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Four-decade record of pervasive grounding line retreat along the Bellingshausen margin of West Antarctica

Frazer D. W. Christie[1], Robert G. Bingham[1], Noel Gourmelen[1], Simon F. B. Tett[1], and Atsuhiro Muto[2]

[1]School of GeoSciences, University of Edinburgh, Edinburgh, UK, [2]Department of Earth and Environmental Science, Temple University, Philadelphia, Pennsylvania, USA

Key Points:

- Grounding line change mapped along Bellingshausen margin of West Antarctica with Landsat and InSAR from 1975 to 2015
- Results show majority of grounding line retreated along entire margin implicating ocean-forced dynamic thinning
- Grounding line at Venable Ice Shelf currently pinned but potential for retreat as at Pine Island Glacier

Supporting Information:

- Supporting Information S1
- Table S1
- Table S3
- Data Set S1

Correspondence to:
F. D. W. Christie,
F.Christie@ed.ac.uk

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Abstract Changes to the grounding line, where grounded ice starts to float, can be used as a remotely sensed measure of ice-sheet susceptibility to ocean-forced dynamic thinning. Constraining this susceptibility is vital for predicting Antarctica's contribution to rising sea levels. We use Landsat imagery to monitor grounding line movement over four decades along the Bellingshausen margin of West Antarctica, an area little monitored despite potential for future ice losses. We show that ~65% of the grounding line retreated from 1990 to 2015, with pervasive and accelerating retreat in regions of fast ice flow and/or thinning ice shelves. Venable Ice Shelf confounds expectations in that, despite extensive thinning, its grounding line has undergone negligible retreat. We present evidence that the ice shelf is currently pinned to a sub-ice topographic high which, if breached, could facilitate ice retreat into a significant inland basin, analogous to nearby Pine Island Glacier.

1. Introduction

Satellite remote sensing of the Antarctic Ice Sheet over the last 25 years has revealed trends of ice loss that are especially pronounced around its coastline [*Shepherd et al.*, 2010; *McMillan et al.*, 2014]. Manifested by thinning of floating ice [e.g., *Paolo et al.*, 2015], dynamic thinning of grounded ice [e.g., *Pritchard et al.*, 2009], and rapid retreat of grounding lines [e.g., *Rignot et al.*, 2014], these ice losses form one of the largest components of contemporary global sea level rise [*Shepherd et al.*, 2012; *Vaughan et al.*, 2013]. However, the processes by which they occur remain imperfectly understood. The spatiotemporal distribution of the observed ice losses strongly suggests a forcing that initiates at ice-sheet margins [*Pritchard et al.*, 2012], and several recent studies have attributed the dynamic thinning of grounded ice to the thinning of ice shelves [*Jenkins et al.*, 2010; *MacGregor et al.*, 2012; *Pritchard et al.*, 2012]. This is, in turn, a possible consequence of the ingress of relatively warm circumpolar deep water (hereafter CDW) across the continental shelf to the sub-ice shelf cavity [*Walker et al.*, 2007; *Jacobs et al.*, 2011; *Bingham et al.*, 2012]. These processes are hypothesized to be responsible for ice losses from the sectors of West Antarctica that drain to the Amundsen and Bellingshausen Seas [*Rignot et al.*, 2008; *Pritchard et al.*, 2012; *Bingham et al.*, 2012] but, while the former has received considerable observational and modeling attention [*Payne et al.*, 2004; *Shepherd et al.*, 2004; *Scott et al.*, 2009; *Joughin et al.*, 2012; *Favier et al.*, 2014; *Joughin et al.*, 2014; *Rignot et al.*, 2014; *Seroussi et al.*, 2014; *Goldberg et al.*, 2015], few observations of glacial change in the Bellingshausen Sea Sector (hereafter BSS) have been published.

It is important to generate improved records of glacial change in the BSS for several reasons. First, the estimated mass lost from the region between 1992 and 2006 (~49 Gt yr^{-1}) comprised 35% of total mass loss from the West Antarctic Ice Sheet [*Rignot et al.*, 2008], second only to contributions from the Amundsen Sea Sector (~64 Gt yr^{-1}; 46% of total loss). Second, a recent study has suggested that mass loss since 2009 has been particularly significant and that the region may have destabilized [*Wouters et al.*, 2015]. Third, although studies from the neighboring Amundsen Sea Sector are offering insights into dynamic thinning and its possible oceanic forcing, the scarcity of glacial change records and supporting information in the BSS (such as sub-shelf and grounded-ice geometry and physical oceanographic data) make it difficult to assess whether changes in the BSS have occurred at similar or different time scales to those in the Amundsen Sea Sector, potentially implicating different or additional forcing or feedbacks.

In order to improve our understanding of glacial change across this region, we generated a comprehensive data set of grounding line (hereafter GL) positional change along the Bellingshausen Sea coastline for the period 1975–2015. We exploited a combination of optical and radar satellite imagery to map GL positions at approximately 5 to 10 year intervals. We find that, over this 40 year period almost the entire length of the coastline has experienced GL retreat, with the greatest retreat typically occurring in regions containing deeply bedded, fast-flowing outlet glaciers and ice streams. We suggest that this pervasive trend implicates widespread access of CDW to the Bellingshausen glacial margin, so that even despite relatively high accumulation inland, the mass balance for much of the region is strongly negative. Only in some special cases has the GL remained relatively stable in position or undergone advance, probably as a result of local topographic pinning that is likely to be overcome with continued regional oceanic forcing.

2. Methodology

2.1. Landsat Grounding Line Mapping

We used Landsat optical satellite imagery as our primary data source due to its unmatched spatial-temporal coverage compared with other remote sensing data sets covering the BSS. The position of the GL was mapped from these images at 5 to 10 year intervals between 1975 and 2015 as follows.

Referring to the conceptual diagram of *Fricker et al*. [2009, their Figure 2], the "true" GL (where ice decouples from the bed and begins to float due to tidal forcing; their "*G*") cannot be identified reliably with static, optical imagery (such as Landsat). Instead, using Landsat imagery, the most suitable proxy for *G* is the break-in-slope, otherwise known as the "inflection point" (hereafter I_b; after notation in *Fricker et al*. [2009]), which is defined as the most seaward continuous slope break detectable in satellite imagery [*Brunt et al*., 2010; following *Scambos et al*., 2007]. Situated within close proximity to *G*, I_b appears as a clearly defined shadow-like change in image brightness on optical imagery [cf. *Bindschadler et al*., 2011] (Figure S1 in the supporting information).

Using GIS tools on georeferenced multispectral Landsat scenes (Table S1), we digitized the position of I_b at approximately 50–100 m intervals along the BSS coastline and coastal islands for years 1975, circa 1985, circa 1990, 2000, 2005, 2010, and 2015. The majority of scenes used were acquired during austral summer (January/February). For circa 1985 and circa 1990 we utilized data from ± 2 years where 1985/1990 scene spatial coverage was poor. Throughout the study, only images with cloud cover ≤10% were utilized. The launch failure of Landsat 6 in 1993 was responsible for a hiatus in Landsat acquisitions throughout the mid-1990s, precluding any mapping of results for circa 1995.

Using a similar I_b mapping technique to that employed by *Bindschadler et al*. [2011], we estimate positional uncertainty (1σ) of I_b = ~103 m for most of the BSS coastline (206 m for 1975 mapping), with the exception of the fast-flowing Ferrigno Ice Stream, where I_b (1σ) = ~502 m [cf. *Bindschadler et al*., 2011; see Text S1 and Table S2]. In terms of the principal objective of this paper, which is to monitor changes in the position of the GL across 5 year intervals, any imprecision in locating I_b (on the order of 10^2 m) is outweighed by changes in its position after 5 years (on the order of 10^2–10^3 m). This uncertainty broadly matches the positional uncertainty associated with other remote sensing techniques, such as interferometric synthetic aperture radar (InSAR) [cf. *Rignot et al*., 2011].

