Multi-decadal glacier surface lowering in the Antarctic Peninsula

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[1] From approximately 400 glaciers of the western Antarctic Peninsula, no in situ records of mass balance exist and their recent contribution to sea level is consequently poorly constrained. We seek to address this shortcoming by using surface elevations from USGS and BAS airborne (1948–2005) and ASTER spaceborne (2001–2010) stereo imagery, combined by using a rigorous semi-automated registration approach, to determine multi-decadal glacier surface elevation changes in the western Antarctic Peninsula for 12 glaciers. All observed glaciers show near-frontal surface lowering and an annual mean lowering rate of 0.28 ± 0.03 m/yr at the lower portion of the glaciers during the $\sim$4 decades following the mid-1960s, with higher rates for the glaciers in the north-west parts of the Antarctic Peninsula. Increased lowering of up to 0.6 m/yr can be observed since the 1990s, in close correspondence to increased atmospheric positive degree days. In all cases, surface lowering reduces to zero within 5 km of the glacier front at around 400 m altitude. This lowering may have been at least partially compensated for by increased high-altitude accumulation. **Citation:** Kunz, M., M. A. King, J. P. Mills, P. E. Miller, A. J. Fox, D. G. Vaughan, and S. H. Marsh (2012), Multi-decadal glacier surface lowering in the Antarctic Peninsula, *Geophys. Res. Lett.*, *39*, L19502, doi:10.1029/2012GL052823.

1. Introduction

[2] Atmospheric temperature in the Antarctic Peninsula (AP) has increased at a rate of more than +3.5°C/century [*Vaughan et al.*, 2003] during the second half of the 20th century. In parallel, widespread glacier acceleration and frontal retreat has been observed and ice shelves in the region have lost substantial parts over the last decades [*Pritchard and Vaughan*, 2007; *Scambos et al.*, 2000]. Recent studies have focused on the understanding of the dynamically-driven changes of many of the east coast glaciers since the 1990s [e.g., *Davies et al.*, 2011; *Rott et al.*, 2011]. The west coast glaciers, and the other east coast glaciers, are less well monitored and no records of multi-decadal mass balance exist for them.

[3] As a result, the AP component of global projections of sea-level rise has a very weak observational basis [e.g., *Lemke et al.*, 2007]. Recent glacier mass balance compilations have resorted to inferring AP mass balance from global averages [*Leclercq et al.*, 2011] or the Canadian High Arctic [*Dyurgerov and Meier*, 2005]; this is unsatisfactory and almost certainly produces biased estimates [*Kaser et al.*, 2006].

[4] Here, we address this lack of mass balance data for AP glaciers by presenting glacier surface elevation change measurements for 12 glaciers in the western AP. We extracted digital elevation models (DEMs) from USGS and BAS archive aerial stereo imagery from the 1960s to 2000s and rigorously combined these with recent elevations derived from ASTER satellite data.

2. Data and Methodology

[5] We identified 38 cloud-free USGS aerial photographs (scanned at 25 micrometer/1000 dots per inch) from the late 1960s covering nine marine-terminating glaciers along the western side of the AP. The selected frames represent the central vertical images of tri-camera photography and are suitable for DEM generation in the direction of flight. Systematic stereo-coverage of the AP is, however, prevented by the wide spacing between flight lines. We also identified BAS-archived historic and modern stereo images of three additional glaciers dating back to the 1940s. This gave a total of 12 primary glaciers plus several other smaller adjacent glaciers all located along the western Peninsula, between 64° and 71°S. To extract DEMs from the aerial stereo-photography we used SOCET SET 5.5.0 together with the Next Generation Automatic Terrain Extraction (NGATE) module. Overlapping ASTER DEMs were generated from Level 1b data using ENVI 4.6.1.

