Ice loss from the Southern Patagonian Ice Field, South America, between 2000 and 2012

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[1] A time-series composed of 156 ASTER derived Digital Elevation Models (DEMs) and a radar-penetration-bias corrected version of the Shuttle Radar Topography Mission (SRTM) DEM is used to derive ice surface height and volume changes at the Southern Patagonian Ice Field (SPI) in southern South America. The observations, made between February 2000 and March 2012, indicate that the ice field is rapidly losing volume at many of the largest outlet glaciers, and in most cases thinning extends to the highest elevations of the ice field. Mass loss is occurring at a rate of $-20.0 \pm 1.2 \ Gt \ a^{-1}$, which, when summed with mass-loss at the adjacent Northern Patagonian Ice Field results in a combined rate of $-24.4 \pm 1.4 \ Gt \ a^{-1}$, equivalent to $+0.067 \pm 0.004 \ mm \ a^{-1}$ of sea level rise. Our decade-long mass loss rates are substantially higher than those derived during the last three decades of the 20th century, but are in good agreement with recent GRACE observations. Our volume loss estimate is sensitive to constraints applied to the amount of thickening in the accumulation zone. New field measurements and a continued DEM time-series will be required to refine our estimates. Citation: Willis, M. J., A. K. Melkonian, M. E. Pritchard, and A. Rivera (2012), Ice loss from the Southern Patagonian Ice Field, South America, between 2000 and 2012, Geophys. Res. Lett., 39, L17501, doi:10.1029/2012GL053136.

1. Introduction

[2] Ice mass loss at the Southern Patagonian Ice Field (SPI) provides a disproportionately large contribution to recent sea level change with respect to the size of the ice field [Rignot et al., 2003]. Most of the glaciers of the ice field are decaying [Aniya, 1999; Casassa et al., 2002; Heid and Kääb, 2012; Lopez et al., 2010] with several undergoing “catastrophic” retreat [Rignot et al., 2003]. The ice field lost volume at a rate of $-13.5 \pm 0.8 \ km^3 \ a^{-1}$ between 1975 and 2000 [Rignot et al., 2003], about eight times faster than the estimated loss rate of $-1.7 \pm 0.4 \ km^3 \ a^{-1}$ that occurred between the Little Ice Age in 1750 and 2010 [Glasser et al., 2011]. Gravity Recovery and Climate Experiment (GRACE) observations suggest a combined mass loss from the Northern Patagonian Ice Field (NPI) and the SPI of $-23.0 \pm 9.0 \ Gt \ a^{-1}$ for the period between 2003 and 2011 [Jacob et al., 2012], $-26.6 \pm 6.0 \ Gt \ a^{-1}$ between 2003 and 2009.25 [Ivins et al., 2011], and $-25.1 \pm 9.9 \ Gt \ a^{-1}$ between 2002.25 and 2006 [Chen et al., 2007], assuming a density of ice of 900 kg m$^{-3}$. These observations indicate the ice fields are undergoing rapid drawdown.

[3] This study provides updated rates of surface height changes through time (dh/dt) over the entire SPI, resulting in a high-resolution map that is used to identify glaciers that are undergoing rapid thinning and to provide insight into the spatial variability in dh/dt rates across the ice field. The dh/dt data are integrated to provide estimates of volume (dV/dt) and mass change rates. We use all suitable Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) scenes collected between February 2000, when the Shuttle Radar Topography Mission (SRTM) Digital Elevation Model (DEM) was acquired, and March 2012.

[4] The SPI is the largest contiguous temperate body of ice in the southern hemisphere [Naruse and Aniya, 1992]. It stretches from about 48.5°S to 51.5°S (Figure 1) and has an average altitude of 1355 m above sea level. The ice field is fed continuously by precipitation-laden westerly winds that flow from over the Pacific Ocean and collide with the north–south trending mountains [Casassa et al., 2002]. Many of the glaciers on the western, Chilean side terminate at sea level as tidewater calving glaciers, while the majority of the glaciers on the eastern side, in both Chile and Argentina, terminate in lakes [Casassa et al., 2002; Warren and Aniya, 1999].

