fahnestock, 2004, 2007).

Evidence for profound past changes in ice stream discharge is found in many locations in the Ross Sea sector of the WAIS. Buried margin crevasses reveal the stagnation of Kamb Ice Stream to have transpired about 160 years ago, as well as the stagnation of Siple Ice Stream (a distributary of Kamb) 420 ± 60 years ago (Catania et al., 2012; Jacobel et al., 2000; Joughin and Tulaczyk, 2002; Retzlaff and Bentley, 1993). Flow features in the surface of the Ross Ice Shelf (RIS) have been used to extend the observational record of ice stream discharge variability beyond the instrumental era (Fahnestock et al., 2000). Because the flow of the shelf must adjust to the volume flux of ice discharged into it, physical tracers within the ice form a time-integrated record of past events (MacAyeal et al., 1988). That record indicates that the recent stagnation events are part of a longer term cycle of stagnation and reactivation on century time scales (Hulbe and Fahnestock, 2007). Whillans Ice Stream (WIS) stagnated about 850 years ago and reactivated 400 years later. MacAyeal Ice Stream experienced a similar cycle 800 to 650 years ago.

Whillans Ice Stream (WIS) and the ice plain across which it discharges into the ice shelf is a likely target for change detection because the direct observational record there is relatively long (Bindschadler et al., 2005). The downstream reach of the ice stream (the lightly-grounded ice plain in particular) has been slowing throughout the period of detailed observation, about the last 40 years. Bindschadler et al. (2005) used a careful survey of flow direction, stress regime, and ice thickness time series to conclude that while local (a few ice thicknesses in the horizontal scale) variations were common, there was no clear systematic regional change associated with WIS deceleration, that is, deceleration was neither caused by nor causing detectable changes in ice plain geometry. The most striking changes were in the vicinity of Crary Ice

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* Corresponding author.
E-mail address: christina.hulbe@otago.ac.nz (C.L. Hulbe).

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Rise (Fig. 1)—primarily thinning along its margins—and upstream (thickening) and downstream (thinning) of some ice rumples.

Here we suggest that looking downstream of ongoing changes on grounded ice is the best way to make sense of observations made over time intervals that are short relative to time scales inherent to the system. By examining changing characteristics in the RIS, in particular speed and thickness change, we may be able to determine if the slowdown of WIS is a recent perturbation to a stream that had for some time prior been at steady state or if it is instead part of an ongoing cycle of slow down and speed up. In the latter, then we may be able to infer the processes underlying that cycle. We may also be able to detect changes associated with other outlets, where the direct observational record is more sparse.

Correct interpretation of the ice shelf signal requires consideration of several time scales. Changes in boundary velocity—for example those due to a change in ice stream speed across the grounding line—propagate nearly instantaneously through the ice shelf while changes in ice thickness—those due to changes in ice flux across the grounding line and to changes in ice divergence—propagate on a slower, advective time scale (MacAyeal and Lange, 1988). Changes in the discharge of adjacent ice streams or the location of the grounding line may also be propagated through the coupled momentum and mass balance.