It is important to recognize that under some conditions the use of I_b as a proxy for the GL may fail [cf. Fricker *et al*., 2009]. The typical contexts in which I_b makes a poor indicator are in areas of fast ice flow, where subglacial bed and surface slopes are shallow and I_b is difficult to define; and/or in ice plains, where multiple breaks-in-slope may be present around the grounding zone, or where the break-in-slope may decouple substantially from the location of the true GL [cf. Corr *et al*., 2001; *Fricker and Padman*, 2006; *Fricker et al*., 2009; *Brunt et al*., 2011]. Therefore, to ascertain the utility of Landsat for monitoring GL change, we additionally employed InSAR mapping techniques to monitor the nature and configuration of grounding zones throughout the BSS and to identify the potential presence of ice plains in this sector.

2.2. Interferometric Synthetic Aperture Radar (InSAR)

Where synthetic aperture radar (SAR) observations have been acquired, it is possible to map the inland limit of tidal flexure acting on the grounding zone, *F*, directly [e.g., *Fricker et al*., 2009, Figure 2; *Brunt et al*., 2010; *Rignot et al*., 2011]. We applied interferometry to SAR data acquired from the European Space Agency's ERS-1 and ERS-2 satellites to delineate *F* along the BSS coastline between 1992 and 2011. The data consist of SAR scenes acquired in 1992, 1994, 1996, and 2011 (Table S4), and the generated interferograms have temporal baselines of either 1 day (1996) or 3 days (1992, 1994, and 2011).

Following *Park et al.* [2013], we employed a double-differential InSAR processing technique, whereby we differenced consecutive interferograms, corrected for the effects of surface topography using a resampled version of the Bedmap2 surface digital elevation model [*Fretwell et al.*, 2013], in order to remove signals related to ice flow and reveal vertical surface motion due to tidal flexure acting upon the grounding zone from floating ice shelves. Such motion, which is represented in double-differenced interferograms as a band of closely spaced fringes across the grounding zone (Figure S2), reveals *F* as the landward limit of tidally induced vertical ice motion.

To quantify the positional uncertainty of InSAR-derived *F*, we compared the tidally variable location of *F* from multiple double-differenced interferograms, where possible across multiple epochs. From this, we estimated tidally induced variation in *F* to range from ~ ± 100 m across areas of steeply bedded, shoaling subglacial topography to ≤ ~ ± 300 m across deeply grounded outlet glaciers with shallow bed slopes. These values align with positional accuracy estimations in other areas of Antarctica [cf. *Rignot et al.*, 2011; *Park et al.*, 2013].

Our determination of InSAR-derived *F* confirms the absence of ice plains and other complex ice-shelf geometries along the BSS. Therefore, unlike for other regions of Antarctica where optical-based I_b mapping campaigns have been prone to failure [e.g., *Fricker et al.*, 2009; *Brunt et al.*, 2011], we are confident that Landsat-derived I_b acts as a good proxy for the GL in this sector.

2.3. Quantifying Grounding Line Position Retreat/Advance

To quantify GL advance or retreat over epochs, we defined the 2015 Landsat-mapped I_b as a baseline. We partitioned this baseline into 30 km segments, and at the limit of each segment defined a normal extending infinitely both landward and seaward from the baseline. These normals intercept the mapped I_b for all years. To calculate advance or retreat of I_b for any 30 km segment of coastline over any period [2015 − *y*], where *y* = year of interest, we defined a polygon bounded by the 2015 baseline, the mapped I_b for year *y*, and the two intercepting normals. We then summed the areas of each polygon, in each case defining whether I_b had retreated or advanced over the epoch of interest, and then divided by 30 km to convert the final figure into a magnitude of I_b advance/retreat over each epoch of interest (see Text S2 and Data Set S1).

3. Results

Figure 1 shows that the most significant changes to the GL position between 1990 and 2015 are located at Ferrigno and Fox Ice Streams (−2.77 ± 0.50 km and −1.79 ± 0.14 km, respectively) and Stange Ice Shelf (−0.92 ± 0.14 km). Since 1990, net GL retreat has been widespread along the BSS coastline and around Thurston Island and has occurred along much of the margin draining to the Amundsen Sea between the Cosgrove Ice Shelf and Pine Island Glacier. In total, between 1990 and 2015, 65.4% of the GL along the mainland BSS experienced net retreat and only 7.4% net advance. Over the same epoch, 29.3% of the GL around Thurston Island experienced net retreat and only 1.8% net advance. Most of the detectable retreat around Thurston Island occurred along its southern (Abbot Ice Shelf-facing) margin; change along its northern margin was often not possible to discern outside Landsat 1σ-error bounds. Figure 1 also shows the tendency for the greatest GL retreat to be associated with regions of fast-flowing, likely deeply grounded ice.

Partitioning the observations into change over 5 to 10 year epochs (Figure 2), the most comprehensive coverage along the entire coastline begins in 1990, from which time it is clear that the pace of GL retreat at Eltanin Bay has been persistently high. There (for segments 22–26 in Data Set S1; locations of Ferrigno and Fox Ice Streams), the GL retreated on average 0.45 ± 0.14 km (at a rate of 45 ± 27 m yr^{-1}) from 1990 to 2000, 0.46 ± 0.14 km (92 ± 27 m yr^{-1}) from 2000 to 2005, 0.50 ± 0.14 km (100 ± 27 m yr^{-1}) from 2005 to 2010, and most recently 0.52 ± 0.14 km (104 ± 27 m yr^{-1}) from 2010 to 2015.

Where data exist from 1975 or 1985 onward, some sites experienced relatively fast GL retreat in the earlier epochs we have analyzed, followed by slower retreat or undetectable change, and then increased retreat toward the present day. One example is the western Stange Ice Shelf, where the GL at Berg Ice Stream retreated 0.54 ± 0.14 km (36 ± 27 m yr^{-1}) between 1975 and 1990, remained stable within Landsat error bounds between 1990 and 2010, then retreated 0.24 ± 0.14 km (48 ± 27 m yr^{-1}) between 2010 and 2015. Farther west, between latitudes 93°W (eastern Abbot Ice Shelf) and 87°W (Wiesnet Ice Stream/Venable Ice Shelf) we also find notable GL retreat between 1985 and 1990 relative to all intervening epochs until 2010–2015. In 2010–2015, the GL along eastern Abbot Ice Shelf then experienced a renewed retreat (distance 0.31 ± 0.14 km; rate 62 ± 27 m yr^{-1}; segments 38–40 in Data Set S1), while the GL also experienced a detectable minor