[5] Observed regional near-surface air temperatures increased at a rate of between 0.4°C to 2.0°C per century between the 1930s and 1990s [Carrasco et al., 2002, 2008; Rosenblüth et al., 1995; Rosenblüth et al., 1997; Villalba et al., 2003]. The NCEP/NCAR model reanalysis indicates similar rates of warming at the 850 HPa altitude [Rasmussen et al., 2007], which at about 1350 m above sea level is near the average height of the ice field [Carrasco et al., 2002]. It is likely that this warming is partly responsible for the ongoing demise of the ice field [Rasmussen et al., 2007] with additional influences due to recent changes in precipitation type [Carrasco et al., 2002; Rosenblüth et al., 1995], ocean temperatures [Gille, 2008] and ice dynamics [Rivera et al., 2012a, 2012b; Sugiyama et al., 2011].

2. Methods

[6] We provide a summary of our methods, which are a modification of Willis et al. [2012].

2.1. DEM Co-registration

[7] 156 “Product 14” DEMs and orthorectified ASTER images [Fujisada et al., 2005] are retrieved from the NASA/USGS Land Processes Distributed Active Archive Center (LPDAAC) facility. The Automatic Registration and Orthorectification Package (AROP) [Gao et al., 2009] is used to
horizontally co-register each 15-m resolution ASTER image (band 3n) to pan-sharpened 14.25 m resolution Global Land Survey Landsat-7 imagery of the SPI that has been previously orthorectified to the 90 m resolution C-band SRTM DEM by the Global Land Cover Facility [Tucker et al., 2004]. When the ASTER imagery is co-registered to the Landsat imagery, the accompanying ASTER DEM is simultaneously co-registered to the SRTM, see Willis et al. [2012] and Melkonian [2011] for full details. Each ASTER DEM is resampled using bicubic convolution and re-projected to have the same 90 meter posting and UTM projection as the SRTM. The maximum root-mean-squared-error allowed by AROP during the co-registration process is set to 0.75 pixels. This precise horizontal co-registration reduces errors associated with slope offsets [e.g. Nuth and Kääb, 2011].

2.2. SRTM DEM Radar Penetration

The uncertainty on the SRTM DEM is set to ±5 m based upon field studies on glaciers at the southern end of the SPI [Rignot et al., 2003]; previous verification studies [Rodriguez et al., 2006] and comparison to ICESat laser altimetry tracks, both published [Carabajal and Harding, 2005] and examined during our work.

In order to estimate the penetration of 5.6 cm wavelength C-band radar into ice and snow [e.g. Rignot et al., 2001] we compare the elevations of the C-band DEM to the simultaneously acquired 3.1 cm wavelength X-band SRTM DEM data that covers 80% of the ice field [Rabus et al., 2003]. We put the X- and C-band DEM’s into the same vertical datum then resample and co-register the C-band DEM to the X-band DEM using the techniques of Nuth and Kääb [2011]. We difference the DEMs and remove a curvature dependent bias [Gardelle et al., 2012] (Figure S2 of the auxiliary material) and find that ice surfaces recorded by the X-band are higher than the C-band data by about 2 m at all altitudes (Figure S2 of the auxiliary material). X-band penetration relative to the ASTER elevations is not expected to be significant on ice [e.g. Gardelle et al., 2012], especially since the ice field surface was wet during the late southern summer [e.g. Monahan and Ramage, 2010], but further work will be necessary to quantify X-band penetration into the firm in the accumulation zone. For this study we assume the 2 m low-bias of the C-band is due to radar penetration and correct the C-band DEM for this bias.

2.3. Constraints on Elevation Change Rates

Field studies within the accumulation zone of the Chico Glacier [Rivera et al., 2005] indicate dh/dt of −1.9 ± 0.14 m a⁻¹ there. Preliminary examination of additional airborne laser altimetry suggests no thickening at the highest elevations of the Jorge Montt accumulation zone. It is unclear whether these sparse observations are representative of behavior elsewhere in the accumulation zone of the SPI. For the purpose of this study we constrain dh/dt rates in the accumulation zone to be between +3.5 m a⁻¹ and −10 m a⁻¹. The upper bound on these constraint is about half the estimated accumulation rate [e.g. Casassa et al., 2002] and provides accommodation for firn densification and ice advection. Our time-series results are sensitive to this upper bound and we explore the effect of varying this parameter in the discussion.