[6] Complete camera calibration data for the USGS single frames are not available. Missing calibration information, in particular unknown fiducial mark coordinates, prevents the use of such data for high quality DEM generation. This limitation was overcome by measuring the fiducial marks from multiple frames to generate new mean coordinates, and, in combination with the known focal length and lens distortion parameters, solving for the camera interior orientation (see auxiliary material for further information).[1] For all the BAS imagery full camera calibration data were available. For each image pair, approximately 50 tie points were used to solve for relative orientation. For the BAS imagery (at three locations) it was possible to determine the absolute orientation from measured ground control points. For the USGS imagery 'artificial' GCPs were extracted from distinctive surface features in the orthorectified ASTER images and their respective DEM.

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[1]Auxiliary materials are available in the HTML. doi:10.1029/2012GL052823.

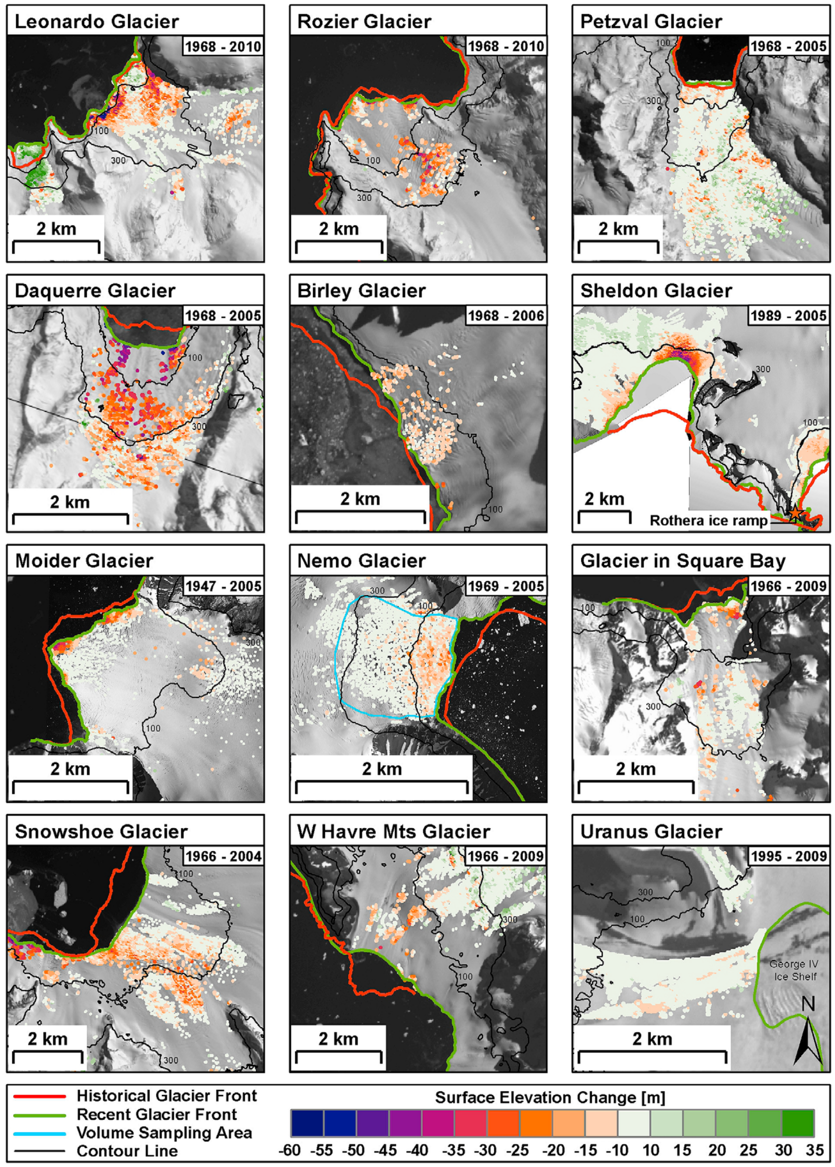

Figure 1. Overview of glacier surface elevation changes. 100 m and 300 m elevation contours lines are shown to delineate the extent of surface change. Zero-mean elevation changes over stable terrain are not shown for clarity.