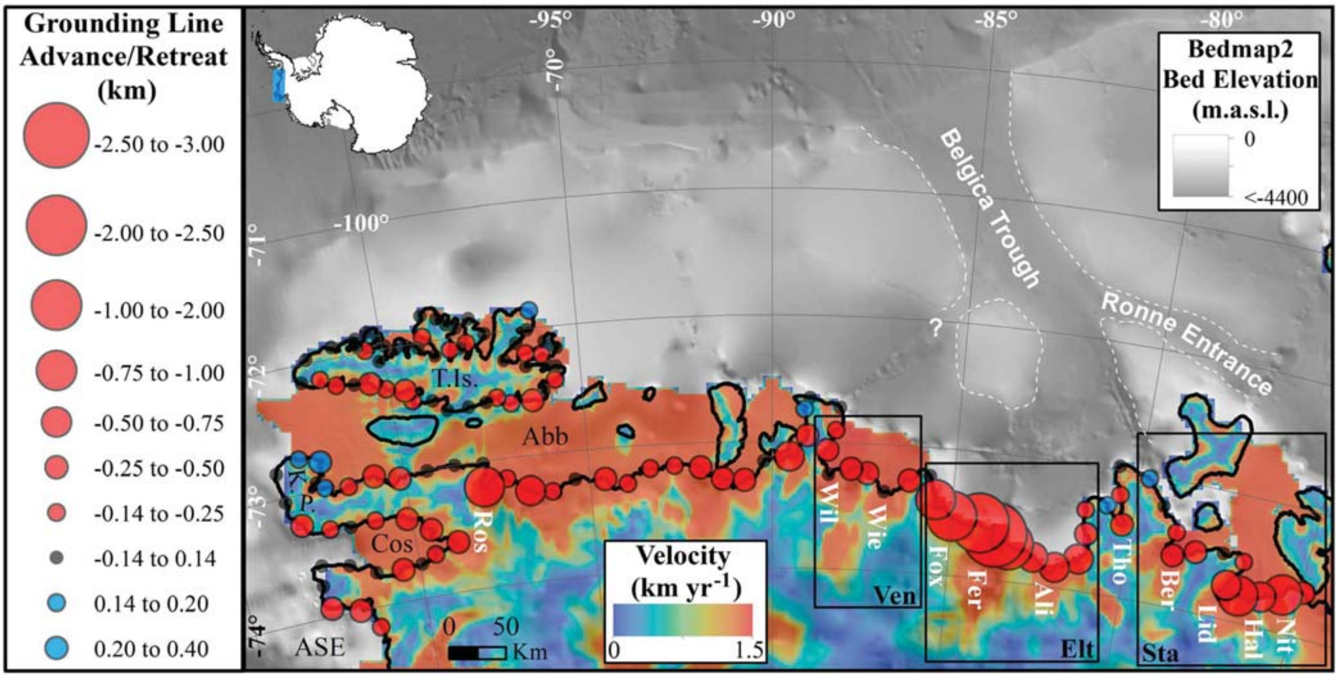

Figure 1. Net grounding line (GL) change along Bellingshausen Sea coastline of West Antarctica between 1990 and 2015 derived from Landsat mapping of I_b. Circle radii denote the magnitude and direction of change (red: *retreat*, blue: *advance*) for every 30 km segment of the BSS GL. Note the nonlinear scaling of change. Small black circles symbolize negligible detectable change within Landsat error bounds (cf. Table S2). Change symbols are superimposed over modeled depth-averaged ice velocity (km yr^{-1}) [*Williams et al.*, 2014] and continental shelf bathymetry [*Fretwell et al.*, 2013]. Inset (top left): study location. Thin broken white lines delineate the dimensions of the Belgica Trough, the Ronne Entrance tributary trough, and the area of deep bathymetry which potentially connects the margin of the Venable Ice Shelf to Belgica Trough. *Sta* and *Ven* denote the *Stange* and *Venable Ice Shelves* (respectively); *Elt*, *Eltanin Bay*; *T.Is.*, Thurston Island; *Abb*, Abbot Ice Shelf; *Cos*, Cosgrove Ice Shelf; *K.P.*, King Peninsula; *ASE*, Amundsen Sea Embayment; *Nik*, Nikitin Glacier; *Hal*, Hall Glacier; *Lid*, Lidke Ice Stream; *Ber*, Berg Ice Stream; *Tho*, Thompson Glacier; *Ali*, Alison Ice Stream; *Fer*, Ferrigno Ice Stream; *Fox*, Fox Ice Stream; *Wie*, Wiesnet Ice Stream; *Wil*, Williams Ice Stream; and *Ros*, Rosanova Glacier.

retreat between the Wiesnet and Williams Ice Streams feeding Venable Ice Shelf (0.23 ± 0.14 km at 45 ± 27 m yr^{-1}; segments 30 in Data Set S1).

4. Discussion

Our observations demonstrate that since 1975/1985 there has been a net retreat of the grounding line along almost all of the BSS (Figures 1, 2, and S3), demonstrating a pervasive trend of ice response along the whole Bellingshausen Sea coastline. Within this trend, several further phenomena are notable. First, both the greatest retreat (Figure 1) and the greatest increase in GL retreat rates from the 1990s to the present (Figure 2) have occurred along the eastern BSS coast, incorporating the glaciers and ice streams draining to Eltanin Bay and Stange Ice Shelf. Second, observed retreat rates in the BSS have varied over time across different locations. Third, while the ice streams feeding the Venable Ice Shelf have experienced limited retreat over the last three decades, retreat rates do not align with the more pronounced glaciological changes occurring more widely across the BSS and nearby Amundsen Sea Embayment.

Ferrigno Ice Stream, where we have observed the greatest net GL retreat from 1990 to 2015 (Figure 1), has been highlighted in several previous studies as a region of especially pronounced ice surface lowering and inferred ice loss [*Rignot et al.*, 2008; *Pritchard et al.*, 2009; *McMillan et al.*, 2014; *Wouters et al.*, 2015]. *Bingham et al.* [2012] suggested that Ferrigno Ice Stream is currently undergoing dynamic thinning as a consequence of the ingress of warm CDW to its ice front along Belgica Trough, a continental-shelf transecting depression formed during glacial maximum conditions [*Ó Cofaigh et al.*, 2005, Figure 1]. Here we add that the GL of Ferrigno Ice Stream has been retreating since at least 1975 (Figure S3) and that the neighboring Alison and Fox Ice Streams have been retreating since at least 1975 and circa 1990 respectively (Figures 2 and S3; no data exist over Fox Ice Stream prior to circa 1990), suggesting that the effects of CDW forcing have been prevalent in Eltanin Bay for at least 25 years. The region's vulnerability is likely exacerbated by shallow bed slopes