In the ablation zone, above 600 m a.s.l., dh/dt is constrained to +3.5 m a⁻¹ and −60 m a⁻¹ to sufficiently record coherent areas of thinning that occur at HPS-12 glacier and thickening on the Pio-Xi trunk. Below 600 m a.s.l.

[8] Each ASTER DEM is vertically co-registered to the SRTM DEM using an iterative routine that minimizes the average residual difference in bedrock altitudes when the SRTM is subtracted from the ASTER DEM. The standard deviation on the spread of residual bedrock altitudes, which is typically 15 to 20 m, is taken as a measure of the elevation uncertainty on the individual ASTER DEM. These procedures result in coregistered stacks of elevations that are used to generate time series of elevation change for each pixel (See Figure S1 of the auxiliary material for an example).

[11] Field studies within the accumulation zone of the Chico Glacier [Rivera et al., 2005] indicate dh/dt of −1.9 ± 0.14 m a⁻¹ there. Preliminary examination of additional airborne laser altimetry suggests no thickening at the highest elevations of the Jorge Montt accumulation zone. It is unclear whether these sparse observations are representative of behavior elsewhere in the accumulation zone of the SPI. For the purpose of this study we constrain dh/dt rates in the accumulation zone to be between +3.5 m a⁻¹ and −10 m a⁻¹. The upper bound on these constraint is about half the estimated accumulation rate [e.g. Casassa et al., 2002] and provides accommodation for firn densification and ice advection. Our time-series results are sensitive to this upper bound and we explore the effect of varying this parameter in the discussion.

[12] In the ablation zone, above 600 m a.s.l., dh/dt is constrained to +3.5 m a⁻¹ and −60 m a⁻¹ to sufficiently record coherent areas of thinning that occur at HPS-12 glacier and thickening on the Pio-Xi trunk. Below 600 m a.s.l.

Auxiliary materials are available in the HTML. doi:10.1029/2012GL053136.
rates are constrained to +10 m a\(^{-1}\) and −60 m a\(^{-1}\) to capture rapid thinning at many glaciers and allow the volume increase at the terminus of the advancing Pio-Xi Glacier [Rivera and Casassa, 1999] to be recorded.

[13] A weighted linear-regression is applied to the time-series of elevations on a pixel-by-pixel basis. Each ASTER elevation point is inversely weighted according to the uncertainty on the DEM from which it is extracted. The gradient of the regression at each pixel is the dh/dt for that pixel, with the one-sigma uncertainty (σ) on the rate derived from the model covariance matrix (see Figure S3 of the auxiliary material for maps of uncertainties, date coverage and the number of images used). dh/dt is obtained by integrating the dh/dt rates over the ice field.

[14] Volume uncertainties are computed at a 95% confidence interval using spatial averaging, 1.96σ/A \sqrt{n}, where A is the area of the region concerned, and n represents the number of uncorrelated pixels in that area [Howat et al., 2008]. An estimate of the correlation across our observations is found using the variance of off-ice dh/dt rates examined using methods developed by Rolstad et al. [2009]. Our measurements decorrelate at length scales greater than 1.8 km (areas of 1.8 km by 1.8 km).

2.4. Glacier Boundaries and Mass Change Calculations

[15] Glacier outlines and internal bedrock polygons are traced from orthorectified ASTER and Landsat imagery, guided by Aniya et al. [1996]. The glaciers in this study cover 12,118 km\(^2\) excluding an area of 432 km\(^2\) of exposed bedrock within the perimeter of the ice field. This glacierized area is about 93% of the size of the area used by Rignot et al. [2003]. It should be noted that our basin sizes and definitions are somewhat different from those used by Rignot et al. [2003] in the northwest of the ice field.