[7] A first comparison of the fit between the historic and recent DEMs over stable terrain showed significant offsets. Mean DEM offsets ranged from −45 to 69 m in plan to −11 to 50 m in elevation. Stable terrain (i.e., rock outcrops) was mapped from the ASTER imagery. To remove the offsets we applied a semi-automatic robust least-squares surface matching technique [*Miller et al.*, 2009]. The underlying algorithm minimizes the Euclidean distances between the surfaces to be matched by estimating the seven parameters of a conformal transformation (three translations, three rotations and one global scaling). Matching was iteratively performed for points on rock outcrops with outliers iteratively downweighted. The final parameter solution was then applied to all points of the historical DEM to provide the best fit with the most recent ASTER DEM, taken as a reference surface. We investigated possible elevation-dependent biases by estimating an additional parameter for scale distortion in the vertical component. We found no evidence of ASTER jitter reported elsewhere [*Nuth and Kääb*, 2010].

[8] The post-matching RMS of DEM differences over stable (generally rough) terrain was ∼15 m, but this overstates the error over the generally flatter glacier surfaces. The mean difference over stable terrain was ±2 m on average, equivalent to an uncertainty in rate of 0.05 m/yr over a typical 40-year span between DEMs. We assigned uncertainties to the ice elevation differences by assessing the semivariograms of elevation differences over stable terrain [*Rolstad et al.*, 2009], which exhibit autocorrelation distances of 200 m. Uncertainties greater than ±2 m were found at sites where the stable terrain is almost entirely represented by steep topography and where shadowing or low image contrast resulted in errors in the ASTER DEM. These areas were manually masked. Points with low correlation values from the DEM extraction in SOCET SET were excluded from the historical DEM. After DEM co-registration, surface elevation change was computed by subtracting the earlier DEM of the later with computed summation over the DEM extent. Changes in glacier length (Figure 1) were surveyed by

Table 1. Datasets and Glacier Change Assessment

Glacier	Lat [°S]	Lon [°W]	Year Historical	Year Recent	Span [y]	LC[a] [m][m/yr]	FEC[a] [m/yr w.e.]	Elev dh_0[b] [m]	Dist dh_0[b] [m]	Total Area [km^2]	Coverage [%]
Leonardo	64.70	61.93	1968	2010	42	49 (1.2)	−0.29 ± 0.06	550	4300	51.3	13
Rozier	64.75	62.17	1968	2010	42	−81 (−2.0)	−0.40 ± 0.13	n/a	n/a	21.1	21
Petzval	64.94	62.91	1968	2005	37	42 (1.1)	−0.10 ± 0.14	600	4000	46.3	20
Daquerre	65.10	63.56	1968	2005	37	−311 (−8.4)	−0.42 ± 0.15	600	2700	17.2	25
Birley	65.93	64.45	1968	2006	38	−130 (−3.4)	−0.32 ± 0.10	n/a	n/a	17.3	33
Sheldon[c]	67.51	68.34	1989	2005	16	−967 (−60.4)	−0.57 ± 0.09	150	1200	191.8	19
Moider[c]	67.70	67.63	1947	2005	58	−299 (−5.2)	−0.12 ± 0.03	200	4000	54.3	17
Nemo[c]	67.71	67.33	1969	2005	36	−165 (−4.6)	−0.24 ± 0.03	300	1700	45.5	7
Square Bay	67.93	66.88	1966	2009	43	−207 (−4.8)	−0.29 ± 0.06	500	4100	22.3	23
Snowshoe	68.31	66.71	1966	2004	38	−682 (−17.9)	−0.28 ± 0.06	400	4800	233.9	24
W Havre Mts	69.24	72.05	1966	2009	43	−662 (−15.4)	−0.22 ± 0.09	450	2300	72.2	20
Uranus[c]	71.37	68.30	1995	2009	14	n/a	−0.04 ± 0.13	n/a	n/a	n/a	n/a
Mean			1970	2007	37	−310 (−11.3)	−0.28 ± 0.03	417	3233	70.3	20

[a]LC = Frontal length change; FEC = Frontal elevation change rate.
[b]Elev = Elevation of zero lowering; Dist = Distance to zero lowering from glacier front; n/a = Measurement not applicable due to limited data coverage or indefinite signal of lowering. Measurements were made along longitudinal profiles for glaciers where a distinct decay in the lowering signal was observable. Note: Full glacier extent is not covered by historical imagery.
[c]BAS datasets for historical data.

averaging at least five measurements perpendicular to each glacier front. We did not quantify changes to nunatak extents.