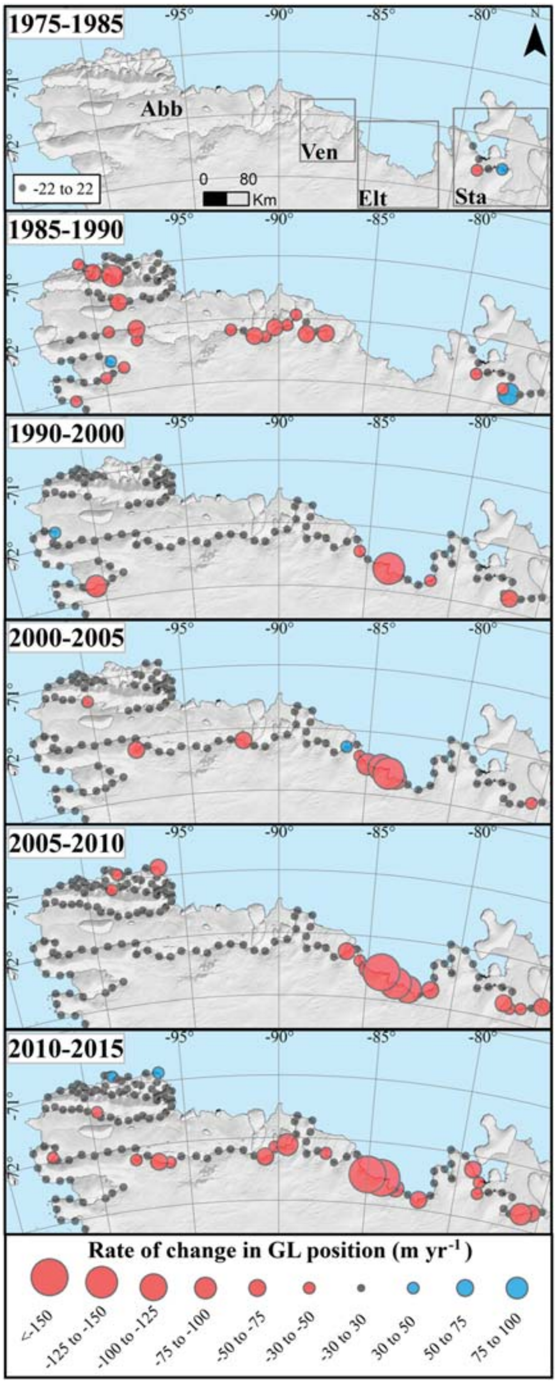

Figure 2. Rate of change in GL position over 5–10 year epochs between 1975 and 2015. Circle radii denote the magnitude and direction of change (red: retreat, blue: advance) for every 30 km segment; black circles indicate changes within Landsat error bounds. Note the nonlinear scaling of change and different scale for 1975–1985 error bounds. Site labels as per Figure 1. Data overlaid upon MOA2004 [Haran *et al*., 2013].

observed near/at the grounding zones of this sector as compared to neighboring regions [*Bingham et al*., 2012, Figure 3], in addition to the lack of an ice shelf to provide a backstress to grounded ice. As such, the general absence of ice shelves throughout this region further signals potentially longstanding ingress of CDW to the bay [*Holland et al*., 2010]. It is also notable that GL retreat here has occurred at a greater rate over the last decade (2005–2015) than previously (Figure 2), a trend which is in general agreement with those reported by *Rignot et al*. [2008], *Pritchard et al*. [2009], *McMillan et al*. [2014], and *Wouters et al*. [2015] with respect to other indicators of accelerating glacial response over this period. Moreover, following the approach of *Park et al*. [2013], GL thinning estimates obtained from our 2010–2015 GL retreat rates are consistent with estimates of inland thinning produced through CryoSat-2 swath processing over the same observational period (Figure S4 and Text S3). Analysis of CryoSat-2-derived rates of surface elevation change reveals extensive thinning inland of the Fox and Ferrigno GLs (maximum $-6.40\,\mathrm{m\,yr^{-1}}$, $\sigma = \pm 0.49\,\mathrm{m\,yr^{-1}}$), which compares well with our Landsat-derived theoretical thinning estimations (mean $= -5.96\,\mathrm{m\,yr^{-1}}$, $\sigma = \pm 3.40\,\mathrm{m\,yr^{-1}}$; Figure S4 and Text S3). There is therefore mounting evidence that the Eltanin Bay component of the BSS is not only prone to marine instability but is already experiencing the early stages of retreat, like nearby Pine Island Glacier [cf. *Jenkins et al*., 2010].

Ice flowing into Stange Ice Shelf is also remarkable for experiencing extensive GL retreat since 1975 (Figures 2 and S3). The pervasiveness and magnitude of GL retreat there suggest that the ice shelf's capacity to buttress the contributing glaciers' flow has been diminishing as a likely consequence of ice-shelf thinning. Thinning across Stange Ice Shelf has been inferred from radar altimetry between 1992 and 2008 [*Holt et al*., 2014] and 1994 to 2012 [*Paolo et al*., 2015], though the rates have varied considerably with location and time [*Holt et al*., 2014]. The greatest thinning of Stange Ice Shelf has consistently occurred on its westward flanks [*Paolo et al*., 2015], but where the western Stange Ice Shelf meets the BSS GL (at Berg Ice Stream, Figure 1), GL retreat between 1975 and 2015 has been suppressed relative to farther east (Figures 1 and 2). An explanation for this

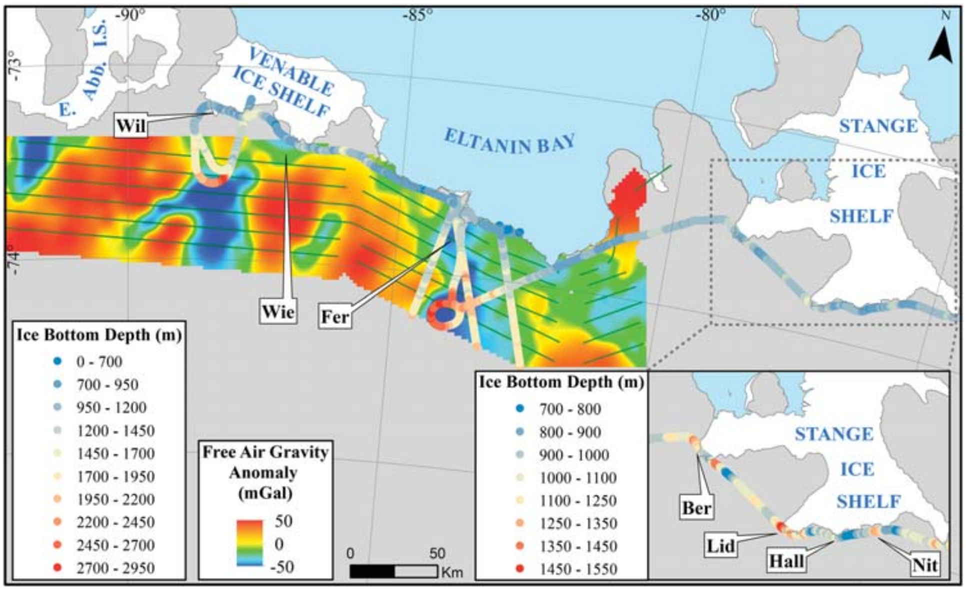

Figure 3. IceBridge radar and gridded free-air gravity anomaly data over the eastern BSS. Radar profiles were acquired on 3 November 2009 and 16 November 2011 [*Leuschen et al*., 2015a, 2015b]; gravity data were acquired on 25 October 2012 (flight lines in green) [*Cochran and Bell*, 2014]. *E. Abb. I.S.* denotes eastern Abbot Ice Shelf. Inset shows enlarged radar-derived ice bottom depths at Stange Ice Shelf. All other site labels as per Figure 1. Black lines delineate the 2008/2009 MOA GL [*Haran et al*., 2014].

apparent discrepancy lies in the geometry of the ice bed at the relative locations: a radar profile approximately following the GL along Stange Ice Shelf hints at a shallower bed, or bed protuberance, underlying Berg Ice Stream's GL compared with at the more eastern outlets feeding the ice shelf (notably Lidke Ice Stream; Figure 3). With respect to the temporal changes in GL retreat, the oscillatory nature of the observed retreat rates (i.e., instances of large-to-small-to-large retreat through time) corresponds with the suppression of ice-shelf thinning *Paolo et al*. [2015] measured for all BSS ice shelves from ~2000 to 2008 relative to the periods immediately preceding and succeeding it (i.e., 1994–2000 and 2008–2012). That the GL retreat and ice-shelf thinning trends follow a similar oscillatory pattern lends credence to the hypothesis that regional changes on and around Stange Ice Shelf are, as with Ferrigno Ice Stream, fundamentally driven by persistent CDW access to the ice-shelf cavity [cf. *Holt et al*., 2014]. Though the sub-shelf and near-ice-front bathymetry of Stange Ice Shelf remain poorly surveyed [*Graham et al*., 2011], the observed behavior of the GL retreat is strongly suggestive that the ice shelf is underlain by a north-south-aligned deep bathymetric trough that allows CDW to flow from the Ronne Entrance beneath Stange Ice Shelf to the GL.