[16] Equilibrium Line Altitudes (ELAs) are needed to guide the application of constraints on our dh/dt regressions. ELAs are from Aniya et al. [1996] and Rivera and Casassa [1999] where available. For the remaining glaciers we digitize the end of summer (February to early-April) snow-to-ice transition on the glacier from 2005–2011 ASTER orthoimagery. We provide the average altitude of this zone from the SRTM DEM as a proxy for the ELA [e.g., Bamber and Rivera, 2007]. Our “picked” snowlines are typically within 100 m of the altitude of previously measured ELAs.

[17] For grounded tidewater- or lacustrine-calving glaciers we estimate the minimum thickness of ice at the front using buoyancy calculations along with sparse bathymetry measurements [e.g., Naruse and Skvarca, 2000; Rivera et al., 2012b]. Area loss rates are measured from ASTER and LANDSAT orthoimagery and coupled to thickness estimates to provide sub-aqueous volume loss rates (see auxiliary material).

[18] Converting the dh/dt into mass changes is uncertain because the density of the material lost or gained is poorly known. We assume a density of 900 kg/m\(^3\) for all material involved [e.g., Berthier et al., 2010; Rignot et al., 2003].

3. Results

[19] Figure 1 shows the average 2000–2012 dh/dt for the SPI. Thinning is extensive at low elevations around most of the circumference of the ice field, extending into the high interior across most of the region. Much of the thinning occurs at large eastern glaciers that terminate in lakes. ASTER DEMs provide coverage of over 96% of the ice field between 2000 and 2012 with the majority of gaps occurring on featureless snow plains in the interior where the optical DEMs perform poorly. Scaling rates up to 100% of the ice-covered portion of the SPI provides an average dh/dt of −21.2 ± 0.5 km a\(^{-1}\). We add sub-aqueous ice-loss, estimated at −1.0 ± 0.8 km a\(^{-1}\) to provide a total dh/dt of −22.2 ± 1.3 km a\(^{-1}\) for the ice field. This equates to an area-averaged dh/dt of −1.8 ± 0.1 m a\(^{-1}\) of ice distributed across the entire SPI. The area averaged dh/dt of the 3593 km\(^2\) NPI, is slower, −1.3 ± 0.1 m a\(^{-1}\), providing a dh/dt there of −4.9 ± 0.3 km a\(^{-1}\) between 2000 and 2011 (Recalculated from Willis et al. [2012], using the same radar correction, spatial averaging and density as this work).

[20] The −22.2 ± 1.3 km a\(^{-1}\) 2000–2012 SPI volume loss rate is considerably faster than the −13.5 ± 0.8 km a\(^{-1}\) rate estimated by Rignot et al. [2003] between 1968/75 and 2000, but considerably slower that their 1995–2000 rate of −38.7 ± 4.4 km a\(^{-1}\) - we discuss possible explanations for this in section 4.

[21] The dh/dt for the ~8600 km\(^2\) accumulation zone of the SPI is −9.0 ± 0.4 km a\(^{-1}\), which equates to a dispersed dh/dt of −1.1 ± 0.1 m a\(^{-1}\) over the area. dh/dt for the ~3500 km\(^2\) ablation zone (not including sub-aqueous volume loss estimates) is −12.2 ± 0.2 km a\(^{-1}\), which equates to a dh/dt of −3.5 ± 0.02 m a\(^{-1}\). Results from all named glacier basins are provided in Table S1 of the auxiliary material.

[22] The SPI dh/dt rate (−22.2 ± 1.3 km a\(^{-1}\)) equates to a mass change rate of −20.0 ± 1.2 Gt a\(^{-1}\) which, when summed with the losses at NPI provides a total rate of −24.8 ± 1.4 Gt a\(^{-1}\) for the combined ice fields. This contribution of +0.067 ± 0.004 mm a\(^{-1}\) to sea level rise is about 50% more than the 1968/75 to 2000 rate of contribution from the NPI and SPI estimated by Rignot et al. [2003].