3. Results

[9] For all main glaciers, net surface lowering is observed near the glacier fronts (Figure 1 and Table 1), with an average total lowering of 20–30 m and a maximum of up to ~50 m over the multi-decadal periods. The spatial coverage (~20%) of each glacier is incomplete due to a combination of limited coverage of the historical data and a lack of surface features at increasing distance from the glacier front. Most of the observed glaciers are heavily crevassed at the front, so that it is much easier to extract measurable points from the frontal areas. It is rare for a glacier to have a uniform density of such points over its entire surface, and glacier-wide mass balance assessment remains challenging. For glaciers where we have sufficient spatial coverage, the lowering decays to zero within a few km or less of the ice front (e.g., Nemo and Sheldon glaciers) and is limited to regions lower than ~400 m elevation (Table 1 and Figure S1). Some glaciers show elevation increases of up to ~20 m at higher elevations (e.g., Sheldon and Petzval glaciers). Glacier fronts are generally stable (e.g., Petzval and Leonardo glaciers) or display retreat of up to 1 km (Table 1). There is some localised variation in these patterns, as is evident on the small glacier adjacent to Leonardo Glacier where glacier advance and thickening of up to 30 m is evident (Figure 1), but for larger glaciers the general pattern of surface lowering and frontal retreat appears robust. Seasonal variations of up to 2 m in snow cover [*van Lipzig et al.*, 2004] between images are within the level of accuracy that can be achieved from the ASTER DEMs.

[10] We computed the mean lowering rate for the glacier fronts using points sampled within a representative 1 km^2 area of each glacier within the 1 km of the front. We express the rates in water-equivalent units, assuming that most of the lowering occurs at the density of ice (917 kg/m^3). The sampling is limited by the lack of surface detail and incomplete coverage of the whole glacier surface. For the Rothera ice ramp we measure a surface lowering of 0.28 ± 0.09 m/yr (1989–2005), which is in relatively close agreement with in situ measurements of 0.32 m/yr (1989–1997) [*Smith et al.*, 1998]. This comparison provides both a partial validation of our measurement accuracy and suggests that the lowering of the ice ramp has continued at a rate of 0.24 m/yr since 1997. The average frontal lowering rate that we measured for all observed glaciers was 0.28 ± 0.03 m/yr over an average period of 37 years (1970–2007). Rates of lowering are greater in the northern Peninsula than in the south (Figure 2). Sheldon Glacier appears to be an exception to this pattern, but the time period of data for this glacier covers only 1989 to 2005. One possible interpretation is that it stands apart from nearby glaciers because lowering rates have increased recently.

[11] Moider and Nemo Glaciers, located near to Sheldon Glacier (Figure 2), provide an opportunity to examine any temporal variability more closely, since elevation data are available on more than two epochs. For an area with sufficient point density along the front of Nemo Glacier we calculated volumetric change on a decadal timescale (Figure 3a) relative to 2005. Also plotted are data for Moider Glacier as given in *Fox and Cziferszky* [2008]. The multi-decadal trend is negative and shows dramatically increased rates of lowering since 1989. The mean elevation change rate of Nemo Glacier has changed substantially in the last two decades from being almost zero at 0.03 m/yr (1969–1989) to −0.44 m/yr (1989–2005). The latter agrees well with the rate for Sheldon Glacier for a similar time period. Together, these measurements suggest a strong increase in glacier lowering since the early 1990s.