GL retreat observed along the margins of the Abbot and Cosgrove Ice Shelves, as well as on southern Thurston Island, has occurred in locations farther from known regions of CDW ingress (Figure 1). However, most of the larger retreats in the western Bellingshausen Sea (e.g., Rosanova Glacier, Figure 1) took place where Bedmap2 [*Fretwell et al*., 2013] indicates deeply bedded grounding zones. Similar to Stange Ice Shelf, parts of this coastline exhibited some oscillatory behavior, with the margin between 100°W and 94° W (including Rosanova Glacier) and 93°W and 89°W (eastern Abbot Ice Shelf) experiencing higher retreat in 1985–1990 relative to any subsequent epoch until 2010–2015, when further enhanced GL retreat was observed. Given these observations, we hypothesize that CDW-forced dynamic thinning is also occurring in many of these locations. The dearth of bathymetric data in the western BSS [*Graham et al*., 2011] leaves open the possibility that throughout the length of the Abbot Ice Shelf, there may be undiscovered access routes for CDW to reach GLs here. Indeed, there is compelling evidence from recently acquired aerogeophysical data that a tectonically rifted basin underlies the majority of the ice shelf [*Cochran et al*., 2014] and that deep topographic lows may be present near the grounding lines of the eastern Abbot Ice Shelf (Figure 3).

The few cases of GL advance we have observed across the BSS are located where ice at the GL is thin and slow, and the bed steepens/shoals inland (Figure 1). For example, the small advances at King Peninsula (Figure 1 and segments 59, 61, and 62 in Data Set S1) occurred where ice thickness is ~200–400 m and modeled depth-averaged ice velocities are $< 100\ \mathrm{m\ yr^{-1}}$. In these locations, we hypothesize that the grounding zone geometry is relatively immune to dynamic thinning and that the GL subsequently advanced in response to the high rates of accumulation experienced by the BSS [*Lenaerts et al.*, 2012].

Finally, we highlight the minor GL retreat experienced by the ice streams feeding Venable Ice Shelf, which appears at odds with recent observations that this ice shelf has experienced significant thinning since the 1990s [e.g., *Pritchard et al.*, 2012; *Paolo et al.*, 2015], likely forced by CDW accessing the ice shelf cavity via Belgica Trough (see Figure 1) [*Graham et al.*, 2011]. Indeed, *Paolo et al.* [2015] report Venable Ice Shelf as having experienced the most dramatic thickness reduction of all Antarctic ice shelves over the period 1994–2012, thinning at a mean rate of 36.1 ± 4.4 m per decade to 82% of its 1994 thickness. Despite this exceptional thinning and known deep bathymetry to the ice shelf's northern margin that potentially connects to the CDW-flooded Belgica Trough, we have observed only limited GL retreat at this location. Here the largest GL change occurred between 1985 and 1990 (0.22 km of retreat at $45\ \mathrm{m\ yr^{-1}}$; segments 30–32 in Data Set S1); and little change in GL position has been observed since. As noted previously, a small retreat is apparent in 2010–2015, but its magnitude (total 0.11 km) is insufficient to support a case that it represents a significant upward or persistent trend in retreat rate [cf. *Wouters et al.*, 2015]. We propose that the discrepancy between oceanic forcing and suppressed GL response observed over the last two decades is a result of the specific bed geometry at/around the grounding zone, presently acting to pin the GL to its current location. Radar data acquired by NASA Operation IceBridge suggest that the bed along the Venable Ice Shelf GL is relatively shallow, including where Wiesnet and Williams Ice Streams feed into the ice shelf (Figure 3). Significantly, however, where the same surveys profiled ice thickness inland, they partly captured a deep basin underlying Williams Ice Stream from ~40 km inland. A subsequent airborne gravity survey of this basin, reaching a farther 50 km inland (Figure 3), showed free-air gravity anomalies 80–100 mGal lower than the surroundings, comparable to those observed over Ferrigno Ice Stream [*Bingham et al.*, 2012], indicating the likely existence of a significant, deeply bedded basin lying beneath Williams Ice Stream. This feature is not identified in Bedmap2 [*Fretwell et al.*, 2013], which did not incorporate these gravity-anomaly data, and renders the setting analogous to both Ferrigno Ice Stream and Pine Island Glacier, which are already experiencing active dynamic thinning. Moreover, in the case of Pine Island Glacier, the current trends are a likely response to the floating section having become ungrounded from a previously longstanding pinning point [*Jenkins et al.*, 2010]. We propose Williams Ice Stream, therefore, as an analogous setting with potential for significant future retreat once the coastal pinning point is breached. This hypothesis calls for much improved knowledge of the subglacial geometry of the GL at Williams Ice Stream/Venable Ice Shelf as an important scientific objective to facilitate numerical modeling of the region's future.

5. Conclusions

Our analysis of Landsat-derived grounding line change reveals that ~65% of the grounding line of the Bellingshausen Sea margin retreated between 1990 and 2015. The changes are pronounced, as expected, in locations where known bathymetric lows across the continental shelf facilitate ocean-driven dynamic thinning (e.g., Ferrigno Ice Stream) but have occurred more pervasively along the entire coastline than has previously been reported, implicating the likely ingress of relatively warm circumpolar deep water to the majority of the Bellingshausen Sea margin. Nevertheless, despite extensive thinning over recent decades, we observe only minimal grounding line retreat at Venable Ice Shelf. Hypothesized to be currently pinned to a sub-ice topographic high, future sustained shelf thinning at or exceeding recently reported rates would breach this barrier and facilitate ice retreat into a significant inland basin, analogous to nearby Pine Island Glacier. Together, these findings warrant the requirement for continued observation of this important and dynamically evolving region of Antarctica.

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Erratum

In the originally published version of this article, several instances of text were incorrectly typeset. The following have since been corrected, and this version may be considered the authoritative version of record.

In section 1, “Antarctic ice-sheet” was changed to “Antarctic Ice Sheet.” Also in section 1, “West Antarctic ice-sheet” was changed to “West Antarctic Ice Sheet.”

In the acknowledgments, “reviews” was changed to “reviewers.”

Geophysical Research Letters

Supporting Information for

Four-decade record of pervasive grounding line retreat along the Bellingshausen margin of West Antarctica

Frazer D.W. Christie[1], Robert G. Bingham[1], Noel Gourmelen[1] Simon F.B. Tett[1], Atsuhiro Muto[2]

[1]School of GeoSciences, University of Edinburgh, Drummond Street, Edinburgh, EH8 9XP

[2]Department of Earth and Environmental Science, Temple University, USA

Contents of this file

Figures S1 to S4
Text S1 to S3
Tables S1 to S3
Dataset S1

Additional Supporting Information (Files uploaded separately)

Dataset S1
Tables S1 and S4

Introduction

This file contains additional information pertaining to the derivation, analysis and interpretation of the results presented in this study. Landsat, ERS-1/2 Synthetic Aperture Radar (SAR), Operation IceBridge and CryoSat-2 data were acquired between 2013 and 2016 from publically available NASA and ESA data repositories. MODIS MOA and Bedmap2 data products were obtained from the NSIDC data portal and the SCAR Antarctic Digital Database, respectively. Detection of grounding line (GL) change was completed using standard GIS software and SAR processing was carried out using commercial Interferometric Synthetic Aperture Radar (InSAR) processing tools. At time of publication, the authors declare no known imperfections or anomalies within the data sets.