3.1. Individual Basins

[23] The 2000–2012 dh/dt of −3.1 ± 0.2 km a\(^{-1}\) at the 834 km\(^2\) Upsala Glacier (Figure 2) is a five-fold increase in ice-loss compared to the 1975–2000 rate of −0.6 km a\(^{-1}\) estimated by Rignot et al. [2003]. Thinning at the glacier front between 2000 and 2005 (−10.0 ± 2.0 m a\(^{-1}\), accelerated to −24.8 ± 2.4 m a\(^{-1}\) between 2005 and 2011. The terminus of the glacier retreated ~3.5 km between February 2000 and August of 2011, while the elevation profile of the glacier evolved from being broadly convex to broadly concave with a flat low-profile terminus area. Rapid retreat of the front is ongoing as of March, 2012. Upsala Glacier accounts for about 15% of the total mass loss from the ice field.

[24] Rignot et al. [2003] observed dh/dt of −1.7 km a\(^{-1}\) and a peak dh/dt of −17.9 m a\(^{-1}\) between 1975 and 2000 for the 389 km\(^2\) Jorge Montt Glacier (Figure S1 of the auxiliary material). The glacier front retreated more than 8 km over the same period [Rivera et al., 2012b]. Our dh/dt rate of −1.8 ± 0.1 m a\(^{-1}\) between 2000 and 2012 is approximately the same but we capture a faster coherent area of rapid thinning (−21.5 ± 0.8 m a\(^{-1}\)) near the present day front. The glacier terminus retreated on average about 200 m a\(^{-1}\) (2.2 km) between 2001 and 2011. The volume loss at the Jorge Montt may be in part due to melting, as passive microwave observations indicate prolonged and continuous surface “wetness” over Jorge Montt over almost the entire 2002–2008 period [e.g. Monahan and Ramage, 2010]. It is,
however, more likely that changing ice dynamics linked to the large calving flux of $2.4 \text{ km}^3 \text{ a}^{-1}$, rapid retreat in to a deep channel and longitudinal stretching at the front of the glacier plays a larger role [Rivera et al., 2012a].

The small 165 km$^2$ HPS-12 glacier (Figure S4 of the auxiliary material) lost volume at a rate of $-0.8 \pm 0.1 \text{ km}^3 \text{ a}^{-1}$ and retreated 4.3 km up-valley between 2000 and 2012. A maximum $\text{dh/dt}$ of about $-28 \text{ m a}^{-1}$ was recorded between 1995 and 2000 [Rignot et al., 2003]. Maximum rates of thinning slowed to $-21.0 \pm 3 \text{ m a}^{-1}$ between 2000 and 2005 before accelerating significantly to $-57.0 \pm 13 \text{ m a}^{-1}$ between 2005 and 2011. Thinning between 2000 and 2011 totaled $-450 \text{ m}$, near the front of the glacier.

The largest glacier of the ice field, the 1269 km$^2$ Pio-Xi Glacier continued its slow advance through old growth forest and shallow coastal waters [Rivera and Casassa, 1999; Warren and Sugden, 1993]. The average $\text{dh/dt}$ at the front region, at altitudes below 300 m a.s.l., is $+2.4 \pm 0.4 \text{ m a}^{-1}$, which compares well to previous measurements [Rivera and Casassa, 1999]. Overall however, the glacier $\text{dV/dt}$ rate of $-1.0 \pm 0.1 \text{ km}^3 \text{ a}^{-1}$ indicates the advance of Pio-Xi is unlikely to be sustained.

4. Discussion and Conclusions

The combined 2000 to 2012 ice-mass loss rate of the SPI and NPI ($-24.4 \pm 1.4 \text{ Gt a}^{-1}$) is about the same as rates estimated during first decade of the 21st century using GRACE [Chen et al., 2007; Ivins et al., 2011; Jacob et al., 2012]. Our rate is much less than the 1995–2000 rate of $-37.7 \pm 4.0 \text{ Gt a}^{-1}$ estimated by Rignot et al. [2003] for the combined ice fields. To investigate whether the ice field mass loss is slowing down between 1995–2000 and 2000–2012 we examine how short time interval trends can differ from longer-term trends when data coverage is only available over a portion of the ice field for shorter time periods.

For example, while we think we can use our time-series approach to create reliable $\text{dh/dt}$ results for individual glaciers over less than decadal intervals, we do not have spatial coverage over short timescales to do this over the entire ice field, especially the accumulation zone. To illustrate this, we find the area and average $\text{dV/dt}$ for 10 m elevation bands for the entire SPI for the sub-periods 2000–2006 and 2006–2012 (Figure S5 of the auxiliary material). We then find the area within each band for which we have coverage and scale the results for each band at each time interval to the appropriate area to provide estimates for the whole SPI.