4. Discussion

[12] All of the glaciers we consider are marine terminating and hence are potentially subject to both atmospheric and oceanic warming. We calculated the number of atmospheric positive degree days (PDD) [*Braithwaite*, 1984] at the nearby stations Rothera and Faraday (Figure 2) since 1955 (Figure 3b). Temperature data were averaged by decade around the mid-decade points. Almost 60% of the PDDs since 1955 have occurred since 1985, with a particularly notable increase in the decade in 1985–1995, at about the time when we observe the onset of more rapid lowering at Moider and Nemo Glaciers. Increased lowering is centred on

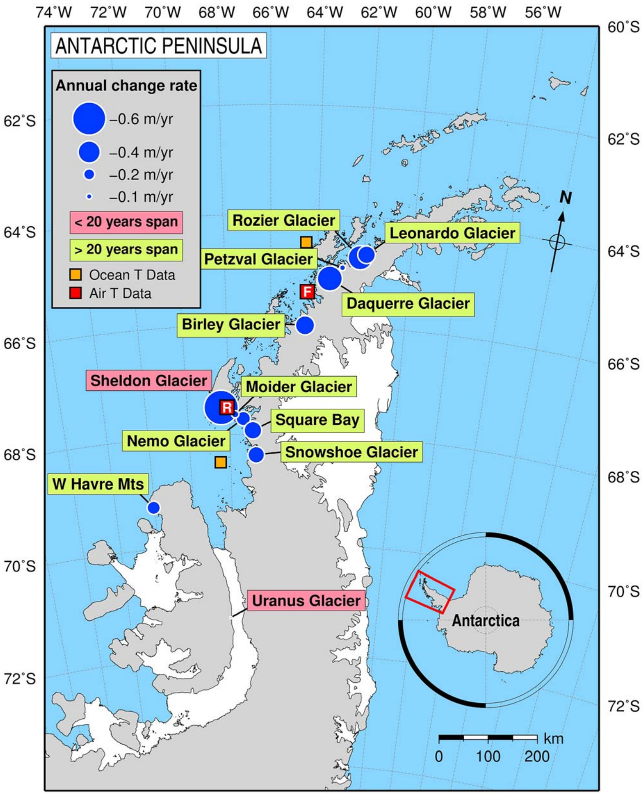

Figure 2. Mean annual surface lowering rates. Colored squares represent locations of ocean and surface air temperature measurements, F = Faraday, R = Rothera. "Span" is the difference between acquisition date of the oldest and most recent dataset for each site.

the period following 1989, whereas PDD sums show a dramatic increase in the decade centred on 1990, suggesting either the 1985–1989 PDD sums are much lower than the decadal average or the lowering is delayed, perhaps due to initial refreezing of surface melt. Total PDDs are greatest in the north (Figure 3) [cf. *Pritchard and Vaughan*, 2007] in agreement with greater surface lowering rates there, and the strong trend in atmospheric PDD further implicates surface melting as a dominant source for the lowering.

[13] Assuming a temperature lapse rate of −0.0082°C/m for the western Peninsula [*Morris and Vaughan*, 2003], total PDD in this region reduces by ~60% at 100 m elevation and ~98% at 400 m elevation. This is in agreement with the reduction in surface lowering to zero at around 400 m (Figure S1). However, some glaciers exhibit quite complex spatial patterns of lowering (e.g., Leonardo and Snowshoe glaciers) as would be expected given that there is not a linear relationship between PDD and surface melt due to, for example, glacier aspect, shadow and cloudiness.

[14] Also plotted on Figure 3b are upper 100 m oceanic temperature anomalies [*Meredith and King*, 2005], represented as the average column temperature between 0 and 100 m depth, for the two grid cells located closest to our glaciers (see locations in Figure 2). The ocean temperatures show a distinct warming although, unfortunately, the data do not span the full period of our elevation data. Nevertheless they do not clearly show a signal which could be related to faster rates of lowering at our more northerly glaciers. Further data are required to allow a complete partitioning of the respective roles of atmosphere and ocean in the observed glacier lowering.