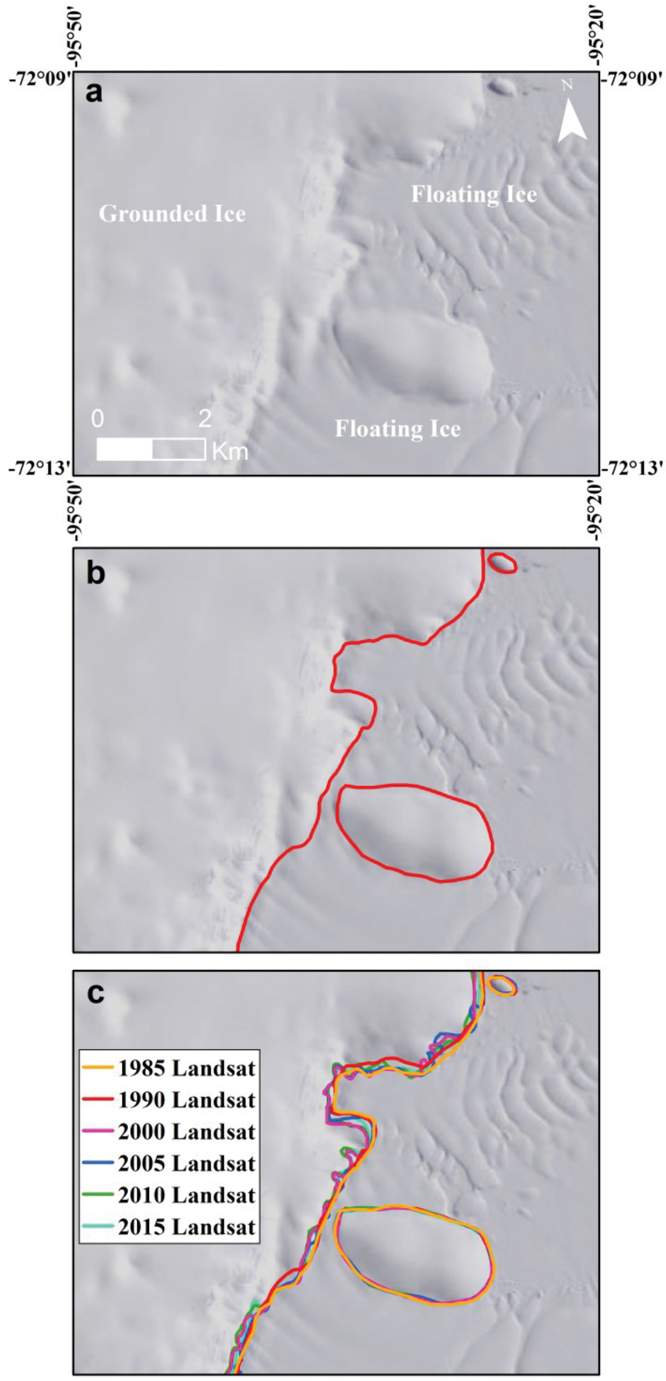

Figure S1. Visual summary of break-in-slope mapping. ***a***, Landsat scene acquired on 20 February 1991. Image shows a clearly identifiable break-in-slope between the grounded ice (eastern Thurston Island) and adjacent, floating ice. Two grounded ice rises are also visible. ***b***, The break-in-slope is digitized for year c.1990, and represents Point I_b. ***c***, This process was carried out along the whole BSS and was repeated for all other epochs, using Landsat imagery acquired during 1975 (not available in this location hence not depicted in the example above), c.1985, 2000, 2005, 2010 and 2015.

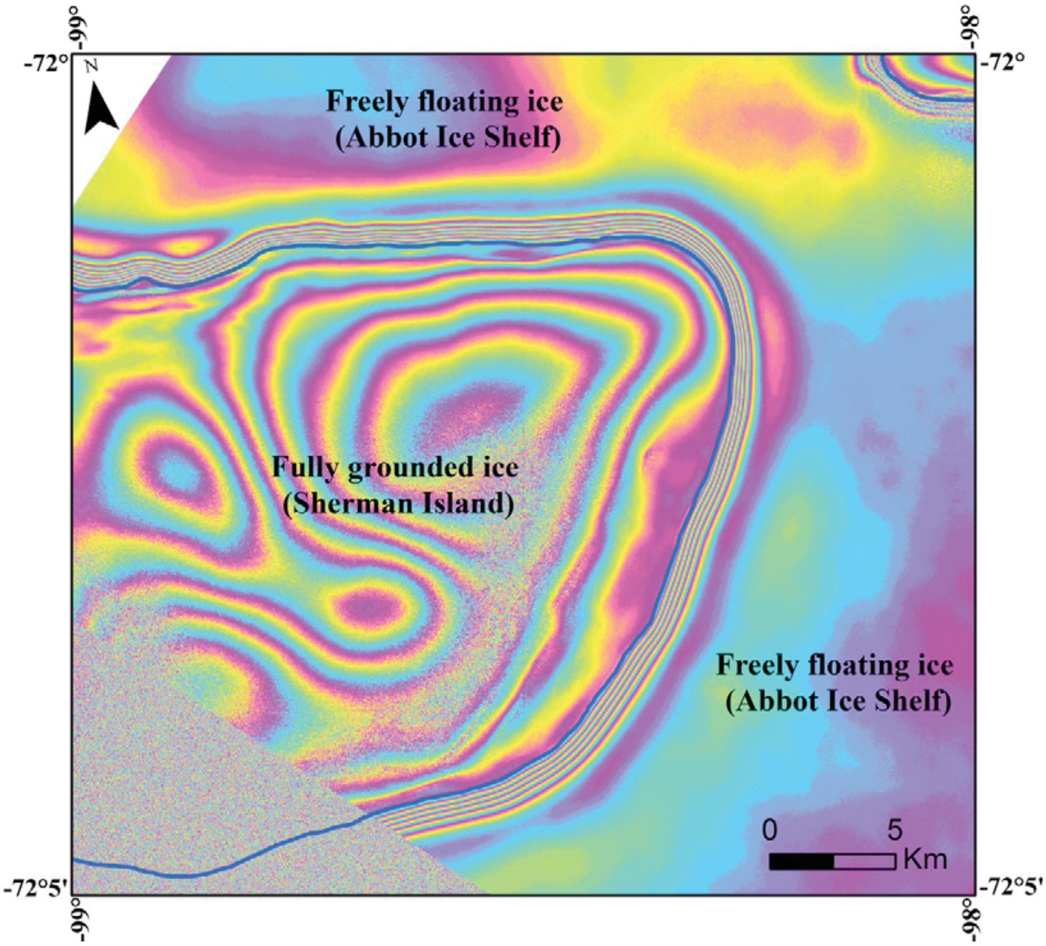

Figure S2. Double differential interferogram showing Abbot Ice Shelf and eastern Sherman Island, processed using SAR imagery acquired on 19920123-19920126 and 19920129-19920201 (3-day temporal baseline). The absolute limit of tidal flexure (*F*; blue line) is delineated by the landward limit of closely-spaced fringes observed between the freely-floating ice shelf and fully-grounded ice [cf. *Fricker* et al., 2009; their Figure 2].