A second and dominant source of uncertainty is the maximum allowable thickening rate constraint that we apply to automatically filter out clouds and DEM errors from our
time-series. Varying this “upper bound” from +5 to +0 m a\(^{-1}\) changes the dV/dt rate for the accumulation zone from \(-6.4 \pm 0.4 \text{ km}^3 \text{a}^{-1}\) to \(-20.1 \pm 0.5 \text{ km}^3 \text{a}^{-1}\), potentially increasing the total SPI dV/dt by 50% to \(-33.3 \pm 1.4 \text{ km}^3 \text{a}^{-1}\). If the “no thickening” constraint is used, then the accumulation zone loses more volume each year than the ablation zone, which seems unlikely. Attempts to tune our constraints using point measurement of thinning made between 1998 and 2000 at the Chico Glacier accumulation zone (\(-1.9 \pm 0.1 \text{ m a}^{-1}\)) [Rignot et al., 2005] are difficult due to persistent clouds in that area. We find that 2000–2012 median dh/dt rates with uncertainties expressed as median absolute deviations on the non-Gaussian distribution, for a 1.8 km radius around the site vary from \(-0.6 \pm 1.6 \text{ m a}^{-1}\) to \(-1.9 \pm 1.0 \text{ m a}^{-1}\), with our chosen constraint (+3.5 m a\(^{-1}\)) providing a local dh/dt of \(-1.0 \pm 1.3 \text{ m a}^{-1}\). Additional field measurements of dh/dt in the accumulation zone on the western and central regions of the SPI could tune our constraints and indicate whether the short-term behavior at Chico Glacier is representative of the larger ice field. Our results in the ablation zone are robust as the imagery is clearer and the signal larger. We note we can experiment with different values of maximum allowed thickening because we are using a time series approach instead of only differentiating two DEMs.

[31] We suggest that the Rignot et al. [2003] estimate of volume loss of the SPI from 1995–2000 (\(-38.7 \pm 4.4 \text{ km}^3 \text{a}^{-1}\)) was too high, mainly due to extrapolating results from coverage of only 43% of the ice field to the entire ice field, the short timescale involved and the lack of a radar penetration correction. However, given the uncertainties we cannot rule out the dV/dt was higher between 1995–2000 than 2000–2012.

[32] We find that the combined NPI and SPI contributed to sea level between 2000 and 2012 at a substantially faster rate (\(-24.4 \pm 1.4 \text{ Gt a}^{-1}\)) than between 1968/75 and 2000 (15.0 \pm 0.7 \text{ Gt a}^{-1}\) [Rignot et al., 2003]. Thinning occurring at the freshwater calving glaciers on the eastern side of the ice field accounts for about 40% of the volume loss signal at the SPI. Atmospheric warming has increased the number of days per year with summer conditions that promote melting at the ice field surface [Monahan and Ramage, 2010]. It is plausible that this melt water is making its way to the base of the glaciers, changing the basal pressures and subsequently affecting local ice dynamics [e.g. Sugiymama et al., 2011]. It is likely that warming is also increasing the amount of rainfall at high elevations.

[33] In summary, strong temporal variability in volume change should be considered when interpreting volume and mass change rates that have been calculated using short timescales and/or with incomplete spatial coverage. Radar penetration, which provides about 10% of our signal over the 2000–2012 timescale, should be considered for studies using the SRTM C-Band radar over glaciers. Our technique of automatically stacking DEMs is sensitive to the applied constraints on accumulation zone thickening. Field measurements of average dh/dt rates in the accumulation zone of the SPI and an extension of the time-series would improve our results.

[34] Acknowledgments. ASTER data is from the LPDAAC USGS/Earth Resources Observation and Science (EROS) Center, Sioux Falls, South Dakota: http://lpdaac.usgs.gov/get_data. LANDSAT and SRTM data are from NASA Global Land Cover Facility (http://www.landcover.org).

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