[15] *Pritchard and Vaughan* [2007] observed that western Antarctic Peninsula glaciers accelerated by 12% on average from 1992 to 2005. They found no direct relationship to PDDs and suggested that the thinning of glacier fronts was bringing them nearer to floating, reducing their effective basal pressure resulting in faster sliding. Our observations confirm that there is widespread glacier front thinning in the western Antarctic Peninsula. Based on a thinning-retreat relation *Pritchard and Vaughan* [2007] calculated that a thinning rate of 5 m/yr could cause the observed acceleration. However, our observed surface lowering is more than an order of magnitude smaller and increased lowering since the 1990s cannot explain the difference. This suggests either that there is another mechanism controlling acceleration, or that a much lower rate of thinning could explain their observed accelerations.

[16] Assuming that glacier lowering is due to a contribution of atmospheric and oceanic melting, we infer that the majority of glaciers on the western AP, approximately 400, are likely to have been thinning at their fronts over recent decades. However, the spatial and temporal variation in the lowering rates suggests that the pattern of surface change is not a simple one and that a regional upscaling is not straight forward. Accumulation has generally increased in the AP over the period of the observations [see *Pritchard and Vaughan*, 2007] and hence reduced mass input cannot be the reason for the observed lowering. Indeed, we observe elevation increase at some higher altitude locations within a few km of the glacier fronts, raising the possibility that the observed surface lowering may be at least partially compensated by higher elevation accumulation increases of the sort inferred by *Nield et al.* [2012] as occurring across the entire Peninsula over this period. Given that glacier thinning is likely linked to acceleration and possible calving, its impact on sea level may be greater than the lowering rates by

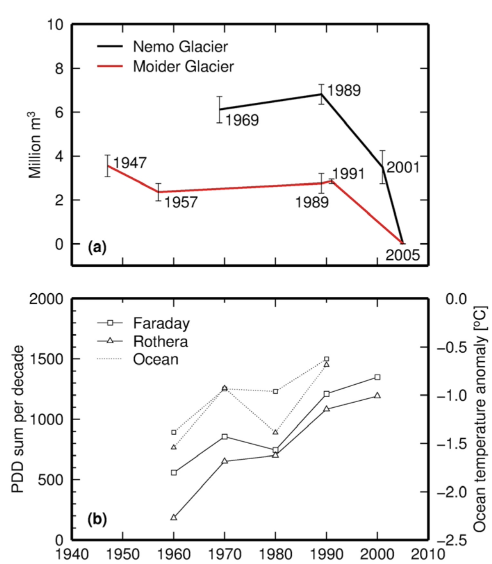

Figure 3. (a) Multi-decadal surface volume change of frontal areas relative to 2005. (b) Positive degree days (PDD) per decade and ocean temperature anomaly.

themselves would suggest. The partial coverage of the historic imagery restricts a glacier-wide mass balance assessment and the net contribution of the western Antarctic Peninsula to recent sea level change will, therefore, remain ambiguous until more spatially and temporally comprehensive measurements are made.

5. Conclusions

[17] Using archival aerial stereo-photography combined with ASTER DEMs we have demonstrated widespread glacier surface lowering in the western Antarctic Peninsula, with lowering confined generally to within 1 km of the front, and below 400 m elevation. Glacier frontal lowering exhibits a latitudinal pattern in accordance with higher surface temperature increases in the north. Two glaciers which have multi-epoch coverage show significantly larger-than-average lowering since about 1990, in close correspondence with an increase in the number of positive degree days in the decade centred on 1990. Low-altitude surface lowering of the glaciers in this study maybe at least partially balanced by increased accumulation at higher elevations. If this is the case, the contribution to sea level change from these glaciers is smaller than the frontal lowering alone would suggest. Ice-dynamical feedbacks especially highlight the need for more comprehensive AP glacier thinning information. Our 12 glaciers represent just 3% of the Peninsula glaciers, and only two of them have more than two epochs of data. Further spatial and temporal coverage is required, and indeed such additional archives of the kind that we exploit here do exist.

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