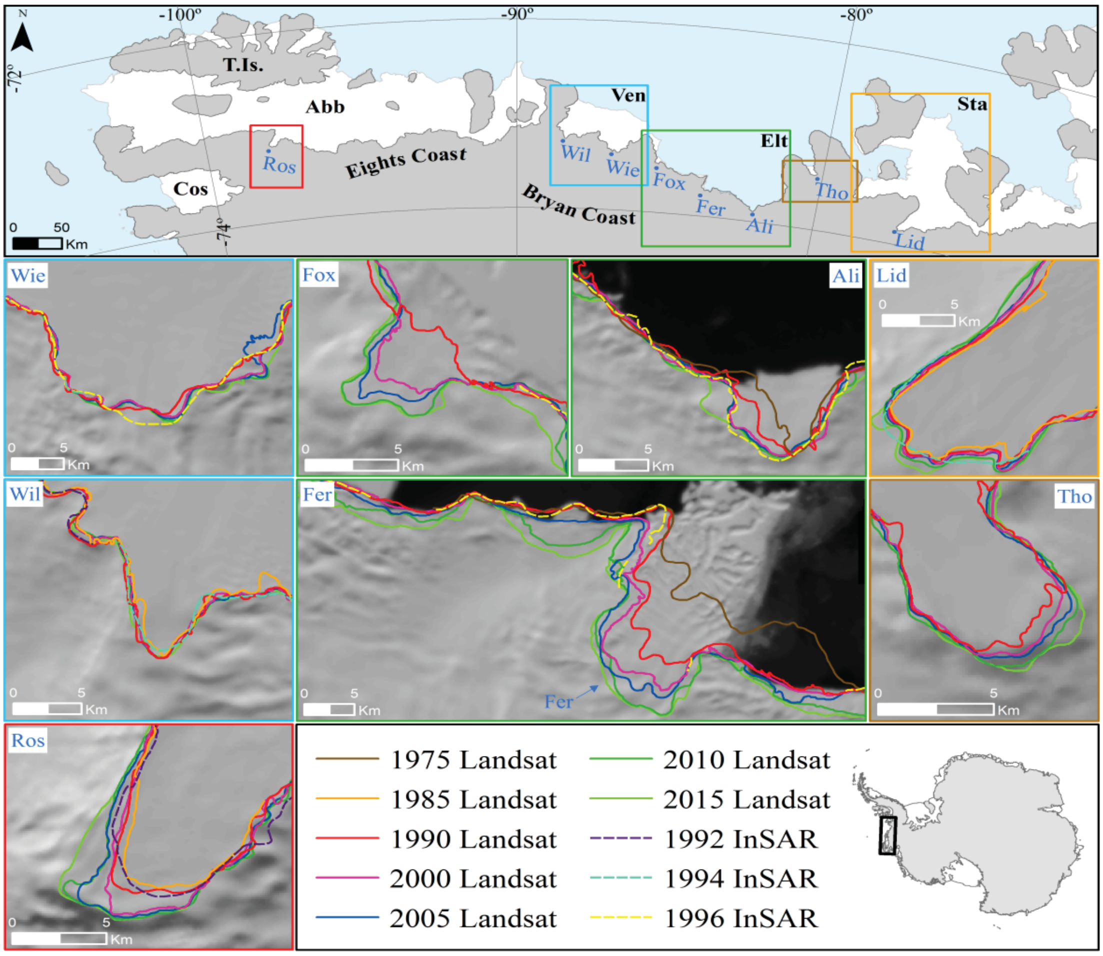

Figure S3. Examples of I_b migration and comparison with InSAR-derived F at selected locations in the BSS. Boxes in top panel are color-coded to correspond with matching color-bordered inset panels. Labelled sites as per Figure 1. 1975/1985 I_b lines are absent in some panels due to the lack of Landsat spatial coverage during these epochs. Top panel is superimposed on the MOA2004 continental outline [*Haran* et al., 2013], and black lines delineate the 2008/2009 MOA GL [*Haran* et al., 2014]. All other panels are superimposed over MOA2004 [*Haran* et al., 2013]. Inset (bottom right): study location.

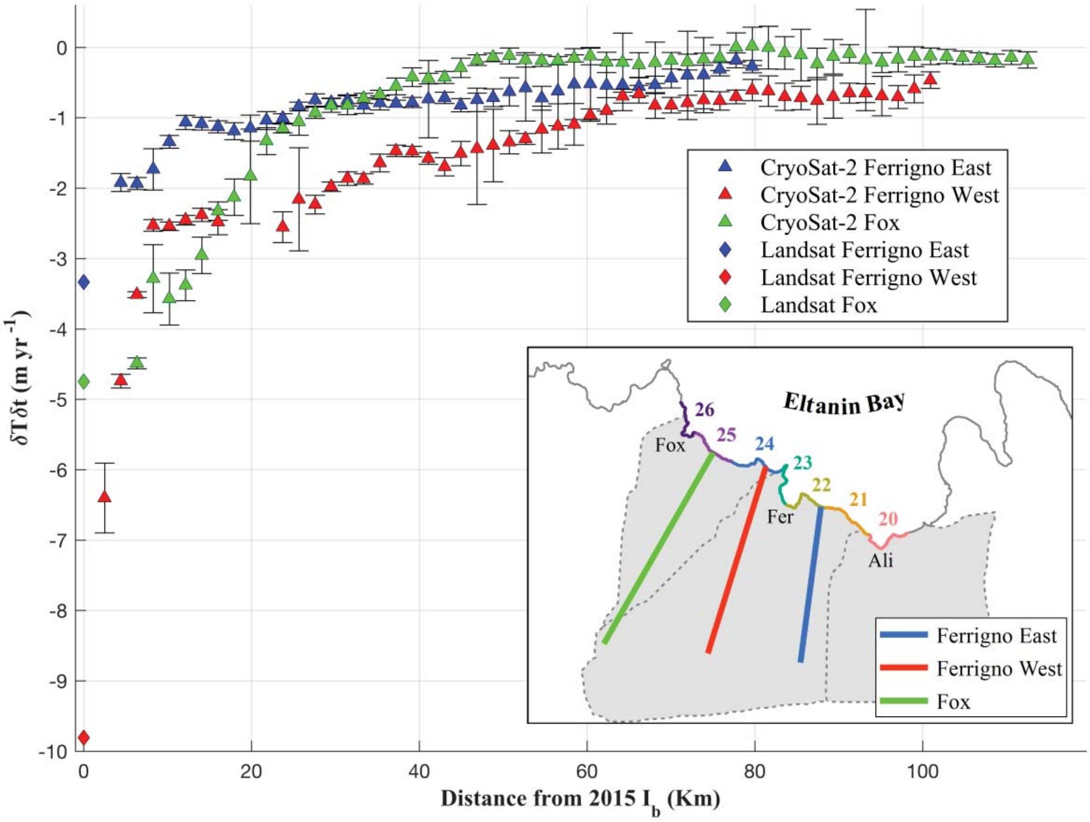

Figure S4. Rates of thickness change (δT/δt) over selected profiles in Eltanin Bay, derived from CryoSat-2 observations and Landsat-derived GL thinning estimations for the period 2010-2015 (refer to Text S3 for more information on estimating Landsat and CryoSat-2 thinning rates). Inset shows Eltanin Bay and locations of profiles shown in graph. Fox, Ferrigno and Alison Ice Stream drainage catchments shown in gray. Colored GLs and corresponding numbers refer to selected 30 km segment sites within the Eltanin Bay region, as discussed in main text and Data Set S1.