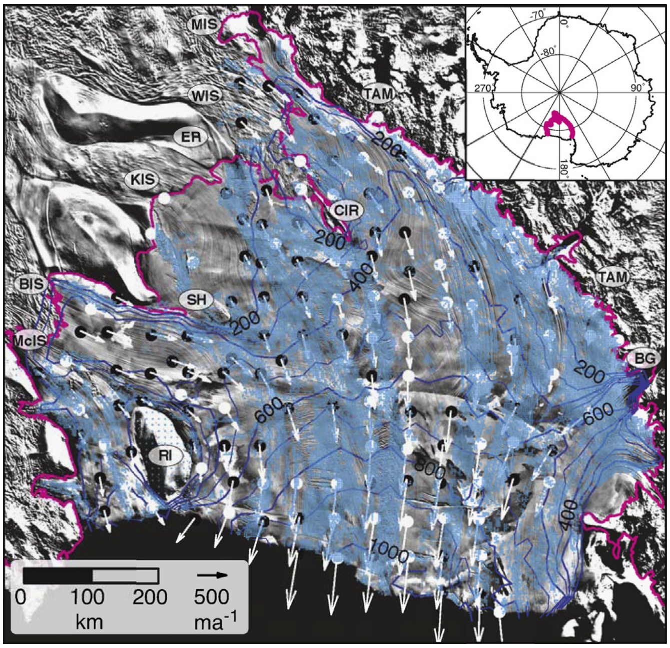

Fig. 1. Ross Ice Shelf with RIGGS and MOA-derived data used in this study. The 1970s RIGGS locations are indicated by black (directly observed) and white (computed using strain rosette measurements) circles with azimuths. The locations of the ~2006, MOA-derived velocities are shown as blue dots and the speed is contoured at an interval of 100 m a^{-1}. The magenta line here and in other figures is the ice shelf grounding line, the transition across which ice goes afloat. BG: Byrd Glacier; BIS: Bindschadler Ice Stream; CIR: Crary Ice Rise; ER: Engelhardt Ridge; KIS: Kamb Ice Stream; McIS: MacAyeal Ice Stream; MIS: Mercer Ice Stream; RI: Roosevelt Island; SH: Steershead ice rise; TAM: Transantarctic Mountains; WIS: Whillans Ice Stream. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

2. Ice velocity

Ice shelves flow by gravity-driven horizontal spreading that transports ice arriving at the grounding line to the calving front. The location of the grounding line is determined by floatation, where the weight of the ice is balanced by buoyancy. Lightly-grounded ice plains at the downstream ends of fast-flowing ice streams broaden the grounding line into a grounding zone. Resistance to ice shelf flow is provided by shear at bay walls, ice rise margins, and the subglacial bed wherever ice runs aground, by compression upstream of ice rises, and by sea water pressure normal to the calving front. At any time, the speed and thickness of the ice depends on the integrated effects of flux across the grounding line, the geometry of the embayment through which the ice flows, mechanical properties of the various materials encountered by the ice, and variations in these over time.

2.1. *Velocity measured at the surface*

The first spatially comprehensive measurements of Ross Ice Shelf flow speed were made during the Ross Ice Shelf Geophysical and Glaciological Survey (RIGGS) between 1973 and 1978 (Bentley, 1984). The RIGGS program made measurements at nearly 200 stations on the RIS and WIS ice plain, including 72 direct measurements of ice velocity and 77 ice velocities inferred from strain rosette surveys. Those observations mark one time point in the present study (Fig. 1). Errors in the velocity measurements depend on the measurement type and range from less than 5 m a^{-1} for control stations at camp sites to about 30 m a^{-1} at the stations interpolated using strain rates. We use the errors reported in appendix Table A1 of Thomas et al. (1984) in our error propagation.

A second epoch of velocities is computed using repeat images of the ice shelf, derived from the MODIS Mosaic of Antarctica mosaic, MOA (Scambos et al., 2007; Haran et al., 2005) spanning late November through February, 2003–2004, and an identically processed mosaic from images acquired in November 2008 through early March 2009 (hereafter, MOA2004 and MOA2009, respectively). The Ross Ice Shelf region is predominantly comprised of images from late December through late January for both mosaics. Image processing details are provided in Scambos et al. (2007), Haran et al. (2005). The range of image acquisition times in each of the two composites (generally 0900 UTC +/− hrs) incorporates a range of sun azimuths, providing a good image representation of subtle ice shelf features having any orientation. Image stacking and other filtering and data handling methods provide nearly seamless, uniform mosaic images of the snow surface with enhanced surface resolution. Typically 10 to 40 images contributed to each grid cell in the composites over the RIS. Both images are gridded at 125 m scale, and resolution is improved to 150–200 m from the 250 m nominal resolution of the individual MODIS scenes. Some artifacts from surface frost and thin fog remain in the images, but are subdued relative to the enhancement of persistent surface topography by the multi-image stacking approach.

The second velocity field is generated using feature tracking software (Scambos et al., 1992) with the two MODIS mosaic images. Motion is determined by measuring the offset of features, identified and matched by the correlation of small image subscenes extracted from the larger images. Several sub-scene sizes are used, ranging from 2 km to 8 km spatial equivalent scale. Small surface features from crevasses, crevasse scars, bottom crevasses, and rift edges move with the ice, and these provide image variations that can be numerically correlated. Errors in the feature tracking depend on pixel size, temporal separation of the two images, and the quality of the correlation matching algorithm. Coregistration errors are estimated at ~50 m for the geolocation of the individual MODIS scenes (which is likely reduced in the composite MOA2004 and MOA2009 mosaics by averaging of individual scene offsets); feature tracking is accurate to 0.25 to 0.5 pixels (Scambos et al., 1992). The time separation is 5 years. No error is attributed to the compositing period (roughly 35 days for the Ross Ice Shelf section of the two mosaics) because the images are blended into a single surface image representation, and smearing due to motion in the compositing period (typically 50 m, maximum 75 m) would be represented as a sub-grid-cell blurring.