Text S1. Quantification of Landsat positional accuracy

Following *Bindschadler* et al. [2011], error in the mapping of Point I_b is dependent on the nature of the grounded-ice boundary and the geometric error associated with each Landsat satellite platform (i.e. the orbital/geo-positional error associated with successive Landsat passes; cf. *USGS* [2015]), assuming tidal variation in the position of the grounding zone to be negligible. Hence, where the boundary transition is from grounded ice to open ocean, we mapped I_b to within one Landsat scene pixel (i.e. a prescribed one pixel error; where pixel value is dictated by the spatial resolution of the satellite sensor used (Table S2)). For grounded-ice/sea-ice transitions, where the boundary is potentially more ambiguous, we mapped to within two Landsat pixels of error. At the deeply bedded Ferrigno Ice Stream, where ice velocities exceed 1 km yr^{-1} [*Rignot* et al., 2008; *Williams* et al., 2014] and the break in slope is more ambiguously located compared to neighboring (more steeply bedded) locations, we assigned a 500 m error boundary in accordance with Bindschadler et al. [2011]. Along the remaining boundaries of the BSS coastline, we prescribed a three-pixel error. These include regions of slow-flowing grounded ice flowing into ice shelves and some outlet glaciers where I_b is more readily identifiable than faster flowing ice streams.

Accordingly, following the discussion in *Bindschadler* et al. [2011], estimations of positional error (1σ) were calculated as the root-sum-square of I_b delineation accuracy and satellite geometric error. These create the positional accuracy estimations found in Table S2.

In order to propagate error between successive I_b observations (e.g. the results shown in Figure 1), we calculated the root-sum-square of the (1σ) positional errors associated with Landsat 8 (i.e. for the 2015 baseline, refer to main text Section 2.3) and the earlier Landsat platform under analysis for a given epoch. Thus, for example, in Figure 1, the majority of accumulative errors in the BSS equal 137 m, as the grounding line boundary type is classed as 'Slow-ice-to Shelf or Outlet Glacier' (Table S2; e.g. $\sqrt[2]{91^2 + 103^2}$). For Figure 2, scaled error approximations were derived by dividing these errors by the temporal period under question.

Text S2. I_b change quantification: Calculation of 5- to 10-year change.

Following completion of the I_b change calculations detailed in the main text (section 2.3), changes in I_b position over 5- to 10-year periods were calculated using the following expression:

$$\frac{\delta I_b}{\delta t} = \frac{\delta I_{b\,[2015-y0]} - \delta I_{b\,[2015-y+1]}}{\delta t} \qquad [1]$$

where $\delta I_b/\delta t$ corresponds to the change in I_b per 30-km segment over 5- to 10-year period; $\delta I_{b[2015\text{-}y0]}$ is the change in I_b over the earliest epoch under examination (e.g. change in 2005 GL to the baseline 2015 GL); $\delta I_{b[2015\text{-}y+1]}$ is the change in I_b over the immediately succeeding epoch (e.g. 2010 to 2015); and δt is the temporal period between the two epochs. Refer to Data Set S1 and Figure 2 for raw data and examples of this process.

Notably, owing to the limited spatial and/or temporal availability of Landsat imagery across some regions of the BSS (especially for 1975), the above processes were not carried out in sectors containing no [2015-*y*] data for one or both epochs under analysis.

Text S3. CryoSat-2-derived elevation change for ice draining into Eltanin Bay, and comparisons with 2010-2015 Landsat-derived estimations of ice thinning at the GL.

Rates of surface elevation change were generated from swath processing of CryoSat-2 data acquired between 2010 and 2015. Swath processing is enabled by the Synthetic Aperture Radar Interferometric (SARIn) mode of CryoSat-2 and enables one to two orders of magnitude more elevation measurements than conventional point-of-closest-approach (hereafter POCA) altimetry techniques, thereby allowing increased spatial and temporal resolution and improved coverage of ice-sheet marginal areas [*Hawley* et al. 2009; *Gray* et al. 2013]. Linear rates of surface elevation change were then derived at 500 m posting using a repeat track approach [*McMillan* et al., 2014]. Across Eltanin Bay, where we detect the greatest changes in GL position over the observational period 2010-2015, we extracted 3 profiles across the Ferrigno and Fox catchment regions (Figure S4). These profiles reveal increasing trends of rates of surface elevation change ($\delta T/\delta t$) towards the GL, with maximum values occurring at Ferrigno West (-6.40 m yr^{-1}, $\sigma = \pm0.49$ m yr^{-1}; Figure S4). Notably, average $\delta T/\delta t$ across all 3 profiles equals -0.84 m yr^{-1} ($\sigma = \pm0.21$ m yr^{-1}), which compares well with previously reported CryoSat-2 POCA observations in this area [*McMillan* et al., 2014; *Wouters* et al., 2015].

In addition to CryoSat-2-derived $\delta T/\delta t$, theoretical thinning rates over the same epoch (2010-2015) were calculated using our Landsat-derived observations of I_b change. Rates were calculated using the empirical formula detailed in *Park* et al. [2013], whereby theoretical $\delta T/\delta t$ is a function of GL retreat, surface topography/slope, subglacial bed topography/slope, seawater density (1027.5 kg m^{-3}) and the density of ice (917 kg m^{-3}). Our calculations utilized surface and bed slopes inferred from BEDMAP2 [*Fretwell* et al., 2013], and produced the theoretical estimations shown in Figure S4.

Overall, our estimations show excellent agreement with CryoSat-2 swath observations at Ferrigno East and Fox, indicating increasing trends of negative $\delta T/\delta t$ towards the GL at these locations. The high Landsat-derived $\delta T/\delta t$ value calculated for Ferrigno West (-9.8 m yr^{-1}) is believed to be subject to error, associated with the dramatic break up of ice and corresponding GL retreat across this sector between 2010 and 2015 (Figure 2; Figure S3). Nonetheless, the close agreement between Landsat and CryoSat-2 thinning values at Ferrigno East and Fox supplements other independent observations of dynamic thinning across the Eltanin Bay region in recent years [*Rignot* et al., 2008; *Bingham* et al., 2012; *McMillan* et al., 2014, *Wouters* et al., 2015], and acts as auxiliary validation of our Landsat I_b mapping.

Table S1. Landsat data used in the present study.

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Platform	*Year Mapped*	*Grounded Ice Boundary Classification*	*Spatial Resolution (m)*	*Prescribed Pixel Error (m)*	*Geometric Error[1] (m)*	*Positional Error (1σ; m)*
Landsat 2 (MSS[2])	1975	Open ocean	60	1 (60)	<100	117
		Grounded-ice/Sea-ice		2 (120)		156
		Slow-ice-to Shelf or Outlet Glacier		3 (180)		206
		Fast flowing, deeply bedded grounded ice		500		510
Landsat 4,5 (TM[3]), 7 (ETM+[4])	1985 1990 2000 2005 2010	Open ocean	30	1 (30)	<50	58
		Grounded-ice/Sea-ice		2 (60)		78
		Slow-ice-to Shelf or Outlet Glacier		3 (90)		103
		Fast flowing, deeply bedded grounded ice		500		502
Landsat 8 (OLI[5])	2015	Open ocean	30	1 (30)	12	32
		Grounded-ice/Sea-ice		2 (60)		61
		Slow-ice-to Shelf or Outlet Glacier		3 (90)		91
		Fast flowing, deeply bedded grounded ice		500		500

[1]Geometric error values derived from *Lee* et al. [2004]; *Tucker* et al. [2004], *Bindschadler* et al. [2011], *Storey* et al. [2014]. [2]*Multispectral Scanner*. [3]*Thematic Mapper*. [4]*Enhanced Thematic Mapper Plus*. [4]*Operational Land Imager*

Table S2. Landsat platform-specific positional accuracy (1σ) estimations for BSS grounded ice boundary classifications.

Table S3. ERS-1/2 Synthetic Aperture Radar imagery used to make double-differential interferograms.

(File uploaded separately as 2016GL068972-ts03.pdf)

Data Set S1. GL change values generated in this study.

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